A RE-EXAMINATION OF THE MONZONITIC COMPLEX AT MOUNT DROMEDARY, N.S.W.

A thesis submitted for the degree of DOCTOR OF PHILOSOPHY in the Australian National University

by

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STATEMENT

The studies incorporated in this thesis were made during the period between January 1961 and January 1964. A total of approximately four months was spent in mapping and collecting specimens from the Mt Dromedary district. The plane table mapping was done by the author with the assistance of Mr I. Faulks and Mr J.K. Boesen. All figures and maps in the thesis were drafted by the author. Dr G.A. Joplin made the modal analyses of the garnet-pyroxene rocks, but the remainder of the microscopical work was done wholly by the writer. Forty-two of the rock and mineral analyses were made by the author; Mr I.D. Martin carried out the spectrographic determinations of some trace elements in some rocks and minerals. The remainder of the analyses were made by Avery and Anderson of Melbourne and Mr A.J. Easton. Dr J.F. Lovering carried out the work on the electron micro-analyzer. The collation and interpretation of the data provided by those named above was the responsibility of the author, and all other work in the thesis was carried out by him.

(Robert Samuel Boesen)
TABLE OF CONTENTS

ACKNOWLEDGEMENTS 1
INTRODUCTION 2
ROCK NOMENCLATURE 8

PART I - THE STRUCTURE OF THE MONZONITIC COMPLEX AT MOUNT DROMEDARY 12

THE STRUCTURAL DATA 14
(1) Previous Investigations 14
(2) Mapping and Collecting of Specimens 14
(3) General Field Relations 15
(4) Contact Relations 19
(5) Internal Structures 24

INTERPRETATION OF THE DATA 34
(1) Summary of the Structural Data 34
(2) Magma Movement 35
(3) Method of Emplacement 37
(4) The Crustal Level of the Intrusion 39
(5) Proposed Structure 42
(6) The Laccolith Hypothesis 45

PART II - PETROGRAPHY 47
(1) Major Intrusions 48
(2) Minor Intrusions 54
(3) Pyroxenitic Masses 61
(4) Nepheline-bearing Intrusions 82
(5) Garnet-pyroxene Rocks and Associated Pyroxenites 93
(6) Andesitic Lava and Tuff 108

PART III - MINERALOGY 112
(1) Determinative Techniques 113
(2) The Clinopyroxenes 119
(3) The Biotites 134
PART IV - PETROCHEMISTRY AND PETROGENESIS

(1) Major Element Variation

(a) The pyroxenite-monzonite-banatite group. 154

(b) The undersaturated, alkalic group. 156

(c) Mafic versus felsic index. 158

(d) Proportions of iron, magnesia and alkalies. 159

(e) Miscellaneous rocks. 160

(2) Petrogenesis

(a) The monzonite-banatite association. 161

(b) The pyroxenites and garnet-pyroxene rocks. 163

(c) The undersaturated alkalic rocks. 166

(d) The petrogenetic role of the clinopyroxenes. 170

(e) Other minerals. 172

(f) The influence of the epi-tectonic locale. 175

BIBLIOGRAPHY 178
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Mount Dromedary is located on the south coast of New South Wales, about 180 miles south of Sydney and 85 miles south-east of Canberra. It lies very near the coast and is a prominent local landmark, rising out of the surrounding coastal lowlands to 2,613 feet above sea level.

The only detailed account of the geology of the monzonitic complex at Mt Dromedary published to date is that of Brown (1930), who greatly modified an earlier sketch-map of Anderson (1892) and gave petrographic descriptions of the igneous rocks, together with eleven analyses of the more important types. She also described a number of rare feldspathoidal rocks including covite, ijolite, shonkinite and jacupirangite and some garnet-bearing pyroxenites, and she concluded that the rocks of the complex were derived from a monzonitic magma chiefly through gravitative differentiation in situ in a laccolithic intrusion. She suggested that the feldspathoidal types might have formed by 'secondary differentiation' in a body of rest magma segregated by some physical disturbance of the complex. The origin of the garnet-bearing pyroxenites was felt to be even more obscure, but possibly related to limestone syntaxis.

Subsequent publications have been limited to a few minor observations: Wilshire and Binns (1961) give an analysis of a pyroxene from the Mt Dromedary district, Evernden and Richards (1962) have determined the absolute age of a biotite from the main body of monzonite, and Robertson (1963) has made some observations on the paleomagnetism of several of the rock-types.
Several of the geologists who have visited the complex felt that it has, in part at least, a stock-like form. Joplin (1962, p.67) stated that it has features characteristic of ring-dykes.

The monzonitic complex at Mt Dromedary appears to belong to a group of rocks which are characterized chemically by fairly high alkalies, with $K_2O > Na_2O$, and rather high lime and alumina. These rocks are emplaced in a late- or post-orogenic environment, and though they occur in only small bulk, they are very widely distributed ('The problem of the potash-rich basaltic rocks': Joplin, in press). However, relatively little is known about the parent magma or its mineralogical trends during differentiation, and one of the main aims of this study is to provide a detailed account of some of the variations observed in a suite derived from that magma.

The thesis is concerned with a re-examination of the complex, with special attention to its structure, variations in the principal mafic mineral series, and the petrology of the ultramafic, alkaline and garnet-bearing rocks. The structure and mechanism of emplacement of the rocks are given in Part I. The descriptive petrography is treated in Part II, and the mineralogy in Part III. Part IV deals with the petrochemistry of the complex and a discussion of the petrogenesis of the various rocks, with special reference to the garnet-bearing and feldspathoidal types.

The structural data described below strongly favors a stock-like form for the great bulk of the complex. This is concluded from the steep,
sharp contacts between the major rock intrusions, and also from the steeply oriented primary igneous flow structures which occur in most of the members of the complex. There are strong indications that the intrusions were injected to a very high crustal level and that the largest intrusive mass, the banatite, was emplaced consequent to the development of a large ring fracture within, and concentric with, the earlier intrusion of monzonite. The pyroxenites are probably slightly older than the other rocks and although their contacts are rarely exposed, it is suggested that they represent large blocks brought up from depth by the younger intrusions. Portions of these blocks of pyroxenite are delicately laminated in a manner which suggests that they were formed originally by crystal accumulation at depth.

The petrography of the complex is treated in considerable detail because of the heterogeneity encountered within many of the rock bodies. Banatite and monzonite, both of which contain a small amount of quartz, occupy about 85% of the total area of outcrop, i.e. 21 square miles (54 sq. km) out of 25 square miles (64 sq. km) and the nepheline-bearing rocks, 'ijolite', shonkinite and nepheline monzonite (with some covite), in all comprise only about 6% (1.5 sq. miles, or 3.9 sq. km). Pyroxenite and the garnet-pyroxene rocks make up about 4% (1 sq. mile, or 2.6 sq. km) and a thin sheet of andesitic lava and tuff covers the remaining 1.5 square miles (3.6 sq. km).

The banatite shows a systematic variation in composition from a quartz-poor, porphyritic margin to a more felsic, even grained core.
The monzonite shows small local variations in the proportion of its minerals, resulting from digestion of the abundant xenoliths found in it.

The nepheline-bearing and the garnet-pyroxene rocks provide striking examples of hybridism between an alkalic magma and ultramafic rock, notably pyroxenite.

A detailed study of one of the exposures of pyroxenite provides evidence of its origin at some deeper level and of the effects of hybridization between monzonitic fluid and the ultramafic rock.

Study of the numerous and varied dykes has shown that the majority are derived almost directly from the two major rock types constituting the complex, banatite and monzonite. The abundant lamprophyre dykes (camptonite) are very nearly isochemical with the monzonite and their distinctive appearance is believed to be dominantly controlled by the differing physical conditions attending their crystallization. There are some basaltic dykes of a somewhat tholeiitic nature, and some which are very highly alkalic. There are also a very few garnet-bearing dykes which have a foyaitic composition and these appear to be related to the alkalic intrusions of covite, etc.

The most important mafic mineral is a lime-rich clinopyroxene which shows a restricted variation in composition between the various members of the complex. The optical and chemical properties of the pyroxenes compare closely with those found in similar intrusive complexes elsewhere and with those considered typical of the alkali basalt magma. Several biotites have been examined in similar detail and unlike the pyroxenes, they show considerable substitution of Fe\(^{++}\) for Mg\(^{++}\). Three
amphiboles and three garnets have also been studied.

The felsic minerals have been examined optically and some microchemical tests have been made. The pyroxenites contain calcic bytownite in the apparently unaltered rock but this is replaced by andesine in the altered (often garnet-bearing) rock, where it may be joined by a number of minerals such as scapolite, cancrinite, thomsonite, nepheline, albite, etc.

In the more felsic rocks the plagioclase is joined by alkali feldspar, often microperthitic, and usually occurs as plates enclosing the plagioclase as in the monzonitic fabric. In the monzonitic and the feldspathoidal rocks there is clear evidence of the replacement of the plagioclase by alkali feldspar. Nepheline, more or less altered to giesecite and cancrinite, is the most abundant feldspathoid, with sodalite common in small amounts. It invariably has a form indicating late crystallization and it very commonly replaces plagioclase, and, to a much lesser extent, alkali feldspar.

Twenty-six new rock analyses are given, including two of the sedimentary country rock (Wagonga Beds). Some trace element abundances have been determined for the alkali-rich rocks and for the biotites. The trend of composition is characterized by a constant small excess of potash over soda. There is little tendency for potash to increase at a greater rate than soda in the more felsic members, although this feature is common in most differentiated series. There is no tendency to absolute iron enrichment, and very little relative enrichment in iron. The feldspathoidal rocks are consistently higher in potash than the
remainder, and although soda is initially lower it increases at a greater rate than potash, and becomes dominant in some dyke rocks. There is no absolute iron enrichment in this group, but slight relative iron enrichment occurs in the intermediate members.

The weak trend to relative iron enrichment (very similar to that shown by the calc-alkaline suite, after Daly, 1935) may be related to $pO_2$ as Osborn (1959) suggests, but the trends of pure differentiation have been partly obscured in these rocks by varying degrees of hybridism involving pyroxenite.

Most of the rocks at Mt Dromedary have either a little modal quartz or nepheline, and in most cases their norms contain little or no hypersthene. The parental magma is believed to be basaltic, lying near the plane of critical undersaturation. The alkalies amount to about 5%, with $K_2O$ nearly equal to $Na_2O$.

The origin of the pyroxenite is not clear, but it has been caught up in the more felsic rocks and hybrids have been produced.

The hypothesis of limestone syntaxis is rejected in regard to the feldspathoidal rocks and the possible role of incorporation of pyroxenite is discussed.
ROCK NOMENCLATURE

The nomenclature used herein differs from that of Brown (1930) and is set out below:

(1) **Banatite**: Johannsen (1932) writes that the name 'banatite' was given by von Cotta (1865) to a group of dioritic rocks in the Banat. Brøgger (1895) used the term to signify those rocks carrying about equal amounts of orthoclase and plagioclase feldspar, with a silica content of 63–66%. He saw the banatites as transitional between the quartz-syenites and the quartz-diorites in regard to feldspar proportions, and between monzonite and adamellite (the acid quartz-monzonites) in silica content. Brown (1930) applied the term in the sense of Brøgger, but noted (p.664) that "syenite might be permissable and useful as a field name".

The term banatite is retained because it conforms to Brøgger's definition and because of the dubious advantages of alternative rock-names, especially 'adamellite'. 'Adamellite' has a similar history to 'banatite' in that the former was used to describe tonalitic rocks of the Adamello batholith (Cathrein, 1890) and was subsequently used by Brøgger to denote monzonitic rocks more acid than the banatites.

In the current 'pigeon-hole' rock classifications erected on a mineralogical basis, such as that of Hatch, Wells and Wells (1952), a similar relation is implied, with quartz monzonites passing into adamellites with increase in modal quartz (>10%). Although the more
siliceous phases of the banatite contain up to 15% of modal quartz, it is likely that the intrusion as a whole contains a little less than 10%, in which case the rock could be called quartz-monzonite. There is, however, a good reason for not so doing, in that the fabric of the rock at Mt Dromedary is not monzonitic (cf. Brown, 1930, p.642).

In summary, the rock conforms adequately to Brøgger's definition; the rock is not strictly an adamellite since it is not siliceous enough; and although mineralogically it has the composition of a quartz monzonite, it is better not so named because it lacks monzonitic fabric.

(ii) Porphyritic banatite: Brown (1930) referred in the text to the porphyritic margin of the banatite as 'porphyritic monzonite', although this phase is shown on her map as 'porphyritic banatite'. Since the porphyritic rock is simply a local modification of the banatite there seems no reason to employ another rock term in describing it.

(iii) Monzonite: This term is retained since it quite adequately describes the rock surrounding the banatite at Mt Dromedary. Smaller intrusions which have monzonitic fabric and the appropriate proportion of the feldspars are described generally as 'monzonitic satellites'. They may contain small amounts of modal olivine or modal quartz.

(iv) Shonkinite: This rock crops out in the immediate vicinity of Tilba Tilba village. Brown (1930) refers to it as olivine-shonkinite (p.654) and has shown that there is a close chemical resemblance between it and the average of three shonkinites given by Pirsson (1900) and the original type of shonkinite from Square Butte (Pirsson, 1905). The rock
at Tilba Tilba appears to be remarkably similar to the Montana shonkinites.

(v) The rocks described by Brown (1930, p.657) as nepheline-bearing and related to monzonite, shonkinite and ijolite, are referred to below in general as nepheline monzonite, with specific mention where necessary of the particular phase. This is because the bulk of the intrusion (if it be a single body) has the composition of nepheline monzonite; the covite and shonkinite mentioned by Brown are very local modifications. The term shonkinite is reserved for the rock exposed at Tilba Tilba.

(vi) The rocks included in Brown's 'gabbro series' are referred to below simply as pyroxenite or pyroxenitic rocks. The use of the term jacupirangite is not justified, in view of the fact that nepheline has not been observed in these rocks.

(vii) Brown (1930, p.666) described the ijolite as being composed chiefly of pyroxene, nepheline and melanite, with a small amount of orthoclase, accessory plagioclase and a little olivine, sphene, apatite, sodalite and iron ore. Johannsen (1938) describes the type rocks (p.314) as usually medium grained, with hypidiomorphic-granular texture, and made up of abundant nepheline, zoned green clinopyroxene, biotite (as inclusions in the pyroxene), long prisms of yellow, green or reddish pleochroic sphene, apatite, locally wollastonite, iiwaarite (andraditic garnet with extensive substitution of Ti for Si; total TiO$_2$ = 16.44) and secondary cancrinite, zeolites and calcite. On the basis of these descriptions, Brown's use of the term ijolite seems justified. However, the specimens collected during the present study are very variable in their composition, but nepheline is
always subordinate to plagioclase and orthoclase, which together make up over half the rock, and there is a considerable amount of green hastingsitic amphibole. Because of the inhomogeneity of the rock and its poor outcrop, however, the mean composition of the rock is not known and so the term used by Brown is provisionally retained; to indicate the uncertainty regarding this rock its name is given in the form 'ijolite'.

(viii) An unusual association of garnet-bearing mafic and ultramafic rocks were collectively termed by Brown (1930, p.667) "(?)kedabekite-melanite-melteigite-jacupirangite"; she stated that they were made up of variable amounts of melanite, strongly pleochroic lime-rich clinopyroxene (yellow to green), and anorthite, minor nepheline, orthoclase and accessories. The present study has shown that the great bulk of the rocks in question consists of variable amounts of melanite (titaniferous andradite-grossular), pale green, very weakly pleochroic, diopsidic salite, up to 47% of feldspar with orthoclase predominating, while nepheline is absent. In view of the abundance of essential nepheline in melteigites (Br"gger, 1921; Johannsen, 1938) and the absence of feldspar in jacupirangite (Washington, 1901) the use of these rock names is unjustified. The present writer regards these rocks as hybrids and they are termed garnet-pyroxene rocks. These, and their associated mafic hybrids, have neither the mineralogy nor textures of kedabekite, which is described by Johannsen (1938, pp.240-242) as a eucritic rock with bytownite (46%), andradite (29%) and green clinopyroxene (25%, calculated as $Ca_{50.6}Mg_{24.6}Fe_{24.8}$); the term is hereby rejected.
The following account of the structure of the monzonitic complex is presented in two major sections, dealing with the structural data and their interpretation respectively.

PART I

THE STRUCTURE OF THE MONZONITIC COMPLEX

AT MOUNT DROMEDARY

The structural interpretation includes:

1. A short summary of the major structural elements;
2. An account of the movement of magma in the intrusive and the extrusive;
3. Discussion of the probable stress conditions during the emplacement of the monzonite, the dyke swarm, and the monzonite.
THE STRUCTURE OF THE MONZONITIC COMPLEX

AT MOUNT DROMEDARY

The following account of the structure of the monzonitic complex is presented in two major sections, dealing with the structural data and their interpretation respectively.

The former includes: (1) a brief summary of previous investigations; (2) an account of the mapping and of the collection of specimens for this study; (3) a description of the local geography, the spatial arrangement of the rocks, their age, and area of outcrop; (4) a detailed account of the contact relations of the important members of the complex; and (5) a detailed description of the mesoscopic (sensu Weiss and McIntyre, 1957, p.557) internal structures of the important members of the complex.

The structural interpretation includes:

(1) a short summary of the major structural elements;

(2) an account of the movement of magma in the banatite and the monzonite;

(3) discussion of the probable stress conditions during the emplacement of the monzonite, the dyke swarm, and the banatite;
(4) discussion of the probable crustal level reached by the complex;

(5) the proposed structure of the complex; and

(6) a comparison of the present interpretation with the proposals of Brown (1930).

THE STRUCTURAL DATA

(1) Previous Investigations

The first attempt to map the igneous rocks in the Mt Dromedary district was made by Anderson (1892) as part of a very general survey of the geology of the south coast of New South Wales.

Brown (1930) considerably modified Anderson's sketch-map in the course of a study of the structure and petrology of the monzonitic complex. She pointed out that several features of the distribution of the igneous rocks could be equally well explained by assuming either a laccolithic or stock-like form for the complex, but she considered that the field, petrographic and petrochemical evidence favored the laccolith hypothesis.

(2) Mapping and Collection of Specimens

A total of about four months was spent in the field during the two-year period 1961-1962. The complex was re-mapped in the course of the structural study. Field data were marked initially on enlarged aerial photographs (scale approximately 1:14,000) and were later transferred to a base map constructed from aerial photographs by the method of the radial line plot. (see Plates I-VI, end of thesis).
Figure 1. View across Tilba Tilba Lake to Mt Dromedary (wooded) and Little Dromedary to the left (east).

Figure 2. Looking due east from the summit of Mt Dromedary. Little Dromedary partly obscured by dead tree. Poole's Point immediately beyond source of smoke.
A narrow strip of well-exposed rock along the coast to the east of Mt Dromedary was mapped accurately, on a scale of 1:240, using a plane table and alidade, to determine the sequence of intrusion of the dykes. The mesoscopic structures observed in the pyroxenitic rocks at Poole's Point were also mapped using a plane table, on a scale of 1:120.

A number of samples were taken from each member of the complex so that a detailed picture could be gained of the compositional variations within each mass. Oriented specimens were taken at intervals along two radial traverses of the banatite intrusion, and also from several of the other bodies.

(3) General Field Relations

(a) Geography.

Mt Dromedary is situated on the south coast of New South Wales, about 170 miles south of Sydney and 85 miles south-east of Canberra. The members of the complex crop out on Mt Dromedary and on the undulating lowlands between the base of the mountain and the coast about 2½ miles to the east. The elevation of the country surrounding the mountain increases steadily in a westward direction from the coast to the highlands of the Great Dividing Range, about 40 miles distant. The country to the east and south of the mountain is moderately dissected and has an average height of about 150 feet above sea level. The country immediately west and north of Mt Dromedary is more deeply dissected and the average height is about 500 feet above sea level. The mountain itself has a rather low profile and stands about 2,300 feet above its surroundings. Its slopes are sharply dissected and the pattern of drainage is radial.
Figure 3. Vegetation on Mt Dromedary.

Figure 4. Little Dromedary, seen from near Tilba Tilba. Wooded portion roughly corresponds to the porphyritic banatite.
The district is covered with dense eucalyptus forest except for the small area between the eastern foot of Mt Dromedary and the sea. This land has been cleared and now supports lush dairy pastures. The vegetation on the mountain is very dense, often impenetrable, and seriously hinders field mapping. The rainfall is high but patchy in distribution, so that different aspects of the mountain support different ecological communities.

Access to the exposures of igneous rock between the foot of Mt Dromedary and the sea is fairly good but elsewhere it is very poor. Two minor roads run roughly north-south, one on either side of the mountain, and a narrow trail runs from Tilba Tilba village north-west to near the summit and then down to the north-east. Narrow fire and timber trails have been cut through the forest around the base of Mt Dromedary but these are of very limited use since they are commonly overgrown or blocked by fallen timber. Further, the canopy is nearly continuous except where the land has been cleared for cultivation, and it was often very difficult or impossible to determine one's exact position on the map.

(b) Geographical distribution of the rocks.

Mt Dromedary is the eroded remnant of a composite igneous intrusion. The older of the two members is composed of monzonite and has a nearly circular outcrop about 4½ miles in diameter. The younger member is a banatite which has a narrow, strongly porphyritic, marginal zone; the outcrop of this body is also nearly circular, with a diameter of 3 miles. The banatite occurs within, and concentric with,
the older monzonite, cropping out from the summit (in plan, approximately the common centre of the intrusions) to near the foot of the mountain where the porphyritic marginal phase intrudes the monzonite of the lower slopes.

Several small, sub-circular intrusions crop out in the country rock within about 2 miles of the southern half of the outer contact of the monzonite at Mt Dromedary. These satellitic bodies have monzonitic compositions, except for several small outcrops of pyroxenitic rock at Poole's Point and immediately to the north of Little Dromedary.

Little Dromedary lies mid-way between Mt Dromedary and the sea and rises sharply to an elevation of about 700 feet. It is composed largely of porphyritic banatite similar to that of the marginal phase at Mt Dromedary. There is a great diversity of rock types in the vicinity of Little Dromedary, but unfortunately this area has been cultivated and the contact relations of the bodies are commonly obscure.

The country rock (Wagonga Beds) consists chiefly of tightly folded beds of buff colored shales and phyllites, and minor dark grey chert. There are a few beds of sedimentary breccia, and thin, conformable intercalations of fine grained igneous rock. Some are undoubtedly tuffaceous but it is not clear in many cases whether they are sills, bedded dykes or flows. Minor sulphide mineralization is associated with some of the intercalations.

(c) Ages of the rocks and their area of outcrop.

The Wagonga Beds contain some radiolaria, but are otherwise
unfossiliferous. They are overlain unconformably by quartzites and slates to the east of Quaama, about ten miles to the south of Mt Dromedary. The slates are believed to be of Upper Ordovician age on the evidence of a graptolite fauna *(Diplograptus, Climacograptus)*, and consequently the Wagonga Beds are tentatively regarded as having a Cambrian age (Brown, 1933, p.335; David, 1950, p.124).

Until quite recently the monzonitic complex at Mt Dromedary was believed, on petrological grounds, to be of Permian age (Brown, 1930, p.692). Evernden and Richards (1962) have estimated the age of the complex to be $94 \times 10^6$ years. The determination was carried out on biotite from the monzonite at Mt Dromedary, using the potassium-argon method. Determinations have since been made on (a) hornblende from the nepheline monzonite near Little Dromedary, and (b) biotite from the laminated pyroxenite at Poole's Point; this data, as yet unpublished, was kindly supplied by Dr I. McDougall.

(a) Specimen number: GA 390 (Nepheline monzonite). Little Dromedary.

Mineral: Hornblende. $K = 1.99\% \ (2.18\% K_2O)$.

\[
\frac{Ar_{40}}{K_{40}} = 5.330 \times 10^{-3}. \quad \text{Air correction} = 12\%.
\]

Age = $89 \pm 3 \times 10^6$ years.

(b) Specimen number: GA 389 (Laminated pyroxenite). Poole's Point.

Mineral: Biotite. $K = 7.28\% \ (7.95\% K_2O)$.

\[
\frac{Ar_{40}}{K_{40}} = 5.481 \times 10^{-3}. \quad \text{Air correction} = 16.9\%.
\]

Age = $92 \pm 3 \times 10^6$ years.

These figures are in good agreement with that of Evernden and Richards, and the complex therefore appears to be of Middle Cretaceous age.
The total present outcrop of the rocks of the complex is about 25 square miles (64 square km), and the percentage outcrop of each rock-type is shown in Table I.

### Table I.

<table>
<thead>
<tr>
<th>Rock-type</th>
<th>Outcrop %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Banatite</td>
<td>40.8</td>
</tr>
<tr>
<td>Porphyritic banatite</td>
<td>8.8</td>
</tr>
<tr>
<td>Monzonite</td>
<td>25.4</td>
</tr>
<tr>
<td>Monzonitic satellites</td>
<td>4.8</td>
</tr>
<tr>
<td>Porphyritic banatite satellites</td>
<td>1.7</td>
</tr>
<tr>
<td>Leucocratic micro-banatite</td>
<td>0.7</td>
</tr>
<tr>
<td>Altered pyroxenite</td>
<td>0.7</td>
</tr>
<tr>
<td>Garnet-pyroxene rock</td>
<td>0.4</td>
</tr>
<tr>
<td>Andesitic lava and tuff</td>
<td>7.6</td>
</tr>
<tr>
<td>Nepheline monzonite</td>
<td>5.9</td>
</tr>
<tr>
<td>'Ijolite'</td>
<td>0.3</td>
</tr>
<tr>
<td>Shonkinite</td>
<td>0.2</td>
</tr>
<tr>
<td>Pyroxenite</td>
<td>2.7</td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td><strong>100.0</strong></td>
</tr>
</tbody>
</table>

Inspection of this Table shows that the intrusive monzonitic rocks are strongly dominant, and that the nepheline-bearing rocks, the ultramafics and the volcanics occur in very minor quantities.

(4) **Contact Relations**

The observations made on the contacts are presented in five parts:

(a) The monzonite-country rock contact;

(b) The banatite-monzonite contact;
(c) The contacts of the minor rock-types;

(d) The dykes; and

(e) The formlines.

(a) The monzonite–country rock contact.

The present writer agrees with Brown’s observations (1930, p.686) that a study of the strike directions of the country rock shows that the igneous mass of Mt Dromedary has forced its way into the old sediments. On the east and west of the mountain the beds bow around the intrusion, while to the north and south the igneous rock is strongly transgressive in its relations to the country rock. Locally, the strike of the beds may be sharply deflected near the outer contact of the monzonite.

Because of the dense undergrowth and the cultivated pastures the actual contact between the Wagonga Beds and the monzonite is only occasionally observed, although its position can always be inferred to within 20 yards. The contact does not appear to be influenced by topography and it runs quite smoothly over broken country. Where it is exposed, the contact is knife-sharp, transgressive, and has a steep dip.

The monzonite has a narrow marginal zone which is much finer-grained than the bulk of the rock.

(b) The porphyritic banatite–monzonite contact.

The contact between these rock types is located near the base of Mt Dromedary and can be recognized from a distance by the sudden steepening of the lower slopes, resulting from the greater rate of erosion of the monzonite as compared to the porphyritic banatite. The dense vegetation and broken nature of the terrain usually obscures the actual contact but
Figure 5. Contact between porphyritic banatite and monzonite, east of Central Tilba. Phenocrysts are about 2–3 cm long.

Figure 6. Contact between banatite and monzonite. Note that monzonite crops out at a considerably higher level than some of the banatite.
Figure 6. Contact between banatite and monzonite. Note that monzonite crops out at a considerably higher level than some of the banatite.
its position can be inferred to within a few yards. Where it is exposed, e.g. in Portion 220, Parish of Narooma, the contact is very sharp (Figure 5), and there is no reason to suppose that the two rock-types are gradational at any point.

As in the case of the monzonite and the country rock, topography appears to have very little control over the contact. At a number of places, monzonite crops out hundreds of feet above nearby porphyritic banatite (Figure 6).

(c) The contacts of the minor rock-types.

In most cases the shaley detritus formed by disintegration of the country rock obscures the contacts of the smaller intrusions, but the steep dip of the contact plane can be inferred from the relation of the lithological boundary to topography. The steeply dipping intrusive contacts of two small bodies are exposed in cuttings along the Princes Highway. The contact between the porphyritic banatite at Little Dromedary and the nepheline monzonite closely resembles the monzonite-porphyritic banatite contact at Mt Dromedary.

The relations between the rocks cropping out on the cleared land north and east of Little Dromedary are very complex and cultivation of this area has obscured many important contacts.

The volcanic rocks crop out on the higher ground and appear to rest upon the monzonitic intrusions. There do not appear to be any blocks of the volcanics caught up in the underlying rocks.

The alteration of the pyroxenitic rocks near Little Dromedary that are surrounded by nepheline monzonite suggests that they are large
xenolithic masses, partially transformed by metasomatic processes. The contacts of the pyroxenites cropping out at Poole's Point and on the shores of Tilba Tilba Lake are completely obscured, by either dune sand, Tertiary gravels, or the lake itself.

(d) The dykes.

Brown (1930, p. 689) noted that dyke-like bodies of porphyritic banatite intrude the surrounding coarse monzonite. The present writer has traced a few of these dykes back into the main mass of the porphyritic banatite. They are up to 8 feet wide, their strike is normal to the local contact, and they dip very steeply. The nature of the country prevented tracing the dykes from the porphyritic banatite, through the monzonite, into the country rock.

Several hundred dykes intrude the country rock surrounding the monzonitic complex, although only a few have been observed intruded into the other igneous bodies. The number, attitude and arrangement of the dykes are difficult to determine accurately because their outcrop is frequently obscured by vegetation and rubble. The best exposures are in road cuttings, particularly along the Princes Highway, and on narrow benches developed at intervals along the coast.

The average width of the dykes is about 3 feet, and a few reach 15 feet. The vast majority appear nearly conformable with the bedding of the country rock; consequently their regional strike is about 160° and their dip, to the east and west, is usually steeper than 60°. There are relatively few dykes radially disposed relative to the complex, but they are still abundant; on the coast east of Mt Dromedary about 60 were observed
Figure 7.
over a distance of a mile.

Their number appears to decrease quite rapidly to the north-east and south-west of the mountain. Most of the dykes make up a swarm about 5 miles across, trending about 160°, and embracing the igneous rocks which form Mt Dromedary. There may also be a subordinate, radial pattern of dyke intrusion.

(e) The formlines.

It has been stated above that the lithological boundaries do not appear to be laterally deflected as they pass across dissected topography. This implies, of course, that there is little or no relation between contour and the trace of a contact. Figure 7 shows the relation between the formlines and the lithological boundaries on Mt Dromedary. It is apparent that:

(i) The lithological boundaries on Mt Dromedary do not coincide with the local contour. The average elevation of a contact may differ from one side of the mountain to the other by more than 400 feet. In general, points on the north-western flank are considerably higher than corresponding points on the south-eastern flank of the mountain.

(ii) In areas of high local relief, common on the northern and western slopes, there is very little relation between the contours and the lithological boundaries.

(iii) The apparent vertical extent of the rock-types varies from place to place over their outcrop. For example, the apparent vertical extent of the monzonite on the west slopes is 400-500 feet, but it is only about 100 feet on the east slope of the mountain.
MT DROMEDARY: Primary igneous structures, dykes

Scale in miles.

Figure 8.
(5) Internal Structures

No orogeny is known to have affected the district since the probable time of emplacement of the complex, and the linear and platy structures observed in the igneous rocks may reasonably be described as primary (sensu Balk, 1937). The intrusive rocks of the complex, and especially those forming Mt Dromedary, were studied for primary igneous structures, particularly those attributed to flow. The observations are recorded in Figure 8.

Unless otherwise specified, the term 'banatite' is used to indicate both the normal banatite and the porphyritic marginal phase.

The three rock-types whose internal structures are most important are the monzonite and banatite at Mt Dromedary, and the pyroxenitic rocks.

(a) Structures of the monzonite.

The primary igneous structures displayed by the monzonite include a combination of platy and linear flow. Both structural elements vary in intensity from place to place, but they are usually not very strongly developed. The strike of the platy flow structure is parallel or sub-parallel to the outer contact and the dip is nearly always steeper than 70°. However, the sense of the dip, toward or away from the contact, varies erratically from place to place. The flow lines lie within the plane of flow, and pitch steeply down it. The platy structure results partly from the sub-parallel alignment of biotite flakes. The other mafic minerals, hornblende and clinopyroxene, have a much more prismatic habit and contribute more to the lineation within the plane of the platy structure than to the platy structure itself.
Figure 9. Biotite-rich schlieren in boulder at Central Tilba quarry.
Figure 10. Same as (9), in face of quarry.
Figure 10. Same as (9), in face of quarry.
The monzonite contains a large number of platy xenoliths of country rock which are strongly aligned in the platy parallelism of the rock. They are thin compared to their lateral dimensions, which are about equal. Their average diameter is about 5 cm, and their thickness is about 0.5 cm. Rarely, as in the quarry at Central Tilba or on the western slopes of Mt Dromedary, the xenoliths are much larger and may have a diameter in outcrop of up to 1,000 feet.

Prominent biotite-rich schlieren are exposed in the quarry at Central Tilba, at the eastern foot of Mt Dromedary. They appear as dark, irregular layers and spindles, and are steeply oriented (Figures 9, 10). The dip of the platy flow structures observed in this area is more variable than elsewhere in the monzonite.

In the vicinity of the quarry at Central Tilba and on the northern slopes of the mountain, two intersecting lineations can be observed on a few sub-vertical rock faces whose strike is normal to the local monzonite contact. The steeper of the two lineations is the trace of the dominant local platy flow structure, while the other cuts across this at about 30° and dips outward toward the contact at about 60°. Good exposures are so rare that it is not clear whether the second lineation is the trace of a linear parallelism, or of another platy parallelism with substantially the same strike as the 'normal' platy structure. Similarly, the time relations between the two structures is not clear. H. Cloos (in Balk, 1937, p.23) has noted a discordance between flow layers and flow lines in a quarry at Strehlen, Germany, where the dip of the flow layers (schlieren) and the linear parallelism differed by about 20°. On the other hand, the
TABLE 2. Variation of specific gravity with depth in cored rock.

<table>
<thead>
<tr>
<th>Depth below collar (feet)</th>
<th>Specific Gravity</th>
</tr>
</thead>
<tbody>
<tr>
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<td>Monzonite</td>
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<tr>
<td>0</td>
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</tr>
<tr>
<td>25</td>
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<tr>
<td>575</td>
<td>3.24</td>
</tr>
<tr>
<td>600</td>
<td>3.22</td>
</tr>
</tbody>
</table>

Mean specific gravity

Temperature during determination

Immersion medium

Approximate weight of sample (grms)
Figure 11. Showing orientation of feldspar phenocrysts and relation to the wall of the intrusion. (See also Figure 12).
Figure 12. Phenocrysts in porphyritic banatite, Little Dromedary. Natural size.
weaker lineation may not be a primary structure; it may be related to the subsequent intrusion of the banatite.

Jointing in the monzonite occurs along two near-vertical planes, one radial and one tangential to the shape of the outcrop. A third or even a fourth sub-vertical joint plane may be present in some localities and these planes strike obliquely across the more regular directions mentioned above. It is very difficult to assess the importance of flat-lying (Q) joints, because well-exposed vertical sections are extremely rare. The joints are seldom occupied by aplitic rock or pegmatites, as the monzonite is rather poor in these phases.

A vertical diamond drill hole was sunk in the monzonite near Tilba Tilba village (Plate I) and core was obtained to a depth of 412 feet. The core has not yet been examined in detail but the rock does not appear to show any significant variation in composition with increasing depth. Table 2 shows the variation in density of the rock with increasing depth.

(b) Structures of the banatite.

The primary flow structures of the banatite are more regular and more striking than those of the monzonite. The porphyritic marginal phase shows a strong platy flow structure resulting from the sub-parallel orientation of the alkali feldspar brachypinacoids, (010). The very steep linear parallelism within the plane of the platy structure is marked by the sub-vertical lengthwise orientation of the phenocrysts, i.e. with the [a] crystallographic axes of the feldspars parallel to the flow lines (Figure 11). The prismatic and platy mafic minerals such as hornblende and biotite which are concentrated in the marginal porphyritic phase are oriented in the
flow structures, but their contribution is relatively inconspicuous, chiefly because of the size and abundance of the alkali feldspar phenocrysts.

The number of phenocrysts and the amount of the mafic minerals decreases rapidly as the porphyritic marginal phase grades into the more even grained core, and the structures marked by the phenocrysts become much less prominent. At the same time, the platy flow structures become slightly weaker and the flow lines slightly stronger, and this pattern persists to the centre of the banatite.

The inaccessible nature of much of Mt Dromedary has limited the number of structural observations, but the data available indicate that the platy flow structures in the core of the banatite trend north-south and may dip steeply either to the east or the west. The flow lines consistently pitch very steeply to the north. It is clear that there is little or no tendency toward formation of arch or dome structures in the banatite.

Xenoliths are very rarely observed in the banatite, in sharp contrast to their abundance in the monzonite.

The banatite is jointed along two near-vertical planes, one radial and one tangential with respect to the circular outcrop. This gives rise to nearly square vertical columns, which are best observed in the porphyritic banatite. There are two other weak joint planes, each of which strikes tangentially to the circular form, but one dips towards the outer contact at about 40°, while the other dips away from the contact at a similar angle. The joints are only occasionally occupied by narrow aplite dykes or pegmatites.
Figure 13. Pyroxenite, Poole's Point.

__50 feet__
(c) **Structures of the pyroxenitic rocks.**

There are several pyroxenitic bodies in the Mt Dromedary district; their outcrop varies in diameter from a few yards to more than half a mile. The pyroxenitic complex at Poole's Point was studied in great detail because it is made up of several different rock types, each of which shows characteristic structures, and because of its complex history. The outcrop is about 600 feet long and at most 120 feet wide. Rock thought to be in situ occupies about three-fifths of this extent; the rest is covered by sand and rubble. The outcrop, bounded on the east by the sea and elsewhere by sand, has been mapped by plane table; the limits of rock thought to be in situ are marked, together with the more prominent dykes and faults (Figure 13).

Certain rocks of the pyroxenitic complex appear to be igneous cumulates (sensu Wager et al., 1960) which have been partially recrystallized. These rocks are layered in a manner which resembles the 'inch-scale layering' in the Stillwater complex (Hess, 1960, pp. 51, 77, 133) except that the layers at Poole's Point are much narrower. Partly because of the thinness of the layers, the present author prefers to speak of 'lamination' and 'laminae' rather than of 'layering' and 'layers'. The term 'lamina' is used herein to mean the thinnest layer possessing a distinct mineralogical character; hence plagioclase-rich laminae, pyroxene-rich laminae.

From place to place the plagioclase-rich and pyroxene-rich laminae are of equal thickness and alternate very regularly over a total thickness of a few inches. This appears in outcrop as a series of parallel, closely
Figure 14. Concentration of plagioclase-rich laminae (a set). Note also the small transverse fault, xenolith, and angular discordance between laminae. Scale mark 1 foot (approx. 30 cm).
Figure 15. Pyroxenite, Poole's Point, showing traces of steeply dipping lamination over area marked in Figure 13.
Figure 16. Ribbed outcrop of laminated dense pyroxenite.
Lens cap is 4 cm in diameter.
spaced, white lines about 2 mm across, on a very dark background; such bands of concentration of plagioclase laminae are referred to below as 'sets'. A typical set is shown in Figure 14.

Samples from the sets have modes similar to those of gabbros. This term is not used, however, since the sets are local zones of plagioclase concentration of relatively small extent enclosed in a large bulk of ultramafic rock, and the use of a different lithological term under these circumstances would obscure the close relations between the plagioclase-rich and plagioclase-poor portions. The term 'feldspathic pyroxenite' is therefore used.

The following structures have been distinguished in the pyroxenitic complex:

(i) **Lamination.** The laminated rock is confined to the southernmost outcrop (Figure 13). Traces of the laminae were mapped using a plane table and the pattern is shown in Figure 15. The laminated pyroxenite has been divided into two types on field criteria.

The more extensive type contains prominent plagioclase-rich laminae and weathers to a fairly smooth surface, while the less extensive, feldspar-poor type weathers to give a ribbed outcrop (Figure 16). The plagioclase-rich laminae of the more extensive type are about 1.5 mm thick and are enclosed in plagioclase-poor pyroxenite of relatively massive appearance. The degree of development of lamination varies widely, resulting in gradations from virtually massive pyroxenite to strongly laminated rock of gabbroic composition.

The less extensive type of lamination is distinguished in the field by its ribbed appearance in outcrop and by its more mafic nature,
Figure 17. Portion of fold in laminated pyroxenite, Poole's Point.

Figure 18. Portion of folded, laminated pyroxenite. Note discordance between laminae. Scale mark is 1 foot (approx. 30 cm).
Figure 19. Discordance of laminae. One-third natural size.
especially in regard to the ore minerals. The alternation of laminae in this case appears to be much more regular and no structures such as sets were noted. The laminae are about 7 mm thick in comparison with 1.5 mm for the feldspar-rich members of the preceding type.

The strike of the lamination observed at Poole's Point varies from place to place, but the dip is always steep.

(ii) **Structures in laminated rock.** The traces of the lamination are shown in Figure 15, from which it is apparent that the laminated rock has suffered some degree of folding. The laminated mafic rock lies in a narrow, gently arcuate belt, slightly offset at a few points by a number of small faults. The folding in the laminated (Fig. 17) feldspathic pyroxenite, only a few feet away, is much more pronounced and has produced a complete sigmoidal fold with two small, asymmetrical folds (similar in form to drag folds) on its limbs.

The traces of laminae in the weakly laminated rock which encloses the prominent fold often do not conform to the shape of the fold. Also, the laminae which are folded pass quite abruptly along their length into apparently massive pyroxenite.

Angular discordances between single laminae, and between sets of laminae, are common (Figures 18, 19).

(iii) **The massive pyroxenite.** The majority of the pyroxenite exposed at Poole's Point appears to be massive; it is younger than, and encloses, the laminated rocks. There is ample evidence of disruption and invasion of the laminated rocks by the massive type (Figures 20, 21). In nearby rubble, irregular fragments of laminated
Figure 20. Laminated rock enclosed in darker, more massive pyroxenite. Lens cap is 4 cm in diameter.
Figure 21. Illustrating the sequence of events: 1. Formation of lamination. 2. Intrusion by darker, coarse rock (note penetration along plane of lamination). 3, 4. Small fault and dyke emplacement.
Figure 22. Faulting marked by the offset produced in a set. The fault surface is clearly visible. Lens cap is 6 cm in diameter.
Figure 23. Platy xenoliths aligned in the plane of lamination. Lens cap is 4 cm in diameter.
pyroxenite a foot or more across can be seen, completely enclosed in the massive rock.

(iv) **Faults.** Faulting is easily recognized by the offset it produces in the lamination. The faults are minor and result in maximum offsets of up to 2 feet. The strike of the faults varies but their dip is consistently steep. In many cases faulting occurred along a single surface and without brecciation (Figure 22), although there are a few zones of intense fracture, up to 2 feet wide, which may be the result of fault movements. Weathering of these zones produces a rock made up of abundant, small pyroxenite nodules set in a friable matrix. Sulphide mineralization occurs along some of the fracture zones.

(v) **Xenoliths.** The laminated feldspathic pyroxenite commonly contains fine grained, elliptical or tabular xenoliths, about an inch long, oriented parallel to the plane of lamination. They do not appear to owe their elongation to shear stress, since their terminations are quite commonly blunt and irregular. (Figure 23).

The largest xenolith observed in the laminated feldspathic rock is a rectangular block of fine grained basaltic material, measuring 18 inches by 30 inches, aligned lengthwise parallel to the local lamination (Figure 14).

(vi) **Dykes and veins.** These are common throughout the complex but are more abundant in the northern outcrops. The dykes are less than a foot wide, and tend to follow the three dominant joint directions, forming triangular networks in places (Figure 24). The wider dykes are fairly straight but the narrower ones are usually sinuous. The
Figure 24. Triangular network of dykes.

Figure 25. Composite dyke.
Figure 26. Feldspathic dyke or vein similar to sample analyzed (RB417). Lens cap is 6 cm in diameter.
dyke-rocks are dark and fine grained, and may closely resemble their host-rock. In the northernmost outcrop some of the dykes are noticeably lighter in color than the rock they intrude, and are irregular in form. Figure 25 shows one of the very few composite dykes, and Figure 26 a feldspathic type.

The veins rarely exceed an inch in breadth, and cannot be traced continuously over more than a few feet. In contrast to the dykes, they are more coarse grained than their host-rock and do not have sharp intrusive boundaries. The shape and orientation of the veins are not obviously related to the joint pattern of the pyroxenite. A rough distinction can be made between feldspar-free veins, which appear as stringers of coarse black amphibole grains embedded in the massive pyroxenite, and much thicker veins, rich in alkali feldspar. It is clear that both types of vein and the lighter colored dykes are intimately related. Figure 27 shows a narrow, irregular dyke of the lighter colored type cutting through coarse, massive pyroxenite. Issuing from it is a narrow feldspathic vein with the coarse, dark grains of amphibole associated with it being clearly visible. About 18 inches from its junction with the dyke, the vein narrows abruptly and the feldspar virtually disappears, and the course of the vein is marked by the dark uralitic amphibole.

(vii) Joints. There are three prominent joint directions in the pyroxenitic complex at Poole's Point (Figure 24). The dip of the joints is nearly vertical, and their effect is to break up the pyroxenitic rocks into small, vertical, often triangular, prisms. There is a suggestion of another, more weakly developed, sub-horizontal joint plane
Figure 27. Dyke-vein relations. Attenuation of dyke is natural and not the result of faulting.
Figure 28. Block of coarse pyroxenite from drill site. The marked difference in amount of light constituents is illusory. The block is rectangular, with an edge nearest the camera, so that the light and dark portions are sides at right-angles to each other. One-third natural size.
Figure 29. Same block as in Figure 28; feldspar-rich face. Natural size.

Figure 30. Same block as in Figure 28; pyroxene-rich face. Natural size.
but this is hard to assess since the surface of the exposure is itself sub-horizontal, and very near sea level.

   (viii) **The drill core.** Pyroxenite which closely resembles the massive type at Poole's Point crops out along the eastern shore of Tilba Tilba Lake. The rock is very dark and coarse grained, with well formed equant pyroxene crystals set in a very sparse matrix of labradorite. Hornblende is patchy in distribution and minor in amount, forming ragged but extensive grains enclosing the pyroxene crystals.

   A vertical diamond drill hole was sunk in the pyroxenite and a complete section of core was recovered from a depth of 25 feet to 602 feet. Although the rock at the surface appeared fairly fresh, no core was recovered for the first 25 feet. Apart from a few narrow dykes, pyroxenite was cored continuously until drilling was stopped at 602 feet. The pyroxenite does not appear to show any systematic variation in texture, grain size or mineralogy with depth. Sections of core from the top and from the bottom of the hole are virtually indistinguishable. Table 2 shows that there is no significant variation of rock density with increasing depth.

   Although the pyroxenite in situ at the drill site seems to be massive, some degree of mineral orientation can be observed in a few loose blocks. Laminar structures similar to those exposed at Poole's Point have not been observed. (Figures 28, 29, 30).

   The small xenolithic masses of pyroxenite near Little Dromedary show fine-scale lamination in a few places, but most of the rock appears to be rather massive.
(1) Summary of the Structural Data

The present writer regards the structural data described above as clear evidence that the majority of the members of the monzonitic complex at Mt Dromedary are stock-like bodies which have been intruded nearly vertically into the Wagonga Beds. The major features shown by the rocks cropping out on Mt Dromedary and by the pyroxenitic rocks found on the eastern shore of Tilba Tilba Lake and on the coast at Poole's Point may be summarized as follows.

(a) The beds around the base of Mt Dromedary have been disturbed, showing that "the igneous mass has forced its way into the old sediments" (Brown, 1930, p.686).

(b) The monzonite appears to be slightly chilled against its outer contact.

(c) Where the contact between the monzonite and the country rock is exposed it is strongly transgressive, knife-sharp, and dips very steeply. The contact usually is not exposed but it can be inferred quite accurately, and it runs smoothly over the dissected basal slopes of the mountain.

(d) The inner contact, between the monzonite and the porphyritic banatite, is also knife-sharp and dips very steeply.

(e) Locally, the correspondence of lithological boundaries and contour is poor, and on the mountain as a whole it is only approximate.

(f) The igneous rocks on Mt Dromedary show linear and platy parallelism of their constituents.
(g) Platy xenoliths, strongly resembling pieces of country rock, are very common in the monzonite, and a large inclusion of basic volcanic rock is exposed on the western slopes of Mt Dromedary. Abundant biotite-rich schlieren occur in the monzonite quarried at Central Tilba. There are, however, very few xenoliths in the banatite, and they appear as small, elliptical patches, distinguished from the host-rock only by their slightly higher biotite content.

(h) On Mt Dromedary, dykes of porphyritic banatite intrude the surrounding monzonite.

(i) The pyroxenite has a complicated geological history, but the bulk of the rock is virtually massive, and it extends to appreciable depths.

(2) **Magma Movement**

The stock-like nature of the monzonitic intrusion at Mt Dromedary is clearly indicated by its sharp, steep contact and the disturbed bedding in the country rock. The monzonite is slightly chilled against its outer contact and a medium grade contact metamorphic aureole has been developed in the beds around it. The steepness of the contact is inferred from its undeflected passage across dissected country; this is best examined on the northern and western slopes of Mt Dromedary. Rapid and repeated changes in elevation of the contact of about 300 feet are common as it cuts across steep ridges, yet there is no lateral displacement of the boundary. The primary igneous flow structures provide strong evidence of vertical, or near-vertical, movement of the monzonite magma.

The contact between the banatite and monzonite resembles the
monzonite-country rock contact in that they are both knife-sharp and steep, where they are exposed, and indicate their steepness otherwise in their apparent disregard for topography. The banatite-monzonite contact differs only in that it can be more accurately mapped, is more regular in plan, and is confined more to areas of high local relief than the monzonite-country rock contact. Hence the same reasoning may be applied in proposing that the banatite, like the monzonite, is a nearly vertical, stock-like intrusion.

The record of flowage in the banatite indicates steep upward movement. Steeply dipping, marginally developed, primary flow structures are common in small stock-like intrusions and a number of such occurrences are described briefly by Balk (1937, p.86 et seq.); he notes, however (p.88), that the interiors of these bodies do not usually show flow structures, whereas they are quite distinct at Mt Dromedary. The syenitic stock at Snowbank Lake (Grout and Balk, 1934) has similar structures and dimensions to the banatite. The intrusion in that case is steep, and the flow structures are conformable with, and more strongly developed near, the contacts. In its interior, steeply dipping flow lines predominate. Dome or arch structures are absent both from the syenitic stock at Snowbank Lake and from Mt Dromedary. As Balk points out (p.115), stock-like intrusions, especially those penetrating to high crustal levels, may not show domed roof structures, so the absence of these structures from the banatite need not imply that the present level of erosion of the complex is deep.

The available evidence points to the emplacement of both the
monzonite and the banatite by upward movement of their respective magmas.

(3) Method of Emplacement

(a) The monzonite. The outcrop of the monzonite is egg-shaped, with the longer axis east-west and the nose towards the east. The beds of the country rock are more strongly deflected around the eastern than around the western contact, and the monzonite appears to have gained much of its room by simply shouldering aside the meta-sediments, which yielded more easily to the east of the intrusion than to the west.

The intrusion cuts sharply across the beds of the country rock, particularly on the northern and southern contacts, and meta-sedimentary xenoliths are common. If the structures of the monzonite now exposed can be taken as representative of the original mass then it is unlikely that piecemeal stoping (sensu Daly, 1933, p.268) was a significant factor in the emplacement of the stock, because the xenoliths are aligned in the flow structures and appear to have been passively transported by the magma during its intrusion. Further, the country rock is less dense than the monzonite; their specific gravities are 2.67 and 2.87, respectively. In general, the xenoliths are only weakly reconstituted, and they have not developed a metamorphic assemblage which would have a greater density than the monzonite.

The monzonite appears to have been emplaced almost wholly by forceful injection.

(b) The dykes. The dykes in the swarm were intruded sub-parallel to the plane of the tension fractures which would tend to form in the
country rock as a result of the east-west elongation of the monzonite body during its intrusion. The dykes may have played an active part by locally lengthening the crust to allow the lateral extension of the monzonite, or they may have been emplaced as a result of tensional stress previously set up by the intrusion during its east-west expansion. In either case, the steeply dipping bedding cleavage of the meta-sediments is a marked plane of weakness and dykes would tend to be emplaced parallel to it. The dyke swarms associated with the intrusions in the Ben Nevis-Ben Cruachan area of the Scottish South-West Highlands also trend parallel to the schists and phyllite of the country rock (Bailey et al., 1916, p.90).

(c) The banatite. There is little doubt that the monzonite was forcefully injected into the Wagonga Beds and it has been shown above that certain of the structures of the monzonite, such as steep contacts and steep flow structures, also occur in the banatite. However, these similarities do not imply that both intrusions were emplaced by the same mechanism. The east-west elongation of the monzonite probably resulted from its forceful expansion in that direction during intrusion. There is no evidence of strong deformation of the monzonite subsequent to its emplacement and it follows therefore that the east-west elongation of the banatite could not be a result of forceful lateral expansion during intrusion.

The rarity of xenoliths of either sedimentary or monzonitic rock in the banatite, the small number of dykes of banatite intruded into the monzonite, the smooth contact surface between the two intrusions, and the scant evidence of cataclasis in the monzonite near the contact, indicate
(i) The banatite may not, unless at considerable depth, have come in contact with the Wagonga Beds.

(ii) The monzonite was very largely or wholly solidified at the time of emplacement of the banatite and the plane of contact between the two rock types has not been altered significantly by stoping.

(iii) The banatite was emplaced under conditions of greatly reduced pressure, relative to the emplacement of the monzonite.

(4) The Crustal Level of the Intrusion

An estimate of the probable crustal level reached by the complex can be made from a consideration of the stratigraphy at the time of its intrusion, that is in the Middle Cretaceous. Accounts of the regional stratigraphy are given by David (ed. Browne, 1950) and Brown (1931, 1933).

The basement rocks in the Mt Dromedary district are the tightly folded Wagonga Beds, and they are overlain unconformably by minor amounts of Upper Ordovician slates and quartzites. The Lower Silurian is not represented and there are no Upper Silurian rocks within thirty miles of the complex. Rocks of Lower and Middle Devonian age appear to be absent but a long, broken belt of Upper Devonian rocks passes to the west of the mountain and extends from south of Eden to Milton, a distance of about 130 miles. The Upper Devonian succession includes acid lavas, interbedded freshwater shales, sandstone and volcanic rocks, and an association of shallow marine conglomerates, quartzites, shales, etc. The degree of folding is fairly mild and the local succession is about 2,200 feet thick. From the close of the Devonian period the south coast of New South Wales was apparently emergent land with low relief.
Further to the north, on the Illawarra coast, there are Permian volcanics (Bumbo Latite) which are chemically very similar to the monzonitic rocks at Mt Dromedary. Because of this similarity, Brown (1930) suggested that both occurrences were derived from a common parent and were essentially contemporaneous.

The movements of the continent during the Mesozoic were essentially epeirogenic, and the Mesozoic closed with general marginal uplift and increased rates of erosion. At the time of intrusion of the complex the country appears to have been relatively stable, with low elevation and relief. The local stratigraphic section would show the basement of folded Wagonga Beds, overlain unconformably by a small thickness of Upper Ordovician rocks, which would in turn be overlain unconformably by a gently folded Upper Devonian sequence with a thickness greater than 2,200 feet (the present local thickness, after some erosion) but probably less than 4,000 feet (the probable thickness of the sequence measured in sections protected from erosion by Permian strata or by younger stages of the same sequence). A working figure of 3,000 feet is adopted in the following discussion.

Any previous cover of the complex has been eroded off, and the Upper Devonian beds are found as discontinuous outcrops some miles to the north, west and south of Mt Dromedary. The general attitude of the base of the Upper Devonian succession can be determined fairly simply by comparison of a lithological with a topographical map of the district. Since the degree of folding of the Upper Devonian rocks is slight, the level so obtained for the base should not be very seriously in error.
Using this method, it is found that the base of the Upper Devonian sequence dips very gently toward the south-east and intersects the present surface of Mt Dromedary at about the 900 feet contour. Since the summit of Mt Dromedary is at present just over 2,600 feet above sea level, it follows that the banatite must have projected at least 1,700 feet (2,600 - 900 feet) into the overlying Upper Devonian rock; that is, the banatite must have been intruded to within 1,300 feet (3,000 - 1,700 feet) of the surface of the crust. Since allowances have to be made for erosion of the Upper Devonian sequence prior to the intrusion of the complex, and for erosion of the igneous rock since its emplacement, then the intrusions must have come well within 1,000 feet of the surface.

The lavas and tuffs which crop out over about 2 square miles of the lowlands to the east of Mt Dromedary also provide some evidence that the complex was intruded to a high crustal level. The volcanic rocks are younger than the intrusive members of the complex, since they have been laid down on the eroded surface of some of the smaller intrusions, but they have been intruded by dykes clearly related on petrological grounds to the Lower Tertiary basalts. The lavas and tuffs were very probably laid down at some time late in the Upper Cretaceous.

If it is assumed for the moment that the banatite crystallized under about 1,000 feet of cover, then about 3,000 feet of erosion would have to take place during the Upper Cretaceous so that the volcanics could occupy their present position (2,500 feet below the present summit). Obviously, if the complex crystallized under more than 1,000 feet of cover, then a correspondingly greater rate of erosion is demanded, and
this is the significance of the volcanic rocks: unless the rate of erosion in Upper Cretaceous time was high (and there appears to be little reason to believe that it was), the complex could not have had much cover. Proposals of greater cover for the complex must account for the increased rates of erosion so demanded.

(5) Proposed Structure

Cauldron subsidence, ring-dykes and cone-sheets are typically sub-volcanic structures associated with conditions of relative crustal stability (i.e. a late- or post-orogenic environment). The classic examples have been described by Clough et al. (1909) and by Bailey et al. (1916) from the Ben Nevis-Glen Coe-Etive district. The Tertiary complexes of Skye, Rhum and Ardnamurchan are famed for their ring structures, and the emplacement of the Mourne granites was explained by Richey (1928) on the basis of successive cauldron subsidences. Ring structures and cauldron subsidence have been described in the Damaraland, e.g. Okonjeje (Simpson, 1954) and Messum (Korn and Martin, 1954); in Northern Rhodesia (Turner, 1963); in New Hampshire (Kingsley, 1931; Billings, 1945) and Vermont (Chapman and Chapman, 1940); and from several localities in Eastern Australia (Hills, 1959).

Unfortunately the terms ring-dyke and ring-intrusion are often applied indiscriminately to igneous bodies which have arcuate outcrops. The danger then arises of assuming that since there appears to be a ring-dyke then cauldron subsidence has taken place. Commonly, stratigraphical evidence of subsidence or direct evidence of the
existence of a submerged cauldron is absent, and identification of the
structure depends on indirect information and comparisons with
authenticated occurrences.

The similarities between the Mt Dromedary intrusions and
recognized cauldron subsidence complexes have been pointed out above:

(i) High crustal level (sub-volcanic).
(ii) Non-orogenic environment.
(iii) Concentricity of outcrop.
(iv) The emplacement of the banatite was much more
permissive than that of the monzonite.

In addition, the contact between the porphyritic margin of
the banatite and the monzonite is smooth, regular, and rarely crossed
by banatite dykes; and it is proposed that a nearly vertical,
cylindrical fracture or fault was induced in the largely or wholly
solid monzonite, either by roof collapse consequent on the effusion of
volcanic material should the monzonite have penetrated the surface, or
from tension on cooling allied with pressure fluctuations in the magma
chamber below.

At room temperature the banatite has a specific gravity of
about 2.66 and the monzonite about 2.87, and if the ring fracture was
associated with roof collapse in the monzonite, the cauldron would
tend to sink in the banatite magma. The cauldron of monzonitic rock
in that case would be cylindrical, rather than conical as proposed by
Anderson (1936), and a very large fraction of the initial intrusion of
monzonite.
Alternatively, the monzonite enclosed by the ring fracture may have been slowly forced upward by the rising column of banatite, weathering away as it appeared at the surface. This hypothesis has the advantages of better accounting for the flow structures in the banatite, and explaining the apparent absence of xenolithic material or corresponding monzonite-banatite hybrids. The cylindrical shape of the cauldron is not the disadvantage to the idea of extrusion that it is to the hypothesis of cauldron subsidence.

The fact that the whole of the banatite shows steep flow structures indicates that the magma was in movement up to the time of freezing, and also that the magma was probably rather viscous. The viscosity and late movement are indicated in thin section by granulation along grain boundaries, bending of the minerals, and $2V's$ of up to $15^\circ$ in the quartz. Viscosity and prolonged movement in the banatite would hinder foundering of the cauldron, but could allow its extrusion.

In summary, the great bulk of the complex was emplaced through upward movement of magma. The monzonite at Mt Dromedary was a forceful intrusion and made room by shouldering aside the country rock. The banatite does not show similar features of forceful intrusion but it appears to have been intruded along the site of a ring fracture in the older monzonite. The pyroxenite at Poole's Point is structurally inhomogeneous but the more massive rock at the drill site on the shore of Tilba Tilba Lake extends to a depth of at least 600 feet without any appreciable variation in structure or composition.
(6) The Laccolith Hypothesis

The proposal by Brown (1930) that the complex has the form of a laccolith, possibly emplaced along the unconformity between the folded basement and the overlying Upper Devonian rocks, fails to account for the following:

(i) The porphyritic banatite-monzonite contact and the monzonite-country rock contact are knife-sharp and steeply dipping;

(ii) Dykes of porphyritic banatite intrude the monzonite;

(iii) The pyroxenite at Poole's Point does not show structures consonant with its formation by gravitative accumulation in situ, and appears to be very thick. In the vertical core, no appreciable change in composition or structure was found over 602 feet;

(iv) Similarly, no variation was noted in the 412 feet of core from the monzonite;

(v) Unless the floor of the proposed laccolith were extremely irregular, the geometry of the body would assure a much greater exposure of pyroxenite and monzonite than exists at this time; and

(vi) There is strong evidence that the banatite and the monzonite were emplaced by vertical movement of magma.

The evidence discussed above favors the high-level emplacement of most members of the complex by forceful injection. The relatively large mass of banatite was probably emplaced, after ring fracture or faulting in the monzonite, either by cauldron subsidence of the cylindrical core or by its slow extrusion (and contemporaneous weathering) during the
ascent of a viscous banatite magma.

At least some of the pyroxenite near Poole's Point appears to have originated by crystal settling on the floor of a magma chamber, but it has clearly been deformed, perhaps by slumping, and caught up in younger ultramafic material. The occurrence of very similar rocks in several of the minor monzonitic intrusions suggests that there may be a common source of pyroxenitic rock in the Mt Dromedary district.
PETROGRAPHY

From the samples collected in the field, over five hundred thin sections were cut and examined. Such a number of thin sections was prepared for two quite different reasons: firstly, in order to examine within-body variation on a small scale; and secondly, because the writer was faced with the combination of complex (and critical) field relations, and very poor exposure in the area north and north-east of Little Boundary. In hand specimens, some of the over- and under-erupted pyroxenitic rocks are virtually indistinguishable, partly because they are relatively fine-grained and their constituents are difficult to recognize. An effort has to be made to, in effect, petrographic mapping.

Petrographic descriptions are given below of the principal rock-types, with special reference to the pyroxenitic rocks cropping out near People's Point and to the altered pyroxenitic rocks found north of Little Boundary.

(a) Major Intrusions

(b) Harzitite at Mt Boundary.

Two series of specimens were collected from Mt Boundary, one section running from the summit east toward Central Tihka, and the other running from the summit north to the periphery of the intrusion. The writer's observations confirmed those of Brown (1939, p.643) that there is a progressive variation in the character of the rock from the core to the contact.
From the samples collected in the field, over five hundred thin sections were cut and examined. Such a number of micro-sections was prepared for two quite different reasons: firstly, in order to examine within-body variation on a small scale; and secondly, because the writer was faced with the combination of complex (and critical) field relations, and very poor exposure in the area north and north-east of Little Dromedary. In hand specimens, many of the over- and undersaturated monzonitic rocks are virtually indistinguishable, partly because they are relatively fine-grained and their constituents are difficult to recognize; recourse has to be made to, in effect, petrographic mapping.

Petrographic descriptions are given below of the principal rock-types, with special reference to the pyroxenitic rocks cropping out near Poole’s Point and to the altered pyroxenitic rocks found north of Little Dromedary.

(1) Major Intrusions

(a) Banatite at Mt Dromedary.

Two series of specimens were collected from Mt Dromedary, one section running from the summit east toward Central Tilba, and the other running from the summit north to the periphery of the intrusion. The writer’s observations confirmed those of Brown (1930, p.642) that there is a progressive variation in the character of the rock from the summit to the contact.
The rock at the top of the mountain is medium to fine grained, pale pink or grey, and speckled with small mafic grains. The grain is fairly even, although there are a few phenocrysts of alkali feldspar which are sometimes found in the shape of a cross, formed by the intergrowth of two grains. Quartz, plagioclase, biotite and hornblende are readily recognized in the hand specimen. The number of phenocrysts and the amount of mafic minerals increase, and the quantity of quartz decreases, toward the margin of the intrusion. The rock then becomes greatly enriched in alkali feldspar phenocrysts of tabular habit, commonly measuring 3 x 2 x 1.5 cms and elongated along the 'c' crystallographic direction with the (010) face oriented sub-parallel to the plane of local contact. There is a concomitant increase in the proportion of mafic minerals, clinopyroxene becoming noticeable, while quartz can seldom be found.

Under the microscope the rock is seen to have a hypidiomorphic granular texture, and is made up of perthitic alkali feldspar, twinned and zoned plagioclase, and varying amounts of quartz, biotite, hornblende, clinopyroxene and accessory minerals. Some thin sections show particularly well the alignment of the alkali feldspar grains (RB260, RB 267) noticeable in some hand specimens. The alignment of these grains and the general texture of the rock is brought out by staining the potash feldspar with sodium cobaltinitrite (Chayes, 1952). The present writer found very little trace of the "strong tendency to monzonitic fabric" recorded by Brown (1930, p.642).

The alkali feldspar occurs as prismatic or even lath-like grains
### TABLE 3. Banatite modes and specific gravities.

<table>
<thead>
<tr>
<th>Mode</th>
<th>RB198</th>
<th>RB260</th>
<th>RB261</th>
<th>RB262</th>
<th>RB264</th>
<th>RB265</th>
<th>RB267</th>
<th>RB268</th>
<th>RB269</th>
<th>RB222</th>
<th>RB109</th>
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<tbody>
<tr>
<td>Quartz</td>
<td>14.4</td>
<td>14.2</td>
<td>14.7</td>
<td>14.8</td>
<td>12.9</td>
<td>14.1</td>
<td>11.2</td>
<td>8.4</td>
<td>10.2</td>
<td>9.4</td>
<td>3.5</td>
</tr>
<tr>
<td>Orthoclase</td>
<td>46.2</td>
<td>36.0</td>
<td>46.5</td>
<td>43.9</td>
<td>37.0</td>
<td>38.6</td>
<td>48.3</td>
<td>47.1</td>
<td>37.8</td>
<td>55.8</td>
<td>66.3</td>
</tr>
<tr>
<td>Plagioclase</td>
<td>37.3</td>
<td>42.7</td>
<td>34.1</td>
<td>37.8</td>
<td>43.6</td>
<td>40.2</td>
<td>34.4</td>
<td>33.7</td>
<td>43.6</td>
<td>28.0</td>
<td>22.4</td>
</tr>
<tr>
<td>Biotite</td>
<td>-</td>
<td>0.4</td>
<td>tr.</td>
<td>-</td>
<td>0.3</td>
<td>0.1</td>
<td>1.5</td>
<td>2.5</td>
<td>2.7</td>
<td>2.0</td>
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<tr>
<td>Hornblende</td>
<td>1.0</td>
<td>-</td>
<td>3.0</td>
<td>1.5</td>
<td>4.2</td>
<td>4.4</td>
<td>1.8</td>
<td>5.9</td>
<td>2.7</td>
<td>3.7</td>
<td>3.2</td>
</tr>
<tr>
<td>Pyroxene</td>
<td>-</td>
<td>0.6</td>
<td>tr.</td>
<td>-</td>
<td>0.3</td>
<td>0.6</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Chlorite</td>
<td>-</td>
<td>3.5</td>
<td>tr.</td>
<td>0.3</td>
<td>-</td>
<td>-</td>
<td>1.0</td>
<td>0.6</td>
<td>0.8</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Opaques</td>
<td>0.9</td>
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<td>1.3</td>
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<td>1.8</td>
<td>1.2</td>
<td>1.7</td>
<td>1.3</td>
<td>1.1</td>
<td>0.8</td>
</tr>
<tr>
<td>Calcite</td>
<td>-</td>
<td>1.2</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.4</td>
<td>-</td>
<td>0.9</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Sphene</td>
<td>0.1</td>
<td>0.3</td>
<td>0.4</td>
<td>0.2</td>
<td>0.5</td>
<td>0.2</td>
<td>0.1</td>
<td>0.1</td>
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<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Specific</td>
<td>-</td>
<td>2.59</td>
<td>2.59</td>
<td>2.63</td>
<td>2.62</td>
<td>2.64</td>
<td>2.63</td>
<td>2.61</td>
<td>2.64</td>
<td>2.64</td>
<td>2.65</td>
</tr>
</tbody>
</table>

All modes in volume per cent.
which are almost invariably perthitic. The exsolved sodic phase (mean composition $\text{An}_{13}$) forms a reticulate network of subparallel veins in the host. In sections near (100) the exsolved oligoclase appears as sub-rectangular grains aligned in two directions normal to one another, giving a grid-like appearance to the orthoclase grain. Twinning on the Carlsbad law is very common. The values for $2V$ and the $\beta$ index showed an appreciable range, with the optic axial angle about 60° and $\beta = 1.526-1.529$. The plagioclase is almost invariably zoned (oscillatory normal) and shows fairly fine lamellar twinning. The most calcic core measured was $\text{An}_{45}$ but the range in mean values for core and rim is from $\text{An}_{28}$ to $\text{An}_{19}$.

Quartz occurs as anhedral interstitial plates, with undulose extinction. Strain $2V$'s of up to 15° are common. The quantity of quartz varies from a maximum of about 15 vol. per cent in the central part of the intrusion (RB262) to about 3 vol. per cent in the porphyritic margin (RB109).

The mafic minerals are usually quite subordinate except in the marginal zone. In the more siliceous phases deuteritic activity has been responsible for the production of pennine, magnetite and calcite from the hornblende and biotite, and the ilmenite of the lamellar intergrowths of magnetite and ilmenite has been converted into sphene which now contains the parallel blades of magnetite. Sphene has also resulted from the penninization of the biotite and hornblende, and in a few of the strongly altered, more mafic border phases (RB166) rutile and a little tourmaline may occur. Deuteritic action on the feldspars has produced
### TABLE 4. Banatite analyses.

<table>
<thead>
<tr>
<th></th>
<th>RB351A</th>
<th>VII</th>
<th>VIII</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>64.02</td>
<td>64.49</td>
<td>59.44</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.36</td>
<td>0.46</td>
<td>0.54</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>17.05</td>
<td>17.48</td>
<td>19.58</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>1.99</td>
<td>1.64</td>
<td>0.31</td>
</tr>
<tr>
<td>FeO</td>
<td>2.17</td>
<td>1.69</td>
<td>3.91</td>
</tr>
<tr>
<td>MnO</td>
<td>0.13</td>
<td>0.11</td>
<td>0.07</td>
</tr>
<tr>
<td>MgO</td>
<td>1.04</td>
<td>0.66</td>
<td>1.27</td>
</tr>
<tr>
<td>CaO</td>
<td>3.08</td>
<td>3.28</td>
<td>3.95</td>
</tr>
<tr>
<td>BaO</td>
<td>0.12</td>
<td>tr.</td>
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</tr>
<tr>
<td>Na₂O</td>
<td>4.73</td>
<td>4.16</td>
<td>3.21</td>
</tr>
<tr>
<td>K₂O</td>
<td>4.67</td>
<td>4.79</td>
<td>6.60</td>
</tr>
<tr>
<td>H₂O⁺</td>
<td>0.62</td>
<td>0.52</td>
<td>0.88</td>
</tr>
<tr>
<td>H₂O⁻</td>
<td>nil.</td>
<td>0.18</td>
<td>0.12</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.10</td>
<td>0.22</td>
<td>0.07</td>
</tr>
<tr>
<td>CO₂</td>
<td>0.23</td>
<td>0.71</td>
<td>0.49</td>
</tr>
<tr>
<td>S</td>
<td>n.d.</td>
<td>0.06</td>
<td>-</td>
</tr>
<tr>
<td>Total</td>
<td>100.31</td>
<td>100.45</td>
<td>100.44</td>
</tr>
</tbody>
</table>

**RB351A** - Banatite, near Mt Dromedary Trig. Anal. R.S. Boesen.


pennine, calcite, muscovite and a little epidote (RB260). In general, deuterite alteration is slight.

Clinopyroxene occurs in appreciable quantity (> 0.5 vol.%) only in the porphyritic marginal phase, and is then usually being replaced by biotite or hornblende. Analyses and optical data for all three minerals are given below.

Table 4 shows three analyses of the banatite, one from the summit (RB351A), one from near the porphyritic margin and one from the margin (VIII and VII respectively from Brown, 1930, p.680). Table 3 gives the modal variation in the section from the summit to the northern contact.

(b) Monzonite.

The monzonitic body has an annular outcrop circling Mt Dromedary, bounded on the inside by the porphyritic banatite and on the outside by the metamorphosed Wagonga Beds.

The color of the rock varies a little from light grey to a darker grey-green, the 'grey granite' and 'black granite' respectively referred to by Brown (1930, p.648). The monzonite is even grained and the average grainsize is about 3 mm, although both alkali feldspar and biotite occur in irregular plates as much as 4 cm in diameter.

Under the microscope the monzonites show typical monzonitic texture. The rock is made up of alkali feldspar which is not usually perthitic, plagioclase, and varying amounts of biotite, hornblende, quartz and clinopyroxene. Accessories include apatite and opaques, chiefly magnetite-ilmenite intergrowths, magnetite and ilmenite grains, pyrite
Figure 31. Monzonite from drill core, 1/4 mile north of Tilba
Tilba, 93 feet below collar. Small opaque grains, dark biotite, and
granular pyroxene are set in alkali feldspar and subordinate
plagioclase. The arrow marks radiating biotite, formed as a late
growth on opaque minerals. Ordinary light, 5 diameters. (The
dark patch covering most of the photograph is caused by the cementing
medium.)
and chalcopyrite (Figure 31).

The alkali feldspar forms large optically continuous plates in which the other minerals are set. It is usually clear and relatively free from alteration, but in most sections it appears to be replacing the plagioclase (RB96). The optic axial angles range from 60° to 70°, but the $\beta$ indices are more constant ( $\beta = 1.526-1.528$).

Plagioclase forms prismatic grains usually replaced in some degree by the orthoclase. The grains are zoned from An$_{50}$-An$_{35}$ and some cores reach An$_{70}$. Lamellar twinning is very common. Vermicular intergrowths of the more acid plagioclase with the orthoclase may make up 20 vol.% of some specimens.

The light brownish-green hornblende is associated with the pyroxene and appears to bear a reaction relation to it. The pyroxene is well formed and may contain inclusions of apatite and magnetite. The biotite forms ragged flakes enclosing the pyroxene or occurring in discrete grains. Inclusions of apatite and opaques are common. Minor alteration to hornblende has taken place in a few grains. The biotite is an olive-yellow color when it is found in the body of the rock, but when it forms as shells or as radiating spines around the opaque grains (Figure 31) it is a strongly pleochroic red variety.

Quartz is a minor constituent, but it is present in every sample, whereas Brown (1930, p.649) did not find quartz in the 'grey granite'. It is interstitial and clearly of late crystallization. It may or may not show strained extinction.

A curious feature of the replacement of the plagioclase by ortho-
### TABLE 5. Modes of monzonites.

<table>
<thead>
<tr>
<th></th>
<th>RB117</th>
<th>RB120</th>
<th>RB121*</th>
<th>RB221**</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quartz</td>
<td>1.8</td>
<td>3.5</td>
<td>7.5</td>
<td>tr.</td>
</tr>
<tr>
<td>Alkali feldspar</td>
<td>44.7</td>
<td>40.0</td>
<td>56.8</td>
<td>43.8</td>
</tr>
<tr>
<td>Plagioclase</td>
<td>24.4</td>
<td>30.5</td>
<td>22.8</td>
<td>28.6</td>
</tr>
<tr>
<td>Biotite</td>
<td>12.9</td>
<td>5.1</td>
<td>2.6</td>
<td>7.6</td>
</tr>
<tr>
<td>Hornblende</td>
<td>1.1</td>
<td>12.0</td>
<td>6.0</td>
<td>-</td>
</tr>
<tr>
<td>Pyroxene</td>
<td>12.5</td>
<td>5.7</td>
<td>2.6</td>
<td>17.6</td>
</tr>
<tr>
<td>Opaques</td>
<td>2.6</td>
<td>3.2</td>
<td>1.7</td>
<td>2.4</td>
</tr>
</tbody>
</table>

*Light monzonite.

**Dark monzonite (analyzed).

RB117 - Monzonite adjacent to large xenolith of volcanic rock, north-west of Mt Dromedary Trig.

RB120 - From road cutting, ¾ mile north along highway from Tilba Tilba.

RB121 - From road cutting, 1½ miles north along Wagonga road from Central Tilba.

RB221 - From Central Tilba quarry.
### TABLE 6. Analyses of monzonites.

<table>
<thead>
<tr>
<th></th>
<th>RB221</th>
<th>I</th>
<th>II</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>53.29</td>
<td>51.09</td>
<td>54.20</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.75</td>
<td>1.02</td>
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<tr>
<td>Al₂O₃</td>
<td>17.32</td>
<td>16.11</td>
<td>15.73</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>2.00</td>
<td>3.11</td>
<td>3.67</td>
</tr>
<tr>
<td>FeO</td>
<td>5.28</td>
<td>6.58</td>
<td>5.40</td>
</tr>
<tr>
<td>MnO</td>
<td>0.14</td>
<td>0.18</td>
<td>0.70</td>
</tr>
<tr>
<td>MgO</td>
<td>3.73</td>
<td>4.69</td>
<td>3.40</td>
</tr>
<tr>
<td>CaO</td>
<td>6.92</td>
<td>9.10</td>
<td>8.50</td>
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<tr>
<td>BaO</td>
<td>0.03</td>
<td>nil</td>
<td></td>
</tr>
<tr>
<td>Na₂O</td>
<td>3.23</td>
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<td>3.07</td>
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<tr>
<td>K₂O</td>
<td>5.81</td>
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<td>4.42</td>
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<tr>
<td>H₂O⁺</td>
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<td>0.50</td>
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<tr>
<td>H₂O⁻</td>
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<td></td>
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<tr>
<td>P₂O⁵</td>
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<td>0.77</td>
<td>0.50</td>
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<tr>
<td>CO₂</td>
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</tr>
<tr>
<td>Total</td>
<td>99.92</td>
<td>100.64</td>
<td>100.49</td>
</tr>
</tbody>
</table>

RB221 – Dark monzonite, quarry, Central Tilba.
Anal. R.S. Boesen.

I – Light monzonite, quarry, Central Tilba.

II – Monzonite, Monzoni, Tyrol. Anal. V. Schmelk.
clase is that a very thin film of more calcic plagioclase is deposited around the pyroxene grains, effectively removing them from a reaction relation with hornblende or biotite. The reaction involved in the replacement and deposition may be written:

\[ K^+ + (Ca^{++} + Na^+) \rightarrow (K^+ + Na^+) + Ca^{++} \]

orthoclase plagioclase alkali anorthite
about An50 feldspar

Table 6 gives the analyses of the 'grey granite' (Brown, 1930, Table 3, p.650) and the darker monzonite ('black granite') from the quarry at Central Tilba, together with the Monzoni rock analyzed by Schmelok (Brogger, 1895). Some modes of monzonite from Mt Dromedary are given in Table 5.

The difference between the light and dark monzonite appears, from comparison of RB121 and RB221 in Table 5, to be chiefly in the pyroxene content. However, other specimens are lower in pyroxene and higher in biotite, and it is probable that the depth of color is related to the biotite content. It is significant that xenolithic material, rich in biotite, is intimately related to the darker monzonite and is drawn out in conspicuous schlieren (see Figures 9, 10). Furthermore Table 6 shows that the amount of K2O in RB221 is much higher than in the rock analyzed by Brown (1930); this would be a consequence of high biotite content, provided the amounts of alkali feldspar in the two rocks were comparable.
(2) Minor Intrusions

A number of small intrusive bodies are spread around the southern aspect of Mt Dromedary. Some of them appear to be hybrids, produced by the action of monzonitic magma on mafic or ultramafic clinopyroxene-rich rocks. Some are just as clearly related in a simple way to the banatite and monzonite at Mt Dromedary. A number of samples from these minor intrusions have been sectioned and examined and some analyses have been carried out, but otherwise little attempt was made to study the problem of their genesis.

Small stock-like bodies.

(i) A small plug crops out on the road to Eurobodalla, and due west of the Mt Dromedary Trig. Brown (1930) has reported the presence in this body of fragments of amphibolized pyroxenites, essexite and recrystallized monzonite, set in a matrix of variable aspect.

On examination, the body was found to be dominantly monzonitic in composition, but inhomogeneities do occur, in particular in its eastern and northern portions through which the road passes. In these outcrops the rock, which resembles that immediately south of Tilba Tilba village, is very dark grey in color, medium to fine grained, and weakly porphyritic in pyroxene. It is composed of abundant grains of clino­pyroxene, optically indistinguishable from the pyroxenes in the monzonite at Mt Dromedary. Olivine of composition $\text{Fa}_{23} \pm 2$ (determined from $2V$ and $\beta$ index, referred to the curves of Poldervaart, 1950) is also present; it is slightly altered and is enclosed in pale, reddish biotite. The
remainder of the rock consists of a fine grained aggregate of zoned and twinned basic andesine occurring in stumpy grains, and interstitial plates of alkali feldspar and quartz.

The rock in the interior of the mass is even more clearly monzonitic, although relict olivine persists. Quartz reaches about 5 vol. per cent.

(ii) The intrusion cropping out three miles south-west of Mt Dromedary Trig is very similar in composition and appearance to the even grained margins of the monzonite at Mt Dromedary. The rock is light to medium grey in color and medium grained, with a rather monzonitic texture. It is composed of alkali feldspar and plagioclase, with clinopyroxene (again optically identical with that in the monzonite at Mt Dromedary), ragged grains of yellow-brown biotite, and a little interstitial quartz.

(iii) The intrusion which crops out in cuttings on the highway about 3½ miles due south of Mt Dromedary Trig has been described by Brown (1930) as olivine monzonite. It closely resembles the more monzonitic portion of the rock described under (i) above; like that rock, the olivine monzonite is probably a hybrid.

(iv) The rock associated in the field with the 'ijolite' is also monzonitic; it is dark grey and even grained, with an average grainsize of about 1-2 mm. It has a somewhat monzonitic texture, and is
composed of tablets of plagioclase set in a groundmass of alkali feldspar plates; pyroxene and biotite are the dominant dark minerals, with subordinate amounts of greensih-brown amphibole, opaque minerals, apatite, and chloritic pseudomorphs after olivine.

(v) Due east of Central Tilba there are three small outcrops of fine grained quartzo-feldspathic rock, given on the geological map as "small micro-banatite intrusions". Two, at least, of these bodies are stock-like in form. They are leucocratic, and have a rather trachytic texture with a tendency to be weakly porphyritic in alkali feldspar laths. Quartz is a common minor constituent, occurring as interstitial or rounded grains. Mafic minerals are rare, but there are a few grains of light brown biotite, opaque minerals and, very rarely, pyroxene. Chemically, the rocks are very similar to the core banatite and the leucocratic dyke rocks.

(vi) Closely associated with (v) in the field are two outcrops of moderately fine grained monzonitic rock similar to the main body of monzonite at Mt Dromedary. They are petrographically very similar, but since they are overlain by lavas and tuffs it is not certain whether the two outcrops are part of the same mass. They are composed of zoned and twinned plagioclase (An₄₀) in tabular grains, set in a sparse groundmass of alkali feldspar. Pyroxene is prominent, frequently rimmed by deep red-brown biotite. A few partly altered grains of olivine are also present, and a very small amount of quartz is found in the interstices
<table>
<thead>
<tr>
<th>Mineral</th>
<th>RB82</th>
<th>RB171</th>
<th>RB192</th>
<th>RB331</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quartz</td>
<td>1.2</td>
<td>1.6</td>
<td>tr.</td>
<td>0.2</td>
</tr>
<tr>
<td>Alkali feldspar</td>
<td>20.4</td>
<td>32.2</td>
<td>23.1</td>
<td>23.1</td>
</tr>
<tr>
<td>Plagioclase</td>
<td>32.8</td>
<td>42.6</td>
<td>39.2</td>
<td>43.7</td>
</tr>
<tr>
<td>Olivine</td>
<td>3.6</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Pyroxene</td>
<td>32.9</td>
<td>14.4</td>
<td>17.2</td>
<td>16.4</td>
</tr>
<tr>
<td>Hornblende</td>
<td>-</td>
<td>-</td>
<td>1.3</td>
<td>2.3</td>
</tr>
<tr>
<td>Biotite</td>
<td>5.3</td>
<td>5.2</td>
<td>15.3</td>
<td>9.3</td>
</tr>
<tr>
<td>Opaques</td>
<td>3.1</td>
<td>3.7</td>
<td>1.7</td>
<td>4.3</td>
</tr>
<tr>
<td>Accessories</td>
<td>0.7</td>
<td>0.3</td>
<td>0.2</td>
<td>0.3</td>
</tr>
<tr>
<td>Chlorite</td>
<td>-</td>
<td>-</td>
<td>2.0</td>
<td>-</td>
</tr>
</tbody>
</table>

RB82 - Stock, 3 miles west of Mt Dromedary Trig.
RB171 - Stock, 3 miles south-west of Mt Dromedary Trig.
RB192 - From monzonitic associate of 'ijolite', 3 miles south-south-west of Mt Dromedary Trig.
RB331 - Monzonite, 1½ miles east-north-east of Central Tilba.
between the other minerals (RB331).

Some modes of the monzonitic satellites are given in Table 7.

(vii) The structural relation of the dykes to the intrusive complex has been described above. The dykes appear to be in a swarm striking west of north, and there may also be a radial pattern of dyke intrusion centred on Mt Dromedary. The outcrop of the dykes is usually very poor, except in deep road cuttings and on the low, wave-cut rock platform bordering the sea between Poole's Point and Cape Dromedary, about two miles north. In this section, the metasediments form a narrow discontinuous bench which has been intruded by a number of dykes. About three-quarters of a mile of the coastline in this area was mapped using the plane table (Plate III), and the dykes cropping out were studied to determine the variety of rock types and their sequence of intrusion. It was felt that the information gained in this study would help to elucidate the sequence of intrusion of the larger bodies of the monzonitic complex.

The country rock consists of tightly folded, thinly bedded phyllitic metasediments and minor cherty bands. The beds are penetrated by irregular veins of white quartz varying in thickness from a few millimetres to 20 cm. Because of the folding the beds dip very steeply, and their strike in this area is very consistently north-south, although immediately beyond the northern boundary the strike swings rapidly to the north-west. The bedding and strike of the country rock are so regular that the hatching shown on the map is a good representation of the actual relations of the metasediments to the dykes.
Figure 32. Looking north-east from Poole's Point; pyroxenite rubble in the foreground and Wagonga Beds forming the low headland beyond. The right extremity of the headland is the southern limit of mapping shown in Plate III (4). Arrow marks the site of a small exposure of monzonitic rock believed to be associated with hybrid dykes in the pyroxenite.
More than seventy dykes and several small rounded intrusions crop out in this coastal section. The great majority of the dykes are quite narrow, ranging from about an inch to three feet thick, with some up to fifteen feet thick; they trend east-west cutting sharply across the strike of the country rock. The dykes dip fairly steeply, but neither the dip nor the strike of any one dyke remains constant over its exposed extent. Several dykes terminate abruptly or lens out into the country rock, and offshoots are a common feature of the larger bodies. Many of the narrow dykes show small dislocations resulting from minor faults.

Two dykes are much larger than their fellows, being about fifty feet thick, and are much more nearly conformable with the metasediments. Consequently, they strike roughly at right-angles to the majority of the dykes and are cut in several places by them.

Virtually all the dykes are composed of either dolerite, camptonite or a quartz-feldspar rock. Two garnet-bearing foyaitic dykes, and three of analcite basalt, were found. The small rounded intrusions are made up of camptonite, and the breccias associated with some of them consist of fragments of the country rock cemented together by a very small amount of camptonitic material.

The dolerites were intruded first, followed by the camptonites and then the quartz-feldspar dykes. The dykes of analcite basalt appear to be younger than the quartz-feldspar dykes. The garnet-bearing foyaitic rocks are younger than the dolerites but their relation to the other dykes is uncertain.
(a) The dolerite dykes are made up of a medium grained, holocrystalline rock which has very sporadic clots of andesine-labradorite feldspar, giving it a glomeroporphyritic appearance. Under the microscope the texture appears intergranular to sub-ophitic, with a pale pink diopsidic clinopyroxene, pale green uraltic amphibole and chlorite, weakly zoned plagioclase, and minor anhedral quartz, numerous tiny grains of opaque minerals, apatite and red-brown biotite. In the coarser grained rocks there is a small amount of interstitial quartz-alkali feldspar intergrowth. It is interesting to note that titaniferous clinopyroxenes are more commonly associated with the alkali basalts, but quartz-alkali feldspar intergrowths are typical of the tholeiites, e.g. Red Hill in Tasmania (McDougall, 1960). A specimen of the dolerite (RB7) was analyzed (Table 29).

(b) The camptonites are fine grained, dark grey rocks with phenocrysts of pyroxene and amphibole. They are holocrystalline and porphyritic, with a microcrystalline groundmass. The texture is rather fluidal. The colorless pyroxene phenocrysts are extensively altered to aggregates of tabular biotite grains. Yellow to greenish-brown amphibole has replaced much of the pyroxene, and also occurs in separate grains. The groundmass is a fine grained mosaic of the mafic minerals and tabular grains of plagioclase and alkali feldspar. Apatite is a common accessory. Inspection of the analysis of RB1 (Table 29) shows its close resemblance to the monzonitic intrusion at Mt Dromedary.

(c) Quartz-feldspar dykes cut all but the analcite basalt dykes in this section. The latter are probably related to the
Lower Tertiary volcanism in eastern Australia and are not discussed herein.

The quartz-feldspar dykes are very light in color, and fine grained; the minerals are strongly aligned, giving a rather trachytic texture to the rock. Mineralogically they are very similar to the microbanatite intrusions described above, and chemically they are closely related to the core banatite at Mt Dromedary (see RB10, Table 29).

(d) The garnet-bearing foyaitic dykes cut the dolerite, but their relations to the other dykes are obscure. They are fine grained, light grey rocks, weakly porphyritic in acid plagioclase, garnet, and soda-rich pyroxene and amphibole. The microscope shows a strongly fluidal texture, with aligned laths of plagioclase (about An₁₅) and alkali feldspar; euhedral, strongly zoned crystals of andraditic garnet (analysis RB16G, Table 26) about 1 mm in diameter; ragged small clots of deep green, weakly pleochroic pyroxene (Z° about 40°); and subordinate, moderately pleochroic amphibole with:

\[
\begin{align*}
X &= \text{yellowish-green} \\
Y &= \text{deep greenish-brown} \\
Z &= \text{deep green}
\end{align*}
\]

\(X < Y < Z, \text{ strong.}\)

A section was stained with sodium cobaltinitrite and a modal analysis showed alkali feldspar 53%, plagioclase 26%, pyroxene and amphibole 7%, and garnet 14%. There is a considerable amount of nepheline in the norm, and much of this has probably been counted in with the alkali feldspar. The fine grain and weathered state of the rock makes mineral identification difficult, and probably accounts for the fact that nepheline was not observed. The chemistry and mineralogy of the rock indicate its genetic link with the nepheline-bearing intrusions.
(3) Pyroxenitic Masses

Several of the intrusions, especially the smaller ones, contain fragments of pyroxenite and of amphibolite derived therefrom. Small pieces of amphibolite, about an inch in diameter, are very common in the more mafic dykes. Wilshire and Binns (1961) have described and discussed the common occurrence of basic and ultrabasic xenoliths from some volcanic rocks in New South Wales.

(a) The pyroxenite complex at Poole’s Point.

The pyroxenite which crops out at Poole’s Point, 5 miles due east of the Mt Dromedary Trig, has been studied in detail, from the point of view of both mesoscopic structures and petrology. Some phases of the pyroxenite have distinctive structures. About three-quarters of the rock in situ is coarse, massive pyroxenite whose dominant constituent is clinopyroxene near salite in composition, with abundant opaque grains, and varying minor amounts of calcic plagioclase, biotite, hornblende, olivine and apatite. The laminated rocks differ in texture from this pyroxenite and the laminated feldspathic type has a distinctly different modal constitution. Certain of the dykes and all the veins contain less pyroxene and opaques than the massive pyroxenite and more amphibole, biotite and a less calcic plagioclase than the types above. Some veins carry as much as 30% of alkali feldspar and other minerals not observed in other parts of the complex, e.g. sphene, chlorite and epidote.

The following types of pyroxenite can be distinguished in the complex at Poole’s Point:

(i) Laminated feldspathic pyroxenite, characterized by the
development of thin laminae rich in plagioclase, and by the prominent folding of these laminae (Figures 15, 17).

(ii) **Laminated dense pyroxenite**, recognized by its longitudinally ribbed, narrow outcrop (Figure 16), and distinguished from (i) by the absence of plagioclase-rich laminae.

(iii) **Massive pyroxenite**, coarser than (i) or (ii), lacking megascopic indication of a laminated structure (Figure 21). This is the type which is the most extensive, and it appears to enclose and intrude (i) and (ii).

Dykes, veins and xenoliths are found in the complex and are relatively insignificant in bulk, but they provide important data concerning the geological history of the complex.

(i) **Laminated feldspathic pyroxenite.** The rocks included in this type are holocrystalline and medium grained with a hypidiomorphic-granular texture, more or less strikingly modified by parallelism of the constituents, particularly plagioclase. The grainsize is uneven, with the tabular pyroxenes averaging about 3.0 x 1.0 x 1.0 mm while the plagioclase grains average about 0.2-0.4 x 0.1-0.2 mm. Biotite flakes reach a diameter of about 10 mm. The rock is composed of calcic plagioclase, clinopyroxene (salite), magnesian olivine, biotite, apatite and certain opaque minerals. The amounts of pyroxene and feldspar vary antipathetically over a wide range (Table 8) and there is a corresponding range in grainsize and texture of the rocks. When feldspar is abundant, thin sections show a very strongly parallel texture (Figure 33) picked
Figure 33. Laminated feldspathic pyroxenite showing alignment of plagioclase. Crossed nicols. Scale mark is 1 mm in length.

Figure 34. Laminated dense pyroxenite, showing pyroxene set in a plexus of amphibole. Plane light. Scale mark is 1 mm in length.
out chiefly by the alignment of the plagioclase twin lamellae. Where the feldspar content is low, the texture is more equigranular and the grainsize increases. The boundaries of the laminae are marked by a sharp decrease in the amount of the locally dominant mineral, which occurs in reduced amount in the adjacent lamina; hence, although there are strong fluctuations in their proportions, no mineral disappears. Each mineral species shows a different pattern of distribution between the laminae. The feldspar and pyroxene show the strongest differentiation into separate laminae while the olivine is more evenly distributed, although a greater part is associated with the pyroxene laminae than with the plagioclase. The opaques and apatite seem to be evenly spread through the rock, but the biotite tends to be more abundant in the plagioclase-rich laminae.

(ii) **Laminated dense pyroxenite.** This pyroxenite is a dense, fairly even grained, coarse rock with a hypidiomorphic-granular texture. Clinopyroxene and opaque minerals make up the bulk of the rock, with minor amounts of plagioclase, olivine, biotite, amphibole and apatite.

The prismatic pyroxene grains are aligned parallel to the plane of lamination of the rock but there is no strong impression of parallelism in the texture, because the very minor plagioclase occurs as anhedral grains confined to the interstices between the pyroxenes. In some sections the pyroxene grains are set in a plexus of either amphibole or biotite which, like the feldspar, are clearly of late crystallization. The lamination in the rock is picked out by sporadic
TABLE 8. Modes of ultramafic rocks from Poole's Point.

<table>
<thead>
<tr>
<th></th>
<th>RB27</th>
<th>RB402</th>
<th>RB32</th>
<th>RB404</th>
<th>RB93</th>
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<td><strong>LAMINATED PYROXENITES</strong></td>
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<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Pyroxene</td>
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<td>41.9</td>
<td>48.1</td>
<td>65.7</td>
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<tr>
<td>Plagioclase</td>
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<td>62.1</td>
<td>33.0</td>
<td>25.3</td>
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<tr>
<td>Olivine</td>
<td>9.1</td>
<td>0.5</td>
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<td>1.3</td>
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<tr>
<td>Biotite</td>
<td>1.9</td>
<td>6.9</td>
<td>9.8</td>
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<td>5.4</td>
</tr>
<tr>
<td>Amphibole</td>
<td>-</td>
<td>-</td>
<td>tr.</td>
<td>0.3</td>
<td>tr.</td>
</tr>
<tr>
<td>Opaques</td>
<td>8.5</td>
<td>12.1</td>
<td>11.4</td>
<td>11.0</td>
<td>22.7</td>
</tr>
<tr>
<td>Apatite</td>
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<td>2.3</td>
<td>1.7</td>
<td>1.1</td>
<td>tr.</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
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<th>RB405</th>
<th>RB406</th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>MASSIVE PYROXENITES</strong></td>
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<td></td>
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<tr>
<td>Pyroxene</td>
<td>67.2</td>
<td>68.5</td>
<td>71.5</td>
<td></td>
<td></td>
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<tr>
<td>Plagioclase</td>
<td>13.2</td>
<td>5.4</td>
<td>4.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Olivine</td>
<td>tr.</td>
<td>tr.</td>
<td>tr.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Biotite</td>
<td>10.2</td>
<td>4.1</td>
<td>1.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Amphibole</td>
<td>0.7</td>
<td>7.6</td>
<td>6.5</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Opaques</td>
<td>7.7</td>
<td>14.3</td>
<td>16.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Apatite</td>
<td>1.0</td>
<td>0.1</td>
<td>tr.</td>
<td></td>
<td></td>
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The modes are listed in order of increasing pyroxene content and decreasing plagioclase content.
Figure 35. Showing nibbled margins of pyroxene grains. Plane light. Scale mark is 1 mm in length.

Figure 36. Blobs of amphibole replacing pyroxene. Plane light. Scale mark is 1 mm in length.
concentrations of opaque minerals, and the lamination is as well
developed as it is in the feldspathic rocks.

(iii) Massive pyroxenite. Thin sections of the massive pyroxenite show only weak grain orientation. In general, the rock is even grained, made up of medium to coarse grains; it is holocrystalline and has allotriomorphic- to hypidiomorphic-granular texture. Pyroxene, in subhedral interlocking grains, is the major constituent, with feldspar anhedral occurring in the interstices. Typically the pyroxene contains numerous small grains of amphibole and biotite (Figure 36). Opaques are abundant, and are found in equant grains distributed evenly throughout the rock.

Both amphibole and biotite are of late formation and replace the pyroxene along grain margins, so that the pyroxene grains may appear to be embedded in plates of either mineral. There is little doubt that of the two, the amphibole is the later mineral, since very often it can be seen enclosing and replacing the biotite (Figure 40). Modes of the principal types are given in Table 8.

The pyroxene in the laminated feldspathic rocks is very pale green, prismatic, subhedral to euhedral, with grains up to 6 x 2 x 3 mm, averaging about 3 x 1 x 1.5 mm. The grains are oriented with their 'c' axis sub-parallel to the plane of lamination. The margins of most grains have a crenulated or nibbled appearance (Figure 35), and a few grains have narrow rims of biotite.
TABLE 9. Chemical analyses of ultramafic rocks from Poole's Point.

<table>
<thead>
<tr>
<th>Chemicals</th>
<th>Dykes</th>
<th>Laminated pyroxenite</th>
<th>Coarse massive pyroxenite</th>
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</thead>
<tbody>
<tr>
<td></td>
<td>RB12A</td>
<td>RB407</td>
<td>RB27</td>
</tr>
<tr>
<td>SiO&lt;sub&gt;2&lt;/sub&gt;</td>
<td>46.26</td>
<td>42.62</td>
<td>38.04</td>
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<tr>
<td>TiO&lt;sub&gt;2&lt;/sub&gt;</td>
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<td>1.63</td>
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<tr>
<td>Al&lt;sub&gt;2&lt;/sub&gt;O&lt;sub&gt;3&lt;/sub&gt;</td>
<td>8.61</td>
<td>9.47</td>
<td>16.86</td>
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<tr>
<td>Fe&lt;sub&gt;2&lt;/sub&gt;O&lt;sub&gt;3&lt;/sub&gt;</td>
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<td>6.00</td>
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<td>FeO</td>
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<td>MnO</td>
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<td>MgO</td>
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<td>7.60</td>
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<td>CaO</td>
<td>15.15</td>
<td>16.56</td>
<td>15.16</td>
</tr>
<tr>
<td>Na&lt;sub&gt;2&lt;/sub&gt;O</td>
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<tr>
<td>K&lt;sub&gt;2&lt;/sub&gt;O</td>
<td>1.74</td>
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<tr>
<td>H&lt;sub&gt;2&lt;/sub&gt;O&lt;sup&gt;+&lt;/sup&gt;</td>
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<td>0.24</td>
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<tr>
<td>H&lt;sub&gt;2&lt;/sub&gt;O&lt;sup&gt;-&lt;/sup&gt;</td>
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<td>P&lt;sub&gt;2&lt;/sub&gt;O&lt;sub&gt;5&lt;/sub&gt;</td>
<td>0.70</td>
<td>1.29</td>
<td>1.34</td>
</tr>
<tr>
<td>Total</td>
<td>99.80</td>
<td>100.43</td>
<td>99.91</td>
</tr>
</tbody>
</table>

Analyses by Avery and Anderson.
Very small, zonally arranged inclusions of apatite, opaque material and plagioclase are common in the pyroxenes. There is also a patchy development of rutile rods, which appear in thin section in one, or commonly two, crystal directions, thus forming a grid (Figure 62). The zoning is described in detail in Part III (Mineralogy); the variation in chemical composition between the zones is slight. Approximately a quarter of the grains are twinned, giving either simple pair twins or repeated pair twins with (100) as composition plane. A few grains have twin lamellae which are discontinuous and terminate within the body of the grain. The well developed prismatic cleavage and the composition plane of twinned grains are sometimes quite distinctly curved. Specimens taken from near the massive pyroxenite which encloses the laminated rock contain pyroxene which is flecked with numerous small patches of olive-colored amphibole. As the amount of plagioclase decreases the pyroxene grains, which are otherwise rarely contiguous, adopt a more anhedral form.

The laminated dense pyroxenite contains abundant tabular grains of pyroxene which have well developed, though slightly rounded, crystal outlines. The grains, which measure about 1.5 x 0.6 x 0.4 mm, are moderately aligned sub-parallel to the plane of lamination. Inclusions, including rutile rods, are much rarer than in the pyroxenes of the more feldspathic rocks, but twinning and zoning by birefringence are developed to a similar degree. Alteration is confined to minor rimming by both biotite and amphibole.

The pyroxene of the massive pyroxenite occurs in subhedral to
anhedral grains of equant form, with nibbled margins. The grainsize is about 2.5 mm, with some grains up to 6 mm. The pyroxene is rich in small inclusions of plagioclase, apatite, biotite and amphibole; rutile rods occur in irregular patches, usually in the core of the grains. The abundant inclusions give these pyroxenes a generally dirty appearance. Weak hour-glass structure is sometimes observed, more easily when the grains are near an extinction position. Weak birefringence zoning is present in most sections; as before, there is a tendency for inclusions to be concentrated in the zones of lower birefringence and especially in the marginal zone, which is often of lower birefringence. There is a strong tendency for the core to be of higher birefringence and to be free of most inclusions except amphibole, which appears to be indiscriminate in its distribution in the grains (Figure 36). The frequency and type of twinning is similar to that described above; occasional offsets, suggestive of micro-faulting, are observed in some twin lamellae. Alteration of the pyroxene to biotite and hornblende is more intense in the more northerly exposures.

**Plagioclase** occurs in the laminated feldspathic pyroxenite as small, tabular grains which have a subparallel alignment within the plane of lamination. In the laminated dense pyroxenite and in the coarse massive rock, the plagioclase is found as anhedral plates filling the interstices between the mafic grains. In both cases the plagioclase is quite limpid, and often contains numerous tiny, rounded blebs of a very dark or opaque mineral. The extinction of most grains is slightly mottled.
Figure 37. Crushed grain of plagioclase in massive pyroxenite. Crossed nicols. Scale mark is 0.2 mm in length.

Figure 38. Faulted and bent twin lamellae in plagioclase. Crossed nicols. Scale mark is 0.2 mm in length.
Where zoning occurs, it embraces a very narrow range of probably not more than 5% An, and is of the progressive normal type. Twinning is found in virtually all of the plagioclase grains. Parallel and complex types (especially Carlsbad-albite twinning) with (Ol0) as the composition plane are strongly dominant; pericline twinning is commonly associated with these types. The number of lamellae in each grain is small, and the twins are of about equal width. Composition planes are commonly vicinal and many grains show the effects of bending, crushing (Figure 37) and microfaulting (Figure 38).

The mean composition of the plagioclase varies slightly from sample to sample and the most calcic mean (An88) is from the laminated dense pyroxenite. Mean values for the massive pyroxenite are close to An86, and the laminated feldspathic pyroxenite plagioclase has a mean of An87. There is no significant difference in the anorthite content of the plagioclase in the various laminae; it is calcic bytownite.

Olivine is found in each of the three types of pyroxenite although the actual amount varies considerably from type to type. It is most abundant in the laminated feldspathic pyroxenite, where there is a rough correlation between the amount of olivine and of plagioclase (Table 8). The olivine typically occurs as colorless, equant anhedra, rarely exceeding 1 mm in diameter. Inclusions of granular opaques are more abundant in the olivine than in the other minerals. It is commonly rimmed by biotite in the laminated feldspathic rocks and, to a lesser extent, by pyroxene. The degree of alteration of the olivine varies with
Figure 39. Poikiloblastic grain of biotite (analyzed sample A) in laminated feldspathic pyroxenite. Plane light. Scale mark is 1 mm in length.

Figure 40. Amphibole replacing biotite. Plane light. Scale mark is 0.2 mm in length.
Figure 41. Feldspar-poor phase of laminated feldspathic pyroxenite. Note biotite replacing plagioclase; pyroxene; abundant inclusions in pyroxene; bending of biotite. Plane light. Scale mark is 1 mm in length.
the rock-type. In the feldspathic laminated rocks it is negligible; in the laminated dense pyroxenite it is distinct, while in the massive type more than half the olivine has been altered to serpentine minerals and, very rarely, bronzite and magnetite. The olivine is not zoned, and determinations of $2V$ and the $\beta$ index (Poldervaart, 1950) indicate a composition of $Fa_{20}$.

The biotite shows three modes of occurrence in the pyroxenites. The most striking of these is seen in sections of the laminated feldspathic pyroxenite which are cut parallel to the plane of lamination. These sections are nearly parallel to the (001) cleavage of the biotite, which appears as large, lacy plates (up to 25 mm) poikilitically enclosing the opaque minerals and plagioclase (Figure 39). In the other pyroxenite phases, biotite marginally replaces and occupies the interstices between pyroxene and olivine. The third mode of occurrence is in ragged, elongate grains, wedged between the other minerals and bearing no obvious replacement relation to them. In some sections, especially those of the massive pyroxenite, it is clear that the amphibole has formed at the expense of the biotite (Figure 40). Without exception, the biotite grains show some degree of bending (Figure 41).

There is considerable variety in the optical properties of the biotites. That from the laminated feldspathic pyroxenite is a light colored magnesian variety, strongly pleochroic from nearly colorless to red; the dominant variety of other types of pyroxenite is darker, and pleochroic from honey-brown to deep red-brown. As one passes into the
Figure 42. Basal section of pyroxene showing replacement by amphibole; note the characteristic cleavages. Crossed nicols. Scale mark is 1 mm in length.
northern exposures of the massive pyroxenite, the biotite acquires an olive tint and in some specimens adjacent grains of the mineral show distinctly different pleochroism schemes, one in red-browns and the other in olive-browns. The values for $2V$ and the $\beta$ index increase with increasing intensity of the olive tint.

Amphibole is not very abundant and is completely absent from much of the laminated pyroxenites. It is moderately pleochroic from olive to deep brown, and $Z^c = 24^\circ$. The amphibole has formed through replacement of pyroxene, usually along grain boundaries, so that in some specimens the mineral is optically continuous over a few centimetres; even at its most extreme, however, it does not replace more than about one-third of its host. Prismatic sections of pyroxene show many irregular patches of optically continuous amphibole, which appears in basal sections as numerous, roughly equant patches showing the characteristic amphibole cleavages (Figure 42). The $[b]$ and $[c]$ crystallographic directions of the pyroxene are retained by the amphibole and the $[a]$ directions are very similar. Replacement of the pyroxene by amphibole took place after the time of twin formation in the former.

The amphibole is often associated with the opaques, and tends to form a plexus enclosing all the earlier formed minerals. Replacement of plagioclase by amphibole is minor. The textural evidence is strong that the amphibole was the latest mineral to be formed; it replaces all the other minerals, including biotite.

In general, the amount of amphibole in the pyroxenites increases from the southern to the northern outcrops.
Figure 43. Lobate, interstitial opaques in laminated feldspathic pyroxenite. Plane light. Scale mark is 1 mm in length.
The opaques are abundant in all phases of the pyroxenite, where they occur as anhedral grains, 0.2-1.0 mm in diameter; or, as frequently happens in the more feldspathic rocks, as tiny and very irregular lobate grains interstitial to the plagioclase (Figure 43). Biotite is very often seen enveloping the opaques. Examination by reflected light reveals the opaque constituents to be mainly magnetite, occurring both as discrete grains and as host to ilmenite, which is the next most important by volume. Ilmenite also forms discrete grains, but it is chiefly found as very well developed exsolution lamellae in the (111) planes of the magnetite. The ratio of magnetite to ilmenite is about 3:1 by volume. Other opaques include the sulphides pyrrhotite, chalcopyrite and pyrite. In a few grains of pyroxene haloes of higher birefringence have developed around inclusions of the opaques.

Apatite is found in quantity only in the laminated feldspathic pyroxenite, where it forms colorless, stumpy euhedra (0.1-0.2 mm in diameter) and numerous small prisms. In the other rocks the apatite crystals are smaller and have an acicular habit. The mineral is very often included in plagioclase and in the outer zones of pyroxene grains.

Zircon and sphene are very rare or absent.
Figure 44. Basic dyke; note fluxion, porphyritic texture.
Plane light. Scale mark is 1 mm in length.
(iv) **Dykes.**

On the basis of microscopical criteria, the dykes which intrude the pyroxenites have been divided into two groups:

(a) **Basic dykes.** with a mineralogy comparable to phases of the laminated feldspathic pyroxenite. There are very few of these and they are found intruding the southernmost body of pyroxenite.

(b) **Hybrid dykes.** These are narrower and more numerous than the basic dykes, and are more common in the northern exposures.

Table 9 gives the chemical analyses of a sample of each type.

(a) **Basic dykes.**

The dyke-rocks are holocrystalline, medium to fine grained, and have a porphyritic texture with phenocrysts of pyroxene and olivine set in a fluidal groundmass composed of lath-shaped plagioclase grains strewn with tiny apatite needles, ragged tabular biotite, equant subhedra of ore minerals, and second generation pyroxene and olivine (Figure 44). The grainsize and mineralogy of the dykes are unchanged up to their borders. There is a slight marginal localization of biotite grains in the dykes which facilitates fracture along the contact surface. The pyroxenite enclosing the dykes shows very little evidence of recrystallization or metasomatism; olivine grains in the pyroxenite abutting against the dykes are unaffected.

The phenocrysts are dominantly pyroxene, pale green in color.
Figure 45. Very fine zoning, and hour-glass structure in pyroxene phenocryst. Crossed nicols. Scale mark is 0.2 mm in length.
The well-formed crystals have a prismatic habit with an average length of 2 mm, and are arranged sub-parallel to the dyke margins. They are free from alteration but they may contain seriate zones of inclusions, parallel to the crystal outlines. The included grains may be of biotite, plagioclase, apatite and opaque minerals. Twinning is of a similar nature and abundance to that of the pyroxenes of the pyroxenitic types (i), (ii) and (iii) above. Fine-scale birefringence zoning and hour-glass structures are common (Figure 45).

The olivine phenocrysts occur in subhedral tabular grains, rarely more than 1 mm in length, and they are oriented sub-parallel to the dyke walls. Measurements of $2V$ and the $\beta$ index indicate a composition of $Fa_{20}$. Some grains show transverse deformation bands. Zoning and twinning are absent.

The groundmass is strongly fluidal and is composed of small, clear, lath-shaped grains of bytownite ($An_{86}$), equant subhedra of olivine, pyroxene and opaques, numerous ragged flakes of magnesian biotite, and abundant tiny needles and prisms of apatite. Many of the plagioclase grains are bent and fractured. Zoning is very weak or absent although weakly mottled extinction does occur, especially in larger grains.

The groundmass is very liberally scattered with small, elongate, ragged grains of biotite which, like the plagioclase, marks out the fluidal fabric and is moulded around the larger grains of the rock. It is quite free from alteration. The biotite replaces both pyroxene and plagioclase to a very minor degree but it is more commonly associated with the opaque grains. The biotite grains are invariably and
conspicuously bent. Pleochroism is strong, from nearly colorless to deep red-brown; optically the mineral resembles the biotite of the laminated feldspathic pyroxenite.

The opaques are found as irregular, anhedral grains, usually less than 0.1 x 0.03 mm, made up largely of ilmenite and magnetite, commonly intergrown. Sulphides were not detected in section but a few chalcopyrite grains were observed in the field. The opaques are frequently moulded around, and occasionally include, plagioclase euhedra, and in general have textures consonant with late crystallization. Their most frequent mineral associate is biotite.
Figure 46. Complex pattern of dykes, northern outcrop, Poole's Point. Xenoliths in dyke near lens cap, at right centre of photograph.
Figure 47. Same as Figure 46, showing the xenoliths in the dyke. Lens cap is 6 mm in diameter.

Figure 48. Biotite corona surrounding recrystallized xenolith. Plane light. Scale mark is 1 mm in length.
(b) Hybrid dykes. The rock of the hybrid dykes is lighter in color than that of the basic dykes, and is less dense. It is a fine grained, dark grey rock but it appears to be very uneven in grainsize, with various sizes of pyroxene, usually not more than 3 mm in length. Biotite is a much more obvious constituent in hand specimens of the rock than in those of the basic dykes. The hybrid dykes are less regular in form than the basic type and they have rafted off pieces of the pyroxenite which they intrude (Figures 46, 47).

The microscope reveals the distinctive uneven-granular texture of the hybrid rocks. The most prominent minerals are pyroxene, plagioclase and biotite, with a small amount of ore minerals and very numerous tiny prisms and needles of apatite. The biotite flakes show weak sub-parallel alignment. Olivine is absent, and amphibole is rare. A few reconstituted xenoliths occur in these dykes. They are generally only a few millimetres in diameter, and are composed of a rounded aggregate of granoblastic grains, chiefly of diopsidic pyroxene, with minor amounts of opaque material and untwinned plagioclase; they have conspicuous flakes of biotite forming coronas around them (Figure 48). The pyroxenite adjacent to the dykes is richer in both biotite and amphibole than is usual, and the plagioclase has been transformed from An$_{86}$ to An$_{57}$.

The grains of pyroxene in the dykes are fairly euhedral and have a large and continuous variation in size from 3.0 mm to less than 0.1 mm. The mineral varies in color from a very pale green to a darker grey-green and is not perceptibly pleochroic. Inclusions of plagioclase, biotite and, to a lesser extent, opaques and apatite are common. These
inclusions often show a seriate arrangement and are commonly confined to the outer parts of the crystals, which may be slightly irregular because of interference by other minerals during their growth late in the crystallization history.

The pyroxenes are prominently zoned; zoning is present in all pyroxenes from the pyroxenite at Poole's Point, but in these rocks the zone types are strongly developed:

(i) Zoning by color in plane light. Crystals usually have either a core of darker grey-green, or a zone of this color surrounding a pale green core. Passing outwards, a broad zone of lighter color is encountered, with a narrow marginal rim of a medium grey-green shade (Figure 66).

(ii) Zoning by inclusions, as described above.

(iii) Zoning by rutile rods. There is a strong tendency for rutile rods to develop jointly with other zonal characters, especially in the zones of lower birefringence. This results in clouding and darkening of the cores, or of the zones between the cores and the crystal margins. This may be a contributory factor in producing the plane light color variations in all but the narrow green rims, which are free of microscopic inclusions.

(iv) Birefringence zoning involves the two styles noted before, namely the relatively large-scale bands of differing birefringence and the very small scale repetitions.

Twinning is less frequent than in other phases of the pyroxenite, The fragmental, xenocrystal nature of much of the pyroxene is indicated
Figure 49. Hybrid dyke, showing xenocrystal pyroxene. Crossed nicols. Scale mark is 1 mm in length.
by rough and abrupt terminations to otherwise quite euhedral grains and
by the termination of the zonal structures against these fracture
surfaces (Figure 49).

Plagioclase occurs as stumpy and platy grains about 0.1 mm, or
less, in width and confined to the groundmass. It is subhedral to
anhedral and tends to be interstitial to the other minerals. It appears
to be quite fresh and unaltered, excepting minor encroachment by
biotite, and it bears inclusions of biotite, sparse opaque granules, and
numerous tiny, acicular apatite crystals.

While the plagioclase of the basic dykes shows very little
evidence of zoning, that of the hybrid dykes is characteristicall
strongly zoned. Passing outward from a calcic core, the plagioclase
composition becomes more sodic until near the grain margins, where the
feldspar is enriched in lime. Broadly speaking, the zoning is progressive
normal; this is particularly true of the smaller grains, many of which
may not be twinned, and whose variation in composition in the zones is
from about An$_{60}$ to about An$_{45}$. The cores of the larger grains have a
mean composition of An$_{66}$ with a few grains as high as An$_{73}$, while the
margins are about An$_{45}$.

Biotite is found in abundant ragged, prismatic grains in the
groundmass. It shows little tendency to replace pyroxene or plagioclase
but is closely associated with the opaques, and has strong olive tones
in its pleochroism. Examination of a concentrate extracted for optical
work revealed that two quite distinct varieties are present, one
pleochroic in reds and browns, and the other in olive brown.
An olive colored amphibole occasionally forms ragged grains similar to some sections of biotite (in plane light) and as a replacement of pyroxene. The amphibole occurs only in trace amounts, and its texture indicates very late formation, contemporaneous with or slightly later than the biotite. Olivine is characteristically absent from these rocks. The amount of opaque material is much reduced; that which is present is found in very small granules enclosed by biotite, or as occasional larger grains included in pyroxene. The apatite forms such minute crystals, especially abundant as inclusions in plagioclase, that no attempt was made to determine their amount accurately by point counting.

(v) Veins.

The veins are more commonly exposed on the surfaces of the rock of the northern outcrops. They are coarse to medium grained with average grainsize being about 2 mm, but with many grains reaching 7 mm. Their texture is hypidiomorphic granular, and the minerals include intermediate plagioclase, alkali feldspar, clinopyroxene, amphibole, biotite, pennine, epidote, clinozoisite, sphenite, apatite, ore minerals and zircon. The mineral abundances vary widely, from amphibole-rich types with minor plagioclase and biotite to types containing about 30% each of alkali feldspar and plagioclase, 20% amphibole, and the remainder composed of the minerals mentioned above.

The more feldspathic veins show a bewildering variety of mineral replacements; clinopyroxene and biotite by amphibole, amphibole
Figure 50. Relict pyroxene and biotite in plate of amphibole.

Plane light. Scale mark is 1 mm in length.
by chlorite, biotite by chlorite, ilmenite by sphene, plagioclase by alkali feldspar, and plagioclase by sericite, clinzoisite and epidote.

The boundaries of the veins are not sharp, but are marked by a more or less rapid transition into the normal mineralogy of the massive pyroxenite. This, together with the sinuous shape of the veins, indicates that the material that formed the veins was highly mobile, since it was introduced either into completely massive rock or along microscopic fissures. The veins, unlike the dykes, are not obviously related to the planes of jointing of the pyroxenitic complex.

The amount of pyroxene contained in the veins is minute and the mineral is found as relict grains enclosed in large plates of amphibole (Figure 50). The optics of the relict grains indicate that they are of similar composition to the pyroxene in the surrounding massive pyroxenite.

The plagioclase of the feldspathic veins forms euhedral prisms up to 1 mm in length. These are usually set in plates of alkali feldspar, giving a rather monzonitic texture (Figure 51). The plagioclase shows lamellar twinning and is strongly zoned. The pattern of zoning is similar to that found in the feldspar of the hybrid dykes, with a more calcic core grading quickly to a more sodic zone just inside the grain boundary and then passing to a lime-enriched rim. In this case, however, the plagioclase is overall more soda-rich, the cores rarely being more calcic than $\text{An}_{50}$. The most sodic zone just inside the grain boundary can be as albitic as $\text{An}_{12}$, with the rim about $\text{An}_{21}$.

Alteration products of the plagioclase include pennine and sericite, the latter occasionally being arranged in zones in the body of the crystal.
Figure 51. Alkali feldspar in transverse veins and patches replacing plagioclase. Black groundmass is alkali feldspar at extinction. Crossed nicols. Scale mark is 1 mm in length.
Alkali feldspar is more abundant in the wider veins, where it may comprise a third of their bulk. It is of rather late crystallization and encloses and replaces the plagioclase (Figure 51). It is not perthitic and is only slightly kaolinized. It has a rather low 2V of 52° and its optical properties suggest that it is a soda-bearing orthoclase cryptoperthite.

Biotite is found in only small amounts, almost invariably associated with either the amphibole or pennine, both of which appear to be replacing the mica (Figure 50). It is pleochroic from yellow to dark red-brown, and resembles the biotite common in the massive pyroxenite.

The amphibole forms large anhedral grains showing prominent prismatic cleavage. It is certain that some at least has formed through replacement of biotite and pyroxene since it encloses relicts of both these minerals (Figure 50). It also contains inclusions of other minerals, mainly plagioclase and apatite. The amphibole is moderately pleochroic:

\[
\begin{align*}
X &= \text{honey-yellow} \\
Y &= \text{olive} \\
Z &= \text{olive.}
\end{align*}
\]

\[
\begin{align*}
X^\circ C &= 25^\circ \\
2V &= \text{ve, high.}
\end{align*}
\]

Pennine replaces the amphibole, but in only very small amounts.

Apatite is unusually abundant in the veins and occurs as large, stumpy prisms, up to 0.5 mm in length. These crystals are frequently included in the amphibole and in the alkali feldspar. The amount of apatite varies within and between the veins, but in some thin sections it is as high as 5% by volume.
The sporadic anhedra of sphene may reach 1 mm in length; they almost invariably enclose opaque minerals. From the parallel arrangement of the blades of opaque inclusions, it would appear that the ilmenite of prior magnetite-ilmenite intergrowths has been made over into sphene, leaving the original blades of magnetite untouched.

The amount of epidote and olinozoisite in the more feldspathic veins locally reaches about 10% by volume. The grains are usually interstitial to the alkali feldspar and also appear as replacement products of plagioclase and amphibole. The grains are very irregular and often exceed 1 mm in length. Pennine is found in an interstitial form, where it appears to be the last formed mineral, and as a replacement product of plagioclase, biotite and amphibole (Figure 52). It rarely makes up more than 1% of the rock. In plane light it is pale green and weakly pleochroic, and under crossed nicols shows strong, anomalous, deep blue interference colors.

(vi) Xenoliths.

The xenoliths enclosed in the laminated feldspathic pyroxenite are fine grained, recrystallized rocks showing poorly developed mineral parallelism in thin section. Some have a granoblastic texture and are composed of weakly aligned, sub-idioblastic laths of limpid calcic plagioclase, between which are scattered xenomorphic grains of diopsidic pyroxene, magnesian olivine and ore minerals. Biotite forms xenomorphic grains and poikiloblastic plates which enclose the ferromagnesian minerals. Apatite occurs as idioblastic needles and prisms scattered
<table>
<thead>
<tr>
<th></th>
<th>RB401</th>
<th>RB408</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pyroxene</td>
<td>44.2</td>
<td>66.3</td>
</tr>
<tr>
<td>Plagioclase</td>
<td>39.3</td>
<td>9.1</td>
</tr>
<tr>
<td>Olivine</td>
<td>3.3</td>
<td>0.3</td>
</tr>
<tr>
<td>Biotite</td>
<td>4.7</td>
<td>7.0</td>
</tr>
<tr>
<td>Amphibole</td>
<td>-</td>
<td>6.2</td>
</tr>
<tr>
<td>Opaques</td>
<td>7.4</td>
<td>9.7</td>
</tr>
<tr>
<td>Apatite</td>
<td>1.1</td>
<td>1.4</td>
</tr>
</tbody>
</table>

RB401 - Fine grained block in laminated feldspathic rock.
RB408 - Coarsely banded xenolith in massive pyroxenite.
through the body of the rock. The plagioclase is not zoned and is
invariably twinned, usually with (010) as the composition plane. Vicinal
planes near (010) are common composition faces. There are only a few
twin sub-individuals in each grain, and simple twins with only two
lamellae are common. Pericline twinning occurs together with the other
types in a few grains. Lamellae are sometimes bent. The pyroxene and
olivine form colorless, equant, xenomorphic grains which are free from
inclusions and alteration, apart from rimming by the biotite. Prismatic
cleavage is not well developed in the pyroxene and it is often difficult
to distinguish this mineral from the olivine. The biotite, both in
texture and in optical properties, closely resembles that in the
laminated feldspathic pyroxenite.

Except for a few minor differences, the banded xenoliths are
identical with the plagioclase-poor parts of the laminated feldspathic
pyroxenite; the most obvious differences are that the xenoliths are
coarser grained and are prominently banded in outcrop. Under the
microscope they resemble the laminated feldspathic pyroxenite, the chief
difference being that replacement of the pyroxene and biotite by
amphibole has resulted in the formation of prominent poikiloblastic
plates of the latter which now enclose these minerals. The optics of
the plagioclase and biotite in the plagioclase-poor phases of the
laminated feldspathic pyroxenite are not significantly different from
those of the same minerals in these xenoliths.

The modes of two xenoliths are given in Table 10.
(4) **Nepheline-Bearing Intrusions**

Of the three bodies included in this group, the nepheline monzonite is by far the largest.

(a) **Ijolite**.

The rock called ijolite by Brown (1930) was described (p. 666) as slightly porphyritic in augite, with a dioritic groundmass. The texture was reported to be poikilitic, and the mineralogy was given as "chiefly pyroxene, nepheline, and melanite, with some orthoclase, plagioclase, olivine, sphene, apatite, sodalite, and iron ore". The rocks examined in the present study differ from those of Brown in several important respects; the amount of nepheline observed (RB188) is very low, both alkali feldspar and plagioclase are present in quantity, and there is abundant green amphibole. The type ijolite was described by Ramsay and Berghell (1891, in Johannsen, 1938), and contains abundant nepheline and pale green augite and small amounts of biotite, sphene and apatite. Brügger (1921) described the Melteig ijolite as carrying some biotite, melanite, sphene, magnetite, pyrite and calcite. Ijolites are rocks composed very largely of nepheline and clinopyroxene (green in most cases, often zoned), the nepheline being somewhat more abundant than the pyroxene, with or without small amounts of melanite, biotite, sphene, opaque minerals, calcite, wollastonite, and alteration products, chiefly of nepheline.

They do not appear to carry more than very small amounts, if any, of alkali feldspar, plagioclase or amphibole, and it is obvious
Figure 53. 'Ijolite'. Note conspicuously zoned pyroxenes, large tablets of alkali feldspar. Ordinary light, 5 diameters.
that the rock specimens taken in the course of this study bear little resemblance to ijolite. The outcrop of the mass in the Mt Dromedary district is very poor and the variations in texture and mineral abundances of the exposed rock are evident even in hand specimens. Consequently the mean composition of the body is not known, and from Brown's descriptions it appears that there are more typically ijolitic phases than those observed by the present writer. Pending more detailed study of this extremely interesting occurrence, the rock is termed 'ijolite'.

The sample selected for rock and mineral analyses (RB188, Figure 53) has a mid-grey body color, being made up of dominantly felsic minerals, speckled with granular clinopyroxene and hastingsitic amphibole. Mafic xenoliths up to several centimetres in diameter are common, and the large grains of clinopyroxene (up to 5.0 mm) in the rock are probably xenocrystal. The phenocrysts of alkali feldspar are flattened on the [b] crystallographic axis and their large (010) faces are aligned sub-vertically, giving the rock a strongly platy structure. The grainsize is variable, with the phenocrysts reaching 10 mm and the plagioclase of the groundmass commonly as small as 0.1 mm. The average grainsize is about 0.5-1.0 mm.

The groundmass has a hypidiomorphic-granular texture and is composed of clinopyroxene, plagioclase, alkali feldspar and green amphibole, with small amounts of pink garnet, sphene, apatite, olivine, analcrite/sodalite, cancrinite and nepheline, and the products of saussuritization and amphibole breakdown, including calcite, muscovite,
epidote, albite, chlorite, biotite, sphene and opaque minerals. A mode of the rock (RB188) gave:

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Vol. %</th>
</tr>
</thead>
<tbody>
<tr>
<td>alkali feldspar</td>
<td>30.3</td>
</tr>
<tr>
<td>salitic clinopyroxene</td>
<td>24.0</td>
</tr>
<tr>
<td>plagioclase</td>
<td>16.5</td>
</tr>
<tr>
<td>hastingsitic amphibole</td>
<td>8.8</td>
</tr>
<tr>
<td>opaque minerals</td>
<td>2.9</td>
</tr>
<tr>
<td>calcite</td>
<td>2.9</td>
</tr>
<tr>
<td>analcite/sodalite</td>
<td>1.9</td>
</tr>
<tr>
<td>apatite</td>
<td>0.6</td>
</tr>
<tr>
<td>rest</td>
<td>12.1</td>
</tr>
</tbody>
</table>

Alkali feldspar: The phenocrysts and larger groundmass crystals are tabular or lath-like, commonly twinned on the Carlsbad law, and are perthitic, especially near the grain boundaries. The interior of the phenocrysts is usually free from exsolution and alteration, and gives zonal extinction. 2V for the limpid cores is as low as 61.5° and rises to about 69° before the exsolved phase becomes microscopically visible. There is no trace of the twinning commonly found in anorthoclase and the feldspar is believed to be orthoclase micro- and crypto-perthite. The very small laths of orthoclase in the groundmass are quite limpid and are of very late crystallization. Their optic axial angle is about 75°.

Plagioclase: Occurs as small well-formed prismatic and tabular crystals enclosed in the homogeneous portions of the larger alkali feldspar grains and scattered through the groundmass. Progressive-normal zoning and lamellar twinning with (010) and vicinal planes near (010) as composition face are prevalent. The mean composition of the cores is An_{60} and the rims are normally about An_{20}. Locally, saussuritization and sodic fluids have formed albitic rims around the plagioclase of the groundmass,
with or without prior resorption of the normal marginal zones.

**Feldspathoids**: These include very minor cancrinite and nepheline and an isotropic mineral which may be either sodalite (cf. Brown, 1930, p. 666) or analcite. It is of very late crystallization and occupies the interstices between the other minerals. It attacks the plagioclase in contact with it, and a very thin film of deep yellow-green pyroxene is developed at its junction with the much lighter colored pyroxene of the rock. It appears to have had very little effect on the potash feldspar and none on the garnet.

**The salitic clinopyroxene** is found in prismatic euhedra about 1.0 mm in diameter. The grains are strikingly zoned, having pale green, weakly pleochroic cores and darker margins, distinctly pleochroic from greenish-yellow to greyish-green. Many crystals show an alternation of light and dark zones between core and rim.

**The amphibole** is strongly pleochroic from greenish-yellow to deep green or deep bluish-green; it is found in subhedral to anhedral plates frequently enclosing and replacing pyroxene. Poikilitic inclusions of apatite, potash feldspar and plagioclase are common. Decomposition yields a finely divided aggregate of iron ore, sphene, calcite, chlorite, epidote and a little green biotite.

**Garnet** is rare; it forms pinkish-brown euhedra, which are isotropic and very weakly zoned. Inclusions and alteration are absent. The mineral has been analyzed (Part III) and is titaniferous andradite-grossular, the type of garnet usually found in foyaitic and phonolitic rocks.

The rare grains of olivine occur only as cores to grains of
Figure 54. Shonkinite, Tilba Tilba. Note grains of olivine, biotite, zoned pyroxene. Ordinary light, 5 diameters.
clinopyroxene. The olivine is colorless, commonly fringed with granules of magnetite, and has a high (>80°) negative optical axial angle which indicates a forsteritic composition. Similar olivine is found in the olivine monzonite and in the shonkinite at Tilba Tilba. The magnesian composition of the olivines contrasts strongly with the alkaline, and even oversaturated, nature of the rocks now enclosing them.

(b) Shonkinite.

This rock crops out at Tilba Tilba village and has been described by Brown (1930, p.654) as olivine shonkinite. Unfortunately its contacts are obscured by soil. Hand specimens are very dark grey and medium- to fine-grained. Pyroxene, biotite and alkali feldspar, some of which may appear opalescent, are the major constituents. Microscopically, the rock has a hypidiomorphic-granular texture and is composed of subhedral interlocking stumpy grains of alkali feldspar in which are set prismatic crystals of clinopyroxene, ragged subhedra of biotite, rare grains of olivine usually enclosed in biotite, small equant crystals of plagioclase, and abundant small grains of opaque minerals and apatite (Figure 54). Brown (p.655) has recorded nepheline, anorthoclase and microcline, none of which have been observed in the sections observed by the present writer.

The salitic clinopyroxene is pale green, weakly zoned, and has inclusions of opaque minerals, apatite and a little plagioclase. Composite grains are very common.

The alkali feldspar is rarely perthitic and forms clear, roughly rectangular plates which have rather undulose extinction. The cleavages
are poorly developed and curved cross-fractures are more often observed. The plagioclase (about An\textsuperscript{45}) has been largely replaced by the alkali feldspar, leaving corroded relics and ghosts. In several cases almost total resorption of plagioclase has produced pseudo-perthitic textures in the alkali feldspar. The optic axial angle of the alkali feldspar is quite variable and ranges from about 50° to over 70°, and the mineral is probably orthoclase cryptoperthite. The blue opalescence of some grains may be caused by the abundant tiny inclusions which lie in irregular planes within the orthoclase.

The biotite is closely associated with the other mafic minerals, especially the opaques (ilmenite-magnetite intergrowths, plus separate grains of each mineral).

The olivine has an optic axial angle of \(-86°\), indicating a composition of about \(\text{Fa}_{30}\).

The shonkinite is texturally and mineralogically very similar to the monzonite (cf. Figures 31 and 54) and would grade into it if most of the olivine and some pyroxene of the shonkinite were replaced by alkali feldspar and acid plagioclase. The appearance of nepheline in the norm of the shonkinite (Brown, 1930, p.656; RB350, Table 30) is probably a consequence of the appreciable amount of biotite in the rock. A similar situation occurs in the case of the monzonites, where a small amount of nepheline appears in the norm (Brown, 1930, p.651; RB221, Table 30) and the rock actually contains a small amount of quartz.
Figure 55. Nepheline monzonite, south-eastern slopes of Little Dromedary. Stained for nepheline. Pyroxene shows mid-grey; nepheline is darker and occurs in plates and in very irregular grains. Darkest mineral is amphibole. Ordinary light, 5 diameters.
(c) Nepheline monzonite.

The nepheline monzonite crops out around Little Dromedary, and it varies considerably in composition from place to place. Several masses of mafic rock, a few of which are garnet-bearing, appear to be inclusions caught up in it, but their contacts are obscured by undergrowth or pasture.

The more typical rock is light grey and medium grained with the mafic minerals, chiefly pyroxene and amphibole, occurring in prismatic grains about 1 x 1 x 3 mm set in a felsic aggregate of similar grainsize and composed of alkali feldspar, plagioclase and nepheline. More mafic types are somewhat finer grained and have a dioritic appearance, with felsic patches 3–4 mm in diameter scattered through the rock. Under the microscope the texture in general is hypidiomorphic-granular, but in many thin sections the texture is monzonitic.

The chief minerals are alkali feldspar, slightly zoned plagioclase (mean composition An_{44}), salitic clinopyroxene, green amphibole (hastingsitic), nepheline, and an isotropic feldspathoid, some at least of which is sodalite, the remainder probably being analcîte. The remainder of the rock is made up of small amounts of forsteritic olivine, biotite, calcite, cancrinite, garnet, sphene, epidote, opaque minerals, apatite and various alteration products of the feldspathoids (Figure 55).

The plagioclase was the first felsic mineral to crystallize and it formed euhedral prismatic grains which show lamellar twinning and weak progressive-normal zoning. Alkali feldspar and nepheline followed, and their textures indicate more or less simultaneous crystallization except in the fine grained marginal phase where nepheline typically
occupies the interstices between the feldspar prisms. The plagioclase is partially replaced by the alkali feldspar and the nepheline, and commonly has narrow albitic rims.

The alkali feldspar in the marginal rock occurs in clear tablets which rarely show cleavage but are commonly cross-fractured. Alteration is absent and there is little or no perthite. The optic axial angle is rather small (mean $2V = 54^\circ$) and the mineral may be transitional between the sanidine- and orthoclase-cryptoperthites. It has very strongly replaced the earlier plagioclase and is in turn somewhat embayed by the later nepheline. In the coarser rocks toward the interior of the intrusion, the alkali feldspar is mildly kaolinized and is sporadically perthitic. The optic axial angle of both the exsolved and apparently homogeneous alkali feldspar is the same ($61^\circ-65^\circ$).

Nepheline has an interstitial form in the marginal rock but also forms rounded or prismatic grains in the coarser types. It is usually very fresh, moderately well cleaved, and in general only slightly altered. In some rocks the place of nepheline is taken by sodalite/analcite (wholly or in part). The nepheline alters to thomsonite, cancrinite and fibrous patches of micaceous material (gieseckite). It is commonly rimmed by albite, especially in those rocks which show the effects of deuteric activity.

The clinopyroxene is very pale green with irregular patches of darker green, usually but not always toward the edges of the prismatic grains. Inclusions of opaque minerals and apatite are common. Marginal replacement of the pyroxene by brownish-green amphibole is widespread (Figure 74).
TABLE 11. Modes of nepheline monzonite.

<table>
<thead>
<tr>
<th></th>
<th>RB240</th>
<th>RB241</th>
<th>RB424</th>
<th>RB427</th>
<th>RB429</th>
<th>RB434*</th>
<th>RB435</th>
<th>RB438a<em>RB438b</em></th>
</tr>
</thead>
<tbody>
<tr>
<td>Orthoclase</td>
<td>38.7</td>
<td>26.0</td>
<td>41.2</td>
<td>44.1</td>
<td>39.5</td>
<td>42.7</td>
<td>41.1</td>
<td>32.0</td>
</tr>
<tr>
<td>Plagioclase</td>
<td>17.1</td>
<td>18.3</td>
<td>24.1</td>
<td>27.4</td>
<td>28.9</td>
<td>17.6</td>
<td>17.7</td>
<td>21.5</td>
</tr>
<tr>
<td>Nepheline</td>
<td>17.8</td>
<td>20.4</td>
<td>12.3</td>
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<td>8.8</td>
<td>5.4</td>
<td>14.2</td>
<td>23.3</td>
</tr>
<tr>
<td>Sodalite</td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td>0.1</td>
<td>1.0</td>
</tr>
<tr>
<td>Analcite</td>
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<td></td>
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<td>1.2</td>
</tr>
<tr>
<td>Cancrinite</td>
<td></td>
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<td>1.0</td>
</tr>
<tr>
<td>Pyroxene</td>
<td>11.6</td>
<td>16.4</td>
<td>5.3</td>
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<td>5.0</td>
<td>8.1</td>
<td>7.6</td>
<td>4.5</td>
</tr>
<tr>
<td>Amphibole</td>
<td>9.8</td>
<td>15.2</td>
<td>11.5</td>
<td>13.5</td>
<td>15.8</td>
<td>22.7</td>
<td>14.6</td>
<td>17.6</td>
</tr>
<tr>
<td>Biotite</td>
<td>3.4</td>
<td>1.1</td>
<td>3.8</td>
<td>0.2</td>
<td>0.4</td>
<td>2.3</td>
<td>1.0</td>
<td>0.9</td>
</tr>
<tr>
<td>Opaques</td>
<td>2.3</td>
<td>2.6</td>
<td>1.2</td>
<td>0.3</td>
<td>&lt;0.1</td>
<td>0.8</td>
<td>1.8</td>
<td>0.1</td>
</tr>
<tr>
<td>Sphene</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>&lt;0.1</td>
<td>0.1</td>
<td>0.3</td>
</tr>
<tr>
<td>Apatite</td>
<td>0.3</td>
<td>&lt;0.1</td>
<td>0.6</td>
<td>0.3</td>
<td>&lt;0.1</td>
<td>0.3</td>
<td>0.8</td>
<td>0.1</td>
</tr>
</tbody>
</table>

'Epidote = 0.1.

*Nepheline-poor and nepheline-rich halves of one thin section.
The amphibole occurs as anhedral plates which are more or less poikilitic in form. The minerals most commonly enclosed by the amphibole are pyroxene, plagioclase, biotite, apatite and opaque grains. The amphibole shows minor variations in depth of color and optic orientation. Alteration is usually absent, but some grains have been transformed to an aggregate of green biotite (as opposed to the olive-brown biotite which is largely replaced by the amphibole) and calcite.

Garnet is found in very small amounts and appears to be associated with the breakdown of amphibole and pyroxene, the albitization of the plagioclase and nepheline, and the deposition of zeolitic infillings (? scolecite). The garnet is pale yellow-brown, distinctly birefringent, and its usual mineral associates are calcite and plagioclase.

Table 11 gives the modes of several specimens of nepheline monzonite collected over an area about 10 feet square (3 m square) on the south-eastern slope of Little Dromedary. The site is the same as that of the specimen of covite analyzed by Brown (1930, p.659). Inspection of the modes shows that the amount of plagioclase does not fall below 17 vol. %, a figure in fair agreement with the appropriate norm (Table 30). The amount of normative anorthite in the so-called covite (Brown, p.659) is comparable with the normative orthoclase (i.e. 29 and 23 wt % respectively) and is in great excess over normative albite (1 wt %) in view of the actual composition of the plagioclase (basic andesine). It is very likely that the figure for $\text{Al}_2\text{O}_3$ is in error owing to some silica passing through into $\text{R}_2\text{O}_3$. If this were so then the normative amounts of anorthite and
nepheline would decrease, and diopside and albite would increase. In any event the figures given by Brown and the modes and analytical data provided herein make it clear that the so-called covite is incorrectly named, since Washington states that the type covite lacks plagioclase (1901, pp. 613, 614, 615). With regard to the modes of the nepheline monzonite (Table 11) it should be noted that all specimens were stained with methylene blue, using the method of Shand (1938), to differentiate the feldspathoids.

**Aplitic veins** are not common in the nepheline monzonite and only three specimens were collected, each from veins about 3 cm wide. Two of these are petrographically quite distinct, as is the rock locally adjoining them.

The texture of the rock enclosing the vein in RB123 is sub-idiomorphic-granular; the rock is made up of subhedral crystals of plagioclase (andesine) and equant anhedral plates of alkali feldspar, some of which are weakly perthitic, minor sub-ophitic hastingsitic amphibole and a few prisms of colorless and pale green clinopyroxene. Inclusions of apatite are common and there are a few grains of calcite, opaque minerals, green biotite, sphene, a few irregular patches of cancrinite, and a little nepheline.

The vein is medium grained and pale pink. The texture is allotriomorphic-granular, and the rock is virtually monomineralic, being made up almost wholly of equant grains of vein perthite, with small interstitial plates of water-clear albite and calcite with subhedral grains of apatite, sphene, and a clinopyroxene with:
\[ \begin{align*}
X &= \text{bright green} \\
Y &= \text{bright green} \\
Z &= \text{yellow-brown.}
\end{align*} \]

with subordinate amounts of plagioclase (usually in the form of lobate relics and intergrowths with nepheline) and clear subhedral crystals of alkali feldspar. The nepheline is very slightly altered to equant plates and arboreal grain boundary anhedra of cancrinite, and occasional stumpy grains of muscovite.

The mafic minerals include ragged subhedra of a colorless clino-
pyroxene which contains abundant inclusions of apatite and opaque minerals, a brown amphibole with fairly weak pleochroism, and a small amount of a pale blue soda amphibole and some green biotite. This is undoubtedly a very alkaline rock but the mafic minerals are much less affected in their composition than are those in much less alkaline phases of the nepheline monzoniite.

The vein has an extraordinarily alkalic composition compared to RB123 above. The texture of this medium grained rock is hypidiomorphic-granular; in form, most of the minerals are tabular or lath-like with the exception of nepheline and the isotropic feldspathoid, which tend to form more equant grains. The rock has not been stained for either feldspathoid or potash feldspar and since it is difficult to distinguish readily between untwinned microcline, nearly pure orthoclase, albite-oligoclase and small laths of nepheline, the following mode should be considered only as an approximation.
IC-feldspar

Albite + oligoclase

Nepheline

Nodalite/analcite

Canorinite

Pale yellow-brown garnet

Muscovite

Olive-brown & green biotites

Sphene

Opaques

Calcite

Apatite

Vol. %

40.5

30.5

19.0

4.5

4.5

1.0

As in the body of the nepheline monzonite, both nepheline and potash feldspar have attacked the oligoclase. A few grains of the potash feldspar show incipient microclinization; this has not been observed by the present writer in any other of the thin sections of rocks of the monzonitic complex.

(5) Garnet-Pyroxene Rocks and Associated Pyroxenites

The rocks discussed below crop out immediately to the north of the porphyritic banatite at Little Dromedary. The field relations of these rocks to each other and to the nepheline monzonite which surrounds them are unfortunately obscured by cultivation. There are a number of small bodies, probably xenoliths carried up from depth, cropping out on the northern slope of Little Dromedary, but the largest and most important body occurs about one mile east of Central Tilba, a little further north
than the small masses. The diameter of this outcrop is a little over half a mile (1 km) and it is not known whether the rock is part of a prior intrusion, sensibly in situ but engulfed in the monzonitic magma, or whether it is simply a large block brought up from depth by the monzonitic magma. This question does not bear directly on the present account and is not considered further, except to note that the nepheline monzonite is the younger igneous body and encloses the pyroxenite. The origin of the ultramafic rock is discussed in Part IV.

Three petrographic units can be distinguished, although all three may be represented in the same small outcrop:

(a) Unaltered pyroxenite, and those altered rocks not within the zone of intense garnetization.

(b) Garnet-pyroxene rocks.

(c) Mafic hybrids.

(a) **Unaltered pyroxenite.**

Unaltered or weakly altered pyroxenite forms a large portion of the rocks under examination. The rock which shows the effects of metasomatism or recrystallization, usually by the development of garnet, makes up perhaps a quarter of the total outcrop of the ultramafic rock.

The unaltered pyroxenites are coarse grained, apparently massive rocks, quite dense and usually dark green in color. Under the microscope they appear to have a coarsely granular texture with equant subhedral to euhedral crystals of pale green salitic clinopyroxene set in a sparse matrix of labradorite or bytownite, with varying amounts of amphibole, biotite, apatite, opaque minerals and forsteritic olivine. The pyroxenites
are texturally, mineralogically and chemically very similar to the massive pyroxenites at Poole's Point. There are small but distinct differences from one specimen to the next with regard to grain size, content of opaques, etc., just as there are at Poole's Point.

Alteration of the pyroxenite from place to place is sporadic, varies in intensity, and has produced various alteration products. The original texture of the rock is never lost, even though alteration may be very extensive, provided that garnet or epidote minerals or scapolite are not abundant.

The earliest change suffered by the pyroxenites results in sericitization of the plagioclase, deposition of abundant prisms of apatite, and the development of a patchy, pale green coloration in the clinopyroxene. At this point, volume changes in the rock produce severe cracking and shattering, especially in the pyroxene.

This is followed by the patchy replacement of pyroxene by a brownish-green biotite, which is usually followed shortly after by further patchy replacement of pyroxene by a yellow-green amphibole and also partial replacement of the biotite by the amphibole. Part of the opaques are consumed in this process, and grains of sphene and calcite are deposited between the pyroxene relicts.

<table>
<thead>
<tr>
<th>Amphibole:</th>
<th>Biotite:</th>
</tr>
</thead>
<tbody>
<tr>
<td>$X = \text{yellow}$</td>
<td>$X = \text{yellow}$</td>
</tr>
<tr>
<td>$Y = \text{olive brown}$</td>
<td>$Y = \text{greenish-yellow}$</td>
</tr>
<tr>
<td>$Z = \text{olive brown}$</td>
<td>$Z = \text{olive brown}$</td>
</tr>
</tbody>
</table>

$X < Y < Z$, moderate. $X \leq Y < Z$, strong.

Replacement of this type is quite advanced in some specimens,
Figure 56. Hydrothermal vein cutting pyroxenite, 1 mile east of Central Tilba. Grains of garnet marked 'G'. Ordinary light, 5 diameters.
the amphibole becoming greenish-blue in color and probably more sodic. The pyroxenite is then made up of pyroxene grains riddled and largely replaced by decussate aggregates of green amphibole, a little relict biotite, sporadic grains of sphene, and even a considerable amount of pennine. Garnet is absent (e.g. RB131, RB138).

Other types (such as RB135) have numerous small patches of biotite and amphibole sieving the pyroxene, while the original plagioclase has been completely replaced by scapolite. Garnet is also absent from these rocks.

In RB140, a pale brown garnet appears in small, rounded grains and in the interstices between the crystals of pyroxene, which are heavily flecked with deep green amphibole. The garnet is optically anisotropic.

Specimens RB141 (Figure 56) and RB143 are samples of garnet-bearing veins. The most abundant minerals in the veins are scapolite, a fibrous mass of muscovitic material, birefringent garnet (2V = 86°; mimetic after plagioclase), thomsonite, calcite and apatite, epidote, zoisite, pennine and sphene.

(b) Garnet-pyroxene rocks.

(i) Field occurrence. The chief exposure of these rocks is on a low hill about a mile north of the summit of Little Dromedary and a mile east of Central Tilba. The small bodies on the northern slopes of Little Dromedary are garnet-bearing in part; since they show very much the same features and reflect the same physico-chemical processes, they are included in the discussion of the main exposure. This is the locality from which Brown (1930, pp. 667-673) described (?)kedabekite-melanite-
melteigite-jacupirangite. The garnetized rock crops out slightly to the south of the centre of the body of pyroxenite and occupies a relatively small fraction of the exposure of ultramafic rock. The erratic distribution of garnet is easily noticed in the field, there being an intimate association of garnet-rich and garnet-poor phases. Veins are rare, and the mass is cut by a few narrow, fine grained, trachytic dykes.

(ii) Internal structures. Banding picked out by garnet-rich and garnet-poor fractions, or by local changes in the grainsize, chiefly of garnet, is clearly visible on the faces of some blocks of the rock. The breadth of these very straight bands varies from one to another but is usually about 3 mm, and each is separated from its neighbor by a band of garnet-poor rock about 10 mm wide. Unfortunately, there are only a few blocks of sufficient size exposed which could be examined for layered structures; mesoscopically the rock in general appears to be fairly massive. There is no doubt, however, that some of the garnet-pyroxene rock has a regular, planar structure (cf. Brown, 1930, p.667). Examination of some hand specimens (RB366) has revealed strong parallelism of the [c] crystallographic axes of the clinopyroxenes. This preferred orientation is a relict structure and has clearly developed prior to, and independent of, the garnetization of the rock. The conventional interpretation of mineral alignment in igneous rocks is that flow processes operated on the rock matter, the lineation being parallel to the direction of greatest local elongation. The layered structure of some of the garnet-pyroxene rock mentioned above may be the result of mimetic recrystallization of previously layered pyroxenite. Alternatively, since
the mineral most clearly marking out the layering is of metamorphic origin (garnet), the structure may have arisen through metamorphic differentiation not related to initial inhomogeneities in the pyroxenite.

(iii) Petrography. Hand specimens vary in color from medium grey-green to dark brownish-green depending on the content of garnet and felsic constituents. Weathered surfaces have a rusty, nodular appearance because of the slower decomposition of the garnet in the rock compared to the other components. Small vugs are scattered through the rock and for the most part are barren. Some have infillings of fibrous, stellate or massive hydrous lime-silicates, and some have allowed the garnet to develop free crystal faces. The garnet-pyroxene rock is very dense, and appears to be medium to fine grained. The rock has a sub-conchoidal fracture and fresh surfaces have a resinous lustre. Cleavage reflections from the clinopyroxene are often very dull because of the abundant tiny grains of garnet which the pyroxene contains.

The microscopical texture of the rock is best described as hypidiomorphic-granular, modified in greater or lesser degree by the development of porphyroblasts of garnet, xenoblastic plates of alkali feldspar and plagioclase, and idioblastic grains of apatite. The texture of the garnet-rich rocks is poikiloblastic, with extensive intergrowth of relict clinopyroxene with feldspar and garnet. Brown (1930, p.667) attributes these textures to overlapping primary magmatic crystallization of the pyroxene and garnet.

The salitic clinopyroxene closely resembles that of the unaltered
and slightly altered pyroxenites. It is pale green, weakly pleochroic, and shows good prismatic cleavage. Inclusions of opaque minerals, amphibole, biotite and sphene are rare.

The intensity of the pale green color of the pyroxene varies between rocks, between grains, and between portions of grains, in a way not obviously related to the presence, absence or proximity of the other minerals. A small proportion of the clinopyroxene is virtually colorless, while a few grains appear to be very faintly pink.

The well-formed prisms of pyroxene in the garnet-poor rocks measure about 1.0 x 1.0 x 2.5 mm, with some crystals up to 5.0 mm in length; they are sieved with numerous small grains of feldspar, garnet and apatite, and have a poikiloblastic appearance. As replacement of the pyroxene progresses, the relict grains become more and more spongy, until finally all that remains of the relatively large original crystal is a group of small separate grains which have the same optic and crystallographic orientation. Replacement of this kind has been more extensive in the feldspar-rich rocks.

Garnet is usually abundant and occurs as irregular porphyroblasts, commonly up to 2 cm in diameter, and as numerous equant grains sieving the entire rock. Crystal faces are only poorly developed. The color varies from light yellowish-brown to pinkish-brown and deep pink, and shows very patchy variations. Inclusions of pyroxene, apatite, calcite and feldspar are common, together with numerous tiny opaque grains. The bulk of the mineral is very weakly birefringent. A sample has been separated and analyzed (see Part III) and the garnet is a titaniferous andradite-grossular.
The third constituent which occurs in appreciable amounts is feldspar which, together with the garnet and pyroxene, makes up 95 vol. % or more of these rocks. Both plagioclase and alkali feldspar are present in nearly all the thin sections examined, although the proportions and total amount of the two minerals varies considerably.

The plagioclase feldspar in the garnet-pyroxene rocks differs from that of the relatively unaltered pyroxenite in that it is not found as plates interstitial to the pyroxene crystals, and is less calcic in composition. The typical mode of occurrence of the plagioclase is as xenoblastic plates up to 5 mm in diameter (average about 0.7 mm) forming a blastophitic intergrowth with the relict pyroxene. The plagioclase very often appears as a multitude of small, optically continuous, lobate grains sieving the pyroxene, and as more extensive plates which lie between the crystals of garnet and clinopyroxene. Some relict grains of the original calcic plagioclase (labradorite-bytownite) persist in some specimens, and are distinguished by their higher refractive indices, their birefringence colors (cream, compared to middle greys) and their intimate veining by the later plagioclase. Both types of plagioclase show very weak zoning and lamellar twinning, frequently with one set of lamellae broad and the remaining units extremely narrow. Pericline twinning is present in some grains and on occasion may be dominant over the complex and parallel types. Cleavage is generally quite poorly developed. Much of the plagioclase in those rocks with alkali feldspar appears to be, and very often is, untwinned, and is easily confused with the alkali feldspar.

Measurements of the β index of a few grains of plagioclase from each section indicated a small range in composition about a mean of An 40.
(middle andesine) and the zone method of Rittmann (1929; in Emmons, 1943) was then employed on the twinned grains to estimate the composition more accurately (see Barber, 1936, p.281). In the cases where one of the sets of lamellae are very narrow, some corroborative determinations were made, following the method of Turner (1947, p.408). The mean range in composition of the andesine was from $\text{An}_{44}$ (cores) to $\text{An}_{36}$ (rims).

The alkali feldspar usually predominates over the andesine, and has the same mode of occurrence. It appears to be later than the andesine, which it has substantially replaced in many cases (RB382, RB389), thus tending to produce a simple pyroxene-alkali feldspar assemblage. The grains are equant and usually xenoblastic with sutured boundaries against like grains. Simple twins have been observed in some sections. The alkali feldspar is very weakly perthitic and is slightly kaolinized. The extinction of most grains is slightly undulose. The optic axial angles are fairly constant, with a mean $2V$ of $65.5^\circ$.

The most common accessory mineral is apatite, which is found in idioblastic lath-like prisms scattered through all the other constituents of the rocks. The crystals tend to concentrate in groups. There is a slight tendency for the apatite to associate with the relict clinopyroxene, especially in the more feldspathic rocks, and trails of crystals have been observed in or near well defined channels along which occurred replacement of calcic plagioclase by andesine and of both of these by alkali feldspar (RB389).

The apatite is colorless, but may contain tiny, zonally arranged inclusions. The $(0001)$ cleavage is weak and some grains have a very small
extinction angle (1°-2°), and a small, positive optic axial angle.

Scapolite has been noted in some specimens; the mineral occurs in the same fashion as the plagioclase, which has partly replaced it. Nepheline has not been observed by the present author although a small amount was reported by Brown (1930, p.670). Opaque minerals are rare. There are some grains of weakly pleochroic, pinkish-brown sphene which are often difficult to distinguish from the garnet, especially in plane light. Traces of emerald-green amphibole occur as alteration products of pyroxene in many specimens. Other minerals found in trace amounts are idocrase, calcite, clinzoisite ± epidote, biotite, prehnite, and a fibrous zeolite (thomsonite?).

Thirty-five hand specimens were taken over an area of about 60 square feet (5.6 sq. metres) and their modes were determined (Table 13) in order to assess the homogeneity of the rock and the mineral paragenesis. A sketch map was made of the outcrop and the site of each sample, and the variation in the rock of each major component was examined by constructing mode 'contours'. On the data available, it appeared that the distribution of minerals was not systematic.

The proportions of garnet, feldspar and clinopyroxene were determined for each specimen in Table 13 and are plotted in Figure 58. The trend of the modes is fairly well defined and represents a large variation in the proportion of the major minerals. The broken line represents the upper limit of pyroxene content, and is also parallel to the trend in composition.

The chemical analyses of a very weakly altered pyroxenite, a garnet-
Figure 57. Garnet-pyroxene rock, with virtually no garnet but abundant alkali feldspar. Pyroxene (darker grey) occurs in very irregular relict grains, optically continuous over several millimetres. Ordinary light, 5 diameters.
### TABLE 12. Analyses of rocks and minerals related to garnet-pyroxene rock.

<table>
<thead>
<tr>
<th></th>
<th>RB338</th>
<th>A</th>
<th>B</th>
<th>RB363AG</th>
<th>7</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>45.00</td>
<td>44.28</td>
<td>40.26</td>
<td>36.04</td>
<td>51.80</td>
</tr>
<tr>
<td>TiO₂</td>
<td>1.08</td>
<td>1.28</td>
<td>1.68</td>
<td>1.34</td>
<td>0.24</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>8.71</td>
<td>10.13</td>
<td>10.95</td>
<td>7.19</td>
<td>1.84</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>5.46</td>
<td>4.04</td>
<td>12.16</td>
<td>20.35</td>
<td>1.82</td>
</tr>
<tr>
<td>FeO</td>
<td>6.74</td>
<td>6.21</td>
<td>3.38</td>
<td>1.31</td>
<td>7.45</td>
</tr>
<tr>
<td>MnO</td>
<td>1.14</td>
<td>0.23</td>
<td>-</td>
<td>0.30</td>
<td>0.47</td>
</tr>
<tr>
<td>MgO</td>
<td>10.99</td>
<td>8.42</td>
<td>2.16</td>
<td>0.50</td>
<td>12.38</td>
</tr>
<tr>
<td>CaO</td>
<td>17.79</td>
<td>21.86</td>
<td>26.30</td>
<td>33.27</td>
<td>22.85</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.96</td>
<td>0.31</td>
<td>0.44</td>
<td>0.11</td>
<td>0.61</td>
</tr>
<tr>
<td>K₂O</td>
<td>1.08</td>
<td>0.69</td>
<td>1.33</td>
<td>0.05</td>
<td>0.03</td>
</tr>
<tr>
<td>H₂O⁺</td>
<td>0.50</td>
<td>1.29</td>
<td>0.64</td>
<td>n.d.</td>
<td>0.26</td>
</tr>
<tr>
<td>H₂O⁻</td>
<td>-</td>
<td>0.29</td>
<td>0.28</td>
<td>n.d.</td>
<td>-</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.65</td>
<td>1.35</td>
<td>0.89</td>
<td>n.d.</td>
<td>0.20</td>
</tr>
<tr>
<td>CO₂</td>
<td>0.12</td>
<td>-</td>
<td>-</td>
<td>n.d.</td>
<td>-</td>
</tr>
<tr>
<td>Total</td>
<td>100.22</td>
<td>100.38</td>
<td>100.47</td>
<td>100.46</td>
<td>99.95</td>
</tr>
</tbody>
</table>

Total iron as FeO 11.66 9.85 14.31 19.61 9.09

*Absence of MnO highly unlikely; see RB363AG, 7.*

**RB338** — Very slightly altered pyroxenite, 3/4 mile east of Central Tilba.
Analyst R.S. Boesen.


**RB363AG** — Garnet from same locality as B. Analysts Avery and Anderson.

**7** — Pyroxene from same locality as B, coexisting with RB363AG. Analyst R.S. Boesen.
poor and a garnet-rich type are given in Table 12, together with the analyses of the garnet and clinopyroxene of these rocks.

The following observations may be noted regarding the garnet-pyroxene rocks:

(i) The texture of the rocks is dominated by extensive replacement intergrowths, chiefly between garnet and pyroxene in the mafic types and, to a lesser extent, between alkali feldspar and andesine and pyroxene (and perhaps garnet) in the more felsic rocks.

(ii) The replacement textures are superimposed on a prior hypidiomorphic-granular texture (with or without preferred orientation).

(iii) The texture and composition of the rocks varies widely over very small distances.

(iv) The garnet-pyroxene rocks are intimately associated in the field with pyroxenite on the one hand, and with nepheline monzonites on the other.

(v) The abundance of apatite, and the presence of minerals such as calcite, scapolite, alkali feldspars and some zeolites indicates the widespread activity of alkaline, volatile-rich fluids. The association of andesine and perthitic alkali feldspar, and the absence of alteration of the garnet or pyroxene to amphibole, mica, chlorite, etc., suggest a moderate to high temperature environment.

(vi) The replacement of pyroxene and calcic plagioclase by garnet is not isochemical, and since there is a considerable difference in the densities of the weakly and strongly altered rocks, considerable quantities of material have been introduced and removed. Examination of
### TABLE 13. Modes of garnet-pyroxene rocks

<table>
<thead>
<tr>
<th>Sample</th>
<th>Garnet</th>
<th>Pyroxene</th>
<th>Feldspar</th>
<th>Amphibole</th>
<th>Calcite</th>
<th>Others</th>
</tr>
</thead>
<tbody>
<tr>
<td>EB370</td>
<td>53.5</td>
<td>46.2</td>
<td>30.6</td>
<td>0.3</td>
<td>0.2</td>
<td></td>
</tr>
<tr>
<td>EB371</td>
<td>45.4</td>
<td>38.9</td>
<td>42.4</td>
<td>0.5</td>
<td>0.3</td>
<td></td>
</tr>
<tr>
<td>EB372</td>
<td>48.2</td>
<td>32.2</td>
<td>34.2</td>
<td>0.5</td>
<td>0.3</td>
<td></td>
</tr>
<tr>
<td>EB373</td>
<td>43.5</td>
<td>31.8</td>
<td>43.5</td>
<td>0.5</td>
<td>0.3</td>
<td></td>
</tr>
<tr>
<td>EB374</td>
<td>38.6</td>
<td>38.4</td>
<td>41.5</td>
<td>0.5</td>
<td>0.3</td>
<td></td>
</tr>
<tr>
<td>EB375</td>
<td>38.8</td>
<td>34.5</td>
<td>51.5</td>
<td>0.5</td>
<td>0.3</td>
<td></td>
</tr>
<tr>
<td>EB376</td>
<td>39.1</td>
<td>32.2</td>
<td>43.5</td>
<td>0.5</td>
<td>0.3</td>
<td></td>
</tr>
<tr>
<td>EB377</td>
<td>38.6</td>
<td>32.2</td>
<td>43.5</td>
<td>0.5</td>
<td>0.3</td>
<td></td>
</tr>
<tr>
<td>EB378</td>
<td>41.3</td>
<td>30.6</td>
<td>54.5</td>
<td>0.3</td>
<td>0.2</td>
<td></td>
</tr>
<tr>
<td>EB379</td>
<td>40.6</td>
<td>58.6</td>
<td>51.5</td>
<td>0.3</td>
<td>0.2</td>
<td></td>
</tr>
<tr>
<td>EB380</td>
<td>31.5</td>
<td>33.9</td>
<td>33.5</td>
<td>0.3</td>
<td>0.2</td>
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<tr>
<td>EB381</td>
<td>24.0</td>
<td>21.2</td>
<td>21.2</td>
<td>0.3</td>
<td>0.2</td>
<td></td>
</tr>
<tr>
<td>EB382</td>
<td>3.4</td>
<td>0.3</td>
<td>0.3</td>
<td>0.3</td>
<td>0.2</td>
<td></td>
</tr>
<tr>
<td>EB383</td>
<td>3.2</td>
<td>3.6</td>
<td>3.6</td>
<td>3.6</td>
<td>3.6</td>
<td></td>
</tr>
<tr>
<td>EB384</td>
<td>3.2</td>
<td>3.6</td>
<td>3.6</td>
<td>3.6</td>
<td>3.6</td>
<td></td>
</tr>
</tbody>
</table>

* Modes by G. A. Joplin.
Figure 58. Modal garnet (G), feldspar (F) and clinopyroxene (Cpx) in the garnet-pyroxene rocks RB361-RB395. V is a garnet-feldspar vein (RB387).
Table 12 shows that large amounts of MgO and smaller amounts of silica are removed during reconstitution of the rocks, while there is addition of CaO, Al₂O₃, alkalies and volatiles, and perhaps TiO₂ and iron. Extensive oxidation of ferrous iron is also effected.

(c) Mafic hybrids.

These rocks are restricted in their occurrence to the north-eastern slopes of Little Dromedary. They are surrounded by nepheline monzonite, and crop out within fifty yards of the porphyritic banatite-nepheline monzonite contact. The grain size varies between and within the masses, some specimens being fine to medium grained (RB276) and others medium to coarse grained (RB238); they range in color from grey to dark green to nearly black. The constituent minerals show preferred orientation in hand specimens. The habit of the crystals in RB276 is more elongate than that of the coarser grained RB238 and the lineation is better defined. The rock has a laminar structure similar to that of the laminated feldspathic pyroxenite at Poole's Point, except that the differentiation of the mafic and felsic minerals into separate laminae is not so well developed. Recrystallized platy xenoliths up to 10 cm in diameter have been observed in the rock and are oriented parallel to the plane of lamination.

The platy structure of RB238 is not very strong, partly because the rock is poor in minerals with a platy habit, but the linear parallelism of the pyroxene prisms is quite clear. In the hand the rock has a distinctively dioritic appearance, being made up of black subhedral to euhedral tabular crystals of pyroxene and small black flakes of biotite, set in
a medium grained granular matrix of pearly and translucent light-colored minerals, dominantly alkali feldspar. In thin section, RB238 has a hypidiomorphic-granular texture very similar to many of the massive pyroxenites, and some of the garnet-pyroxene rocks carrying appreciable amounts of alkali feldspar.

The most abundant mineral in RB238 is clinopyroxene, which occurs in stumpy euhedra which measure about 1.0 x 1.5 x 2.0 mm. They are very weakly colored, usually pale green, but some grains are pale yellow and others very pale pink. Pleochroism is virtually absent in most crystals, although some have

\[
\begin{align*}
X &= \text{very pale yellow or pink} \\
Y &= \text{pale greyish-green} \\
Z &= \text{pale green.}
\end{align*}
\]

The margins of most grains have a slightly deeper green tint and the optic orientation and birefringence of the more strongly colored portions differ slightly from those of the cores. The crystals carry abundant small inclusions of opaque minerals, apatite, plagioclase and dark olive biotite, zonally arranged, and weak schiller structure occurs in the cores of some pyroxenes.

Prismatic cleavage is strongly developed, especially in the outer portions of the crystals, and slight alteration along the cleavage planes makes them appear dark. The pyroxenes in the mafic hybrid rocks are characterized by cleavage parallel to (100) and (010), either with or without the more usual (110) set.

The pyroxene grains are set in equant plates of weakly perthitic alkali feldspar and andesine, both of which have nearly completely
replaced a prior, more calcic plagioclase. A small amount of scapolite occurs as interstitial plates between the pyroxene crystals. Abundant prisms of apatite are scattered through the rock, usually associated with the alkali feldspar and frequently forming trails in it. Coarse, opaque polygonal grains have a similar occurrence. Pink, slightly pleochroic sphene is abundant and of very late crystallization; it occurs both as interstitial anhedra, sometimes pakilitically enclosing clusters of apatite crystals, and as lobate grains intergrown with, and apparently replacing, pyroxene. The biotite forms ragged grains and stumpy prisms closely associated with and marginally replacing the pyroxene. It also forms thin rims around the opaque minerals, and appears to be stably associated with the andesine and alkali feldspar. Its pleochroism is very strong, from light yellow-brown to very dark greenish-brown.

Garnet was not observed, but the form and textures of the coarse opaque grains suggest its former occurrence.

RB276 is composed of approximately equal amounts of pleochroic, mid-green pyroxene and laths and rounded crystals of twinned and strongly zoned andesine. Other minerals include amphibole, traces of red-brown biotite, apatite, sphene, alkali feldspar and opaques.

The pyroxene forms lath-like, prismatic and equant euhedra and subhedra which vary considerably in size. The more elongate grains are the largest and average about 1.5 mm in length, with some up to 4 mm; the more granular crystals average about 0.3 mm in diameter. Granular aggregates of pyroxene crystals are common. The dominant color of the pyroxene is greyish-green, and the pleochroism is:
\[
\begin{align*}
X &= \text{pale yellow} \\
Y &= \text{greyish-yellow} \\
Z &= \text{grey-green or slightly blue-green.} \\
X &= Y \leq Z, \text{ moderate.}
\end{align*}
\]

Several grains have colorless, non-pleochroic cores, and zoning by one or more of color, birefringence and inclusions is universal. The pyroxene bears a close resemblance to that in the 'ijolite' (q.v.). The dispersion about the B optic axis is strong and gives rise to bright orange and violet colors near the extinction position.

Like the pyroxene in RB238, cleavage takes place on (110), (100) and (010), although the (110) predominates.

The plagioclase laths and tablets are zoned from acid labradorite to acid andesine and oligoclase. The crystals almost invariably have prominent zones of tiny inclusions and alteration products.

Large, pseudomorph grains of green and blue-green hastingsitic amphibole enclose and vein the pyroxene and most of the coarsely granular opaque minerals. Some of these lacy grains have a diameter of about 15 mm and produce the lustre-mottling seen in the hand specimens. The amphibole has formed chiefly by replacement of pyroxene, biotite, opaques and plagioclase. The rock contains no feldspathoids.

The manner in which the coarse opaque grains are occasionally intergrown with the pyroxene and with sphene, and their common, straight, polygonal outlines against the pyroxene and other minerals in these rocks indicate that the opaques are pseudomorphs after andraditic garnet. This conclusion is supported by the observation in RB276 of a tiny grain of clear red-brown garnet embedded in a small plate of clinozoisite.
Figure 59. Lithic tuff, 1½ miles east of Central Tilba. Ordinary light, 5 diameters.
The late stage of crystallization and the abundance of sphene, especially in RB238, is in accord with its derivation in part by breakdown of titaniferous andradite.

A particularly striking rock (RB485), unfortunately not found in situ, was collected from the same general area as the above samples. It is composed of zoned granular pyroxene, pleochroic in yellows and grey-greens much like that of RB276 but slightly more strongly colored, and abundant deep red-brown subhedral garnet, set in large clear plates of andesine.

The pyroxene and garnet both form independent grains but are more usually associated with one another in a coarse granular intergrowth. Large subhedra of pale pink sphene are also associated with the pyroxene and garnet. Apatite and opaque minerals form numerous small grains, often as inclusions in the pyroxene, but they do not occur in quantity. Feldspathoids have not been observed in this rock.

(6) Andesitic Lava and Tuff

This sequence of rocks crops out to the east of Mt Dromedary, and includes most of the exposed rock lying immediately west of Tilba Tilba Lake. It is difficult to judge the orientation of the volcanic rocks, but they appear to dip gently to the east and they extend to the coast at the mouth of the lake, forming low black cliffs. In the coastal strip a mile north of the cliffs, some of the lamprophyric rocks lie sub-horizontally and may be shallow, sill-like members of the same magma. Brecciation of some of the rocks has been noted by Brown (1930, p.674), and these are probably tuffaceous members. The rock forming the cliffs south of Tilba Tilba Lake has been intruded by a number of dykes and small,
irregular bodies, most of which have monzonitic compositions. These intrusions have produced weak local recrystallization of the lava and tuff.

Hand specimens of the lava are very dark grey, holocrystalline, and have a porphyritic texture with stumpy black laths of amphibole and pyroxene, usually about $1 \times 1 \times 2$ mm, set in a dark aphanitic groundmass. The microscopical texture of the groundmass is pilotaxitic and moderately fluidal, with twinned and weakly zoned andesine and labradorite laths, granular or interstitial alkali feldspar, tiny rounded grains of amphibole, pyroxene, and opaque minerals.

The pyroxene may be colorless, pale green or pale pink, and may or may not be very weakly pleochroic. The optical properties indicate that it is a salitic clinopyroxene. Mild birefringence zoning occurs in the phenocrysts, and the prismatic cleavages are well developed.

The amphibole is a pale green, faintly pleochroic, uralitic variety with:

- $X = \text{very pale yellow}$
- $Y = \text{pale yellowish-green}$
- $Z = \text{light green}$

$X < Y < Z$, weak.

$Z^\circ = 28^\circ$

$2V (-ve) = \text{approx. } 75^\circ$.

It has replaced the pyroxene very slightly in some rocks and almost entirely in others (RB323). Its origin is indicated in the latter by the common preservation of the pyroxene cleavages. The alteration is accompanied by the appearance of small amounts of calcite, quartz and tiny grains of either sphene or epidote. The mode of the analyzed rock (RB323) is:
### Chemical Composition

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Vol. %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Amphibole</td>
<td>23.7</td>
</tr>
<tr>
<td>Plagioclase</td>
<td>49.8</td>
</tr>
<tr>
<td>Alkali feldspar</td>
<td>19.2</td>
</tr>
<tr>
<td>Opaques</td>
<td>5.1</td>
</tr>
<tr>
<td>Apatite, sphene, calcite &amp; quartz</td>
<td>2.2</td>
</tr>
</tbody>
</table>

**RB490** is less altered than RB323, and has phenocrysts of colorless and very pale pink clinopyroxene, partly uralitized, and acid labradorite set in a pilotaxitic groundmass consisting of tabular and lath-like euhedra of zoned andesine-labradorite, anhedral alkali feldspar, abundant tiny opaque grains and irregular granules of sphene. In the groundmass there are several small, clear, polygonal (usually hexagonal) plates which are quite free from the tiny opaque grains; these colorless areas are about 0.1 mm in diameter. Under crossed nicols they appear as decussate aggregates of weakly birefringent, polysynthetically twinned grains. In view of the relatively alkali-rich, potash dominant chemistry of these rocks, these aggregates may be twinned leucite or potash analcite, or feldspar pseudomorphs thereof.

The tuffs are massive, light to dark grey or green rocks made up very largely of fragments of holocrystalline intermediate and basic lavas, some coarser grained monzonitic rock, and fractured crystals of feldspar (Figure 59). The volcanic fragments are usually very fine grained, porphyritic in plagioclase and clinopyroxene, with pilotaxitic and fluidal groundmass textures marked out by very small feldspar laths. Cryptocrystalline groundmasses charged with abundant opaque grains are common. Amphibole and brown biotite occur in ragged flakes between the
clastic grains, and some samples of tuff are slightly epidotized. Neither glass shards nor glassy groundmasses were observed.

Anderson (1892) gives an analysis (by Mingaye) of a hornblende andesite collected a mile west of the entrance to Tilba Tilba Lake. Comparison with the analysis of RB323 (Table 29) which was collected about a mile north of the source of Anderson's sample shows that the two rocks are clearly related. RB323 contains less $SiO_2$, $Al_2O_3$ and alkalies, and more total iron, CaO and MgO, and therefore represents a less differentiated phase.

Brown (1930, pp.675, 676) suggested that the composition of the parent magma of the complex was similar to that of the andesite analyzed by Mingaye. The present writer is in general agreement with this view, but proposes that the magma was more basic than RB323; that is, the parent was a truly basaltic rock with the general characteristics of the analyzed lavas, namely relatively high total alkalies, with $K_2O > Na_2O$, high CaO compared to MgO, and moderately high $Al_2O_3$.

Intrusive and extrusive rocks with these features are found elsewhere in eastern Australia and overseas, notably Yellowstone Park and the Sierra Nevada where they are described as latites and shoshonites. The tectonic environment, composition and differentiation of these rocks are discussed by Joplin (in press), who proposes that they represent a distinct magma-type (shoshonite).
PART III

MINERALOGY

Detailed optical and chemical studies have been made of the
pyroxenes and biotites from the more important rock types, and of these
harzburgites and three garnes. Olivines, feldspars and other minerals
were examined optically only.

The chemical and optical properties were determined using pure
samples of the respective minerals separated from the rock, and in thin
sections. In the case of the feldspars, olivines and opaque minerals,
the optical measurements were made almost entirely in thin and polished
sections and on grains picked out from them.

The techniques of separation and analysis of the minerals were the
same as those used for the rocks, and are described below.

(b) Determinative Techniques

(a) Sample preparation.

All the rocks which were analysed were crushed in a pestle and
mortar to pass a 100 mesh screen and were then very finely ground under
acetone (A.R. grade) by hand in an agate mortar. The minimum size of
the sample was determined by the grainsize of the particular rock. In
analogy with the practice in modal analysis, a volume of rock was taken
which was greater than 100 times that of the largest mineral grain. In
these cases where there was obvious inhomogeneity in the rock, several
samples were taken from a small area, and the resultant large sample was
ground and quartered. This was not done in the case of the sample of
MINERALOGY

Detailed optical and chemical studies have been made of the pyroxenes and biotites from the more important rock types, and of three hornblendes and three garnets. The olivines, feldspars and other minerals were examined optically only.

The chemical and optical properties were determined using pure samples of the respective minerals separated from the rock, and in thin sections. In the case of the feldspars, olivines and opaque minerals, the optical measurements were made almost entirely in thin and polished sections and on grains picked out from them.

The techniques of separation and analysis of the minerals were the same as those used for the rocks, and are described below.

(1) Determinative Techniques

(a) Sample preparation.

All the rocks which were analyzed were crushed in a percussion mortar to pass a 100 mesh screen and were then very finely ground under acetone (A.R. grade) by hand in an agate mortar. The minimum size of the sample was determined by the grain size of the particular rock. By analogy with the practice in modal analysis, a volume of rock was taken which was greater than 100 times that of its largest mineral grain. In those cases where there was obvious inhomogeneity in the rock, several samples were taken from a small area and the resultant large sample was ground and quartered. This was not done in the case of the sample of
pyroxenite from the bottom of the drill hole, and a sample of about 50 gms was chosen which appeared to be representative of the cored rock. In view of the coarse grain of the rock, this analysis may be subject to sampling error.

The minerals were separated from a portion of the same piece (or pieces) of the rock crushed for analysis. The crushed rock was screened at frequent intervals on a closely graded series of sieves. In this way certain minerals would tend to concentrate in one or more of the various fractions, according to their brittleness, lack of cleavage, characteristic form, etc. Both biotite and garnet tend to remain on the coarser screens because of the flaky nature of the biotite and the hardness and lack of cleavage of the garnets.

In general it was found convenient to work with a crush of about 120-150 mesh. The minerals were separated from this fraction using an iso-dynamic magnetic separator and heavy liquids. Some difficulty was experienced in separating pyroxene (especially when strongly zoned) from hornblende, and hornblende from biotite. Each mineral shows a range in magnetic properties and density (and, presumably, chemistry), and overlapping of these ranges with those of other minerals makes their complete separation using the above methods virtually impossible. Accordingly, several samples were finally purified by hand-picking. The samples were then finely ground under acetone, as for the rocks. Nylon and silk screens were used throughout.
(b) **Chemical analyses.**

Prior to analysis, all samples were dried at 105°C for four hours (two hours for the biotites).

Silica was determined gravimetrically in the conventional way, and the amount of silica passing into the filtrate was estimated colorimetrically after development of the molybdenum-blue complex.

Alumina in the minerals was determined gravimetrically after its separation from the $R_2O_3$ precipitate by means of an ion exchange column. The results were compared with the figures for alumina determined before and after the separation, using the colorimetric method described by Riley (1958). The agreement was found to be quite satisfactory, and the colorimetric method alone was used in most of the rock analyses.

Total iron was determined colorimetrically after development of the red dipyridyl complex. Ferrous iron was found by the conventional method of titration with potassium dichromate, using barium diphenylamine sulphonate as indicator.

Titanium, manganese and phosphorus were determined colorimetrically. Both peroxide and Tiron were used for titanium, depending on its level in the sample. Manganese was measured as permanganate, using the method of Riley (1958). The yellow molybdovanadophosphoric acid complex was developed to determine phosphorus.

Lime and magnesia were found using the normal gravimetric methods in all the mineral analyses. The results compared satisfactorily with those obtained from titrimetric techniques using EDTA with calcium indicator for lime, and EDTA with solochrome (eriochrome black) indicator
for lime plus magnesia after precipitation of $R_2O_3$ from an HF dissolution. This method alone was used in the rock analyses except where the titrimetric figure for lime was less than 1.0%, in which case a gravimetric determination was also made. It was found that only at quite low levels of lime did the two results differ significantly.

The alkalies were measured using a Perkin-Elmer flame photometer with lithium as internal standard. Low amounts of alkalies (in the pyroxenes) were measured on a Beckman model DU spectrophotometer fitted with a flame photometric attachment.

Combined water was determined using the modified Penfield method set out in Washington (1930).

The trace elements (including $BaO$) were determined by emission spectrography, using a Jarrell-Ash and Co. Ebert spectrograph equipped with a grating of 15,000 lines per inch. The wave-length range considered was from Boron 2496 Å to Barium 4934 Å. The technique used was that of Kolbe and Ahrens (in press) except that a value of 31.5 p.p.m. was assumed for the nickel standard and the wave-length of the line measured was 3050.82 Å. The standard used for boron was the C.A.A.S. syenite Sy-1, for which a figure of 66 p.p.m. was used.

(c) Specific gravities.

The specific gravities were determined from the loss in weight of a sample on immersion in toluene. The sample varied in size from about 20 mgm for the minerals to 200 gms in some rocks. All mineral densities were determined at 20°C.
(d) **Optical methods.**

The optical properties were determined both from crushes of the analyzed material and from oriented grains picked from thin sections. It was found to be very convenient to make a duplicate section in the normal way and then soak it in a solvent solution so that a clean thin slice of rock could be obtained. Very few slices disintegrate in the process and they can be handled, provided great care is taken. The slices are transferred to a clean glass slide and moistened with oil. Clean, appropriately oriented grains may be then broken out of the slice much more easily than from a normal thin section.

Refractive indices were measured using ordinary immersion techniques, taking care to minimize the effect of temperature variations during the determinations. Oils graduated in refractive index by steps of 0.002 were used. The index being measured was found within the range of two such liquids and the exact match was achieved by mixing some of each oil. The matching oils were immediately measured using a Leitz Abbé-type refractometer (or a Leitz-Jelly instrument for high index oils). Birefringence was found by subtraction of the $\alpha$ and $\gamma$ indices.

Optic axial angles were measured conoscopically (after Hallimond, 1950) by rotation from optic axis to optic axis, and the appropriate corrections for refractive index variations were applied.

All final measurements were made in sodium light. The error in the determinations of indices, optic axial angles and extinction angles increases with the absorption of the mineral in question, but it is probably less than $\pm 0.002$ (indices) and $\pm 0.5^\circ$ (2V) for the pale minerals.
(pyroxenes) and ± 0.008 and ± 2.0° for the darkest hornblends. The error of refractive index of the garnets is probably higher, because of the rapid decomposition of the immersion media.

In the pyroxenes, β indices were measured on (100) parting tablets; α and γ were measured on grains (usually from thin sections) giving centred Bxα and Bx0 interference figures respectively. Z°C was measured only on twinned grains (Hess, 1949). Working on oriented grains from thin sections was preferred, because of the value of information that can be gained from single crystals. The method of determining α and γ by minimum and maximum indices of crushed material gives results which are not typical of the sample and the common small difference in optical properties between grains is obscured.

The refractive indices and optic axial angles of the biotites and hornblends were generally difficult to determine because of their strong absorption and common small inclusions of other minerals. An attempt was made to determine the indices of the analyzed samples when supported by admixed crushed glass, but with unsatisfactory results. Some measurements were made on grains broken out of sections carefully ground to less than usual thickness.

The plagioclase feldspars were examined chiefly by using a four axis universal stage, and little attention was paid to their refractive indices since compositions derived from ordinary index determinations are subject to considerable error arising from the probable error in the value determined for the index. Accordingly, plagioclase compositions were determined using the universal stage and following the methods of
Figure 60. Feldspathic pyroxenite (RB402), Poole's Point, from which pyroxene 1 was separated and analyzed. Plane light. Scale mark is 1.0 mm in length.

Figure 61. Hybrid dyke (RB12A), Poole's Point, from which pyroxene 3 was separated and analyzed. Crossed nicols. Scale mark is 1.0 mm in length.
Turner (1947) and Rittmann (1929; as set out in Emmons, 1943). The alkali feldspars were examined to find the value of the $\beta$ index and the optic axial angle.

Some minor observations made of the indices and optic axial angles of several other minerals have been given above in the petrography.

(2) The Clinopyroxenes

(a) Introduction.

Thirteen pyroxene fractions, separated from ten rock-types, have been analyzed. Except for the pyroxene found in the foyaitic dykes (described in the coastal dyke section), the thirteen samples quite adequately cover the range in composition and optics of all the pyroxenes encountered in the rocks of the complex. The pyroxene assemblage consists entirely of a lime-rich monoclinic variety; sub-calcic augite, pigeonite and orthopyroxene have not been observed as primary phases, and only a very few grains of orthopyroxene were found as an alteration of olivine in some specimens of pyroxenite.

The pyroxenes were separated from:

(1) Laminated feldspathic pyroxenite at Poole's Point.

(2) Basic dyke, Poole's Point, comparable with the laminated pyroxenite.

(3) Hybrid dyke, Poole's Point.

(4) Monzonite, from the quarry at Central Tilba.

(5) Porphyritic banatite, half a mile north of Tilba Tilba.
Same location as the rock analyzed by Brown (1930).

(6) Pyroxenite, one mile east of Central Tilba.

(7) Garnet-pyroxene rock, same locality.

(8) Shonkinite, Tilba Tilba.

(9) 'Ijolite', near the highway, two miles south of Mt Dromedary Trig.

(10) Nepheline monzonite, south-eastern flank of Little Dromedary. Same location as the rock analyzed by Brown (1930).

The pyroxenes are referred to below by these numbers; e.g. the pyroxenes 9A and 9B are the iron-poor and iron-rich fractions of the bulk pyroxene separated from the 'ijolite'.

The form of the clinopyroxenes varies from one rock-type to another, according to the particular conditions of crystallization. For example, the pyroxene grains in the laminated pyroxenite are roughly prismatic in form but have nibbled outlines which, together with other microscopic features, indicate slight recrystallization of the rock. The pyroxene in the massive pyroxenite has almost identical optical properties to that of the laminated pyroxenite, but lacks the recrystallization textures. In the monzonite and banatite the mineral occurs as rounded subhedral grains, at times enclosed by biotite or hornblende. The 'ijolite' and shonkinite contain equant euhedra, but the pyroxene of the nepheline monzonite occurs as subhedral grains which are often enveloped by the other mafic minerals. The pyroxene textures in the garnet-pyroxene rock are typically metamorphic, and poikiloblastic plates
are common in thin sections.

(b) **Optical properties.**

The pyroxenes of the undersaturated alkalic rocks show distinct marginal iron enrichment; the other pyroxenes are relatively homogeneous and optically very similar. They vary from almost colorless to a pale apple-green, and pleochroism is weak or absent. The cores of the pyroxene grains in the 'ijolite' and nepheline monzonite have a yellow tint which grades rapidly to olive-green at the borders. Pleochroism is moderate, with $X =$ yellow, $Y =$ olive-yellow and $Z =$ olive-green; absorption is $X < Y < Z$.

The optic axes show inclined dispersion which is distinct in pyroxenes from the pyroxenites, monzonite and banatite, and strong for those from the alkaline rocks. The color fringes are much more pronounced for the 'B' optic axis emerging near $[c]$, in (001). Twinning, either simple or multiple (Figure 62), with (100) as the composition plane, is not uncommon.

The optical measurements of the analyzed pyroxenes are listed in Table 15. The refractive indices and $2V$ are consistently higher than those shown in Hess (1949), but the birefringences are comparable.

The effect of minor elements on the optical properties is not fully understood. Titania apparently causes an increase in the principal refractive indices (Segnit, 1953; Murray, 1954; Wilkinson, 1957; Challis, 1963), and the acmite molecule appears to give higher values for $2V$ (Yagi, 1953). Although the increase in $2V$ with increasing soda is
apparent in the pyroxenes of this study, their titania content is usually only slightly greater than the 0.4% considered by Hess (1949) in drawing up his determinative curves.

The pyroxene in the garnet-bearing foyaitic dykes is mid- to dark-green and moderately pleochroic, and resembles the most iron-rich rims of the pyroxene in the 'ijolite' and nepheline monzonite. The grains are too small to permit exact measurements of refractive indices, 2V or 2^\(\alpha\). The β index appears to be considerably higher than 1.72, and the optic axial angle is positive and large. The extinction angle is about 35°. From these observations it would appear to be even more iron- and soda-rich than the pyroxenes 9B and 10B.

No other pyroxenes have been observed in any member of the complex which are noticeably enriched in iron or soda, and it may safely be assumed that all the pyroxenes have a salitic composition comparable with the analyzed samples.

(c) Zoning.

The relatively large-scale and irregular core and rim structure of pyroxenes from the 'ijolite' and nepheline monzonite has been mentioned above and is not dealt with further. However, the pyroxenes of these rocks show zone phenomena common to the other pyroxenes and are included in the following account.

Except in cases where there is evidence of considerable post-consolidation metamorphism, such as in the garnet-pyroxene rock, the
Figure 62. Twinned pyroxene crystal showing zoning by inclusions. Note abundant plagioclase inclusions in the outer zone and the narrow bands of rutile rods in a sagenitic web. Crossed nicols. Scale mark is 1 mm in length.
Figure 63. (a) Pyroxene grain showing broad outer zone relatively free from inclusions. Abundant tiny rods of rutile (?) are zonally arranged in the inner portion of the grain, and impart a darker color to it. Plane light. Scale mark is 0.2 mm in length.

(b) Same, near extinction position. Note difference in extinction angle between the core and the zone free of inclusions.
zones are oriented parallel to the crystal faces or, where the texture
is such that these are suppressed, parallel to recognizable crystallo-
graphic directions. There are three distinct types of zoning, all of
which may be developed in the same grain.

In the first type, the zones are picked out by minerals included
in the pyroxene, oriented in rows parallel to the crystal outlines. The
minerals more commonly found as inclusions are plagioclase, ilmenite-
magnetite, apatite, alteration products such as biotite or hornblende,
and sagenite (Figure 62). The zones free of inclusions may or may not
lie nearer the crystal margins (Figures 62 and 63).

The second type is characterized by small local variations in
color, birefringence and optic orientation of the pyroxene itself, and
the zones are relatively broad (Figure 63). In exceptional cases these
can be seen in thin section without the aid of a microscope. Their
average width is about 0.1 mm, with a few reaching 0.5 mm. The difference
in optical properties between the zones is slight; the lighter colored
portions have slightly lower refractive indices and higher birefringence
(up to 0.003) but 2V and Z°c are virtually unchanged throughout the
grain. There is a tendency for the sagenite to be developed in the zones
of lower birefringence.

The third type is not usually visible except between crossed
nicols, and preferably very near an extinction position. Under these
conditions the zoning appears as an alternation of very narrow, closely
spaced, light and dark striae parallel to the crystal margins. Very often
it is only the small difference in dispersion between the zones that
Figure 64. Shonkinite, with pyroxene crystal very near an extinction position showing lamellar zoning. Area marked is shown below. Convergent light, crossed nicols. Scale mark is 1 mm in length.

Figure 65. Detail of pyroxene grain in shonkinite, showing fine lamellar zoning. Convergent light, crossed nicols. Scale mark is 0.2 mm in length.
makes them visible near their extinction position (Figures 64 and 65). The thickness of each zone is of the order of $10^{-4}$ cm. Hour-glass structure is sometimes associated with this zone type. The zoning closely resembles the fine-scale, oscillatory type common in the plagioclase phenocrysts of andesitic lavas. Examination on the universal stage has shown that although discontinuities in the zoning are very rare, the adjoining zones grade quickly into one another. There are very few structures analogous to those resulting from resorption during the growth of plagioclase crystals, and the zoning is characterized by its regularity. No measurements of $2V$, refractive indices, $2\phi C$, etc. could be made, because of the very small scale of this zoning. The interference figures obtained on finely-zoned grains were as sharply defined as those obtained from apparently unzoned grains.

Grains of pyroxene showing both types of zoning were examined using an electron micro-analyzer in an attempt to determine the chemical differences between the zones.

A polished thin section of the shonkinite (pyroxene 8) was examined microscopically to locate a favorably oriented grain of pyroxene which was relatively free of inclusions, fractures, pits etc., and displayed clearly both the relatively broad and the very fine zones. A detailed, accurate plan of the crystal was prepared (Figure 66) and a traverse path, normal to the traces of the zones, was chosen. The section was transferred to the electron micro-analyzer and properly oriented by means of the sketch; traverses were then made with the detectors of the instrument adjusted to the different elements before each run so that
Figure 66. Diagram of zoned pyroxene studied using electron microanalyzer. Area marked 'A' is shown in Figure 67 and area marked 'B' in Figure 68.

Figure 67. Scintillation picture showing distribution of Fe over area marked 'A' in Figure 66. Broken lines enclose Fe-poor zone.
elemental variations between the zones could be estimated. The results of this approach supported the indications of the optical studies that the variation in chemistry was slight. It also became clear that the lamellar zoning was on much too fine a scale to warrant further investigation in this study. Accordingly, the instrument was set to scan repeatedly an area about 0.5 mm square which contained the broader type of zone. The signal from the X-ray detectors was passed to a cathode ray oscilloscope, whose screen was photographed over a period of minutes. In this way a scintillation picture is obtained of the distribution of any particular element in the area scanned. The relative abundances of an element are roughly proportional to the density of the points on the photograph, and if the bulk chemistry is known, as it is in this case, then the absolute variation in the elements can be estimated. Since this method gives a picture of relative abundances, a variation of 1% in the content of silica in the pyroxene would be difficult to detect, whereas a 1% variation in the alumina content would be much more apparent.

The best contrast in distribution between the zones was shown by iron and aluminium (Figures 67 and 68), while titanium, manganese, chromium and magnesium vary weakly or not at all. The micro-analyzer was not capable of determining sodium, and results with the relatively abundant silicon and calcium were, as expected, noncommittal.

In summary, the pyroxenes of the monzonitic complex differ texturally, depending on their rock-type, but appear to have very similar optical properties. Very fine, lamellar zoning is characteristic of the pyroxenes, but the structure usually is not visible except between crossed
Figure 68. Copy of scintillation pictures showing the distribution of Al over the area marked 'B' in Figure 66. Broken lines enclose the Al-poor area.
nicols and very near an extinction position. Appreciable compositional zoning occurs only in the pyroxenes from the 'ijolite' and nepheline monzonite, both nepheline-bearing rocks. Relatively broad zones of lighter color than the major part of the grain commonly occur within and parallel to crystal outlines. Preliminary studies of the chemical variation between the zones, using an electron micro-analyzer, indicate that the lighter colored areas contain considerably less Al and Fe, and approximately constant amounts of Ti, Mn, Cr and possibly Mg. The variation in Si, Na and Ca was not determined.

Pyroxenes more diopsidic than that of the shonkinite (e.g. 1, 2 and 3 in Table 14) are also lighter colored. Examination of Table 14 shows that the more diopsidic pyroxenes are poorer in Al as well as Fe, so it is very likely that the lighter colored, broad zones are more diopsidic than the remainder of the grain, and thus represent reverse zoning.

Kuno and Sawatari (1934) described crystals of augite which are zoned in a very similar way to the pyroxenes from the monzonitic complex at Mt Dromedary. The augites are made up of numerous concentric zones, sometimes of an oscillatory type, often combined with hour-glass structure. The zones are much broader than the lamellar type described in this study, and the authors were able to make detailed measurements of the optical variation between the zones. Their tentative conclusion was that crystallization tended towards decrease in CaO, Fe₂O₃, Al₂O₃ and TiO₂, and increase in MgO. It is interesting to note that these pyroxenes are very pale green, weakly pleochroic, and chemically very
similar to those at Mt Dromedary.

The explanations proposed for reverse and oscillatory zoning in pyroxenes are virtually identical with those put forward to account for similar zoning in plagioclase feldspar, and include intermittent additions of the parent magma, movement of the crystallizing grain within the magma, diffusion between grains of differing composition, and fluctuations in volatile concentration.

Microscopical and field evidence leave no doubt that the various pyroxenes have crystallized from their host-rocks, except in the case of the altered pyroxenites. It follows that the conditions leading to the zonal growth of the pyroxenes persisted throughout the differentiation and emplacement of the complex.

(d) The pyroxene analyses.

The formula for clinopyroxene, after Warren and Briscoe (1931), Berman (1937), Hess (1949) and Kuno (1955), is \( W_{1-p}(X,Y)_{1+p}Z_{2}O_{6} \), where

\[
W = Ca^{2+}, Na^{+}, K^{+}
\]
\[
X = Mg^{2+}, Fe^{2+}, Mn^{2+}, Ni^{2+}
\]
\[
Y = Al^{3+}, Fe^{3+}, Cr^{3+}, Ti^{4+}
\]
\[
Z = Si^{4+}, Al^{3+}, Ti^{4+}, Fe^{3+}
\]

The chemical analyses are shown in Table 14, and are recalculated on the basis of six oxygen atoms in Table 16.

Although the optical study which preceded the separation and analysis of the pyroxenes gave the first indication that there was probably little variation in their chemistry, such close similarity of the
TABLE 14. Analyses of clinopyroxenes from the monzonitic complex at Mount Dromedary.

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<td>0.01</td>
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</tr>
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<td>0.29</td>
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<td>0.67</td>
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<td>0.02</td>
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<td>0.04</td>
<td>0.02</td>
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<td>0.25</td>
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<td>H₂O²⁺</td>
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<td>0.21</td>
<td>0.23</td>
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<td>n.d.</td>
<td>0.14</td>
<td>0.26</td>
<td>0.17</td>
<td>0.17</td>
<td>0.28</td>
<td>0.42</td>
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<tr>
<td>Total</td>
<td>100.10</td>
<td>100.24</td>
<td>100.26</td>
<td>100.07</td>
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<td>100.21</td>
<td>100.17</td>
<td>100.25</td>
<td>100.15</td>
<td>100.21</td>
</tr>
</tbody>
</table>

*Analyst A.J. Easton; remaining analyses by R.S. Boesens.

1 Pyroxene from laminated feldspathic pyroxenite (RB402), Poole's Point. 2 Pyroxene from basic dyke (RB407), Poole's Point. 3 Pyroxene from hybrid dyke (RB12A), Poole's Point. 4 Pyroxene from coarse monzonite (RB221), Central Tilba quarry. 5 Pyroxene from porphyritic banatite (RB158), ½ mile north of Tilba Tilba. 6 Pyroxene from pyroxenite (RB363B) enclosed by nepheline monzonite, 1 mile east of Central Tilba. 7 Pyroxene from garnet-pyroxene rock (RB363A), 1 mile east of Central Tilba. 8 Pyroxene from shonkinite (RB350), Tilba Tilba. 9A, 9B Core and rim respectively of pyroxene from 'ijolite' (RB188) near highway, 2 miles south of Mt Dromedary Trig. 10, 10A, 10B Bulk pyroxene, core and rim pyroxene respectively from nepheline monzonite (RB432), south-east flank of Little Dromedary.

N.B. Analyses grouped in rock types: 1-3 pyroxenites; 4-5 monzonite and banatite; 6-7 altered pyroxenites; 8-10 nepheline-bearing rocks.
analyses was not anticipated, since silica ranges from 38% in the pyroxenite to more than 62% in the banatite.

**Silica** remains fairly constant except in the 'ijolite' and nepheline monzonite pyroxenes, where it decreases slightly. The figures for silica compare fairly closely with those of calcic tholeiitic clinopyroxenes with a similar degree of iron substitution (Hess, 1949). The Garbh Eilean pyroxenes (Murray, 1954) contain about 1% less silica, and those from the Black Jack sill (Wilkinson, 1957) have 2-3% less.

**Titania** is relatively low, resembling the amount contained in tholeiitic pyroxenes much more closely than the alkali basalt types. There is no apparent relation between the titania content of the pyroxenes and of their host rocks, nor is there any systematic increase or decrease of this component through the analyses. The zoned pyroxenes in the 'ijolite' and nepheline monzonite show, respectively, strong marginal enrichment and moderate marginal depletion of titania (Table 14; analyses 9A, 9B and 10A, 10B).

The amount of alumina varies considerably, but is more abundant in the pyroxenes of the more alkaline rocks. The alumina figures are rather higher than is usual in tholeiitic clinopyroxenes, but are very similar to those quoted by Murray (1954) for the teschenitic Garbh Eilean sill.

**Manganese** follows ferrous iron, with the ratio FeO/MnO approximately twenty to one, except for analyses 6 and 9B which are low in MnO.

The figures for lime, magnesia and ferrous iron are very similar to those of Murray (1954) and Wilkinson (1957). Ferrous iron enters the pyroxene structure chiefly at the expense of magnesium. The range in
### Table 15: Chemical and optical properties of the clinopyroxenes in Table 14.

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<td><strong>Composition, atomic per cent</strong></td>
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<td></td>
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<td></td>
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<tr>
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<td>45.6</td>
<td>43.5</td>
<td>43.9</td>
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<td>46.3</td>
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<td>45.8</td>
<td>44.7</td>
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<tr>
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<td>40.4</td>
<td>40.6</td>
<td>36.8</td>
<td>37.2</td>
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<tr>
<td>Fe</td>
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<td>13.7</td>
<td>13.8</td>
<td>19.7</td>
<td>18.9</td>
<td>13.4</td>
<td>15.7</td>
<td>15.2</td>
<td>14.1</td>
<td>25.1</td>
<td>16.6</td>
<td>21.4</td>
<td>25.0</td>
</tr>
</tbody>
</table>

| **Specific gravity** | 3.38 | 3.36 | 3.36 | 3.40 | - | 3.34 | 3.35 | 3.36 | 3.35 | 3.41 | 3.36 | 3.38 | 3.40 |

| **Mean refractive indices** |    |    |    |    |    |    |    |    |    |    |    |    |    |
| α  | 1.694 | 1.693 | 1.696 | 1.696 | 1.694 | 1.695 | 1.695 | 1.695 | 1.702 | 1.695 | - | 1.702 |    |
| β  | 1.700 | 1.699 | 1.701 | 1.701 | 1.700 | 1.702 | 1.701 | 1.703 | 1.709 | 1.702 | - | 1.709 |    |
| γ  | 1.721 | 1.720 | 1.723 | 1.723 | 1.722 | 1.724 | 1.723 | 1.723 | 1.723 | 1.724 | - | 1.730 |    |

| **Mean 2Vε** | 54.0° | 54.0° | 54.0° | 53.5° | 53.5° | 58.5° | 58.5° | 54.5° | 55.5° | 60.0° | 56.0° | - | 61.5° |
| Δ - α | 0.027 | 0.027 | 0.027 | 0.028 | 0.027 | 0.028 | 0.028 | 0.028 | 0.028 | 0.028 | - | 0.028 |    |

| **FeO + Fe₂O₃** | x 100; |
| **FeO + Fe₂O₃ + MgO** |    |
| Pyroxene | 34.7 | 37.3 | 37.6 | 48.1 | 47.1 | 38.4 | 42.8 | 41.5 | 38.4 | 58.6 | 44.4 | 53.8 | 59.2 |
| Rock    | 70.5 | 53.6 | 49.0 | 66.6 | 76.3 | 67.3 | - | 58.1 | 73.8 |    |    |    |    |

* Fe = Fe³⁺ + Fe²⁺ + Mn²⁺.

Specimens numbered as in Table 14.
values for ferrous iron in these analyses is small compared to that of pyroxenes from the tholeiitic rocks showing a similar degree of fractionation (Brown and Vincent, 1963).

Soda is fairly high throughout, and increases sharply in the pyroxenes from the nepheline-bearing rocks. The increase in soda is reflected in the increase in optic axial angle (Table 15). A similar but much stronger trend is recorded by Yagi (1953) for the alkalic rocks of the Morotu district.

Kushiro (1960) and Le Bas (1962) have described and attempted to account for the amount and distribution of Si, Al and Ti in calcic igneous clinopyroxenes. Kushiro has shown that when Al is plotted against Si or Ti, the pyroxenes of tholeiitic, feldspathoidal and non-feldspathoidal alkalic rocks fall in three roughly defined fields. A more detailed study along similar lines was made by Le Bas, who effected a clearer distinction between the pyroxenes of the major magma-types by considering only groundmass pyroxene. Le Bas discusses the relation of the amount of aluminium in tetrahedral co-ordination \( \text{Al}_2 \) to the total aluminium and to \( \text{TiO}_2 \); his plot of \( \text{Al}_2 \) against \( \text{TiO}_2 \) weight per cent (p. 280) indicates fundamentally different trends, during differentiation, of pyroxenes from the alkaline rocks and from the tholeiitic, high-alumina and calc-alkaline rocks.

Comparison of the data from the present study with those of Kushiro and Le Bas shows:

1. The pyroxenes of the monzonitic complex have relatively small figures for \( \text{Al}_2 \) compared to other pyroxenes with similar
TABLE 16. Analysis of Table 14 recalculated on the basis of six oxygen atoms.

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<td>Si</td>
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<td>1.874</td>
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<td>1.850</td>
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<td>0.150</td>
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<td>0.039</td>
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<td>0.034</td>
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<td>0.063</td>
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<td>0.054</td>
<td>0.080</td>
<td>0.050</td>
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<td>0.080</td>
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<td>2.00</td>
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</table>

% Al in Z: 5.0 6.3 5.6 4.4 5.8 6.5 2.5 6.4 7.0 7.5 5.0 6.4 6.3
### TABLE 17. C.I.P.W. Norms of Analyses in Table 8.

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<td>0.22</td>
<td>0.45</td>
<td>0.95</td>
<td>0.17</td>
<td>0.17</td>
<td>0.28</td>
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<td>0.61</td>
<td>0.39</td>
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<td>0.95</td>
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<tr>
<td>Ab</td>
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<td>1.42</td>
<td>4.83</td>
<td>4.04</td>
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<td>8.82</td>
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<td>1.08</td>
</tr>
<tr>
<td>Il</td>
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<td>1.33</td>
<td>0.91</td>
<td>0.85</td>
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<td>1.33</td>
<td>0.46</td>
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<td>1.20</td>
<td>3.92</td>
<td>1.79</td>
<td>1.56</td>
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</tr>
<tr>
<td>Mt</td>
<td>3.47</td>
<td>2.45</td>
<td>3.25</td>
<td>2.57</td>
<td>2.69</td>
<td>4.17</td>
<td>2.64</td>
<td>3.91</td>
<td>4.12</td>
<td>5.79</td>
<td>5.79</td>
<td>6.19</td>
<td>-</td>
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<td>Cm</td>
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<td>0.11</td>
<td>0.13</td>
<td>0.02</td>
<td>0.09</td>
<td>-</td>
<td>-</td>
<td>0.07</td>
<td>0.02</td>
<td>-</td>
<td>0.02</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>Total</td>
<td>100.01</td>
<td>100.04</td>
<td>100.11</td>
<td>99.79</td>
<td>100.26</td>
<td>100.26</td>
<td>99.57</td>
<td>100.00</td>
<td>99.98</td>
<td>100.01</td>
<td>99.64</td>
<td>99.61</td>
<td>99.53</td>
</tr>
</tbody>
</table>

Numbering as in Table 14.
proportions of \((Ca^{2+} + Na^+ + K^+)\), \(Mg^{2+}\) and \((Fe^{3+} + Fe^{2+} + Mn^{2+})\). This is a consequence of relatively high \(SiO_2\), and has been noted above.

(ii) The proportion of the total Al that is in tetrahedral co-ordination is consequently low, and compares more closely with the values for pyroxenes from tholeiitic, high-alumina and calc-alkalic rocks than with those of the alkali types.

(iii) In the diagrams of \(Al_z\) versus \(TiO_2\) wt per cent and of \(Al_2O_3\) wt per cent versus \(SiO_2\) wt per cent, the pyroxenes plot in the field of the tholeiitic, high-alumina and calc-alkaline types.

The relations between oversaturated tholeiite, undersaturated tholeiite and critically undersaturated alkali basalts are illustrated by Yoder and Tilley (1962, p.352). They point out that the norms of the pyroxenes from the alkali basalts studied by them contain nepheline, while the norms of the pyroxenes from the tholeiitic rocks contain considerable hypersthene or even quartz (p.365, Table 5).

All the pyroxenes in the present study contain normative nepheline except 5 and 9B (Table 17); the amount of hypersthene in the norms of these two pyroxenes is small, and indicates only a very slight excess of silica over that required to satisfy the \(Na_2O\) for albite. Although normative hypersthene appears in 9B, the amount of nepheline in the norm of 9A ensures that the bulk pyroxene would also have nepheline in its norm. The absence of normative nepheline in pyroxene 5 is hardly surprising since the host rock (porphyritic banatite) contains about 60 wt per cent \(SiO_2\), and about 5% of quartz appears in its norm.
Figure 69. Plot of analyses of the pyroxenes from the monzonitic complex on conventional Ca-Mg-Fe atomic per cent diagram (Fe = Fe$^{3+}$ + Fe$^{2+}$ + Mn$^{2+}$). Solid circles represent analyses of pyroxenes from nepheline-bearing rocks. Numbers correspond to those of the analyses in Table 14.
(e) **Trend of pyroxene composition.**

The analyzed clinopyroxenes are plotted in Figures 69 and 70 on the basis of atomic per cent Ca–Mg–Fe, where \( \text{Fe} = \text{Fe}^{3+} + \text{Fe}^{2+} + \text{Mn}^{2+} \).

Figure 69 shows the detailed relations of the various pyroxene fractions. The upper and lower limit of iron substitution occurs in pyroxenes 9B and 1 respectively. The trend of pyroxene composition is from \( \text{Ca}_{46.1}\text{Mg}_{40.8}\text{Fe}_{13.1} \) to \( \text{Ca}_{43.0}\text{Mg}_{31.9}\text{Fe}_{25.1} \). The trends of zoning in the 'ijolite' (9A, 9B) and mepholine monzonite (10A, 10, 10B) pyroxenes are very similar to each other, and to the pyroxenite–banatite trend (2, 5 respectively). The plots of the three pyroxenes from the pyroxenitic rocks at Poole’s Point (1, 2, 3) nearly coincide. The pyroxenes from the altered pyroxenite (6, 7) are more lime-rich than 1, 2 and 3; while 7, separated from the garnet–pyroxene rock, is noticeably poor in titania and alumina and rich in silica.

The pyroxenes from the monzonite and banatite (4, 5) plot very close to one another, slightly below the mid-point of the line joining the analyses of the core and rim of the 'ijolite' pyroxene. The plots of 4 and 5 are curious in that they show the pyroxene from the banatite to be slightly more diopsidic than that from the more basic monzonite.

The trends for the pyroxenes from the tholeiitic intrusion at Red Hill (McDougall, 1961) and for those of the alkali basalt magma (Wilkinson, 1956) are compared in Figure 70 with the pyroxenes studied herein.

(f) **Pyroxenes in other monzonitic rocks.**

The analysis of a pyroxene from the Mt Dromedary area given by
Figure 70. Plot of analyses of pyroxenes from the monzonitic complex on conventional Ca-Mg-Fe atomic per cent diagram (Fe = Fe$^{3+}$ + Fe$^{2+}$ + Mn$^{2+}$), together with the trend for pyroxenes from alkali basalt magma (A - A) after Wilkinson (1956), and the tholeiitic Red Hill intrusion (T - T) after McDougall (1961).
Wilshire and Binns (1961, p.198) has not been used in this study because the present writer was unable to ascertain either the exact location or the nature of the host rock.

Other examples of relatively potash-rich monzonitic intrusions emplaced in a late- or post-orogenic environment occur at Fort Cygnet, Tasmania; Milton, about 100 miles north of Mt Dromedary; Central Montana; the kentallenites of the South-West Highlands of Scotland; and possibly at the type locality for monzonite, Monzoni in the Tyrol.

The pyroxene found in such intrusions is typically a lime-rich monoclinic variety. It is only rarely ophitic in form, and is colorless or pale green and very weakly pleochroic. In the highly alkalic rocks the pyroxene grains may have a deeper green rim or, more rarely, occur as aegirine-augite. Zoning by inclusions, color and birefringence is not uncommon (Hill and Kynaston, 1900, Plate XXIX; Larsen et al., 1941, p.1843). The striking constancy in composition of the pyroxenes from this type of association has been commented upon by Larsen et al. (1941), Pirsson (1905a) and others. Larsen et al. (1941), writing on the mineralogy of the igneous rocks of the Highwood Mountains, state (pp.1843-1844) "the augites are pale green and characteristically zoned with recurrent zones of somewhat different color and extinction angle but with not much difference in composition ... The augites show a remarkable uniformity in their optical properties in nearly all the rocks".

The analyses of pyroxene from the shonkinite at Square Butte in the Highwood Mountains district (Pirsson, 1905b, p.98) and of that from the shonkinite at Mt Dromedary are compared in Table 18.
The alkali basalt magma has a definite preponderance of soda over potash which is maintained in the strongly fractionated rocks, despite the fact that in the more felsic differentiates potash increases at a greater rate than soda. The igneous rocks at Mt Dromedary, on the other hand, have a preponderance of potash over soda in all but the ultramafic types, and there is a weak trend toward relative enrichment in potash. This difference in potash-soda relationship between the alkali basalts and the monzonitic suite (shoshonites) is reflected in the pyroxenes. The soda content of the pyroxenes from monzonitic and syenitic rocks derived
from alkali basalt (Yagi, 1953) is greater than that of the pyroxenes from comparable rocks at Mt Dromedary and similar occurrences.

(2) The Biotites

(a) Introduction.

Seven biotites were separated from the more important members of the complex; they have been chemically analyzed and their optical properties and specific gravities determined. The rocks from which the biotites were separated are:

(A) Laminated feldspathic pyroxenite, Poole's Point.
(B) Pyroxenite, one mile east of Central Tilba.
(C) Shonkinite, Tilba Tilba.
(D) Monzonite, Central Tilba quarry. Same locality as rock analyzed by Brown (1930, p.650).
(E) Nepheline monzonite, south-eastern flanks of Little Dromedary. Same locality as rock analyzed by Brown (1930, p.659).
(F) Recrystallized xenolith, Central Tilba quarry.
(G) Porphyritic banatite, half a mile north of Tilba Tilba. Same locality as rock analyzed by Brown (1930, p.647).

The biotites are referred to below by the same letters as their host-rocks, immediately above.

Although the biotites have different origins, some being metamorphic and some of primary crystallization, they are chemically very closely related and are treated as a group. The range of composition
TABLE 19. Optical properties of the analyzed biotites.

<table>
<thead>
<tr>
<th>Refractive Indices</th>
<th>2V</th>
<th>Pleochroism</th>
<th>Absorption</th>
<th>Specific Gravity</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>A</strong> 1.590-1.592</td>
<td>3°-5°</td>
<td>X = very pale yellow</td>
<td>X &lt;&lt; Y &lt; Z</td>
<td>2.92</td>
</tr>
<tr>
<td>( \alpha ) 1.627-1.630</td>
<td>( \beta )</td>
<td>Y = red-brown</td>
<td>( \gamma )</td>
<td>Z = darker red-brown</td>
</tr>
<tr>
<td>( \gamma )</td>
<td>( \beta )</td>
<td>( \alpha )</td>
<td></td>
<td>( \beta )</td>
</tr>
<tr>
<td>( \gamma )</td>
<td>( \beta )</td>
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</tr>
<tr>
<td>( \gamma )</td>
<td>( \beta )</td>
<td>( \alpha )</td>
<td></td>
<td>( \beta )</td>
</tr>
<tr>
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<td>X &lt;&lt; Y &lt; Z</td>
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</tr>
<tr>
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<td>1.641</td>
<td>Y = brown, green tint</td>
<td>( \gamma )</td>
<td>strong</td>
</tr>
<tr>
<td>( \gamma )</td>
<td></td>
<td>( \beta )</td>
<td></td>
<td>Z = olive-brown</td>
</tr>
<tr>
<td>( \gamma )</td>
<td></td>
<td>( \beta )</td>
<td></td>
<td>( \alpha )</td>
</tr>
<tr>
<td>( \gamma )</td>
<td></td>
<td>( \beta )</td>
<td></td>
<td>( \alpha )</td>
</tr>
<tr>
<td><strong>C</strong> 1.659</td>
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<td>X = yellow</td>
<td>X &lt;&lt; Y &lt; Z</td>
<td>3.00</td>
</tr>
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<td>( \gamma )</td>
<td>strong</td>
</tr>
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<td>( \gamma )</td>
<td>( \beta )</td>
<td>( \alpha )</td>
<td></td>
<td>( \beta )</td>
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<td>( \beta )</td>
<td>( \alpha )</td>
<td></td>
<td>( \beta )</td>
</tr>
<tr>
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<td>( \beta )</td>
<td>( \alpha )</td>
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<td>( \gamma )</td>
<td>( \beta )</td>
<td>( \alpha )</td>
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<td>Y = brown</td>
<td>( \gamma )</td>
<td>strong</td>
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<td>( \beta )</td>
<td>( \alpha )</td>
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<td>( \beta )</td>
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</tr>
<tr>
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<td>Y = brown, green tint</td>
<td>( \gamma )</td>
<td>strong</td>
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<td>( \gamma )</td>
<td>( \beta )</td>
<td>( \alpha )</td>
<td></td>
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<td>( \beta )</td>
<td>( \alpha )</td>
<td></td>
<td>( \beta )</td>
</tr>
<tr>
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<td>3.12</td>
</tr>
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<td>1.670</td>
<td>Y = deeper yellow-green</td>
<td>( \gamma )</td>
<td>strong</td>
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<tr>
<td>( \gamma )</td>
<td>( \beta )</td>
<td>( \alpha )</td>
<td></td>
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</tr>
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<td>X &lt; Y &lt; Z</td>
<td>3.12</td>
</tr>
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<td>( \beta )</td>
<td>1.667</td>
<td>Y = deeper green</td>
<td>( \gamma )</td>
<td>strong</td>
</tr>
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<td>( \gamma )</td>
<td>( \beta )</td>
<td>( \alpha )</td>
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<td>( \beta )</td>
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<td>Y = green</td>
<td>( \gamma )</td>
<td>strong</td>
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<td>( \beta )</td>
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<td>( \beta )</td>
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<td>X &lt; Y &lt; Z</td>
<td>3.12</td>
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<td>( \beta )</td>
<td>1.667</td>
<td>Y = green</td>
<td>( \gamma )</td>
<td>strong</td>
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<tr>
<td>( \gamma )</td>
<td>( \beta )</td>
<td>( \alpha )</td>
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<td>X &lt; Y &lt; Z</td>
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<td>1.667</td>
<td>Y = green</td>
<td>( \gamma )</td>
<td>strong</td>
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<td>( \gamma )</td>
<td>( \beta )</td>
<td>( \alpha )</td>
<td></td>
<td>( \beta )</td>
</tr>
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<td><strong>M</strong> 1.666</td>
<td>23°-25°</td>
<td>X = brown</td>
<td>X &lt; Y &lt; Z</td>
<td>3.12</td>
</tr>
<tr>
<td>( \beta )</td>
<td>1.667</td>
<td>Y = green</td>
<td>( \gamma )</td>
<td>strong</td>
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<tr>
<td>( \gamma )</td>
<td>( \beta )</td>
<td>( \alpha )</td>
<td></td>
<td>( \beta )</td>
</tr>
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<td>25°-28°</td>
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<td>X &lt; Y &lt; Z</td>
<td>3.12</td>
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<tr>
<td>( \beta )</td>
<td>1.667</td>
<td>Y = green</td>
<td>( \gamma )</td>
<td>strong</td>
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<td>( \gamma )</td>
<td>( \beta )</td>
<td>( \alpha )</td>
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<td>( \beta )</td>
</tr>
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<td><strong>O</strong> 1.666</td>
<td>28°-30°</td>
<td>X = brown</td>
<td>X &lt; Y &lt; Z</td>
<td>3.12</td>
</tr>
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<td>( \beta )</td>
<td>1.667</td>
<td>Y = green</td>
<td>( \gamma )</td>
<td>strong</td>
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<td>( \gamma )</td>
<td>( \beta )</td>
<td>( \alpha )</td>
<td></td>
<td>( \beta )</td>
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<td>30°-33°</td>
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<td>X &lt; Y &lt; Z</td>
<td>3.12</td>
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<tr>
<td>( \beta )</td>
<td>1.667</td>
<td>Y = green</td>
<td>( \gamma )</td>
<td>strong</td>
</tr>
<tr>
<td>( \gamma )</td>
<td>( \beta )</td>
<td>( \alpha )</td>
<td></td>
<td>( \beta )</td>
</tr>
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<td>X &lt; Y &lt; Z</td>
<td>3.12</td>
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<tr>
<td>( \beta )</td>
<td>1.667</td>
<td>Y = green</td>
<td>( \gamma )</td>
<td>strong</td>
</tr>
<tr>
<td>( \gamma )</td>
<td>( \beta )</td>
<td>( \alpha )</td>
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<td>( \beta )</td>
</tr>
<tr>
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<td>35°-38°</td>
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<td>X &lt; Y &lt; Z</td>
<td>3.12</td>
</tr>
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<td>( \beta )</td>
<td>1.667</td>
<td>Y = green</td>
<td>( \gamma )</td>
<td>strong</td>
</tr>
<tr>
<td>( \gamma )</td>
<td>( \beta )</td>
<td>( \alpha )</td>
<td></td>
<td>( \beta )</td>
</tr>
<tr>
<td><strong>S</strong> 1.666</td>
<td>38°-40°</td>
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<td>X &lt; Y &lt; Z</td>
<td>3.12</td>
</tr>
<tr>
<td>( \beta )</td>
<td>1.667</td>
<td>Y = green</td>
<td>( \gamma )</td>
<td>strong</td>
</tr>
<tr>
<td>( \gamma )</td>
<td>( \beta )</td>
<td>( \alpha )</td>
<td></td>
<td>( \beta )</td>
</tr>
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<td>40°-42°</td>
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<td>X &lt; Y &lt; Z</td>
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<td>( \beta )</td>
<td>1.667</td>
<td>Y = green</td>
<td>( \gamma )</td>
<td>strong</td>
</tr>
<tr>
<td>( \gamma )</td>
<td>( \beta )</td>
<td>( \alpha )</td>
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<td>( \beta )</td>
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<td>Y = green</td>
<td>( \gamma )</td>
<td>strong</td>
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<td>( \beta )</td>
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<td>1.667</td>
<td>Y = green</td>
<td>( \gamma )</td>
<td>strong</td>
</tr>
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<td>( \gamma )</td>
<td>( \beta )</td>
<td>( \alpha )</td>
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<tr>
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<td>X &lt; Y &lt; Z</td>
<td>3.12</td>
</tr>
<tr>
<td>( \beta )</td>
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<td>Y = green</td>
<td>( \gamma )</td>
<td>strong</td>
</tr>
<tr>
<td>( \gamma )</td>
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<td>48°-50°</td>
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<td>X &lt; Y &lt; Z</td>
<td>3.12</td>
</tr>
<tr>
<td>( \beta )</td>
<td>1.667</td>
<td>Y = green</td>
<td>( \gamma )</td>
<td>strong</td>
</tr>
<tr>
<td>( \gamma )</td>
<td>( \beta )</td>
<td>( \alpha )</td>
<td></td>
<td>( \beta )</td>
</tr>
</tbody>
</table>
and optical properties shown by these analyzed samples covers all the biotites found in the complex; in fact, the only biotite not closely similar to one or other of the seven is found in a recrystallized basic rock in the contact metamorphic aureole.

(b) **Optical properties and the effects of major element variation.**

The chemical analyses are listed in Table 20 and the optical properties and densities of the analyzed biotites are shown in Table 19. The biotite fractions C and G each consist of two varieties which differ slightly in optical properties. In each case the younger modification occurs in quite small amounts, either as marginal patches on the older biotite, or as separate small grains, or both. Biotite F is similarly associated with a small quantity of a pleochroic clino-amphibole, which has

\[
\begin{align*}
X &= \text{straw-yellow} \\
Y &= \text{yellow-green} \\
Z &= \text{blue-green}.
\end{align*}
\]

In all tables the biotites are listed in order of decreasing MgO wt per cent. The following variation in physical properties of the biotites has been observed as the amount of MgO decreases:

(i) The color for the Y or Z directions changes from red to red-brown to green-brown.

(ii) The absorption parallel to X, Y and Z increases.

(iii) The refractive indices increase.

(iv) The birefringences increase.

(v) \(2V\) tends to increase.
Figure 7. Absorption picture of some biotites.
(vi) Dispersion of the optic axes increases.
(vii) The density of the mineral increases.

The color of the biotites compares closely with those studied by Hall (1941a) with similar proportions of FeO, MgO and TiO₂. The refractive indices estimated from the indices in Table 19 have been plotted against FeO + 2(Fe₂O₃ + TiO₂) and the resulting curve is in good agreement with that of Heinrich (1946).

It was stated above that as magnesia decreased the absorption for each of the X, Y and Z directions increased. The absorption schemes of several of the biotites have been compared graphically in Figure 71. The curves were drawn from points obtained by plotting the readings of an exposure meter inserted in the ocular of a petrographic microscope against the rotation, in degrees, of a biotite grain from minimum to maximum absorption.

In Figure 71, the ordinate is linear in scale and the figures on it are inversely proportional to the intensity of the light received at the photocell, so that high values correspond to low light intensity, i.e. to high absorption. The abscissa records the stage rotation in degrees, where 0° corresponds to the position of minimum absorption, i.e. with the trace of the cleavage normal to the plane of polarization of the lower nicol, and 90° corresponds to the position of maximum absorption.

Care was taken to use a light source of standard intensity, and to measure only well-oriented grains in sections of standard thickness. Variations in light intensity and grain thickness merely change the vertical scale of the curve while its form remains characteristic for
### TABLE 20a. Chemical analyses of biotites.

<table>
<thead>
<tr>
<th></th>
<th>A</th>
<th>B</th>
<th>C</th>
<th>D</th>
<th>E</th>
<th>F</th>
<th>G</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>37.36</td>
<td>36.60</td>
<td>36.80</td>
<td>37.01</td>
<td>36.02</td>
<td>35.90</td>
<td>37.96</td>
</tr>
<tr>
<td>TiO₂</td>
<td>3.68</td>
<td>3.28</td>
<td>3.70</td>
<td>3.96</td>
<td>3.67</td>
<td>4.55</td>
<td>3.26</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>16.60</td>
<td>15.98</td>
<td>14.27</td>
<td>14.22</td>
<td>14.56</td>
<td>14.82</td>
<td>12.79</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>1.61</td>
<td>2.80</td>
<td>2.85</td>
<td>2.88</td>
<td>3.37</td>
<td>3.12</td>
<td>2.31</td>
</tr>
<tr>
<td>FeO</td>
<td>8.90</td>
<td>14.35</td>
<td>14.30</td>
<td>17.78</td>
<td>18.64</td>
<td>18.50</td>
<td>21.10</td>
</tr>
<tr>
<td>MnO</td>
<td>0.09</td>
<td>0.14</td>
<td>0.15</td>
<td>0.20</td>
<td>0.42</td>
<td>0.22</td>
<td>0.43</td>
</tr>
<tr>
<td>MgO</td>
<td>17.92</td>
<td>14.00</td>
<td>13.14</td>
<td>10.83</td>
<td>10.42</td>
<td>9.84</td>
<td>8.49</td>
</tr>
<tr>
<td>CaO</td>
<td>1.23</td>
<td>1.21</td>
<td>1.62</td>
<td>1.09</td>
<td>0.73</td>
<td>0.50</td>
<td>0.85</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.38</td>
<td>0.26</td>
<td>0.11</td>
<td>0.10</td>
<td>0.21</td>
<td>0.28</td>
<td></td>
</tr>
<tr>
<td>K₂O</td>
<td>8.38</td>
<td>8.36</td>
<td>7.68</td>
<td>8.70</td>
<td>8.71</td>
<td>8.89</td>
<td>8.84</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.19</td>
<td>0.00</td>
<td>0.10</td>
<td>0.03</td>
<td>0.00</td>
<td>0.03</td>
<td>0.08</td>
</tr>
<tr>
<td>H₂O⁺</td>
<td>2.79</td>
<td>2.98</td>
<td>3.96</td>
<td>2.66</td>
<td>2.88</td>
<td>2.93</td>
<td>2.84</td>
</tr>
<tr>
<td>Total</td>
<td>99.93</td>
<td>100.29</td>
<td>99.13</td>
<td>99.79</td>
<td>100.39</td>
<td>100.28</td>
<td>100.08</td>
</tr>
</tbody>
</table>

* Determined spectrographically.

Analyses arranged in order of decreasing MgO. Anal. R.S. Boesen.

A Biotite from laminated feldspathic pyroxenite, Poole's Point.
B Biotite from pyroxenite body enclosed by nepheline monzonite, 1 mile south-east of Central Tilba.
C Biotite from shonkinite, Tilba Tilba.
D Biotite from monzonite, Central Tilba quarry.
E Biotite from nepheline monzonite, south-eastern slope of Little Dromedary.
F Biotite from xenolith, Central Tilba quarry.
G Biotite from porphyritic marginal phase of banatite, 1 mile north of Tilba Tilba.
<table>
<thead>
<tr>
<th>Detection limit</th>
<th>A</th>
<th>B</th>
<th>C</th>
<th>D</th>
<th>F</th>
<th>G</th>
</tr>
</thead>
<tbody>
<tr>
<td>B</td>
<td>10</td>
<td>10-20</td>
<td>120</td>
<td>10-20</td>
<td>41</td>
<td>24</td>
</tr>
<tr>
<td>Co</td>
<td>1</td>
<td>105</td>
<td>120</td>
<td>127</td>
<td>135</td>
<td>18</td>
</tr>
<tr>
<td>Cr</td>
<td>2</td>
<td>5</td>
<td>150</td>
<td>120</td>
<td>135</td>
<td>8</td>
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<td>Cu</td>
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<td>105</td>
<td>120</td>
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<td>Nb</td>
<td>5</td>
<td>105</td>
<td>105</td>
<td>5</td>
<td>-</td>
<td>7</td>
</tr>
<tr>
<td>Ni</td>
<td>2</td>
<td>105</td>
<td>330</td>
<td>5</td>
<td>-</td>
<td>7</td>
</tr>
<tr>
<td>Sc</td>
<td>1</td>
<td>17</td>
<td>123</td>
<td>38</td>
<td>7</td>
<td>14</td>
</tr>
<tr>
<td>V</td>
<td>3</td>
<td>350</td>
<td>360</td>
<td>360</td>
<td>225</td>
<td>620</td>
</tr>
<tr>
<td>Zr</td>
<td>5</td>
<td>tr.</td>
<td>10</td>
<td>10</td>
<td>85</td>
<td>60</td>
</tr>
</tbody>
</table>

Other elements looked for but not found: Be, La, Mo, Nd, U, W, Y.

*Anal. I.D. Martin.*
that grain. The curves for biotites A, C, D, E and F are surprisingly reproducible and appear to be characteristic for each particular species of biotite.

Diagrams of this type are very simple to construct and they are a useful adjunct to the conventional statement of schemes of pleochroism and absorption.

(c) The chemical analyses.

The analyses of the major and trace constituents of the biotites are contained in Tables 20a and 20b respectively. The range of iron and magnesium oxides is quite considerable and the substitution between these two components is the most extensive chemical variation of the group. Alumina is rather low, and decreases slowly as iron increases. Titania is fairly high, and remains relatively constant, except for biotite F in which it is more abundant. Ferrous oxide ranges from 8.90% to 21.10% and the ratio of FeO to FeO is relatively constant at about 5.5 except for biotite G where it is 9.1. MnO tends to follow ferrous iron. The biotites contain considerable amounts of baria and lime. Baria is highest in the magnesia-rich and magnesia-poor samples and lowest in the intermediate members. Potash is less abundant than is usual because of the relatively high content of lime, baria and soda.

In a study of the major elements of some biotites, Vistelius and Sarmanov (1961) conclude that there is a negative correlation between Fe$^{2+}$ and Mg, Ti and Mg, and Fe$^{3+}$ and Mg; and positive correlation between Al$^4$ and Fe$^{2+}$, Ti and Fe$^{2+}$, and K and Al$^4$. Inspection of the atomic ratios of the biotites of the present study (Table 21) shows that the
Figure 72. Plot of wt% MgO - Fe$_2$O$_3$ + TiO$_2$ - FeO + MnO compared to some paragenetic fields. (After Heinrich, 1946.)
correlation of Fe$^{2+}$ and Mg is negative in all the biotites, the relationship between Fe$^{2+}$ and Mg and between Ti and Mg in the biotites of undoubted igneous origin (C, D, G) agrees with that described by the above authors, but Ti and Fe$^{2+}$ appear to be related in the opposite sense.

The order in which the biotites are listed corresponds to the order of increasing fractionation in the host-rocks (on the criteria of Simpson's mafic and felsic indices, 1954), with the exception of the xenolithic rock from which biotite F was separated. The ratio of FeO to MgO in the biotites increases during differentiation and the ratio of RO to $R_2\text{O}_3$ decreases. A similar variation has been noted by Deer (1937), Brammall and Harwood (1932), and others.

Figure 72 shows the analyzed biotites plotted on the basis of MgO, FeO + MnO, and Fe$_2\text{O}_3$ + TiO$_2$ wt per cent after Heinrich (1946), with his fields for biotites from gabbroic, dioritic, syenitic and granitic rocks included for comparison. The trend of composition of the biotites A–G agrees with that proposed by Heinrich, but individual biotites do not plot in the appropriate fields (e.g. A, B and C). The relation between MgO, Al$_2\text{O}_3$ and total iron (as FeO) of the biotites is similar to that described by Nockolds (1947) for the biotites of alkaline igneous rocks.

(d) **Structural formulae.**

The structure of the mica group has been determined through the investigations of Mauguin, Pauling, Jackson and West, Hendricks, and others. The general formula of the micas may be written (after Berman, 1937) as $W_4(X, Y)_{8-12}Z_{16}O_{40}(O, OH, F)_8$, where
### TABLE 21. Atomic ratios and charge distribution of analyzed biotites.

<table>
<thead>
<tr>
<th>ATOMIC RATIOS</th>
<th>A</th>
<th>B</th>
<th>C</th>
<th>D</th>
<th>E</th>
<th>F</th>
<th>G</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Tetrahedral</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Si</td>
<td>2.68</td>
<td>2.70</td>
<td>2.77</td>
<td>2.78</td>
<td>2.71</td>
<td>2.71</td>
<td>2.90</td>
</tr>
<tr>
<td>Al</td>
<td>1.32</td>
<td>1.30</td>
<td>1.23</td>
<td>1.22</td>
<td>1.29</td>
<td>1.29</td>
<td>1.10</td>
</tr>
<tr>
<td></td>
<td>4.00</td>
<td>4.00</td>
<td>4.00</td>
<td>4.00</td>
<td>4.00</td>
<td>4.00</td>
<td>4.00</td>
</tr>
<tr>
<td><strong>Octahedral</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Al</td>
<td>0.09</td>
<td>0.08</td>
<td>0.03</td>
<td>0.04</td>
<td>0.01</td>
<td>0.03</td>
<td>0.06</td>
</tr>
<tr>
<td>Ti</td>
<td>0.20</td>
<td>0.18</td>
<td>0.21</td>
<td>0.21</td>
<td>0.21</td>
<td>0.26</td>
<td>0.19</td>
</tr>
<tr>
<td>Fe&lt;sup&gt;3+&lt;/sup&gt;</td>
<td>0.09</td>
<td>0.16</td>
<td>0.16</td>
<td>0.16</td>
<td>0.19</td>
<td>0.18</td>
<td>0.13</td>
</tr>
<tr>
<td>Fe&lt;sup&gt;2+&lt;/sup&gt;</td>
<td>0.54</td>
<td>0.88</td>
<td>0.90</td>
<td>1.12</td>
<td>1.18</td>
<td>1.17</td>
<td>1.35</td>
</tr>
<tr>
<td>Mg</td>
<td>1.92</td>
<td>1.53</td>
<td>1.47</td>
<td>1.21</td>
<td>1.17</td>
<td>1.10</td>
<td>0.98</td>
</tr>
<tr>
<td>Mn</td>
<td>0.01</td>
<td>0.01</td>
<td>0.01</td>
<td>0.01</td>
<td>0.03</td>
<td>0.02</td>
<td>0.03</td>
</tr>
<tr>
<td></td>
<td>2.85</td>
<td>2.78</td>
<td>2.78</td>
<td>2.75</td>
<td>2.79</td>
<td>2.76</td>
<td>2.74</td>
</tr>
<tr>
<td><strong>Large cation</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ca</td>
<td>0.10</td>
<td>0.10</td>
<td>0.13</td>
<td>0.09</td>
<td>0.06</td>
<td>0.04</td>
<td>0.07</td>
</tr>
<tr>
<td>Ba</td>
<td>0.01</td>
<td>0.01</td>
<td>-</td>
<td>0.01</td>
<td>-</td>
<td>0.01</td>
<td>0.01</td>
</tr>
<tr>
<td>Na</td>
<td>0.11</td>
<td>0.05</td>
<td>0.07</td>
<td>0.05</td>
<td>0.14</td>
<td>0.11</td>
<td>0.04</td>
</tr>
<tr>
<td>K</td>
<td>0.77</td>
<td>0.79</td>
<td>0.74</td>
<td>0.83</td>
<td>0.84</td>
<td>0.85</td>
<td>0.86</td>
</tr>
<tr>
<td></td>
<td>0.99</td>
<td>0.95</td>
<td>0.94</td>
<td>0.98</td>
<td>1.04</td>
<td>1.01</td>
<td>0.98</td>
</tr>
<tr>
<td><strong>OH</strong></td>
<td>1.34</td>
<td>1.46</td>
<td>1.98</td>
<td>1.33</td>
<td>1.45</td>
<td>-1.48</td>
<td>1.45</td>
</tr>
</tbody>
</table>

Octahedral charge  
-0.23  +0.27  +0.17  +0.17  +0.18  +0.23  +0.03
Tetrahedral charge -1.32  -1.30  -1.23  -1.22  -1.29  -1.29  -1.10
Inherent layer charge -1.09  -1.03  -1.06  -1.05  -1.11  -1.06  -1.07

Interlayer Charge  
<table>
<thead>
<tr>
<th>Cations</th>
<th>Position</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ca</td>
<td>0.99</td>
</tr>
<tr>
<td>Ba</td>
<td>0.95</td>
</tr>
<tr>
<td>Na</td>
<td>0.94</td>
</tr>
<tr>
<td>K</td>
<td>0.98</td>
</tr>
<tr>
<td></td>
<td>1.04</td>
</tr>
<tr>
<td></td>
<td>1.01</td>
</tr>
<tr>
<td></td>
<td>0.98</td>
</tr>
</tbody>
</table>

*Calculations are based on 44 cationic and anionic equivalents of the idealized unit cell (Foster, 1960), but are shown here in terms of the half-cell.*
Figure 73. Plot of atomic $\text{Mg} - \text{Al} + \text{Fe}^3 + \text{Ti} - \text{Fe}^2 + \text{Mn}$ in the octahedral group, compared to some paragenetic fields (after Foster, 1960).
\[ W = K, Na, Ca, Ba \]
\[ X = Mg, Fe^{2+}, Mn^{2+} \]
\[ Y = Al, Fe^{3+}, Ti, Mn^{3+} \]
\[ Z = Si, Al. \]

Where there is not enough Al to satisfy the Z group, various workers allot either Fe\(^{3+}\) or Ti. Various end members have been proposed for the biotites by Winchell (1935), Stevens (1946), Foster (1960), Eugster and Wones (1962), and others. Various methods of calculation of structural formulae are also used. The method followed here is described by Foster (1960), and the results of the calculations are given in Tables 21 and 22. The structural formulae show that all the biotites have an octahedral occupancy less than 3.00, and are consequently not truly trioctahedral. The octahedral group shows a positive charge, which is a result of substitution of \( R^{2+} \) by \( R^{3+} \) cations. This positive charge is neutralized by the negative charge associated with the vacant octahedral sites and with substitution of Al for Si in the tetrahedral group.

Substitution of Ca\(^{2+}\) and Ba\(^{2+}\) for \( K^+ \) in the large cation group tends to increase the interlayer cationic charge which has been appropriately compensated in these biotites by reduced occupancy. The OH values are appreciably lower than the theoretical 2.00 and this is probably because fluorine was not determined. The ratio of ferric to ferrous iron is low and it is unlikely that the OH deficiency indicates the presence of oxybiotite.

Figure 73 illustrates the variation of the octahedral cations \( R^{3+} - Mg - Fe^{2+} + Mn^{2+} \) for the biotites of this study compared with some
**TABLE 22. Structural formulae* of the analyzed biotites.**

<table>
<thead>
<tr>
<th></th>
<th>Structural Formula</th>
<th>2.68 Al$_{1.32}$</th>
<th>0$_{10}$(OH)$_2$</th>
<th>-1.32</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>[Al$<em>{0.09}$Ti$</em>{0.20}$Fe$<em>{2+}^{3+}$]$</em>{0.09}$</td>
<td>0.54Mg$<em>{1.22}$Mn$</em>{2+}^{2+}$</td>
<td>0.01Si$_{2.68}$</td>
<td>1.09(K, Na, Ca, Ba)$_{2}^{2+}$</td>
</tr>
<tr>
<td></td>
<td>0.03Fe$^{2+}$</td>
<td>2.05</td>
<td>1.04</td>
<td></td>
</tr>
<tr>
<td>B</td>
<td>[Al$<em>{0.08}$Ti$</em>{0.16}$Fe$<em>{2+}^{3+}$]$</em>{0.16}$</td>
<td>0.88Mg$<em>{1.53}$Mn$</em>{2+}^{2+}$</td>
<td>0.01Si$<em>{2.70}$Al$</em>{1.30}$</td>
<td>0$_{10}$(OH)$_2$</td>
</tr>
<tr>
<td></td>
<td>1.18Fe$^{3+}$</td>
<td>1.06</td>
<td></td>
<td></td>
</tr>
<tr>
<td>C</td>
<td>[Al$<em>{0.05}$Ti$</em>{0.21}$Fe$<em>{2+}^{3+}$]$</em>{0.21}$</td>
<td>0.90Mg$<em>{1.47}$Mn$</em>{2+}^{2+}$</td>
<td>0.01Si$<em>{2.77}$Al$</em>{1.23}$</td>
<td>0$_{10}$(OH)$_2$</td>
</tr>
<tr>
<td></td>
<td>1.10Fe$^{3+}$</td>
<td>1.06</td>
<td></td>
<td></td>
</tr>
<tr>
<td>D</td>
<td>[Al$<em>{0.04}$Ti$</em>{0.21}$Fe$<em>{2+}^{3+}$]$</em>{0.21}$</td>
<td>0.90Mg$<em>{1.21}$Mn$</em>{2+}^{2+}$</td>
<td>0.01Si$<em>{2.78}$Al$</em>{1.22}$</td>
<td>0$_{10}$(OH)$_2$</td>
</tr>
<tr>
<td></td>
<td>1.10Fe$^{3+}$</td>
<td>1.06</td>
<td></td>
<td></td>
</tr>
<tr>
<td>E</td>
<td>[Al$<em>{0.01}$Ti$</em>{0.21}$Fe$<em>{2+}^{3+}$]$</em>{0.21}$</td>
<td>0.19Mg$<em>{1.17}$Mn$</em>{2+}^{2+}$</td>
<td>0.03Si$<em>{2.71}$Al$</em>{1.29}$</td>
<td>0$_{10}$(OH)$_2$</td>
</tr>
<tr>
<td></td>
<td>1.10Mn$^{2+}$</td>
<td>1.04</td>
<td></td>
<td></td>
</tr>
<tr>
<td>F</td>
<td>[Al$<em>{0.03}$Ti$</em>{0.26}$Fe$<em>{2+}^{3+}$]$</em>{0.26}$</td>
<td>0.18Mg$<em>{1.10}$Mn$</em>{2+}^{2+}$</td>
<td>0.02Si$<em>{2.71}$Al$</em>{1.29}$</td>
<td>0$_{10}$(OH)$_2$</td>
</tr>
<tr>
<td></td>
<td>1.10Mn$^{2+}$</td>
<td>1.01</td>
<td></td>
<td></td>
</tr>
<tr>
<td>G</td>
<td>[Al$<em>{0.06}$Ti$</em>{0.19}$Fe$<em>{2+}^{3+}$]$</em>{0.19}$</td>
<td>0.13Mg$<em>{0.98}$Mn$</em>{2+}^{2+}$</td>
<td>0.03Si$<em>{2.90}$Al$</em>{1.10}$</td>
<td>0$_{10}$(OH)$_2$</td>
</tr>
</tbody>
</table>

*Structural formulae are of the half-cell, using the notation of Foster (1960).
of the paragenetic groups shown by Foster (1960, Figure 12). The variation in the octahedral cations is dominated by substitution of Mg by \( \text{Fe}^{2+} + \text{Mn}^{2+} \), with only small changes in \((\text{Al} + \text{Fe}^{3+} + \text{Ti})\). The plots of the biotites in Figures 72 and 73 are nearly superposable, and this is partly a result of the fact that only a small amount of Al appears in six-fold co-ordination.

(e) **Summary.**

The biotites separated from some members of the monzonitic complex at Mt Dromedary range in composition from Mg-dominant to weakly Fe\(^{2+}\)-dominant. Their content of titania is fairly high and relatively constant, and they contain considerable amounts of lime, baria and soda. The most important compositional trend is that of substitution of Mg by Fe\(^{2+}\).

The composition of the biotites is closely related to the bulk composition of the host-rock and does not appear to be influenced by the accompanying mafic minerals. The relations between chemical composition, optical properties and paragenesis are similar to those proposed for igneous biotites by previous workers.
(a) Introduction.

Amphibole was found in greater or smaller amount in virtually every thin section examined of the rocks of the monzonitic complex. It occurs in significant amounts in the nepheline monzonites around Little Dromedary, in the 'ijolite', in some phases of the monzonite at Mt Dromedary (e.g. RB120, 12 vol.%), and in the banatite, especially the porphyritic marginal phase.

Three samples of amphibole, one from each of the nepheline monzonite, 'ijolite' and banatite, were separated and analyzed, and their optical properties and specific gravities were determined. The analyzed samples most probably do not embrace the range of composition of all the amphiboles in the complex, there being less soda-rich types in the pyroxenite at Poole's Point, and more soda-rich types in highly alkaline phases of the nepheline monzonite, but they are probably representative of the bulk of the amphibole of the complex, and they allow a comparison of the minerals from under- and over-saturated alkaline rocks.

(b) Nomenclature.

The classification and nomenclature of the lime-rich amphiboles has been studied recently by Hallimond (1943), Winchell (1945), Smith (1959), Wilkinson (1961), and others. The general formula for the lime-soda amphiboles can be written $W_{2-3}(XY)_{5}Z_{2}O_{22}(OH)$, where
W = Ca, Na, K
X = Mg, Fe\(^{2+}\), Mn\(^{2+}\)
Y = Al, Fe\(^{3+}\), Ti
Z = Al, Si.

The four generally accepted end-members and their iron analogues for those amphiboles with Ca = 2 (atoms) are (after Sundius, 1946):

(i) Tremolite → Actinolite
\[ \text{Ca}_2\text{Mg}_5\text{Si}_8\text{O}_{22}(\text{OH})_2 \]

(ii) Pargasite → Hastingsite
\[ \text{Ca}_2\text{Na}_2\text{Mg}_4\text{Al}_3\text{Si}_6\text{O}_{22}(\text{OH})_2 \]

(iii) Edenite → Ferro-edenite
\[ \text{Ca}_2\text{Na}_2\text{Mg}_5\text{Al}_7\text{Si}_6\text{O}_{22}(\text{OH})_2 \]

(iv) Tschermakite → Ferro-tschermakite
\[ \text{Ca}_2\text{Mg}_3\text{Al}_4\text{Si}_6\text{O}_{22}(\text{OH})_2 \]

(c) The chemical analyses.

The chemical analyses of the three amphiboles are given in Table 23a, and the atomic ratios calculated on the basis of 24(O, OH, F) are given in Table 23b. In each of the three, Si is very close to 6.00 and the Z group is brought to 8.00 by Al\(^{IV}\), leaving a small amount of Al\(^{IV}\) in the Y group. Ca fills nearly 2.00 of the large cation sites in RB188Hb and RB108Hb but falls to the rather low level of 1.71 sites in RB432Hb.

In bulk chemistry the three amphiboles are very similar, with the most noticeable variation taking place in the moderate amounts of titania and in the total alkalies. The two samples from the nepheline-
Figure 74. Nepheline monzonite from same locality as 'covite' analyzed by Brown (1930, p.659), showing deep green amphibole (RB432Hb) enclosing pale green pyroxene. Note abundant inclusions of apatite in pyroxene. Plane light. Scale mark is 1 mm in length.
bearing rocks differ appreciably in their content of silica, titania and total iron. The sample from the banatite is distinguished by a relatively low amount of alkalies and higher titania. Its ratio of \((\text{FeO} + \text{MnO})/\text{MgO}\) is 1.78, compared to 1.10 for both the amphiboles from the undersaturated rocks. \(\text{K}_2\text{O}\) is rather high compared to \(\text{Na}_2\text{O}\) in RB432Hb and RB188Hb.

The three amphiboles are titaniferous members of the pargasite-hastingsite series. The two samples from the undersaturated rocks are very nearly midway between the two end members, having about equal amounts of \(\text{Mg}^{2+}\) and \(\text{Fe}^{2+}\), while the amphibole from the oversaturated rock is more hastingsitic.

Billings (1928) believes that the dominant substitution in the hastingsite group is \(\text{Fe}^{2+}\) for \(\text{Mg}^{2+}\) and accompanies falling temperature and progressive differentiation of the host-rock and accordingly, RB108Hb should represent a later stage of differentiation relative to the other amphiboles. This appears to be true in this case. The mafic indices of amphibole and rock are closely similar for each of RB432Hb, RB188Hb and RB108Hb. Larsen and Draisen (1950) have noted a similar relation for rocks and minerals of the southern California batholith.

(d) **Optical properties, specific gravities.**

Table 24 contains a summary of the optical properties and the specific gravities of the analyzed specimens.

The lack of obvious correlation between chemical and optical properties has been recognized by many workers (Buddington and Leonard,
## TABLE 23a. Chemical analyses of amphiboles.

<table>
<thead>
<tr>
<th></th>
<th>RB432Hb</th>
<th>RB188Hb</th>
<th>RB108Hb</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>39.17</td>
<td>37.65</td>
<td>38.70</td>
</tr>
<tr>
<td>TiO₂</td>
<td>2.04</td>
<td>1.49</td>
<td>2.36</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>13.19</td>
<td>12.44</td>
<td>13.45</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>5.50</td>
<td>8.32</td>
<td>3.85</td>
</tr>
<tr>
<td>FeO</td>
<td>15.45</td>
<td>14.67</td>
<td>18.25</td>
</tr>
<tr>
<td>MnO</td>
<td>0.52</td>
<td>0.44</td>
<td>0.87</td>
</tr>
<tr>
<td>MgO</td>
<td>8.09</td>
<td>7.61</td>
<td>6.00</td>
</tr>
<tr>
<td>CaO</td>
<td>10.30</td>
<td>12.06</td>
<td>12.07</td>
</tr>
<tr>
<td>Na₂O</td>
<td>2.31</td>
<td>2.07</td>
<td>1.76</td>
</tr>
<tr>
<td>K₂O</td>
<td>2.02</td>
<td>2.08</td>
<td>0.96</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.04</td>
<td>0.02</td>
<td>0.05</td>
</tr>
<tr>
<td>H₂O</td>
<td>1.27</td>
<td>1.27</td>
<td>1.92</td>
</tr>
<tr>
<td>Cr₂O₃</td>
<td>tr.</td>
<td>n.d.</td>
<td>0.01</td>
</tr>
<tr>
<td>Total</td>
<td>99.90</td>
<td>100.22</td>
<td>100.25</td>
</tr>
</tbody>
</table>

## TABLE 23b. Analyses of Table 23a recalculated on the basis of 24(O,OH,F).

<table>
<thead>
<tr>
<th></th>
<th>RB432Hb</th>
<th>RB188Hb</th>
<th>RB108Hb</th>
</tr>
</thead>
<tbody>
<tr>
<td>Z(Al)</td>
<td>604</td>
<td>590</td>
<td>596</td>
</tr>
<tr>
<td>Z(AIV)</td>
<td>196</td>
<td>210</td>
<td>204</td>
</tr>
<tr>
<td>Y(AlV)</td>
<td>44</td>
<td>20</td>
<td>42</td>
</tr>
<tr>
<td>Y(Ti)</td>
<td>24</td>
<td>18</td>
<td>27</td>
</tr>
<tr>
<td>Y(Fe³⁺)</td>
<td>63</td>
<td>98</td>
<td>44</td>
</tr>
<tr>
<td>X(Mn)</td>
<td>200</td>
<td>192</td>
<td>235</td>
</tr>
<tr>
<td>X(Mg)</td>
<td>7</td>
<td>6</td>
<td>10</td>
</tr>
<tr>
<td>W(Ca)</td>
<td>187</td>
<td>180</td>
<td>138</td>
</tr>
<tr>
<td>W(Na)</td>
<td>171</td>
<td>202</td>
<td>200</td>
</tr>
<tr>
<td>W(K)</td>
<td>69</td>
<td>62</td>
<td>52</td>
</tr>
<tr>
<td>W(H)</td>
<td>39</td>
<td>41</td>
<td>19</td>
</tr>
<tr>
<td>W</td>
<td>132</td>
<td>134</td>
<td>198</td>
</tr>
</tbody>
</table>

RB432Hb - Amphibole from 'covite', south-eastern slope of Little Dromedary. Same locality as rock analyzed by Brown (1930, p.650).

RB188Hb - Amphibole from 'ijolite', 3 miles south-south-east of Mt Dromedary Trig.

RB108Hb - Amphibole from porphyritic banatite, ½ mile north of Tilba Tilba. Same locality as rock analyzed by Brown (1930, p.647).

Analyses arranged in order of decreasing MgO.
TABLE 24. Optical properties and specific gravities of amphiboles in Table 23.

<table>
<thead>
<tr>
<th>Refractive indices</th>
<th>RB432Hb</th>
<th>RB188Hb</th>
<th>RB108Hb</th>
</tr>
</thead>
<tbody>
<tr>
<td>α</td>
<td>1.673</td>
<td>1.675</td>
<td>1.665</td>
</tr>
<tr>
<td>β</td>
<td>1.695</td>
<td>1.702</td>
<td>1.687</td>
</tr>
<tr>
<td>δ</td>
<td>1.698</td>
<td>1.703</td>
<td>1.694</td>
</tr>
</tbody>
</table>

Fleuchroism

<table>
<thead>
<tr>
<th></th>
<th>RB432Hb</th>
<th>RB188Hb</th>
<th>RB108Hb</th>
</tr>
</thead>
<tbody>
<tr>
<td>X</td>
<td>Yellow-brown</td>
<td>Yellow-brown</td>
<td>Pale yellow</td>
</tr>
<tr>
<td>Y</td>
<td>Yellow-brown</td>
<td>Yellow-brown, green tint</td>
<td>Light brown</td>
</tr>
<tr>
<td>Z</td>
<td>Mid to dark green</td>
<td>Deep olive-green</td>
<td>Mid green</td>
</tr>
</tbody>
</table>

2V (-ve) 30°-33°
Z°C 22°-26°

Specific gravity

<table>
<thead>
<tr>
<th></th>
<th>RB432Hb</th>
<th>RB188Hb</th>
<th>RB108Hb</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.786</td>
<td>2.812</td>
<td>2.797</td>
<td></td>
</tr>
</tbody>
</table>

Molecular (FeO + MnO)/MgO

<table>
<thead>
<tr>
<th></th>
<th>RB432Hb</th>
<th>RB188Hb</th>
<th>RB108Hb</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.10</td>
<td>1.10</td>
<td>1.78</td>
<td></td>
</tr>
</tbody>
</table>
In particular, the influence on the optical properties of TiO$_2$ is likely to be considerable (cf. titaniferous clinopyroxenes - Wilkinson, 1957) since in many kaersutites titania is 8-10% (Yagi, 1953, Table 8; Wilkinson, 1961, Table 3). With regard to the hastingsite group, Billings (1928, p.292) states that increase in magnesia results in an increase in 2V and a decrease in the principal refractive indices and the specific gravity. The properties of the amphiboles listed in Table 24 do not show these relations.

(e) **Comparison with amphiboles from similar occurrences.**

An account of the amphiboles from the Highwood Mountains is given by Larsen *et al.* (1941) and an analysis of a barkevikite is quoted from Lindgren. The mineral is not very similar to those from Mt Dromedary as it is much lower in MgO, and higher in Al$_2$O$_3$ and total iron. The amphiboles from the syenitic rocks at the Highwood Mountains have similar optical properties to those given herein (Larsen *et al.*, 1941, p.1842).

Pecora and Fisher (1946) describe an amphibole from vugs in a monzonitic dyke in the Bearpaw Mountains. The analysis and optical properties of this sample compare closely with those from the rocks at Mt Dromedary (Table 23). Two hornblendes from the Port Cygnet alkali complex (Edwards, 1947) are included in Table 25 for comparison.

The factors likely to affect the particular type of amphibole crystallizing from alkalic undersaturated rocks have been discussed by Wilkinson (1961) and to a lesser extent by Billings (1928). It should be noted that the two amphiboles separated from the nepheline-bearing rocks
<table>
<thead>
<tr>
<th></th>
<th>A</th>
<th>B</th>
<th>C</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>38.92</td>
<td>39.50</td>
<td>39.20</td>
</tr>
<tr>
<td>TiO₂</td>
<td>1.51</td>
<td>0.25</td>
<td>1.39</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>12.68</td>
<td>17.45</td>
<td>16.06</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>7.42</td>
<td>9.60</td>
<td>11.42</td>
</tr>
<tr>
<td>FeO</td>
<td>14.64</td>
<td>11.52</td>
<td>8.80</td>
</tr>
<tr>
<td>MnO</td>
<td>0.09</td>
<td>0.15</td>
<td>0.72</td>
</tr>
<tr>
<td>MgO</td>
<td>7.86</td>
<td>7.60</td>
<td>3.76</td>
</tr>
<tr>
<td>CaO</td>
<td>11.24</td>
<td>12.20</td>
<td>15.50</td>
</tr>
<tr>
<td>Na₂O</td>
<td>2.38</td>
<td>n.d.</td>
<td>1.33</td>
</tr>
<tr>
<td>K₂O</td>
<td>1.58</td>
<td>n.d.</td>
<td>1.82</td>
</tr>
<tr>
<td>H₂O⁺</td>
<td>0.93</td>
<td>n.d.</td>
<td>n.d.</td>
</tr>
<tr>
<td>H₂O⁻</td>
<td>n.d.</td>
<td>n.d.</td>
<td>n.d.</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>n.d.</td>
<td>0.15</td>
<td>0.61</td>
</tr>
<tr>
<td>F</td>
<td>1.20</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Less O for F</td>
<td>0.51</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Total</td>
<td>99.94</td>
<td>98.42</td>
<td>100.69</td>
</tr>
</tbody>
</table>


at Mt Dromedary are green, as opposed to the brown barkevikites often found in alkalic rocks, especially those of alkali basalt parentage.

(5) The Garnets

(a) Introduction.

Garnets were found in, or associated with, only the undersaturated alkalic rocks, with one exception, namely the basic volcanic block enclosed by the monzonite on the west slope of Mt Dromedary (RB112A; rock RB114 analyzed, see Table 29). The garnet is most abundant in some phases of the garnet-pyroxene rocks, reaching 58 vol. per cent (RB36]). The 'ijolite' and nepheline monzonite have only accessory amounts, and it is very frequently absent in thin sections of the latter rock. The garnet-bearing dykes cropping out on the coast half a mile north of Poole's Point contain up to about 14 vol. per cent of garnet (RB16).

Three garnet samples were obtained, one from each of the garnet-pyroxene rock RB363A (garnet RB363AG), the 'ijolite' RB188 (garnet RB188G) and the foyaitic dyke RB16 (garnet RB16G). The samples were analyzed and their specific gravities and optical properties were determined.

Some observations were made on the garnet in the garnet-pyroxene rock using the electron micro-analyzer. The cell edge of RB16G was determined by means of X-ray powder photography.

Optically, the garnets observed in the rocks of the complex are similar, and the analyzed samples are thought to be representative of the
Figure 75. Thin section (RB363A) showing garnet with color zoning in patches and lamellae (A), and very fine zoning marked by differences in refractive index (B). Plane light. Scale mark is 0.2 mm in length.
(b) **Optical properties of the garnets.**

The form and color of the garnets vary from one rock-type to another. In the foyaitic dyke the grains of garnet are very well formed, and the grain boundaries are sharp, with only occasional rimming by dark green amphibole. The grains reach about 1.0 mm in diameter, and usually the crystals are made up of concentric zones of alternating lighter and darker colors. The color of the garnet varies from pale pink or brown to a strong brownish-pink.

The garnet in the 'ijolite' is subhedral to euhedral, pale pinkish-brown, and free from alteration. It shows better formed faces against the felsic minerals and can be seen apparently replacing the pyroxene. Color zoning is slight and there are few inclusions.

The nepheline monzonite has very little garnet, and it is found in subhedral grains, up to 0.5 mm, which are pale brown in color. Inclusions are rare, and the mineral appears not to have particular associates, except perhaps plagioclase. Zoning is absent.

The rare grains of garnet found in the large xenolith on the west slopes of Mt Dromedary reach 1 cm in diameter, and they are xenoblastic. They are pale pinkish-brown and free from alteration. Patchy color zoning occurs. The minerals associated with the garnet include calcite, quartz, diopside, a bright green clinoamphibole (Z°C = 28°), and opaques.

The garnet in the garnet-pyroxene rocks and associated veins is
<table>
<thead>
<tr>
<th>Chemical analyses of garnets</th>
<th>RB363AG</th>
<th>RB188G</th>
<th>RB16G</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>36.04</td>
<td>35.66</td>
<td>36.32</td>
</tr>
<tr>
<td>TiO₂</td>
<td>1.34</td>
<td>2.60</td>
<td>3.05</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>7.19</td>
<td>9.03</td>
<td>6.57</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>20.35</td>
<td>17.36</td>
<td>18.29</td>
</tr>
<tr>
<td>FeO</td>
<td>1.31</td>
<td>1.89</td>
<td>2.03</td>
</tr>
<tr>
<td>MnO</td>
<td>0.30</td>
<td>0.48</td>
<td>0.18</td>
</tr>
<tr>
<td>MgO</td>
<td>0.50</td>
<td>0.57</td>
<td>2.25</td>
</tr>
<tr>
<td>CaO</td>
<td>33.27</td>
<td>32.18</td>
<td>30.74</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.11*</td>
<td>0.12</td>
<td>0.16</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.05*</td>
<td>0.08</td>
<td>0.14</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>n.d.</td>
<td>0.01</td>
<td>n.d.</td>
</tr>
<tr>
<td>H₂O⁺</td>
<td>n.d.</td>
<td>0.19</td>
<td>n.d.</td>
</tr>
<tr>
<td>Cr₂O₃</td>
<td>n.d.</td>
<td>0.02</td>
<td>n.d.</td>
</tr>
<tr>
<td>Total</td>
<td>100.46</td>
<td>100.19</td>
<td>99.73</td>
</tr>
</tbody>
</table>

RB363AG - Garnet from garnet-pyroxene rock, 1 mile south-east of Central Tilba.
RB188G - Garnet from 'ijolite', near the highway, 2 miles south of Mt Dromedary Trig.
RB16G - Garnet from garnet-bearing foyaitic dyke, ½ mile north of Poole's Point.

Other analyses by R.S. Boesen.
pale pink to a fairly strong red-brown. The form of the mineral is typically xenoblastic, and it replaces pyroxene and plagioclase forming lacy networks through the rock. The size of the grains is very variable, ranging up to 2-3 cm. Color zoning is of two kinds, one showing in irregular patches with rapid, but not discontinuous, gradations into lighter or darker portions; the other appears as fine, lamellar repetitions of lighter and darker zones very similar in type and scale to the lamellar zoning in the clinopyroxenes. In addition, the garnets may show an onion-skin structure parallel to grain boundaries. This is not delineated by color variations, but by the Becke line effect. The origin of this type of zoning is not known. Figure 75 shows examples of these zone effects. Refractive index variation across the zones is slight.

The garnets of the 'ijolite' and xenolith are quite isotropic in thin section, but the remaining garnets usually show slight birefringence. In outstanding cases, the birefringence reaches about 0.005, and the grains appear to be made up of lamellae and sections of differing optic orientation. Thin section RB141 (Figure 56) contains garnet which shows twinning and grain structure clearly palimpsest after plagioclase. 2E's of about \(-86^\circ\) have been noted in several sections.

The indices of refraction commonly found are similar to those in Table 27.

(c) Chemical analyses.

The chemical analyses of the three garnet samples are given in
TABLE 27. Atomic ratios*, composition, and some properties of analyzed garnets.

<table>
<thead>
<tr>
<th>Atomic Ratios</th>
<th>RB363AG</th>
<th>RB188G</th>
<th>RB16G</th>
</tr>
</thead>
<tbody>
<tr>
<td>Si</td>
<td>294</td>
<td>289</td>
<td>296</td>
</tr>
<tr>
<td>Ti</td>
<td>6</td>
<td>11</td>
<td>4</td>
</tr>
<tr>
<td>Ti</td>
<td>2</td>
<td>5</td>
<td>15</td>
</tr>
<tr>
<td>Al</td>
<td>69</td>
<td>87</td>
<td>63</td>
</tr>
<tr>
<td>Fe&lt;sup&gt;3+&lt;/sup&gt;</td>
<td>124</td>
<td>106</td>
<td>112</td>
</tr>
<tr>
<td>Fe&lt;sup&gt;2+&lt;/sup&gt; + Mn</td>
<td>11</td>
<td>16</td>
<td>15</td>
</tr>
<tr>
<td>Mg</td>
<td>6</td>
<td>7</td>
<td>26</td>
</tr>
<tr>
<td>Ca</td>
<td>291</td>
<td>279</td>
<td>269</td>
</tr>
<tr>
<td>(Si + Ti)</td>
<td>300</td>
<td>300</td>
<td>300</td>
</tr>
<tr>
<td>R&lt;sup&gt;3&lt;/sup&gt;</td>
<td>1.97</td>
<td>1.98</td>
<td>1.90</td>
</tr>
<tr>
<td>R&lt;sup&gt;2&lt;/sup&gt;</td>
<td>3.08</td>
<td>3.02</td>
<td>3.10</td>
</tr>
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</table>

Composition in mol. per cent:

<table>
<thead>
<tr>
<th></th>
<th>RB363AG</th>
<th>RB188G</th>
<th>RB16G</th>
</tr>
</thead>
<tbody>
<tr>
<td>Andradite</td>
<td>63.6</td>
<td>59.6</td>
<td>58.8</td>
</tr>
<tr>
<td>Grossularite</td>
<td>29.2</td>
<td>29.2</td>
<td>18.4</td>
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<tr>
<td>Titanium garnet</td>
<td>1.0</td>
<td>2.6</td>
<td>7.9</td>
</tr>
<tr>
<td>Almandine</td>
<td>4.1</td>
<td>6.0</td>
<td>5.4</td>
</tr>
<tr>
<td>(+ spessartite)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pyrope</td>
<td>2.1</td>
<td>2.6</td>
<td>9.5</td>
</tr>
<tr>
<td>Excess</td>
<td>Ca&lt;sub&gt;2&lt;/sub&gt;(SiTi)</td>
<td>Ca&lt;sub&gt;2&lt;/sub&gt;(SiTi)</td>
<td>Ca&lt;sub&gt;2&lt;/sub&gt;(SiTi)</td>
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<tr>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>Refractive index</td>
<td>1.830</td>
<td>1.828</td>
<td>1.828</td>
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<tr>
<td>Specific gravity</td>
<td>3.702</td>
<td>3.682</td>
<td>3.714</td>
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<tr>
<td>Cell edge (Å)</td>
<td></td>
<td></td>
<td>11.98</td>
</tr>
</tbody>
</table>

*Atomic ratios on the basis of 12(0).*
Table 26, and their atomic ratios and some optical and physical data are given in Table 27.

Chemically, the three garnets are remarkably similar, and perhaps the more so because of their differing modes of occurrence. RB363AG has been formed through the action of hydrothermal fluids on pyroxenite, RB188G appears to be of late or very late crystallization in a rather mafic alkalic rock, while RB 16G has the features of a mineral of primary crystallization, and occurs in a felsic dyke.

The more significant variations in the analyses concern titania, which is more abundant in the dyke-garnet; magnesia, which is very low in RB363AG and RB188G and notably higher in RB16G; and lime, which shows a close inverse relation to magnesia. The amount of total iron is relatively constant, and the ratio FeO/Fe₂O₃ is very low. Manganese does not follow ferrous iron as might be expected from a consideration of the similarity in charge, ionic radius and electronegativity of the two ions (Wager and Mitchell, 1951; Ringwood, 1955). This apparent lack of relation could be due to inaccuracies in the determination of ferrous iron.

Alumina varies in the same sense as manganese.

The accepted formula of the garnet group may be written

\[ 3R''O.R''''O₃ \cdot 3SiO₂ \], where \( R'' \) = divalent metal ions, usually Ca, Mg, Fe and Mn; and \( R'''' \) = trivalent metal ions, usually Al, Cr, Fe (± Mn³⁺) and Ti.

In the event of a deficiency in silica, Ti and Al may be apportioned to fill the group. Theoretically the ionic proportions of the three groups should be 3:2:3, and Tröger (1959, p.2) advocates redistribution of Fe₂O₃ to bring the R groups to a ratio of 2:3, pointing
out that errors in determining ferrous and ferric iron are probably a source of departure from the ideal. The deficiency in $R''$ in apparently excellent analyses of garnets is very common (A.J. Easton, pers. comm.) and may reflect a real structural feature, rather than analytical error.

The three garnets are dominantly andradite-grossular (molecular compositions). The agreement between the relations of composition, refractive index and specific gravity of the garnets with those of Winchell (1958) is quite good. The cell edge of RB16G was determined from X-ray powder photographs.

The zoning of the garnets was studied on a sample from the garnet-pyroxene rock, using the electron micro-analyzer. The procedure was identical with that followed during the examination of the pyroxene zoning. The lamellar zones were beyond the limit of resolution of the instrument, and a series of scintillation pictures was obtained for Ti, Al, Mg, Cr, Mn and Fe over an area of high color contrast. Surprisingly, the results indicated only very small variations in element abundances. The pink and brown coloration of the garnets has been widely ascribed to titania, and although the darker garnet is richer in that element (and also in Al and Mn), the difference is very slight compared to the change in color. The difference between the zones may lie as much in the oxidation states of some elements as in their actual abundances.

(d) Relation of composition to rock type.

Several attempts have been made to correlate compositions and parageneses for the garnet group (Ford, 1915; Heritsch, 1927; Wright, 1938).
<table>
<thead>
<tr>
<th></th>
<th>RB16G</th>
<th>A</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>36.32</td>
<td>36.00</td>
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<tr>
<td>TiO₂</td>
<td>3.05</td>
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<td>Al₂O₃</td>
<td>6.57</td>
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<tr>
<td>Fe₂O₃</td>
<td>18.29</td>
<td>17.07</td>
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<tr>
<td>FeO</td>
<td>2.03</td>
<td>4.17</td>
</tr>
<tr>
<td>MnO</td>
<td>0.18</td>
<td>1.24</td>
</tr>
<tr>
<td>MgO</td>
<td>2.25</td>
<td>0.34</td>
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<tr>
<td>CaO</td>
<td>30.74</td>
<td>28.80</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.16</td>
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</tr>
<tr>
<td>K₂O</td>
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<tr>
<td>P₂O₅</td>
<td>n.d.</td>
<td>0.72</td>
</tr>
<tr>
<td>Total</td>
<td>99.73</td>
<td>98.08</td>
</tr>
</tbody>
</table>

**RB16G** - From foyaitic dyke, half a mile north of Poole's Point. Analyst A.J. Easton.

The garnets studied by these and other workers were selected partly on the basis of their relatively simple constitution and the usefulness of the results is accordingly limited. A considerably more detailed study was made by Tröger (1959), who defined 28 paragenetic groups and gave estimates of within-group variation.

The garnets of this study compare closely with the foyaite-ijolite-melteigite and phonolite paragenetic groups (Tröger, 1959, p.11; columns V and VII). Garnets from metamorphosed mafic and ultramafic rocks have distinctly different compositions; they are richer in pyrope, almandine and spessartite, and poorer in titania. This indicates that the composition of the garnet in the garnet-pyroxene rocks was influenced far more by the composition of the hydrothermal fluids effecting the transformation than by the composition of the rock being transformed.

The occurrence of melanite (titan-andraditic garnet) is more common in those undersaturated alkalic rocks which have soda dominant over potash, and alkalies in excess of alumina (e.g. Loch Borolan, Sekukuniland, Iwaara, etc.). The alkalic rocks similar to those at Mt Dromedary, which have soda and potash about equal and soda plus potash less than alumina, have not attracted the same amount of attention from geologists; data on the garnets of the intrusive rocks are scanty. Larsen et al. (1941, p.1854) noted the occurrence in the Middle Peak stock and in a few other rocks of a brown melanite garnet with a refractive index of 1.82. Edwards (1947, p.110; Table 6) gives the analysis of a garnet from a hauyne-sanidine-garnet porphyry dyke at Port Cygnet, Tasmania; this is compared in Table 28 to the garnet from the foyaitic dyke (RB16G).
PART IV

PETROCHEMISTRY AND PETROGENESIS
<table>
<thead>
<tr>
<th>Element</th>
<th>Rb</th>
<th>Sr</th>
<th>Ba</th>
<th>Ca</th>
<th>Mg</th>
<th>Fe</th>
<th>Mn</th>
<th>SiO2</th>
<th>TiO2</th>
<th>Al2O3</th>
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<th>Total</th>
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<td>52.39</td>
<td>63.64</td>
<td>64.04</td>
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<td>1.00</td>
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<td>Cl02</td>
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<td>17.65</td>
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<td>18.00</td>
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<td>0.15</td>
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<td>0.15</td>
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</tr>
<tr>
<td>V</td>
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<td>0.15</td>
<td>0.11</td>
<td>0.69</td>
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<td>0.10</td>
<td>0.17</td>
<td>0.19</td>
</tr>
<tr>
<td>Mn</td>
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<td>0.02</td>
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<td>0.02</td>
<td>0.02</td>
<td>0.02</td>
<td>0.02</td>
<td>0.02</td>
</tr>
<tr>
<td>Total</td>
<td>99.91</td>
<td>100.26</td>
<td>99.72</td>
<td>100.49</td>
<td>100.22</td>
<td>100.21</td>
<td>100.29</td>
<td>100.31</td>
<td>99.99</td>
<td>100.43</td>
<td>99.80</td>
<td>100.13</td>
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</table>

Other elements searched for but not found: Be, La, Mo, Nd, U, W, Y.

@Analyst: Avery and Anderson. *Analyst: J. R. Martin.
All other analyses: R. S. Boesen.
### Key to Table 29.

<table>
<thead>
<tr>
<th>Code</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>RB27</td>
<td>Laminated feldspathic pyroxenite, Poole's Point.</td>
</tr>
<tr>
<td>RB402</td>
<td>Laminated feldspathic pyroxenite, Poole's Point; less pyroxene and olivine.</td>
</tr>
<tr>
<td>RB412</td>
<td>Coarse massive pyroxenite, Poole's Point.</td>
</tr>
<tr>
<td>RB363</td>
<td>Pyroxenite associated with garnet-pyroxene rock, 1 mile east of Central Tilba.</td>
</tr>
<tr>
<td>RB338</td>
<td>Pyroxenite associated with garnet-pyroxene rock, 1 mile east of Central Tilba; apparently quite unaltered.</td>
</tr>
<tr>
<td>Drom</td>
<td>Pyroxenite from core, 600 feet below sea level; east shore of Tilba Tilba Lake.</td>
</tr>
<tr>
<td>RB64</td>
<td>Monzonite, near pyroxenite, Poole's Point.</td>
</tr>
<tr>
<td>RB331</td>
<td>Monzonitic satellite, 1½ miles east-north-east of Central Tilba.</td>
</tr>
<tr>
<td>RB221</td>
<td>Monzonite ('black granite'), Central Tilba quarry.</td>
</tr>
<tr>
<td>RB227</td>
<td>Leuocratic quartz-feldspar intrusion, 1½ miles east of Central Tilba.</td>
</tr>
<tr>
<td>RB351</td>
<td>Banatite, Mt Dromedary Trig.</td>
</tr>
<tr>
<td>RB323</td>
<td>Andesitic lava, 1½ miles east of Central Tilba.</td>
</tr>
<tr>
<td>RB407</td>
<td>Basic dyke, Poole's Point.</td>
</tr>
<tr>
<td>RB12A</td>
<td>Hybrid dyke, Poole's Point.</td>
</tr>
<tr>
<td>RB7</td>
<td>Dolerite dyke, 1 mile north of Poole's Point.</td>
</tr>
<tr>
<td>RB16</td>
<td>Garnet-bearing foyaitic dyke, ½ mile north of Poole's Point.</td>
</tr>
<tr>
<td>RB1</td>
<td>Camptonitic dyke, 3¼ mile north of Poole's Point.</td>
</tr>
<tr>
<td>RB417</td>
<td>Feldspathic vein in pyroxenite, Poole's Point.</td>
</tr>
<tr>
<td>RB10</td>
<td>Leuocratic quartz-feldspar dyke, 3¼ mile north of Poole's Point.</td>
</tr>
<tr>
<td>RB188</td>
<td>'Ijolite', near highway, 3 miles south-east of Mt Dromedary Trig.</td>
</tr>
<tr>
<td>RB350</td>
<td>Shonkinite, Tilba Tilba village.</td>
</tr>
<tr>
<td>RB424</td>
<td>Nepheline monzonite, same location as the covite of Brown (1930), 1½ miles east of Tilba Tilba village.</td>
</tr>
<tr>
<td>RB114</td>
<td>Recrystallized basaltic xenolith, 2 miles west-north-west of Mt Dromedary Trig.</td>
</tr>
<tr>
<td>RB201</td>
<td>Recrystallized biotite-rich xenolith, Central Tilba quarry.</td>
</tr>
<tr>
<td>RB472</td>
<td>Buff colored pelitic shale (Wagonga Beds), 1½ miles north-east of Central Tilba.</td>
</tr>
<tr>
<td>RB473</td>
<td>The same, but slightly coarser and with more quartz.</td>
</tr>
<tr>
<td></td>
<td>C.I.P.W. norms of analyses in Table 29*</td>
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<tr>
<td>----------</td>
<td>---------------------------------------</td>
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<tr>
<td></td>
<td>RB27</td>
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<td>Quarts</td>
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<td>Ilmenite</td>
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<td>Apatite</td>
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<tr>
<td>Calcite</td>
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<tr>
<td>Felsic Index</td>
<td>8.1</td>
</tr>
<tr>
<td>Mafic Index</td>
<td>68.6</td>
</tr>
<tr>
<td>Anorthite content of normative plagioclase</td>
<td>92.9</td>
</tr>
</tbody>
</table>

*Corundum.  **Wollastonite.  ***Calcium orthosilicate.

*Excluding sediments and xenoliths.
Brown (1930) described the petrogenesis of the complex at Mt Dromedary in three main parts, dealing respectively with the monzonitic, the nepheline-bearing and the melanite-bearing rock series.

The major units of the 'normal monzonitic' series were pyroxenite (Poole's Point), monzonite, porphyritic banatite and banatite, all of which were fractionated in situ from monzonitic magma by means of simple crystal settling. The nepheline-bearing rocks were thought to be formed by either segregation and 'secondary differentiation' of a rest fluid from the crystallizing monzonitic rocks (p.689) or by desilication by incorporation of limestone (pp.690, 691). The origin of the garnet-bearing rocks was left to some even more obscure process, possibly related to 'secondary differentiation' of a phase of the pyroxenite or to incorporation of limestone.

The present study has shown that the structure of the complex, at least at its present level of erosion, cannot be that of a laccolith and accordingly Brown's petrogenetic model must be discarded. The study of silicate systems alone renders it highly unlikely that the nepheline monzonite could be derived from a rest fluid filter-pressed from what is now a quartz-bearing rock, especially considering the relative volumes of the two bodies.

New interpretations of the petrology must account for the pyroxenite at Poole's Point (and elsewhere), the differentiation of the monzonitic rocks, the origin of the nepheline-bearing types, and the garnet-pyroxene
bodies. The association of nepheline-bearing and quartz-bearing rocks deserves close study, especially since there is strong evidence that the two series are coeval.

The petrochemistry of the members of the complex is described below, and is primarily concerned with the variations observed in rocks of dominantly magmatic origin. Those rocks which are probably or obviously strongly hybridized, or xenoliths, sediments etc., are not shown on the variation diagrams. The pyroxenites near Poole's Point are provisionally grouped with those rocks without, or with only traces of, nepheline in their norms. The nepheline-bearing rocks, including the shonkinite at Tilba Tilba village, are also discussed as a group. Relevant analyses by Brown (1930) have been incorporated into the diagrams and discussion.

(1) **Major Element Variation**

The new analyses of rocks from the Mt Dromedary district are set out in Table 29 and the C.I.P.W. norms of all but the sediments and xenoliths are given in Table 30. The analyses are grouped according to rock-type and mode of occurrence, and arranged within the groups in order of increasing silica.

The major element variations of the analyzed rocks are shown in Figures 76, 77 and 78, in which the weight per cent of the oxides is 

\[
\frac{\text{FeO} + \text{Fe}_2\text{O}_3 \times 100}{\text{FeO} + \text{Fe}_2\text{O}_3 + \text{MgO}} \quad \frac{\text{Na}_2\text{O} + \text{K}_2\text{O} \times 100}{\text{Na}_2\text{O} + \text{K}_2\text{O} + \text{CaO}}
\]

plotted against the ratios, the mafic and felsic indices respectively (Simpson, 1954a). The mafic index, first employed by Wager and Deer (1939), gives a measure of the variation
Figure 76.

Graph showing the variation of various oxide contents with mafic index. The oxides plotted are SiO₂, Al₂O₃, Fe₂O₃, FeO, MgO, CaO, Na₂O, K₂O, and TiO₂. The mafic index is calculated as Fe₂O₃/FeO x 100/(Fe₂O₃ + FeO + MgO).
of the mafic mineral series (the discontinuous reaction series of Bowen, 1922), while the felsic index is related to the variation of the felsic minerals (the continuous reaction series).

The crystallization of tholeiitic and alkali basalts is broadly involved with the mafic and felsic mineral groups (Walker and Poldervaart, 1949; Tilley, 1950; Wilkinson, 1958); the complete course of fractionation may thus be represented by the use of both indices. Simpson points out (1954a, p. 240) that the mafic index is best used to portray the early and middle stages of fractionation, while the felsic index gives a better picture of the later stages.

(a) The pyroxenite-monzonite-banatite group.

The oxides of these rocks are plotted against the mafic index in Figure 76, and the felsic index in Figure 77. The variation of the oxides in both diagrams is remarkably regular, there being a progressive increase during fractionation of silica, alumina and alkalies which is sustained by a steady decrease in ferrous and ferric iron, magnesia, lime and titania.

The monzonitic and banatic rocks tend to lie along two separate lines, each of which represents a specific value of an oxide. For example, the four most fractionated samples are the equivalents of leucocratic quartz banatite, and the centre group of six samples are monzonitic rocks and equivalent lavas. Reference to the curve for either silica, soda or total iron (Figure 77) shows that each group tends to have its own particular amount of these oxides. The curves have been drawn for the smoothest best fit, and it is possible that the transition
Figure 77.

![Graph showing the distribution of various minerals and compounds across different felsic index values. The graph includes data for SiO$_2$, Al$_2$O$_3$, Fe$_2$O$_3$, FeO, Total Fe as FeO, MgO, CaO, Na$_2$O, K$_2$O, and TiO$_2$. Each mineral or compound is plotted against the felsic index, which ranges from 0 to 90.]
from one group to another is marked by too gentle an inflexion.

Silica shows the greatest variation in the pyroxenite-monzonite-banatite group, and the greatest difference is not between the pyroxenites and monzonites but between the monzonites and banatites. Alumina increases sharply from the pyroxenites to a very stable level at 17-18 wt %. This is maintained chiefly by the exchange of alumina between the decreasing amount of anorthite in the plagioclase and the increasing amounts of soda and potash going to form alkali feldspars and biotite. Figure 77 shows that the decrease in lime through the series is very nearly linear.

The curve for magnesia is convex downward, and shows a greater rate of decrease than lime in the early rocks, but a smaller rate in the most felsic. This results primarily from fractionation of plagioclase, and to a smaller degree from the slight decrease in magnesia in the later pyroxenes. The weaker relation between magnesia and lime in the felsic rocks reflects the entry of magnesia into the lime-poor mineral, biotite. The amphibole has a CaO/MgO ratio very similar to that of the associated pyroxene, and the growth of one at the expense of the other would have little effect on the relative proportions of the two oxides.

The curves for FeO, Fe$_2$O$_3$ and TiO$_2$ are very similar, with the highest values for the oxides found in the mafic rocks, and the lowest in the felsic types. With increasing fractionation, the curves fall fairly sharply and form a plateau of constant or slightly diminishing composition which extends over the middle range of fractionation. With progression into the highly fractionated rocks, the oxide content falls sharply to low, constant levels.
Figure 78.

The diagram shows the variation of various elements with the Felsic Index. The Felsic Index is defined as

\[
\text{Felsic Index} = \frac{\text{Na}_2\text{O} + \text{K}_2\text{O}}{\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O}} \times 100
\]

The elements plotted are:

- SiO₂
- Al₂O₃
- Fe₂O₃
- FeO
- Total Fe as FeO
- MgO
- CaO
- Na₂O
- K₂O
- TiO₂

The data points are represented by black circles, and the lines connect them to show the trend as the Felsic Index changes.
Absolute iron enrichment does not occur but a comparison of the trends for total iron and magnesia in Figure 77 shows slight relative enrichment in the monzonitic rocks.

Alkalies increase rapidly from the pyroxenites, and the enrichment of each is moderate and regular in the range from the monzonites to the most felsic rock. Potash is dominant, except in some pyroxenites, and is slightly enriched relative to soda during fractionation (see Bowen, 1928). It should be noted that the scales for K₂O and Na₂O in Figures 76 and 77 are slightly different.

(b) The undersaturated, alkalic group.

Figure 78 shows these rocks plotted against the felsic index. The broken line corresponds to the trend of the components of the pyroxenite-monzonite-banatite group.

The plots for all oxides except alumina fall along simple curves, which cover a smaller range in felsic fractionation than is shown by the pyroxenite-monzonite-banatite group. The components which show the greatest difference between the two rock-series are silica, which is consistently less abundant in the undersaturated alkalic rocks, total iron, lime and potash, all of which are more abundant. Alumina and soda are both lower in the more mafic rocks, but higher in the more strongly fractionated types than rocks of the pyroxenite-monzonite-banatite group showing the same degree of felsic fractionation.

Minor absolute enrichment in ferric iron occurs early in the fractionation series; the trend of total iron is similar but its degree of enrichment is negligible. Enrichment in iron relative to magnesia is
Figure 79. Plot of mafic against felsic index for pyroxenites (solid circles) and monzonitic rocks (open circles).

Figure 80. Plot of mafic against felsic index for rocks at Mt Dromedary compared with trends from other differentiated series: ●●●● trend of Figure 79; ●●●● trend of undersaturated alkalic rocks at Mt Dromedary; --- calc-alkaline trend, after Daly; x-x-x Skaergaard trend.
much more apparent in this series because of the greater initial decrease in magnesia compared to that of the pyroxenite-monzonite-banatite rocks. The higher amounts of lime and total iron, and of magnesia in only the early members of the undersaturated alkali rocks, is partly a result of a higher content of pyroxene in the earlier rocks; in the later members the pyroxene is represented by hastingsitic amphibole, iron-enriched pyroxene, and ultimately lime-iron garnet, all of which are poorer in magnesia (relative to Ca and Fe) than the early salitic pyroxene.

The relationship between the abundant alkalis is the reverse of that normally observed, in that potash shows strong early dominance over soda, but the latter is enriched relative to potash as fractionation progresses, and in the final fraction soda is dominant.

The crystallization sequence in calc-alkaline magma includes minerals which soda can enter. The preferential incorporation of soda relative to potash is determined by energy considerations, closely related to the difference in size of the two ions. Potash is thereby concentrated in the rest magmas and ultimately crystallizes as dominantly potassic minerals such as alkali feldspar and biotite.

The rocks in this group show a mineral paragenesis comparable to that of the calc-alkaline rocks; that is, the mode of crystallization of the nepheline-bearing rocks is miaskitic, and although the minerals of these rocks are more sodic than their monzonitic counterparts, the differences are small. Consequently, one would expect a more normal trend for the alkalis.
Figure 81. Mafic (M) versus felsic (F) indices for the undersaturated alkalic rocks. The trends calculated for the alkali basaltic rocks of Yagi (1953) and Wilkinson (1958) are included for comparison.
(c) **Mafic versus felsic index.**

The plot of the mafic index against the felsic index for the rocks of a differentiated series is useful in assessing the relative importance of each type of fractionation.

Figure 79 shows the relation between the mafic and felsic indices for the pyroxene-monzonite-banatite group. The flat slope of the curve indicates the relatively powerful fractionation of the felsic minerals compared to that of the mafic series. The sluggish rate of mafic fractionation is reflected mineralogically by the occurrence in most of the rocks of clinopyroxenes with very restricted composition.

The trend for the undersaturated alkalic rocks is shown in Figure 80, together with the course of fractionation shown in Figure 79. The trends for rock-series with and without absolute iron enrichment are shown for comparison (Skaergaard and calc-alkaline trends respectively).

The least fractionated member of the undersaturated alkalic group (shonkinite: Brown, 1930, p.656) has roughly the same mafic index as the least fractionated member of the pyroxene-monzonite-banatite group, but the felsic indices differ quite strongly. The slope of the variation curve is very close to 45° over much of its course. Such a trend would be developed in a rock series by strong, non-selective fractionation of both mafic and felsic constituents (Simpson, 1954a, p.242).

The late stage members of this group of rocks were apparently more strongly influenced by selective felsic fractionation, since the variation curve tends to flatten out parallel to the felsic index axis.

The undersaturated alkalic rocks are compared (Figure 81) with some alkali basalt derivatives described by Yagi (1953) and Wilkinson (1958).
Figure 82. Plot of wt % FeO - MgO - Na2O + K2O for rocks from the monzonitic complex at Mt Dromedary, compared with trends from the tholeiitic Red Hill (R→R) and Skaergaard (S→S) intrusions and the calc-alkaline rocks (C→C) after Daly.
Absolute iron enrichment does not occur in the Morotu rocks but has been reported from the Black Jack sill. Distinct enrichment in iron relative to magnesia is found in both occurrences.

(d) Proportions of iron, magnesia and alkalies.

All the analyses used in the variation diagrams, plus the accumulative rocks, are plotted in Figure 82 to show the variation in the proportions of ferrous iron, magnesia and total alkalies. Included for comparison are the differentiation trends for the Skaergaard intrusion (Wager and Deer, 1939), the tholeiitic Red Hill intrusion (McDougall, 1962, p.302) and the calc-alkali trend (after Daly, 1933).

The pyroxenites all plot between the Skaergaard trend line and the FeO-MgO sideline and they show some variation in their FeO-MgO ratios. This scatter of plots about the basic end of an otherwise regularly varying series is very commonly observed in strongly fractionated associations, and the scattering is usually taken to indicate that the corresponding rocks were formed, in part at least, by crystal accumulation. The parent of the association is assumed to have the composition of the most basic rock lying along the curve of continuous variation.

It is clear from Figure 82 that the rocks, viewed as a whole, show no absolute or relative enrichment in ferrous iron, and the trend of composition is very similar to that of the calc-alkali association.

When total iron is substituted for ferrous iron the plots all migrate toward the iron apex, but their relative positions are maintained. The points most affected are those which represent the pyroxenites, largely because these rocks are rich in magnetite.
Figure 83. Variation of total iron as FeO (F), K₂O + Na₂O (A) and MgO (M) for the undersaturated alkalic rocks. The trends calculated for the alkali-basaltic rocks of Yagi (1953) and Wilkinson (1958) are included for comparison.
Figure 83 shows the undersaturated alkalic rocks from Mt Dromedary plotted on the basis of weight per cent FeO (total) $- \text{MgO} - \text{Na}_2\text{O} + \text{K}_2\text{O}$. The variation curves for the alkalic rocks of the Morotu district (Yagi, 1953) and the Black Jack sill (Wilkinson, 1958) are included for comparison.

(e) Miscellaneous rocks.

The large block of volcanic rock enclosed by the monzonite to the west of Mt Dromedary trigonometrical station is composed in part of tuffaceous basaltic rock, which has been slightly recrystallized. A sample was analyzed (RB114, Table 29) and it is chemically distinct from the lava to the east of Mt Dromedary (RB323, Table 29). The occurrence of amygdaloidal basalts and melaphyres in the local Upper Devonian sequence has been reported by Brown (1931; 1933) and the sample from the inclusion is not unlike them. It is proposed that this large mass has been stoped from the Upper Devonian rocks during their invasion by the monzonite. The recrystallized biotite-rich xenolith in the Central Tilba quarry (RB201, Table 29) may have a similar origin.

Two specimens of the country rock (Wagonga Beds) were analyzed with the principal aim of determination of their content of lime, magnesia and carbon dioxide. The analyses show very little lime, fairly low magnesia and no carbonate. They contain quite considerable amounts of potash, silica and alumina. Their norms contain 28.6% and 60.7% quartz; 11.2% and 4.6% corundum; 41.3% and 23.7% total feldspar, with orthoclase greatly dominant.

RB7, the dolerite dyke, is chemically distinguished from the other igneous rocks in the district by its relatively high TiO$_2$ and Na$_2$O, and
low CaO, MgO and K₂O. It is thought to be unrelated to the monzonitic complex.

(2) Petrogenesis

(a) The monzonite-banatite association.

These rock-types and their dyke and extrusive equivalents together make up about 90% of the outcrop of the complex at Mt Dromedary. Their chemical variations have been discussed above and would indicate that mechanisms of dominantly felsic fractionation have been operative in producing monzonitic, banatitic and quartz-banatitic members from an andesitic or basaltic parent. The relatively constant levels of the oxides in each of the monzonitic sub-group and the banatitic sub-group when plotted against the mafic or felsic indices, compared to the jump in the levels between the sub-groups suggest that there was an absence of extrusive activity between the time of emplacement of the two sub-groups. This is also suggested by:

(i) the abundant large phenocrysts in the marginal phase of the banatite;

(ii) the sharp textural and mineralogical differences between the two major rock types.

It is of interest to note the different stress conditions attending the emplacement of these major bodies.

The strongly porphyritic banatite at Mt Dromedary occupies about 20% of the total outcrop of the intrusion (Table 1). The transition from the porphyritic to the non-porphyritic banatite is noticeable enough
in the field to map approximately. The width of the zone of abundant phenocrysts remains fairly constant around the circumference of the banatite.

It is clear from the near-perfect concentricity of the contacts of the monzonite and banatite and their common axis of intrusion that the two bodies are structurally very closely related. Furthermore, the rocks were intruded in order of increasing acidity, so it appears very likely that the monzonite and banatite are successive differentiates of the same body of magma. The alkali feldspar phenocrysts are of intratelluric origin (applying the criteria of Watson, 1901) and they probably crystallized during the interval between the intrusion of the monzonite and banatite. The specific gravity (at room temperature) of the phenocrysts is 2.56, and the mean specific gravity of the banatite including the phenocrysts is 2.63; without the phenocrysts it is calculated to be a little higher, 2.66. It would appear then that the phenocrysts would float in a banatite melt, provided that the decrease in density accompanying the melting of the banatite was small and the melt was fairly dry.

After a period of quiescence, during which appreciable fractionation took place, probably accompanied by formation of a layer of large alkali feldspar crystals at the roof of the magma chamber, ring fracture occurred and the banatite was emplaced. The marginal concentration of the phenocrysts was probably brought about as a consequence of the movement of the magma, rather than by chemical processes such as migration of potash to the margins. This is indicated by the intratelluric nature of the alkali feldspar crystals. The porphyritic banatite
must therefore be regarded as an accumulative rock.

The genetic relations of the analyzed rocks is indicated (but not necessarily proven) by their regular chemical variations and similarities. Dominantly felsic fractionation at a relatively shallow depth of a moderately potassic, basic andesitic or basaltic parent magma would account adequately for the broad relationships observed.

(b) **The pyroxenites and garnet-pyroxene rocks.**

(i) **The unaltered pyroxenites.** The rocks exposed near Poole's Point are ultrabasic, and show a variety of structures and composition which clearly indicates that it has diverse origins.

The laminated feldspathic pyroxenite has structures which indicate its formation by rhythmic precipitation in a floored magma chamber at depth. Features typical of current action may be observed in much of the rock. It has been subsequently deformed and intruded by younger ultramafic material with very similar chemical and mineralogical characteristics, and this rock in turn appears in some exposures to be more or less broken up and enclosed in yet younger pyroxenite. The mode of emplacement of this corporate mass is not known since its contacts are obscured by alluvium, but two alternatives are suggested; the pyroxenite is dominantly either intrusive or xenolithic.

The pyroxenite and altered pyroxenitic rocks have not been observed in contact or nearly so with the sedimentary Wagonga Beds.

At those points where the pyroxenite crops out near the metasediments
monzonitic rock is associated with it (RB64, analysis in Table 29),
and minor displays of hybrid rocks may be seen (RB12A, analysis in
Table 29). Pyroxenitic rock is very commonly found as xenoliths
in the coastal dykes and the minor monzonitic intrusions around Mt
Dromedary. The most notable examples are the garnetized masses
surrounded by nepheline monzonite near Little Dromedary. Some of
this rock is laminated and local variations in texture and composition
are very similar to those observed at Poole's Point.

The wide distribution of these rocks, their range in size, and
their apparently universal association with other igneous rocks present
grave difficulties in regard to the hypothesis of their dominantly
intrusive nature.

The alternative hypothesis, that of a dominantly xenolithic
origin, accounts for the observed relations quite easily, except in
the matter of size. The body of partially garnetized rock north of
Little Dromedary has a diameter of nearly half a mile, while the
vertical extent of the pyroxenite near Poole's Point exceeds 600 feet.

Reference to Plates I and II shows that, compared to the
probable size of the monzonitic complex, a block of pyroxenite which
measured as much as half a mile on each edge is still quite small.
However, the difference in density of the pyroxenite and the nepheline
monzonite is quite considerable, and the rate of sinking of large
blocks of pyroxenite in a hydrous monzonitic melt would be expected to
be high, especially if Stokes' law applies.
In thin section the nepheline monzonite is not characterized by a hydrous mineral assemblage; on the contrary, there is abundant evidence of early loss of volatiles (replacement of biotite by amphibole), while the textures of the felsic minerals are dominated by replacements which have apparently taken place in nearly or completely solid rock (nepheline and orthoclase after plagioclase). Pegmatites are rare or absent and only a few narrow felsic veins have been observed. Therefore, it is likely that the nepheline monzonite magma was rather viscous, with the result that sinking of the pyroxenite was greatly hindered.

In view of the poor exposures in the contact zones and the lack of precise data concerning the viscosity of the monzonitic magma, the form of the pyroxenitic rocks remains a matter of doubt. It is clear, however, that some phases of the pyroxenite are of accumulative origin, and that the pyroxenite as a whole is at least slightly older than the other rocks of the complex.

(ii) **Garnet-pyroxene rock.** Field, petrographic and chemical data indicate quite clearly that the garnet-pyroxene and associated rocks have formed from pyroxenites through metasomatism related to the surrounding nepheline monzonite. The replacement was very patchy and it produced assemblages usually rich in a pink andraditic garnet, with widely varying amounts of alkali feldspar, andesine and relict lopolitic clino-pyroxene. The extensive introduction of alkali feldspar can be correlated with a decrease in the amount of garnet, so that rocks composed dominantly of garnet, pyroxene, minor calcic plagioclase; pyroxene, alkali feldspar + andesine, minor garnet; and pyroxenite, are intimately associated in the field.
Replacement of the pyroxene, plagioclase and opaque minerals by garnet appears to have been volume for volume, and was not accompanied by deformation of the original rock. The original textures of the pyroxenites are easily deduced from the relict pyroxenes.

The development of garnet, etc., resulted in an increase in specific gravity of the rocks by about 10%, and this must be taken into account in determining the net change in chemistry brought about by garnetization. From the analyses in Table 12 (A and B) the net change in the principal oxides can be calculated, assuming volume for volume replacement and an increase in specific gravity of 10%. To derive sample B from sample A the following proportions by weight of the major oxides must be added:

- TiO₂: 0.6
- Al₂O₃: 2.0
- Total iron: 6.7
- Lime: 7.0
- Alkalies: 1.2

The weight proportion of oxide removed is MgO, 6.0.

Clearly, the trend is one of introduction of alkalies and alumina to form the feldspars, plus CaO and iron for the garnet. The most important oxide removed is magnesia. Nearly all of the iron in the garnet is in the oxidized state. The atomic proportions of lime, alkalies and alumina (+ iron) introduced exceed that of the magnesia removed, so that the transformation of the pyroxenite may actually result in enrichment of silica in the enclosing rock.

(c) The undersaturated alkalic rocks.

The magmatic nature of the nepheline monzonite is indicated by its
restricted occurrence, sharp contacts with finer grained rock against the Wagonga Beds, and primary igneous flow structures. The contact relations of the 'ijolite' are obscure, but the rock shows a very strong platy flow structure and it is also thought to be of igneous origin.

The total outcrop of the nepheline-bearing rocks is a little over 6% of that of the entire complex, 90% of which is taken up by oversaturated and nearly saturated monzonitic rock.

The Harker diagram is of very limited use because of the relatively constant amount of silica within groups of similar rocks compared to the variations in the remaining oxides (see Figures 76, 77, 78). The shapes of the curves of major element variation are consequently ill-defined. However, the available data show that the alkali-lime index (Peacock, 1931) of the pyroxenite-monzonite-banatite group is about 53, and for the nepheline-bearing group it is about 50. Using Peacock's terminology, the former group is alkali-calcic, while the latter group is barely alkalic (alkalic, less than 51; alkali-calcic, 51-56; calc-alkalic, 56-61; calcic, greater than 61).

None of the rocks of the monzonitic complex are so undersaturated with respect to silica that the place of the feldspars is wholly or largely taken by feldspathoid.

The nepheline-bearing rocks of the complex are essentially coeval with the main intrusion of the monzonite. Banatite intrudes the latter at Mt Dromedary, and the former at Little Dromedary, and consequently the complex may be referred to those alkalic intrusive associations which include nepheline-syenite but close with quartz-syenite (Tilley, 1958, p.332).
The actual chemical differences between the monzonite at Mt Dromedary and the nepheline monzonite (cf. RB221, RB424; Tables 19, 20) are not great; their norms, in particular, compare rather closely. The chief differences between the nepheline-bearing and the monzonitic rocks lie in the slightly higher total iron, lime and potash of the former.

The trend toward soda enrichment (Figure 78) is based upon the plot of one sample, the garnet-bearing foyaitic dyke, and if this were omitted the trend of the alkalies would be very similar for both groups of rocks.

It is proposed that the nepheline-bearing rocks, in particular the nepheline monzonite and 'ijolite', were derived from essentially the same parent magma as the monzonitic rocks, and that the different trends of differentiation were initiated at a relatively early stage.

A similar relation of over- and undersaturated rocks has been described by Simpson (1954), who believes that the basic parent magmas of the alkalic and tholeiitic series of the Okonjeje complex were only slightly different, and were both related to a common parent. Barth (1945, in Turner and Verhoogen, 1960, p.395) proposed that in the Oslo complex both over- and undersaturated felsic differentiates were derived from a saturated syenite-monzonite magma.

The limestone syntaxis hypothesis propounded by Daly and Shand appears to be of limited application in the production of the nepheline monzonite or 'ijolite' for the following reasons:

(i) The Wagonga Beds do not appear to have calcareous members. This has been pointed out by Brown (1930) and has been verified by the present writer. In order to assess the effect of assimilation of
the phyllitic country rock, two samples (RB472, RB473; Table 19) were chemically analyzed. Lime is notably low in these rocks, while potash, silica and alumina are rather high.

(ii) The nepheline-bearing rocks do not show a distinctive lime-silicate mineralogy, and on the whole they are poorer in CO₂ than many other rocks of the complex. The garnet-pyroxene rocks, though superficially similar to calc-silicate hornfels, are almost certainly altered igneous rock.

(iii) Incorporation of limestone subsequent to the formation of the early pyroxene in these rocks is highly unlikely, since the zoning which the pyroxenes display is toward less, not more, diopsidic compositions.

(iv) The much larger intrusions of monzonite and banatite show no trace of carbonate assimilation.

(v) The igneous rocks as a whole are characterized by constant chemical features, in particular the near equality of Na₂O and K₂O, and if such processes have operated then they must have acted in such a way as to preserve these features, yet give undersaturated rocks on one hand and quartz-bearing types on the other.

In view of the probable parent magma for the monzonitic complex (a near-critically undersaturated potash-basalt), the derivation of the nepheline-bearing rocks can be adequately accounted for by normal differentiation, analogous to that of the alkali basaltic rocks.
(d) **The petrogenetic role of the clinopyroxenes.**

The close optical and chemical resemblance of the suite of pyroxenes provides one of the stronger arguments in favor of all the rocks, including the pyroxenite, being regarded as derived from the same parent magma. This hypothesis affords at once an explanation of the source of the pyroxenite and an indication of some possible mechanisms of differentiation.

The pyroxenites at Poole's Point are very basic rocks, with silica falling as low as 38.04%. This is partly a consequence of their high content of magnetite and ilmenite, which commonly sum to more than 10% by weight. The structures of these rocks have been described in detail (Part I, II) and it is sufficient at this point to note that they have formed in part by crystal accumulation under the influence of gravity, and show structures indicative of current action in the magma chamber.

Tables 14 and 17 give the analyses and norms respectively of the analyzed pyroxenes. Except in the cases of the samples from the banatite (5) and the ferriferous rim of the 'ijolite' pyroxene (9B), all the pyroxenes are nepheline-normative; they are all olivine-normative. The small amount of normative hypersthene in (5) may be related to the saturated nature of the host-rock, whereas the hypersthene in the norm of (9B) is a consequence of its high TiO₂ content.

Separation of a nepheline- and olivine-normative pyroxene from a basaltic magma lying near the plane of critical undersaturation will tend to enrich the residual liquid in silica, and increase its ratios of Na₂O to CaO, K₂O to Na₂O and CaO, and FeO to MgO, only provided that the phase which is separating is relatively enriched in CaO, (Na₂O + CaO) and MgO respectively. Continued withdrawal of pyroxene or another phase (or
Figure 84. Variation of some normative minerals between mean pyroxenite and lava RB323. Shaded fields represent silica ($N_{e_o}$) less than, and in excess of, the amount required to convert all the available soda to nepheline. P.C.U. = plane of critical undersaturation.
phases) appearing with it on the crystallization surface requires the co-
precipitation of that phase (or phases). Consequently, the relation 
between successive differentiates in natural systems is rarely, if ever, 
so simple that it can be adequately described in terms of simple 
subtraction of a single stoichiometric component. Although the separation 
of pyroxene of a similar composition to those analyzed should increase the 
magmatic proportions of silica, potash, etc., the effect of co-precipita-
tion of the opaque minerals and calcic plagioclase must be considered in 
subtraction calculations.

Figure 84 illustrates the effect, in terms of the normative 
minerals, of the separation of rock matter of the composition $P$ (mean 
pyroxenite calculated from Table 20) from rock whose composition is that 
calculated from RB323 and VI, the andesite quoted by Brown (1930). As separation of the pyroxenite proceeds, the nepheline of $P$ is consumed 
to yield albite, and at the point that there is no normative nepheline 
represented, normative hypersthene appears. The point of transition 
between these two antipathetic minerals is analogous to the critical 
plane of silica undersaturation demarking those basaltic rocks tending 
to a saturated differentiate (the various tholeiitic types) and those tend-
ing to an undersaturated residuum (the alkali basaltic types). The 
variation of the normative minerals shown in Figure 84 is intended to be 
illustrative, not quantitative. Diagrams of this kind may be very 
neearly faithful representations of actual processes, because of the small 
variation in composition of the major subtracted phase, i.e., pyroxene. 
Hence the clinopyroxene of the pyroxenite-monzonite-banatite group affords 
a mechanism of passing from the critically undersaturated trend to the 
olivine tholeiite trend of differentiation, without recourse to incongruent
melting relations of certain minerals, as suggested by Tilley (1958, pp. 332, 333).

The researches of Schairer and Yoder (1960), Yoder and Tilley (1962) have made it clear that very small differences in composition (especially the amount and distribution of $\text{Al}_2\text{O}_3$) of the basaltic pyroxenes may commit their residual liquids toward an over- or undersaturated end-product. The pyroxenes and the rocks of the monzonitic complex at Mt Dromedary may offer an example of the effects of these subtle differences in initial composition. It should be noted, however, that the pyroxenes of the more alkalic rocks at Mt Dromedary are richer in $\text{Al}_2\text{O}_3$ than those of the less alkalic rocks, the reverse of the relation discussed by Yoder and Tilley (1962, p. 413).

(e) Other minerals.

(i) Olivine. Olivine occurs as a primary phase in the more feldspathic pyroxenites and in small monzonitic intrusions of hybrid aspect, including perhaps the shonkinite at Tilba Tilba village. The mineral has been recorded by Brown (1930) in the nepheline monzonite and has been noted by the writer in the 'ijolite' (Figure 53). It is a magnesian variety and except in the pyroxenites it does not appear to be in equilibrium with its host; it is very commonly jacketed by biotite and thereby effectively removed from reaction.

The olivine may be associated with clinopyroxene in the igneous rocks but there is strong evidence of a reaction relation between the two minerals as there usually is in tholeiitic rocks where olivine reacts to give orthopyroxene. In the 'ijolite' the rare grains of olivine are mantled by diopsidic clinopyroxene and this may be a consequence of a
reaction relation pertaining in an alkalic environment as suggested by Schairer and Yoder (1960).

(ii) Biotite and amphibole. The paragenetic sequence in several rocks containing both biotite and amphibole is (olivine)-pyroxene-brown biotite-olive amphibole-(green biotite). There is widespread replacement of the common red-brown and olive-brown biotite by the amphibole, and this is best observed in the pyroxenites and the nepheline monzonite. This reversal of the normal crystallization sequence has been noted in a few other localities, usually in rather alkaline rocks.

Dmitriyev et al. (1962) describe such an occurrence in the Kaib massif of Central Kazakhstan. The rocks concerned are alkali granites with associated mafic hybrids, and it is interesting to note that the ratio of Na₂O to K₂O is very nearly unity in these rocks. The decrease in stability of the biotite structure as the alkalinity of the rock increases is ascribed to weakening of the Al⁴⁺ bonding as the content of Fe³⁺ of the biotite increases (p.258). Eventually the sheet structure gives over to the chain structure of pyroxene and amphibole.

The precipitation of amphibole may be accompanied by a new biotite, which is poorer in Fe³⁺ because of the entry of that ion into magnetite and the amphibole. This hypothesis appears to account quite adequately for the replacement of biotite by amphibole, and the concurrent or subsequent crystallization of a distinctly different biotite in the nepheline monzonite. However, this mechanism is much harder to envisage in the case of the pyroxenitic rocks.

Replacement could also be brought about by either or both increase in temperature and decrease in pressure such that biotite ceased to be stable (Yoder and Eugster, 1954), its place being taken by the more
stable amphibole. Temperature increase is preferred to pressure decrease as a cause, since biotite has crystallized subsequently, a feature in accord with cooling, but more difficult to explain in terms of another pressure change.

Regardless of the mechanism, replacement of the biotite would yield a considerable quantity of a hydrous fluid rich in potash. If this is associated as Dmitriyev suggests with increase in alkalinity in the nepheline monzonite, it offers a source of the alkalic fluids which have permeated the pyroxenite xenoliths near Little Dromedary.

Since the biotite was obviously unstable at the time of its replacement, it is not surprising that the subsequent alkali metasomatism did not produce biotite in the garnet-pyroxene rocks.

(iii) Garnet. It has been suggested above that the chemical similarity of the analyzed garnets reflects similar chemical (and physical) conditions of crystallization, despite the differences in the host rocks. The garnets appear to have crystallized in a hydrous alkalic environment, probably at quite low temperatures. The precipitation of andradite garnet in the 'ijolite' and the foyaitic dyke in place of lime feldspar and magnetite or hastingsitic amphibole may be a consequence of combined low temperature and high alkali content.

If the rocks contained biotite and/or hastingsitic amphibole, and these minerals became unstable in response to decreasing temperature, pressure and increasing alkali content, their breakdown would yield lime, which could not be precipitated as anorthitic plagioclase because of the low temperatures, but would tend to form lime garnet, taking up the iron and titania of the pre-existing mafic minerals to give andradite. It is
perhaps significant that garnet and biotite are not found together in these rocks. A similar explanation has been offered for the garnet at Port Cygnet by Edwards (1947, p.96).

(f) The influence of the epi-tectonic locale.

Figures 79 to 83 show that there are two principal lines of variation, that of the pyroxenite-monzonite-banatite group, which is closely similar to the calc-alkali trend (of Daly), and that of the undersaturated alkalic rocks, similar but by no means identical with the trends for derivatives of critically undersaturated basalts.

In general, absolute iron enrichment is absent, as it is in the continental calc-alkaline rocks. Osborn (1959) has discussed the role of oxygen pressure in the crystallization of basaltic magma, and the rocks at Mt Dromedary can be profitably discussed in the light of his findings.

He proposes that fractional crystallization of olivine basalt under conditions of more or less constant $pO_2$ leads to little iron enrichment compared to fractional crystallization at constant total composition. This is brought about by fairly early precipitation of magnesio-ferrite in the former case. The trend is dominantly toward silica enrichment, and a considerable proportion of the fractionated magma crystallizes at the ternary eutectic pyroxene-magnesioferrite-silica. Osborn states (p.627) that fractional crystallization serves to purge the liquid of olivine and some pyroxene and magnesioferrite crystals, and the liquid then terminates at the eutectic in large volumes.

This is very similar to the situation at Mt Dromedary, where the pyroxenites could represent this early purging, while the final products,
monzonite and banatite, occur in very large volume compared to the pyroxenites, although there is no way to estimate the original volume of the pyroxenite since only small masses have been observed which have probably been brought up from depth.

Osborn believes the condition of relatively constant \( pO_2 \) may be induced by the comparatively high content of water in rocks of the orogenic belts. The monzonitic complex at Mt Dromedary intrudes only slightly altered argillaceous sediments which have from 2-5\% by weight of water, and this would be adequate to give the conditions described by Osborn. Reference to Figures 76 and 77 shows the progressive depletion in iron oxides and the rock and mineral analyses in general show constant or slightly increasing values for \( Fe_2O_3/FeO \).

The affinities of the rocks and minerals with those of the alkali basaltic rocks have been pointed out in several places above. The different chemical features of the proposed parent magma of the monzonitic complex and its trends of differentiation show some resemblance to the calc-alkaline rocks, and it is suggested that these resemblances stem from the continental location of the intrusion; the chemical characteristics of the complex may possibly be related to assimilation of a small amount of sialic material, or to some tectonic control of the kind often cited as operative in the production of alkaline rocks of the platform regimen. In this regard, it is interesting to note that palaeomagnetic observations (Robertson, 1963) indicate an abrupt and considerable shift in palaeogeo-latitude of eastern Australia dating from about the time of intrusion of the complex at Mt Dromedary.

The orthodox interpretation of these observations by those who
subscribe to theories of continental drift is that the position of the continent changed relative to the axis of revolution of the earth at that time. This interpretation is based on two assumptions: firstly, that the earth's magnetic field is, and has been, that of the geocentric axial dipole; and secondly, that the geomagnetic and geographical axes have remained essentially coincident.
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PLATE III

Plane table maps of the coastal strip east of Tilba Tilba Lake, showing dyke relations. Figure at left is a sketch map showing the relations between the sectional maps 1, 2, 3 and 4.

Rock Types.

(i) Tightly folded Wagonga Beds; hatching is very nearly parallel to the trend of the beds.

(ii) Dolerite.

(iii) Hornblende lamprophyre (camptonite) and associated breccia.

(iv) Quartz-bearing feldspathic rocks.

(v) Garnet-bearing foyaitic rocks.

(vi) Miscellaneous intermediate to basic dyke-rocks, usually very deeply weathered (black in sectional maps).

The sand adjoining the coastal strip is shown stippled, and the sea (to the east) is left unshaded. Unless otherwise indicated, the intrusions are nearly vertical.
A. Cross-section after Brown (1930).

B. Proposed cross-section, showing present profile compared to reconstructed base and top of Upper Devonian rocks at the time of intrusion of the complex (upper Middle Cretaceous).
GEOLOGY
of the
MOUNT DROMEDARY DISTRICT,
N.S.W.

LOCALITY MAP

MOUNT DROMEDARY

SCALE IN MILES

- - - - Approximate limit of spotting in aureole
- - - - Diamond drill hole

Garnet-pyroxene rock
Monzonic satellites
Felspathoid-bearing rocks
Pyroxenite
Lava & tuff andesitic
Older basic volcanic rocks
Garnet-pyroxene rock

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