

SOME CONTACT METAMORPHIC
AND MANGANIFEROUS ROCKS FROM SOUTHERN
NEW ENGLAND

by

FRANK HOWARTH

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SUMMARY

Some contact metamorphic rocks and manganese deposits were examined from an area near Tamworth. The manganese deposits were examined and manganese silicates and oxides are described from them. It was determined that the manganese deposits are probably of sedimentary origin, and may represent allochthonous blocks carried into the present sediments from some outside source. The contact metamorphism of the rocks and manganese deposits is examined. Conditions of metamorphism up to the high hornblende hornfels facies were detected in the sediments and volcanic rocks. The metamorphism on the manganese deposits is thought to have caused some decarbonation in two of the deposits.

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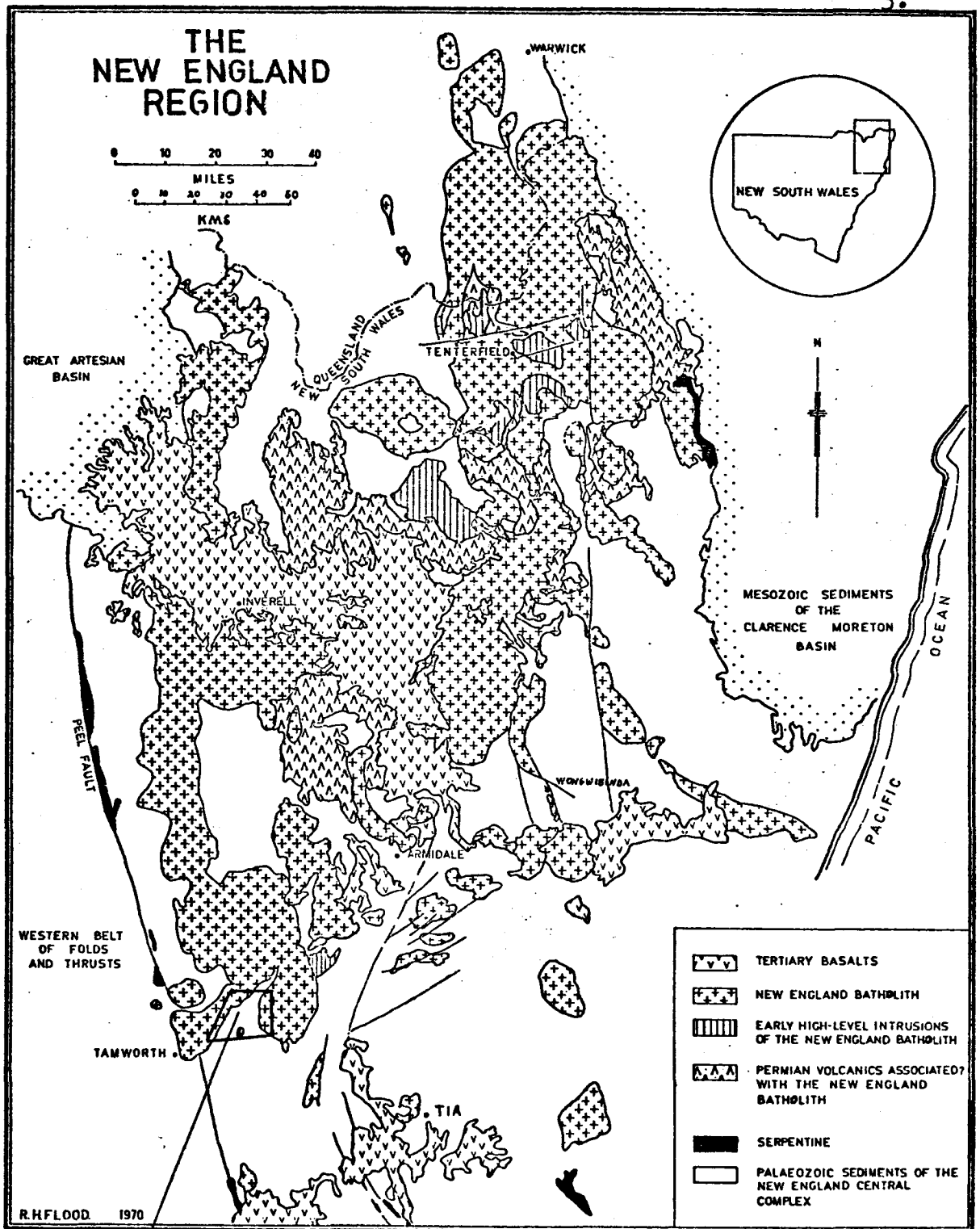
INTRODUCTION

This thesis deals with the description and discussion of some contact metamorphic and manganiferous rocks from part of the Woolomin Beds in southern New England.

The area studied is approximately 30 kms. north-east of Tamworth and includes parts of the thermal aureoles of the Walcha Road Adamellite, Bendemeer Adamellite, Moonbi Adamellite, Bannalasta Adamellite, and Limbri Leucoadamellite. The study area is shown on the locality map (figure 1) and in detail on the accompanying large scale map. Sampling was carried out using New South Wales standard 2 inches to 1 mile Topographic Maps (sheet names: Bendemeer, Haning, Kentucky, and Walcha Road) and aerial photographs (Lands Department: Nundle 1458 and Bendemeer 1676 series). The large scale map was compiled from the above sources, plus the Central Mapping Authority provisional sheet Nundle North (1: 2500 scales).

The aim of this thesis is to examine the history of the rocks in the study area, from the formation of the Woolomin Beds to the intrusion of the New England Batholith causing the contact metamorphism. A considerable part of the thesis examines the likely origin of the manganiferous rocks as these rocks show many features apparently uncommon to most manganese deposits. After discussing the origin of

these rocks the effects of contact metamorphism in the study area are examined. However due to the wide variety of rocks in the study area this discussion is in places brief and of a reconnaissance nature, considering only some aspects of the contact metamorphism in detail. The effects of contact metamorphism on the manganiferous rocks is discussed and related to the surrounding hornfelses.



THIS STUDY

Figure 1

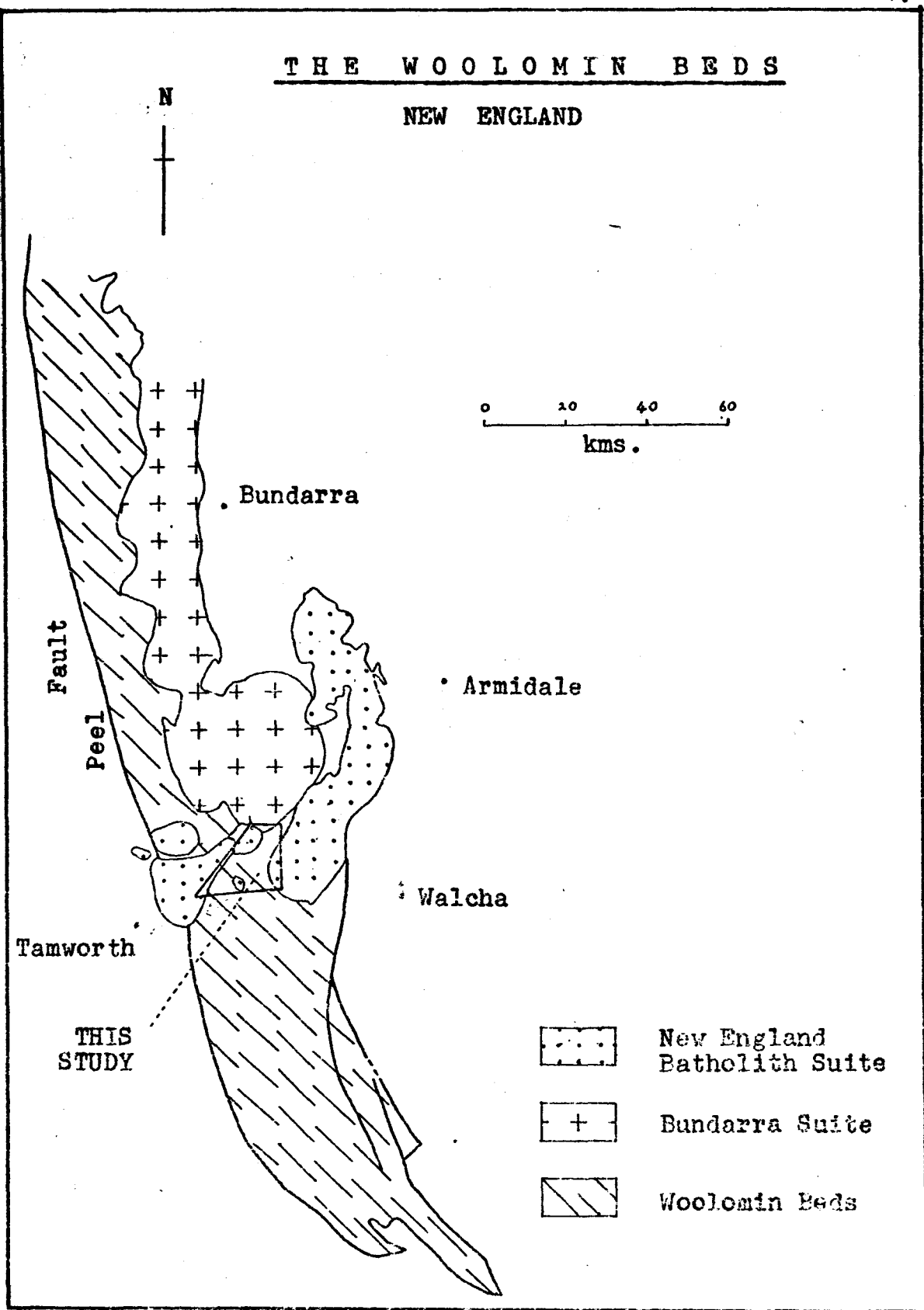


Figure 2

CHAPTER 1

REGIONAL GEOLOGY.

The contact metamorphic and manganiferous rocks examined in this study form a small part of the Woolomin Beds, an undifferentiated sequence of moderately deformed Paleozoic sediments from the New England region of New South Wales.

The main elements of the New England region are shown in figure 1 , while a more detailed map of the Woolomin Beds is presented in figure 2 . The main elements are: The Paleozoic sediments of the Central Complex of north-eastern New South Wales (Voisey 1959) of which the Woolomin Beds are a part; the New England Batholith which intrudes the Central Complex; and the Western Belt of Folds and Thrusts (Voisey 1959) which adjoins the Central Complex across the Peel Thrust Zone. In this chapter these elements will be briefly described, with a more detailed description of the Woolomin Beds, and the following Chapter will discuss the likely paleo-environment of the Woolomin Beds and briefly outline the likely geological history of the region.

The nomenclature that has developed for the sediments of the Central Complex is based mainly on locality or degree of deformation rather than stratigraphical subdivision, thus the Woolomin Beds represent mainly the western belt of the Central Complex and not an age or facies

subdivision of it. However within the Central Complex three broad lithological associations are recognisable, the oldest of these associations includes the Woolomin Beds, and also some rocks around Armidale and south and east of Tia. The description following of the Woolomin Beds can be taken to cover this association, the other two associations will be described after the Woolomin Beds.

The Woolomin Beds were first described in detail by Benson (1913-1918), and have since been described in part only, this work being summarised by Packham (1969). Aspects of the contact metamorphism of parts of the Woolomin Beds have been discussed by Spry (1953, 1956) and Binns (1966); Lewington (1973) has described the regional burial metamorphism and structure immediately south of the area studied in this thesis, while Segnit (1962) described some of the manganese deposits in the study area.

The rocks of the Woolomin Beds can be divided into three broad, but overlapping types: (i) Cherts and Jaspers, (ii) Clastic and Pyroclastic rocks, (iii) Igneous rocks.

(i) Massive cherts and jaspers form a relatively small proportion of the total Woolomin Beds, however they occur relatively frequently as interbeds varying from a few centimetres to tens of metres thick within the other sediments. The cherts are grey-green to grey in handspecimen, consisting

mainly of quartz and minor opaques, probably mainly magnetite, ilmenite and some carbonaceous material. The red jaspers usually contain abundant radiolaria in bands, and very fine quartz veinlets. The lack of red stain on the radiolaria suggests that the jaspers are primary, and not stained red subsequent to formation.

(ii) One extreme of the clastic rocks is represented by relatively abundant lithic sandstones ("greywackes"). The rock fragments are dominantly mudstone, with chert and volcanics also present, although often the rock is mainly matrix with only minor fragments. Plutonic rock fragments are absent. Minor lenses of pyroclastics occur interbedded in the sediment, and are mainly lithic tuffs.

The two above classes represent the end members in the sediment series present in the Woolomin Beds, the dominant rock type consisting of what has variously been described as "siliceous semi-pelites", "argillites", and "banded cherts". The latter name gives the best description of the rock, as it consists mainly of silica-rich (chert) and silica-poor detrital interbeds. At one extreme the silica-rich interbeds form massive cherts and jaspers, while at the other the detritus rich interbeds form massive lithic sandstones or massive shaly or phyllitic interbeds. However the dominant rock type is the banded chert, consisting of thin (microscopic to several centimetres) alternating layers of silica-rich and detritus-rich material, extremes

of each type dominating the different areas. For the purpose of concise reference from now on in the text, the term "cherts" will be used to describe rocks consisting of approximately 95% or more quartz, while "pelite" will be used for the fine grained shaly rocks. "Banded chert" will be used to describe interlayering of the two above types, while "lithic Sandstone" will be used for rocks consisting of rock fragments and sedimentary matrix.

(iii) Igneous rocks present in the Woolomin Beds (not including acid intrusives associated with the New England (Batholith) are dominantly basic (basaltic) volcanics and dykes, with minor trachytes, dacites, rhyodacites and andesites. The textures of these rocks range from fine grained porphyritic to doleritic and as their outcrop is usually only sporadic it is difficult to tell in most cases whether they are extrusive or intrusive, unless features such as pillow structures or vesicles are present, suggesting extrusion. Again, for the purposes of concise reference the term "volcanics" will be used for these rocks unless some feature present in a particular specimen allows a more definite name. The volcanics are interbedded with the other sediments, the proportion of the two types being variable.

Limestones have been found in only one small area of the Woolomin Beds, in the Attunga region just east of the Peel Thrust where they occur as small masses varying from a

single outcrop to discontinuous outcrop of 0.5km. They were described by Benson (1915b) and by Chappell (1961) who found Siluro-Ordovician fossils in the limestones. The origin of these limestones will be discussed in the next Chapter.

The two other lithological associations making up the Central Complex are thought to be younger than the Woolomin Bed association, although the age of the Woolomin Beds is problematical and will be discussed later. Both other types contain Permian fossils, and the dominant type consists mainly of lithic sandstones and pelites with rare basic volcanics and lacking cherts and jaspers, while the third association, containing the terrestrial Glossopteris - Gangamopteris flora, consists mainly of conglomerates, pelites and minor acid volcanics and pyroclastics. This last and probably youngest association is found immediately north-west of Armidale and is less deformed than either of the other two associations, while the second association forms the bulk of the Central Complex. Binns (1967) states that these associations are found either in fault contact or separated by parts of the New England Batholith. Stratigraphic relationships as such have not been determined.

Two regions of the Central Complex sediments have been regionally metamorphosed to almandine-amphibolite grade, forming sillimanite, cordierite and almandine bearing

schists. Binns (1966) describes one of these areas, termed the Wongwibinda Complex, just west of the township of Wongwibinda, while the other area, the Tia Complex is found south of the village of Tia. Most of the Woolomin Beds have been metamorphosed to the prehnite-pumpellyite meta-greywacke facies (defined in Turner, 1968), Lewington (1973) describes the development of this facies in the area immediately south of study area for this thesis.

Almost entirely within the Central Complex is the New England Batholith, extending from the Peel Thrust zone to the coast, and under the Mesozoic sediments of the Clarence - Moreton Basin. Several recent summaries of previous work and detailed studies of parts of the New England Batholith are available, particularly Binns (1967), Leitch (1968), Packham (1969), and Flood (1971) and as only the broad aspects of the New England Batholith will be described here, the reader is referred to these works for further details.

A two-fold division of the New England Batholith into two suites has been recognised, while Flood (1971) suggests a three-fold division. These suites are:

The Hillgrove Suite (Binns 1967). This group comprises a relatively small group of pre and synkinematic intrusions located east and south of Armidale, and are thought to be the oldest intrusions in the New England Batholith, ranging in age from 269 - 250 my. (Binns 1966).

They are either foliated or show protoclastic textures, are generally elongate parallel to the strike, and are associated with high grade regional metamorphism (the Wongwibinda Complex, Binns 1966). Granodiorites are the dominant rock type, and show relatively low $\text{Fe}_2\text{O}_3/\text{FeO}$ ratios and low CaO compared to the other suites.

The New England Batholith Suite. These intrusions form the majority of the overall New England Batholith and are younger than the Hillgrove Suite, ranging from 248 - 220 my. in age. Flood (1971) recognises a further division of the New England Batholith Suite, forming the Bundarra Suite. He suggests that these intrusions form the older part of the New England Batholith Suite but are younger than the Hillgrove Suite, and form a belt of intrusions extending from the north of New England parallel to the Peel Thrust and ending north-east of Tamworth. They are characterised by coarse grain size, little hornblends, some muscovite, low $\text{Fe}_2\text{O}_3/\text{FeO}$ ratio, and are mainly leucoadamellites. In contrast the New England Batholith Suite shows higher $\text{Fe}_2\text{O}_3/\text{FeO}$ ratios and are mainly adamellites, although they range from leucoadamellites through granodiorites, diorites, monzonites and gabbros. Flood (1971) gives the approximate areas of the three suites making up the New England Batholith as: Hillgrove Suite 1400 sq. km., Bundarra Suite 3440 sq. km., New England Batholith Suite 11,000 sq. km.

The five plutons bordering the study area of this thesis and responsible for the contact metamorphism described later are shown in figure 3 and on the large scale locality map.

The oldest of these intrusions is the Bannalasta Adamellite (part of Binns' 1967 Kiaora Adamellite, which has since been recognised as a composite batholith) and forms the southern end of Flood's Bundarra Suite and is typical of that Suite in being coarse grained (average 1 cm.) discordant, and muscovite bearing. Flood (1971) suggests that it is one of the oldest intrusions in southern New England, probably 245 - 250 my. old. The presence of muscovite and tourmaline bearing acid pegmatites suggests that it was probably a relatively "wet" (high water content) deep intrusion.

The other four intrusions are all part of the New England Batholith Suite and are the Bendemeer Adamellite, Moonbi Adamellite, Walcha Road Adamellite, and Limbri Leucoadamellite and are all hornblende-biotite bearing. The Bendemeer Adamellite is older than the Moonbi Adamellite, however the relative age of this pair and the Walcha Road Adamellite is uncertain. The Limbri Leucoadamellite is probably the youngest of the intrusions, and the lack of hydrous phases suggests it was probably a relatively shallow intrusion. The Walcha Road Adamellite shows markedly different contact relations with the intruded

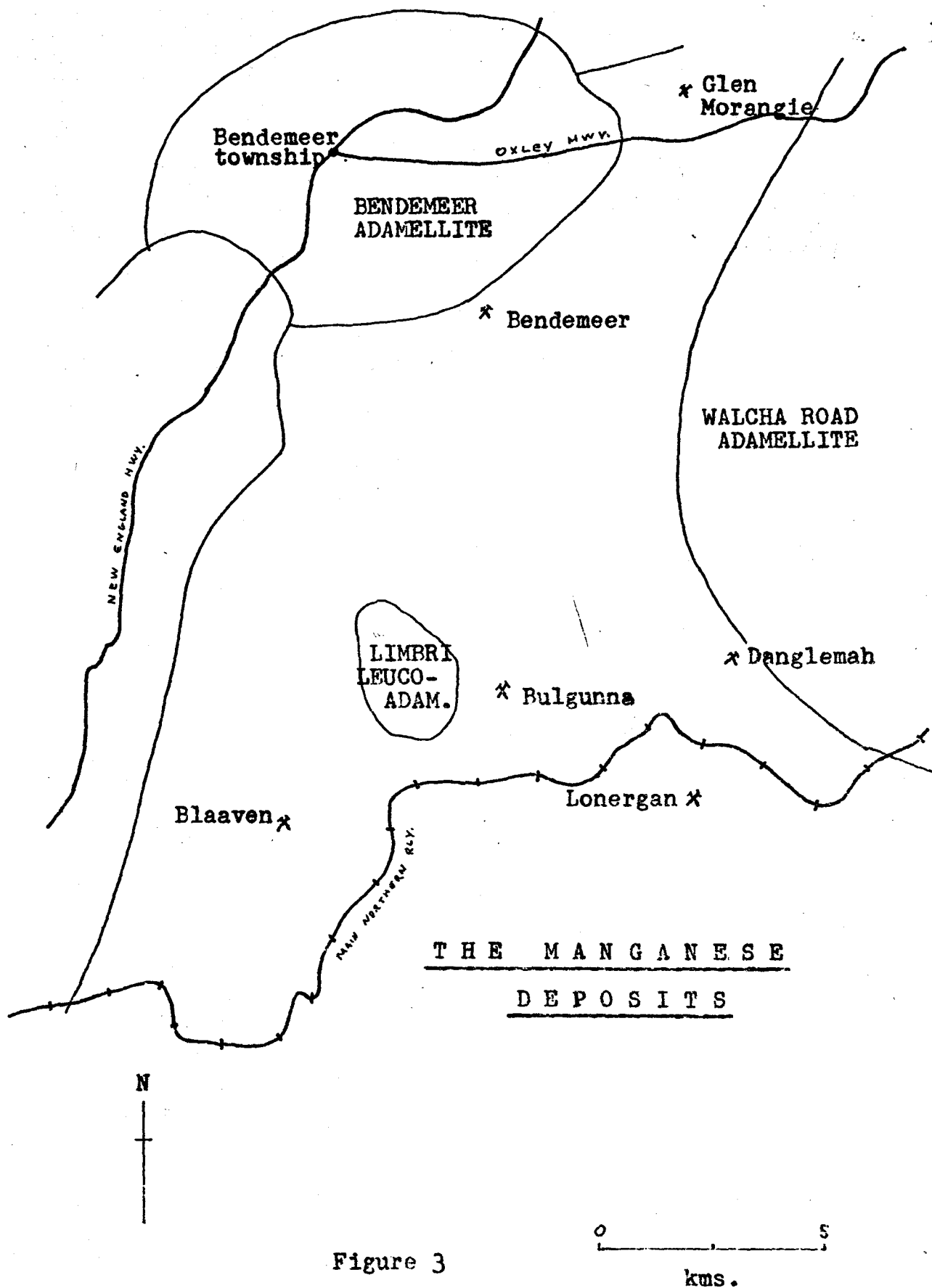


Figure 3

sediments than those of the other intrusions. Whereas the others are all "disharmonious" (Walton 1955) and do not strongly alter the structure of the contact rocks, the Walcha Road Adamellite shows signs of forceful emplacement. Flood (1971) and Lewington (1973) describe the change in regional trend to one parallel to the contact as the contact is approached, until within 10m. of the contact the rock is sheared to such an extent that only disconnected fold hinges are visible. The petrographic effects of this foliation will be discussed in the chapters on the contact metamorphism. Table 1 gives analysis and modes for four of the intrusions, none being available for the Limbri Leucoadamellite.

The Western margin of the Central Complex is, as stated earlier, the Peel Thrust Zone. The thrust extends from the north under the Great Artesian Basin to the region of Nundle, south east of Tamworth. Recent mapping in the Nundle area suggests that the thrust consists of a band of arcuate faults rather than a single fault line. Numerous alpine-type serpentinite masses occur along the thrust zone, and were probably intruded along it. As the Inlet Monzonite cuts the fault zone, its age of 248 my. places a minimum age limit on the thrust, while its maximum age is defined by the Carboniferous sediments it cuts (approximately 340 my.).

The Paleozoic sediments to the west of the thrust were termed by Voisey (1959) the "Western Belt of Folds and

Thrusts" and the easternmost of these sediments have been termed the Tamworth Trough. They are less deformed than the Central Complex sediments and show no high grade regional metamorphism, some evidence of burial metamorphosis being present however. The sediments vary in age from lower Paleozoic (doubtfully), certainly Devonian, to middle Carboniferous.

TABLE 1 (a)

ANALYSES OF ADAMELLITES SURROUNDING THE STUDY AREA.

	Bendemeer Adamellite	Bannalasta Adamellite	Moonbi Adamellite	Walcha Road Adamellite
SiO ₂	68.84	72.83	66.64	69.13
TiO ₂	0.42	0.26	0.54	0.48
Al ₂ O ₃	14.41	14.31	14.71	14.04
Fe ₂ O ₃	0.84	0.13	1.01	0.87
FeO	1.87	1.60	2.51	1.62
M ⁿ O	0.07	0.04	0.09	0.07
MgO	1.71	0.54	1.85	1.72
CaO	2.92	1.56	3.60	2.73
Na ₂ O	3.14	3.08	3.08	3.55
K ₂ O	4.76	4.74	4.58	3.93
P ₂ O ₅	0.19	0.15	0.25	0.16
H ₂ O ⁺	0.74	0.48	0.60	0.60
H ₂ O ⁻	0.18	0.10	0.14	0.14
CO ₂	0.04	0.10	0.14	-
BaO+S ^r O	0.16	0.10	10.14	-
TOTAL	100.29	100.02	99.9-	98.99
PLAG An%	30	21	35	21

(Walcha Road analysis from Flood 1971, others from Packham 1969)

TABLE 1 (b)

MICROMETRIC ANALYSES OF ADAMELLITES
SURROUNDING THE STUDY AREA

	Bendemeer Adamellite	Bannalasta Adamellite	Moonbi Adamellite	Walcha Road Adamellite
Quartz	28.8	36.8	25.5	27.3
Orthoclase	23.9	23.5	28.6	23.7
Plagioclase	37.5	34.9	30.8	37.9
Muscovite	-	1.6	-	-
Biotite	6.0	3.2	9.2	6.8
Hornblende	3.6	t ^r .	5.4	6.8
Sphene	0.2	-	0.4	0.5
Opaques	t ^r .	t ^r .	0.1	0.3
Apatite	t ^r .	t ^r .	t ^r .	-
Zircon	t ^r .	t ^r .	t ^r .	-

(Micrometric analyses are for the same specimens as are shown analysed in Table (a); sample numbers are Bendemeer Adamellite, G225; Bannalasta Adamellite, G235; Moonbi Adamellite, G38; (all from Packham 1969); Walcha Road Adamellite WR1; (from Flood 1971).

CHAPTER 2.

PALEOENVIRONMENT AND REGIONAL HISTORY.

The following chapters on the origin of the manganese deposits in the study area require a brief discussion of the likely paleoenvironment of the Woolomin Beds and surrounding Paleozoic sediments. This will be followed by a brief outline of the likely regional geological history, from the deposition of the sediments to the emplacement of the final parts of the New England Batholith, as this is necessary background for discussion of the contact metamorphism in the study area.

The lithological and sedimentary characteristics of the Woolomin Beds suggest deposition in a deep-water marine environment in a volcanically active area. Graded beds in the lithic sandstones suggest turbidity current activity, while the "banded" nature of the sediments in many places was probably due to alternating turbidity current transported clastic deposition, and chemical or biological deposition of silica in the quiescent periods. The intense and apparently incoherent deformation visible in some of the sediments was probably partly due to soft sediment deformation associated with slumping of sediments.

The lack of significant carbonate deposits suggests that the sediments may have been deposited below the carbonate compensation depth (approximately 4,500 m.). The small

scattered limestone outcrops described by Chappell (1961) near the Peel Fault may also be the result of mass movement processes. Chappell (1961) notes that the outcrops examined appeared to be small localised occurrences in otherwise normal Woolomin Beds' sediments, (banded cherts, lithic sandstones) but that bedding relationships were obscured. The existence of these fossiliferous (the only fossils found in the Woolomin Beds) probably shallow water limestones in one small area in a deep water sequence appears odd.

Edgecombe (1972) and Conaghan (1973) describe limestones in a similar environment in the Nubrigyn Formation in the Mumbil-Stuart Town area of the central west of New South Wales. They crop out in a similar discontinuous sporadic manner to those in the Woolomin Beds, and are thought to be allochthonous blocks ("olistostromes") rafted into their present position by submarine landslides or slumping (such a process is described by Jones 1967 from the New Hebrides Islands, and by Cox and Pratt 1973 from California.). The age of the fossils in the limestone is therefore the age of the limestone only, and the limestones may be considerably older than the sediments in which they now reside.

This same explanation is likely for the limestones in the Woolomin Beds, and as the accepted age of the whole Woolomin Beds has been based on the fossil ages of the

limestones, Siluro-Ordovician, then the Woolomin Beds may be considerably younger than was thought.

Much of the dating of the apparently older members of the Tamworth Trough (e.g. the Trelawney Beds, Packham 1969) has been based on similar limestones in deep water (?) sediments. Whether an extensive lower Paleozoic sequence exists in the New England region is therefore doubtful; other ages not based on limestone fossils indicate Devonian or younger age for most of the sediments.

The main change across the Peel Thrust from east to west is one of decreased intensity of deformation, with a lack of massive jaspers and cherts on the western side. Otherwise the lithology of the sediments, and probably the paleoenvironment are similar on either side of the thrust. This type of volcanic sedimentary association is characteristic of volcanic island arc regions, such as those around the margins of the Pacific Ocean. The meagre provenance direction data available for some of the sediments in the Tamworth Trough suggests the source of the sediments lay (closely) to the West. There is little evidence for an eastward bounding high having existed during the deposition of the Tamworth Trough sediments and the Woolomin Beds (and the other sediments of this type in the Central Complex).

The history of sediment deposition and volcanic activity during the Paleozoic is probably one of eastward accretion. At any one period the sediment probably

changed in type to deeper water facies towards the east, or as time passed the shallow water or terrestrial facies moved east. Thus terrestrial(?) sediments were being deposited in the Tamworth Trough in the upper Devonian, and in the Central Complex in the upper Permian.

Major tectonic activity probably began in the lower Permian in the Central Complex with the emplacement of the Hillgrove Suite, and associated deformation and regional metamorphism. Although the date of the beginning of movement on the Peel Thrust has not been determined, it was almost certainly active during this period. The areas of high grade regional metamorphism associated with the Hillgrove Suite (Binns 1966) were probably reflected in most of the other Paleozoic sediments of the Central Complex by regional burial metamorphism to lower prehnite pumpellyite metagreywacke facies. However Leitch (1969) suggests that there is evidence that burial metamorphic assemblages had developed before the regional deformation and high grade metamorphism.

The implication of this is that the contact metamorphism in the study area, mostly caused by members of the New England Batholith Suite, occurred well after the burial metamorphism was at its maximum intensity and after the regional deformation. The effects of this will be examined in the Chapters on the contact metamorphism.

Throughout the period of intrusion of the New

England Batholith (approximately 270 to 220 my. ago) the New England region was probably accreting eastwards. For detailed discussion of this process and the regional history the reader is referred to Scheibner (1972), who presents his ideas in terms of plate tectonics, and to Leitch (1969) and Voisey (1959) for discussion in geosynclinal terms.

In summary then the development of New England probably began in the lower to middle Paleozoic with deposition in a frontal island arc environment, and with gradual movement of the arc eastwards. Deformation and intrusion began in the lower Permian with deformation decreasing in the Central Complex in middle Permian times, but with intrusion of the New England Batholith continuing until the upper Permian (approximately 220 my. ago).

CHAPTER 3.

MANGANESE IN THE WOOLOMIN BEDS.

Numerous small manganese deposits have been found in the Woolomin Beds. Most of them were mined for manganese early this century and Segnit (1962) estimates that 6,400 tonnes of ore were produced during 1918, the year of maximum production. Several of these small deposits from the Bendemeer - Danglemah - Limbri area are described and discussed in this study. The following chapters present a discussion of the probable origin of these deposits and the low grade regional metamorphism effected them. Types of ancient manganese deposits and types of modern ocean floor manganese accumulations are examined and related to the environment and types of manganese deposits found in the Woolomin Beds.

Six manganese deposits were briefly examined and sampled, two of the deposits consisting each of two separate bodies a short distance apart. The locations of the various deposits are shown on

FIGURE 3

and accompanying large scale map. All the deposits have at some stage been mined for manganese or for manganese silicates (rhodonite especially). This section will be purely descriptive, the implications and probable explanations for the form of the deposits and mineral assemblage will be discussed in later sections. The names used for

the deposits, with the exception of the Danglemah mine, have been devised by the writer and are those of the properties on which the deposits are located. The order of description following is one of increasing distance from the (horizontally) nearest adamellite contact.

(i) The Danglemah Mine:

This deposit was at the time of sampling held under a mining lease although it was not being worked, Segnit (1962) described the mine history and mineral assemblage. At the time of sampling access to the ore body itself was prevented by safety gates blocking off the water filled shaft, sampling was therefore carried out from the large quantities of rubble adjacent to the mine.

The ore body is lens shaped to ovoid in plan, approximately 5m. wide and 17m. long. The central part of the body has been mined leaving only the two ends of the body visible at the surface (see Plate 1). Segnit (1962) states that the body "had a central core of hard oxides, while soft oxides occurred near the surface and against the hanging wall," the silicates were found around the central oxide cone. The contact of the ore body, with the country rock is sharp although Segnit (1962) suggests that some minor redistribution of secondary manganese oxides in quartz veins occurred. The country rock at Danglemah, unlike the other deposits examined, consists of grey-green phyllite, the bedding and foliation of which trend around

the ore body itself. There appears to be a thin layer of chert on the hanging wall side of the deposit but other than this the quartz rich rocks characteristic of the other deposits appear to be lacking. Although the deposit is less than 0.5 km from the Walcha Road adamellite, there do not appear to be any signs of metasomatic activity in the vicinity of the deposit.

The main mineral types in the deposit are silicates, oxides, and carbonates. No optical attempt was made to identify the oxides because of the difficulty of distinguishing different manganese oxides in reflected light, hausmannite was however identified by X-ray diffraction, and Segnit (1962) identified pyrolusite and psilomelane as the main constituents of the secondary oxides.

The dominant manganese silicate assemblage at Danglemah consists of pink quartz-rhodonite rock; the quartz generally forms small to medium sized (up to 2mm.) grains, some showing undulose extinction and sutured grain boundaries; and rhodonite forms small irregularly shaped to larger subhedral tabular grains in patches and bands within the quartz.

Rarely large angular plagioclase grains (approximately An 40) showing deformation twins and included boundaries are found in the quartz (see plate 2a).

Segnit (1962) refers to a "brown rock of

quartzitic appearance and glassy fracture" which he describes as "quartz-knebelite rock". However the present study indicates that the knebelite-like mineral is acmite (referring to the brown variety of aegirine, identified by x-ray diffraction), forming small tabular grains in bands and small patches within the quartz-rhodonite rock (see plate 2a).

Although Segnit (1962) describes tephroite from Danglemah, as part of a "dull olive green rock with a sacharoidal texture", none was found in the samples collected by the writer. However Segnit (1962) also points out that this mineral was a valuable manganese ore due to its high manganese content and was therefore thoroughly removed during mining.

Carbonate was found at Danglemah in a sacharoidal pink rock in which rhodonite forms large (up to 5mm.) Poikiloblastic grains with smaller rounded carbonate grains. Segnit (1962) suggests that the carbonate is a low magnesium mangandolomite although the writer also identified calcite occurring as a large patch in quartz, some grains showing oxide along carbonate cleavage.

One handspecimen showed a grey-green patch approximately 3 cm. in diameter in quartz-rhodonite rock; the patch consists of quartz with abundant very small acicular to tabular inclusions of what appear to be an alkali amphibol (moderate relief, bright blue to turquoise-green

pleochoism, the other optical features being obscured by the quartz). Minor inclusions of oxide, epidote, and rhodonite occurred in the patch as well.

Small grains of barite were also noticed in most specimens.

(ii) The Bendemeer Mine:

This deposit was at the time of sampling being mined for manganese silicate rock (for display and gem purposes) and was consequently obscured in large part by rubble. The ore body itself appeared to be roughly circular in shape, being exposed in the side of a hill above the contact of the Bendemeer adamellite. The rock enclosing the deposits appears to be mainly massive to banded cherts, although mafic volcanics outcropped approximately 50m. to the east of the deposit. Manganese silicates and quartz appeared to make up most of the deposit, with secondary(?) oxides present in cracks and on joint surfaces.

The mineral assemblage of the deposit appears to be similar to Danglemah, with tephroite being more abundant (now) than at Danglemah. The main constituent of the deposit is again pink quartz-rhodonite rock, with the rhodonite sometimes developing cross-crossing networks of subhedral tabular grains.

Tephroite appears as two main forms at Bendemeer. Firstly as part of an olive green rock where it forms small rounded grains with larger rounded carbonate grains,

and sometimes forms large (up to 8 mm.) long thin or ovoid porphyroblasts in the tephroite-carbonate rock (see plate 2b). The carbonate appears to be rhodocrosite (on the basis of x-ray diffraction). The second form of tephroite is in lensoidal olive green-grey patches (up to 1 cm. wide) in rhodonite-quartz rock. These patches are zoned, having an outer rim of rhodonite, then a thin patchy rim of carbonate, followed by a core of tephroite-hausmannite minor-carbonate (see Plate 3a). As will be discussed later, these patches probably reflect the gradual reaction of manganese rich areas with quartz, the silicate mineralogy reflecting decreasing manganese content outwards. Rhodonite also forms large (up to 1 cm.) poikiloblastic grains with carbonate, (see Plate 3b) as a Danglemah. Plagioclase and barite also occur as accessories.

(iii) The Bulgunna Deposit:

Two small bodies approximately 0.5 km. from the Limbri leucoadamellite and 60 m. apart constitute the Bulgunna deposit. One of these bodies is distinctly lenticular, about 10 m. long and 4 m. thick, while the other appears block-like, but is obscured by rubble. Both bodies are in banded chert, large blocks of massive chert outcrop nearby.

Both deposits here appear to contain less manganese than either of the two described earlier, the deposits again consisting mainly of rhodonite in quartz, but with

less rhodonite. Some of the rocks from one of the deposits consist of a medium to dark brown banded rock, which consists in thin section of bands of very fine brown oxide in quartz. Thin bands and patches of rhodonite occur in the rock, and where these cut the oxide bands the oxide appears to have been removed (see Plate 4a). The rhodonite in the deposits is dominantly fine grained and irregular in shape, sometimes developing patches of coarser irregular grains.

Large grains of carbonate occur sporadically through the rock in the rhodonite rich areas, with minor barite also present.

(iv) The Glen Morangie Deposit:

This small deposit is located at the top of a Chert hill and is partly exposed in a small costein dug by miners. The actual shape of the body is not apparent, apart from being small and probably tabular (approximately 8m. long). The deposit consists mainly of manganese oxide in quartz, with minor development of rhodonite which occurs as bands and patches with minor veinlets of carbonate. Some of the quartz shows evidence of deformation tending, to be polygonal and sometimes showing sutured grain boundaries and undulose extinction.

(v) The Blaaven Deposit:

The only evidence for the presence of this deposit is two small shafts 4 m. apart and manganese oxide scree. It is found within banded chert, and appears to consist only

of oxide and quartz. Its size and vertical extent are unknown.

(vi) The Lonergan Deposit:

This deposit, which consists of two separate bodies about 150 m. apart, is again developed in banded cherts, although some jasper is present. Although it is furthest from any of the adamellite batholiths, it shows a similar mineral assemblage to the Danglemah and Bendemeer mines.

The deposit is mainly quartz-rhodonite rock, but develops tephroite and carbonate in lenses as at Bendemeer (see Plate 4b). However the whole assemblage is finer grained than at either of the other two deposits and the olive-green tephroite-carbonate rock is not present.

Acmite was identified first in the present study from this deposit, where it occurs as a very fine grained mass approximately 1 cm. thick and at least 4 cm. wide (see Plates 5a and b), handspecimen it appears to form a semi-continuous layer roughly conformable with rhodonite bands. In other specimens it occurs as thin bands, but not disseminated as at Danglemah and Bendemeer. (Further information on the acmite is provided in the appendix.)

Some hand-specimens from one of the deposits show jasper in contact with rhodonite quartz rock. A complex minerology is developed in these areas. The jasper appears to contain radiolaria, and consists of quartz, hematite, minor epidote(?) and a fine bladed to tabular mineral which

may be an epidote (see Plate 6a). The latter mineral also occurs as clusters and disseminated in patches in the quartz-rhodonite rock near the jasper contacts (i.e. within approximately 10 cm.

Quartz veins carrying rhodonite, barite, and plagioclase(?) cut the rock in places. Barite also occurs as scattered grains in the rock.

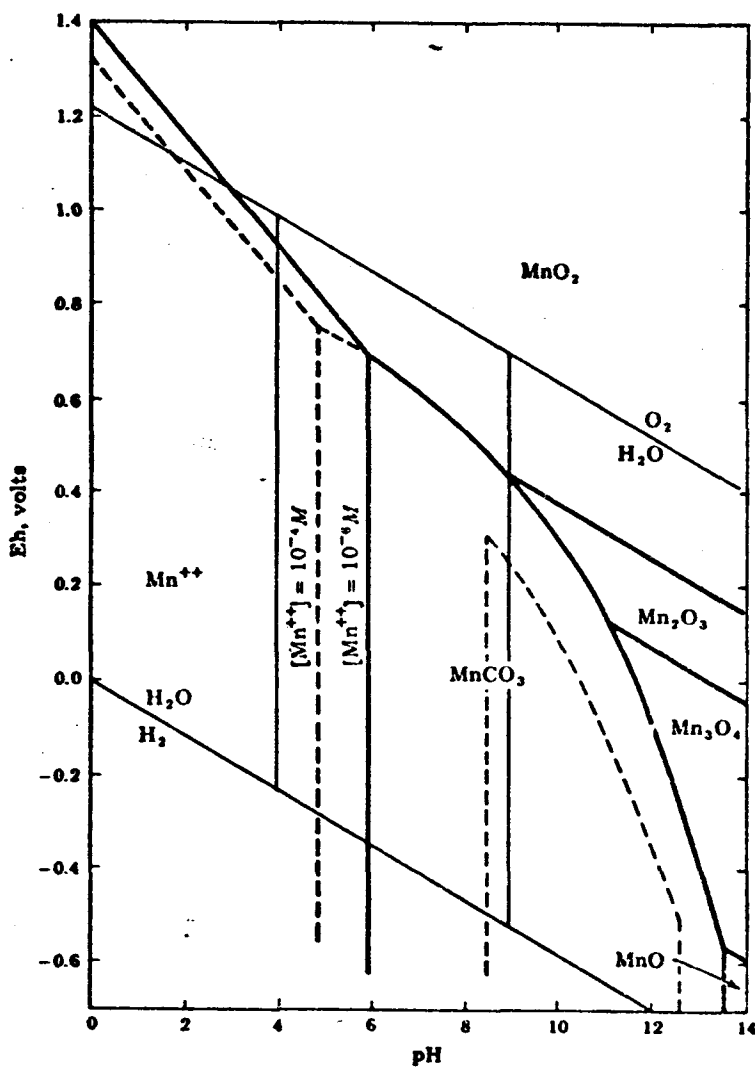
The six manganese deposits described above although broadly similar in mineralogy, vary in size and probably in overall manganese content. Danglemah and Bendemeer appear to be the most manganese rich of the deposits having relatively more manganese minerals in the deposits, Bulganna and Lonergan being smaller in size and having less manganese-rich minerals (i.e. less tephroite and carbonate). The two smallest deposits may be only manganese rich patches in the banded chert, as their outlines being obscured; however the other deposits appear to have sharp boundaries.

During sampling of the deposits and of the study area as a whole no occurrences of thin bedded or broadly disseminated manganese minerals were found. The only occurrence of manganese in the Woolomin Beds is in the types of deposits described above, very small and very large deposits being unknown. Analyses of sediments and volcanics from the Woolomin Beds show the rocks to have low to normal manganese contents; table 2 compares values for the Woolomin Beds to average manganese contents for those rock types.

CHAPTER 4.

THE GEOCHEMISTRY OF MANGANESE.

To facilitate the discussion of the origin of these manganese deposits some significant features of the geochemical behaviour of manganese will be outlined here. As with iron, manganese is carried in solution as Mn^{2+} and stays in solution as long as the solution is slightly acid and not strongly oxidising (the left hand side of figure 4). The manganese will precipitate if the solution becomes too oxidising, or saturated in dissolved manganese. The solubility of the manganese ion decreases as the Eh and pH increase. The manganese compound precipitated depends on the concentration of dissolved carbonate, silicate and sulphide in the solution. Figure 4 shows the field of MnCO_3 for a relatively high concentration of dissolved carbonate (1M.), while the dashed field is for carbonate concentration of only 0.001M. If silica content is high in the solution, manganese silicate can occupy a large part of the right hand side of the diagram. The sulphide only rarely forms, in strongly reducing conditions. If the content of silica, carbonate, and sulphide are relatively low (as is normally the case) the oxide and more rarely the hydroxide, will form. Manganese will often precipitate initially as a colloid which may itself be transported long distances before flocculating. Manganese colloids are



Eh-pH diagram showing stability fields of common manganese minerals. Assumed concentrations are total carbonate, 1M; total sulfur, 10⁻⁶M.

Figure 4

unusual in carrying a negative charge which tends to adsorb cations, especially K^+ , Na^+ , Ni^{2+} , CO^{2+} , Pb^{2+} , Ba^{2+} . and Cu^{2+} , which probably accounts for the relatively high content of minor elements in manganese minerals and deposits.

Not only must these factors be taken into account in attempting to determine the mode of origin of manganese deposits, the effects of subsequent metamorphism on the mineral assemblage must be considered, in particular, loss of carbonate by decarbonation reactions (Burt 1972, Goldsmith and Graf 1956) after forming silicates.

CHAPTER 5.

THE DEPOSITS - SYNGENETIC OR EPIGENETIC?

The early geologists' and prospectors' view of the genesis of the manganese deposits was that they were obviously derived from the nearby adamellite batholiths, and if not from them, then by some hydrothermal mechanism. However it appears more likely that the deposits are syngenetic, having been deposited with the enclosing sediments or incorporated into the sediments in a solid form. The reasons for this are:

- (i) The low to normal manganese content of the adamellites and surrounding sediments (see table 2). These may be compared to the average values given in Stanton (1971) and also in the table.
- (ii) The absence of any significant metasomatic activity at the adamellite contacts; the absence of other types of hydrothermal deposits in the area; the absence of significant mobilisation of quartz or other minerals around the deposits, or within the Woolomin Beds generally.
- (iii) The tendency for manganese not to be concentrated in late stage magmatic activity; rather it tends to enter the early formed mafic phases.
- (iv) The deposits themselves appear either lensoidal or podlike with apparently sharp boundaries with the enclosing rock, and show no sign of significant veining or irregularities.

TABLE 2

AVERAGE MANGANESE CONTENT OF SOME
COMMON ROCK TYPES

	<u>Study Area</u> *	<u>Stanton (1971)</u> **
Rhyolite	-	0.08
Andesite	-	0.15
Basalt	0.19 (av. 17 anal)	0.23
Adamellite	0.07 (av. 4 anal)	0.08
"Greywacke"	0.11 (av. 10 anal)	0.08
Pelites	0.07 (av. 3 anal)	-
Jasper	0.04 (1 anal)	-

* Pelites, jasper, and six of the basalt analyses were made by the writer . Other analyses were by Lewington (1973).

** References for average analyses are given in Stanton (1971)

(v) The proximity of some of the deposits to the adamellite contacts appears to be coincidental, the overall distribution of the deposits being apparently random.

On the basis of these points it seems unlikely that the deposits are other than syngenetic. Therefore the next section will examine the nature and occurrence of sedimentary manganese deposits. There are three main ways of supplying manganese to sedimentary environments:

- (i) Continental erosion.
- (ii) Volcanic exhalation.
- (iii) Leaching of basic volcanics.

The relative magnitudes of each and the relevance to the New England deposits will also be discussed in the next two sections.

CHAPTER 6.

SEDIMENTARY MANGANESE DEPOSITS.

A wide variety of sedimentary manganese deposits exist, both in the mineral assemblage of the deposits and in the environments in which they are found. This section presents a survey of these deposits, and of the systems used to classify them. The results of the survey will then be linked to the New England deposits, with the overall aim being to gain ideas on the likely genesis of the deposits.

Two main systems are used in classifying sedimentary manganese deposits, the first developed primarily from the Indian deposits is based mainly on the mineral assemblage of the deposits and hence the results of all the conditions the deposits have been subject to. The second system is based on the rock sequences or "formations" in which the deposits occur and is therefore environmental in its approach.

CLASSIFICATION A.

This classification is based mainly on regionally and contact metamorphosed deposits, and is summarised by Roy (1965). The first type of deposit in this classification is called GONDITE, and they occur in regionally metamorphosed terrains. Roy (1965) describes them as the "metamorphosed equivalent of non-calcareous, manganiiferous, arenaceous and argillaceous sediments and consist of

spessartite and quartz as essential minerals, commonly with rhodonite and with other manganese silicates". Some lower oxides of manganese may occur, but the essential factors in labelling a deposit a Gondite type is the lack of carbonate and presence of spessartite and quartz. The form of the deposits is considered not to be important, probably having been determined by the regional metamorphism and folding.

The second type of deposit was defined in Brazil and has been termed (by Roy 1965 from Dorr et al) QUELUZITE. These rocks are regionally metamorphosed to almandine amphibolite grade but differ from Gondites in having abundant rhodocrosite (carbonate) with essential alabandite (sulphide) and graphite and some silicates, but practically devoid of high temperature lower oxides. It is suggested by Roy (1965) that these deposits probably formed in a strongly negative Eh environment, whereas the Gondites were thought to have been deposited under oxidising conditions.

Roy (1965) also suggests an intermediate Gondite - Queluzite type characteristic of some deposits in the U.S.S.R. These deposits contain oxides, carbonate and silicate, but without sulphide and graphite. Roy (1965) considers that this assemblage indicates conditions of formation intermediate between the Gondite and Queluzite, i.e. neither strongly oxidising or reducing.

A fourth type of deposit occurs in calc-granulite marble terrains, and consists of lower oxides only.

Roy (1965) considers this to indicate a change in pH and Eh during deposition, however, other possible reasons for the apparent anomaly of oxides in a carbonate sequence exist - these will be discussed later.

The last type in Roy's classification are termed KODURITES and are formed in contact metamorphic terrains. They must have spessartine-andradite garnets in the mineral assemblage, and often have sodic minerals such as acmite and feldspars present which Roy (1965) takes to indicate that contact metasomatism has occurred. However, these sodic minerals may exist in non-contact metamorphic or metasomatic environments, so that using them to indicate metasomatism seems unwise.

As was pointed out in the geochemistry section, more factors than merely depositional ones influence the mineral assemblages produced in metamorphosed manganese deposits. Therefore attempting to draw conclusions on the conditions of formation merely on the basis of mineral assemblage seems dangerous. For example, the lack of carbonate in any of the deposits, such as the Gondites (which are generally strongly metamorphosed) may well be due to decarbonation during the metamorphism. It has also been suggested that the sulphides in the Brazilian Queluzite deposits are due to later hydrothermal activity and are not a primary feature. On a purely mineralogical basis the New England deposits fit the classification in the intermediate

Gondite-Queluzite class, but this does not reveal any information on the genesis of the deposits.

If the mineral assemblage will not reveal anything, then the environment of the deposits must be examined - this is the basis of the second classification. This classification was developed initially by Russian workers (Varentsov 1964, Shatskiy 1966, Sokolova 1971) and has been termed the "Formational Method". Sokolova (1971) defines formations as "natural associations of rocks or lithological facies which are related to each other paragenetically both in lateral direction and vertical stratigraphic succession" and also as "geological bodies which can be identified with neither stratigraphic suites nor the rocks of a certain petrographic composition." To understand the genesis of manganese deposits, the deposits must be treated as part of the formation as a whole and not as separate entities. The formational approach leads to four main types of manganese deposit formations:

(i) The Orthoquartzite - Glauconite - Clay Association.

The deposits in this association form approximately 70% of the worlds known manganese reserves (excluding sea floor accumulations) and are associated with quartzites, glauconitic beds, marls, limestones, conglomerates, and carbonaceous shales. The deposits themselves are conformable, forming thin lenses and extensive beds, and are thought to have formed in shallow marine or estuarine conditions.

This formation is found over stable basements, the deposits being either in contact with the basement or separated by sedimentary sequences and the manganese is thought to have been supplied by erosion of the basement. Oxide ores grade into carbonate ones as the depth of water in the depositional basin increased, reflecting a change in the condition of deposition.

(ii) The Limestone - Dolomite Association.

The ones in this association are related to marine carbonate sequences, and can be divided into two main types:

- (a) those developed on stable basements platforms;
("Morroccan type")
- (b) those developed in geosynclinal zones;

The Morroccan type deposits usually form in a limestone dolomite unit between sequences of terrigenous material. They are thought to have formed in platform basins transitional between the carbonate basinal facies and the terrigenous shore facies. The manganese is again thought to have been supplied by continental erosion.

The geosynclinal facies are thought to have formed in a less stable environment, the ore being within limestone dolomite sequences, usually overlying volcanic sequences, and often grading laterally into dark pyritic shales. Variability and discontinuity of sedimentation units are characteristic. The ore itself forms laminae and oolites of rhodocrosite within the limestones. The source of the

manganese in this association is debated, some writers favouring terrigenous sources, others favouring volcanic sources.

(iii) The Jaspillite Association.

The manganese ones in these deposits form part of jaspillite sequences, the manganese tending to increase relative to iron towards the margins of the jaspillites (from less than 1% Mn within the jaspillites to 5% or more at the margins). The source of the manganese and mode of formation of these deposits is involved with the whole complex jaspillite problem, and as the New England deposits are obviously not of this type nothing further will be said here.

(iv) The Volcanogenic - Sedimentary Formation.

Manganese deposits in this formation are the most numerous of all types, although economically the least important. Essentially they represent deposits associated with marine volcanic sedimentary sequences and it may well be that the New England deposits fall in this formation, considering the nature of the Woolomin Beds. Two main ways have been suggested for subdividing and classifying Volcanogenic - Sedimentary formations, the first by Shatskiy (1966) and the second by Sokolova (1971).

Shatskiy (1966) subdivides this formation into two groups on the basis of the average type of the associated volcanics, the groups being:

I The "greenstone - siliceous group";

II The "prophyry - siliceous group";

Group I formations are far more extensive than Group II and contain more manganese deposits, which is probably a reflection of the relatively low content of manganese in acid volcanism. Both groups are further subdivided into four and three formations respectively, these being:

- (1) Volcanogenic formation - Groups I and II;
- (2) The Jasperoid formation - Group I only;
- (3) The Siliceous Shaly formation - Groups I and II;
- (4) The Remote Siliceous formation - Groups I and II;

The first three of the formations above represent changes in the dominant lithology within the overall volcanic sedimentary formation, and as Shatskiy suggests, "constitutes a single group of formations, a widespread and very common one".

The volcanics within the formations include volcanic derivatives such as tuffs and lithic sandstones as well as flows and minor intrusives, these being mainly basaltic-mafic in Group I and intermediate-acid in Group II. Of the four formations listed above, the Jasperoid formation is dominant, being characterised by abundant red jaspers intimately associated with the volcanics. The Siliceous - Shaly formation reflects the increasing proportion of yellow-grey siliceous shales interbedded with clayey shales and

carbonaceous shales, plus the necessary volcanics. The Volcanogenic type is of course where volcanics themselves dominate the sequence. The Remote Siliceous formation is where volcanically derived silica is found in otherwise non-volcanic terrains, eg. pelagic sequences, but as Shatskiy points out, it is hard to identify and prove many of these formations.

The Volcanagenic Sedimentary formations are characterised by relatively high manganese values throughout the sequence, the deposits forming where the high background becomes concentrated. (This feature is absent from the New England deposits and Woolomin Beds, the background values of manganese being low to normal, see table 2). The deposits are concordant, lensoidal or semicontinuous, and usually restricted to one horizon within the sediments. In unmetamorphosed to low grade metamorphic deposits oxides dominate the ore types, with some silicate and carbonate ores present, and indicate, along with the jaspers, that the sequence is generally highly oxidised (except for the carbonaceous shales). It is almost certain that the source of the manganese lies with the volcanism, the manganese being supplied either by direct exhalation from vents or fumaroles or by leaching of volcanics by seawater. The relatively high iron content with the manganese suggests that both were deposited close to the source, the manganese not having separated from the iron. This seems to rule out

erosion of land masses as the source of the manganese.

The second classification of Volcanogenic - Sedimentary formations, that of Sokolova (1971), emphasises that these formations are both volcanic and sedimentary, and that the proportions and nature of each component varies. The volcanics vary from acid to basic, the sediments from marine-carbonate to terrigenous, and the proportion of volcanics to sediments varies. Considering the formation in four parts:

- (1) Basic Volcanic-Carbonate formation: the rock types being limestones, clayey-siliceous carbonates, jaspers, effusives, tuffs, iron and manganese ones. The ores are oxide and carbonate for both manganese and iron.
- (2) Acid Volcanic-Carbonate: The rock types being limestone, siliceous limestones, micaceous-siliceous chloritic slates, "porphyries", quartz albitophyres, tuffs, and manganese ores. The ores are carbonate and oxide.
- (3) Basic Volcanic-Terrigenous formation: the rock types being lenses of limestone, clays, siliceous shales, tuffs, jaspers, spillites, basalts, manganese and iron ores. The ores are again oxides and carbonates of manganese and iron.
- (4) Acid Volcanic-Terrigenous formation: the rock types are mainly sandstones, siltstones, and shales, with subordinate sandy limestones, "felsites", "quartz porphyries", acid tuffs and manganese ores. The ores are again carbonate

and oxide, with no associated iron.

The first two formations are related to continuous marine series whereas the last two appear to be marginal marine to shore facies. The ore deposits are correspondingly slightly more continuous in the first two types, consisting of series of conformable lenses to semi-continuous beds, whereas in the latter types the deposits tend to be slightly more erratic and discontinuous.

In both Shatsiy's (1966) and Sokolova's (1971) classifications it is suggested that the shape of the deposits (in relatively undeformed sequences) is related to the topography of the basin or ocean in which they are deposited; lensoidal deposits suggesting that the manganese ores filled depressions or hollows in the sea floor, while more continuous deposits suggest a flatter topography.

The associations of rock types described above bear many resemblances to the New England deposits, particularly Shatskiy's Jasperoid formation (except that the siliceous rock in the Woolomin Beds is dominantly grey chert rather than red jasper) and Sokolova's Basic Volcanic-Terrigenous formation (except that the terrigenous component is probably low in the Woolomin Beds). However some significant differences are apparent:

- (1) The deposits in the Woolomin Beds appear to be far more erratic and discontinuous than those described above. Also they don't appear to be related to one sediment

horizon or rock type.

(2) The one bearing horizons normally tend to be relatively high in manganese, the deposits often being particularly rich patches or beds in an otherwise manganese rich horizon. However the host rocks in the Woolomin Beds (near and far from the deposits) tend to be low to normal in manganese.

(3) Carbonates (eg. limestones) usually occur in the described formations, especially if carbonate is present in the ore. These are rare in the Woolomin Beds, although carbonate ore occurs.

It is apparent then that the New England deposits don't neatly fall into a category of known manganese deposits, although they broadly fit the Volcanogenic - Sedimentary formation. This at least suggests that the manganese was supplied by volcanism, and deposited in relatively oxidising conditions. In the next section modern marine manganese deposits, currently forming, will be examined and an attempt made to explain the above anomalies.

CHAPTER 7.

MANGANESE IN MODERN OCEAN SEDIMENTS.

Since the Challenger Expedition of 1873 it has been known that manganese is accumulating in various parts of ocean basins, particularly in the Pacific Ocean. In this section the source, form, and likelihood of preservation of these accumulations will be discussed and related to ancient manganese deposits, particularly those in the Woolomin Beds. Figure 5 presents a synthesis or summary of the material presented in this section, and will be explained in the following pages.

There are three main sources for the supply of manganese into oceans and seas.

- (i) continental runoff;
- (ii) Volcanic exhalation - hydrothermal;
- (iii) leaching of basic volcanics by seawater;

There is much debate over the relative magnitude of each of these sources, most workers agreeing that the first two probably supply most manganese into the oceans, however recent experiments on the leaching of manganese from basic volcanics, particularly pyroclastics, (Varentsov 1971) suggests that this source might be at least as important as continental runoff. Nevertheless, the amount of manganese at the source does not directly relate to the size or grade of deposits formed, this being dependent on several factors.

MANGANESE ACCUMULATION AND

DEPOSIT FORMATION

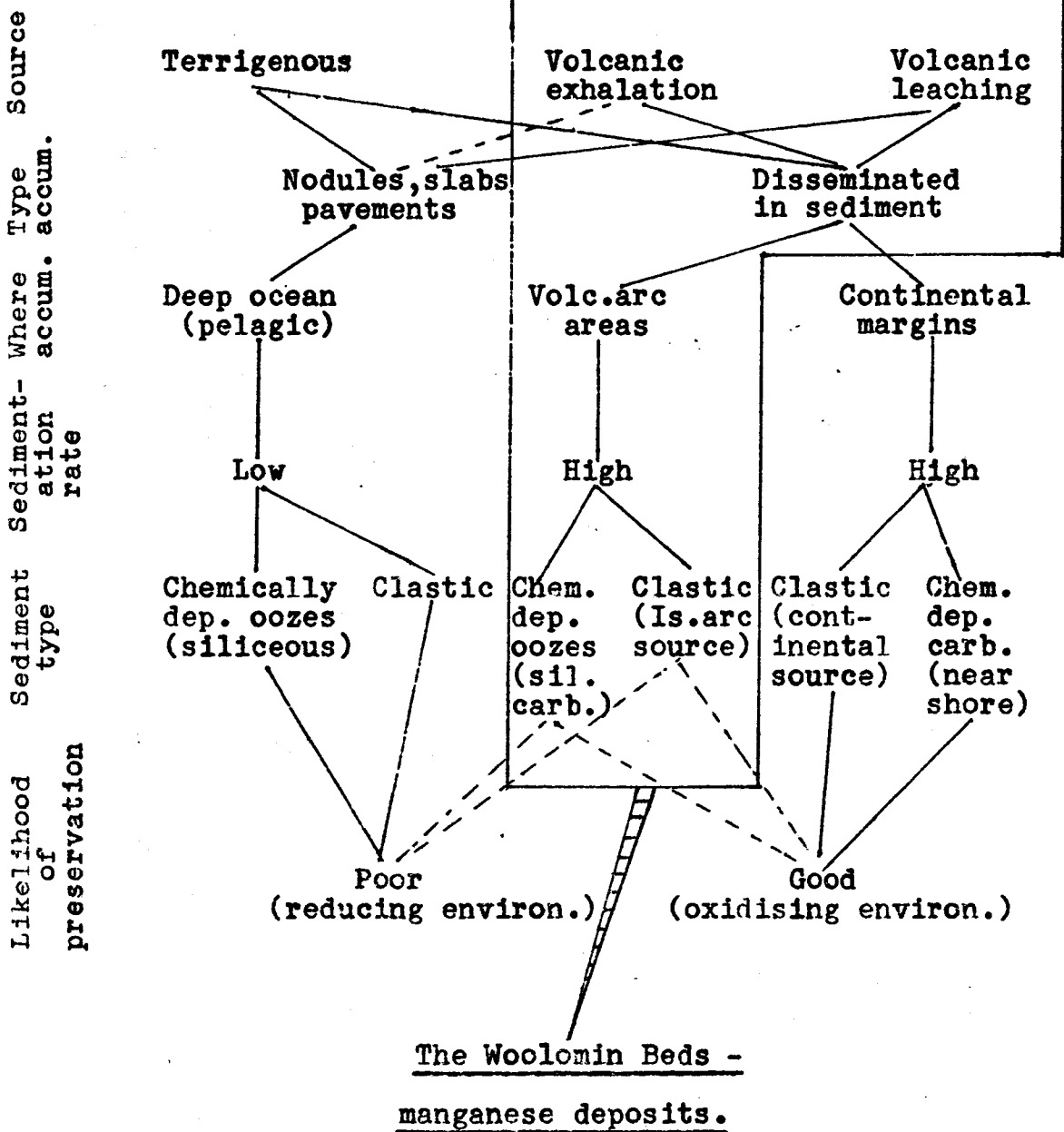


Figure 5

Continental run-off supplies manganese to the ocean margins; however much sediment is associated with the manganese and if most of both deposit near the shore, as appears most likely, then the total sediment formed will have a relatively low manganese content. Some of the manganese stays in solution or as a colloidal suspension to be deposited, very slowly, in the central ocean area where very little sediment is depositing.

The manganese derived by leaching of volcanics probably also travels relatively large distances and contributes to the overall dissolved manganese in the ocean.

However the majority of the manganese supplied by volcanic exhalations (from eruptive vents, and from fumaroles) arrives in the ocean in an oversaturated form, usually as part of hot silica-rich solutions, and therefore most of it probably precipitates close to its point of exhalation, with the remainder going into the overall dissolved manganese in the ocean.

The two key factors in producing manganese concentrations on the sea floor are then the relative rates of manganese and sediment deposition. The manganese will only accumulate if it is not too diluted by sediment; the form of manganese accumulation produced depends on the relative supplies of each.

These accumulations take two main forms, firstly as nodules, slabs and pavements; secondly as manganese

rich sediments. The manganese content is a maximum in the first form, and decreases in the second.

Nodules, slabs and pavements tend to form in areas of least sedimentation (eg. pelagic regions), and relatively low manganese supply, whereas when a certain sedimentation threshold is reached, the manganese begins to deposit as microparticles disseminated in the sediment. Nodules are the most well known of all the forms, they range in size from an average of 1 to 25 cm. diameter, density from 2.1 to 3.1, and can take almost any shape but favour rounded forms. Slabs are essentially large flattened nodules (up to several square meters in area, and up to 0.3m thick) and pavements are hardened manganese rich crusts on the sediment. Slabs and pavements favour high current - low sediment areas to form, such as seamounts and plateaus (eg. the Blake Plateau), whereas nodules tend to occur on abyssal plains.

The composition of all three forms is broadly the same (table 3 gives the average composition of nodules from the Pacific) and they are composed chiefly of three manganese oxides:

delta MnO_2 (birnessite), 10\AA manganite, and 7\AA manganite, with other elements and groups such Fe^{3+} , OH^- , and H_2O within their structure. A less abundant but important fourth phase is todorokite, $(\text{Mn}^{2+}, \text{Mg}^{2+}, \text{Ba}^{2+}, \text{Ca}^{2+}, \text{K}^+, \text{Na}^+)_2 \text{Mn}_5^{4+} \text{O}_{12} \cdot 3\text{H}_2\text{O}$, which can carry large quantities of trace and

Maximum, minimum, and average weight percentages of 27 elements in manganese nodules from the Pacific and Atlantic Oceans. (From Mero, 1965, p. 120)

Element	Weight percentages (dry-weight basis)†					
	Pacific Ocean, 54 samples			Atlantic Ocean, 4 samples		
	Maximum	Minimum	Average	Maximum	Minimum	Average
B	0.06	0.007	0.029	0.05	0.009	0.03
Na	4.7	1.5	2.6	3.5	1.4	2.3
Mg	2.4	1.0	1.7	2.4	1.4	1.7
Al	6.9	0.8	2.9	5.8	1.4	3.1
Si	20.1	1.3	9.4	19.6	2.8	11.0
K	3.1	0.3	0.8	0.8	0.6	0.7
Ca	4.4	0.8	1.9	3.4	1.5	2.7
Sc	0.003	0.001	0.001	0.003	0.002	0.002
Ti	1.7	0.11	0.67	1.3	0.3	0.8
V	0.11	0.021	0.054	0.11	0.02	0.07
Cr	0.007	0.001	0.001	0.003	0.001	0.002
Mn	41.1	8.2	24.2	21.5	12.0	16.3
Fe	26.6	2.4	14.0	25.9	9.1	17.5
Co	2.3	0.014	0.35	0.68	0.06	0.31
Ni	2.0	0.16	0.99	0.54	0.31	0.42
Cu	1.6	0.028	0.53	0.41	0.05	0.20
Zn	0.08	0.04	0.047			
Ga	0.003	0.0002	0.001			
Sr	0.16	0.024	0.081	0.14	0.04	0.09
Y	0.045	0.016	0.033	0.024	0.008	0.018
Zr	0.12	0.009	0.063	0.064	0.044	0.054
Mo	0.15	0.01	0.052	0.056	0.013	0.035
Ag	0.0006		0.0003‡			
Ba	0.64	0.08	0.18	0.36	0.10	0.17
La	0.024	0.009	0.016			
Yb	0.0066	0.0013	0.0031	0.007	0.002	0.004
Pb	0.36	0.02	0.09	0.14	0.08	0.10
L.O.I.¶	39.0	15.5	25.8	30.0	17.5	23.8

† As determined by x-ray emission spectrography.

‡ Average of 5 samples in which Ag was detected.

¶ L.O.I. = Loss on ignition at 1100°F for 1 hour. The L.O.I. figures are based on a total weight of air-dried sample basis.

TABLE 3

minor elements. Barite, celestite and anatase are other authigenic minerals, while detrital minerals present (up to 20% of the total mass) include clays, feldspars, micas, quartz, and phillipsite. All the phases except the detrital minerals are cryptocrystalline, the oxides occurring as layers with the other phases between them.

Nodules, slabs and pavements tend to accumulate very slowly, in the order of 1 mm. per $10^3 - 10^5$ years, therefore the associated sedimentation rate must be less than or equal to this value; if the sedimentation rate exceeds the accretion rate of the manganese minerals they will accumulate as microparticles disseminated in the sediment.

It has already been suggested that nodules, slabs and pavements form in pelagic areas of the ocean, and that the source of the manganese is the dissolved or suspended reservoir in the ocean itself, having been originally derived from all of the three suggested sources (continents, volcanic exhalation, and leaching). The paleo environment of the Woolomin Beds has been shown to be part of an Island-arc complex, and therefore subject to reasonably voluminous, if erratic, sedimentation. It is unlikely then that nodules, slabs or pavements were the original form of the New England deposits. Some writers have suggested that it is unlikely that most of the present nodule accumulation will ever form deposits, due to the tendency of nodules to dissolve if buried. This is apparently due to the reducing conditions

prevailing at some (usually shallow) depth in the sediment column. Dredging and coring in abyssal regions show that nodules usually exist at the sediment surface or only to very shallow depths (less than 1m.) within the sediment.

The main regions where disseminated manganese-rich sediments are found are some small areas near the continental margins, especially semi-closed seas such as the Baltic, and in active island-arc complexes, such as the Phillipine-Marianas systems. They are also known to occur on the flanks of active volcanoes, such as Banu Wuhu in Indonesia (Bonatti et al in Horn, 1972).

The continental margin manganese disseminations are usually in terrigenous clastic sediments and near shore carbonate sediments; it is these accumulations which probably form the ortho-quartzite-glaucinite-clay and limestone-dolomite formations described earlier.

However the analogues of the New England deposits must lie in the island-arc and volcano accumulations. It was suggested earlier that when silica rich manganese (and often iron) bearing solutions debouch into the ocean they are usually relatively hot and saturated in silica, manganese, and iron, so that these components begin to precipitate almost immediately, unless the environment is strongly reducing (such as in the Red Sea brine pools). It is suggested (Bonatti et al in Horn 1972) that a manganese rich silica gel then forms and collects in depressions in

the sea floor close to the source of the solutions, or forms flat continuous layers if the region is fairly level. This type of accumulation probably explains the form of most volcanogenic-sedimentary deposits, which are lenses to semi-continuous layers within the sediment. This type of area would be subject to a continuously high precipitation of manganese with the normal sediment, giving a sequence that is as a whole rich in manganese, occasional intense outpourings forming the actual deposits. The associated sediments in this environment would be dominantly siliceous oozes (forming the cherts or jaspers) with the volcanics (flows and pyroclastics) or turbidites (forming the lithic sandstones and shales). Limestones would probably also form if the area was not below the carbonate compensation depth.

The above environment has strong resemblance to that suggested for the Woolomin Beds. However the main lithology in the Woolomin beds is chert, rather than jasper, indicating less iron and probably reducing conditions, and limestone is almost absent. Even though this reveals the probable environment in which the New England deposits were produced, and broadly how they were produced (from exhaled manganese and silica) it still leaves the anomalies discussed at the end of the last chapter unresolved. The Woolomin Beds are not as a whole sequence high in manganese, nor are the rocks immediately enclosing the deposits high

in manganese. The deposits themselves are small and isolated and don't appear to be restricted to one stratigraphic horizon or one host rock type.

CHAPTER 8.

THE ORIGIN OF THE DEPOSITS - A SYNTHESIS.

The New England manganese deposits are almost certainly syngenetic, a part of the original sediments, yet in detail they appear out of place and not in accord with their environment. The Chapter on ancient manganese deposits showed that the deposits forming in a volcanogenic-sedimentary environment, such as the Woolomin Beds, are normally conformably lensoidal (their width being much greater than their thickness) to continuous beds, usually in a sequence that is high in manganese. Study of modern formative manganese deposits in volcanic areas, such as island arcs, show how these deposits form predominantly by volcanic exhalation and formation of manganese-rich silica oozes or gels. The explanation of the anomalies between this and the New England deposits might be explained in two ways: Firstly, that the New England deposits merely represent an unusual case of the normal volcanogenic-sedimentary deposit but were formed the same way as normal types of these deposits; secondly, that some other process active in this environment, such as slumping or turbidity currents, has caused the deposits to be moved from their place of formation to their current location.

(i) An unusual case of the normal process:

The degree of localisation of manganese accumu-

ation that can be produced by, say, a single fumarole issuing onto the sea bed, is relatively uncertain. Sampling of submarine volcanoes and the surrounding sea floor is done, for practical purposes, on a grid too large to pick up accumulation the size of the New England deposits; if one or two of these manganese-rich patches were detected (if they can exist) then there is no easy way of telling the exact size of the deposit. Therefore if one fumarole were to exhale a small amount of manganese-silica rich solution and it precipitated almost immediately then deposits the size of the New England deposits might conceivably result. However, there are a number of reasons why this seems unlikely.

Firstly, the likelihood of widely scattered fumaroles (if they exist) suddenly producing manganese rich solutions for short periods, and otherwise very little manganese, seems unlikely.

Secondly, even if these fumaroles producing sporadic manganese did exist, then the deposits formed would still be expected to be relatively thin lenses, rather than blocks. A more likely explanation of the deposits might be that they represent the remnants of larger manganese-silica accumulations that were destroyed or diluted by turbidites or currents. The deposits may be remnants that were in hollows or depressions in the sea floor.

If any of the deposits represent local hydrothermal manifestations then it is probably the smaller

deposits such as Blaaven and Glen Morangie, where the outlines or extent of the deposits may be more diffuse than they look. However, as a whole this explanation is not attractive and seems unlikely to be feasible.

(ii) Allochthonous blocks:

It was noted earlier that there appears to be evidence in the Woolomin Beds of turbidite activity and probably slumping leading to soft sediment deformation. It was also noted that the only known limestones in the Woolomin Beds are some isolated occurrences in the west of the Beds near the Peel Fault (Chappell 1961). These limestones occur in a manner similar to those in the Nubrigyn Formation in the Tamworth Trough, which are almost certainly allochthonous blocks rafted into the sediment by a slumping or turbidity currents.

Volcanic islands were probably the source of the limestone blocks, which were detached or slumped in a manner similar to that described from Espiritu Santo Island by Jones (1967). The lack of limestone blocks in the rest of the sequence probably reflects a lack of volcanoes or seamounts that reached the surface or came sufficiently near to form reef limestones. However the existence of lava flows and tuffs testifies to the existence of volcanoes in the region. Some of the tuffs occur in a manner very similar to the manganese deposits, as isolated outcrops within banded cherts and it seems highly unlikely that they

could have formed in situ.

Manganese accumulations on the flanks of undersea volcanoes are known in several localities (Bonnatti et al in Horn 1972) where they are thought to form silica-manganese concentrations as described earlier. These concentrations are probably less than a square kilometre in area, and relatively thick. If one of these concentrations were disrupted by mass flow of some kind (tectonically or gravity induced) after it had partially consolidated or was sufficiently coherent not to be totally dispersed, then remnants or fragments of it might resemble the form of the New England deposits, ie. as small blocks or lenses within otherwise barren sediment. This explanation would explain the unusual shapes of some of the deposits, their apparently sharp outline and block-like form, and also the lack of other disseminated manganese in the sediment. The lack of jaspers and the presence of some carbonaceous shales in the Woolomin Beds suggests that the environment was not oxidising, so that any manganese-rich unconsolidated sediment with the blocks may have had the manganese dissolved and removed by leaching, whereas if the blocks themselves were semi-consolidated the leaching might not have effected them.

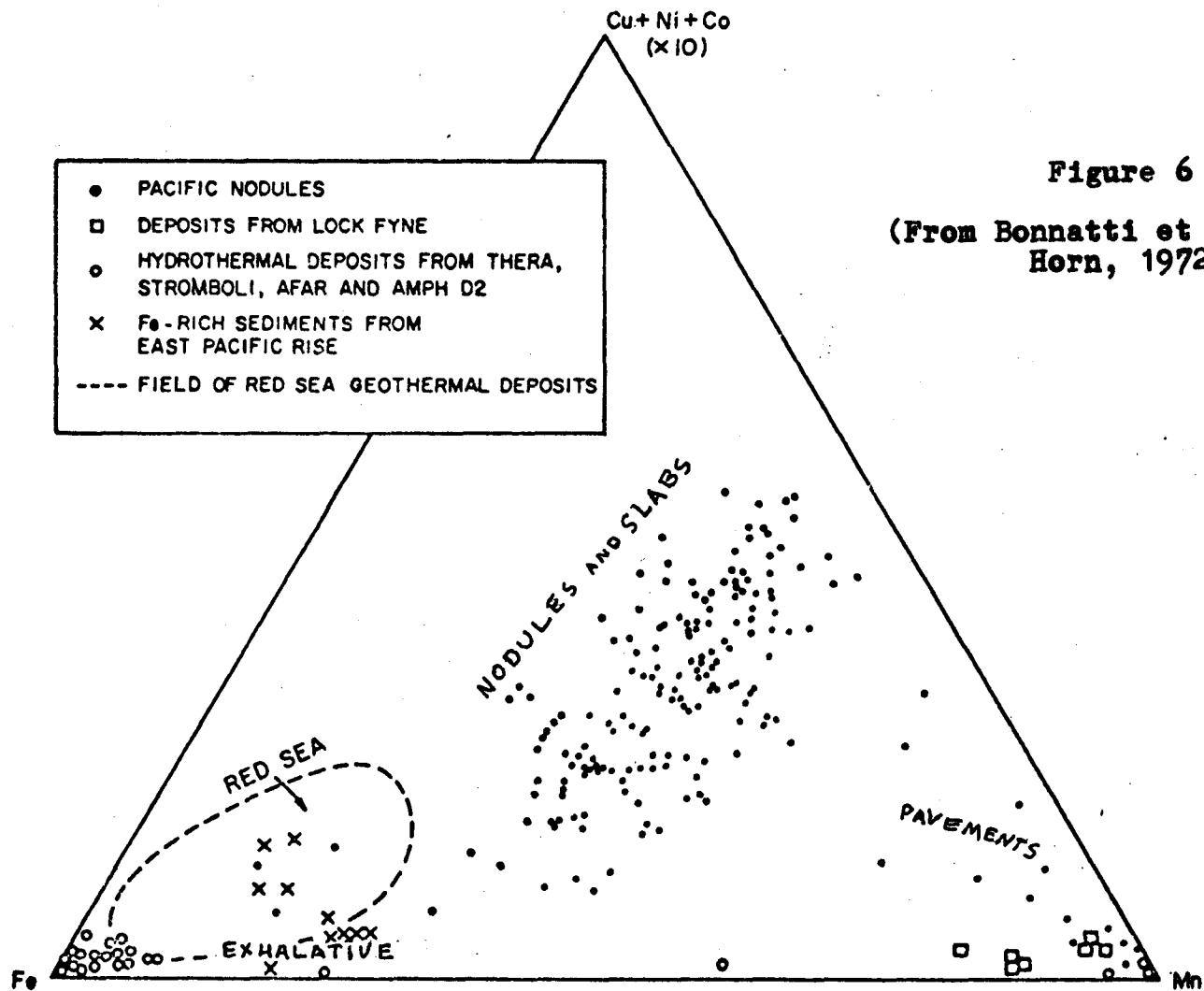
As with the first explanation the allochthonous block idea has drawbacks. No obvious breccias were noticed around the deposits. Although at the time of sampling they weren't searched for, these might be expected if the

deposits were part of a submarine landslide, unless the only other sediments associated with the deposits before slumping were oozes or muds that dispersed during the slumping process.

Clearly, to resolve the origin of the deposits further detailed work is required. There are two methods, not available to the writer at the time of the study, that might provide more conclusive evidence:

(i) In many modern ocean floor manganese accumulations regular trace element variations have been detected between different types of deposits (Bonnatti et al in Horn 1972; Bostrom et al 1971, 1973; Bender et al 1971). Volcanic exhalative deposits tend to be higher in Fe and lower in CU+Ni+Co than nodular, slab or pavement types (see Fig. 6). Exhalative deposits also appear to have higher U/Th ratios. A similar pattern is suggested for the sediments surrounding the deposits, therefore the various minor and trace element patterns for the deposits could be compared to those of the enclosing sediment; if they are widely different it might suggest that the deposits are allochthonous. As far as the writer is aware, this method has not been used on ancient deposits and their host rocks, therefore the reliability of the method is uncertain.

(ii) Close examination of the deposits and their host rocks for evidence of discordancy, sediments deformation and brecciation. Geopetal structures in the deposits, such



Ratio Fe/Mn/Cu+Ni+Co in ferromanganese deposits.

as fractures filled with sediment, might indicate the relation of the bedding inside and outside the deposit.

As more information on processes active in volcanic regions of the ocean, especially mass movement processes and manganese deposit formation processes becomes available and more detailed study of the paleo environment of the Woolomin Beds is carried out, then the problems of genesis of the New England manganese deposits might be solved with more certainty than at present.

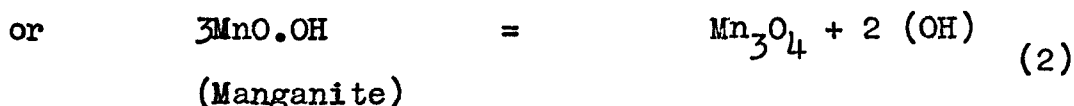
CHAPTER 9.

SINCE FORMATION - DIAGENESIS AND METAMORPHISM.

The mineral assemblages now visible in the deposits are the combined result of the assemblage at the time of formation and the effects of diagenesis, low grade regional burial metamorphism and, for two of the deposits, contact metamorphism.

The mineral assemblage at the time of formation has, to a certain extent, already been discussed. The abundant quartz in the deposits shows that silica was precipitating during formation and was probably present initially as amorphous opaline silica. The geochemical behaviour of manganese showed that it can precipitate from sea water as the oxide, hydroxide, carbonate and silicate; all except the hydroxide are now present in the deposits, but the problem is which were there initially.

Loneragan is the deposit least effected by contact metamorphism, therefore its assemblage is probably the most primary. It contains oxide, as the lower valence higher temperature form hausmannite and as finely disseminated grains in bands in quartz. The disseminated oxide form is probably primary, although the oxide has probably changed structurally from the higher valence forms precipitated from solution, to a more stable lower valence form. The hausmannite may be primary having formed from the recrystallisation of higher valence forms, for example:



(These reactions are general, the formulae for birnessite and manganite being only approximately as shown, as they both tend to be non-stoichiometric.) Alternatively the hausmannite may have formed from the breakdown of rhodocrosite, as discussed later. Literature on other deposits in similar environments also strongly suggest that oxides are part of the primary mineral assemblage. Hydroxides, such as manganite, were almost certainly present as well, although these become unstable during diagenesis and probably recrystallise to a dehydrated form, such as shown in reaction (2).

Three of the deposits contain significant carbonate, as rhodocrosite and calcite, (Danglemah, Bendemeer and Bulgunna) while Lonergan and Glen Morangie show small amounts. As carbonate does not normally form during diagenesis and metamorphism, unless metasomatically, then it is also probably primary.

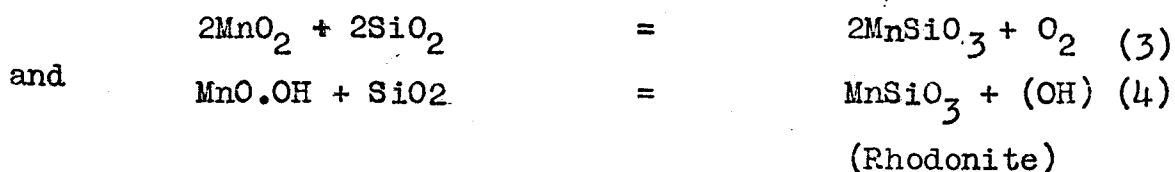
The manganese-bearing silicates, rhodonite, tephroite and acmite, pose a greater problem. They are present in the three main deposits, and their presence in the Lonergan deposit demonstrates that they had at least

started to form before the contact metamorphism had effected Danglemah and Bendemeer.

Considering each of the silicates in order of abundance:

(i) Rhodonite: It can precipitate directly from a manganese-silica rich solution and it can form from silica (quartz) and manganese oxide or carbonate during diagenesis, regional, and thermal metamorphism. Although the direct precipitation of rhodonite is chemically possible, it is usually absent from low grade oxide-silica deposits, suggesting that direct silicate precipitation is rare in nature.

The rhodonite probably began to form during the crystallisation of the amorphous silica to quartz and the breakdown of the unstable primary oxides, the approximate reactions being:

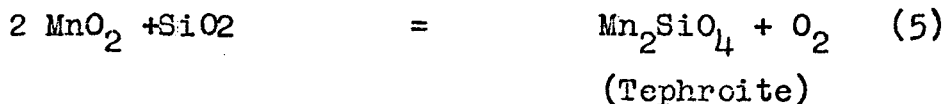


This process probably had most effect with the onset of the low grade regional metamorphism.

The brown banded oxide-quartz rock should theoretically have changed completely to rhodonite and quartz, however natural rhodonite is never pure MnSiO_3 , it always contains some calcium. Normally the calcium needed to

form the rhodonite is present in the manganese oxide or is removed from traces of calcite or plagioclase usually present, but if the oxide in the brown banded rock is calcium free then rhodonite probably could not form. The bands or lines where it has formed probably represent relict calcite veinlets, the oxide having been extracted from zones around the veinlets (see Plate 4a).

(ii) Tephroite: Hewitt et al (1961) suggests that tephroite can form before rhodonite in oxide-silica deposits; its form of occurrence at the Lonergan deposit suggests that it probably began to form at least with the rhodonite. The tephroite occurs with hausmannite in lenses surrounded by rhodonite and some carbonate, then by rhodonite and quartz. These lenses probably consisted originally of particularly rich manganese oxide (plus some silica and carbonate) patches in the normal manganese-silica gel. The areas where silica was in excess formed the relatively low manganese content silicate, rhodonite (approximately 40% Mn) whereas the oxide-excess regions formed tephroite (approximately 60% Mn) eg.:



and hausmannite, as in reactions (1) and (2).

The duration of these processes is uncertain, probably having started during burial metamorphism and

apparently having gone almost to completion to form an assemblage stable under the low grade regional metamorphism.

The lack of evidence of deformation, apart from some undulose and sutured quartz grains, might suggest that recrystallisation took place after the regional deformation had reached its peak; however it is probably more likely that the manganese deposits, being dominantly chert blocks, were little effected by the deformation.

(iii) **Acmite:** The occurrence of this mineral in low grade metamorphic deposits is rare, not to the writer's knowledge having been reported in a similar environment in Australia before. However it is known in several overseas manganese bearing assemblages and deposits. Roy (1965) and Lahiri (1971) cite examples of acmite bearing regionally metamorphosed Gondite-type deposits in India, and De Villiers (1951) mentions acmite from the Otjosondu deposits (Gondite-type) in South Africa. It is also known to occur in manganese jaspillites from Canada (Threadgold, pers. comm.) and Minnesota (Grout 1946), both of which are regionally metamorphosed. Most of the writers suggest that the **acmite** is the product of the regional metamorphism or metasomatism, however it is known to have formed authigenically in the unmetamorphosed Green River Formation from Wyoming, (Milton and Eugster 1959) which is a hypersaline lacustrine deposit. The acmite in the Lonergan deposit was probably not authigenic, but may have formed early in

the history of the deposit. Chemically, the presence of the acmite is relatively easy to explain in that most manganese accumulations contain appreciable iron and sodium (see the analysis of nodules, table 3) and its occurrences as an authigenic mineral in the Green River Formation testifies to its stability at low temperatures and pressures. The fine grained, concentrated form is harder to explain, normally it would be expected to be disseminated in the rock, unless the layer it now occupies was particularly rich in sodium and iron. (The disseminated format Danglemah and Bendemeer is probably due to thermally induced recrystallisation). Alternatively the acmite, as an iron-sodium rich solution, may have migrated to that layer during diagenesis and crystallised as a fine grained mass.

Of the other minerals present in the Lonergan deposit, the barite is probably primary, as barium and sulphate are produced by the exhalation. The plagioclase probably recrystallised from detrital fragments or formed from minor calcium, sodium and iron present in the sediment. The minor quartz, rhodonite, plagioclase, and carbonate bearing veinlets present in all the deposits suggest that some minor redistribution of material took place after the silicates had formed.

The effects of the contact metamorphism on the Danglemah and Bendemeer deposits will be discussed after

consideration of the main contact metamorphic assemblages
in the study area.

CHAPTER 10.

THE CONTACT METAMORPHISM - A PRELIMINARY DISCUSSION.

The previous chapters considered the origin and history of the manganese deposits to the stage of the low grade regional metamorphism effecting most of the Central Complex, and the history of the Woolomin Beds was briefly described in the first chapters. The following chapters now consider the effects of contact metamorphism on the rocks, including the manganese deposits, in the study area, and this chapter forms an introduction to this section.

Aspects of contact metamorphism in parts of the Woolomin Beds have been considered, in particular by Spry (1953, 1956), Binns (1965), and Flood (1971). Spry (1953) examined rocks from the Puddledock area, near Armidale, intruded by a quartz monzonite body, and described sediments and volcanics (some spillitic) metamorphosed to the hornblende hornfels facies, and in places probably the low pyroxene hornfels facies.* In 1956 Spry described rocks from an area west of the previous area intruded by a monzonite body. Rocks in that aureole again appeared to develop hornblende hornfels facies assemblages. Spry in both cases above noted considerable lime metasomatism of the volcanics, particularly the spillites, with some scapolitisation.

* Facies names used in this text are those of Turner (1968) unless otherwise indicated.

The pelitic rocks in both areas were relatively siliceous and failed to develop cordierite.

Binns (1965) described some mafic rocks from the southwest edge of the area studied in this thesis, with emphasis on the change in composition and texture of amphiboles with increasing contact metamorphic grade. The maximum grade he found in the mafic rocks was hornblende hornfels facies; extensive reference to his work will be made later.

Flood (1971) briefly described the contact aureole of the Walcha Road Adamellite and suggested that again hornblende hornfels facies rocks had developed, although he notes finding some small outcrops of orthopyroxene bearing volcanics north of the study area. The specimens were found in a narrow band of sediments between the Bannalasta and Walcha Road Adamellites, the band being approximately 100 m. wide at that point. He suggests that these pyroxene hornfels rocks may have been produced by the overlapping of the two aureoles, or be fragments of higher grade material brought up by the intrusives.

The above studies and most studies of contact metamorphism suggest that batholiths of adamellite composition develop contact aureoles showing assemblages up to hornblende hornfels facies. The occurrence of pyroxene hornfels facies assemblages in areas intruded by adamellite batholiths appears to be restricted to closely overlapping

aureoles or roof pendants such as described by Loomis (1966), in the Sierra Nevada Batholith.

However understanding of some aspects of the assemblages produced during the contact metamorphism in the study area, particularly of the manganese deposits, requires consideration of the load pressure on the host rocks during the period of contact metamorphism. In the chapter on the geological history it was noted that most of the Central Complex shows evidence of varying degrees of regional metamorphism, the time of maximum intensity of the metamorphism thought to have been associated with intrusion of the Hillgrove Suite, the oldest suite in the New England Batholith (approximately 250 to 270 my. ago, Binns 1966). Leitch (1973) considered that the Woolomin Beds show evidence of varying degrees of low grade regional metamorphism; the southern part, including the study area being in the low part of the prehnite-pumpellyite facies. He considered this metamorphism to be of regional rather than burial nature, as in many cases younger rocks show higher grades of metamorphism than other ones. Lewington (1973) examined the development of the prehnite-pumpellyite facies in the region immediately south of the study area. The development of this facies may be used to estimate the regional temperature and pressure at the time of the regional metamorphism.

Several writers have estimated the temperature and pressure range of the prehnite-pumpellyite facies, from experimental data and natural occurrence Liou (1971), and Bishop (1973) have developed pressure-temperature fields for this facies, based on experimental breakdown of phases, which they feel are compatible with field occurrence. Liou (1971) suggests a temperature range of 250°C - 380°C at 3kb. for the prehnite-pumpellyite facies, with a pressure range of 1.5kb-3.5kb. Bishop (1973) suggests temperatures from 200°C to 310°C and pressures from 3 to 5 kb., with the broadest temperature range at the lowest pressure.

As the metamorphism near the study area is considered to be of the low prehnite-pumpellyite facies, while Leitch (1973) also suggests that the overall regional metamorphism was of a relatively low pressure type (lack of kyanite and abundance of sillimanite in the high grade areas), then the lower temperature and pressure stability range of the prehnite-pumpellyite facies should provide a reasonable estimate. Therefore the load pressure at the time of regional metamorphism was probably 2 to 3 kb., with temperature from 200° to 300°C .

However, as was indicated earlier, the regional metamorphism probably occurred well before intrusion of the New England Batholith Suite, therefore the temperature and pressure estimate provided by the prehnite-pumpellyite facies can only be taken as a maximum developed some time

before the intrusion.

The following chapters will consider the effects of the contact metamorphism in the study area, treating the sediments (including the lithic sandstones), igneous, and manganese rocks separately. Some problems were experienced in sampling the study area due to the more abundant outcrop of some rock types in one area. The large map at the back of the thesis shows the location of samples and their broad rock type. The writer was also hampered to some extent by lack of knowledge of compositional variation in some phases within the area, and by fine grain size of many rocks. The main results from the examination of each lithology will be discussed in the concluding chapter.

CHAPTER 11.

CONTACT METAMORPHISM - THE SEDIMENTS

Sediments in the study area consist of cherts, banded cherts, pelites, and lithic sandstones, as described in earlier chapters. The contact metamorphic effects on the sediments will be considered in three sections: (i) the pelites, silicious pelites, and banded cherts, (ii) cherts and jaspers, (iii) lithic sandstones.

(i) Metamorphic assemblages found in the pelitic and siliceous pelitic rocks are those of the hornblende hornfels facies near the contacts of the adamellites (except the Limbri Leucadamellite) with some evidence of albite epidote hornfels facies further away. Assemblages developed in the rocks are commonly:

- | | |
|---|-----|
| quartz-biotite-plagioclase (minor) | (1) |
| quartz-biotite-muscovite-plagioclase (minor) | (2) |
| quartz-biotite-muscovite-cordierite-plagioclase (minor) | (3) |
| quartz-biotite-cordierite-plagioclase (minor) | (4) |

and more rarely:

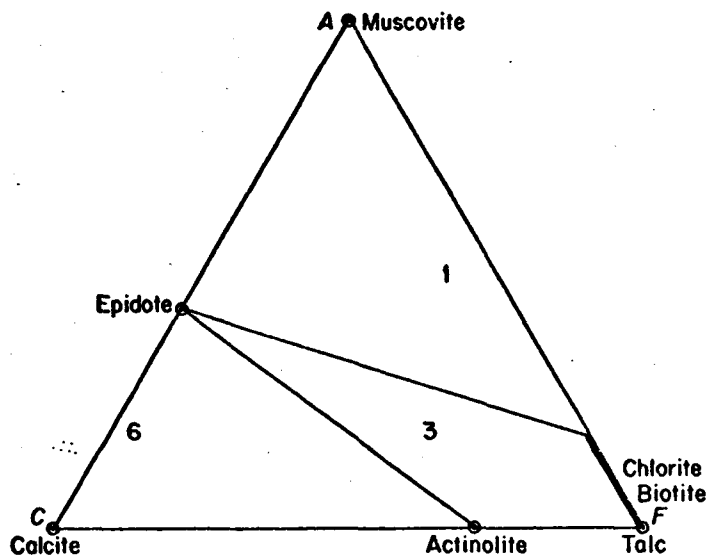
- | | |
|---|-----|
| quartz-biotite-muscovite-andalusite | (5) |
| quartz-biotite-muscovite-cordierite-chlorite | (6) |
| quartz-biotite-muscovite-cordierite-alkali feldspar | (7) |

ACF and A'KF diagrams for pelitic rocks in the albite epidote hornfels facies and hornblende hornfels facies are

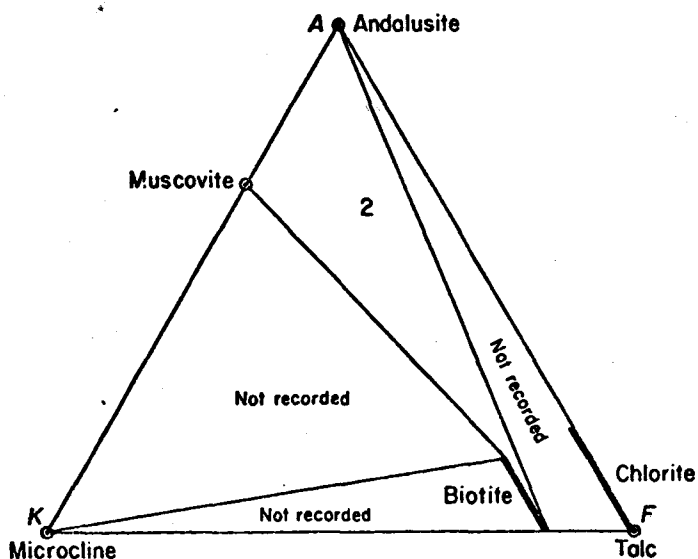
shown in figures 7,8 . Analyses of three pelites and one jasper were made and are shown in table 4 , the first is outside the contact aureole (shows no sign of recrystallisation) and the second and third are cordierite bearing. They are also plotted on the ACF and A'KF diagrams. Reference will be made to these diagrams, the analyses, and the assemblages listed above, in the following discussion.

The first visible sign of metamorphism in the pelitic and banded pelitic rocks (banded cherts) is generally a coarsening of grain size and development of assemblages (1) and (2) listed before. The lowest grade or unmetamorphosed pelitic and banded pelitic rocks in the study area are generally extremely fine grained and often the assemblage in these rocks is not discernable in thin section. In handspecimen the unmetamorphosed rocks appear as grey or grey brown shaly to massive rocks, sometimes showing banding and lighter coloured patches. In thin section these rocks may show essentially parallel fine layering, or show a deformed patchy appearance, (Plate 7a) with relatively quartz-rich and quartz-poor areas.

The appearance of the pelitic rocks as the various contacts are approached varies considerably. It was noted in the regional geology chapter that the contact zone of the Walcha Road Adamellite shows signs of considerable deformation and this is reflected most strongly in the pelitic rocks. As this contact is approached in the south-eastern

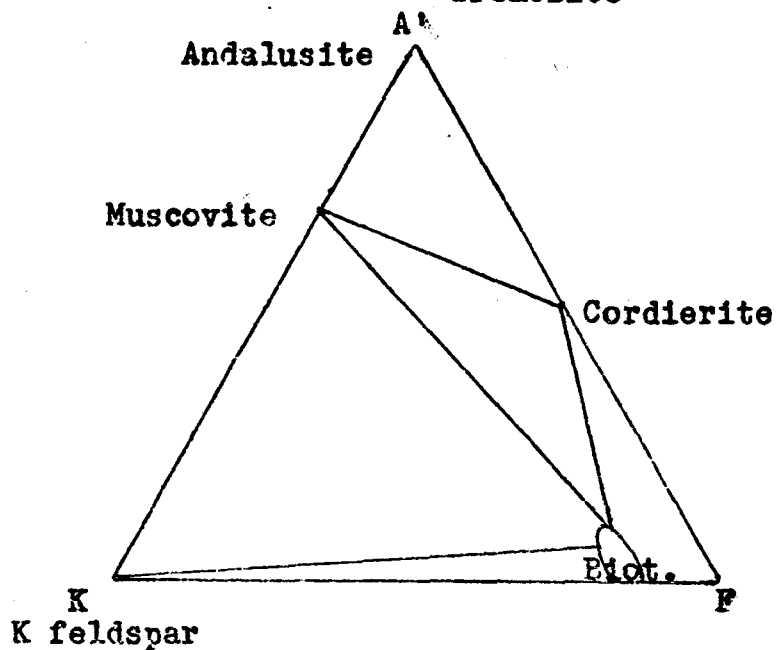
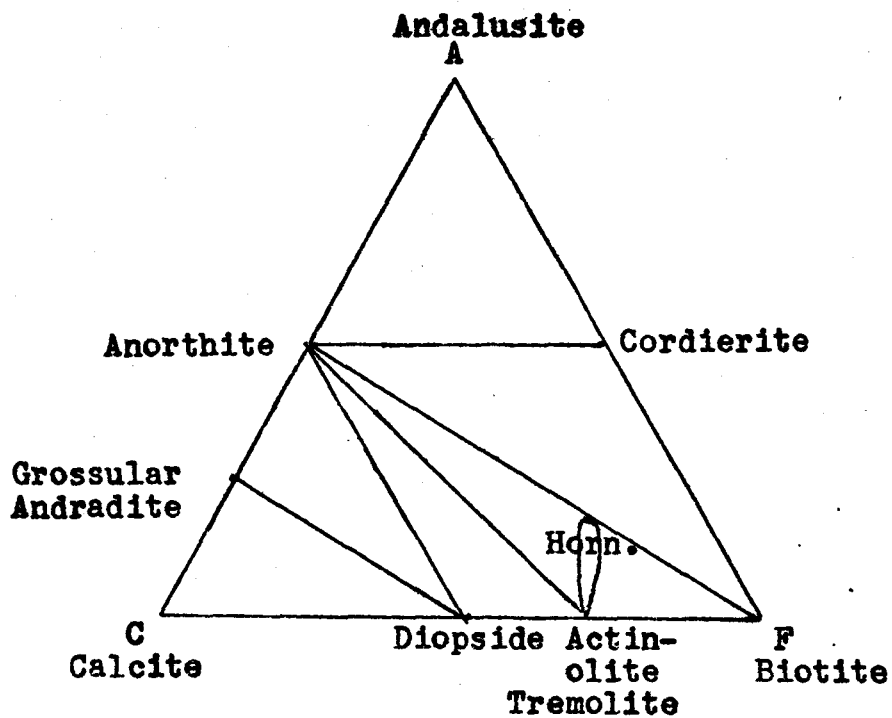


Albite-epidote-hornfels facies. *ACF* diagram for rocks with excess SiO_2 and K_2O . Quartz, albite, and microcline are possible additional phases.



Albite-epidote-hornfels facies. *AKF* diagram for rocks with excess SiO_2 and Al_2O_3 (triangle 1 of Fig. 6-1). Quartz, epidote, and albite are possible additional phases.

Figure 7



HORNBLENDE HORNFELS FACIES

Figure 8

TABLE 4

PARTIAL ANALYSES OF PELITIC ROCKS

SPECIMEN NO	38	12	D6	L2*
SiO ₂	77.40	77.28	73.96	90.07
TiO ₂	0.45	0.39	0.36	0.34
Al ₂ O ₃	10.09	9.31	11.17	2.78
Fe ₂ O ₃ **	4.00	4.19	4.96	3.77
MnO	0.06	0.09	0.06	0.04
MgO	1.29	1.40	1.58	0.08
CaO	0.11	0.04	0.13	0.03
Na ₂ O	0.74	0.15	0.30	0.01
K ₂ O	2.19	2.81	3.96	0.11
P ₂ O ₅	0.07	0.02	0.07	0.01
DIFF.+	<u>3.60</u>	<u>4.32</u>	<u>3.25</u>	<u>2.75</u>
TOTAL	100.00	100.00	100.00	100.00

* L2 is a jasper

** Total Fe as Fe₂O₃

+ DIFF. is % required to make total 100%.

(see appendix on analytical methods)

part of the study area the pelitic rocks develop a phyllitic appearance within approximately 1 km. of the contact, with porphyroblasts becoming visible in the handspecimens within $\frac{1}{2}$ km. of the contact (Plate 1 shows the phyllites enclosing the Danglemah orebody). Within 50 m. of the contact the pelitic and siliceous rocks become schistose and coarsely banded, with the schistosity paralleling the contact. However the pelitic rocks close to the Moonbi and Bendemeer Adamellite contacts are generally dark and relatively coarsely crystalline, developing only a mild schistosity right at the contact with no sign of phyllitic appearance. No pelitic rocks were collected sufficiently close to the Limbri Leucoadamellite to detect any change in texture or appearance.

In this section the first sign of contact metamorphic effect was generally some degree of recrystallisation, however there was no systematic relation between this and distance from the contact, as many of the pelitics were too fine grained to detect the first appearance of muscovite and biotite accurately. The unmetamorphosed pelites appeared to consist of quartz (matrix and radiolaria), clays, minor chlorite and minor plagioclase where coarse enough to tell. Biotite and muscovite were the first recrystallised phases to be definitely identified. The first appearance of cordierite appeared systematically related to distance from the contact and this is

shown in figure 9 , as the cordierite isograd. (The significance of this will be discussed later.) Because of the diversity of form and texture of the recrystallised pelites each significant mineral phase is described below:

Quartz: Occurs as radiolaria and as part of the matrix of the rock, becoming granoblastic to polygonal as the contacts are approached, coarser grained in relatively mica free areas.

Biotite: Occurs as small stubby to larger xenoblastic grains, after oriented parallel to the foliation or bedding. Colour varies from pale yellow brown to red brown, however it was too fine grained to determine the optical orientation of the pleochroism.

Muscovite: Generally has three habits; in low grade and higher grade rocks occurs as fine grains parallel to the foliation (Plate 9a); as fine grains in poikiloblastic cordierite or as fine grains in retrograde(?) porphyroblasts (Plate 10b); as relatively large subidioblastic grains in higher grade rocks (Plate 10a).

Chlorite: Relatively rare, as fine grains in unmetamorphosed or low grade rocks, and as larger grains in chlorite-muscovite patches close to the contacts (Plate 10b).

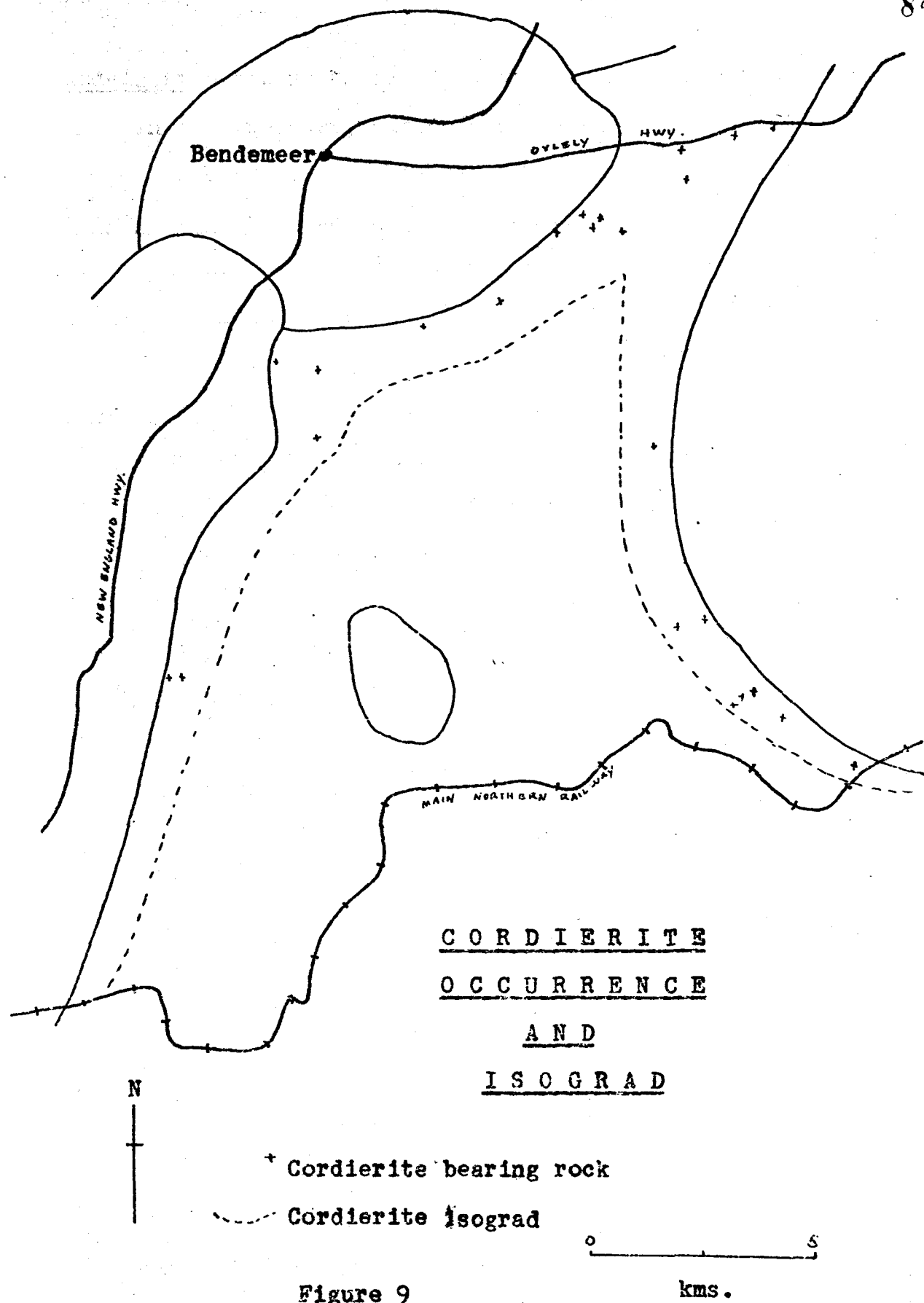


Figure 9

Andalusite: Very rare; as a poikiloblastic porphyroblast in one specimen, and possibly a very small stubby grains in some specimens.

Plagioclase: As moderately scarce small grains in high and low grade rocks; too fine grained to detect composition although it appeared to be approximately andesine.

Alkali feldspar: Detected by staining in some specimens with the cordierite isograd near the Moonbi Adamellite, occurring as very small granoblastic grains generally in the mica poor areas, but within mica rich areas as well. Rocks showing alkali feldspar also contain cordierite and muscovite.

Cordierite: Although it was detected in a relatively large number of specimens its identification was generally difficult as it was often sieve textured or altered to yellow nearly isotropic pinite. Attempts were made to stain for cordierite using amaranth and trypan blue stains, however, these were not successful (see appendix on analytical and optical methods) probably because of slight alteration of the cordierite. The cordierite occurred as relatively large porphyroblasts, always larger than other grains in the rock, and generally ovoid or irregular in shape. Larger porphyroblasts appeared to develop in mica rich fine grained rocks, Plate 8a shows sieve textured

poikiloblastic grains in the quartz rich areas (not visible in the Plate). The maximum size of the cordierite porphyroblasts was 5 mm. In the aureole of the Walcha Road Adamellite cordierite appeared as larger sieve textured porphyroblasts in phyllite with the micaceous folia trending around the porphyroblasts, (Plate 9a and b) and also around a smaller poikiloblastic grain in less foliated slightly coarser grained rock (Plate 8b).

Reference to the ACF and A'KF diagrams in figure for the Albite-epidote hornfels facies suggests that pelitic rocks should develop the assemblages:

Andalusite-muscovite-biotite \pm quartz

or muscovite-epidote-biotite (chlorite) \pm quartz

In field occurrence the assemblages most often developed are:

Quartz - muscovite-biotite-(chlorite)

and Quartz - muscovite-biotite-andalusite

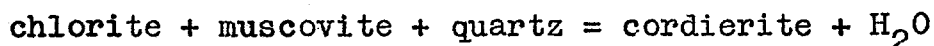
Reference to the corresponding diagrams for the hornblende hornfels facies suggest that the significant changes when developing hornblende hornfels facies conditions are the disappearance of epidote and chlorite and appearance of plagioclase and cordierite. The only one of the changes that appeared to occur systematically in the study area was the appearance of cordierite (shown as the cordierite isograd in figure 9 .) The first appearance of biotite in the

pelitic rocks could not easily be detected due to their fine grain size. The lack of andalusite is probably indicative of low Al_2O_3 and relatively high silica in the pelite rocks, reference to the three analyses in table , and the location of these compositions supports this. The first analysis is of a low grade shaly pelite from the south end of the study area, while the other two are cordierite bearing rocks from close to the Bendemeer and Moonbi Adamellites respectively.

The first appearance of cordierite appears to be the most reliable indicator of metamorphic grade in the pelitic rocks. It is characteristic of the hornblende hornfels facies (Turner, 1968) so that its first appearance should approximately outline the extent of hornblende hornfels development in the pelitic rocks in the study area. The conditions required for the formation of cordierite are dependent on the reaction or reactions forming it. Turner (1968) and Reverdatto (1972) both suggest there is probably a number of relatively complex reactions that form cordierite but that little is known about them. They also suggest however that in rocks of similar composition, as the pelites in the study area appear to be, cordierite should begin to form over a relatively narrow temperature-pressure field. Although experimental data on the formation of cordierite is scarce, Schreyer and Yoder (1964) have attempted to determine the stability of Mg-cordierite, and Turner (1968) has

combined this work with the natural occurrence of cordierites to suggest that it probably begins to form at approximately 450°C at 3 kb. pressure. Therefore the cordierite isograd shown in figure 9 probably approximates to this temperature, bearing in mind the lack of samples of some areas.

It was noted earlier that chlorite occurs in patches with muscovite (Plate 10b) in some rocks above the cordierite isograd, these patches probably represent retrograded cordierite, and are a result of the general reaction:



The presence of subhedral tabular muscovite grains in some cordierite-bearing rocks (Plate 10a) may also be due to retrograde effects, as the large muscovite grains are randomly orientated and not related to the fine-muscovite bearing folia in the rock, and the cordierite is pinitised and mantled by muscovite. These retrograde effects may be due to locally high water pressures at some stage after the initial cordierite forming metamorphism, accelerating the muscovite and chlorite forming hydration reactions. The large tabular muscovite grains may merely represent recrystallised muscovite formed after the peak of metamorphism; the tabular-muscovite bearing rocks were found midway between the Walcha Road and Bendemeer Adamellites on the Oxley Highway, where the contact aureoles probably overlap

considerably (note cordierite isograd) so that the first intrusion may have formed the cordierite, and the second intrusion caused a rise in water pressure which stimulated the recrystallisation of the muscovite.

Binns (1965) noted the occurrence of alkali feldspar with cordierite and muscovite close to the Moonbi Adamellite and its occurrence was also noted in this study. It probably reflects the beginning of the breakdown of muscovite and suggests that upper hornblende hornfels facies conditions exist close to the Moonbi Adamellite. Alkali feldspar was not detected in low grade pelites or near any of the other intrusions.

Pelitic rocks in the study area show evidence of three broad phases of ^{deformation} metamorphism: the first appears to have been soft sediment deformation forming a patchy angular "breccia" of cherty and pelitic areas, of the type shown in Plate 7a and b. The second phase was probably associated with the regional metamorphism and is illustrated by deformed and faulted quartz veins in the unmetamorphosed pelites. The third phase was associated with intrusion of the adamellites, particularly the Walcha Road Adamellite, and appears to have occurred in large part after the contact metamorphism. Most of the cordierite porphyroblasts in the Walcha Road aureole show pelitic folia trending around them, rather than through them, suggesting syn or postkinematic growth of the porphyroblasts (Plate 9a and b). In

particular one rock (shown in Plate 9a) contains porphyroblasts with an internal or preferred orientation at approximately 45° to the external foliation. Although some rotation of the porphyroblasts probably occurred during the deformation, the internal preferred orientation may reflect an earlier foliation at an angle to the later one. The relict foliation may have been caused by the regional deformation, or be an early stage of the intrusive deformation. Deformation after the cordierite had formed suggests that intrusion of the Walcha Road Adamellite continued after maximum contact metamorphism (and therefore probably maximum heat) had developed. The latter phases of intrusion may have been in the solid state.

(ii) Cherts and jaspers: These rocks occur as thin to thick (10 m) interbeds or pods in banded cherts, jaspers being far less common than grey chert. Table 4 includes an analysis of a jasper occurring near the Lonergan manganese mine, and shown in Plate 12a; most of the jaspers contain bands, patches or scattered radiolaria. The cherts usually contain varying amounts of opaques, probably magnetite and carbonaceous material.

Progressive metamorphism of the cherts is shown by increase in grain size to a granoblastic texture near the contacts, and by increasing disruption of the opaque bands and patches by quartz veins. Plate 11 shows a concentric banding structure in a chert near the Bendemeer Adamellite

contact. The outer margins of the banding appear to be related to stylolites present in the rock, and it is suggested (Conaghan, pers. comm.) that the concentric bands were formed by some pressure solution phenomenon associated with the stylolites, causing concentric deposition of the opaque material. It is thought that the structure is not a relict sedimentary feature but probably formed during the contact metamorphism.

The two readily recognisable jaspers collected come from low grade regions of the study area, however a rock collected near the Bendemeer Adamellite contact (80B) is probably a metajasper. It appears in handspecimen as a pink friable saccharoidal rock, and in thin section as a granoblastic aggregate of quartz with areas of euhedral radiating hematite crystals, and some dark brown euhedral grains that are probably biotite. A similar rock was described by Spry (1956) which contained magnetite rather than hematite.

(iii) The lithic sandstones: Lithic sandstones were found relatively widely over the study area, their distribution is shown on the large map. They consist essentially of rock and mineral fragments ranging in size from sand to cherts up to 20 cm. long, in a generally fine grained matrix. The clasts in the samples collected are generally less than 5 mm. long.

The clasts types include quartz, plagioclase (generally andesine), graphic quartz-feldspar intergrowth, quartzite, spillite (?consisting of calcite-epidote-plagioclase), calcarous tuff, pelite, rare basaltic fragments, and no plutonic fragments. Some of the lithic sandstones may be transitional to lithic tuffs. The matrix is generally fine grained and appears to consist of clays, calcite, chlorite (minor), zeolites (?minor), epidote, opaques, and some hematite, although the fine grain size made mineral identification difficult. Plate 13a shows a typical unmetamorphosed lithic sandstone.

The effect of metamorphism is generally shown by formation of biotite or amphibole, sometimes both. The first appearance of these minerals was generally not closely related to distance from an adamellite contact.

Biotite occurred as radiating aggregates (as in Plate 14a, with amphibole) or, as stubby tabular small crystals, (or in Plate 13b) and as irregular rounded to angular plates, especially when forming in quartz rich areas (as in the centre of Plate 13b). The type of amphibole forming in the lithic sandstones was hard to determine (this problem will be discussed in the next chapter) as it generally occurred as acicular to radiating grains and varied in colour from green-brown to green-blue.

In most cases the biotite and amphibole appeared to be forming in the matrix and growing into the clasts, in cases where more mafic fragments were present amphibole was abundant in the fragments (Plate 14b). Biotite was also generally present in the matrix. It was noted earlier that after only biotite or amphibole appeared to be present, and only in some cases both occurred together; in some cases there appeared to be no effect in the lithic sandstones even close to the contact. This variety of response to the metamorphism is almost certainly compositional, the biotite bearing rocks probably having more available potassium, the amphibole bearing rocks have more calcium.

Facies distinctions in the lithic sandstones was generally not possible on the basis of their lithology alone, although one specimen contained the assemblage biotite-chlorite-calcite-amphibole suggesting that the facies there is albite-epidote hornfels and the amphibole is probably an actinolite.

As many of the features of the lithic sandstones are similar to some of the igneous rocks, further discussion in the next chapter will amplify the above discussion.

CHAPTER 12.

CONTACT METAMORPHISM - THE "VOLCANICS"

The nature of the "volcanic" rocks in the study area, and in the Woolomin Beds in general, was discussed in Chapter 1. These rocks represent the dominantly basic volcanism associated with an island arc tectonic environment. The term "volcanic" is not strictly correct as many of the rocks may either be extrusive or intrusive, their textures not being sufficiently diagnostic to tell; the term is used mainly for convenience, and to distinguish these igneous rocks from those associated with the adamellite batholiths. The distribution of the volcanic rocks in the study area is shown on the large map at the back of the thesis. It will be noted that a large number of the volcanic rocks occur in a band extending from south of the Limbri Leucoadamellite, on either side of it, and north to the contact of the Bendemeer Adamellite. Several of the volcanics near the Bendemeer contact were analysed to determine whether an isochemical series existed, these analyses are shown in table 5 (a) and C.I.P.W. norms for them are shown in table 5 (b). The significance of these will be discussed later. A pyroclastic(?) rock from just south of Danglemah railway station was also analysed, the results being shown in table 6 .

It was hoped initially that the volcanic rocks in the study area would provide the most information on the

TABLE 5 (a)

PARTIAL ANALYSES OF BASIC IGNEOUS ROCKS

SPECIMEN NO.	<u>18</u>	<u>306</u>	<u>309</u>	<u>310</u>	<u>311</u>	<u>312</u>
SiO ₂	47.70	48.34	48.52	46.71	49.50	47.77
TiO ₂	2.18	1.90	1.64	2.32	1.43	1.61
Al ₂ O ₃	13.37	13.29	13.20	12.26	14.98	13.92
Fe ₂ O ₃ *	16.72	14.83	15.12	16.97	13.17	12.50
MnO	0.24	0.20	0.24	0.25	0.19	0.18
MgO	4.02	7.04	6.55	6.70	5.15	6.68
CaO	11.40	9.37	10.21	11.25	11.17	13.11
Na ₂ O	2.40	2.95	2.72	2.13	3.18	2.62
K ₂ O	0.25	0.14	0.20	0.19	0.24	0.25
P ₂ O ₅	0.19	0.17	0.14	0.20	0.13	0.12
DIFF. +	<u>1.53</u>	<u>1.77</u>	<u>1.46</u>	<u>1.02</u>	<u>0.86</u>	<u>1.24</u>
TOTAL	100.00	100.00	100.00	100.00	100.00	100.00

* Total Fe as Fe₂O₃

+ DIFF. is % required to make total 100% (see appendix on analytical methods).

TABLE 5 (b)

C.I.P.W. NORMS OF BASIC IGNEOUS ROCKS

Specimen No.	<u>18</u>	<u>306</u>	<u>309</u>	<u>310</u>	<u>311</u>	<u>312</u>
Quartz	2.00	0.00	0.00	0.00	0.00	0.00
Orthoclase	1.48	0.83	1.18	1.12	1.42	1.48
Albite	20.30	24.95	23.00	18.01	26.90	22.16
Anorthite	24.97	22.61	23.22	23.33	25.90	25.49
Diopside	25.52	18.71	21.89	25.77	23.75	31.75
Hypersthene	13.57	18.65	18.64	17.78	10.44	0.91
Olivine	0.00	3.06	1.71	1.95	2.96	9.12
Magnetite	6.06	5.42	5.45	6.15	4.77	4.52
Ilmenite	4.14	3.61	3.11	4.41	2.72	3.06
Apatite	0.44	0.39	0.32	0.46	0.30	0.28
DIFF.	1.53	1.77	1.46	1.02	0.86	1.24
TOTAL	100.00	100.00	100.00	100.00	100.00	100.00
FLAG. AN	55.16	47.54	50.23	56.43	49.05	53.49

(Norms calculated on CO₂ free basis, and assuming FeO to Fe₂O₃ ratio of 3 to 1. see appendix on analytical methods.)

TABLE 6

PARTIAL ANALYSIS OF A PYROCLASTIC ROCK

SPECIMEN NO.	<u>375</u>
SiO ₂	47.74
TiO ₂	1.03
Al ₂ O ₃	13.54
Fe ₂ O ₃ *	6.60
MnO	0.23
MgO	2.87
CaO	11.84
Na ₂ O	4.00
K ₂ O	1.66
P ₂ O ₅	0.26
LOSS * *	8.16
DIFF. ⁺	<u>2.09</u>
TOTAL	100.00

* Total Fe as Fe₂O₃

** Loss at 1100°C for 30 minutes.

+ DIFF. is % required to make total 100% (see appendix on analytical methods).

contact metamorphic conditions. Theoretically the volcanic rocks, especially those of mafic composition, should develop diagnostic assemblages of the various facies, in this case it was hoped to accurately delineate the albite-epidote hornfels facies and hornblende hornfels facies. However it became clear that a major difficulty in doing this was the lack of compositional data of the various phases, especially amphiboles. Although many writers use the terms "hornblende" and "actinolite" freely in describing contact metamorphic facies it became apparent in this study that distinction between these two phases on grounds other than chemical analysis is difficult, and probably inaccurate. This problem will be discussed in detail later; in the descriptive sections the group or non-specific name will be used for phases such as amphiboles, pyroxenes and feldspars, unless some feature suggests that a more specific name is justified.

Because of the diversity of volcanic rock types in the study area each broad type will be briefly described and important aspects discussed in detail, and related to contact metamorphic facies if possible. Conditions of metamorphism suggested by these rock types or associations will be discussed later.

A number of rocks that may be spillitic were found in the southern part of the study area. They consist dominantly of calcite-chlorite-epidote-plagioclase (albite?-

no twinning was apparent) forming a dense pale to dark green rock. Some of the rocks contained clinozoisite rather than epidote, and a pale green fibrous amphibole was also sometimes present. The grain size varied from very fine granular, to rock containing laths or ragged calcite grains (Plate 20a). Three of these rocks (sample Nos. 157, 160 and AX2) contained veins. The vein in 160 consisted of fine granular epidote and larger ragged calcite in quartz and shapeless plagioclase (albite?-no twins visible). A large vein in 157 consisted of fine brown granular clinozoisite and coarser calcite, cut by veinlets of coarse angular ragged calcite, quartz, plagioclase, and axinite. The veins in sample AX2 consist of ragged undulose coarse quartz grains, large (up to 1 cm.) deformed calcite grains and abundant subhedral to euhedral axinite. The appearance of the rock in hand-specimen is striking, as it consists of a dark lime green rock cut by veins of dominantly mauve axinite. This sample is one of a number collected from this locality which contains abundant axinite veinlets, and will be discussed in detail.

The green, spillitic(?) rock in which the axinite veins occur forms an outcrop approximately 20 m. by 10 m. in a hillside consisting dominantly of banded and deformed cherts. Numerous axinite veins (up to 3 cm. wide) and patches (up to 10 m. dia.) cut through the rock, some of the veins contain an asbestiform material that may be fibrous actinolite. (Also in the hillside near the

axinite rich rock are a number of pods or lenses of pyroclastic rock, consisting of calcite pods in a matrix of fine edpidote, chlorite, plagioclase, and minor alkali feldspar; an analysis of one of the rocks is presented in table .

Benson (1915) first noticed axinite in spillitic rocks in the Nundle district, and attributed the source of the boron to metasomatic activity, and Vallance (1966) described axinite in contact metamorphosed rocks near Quéanbeyan, N.S.W., and also favoured an extraneous source for the boron. However Carstens (1965) described axinite occurrence in the Norwegian Caledonides (in rocks very similar to those in the study area) metamorphosed to prehnite-pumpellyite and greenschist grade. He considered that boron mobilisation to form axinite (often associated with edpidote and actinolite) was a normal part of the host rock alteration, the veins consisting of mobilised alteration products, and was not due to any extraneous inflow of boron-rich fluids, ^t the boron being supplied by the volcanism. It seems more likely that the latter explanation accounts for the axinite in the rocks in the study area, the boron mobilisation having occurred with the deformation and low grade regional metamorphism. There is no evidence in or near the axinite bearing rocks to suggest that the boron was introduced by hydrothermal means.

Referring back to the green spillitic(?) rocks in the study area as a whole. The pyroclastic rock analysed

(sample No. 375 Plate 12b) may be a spillitic tuff, and its assemblage (calcite-epidote-chlorite-plagioclase) is almost certainly not due to the contact metamorphism. The presence of minor amphibole in some of these rocks is probably due to the regional rather than contact metamorphism. Two rocks of apparently similar composition were found near the Moonbi Adamellite contact. One (sample No. 353) consisted of subhedral pale brown-green amphibole in patches of calcite-chlorite-amphibole in a groundmass of feldspar. Some traces of clinopyroxene were found in the calcite-chlorite patches, suggesting that these patches may be relict phenocrysts. Although the composition of the groundmass feldspar and amphibole was not known, the mineral assemblage is suggestive of the albite-epidote-hornfels facies. The other rock was found in the north of the study area close to the Moonbie contact (and therefore should be in the hornblende hornfels facies) and had a similar texture to the spillitic tuff; it consisted of ovoid patches of pale green fibrous and radiating amphibole with interstitial plagioclase (An 70) in a mass of disseminated epidote, scattered euhedral sphene and abundant bands of fine opaques. Some patches and bands of fine granular clinopyroxene were also present, it was uncertain whether these were formed by the metamorphism or were primary.

Several alkali feldspar bearing rocks were found in the study area. Sample No. D1 from near the Danglemah

deposit consists of euhedral (often hexagonal) biotite and small patches of calcite-chlorite in an aphanitic matrix of alkali feldspar (Plate 18a). It is uncertain whether this rock is older than the Walcha Road adamellite, it may be a lamprophyric dyke (minette) from the intrusion. A sample (No. 125) from near the Limbri Leucoadamellite showed a relict spherulitic texture, the spherulites consisting of plagioclase, alkali feldspar, and quartz, in a groundmass of the same minerals, with widespread but minor muscovite. This rock probably represents a metamorphosed acid volcanic. It was noted in several samples that relict plagioclase phenocrysts retained strong zoning patterns (Plate 18b), even though the calcic cores had often broken down to fine calcite, epidote, and presumably albite.

Albite-epidote-hornfels facies assemblages appeared to be well developed in some areas, (e.g. sample No. 130, 137) the rocks consisting of two generations of feldspar (after ablated cores in relict phenocrysts) abundant clinozoisite or epidote, chlorite, calcite, biotite and amphibole. The habit of the amphibole varied from small euhedral grains to ragged fibrous patches, to small stubby grains.

Garnet was found only in one specimen (124), near the Limbri Leucoadamellite, where it formed a vein filling with minor epidote, in a rock consisting of opaques,

feldspar and epidote. Its origin may be metasomatic, although it is probably related to veinlets of the type described in the spillitic(?) rock.

The majority of volcanic rocks in the study area appear to be of basaltic composition, and consist dominantly of varying amounts of amphibole, plagioclase, quartz, and clinopyroxene. Very little biotite was found. Clinopyroxene occurs as two generations, as relict grains altering to amphibole (Plates 17b and 16a) and as apparently stable grains in rocks near the adamellite contacts (e.g. Plate 19). Olivine was found in one specimen (Plate 17a) where it was altering to amphibole, as was the pyroxene also present in the rock.

Plagioclase - amphibole was however the main assemblage in the basaltic volcanic rocks. Binns (1965) examined the composition of amphiboles in rocks of this type in the Moonbi contact aureole, basing his results on analyses from four locations, three of them close to the contact and the other approximately 1 km. away. Binns notes a textural change to more granoblastic texture close to the contact, and suggests a colour change as well, from pale blue green ragged amphibole to "a deeper bluish-green variety and then a deeply coloured brownish type with granular habit". He suggests that the alkali content of the amphiboles increases with grade, but that alumina content is relatively constant, indicating that the

amphiboles are true hornblendes, even with the "ragged actinolitic habit". Binns does not include whole rock analysis data so it appears uncertain as to how close the amphibole composition is related to whole rock composition. It was noted in the rocks collected by the writer that over the whole study area amphibole colour varied widely, and could not be related to distance from the contact. Habit of the amphiboles in the study area also could not be related to distance from the contact, ragged, massive, and fibrous types being present at all stages. The analyses and norms in table 12 are of a series of rocks approaching the Bendemeer Adamellite contact, they were made to compare amphibole appearance with whole rock composition. It was noted that no systematic relation existed between the colour of the amphibole and the composition except in rock No. 312 (Plate 21a) which shows signs of calcic metasomatism in the form of diopside-calcite-quartz-blue amphibole (tremolite?)-chalcopyrite veinlets. Here the vein amphibole is considerably more blue than the host rock amphibole. Plates 20b, 21 and 22 show textures of some of these rocks.

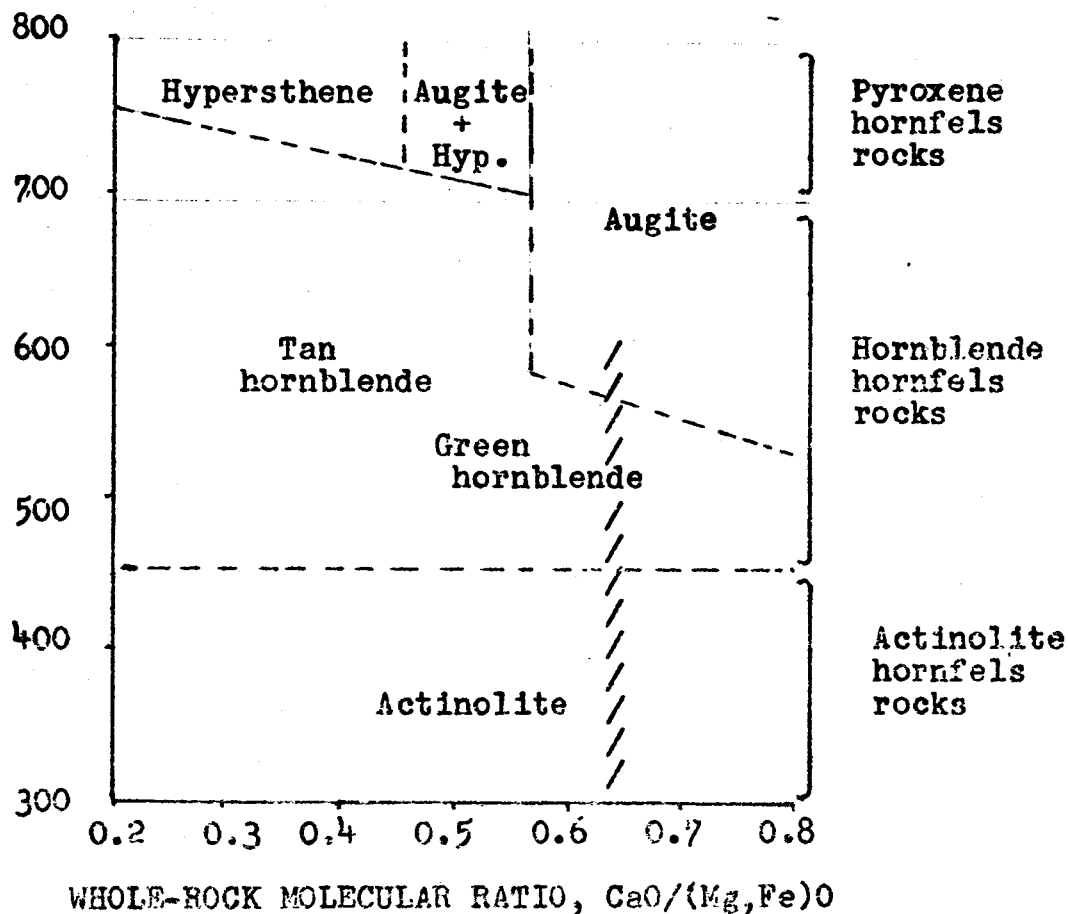
As was noted earlier, definition of the hornblende hornfels facies in the plagioclase amphibole hornfelses has not generally been possible due to lack of data on the composition of the amphiboles. A similar problem was found with clinopyroxenes present in the higher grade rocks

(Plate 21a and 19) it not being possible to distinguish completely between diopsidic and augitic pyroxenes. Pale green pyroxenes present in calcite bearing veinlets near the contacts were assumed to be diopsidic. Clinopyroxenes present in some rocks as part of the overall assemblage (near the contacts) suggest that upper hornblende hornfels facies conditions have been developed.

Loomis (1966) has suggested a temperature composition relationship for mafic minerals in the Mt. Tallac roof remnant in the Sierra Nevada batholith. The assemblages in this remnant are similar to those in the study area, except that the pyroxene hornfels facies is developed. Loomis' diagram is reproduced in figure , and the likely position of the study area mafic hornfels shown on it. This suggests that temperatures of up to 600°C were developed near the contact. Loomis suggests a broad relation between the amphibole colour and the whole rock $\text{CuO}/\text{MgO} + \text{FeO}$ ratio for hornfels in the Mt. Tallac remnant, and this appeared to hold broadly for the amphiboles in the analysed rocks. No tan amphiboles were found in hornblende hornfels rocks, but the amphibole colour varied from brown-green to green and blue green, the $\text{CaO}/\text{MgO} + \text{FeO}$ ratio was high in the analysed rocks.

Lack of time has prevented more detailed examination of the "volcanic" hornfels rocks in the study area, the treatment here being only of a reconnaissance nature.

Figure 10



Estimated temperature-composition relationships in
Mt. Tallac mafic metamorphic rocks.

(from Loomis, 1966)

// Shows approximate position of study area mafic rocks

Given the ability to accurately determine phase compositions, the hornfels in the study area would yield a large amount of information on the development of contact metamorphic facies, particularly the relatively unknown or poorly described albite-epidote-hornfels facies. This facies appears to have been well developed in some parts of the study area.

CHAPTER 13

CONTACT METAMORPHISM - THE MANGANESE DEPOSITS.

Four of the manganese deposits in the study area lie within 1 km. of adamellite contacts, they are the Bendemeer, Danglemah, Bulgunna and Glen Morangie deposits. The two closest of these, Bendemeer and Danglemah, show effects of contact metamorphism.

The mineral assemblages of the various deposits were described in Chapter 3. It was noted that the Lonergan deposit, furthest from any of the adamellite intrusions, contains the same mineral types as the two closest deposits (rhodonite, quartz, tephroite, acmite, carbonate) although the proportions or amount of the various minerals varied. The Danglemah and Bendemeer deposits contained relatively less quartz, and more rhodonite, tephroite and carbonate.

The texture of the Bendemeer and Danglemah deposits was also noted to be different, the assemblages in these two deposits being generally coarser grained with the rhodonite forming large poikiloblastic to skeletal grains with carbonate, and the tephroite forming large subhedral and small granular grains with carbonate. In the areas with quartz and rhodonite there was less difference, the rhodonite forming slightly larger grains and the quartz being more polygonal. Acmite in the Bendemeer and Danglemah deposits was fine granular and disseminated rather than in dense very fine grained masses as at Lonergan. Zoned patches

and bands containing quartz, rhodonite, tephroite, carbonate, and hausmannite were present in the Bendemeer, Danglemah, and Lonergan deposits.

The contact metamorphic effects discussed in the two previous chapters suggest that the Bendemeer and Danglemah deposits lie within rocks showing hornblende hornfels facies assemblages. Temperature predictions were made on the basis of these assemblages. It was suggested that the cordierite isograd indicated temperatures of approximately 450°C , while the presence of clinopyroxene in the mafic rocks near Danglemah suggest that temperatures up to approximately 600°C were probably attained during the contact metamorphism. The lack of pyroxene hornfels facies rocks suggested that temperatures probably did not reach 700°C , although this is dependent on water pressure in the rocks. In Chapter 9 it was suggested that the load pressure at some time before the intrusion of the adamellites was approximately 2 to 3 kb. The Bendemeer and Danglemah deposits were probably effected by temperatures of 500 to 550°C , at a maximum of 3 kb. pressure.

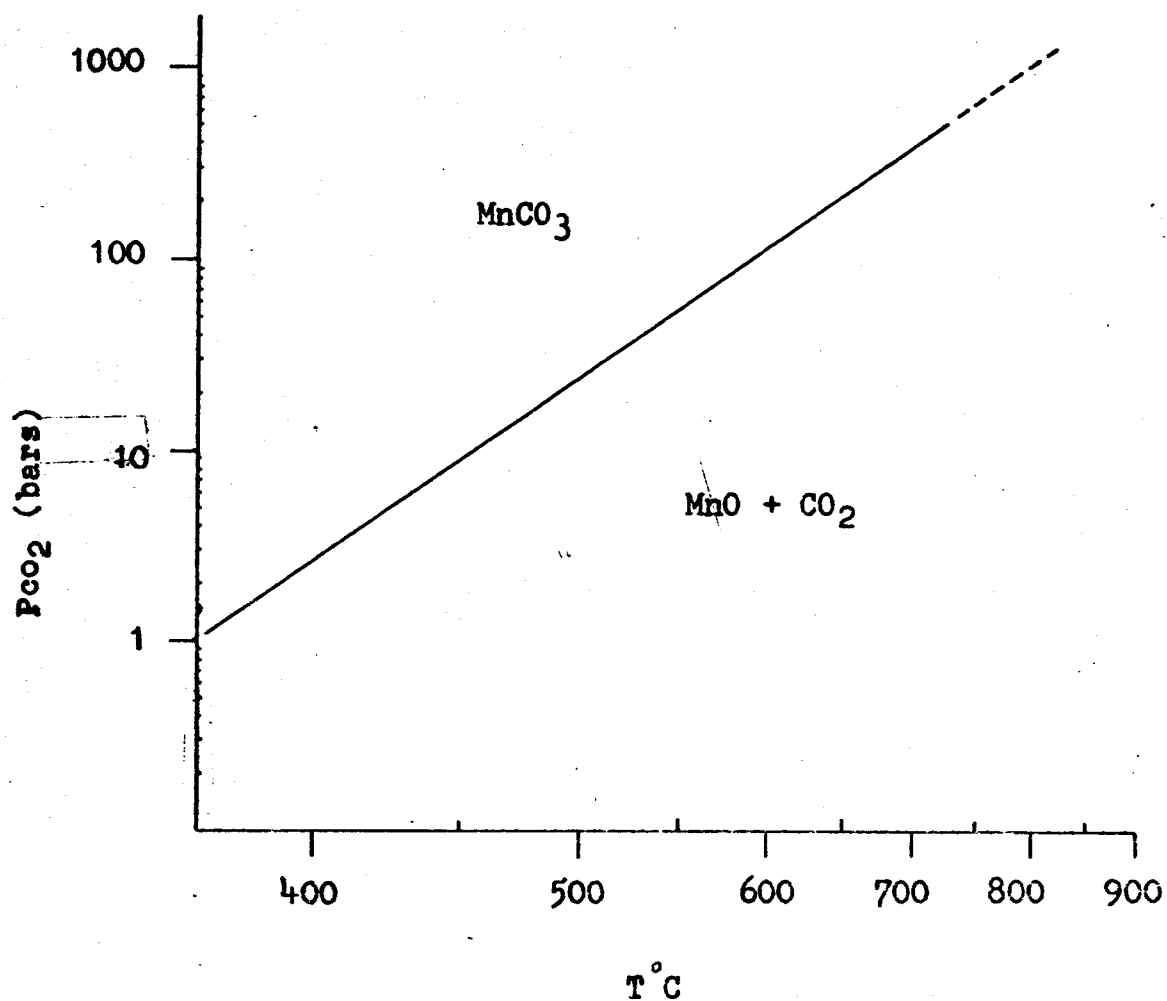
It was noted earlier that the Bulgunna and Glen Morangie deposits are close to intrusions. The Bulgunna deposit is approximately 1 km. from the Limbri Leuco-adamellite and may be in the lower albite-epidote hornfels facies, however it contains essentially only fine grained rhodonite and quartz and does not show any signs of

recrystallisation (i.e. similar texture to the Lonergan deposit.) The Glen Morangie deposit is mainly quartz and manganese oxide, with only very minor rhodonite.

Knowing then that no new phases were found in the metamorphosed deposits (Bendemeer and Danglemah particularly) it must be considered to what extent the textures and assemblages in these deposits are the result of the contact metamorphism. Some of the textures in these two deposits are suggestive of rapid mineral growth, particularly the poikiloblastic to skeletal rhodonite grains (some up to 1 cm. long) occurring with carbonate, and mantling tephroite-hausmannite areas (Plate 3a). It was also noted that quartz was not found with rhodocrosite-rich carbonate in the Bendemeer or Danglemah deposits, (the approximate carbonate composition was determined by x-ray diffraction indicating only whether it was manganese or calcium rich).

Two important studies have been made on oxide components common in manganese deposits. The first of these was carried out by Goldsmith and Graf (1957) on the system $\text{CaO}-\text{MnO}-\text{CO}_2$. They examined the solid solution relationships of CaCO_3 and MnCO_3 and the thermal breakdown of MnCO_3 to MnO and CO_2 . They concluded that a complete solid solution between MnCO_3 and CaCO_3 exists only above 550°C at CO_2 pressures sufficient to prevent decomposition, below this temperature a miscibility gap exists near the Mn end. On the basis of experimental results they produced the

Figure 11

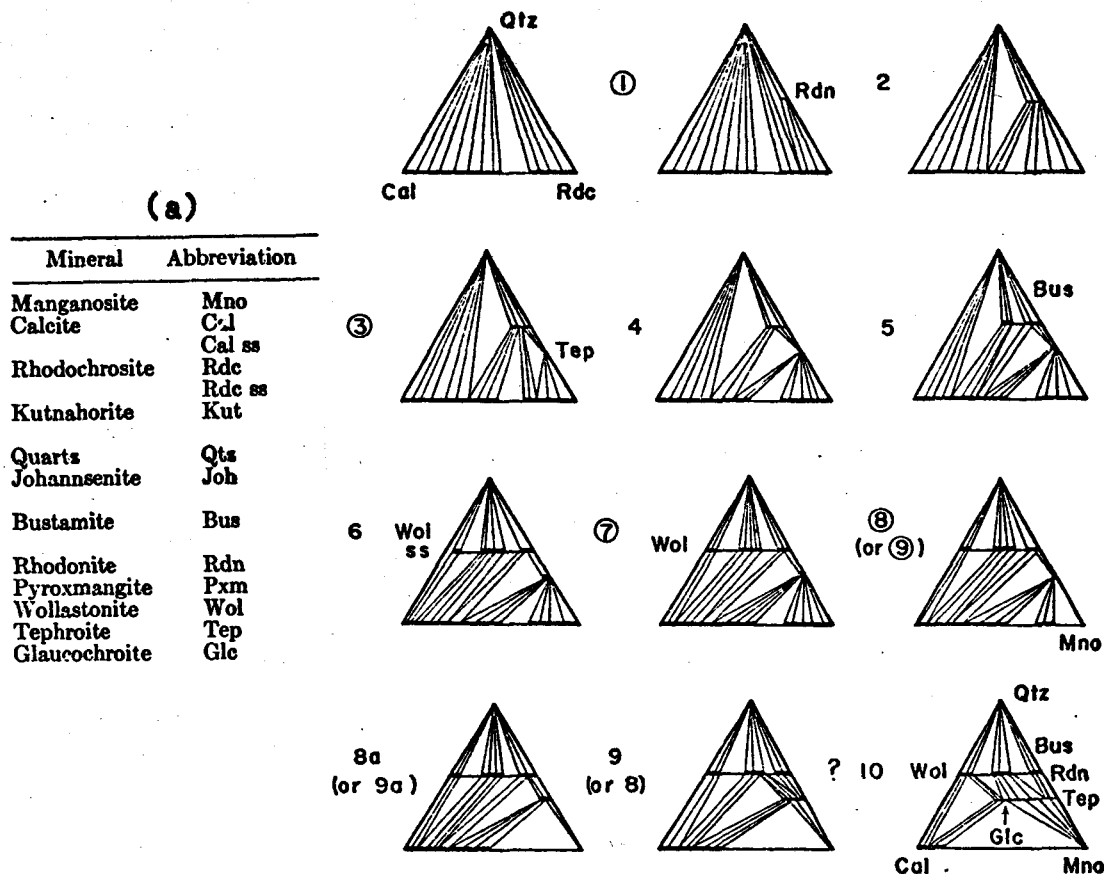


Equilibrium thermal decomposition of MnCO_3
(Goldsmith and Graf, 1957)

breakdown curve for MnCO_3 shown in figure 11 . The important factor here is that only at low P.CO_2 levels (50-60 bars) would MnCO_3 be unstable at approximately 500°C , the suggested metamorphic temperature for the two deposits. It was noted in the chapter on the volcanic contact metamorphism that the P.CO_2 was probably high near the contacts, as indicated by the lack of wollastonite and presence of calcite-diopside-quartz veinlets near the contacts.

Burt (1972) examined the system $\text{CaO-MnO-SiO}_2\text{-CO}_2$ using data from natural occurrences of assemblages and experimental data on behaviour of phases. Figure (a) lists the minerals present in this system. Burt formulated a decarbonation sequence for the system, and this is shown in figure (b) for about 500°C , starting with calcite, rhodocrosite and quartz. The reactions involved in this process are shown in figure (c). Burt also suggests a schematic P-T diagram for the system, shown in figure . The P.CO_2 dependent nature of these reactions is emphasised.

The applicability of this system ($\text{CaO-MnO-SiO}_2\text{-CO}_2$) to the manganese deposits is close, and by examination of the phases in the deposits they may be related to the system. It was noted earlier that rhodocrosite-rich carbonate and quartz don't occur together in the Bendemeer or Danglemah deposits, suggesting that reactions (1) and (2) shown in the figures probably have gone to completion. The poikiloblastic nature of the rhodonite is probably the result of this



(b) Progressive decarbonation of a calcite-rhodochrosite-quartz mixture at about 500°C. Numbers refer to decarbonation reactions ; circled numbers identify reactions that have been studied experimentally.

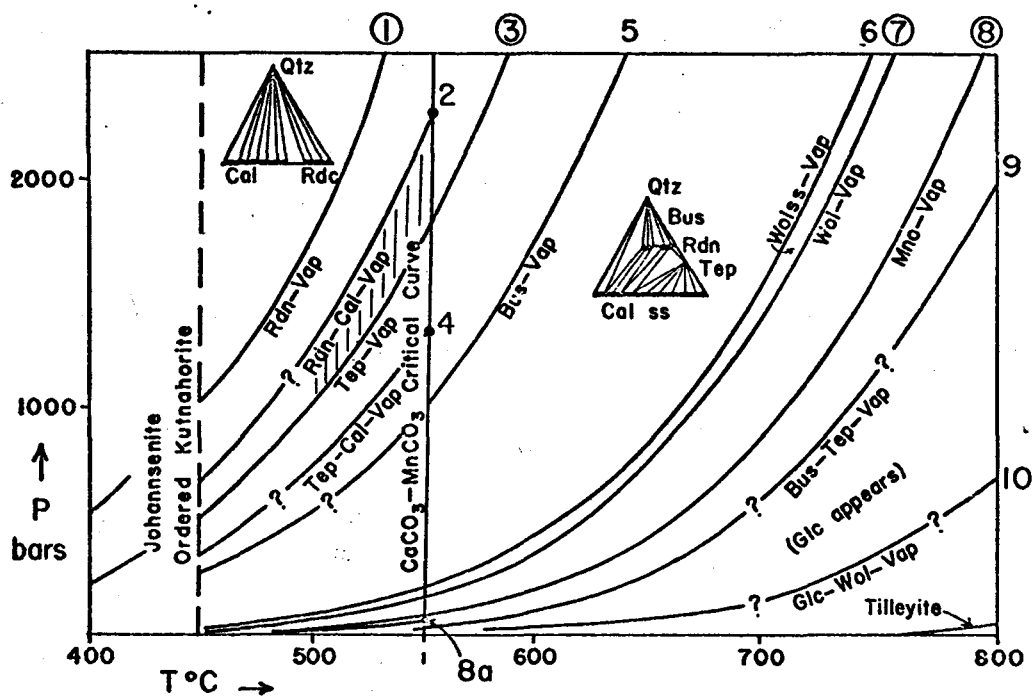
(c) Decarbonation Reactions in the System $\text{CaO-MnO-SiO}_2\text{-CO}_2$

	Experimental Studies
1. $\text{Rdc} + \text{Qtz} = \text{Rdn} + \text{Vap}^*$ $\text{MnCO}_3 + \text{SiO}_2 = \text{MnSiO}_3 + \text{CO}_2$	Peters, 1971; Huebner and Eugster, 1968
2. $\text{Rdc ss} + \text{Qtz} = \text{Rdn} + \text{Cal ss} + \text{Vap}$	
3. $\text{Rdc} + \text{Rdn} = \text{Tep} + \text{Vap}$ $\text{MnCO}_3 + \text{MnSiO}_3 = \text{Mn}_2\text{SiO}_4 + \text{CO}_2$	Huebner and Eugster, 1968
4. (or 5) $\text{Rdc ss} + \text{Rdn} = \text{Tep} + \text{Cal ss} + \text{Vap}$	
5. $\text{Rdn} + \text{Cal} + \text{Qtz} = \text{Bus} + \text{Vap}$	
6. $\text{Cal} + \text{Qtz} + \text{Bus} = \text{Wol ss} + \text{Vap}$	
7. $\text{Cal} + \text{Qtz} = \text{Wol} + \text{Vap}$ $\text{CaCO}_3 + \text{SiO}_2 = \text{CaSiO}_3 + \text{CO}_2$	Greenwood, 1967b; Harker and Tuttle, 1956
8. (or 9) $\text{Rdc} = \text{Mno} + \text{Vap}$ $\text{MnCO}_3 = \text{MnO} + \text{CO}_2$	Goldsmith and Graf, 1957; Huebner, 1969
8a. (or 9a) $\text{Rdc ss} = \text{Mno} + \text{Cal ss} + \text{Vap}$	
9. (or 8) $\text{Rdn} + \text{Cal ss} = \text{Bus} + \text{Tep} + \text{Vap}$ (Glaucochroite appears)	
10. $\text{Bus} + \text{Cal} = \text{Wol} + \text{Glc} + \text{Vap}$	

* Vap, vapor;

(full references in Burt, 1972)

Figure 12



Schematic P - T diagram of the system CaO-MnO-SiO_2 , in the range 0 to 2.5 kb, 450°–800°C.

(from Burt, 1972)

Figure 13

reaction, the original assemblage being mainly calcite and quartz. The next reaction in the sequence, reaction (3), between rhodocrosite and rhodonite probably hasn't occurred as these two minerals appear stable together. The tephroite rich area probably represent more manganese rich patches, the tephroite having formed from manganese oxide plus silica, there not being enough silica available to form rhodonite (as discussed in Chapter 9). No bustamite was found in the deposits, suggesting also that reaction (5) has not occurred.

Using the above constraints (i.e. that reaction 2 occurred, but that reaction 3 didn't) the Bendemeer and Danglemah assemblages occupy the lined area on figure , assuming the temperature to have been between 500 and 550°C. Bearing in mind the schematic nature of this diagram the $P.CO_2$ was probably between 1000 and 2300 bars. This is in accord with the high $P.CO_2$ suggested from the volcanic assemblages discussed earlier.

The lack of knowledge of the compositions of the phases in the deposits (especially the carbonates) limits the investigation of the metamorphic effects involving solid solution phenomena. If evidence was found for a complete solid solution having existed in the carbonate, this would suggest the temperature attained during the metamorphism was greater than 550°C; the pyroxmangite content of the rhodonite is also thought to be an indicator of pressure; the higher the pressure the more pyroxmangite is present, until

at approximately 20 kb. pyroxmangite is the stable phase (Burt, 1972). Further investigation along these lines should provide more data on the contact metamorphism of the manganese deposits.

THE CONTACT METAMORPHISM - CONCLUSION

The contact metamorphic rocks in the study area, including the manganese deposits, have been briefly examined in the three preceeding chapters.

The sedimentary rocks in the area showed varying response to the metamorphism. The pelitic and siliceous pelitic rocks, while not showing much effect at low grades, developed cordierite on a relatively widespread scale allowing the broad delineation of the cordierite isograd shown in figure . It was suggested that the cordierite isograd probably represents a temperature of approximately 450°C at 2-3 kb. (the maximum pressure likely to have been prevailing in the region). The development of alkali feldspar near the contact of the Moonbi Adamellite suggests that upper hornblende hornfels conditions were achieved during the metamorphism. The nature of the cordierite porphyroblasts in the Walcha Road aureole suggests that the maximum temperature of metamorphism was probably achieved before the deformation associated with the intrusion had ceased. The banded cherts and massive cherts showed little response to the metamorphism, apart from forming granular quartzites, while the jaspers appeared to form a pink granular quartzite near the contact. The lithic sandstones were characterised by the development of amphibole, biotite, or both, at varying distances from the contact.

A wide range of "volcanic" rocks was found in the study area, and these showed to varying extents the development of hornblende hornfels and albite-epidote hornfels facies assemblages. The study of these rocks was hampered by lack of knowledge of the composition of key phases in rocks, especially amphiboles and feldspars. It was suggested that there was little relation between the habits of amphiboles and their composition or metamorphic grade, and that amphibole colour is dependent more on bulk rock composition rather than metamorphic grade. Temperatures in the order of 600°C were suggested by the development of clinopyroxenes in some hornfels near the contacts, while some calcic metasomatism is suggested by the presence of diopside-calcite-quartz-tremolite veinlets (and reflected in the analysis of rock No. 312, in table 5 (a), and shown by the high normative diopside and olivine in the norm for the same rock). The analysed mafic rocks were high calcium-low potassium basalts, developing the assemblage hornblend-plagioclase near the contact.

The two main metamorphosed manganese deposits, Danglemah and Bendemeer, showed signs of the development of a decarbonation series as described by Burt, 1972. The stability of rhodonite-carbonate and lack of the assemblage carbonate-quartz suggested that a $P.CO_2$ of approximately 1000 to 2300 bars was present at the time of metamorphism. This is in accord with the calcium metasomatism suggested in the volcanic metamorphics.

It is concluded that the metamorphism in the study area was of similar intensity to that described for most comparable adamellite or acid batholiths. The overlapping of the metamorphic aureoles is reflected by the joining of the cordierite isograds in the north of the study area, however no pyroxene hornfels facies assemblages were detected in the study area.

APPENDIX I

OPTICAL AND ANALYTICAL METHODSOptical Methods:

Optical investigations were carried out on Leitz binocular normal stage microscopes. Optic axial angles were estimated and optic signs determined from interference figures, universal stage methods were not used.

Plagioclase feldspar composition was estimated from maximum extinction angles of albite twins in sections normal to (010).

Staining for alkali feldspar was carried out using sodium cobaltinitrite solution. Staining for cordierite was attempted without success, the methods used being those of Boone and Wheeler (1968), both cobaltinitrite-amaranth and trypan blue stains were used. The lack of success was probably due to slight secondary alteration of the cordierites.

Analytical Methods:

Chemical analyses on rocks and minerals were performed using a combination of x-ray fluorescent and wet chemical techniques. The major oxides SiO_2 , TiO_2 , Al_2O_3 , total iron as Fe_2O_3 , MnO , CaO , MgO , K_2O , and F_2O_5 were determined by x-ray fluorescence following the method of Norrish and Chappell (1966). The x-ray machine used was a Siemens S.R.S. spectrometer with a 4 kilowatt generator and standard counting rack. Relevant U.S.G.S. rock

standards were run with unknowns.

Sodium was determined using standard flame photometric methods with closely spaced artificial standards.

Ferrous iron determinations were not carried out, however for calculations of C.I.P.W. norms arbitrary Fe^{2+} to Fe^{3+} ratios were used, based on known ratios in similar rock types.

In most cases loss on ignition determinations were not made. The word "DIFF." shown in most analyses is the difference between the percentage total of major elements and the theoretical whole rock total of all components, 100%. It is therefore an estimate of the total volatiles, minor and trace elements present.

A Siemens x-ray diffractometer was used for mineral identification.

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Norrish, K., and B.W. Chappell: X-ray Fluorescence Spectrography, in "Physical Methods in Determinative Mineralogy," Academic Press, London, 1966.

APPENDIX II

ACMITE

Acmite was detected in the manganese silicate-oxide rocks from the Lonergan manganese deposit. It was first identified as an aegirine-augite, on the basis of x-ray diffractometry using powdered rock mounted in "Tarzan's Grip" on a glass slide. As the composition of the aegirine-augite series effects x-ray diffractograms only in peak intensity rather than shifting the peaks, the unknown material was again subject to x-ray diffraction, using a compressed powder mount to prevent preferred orientation of the crushed material. The material was then identified as an acmite, using the A.S.T.M. card index. The acmite was then analysed by x-ray fluorescence techniques as described in Appendix I, and the results are shown in Table 7. It was known at the time of analysis that the sample contained some quartz, and this is reflected in the high SiO_2 content in the first analysis in the table. The "corrected" analysis was obtained by reducing the SiO_2 content to 52% (the average SiO_2 content for acmites) and distributing the 5% excess silica among the other components, thus giving a better idea of the true composition of the acmite. The high manganese content suggests that the mineral should more correctly be called a "blanfordite", however this name was not used in the text as the acmite identified (by x-ray

diffraction) in the Danglemah and Bendemeer deposits was not analysed. The low total of the analysis may be due to a high content of minor elements in the mineral, as acmite pyroxenes are known to be able to take up considerable amounts of minor elements. Suitable U.S.G.S. standards were run and the x-ray machine re-calibrated just before the acmite analysis was carried out.

TABLE 7

ACMITE ANALYSIS

	Original	Corrected*
SiO ₂	57.01	52.00
TiO ₂	0.14	0.16
Al ₂ O ₃	0.61	0.68
Fe ₂ O ₃ **	21.06	23.59
MnO	5.38	6.03
MgO	0.34	0.38
CaO	0.49	0.55
Na ₂ O	9.88	11.07
K ₂ O	0.03	0.04
P ₂ O ₅	0.03	0.04
L.O.I ***	<u>0.60</u>	<u>0.60</u>
	95.57	95.14

Refer to appendix on analytical methods for full explanation of above analysis.

* corrected for 5.01% quartz impurity

** Total Fe as Fe₂O₃

*** Loss at 1100°C for 30 minutes.

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PLATES

Abbreviations used in plate explanations :

S.N. = Specimen Number

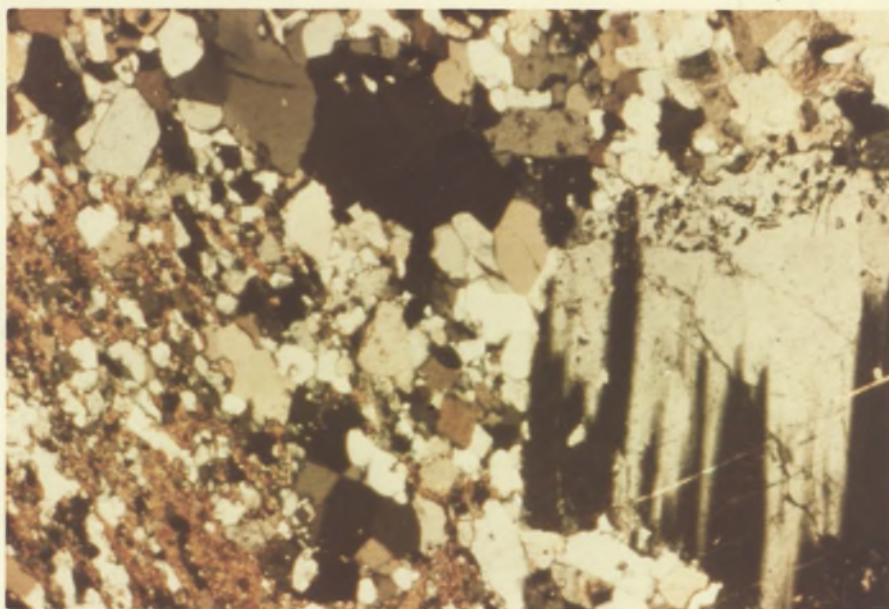
P.P. = Plane Polarised

C.N. = Crossed Nicols



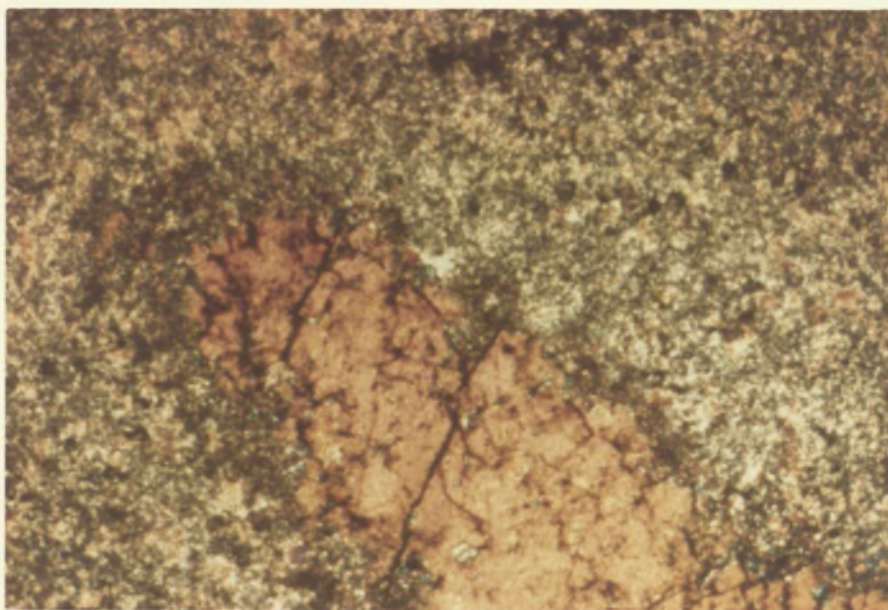
The Danglemah Orebody.

(a)



Danglemah deposit: plagioclase, quartz, and acmite.
(S.N. D4) C.N., x 100

(b)



Bendemeer deposit: porphyroblast of tephroite in
rhodonite and quartz. (S.N. MN7) C.N., x 100

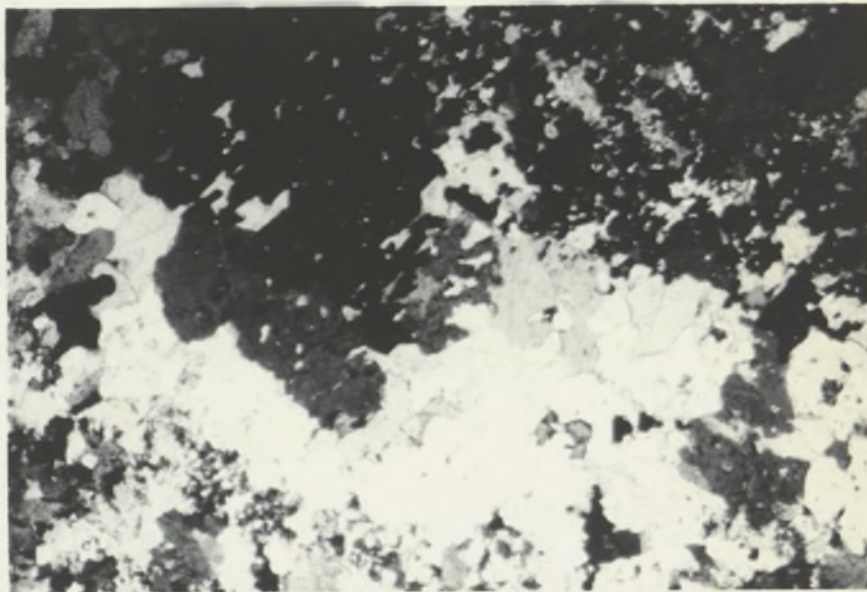
(a)



OCT 73

Bendemeer deposit: zoned patch, quartz to rhodonite to rhodonite + carbonate to tephroite + hausmannite (S.N. NM5) P.P., x 130

(b)

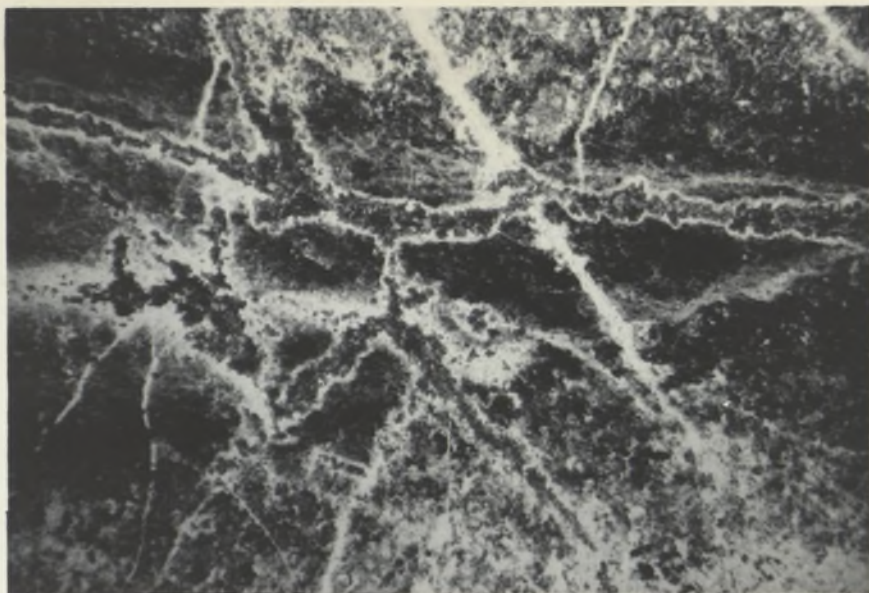


OCT 73

Bendemeer deposit: poikiloblastic rhodonite grains in carbonate. (S.N. NM 2) C.N., x 100

(a)

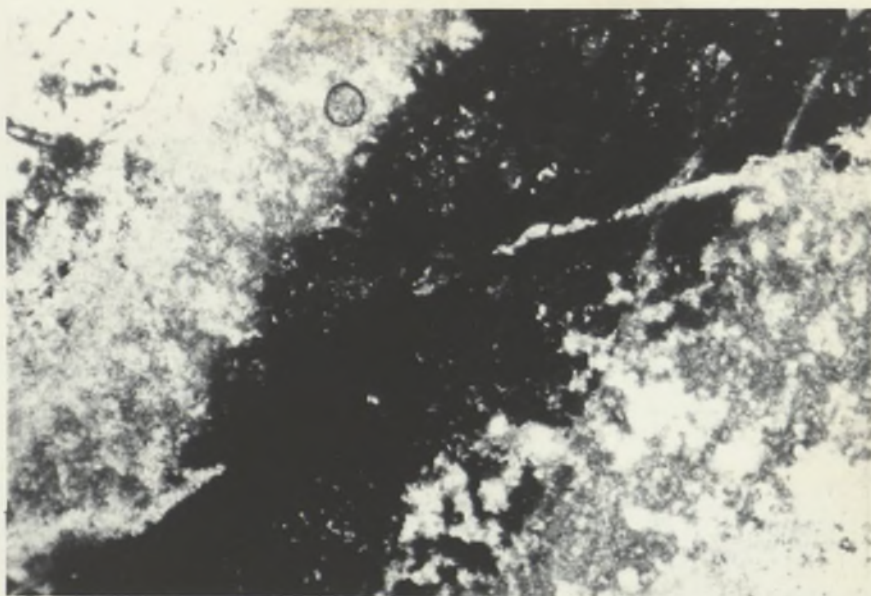
OCT 73



Dulgunna deposit: rhodonite-quartz bands cutting fine oxide quartz rock. (S.N. BUL 5) P.P., x 130

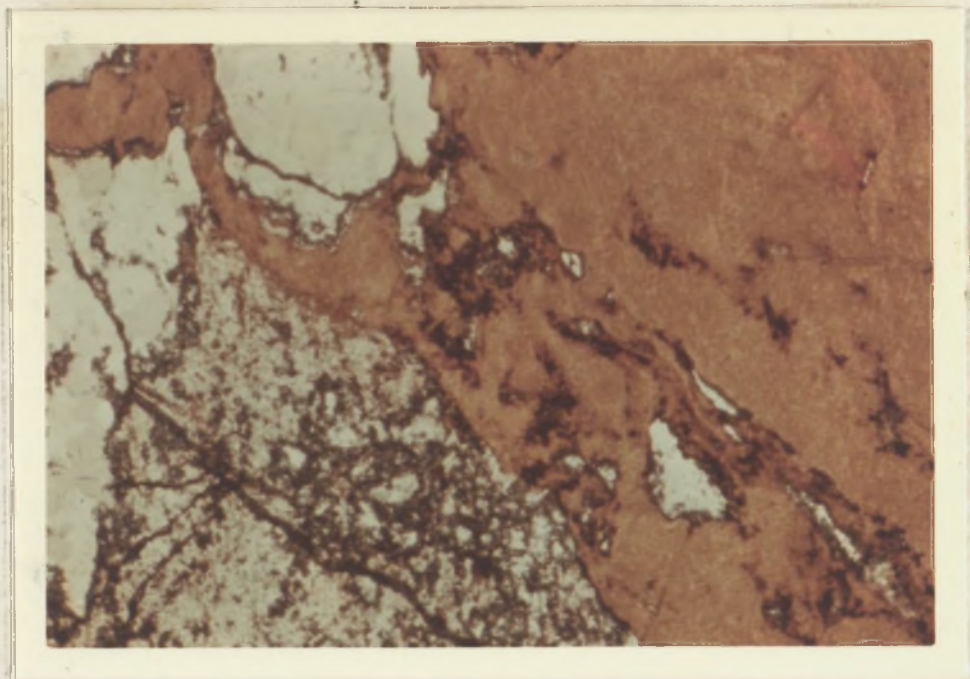
(b)

OCT 73



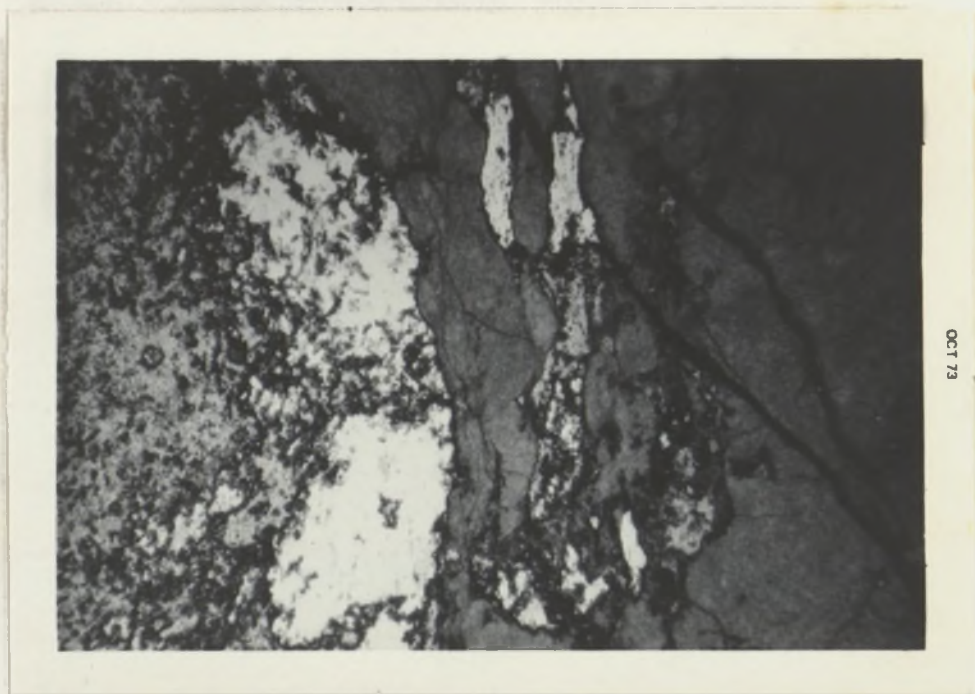
Loneragan deposit: hausmannite-tephroite band surrounded by rhodonite and quartz (S.N. L1-3) P.P., x 130

(a)



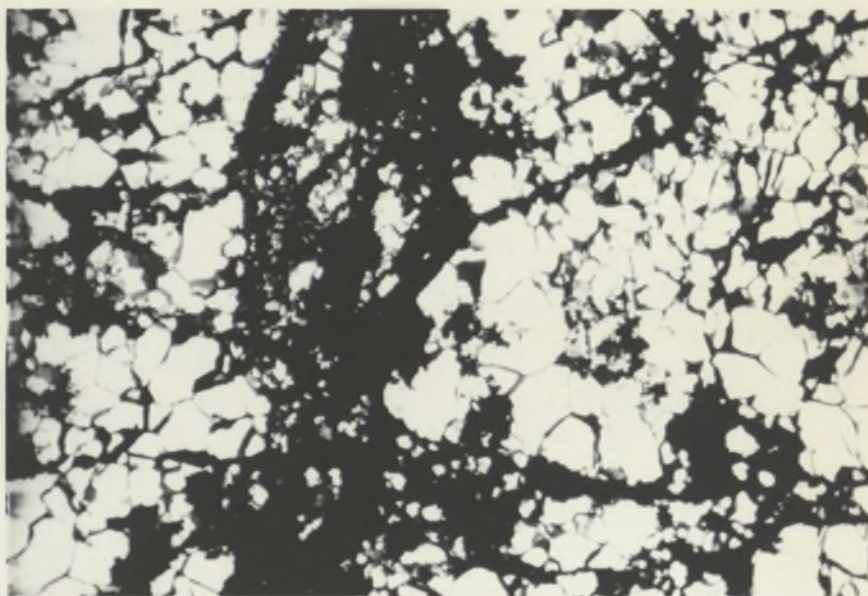
Loneragan deposit: edge of acmite band, in quartz and rhodonite. (S.N. L2-1a) P.P., x 100

(b)



Loneragan deposit: as above, quartz at edge of acmite, with rhodonite surrounding. (S.N. L2-3a) P.P. x 130.

OCT 73



OCT 73

Blaaoven deposit: manganese oxide in quartz.
(S.N. 156) P.P., x 130

(a)

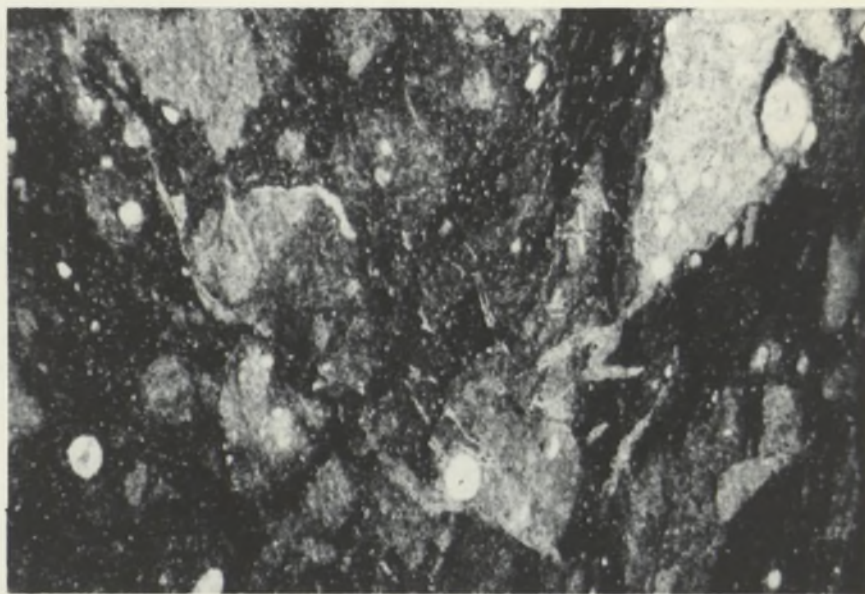
OCT 73



Radiolaria and quartz rich patches in deformed pelitic rock
(S.N. 180) P.P., x 130

(b)

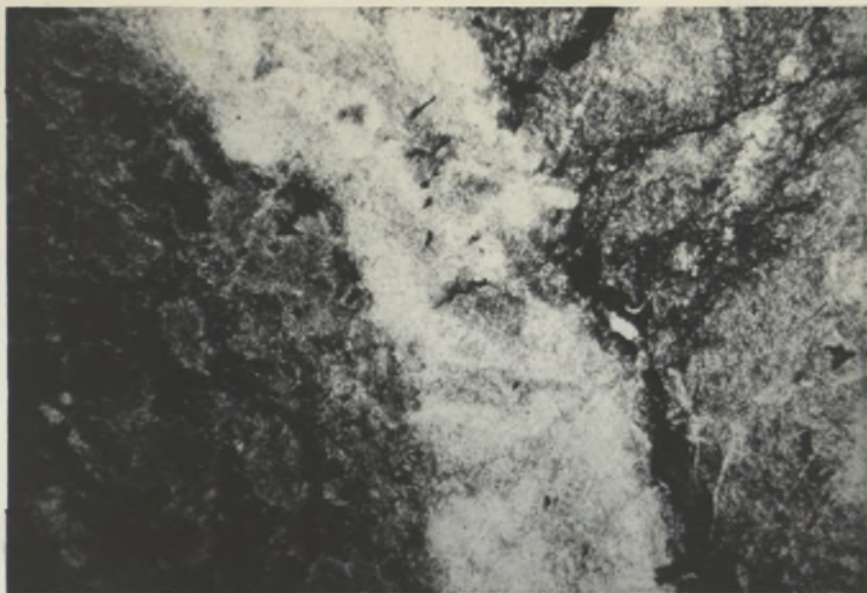
OCT 73



Pelite showing sedimentary brecciation forming quartz and
biotite rich patches. (S.N. 67) P.P., x 100

(a)

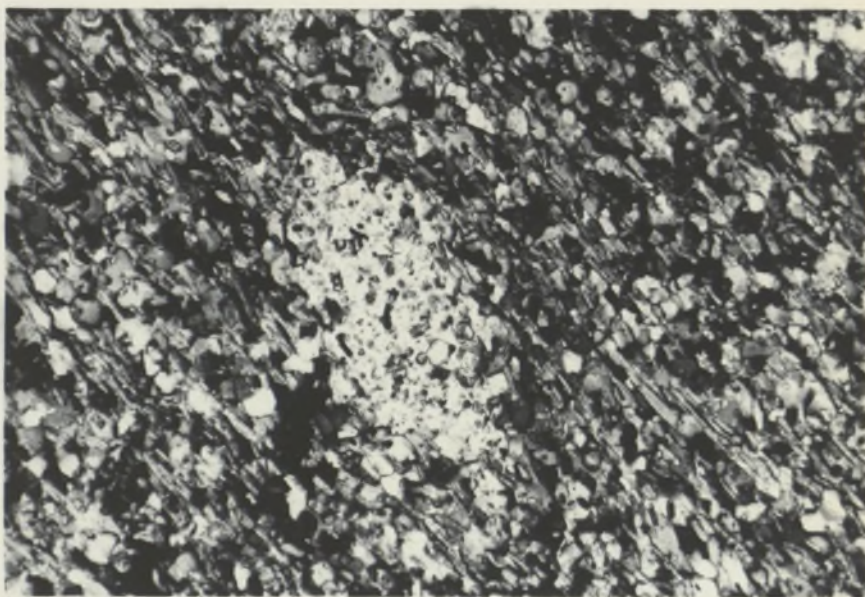
OCT 73



Banded Chert showing cordierite porphyroblasts in pelite area, right-hand side. (S.N. 7) P.P., x 130

(c)

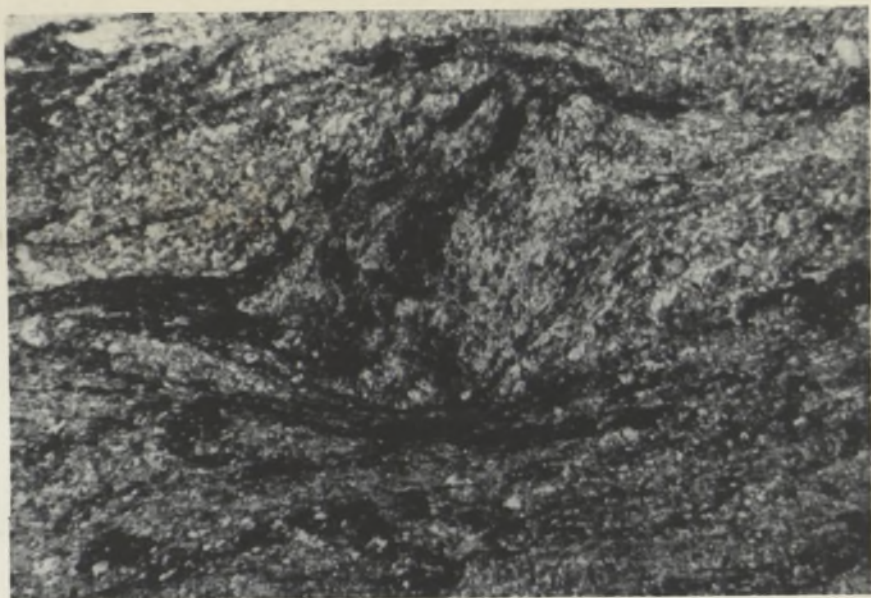
OCT 73



Cordierite porphyroblast in metapelite (S.N. 27) P.P., x 400

(a)

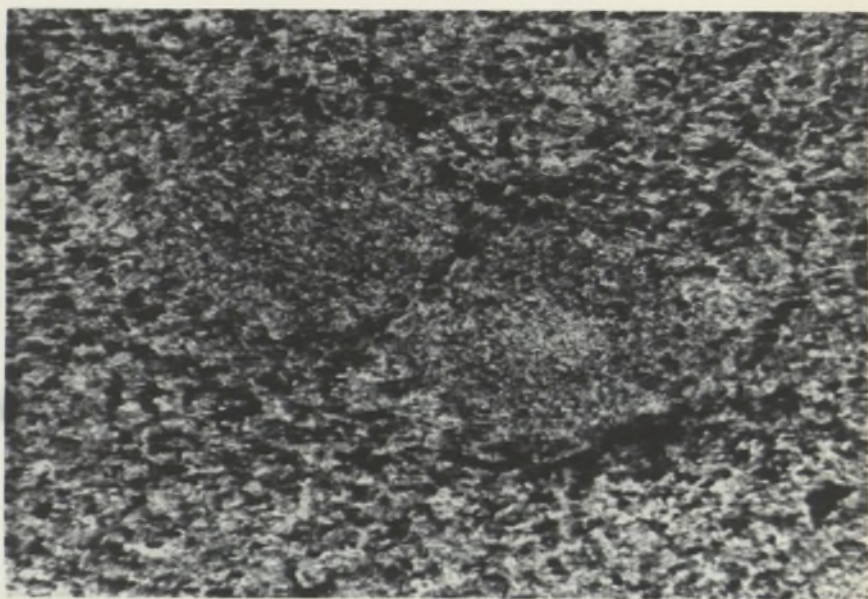
OCT 73



Cordierite porphyroblast, note internal foliation at approx. 45° to surrounding foliation. (S.N. 180)
P.P., x 400

(b)

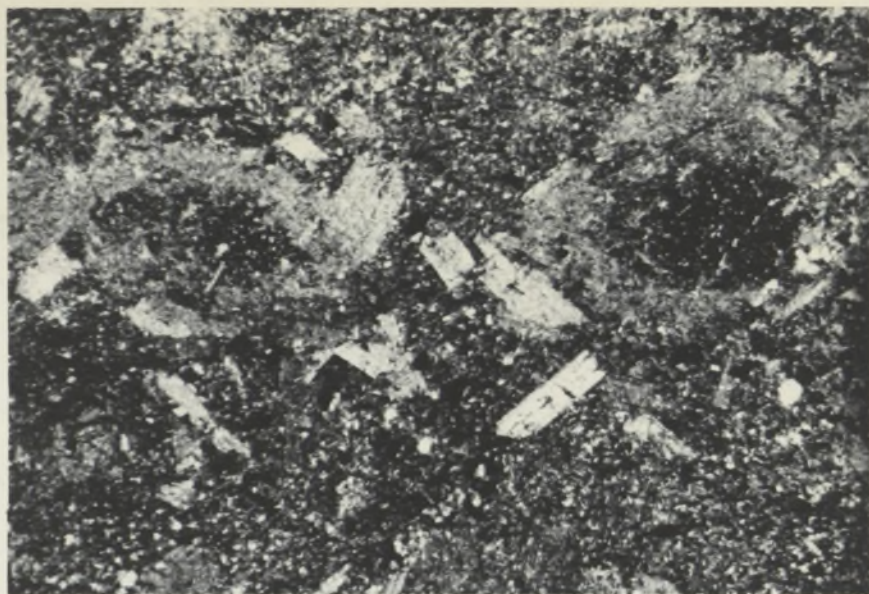
OCT 73



Cordierite porphyroblasts in knotted phyllite
(S.N. D6), P.P., x 400

(a)

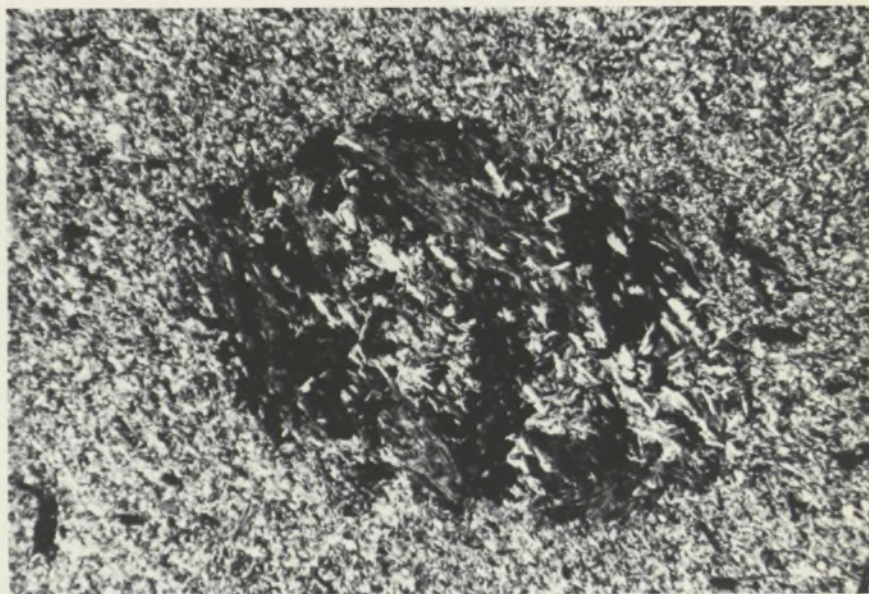
OCT 73



Large tabular muscovite grains and fine muscovite mantling pinitised cordierite porphyroblasts. (S.N. 52B), C.N., x 100

(b)

OCT 73



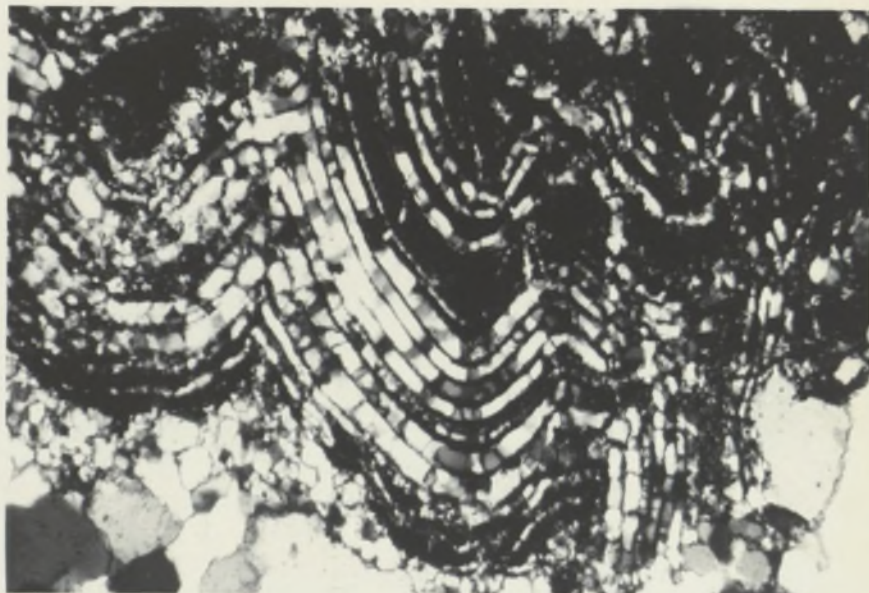
Muscovite-chlorite patch (retrograded cordierite?) in metapelite. (S.N. 190), C.N., x 400

NOT AVAILABLE DUE TO
PROCESSING ERROR

(a)

Concentric banding of opaques in metachert.
(S.N. 55a), P.P., x 130

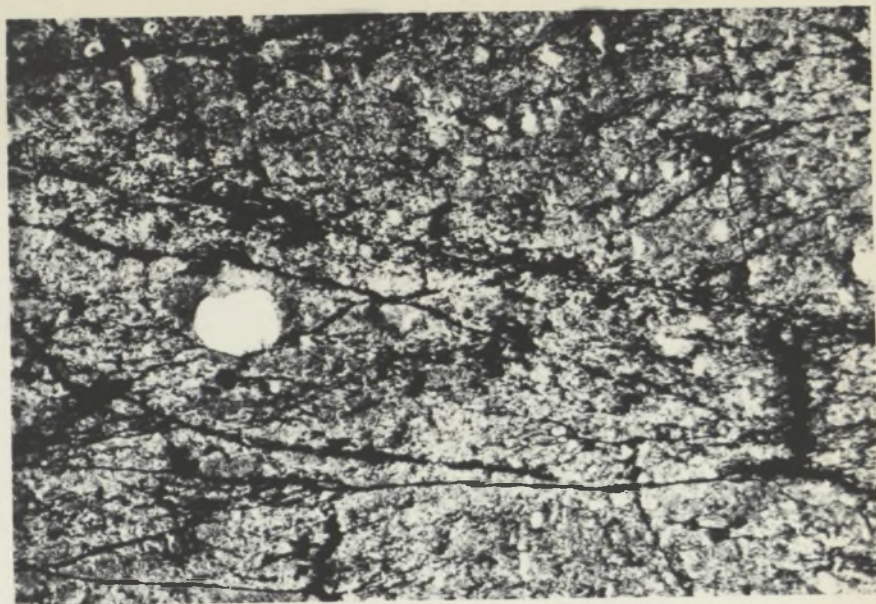
(b)



OCT 73

Detail of banding showing quartz grain shapes
(S.N. 55a), C.N., x 400

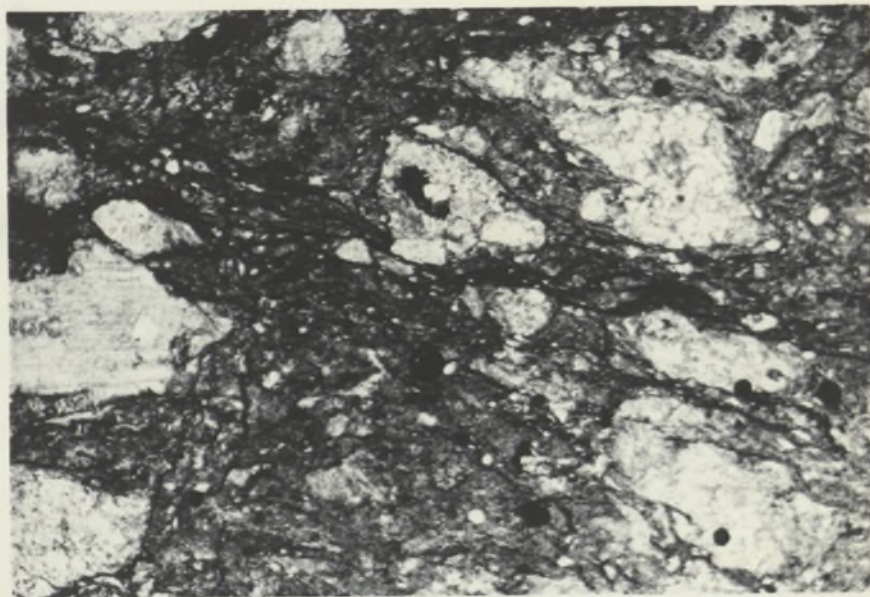
(a)



OCT 73

Jasper, showing some radiolaria.
(S.N. L2), P.P., 100 x.

(b)



OCT 73

Tuff(?), showing patches of calcite in chloritic matrix.
(S.N. 375), P.P., 100 x

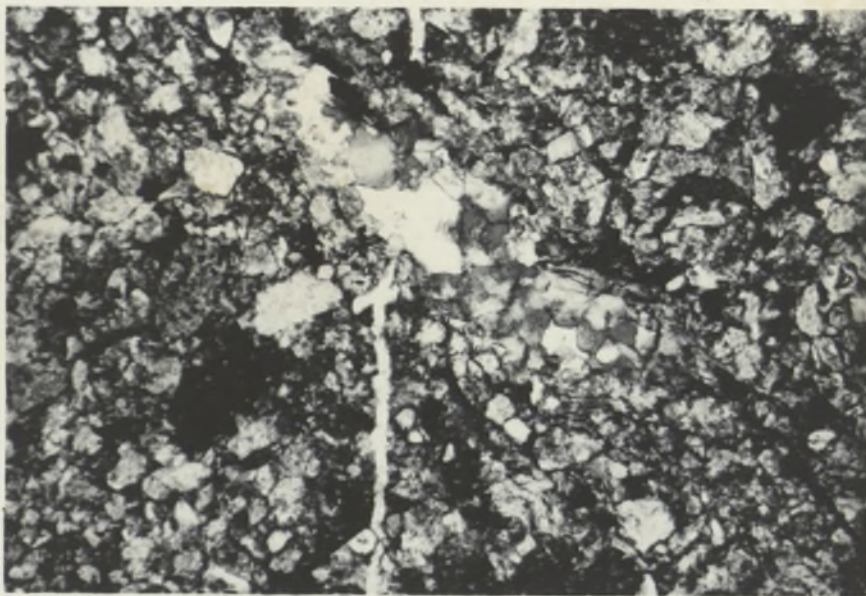
(a)



OCT 73

Unmetamorphosed lithic sandstone showing rock and mineral fragments in fine matrix. (S.N. 146), P.P., x 100

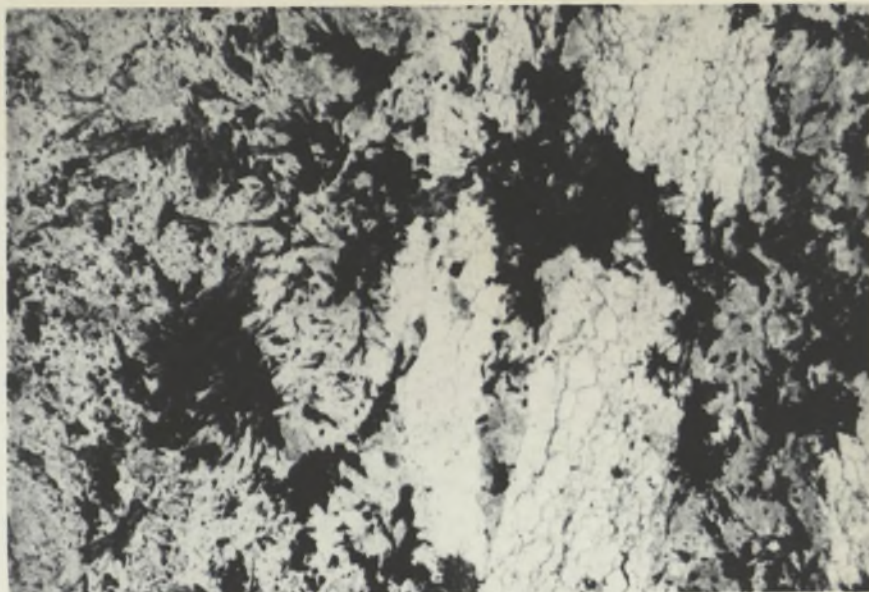
(b)



OCT 73

Metamorphosed lithic sandstone showing biotite in patch in centre. (S.N. 135), P.P., x 100.

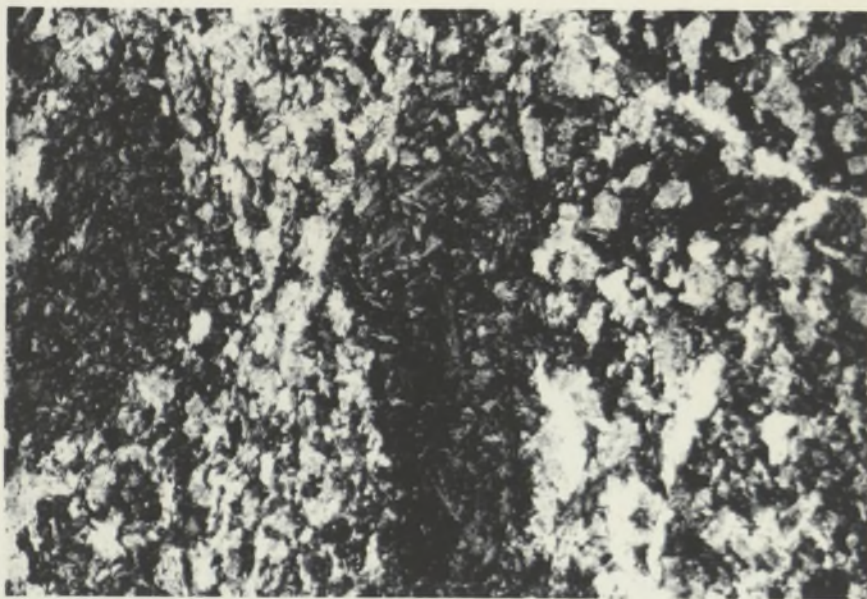
(a)



OCT 73

Ragged and radiating actinolite and biotite in veined metamorphosed lithic sandstone. (S.N.19, P.P., x 100)

(b)



OCT 73

Amphibole (Hornblende?) in volcanic fragments in metamorphosed lithic sandstones. (S.N. 72), P.P., x 100

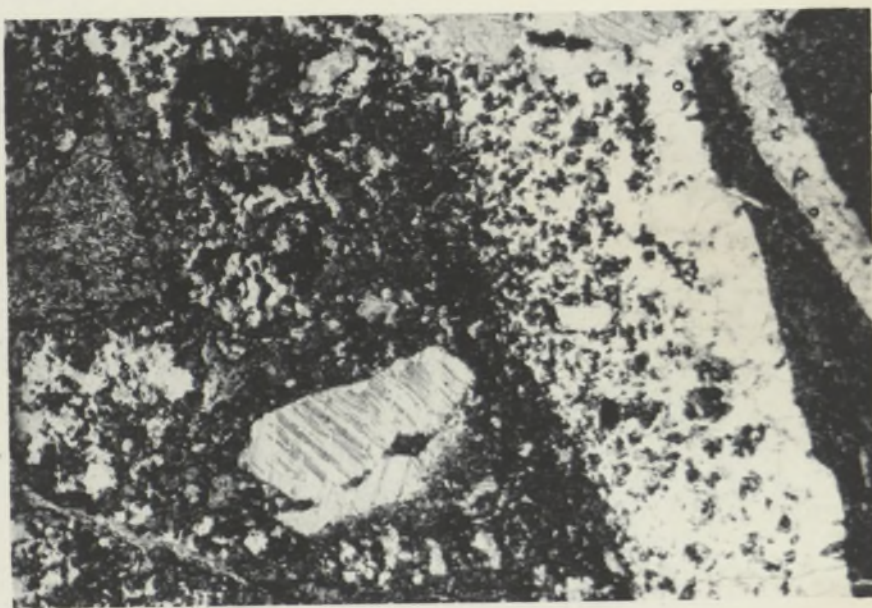
(a)



OCT 73

Albitised(?) relict plagioclase phenocrysts in metabasalt. (S.N. 287), C.N., x 100

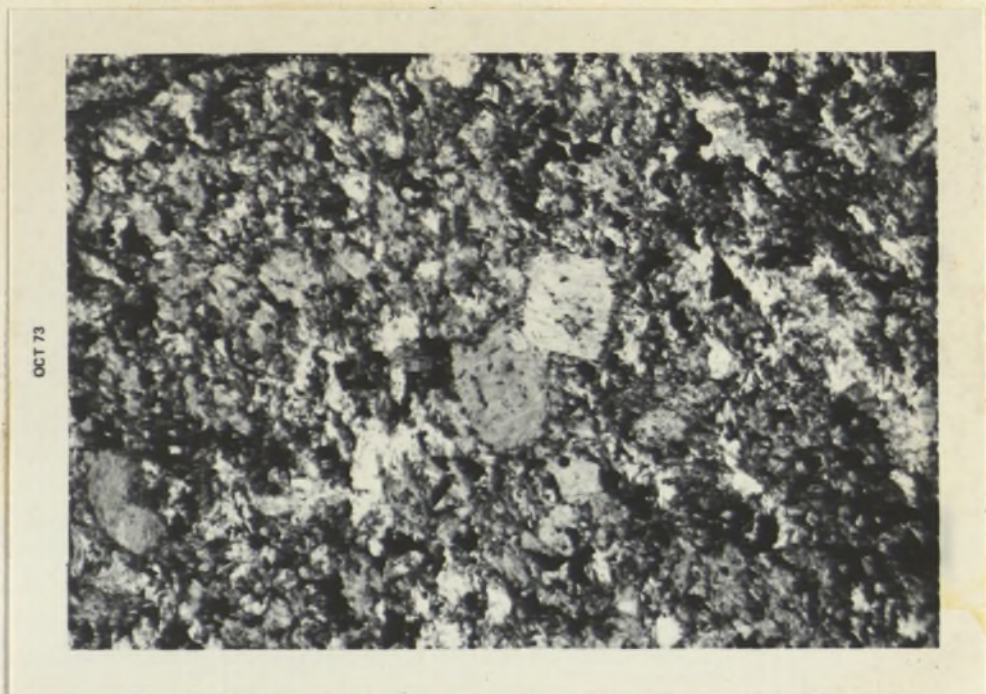
(b)



OCT 73

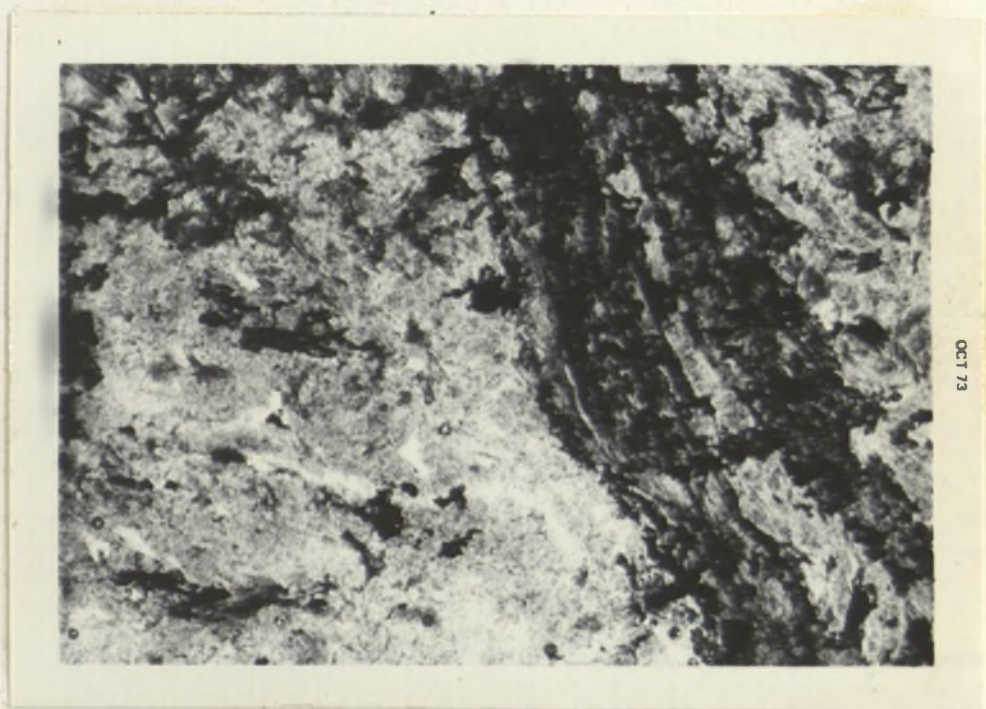
Calcite-epidote-quartz plagioclase vein in epidote-quartz-chlorite rock. (spillite?) (S.N. 160), P.P., x 100

(a)



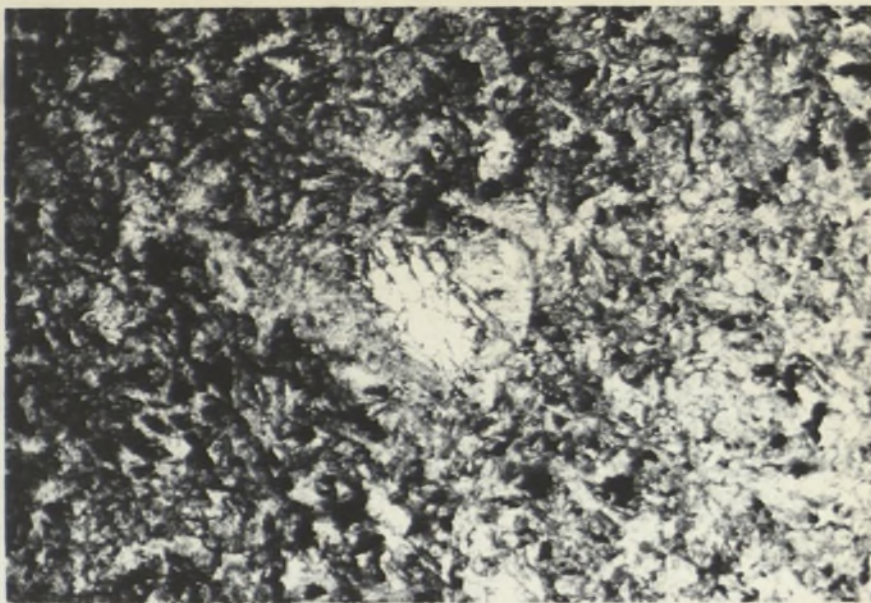
Relict clinopyroxene (centre) in mafic hornblende hornfels
(S.N. 228), P.P., x 100

(b)



Hornblende replacing plagioclase phenocryst in meta basalt.
(S.N. 80D), P.P. x 100

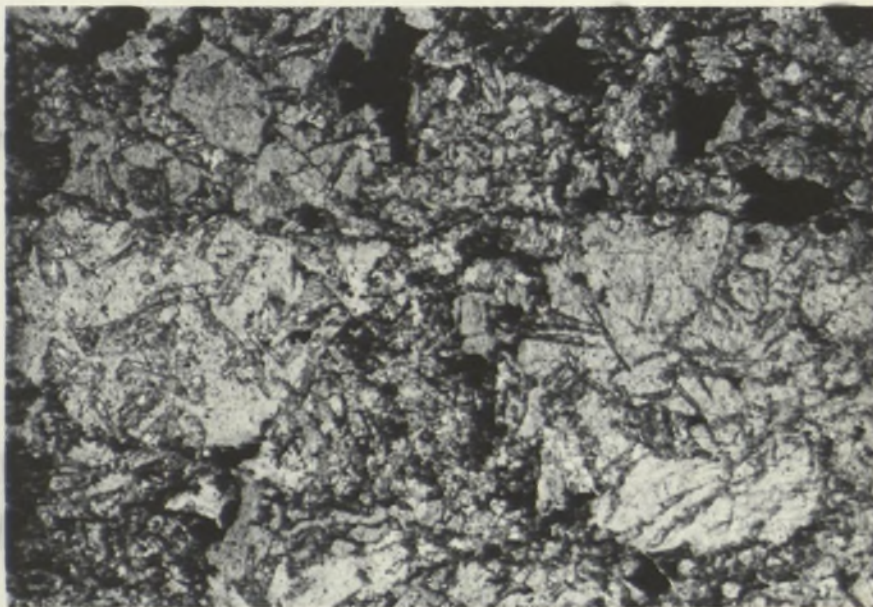
(a)



OCT 73

Relict olivine grain (centre) mantled by amphibole in meta-basalt. (S.N. 284), P.P., x 100

(b)

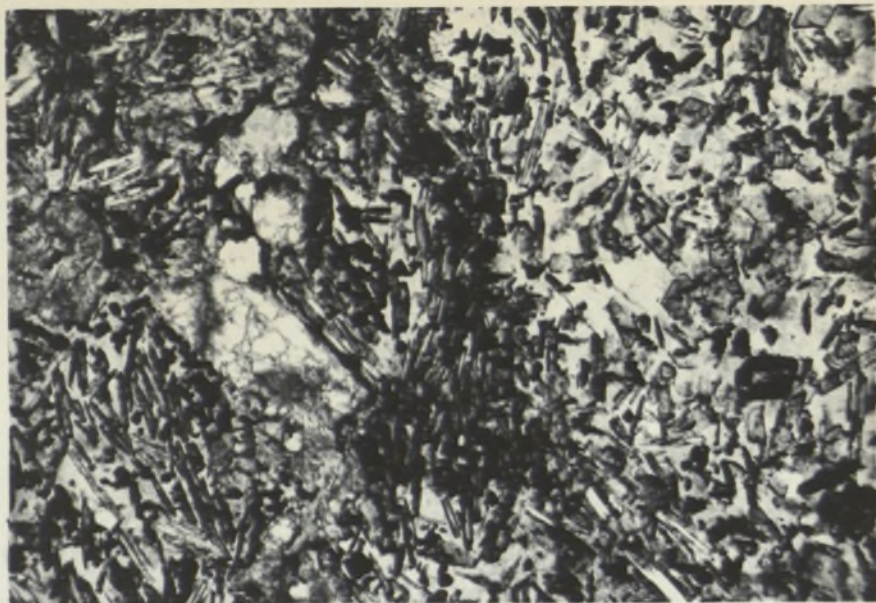


OCT 73

Large relict pyroxene grains in altered gabbro(?) (S.N. 161), P.P., x 100.

(a)

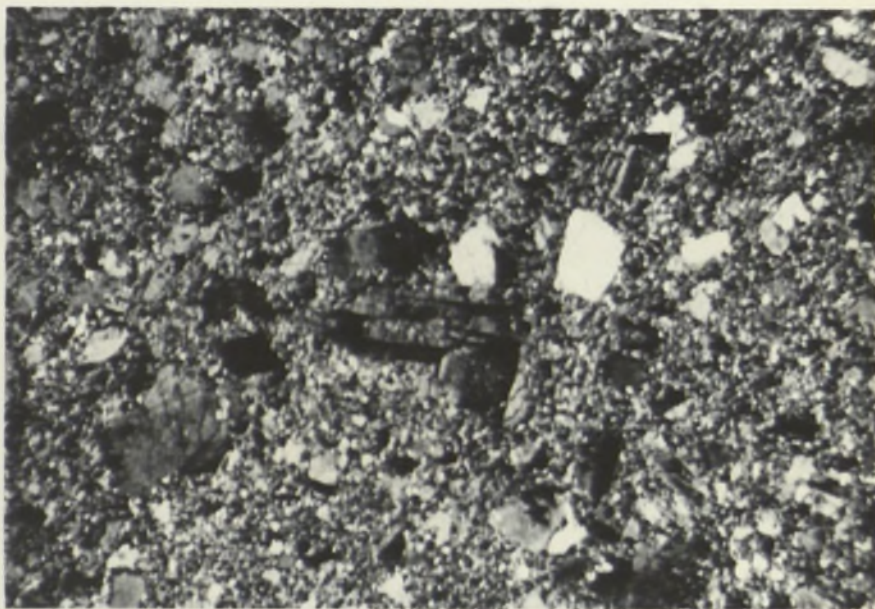
OCT 73



Lamprophyre(?) showing euhedral biotite and calcite-chlorite inclusion in alkali feldspar groundmass.
(S.N. D1), P.P., x 100.

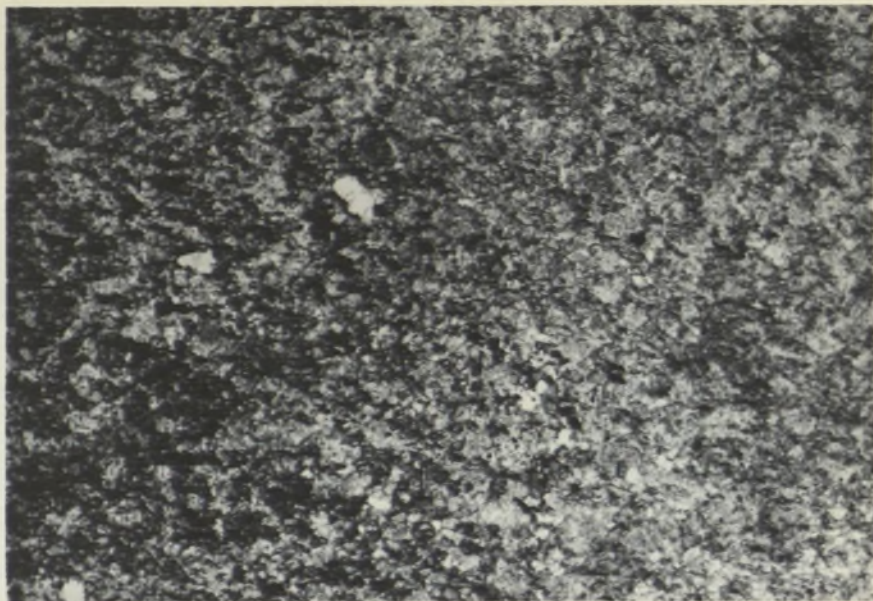
(b)

OCT 73



Relict zoned plagioclase phenocrysts in acid volcanic rock.
(S.N. 52C), C.N., x 100

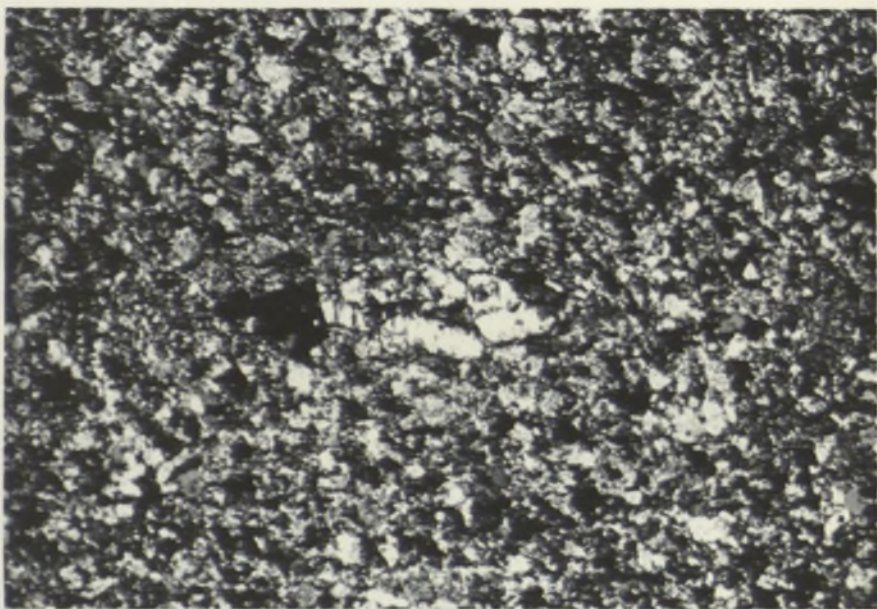
(a)



OCT 73

Boundary of pyroxene (diopside?) rich area (on right) and amphibole rich area in hornblende-hornfels rock. (S.N. 176). P.P., x 100

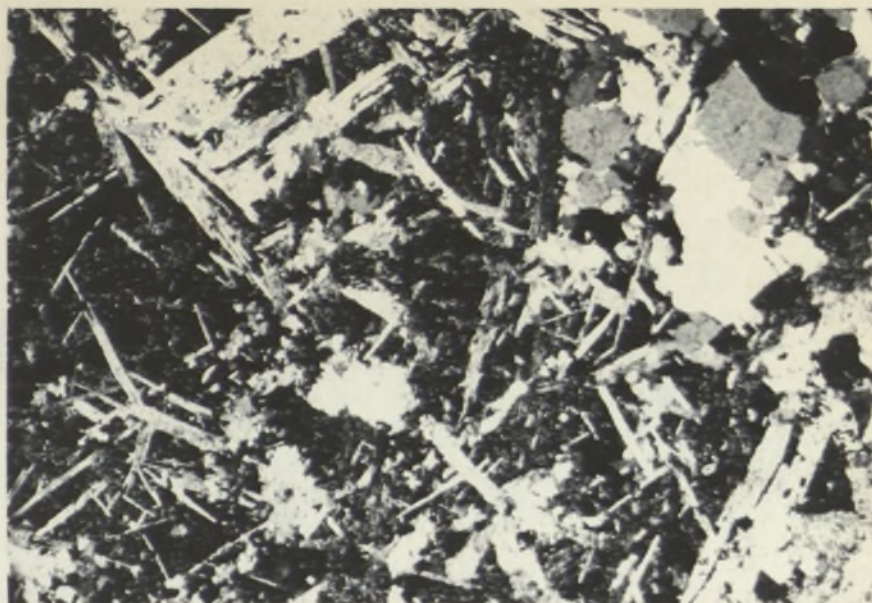
(b)



OCT 73

Clino-pyroxene grains in same rock. (S.N. 176), P.P., x 100

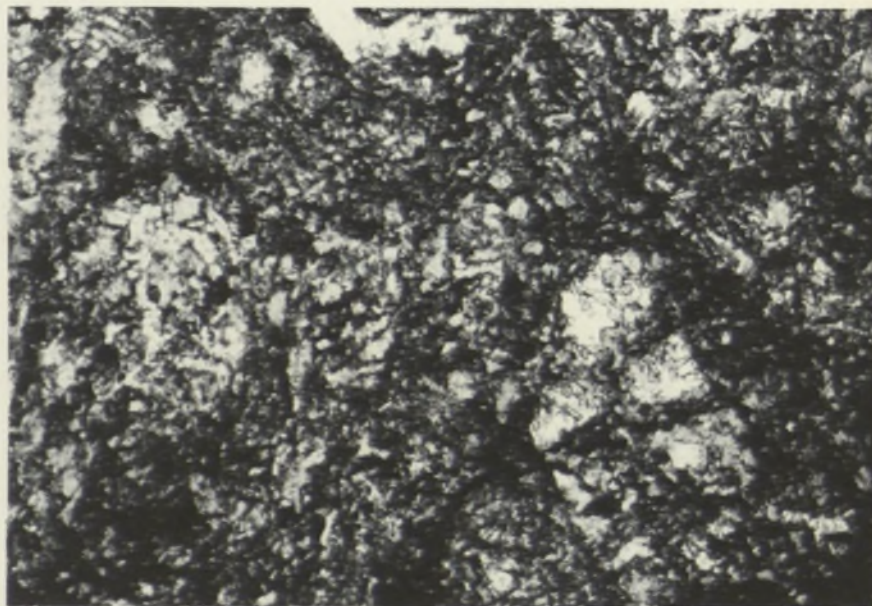
(a)



OCT 73

Calcite laths in chlorite-epidote-quartz rock (spillite?)
(S.N. 159), C.N., x 100

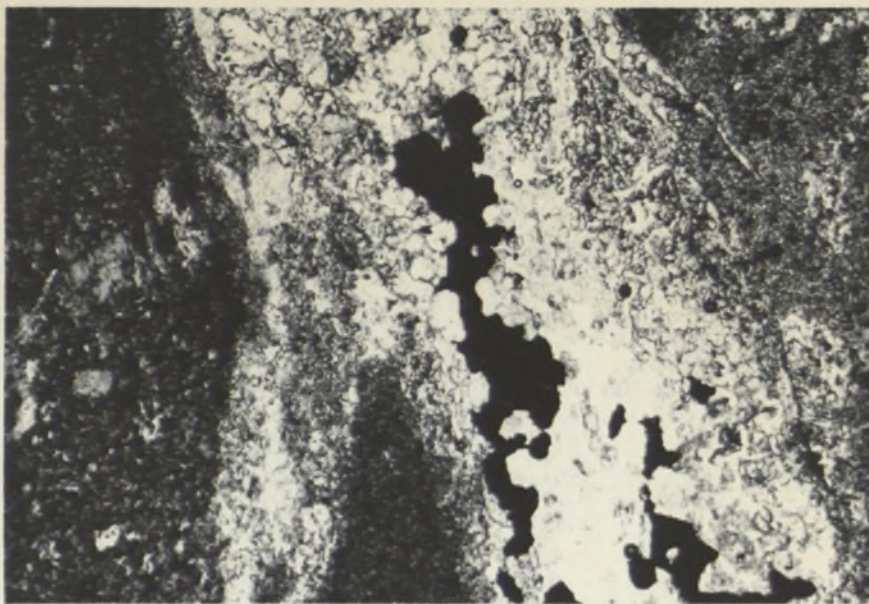
(b)



OCT 73

Hornblende hornfels texture in mafic rock near Moonbi
Adamellite. (S.N. 309), P.P., x 100

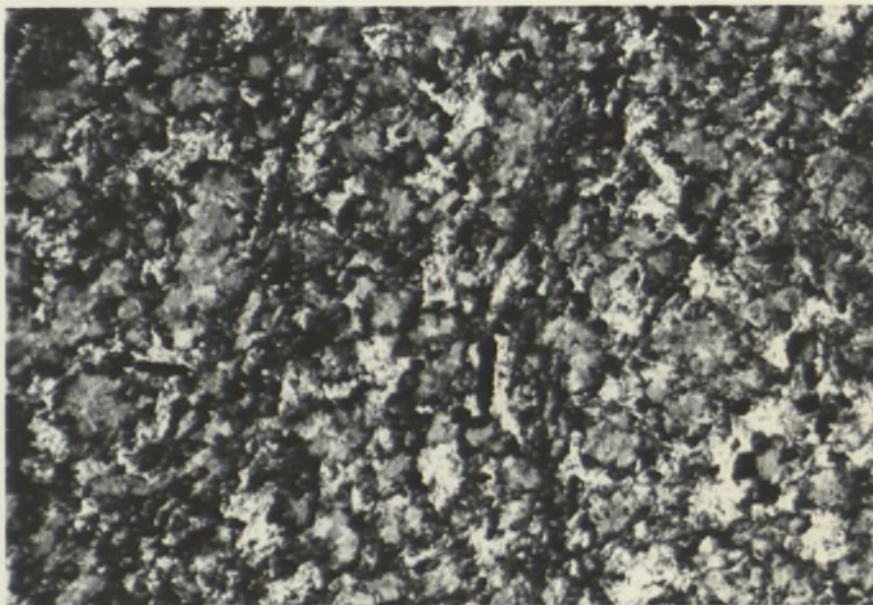
(a)



OCT 73

Diopside-quartz-calcite-chalcopyrite vein in mafic rock at Moonbi contact. (S.N. 312), P.P., x 100

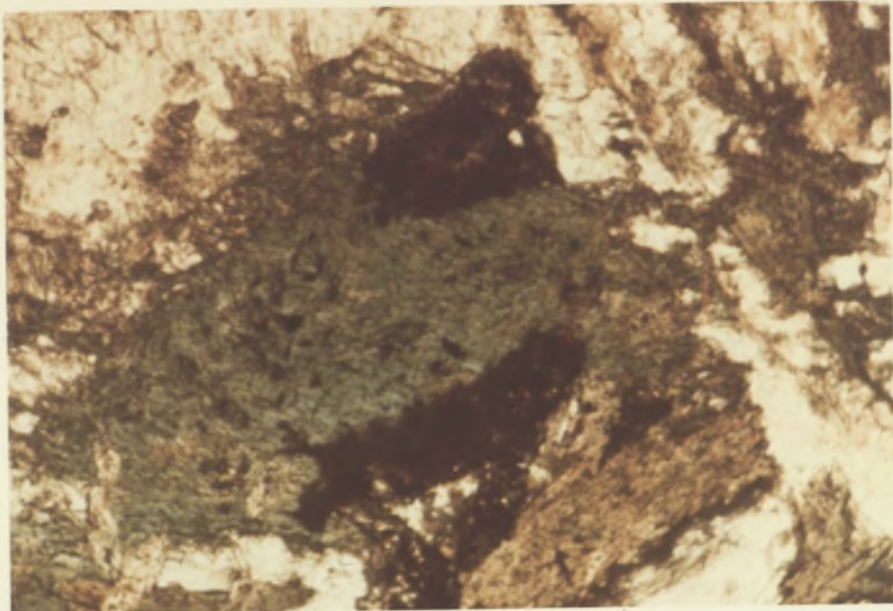
(b)



OCT 73

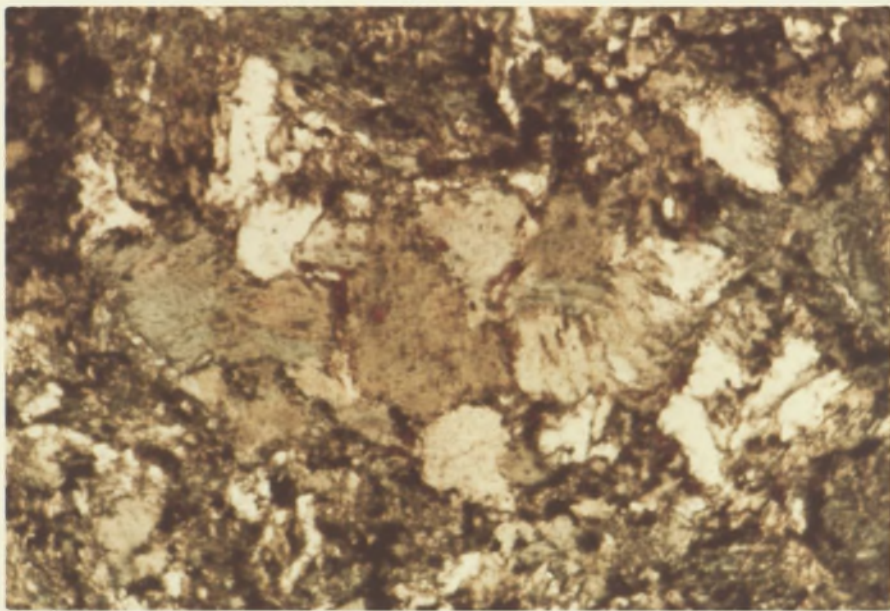
Granoblastic texture of mafic hornblende hornfels at Moonbi contact. (S.N. 313) P.P., x 100.

(a)



Amphibole in mafic hornfels approx. 2 miles from Moonbi contact. (S.N. 18(2)) P.P., x 100.

(b)



Hornblende grains in mafic hornfels close to Moonbi contact. (S.N. 310), P.P., x 100.

Pres 13
QE
341
H63
copy 1





SAMPLE LOCALITY MAP

SAMPLES COLLECTED BY F. R. HOWARTH

1973

1 31680

0 1 2 3

KILOMETRES

0 1 2

MILES

SCALE

66°00'00" N

131°00'00" E