CHAPTER 3: METAMORPHISM IN THE ROCKVALE - COFFS HARBOUR REGION

A. INTRODUCTION

The sediments of the Rockvale - Coffs Harbour region have suffered several different types of metamorphism. The types of metamorphism and their areal extent are stated briefly below, and then discussed in more detail.

1. Regional metamorphism

Regional metamorphism is of two types. The first type produced low-grade metamorphism of prehnite-pumpellyite facies to lower greenschist facies that has affected rocks of the Coffs Harbour and Dyamberin Blocks, and the Girrakool Beds of the Rockvale Block. The other type of regional metamorphism produced schists of grade up to the highest amphibolite facies and is associated with migmatites and gneissose granitic intrusions of the Wongwibinda Complex.

2. Dynamic metamorphism

Mylonites, cataclasites and phyllonites associated with the faults of the Wongwibinda Complex (Glen Bluff, Fishington and Wongwibinda faults), the Demon Fault and the Bellinger Fault are products of dynamic metamorphism.

3. Contact metamorphism

Lithologies of basic igneous composition are not present in the field area but the assemblages recorded in the pelitic and psammitic rocks indicate that hornblende hornfels facies grade of metamorphism was reached in most of the thermal aureoles around the granitic intrusions.

4. Regional-scale thermal metamorphism

The southern part of the Coffs Harbour Block, covering an area of approximately 2500 sq. km of mainly Brooklana and Moombil Beds, has suffered polymetamorphism. Low-grade regional metamorphism has been overprinted by a regional thermal event which produced randomly oriented biotite crystals similar to those observed in contact aureoles.

This chapter will concentrate on the low-grade regional metamorphism (labelled M1) and the regional scale thermal metamorphism in the sediments (labelled M2). The contact metamorphism, the regional metamorphism of the Wongwibinda Complex and some of the dislocational metamorphism have been described in detail by earlier writers and will be only briefly mentioned here.

B. COFFS HARBOUR BLOCK

The extent of metamorphic recrystallisation in the Coffs Harbour Block increases towards the Bellinger Fault, as does the structural deformation of the rocks. This metamorphism is a part of a low-grade regional metamorphism principally associated with a Permian orogeny occurring over a wide area of New England and reported by McKee and Leitch (1971). The details of the metamorphic history of these rocks will be deduced from the relationships of metamorphic minerals to structural elements, and textural relationships of coexisting minerals. Using these features two major metamorphic episodes have been recognised in the Coffs Harbour Block.

M1 Metamorphism

The first major episode produced low-grade regional effects throughout the whole of the Coffs Harbour Block and coincided with a regional deformation which mainly produced mesoscopic folding and an axial plane cleavage. At higher grades, the alignment of white mica and a foliation formed by segregated bands of minerals is parallel to the axial plane cleavage of the first period of folding.

The lower-grade parts of the block have been affected only by M1. It seems to have been caused by rapid accumulation of the geosynclinal pile, which caused a rapid increase in H_2O pressures accompanied by a low geothermal gradient. The alignment of minerals can be attributed to a high directed pressure accompanying the deformation. The products of M1 are atypical in that two diagnostic minerals in rocks presumed to be of similar grade of metamorphism elsewhere have not developed in the Coffs Harbour area. They are actinolite and stipnomelane and their absence may be related to the bulk chemistry of the rocks.

M2 Metamorphism

M2 occurs in the southern part of the Coffs Harbour Block and is a product of a regional-scale static thermal event which allowed the development and growth of randomly-oriented biotite and, less commonly, white mica. These micas occur as fine cross-cutting porphyroblasts overprinting the earlier aligned white micas of M1. The growth of the micas occurred in the absence of directed pressure and M2 is therefore post-tectonic in origin. It has only affected the higher-grade rocks of the M1 episode. The new metamorphic minerals developed in an area of low confining pressure and high heat flow. It is possible that the heat flow was from a large concealed batholith but there is no real evidence for or against this suggestion. The age of the thermal event is not known but is assumed to be late Palaeozoic. It has not affected the Nambucca Slate Belt and other regions to the south, which suggests that it was not an accompaniment of the Mesozoic rifting of the Lord Howe Rise from Australia and the formation of a spreading ridge in the Tasman Sea.

M1 and M2 Metamorphic Zones

The overprinting by thermal biotite and associated minerals has masked, to a considerable extent, the minerals and textures produced by M1. Recognition of mineral zonations in these rocks is also hampered because mineral phases with restricted stability are not developed to any great extent, probably because of the limited occurrence of suitable lithologies, such as metabasites. All volcanic rocks are very limited in areal extent and were only observed in the least metamorphosed parts of the sequence. Nevertheless, it has been possible to divide the rocks of the block into four mineralogical zones of metamorphism (Map 2). In order of increasing grade these are:

Zone I : prehnite-pumpellyite
Zone II : epidote-chlorite-white mica
Zone IIIa : biotite-epidote-chlorite-white mica
Zone IIIb : biotite-chlorite-white mica
Zone IV : biotite-white mica

Mineralogical Definition of the Zones

The metamorphic zones have been formulated using the observed ranges of significant phases in the metaclastic rocks (Fig. 13A).

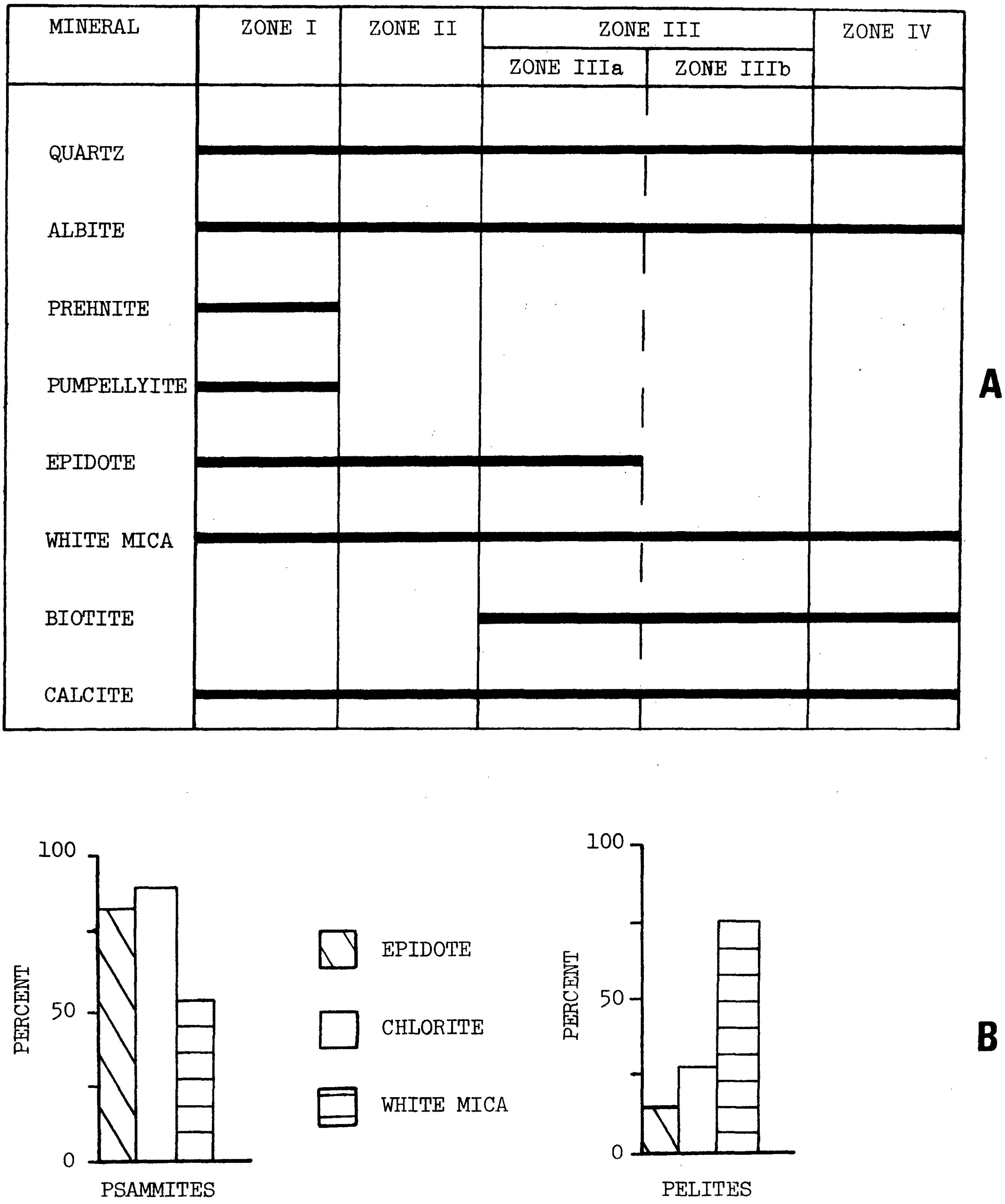


Fig. 13 A: Zonal distribution of metamorphic mineral phases in metaclastic rocks from the Coffs Harbour Block.

B: Development of metamorphic minerals in psammitic and pelitic lithologies from the Coramba Beds.

Zone I: Critical assemblages all include prehnite or pumpellyite or both. This zone is developed only in the uppermost portion of the Coramba Beds.

Zone II: This is a transitional zone lacking diagnostic minerals. Prehnite and pumpellyite have disappeared, epidote, chlorite and white mica being the main metamorphic minerals.

Zone III: The appearance of biotite heralds the start of Zone III and this isograd marks the most northerly extent of M2 overprinting M1. This zone is subdivided tentatively into two subzones:

> Subzone IIIa: Epidote, chlorite and white mica are all present along with randomly oriented biotite blebs.
>
> Subzone IIIb: This subzone differs from IIIa in that epidote has disappeared. This may be lithologically controlled, and not a grade-dependent feature.

Zone IV: Chlorite is absent and definitive assemblages contain only biotite and white mica as diagnostic minerals.

Discussion of the zonal scheme

Isograds and zone boundaries used to construct the zones are defined by:

(i) the disappearance of prehnite and pumpellyite
(ii) the appearance of biotite
(iii) the disappearance of chlorite.

Sampling problems over the Coffs Harbour Block (approximately 3750 sq. km) and the absence of critical assemblages in some rocks has made accurate mapping of zone boundaries difficult, and in some places they have been inferred.

(i) Disappearance of prehnite and pumpellyite

The disappearance of prehnite at a stage within the prehnite-pumpellyite facies, and the disappearance of pumpellyite at the upper boundary of the facies, have been recorded in the Sanbagawa Schists (Seki *et al.* 1971), in the Otago Schists (Bishop 1972) and in the Nambucca Slate Belt (Leitch 1972). Other workers (Seki *et al.* 1969, in the Tanzawa Mountains; Smith 1969, in Lachlan Geosyncline) have recorded the disappearance of prehnite and pumpellyite simultaneously, which possibly indicates that reactions involving the two phases are coupled.

With more detailed sampling in parts of the block it may be possible to recognise two isograds, one marking the disappearance of prehnite, and the other marking the disappearance of pumpellyite. Because the psammitic rocks on either side of the prehnite-pumpellyite isograd appear to be of similar composition, it is inferred that the disappearance of the two minerals is controlled by metamorphic grade.

(ii) Appearance of biotite

The biotite isograd is known to occur in both regional rocks of greenschist facies and in contact rocks of albite-epidote hornfels facies. Barrow (1912) first described a biotite zone in the Dalradian Schists of Scotland and this has been confirmed by many workers including Tilley (1925), Kennedy (1949) and Mather (1970). Biotite produced by M2 appears to develop in both the matrix of psammitic rocks and in pelitic rocks at approximately the same place and hence appears to be grade dependent with little lithological influence. By contrast, Mather (1970) found that in greywackes in Dalradian rocks biotite developed at a lower grade than in pelites, and therefore in that region it is chemically controlled as well as grade dependent.

(iii) Disappearance of chlorite

The disappearance of chlorite after the disappearance of epidote in the Coffs Harbour Block contradicts most recorded assemblages. James (1955) inferred, however, that biotite probably forms from chlorite because the amount of biotite increases as the amount of chlorite decreases. Hence the disappearance of chlorite at the Zone IV boundary is thought to be a grade-dependent event but the possibility of lithological dependence cannot be overlooked.

(iv) The disappearance of epidote within Zone III

The subdivision of Zone III into two subzones is based on the absence of epidote in the higher-grade subzone. This situation is unusual because in many metamorphic terrains epidote exists at higher grades after the disappearance of chlorite (Miyashiro 1958, in Abukuma Plateau; Banno 1964, in Sanbagawa Schists; James 1955, in Michigan; Seki 1957, in Arisu contact aureole, Kitakami Mountains). However the converse, described here, has also been noted by Seki *et al.* (1969) in the Tanzawa Mountains, where regionally metamorphosed rocks have also, in part, been thermally influenced by the intrusion of a quartz diorite.

The more frequent development of epidote and chlorite in psammitic rocks than in pelitic rocks of the Coramba Beds shows that the development of those minerals is controlled by lithology to a certain extent (Fig. 13B). The Coramba Beds were examined closely in a large number of samples (432) partly because of the virtual absence of the effects of M2 overprinting M1. The development of epidote is concentrated in psammitic rocks, occurring in 82% of them but in only 15% of the pelites. The disappearance of epidote at the subzone boundary can be related to the parent lithologies in the three formations. It has been already shown (Fig. 4) that a progressive coarsening of the grainsize of the sediments occurs from south to north. Psammitic lithologies dominate in the Coramba Beds but constitute only 12% of the Moombil Beds.

Hence epidote, which is lithologically confined to mainly psammitic rocks, would be rare in the pelite-dominated Brooklana Beds and Moombil Beds. Therefore the subzone boundary is probably controlled by lithology with little or no dependence on metamorphic grade.

Metamorphic Petrography of the Coffs Harbour Block

Effects of the low-grade regional metamorphism (M1)

The rocks of the Coffs Harbour Block will be discussed in terms of their parent lithology. The metaclastic lithologies show a wide variation in the degree of textural reconstitution but, because of the lack of metabasic rocks in the sequence no really diagnostic minerals are developed. Hence it is difficult to demonstrate that an increase in metamorphic grade closely accompanies textural reconstitution. The modification of the detrital fabric, and textural reconstitution is related to the parent rock type.

Greywacke-type Psammites

There is no distinct boundary between subgreywackes with 2% matrix and greywackes with 40% matrix. Hence all psammitic rocks of this type are grouped under the term "greywacke-type". The coarser fraction of the psammites originally shows little microscopic evidence of deformation or recrystallisation, the onset of metamorphism being indicated mainly by the progressive recrystallisation of the silty and clay-rich matrix to a finely divided mixture of granoblastic quartz and albite, small granules of epidote, thin threads of chlorite and white mica and rare opaque minerals. At the lowest grade (Zone I) there is no preferred orientation of the new mineral phases. The paucity of muddy matrix in some parent lithologies is reflected in the relative impoverishment of white mica and other secondary minerals. Veins consisting of variable combinations of quartz, albite, chlorite, prehnite, calcite and epidote have formed in many rocks.

Initially, detrital grains are little modified with restricted albitisation of plagioclase, development of minor epidote, hornblende partly altered to chlorite and some volcanic lithic fragments devitrified to fine-grained granoblastic quartz and feldspar. The detrital grain boundaries still retain their original sharp outlines (Plate 8A). Prehnite occurs both as a vein mineral and as a spongy aggregate in the matrix. Pumpellyite is confined to the matrix as tiny randomly-oriented flakes, usually associated with epidote which appears to have developed as a reaction corona. Chlorite is commonly associated with both prehnite and pumpellyite.

With increasing grade and the disappearance of prehnite and pumpellyite, the micaceous groundmass becomes coarser and starts to develop a preferred orientation. Granular phases such as epidote tend to coarsen in the matrix. The coarse detrital fragments start to develop a dimensional orientation parallel to the preferred orientation of white mica (Plate 8B) but still show little marginal alteration. Much of the lithic debris has broken down to a granoblastic quartz-feldspar aggregate obliterating much of the relict volcanic textures. Mudstone fragments tend to develop minute flakes of white mica.

With further increase in grade some of the textures are masked by the overprinting of thermal biotite and associated minerals. The low-grade regional features appear to be marked by the modification of detrital grain boundaries with both albitised plagioclase and quartz being embayed and fine-grained aggregates developing at the margins. The rocks consist of relict detrital fragments "floating" in a quartz-albite rich metamorphic groundmass with an average grain size of about 0.03 mm (Plate 8C). White mica flakes have become much coarser and show a strong preferred orientation which occasionally swirls around the relict grains. Detrital quartz grains are breaking down to granoblastic aggregates of finer crystals.

The coarsening of groundmass phases and continued modification of detrital grains with preferential destruction of plagioclase detritus is

PLATE 8

A. Greywacke with detrital grains retaining their original sharp outline. Unit D, Coramba Beds. S32383, magnification x 20, plane polarised light.

B. Greywacke with detrital grains developing a preferred orientation parallel to the orientation of white mica. Unit D, Coramba Beds. S32413, magnification x 20, plane polarised light.

C. Lithic greywacke showing breakdown and reconstitution of lithic fragments to quartz-albite-white mica with a preferred orientation. Unit D, Coramba Beds. S32373, magnification x 50, crossed nicols.

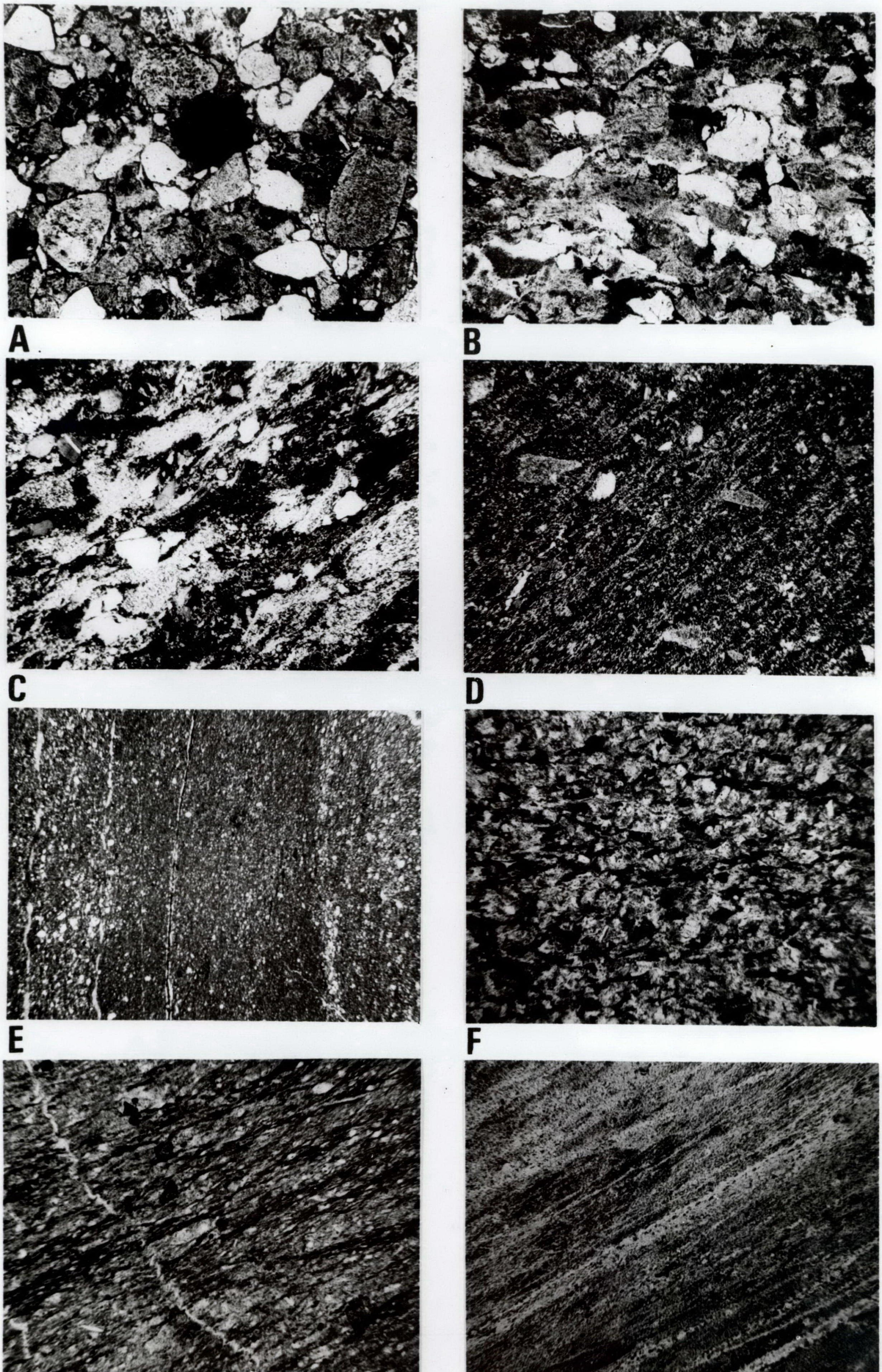
D. Almost complete reconstitution of a fine-grained greywacke with rare relict detrital grains of quartz and feldspar still present. Brooklana Beds. S32594, magnification x 50, crossed nicols.

E. Pelitic rock showing original sedimentary laminations and consisting of fine grains of quartz and feldspar set in a dark-brown matrix. Unit D, Coramba Beds. S32410, magnification x 20, plane polarised light.

F. Development of a weak foliation due to flattening of some grains in a pelite. Unit A, Coramba Beds. S32555, magnification x 50, plane polarised light.

G. Pelite with very fine-grained white mica enclosing small relict detrital quartz which show a marked orientation parallel to the white mica. Brooklana Beds. S32628, magnification x 50, plane polarised light.

H. Pelite suffering highest metamorphic grade and exhibiting development of a foliation due to segregation of minerals, and almost complete reconstitution of detrital material. Moombil Beds. S32664, magnification x 50, plane polarised light.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indicative of the highest grades observed in the psammitic rocks. Complete reconstitution of the rock does not occur and it is still possible to observe some remains of a relict sandstone texture (Plate 8D).

Pelitic Rocks

Non-siliceous pelitic rocks have been reconstituted mainly to micaceous phases; they resemble the matrix of the psammites and differ from the siliceous pelites which verge on cherts. With increasing metamorphic grade these rocks show a textural progression from laminated mudstones to cleaved slates.

The least affected pelites consist of very fine detrital grains of quartz and feldspar set in a dark-brown matrix (Plate 8E) consisting of tiny shreds of white mica, which appears to be dominant over chlorite. Epidote and prehnite are rare but calcite may occur as spongy xenoblastic masses within the rock. Veins of quartz with or without other minerals such as chlorite, calcite, epidote and prehnite frequently occur. Some flattening may have occurred and a weak foliation may be present (Plate 8F).

With increase in grade a network of lepidoblastic white mica crystals shows a well developed foliation occasionally overprinted by spongy aggregates of xenoblastic calcite. The white micas completely enclose the small relict detrital quartz fragments which now show a marked dimensional orientation (Plate 8G).

The highest grade observed in the pelites is the development of a foliation due to the segregation of minerals into bands of quartzo-feldspathic layers (Plate 8H) and this has involved the almost complete reconstitution of detrital material.

Siliceous Pelites

These rocks were originally siliceous mudstones to cherts and consisted dominantly of cryptocrystalline silica with minor detrital quartz and mud (Plate 9A). The mud has been converted to white mica, and if sufficiently abundant produces an incipient foliation. Veining is common, particularly quartz, with or without calcite (Plate 9B).

Increased metamorphism is marked by the coarsening of the mineral phases. Detrital grain boundaries of quartz become blurred and the cryptocrystalline groundmass coarsens so that individual grains can be discerned

PLATE 9

A. Relatively unmetamorphosed chert consisting dominantly of cryptocrystalline silica. Unit D, Coramba Beds. S32816, magnification x 50, plane polarised light.

B. Slightly recrystallised chert containing coarser-grained quartz veins. Unit D, Coramba Beds. S32401, magnification x 50, crossed nicols.

C. Coarsening of quartz grains due to increasing grade of metamorphism. Individual grains can now be distinguished microscopically. Brooklana Beds. S32604, magnification x 50, crossed nicols.

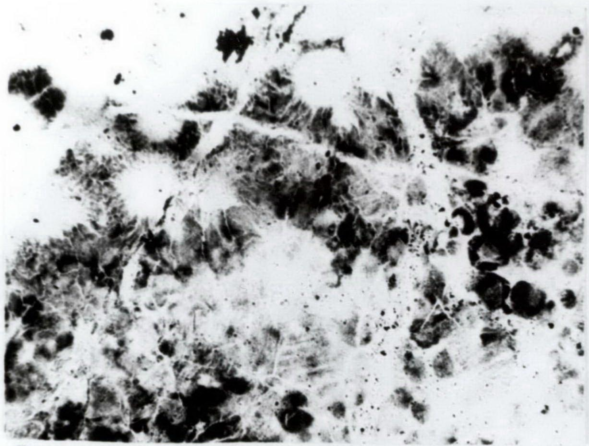
D. Chert showing coarse granoblastic aggregates of quartz due to recrystallisation at the highest grade of metamorphism. Moombil Beds. Magnification x 50, crossed nicols.

E. Silt-sized fraction in a quartz-rich psammite exhibiting a strong foliation due to the alignment of white mica. Unit D, Coramba Beds. S32819, magnification x 50, crossed nicols.

F. Enlargement of portion of Plate 9E. Magnification x 200, crossed nicols.

G. Randomly-oriented flakes of metamorphic biotite due to alteration of the matrix of a lithic greywacke. Unit D, Coramba Beds. S32404, magnification x 200, plane polarised light.

H. Metamorphic biotite growing preferentially at boundaries between detrital grains. Unit B, Coramba Beds. S32534, magnification x 50, plane polarised light.



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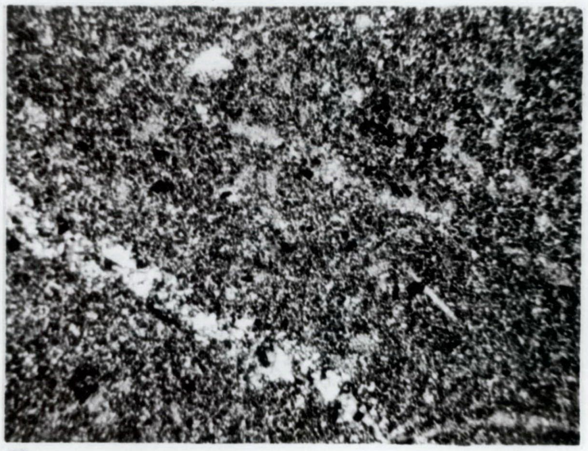

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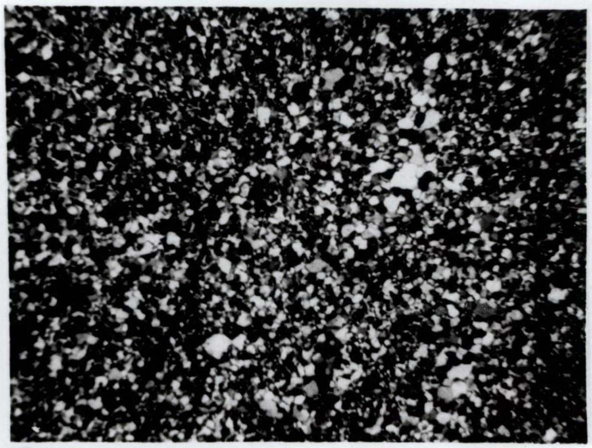

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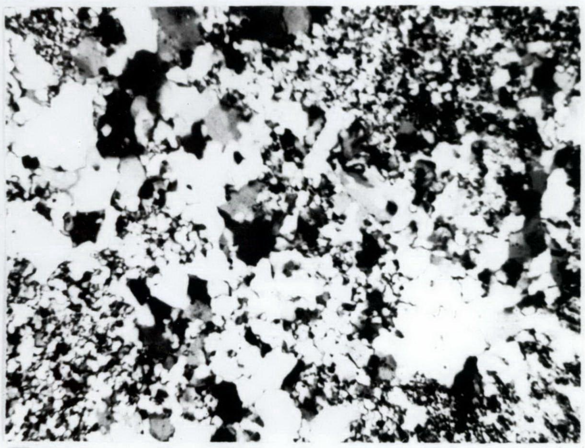

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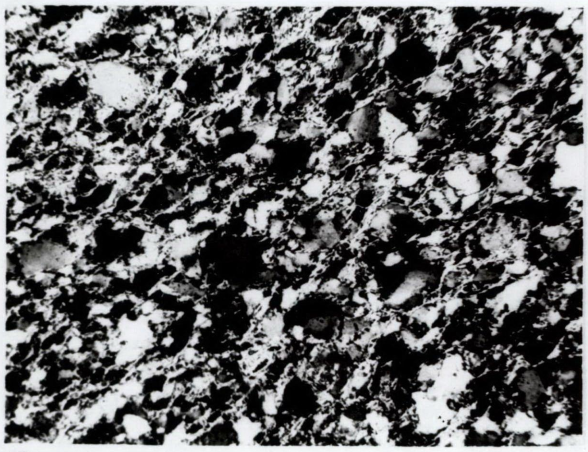

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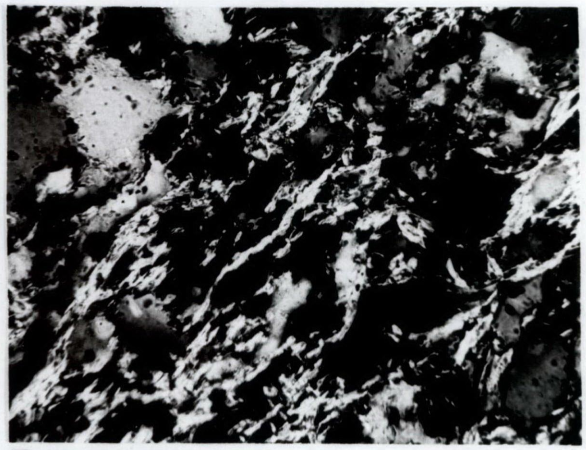

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microscopically (Plate 9C). The results of the highest-grade are granoblastic quartz-rich aggregates containing rare albite and white mica (Plate 9D). A foliation is either absent or only very poorly developed.

Quartz-rich Psammite

Only one specimen of this rock type (S32819) was found and it occurs in the lower part of Zone II. It has a psammitic and a silt-grade fraction. The coarse fraction consists of angular to rounded detrital quartz (0.3 - 0.7 mm) with rare feldspar, zircon and sphene grains. The quartz grain-boundaries are often irregular and are being broken down and incorporated in the matrix. The clay portion has been converted to white mica and exhibits a weak foliation.

The silt-sized fraction (0.05 - 0.1 mm) has a higher proportion of matrix which exhibits a strong foliation (Plates 9E and 9F) due to the alignment of white mica. The bedding contact between the silt and psammite fractions is sharp and is cut at an angle of 20° by the mica foliation.

Acid Volcanic Rock

An acid lava with relict flow banding occurs in Zone II. In thin section (S32810, S32811) a very irregular devitrification texture (Plate 6A) of quartz and feldspar has developed. Circular pods of calcite occur abundantly in S32811 but are rare in S32810. The only other mineral developed is a faint brownish-green chlorite.

Metabasic Volcanic Rocks

Metabasic rocks have been observed at two localities in Zone II, and have suffered extensive alteration with the development of chlorite and calcite. In S32812 a relict intergranular texture is still obvious and the assemblage is plagioclase-calcite-chlorite-opaques. S32818 is a highly sheared propylitised intermediate to mafic volcanic with an assemblage of calcite-chlorite-talc-quartz. The talc is possibly replacing orthopyroxene.

Textural Modification of the Metaclastic Rocks

The progression from slightly altered to almost completely reconstituted rocks has been described above. The relationships of the important textural changes to metamorphic zones are set out below. The zones listed record the first occurrence of the features which do not necessarily exist in a similar rock at a higher grade:

	Textural Change	Psammite	Pelite
1.	Alignment of sheet silicate	II	I
2.	Development of foliation	IV	IIIb
3.	Alignment of detrital grains	II	II
4.	Modifications of margins of quartz grains	II	II
5.	Reconstitution of lithic grains	IIIb	-

Turner (1935) and Hutton and Turner (1936) recognised four units of progressive textural change in greywackes from the Alpine and Otago Schists of New Zealand. Turner (1938) subdivided the chlorite zone into four textural subzones (Chl 1 - Chl 4). The scheme has been used extensively in New Zealand (e.g. Reed 1958; Mason 1962; Grindley 1963). Leitch (1972) equated those subzones with the mineral zones he identified in the Nambucca Fold Belt. Similarly in the Coffs Harbour Block, Zone I is equated with Chl 1 subzone, Zone II with Chl 1-2, Zone III with Chl 2 and Zone IV with Chl 2-3. The positions of the textural subzone boundaries and mineral zone boundaries do not correspond. However it is clear that an increase in textural evolution is accompanied by an increase in the metamorphic grade. Because it has been possible to relate the rocks of the Coffs Harbour Block and the Nambucca Fold Belt with the subdivisions of Turner, it is inferred that the conditions during metamorphism in New England must have been similar to those that produced the Otago and Alpine Schists of New Zealand. This is confirmed by similar mineral assemblages in rocks from New England and New Zealand.

The Development of the Zone of Thermal Biotite (M2)

Biotite appears in both pelitic rocks and in the matrix of psammites at approximately the same grade. In pelites it consists of randomly oriented single grains and small clusters of lepidoblastic crystals. In psammites biotite occurs as large clusters (up to 2 mm) of numerous small randomly oriented flakes in the matrix (Plate 9G) and is located preferentially at boundaries between detrital grains and the matrix (Plate 9H). At this early stage there is some breakdown of the margins of detrital grains, particularly of quartz.

The biotite rapidly becomes a major component of the matrix of the psammites and the remainder consists of granoblastic quartz and feldspar with some chlorite, epidote and white mica probably remaining from the earlier M1.

In the highest grades biotite is accompanied by large granoblasts of randomly-oriented muscovite which differs markedly, in its form, size and orientation, from the white mica produced by the low-grade regional event.

In many higher grade rocks a foliation produced by the alignment of white mica is still present but has been overprinted by randomly-oriented biotite grains which tend to grow preferentially around boundaries of any grains that have not been reconstituted into the granoblastic groundmass (Plate 10A). In some thin sections spongy blebs of calcite developing simultaneously with the biotite, grew across the white mica foliation (Plate 10B). Only in rocks of fine-grained parentage has the reconstitution been almost complete and here the biotite often occurs as stringers, presumably being controlled by the bands of parent mineral(s) from which the biotite is growing.

Metamorphic Minerals

For the more common phases of metamorphic minerals produced by M1 and M2, an attempt is made to determine any grade-dependent changes, using optical and X-ray parameters. The properties of the minerals are compared with those of other low-grade terrains, particularly the adjacent Nambucca Fold Belt. Minerals from two specimens in the Dyamberin Block (R32705, R32724) and from three specimens of the Rockvale Block (R32764, R32785, R32801) have been examined also, and their properties are extremely similar to those of minerals from the Coffs Harbour Block.

Quartz

Quartz occurs in most metamorphic assemblages examined and is also the dominant vein mineral present. Dennen *et al.* (1970) showed that the Al content of quartz could be used as a geothermometer, because there is an increase of 1 ppm Al per 3.6°C rise in temperature. No chemical work for quartz was undertaken but the (100) and (101) peaks on X-ray diffractometer charts were examined to investigate possible trends. No noticeable variation exists in the position of the two quartz peaks with increasing grade (Table 11, Figs. 14A and B). This suggests that there is only a small increase in temperature of metamorphism from Zone I to Zone IV.

Because of the demonstrated stability of quartz within the four zones, the (101) peak of this mineral was used as the standard from which peaks of

PLATE 10

A. Randomly-oriented biotite grains growing preferentially around boundaries of detrital grains and overprinting an earlier foliation. Brooklana Beds. S32597, magnification x 50, plane polarised light.

B. Randomly-oriented blebs of biotite overprinting a white mica foliation. Unit A, Coramba Beds. S32578, magnification x 20, plane polarised light.

C. Well developed idioblastic crystals of epidote altering from detrital plagioclase in a lithic greywacke. Unit D, Coramba Beds. S32366, magnification x 20, plane polarised light.

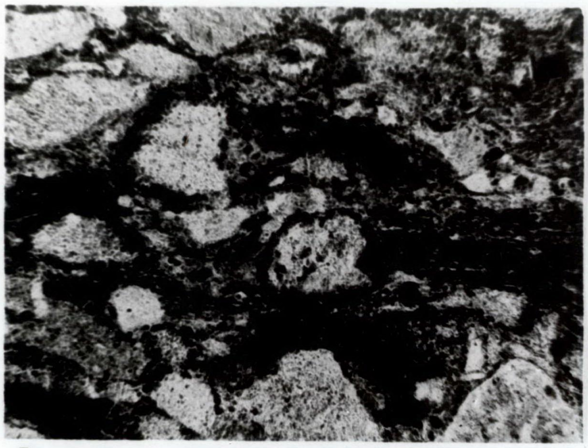
D. Veins of prehnite in a fine-grained siltstone. Unit D, Coramba Beds. S32395, magnification x 20, plane polarised light.

E. Xenoblastic crystals of metamorphic garnet in a fine-grained siltstone. Brooklana Beds. S32645, magnification x 50, plane polarised light.

F. Radiating needles of secondary tourmaline in a coarse greywacke-breccia. Girrakool Beds. S32767, magnification x 200, plane polarised light.

G. Idioblastic pyrite crystals showing distinctive cubic outlines. Dyamberin Beds. S32695, magnification x 50, plane polarised light.

H. Typical slate from Dyamberin Beds showing well developed foliation. S32705, magnification x 50, plane polarised light.



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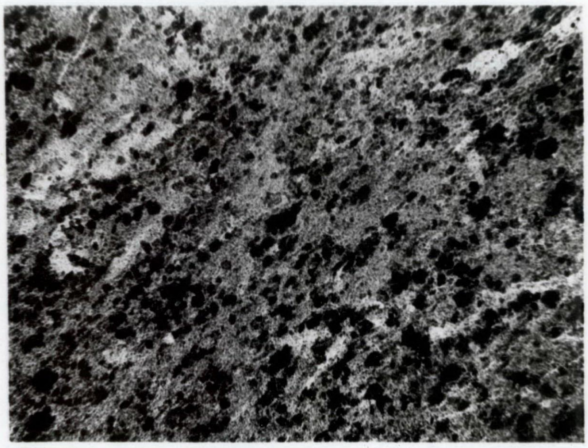

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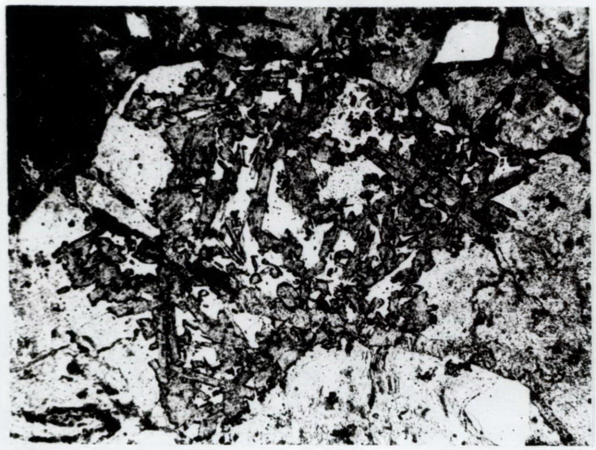

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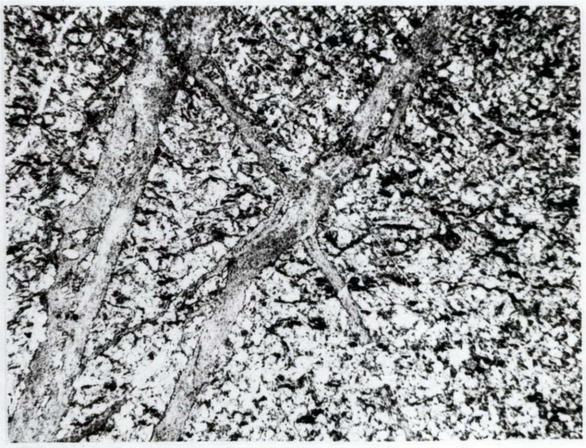

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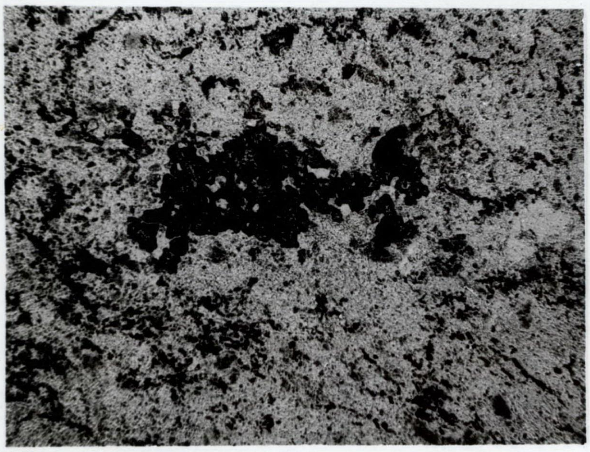

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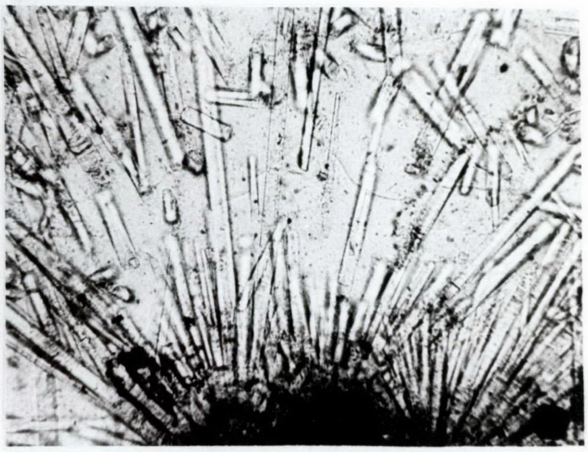

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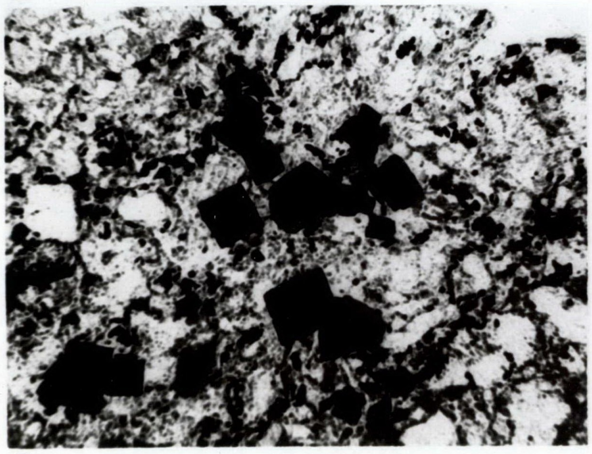

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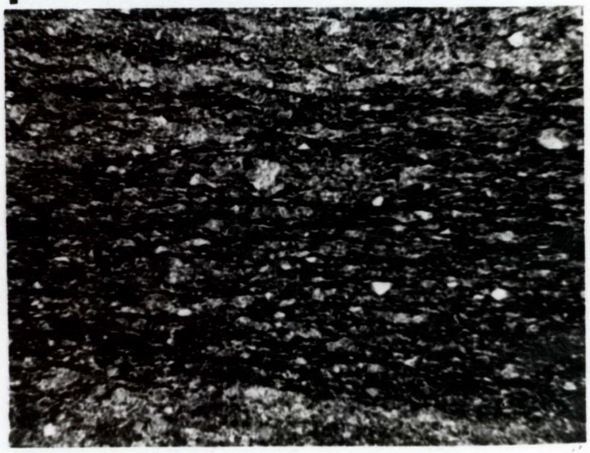


H

Table 11: X-ray diffraction data for Quartz

Specimen	Parent Lithology	Zone	d(100)A°	d(101)A°
32375	Psammite	I	4.271	3.348
32446	Psammite		4.267	3.352
32439	Psammite		4.271	3.353
32395	Pelite		4.279	3.356
32361	Psammite		4.287	3.359
32388	Psammite	II	4.251	3.338
32434	Pelite		4.255	3.341
32415	Pelite		4.251	3.346
32724	Metabasite		4.263	3.346
32387	Pelite		4.259	3.351
32454	Psammite		4.271	3.353
32705	Pelite		4.271	3.353
32819	Quartz-psammite		4.261	3.356
32441	Psammite		4.275	3.356
32373	Psammite		4.287	3.358
32371	Pelite		4.283	3.358
32452	Psammite		4.281	3.361
32378	Pelite		4.283	3.361
32498	Psammite	IIIa	4.239	3.330
32419	Psammite		4.259	3.347
32633	Pelite		4.271	3.351
32380	Psammite		4.265	3.353
32764	Psammite		4.279	3.358
32785	Psammite		4.283	3.363
32578	Pelite		4.281	3.366
32534	Psammite		4.287	3.373
32614	Pelite	IIIb	4.259	3.347
32600	Psammite		4.263	3.353
32612	Pelite		4.271	3.353
32609	Pelite		4.271	3.353
32615	Pelite		4.271	3.353
32583	Psammite		4.275	3.356
32672	Pelite		4.279	3.356
32616	Pelite		4.279	3.358
32594	Psammite		4.275	3.361
32588	Psammite	IV	4.271	3.351
32694	Pelite		4.263	3.351
32686	Pelite		4.275	3.356
32673	Pelite		4.283	3.358
32679	Pelite		4.283	3.361
32682	Pelite		4.267	-
32681	Pelite		4.283	-
32801	Psammite		4.291	-
32587	Pelite		4.293	-

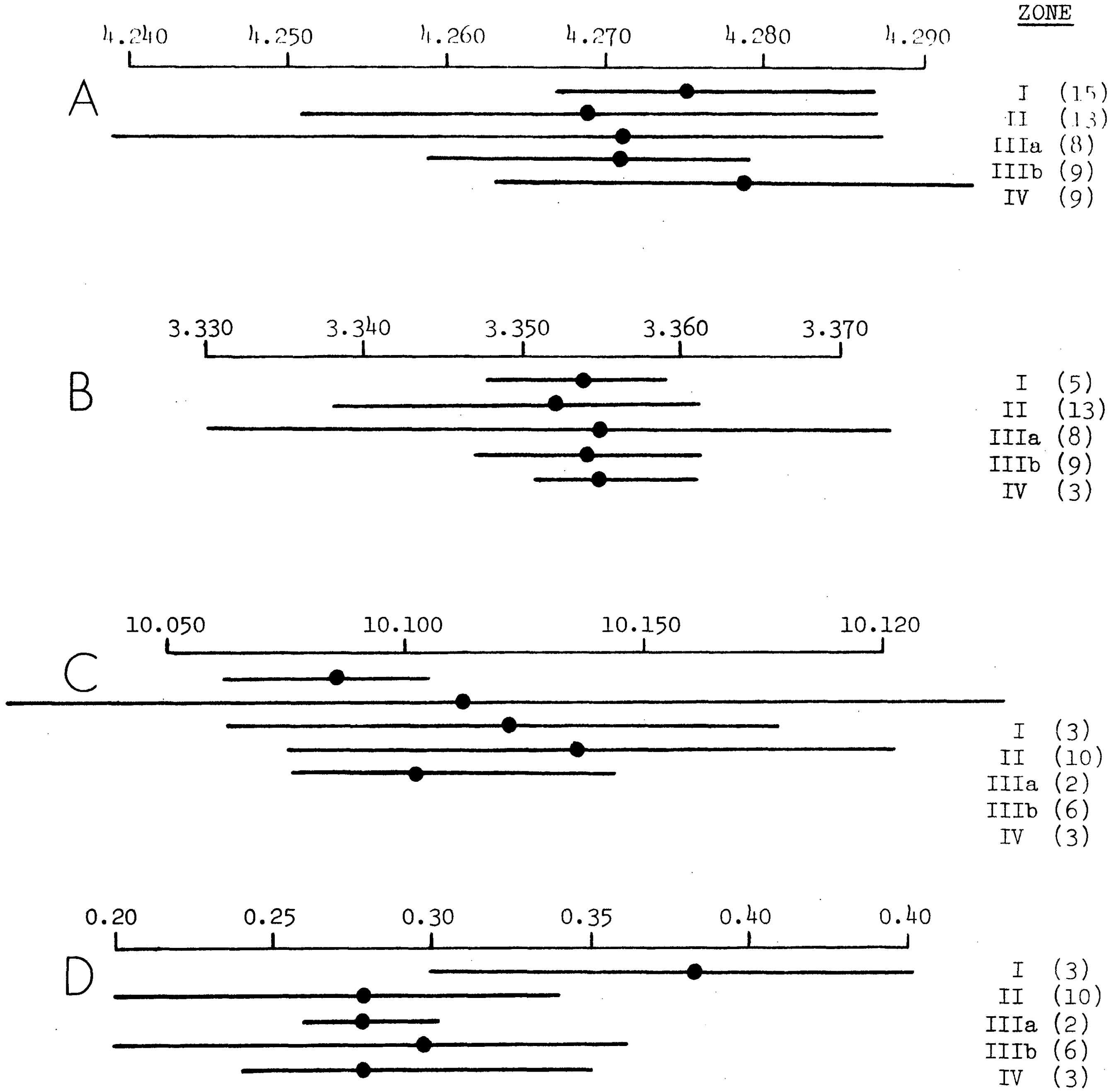


Fig. 14 A: Range of d (100) A° quartz peaks within the metamorphic zones.

B: Range of d (101) A° quartz peaks within the metamorphic zones.

C: Range of d (002) A° basal spacings for white micas within the metamorphic zones.

D: Range of white mica crystallinity within the metamorphic zones.

For each feature the bar indicates the range, and the filled circle is the mean. The number in brackets after each zone number refers to the number of samples examined.

all other minerals were measured.

Albite

The extremely fine grain-size of plagioclase of definite metamorphic affinity has prevented the determination of a reliable refractive index. However the mineral is assumed to be albite, in accordance with that found in other metamorphic terrains of similar grade. Measurement of the factor $2\theta\,(131) - 2\theta\,(1\bar{3}1)$ for several specimens gave values of 1.05 - 1.57° 2θ, but because of the presence of detrital feldspars in most specimens, the validity of these values is suspect.

White Mica

The colourless to extremely pale green micaceous mineral in the rocks from the Coffs Harbour Block is referred to here as "white mica", because of the confused state of the nomenclature of mica phases in very low-grade metamorphic rocks, and particularly of the term illite (see Dunoyer de Segonzac 1970, p.312). Workers in other field areas of comparable grade have shown that the white mica has phengitic affinities (Brown 1967; Leitch 1972).

White mica occurs in all lithological types but is more common in pelitic rocks (Fig. 13B) and is the main mineral which defines the foliation developed in some of the rocks from higher-grade zones. The fine grain-size and frequent occurrence of minute inclusions has rendered refractive index determinations difficult and unreliable.

The means for basal spacings d(002) for white micas from each zone (Table 12, Fig. 14C) show a distinct increase from Zone I to Zone IIIb, but with a decrease from a mean of 10.136°A in Zone IIIb to 10.102°A in Zone IV. Banno (1964) records a slight increase in the basal spacings of muscovite from the Sanbagawa Schists, which he relates to an increase in the paragonite component. The size and range of basal spacings measured by Banno are smaller than those from the Coffs Harbour Block (9.90 - 9.98°A cf 10.01 - 10.22°A). Zen and Albee (1964) have related chemical composition (Na-content) to the d(002) spacing using the regression equation : $d(002) = 10.034 - 0.427\ N_{pa}$, where N_{pa} is the mole fraction of paragonite. The high basal spacings recorded here indicate the white mica approaches muscovite in composition with the mole fraction of paragonite reaching a maximum of 0.04 in R32387. This result agrees with Zen and Albee (*op. cit.*) who state that paragonite has not been recorded in rocks of lower grade than the biotite zone in regional

TABLE 12: X-ray diffraction data for white micas

Specimen	Parent Lithology	Zone	d(002)A°	Half-peak width
32361	psammite	I	10.063	0.45
32446	psammite		10.086	0.38
32358	psammite		10.109	0.30
32387	pelite	II	10.017	0.34
32434	pelite		10,065	0.30
32441	psammite		10.086	0.32
32705	pelite		10.086	0.34
32819	quartz-psammite		10.097	0.23
32373	psammite		10.109	0.20
32454	psammite		10.120	0.33
32371	pelite		10.132	0.28
32378	pelite		10.178	0.22
32452	psammite		10.225	0.23
32414	psammite	IIIa	10.063	0.26
32578	pelite		10.178	0.30
32600	psammite	IIIb	10.075	0.20
32616	pelite		10.075	0.32
32612	pelite		10.132	0.36
32615	pelite		10.155	0.32
32609	pelite		10.178	0.32
32594	psammite		10.202	0.28
32673	pelite	IV	10.075	0.25
32694	pelite		10.086	0.24
32588	psammite		10.144	0.35

metamorphic rocks. However at the present time, there are diverse opinions concerning the relationship between metamorphic conditions and white mica compositions in low-grade rocks (e.g. Brown 1967, 1968; Velde 1967, 1968).

Weaver (1960, 1961) noted the 10°A peak of white micas on X-ray diffraction charts became sharper with increasing grade of metamorphism. This relationship, termed "crystallinity", is quantified by measuring the width of the 10°A peak at half-peak width, and has been studied by numerous workers (e.g. Kubler 1966, 1968; Dunoyer de Segonzac *et al.* 1968; Leitch 1972) and reviewed by Dunoyer de Segonzac (1970, p.315). The results from the Coffs Harbour Block (Table 12, Fig. 14D) are not as conclusive as those of previous workers in other areas. There is an increase in the sharpness of the peak between Zones I and II but the crystallinity remains relatively constant from Zone II to Zone IV. This may be a result of M2 overprinting M1 for Zones III and IV. The increase in the peak sharpness has been ascribed to an increase in the "crystallinity" of the mica (Kubler 1966) and is probably related to an increase in the size of the reflecting crystals (Weaver 1961) and to an increase in ordering of the mica structure.

Chlorite

Chlorite is widespread in rocks from Zones I to III and the disappearance of it marks the commencement of Zone IV. Chlorite tends to be more abundant in psammitic rocks than pelitic varieties (Fig. 13B), and this also is valid for the Nambucca Fold Belt (Leitch 1972). In a few rocks from Zone IV the presence of chlorite was indicated by an extremely weak 14°A peak of the (001) basal reflection, although the mineral was not observed microscopically. Hence the commencement of Zone IV is marked by the microscopic disappearance of chlorite.

Chlorite occurs as small aggregates in the pelites and matrix of the psammites, and often occurs as "plates". It is generally pleochroic from almost colourless to pale green but a pale green-brown variety less commonly occurs.

Basal spacings d(001) were determined according to the following formula of Leitch (1972): $d(001) = [2.d(002) + 3.d(003) + 4.d(004)]/3$. The fields for basal spacings from each zone overlap and no sensible trend can be determined from the means (Fig. 15B). The chlorites therefore remain relatively constant with respect to their basal spacing, with increasing grade of metamorphism.

It is possible to calculate approximate compositions of chlorites if the following formula (from Albee 1962, modified by Leitch 1972) is assumed: $(Mg_{6-x-y}, Fe^{2+}_{y}, Al_{x})\ (Si_{4-x}\ Al_{x})\ O_{10}\ (OH)_{8}$. Only semi-quantitative data on compositions can be obtained by this method, which depends on the following assumptions:

(1) Cations such as Fe^{3+}, Cr, Mn are absent,

(2) Al is equally divided between the octahedral and tetrahedral lattice sites,

(3) There are no vacant lattice positions.

The degree to which the above conditions occur in analysed chlorites has been discussed by Hey (1954), Albee (1962) and Petruk (1964). The Al content is related to the (001) basal spacing (Hey 1954). Brindley (1961) modified the regression equation of Hey (*op. cit.*) to: d(001) = 14.55 - 0.29 x tetrahedral Al. The Fe^{2+} content was determined using the method of Petruk (1964) based on relative intensities of the (002), (003) and (004) reflections. The chlorite formulae (Table 13) can then be determined using the tetrahedral aluminium (x) and Fe^{2+}(y) values. The results are plotted on the nomenclatural grid of Hey (1954) with chlorites from the Nambucca Fold Belt (Leitch 1972) for comparison (Fig. 15A).

The chlorite compositions are plotted on an Al-Fe-Mg triangular diagram (Fig. 15C) and chlorites from psammitic rocks exhibit a more uniform composition than do those from pelitic rocks. Comparison of the field with those from the Nambucca Fold Belt (Leitch 1972) and Sanbagawa Schists indicates a very similar range of composition is developed despite some differences in grade (Fig. 15D). The effect of grade on chlorite compositions is not fully understood and several contrasting results have been published. Iwasaki (1963), Banno (1964) and Horikoshi (1965) all describe an increase in Fe content with grade in Sanbagawa metabasic rocks but Katada (1965) described a reverse trend for the basic rocks of the Ryoke Belt. Banno (*op. cit.*) believes a decrease in Fe content occurs with increasing grade for pelitic rocks and Tagiri (1973) describes a decrease in the Fe content of chlorites from Abukuma Plateau with increasing grade. In the pelites from the Coffs Harbour Block a random pattern was observed.

Mg/Fe ratio varied widely while Al/Mg + Fe remained relatively uniform for greenschist facies chlorites of Eastern Otago (Brown 1967). A similar conclusion may be drawn for chlorites from the Coffs Harbour Block (see Table 13). Ernst *et al.* (1970) showed the ratio Si/Si + Al decreased with an

Table 13: X-ray diffraction and semi-quantitative chemical data for chlorites

Specimen	Parent Lithology	Zone	d(001)°A	Mg	Fe^{2+}	Al^{IV}	Si	Al^{VIII}	Mg/Fe	Al/Mg + Fe	Si/Si + Al
32395	Pelite	I	14.197	1.4	3.4	1.2	2.8	1.2	0.41	0.50	0.54
32439	Psammite		14.213								
32446	Psammite		14.221	2.6	2.3	1.1	2.9	1.1	1.13	0.45	0.57
32358	Psammite		14.222								
32361	Psammite		14.261								
32387	Pelite	II	14.105								
32434	Pelite		14.126								
32388	Psammite		14.173	2.2	2.5	1.3	2.7	1.3	0.88	0.55	0.51
32415	Pelite		14.182								
32724	Metabasite		14.192	2.7	2.1	1.2	2.8	1.2	1.29	0.50	0.54
32373	Psammite		14.196	1.9	2.9	1.2	2.8	1.2	0.66	0.50	0.54
32452	Psammite		14.221								
32705	Pelite		14.230	1.1	3.8	1.1	2.9	1.1	0.29	0.45	0.57
32378	Pelite		14.279	3.7	1.4	0.9	3.1	0.9	2.64	0.35	0.63
32371	Pelite		14.287								
32819	Quartz-Psammite		14.298								
32414	Psammite	IIIa	14.192	2.6	2.2	1.2	2.8	1.2	1.18	0.50	0.54
32633	Pelite		14.218								
32534	Psammite		14.222								
32785	Psammite		14.231	3.3	1.6	1.1	2.9	1.1	2.06	0.45	0.57
32380	Psammite		14.241								
32764	Psammite		14.264								
32612	Pelite	IIIb	14.204								
32615	Pelite		14.226	2.4	2.5	1.1	2.9	1.1	0.96	0.45	0.57
32616	Pelite		14.238	1.8	3.1	1.1	2.9	1.1	0.58	0.45	0.57
32583	Psammite		14.250								
32672	Pelite		14.287	3.9	1.2	0.9	3.1	0.9	3.25	0.35	0.63
32694	Pelite	IV	14.182								
32673	Pelite		14.188	2.8	2.0	1.2	2.8	1.2	1.40	0.50	0.54
32686	Pelite		14.204								
32588	Psammite		14.211	2.5	2.3	1.2	2.8	1.2	1.09	0.50	0.54
32681	Pelite		14.241								
32587	Pelite		14.310								

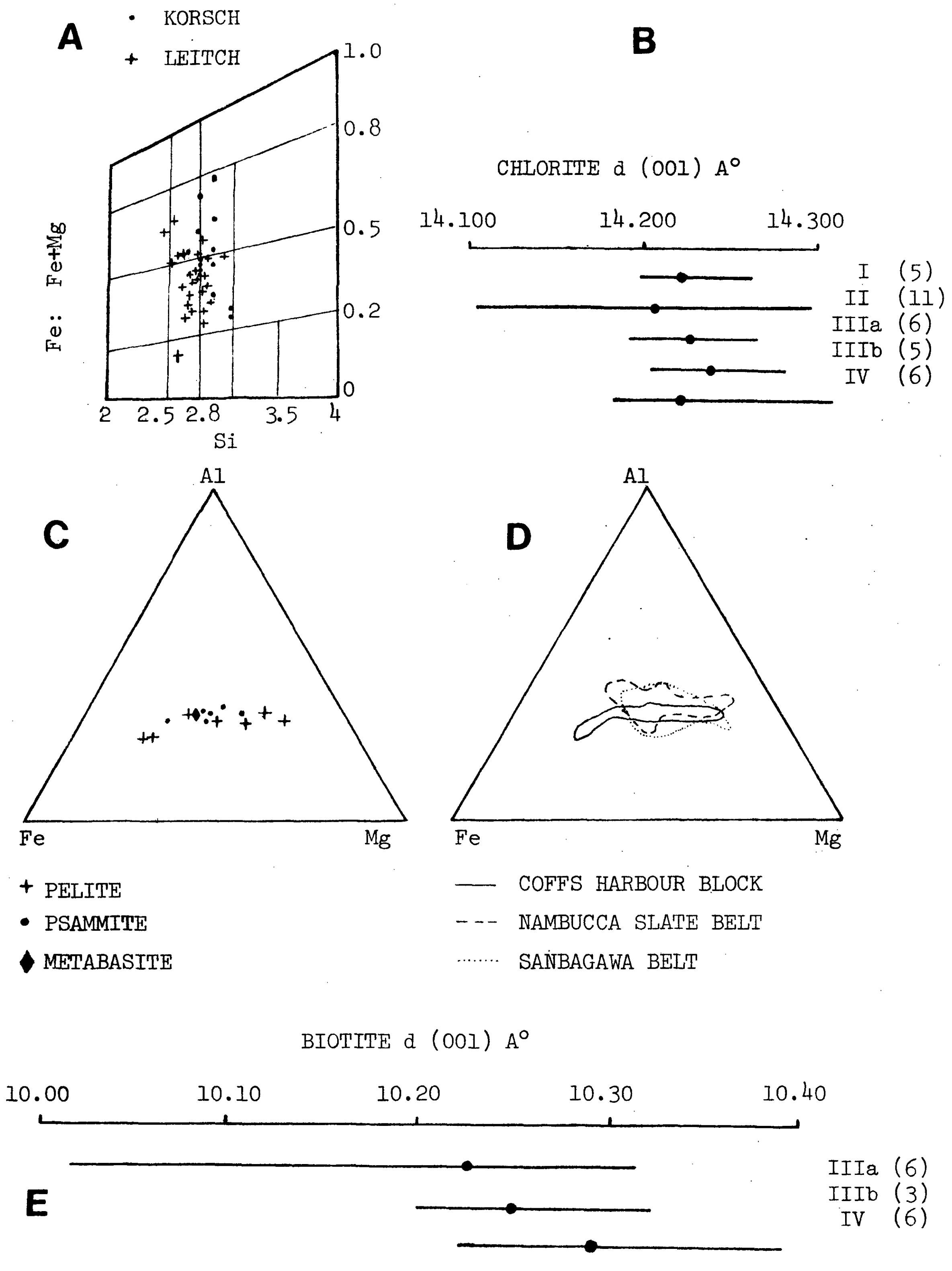


Fig. 15 A-D: Diagrams for chlorite. For descriptions see Appendix I pp. 63-65 and Table 13.

E: Range of d (001) A° basal spacings for biotite within the metamorphic zones. Bar indicates range and filled circle is the mean.

increase in grade in the Sanbagawa Schists. Leitch (1972) confirmed this for chlorites from the Nambucca area. Mean values for chlorites from the Coffs Harbour Block are: Zone I - 0.56; Zone II - 0.56; Zone III - 0.59; Zone IV - 0.54, and the results are inconclusive.

Data presented above show that chlorites from metasediments of the Coffs Harbour Block are very similar to chlorites in other areas of comparable grades of metamorphism.

Biotite

Biotite is the index mineral for Zone III, occurring in all lithological types but is more abundant in pelitic than psammitic rocks, particularly in Zone IIIa. The pleochroic schemes show a wide range. The two most common schemes are α = pale straw; $\beta\gamma$ = red-brown (e.g. S32534) and α = yellow-brown; $\beta\gamma$ = drab brown (e.g. S32533). In S32647 two pleochroic schemes (α = pale straw, $\beta\gamma$ = khaki-brown and α = pale green, $\beta\gamma$ = green) are end members of a series of pleochroic schemes.

White (1964) described stilpnomelane from the Brisbane Metamorphics which had previously been identified as biotite. Several authors (e.g. Iwasaki 1963, Brown 1967, Leitch 1972) have described stilpnomelane from rocks of similar metamorphic grade to the Coffs Harbour Block. Hence several samples were crushed and the brown mica separated to a concentrate which was X-rayed in an attempt to distinguish between stilpnomelane and biotite. All fifteen samples (Table 14A) examined showed a strong 10°A peak indicating the presence of biotite. Stilpnomelane was not observed in any specimen from the Coffs Harbour Block.

The means of the basal spacings (Fig. 15E) show a progressive increase from Zone IIIa to Zone IV even though there is considerable overlap in the fields of each zone. This may indicate a relationship between metamorphic grade and the basal spacings of biotite. A complete X-ray pattern for one biotite is listed with a biotite from the greenschist facies of eastern Otago (Brown 1967) for comparison (Table 14B). Apart from the (001) and (110, 020) reflections these two samples show a remarkable similarity to each other.

Epidote Group

Minerals of the epidote group occur in all rock types of Zones I to IIIa and are more abundant in the psammites than the pelitic rocks (Fig. 13B). The boundary between subzones IIIa and IIIb was delineated by the absence of

TABLE 14A: X-ray data for biotites

Specimen	Parent Lithology	Zone	d(001)°A
32498	psammite	IIIa	10.017
32633	pelite		10.202
32764	psammite		10.249
32380	psammite		10.249
32785	psammite		10.309
32534	psammite		10.345
32614	pelite	IIIb	10.155
32672	pelite		10.273
32583	psammite		10.321
32686	pelite	IV	10.225
32679	pelite		10.249
32682	pelite		10.297
32681	pelite		10.297
32801	psammite		10.297
32587	pelite		10.394

TABLE 14B: X-ray data for biotite sample 32534 with one listed by Brown (1967) for comparison

	Sample 32534		668-60 (Brown)	
hkl	A°	intensity	A°	intensity
001	10.35	100	10.15	strong
110,020	4.51	3	4.62	weak
			4.04	very weak
003	3.37	100	3.36	moderate
112	3.17	6	3.16	very weak
11$\bar{3}$	2.95	5		
20$\bar{1}$, 130	2.64	12	2.63	strong
004, 130	2.53	19	2.52	weak
201	2.45	9	2.45	moderate
040, 132	2.28	5	2.29	very weak
	2.19	9	2.18	moderate
	2.02	14		
	1.92	3		
			1.68	weak
			1.54	moderate

epidote, which is probably lithologically controlled. Rare epidote grains occur in S32602 and S32584, both of which are psammites and occur in Zone IIIb near the Zone IV boundary. Leitch (1972) reported epidote in a psammite (S15718) from Zone IV. It is not possible to determine whether the epidote is detrital or metamorphic but Leitch believes it could have formed at the same time as the development of the biotite.

Both epidote and clinozoisite occur but the colourless clinozoisite, often with anomalous blue birefringence colours, is confined mainly to veins within the sediments of Zone II and is often associated with quartz and chlorite. Epidote varies from colourless to pleochroic yellow to pale green and has higher birefringence colours than the clinozoisite. Both species occur as idioblastic to xenoblastic crystals (Plate 10C) and are often difficult to distinguish from detrital epidote, particularly in the lowest-grade rocks. Several workers (e.g. Miyashiro and Seki 1958; Banno 1964) have shown that epidote composition changes as the grade of metamorphism increases. However no optical or chemical work was undertaken on epidotes from the Coffs Harbour Block.

Prehnite

Prehnite, along with pumpellyite, is the index mineral for Zone I and occurs commonly as spongy disoriented aggregates up to 1 mm in diameter and is also a very common vein mineral (Plate 10D), often associated with quartz in veins up to 3 mm wide. It has a milky white appearance in thin section and often exhibits undulose extinction. Prehnite occurs in most of the psammitic rocks from Zone I but is not as common in the pelitic varieties.

Prehnite was recorded in two specimens from Zone IV (S32671, S32679) where it occurs in veins approximately 0.3 mm wide. The veins truncate biotite flakes and hence is a late stage event occurring after M2 and cannot be related in time to the prehnite developed in Zone I which formed during M1.

Pumpellyite

Pumpellyite occurs only in Zone I and is less common than prehnite. All pumpellyite recognised exhibited characteristic blue-green pleochroism and occurred as disoriented aggregates of tiny xenoblastic crystals. A change in the colour of pumpellyite from colourless to green was recorded in the Lachlan Geosyncline by Smith (1969) in his Zone 3 (prehnite-pumpellyite zone). This is the reverse of the situation recorded by Seki (1961), Iwasaki (1963) and Bishop (1972) who all record that clear pumpellyite is stable at a higher grade than the green variety. In the Lachlan Geosyncline the two varieties occur in different lithologies and bulk rock composition may be the critical factor.

No colourless pumpellyite was found in rocks of higher-grade than the ones in which the blue-green pumpellyite occurs. Rocks of lower grade than those containing blue-green pumpellyite do not occur in the field area and hence the findings of Smith (1969) cannot be conclusively confirmed.

Minor Metamorphic Minerals

Actinolite is rare in the metasediments of the Coffs Harbour Block, being a mineral more characteristic of metabasic rocks. It occurs in a Zone II psammite (S32381) and Leitch (1972) reported the occurrence of it from a Zone IV psammite (S15719) near the Bellinger Fault. Actinolite is found in a Zone I rock from the Rockvale Block (S32806) in which prehnite also occurs. The presence of actinolite is largely restricted to rocks having detrital hornblende, where the actinolite occasionally occurs as minute fringing needles. Usually, however, hornblende alters to chlorite. Hence in contrast to the Nambucca Fold Belt (Leitch 1972) the actinolite in sediments from the Rockvale - Coffs Harbour region, because of its scarcity, cannot be used as an index mineral.

Garnet occurs as minute colourless xenoblastic crystals (Plate 10E) in two siltstone specimens (S32612 in Zone IIIb, S32645 in Zone IV). Both parent rocks are siliceous but the garnet content is less than 1% and very fine grained preventing optical or X-ray examination. Smith (1969), Jolly (1970) and Leitch (1972) reported garnet or hydrogarnet in rocks of comparable grade to those of the Coffs Harbour Block.

Calcite has a random distribution in all lithological types. Very similar parent material can vary widely in their content of this phase. Calcite is also a common vein mineral and sometimes shows evidence of strain in the form of bent twin lamellae and undulose extinction.

Tourmaline is rare and usually occurs as outgrowths from fractured detrital grains of tourmaline. However one specimen from the Rockvale Block (S32767, Plate 10F) has radiating needles of secondary tourmaline up to 0.5 mm long which vary from colourless to pleochroic green-brown varieties.

Sphene is a rare accessory mineral and occurs associated with chlorite in the breakdown of detrital biotite in psammitic rocks. It also occurs in association with the secondary biotite in Zones III and IV.

Opaque Minerals are present in most thin sections with pyrite and haematite being the two most common species. Pyrite occurs in most rock types, often with distinctive cubic outlines (Plate 10G) but is common in pelitic and siliceous rocks. Haematite occurs mainly in siliceous rocks which show a characteristic red colouration.

Metamorphic Assemblages and Mineral Reactions

The metamorphic episodes outlined above will now be discussed in terms of their metamorphic assemblages and possible metamorphic reactions. Assemblages and their observed distribution in the zonal scheme are shown in Table 15. Restriction of assemblages to certain zones does not necessarily imply that they are critical to the recognition of that zone. Restriction of a number of these assemblages (e.g. 23, 28, 30, 38) to one zone might be a result of incomplete sampling and some assemblages can be expected to range over several zones. It is obvious that the assemblages which are the least zonally restricted are also the most common (e.g. 17, 21, 34, 35, 50, 52).

Many of the metamorphic assemblages can be considered in terms of the appearance or disappearance of the phases prehnite, pumpellyite, chlorite, epidote, white mica, calcite and biotite. A triangular diagram with the coigns (Al_20_3 + Fe_20_3), Ca0, (Fe0 + Mg0) has been used (Fig. 16) to illustrate relationships between the assemblages for Zones I to IV. By constructing tie lines between co-existing phases it is possible to see chemical relationships between assemblages. Mineral compositions listed by Deer *et al.* (1962-1963, Vols. 1-5) are utilized in determining mineral positions in Fig. 16. Chlorite is the one exception where semi-quantitative data indicate that chlorite compositions are in the range of A = 17-22, F = 78-83, C = 0. In drawing tie lines it is assumed that chlorite has a composition halfway between the two end members.

Leitch (1972) made several assumptions as to the state of the rock systems that he studied, and they are adopted in this study also:

1. The assemblages are equilibrium assemblages.
2. All assemblages contain quartz and hence $Si0_2$ is an excess component.

Table 15: Metamorphic assemblages in the metasedimentary rocks of the Coffs Harbour Block

	Assemblage	Zone I	Zone II	Zone IIIa	Zone IIIb	Zone IV
1.	quartz-albite-prehnite-calcite	X				
2.	quartz-albite-chlorite-prehnite-calcite	X				
3.	quartz-albite-chlorite-epidote-prehnite-(± opaques)	X				
4.	quartz-albite-chlorite-epidote-prehnite-calcite	X				
5.	quartz-albite-chlorite-epidote-white mica-prehnite	X				
6.	quartz-albite-chlorite-epidote-white mica-prehnite-calcite	X				
7.	quartz-albite-epidote-prehnite-calcite	X				
8.	quartz-albite-white mica-prehnite	X				
9.	quartz-albite-chlorite-epidote-prehnite-pumpellyite	X				
10.	quartz-albite-chlorite-epidote-pumpellyite	X				
11.	quartz-albite-chlorite-epidote-white mica-pumpellyite	X				
12.	quartz-albite-chlorite-epidote-white mica-pumpellyite-calcite	X				
13.	quartz-albite-epidote-pumpellyite	X				
14.	quartz-albite-chlorite	X	X		X	
15.	quartz-albite-chlorite-epidote	X	X	X		
16.	quartz-albite-chlorite-epidote-white mica	X	X	X		
17.	quartz-albite-chlorite-white mica	X	X	X	X	
18.	quartz-albite-chlorite-epidote-calcite	X	X			
19.	quartz-albite-chlorite-epidote-white mica-calcite	X	X	X		
20.	quartz-albite-chlorite-white mica-calcite-opaques	X		X		
21.	quartz-albite-white mica	X	X	X	X	X
22.	quartz-albite-chlorite-epidote-calcite-actinolite		X			
23.	quartz-albite-chlorite-epidote-opaques		X			
24.	quartz-albite-chlorite-epidote-white mica-(± opaques)		X			
25.	quartz-albite-chlorite-epidote-tourmaline		X			
26.	quartz-albite-chlorite-epidote-calcite-(± opaques)		X			
27.	quartz-albite-chlorite-calcite-(± opaques)		X			
28.	quartz-albite-epidote		X			
29.	quartz-albite-epidote-white mica-(± tourmaline)-(± opaques)		X			
30.	quartz-albite-calcite-opaques		X			

Table 15 (continued)

Assemblage	Zone I	Zone II	Zone IIIa	Zone IIIb	Zone IV
31. quartz-albite-chlorite-epidote-white mica-tourmaline		X			
32. quartz		X	X	X	X
33. quartz-albite		X	X		X
34. quartz-albite-chlorite-white mica-calcite		X	X	X	
35. quartz-albite-white mica-calcite		X	X	X	X
36. quartz-albite-calcite		X			X
37. quartz-albite-white mica-calcite-opaques			X		
38. quartz-albite-epidote-white mica-calcite			X		
39. quartz-albite-chlorite-biotite			X		
40. quartz-albite-chlorite-epidote-biotite			X		
41. quartz-albite-chlorite-epidote-white mica-biotite-(± tourmaline)			X		
42. quartz-albite-chlorite-epidote-white mica-calcite-biotite			X		
43. quartz-albite-chlorite-epidote-calcite-biotite			X		
44. quartz-albite-chlorite-epidote-biotite-sphene			X		
45. quartz-albite-chlorite-white mica-biotite-sphene			X		
46. quartz-albite-epidote-biotite			X		
47. quartz-albite-epidote-white mica-biotite			X		
48. quartz-albite-epidote-white mica-calcite-biotite			X		
49. quartz-albite-epidote-biotite-sphene			X		
50. quartz-albite-biotite			X	X	X
51. quartz-albite-chlorite-white mica-biotite			X	X	
52. quartz-albite-white mica-biotite			X	X	X
53. quartz-albite-white mica-calcite-biotite			X	X	X
54. quartz-albite-white mica-biotite-opaques			X	X	
55. quartz-albite-chlorite-white mica-calcite-biotite				X	
56. quartz-albite-white mica-biotite-garnet				X	
57. quartz-albite-chlorite-white mica-garnet				X	
58. quartz-albite-white mica-pyrite				X	
59. quartz-albite-calcite-pyrite				X	
60. quartz-chlorite-white mica				X	
61. quartz-opaques				X	X

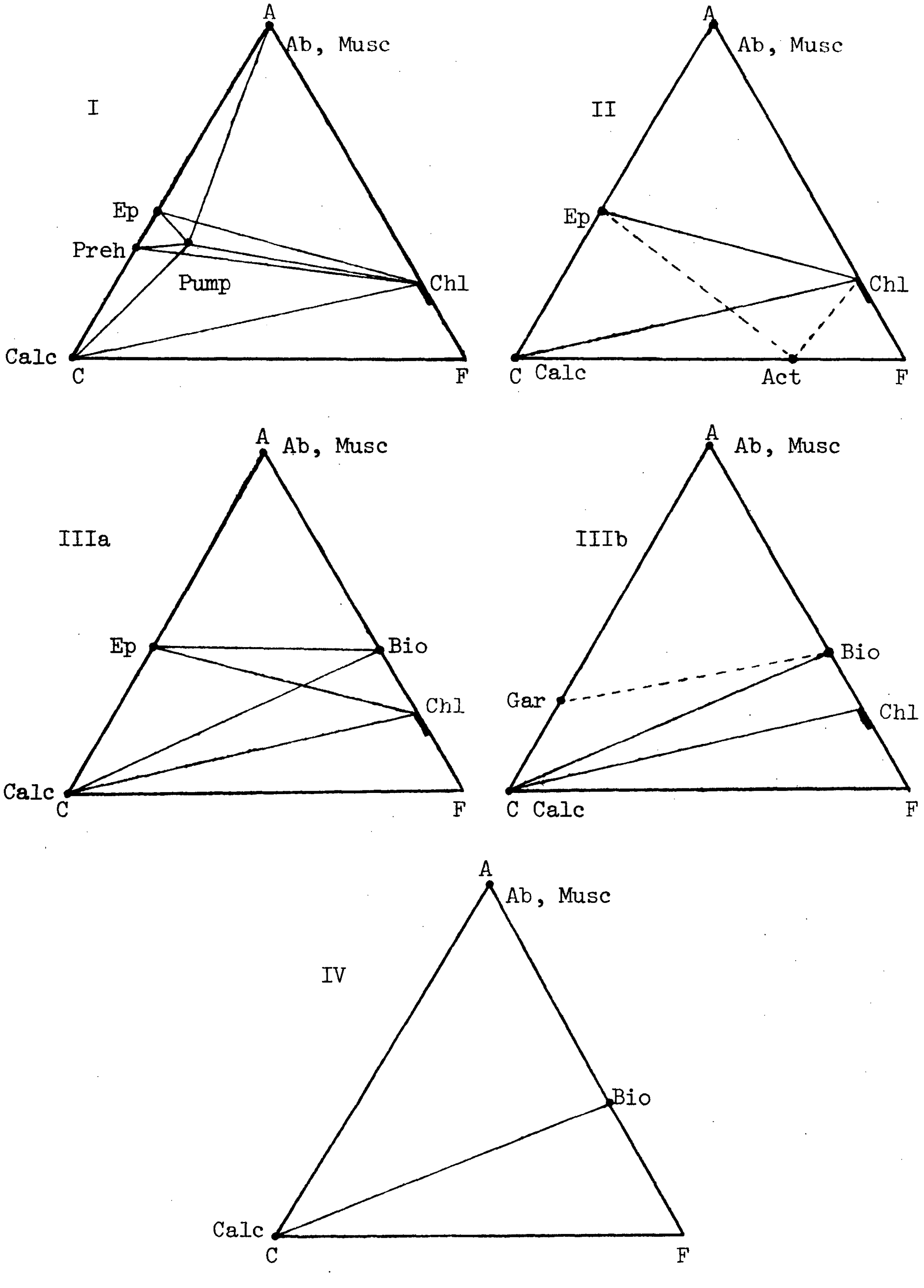


Fig. 16: ACF diagrams showing relationships between assemblages for metamorphic zones I to IV.

Key: A = Al_2O_3 + Fe_2O_3, C = CaO, F = FeO + MgO, Ab = Albite, Musc = Muscovite, Ep = Epidote, Preh = Prehnite, Pump = pumpellyite, Calc = Calcite, Chl = Chlorite, Act = Actinolite, Bio = average composition for biotite, Gar = Garnet.

3. FeO and MgO are isomorphous. Chlorite and biotite, the major Fe^{2+} and Mg silicates in the assemblages both allow considerable isomorphous exchange between these components. Stilpnomelane, which does not allow isomorphous exchange, was not recognised in any rock.
4. The system is closed to oxygen. If this is not assumed Fe_2O_3 cannot be considered distinct from FeO because the reaction $4FeO + O_2 \rightarrow 2Fe_2O_3$ would be externally controlled.
5. Al_2O_3 is sufficiently abundant to have combined with all Na_2O in the rock to form albite and with K_2O to produce muscovite or, at a higher grade, biotite. This assumption is justified by the assemblages because albite appears to be the only Na-bearing phase present, and white-mica and biotite the only K-bearing phases. Minor K-feldspar is assumed to be of relict detrital origin.
6. H_2O was a mobile and excess component.
7. CO_2 played an important role in determining phase assemblages. CO_2 can be absent, or present as an inert compound both in excess and not in excess, or can be a perfectly mobile component. There is little reason to expect that CO_2 played a similar role in all rocks but its influence on the development of a particular assemblage is difficult to identify with certainty.

Metamorphic Reactions

Assuming that the detrital components consisted of quartz, plagioclase, K-feldspar, hornblende and acid-intermediate volcanic lithic fragments, with a clay matrix, it is possible to present a series of hypothetical reactions which could have occurred to produce the metamorphic assemblages that are observed.

Zone I

The metamorphic minerals in this zone possibly arose from a series of reactions including the albitisation of plagioclase, conversion of illites to white mica and the development of chlorite. Frey (1970) showed that mixed layer illite/montmorillonite produced phengite and Al-rich chlorite in a transition zone between the diagenesis and greenschist facies, whereas Dunoyer de Segonzac (1970) showed that montmorillonites were transformed into chlorites in an Mg-rich environment and into illites in a K-rich environment. The illites were gradually converted into white mica. The zone of stability for the mixed-layer minerals has been shown by Muffler and White (1969) to be 80 - 200°C with depths of 1 - 3 km in the Salton Basin of California, where the geothermal gradient is exceptionally high. Hence it is possible to form the chlorites and white micas by the transformation of clay minerals.

The detrital plagioclases in the Coffs Harbour Block have an average original composition of An_{40} and consequently their Na-content can be used to produce albite. The extra Na required may be derived from clay minerals. Reaction (1) below (modified from Leitch 1972) is more likely than reactions (2) and (3):

(1) $2CaAl_2Si_2O_8 \underset{\uparrow}{\cdot} 3NaAlSi_3O_8 + H_2O + 5SiO_2 + Na_2O$

$$\text{(plagioclase An40)} = \underset{\text{(albite)}}{5NaAlSi_3O_8} + \underset{\text{(prehnite)}}{Ca_2Al_2Si_3O_{10}(OH)_2}$$

(2) analcime + quartz = albite + water (Coombs 1961)

(3) laumontite + calcite = prehnite + quartz + H_2O + CO_2 (Winkler 1967).

Reactions (2) and (3) are dependent on the presence of zeolite minerals of kinds having lower T and P stability conditions than those recorded in the Coffs Harbour Block. (They might have occurred however in rocks that have been removed by erosion).

Several workers have proposed reactions to produce pumpellyite but most involve the consumption of prehnite:

(4) laumontite + prehnite + chlorite = pumpellyite + quartz + H_2O (Coombs 1961)

(5) laumontite + calcite + chlorite = pumpellyite + quartz + H_2O + CO_2 (Martini *et al.*, *in* Winkler 1967, p.157).

(6) 4 prehnite + amesite – chlorite + H_2O = 2 pumpellyite + quartz (Seki 1961, p.420).

(7) prehnite + chlorite + quartz = pumpellyite + actinolite + H_2O (Hashimoto 1966)

(8) prehnite + chlorite + H_2O + CO_2 = pumpellyite + calcite + quartz (Coombs *et al.* 1970)

(9) prehnite + kaolinite = pumpellyite + quartz + H_2O (Liou 1971)

(10) prehnite + chlorite + H_2O = pumpellyite + actinolite + quartz (Nitsch 1971).

Because actinolite is virtually absent in Zone I, reactions (7) and (10) are unlikely. In most assemblages where pumpellyite occurs calcite is absent, and laumontite and kaolinite have not been detected. Consequently the only really valid reaction for the Coffs Harbour rocks is (6). Textural evidence supports this reaction because in S32368 chlorite has a rim of pumpellyite which in turn is surrounded by a rim of epidote. Prehnite occurs very close to these composite aggregates.

Epidote occurs throughout Zone I and may be produced by the breakdown of prehnite or pumpellyite or both. Some possible reactions (Nitsch 1971) are:

(11a) pumpellyite + quartz = prehnite + epidote + actinolite

(11b) prehnite + chlorite = pumpellyite + epidote + actinolite

(11c) pumpellyite + quartz = prehnite + epidote + chlorite.

Only (11c) is thought to have occurred, because actinolite is absent in Zone I.

Calcite in Zone I is associated with prehnite (assemblages 1, 2, 4, 6, 7), pumepllyite (assemblage 12) and also chlorite and epidote, with or without white mica (assemblages 18, 19). Possible reactions are:

(12) $\underset{\text{prehnite}}{Ca_2Al_2Si_3O_{10}(OH)_2} + \underset{\text{K-feldspar}}{KAlSi_3O_8} + 2CO_2 = \underset{\text{muscovite}}{KAl_3Si_3O_{10}(OH)_2} + 2CaCO_3 + 3SiO_2$

(13) prehnite + CO_2 = clinozoisite + calcite + quartz (Coombs *et al.* 1970).

It is probable that both reactions (12) and (13) as well as (8) have occurred because calcite and pumpellyite do not always co-exist and not every assemblage containing calcite contains muscovite or epidote or both.

Zone II

Entry into this zone is marked by the disappearance of both prehnite and pumpellyite. Several reactions have been proposed for the consumption of prehnite (reactions 4, 6, 7, 8, 9, 10, 11b, 12, 13). Because pumpellyite seems to disappear coincidently with prehnite, most of the above reactions would cease to occur at the boundary between Zones I and II. Favoured reactions for prehnite consumption are (12), (13) and:

(14) iron oxide + prehnite = epidote + H_2O (Leitch 1972, p.249).

Pumpellyite might disappear by one of the following reactions:

(15) pumpellyite + calcite + haematite + quartz = epidote + actinolite + H_2O + CO_2 (Iwasaki 1963, p.27).

(16) pumpellyite + quartz + CO_2 = epidote + actinolite + calcite + H_2O (Miyashiro and Seki 1958, *in* Bishop 1972).

(17) pumpellyite + chlorite + quartz = actinolite + clinozoisite + H_2O (Banno 1964, p.292).

(18) pumpellyite + antigorite chlorite + quartz = actinolite + epidote + H_2O (Seki 1969).

(19) pumpellyite + CO_2 = clinozoisite + chlorite + quartz + calcite + H_2O (Leitch 1972, p.231).

Because of the paucity of actinolite in Zone II reaction (19) is the most favoured. However, actinolite was observed in one specimen (S32487) where it was recrystallising from detrital hornblende, along the lines of the following equation:

(20a) $$\underbrace{NaCa_2(Mg, Fe, Al)_5(Si, Al)_8O_{22}(OH)}_{\text{hornblende}} + \underbrace{3SiO_2}_{\text{quartz}} + O_2 = \underbrace{Ca_2(Mg, Fe)_5\,Si_8O_{22}(OH)_2}_{\text{actinolite}} + \underbrace{NaAlSi_3O_8}_{\text{albite}}$$

This reaction is not common and hornblende is usually observed altering to chlorite, by two possible reactions:

(20b) hornblende + CO_2 + H_2O = chlorite + calcite

(20c) hornblende + H_2O = chlorite + epidote

Zones III and IV

The entry to Zone III is marked by the first appearance of biotite, plus minor muscovite. Several reactions have been suggested for the development of biotite:

(21a) phengite + chlorite = biotite + muscovite + SiO_2 + H_2O (Ernst 1963, p.1367).

(21b) phengite = muscovite + biotite + microcline + quartz + H_2O (Ernst 1963, p.1367).

(22) chlorite + quartz + KOH = biotite + muscovite + H_2O (McNamara 1966, p.408).

(23) microcline + chlorite = biotite + muscovite + quartz + H_2O (Winkler 1967, p.99).

(24a) phengitic-muscovite + chlorite = muscovite + chlorite + biotite + H_2O

(24b) microcline + chlorite + phengite = biotite + muscovite + quartz + H_2O

(24c) K-muscovite + chlorite = quartz = biotite + less K-rich muscovite + H_2O

(24d) 3 muscovite + 3 prochlorite = 3 biotite + 4 Al-rich chlorite + 7 quartz + 4 H_2O

Reactions 24 a - d are from Mather (1970).

(25) chlorite + microcline = biotite + muscovite (Brown 1971)

(26) chlorite + K-feldspar = biotite + muscovite + quartz (Hoschek 1973).

The existence of secondary microcline has not been recognised in Zone III and Zone IV rocks, and detrital K-feldspar is also rare. Hence reactions 21b, 23, 24b, and 25 possibly did not occur. Because chlorite disappears at the boundary between Zones III and IV, reactions having chlorite as a product (24a, 24d) are not valid and hence the most likely reactions appear to be 21a and 24c. For reaction 21a, the low-grade white mica is assumed to be phengite.

Subzones IIIa and IIIb are separated on the basis of the absence of epidote which is lithologically controlled and hence is not related to a mineral reaction.

The presence of garnet in two silica-saturated rocks from subzone IIIb contrasts with Leitch (1972) who concluded that the presence of garnet in some rocks of his Zone 2 was related to a silica-poor environment and that the garnet (thought to be grossular) may have been produced by either of:

(27) prehnite + CO_2 + H_2O = grossular + calcite

(28) prehnite = grossular + clinozoisite + quartz + water (Strens 1968).

These reactions have probably not occurred in Zone IIIb because of the excess silica present and the isograd marking the disappearance of prehnite is a significant distance (15 km) from the noted occurrences of garnet. Brown (1967) thought that garnets in the Otago Schists could have been due to either the onset of garnet-producing reaction in quartzo-feldspathic schists at this metamorphic grade, or slightly higher spessartine and grossular components in the rocks of the zone. The isolated occurrence of garnets in the Coffs Harbour Block supports the second hypothesis of Brown and that the host rock chemistry was suitable for the development of garnets at the Zone IIIb grade of metamorphism.

Opaque minerals, particularly haematite and pyrite, occur in rocks of all four zones and it is possible that haematite was produced by:

(29) epidote + actinolite + H_2O + CO_2 = chlorite + haematite + calcite + quartz (Leitch 1972).

Metamorphic Facies

Most reactions discussed above are considered to have helped to produce the observed zonal sequence of minerals. The concept of progressive metamorphism (Miyashiro 1973) has been accepted, and it implies that the higher-grade assemblages developed from their parent lithologies by way of the appropriate mineralogies in the lower grades. The reactions are a response to progressive regional changes in physical conditions.

Zone I

The presence of prehnite and pumpellyite and the absence of zeolites and actinolite place these rocks in the prehnite-pumpellyite facies of Coombs (1961). The presence of epidote in most assemblages from this zone indicates that the rocks occur in the upper part of the prehnite-pumpellyite facies where both prehnite and pumpellyite are decomposing to epidote (Miyashiro 1973, p.165).

Zone II

The absence of both prehnite and pumpellyite in rocks of similar composition to those of Zone I suggests that these rocks do not belong to the prehnite-pumpellyite facies. The typical assemblage of quartz-albite-chlorite-epidote-white mica indicates that the rocks occur in the lowest part of the greenschist facies below the biotite isograd.

Actinolite is not developed and pumpellyite disappears at the same time as prehnite. Hence the transitional pumpellyite-actinolite facies of Hashimoto (1966) is considered to be absent. It is possible that the pumpellyite-actinolite facies was not recognised because of the absence of metabasic lithologies in which actinolite usually develops. Seki *et al.* (1969) and Coombs *et al.* (1970) both consider the pumpellyite-actinolite facies to be suppressed in areas of low pressure-high geothermal gradient conditions and this has been confirmed experimentally by Nitsch (1971) and Liou (1971).

Texturally this zone is equivalent to Chl 1 - Chl 2 and hence is atypical of greenschist facies textural reconstitution and further suggests that the zone is of lowest greenschist facies. However Bishop (1972) reports Chl 1 - Chl 2 from the prehnite-pumpellyite or pumpellyite-actinolite facies in the Otago Schists.

Zone III

The appearance of biotite overprinting the previously developed white mica foliation presents a problem of designating the rocks to a facies. As far as can be determined all rocks of Zones III and IV were metamorphosed by M1 to greenschist facies of a similar to slightly higher grade than the rocks of Zone II. M2 is more akin to the facies of metamorphism observed around a contact aureole than to regional metamorphic facies. The presence of biotite is indicative of either upper greenschist facies or albite-epidote hornfels facies (Turner 1968). Because the rocks are incompletely reconstituted and because of the inferred static thermal conditions, the rocks are more akin to the latter facies. However the regional scale of this metamorphic event weighs against inclusion in the albite-epidote hornfels facies of Turner (1968) and it is preferred to place the M2 of Zone III in the low-pressure greenschist facies of Miyashiro (1973).

It is considered that hornblende-hornfels facies or amphibolite facies conditions were not reached because cordierite is not developed and the rocks are incompletely reconstituted.

Zone IV

Rocks of this zone are similar to Zone III and probably suffered greenschist facies regional metamorphism (M1) which was later overprinted by a regional-scale thermal metamorphism (M2) of low-pressure greenschist facies grade.

Facies Series

The concept of facies series of Miyashiro (1961) and extended by Seki (1969) can be applied to rocks of the Coffs Harbour Block. The sequence of facies produced by M1 of prehnite-pumpellyite to greenschist, with the absence of pumpellyite-actinolite is indicative of Low Pressure II Facies Series of Miyashiro (1973, p.298) and may be compared with the Tanzawa

Mountains (Seki *et al.* 1969). Actinolite may be absent only because of the lack of metabasic rocks and M1 may belong to the medium-pressure facies series similar to that described by Leitch (1972) from the Nambucca Fold Belt. However, stilpnomelane, indicative of medium-to high-pressure (Miyashiro 1973, p.263), is absent, possibly supporting the placement of M1 in the Low Pressure II Facies Series.

The M2 metamorphism of albite-epidote hornfels or low-pressure greenschist facies belongs to the low-pressure I facies series of Miyashiro and may be compared with the rocks from the Iritono area of the central Abukuma Plateau (Shido 1958).

Conditions of Metamorphism

Experimentally-determined equilibrium curves for reactions that help delineate the physical conditions of low-grade metamorphism are plotted on Fig. 17. Because these data have been determined for simple compositional systems and from synthetic minerals the results must be considered tentative. H_2O is the only fluid phase present in the experimental determinations and hence these results can be applied only to rocks which lack calcite.

M1 Conditions

The only usable boundary produced by M1 is the Zone I-II boundary, which is considered to mark the conversion from prehnite-pumpellyite to greenschist facies. Several workers (e.g. Landis and Rogers 1968; Hinrichsen and Schurmann 1969; Nitsch 1971; Liou 1971) have experimentally determined conditions involving the breakdown of prehnite and pumpellyite (see Fig. 17 for experimental curves). In the above investigations the breakdown of prehnite resulted in the development of actinolite, clinopyroxene or grossularite, all of which are not developed to any marked extent in the Coffs Harbour Block. Seki *et al.* (1969) showed that while actinolite was common in the Tanzawa Mountains, it did not occur in the presence of prehnite or pumpellyite, but appeared just after these two minerals had disappeared. There, actinolite is developed mainly in basic volcanic rocks and hence its absence from the Coffs Harbour Block may be lithologically controlled. Greywackes and other clastic lithologies typical of the Coffs Harbour Block do not occur in the Tanzawa Mountain Sequence (Seki *et al.* 1969).

Only one curve of Nitsch (1971) seems appropriate for conditions in the Coffs Harbour Block (curve C, Fig. 17). Here, the breakdown of pumpellyite plus quartz produced prehnite, epidote and chlorite but no actinolite and indicates a geothermal gradient of over 40°C/km. A lower geothermal gradient would result in the development of actinolite from the breakdown of pumpellyite. The work of Nitsch also indicates that actinolite results from the breakdown of prehnite (Curve D, Fig. 17). Therefore the lack of actinolite in the Coffs Harbour Sequence suggests that the application of Nitsch's results is not valid. However the absence of experimental data not involving the development of actinolite makes it necessary to use Nitsch's results to give the best approximation of M1 conditions. Because prehnite and pumpellyite disappear together the conditions must be close to those of the intersection of curves C and D (Fig. 17). Nitsch showed that the prehnite-pumpellyite-chlorite-quartz paragenesis at 2kb, can be stable up to 345°±20°C. A geothermal gradient of 40°C/km is consistent with the conclusion that M1 is a low pressure facies series, but would be inconsistent if the metamorphism was of medium pressure similar to that described by Leitch (1972) for the Nambucca Fold Belt, and Bishop (1972) from Otago. Both these authors require a geothermal gradient of 15° - 25°C/km to produce their metamorphic assemblages.

Conditions for the Zone I-II boundary for a geothermal gradient of 40°C/km are presented in Table 15B. The results of Liou (1971) agree reasonably well with those of Nitsch (1971). However the conditions for the lower limit of the greenschist facies postulated by Turner (1968) are much lower than those determined experimentally by Nitsch and Liou. The upper limit of the greenschist facies was not reached by M1 but the conditions postulated by Turner (*op. cit.*) appear to be too low if the experimental results of Nitsch and Liou are valid. The work of Landis and Rogers (1968) and Hinrichsen and Schurmann (1969) is not applicable to conditions for M1.

M2 Conditions

The appearance of biotite marks the Zone II-III boundary and the start of M2 conditions. The appearance of biotite in pelitic rocks is thought to be due to reactions between low-grade muscovites and chlorites (e.g. Tilley 1926; Brown 1967).

Winkler (1957) experimentally produced muscovite-chlorite-quartz from illite-bearing clays at 400°C and P_{H20} of 2 kb, and biotite at about 550°C and

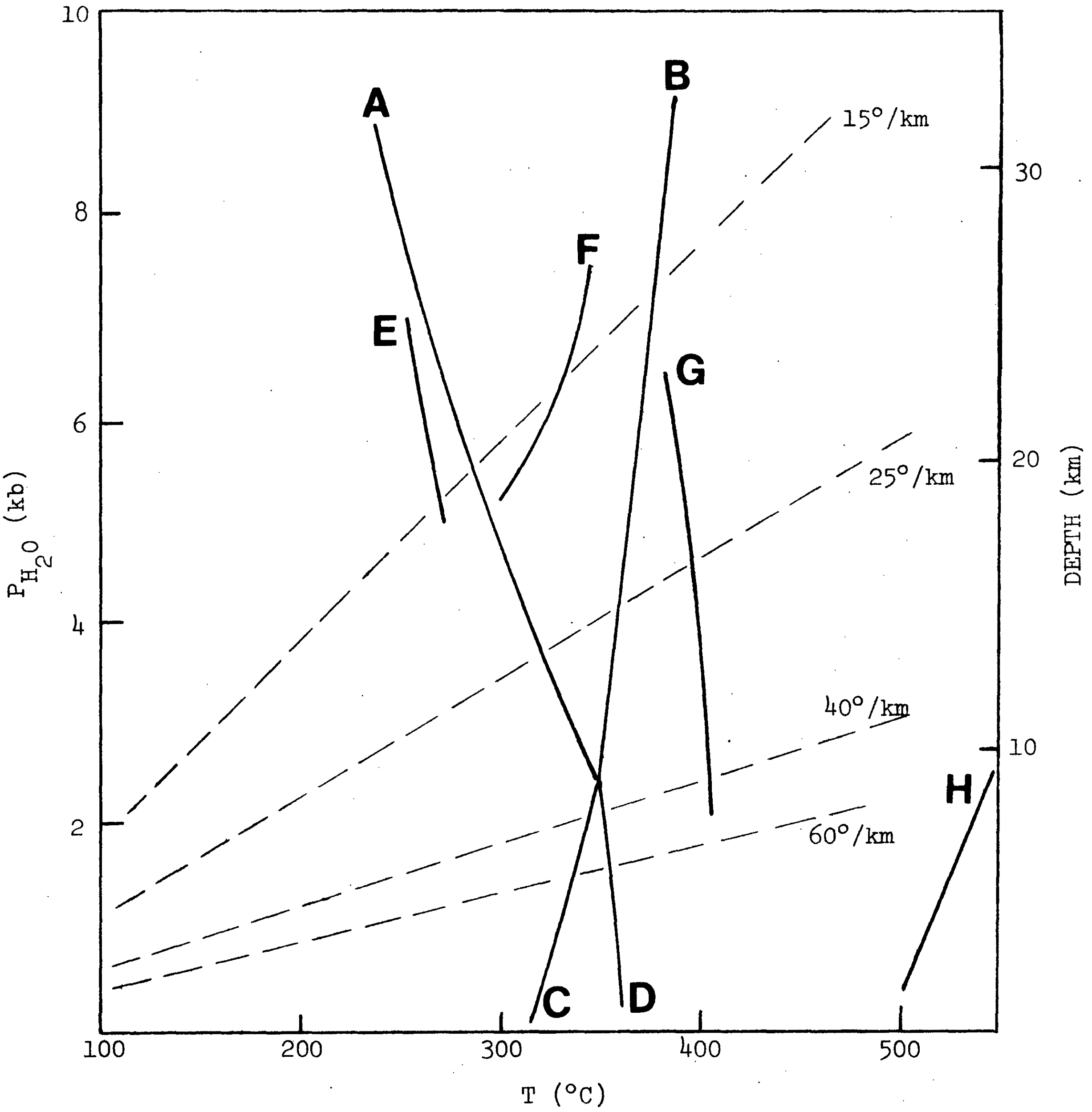


Fig. 17: Experimentally determined equilibrium curves for reactions used in defining P, T conditions during M1.

A: Preh + Chl + H_2O→Pump + Act + Qz (Nitsch 1971).
B: Pump + Chl + Qz→Ep + Act + H_2O (Nitsch 1971).
C: Pump + Qz→Preh + Epi + Chl (Nitsch 1971).
D: Preh + Chl + Qz→Epi + Act (Nitsch 1971).
E: Preh + Chl + H_2O→Pump + Qz (Hinrichsen & Schurmann 1969).
F: Pump→Zois + Gross + Chl + Qz (Hinrichsen & Schurmann 1969).
G: Preh→Zois + Gross + Qz + H_2O (Liou 1971).
H: Pump + Qz→An + Cpx (Landis & Rogers 1968).

Key: Qz - quartz, Preh - prehnite, Pump - pumpellyite, Chl - chlorite, Act - actinolite, Ep - epidote, Zois - zoisite, Gross - grossularite, An - anorthite, Cpx - clinopyroxene.

TABLE 15a: Approximate T, P and depth of burial suggested for M1 from the graphs of Liou (1971), Nitsch (1971) and Turner (1968), assuming a geothermal gradient of 40°C/km

Author	Nitsch	Liou	Turner	
Event	Disappearance of prehnite-pumpellyite (Zone I-II boundary)	Disappearance of prehnite (Zone I-II boundary)	Lower limit of green-schist facies (Zone I-II boundary)	Upper limit of green-schist facies (condition not reached)
T (°C)	360	402	260	440
P (kb)	2.2	2.4	1.9	2.8
Depth of burial (km)	9	10	6.5	11

the same pressure. Turner (1968, p.117) criticised these results because of evidence of disequilibrium and stated that the only possible conclusion was that biotite first appeared in pelitic rocks at P_{H2O} of 2 kb at some unknown temperature below 550°C. The true nature of the critical reaction at the biotite isograd is still unknown (cf Brown 1967) and hence P-T conditions for the appearance of biotite are not known. Turner (1968, p.118) considers that the small amount of available experimental data are consistent with the first development of biotite between 300°C and 400°C and a few kb of water pressure. The work of Garlick and Epstein (1967) and Epstein and Taylor (1967) on oxygen isotopes has shown that, for medium-pressure rocks of the Appalachians, biotite occurs at 350 - 400°C. It would be expected that for the low pressure M2, biotite would possibly appear at a lower temperature than those recorded for the Appalachian rocks.

If the first appearance of biotite can be used to define the lower limit of the albite-epidote hornfels facies then the following conditions derived from Turner (1968, p.258) may have occurred. For a geothermal gradient of 40°C/km T = 430°, P = 3.0 kb and depth of burial = 10.8 km; and for a geothermal gradient of 60°C/km T = 390°, P = 1.8 kb and depth of burial = 6.5 km.

Conditions observed in the Coffs Harbour Block did not reach the hornblende-hornfels facies and hence the conditions for the albite-epidote hornfels facies upper boundary of T = 480°, P = 2.1 kb depth = 7.5 km for a geothermal gradient of 60°C/km were not reached. It is possible that the extensive belt of biotite-grade rocks (up to 40 km wide) may be due to a sub-horizontal biotite isograd produced by a subsurface source of heat such as an extensive, continuous concealed batholith.

The Coffs Harbour Block metamorphism occurred in a synclinal structure complicated by mesoscopic folding and minor reverse faulting. The geological history of the region does not favour stratigraphic thicknesses of 14 - 27 km required for geothermal gradients of 15 - 25°C/km. The depths of burial required for a geothermal gradient of 40°C/km (see Table 15B) appear more probable than those of lower gradients.

Age of Metamorphism and relation to deformation

In the southern part of the Coffs Harbour Block the slaty cleavage results, in part, from the preferred orientation of metamorphic phases, particularly white mica. Consequently there is a close temporal relationship between M1 and D1 episode of deformation. M2 is post-tectonic, occurring as a static event producing randomly-oriented biotite after the directed pressure of D1 had been removed.

There is evidence that, although most of the metamorphic phases occurred during and after D1, crystallisation extended over a longer period. Quartz veins, varying in age from pre-M1 to post-M2 have been recognised. It is considered that M1 is contemporaneous with the metamorphism in the Wongwibinda Complex (Binns 1966) and the Nambucca Fold Belt (Leitch 1972) and hence is probably Middle Permian. Leitch (*in prep.*) has determined an age of 250 m.y. for the metamorphism in the Nambucca Fold Belt. M2 is possibly associated with the intrusion of the New England Batholith in Late Permian time.

C. DYAMBERIN BLOCK

The rocks of the Dyamberin Block have suffered incipient to low-grade regional metamorphism and the following assemblages were observed:

(i) quartz-albite-chlorite ± calcite (2 specimens)
(ii) quartz-albite-chlorite-white mica ± opaques (9 specimens)
(iii) quartz-albite-chlorite-epidote-calcite (1)
(iv) quartz-albite-white mica (3)
(v) quartz-albite-chlorite-white mica-calcite (8)
(vi) quartz-albite-white mica-calcite (5)
(vii) quartz-albite-calcite (1)

All specimens examined were psammites and pelites except one metabasite (S32724, assemblage iii) and one acid volcanic (S32718, assemblage vi).

No diagnostic minerals such as prehnite, pumpellyite or actinolite were observed. Hence these rocks are tentatively correlated with Zone II of the Coffs Harbour Block and accordingly are placed in the lower greenschist facies. However calcite is a widespread phase, the chemical potential of CO_2 was high and consequently prehnite has not developed. Therefore it is possible that these rocks may be isofacial with Zone I from the Coffs

Harbour Block.

Biotite, presumably metamorphic, was found in 6 out of 35 specimens. No pattern of regional significance could be recognised and the following assemblages were recorded:

1. quartz-albite-biotite-white mica (S32713, S32714, S32699)
2. quartz-albite-biotite-white mica-chlorite (S32698, S32703)
3. quartz-albite-biotite-white mica-chlorite-calcite (S32733)

The biotite-bearing samples occurred in three main areas: within 1 km of the Wongwibinda Fault; less than 2 km west of the Demon Fault; and, near Dyamberin mine where Cumming (1971) found evidence of movement along joint planes and a later incoming of ore material. It is considered that the biotite is not related to the regional metamorphism, but may be the result of movement along faults, or less likely, the relict of a higher-grade metamorphic episode which has escaped subsequent downgrading.

X-ray data on minerals from R32705 (slate, Plate 10H) and R32724 indicate they are similar in properties to those described previously from the Coffs Harbour Block.

The age of the regional metamorphism is not known, but must post-date the deposition of the late Carboniferous-Permian Dyamberin Beds, and is probably middle Permian and associated with the orogenesis that occurred throughout the New England area.

D. ROCKVALE BLOCK

Metamorphism in this block can be divided into three types:

1. Low-grade regional metamorphism, and an associated metamorphism of apparent thermal origin, similar to that recorded from the Coffs Harbour Block.
2. Medium-to high-grade metamorphism producing schists and migmatites with associated gneissic granites in the Wongwibinda Complex.
3. Contact metamorphism around granitic intrusions.

1. Low-grade regional metamorphism

Sediments of the Rockvale Block can be subdivided in zones similar to the Coffs Harbour Block (Map 2) and hence will be briefly discussed because of the comprehensive treatment already given to the Coffs Harbour Block.

Zone I

Rocks of zone I (prehnite-pumpellyite facies) occur in the south-western corner of the field area adjacent to prehnite-bearing rocks described by Traise (1973). Actinolite altering from detrital hornblende occurs in one thin section (S32806, assemblage : quartz-albite-chlorite-epidote-actinolite-prehnite) and hence cannot be related to the breakdown of prehnite. Other zone I assemblages observed are: 4, 6, 14, 17, 19 (numbers refer to Table 15).

Zone II

Rocks similar to zone II of the Coffs Harbour Block occur in the Girrakool Beds north of the Rockvale Adamellite-Granodiorite. They appear to grade eastwards into the Zone of Transitional Schists, which marks the western Boundary of the Wongwibinda Complex. Binns (1966, p.26) concluded that no outer zone of greenschist facies existed at the edge of the Wongwibinda Complex and attributed this to the Girrakool Beds suffering widespread heating by intrusive rocks while the deformation and metamorphism were occurring in the Wongwibinda area. The zone II portion of the Girrakool Beds contains mineral assemblages typical of greenschist facies. However the amount of internal deformation and textural reconstitution is slight, being similar to that found in the Coffs Harbour Block (see Plates 11A and 11B). The absence of prehnite is grade-dependent because zone II rocks are lithologically very similar to those of zone I.

Zones III and IV

These two zones have been grouped together because they have been difficult to separate in the Rockvale Block. Rocks of this zone occur in three different areas within the Rockvale Block (Map 2) near "Girrakool" (GR 5008 2408), "Clifton" (GR 4870 2481) and west of the Abroi Granodiorite, south of the Rockvale Adamellite-Granodiorite.

Typical assemblages (from 46 thin sections) are: 21, 40, 46, 47, 48, 50, 51, 52, 53, 55 (refer to Table 15) as well as

1. quartz-albite-white-mica-tourmaline (S32767), and
2. quartz-albite-chlorite-white mica-biotite-tourmaline (S32784).

PLATE 11

A. Weakly developed foliation in a fine-grained siltstone. Girrakool Beds. S32790, magnification x 20, plane polarised light.

B. Recrystallised matrix producing white mica with a pronounced preferred orientation in a lithic greywacke. Girrakool Beds. S32781, magnification x 50, plane polarised light.

C. Mylonite produced by movement on the Wongwibinda Fault at the Kangaroo Creek crossing on the Guyra - Ebor Road. Magnification x 20, plane polarised light.

D. Effect of movement on the Demon Fault in a granitic rock from the Chaelundi Complex producing a cataclasite with severe brecciation and granulation to produce an amorphous rock flour. S32835, magnification x 20, plane polarised light.

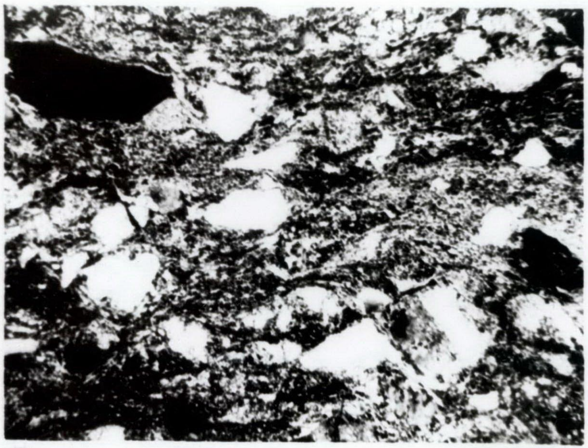
E. Effect of movement on the Bellinger Fault transposing bedding and cleavage along closely-spaced slip surfaces in the Brooklana Beds. S32653, magnification x 20, plane polarised light.

F. Brecciation and subsequent quartz veining in a pelite induced by movement on the Bellinger Fault. Brooklana Beds. S32639, magnification x 20, plane polarised light.

G. Augen structures with slightly shredded outline surrounded by fine-grained micaceous material produced by movement on the Bellinger Fault. Brooklana Beds. S32644, magnification x 2½.

H. Cordierite-bearing hornfels near the contact of the Kellys Creek Leucoadamellite. S32560, magnification x 20, plane polarised light.

A



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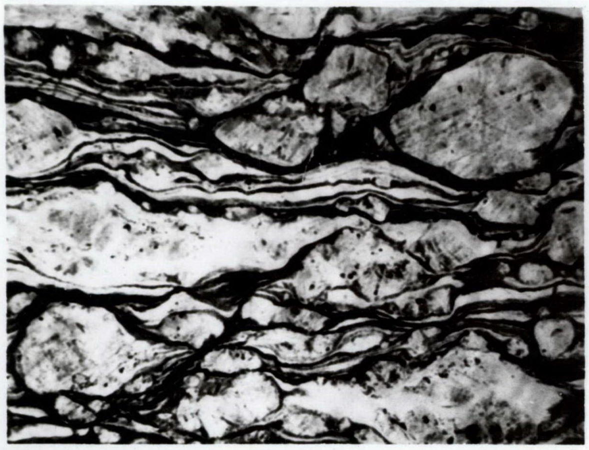

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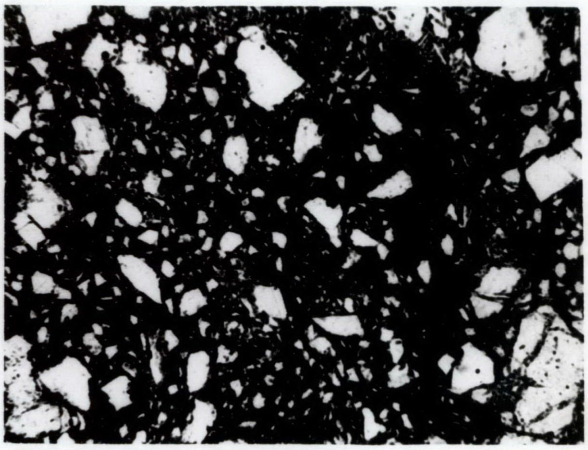

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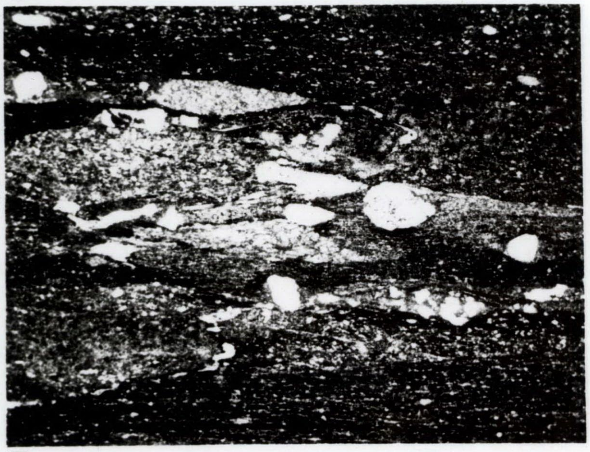

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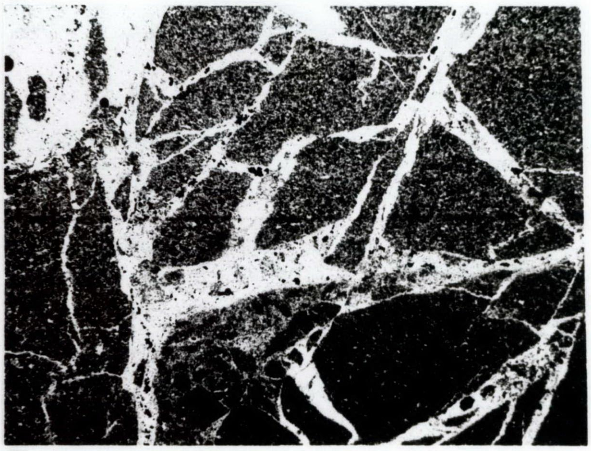

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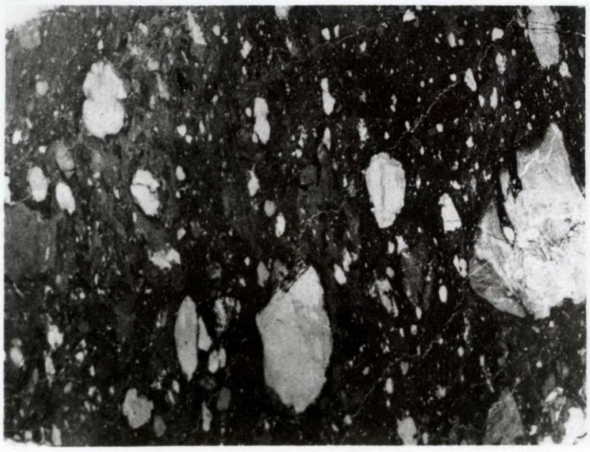
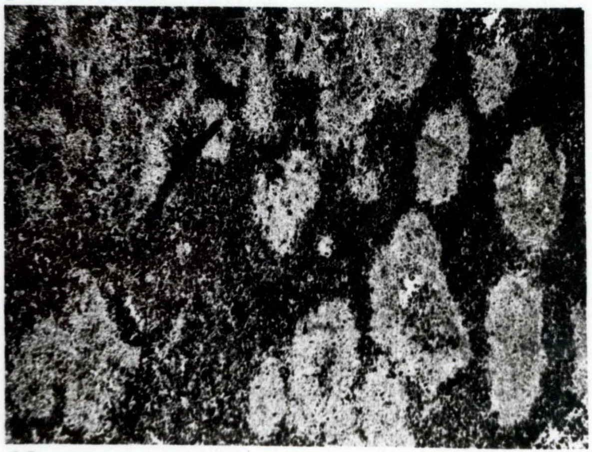


G



H

These assemblages are very similar to those of Zones III and IV from the Coffs Harbour Block.

Biotite-Cordierite Zone

Neilson (1965) mapped a zone of biotite-cordierite-bearing rocks occurring 3 km south of the contact with the Tobermory Adamellite, and is separated from contact hornfels of similar grade by a belt of lower-grade biotite-bearing rocks in which cordierite is absent. Neilson invoked an underlying apophysis of the Tobermory Adamellite, which has not yet been exposed by erosion, to produce this "outlier" and recorded the following assemblage: quartz-plagioclase-K-feldspar-cordierite-biotite-muscovite ± opaques. The work of Neilson supports the conclusion drawn here, that the biotite-bearing rocks of Zones III and IV have been produced by a static thermal event which was probably an underlying intrusion with a high heat flow.

Binns and Richards (1965) report a K/Ar age of 250 m.y. for a psammitic schist from the Rampsbeck Schists and it is considered that the low-grade metamorphism further to the west also occurred at this time. The Girrakool Beds are Permian in age (p.10) and hence the metamorphism must have occurred after the deposition of these sediments.

2. Metamorphism in the Wongwibinda Complex

The Wongwibinda Complex (Binns 1966) consists of a NNW-trending belt of high-grade schists, migmatites and gneissose granitic intrusions covering an area of about 270 sq. km, and truncated on the east by the Wongwibinda Fault. The Glen Bluff Fault marks the western extent of the completely recrystallised Rampsbeck Schists. To the west the Zone of Transitional Schists merges into the Girrakool Beds.

Zone of Transitional Schists

This narrow zone 2 km wide consists of partly recrystallised greywackes and pelitic rocks. Binns (1966) described aligned flakes of biotite and muscovite, with scattered scales of graphite. Detrital grains of quartz and feldspar are retained but the edges of the grains have been recrystallised and merge into the schistose matrix. Binns (*op. cit.*) placed this zone in the amphibolite facies because of recrystallisation of detrital plagioclase to clear calcic plagioclase rather than albite and epidote.

Rampsbeck Schists

These were divided by Binns (*op. cit.*) into a low-grade zone adjacent to the Transitional Schists and a high-grade zone near the Zone of Migmatites, on textural and mineralogical evidence. However, he mapped no isograds. The low-grade zone is entirely reconstituted and an assemblage of biotite-muscovite-plagioclase-quartz is characteristic.

The high-grade Rampsbeck Schists are characterised by an increased grain-size and the occurrence of orthoclase, almandine, cordierite and sillimanite. Herbert (1963) reported sillimanite and almandine from schists near Fishington. Binns listed the following assemblages as the principal ones:

Pelitic Schists

1. biotite-(sillimanite)-plagioclase-(orthoclase)-quartz
2. biotite-cordierite-sillimanite-plagioclase-orthoclase-quartz
3. biotite-(cordierite)-(sillimanite)-almandine-plagioclase-(orthoclase)-quartz

Semipelitic and Psammitic schists

4. biotite-plagioclase-(orthoclase)-quartz
5. biotite-almandine-plagioclase-orthoclase-quartz

The absence of basic or aluminous lithologies hindered the accurate determination of grades in the Wongwibinda Complex but Binns considered that conditions equivalent at least to the higher-grade subdivisions of the amphibolite facies (Turner 1968) occurred. These conditions are indicated by the occurrence of cordierite, almandine, sillimanite and orthoclase in the metasedimentary schists and diopside, grossular and clinozoisite in calcareous nodules. The highest grade in the Wongwibinda Complex probably did not exceed the amphibolite facies.

Fisher (1968) described the petrography of three thin (up to 9 m wide) amphibolite masses and associated quartzites which had been previously unrecorded. They consist of a series of discontinuous lenses occurring over a strike length of about 3 km. The three amphibolites all occur within the high-grade zone of Binns (*op. cit.*) and have differing mineralogies between each other. Two also have internal variations in mineralogy, grainsize and degree of foliation. Intimately associated with the amphibolites are banded quartzites exhibiting a varied mineralogy. Fisher (*op. cit.*) supports the conclusions of Binns (*op. cit.*) that the Rampsbeck Schists suffered conditions of highest-grade amphibolite facies and that granulite facies conditions were not reached. A low pressure facies series of the Abukuma type is indicated.

Binns and Richards (1965) reported a radiometric (K/Ar) age of 250 m.y. for a garnetiferous psammitic schist from the Rampsbeck Schists.

A group of rocks occurring as an elongated north-south trending inlier, within the Abroi Granodiorite approximately 1 km west of "Maryburn" Station, has been described in detail by Ransley (1964). A typical mineral assemblage is: biotite-quartz-muscovite-K-feldspar-oligoclase-(garnet)-(tourmaline), and Ransley assigned the rocks to the amphibolite facies. Ransley (*op. cit.*) considered that the Maryburn inlier had suffered retrogressive metamorphism with pale-green muscovite and quartz pseudomorphs replacing cordierite and andalusite.

Marriott (1973) described garnet and tourmaline from xenoliths of the Abroi Granodiorite and Rockvale Adamellite-Granodiorite and suggested that the Maryburn inlier is possibly a macroxenolith representing original parent material of the two intrusions. However the present author concurs with Ransley (1964) in that the Maryburn inlier represents a small portion of the Rampsbeck Schists enclosed in granitic rocks.

Zone of Migmatites

This narrow zone, approximately 1-2 km wide, occurs between the Abroi Granodiorite and the Rampsbeck Schists and has been described in detail by Binns (1966). The metasedimentary rocks in the zone resemble the high-grade Rampsbeck Schists in mineralogy but are slightly coarser than the true schists. These rocks are intruded by abundant veins and Binns recognised three different types. The oldest are thin felsic bands, mineralogically very similar to the enclosing schists. The second type is an intrusive leucoadamellite and the third vein-type is a biotite-rich adamellite. Certain metasedimentary schists and all vein types exhibit post-crystallisation microscopic deformation and this supports field evidence of repeated deformation during emplacement of the three types of veins.

E. DYNAMIC METAMORPHISM

In the Rockvale - Coffs Harbour region several faults have associated dynamic metamorphic products. These faults include those of the Wongwibinda Complex (Glen Bluff Fault, Fishington Fault and Wongwibinda Fault), the Demon Fault and the Bellinger Fault. Nomenclature for rocks produced by dynamic metamorphism is very confused with several schemes being proposed (Hsu 1955, Christie 1960, Reed 1964, Higgins 1971). The terms used here mainly follow Spry (1969 p.229).

Faults of the Wongwibinda Complex

Binns (1966) considered that movement along the Wongwibinda Fault after the emplacement of the Abroi Granodiorite led to the development of cataclastic effects in the granodiorite which are noticeable up to 500 m from the fault and a mylonite is produced in the actual fault plane (Plate 11C). Binns records shattered and sheared sediments for several hundred metres east of the Wongwibinda Fault in the Dyamberin Beds.

The Fishington Fault has produced sheared and contorted rocks in the Dyamberin Beds and Abroi Granodiorite. The Glen Bluff Fault has produced shredded biotites and cataclastic breakdown of felsic constituents in the Tobermory Adamellite. The intensity of deformation of Glen Bluff and Fishington Faults has not been as severe as that which produced the mylonites of the Wongwibinda Fault (Binns 1966).

Demon Fault

For a structure over 200 km in length and having a strike-separation (probably a true strike-slip) of at least 30 km, the Demon Fault is extraordinary in that the dynamic metamorphic effects along it are limited to a very narrow zone about 50 m wide.

Gradational effects were observed in the Chaelundi Complex east of the Demon Fault. The first noticeable effects are the development of strained extinction in quartz, growth of secondary calcite and fracturing of plagioclase with minor granulation at some feldspar boundaries (S32840). The next stage produces severe fracturing and a crush breccia (S32837). There is some veining, biotites have been chloritised, and severely kinked, feldspars have been fractured and quartz shows strain extinction.

This rock grades into a protocataclasite (S32838) where there has been chloritisation, kinking and shredding of biotite and fracturing and granulation of feldspar. A rock flour is beginning to develop. Adjacent to the fault cataclasites (S32835, Plate 11D, S32836) occur. There has been severe brecciation, fragmentation and granulation of quartz and feldspar to produce a very fine-grained green matrix. Minor relict quartz and feldspar are enclosed in the amorphous rock flour.

The first noticeable effect of the Demon Fault in the Sara Beds is kinking of cleavage in the slates. Closer to the fault there is a pronounced swing in strike from about 144° to 110°. Adjacent to the fault there is severe granulation of the sediment (S32720) to produce a cataclasite-like rock, which still retains some original texture.

North of the field area in the Gibraltar Range region the Demon Fault truncates the Dandahara Creek Granite producing cataclasites with a green matrix, adjacent to the fault. 100 m east of the fault kinked biotite and strained quartz are the only noticeable effects. The sediments west of the fault have been ground up to a cataclasite-like rock similar to S32720 in the Sara Beds.

Bellinger Fault

The Bellinger Fault separates the Coffs Harbour Block from the Nambucca Slate Belt. Outcrop is poor and the only deformational effects observed were at the headlands of Bonville and Boambee. Deformational effects of this fault are not as severe as those produced by the Wongwibinda or Demon Faults. Siliceous argillite, slate and greywacke of the Brooklana Beds have been altered to produce phyllonitic-type rocks as defined by Turner (*in* Williams, Turner and Gilbert 1954).

Bedding and slaty cleavage strike approximately east-west at a very small angle to the strike of the fault. These planes appear to have been transposed by differential movement on closely-spaced slip surfaces (e.g. S32653, Plate 11E, S32652, S32651) which are nearly parallel to the original bedding and cleavage. There has been attenuation and brecciation of the sediment along the slip surfaces and rocks produced include highly cleaved slates with mica crystallising along slip surfaces, coarse-grained phyllonitic rocks and breccias (Plate 11F). Bedding has been frequently obliterated and it is difficult to distinguish between original slaty cleavage and the subsequent slip surfaces.

Numerous quartz veins invade the deformed rocks and are often ptygmatically folded. Some veins are truncated, and broken pods of quartz occur, indicating later movement after emplacement. The slip surfaces vary from 0.2 mm in muds to 5 mm or more in coarser-grained or siliceous rocks. Mesoscopic and microscopic kink bands are common (S32650).

The phyllonites (S32653, S32652) consist of alternating bands of pelitic and quartzose material and are often sheared and irregular in outline. Randomly-oriented biotite possibly developed, after the differential movement, in association with M2. Flaser structures (lenticular aggregates of fine-grained recrystallised quartz and feldspar) are commonly developed in originally fine grained rocks (S32649). Augen structures are relicts of original material with a shredded outline around which closely spaced foliated layers swirl (S32644, Plate 11G) and appear to be aligned in the matrix. Spry (1969) considers that these augens could possibly be mechanically rotated into parallelism by movements of the matrix.

The zone of deformation associated with the Bellinger Fault appears to be wide, because phyllonitic rocks and augen structures occur at Boambee Headland, approximately 3 km north of the inferred position of the fault.

F. CONTACT METAMORPHISM

Contact metamorphic aureoles occur around most of the granitic intrusions in the Rockvale - Coffs Harbour region. The grade of metamorphism varies from albite-epidote-hornfels facies to hornblende-hornfels facies of Turner (1968). Pyroxene-hornfels facies have not been recognised in the field area but have been described by Binns (1965) 25 km to the west in basic rocks intruded by the Duval Adamellite.

Albite-epidote-hornfels facies is difficult to distinguish in sediments from the field area because of the low-grade regional metamorphism that the rocks have undergone. This is particularly so in the southern part of the Coffs Harbour Block where thermally-produced biotite has developed on a regional scale. Albite-epidote-hornfels facies has been previously described in incompletely reconstituted sediments near the margins of the Emerald Beach Leucoadamellite by Korsch (1971).

Thermal zones around the plutons (Map 2) define the limits of reconstitution of sediments into hornfels and do not represent the lower boundary of the hornblende-hornfels facies.

Coffs Harbour Block

Typical mineral assemblages observed in completely reconstituted hornfels include:

1. quartz-albite-biotite-muscovite-(opaques) (e.g. S32658, S32660)
2. quartz-albite-muscovite-biotite-cordierite-(opaques) (e.g. S32659, S32691)
3. quartz-albite-biotite-muscovite-garnet (S32690)

and are typical of hornblende-hornfels facies for pelitic and semi-pelitic rocks. Specimen S32690 occurs between Sheep Station Creek Complex and Dundurrabin Granodiorite and possibly has been affected by thermal episodes from both plutons.

Cordierite-bearing hornfels (Plate 11H) were observed near the contact of the Kellys Creek Leucoadamellite, the Dundurrabin Granodiorite and the small intrusion north of Dorrigo. Cronk (1973) records a cordierite zone of approximately 500 m to the south of Sheep Station Creek Complex.

Dyamberin Block

Collerson (1967) described an aureole 400 m wide around the northern margin of the Round Mountain Leucoadamellite in the Dyamberin Block. Collerson records assemblages of quartz-biotite-plagioclase-muscovite with cordierite, orthoclase and microcline in the innermost zone of the aureole. These minerals are typical of the hornblende-hornfels facies for pelitic and psammitic rocks.

Rockvale Block

Zones of cordierite-biotite hornfels occur around most intrusions of the Rockvale Block within the field area and vary in width from 50 m on the south side of the Rockvale Adamellite-Granodiorite (Marriott 1973) to 800 m around the Abroi Granodiorite (Collerson 1967). Garnets were recorded close to the contact of the Abroi Granodiorite by Collerson (*op. cit.*). Ransley (1964) and Marriott (1973) both record hornblende developed in hornfels south of the Rockvale Granodiorite.

The thermal aureoles developed in the Rockvale - Coffs Harbour region are very similar to those described elsewhere in New England by Spry (1953, 1955), Vernon (1961) and in several unpublished theses (e.g. Ransley 1964, Collerson 1967, Neilson 1965, Flood 1964).