THE STRUCTURAL PETROLOGY

of an area

EAST OF SPRINGTON, SOUTH AUSTRALIA

by

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All specimens referred to in this thesis are catalogued in the museum of the department of Geology of the University of Adelaide. The specimen numbers are prefixed by the numerals A 185- . The catalogued information includes the name of the rock, the catalogue number and the approximate location. The exact location of each specimen is given as a national grid reference as drawn on the Cambrai 1:63,360 Sheet (1940) (Fig. 2) and reproduced on Plate 1.

SUMMARY.

The structure and petrology of 50 square miles of folded and metamorphosed early Palaeozoic sediments on the eastern edge of the Mount Lofty Ranges, 40 miles east-north-east of Adelaide, South Australia, have been studied in detail. The purpose of this investigation has been to ascertain the character of the metamorphism and the time relation of metamorphism to deformation in the eastern portion of the Mount Lofty Range metamorphic belt. The rock units, showing a wide range of chemical variety, have been demonstrated to cross the metamorphic zones from the biotite zone to the sillimanite-potash felspar zone.

A variety of sedimentation structures have been recognised within the rock sequence.

Zonation of the metamorphism has been achieved through the establishment of metamorphic mineral facies boundaries. The metamorphism is of the low to intermediate pressure andalusite-staurolite-cordierite-sillimanite type. The appearance of kyanite along with andalusite and sillimanite in the aluminous pelitic schists on the western edge of the area, and the apparent repression of a number of metamorphic reactions releasing volatile components in the associated rocks, suggests that pressures were higher in the western part of the area during metamorphism. The progressive metamorphism of quartzo-

felspathic schists, pelitic schists, aluminous pelitic schists, cordierite and anthophyllite schists, marbles and calc-silicate rocks is described and various metamorphic mineral reactions are discussed. An optical study has been conducted on the plagioclase-epidote equilibrium relations in calc-silicate rocks.

Metamorphic segregation veins are common throughout the whole area.

A number of rocks of metasomatic origin have been described. Of these, fine-grained quartz-albite rocks replacing quartzo-felspathic schists, marbles and calc-silicate rocks are the most abundant. Narrow zones of quartz-potash felspar rocks on the western edge of the area are closely associated with potash deficient schists containing anthophyllite, chlorite and minor muscovite in place of the biotite and potash felspar of the normal quartzo-felspathic schists.

A variety of intrusive rocks in the form of small widely distributed bodies include dolerites, diorites, syenites and granodiorites. Swarms of aplite and pegmatite dykes appear to have been derived from the granodiorites.

Two phases of folding are recognised, the earlier having slaty cleavage as axial surface and the later crenulation cleavage as axial surface. Two apparently coeval crenulation cleavages of differing orientation are found in different parts of the area. Petrofabric studies were conducted on quartz, calcite, mica and cordierite.

A study of porphyroblast growth has indicated that metamorphism occurred throughout, and outlasted, the two phases of folding.

In the dying stages of metamorphism a compound fault system displaced the metamorphic mineral isograds. The recently active Milendella Fault is considered to have a complex history commencing in Palaeozoic times.

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

ACKNOWLEDGMENTS

This mapping project and the associated petrological and structural study was initially suggested to the author by Dr. A. W. Kleeman.

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A summary of the results of this work was presented as an address to the Geological Society of Australia (South Australian Branch) at the November meeting, 1964.

Dr. A. W. Kleeman read and criticized the whole manuscript. Dr. J. L. Talbot read and criticized the structure section.

Mr. J. Lorenzin prepared the thin sections used in the petrological study. Mr. J. Biddle assisted in the interpretation of the X-ray powder photographs. Miss A. M. C. Swan offered advice on the lay-out of some of the maps and line drawings. Miss B. G. Storer typed the figure descriptions. The thesis was typed at the Perfection Copying Office, Adelaide. Harding and Halden Pty. Ltd. photographically reduced and reproduced the line drawings.

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INTRODUCTION.

Purpose and Aims of the Present Study.

The Mount Lofty Ranges near Adelaide, South Australia, expose an upper Precambrian to lower Palaeozoic geosynclinal sequence forming portion of the Adelaide Geosyncline (Fig. 1). The eastern portion of the Mount Lofty Ranges has been metamorphosed, the highest grade parts reaching the upper amphibolite facies adjacent to nodes of granitic paragneiss. The purpose of the present study is to elucidate some of the metamorphic and structural history of the Mount Lofty Range metamorphic belt through the detailed investigation of a small representative area. A previously unmapped area of approximately 50 square miles on the eastern edge of the ranges, situated west of the town of Cambrai (Cambrai 1:63,360 Military Sheet) and 39 miles east-north-east of Adelaide, was chosen because of the good exposure of a variety of rock types displaying a considerable range of metamorphic grade. The aims of the investigation have been to construct, with the aid of aerial photographs and field mapping, a detailed geological map; to collect and compile data on sedimentation and tectonic structural features; and to conduct a contemporaneous laboratory study of the metamorphic and structural petrology.

Relation of the Cambrai Area to the Geological Pattern of the Mount Lofty Ranges.

The location of the area mapped in this study is indicated in Figs. 2 and 3. Fig. 2 also shows the major stratigraphic subdivisions of the Mount Lofty Ranges. Several older Precambrian basement inliers are exposed in the core of the

FIG. 1.

Geological map of South Australia.

(Courtesy of the Geological Survey of South Australia).

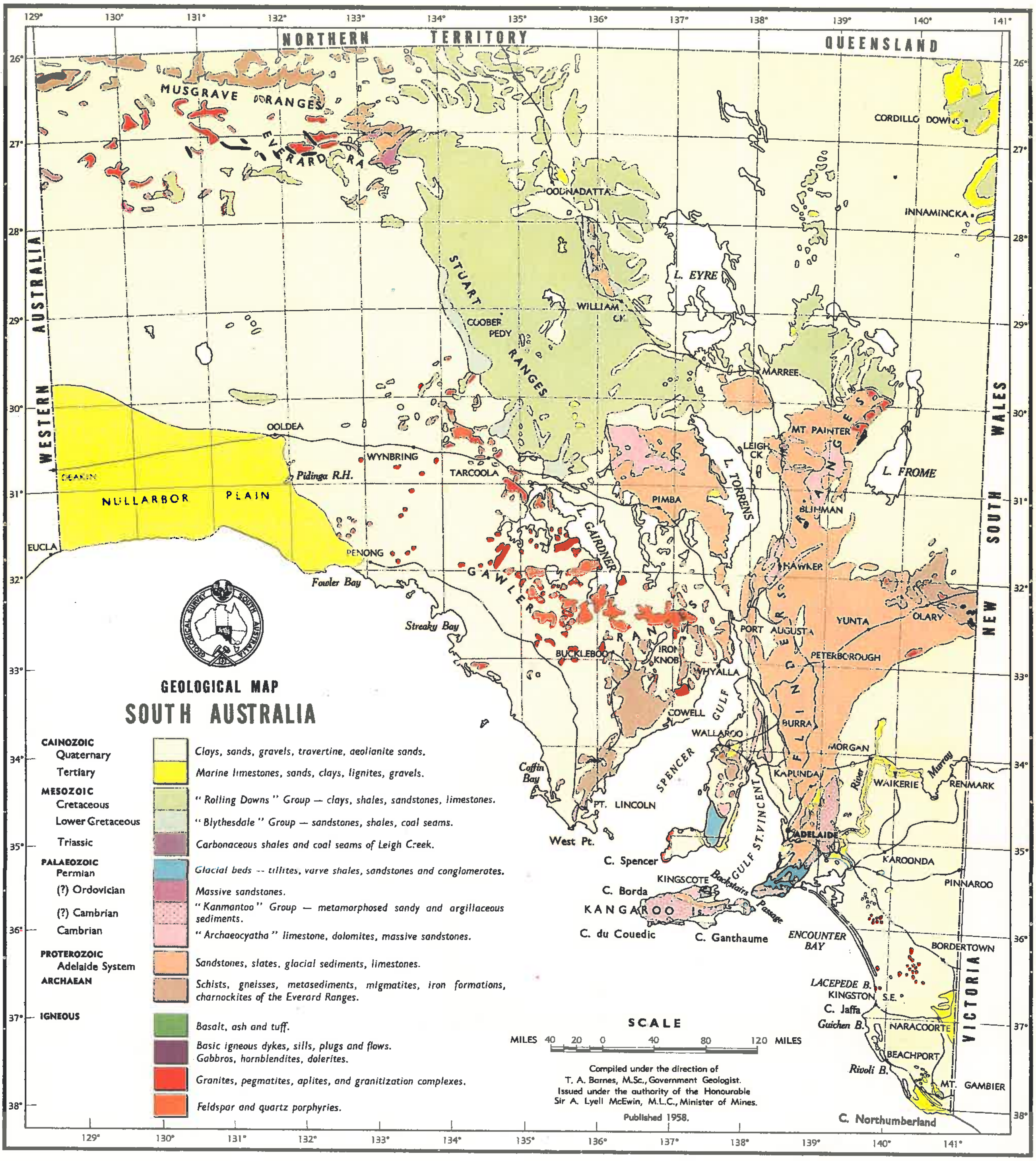

NORTHERN TERRITORY
QUEENSLAND
WESTERN AUSTRALIA
NEW SOUTH WALES
VICTORIA
MUSGRAVE RANGES
EVERARD RA
CORDILLO DOWNS
OODNADATTA
INNAMINCKA
STUART RANGES
L. EYRE
WILLIAM CK.
COOBER PEDY
MARREE
MT PAINTER
LEIGH CK
L. FROME
L. TORRENS
OOLDEA
WYNBRING
Pidinga R.H.
TARCOOLA
DEAKIN
NULLARBOR PLAIN
PIMBA
BLINMAN
EUCLA
PENONG
L. GAIRDNER
HAWKER
Fowler Bay
GAWLER RANGES
Streaky Bay
PORT AUGUSTA
YUNTA
OLARY
BUCKLEBOO
IRON KNOB
PETERBOROUGH
WHYALLA
FLINDERS RANGES
COWELL
BURRA
WALLAROO
SPENCER GULF
MORGAN
Coffin Bay
KAPUNDA
River Murray
WAIKERIE
RENMARK
PT. LINCOLN
West Pt.
ADELAIDE
GULF ST. VINCENT
C. Spencer
KARAONDA
PINNAROO
KINGSCOTE
C. Borda
Backstairs Passage
KANGAROO I.
C. du Couedic
C. Ganthaume
ENCOUNTER BAY
BORDERTOWN
LACEPEDE B.
KINGSTON S.E.
C. Jaffa
Guichen B.
NARACOORTE
BEACHPORT
Rivoli B.
MT. GAMBIER
C. Northumberland
GEOLOGICAL SURVEY SOUTH AUSTRALIA
GEOLOGICAL MAP
SOUTH AUSTRALIA
CAINOZOIC
Quaternary
Clays, sands, gravels, travertine, aeolianite sands.
Tertiary
Marine limestones, sands, clays, lignites, gravels.
MESOZOIC
Cretaceous
" Rolling Downs " Group — clays, shales, sandstones, limestones.
Lower Cretaceous
" Blythesdale " Group — sandstones, shales, coal seams.
Triassic
Carbonaceous shales and coal seams of Leigh Creek.
PALAEOZOIC
Permian
Glacial beds -- tillites, varve shales, sandstones and conglomerates.
(?) Ordovician
Massive sandstones.
(?) Cambrian
" Kanmantoo " Group — metamorphosed sandy and argillaceous sediments.
Cambrian
" Archaeocyatha " limestone, dolomites, massive sandstones.
PROTEROZOIC
Adelaide System
Sandstones, slates, glacial sediments, limestones.
ARCHAEAN
Schists, gneisses, metasediments, migmatites, iron formations, charnockites of the Everard Ranges.
IGNEOUS
Basalt, ash and tuff.
Basic igneous dykes, sills, plugs and flows. Gabbros, hornblendites, dolerites.
Granites, pegmatites, aplites, and granitization complexes.
Feldspar and quartz porphyries.
SCALE
MILES 40 20 0 40 80 120 MILES
Compiled under the direction of T. A. Barnes, M.Sc., Government Geologist.
Issued under the authority of the Honourable Sir A. Lyell McEwin, M.L.C., Minister of Mines.
Published 1958.

FIG. 2.

Generalized geological map of the Mt. Lofty Ranges (base data from Plate A, "Geology of South Australia", 1958 - slightly modified).

Grid indicates locations of 1:63,360 sheets

1 - Eudunda	7 - Echunga
2 - Truro	8 - Mobilong
3 - Gawler	9 - Yankalilla
4 - Cambrai	10 - Milang
5 - Adelaide	11 - Jervis
6 - Mannum	12 - Encounter

The situations of various towns and localities mentioned in the text are also indicated.

A	Adelaide	MB	Murray Bridge	S	Springton
An	Angaston	Mc	Macclesfield	Se	Sedan
As	Ashbourne	MC	Mt. Crawford	SH	Sellick Hill
BH	Black Hill	MM	Mt. Magnificent	St	Strathalbyn
C	Cambrai	Mo	Monarto	T	Truro
G	Gumeracha	N	Nairne	TC	Tanunda Creek
H	Harrogate	O	Oakbank	Tu	Tungkillo
IH	Inman Hill	P	Palmer	VH	Victor Harbour
K	Kanmantoo	R	Rockleigh	W	Williamstown
M	Milendella	RC	Reedy Creek		
Ma	Mannum	RG	Rocky Gully		

The location of the area mapped for the present study is outlined.

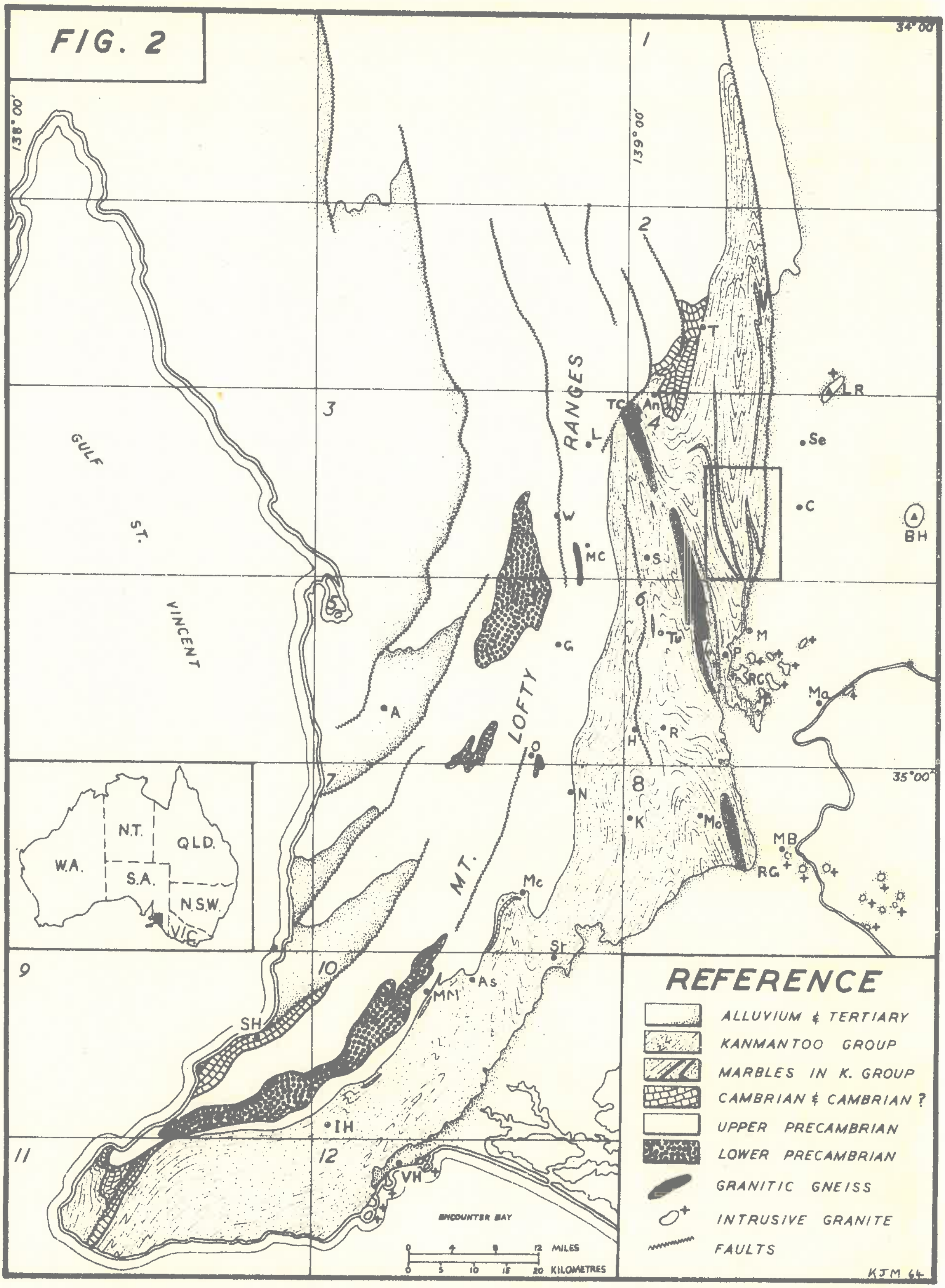


FIG. 2
138°00'
139°00'
34°00'
35°00'
GULF
ST.
VINCENT
MT.
LOFTY
RANGES
ENCOUNTER BAY
W.A.
N.T.
QLD.
S.A.
N.S.W.
VIC.
REFERENCE
ALLUVIUM & TERTIARY
KANMANTOO GROUP
MARBLES IN K. GROUP
CAMBRIAN & CAMBRIAN ?
UPPER PRECAMBRIAN
LOWER PRECAMBRIAN
GRANITIC GNEISS
INTRUSIVE GRANITE
FAULTS
0 4 8 12 MILES
0 5 10 15 20 KILOMETRES
KJM 64

FIG. 3.

Portion of Cambrai 1:63,360 Military Sheet embracing the limits of the mapped area.

ranges. These inliers consist of strongly metamorphosed schists and gneisses showing considerable retrogressive metamorphism associated with phyllonitization, which partially pre-dated and partially accompanied the deposition and deformation of the overlying upper Precambrian sediments of the Adelaide Geosyncline (Talbot, 1962). The basement rocks have been named "Houghtonian" (Howchin, 1926), "Barossian" (Woolnough, 1908) and "Houghton Complex" (Talbot, 1963), and have been considered to be "Archaean" (David, 1922), although there are no age datings in support of this.

The basement rocks are overlain with strong unconformity by upper Precambrian geosynclinal sediments which are only weakly metamorphosed near Adelaide, where they preserve ample evidence for a shallow water facies (Sprigg, 1952). These sediments have been referred to as the "Adelaide Series" (David, 1922), the "Adelaide System" (Torrensian, Sturtian and Marinoan Series) (Mawson and Sprigg, 1950), and as the "Adelaide Supergroup" (Torrens, Sturt and Marino Groups) (Daily, 1963). Regional mapping of the Adelaide Geosyncline, especially in the Flinders Ranges, has lead officers of the State Geological Survey to propose a new stratigraphic subdivision (Thomson et al., 1964).

South of Adelaide, in the Sellick Hill region (and in the Flinders Ranges) the upper Precambrian sequence passes upward without marked unconformity into rocks containing a lower Cambrian fauna. In the Delamere region, south of Sellick Hill, the fossiliferous Cambrian sequence has been found to pass upward conformably into a thick non-fossiliferous sequence of greywackes and impure arkoses of the Kanmantoo Group (Daily, 1963). The Cambrian-Precambrian boundary in the eastern Mount Lofty Ranges has been discussed by Horwitz, Thomson and Webb (1959). The Kanmantoo Group (Sprigg and Campana, 1953) consists of a thick monotonous deep-water sequence of flysch facies rocks, believed to be deposited in a rapidly sinking trough-like basin in Cambro-Ordovician time (Sprigg, 1952). The metasediments mapped in the present study are correlated with the basal portion of the Kanmantoo Group.

Metamorphism, Orogenesis and Age.

The older Precambrian basement rocks were folded, strongly metamorphosed and partially phyllonitized prior to the deposition of the overlying upper Precambrian sediments. The age of this metamorphism and orogeny is not yet known. Regional age studies suggest a middle Proterozoic age (1500 m.y. (?) - Wilson et al., 1960). Thomas (1924) obtained an age of 1000 m.y. for monazite in a pegmatite at Normanville. Biotite from a pegmatite at Myponga has yielded a Rb/Sr age of 500 m.y., a K/A of 565 m.y. and associated uraninite has yielded a U/Pb age of 520 m.y. (Wilson et al., 1960).

The younger Precambrian sediments are believed to be of upper Proterozoic age (David, 1932) because they pass upward into a fossiliferous Cambrian sequence. Recent Rb/Sr age determinations by Dr. W. Compston at Australian National

University, Canberra (Compston, in press) suggest that the upper Precambrian sediments of the Adelaide Geosyncline may be 600-1300 m.y. old. Dr. Compston has also given a Cambrian Rb/Sr age to spilitic volcanic rocks at the base of the Kanmantoo Group near Truro.

Although there is some evidence of middle Cambrian orogenic activity on Kangaroo Island, the main conformable folding of the upper Precambrian, Cambrian and Kanmantoo sequences occurred during an early Palaeozoic orogeny. This orogeny was accompanied by metamorphism and igneous activity in the eastern Mount Lofty Ranges, and was probably largely extinct before the intrusion of the bathylithic granites outcropping in an arcuate belt on the eastern margin of the ranges (Fig. 2). These granites have been given an Ordovician age (457 m.y., K/A on biotite, Victor Harbour granite, Evernden and Richards, 1962; 420 m.y. and 440 m.y., Victor Harbour and Cold and Wet granites, U/Pb on zircons, Fander, 1961; 460 m.y., Palmer and Anabama granites, Rb/Sr, Compston, in press).

Although remaining a zone of positive uplift throughout the Palaeozoic and Mesozoic (Campana, 1955), the Mount Lofty Ranges have been strongly uplifted in Tertiary to Recent times. Tertiary seas overlapped onto the bedrock on the western and eastern sides of the ranges but the central portion has remained emergent throughout the Tertiary period (Glaessner, 1953).

The Kanmantoo Group.

The metasediments of the present area lie within the Kanmantoo Group as defined by the State Geological Survey. Recent mapping by the Survey has delineated the extent of the Kanmantoo Group (Fig. 2). Stratigraphic correlations (Horwitz, Thomson and Webb, 1959, Daily, 1963) and a recent age determination on the Truro Volcanics (Compston, in press) has indicated that the Kanmantoo Group rocks are of Cambrian age or younger.

Tate (1897) gave a Silurian age to rocks of the Mount Lofty Ranges and considered that the rocks on the eastern side of the ranges were the youngest. Brown (1886) produced the first geological map of South Australia, which showed the rocks of the western Mount Lofty Ranges as a single sequence of Silurian age. In 1893 Tate considered that the rocks were Precambrian rather than Silurian, and remarked on the anomaly of the younger rocks to the east being more metamorphosed than the older rocks to the west.

With the discovery of a Cambrian fauna at Sellick Hill (Howchin, 1897), Howchin (1906) proposed that the rocks of the Mount Lofty Ranges should be divided into two sequences, a "lower Cambrian" sequence overlying the metamorphosed Precambrian basement complex occurring as inliers in the core of the ranges. Woolnough (1908) considered that the rocks on the eastern side of the ranges, being highly metamorphosed, were also part of the Precambrian basement. He proposed the term "Barossian" or "Barossian Series" for these rocks. Benson (1909), discussing the petrology of various metamorphosed rocks in the Mount Lofty Ranges, used the term Barossian Complex to embrace both Woolnough's Barossian Series and the rocks of the basement inliers forming the core of the ranges.

Howchin (1926) showed that Woolnough's "Barossian" was the metamorphosed equivalent of his "Cambrian" sequence on the western side of the ranges. He emphasized the older character of the rocks at Houghton, which he termed the "Houghtonian", "Archaean" or "Fundamental Complex". Howchin (1929) considered that the Barossian Complex Kanmantoo Group was a deep-water facies of the Precambrian Adelaide Series (David, 1922), which overlay the Fundamental Complex at Houghton. Hossfeld (1935) after a reconnaissance geological investigation of the northern Mount Lofty Ranges rejected Howchin's (1926) ideas and followed Woolnough's thesis that the Barossian on the eastern side of the ranges, and Howchin's Houghtonian were the same series.

The term Kanmantoo Group was first used in mapping the Adelaide 1 mile sheet (Sprigg, Whittle and Campana, 1951). In mapping the Gawler 1 mile sheet Campana (1951) disagreed with Hossfeld, and considered the Barossa Ranges to be composed partially of Adelaide System rocks and partially Kanmantoo Group rocks.

Sprigg (1952) showed that sedimentation in the Adelaide Geosyncline took the form of a continental terrace outgrowth from the Western Australian shield, and that sedimentation continued during the early Palaeozoic, in the Kanmantoo trough, after ceasing elsewhere. Sprigg and Campana (1953) formally defined the Kanmantoo Group for that monotonous sequence of non-fossiliferous evenly bedded phyllites, siltstones and impure arkoses, more than 30,000 feet thick, on the eastern side of the Mount Lofty Ranges. They showed that this sequence conformably overlay the fossiliferous lower Cambrian of the Sellick Hill region, but was separated from the Adelaide System on the eastern side of the ranges by a fault (the Nairne fault), which has since been partially discredited. Campana and Horwitz (1955) considered that the Kanmantoo Group was transgressive over the Adelaide System sediments, the Nairne fault being regarded as an overlap with some local movement. Kleeman and Skinner (1959), working in the Strathalbyn-Harrogate region proposed that the base of the Kanmantoo Group be placed at the base of the Nairne Pyritic horizon, which the State Geological Survey considered to lie within the Kanmantoo sequence. Horwitz, Thomson and Webb (1959) have summarized the present position of the Survey regarding the base of the Kanmantoo Group.

<u>Previous Investigations in the Present Area.</u>

Brown (1908) has recorded the early activities of the three most important mining ventures in this area. Between 1849-1851 and 1859-1863 the North Rhine Mine was worked for copper. Twenty tons of 20% copper ore (azurite and malachite) were sent to England, but later more extensive operations at depth proved this field to be uneconomic. Between 1890-91 the Mount Rhine Silver Mining Company operated the Paint Mine for silver and traces of gold in galena, some ochre also being raised. The Kanappa Mine was operated intermittently between 1867 and 1902, some 400-500 tons of copper ore being removed. A small parcel of azurite was recently obtained by a local prospector from a recently exposed copper show south of the Kanappa Mine, discovered during the present mapping programme.

Hossfeld (1935) included portion of the present area in his regional geological map of the northern Mount Lofty Ranges. In 1951 a number of Honours students (Markham, Rowley, Harms, Kaewbaidhoon) attached to the Department of Economic Geology, undertook brief mapping projects using aerial photographs, including portions of the present area. Jack (1923) and Johns (1963) have commented on the economic potential of marbles in the Cambrai area. The Cambrai 1 mile geological sheet, on which the present area is situated, has not yet been published by the Geological Survey. Chinner (1955) and White (1956) have carried out extensive petrological studies to the south and west of the present area.

Physiography and Geography.

The eastern side of the bedrock exposure is bounded by the Milendella fault, an old Palaeozoic fault rejuvenated during

Tertiary to Recent movements, which, by analogy with eastern Australia, have been attributed to the Kosciuskan Orogeny (Howchin, 1929; David and Browne, 1950). Prior to the recent uplift the Mount Lofty Ranges had reached a state of peneplanation, so that the bedrock physiography of the present area is that of a heavily dissected uplifted plateau. Remnants of the plateau lie at 12-1400 feet in the mapped area, 1000-1200 feet above the general level of the pre-Tertiary surface east of the Milendella Fault. The eastern side of the area is dominated by a strongly eroded scarp topography, with numerous steeply graded subsequent watercourses of endoreic character. These streams are most effective as erosive agents during sharp summer thunderstorms or heavy winter rains when loose bedrock material is scoured from their channels, the bulk being discharged on pediment fans at the base of the scarp, and the water with suspended matter ending in alluvial pans a few miles east of the scarp. The larger weakly graded antecedent streams (Marne River, Somme River, Saunders Creek) effectively drain the higher rainfall areas to the west. Because bottom erosion in these streams kept pace with the Tertiary uplift, deep meandering gorges have been cut through rocks resistant to erosion by subsequent streams. These larger streams are also seasonal, but during the high rainfall winter months they may discharge their waters into the River Murray.

The climate is essentially mediterranean, with cool rainy

FIG. 4.

General view to the south-east across the mapped area from the western side (location 154-145) looking towards Saunders Creek. The Murray Plains near Mannum are seen in the far distance. Area in the foreground and levelled hill-tops in the middle distance outline the old Tertiary peneplained surface at about 1250' above present sea level. Small monadnock in right distance reaches 1350'. Blocky outcrops of well-bedded meta-arkose formation in foreground.

FIG. 5.

General view of youthful erosional topography on the eastern scarp south of Saunders Creek, looking south from 198.080. Smooth barren hills with sparse grass and occasional stunted sheoaks, and gently graded larger watercourses lined with gum trees are typical features of the eastern side of the area.

FIG. 4

FIG. 5

winters and hot dry summers, the pattern being broken only by rare summer thunderstorms of monsoonal origin. The mean annual rainfall varies from 25 inches on the western side of the area to 15 inches at the base of the scarp to 12 inches at Sedan, a small town in the rain shadow on the plains east of the scarp. (Jessup, 1948).

In the more eroded eastern portion of the area the bedrock exposure is excellent. The more mature topography to the west harbours numerous areas of shallow alluvium overlying softer bedrock units.

Most of the mapped area is open savannah woodland (Jessup, 1948). In the skeletal soils, sparse sheoaks (Casuarina stricta) dot rolling hills of open grassland. Stands of red gum (Eucalyptus camaldulensis) and blue gum (Eucalyptus leucoxylon) are limited to the more permanent watercourses and springs and to the higher rainfall areas to the west.

SECTION A. - SEDIMENTATION.

In the metamorphosed rocks of the Cambrai area, sedimentary grains and textures and sedimentation structures may be preserved, or persist in a modified form, in schists and gneisses of high metamorphic grade. The style and course of tectonic deformation differs in rocks of dissimilar lithologies. The importance of original lithology in determining the relative response of a sediment to both metamorphism and tectonism has been emphasized by Reed (1962). The purpose of this preliminary section is to outline those features of the stratigraphic sequence in the Cambrai area which have a bearing on the subsequent discussions of metamorphism and tectonic activity.

CHAPTER 1.

STRATIGRAPHY AND SEDIMENTATION.

(a) The Stratigraphic Sequence.

The majority of sedimentary rocks in the area under consideration are monotonous fine-grained grey quartz-felspar-biotite sandstones with prominent biotite bedding laminations. Within the thick strongly folded sequence, several minor pelitic beds, numerous thin calcareous pelites and a calcitic limestone, now metamorphosed in varying degrees to phyllites, schists, gneisses, calc-schists, calc-silicate rocks and marbles, serve as stratigraphic marker units. A general outline of the stratigraphic sequence is presented in Table 1, and in the legend of Plate 1. The table also presents estimates of the thicknesses of various stratigraphic units in four portions of the mapped area which are not greatly complicated by fold closures. The results indicate a large variation in the thickness of some units. The maximum total thickness of the exposed sequence is estimated at 22,300', but this figure is probably excessive.

The overall effect of deformation on the stratigraphic thickness has not been determined. Observations on the distortion of sedimentation structures in folded beds has indicated that the distortion of bed thicknesses by pure shear and compressive effects may be considerable. These tectonic effects cause a thinning of the beds within the fold limbs,

TABLE 1.

STRATIGRAPHIC SUCCESSION - CAMBRAI AREA.

Rock type	(1)	(2)	(3)	(4)
YOUNGEST				
Quartzo-felspathic schists and gneisses	-	3730'	-	-
Cross-bedded meta-arkose (Probable equivalent of Inman Hill Formation)	-	3940'	6510' plus	-
		7670'	6510' plus	
Micaceous schists	260'	608'	382'	-
Quartzo-felspathic schists	1090'	1100'	3280'	800'
Calc-silicate bed	-	40'	134'	64'
Quartzo-felspathic schists	560'	816'	2530'	-
	1910'	2570'	6320'	
Impure marble	-	-	190'	-
Quartzo-felspathic schists with minor greywacke pebble beds and calc-silicate lenses	-	-	-	2110'
Quartzo-felspathic schists with numerous calc-silicate beds	-	-	-	5250'
Calcareous schists OLDEST	-	-	-	1950' plus
				9500' plus

(1) Section in north-west corner of mapped area.
(2) Section in south-west corner of mapped area.
(3) Section in central portion of mapped area.
(4) Section on east side of mapped area.

and a thickening within the fold closures. Thus, the thickness of a measured sequence will depend on its position within the major fold structure. The apparent thickness of a sequence in which the bedding planes generally lie at an appreciable angle to the axial plane cleavage (e.g. Col. (3), Table 1), will be expected to be greater than those sequences in which the angle between the bedding planes and the axial plane cleavage is not appreciable (Cols. (1), (2), (4), Table 1). The sequence indicated in column (3), in which the section line may lie closer to the circular section of the deformation ellipsoid in the area, may be considered to be a better estimate of the original stratigraphic thickness than those of columns (1), (2) and (4), which were taken from the extreme limb structures. However, the thickness variations obtained are not considered to be entirely due to tectonic effects; some original sedimentary variations may be present.

It should be noted that some beds, for example, thin calc-silicate marker units, are remarkable for their continuity and uniformity of thickness over long distances; beds less than one foot thick have been traced along strike for many hundreds of yards. Visible facies changes and rapid changes in bedding thickness are rarely observed.

(b) Correlation of the Cambrai stratigraphy with other portions of the Mt. Lofty Ranges.

Regional mapping by the South Australian Geological Survey has indicated that the meta-sediments of the present area lie within, and near the base, of the Kanmantoo Group as currently defined by the Survey. The thick meta-arkose unit and the marble horizon have recently been mapped on the Truro 1-mile sheet (Coats and Thomson, 1959). The marble was mapped by White (1956) in its continuation south of the present area to Milendella, and was considered at that time to be a possible equivalent of the Angaston, Tungkillo and Macclesfield marbles, which were tentatively equated with the Archaeocyatha limestones of the Sellick Hill region, on the unmetamorphosed western side of the Mt. Lofty Ranges. Recent mapping by the Survey in the Macclesfield-Ashbourne-Mt. Magnificent region has led to a more detailed correlation of the Cambrian-Precambrian boundary relationships on the eastern side of the Mt. Lofty Ranges (Horwitz, Thomson and Webb, 1959). A black phosphatic and pyritic shale, traced intermittently on the eastern side of the ranges, has been equated with the fossiliferous phosphatic Heatherdale Shales at Sellick Hill. The top of this shale is now defined as the base of the Kanmantoo Group. This has placed the Macclesfield marble within the basal portion of the Kanmantoo sequence. The marble unit mapped in the present area is considered to be approximately equivalent to the Macclesfield marble and the sequence above and below this marble are considered to belong to the Kanmantoo Group. In the 9000' sequence below the marble in the Cambrai area no dark phosphatic phyllites have been observed which might indicate that the base of the Kanmantoo Group had been reached.

The bedded acid volcanics now recognised a few miles north of this area, and extending northwards onto the Truro sheet, situated at or near the base of the Kanmantoo Group, have not been observed in the present area. The thick meta-arkose, with its characteristic cross-bedded and foreset slump structures, has been found to be a thick and persistent unit near the base of the Kanmantoo Group throughout the eastern Mt. Lofty Ranges, and can be satisfactorily correlated with the Inman Hill Formation (Forbes, 1957).

Kleeman and White (1956) have published a general succession for the Kanmantoo Group in the metamorphosed Palmer-Tungkillo-Harrogate area. The stratigraphy of the present area may be correlated with their units 1-5.

(c) Sedimentation Structures in the Cambrai Succession.

Several kinds of sedimentary structures have been observed through the sequence, and because they are preserved in a modified form in the deformed and metamorphosed rocks, they must be distinguished from structures produced solely by metamorphic or tectonic activity. It is not always possible to determine an unequivocal origin for certain structures. All sedimentation structures have been modified to some degree by imposed tectonic strains, and in many cases metamorphic recrystallization has caused sedimentary structures to become more or less obvious than in the original sediment.

(1) Bedding and bedding lamination.

The majority of quartzo-felspathic schists are well-bedded and show weak micaceous separation planes which cause platy weathering on exposure. Individual beds may vary considerably in thickness, the thicker beds being generally coarser grained. Although some bedding units may be massive, the majority are subdivided by numerous thin biotite laminae. In the finer grained rocks these laminations may be very closely spaced and show remarkable continuity throughout a rock exposure.

The marble unit is an impure calcite marble, generally containing no more than 80% calcite, which possesses a strong lamination or layering at all metamorphic grades. In the low grade marbles the layering consists of fine-grained phyllitic laminae of undoubted sedimentary origin.

(2) Heavy mineral lamination.

In thin sections heavy minerals are often seen to be concentrated in the biotite laminae. Rare layers exceptionally rich in biotite and heavy minerals up to 6" in thickness have been observed within the more arkosic rocks. In some instances the heavy mineral bands have clearly filled wash-out structures cut into underlying laminated arkose (Fig. 7), while in other instances long continuous layers of heavy minerals

FIG. 6.

Uniform lamination and compositional banding in laminated quartzo-felspathic schist below thick calc-silicate bed (Marker 2). Location 169.217. Coin diameter 2.8 cm.

FIG. 7.

Biotite bearing heavy mineral bands (artificially outlined by pencil) filling wash-out structure in laminated quartz-felspar-biotite meta-arkose. A second heavy mineral band occurs in bottom right-hand corner. *Note small-scale displacements on joints in* bottom half of picture. Location 187.170. Coin diameter 2.8 cm.

FIG. 6

FIG. 7

occur as topsets to coarse cross-bedded units. In hand specimens these bands are black, heavy and slightly radioactive. An analysis of a typical heavy mineral band is presented in Table 2. Zircon is a common constituent as rounded clear to cloudy grains. Rounded to angular pale yellow monazite lends the rock its radioactivity. The concentration of biotite in these bands suggests that biotite or its precursor was an original sedimentary constituent. There is no indication that this biotite was formerly amphibole or pyroxene. Biotite bands have been observed in non-metamorphosed sediments by Tucker (1960).

(3) Cross-bedding.

There are two types of cross-bedding in rocks of the Cambrai area.

(a) In the finer grained rocks a small scale cross-lamination is seen to be closely associated with asymmetric ripple marking. Beds up to several feet thick may exhibit festoons of small scale cross-bedding associated with successive layers of ripple marks.

(b) In the coarse-grained arkoses cross-bedding on a medium to large scale is ubiquitous, and is especially notable in the excellent exposures of the Marne Gorge. Foreset units up to 3-4 feet

TABLE 2.

MECHANICAL ANALYSIS OF HEAVY MINERAL BAND FROM ARKOSE

Weight of sample before treatment	...	513 grm.
Weight after crushing and washing	...	440 grm.
Percentage weight loss	...	16%

Composition.	Weight percent
Hematite	78.9
Biotite	9.0
Magnetite	5.4
Zircon	4.2
Muscovite	1.1
Monazite	0.9
Xenotime	0.3
Quartz	0.2
	100.0

FIG. 8.

Micro-cross-bedding in quartz-felspar-biotite arkosic schist, viewed in longitudinal section. Location 182.155. Coin diameter 2.8 cm.

FIG. 9.

Large scale cross-bedding in cliff outcrop of meta-arkose in Marne River gorge at 200.176. Scale is 1-foot (30 cm.) rule. Note tapering out of bottom sets.

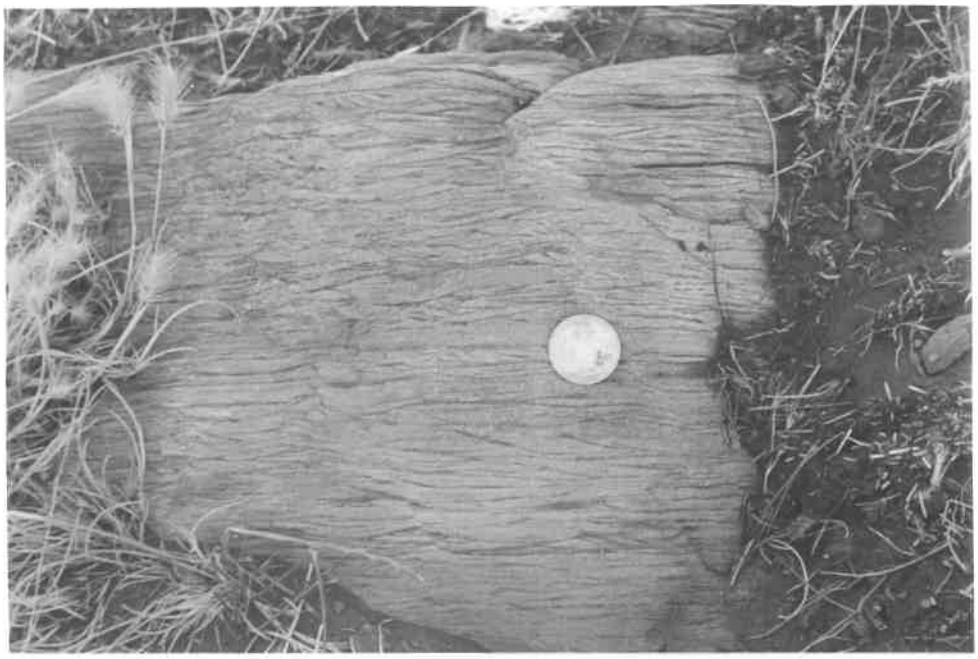


FIG. 8

FIG. 9

thick are not uncommon. Occasionally broad trough or festoon bedding appears where the foreset units are not separated by bedding surfaces.

Common to all coarse cross-bedded arkoses is the occasional appearance of a cross-bedded unit whose foresets appear to have slumped forward down the foreset slope, producing, in cross-section, a resemblance to plastic folding (Fig.11). This type of overturned cross-bedding has been recorded by numerous workers, both in old and recent sediments (e.g. Harms, Mackenzie and McCubbin, 1962, and Prentice, 1960), excellent examples have been figured by Wilson (1956) and Potter and Pettijohn (1963, Pl 6B), and similar structures have been produced experimentally by McKee, Reynolds and Baker (1962). The presence of the thin biotite laminae in the Cambrai examples may have encouraged the formation of these structures. As the tops of the overturned foresets have usually been eroded, or transected by an overlying layer, they may have the appearance of sheared out (transposed) folds. In the low grade rocks there is no doubt that these structures are of sedimentary origin. In the higher grade portions of the area some tectonically transposed folds have been observed, and

FIG. 10.

Coarsely cross-bedded arkose overlain by thin tapering non-laminated greywacke layer, a thin bed with micro-cross-bedding, a second thin unsorted greywacke and laminated arkose. Location 203.177. Coin diameter 2.8 cm.

FIG. 11.

Slumped cross-bedding in arkose. Note especially the slurring of the lamination in the slumped portion. Bottom parts of the fore-set slopes appear undisturbed. Location 181.243. Coin diameter 2.8 cm.

FIG. 10

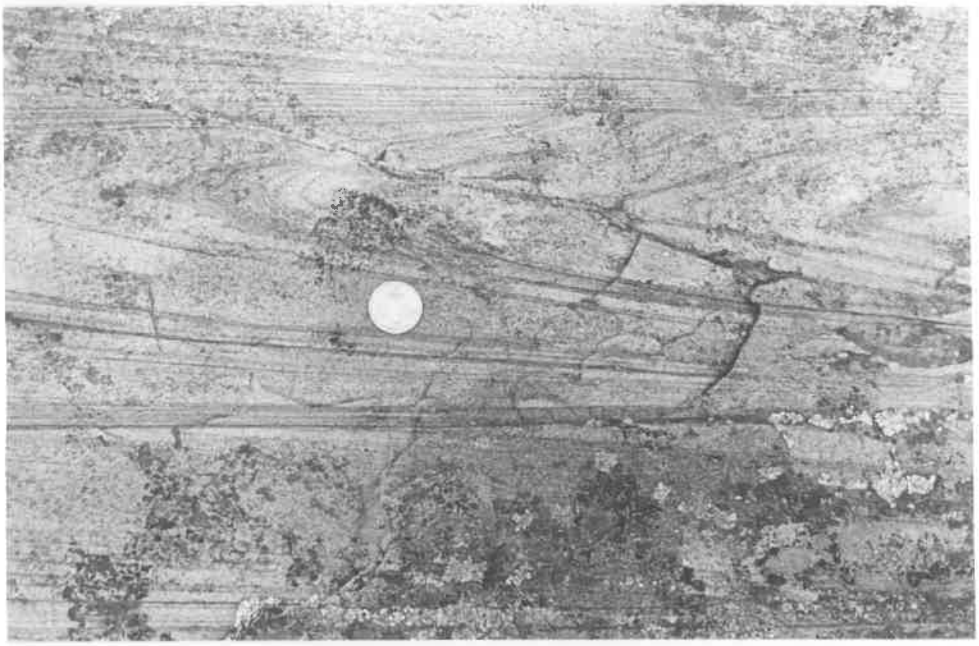


FIG. 11

and the distinction between sedimentary cross-bedding and tectonic transposition is often indefinite. On the Mannum sheet White (1956) found many apparently cross-bedded structures which he attributed to transposition in folding. The abundance of overturned cross-bedding in the lower grade parts of the present area suggests that many of the apparent transposed folds in the high grade rocks may also be of sedimentary origin. Similar structures have been recorded by Horwitz, Thomson and Webb (1959) from the Inman Hill Formation in the southern part of the ranges.

(4) Ripple marking.

In many of the finer grained sandstones, asymmetric ripple marking is a commonly observed feature. The ripple crests are usually parallel and show little bifurcation. Associated cross-bedding structures indicate that these ripples are of the transverse type. The less deformed ripples show sufficient asymmetry to demonstrate a current sense. The importance of pure and simple shear in the deformation of fine-grained ripple marked rocks is reflected in the considerable distortion of the ripple mark profiles. In the more recrystallized rocks, where the small scale cross-lamination may be destroyed, the distinction of ripples from pseudo-ripples (Ingerson, 1940) is

FIG. 12.

Top surface of ripple marked bed. Current sense from bottom to top of outcrop, initially west to east. Location 211.181. Coin diameter 2.8 cm.

FIG. 13.

Outcrop of intraformational pebble bed. Quartzo-felspathic pebbles in unsorted quartz-felspar-biotite matrix. Location 222.141. Coin diameter 2.8 cm.

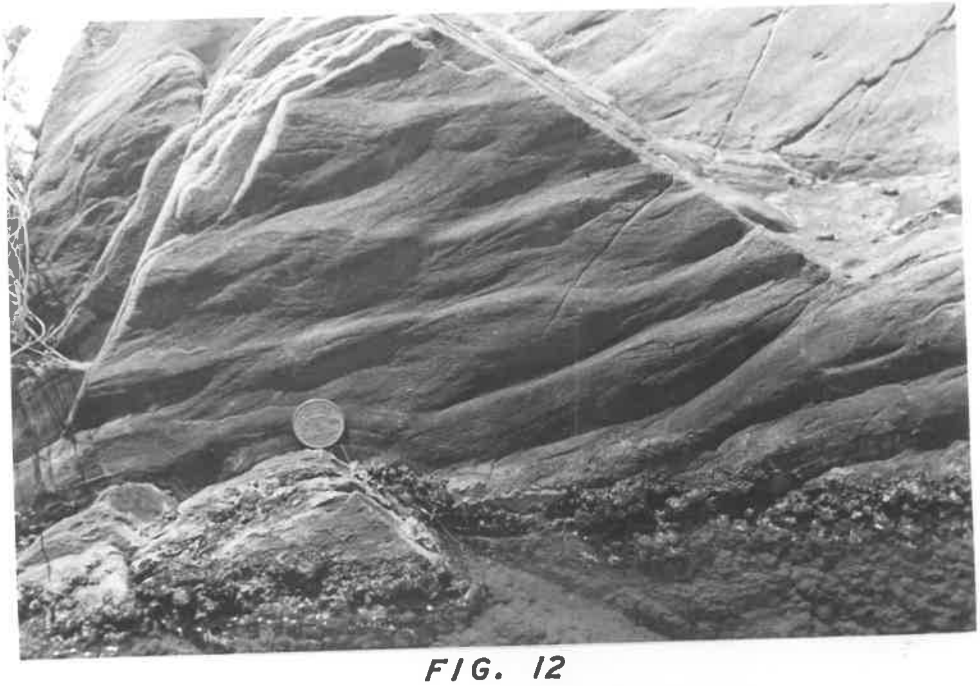


FIG. 12

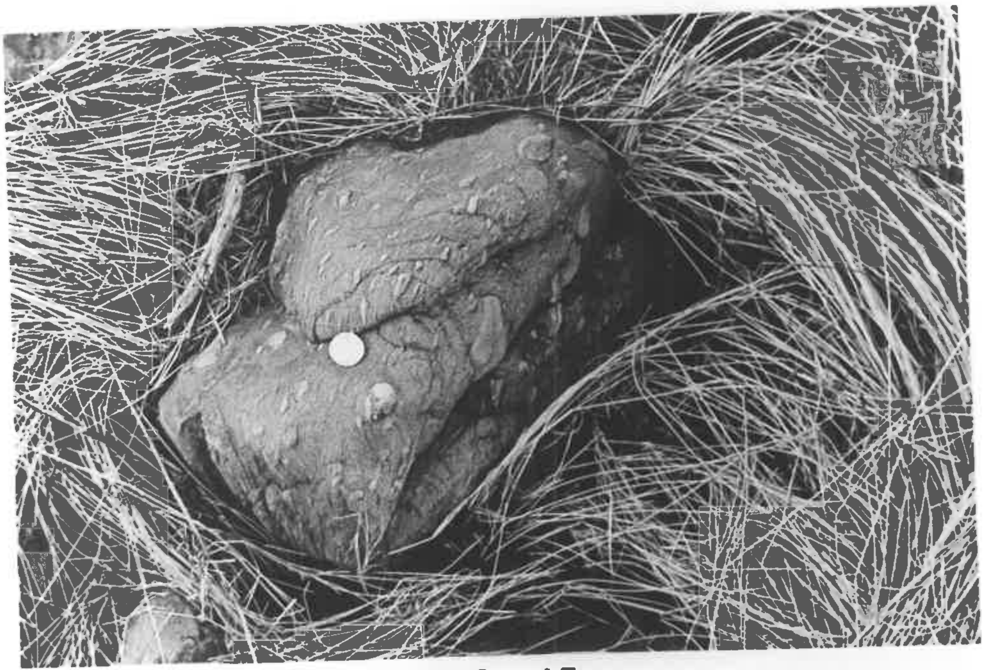


FIG. 13

difficult, especially as the ripple crests are oriented in a north-south direction sub-parallel to the tectonic axes. However, it is suspected that most of the distorted tectonic looking rucking structures in the high grade rocks are, in fact, distorted ripple marking. Ripples of the symmetric type have not been observed.

(5) Other miscellaneous structures.

Several other less common sedimentation structures have been observed. Occasional small-scale washout structures have been noted in the coarser arkoses. Convolute bedding has been seen in several fine-grained sandstone beds. Load cast features are not uncommon where fine-grained sandstones overlie more pelitic units. Several minor lenticular greywacke pebble beds occur intermittently below the limestone unit. Some of these have allochthonous pebbles, while others are intraformational conglomerates produced by currents stirring up pre-existing sedimentary layers. Fig. 14 shows sedimentary breccia fragments of pelitic schist in the base of the massive meta-arkose unit. These fragments were derived from the pelitic beds lying immediately below the arkose.

(d). Palaeocurrent Data.

While mapping the arkoses of the Marne gorge, an opportunity was taken to measure some of the excellent exposures of

FIG. 14.

Outcrop near base of uniform well-laminated meta-arkose formation stratigraphically overlying mica schist beds (Marker 1). Note fragments of mica schist caught up in the arkose. Location 182.241. Coin diameter 2.8 cm.

FIG. 15.

Well laminated arkose showing scour structures filled with unsorted greywacke. Location 203.177. Coin diameter 2.8 cm.

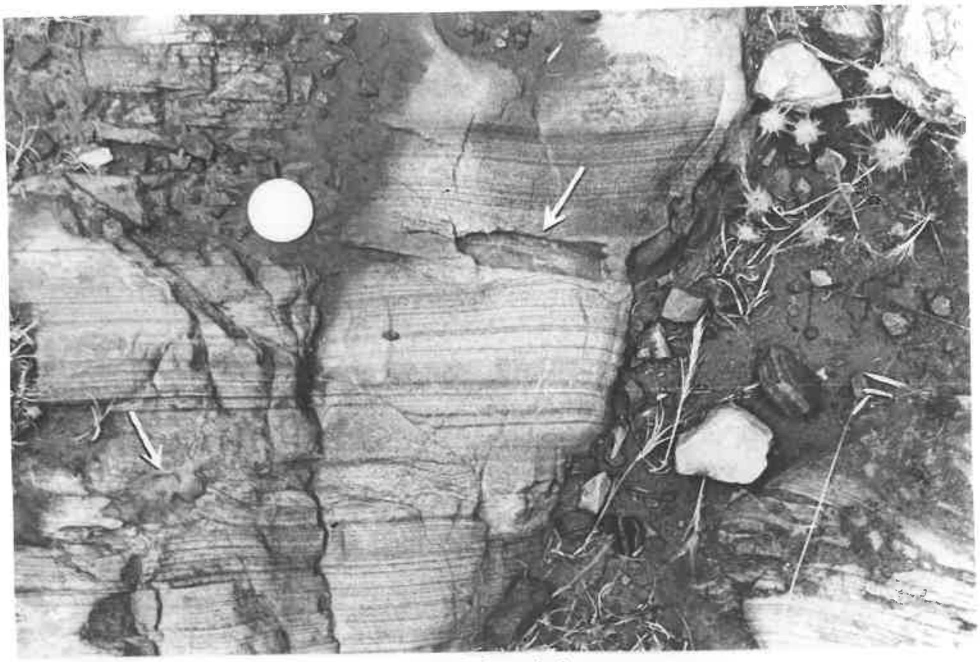


FIG. 14

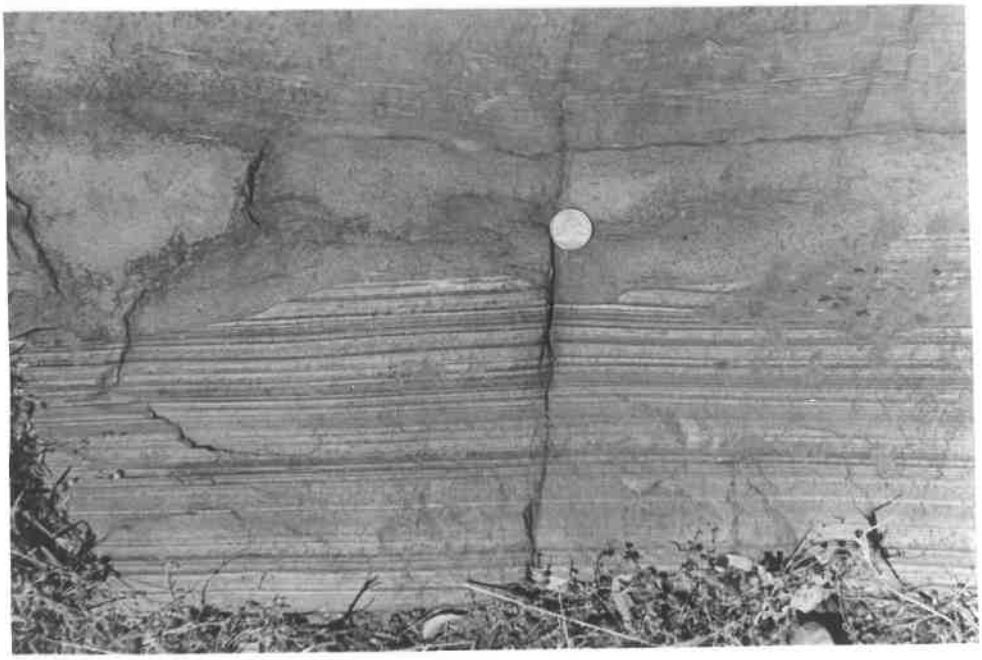


FIG. 15

both topset and foreset planes of cross-bedded units in order to obtain some preliminary data on the palaeocurrents in this portion of the Kanmantoo Trough. Little work of this type has yet been attempted either in the Kanmantoo Trough or in the Adelaide Geosyncline as a whole, although there is an abundance of field material for this line of research.

In this area of closely appressed plunging folds it is necessary both to un-plunge and un-tilt the bedding before the palaeocurrent directions can be estimated. At each station the dip and strike of the foresets, bedding, cleavage and fold plunge are required. A stereographic method similar to that of Ramsay (1962) was used for unfolding the data. The cleavage, being always steep (70-80°), was assumed to be vertical, so that the fold plunge (usually less than 40°) was brought to the horizontal in a vertical plane. This assumption simplified the analytical procedure, and was found to have very little effect on the final azimuth. Having un-plunged the structures the bedding was then brought to the horizontal and the current direction determined. A similar method was adopted in ripple mark reconstruction. This method assumes simple flexural folding, which may have been approximated in the coarse arkosic units from which the data was collected. However, both pure and simple shear have also been effective in these rocks. Because of the uncertainties involved (Ramsay, 1962), the quantitative effect of these forms of deformation on the sedimentary structures has not been determined.

One hundred cross-bedding sets were measured in each of five sub-areas (Fig. 16), and 113 ripple axes were measured in a cliff face at 211.182. The results are presented in figure 17 where a remarkable consistency of data is evident. Both the cross-bedding and ripple marking data indicate a strong west-east current trend, with a current sense from west to east. Allowing for the approximations made in unfolding the beds, there is no appreciable variation in current trend in the five sub-areas and over at least 8000' of stratigraphic section. The average reconstructed foreset dips in sub-areas 1-5 are 17.32, 16.29, 16.07, 17.26 and 18.65 respectively. With respect to the major synclinal structure in the area from which the data were taken, 458 west limb readings yielded an average reconstructed foreset dip of 16.82°, and 42 east limb readings 20.38° (with some individual reading up to 55°). The difference between these two values gives some indication of the effects of shear folding on the cross-bedding sets. As the foresets are predominantly east dipping the larger angle on the eastern synclinal limb is to be expected (Fig. 16).

<u>Conclusions</u>.

The ultimate object of sedimentation structure studies is to reveal something of the original environment of sedimentary deposition. The continuity and evenness of bedding of the majority of the Cambrai rocks suggests sedimentation in a gently inclined smooth bottomed submarine trough. The absence of quartzites, pure arkoses and other sorted rock types

FIG. 16.

A. Location of sub-areas in which palaeocurrent data was collected.

Sub-areas 1-5: Cross-bedding data.

Sub-area 6: Current ripple data.

B. Distortion of cross-bedding fore-sets in a syncline under the influence of flexural bedding plane slip.

Original cross-bedding (dotted) distorted to new orientation (solid lines).

(After Pettijohn, 1957, p.169).

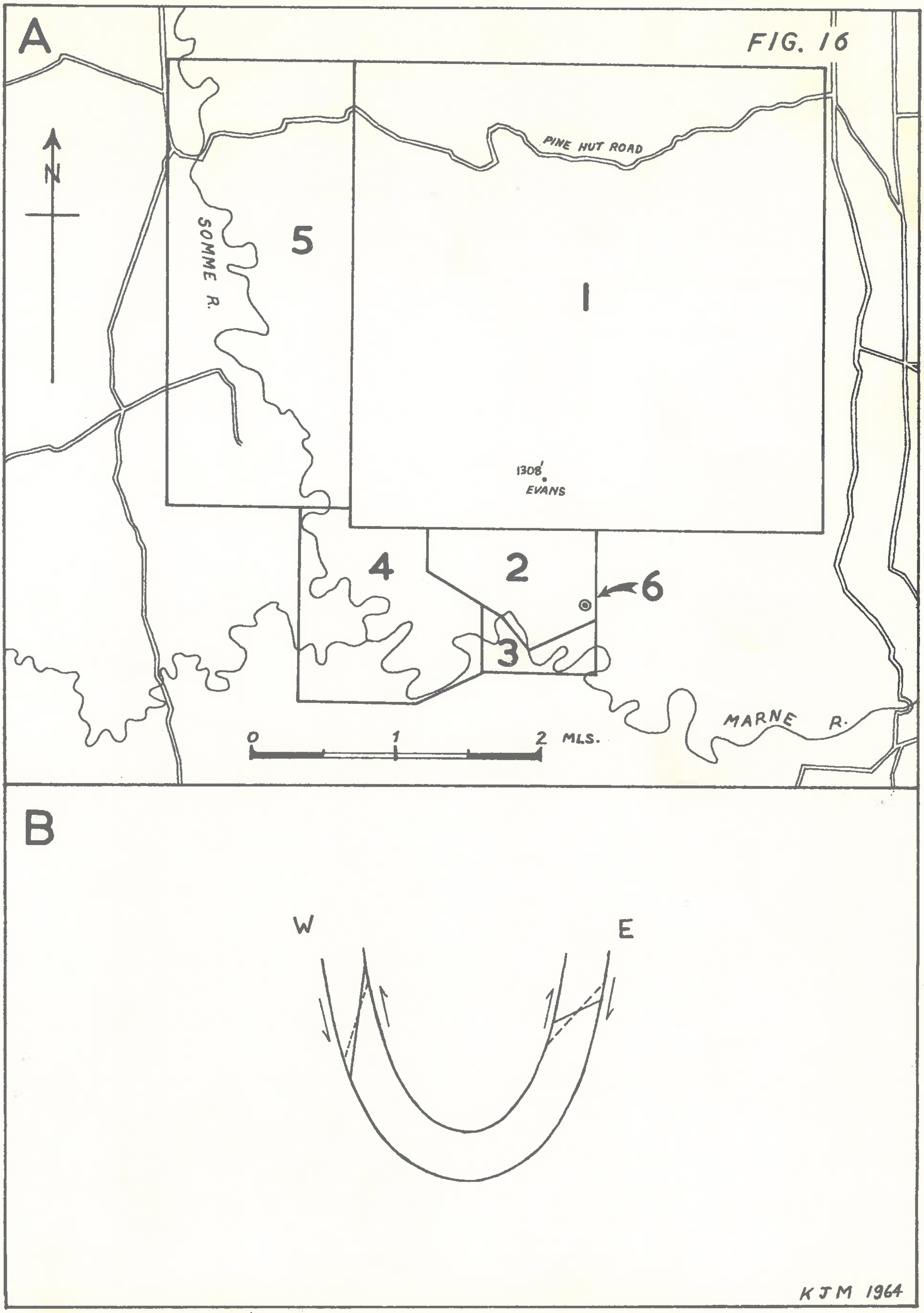

A
FIG. 16
N
PINE HUT ROAD
SOMME R.
5
1
1308'
EVANS
4
2
6
3
MARNE R.
0
1
2 MLS.
B
W
E
KJM 1964

FIG. 17.

Rose diagrams of palaeocurrents determined in the Cambrai area.

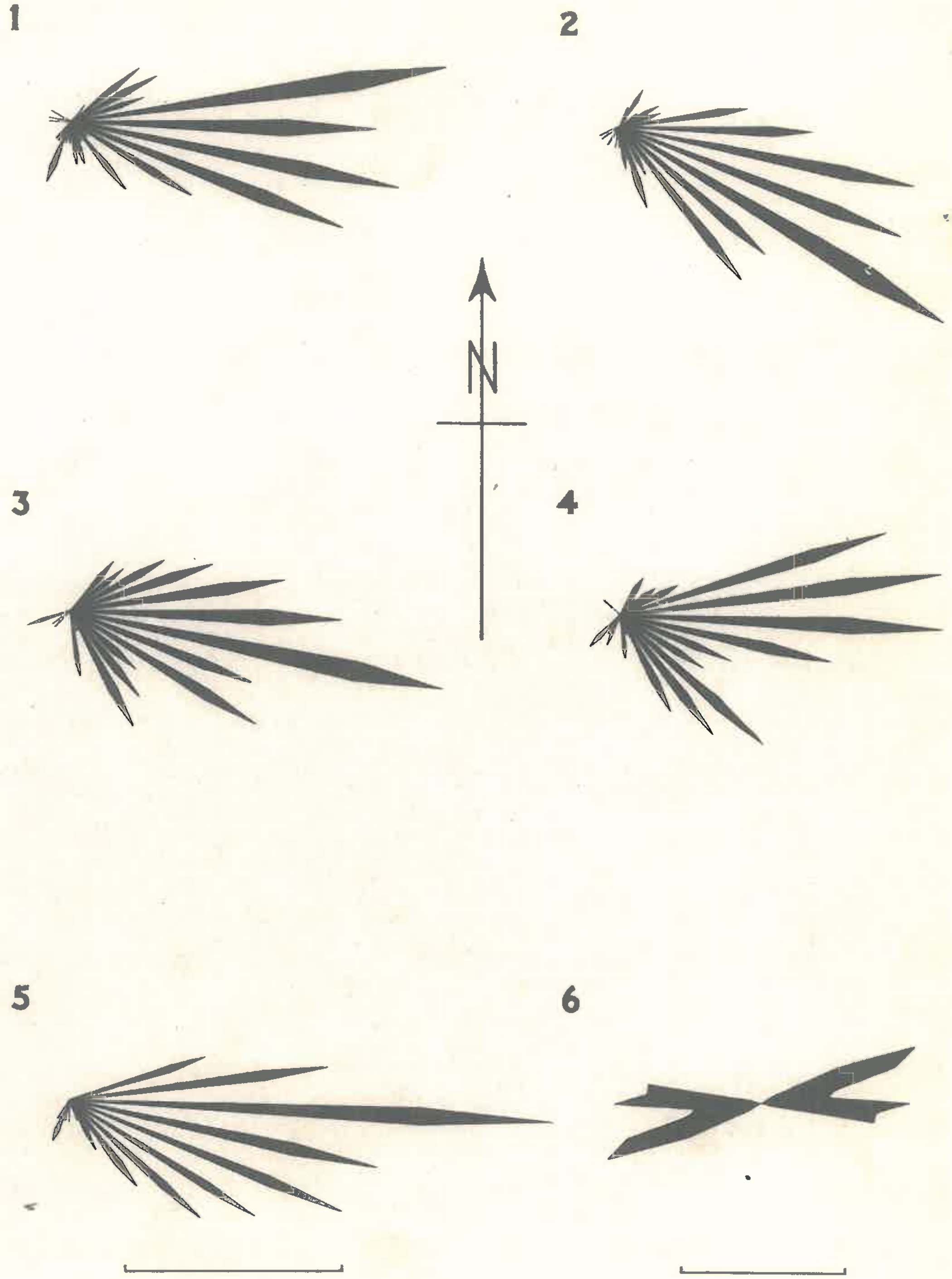
Diagram numbers correspond to respective sub-areas of Fig. 16.

Diagrams 1-5. Palaeocurrents determined from reconstituted cross-bedding fore-sets.

Diagram 6. Palaeocurrents determined from reconstituted current ripple mark axes.

Each diagram contains 100 current directions. In diagrams 1-5 the length of each arm of the rose is proportional to the number of readings within a ten degree compass interval (length of scale below diagram 5 is equivalent to 10 readings). In diagram 6 the arms were initially drawn at 5° intervals (length of scale below diagram 6 is equivalent to 20 readings).

FIG. 17



K.J.M. 1964.

indicates deposition below the depth of influence of surface wave action. The current ripples and cross-bedded structures could have been produced by submarine currents at considerable depths. The presence of convolute and slumped bedding structures suggest a sloping sea floor. Rare greywackes, intraformational conglomerates and sedimentary brecciation may be attributed to turbidity currents and severe current abrasion. The cross-bedded arkoses indicate sedimentation of the coarser material by traction currents, while the finer grained rocks containing abundant and often very delicate laminations are best considered as a result of settling of sediment in a relatively still environment from sediment carrying currents at higher levels in the ocean.

In the coarsely cross-bedded rocks a general sedimentary cycle can often be recognised. These cycles apparently begin with severe current abrasion causing wash-out structures in well-laminated arkose. The wash-out structures may be covered with a layer of heavy minerals or filled and overlain by a totally unsorted arkosic greywacke. This massive greywacke passes upward into coarsely cross-bedded arkose, apparently produced under moderate currents with ample sediment supply. Above the cross-bedded units the sediment becomes finer grained and coarse cross-bedding gives way to cross-lamination and current rippling. At this stage the currents have apparently become gentle and the sediment supply weak. The top of the cycle is represented by evenly laminated micaceous silts and arkoses apparently sedimented under quiescent conditions by settling out of sediment from higher level currents. The next cycle may then commence with wash-out structures and the sequence is repeated. The thickness of each cycle unit is variable and may average about 10'.

A brief examination of the sedimentation structures has revealed the necessity to distinguish slump folds and ripple axes from transposition folds and tectonic rucking. The problem is made especially acute in this area because the axes of slump folds and ripple crests are oriented north-south and sub-parallel to the tectonic axes.

SECTION B. - METAMORPHISM.

In the vast area occupied by the Adelaide Geosyncline, perhaps 70-100 thousand square miles, there are only a few small areas in which rocks have reached a grade of metamorphism equivalent to the biotite zone. The largest of these is the Mt. Lofty Range metamorphic belt, a north-south elongate zone in which the metamorphism reaches the sillimanite grade. Other smaller areas of biotite zone metamorphism, resulting in lustrous phyllites, scapolite and tremolite dolomites and schists, are known at and near the base of the Adelaide System rocks adjacent to crystalline older Precambrian basement in the Mt. Painter and Olary districts (Mawson and Dallwitz, 1945; Talbot, 1962). Elsewhere in the geosyncline pelitic rocks have undergone only slight dynamothermal metamorphism with the formation of pencil shales and fissile slates.

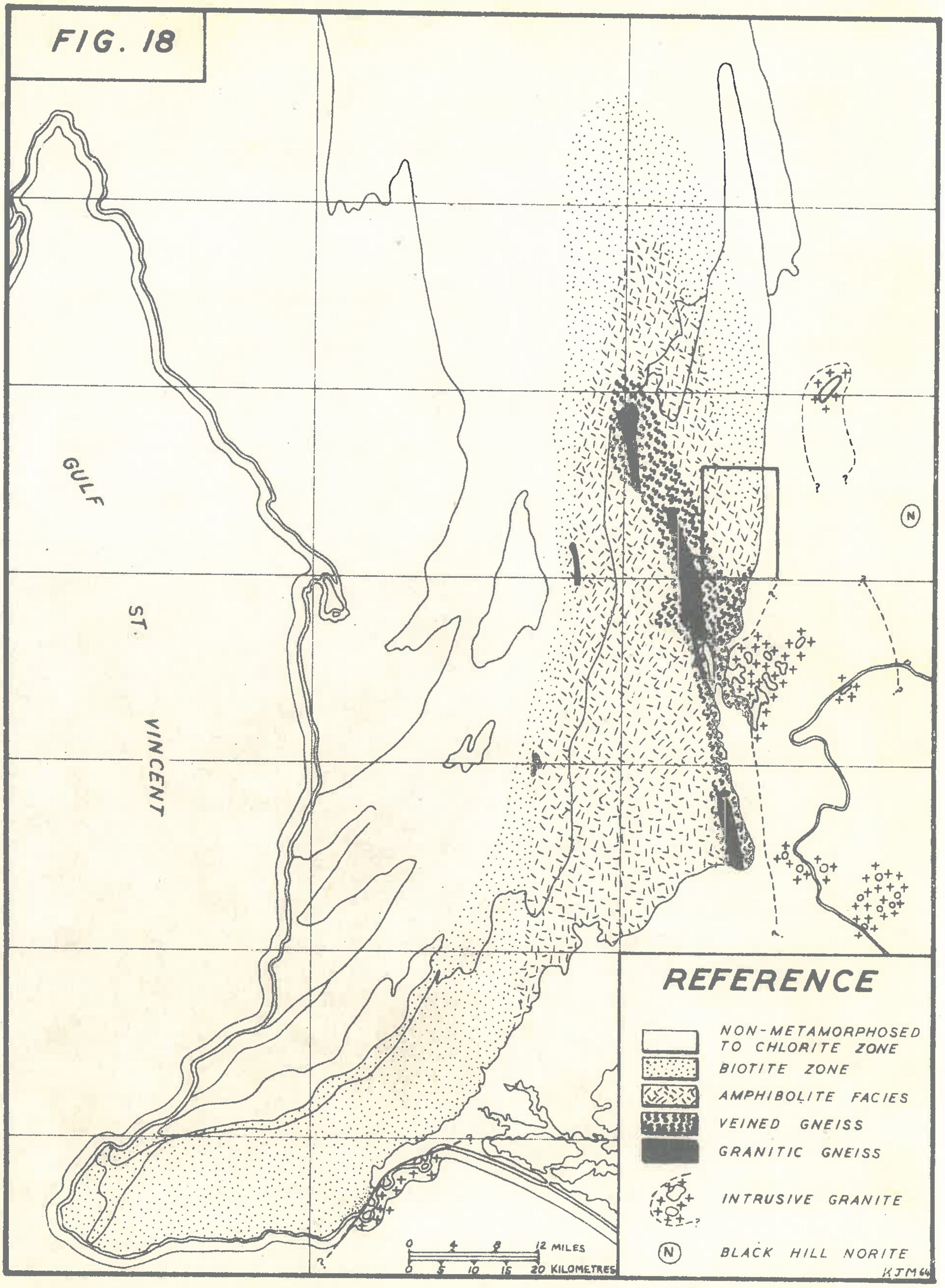
Rocks of the biotite zone and amphibolite facies metamorphism in the eastern Mt. Lofty Ranges occupy an exposed area of about 5000 square miles, and extend below the Tertiary sediments of the Murray Basin south-east of the ranges. Biotite zone rocks outcrop from Kapunda in the north, to Kangaroo Island in the south, a distance of about 200 miles. Fig. 18 indicates areas of biotite zone metamorphism in the Mt. Lofty Ranges. The western portions of the Ranges, including Mawson and Sprigg's (1950) type sections of the Adelaide System near Adelaide, and the Cambrian sequence at Sellick Hill,

FIG. 13.

Generalized map of the Mt. Lofty Range metamorphic belt, showing the location of the area studied in relation to regional metamorphic zonation.

The major geological outlines are taken from Fig. 2.

FIG. 18


GULF
ST.
VINCENT
N
0 4 8 12 MILES
0 5 10 15 20 KILOMETRES
REFERENCE
NON-METAMORPHOSED TO CHLORITE ZONE
BIOTITE ZONE
AMPHIBOLITE FACIES
VEINED GNEISS
GRANITIC GNEISS
INTRUSIVE GRANITE
BLACK HILL NORITE
KJM 64

are relatively unmetamorphosed.

The central portion of the Mt. Lofty Range metamorphic belt is occupied by amphibolite facies metamorphic rocks. These higher grade rocks occupy a north-south elongate area extending from Tanunda in the north, southward through Tungkillo, Palmer, Rockleigh, Rocky Gully, and extend below Tertiary cover east of Strathalbyn (Fig. 18). The core of this belt of high grade metamorphism is occupied by granitized gneisses: the Tanunda Creek gneiss (Hossfeld, 1925; Chinner, 1955), the Rathgen gneiss (Rattigan and Wegener, 1951; White, 1956), the Summerfield gneisses (Sando, 1955; Hopkins 1951) and the Monarto granite gneiss (Kruger and Johns, 1949). These gneissic areas are surrounded by a "zone of veins" consisting of migmatitic gneisses containing both intruded and exuded veins.

The present contribution is a metamorphic and structural study of a small portion, approximately 50 square miles, of the eastern flank of the Mt. Lofty Range metamorphic belt. Within this area a wide range of rock types and metamorphic grades has enabled a detailed reconstruction of the metamorphic character of the eastern flank of this recrystallized belt. Previous detailed investigations of the eastern Mt. Lofty Range metamorphism have been carried out by Chinner (1955) in the Tanunda Creek region, 10 miles to the north-west, and by White (1956) on the Mannum one-mile sheet to the south and south-east of the present area. These studies were concentrated on the amphibolite facies granitized gneisses forming the core of the metamorphic belt. The present investigation examines the metamorphic reconstitution in various rock types under progressive metamorphism from the biotite zone to the migmatite zone adjacent to the granitized gneisses.

In studies of progressive metamorphism it has often been found convenient to map mineral isograds, based on the first appearance of certain index minerals, which give an indication of the grade of metamorphism reached in various portions of an area. This method of mapping metamorphic zones was first used successfully by Barrow (1893), and later Tilley (1925), in the Scottish Highlands. Although this method has met with some justified criticism, mainly from experimentalists in the field of metamorphic petrology (e.g. Yoder, 1952; Greenwood, 1962), it still remains the most satisfactory and convenient means of zonal mapping, especially in preliminary work (Green, 1963; Billings, 1956; Lyons, 1955). The mineral isograd method has been adopted here because it is readily adaptable to both field and laboratory study. Turner's facies classification (Fyfe, Turner and Verhoogen, 1958; Turner and Verhoogen, 1960), although having a stricter theoretical basis than mineral zonation, is not sufficiently general enough to make allowance for regional andalusite-staurolite metamorphism, and it has been found difficult to consider the 'unusual' metamorphic assemblages in the Cambrai area, in terms of the existing facies concepts.

In the metamorphism of the Cambrai area a number of useful mineral isograds have been defined. These are illustrated in Plate 2 where the effects of the post-metamorphic faulting in the central part of the map has been ignored.

It is desirable that isograds should be as free as possible from the effects of intensive variables other than total pressure and temperature. The sillimanite isograd, based on the polymorphic phase transformation andalusite (kyanite) sillimanite, depends only on total pressure and temperature, assuming that foreign ions (e.g. Mn in andalusite) have a negligible effect, and that the kinetics of transition allow the phase change to proceed as soon as the sillimanite field of stability is entered. Isograds dependent only on total P and T are rare. Isograds based on discontinuous reactions releasing one volatile (e.g. water) are subject not only to total P and T, but to the partial pressure of the volatile component (e.g. the first appearance of andalusite or staurolite and the sillimanite-potash felspar isograd). In mapping such isograds it must be assumed either that the partial pressure of water vapour is constant, or varies regularly across the area, as might P and T. Isograds based on discontinuous reactions releasing more than one volatile (e.g. reactions in calcareous rocks), are even more subject to unknown quantities (Greenwood, 1962), although, when used in conjunction with other isograds, they may yield valuable information regarding variations in the fugacities of the volatile components. Isograds based on continuous reactions (e.g. composition of plagioclase in equilibrium with epidote) have been used with success in some metamorphic areas, although, as pointed out by Kretz (1963), the equilibrium reaction should be precisely defined, so that the effects of intensive variables other than total P and T may be determined.

The most useful isograds in defining the metamorphic grade at Cambrai are -

(1) the first appearance of andalusite and staurolite (probably almost coincident)

(2) the first appearance of sillimanite in rocks having an excess of Al.

(3) the sillimanite-potash felspar isograd (based on the decomposition of muscovite).

Using these isograds the metamorphic sequence has been subdivided for subsequent discussions into -

(1) Biotite zone.

(2) Andalusite-staurolite zone.

(3) Sillimanite-muscovite zone.

(4) Sillimanite-potash felspar zone.

Reference is also made to an andalusite-staurolite-kyanite zone forming a narrow wedge at the base of the sillimanite-muscovite zone in the north-west corner of the area, and to a biotite-andalusite zone on the eastern side of the area where andalusite has been observed to appear before staurolite. Although sparse garnets occur in the biotite zone, a separate garnet zone has not been recognized because of the difficulty of defining a garnet isograd.

Variation in the metamorphic character of the eastern and western portions of the area, in rock types where metamorphic reactions are strongly dependent on pressure, has lead to a sub-division of these rocks into <u>eastern</u> and <u>western</u> metamorphic sequences, the division being defined by the prominent post-metamorphic thrust-wrench fault (Plate 3).

CHAPTER 2.

QUARTZO-FELSPATHIC SCHISTS.

Introduction.

The term quartzo-felspathic schist has been used by White (1956) to describe the ubiquitous quartz-felspar rich rocks from the Mannum sheet area, and will be adopted here, with slight modification. Whereas White has used the term to cover both semi-arkosic and semi-pelitic quartz-felspar bearing schists, the present author prefers to define quartzo-felspathic schists as including only those rocks with less than 20% mica minerals. Rocks with 20-50% mica minerals are considered as semi-pelitic schists, and those with more than 50% mica as pelitic. As approximately 70-80% of the rocks outcropping in the present area lie within the quartzo-felspathic schist category, as defined here, it has been found particularly convenient to treat these rocks separately from the semi-pelitic and pelitic schists which are of lesser importance in the metamorphic sequence, and usually contain in addition such minerals as garnet, staurolite or the aluminosilicates.

(a) Appearance and Structure.

The quartzo-felspathic schists are fine-grained light coloured rocks of semi-arkosic appearance in which bedding is well preserved as fine laminations rich in biotite. Even in the highest metamorphic grades these rocks have a grain-size less than 0.5 mm., and the individual quartz and felspar grains are only seen with difficulty with the naked eye. Although the sedimentary grain textures are still well preserved in the lower grade rocks, there are no larger grains of quartz or felspar such as those commonly found in shallow water arkoses. Likewise conglomerate beds are absent, although rare layers of greywacke aspect, containing heterogeneous pebbles have been found. Pure quartzites are absent and quartz-felspar rich rocks lacking mica are rare. The macro-sedimentary structures preserved in these rocks have already been treated in the previous section. The most important sedimentation structure affecting the metamorphism of these rocks is the micaceous lamination. In some rocks these laminations are faint or absent while in others they are extremely well developed. Examination of the lower grade rocks suggests that these laminations are of two kinds. Some laminations are characterized not only by an abundance of biotite, but also heavy mineral concentrates. Other laminations are units of graded bedding; large quartz and felspar grains near the base of the laminae, grade upward into finer quartz and felspar grains with a progressive increase in interstitial mica to the top of the laminae, where there is an enrichment of mica, especially biotite. These mica laminations act as planes of weakness, causing laminated arkoses to have a platy form of outcrop.

A moderate to pronounced schistosity has pervaded all rocks

FIG. 19.

Outcrop of meta-arkose (quartzo-felspathic schist) in the Marne River gorge. Many of these exposures display well preserved coarse scale cross-bedding. Location 200.176.

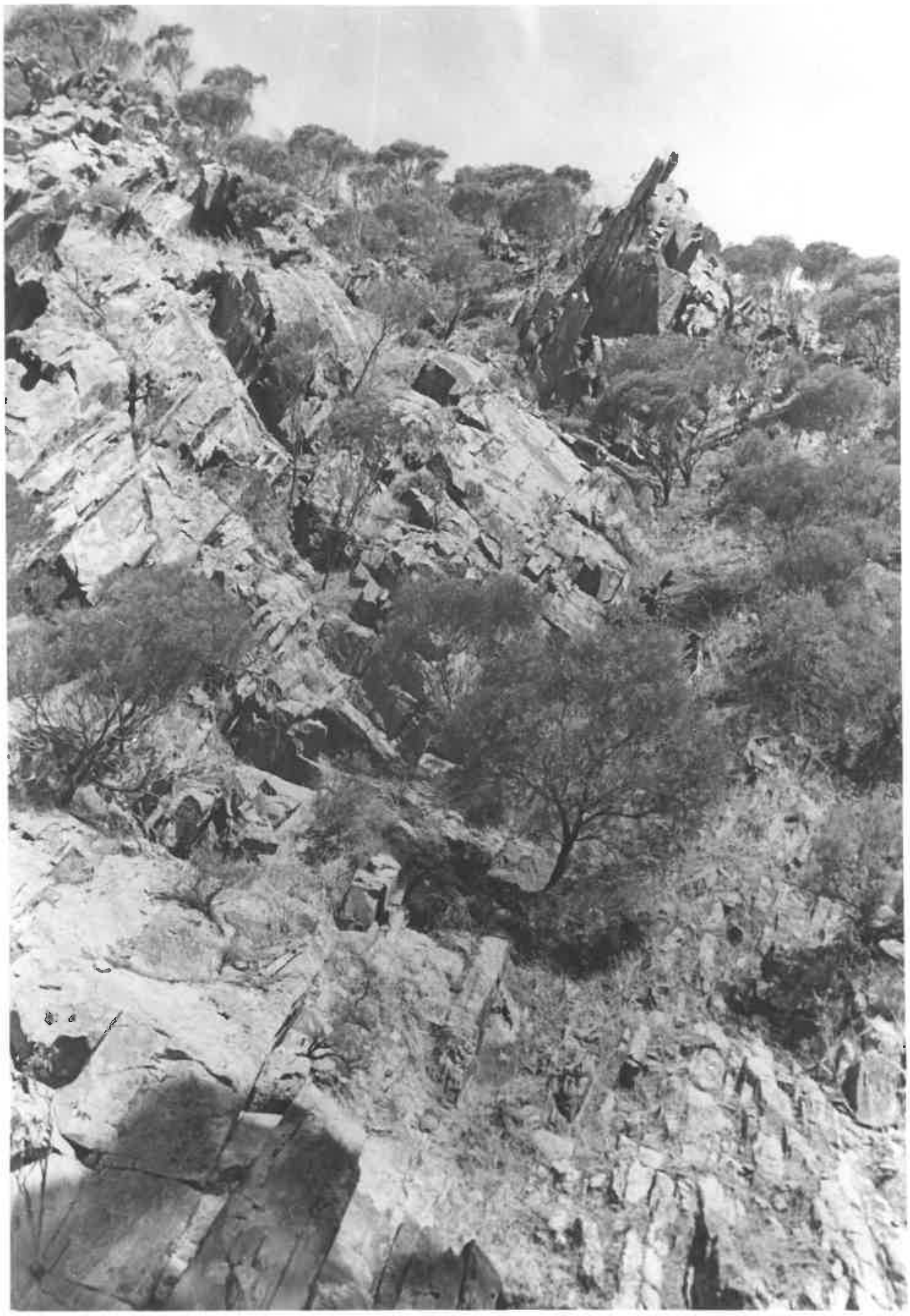


FIG. 19

irrespective of metamorphic grade. The mica elements have been moderately to strongly oriented within this imposed foliation. Where the schistosity is seen to cut the sedimentary laminations at a moderate to high angle, there is little tendency for the mica plates to lie parallel to the laminations. The intersection of the bedding planes and the schistosity has resulted in a pronounced lineation, especially where these planes are at high angles.

The quartz-felspar-biotite granulites of the Moine Series (Read, 1931, p.37) are apparently similar to those quartzo-felspathic schists described here, although the Moine granulites generally contain epidote.

(b) Mineralogy and Mineral Assemblages.

The mineralogy of these rocks is relatively simple. The major minerals are quartz, plagioclase, potash felspar, biotite, muscovite, sillimanite and chlorite, and the lesser constituents zircon, monazite, apatite, iron ore, tourmaline, xenotime and rarely sphene or rutile. The following mineral assemblages have been observed.

(1) Quartz-plagioclase-biotite-muscovite-chlorite
(2) Quartz-plagioclase-potash felspar-biotite-muscovite
(3) Quartz-plagioclase-potash felspar-muscovite
(4) Quartz-plagioclase-biotite-muscovite
(5) Quartz-plagioclase-biotite-muscovite-sillimanite-potash felspar.

Assemblage (1) is confined to biotite zone rocks, assemblage (5) to sillimanite-potash felspar zone rocks.

Thirteen micrometric analyses of quartzo-felspathic schists, as defined above, are given by White (1956). These micrometric analyses indicate the following variation in composition in the quartzo-felspathic schists; quartz 30-50%, plagioclase 25-60%, potash felspar 0-25%, biotite 0-20% and muscovite 0-10%, the major accessories being magnetite and apatite. Fig. 20, determined from White's micrometric data, expresses the composition of the quartzo-felspathic schists in terms of quartz-felspar-others and quartz-plagioclase-potash felspar variation. It is interesting to note the composition range of these rocks, in particular their approximation to the composition of granodiorites (Bowen and Tuttle, 1958). Although no micrometric analyses were made on the quartzo-felspathic schists of the present area, it is apparent from thin section study that these rocks also fall within the composition range determined by White.

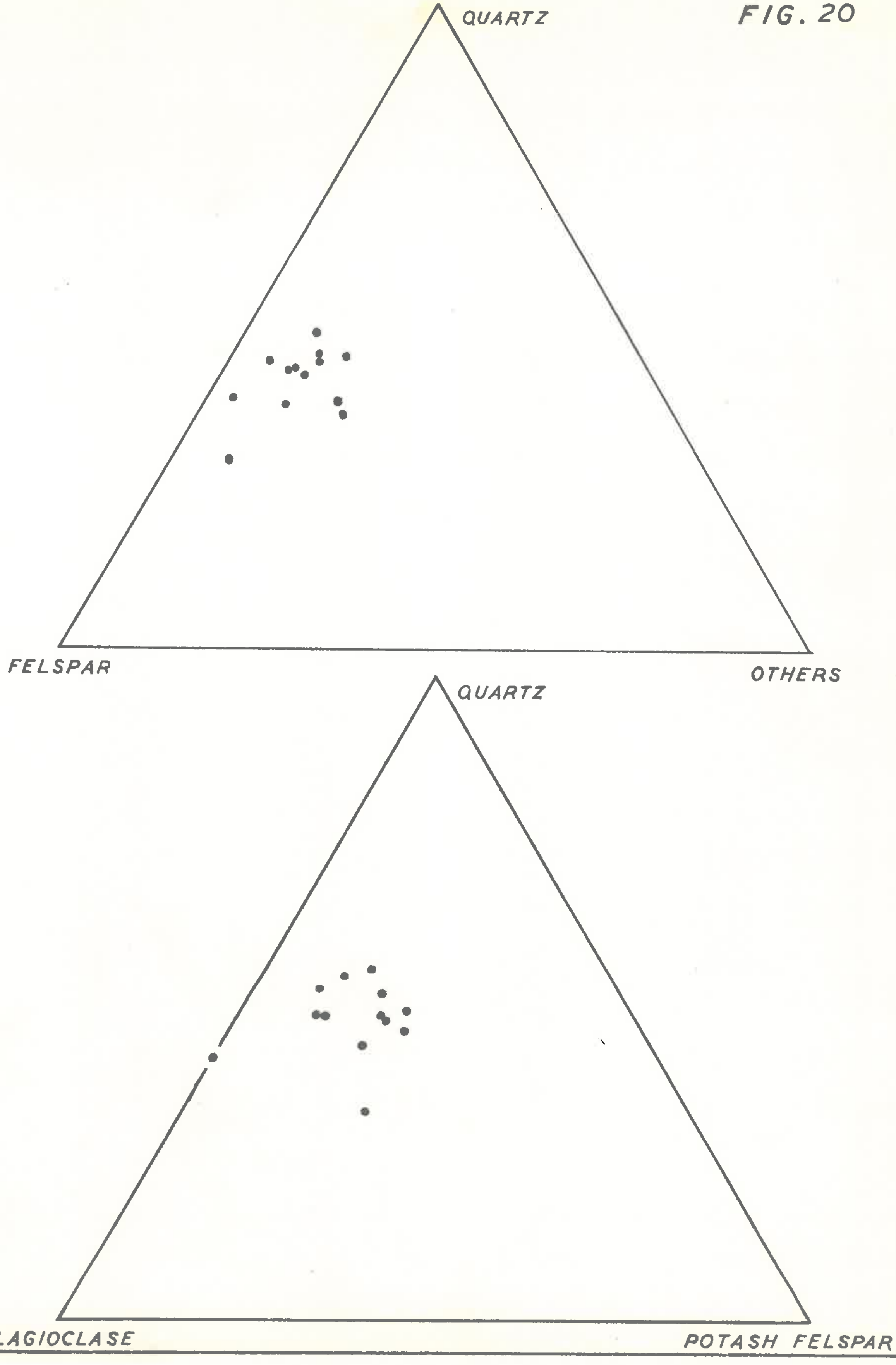
(c) Progressive Metamorphism of Quartzo-felspathic Schists.

Introduction. The changes observed within these rocks with increasing progressive metamorphism is largely textural. The only mineralogical reaction occurs in the highest grade rocks, where muscovite has begun to decompose to sillimanite and potash felspar. The response to metamorphism (Reed, 1962) is very low in these rocks. The original sedimentary felspars

FIG. 20.

Mineralogical composition of quartzo-felspathic schists expressed in terms of Quartz-Felspar-Others and Quartz-Plagioclase-Potash felspar variation (after micrometric analyses of White, 1956).

FIG. 20



are not entirely destroyed until the sillimanite zone is reached. The low metamorphic response in these rocks may be related to a lack of metamorphic dehydration reactions which might supply additional water to the fluid phase to aid re-crystallization.

(1) Biotite zone.

The arkoses of this grade have a clastic texture in which the original matrix between the quartz and felspar grains has recrystallized to micas. The newly formed micas weave and mold around relic clastic quartz and felspar grains in striving for parallelism within the tectonic foliation. The quartz and felspar grains have irregular grain contacts, in particular the felspar in which the degree of irregularity increases as the grains recrystallize to new metamorphic felspar granules, at the old grain boundaries, with increasing metamorphism. In these low grade rocks the quartz and felspar grains range in size to 0.4 mm., while the biotites and muscovites or the matrix between these grains seldom exceed 0.1 mm. This relationship is reversed in the completely recrystallized rocks, where the micas exceed the quartz and felspar in grain size.

The quartz grains are generally clear, xenoblastic to granblastic and weakly strained. The relatively weak strain effects shown by quartz grains, in contrast to the felspars, coupled with the development of planar grain contacts and triple points between adjacent quartz grains, suggests that the original clastic quartz has a greater tendency to polygonize after deformation than to sustain a high strain and then undergo recrystallization (see also Voll, 1960, p.517-519).

Plagioclase, the most abundant felspar, is characterized by a strong show of albite twinning. The grains are undulose and the twins are strongly bent and twisted. Chess-board and disrupted twins are common in the more deformed grains. Each grain is clouded by numerous minute inclusions of sericite, and a fine-grained high relief mineral, probably clinozoizite. The grain rims are clear of inclusions, the thickness of these rims increasing with the metamorphic grade. The composition determined by the five axis method (Emmons, 1959), ranged from An 0-10, with considerable variations from grain to grain. Such variations could be the result of (a) reaction rates being too slow for adjustment to metamorphic equilibrium, allowing some original plagioclase to persist, or, (b) various local equilibria being set up within the scale of a thin section.

As pointed out by White (1956), potash felspar is particularly irregular in its occurrence within the quartzo-felspathic schists, being abundant in some rocks and almost absent in others. The potash felspar grains in the biotite zone rocks are clouded, and show poorly developed cross-hatched twinning. Most of the grains are relic perthites, the included albite blebs making up 50% of some grains. With increase metamorphism the albite migrates to the margins of the grains, and eventually forms separate albite granules.

Biotite and muscovite form a matrix between the relic clastic quartz and felspar grains. The biotite is pleochroic green-black to pale lemon in the lowest grade rocks, the colour approaching mustard brown with increasing grade. The small stumpy biotite flakes have a tendency to clot.

In the lowest grade rocks a green chlorite may be present in association with biotite. This mineral disappears within the upper part of the biotite zone.

Muscovite was present in most of the rocks examined. A few large highly bent flakes are scattered between the quartz and felspar grains in most of the low grade rocks. These, like the felspars, have evidently survived from the original sediment. Their corroded grain edges suggests some disequilibrium with the fine flakes of newly formed biotite and muscovite in which they lie. These relic flakes are quite distinct from the non-strained disoriented metamorphic muscovite xenoblasts found in some of the higher grade quartzo-felspathic schists. Small sericitic muscovite flakes of metamorphic origin are intergrown with biotite in the micaceous matrix of the biotite zone rocks.

(2) Andalusite-staurolite zone.

The quartzo-felspathic schists of this zone have a similar appearance and mineralogy to those of the biotite zone, chlorite now being absent. At this grade remains of the original quartz, felspar and muscovite grains are still present, and in no rocks could the textures be considered indicative of complete metamorphic equilibrium. The degree of recrystallization in these rocks, considering destruction of the original clastic textures, varies from about 30% near the base of this zone, to about 60% near the top. The grain size is variable due to imcomplete recrystallization. Quartz is generally preserved as larger grains than the newly formed granules of recrystallized felspar. Within the andalusite-staurolite zone the biotite increases in grain size until it exceeds that of the quartz and felspar. The relic clastic plagioclase is characterized by an abundance of albite twins, although the small inclusions of sericite and epidote common in the low grade rocks have disappeared. The potash felspar grains still contain some of their original perthitic albite stringers. The large strained relic muscovites are still present in most rocks, while the newly formed metamorphic flakes intergrown with biotite have increased slightly in grain size. The biotite of this zone is notably darker than that of the biotite zone, being pleochroic brown-black to pale lemon. The poor degree of recrystallization in these intermediate grade rocks suggest that diffusion during metamorphism was very limited. These rocks may have been largely closed to H_2O migration during metamorphism (Thompson, 1955).

(3) The sillimanite-muscovite zone.

The quartzo-felspathic rocks have almost lost all traces of their original clastic texture before the upper part of the sillimanite-muscovite zone is reached. Sillimanite does not appear within the quartzo-felspathic schists of the sillimanite-muscovite zone, and muscovite is stable with biotite and potash felspar. The biotite and muscovite have further increased in grain-size, and are now well oriented within the tectonic foliation. The metamorphic quartz and felspar grains show a small, though variable, grain-size (0.02 - 0.05 mm.) within the lower part of this zone, but rapidly develop an even grained texture at higher grades. In the absence of calc-minerals

FIG. 21A.

Photomicrograph of low grade (biotite zone) quartzo-felspathic schist.

Rock A185-5. Location 186.238. Crossed polars.

Slightly undulose quartz, cloudy plagioclase with some simple twinning and rarer potash felspar grains in an interstitial matrix of fine-grained biotite and muscovite. There are rare deformed relic sedimentary muscovite flakes (arrow). The rock has a moderate imposed tectonic foliation (NW to SE).

FIG. 21B.

Photomicrograph of high grade (sillimanite-muscovite zone) quartzo-felspathic schist.

Rock A185-790$_{(1)}$. Location 170.078. Crossed polars.

This rock has suffered complete recrystallization and is approaching an equilibrium metamorphic texture. The slide shows slightly undulose quartz, well twinned plagioclase, and potash felspar with cross-hatched twinning. Biotite and rarer muscovite flakes are weakly oriented.

FIGS. 21A and 21B illustrate the degree of recrystallisation in the metamorphism of arkosic rocks from the biotite zone to the sillimanite-muscovite zone.

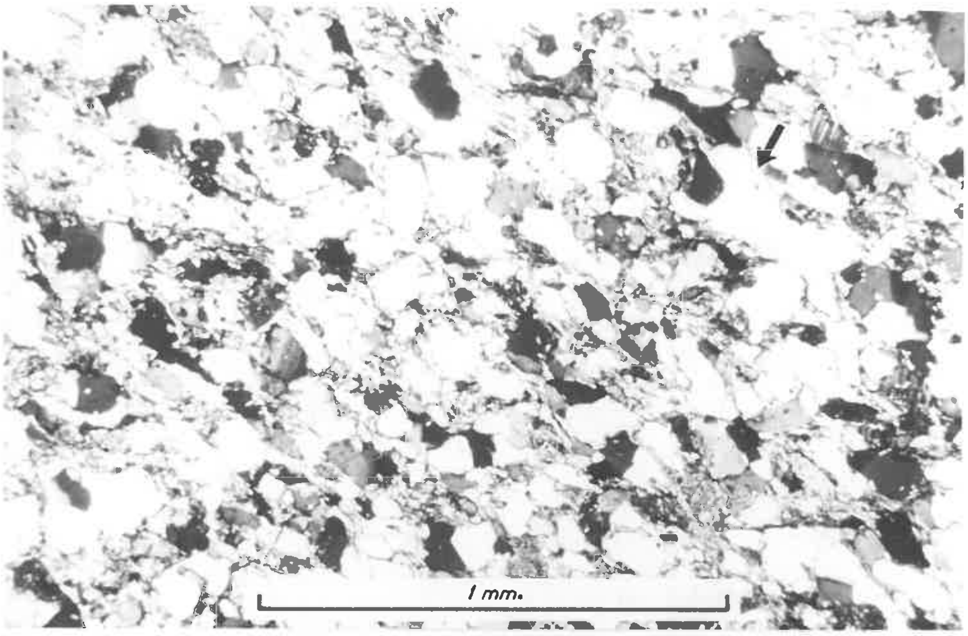


FIG. 21A

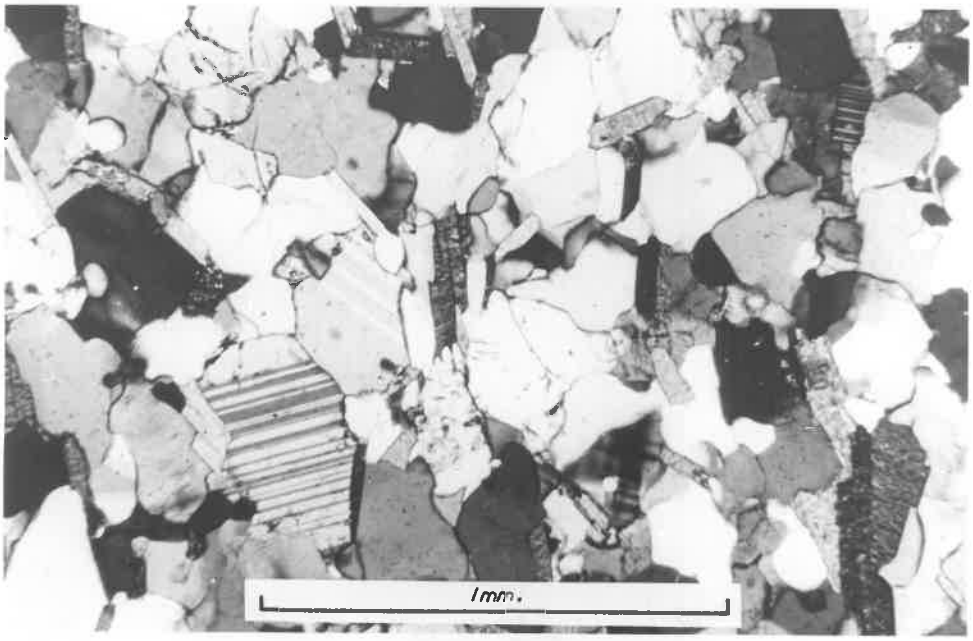


FIG. 21B

the plagioclase composition is near pure albite.

In the upper part of the sillimanite-muscovite zone the quartzo-felspathic schists have an equilibrium crystalloblastic texture. The quartz, potash felspar and plagioclase are even grained at about 0.3 mm. Large 1 mm. biotite and muscovite flakes are well oriented in the schistosity. The potash felspar has taken up some albite molecule which has exsolved on cooling with the formation of a fine hair perthite structure. Cross-hatched twinning is incipiently developed in all grains. The plagioclase has a composition of An 10-12, and twinning is rare and simple. A faint reverse zoning of the plagioclase suggests that albite from the plagioclase grains has been absorbed by the potash felspar at the higher temperatures of the upper part of the sillimanite-muscovite zone.

(4) The sillimanite-potash felspar zone.

Within the sillimanite-potash felspar zone the muscovite begins to decompose to sillimanite and potash felspar (the second sillimanite isograd of Heald, 1950). As pointed out by Guidotti (1963) the decomposition of muscovite is probably not discontinuous, but may involve an equilibrium reaction between muscovite, potash felspar, sillimanite and plagioclase. The reaction may be written

[quartz + muscovite + plagioclase $\rightleftharpoons$ sillimanite + soda rich potash felspar + more calcic plagioclase + water.]

This reaction depends not only on total pressure and temperature, but also the activity of water vapour and the anorthite content of the plagioclase. Guidotti has devised the above reaction to explain the perthitic component of the potash felspar. It must be pointed out, however, that not all potash felspar is formed through the decomposition of muscovite, and original potash felspar may absorb albite from the plagioclase prior to the decomposition of muscovite, as in the present area.

Near the Kanappa Mine both biotite and muscovite are present as large flakes within the quartzo-felspathic schists. Sillimanite fibres are present in most muscovite flakes and a few flakes have been completely replaced by fibrolite. Sillimanite does not replace biotite in these rocks. This observation contrasts with the commonly observed replacement of biotite by sillimanite, arising from the decomposition of former aluminosilicates, within the more aluminous pelitic schists. Chinner (1961) has considered that the apparent replacement of biotite by sillimanite in such rocks may be a result of nucleation of newly formed sillimanite within a structurally favourable mica host. The stability of biotite depends strongly on its composition, iron rich varieties decomposing before magnesium rich ones (Yoder and Eugster, 1954; Eugster and Wones, 1958). In the Cambrai quartzo-felspathic schists no evidence has been found for the decomposition of biotite prior to muscovite.

In the Kanappa Mine region a reversal of the reaction bringing about a decomposition of muscovite seems to have occurred at a late stage in the metamorphism, for a small amount of secondary muscovite has epitaxically replaced the rims of some biotite flakes.

The potash felspar in these rocks has well developed hair perthite structure but no distinct cross-hatched twinning. A wavy extinction in many grains, especially about quartz inclusions, suggests that submicroscopic twinning may be present. Plagioclase has a composition near An 12 and is rarely twinned.

Muscovite will commence to decompose much earlier in a "dry" metamorphic environment, where the activity of water is low. The well developed sillimanite-potash felspar zone south of the Kanappa Mine suggests that the activity of water in metamorphism was lower there than in the south-west corner of the area, where the decomposition of muscovite is only rarely observed even in rocks which are decidedly migmatitic. The wetter conditions of metamorphism indicated by the higher grade western rocks, is also supported by textural and mineralogical evidence from the pelitic and calc-siliceous rocks.

(5) Migmatites.

In the south-west corner of the area the quartzo-felspathic schists take on a migmatitic aspect (the "zone of veins" of White, 1956). The original biotite laminae have become accentuated while the quartz-felspar rich layers increase notably in grain-size, and commence to segregate into veins. These veins, at first, lie strictly parallel to the bedding laminations, but with increased mobilization they may become discordant. In the most highly migmatized rocks the biotite is segregated into irregular shlieren bordering broad zones of quartzo-felspathic material. These rocks have an appearance which suggests that they had become partially molten. The experimental work of Winkler and von Platen (1961a and b) suggests that migmatites are a result of partial liquification of the quartz-felspar fraction of the rock (see also Holmquist, 1920; Sudovikov, 1955). In the present area partial melting may have been aided by the late synmetamorphic intrusion of numerous small bodies of granodiorite accompanied by veins of aplite.

Muscovite occurs in most of the migmatitic rocks in the south-west corner of the mapped area. In some muscovite flakes fibres of sillimanite have appeared, suggesting that the sillimanite-potash felspar zone has been barely reached here. The absence of a sillimanite-potash felspar zone under conditions of high water vapour activity is readily seen from Fig. 22. The experimental curves for the stability of muscovite in the present of quartz at $P_t = P_{H_2O}$ (Segnit and Kennedy, 1961), and for the minimum melting relations of wet granites (Tuttle and Bowen, 1958) cross each other at about 2000 bars and 700°C. Thus, under maximum a_{H_2O} conditions a metamorphic rock with a granitic composition in the sillimanite zone would be largely molten before muscovite could decompose to sillimanite and potash felspar in the presence of quartz. Under conditions of $P_{H_2O} < P_t$ the muscovite + quartz stability curve becomes a divariant band opening out towards lower temperatures with decreasing partial pressure of water vapour (Yoder, 1955). Once muscovite enters the P-T region covered by this band it begins to decompose, and continues to disintegrate with increasing temperature, until the curve for

FIG. 22

Stability relationships of muscovite. Curves 1 and 3 for the reaction [quartz + muscovite $\rightleftharpoons$ potash felspar + sillimanite + H_2O.]

Curve 1: Curve for the commencement of muscovite decomposition at $P_{H_2O} < P_{total}$.

Curve 2: Minimum melting curve for wet granites (Tuttle and Bowen, 1958).

Curve 3: Stability curve for muscovite in the presence of quartz at $P_{H_2O} = P_{total}$ (Segnit and Kennedy, 1961).

Area between curves 1 and 3 represents a divariant band in the P-T field over which muscovite decomposes at $P_{H_2O} < P_{total}$.

Shaded area: P-T range over which muscovite decomposes before the commencement of melting in rocks of granitic composition.

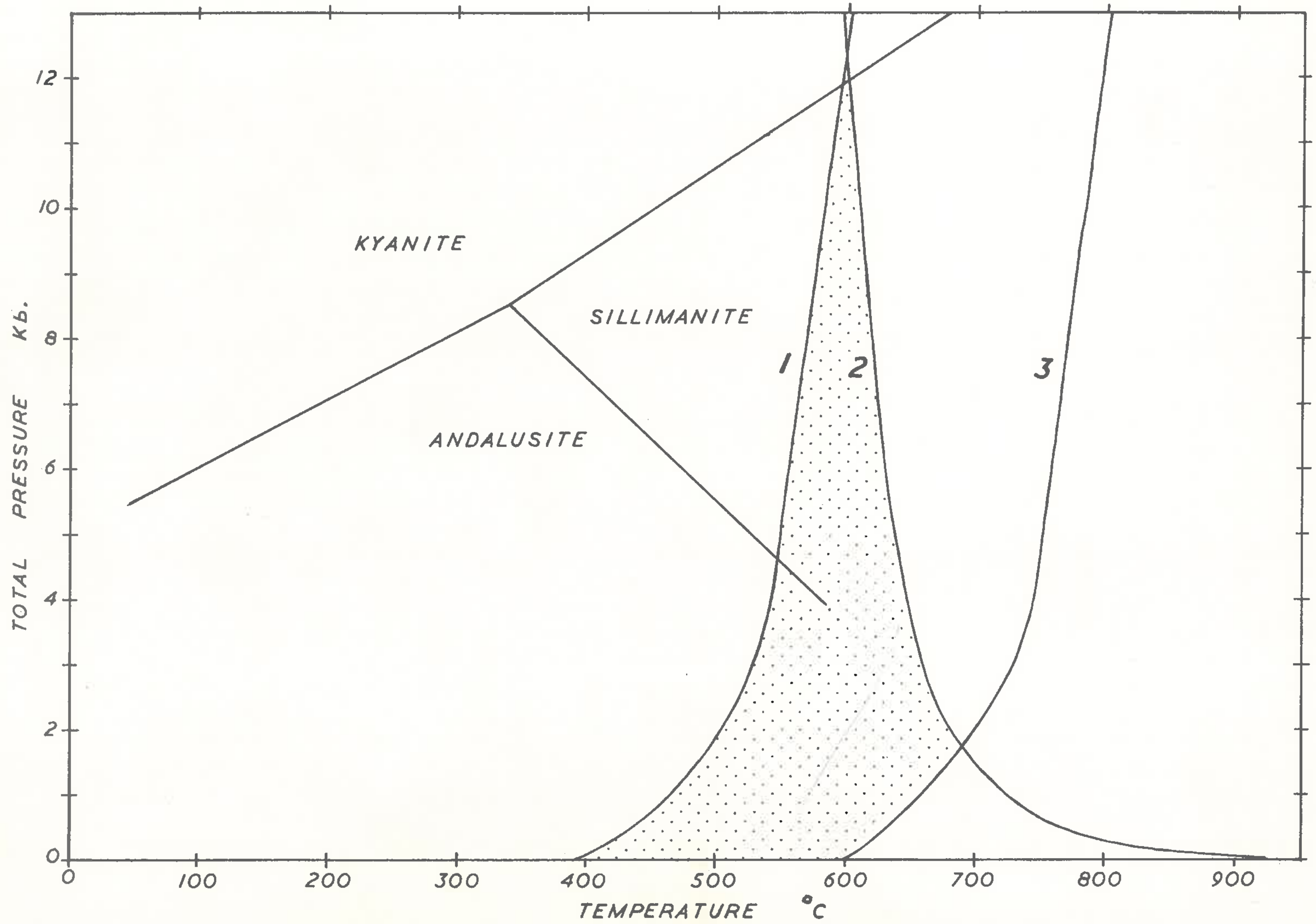

KYANITE
SILLIMANITE
ANDALUSITE
1
2
3
TOTAL PRESSURE Kb.
0
2
4
6
8
10
12
0
100
200
300
400
500
600
700
800
900
TEMPERATURE °C

FIG. 22

muscovite + quartz stability for maximum P_{H_2O} is reached, where the muscovite disappears entirely. Within the P-T field indicated by the band the assemblage muscovite-quartz-potash felspar-sillimanite is stable. The width of the band gives some quantitative estimate of the degree of water deficiency. Under water deficient conditions the curve of minimum melting of a rock remains fixed, but the proportion of liquid to crystals at any temperature above this curve is considerably decreased (Yoder, Stewart and Smith, 1957), and migmatites may not become prominently developed until later in the metamorphic sequence (see also discussion by Schreyer, Kullerud and Ramdohr, 1964). At a sufficiently low activity of water vapour a P-T region becomes available in which muscovite begins to decompose before the rock commences to melt. A zone of non-migmatitic rocks in which sillimanite and potash felspar are stable is then possible. In the Kanappa Mine region sillimanite and potash felspar are stable in non-migmatized rocks. However, south of the Kanappa Mine the rocks eventually become migmatitic and presumably the curve for minimum melting of granite has been passed. The formation of some sillimanite fibres in muscovite within the highest grade migmatites of the south-west corner of the area indicates that under higher P_{H_2O} conditions the sillimanite-potash felspar zone may have just been reached, but within the P-T field of partial melting.

(6) Abnormal quartzo-felspathic schists.

White (1956) has described potash deficient schists, which appear to have replaced normal potash bearing quartzo-felspathic schists, in the Rockleigh area, south east of Palmer. He concluded that these rocks, with their high soda content, have probably suffered metasomatism involving a removal of potash. Some of the Rockleigh schists contain anthophyllite, while biotite is rather rare. Similar rocks have been described by Chinner (1955) from the Tanunda Creek area, and by Kruger and Johns (1949) from Rocky Gully.

Similar rocks occur in the present area. Here they replace the normal meta-arkosic quartzo-felspathic schists on the western margin of the mapped area, in the lower portion of the sillimanite-muscovite zone.

These rocks are light coloured fine even grained, usually non-laminated, quartzo-felspathic schists, having a moderate to strong foliation outlined by muscovite or chlorite orientation. Like the normal quartzo-felpathic schists near the base of the sillimanite-muscovite zone, the quartz-felspar matrix has not reached an equilibrium metamorphic texture,

FIG. 23A.

Photomicrograph of potash deficient schist (sillimanite-muscovite zone).

Rock A185-715. Location 151.199. Crossed polars.

Well oriented texture outlined by muscovite and chlorite flakes and shape orientation of quartz and felspar grains. Matrix of quartz and twinned plagioclase has recrystallized but has irregular grain-size.

FIG. 23B.

Photomicrograph of heavy mineral bands in arkose.

Rock A-185-366. Location 188.156. Ordinary light.

Layers of opaque iron ores with accessory zircon, monazite and apatite associated with layers enriched in biotite. Matrix consists of quartz and felspar.

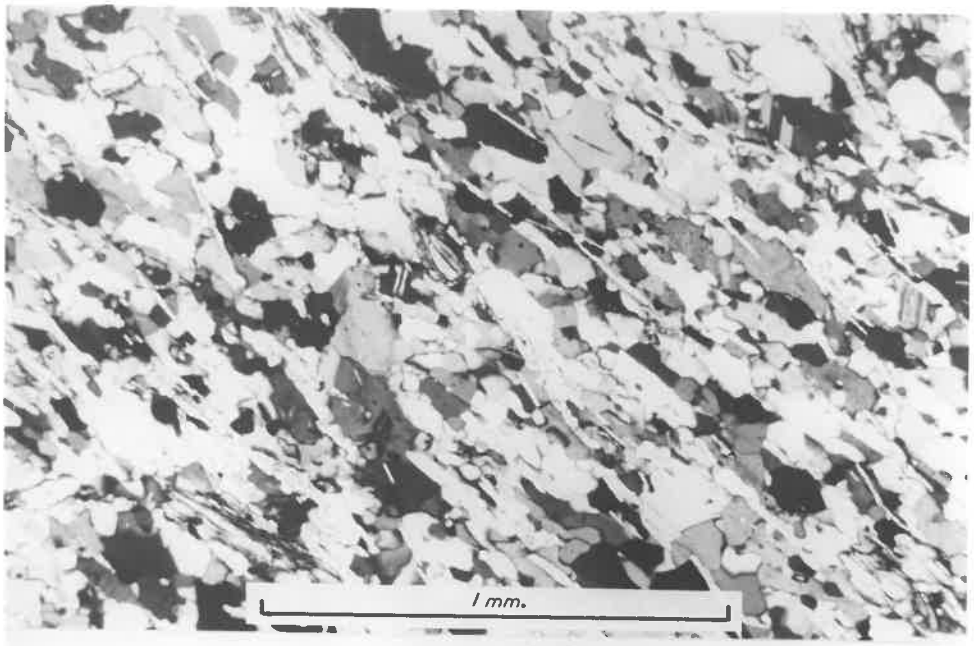


FIG. 23A

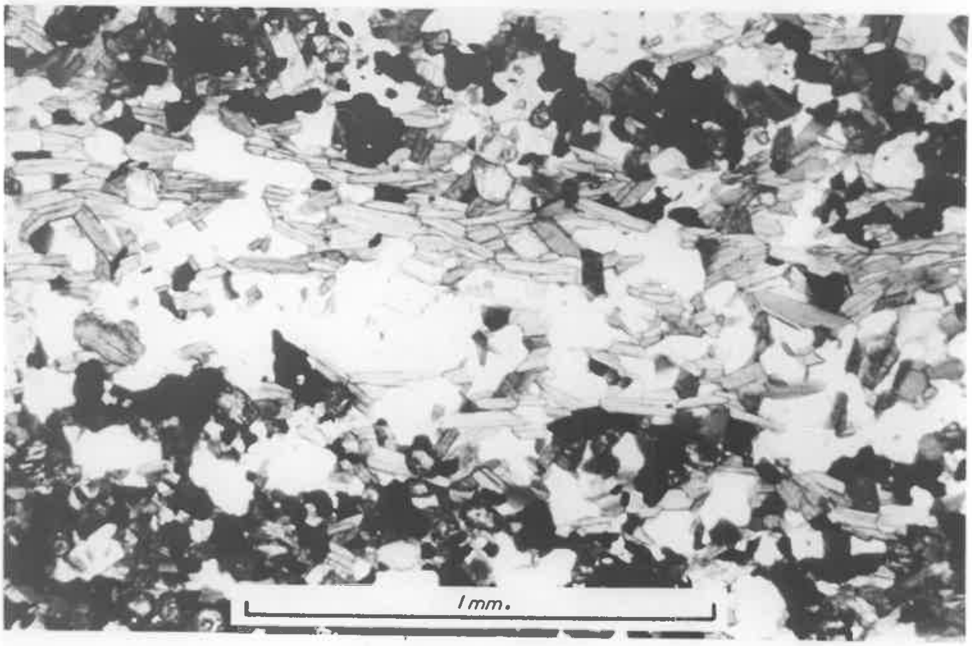


FIG. 23B

and some relic clastic plagioclase grains are still present. In all rocks examined the plagioclase had a composition near pure albite. Biotite is usually absent, although a few pale-coloured flakes are present in some rocks. In place of biotite most rocks have large moderately oriented chlorite flakes which tend to aggregate into clumps. The chlorite has a birefringence near 0.006 and is optically positive. Thin oriented muscovite flakes are not uncommon in some rocks. Anthophyllite, altering to chlorite, may also occur, taking the place of chlorite or biotite. These rocks have a very low potash content, and with the stratigraphic evidence indicating that they have replaced the normal biotite-potash felspar bearing arkoses, it must be concluded that potash has, by some means, been removed during metamorphism.

While apatite and zircon are common accessories, tourmaline is very rare. Ilmenite and pyrite have been observed in some rocks but in others red rutile apparently takes the place of ilmenite. Rutile has not been found among the accessory minerals of the normal meta-arkoses. It must be assumed, in some cases at least, that iron oxide has also been effectively removed; magnesium and titanium, in chlorite, anthophyllite and rutile, being left behind.

The genesis of these rocks remains a mystery, although it is evident that allochemical metamorphism has played some part in their formation. Further considerations on the genesis of these rocks will be outlined in Chapter 10.

CHAPTER 3.

SEMI-PELITIC AND PELITIC SCHISTS.

The semi-pelitic and pelitic schists of the Cambrai area may be conveniently sub-divided into three separate groups based on their mineralogy and bulk chemistry. Pelitic rocks maintaining an excess of alumina, and distinguished by the presence of the aluminosilicate minerals staurolite, andalusite, kyanite or sillimanite, are described in Chapter 4. Schists high in alumina and magnesia, and characterized by the occurrence of chlorite, cordierite and anthophyllite, and treated separately in Chapter 5. The remaining micaceous and garnetiferous pelitic schists are considered in the present chapter.

(a) Minerals and Mineral Assemblages.

The following major mineral assemblages have been observed.

Assemblages below the sillimanite-potash felspar zone.

(1) Quartz-plagioclase-biotite-muscovite-chlorite (lowest grades only)

(2) Quartz-biotite-graphite-potash felspar (rare)

(3) Quartz-plagioclase-biotite-muscovite-potash felspar (and all quartz-biotite bearing sub-assemblages)

(4) Quartz-plagioclase-biotite-muscovite-garnet (and all quartz-biotite bearing sub-assemblages)

(5) Quartz-plagioclase-biotite-muscovite-potash felspar spessartite

Assemblages transitional to sillimanite-potash felspar zone.

(6) Quartz-plagioclase-biotite-muscovite-sillimanite-potash felspar (and quartz-biotite bearing sub-assemblages)

(7) Quartz-biotite-sillimanite-garnet

Assemblages within the sillimanite-potash felspar zone

(8) Quartz-plagioclase-biotite-potash felspar-sillimanite-garnet (and quartz-biotite bearing sub-assemblages).

(b) Micaceous pelitic schists.

(1) Biotite zone.

The lowest grade pelitic rocks in the present area are grey micaceous phyllites. These rocks have been completely reconstituted in metamorphism, and consist of well oriented (0.1 mm.) flakes of pale brown biotite, muscovite and lesser pale green chlorite, embedded in a fine-grained (0.05 mm.) quartz-plagioclase matrix. Bedding lamination, if originally present, has been totally destroyed by the superposition of the strong tectonic foliation.

The semi-pelitic schists show features intermediate between those of the predominantly clastic arkoses and the completely recrystallized phyllites. In some rocks clear strained relic sedimentary quartz grains are set in an interstitial metamorphic matrix of biotite, plagioclase and potash felspar. The potash felspar, readily identified by its tartan twinning, has a tendency to develop poikiloblastic crystals which enclose the stumpy biotite flakes and small twinned plagioclase grains of the matrix.

(2) Andalusite-staurolite zone.

Through this zone the grain size of the pelitic schists is progressively enlarged. Chlorite is no longer a stable component. The intergrown biotite and muscovite micas, while still maintaining an orientation within the schistosity, tend to become more clumped together as the finer interstitial quartz and plagioclase grains disappear with the enlargement of adjacent grains favoured for survival.

(3) Sillimanite-muscovite zone.

Before this zone is reached the semi-pelitic rocks have undergone complete recrystallization to an even crystalloblastic texture, although sedimentation structures, such as micro-cross-bedding and biotite lamination may be well preserved. Well oriented orange-brown biotite, with prominent black pleochroic halos, is set in an even grained (0.1 mm.) granoblastic quartz-felspar matrix. The metamorphic plagioclase is rarely twinned. The smooth equilibrium contacts between the quartz and felspar grains is largely responsible for the sandy nature of the weathered rocks.

Within the sillimanite zone the grain size of the more pelitic schists continues to increase. In the mica rich schists the intimately intergrown biotite and muscovite flakes become more elongate and grouped together, while the quartz and plagioclase grains have increased to 0.3 mm. Muscovite is stable through the sillimanite-muscovite zone and sillimanite is absent.

The semi-pelitic rocks generally show a greater abundance of biotite than muscovite. Some biotites have a dark green-black to pale yellow pleochroism while others have a deep red Z-axial colour. The muscovite, occurring as large, clear, somewhat stumpy flakes, is widely dispersed. Within this zone the grain size of the pelitic schists (0.3 mm.) exceeds that of the more quartzo-felspathic semi-pelitic schists (0.2 mm.).

(4) Sillimanite-potash felspar zone.

The appearance of thin sillimanite fibres within the muscovite plates marks the commencement of muscovite break-down and the entrance of the sillimanite-potash felspar zone. The decomposition of muscovite is rather erratic in its occurrence; muscovites in some parts of a thin slice are more strongly attacked than others suggesting local disequilibrium on the scale of a thin section (see also Phinney, 1963). A similar irregularity in the fibrolitization of mica minerals has often been noted (e.g. Chinner, 1961). As the muscovite becomes replaced by fibrous sillimanite, spears of sillimanite penetrate into adjacent quartz, plagioclase and biotite. However, there is no indication of the instability of adjacent biotite, or of nucleation of sillimanite within this mica.

In the high grade pelitic schists of the sillimanite-potash felspar zone south of the Kanappa Mine, lens shaped eyes of fibrous sillimanite oriented within the strong rock schistosity, seem to have replaced former muscovite megacrysts. This fibrolite has been strongly replaced at a later stage by skeletal muscovite.

In the migmatite zone in the south-west corner of the area the pelitic gneisses have begun to develop a more foliate structure, suggestive of partial melting and the separation of the micaceous elements into schlieren. Large muscovites, 1-3 mm. in length, either intergrown with biotite or having a stumpy disoriented form, contain a few sillimanite fibres indicative of the commencement of muscovite decomposition. As with the quartzo-felspathic schists, muscovite appears to be stable within the south-western migmatite zone, while undergoing complete decomposition a little prior to the development of really migmatitic rocks in the south-eastern portion of the area. This again suggests slightly higher P_{H_2O} conditions in the western rocks during metamorphism, thus enabling the preservation of muscovite to higher temperature conditions above the minimum melting curve of granites (Bowen and Tuttle, 1958; Fyfe, Turner and Verhoogen, 1958).

(c) Porphyroblastic Mica Schists.

Pelitic and semi-pelitic schists containing large silvery muscovite porphyroblasts, generally disoriented and transecting the rock schistosity in diverse directions, are well displayed in parts of the eastern Mt. Lofty Ranges. In the present area these rocks have a wide distribution, but are particularly common within the pelitic and aluminous pelitic schists. The muscovite porphyroblasts are usually large (1-3 mm.) randomly-oriented parallel-sided crystals. Although containing some quartz and plagioclase inclusions, much of the quartz and plagioclase, and all of the biotite of the original matrix now occupied by the porphyroblast, has been replaced. Only the larger flakes show slight forcing aside of the matrix in a manner similar to that produced experimentally by Schuiling and Wensink (1962) during the static growth of copper sulphate in sand. This porphyroblastic muscovite is considered to be of metasomatic origin and will be considered further in Chapter 10.

(d) Graphitic Schists.

Graphitic schists occur within the low grade pelitic sequence below the marble horizon several miles north of Pine Hut. In the present area few graphitic schists have been found. The rarity of graphitic schists in the Cambrai area may be a result of slight sedimentary facies changes, or to oxidation of the graphite in metamorphism. A small area of unusually fine grained grey rocks at 230.187 contain graphite. These rocks are composed predominantly of fine-grained (0.015 mm.) potash felspar, with lesser quartz and stumpy grains of orange-brown biotite. Fine dust-like graphite occurs as inclusions within the potash felspar grains and is particularly abundant at the grain contacts.

(e) Garnetiferous Schists.

Garnet appears to be stable in rocks of widely varying bulk composition. Minute garnets are found scattered through low and intermediate grade pelitic schists, and are particularly common in the more aluminous pelitic schists to be discussed in the following chapter.

(1) Almandine-mica schists.

It has not been found possible to map a garnet isograd in this area, although garnet bearing pelitic rocks are not uncommonly found in the biotite zone below the isograd marking the first appearance of andalusite and staurolite in the aluminous pelitic schists. Garnets occurring in the biotite zone rocks are rare, scattered and of minute size, often less than 0.2 mm. in diameter. These garnets show a sub-idioblastic to idioblastic form, and have minute quartz inclusions near their cores. The biotite and muscovite flakes immediately adjacent to the garnet crystals show little or no deflection from their well oriented course, suggesting that at least the greater part of the garnet crystals were formed after the orientation of the mica foliae. In the semi-pelitic schists garnets generally have a more irregular xeno-blastic crystal form, and are confined to the biotite rich laminations. A closer examination of the low grade phyllitic schists over a wider area may eventually lead to the location of a garnet isograd.

Within the sillimanite zone medium-grained quartz-biotite-muscovite schists may contain up to 20% garnet as crystals to 1 mm. across. Some of these garnets have a textural relationship with their enclosing matrix suggestive of post-tectonic replacement, while others show distinctive internal sigmoidal rotation features, signifying growth syntectonic to the development of the slaty cleavage, and post-tectonic outer idiomorphic rims. Small quartz and iron ore inclusions are common. Within the sillimanite-potash felspar zone the assemblage quartz-biotite-sillimanite-potash felspar-almandine has been observed.

Although garnet is plentiful in the high grade pelitic schists and gneisses of the Kanappa Mine region, only one garnet locality (152.116) is known from the migmatite zone in the south-west corner of the area. Here a small outcrop contains numerous large 5-7 cm. garnet dodecahedra embedded in a biotite rich matrix. Read (1931, p.111) has described similar sparse occurrences of large garnets in the inner part of the "zone of veins" in the Loch Choire injection complex of Central Sutherland.

(2) Spessartite-mica schists.

Just below and within the sillimanite-muscovite zone on the western side of the area a number of localities of spessartite schists have been observed (e.g. 164.155, 157.213). These rocks are chiefly recognized in the field by a distinctive black manganese oxide coating on joint surfaces, and the presence within the rocks of layers of fine-grained brownish-pink garnets. These layers, which are of variable thickness up to 1 cm., and may contain up to 80% garnet, are sometimes strongly folded and deformed. The spessartite crystals are usually very small, seldom greater than 0.1 mm. even in the sillimanite zone, and contain very minute inclusions indicative of early growth in progressive metamorphism (Tilley, 1923, Miyashiro, 1953). The matrix may contain quartz, plagioclase, biotite and potash felspar. In some rocks spessartite is stable with potash felspar below the sillimanite isograd. This felspar forms abundant granules in the matrix, and has incipient microcline twinning. Muscovite may occur either as oriented flakes intergrown with the biotite or as stumpy cross-cutting porphyroblasts. The biotite usually has a dark sepia to black Z-axial colour.

Spessartite garnets in metamorphic rocks are usually very fine-grained, even when the rocks are of high metamorphic grade (e.g. the spessartite sand-stones of Broken Hill, Stillwell, 1953, p.617), although in pegmatites spessartites may grow to a large size. The small size of manganese garnets in comparison to adjacent almandine garnets has been commented upon by Chinner (1960). Chinner suggests that the spessartites may nucleate more readily, thus preserving a fine grain-size. An alternative possibility is that the spessartite garnets have grown at a much earlier stage of the metamorphism, indicated by the much finer internal inclusions, and that once formed, there may be no tendency for the crystals to further enlarge at any greater rate than the recrystallization of the rock matrix as a whole. Almandines, however, growing at a later stage of metamorphism under

FIG. 24A.

Photomicrograph of spessartite-mica schist.

Rock A185-361. Location 164.155. Ordinary light.

Portion of sharply folded layer of fine spessartite crystals intermingled with biotite and rarer muscovite in quartz-felspar matrix. Note very fine inclusions in spessartite.

FIG. 24B.

Photomicrograph of porphyroblastic felspar schist, Kanappa Mine.

Rock A185-428b. Location 216.135. Ordinary light.

Porphyroblasts of plagioclase (clear with rare simple twinning) and potash felspar (with fine hair perthite) containing small oriented inclusions of biotite and rarer muscovite. Between the porphyroblasts the biotite is strongly clustered.

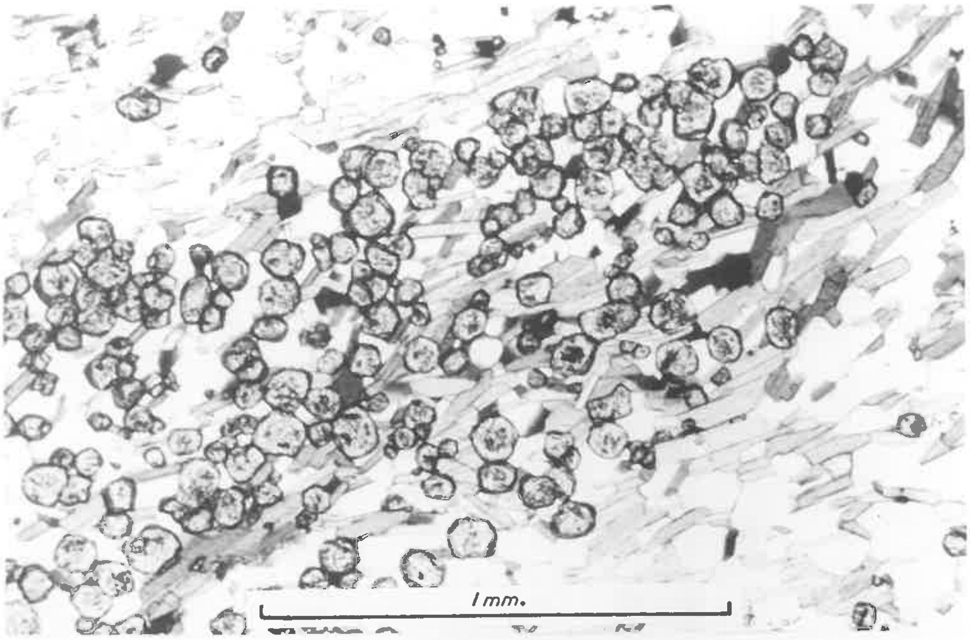


FIG. 24A

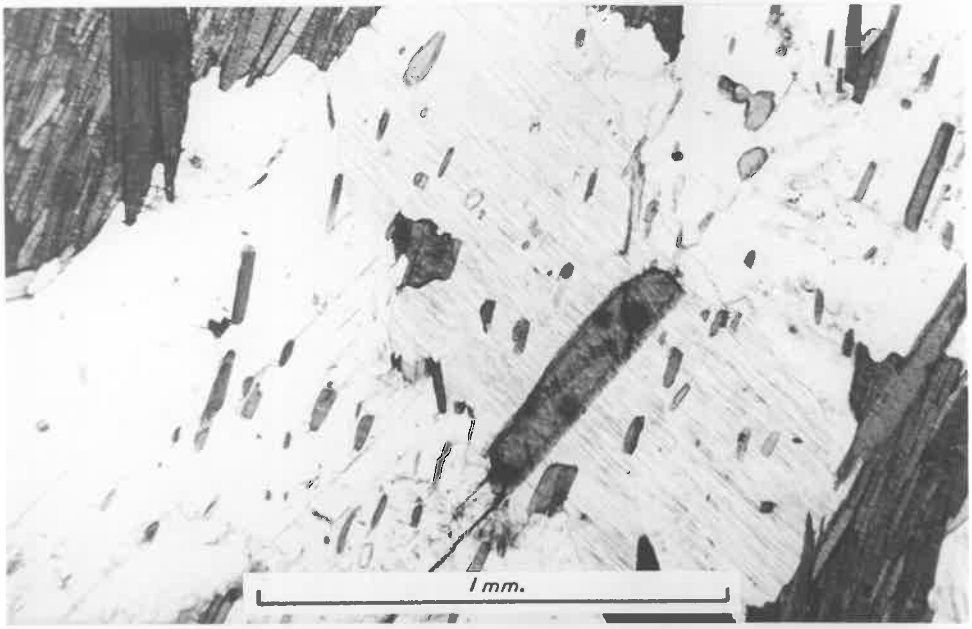


FIG. 24B

conditions of more extensive migration, may have opportunity to grow much larger porphyroblasts.

(f) Porphyroblastic Felspar Schists of the Kanappa Mine Region.

Xenoblasts of potash felspar and albite-oligoclase are common in certain of the more pelitic rocks of the Kanappa Mine region. Individual porphyroblasts seldom exceed 5 mm. in diameter. These felspars are often grouped into layers parallel to the foliation, and are surrounded by an accumulation of 'rejected' matrix components, especially biotite. There is little indication that these rocks have formed through the metasomatic introduction of alkalies since optical examination has indicated that non-porphyroblastic rocks of similar composition may occur in adjacent layers. Even on the scale of a single hand specimen a rapid, but gradational, change from normal non-porphyroblastic schists to a coarsely porphyroblastic rock may be observed. Similar variations are commonly encountered in the field across strike. Zones of porphyroblastic rocks may be traced for many hundreds of yards along the trend of the schistosity. These narrow zones appear to be more closely related to the tectonic foliation than to the original bedding. Grain size variations across zones of porphyroblastic schists suggests that intense local grain growth has occurred within the zones. Variations in the activity of water vapour, perhaps caused by local influxes of aqueous fluids soaking along the tectonic foliation, might account for the observed effects. The growth of the felspar porphyroblasts does not seem to have affected the metamorphic status of these rocks, for north of the Kanappa Mine muscovite is stable within both normal and porphyroblastic schists, while further south, on entering the sillimanite-potash felspar zone, muscovite has commenced to decompose in both rocks.

Both potash felspar and plagioclase porphyroblasts may occur together in these rocks, although potash felspar is by far the more abundant. Some porphyroblastic rocks contain up to 80% potash felspar porphyroblasts. The potash felspar xenoblasts are grey coloured in hand specimen, and are filled with small rounded quartz and plagioclase and oriented biotite and muscovite inclusions. The continuity of the oriented micas inside and outside the porphyroblasts is indicative of static post-tectonic crystal growth. In a few rocks a later deformation, causing broad crenulations within the micaceous rock matrix, has bodily rotated the porphyroblasts up to 30°. Evidently this phase of movement occurred after the main peak of metamorphic recrystallization and grain growth, but before the growth of secondary muscovite occurring as a replacement of sillimanite. Exsolution features are commonly observed within the potash felspar, although the amount of albite exsolved is small. Exsolution has resulted in the formation of hair perthites, vein perthite being very rare. These hair perthites consist of fine closely spaced parallel hairs of albite aligned along the a-axis of the potash felspar lattice, corresponding to the intersection of the $\{010\}$ and $\{001\}$ cleavage planes. A perthitic felspar crystal with hair like exsolution, similar to that observed here has recently been figured by Scharbert (1964, Pl. 6, Fig. B). The Cambrai potash felspar porphyroblasts have 2Vx values ranging from 65-72^{o}, microcline twinning being incipiently developed, and largely sub-microscopic. Myrmekite and quartz-biotite symplektite have been rarely observed. Within the sillimanite-potash felspar zone sillimanite spindles may appear enclosed within the potash felspar porphyroblasts. This sillimanite has presumably replaced former muscovite inclusions which are not uncommon in the lower grade rocks. Guidotti (1963) has described similar perthitic microcline megacrysts from the pelitic rocks of the Bryant Pond Quadrangle, Maine.

The rarer plagioclase porphyroblasts are clear, sharply twinned and have a composition between An 7 and An 22. Slight reverse zoning is observed in some grains, and borders against the perthitic potash felspar have a narrow albite rim.

Small garnet porphyroblasts are also present in some rocks and the garnet-sillimanite-potash felspar assemblage is known from the higher grade rocks. However the sillimanite-mangerite subfacies as defined by de Waard (1963) has not been attained.

The matrix between the felspar porphyroblasts consists of granoblastic quartz and plagioclase and oriented red-brown biotite with numerous dark halos. Primary muscovite may be present in the lower grade rocks (sillimanite-muscovite zone). Secondary muscovite is common as a replacement product of fibrolite in the higher grade rocks.

CHAPTER 4.

ALUMINOUS PELITIC SCHISTS.

The aluminous pelitic schists are particularly important because they furnish the aluminosilicate marker minerals used in grading the metamorphism of this area. Because of the appearance of kyanite in the western aluminous pelitic schists, rocks from the eastern and western portions of the area are treated separately.

A. Aluminous Pelitic Schists from the Eastern Portion of the Area.

(a) Minerals and Mineral Assemblages.

The major minerals observed in these schists are quartz, plagioclase, biotite, muscovite, chlorite, garnet, epidote, staurolite, andalusite, sillimanite. Accessory minerals are iron ore, zircon, apatite, sphene, tourmaline, monazite and rutile. As all rocks examined contained quartz, biotite and plagioclase, these minerals will be assumed to occur in all assemblages quoted below.

Biotite zone.

(1) Muscovite-chlorite(porphyroblasts)-garnet epidote

(2) Muscovite-biotite(porphyroblasts)

(3) Garnet-biotite(porphyroblasts)

Andalusite-staurolite zone.

(4) Muscovite-chlorite(porphyroblasts)-andalusite-

garnet (and various sub-assemblages)

(5) Muscovite-chlorite(porphyroblasts)-staurolite-garnet

(6) Muscovite-muscovite(porphyroblasts)-chlorite (porphyroblasts)

(7) Muscovite-andalusite-staurolite

Sillimanite-muscovite zone.

(8) Staurolite

(9) Muscovite-andalusite-sillimanite (and sub-assemblages)

Sillimanite-potash felspar zone

(10) Sillimanite

(b) Biotite zone

When andalusite and staurolite schists are traced into the biotite zone they are seen to grade into fine-grained strongly foliated phyllitic schists containing small white knots of albite and muscovite, and numerous small (< 1.0 mm.) disoriented circular plates of deep-green chlorite. The white knots become more abundant as the andalusite isograd is approached, and it is within these knots that andalusite is first observed. Unlike the arkoses or the semi-pelitic schists these low grade phyllitic schists contain no sedimentary grain relics, and generally there is little evidence for bedding lamination, except for occasional layers of heavy minerals. A strong schistosity is expressed by the orientation of abundant small (< 0.1 mm.) biotite and muscovite flakes. The biotite flakes tend to show a stumpier form than the muscovite, and show a variety of Z-axial colours ranging from pale-brown to sepia to very dark brown. Very fine (< 0.05 mm.) quartz and albite (An 1-2) grains are abundant between the individual mica flakes. Small skeletal garnets, often < 0.1 mm., are sparsely distributed through many of the chlorite porphyroblast bearing rocks. These garnets are commonly concentrated within iron ore bearing heavy mineral bands, and may occur as inclusions within the chlorite megacrysts. Compared with the fine grain size of the matrix the chlorite porphyroblasts are large. The chlorites have a stumpy form although the basal crystal faces are often developed. Several individual flakes may be aggregated into a single clump. The chlorites have a random orientation within the rock matrix. Post-tectonic crystal growth is indicated by the presence of parallel trains of inclusions which pass uninterrupted into the strongly aligned schistose matrix. The chlorite is pleochroic from pale yellow to pale green, has numerous pleochroic halos and small iron ore inclusions, and shows multiple twinning.

A close scrutiny of these rocks for chloritoid failed to reveal the presence of this mineral. Chloritoid has not yet been recorded from the Mt. Lofty Ranges. Green (1963) considers that chloritoid may be more characteristic of kyanite-sillimanite metamorphism. The optical properties of a number of chlorite porphyroblasts from the present area are summarized in Table 3 and Fig. 25. The β refractive index varies from

TABLE 3.

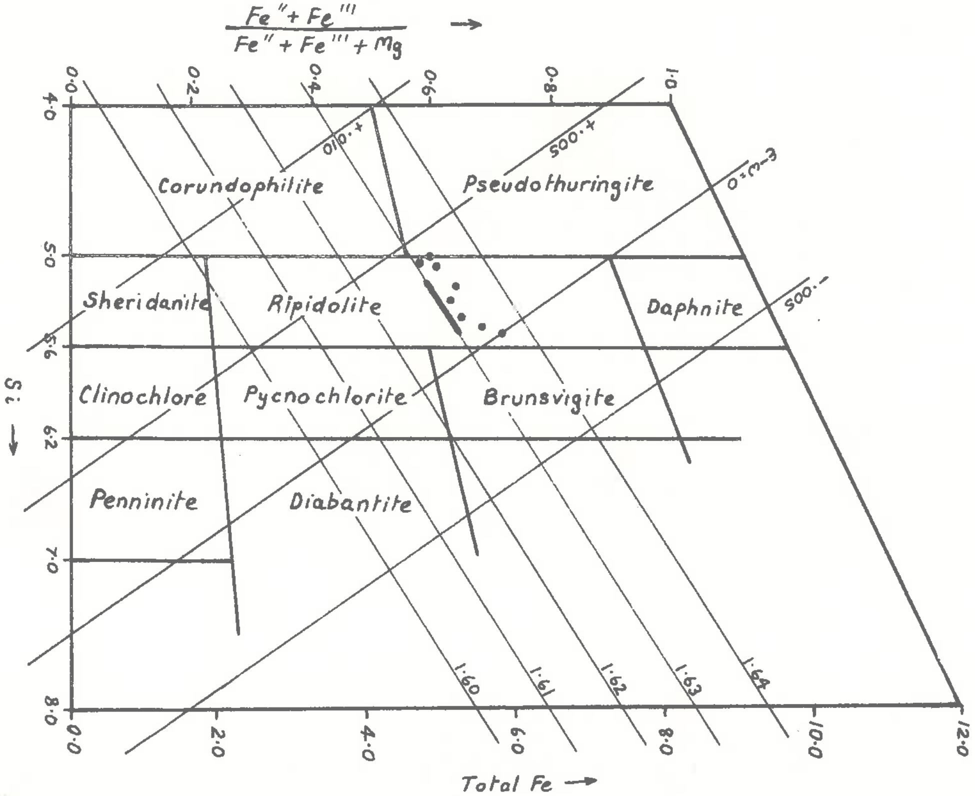
PROPERTIES OF CHLORITES.

Rock No.	Location.	Chlorite properties				
		β R.I.	Optic Sign	Biref.	Alteration	
Chlorite in vughs. (no thin sections)						
577	Pine Hut	1.630	+	n.d.	Fresh.	
576	233.240	1.630	+	n.d.	Fresh.	
599	228.243	1.630	+	n.d.	Fresh.	
Chlorite porphyroblasts in schists.						
588	227.238	1.633	+	.004-5	Fresh.	
586	232.232	1.632	+	.001-2	At edges	1.
(579b	236.227	1.647	−	high	Biotite.)	
581	235.226	1.633	+	.003-4	Fresh.	2.
490	232.215	1.636	+	.000	At edges.	3.
617	189.216	1.631	+	.004	Fresh.	
625	194.202	1.632	+	.002	Fresh.	
626	197.197	1.632	+	.002-3	At edges.	4.
627	196.196	1.634	+	.003	At edges.	5.

1. Alteration to - ve, birefringence .021 (biotite ?).
2. Pleochroic haloes - outer rims isotropic
   inner portion -ve, biref. .009.
3. Alteration to - ve, birefringence low.
4. Alteration to - ve.
5. Alteration to - ve, birefringence high.

FIG. 25.

Composition variation of chlorites from low grade aluminous pelitic schists, Cambrai. Refractive index-birefringence data plotted on diagram from page 151, Deer, Howie and Zussman, 1963, Vol. 3.



Fe'' + Fe''' / Fe'' + Fe''' + Mg →
0.0
0.2
0.4
0.6
0.8
1.0
4.0
5.0
5.6
6.2
7.0
8.0
Si ←
Corundophilite
Pseudothuringite
Sheridanite
Ripidolite
Daphnite
Clinochlore
Pycnochlorite
Brunsvigite
Penninite
Diabantite
+.010
+.005
0=(+-)
-.005
1.60
1.61
1.62
1.63
1.64
0.0
2.0
4.0
6.0
8.0
10.0
12.0
Total Fe →

FIG. 25

1.631-1.636, the mineral is optically positive and the birefringence varies from 0.000 to 0.005. According to Deer, Howie and Zussman, (1963, vol. 3, p.151) these chlorites are ripidolites with Fe total/Fe total + Mg of 0.52-0.60. Fig. 25 illustrates this composition range. Identical chlorite porphyroblasts have been described from Vermont by Woodland (1963).

In many of the fine-grained phyllitic and calc-phyllitic schists near Pine Hut small tension (?) gashes are filled with coarse chlorite, muscovite, quartz, calcite, pyrite, etc. On exposure the calcite is readily removed leaving vugh-like openings coated with euhedral chlorite crystals. As Table 3 shows, these chlorites differ little in optical properties from the chlorite porphyroblasts in the schists.

At one location (rock 579, 236.227) small bronze-coloured porphyroblasts of similar habit to the green chlorite porphyroblasts, were found to consist of golden-brown biotite (β 1.647). This biotite appears to have been formed during a late alteration of the original chlorite porphyroblasts (such alteration has been observed at the edges of some chlorite porphyroblasts in other rocks), and is quite distinct from the sepia-brown oriented biotite of the matrix.

Small (0.5 mm.) knots consisting of albite, disoriented muscovite and lesser quartz are scattered through the matrix of some rocks. Well oriented muscovite is strongly concentrated

FIG. 26.

Tension (?) cracks filled with quartz, calcite, chlorite and muscovite and surrounded by albite rich rims in fine-grained quartzo-felspathic schist.

Location 234.239. Largest cracks 5 cm. in length.

FIG. 27.

Outcrop of andalusite-staurolite-biotite schist.

Location 221.147. Coin diameter 2.8 cm.

Note relic sedimentary lamination outlined by layers richer in biotite.

FIG. 26

FIG. 27

adjacent to these knots, and is seen to sweep around the knots in a syn-tectonic fashion, although there is no evidence that the knots themselves have been rotated.

(c) Andalusite-Staurolite Zone.

As the andalusite zone is entered the white albite-muscovite knots seen in the lower grade rocks take on more distinctive prismatic forms which weather out on exposed rock surfaces. In thin section these prisms are seen to be still largely made up of albite, muscovite and lesser quartz, although rare relic granules of andalusite have been observed encased in some quartz grains. Evidently the reaction forming the first andalusite prisms reversed slightly after the metamorphic peak had been exceeded, causing the andalusite to be pseudomorphed by its reactants.

In most areas in which andalusite or kyanite schists have been observed, the reaction by which these minerals first appear remains a mystery. It has generally been concluded that pyrophyllite or paragonite are the most likely reactants involved in the formation of the aluminosilicates, but this has been seldom verified. The nature of the 'muscovite' in the low grade aluminous schists in the present area has not yet been investigated, but the abundance of albite in association with the early andalusite crystals does suggest the break-down of paragonite in the presence of quartz as one possibility for the origin of andalusite. However, Rosenfeld (1956) considers that paragonite decomposes in the sillimanite zone. A reaction involving muscovite and quartz with the formation of andalusite and potash felspar (Turner and Verhoogen, 1951, p.447), although occurring in some low pressure contact aureoles, is not tenable here.

It has not been possible to determine whether andalusite or staurolite first becomes stable within the metamorphic sequence. In some areas of andalusite-staurolite metamorphism, the staurolite has been observed to appear slightly before the andalusite (e.g. Hietanen, 1956; Green, 1963; Zwart, 1963), while in others andalusite and staurolite appear in the metamorphic sequence at about the same time. The problem of relative andalusite and staurolite stability is, however, not so important as the stability of hydrous versus anhydrous aluminosilicate phases. In a dry environment, andalusite, and perhaps staurolite, are stable under atmospheric P-T conditions (Bell, 1963). The first appearance of andalusite and staurolite within the metamorphic sequence is more closely related to dehydration during progressive metamorphism than an entrance into the stability fields of these minerals.

Staurolite is apparently limited to rocks of a certain range of bulk composition. This mineral has been considered to form from chloritoid (Harker, 1939, 2nd Ed., p.225; Halferdahl, 1961), but this has been seldom observed. The first staurolite at Cambrai appears within rocks which also contain chlorite porphyroblasts. A direct replacement of the chlorite by staurolite is never observed, although at slightly higher grades staurolite becomes quite abundant in some rocks, and the chlorite porphyroblasts have entirely disappeared. With the disappearance of the chlorite, the white mica of the matrix also becomes less abundant. Some staurolites are surrounded by a biotite poor corona. These changes are suggestive of a reaction such as

$$\left[\text{muscovite} + \text{chlorite} + \text{biotite 1} + \text{quartz} \rightleftharpoons \text{staurolite} + \text{biotite 2} + (\text{garnet}) + \text{water}\right]$$

Before the sillimanite isograd both andalusite and staurolite become prominent constituents of some pelitic schists. Both minerals stick out on the exposed and weathered rock surfaces, the xenoblastic andalusite forming rounded to irregular white knots, the staurolite, prismatic golden brown idioblasts. Except for minor post-tectonic outgrowths the andalusite seldom shows its crystal form, preferring to grow as large irregular poikiloblasts sieved with small matrix inclusions. These skeletal crystals may reach 1-2 cm. in size. In rocks which have suffered a mild second deformation resulting in crenulation of the rock schistosity, the andalusites may form large rambling skeletons showing true helicitic structure, indicative of crystal growth after the second deformation. Small iron ore inclusions are common within the andalusites but absent from the biotite bearing rock matrix.

Staurolite may occur separately or with andalusite, although there is no indication of the replacement of staurolite by andalusite (see Akaad, 1956). Staurolite has a strong tendency to present an idioblastic outline, although, like andalusite, the crystals are always sieved with quartz and iron ore inclusions. These inclusions often outline internal S-planes which may show simple sigmoidal, planar or more complex irregular helicitic forms. The staurolite porphyroblasts seldom attain a size comparable with the andalusite crystals, although crystals up to 3-4 mm. are not uncommon.

The bedded nature of andalusite and staurolite rocks is clearly seen in those rocks where the porphyroblasts are concentrated within folded bands of laminae representing the original bedding. A late stage alteration of andalusite to shimmer aggregates (Barrow, 1893) and staurolite to pinitic chlorite-muscovite knots has been occasionally observed. These knots preserve the internal S-planes seen within the unaltered crystals. However, the majority of schists are free of such alteration effects.

Rare isolated garnets may occur within the andalusite and staurolite schists. As in the biotite zone, these are small (generally < 0.2 mm.) and may have a spongy, rounded or idioblastic form. Garnets are sometimes included within andalusite or staurolite, which confirms their earlier growth within progressive metamorphism. Both biotite and muscovite are well oriented within the quartz-plagioclase rock matrix.

FIG. 28A.

Outcrop of staurolite schist showing growth of staurolite crystals in bedding laminations. Location 147.247. Pen 15 cm. (6").

FIG. 28B.

Photomicrograph of staurolite-mica schist. Rock A185-740. Location 155.221. Ordinary light. Equidimensional crystals of garnet and large xenoblasts of staurolite, both with oriented inclusions, in matrix of biotite and quartz.

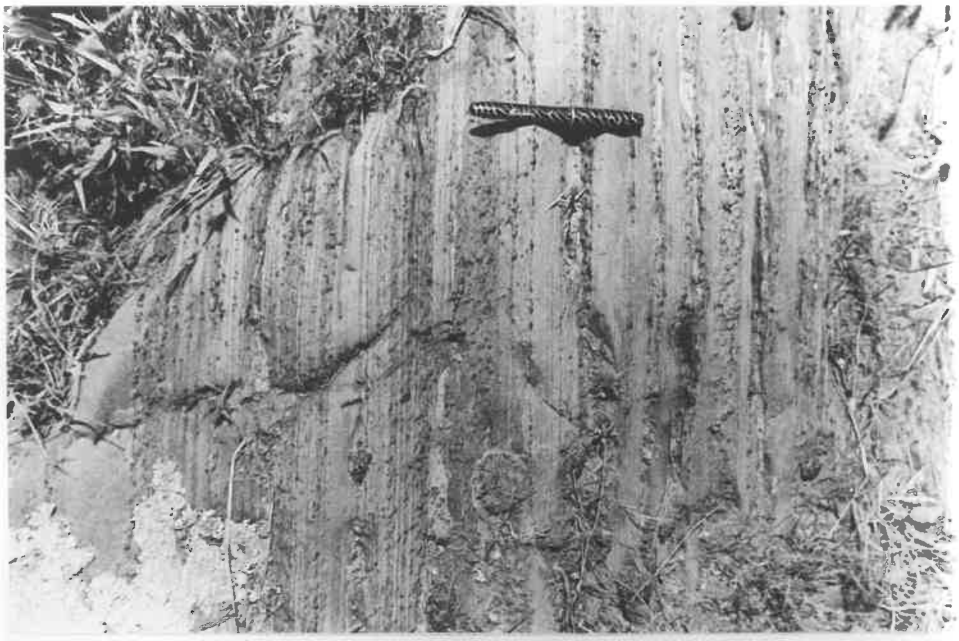


FIG. 28A

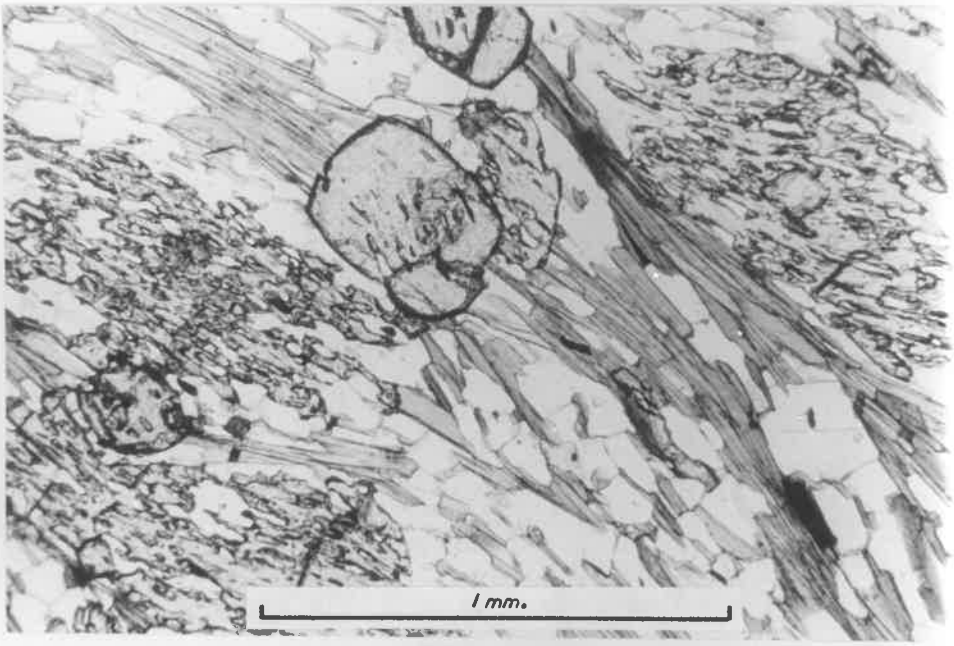


FIG. 28B

With the growth of andalusite and staurolite the matrix minerals have increased their grain-size to about 0.1-0.15 mm. Post-deformation segregation veins of all sizes are observed within these rocks. These will be considered in greater detail in Chapter 9.

(d) Sillimanite-muscovite Zone (First sillimanite isograd of Billings, 1956).

In the lower part of the sillimanite zone the andalusite and staurolite have reached their maximum development. The sillimanite isograd is defined by the first appearance of tufts of fibrous sillimanite adjacent to the andalusite or staurolite crystals. Advancing to higher metamorphic grades the andalusite and staurolite are progressively replaced by sillimanite, until this is the only remaining aluminosilicate mineral. In the more magnesian rocks the staurolite and andalusite may be replaced by cordierite as described in Chapter 5. Green (1963, p.1003) has found that staurolite decomposed before the andalusite-sillimanite transition in New Hampshire. This relationship does not hold in the present area, except where staurolite is replaced by cordierite by a reaction commencing just before the sillimanite zone is reached.

In the lower part of the sillimanite zone large andalusite xenoblasts may reach 2-3 cm. across. Quartz and plagioclase inclusions increase in size progressively from cores to rims indicating growth of the andalusite during a considerable period of grain growth within the rock matrix. These large crystals may also show a variation in internal structure from cores to rims, preserving a record of time relations between crystal growth and deformation (see Chapter 14). Some crystals have also been bodily rotated in their rock matrix after crystal growth ceased.

Within the sillimanite-muscovite zone the matrix has increased its grain-size to 0.1-0.3 mm. The quartz and plagioclase are granoblastic, the felspar being rarely twinned, while the biotite, usually a red-brown colour, is less oriented than in the lower metamorphic grades. Biotite may be slightly concentrated near the margins of some andalusite crystals. In a few rocks large (to 2 mm.) irregular well twinned plagioclase xenoblasts are intergrown with the margins of skeletal andalusites in a diablastic texture (Harker, 1932, p.40).

Fibrolitic sillimanite first appears as minute tufts nucleated in biotite adjacent to andalusite crystals. There is no indication that biotite has broken down with the formation of fibrolite away from the andalusite porphyroblasts. Although the fibrous sillimanite has clearly formed at a later stage in the progressive metamorphism than the andalusite, there is no evidence for a widespread formation of fibrolite throughout these pelitic rocks (c.f. Watson, 1948). The limitation of the fibrous sillimanite to the andalusite borders indicates that the P-T conditions have entered the stability field of sillimanite, and that the andalusite margins are beginning to dissolve, with the nucleation of newly formed sillimanite nearby. Some of the larger andalusite porphyroblasts are surrounded by vague concentric shells. The andalusite with its numerous quartz and plagioclase inclusions is immediately surrounded by a shell of biotite, plagioclase, quartz and secondary muscovite. This is succeeded by a zone of coarse plagioclase-biotite, and then a quartz rich shell in which fibrous sillimanite crowds the quartz in fine streams radial to the margins of the andalusite crystal.

Once andalusite has clearly become unstable and fibrous sillimanite has appeared in some quantity at its borders, stout sillimanite prisms appear within the andalusite. These prisms have an idioblastic form and are aligned parallel to the intersecting andalusite cleavages. As each andalusite has a different orientation, the prisms, also vary their orientation accordingly. A direct transformation of andalusite to sillimanite is evident. This transformation is further discussed in the succeeding chapter.

Higher in the sillimanite-muscovite zone andalusite and staurolite are completely destroyed, although the former presence of andalusite is indicated by the bunches of oriented sillimanite prisms surrounded by, and intergrown with, fibrolite. A few andalusite or staurolite relics may remain as inclusions within quartz. In some rocks these sillimanite knots have undergone considerable alteration to shimmer aggregate.

(e) Sillimanite-potash Felspar Zone.

In this zone the aluminous pelites are represented by knotted sillimanite schists. Within this zone muscovite becomes unstable and breaks down yielding further fibrous sillimanite and a little potash felspar.

B. Aluminous Pelitic Schists from the Western Portion of the Area.

(a) Minerals and Mineral Assemblages.

The minerals observed in these schists are similar to those of the eastern side of the area, except that kyanite may appear in the intermediate metamorphic grades. The following mineral assemblages have been observed. Quartz, plagioclase and biotite were present in all rocks examined, and are to be assumed present in all the assemblages quoted below.

Andalusite-staurolite zone.

(1) Chlorite-staurolite-muscovite

(2) Muscovite-andalusite-staurolite

(3) Muscovite-staurolite-garnet

Sillimanite-muscovite zone.

(4) Muscovite-andalusite-sillimanite (and all sub-assemblages)

(5) Muscovite-andalusite-kyanite-sillimanite-staurolite (and various sub-assemblages)

(6) Andalusite-kyanite-sillimanite-staurolite-garnet

(7) Muscovite-andalusite-sillimanite-garnet

Sillimanite-potash felspar zone.

(8) Sillimanite

(b) Biotite Zone.

The biotite zone aluminous schists are represented by chlorite porphyroblastic phyllites, which are similar in all respects to those described from the eastern portion of the area.

(c) Andalusite-staurolite zone.

The andalusite-staurolite schists are similar in most respects to those on the eastern side of the area. Large andalusite and staurolite and smaller garnet and chlorite porphyroblasts are developed in a fine-grained schistose biotite-muscovite-quartz-felspar matrix. The quartz and felspar granoblasts of the matrix progressively increase in size from 0.02 to 0.2 mm. between the andalusite and the sillmanite isograds.

The relation between deformation and crystal growth in these schists is interesting. Some rocks have a simple planar foliation with evenly spaced well oriented micas. Many rocks have been affected to varying degrees by a second deformation resulting in crenulation cleavage. In some rocks the crenulation cleavage has completely transposed the older schistosity into a new foliation in which the matrix micas are again well oriented. Complete transposition is indicated by an uneven mica distribution in the matrix, by the 'tectonic segregation' of minerals into lenses parallel to the new foliation, and by the preservation of true helicitic crenulations, outlined by traces of inclusions, in the aluminosilicate porphyroblasts. As in the eastern rocks both the andalusite and staurolite porphyroblasts are usually highly sieved with matrix inclusions, including iron ore granules, the staurolite tending to form smaller crystals with a more idioblastic shape. Interpenetration twinning is commonly observed within the staurolite porphyroblasts.

Small garnets are present in many of the andalusite and staurolite schists, some being included within the aluminosilicate minerals. Chlorite appears to be a primary constituent in several staurolite schists. The chlorite may occur either as flakes oriented parallel to the adjacent biotites of the matrix or stumpy porphyroblasts cross-cutting the matrix foliation.

(d) Sillimanite-muscovite Zone.

In the western side of the Cambrai area the sillimanite isograd has been accurately defined by a large number of samples. Sillimanite first appears as stumpy tufts of fibrolite nucleated within biotite adjacent to andalusite and staurolite porphyroblasts. Although andalusite and staurolite may be present well into the sillimanite zone, fibrolite appears in all aluminosilicate bearing schists as soon as the sillimanite isograd is crossed. The growth of sillimanite depends only on the intensive variables, total pressure and temperature and independent of partial fluid pressures. Thus, even though the kinetics of transformation may be such as to prevent a rapid growth of sillimanite from the metastable aluminosilicates, some sillimanite may commence growing as soon as the P total-T conditions enter the field of sillimanite stability. The sillimanite isograd is, therefore, one of the most definitive P-T markers in the metamorphism of pelitic schists.

In the north-west corner of the area kyanite appears in stable equilibrium with andalusite, staurolite, sillimanite and garnet. Here the P-T conditions at the peak of metamorphism were apparently close to those of the triple point of the aluminosilicate equilibrium curves (P 8000 bars, T 350°C). The small xenoblasts of kyanite, generally < 0.5 mm., distinctive because of their high relief and birefringence and prominent cleavages, occur in stable association with andalusite and staurolite. There is no textural evidence suggestive of replacement between these minerals. Fibrolite in the matrix apparently appeared at the peak of metamorphism when the P-T conditions temporarily entered the sillimanite stability field. Although kyanite is relatively rare within these schists,

FIG. 29A.

Photomicrograph of andalusite-kyanite-sillimanite schist.

Rock A185.750. Location 147.237. Ordinary light.

Strongly cleaved kyanite at edge of andalusite xenoblast in quartz-felspar-biotite schist. Some fibrous sillimanite near andalusite and outlining small fold in south-west corner of field of view.

FIG. 29B.

Photomicrograph of staurolite-andalusite-sillimanite schist.

Rock A185-716a. Location 152.212. Ordinary light.

Portions of large porphyroblasts of staurolite (left) and andalusite (right) with small aggregate of fibrolite (centre) in quartz-plagioclase-biotite matrix. Iron ore inclusions in staurolite outline simple internal structure. Skeletal form of andalusite xenoblast preserves an earlier crenulation cleavage not preserved in coarse-grained matrix.

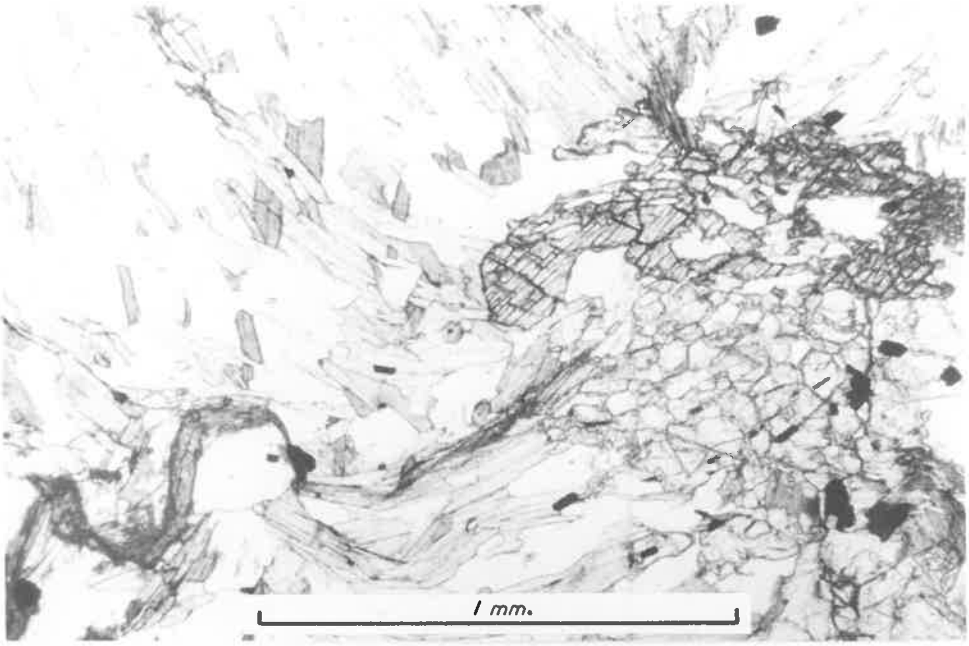


FIG. 29A

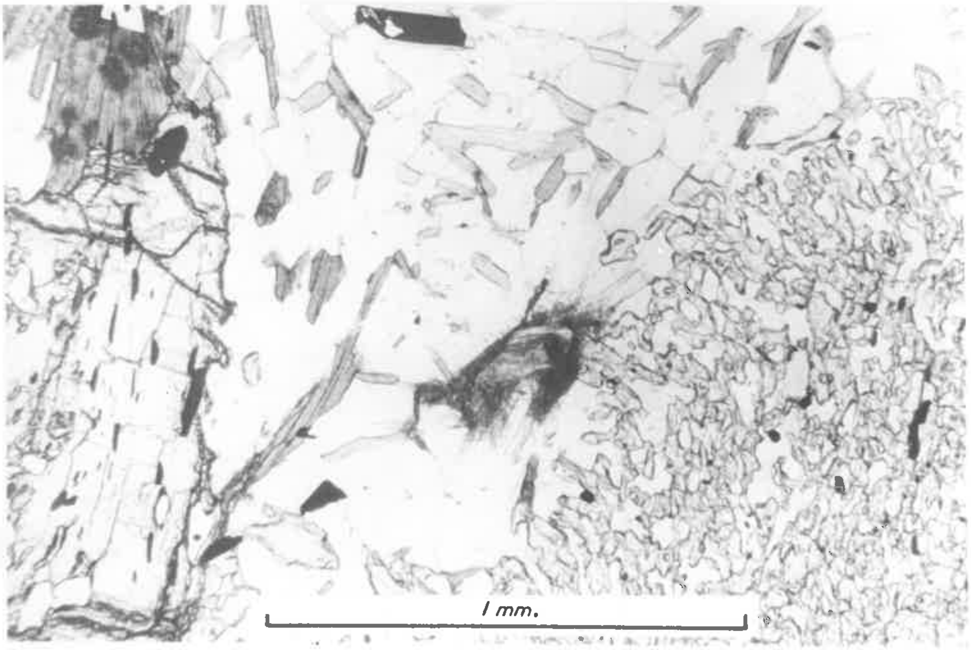


FIG. 29B

large blue blades of kyanite are well developed within associated segregation veins together with andalusite and sillimanite (see chapter 9). The presence of kyanite in the western rocks suggests that a higher pressure of metamorphism was reached in the western portion of the area than in the eastern.

As in the eastern rocks, the andalusite porphyroblasts reach their maximum growth on entering the sillimanite-muscovite zone. They may show complex internal and external textural relations. Much of the andalusite has grown during the early stages of crenulation cleavage development associated with a second phase of deformation. This is indicated by the preservation of well developed crenulations, outlined by trains of matrix inclusions. Both staurolite and andalusite are progressively replaced by both fibrous and prismatic sillimanite within the sillimanite-muscovite zone.

(c) Sillimanite-Potash Felspar Zone.

Within the sillimanite-potash felspar zone sillimanite is the only aluminosilicate present in the aluminous pelitic schists.

C. Discussion.

(a) The Status of Sillimanite as an Index Mineral of Progressive Metamorphism.

The close association of sillimanite with zones of injection and granitization has given rise to two main schools of thought regarding the status of sillimanite as an index mineral of progressive metamorphism.

Some workers (e.g. Watson, 1948; Wychoff, 1952) have considered that sillimanite is a metastable phase formed during the introduction of late stage pegmatitic solutions expelled from zones of intrusion and granitization. The major points in favour of this theory are

(1) the late stage textural relationships which sillimanite often displays in high grade pelitic rocks,

(2) the formation of sillimanite in the inner aureoles of granite intrusions,

(3) the occurrence of sillimanite in pegmatite and aplite veins in the 'zone of veins' surrounding some injection complexes.

Korikovskii (1963) has described the growth of fibrolite during the granitization and migmatization of granulitic gneisses in the Stanov Mountain range. He considered that fibrolitization occurred in rocks not already having an excess of alumina, by a process of acid lixivation connected with the passage of hot acidic fluidal waves, having a low activity of potash, through rocks bordering gneissose granite areas. These solutions are considered capable of leaching alkalies out of biotite, plagioclase and microcline, and precipitating Al and Si in the form of sillimanite and quartz. Korikovskii concludes that the value of sillimanite as an index mineral of progressive metamorphism has been exaggerated.

Other metamorphic petrologists (e.g. Weiss, 1949; Sanders, 1954) have concluded that sillimanite is a purely metamorphic mineral. This conclusion is supported by

(1) the existence of mappable sillimanite isograds cross-cutting tectonic zones which might be supposed to be favourable for the influx of solutions,

(2) the great width of some sillimanite zones in areas of high grade regional metamorphism.

The thesis that sillimanite is only stable within a particular P-T field is supported by recent experimental work on the aluminosilicate minerals (Clark, Robertson and Birch, 1957; Clark, 1961; Bell, 1963; Khitarov et.al., 1963). These experiments have indicated the sluggishness of transformation of these minerals, also suggested by petrographic study, and have shown that aluminosilicate minerals may <u>persist</u> metastably outside of their respective fields of stability. However, the possibility of the <u>formation</u> of an aluminosilicate mineral outside of its own stability field, must be considered with extreme caution.

Korikovskii (1963) states that in many rocks biotite is "replaced with sheaf-like needles of fibrolite; the process begins at the edge of the crystals, then along the cleavages, and, at last, only yellow pleochroic remnants of biotite remain within the fibrolite aggregate" (p.187). A similar replacement of biotite is described by Tozer (1955) from gneisses adjacent to the Main Donegal Granite, and by Chinner

(1961). Chinner has also considered the possibility of nucleation of sillimanite within biotite as a favourable host. In the Cambrai area rocks not having an excess of alumina show a replacement of muscovite by sillimanite in the highest grade rocks. This fibrolite may penetrate into adjacent grains, including biotite, but there is no evidence to suggest that biotite has commenced to break down to form fibrolite. In rocks having an excess of alumina, fibrous sillimanite first appears as tufts and sheaves of needles embedded within biotite adjacent to aluminosilicate porphyroblasts. Such rocks are bounded from sillimanite free andalusite and staurolite schists by a sharply defined sillimanite isograd. Within the sillimanite zone there is no indication of the fibrolitization of biotite well away from the andalusite and staurolite porphyroblasts. The fibrolitization of biotite in the present area is best considered as a nucleation of sillimanite within biotite once the rocks having an excess of alumina entered the P-T stability field of sillimanite.

The formation of coarse sillimanite prisms within andalusite porphyroblasts, by the transformation of andalusite to sillimanite within the mapped sillimanite zone, is further evidence that the sillimanite zone rocks of Cambrai have, in fact, entered the P-T stability field of sillimanite, and that fibrolite is not formed here as a metastable phase.

It may be noted here that coarse sillimanite crystals have been observed in some of the highly differentiated pegmatitic aplite intrusions within sillimanite zone rocks north of the Kanappa Mine (see Chapter 11), but there is no indication of the development of fibrous sillimanite adjacent to these veins. Sillimanite in metamorphic segregation veins is also confined to the sillimanite zone as defined by the first appearance of sillimanite within the aluminous pelitic schists.

(b) P-T Conditions indicated by the Aluminosilicate phases.

Until the recent experimental determination of the P-T stability fields of the aluminosilicates andalusite, sillimanite and kyanite (Bell, 1963; Khitarov et.al., 1963), a number of P-T diagrams had been produced based on thermodynamic and petrographic principles (Miyashiro, 1949; Thompson, 1955; Hietanen, 1956; Schuiling, 1957; Schuiling, 1962). The temperature values were determined from existing experimental data on a number of metamorphic reactions, and pressure values were based on estimates of the stratigraphic depths of burial during metamorphism. Limited thermodynamic data on the aluminosilicate minerals has lead Skinner, Clark and Appleman (1961) to estimate the slopes of the phase boundaries separating the three phases. The experimentally determined slope of the kyanite-sillimanite curve 13.2 bars/^{o}C (Clark, 1961), is considerably smaller than that predicted, 21.2 bars/^{o}C, thus indicating the necessity to consider horizontal P and T gradients in progressive metamorphism (Clark, Robertson and Birch, 1957, 1958) to explain the commonly observed kyanite-sillimanite succession in many metamorphic belts. This a departure from the geothermal gradient concept advocated by some metamorphic petrologists to explain progressive metamorphic zonation (e.g. Schuiling, 1958). One of the most startling revelations of the new experimental work is the high pressures necessary for the formation of kyanite. The triple point of the alumino-silicate diagram lies near T 350°C and P 8.5 Kb. Thus, at least 8 Kb., equivalent to about 30 Km. depth of burial or 100,000 feet of stratigraphic thickness, is necessary for kyanite to form at all. These pressures are far in excess of those estimated from field data (e.g. Schuiling, 1957; Haller, 1962). Read (1962) has pointed out that actual depths of 30 Km. for the formation of kyanite are unacceptable, and has emphasized the importance of "geological depth".

Clark (1961) has considered the possibility of tectonic overpressures, that is, pressures exerted during rock deformation, to help explain the excessive pressures necessary for the formation of kyanite. Clark's calculations suggest the possibility of an overpressure of about 1 Kb. (approximately 4 Km. overburden) in rocks under a total pressure of about 10 Kb. The presence of post-tectonic kyanite and kyanite in veins in some kyanite bearing areas, including the present one, suggests that tectonic overpressures do not offer a satisfactory solution in all cases. The experimental work has also demonstrated that shearing stress is not necessary in the formation of kyanite (Clark, Robertson and Birch, 1957).

The aluminosilicates, being anhydrous polymorphs of a one-component system, have phase fields which are defined only by total pressure and temperature and are not affected by partial fluid pressures, except in so far as these limit the stability of the anhydrous phases at low temperatures. The total pressure may be considered to be made up of three components; load pressure, tectonic overpressure and fluid pressure. It seems likely that fluid pressures in metamorphism are much higher than have generally been considered, and that these have made a considerable contribution to the total pressure of metamorphism.

From the aluminosilicate diagram it can be seen that the lowest temperature at which sillimanite can form will be at the P-T conditions of the triple point (T 350°C and P 8.5 Kb.). The occurrence of the three polymorphs in the north-western portion of the mapped area, indicates that the P-T conditions of the triple point were reached here. It is interesting to note that the plagioclase in equilibrium with epidote in calc-silicate rocks adjacent to these andalusite-kyanite-sillimanite bearing schists has a composition of An 27. If this value is applied to Ramberg's (1949) plagioclase-epidote equilibrium curve a temperature of 350°C is indicated, in remarkable, although, perhaps fortuitous, agreement with the temperature of the triple point.

As the sillimanite isograd is followed away from the north-west corner of the area kyanite becomes progressively less stable, until only andalusite and sillimanite are present in the eastern portion of the area. According to the steep negative slope of the andalusite-sillimanite curve in the equilibrium diagram, the sillimanite isograd in the eastern part of the area indicates a temperature in excess of 350°C, perhaps 400°C, and pressures less than 8.5 Kb. perhaps 7-8 Kb. Because of the small size of the area and the lack of evidence for more than one major pulse of metamorphism, it is unlikely that the lower total pressures indicated by the andalusite-sillimanite metamorphism on the eastern side of the area can be attributed to lesser depths of burial, or smaller tectonic overpressures. It is concluded that fluid pressure variation must explain the increasing total pressure towards the western side of the area. This conclusion is also supported by the marked suppression of reactions releasing volatiles in passing from the eastern to the western rocks. (See Chapters 6 and 7).

(c) Metastable Aluminosilicate Phases.

Theoretically two aluminosilicate minerals should not co-exist except at the P-T conditions of the phase boundaries. The common observation of two aluminosilicates in association over a wide portion of the metamorphic sequence in the present area, is by no means unique, and has been noted by numerous workers in other metamorphic belts. The sluggishness of the polymorphic transformations is well documented by the experimentalists (Clark et al., 1957; Bell, 1963). Chinner (1961) has considered that this sluggishness is related to a band of indifference, induced by the low free energies of phase change in these minerals, in which one polymorph may persist metastably within the stability field of another. The presence of two phases in association may also be aided by the presence of certain foreign ions, such as Mn in andalusite. In the present area andalusite and staurolite may persist well into the sillimanite zone, especially if these minerals are protectively enclosed within quartz or felspar.

The occurrence of the three aluminosilicates, kyanite, andalusite and sillimanite, together with staurolite, in the aluminous schists in the north-west corner of the present area must also indicate a certain degree of metastability under P-T conditions close to the triple point. Similar assemblages have been described by Hietanen (1956, 1961) from Idaho, Chakravarty (1960) from India, and Woodland (1963) from the Burke area, Vermont.

(d) Comparison with Other Areas of Andalusite-staurolite Metamorphism.

A number of close similarities between the metamorphism of the aluminous schists at Cambrai and some other andalusite-staurolite-sillimanite metamorphic belts, emphasizes the necessity to consider andalusite-staurolite-sillimanite metamorphism as a unique metamorphic facies series (the low pressure intermediate type of Miyashiro, 1961). Woodland (1963) has described the aluminous schists of the Burke area, Vermont. The presence of chlorite porphyroblasts in the low grade rocks, the partial inversion of andalusite to prismatic sillimanite, with the growth of fibrolite in nearby biotite (which Woodland attributes to biotite decomposition), the poor degree of grain growth in the matrix until the sillimanite zone (although the metamorphic minerals appear to demonstrate approximate equilibrium), the association of granitoid intrusions with metamorphism, and the development of the metamorphic climax during and after a second deformation phase, are features common to both the Vermont and Cambrai areas.

The recognition of various facies series based on varying total pressures during metamorphism (Miyashiro, 1961) has discouraged the strict sub-division of metamorphism into "contact" and "regional" types (Turner and Verhoogen, 1951, 1960). Regional sequences of andalusite-sillimanite and andalusite-staurolite metamorphism, known as the Buchan Type (Read, 1952) or the Abukuma Type (Miyashiro, 1961), are now recognised throughout the world. Regional andalusite-sillimanite metamorphism is described from the Abukuma Plateau, Japan (Miyashiro, 1958, Shido, 1958); Cooma, New South Wales (Joplin, 1942); Albury-Adelong-Wantabadgery-Tumbarumba region of New South Wales (Vallance, 1953); Celebes (Miyashiro, 1961); and southern California. A slightly higher pressure andalusite-staurolite metamorphism is recognised in Scotland (Read, 1923, 1952); Finland (Eskola, 1927); Korea (Yamaguchi, 1951); Japan (Ishioka and Suwa, 1956); New England (Green, 1963; James, 1955; Woodland 1963; etc.); Pyrenees (Zwart, 1963) and the Olary Province, South Australia (Campana and King, 1958).

The eastern side of the Mt. Lofty Ranges as exemplified by the present area, has been metamorphosed under P-T conditions typical of andalusite-staurolite-cordierite-sillimanite Buchan Type metamorphism. We have seen that kyanite appears in the aluminous schists on the western side of the Cambrai area, indicating increasing total pressures to the west. West of the present mapped area, the central axis of the Mt. Lofty Range metamorphic belt is occupied by sillimanite grade gneisses and zones of granitization. On the western flanks of the metamorphic belt in the Williamstown region the presence of andalusite has not been authenticated, and the metamorphism appears to be of the kyanite-sillimanite type (Alderman, 1942; Mills, 1963). At least two other metamorphic belts (New England and the Scottish Highlands) are known to show gradations between the andalusite-staurolite-sillimanite sequence and the kyanite-sillimanite sequence (Chinner, 1962).

CHAPTER 5.

ANTHOPHYLLITE AND CORDIERITE SCHISTS.

Introduction.

Anthophyllite is a widespread metamorphic mineral throughout the higher grade portions of the Mt. Lofty metamorphic belt. Extensive occurrences of anthophyllite bearing rocks have been recorded by Chinner (1955) in the Tanunda Creek area, and by White (1956) on the Mannum one-mile sheet (White and Thatcher, 1957). Anthophyllite schists are also common in the Rocky Gully-Monarto area west of Murray Bridge (Johns and Kruger, 1949).

Judging from the literature cordierite is of much more restricted occurrence in the Mt. Lofty Ranges. Cordierite bearing rocks have not been recorded from the central or western portions of the metamorphic belt, and seem to be confined to a narrow zone hugging the eastern side of the ranges, extending from the present area through Palmer, Reedy Creek and Rocky Gully.

Knotted cordierite and andalusite schists are well developed in bands within hornfelsed quartz-biotite schists adjacent to the Victor Harbour-Encounter Bay granite (Browne, 1920; Bowes, 1954). Away from the granite contact the country rocks, mainly folded quartz-felspar-biotite schists, belong to the biotite zone of regional metamorphism. Closer to the granite contacts these rocks have been recrystallized to a more decussate hornfelsic texture, and contain local bands of cordierite, andalusite, albite and chlorite bearing rocks. Bowes (1954, p. 199) considers that these rocks have resulted from the overprinting of the regionally metamorphosed schists by a later contact metamorphism associated with the intrusion of the granite bathylith. Bowes (p.195) found no evidence pertaining to the order of appearance of the minerals, and the cordierite and andalusite seem to have grown contemporaneously. A stable cordierite-andalusite assemblage within some of the contact hornfelses indicates a low pressure metamorphic environment characteristic of hornfelses adjacent to high-level plutonic intrusions (Miyashiro, 1961).

The cordierite rocks of Rocky Gully, Reedy Creek and the present area are different from those adjacent to the Victor Harbour-Encounter Bay granite. On the eastern side of the ranges the cordierite rocks have a more regional distribution, are limited to the sillimanite zone of metamorphism, and the association of andalusite with cordierite is unstable. Evidence from the present area indicates that the cordierite has grown through reactions involving the break-down of former andalusite and staurolite near the lower boundary of the sillimanite zone. The metamorphism in these areas is similar to the Buchan type of Aberdeenshire (Read, 1952).

Johns and Kruger (1949) have recorded a number of cordierite bearing assemblages from the Rocky Gully-Monarto area, but do not discuss the cordierite paragenesis. These rocks, which occur near the western margin of the Murray Bridge granite bathylith, are described as being of high metamorphic grade, and are associated with partially migmatitic gneisses and schists. The authors seem to regard these high grade rocks as part of the contact aureole of the granite, but it is not clear what actual effects the granite intrusion has had on the high-grade regionally metamorphosed rocks.

A similar problem arises in the Reedy Creek area, south of Palmer, where Sando (1957) has described sparse but large knobs of cordierite within high grade gneisses and cordierite-anthophyllite schists adjacent to the western contact of the Reedy Creek granodiorite-tonalite mass. Here too, it is not yet clear what metamorphic effects the intrusion of the granodiorite mass has had upon the high grade regionally metamorphosed rocks, and the status of cordierite as a "contact" mineral is in doubt. White (1956) does not record cordierite in the Palmer area, although it seems likely, from the present study, that cordierite might be found in the Palmer-Milendella region.

(a) Anthophyllite and Cordierite Rocks from the Cambrai Area.

Anthophyllite and cordierite bearing rocks appear just before the sillimanite zone, and are stable to the highest grades of metamorphism reached in the present area. These rocks are usually limited to sparse and narrow stratigraphic horizons, but anthophyllite is more widespread in the western part of the area where it occurs in place of biotite within parts of the main arkose formation, as described previously. Cordierite assemblages are much less common than those containing anthophyllite. In some rocks anthophyllite and cordierite form a stable association. The cordierite has a regional distribution unrelated to local intrusive activity, and this mineral is notably absent from the kyanite zone rocks in the north-west corner of the area.

The following assemblages have been observed:

Anthophyllite bearing rocks.

(1) Quartz-plagioclase-biotite-anthophyllite-garnet.

(2) Quartz-plagioclase-biotite-anthophyllite.

(3) Quartz-plagioclase-anthophyllite.

(4) Plagioclase-anthophyllite.

Cordierite bearing rocks.

(5) Quartz-plagioclase-biotite-muscovite-cordierite-sillimanite.

(6) Quartz-plagioclase-biotite-muscovite-cordierite.

(7) Quartz-plagioclase-biotite-cordierite-sillimanite.

(8) Quartz-plagioclase-biotite-cordierite.

Anthophyllite-cordierite bearing rocks.

(9) Quartz-plagioclase-biotite-cordierite-anthophyllite-garnet-chlorite.

(10) Quartz-plagioclase-biotite-cordierite-anthophyllite-chlorite.

(11) Quartz-plagioclase-biotite-cordierite-anthophyllite-garnet.

(12) Quartz-plagioclase-biotite-cordierite-anthophyllite.

(13) Quartz-plagioclase-biotite-cordierite-anthophyllite-dravite-chlorite.

Each of these assemblages may be accompanied by the accessory minerals tourmaline, apatite, zircon, iron ore, sphene, rutile or monazite. It must be noted that the term anthophyllite has been used here in its general sense, covering both true anthophyllite and gedrite compositions. From the aluminous nature of the associated minerals it seems probable that the anthophyllites approach a gedritic composition.

(b) Anthophyllite and Cordierite Rocks from the Eastern Part of the Area.

(1) Biotite zone and andalusite-staurolite zone.

Within the well defined progressive metamorphic sequence on the eastern side of the area, anthophyllite and cordierite appear more of less simultaneously just prior to the initial appearance of sillimanite in the aluminous pelitic schists. At lower grades these rocks may be represented by biotite-chlorite, biotite-staurolite and biotite-andalusite schists. Without chemical anlyses it is not possible to determine which particular low grade schists are liable to yield cordierite and anthophyllite at a higher grade of metamorphism.

The lowest grade rock to contain cordierite (specimen 505, location 227.188, assemblage (9)) is a fine-grained uniform non-laminated grey quartz-felspar-biotite schist containing numerous dark-grey ovoid cordierite xenoblasts to 4 mm. in length, elongated parallel to the schistosity. The rock was found at the northern margin of a large dolerite mass, but appears to represent the first appearance of cordierite within the regionally metamorphosed sequence. In thin slice the rock matrix has a grain size of 0.05-0.15 mm., with small strongly aligned dark brown biotites evenly spread in granular quartz-albite. Small pale-green pseudo-isotropic chlorite flakes are dispersed through the biotite matrix and appear to belong to the primary assemblage. Diffuse ovoid xenoblasts of cordierite are highly sieved by matrix inclusions, and are commonly enveloped in thin rims of isotropic alteration products. Small granules of staurolite are closely associated with the cordierite in three ways. Firstly as ragged relics within the cores of aome cordierites. Secondly as small granules usually enclosed within the isotropic clay-like alteration products rimming the cordierites. Thirdly as trains of small granules along sparse syn-metamorphic micro-fractures where these cut the cordierite poikiloblasts. Following these shears into the matrix it is seen that fresh biotite is oriented within them. Some pale green chlorite flakes appear within the shears adjacent to the cordierite. Within the matrix rare groups of anthophyllite prisms are surrounded by chlorite free marginal zones, suggesting that the anthophyllite may have formed from this mineral. Small (0.2 mm.) pink isotropic idioblasts of garnet are sparse, but widely dispersed through both the matrix and cordierite xenoblasts. There is no evidence for the disequilibrium of cordierite and garnet. The schistosity is not deflected by either the cordierite, anthophyllite or garnet, all of which appear to be of post-tectonic growth.

The much reduced quantity of biotite and the presence of staurolite relics within the cordierite xenoblasts, indicates that the cordierite has formed by a reaction involving both of these minerals, and perhaps chlorite also. A slight reversal of this reaction under falling temperature would account for the appearance of staurolite granules at the cordierite margins, and within the shear planes. These relations are expected where the P-T conditions are on the threshold of the reaction forming cordierite from staurolite and biotite.

A nearby cordierite-anthophyllite-chlorite schist (specimen 525, location 226.176, assemblage (10)), illustrates the formation of anthophyllite from chlorite. This is a light grey rock with compositional layering outlined by alternating coarse (0.2 mm.) and fine (0.05 mm.) quartz-albite layers and bands rich in pearly honey-coloured anthophyllite. The anthophyllite forms large radiating sheaves and prisms set in a fine granular matrix of quartz and plagioclase, in which pale chlorite and platy iron ore define a strong foliation. The pale, almost colourless, anthophyllite is often closely intergrown with the near colourless to pale green chlorite (birefringence .009, + ve). Progressive gradations from an intimate mixture of poorly crystallized anthophyllite and chlorite, through a more recrystallized mottled anthophyllite with rare chlorite relics, to well-formed radiating sheaves of tapered prisms of anthophyllite cleared of chlorite, may be observed. These relations indicate the formation of anthophyllite from original chlorite. A depletion of quartz adjacent to the newly formed anthophyllite prisms suggests that this mineral is also used in the reaction. Rare pale green-brown biotite flakes are also present in the matrix, but these play no part in the formation of the amphibole. Flattened ovoid xenoblasts of cordierite to 3mm. in length, sieved with matrix inclusions, have appeared in some layers, but there is nothing to suggest the manner in which this cordierite may have formed. The cordierites have undulose extinction and distinct yellow pleochroic halos, but no twinning.

(2) Sillimanite-muscovite zone.

South of the Marne River and north of the Kanappa Mine, within the sillimanite-muscovite zone of metamorphism, several narrow beds of knotted cordierite schists have been observed. These occurrences contain the freshest and best developed cordierite in the area, some rocks having glassy grey ovoid cordierite xenoblasts 1 cm. in diameter. Compared with the rocks described above, the matrix grain-size in these rocks has increased noticeably, so that the biotite flakes and interstitial quartz and plagioclase grains are readily discerned in hand specimens. The biotite flakes are moderately oriented into a preferred schistosity and are commonly concentrated into foliae, or have an irregular distribution which may be related to either original bedding lamination or metamorphic segregation. In some rocks a slight wavyness of the foliation is observed. This is expressed by irregular shallow folds in the matrix and diversion of the foliation about the large cordierite porphyroblasts. Open

FIG. 30A.

Photomicrograph of cordierite-anthophyllite schist.

Rock A185.413a. Location 222.139. Crossed polars.

Bladed anthophyllite penetrating into large cordierite poikiloblast. Matrix consists of quartz plagioclase and biotite.

FIG. 30B.

Photomicrograph of cordierite schist showing twinning in large cordierite poikiloblast.

Rock A185-829. Location 221.142. Crossed polars.

Other minerals present are plagioclase, andalusite and tourmaline.

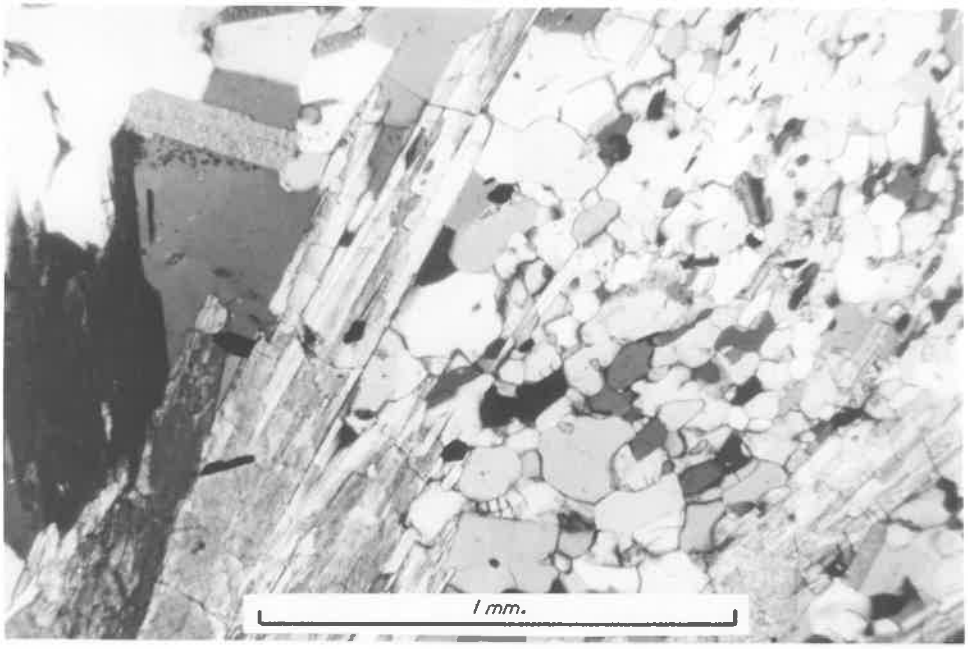


FIG. 30A

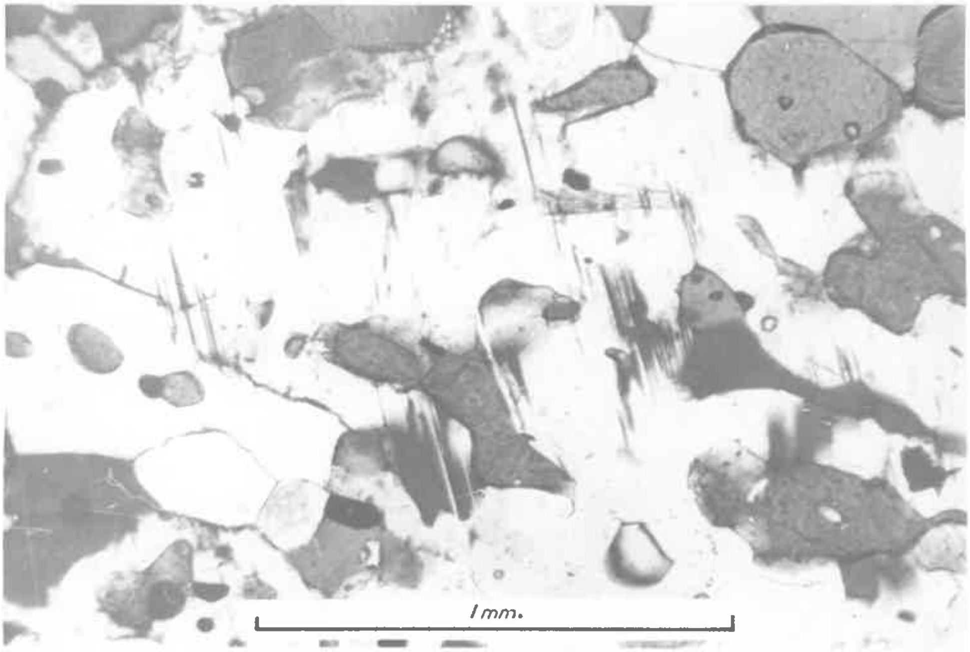


FIG. 30B

S-shaped trains of matrix inclusions within the cordierites in such rocks suggest slight syn-tectonic rotation during growth. In some single cordierite crystals the inclusion trains show rotations of up to 90°. The possibility of syn-tectonic cordierites has been considered by Schreyer (1959). These structures may be attributed to the imposition of a weak crenulation cleavage on the rock during cordierite crystal growth. Other rocks have apparently escaped this late deformation and planar traces of matrix inclusions in the cordierite pass into the matrix foliation without disturbance.

The quartz and plagioclase granoblasts of the matrix have a variable grain-size of 0.05-1.0 mm. from rock to rock, the average grain-size being near 0.2 mm. In some rocks the quartz and plagioclase increase in grain-size adjacent to the cordierite knots. Both quartz and plagioclase generally lack evidence of post-tectonic strain. The plagioclase is albite to albite-oligoclase in composition. Moderately oriented stumpy biotite flakes are abundant in the matrix, although decreasing in amount adjacent to cordierite. The biotite is little altered and has a wide variety of pleochroic schemes, the Z-axial colours ranging from mustard-green, to mustard-brown to mid-brown, to a deep red-brown. In the strongly pleochroic red biotites, black pleochroic halos are common.

Andalusite and staurolite relics are preserved as cores within many of the larger cordierite porphyroblasts. The andalusite and staurolite are consistently enclosed within the cordierite, and there is no evidence for the stable association of cordierite with either of these minerals. The degree of persistence of these relics is related to the availability of quartz and/or biotite, which are used in the formation of cordierite. Thus quartz, and sometimes biotite as well, is depleted from the matrix adjacent to the cordierite porphyroblasts, and with limited diffusion through the cordierite, the reaction ceases. The aluminosilicate relics are sometimes quite large, reaching almost 1 c.m. in width in some porphyroblasts. In other cases only minute granules may be preserved within the cores of the cordierite crystals. The relics usually contain trains of minute inclusions, much smaller than those in the surrounding cordierite, thus indicating an earlier stage of crystal growth for the aluminosilicates. The minute inclusions are often arrayed in planes, indicating the growth of the aluminosilicates after the formation of the cleavage associated with the first, and major, phaseof folding.

Being within the sillimanite zone both the andalusite and staurolite relics have begun to break down to sillimanite. This is evidenced by the growth of coarse sillimanite prisms within, and near, the aluminosilicate relics encased within the cordierite porphyroblasts, and by the growth of fine sillimanite fibres in the matrix without.

The sillimanite prisms may reach 1 mm. in length and 0.1 mm. in diameter, and are invariably idioblastic towards the enclosing minerals. In cross-section the prisms have a strong $\{010\}$ cleavage, and are seen to possess simple $\{110\}$ crystal faces, unmodified by the presence of $\{010\}$ pinacoids.

The sillimanite prisms within the staurolite have a seemingly random orientation, with no relation to the staurolite lattice structure. With further growth of cordierite and corrosion of the staurolite, the sillimanite prisms are left embedded within the cordierite, where they show no observable corrosive features, and appear to be in stable association with the cordierite.

The sillimanite prisms within the andalusite relics, however, have a unique orientation, which is clearly related to the andalusite lattice. The prisms, although usually isolated from each other, are in perfect parallel lattice alignment. Where the andalusite is still preserved it is seen that the c-axes of the sillimanite and the c-axes of the andalusite are parallel. This topic will be discussed further at a later stage.

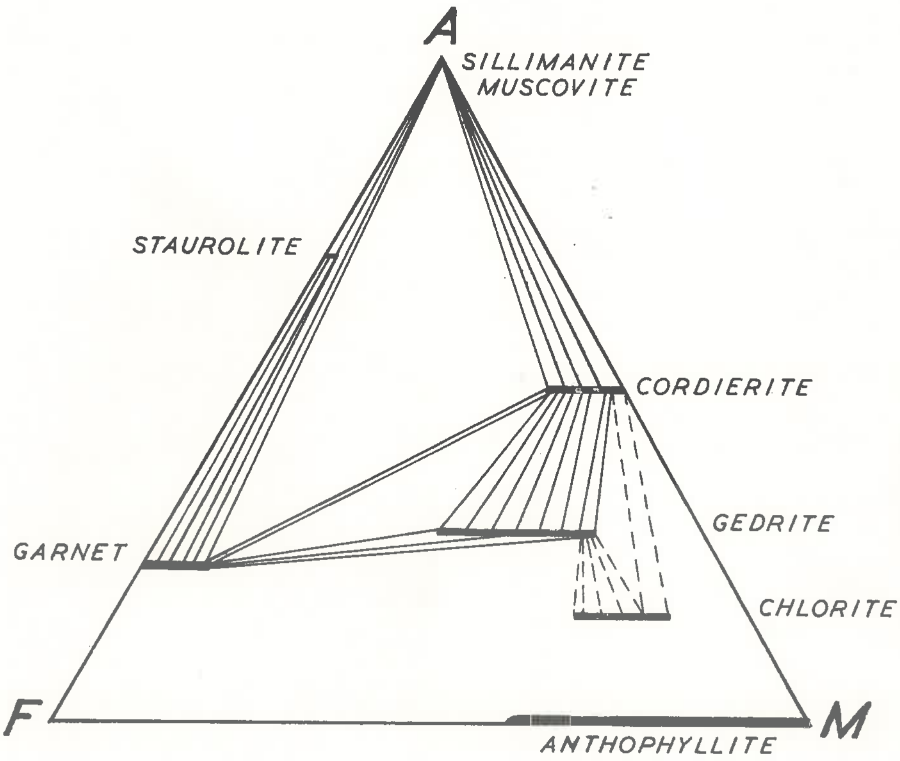
Thus, although andalusite and staurolite are clearly metastable, sillimanite appears to share a stable relationship with the cordierite. This is indicated not only by the lack of corrosion of the sillimanite prisms, but also by the preservation of fine sillimanite fibres near the edge of the cordierite porphyroblasts. Within rocks containing anthophyllite, sillimanite is absent. This is to be expected from a consideration of the AFM diagram (Figure 31).

The cordierite porphyroblasts are large and xenoblastic and often show a preferred growth along the weak rock foliation. A few rocks have large cordierite xenoblasts with no aluminosilicate relics. These cordierites, although containing trains of small inclusions of biotite, quartz and felspar, are unstrained, and show sharp uniform extinction. In those rocks with aluminosilicate relics within the cordierite, the cordierite porphyroblasts may be structurally divided into two portions. An inner zone adjacent to the aluminosilicate relics consists of a polygonized mozaic of cordierite sub-grains in semi-parallel

FIG. 31.

AFM diagram for cordierite-anthophyllite rocks showing the absence of a stable assemblage of sillimanite-anthophyllite.

FIG. 31


A
SILLIMANITE
MUSCOVITE
STAUROLITE
CORDIERITE
GEDRITE
GARNET
CHLORITE
F
M
ANTHOPHYLLITE

orientation with the main outer portions of the porphyroblasts, which, although having biotite and felspar inclusions, have an unstrained appearance. Polygonized mozaics within large crystals are usually a result of recrystallization after strain. The absence of strain in the outer portions of the porphyroblasts reflects a lack of post-tectonic strain in the rock as a whole. The consistent occurrence of the mozaic adjacent to the aluminosilicates suggests that the strain may have resulted from thermal contraction effects on cooling (Rosenfeld and Chase, 1961).

Weakly developed hour-glass twins are rarely developed within the cordierite; they occur most commonly within the polygonized grains. The yellow pleochroic halos, so typical of cordierite, are usually absent.

Alteration of the cordierite is common. In hand specimens glassy cordierites are often rimmed by a brown halo of an isotropic alteration product. Rarely the alteration product consists of a flaky clay-like mineraloid with a very faint birefringence. A similar alteration of cordierite is described by Schreyer and Yoder (1961) and Layton (1963). In other rocks the cordierite is replaced by a mat of fine sericite and chlorite. This sericitic mat often shows relic cleavage structures inherited from the cordierite, and in many cases the sericite flakes are sub-aligned within a particular direction in relation to these relic crystal planes.

In rocks lacking aluminosilicates, anthophyllite may appear as sheaves of idioblastic prisms stable with both biotite and cordierite. Muscovite appears to be stable within these cordierite rocks but has been observed only once. Then it occurred both as small stumpy grains within biotite in the matrix, and as post-tectonic porphyroblasts near, and embedded within, the margins of the cordierite porphyroblasts.

A late formed pale-green chlorite is common, in small amounts, in many of these cordierite rocks. It appears as a direct alteration product pseudomorphing biotite flakes and as cross-cutting flakes replacing biotite adjacent to the cordierite porphyroblasts. It is doubtful whether this chlorite has the status of a stable primary mineral within the rock assemblage.

Of the minor constituents iron ore and tourmaline are particularly common. The ore occurs as abundant inclusions within the cordierite, perhaps in part inherited from the former aluminosilicates, but is rare within the rock matrix. Tourmaline is abundant as small zoned grains with various pleochroic schemes in

FIG. 32A.

Photomicrograph of oriented prisms of sillimanite replacing large xenoblast of andalusite with cordierite and some plagioclase forming a general background.

Rock A185-413c. Location 222.139. Crossed polars.

FIG. 32B.

Photomicrograph of staurolite porphyroblast rimmed by corona of cordierite. The matrix consists of quartz felspar and biotite.

Rock A185-829. Location 221.142. Ordinary light.

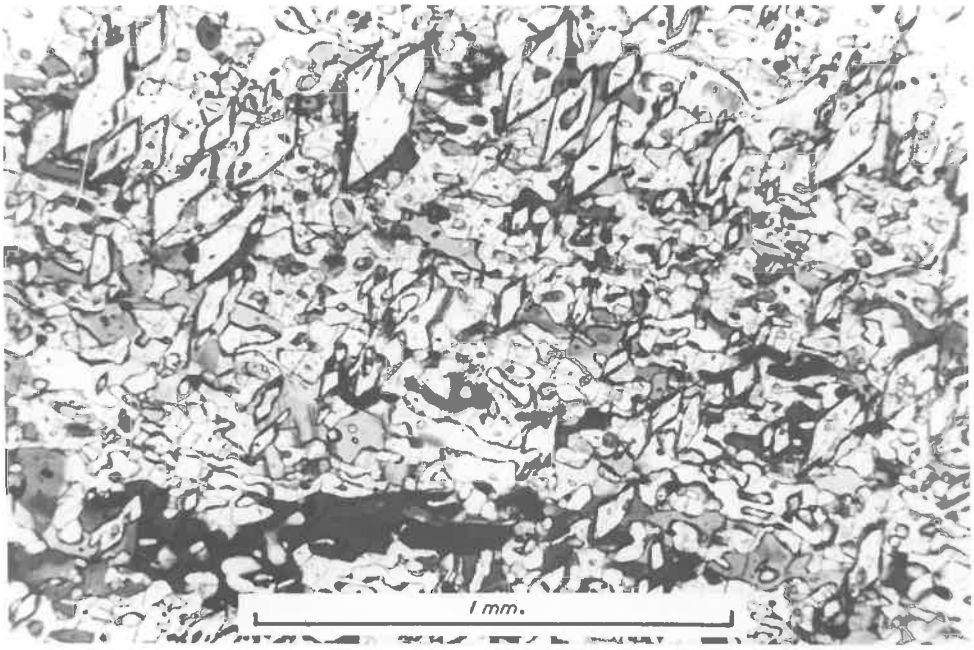


FIG. 32A

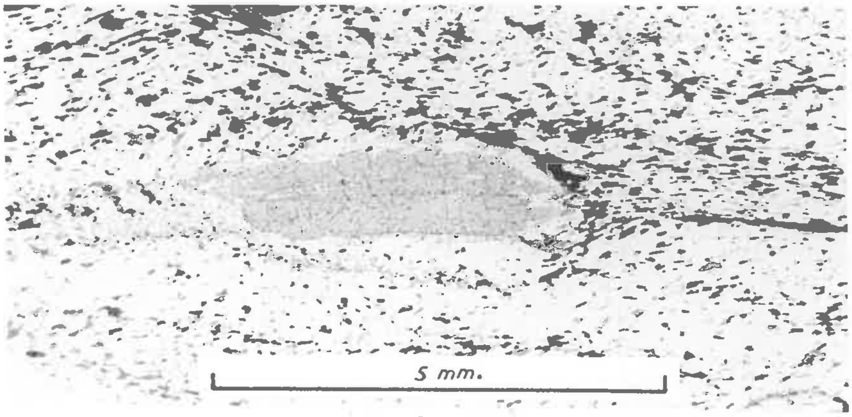


FIG. 32B

blue-green and yellow. In some rocks the abundance of tourmaline suggests a partial metamorphic origin. In other cases the close association of the tourmaline grains with distinct zircon-iron ore heavy mineral laminations reflects a sedimentary origin. The abundance of dravitic tourmaline within rock 375 (location 232.165, assemblage (13)) suggests a metamorphic or metasomatic origin. A fuller description of this rock is presented in chapter 10 under "Tourmaline metasomatism". The commonly reported occurrence of tourmaline in cordierite bearing rocks (e.g. Tilley and Flett, 1930) suggests that the original sediments giving rise to these rocks might be somewhat more enriched in boron than the average shale.

(3) Sillimanite-Potash Felspar zone.

South of Kanappa Mine, towards Saunders Creek, within the sillimanite-potash felspar zone, the assemblage cordierite-potash felspar, although probably stable, has not been observed. The rocks of this area have been altered, after shearing, through a late stage pervasion by hydrous fluids. Fresh cordierite has not been found, although sericitic pseudomorphs of this mineral are common, especially within anthophyllite bearing rocks. In most cases anthophyllite has also suffered fracture and extensive alteration through various stages to pale-green chlorite. Even when completely altered the prismatic shapes are distinctive. A somewhat less altered light coloured rock from just south of the Kanappa Mine (specimen 462, location 217.106, assemblage (1)) has a weak foliation and a pronounced lineation due to the alignment of red-brown biotite flakes and anthophyllite prisms within a fine-grained quartz-felspar matrix. The rock is studded with widely dispersed small (0.3 mm.) ruby-red garnet dodecahedra.

Garnet is also a common constituent of many of the more altered anthophyllite rocks from further south. These garnets, although having a few quartz inclusions in their cores, are usually idioblastic in outline and may reach 2 mm. in diameter. Occasionally within the one thin section a progressive alteration sequence of garnet is seen. Fresh idioblastic garnets are fractured and become partially, and then wholly, replaced by fine matted chlorite. A further state of alteration, seen in some rocks, is a decomposition of the chlorite to fine feathery sericite, with the expulsion of iron as limonite. At this stage the idioblastic garnet outline is still perfectly preserved, indicating static replacement. Biotite, although sometimes replaced by chlorite, is generally not much affected by this late stage alteration. The plagioclase may become slightly clouded. The late stage pervading solutions appear to have been rich in potash.

(c) Anthophyllite and Cordierite Rocks from the Western Part of the Area.

(1) Andalusite-staurolite zone.

Cordierite first appears in the western portion of the area just before the first appearance of sillimanite. Again anthophyllite enters almost simultaneously. The presence of staurolite relics in some cordierites suggests the cordierite has again formed in part from this aluminosilicate. The first appearance of cordierite in the western rocks is marked by specimen 340c (location 174.159, assemblage (8)). This is an exceptionally fine-grained massive buff-coloured rock in which closely packed ovoid cordierite xenoblasts are barely distinguishable in hand specimen. In thin section the cordierite appears as 1 mm. poikiloblasts covering about 60% of the slice. However, the poikiloblasts are so highly sieved with fine matrix inclusions that only about 10-15% of the slice is actually cordierite. The cordierites are embedded in a foliated quartz-felspar-biotite matrix of exceptionally fine grain (0.05 mm.). The porphyroblasts are marked off by a thin isotropic alteration rim and by the lesser abundance of pale-brown biotite among the inclusions compared with the matrix.

(2) Sillimanite-muscovite zone.

Within the sillimanite zone, sillimanite may appear in the cordierite rocks lacking anthophyllite, (e.g. rock 323, location 181.126, assemblage (5)). Rock 323 is a light-grey bedded slightly schistose granofels in which ovoid glassy grey cordierite porphyroblasts, to 4 mm. across, are flattened within the schistosity. In thin slice the texture appears rather even, and at first sight there is no sign of the cordierite. The presence of distinct pleochroic halos reveal the cordierite as very diffuse skeletons interstitial to the quartz-felspar matrix. Within the cordierite poikiloblasts the red-brown biotite is considerably depleted. A few andalusite relics occur within the cordierite and fibrous sillimanite is common within the matrix, where it has nucleated in biotite, especially in those flakes near cordierite containing andalusite relics. A few muscovite flakes in the matrix also appear primary. The quartz-felspar matrix of this rock, and several of the other low grade cordierite and anthophyllite rocks, is very fine-grained and weakly recrystallized. The fine grain-size of the matrix is in marked contrast to the large skeletal cordierites and prismatic anthophyllites. Strings of rutile, apatite, tourmaline and zircon outline bedding plane laminations.

Rocks 397 and 302 illustrate features characteristic of cordierite-anthophyllite rocks within the sillimanite zone. Rock 397 (location 164.138, assemblage (12)) is a green-black coarse-grained moderately foliated cordierite-anthophyllite-biotite gneiss. Large sheaves of thin anthophyllite prisms, scattered biotites and occasional cordierites are set in a fine to medium-grained quartz-felspar matrix. There is a suggestion of minor folding of some of the anthophyllite and biotite rich portions of the rock, which may be related to a weakly developed crenulation cleavage. The anthophyllite crystals reach 1 cm. in length and generally occur in aggregates of crystals in subparallel orientation. The amphibole is pleochroic X lemon, Y green-yellow, Z blue-grey and rare halos occur about small radioactive inclusions. The anthophyllite is in stable association with a deep mustard-green biotite. A granular quartz-plagioclase matrix makes up some 60% of the rock. The cordierite poikiloblasts, which reach 1 cm. across, show a tendency for preferred growth in the schistosity. Some of the xenoblasts have a rather twisted form indicating growth of the crystals around shallow micro-folds in the matrix, or slight rotation in growth. Iron ore, biotite, quartz and rare anthophyllite and staurolite relics occur as inclusions. Although the cordierite may be strongly sericitized, the enclosed biotite and anthophyllite remain unaffected. Anthophyllite inclusions within the cordierite have a corroded appearance when compared with the sharply edged prisms in the matrix. This suggests that the anthophyllite may be involved in the formation of the cordierite. Rock 319 (location 169.102, assemblage (2)) contains large (2 mm.) poikiloblasts of what, at first, appear to be fresh cordierite. The presence of some fine lamellar twinning also suggests plagioclase. X-ray powder photographs indicated that these porphyroblasts were plagioclase, and that no cordierite was present in the rock.

Rock 302 (location 174.104, assemblage (12)) is a light coloured platy quartzo-felspathic gneiss, in which pale biotite and anthophyllite, within the gneissic layers, outline a distinct foliation and lineation. This rock would be unremarkable but for the fact that the cordierite has a strong preferred crystallographic orientation. This orientation has been studied and the results will be presented in chapter 13. In thin slice thin layers of anthophyllite and cordierite outline a gneissic texture in the quartz-felspar-biotite matrix. The anthophyllite is faintly coloured and almost imperceptibly pleochroic. An alignment of the prisms within the foliation gives the rock a pronounced lineation. Within other gneissic foliae cordierite forms flat lensoidal grains up to 5 mm. in length. The cordierite has suffered partial alteration to sericite and chlorite. Late decussate flakes of a very pale chlorite (+ve, Δ=.009) are associated with the cordierite and anthophyllite. Again there is some doubt concerning the primary or secondary status of this mineral. The matrix (85%) is largely granoblastic quartz and albite-oligoclase with a grain size near 0.15 mm., containing sparse flakes of a pale mid-brown biotite.

(3) Sillimanite-potash felspar zone.

Only two occurrences of knotted cordierite gneisses are known from the migmatite zone in the south-west. These are located at 152.121 (specimen 226) and at 158.112. Both of these rocks have suffered almost complete sericitization of their original cordierite, an X-ray examination of the centre of a large 3 cm. nodule from specimen 226 indicated the presence of cordierite. The cordierite nodules are embedded within a biotite rich gneissose matrix. The biotite occurs in coarse (0.5 mm.) stumpy flakes which have a tendency to segregate into irregular schlieren parallel to a foliation within the quartz-oligoclase matrix. The biotite has been largely altered from a red-brown to a green-grey variety. The green biotite is in the process of further alteration to chlorite. A distinct normal zoning characterizing the plagioclase grains is, no doubt, also connected with these late retrogressive changes.

Several beds of dark foliated, and sometimes lineated, anthophyllite amphibolites have been observed in the western portion of the area. Some of the freshest, and best developed, anthophyllite specimens come from a number of beds in an anthophyllite rich zone lying stratigraphically above the main arkose unit on the western margin of the area. Some of these rocks are made up entirely of plagioclase and anthophyllite, while others also have quartz. Apatite, iron ore and sphene are common accessory minerals. The amphibole forms clear idioblastic individuals with few inclusions. The anthophyllite has well developed closely spaced $\{210\}$ cleavages, and usually have a simple $\{210\}$ prismatic form. Cross-fractures, near basal partings, are often present. The granular plagioclase matrix is usually of albite-oligoclase composition, but rarely becomes labradorite.

An unusual dark-green fine-grained nematoblastic garnet-anthophyllite rock at 165.131 (specimen 276, assemblage (1)) shows evidence of strong micro-folding of the anthophyllite during a phase of crenulation cleavage development. Some of the numerous idioblastic garnets through the anthophyllite-quartz matrix show some evidence of rotation during growth. The anthophyllite spindles, embedded in quartz, are seen to stream around small, often rootless, folds within microlithons between planes of crenulation cleavage. Some irregular albite xenoblasts with numerous quartz and amphibole inclusions, show evidence of both syntectonic and post-tectonic growth. In some cases the crenulation cleavage planes diverge around the albite porphyroblasts, while in others a true helicitic structure of crenulated anthophyllites is well preserved.

The replacement of biotite by anthophyllite in some areas of arkose on the western side of the area casts some doubt on the surmise that these anthophyllite- rich rocks represent sedimentary horizons. Further suspicions are aroused where anthophyllite-garnet rocks can be traced along strike into biotite-garnet rocks.

(d) Mineralogy.

Because of the difficulties involved in the optical verification of some cordierites, all rocks suspected of containing cordierite were subjected to X-ray powder diffraction. Selected porphyroblastic knots were separated from the rocks, crushed to pass a 120 mesh sieve, and passed through a Franz Isodynamic Separator. Cordierite was found to be weakly magnetic, the order of decreasing magnetic susceptibility of the commoner minerals being: Chlorite, biotite, tourmaline, muscovite, cordierite, apatite, sphene, rutile, quartz and plagioclase. The presence of cordierite within the appropriate magnetic fraction was verified by the use of the distinctive high intensity, low angle reflection at 8.54 Å, corresponding to the 110 or 200 planes of the crystal lattice. This was distinguished from other low angle reflections such as biotite 10.1 Å, anthophyllite and gedrite 8.27 Å and 9.00 Å, muscovite 10.1 Å, chlorite 7.15 Å and staurolite 8.2 Å.

The results of a U-stage optical study of some cordierite bearing rocks is presented in tables 4 and 5. Cordierite compositions in 10 rocks were determined by measurement of the optic axial angle, it being assumed that the cordierite is in the low structural state (Miyashiro, 1957). The results indicate a composition variation of 11-19% of the iron molecule. There is no systematic variation of composition with grade of metamorphism. The variation is related to the original rock composition. Biotite rich rocks have the more iron rich cordierites. Cordierite associated with pale biotites (e.g. 323, 321) is iron poor. Rock 505 is the only rock containing garnet and unaltered cordierite. The cordierite has the maximum iron content (19% iron molecule).

The plagioclase of these rocks ranges in composition from

TABLE 4.

PROPERTIES AND COMPOSITIONS OF SOME ANTHOPHYLLITES AND CORDIERITES.

Rock No.	Anthophyllite	Cordierite	
	2Vz	2Vx	Composition.
13b	-	74.3	Fe 19
413c	-	74.9	Fe 18-19
413a	70-72	73.2	Fe 19
302	74-76	76.5	Fe 18
389	-	73.5	Fe 19
505	n.d.	74.0	Fe 19
321	-	87.1	Fe 11
323	-	87.5	Fe 11
375	68-70 ?	83.7	Fe 13
397	84-88	80.9	Fe 15

TABLE 5.

PLAGIOCLASE COMPOSITIONS FROM ANTHOPHYLLITE AND CORDIERITE ROCKS.

Rock No.	2Vz	Plagioclase compositions from 2Vz	5-axis method.
369	83.6	An 8-9	-
557	-	-	An 6
90a	84.2	An 9	An 6
90b	87.5	An 13	An 8
89	80.4	An 5	An 4
84b	87.0	An 12	An 8
83b	84.6	An 10	An 4
64	84.8	An 10	An 4
63f	80.7	An 5	An 4
63c	82.8	An 8	An 4
63b	83.7	An 8	An 4
63a	81.9	An 6	An 1
13b	82.6	An 7	-
413c	82.8	An 8	-
413a	88.6	An 15	-
302	83.5	An 9	-
389	84.8	An 10	-
505	88.8	An 15	-
321	84.0	An 9	-
323	84.6	An 10	-
375	85.0	An 10	-
397	83.4	An 9	-

-----

albite to oligoclase - only rarely have more basic compositions been recorded. In the ten cordierite-bearing rocks examined the plagioclase compositions determined from 2V measurements varied from An 7-15. In 12 other anthophyllite bearing rocks the compositions determined from optic axial angle measurements ranged from An 5-13, while compositions determined by the 5-axis technique, using Emmons' (1959) migration curves, lay in the range An 1-8. The individual compositions, presented in table 5, were determined by using the average of five readings from different grains in the same rock. The values determined by the 2V measurements are consistently slightly higher than those determined by the 5-axis technique. This may suggest a slight departure from the low temperature plagioclase structure. The twin laws encountered were predominantly albite, rarer pericline and a few Carlsbad. This simplicity of twinning is typical of metamorphic rocks (Gorai, 1951; Turner, 1951).

Optics of the anthophyllites have not been studied in detail. Four anthophyllites gave 2Vz variation from 70-88°, the value of the optic axial angles increasing with the absorption colours. The pleochroic schemes of the anthophyllites range from almost colourless, through neutral yellows, honey browns to steel greys, with increasing absorption.

The muscovite porphyroblasts margining the cordierites in rock 389 have a low optic axial angle (2Vx 36.5), suggesting a high magnesium content (Tröger, 1956, p. 83).

The late formed chlorite flakes are consistently pale green to almost colourless, are optically positive and have a birefringence similar to quartz. A similar chlorite in the cordierite hornfelses of Kenidjack, Cornwall has been attributed by Tilley and Flett (1930) to "chloritization".

Mention has previously been made of an observed crystallographic relation between andalusite and its replacing dimorph sillimanite. In order to examine these relations more closely a number of andalusite, sillimanite, staurolite and cordierite crystallographic orientations were measured in various rocks. Some of these results are presented in Fig. 33. The results indicate that there is no consistent orientation relationships between the staurolite lattice and the replacing sillimanite or cordierite. Nor are there any crystallographic relationships between the cordierite lattices and the enclosed sillimanite or andalusite. However, a close parallelism between the andalusite and sillimanite orientations was often observed. Consistently X = c of andalusite was parallel to Z = c of the sillimanite, Y = b of andalusite was parallel to X = a of sillimanite and Z = a of andalusite was parallel to Y = b of sillimanite. Thus the optic axial plane of andalusite is perpendicular to the optic axial plane and the $\{010\}$ cleavages of sillimanite. Furthermore the $\{110\}$ prism faces of the sillimanite are sub-parallel to the $\{110\}$ prismatic cleavages of the andalusite. These relations are figured in Fig. 34.

FIG. 33.

Measured orientation relationships between sillimanite andalusite and cordierite in specimens A185-13 and A185-413c. The close tie between the orientation of the sillimanite and andalusite lattices is clearly indicated.

FIG. 33

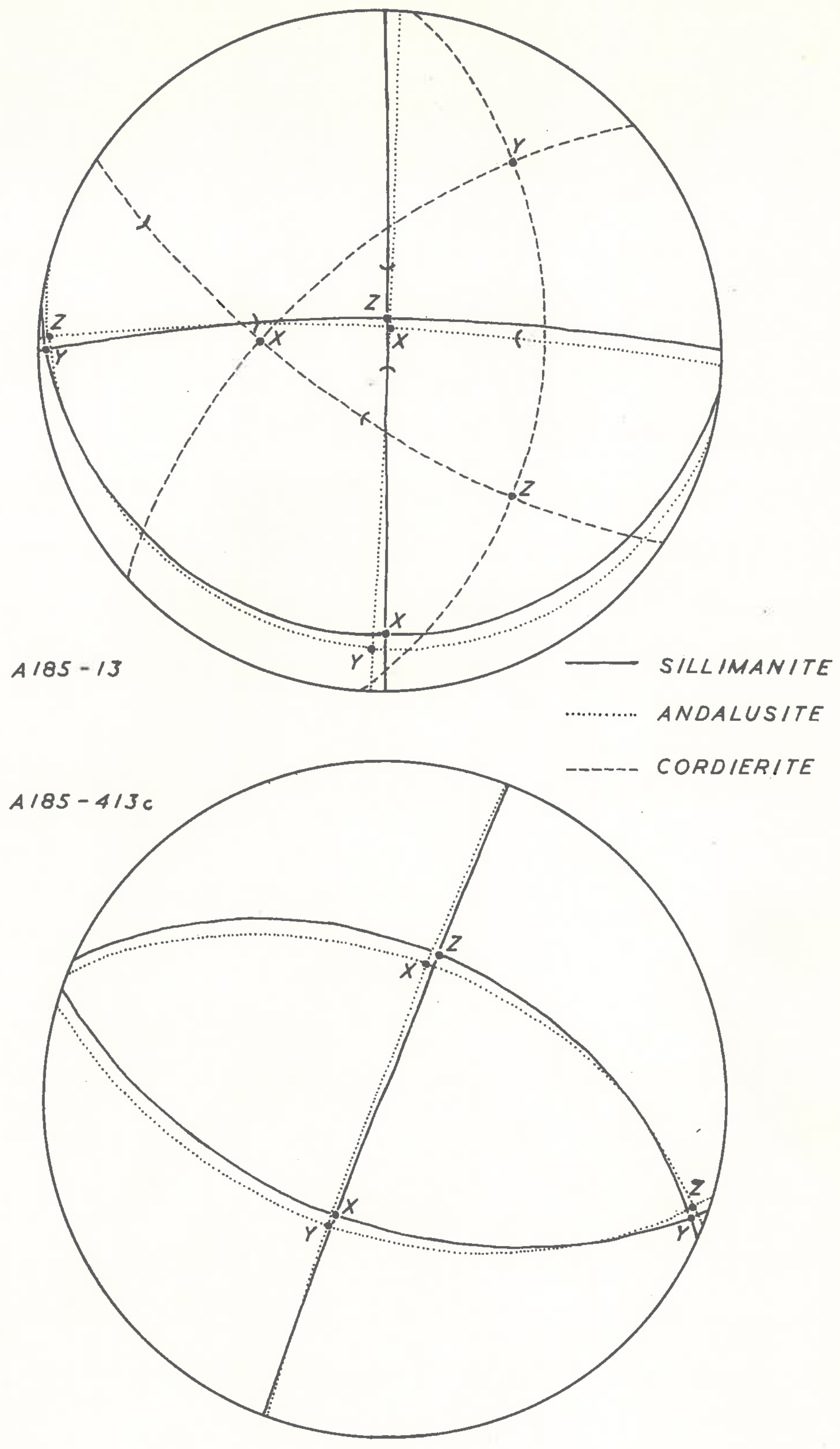


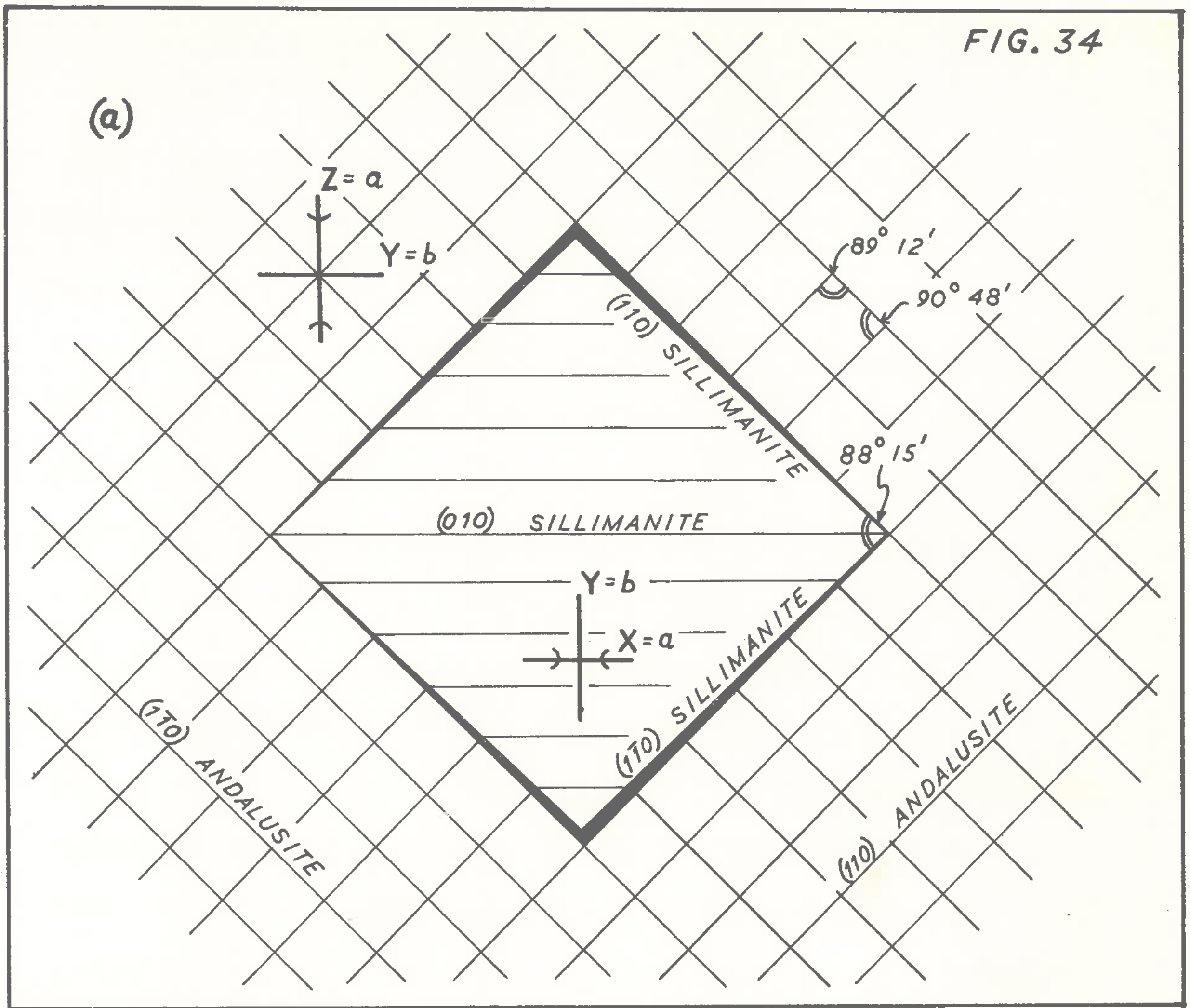
FIG. 34.

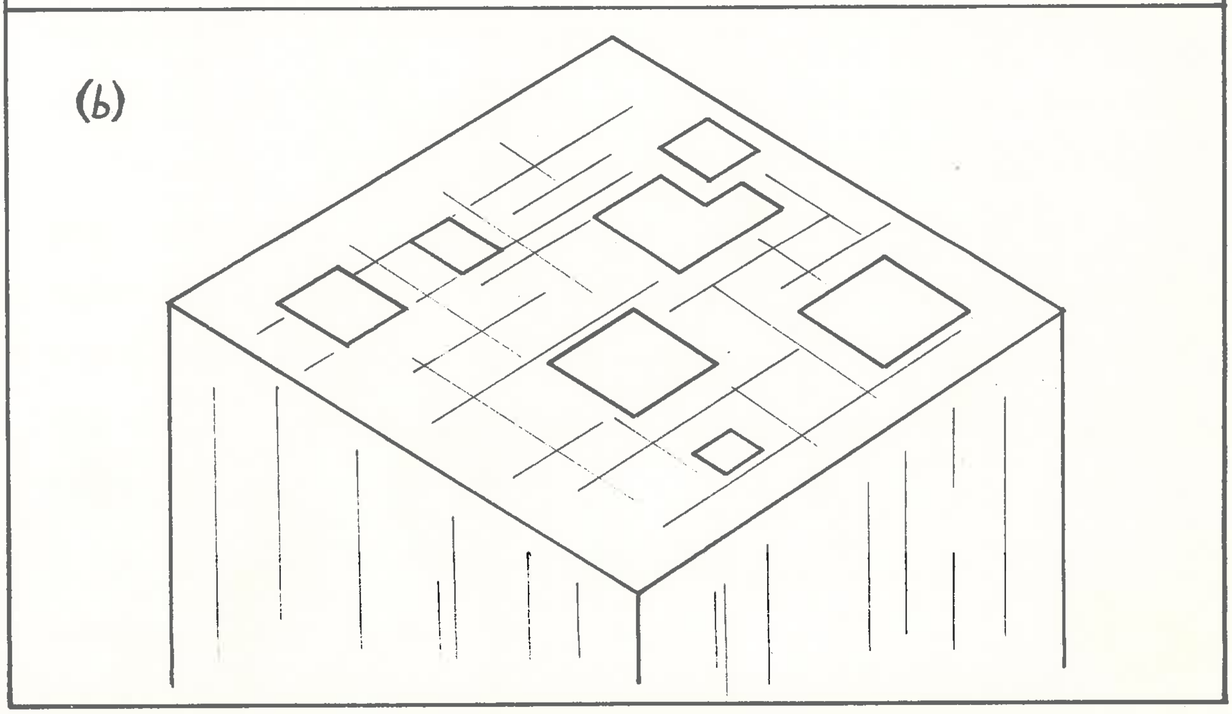
Crystallographic relationships between andalusite and its replacing dimorph sillimanite.

(a) Projection along c-axes of andalusite and sillimanite showing prismatic andalusite cleavage and sillimanite cleavage. The crystallographic angular relationships and orientation of the principal optic axes is indicated.

(b) Idealized block diagram showing sillimanite prisms replacing andalusite crystal.

FIG. 34


(a)
Z = a
Y = b
89° 12′
90° 48′
(110) SILLIMANITE
88° 15′
(010) SILLIMANITE
Y = b
X = a
(1̄10) SILLIMANITE
(1̄10) ANDALUSITE
(110) ANDALUSITE


(b)

Dana-Ford (1947, p. 615, 616) record angles of 89° 12' for $\{110\}^{\wedge}\{1\bar{1}0\}$ in andalusite, and 88° 15' for $\{110\}^{\wedge}\{1\bar{1}0\}$ in sillimanite. From Fig. 34 it can be seen that the theoretical angles between the andalusite cleavages and the sillimanite crystal faces, for the orientation observed, are 90° 48' and 88° 15' respectively; a difference of 2° 33'. For the orientation observed this would result in a small gap between the andalusite cleavages and the replacing sillimanite prisms. Assuming the angular relations given above the size of this gap may be calculated. If 100 cc. of andalusite were completely converted to sillimanite prisms in the manner described, the gaps would occupy 4.35 cc. Assuming densities of 3.13 for andalusite and 3.23 for sillimanite, theoretically, 100 cc. of andalusite completely converted to sillimanite would leave a space of 3.1 cc. The small difference between these figures, an excess of 1.25 cc. of sillimanite, could be accommodated by the migration of some of the sillimanite forming components to the ends of the growing prisms, or by the non-uniform growth of the $\{110\}$ sillimanite prism faces within the andalusite. In practice, within the cordierite rocks, the sillimanite prisms are seldom entirely enclosed within the andalusite, and the gap between the andalusite and sillimanite becomes occupied by cordierite.

It is concluded that the observed angular relations between andalusite and the enclosed sillimanite prisms, in rocks undergoing the dimorphic phase change andalusite to sillimanite, is a consequence of the growth of the sillimanite prisms along the intersecting $\{110\}$ cleavages in the andalusite, governed by favourable angular relationships between the sillimanite crystal faces and the andalusite cleavages. The existence of a narrow theoretical gap between the andalusite cleavages and the sillimanite prism faces could be occupied during the transformation, by fluid through which ions from the corroding andalusite could be transferred and added in layers to the growing $\{110\}$ crystal faces of the sillimanite, thus encouraging a large idioblastic growth of the prisms.

Strangely, few examples of this type of replacement have been recorded in the literature. An example from the Palaeozoic of the Pyrenees has been figured by Lacroix (1893-6, Vol. 1, p. 42, Fig. 21) and from Vermont by Woodland (1963, Fig. 4, p. 358). The occurrence of sillimanite prisms embedded within cordierite has been commonly reported (e.g. Scotford, 1956; Parras, 1958; Schreyer, Kullerud and Ramdohr, 1964, Pl. 1, Fig. 3) but there is seldom evidence that these sillimanites have inverted from other aluminosilicates.

(e) Discussion.

(1) Reactions involved in the formation of cordierite and anthophyllite.

A short review of a number of metamorphic reactions important in the formation of cordierite is presented by Schreyer and Yoder (1961). Several metamorphic reactions seem to be important in the development of cordierite in the present area. Cordierite seems to be most commonly formed by a reaction involving staurolite, biotite and quartz.

e.g. $\left[\underset{\text{biotite}}{K_2\,(Mg_{0.8}Fe_{0.2})_4\,Al_4\,Si_6\,O_{22}\,(OH)_2} + 2\,\underset{\text{quartz}}{SiO_2} + \right.$

$2\,\underset{\text{staurolite}}{Al_2SiO_5.Fe(OH)_2} \rightleftharpoons 2\,\underset{\text{cordierite (20\% Fe mol.)}}{(Mg_{0.8}Fe_{0.2})_2\,Al_4\,Si_5\,O_{18}} +$

$\left. FeO + H_2O + KOH \right]$

A rather magnesian biotite is necessary for the formation of cordierite, so that in rocks having iron-rich biotite and staurolite, cordierite may not form, the staurolite becoming replaced by sillimanite in the sillimanite zone as described in the previous chapter. At 178.141 alternations of staurolite and cordierite bearing beds outcrop in a watercourse. The stable occurrence of staurolite in such rocks is probably related to the biotite composition. Hietanen (1961) has noted similar alternations of staurolite and cordierite bearing beds west of the Idaho Bathylith. In the above reaction it is assumed that the staurolite and cordierite have the same Mg/Fe ratio. This is, of course, an oversimplification. A reaction relation must exist such that the biotite becomes more iron-rich with the formation of cordierite. Eventually, at high grades, biotite, sillimanite and cordierite form a stable association, as observed in many of the sillimanite zone cordierite rocks. The FeO and KOH released in the above reaction may be reabsorbed in the new equilibrium biotite or go to form iron ore granules (common in the cordierite, but absent from the matrix), orthoclase (not observed) or muscovite (late muscovite has been observed adjacent to cordierites in some instances).

In other rocks the cordierite has formed from andalusite by a reaction such as the following.

$$\left[\underset{\text{biotite}\quad(309\text{ cc.})}{K_2(Mg,Fe)_4(Al,Fe)_4Si_6O_{22}(OH)_2} + \underset{\substack{\text{andalusite}\\(104\text{ cc.})}}{2Al_2SiO_5} + \underset{\substack{\text{quartz}\\(46\text{ cc.})}}{2SiO_2}\right.$$

$$\left.\rightleftharpoons \underset{\text{cordierite}\quad(462\text{ cc.})}{2(Mg,Fe)_2(Al,Fe)_4Si_5O_{18}} + 2\,KOH\right]$$

then

$$\left[2KOH + \underset{\text{andalusite}}{3Al_2SiO_5} + \underset{\text{quartz}}{3\,SiO_2} \rightleftharpoons \underset{\text{muscovite}}{2KAl_2Si_3AlO_{10}(OH)_2}\right]$$

The corroded staurolite and anthophyllite relics in the cordierite of rock 397 suggests the following reactions.

$$\left[3.2MgO \cdot 2.8FeO \cdot (Al,Fe)_2O_3 \cdot 7SiO_2 \cdot H_2O + \right.$$

gedrite (47% iron end member) (265 cc.)

$$3\ Al_2SiO_5.Fe(OH)_2 \rightleftharpoons$$

staurolite (201 cc.)

$$\left. 2(\ 2(Mg_{0.8}Fe_{0.2})O \cdot 2Al_2O_3 \cdot 5SiO_2\ ) + 5\ FeO + 4H_2O \right]$$

cordierite (462 cc.) (62.5 cc)

A reaction involving the formation of cordierite show a slight increase in rock volume, and are not favoured by increasing rock pressures.

In the lower grade rocks anthophyllite appears to form from chlorite and quartz in the following manner.

$$\left[7\ (Mg,Fe)_3\ Si_2\ O_5\ (OH)_4 + 10\ SiO_2 \rightleftharpoons \right.$$

chlorite quartz

$$\left. 3\ (Mg,Fe)_7\ Si_8O_{22}(OH)_2 + 11\ H_2O \right]$$

anthophyllite

Thus both quartz and chlorite become depleted adjacent to the newly formed anthophyllite prisms. The above reaction is a simplification, the actual reaction would involve aluminous chlorite and gedrite.

(2) Comparison with the Reedy Creek cordierite.

The cordierite nodules of Reedy Creek (Sando, 1957) have features which are reminiscent of some of the coarser cordierite knots of the present area. The Reedy Creek nodules are usually zoned in the following manner. An inner zone, consisting of idioblastic sillimanite prisms in parallel orientation and embedded in cordierite, is enclosed by a zone containing granular corundum, magnetite and spinel within cordierite. Sando considers that there are two kinds of nodules, one having a predominance of corundum and magnetite, within this silica deficient zone, the other spinel and magnetite. These inner zones are encased in a thick shell of cordierite, often showing a mozaic structure, consisting of sub-grains in near parallel orientation, and containing small idioblastic inclusions of biotite. Each nodule is encased within a quartz depleted, and biotite poor, outer zone of granoblastic plagioclase, which grades off away from the nodule into a quartz-plagioclase-biotite rock matrix. Sando considered that these nodules might have originated by the metamorphism of former clay pellets or as segregations within a favourable matrix. It now seems more probable that these peculiar nodules represent a further stage in the metamorphism of nodules similar to those of the present area. The parallel growth of sillimanite prisms in the cores of some nodules is important, for this indicates the presence of a pre-existing structural control. Evidence from the present area suggests that such a sillimanite orientation can result from the replacement of former andalusite during the dimorph phase change favoured by increasing temperature and pressure. Neither corundum of spinel have been observed within the cordierite knots of the present area. The presence of these silica deficient minerals at Reedy Creek, suggests that the cordierite there, continued to form at high (higher ?) temperatures, with the availability of sufficient Mg, Fe and Al from biotite and sillimanite, but a deficiency of silica, evidenced by a depletion of quartz around the nodules. This facilitated the breakdown of sillimanite to corundum and silica. Sando's observation that some nodules have chiefly corundum and magnetite, and others spinel and magnetite, suggests that the cordierite nodules have grown, in one case, from former andalusite porphyroblasts, and in the other, staurolite, the latter breaking down to sillimanite and spinel at sufficiently high temperatures (suggested by the experimental work of Halferdahl (1956) on chloritoid stability).

The lack of corundum and spinel in the cordierite rocks of the present area is probably a result of the smaller size of the original andalusite and staurolite porphyroblasts. Thus, the resultant cordierite knots, usually less than a centimeter in diameter, are too small to sustain an inner silica deficient zone.

(3) The significance of cordierite-garnet assemblages.

Chinner (1962) has shown that the assemblage cordierite-garnet may be used as an indicator of the geological pressures involved in metamorphism. Only one sample (505) containing fresh garnet and cordierite was found in the present area. The composition of the cordierite (19% iron molecule) is equal to the maximum iron content of cordierites in this area. This is the critical value for geobarometric considerations. Some granulites have a cordierite iron content less than 19% iron molecule, while the cordierites of contact metamorphosed rocks have a distinctly higher range of iron values (Chinner, 1959, 1962). The maximum iron content of cordierites from this area is consistent with the values expected for a regional Buchan-type metamorphism.

(4) The absence of cordierite in the kyanite zone.

The absence of cordierite rocks among the kyanite bearing schists in the north-west corner of the present area is in accord with the findings of Hietanen (1956, 1961), who, in the Boehls Butte quadrangle, Idaho, found that the field of cordierite stability may meet, but does not overlap, the field of kyanite stability near the triple point of the aluminosilicate diagram (8.5 Kb. and 350°C).

(5) The absence of a stable cordierite-andalusite assemblage.

The absence of the stable assemblage cordierite-andalusite in the present area is of considerable interest. This assemblage is quite stable in areas of lower pressure contact metamorphism. The stability curves for magnesium cordierite (after Schreyer and Yoder, 1959) and the aluminosilicate polymorphs (modified from the results of Bell, 1963 and Khitarov, Pugin, Pin Chao and Slutskii, 1963) are presented in P-T projection in Fig. 35. The aluminosilicate polymorph transitions are invariant with respect to P_{H_2O}, but this is not so for cordierite stability. The stability curve of cordierite has been derived under the condition of $P_{H_2O} = P_{total}$, a condition unlikely in natural metamorphism. This curve represents the highest temperatures at which cordierite might form in rocks of favourable composition. A second hypothetical curve has been drawn to indicate an allowance for $P_{H_2O} < P_{total}$, and the effects of the iron-cordierite molecule, which is not believed to be very great (Schreyer and Yoder (1959)). This curve is made to pass near the aluminosilicate triple point to satisfy the petrographic data of both the Boehls Butte quadrangle and the present area. The curve does not enter the kyanite field because of the entrance of a

FIG. 35.

The stability relationships of cordierite in P-T projection.

(1) Hypothetical curve for the formation of iron bearing cordierite by a reaction such as:

$$\left[\begin{matrix}\text{staurolite}\\ \text{andalusite}\end{matrix} + \text{biotite} + \text{quartz} \rightleftharpoons \text{Mg-Fe cordierite} + \text{vapour}\right] \text{ at } P_{H_2O} < P_t.$$

(2) Experimentally determined curve for the formation of magnesian cordierite by the reaction:

$$\left[\text{Mg cordierite} + \text{vapour} \rightleftharpoons \text{chlorite} + \text{quartz} + \text{"sillimanite"} + \text{vapour}\right]$$ (after Schreyer and Yoder, 1959).

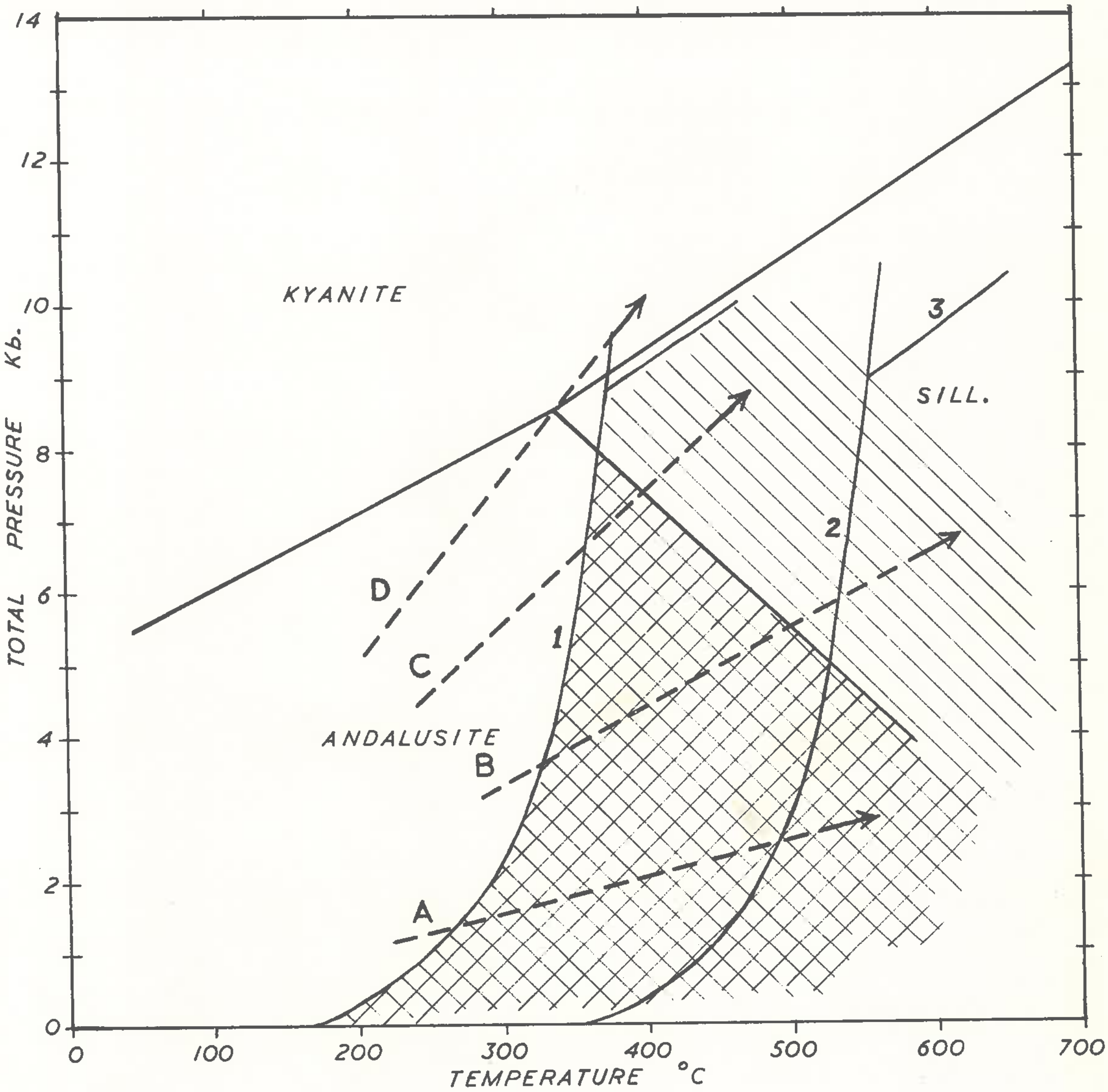
The curves for the aluminosilicate polymorphic phase changes are also shown for comparison.

Shaded area - P-T field in which cordierite may occur in equilibrium with an Al_2SiO_5 mineral.

Cross-hatched area - Field where cordierite-andalusite may be in stable equilibrium.

Curves A-D - possible P-T paths in metamorphism (see text).

FIG. 35



14
12
10
8
6
4
2
0
TOTAL PRESSURE Kb.
0
100
200
300
400
500
600
700
TEMPERATURE °C
KYANITE
ANDALUSITE
SILL.
A
B
C
D
1
2
3

new phase field at a pressure near 10 kb., the field of chlorite + quartz + "sillimanite" + vapour in the experimental studies of Schreyer and Yoder (1959). In nature, cordierite rocks may be replaced at the higher pressures of the kyanite field by a kyanite-gedrite paragenesis (Tilley, 1939), or by the assemblage kyanite-muscovite-biotite (Harker, 1954). Hietanen (1961) believes that the field of staurolite stability may be divided into two parts; a staurolite-kyanite field, and a staurolite-cordierite field. The conditions of metamorphism in the present area appear to be lying on the boundary of these two fields.

In Fig. 35 Curve A and Curve B depict possible metamorphic paths in contact metamorphism. These curves pass through a wide field in which andalusite and cordierite are stable in rocks of appropriate composition. In contact metamorphism both andalusite and cordierite appear so early that Harker (1939) considered that both of the minerals were unsuitable for the definition of metamorphic isograds. Curve C depicts the metamorphic path for the rocks on the eastern side of the present area. This curve illustrates the formation of andalusite earlier than cordierite, and the appearance of cordierite just before the entrance into the sillimanite zone. The assemblage andalusite-cordierite could be stable only under a very limited P-T range. At higher temperatures the curve enters the field of cordierite-sillimanite stability. Curve D illustrates the metamorphic path for rocks from the north-west corner of the present area. Andalusite is stable in the earlier rocks. The curve passes near the triple point, allowing the appearance of sparse kyanite and then may pass into part of the sillimanite field where cordierite is not stable.

In the type area of Buchan metamorphism in the Ythan Valley of Aberdeenshire, Read (1952) found a stable association of cordierite and andalusite even in the lowest grade knotted mica phyllites. Within the sillimanite zone andalusite in the cordierite assemblage is replaced by sillimanite yielding the assemblage cordierite-sillimanite-mica-felspar-(almandine). Fyfe, Turner and Verhoogen (1958) regard this assemblage as transitional from the hornblende hornfels facies of contact metamorphism to the almandine amphibolite facies of regional metamorphism (see also Oki, 1961). Similar assemblages have recently been described from the Westport gneisses of Ontario (Wynne-Edwards and Hay, 1963) and from the Bodenmais sulphide deposit, Bavaria (Schreyer, Kullerud and Ramdohr, 1964). The occurrence of the assemblage andalusite-cordierite in the lower metamorphic grades in Aberdeenshire indicates that the present area has been metamorphosed under a higher pressure gradient than the Buchan type area. Curve B in Fig. 35 may be considered to represent the P-T path of metamorphism of the Aberdeenshire rocks. It is concluded that the metamorphism of the present area has followed a P-T path intermediate between that of Buchan type metamorphism (Miyashiro's low pressure intermediate facies series) and Dalradian metamorphism (Tilley, 1925).

(6) Turner's facies classification.

Turner (1948, 1951) introduced a cordierite-anthophyllite subfacies within the amphibolite facies. However, the relative rarity of rocks containing cordierite and anthophyllite, the doubtful isochemical metamorphic status of these minerals in some areas, and the wide P-T range over which they are stable, has made this subfacies obsolete. Turner (1960, p.552) considers that cordierite-sillimanite and cordierite-almandine assemblages are transitional from the hornblende-hornfels facies to the almandine amphibolite facies.

(7) Origin of the cordierite and anthophyllite rocks.

The rarity of sediments of a composition suitable for the development of anthophyllite and cordierite-anthophyllite assemblages has lead many workers to the conclusion that the occurrence of these rocks is related to a strong metasomatic influence. In the classic Orijärvi region of south-west Finland Eskola (1914, 1915) has attributed anthophyllite bearing rocks to the magnesium metasomatism of original leptites (acid lavas and tuffs). Despite the inherent difficulties involved in widespread magnesium and iron metasomatism, a similar conclusion was reached by Tilley (1935) for the anthophyllite rocks of the Kenidjack and Botallack areas, Cornwall, after a consideration of various other possible origins (Tilley and Flett, 1930) and a visitation to the classic Orijärvi area. The magnesium metasomatism hypothesis has also been adopted by Osborne (1939) for the anthophyllite-cordierite gneisses of Montauban, Quebec. Tilley (1937, p.307) considers that an anthophyllite-staurolite-cordierite assemblage from the Lizard, Cornwall, has resulted from the regional metamorphism of argillaceous sediments. Simpson (1938) and Prider (1940) have described cordierite-anthophyllite rocks, from Western Australia, formed through the metamorphism of contaminated hypersthenic rocks.

Some doubts concerning the origin of some of the anthophyllite-cordierite and anthophyllite rocks of the present area have already been expressed. It is possible that some of these rocks have been isochemically metamorphosed, but the relative abundance of anthophyllite rocks in the western part of the area, and their stratigraphic relations, suggests a strong metasomatic influence. There is, however, no necessity to consider magnesium or iron metasomatism in the formation of these rocks, for these elements were probably already present in former biotite. All that is necessary is a metasomatic leaching-out of potash. The manner in which this process may occur is outlined in Chapter 10.

CHAPTER 6.

METAMORPHISM OF LIMESTONES.

Introduction.

Among the metamorphic rocks of the Cambrai region, marbles are the most distinctive, and, over most of their exposure, may be readily traced out on aerial photographs. The marble beds crop out almost continually on the eastern scarp of the ranges, where their massive nature resists mechanical disintegration, and promotes the formation of prominent ridges deeply dissected by the steep rugged ravines of cross-strike consequent streams. On the more mature uplands to the west, where mechanical weathering is less effective, the marbles crop out poorly, generally forming residual hills, large whalebacks or low pavements, partially or wholly surrounded by soil flats. Although forming an intricate outcrop pattern in plan, detailed structural mapping has indicated that the marbles comprise a single stratigraphic horizon. The lower grade marbles are characterized by abundant intercalations of phyllitic schist, representing mud seams in the original carbonate sediment. The proportion of silicate minerals in the marble varies from 5-50%, but usually averages about 30%, and "pure" carbonate marbles do not occur. These marbles have been examined by the South Australian Geological Survey as a potential source of carbonate rock for commercial use. About 3 miles south of the present area near Milendella "a thin, medium grained pale-grey, somewhat micaceous marble was sampled over a width of 110 ft., having an analysis as below:

$CaCO_3$	$MgCO_3$	SiO_2	Fe_2O_3	Al_2O_3	P_2O_5
79.3%	0.82%	9.10%	1.17%	7.25%	0.05% "

(Johns, 1963, p.31). The continuation of this marble into the marbles of the present area was mapped but not sampled. Mechanical analyses of selected typical low grade and high grade marbles from the present area are presented in Table 6. Occasional interbeds and border zones containing more than 50% silicate minerals are here termed "calc-silicates", and are similar in all respects to the calc-silicate beds interbedded with quartzo-felspathic schists above and below the marble horizon. These rocks will be considered in a subsequent section.

Even strongly reconstituted marbles show some compositional inhomogeneity mimicing the argillaceous seams in the original sediment. These seams are usually well preserved as phyllitic schist bands at low metamorphic grade, although they may be highly folded. With increasing metamorphic grade this original foliation is progressively destroyed by increasing rates of grain growth and the appearance of new minerals, but, even in the high grade coarse grained marbles, scapolite and clinopyroxene poikiloblasts outline relics of the original layering. This effect may be especially observed in the cores of small folds. There is no tendency within these rocks for the formation of new metamorphic layering.

TABLE 6.

MECHANICAL ANALYSES OF SELECTED TYPICAL MARBLES.

Low grade marble. Specimen 30. Location 220-214.

	Weight percent.
Carbonate (calcite)	66.0
Scapolite, quartz	19.3
Muscovite	4.4
Biotite	7.1
Potash felspar	2.5
Pyrite	0.7
	100.0

High grade marble. Specimen 18. Location 194-086.

	Weight percent.
Carbonate (calcite)	66.6
Scapolite	14.6
Diopside	10.1
Potash felspar	8.3
Phlogopite	0.3
Sphene	0.1
Hornblende	0.03
	100.03

The major penetrative structure in the metamorphic limestone is the axial plane cleavage of the first and major deformation. This foliation is clearly outlined by the shape orientation of "flattened" calcites and, especially in the lower grade rocks, by a strong alignment of the sheet silicates. A pronounced lineation, a ribbing or microfolding, within this foliation is the result of the intersection of the axial plane cleavage with the argillaceous seams and the original bedding laminations. There are no lineations in the marbles caused by mineral elongations, and the calcite grains may be shown to be "flattened" but not elongate in form. The superposition of a rarer later foliation, equivalent to the crenulation cleavages in the adjacent pelitic schists, has the effect of reorienting the former rock fabric. More details of the structural geometry of the marbles will be presented in a later section.

In the crystalloblastic marbles of the western belt, the well oriented calcite fabric shows little post-crystallizational strain. Weak twinning of the calcite is attributed to a late faulting stage (Chapter 15). However, post-crystalline strain is much more pronounced in the eastern belt of marbles where deformation associated with both folding and faulting outlasted recrystallization. Here bent and twisted twin lamellae are numerous in all calcite grains, and polygonization recrystallization (annealing) has resulted in mortar and "mylonitized" augen textures. These textures have

FIG. 36A.

General view of marble exposure looking south from 212.154.

FIG. 36B.

Outcrop of marble with strongly folded compositional layering. Coin diameter 2.8 cm. Location 165.183.

FIG. 36A

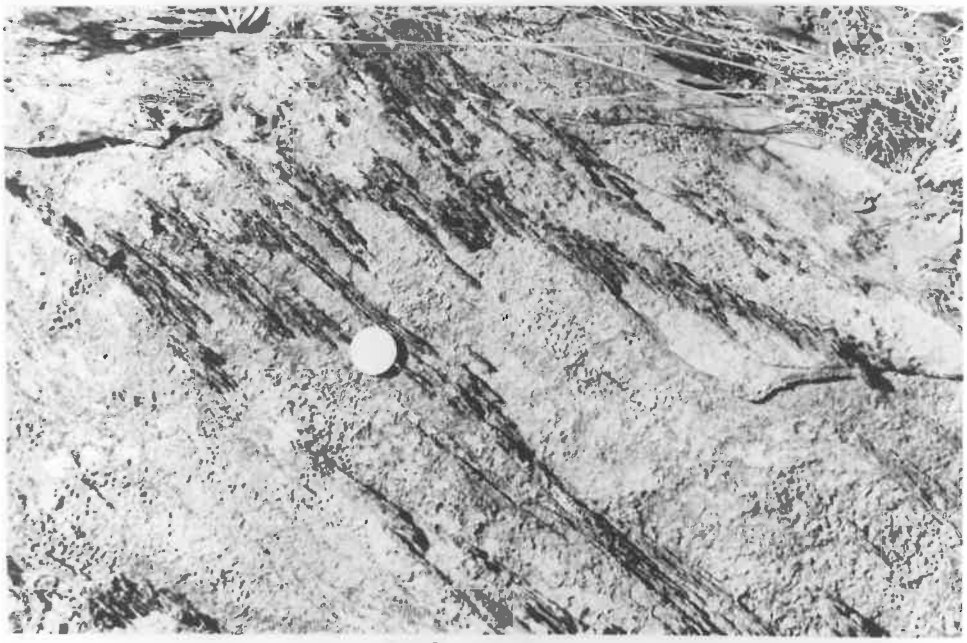


FIG. 36B

been described also in the marbles of Milendella, a few miles to the south (White, 1959). The silicate minerals also show strong evidence of strain. Quartz and calcite have undulose extinction, phlogopite flakes are micro-folded, clinopyroxene grains have become cloudy and have numerous $\{100\}$ polysynthetic twin lamellae sub-parallel to the axial plane foliation, and all the brittle silicate minerals may show numerous tension fractures perpendicular to the axial plane foliation.

The minerals of the primary paragenesis observed in the more common silica-potash sufficient marbles are quartz, calcite, plagioclase, scapolite, biotite-phlogopite, muscovite, potash felspar, clinopyroxene and grandite (grossular-andradite), with accessory tourmaline, pyrite, apatite and sphene. In addition in the potash deficient rocks chlorite, actinolite, hornblende and brucite may occur.

A secondary paragenesis resulting from the retrograde metamorphism of the higher grade assemblages consists of actinolite, hornblende, epidote, calcite, quartz, muscovite and grandite(?) (see also White, 1959). These secondary assemblages are commonly developed in those marbles showing mortar structure and annealing of the originally highly strained calcite fabric.

As with the pelitic rocks, it has been found useful to treat the metamorphism of the eastern and western marbles separately.

A. Marbles of the Eastern Part of the Area.

The eastern marbles have notable changes in grain size, texture and mineralogy with increasing metamorphic grade. The primary assemblages observed, in order of appearance with increasing grade, are:

Biotite in pelitic schists.

(1) Quartz-calcite-plagioclase-biotite-muscovite.

Biotite-andalusite in pelitic schists.

(2) Quartz-calcite-plagioclase-biotite-muscovite-potash felspar.

Biotite-andalusite-staurolite in pelitic schists.

(3) Quartz-calcite-plagioclase-biotite-muscovite-potash felspar-scapolite.

Anthophyllite, cordierite and sillimanite appear in pelitic schists. Diopside forms from amphibole in some calc-silicate rocks.

(4) Quartz-calcite-biotite-potash felspar-diopside-scapolite.

(5) Quartz-calcite-plagioclase-phlogopite ($\pm$ diopside, scapolite, potash felspar).

Potash felspar-sillimanite and potash felspar-cordierite assemblages stable in pelitic schists.

(6) Quartz-calcite-plagioclase-phlogopite-diopside-scapolite-potash felspar-grandite.

(7) Quartz-calcite-plagioclase-diopside- ($\pm$ scapolite, potash felspar).

(8) Quartz-calcite-plagioclase-phlogopite-muscovite-scapolite.

All assemblages may have the accessory minerals sphene, apatite, tourmaline or pyrite.

It must be noted that all of these assemblages have excess quartz, and many have excess potash as potash felspar or muscovite. These characters reflect the original argillaceous nature of the sediment. Because of the large number of chemical components in these argillaceous limestones it is difficult to represent the assemblages on a metamorphic facies diagram. Francis (1958) has summarized assemblages in the calcareous hornfelses of Glen Urquhart in sub-tetrahedra of the ACKF tetrahedron. His classification, although fitting the Glen Urquhart rocks, does not allow the assemblage potash felspar-diopside, and is therefore inapplicable here. An attempt to represent a lower and a higher grade facies in an ACKF tetrahedron is presented as Fig. 37.

(a) Descriptive Petrography.

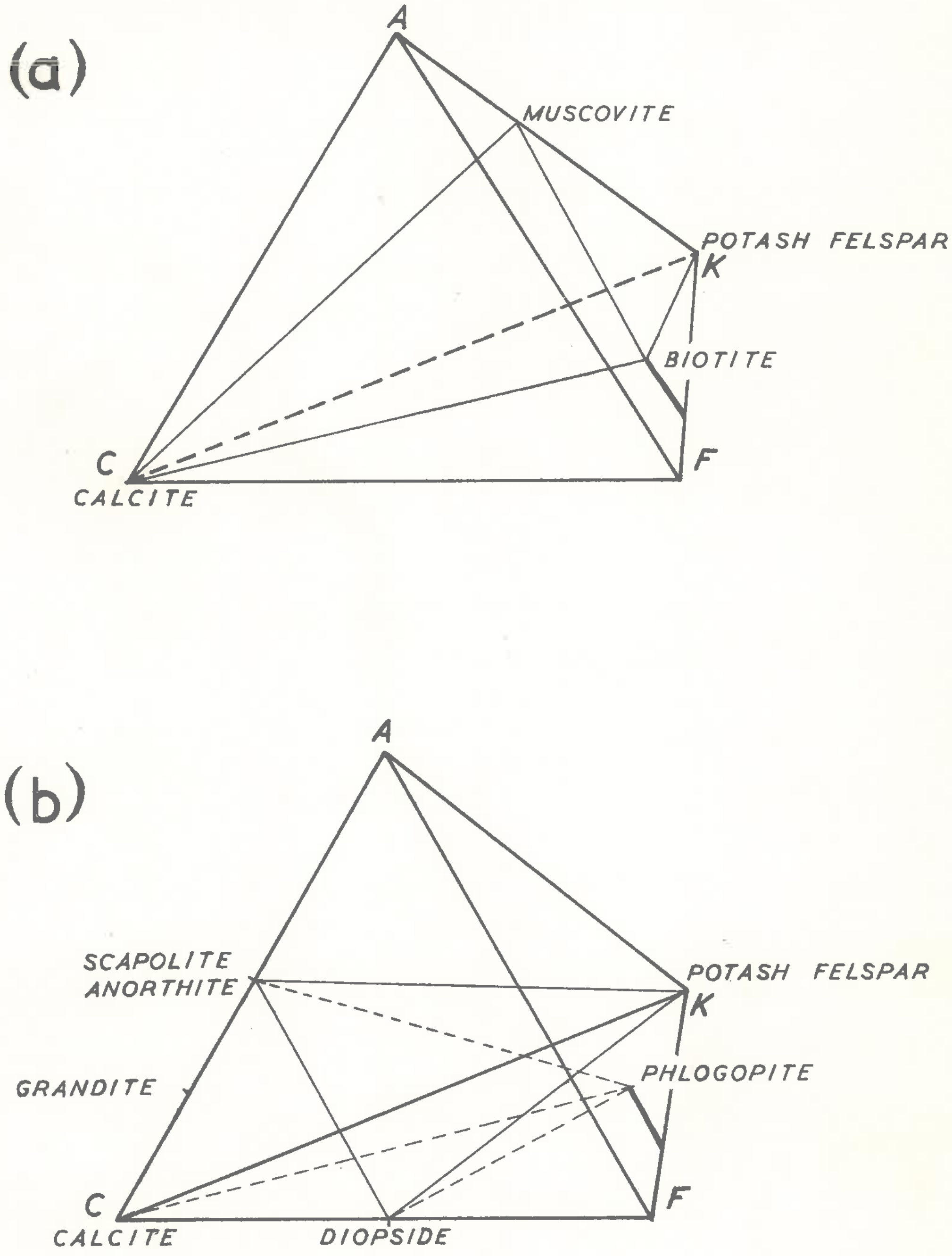
(1) Biotite zone.

The marbles in the biotite zone of regional metamorphism, before the appearance of andalusite and staurolite in the aluminous pelitic schists, are very fine-grained and light grey in colour. Coarser grained (0.10 mm.) lighter bands consisting largely of calcite are interspersed with thin grey silky phyllitic seams composed of muscovite, biotite, quartz

FIG. 37.

ACKF diagrams for marbles showing mineralogical equilibrium at a lower grade (a) and a higher grade (b). SiO_2 and Na_2O are considered to be in excess and quartz and albite may be additional phases.

FIG. 37


(a)
A
MUSCOVITE
POTASH FELSPAR
K
BIOTITE
C
CALCITE
F
(b)
A
SCAPOLITE
ANORTHITE
POTASH FELSPAR
K
PHLOGOPITE
GRANDITE
C
CALCITE
DIOPSIDE
F

and sodic plagioclase with an average grain size of 0.02 mm. The phyllitic seams are strongly folded. The folds have an axial plane cleavage defined by calcite grain flattening (caused in part by twin gliding), and a strong dimensional orientation of minute biotite flakes in the argillaceous seams. Calcite, composing 80-95% of the rock, occurs as lens shaped grains having elongations in section of 2:1. Some twin lamellae are visible in the calcite. Quartz and felspar grains, generally too fine for optical tests, are abundant in the argillaceous bands. Biotite, pleochroic X-pale lemon to YZ-mid-red-brown, is also concentrated in the argillaceous seams. Muscovite forms rare stumpy partially oriented flakes in the carbonate rich portions, and larger (0.2 mm.) porphyroblasts in the argillaceous seams. Accessory sphene is common as fine granules and dust.

(2) Andalusite-staurolite zone.

No new minerals appear in the marbles until the adjacent pelitic schists contain abundant andalusite and staurolite. By this stage the grain size of both the carbonate and argillaceous layers of the marble has increased ten fold by volume. Then numerous small spots appear, in which muscovite and calcite grains reach 1 mm. in diameter. Within these spots skeletal scapolites first occur. Concomitant with the growth of scapolite is the appearance of potash felspar and more calcic plagioclase, and the destruction of muscovite.

(3) Sillimanite-muscovite zone.

After cordierite and anthophyllite have grown in the potash deficient pelites, and at the grade where sillimanite is first found as tufts associated with muscovite in the more aluminous pelitic schists, clinopyroxene makes its first appearance in the marbles. This clinopyroxene forms by a reaction which removes much of the initial biotite, and increases the content of potash felspar. With the growth of the new minerals scapolite and diopside, grain size increases rapidly. Both scapolite and diopside may form large skeletal xenoblasts, to 4 mm., which usually replace relics of the more argillaceous seams in the marble. The growth of clinopyroxene does not remove biotite entirely; some phlogopite always remains. In the pyroxene-scapolite marbles, quartz, non-twinned potash felspar, and plagioclase (composition near An 40) are present as optically determinable granoblasts embedded in a matrix of calcite with an average grain size of 0.3 mm.

(4) Sillimanite-Potash Felspar zone.

Except for the local appearance of grossular-andradite in the marble near the Kanappa Mine, no new

FIG. 38A.

Photomicrograph of marble from eastern side of the area. Calcite shows well developed deformation twins. Other minerals present are quartz, felspar and biotite.

Rock A185-807a. Location 204.148. Ordinary light.

FIG. 38B.

Photomicrograph of marble from eastern side of the area showing strong annealing recrystallization at grain boundaries.

Rock A185-802. Location 202.129. Ordinary light.

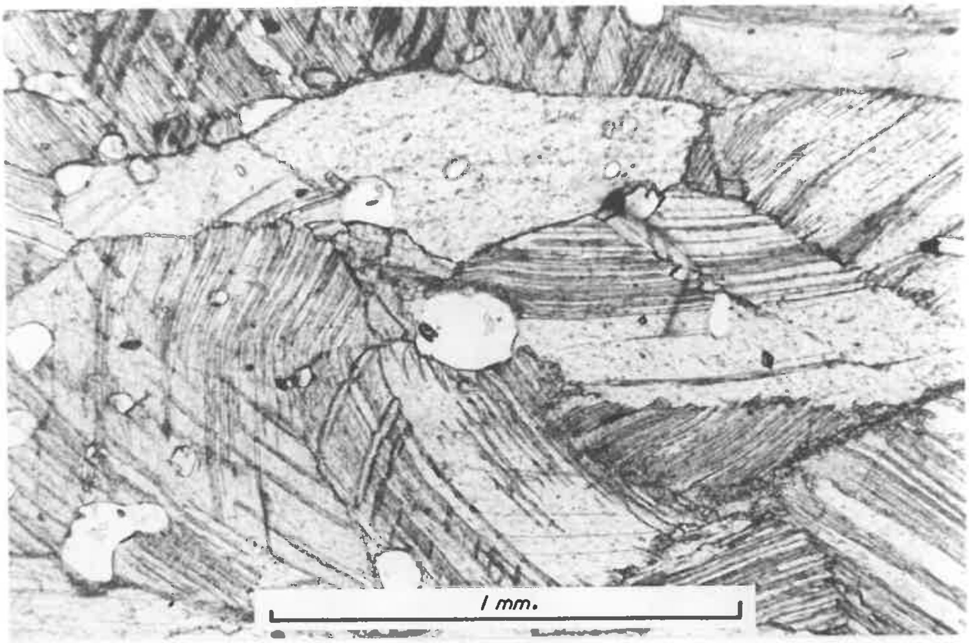


FIG. 38A

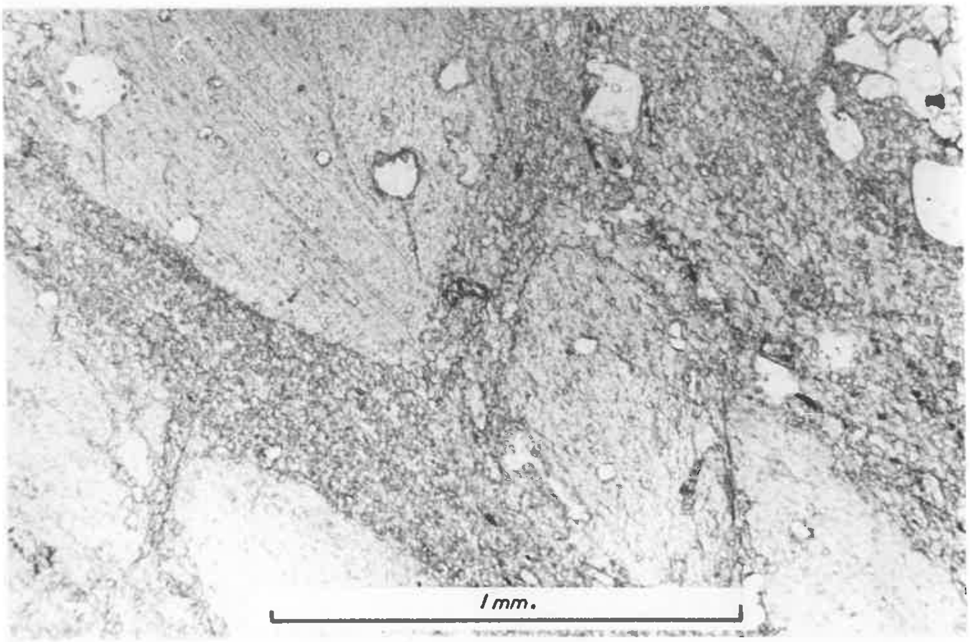


FIG. 38B

minerals appear within these rocks up to the highest metamorphic grades where sillimanite is stable with potash felspar and cordierite in the pelitic schists, and andalusite and staurolite have been completely destroyed. However, several progressive changes in both the texture and the mineralogy of the marbles can be observed in this high grade interval. The grain size progressively increases until the calcite grains reach 1 mm. or more. With increasing grade the early formed large scapolite and clinopyroxene poikiloblasts become strained, recrystallized to a mozaic of polygonal grains, broken apart, and eventually recrystallized as clear, inclusion-free granoblasts interspersed with quartz, potash felspar and plagioclase in the calcite matrix. The well oriented phlogopite flakes in the early scapolite-pyroxene marbles become enlarged and more isolated with increasing grade, and eventually form large skeletal xenoblasts to 1 mm. in diameter enclosing quartz and calcite inclusions.

(5) Minerals of a secondary Paragenesis.

In a coarse-grained diopside-scapolite-phlogopite marble at the Kanappa Mine some pale pink garnets have been rarely observed. These occur in webs to 2 mm. enclosing quartz, calcite and felspar inclusions. Because of the absence of primary epidote in the lower grade marbles, and the occurrence of some secondary epidote in the garnet bearing rock, there is some doubt that this garnet belongs to the primary assemblage, and it may form part of the secondary paragenesis described by White (1959).

This secondary paragenesis is well represented in the higher grade marbles, especially those which have suffered some annealing recrystallization of their highly strained calcite. Secondary blue-green hornblende may form by epitaxic replacement of the clinopyroxene, as described by White, 1959. Individual crystals of actinolite, along with skeletal muscovite, may appear within the polygonized granular aggregates of calcite. The secondary nature of these minerals is shown by the juxtaposition of calcite and quartz. Epidote, usually in a poorly recrystallized form showing variable birefringence, occurs as an alteration product of scapolite. The destruction of the marialite molecule of the scapolite may contribute to the sodic rims on some plagioclase grains. Some secondary calcite and quartz may accompany the retrograde changes, the quartz having a tendency towards idioblastic growth. In many cases there appears to be a close tie between the occurrence of the secondary paragenesis and the granular aggregates of polygonized calcite, but it is not clear whether these processes are contemporaneous. As there is a direct correlation between the degree of strain of the original primary calcite and the amount of polygonization, this recrystallization may be attributed to annealing (Griggs, Paterson, Heard and Turner, 1958). All gradations may be observed from the growth of minute granules of secondary calcite on the grain boundaries in weakly strained calcite fabrics, through mortar structures in which large highly strained primary calcite grains are wholly surrounded by a polygonized matrix, to "mylonitized" structures in which anastomosing streams of granules weave around floating relics of highly twisted primary calcite. The degree of polygonization also increases with metamorphic grade, suggesting a weak temperature gradient at the time of annealing.

The muscovites in the marble specimen yielding assemblage (8) have been bent and folded in a similar manner to the associated phlogopite flakes. Here the muscovite appears stable even though calcite and quartz are present. In this high grade marble it is supposed that locally high P_{H_2O} may have allowed the preservation of the muscovite.

(b) Progressive Metamorphic Reactions.

(1) Decomposition of muscovite.

It is evident from the above description of the metamorphic evolution of the limestones that several important reactions are involved in the progressive metamorphism. The first of these involves the formation of scapolite and the destruction of muscovite. The following reaction is proposed:

$$\left[\underset{\text{muscovite}}{K_2O.3Al_2O_3.6SiO_2.2(H,Cl)_2O} + \underset{\text{calcite}}{2CaCO_3} + \underset{\text{quartz}}{4SiO_2} \rightleftharpoons \right.$$

$$\left. \underset{\text{orthoclase}}{K_2O.Al_2O_3.6SiO_2} + \underset{\text{anorthite}}{2CaO.Al_2O_3.2SiO_2} + 2CO_2 + 2(H,Cl)_2O \right] \quad ....(1)$$

Yoder and Eugster (1955) have derived a univariant equilibrium curve for the decomposition of muscovite. The decomposition temperature is lowered some 50°C by the presence of quartz (Turner and Verhoogen, 1960, p.134) and presumably, in the presence of both quartz and calcite at a lower temperature still (Harker, 1932 p.254). Thus muscovite is destroyed and potash felspar and the anorthite molecule appear. The anorthite may, in part, calcify the plagioclase already present, and in the presence of the volatiles Cl, F and SO_2, either in the pore solutions or released from the muscovite, scapolite may be free to form. There is no evidence for the introduction of volatiles for the formation of scapolite. The findings of the present author are in agreement with the conclusions of White (1959) that the scapolite in this metamorphic belt has formed through the isochemical metamorphism of an original chlorine bearing sediment.

(2) Decomposition of biotite.

A second major reaction occurs with the formation of clinopyroxene. Ramberg (1952, p.152) believed that the following reaction was important in the formation of pyroxene from biotite in marbles.

$$\left[\underset{\text{phlogopite}}{KMg_3AlSi_3O_{10}(OH)_2} + \underset{\text{calcite}}{3\ CaCO_3} + \underset{\text{quartz}}{6\ SiO_2} \rightleftharpoons \underset{\text{potash felspar}}{KAlSi_3O_8} + \underset{\text{diopside}}{3\ CaMgSi_2O_6} + 3\ CO_2 + H_2O\right] \quad \ldots\ldots\ldots(2)$$

However, the reaction cannot be as simple as this, for biotite is not entirely destroyed with the appearance of diopside, and the assemblage phlogopite-calcite-quartz is stable to the highest grades of metamorphism observable in this area. This is also in accord with the findings of Francis (1958). Evidently the reaction has the form

$$\left[\text{biotite I} + \text{calcite} + \text{quartz} \rightleftharpoons \text{biotite II} + \text{diopside} + \text{potash felspar} + CO_2 + H_2O\right] \quad \ldots\ldots\ldots(3)$$

where biotite II has a higher Mg/Fe ratio. In order to test this, refringence determinations on a number of biotites and pyroxenes were made, and it was found that there were progressive composition changes in both biotite and pyroxene with increasing metamorphic grade.

Until clinopyroxene first appears in the marbles the biotite usually has a strong pleochroic scheme X-very pale lemon Y-Z-orange-brown or orange-red-brown or sepia-orange-brown. After the appearance of clinopyroxene the pleochroic scheme changes progressively through

X-pale yellow	Y-Z-pale brown with a yellow tinge
to X-very pale lemon	Y-Z-deep yellow
to X-colourless	Y-Z-pale lemon with brownish tinge.

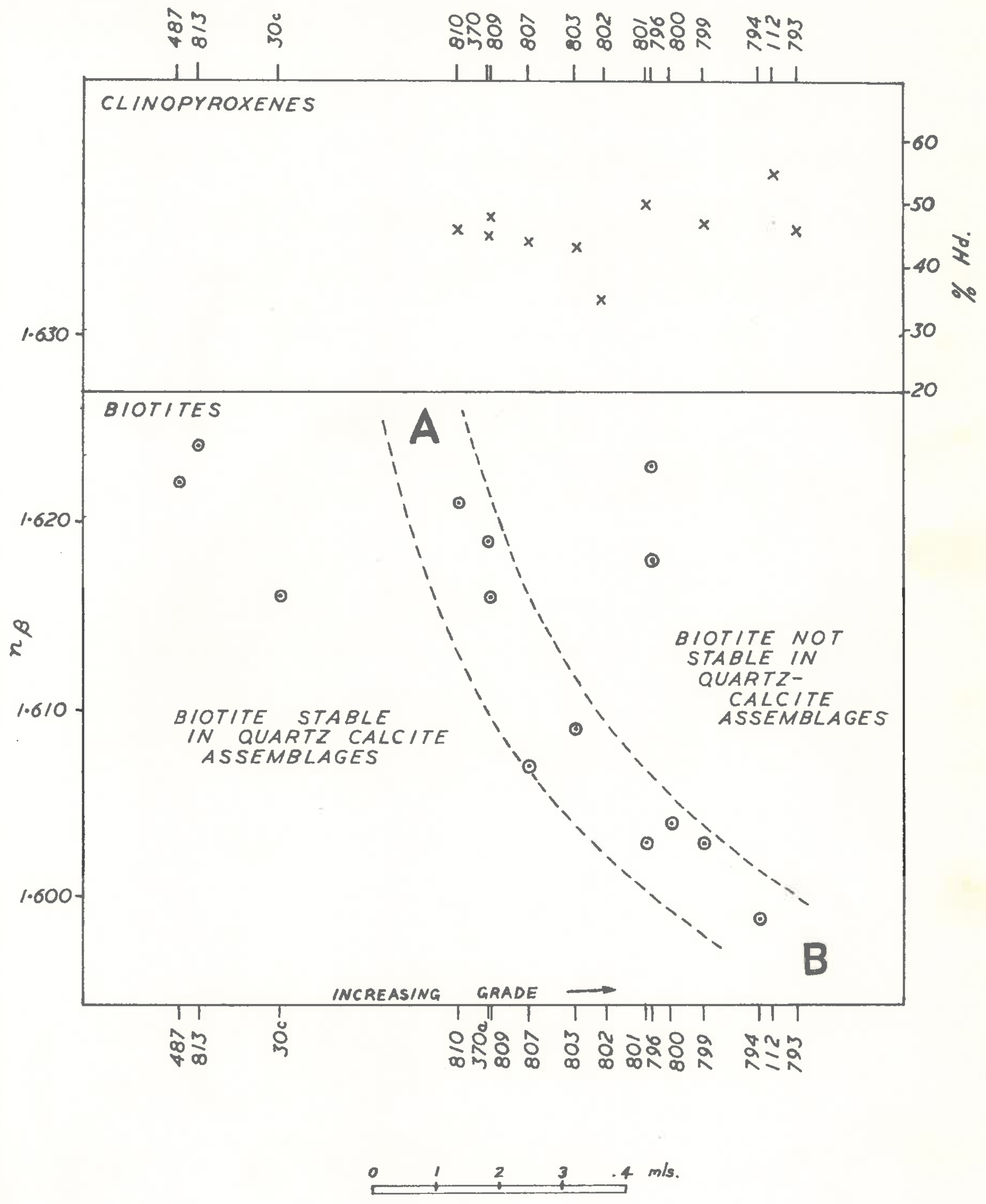
Accompanying these colour changes the β refractive index decreases from 1.624 to 1.599 (15 measurements) indicating a composition change from 35% annite to 15% annite, assuming phlogopite-annite end members only (Eugster and Wones, 1958; Wones, 1963) or approximately 20% iron molecule to approximately 10% iron molecule in the phlogopite-biotite series (Heinrich, 1946). Both pleochroic scheme and refractive index changes suggest a progressive increase of Mg/Fe of the biotite with grade. The associated clinopyroxenes have β refractive indices ranging from 1.697 to 1.710 (10 readings) indicating a composition range from 35-55% hedenbergite molecule (Hess, 1949).

The results are recorded in Appendix III and are graphed in Fig. 39. This graph shows the refractive indices of the biotite and clinopyroxene plotted against "increasing grade" measured as a function of the linear horizontal distance in a north-south direction (scale 1 cm. = 1 mile). This line of increasing grade along which the specimen numbers have been plotted, may be considered as an axis of increasing temperature. If we ignore the three Kanappa Mine specimens, which seem to have abnormal biotite properties, there is a regular relationship between the biotite and clinopyroxene compositions.

This figure is essentially a T-X diagram. To the left of the band AB, biotite is stable in quartz-calcite assemblages, but not clinopyroxene. To the right of the band AB (higher temperature), biotite is not stable in quartz-calcite assemblages. Within the band AB lie biotite compositions in equilibrium with pyroxene. Thus, iron-rich biotites such as 813 would remain stable with increasing temperature in a calcite-quartz milieu until the band AB was reached, when pyroxene would begin to form. With further temperature increase the biotite would become progressively more magnesian, following the band AB towards B. A more magnesian original biotite, for example 30c, would first form pyroxene at a slightly higher temperature.

FIG. 39.

Biotite and clinopyroxene composition variation in marbles from the eastern side of the area, with increasing grade of metamorphism. The measured data is recorded in Appendix III. For explanation see text.



487
813
30c
810
370
809
807
803
802
801
796
800
799
794
112
793
CLINOPYROXENES
60
50
40
30
20
% Hd.
1·630
BIOTITES
A
B
1·620
1·610
1·600
nβ
BIOTITE STABLE IN QUARTZ CALCITE ASSEMBLAGES
BIOTITE NOT STABLE IN QUARTZ-CALCITE ASSEMBLAGES
INCREASING GRADE
370a
0
1
2
3
4
mls.

In rock 807 pyroxene is just beginning to form, the original biotite being more magnesian still. Rock 794, although containing phlogopite, has no pyroxene. The original biotite in this rock was presumably magnesian rich from the start, and the biotite has only just reached the temperature where it might become unstable and yield pyroxene. Thus the temperature of the reaction forming clinopyroxene from a biotite-quartz-calcite assemblage depends on the original biotite composition, and on the greater thermal stability of magnesium rich biotites compared with their iron rich analogues. (Yoder and Eugster, 1954; Eugster and Wones, 1962).

The abnormal properties of the biotites in the Kanappa Mine marbles may be related to

(1) Higher $Ti^{III or IV}$ or Fe^{III} contents.

(2) Higher P_{H_2O} restraining the reactions forming pyroxene.

(3) A different P_{O_2} value affecting the stability of iron biotite.

The second hypothesis is favoured since, as considered previously, it is likely that P_{H_2O} has been locally high in the Kanappa Mine area.

In the marbles, MgO and FeO are essentially acting as separate components, and the biotite-clinopyroxene reaction relations must be shown on an ACFM tetrahedron (Fig. 40). The grade at which diopside first appears depends on the Mg/Fe ratio of the original biotite, which reflects directly the initial Mg/Fe ratio of the sediment. The first pyroxene to appear with the decomposition of biotite will presumably have an Mg/Fe ratio similar to that of the biotite (Fig. 40a). Then with a further increase of temperature the Mg/Fe ratio of the biotite increases, and the Mg/Fe ratio of the pyroxene may decrease (Fig. 40b). However, once the biotite is almost used up, the Mg/Fe ratio of the clinopyroxene must approach that of the original biotite. In the highest grade marbles 112 and 793 clinopyroxene is present, but biotite has disappeared entirely.

The composition of the phlogopite in equilibrium with diopside-quartz-calcite could be unique if the pressure variables P_t, P_{H_2O}, P_{O_2}, P_{CO_2} could be fixed. Because of the difficulty of fixing the pressure variables, it is unlikely that this reaction relation could be used as a geothermometer. The composition of clinopyroxene in such an assemblage could be variable, depending largely on the original biotite composition.

(3) Formation of garnet and secondary paragenesis.

If the grandite of the Kanappa Mine marble is primary, it may have formed from original epidote

FIG. 40.

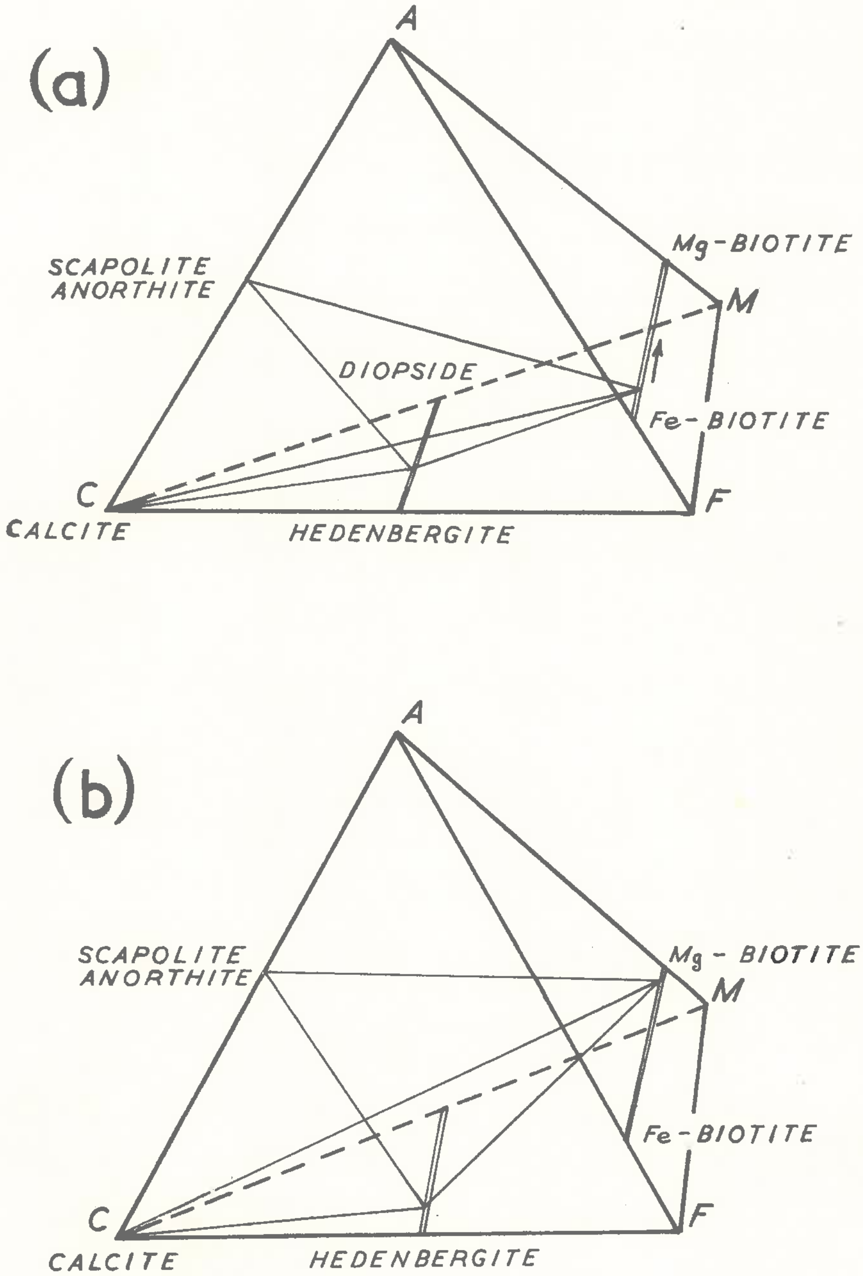
ACFM diagrams for marbles.

(a) Mineralogical equilibrium relationships in intermediate grade marble in which iron-rich biotite is stable.

(b) Mineralogical equilibrium relationships in high grade marble where only magnesium-rich biotite is stable.

SiO_2, Na_2O and K_2O may be in excess so that quartz, albite and potash felspar are possible additional phases.

FIG. 40


(a)
A
SCAPOLITE
ANORTHITE
DIOPSIDE
Mg-BIOTITE
M
Fe-BIOTITE
C
CALCITE
HEDENBERGITE
F
(b)
A
SCAPOLITE
ANORTHITE
Mg-BIOTITE
M
Fe-BIOTITE
C
CALCITE
HEDENBERGITE
F

thus:

$$\left[Ca_4(AlFe)_5Si_6O_{25}.H_2O + 5\,CaCO_3 + 3\,SiO_2 \rightleftharpoons \right.$$

epidote calcite quartz

$$\left. 3\,Ca_3(AlFe)_2Si_3O_{12} + 5CO_2 + H_2O \right]$$

grandite ....(4)

In the absence of any primary epidote in the marbles, it is perhaps more probable that this garnet is secondary and has formed through Fe metasomatism, or by a secondary reaction between meionite and pyroxene as suggested by White (1959, p.304).

The formation of the secondary actinolite and epidote paragenesis depended on the availability of water and CO_2 in the late stage fluids. The following reactions were no doubt important.

$$\left[5\,Ca(Mg,Fe)(SiO_3)_2 + H_2O + 3\,CO_2 \rightleftharpoons \right.$$

clinopyroxene

$$\left. Ca_2(Mg,Fe)_5Si_8O_{22}(OH)_2 + 3\,CaCO_3 + 2\,SiO_2 \right]$$

actinolite calcite quartz

..........(5)

$$\left[3\,(CaAl_2Si_2O_8).CaCO_3 + H_2O \rightleftharpoons \right.$$

meionite

$$\left. 2\,Ca_2Al_3Si_3O_{12}(OH) + CO_2 \right]$$

clinozoizite ....(6)

$$\left[3\ (NaAlSi_3O_8).NaCl + H_2O \rightleftharpoons \underset{\text{albite (rims on plagioclase)}}{3\ NaAlSi_3O_8} + \underset{\text{solution}}{NaCl + H_2O}\right] \qquad \text{(marialite)} \qquad \ldots\ldots\ldots(7)$$

(c) Mineralogy of the Eastern Marbles.

Calcite is the only carbonate mineral observed in these marbles and is stable with quartz at all grades of metamorphism. From the biotite zone to the mid-sillimanite zone the calcite has progressively increased in grain size over 1000 times by volume. Small granoblasts of quartz and plagioclase were present in most specimens. The fine grain size of the plagioclase in the lower grade marbles makes composition determinations difficult. The composition of 5 plagioclases from scapolite bearing marbles from various parts of the sillimanite zone fell in the narrow range An 37-An 39. It is concluded that

(1) plagioclase does not increase in anorthite content above the scapolite "isograd", in accord with reaction (1) above, the only anorthite producing reaction.

(2) the Na/Al ratio in the original sediment lay within a narrow range.

The first scapolite to appear in the marbles (specimen 30c location 220.214) has $n_o = 1.567 \pm .001$, indicating a composition near Me 45 (Fig. 43). In the low grade marbles muscovite has a low optic axial angle, $2V_x$ 30-36°, suggesting the presence of some phengite molecule. Potash felspar within the marbles never shows cross-hatched twinning. In one specimen $2V_x$ ranged from 61-64°. White has recorded $2V_x$ as low as 45° in potash felspar of the Milendella marbles. The properties of biotite and clinopyroxene have been described above. The clinopyroxene often shows strong strain effects resulting in polysynthetic twinning on $\{100\}$ (see also Drever, 1939) and undulose extinction, and it was found that measured $2V_z$ values varied from 34-61° across a single xenoblast in one rock.

Among the accessories, sphene and apatite are the most common. Isolated grains of tourmaline are very rare, but nevertheless, widespread in occurrence. The unusual pleochroic scheme E-colourless and O-rich indigo blue or sky blue is distinct from the green and yellow colours of the accessory tourmalines of the argillaceous and arenaceous metasediments. It is concluded that these tourmalines have grown during metamorphism, using up traces of boron trapped within the original sediment.

Notable absences within these marbles are the epidote minerals (and hence the rarity of grandite garnets), wollastonite and vesuvianite, characteristic of lower pressure environments. Because of silica sufficiency, and low magnesia content, forsterite and dolomite are also absent.

B. Marbles of the Western Part of the Area.

The northerly trending western limestone beds obliquely intersect the north-westerly trending metamorphic isograds based on the mineral assemblages within the pelitic schists (Plate 2). Thus the range of metamorphic grade encountered by the western marbles is much more limited compared with that of the eastern marbles. The lowest grade marbles of the western belt are associated with pelitic schists containing andalusite, staurolite and muscovite, slightly below the grade at which sillimanite first appears. The highest grade marbles occur in the middle part of the muscovite-sillimanite zone of the pelitic schists.

(a) Mineral and Mineral Assemblages.

The majority of the western marbles lie between the mineral isograds based on the first appearance of scapolite in the muscovite-calcite-quartz assemblages, and the first appearance of clinopyroxene in the biotite-calcite-quartz assemblages. The reactions involving the break-down of the micas, in particular the biotite, appear to have been delayed in the western marbles. This is considered to be related to increasing values of the pressure factor as one proceeds westwards. As well as the more common potash sufficient biotite-calcite-quartz marbles, a few potash deficient chlorite bearing marbles have been encountered. In the presence of excess quartz these develop first amphibole, and then clinopyroxene, with increasing metamorphic grade. One marble deficient in both potash and quartz, having the assemblage calcite-actinolite-brucite, was found. No grossular, vesuvianite or wollastonite have been found in these marbles.

The assemblages encountered with increasing grade are as follows.

Potash sufficient marbles.

(1) Quartz-calcite-biotite-plagioclase-muscovite.

(2) Quartz-calcite-biotite-plagioclase-muscovite-potash felspar-scapolite.

(3) Quartz-calcite-phlogopite-clinopyroxene-scapolite-potash felspar.

Potash deficient marbles.

(4) Quartz-calcite-biotite-chlorite-muscovite-scapolite.

(5) Quartz-calcite-biotite-plagioclase-amphibole-chlorite-scapolite.

(6) Quartz-calcite-amphibole-clinopyroxene-scapolite-epidote.

Silica-potash deficient marble.

(7) Calcite-amphibole-phlogopite-brucite.

All assemblages may have accessory sphene, apatite, tourmaline or pyrite.

The grain size of the lower grade western marbles is somewhat greater than their eastern counterparts and there is no consistent grain size change with increase of grade. The grain size of the calcite is closely related to the amount of dispersed fine-grained silicate impurities. The complete lack of preservation of intense strain effects related to folding in the western marbles is in marked contrast to the highly strained eastern marbles. The calcite textures in the western marbles range from granoblastic to a somewhat interlocking jigsaw pattern. The calcite grains may be equidimensional or distinctly flattened into the axial plane foliation. The c-axes have a strong preferred orientation (Chapter 13). In the western marbles recrystallization clearly outlasted deformation associated with folding. This is reflected in the absence of undulose extinction and bent twin lamellae in the calcite, and in the static idioblastic growth of the new minerals scapolite, potash felspar, amphiboles and pyroxene. Fine twin lamellae in the calcite are related to post-crystallizational faulting. However, marbles approaching the eastern side of the area, such as those forming the anticline in Saunders Creek to the south, show a moderate post-crystallizational strain. A block of marble located 1 mile NE of Whites trigonometric station and just west of the main fault dividing the eastern and western parts of the area, was the source of some of the most highly strained marble specimens found. These have a strongly foliated true mylonitic texture. Large calcites with a patchy undulose extinction have become squeezed out parallel to the foliation so that in section their length is many times their width.

(b) Descriptive Petrography of the Potash Sufficient Marbles.

The lowest grade marbles contain bedded compositional layers of fine-grained quartz, oligoclase and oriented red-brown biotite in a coarser calcite rich matrix. Muscovite appears as stumpy flakes distributed throughout both the calcite and quartzo-felspathic bands. With increased grade muscovite is destroyed with the generation of potash felspar and anorthite, which calcifies the plagioclase to An 30-An 40. Scapolite may also appear with the disappearance of muscovite, while the orange-red-brown biotite remains stable. The scapolite and untwinned potash felspar are usually porphyroblastic, the scapolite in some cases showing an idioblastic tetragonal form. Both minerals commonly have numerous quartz, calcite and biotite inclusions, those in the potash felspar being particularly fine and concentrated near the cores of the grains.

After potash felspar and scapolite have formed, some muscovite may still persist, and seems reluctant to leave the assemblage. In some cases the assemblage potash felspar-scapolite-muscovite appears stable. This may suggest a degree of divariance for the reaction

$$\left[\text{muscovite} + \text{calcite} + \text{quartz} \rightleftharpoons \text{orthoclase} + \text{anorthite} + H_2O + CO_2\right].$$

If so, the reaction may not go to completion at a particular P-T value but within a narrow range of P-T. As the reaction contains two volatiles this divariance is possible in a closed system (Walter, Wyllie and Tuttle, 1962). If $P_{volatiles}$ is high, then the system may be considered as partially closed and a degree of divariance is introduced. The presence of secondary muscovite porphyroblasts in many of the western marbles suggests that under falling temperature, and with the availability of some H_2O and CO_2, the above reaction was readily reversed. The formation of clinopyroxene from biotite only occurs in the highest grade marbles to the south (assemblage (3)).

(c) Descriptive Petrology of the Potash-deficient Limestones.

The lowest grade marble in this category is represented by assemblage (4). In this rock scapolite has begun to appear through the destruction of muscovite. Both biotite and chlorite appear to be in stable equilibrium. The biotite is a pale brown variety, and the chlorite pale green with a very low birefringence. The chlorite flakes are generally larger than the biotite and may enclose some biotite flakes as inclusions. A slightly higher grade is represented by assemblage (5). Here muscovite has disappeared and chlorite is in the process of disintegration, with the concurrent growth of amphibole. The following reaction is indicated.

$$\left[5\ Mg_6Si_4O_{10}(OH)_8 + 12\ CaCO_3 + 28\ SiO_2 \rightleftharpoons 6\ Ca_2Mg_5Si_8O_{22}(OH)_2 + 19\ H_2O + 12\ CO_2\right]$$

chlorite, calcite, quartz; tremolite

..........(8)

The amphibole, occurring as bunches of irregular porphyoblasts with numerous quartz-felspar inclusions, is pleochroic X-pale lemon, Y-mid-green, Z-blue-green. Pale biotite is often closely associated with the amphibole. Dark halos are common around small sphene inclusions in both amphibole and biotite. A few rare chlorite flakes are still preserved near the amphibole, these being pleochroic from pale lemon to a distinct, but pale, green and have anomalous lavender interference colours.

With a further increase of metamorphic grade the chlorite reacts completely. Then amphibole becomes unstable in the presence of quartz and calcite and the following reaction ensues.

$$\left[Ca_2(Mg,Fe)_5Si_8O_{22}(OH)_2 + 3\ CaCO_3 + 2SiO_2 \rightleftharpoons 5\ Ca(Mg,Fe)Si_2O_6 + 3\ CO_2 + H_2O\right]$$

actinolite, calcite, quartz; clinopyroxene

..........(9)

Rock 707 (location 137.206) has assemblage (6). Small granules of pale green pyroxene have appeared, while some unreacted amphibole still remains. Biotite and plagioclase are absent, and since pure anorthite, or grossular, are not stable at this grade a few primary epidote grains are present.

(d) Mineralogy of the Western Marbles.

Quartz and plagioclase were present as small grains in most of the marbles examined. Estimation of the composition of the plagioclase in four specimens by the $\perp a$ extinction method gave compositions ranging from An 31-An 42 (using curves in Troger, 1959, p. 111). Allowing for some zoning the composition range is narrow, and compares with the plagioclase composition in the eastern marbles. Again the major contributing factor to the anorthite content of the plagioclase is the destruction of the early muscovite. Potash felspar may occur as small grains, but is more common as large porphyroblasts. Again no cross-hatched twinning could be discerned and the optic axial angles are low. In the high grade marble, specimen 18, location 194.086, potash felspar has 2Vx $54-63^{\circ}$ (variation due to strain) and $n\beta = 1.524 \pm .002$. Potash felspar porphyroblasts in specimen 256, location 169.148, gave 2Vx $64^{\circ} \pm 1^{\circ}$. Scapolite granoblasts in specimen 18 gave $n_{\omega} = 1.577 \pm .001$, indicating a composition near Me 59, and porphyroblasts in specimen 240, location 170.130, gave $n_{\omega} = 1.579 \pm .001$, Me 61. Biotite is stable throughout the western marbles and usually has a red-brown or orange-brown Z-axial colour. The biotite in specimen 18 gave $n_{\gamma} = 1.625 \pm .002$ and had the pleochroism X-dark yellow, YZ—yellow green. The associated clino-pyroxene in this specimen has the properties $n_{\beta} = 1.705 \pm .002$, 2Vz $57^{\circ} \pm 1^{\circ}$, $Z \wedge c$ $43 \pm 1^{\circ}$ and is pleochroic pale yellow to pale green, indicating a composition near Ca:Mg:Fe of 47:29:24, salite (Hess, 1949).

C. Comparison of the Western and Eastern Marbles.

The western and eastern marbles have a similar composition and pass through a similar series of metamorphic reactions with increasing grade. Most marbles are sufficiently rich in potash and alumina for chlorite and amphibole to be absent, being replaced by biotite. A few chlorite and amphibole bearing marbles have been encountered in the western belt. The eastern marbles exhibit a wider range of metamorphic grade than their western counterparts. Reactions involving the destruction of muscovite and biotite, with the release of the volatiles H_2O and CO_2 seem to have been delayed in the western marbles. This may be attributed to increased total pressure to the west, indicated by the occurrence there of kyanite bearing schists. The most notable difference between the western and eastern marbles is textural. Recrystallization in the western marbles is post-tectonic, while in the eastern marbles tectonic activity outlasted recrystallization. Either recrystallization ceased earlier or deformation proceeded to a later stage in the eastern marbles. Petrofabric work in the eastern marbles suggests that much of the strained calcite fabric is related to the first and major deformation period, which is likely to have occurred simultaneously throughout the area. It is considered more likely that recrystallization ceased earlier in the eastern marbles. The more highly

FIG. 41.

Photomicrograph of marble from western side of the area showing directed crystalloblastic texture and characteristic simple twinning of calcite. Also within the field of view are small grains of quartz, plagioclase and biotite, and portions of several potash felspar porphyroblasts filled with minute inclusions.

Rock A185-731. Location 153.231. Ordinary light.

FIG. 42.

Photomicrograph of brucite-tremolite dolomite.

Rock A185-555. Location 234.210. Crossed polars.

Partially altered ragged prisms of tremolite (left) and slightly deformed brucite flakes are set in a granular dolomite matrix. Brucite flakes having the orientation shown by the flake in the centre of the field of view show the development of strong deformation bands parallel to their c-axes.

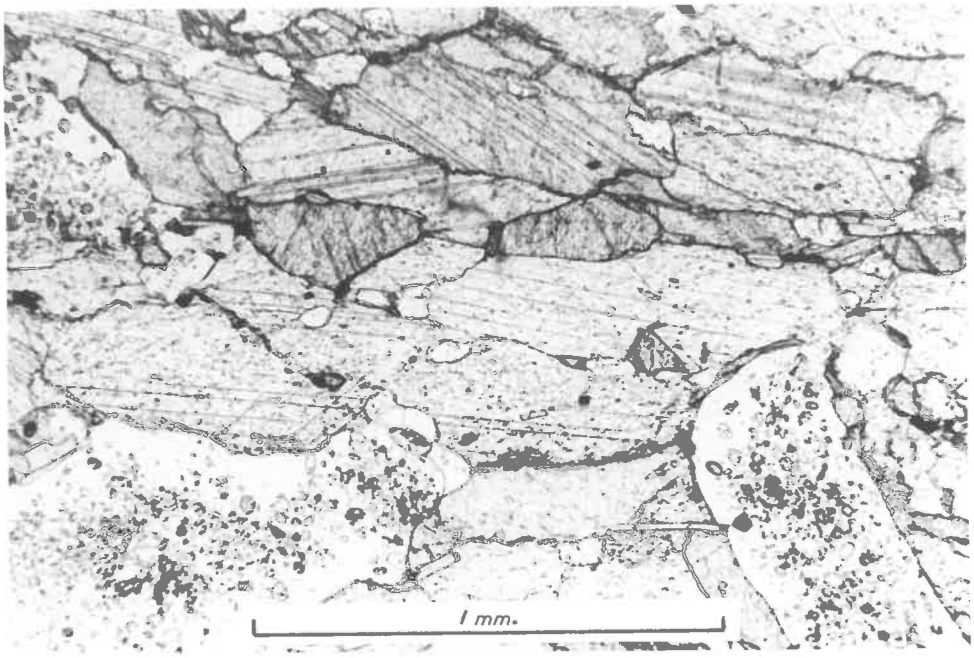


FIG. 41

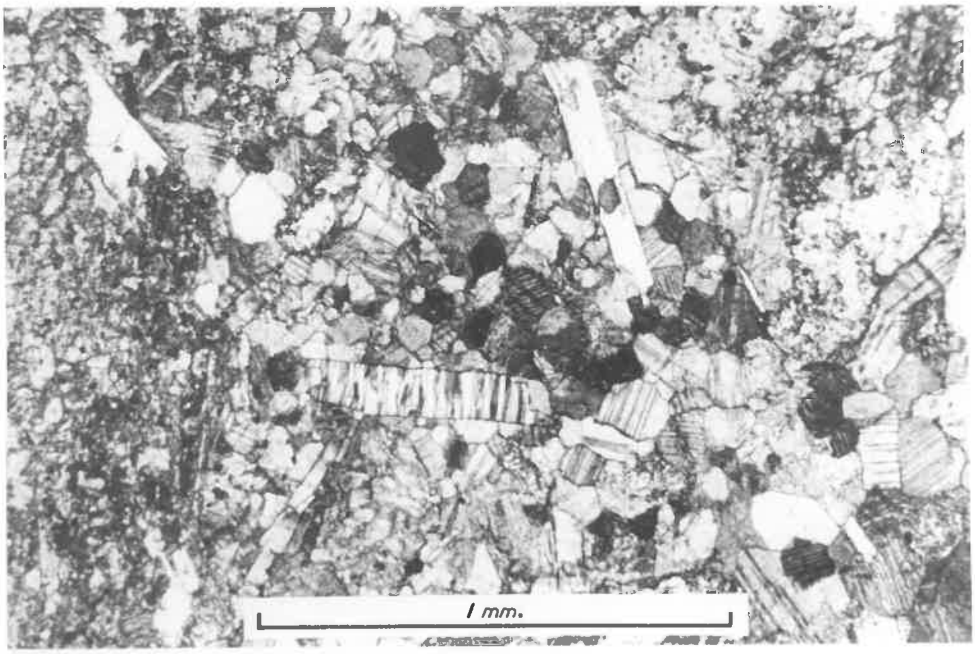


FIG. 42

strained fabrics have suffered some annealing recrystallization. Greater activity of water during the metamorphism of the western marbles would have enabled recrystallization to outlast deformation associated with folding. This picture is conformable with the conclusions reached from the study of the metamorphic recrystallization in the pelitic and arenaceous rocks.

D. Silica Deficient Magnesian-Carbonate Rocks.

Rocks of this category are characterized by the association of brucite and a magnesian amphibole (tremolite or magnesian hornblende). Dolomite, calcite, phlogopite and scapolite may also be present with the accessories sphene, apatite and iron ore. Only three locations are known and the rocks have the following assemblages:

555 Dolomite-tremolite-brucite-scapolite-(sphene)

406 Dolomite-Mg hornblende-brucite-calcite-phlogopite-(apatite-sphene-iron ore)

258 Calcite-actinolite-brucite-phlogopite-(sphene).

Specimen 555 (location 234.210) was taken from a narrow steeply dipping lenticular band outcropping over a length of 50 feet in a thick sequence of fine-grained micaceous quartzo-felspathic schists of staurolite grade. The rock is a uniform white to buff-coloured fine-grained dolomite with a faint foliation banding outlined by tremolite prisms and brucite flakes weathering out on the exposed surfaces. Eighty percent of the rock is a fine (0.1 mm.) even-grained granoblastic matrix of dolomite in which is embedded numerous prisms of tremolite and flakes of brucite with rarer anhedral scapolite and granules of sphene. The tremolite (2Vx 80^{o}, $Z \wedge c$ 15) forms thin colourless prisms to one millimeter in length, which have ragged edges surrounded by fine granular secondary dolomite resulting from retrograde alteration. Brucite appears in well-formed colourless tabular crystals reaching 0.5 mm. in length. These flakes are uniaxial positive, have a weak cleavage, and a birefringence similar to quartz. In those flakes at a high angle to the foliation, deformation by gliding on kink lamellae parallel to [001] is common. The late deformation causing this kinking has also caused fine f twins to appear in the dolomite and basal fractures in the tremolite prisms.

Rock 406 (location 204.133) occurs at the edge of a marble bed adjacent to a large mass of metasomatic quartz-albite rock. The rock is a coarse grained (1-2 mm.) grey, mica-amphibole rich, non-foliated marble. The rock has an even granoblastic texture, with some coarser prismatic growth of the amphibole. Like all the eastern marbles this rock has suffered some post-crystallization strain, and the presence of widespread twin lamellae in the carbonates has established that calcite is in excess of dolomite. Phlogopite forms 1 mm. squat crystals with a pleochroic scheme X-colourless Y,Z-lemon. Brucite is rare and has commonly grown as thin colourless flakes attached to the {001} faces of the phlogopite crystals. There is, however, no evidence for phlogopite-brucite instability. Late deformation causing twinning in the carbonates has bent the phlogopite flakes and caused deformation kinks in the brucite similar to those of rock 555. A pale green slightly pleochroic amphibole is abundant as anhedral stumpy prisms. A high $Z \wedge c$ angle (24°) suggests magnesian hornblende. Most of the prisms are zoned with an increased colour intensity and extinction angle value towards their rims.

Rock 258 (location 168.145) consists of numerous randomly arranged slender prisms of pale green amphibole (2Vz $87\frac{1}{2}^\circ$, $Z \wedge c$ $17\frac{1}{2}^\circ$) with rare flakes of phlogopite and brucite embedded in a coarse grained matrix of untwinned calcite.

In all these rocks brucite occurs as well-formed primary crystals and there is no evidence for former periclase. Recent experimental work on the system MgO-CO_2-H_2O (Walter, Wyllie and Tuttle, 1962) has indicated that brucite is not stable in the presence of aqueous vapours which contain more than a few weight percent CO_2, and that high P_{H_2O} favours the growth of brucite in place of periclase. The brucite rocks at Cambrai have presumably formed under high P_{H_2O} relative to P_{CO_2} conditions.

Rock 555 occurs at a lower grade than the first diopside, while rock 406 occurs within the sillimanite-muscovite zone, where diopside is common in the adjacent calc-silicate rocks. Weeks (1956) has shown that for a given total pressure forsterite appears at a temperature some $80^\circ C$ higher than that of diopside. The stability of amphibole and dolomite in rock 406 indicates that the temperature has not been reached where forsterite could form by reaction

$$\left[\underset{\text{tremolite}}{Ca_2Mg_5Si_8O_{22}(OH)_2} + \underset{\text{dolomite}}{11\ CaMg(CO_3)_2} \rightleftharpoons \underset{\text{forsterite}}{8\ Mg_2SiO_4} + \underset{\text{calcite}}{13\ CaCO_3} + 9\ CO_2 + H_2O\right] \quad \ldots\ldots\ldots\ldots(10)$$

CHAPTER 7.

METAMORPHISM OF THE CALC-SILICATE ROCKS.

Introduction.

The calc-silicate rocks constitute a common and widespread group of metamorphic rocks in the Cambrai area, occurring as numerous beds of variable thickness within the quartzo-felspathic schist sequence above and below the marble horizon. In the lower grades of metamorphism the calc-silicate rocks are represented by calcite-mica schists, which, like the marbles, are strongly banded by alternating calcite and mica rich layers. Widespread quartz and felspar grains within the calcite rich layers suggests a clastic origin for these rocks. With recrystallization and the growth of new calc-silicate minerals at higher metamorphic grades, the beds outcrop conspicuously, and serve as strong local stratigraphic marker horizons within the monotonous quartzo-felspathic schist sequence. Some of the thicker beds reach 60-100 ft. in width and have been mapped and plotted in Plate 1. As well as these larger beds, numerous thin bands and laminations of calc-silicate composition are dispersed through parts of the quartzo-felspathic schist sequence. The calc-silicate rocks show a wide range of compositional, mineralogical and textural variation. The mineralogical changes depend largely on reactions occurring with increasing metamorphic grade. As in the marbles, these reactions, which involve release of the volatiles H_2O and CO_2, are retarded in the western rocks, suggesting increasing pressure, especially P_{H_2O}, to the west. Calc-silicates from the western margin of the area have a distinctive mineralogy and are considered as a separate group.

A. Calc-silicate Rocks from the Eastern Part of the Area.

The majority of the calc-silicate rocks from the eastern side of the area form distinct beds within the quartz-felspar-biotite schist sequence stratigraphically below the marble. The more important of these beds have been mapped and plotted in Plate 1. On the eastern scarp of the ranges north of the Marne River, a thick sequence of calc-schists within the staurolite grade of metamorphism, contain widespread scapolite and amphibole. In general, with increasing grade, the rocks change from calcite-biotite schists, to scapolite-amphibole calc-silicates, to diopside scapolite calc-silicates.

The list of assemblages given below is, by no means, a complete list of the assemblages actually observed, but comprises selected critical assemblages arrayed in order of appearance with increasing metamorphic grade. In actual assemblages secondary minerals are often present, but these have been ignored here. All assemblages may contain in addition the following accessory minerals: pyrite, iron ore, sphene, apatite, tourmaline and zircon.

Biotite zone in pelites. Plagioclase to An 8.

(1) Quartz-calcite-biotite-muscovite-plagioclase.

Biotite-andalusite zone in pelites. Plagioclase to An 15.

(2) Quartz-calcite-biotite-muscovite-plagioclase-potash felspar.

(3) Quartz-calcite-biotite-amphibole-plagioclase-chlorite.

(4) Quartz-calcite-biotite-plagioclase-scapolite-potash felspar.

Biotite-andalusite-staurolite zone in pelites. Plagioclase to An 60.

(5) Quartz-calcite-scapolite-biotite-amphibole-plagioclase-potash felspar.

(6) Quartz-calcite-scapolite-clinopyroxene-biotite-plagioclase-potash felspar.

(7) Quartz-scapolite-amphibole-plagioclase-potash felspar-biotite-clinopyroxene.

Sillimanite-muscovite zone in pelites. Plagioclase to An 86.

(8) Quartz-amphibole-biotite-epidote-garnet-plagioclase-potash felspar.

(9) Quartz-amphibole-clinopyroxene-garnet-plagioclase-scapolite-potash felspar.

(10) Quartz-amphibole-clinopyroxene-garnet-plagioclase-biotite-potash felspar.

Sillimanite-potash felspar zone in pelites. Plagioclase to An 93.

(11) Quartz-scapolite-biotite-amphibolite-clinopyroxene-plagioclase-potash felspar.

(12) Epidote-calcite-diopside-scapolite.

(13) Quartz-calcite-diopside-scapolite.

(a) Descriptive Petrography of the Calc-silicates.

(1) Calc-silicates of the Biotite Zone.

Within the biotite zone, and before the appearance of andalusite in the pelitic schists, the calc-silicates are represented by fine-grained carbonate schists. Except for an increased content of silicate constituents, giving them a darker more schistose appearance, these rocks closely resemble the low grade marble. Like the marble they are composed of alternating carbonate and silicate rich bands, 1 mm. to 2 cm. thick, which are often intricately micro-folded about a strong axial plane foliation defined by biotite and calcite orientation. The calc-schists have the same mineralogy as the marbles and are of similar grain-size. The carbonate rich bands, distinctly coarser grained than the phyllitic bands, carry abundant sub-rounded and strained quartz grains. The lack of strain in the calcite suggests that the quartz may be relic detrital grains which have not yet recrystallized. Plagioclase in one rock was found to have a composition near An 8. Muscovite is commonly xenoblastic, as in the limestones, while red-brown biotite is well oriented into the axial plane foliation.

A thin border zone of scapolite schist adjacent to copper bearing quartz-tourmaline veins in the Pine Hut copper prospects is the lowest grade occurrence of scapolite known in this area. This rock is a dark-grey fine-grained quartz-felspar-biotite-muscovite schist containing numerous 0.5-1 mm. xenoblasts of scapolite crowded with minute quartz inclusions. Biotite and muscovite are uniformly abundant in the matrix right up to the edge of the scapolite crystals. Calcite is absent from this rock. If suitable volatiles were available the scapolite may have crystallized directly from plagioclase. The scapolite has $n_{\omega} = 1.569$, suggesting a composition near Me 49.

(2) Calc-silicates of the biotite-andalusite zone.

Soon after the appearance of andalusite in the pelitic schists, the assemblages chlorite-calcite-quartz and muscovite-calcite-quartz become unstable, the first yielding amphibole and the second yielding anorthite (scapolite) and potash felspar by reactions similar to those described for the marbles.

The first rock to contain amphibole (specimen 584, location 232.225) consists largely of granoblastic calcite, quartz and plagioclase (An 5-15), with lesser biotite, amphibole and chlorite. The amphibole, forming small isolated irregular stumpy poikiloblasts with numerous quartz and calcite inclusions, is pleochroic X-pale yellow, Z-pale green, and has the properties 2Vx 72°, $Z \wedge c$ 15½°. According to Tröger (1959, p. 77) these properties suggest an intermediate hornblende with 60% Mg-molecule. Small pale green chlorite flakes with anomalous dark lavender pink interference colours are apparently primary. Squat flakes of biotite are pleochroic golden-brown to lemon-yellow.

Specimen 812, location 237.219, marks the first appearance of scapolite in a carbonate rock. In thin slice the rock consists of aggregates of coarse calcite and large (1 mm.) syntectonically rotated subidioblastic porphyroblasts of scapolite sieved with fine quartz, felspar and calcite granules, immersed in a fine foliated matrix of quartz, felspar and biotite. Some porphyroblasts have retrograded to a fine mat of sericite, calcite and quartz.

<u>(3) Calc-silicates of the biotite-andalusite-staurolite zone.</u>

On the eastern edge of the ranges, north of the Marne River, a slightly calcareous schist sequence, not represented elsewhere in this area, contains abundant scapolite, amphibole and potash felspar.

These schists, usually carbonate deficient, show a considerable variety of textures depending on the amounts of scapolite, amphibole and potash felspar present. These minerals, grown as porphyroblasts, are set in a fine-grained granular and foliated matrix of quartz felspar and biotite. Where the foliated matrix is dominant, it may control the shapes of the post-tectonic porphyroblasts. Scapolite xenoblasts, in particular, are commonly flattened by faster crystal growth parallel to the foliation. On the other hand, the amohiboles commonly show randomly oriented, bladed, radiate or "bowtie" idioblasts, which sharply cross-cut the foliated matrix. Where the amphibole is a major component of the rock it tends to develop large web-like skeletal porphyroblasts which grow more favourably along certain original compositional layers of the rock.

Rock 494, location 236.212, a fine-grained grey calc-schist spotted with white scapolite and green amphibole porphyroblasts, is typical of the lower grade rocks of this sequence. Porphyroblasts of scapolite, to 1 mm., sieved with small quartz, biotite and amphibole inclusions; ragged pale-green amphibole, to 0.3 mm., with 2Vx 73^{o} and Z $\wedge$ c 16½^{o} (60% Mg-hornblende); and potash felspar with incipient microcline twinning, are dispersed through a quartz-plagioclase-biotite matrix having an average grain-size of 0.02 mm. Fractured and strained, rounded quartz grains to 0.3 mm. may be detrital. Lozenge shaped and granular sphenes are numerous. The plagioclase composition is An 30-40.

At 238.212 rocks similar to 494 are cut by quartz-scapolite-hornblende-potash felspar bearing segregation veins to several centimetres in width (specimens 24 a-d). Occasional veins and lensoidal pods are zoned; a core composed of coarse scapolite prisms immersed in quartz being surrounded by a coarse grained amphibole rich rim, in turn followed by a light coloured outer rim of very fine-grained quartz and plagioclase.

In the middle staurolite zone, the more massive calc-silicates resemble fine-grained non-foliated syenites in hand specimen, for example specimen 540, location 238.197. These rocks are principally composed of potash felspar amphibole and plagioclase. The potash felspar shows incipient microcline twinning, the plagioclase has a composition near An 36-39, and the amphibole, occurring as diversely oriented anastomosing crystals embedded in the potash felspar, is a pale-green hornblende (2Vx 76°, $Z \wedge c$ 16°). However, the more usual rock at this grade is a calc-schist, represented by specimen 495, location 239.186, consisting of scattered ovoid scapolite xenoblasts to 2 mm., with lesser pale-green amphibole and potash felspar porphyroblasts, dispersed through a fine-grained foliated matrix of quartz, felspar and biotite.

The potash felspar in these rocks regularly shows incipient cross-hatched twinning and some grains are zoned. In specimen 24, cross-hatched cores have 2Vx 72-80°, while non-twinned rims have 2Vx 60-74°. At a slightly higher grade, for example 494, the potash felspar has 2Vx 57-62°.

Optical measurements on a number of amphiboles in these rocks suggests that they are common hornblendes. 2Vx ranges from 69-77°, while $Z \wedge c$ values lie between 15-18°. The results have been plotted graphically in Fig. 43a (black crosses), along with results from other amphiboles from calc-silicate rocks in other parts of the area. Japanese workers (e.g. Shido, 1958) have successfully used 2Vx versus n_{γ} plots for the optical distinction of amphiboles belonging to separate amphibole series. The chemistry of amphiboles cannot be readily related to optic parameters (for example, actinolites; Deer, Howie and Zussman, vol.2, pp. 251-253), and the 2Vx versus $Z \wedge c$ plots used here give only a rough indication of the amphibole chemistry. The curves of Fig. 43 have been derived from published 2Vx and $Z \wedge c$ curves as indicated. The

FIG. 43.

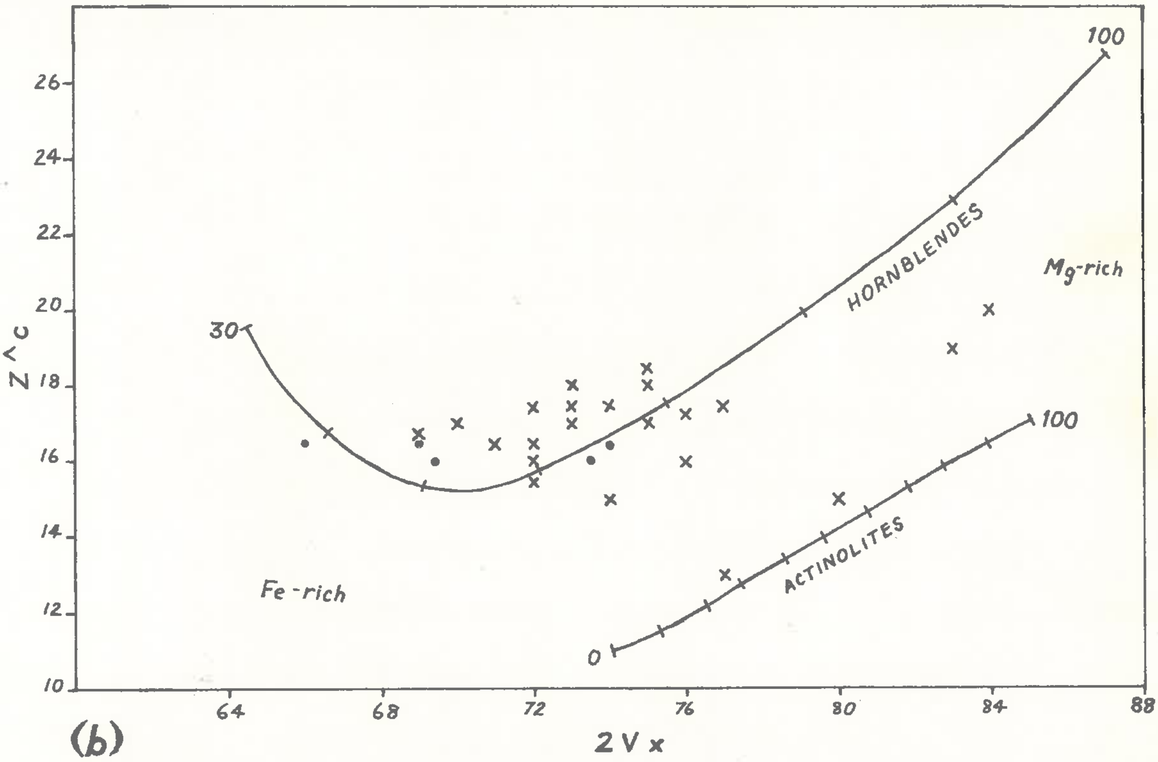
Optics of amphiboles from the Cambrai area.

(a) Variation of $Z^{\wedge}c$ with $2V_x$ in amphiboles from metasediments on the eastern side of the area (crosses) and the western side (dots). The curves for actinolite and hornblende composition variation were derived from Tröger, 1956, pages 72 and 77.

(b) Similar plot of amphibole optic variation for amphiboles in rocks of igneous parentage.

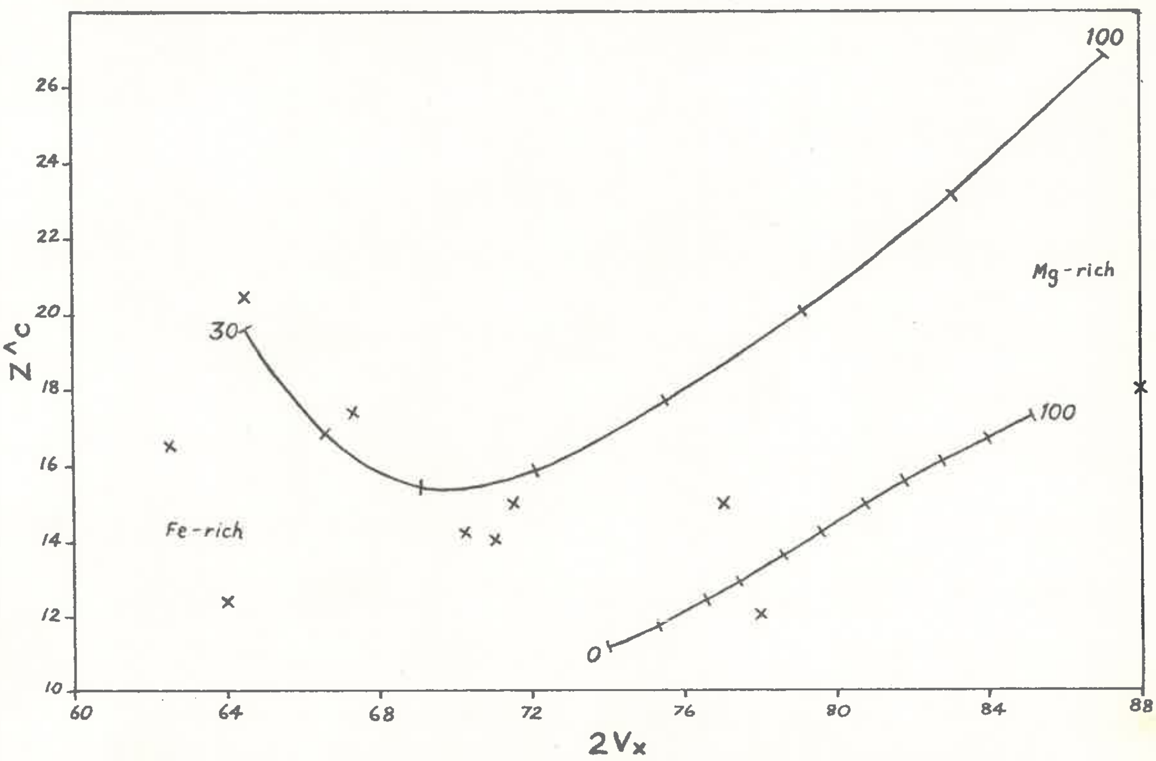
FIG. 43

(a)



100
HORNBLENDES
Mg-rich
30
100
ACTINOLITES
Fe-rich
0
Z^c
2Vx

(b)



100
Mg-rich
30
100
Fe-rich
0
Z^c
2Vx

metamorphic amphiboles from the Cambrai calc-silicates tend to fall into two groups corresponding to either the actinolite or common hornblende series (hastingsitic amphiboles from the western calc-silicates, and cummingtonites, have been omitted). Amphiboles from rocks of igneous parentage (Fig. 43b), plotted in a similar manner, show no consistent relations.

An unusual dark-green amphibole-rich calc-silicate from the commencement of the staurolite zone (specimen 559, location 233.212), consists largely of deeply coloured blue-green actinolite (2Vx 77°, Z^c 13°, X-pale lemon, Y-mid-green, Z-deep blue-green, 30% Mg-actinolite - Tröger, 1959, p. 72) and coarse calcite, interspersed with minor iron ore and biotite, and disrupted relics of fine-grained phyllitic schist bands.

In the calcite sufficient calc-schists containing amphibole and quartz, diopside appears soon after the first appearance of staurolite in the pelitic schists.

(4) Calc-silicates of the sillimanite-muscovite zone.

Soon after the first appearance of fibrolitic sillimanite in the pelitic schists, garnet appears in some calc-silicates. Like the garnet in the limestone at the Kanappa Mine, the primary nature of this mineral is in doubt, for in many rocks secondary epidote and calcite are also present. As the garnets are usually surrounded by, and have inclusions of, both epidote and calcite, its origin by reaction (4) (Chapter 6) is indicated.

In rock 14, location 222.166, garnet appears to be a primary constituent. This rock consists of thin amphibole-garnet-quartz-felspar bands dispersed through fine-grained quartz-plagioclase-biotite schist. The schist consists of 0.1 mm. granoblasts of quartz set in fine (0.01 mm.) granular bytownite and moderately oriented orange-brown biotite. Secondary clinozoizite granules have appeared at numerous points. The more calcareous layers consist of large, 1-5 mm., interlocking skeletal webs of hornblende of intermediate composition (2Vx 70°, Z $\wedge$ c 17°, X-pale-green, Y-mustard-green, Z-blue-green). The amphibole is intergrown with small irregular to skeletal garnets and occasional epidotes, immersed in a matrix of quartz, potash felspar and bytownite (An 73-80). Some of the epidote is associated with pale green chlorite, forming from original biotite, and may be secondary. Another rock of similar appearance (spec. 11, location 229.148) contained a more sodic plagioclase (An 46-53) and no garnet.

In rock 15, location 225.155, minor garnet, enclosed in scapolite which has in part altered to calcite and epidote, is of more doubtful primary origin. This rock is typical of the normal calc-silicate rocks in the lower part of the sillimanite zone. Despite complete recrystallization to coarse interlocking skeletons of amphibole, diopside, scapolite, potash felspar and quartz, the rock shows the preservation of small scale fold structures. In thin section the texture is complex. Large webbed amphibole and diopside skeletons are intricately intergrown with large sieved scapolites. In places plagioclase (An 38-46), potash felspar and quartz form large porphyroblasts. Quartz and felspar also occur as small granoblastic inclusions within the amphibole, diopside and scapolite. Various sized lozenges of sphene are distributed throughout. The texture is further complicated by secondary alteration of diopside to pale actinolite, and scapolite to epidote and calcite.

Large, 1-5 mm., sub-idioblastic garnets, surrounded by secondary epidote and chlorite, are abundant in rock 388 (location 222.150). With its complex intergrown texture this rock is similar to rock 15, but contains abundant randomly oriented interlocking poikiloblasts of amphibole and no scapolite. The plagioclase is basic (An 86), and some epidote may be primary.

Another type of calc-silicate rock encountered in this zone of metamorphism is exemplified by specimen 430, location 221.144. This is a well-bedded, faintly

FIG. 44.

Photomicrograph of amphibole-quartz-bytownite calc-silicate.

Rock A185-430. Location 221.144. Ordinary light.

Decussately arranged prisms of common green amphibole set in a fine granular matrix of quartz and bytownite.

FIG. 45.

Photomicrograph of clinopyroxene-scapolite rock, Kanappa Mine.

Rock A185-10. Location 216.118. Ordinary light.

Several grains of sphene are also present.

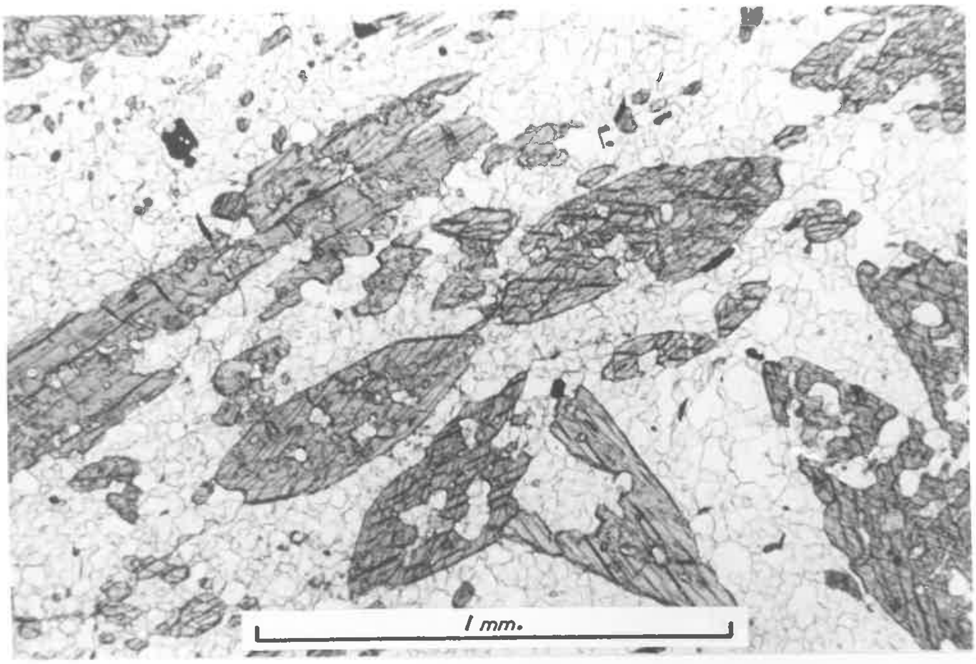


FIG. 44

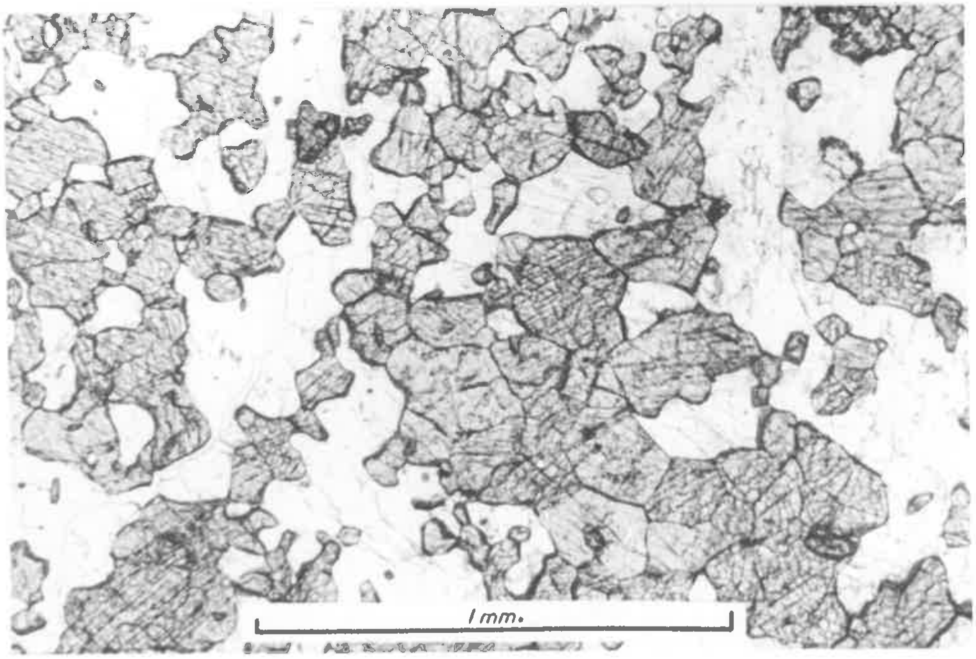


FIG. 45

layered rock, consisting of randomly oriented prismatic hornblende poikiloblasts embedded in a fine quartz-bytownite matrix. Granules of sphene, iron ore and apatite are common as accessories.

(5) Calc-silicates of the sillimanite-potash felspar zone.

In the calc-silicates of the sillimanite-potash felspar zone, the reaction involving the formation of diopside from amphibole, calcite and quartz has gone to completion, and amphibole is only preserved in the quartz or calcite deficient assemblages. Biotite, too, has almost disappeared from the quartz-calcite assemblages, with consequent enrichment in clinopyroxene. The highest grade quartz-calcite bearing calc-silicates may have the assemblage quartz-calcite-clinopyroxene-scapolite-plagioclase-potash felspar-sphene, with minor phlogopite. Some secondary alteration, with the formation of epidote and calcite from scapolite, and amphibole and chlorite from clinopyroxene, is ubiquitous. In the Kanappa Mine calc-silicates, it is difficult to distinguish the secondary and primary minerals because of a more intense recrystallization accompanying the secondary paragenesis. Here well crystallized skeletal garnet has appeared in clinopyroxene-scapolite rich calc-silicate rocks containing much secondary epidote, calcite, quartz and amphibole.

Numerous small prisms of primary epidote occur in rock 99 (location 207.097). This is a quartz deficient rock containing abundant granoblastic calcite dispersed through anastomosing skeletal porphyroblasts of scapolite. Scapolite rich layers contain numerous deep-green hedenbergite granules. The epidote occurs as well-formed idioblastic prisms enclosed in calcite and scapolite.

In these high grade rocks plagioclase may have any composition. Rock 833 (location 208.087) consists of randomly oriented prismatic porphyroblasts of a pale-green magnesian amphibole (2Vx 83°, $Z \wedge c$ 19°), set in an even grained granoblastic matrix of non-twinned anorthite (An 94 $n_{\beta}=1.582$). Sphene is a common accessory. Clinozoizite has begun to form as a secondary alteration product.

In a clinopyroxene-scapolite calc-silicate bordering an aplite vein, the plagioclase was found to be strongly zoned from cores of An 80 to rims of An 16.

Textural changes are important in the high grade calc-silicates. In the lower portion of the sillimanite zone the major minerals of the calc-silicates, clinopyroxene, scapolite and amphibole, form large skeletal porphyroblasts, sometimes up to 1-2 cm. in diameter. In the highest grade rocks these minerals have largely recrystallized to an even granoblastic texture. All stages in this polygonization can be observed. The large crystals of the lower grade rocks generally show undulose extinction resulting from bending of the crystal lattice under a late phase of deformation. In the higher grade rocks chemical activity was strong enough to enable the strain in the porphyroblasts to be alleviated by recrystallization after the deformation had ceased. This process is similar to the granular recrystallization of cordierite described earlier, and has certain similarities to the annealing recrystallization in the strained marbles, but, whereas the calcite annealing appeared to be associated with the secondary paragenesis, the polygonization of the calc-silicate minerals seems to have been independent of the secondary paragenesis, and may be an earlier event. In some rocks cores of strained clinopyroxene or scapolite porphyroblasts are surrounded by an aggregate of polygonal unstrained granoblasts. These granoblasts may preserve a crystallographic orientation related to the original porphyroblast, which, even in the most recrystallized rocks, may not be totally destroyed. Thus, in the recrystallized granoblastic calc-silicates, the grains may have a sub-parallel optic orientation over an area of several square centimetres, depending on the size of the original porphyroblasts. If a petrofabric diagram was prepared for any of these minerals in such a granoblastic rock, a fair orientation of the optic axes may be recorded, which, however, would not be the result of tectonic orientation. Gellatly (1964) has described similar features in metamorphosed porphyritic nepheline syenites from Darkainle, Somali Republic. Strained relict cores of the magmatically oriented nepheline porphyroblasts are enclosed by "recrystallized polygonal interlocking sub-grains". Petrofabric analysis of the mozaic grains revealed a tendency for local domains made up of sub-grains in near parallel orientation. Scapolite has yielded "fair looking" petrofabric diagrams from some rocks of the Mt. Lofty Ranges (e.g. White, 1956). However, it needs to be demonstrated that such orientation diagrams record the result of true tectonic orientation of the grains and not a mimetic orientation resulting from the polygonization of original large porphyroblasts.

(b) Conclusions.

The eastern calc-silicates have evolved through metamorphism in a similar manner to the adjacent marbles, and many of the reactions important in the metamorphism of the marbles apply to the calc-silicates also. Parras (1958) has described calc-silicate assemblages from the West Uusimaa complex of south-west Finland which are similar, in many respects, to the higher grade calc-silicates of the present area. In West Uusimaa clinopyroxene-scapolite calc-silicates grade into rocks of the charnockite facies.

B. Calc-silicates from the Western Part of the Area.

The western calc-silicate rocks may be divided into two distinct groups, both petrologically and geographically. Associated with the marble of the western anticlinal zone, and to the east, stratigraphically below the impure arkose unit, are calc-silicate rocks of a similar character to those described from the eastern side of the area; that is, rocks, which, beginning in the lower grades as calcite-mica schists, progressively develop scapolite, amphibole and clinopyroxene with increasing metamorphic grade. Several thin calc-silicate horizons outcropping stratigraphically above the limestone, to the west of the anticlinal zone, however, contain a distinctive dark-green hastingsitic amphibole, abundant epidote in equilibrium with plagioclase, clinopyroxene and quartz. In these rocks scapolite is rare and normally absent.

Therefore, in considering the petrology of the western calc-silicates, it is necessary to subdivide them into the following groups,

(1) Clinopyroxene-scapolite calc-silicates,

(2) Hastingsite-plagioclase calc-silicates.

The latter group occurs on the western margin of the area, and the geographic boundary between the two groups is marked by the westernmost limestone-schist boundary.

(a) The Clinopyroxene-Scapolite Calc-silicates.

(1) Minerals and Mineral Assemblages.

The following assemblages, listed in order of appearance with increasing grade, have been observed.

Biotite-andalusite-staurolite zone in pelitic schists

(1) Quartz-calcite-biotite-muscovite-plagioclase.

(2) Quartz-calcite-biotite-muscovite-plagioclase-scapolite-potash felspar.

Sillimanite-muscovite zone in pelitic schists.

(3) Quartz-calcite-biotite-clinopyroxene-potash felspar-amphibole-epidote.

(4) Quartz-calcite-biotite-scapolite-plagioclase-potash felspar-epidote-clinopyroxene-amphibole.

(5) Quartz-biotite-scapolite-garnet-potash felspar.

(6) Quartz-calcite-biotite-scapolite-clinopyroxene-potash felspar.

(2) Descriptive Petrography.

The lowest grade calc-silicates (e.g. specimen 746, location 164.245, assemblage (1)) are fine-grained banded calcite-mica schists similar in most respects to the low grade eastern calc-schists.

At a slightly higher grade, scapolite first appears in close association with muscovite xenoblasts (e.g. specimen 649, location 184.176, assemblage (2)). At higher grades the muscovite progressively disappears, being replaced by scapolite. The scapolite usually has a sub-idioblastic form and encloses abundant small inclusions of calcite, biotite, quartz and felspar. Concomitant with the growth of the scapolite porphyroblasts is the appearance of potash felspar, calcification of the plagioclase and a general increase in grain size, especially of calcite. The reaction involving the break-down of muscovite in the presence of calcite and quartz seems to again satisfy the requirements of these changes.

Porphyroblastic scapolite schists, and scapolite-garnet schists without carbonate, are occasionally met with (assemblage (5)). These rocks are transitional between pelitic schists and calc-schists. At a lower grade they may have had minor free carbonate since used up in the formation of scapolite. In these rocks, the abundance of scapolite and the absence or rarity of potash felspar, indicates that a reaction such as that proposed for the development of scapolite in the calcareous rocks, is not tenable. Some scapolite rich schists appear below the grade at which scapolite first appears in rocks with a calcite milieu. No observations have been made which suggest the manner in which this scapolite may have formed. A typical rock (specimen 651, location 185.179) consists of closely packed coarse (1-2 mm.) scapolite porphyroblasts separated by a small amount of quartz-biotite matrix. The biotite flakes are commonly concentrated near the margins of the scapolite crystals. In these rocks it would appear that the abundant scapolite has grown directly from scapolite forming constituents, perhaps plagioclase, with minor carbonate and trapped chlorides. An introduction of volatiles for the formation of scapolite in these schists might be considered. However, all of the scapolite schists of this type are restricted to thin stratigraphic horizons, and it is more likely that the volatiles needed for the growth of scapolite were present in the original sediment.

Rock 682 (location 165.175) is a fine-grained (0.02 mm.) foliated quartz-felspar-biotite schist containing large isolated 2 mm. "flattened" scapolite porphyroblasts. Some calcite is present in the matrix of red-brown biotite and high relief plagioclase. Muscovite xenoblasts are dispersed among the scapolite porphyroblasts. No potash felspar has been produced by the growth of scapolite, and again it seems likely that the scapolite has grown directly from plagioclase. Birefringence changes within the scapolite porphyroblasts indicate a steady change of composition from the cores (Me 70) to the rims (Me 45). This was the only example of strong scapolite zoning observed, and the composition variation spans the range determined from all other scapolites from the area.

Rock 362 (location 167.156) contains scapolite and rarer garnet porphyroblasts in a biotite quartz matrix. Minor potash felspar may occur in the matrix, but again, the abundance of scapolite suggests direct formation from original plagioclase.

Just before the sillimanite isograd is reached clinopyroxene has commenced to appear in amphibole-calcite-quartz assemblages. Rock 329 (location 194.146) has largely randomly oriented skeletons of pale-green amphibole, to 1 cm., set in a fine-grained matrix of quartz, calcite, potash felspar and biotite (largely altered to chlorite). Epidote is present as numerous small poikiloblasts, while the clinopyroxene has begun to grow as large interstitial skeletons (assemblage (3)).

Within the sillimanite zone, diopside progressively increases at the expense of amphibole, calcite and quartz. The occurrence of these four minerals together, in many of the rocks examined, seems to be related to a lack of extensive diffusion of material over more than a few millimetres. Thus, in some rocks, amphibole and quartz may be stable well within the sillimanite zone, although grains of calcite may be only 1-2 mm. distant. Other factors, such as the amphibole composition, sluggishness of reaction and variable P_{H_2O} may also be important in preventing the reaction from proceeding rapidly.

Typical of the rocks in which amphibole has not yet been completely destroyed, by reaction with quartz and calcite, is specimen 251 (location 174.145), a massive banded calc-silicate rock in which finely skeletal webs of pale-green clinopyroxene, to several millimetres in width, are closely interlocked. Some primary amphibole is still preserved where calcite is lacking. Scapolite occurs as small irregular porphyroblasts throughout. Calcite, potash felspar ($2V_x$ 62°), quartz and plagioclase (An 38) form a granoblastic matrix.

A nearby rock (specimen 336b, location 185.140), of a similar metamorphic grade, contains quartz-calcite-amphibole as a stable assemblage and no pyroxene has yet appeared. The amphibole occurs as large 1-2 cm. skeletons, enclosing skeletal scapolites, calcite, porphyroblasts of potash felspar and granules of quartz, plagioclase and biotite. Primary epidote is also common within this rock as small xenoblasts filled with minute quartz inclusions and enclosed entirely within the scapolite porphyroblasts. This relationship suggests that, in this case, the scapolite has formed by a reaction between epidote, plagioclase and calcite.

In the highest grade rocks, assemblage (6), biotite becomes very scarce, is phlogopitic, and may have begun reacting with quartz and calcite to produce clinopyroxene.

(3) Progressive mineralogical changes in the western calc-silicates.

Progressive mineralogical changes in the diopside-scapolite calc-silicates of the western group are not very marked. Biotite is usually a red-brown variety up to the highest grades, where it is replaced by a paler phlogopitic variety. Scapolite compositions lie within the range Me 45-Me 70. Potash felspar consistently exhibits incipient cross-hatched twinning, and, in two rocks, had 2Vx values between 60-63°. The amphiboles are common blue-green to green hornblendes, which, in four specimens, showed the following range of optical properties: 2Vx 63-73°, $Z \wedge c$ 16-18°. The clinopyroxenes are pale-green varieties (salites), deep green hedenbergitic clinopyroxenes are absent. Because of the fineness of grain-size, plagioclase compositions could not be obtained from the lowest grade rocks. Near the base of the sillimanite-muscovite zone, plagioclase has a composition near An 40. Rock 121 (location 204.083), a higher grade calc-silicate, contains bytownite in equilibrium with epidote. The specimen is part of a narrow 4 cm. calc-siliceous band embedded within pelitic quartzo-felspathic schists. The plagioclase (An 80-94) occurs as small rarely twinned interstitial grains between larger granoblasts of quartz, skeletal porphyroblasts of a deep-green amphibole (2Vx 63°, Z $\wedge$ c 18°), sparse flakes of golden-brown biotite and rare xenoblastic grains of epidote.

An unusual plagioclase was noted in rock 285, location 179.119. The specimen shows several 1 cm. mottled green and white amphibole-plagioclase-quartz-potash felspar rich bands embedded in a fine-grained grey quartz-felspar-biotite schist. The bands consist of skeletal porphyroblasts of green hornblende (2Vx 73°, Z $\wedge$ c 16°), intergrown with large 0.4 mm. xenoblasts of plagioclase and potash felspar, peppered with small grains of quartz. The plagioclase is quite fresh and clear, and has a composition range from An 45-An 53 (Emmons' 5-axis method). The composition variation is due to the presence of fine oscillatory zoning. In appropriately oriented grains, the zones are seen to be very narrow, and to outline delicate concentric euhedral crystal forms within the grains.

The zones are separated by sharp compositional changes (Becke lines), and appear to be made up of two plagioclase compositions, one near An 45 and the other near An 53. As the rock is of definite sedimentary-metamorphic origin, the euhedral zones must have been produced in the solid state after the plagioclase grains had grown. The rims of the grains are of a similar composition to the cores, and a theory involving diffusion of more albitic plagioclase in from the rims must be discarded. The most likely alternative is the exsolution of two plagioclases, one near An 45 and the other near An 53. This would imply that there is an immiscibility gap in the plagioclase series over this range of composition. Miyashiro (1958) has proposed an immiscibility gap in the plagioclase series in the range An 40-50 on the basis of observations in the Gosaisyo-Takanuki area of the Abukuma Plateau, Japan. It might be supposed, for the present case, that the plagioclase was once homogeneous, and that during slow cooling, phases near An 45 and An 53 separated out into narrow zones within the crystal structure. Misch (1954) has recorded the occurrence of recurrent zoning in the plagioclase of mica schists and calc-silicate granulites, and he has described an excellent example of sharply euhedral recurrent zoning in the plagioclase of a calc-silicate rock from Nanga

FIG. 46.

Photomicrograph of recurrent oscillation zoning in metamorphic plagioclase of thin amphibole-quartz-labradorite layer in calc-schist.

Rock A185-285. Location 179.119. Crossed polars.

FIG. 47.

Photomicrograph of portion of thin slice of hastingsite-plagioclase calc-silicate from near the base of the sillimanite zone showing granoblastic texture and compositional banding.

Rock A185-704. Location 153.180. Ordinary light.

Central band contains hastingsite, clinopyroxene and epidote with minor plagioclase and iron ore. The adjacent bands are alternately enriched in plagioclase and epidote.

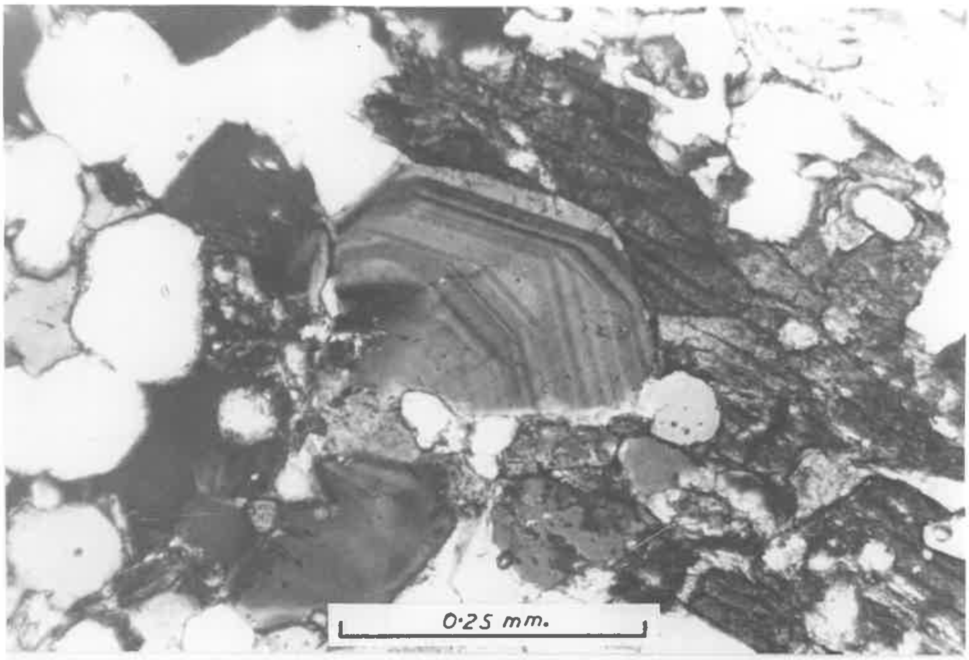


FIG. 46

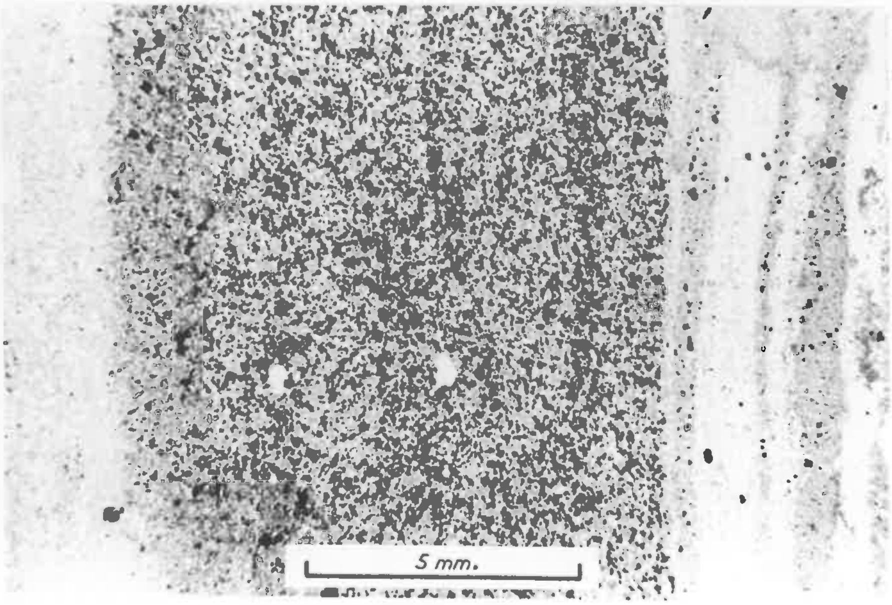


FIG. 47

Parbat, North-west Himalayas (Misch, 1964). In this plagioclase up to 12 recurrences were observed, the alternate zones being in the range An 60-63 and An 84-91. Although of some interest, the cause of this type of zoning remains obscure.

Like the eastern calc-silicates these rocks also show some progressive textural changes with increasing metamorphic grade. The most notable is the polygonization of large skeletal porphyroblasts of scapolite, amphibole and clinopyroxene in the higher grade rocks. This process proceeds in a similar manner to that already described for the eastern calc-silicates.

(b) The Hastingsite-Plagioclase Calc-silicates.

The calc-silicate rocks now to be described are readily distinguished from the clinopyroxene-scapolite rocks by an almost universal occurrence of plagioclase with a dark green-black amphibole, which, from optic data, is considered to belong to the hastingsite series. The more calcic varieties contain, in addition, abundant epidote and clinopyroxene.

The hastingsite-plagioclase calc-silicates outcrop as several closely related long thin beds intercalated within a uniform steeply dipping sequence of quartzo-felspathic schists, and, stratigraphically, lie several hundred feet above the marble bed (Plate 1). These beds may be correlated with several calc-silicate beds, intercalated with quartzo-felspathic schists, above the marble on the eastern limb of the main western anticlinal zone. However, specific correlation of individual beds across the anticlinal zone cannot be attempted. When traced along strike, the hastingsite-plagioclase calc-silicate beds are found to be of variable thickness and composition, and to undergo minor sedimentary facies variations. Well defined beds, rich in calc-silicate minerals, may grade off into more felspathic varieties, and, in some sections, the calc-silicate beds have disappeared from the quartzo-felspathic sequence.

A well defined sedimentary lamination, expressed by mineral banding, is present in almost all beds, each lamination being enriched in a particular mineral or association of minerals. Some rocks are particularly colourful; alternating white, yellow, green and black bands being enriched in felspar, epidote, clinopyroxene and hastingsite respectively. Individual bands are seldom more than a few millimetres in thickness, and because most of the rocks are exceptionally fine-grained, very fine closely spaced laminations may be well preserved. Most of the bands are uniform in thickness, and are continuous over the scale of the hand specimen; but occasionally a band may thicken into a lens, or show discontinuity.

As these beds are located on the western limb of the isoclinally folded anticlinal structure, the tectonic foliation (axial plane foliation of schistosity) is sensibly parallel to the relic sedimentary banding. Only rarely are minor folds observed where the tectonic foliation departs appreciably from the bedding structure.

The majority of rocks have a fine even grained crystalloblastic texture, and only in the higher grade rocks to the south are coarser porphyroblastic varieties conspicuous. In the crystalloblastic rocks the tectonic foliation is clearly outlined by the shape orientation of quartz and felspar granules, and by dimensional orientation of the amphibole prisms. Rarely the amphibole prisms also outline a faint lineation.

The calc-silicate beds outcrop on a constant strike, trending north-north-west for a distance of over 10 miles, and cut obliquely across metamorphic boundaries from mid-staurolite grade in the north, to mid-sillimanite grade in the south. This simple relationship between the strike of the beds and the metamorphic isograds, allows ready comparison of samples of progressively changing grade, the assumption being made that the grade changes progressively and steadily with distance along the beds. A study of mineralogical equilibrium within these rocks, to be elaborated in the following chapter, has confirmed the validity of this assumption.

(1) Minerals and Mineral Assemblages.

The major minerals encountered in these rocks are quartz, plagioclase, epidote, clinopyroxene, hastingsite and potash felspar. Hornblende, biotite, garnet, scapolite and calcite are relatively rare, being found in only a few of the samples examined.

Iron ore (usually magnetite), sphene, apatite and zircon are the commoner accessory constituents. Apart from progressive mineralogical changes, the only new mineral to appear during the metamorphism of these rocks is grossular-andradite garnet. This garnet has been encountered in one specimen from the mid-sillimanite zone. These rocks are consistently found to be remarkably fresh in thin section, and there is generally no evidence for a secondary paragenesis.

The following major assemblages have been observed within these rocks.

<u>Andalusite-staurolite-kyanite zone in pelitic schists</u>. An 26.

(1) Quartz-plagioclase-epidote-hastingsite-clinopyroxene.

<u>Sillimanite-muscovite zone in pelitic schists</u>. An 27.

(2) Quartz-plagioclase-epidote-hastingsite-clinopyroxene-potash felspar.

(3) Quartz-plagioclase-epidote-hastingsite-clinopyroxene-calcite.

(4) Quartz-plagioclase-epidote-hastingsite-clinopyroxene-scapolite.

(5) Quartz-epidote-clinopyroxene-hastingsite-calcite-scapolite.

(6) Quartz-plagioclase-hastingsite-biotite-potash felspar.

(7) Quartz-plagioclase-epidote-hastingsite-clinopyroxene-garnet-scapolite-potash felspar. (Plagioclase An 50).

Most mineral assemblages found in these calc-silicate rocks may be expressed by assemblage (1), or its sub-assemblages. In the upper andalusite-staurolite-kyanite zone and the lower part of the zone of sillimanite in pelitic schists, assemblage (1) is commonly encountered, the minerals being either intimately mixed, or concentrated into alternate bands. Sub-assemblages derived from assemblage (1) make up the majority of rocks examined from all grades. Rocks containing plagioclase and epidote invariably contain clinopyroxene in addition. Assemblages without epidote are especially common in the higher grade rocks, where much of the epidote has disappeared through reactions involving plagioclase and clinopyroxene. Assemblages lacking both epidote and pyroxene are common, and usually have a striking black and white banded appearance due to alternating hastingsite and plagioclase rich bands. Quartz free assemblages are not uncommon. Hastingsite or plagioclase free assemblages are very rare. Rarely hastingsite may be replaced by an amphibole of the hornblende series.

Potash felspar bearing rocks are rare, and are limited to the more felspathic varieties. These rocks may be expressed by assemblage (2), and its sub-assemblages. Sodium cobaltinitrite staining of polished sections has revealed that the potash felspar is commonly concentrated into bands parallel to the bedding. Rocks containing calcite and scapolite are rare, and are represented by assemblages (3), (4), and (5). Biotite bearing calc-silicate rocks, assemblage (6), are transitional towards pelitic schists. Assemblage (7) was found in one rock from the highest grade. Here grossular-andradite has appeared in a quartz-epidote bearing rock. At a slightly lower grade the assemblage quartz-calcite-epidote is stable. Thus garnet has formed within these rocks well within the sillimanite zone, and the epidote-quartz-calcite assemblage appears to be stable to a higher temperature in these rocks than in the calc-silicates to the east, a further observation suggesting increased pressure to the west.

An optical-mineralogical study of these rocks has been undertaken to evaluate epidote-plagioclase equilibrium relationships, and the detailed results will be presented in a following chapter. The epidote has been found to be of almost constant composition, approximately one third of the aluminium positions in the structure being replaced by ferric iron (see also Miyashiro and Seki, 1958). Plagioclases in equilibrium with epidote show a consistent and steady increase in anorthite content, from An 26 at the lowest grade (andalusite, staurolite and kyanite in adjacent pelitic schists), to An 27 (first sillimanite noted in pelitic schists), to An 40 (mid-sillimanite-muscovite zone). Epidote-plagioclase bearing rocks are very rare in the mid-sillimanite zone, but plagioclase in some adjacent rocks suggest that the anorthite content may increase more rapidly once a composition of An 40 has been exceeded. The clinopyroxenes fall within the range 36-70% hedenbergite, and in the epidote bearing assemblages there is a progressive increase of hedenbergite content with grade. The refractive index n_α of the hastingsites ranged from 1.638-1.698. A close correlation between n_α, colour intensity and magnetic susceptibility, suggests that increasing refractive index may be related to increasing iron content. Although there is a wide range of n_α values for hastingsites at any one grade, the n_α values progressively increase with grade, presumably reflecting increasing iron content. Measured $Z^\wedge c$ values for the hastingsites ranged from 13-21°, but there is no apparent correlation with n_α. Optic axial angles across X, for the hastingsites with n_α in the range

1.682-1.698, varied from 19-45°, again with no consistent relation to refringence, although, overall, the 2Vx values decrease with increasing n_{α} values.

(2) Petrography of the Hastingsite-Plagioclase Calc-silicates.

Rock 761 (location 145.241) is typical of the lower grade hastingsite-plagioclase calc-silicates. The rock is a fine green-black to grey laminated crystalloblastic calc-silicate schist, with alternating dark and light layers. The dark layers consist of green, faintly pleochroic clinopyroxene, as small 0.5 mm. porphyroblasts, intermingled with fresh sub-idioblastic prisms of blue-green magnesium-rich hastingsite, with the pleochroic scheme X-pale lemon, Y-clear mustard-green, Z-fresh blue-green and n_{α} 1.638, and rare pale yellow granules of epidote. A small amount of quartz forms an interstitial matrix to the dark minerals. The light coloured laminations are principally composed of fine (0.02 mm.) granoblastic quartz and plagioclase, full of minute inclusions of diopside, hastingsite and sphene. The plagioclase (An 26) is only distinguished from the quartz with difficulty. Apatite and iron ore occur as accessory constituents. The epidote granoblasts contain minute inclusions of indeterminate character. This feature is common to the higher grade epidotes also, and indicates that the epidote grew at an early stage in the metamorphic history, replacing a fine-grained matrix, while clinopyroxene and hastingsite, which contain only rare inclusions of a larger size, are considered to have grown later. In the higher grade rocks the epidote grains have clear rims resulting from increased grain growth.

In some rocks the dark bands are alternatively enriched in epidote and clinopyroxene, hastingsite appearing in small amounts at the boundaries of the epidote and pyroxene rich layers. The lighter coloured quartz-felspar bands are consistently fine-grained (0.02-0.04 mm.), granoblastic, and filled with minute inclusions of the mafic constituents of the rock.

In other rocks the light and dark minerals are intimately intermingled, and distinct light and dark layers are not clearly discernible, although a good grain shape orientation of the amphibole outlines the tectonic foliation.

Another variety of calc-silicate consists of alternating white quartz-plagioclase bands and coal black iron-rich hastingsite bands. Sub-idioblastic magnetite is common throughout both light and dark bands.

At the grade where staurolite has disappeared, and sillimanite is common in the pelitic schists, the grain-size of the calc-silicates has increased, and some of the clinopyroxene and hastingsite tends to grow as porphyroblasts.

Layers containing calcite granoblasts occur in rock 695 (location 154.183). In this rock, alternate layers contain clinopyroxene and epidote, and again hastingsite has grown at the contact between these layers, suggesting that the amphibole has formed by a reaction involving the epidote and pyroxene. This reaction is controlled by the plagioclase composition, and epidote and clinopyroxene grains may occur side by side in equilibrium.

Rock 697 (location 155.174) is a crudely banded black and white rock in which irregular slightly polygonized xenoblasts of scapolite are sparsely distributed. The grain size of this rock is considerably larger than any lower grade rocks, the quartz and felspar bands having an average grain size of 0.2 mm., while the darker bands have coarse hastingsite porphyroblasts up to 5 mm. across. Clinopyroxene commonly occurs as optically continuous granules enclosed within the hastingsite. This indicates that the amphibole has overgrown and replaced former skeletal pyroxene xenoblasts. This relationship has also been commonly observed in other rocks containing coarse hastingsite xenoblasts. Narrow pale-green rims of a higher birefringence secondary amphibole have been observed on some hastingsites.

A colourless to pale-green secondary actinolite is common in rock 355 (location 156.170), a well crystallized granoblastic quartz-epidote rock containing strings and patches of quartz, felspar and hastingsite. Near patches of limonite replacing original pyrite, the hastingsite has become bleached and recrystallized to actinolite. The actinolite has a lower $Z^\wedge c$ value and a higher birefringence than the hastingsite.

Rock 320 (location 166.119) is also a granoblastic quartz-epidote rock, which contains, in addition, clots of clinopyroxene and hastingsite granules and large skeletons of scapolite undergoing polygonal recrystallization. The presence of primary calcite indicates that grossular-andradite has not yet formed in this quartz-epidote-calcite assemblage.

Rock 147 (location 179.086), from a slightly higher grade, contains small sub-idioblastic grains of fresh pale orange-pink garnet (a grossular-andradite, $n = 1.796 \pm .001$). The rock is a dense green-grey and pink speckled calc-silicate, with a moderate foliation and a vague layering. Large porphyroblasts of clinopyroxene, to 3-5 mm., and rarer hastingsite porphyroblasts to 1 mm., are embedded in a semi-granoblastic quartz-plagioclase-scapolite-clinopyroxene-hastingsite matrix. The garnets, along with sub-idioblastic, isolated, or bunched crystals of epidote, are evenly distributed throughout. Calcite is absent. The epidote is pleochroic from pale-yellow to pale-green, has anomalous bright interference colours, and in the outer portions of the crystals, a fine, but strongly developed, euhedral zoning. The plagioclase grains are also zoned, with cores of An 40 and rims of An 54. The cores are often separated from the rims by a sharp compositional break, again suggesting a possible immiscibility gap in the region of An 40-50 (Miyashiro, 1958).

Rock 791 (location 188.080) is typical of rocks transitional from hastingsite calc-silicates to pelitic schists. Rare hastingsite granoblasts are dispersed through a moderately foliated quartz-felspar-biotite matrix. The biotite has the pleochroic scheme X-bright yellow, YZ-very dark brown, and the hastingsite X-pale yellow, Y-mid-green, Z-very deep green.

Metamorphic segregation veins may be considered absent from the hastingsite-plagioclase calc-silicates, but thin veins of coarse bladed hastingsite cross-cut one low grade outcrop along joint directions (rock 776, location 148.222). The blades of hastingsite are approaching 2 cm. in length, have a strong tendency to grow perpendicular to the walls of the vein, and enclose minor plagioclase in their interstices. The contacts of the vein with the fine-grained granoblastic clinopyroxene, quartz, plagioclase, hastingsite calc-silicate, are irregular, and are marked by a selvage of fine-grained hastingsite. Within the vein the large closely grown prisms of amphibole have the following properties: n_{α} 1.665, X-pale lemon, Y-deep mid-green, Z-deep green-blue. The colour of the mineral masks the birefringence. Faint colour zoning may occur within the prisms, and some show twinning on $\{100\}$. The $\{110\}$ cleavages are closely spaced and well developed. The edges and the ends of the prisms are epitaxically overgrown by a thin rim of a higher birefringent, pale green to neutral, non-pleochroic amphibole, with a similar extinction angle to the hastingsite. Some irregular fractures across the prisms within the vein, also contain this pale green amphibole.

Conclusions.

The hastingsite-plagioclase calc-silicates are characterized by the assemblage hastingsite-plagioclase, in many cases accompanied by clinopyroxene, epidote, quartz and potash felspar. Epidote decreases in amount with increasing grade. Calcite and scapolite are rare constituents. The textures are usually fine-grained, directed and granoblastic, becoming slightly coarser grained and more porphyroblastic at higher grades.

The only distinctive reaction occurring with increasing grade is the formation of grossular-andradite garnet high in the sillimanite zone. A complex progressive reaction between epidote, clinopyroxene and plagioclase, proceeding under increasing temperatures, leads to the calcification of the plagioclase, destruction of epidote, and perhaps some clinopyroxene, and the growth of hastingsite. Both clinopyroxene and hastingsite become more enriched in iron with grade, while the epidote remains constant composition, with one third of its possible aluminium content replaced by ferric iron. Table 7 compares the main features of the hastingsite-plagioclase calc-silicates with the clinopyroxene-scapolite calc-silicates.

As these hastingsite-plagioclase calc-silicate rocks are stratigraphically equivalent to diopside-scapolite calc-silicate rocks to the east, it is interesting to consider what has caused the mineralogical differences between these rock types. The rarity of scapolite in the hastingsite-plagioclase rocks might, at first, suggest that the volatiles necessary for the production of scapolite are absent.

TABLE 7.

COMPARISON OF HASTINGSITE-PLAGIOCLASE CALC-SILICATES AND CLINOPYROXENE-SCAPOLITE CALC-SILICATES.

Hastingsite-plagioclase calc-silicates.	Clinopyroxene-scapolite calc-silicates.
Mafic components.	
Hastingsite-clinopyroxene-epidote.	Hornblende-clinopyroxene-biotite.
Felsic components.	
Quartz-plagioclase-(potash felspar).	Quartz-scapolite-(plagioclase)-(potash felspar).
Rarity of scapolite and calcite.	Abundance of scapolite and calcite.
Rarity of potash felspar.	Abundance of potash felspar.
Abundance of plagioclase and epidote.	Rarity of plagioclase and epidote.
Simple directed granoblastic texture.	Complex non-directed intergrown porphyroblastic textures.
Fine bands of variable composition well preserved to high grades.	Fine bands of low grade calc-schists become progressively destroyed at higher grades.

However, hastingsites may contain high proportions of chlorine and other volatiles, and Mr. J. Hutton of the C.S.I.R.O., Division of Soils (personal communication), has shown that a very dark ferrohastingsite from Tungkillo (n_{α} 1.711) has a chlorine content of 1.6%, and that other hastingsites, from both the Tungkillo and the present area, are chlorine bearing. Perhaps soda, potash or ferric iron contents are more important in promoting the differences between these rocks. If the adjacent arkosic sequence has had much of its potash leached out and replaced by soda, as was concluded previously, then it is possible that the western calc-silicate rocks were also affected, leading to an enrichment in soda and a consequent production of hastingsite in place of biotite or hornblende. Until a suite of chemical analyses becomes available for both rocks, this problem will not have a satisfactory solution.

CHAPTER 8.

MINERALOGICAL STUDIES IN CALC-SILICATE ROCKS.

A. Scapolite.

Introduction.

Shaw (1960) has recently summarized existing data on scapolite structure, occurrence and composition. It is evident that scapolite has a wide composition range, and physical properties can only be related to chemical composition with considerable approximation. Although a large number of scapolite analyses are now available, it is still not clear why certain composition ranges are favoured within different environments. Barth (1952, p. 284) has presented a tentative diagram illustrating plagioclase-scapolite equilibrium with temperature, the plagioclase being considered to be always more sodic than the scapolite with which it is in contact. There is little evidence to support the existence of such a relationship (Turner and Verhoogen, 1951, p. 590; Shaw, 1960, p.279).

The object of the present study was to determine the scapolite composition range in various rocks and to determine whether there was any consistency between scapolite composition and metamorphic grade, rock type or associated plagioclase composition.

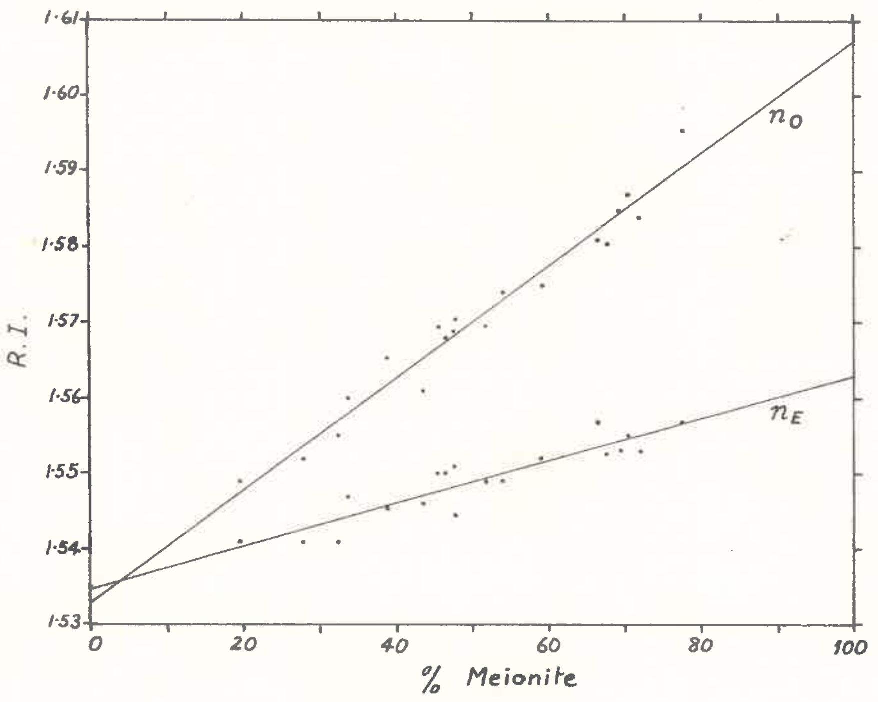
(a) Mineralogical Data.

Shaw (1960) has recommended that the mean refractive

FIG. 48.

Graph of scapolite composition versus refringence variation. The regression curves for n_O and n_E were drawn from Shaw's "reliable" optic data on scapolites.

FIG. 48


1·61
1·60
1·59
1·58
1·57
1·56
1·55
1·54
1·53
R. I.
n_O
n_E
0
20
40
60
80
100
% Meionite

indices be used in optic composition determinations, although his conclusions on page 279 suggest that this is no better than using n_o. The refractive index n_o was determined for 58 scapolites from various rocks in the present area. The results are presented in Appendix I. Using the 'reliable' analyses of Shaw (1960), a graph relating R.I. to composition in terms of percent marialite and meionite molecules was obtained (Fig. 48). The method of least squares yielded the equations

$$n_o = 1.5333 + 0.00074019 \text{ \% meionite}$$
$$n_E = 1.5349 + 0.0002806 \text{ \% meionite}$$

Using the n_o curve a composition range of Me 43-70 is indicated for the 58 samples studied, most specimens falling in the range Me 45-60. Fig. 49 illustrates the geographic location of the various compositions. There is evidently no consistency of composition with grade of metamorphism. Data presented in Appendix I also indicates that there is no relation between scapolite composition with rock type or associated plagioclase composition.

The limited scapolite composition range, Me 45-60 for the majority of scapolite, is interesting. White found a composition range of Me 55-60 in the Tungkillo and Milendella areas. As a wide variety of rocks have been examined in the present study, including scapolite bearing pelitic schists, calc-schists, calc-silicate rocks, limestones, scapolite veins and replacement scapolite in metadolerites, etc., the

FIG. 49.

Geographic location of determined scapolite compositions. The compositions are given in terms of percent meionite determined from the n_o refractive indices and Fig. 48.

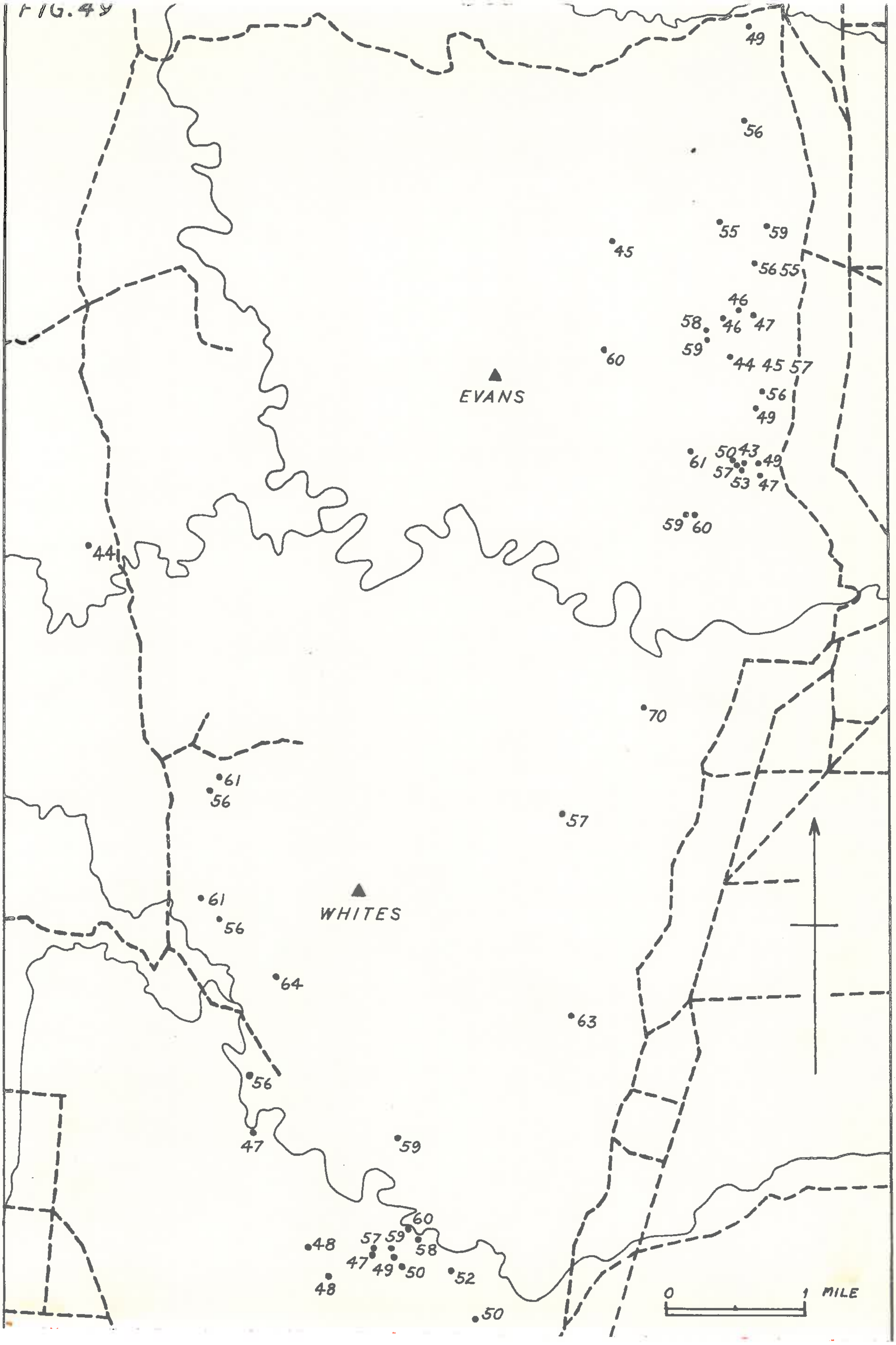


FIG. 49
EVANS
WHITES
0
1 MILE
49
56
55
59
45
56 55
46
58
46
47
59
60
44 45 57
56
49
61
50 43
49
57
53
47
59
60
44
70
61
56
57
61
56
64
63
56
47
59
60
48
57 59
58
47
49
50
52
48
50

limited composition range cannot be related to bulk composition. Perhaps the limited range may be related to the stabilizing influence induced by a particular scapolite structure favoured under metamorphic crystallization conditions. However Parras (1958) records 12 refringence values for scapolites from the West Uusimaa Complex, Finland. These indicate a composition range, using the present n_o curve, of Me 72-85. These scapolites have formed under conditions of the granulite facies. Scapolites from the Cloncurry District, Queensland (Edwards and Baker, 1953) have a composition range of Me 25-30. Perhaps pressure variables, such as the relative partial pressures of CO_2 and Cl_2 are important factors in controlling the scapolite composition range within a given metamorphic belt.

Scapolite has often been considered as a pneumatolytic mineral (e.g. Edwards and Baker, 1953; Sundius, 1915), but some authors have concluded that scapolite can be of purely metamorphic origin (e.g. Parras, 1958; von Knorring and Kennedy, 1958). Shaw (1960) has indicated that scapolite is stable over the whole range of metamorphic P-T conditions. White (1959) has considered that scapolite in calc-silicates and marbles in the Milendella and Tungkillo area, south and south-west of the present area, has a sedimentogenous metamorphic origin, there being no indication of the introduction of volatiles during metamorphism.

The uniformly bedded nature of the scapolite bearing rocks in the present area is in support of a sedimentary-metamorphic origin for the scapolite at Cambrai. There is no relation at all between the occurrence of igneous or metasomatic rocks and the occurrence of scapolite. The appearance of scapolite in joint veins and as a replacement mineral of basic plagioclase in those metadolerites intruding calc-silicate rocks, is regarded as a result of the mobilization of volatiles from the adjacent intruded scapolite bearing rocks (which show a slight coarsening of grain at the metadolerite contacts). Metadolerites intruded into rocks free of scapolite are scapolite free. It is interesting that scapolite is also widespread in the metamorphosed portions of the Adelaide System rocks (e.g. Williamstown, Mills, 1963; Yankaninna, Mawson and Dallwitz, 1945). In conclusion the present writer is in agreement with White with regard to the origin of scapolite in this region.

B. Plagioclase-Epidote Equilibrium Studies.

Introduction.

A reaction relationship between plagioclase and epidote has been known to exist for many years (e.g. Ramberg, 1944), although relatively few field and laboratory studies have been specifically directed towards this problem. A recent study by De Waard (1959) has sparked off a number of theoretical papers on this topic (e.g. Christie, 1959; Rutland, 1961; Noble, 1962; Christie, 1962; Rutland, 1962; Brown, 1962; Sen, 1963; Kretz, 1963), although field and laboratory data is still sadly lacking. Brown (1962) has reviewed the existing data on plagioclase-epidote relationships in regard to type and degree of metamorphism, especially for the peristerite region.

(a) The aims of the present Study.

The present study attempts to add further practical data to the existing knowledge of plagioclase-epidote equilibrium relationships. The Cambrai area is essentially a sodic meta-sedimentary province. Sodic plagioclase is one of the major minerals of the area, being common in all rock-types except marbles and calc-silicates. Primary epidote in stable equilibrium with plagioclase is confined to a narrow belt of calc-silicate beds on the western side of the western belt of marble outcrops. These calc-silicate rocks are also characterized by the presence of a very dark amphibole which appears, from optic data, to be hastingsite. The petrology of the hastingsite-plagioclase calc-silicates has already been described (Chapter 7). The purpose of this study is to elucidate plagioclase-epidote equilibrium relationships and changes in the associated minerals with changes in metamorphic grade.

(b) The Hastingsite-Plagioclase Calc-silicates.

(1) Occurrence and relation to metamorphic grade.

Unfortunately the epidote bearing beds do not extend much below the sillimanite isograd in the mapped area, and most of the rocks studied lie within the sillimanite-muscovite zone. The lowest grade rocks have been recrystallized under P-T conditions close to the triple point of the aluminosilicates ($T\ 350^{o}C$ P 8.5 Kb., see Chapter 4). The hastingsite-plagioclase calc-silicates crop out along a constant strike for over 10 miles and lie obliquely across linear metamorphic isograds (e.g. the sillimanite isograd) from the top of the andalusite-staurolite zone to near the top of the sillimanite-muscovite zone. This simple relationship between the strike of the beds and the metamorphic grade allows ready comparison of samples of progressively changing grade, the assumption being made that the grade changes progressively and steadily with position along the beds. Where possible samples of well reconstituted metadolerite intrusions, adjacent to the calc-silicate beds, were also taken along with the calc-silicate samples.

The mineral assemblages and petrology of these rocks has been fully described in chapter 7. There it was pointed out that all assemblages having plagioclase in equilibrium with epidote also have clinopyroxene as an essential mineral. The dark-green hastingsitic amphibole is also present in most rocks and appears to partially replace the clinopyroxene with increasing grade. It was also noted that although these rocks also have a considerable range of bulk composition the only discontinuous mineral change encountered in the range of metamorphism studied, apart from the gradual equilibrium changes within the mineral groups, was the appearance of garnet in an epidote bearing rock high in the sillimanite-muscovite zone.

The rocks under study are generally strongly banded or layered, each band presenting concentrations of different minerals. These bands appear to represent relict sedimentary laminations, which, through metamorphic recrystallization and differentiation, have evolved into a pronounced gneissic structure. The rock textures are granoblastic with a strong tendency for planar grain contacts between minerals. Lack of zoning in all minerals but those of a few rocks, and lack of evidence for a secondary mineral development, suggests good mineralogical equilibrium. Except for the high grade garnet bearing rock, there is no evidence for plagioclase-epidote disequilibrium within these rocks.

(2) Sampling.

The collection of samples was highly selective, especially at the higher grades of metamorphism. The main object during the sampling was the collection of rocks having both epidote and plagioclase as major constituents. Some samples were also attractive because of their abundance of amphibole. Within the lower part of the sillimanite-muscovite zone both epidote and plagioclase were common in most outcrops, and sampling was more random. Higher in the sillimanite zone, because of the rarity of epidote, all locations where epidote was found were sampled.

At some locations several samples were collected and studied. Where these samples were found to show different compositions for a particular mineral they were graphed as separate points in the figures, but where compositions were identical for a particular mineral they were graphed as only one point. That is, there are slightly fewer points on the figures than are recorded in the appendices.

(3) The data.

During this investigation 168 refractive index determinations, mostly considered accurate to $\pm$ .001, have been made. The results are recorded in Appendix II, and have been graphed in Fig. 50 where ordinates represent mineral composition, and abcissae, grade of metamorphism (largely temperature increase).

(4) Mineralogical determination and accuracy.

Because of the fine grain size of many rocks (e.g. 0.025 mm.), and the lack of twinning in the plagioclase grains, refractive index determinations were chosen in preference to universal stage determinations. For each mineral refringence range, oils were prepared sufficiently close to give a refractive index to ± .001 in normal clear grains. White light and sodium light were used in conjunction. Conditions were fairly constant during the determinations and oils were checked after each match with a refractometer kept next to the microscope. Periodic checks were made on the constancy and accuracy of the refractometer determinations.

Although subject to slight errors, such as lack of absolute temperature control, the method used was found to be much more rapid than the double variation technique. With the precautions taken most determinations are believed to be accurate to ± .001 and are of sufficient accuracy for the type of work being done. In some cases the presence of alteration products, numerous minute inclusions or grain coatings has introduced errors in excess of ± .001. Where errors were likely to be higher than ± .002 the determinations were abandoned.

For the plagioclase, n_β was measured and compositions determined using curves from Hess (1960). n was chosen because most determinations were made in the presence of small to large, and sometimes unknown quantities of quartz, potash felspar, scapolite as well as some mafic minerals. Centred optic axis figures could be found and readily identified as belonging to plagioclase by the almost straight isogyre (high 2V). Such grains did not need orientation for measurement and have the advantage of being able to be rotated during the oil match determination. Although n_α or n_γ may show a more consistent refringence variation with composition, interference figures containing n_α or n_γ would be difficult to recognise as belonging to plagioclase. Cleavage curves could not be used because of the general lack of cleavages and twinning.

For the epidote and clinopyroxene, optic axis figures were also employed. Because of the poor cleavage development and the equidimensional shape of most grains, little trouble was experienced in finding centred optic axis figures. Epidote and clinopyroxene optic axis grains were readily identified by colour and by faint anomalous blue interference colours in the epidote figures.

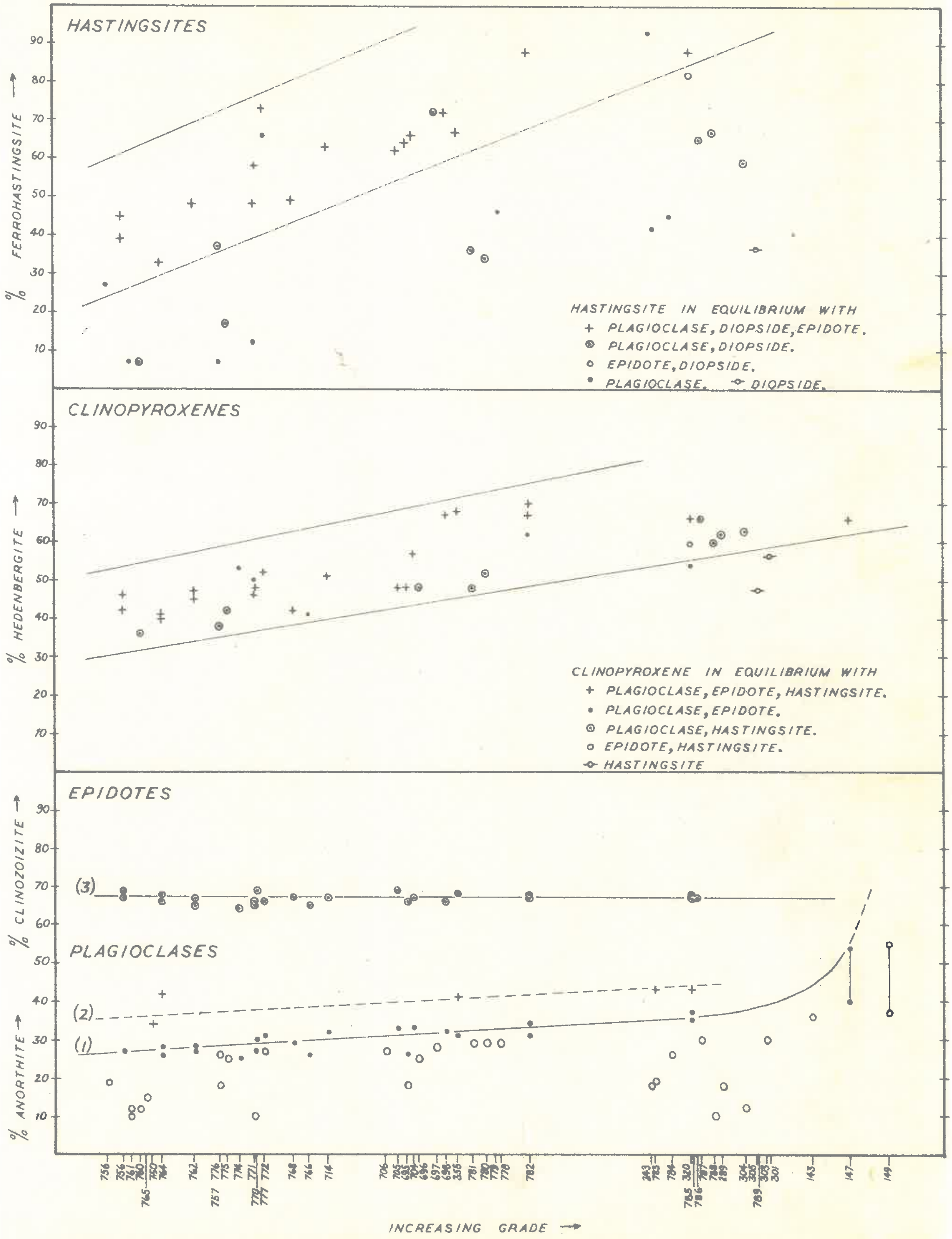
For the amphiboles, n_α was chosen and measured in flash figures. Because of the extremely dark colour and the strong absorption in the Y and Z directions, it was impossible to obtain n_β or n_γ in most grains. Flash figures were not uncommon and light clearly penetrated the Y direction allowing an oil match to be determined.

For epidote, diopside and hastingsite the curves

FIG. 50.

Optical composition variation in minerals of the hastingsite-plagioclase calc-silicates. The sample numbers are listed in order of increasing grade (temperature) along the beds. For explanation and discussion see text.

FIG. 50


HASTINGSITES
% FERROHASTINGSITE →
HASTINGSITE IN EQUILIBRIUM WITH
+ PLAGIOCLASE, DIOPSIDE, EPIDOTE.
PLAGIOCLASE, DIOPSIDE.
EPIDOTE, DIOPSIDE.
PLAGIOCLASE.
DIOPSIDE.
CLINOPYROXENES
% HEDENBERGITE →
CLINOPYROXENE IN EQUILIBRIUM WITH
+ PLAGIOCLASE, EPIDOTE, HASTINGSITE.
PLAGIOCLASE, EPIDOTE.
PLAGIOCLASE, HASTINGSITE.
EPIDOTE, HASTINGSITE.
HASTINGSITE
EPIDOTES
% CLINOZOIZITE →
(3)
PLAGIOCLASES
% ANORTHITE →
(2)
(1)
INCREASING GRADE →

given in Troger (1956) have been used to estimate mineral composition.

(5) Examination of results.

Fig. 50 shows three curves,

(1) composition of plagioclase with increasing grade in the metasediments

(2) composition of plagioclases with increasing grade in metadolerites

(3) composition of epidotes with increasing grade in metasediments.

a. Plagioclase from metasediments.

1. Plagioclase in equilibrium with epidote.

These determinations have been plotted as black dots in Fig. 50, curve (1). The points outline a well defined curve, which has been drawn by inspection for best fit. This curve represents portion of the sub-solidus curve for plagioclase-epidote equilibrium in the amphibolite facies of the Cambrai area.

Two points may be noted concerning this curve.

(a) The curve appears to be a straight line between An 26-An 35. No values fall more than 5% off the empirical curve.

(b) Above An 35 the curve begins to increase in gradient, but cannot be accurately defined because of the extreme rarity of epidote bearing rocks at this high metamorphic grade. Only one rock was found having epidote with plagioclase > An 35. In this specimen, the epidote, although strongly zoned, is present in sufficient quantity to suggest possible equilibrium with the plagioclase. However, some secondary epidote is also present. The plagioclase is strongly zoned in reverse, with rims of An 54 and cores of An 40. This indicates rapid changes in the mineral equilibrium. The plagioclase anorthite values of some nearby amphibole bearing metasediments of this grade reach An 84.

2. <u>Plagioclase not in equilibrium with epidote</u>.

These are plotted as circles in Fig. 50. All fall below the equilibrium curve as expected. Two main conclusions may be stressed,

(1) no evidence for a subsolvus gap down to An 10,

(2) in some rocks the plagioclase is

notably zoned by

(a) reverse zoning in high grade specimens or

(b) normal zoning in several specimens in the middle sections of the graph. This may be attributed to a local late introduction of soda.

b. Plagioclases from metadolerites.

These plagioclases have been plotted as crosses in Fig. 50 curve (2). They appear to define a curve lying at higher anorthite values (9% higher) than that of the associated metasediments. In most rocks the plagioclase is somewhat normally zoned, suggesting that these rocks have not quite reached equilibrium with their metamorphic environs. Primary epidote is not present in these rocks.

c. Epidotes from the metasediments.

These have been plotted in terms of the ferric iron molecule in Fig. 50, curve (3). Throughout the grades of metamorphism studied no real variations were perceptible. All values of n_β fell near the iron end of Tröger's curves at approximately 33% iron molecule. Slight variations in different assemblages can be recognised.

(a) Epidotes in equilibrium with plagioclase-clinopyroxene-hastingsite have compositions ranging from 31-35% Fe''' molecule (20 determinations), with an average composition of 33% Fe''' molecule.

(b) Epidotes in equilibrium with clinopyroxene-hastingsite (quartz or scapolite) have similar compositions (32% Fe''' molecule) to those also having plagioclase in their assemblages.

(c) Epidotes in equilibrium with plagioclase and clinopyroxene (but no hastingsite) are slightly higher in iron contents (34, 35, 36% Fe''' molecule).

d. <u>Clinopyroxenes</u>.

Fig. 50 shows a plot of clinopyroxene composition versus grade of metamorphism. Several features may be noted.

(a) A strongly limited composition range (20-25%) for a particular grade of metamorphism.

(b) A general rise of iron content with grade up to 45-70% hedenbergite, and constancy of composition thereafter.

This figure also presents a subdivision of the clinopyroxenes into types depending on the equilibrium association.

Clinopyroxenes in equilibrium with plagioclase-hastingsite-epidote seem to show a general rise in iron content with grade (within the lines indicated).

Clinopyroxenes in equilibrium with plagioclase and hastingsite but no epidote generally show slightly lower iron contents than those which also contain epidote.

Clinopyroxenes in equilibrium with only hastingsite or hastingsite-epidote are rare.

e. Hastingsite amphiboles.

Fig. 50 shows the compositions of hastingsites in terms of metamorphic grade. There appears to be a general increase in iron content of the hastingsites with grade. However there is a wide band of compositional possibilities for any one grade.

Hastingsites with more than 80% Fe''Fe''' molecule are limited to higher grades.

Hastingsites with less than 30% Fe''Fe''' molecule are limited to lower grades.

This figure also shows the compositions of hastingsites subdivided according to the equilibrium associations.

Hastingsites in equilbrium with plagioclase-clinopyroxene-epidote show a much restricted range of composition for a particular metamorphic grade (within the lines indicated).

Hastingsites in equilibrium with plagioclase and clinopyroxene but no epidote, show a tendency for low iron content compared with those which also have epidote.

The validity of correlating iron content of the hastingsites with the refractive indices is indicated by a close correlation between the magnetic susceptibility and the refringence values (Fig. 51).

(6) Discussion.

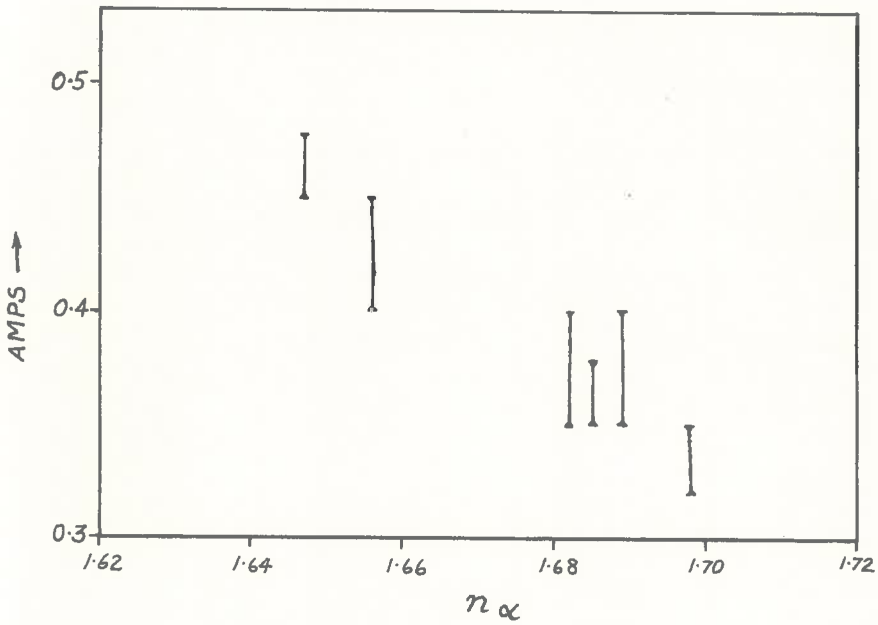
a. Plagioclase.

The elucidation of the plagioclase-epidote equilibrium relationship is basically a field problem, with adequate field control on the collection of samples in relation to metamorphic grade an essential pre-requisitie to the study. Important field studies of plagioclase-epidote

FIG. 51.

Relation between magnetic susceptibility and refractive index in hastingsitic amphiboles. The magnetic susceptibility was measured as a function of the range of amperage setting on a Franz Isodynamic Magnetic Separator, over which hastingsitic amphibole was removed as a magnetic fraction when the separator had a forward slope of 30° and a side slope of 15°.

FIG. 51



equilibrium have been made by De Waard (1959), Compton (1955), Lyons (1955), Lambert (1959), Ambrose (1936) and others. Our present knowledge of plagioclase-epidote equilibrium is not only hampered by a general lack of field-laboratory data, but also by the poor presentation of much of the information vital to the study. Because plagioclase-epidote equilibrium is deeply involved with the other minerals in the rock assemblage, it is imperative that assemblages, textures and possible mineral reactions be completely described (Kretz, 1963).

Considerable advance has been achieved on a theoretical front. The hypothetical equilibrium curve introduced by Ramberg (1944) has been considerably modified through laboratory work on plagioclase-epidote stability and subsolidus relations, structure analyses of plagioclases, order-disorder studies and general knowledge from field data (Christie, (1959, 1962); Noble, 1962; Rutland, 1962; Brown, 1962; and Sen, 1963).

The peristerite subsolvus region near sodic oligoclase is believed to correspond to a gap in the plagioclase composition range which has often been recognised in petrological studies of regionally metamorphosed rocks (De Waard, 1959;

Ambrose, 1936; Turner, 1933; Miyashiro, 1958; Seitsaari, 1952 and many others). The sizes of the immiscibility and composition gaps have been considerably debated (e.g. Brown, 1962; Sen, 1963). It has been fairly well established that the composition gap in the plagioclase series in areas of low pressure metamorphism is small or non-existent (e.g. Compton, 1958; Rutland, 1962; Brown, 1962). Christie (1962) has considered the possibility of a second miscibility gap in the region An 35-50, but so far there is insufficient evidence in support of this.

Data from the Cambrai area can only apply to that part of the equilibrium curve between An 27 and An 35. Over this range the curve is almost a straight line. No values fall more than 5% An off the curve. This confirms the close approach to equilibrium suggested by the rock textures. Lyons (1955) studied 75 rocks from the Hanover Quadrangle (New Hampshire-Vermont) and found some values up to 30% anorthite off the predicted curve of Ramberg. Ramberg (1944, p.54) states that the presence of calcite will favour sodic plagioclase and that variations in K,Al,H_2O, or PP.H_2O will also affect the equilibrium curve. In the Cambrai samples the most aberrant plagioclase (695) contains minor calcite. This sample has deviated towards more sodic plagioclase and may support Ramberg's statement concerning the presence of calcite. A sharp change in the gradient of the equilibrium curve may begin at An 35, with plagioclase becoming rapidly more calcic within a narrow temperature interval.

Noble (1962) graphed Lyons' (1955) data for plagioclase not in equilibrium with epidote and found that the immiscibility gap around sodic oligoclase was reflected in these plagioclases as well as in those plagioclases in equilibrium with epidote. No obvious gap exists in the Cambrai data for those plagioclases not in equilibrium with epidote. In any case, at this grade of metamorphism the equilibrium curve may be above the unmixing solvus. Being a zone of low pressure intermediate type metamorphism, perhaps the gap is too small to recognise with the number of determinations made.

In the range An 27-35 in the metasediments, the plagioclases of the associated basic schists (metadolerites) are constantly higher in anorthite content (9%). De Waard (1959) found that in the range An 10-34, the pelitic schists are higher in anorthite than the basic schists, and that at An 34 the curves crossed. The differences in the Cambrai data may be due to incomplete equilibrium in the rocks.

b. Epidote.

A number of workers have reported on epidote compositions in metamorphic rocks of varying grades (Tilley, 1923; Vogt, 1927; Sugi, 1931; Turner, 1933; Ambrose, 1936; Hutton, 1940, Lyons, 1955; Shido, 1958; and Lambert, 1959). In all these cases except Lyons and Shido, the lowest grade epidotes have been iron rich, with the iron content decreasing with metamorphic grade. Ambrose and Hutton found variable epidote compositions in low grade rocks. Lyons found up to 20% iron molecule ; Shido, variations from 8% to greater than 40% iron molecule, but neither found systematic variations with grade. Lambert (1959, p. 576-7) states "it is only possible to conclude that the iron content of epidote may only be used with severe restriction to indicate the metamorphic grade."

Results from the Cambrai area seem to be unique. Epidotes from all grades have a remarkably constant composition (33% Fe''' molecule 3%). Such constancy in composition is notable when the variability in composition of the associated clinopyroxenes and amphiboles is considered. Miyashiro and Seki, (1958) have shown, on structural grounds, that epidote with a composition of 33-1/3% iron molecule should be more stable than other epidote compositions. They also present evidence that epidote enlarges its composition field with metamorphic grade, but there is no indication of this in the present area. The constancy of epidote composition also suggests that the hastingsite-plagioclase rocks closely approach equilibrium.

c. Clinopyroxenes.

Refringence determinations of clinopyroxenes from these rocks indicate a composition range from Hd 37-70. There is a strongly limited range of composition for a particular grade of metamorphism.

White (1959) records a pyroxene from the Milendella marble with 40% hedenbergite, and a variation of 15-80% hedenbergite in the Tungkillo calc-silicates. He concludes that there is "no systematic variation from one assemblage to another or from one environment to another."

The limitation of the composition of the clinopyroxenes in the hastingsite-plagioclase rocks in the present area is believed to be closely tied to the equilibrium reactions involving plagioclase and epidote.

d. Amphiboles.

Wiseman (1934) demonstrated that amphiboles in the Grampian Highlands increased in Fe''/Mg with grade and that associated chlorite and actinolite varied systematically in composition. Ramberg (1952, p. 344-349) expected from empirical and theoretical reasoning that Fe''/Mg and Al^{IV} (replacing Si^{IV}) in amphiboles would increase with metamorphic grade. Lyons (1955) found no systematic variation with grade of 34 amphiboles from the Hanover Quadrangle. The progression actinolite - blue-green hornblende - green-brown hornblende is well known from many metamorphic regions and is used widely in Japan to grade metamorphic rocks. Shido (1958) and Miyashiro (1958) have considered compositional variations in some metamorphic calc-amphiboles from the Abukuma Plateau, Japan. Miyashiro found no evidence of systematic composition variations within any one amphibole group with metamorphic grade. Shido (1958, p. 204-209) found a tendency for the refractive index to increase with grade. She found that alkalies and titanium increased with grade, although Al^{IV} did not.

No studies on the variations in hastingsite composition with grade are known to the writer. n_{α} of hastingsites from the present area vary from 1.638 - 1.698 (7-93% Fe''Fe'''mol. according to Troger's curves, 1956). There is a tendency for these amphiboles to increase in refractive index with grade, but there is a large range at any one grade.

e. <u>The plagioclase-epidote equilibrium reaction.</u>

Many, but few satisfactory reactions have been proposed to explain the reaction relationship between epidote and plagioclase (e.g. Ambrose, 1936; Turner, 1948; Ramberg, 1949; Barth, 1952; Shido, 1958; Fyfe, Turner and Verhoogen, 1958; Kretz, 1963).

Epidote-plagioclase equilibrium has been closely approached in the Cambrai area. Except for the formation of garnet in the highest grade rocks, no new minerals appear in the assemblages during the equilibrium reaction. The absence of quartz in some of the Cambrai rocks invalidates many of the proposed reactions. The equilibrium reaction must occur through continuous adjustment of the composition of minerals involved in the assemblages. Thus the change in clino-pyroxene and amphibole with grade must be closely tied with the calcification of the plagioclase and the destruction of epidote. Epidote, of constant iron content becomes rarer as the grade increases. The destruction of the epidote not only involves the formation of the anorthite molecule, but also the release of the iron epidote molecule. The pyroxenes and amphiboles appear to increase in ironcontent with grade. Because the intricate chemical changes involved in the mafic minerals (especially the amphibole) have not yet been investigated, the chemistry of the epidote-plagioclase reaction cannot be completely evaluated at this stage.

(c) Other Plagioclase-Epidote Data from the Cambrai Area.

On Plate 2 (metamorphic mineral isograds) a number of plagioclase determinations from various rocks in the Cambrai area have been plotted. Those rocks considered to have epidote and plagioclase in equilibrium are indicated. The rarity of plagioclase in equilibrium with epidote in rocks on the eastern side of the area is notable, as is the more rapid increase in anorthite content of plagioclase in these rocks. This is further evidence suggestive of increased pressures in metamorphism in passing from the eastern to the western portions of the mapped area.

CHAPTER 9.

METAMORPHIC SEGREGATION VEINS

Introduction.

Throughout the Cambrai area, numerous veins and pods of quartz, carrying metamorphic minerals similar to those in the wall-rock, have been observed. These are believed to be metamorphic segregations which bear no relation to the intrusive rocks to be described in chapter 11. Several kinds of metamorphic segregation veins have been recognised. Aluminosilicate-quartz segregations are the most common, these being confined to the aluminium rich micaceous schists. In some micaceous garnet schists garnet-quartz segregations have been observed. Within the calcite-mica schists small muscovite-quartz-calcite segregations may form lenses parallel to the banding. Throughout all grades of metamorphism potash felspar-quartz-iron ore veins have been observed within the massive meta-arkoses.

(a) Aluminosilicate Segregations.

The aluminosilicate segregations are confined to the andalusite, staurolite, kyanite or sillimanite bearing schists. There is a tendency for these segregations to occur in widely spaced pods rather than as small closely spaced veins. Some of the larger pods may reach several feet in width and 20 feet in length, and are large enough to be represented on the geological map (e.g. area around Jutland on Plate 1.). Their elongate lens-like form is always parallel to the main foliation of the schists in which they are contained.

With their coarse, pink, prismatic andalusite; blue bladed kyanite; white fibrous sillimanite; red-black biotite; white muscovite and plagioclase; occasional large prisms of yellow-green apatite; and rarer small pink garnets and deep blue corundum set in an ample matrix of milky-white quartz, these rocks have a most exotic appearance in the field.

The segregation veins undergo mineralogical changes with metamorphic grade parallel with similar changes in the wall-rock schists. The segregations are sensitive indicators of the first appearance ofthe various aluminosilicates in the metamorphic sequence and are useful in the determination of the aluminosilicate mineral isograds. They are, however, more insensitive than the schists to the disappearance of the un-stable phases. Thus andalusite may be preserved metastably within the segregations when the andalusites of the adjacent schists have been completely replaced by sillimanite. The metastability of the andalusite is aided by its large crystal size in the segregations.

The segregations show a different mineralogy on the eastern and western sides of the area, in character with the changing nature of the metamorphism. For this reason the segregations in the eastern and western schists will be treated separately the arbitary boundary again being the major thrust fault.

A. Aluminosilicate Segregations of the Eastern Part.

(1) Biotite zone.

Before the first appearance of andalusite, occasional barren quartz segregations up to several inches in width occur within the fine-grained, chlorite-knotted biotite schists. Beds of these schists may be traced to higher grades where andalusite develops.

(2) Andalusite zone.

With the first sign of knots representing the growth of andalusite in the schists, pink andalusite crystals appear in the quartz segregations. At 234.219 a large 15' x 4' quartz pod in fine-grained mica schist consists of milky quartz loaded with irregular bunches of pink andalusite crystals up to several inches across (Fig. 52). The andalusites interpenetrate each other, but form euhedral orthorhombic prisms against the quartz. Late or secondary muscovite occurs at the boundary of the andalusite prisms. Rare stumpy blue corundum crystals to 2 mm. are enclosed in the andalusite.

(3) Andalusite-staurolite zone.

Sillimanite first appears in the segregations at 229.200, although sillimanite is not found in the adjacent staurolite-andalusite rich schists there. The sillimanite occurs as fibrous mats and stumpy prisms to 2 mm. in length, penetrating the surrounding

FIG. 52A.

Massive quartz-andalusite pod with minor corundum as small inclusions in andalusite. Scale is 1 foot (30 cm.) rule. Location 235.218.

FIG. 52B.

Quartz-kyanite-sillimanite-andalusite-(garnet) segregation in micaceous schist. Coin diameter 2.8 cm. Location 158.183.

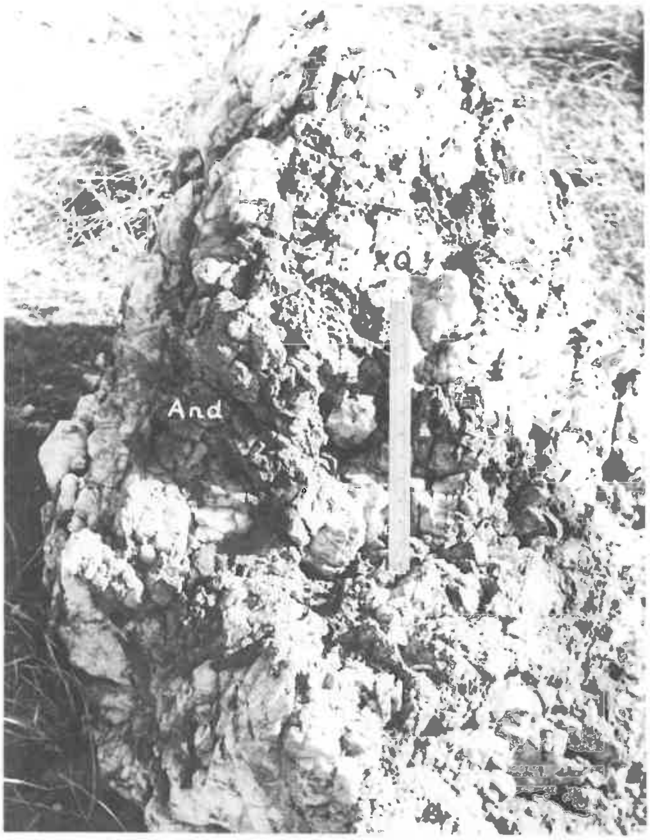


FIG. 52A

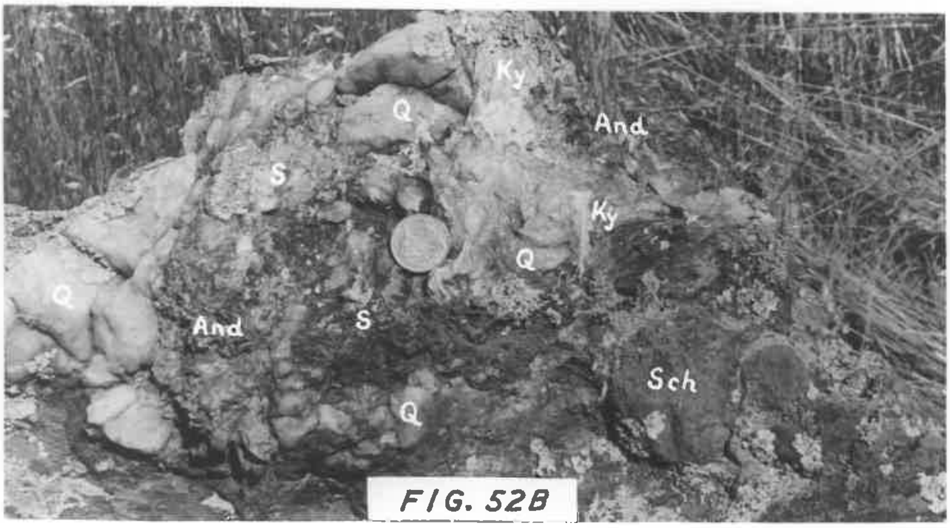


FIG. 52B

andalusite, quartz, muscovite and subhedral plagioclase, and appears to be of late origin. Minor constituents of the vein are ragged blue corundum grains enclosed in the andalusite; biotite; ilmenite with coating of leucoxene; and small prisms of tourmaline. Some sericite recrystallizing to muscovite and chlorite is present as a secondary mineral.

The borders of the segregations in this metamorphic grade are usually enriched in biotite and andalusite. Andalusite may occur within the core of the vein, but staurolite is restricted to the biotite-andalusite selvages, where it occurs as small granules within the andalusite. Staurolite cannot be regarded as a particularly stable mineral within these segregation veins. The biotite-andalusite selvages are normally sharply bounded against the fine-grained strongly foliated quartz-oligoclase-biotite-muscovite-(andalusite-staurolite) schists of the wall-rock. Apart from a slight increase in biotite within 1 cm. of the contact, the schist maintains its grain-size and general character right up to vein, the contact of which is marked by the growth of large andalusite crystals. These crystals, growing out into the schist, enclose biotite which preserves its original tectonic orientation. Portions of schist trapped between the growing andalusite crystal recrystallize, losing their quartz and felspar, presumably to the vein, and the biotite, preserving its original optic character, becomes coarse-grained, the new crystals partly mimicing the pre-existing foliation and partly forming new random grains. The dark andalusite-biotite selvage may be a fraction of a centimetre to several centimetres in width. As the centre of the vein is approached biotite decreases in amount and clots rich in andalusite, muscovite, plagioclase and sillimanite are predominant within the quartz rich core. Muscovite is not uncommonly intergrown with the biotite in the selvages of the segregations and may also occur as sparse cross-cutting flakes dispersed through the fine-grained schist near the vein contact. Staurolite only occurs in the selvages of the vein, included, with biotite and iron ore granules, in the andalusite. Finely fibrous sillimanite, occurring at the sharp contact between the vein and the schist and as rims or coronas on the andalusite crystals, has formed late within the crystallization history of the vein. This occurrence of sillimanite is interpreted as an indication that the culmination of the metamorphism occurred after the main bulk of the vein had formed.

(4) Sillimanite-muscovite zone.

Within the sillimanite-muscovite zone, where andalusite and staurolite have become unstable within the aluminous schists, having been replaced by sillimanite, biotite or cordierite, andalusite may persist as a major mineral within the segregations, but is strongly pierced by well-formed sillimanite prisms and spindles. The sillimanite here is not so often in the form of fibrolite as in coarse "spears" transecting both plagioclase and andalusite. Potash felspar is absent, but muscovite may be present as large primary flakes as well as a secondary alteration of the andalusite and plagioclase. Rutile, corundum and apatite appear in accessory proportions. The sequence of crystallization shown by these segregations is

(1) quartz-andalusite plagioclase-muscovite-(corundum-apatite-biotite),

(2) sillimanite,

(3) secondary muscovite.

(5) Sillimanite-potash felspar zone.

Muscovite and andalusite do not occur as primary constituents within the segregations of this zone. Sillimanite is abundant as grey-brown fibrous masses with individual needles streaming through the common quartz, plagioclase, biotite and potash felspar of the vein. Deep red biotite occurs as large ragged individuals in part strongly replaced by sillimanite fibres. The quartz of these high grade segregations is notably more glassy than in the lower grade ones.

The plagioclase (composition An 0-26) forms large subhedral unzoned crystals with well developed albite and pericline twinning. Potash felspar, as large interstitial grains strongly penetrated by sillimanite spindles, takes the place of muscovite. This felspar is finely perthitic; the thin oriented films of albite are best seen when the grains are near extinction under crossed polars.

B. Aluminosilicate Segregations of the Western Part.

(1) Staurolite-andalusite-kyanite-(sillimanite) zone.

Blue bladed kyanite is a characteristic constituent of many aluminosilicate segregations of this metamorphic zone. The rarity of kyanite in the adjacent schists makes the appearance of this mineral in the segregations a useful indicator of kyanite zone metamorphism.

Kyanite may be the predominant aluminosilicate in some segregations, but is more usually accompanied by andalusite and sillimanite, and, in those segregations approaching the outer limit of the kyanite zone, the kyanite appears as small irregular isolated grains.

A typical kyanite bearing segregation may be seen east of the road at 158.185 (Fig. 52). The segregation, in the form of a lens parallel to the schistosity, several feet long and up to one foot in thickness, occurs in medium-grained quartz-felspar-biotite schists. Near the edge of the vein, and projecting into the quartz-rich core, are pockets of pink andalusite crystals to 3 cm. and blue well crystallized blades of kyanite to 5 cm., surrounded by white fibrous sillimanite studded with small pink garnets. Biotite and muscovite are also important constituents.

In the segregation at 157.185 (specimen 676) the aluminosilicate rich portion consists of abundant randomly oriented well-formed blades of clear kyanite to 1 cm., enclosed by interstitial red-brown biotite, muscovite and plagioclase. $\{100\}$ twinning is common in the kyanite. Andalusite is absent, but fibrolitic sillimanite occurs as irregular vein-like mats between groups of kyanite crystals and as numerous small needles in muscovite and, more rarely, biotite. The kyanites are rarely seen to be penetrated by sillimanite, but at the ends of the blades sillimanite spindles are sometimes observed in alignment with the kyanite c-axes. Here, too, sillimanite is a latecomer to the primary kyanite-muscovite-biotite-plagioclase assemblage.

Rock 673 (location 158.172) is an example of a segregation in which andalusite, kyanite and sillimanite are all well displayed in a matrix of plagioclase and quartz. In this rock, the kyanite appears as blue prisms which have become enclosed in plagioclase and

and andalusite. In both instances the kyanites have been considerably corroded along their c-axes suggesting that they were partially replaced by the plagioclase and the andalusite. The outer portions of both the andalusite and the kyanite crystals have been replaced by sillimanite growing as prisms aligned parallel to the c-axes of both the earlier aluminosilicates. Later trains of fibrolite developed interstitial to the earlier formed andalusite and kyanite. The sequence of crystallization of the aluminosilicate minerals in this segregation is thus: kyanite, joined later by andalusite, followed by prismatic sillimanite and then later by fibrous sillimanite.

Other segregations are rich in andalusite with kyanite subordinate or absent. Specimen 259 (location 168.148) of an Al-rich portion of a segregation vein consists of large, clear pink andalusites enclosed in unstrained quartz. The andalusites are surrounded by thin coronas of late-formed, fibrous sillimanite. Kyanite occurs as rare inclusions in the andalusite. Minor biotite, altering to secondary chlorite, muscovite and apatite are also present. Similar segregations are richer in plagioclase (albite-oligoclase).

A segregation pod at 178.135 is, in portions, rich in large apatite prisms (to 6" x 1"). The aluminosilicate rich portions of the vein contain much andalusite, minor kyanite and interstitial late sillimanite. Euhedral rutile and tourmaline are minor accessories. The andalusite of this rock is of particular interest because of its bladed crystal form. In appearance these blades resemble kyanite (Fig. 53). Some of them are "twinned", the twin individuals being occupied by andalusite in different optic orientations. The "composition planes" are irregular and blurred. Each crystal of andalusite is made up of many small individuals with slight differences in their optic orientation. Thus the "supercrystal" has a mottled appearance near extinction under crossed polars. These observations indicate that the andalusite has replaced original kyanite, small relics of which still remain in the rock.

At 152.212 several staurolite-quartz segregations occur in staurolite rich schists. This may indicate a greater stability of staurolite in the segregation veins under kyanite zone conditions, as opposed to the andalusite-sillimanite conditions of the eastern portion of the area.

(2) Sillimanite zone.

In the western part of the area the grade is not reached where the segregations contain sillimanite

FIG. 53.

Photomicrograph of andalusite pseudomorphing former twinned kyanite in aluminosilicate segregation. The matrix consists of fibroidal secondary muscovite.

Rock A185-280a. Location 178.135. Crossed polars.

FIG. 54A.

Photomicrograph of fresh kyanite crystals with a slightly corroded appearance embedded in a large andalusite crystal in an aluminosilicate segregation vein.

Grey patch to left - fibrolite replacing andalusite along c-axis of andalusite crystal.

Grey patch bottom right - crowd of oriented sillimanite prisms replacing andalusite.

Rock A185-673. Location 158.172. Crossed polars.

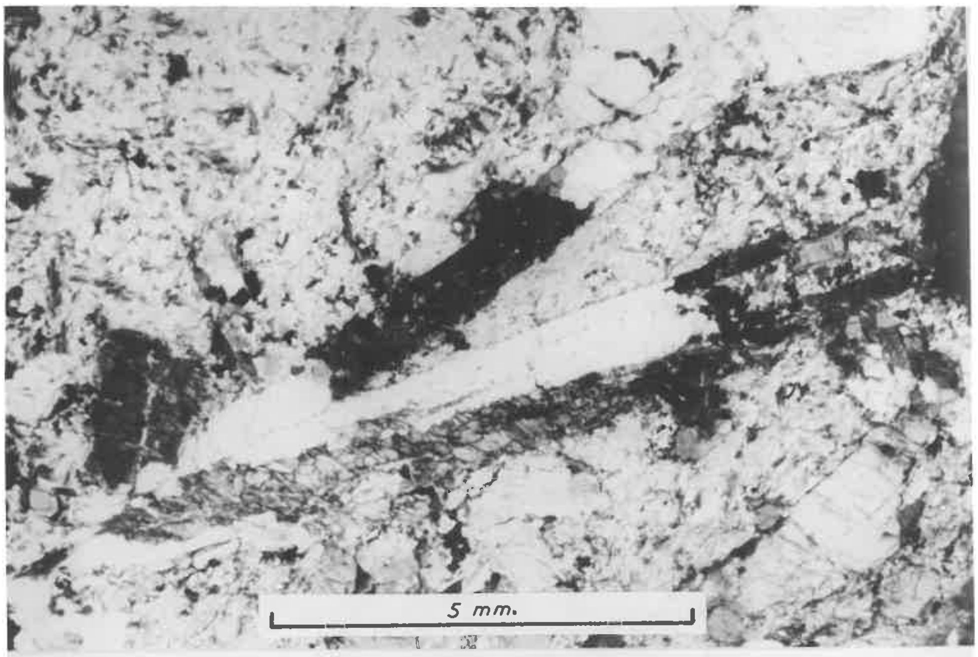


FIG. 53

FIG. 54A

and potash felspar in stable equilibrium, and so, muscovite along with red-brown biotite are still stable constituents. Fibrous and prismatic sillimanite is the major aluminosilicate, with some andalusite surviving by virtue of its enclosure in quartz. In some of these segregations (e.g. 307, location 186.100) a very pale chlorite (corundophilite), penetrated by sillimanite prisms, appears to be a stable constituent.

Discussion and Conclusions.

The mineralogy of the aluminosilicate veins closely reflects that of the enclosing schists, and is sympathetic to changes of metamorphic grade. Thus the veins may be used in the same way as the schists to determine mineral isograds reflecting changing metamorphic grade. This observation, and the apparent lack of connection of these veins with an igneous source, suggests that they are the result of metamorphic excretion or segregation.

Veins of this type, where the mineral components faithfully reflect the mineralogy of the host rocks, are apparently quite common in aluminosilicate bearing schists in various parts of the world. Some workers have concluded that they owe their origin to metamorphic differentiation, while others admit the possibility of introduced igneous or extraneous fluids aiding the formation of the initial quartz or pegmatite vein. Read (1933) has described some excellent kyanite-quartz segregation veins in the kyanite-chloritoid schists of Unst, Shetland Islands. Bosworth (1910) recorded pre-granitic kyanite-tourmaline veins in the outer parts of the aureole of the Ross of Mull Granite. These veins are somewhat similar to the kyanite-tourmaline veins of the Bhandara District, India (Chatterjee, 1931). Kyanite-quartz veins also occur in Korea (Miyashiro, 1951), and in the Appalachians (Stuckey, 1932; Espenshade and Potter, 1960). Alderman (1948) has described kyanite pegmatites cutting sillimanite-quartz masses in the Williamstown District, South Australia, 14 miles west of the present area. McKinstry (1949, p. 888) has discussed the varying mineralogy of veins at differing metamorphic grades in South-east Pennsylvania, culminating at the highest grades with the appearance of kyanite, tourmaline and garnet. Woodland (1963) has described a kyanite-andalusite-sillimanite-oligoclase-quartz pegmatite, with biotite rich selvages against a staurolite-fibrolite schist, from North-east Vermont, which is strikingly similar to several veins in the present area. Andalusite-quartz veins have been recorded in areas of low pressure andalusite type metamorphism, and within the aureoles of some intrusive granitic masses in California (Rose, 1957), in the Blue Ridge and Piedmont Provinces of the Appalachians (Espenshade and Potter, 1960) and in the aureole of the Donegal Granite (Pitcher and Read, 1960). Sillimanite veins have been described by Chinner, 1961, and Alderman, 1948. Staurolite is less commonly encountered in veins, but some excellent examples from western New Hampshire have been discussed by Chapman (1950). An example has also been recorded from Karelia (Eskola, 1932, p.58).

Kyanite-quartz, sillimanite-quartz, andalusite-quartz and staurolite-quartz veins are all represented in various parts of the present area. The major mineralogical changes observed in the veins in the eastern part of the area are,

(a) appearance of andalusite in the quartz segregations at the first signs of andalusite knots in the schists,

(b) the appearance of sillimanite a little before it is found in the adjacent schists in the andalusite-staurolite zone,

(c) an increase of sillimanite and a decrease of andalusite with grade in the lower part of the sillimanite zone,

(d) the instability of muscovite and the appearance of the stable pair potash felspar-sillimanite in the higher part of the sillimanite zone.

In the western part of the area the changes are less marked, but kyanite is an important constituent of the veins in the middle grades of metamorphism. The major changes are,

(a) kyanite and andalusite in the segregations is replaced by late sillimanite in the lower parts of the sillimanite zone,

(b) a progressive increase in sillimanite by the replacement of kyanite and andalusite until, in the highest grade segregations, sillimanite is the only Al_2SiO_5 polymorph remaining, being in equilibrium with muscovite but not potash felspar. The replacement of original kyanite by andalusite has been described from a segregation vein in the central part of the area.

These observed changes may be hypothetically represented in a stability diagram of the Al_2SiO_5 polymorphs (Fig. 54 after Clark, Robertson and Birch, 1957; with modifications by Clark, 1961, Khitarov et al. 1963, and Bell, 1963). On this diagram, Curve I illustrates an isobaric migration path from the stability fields of hydrous phases through andalusite to sillimanite with increased temperature, and represents the observed changes in the eastern portion of the area. Curve II shows an isobaric path, at a higher pressure, from the stability field of kyanite to that of andalusite, representing the segregation from the central part of the area containing andalusite pseudomorphs after kyanite. Curve III illustrates an isobaric migration at a still higher pressure near the triple point. This curve represents the replacement of kyanite by sillimanite in those segregations in the western part of the area.

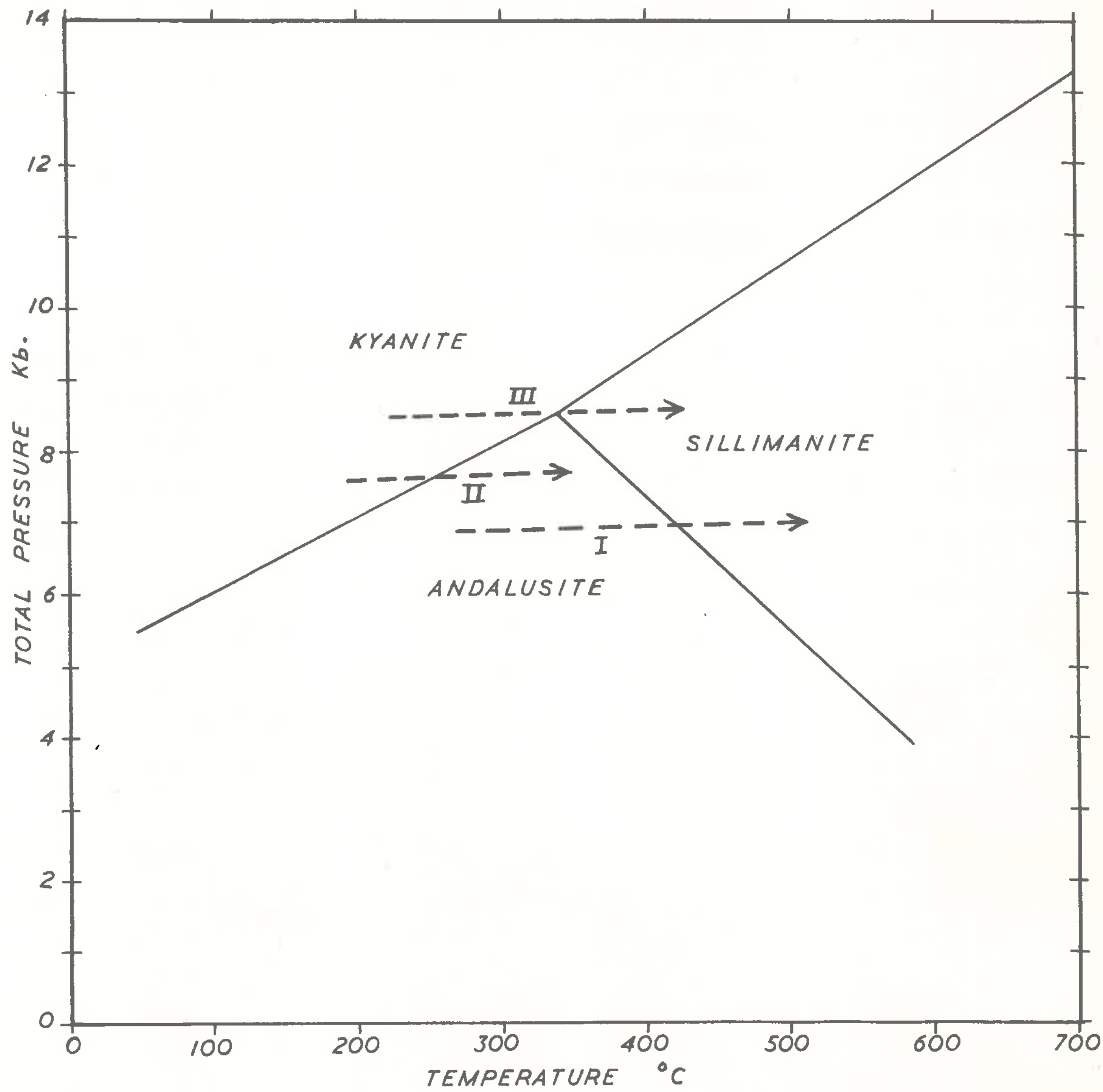
Thus the observed changes in the mineralogy of the segregation veins from east to west can be attributed to an increase in the total pressure. This conclusion confirms the evidence gained from a study of the mineralogical changes in the metamorphic rocks, that the pressure during metamorphism increased westwards.

Under these changing P-T conditions, the stability of the various aluminosilicate minerals can be studied in the natural

FIG. 54.

Stability relationships of Al_2SiO_5 polymorphs (see text).

FIG. 54



14
12
10
8
6
4
2
0
TOTAL PRESSURE Kb.
KYANITE
III
II
I
SILLIMANITE
ANDALUSITE
0
100
200
300
400
500
600
700
TEMPERATURE °C

geological environment. It has already been pointed out that the first appearance of both andalusite and sillimanite in the segregations is apparently quite sensitive to rising metamorphic grade (temperature), as, with the first signs of andalusite or sillimanite in the schists, these minerals develop strongly in the segregations. Thus as soon as the metamorphic conditions attending an andalusite or kyanite segregation enter the stability field of sillimanite, small fibres of sillimanite appear at the boundaries of the andalusite and kyanite crystals. Heinrich (1952) has described similar "late" sillimanite in kyanite and andalusite segregations and has supposed that the sillimanite formed "at lower temperatures". However, in the Cambrai segregations there appears to be a "wide band of indifference" in which andalusite and kyanite are metastable well within the sillimanite zone. This metastability is no doubt aided by the large size of the andalusite and kyanite crystals, especially where they are enclosed in quartz. Andalusite and kyanite also coexist, sometimes with sillimanite and staurolite, in many of the segregations in the western part of the area, and even in the case where the andalusite had extensively replaced the kyanite, some kyanite still remained as relics. Assuming that the P-T conditions changed uniformly with increasing grade, this would indicate considerable sluggishness for the kyanite-sillimanite, andalusite-sillimanite, and andalusite-kyanite inversions. It is, of course, possible that the P-T conditions fluctuated during the growth of the veins. These incomplete polymorphic transformations in segregation veins have been commented upon by Espenshade and Potter (1960), who considered that they were the result of fluctuations in physical conditions during the metamorphism. However, since the metamorphism here is of a regional type, it is unlikely that the fluctuations would be sharp.

Experimental work on the Al_2SiO_5 polymorphs (Clark, Robertson and Birch, 1957; and Clark, 1961) has indicated that the transformation from one polymorph to another is particularly sluggish. The reluctance for transformation to occur may be related in part to low free energy differences between the polymorphs (Miyashiro, 1960; Chinner, 1961). Chinner (1962) concluded that the stability fields of the Al_2SiO_5 polymorphs overlap, resulting in a band of indifference in which two or three polymorphs may be in metastable equilibrium. This is apparently as true for natural geological conditions as for those of the laboratory.

The mechanisms of polymorphic transformation are also important in considerations of the sluggishness of the inversions. In the Cambrai segregations, the observation in one case of large andalusite pseudomorphs after kyanite, suggests that the mechanisms of the kyanite-andalusite transformation are relatively simple, rearrangement of the atomic structure being of prime importance and physical transport at a minimum. The mechanisms of transformation of andalusite and kyanite to sillimanite, however, appear to involve not only atomic rearrangement, but also nucleation and short range migration of material, for in most cases sillimanite formsmany separate prisms, bordering, and sometimes growing well away from, the andalusite and kyanite crystals. Intergrowths of andalusite and sillimanite with parallel c-axes, as observed in the schists, are not so commonly observed in the segregations.

The greater instability of staurolite relative to the anhydrous Al_2SiO_5 minerals in segregations is worthy of comment. Chapman (1950) has noted that staurolite is restricted to the border zones of the quartz veins in New Hampshire and he contrasts this with the more dispersed nature of the kyanite in the veins described from Unst (Read, 1933). At Cambrai staurolite is quite a stable component of the schists in the eastern side of the area but is unstable within both the quartz veins and their border zones. In the higher pressure kyanite metamorphism to the west some staurolite-quartz segregations have been noted, but staurolite again does not occur within the kyanite-andalusite bearing veins.

Besides the aluminosilicate minerals, other major minerals of the veins also undergo some changes with increasing metamorphic grade. Quartz loses its translucent milky appearance and becomes more glassy with increasing grade. Plagioclase compositions within the veins range from An 0-26, and there is no progressive increase in the An content with grade, although the higher An contents are observed in the higher grades. Biotite is stable in the segregations throughout all the grades of metamorphism observed here, and, like the biotite of the adjacent schists, is usually a red-brown colour. This red-brown colour intensifies within the sillimanite zone. Muscovite is stable as a primary constituent of the veins, except in the highest grades, where it is replaced by slightly perthitic potash felspar.

The minor minerals of the veins are rutile, tourmaline, corundum, apatite, sphene, iron ore and garnet. Rutile usually appears to result from the alteration of biotite to chlorite. Rutile is a common mineral in association with the quartz-sillimanite veins of Williamstown (Alderman, 1948). The small amount of tourmaline in the segregations is in accord with the sparse, but widespread accessory tourmaline of the schists. Tourmaline has been commonly observed in segregation veins elsewhere, and is particularly common in those kyanite veins described by Bosworth (1910). Corundum, although sparse, is of particular interest because of its occurrence in many of these quartz rich segregation veins. Surveying the literature, it would appear that corundum is a common accessory constituent of most veins of this type (e.g. Rose, 1957; Alderman, 1948; Espenshade and Potter, 1960; Kerr, 1932; Stuckey, 1932 etc.). The corundum is usually described as small euhedra included within the andalusite and kyanite, but, of course, not in contact with quartz. Rose (1957) considered that the corundum was secondary, being formed by a reaction such as;

$$\left[6Al_2SiO_5 + \underset{\text{(introduced)}}{2H_2O + K_2O} \rightleftharpoons 2KAl_3Si_3O_{10}(OH)_2 + 3Al_2O_3 \right]$$

In all cases he observed euhedral corundum separated from the andalusite by an alteration rim of muscovite. Kerr (1932), however, described corroded euhedral corundum embedded in andalusite. He concluded that the corundum was one of the first constituents to crystallize and was later partially corroded and replaced by andalusite. Stuckey (1932) described corundum crystals enclosed in kyanite. In the Cambrai segregations the corundum is enclosed in andalusite, and although secondary muscovite sometimes obscures the andalusite-corundum contacts, this is not invariably the case, and the corundum is considered a primary constituent. The mechanism of formation of these corundum crystals within such siliceous veins is still not understood, but presumably during the growth of the vein, some portions became so depleted in silica, that corundum was able to form. There is also the possibility of exsolution of the corundum from the andalusite or kyanite.
Apatite is a rather enigmatic constituent of the segregation veins; large crystals are frequently encountered, although in the schists the apatite is only an accessory constituent. It would appear that, in many cases, the apatite is more abundant in the veins than in the adjacent schists, although this is difficult to verify. An observation at 150.138 suggests that apatite is particularly mobile during metamorphic

FIG. 55.

Photomicrograph of kyanite-sillimanite segregation.

Rock A185-676. Location 157.185. Crossed polars.

Coarse bladed kyanite with lesser red-brown biotite and oligoclase (left), bunches of fibrous sillimanite (centre) and matted aggregate of muscovite penetrated by sillimanite fibres (right).

FIG. 56.

Photomicrograph of sillimanite segregation (sillimanite-potash felspar zone).

Rock A185-22a. Location 209.088. Ordinary light.

Fibrous sillimanite with intermingled red-brown biotite in a matrix of coarse quartz, oligoclase and potash felspar.

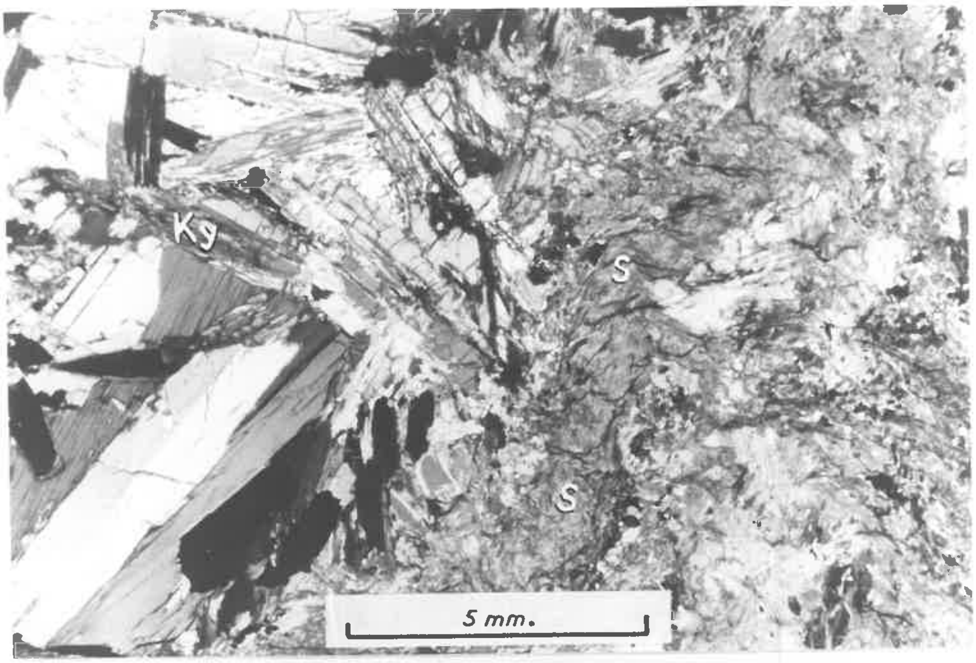


FIG. 55

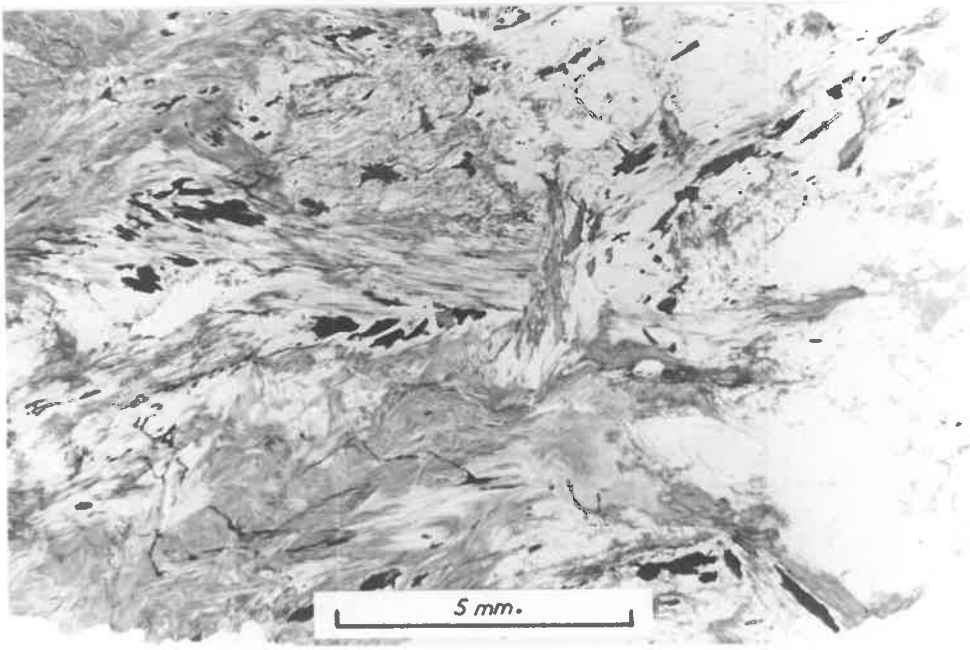


FIG. 56

differentiation and may develop high concentrations in veins. Here a small quartz vein filling a cross-cutting tension gash intersects a sedimentary heavy mineral band in arkose. Against the heavy mineral band the gash is filled with apatite. Thus apatite, like quartz, is apparently a particularly mobile mineral during the growth of segregation veins, and so, may develop concentrations within the veins in excess of that in the adjacent schists. Topaz and lazulite, observed in similar veins in Carolina and California, have not been recognised here.

Origin of the veins, their age and their mechanism of formation.

Although the mineralogy of the veins and their basic selvages suggests that these veins are of metamorphic segregation origin, one cannot be sure that this is entirely true, and must allow for the possibility of introduced extraneous fluids derived from some more deep-seated source. Woodland (1963) considers that the derivation of the aluminosilicate minerals from the adjacent aluminous schists is catalysed by the introduction of extraneous fluids. Even if the modal composition of a vein could be determined to test this, it is unlikely that it would compare with the mode of the adjacent schist because of the differential rates of migration of material, generally leading to a greater abundance of quartz and apatite than in the adjacent schists, and the likelihood of longitudinal migration. In the absence of evidence for a connection with any igneous activity it is concluded that the veins were formed by metamorphic differentiation.

The presence of unstrained to weakly strained quartz, random mineral orientation, and biotite porphyroblasts cross-cutting the foliation in the segregation borders, indicate that these veins were formed after the deformation associated with the formation of both the main schistosity and the crenulation cleavage. The minerals of the veins are similar to those of the adjacent schists, suggesting that the veins were formed at the height of the metamorphism. Thus the veins provide evidence that the metamorphism reached its peak after the deformation associated with folding and the formation of schistosity and crenulation cleavages. This conclusion has been confirmed by a study of the mineral textures in the schists (Chapter 14). Chapman (1950) reached a similar conclusion for the New Hampshire staurolite segregations, but allows their formation over an extended time range. Unlike the veins described by Chapman, the segregations of this area are both large and widely spaced. The initial formation of a vein is the important problem. Once initiated Eskola's (1932 b) solution principle is likely to be much more effective than his concretion principle in increasing the size of the vein. As the veins are post-folding, perhaps the release of pressures associated with the folding caused longitudinal fractures to open, allowing the initiation of segregation veins. Once a chemical or pressure gradient was established, the vein would enlarge by lateral secretion and the replacement of the adjacent schist.

(b) Garnet-Quartz Segregations.

Occasional garnets have been noted in the aluminosilicate veins described previously, where they occur as small euhedra enclosed in fibrous sillimanite. Within the micaceous garnet bearing schists rare quartz veins with garnet rich selvages have been observed. There are two varieties of these distinguishable in the field. One variety is exemplified by a small 2-3" wide glassy quartz vein parallel to the schistosity of garnet bearing mica-quartz schists at 158.169. The vein has a massive deep-red garnet selvage bounded irregularly against the schist. Apart from a little biotite caught up in the garnet, no other major minerals are present. The other variety, also rare, occurs prominently at 195.192 and 223.141. In this variety the quartz vein core is small compared with the size of the selvage. The selvage consists predominantly of light coloured plagioclase speckled with black biotite and red garnet. These veins are considered to have originated in a similar manner to, and contemporaneous with, the aluminosilicate segregations, the host rocks in this case, however, containing garnet, but no Al_2SiO_5 minerals.

(c) Muscovite-Calcite-Quartz Segregations.

Within the calcite-mica schists, the low grade equivalents of the calc-silicate rocks, many small quartz veinlets sometimes appear parallel to the banding. The veins contain coarse calcite and randomly oriented muscovite flakes at their edges and as clumps projecting into their quartz rich cores. The widespread occurrence of these veins in the middle grades of metamorphism also suggests an origin by metamorphic excretion.

(d) Scapolite Segregations.

In the calc-silicate rocks and marbles of the sillimanite zone occasional pods rich in scapolite are found. It is from these that the best and largest scapolite crystals may be obtained, the scapolite prisms growing into calcite filled vughs and as bunchy masses. Scapolite prisms to 4" in length occur in aggregates in the segregation at 196.084 (spec. 137) and some better formed but smaller prisms were found in a prominently outcropping pod at 172.127 (spec. 271). Both of these segregations are enclosed in marble and have coarse diopside and dark green amphibole associated with the scapolite.

Similar segregations also occur in calc-silicate terrain; for example, a scapolite-potash felspar-quartz pod in scapolite schists at 236.186 (spec. 509). This rock consists essentially of a mass of interlocking white microcline and quartz penetrated in diverse directions by large (1cm.) ragged prisms of scapolite with a peculiar fibrous appearance similar to that usually adopted by prismatic sillimanite. Chlorite, after former biotite, and late muscovite are also important constituents.

(e) Potash Felspar - Quartz Veins.

These veins are largely limited to the meta-arkose unit where they appear as minor infillings in tension gashes and joints. They have a widespread occurrence in all grades of metamorphism from biotite grade upwards and are characterized by the presence of subhedral to euhedral creamy-pink potash felspar crystals to 2-3" in diameter, embedded in milky to glassy unstrained quartz. Platy iron ore may appear as an accessory constituent. The arkose contacts are usually enriched in albite. Clots of arkose, sometimes seen embedded in the quartz vein as "xenoliths," might suggest an intrusive origin, but the minor and widespread occurrence of the veins at all grades of metamorphism supports an origin by metamorphic segregation. In the higher grade micaceous schists some quartz and potash felspar rich veins with biotite and garnet rich selvages also appear to have been formed by sweating out during the metamorphism.

Conclusions.

In the Cambrai area veins attributable to an origin by metamorphic segregation may be found at all grades of metamorphism and in all rock types, whether they be quartzo-felspathic, micaceous or lime rich. In all cases, neglecting the late stage alteration products, the major minerals closely reflect those in the adjacent country rock. The mineralogy and structure of the veins suggests a formation in the period just before, and at the height of, the metamorphism, and after the main episodes of structural deformation. The segregation veins form an integral part in the metamorphic history of the

Cambrai area, and are useful guides to the physical conditions attending the metamorphism and as indicators of metamorphic grade.

CHAPTER 10.

METASOMATISM.

Introduction.

In this chapter non-intrusive rocks which cannot be satisfactorily attributed to simple isochemical metamorphic recrystallization of sediments will be discussed. Without a detailed chemical study it is not possible to explicitly establish the nature of the chemical changes involved in the metasomatism, but some general trends can be obtained from petrological examination. The petrological data which follows is essentially a preliminary step towards a knowledge of the metasomatic processes; a more complete understanding must await a detailed chemical study. Several kinds of metasomatic rocks are recognized and these will be dealt with in turn. Following the petrographic description and discussion of these more distinctly metasomatic rocks, some broader metasomatic aspects of the Cambrai area will be considered and some tentative conclusions will be drawn concerning the relative roles of the processes of metamorphism and metasomatism in the region.

(a) Quartz-Albite Rocks.

(1) Field occurrence.

Fine-grained light coloured quartz-albite rocks are widely distributed through the Cambrai region, and are especially common in the Saunders Creek-Jutland area. Although field evidence indicates that these rocks were formed by replacement of a variety of metasediments, their contacts with the adjacent rocks are sufficiently sharp to yield a reliable map pattern, and they have been plotted on Plate 1 as "Quartz-Albite rocks". The purpose of this section is to outline the major features of these rocks and to consider their mode of formation. Unfortunately no chemical analyses of the quartz-albite rocks or their adjacent contacts have yet been made, but some concept of the trend of the chemical changes involved in the formation of these rocks can be gained from petrological observations.

In the field the quartz-albite rocks are characterized by a white or light grey, or, if altered, a pale pinkish colour. The massive and arkosic nature of these rocks usually results in blocky jointed outcrops which stand above the general level of erosion of the adjacent metasediments. The quartz-albite rocks are readily distinguished from the quartzo-felspathic schists by their much finer grain size (0.02 to 0.10 mm.). The contacts, although usually sharp enough for reproducible mapping on the scale of Plate 1, are gradational in detail. The contacts between quartz-albite rocks and quartzo-felspathic schists may be gradational over a wide border zone, and the mapped boundaries are less definite. It is within these wider border zones that the stages involved in the formation of the quartz-albite rocks are most clearly seen. Bedding features such as biotite rich bands, heavy mineral lamellae or calcareous layers may be preserved well within these border zones, and it is evident that both tectonic and sedimentary structures within the original metasediments may be preserved intact until the replacement becomes so complete that the relic structures have entirely disappeared. There is no evidence for volume change or distortion resulting from the formation of the quartz-albite rocks, although one cannot be entirely sure that a volume change has not occurred (Poldervaart, 1953). These rocks have been formed by metasomatic soaking involving equal volume ion for ion replacement.

Perusal of Plate 1 will indicate the general size and shape of the quartz-albite bodies. In size they vary from small pods, only a few feet across, to large masses, such as that forming the high prominence, one mile in length and 500 feet wide, west of Kanappa Mine. The shape of the quartz-albite bodies generally reflects some structural control. The bodies are usually elongated parallel to the main schistosity, a result consequent upon the replacement of rocks having a pronounced penetrative foliation (Ramberg, 1952).

Although quartz-albite rocks are more common in the higher grades of metamorphism, it must be noted (Plate 1) that these rocks are by no means confined to any particular metamorphic grade, and indeed, they even occur below the andalusite isograd in biotite zone rocks at Pine Hut. It is significant that the quartz-albite rocks are entirely absent from the "zone of veins" migmatic area in the south-west. Nor are quartz-albite rocks confined to particular rock types, for, besides replacing the quartzo-felspathic schists, they also replace marble, calc-silicates and rarely metadolerites. The distribution of quartz-albite rocks is strongly influenced by structural control and the majority of these rocks are found within anticlinal folds. The Saunders Creek-Jutland group of quartz-albite rocks have formed within and near the core of a tightly appressed major anticline. The quartz-albite rocks west of the Kanappa Mine are confined to the roof of a closed domal anticline where they have replaced quartzo-felspathic schists below the marble bed. In this case original bedding has also determined the disposition of the quartz-albite rocks.

(2) Mineralogy and Petrology.

Quartz and albite are the major constituents, but other minerals may also be present in minor amounts. Potash felspar and muscovite are stable components of some quartz-albite rocks. Biotite and chlorite, while often present, usually appear to be unstable remnants surviving from the original schist. Opaque iron ore, sphene and rutile occur as variable accessories which have originated, for the most part, during the formation of the quartz-albite rocks. Apatite, zircon and tourmaline are minor accessories believed to be relics from the original sediments.

Quartz may comprise between 2 and 50% of the quartz-albite rocks. It occurs as small 0.05-0.1 mm. clear granoblasts or irregular anhedra evenly distributed through the albite, and in the completely replaced rock, there is little or no tendency for the quartz to be more common in certain bands. In some quartz-albite rocks which have suffered strain, deformation bands and Boehm lamellae appear in the quartz.

Plagioclase may form 35-80% of the quartz-albite rocks, the grains, like quartz, generally appearing as small granoblasts. In some rocks the felspar may be slightly elongate in $\{010\}$. This felspar is sometimes clouded by a slight alteration. Twinning is simple and rare, but in those rocks with strained quartz, chess-board twinning is common. The relief is always distinctly less than quartz and measured compositions (Emmons 5-axis technique) lie close to An 2. The majority of twins follow the albite law, a few the

Carlsbad.

Potash felspar is a stable component of some quartz-albite rocks within the sillimanite zone, where it may comprise 10% of the rock. It invariably has well developed cross-hatched twinning and is either granoblastic or intimately intergrown with albite such that $\{010\}$ is common to both minerals. The albite-potash felspar intergrowths are especially common near unstable biotite flakes, and it is suspected that the potash felspar has been derived in part from the alteration of original biotite (Fig. 57).

Muscovite, always in the form of thick irregular skeletal flakes, is often present in the quartz-albite rocks as a stable constituent. It is more common in those quartz-albite rocks replacing the lower grade metasediments, where it may comprise up to 15% of the rock. The muscovite and the potash felspar may form a reciprocal relationship, the mica occurring in quartz-albite rocks replacing the low grade metamorphics and the potash felspar in those rocks in the higher grade zones. This is the only apparent relationship between the quartz-albite rocks and grade of metamorphism.

Biotite, pleochroic from pale lemon to dark brown, is present in amounts up to 5% in some quartz-albite rocks. However, it rarely appears to be a stable

FIG. 57A.

Photomicrograph of quartz-albite rock.

Rock A185-148b. Location 183.083. Crossed polars.

Strongly ferruginized and corroded biotite flakes in a matrix of quartz, albite and potash felspar (cross-hatched twinning).

FIG. 57B.

Photomicrograph of quartz-albite rock.

Rock A185-454. Location 211.103. Crossed polars.

Granular quartz and rarely twinned albite. Occasional larger potash felspar porphyroblasts (bottom left).

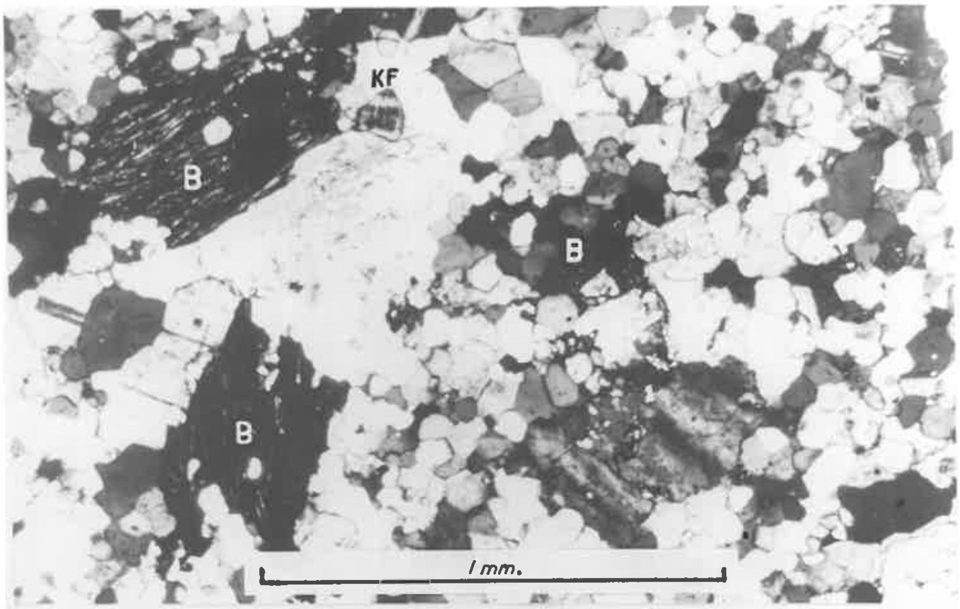


FIG. 57A

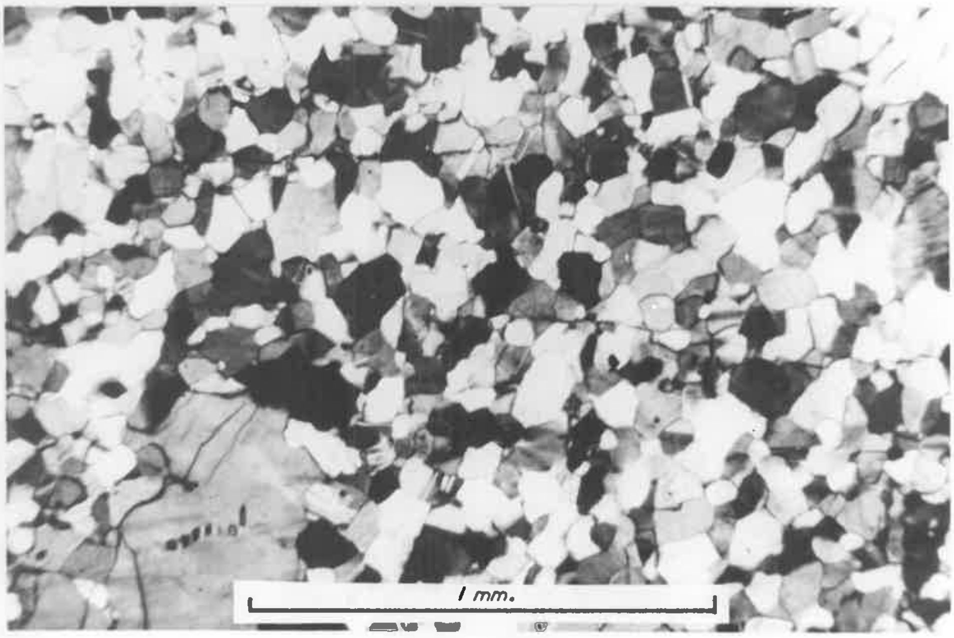


FIG. 57B

constituent and generally has a corroded appearance. Chlorite may appear as an alteration product of the biotite, but this, too, appears metastable.

Sphene, as small rounded grains is the most consistently developed accessory. It is sometimes accompanied by rutile. Minute grains of ilmenite also appear in some rocks. The sporadic occurrence of accessory apatite, zircon and tourmaline suggests that these are relics from the original metasediment.

The rock fabrics range from equilibrium granoblastic textures to textures in which unevenly sized amoeboid quartz and albite grains form an interlocking jigsaw. Rarely, as in specimen 265 (location 168.135), a quartz-albite rock replacing marble, the plagioclase takes on a slightly lath-like appearance, although there is no suggestion of a flow structure.

The quartz-albite rocks are usually massive, but, in one potash-felspar bearing example, the potash felspar outlines distinct layers suggestive of relic bedding. In another quartz-albite rock (specimen 454, location 211.103) secondary chlorite pseudomorphs prismatic forms resembling amphibole. This may have originally been an anthophyllite (?) bearing quartz-albite rock.

It is concluded that the essential constituents of the quartz-albite rocks are quartz and albite, with lesser muscovite and potash felspar, and accessory sphene, ilmenite and rutile. Thus, whatever the nature of the metasomatism producing these rocks, the stable chemical components in the final product are SiO_2, Al_2O_3, Na_2O, K_2O and TiO_2. A more specific knowledge of the chemical changes involved in the formation of these rocks can be gained from a consideration of their contacts with the adjacent metasediments.

(3) Partially altered rocks and gradational border zones.

Most of the quartz-albite rocks have replaced quartz-felspar biotite schists. At 230.187 a small sharply bounded pod of quartz-albite rock has replaced quartz-felspar-biotite schist, the bedding and foliation traces clearly passing through the altered rock. Two specimens (825a and b), one of quartz-albite rock and the other of unaltered schist, were collected three feet apart along the strike of the bedding. The schist consists of sodic plagioclase (25%) and quartz (45%) in an even-grained granoblastic texture penetrated by well oriented biotite (25%). The rock is cut sporadically by flakes of late muscovite (3%). Iron ore, tourmaline and zircon are accessories. In the quartz-albite rock the plagioclase has increased to 60%, quartz decreased to 25% and muscovite increased to 15%. Bleached and partially destroyed biotite altering to chlorite appears as minor unstable relics. Rutile and sphene have joined the accessory minerals. The grain-size is irregular, the quartz forming grains of similar size to those of the original schist, while the albite appears as an abundant fine-grained interstitial matrix. Thus, in the formation of the quartz-albite rocks from the schist, the biotite has been destroyed and some quartz has been used up concomitant with a marked increase in the proportions of albite and muscovite, and the appearance of sphene and rutile among the lesser constituents. During the albitization the even-grained crystalloblastic texture of the schist has become finer grained and uneven. Twinned plagioclase is absent in the schists but is common within the quartz-albite rock.

Similar changes have been observed across the quartz-albite rock-schist contact at 210.118. Three specimens (411, a,b,c) were collected across the contact, which, at this point, is parallel to the relic sedimentary banding of the schists. However the schists have a uniform appearance and it may be assumed that the three specimens had similar initial compositions. Again, on passing from the quartzo-felspathic schists into the quartz-albite rock, the quartz content decreased (70-30%), albite increased

(20-70%), while biotite decreased (10-2%). Neither muscovite or potash felspar are present in these rocks. Iron ore granules are absent from the schists but are common in the quartz-albite rock. In passing from the schist into the quartz-albite rock the grain-size diminishes. Again the general trend is clear; an increase in the amount of albite, partially at the expense of quartz, accompanied by the destruction and disappearance of biotite with the formation of iron ore and sphene (cf. Leedal, 1952).

The replacement of marble by quartz-albite rock apparently involves the removal of almost all the original marble components. The composition of the quartz-albite rocks replacing the marble should then give a good indication of what has been introduced into these rocks. Specimen 265 (location 168.135) is typical of a small quartz-albite rock enclosed in marble. The rock consists of albite (60%), quartz (35%) and muscovite (3%), with traces of chlorite iron ore and sphene. Here SiO_2, Al_2O_3 and Na_2O were introduced while the original MgO, CaO and FeO of the marble have been removed almost entirely.

The albitization of a metadolerite body has been observed at 232.188. Three specimens (516 a,b,c) were collected; one of the normal metadolerite, one of the partially altered, and one of the most altered metadolerite. The unaltered metadolerite consists of hornblende (55%), plagioclase (An43-57 - 30%) and scapolite (10-15%) with accessory biotite, epidote, apatite, sphene and iron ore. The plagioclase is well zoned (rims to An20), a feature common within these metadolerites (see Chapter 11). In the intermediate rock hornblende has decreased to 15% and scapolite to an accessory role while plagioclase (An 15 - mainly twinned on {010}) has increased to 80%. The coarse-grained blastophitic texture of the metadolerite has been replaced by a fine-grained granular texture. The amphiboles are more isolated, recrystallized and skeletal. The plagioclase still preserves its "igneous" twinning but is changed in composition, and the crystal edges have recrystallized to small granules. The most altered rock contains 10% diopside as clear skeletal grains, 5-10% pale green actinolite and 80% plagioclase (An12). Scapolite has entirely disappeared. No signs of the former blastophitic texture remain, the plagioclase being fine-grained and granular and showing only rare simple albite twinning. Here no quartz has been produced, although the marked increase in plagioclase content indicates that SiO_2 must have been introduced along with Na_2O. Edwards and Baker (1953) have described the disappearance of scapolite and the replacement of hornblende by actinolite during the albitization of scapolite-hornblende-plagioclase schists in the Trekelano area, Cloncurry District, Queensland. Although they consider that scapolitization and albitization may be closely related processes in the Cloncurry District, there is certainly no connection between the occurrence of scapolite and the albitization in the present area, and indeed, scapolite never appears within the quartz-albite rocks, even where they replace scapolite bearing metasediments.

The Gumeracha-Lyndoch albite rocks (Dickinson, Sprigg, et. al. 1951) differ from those of the present area in having much talc and lesser pyrite as constituents. Stillwell and Edwards, 1951, consider that the talc at Gumeracha has formed from the breakdown of biotite in the original schists during the albitization. Although some schists replaced by the quartz-albite rocks in the Cambrai area contain up to 25% biotite no talc has been observed. At Broken Hill, Vernon (1961) found that, during the albitization of original biotite bearing rocks, the biotite became progressively paler in colour and analyses revealed an increase in the Mg/Fe ratio but no talc was observed.

In all the examples considered above, the formation of the quartz-albite rocks involves an increase in the SiO_2, Al_2O_3, and Na_2O components and a decrease in the MgO, CaO, and FeO components. Much of the K_2O, TiO_2 and some of the FeO of the original rock (mostly in the biotite) is possibly preserved in the muscovite, ilmenite, sphene and rutile. Some MgO may be preserved in the muscovite as the phengite molecule, a low 2Vx of 35-37^{o} supporting this, but not all the MgO from the original rock could be preserved in this way. It is therefore concluded from the petrological evidence that the quartz-albite rocks are the result of metasomatic replacement involving the introduction of soda, alumina and silica and the removal of magnesia, lime and ferrous oxide.

(4) Hypotheses of origin of the quartz-albite rocks.

If the quartz-albite rocks have resulted from the introduction of material, a question arises as to the source of this material. Albitization, accompanied by the formation of albite and quartz-albite rocks, is a common feature of many igneous and metamorphic terrains. Some albite rich rocks are closely associated in space and time with igneous rocks which have been regarded as a source of the albitizing solutions. There is, however, much diversity in these occurrences. Albitites are commonly associated with acid igneous rocks. Locally, small albitite masses have been described from the borders of the granite at Cape Willoughby, Kangaroo Island, South Australia (Tilley,

1919) and the contemporaneous Victor Harbour granite, Encounter Bay, South Australia (Browne, 1920). The formation of albitites is usually considered to be a late stage process, the albitizing agents being, perhaps, sodic rest solutions expelled in the late stages of crystallization of the granite (e.g. Tilley, 1919). Phillips (1955), describing quartz-albite rocks near the edge of the Criffell-Dalbeatie granodiorite, concludes that these rocks have arisen through the action of a wave of metasomatism ascending ahead of the rising magma, because of intrusive contacts of the granodiorite against the quartz-albite rocks.

Albitization phenomena are not only restricted to acid igneous rock associations but have been described in connection with diorites (e.g. Gilluly, 1933) and even gabbros and dolerites (e.g. Agrell, 1939), and in these cases there is also severe chemical alteration of the basic rocks. In the formation of albitites, adinoles and spilosites it is usually concluded that soda, along with silica and alumina, have been introduced while lime, magnesia and iron have been removed.

In other areas, soda metasomatism, often not closely related to any obvious igneous activity, has lead to granitization phenomena (e.g. Misch, 1949), sometimes at relatively low temperatures high in the crust (e.g. Coombs, 1950). During granitization, soda may be introduced with the expulsion of potash, or vice versa. Harme (1958, 1959), however, considers that in alkali metasomatism leading to granitization, one alkali does not necessarily expel the other, and that there is a tendency towards a particular ratio of alkalies corresponding to an "ideal" granite composition. In the Inyo Range, California-Nevada, Anderson (1937) has described soda metasomatism leading first to granitization and later to the formation of quartz-albite rocks quite similar to those described here. Anderson considers that in the Inyo Range the Boundary Creek granite has been the source of the sodic solutions leading to the formation of the overlying Pellisier granite and the quartz-albite rocks. In some metamorphic terrains quartz-albite rocks may have a widespread occurrence at all metamorphic grades (as in the Broken Hill District - Vernon, 1961), and the albitizing solutions have presumably migrated considerable distances from some unknown source at depth. In zeolite facies metamorphism in New Zealand, Coombs (1954) has described minor metasomatic quartz-albite rocks which may have arisen as an integral part of the metamorphism, through the circulation of alkali charged connate waters. Thus, although quartz-albite and albite rocks are common throughout the world, there is great diversity in their environment of formation and in the source of the albitizing solutions.

The quartz-albite rocks at Cambrai are not visibly related to any igneous intrusive rocks. They are younger than the metadolerites which they may replace, and are older than the aplites, dykes of which may cross-cut them. The fine uneven grained texture of many of the quartz-albite rocks suggests that they were formed late in the metamorphic history, and, except in the highest grade zones, have not reached metamorphic equilibrium with the unaltered metasediments. The quartz-albite rocks may be roughly contemporaneous with the granodiorites. The relation of the quartz-albite rocks to the intrusive rocks will be discussed further after the igneous rocks have been considered (Chapter 11). The widespread occurrence of the quartz-albite rocks in all grades of metamorphism, the concentration of these rocks in the highest grades of metamorphism, but their absence from the migmatite zone, are among the most important facts to be considered in deciding their origin.

In considering an origin for these rocks, the albite and quartz-albite rocks of the Gumeracha and Lyndoch-Fewsey Vale Peak Districts of the central

Mount Lofty Ranges must be taken into account. These rocks occur in medium and high grade metasediments west of the axis of granitization and the zone of veins marking the centre of the metamorphic zonation in the eastern Mt. Lofty Ranges. Their petrology has been discussed in the Gumeracha area by Stillwell and Edwards (1951) and in the Pewsey Vale Peak area by Chinner (1955). Some aspects of the albitization in the Pewsey Vale Peak area are at present being studied by Mr. R. Offler. This western belt of albitization is almost 30 miles in length, extending in a north-north-east direction from Oakbank to Lyndoch and is characterized throughout its length by unique talc-albite rocks, including the economic talc deposits of Gumeracha and Lyndoch. There are no igneous rocks in the Gumeracha-Lyndoch area to which the source of the albitizing solutions can be clearly ascribed. However, a small body of metasomatic gneissic granite, synkinematic to a second crenulation cleavage fold phase, south of the Warren Reservoir, Williamstown (Mills, 1963) and a similar granitic gneiss at Oakbank lie within this zone of talc-quartz-albite rocks.

There is no evidence to establish contemporaneity of the Gumeracha-Lyndoch rocks with the quartz-albite rocks of the Cambrai area. In fact, the albitizing solutions in each area may have had a different source.

Mr. R. Offler (personal communication) has stated that some albite rocks are post-faulting in the Pewsey Vale Peak area, while the Cambrai quartz-albite rocks are certainly pre-faulting. It seems likely that the Cambrai quartz-albite rocks are older than those of the Gumeracha-Lyndoch area.

The Cambrai quartz-albite rocks seem to have been formed late in the metamorphic history and therefore it is unlikely that they have originated by the action of ascending alkaline solutions driven ahead of the rising migmatite-granitization core of the metamorphic belt, although the absence of the quartz-albite rocks from the migmatite zone might suggest this. Thus it must be concluded that a satisfactory origin for the quartz-albite rocks is not established.

(b) Quartz-Potash Felspar Rocks.

(1) Introduction.

In the wide sequence of steeply dipping potash deficient meta-arkoses west of the Marne River Reserve, bands of quartz-potash felspar rocks have been separately mapped and plotted in Plate 1. These rocks may have no direct connection with the quartz-albite rocks described above; textural evidence indicates that they are syn-metamorphic and therefore older than the quartz-albite rocks. The quartz-potash felspar rocks occur in bands parallel to the bedding and schistosity of the adjacent meta-arkoses, but their excessive abundance of potash-felspar is far greater than would be encountered in any normal sediment (Pettijohn, 1957, pp. 369-370). The highly potassic composition of these rocks, along with the textural evidence, suggests that they are not isochemically metamorphosed sediments, but are the result of metasomatism in which potash rich fluids have been especially active.

(2) Petrography.

Those portions of the quartz-potash felspar band south of the Marne River consist of a very fine-grained pale pink gneissic rock with irregular subparallel laminae outlined by minute iron ore granules. The most striking feature is the grain-size, which is considerably finer than that of the adjacent meta-arkoses. The rock is composed largely of microcline and quartz (in the ratio 60:40) with minor muscovite and dusty iron ore. The microcline occurs as crowds of minute grains (0.05-0.10 mm.) filtered with fine iron ore dust. The quartz, usually free of opaques, occurs as larger xenomorphic, slightly corroded grains arranged in thin gneissic layers. Broad to thin muscovite flakes also show a parallelism to this layering.

Somewhat similar fine-grained but less gneissic rocks occur within the mapped band north of the Marne

River. Specimen 702 (location 146.183) again shows an abundance of fine-grained twinned microcline crowded with opaque dust and granules and interspersed with lesser quartz. Dispersed through the microcline quartz matrix are lensoid clots of coarser quartz grains containing included biotite in a poor state of preservation. These quartz-biotite clots may be micro-skialiths (Goodspeed, 1959) which have escaped replacement by the fine-grained microcline and quartz.

The usual rock composing the mapped band at the Marne River and northwards is a fine-grained granitic gneiss similar in general appearance, although with a finer grained quartzo-felspathic matrix, to parts of the granitic gneiss in Tanunda Creek or the Mt. Crawford granitic gneiss. This granitic gneiss is a pale pink coloured rock with a distinct foliation outlined principally by thin isolated disc-shaped patches (to 1 cm. in diameter) of well oriented biotite flakes with rare intermingled muscovite. The biotite may also outline a faint flat lying lineation. Biotite shows the common pleochroism pale yellow to grey brown. An ample fine-grained microcline-quartz matrix surrounds and isolates the patches of oriented micas, the well twinned microcline being in excess of the quartz. In some of the lesser reconstituted granite gneisses albite may also be present as slightly larger highly corroded grains and quartz may form distinct strings or layers of grains within the quartz-microcline matrix. Rarely secondary cross-cutting muscovite flakes are also present as well as the syntectonic muscovite. Iron ore granules are again a common accessory.

Within the mapped band in the north there are a number of distinctive rocks which contain no potash felspar, but which occur in close association with the microcline rocks. These are quartz or quartz-albite rich rocks which contain, in parts, much iron ore and amphibole. The iron ore and the amphibole, which are often closely associated, have a remarkably patchy development which cannot be related in any way to original bedding in the adjacent metasediments. In any case the iron ore rich rocks do not have any greater monazite or zircon content than adjacent portions of the rock lacking iron ore. This is in contrast to the concomitant increase in the "colourless heavies" in the true sedimentary heavy mineral bands found elsewhere in the arkose. The iron ore and amphibole rich rocks were therefore formed by metamorphism and metasomatism and cannot be related directly to an original sedimentary heavy mineral source.

Specimen 701 (location 147.181) is one variety of the iron rich rock in which abundant thin ilmenite plates, associated with more equidimensional haematite crystals, outline a pronounced schistosity within a crystalloblastic quartz-albite matrix. Muscovite is also present as a major mineral, the flakes being oriented parallel to the foliation outlined by the ilmenite plates. The muscovite is characterized by an extremely thin euhedral form, some length-breadth ratios reaching the value of 60:1. The absence of biotite from this rock may be due to the high oxidation ratio indicated by the ilmenite-haematite assemblage (Chinner, 1959). Specimen 702, from the same location, is very similar, but contains in addition large (1 cm.) disoriented porphyroblasts of pale green slightly pleochroic amphibole (2Vx 76-80).

(3) Origin.

It is concluded that neither field observation or laboratory examination can support a simple metasedimentary origin for these rocks. They are unusual in composition and it appears that some form of metasomatism was involved in their formation. It is suggested that potash introduction was important in forming the microcline rocks but the author can find no explanation for the close association of the potash rich and potash deficient rocks. It is believed that the formation of these rocks is closely connected with the removal of potash from the adjacent meta-arkoses, an important problem to be considered further later in this chapter.

(c) Hydrous Metasomatism of Marbles and Calc-silicates with the Formation of Actinolite, Chlorite and Talc.

(1) Occurrence and petrography.

A distinctive fine-grained chlorite schist containing long tapered pale brown prisms of amphibole occurs at several localities near the edge, or within, marble and calc-silicate beds in the Jutland-Marne River Reserve area. These rocks are often associated with others rich in actinolite or talc and some of the chlorite-actinolite schists are altering to talc by surface alteration. Immediately north of the Marne River Reserve at 158.174 the scapolite-amphibole rich calc-silicate rock on the edge of the marble has been locally replaced over a distance of 10-15 yards by a fine-grained chlorite schist containing rosettes of long, 1-3 cm., tapered spindles of amphibole. The schistosity outlined by the chlorite flakes is parallel to that in the adjacent schist and controls the growth and orientation of the platy rosettes of amphibole.

An identical rock occurs 2 miles south at Jutland (172.143) where the rock is seen to be altering, apparently by surface weathering, to talc. 500 yards east of this occurrence (at 201.119, specimen 325) an identical rock is embedded in marble.

Microscopically these chlorite-actinolite schists consist of 80-90% of fine (0.1 mm.) flakes of pale optically positive chlorite outlining a strong foliation, in which isolated euhedral prisms of colourless amphibole, with optic axial angle close to 90°, are embedded. Small tablets of ilmenite and grains of apatite are widely distributed through the chlorite, and small prismatic rutiles are enclosed within the amphibole. The whole rock may be cut by a late crenulation cleavage into which the chlorite flakes and ilmenite tablets are bent.

Actinolite, actinolite-chlorite and actinolite-biotite rocks at 169.148 (specimen 257) may be petrogenetically related to the chlorite-actinolite schists. The structure here is obscure but these well-recrystallized rocks seem to have entirely replaced the original marble bed.

Massive pale tremolite rocks also occur at 175.129 (specimen 269) where they replace the original calc-silicate rocks bordering the marble. At this locality a body of crenulated foliated talc occurs within the actinolite rocks. In thin slice these tremolite and actinolite rocks also show some interstitial alteration to talc.

(2) Genesis.

The distinctive and unusual character of the chlorite-actinolite rocks suggests that these were formed contemporaneously under similar conditions at the three widely spaced localities. Field relations indicate that these, and the associated actinolite rich rocks, were formed by local replacement of the original calc-silicates and marble. The fine grain-size of the chlorite in the chlorite-actinolite schists, and the general lack of strong tectonic orientation of the actinolite, suggests that these rocks were formed late in the metamorphic and tectonic history, although apparently before a late phase of deformation associated with the formation of a crenulation cleavage. These rocks may be more or less contemporaneous with the nearby fine-grained quartz-albite rocks described above, and there may be a genetic connection between them. If much H_2O was evolved during the formation of the quartz-albite rocks, it is conceivable that this could escape to higher levels and be channelled against the more impervious calc-silicates and marble where local addition and removal of material could result in the hydrous chlorite, actinolite and talc bearing assemblages.

The talc deposits of the Gumeracha-Lyndoch area are more closely associated with quartz-albite rocks and are not directly connected with marbles or calc-silicates. Stillwell and Edwards (1951) have proposed that the Gumeracha talc-albite rocks have formed by the alteration of original biotite schists, a conclusion also supported by Whittle (1957) in the Oakbank area.

On the western edge of the marble bed one half mile east of the Marne River Reserve are a number of large patches of highly kaolinized and altered schists. The cause of the kaolinization is not clear, but in several of these altered zones small quartz veins and stringers carry some copper sulphide mineralization. The marbles and calc-silicates here have not been affected. Here, too, it appears that the marble has acted as an impervious barrier to ascending hydrous solutions. These alteration zones are apparently younger than the chlorite, actinolite and talc rocks described above.

(d) Skarns and other Rocks ascribed to Alteration and Replacement.

(1) Massive garnet veins and pods.

White (1956) recorded several massive garnet pods from Milendella, immediately south of the present area. Only one similar occurrence has been found at Cambrai (location 216.143). Here a massive vein of red-brown garnet, one foot wide and 20-30 feet in length, outcrops at a marble-schist junction. Some small vughs within the vein, outlined by euhedral garnet crystal faces, are coated or entirely filled with bright green epidote. Slight copper mineralization occurs adjacent to the vein in coarse-grained scapolite-diopside and scapolite-quartz rocks.

(2) Garnet and amphibole skarns of the Kanappa Mine.

During metamorphism the copper deposits of the Kanappa Mine were introduced adjacent to tightly folded schist-marble contacts. Apparently associated with the ore formation was the extensive grain growth in the adjacent schists, gneisses and marble, leading to a much coarser grain-size than usual, and also to the appearance of garnet and dark amphibole in the marble in association with enlarged grains of deep green diopside. This is the only locality for garnet bearing marble in the Cambrai area.

Some sulphide is associated with minor amphibolite (metadolerite) pods intruded near the marble-schist contacts, but the majority of the ore appears to have been won from replacive quartz-sulphide veins embedded in quartz-felspar-biotite-garnet schists adjacent to the marble. The quartz-sulphide ore is enclosed in a distinctive quartz-biotite-garnet-cummingtonite-anthophyllite gneiss and all stages in the gradual replacement of this gneiss by the sulphide bearing quartz can be observed. Thus the "vein" contacts are indefinite, and the garnet-amphibole gneiss, being an integral part of the ore vein, may be considered as a skarn. The garnet-amphibole gneiss is medium-grained and has a pink-grey colour. Abundant pink anhedral garnets interspersed with aggregates of small pale prismatic amphiboles and red biotites are set in a matrix of quartz and lesser plagioclase. The garnets, which reach 2 mm. across, carry numerous inclusions of quartz, plagioclase and amphibole. The amphiboles are nearly colourless in thin section. One is a cummingtonite ($2V_z = 79^\circ$, $Z \wedge c = 19^\circ$, which corresponds to 60% cummingtonite molecule (Deer, Howie & Zussman, Vol. 2, p.242)) with a high birefringence and good lamellar $\{100\}$ twinning. The other is a non-twinned anthophyllite ($2V_z = 73^\circ$, $Z = c$). The biotite is pleochroic from pale lemon to a bright orange brown. The plagioclase has a composition near An 25 and is twinned on the Ala A, Acline and pericline laws but rarely on the albite law. Tourmaline, apatite and iron ores are accessory constituents.

A somewhat similar rock occurs near the marble-schist contact 100 yards east at 217.118 (specimen 437). This rock is also stained by minor limonite and malachite and has limonite pseudomorphs after original chalcopyrite. The rock is strongly foliated and contains abundant pale pink garnet, deep-green hornblende and white cummingtonite embedded in quartz and basic labradorite. The garnet is anhedral and poikiloblastic as in the previous rock. The cummingtonite has the properties $2V = 90^o$, $Z \wedge c$ 16^o, which correspond to 27% cummingtonite molecule (Deer, Howie and Zussman, Vol. 2, p. 242). The hornblende has the properties $2Vx = 75^o$, $Z \wedge c$ 16^o. Both this and the previous rock are considered to have a sedimentary origin.

The cummingtonite-hornblende association is common in both para- and ortho-amphibolites in low pressure regional metamorphic terrains (e.g. Eskola, 1950; Watters, 1959; Shido, 1958). The association cummingtonite-anthophyllite (or gedrite) is less common but has been reported, for example, by Eskola and Kervinen (1936) in an amphibolite from Isopää, Finland, and in schists and granofels of the Moine Series (Collins, 1942). The problem of the relationship of these isochemic species has been discussed by Layton and Phillips (1962) and it is evident that more chemical data is needed to establish the degree of overlap of the chemical stability ranges of these amphiboles.

Boyd (1959) has presented an hypothetical T-X diagram for the join $Mg_3Si_4O_{10}(OH)_2$ - $Fe_3^{''}Si_4O_{10}(OH)_2$ which gives an explanation of the association cummingtonite-anthophyllite having a limited chemical range from 30-40% of the iron molecule. The cummingtonite from the cummingtonite-anthophyllite association at the Kanappa Mine, which has a composition estimated from the optical properties of 40% of the iron molecule, is in the composition range where, to our present knowledge, the stable association of cummingtonite and anthophyllite is permissible. There is certainly no evidence for any instability between the two amphiboles in this rock.

In the Kanappa region sulphide mineralization also occurs at the junctions of marble and small intruded amphibolite (metadolerite) pods. Here the sulphides, pyrite, chalcopyrite and pyrrhotite are associated with secondary epidote (clinozoizite) and actinolite. Skarn zoning over a distance of 1-2 cm. has been observed at the amphibolite-marble contacts (specimen 442, location 216.117). The marble consists of aggregates of diopside granules enclosed in a matrix of larger anhedral calcites with minor secondary plagioclase and clinozoizite. Toward the amphibolite a zone of amphibole marble is first encountered in which hornblende and calcite are the major constituents with lesser diopside and secondary actinolite. This zone is followed by an amphibole rich sulphide bearing zone in which hornblende and lesser diopside are present with minor secondary actinolite and sulphide. A clinozoizite rich sulphide zone succeeds the amphibole rich sulphide zone. Here clinozoizite (2V = 90°) and diopside are the major minerals, actinolite and sulphides are lesser and hornblende and sphene are accessories. Finally the amphibolite is reached, where hornblende is the major mineral accompanied by lesser diopside and epidote. In these zones the deep-green hornblende (2Vx = 75°, $Z \wedge c$ $18\frac{1}{2}°$) is often replaced by homoaxial overgrowths of the younger pale green actinolite (2Vx = 80°, $Z \wedge c$ 15°). The presence of euhedral crystals of both epidote and actinolite embedded within and intergrown with the sulphides indicates that these were formed contemporaneously late in the metamorphic history, the amphibolite-marble contacts presumably acting as suitable structural and chemical sites for ore deposition.

(3) Limonite-haematite skarns.

Rocks of this category occur within and adjacent to the marble in the western part of the Cambrai area. They are best displayed at two locations in the Saunders Creek area. At 196.082 a large outcrop of flat dipping foliated marble has been partially replaced by fine-grained haematite and limonite. The contact between the replaced and the unaltered marble is sharp and steeply dipping, thus transecting the foliation which can be traced into the altered zone without structural disturbance. A similar and somewhat better exposed occurrence at 178.098 is illustrated in Fig. 59. Here it is clear that the replacement of the marble is controlled by jointing but the foliation of the marble may again be traced through the altered rock. These rocks have formed by ionic replacement without any structural disturbance.

These red-brown limonite-haematite rocks are rather porous, but the extent to which this is due to surface weathering is not certain. In thin section the original carbonates and silicates of the marble are seen to be entirely removed and replaced by fine dusty haematite, quartz and sericite. A typical specimen was found to consist of 30% haematite, 25% quartz, 5% sericite and 40% pore space. The haematite occurs as cryptocrystalline to microcrystalline translucent flakes, which still preserve a palimpsest representation of the original marble texture by edging spaces previously occupied by calcite which has since been entirely removed. The quartz occurs in

FIG. 58.

Photomicrograph of tourmaline-chlorite-quartz rock.

Rock A185-311. Location 186.115. Ordinary light.

Aggregates of zoned tourmalines, bunches of pale green chlorite and quartz spattered with ragged chlorite and biotite inclusions.

FIG. 59.

Limonite-haematite skarn replacing marble.

Location 178.098.

Well defined contact (perhaps a former joint) between limonite-haematite-quartz skarn (top) and normal marble (below). Folds in compositional bands in the normal marble pass uninterrupted into the skarn. Coin diameter 2.8 cm.

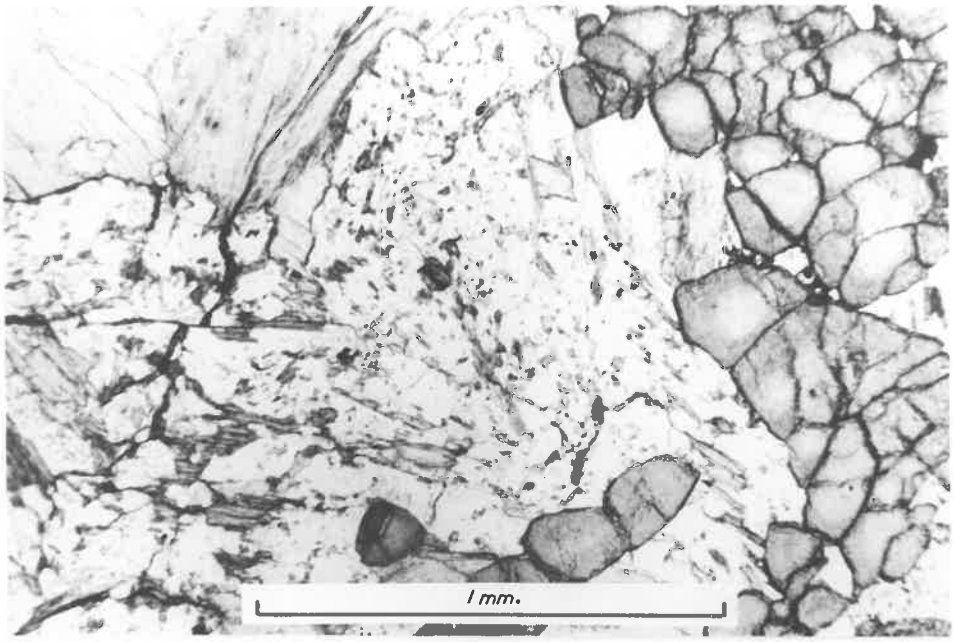


FIG. 58

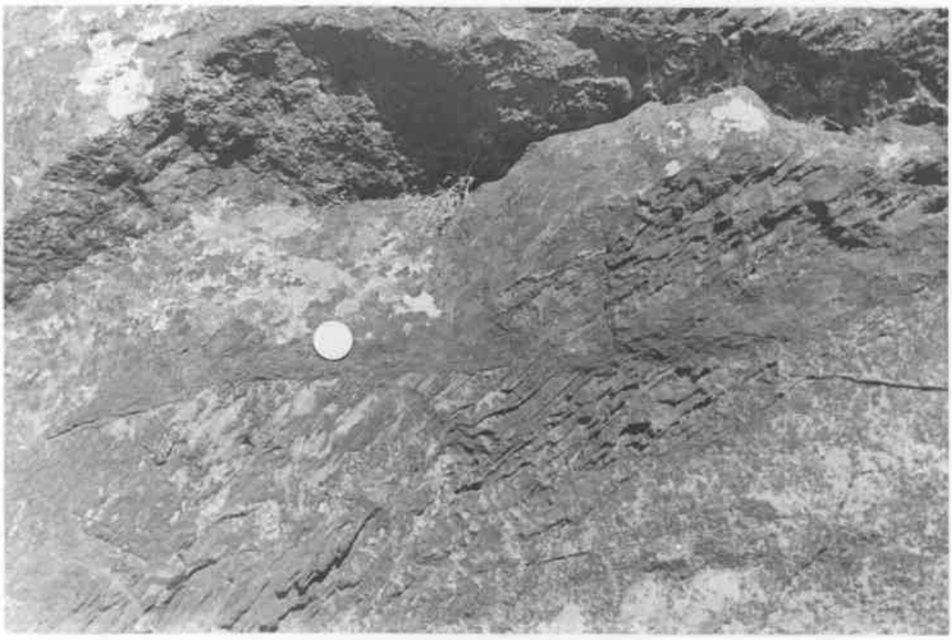


FIG. 59

clumps of subhedral to anhedral grains or as quite euhedral isolated prisms enveloped in the haematite. Disoriented flakes of sericite appear in small dispersed groups.

At both localities the proportion of marble replaced is very small. There is no direct evidence to indicate the origin of these rocks. It is possible that they may represent deposited iron oxides expelled during the replacement of nearby marble with quartz-albite rocks, a replacement occurring extensively in both localities where the haematite rocks occur. However, the occurrence of similar rocks replacing the marble on a larger scale north of the Marne River Reserve, and in the north-west corner of the area where the quartz-albite rocks are rare and do not replace the marble, is not in favour of this hypothesis. Unfortunately outcrop adjacent to these larger haematite-quartz bodies is poor and the contacts with the marble and schists are not exposed.

(e) Tourmaline Metasomatism.

Tourmaline has been noted as an accessory constituent of many metasediments in the Cambrai area; here the tourmaline is considered to represent an original relic sedimentary component which has, in some cases, been recrystallized or added to during the metamorphism. The concentration of tourmaline in the heavy mineral layers indicates that much tourmaline dispersed in the sediments is of clastic origin. However, some accessory tourmaline may be purely metamorphic in origin, having formed during the recrystallization of the original clay constituents of the sediment (Frondel and Collette, 1957). Tourmaline appears also in some acid intrusive rocks; the quartz veins, aplites and pegmatites to be discussed in a subseqent chapter. The metasediments bordering tourmaline bearing pegmatites may be loaded with introduced tourmaline, and there is clearly a genetic tie between the tourmaline and the introduced vein. There are a few minor occurrences of tourmaline rich rocks in which the source of the tourmaline is not apparent and these will be described here.

At 186.115 (specimen 311) a thin "vein" of tourmaline rich rock is embedded in quartzo-felspathic schists. The rock has a dark green black colour and consists of thin black tourmaline prisms arrayed in lineated rosettes piercing bunches of coarse pale green chlorite in a matrix of quartz. In thin section the tourmaline is seen to be quite euhedral and strongly zoned, with the crystal cores being coloured a deep sky blue, passing outwards through yellow zones to deep green rims. The quartz of the matrix encloses many small flakes of biotite and this suggests that the chlorite (β = 1.604 $\pm$ .003, positive, $\triangle$ = .008; a ripidolite near corundophilite) has formed partly from the alteration of biotite during the tourmalinization, the minor rutile present being a bi-product of the reaction. Similar yellow-green tourmalines occur in an altered biotite-andalusite schist at 212.181 (specimen 532) as streams of small stumpy euhedral prisms.

An unusual tourmaline bearing rock at 232.165 (specimen 375) appears to form an 8" thick stratigraphic layer embedded within alterating staurolite, andalusite and calcareous schists. This rock is composed of interlocking cone-like aggregates of golden-brown tourmaline and pale pearly anthophyllite in a matrix of chlorite, cordierite, albite and quartz. In thin section the tourmalines are strongly pleochroic from very pale lemon to a rich golden-yellow. Although the birefringence is uniform throughout there are patches within the tourmaline which are colourless and non-pleochroic. Small ilmenite and rutile crystals are included within the tourmaline and thin fibres of anthophyllite resembling sillimanite penetrate the edges of the crystals and extend as fine needles well into their cores. An identical feature has been described by Tilley and Flett (1930) in a "tourmalinized" rock from Kenidjack, Cornwall. The tourmaline has a very low magnetic susceptibility, suggesting a low iron content and a dravitic composition. Fan-like arrays of ripidolite, similar to those in specimen 311 described above, appear to have formed after an original pale biotite which occurs as relics within the large anhedral quartz grains. Large irregular grains of cordierite are intergrown with albite and quartz. In this case it is uncertain whether the tourmaline is a result of metamorphism of a boron rich shale or is the result of boron metasomatism.

(f) Post-tectonic Muscovite Porphyroblasts.

Stumpy post-tectonic muscovite porphyroblasts, cross-cutting or mimicing syn-tectonic structures are common in the more pelitic and semi-pelitic rocks of the Cambrai region. These muscovite flakes are apparently common throughout the whole metamorphic belt in the Eastern Mt. Lofty Ranges, for they have been a subject of comment by most previous petrological investigators in this region. In the present area they have been noted within the pelitic and semi-pelitic rocks, especially those rich in alumina, in all grades of metamorphism above the andalusite isograd, but the coarser and more prominently developed muscovites are found within the sillimanite zone. The greater abundance of these muscovites in the more pelitic members gives their distribution a stratigraphic appearance, and only rarely are sufficiently large and fresh outcrops exposed where the true relationship of these cross-cutting flakes to the original bedding can be observed.

The relation of the distribution of the muscovite flakes to bedding and foliation is clearly displayed on the walls of a cave in a uniform semi-pelitic schist at 207.123. Here the muscovite flakes have a patchy distribution, the boundary of the individual patches being quite irregular and having no relationship to either the bedding or schistosity traces. Thus the distribution of the muscovite flakes cannot be related specifically to the original sedimentary banding.

The confinement of the muscovite flakes to the more pelitic schists, especially those with excess alumina, is due to the influence of initial composition. It is considered that the most likely origin for these post-tectonic muscovite flakes is through introduced potassic solutions soaking through the metasediments late in the metamorphic history, the muscovite forming by the fixation of introduced potash in those pelitic rocks where alumina was available. A possible source for the potassic solutions is considered in the next section.

Billings (1938) has described similar sporadic post-tectonic muscovite from Western New Hampshire, where this mineral shows all stages in the pseudomorphic replacement of sillimanite. Analyses indicated that potash had increased in the rocks containing the late muscovite and Billings concluded that potash and water had been introduced from some extraneous source, perhaps a muscovite bearing granite.

(g) Potash Deficient Schists.

The petrography of the potash deficient schists, in particular anthophyllite and chlorite-anthophyllite schists, has been described in a previous section. Although some of these rocks may be the result of isochemical recrystallization of initially potash deficient sediments, it is difficult to explain the widespread occurrence of anthophyllite-chlorite-quartz-albite schists in the meta-arkose sequence on the western side of the area in this way. The normal meta-arkose occurring in the syncline in the central and eastern side of the area contains biotite and some potash felspar, but no potash deficient anthophyllite or chlorite rocks have been observed there. This is in marked contrast to the wide belts of stratigraphically equivalent anthophyllite and chlorite bearing quartzo-felspathic schists of the western side of the area, in which potassic minerals are entirely lacking. The patchy occurrence of these potash deficient rocks within the well-bedded meta-arkoses and the local gradation within narrow beds from biotite rich schists to anthophyllite rich schists over a distance of a few feet along the strike has been described. These observations cannot be attributed to original sedimentary facies differences, and some allochemical process involving widespread leaching-out of potash in diagenesis or metamorphism must be invoked.

Some of the other peculiarities of the metamorphism of the sediments in the western side of the area have also been described, in particular, the increased degree of recrystallization and grain growth in the western rocks and the different petrological character of the calc-silicates. Another unusual feature is the occurrence of the quartz-potash felspar rocks, described earlier in this chapter, within the potash deficient meta-arkose.

It is possible that all these unusual features are interconnected and are basically the result of a single widespread metasomatic process affecting the whole of the western side of the area. Whatever the nature of this metasomatic process, potash has been effectively removed from large belts of meta-arkose having a low, but appreciable, initial potash content. This removed potash, if reprecipitated within select zones, might supply a ready explanation of the quartz-potash felspar rocks.

A similar problem involving an intimate mixture of potash deficient and potash enriched rocks was met by Chinner (1956) in the Tanunda Creek area. He concluded that it was likely that potash removed from the potash deficient schists was re-deposited in the granitic gneisses as a commonly observed late replacive microcline. Extensive zones of potash deficient rocks have also been described by White (1956), and it must be concluded that zones of potash deficiency are characteristic of the high grade metamorphic rocks surrounding the granitized core of the metamorphic belt in the eastern Mt. Lofty Ranges.

The close association of zones of granitization with abundant and widespread potash deficient rocks in the high grade metamorphics of the eastern Mt. Lofty Ranges is, indeed, one of the major unsolved riddles of this metamorphic province. In the consideration of this problem two questions are vital;

(1) has potash been removed from the potash deficient schists?

(2) if potash has been removed, has it been redeposited in the granitized gneisses, or carried to lower temperature zones?

The solution to both of these questions requires considerable chemical data and the problem as a whole is outside the scope of the present work.

The conditions leading to a complete removal of potash from some rocks are not understood, but the experimental work of Orville (1959) may have application here. Orville (1959) has investigated alkali ion exchange between alkali salt solutions and felspars, and has found that the reaction

$$\left[\text{K-Ca felspar} \;+\; Na^{+} \rightleftharpoons \text{Na-Ca felspar} \;+\; K^{+}\right]$$

tends to proceed to the right. It is almost universally accepted that hydrous solutions (including original connate waters) are expelled during increasing metamorphism in a sedimentary pile. If the sediments are marine, the expelled waters might be highly charged with soda derived from initially trapped sea water. These solutions, ascending upwards away from the core of the metamorphism would cause ion exchange reactions such as those described by Orville. This could result in widespread potash deficiency in the higher grade metasediments similar to that found at Cambrai and elsewhere in the Mt. Lofty Range metamorphic belt. The potash rich solutions passing on away from the potash deficient zones may either form local concentrations of potash rich rocks (such as the quartz-potash felspar rocks ?), or become dispersed in the lower temperature zones where they result in the growth of late muscovite in rocks containing excess alümina.

Dunn (1942) has given the name "diabrochomorphism" to permeation metamorphism and metasomatism caused by solutions (connate or otherwise) driven up from depth by rising geothermal gradients. The above picture, although quite hypothetical at present, could explain many of the more puzzling features of the metamorphism and metasomatism of the Cambrai area.

Conclusions.

Metasomatism has played an important part in the metamorphic evolution of the Cambrai area, both in the formation of local metasomatic rocks and on a wider scale. Korzhinsky (1964) considers that a single ascending wave of solutions may produce diverse metasomatic effects and that it is not necessary to propose separate waves of metasomatism for each kind of metasomatic product. Thus the various metasomatic products described in this chapter could be more inter-related than might be supposed from their diversity.

CHAPTER 11.

INTRUSIVE ROCKS.

Introduction.

Intrusive rocks form only a small proportion of the rocks exposed in the Cambrai area. Although insignificant in bulk, they are of widespread occurrence and their distribution mineralogy and petrology are of particular interest. Unlike the large plutons of granite, tonalite and norite (Murray Bridge granite (Kleeman, 1934), Mannum granite (Goode, Alderman, 1927; 1929), Reedy Creek granodiorite and tonalite (Sando, 1957; Alderman, 1927), Long Ridge granite and Black Hill norite (Alderman, 1927)) located in a north-south belt east of the present area, the intrusive bodies here are always small, and take the form of dykes, plugs and pipes discordant, and lensoidal dykes, plugs and sills concordant, with the bedded metasediments into which they have been emplaced.

Where the intrusive activity has been strong during the late stages of the regional metamorphism, the intrusions pepper the metasediments rather than form a large single body irrespective of the nature of the intrusive rock. This form of intrusion appears to be related to the nature of the sedimentary medium and the physical conditions during the intrusion rather than the nature of the intruding material. Moderate temperatures and high confining pressures during the waning stages of metamorphism, and structural inhomogeneities within the metasediments, are believed to be major factors deciding the form of intrusion. Younger intrusives; aplites, quartz veins and some dolerites, emplaced during the last waning stages of metamorphism, have their intrusion forms strongly controlled by tension fractures, joints and faulting.

The intrusive rocks may be classified into four major groups:-

1. Metadolerites.
2. Granodiorite - aplogranite - aplite - pegmatite suite.
3. Syenites, diorites and other intermediate rocks.
4. Quartz veins.

Basic rocks of the dolerite-gabbro clan are widely distributed as small pods and lenses throughout the western part of the area, but have their most spectacular development in a small area on the eastern scarp, west from the town of Cambrai. The granodiorites are confined to the areas of highest grade of regional metamorphism, namely, in the south-west corner of the area and near the Kanappa Mine. Granites, aplites and minor pegmatites, which appear to be the later differentiates of the granodiorites, are found within the granodiorite belts and as streams of anastomosing cross-cutting dykes in the southern central portion of the area. Members of this acid series are not found north of the Marne River. Intermediate rocks of the syenite-diorite group are distributed widely in the central portion of the area. Although rare, isolated, and insignificant in amount relative to the other intrusives, their mode of intrusion and variable petrology is of considerable interest. Most quartz veins appear to be of the metamorphic segregation type and have been discussed elsewhere. Quartz-tourmaline veins in the Pine Hut area, however, cannot be attributed to segregation.

Three major factors, metamorphic grade, structure and physical nature of the metasediment intruded, appear to affect the distribution of the intrusives. The acid intrusives are confined to the deeper metamorphic levels. The intermediate and basic rocks are notably confined to the less competent metasediments, and, as can be seen from Plate 1, do not penetrate the thick massive arkose forming the centre of the syncline in the northern part of the area. The intrusions do not seem to be controlled by pre-existing folds in the metasediments, although weaknesses along bedding planes and axial plane foliation have facilitated intrusion in many instances.

Rarely, the relative ages of the various intrusive types can be observed directly from cross-cutting relationships, but, in general, the degree of metamorphic reconstitution gives the only indication of age. The degree of metamorphic recrystallization depends not only on the relation of the time of intrusion to the time of metamorphism, but also on the metamorphic grade of the metasediments into which the rocks are emplaced. It is therefore difficult to compare the relative ages of similar rocks intruded into high and low grade areas. In general, dolerites were the first rocks to be emplaced and all have suffered mild to severe metamorphic reconstitution. It is likely, considering the various degrees of metamorphic recrystallization within some adjacent dolerites, that there were several periods of dolerite intrusion, ranging in age from the height of the metamorphism to a late metamorphic stage. The slight metamorphic reconstitution of the granodiorite-aplite suite suggests that these rocks were emplaced late in the metamorphic history. The syenites and diorites are not known to cut other intrusive rocks, but some contain dolerite xenoliths. They are known to be pre-faulting, and from their degree of metamorphism, are probably younger than the granodiorite-aplite suite. A porphyritic dolerite dyke intruded along a fault in the south-west corner of the mapped area cross-cuts granodiorite and aplite, and is the youngest intrusive known.

In general the intrusions are late metamorphic, and their degree of recrystallization and metamorphic alteration has not been sufficient to destroy their igneous textures. A few small amphibolites in the Saunders Creek area show insufficient evidence for intrusive contacts or relic igneous textures to decide an unequivocal origin.

Contact metamorphic effects are rarely observed. The intrusives being small, the weak contact effects were probably transient, being overprinted and dissipated by the effects of the waning regional metamorphism. Contact effects are preserved in some of the thin contact zones of the larger dolerite bodies and some of the syenite-diorite pipes, where a fine-grained recrystallized hornfelsic texture, but no new minerals, appears.

(a) Metadolerites.

Introduction.

The term "metadolerite", although used synonymously with the old term "epidiorite", is preferred by the present writer. As the origin of most of the basic rocks considered here is not in doubt, the term "amphibolite" is reserved for those amphibole rich rocks of dubious origin.

The metadolerites in this area may be divided into three groups based on geographic location:-

a. The Cambrai dolerite field.

b. Metadolerites of the western group.

c. Metadolerites of the southern group.

The metadolerites of group a. occur within the andalusite-staurolite zone of regional metamorphism; those of group b. within the andalusite-staurolite-kyanite and lower sillimanite zones; and those of group c. within the higher part of the sillimanite zone.

(1) The Cambrai dolerite field.

This field, located on the scarp west of Cambrai, contains over 600 separate dolerite intrusions within an area of 2½ square miles (Plate 1). The majority of these are small pods and lenses a few yards wide and 10-20 yards in length. Some of the smallest intrusions are only one foot in width. The largest intrusions tend to form elongate lenses and steeply dipping sills, up to one mile in length, injected along bedding planes in the well-bedded evenly dipping metasedimentary sequence. In this way there is minimum disruption of the continuity of the bedded sequence, and contact effects such as brecciation are, in general, totally lacking. Some sills, especially in the central portions of the field, obliquely transect the bedded sequence for part of their length. Several large irregular discordant plugs on the western and northern edges of the field show considerable brecciation of the quartzo-felspathic schists near their contacts. Quartzo-felspathic schists south of the large mass at 229.188 (Plate 1) have been crushed and folded up to 300 yards from the edge of the body. The irregular folding caused by the forceful intrusion is outlined by the mapped portions of several calc-silicate marker beds. Contact metamorphic effects are rarely observed.

The degree of recrystallization suffered by these dolerites suggests that they were intruded soon after the climax of the regional metamorphism. The original pyroxene has been replaced by amphibole even in the centre of the largest masses. The plagioclase, however, generally preserves its original twinning and lath-like form, so that a palimpsest igneous texture is apparent.

There is a complete gradation between metadolerites having an abundance of plagioclase phenocrysts (up to 60%) to metadolerites totally lacking them. Although the average size of the phenocrysts is 0.5 mm., some may reach 2-3 cm. There is no relation between the size of the intrusion and the abundance of phenocrysts. Many bodies are composite, some portions having an abundance of phenocrysts and others lacking them. There appears to be no simple pattern in the distribution of the phenocryst rich dolerites, although it was observed that most of the intrusions on the south-west edge of the field were markedly porphyritic. No evidence was found for original pyroxene phenocrysts.

Igneous structures are rare. A weak alighment of phenocrysts, plagioclase laths and iron ore granules near the contacts of some intrusives has been observed.

Xenoliths are lacking.

The predominant structures are metamorphic. Although the centre of the intrusions resist any external structural impositions, the contacts often take on a schistose foliation, defined by the orientation of amphibole prisms, parallel to the axial plane foliation in the adjacent schists. Within these schistose contacts the palimpsest igneous texture has been completely replaced by a granoblastic metamorphic one. In some of the schists in the area covered by the dolerite field a second foliation (a crenulation cleavage here termed S_4 - see Chapter 12) transects the main schistosity producing steeply west plunging crenulations. This newer foliation may also transect the schistose portions of the metadolerites, resulting in a steep westerly plunging lineation defined by the amphibole prisms. Occasionally a porphyritic dolerite has become schistose. The phenocrysts, recrystallized to a granoblastic aggregate of andesine and scapolite, become flattened in the foliation plane. If intersected by the newer foliation the "phenocrysts", now granoblastic aggregates of small grains, become elongate parallel to the amphibole prism lineation. The resultant rock resembles a lineated augen gneiss. The implications of these structures will be discussed further in the next chapter.

The metadolerites crop out as tough massive jointed blocks highly resistant to weathering. Unlike the adjacent metasediments they have strongly resisted crushing associated with the faulting. Fresh broken surfaces expose the dark green-grey colour of the groundmass and the partially altered grey-white plagioclase phenocrysts. Generally the groundmass is coarse enough for the plagioclase laths, amphibole grains and the relic ophitic texture to be clearly discerned. Grainsize of the groundmass varies with the size of the intrusion and ranges from 0.2 mm. or less, for the contact phases of small bodies, to 1 mm. in the larger bodies. Rarely the grain-size may exceed this and reach 5 mm., the rock then having the appearance of a medium grained gabbro.

In their mineralogy the metadolerites are relatively simple. The major minerals are amphibole, plagioclase and scapolite. Biotite, often partially altered to chlorite, is always present, although generally sparse. Iron ore, sphene and apatite are constant accessories. Epidote and sericite-muscovite, although unstable within the completely reconstituted assemblage, are always present as transient alteration products of plagioclase phenocrysts. Potash felspar calcite and tourmaline are rare, minor constituents.

Despite the small number of major mineral phases, the mineral textures of the metadolerites are often complex and variable. Except in the more recrystallized schistose metadolerites, where the textures are essentially metamorphic, palimpsest ophitic and porphyritic, replacement and metamorphic textural features are combined in the overall metadolerite texture. Replacement textures arise from the growth of scapolite from the plagioclase. Where scapolite is absent, the palimpsest ophitic textures pass to granoblastic metamorphic textures by granular recrystallization and grain growth of the amphibole and plagioclase. Where scapolite is present, the plagioclase is progressively replaced by large skeletal crystals of scapolite. With the development of foliation, these large scapolite crystals become granular mosaics consisting of numerous granules of slightly varying optical orientation, and finally, granular crystalloblastic aggregates interspersed with recrystallized plagioclase granules and amphibole prisms. In the porphyritic metadolerites the phenocrysts may be seen in all stages of replacement by coronas or irregular penetrating fingers of scapolite. Concomitant with the scapolitization of the phenocrysts is the growth of sericitic mica and spongy clinozoizite aggregates within the remaining central portion of the phenocrysts. With the complete replacement of the original plagioclase phenocrysts, the epidote-sericite sponges may recrystallize, but are not stable with the recrystallized groundmass of the rock. Eventually, when equilibrium is more closely attained, the epidote disappears with the attendant increase of the anorthite content of the new metamorphic plagioclase, and the white mica presumably contributes to a slight increase of the minor biotite content of the rock. However, where the plagioclase is almost totally replaced by scapolite, epidote may become a stable phase, although present in minor amount.

Amphibole normally forms 35-60% of the metadolerites. Although variable in grain size up to 5 mm. in various rocks, it has a rather uniform character. The grains are usually equidimensional to slightly elongate and have very ragged edges. Myriads of iron ore granules and dust-like inclusions are present except in the most recrystallized rocks. The normal pleochroism is

X = pale lemon

Y = mid green to mustard green

Z = deep green with a blue tinge,

but the intensity of colour is not great. Patchy colouration and a slight colour zoning, with deeper green rims, is not uncommon (see also Read, 1931, p. 73). Twinning is rare and undulose extinction is a general rule. In the lower grade northern part of the area (commencement of the andalusite zone) the pleochroism is

X = neutral

Y = pale green

Z = pale blue-green,

and the grains have a stronger tendency for a fibroidal, bowtie structure.

Biotite, in the form of small stumpy diversely oriented flakes, is closely associated with the amphibole, with which it appears to be in stable equilibrium. It is present in all rocks in amounts ranging up to 5% (25% in one exceptional instance, rock specimen 25, location 236.212). In different rocks the pleochroism is variable with

X = very pale lemon to yellow

Y - Z = pale gold brown to sepia

Partial alteration to pale green slightly pleochroic chlorite, showing anomalous deep purple or mauve-grey interference colours, is common, but not universal. Metadolerites from the lower grade northern part of the field contain some chlorite resulting from the alteration of amphibole.

<u>Iron ore</u> and <u>sphene</u> are also closely associated with, and included in the amphibole. The iron ore appears as numerous, small scattered granules or aggregates of granules, commonly rimmed by granular sphene. Occasional brown halos in the amphibole have granules of sphene at their cores. <u>Apatite</u> is an ubiquitous accessory, occurring as minute clear prisms embedded in the relic igneous plagioclase. <u>Calcite</u> is an uncommon constituent occurring only in those metadolerites of lower metamorphic grade. Rare grains of <u>potash felspar</u>, with incipient cross-hatched twinning, may be found in the more recrystallized metadolerites.

The <u>plagioclase phenocrysts</u> exhibit various degrees of replacement by scapolite with concomitant alteration to epidote and sericite. Some phenocrysts have clear unaltered cores which display well developed twinning on albite, Carlsbad-albite, Carlsbad, pericline and Ala B laws and distinct normal zoning. Rare instances of thin oscillatory zoning have been observed. A phenocryst extracted from rock specimen 27b gave the refractive indices $\alpha = 1.561$ $\beta = 1.567$ $\gamma = 1.571$ ($\pm .003$) indicating a composition of An 62 (labradorite-bytownite). Table 8 illustrates plagioclase composition measurements determined on a 5-axis Leitz Universal Stage using the Emmons 5-axis technique (Emmons, 1943)

TABLE 8.

COMPOSITION OF PLAGIOCLASE IN METADOLERITES OF THE CAMBRAI DOLERITE FIELD.

Rock No.	Location	% An cores	% An rims	Twin laws	Method
Plagioclase phenocrysts.					
501	239-188	An 80			⊥a
27b	230-212	An 82	An 62	Albite, Carlsbad-albite	Emmons
571a	236-220	An 55-78	An 38-47	Albite, Carlsbad	Emmons
571b	236-220	An 60			⊥a
Groundmass plagioclase (relic igneous).					
501	239-188	An 58	An 20		⊥a
516a	232-188	An 43-57	An 20		⊥a
27b	230-212	An 61-68	An 32-40	Albite, Carlsbad-albite with rarer Pericline, Carlsbad	Emmons
27c (non-porphyritic)	230-212	An 57-61	An 31-43	Albite, Carlsbad, Pericline, Ala B	Emmons
28	229-211	An 37	An 25	Albite, Carlsbad-albite, Carlsbad, Pericline	Emmons
Matrix plagioclase (metamorphic).					
823	235-171	An 30-38 (unzoned)			⊥a

and measurement of the angles X' ^ (010) or (001), observed along the a-axis (using curves in Troger, 1956, page 111). The cores of some phenocrysts approach An 82. The groundmass plagioclase, while preserving its original lath-like form and twinning, always displays strong normal zoning. The wide range of composition, An 20 - An 68, is a reflection of the degree of disequilibrium. In the metamorphically reconstituted dolerites the plagioclase is nonzoned. The variation of composition of the plagioclase in rock specimen 823 (Table 8) is related to the presence of epidote marking the centres of former phenocrysts, which, as yet, have not reacted with the matrix to establish complete equilibrium.

Spongy epidote and associated sericitic mica are formed during the alteration of the plagioclase phenocrysts, especially when these are being replaced by scapolite. The epidote, when recrystallized, has the optic properties of an iron free clinozoizite. In the early, poorly recrystallized, stages of its formation two phases may be present; one, with deep grey interference colours, being embedded in a matrix having anomalous Berlin blue interference colours. In ordinary light both have the same relief and appearance. With complete metamorphic reconstitution both the epidote and the sericite disappear from the assemblage.

<u>Scapolite</u>, although a major mineral phase in most of the metadolerites, is very erratic in occurrence. As pointed out previously, the composition of the scapolite replacing the plagioclase phenocrysts of the metadolerites is not significantly different from that of scapolites extracted from other diverse rock types in the area. Scapolite replacing a phenocryst in rock 25 yielded $n_{\omega} = 1.566$ ($\pm .001$), and in rock 27b, $n_{\omega} = 1.578$ ($\pm .001$), indicating compositions of Me 45 and Me 60 respectively. Before the more advanced stages of the metamorphic recrystallization to a granoblastic aggregate, the scapolite is observed in all stages of replacement of plagioclase. The genesis of the scapolite is essentially concerned with the source of the chlorine and other volatiles necessary for its formation. Some metadolerites are devoid of scapolite while others may carry up to 40%. It is therefore unlikely that the volatiles were derived from the dolerites themselves. Some of the bedded metasediments intruded by the dolerites are rich in scapolite, and it is consistently observed that dolerites intruding a scapolite rich bed are also rich in scapolite, while scapolite is absent from those dolerites injected into quartzo-felspathic schists lacking scapolite. This observation also

holds for metadolerites in other parts of the Cambrai area. It is concluded that the scapolite of the metadolerites results from the local absorption of chlorine and other volatiles from the injected meta-sediments. This absorption may have occurred most freely during the cooling of the intrusions and their subsequent amphibolitization at the elevated temperatures of the regional metamorphism.

In the south-western portion of the field, a number of metadolerites are traversed by white, medium-grained plagioclase veins. These veins, up to several inches in width, are confined to certain joint directions and are composed essentially of nonzoned, granoblastic plagioclase (An 36), minor irregularly distributed amphibole, interstitial epidote and rare euhedra of pale brown sphene. Rare vughs are lined with euhedral plagioclase. The plagioclase, with well developed lamellar twinning on the albite and pericline laws, has undulose extinction, perhaps due to strain, and abundant clay-like alteration products which yield the characteristic white colour of the rock. The origin of these veins is not known, although they seem to represent a late pegmatitic phase of the dolerites injected into early joints. They have not been observed cutting the surrounding metasediments.

In the more scapolitized metadolerites, scapolite, sometimes with quartz, may occur as joint infillings. The refractive index, n_ω, of the scapolite in occurrences of this type ranges from 1.5675 to 1.5775 ($\pm$ .001) (seven determinations) indicating a composition range from Me 46 to Me 60.

Contact metamorphic effects within the metasediments adjacent to the metadolerites is seldom apparent. At the southern contact of the large dolerite plug at 229-188 the quartzo-felspathic schist has an exceedingly fine-grained (0.04 mm.) granular texture, with several thin (5 mm.) veins, obliquely transecting the bedding traces, consisting of coarser (0.10 mm.) plagioclase, severely altered to distinct sericite-muscovite flakes; abundant zoned tourmaline, forming 20% of the vein; and accessory rutile. The tourmaline occurs as euhedral prisms oriented parallel to the vein walls. Tourmaline poikiloblasts occur as an accessory in the biotite rich metadolerite, specimen 25, location 236-212.

A peculiar, structureless, green spotted hornfels at the northern edge of the large dolerite plug at 236-220 appears to be of contact metamorphic origin. The evenly distributed spots consist of aggregates of small clear diversely oriented pale green actinolite poikiloblasts with lesser pale brown biotite, while the pale matrix consists of abundant quartz grains embedded in fine granular albite. Iron ore is a common accessory and rounded, "water-worn" zircons suggests that the rock is basically of sedimentary origin.

A rock of somewhat similar appearance has formed as an intermediate stage of albitization of a small dolerite lens at 232.188. This example of albitization has been dealt with in chapter 10 where it was concluded that the albitization was not associated with the dolerite intrusion, and has no connection with the scapolitization of the metadolerites.

(2) Metadolerites of the western group.

Small dykes and lenses of metadolerite are widely, but sparsely distributed in the major anticlinal zone in the western part of the mapped area (Plate 1). The majority of occurrences are small, being only a few tens of feet in length. A thin dyke south-east of Moralinga, however, has a length of one mile, and west of this two lenses adjacent to a marble bed are up to half a mile in length and, in places, 300 feet in width. The intrusions of this group are almost invariably oriented in the tectonic trend, being injected into the metasediments parallel to the north-south foliation, the axial plane cleavage of the first and major deformation. Owing to the greater degree of regional metamorphism suffered by the majority of these metadolerites, contact metamorphic and structural effects have not been observed. The degree of metamorphic and structural reconstitution is generally strong and the relative ages of the different dolerites is impossible to determine, although the majority appear to have been intruded near the height of the metamorphism. The one mile dyke south of Moralinga still preserves a blastophitic texture, although the original pyroxene is entirely altered to amphibole, and appears to be one of the youngest dolerites represented. Most of the other metadolerites intruded into the schists have taken on a strong foliation, but rarely any lineation. Small pods injected into marble have suffered considerable metasomatic alteration, recrystallization and retrograde effects. Porphyritic metadolerites have not been observed in this area.

The textures range from blastophitic to schistose, through varying degress of granular recrystallization and foliation development. The highly altered rocks intruding the marbles have coarse irregular textures.

The long dyke south-east of Moralinga zonsists of massive green-grey fine grained metadolerite in which plagioclase, preserving its original twinning and lath-like form, shows a relic ophitic textured relationship with the amphibole which has totally replaced the original pyroxene. The amphibole has the pleochroic scheme X = pale lemon, Y = mustard-green, Z = blue-green, the optic properties $2V_x$ 70° ($\pm$ 2°) and $Z \wedge c$ 14° ($\pm$ 1°). It contains numerous iron ore and other inclusions which sometimes preserve structures that may be related to the former pyroxene lattices. The iron ores, in particular, often form large skeletal grains seeming to ghost original pyroxene crystal directions. Small isolated specks of biotite are closely associated with the amphibole. The plagioclase is zoned, having cores of An 49 and rims of An 30, is slightly sericitized and encloses numerous apatite needles. A small amount (3-5%) of quartz occurs as granules. Scapolite, calcite and epidote are absent.

The metadolerites represented by specimens 750 (location 147-237) and 764 (location 146-235) are of similar composition to that described above, but show stages in the development of a foliated crystalloblastic texture. The original plagioclase laths have recrystallized to fine nontwinned granules, with a composition almost in equilibrium with the metamorphic grade (upper staurolite zone). The amphibole has recrystallized to inclusion free, oriented prisms; although some clots, about which the foliation diverges, consisting of inclusion crowded unoriented coarse amphiboles, still remain. In rock 750 the plagioclase has a composition of An 34 (R.I. and Emmons 5-axis method) and the amphibole has $2V_x$ 71° (± 3°) and $Z \wedge c$ 14° (± 1°). In rock 764 the plagioclase has a composition of An 41 (R.I. and Emmons 5-axis method) and the amphibole $2V_x$ 64° (± 1°) and $Z \wedge c$ 12.5 (±1°).

Those few metadolerites intruded into scapolite bearing metasediments may have much of their felspar replaced by scapolite.

Metadolerites intruding the marbles may be so recrystallized and metasomatically altered that their original nature becomes obscured. A series of small pods, some only one foot in width, of coarse-grained pale green-grey rock in the vicinity of 175-174 consist largely of plagioclase, chlorite and calcite with minor amphibole, phlogopite and quartz and accessory apatite, iron ore and sphene. The minerals are regularly distributed in a non-foliated texture. The plagioclase occurs as randomly oriented laths undergoing replacement by irregular calcite grains. The laths preserve their original complex "igneous" twinning and have compositions ranging from An 60-70 in their non-zoned cores, to An 24-59 on their rims. The chlorite (Δ .004, +ve) forms large (to 3 mm.), randomly oriented, pale green flakes with which are associated large irregular iron ore aggregates and rare pale brown phlogopite books. The green actinolite prisms are confined to, and embedded in evenly distributed quartz rich aggregates. Scapolite is absent from the rock, although associated with calcite and quartz in joint infillings. A reaction such as

$$\left[\text{hornblende} + \text{epidote} + CO_2 + H_2O \rightleftharpoons \text{chlorite} + \text{quartz} + \text{calcite}\right]$$

(Tilley, 1938) appears to have occurred in this rock. Miyashiro (1958) and Shido (1958) consider that, in the greenschist facies, the presence of local high water and CO_2 concentrations would promote the formation of calcite-quartz-chlorite-epidote-albite rocks from actinolite-chlorite-epidote-albite rocks. The presence of basic plagioclase in this rock, in place of epidote-albite, may be related to a higher temperature during the chloritization.

An apparently intrusive pod near 153-230 has sideritized the surrounding marble. The rock consists of numerous, 1-2 mm., pink garnet xenoblasts embedded in a matrix of close-packed green and white amphiboles. The pale pink garnet contains numerous fine inclusions arranged in an unusual and distinctive "brain structure". The white amphibole is cummingtonite with common polysynthetic twinning on $\{100\}$, $2V_z$ 80° ($\pm 1^\circ$)

and $Z \wedge c$ 18° ($\pm 2^{\circ}$). This amphibole merges into patches of a green amphibole with the pleochroic scheme X = pale lemon, Y = mid-green, Z = greenish blue, $2V_x$ 74° and $Z \wedge c$ 12°.

The two amphiboles may form separate crystals in non-parallel growth, crystals in parallel growth (with $\{110\}$ as common planes) or an irregular patchy inter-merging of the two phases within the same crystal. Iron ore is a common accessory. No felsic minerals are present. Although the rock is an intrusive, no evidence of the original nature of the rock is preserved.

This greater degree of recrystallization and alteration of metadolerites injected into the marbles might be attributed to reactions occurring during slow cooling under high confining pressures of H_2O and CO_2 vapour trapped within the intruding body by the impervious marble host. It is evident that much interchange of material between the metadolerite and the marble host has occurred.

(3) Metadolerites of the southern group.

Some 30-40 irregular dykes and bodies of metadolerite occur in close spatial, but apparently genetically fortuitous, association with quartz-albite rocks in the Saunders Creek area, and small lenses up to 100 yards in length are sparsely distributed throughout the granodiorite field in the south-west corner of the area (Plate 1).

Both porphyritic and non-porphyritic varieties occur, showing varying degrees of destruction of their blastophitic textures. The most recrystallized meta-dolerites were presumably intruded earlier in the metamorphic history. A thin dyke of porphyritic dolerite injected into a fault plane in the south-west of the area has been uralitized, but otherwise preserves its original texture.

Dolerites intruding limestones in the Saunders Creek area generally outcrop in round pod-like forms. Their contacts with the marble and the occasional marble xenoliths have reactions zones up to 2-3 feet in width. These reaction zones comprise an inner zone (near the dolerite) of finer grained amphibole schist with numerous large scapolite porphyroblasts and an outer zone of altered marble containing abundant large amphibole prisms.

Not all metadolerites intruding the marbles show these skarn-like reaction zones, however. In some cases the boundaries are knife sharp and strong boudinage of the intrusives is often apparent (Fig. 61).

FIG. 60.

Photomicrograph of porphyritic metadolerite.

Rock A185-501. Location 239.188. Crossed polars.

Relic phenocrysts with coronas of scapolite in a recrystallized matrix of amphibole plagioclase and scapolite.

FIG. 61.

Boudinaged fine-grained metadolerite intruded into marble. Location 169.131. The marble shows plastic folding near the boudins and clots of secondary scapolite, amphibole, quartz and mica have formed between the boudins.

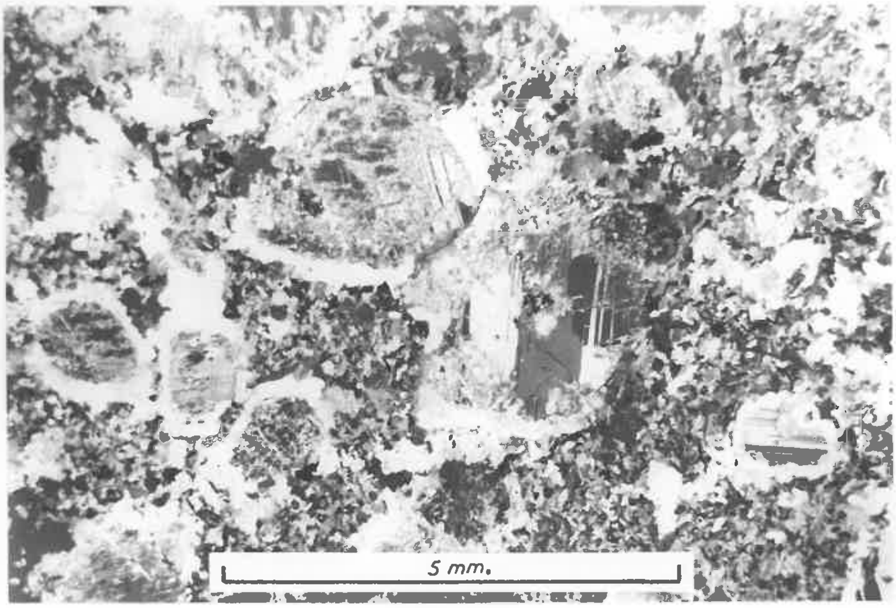


FIG. 60

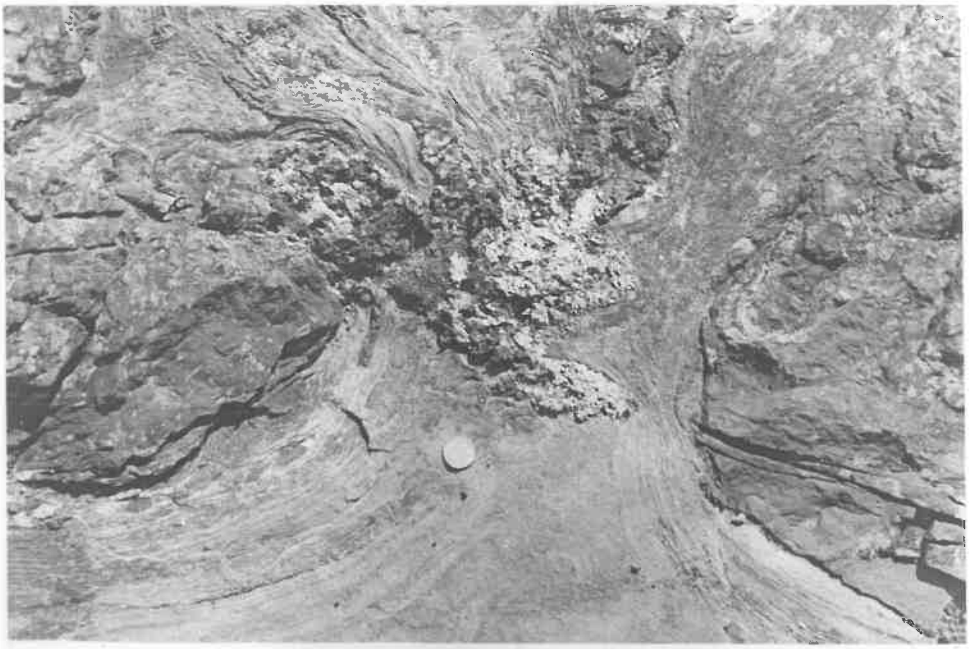


FIG. 61

A typical non-porphyritic metadolerite of this high grade area is represented by specimen 139 (location 204-081), taken from a narrow, half-mile long dyke intruded parallel to the foliation in micaceous quartzo-felspathic schists. The rock is a slightly foliated and lineated, dark green-grey metadolerite. The lineation seems to be an expression of a second foliation (crenulation cleavage) developed in the adjacent mica schists where it has produced crenulation axes parallel to the amphibole prism axes in the metadolerite. The texture is crystalloblastic and even-grained. The rock consists of 50% amphibole, 40% plagioclase, 10% quartz with sparse biotite altering to chlorite and accessory iron ore, with granular coronas of sphene, and apatite. Epidote and potash felspar occur in cross-cutting veinlets. The amphibole in this higher grade of regional metamorphism (sillimanite zone) has lost its blue colour.

X = mustard lemon

Y = brownish green

Z = deep green and has $2V_x$ 78° (± 1°)

and $Z \wedge c$ 13° (± 2°). The plagioclase has a composition of An 38 (± 5% An due to the zoning). Those plagioclase grains showing twinning (about 50%) are twinned on the albite and rarely the Carlsbad laws, their complex "igneous" twinning having been completely destroyed in recrystallization.

Several porphyritic metadolerites are known in this area. Rock 133 (location 183-093) has an oriented flow texture with lath-like phenocrysts of both plagioclase (now highly sausseritized) and amphibole up to 1 cm., in a fine-grained 0.1-0.2 mm. groundmass of granoblastic plagioclase (An 37 cores, to An 12 rims) and green amphibole ($2V_x$ 71.5° (± 2°) and $Z \wedge c$ 15° (± 2°)). Optic axial measurements on the plagioclase gave consistently higher results than expected for plagioclases of the composition determined from the low temperature migration curves, suggesting that the plagioclase has not fully attained a low temperature structural state.

The porphyritic metadolerite 152 (location 184-086) has intruded the quartzo-felspathic schists as a series of massive bosses. Most of the rock consists of clear unaltered plagioclase phenocrysts showing well preserved complex twinning. The unzoned cores, often with a euhedral crystal form against a thin, normally zoned more sodic rim, have a composition near anorthite ($X' \wedge 010$ = 43-45°, $\perp a$). The groundmass consists of recrystallized, clear, granoblastic amphibole and rarely twinned, clear, anhedral metamorphic plagioclase (An 42-79). Apart from rare incipient saussuritization and the growth of minor scapolite, there is little sign of alteration of the phenocrysts. The apparent stability of the phenocrysts in the well recrystallized matrix is somewhat anomalous, but might be explained by a rapid loss of water vapour to the surrounding schists during the waning stages of the high grade metamorphism, thus considerably lowering the diffusion rate and preserving the phenocrysts intact.

Rock 146 (location 182-085) is of some interest because it is one of the rare metadolerites in which the two amphiboles hornblende and cummingtonite are in stable equilibrium. The rock is fine-grained, foliated and slightly lineated. The plagioclase has been arrested in an intermediate stage of recrystallization, some portions of the rock consisting of small untwinned granules of plagioclase in crystalloblastic texture with the amphibole, while others yet retain their blastophitic texture, where the plagioclase preserves its original lath-like form and complex igneous twinning and is strongly zoned (An 57-73), the rims being the more sodic. Rare plagioclase phenocrysts up to 2 mm. have cores approaching the composition An 81. Albite-Ala B twin laws were found to be commonly preserved. The hornblende is in considerable excess of the cummingtonite (5:1). The hornblende has the optic properties $2V_x$ 77° ($\pm 2°$) and $Z \wedge c$ 15° ($\pm 1°$) and the pleochroic scheme: X, pale lemon; Y, mid-green; Z, green with bluish tinge. The cummingtonite is colourless and has the properties $2V_z$ 77½° ($\pm 2°$) and $Z \wedge c$ 19½° ($\pm 1°$). The cummingtonite properties suggest a composition near 65% of the magnesium molecule (Deer, Howie and Zussman, Vol. 2 p.242). The cummingtonite generally occurs in homoaxial intergrowth with the hornblende, although it may form separate individuals. The intergrowths may be irregular, or regular, using $\{110\}$ as boundary planes between the two phases. It is also not uncommon to find grains with simple twinning on $\{100\}$, one side of the composition plane being cummingtonite and the other hornblende. Minor quartz and iron ore and accessory apatite complete the assemblage. Sphene and biotite are absent.

The thin fault zone dolerite near 162-077 has a distinct ophitic texture affected only by a slight sericitization of the plagioclase and the complete uralitization of the original pyroxene to a pale green, weakly pleochroic amphibole of uncertain fibrous character. The sparse plagioclase phenocrysts (An 57-75) show clear cut complex twinning and delicate oscillation zoning. The lath plagioclase of the groundmass has a composition range of An 65, for cores, to An 29, for rims. Later solutions in the fault zone have altered the dyke in some portions to a fine sericitic mat in which is preserved, in detail, the former mineral shapes and textures.

(4) Amphiboles of Uncertain Origin.

The rocks dealt with here are largely limited to the Saunders Creek area where the high grade of metamorphism, metamorphic complexity, degree of recrystallization and paucity of outcrop contribute to their genetic obscurity. These amphiboles may occur as either:-

a. Dark amphibole-rich schists and gneisses along portions of the marble-schist contact.

b. Poorly exposed finer grained amphibole schists apparently embedded within semi-pelitic schists.

a. Amphibolites at marble-schist contacts.

The marble-schist contacts are normally marked by a thin diopside-scapolite calc-silicate horizon, which in places may become rather amphibolitic. Locally amphiboles bearing a resemblance to a metamorphosed basic igneous rock appear, and while it is possible that they represent a facies variation of the normal calc-silicate rock at the marble-schist contact, it is also possible that they are metamorphosed basic intrusions. The marble-schist contact elsewhere in the Cambrai region is known to be favourable for the structural control of intrusive rocks.

At 210-092 (specimen 103) there are several small exposures of a banded amphibolite on the marble-schist contact. The rock has a uniform appearance throughout, being composed of a light-grey fine-grained felspar-amphibole matrix, through which is regularly dispersed thin discontinuous lensoidal plates, up to 3 cm. in diameter and 3 mm. thick, composed of numerous dark-green amphiboles. The plates, which outline a pronounced foliation plane, appear like regularly spaced tension cracks which have opened and filled with amphibole. Over half of the rock is green amphibole, $2V_x$ 75°(± 3°) and $Z \wedge c$ $18\frac{1}{2}°$, occurring in recrystallized diversely oriented prisms containing small inclusions of sphene, apatite and plagioclase. Rarely twinned anhedral granoblasts of plagioclase, twinned only on the albite law, are strongly sericitized throughout. Pistacitic epidote is a minor constituent, appearing in sparse crystals, though generally well crystallized and equal in size to the plagioclase. Pale green ripidolitic chlorite flakes, also a minor constituent, have resulted from the alteration of original biotite, relics of which are still occasionally found. Sphene is abundant throughout as small pale-brown rounded granules, sometimes with cores of iron ore. The rock thus has an overall mineralogical composition similar to amphibolites derived from normal metadolerites.

At 207-088 (specimen 111) a dark-green, almost black, amphibolite, slightly speckled with white felspar, appears as sparse float and outcrop near a marble-schist contact. The felspar specks, composed of numerous fine granules, outline a weak banding parallel to the moderate foliation formed by the orientation of the amphibole prisms. The amphibole, in clear anhedral to euhedral interpenetrating prisms up to 2 mm. in length, forms 70% of the rock. As in specimen 103 the amphibole is pale in thin section and weakly pleochroic,

X = pale cream

Y = pale yellow-green

Z = pale mustard green.

Other optic properties are $2V_x$ 83° (± 3°) and $Z \wedge c$ 19°. Rare sphene and orthite(?) inclusions are surrounded by green-brown to colourless pleochroic halos. The interstitial matrix consists of close packed nontwinned slightly zoned granular (0.05-0.1 mm.) anorthite ($2V_x$ 78°, $\beta = 1.581$ ($\pm$ .002), An 89-93). The well defined triple point contacts of the anorthite grains suggests a close approach to metamorphic equilibrium. Minor sphene granules about decomposed iron ore complete the assemblage.

A similar dark amphibolite (specimen 70, location 207-082) a few hundred yards to the south is of similar occurrence, but, although moderately foliated, does not have a banded appearance. The amphibole has similar optic properties to the amphibole of specimen 111, but occurs as radiating sheaves rather than single prisms. The plagioclase has been entirely altered to cloudy clinozoizite, which has migrated into fractures in the cloudy amphiboles caused by shearing associated with a nearby major fault.

b. Amphibolites within schist.

At 205-086 (specimen 150) an irregular outcrop of amphibolite is associated with quartz-albite rock (albitized schist). In outcrop large scapolite and diopside crystals are seen in some portions of the rock, while other portions consist of dark fine-grained amphibolite. The amphibolite in slice has a fresh well-recrystallized granoblastic texture. Well-formed hornblende prisms, $2V_x$ 74°, $Z \wedge c$ $16\frac{1}{2}°$, X, pale lemon; Y, mustard green; Z, deep green, form aggregates in a matrix of clear anhedral equidimensional plagioclase grains. The plagioclase has slight reverse zoning and a composition range of An 43-63. The twinning, on the albite and pericline laws, is simple. A small amount of diopside is present within the hornblende aggregates and a small amount of scapolite has begun to replace the plagioclase. Sphene containing specks of iron ore is a common accessory. Biotite is absent. A few irregular patches and veins of andesine (An 40), containing abundant sphene inclusions, cut the amphibolite. The felspar grains are anhedral and rarely twinned (albite laws only). Some portions are being replaced along the grain boundaries by sprawling poikiloblasts of scapolite. These veins are reminiscent of similar andesine veins cutting metadolerites in the southern part of the Cambrai dolerite field.

On a saddle at 207-074 and immediately west of the marble, the boundary of which is believed to mark a major fault, there crops out, for 20 feet across strike, amphibolites of variable character. Six specimens (131 a - f) were taken. The first, 131a, located next to the limestone and the fault, is crushed beyond recognition but appears to have been an anthophyllite schist. The second, 131b, 2 feet west of 131a, also slightly crushed, is a light coloured lineated oligoclase-anthophyllite schist, with minor quartz and accessory rutile, iron ore and apatite. The third, 131c, 4 feet west of 131b, is a medium-grained, grey amphibolite largely composed of pale green amphibole, highly sericitized grano-blastic labradorite, minor white cummingtonite and accessory iron ore and apatite. The plagio-clase is rarely twinned. The clear green amphi-bole is unaltered and is pleochroic X= very pale lemon, Y = pale yellow green and Z = pale mid green. The fourth, 131d, one foot west of 131c, is a light green-grey amphibolite in which trains of iron ore granules and oriented amphibole prisms outline a pronounced lineation. The rock is composed almost entirely of amphiboles, both green and white, and lesser sericitized plagio-clase. Here the cummingtonite is more abundant than the green amphibole, the latter being of similar appearance to that in 131c. Both amphi-boles generally occur as separate crystals, although they may show homoaxial growth. In any case, they both appear to be in stable metamorphic equilibrium. The fifth, 131e, $5\frac{1}{2}$ feet west of 131d, is a light grey amphibolite, which, except for finer grain and stronger lineation, is similar in most respects to 131c. The plagioclase, however, is fresh and displays rare simple twinning and common reverse zoning. The sixth, 131f, $1\frac{1}{2}$ feet west of 131e, is a striking honey brown amphibolite in which glistening prisms of anthophyllite and cummingtonite outline a moderate lineation. Except for the occasional sericitic nuclei in the plagioclase, the rock is entirely fresh. The anthophyllite and cummingtonite form closely associated clear fresh prisms which often interpenetrate and are apparently quite stable. Rarely one amphibole is seen overgrowing a smaller prism of the other with their c-axes parallel. The anthophyllite is faintly pleochroic in pinks and yellows. The granoblastic plagioclase is clear, rarely twinned, and contains numerous small prisms of apatite. Iron ore is again a common accessory. Other mineralogical data are summarized in Table 9. In addition a zoning of some cummingtonites in 131d has been discerned, the cores having $2V_z$ $75\frac{1}{2}^\circ$, $Z \wedge c$ 21°

TABLE 9.

MEASUREMENTS OF OPTICAL PROPERTIES OF MINERALS IN ROCKS 131 b - f.

Plagioclase. Compositions using Emmons 5-axis method.

Rock No.	Cores.	Rims.	Twin laws.
131 b	An 21 (unzoned)		Albite only.
131 c	An 39	An 46	Albite and pericline or acline.
131 d	An 40-44	An 61-69	Albite and rare acline.
131 e	An 43	An 49	Albite and pericline or acline.
131 f	An 37-47	An 44-48	Albite, pericline, Carlsbad, Carlsbad-albite.

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Amphiboles.

Rock No.	Ortho-amphibole	Clino-amphiboles	
		White	Green
131 b	$2V_z$ $81\frac{1}{2}^\circ$	none	none
131 c	none	n.d.	$2V_x$ 83°, $Z \wedge c$ 16°
131 d	none	$2V_z$ 76°, $Z \wedge c$ 21°	$2V_z$ $87\frac{1}{2}^\circ$, $Z \wedge c$ 15°
131 e	none	n.d.	$2V_x$ 83°, $Z \wedge c$ 17°
131 f	$2V_z$ 74°	$2V_z$ 74°, $Z \wedge c$ 23°	none

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(74% Mg molecule) and the more gruneritic rims $2V_z$ $78\frac{1}{2}^o$, $Z \wedge c$ 20^o (65% Mg molecule; Deer, Howie and Zussman, Vol. 2, p.242). This amphibolite is unique: the central portion bears no similarity to any metasediment within the area but has a possible igneous composition; the border rocks however, could hardly be equated with any igneous source, although they are similar to other meta-sedimentary rocks within the area. Although providing considerable mineralogical interest the origin of these rocks must be considered unsolved.

It remains to mention an unusual dark pod of foliated biotite-quartz schist appearing to intrude uniform meta-arkose at 166-110 (specimen 316). The total lack of heavy minerals, and the uniformity of composition throughout, suggests that this might be a highly reconstructed basic rock.

(5) Discussion and conclusions.

Although it is apparent from the foregoing descriptions that metamorphic equilibrium is seldom attained within the metadolerites, certain general conclusions regarding their reaction to metamorphism can be ascertained.

As pointed out by Wiseman (1934, p.397), where the metamorphic equilibrium of a metadolerite has not been completely attained, the extent to which the mineral assemblage has been produced by prograde or retrograde reactions is not always apparent. This is especially true of those more massive metadolerites of the present area that still retain their blastophitic texture. Foliation development within the metadolerites clearly aids nucleation, recrystallization and the achievement of close mineralogical and textural equilibrium. The sensitivity of a dolerite to metamorphism is thus partially dependent on the development of a contemporaneous foliation. Of greater importance, however, is the availability of volatiles, especially H_2O. It has been noted that in some metadolerites intruded into the higher grades of metamorphism there is a remarkable preservation of phenocrysts, although the matrix may be completely recrystallized, while at lower grades the phenocrysts may be more readily destroyed. Such features may be related to the greater availability of volatiles in the lower grades of metamorphism.

Although it is not possible, in the limits of this study, to consider the complex chemical changes involved in the reconstruction of the metadolerites, several general conslusions may be made regarding the most likely mineral reactions occurring. The end-products of metamorphism of a non-porphyritic dolerite are hornblende, plagioclase with minor quartz, biotite, sphene and iron ore. This assemblage has presumably originated from the augite, plagioclase and ilmenitic magnetite of the original dolerite. The uralitization of the pyroxene is apparently an easy reaction, involving little introduction or removal of material and, in which, the plagioclase plays little or no part.

The first stage, then, is the uralitization of the original pyroxene, while the original felspar remains intact. A reaction such as

$$\left[\underset{\substack{\text{pyroxene}\\ \text{(augite)}}}{4\ Ca_{0.75}(Mg,Fe)_{1.25}Si_2O_6} + H_2O \rightleftharpoons \underset{\text{actinolite}}{Ca_2(Mg,Fe)_5Si_8O_{22}(OH)_2} + CaO\right] \quad \ldots\ldots.1$$

may be dominant.

The next stage is the recrystallization of the uralite to a clear hornblende and the beginning of strong normal zoning of the plagioclase laths near their rims, indicating a reduction of anorthite content. At this stage a blastophitic texture is still apparent but the plagioclase laths are losing their external form, the edges being replaced by intergrowing hornblende and sodic plagioclase. A reaction such as

$$\left[\underset{\text{actinolite}}{Ca_2(Mg,Fe)_5Si_8O_{22}(OH)_2} + \underset{\text{anorthite}}{CaAl_2Si_2O_8} \rightleftharpoons \underset{\text{hornblende}}{Ca_2(Mg,Fe)_4(Al,Fe)(Si_7AlO_{22})(OH)_2} + CaO + 3SiO_2 + (Mg,Fe)O \right] \quad \ldots\ldots\ldots 2$$

is indicated.

Shido (1958) has shown that in the low pressure metamorphism of the Abukuma Plateau, Japan, the hornblende is relatively rich in the edenite molecule. She suggests the reaction

$$\left[\underset{\substack{\text{actinolite} \\ \text{271 cc.}}}{Ca_2Mg_5Si_8O_{22}(OH)_2} + \underset{\substack{\text{albite} \\ \text{102 cc.}}}{NaAlSi_3O_8} \rightleftharpoons \underset{\substack{\text{edenite} \\ \text{267 cc.}}}{NaCa_2Mg_5AlSi_7O_{22}(OH)_2} + \underset{\substack{\text{quartz} \\ \text{90 cc.}}}{4\ SiO_2} \right] \ldots\ldots\ldots 3$$

which involves a reduction of volume of more than 4%. Thus both the albite and anorthite molecules of the plagioclase may be used in the formation of the hornblende.

These reactions release CaO, SiO_2 and (Mg,Fe)O. Some SiO_2 crystallizes as quartz. The other oxides are most likely used up in the subsidiary reactions

$$\left[\underset{\substack{\text{ilmenite} \\ \text{molecule in} \\ \text{iron ore}}}{FeTiO_3} + CaO + SiO_2 \rightleftharpoons \underset{\substack{\text{sphene} \\ \text{with remaining} \\ \text{magnetite}}}{CaTiSiO_5 + FeO} \right] \ldots\ldots 4$$

and

$$\left[\underset{\text{"orthoclase"}}{2KAlSi_3O_8} + 6(Mg,Fe)O + H_2O \rightleftharpoons \underset{\text{phlogopite}}{K_2(Mg,Fe)_6(Si_6Al_2O_{20})(OH)_2} \right] \ldots\ldots.5$$

In the latter reaction the orthoclase molecule may be in solid solution with the plagioclase, and the reaction would account for the formation of zones of small biotite flakes near the edges of some large phenocrysts (see also Wiseman, 1934, p.385).

The transient formation of white mica and epidote during the recrystallization of basic phenocrysts has already been noted. Ramberg (1949) suggested the possible reaction

$$\left[\underset{\text{anorthite}}{4\ CaAl_2Si_2O_8} + \underset{\text{orthoclase}}{KAlSi_3O_8} + 2H_2O \rightleftharpoons \underset{\text{muscovite}}{KAl_3Si_3O_8(OH)_2} + \underset{\text{clinozoizite}}{Ca_2Si_3Al_3O_{12}(OH)} + \underset{\text{quartz}}{2\ SiO_2} \right] \ldots\ldots\ldots6$$

If there is sufficient orthoclase molecule in the plagioclase this reaction would account for the observed products. This saussuritization is always a biproduct when the phenocrysts are being replaced by scapolite. The association might suggest that the epidote and white mica were produced in a scapolite forming reaction. Although this might be at least in part true for the epidote, in view of the small amount of epidote and white mica produced and their transient nature, it is more likely that the introduction of volatiles during the scapolitization reaction provided H_2O to enable equation 6 to become effective. With complete metamorphic reconstruction the epidote and white mica disappear, presumably aiding the calcification of the remaining metamorphic plagioclase and the formation of more biotite.

In the low grade metadolerites, and in some metadolerites enclosed in marble, chlorite has appeared. A reaction such as

$$\Big[\underset{\text{tschermakite}}{5Ca_2(Mg,Fe)_5Al_4Si_6O_{22}(OH)_2} + 4\,H_2O \rightleftharpoons$$

$$\underset{\text{actinolite}}{Ca_2(Mg,Fe)_5Si_8O_{22}(OH)_2} + \underset{\text{chlorite}}{2Mg_5Al_2Si_3O_{10}(OH)_8} +$$

$$8CaAl_2Si_2O_8\Big] \quad \text{(Shido, 1958)} \quad ....7$$

would be promoted at lower temperatures of higher water vapour pressures.

Throughout the medium grades of metamorphism, the usual assemblage in metadolerites is amphibole and plagioclase with minor quartz, biotite, sphene, iron ore and apatite. Progressive changes may occur in the composition of the amphibole and plagioclase with increasing metamorphic grade. In the amphibole there may be a marked change in the Z-axial colour from clear blue-green to deep blue-green to deep green to brownish green to brown (Wiseman, 1934; Miyashiro, 1958; Shido, 1958; Layton, 1963). In Japan this has been used as a convenient index of metamorphic grade. In lower pressure metamorphism the brown colour may appear earlier, and in the contact zones of the Iritono District, Japan, the blue-green amphibole is entirely absent (Shido, 1958). Shido has found that the amphiboles increase in Al, Ti and alkali content with grade.

In the present area the amphiboles in the lowest grade chlorite bearing metadolerites have a Z-axial colour of pale blue-green. In the andalusite-staurolite-garnet zone the colour is deep green with a blue tinge. In the slightly higher pressure andalusite-staurolite-kyanite-sillimanite zone the colour is a deep blue-green and in the sillimanite zone a deep green. There is some variability of colour in the sillimanite zone, and in the amphibolites of the highest grade, the amphiboles are much paler in colour and show shades of yellow-green, mustard-green and brownish-green. If the hornblende-anorthite amphibolites are metadolerites then the amphibole must have absorbed most of the albite molecule from the plagioclase.

The increase of the anorthite content of the plagioclase with metamorphic grade has been emphasized by many writers (e.g. Wiseman, 1934; Shido, 1958).

The major changes occur at the lower grades where epidote is present. Once the epidote has been destroyed and the plagioclase reaches an andesine or sodic labradorite composition, the changes are much less spectacular. In the metadolerites of the Cambrai area it is not easy to access changes in the plagioclase composition with grade because

(a) the metadolerites show a grade of metamorphism higher than that at which epidote is stable, and the composition changes with the higher grades will be small,

(b) the plagioclase tends to preserve its original form, and in the absence of mechanical shearing is rather sluggish in attaining an equilibirum composition,

(c) the bulk composition of most of the metadolerites is variable because of the varying proportion of basic plagioclase phenocrysts.

In several well recrystallized metadolerites in the western group the plagioclase has almost attained equilibrium, and there is a steady increase of anorthite content with grade from An 35 in the andalusite-staurolite zone, to An 45 in the sillimanite zone (see Fig. 54, Chapter 8). In the absence of epidote these changes must be related to changes in the composition of the hornblende (e.g. increasing edenite molecule). It is significant that over the same interval of metamorphic grade the adjacent epidote bearing calc-silicates show plagioclase composition changes from An 26 to An 36. This suggests that epidote formation has been "held back" in the metadolerites by the apparent stability of the slightly more calcic plagioclase.

The absence of garnet and epidote as stable constituents in the Cambrai metadolerites is in marked contrast to their ubiquitous occurrence in the Southwest Highlands (Wiseman, 1934). In this respect the Cambrai metadolerites are similar to Wiseman's "abnormal" epidiorites of Banffshire and Aberdeenshire. These epidiorites, characterized by the absence of chlorite, epidote and garnet, have been described by Read (1923, p. 90-101 and Sutton and Watson, 1951). Shido (1958), describing basic metavolcanics of the Abukuma Plateau, attributes the absence of garnet and epidote in the middle grades of metamorphism there, to lower solid pressures during the metamorphism. This conclusion is also applicable to the present area, and adds further weight to the conclusion made from the study of the metasediments, that the metamorphism is of a low pressure intermediate type. In the higher pressure kyanite-sillimanite metamorphism of the Williamstown district, 12 miles west of the present area, the author has noted sparse red garnets in strongly recrystallized metadolerites (Mills, 1963, p.175).

From the literature it would seem that biotite is stable in small amounts in most epidiorites. Wiseman has noted that biotite is absent from the epidiorites in the kyanite and sillimanite zones of the Dalradians. Biotite appears to be stable in some metadolerites of the sillimanite zone at Cambrai, but is absent from the most recrystallized amphibolites. Dolerites carry a small amount of potash; at lower grades this apparently forms its own phase, biotite, while at higher grades it enters into solid solution with the plagioclase and hornblende.

The appearance of diopside within one high grade amphibolite at Cambrai is not inconsistent with Wiseman's observation of sparse diopside in the sillimanite zone.

At Cambrai, cummingtonite occurs in intimate association with green hornblende in some metadolerites of the sillimanite zone. It has also been observed in association with green hornblende and garnet from a highly altered metadolerite in marble of staurolite-kyanite-sillimanite grade. This observation that cummingtonite in basic schists is limited to the sillimanite zone is in agreement with the findings of Shido (1958), who also describes the occurrence of cummingtonite in relation to types of metamorphism. In basic rocks the cummingtonite is apparently limited to the lower pressure types of metamorphism (in the higher pressure types it may be replaced by orthopyroxene). Shido considers that the cummingtonite molecule may be in solid solution with the hornblende at low grades, but at higher grades may exsolve or form by a reaction such as

$$\left[\underset{\substack{\text{tschermakite} \\ 1820\text{ cc.}}}{7Ca_2Mg_3Al_4Si_6O_{22}(OH)_2} + \underset{\substack{\text{quartz} \\ 224\text{ cc,}}}{10\ SiO_2} \rightleftharpoons \underset{\substack{\text{anorthite} \\ 1414\text{ cc.}}}{14\ CaAl_2Si_2O_8} + \underset{\substack{\text{cummingtonite} \\ 807\text{ cc.}}}{3\ Mg_7Si_8O_{22}(OH)_2} + 4\ H_2O\right] \qquad \ldots\ldots\ldots 8$$

The cummingtonite-hornblende association in amphibolites has also been discussed by Eskola (1950) and Watters (1959).

The cummingtonite-anthophyllite assemblage in the rock 131f is also of some interest. Unfortunately the origin of this rock is uncertain. A cummingtonite-anthophyllite assemblage in an amphibolite from Isopaa, Finland has been described by Eskola and Kervinen (1936), and the problems of the cummingtonite-anthophyllite association have been discussed by Layton and Phillips (1962).

(6) Comparison with other metadolerites in the Mt. Lofty Ranges.

Mawson (1926) mentions the occurrence of petrologically old looking basic dykes in the Milendella region, immediately south of the present area.

Alderman (1931) has described porphyritic metadolerites from the Woodside district, 20 miles south-west of the Cambrai area. Some of these have scapolite replacing the plagioclase, and are similar in most respects to the metadolerites of the Cambrai dolerite field. More recent mapping by the State Geological Survey has indicated the occurrence of a swarm of cross-cutting dykes in the Woodside area.

Chinner (1956) describes metadolerites from the Tanunda Creek area, 12 miles to the north-west. There are many similarities of these rocks to those of the Cambrai area. Some carry plagioclase phenocrysts and preserve a blastophitic texture, while others have become lineated schistose amphibolites. Cores of relatively unaltered dolerite occur in some masses, and relic olivine was noted in one instance.

White (1956) has described many similar and wide-spread occurrences from the Mannum Sheet. The largest masses have unaltered dolerite cores and foliated, lineated margins. Preservation of a blastophitic texture is common, and both plagioclase and augite phenocrysts may occur. Scapolite, however, was rarely observed, even when there were scapolite bearing calc-silicates nearby.

The well recrystallized foliated and lineated amphibolite dykes of the Williamstown district (Mills, 1963) are similar to the more foliated metadolerites of the sillimanite grade in the Cambrai area.

Metamorphosed dolerite intrusions are therefore common throughout the whole of the metamorphosed portion of the Mt. Lofty Ranges, and except for the rare garnets in the Williamstown examples, garnet and epidote are not described as stable minerals. These metadolerites must therefore be classed with the Banffshire and Aberdeenshire types rather than with those of Perthshire.

(b) The granodiorite-aplogranite-aplite-pegmatite suite.

(1) Granodiorites.

a. Field characteristics.

Most granodiorites in the Cambrai area may be classified as belonging to either the

a. the granodiorite belt of the south-west corner of the mapped area, or

b. the Kanappa Mine region.

1. The granodiorite belt of the south-west.

In the south-west belt intrusions of biotite granodiorite are very numerous; hundreds of small bodies ranging from small pods a few feet in length, to irregular shaped intrusions up to one mile in length, have been mapped (Plate 1). The shape and distribution of the intrusions is governed by the tectonic trend of the high grade gneisses into which they have been emplaced. The predominant forms of intrusion are lenses, dykes, plugs and irregular "jelly-fish" shaped masses. These intrusion forms have enabled the injection of a large amount of granodioritic magma, in some parts 20-30% of the total rock volume, without appreciable disturbance of the tectonic trend of the surrounding gneisses. The distribution of the granodiorite bodies is of some interest; in particular, the sharp eastern boundary of the intrusive belt. This boundary also reflects lithological changes in the metasediments, the arkoses to the east of the boundary being more massive, less micaceous, less recrystallized and finer grained than the micaceous gneisses, veined gneisses and migmatites to the west. There is, however, no evidence of a tectonic disruption at this boundary.

The granodiorite-schist contacts are generally quite sharp, and, owing to the difference in grain-size and texture of the two rocks and the presence of lamination (relic bedding) in the schists, the contacts are readily discernible. There is no recognisable change in the grain-size of the quartzo-felspathic schists as the granodiorite contacts are approached. Where assimilation has occurred the contacts may be more gradational in detail. The contaminated rocks are distinguishable by their irregular grain-size and blotchy appearance. Metasediment xenoliths, showing varying degrees of diversion from their original orientations, are not uncommon, especially near the contacts of the intrusions. The xenoliths, having a composition not greatly different from that of the granodiorite, are soon assimilated. In many cases, however, small disoriented clots of biotite remain. These are sparsely distributed and are more abundant in the more highly contaminated portions of the granodiorites. They do not seem to represent remnants of basic xenoliths, and were apparently grown during the assimilation processes.

The biotite of these basic clots is many times the grain-size of the biotite in the quartzo-felspathic schists being assimilated, and is more comparable to the biotite within the granodiorite. In the assimilation of the more migmatized gneisses, the basic biotite rich bands are less readily assimilated, and may persist near the borders of the body as streaky xenoliths and schlieren. Rarely a minor layering has been noted near the granodiorite contacts, where assimilation was presumably accompanied by movements within the magma.

Within the country rocks near the granodiorites the most notable features are veined injection and distortion of the original foliation structures. Cross-cutting granite, aplite, pegmatite and quartz veins are common at some granodiorite contacts. These veins often merge into the granodiorite, suggesting that they represent the last phase ejected by the crystallizing granodiorite magma. Forceful injection of the granodiorites is often indicated by the irregular plastic folding and distortion of the foliation structures within the gneisses at the granodiorite contacts. The presence of these plastic structures in place of brecciation indicates that these metasediments were at a high temperature during the intrusion. This is also confirmed by the lack of contact metamorphic and alteration effects and the imposition of a penetrative foliation (gneissosity) on the granodiorites. This moderate to faint foliation is seldom strong enough to be measured in outcrop, but is quite obvious as a moderate biotite orientation in hand specimen and thin section. Kleeman (1954, p. 16) has tested the orientation of the biotite in one of these granodiorites and has found that a significant biotite orientation exists. Rarely the foliation becomes almost as strong as that in the Rathgen gneiss (a paragneiss, White, 1956) outcropping a few miles to the west. It must be concluded that the granodiorites were intruded while the metasediments were still at a high temperature, probably soon after the climax of the metamorphism and in the waning stages of the tectonic activity.

In the field two distinct types of

granodiorites may be found, one a fine-grained variety, the other medium grained, but both of similar mineralogy. Intrusions of both varieties are intimately mixed but are not seen to grade from one to the other within the same body. Rarely granodiorites with intermediate grain sizes occur. The finer grained varieties might be supposed to be of younger age, and the more dyke-like character of some intrusions of this variety is considered in support of this.

Within any one body, apart from the assimilative contact effects, the granodiorite is very uniform in both texture and mineralogy. Porphyritic varieties are absent, although in some of the coarser rocks the rectangular form of some of the larger plagioclases may be distinct. In the vicinity of 153.107 the metasediments have been granitized to a fine-grained rock barely distinguishable from the fine-grained intrusive granodiorite also present.

2. Granodiorites of the Kanappa Mine region.

The granodiorites of the Kanappa Mine area are similar to the finer grained variety of the south-west belt, and generally occur as dykes parallel to the trend of the main north-south foliation. These dykes may vary in thickness from 1-40 feet and range up to 1000 feet in length (Plate 1). The contacts with the metasediments are again sharp, and xenoliths are not uncommon. One of these fine-grained granodiorites has intruded marble, the marble-granodiorite contact being marked by an epidote skarn. The only exception to the fine-grained dyke-like granodiorites of the Kanappa region is a medium-grained tonalite-diorite (location 214.109). This rock is richer in biotite and poorer in quartz than the medium-grained granodiorites of the south-west belt, and appears to be a more basic variant of the granodiorite suite.

b. Petrology of the Cambrai granodiorites.

Mineralogically the granodiorite suite is relatively simple and uniform throughout. The major minerals are plagioclase, quartz, potash felspar and biotite. Muscovite, opaques, apatite, sphene, zircon, epidote, monazite and tourmaline form only accessory proportions. Muscovite (recrystallized from sericite), chlorite, and some epidote and sphene occur as late to post-magmatic products of alteration of plagioclase and biotite.

Texturally the rocks are uniform within any one intrusion, but may vary from one intrusion to another in both grain-size and degree of foliation development. Reference to finer and coarser grained varieties has already been made. The finer variety has a grain size averaging 0.05-1 mm. and the coarser 1-4 mm. Rarely the biotite flakes may reach 4-5 mm. in width in the coarser rocks. Apart from the grain size variation there is no consistent petrological differences between the finer and coarser grained varieties.

In the least foliated varieties the original hypidiomorphic texture is well preserved. Subhedral plagioclases with complex twins, delicate oscillation and strong normal zoning, and irregular plates of almost randomly oriented biotite, are surrounded by anhedral quartz and interstitial potash felspar. With the imposition of the foliation the biotites become more aligned, the plagioclases become bent, and fine close-spaced deformation twins appear, the quartz becomes highly strained and starts to fracture into superindividuals along bands of pronounced undulose extinction, and the rock begins to lose its hypidiomorphic texture and takes up a more crystalloblastic one. A further stage is the development of distinct bands along which the crystalloblastic texture is more complete, and in which the biotite is strongly aligned. At this stage the rock may take on a distinct lineation defined by the biotite flakes. The stage is not reached where the recrystallisation is so complete that the original igneous texture of the rock cannot be recognised.

The major mineral of the granodiorites is plagioclase, forming 35 - 60% of the rock. In both the fine and coarse grained rocks the composition of the plagioclase is consistently An 25 for non-zoned grains (in one case An 15, specimen 423, location 219.119). In the more zoned plagioclases the cores range from An 25 - An 42 and the rims are consistently An 20. The normal zoning becomes strongest at the boundary of the rim and the core of the grains. The largest part of the grain, the core, may have delicate oscillation zoning which preserves the euhedral outlines of the plagioclase crystals at various stages of growth (see also Leedal, 1952). In the cores twinning is well preserved, but commonly fades out at the oligoclase rims. Many common twin laws are represented; albite, Carlsbad, pericline, albite-Carlsbad, Ala B, albite-Ala B, and

Manebach were observed. In the foliated rocks strain is expressed in the plagioclase by crystal bending and the development of abundant fine albite and pericline twins (Vance, 1961). The plagioclase cores are generally sericitized, and the sericite may recrystallise to ragged crystals of muscovite.

Quartz constantly forms about 35% of the rock. The grains are always anhedral and may carry many small euhedral prisms of apatite and zircon and rare tourmaline. In deformation the quartz becomes undulose, and eventually fractures into strips along the deformation bands.

Potash felspar is a variable constituent; it may be absent or it may form up to 20% of the total rock. Cross-hatched twinning is usually present, especially in the more deformed granodiorites (see also Harker, 1954). Exsolution lamellae are lacking, but myrmekite is common at the plagioclase grain boundaries. The myrmekite consists of sodic plagioclase (generally oligoclase) containing small worms or graphic blebs of quartz. The origin of myrmekite has been debated by many workers. Three main theories of the origin of myrmekite have been proposed:

(a) A bi-product of the replacement of potash felspar by plagioclase (e.g. Becke, 1908; Sederholm, 1916; Leedal, 1952).

(b) The last drops of liquid to crystallise during the cotectic crystallisation of quartz and plagioclase. (e.g. Rogers, 1961).

(c) A product of exsolution during the slow cooling of potash felspar (e.g. Spencer, 1945; Shelley, 1964; Carman and Tuttle, 1963, have given some experimental evidence for this).

The exsolution theory is favoured by the present writer. The formation of a myrmekite-like structure in steels, where austenite solid solution exsolves ferrite and cementite, is well known to metallurgists (Azaroff, 1960, p.200). Further experimental work on the mechanism of myrmekite formation is necessary before any of the above theories can be discarded.

Biotite, in aggregates of flakes, may comprise up to 15% of the normal granodiorites. It is a highly coloured variety pleochroic from pale lemon yellow to dark brown black; some biotites from the Kanappa granodiorites are more reddish brown in colour. Inclusions of zircon and monazite, surrounded by pleochroic halos are not uncommon.

Apatite and epidote may be occluded near the edge of the grains.

In some rocks the biotites have altered to chlorite (pleochroic pale lemon to green), muscovite and minor sphene. Where the biotite was originally red in colour the derived chlorite is bright green. With the biotite in specimen 412 (location 220.123), an amphibole pleochroic from pale lemon to mid blue-green is a major component. In this same specimen sphene is present as large lozenge shaped crystals, whereas in the normal granodiorites it usually has a more granular habit.

Strongly birefringent and pleochroic epidote is usually present in minor amounts occluded in the biotite, where it appears to be a primary mineral phase. Shido (1958) has observed primary epidote in the granodiorites of the Tabito Igneous Complex, Nakoso, Japan. She considers that the epidote may be in equilibrium with the plagioclase (An 31-40), a high P_{H2O} in the crystallising rock promoting the stability of epidote at a higher temperature than is normal in a metamorphic rock. Zircon occurs as minute euhedral prisms; occasional larger rounded zircons

FIG. 62.

Exposed surface of medium-grained granodiorite showing slightly directed texture outlined by biotite flakes. Location 156.125. Coin diameter 2.8 cm.

FIG. 63.

Photomicrograph of biotite-tonalite.

Rock A185-463. Location 214.109. Crossed polars.

Mostly biotite and plagioclase (well zoned and twinned) with lesser quartz. Epidote is common as inclusions in the biotite.

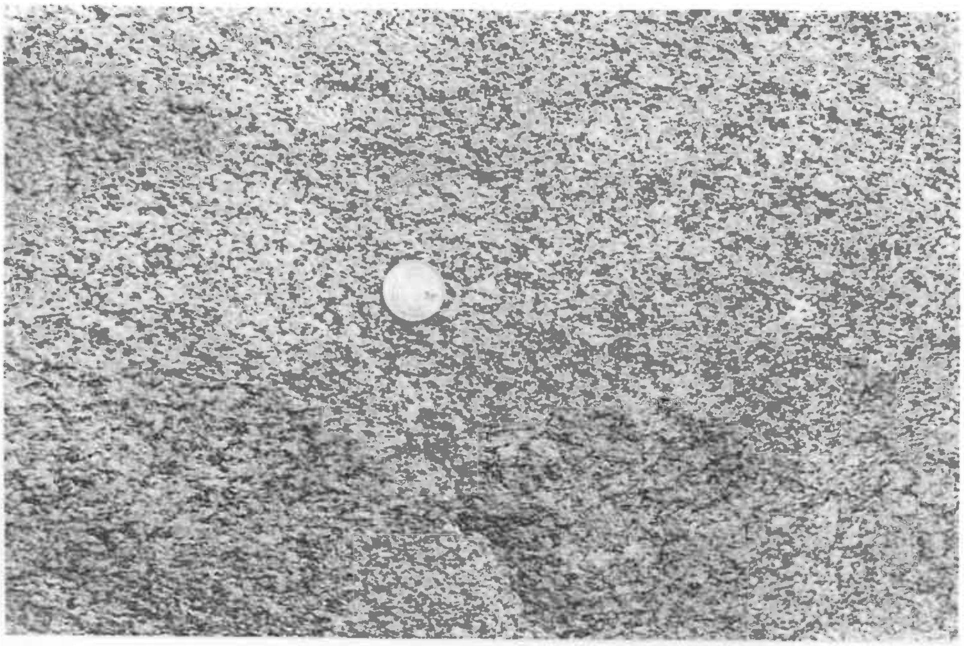


FIG. 62

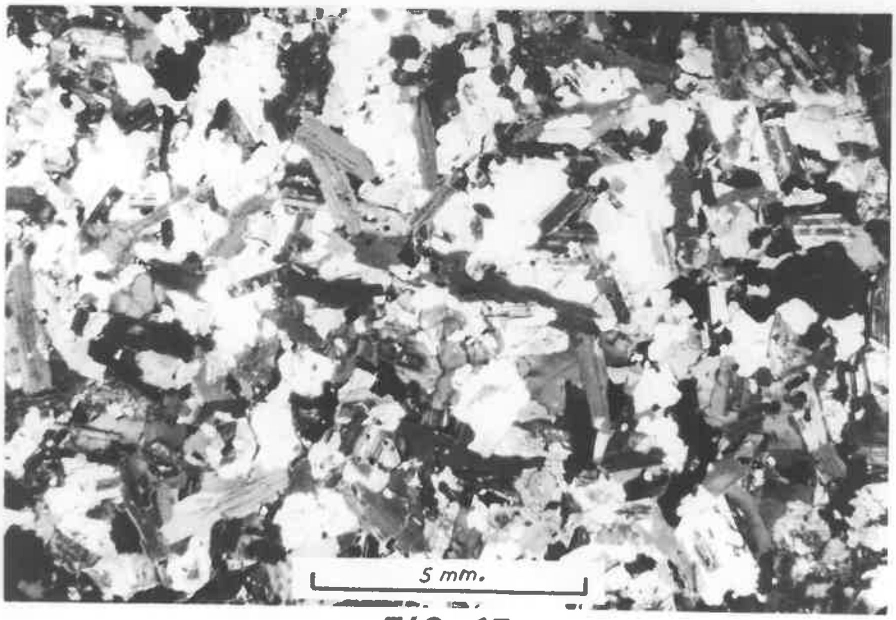


FIG. 63

may occur.

A somewhat more basic member of the granodiorite series is represented by the unique rock outcropping at 214.109 (spec. 463). This rock is a tonalite approaching a dioritic composition, potash felspar and quartz are rare, while plagioclase and biotite are abundant. The composition of the plagioclase is An 38 (An 58, cores, to An 33, rims, for zoned crystals), which is considerably more calcic than in the other granodiorites.

It has been convenient to use the name granodiorite as a field term for these rocks. Since the average grain size is generally less than 2 mm., the quartz content is greater than 10% and the potash felspar content ranges from 0 - 30% of the total felspar, the rocks are more rightly termed microgranodiorites and microtonalites. Three micrometric analyses of the granodiorite rocks from the Kanappa Mine region are presented in Table 10. The first two are typical of the fine-grained varieties, and the third is the unusual rock, 463, which approaches a dioritic composition. Granodiorites of the western area are similar in composition to those of the first two analyses.

TABLE 10.

MODAL COMPOSITIONS* OF GRANODIORITES OF THE KANAPPA MINE REGION.

Field Name	1. Fine-grained granodiorite	2. Fine-grained granodiorite	3. Medium-grained tonalite
Rock No.	439	423	463
Location	216.117	219.119	214.109
No. of points counted	2000	2018	3657
Plagioclase	48.6	44.7	58.8
Potash felspar	1.4	10.6	1.6
Quartz	38.1	35.1	13.8
Biotite	11.5	8.1	23.5
Muscovite	0.1	1.5	0.8
Opaques	0.3	-	0.3
Apatite	-	-	1.0
Chlorite	-	-	0.2
	100.0	100.0	100.0
B.M.R. Name	Micro-tonalite	Micro-granodiorite	Micro-tonalite approaching micro-diorite

"Micro" signifies grain-size 1-5 mm.

* Not calculated to weight %.

c. Further evidence that the granodiorites are magmatic.

Introduction.

In an attempt to supplement the evidence from field observation which suggests that the granodiorites are intrusive rocks, heavy mineral analyses were conducted on specimens of a normal medium grained granodiorite (location 166.086) and a normal quartzo-felspathic schist near the granodiorite contact. The granodiorite contacts here are sharp, and the schist adjacent to the granodiorite contact shows some plastic distortion. The specimens were collected three feet apart, and aligned parallel to the north-south foliation of the gneiss and the faint gneissosity of the granodiorite, and were of sufficient bulk (10-14 lb.) to be representative of the heavy mineral content.

1. Experimental Procedure.

The specimens were deliberately chosen in a slightly weathered and crumbly condition to enable them to be readily reduced to sand on mild crushing in a steel mortar. By repeated sieving and crushing the rocks were reduced to their individual grains. The finest fraction (approximately 1/10th of the crushed material) was then sieved off, and the coarser material discarded. In the writer's experience with schists, gneisses and granites, the accessory mineral fraction is almost entirely confined to this finest fraction. After elutriation in water to remove clay sized particles, the fine sand was warmed in dilute HCl. This treatment removes iron oxide coatings from the grains and dissolved the apatite (but has no effect on the rare-earth phosphates; monazite and xenotime). By repeated vibration and swilling in a small dish, and using a pipette adapted for the subaqueous removal of grain fractions, an almost pure heavy mineral concentrate was obtained. This method has been found most effective in the removal of heavy minerals from sands; the panning process being repeated until the yield of heavy minerals became negligible. The final purification of the accessory fraction consisted of floating off the remaining light minerals using methylene iodide. For convenience of study the accessory mineral fraction was subdivided using a Franz Isodynamic Magnetic Separator.

2. Results.

The weight percent of the accessory minerals extracted from the two rocks were:

granodiorite	0.02%
quartzo-felspathic schist	0.20%

Using the method described above these figures are not likely to be more than 20% smaller than the actual accessory mineral content of the specimens. Thus there is ten times as much accessory mineral in the schist as in the granodiorite.

The results of the heavy mineral analyses are presented below.

	Granodiorite	Quartzo-felspathic schist
Monazite	59.2	53.1
Opaques	17.2	12.6
Zircon	15.1	33.6
Epidote	6.8	-
Xenotime	1.3	-
Tourmaline	0.3	-
Rutile	0.1	0.7
	100.0	100.0

Ignoring the epidote the results of these analyses are not greatly dissimilar. The most notable feature is the abundance of monazite in both rocks.

However there is a considerable difference between the morphology of the accessory minerals in these two rocks, in particular the zircon. Fig. 64 presents camera lucida drawings of typical zircon populations from the granodiorite and the schist. There are notable differences in both the size and morphology of the grains. Zircons from the granodiorite are in the form of small zoned sharp doubly terminated prismatic euhedra, while those of the schist are large, anhedral, angular to rounded grains. The few more rounded zircons in the granodiorite may be xenocrysts (see also Baker, 1942; and Verspyk, 1961). The zircons in both rocks are clear and colourless, but some dark metamict malacons occur in the schist. Grain overgrowths were not observed in the schist zircons but were not uncommon in the granodiorite zircons.

The monazite in both rocks is citron yellow to honey yellow, with a small percentage of semi-opaque, more magnetic, metamict grains containing small brown blebs. The granodiorite monazite is much larger (0.2 mm.) than that of the associated zircons, the grains

FIG. 64c

Representative zircon populations from granodiorite (top) and meta-arkose (bottom). Location of specimens 166.086. For discussion see text.

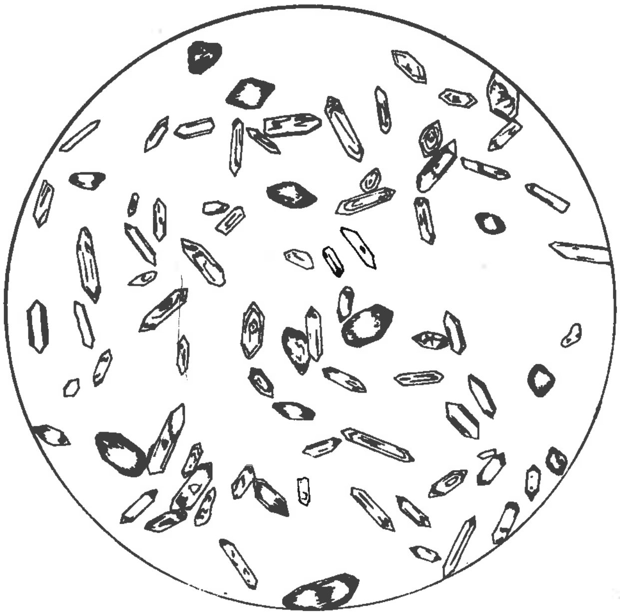
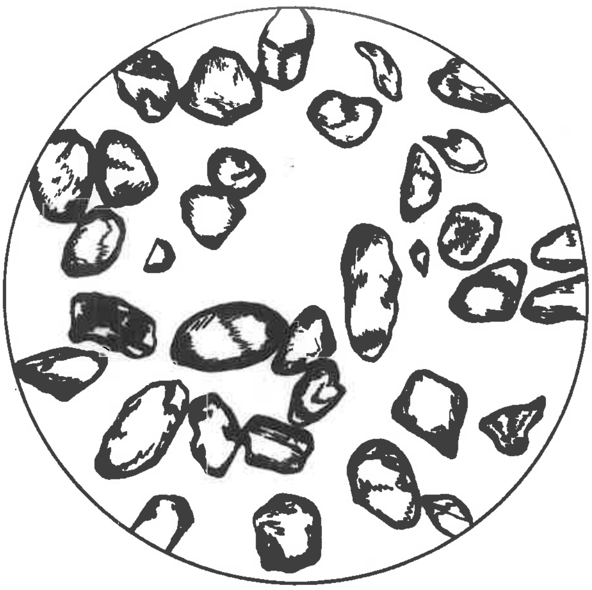


FIG 64

1 mm.



have distinct crystal faces, inclusions of small round to euhedral zircons and rare lamellar twinning. The schist monazite grains are smaller (0.05 mm.) and well-rounded.

Small grains and prisms of pale yellow to orange yellow rutile is present in both rocks. The epidote of the granodiorite is a distinct yellow-green pistacitic variety, the xenotime is distinguished from the zircon by its slight magnetism and lower relief, and the tourmaline is pleochroic pale pink to grey-pink.

3. Discussion and conclusions.

Derby (1891) was the first to recommend the use of zircon in the determination of rock origin, and later, Trueman (1912) advocated the use of zircon in the determination of the origin of gneisses. He also suggested monazite, but noted that in some metamorphics the monazite had recrystallised. In using zircon as a criterion of rock origin three basic assumptions were made:

(a) the zircons of igneous rocks are euhedral,

(b) zircons become rounded during erosional transport,

(c) zircons do not recrystallise in metamorphism.

Armstrong (1922), having found sausage-shaped zircons and secondary overgrowths on zircons in a high grade gneiss, considered that deformation might affect the zircon morphology and discouraged the use of zircon shape as a criterion of rock origin. Graves (1927), described euhedral and corroded zircons in various granitic rocks from the Channel Islands, but did not consider the origin of these rocks in relation to the zircon. Tyler (1940), had some success with zircon as a criterion of origin and provenance in certain rocks including ores. In 1949 Poldervaart and Backstrom presented evidence that under ultra-metamorphic conditions zircon could recrystallise, and they described various fused crystals and crystal outgrowths (see also Poldervaart and Eckelmann, 1955) on zircons from gneiss domes in Kakamas (Cape Province). Poldervaart (1950) advocated the use of zircon as a criterion of granitization on the basis that zircons in granitized rocks are rounded (or may have overgrowths or outgrowths) and had an elongation of less than 2, compared with the zircons of intrusive granites which were usually terminated and had a crystal elongation greater than 2. This method has been adopted with some success by several workers (Vitanage, 1957; and Verspyk, 1961). Verspyk (1961), studying the metamorphic and intrusive rocks of the Aston and Hospitalet massifs, Central Pyrenees, has found a distinct difference between the habit of zircons of the intrusive granodiorites, and the gneisses and migmatites. He considered zircon useful in the determination of the origin of veined migmatites. Murthy and Siddique (1964) found no perceptible recrystallisation of zircons in the garnet-sillimanite gneisses of Orissa and Andhra, Pradesh, India, thus indicating that even under granulite conditions zircon may not recrystallise.

It is now well established that the zircons of intrusive granites are euhedral in form, this being attributed to zircon being one of the first solid phases to crystallise from the granitic melt (Poldervaart, 1956). Larsen and Poldervaart (1957) have placed zircon morphology study on a more quantitative footing using statistical evaluation of the reduced major axes of the zircon crystals from igneous rocks, assuming that the zircon morphology is governed by the physico-chemical environment of the melt. Alper and Poldervaart (1957) have had some success in establishing the consanguinity of various phases of the Animas stock, New Mexico using these statistical methods, and they conclude "zircon studies . . . . can establish beyond doubt that a granitic rock is of magmatic origin."

Evidence from the Cambrai granodiorites establishes that these are magmatic in origin; the zircon morphology of the granodiorite and schist is quite dissimilar. There is no evidence from the zircon morphology in favour of the granodiorites being granitized sediments. However the similarity of the heavy mineral assemblages of the granodiorite and the schist, and the high ratio of monazite to zircon in both rocks (4:1 for the granodiorite and 5:3 for the schist), are features which suggest that the granodiorites might have resulted from the complete melting of sedimentary materials at greater depths.

The sedimentary, metamorphic and intrusive rocks of the Mt. Lofty Ranges are characterised by a high proportion of monazite in their accessory mineral fractions. Monazite bearing pegmatites have been reported from the older Precambrian rocks in the core of the ranges (Thomas, 1924; Webb, 1954) and in the younger pegmatites intruding the Adelaide Super-group (Wilson, 1945). The writer has reported monazite in pegmatites cutting Torrens Group rocks in the Gumeracha and Williamstown district, and has found that monazite is one of the most common accessory minerals in the schists and gneisses in the Torrens Group near Williamstown (although in arkoses and sandstones zircon is by far the major accessory mineral), (Mills, 1963). Fander (1961), in a regional study of accessory minerals in some South Australian granitic rocks reported the presence of monazite in some granitic rocks of the Mt. Lofty Ranges (Victor Harbour-Port Elliot granite, Murray Bridge granite (intrusives) and the Monarto Granite (paragneiss)). Monazite has also been reported commonly in rocks of the Olary Province (Campana & King, 1958).

The abundance of monazite in rocks of the Mt. Lofty-Olary Arc is in marked contrast to its absence in the granite and granodiorite intrusives of Victoria. In a detailed study of 109 occurrences of granitic rocks in Victoria Baker (1942) reported no monazite, although apatite was abundant. The Mt. Lofty Ranges may therefore be considered as a "monazite rich" province.

Kurovets (1963), after studying the accessory minerals of the Adamovsk igneous complex, Southern Urals has introduced a classification of granites into:

(a) apatite granites (supersaturated in aluminium and rich in pneumatolytic minerals, beryl, fluorite, molybdenite etc.),

(b) zircon granites (rich in rare-earths and thorium).

The granodiorite-aplite suite at Cambrai fits well into his "zircon granite" category, but it may be noted that rocks of the Victor Harbour-Murray Bridge suite (containing both rare-earth and pneumatolytic minerals) and the Victorian granodiorite suite (apatite abundant but no monazite or pneumatolytic minerals) do not fit into his classification at all.

d. Conclusions.

The granodiorites are more rightly termed biotite micro-granodiorites and biotite micro-tonalites. Several lines of evidence confirm a magmatic origin;

(1) the zircons of the granodiorites are euhedral and quite different in their morphology to those of the adjacent metasediments,

(2) the textures are relic hypidiomorphic, and igneous features such as fine oscillation zoning is preserved in the plagioclase, and

(3) the field observations of disoriented xenoliths and plastic distortion of the schists at the contacts support forceful injection of a magma.

The presence of a slight foliation, the absence of contact effects (except in the limestone) and the limitation of the intrusions to sillimanite grade metamorphism, places these granodiorites with Buddington's (1959) "mesozone" granites.

It is possible that these rocks were formed by the injection of magma derived from the anatexis of sedimentary material at greater depths, or it may be that they are related to the bathylithic mass of tonalite occurring east of Palmer.

(2) Aplogranites, aplites and pegmatites.

a. Field Characteristics.

Transgressive veins of aplogranite, aplite and pegmatite are closely associated with the granodiorite intrusions of the south-west belt. All gradations from granodiorite, to leucogranite, to aplite, to aplopegmatite, to pegmatite, to quartz veins may be observed here, and it is clear that these rocks form a differentiation series whose parent is the granodiorite. The swarms of aplite dykes in the Saunders Creek-Kanappa area cannot be so directly related to a granodiorite parent, although it is likely that these aplites, too, are later differentiates of the granodiorite magma existing at deeper levels. These aplitic rocks show a progressive change in character from south to north. The leucogranites south of Saunders Creek pass northwards to aplites and aplopegmatites east of Whites Trig., and these, in turn, to pegmatites, which finally dwindle and terminate in the north at the Marne River.

The leucogranites occur as wide dyke-like bodies intruding both marble and quartzo-felspathic schist in the Saunders Creek area (coloured red on Plate 1). They are light coloured medium grained rocks rich in felspar and quartz and poor in mica (biotite plus muscovite or muscovite only). In some of these, potash felspar may exceed the plagioclase. The leucogranites grade northwards to lighter coloured aplites with little or no mica, but with increasing amounts of potash felspar. Some of these aplites grade to aplopegmatites, in which clots of coarser pegmatitic material has developed within the main aplitic mass. The texture and grain size of the aplopegmatites is extremely variable. Further north the dykes become sparser, narrower and more continuous, and the aplitic texture changes to a fine-grained pegmatitic one. The cores of some of these pegmatites may be coarse-grained and contain small amounts of tourmaline and garnet. Some pegmatites have a central core of quartz pierced by black radiating prisms of tourmaline. The ultimate differentiates of the aplite-pegmatite suite are rare thin ($< 6''$), medium-grained to coarse-grained pegmatites, in which quartz, potash felspar, muscovite and tourmaline are the major constituents, with rarer garnet, apatite, sillimanite and very rare beryl.

There is abundant evidence for forceful injection of these rocks. In some places the veins have followed pre-existing joint patterns (Fig. 65), and in the northernmost dykes there is a strong tendency for intrusion along the well defined bedding planes. In all cases the veins have forced an opening; structures on either side of the veins, although displaced, match exactly. These are true dilation veins. Disrupted xenoliths are not uncommon, even in the thinnest pegmatites. Rarely in the larger aplite masses xenoliths may be abundant, the rock then taking on the appearance of agmatite.

There is generally not much evidence for assimilation at the vein walls but some contact effects have been observed. At 197.092 a leucogranite-quartzo-felspathic schist contact is well exposed. As the contact of the dyke is approached the leucogranite changes through aplogranite to aplite by progressive loss of mica. The dyke is sharply bounded against the schist, but the schist near the contact has been strongly brecciated. Projecting through this brecciated zone, and extending for a short distance into the unbroken schist, are thin veins of fine-grained aplite feathering off at their ends into quartz veinlets. The schist has been bleached and silicified within a few inches of some of these veins. Thin calcareous layers in the schist have recrystallised to a coarser grain near the dyke. Within the marbles and calc-silicates epidote or scapolite skarns may be formed at the dyke contacts. Many of the thin pegmatic veins in the marble, however, do not show contact effects (Fig. 66). The contacts of the thinner dykes of aplite and pegmatite are commonly knife-sharp against the quartzo-felspathic schist, and chilled contacts up to 1 cm. thick may occur. These chilled contacts may sometimes consist of two zones; an outer very fine grained zone 1-2 mm. thick, followed inwards by a wider zone in which crystals of felspar, quartz and muscovite have grown inwards perpendicular to the contact. Swarms of small tourmaline prisms may grow within schist at the contact, and within the schist xenoliths of tourmaline bearing pegmatites. In many cases, however, the tourmaline pegmatites, like the aplites, also have knife-sharp contacts with quartzo-felspathic schist.

The presence of a moderate foliation in some of the leucogranites indicates that these rocks are of similar age to that of the granodiorites. The pattern of the dyke swarm in the Saunders Creek area establishes that the aplites were

FIG. 65.

Aplite vein injected into laminated quartzo-felspathic schist with disrupted xenolith of schist. Location 222.141. Coin diameter 2.8 cm.

FIG. 66.

Narrow folded pegmatite vein in marble. Location 212.087. Scale is 1 foot (30 cm.) rule.

FIG. 65

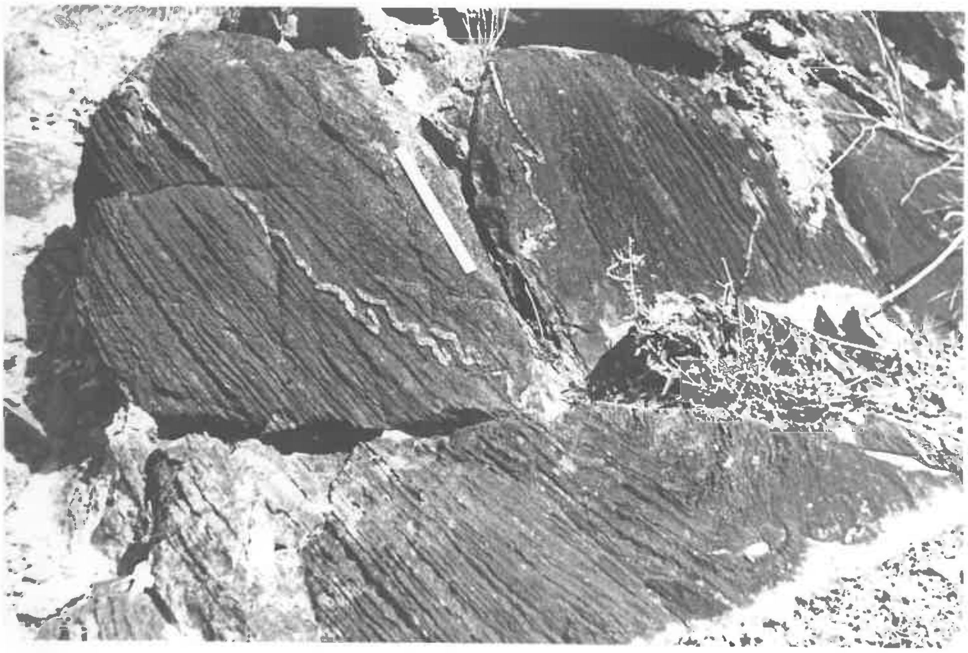


FIG. 66

intruded after the main folding. However aplites intruding the more micaceous schists and marbles here, may show ptygmatic folding (even on a large scale, 207.093 Plate 1). Evidently penetrative movements within the less competent rocks had not reached completion when the dykes were intruded.

b. Mineralogy and Petrology.

These rocks vary widely in both texture and composition. The major minerals are quartz, plagioclase, microcline and muscovite. In the leucogranites rare red-brown biotite may also be present, but its ragged form suggests that it is not particularly stable. In contrast with the granodiorites, muscovite is a stable mineral within the leucogranites, and may outline a prominent foliation due to grain alignment. Muscovite may be rare to absent in the aplites, but is again important as plumose aggregates within the coarser grained pegmatites. Quartz occurs in anhedral highly strained grains in the aplites and leucogranites, and as less strained slightly milky masses in the coarse pegmatites. Potash felspar is a major component of most aplites and pegmatites. It is a light flesh-coloured microcline with well developed Carlsbad and cross-hatched twinning and an optic axial angle $2V_x$ 74-76°. The plagioclase, unlike that in the granodiorites, is never zoned. Its composition varies from An 24 in the leucogranites, to An 13-9 in the aplites, to An 3 in the pegmatites. Carlsbad and albite twin laws were the only ones observed. The albite twins are often fine and closely spaced and may be in part mechanical twins, especially in the more foliated aplites where the felspars have become bent and fractured.

At 208.130 a 10 cm. wide pegmatite vein contains coarse silvery muscovite through a quartz-microcline matrix. Well formed 3 cm. coal-black tourmaline prisms are common, while pink garnet and stumpy smoky green apatite crystals are rarer. A small projection of quartzo-felspathic schist into the vein has small pink garnets developed within it, although garnets do not occur in the same bed outside the vein.

At 211.135 a thin sillimanite-tourmaline-spessartite pegmatite transects the quartzo-felspathic schist. The pegmatite is 7 cm. thick, with narrow (3 mm.) chilled margins and knife-sharp contacts with the schist. The major minerals are quartz, muscovite and microcline, with an average grain-size of 1 cm. One fifth of the rock consists of sillimanite, whose stumpy anhedral to euhedral crystals to 5 cm. have a fibrous, almost asbestiform, internal structure. Black tourmaline is common in the centre of the vein, where it forms irregular prisms up to 5 cm. in length embedded in quartz. A less common constituent is subhedral reddish-brown spessartite in crystals approaching one centimeter in diameter.

Beryl is extremely rare in these rocks. A single 1 cm. x 2 mm. pale green prism was noted in a pegmatite at 222.150. Rare prismatic plates of gold-brown vermiculite occur within some of the thin pegmatites intruding marble. These pegmatites are also characterised by a peculiar blue-white microcline. Roy and Romo (1957) conclude that vermiculite may form at temperatures less than 300°C at PH_2O 10000 lb./in^2.

Table 11 presents the micrometric analyses of a typical aplite and a fine-grained pegmatite. The most notable feature is the marked increase in microcline relative to plagioclase in the pegmatite, the plagioclase changing composition from An 9 in the aplite to An 3 in the pegmatite.

TABLE 11.

MICROMETRIC ANALYSES OF APLITES AND PEGMATITES.*

Rock No.	417	379
Rock type.	aplite	pegmatite
Location	213.134	228.152
No. of points counted.	2000	2500
Plagioclase	28.5	8.6
Microcline	33.5	62.0
Quartz	35.2	25.6
Muscovite	2.7	3.4
Tourmaline	-	0.4
Apatite	Tr.	-

* not calculated to weight percent.

c. Discussion and Conclusions.

The leucogranite-aplite-aplopegmatite-pegmatite suite represents a continuous differentiation series whose parent is granodiorite. These rocks were forcibly injected as magmatic liquid; this liquid, being at a slightly higher temperature than the country rock, gave rise to dykes with slightly chilled contacts. The injection took place before folding in the less competent rocks was complete. The rarity of pegmatite relative to aplite suggests that the magmatic liquid giving rise to these rocks was rather "dry." The presence of sillimanite in pegmatites has been considered by Ramberg (1949) to be indicative of "dry" conditions in the crystallising liquid. The lack of any hydrous alteration near most of the dykes, even the coarse pegmatites, suggests that very little water escaped during the crystallisation.

(c) Syenites and diorites.

The rocks to be treated here form a distinctive group apparently unrelated to the dolerites or the granodiorite suite.

The syenites and diorites are distinctive in both their petrology and field characteristics. Several micrometric analyses of these rocks, presented in Table 12, give an indication of their composition and variability. Rock 378a (column 2) is strongly porphyritic (biotite and plagioclase phenocrysts) and a recalculated analysis of the groundmass is presented in column 3. The major characteristic of the syenites and diorites is their high felspar and low quartz contents. Both biotite and hornblende are usually present, although in variable amounts. These rocks are expected to be rather contaminated by the assimilation of country rock material. In Table 12 specimens 327a and 332c are probably the least contaminated.

In the field the syenites and diorites may occur as either small intrusive pipe-like plugs, or thin (less than 3 feet) dykes, and are characterised by their isolation and widespread occurrence. They have no visible spatial relationships to either the granodiorites or the dolerites, and their only spatial limitation in the area appears to be their confinement to the intermediate zones of regional metamorphism (andalusite, staurolite and low sillimanite grade).

The best exposed occurrence of the

TABLE 12.

MICROMETRIC ANALYSES OF SYENITES AND DIORITES.*

Rock No.	327a	378a total	378a groundmass only	378b	405	332c
No. of points counted.	2000	2018	-	2000	1600	1000
Plagioclase	27.4	73.0	52.3	60.0	31.1	20.7
Potash felspar	29.4	5.4	10.7	18.4	17.3	30.5
Quartz	10.8	5.1	10.4	6.5	19.2	9.1
Biotite	15.0	10.8	14.9	7.8	5.3	9.7
Hornblende	14.5	1.1	2.3	3.4	25.3	26.2
Sphene	2.0	1.1	2.2	0.9	1.2	1.6
Apatite	0.8	1.1	2.3	0.3	0.3	2.1
Epidote	-	0.5	1.0	1.6	-	0.1
Pyrite	0.1	-	-	-	0.3	-
Opaques	-	1.9	3.9	1.1	-	-

* not calculated to weight percent.

Location of specimens given on Table

syenite-diorite plugs is at 236.172 (specimens 378a, b, 821 and 822) where two subvertical pipes, circular in cross-section and 50 feet in diameter, 200 feet apart, outcrop in the base of a forked watercourse, (Fig. 67). The pipes have cleanly cut through steeply dipping well-bedded quartzo-felspathic schists and interbedded dolerites without any visible wedging or lateral distortion, although at the immediate contacts of the intrusion the schists have been strongly brecciated, bleached and altered. The southern intrusion consists of massive hornblende diorite with an irregular patchy intergrowth of fine and coarse-grained phases. There are no foliations or other penetrative structures, xenoliths or cross-cutting veins. The northern intrusion has a partial outer ring, up to 3 feet thick, of diorite with coarse hornblende prisms. The centre of the intrusion is occupied entirely by close packed xenoliths cemented by fine-grained interstitial contaminated dioritic material. The xenoliths range through all sizes up to several feet across, and consist of a jumbled mixture of many varieties of schist and meta-

FIG. 67A.

General view of two diorite pipes intruded into quartzo-felspathic schists near the edge of the ranges north of the Marne River. The pipe in the middle distance consists of dioritic rock (D). The pipe in the foreground adjacent to a steeply dipping metadolerite sill (Dol) consists of breccia cemented with diorite (DB) with a partial rim of diorite (D). Photograph taken from 235.173 looking south.

FIG. 67B.

Exposure of diorite from the middle distance pipe of Fig.67A (location 236.172) showing variation in texture and grain-size on the scale of the outcrop.

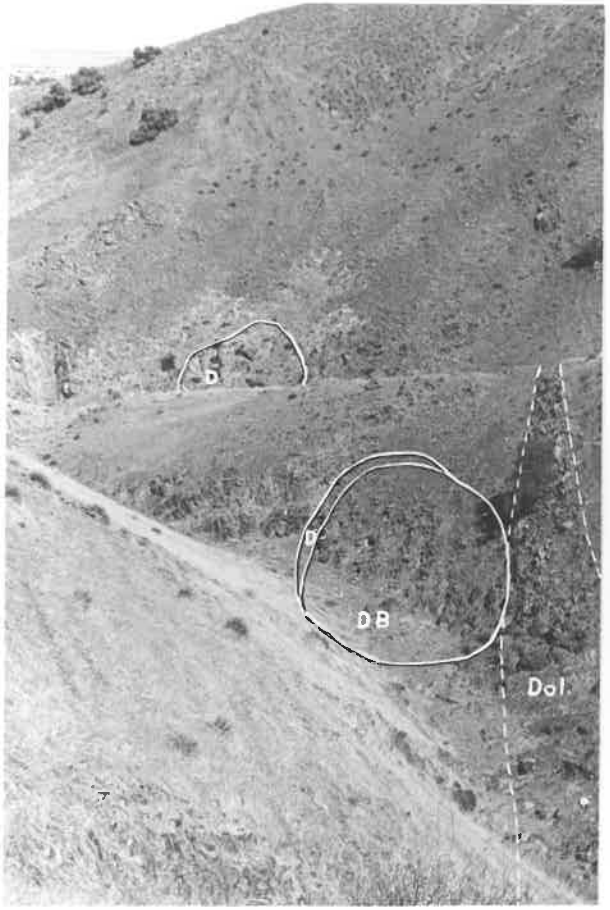


FIG. 67A

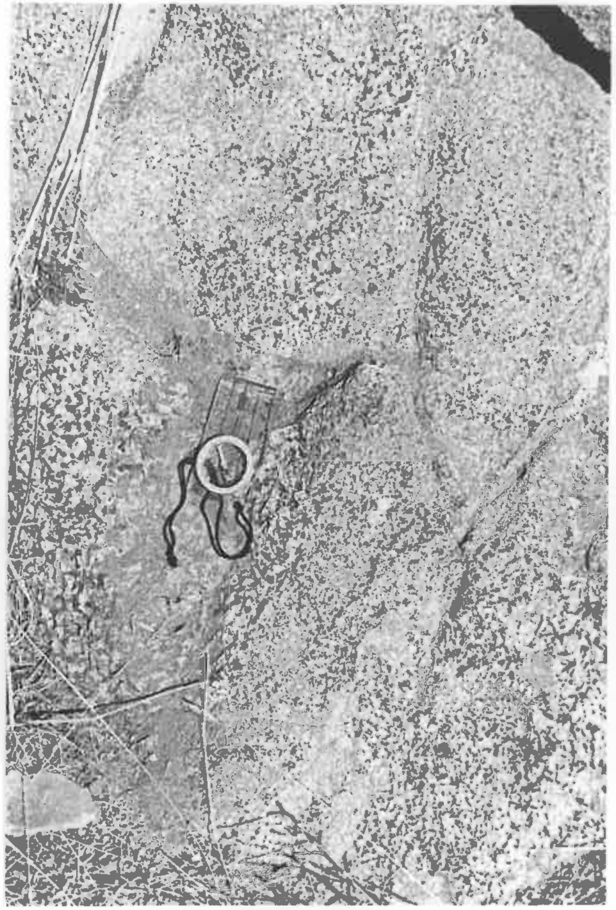


FIG. 67B

dolerite (porphyritic, non-porphyritic and schistose etc.) similar to those rocks exposed in the area adjacent to the intrusions, as well as rocks, such as fine-grained granodiorites, not exposed in the immediate vicinity. Most of the banded schist xenoliths show evidence of plastic distortion by confinement between the adjacent xenoliths. The contacts of many of the schist xenoliths with the interstitial igneous material have been fogged by assimilation. The diorite pipe must have been forced through many thousands of feet of metasediments and metadolerites to have accumulated such a variety of xenoliths. The injection must have taken place rapidly, for heat loss to the walls of the pipe would soon lower the temperature of the magma in such a narrow intrusion. The mechanism of the intrusion is a difficult problem. Perhaps explosive churning similar to that proposed for some diatreme conduits (eg. Williams, 1954) may have been an important mechanism, especially as there is no evidence for the wedging aside of the well bedded sediments or of ring fracture stoping.

A similar plug is exposed at 200.131

(specimen 327a). The rock is a coarse grained biotite-hornblende syenite (column 1, Table 12) and is uniform throughout the intrusion. No xenoliths are present here. At the contact the massive quartzo-felspathic schist has taken on a knotted hornfelsic texture, the uniformly distributed knots being skeletal muscovite, biotite and quartz in a peculiar worm-like intergrowth.

At 157.211 what might be the top of a diorite conduit is exposed. Here a dark, massive,biotite-plagioclase-(An 8)-quartz tonalite plug a few feet across is surrounded by several feet of severely brecciated, hornfelsed, quartzo-felspathic schist, the disrupted blocks of which are cemented by interstitial, contaminated igneous material.

Several syenites and diorites occur in the form of fine-grained dykes. Specimen 332c (location 180.146) is representative of one such dyke. The rock is similar in composition (Table 12) and appearance to the much coarser grained syenite 327a. Another dyke at 170.129 (specimen 268) cross-cuts marble and has been considerably altered, epidote

FIG. 68A.

Photomicrograph of syenite-diorite.

Rock A185-405. Location 208.128. Crossed polars.

Non oriented prisms of amphibole, lesser biotite and minor lozenges of sphene in a coarse-grained groundmass of plagioclase quartz and lesser potash felspar.

FIG. 68B.

Photomicrograph of porphyritic diorite.

Rock A185-378a. Location 236.172. Crossed polars.

Strongly zoned phenocrysts of plagioclase set in a matrix of plagioclase and biotite with minor quartz, sphene, epidote and amphibole.

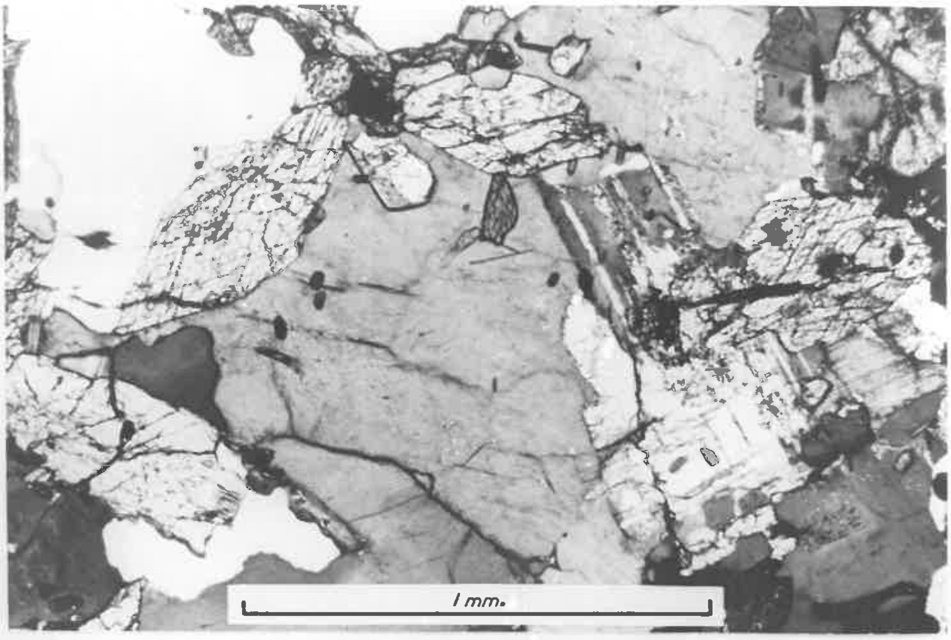


FIG. 68A

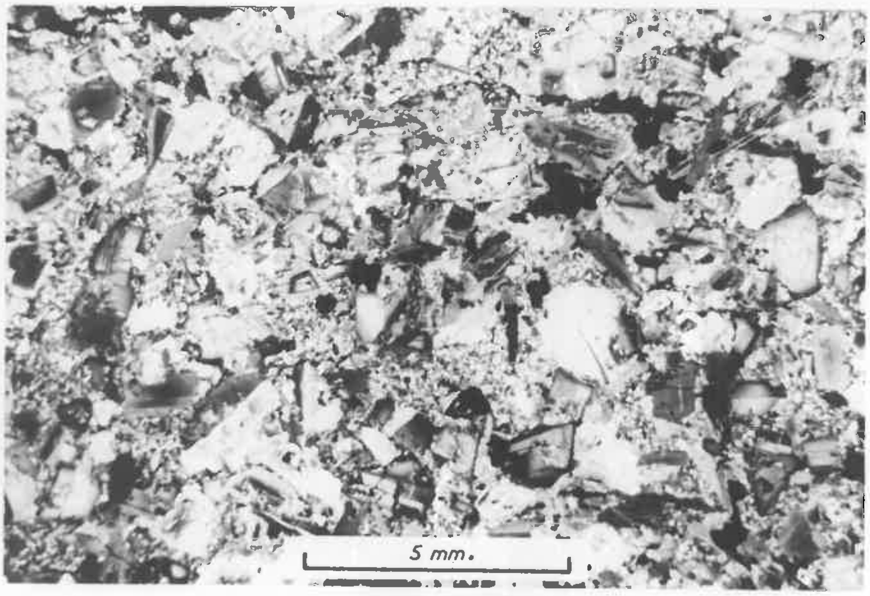


FIG. 68B

and scapolite being major secondary constituents replacing the original plagioclase.

Plagioclase in the coarser rocks is usually subhedral to anhedral and is strongly twinned and zoned. In the finer dyke rocks the plagioclase has a lath-like form and these laths, combined with the biotite plates and the hornblende prisms, may in some cases, outline a moderate igneous foliation. Table 13 presents the plagioclase composition variation in several syenites and diorites along with the twin laws that have been recognised in each. The plagioclase cores are usually andesine, more rarely they may be labradorite or oligoclase. Strongly zoned, euhedral labradorite phenocrysts are abundant in the diorite 378a. These range from An 20 at their rims to An 70 at their cores. The narrow sodic outer rims are believed to be formed by the exsolution of sodic plagioclase from the potash felspar.

Potash felspar is present in all these rocks in varying amounts. The optic axial angle, $2V_x$ varies from 63-68° and tartan twinning may be strong, weak or absent (Table 14a), with no direct relation to the optic

TABLE 13.

COMPOSITION OF PLAGIOCLASE IN SYENITES AND DIORITES.

Rock No.	Location.	%An cores	%An rims.	%An outer rims. ∅	Twin laws*
378a	236.171	51-72	22-30	20	Ab,Pr,M,C.
378b	"	54 (non-zoned An29-37)	29		Ab,C,M.
821	"	(non-zoned An32)			
822	235.172	(non-zoned An39)			
327a	200.131	44-49	20	15	Ab,C, Ab-C.
405	208.128	42-46	34-38		Ab,Ac,Pr, Ab-AlaB.
332c	180.146	23	20		Ab,C.
655	182.185	53-58	34		Ab-C,Pr.
91	210.096	43	34		Ab,Ab-C,Pr, AlaB.
268	170.129	49-55	20-27		

* Twin laws determined by Emmons 5-axis technique.

(Ab = albite, C = Carlsbad, Ab-C = albite-Carlsbad,
Pr = pericline, Ab-AlaB = albite-Ala B, M = Manebach
Ac = acline, AlaB = Ala B.)

∅ Thin outer rims on plagioclase formed by exsolution from potash felspar.

axial angle (c.f. Shido, 1958). In the incipiently twinned grains the tartan twinning is best developed near inclusions. The potash felspar occurs as irregular interstitial grains, and is evidently one of the last minerals to crystallise. Perthite stringers within the felspar are rare.

Quartz is a minor constituent. The interstitial grains are usually strained, and where the rocks have been somewhat crushed (late magmatic injection forces ?) the quartz may fill fractures in the felspars.

Myrmekite is sometimes observed at the plagioclase-potash felspar boundaries. The plagioclase of this myrmekite may have compositions up to An 38. If the myrmekite was formed by exsolution from the potash felspar, then the plagioclase originally held in solid solution in the potash felspar phase must have been more anorthite rich than that held in the potash felspar phase in the granodiorites or aplites.

The amphiboles (some of whose properties are summarised in Table 14b) have a subhedral prismatic crystal form. The usual pleochroism

TABLE 14.

a. PROPERTIES OF POTASH FELSPARS IN SYENITES AND DIORITES.

Rock No.	$2V_x$	Twinning.
378a	67.5	Incipient tartan
327	65	Incipient tartan
405	63.5	Good tartan
332c	63	None

b. PROPERTIES OF AMPHIBOLES IN SYENITES AND DIORITES.

Rock No.	$2V_x$	$Z \wedge c$	Pleochroism X	Y	Z
378a	38	18	pale yellow	green-blue	blue
378b	48	15	pale yellow	mid-green	blue
405	67.5	17.5	v. pale grn.	mid-green	dark-blue-green.
332c	62.5	16.5	pale yellow	pale green	pale greenish-blue.
327 (1)	117	11.5	non-pleochroic		
(2)	57-68	16.5	pale yellow	mid-green	bluish-green.

----

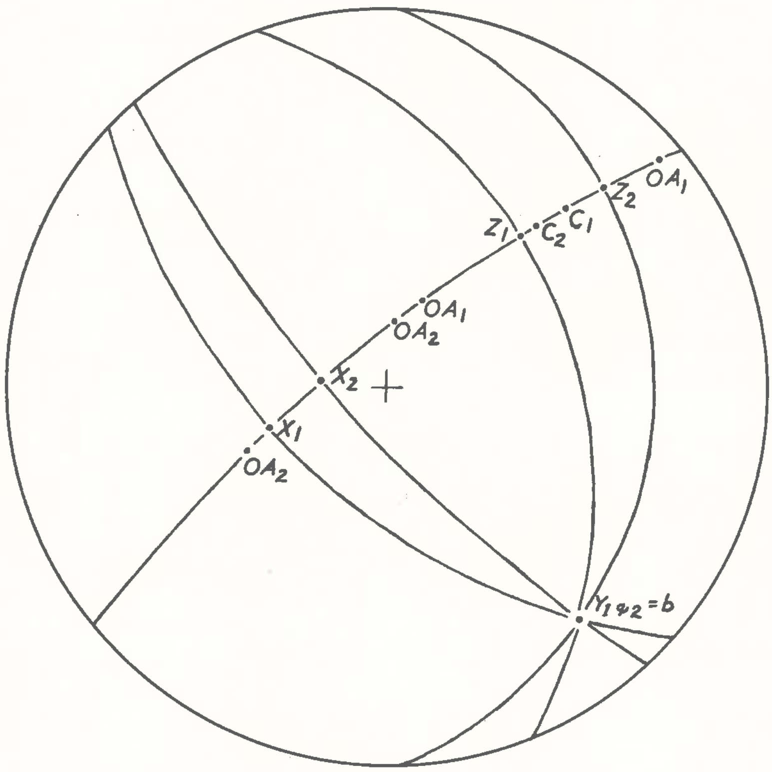
is from pale yellow to mid green to a deep blue-green. The optic axial angle is less in those with a distinct blue Z-axial colour. In some rocks (e.g. 327a) two amphiboles are present. A non-pleochroic light yellow-green cummingtonite occurs as ragged relic cores within the green amphibole, which is strongly colour zoned from pale green near the cummingtonite to mid-green at the outer rims. The green amphibole and the cummingtonite are usually homoaxially intergrown. This homoaxial relation is illustrated for one pair of amphiboles in Fig. 69. A similar replacement of cummingtonite by hornblende has been observed by Shido in the Tabito Igneous Complex.

A common feature of the amphiboles is the inclusion of stumpy biotite flakes whose platy sides are aligned subvertical to the c-axes of the amphibole prisms. As the biotite and the hornblende appear to be in good chemical equilibrium, this structure is supposed to be the result of the incorporation of the platy biotite crystals floating in the magma on the more rapidly growing ends of the amphibole prisms. This would be aided by the

FIG. 69.

Homoaxial intergrowth between two amphiboles in syenite. Rock A185.327. Location 200.131.

FIG. 69



1 WHITE AMPHIBOLE
$2V_z = 63°$, $Z \wedge c = 11\frac{1}{2}$.

2 GREEN AMPHIBOLE
$2V_x = 57°$, $Z \wedge c = 16$.

attraction between solid particles within the liquid, and by the greater movement of ions towards the more rapidly growing prism ends. The rarer biotite flakes aligned parallel to the c-axis are probably attracted to the growing amphiboles in a similar way, and are incorporated during the slower growth of the sides of the prisms.

The biotite is pleochroic from pale yellow to mustard brown. Biotite may occur as small stumpy flakes embedded in larger amphibole prisms, or as large phenocrystal flakes incorporating small amphibole grains on their edges. In both cases the biotite crystallised before or at the same time as the amphibole.

Iron rich epidote is often present as a stable primary mineral. These rocks may be considered "saturated" in epidote, for, with slight post-magmatic alteration, epidote is produced in abundance as a secondary constituent replacing the plagioclase.

Both apatite and sphene are common accessory minerals. Euhedral zircon is rare. Opaque iron oxides and pyrite may also be present.

In these rocks a paragenetic mineral sequence can be readily obtained through a study of the mineral textures. The usual sequence of crystallisation is:

zircon; sphene-apatite; biotite-hornblende-epidote; plagioclase; quartz-potash felspar; exsolution of potash felspar-myrmekite.

Near the Marne River Reserve several small pipes of sodaclase diorite (Johannsen 1923) have intruded marble and schist. These are massive, fine-grained (0.1 - 0.3 mm.), non-foliated, light coloured rocks consisting of 70-80% sodic plagioclase (An 6-11), 5-15% quartz, 10-15% hornblende with minor biotite, sphene apatite and opaques. Isolated plagioclase phenocrysts to 3 mm. are seen in the fresh, fractured surfaces. The plagioclase occurs as closely spaced, decussate, elongate laths well twinned on {010}, sometimes outlining a faint igneous flow structure, as well as non-zoned phenocrysts of the same composition. Much of the polysynthetic twinning is discontinuous across the grains and may be in part mechanical in origin. Between the laths minute granules of non-twinned plagioclase have begun to appear. The amphiboles occur in a granular, slightly skeletal form, and are pleochroic from X-pale yellow to Y-mid green to Z-bluish-green. The biotite is pheochroic X-pale yellow to Z-dark brown-black and is a rarer constituent. Small grains of quartz are interstitial and potash felspar is absent.

Conclusions - Syenite-diorite Group.

The syenite-diorite group is distinct in time and space from the granodiorite-aplite suite and the dolerite group. Some diorites and syenites have intruded as narrow pipes, others as thin dykes. Their isolation and the presence of a variety of xenoliths in one of these pipes suggests that these intrusions have penetrated many thousands of feet of metasediments. In time, the syenites and diorites are older than the faulting, but younger than the folding. They have suffered only very slight metamorphic alteration and a metamorphic foliation has not been developed within them. In the absence of cross-cutting relations, the age of the syenites and diorites relative to the later differentiates of the granodiorite suite is not known.

The peculiar sodaclase diorites of the Marne River Reserve have mineralogical features relating them to this group of rocks. They, too, lack metamorphic foliation, although a faint igneous foliation defined by the orientation of the sodaclase laths may be present. They are considered to be contemporaneous with the syenite-diorite group.

(d) Quartz Veins.

Most of the quartz veins and pods in the Cambrai area are believed to be of metamorphic segregation origin and have been dealt with elsewhere. However, a number of quartz veins of intrusive origin are known. Quartz veins and pods are not uncommon in the high grade migmatite-granodiorite area in the south-west. Most of these, with their basic biotite selvages, are no doubt of segregation origin; a few, however, may have been derived from the differentiation of aplites and pegmatites.

At 174.116 a 6 foot wide milky quartz vein carrying some pink felspar has penetrated marble. The marble within one foot of the dyke has recrystallised to a sugary, crystalloblastic texture with a coarsening of grain-size to 1 cm. The quartz-marble contacts are very irregular and jagged. Several thin quartz veins intruding marble in the Saunders Creek area have become strongly boudinaged in the late stages of the marble deformation.

Many small quartz veins with accessory tourmaline occur near the edge of the ranges in the Pine Hut Creek area. They sometimes carry accessory sulphides, including chalcopyrite, and have been prospected for copper in several places. The large vein on which the old North Rhine Copper Mine is situated also carries tourmaline, and is possibly of this type. Parts of this vein are rich in pyrite and lesser chalcopyrite. The quartzo-felspathic, arkosic schist wall-rock of the vein has been kaolinized over a distance of several feet from the vein contact. The abundance of these veins in the area around Pine Hut and northwards suggests that they may be related to the Long Ridge Granite, a late to post metamorphic bathylith known to exist in bores 4-5 miles to the east and outcropping as a "Tertiary island" in the Murray

Plains at Long Ridge, 8 miles north-east of Pine Hut.

General Conclusions.

The aim of this chapter has been to present a brief discussion of the petrology and petrography of the intrusive rocks with a view to further clarifying the geological history of the Cambrai area, and to determine the relationship of these rocks to the metamorphism and metasomatism, which have already been discussed, and to the structure which is to be considered presently.

There are several distinct groups of igneous rocks, and their time relationship to metamorphism, metasomatism and structure may be tentatively correlated as follows.

1. Metadolerites.

Intruded during the main folding and after the height of the metamorphism. These are older than the quartz-albite metasomatism.

2a. Granodiorites and Leucogranites.

Intruded during the last stages of the folding and in the dying stages of the metamorphism. Their relation to the quartz-albite metasomatism is not known.

2b. Aplites and Pegmatites.

Some folding still in progress during intrusion but metamorphism almost extinct. These rocks are younger than the quartz-albite rocks.

3. Syenites and diorites.

Post-folding and pre-faulting. Very late to post-metamorphic.

4a. Quartz-tourmaline Veins.

Pre-faulting and late or post-metamorphic.

4b. Other Quartz Veins.

May be of similar age to the aplites and pegmatites.

5. Altered Dolerite.

A post-faulting dolerite has been intruded in the south-west corner of the area.

Of special interest is the relation of the igneous intrusions to metamorphism and metasomatism. Is the metamorphism or the metasomatism related to the igneous activity? As the granodiorite suite, the syenite-diorite group and the quartz veins are late to post-metamorphic they could not have caused the regional metamorphism, and, at most, they could only have helped prolong recrystallization by locally supplying heat and increasing the water vapour pressure.

It is possible that the coarsening of the grain-size within the migmatitic schists and gneisses in the zone of strong granodiorite intrusion in the south-west was aided by local prolongation of a high temperature and increased P_{H_2O} as a result of the granodiorite intrusion. However, it is equally likely that these migmatitic mica gneisses were the cause of localisation of the granodiorite intrusion belt.

The dolerites are older than the metasomatic solutions causing the quartz-albite rocks and it is unlikely that there is any connection between these solutions and the metadolerites. The aplite dykes can be observed transecting the quartz-albite rocks, and are therefore younger. Unfortunately the relative ages of the quartz-albite rocks and the granodiorites are not known. The possibility of sodic metasomatic fronts preceding granodiorite intrusion has been proposed (e.g. Phillips, 1955). The abundance of the quartz-albite rocks in the

Saunders Creek area in a parallel trend to the granodiorite belt is the only evidence that can be mustered here in support of a causal relation between the quartz-albite rocks and the granodiorites. In the absence of any further evidence, the possibility of deriving metasomatic emanations from the granodiorites must be treated with extreme caution.

It is concluded that none of the intrusive rocks described in this area could have been responsible for the regional metamorphism, for the majority were injected at a very late metamorphic stage. It is possible that the intrusive activity has contributed to the prolongation of the metamorphism locally.

A direct connection between any of the intrusive rocks and the various effects attributed to metasomatism in the Cambrai area has yet to be demonstrated. It is possible that the quartz-albite metasomatism is indirectly connected with the regional intrusion of granodioritic magma but there is no evidence in favour of this. The metasomatic solutions might well have been derived from a more deep-seated source.

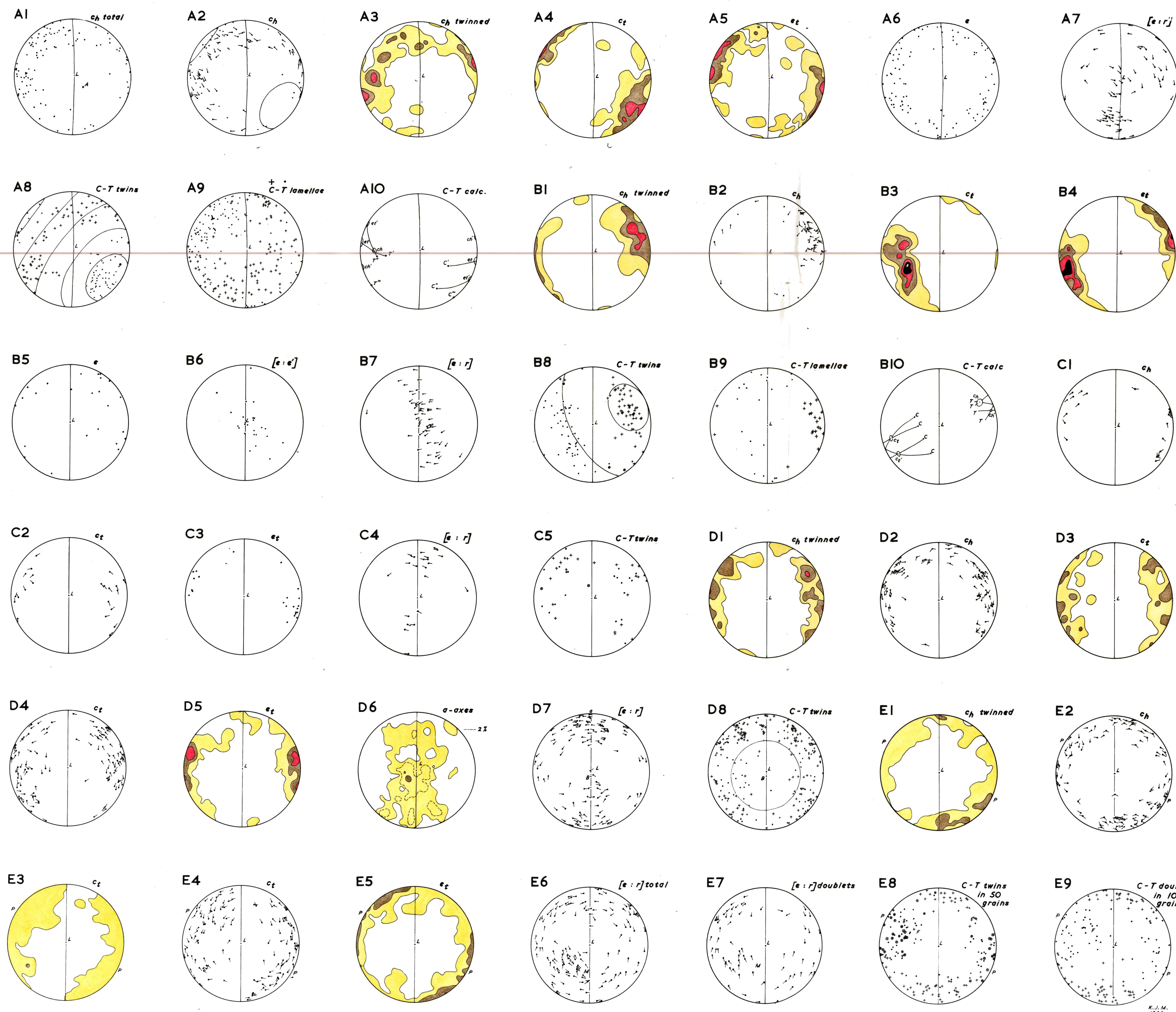
APPENDIX PLATE IA

CALCITE PETROFABRIC DIAGRAMS

DIAGRAMS A1 TO E9

EQUAL AREA SCHMIDT-LAMBERT PROJECTION

CONTOURS 1, 4, 8, 12, 16 % PER 1 % AREA



APPENDIX PLATE IB

CALCITE PETROFABRIC DIAGRAMS

DIAGRAMS F1 TO J7

EQUAL AREA SCHMIDT-LAMBERT PROJECTION

CONTOURS 1, 4, 8, 12, 16 % PER 1 % AREA

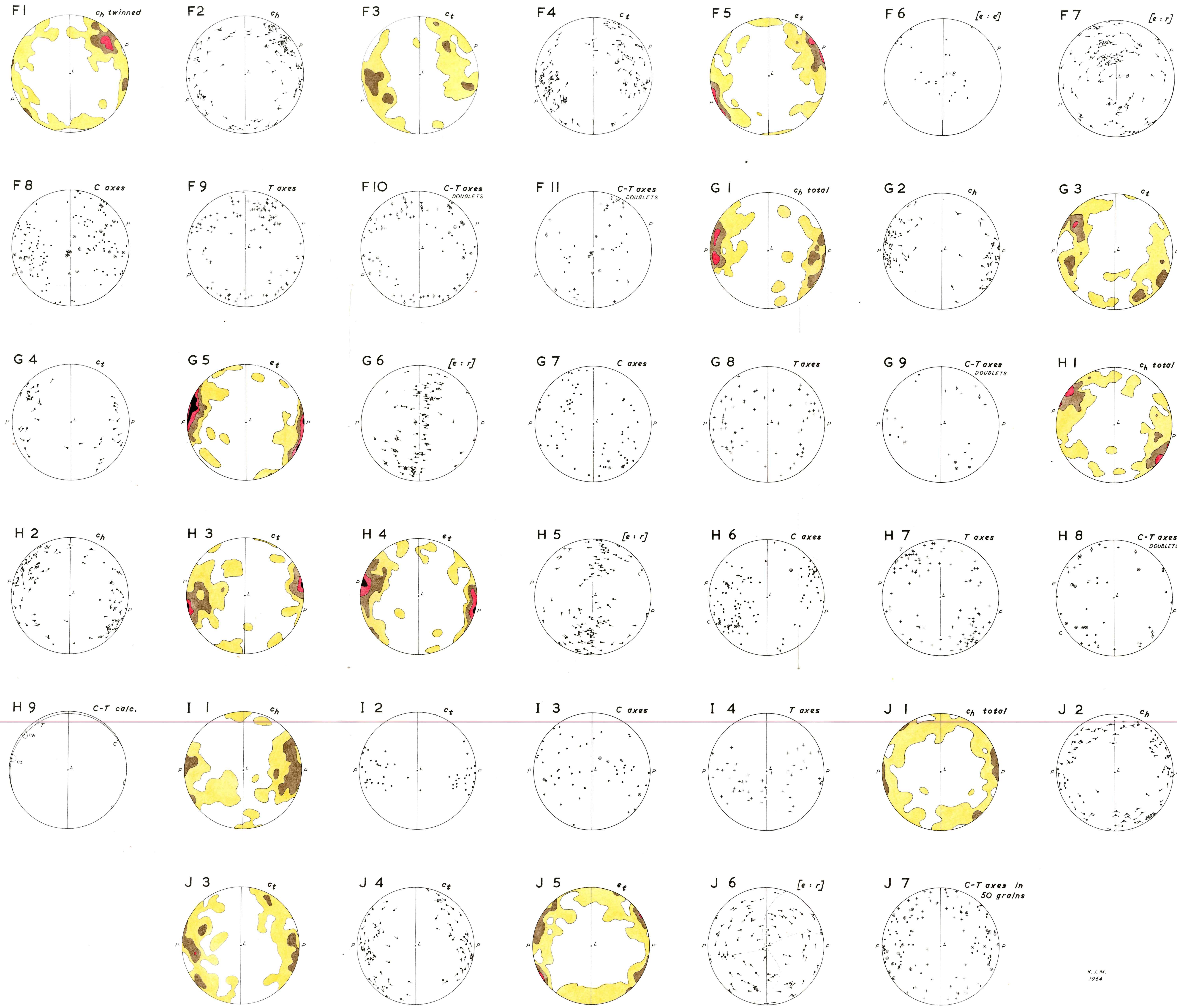
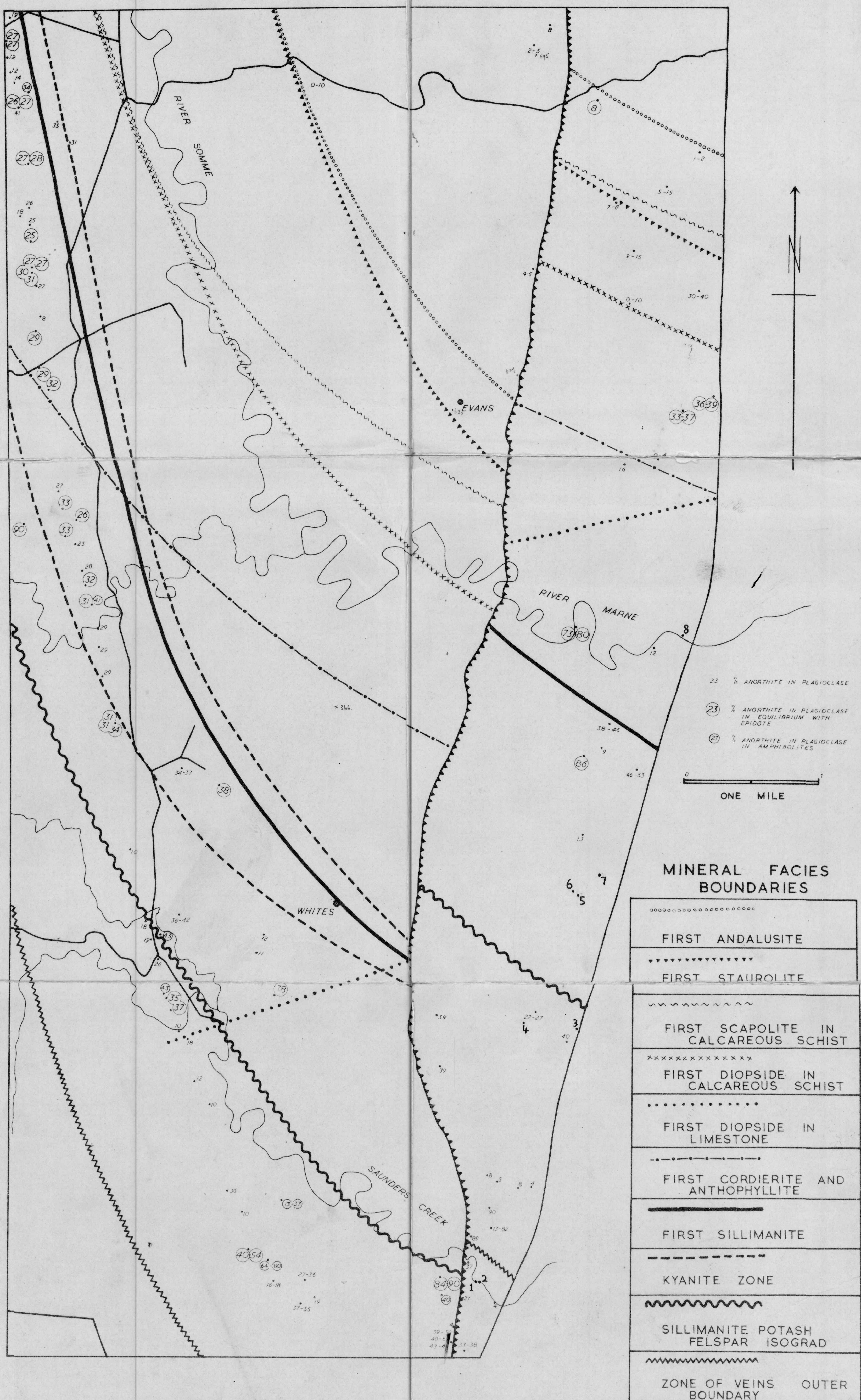


PLATE 2



RIVER SOMME
EVANS
RIVER MARNE
WHITES
SAUNDERS CREEK
N
23 % ANORTHITE IN PLAGIOCLASE
% ANORTHITE IN PLAGIOCLASE IN EQUILIBRIUM WITH EPIDOTE
% ANORTHITE IN PLAGIOCLASE IN AMPHIBOLITES
0
1
ONE MILE
MINERAL FACIES BOUNDARIES
FIRST ANDALUSITE
FIRST STAUROLITE
FIRST SCAPOLITE IN CALCAREOUS SCHIST
FIRST DIOPSIDE IN CALCAREOUS SCHIST
FIRST DIOPSIDE IN LIMESTONE
FIRST CORDIERITE AND ANTHOPHYLLITE
FIRST SILLIMANITE
KYANITE ZONE
SILLIMANITE POTASH FELSPAR ISOGRAD
ZONE OF VEINS OUTER BOUNDARY

PLATE 3.

STRUCTURAL GEOLOGY

OF PORTION OF

CAMBRAI 1:63360 SHEET

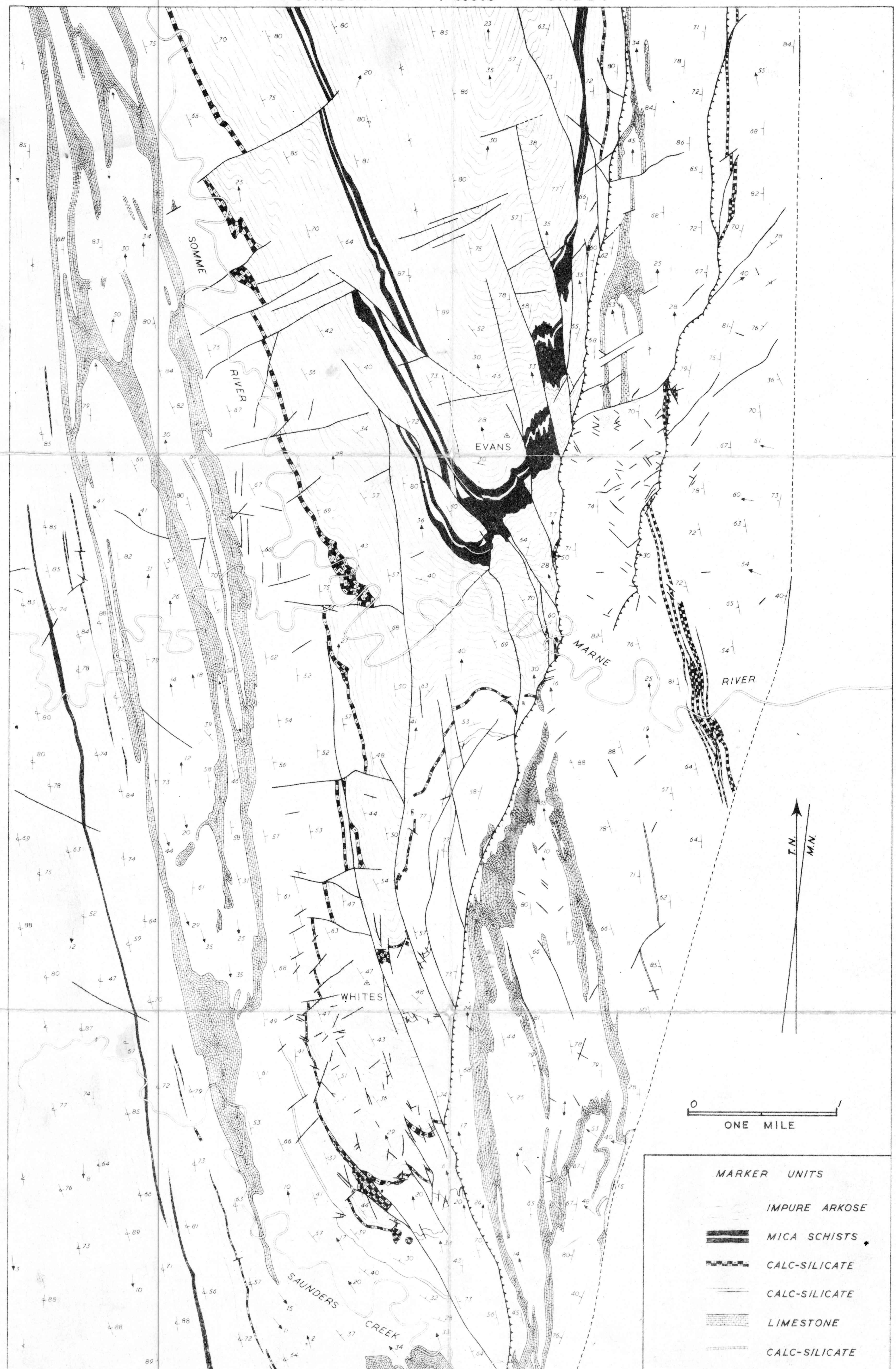


PLATE 4.

STRUCTURAL GEOMETRY OF PORTION OF CAMBRAI 1:63360 SHEET

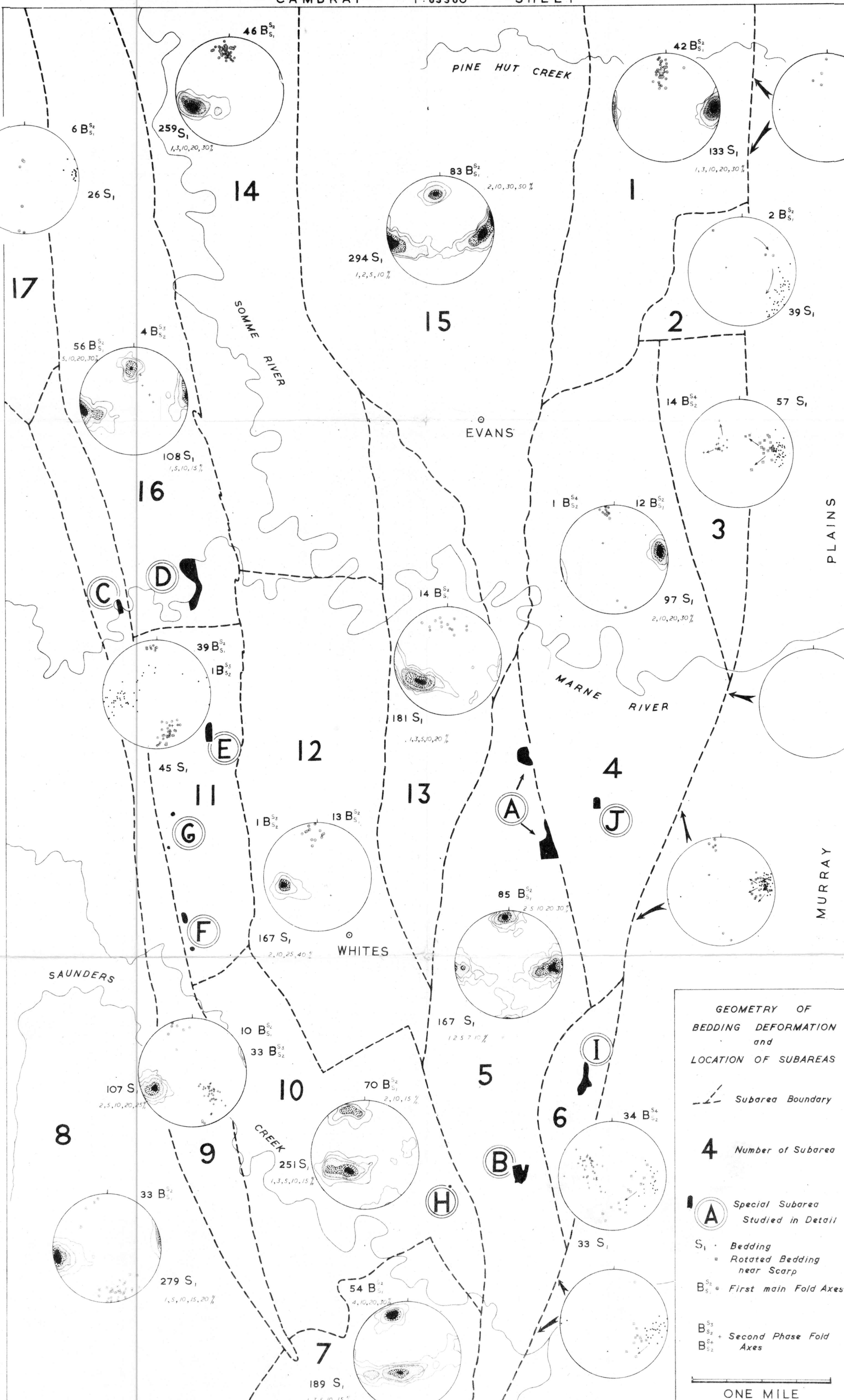
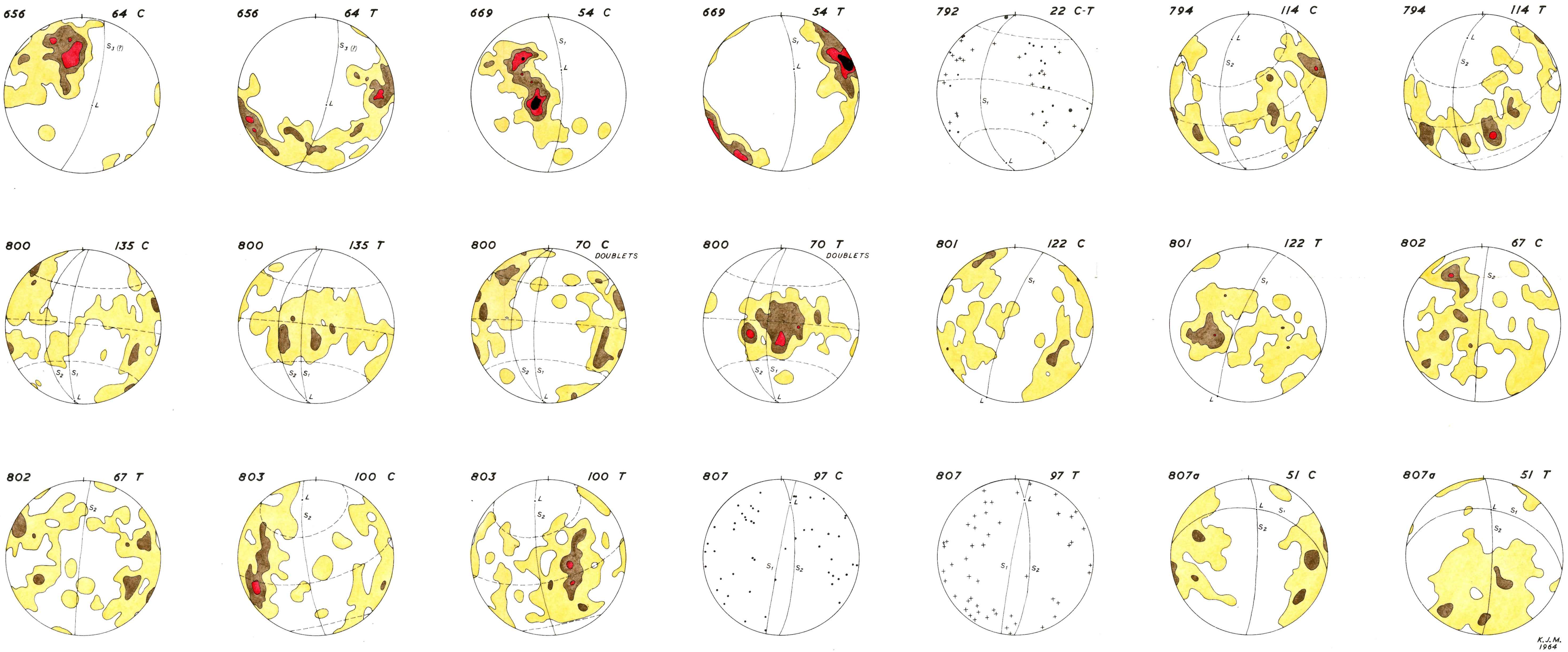


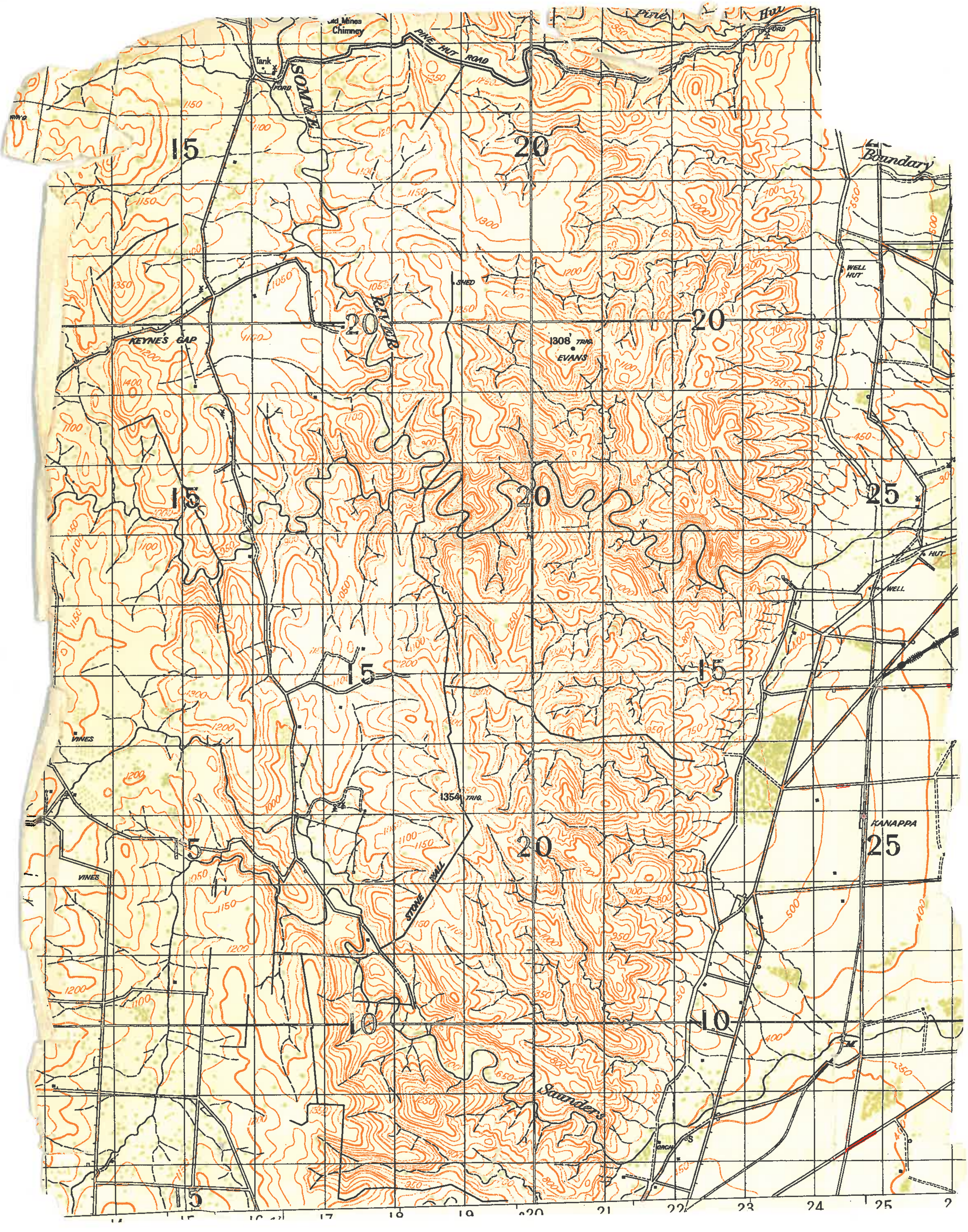
PLATE 5

CALCITE PETROFABRIC DIAGRAMS

COMPRESSION - TENSION DIAGRAMS ROTATED TO GEOGRAPHIC HORIZONTAL

EQUAL AREA SCHMIDT-LAMBERT PROJECTION TRUE NORTH AT TOP CONTOURS 1, 4, 8, 12, 16 % PER 1 % AREA





Old Mines
Chimney
Tank
FORD
SOMME
RIVER
PINE HUT ROAD
Pine
Hut
FORD
Boundary
WELL
HUT
SHED
KEYNE'S GAP
1308 TRIG.
EVANS
HUT
WELL
VINES
1354 TRIG.
STONE WALL
VINES
KANAPPA
Saunders
15
20
20
20
25
15
20
15
15
20
25
10
10