MINERALIZATION ASSOCIATED WITH SPILITES AND KERATOPHYRES IN WEST TASMANIA

The ore deposits of the West Coast mineral field may be roughly classified according to their spatial relationship to magmatic rocks (as in Hall and Solomon, 1962, p. 289-290; Solomon, in press) as follows:

(a) Intramagmatic: cassiterite stockworks in granite (Heemskirk); copper-nickel ores in ultrabasics (Cuni), magnetite in amphibolites (Savage River), etc.

(b) Contact metasomatic: scheelite in the aureole of granite (King Island).

(c) Adjacent to granite stocks and quartz porphyry dykes: cassiterite-pyrrhotite lenses at Renison Be II and Mt. Bischoff and minor lead-silver-zinc fissure veins; in this group are the cassiterite-wolframite fissure systems of Rossarden and Moina on the East and North Coasts respectively.

(d) Not obviously related to intrusive rocks but within volcanics: copper sulphide ores (Mt. Lyell) and galena-sphalerite-chalcopyrite ores (Hercules and Rosebery); some silver-lead-zinc fissure systems (Tullah and possibly Zeehan and Magnet).
The first three types seem clearly magmatic (or hydrothermal) but the last group is of uncertain origin. Most of the deposits of this group occur within the Mt. Read Volcanic Arc and form a longitudinal zone of mineralization extending for 40 miles from Mt. Darwin to north of Rosebery and Tullah. The mineralization is markedly different from that in the remainder of the West Coast mineral field.

It is interesting to note that almost all (95%) of those West Coast ore deposits that are not in the Mt. Read Arc occur at the same stratigraphic horizon, i.e., close to the base of the Crimson Creek formation (or the top of Success Creek phase), and also that most of the deposits are close to areas of spilite development. The Magnet ores, for instance, occur at the same horizon as a thick succession of spilite flows and 85% of production from the Zeehan field comes from the seven mines in which spilites are reported (the Western, Montana, Spray, Grubbs, Oonah, Queen and Queen Extended - see Fig. 22). The presence of dolomite has localised ore deposition in many examples (e.g., Renison Bell, Mt. Bischoff, King Island) and these are clearly related to granitic intrusive rocks but the ores at Magnet and Zeehan are some distance from granites and have a much closer association with spilites. However, this thesis is concerned only with the deposits in the Mt. Read Arc.
Most of the workings within the Mt. Read Volcanics are shown in Fig. 3. They fit into a number of geographical zones, all aligned approximately N-S:

(a) Mt. Darwin to Comstock: the Mt. Lyell Copper Field;
(b) Lake Dora to the Anthony River;
(c) Mt. Tyndall to Red Hills;
(d) Sterling River to Tullah;
(e) Hercules to Rosebery and Silver Falls.

Almost all these deposits are associated with potassic quartz keratophyres, some like the Darwin keratophyre and some of pyroclastic origin. The only operating mines are at Mt. Lyell, Hercules, Rosebery, and Mt. Farrell and these account for most of Tasmania’s mineral production. Figures showing mineral production for these mines to the end of 1963 are given in Table 16.

**STRUCTURAL SETTING OF THE MT. READ VOLCANIC ARC**

As already mentioned, the Mt. Read Arc developed alongside the Tyennan Geanticline during the filling of the Cambrian eugeosyncline (see Fig. 2). It is suggested that the volcanics were erupted in a basin or trough of tensional origin that formed during
Table 16

Production to the end of 1963 from the Mt. Lyell, Rosebery-Hercules and Mt. Farrell Mines

<table>
<thead>
<tr>
<th></th>
<th>Cu (tons)</th>
<th>Pb (tons)</th>
<th>Zn (tons)</th>
<th>Ag (ozs)</th>
<th>Au (ozs)</th>
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<tr>
<td>Mt. Lyell</td>
<td>567,890</td>
<td></td>
<td></td>
<td>16,008,623</td>
<td>594,521</td>
</tr>
<tr>
<td>Rosebery and</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hercules</td>
<td>22,711</td>
<td>236,754</td>
<td>716,700</td>
<td>27,953,860</td>
<td>332,252</td>
</tr>
<tr>
<td>Mt. Farrell</td>
<td></td>
<td></td>
<td></td>
<td>10,228,236</td>
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</table>
the early stages of the sagging and stretching associated with the Cambrian eugeosyncline. The western edge of this trough was probably outlined by faulting, and the Rosebery Fault Zone, for example, probably dates back to this time.

In the early Ordovician a new tensional trough formed adjacent to the Tyennan Geanticline. This trough (the Owen basin or rift valley) was narrower than the Mt. Read trough and formed on the eastern side of the volcanic arc. In contrast to the Mt. Read structure it seems to have been associated not with depression of the geosyncline but with elevation of the Tyennan Geanticline. As a result it was filled by coarse detritus derived largely from the rocks of the geanticline. Fig. 31 illustrates the extent of the Mt. Read Arc and the Owen trough. The faults of the Owen trough may have been present in the Cambrian and acted as feeders for the Mt. Read magma.

Throughout the West Coast the Ordovician sediments are generally concordant with underlying Cambrian rocks. However, just east of Tullah, a 20° discordance in strike has been revealed (Fig. 23) and again south of Queenstown (Fig. 28) there is evidence of discordance. There appears to be a 15° - 20° discordance in the Lake Dora-Lake Spicer area (Fig. 24) and Campana and King (1963, p. 28) referred to unconformity near Red Hills.
Fig. 31: Sketch map to show the locations of the major geanticlinal areas, the Owen basin, the Mt. Read Volcanic Arc, and the major TF structures.
Bradley (1954) has pointed out that the supposed unconformity at Mt. Jukes (Hills, 1914) is due to unsheared horizontal Owen Conglomerate overlying vertically cleaved volcanics. The Jukes Conglomerate testifies to uplift and erosion of Mt. Read Volcanics along the site of the later Owen deposition.

It has been suggested that the pre-Owen (Jukesian Orogeny) movements consisted of gentle folding with faulting (Banks, 1956; Solomon, 1960; Banks and Solomon, 1961) thus giving rise to slight discordance in strikes and localised strong uplifts, particularly on faults flanking the Owen basins.

Campana and King (1963) believed that the Mt. Read Volcanics are separated from the Jukes formation by a long period involving a pre-Dundas Group orogeny, and hence Jukes and Owen Conglomerates rest unconformably on the volcanics but conformably on the Dundas Group and correlates. Where apparently conformable relationships exist between the Mt. Read Volcanics and the Jukes Conglomerate, Campana and King (p. 32) suggested the unconformity has been welded and obscured by diagenesis and subsequent metamorphism.

The next important phase of orogenesis was the Tabberabberan Orogeny, which consisted essentially of two phases of deformation (\(T_{F1}\) and \(T_{F2}\)). The first phase involved an
accentuation of the Cambrian structures with renewed movement on earlier faults, and development of long-wavelength folds on trends parallel to the Owen basin and the Mt. Read Arc. One of these was a synclinal structure passing through Dubbil-Barril and Zeehan, more or less coinciding with a zone of thick sedimentation in the Cambrian basin, the Dundas Trough. To the east (Fig. 31 and Fig. 32) there developed an anticlinal zone comprising the West Coast Range Anticlinorium and the Dundas Anticlinorium, and between this and the Tyennan Geanticline a narrow synclinal structure, represented by the King and Sophia Synclinoria. These anticlinoria and synclinoria have an east-facing asymmetry and imply transport towards the Tyennan Geanticline. This is probably an expression of the wholesale rising and regurgitation of the geosynclinal basin, in which several kilometres of sediments and volcanics had accumulated since Pre-cambrian time.

The second folding phase, $T_F^2$, appears to have been largely independent of the well-established structural trends and has a persistent NW to NNW trend. The folds have a shorter wave length than the $T_F^1$ folds and are clearly superimposed on the earlier structures. Interference between the two fold systems has produced doubly plunging synclines and anticlines like the Zeehan
Fig. 32: Cross-sections to show the geological structure of the West Coast: (a) through Zeehan and Mt. Murchison (about ENE).
Fig. 32: Cross-sections to show the geological structure of the West Coast: (b) through Strahan and Queenstown (East).
and Huskisson Synclines (Fig. 3); the rectangular pattern of these folds is very similar to those produced experimentally by two phases of folding (O’Driscoll, 1963, p. 153).

In the Mt. Dundas area the TF₁ structures appear to be separated from TF₂ structures by major thrust faults of NNE trend and these merge northward into the Rosebery-Hercules Fault Zone (Fig. 3).

Axial surface cleavage is common in relatively incompetent rocks like the Mt. Read Volcanics.

Associated with TF₂ deformation are zones of strike-slip faults of W to WNW trend e.g. at Queenstown (the Linda Fault Zone), between Mt. Jukes and Mt. Huxley, and at Zeehan.

Tectonic Analysis of the Tabberabberan Orogeny

The TF₁ structures involved differential vertical uplift (with some lateral transport), probably related to the regurgitation of the geosyncline. The TF₂ structures have been previously explained as due to dextral shearing of N-S trend (Carey, 1953) or NE-SW compression (Bradley, 1956; Solomon, 1957). The presence of major WNW fracture zones of considerable extent (the Linda Fault Zone is at least 25 miles long) has led the writer to suggest that transcurrent sinistral shearing of E-W trend may have
been the dominating influence during $\text{T}_2$ deformation (Solomon, in press, Appendix A, Paper 1). Fig. 33 illustrates the suggested $\text{T}_2$ stress and strain patterns.

**THE MT. LYELL COPPER FIELD**

All the known economic deposits are at the north end of this mineralized zone and lie between Comstock and Great Lyell (Fig. 34). They are mainly on the divide between Mt. Lyell and Mt. Owen and are worked by the Mt. Lyell Mining and Railway Co. Ltd. The regional geology of the Mt. Lyell copper field has been described by the writer in a previous thesis (1957), in which it was established that the host rocks to ore at Lyell were mainly volcanic. The geological succession in the mine area is as follows:

**Ordovician:** (305 m.) **Gordon Limestone,** with dark grey shales, carrying a rich Lower Ordovician fauna (Hill, 1955).

**Owen Conglomerate:**

**Upper Owen:** Pioneer Beds: 0-9 m.  
Grey, medium-grained sandstones  
Grey, coarse-grained sandstones,  
pink and grey medium-grained
Fig. 33: Suggested Stress and Strain diagrams for West Tasmania during the later phases of the Tabberabberan Orogeny.
conglomerates (31224, 32566, 
31225, 32720).
Grey, coarse- and medium-grained 
cross-beded sandstone with bands 
of alumina-rich chromite (by 
chemical analysis). Specimens 
319696, 30993.

(0-7.5 m.) Basal, fine to medium pebble 
conglomerate, locally with boulders 
of hematite.

Haulage unconformity between 
Comstock and the Iron Blow -----

(61-107 m.) "Chocolate Sandstone": Purple, 
coarse to medium-grained hematitic sandstone with characteristic 
vermicular bodies (31213, 32720).
Some 15 m. below the top of these 
beds are two lenses of oolitic 
hematitic sandstone between 60 and 
100 cm. thick (30954, 32742, 32939, 
30913). They are composed of 60% 
to 90% subangular quartz grains
varying from 0.5 to 0.1 mm. diameter, the remainder consisting of concentrically banded discs of hematite arranged at random (Plate 45, No. 1).

**Middle Owen:**

(122-244 m.) **Hematite Conglomerate:**

Coarse pebble to boulder conglomerate with chert and hematite pebbles; more finely grained and less hematitic away from the mine area.

(61-122 m.) **Pink or red, slightly hematitic sandstones.**

**Lower Owen:**

(+ 244 m.) **Mainly yellowish grey, pebble-to-boulder conglomerate with minor pink or grey sandstones.** Parts of this conglomerate are extremely coarse and chaotic (Plate 45, No. 2).

**Jukes Breccia or Conglomerate:**

(0-61 m.) **Paraconglomerate of mainly volcanic**

Plate 45 No. 2: - Coarse phases of the Lower Owen Conglomerate, near Mt. Sedgwick.
pebbles, cobbles and boulders
in greywacke matrix.

_**Cambrian (?) Mt. Read Volcanics:**_

(+ 1,525 m.) Keratophyres, finely banded tuffs,
thin sandstones and mudstones, etc.

The _Jukes Breccia (or Conglomerate)_ is absent or very
thin in the mine area but it is clear from sections on Mt. Owen and
Mt. Sedgwick that it thickens rapidly eastward to over 500 metres.

Typical conglomeratic phases consist of boulders of keratophyric
lavas and fine-grained tuffs scattered in open framework in a poorly
sorted pebbly greywacke matrix consisting of small volcanic
fragments, albite and volcanic quartz crystals, and sericite and
chlorite (31104, 31105, 31106, 32570, 32875). At the eastern end
of the Linda Valley are coarse, poorly sorted sandstone lenses
within the conglomerate and some lenses contain bands several mm.
thick that are rich in magnetite (by X-ray diffraction). This occurs
as rounded grains up to ½ mm. diameter in a quartz-sericite matrix.

In some specimens, small pebbles of volcanic rock much like the
Darwin keratophyre are prominent. These, together with the rich
magnetite lenses, suggest a source such as the keratophyric rocks
on Mt. Sedgwick that are intruded by hematite-magnetite veins.

The only probable occurrence of the _Jukes Conglomerate_ in
Fig. 34: Topographic map of the Mt. Lyell mining area.
Fig. 35: Geological map of the Mt. Lyell mining area.
Pleistocene Moraine

Copper Clays

Goethite, Siderite, Native Copper

Upper Owen Conglomerate

Pioneer Beds Grey Sandstone

Pioneer Beds Conglomerate, Sandstone

Haulage Unconformity

Chocolate Sandstone with Hematitic Oolite Beds

Middle Owen Conglomerate

Lyell Schists

Chlorite-Quartz Schist

Outcrop of Orebody

Outcrop of Pyritic Schist

Volcanics

Lahprophyre Dyke

Conglomerate Including Hematite, Hematite

North Lyell Conglomerate

Upper Owen Conglomerate

Coarse Grained Conglomerate

Unoriented Owen Conglomerate

Trace of Cleavage
Fig. 36: Geological map of the Queenstown district.
the mine area is near Comstock. The workings of the Comstock mine have exposed a body of rather massive rock adjacent to the conglomerate contact and thickening in depth (see Fig. 57). This body was described by Edwards (1939) as a quartz-porphyry dyke but examination of specimens on dumps (32691, 32899, 32774, 32782) shows it to be fragmental, partly conglomerate and partly coarse sandstone. It is poorly sorted and contains many chert and volcanic rock fragments, volcanic quartz and altered albite crystals, all angular or subangular and distributed in a sericitic or chloritic matrix. Some specimens reveal considerable granulation and also orientation of sericite by cleavage development, though in hand specimen the rock is quite massive and tough.

The thickening in depth indicates thickening to the east, and the rock is interpreted as being equivalent to the Jukes Conglomerate.

**Deposition of the Owen Conglomerate**

Available information on thicknesses in the mine area has been summarised by Banks (1962 b, p. 155, from Wade and Solomon, 1958) and is reproduced here in Fig. 37.

The mine area marks the western margin of the Owen trough and at Queenstown only about 6 m. of quartz sandstone overlies the
Fig. 37: Columns to show thickness and rock-type variations in the Owen basin near Queenstown.

Modified from Banks (1962b).
1 QUEENSTOWN
2 WEST END OF MT. LYELL
3 NORTH LYELL
4 RAZORBACK
5 WEST END OWEN SPUR
6 WEST SIDE MT. OWEN
7 LINDA SPUR
8 CENTRAL OWEN SPUR
9 CENTRAL MT. OWEN
10 MT. LYELL
11 EAST END MT. LYELL

- POINTS WITHIN COLUMNS ARE AT GEOGRAPHICALLY CORRECT POSITIONS AND ARE ARBITRARILY PLACED AT BASE OF UPPER OWEN. LINES JOINING THESE POINTS PROBABLY REPRESENT HORIZONTAL LINES AT BEGINNING OF DEPOSITION OF UPPER OWEN
Mt. Read Volcanics. Within the trough the Owen (particularly the Lower Owen) is thickest on the west side and on Mt. Sedgwick reaches a thickness close to 1,200 m. (4,000 ft.) even allowing for possible repetition by faulting (Fig. 38). At Queenstown it thins at a rate of about 1 m. for every 3 m. horizontal distance and this must surely require a barrier to the west in this area (cf. Banks, 1962b). The filling of the Owen basin is illustrated in a series of sections shown in Fig. 39.

There seems little doubt that the upper half of the Owen Conglomerate is marine (see Banks, 1962; Solomon, 1959) but the lower half, and the Jukes Conglomerate, may well be of terrestrial origin as suggested by Campana et al. (1958).

The Hematite Conglomerate and its Significance

A particular feature of the Owen Conglomerate in the Lyell area along the Upper Owen-Mt. Read contact is the presence of thick, coarse conglomerates, very poorly sorted, that contain subangular cobbles and pebbles (and rare boulders) of chert, hematite, and hematite with barite, in addition to the usual vein quartz, quartzite and quartz schist (Plate 46, No. 1). The irregularity in shape and lack of rounding are illustrated in this plate, in which most of the
Fig. 38: East-west geological cross-section of Mt. Sedgwick. This illustrates the upturned Owen Conglomerate on the west side of the Owen Basin, and the thickness variations in the Jukes and Lower Owen Conglomerates.
Fig. 39: Sections to illustrate the development of the Owen basin in the Mt. Lyell area.
DUNDAS RIDGE

SL

Lateritic surface

UPPER OWEN BASIN

UPLIFT IN WEST, MUCH HEMATITE TO BASIN, SLOWER MARINE DEPOSITION

SL

MIDDLE OWEN

SLOWER DEPOSITION. LOCALLY MATERIAL FROM WEST REPRESENTING RENEWED UPLIFT. DETRITAL HEMATITE INTO BASIN.

SL

LOWER OWEN

LOWER OWEN CONGLOMERATE RAPID DEPOSITION, IMPROVED SORTING, NO DETRITUS FROM WEST

SL

JUKES

JUKES CONGLOMERATE - BRECCIA RAPID DEPOSITION, POOR SORTING, DETRITUS FROM EAST & WEST
Plate 46 No. 1:— Sheared chert-hematite conglomerate of the Middle Owen Conglomerate, north end of Tharsis Ridge.

Plate 46 No. 2:— Coarse conglomerate with very hematitic matrix overlying normal quartz conglomerate, south end of Razorback Ridge.
pebbles are hematite and chert. Parts of this horizon consist of chert, quartz, and quartzite pebbles in a very hematitic sandstone matrix (Plate 46, No. 2).

The hematite pebbles (31025-34, 31075, 31079) consist of varying proportions of hematite, quartz and barite. Some are an irregularly crystallised aggregate of quartz, hematite and barite, others consist of botryoidal masses of hematite with barite filling cores and also interstitial pools. Some hematitic pebbles (31080-82) consist mainly of a very fine-grained cherty quartz mosaic cut by hematite veinlets and containing irregular patches and rings of hematite. Most contain barite in the quartz mosaic. These are identical to the hematite-veined chert outcropping in the North Lyell and Comstock areas. A chemical analysis of a typical pebble from Razorback Ridge is given in Table 17, No. 8; analyses of other pebbles by the writer gave 3.15, 4.60, and 2.23% BaO.

The matrix to the pebbles is generally a poorly sorted sandstone of hematite and chert fragments in quartz grains cemented by hematite.

At Comstock the Hematite Conglomerate (chert and rare hematite pebbles in hematitic quartz matrix Plate 47, No. 1), forms
Plate 47 No. 1:—Hematitic conglomerate, 300 m. south of the Comstock Open Cut.

Plate 47 No. 2:—North Lyell hematite, showing bedding, and pebbles of chert. The pencil is 12 cm. long.
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<td>Total</td>
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Table 17: Hematites, Mt. Lyell
Table 17: Hematites, Mt. Lyell (continued)

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1. Iron Blow hematite (31045).
2. Iron Blow hematite, No. 2 level (32696).
3. North Lyell hematite (31182).
4. North Lyell hematite (32940).
5. North Lyell hematite (31175).
7. North Lyell hematite near base (31734).
8. Pebble of hematite in Razorback Conglomerate (31025).

Oxide analyses by Dept. of Mines, Tasm.
Trace element analyses by Aust. Mineral Development Labs.
a wedge thickest near the ore outcrops and thinning to zero about 400 m. away (Fig. 35). The ore is overlain by a pipe-like body of chert cut by hematite veins, and the conglomerate is clearly detritus from this material. Richly hematitic conglomerate with hematite pebbles reappears between North Lyell and the Iron Blow and thins out rapidly to the east and south. 300 m. south of the Iron Blow (on the west end of the Owen Spur) the hematite is almost entirely in the matrix and this hematitic material passes to normal pink sandstones and conglomerates within a few hundred metres.

The Hematite Masses of North Lyell and the Iron Blow

Massive bodies of hematite occur at North Lyell and the Iron Blow. The Blow mass has been removed by mining but the other is largely untouched. Both lie immediately against, and stratigraphically under, the Upper Owen "Chocolate" sandstone and are underlain by schistose Mt. Read Volcanics (Fig. 55, 45).

Both bodies are similar in size and shape, being tabular, about 90 m. long and up to 12 m. thick in plan, with vertical extents of 120 m. or so; they each amount to some 400,000 tons. They peter our in depth, which means, after allowing for unfolding the steeply dipping adjacent sediments, that they originally wedged
out and vanished to the east.

The Iron Blow hematite (30997, 31043 and 31044) can now only be collected from a dump near the open cut. It is steely grey but with white patches of barite that occurs irregularly distributed or in crudely defined bands; barite content varies from almost nil to at least 50%. The barite and hematite appear to have crystallised together. Some specimens show a crude banding in the hematite which appears to be due to elongate lenses.

The Iron Blow formed a bold tor overlooking the Linda Valley and was assumed to be the source of the alluvial gold found further downstream in 1883. However, after mining had commenced it was soon realised that the gold was derived from a 10 m. layer of limonitic gossan lying immediately east of the hematite (Fig. 55). However, Peters (1893), Johnston (1887) and Thureau (1886) referred to gold in the hematite and Allan (1893) quoted 16 dwt. Au, and 1 oz. 14 dwt. Ag per ton of hematite. A specimen from No. 2 level (32696, collected in 1910) shows relatively high Cu and Ag (Table 17 No. 2). Peters (1893) and Gregory (1905) thought the Iron Blow originally contained copper and silver which had been leached out and deposited in the adjacent sulphide ore.

The North Lyell hematite (31075, 31166-82, 31734) differs in that it contains less barite and shows distinct conglomeratic or
breccia texture, with locally some crude compositional layering parallel to the contiguous Owen sandstones (Plate 47, No. 2). The contact with the Owen sandstones is sharp and parallel to sandstone bedding. The pebbles of the conglomeratic phases, which generally only show on weathered faces, are chert and hematitic chert from 10 cm. to microscopic size (Plate 47, No. 2). The matrix is composed of quartz and hematite apparently somewhat recrystallised to a confused aggregate. In parts of the mass the rich hematite matrix has pronounced colloform or microbotryoidal texture, in some cases centred on chert fragments (Plate 48, No. 1). Where cleavage is well developed, as it is in places, the botryoids are drawn out to give a crude banded effect parallel to the cleavage or are cut off by cleavage surfaces (Plate 48, No. 2).

Barite occurs in blebs or lenses in the hematite-quartz rock, partly in interstitial pools between colloform masses of hematite (Plate 48, No. 1) and partly intergrown with quartz and hematite.

Edwards (1939) stated that the North Lyell mass carried 0.17 grains Au/ton. Analyses made for the writer (Table 17) show that Au, Ag and Cu occur only in very small amounts except in 31166. However, this specimen is from close to the schist contact and the copper may be introduced. The high phosphorus contents are note-
Plate 48 No. 1:— Colloform hematite with interstitial dark "mottled" barite. X5.

Plate 48 No. 2:— Pre-cleavage, botryoidal hematite in North Lyell mass. X15.
worthy and will be referred to later.

Edwards (1939) noted that this hematite was slightly magnetic and polarised and the writer has confirmed that parts of the rock are slightly magnetic. Chemical compositions show up to 2.7% FeO (Table 17) and in some X-ray patterns (31739) possible magnetite lines were read. Although nothing but hematite was revealed in polished sections it is possible that small amounts of magnetite or maghemite exist. A few percent of maghemite would account for the FeO content, the magnetite lines and the relatively high magnetic intensity.

Dr. R. Green (University of Tasmania) kindly determined the direction of the polarity in six North Lyell hematite specimens and proved it to correspond to a Tertiary axis. Presumably this is some near-surface effect produced by chemical changes due to prolonged weathering (chemical remanent magnetism). A few hematite specimens were gathered from underground in the North Lyell fault area. These gave variable (but not Tertiary) results, perhaps as a result of tectonic disturbances.
The Origin of the Hematite Bodies

Earlier writers have expressed widely divergent opinions as to the origin of these hematite masses, including "segregation" fissure lodes (Power, 1892), oxidation of pyrite in situ (Peters, 1893); dehydration of limonite derived from oxidation of pyrite (Thureau, 1886; Johnston, 1887; Ward, 1890; Montgomery, 1893); deposition from ascending solutions carrying ferrous sulphate leached from pyrites, meeting descending meteoric (?) waters (Gregory, 1905, p. 99); leaching by vadose waters of hematite from the adjacent Owen Conglomerate during Devonian earth movements (Hills, 1927); deposition from hydrothermal "ore" solutions, the iron being primary magmatic or leached from the adjacent porphyries (Edwards, 1939); deposition of a "basic", ferruginous "front" as part of the sulphide mineralization (Bradley, 1954, 56, 57); transfer of iron from adjacent volcanics and ferruginous sediments during sulphide mineralization (Wade and Solomon, 1958). Nye et al. (1934) suggested the hematite is "evidently therefore of primary and secondary origin".

Bradley (1954, p. 230) refers to a suggestion by Dr. Opik that the hematite pebbles in the Owen Conglomerate are derived from erosion of a lateritic surface.

A source for the North Lyell hematite mass is not hard to
find. Reference has already been made to hematite-magnetite veins in the Darwin keratophyre of Mt. Sedgwick and similar veins have been found in chert at Comstock and on/northern end of the Tharsis Ridge. Those at Comstock (Plate 49, No. 1) consist largely of hematite (Table 18 No. 2), with barite and stringers of quartz and chert (31093, 32941, 31199). The veins are crumpled (Plate 49, No. 2) and locally sheared. They occur in a large pipe-like mass of fractured chert that is the source of chert pebbles and fragments in the nearby Owen Conglomerate. All the specimens have a proportion of strongly magnetic material and X-ray patterns show the strong magnetite lines. The presence of several magnetite lines and high magnetic intensity in a rock carrying little FeO suggests perhaps the presence of maghemite that may be formed as an oxidation product of magnetite under conditions of rapid oxidation (Lepp, 1957).

The veins of Tharsis Ridge (Plate 50, No. 1) are up to 150 cm thick and composed of hematite (31195, 31196, 31197, 31095, Table 18, No. 1) with little or no gangue except for very small patches of barite. Much of the hematite shows botryoidal texture with radial and concentric structure, individual botryoids being up to 1 cm across (Plate 50, No. 2) and generally elongated in the strong Devonian cleavage. This texture, then, is at least pre-
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Cu (p.p.m.) | 700 | 600 | 200 | >10,000  |

Au   | 3   | 3   | 5   | 3

Ag   | 1   | 0.1 | 1   | 10

Mn   | 200 | 300 | 400 | 300

Ti   | 100 | 100 | 100 | 200

p    | 40  | 200 | n.dt. | <2,000  |

(from above)
Table 18 (continued)

1. Tharsis Ridge hematite vein (31196).
   Analyst: Dept. of Mines, Tasm.

2. Comstock hematite vein (31199).
   Analyst: Dept. of Mines, Tasm.


   Trace elements by X-ray spectrograph,
Plate 49 No. 1:-- Hematite veins in chert at Comstock.

Plate 49 No. 2:-- Crumpled hematite veins in Comstock chert.
Plate 50 No. 1: Vein of hematite in chert on Tharsis Ridge. The lens box measures 5 x 5 cm.

Plate 50 No. 2: Botryoidal texture in hematite veins on Tharsis Ridge, X15.
Devonian. Within kink bands or shear zones in the cleavage the hematite appears to be recrystallised to specularite; where these zones cut chert, recrystallisation to a coarser grain has occurred.

The lower FeO content of the Tharsis material and the very low intensity of the magnetite lines in X-ray powder photographs indicate that less magnetite (or maghemite) is present than in the Comstock veins.

Higher percentages of magnetite occur as residual cores in the hematite-magnetite veins in the Darwin keratophyre on Mt. Sedgwick (Plate 28, No. 2). Polished sections (31254, 31255, 31256) and X-ray photographs indicate approximately equal quantities of hematite and magnetite, with hematite ringing magnetite cores and replacing magnetite along preferred crystallographic directions. The hematite could have been formed by due oxidation/to passage of volcanic gases at fairly high temperature, or to prolonged weathering.

The botryoidal textures in the Tharsis veins show both concentric banding and radial crystals (Plate 50, No. 2) and are similar to colloform textures. Generally, however, the bands are thick, are not scallopped and the radial crystals are confined to individual layers, so that they are more likely to be crustification
textures (Edwards, 1956, p. 21-28). Whether formed from a gel or not, they develop from solutions. Iron hydroxides commonly precipitate as gels to form amorphous limonitic material, or crystallise to goethite or even hematite (see Edwards, 1956, p. 20; Lindgren, 1935). "Colloform" hematite is rare according to Edwards (p. 28) but may develop by dehydration of iron hydroxides during heating to about 125°C in neutral conditions (Smith and Kidd, 1949), or even lower temperatures (about 100°C) if metamorphism is prolonged or conditions slightly acid (Tunnel and Posnjak, 1931). Hematite likely to have been derived from "limonite" has been described from the Rawlins iron ore deposit in Wyoming (Lovering, 1929). Much of the hematite is mixed with limonite and is powdery or loose and "soft" ores of this type occur widely in the Appalachians (U.S.A.). Lindgren (1933, p. 356) and others have suggested they are residual after siderite but Chance (1908) suggested that many of them were fossil gossans. Lovering (1929, and pers. comm. 1961) believed much of the soft hematitic ore at Rawlins was formed by dehydration of limonite that had developed from weathering of siderite, glauconite, etc. The dehydration hypothesis helps to explain what seems to be an abnormally high concentration of hematite for a purely residual deposit.
Colloform or crustification hematite is likely to be supergene but could be hypogene. For example there is some controversy over the origin of the Cumberland (England) hematite deposits which show world-renowned examples of botryoidal hematite ("kidney ore"). Trotter (1945), Dixon (1927) and others favoured a magmatic origin (lateral migration of ore solutions through Carboniferous limestone) but Smith (1924, 1927), Holland (1962) and others favoured deposition from iron-rich solutions percolating down from the Permo-Trias cover. Trotter believed the hematite was deposited from ferric chloride solutions along with barite and iron, and copper and lead sulphides. Both Trotter and Smith suggested a post-Triassic, possibly Tertiary, age for the mineralization.

Despite the presence of sulphides in association with hematite no one has suggested that the hematite might be secondary after sulphides and derived from goethite developed prior to, or during the early stages of, deposition of the Permo-Trias cover.

Some support for this idea is given by Dubois (1962) who found the hematite possessed a chemical (?) remanent magnetism of Permo-Triasic polarity.

Colloform or crustified hematite (partly specularite) occurs in a lead-zinc mine at Balmat (U.S.A.) and Brown (1936) believed
this hematite to be part of a fossil gossan.

On the Tharsis Ridge, Owen sandstones lap onto the veined chert, and as at Comstock, bouldery chert-hematite breccias wedge away from the veins. Hence the present outcrops were probably exposed in Owen time and the botryoidal textures may well have developed in the near-surface parts of the veins by weathering, with the development of a "limonite" crust containing small quantities of insoluble barite, and a small percentage of residual magnetite. Possibly around such outcrops there developed a limonitic chert scree, which, where particularly ferruginous, gave rise by dehydration to hematitic bodies like that at North Lyell. Boulders of limonite-barite crust may well have been shed from the vein outcrops to join the surrounding quartz and quartzite gravel.

The Iron Blow is rather different from the North Lyell mass; it is richer in barite, is not conglomeratic (and hence is low in silica) and has no pronounced textural features other than a crude banding due to elongated ovoid structures (botryoids ?). Its stratigraphic position is identical, judging from the early geological reports. No firm conclusions can be drawn about the origin of this deposit but a reasonable hypothesis is that it is a dehydrated limonitic mass derived from the weathering of an iron-rich source (iron-oxide veins ?
iron sulphides ?). The implied climatic conditions would certainly favour lateritic soil development over a wide area and erosion of such material may have produced the ferruginous cement widely present in the Middle and Upper Owen.

In order to further test the suggestions made above, some trace element variations were briefly studied.

**Phosphorus** tends to be enriched in gossans compared to the source material (Williams, 1934; Hill, 1962) and as shown in Tables 17 and 18 the P content of the North Lyell and Blow hematites is at least 10 times that of the veins (and incidentally, of the nearby sulphide bodies), which contain less than 2,000 ppm.

**Titanium and Manganese** are likely to be enriched by weathering (Rankama and Sahama, 1950, p. 563 and 649) and a tendency in this direction is noticeable in the North Lyell hematite, particularly with titanium. Again the enrichment is strong relative to the concentrations of these elements in the sulphide ores.

Reworking and recrystallisation of barite is likely to reduce its **Strontium** content and spectrographic analyses of Sr contents were made on the barites from several primary veins and from the North Lyell and Iron Blow hematite bodies. It can be seen (Table 19) that there is a tendency for the Sr in supposed secondary rocks to be lower but the test is hardly conclusive.
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<th>Sr% in Barite in Secondary Bodies</th>
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<tr>
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Barite vein from 31189 Philosophers Ridge. From Iron Blow.

Barite vein from 31186 North Lyell mine. From North Lyell hematite.

Barite vein from 32499 Intercolonial Spur. Pebbles in Razorback Conglomerate.

Analyses by X-ray spectrograph, Aust. Mineral Development Labs.
Probably the most interesting point about the two hematite bodies is their distribution in relation to the sulphide ores. The richest concentrations of sulphides in the Lyell area are the Iron Blow pyrite-chalcopyrite orebody, the tightly bunched group of pyrite-bornite-chalcopyrite orebodies at North Lyell and the group of en echelon orebodies at Comstock, all three areas being the sites of ferruginous bodies. The areal coincidence may mean that these areas were favourable both to development of iron oxide veins in the Cambrian and sulphides in the Devonian. There is also the possibility that these places were the sites of sulphide accumulations in the Cambrian and that the weathering of these gave rise to limonitic screes and/or gossans. This suggestion might apply particularly to the Iron Blow which is apparently not a conglomerate and is an area free of iron oxide veins.

The Chert or Chert Breccia of North Lyell and Comstock

The Comstock chert (31094, 31744) forms a pipe-like mass adjacent to the Comstock orebodies; it is some 600 (185 m.) x 400 ft. (120 m.) in plan and known to have a dip-length of at least 1,000 ft. (300 m.). The North Lyell mass (30957, 30958) forms the hanging wall to many of the North Lyell orebodies (Fig. 48) but is of no regular shape and its partial erosion in Owen times and subsequent
fracturing and disruption prevent determination of its original shape. In general, the cherts have not proved favourable hosts for sulphide deposition though they occur within, or adjacent to, mineralised schists.

The chert is milky grey or pinkish when fresh and weathers white. It has a characteristic finely brecciated or closely fractured appearance and in many areas (e.g. the North Lyell Open Cut) these fine fractures have a film of hematite. Parts are veined by hematite and at North Lyell it is traversed by many crush-breccias with hematite matrix (e.g. 32713) varying from microscopic to several tens of cm. wide.

In thin section, the chert is an interlocking, sutured quartz aggregate (individuals averaging 0.02 mm. diameter) with clear, recrystallised patches (< 0.15 mm. across) of coarser grain (Plate 57, No. 2). Many of these have a fringe of hematite and include needles of hematite. Barite occurs in patches and in some specimens there are tiny crystals of pleochroic chlorite.

Rather coarser aggregate (average 0.03 mm.) forms a rock underlying Upper Owen sediments on the Whaleback Ridge (31145) and also in Waterfall Gully (31124). It is hematitic and much like the North Lyell chert with coarser recrystallised patches.

Similar chert is common as fragments in the Owen Conglomerate and several pebbles were clearly fractured and veined by
hematite prior to Owen deposition (e.g. 30951 from Tharsis Ridge). Coarse chert breccias with pink sandstone matrix occur where the Chocolate sandstone rests directly on the North Lyell chert some 100 m. east of the old North Lyell Shaft, and chert-hematite breccias occur on the north end of the Tharsis Ridge, where older Owen Conglomerate rests on chert.

There are pebbles in the Hematite Conglomerate that consist of a slightly hematitic, slightly sericitic quartz aggregate similar to, but coarser than, all but the recrystallised patches of the chert. Some of these pebbles are difficult to distinguish from slightly more sericitic quartz aggregates that contain euhedral quartz crystals and look very much like the boulders of the Philosophers Ridge schists.

Apparently all these highly siliceous rocks, many with hematite, were developed prior to the Owen Conglomerate.

There is little to indicate the origin of the cherts. Their form and lack of banding point away from a sedimentary origin but the association with volcanic rocks points to a volcanic origin. Possibly they represent siliceous material developed in and around geyser zones during the later phases of Mt. Read volcanism in this area.

Cherts of this nature have not been found elsewhere in the Mt. Read Volcanics and they (as with the hematite veins) appear to
be closely connected with sulphide development.

The Mt. Read Volcanics and the Lyell Schists

The host rocks to the Lyell orebodies are a diverse group of sericite-chlorite-quartz rocks known locally as the Lyell Schists. They crop out over an area of one sq. mile and occupy an aureole of hydrothermal alteration (Fig. 35 and 36) in the Mt. Read Volcanic Arc. They merge at the aureole margins to unaltered rocks of the Mt. Read Volcanics. On regional evidence and a reconnaissance microscopic examination the schists were identified as volcanics and greywackes in a previous thesis (1957).

Gregory (1905) was the first to suggest the schists were volcanics and his views differed from all preceding opinions which ranged from sheared intrusives (Twelvetrees, 1902) to "volcanic muds" (Thureau, 1886). Although Hills (1927) and Hills and Carey (1949) repeated Gregory's views, Edwards (1939), Nye et al. (1934), Conolly (1947) and Alexander (1953) considered the schists were intrusive porphyries. Bradley (1954, 1956, 1957) believed the schists were mainly metasomatised sediments,( including Owen Conglomerate) but Wade and Solomon (1958) returned to the volcanic viewpoint. A somewhat similar swing in opinion took place at Rosebery where, for the Rosebery Schists, Twelvetrees and Petterd
(1899) initiated a volcanic viewpoint, Dallwitz (1946) and Finucane (1932) then favoured an intrusive origin, and Hall et al. (1953) returned to a volcanic origin.

Detailed mapping (1:1200 and locally 1:600) of the schists has revealed distinct lithological boundaries and enabled certain textural and compositional rock types to be traced for short distances. Boundaries are generally only visible on deeply weathered surfaces.

Quartz Chlorite Schists

Distinct chlorite-quartz "beds" may be seen in, and south of, the West Lyell Open Cut interbedded with banded sericitic schists (Fig. 35, Plate 51, No. 1). The chloritic beds are up to 6 m. thick and individual beds can be followed for several hundred metres. They are featureless, dark green rocks (30924, 30938, 30940, 31001, 31190, 32744, 30920, 31023, 31701 etc.) consisting mainly of a quartz-chlorite mosaic. The average grain diameter of the quartz is between 0.05 and 0.1 mm. and scattered randomly through this aggregate are slightly rounded to almost circular kernels or patches of coarse quartz aggregate, up to 2½ mm. across and 5 mm. in length in section giving a crude porphyritic appearance to the rock in thin section (Plate 51, No. 2). Most of these coarse
Plate 51 No. 1:— Chloritic schist "beds" in sericitic schist, south of West Lyell Open Cut.

Plate 51 No. 2:— Lenticles of quartz aggregate in quartz-chlorite schist (31700). X15.
kernels are fairly clear but a few are cloudy and have a lath-like form as if replacing feldspar. Although the rocks are cleaved there appears to have been considerable recrystallisation following deformation and many of the quartz crystals show sharp extinction.

The bulk of the chlorite is pleochroic from medium green to pale, yellowish green but some of the chlorite is dark green. The $\beta$ refractive index of the more strongly coloured material is approximately 1.648, the paler about 1.626. The birefringence is low to nil and three $2V$ measurements were between 5 and $10^\circ$ negative. The analyses (Table 20) of these schists clearly show that the chlorite is rich in iron.

As it proved impossible to achieve a perfect separation of the chlorite, attempts were made to analyse it by a semi-quantitative X-ray technique suggested by Schoen (1962). The atomic scattering factors for the Si-Al-Mg group are half as great as those for the Fe-Mn-Cr-Ti group, thus changes in the Mg and Fe contents cause marked changes in peak intensities. Using a theoretical formula of $(\text{Mg, Al})_{12-X}(\text{Fe, Mn, Cr, Ti})_X(\text{Si}_{8-y} \text{Al}_y)_0 20(\text{OH})_{16}$, values for $X$ may be determined by comparing theoretical and calculated (from diffractogram) structural factors for the (001), (002) and (003) reflections.

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Table 20: Chlorite Schists (continued)

1 and 2. Quartz-chlorite schist, near mineralized zone in Conglomerate Creek, 1 upstream from 2, specimens 31699 and 31700.


5. Dark green chloritic "bed", south of West Lyell Open Cut (31112).


8. Average of analyses 1 - 7.
in an aluminium sample holder measuring 4 cm. x 2 cm. x 1 mm. The diffractometer was completely re-aligned and standardised before the chlorites were beamed. The Lorentz-Polarisation factors were obtained from the International Tables for X-ray Crystallography V. 2 (1959).

Three analysed chlorites were available to the writer (31004, 31801 and 32871) and the X-ray results for these chlorites were within the accuracies obtained by Schoen (Table 21). All three would be classified as ripidolites by Foster (1962).

Determinations of chlorites in the Lyell Schists gave 8.3 Fe 3.7 Mg (31112, 31699), 8.4 Fe 3.6 Mg (31108) and 6.7 Fe 5.3 Mg (30940), confirming that they are iron-rich varieties. Though their refractive indices show a considerable variation, they are higher than any recorded in the unaltered volcanics.

In horizontal sections (which are perpendicular to the cleavage) some of the flakes of the chlorite aggregates have parallel orientation but many are arranged at random. In steep sections cut cleavage a more pronounced parallelism is visible due to streaked lenses of chlorite aggregate but even within these lenses a considerable portion of the mineral plates shows no preferred orientation. This suggests that some of the chlorite has crystallised after cleavage development. Very fine-grained sericite is included
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**X-ray Chemical Analysis**

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1. Vein in keratophyre, Comstock tram. Specimen 31004.
2. Chlorite in quartz vein, Gormanston-West Lyell road. Specimen 31801.

Analyses by Dept. of Mines Tasm.
in the chloritic streaks, the proportion of sericite to chlorite varying considerably.

Hematite and magnetite occur in these rocks as needles, irregular patches or veinlets. In many rocks (e.g. 31143) veinlets have been disrupted by the cleavage and iron-chlorite has crystallised in and around iron oxide-rich areas.

In some rocks it is clear that the early quartz kernels have grown by secondary crystallisation of quartz, partly replacing earlier crystals. For example, in 31074 a euhedral quartz crystal outline may be seen in the centre of a kernel but the individual units of the kernel have crystallised independently of the euhedral crystal margins.

Near the Glen Lyell workings, a thin dark chloritic schist bed (32744) reveals a fragmental texture in which quartz grains up to 1 mm. diameter are scattered through a chlorite-quartz base. Most of the grains are of no particular shape but some are like embayed quartz crystals in volcanic rocks and several have perfect lath shapes, as if they had replaced feldspar. A few chloritic patches have well defined lath-form and may also be replaced feldspars. The fragmental texture, the lath outlines and the embayed quartz crystals suggest the rock may originally have been a crystal tuff or a greywacke.
Near 6600S/300W (Fig. 35), a bed of rather pale green rock (31052) consists of a chlorite-sericite matrix containing subangular, even grained (0.1 mm. across) crystals of relatively fresh albite and quartz. This rock may have been a lava. Small patches of the matrix are very fine-grained quartz aggregate and it appears that the coarser quartz-chlorite aggregates of such rocks as 30938, 30940, 30924 etc., are more advanced stages of this crystallisation.

These fragmental rocks grade to beds which are almost certainly quartzose greywackes (30919, 30920, 31023). 30919 is a poorly sorted sandstone of subangular quartz grains in a chlorite-iron oxide matrix (Plate 52) and it forms a bed 15 cm. thick and extending for several metres in sericitic schists. Specimen 30920, from the same bed, is less obviously fragmental, being coarser grained and with a more plentiful chloritic matrix. Magnetite crystals up to 0.3 mm. diameter are scattered throughout (magnetite confirmed by X-ray). Specimen 31023, from a similar bed nearer the Iron Blow, is very similar to 30920.

All these chloritic rocks form distinctive sub-parallel beds on Philosophers Ridge and environs, and at least some of them appear to be greywackes or tuffaceous rocks.
Plate 52: Quartzose greywacke (?) forming a bed in quartz sericite schist, south of West Lyell Open Cut. X55.
Sericite-Fleck Schists

Somewhat similar chloritic rocks that in places form discrete "beds" have been mapped as "sericite-fleck schists". They are dominantly medium or dark green but contain pale grey streaks and lenses of sericite (31112, 31151, 31709, 32740). They occur on Philosophers Ridge, in the West Lyell Open Cut and near North Lyell.

In 311709 from West Lyell, and 31151 from Philosophers Ridge, thin sections cut horizontally show crude porphyritic texture with lath-like masses or lenses of sericite up to 1 x 0.5 mm. in size, randomly arranged in a chlorite-sericite matrix with a little quartz (Plate 53, No. 1). The individual sericite flakes may be random or show parallel orientation. In thin sections cut parallel to the cleavage, the sericitic lenses are considerably stretched out in the plane of the cleavage with a steep pitch (Plate 53, No. 2).

In 32943, also from West Lyell, many of the sericitic lenses are seen to be sericitised albite crystals up to 3 x 5 mm., in which the twinning is still visible. The matrix in this rock consists of albite, chlorite, sericite and fine hematite dust. Most of the feldspar crystals have their long axes sub-parallel and in the plane of the cleavage or are cracked and slightly crushed.

These rocks appear to be more like lavas than the chlorite
Plate 53 No. 1:— Lenses of sericite in sericite fleck schist (31151), section cut horizontal (perpendicular to cleavage). X22.

Plate 53 No. 2:— Lenses of sericite in sericite fleck schist (31151), section cut perpendicular and parallel to cleavage. X5.
schists previously described and their relatively coarse grain indicates they may have been keratophyres.

Quartz Sericite Schists

The craggy outcrops along Philosophers Ridge consist of thick bands of quartz sericite schists interbedded with the relatively thin, dark green chloritic beds just described. The sericitic schists are more common than the chloritic schists. In places along this Ridge and also in part of the West Lyell Open Cut the sericite schists show a fine banding that is generally tightly folded.

The most common sericite schist (31048, 31049, 31103, 31118-20, 31148, 31129, 31161-64, 31183) consists largely of an interlocking quartz mosaic of rather uneven grain, individual crystals varying from about 0.03 to 0.07 mm. diameter. In some rocks (e.g. 31049) sericite is distributed fairly evenly and randomly through the quartz mosaic but in others (e.g. 31048, 31073) the mosaic has clearly been broken up by a crude cleavage, and sericite fills the rather irregular fractures. These more highly cleaved rocks also contain fracture or flowage zones parallel to cleavage that are filled with sericite. These zones tend to break the mosaic into lenticles, producing augen structure. The individual augen vary from microscopic size to several cm. wide and the coarsest
varieties of sericite schist are rather like deformed conglomerates.

Some quartz crystals of these rocks show undulose extinction but the majority do not and it is clear that crystallisation followed, or continued after, deformation. Under high magnification the sericite crystals within these flowage and fracture zones are not strongly oriented and a high proportion have a random arrangement. A little chlorite may be associated with the sericite.

The proportion of sericite is variable; where considerable, the degree of preferred orientation is strong and a schistose appearance more marked.

Several examples (31048, 31049, 31183, 31184, 31246) contain embayed and/or euhedral quartz crystals (Plate 54, Nos. 1 and 2) up to 1/2 mm across and identical to those of the quartz keratophyres. Many of them (e.g. in 31048) have secondary quartz fringes in optical continuity with the primary crystal.

Lenticular, almost lath-like, patches of sericite and chlorite occur in 30952 and may be relics of feldspar crystals and in one specimen (32762 from the West Lyell Tunnel at 2,025 ft.) part of a plagioclase crystal has survived alteration.

Barite crystals up to 0.1 mm are found occasionally in the mosaic.

The banding in these schists is generally only visible on
Plate 54 No. 1:-- Embayed quartz crystal with rim of secondary quartz aggregate in sericitic schist (31246C). X60.

Plate 54 No. 2:-- Embayed quartz crystals in sericitic schist (31049A). X50.
weathered faces (e.g. Plate 55, No. 1). The bands are identical, vary from 2 to 10 mm. thick and are separated by sericitic partings (visible in thin section) that tend to weather out. The banded schists are particularly well exposed on Philosophers Ridge immediately south of West Lyell Open Cut but extend into the east wall of the cut from where fresh material was obtained for analyses (Table 22, Nos. 1 and 2). Much of the banding is parallel to the chlorite beds just described but in many places rather complex, tightly appressed folds are visible. These, which in large part are more or less syngenetic, will be discussed under "Structure".

Some of the quartz sericite schists exposed in the West Lyell Open Cut, and particularly in the vicinity of the West Lyell No. 1 (Honeypot) orebody, have a distinctive pink tinge and much of it is coarsely augen. The quartz aggregate has a very "dusty" appearance, apparently due to the large number of inclusions (of iron oxide?), and this is probably the cause of the pink colour.

Another feature of these sericitic schists is the presence in places of "boulders" and "pebbles" (30921, 30922, 31050, 31051, 31247, 31688, 31687). These range from a cm, or so to about 50 cm, across and the large ones are irregular in shape and subangular (Plate 55, No. 2 and 56, No. 1). These bodies, a few of them
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<td>99.57</td>
<td>99.59</td>
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Table 22: Sericite Schists (continued)


2. Pinkish quartz sericite schist (Honeypot schist) from same area, 1956, with siderite.

3. Quartz sericite schist (31634) Conglomerate Creek, from mineralized zone.

4. Boulder in quartz sericite schist (31718), south of West Lyell Open Cut.

5. Average of analyses (1) to (3), after correcting for FeS$_2$ in (1) and (3).

All analyses by Dept. of Mines, Tasmania.
Plate 55 No. 1:— Banded quartz-sericite schist, from Philosopher's Ridge. X1.25.

Plate 55 No. 2:— Boulder in sericitic schist, south of West Lyell Open Cut.
Plate 56 No. 1:- Conglomeratic schist, Philosophers Ridge.

Plate 56 No. 2:- Banded quartz-sericite schist fragment in schist, Philosophers Ridge.
banded (Plate 56, No. 2) are scattered haphazardly in typical quartz sericite schist and are seldom in contact. Some of these outcrops are much like agglomerates (e.g. Plate 57, No. 1). In thin section, they consist of quartz aggregate of about 0.05 to 0.07 mm. diameter, clouded slightly by sericite and containing slivers and patches of chlorite. The sericite and chlorite is largely unoriented and the quartz mosaic rather clearer than in the matrix schists. Volcanic quartz "phenocrysts" are characteristic and some pebbles show sericite-chlorite laths (e.g. 30921 and 30922). A characteristic feature of the matrix is the presence of coarser "clots" that appear to be recrystallised patches of mosaic. Judging by the chemical analyses of one of them (31688 in Table 22, No. 4) they are considerably more siliceous than their schistose matrix and this matches their relative resistance to weathering and shearing. They are clearly not the products of cleavage deformation and apparently are of pre-Owen age because very similar but less sericitic material forms pebbles several cm. across in the Hematite Conglomerate (30918, 30925, 30970, 31006-8, 31211).

Similar bouldery and pebbly rocks (31242-31247) occur just above the Gormanston-Queenstown road, 2,000 ft. (600 m.) southeast of the mill, and these are close to the outer limit of
Plate 57 No. 1: Quartz keratophyre agglomerate (?) Waterfall Gully.

Plate 57 No. 2: North Lyell chert (30968), crossed nicols. X145.
hydrothermal alteration. Some of the pebbles are identical to 31688 but some have clearer, randomly oriented, lath-like masses of sericite (and a little chlorite) that are almost certainly altered feldspars. The fragments are scattered randomly through the matrix and form some 20% or so of the rock. The matrix is more finely ground, less schistose and rather more sericitic than the quartz sericite schist of Philosophers Ridge. These pebbly rocks are interbedded with banded tuffaceous (?) rocks and have an approximately longitudinal strike. The pebbly rocks may be nodular rhyolites (probably quartz keratophyres) or possibly agglomerates rather like those of Whip Spur and Waterfall Gully (see Plate 30, No. 2; 57, No. 1 and Plate 57, No. 1). The example from the latter area (31272, 31273), some 400 m. along strike to the east, consists of pebbles of pinkish quartz keratophyre in a streaky, fragmental base of quartz, feldspar, sericite and chlorite.

Albers et al. (1961, p. 51-52) have described slightly silicified quartz keratophyres in which the groundmass has been replaced by anhedral microcrystalline quartz, leaving quartz phenocrysts unaltered, and they believe this alteration may be the same age as the albitisation and other metasomatic processes. Possibly the Lyell rocks were silicified during Jukesian movements or perhaps soon after eruption.
Sericite, Batchelorite, Pyrophyllite and Paragonite

In the more intensely mineralised areas, and particularly at Comstock and North Lyell, there are concentrations of sericite-like material in fracture zones, and massive specimens of pale green or pale brownish material may be obtained. This pale green material is slightly translucent when thin, has a rather greasy feel, is soft and generally crudely foliated. It is extremely fine-grained, and irresolvable under the microscope. The brownish material differs in colour and is not translucent. The green material has been previously referred to as batchelorite, delessite, fuchsite, damourite, margarodite and hydromica, and the brown material as agalmatolite or pyrophyllite.

Recent work by Threadgold and Edwards (1957), Bothwell and Moss (1957), Hale (1958) and the writer has shown that most of this material is sericite and some is pyrophyllite, in some cases mixed with chlorite. Sericite includes fine-grained muscovite and paragonite according to Yoder and Eugster (1955) but in this thesis is taken to be fine-grained muscovite only.

Threadgold and Edwards, after examining six specimens submitted to them by the writer in 1956, made the following determinations by X-ray powder methods:
32961: Sericite.
32887: Poorly crystalline sericite.
32889: Sericite and pyrophyllite (7-10%).
32958: Pyrophyllite.
32959: Pyrophyllite and sericite (10%) and chlorite (5-7%).
32962: Sericite.

Available chemical analyses on these and other specimens are shown in Table 23. Threadgold and Edwards pointed out that 32961 is actually richer in Na$_2$O than K$_2$O and concluded that sodium ions had replaced potassium but not to sufficient extent to cause changes in cell size. According to Eugster and Yoder (1955) muscovite is only capable of taking up to about 2% Na$_2$O into solid solution so that higher Na$_2$O contents probably indicate the presence of paragonite mineral. Specimen 32963 with 5.52% Na$_2$O, and also 32888 with 4.37% Na$_2$O, failed to yield any lines characteristic of paragonite (using the data of Yoder, 1959) yet it seems likely that over 5% paragonite is present. Failure to detect the mineral may be due to its very low order of crystallinity, a feature of most of the sericites.

The writer’s X-ray work on part of specimen 32887 (No. 10 of Table 23) showed the presence of pyrophyllite in addition to the
Table 23: Sericite and Pyrophyllite

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Table 23: Sericite and Pyrophyllite (continued)

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<td>98.25</td>
<td>99.99</td>
<td>98.25</td>
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Table 23: Sericite and Pyrophyllite (continued)

5. Sericite, North Lyell (32962). Analyst: Mt. Lyell Co. in 1910, alkalies determined as $K_2O:9.74$, $Na_2O:0.21$ in 1957 by CSIRO, Mineragraphic Investigations, Melbourne.
10. Sericite and pyrophyllite, North Lyell (32887). Analyst: Mt. Lyell Mining Co. Ltd., 1910 (?). Alkalies determined as $K_2O:3.70$, $Na_2O:3.70$ by CSIRO, Mineragraphic Investigations, Melbourne.
sericite recorded by Threadgold and Edwards.

Bothwell and Moss (1957) analysed and X-rayed a sample of batchelorite from Mt. Lyell, and identified it as sericite, thus questioning the validity of the name batchelorite. Their chemical analysis is given in Table 23 (No. 2). Hale (1958), using reconnaissance X-ray methods, pointed out the similarity between muscovite (and chrome mica) patterns and those for samples labelled batchelorite, agalmatolite and fuchsite from Mt. Lyell.

Threadgold and Edwards were unable to find Petterd's original material, from which batchelorite was named, and hence were unable to definitely disprove the validity of the term. However, it is extremely unlikely that batchelorite is a distinctive mineral.

X-ray diffraction patterns were made of several other specimens of "batchelorite-like" material (32958a, b, 32979, 32768) and they proved to be sericites. These and the other sericites already mentioned, are similar to 2M$_1$ muscovites and are almost certainly not 1M$_1$ muscovites, by comparison with Yoder and Eugster's results and those of Radoslovich (1960). Threadgold (1959) suggested that a specimen from the Lyell Comstock mine was a 2M$_2$ (6M) polymorph but Radoslovich (1960) showed by powder photograph comparisons that it was closer to a 2:1 mixture of 1M and 2M$_1$ muscovite.
One of the sericites (Table 23, No. 3) contains almost 2% MgO, probably by substitution of Mg for Al.

Some of the analysed sericites are low in alkalies and high in H$_2$O +, probably due to the replacement of K$^+$ by H$_3$O$^+$ ions (see Brown and Norrish, 1952). Taboadela and Ferrandis (1957), following work by Mackenzie, defined hydromuscovite as shown in Table 24 (based on a structural unit of 20 oxygen and 4 hydroxyl ions). As can be seen from the same table the Lyell examples for which suitable analyses are available show no hydration or a small amount only, so that they are muscovite or fall between muscovite and hydromuscovite.

The greenish colour is probably largely due to chromium. Though 32965 and 32889 proved to have no chromium (see Table 23), Threadgold (1959) reported 0.25% and Bothwell and Moss reported 0.3% Cr$_2$O$_3$. The following results were obtained from specimens analysed by X-ray spectrographic methods:

1. Agalmatolite, Mr. Lyell (University of Tasmania) 0.9% Cr$_2$O$_3$
2. Fuchsite " " " " " 0.9% "
3. Sericite with a little pyrophyllite (32958a) 0.9% "
4. Sericite, Crown Lyell Mine No. 6 level 0.3% "

Samples 1, 2 and 3 are distinctly greenish (10 G Y 4/4, 10 G Y 4/4 and 5 G Y 7/4 respectively), sample 4 is yellowish (5 Y 6/4), and
<table>
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<tr>
<th>Specimen</th>
<th>$K_2O + Na_2O$ (%)</th>
<th>$K^+ + Na^+$ Ionic Ratio</th>
<th>$H_2O^+$ Ionic Ratio</th>
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<td>2.</td>
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<td>6.</td>
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<td>7.</td>
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<td>8.</td>
<td>6.42</td>
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<td>Muscovite</td>
<td>&gt; 9</td>
<td>1.6-4+</td>
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</tbody>
</table>

Specimen numbers from Table 23. Hydromuscovite and muscovite as defined by Taboadela and Ferrandis (1957).
the analyses confirm that chromium is the cause of the colouration. Whitmore et al. (1946) found that chromium micas are fairly common in areas of hydrothermal alteration and that the chromium is generally believed to be of magmatic origin.

Sericite samples concentrated from quartz sericite schists were from the West Lyell area (31636, 31634 and 31693)/analysed by X-ray diffraction and proved to be identical to the 2M1 sericites just described.

Pyrophyllite is the major component of specimens 32958, 32952, 32960 (see Table 23) and occurs in small amounts in others. It is confined to the pods and lenses of "batchelorite" and has only been found within highly mineralized zones.

Other Schist Types

Chlorite and sericite schists form 80-90% of the schist types of the mine areas and the following cases are to some extent rarities.

Relatively unaltered sandstones and mudstones, consisting of equigranular quartz in a chloritic base, occur in thin, impersistent lenses in several places (30926, 32682, 32686, 32757). Even those from the mineralized zone of the West Lyell Open Cut are clearly siltstones, both in the hand specimen and in thin section,
though puckered by the cleavage and somewhat sericitised.

Conglomeratic (agglomeratic ?) phases of the schists are visible in the field and particularly in diamond drill core (e.g. 32884, 32670, from North Lyell, 32683 from the Royal Tharsis area, 32786, 30972, 32904 and 32687 from West Lyell). 32786 consists of a chaotic arrangement of fragments from microscopic size up to 4 cm. diameter, of vein quartz, quartzite or sandstone, laminated siltstone and altered volcanic (?) rocks in a streaky chlorite-sericite-quartz matrix.

Even where sulphide mineralization is intense, the original fragmental texture is in many cases clearly revealed in polished and thin section (e.g. 32904).

Marginal Schists

Most of the marginal area of hydrothermal alteration is of low relief and deeply weathered but fresh material has been obtained from occasional good surface exposures and from the service tunnels (North Lyell and West Lyell Tunnels) that take ore from passes below the workings to the treatment works near the Queen River (see Fig. 34). Textural features show up on weathered faces and banded rocks have been followed across the area west of the Mt. Lyell Reserve workings.
The face of the West Lyell tunnel was examined regularly (with Mr. M.L. Wade, then Chief Geologist of the Mt. Lyell Co.) during its construction and a series of specimens collected (32722 to 32966 and 31696). As the tunnel runs almost normal to the strike of the country rock, rapid variation in schist type was encountered. For the first 1,100 ft. or 3,400 m., many of the rocks appear to be altered albite keratophyres. Thus 23966, 50 ft. (15 m.) from the portal, consists of "augen" of calcite and feldspar wrapped around by a microcrystalline sericitic base containing stringers of pale green chlorite. The augen are mainly calcite mosaics with some chlorite but remnants of the original albite crystals from which they developed are still visible. The crystals are up to 1 x 0.5 mm. but are generally near 0.3 mm. long, and some are broken and cracked by the cleavage. The albite is of low temperature form and apparently identical to that in the volcanics.

Calcite and chlorite replacement of albite and the development of a quartz-chlorite-calcite base is also well displayed in 32722 at 200 ft. (60 m.) Sericitisation is more extreme in 31696, at 230 ft. (70 m.), and fewer albite relics are visible (see Table 25, No. 1). Undulose extinction is developed in several of the feldspars. In 32937 and 32937a (see Table 25, No. 2) much of the rock is a crudely developed quartz-sericite aggregate (plus partly altered feldspar,
### Table 25: Marginal Schists

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**Ig. loss**

<p>| | | |</p>
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<tr>
<td>Total</td>
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<td>99.78</td>
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</table>

1. Altered keratophyre (?), 230 ft. (70 m,) from the West Lyell Tunnel portal (31696).
2. Altered keratophyre (?), 540 ft. (165m) from the West Lyell Tunnel portal (32937).
3. Altered keratophyre (?), 1,712 ft. (522 m,) from the West Lyell Tunnel portal (32934).

**Analysts:** Japan Analytical Chemistry Research Institute.
chlorite, calcite, etc.) and this appears to have grown out of a microcrystalline groundmass or matrix.

Up to about 1,100 ft. (330 m.) relics of albite are visible in some specimens (e.g. 32741, 32740) but others (e.g. 32948, 32935) are merely quartz-sericite-chlorite schists, generally fine-grained but in places containing unaltered embayed quartz phenocrysts.

Closer to the West Lyell Open Cut the schists become more recrystallised and coarser in grain, though very variable, and tend towards the chlorite or sericite schist already described. Thus 32934 (at 1,712 ft. (522 m.) - see Table 25 analysis 3) differs from typical quartz sericite schist only in having a relatively high carbonate content (part siderite ?). Typical quartz-chlorite schists are relatively rare until after about 2,000 ft. (600 m.).

Summary of Schist Types and their Origin

Most schists in the mineralized area are chloritic or sericitic and the alternation of these types in bands indicates that original differences in composition or physical properties must have exerted a very strong influence during alteration.

Study of the marginal zone of alteration shows that at least some of the schists have developed by disruption and replacement of quartz-feldspar and feldspar porphyries (keratophyres) by
sericite, chlorite and carbonates, and the mineralogy is typical of propylitic alteration. In the more intensely altered schists there has been recrystallisation and mobilisation of quartz to form relatively coarse quartz aggregates that form the bases of most schists. However, primary volcanic quartz phenocrysts have survived the most intense alteration, or at least have their outlines preserved despite recrystallisation of the quartz. The only minerals of clearly secondary origin are quartz, sericite (part chromium), pyrophyllite and chlorite; alunite or clay minerals have not been found.

It has proved impossible to trace any one particular rock unit across the boundary of the alteration aureole and as a result the original nature of most of the schists can only be surmised. The banded quartz sericite schists may be tuffaceous, with agglomeratic phases, or may represent flow banded and nodular rhyolitic lavas. Some of the chloritic types are clearly fragmental and may have been tuffs or greywackes but others may have been lavas. Agglomerates (?) and mudstones have also been identified.

Chemical Changes

If the quartz sericite schists were originally quartz keratophyres then they now have a lower total alkali content (the
local acid keratophyres have $> 6\%$ alkalies). The generally low soda content of these schists may be due to loss of soda or an original low soda content (in which case loss of potash must have taken place). The schists have gained water to the extent of about 0.5%. If these schists were originally banded tuffs or tuffaceous mudstones such as occur in Conglomerate Creek then analyses of Table 22 and Table 10 suggest little change. Derivation from keratophyre seems likely because of the presence of volcanic quartz phenocrysts, possible feldspar laths, and banded and agglomeratic textures that have been found in quartz keratophyres.

For the quartz chlorite schists, spilites are probably the nearest primary rocks in composition, though these generally have lower water and FeO and higher CaO and alkalies.

For the marginal schists in the first few hundred metres of the West Lyell Tunnel it seems fairly clear from chemical and microscopic evidence that these are carbonated, slightly hydrated and sericitised keratophyres. The marginal schist at 1,721 ft. (522 m.) in the West Lyell tunnel (32934) which contains quartz phenocrysts, still retains soda and a low water content. It is typically marginal with its high carbonate content and may well represent an intermediate phase of the alteration to normal quartz sericite schist.
In Table 26 the average sericite schist is compared to the average Tasmanian quartz keratophyre, chlorite schist is compared to spilite and finally, a theoretical bulk composition of the alteration aureole is compared to a possible pre-alteration composition. These bulk compositions appear to be the most reasonable and variations on them (e.g. 5 sericite, 1 chlorite) do not affect the conclusions below.

In examining the three pairs of average analyses it is clear that in each case there is loss of Ca, Na and some Al, and gain of H₂O and Fe²⁺. The uniformity of the comparisons rather suggests that these changes may be close approximations of the actual gains and losses but it is clear that no quantitative estimates can be made. For the marginal schists the available information is scarcely sufficient to make even generalisations but it is likely that there has been loss of lime and soda and slight gain of water and possibly potash.

Loss of soda and lime and gain in water are characteristics of the sericitic zones of "porphyry copper" hydrothermal alteration, although this usually also involves gain of potash (e.g. Schwartz, 1959; Creasey, 1959; Anderson et al., 1955 on the San Manuel, Bagdad, Ajo, Ely, Bingham, Castle Dome, Santa Rita and Chuquicamata deposits). Sericitic alteration also generally involves
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<td>1.37</td>
<td>4.86</td>
<td>4.69</td>
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<td>3.47</td>
<td>6.11</td>
<td>12.61</td>
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<td>3.88</td>
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<td>&lt;1.00</td>
<td>0.18</td>
<td>0.49</td>
<td>0.62</td>
<td>1.43</td>
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</table>

1. Average Tasmanian quartz keratophyre (11 analyses of Table 9).
2. Average quartz sericite schist.
3. Average Tasmanian spilite (17 analyses of Table 5).
4. Average quartz chlorite schist.
5. Composition: 3 x quartz keratophyre, 1 x keratophyre,
   1 x spilite.
6. Composition: 3 x sericite schist, 1 x chlorite schist,
   1 x marginal schist.
gain of silica and loss of ferrous oxide, alumina and magnesia but the gain and losses are relatively small and there are exceptions. The presence of distinct argillic zones produces local variations in the chemical changes described for sericitic zones.

Chloritic alteration is less prevalent in the porphyry coppers, particularly in the more altered zones, due in part to the relative lack of ferromagnesian minerals in the unaltered magmatic rocks.

From the limited information available it appears that characteristic features of marginal prophylitic alteration of porphyry coppers are gains in magnesium and calcium, due to increased chlorite and carbonates.

In very general terms, then, the Lyell deposits show many of the characteristics of hydrothermal alteration associated with the disseminated sulphide ore deposits of "porphyry copper" type.

Remarkably little is known about the fairly common alteration of volcanic rocks associated with sulphide deposits, and particularly disseminated ores. Kurshakova (1958) has shown that in the Buribay (Southern Urals) chalcopyrite-pyrite deposit the host spilites are converted to chloritic schists by addition of Mg, Fe²⁺, Al and H₂O with loss of Na, Ca and Si. Reber (in Schwartz,
1959) found an increase of MgO from 1.48 to 18.16% in the chloritisation of a quartz porphyry in Arizona.

Johnston (1940) has described the conversion of a uralitic diabase to a rock composed mainly of quartz, calcite and sericite, with epidote present locally.

Adjacent to the tin-silver veins of Oruro, Bolivia, (Chace, 1948), a quartz latite porphyry (with "volcanic affinities") is converted to sericite and pyrite and the original texture destroyed, while the marginal alteration is characterised by slight sericitisation, chloritisation of biotite and preservation of texture.

Albers et al. (1961, p. 49-52) have described silicification and sericitisation of rhyolitic flows, and chloritisation of intermediate and mafic volcanics in East Shasta Co., California. Some of this alteration is clearly related to mineralization (e.g. at the Rising Star Mine) but some is ascribed to fluids derived from trondhjemitic differentiation (see also Gilluly, 1935) or to geosynclinal connate waters; albitisation is supposed to be coeval with sericite and chlorite growth.

Baird (1960) mentioned chloritisation, sericitisation and silicification of basic and intermediate Ordovician volcanics associated with massive sulphide deposits in Newfoundland, and McAllister (1960) gave more details of chloritic and sericitic rocks
associated with similar deposits in New Brunswick. He suggested some of the chloritic schists were pyroclastics and some were a facies of the local sedimentary "iron formation". The source of the Fe and Mg required for the chlorite is unknown, as most of the associated rocks are acidic. The sericitic schists are derived from acid volcanics.

Analyses of sericitised and unaltered "trachyte" adjacent to cinnabar veins from the Rochester District, Nevada (Knopf, 1924) indicate addition of water, silica and alumina and loss of potash, soda and iron oxides. Adjacent to silver veins silicification is dominant, with an outer sericite zone.

Williams (1934, p. 601) has quoted analyses of altered and fresh quartz porphyry (extrusive ?) which indicate addition of carbonate and potash and significant loss of silica, iron oxides and soda.

Tertiary rhyolites and andesitic tuffs at Telluride, Colorado (Hurst, 1922) show alteration adjacent to sulphide fissure veins. Initial development of chlorite (from ferromagnesian minerals) and quartz, was followed by development of sericite and quartz, all earlier than pyrite. An outer zone of pyropylitic alteration is characterised by chlorite, calcite and pyrite, with sericite rare. Similarly at the Iron King Mine, Arizona (Creasey,
1952) Precambrian rhyolites and andesites cut by sulphide veins are altered to rocks composed of sericite, quartz, ankerite and pyrite and retaining only relics of primary textures.

For volcanic rocks, then, the alteration is essentially confined to sericitisation and chloritisation, with strong control by the original composition shown in some areas but not in others. There is insufficient information available, and probably too much variation, to generalise on the chemical changes involved except to note that loss of soda and gain of water are common features.

Conditions of Alteration

In recent years attempts have been made to define facies or types of hydrothermal alteration in porphyry coppers. In 1959 Greasey distinguished three types: the propylitic (with chlorite, muscovite, epidote and calcite); the argillic (with kaolinite, montmorillonite and muscovite); and the K-silicate (with muscovite, K-feldspar and biotite). He found a close approach to equilibrium using AKF and ACF diagrams, excluding water and silica as "mobile" components (in Korzhinskii's sense).

Burnham (1962) identified two facies: the argillic (with propylitic minerals, plus sericite, chlorite, kaolinite and
montmorillonite and K-feldspar) and the phyllic (with muscovite or biotite, K-feldspar, pyrophyllite, andalusite, topaz, and chlorite). Note that he used sericite to indicate hydromuscovite, illite etc. and not fine-grained muscovite. He modified the traditional AKF diagram for his two facies so as to include water. His inclusion of propylitic alteration in the argillic facies is sound, as the stability fields of propylitic minerals are similar to the argillic, but it is useful to separate the chemically different propylitic type. The addition of water to the AKF diagram has been criticised by Hemley and Hostetler (1964) who pointed out that water is likely to be a "mobile" component as well as silica.

Hemley and Jones (1964) tentatively subdivided hydrothermal alteration into incipient (propylitic), intermediate (sericite, chlorite etc.) and advanced (kaolinite, alunite, jarosite, etc.). Hemley (1959) and Hemley and Jones (1964) have suggested that the main controls in hydrothermal alteration are the temperature, pressure, metal cation/hydrogen ion ratio, and the composition of the wall rock. They have determined the reactions in the Na$_2$O- and K$_2$O-Al$_2$O$_3$-SiO$_2$-H$_2$O systems and outlined in part the stability fields of the alteration minerals in cation chloride solutions of varying concentration and at various temperatures (Fig. 40). Increased pressure moves the reaction curves of Fig. 40 to the left.
Fig. 40: Reaction curves for the K$_2$O-Al$_2$O$_3$-SiO$_2$-H$_2$O and Na$_2$O-Al$_2$O$_3$-SiO$_2$-H$_2$O systems, after Hemley (1959) and Hemley and Jones (1964). Experiments made in the presence of quartz and at 1000 atmospheres total pressure.
From Hemley's curves it is clear that pyrophyllite (or boehmite) is a key mineral in indicating the temperature of alteration under certain conditions. According to Roy and Osborn (1954) pyrophyllite may be synthesised readily between 420°C and 575°C at varying water pressures and if insufficient water is present to form higher hydrates, it may form between 275°C and 405°C e.g. Gruner (1944) produced pyrophyllite from albite at 300°C. Hemley's experiments indicate that the kaolinite-pyrophyllite boundary is near 350°C. Lack of boehmite, which is stable over a similar range to pyrophyllite, is probably due to abundance of SiO₂. Andalusite has been synthesised between 450°C and 600°C and water vapour pressures of 700 and 2,000 atmos. (Roy, 1954) and the lack of andalusite at Lyell suggests alteration temperatures never reached much above, say 500°C.

The nature of the mica in the aureole may be a useful guide because Burnham has indicated that the muscovite of his phyllic facies is different from the sericite (in his sense) of the argillic facies. Yoder and Eugster (1955) were able to transform 1M muscovites to 2M structures by heating, indicating a possible correlation of polymorph and temperature of formation. In 1959, Yoder suggested the mica growth sequence in conditions of rising temperature is 1Md→1M→2M. 1M micas with interlayered clay
minerals (particularly montmorillonite) are characteristic of the argillic facies in porphyry copper deposits but the muscovites without clay minerals are 2M types. Although Radoslovich suggested there may be both 2M and 1M types at Lyell, the writer's own observations suggest a dominance of 2M₄ polymorphs.

Burnham has also pointed out that the degree of hydration of the micas appears to be inversely related to the temperature, illite and hydromuscovite being characteristic of low temperature environments and muscovite of higher temperature. The patchy and small amount of $\text{H}_3^0^+$ - $\text{K}^+$ substitution in the Lyell micas indicates a relatively high temperature of formation.

Considerable quantities of paragonite might be expected in the Lyell alteration zone because relics of albite have been observed in the schists. In general, sodium appears to have difficulty in fixing in mica when freed from feldspar e.g. according to Morey and Chen (1955) albite held in water at 350°C and 10,000 p.s.i. yielded a film of 1M muscovite, "some paragonite" and much analcite, and at lower temperatures no paragonite was detected. The potash in the muscovite was derived from the 0.2% in the albite. Part of the explanation may be in the relatively small stability field of paragonite (Yoder, 1959; Hemley and Jones, 1964; Fig. 40).

Chlorite is a stable mineral in all alteration types. Mg-
chlorite may persist up to 720°C at 2,000 bars (Yoder, 1952) but Fe-chlorites break down above a maximum of about 600°C (Turnock, 1960).

The Lyell Schists do not contain clay minerals and must therefore fit into Creasey's K-silicate type or Burnham's phyllic facies. They are dominantly muscovite of 2M₁ type with a low degree of hydration and are associated with small amounts of pyrophyllite. They probably represent a common alteration assemblage that appears to be a "low grade" or marginal type of the phyllic facies or K-silicate type of alteration and which might well be called the "muscovite-chlorite" type. This distinguishes it from altered rocks in which biotite is stable.

In porphyry copper and other types of alteration there seems to be good reason for believing that chlorite alteration is replaced by sericite alteration (see Creasey and Burnham) and Burnham suggests that the alteration proceeds towards higher degrees of hydration. There is no evidence for replacement of chlorite by sericite at Mt. Lyell and in fact the two main schist types appear to have formed together and remained stable. Possibly at Lyell any trends of the sort realised elsewhere were swamped by primary differences in composition.

It is possible to determine very approximately the temperature
of alteration at Lyell. The overburden consisted of Ordovician and Siluro-Devonian sediments estimated to be at least 8,000 ft. (2,400 m.), producing a pressure of about 700 bars. As the region was tectonically active at this time, stresses due to earth movements may have overwhelmed the load pressure. The crudeness of the overburden estimation is not so vital as might appear, because the effect of pressure changes on the free energy of reaction is relatively minor.

At 700 bars, muscovite is unstable above about 620°C (Yoder and Eugster, 1955), Fe-chlorite above about 560°C and Mg-chlorite above about 675°C (so the Lyell chlorite's upper stability limit might be, say, 600°C), pyrophyllite above about 575°C and paragonite above 600°C.

Thus the maximum possible figure for the Lyell alteration is about 575°C and the lack of andalusite indicates a maximum temperature nearer 500°C.

The lack of clay minerals suggests a minimum of about 300°C for the sericite-chlorite alteration though it is possible to produce this assemblage at lower temperature (down to about 200°C) provided the $K^+/H^+$ ratio is restricted within narrow limits (Fig. 40). Conditions within the sericite-chlorite aureole were probably essentially isothermal and isobaric, and changes in the metamorphic
assemblages were probably due more to changes in the $K^+ / H^+$ ratio and variations in composition of the wall rocks.

Because the indicated temperature of alteration is relatively high and pyrophyllite is present, and because it is likely that there was no gain of potash, it is probable that the $K^+ / H^+$ ratio was fairly low and the solutions fairly acid (see Fig. 40). The general absence of feldspar in the schists also suggests that alteration continued long enough for equilibrium conditions to be closely approached.

**Timing of the Hydrothermal Alteration**

Pebbles in the Hematite Conglomerate show that some silicification of quartz keratophyres (to produce a quartzose groundmass mosaic) had occurred prior to the Ordovician. Whether this alteration or recrystallisation was syngenetic or related to the Jukesian Orogeny is unknown, but at all events it preceded the Devonian cleavage.

Many schists show stages in the break up of the quartz mosaic by movement on cleavage; this movement phase produced some degree of preferred orientation in newly developed sericite and chlorite and involved some quartz recrystallisation, with coarsening of grain size and dimensional orientation. Crystallisation
persisted until after deformation had ceased and some part of the preferred orientation of sericite may be due to mimetic crystallisation. Crystallisation continuing after cleavage development (or a later crystallisation) produced sericite and chlorite with random orientation, and quartz with sharp extinction.

Sulphide deposition caused localised recrystallisation of the schists and where mineralization is intense (say, > 20% sulphides) the matrix immediately around the sulphides is more coarsely grained than normal (e.g. 31183 and Plate 58, No. 1). However, in low-grade disseminated ores there is generally no recrystallisation at all.

Where recrystallisation does occur around sulphide crystals, the quartz, sericite and chlorite involved may be quite randomly oriented, producing a coarse felt, or it may be aligned in a fibrous growth with tendencies to parallel the cleavage direction and/or to form at right angles to the sulphide crystal faces. The best examples (e.g. 31639, 32687) form where these two factors act together i.e. where the crystal faces are at right angles to the cleavage, as in Plate 58, No. 2.

The 'pressure shadows' appear to be post-deformation but somewhat similar features occur in the Rosebery schists and in these
Plate 58 No. 1:– Recrystallised schist around pyrite crystals (31183). X40.

Plate 58 No. 2:– "Pressure shadows" of quartz around pyrite in sericite schist. X70.
there is some indication that they form on pre-existing pyrite crystals during deformation. These features are discussed later.

Nomenclature of the Lyell Schists

The schists do not readily fit into any particular category of metamorphic rock. Because they were formed in a hydrothermal environment they have undergone considerable recrystallisation unrelated to deformation. They are not like crystalline schists which show microfabrics, crinkled cleavage and extensive transposition of earlier surfaces, but they have some similarities to phyllites. They have been in part granulated during deformation and possess features due to dislocation metamorphism but without reaching the "grade" of a phyllonite (as defined by Knopf, 1931).

They are thus part phyllite, part phyllonite, and might best be described as schistose or cleaved volcanics. However the term Lyell Schists is well established and is convenient in the present study.
Relationship between Schist Types and Mineralization

As most drill cores are assayed for copper in five-foot lengths, it is relatively easy to determine the distribution of copper with respect to the dominant schist types, and studies to this end were made on a large number of drill holes from the West Lyell Open Cut.

It is clear that the copper distribution has no relation to schist type when considering all copper values, or only those over 0.5% or only those over 1%. This conclusion fits the cruder observations made while mapping the West Lyell Open Cut, from which it appears that schist type cannot be used as a guide for copper distribution. Though the Honeypot orebody is (or was) largely in sericitic schist, the Prince Lyell orebody is, in considerable part (particularly at the south end), in chloritic schist.

For pyrite, there is very much less information available and studies were made only of West Lyell drill holes numbered 89, 95, 96, 97, 100, 101, 112, 124, 125, 129, 133 and 136, for which FeS₂ assays have been made on core lengths of 10 ft. or more. Several high values (\( > 18\% \) FeS₂) occur in chlorite schists though the majority (85%) are in sericitic schist, yet the lower values of pyrite do not seem to prefer either schist. Mapping the Open Cut
indicates that high pyrite concentrations occur mainly in sericite schists but this may reflect merely the abundance of sericitic types. Without a very detailed study and more FeS$_2$ determinations, it is perhaps unwise to draw any conclusions.

**Structure of the Mt. Lyell Area**

The general picture has been outlined by Gregory (1905), Hills (1927), Conolly (1947), Bradley (1956), and the writer (1957) and modifications have been presented in several subsequent papers (Wade and Solomon, 1958; Solomon, 1962; Solomon and Elms, in press, Appendix A, paper 2).

**Jukesian Structures**

No structural discordance between Ordovician and Cambrian rocks is visible at Lyell but slight discordance is revealed near Lynchford, 2 miles south of Queenstown (Fig. 28).

**Ordovician Structures**

Between Comstock and the Blow, the topmost member of the Owen formation, the Pioneer Beds, overlies the older members with angular unconformity. This feature was mapped by Conolly and
his staff in 1940 and was mentioned by Bradley (1954, p. 209).

Good exposures of the unconformity, known as the Haulage Unconformity, occur on Linda and Pioneer Spurs and below the Blow mine at the end of the old ore haulage to Queenstown (Plate 59, No. 1). The underlying beds, mostly the Chocolate Sandstone, are nearly vertical and in places are tightly folded. "Unfolding" the unconformity (allowing for plunge where necessary) shows that in pre-Pioneer time there existed a N-S strip, some 300 m. wide, of tightly folded beds and that the folds and steep dips passed rapidly to horizontal on either flank (Fig. 41). The Haulage folds are best exposed on the Whaleback, at approximately 1600S/50E, where they strike north with more or less horizontal plunge, are asymmetrical with axial surfaces dipping west, and have wavelengths of 2 to 4 m. Despite the "tightness" of the folds, cleavage is developed only very locally and changes in thickness of individual layers is very small except in some thin mudstone layers (Plate 59, No. 2).

On Pioneer Spur the folds are similar except for steeper axial surfaces. In the Blow open-cut the Chocolate Sandstones are almost isoclinally folded and tilted to the east and similar folds may be seen at the waterfall in Waterfall Gully.

All these folds are quite different from the obviously Tabberabberan folds, in which cleavage developed readily and quartz
Plate 59 No. 1:– Haulage Unconformity, below east end of old Mt. Lyell Co. haulage.

Plate 59 No. 2:– Folds in Owen sediments on Whaleback Ridge, looking north.
Fig. 41: Structural picture after unfolding the Haulage Unconformity.
veining and minor faulting are common.

West of the folded strip, the Pioneer beds overlie Middle Owen sediments and almost certainly overlapped onto the Cambrian rocks (e.g. at the west end of Linda Spur). The cross-sections derived from the detailed mapping indicate that there was little or nor cover over the folded strip at the time of folding and it is clear that the crumpling involved considerable shortening.

Bradley (1954, p. 209) suggested that the folding was the result of slumping and this idea is tentatively accepted by the writer. Continued rise of the west wall of the Owen basin (the Dundas Ridge) is a feature of the structural history of the Mt. Lyell area and gravitational collapse into the basin is a likely result. It is suggested that a surge of uplift in the Upper Owen caused a subaqueous avalanche to sweep into the basin, partly/sliding over the Owen beds and partly by pushing them in front of the avalanche nose. This suggestion is outlined in sections (a) and (b) of Fig. 42. The unusual structure of the Waterfall area is believed to be the result of Haulage folding.

The resulting pattern of Cambrian material locally overlying the crumpled Owen sediments is well exposed on the Whaleback and has led to great complexity in the schist-conglomerate contact in this and other areas.
Fig. 42: Sketches to illustrate the development of the Haulage folds and the Tharsis Schist Zone:

(a) The Owen basin prior to Haulage movements.

(b) The same area during deposition of the Pioneer Beds.

(c) The same area after Tabberabberan faulting and upturning.

(d) Sketches to show the possible development of the Tharsis Schist Zone as a result of two phases of N-S folding.
(a) Hematitic phase

UPPER OWEN

MIDDLE OWEN

LOWER OWEN

(b) Pioneer beds

(d) 1st phase of folding

OWEN CONGLOMERATE

2nd phase of folding

Generalised combination of two phases
The incursion of Cambrian material, both schist and chert, into the Owen basin may be the cause of the schist zone (the Tharsis Schist Zone) that appears to lie within the Owen (Fig. 45). This tongue of schist was regarded by Conolly (1947) as a porphyry intrusion but Wade and Solomon (1958) suggested that it represented a conglomerate phase of the Owen, rich in volcanic debris. This suggestion is now extended to involve slump material, including large blocks of chert. The final form of this schist zone is interpreted as due to further upturning and reverse faulting in the Devonian (see Fig. 42 (c)). As an alternative explanation of the Tharsis Schist Zone, Professor Carey (pers. comm.) has suggested that it may be the result of an early tectonic folding with marked east-facing asymmetry followed by later (Devonian) uplift. This sequence is outlined in Fig. 42 (d), following constructions developed by Carey (1962).

Further evidence of local Ordovician instability is provided by a swarm of sandstone dykes in the Lyell Schists and the Middle Owen near the Owen contact (Bradley, 1954, p. 209). They are best exposed just west of Razorback Ridge, where a dozen or so sub-parallel dykes penetrate schist (Plate 60, No. 1); poorer examples occur near the Blow Open Cut. They vary from a few cm. to 20 cm. wide and are composed of slightly hematitic quartz sandstone (32950, 32702) similar to the Middle Owen sandstones.
Plate 60 No. 1: -- Sandstone dyke penetrating schist, north end of Razorback Ridge.

Plate 60 No. 2: -- Cleavage augen in Upper Owen sandstone, North Lyell area.
They are considerably crumpled and distorted by the cleavage and near the Razorback shaft are smeared against the Owen-schist contact. The three examples in the Middle Owen Conglomerate of the Razorback Ridge are perpendicular to bedding, trend roughly E-W and are only locally distorted. The sandstone dykes in the schist must have been filled from above (as this is the only source), and possibly the cracks formed during earth movements associated with Haulage folding.

The Haulage folding, the sandstone dykes and the coarse Middle Owen breccias derived from Cambrian outcrops all point to localised but repeated disturbances along the west wall of the Owen basin near Lyell. The Pioneer Beds thin markedly over this wall, and may have never covered it, but it is likely that the succeeding Gordon Limestone was deposited over the entire area. The marine sediments of Silurian-Devonian age probably also covered the entire area, burying it under at least 7,000 ft. of overburden. Siluro-Devonian sedimentation was relatively quiescent on the West Coast though the presence of coarse sandstone and grit at the base of the Silurian (the Crotty Quartzite) has been mentioned as evidence of renewed uplift of the source areas (see Solomon, 1962, p. 322).
Tabberabberan Structures

Early N-S Structures (TF₁)

The principal structures of the area are of Tabberabberan age and N and NW-WNW trend (Fig. 43). The dominating feature is the N-trending West Coast Range Anticlinorium, which probably continues north as the Dundas Anticlinorium. It is flanked to east and west by major synclinoria, the King and Dubbil-Barril structures respectively (Figs. 31, 32).

Movement almost certainly occurred at this stage along the west wall of the Owen basin. From Mt. Owen to Mt. Geikie and beyond, this zone is marked by a sharp upturn of the lower Owen with overturning and reverse faulting in places. For example, west of Mt. Sedgwick, the basal Owen sediments are overturned and lie on a clean-cut, west dipping fault surface (Fig. 38). Along strike to the north, the fault disappears and the structure becomes a syncline (e.g. north of the Lake Margaret dam). On the west flank of Mt. Owen, a similar syncline with flattish plunge is visible from the mine area. Many of these N-S structures were observed by Gregory (1905, p. 77-84) who referred to this zone of disturbance as the Great Lyell Fault. Later workers used the term Lyell Shear (Carey, 1953, p. 1124; Bradley, 1956, p. 80) and Campana et al. (1958) the term West Owen Rift Fault. Hall and Cottle (1959)
Fig. 43: Principal Devonian structures of the Mt. Lyell area.
AREA OF MINERALISATION AT MT LYELL

FAULT

ANTICLINE

SYNCLINE

Z

WEST-COAST RANGE ANTICLINALION

GREAT LYELL FAULT

NORTH LYELL FAULT

QUEENSTOWN

DUNDAS ANTICLINALION

ELD INDIAN ANTICLINALION

SOUTHERN ANTICLINALIUM

DARWIN ANTICLINALION
referred to Gregory's term in discussing Campana's paper and the writer has since used the term Great Lyell Fault Zone (1959, 1962) to include several parallel and/or en echelon faults and the associated folding. At Lyell, Conolly (1947) referred to the upturned structure as the Razorback fold.

It is possible that part of the faulting and upturning of this zone took place during Haulage movement but it is clear from the detailed mapping at Lyell that the Pioneer Beds attain fairly steep easterly dips and that at least a major part of this phase was Tabberabberan.

In the Lyell area steeply dipping north-striking sediments are visible near Comstock and on the Tharsis Ridge and in both areas many current bedding facings indicate that the top is to the east. The upturning is in all cases very sharp and the actual axis cannot be observed in the mine area. Even though the North Lyell Tunnel intersects the structure the precise form of the upturn is not clear. Detailed mapping of the tunnel (Fig. 44) shows steeply west-dipping sediments for 180 m. east from the Owenschist contact and these are succeeded eastward by gently east-dipping beds. Little is revealed in the "hinge" area, which may be interpreted as a synclinal axis (Conolly, 1947) or a fault. Support for faulting is seen in the fairly intensive quartz veining in the conglomerates and the orientation of quartz veins in the gently
Fig. 44: Geological map of the North Lyell Tunnel showing
(a) stereographic plot of the small fold axes
(b) suggested stress system to produce secondary shears.
dipping beds. Here the bedding planes have been filled by quartz veins with associated feather fractures, all of which indicate movement top side west (see details in Fig. 44). This is the reverse of the movement expected from a syncline but is similar to that on secondary shears that might result from reverse fault movement, west side up (see Fig. 44, inset (b)).

Minor folds and monoclinal kinks in sandstone beds of the steeply dipping zone have both east and west facing asymmetry and plunge both northerly and southerly. The variable plunge may be due to later folding on WNW, east plunging axes (see stereographic net, Fig. 44, inset (a)). The fault interpretation is shown in Fig. 45. On the structural sequence outlined in Fig. 42 (d) the gently dipping sediments east of the fault are likely to be overturned; unfortunately no facing criteria were observed.

It is likely that the schist-Owen contact was faulted at this stage but this faulting has been obscured in places by later post-cleavage faulting.

It is concluded that the Great Lyell Fault Zone in this area is a fault or a series of sub-parallel faults, of thrust or reverse type, that achieve the same effect as the sharp upturning seen elsewhere along the Zone. The displacements are of the order of 100 m, and are probably of dip-slip type. No evidence was found in the mine
Fig. 45: East-west cross-section (on grid line 400 S) of the North Lyell area.
area (or elsewhere) for the horizontal (dextral) movement mentioned by Carey (1953) and Bradley (1956, p. 80) but in the stress diagram deduced for the TF₂ phase (Fig. 33) transcurrent movement on N-S structures might be expected during TF₂ folding and faulting.

It must be emphasised that uplift along the Great Lyell Fault Zone did not cease before the TF₂ folding and mention of post-TF₂ movement has already been made. Though the division into TF₁ and TF₂ is generally applicable N-S uplift quite probably took place in TF₂ time. Faulting that is post-TF₂ cleavage is described later.

WNW-NW Structures (TF₂)

The TF₁ features described so far are displaced or folded by a number of later WNW-NW structures (TF₂), which are probably broadly synchronous.

The main TF₂ structures are:

(a) Faults with dip-slip and strike-slip movement, and the Linda Fault Zone.

(b) Folds.

(c) Cleavage related to the folds, with a steeply pitching lineation, Tl₁.

(d) Lineation in the cleavage with shallow pitch, Tl₂.
(e) Boudinage in the schists.

Kink bands and other late structures may belong to this phase.

Fig. 46 illustrates the relationships between the small scale structures.

(a) Faults

Dominating this phase are a number of faults that cut the Owen Conglomerate and also the Precambrian quartzites and schists of the Tyennan Geanticline or Block (Fig. 3). All these faults are steeply inclined and have vertical displacements up to 2,500 ft. (750 m.). The faults of the Queenstown area represent a local concentration of these structures, which are of minor significance to north and south. The Queenstown faults appear to form part of a major E-W disturbance known as the Linda Fault Zone (Wade and Solomon, 1958; Solomon, 1962) or the Linda Disturbance (Bradley, 1956, p. 74). This zone forms a graben-like strip extending from west of Queenstown to several miles east of Bubbs Hill but appearing only in the more competent rock types. The nature of the fault zone is revealed by a section from Mt. Sedgwick to Mt. Owen (Fig. 47) and also at Bubbs Hill, where Gordon Limestone and Crotty Quartzite (Silurian) are faulted against the Precambrian (see cross-section in Carey and Banks, 1954, p. 256).
Fig. 46: To show relationships between various minor structures of the later phases of the Tabberabberan Orogeny at Mt. Lyell.
Fig. 47: Geological section from Mt. Sedgwick to Mt. Owen to show the form of the Linda Fault Zone.
Particular features of the NW fault phase were the formation of the graben-like structures of the Comstock and Linda Valleys. Hills (1927) referred to the Linda structure as the Linda saucer and this describes the form of the base of the Gordon Limestone fairly exactly. Conolly (1947) and co-workers mapped most of the principal faults and referred to them as "fault-folds", because they appeared to be the sheared out steep limbs of asymmetrical folds; they are thus akin to thrust faults. Hills (1927) recognised that shortening was involved and called them "lateral thrusts".

The faults are in places accompanied by intensive quartz veining and silicification of the adjacent sandstones but the fault surfaces are in general clear-cut. Little shearing of the Owen sediments is evident along the faults except in the Great Lyell area, at the western extremity of the South Owen Fault, and details of this are given in a later section.

Several of the faults show a considerable horizontal sinistral displacement that is vastly greater than can be reasonably attributed to vertical movement. On these faults the Great Lyell Fault or upturn is offset by distances up to 5,000 ft. (1,500 m.) and this is the principal evidence for regarding these faults as $T_F^2$. The maximum displacement is observed on the Sedgwick Fault and
smaller displacements on the North Lyell and South Owen Faults.

It is not clear whether the fault movement involved oblique-slip or separate horizontal and vertical phases though some support for a late horizontal phase is to be found in the Lyell Schists, in which horizontal slickensides on cleavage faces ($T_{12}$) are later than a steeply plunging lineation ($T_{11}$) related to the development of the WNW folds. Similar slickensides occur in the North Lyell chert and on the South Owen fault.

A considerable amount of information is available on the North Lyell Fault, one of the major WNW structures with pronounced sinistral displacement. The fault face is quite sharp, though with some gouge in places (found mainly in mine workings). In the North Lyell mine area, the south side of the fault is cut by two or three parallel faults which have produced a depressed, synclinal or graben-like strip known as the North Lyell Corridor. Within this zone, some 300 m. wide at the surface, the west dipping Owen contact has been dropped several hundred metres, causing the outcrop of the contact to be displaced eastward and the schists apparently to penetrate east along the Corridor. The severely brecciated schists and chert of this area are the home of numerous irregular pipes and pods of bornite-chalcopyrite ore (Fig. 48). Rather similar eastward penetrations of Cambrian material due to
Fig. 48: Geological cross-section of the North Lyell Fault.
graben-like structures occur at Gormanston and Comstock.

The fault surface in the North Lyell area has been proved by mine openings and diamond drilling to dip steeply south (Fig. 48).

On Mt. Owen the faults are vertical or dip south and are reverse-type. The North Lyell Fault has the opposite throw (south-side-down) to the Owen faults but, at least at its west end, it also dips south. Similarly the Comstock fault is known to dip north yet has north-side-down displacement. Thus the Mt. Lyell ridge appears to stand as a typical tensional horst structure, flanked by normal faults. The Owen faults and crumpling in the Linda saucer all point to shortening and it is difficult to reconcile the Mt. Lyell structure with this picture.

It is possible to explain away the North Lyell dip as the result of horizontal displacement. The main movement of this stage involved rotation about a vertical axis and this is likely to induce rotation about a horizontal axis. This rotation could twist the fault from north-dipping to south-dipping. However, this explanation does not apply to the Comstock fault.

Another explanation involves the assumption of an earlier period of movement on these faults. Possibly during $T_F1$ time tensional faults developed at right angles to the major N-S fold axes,
forming a series of horst and graben structures in the Mt. Lyell area. Later movements, in TF2 time, took place along these earlier structures.

The sharp bend in the N-S upturn between Comstock and Cape Horn is attributed to sinistral movement on the North Lyell Fault and not to the "skew warping" mentioned by Bradley (1957, p. 84 and 85).

The pronounced displacements seen on these TF2 faults in the Owen Conglomerate (and in Precambrian quartzite assemblages - Fig. 3) are not seen in the Mt. Read Volcanics or in the shaly Siluro-Devonian sequences. This is probably because the movements are dispersed along numerous cleavage surfaces in the less competent rocks.

Some late faults apparently related to the TF2 phase swing, or deflect, from WNW to NW trend e.g. the reverse fault separating schist and Owen Conglomerate in the Blow open-cut. From the suggested stress pattern for the TF2 structures (Fig. 33) it is clear that any N-S faults might be expected to show considerable transcurrent movements and faults approaching this trend should show a transcurrent component.

(b) Folds

In the Linda Saucer the Owen beds are severely crumpled into
slightly asymmetrical folds that plunge east near the mine and west near the King River. The majority have their north limbs faulted out and thus are similar in form to the Mt. Owen fault-folds. Their form is revealed in a N-S section from Mt. Lyell to Gormanston (Fig. 49). Most of these folds are of essentially concentric style (e.g. at 5900 S/3000 E on the Linda-Gormanston road, Fig. 35) but near North Lyell, axial surface cleavage is quite strongly developed in sandstones and fine conglomerates and the folds tend to a "similar" style. For the concentric-type folds at least, considerable shortening at right angles to the axial surface must have occurred.

(c) Cleavage Related to TE2 Folding, and Lineation T1

The axial surface cleavage extends into the Lyell Schists and swings, with variations, to about NW as it passes west of the mineralized ground (Fig. 35).

The Lyell Schists show a prominent, crudely penetrative lineation due to dimensional orientation of quartz augen and of mica and chlorite flakes; it lies in the cleavage plane with a westerly pitch of 75-90° (Fig. 50). This lineation is particularly obvious in the sericite-fleck schists in which sericitic lenses (which appear to be altered feldspars) are elongated steeply (Plate 53, Nos. 1 and 2).
Fig. 49: North-south cross-section from North Lyell to the Blow, showing the form of the TF$_2$ folds, and also the native copper-goethite deposits.
Fig. 50: Stereographic projection (lower hemisphere, equal area) of $T_1$ and $T_2$ in the North Lyell-West Lyell area. The inclined plane is an average cleavage orientation.
The shape of the flecks (in terms of three axes) has been measured in three rocks (32740, 31152, 31107) and results for two of them are shown in Fig. 51. The longest axis (A) coincides with Tl, the intermediate axis (B) is almost horizontal and within the cleavage, and the shortest (C) is at right angles to the other axes and the cleavage. These measurements were made in horizontal and vertical (perpendicular and parallel to cleavage) thin sections and therefore all three axes for a particular individual could not be found. The results show only the statistical distribution of values (A in vertical thin sections, B in horizontal thin sections and C in both sections, there thus being more values of C than A or B) but give the general shape of the flecks. Averages of A, B and C are also shown in Fig. 51.

A convenient measure of the deformation is the A/C ratio and this can be measured for each individual seen in vertical thin sections; Fig. 51 shows the distribution of such ratios, the average figures from 31107 and 31152 being 8.5 and 4.7 respectively. The higher figure need not indicate a greater degree of deformation of the rock as a whole because the physical nature of the altered feldspars may have varied from place to place.

The unaltered feldspars of the keratophyres are stumpy laths in which the maximum A/C ratio observed was 2:1. Hence,
Fig. 51: Frequency distributions of A, B and C (axes of sericitic flecks) and distribution of A/C ratio for (a) 31152 and (b) 31107. Note that 15 divisions = 1 mm.
(a) FOR 31152
(b) FOR 31107
assuming these flecks represent original tabular feldspars, the
deformation of some has been considerable and has involved
shortening normal to the cleavage and extension in the lineation
direction. A few flecks are roughly square in section and
apparently undeformed, others show a little deformation (low A/C
ratios) and others apparently are considerably deformed, so that
the deformation is apparently inhomogeneous.

Wilson (1961, p. 487) has recently given a summary of
work on elongation of megascopic bodies during deformation.
Quantitative work has been done on belemnites (Heim, 1878);
ooliths and crinoids (Cloos, 1947), brachiopods, and particularly
spiriferids (Haughton, 1856, Sharpe, 1847, 1849, Sorby, 1856);
lava pillows, amygdules, agglomerate fragments (Wilson, 1951;
Cloos and Heitanen, 1941); and "eyes" in slates (Green, 1917;
Wilson, 1951). Wilson did not mention the preliminary work of
Bryan and Jones (1955) on deformed radiolaria in Queensland
sediments, or the work of Hellmers and Kurtman on crinoid stems,
as described by Nissen (1964).

A number of workers (e.g. Flinn, 1961; Elwell, 1955) have
studied elongation of pebbles in deformed conglomerates.

It is noted that Cloos found considerable variation over
small areas in the degree of distortion of the South Mountain ooliths,
very similar to the variations found in the Lyell rocks. Similarly, Flinn (1961) referred to varying degrees of elongation of pebbles in a deformed conglomerate.

In most cases, the direction of elongation is at right angles to the fold axis but in some it is parallel to the fold axis (e.g. Haughton, 1956) and particularly in deformed conglomerates (e.g. Flinn, 1961) and crystalline schists (Wilson, 1961, p. 491). Dr. A. Spry (pers. comm.) has suggested from a study of deformed conglomerates that elongation is generally parallel to the fold axis.

Elongation of bodies during folding may reasonably be likened to the parallelism of freely moving logs in a stream, and Jeffery (1921), studying the theoretical behaviour of ellipsoidal bodies in a viscous fluid in motion, concluded that they would line up parallel to the direction of movement and rotate about an axis parallel to that direction. Folds commonly form with axes perpendicular to a, the direction of transport, but may also form parallel to a, as indicated by studies of salt domes, both in the laboratory (Escher and Kuenen, 1929) and in the field (Muehlberger, 1959).

In the Lyell Schists the orientation of the fold axes is likely to be variable (as they are in part at least superimposed on the earlier N-S structures) but the elongation direction (T1) is at right angles to the WNW fold axes in the Owen sediments and the
south-westerly plunge of $\mathbf{T}_1$ reflects the asymmetry of most WNW folds.

The production of lineations in a and of flow cleavage is related essentially to rock elongation which may or may not involve shortening or compression normal to the cleavage. Sorby (1856) and Sharpe (1847, 49) related elongation to compression and many more recent workers have referred only to compression as the cause of cleavage (e.g., de Sitter, 1956). However, Fourmarier (1949) and Wilson (1961, p. 483) have emphasised the importance of stretching and Carey (1954, p. 196) has shown that severely appressed similar folds (with or without cleavage) may develop without any compression normal to the fold axes.

Because some shortening seems inevitable in the WNW folds in Owen sediments, it probably also was involved in the folds of the Lyell Schists. The schists have undergone recrystallisation in the presence of fluids (juvenile or connate) and recrystallisation has probably been important in some schists in cleavage and lineation development. However, rotation of feldspars particularly may well have occurred in the early phases of deformation, in a sequence of rotation and recrystallisation such as that envisaged by Collette (1958) for the development of cleavage. In the augen types, and in fact in many of the sericitic schists, oriented sericite
flakes tend to be aligned in two directions which make a smaller angle of 30° to 45° with each other. A similar "double" orientation has been noticed in many mildly altered keratophyres. In most cases there are clearly two distinctive fracture trends visible on both the field and the microscope scales, with their acute angle enclosing the direction of elongation. Commonly the fractures do not have parallel strikes, in which case the acute angle produced encloses the general cleavage strike. Similar features may be seen in the Owen Conglomerate on a coarser scale (Plate 60, No. 2).

These shears outline the augen in the schists just described, the size of the augen being controlled by the angle and frequency of the shears. Increased frequency of shears goes with reduced acute angle and though no continuum has been observed, presumably the end product is a finely granulated rock with sub-parallel cleavage surfaces. This apparent development of the cleavage is similar to that postulated by Becker (1904) who suggested conjugate shears that were formed by compression rotated to sub-parallel alignment on surfaces roughly perpendicular to the compression direction. Actually no fractures rotate in the strict sense but as new fractures form and frequency of fracturing increases, the acute angle decreases about the elongation direction and the rock becomes
more and more granulated. This scheme of development probably only applies to competent rocks; in incompetent rocks cleavage may be largely due to rotation and recrystallisation as envisaged by Collette. The sericite flakes may form in the fractures as they develop or, perhaps more likely, may be mimetic.

Cleavage in the Lyell Schists, then, may develop in a variety of ways, apparently depending on rock type. The processes indicated include changing directions of shear planes, rotation of crystals and pebbles, and recrystallisation of mica and chlorite. Probably both elongation and compression are involved.

Much of the variation and crudeness of the cleavage and the steep lineation is due to the effects of the post-cleavage recrystallisation previously described in the schists. This recrystallisation may have destroyed orientation in the sericite crystals, leaving only a coarse orientation to slivers and bunches of sericite, and may have destroyed any quartz microfabric that had developed. The sericite crystals are too fine-grained to study optically but the quartz crystals of a few schists are sufficiently large to measure accurately the optic axis orientations on the Universal stage at magnifications of about 300. The results of several such examinations showed a random orientation.
(d) Lineation $Tl_2$

A second lineation ($Tl_2$) is apparent on many discrete cleavage surfaces throughout the mine. It consists of fine slickensiding (Plate 61, No. 1) with flattish plunge (Fig. 50) and clearly is later than $Tl_1$. It is probably due to transcurrent movements related to the sinistral phase of the major WNW faults and is evidence that this phase followed the vertical displacement associated with $Tl_1$.

(e) Boudinage in the Schists

Large scale boudinage of the Schists (Plate 61, No. 2 and Plate 62, No. 1) may be seen in the north-west part of the West Lyell Open Cut. The axes of the boudins are approximately horizontal and individual boudins measure from 30 cm. to 9 m. in width and 20 cm. to 4 m. in thickness (using the terminology of Wilson, 1961, p. 497-8). No measurements of length were possible and the separation varies from almost nil to a metre or so (Plate 63, No. 1). The cleavage wraps around the boudins in classical style and many are almost oval in cross section; they appear to have formed by necking as a result of stretching (as in Rast, 1956) in the dip of the cleavage i.e. in the direction of $Tl_1$. The boudins presumably developed in more competent bands in the schists during cleavage development, the bands representing
Plate 61 No. 1:– Horizontal slickensides on a cleavage face, looking north easterly. The pencil at bottom right is approximately 15 cm. long.

Plate 61 No. 2:– Boudins in the Lyell Schists, looking northerly, from the north west part of the West Lyell Open Cut.
Plate 62 No. 1:— Boudins in the Lyell Schists, looking northerly, from the north-west part of West Lyell Open Cut.

Plate 62 No. 2:— Crush zone close to the schist-Owen contact, Royal Tharsis Open Cut, looking north.
Plate 63 No. 1:– Sub-horizontal kink-bands in the Lyell Schists, West Lyell Open Cut.

The face is about 5 m. high.

Plate 63 No. 2:– Close up of kink band in schist, looking southeasterly almost along the kink axis.

The match box is approximately 5 cm. long.
inhomogeneities that were original or that developed by localised recrystallisation during hydrothermal alteration. The development of the boudinage was probably facilitated by the presence of fluids causing the alteration.

A study of the literature on boudinage indicates that most boudins develop with their lengths parallel to $b$ but that some are parallel to $a$ (e.g. Coe, 1959). At Lyell, the boudin lengths are more or less at right angles to $T_{11}$, which is almost certainly $a$, so they are essentially $b$ structures.

Kink Bands and other late-Stage Structures

The Owen-schist contact was faulted at some post-cleavage stage. The contact is exposed on the east wall of the Royal Tharsis Open Cut (an offshoot of the West Lyell Cut) and the schists close to the contact are cut by a crush zone about 40 cm. wide. The fault zone dips westerly, is filled with crushed quartz, and the adjacent schists show associated tensional, quartz filled "feather fractures". These are horizontal and indicate, along with shear fractures in the crush zone, that the fault is reversed (Plate 62, No. 2). This late contact faulting clearly cuts a WNW fault in the middle of Tharsis Ridge (see Fig. 35) but does not appear to be important elsewhere.
Another feature of the schists in the West Lyell Open Cut is the presence of numerous, sub-parallel, more or less horizontal kink-bands (Plate 63, No. 1). These appear to be typical late-stage kink-bands as described by Anderson (1964). The axes of the angular bends in the kink-bands are close to horizontal and trend between $110^\circ$ and $125^\circ$ with the majority at $120^\circ$, i.e., close to the strike of the cleavage. The kinks result in a shortening in the direction of $T_{11}$ in the plane of the cleavage, representing the reverse of the movement during cleavage development. The majority (e.g., Plate 63, No. 2) show sinistral displacement (or top side west viewed along kink axes) but a few show dextral displacement. These two types are almost parallel and can hardly represent conjugate shears. The width of the bands varies little from 4 or 5 cm, and individual bands may be traced for several hundred metres.

Numerous post-cleavage joints and faults cut the schists. The majority are roughly parallel to cleavage and probably formed with that orientation because of the pre-existing planar structure in the rock.

A few N-S to NE faults cut the WNW structures in the Queenstown district and may be partly of tensional origin. A major NNE fault (the King River Fault) cuts off the east ends of
Mt. Owen and Mt. Lyell and a possible parallel fault may have influenced the development of the East Queen River (see Fig. 3).

Folds in the Sericite Schists

A feature of the quartz sericite schists interbedded with chlorite schists on Philosophers Ridge is the presence of banding displaying extreme contortion. In places the banding is parallel to the chlorite beds, and folded with these beds, but in other places it is tightly folded while adjacent chlorite beds are not. The folded layers between unfolded chlorite beds may be only a few metres thick.

A typical exposure is shown in Fig. 52 a and c, from which the following observations can be made:

(a) There is a crumpling about roughly E-W axial surfaces; this crumpling may be an expression of TF₂ cleavage folding though considering that it is confined to a small area there is a possibility that it is due to pre-cleavage folding.

(b) Subtracting this crumpling (see Fig. 52 b) reveals the presence of small doubly plunging folds that are paraboloidal in form. The "eyed" or "onion-peel" folds are earlier than those of phase (a).

(c) In vertical section (Fig. 52 c and Plate 64, No. 1) two stages of folding are again evident (Fig. 52 d). There appears to be a late
Fig. 52:  Folds in sericitic schists of Philosophers Ridge:

(a) Plan of outcrop at 6000 S/00 E (Fig. 35).

(b) The same area, showing traces of axial surfaces of folds.

(c) Vertical section of same exposure, looking north.

(d) The same area showing traces of axial surfaces of folds.
(a) Plan

(b) Axial traces of E-W trend
Axial traces of earlier folding
Possible traces of earlier folding

(c) Vertical section, looking north at face A B of (a)

(d)
Plate 64 No. 1:— Photo of exposure shown in Fig. 52 (c); vertical face, looking north. Pencil approximately 15 cm. long and pointing upward.

Plate 64 No. 2:— "Eye" folds in sericite schist of Philosophers Ridge.
folding with axial surfaces parallel to the general strike and
dipping steeply, and this folding produces the hook structures
seen in Fig. 52 c. The earlier folds may be sections of the
paraboloidal folds of (b).

To conclude, there appear to be three stages of folding:
(1) paraboloidal (2) on trends roughly N-S, and (3) on trends
roughly E-W.

(1) "Eyed" folds are fairly common/Philosophers Ridge (Plate
64, No. 2) and are generally elongate in plan with their elongation
in the plane of the adjacent banding. There are also normal
open folds that crumple the banding and are unrelated to cleavage
and these may belong to the same phase of deformation. The
axial surfaces of these folds are roughly perpendicular to the
general banding and their axes are steep.

The folds of phase (1) may well be pre-consolidation
structures much like those observed in the banded tuff (?) of Whip
Spur (Plate 41, Nos. 1 and 2), their elongation parallel to the
banding being caused by later deformation (phase (2) ?). In places
(e.g. at 52°00'S / 52°E ) there are exposures of apparently
chaotic folds of varying shape and orientation that are pre-cleavage
and probably also pre-consolidation.

(2) Folding of this phase is not much in evidence though in places
both the chlorite beds and the sericite schists are crumpled on small N-S trending folds. These folds may be Jukesian or TF$_1$.

(3) The very common crumpling on roughly E-W axes is probably related to TF$_2$ cleavage.

Shearing and Metasomatism of the Owen Conglomerate

Over very small areas of the schist contact, the Owen Conglomerate is deformed and apparently sericitised. For example, near the southern end of the Tharsis Ridge, there are two or three exposures of the contact and for several cm. in from the schist the Owen is crudely schistose. There appears to have been localised elongation or shearing in a near-vertical direction, producing elongation of some pebbles and, in one case, separation of two halves of a pebble (Plate 65, No. 1). The direction of movement is approximately parallel to Tl$_1$. Lenses and slivers of sericitic schist intrude the conglomerate along certain cleavage surfaces on macroscopic (Plate 65, No. 2) and microscopic scale. Study of thin sections of the sheared sediments shows considerable variation. Some show pronounced, sericite-filled fractures cutting undeformed sandstone, others show closely spaced fracture surfaces, penetrative disruption of the sandstone texture and undulose extinction in the quartz grains, and still others have apparently
Plate 65 No. 1:- Separation of the two halves of a quartz pebble in the direction of T1 in Owen Conglomerate, south end of Tharsis Ridge. The coin is approximately 3 cm, diameter.

Plate 65 No. 2:- Schist "intruding" Owen Conglomerate near schist contact, south end of Tharsis Ridge.
undergone little deformation but consist of quartz grains in a randomly oriented sericitic matrix.

Though the sericite matrix of the latter types might be interpreted as a replacement of quartz, there is no direct evidence of replacement; the sericite may in fact represent recrystallised clayey material that formed in localised, muddy sandstone at the base of the Owen, or may be sericite intimately mixed with disrupted Owen sandstone during tectonic movements and later recrystallised to produce a random arrangement.

Bradley (1957) attempted to explain these features by hydrothermal metasomatism. In Plate A of his paper he showed several photomicrographs of Owen sediments and two of sheared, sericitic Owen sediments, all gathered close to the schist contact in the Blow Open Cut. He suggested that they showed a gradual transformation from sediment to schists. In support of this he quoted (p. 174) alkali analyses of 0.1% Na₂O and 1.6% K₂O for schist close to the contact and 0.4% Na₂O and 4.2% K₂O further from the contact. Examination of this contact, which is a NW trending reverse fault, shows a very narrow zone up to 10 cm. thick, of apparently sheared, sericitic Owen sediment separating sericite schist and Owen sandstone. Samples from this zone (31718, 31718 a) consist of quartz grains in a sericitic base and, like some material from the Tharsis contact, may be the result of tectonic mixing and
later recrystallisation. The increased amounts of each alkali, noticeably in the same ratio, in the schists as compared to the contact material is merely evidence of an increase in mica content. Bradley (1954, 1957) believed that many hundreds of feet of Owen sediments were metasomatised and that the contact represents the eastern limit of the sericitisation "front". However, no evidence of sericitisation of quartz has been found and the writer suggests that the apparent sericitisation is largely (though possibly not entirely) a structural phenomenon.

The largest area of deformed Owen Conglomerate is at Great Lyell, on the downthrown side of the South Owen Fault and at the western limit of a long tongue of Owen Conglomerate (Fig. 36). Here the Lower Owen Conglomerate, for several metres from the fault surface, shows a crude cleavage and parallel elongation of pebbles (Plate 66, Nos. 1 and 2). The pebbles are up to 25 cm. long and vary in shape; some are cigar-like or very prolate spheroids, others tend to parallelepiped form, and a very few of the smaller ones are roughly equidimensional. A few are twisted, possibly due to interference with adjacent pebbles. The matrix of relatively unsheared coarse quartz sandstone is so sparse that many pebbles are in contact.

The shapes of the pebbles are conveniently summarised by
Plate 66 Nos. 1 and 2:– Deformed Lower Owen Conglomerate on the South Owen Fault, looking easterly slightly oblique to the cleavage. The pencil is approximately 12 cm. long.
plotting on a triangular diagram according to the technique used by Sneed and Folk (1958), who used the parameter $\frac{C}{A} \cdot \frac{A-B}{A-C}$ and $3 \sqrt{\frac{C^2}{AB}}$ (Fig. 53), where $A$ is the longest axis, $B$ the intermediate axis and $C$ the shortest axis.

For comparison with undeformed pebbles in the Owen Conglomerate, measurements were made at several points in undeformed Lower Owen Conglomerate on Mt. Owen, and the results are also given in Fig. 53. From this comparison it is clear that elongation has taken place at Great Lyell but to varying degrees.

The Great Lyell pebbles show a pronounced parallelism of their long axes (Fig. 54) which pitch steeply south-west in the cleavage, in much the same manner as $T_1$ in the Lyell Schists. In comparison, the orientation of the long axes in undeformed conglomerate is essentially random (Fig. 54) though with some tendency to lie in the bedding plane.

The South Owen Fault has a vertical component of displacement of 300 m, and the elongation of pebbles probably reflects the direction of movement, i.e., they are parallel to $a$. The fault also displays horizontal movement but this appears to have been confined to the pre-existing fault surface, on which gently plunging slickensides may be seen in places.
Fig. 53: Shape distributions in undeformed and deformed Owen Conglomerate from the Great Lyell area, using the method of Sneed and Folk (1958).
3√S² / LI

L-I
L-S  256 undeformed pebbles

S / L

L-I
L-S  240 stretched pebbles
Fig. 54: Stereographic projections (lower hemisphere equal area) of pebble axes from Owen Conglomerate. The upper diagram (a) shows orientation of A and B axes of deformed pebbles at Great Lyell, the lower diagram (b) shows plots of the long axes of undeformed pebbles from Mt. Owen, all plotted with respect to the bedding shown. Contours are at 2 and 5%.
(a)

Area encloses 243 points for A axis

Area encloses 193 points for B axis

(b)

15 Points

3 Points

1 Point

2 Points
The relatively small amount of elongation of the pebbles and the relatively unsheared nature of the sandstone matrix suggests the possibility that the pronounced parallelism of the pebbles is partly due to rotation into the movement direction. Many of the pebbles were elongate initially and stretching may have affected only a minor proportion of the pebbles.

The direction of the long axes of the pebbles is parallel to $T_{1\perp}$ in the schists and probably reflects the same kinematic pattern.

**Mineralization at Mt. Lyell**

The salient features of the Lyell orebodies have been described in several earlier papers e.g. Alexander, 1953; Wade, 1957; Wade and Solomon, 1958; Solomon and Elms, in press, Appendix A, no. 2. They are largely post-cleavage replacement bodies and divide roughly into four types:

(a) Massive pyrite-chalcopyrite ores with no more than 1% Cu (the Blow or Mt. Lyell Orebody, South Lyell and the Eastern Orebody).

(b) Disseminated pyrite-chalcopyrite ores with a higher maximum copper content (the Prince Lyell, West Lyell No. 1 or Honeypot Orebody, and the Comstock orebodies).
(c) Chalcopryite-bornite ores with high copper content (the orebodies of the North Lyell area).

(d) Banded pyrite-sphalerite-galena lodes east of Comstock (the Tasman and Crown Lyell orebodies).

Deposits of goethite carrying native copper and cuprite occur on the lower eastern slopes of the Mt. Lyell-Mt. Owen divide within the Gordon Limestone (Figs. 35, 49). These deposits have been the subject of a separate study, which has shown them to be oxidation products of mineralization in limestone combined with deposition from copper-bearing solutions draining from exposed sulphide ores.

(a) The massive pyrite-chalcopyrite ores contain up to nearly 90% pyrite and the copper content ranges from nil to a little over 1%. The South Lyell orebody is typical of the low-copper ores and was mined largely as a flux for the siliceous ores of North Lyell. In plan it is elongate along the schistosity and its longest axis dips SW at about 75° i.e., parallel to T1 (Fig. 55).

The Blow orebody, 60 m. away and at the schist-Owen contact, carried significant chalcopryite. The discovery of the limonite-barite gossan, adjacent to the Iron Blow, led to the development of the Mt. Lyell field in
Fig. 55: Cross-section of the Blow Orebody, showing reconstructed original surface and the estimated positions of the limonite-barite gossan and the Iron Blow.
the final years of the 19th century. The orebody moulds itself on the contact and pinches out as the contact flattens some 240 m. below the original surface (Fig. 55). This "ledge" may represent the equivalent of the Tharsis Schist Zone. The ore is essentially massive pyrite with minor amounts of chalcopyrite, enargite, tetrahedrite, bornite, chalcocite, galena and sphalerite. The upper levels averaged 2.85% Cu but grade declined with depth and the remaining 1.5 million tons average only 0.5% Cu. The walls of the ore are well defined though veins of massive pyrite up to about 20 cm. thick penetrate the surrounding schists. Of the small, rich shoots within the body the most important was the Mt. Lyell bonanza, which yielded 850 tons assaying 21.3% Cu and 1,023 oz. Ag per ton, and contained stromeyerite, bornite, chalcopyrite and gold (Petterd, 1910).

Many of the pieces of pyrite ore remaining on dumps (31191a, b, c) show banding highlighted by variations in the grain size of the pyrite (Plate 67, No. 1) and possibly inherited from the cleaved schists.

(b) The disseminated pyrite-chalcopyrite ores seldom contain more than 35% sulphide minerals. They are of larger size and are further from the Owen-schist contact, with the copper content an inverse function of the distance from the contact. The orebodies are lenticular in plan and elongate roughly parallel to the cleavage but persist
Plate 67 No. 1: Banding in massive pyrite ore,
Blow Open Cut, X 1.

Plate 67 No. 2: Aerial view of the West Lyell Open Cut, looking north east. Mt. Lyell in background, Royal Tharsis Open Cut at centre left.
in depth for considerable distances, pitching steeply and
tending to dip slightly more steeply than the cleavage; as a
result they approach the contact in depth. As one orebody
dies out in depth, an orebody develops en echelon further
from the contact. The mineralogy is very simple, 98% of
the ore minerals being pyrite and chalcopyrite.

A group of these disseminated ores is currently mined
in the West Lyell Open Cut, which yields some 2,000,000
tons per annum of ore carrying about 0.7% Cu. An aerial
view of the pit is shown in Plate 67, No. 2. The spread
of copper values makes delineation of orebodies a somewhat
subjective process but Wade (1957) was able to outline two
major areas of more intense mineralization viz. the West
Lyell No. 1 (Honeypot) Orebody and the Prince Lyell Orebody.

Subsequent drilling below the open-cut has shown that the
walls of these bodies became rather better defined in depth.
Mineralization has now been proved to have a vertical extent
of over 700m., without significant change in primary miner-
alogy or nature of wall-rock alteration. For example, the
chlorite schist of 31002 (from diamond drill hole WL 112,
approximately 700 feet below sea level) is similar to specimens
gathered at the surface (using Schoen's method of analysis)
and the sericite at depth (31193, from WL 112 at sea level) gives a similar X-ray pattern to those at the surface.

In addition to pyrite and chalcopyrite occasional patches of bornite and chalcocite have been found in drill cores and molybdenite occurs on fracture surfaces in the Honeypot orebody. Enargite, tennantite and tetrahedrite occur generally on the microscopic scale but in the Mt. Lyell Reserve workings, large specimens of tetrahedrite were discovered. Hematite and some magnetite may be seen in small patches and are invariably pre-sulphide in age (31128, 32699). The sulphides are generally unsheared (e.g. 31711A, 31163, 31193) and clearly post-cleavage but on discrete surfaces showing Tl2 the sulphides are finely striated or polished by the sub-horizontal striae.

The Royal Tharsis orebody, closer to the contact, is a fairly well-defined lens of similar type and form to the Honeypot and Prince Lyell orebodies, though carrying 1.6 to 2.0% Cu. It is shown in slightly oblique section in Fig. 56.

The Comstock orebodies are rich in chalcopyrite relative to pyrite but the proportions are not reliably known as the mines have been closed since the second World War. The presence of bornite and relatively low pyrite content
Fig. 56: Cross-section of the Royal Tharsis orebody, on grid line 2000 S (slightly oblique to the orebody).
indicates that these ores are transitional to the bornite-chalcopyrite group. There are four separate orebodies which are arranged roughly en echelon when projected onto a section (Fig. 57) and lie alongside the Comstock chert (Fig. 35). In the upper levels the orebodies trend approximately NE but swing in depth to N, roughly copying the behaviour of the Owen-schist contact. Edwards (1939) noted that in the upper levels the local strikes of the orebodies are parallel to two rather variable cleavage directions. Besides bornite, chalcopyrite and pyrite, Edwards (1939) found magnetite (probably pre-ore), galena and free gold. Patches of sphalerite-galena-pyrite ore, locally with bornite (31743, 32705, 32769), may be seen in the open-cut.

Chalcopyrite-rich ore relatively low in pyrite forms the Crown Lyell No. 1 orebody which lies in the North Lyell Corridor close to the North Lyell bornite-chalcopyrite ore. It is a discrete orebody that strikes N and plunges about 60° to the NW.

Chalcopyrite ore with bornite and galena occurs in the Lyell Tharsis orebody, just south of the North Lyell Open Cut. Some of the workings followed rich, late-stage veins of chalcopyrite-galena-sphalerite (31206).
Fig. 57: *Cross-section of the Comstock mine area,*
on grid line 5600 N.
West and north-west of the Tharsis Ridge are large zones of pyritic schist carrying less than 0.5% Cu and similar lenses occur south and south-west of the West Lyell Open Cut e.g. in the Glen Lyell area and in Conglomerate Creek (Fig. 35, 36).

All these disseminated ores show splashes or "bursts" of quartz that disrupt the cleavage and many of these carry siderite, chlorite, chalcopyrite and a few also carry pink or reddish albite. The time relationship between these veins and the disseminated sulphides is not clear.

(c) Bornite-chalcopyrite ores only occur significantly in the North Lyell area, which has yielded some 5 million tons of ore averaging 5% Cu, 1.12 ozs. Ag per ton and 0.013 ozs. Au per ton. The ore occurs in irregular pipes, pods and lenses that commonly occur at the boundaries of chert and chloritic or sericitic schist (see Fig. 48). The main mass of chert lies adjacent to the Owen Conglomerate and tends to act as the hanging wall of the mineralized ground, though it is highly mineralized in places.

The orebodies tend to strike either WNW or N-S, parallel to the Owen (or chert)-schist contact. Their dip varies with the attitude of the schist contact.
The rich orebodies are generally composed of bornite, bornite-chalcopyrite, or chalcopyrite with minor amounts of pyrite. However, within the corridor there are also highly pyritic lenses (e.g. the Eastern Orebody) and large pyrite-chalcopyrite lenses (Crown Lyell No. 3) similar to the disseminated types west of the schist contact; several million tons of 1-3% Cu ore have been outlined during the last ten years.

Edwards (1939) and Markham (1963) have recorded magnetite, sphalerite, galena, tetrahedrite, stannite, orange stannite, digenite, chalcocite, sulphosalts of copper, pentlandite, linnaeite, gold and electrum. Magnetite is earlier than all other minerals (31201, 31194).

A small bornite-rich lode occurs in schists east of Comstock in the Tasman and Crown Lyell workings, within about 70 metres of an argentiferous galena-sphalerite-pyrite-chalcopyrite-tetrahedrite lode. This ore, exposed in dumps near the old Tasman shaft is in part finely banded, and the banding is commonly folded (32693, 31261, 39744A, 39745A). It is very similar to the ore of the Rosebery and Hercules Mines.
Mineralogy and Temperature of Ore Deposition

The mineralogy of the sulphide ores has been described by Petterd (1910), Edwards (1939) and Markham (1963) and the following minerals have been recorded by these workers: pyrite, chalcopyrite, bornite, chalcocite, digenite, magnetite, hematite, galena, tetrahedrite, tennantite, enargite, molybdenite, gold, electrum, pentlandite, stromeyerite, linnaeite, arsenopyrite, stannite and "orange stannite". Covellite and native copper are regarded as secondary. Magnetite and hematite are probably pre-Tabberabberan and there is a possibility that covellite intersected in drill holes near sea level below the West Lyell Open Cut is primary (cf. Markham, 1963, p. 137).

Markham concluded that the paragenesis of the ores was as follows:

- pyrite (earliest),
- pyrite and chalcopyrite, pyrite and bornite,
- chalcopyrite and bornite,
- bornite and chalcocite,
- bornite and digenite.

This sequence involves increasing Cu: Fe and Cu:S ratios.

Lack of experimental data (particularly below 400°C) prevented Markham from determining the conditions of deposition but suggested
the sequence above would follow declining temperature and sulphur vapour pressure.

The sphalerite geothermometer (Kullerud, 1953, 1959; Barton and Kullerud, 1958) is probably of limited use at Mt. Lyell because (a) sphalerite is rare (except at Tasman and Crown Lyell and (b) pyrrhotite is absent and (c) in all specimens examined by the writer (e.g. 31709A, 32693, 31261 etc.) sphalerite appeared to replace pyrite and equilibrium could not be proven and (d) the sulphur vapour pressure is unknown.

The available chemical information for Mt. Lyell sphalerites is summarised in Table 27. In order to extend the study to microscopic quantities some determination of the cell edges of sphalerite were made by X-ray diffraction. The cell edge varies with the Fe, Mn and Cd contents (Skinner, 1961), so that 

$$a = 5.4093 + 0.000456X + 0.00424Y + 0.00202Z$$

where X, Y and Z are respectively the FeS, CdS and MnS contents in mole per cent.

The effects of CdS and MnS on the cell edge are considerable (particularly CdS) and as these components may be to some extent dependent on temperature of deposition or at least FeS content (see, for example, Edwards, 1955, Bradbury, 1961 and Table 27) the estimation of FeS content from X-ray data is clearly open to considerable error. For example, Groves (1963) found variations in CdS in sphalerite from Mt. Bischoff, Tasmania were
<table>
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<tr>
<th>Mineral Association</th>
<th>Fe</th>
<th>Mn</th>
<th>Cd</th>
<th>a of unit cell (meas.)</th>
<th>a (calculated)</th>
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<td>32664a. Blow Mine</td>
<td>7.10</td>
<td>0.49*</td>
<td>0.3</td>
<td>5.4174 ± 0.0005</td>
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<td>32664b. Blow Mine</td>
<td>6.5</td>
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<td>32980a. Blow Mine, No. 6 level</td>
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<td>32980b. Blow Mine</td>
<td>1.07</td>
<td>0.005*</td>
<td>0.18*</td>
<td>5.4120 ± 0.0005</td>
<td>5.4109 (average)</td>
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<td></td>
<td>1.02*</td>
<td>0.005*</td>
<td>0.24+</td>
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<tr>
<td>31709A. Lyell-Tharsis</td>
<td>0.54*</td>
<td>0.009*</td>
<td>0.27*</td>
<td>5.4111 ± 0.0005</td>
<td>5.4170</td>
</tr>
<tr>
<td></td>
<td>0.55*</td>
<td></td>
<td>0.28+</td>
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<tr>
<td>32705. Comstock</td>
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<td></td>
<td>5.4201 ± 0.0005</td>
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<td>31206. Lyell Tharsis</td>
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<td>32693. Tasman and Crown Lyell dump</td>
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<td>32981. Great Lyell</td>
<td>1.70</td>
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<td>5.4101 ± 0.0005</td>
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</table>

* X-ray Spectrograph Analyses.  
+ Polarographic  
◦ Colorimetric  

Analyst: Aust. Min. Develop. Labs. except for 32664b (Mt. Lyell Mining Co.) and 32693 (M. Solomon).
large and liable to render cell edge measurements useless as indicators of FeS content.

However, the cell edge measurements on Lyell sphalerites, which broadly match the colour variations, reveal a wide range in the \((\text{FeS} + \text{MnS} \text{ and } \text{CdS})\) contents and also indicate that an FeS content of 7.25\% is the maximum value. This would suggest a temperature of deposition of 450 to 720°C, depending on sulphur vapour pressure (Sims and Barton, 1961).

A geothermometer of possible application to Mt. Lyell involves variation in the thermo-electric potential of pyrite (Smith, 1947). Smith's premises concerning the variation in thermo-electric potential with temperature of deposition have been criticised in recent years as a result of laboratory investigations on pyrite, particularly in Japan (e.g. Sasaki, 1955; Suzuki, 1963). Sasaki found that the sign of the thermal EMF changed rapidly within single crystals, and Fischer and Hiller (1956) and Suzuki (1963) have shown that the electrical properties are largely dependent on the trace elements in the lattice, and these are probably controlled essentially by the composition of the ore-forming solutions. Suzuki concluded, after exhaustive testing, that the thermo-electric potential is most directly affected by the ratio \(X/M\), where \(X=\text{total sulphur} + \text{anion non-metallic impurities}\) and \(M=\text{total}\)
Examination of the thermo-electric potential of pyrite at Lyell was carried out using steel probes 2 mm. apart, one heated to about 100°C, the thermal EMF being measured on a high resistance galvanometer. Two standards were kindly supplied by Professor F. G. Smith, one from Bevcourt Mine, Quebec, supposedly formed at 600°C and the other from El Potosi, Mexico, supposedly formed at 150°C. Study of a large number of pyrite specimens, both single crystals and polished and unpolished aggregates gave the following results:

(a) Some single crystals and also aggregates showed a wide variation in readings while others gave a small variation.

(b) Most, but not all, specimens (single crystals or otherwise) gave modal peaks after a number of readings had been taken.

(c) All readings were positive or zero, positive being used when the current flows in the pyrite in the same direction as the thermal gradient.

(d) The Mt. Lyell readings, taken together, lie between the standards and give a modal peak near the lower temperature ore (Fig. 58).
Fig. 58: Distribution of thermo-electric potentials of Lyell pyrites (thin continuous line), as compared to the 600°C standard (dashed line) and the 150°C standard (thick continuous line). Convert galvonometer scale readings to potential (millivolts) by dividing by 7.5.
In summary, the temperature of deposition of the Mt.
Lyell sulphides is uncertain but probably took place below about
700°C. The discussion on hydrothermal alteration would indicate
a temperature below 600°C.

Little information is gained from the gangue minerals,
many of which are stable over a wide temperature range. Smoky
quartz and mauve fluorite may indicate deposition below 260°C
and 175°C respectively (Bateman, 1956, p. 42-43), probably in
the late stages of the mineralization.

Zoning

A rather poorly defined zoning of sulphide assemblages
has considerably influenced exploration at Mt. Lyell for many years.
Massive pyrite-rich deposits dominate the area south of West Lyell,
pyrite-chalcopyrite ores dominate the central part of Philosophers
Ridge, while bornite-rich, low-pyrite ores are dominant in the
Comstock-North Lyell area. Galena and sphalerite are concentrated
near Comstock and are found at Queenstown and on Little Owen,
south of the Great Lyell area; they thus form a sort of peripheral
fringe.

The zoning is poorly defined because pyritic zones are found
at North Lyell and near Comstock, and galena is known from the
North Lyell and Blow mines.

A similar zoning is present in the mineralized belt at Lake Dora (Fig. 24) where the ores are entirely pyrite in the south, pyrite and chalcopyrite in the centre, and sphalerite-chalcopyrite-galena in the north.

The significance of this zoning is not fully understood.

Geochemical Criteria Related to the Origin of the Ores

The S:Se ratios of West Lyell pyrite concentrates have been determined by Edwards and Carlos (1954) and range from 9,601 to 10,892, and chalcopyrite values are 8,253 and 8,075. These values suggest a magmatic hydrothermal origin.

Similar indications are given by the Co:Ni ratios for the pyrite concentrates. A value of 8 (250:30 and 200:25) was obtained by X-ray spectrographic analysis of two samples of pyrite concentrate, suggesting a magmatic origin (Davidson, 1962). Neither of these two chemical criteria are foolproof (particularly the Co:Ni ratio) and care must be taken in accepting them as evidence of genesis.

Sulphur isotope analyses currently being undertaken by Professor Jensen (Yale University) on both Rosebery and Lyell ores are not yet available.
A number of lamprophyre dykes intersect the mineralized zone at Lyell. Three of these have been found in the mine area (Figs. 35, 36; Plate 68, No. 1) and though they are deeply weathered, fresh material has been obtained from subsurface workings and drill cores. The dykes are unmineralized and unsheared, have chilled margins and occupy dilational cracks parallel to the Tabberabberan cleavage, and they appear to be later than all the visible tectonic features. They vary from a few cm. to almost 4 m. in thickness.

Fresh material has been obtained from the dyke in the West Lyell Open cut (32679, 32784, 32778, 32779, 31712A, 31713A) and from the one on the Gormanston-Queenstown road (ML 462). All specimens are a dark brownish-grey porphyry which in thin section is seen to consist of olivine, augite and biotite phenocrysts in a groundmass of dark material that is partly fibrous and weakly birefringent (serpentine, chlorite ?), and partly glassy containing pale birefringent patches that may be feldspar or feldspathoid.

The olivines are up to 3 x 2 mm. in section and grade down to 0.3 mm. across. They vary from anhedral to euhedral, are cracked and partly serpentinised, and in some rocks (e.g. 32679)
Plate 68 No. 1:—Lamprophyre dyke cutting schist and quartz vein, West Lyell road near Smelters. Scale in inches. Photo by M. L. Wade.

Plate 68 No. 2:— Agglomerate (?) in Quartz Schist, Hercules Mine.
are replaced by calcite and serpentine (?). The augites occur as colourless, fairly fresh, subhedral/euhedral laths up to 1 x 0.3 mm. in size. Of smaller size and merging to the ground are numerous, randomly oriented brown biotite flakes up to 0.3 mm. long.

Two analyses of the dykes are given in Table 28, and notable features are the high potash, low titania and high magnesia.

Very similar biotite-rich dykes are found cutting the spilites at King Island (see also Scott, 1951a) and lamprophyre dykes and sills cut Devonian sediments north of Point Hibbs, and Cambrian (?) spilitic volcanics south of Point Hibbs, near the Spero River mouth.

The King Island dykes are possibly related to the granite at Grassy, some 5 miles to the south but the nearest granite to the Point Hibbs intrusions is some 6 miles to the north and this may be of Cambrian age.

Gee (1963) has described slightly different (no olivine) lamprophyres of probable Devonian age from the Raglan Range, ten miles east of Queenstown, and even further from granite than those at Mt. Lyell.

According to Turner and Verhoogen (1960, p. 250-256)
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<td>BaO</td>
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<td><strong>Total</strong></td>
<td><strong>99.64</strong></td>
<td><strong>100.17</strong></td>
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Table 28: Lamprophyres (continued)

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<td></td>
<td>qu</td>
<td>1.26</td>
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<tr>
<td>or</td>
<td>18.35</td>
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<tr>
<td>ab</td>
<td>7.86</td>
<td>8.91</td>
</tr>
<tr>
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<td>17.25</td>
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<td>fer</td>
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<td>2.51</td>
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<td></td>
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<td>fay</td>
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<td>mag</td>
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<td>ap</td>
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</tr>
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</table>

Total 93.22 94.82

1. Whole dyke intersection (6 m.) in drill hole West Lyell no. 125.
2. Whole dyke intersection (2 m.) in drill hole West Lyell no. 124, on Queenstown-Gormanston Road.

Analyst: Japan Analytical Chemistry Research Institute.
lamprophyres are associated with some granitic intrusions and appear to be later than the granites. However, the origin of lamprophyres is uncertain and Turner and Verhoogen lean toward derivation from an alkaline olivine basalt magma.

THE ROSEBERY AND HERCULES MINES

The Hercules mine lies 5 miles south of Rosebery and its geology is similar to that at the Rosebery mine, so both are treated together in the following discussion.

The geological succession for the mines has already been described in some detail under Fragmental Volcanics and only a brief outline is given below; all rocks are in the Mt. Read Volcanics.

1. Albite-rich quartz keratophyres and agglomerates, the Massive Pyroclastics of Hall et al. (1953).

2. Grey, finely banded mudstones, tuffaceous mudstones and tuffs with minor magnesian calcite and calcite lenses - the "slates" of Hall et al.

3. Mainly fine-grained tuffs, the "Host Rock" of Hall et al.

4. Altered potassic quartz keratophyres, ash flows and agglomerates (the "Quartz Schist") merging to unaltered similar rocks of the Primrose Volcanics that also
contain lenses of finely banded mudstones or tuffs.

5. Quartz sandstones, dolomitic sandstones, volcanics, banded mudstones etc. of the Success Creek Phase.

The Success Creek rocks are most probably of aqueous origin, as also are the banded rocks of the Mt. Read Volcanics. However, the presence of pumice-rich rocks and ignimbrites (?) in the Primrose Volcanics suggests the environment was sub-aerial for at least part of the volcanic phase.

The age of the entire succession is unknown though it is suspected that the Success Creek rocks are near the base of the Cambrian. The problems of dating these rocks and of correlating them with other similar successions in West Tasmania have already been discussed.

**The Read-Rosebery Schists**

The "host rock" and the "quartz schist", are hydrothermally altered volcanic rocks known as the Read-Rosebery Schists (Hills, 1914b, 1915). The alteration is most intense at the Rosebery and Hercules mines but zones of alteration occur between the mines, and quite large areas of the Primrose Volcanics show a weak hydrothermal-like alteration.
These schists have been described as altered volcanics (Hall et al., 1953), intrusive porphyries (Dallwitz, 1946; Finucane, 1932) and sediments with volcanics (Hills, 1914b, 1915). A similar variety of origins has been postulated for the unaltered Mt. Read Volcanics. The schists are similar in form to the less altered Lyell Schists and are a cross between phyllites and phyllonites. They are largely sericitic though small bodies of chloritic schist do occur. Relics of volcanic texture (pumice fragments, shards, etc.), and the presence of volcanic quartz and K-feldspar phenocrysts all point to their being volcanic rocks.

Typical specimens of the altered tuffs which act as hosts for the orebodies are 31769 to 31772 and 32912 to 32917 from the Rosebery Mine, No. 12 level. Most of these consist of a very fine-grained, dark matrix replaced to varying degrees by sericite flakes generally less than 0.02 mm. long. In some specimens these show no particular orientation, in others many of the flakes lie along two directions at angles of 20 to 30° (centred on the average cleavage direction for the area). This double alignment is similar to that seen in some of the sericitic schists at Mt. Lyell and probably has the same origin.

Coarse-grained phases of the host rock tuff carry K-feldspar
and quartz crystals up to 0.5 mm. long, the quartz crystals being in many cases fractured. The feldspars vary from fresh to intensely altered and it is clear that in these rocks, the assemblage never reached an equilibrium condition. In 32914 the matrix is slightly coarser than usual and there appears to have been some recrystallisation and secondary growth of quartz on original crystals.

Specimen 31784A, from the base of the host rock on No. 8 level (Rosebery Mine) has several per cent of iron-rich chlorite (see analysis Table 29, No. 1) flecks in a completely recrystallised quartz rock. Chlorite in material analysed for Finucane (Table 29, No. 2), from the same level, appears to be richer in Mg. Carbonates (calcite and siderite) are present in several specimens, both in irregular patches replacing the tuff, or in veinlets.

Alteration decreases rapidly as the host rock merges to mudstones (largely siltstones) and tuffaceous siltstones of the overlying slates. In these rather more competent rocks and in the coarser tuffs, disruption and flowage along crudely formed cleavage surfaces is apparent (e.g. 32982, 31775). In 32982 the cleavage has torn up finely-banded mudstone, leaving lenticles of crumpled micaceous material wrapped around by sheaves of sericite. These crumpled relics were thought by Dallwitz (1946) to be xenoliths.
Table 29: Rosebery Schists

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<td>Tr</td>
<td></td>
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<td></td>
</tr>
</tbody>
</table>

Total | 100.11| 100.59| 100.15| 99.94| 99.88| 100.49| 101.503|
Table 29: Rosebery Schists (continued)

1. Sericite-chlorite schist at base of host rock, No. 8 level, 30 m. from mine datum, Rosebery Mine.
   
   Analyst: Japan Analytical Chemistry Research Institute.

2. Chlorite schist in host rock. Ore pass drive, No. 8 level, 300 ft. (90 m.) north of mine datum point, Rosebery Mine.

3. Quartz sericite schist (host rock), end of E cross-cut, south end of No. 7 level, Rosebery Mine.

4. Quartz schist, No. 8 level, 100 m. from mine datum, Rosebery Mine. Analyst: Japan Analytical Chemistry Research Institute.

5. Altered albite porphyry, 30 m. east of Primrose No. 2 tunnel, Rosebery Mine.

6. Altered K-feldspar porphyry, main adit, No. 8 level, 1,020 ft. (306 m.) from datum, Rosebery Mine.

7. Sheared and sericitised K-feldspar porphyry, main adit, No. 8 level, 1,100 ft. (330 m.) from datum, Rosebery Mine.

   Analyses 2, 3, 5, 6 and 7 from Finucane (1932).
within intrusive porphyry.

A chemical analysis of sericitised host rock is given in Table 29, No. 3. From the three analyses of host rock it would appear that there is considerable variation in composition at this horizon. As the composition of unaltered host rock is unknown, no estimate of the chemical changes during alteration can be made.

The Primrose Volcanics, for 30 to 50 m. below the Host Rock, are altered to siliceous schists (31784A-31786A, 32805, 32806). These are typically recrystallised (sutured boundaries) quartz mosaics, highly variable in grain size and containing relict volcanic quartz crystals up to 0.3 mm. across. There are outlines of similar-sized feldspar crystals, now replaced by sericite and quartz and in some specimens there are patches of the original feldspar. In 31786A, at No. 8 level (Table 29, No. 4) there are clear quartz mosaics apparently associated with pyrite and containing clear albite that seems to be a gangue mineral. Some of these quartz schists are broken by the cleavage to form coarse augen schists similar to those at Mt. Lyell.

Carbonates are distributed irregularly in the quartz schist and in the host rock. Some of this carbonate is finely disseminated through the schists and appears to have formed with the sericite but the major part is coarsely crystalline and closely associated with sulphides or in discrete veinlets. There may well be two
phases of secondary carbonate growth, one associated with hydrothermal alteration and one associated with mineralization. As can be seen from Table 30 these secondary carbonates are quite different from the cleaved, syngentic lenses seen in the "slate bed".

The quartz schist reveals little of its original texture to the field observer though in road cuttings at the Hercules Mine, distinct pebble-like bodies are scattered through the schist (Plate 68, No. 2). These bodies are pre-cleavage because the cleavage wraps around them, disrupting and shearing out the matrix. The latter consists of a quartz aggregate with sericite and carbonate and the "pebbles" consist of coarsely crystalline carbonate with a little interstitial quartz (32806). The rock may be an altered agglomerate.

Several tens of metres from the ore horizon the Primrose Volcanics (e.g. 32983) show only mild sericitisation, some of which may be of volcanic origin.

Compositions of altered Primrose Volcanics are given in Table 29. Numbers 5, 6 and 7 are 300 m. or so below the ore horizon. Comparison with unaltered Primrose Volcanics (Table 11) reveals no significant changes.

The sulphide minerals are generally post-cleavage and
where concentrated cause localised recrystallisation of the schist (e.g. 31784A). "Pressure shadows", similar to those in the Lyell Schists, occur around sulphide grains (particularly pyrite) and are composed of quartz and/or carbonate and chlorite (e.g. 31775). In most cases the growth is at right angles to the crystal margins but in some the pressure shadow is curved as if the crystal rotated during development of the shadow (Plate 69, No. 1). Hills (1963, p. 130) and Pabst (1931) would probably regard some of the pressure shadows as indicative of pre-cleavage pyrite but Knopf (1929) believed the pyrites described by Pabst were post-cleavage, and there is clearly considerable difficulty in interpreting age relationships from pressure shadows. However, it is possible that some of the pyrite crystals at Rosebery (and also Lyell) were pre-cleavage.

**Conditions and Timing of the Hydrothermal Alteration**

No clay minerals have been found in the Rosebery schists (despite the reference by Hall *et al.*, 1953, p. 1148, to kaolinisation) and the sericite appears from X-ray data to be similar to the dominant 2 M₁ polymorph at Mt. Lyell. Though fewer mineral indications are available at Rosebery, the conditions of alteration may have been very similar to those at Mt. Lyell. Overburden
Plate 69 No. 1: Pressure shadows of chlorite on a pyrite crystal (31775), Rosebery Mine. X125.

Plate 69 No. 2: Lens of massive pyrite-sphalerite-galena ore in Rosebery schists, looking south from the north end of the Open Cut.
pressure cannot be estimated with any accuracy but might reasonably be expected to have been no greater than 800 bars, the overburden being similar to that at Mt. Lyell. However, tectonic stresses may have been of more importance during alteration.

The alteration does not appear to have been quite as severe as at Lyell, except very locally. It is later than, or penecontemporaneous with, the cleavage and in large part preceded sulphide deposition, a sequence similar to that established at Mt. Lyell.

**Structure of the Rosebery-Hercules Area**

The surface geology of the district is shown in Fig. 3 and that of the mine areas in Fig. 11. The orebodies occur in two well-defined tuff lenses that do not visibly connect on the surface. However, the geological succession at each mine is very similar and, by interpolating structure contours for the tuff between the mines, it is evident that the lenses could well be at roughly the same horizon. It is also possible that the Rosebery section is offset by a NE dextral fault as indicated in Fig. 11.

The Rosebery tuff continues to dip east for a dip length of at least 500 m. but the Hercules tuff appears to be cut off by a steep fault (Figs. 59, 60). The Rosebery tuff cuts out north of
Fig. 59 (a): Plan of 12 level of the Rosebery mine, grid lines approximately true east, showing drill holes R1028, R918 and R844.
Fig. 59 (b): Cross-section of the Rosebery mine on grid line 500 N (Rosebery grid).

Main geological boundaries by courtesy of the Electrolytic Zinc Co.
Fig. 60: East-west cross-section of the Hercules mine, on grid line 1200 N (Hercules grid), to show Number 4 level cross-cut and drill hole H391.
Rosebery but very similar tuffs appear again in volcanics at the Pinnacles Mine (4 miles north of Rosebery), also dipping east. South of Hercules, finely banded mudstones and tuffs with N strike and steep E dip crop out sporadically.

The sharp N-S junction between the Mt. Read Volcanics and Success Creek and other sediments west of Rosebery and Hercules has already been mentioned (p. 211). Near Pinnacles and Chester the line is marked by minor crumpling of the sediments and sudden disruption of folds, and the line is almost certainly a major fault. To the west, sediments of the Crimson Creek or older formations dip east fairly consistently. At Rosebery the sediments, which are Success Creek rocks (brought up by the northerly plunge ?), dip west and Campana and King (1963) have interpreted this to mean they are overturned (Fig. 12). This is supported by the apparent gradual steepening and flattening observed in the Primrose Volcanics in traverses run from west to east towards the Rosebery Mine, and by the gradual steepening to vertical and then the gradual change to east dip, seen while traversing north from Rosebery along the railway line.

Near Hercules the contact is poorly exposed on the haulage and appears to be a fault, and small scale, N-striking folds may be seen in the Success Creek rocks in the Stitt River.
The rather meagre evidence thus indicates the presence of a major N-S fault (or series of faults) which near Rosebery takes the form of an overturned, east-facing fold. The structure, called here the Rosebery Fault Zone, is remarkably similar to the Great Lyell Fault Zone at Lyell and it also has quite likely had a complex history. At Tullah (Fig. 23) banded mudstones and tuffs dip steeply west indicating a syncline between Rosebery and Tullah. However, the Tullah "slates" may be overturned and the structure at least partly anticlinal, with Rosebery forming part of a small syncline (see Figs. 3 and 32).

The major N-S/NNE structures were largely developed during the early phases of the Tabberabberan Orogeny (TF₁) and were later crossed by smaller cleavage folds of NNW trend (TF₂). Hills (1914b, 1915) was the first to recognise two sets of folding in this area (his Alpha and Beta folds) but believed the two sets had N and W trends. Conolly and co-workers (in Hall et al., 1953, p. 1156), after mapping the Hercules mine, also concluded that the two fold groups had N and W trends though their maps indicate that they recognised the TF₂ cleavage.

Hall et al., (p. 1154) described at Hercules a 340° cleavage cutting N-striking, bedded rocks dipping east at an average 30°, though the writer's mapping (at 1:7,200) indicates considerable variation in dip. At the Rosebery mine the cleavage strikes a few
degrees east of, and dips slightly steeper than, the bedding which strikes about 340° and dips about 45° E (Fig. 59). The cleavage thus swings some 20° in strike between the Hercules and Rosebery Mines. Cleavage is found throughout the area and is parallel to the axial surfaces of folds involving Siluro-Devonian sediments. It is, however, unusual near Rosebery and Hercules in that it dips easterly whereas elsewhere on the West Coast it has a westerly dip or is vertical. The easterly dip may be due to the influence of the earlier N-S fault zone, the normal direction of elongation of the rocks being deflected by the pre-existing anisotropy. No evidence has been found to support the contention of Campana and King (1963, p. 49) that the cleavage is Cambrian.

A poorly defined lineation (Tl) with 90° pitch may be seen in places but no Tl2 lineations were observed.

Minor folds about the cleavage are admirably exposed in the main cross-cut of 12 level in the Rosebery mine and the small-scale folds indicated in the Hercules cross-section (Fig. 60) may be of similar origin. A major N-S fault appears to cut off the Hercules ore bed (Fig. 60) and determination of the nature of this feature is clearly of economic importance. Though no detailed study of this problem has been made by the writer, the occurrence
of pumiceous, fragmental volcanics like the Primrose Volcanics in drill hole H 391 (which penetrated the rocks on the east side of the fault - Fig. 60) rather points to an east side up displacement.

However, the apparent cut-off, which seldom shows any evidence of movement, may be due to folding rather than faulting. Drilling indicates considerable variation in the upper quartz schist boundary and surface mapping shows wide variation in dip, and it is suspected these features may reflect quite intense TF$_2$ folding. Thus the Hercules tuff may represent a tight synclinal core, as indicated in Fig. 60, with the Primrose Volcanics to east and west.

A few NW faults (e.g. the Bakers Creek Fault) may be of much the same age as the cleavage. Post-cleavage movements were not observed by the writer but Hall et al. (1953, p. 1152) refer to minor post-ore (and therefore post-cleavage) movements on NW faults.

Mineralization at Rosebery and Hercules

The ore mineral assemblage at both mines is similar, the principal minerals being pyrite, sphalerite and galena. Pyrite extends beyond the orebodies and much of the ore horizon contains a little pyrite.
The orebodies are well-defined and form narrow lenses up to 150 m. long and from 1.5 to 18 m. thick (Hall et al., 1953, p. 1152). In the Hercules Mine their orientation appears to be controlled mainly by the cleavage so that they dip more steeply than the bedding, and the lenses tend to have a crude echelon arrangement, both in plan and section. The mine geologists, however, believe that in detail a bedding control may be more important than the cleavage and the possibility of tight folding has already been mentioned. In much of the Rosebery mine bedding is almost parallel to cleavage and the controlling factor is not clear. Plate 69, No. 2 shows a typical lens of massive ore exposed in the Rosebery open cut and lying roughly parallel to both bedding and cleavage. Slight disturbance of normal cleavage orientation occurs on the flanks of the lens indicating there may have been some post-ore movement along the cleavage, or perhaps local volume changes during sulphide emplacement.

Much of the ore is conspicuously banded, the individual bands consisting mainly of pyrite, or sphalerite, or galena, and lying parallel to the walls of the lenses. The bands vary from thick about 0.1 mm. to 1 cm., and tend to be impersistent; they are not as uniform as the bands in the Mt. Isa ore. In 32918a, sub-parallel bands of pyrite and sphalerite show intense convolution
reminiscent of intraformational crumpling. Specimen 32918b shows crude banding of sphalerite, pyrite and galena, both galena and pyrite appearing to flow around elongate kernels of pyrite (Plate 70, No. 1). Further evidence of deformation of ore is indicated by 31859, which shows wisps of galena parallel to the cleavage (Plate 71, No. 1) and in which sphalerite kernels up to 1 x 0.5 cm. possess "pressure shadows" of galena and sphalerite (Plate 71, No. 2) and pyrite grains have pressure shadows of sphalerite. Crumpling of the banding is a feature of the ore (e.g. Plate 70, No. 2). Though curved cleavage planes are occasionally seen in galena, the ore generally shows no internal evidence of deformation (e.g. anisotropy).

The banding may be due to one or more of the following:

(a) Primary sedimentary layering, with some modification by Devonian tectonism followed by annealing.
(b) Deformation and metamorphism, probably Devonian, followed by annealing.
(c) Hydrothermal replacement of sedimentary (and folded) layering.

The presence of small folds in the ore would appear to rule out replacement of cleavage, and also the production of bands by cleavage deformation, because the cleavage is in general not folded.
Plate 70 No. 1:— Banded pyrite-sphalerite-galena ore, Rosebery Open Cut. Scale in mm.

Plate 70 No. 2:— Fold in Rosebery ore, north end of the Open Cut. Scale in mm.
Plate 71 No. 1:— Deformed (?) sphalerite-galena-pyrite ore, Hercules Mine. Kernels of sphalerite (medium grey) have shadows of galena (light grey); the small, bright grains are pyrite and the dark matrix is schist. X4.

Plate 71 No. 2:— Enlargement of part of Plate 71 No. 1. X6.
However, it is possible that the folds developed under these conditions because of the peculiar nature of the sulphide aggregate. With our present limited knowledge of the behaviour of sulphides under stress it would seem unwise to draw any firm conclusions. That deformation of the ore has occurred seems likely but how significant this is in producing the widespread banding is not known.

Mineralogy and Conditions of Deposition

Besides pyrite, sphalerite and galena, the ores contain chalcopyrite, arsenopyrite, bournonite, tetrahedrite and gold (Stillwell, 1934), and cassiterite, molybdenite, stannite, electrum, meneghinite, boulangerite, pyrargyrite and pyrrhotite (Williams, 1960).

The ores have a fairly uniform metal content and average about 20% Zn, 6% Pb and 0.5% Cu, together with some silver and gold (Stillwell, 1934; Hall et al., 1953).

Stillwell established that pyrite and arsenopyrite were the earliest minerals to form, and that the remaining major minerals were all later than pyrite.

Gangue minerals include carbonate, quartz, and barite. Many of the gangue carbonates are manganiferous (rhodocrosite) as shown in Table 30. These late carbonates are coarsely
### Table 30: Carbonates at Rosebery

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<td>Total</td>
<td>101.06</td>
<td>100.15</td>
<td>95.00</td>
<td>100.393</td>
</tr>
</tbody>
</table>
Table 30: Carbonates at Rosebery (continued)

1. R.283: Carbonate gangue, no. 4 level, X-cut opposite 15N rise Rosebery Mine.
2. R.282: Carbonate gangue, no. 8 level, 50ft. north of 13N rise Rosebery Mine.
3. 225: Massive carbonate rock, no. 8 level, 7N-8N rises, Rosebery Mine.

From Finucane (1932);
Analyst: L.H. Bath for 1, 2 and 4
G. Ley for 3.
crystalline and uncleaved, in contrast to earlier carbonates. Stillwell noted that some sphalerite grains enclosed coarse crystals of muscovite and the writer has observed rims of relatively coarsely crystalline (up to 1 mm. long) muscovite around pods of sphalerite in core from the Hercules mine (31838). This material may be recrystallised sericite of the hydrothermal alteration phase.

No attempts were made to use the pyrite geothermometer partly because of the relatively fine grain and rather scattered nature of the pyrite, and partly because the results at Lyell proved to be of doubtful value.

The lack of equilibrium relations between pyrite and sphalerite makes use of the sphalerite geothermometer of doubtful value. Specimens of sphalerite from the mines vary widely in colour and Stillwell (1934) quoted an average 4% Fe but a maximum 10% Fe. According to Sims and Barton (1961) this maximum might indicate a temperature of deposition between 500° and 725°C.

Because arsenopyrite and pyrite appear to have formed in equilibrium, the maximum temperature of mineralization may not have been above 491°C (Clark, 1959).

A magmatic origin for the ores is indicated by the S:Se ratios of 11,960 and 10,573 recorded for pyrite by Edwards and Carlos (1954).
The Age of the Rosebery and Hercules Ores

Determinations of the isotopic composition of lead from the Rosebery (2, 3) and Hercules (1) mines were made by Professor Farquhar of the University of Toronto in 1958, with the following results:

<table>
<thead>
<tr>
<th></th>
<th>204 Pb</th>
<th>206 Pb</th>
<th>207 Pb</th>
<th>208 Pb</th>
</tr>
</thead>
<tbody>
<tr>
<td>(1)</td>
<td>1.39 + 0.02</td>
<td>25.13 + 0.05</td>
<td>21.37 + 0.08</td>
<td>52.11 + 0.05</td>
</tr>
<tr>
<td>(2)</td>
<td>1.34 + 0.06</td>
<td>25.16 + 0.06</td>
<td>21.38 + 0.12</td>
<td>52.12 + 0.07</td>
</tr>
<tr>
<td>(3)</td>
<td>1.40 + 0.04</td>
<td>25.16 + 0.05</td>
<td>21.37 + 0.13</td>
<td>52.17 + 0.08</td>
</tr>
</tbody>
</table>

The variation in Pb (analytical error?) results in a range of ages from Cambrian to Tertiary. If the values should be all near 1.40 (as is possible under the given accuracy limits) then the leads are mildly anomalous and appear to be slightly enriched in uranium-lead (using the criteria of Brown, 1962). They are also then likely to be on the lower part of the age range. Such contamination could perhaps be accomplished during remobilisation of Cambrian ores during Devonian alteration. It is clear that further analyses are required.
OTHER DEPOSITS IN THE MT. READ VOLCANICS

(a) The Rosebery-Hercules Fault Zone is the site of several minor sulphide concentrations apart from the two mines already discussed. Weak sulphide mineralization occurs south of Hercules towards Whip Spur and there are several minor deposits between Rosebery and Hercules (see Fig. 11). North of Rosebery and the Pieman River is the old Chester Mine, from which pyrite was removed for use as a flux by the Mt. Lyell Mining Co. The main orebody is almost entirely pyrite with very small amounts of later chalcopyrite, galena and sphalerite; it is probably much like the South Lyell orebody at Queenstown. It is lenticular, elongated along the cleavage and occurs in hydrothermal altered quartz keratophyres. Reid (1918) reported pyrophyllite (by chemical analysis) and sericite in the alteration aureole, increasing the similarity to the Lyell mineralization.

Drilling by the Electrolytic Zinc Co. at the Pinnacles Mine (north of Chester) has shown low grade sphalerite-galena-chalcopyrite-pyrite mineralization in tuffaceous bands within altered siliceous keratophyres that strike north and dip east.

The most northerly mineralization known along this fault zone is at Silver Falls, where argentiferous galena and sphalerite occur disseminated weakly with carbonate gangue in quartz keratophyre.
(b) Fissure lode deposits occur within, and close to, the belt of west-dipping mudstones and tuffaceous sediments extending from Tullah to the Sterling Valley Mine (Figs. 3 and 23). This zone is galena-rich, with pyrite becoming abundant at the southern end. Most of the important deposits are fissure fillings and are thus quite different from the replacement ore-bodies at Rosebery and Mt. Lyell. They have not been studied in detail but a summary of the geology of the deposits has been presented (in press, Appendix A, No. 4), based on work by the writer and Brooks (1962). Several small deposits south of the Pieman River occur in, or close. to a flow of Darwin keratophyre.

The deposits lie parallel and close to the western margin of the Owen basin and faulting probably took place along this zone in Jukesian times and certainly in early Tabberabberan times. The upturned contact is much like that on Mt. Sedgwick (Fig. 38) and is called here the Mackintosh Fault.

(c) At Red Hills, minor copper mineralization is present over a considerable area in Darwin keratophyre cut by hematite-magnetite veins (Twelvetree's, 1900a). Chalcopyrite and pyrite ramify chloritised keratophyre in small, disconnected patches.

(d) At Lake Dora, a narrow N-trending line of sulphide mineralization occurs in massive potassic and sodic keratophyres
(Fig. 24). The most southern prospect of any size is almost entirely pyrite (with a little galena and chalcopyrite, 31795), the intermediate prospects are pyrite and chalcopyrite (31792 and 31794), and the northern workings are in sphalerite-chalcopyrite-galena ore (31706A). The sulphides occur in small, irregular disseminated patches or in crudely defined veins, flanked by chloritic and/or sericitic schists.

(e) The zone of mineralization at Mt. Lyell continues south for 11 miles, more or less along the axis of the West Coast Range. The deposits, which are almost entirely of copper and iron sulphides, make up what has been termed the Mt. Lyell Copper Field though all the economic deposits are near Queenstown.

Almost all of the prospects occur in or close to Darwin keratophyres, which, south of Mt. Huxley, form a narrow N-S belt terminating in the Darwin Granite (Fig. 3). This line of mineralization has been regarded as the southerly extension of the Great Lyell Fault Zone (Wade and Solomon, 1958) but in fact the amount of deformation of N-S trend in much of this zone is very small, and it appears that the distribution of the volcanics was a more important control over the mineralization. The Great Lyell Fault zone steps to the west of the line of deposits.

From south to north the more significant deposits are as
follows (see also Hills, 1914a; Bradley, 1957): Prince Darwin (chalcopyrite and pyrite in a magnetite-hematite lode within Darwin keratophyre): East Darwin (an area some 300 x 200 m. with poorly defined patches of pyrite-chalcopyrite mineralization in sericitic schists derived from alteration of keratophyric rocks); Lake Jukes (irregular, discontinuous veins and splashes of bornite and chalcocite in a prominent knob (32492-32494) composed of quartz keratophyre, partly granophyric and partly consisting of a boulder-breccia that is probably an agglomerate); Jukes Proprietary (chalcopyrite and pyrite in irregular veinlets with carbonates over a large area of Darwin-like keratophyre and agglomeratic rocks - Launceston Museum OS/33/1063, 31768A, 31769A); Great Lyell and Duke Lyell (along the South Owen Fault, workings reveal weak pyrite-chalcopyrite disseminations in sericitised and chloritised volcanics).

A SUMMARY OF THE PRINCIPAL FEATURES OF THE ORES OF THE MT. READ VOLCANICS

The orebodies of the Mt. Read Volcanics possess several characteristics:

(a) There is a common association of ores with potassic quartz-ose volcanics e.g. the Primrose Volcanics and the Rosebery-
Hercules ores; the Darwin keratophyre and mineralization south of Mt. Lyell, at Red Hills, and near Tullah; and the potassic volcanics at Lake Dora. The original nature of the sericitic schists at Lyell is not known but they may have been potassic quartz keratophyres.

(b) The formation of some orebodies was clearly controlled by the orientation of certain strata (e.g. Rosebery, Hercules and Pinnacles) but in many cases (including Mt. Lyell) stratigraphic controls at the mine scale appear to have been insignificant. For the deposits north of Queenstown a stratigraphic control may be present on a much bigger scale. Thus the Lake Dora-Rolleston and Rosebery-Hercules ores occur several hundred metres above the base of the Mt. Read Volcanics and so also may the Tullah and Red Hills deposits. The stratigraphic position of the Mt. Lyell field is uncertain but the Darwin keratophyre south of Queenstown is clearly well down in the volcanic succession.

(c) For none of the ores is there an obvious magmatic source. The nearest outcrops of Devonian granite to the Mt. Lyell ores are about 20 miles to the north-west, at Trial Harbour (the Heemskirk Granite). Even the Darwin Granite, of Cambrian age, lies 10 miles south of the economic deposits. The lamprophyre dykes at Lyell might be regarded as indicators of a granite stock by some
geologists but as has already been pointed out the connections between lamprophyres and granites is a rather tenuous one, and they may well be derived from basic magma.

A post-ore basalt dyke has been found in the Rosebery mine (Fig. 59) but this may well be of Tertiary age, and the nearest Tabberabberan granite intrusion is some 14 miles away. The Granite Tor stock, east of Tullah and some 8 miles away, is believed to be Cambrian or earlier (Solomon, in press, Appendix A, No. 1). The small body of quartz porphyry near Renison Bell (Fig. 3) and about 4 miles from Rosebery and Hercules is the only intrusion of possible genetic significance.

(d) They are mainly replacement bodies and a phase of deposition followed cleavage development and also hydrothermal alteration involving mainly chloritisation and sericitisation. However there is evidence suggesting that some of the ores may be older than the cleavage, and lead isotope values allow an age similar to the volcanics.

(e) The cleavage clearly controls the local orientation of most orebodies and structural control is very strong on a bigger scale. The coincidence of E-W and N-S structures at Lyell appears to be significant and the major structures like the Great Lyell, Rosebery and Mackintosh Fault Zones are of obvious importance.
these has probably also had a significant influence on the development of the Mt. Read Volcanic Arc and the Owen basin or rift valley.

(1) A distinctive zoning of the metal contents is observed on large and small scale. Thus the northern part of the arc contains largely Zn-Pb ores and the southern part almost entirely copper ores. A similar arrangement on a smaller scale may be seen in the Tullah-Sterling River zone, in the Lake Dora zone, and at Mt. Lyell. The suggestion by Campana et al. (1958) that the large scale zoning is centred on Mt. Bischoff (Waratah) has been criticised by the writer (1959).

THE PETROGENESIS OF THE TASMANIAN SPILITES AND KERATOPHYRES

General

The origin of spilites has been debated for well over a century and excellent summaries of earlier views have been given by Gilluly (1935), Battey (1955), Amstutz (1958), Turner and Verhoogen (1960) and Vallance (1960). On the subject of keratophyres less has been said and this is reflected in the narrower range of theories presented.
On the subject of spilites, most authors conclude by favouring one of the four main theories suggested for their origin. These involve:

(a) Existence of a primary spilite magma and primary crystallisation of the spilitic minerals (Burri and Niggh, 1945; Amstutz, 1954, 1958; Brunn, 1954; Semenenko, 1955).

(b) Alteration of basaltic lavas by solutions derived from the same magma, during or soon after eruption i.e., deuteric or hydrothermal (Dewey and Flett, 1911; Flaherty, 1934) or long after consolidation (Gilluly, 1935).

(c) Derivation from basaltic magma by assimilation and/or reaction with sea water and sediments (Beskow, 1929; Rittmann, 1962).

(d) Alteration of basalts during diagenesis or by burial metamorphism involving circulation of connate water under load (Park, 1946; Scott, 1954; Battey, 1956).

Other, now discarded, theories include derivation of spilites by surface weathering (Termier, 1898) and the all-embracing granitisation concepts of Perrin and Roubault (1941).

Several reviewers have found it impossible to favour any particular theory. After a lengthy discussion of spilites, Vallance (1960, p. 43) stated "I cannot see how any single hypothesis
can account for all the occurrences of spilitic rocks.

Turner and Verhoogen (1960, p. 270) also believed that several factors control the development of spilites (and keratophyres) including "differentiation of the parent magmas, assimilative reaction with rocks situated in the basal levels of the geosyncline, concentration of soda in late-magmatic aqueous extracts, and chemical activity induced by entrapped sea water and rising connate waters squeezed up from deeply buried sediments".

Any satisfactory theory must take into account those features that appear to be characteristic of spilites and keratophyres, viz:

(a) **Mineralogy**

The most pronounced feature is the presence of albite, either primary or secondary, and of low-temperature form (see for example, Turner, 1948, p. 124, and Turner and Verhoogen, 1960, p. 269 on spilites). In some volcanics a secondary origin seems proven (e.g. Gilluly, 1935) but in others there is conflicting evidence (e.g. Sundius, 1915, 1930 and Geijer, 1916 on the Kiruna examples) and in some the mineral appears primary (Battey, 1956 and Tasmanian spilites). Although in none is the case for a primary origin incontrovertible, many studies reaching the conclusion that the albite is secondary, appear to do so without
compelling evidence (e.g. Scott, 1954).

Many spilites contain augite, in most cases relatively unaltered, but in many chlorite is the only ferromagnesian mineral. The chlorite in many cases is primary (e.g. Amstutz, 1954). In keratophyres, a little augite and/or hornblende may be present (e.g. Gilluly, 1935; Williams, Turner and Gilbert, 1954, p. 101) but generally there are only small amounts of chlorite of apparently primary but late development (e.g. Lehmann, 1949).

Also characteristic are late stage epidote, calcite and other lime minerals, also chlorite, albite, etc. These late minerals replace feldspars, fill amygdules and veinlets, and in places replace the entire rock. Areas of intense alteration are typical.

(b) Chemical Composition

Spilites are extremely variable and diverse opinions have been held as to their chemical characteristics. A high water content appears to be the one consistent feature that separates spilites from basalts. The claims by Sundius (1930) for consistently high ferrous iron and titania and low potash, and by Wells (1923) for a low Fe$^{3+}$/Fe$^{2+}$ ratio, have proved invalid, e.g. the Tasmanian spilites with high Fe$^{2+}$/Fe$^{3+}$ ratio contrast strongly with those from Glarner Freiberge with high Fe$^{3+}$/Fe$^{2+}$ (Amstutz,
Most spilites are soda-rich but they may be closely associated with relatively low-soda rocks which have been called spilite (see Vallance, 1960, p. 32-37). This problem arises from the close field association of sodic and non-sodic rocks (e.g. King Island) and the writer has suggested earlier in this thesis that such an assemblage should be regarded as spilitic because the variation in composition is probably of genetic significance.

Tane (1962) and Michel et al. (1960) have described potash-rich spilites from Pelvoux.

Identification of the parent magma is of course highly controversial and opinions cover the range of primary spilite, and alkaline, tholeiitic, calc-alkaline basalts.

Apart from wide variations in type of alkali on both large and small scales, and high water content of intermediate types, no particular chemical features distinguish the keratophyric group. The quartz keratophyres, for instance, are much like alkali rhyolites.

It appears to be widely accepted that contiguous spilites and keratophyres are genetically related and in many cases that keratophyres are differentiates from spilites. However, derivation from magmas independent of the spilite parent have been
suggested and these magmas are generally regarded as acidic.

(c) **Tectonic Environment**

The majority of spilites and keratophyres form in marine conditions and generally in active geosynclines; they are rare in shield areas. In general, spilites occur in the earlier and middle phases of a geosyncline's history but the time range is considerable and there may well be several phases of spilitic activity during the active life of a geosyncline e.g. in Eastern Australia's Tasman Geosyncline, the ophiolitic suite appears in the Cambrian in Tasmania and again in the Siluro-Devonian of New South Wales. Vuagnat (1954) suggested there may be two phases for the Alpine chain. In Cornwall (Dewey, 1948) spilites do not appear until the Middle Devonian, though sedimentation was continuous from the Lower Devonian.

In many areas, effusion of spilitic or keratophyric magma appears to accompany the initiation of new basins of sedimentation. For example, Rippel (1953) has shown that the Lower Devonian geosynclinal sedimentation in Sauerland (Germany) began with the development of relatively small troughs which were rapidly filled by the products of localised keratophyric eruptions. Succeeding sediments and basic and keratophyric volcanics spread beyond the limits of these troughs throughout the geosynclinal basin.
In all probability, the Mt. Read Volcanics represent the infilling of a distinct longitudinal trough marginal to the Tyennan Geanticline and the centres of spilitic eruption may represent smaller troughs.

(d) **Associated Igneous Rocks**

Spilites clearly occur in two igneous rock suites - the ophiolitic and non-ophiolitic (see Vuagnat, 1949). The first, frequently taken as the general type, involves ultra-basics (particularly serpentinites) and gabbros, the former as thick sheets and the latter as small plugs. The areal distribution of the three phases tends to be similar in any one ophiolitic province. Ophiolitic suites are typical of several Tertiary mountain chains, e.g. the Alps, New Zealand and Indonesia. They occur in the Tasman Geosynclinal Zone in central New South Wales (the Great Serpentine Belt) and in western Tasmania. Small (e.g. New South Wales) or large (e.g. Tasmania, New Zealand) amounts of keratophyre may be present. Time relations between members of the ophiolite suite are variable though possibly spilites generally precede serpentinites (Turner and Verhoogen, 1960, p. 258). In the Alps, Vuagnat (1954) suggested serpentinites are contemporaneous or later than diabases but Brunn (1954) believed the serpentinites were the first to form. In California, serpentinites and spilites
were roughly contemporaneous (Turner and Verhoogen, 1960, p. 258).

The non-ophiolitic association comprises spilites and keratophyres with basalts, andesites, etc. and related intrusives; keratophyric rocks are important and serpentinites rare. Generally, the quartz keratophyres are far more abundant than the keratophyres and in some areas the intermediate types appear to be absent (e.g. Davies, 1959, on Cader Idris, Wales). The proportion of spilites varies considerably and the igneous rocks are localised. The Ordovician phase of the Welsh Caledonian geosyncline seems typical (Williams, 1927; Thomas and Thomas, 1956; Smith and George, 1948; Pringle and George, 1948);/Oregon (Gilluly, 1935); the Tertiary of the Olympic Peninsula, Washington (Park, 1946; Waters, 1955); the Devonian and Permo-Trias of Shasta Co. California (Kinkel et al., 1956; Albers and Robertson, 1961) and the Lahn area, Germany (Gotz, 1939; Lehmann, 1949, and other papers).

A feature of many spilite and keratophyre occurrences is the intrusion of albitic and acidic plugs into the volcanics but at a much later stage than the volcanicity. Gilluly (1935) described albite granites and diorites from eastern Oregon that are probably genetically related to spilites and keratophyres, from which the
intrusives are separated by an orogenic period. Similar rocks intrude spilites in northern Norway (Gjelsvik, 1958) and in Scotland (Bloxam, 1960).

In some spilite provinces, basic volcanism was preceded by acidic eruption, e.g. in Anglesey (Smith and George, 1948), but in most areas where a sequence can be established, the reverse relationship is true. Amstutz (1954) was very tentative about his suggestion that the acidic rocks are younger than the spilites. However, Harrington and Hay (1956) believed acidic effusives follow basic in the Mt. Camel area of New Zealand and Kinkel et al. (1956) seem sure that the Balalakla rhyolite (quartz keratophyre ?) follows spilites, andesites and keratophyres. There also appears to be a basic to acid sequence in the Kiruna effusives (Geijer, 1916) and in the volcanics of the Blyava region of the Urals (Zavaritsky, 1960).

(e) Associated Sedimentary Rocks

The spilites and keratophyres of the ophiolite association are commonly emplaced in greywackes, poorly sorted mudstones and cherts typical of synorogenic sediments in eugeosynclinal troughs. Limestones are absent or rare in some cases but are common on the Olympic Peninsula (Park, 1946) and occur with spilites on Anglesey (Smith and George, 1948) and in New South Wales near Tamworth (Benson, 1913) and Woodstock (Stevens, 1952).
Bailey (1936) referred to the association of ophiolites and limestones. Cherts (generally radiolarian) occur with spilites and serpentinites, an association made famous by Steinmann.

Non-ophiolite spilites occur in successions of limestone (e.g. the Matlock area, Derbyshire) and/or sandstones, and Vuagnat (1949) described continental or epicontinental sediments with the Aar and Pelvoux spilites. Amstutz (1954) described red sediments and arkoses in the Verrucano of Glarner Freiberg though his arkoses may be tuffaceous sandstones.

Most spilites and keratophyres developed in an aqueous and mainly marine environment though some may be of terrestrial origin.

**The Tasmanian Spilites and Keratophyres**

Viewing the Tasmanian volcanics in the light of these generalisations reveals few anomalous features. Mineralogically they are rather more altered than many assemblages and belong to the group in which both chlorite and albite appear to be primary. Chemically they are slightly unusual with the high alumina and low titania contents of some of the spilites and the wide variation in composition of the quartz keratophyres. The spilites are part of an ophiolite association associated with the early stages
of geosynclinal, greywacke sedimentation and the large volume of associated keratophyres completes a close comparison with most of the New Zealand spilite-keratophyre occurrences. The associated granitic intrusives appear to be hypabyssal or plutonic equivalents of the quartz keratophyres. Of the associated sediments perhaps the only unusual features are the lack of manganiferous sediments.

Scott (1954), Bradley (1957) and Spry (1962b) are the only workers who have given possible detailed discussions on the origins of the Cambrian spilites and keratophyres. Scott believed they were derived from basalts by the action of connate waters activated during the Jukesian Orogeny and she also hinted that some keratophyres might be metasomatised sediments. Differentiation from primary and probably independent magmas (olivine basalt for spilites, acid or intermediate for keratophyres) was favoured by Spry. Bradley believed that spilites, pyroclastics and greywackes were metasomatised during the Devonian (Tabberabberan) Orogeny to produce keratophyric rocks.

The reasons for believing the Tasmanian acid and intermediate rocks to be volcanic have already been given.
Burial and Regional Metamorphism

Scott's suggestions are similar to those made by several other recent workers on spilite-keratophyre rocks (e.g. Battey, 1956; Vallance, 1960) who have emphasised the possible importance of the effects of burial metamorphism. Scott apparently envisaged a combination of the effects of burial and those of heating and crustal disturbance associated with an orogenic phase but without introduction of juvenile material. She attributed to this phase the development of albite and quartz, the latter forming the quartz spherulite rock, jasper veins etc. in the spilites and quartz phenocrysts in the quartz keratophyres. The quartz spherulite rock has already been discussed in this thesis, with the conclusion that it is of volcanic origin. Siliceous amygdules, clots and veinlets are common in some individual flows on King Island (e.g. one-mile north of City of Melbourne Bay) and jasper clots and veins are associated with epidote veinlets in massive spilites near Penguin. All these features are confined to lava flows and are by no means of general occurrence; they are believed to be typical associates of the Cambrian volcanism. The quartz phenocrysts in the keratophyres are also believed to be primary and it is suggested that the soda and silica metasomatism described by Scott (1954, p. 147) "along a major geanticlinal
structure" actually refers to the distribution of keratophyric lavas along the Mt. Read Volcanic Arc.

Vallance (1960, p. 43) has rightly pointed out that sediments associated with metamorphosed volcanics should also reveal any signs of metamorphism seen in the volcanic rocks. The grey-wackes of Tasmania possess fragments of spilites and detrital crystals of albite along with chlorite and iron ores but these components could be either due to diagnostic replacement or derived from pre-existing volcanics; they are not therefore of much assistance to the problem. The presence of detrital microcline with albite in greywackes or tuffaceous sandstones near Waratah is much stronger evidence for a pre-depositional origin of the feldspars.

In general, a lengthy period of burial metamorphism would surely lead to homogeneous assemblages, especially in the deeper parts of the sedimentary pile but Battey (1955) attributed marked variations in alkali content within a single flow to recrystallisation and limited mobility of certain elements under deep burial. Dickinson (1962) also attributed local potash and general soda metasomatism in a tuff to action of connate waters under heavy overburden; in both these cases the interbedded sediments are not affected. Why these features cannot be magmatic or deuteric is not at all clear. Vallance (1960, p. 43) has an interesting problem
in the dominance of calcic feldspars in the sediments associated with the albite spilites in New South Wales. An analogous problem in Tasmania is the presence of gabbros with calcic feldspar along with albite gabbros, and the presence of calcic feldspar in some andesites. Vallance has no explanation for the New South Wales problem but comments that the variation in feldspar does not support the burial metamorphism theory.

In summary, it is suggested that burial effects are not the sole causes of the albite in the Tasmanian spilites and keratophyres, nor of the siliceous bodies in the spilites or the quartz phenocrysts of the quartz keratophyres. However, burial of the spilites may have contributed to the growth of albite and particularly chlorite, especially where augite and hornblende were present in the rocks; the partial alteration to chlorite of these ferromagnesian minerals might well be due to burial and/or Jukesian metamorphism.

Parental Magma

Referring to Spry's suggestions, there seems little doubt that the spilites are derived from an olivine basalt magma but it has already been shown that they vary in type and exhibit tendencies associated with alkaline, tholeiitic and calc-alkaline types. It is tempting to suggest that the picrite basalt is the primary magma
yet by analogy with Hawaiian lavas this might represent the accumulation of olivines from a tholeiitic magma of composition near A in Fig. 20, from which the remaining types have developed by feldspar accumulation, iron enrichment, etc. Concentration of feldspars, perhaps in combination with assimilation, may have tended to develop the alumina-rich lavas and particularly the porphyritic types. These are more like the basic members of the calc-alkaline assemblages such as those of Lassen Peak, California.

The small amounts of keratophyre and quartz keratophyre associated with the spilites (e.g. at Waratah and also the Wanderer River) could be due to a tholeiitic type of differentiation, in which siliceous, alkali-rich residues develop in small quantities (e.g. McDougall, 1962, on the Tasmanian dolerite magma). However, when the dominant role of the Cambrian keratophyric rocks in Tasmania is considered, it is difficult to see how these rocks could differentiate from tholeiites. The spilite: keratophyre ratio, estimated from field occurrence and structural interpretations, is in the order of 1:10 (the method of measuring the areas of outcrop for comparison, used by Ronov (1946), is considered a little inaccurate in a structurally complex area). In fact the whole series takes on similarities to the calc-alkaline, basalt-andesite-
rhyolite association, but with the rhyolitic rather than andesitic types dominant. The genesis of this association is obscure but a scheme involving differential fusion of the crust to produce contiguous but quite different magmas appears likely. Spry (1962b) suggested the possibility of independent sources for the Tasmanian spilite and keratophyric magmas and further support for this is found in the areal separation of keratophyric types (the Mt. Read Volcanics) and spilitic types (Fig. 1).

The quartz keratophyres of northern New Zealand were assumed by Battey (1956) to be derived by differentiation of a tholeiitic magma but he took no account of the relative quantities of quartz keratophyres and spilites, nor of the lack of intermediate members of the series. His field maps (Battey, 1950, 1951) indicate that spilites form some 60% of the total lavas on Great Island (New Zealand) but that spilites are subordinate to quartz keratophyres on the Rangiawhia Peninsula. Harrington and Hay (1956) also found that keratophyres are much more abundant than the spilites on Mount Camel (New Zealand).

The Tasmanian association has some similarities with that in the Ordovician of the Cader Idris area (Davies, 1959) in which spilites and quartz keratophyres predominate and to that in the Verrucano of Glarner Freiberg (Amstutz, 1954) in which the acid
and basic types are again the most important, with intermediate
types rare. Amstutz showed differentiation curves from basic
to acid members but considered the possibility of there being two
differentiation series - a spilitic and a granitic.

In previous discussion it was shown that the spilites showed
various basaltic magma tendencies but that the andesite and
keratophyres are essentially calc-alkaline. Adding the available
analyses of tuffs and related intrusive rocks confirms these
generalisations (see Figs. 15-20). In several of the variation
diagrams a tendency for the spilite lines to merge into the kerato-
phyre lines is evident but there is still a strong impression that
two distinct trends of differentiation are present and that differentia-
tion of the whole suite from a basaltic magma seems unlikely.
Derivation from an andesitic magma (differentiating in basic and
acidic directions) is possible but the two major groups may develop
from independent but closely associated magmas.

Discussion

Other workers on spilites and keratophyre have concluded
that their rocks show general tendencies similar to certain magma
types but with departures from the normal trends that demand
peculiar conditions of crystallisation, assimilation, metamorphism,
etc., involving the presence of large quantities of water and low temperature.

The theory of spilitization by the action of sea water on basaltic lava has been revived recently by Dietz (1963) and Rittmann (1962). Though there are few incontrovertible examples of terrestrial spilite, there are several cases of marine basalts that are not spilitic (see Amstutz, 1958). These are sufficient to cast doubt on the theories of Rittmann and Dietz.

It is possible that reaction of lava with sea water could produce part of the spilite-keratophyre assemblage, particularly chlorite, sericite and albite (Yoder and Tilley, 1962, p. 468) but it is doubtful whether sea water could permeate the flows sufficiently to be the main cause of the mineral assemblage.

Deuteric, post-extrusion alteration may be an important factor in developing the spilite and keratophyre mineralogy and it has already been suggested that the tuffs of King Island have been affected by such alteration. However the augitic pillow lavas show little or no sign of deuteric activity and it is difficult to see how the large pillows on King Island could be significantly altered by late-stage fluids.

Hydrothermal activity below the surface in the later stages of vulcanism may give rise to spilitic and keratophytic mineral
assemblages, as shown on p. 383, and some of the silicification and K-feldspar development in the Tasmanian rocks might be of this origin. However, there should be ample evidence of replacement of earlier assemblages if such action was solely responsible for spilites and keratophyres and such evidence is lacking in the majority of the Tasmanian rocks.

Alteration of basalts to spilites by deuteritic activity has been postulated by many authors, e.g. Dewey and Flett, 1911; Bailey and Grabham, 1909; Geijer, 1916; Eskola, 1925. Most of these writers have implied that the volatiles are residues of differentiation or are components of a primary, volatile-rich magma, and the problems of genesis thus revert to the magmatic history. Pre-extrusion processes fall into two groups: (a) those in which "normal" magmas are altered during their passage to the surface and (b) those in which hydrous spilitic or keratophyric magmas are of primary origin. As both of these seem to offer reasonable possibilities they are discussed in some detail below.

(a) Basaltic magma probably originates somewhere near 50 to 60 km below the surface (Kushiro and Kuno, 1963; Yoder and Tilley, 1962, p. 520; Eaton and Murata, 1960) but the depth will vary according to local conditions. Direct evidence comes from Eaton and Murata (1960) who derived a depth of 45-60 km.
for Hawaiian magma from seismological evidence and on calculations of the necessary height of the lava column. The Hawaiian lavas pause "a few kilometres" below the surface and erupt some weeks or months later when "sufficient pressure has been built up". Probably this shallow ephemeral reservoir forms a marked change in the physical state of the rocks. Extrapolating this picture to the Cambrian of Tasmania, the magma may well have paused in its rise to the ocean near the base of the Success Creek phase, before proceeding slowly to the ocean floor. Possibly at the pause, some limited settling of olivine crystals from basaltic magma may have occurred to produce picritic basalts. It is suggested that during the pause and thereafter the magma conduit absorbs large quantities of gases (largely steam) either from surrounding material or from juvenile sources, or both. It is possible for steam, etc. to be derived from the wall rocks only if it is unable to escape by any other passage. If connate waters are heated a few hundred degrees Centigrade the water vapour pressure will increase to several thousand atmospheres (log \( P = \frac{-2.3}{T} + \text{Const} \)) and steam may then be forced into the magma column along with elements dissolved from the wall rocks, particularly potash and soda.

The magma may thus stew for a time in these gases, the period varying considerably and probably being one of the major
factors in controlling the mineral assemblages developed later. The volatiles tend to concentrate in the upper parts of the magma chamber because these will be relatively cool and the vapour there will exert less pressure per unit volume. As the volatile content increases pressure is built up and finally the magma is forced to the surface, with release of a large percentage of the volatiles. The first eruptions are likely to be explosive and volatile-rich (see also Kennedy, 1955) and it is interesting to note that coarse agglomerates occur at the base of the volcanic succession on King Island and that the Primrose Volcanics, which are largely fragmental and believed to have been highly gaseous in eruption, occur at the base of the Mt. Read Volcanics.

The spilitic and keratophyric magmas were probably somewhat hydrous and the presence of water in the magmas must profoundly affect the ensuing crystallisation. If crystallisation had commenced in the "dry" state then the early minerals will become unstable. Thus if olivine had commenced to form and segregate gravitationally in a basalt melt, it would be replaced by a hydrous ferromagnesian mineral; such a history may explain the olivine pseudomorphs in the picritic spilites.

Ringwood (1959) suggests that the presence of $\text{OH}^{-1}$ ions in basaltic melts will lower the temperature of crystallisation
(particularly feldspar) and cause preferential formation of pyroxene and amphibole over anorthite. CaO will enter pyroxene and when feldspar crystallisation commences, albite-rich plagioclase will form. Thus in the diopside-anorthite system, the diopside field is widened and the anorthite field reduced. Eutectic crystallisation takes place at a reduced temperature, with formation of sodic feldspar from the Al$_2$O$_3$-rich residuum. Osborne (1959) also believes high water pressure may be a factor in delaying crystallisation of plagioclase.

Unless the water pressure falls off rapidly the diopside would convert to amphibole or biotite. Experiments by Yoder and Tilley (1956, 1962) confirm that at high water pressures (plus 3-4000 bars) amphibolite is the stable assemblage, hence the water pressure (and content) of the Tasmanian diopside lavas cannot have been very high. That water pressure was variable is indicated by the presence of both hornblende- and diopside-gabbros and both diopside- and chlorite-spilites. The lack of hornblende in spilites is a surprising feature; if the chlorite forms because of high water pressure, then intermediate phases with amphibole are to be expected.

In the keratophyres, the presence of OH$^{-1}$ ions produces Ca-rich ferromagnesian minerals rather than Na-rich minerals,
thus directing the path of crystallisation towards albite and preventing development of riebeckite, aegirine, etc.

High oxygen pressure in the magma is likely to cause early crystallisation of magnetite (Yoder and Tilley, 1962, p. 384) and in most Tasmanian rocks, magnetite appears to be early.

It is possible, then, to explain the presence of albite and diopside, and early magnetite, in terms of relatively high water content. An anomalous feature is the presence of chlorite as a late mineral that in many rocks does not pseudomorph amphibole or pyroxene and appears to be primary. Apparently the high water content (at fairly high pressures?) of some magmas or parts of the magmas delays crystallisation of ferromagnesian minerals to produce the unusual sequence: magnetite-albite-chlorite.

If Ca is taken into diopside in the early crystallisation stages it is difficult to understand why there should be such an abundance in late-stage volatiles. It is possible that albite developed by replacement of a more calcic plagioclase, thus releasing Ca into the residual melt, and several authors have suggested such a process to explain the abundance of calcite, epidote, etc. That such reactions are possible has been illustrated by the famous experiments of Eskola et al. (1937) in which calcic plagioclase was "heated" in sodium carbonate to temperatures between
230° and 460° C, to produce albite and calcium carbonate.

However if crystallisation followed the late-chlorite sequence (suppressing anorthite and diopside) then calcium might well be freely available in the later phases of crystallisation and be circulated freely in residual fluids.

(b) Another possible theory that is allied to the previous discussion, is one involving the primary development of spilitic and keratophyric magmas. Amstutz (e.g. 1958), in particular, has favoured such an origin for many years. One of the characteristics of the spilite-keratophyre association is its occurrence in geosynclines. The development of such basins involves sagging, stretching and probably thinning of the crust beneath the basin. Secondary effects are presumably a steeper temperature gradient, and the melting of basal crustal material and the mantle at relatively low pressures (in the initial geosynclinal phases at least). Is it possible that under these conditions relatively hydrous magmas can be produced which find ready access to the surface because of the tensional regime of the geosyncline? This might explain the production of serpentinites as well as spilite and keratophyre magmas.

Some "primary" process such as this would explain more satisfactorily those occurrences (such as Anglesey) in which spilites and keratophyric lavas were extruded directly onto crystalline
basement (in which the opportunities for late-stage hydration would be small). The ensuing crystallisation of the primary magma would follow similar lines to those just outlined under (a).

To conclude, it is suggested that the Tasmanian spilites and keratophyres may be derived from independent magmas of spilitic and keratophyric composition and that they developed that composition on initial melting or during passage to the surface. Hydrothermal activity probably resulted in some alteration at the surface or after burial.

THE RELATIONSHIP BETWEEN SPILITES, KERATOPHYRES AND ORE DEPOSITS

Ores of the Rosebery and Lyell types occur in many geosynclines in association with volcanics.

Along scores of kilometres of the Urals there are areas of pyrite-chalcopyrite mineralization confined to zones of Palaeozoic greenstones (spilites) and keratophyres (Zavaritsky, 1943a, b and other papers; Kurshakova, 1958; Radkevich, 1961; Ivanov, 1962). In the northern Urals folding has been intense and the ores are post-cleavage and hydrothermal alteration (chloritisation and sericitisation), but major structural controls are not obvious. In the southern Urals, folding is gentle and alteration slight.
Zavaritsky favoured a syngenetic origin for the pyrite but Radkevich and Ivanov looked to deep-seated solutions of magmatic origin (related to the volcanics) to explain the alteration and the sulphides.

Pyrite and magnetite lenses characterise spilite-keratophyre assemblages in the Caledonian geosynclinal rocks of Norway and Oftedahl (1958) has interpreted many of the acidic rocks as ignimbrites and the ores as of sedimentary-exhalative origin. Vokes (1962) recognised two types of ore, the pyritic (with pyrite, sphalerite and in some cases galena) and the pyrrhotitic (with chalcopyrite and sphalerite), corresponding to the two types recognised by Stanton (1960) as typical of Palaeozoic volcanic geosynclines. The pyritic types appear to have formed at much the same temperature as the enclosing metamorphics (using the sphalerite geothermometer) and Vokes believes they were recrystallised during folding. Some of the ores are banded like the Rosebery ores.

Oftedahl has suggested a sedimentary-exhalative origin for the Rio Tinto pyrite-chalcopyrite ores in Spain, which for many years were regarded as related to intrusive porphyries (Helm, 1935; Williams, 1934). In a new interpretation Williams (1962) has described the succession as follows:
Purple and dark grey slates
Acid pyroclastics (tuffs and breccias)
Rhyolite (porphyry)

The sulphides lie within the pyroclastic zone over a wide area and the ores contain pyrite, arsenopyrite, chalcopyrite, sphalerite, galena and tetrahedrite. Because they replace the cleaved sediments and volcanics both Oftedahl (1958) and Williams (1962) favoured remobilisation due either to hypogene solutions or to folding. Kinkel (1962) likewise believed the ores were sedimentary-exhalative and somewhat remobilised by intrusion of the granite that lies north of Huelva, and by stresses set up during folding. Chloritisation and sericitisation have occurred near the massive sulphide lenses but the alteration is apparently not particularly extensive. Banding may be seen in some orebodies due to alternations of chalcopyrite and sphalerite (Williams, 1934, p. 613). Though no modern petrographic descriptions of the volcanics are available to the writer some of them appear from summary descriptions by Williams (1934, p. 601) to be quartz keratophyres, both potassic (see analysis on p. 601) and sodic.

Shasta County, California is another area in which keratophyric rocks act as hosts to sulphide ores (Kinkel et al., 1956;
Pyrite-chalcopyrite-sphalerite lenses occur in the Bully Hill and Balaklala rhyolites, of Triassic and Devonian age respectively. These rhyolites are quartz keratophyres interbedded with spilites, pyroclastics and shales, and are altered over wide areas to quartz-sericite schists. The two favoured origins for the ore are deposition from solutions of magmatic origin, and deposition from solutions generated in the geosynclinal column during orogenesis.

Another area that has become of particular economic importance in recent years forms the northern end of the Appalachian Geosyncline extending NNE through Newfoundland, Nova Scotia and New Brunswick in Eastern Canada and south into Vermont and New Hampshire. In this area Silurian and Ordovician sediments were folded in the Devonian and intruded by gabbroic and granitic stocks. Stanton (1961) has shown a zone of volcanics - the New Brunswick Geanticline - passing along the axis of this geosyncline. In Newfoundland, replacement pods and lenses of pyrite-chalcopyrite and sphalerite-galena-chalcopyrite-pyrite ores occur in Ordovician basic pillow lavas and pyroclastics that have undergone chloritisation, sericitisation and silicification (Baird, 1960; Swanson and Brown, 1962; Anger, 1963; Williams, 1963). The pyrite-chalcopyrite ores are mainly in weakly metamorphosed,
hydrothermally altered basic and intermediate volcanics and breccias, intruded by granitic stocks. Anger (1963) has suggested, mainly by comparison with Rammelsberg, that the ores are volcanic-exhalative, and deformed and modified by later folding and granitic intrusions.

In Nova Scotia, pyrite-sphalerite-chalcopyrite orebodies occur in pre-Carboniferous sediments and volcanics (Keating, 1960).

In the Bathurst-Newcastle district of New Brunswick, Ordovician sandstones and greywackes are interbedded with spilites and greenstones and acid lavas and pyroclastics, the latter in many areas converted to sericitic and chloritic schists (Skinner, 1956; Smith and Skinner, 1958; Lea and Rancourt, 1958; McAllister, 1960).

There is some doubt as to the origin of certain altered quartz porphyries that appear to be much like the sheared quartz keratophyres of west Tasmania and many of the acid volcanics and intrusives described by Skinner (1956) have keratophyric tendencies.

The ores occur in volcanics and sediments and include the pyritic and pyrrhotitic types. Sulphur isotope ratios (DeChow, 1960; Tupper, 1960) indicate a magmatic origin for the ores, though DeChow suggests they may be remobilised Ordovician deposits.

In Australia detailed information on the many volcanics
associated with copper-lead-zinc ores is surprisingly rare, even for such important orebodies as those at Mt. Isa. Underlying the Mt. Isa "shale", which contains conformable lead-zinc and slightly discordant chalcopyrite ores (Carter, 1953; Murray, 1961 and others), are the Eastern Creek Volcanics which are metamorphosed (chlorite to biotite grade) basalt (Joplin, 1955). These rocks have been examined by the writer to the north of the Mt. Isa leases (in the Spring Creek area) where they are mainly carbonate and epidote-rich albite basalts (spilites ?). Acid tuffs occur in the Mt. Isa "shale" (Dr. N. J. W. Croxford - pers. comm.) and chloritised spilitic (?) volcanics (the Western Volcanics - Dr. P. Solomon, pers. comm.) overlie the shale. Unfortunately, no detailed work on these volcanics has yet been published.

Similarly, near Bathurst, New South Wales there are about forty orebodies that occur primarily in, or close to, two volcanic horizons (Stanton, 1955). The volcanics are described as andesitic but at both horizons there are schistose quartz-feldspar-porphyries which are in part fragmental. Though Stanton regarded these as of metasomatic origin the descriptions suggest that they are volcanic. They are clearly very similar to schistose rocks near Bathurst, New Brunswick and to some of the Mt. Read Volcanics. Though granites intrude the area, Stanton believed the ores to be volcanic-exhalative.
Similarly at the Lake George mine, near Captains Flat in New South Wales, pyrite-sphalerite-galena-chalcopyrite ores occur within folded Ordovician volcanics varying from dacite to rhyolite (Mine Staff, 1953; Edwards and Baker, 1953). However, the host rocks are considerably altered to sericitic and chloritic schists and no petrological details are given. The mine staff believed the ores to be related to granitic intrusions cropping out five miles south of the mine.

In all these Australian examples there is no clearcut relation to granites or other intrusive bodies, though in each case post-tectonic granitic intrusions are present in the district. There seems to be little doubt that at least the bedded, lead-zinc ores at Mt. Isa are syngenetic (Dr. P. Solomon, pers. comm.) but the origins of the other Australian ores are uncertain.

These ores, and also the other examples from many parts of the world, have certain common features. They are closely associated in space with volcanic rocks, which in the majority of cases cited, are spilitic or keratophyric. The host rocks are generally typical of eugeosynclines and have been folded and metamorphosed to varying but generally slight degrees. Within the vicinity there are generally post-tectonic granitic stocks and the ores are also at least partly post-tectonic, yet not obviously
related to the intrusives. Sodic granites occur in several areas.

In general terms (with Mt. Isa an exception) three types of ore have been found: (a) massive pyrite-chalcopyrite ores, (b) massive or banded "pyritic" ore (pyrite-sphalerite-galena-chalcopyrite) and, (c) massive "pyrrhotitic" ore (chalcopyrite-pyrrhotite).

Stanton (1960) has recognised a common association of pyritic and pyrrhotitic ores with volcanics of generally andesitic to basaltic composition but he believed the associated acid rocks were of metasomatic origin (much as Bradley and Scott believed for the Tasmanian acid rocks), and he did not consider the common relationship between pyrite-chalcopyrite ore and volcanics.

Oftedahl (1958) was inclined to relate all these ore types to welded tuffs but his views have been strongly criticised (Marmo, 1958; Kautsky, 1958; Kullerud et al., 1959). In a reply to discussion by Marmo, Oftedahl (1959) accepted the common association of soda-rich volcanics and ores but believed the soda enrichment to be due to reaction with sea water, and therefore irrelevant.

Amstutz has in several papers (e.g. 1958) affirmed the common association of spilitic rocks and mineral deposits.

Other types of ore occur with spilitic and keratophyric rocks, particularly stratiform hematite-magnetite and manganese deposits such as in the Lahn area of Germany and in Sweden and Norway.
Conversely, the three ore types under discussion occur with volcanics other than spilites and keratophyres and it is difficult to assess the importance of the spilitic rocks relative to other volcanics because of the lack of detailed studies in many areas. However, it is clear that the ore types of Rosebery, Hercules and Mt. Lyell are commonly found in many parts of the world in host rocks similar to the Mt. Read Volcanics and in similar geological environments, so that the problem of the origin of the ores appears to be a fairly general one.

**THE GENESIS OF ORES BY VOLCANIC PROCESSES**

Theories presented for the types of deposits under discussion include:

(a) Exhalative-sedimentary

(b) Hydrothermal and derived from sub-volcanic magma chambers soon after volcanism

(c) Hydrothermal and derived from "independent" and much younger granitic intrusives.

Stanton in particular has developed a scheme of sulphide sedimentation during volcanism "in off-reef facies (or equivalent, where there are no reefs) in seaboard volcanic areas, about volcanic islands or along seaboard volcanic zones of arc structures" (1958, p. 485). His optimum conditions for sulphide build-up are in the
offshore troughs of typical volcanic island arcs. He imagines large supplies of sulphate being delivered by volcanic emanations into a reducing environment, the metal content being syngenetic (and volcanic) or derived from outside sources during "diagenesis, metamorphism, or hydrothermal activity" (1958, p. 485). Oftedahl (1958) however, derives all the metals directly from volcanic exhalations.

The main criticism of Oftedahl's theory (and partly of Stanton's) concerns the likelihood of volcanic exhalations ever being able to supply sufficient metals to form orebodies.

Most fumarole gases and geyser waters carry metals as well as a great variety of elements (see Bateman, 1956, pp. 62 and 63) but information gathered by Zies (1929), White and Waring (1963) and others shows that the amounts of low-volatile compounds are extremely small. White and Waring (p. K24) quote 0.5 pp.m. Zn, 0.03 ppm Cu and 0.03 ppm Pb in the fumarolic gases of Showa-shinzan, Japan.

The transport of significant quantities of the base metals is probably only possible at considerable depths under higher temperatures and pressures. In this regard, the 5,232 ft. hole drilled in a geothermal area in Southern California is of particular interest (White, Anderson and Grubbs, 1963). This hole penetrated
altered shale and siltstone and tapped a hydrothermal brine believed to be at least partly juvenile; related rhyolite domes occur in the area. The water carries approximately 5.4% Na, 2.38% K, 0.032% Li, 4% Ca, and 18.4% total halides. The gases contain high percentages of methane and CO₂, and H₂S content is from zero to 0.35%. The evaporated residue (33.2%) contains, in parts per million, 3,000 Fe, 1,000 Mn, 2 Ag, 500 B, 200 Ba, 0.5 Cr, 20 Cu, 2 Ni, 100 Pb, 2,000 Sr, 500 Zn and many other elements. Encrustation on drainage pipes carrying the brine reach as high as 20% Cu. The mineral assemblages in the altered sediments are of particular interest and are recorded by White et al. in their Table 2, reproduced below:

<table>
<thead>
<tr>
<th>Depth in feet</th>
<th>Minerals, in order of decreasing abundance:</th>
</tr>
</thead>
<tbody>
<tr>
<td>4,477</td>
<td>Chlorite, K-feldspar, K-mica, quartz and albite; epidote and pyrite in veinlets.</td>
</tr>
<tr>
<td>4,484</td>
<td>K-feldspar, quartz, chlorite, K-mica, albite; epidote and pyrite in veinlets.</td>
</tr>
<tr>
<td>4,662</td>
<td>Quartz, K-feldspar, chlorite, albite, pyrite; veinlets of quartz and pyrite.</td>
</tr>
<tr>
<td>4,917</td>
<td>Epidote, quartz, feldspar (?), pyrite.</td>
</tr>
<tr>
<td>4,923</td>
<td>Quartz, chlorite, albite, epidote, pyrite.</td>
</tr>
</tbody>
</table>

These assemblages indicate how spilitic and keratophyric rocks
might be produced in parts of a volcanic pile during late hydrothermal activity. The high metal content of the solutions also indicates how considerable base metal deposits could be built up during this phase. Whether the solutions are entirely volcanic or partly meteoric, the metals are surely of magmatic (volcanic) origin (perhaps with a contribution leached from syngentic metals in the enclosing lavas). The solutions are presumably more or less confined to channels and by precipitation and reaction with wall-rocks, lead to localised concentrations of metal, mainly as sulphides. Beds of permeable tuff or interbedded sediment might well be particularly favourable zones for metal deposition, particularly near volcanic feeders and/or structural fissures, thus giving rise to conformable orebodies. Deposits of this type have been recognised in breccia vents and at shallow levels in volcanic piles, but are generally in discordant fissures (e.g. the Braden ore body in Chile). Alteration of the wall rocks to form chloritic and sericitic rocks is also well documented in active volcanic areas (e.g. Naboko, 1960) and some of the alteration commonly found near the orebodies under discussion might well be of volcanic origin.

The resulting ores, which may be re-deposited during and after tectonic deformation, would appear to be hydrothermal-
magmatic to the present day observer. The sulphur isotope ratios, for instance, would not distinguish them from those derived from independent granitic magmas. Biogenic isotope ratios would indicate a syngenetic or early diagenetic origin, but even in these types the metal content may have been added to biogenic, syngenetic sulphur (probably in pyrite) by later hydrothermal volcanic activity in discrete channels. Such a process might explain the economic concentrations of copper, lead and zinc in strata within a volcanic sequence and containing widespread but small amounts of syngenetic sulphides (mainly pyrite).

It is not clear why spilites and keratophyres rather than other volcanics should be related to "pyrite", "pyrrhotite" and pyrite-chalcopyrite orebodies. Possibly the answer lies in the geosynclinal environment that is common to both volcanics and ores, with its development of unique magmas.

THE ORIGIN OF THE MT. LYELL AND ROSEBERY-HERCULES ORES

Previous views on the origin of the Mt. Lyell deposits have generally involved hydrothermal-magmatic processes, either associated with intrusive porphyries or unseen granitic stocks, of Devonian age (e.g. Gilbert and Pogue, 1913; Hills, 1914a; Edwards,
1939; Conolly, 1947). Lindgren (1933, p. 622) and Wade and Solomon (1958) classified the ores as of mesothermal replacement type. Bradley (1957) related mineralization to a regional Devonian metasomatism but indicated that regeneration of pre-existing deposits may have occurred.

Peters (1893) suggested the Blow orebody was sedimentary and deposited in a swamp under reducing conditions but this theory was not generally accepted.

For the Rosebery-Hercules ores, all workers have related the ores to hydrothermal-magmatic processes of replacement, Hall et al. (1953) classifying the ores as mesothermal.

In recent years several writers have touched on the possibility of a volcanic origin (e.g. Campana et al. 1958; Hall and Solomon, 1962; Pereira, 1963; and Solomon, in press, Appendix A, paper 1), a suggestion first put forward by Thureau (1886) for the Blow orebody at Mt. Lyell. Gregory (1905, p. 131) believed the ores at Lyell were concentrated from igneous rocks which, from his account, are inextricably mixed with the pre-Ordovician volcanics.

There is little doubt that there was a Tabberabberan phase of mineralization involving hydrothermal alteration between about 300 and 600°C (acid solutions ?) followed by ore deposition (basic
However, there is some evidence for a pre-Ordovician phase of mineralization. The hematite-barite veins in the quartz keratophyres and the chert at Mt. Lyell are pre-Ordovician and the hematite bodies at Mt. Lyell may be fossil gossans. At least some of the hydrothermal alteration at Lyell is of pre-Ordovician age, judging by pebbles in the basal Owen beds. Boudinaged quartz veins and deformed ores at Rosebery also testify to a pre- tectonic mineralization. Campana et al. (1958) and Hills (1914a, p. 44) referred to schistose volcanic pebbles in the Jukes Conglomerate, indicating pre-Ordovician shearing, but the writer has not been able to establish that the cleavage in similar pebbles was not Tabberabberan.

Pre-existing sulphide ores subjected to deformation of the Tabberabberan type (particularly TF₂) in the presence of hot, aqueous solutions would almost certainly be re-distributed, by diffusion or perhaps volatilisation at the higher temperatures (see discussion by Sorensen (1963)).

Is it possible that the Tabberabberan solutions might have culled their metals from the Mt. Read Volcanics during folding and hydrothermal alteration? To test this query, samples of unaltered lavas, tuffs and granite were collected between Rosebery and South
Darwin and analysed for Cu, Pb and Zn (Table 31). It is immediately obvious from Table 31 that there is no zoning to compare with the distribution of the Cu, Pb and Zn orebodies along the Mt. Read Arc and in fact, the highest Pb and Zn values were obtained near Queenstown. Many of the values are similar to those in volcanic rocks elsewhere in the world but both Cu and Zn contents are rather high (see also Table 13) and there are local concentrations of all metals (e.g. 31657, 31020, 31786) in the acidic rocks. Some of these contents may be introduced but in 31020 for instance, the metals are probably primary because they occur mainly in amygdules (Plate 26, No. 2).

The conclusion is that some of the Mt. Read Volcanics were derived from magma with high metal content but that their distribution is not compatible with the suggestion that the Tabberabberan ores were culled from the volcanics.

Another possibility is that the metals were derived from rocks older than the volcanics during regional metamorphism. Goodspeed (1952), De Vore (1955) and Boyle (1959) have outlined processes whereby solutions may be generated during metamorphism and concentrated in structural traps where ore deposition may follow. These, and similar suggestions, have been ably summarised by Williams (1960). In Tasmania, the Success Creek rocks
Table 31: Copper, lead and zinc in the Mt. Read Volcanics

**QUARTZ KERATOPHYRES, ACID TUFS AND GRANITES IN THE MT. READ ARC**

<table>
<thead>
<tr>
<th></th>
<th>Pb (p.p.m.)</th>
<th>Zn (p.p.m.)</th>
<th>Cu (p.p.m.)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>From South to North</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Darwin Granite (31761)</td>
<td>25</td>
<td>160</td>
<td>97</td>
</tr>
<tr>
<td>Lynch Creek tuff (31657)</td>
<td>23</td>
<td>360</td>
<td>460</td>
</tr>
<tr>
<td>Lynch Creek tuff (31659)</td>
<td>26</td>
<td>130</td>
<td>79</td>
</tr>
<tr>
<td>Whip Spur (31676)</td>
<td>21</td>
<td>55</td>
<td>65</td>
</tr>
<tr>
<td>Great Lyell (31702)</td>
<td>12</td>
<td>24</td>
<td>70</td>
</tr>
<tr>
<td>Comstock Tram (31751)</td>
<td>45</td>
<td>310</td>
<td>90</td>
</tr>
<tr>
<td>Comstock Tram (31752)</td>
<td>29</td>
<td>120</td>
<td>62</td>
</tr>
<tr>
<td>Comstock Tram (31753)</td>
<td>30</td>
<td>150</td>
<td>85</td>
</tr>
<tr>
<td>Comstock Tram (31754)</td>
<td>37</td>
<td>125</td>
<td>75</td>
</tr>
<tr>
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**ANDESITE**

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**SPILITE**

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underlying the volcanics appear to be the only possible source because the underlying Older Precambrian metamorphics reached their present metamorphosed condition prior to Success Creek deposition. The Success Creek beds are relatively thin below the Mt. Read Volcanic Arc (Fig. 2) and are only mildly metamorphosed, so that they hardly provide a suitable source of metal.

It is far more likely that the ores are of magmatic origin and that any pre-Tabberabberan mineralization was related to volcanic activity. Following the scheme already outlined it is tentatively suggested that such mineralization occurred during Mt. Read effusion and was derived from juvenile-meteoric waters circulating at depth in permeable zones related to early fractures and magmatic conduits (see Fig. 2). Deposition was concentrated in certain acid volcanics which may have been suitable repositories by virtue of their chemical composition or physical properties (e.g. permeability).

 Renewed movement along these old zones of weakness in the Devonian, combined with the rise of hot, aqueous solutions, may have regenerated the earlier deposits to varying degrees. This Tabberabberan phase, presumably due to deep-seated magmatic activity in the major fracture zones, may have contributed much of the metallic content of the ores.
The long history of tectonic and magmatic activity along the western margin of the Tyennan Geanticline may well have been paralleled by repeated phases of mineralization and the tectonic history may provide the link between mineralization, vulcanism and plutonic activity.
REFERENCES


-------------, and Grabham, G. W., 1909: Albitization of Basic Plagioclase Feldspars. Geol. Mag., 6, 250-256.


Banks, M.R., 1956: The Middle and Upper Cambrian Series (Dundas Group and its Correlates) in Tasmania:

----------, 1962a: The Cambrian System in Geology of Tasmania,

----------, 1962b: The Ordovician System in Geology of Tasmania,

----------, 1962c: The Silurian and Devonian Systems in Geology
of Tasmania. J. geol. Soc. Aust., 9, 177-188.

----------, and Solomon, M., 1961: Cambrian Succession in West

Barth, T., F.W., 1962: Theoretical Petrology (2nd ed.). John
Wiley and Sons.


Bartrum, J.A., 1929: Igneous Rocks at Mount Camel, Hohoura,

----------, 1956: Spilitic Rocks in New Zealand. Geol. Mag.,
73, 414-423.


Battey, M.H., 1950: The Geology of Rangiawhia Peninsula, Doubtless

----------, 1951: Notes to Accompany a Topographical Map and a
Provisional Geological Map of Great Island, Three

----------, 1955: Alkali Metasomatism and the Petrology of
some Keratophyres from New Zealand. Geol. Mag.,
92, 104-126.

----------, 1956: The Petrogenesis of a Spilitic Rock Series from
New Zealand. Geol. Mag., 93, 89-110.


Boyle, R. W., 1959: The Geochemistry, Origin and Role of Carbon Dioxide, Water, Sulphur and Boron in the Yellowknife Gold Deposits, Northwest Territories, Canada. Econ. Geol., 54, 1506-1524.


----------, 1959: Some phase Relations in Hydrothermally Altered Rocks of Porphyry Copper Deposits. Econ. Geol., 54, 351-373.


Fuller, R. E., 1940: Ellipsoidal Structure as the Gigantic Disperse Phase of an Emulsion (abs.) Bull. geol. Soc. Amer. 51, 2022.


Knopf, A., 1949: Recent Results of Investigations on the Feldspars. J. Geol., 57, 592-599.


Kohler, A., 1949: Recent Results of Investigations on the Feldspars. J. Geol., 57, 592-599.


---------- and Fuller, R. E., 1928: Chlorophaeite, Sideromelane and Palagonite from the Columbia River Plateau. Amer. Min., 13, 360-382.


Pringle, J., and George, T. N., 1948: South Wales (2nd edit.)
Brit. reg. Geol.


Rast, N., 1956: The Origin and Significance of Boudinage. Geol. Mag., 93, 401-408.


APPENDIX A

Papers referred to in the Text

1. Geology and Mineralization of Tasmania.
2. The Copper Ore Deposits of Mt. Lyell (with R. G. Elms).
3. The Tin Ore Deposits of Mt. Bischoff.
4. The Silver-Lead-Zinc Ore Deposits of Mt. Farrell.
5. Geology of the Mt. Bischoff District (with D. I. Groves).
6. Counting and Sampling Errors in Modal Analysis by Point Counter.
7. The Tectonic History of Tasmania.
8. The Metallic Ore Deposits of Tasmania (with G. Hall).
9. Rib and Hackle Marks on Joint Faces at Renison Bell, Tasmania (with P. A. Hill).
11. The Dundas Group in the Queenstown Area.
12. The Mineralized Rift Valleys of Tasmania.
Geology and Mineralization of

Tasmania.

by

M. Solomon

The pre-Devonian rocks of Tasmania have been affected by several phases of deformation, culminating in the Tabberabberan Orogeny, the most important metallogenetic epoch in the island's history. Post-Devonian sediments lie more or less undisturbed on the older strata and are unmineralized. The mineral deposits lie largely in the west and north parts of Tasmania (Fig. 1), where older rocks are exposed, these being largely covered in the central to eastern and south eastern parts by the post-Devonian rocks.

The most important ore deposits contain Sn, W, Cu, Pb, Zn, Au and Ag and the bulk of production to date has come from ores in rocks of early Cambrian or late Proterozoic age. Production figures are given in Table 1. The principal operating mines are at Mt. Lyell (copper), Rosebery (lead, zinc), Rossarden (tin, wolfram), King Island (wolfram), Renison Bell (tin) and Tullah (silver, lead).

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GEOLOGICAL HISTORY

Precambrian

The oldest rocks in Tasmania are probably those outcropping in the Central Highlands and parts of the north and west coasts. They consist of deformed, low to medium grade quartzose schists (with muscovite, garnet and albite), schistose quartzites, phyllites and amphibolites and are referred to as the Older Precambrian. Spry (1963) suggests these rocks have undergone at least two pre-Cambrian deformation and have been involved in large-scale recumbent folds. The relationship between these rocks and the so-called Younger Precambrian is not at all clear. The latter consists of quartzites and slates with minor dolomites, conglomerates and volcanics; they are relatively undeformed but are of similar primary lithology to the Older Precambrian. Though there is some reason for believing them to overlie the Older Precambrian unconformably, there is also reason to believe that at least in part the Older Precambrian rocks are severely deformed varieties of the Younger Precambrian. Spry (1962) has tentatively concluded that the two rock units are separated by an orogeny, named the Frenchman Orogeny.

Younger Precambrian rocks form successions several thousand feet thick on the west and north coasts. Their distribution indicates that in Younger Precambrian time the Older
Precambrian formed a geanticline (the Tyennan Geanticline) in the Central Highlands area and that the surrounding basin formed part of a large miogeosyncline that probably extended to the mainland of Australia (Fig. 2). The margins of the Geanticline have had a significant control over fold trends in succeeding orogenies.

**Penguin Orogeny**

Deposition was then interrupted by the Penguin Orogeny, with at least two stages of folding, and intrusion of dolerite dykes e.g. at Burnie and possibly in the Savage River area. The tentative dating of the Cooee dolerites at 700 million years (see Spry, 1962) indicates a late Proterozoic age for this Orogeny. The Granite Tor stock, east of Tullah, may also have been intruded at the end of Penguin folding.

**Early Cambrian or late Proterozoic**

The geological history immediately following the Penguin Orogeny is uncertain. Tentatively, it appears that following the Penguin Orogeny several thousand feet of sandstones, siltstones and dolomite were deposited over the Younger Precambrian miogeosyncline, overlapping for several miles onto the Tyennan Geanticline (Fig. 3). The sediments of this transgressive phase include the Carbine Group at Dundas (east of Zeehan), the Smithton and Jane Dolomites, the Success Creek Group (Taylor, 1954), the sandstones and dolomite at Mt. Bischoff, and the calcareous sequence below spilites on King
Island and at Smithton. In the Zeehan district Blissett (1962) includes many rocks of this phase (called here the Success Creek phase) in the Oonah Formation, considered to be Younger Precambrian (Spry, 1962), and Blissett believes the Oonah Formation is succeeded conformably by the Crimson Creek sediments. Thus the relation of many of the Younger Precambrian and Success Creek rocks to the Penguin Orogeny is not clear.

The age of the Success Creek phase is uncertain but is estimated to be early Cambrian or late Proterozoic.

**Cambrian**

In places (e.g., Zeehan) the upper parts of the Success Creek phase contain volcanic rocks which herald a pronounced change in the tectonic stability and form of sedimentation within the geosyncline. For the remainder of the Cambrian the Tyennan Geanticline, now considerably reduced in size, probably remained emergent or under a very shallow sedimentary cover. Similar but smaller "islands" may have formed on an arc along the Rocky Cape Geanticline, ringing the Tyennan mass to north and west (Fig. 4). The relatively narrow arcuate basin between these two structures is known as the Dundas Trough.

At this time Tasmania formed part of the Tasman Geosyndinal
Zone of Eastern Australia, the early stages of which were marked by widespread basic and spilitic submarine vulcanism in Tasmania and Victoria (Thomas and Singleton, 1957). On King Island several thousand feet of spilites and minor pyroclastics conformably overlie mudstones and dolomite breccias (tillite ?) and west of Waratah, over a thousand feet of spilites, pyroclastics and greywackes overlie the Bischoff sandstones. Spilites occur immediately above the Success Creek phase at Smithton and within the Success Creek phase near Zeehan but on parts of the North coast extend to the top of the Middle Cambrian (Banks, 1962a).

The initial extrusion of spilites accompanied a deepening of the sedimentary basin and a change to synorogenic sedimentation characteristic of eugeosynclines; accompanying igneous activity was typical of the ophiolite association. Filling of the sinking basin was accomplished by (a) transport of fine detritus from the shores, (b) occasional inrushes of coarse material from the shores, possibly by density currents (c) accumulation of lavas and considerable quantities of tuff and (d) the products of erosion of volcanic piles and tectonically rising ridges.

The development and decay of localised, ephemeral ridges and troughs was probably a feature of the Cambrian in Tasmania. The resulting sediments are variable and impersistent, making mapping and correlation extremely difficult. The main rock types,
other than volcanics, are paraconglomerates, greywackes and mudstones with minor cherts and limestones. The quantity of volcanic material decreases upwards in the succession and fossils are present only in the upper parts. The rocks succeeding the Success Creek phase are estimated to be 15\textasciitilde20,000 feet thick on the West Coast and comprise the Crimson Creek Argillite (unfossiliferous) and the Dundas Group (lower Middle Cambrian to middle Upper Cambrian).

Campana and King have compared the Crimson Creek and Dundas sediments to the Flysch facies (argillite and greywacke types respectively) and certainly they have the orogenic environment of the Flysch. However, the Tasmanian sediments are not similar to the European Flysch lithologies which are monotonous, well bedded and graded sandstones and mudstones with some conglomerates and limestones.

**Cambrian Igneous Activity**

Associated closely with spilitic lavas and keratophyric tuffs in the Crimson Creek Argillite are thin sills and small plugs of saussuritised albite gabbro of spilitic composition. Large sheet-like bodies of serpentinite and serpentinised pyroxenite occur within the Cambrian sediments, particularly in the Zeehan-Rosebery area, at Heazlewood, Adamsfield, and near Beaconsfield. Where their stratigraphic relationships can be observed these bodies form concordant sheets along the Dundas Group-Crimson Creek boundary.
At Adamsfield Upper Cambrian sediments contain boulders of serpentinite. The ultrabasics may have been intruded under a shallow cover of Middle Cambrian sediments, becoming exposed at Adamsfield by a local Upper Cambrian uplift (Banks, 1962a) or, alternatively, they may have been extruded on the sea floor prior to Dundas Group deposition and locally eroded prior to being covered. The serpentinites are hosts for small copper-nickel orebodies, magnetite, chromite, and osmiridium. This ophiolitic association is typical of sedimentation in a tectonically active eugeosynclinal belt.

An important feature of the Cambrian basin in Tasmania was the development of a thick volcanic pile - the Mt. Read Volcanic Arc - around the north and west margins of the Tyennan Geanticline (Fig. 2). Near Mt. Read it consists of about 10,000 feet thick of basal, potash-rich, brecciated rhyolites, agglomerates and tuffs overlain by mainly sodic volcanic breccias, tuffs, keratophyres and quartz keratophyres. These volcanics are genetically related to the spilites outside the arc and are probably mainly submarine. The age of the volcanics is an intractable problem but it appears that the Mt. Read Volcanics overlie the Success Creek rocks and may have continued to form well up into Dundas Group time. Campana and King (1963), however, suggest they are overlain unconformably by the Crimson Creek and Dundas beds, and that the volcanics developed between Success Creek and Crimson Creek deposition. The volcanic pile appears to have reached its greatest development between the Pieman River and Mt. Darwin,
and certain horizons within the pile act as hosts for the largest of Tasmania's sulphide deposits.

The Mt. Read Volcanics continue north-east of Rosebery to reappear near Ulverstone as the Lobster Creek Volcanics. Keratophyric volcanics also extend parallel to the north coast between Moina and Beulah (the Minnow Keratophyre, Beulah Formation, etc.) though these volcanics are largely in the Upper Cambrian (Burns and Jennings in Banks, 1962a). They may be younger than, or equivalent to the upper part of, the Mt. Read Volcanics. At Tullah and Mt. Darwin the Mt. Read Volcanics were intruded by sodic and potassic granite and adamellic of similar composition to the enclosing quartz keratophyres and rhyolites. The granites are crudely concordant and were probably intruded into the core of the volcanic pile towards the end of volcanic activity.

Near Mt. Read, at Mt. Sedgwick, Mt. Darwin and other localities a particular type of potassic lava was intruded by veins up to 200 feet wide of magnetite-hematite-barite, these components being primary constituents of the magma.

Jukesian Orogeny (= Tyennan Orogeny of Browne, 1949b)

Sedimentation ceased abruptly in the Upper Cambrian with the onset of the Jukesian Orogeny. This produced gentle folding of the Cambrian and older sediments on trends parallel to the margin of
the Tyennan Geanticline, as evidenced by unconformities south of Queenstown, near Tullah, Moina, Ulverstone and other places. The dominating feature however, was the major faulting of similar trend which uplifted the Tyennan and Rocky Cape Geanticlines and produced an inter-geanticline depression. This depression was split by an axial ridge of Cambrian rocks (the Dundas Ridge of Bradley, 1954 or the Porphyroid Anticlinorium of Carey, 1953) into two elongate basins of which the biggest was the Owen basin on the west and north flank of the Tyennan mass. The Owen basin is regarded as a rift valley (the Owen rift valley) by Campana et al. (1958). The Dundas Ridge forms a sharp western wall to the Owen basin from Queenstown to Tullah, the wall marking the line of the prominent West Owen Rift or Great Lyell Fault Zone. This faulting probably followed Jukesian folding. The form of the inter-geanticlinal depression is indicated in Figure 3 which pictures conditions at the close of Owen deposition. On its western side (e.g., the Zeehan basin) there seems to have been only one fault wall, the basin being similar to a fault-angle depression. Other basins formed on the east flank of the Tyennan Geanticline (e.g., at Adamsfield and near Beaconsfield). The basin and ridge movements of the Jukesian Orogeny appear to be due to a continuation of the tension stresses causing the initial sag of the Cambrian basin.

Reference has already been made to the Murchison and Darwin
Granites, which may be more related to Jukesian Movement than to Mt. Read vulcanicity. Pre-Ordovician ages have been suggested for the Dove Granite (near Lorinna) by Twelvetrees (1913) and for the Granite Tor stock by Bradley (1957).

**Ordovician - the Junee Group**

The Owen Basin was initially infilled by the Jukes Conglomerate, derived from erosion of the Mt. Read Volcanics. This was succeeded by up to 3,000 ft. of quartzose conglomerate and sandstones (the Owen Conglomerate) derived largely from erosion of Precambrian rocks. The Lower Owen consists of grey and yellowish coarse conglomerates of possible fluviatile origin, the Middle Owen consists of reddish, medium-grained conglomerates and coarse sandstones while the Upper Owen consists of red, finer grained sandstones with marine fossils and fine-grained conglomerates. This gradual upward diminution in grain size was accompanied by lateral transgression until by the close of Owen time, thin marine sandstones covered much of the Dundas Ridge and parts of the Geanticlines (Fig. 3). The Caroline Creek Sandstone is correlated with the Upper Owen and carries Arenigian marine fossils (Banks, 1962b). At Queenstown, erosion of barite-magnetite veins (possibly containing sulphide also) produced localised thick lenses of limonite-hematite rock on the flank of the Owen basin.

The marine transgression continued during the deposition of shales and limestones (e.g. the Gordon Limestone) which were deposited
over a wide area of Tasmania. These beds are up to 5,000 feet thick and of Middle Arenigian to Upper Ordovician age. They represent a return to miogeosynclinal conditions.

Silurian and early Devonian - the Eldon Group

The basal bed of the Silurian succession, the Crotty Sandstone, is a quartz sandstone or grit that contains detrital chromite and probably is the result of a relatively minor uplift of the source areas. Minor movements at this stage are indicated by a disconformity at Flowery Gully and may be correlated with the Benambran Orogeny of eastern Victoria.

The Crotty Sandstone heralded a prolonged phase of tectonic quiescence during which some 10,000 feet of sandstones, mudstones and limestones were deposited in a miogeosynclinal environment (the Florence Sandstone, Amber shale etc.). The sediments are known as the Eldon Group and extend into the Lower Devonian. They exhibit a gradual diminution in grain size in combination with fairly regular alternations of sandstone and mudstone; representing a gradual decrease in the rate of uplift of the source areas. The extent of the Eldon Group is not known but it clearly covered much of the island and may have covered the Tyennan and Rocky Cape Geanticlines. There is no sign of the tectonic activity associated with the Bowning Orogeny in New South Wales. Part of the Mathinna Beds of eastern Tasmania are correlated with the Eldon Group.
Post-early Devonian

The most important orogenic phase in Tasmania was the Tabberabberan Orogeny. Evidence from Point Hibbs possibly indicates that earth movements had begun to effect the miogeosyncline in the Lower Devonian (Banks, 1962c) and that sedimentation ceased during the Middle Devonian. At Eugenana, undisturbed late Middle Devonian cave deposits in deformed Gordon Limestone indicate an upper limit for the age of the main tectonic activity. Most of the structures visible today were developed during the Tabberabberan Orogeny and deformation was followed by a major phase of granite intrusion and mineralization.

Sedimentation did not resume until early in Permian times and it is likely that earth movements and possibly igneous activity continued into the Kanimblan Orogeny that was so strongly developed in Victoria and New South Wales. From the early Permian to the present day, tectonic activity has been largely epeirogenic. During the Permian and Triassic there were deposited several thousand feet of mudstones, sandstones, conglomerates and limestones comprising two cycles of marine sediments separated by three freshwater sequences. In the Middle Jurassic these sediments were intruded by large and thick sheets of dolerite, the intrusion being accompanied by tensional faulting. Very minor gold mineralization was associated with the Cygnet syenite porphyry of Cretaceous age.
Faulting on approximately NW trends occurred through the Tertiary to form large and small scale horst and graben structures. During this time Tasmania lay on the southern margin of the marine basin that extended into Victoria and South Australia and localised terrestrial deposition took place in some of the graben. Outcropping orebodies like those at Mt. Bischoff were oxidised during the moist warm phases of the Tertiary and bauxite developed locally on dolerite and basalt. Most of the tin and gold alluvial deposits were formed during the late Tertiary and Quaternary.

**TABBERABBERAN OROGENY**

**History**

It is tentatively suggested, as discussed by Solomon (1962), that the Tabberabberan deformation took place in two stages. The earliest deformation consisted of differential vertical uplift to form long wavelength, arcuate synclinoria and anticlinoria on trends already established by preceding movements and controlled by the Tyennan and Rocky Cape Geanticlines (Fig. 5). One of the principal structures (the Dundas-Deloraine anticlinorium) more or less coincides with the axis of the Dundas Trough and also the Dundas Ridge. The mobility of this zone has clearly been one of the main features of Lower Palaeozoic tectonics. Many of the major NW folds in north-east Tasmania may have developed at this time.
The structural pattern at this stage is basically similar to that of the Jukesian Orogeny and it is possible that development of these folds began during the Silurian and early Devonian with a tendency to isolate smaller basins within the miogeosyncline. The Tyennan Geanticline, though acting as a relatively rigid nucleus, also participated in this broad folding; with passage of time (and increasing overburden) its effects on the pattern of deformation decreased. This first stage of folding was accompanied by vertical movements on pre-existing faults near the Geanticlinal margins such as the Great Lyell Fault Zone, which had been active in the Jukesian Orogeny.

In the succeeding stage of the Tabberabberan Orogeny the influence of the Geanticlines was negligible and the orogeny was dominated by structures of approximately NW trend (WNW to NNW). This style of deformation applies particularly to the Zeehan-Queenstown district (Fig. 6). Superimposition, and interference with, earlier folds produced severe local complications. The NW folds are generally of smaller wavelength than the arcuate folds and in most cases superimposition of the NW folds on the earlier structures resulted in marked changes of plunge, e.g. in the Huskisson and Zeehan synclines. A pronounced axial surface cleavage with steep dip is prevalent in all but the most competent rocks such as the Owen sediments and Precambrian quartzites in which breakthrusts are
common. The assertion by Campana and King (1963) that the NNW cleavage at Rosebery is of Cambrian age is untenable for there is nothing to distinguish the cleavage at Rosebery from surrounding Tabberabberan cleavage. In some areas interference between the fold systems has resulted in major fractures between blocks of different fold trend. Jennings (in Solomon, 1962) describes, from the North Coast, thrusts and wrench faults that are marginal to NW folds cutting across earlier E-W folds. Further movement took place at this stage on pre-existing Jukesian structures such as the Henty Fault near Rosebery (Fig. 6).

In some areas the NW folds are bent towards an E-W trend (e.g. at Queenstown and Zeehan) and these zones are characterised by approximately E-W faults with pronounced vertical and transcurrent movements. The finest example is the Linda Fault Zone which has been traced for some 20 miles and is several miles wide. The faults of this zone are most obvious in the Owen and Precambrian formations. In the shallow Devonian sediments the faults swing to NE trend and assume mainly tensional characteristics. The Mr. Lyell copper deposits lie within the Linda Fault Zone. Some early E-W folds, like the Bischoff anticlinorium (Fig. 6), may be due to repeated movement on pre-Tabberabberan deep-seated structures.
Igneous Activity

A number of stocks of granite were intruded late in the Tabberabberan Orogeny or during the Kanimblan Orogeny. These stocks are post-folding but have been faulted in places (e.g., at Heemskirk and Pieman Heads). They appear to be intruded along large scale anticlinal structures (Fig. 4) as suggested by Carey (1953); the stocks of north eastern Tasmania in particular tend to be aligned NW. All are characterised by late stage mineralization near to the margins of the intrusion. Browne (1949a) has suggested that there are two ages of granite, one Tabberabberan and one Kanimblan, the latter types being mineralized. Preliminary dating indicates a Lower Carboniferous age for the Heemskirk Granite (Evernden and Richards, 1962) and an Upper Devonian age for the Coles Bay granite (dating by G. Davis) but two distinct periods of intrusion have not yet been established.

At Waratah and Renison Bell, small plugs and dykes of quartz-porphyry intruded along anticlinal axes and are associated with cassiterite-pyrrhotite mineralization.

Tectonic Analysis of the Lower Palaeozoic Orogenies

Precambrian tectonics are not well known and are of little interest in relation to mineralization. The early Cambrian to early Devonian history in Tasmania appears to have been one of essentially
tensional stresses. These produced mainly basin and ridge topography within the Tasman Geosyncline, sedimentation in the basins being accompanied in the early stages by rise of magma to the surface or to shallow depths.

The Tabberabberan Orogeny was the first major interruption to sedimentation and probably consisted of two phases. The first represents a continuation of earlier tectonic movements with differential vertical movements producing broad folding on pre-existing trends. The second involved development of a new strain pattern dominated by NW folds and WNW to E-W transcurrent fractures which also show vertical movement. The suggested tectonic framework for the Linda Fault Zone is shown in Figure 7, in which the dominant stresses are shown as due to E-W directed transcurrent shear with north-side-west movement. The conjugate N-S transcurrent faults are rare though a possible example occurs five miles west of Queenstown. The indicated stress field, which is a modification of that proposed by Carey (1953), may be of regional significance.

Post-Permian tectonics are essentially tensional and epeirogenic, the principal structures being NW gravity faults possibly in conjunction with N-S transcurrent faults. Pre-existing Tabberabberan structures probably locally controlled the post-Permian trends. Carey (pers. comm.) believes the faults originate by north side east movement on E-W transcurrent faults in southern Victoria.
MINERALIZATION

The metallic mineral deposits occur largely in the west, north and north east of Tasmania (Fig. 1) and the most important mineral districts are:

1. The Darwin-Lyell line (Cu)
2. The Hercules-Rosebery-Pinnacles line (Pb-Zn Cu)
3. The Heemskirk Granite area (Sn and Ag-Pb-Zn)
4. The Dundas-Renison Bell area (Sn and Ag-Pb-Zn)
5. The Mt. Cleveland-Waratah area (Sn and Pb-Zn)
6. The Moina area (Sn-W-Au)
7. The Rossarden-St. Helens area (Su, Sn-W and Au)
8. King Island (W).

The deposits can be roughly classified according to their spatial relationship to igneous rocks:

(a) **Intramagmatic**: cassiterite in granite (Heemskirk); chalcopyrite and pentlandite in serpentinite (Cuni); osmiridium in pyroxenite (Adamsfield).

(b) **Contact Metasomatic**: scheelite in granite aureole (King Island).

(c) **Adjacent to granitic intrusions**: cassiterite and wolframite in discordant veins (e.g. Aberfoyle, Moina); cassiterite, pyrrhotite and pyrite in concordant replacement lenses (Mt. Bischoff, Renison Bell); silver-rich lead and zinc in fissures (Magnet, Zeehan).
(d) **Deposits not obviously related to intrusions but associated with volcanic rocks:** bornite, chalcopryite and pyrite in volcanics (Mt. Lyell); sphalerite and galena and chalcopryite in volcanics (Rosebery, Hercules).

(e) **Not obviously related to igneous rocks:** gold in quartz veins (Mathinna, Beaconsfield).

**Pre-Devonian**

Magnetite lenses are found in amphibolite dykes intruding Older Precambrian rocks in the Savage River area, 15 miles west of Waratah. The magnetite lenses may be the result of segregation in situ (Hughes, 1958) or of segregation and injection (Hall and Solomon, 1962). Hughes believes they are Cambrian but Spry (1964) suggests they are Older Precambrian.

Nickel and copper-nickel mineral assemblages occur in several ultrabasic masses of Cambrian age but are not of economic importance (Williams, 1958). These assemblages are probably injected magmatic segregations. Hall and Solomon (1962) and Campana and King (1963) have tentatively suggested that the two principal mines (Mt. Lyell, and Rosebery) exploit ores of exhalative-volcanic or similar origin. These orebodies, and also those at Mt. Farrell, occur within the Mt. Read Volcanics and are apparently unrelated to any granite bodies. They appear to be at a level in the Mt. Read Volcanics not far above the top of the Success Creek phase. The Mt. Lyell ores
are composed of bornite, chalcopyrite and pyrite with very minor sphalerite and galena, the Rosebery ores consist of sphalerite, galena and chalcopyrite, and the Mt. Farrell ores are galena with minor sphalerite and chalcopyrite. They were all deposited in their present form later than the NW cleavage, and almost certainly late in the Tabberabberan Orogeny. Though there is little evidence of earlier mineralization at Rosebery or Tullah, there is some indication (in the form of Ordovician iron oxide screes and other evidence) of pre-Ordovician mineralization at Mt. Lyell. Hills (1914) suspected that the copper ores of the Jukes-Darwin field were related to the Darwin Granite, of Cambrian age.

It is of interest to note that 90% of Tasmania's mineral production has come from deposits contained in rocks near the top of the Success Creek phase. In some cases, deposition at this horizon has been facilitated by the presence of carbonates, e.g. dolomites at Renison Bell, Mt. Bischoff and King Island. However, in others, deposition seems more closely related to the presence of spilites e.g. 85% of the production from the Zeehan field (King, 1961), and also the Magnet Mine, west of Waratah.

Tabberabberan Orogeny

Associated with the post-orogenic granite stocks are deposits of cassiterite and wolframite, with minor molybdenite and bismuthinite. They occur near the granite margins as fissure systems and small stockworks, and are clearly of magmatic origin. Typical examples are
the Aberfoyle, Storeys Creek, Shepherd and Murphy, and Heemskirk ores. Some lead-zinc orebodies are also closely related to granite bodies, e.g., at Round Mountain.

Scheelite occurs as a replacement of early Cambrian or late Proterozoic carbonate in the metamorphic aureole of a granodiorite on King Island. The carbonate at a similar horizon acts as host for the Mt. Bischoff and Renison Bell pyrrhotite-cassiterite ores which are derived from quartz porphyry dykes and plugs. These possibly post-date the granite stocks and might well be of Kanimblan age. Bilibin (1955) has found from a world-wide study that this type of deposit is rare in pre-Permian rocks.

Lead-zinc-silver haloes ring sulphide-cassiterite ores at Mt. Bischoff and Renison Bell. It is interesting to note that the Dundas and Zeehan silver-lead-zinc fields join with the Renison Bell, Rosebery-Hercules and Tullah ores to form a narrow, NE-trending mineralized zone that contains all the important Pb-Zn mineralization in Tasmania. When this zone is combined with the N-S line joining Rosebery and the Mt. Lyell copper field, it includes over 80% of Tasmania's metal production. The NE zone includes three small centres of tin production (Renison Bell, Razorback and Montana) and it is suspected that the Pb-Zn fields represent haloes about these centres. Possibly the NE trend reflects some pre-Tabberabberan tensional
fracture zone that has controlled mineral deposition in a later period. Almost all of the important deposits of this zone occur near the base of the Crimson Creek Group.

This picture casts considerable doubt on the supposed zoning of ores around the Heemskirk Granite as proposed by Ward (1911) and amplified by Edwards (1953). Magnetite occurs between the tin veins of the Heemskirk Granite and the silver-lead Zeehan lodes, which themselves show a crude zoning of gangue constituents. However, the magnetite is related to a serpentinised ultrabasic (Hughes, 1959, Blissett, 1962) and the Zeehan ores form part of the Zeehan-Tullah zone mentioned above. Thus the magnetite and silver-lead ores do not necessarily form zones around the Heemskirk Granite.

The gold deposits at Beaconsfield, Lefroy etc. are probably late-stage hydrothermal deposits.

The possibility of the Rosebery and Mt. Lyell fields being of essentially syngenetic origin has already been mentioned. However, it is clear that final deposition was at least late-Tabberabberan because the ores are post-cleavage and show strong orientation by Tabberabberan structures. Whether or not these ores are regenerated is a subject for debate.

Relation of Mineralization to Tectonics

Recent work by Russian geologists has summarised the relation
of ore emplacement to the history of geosynclines. Bilibin (1955) has given a generalised igneous, metallogenic and tectonic cycle that compares closely with the history of the Tasman Geosyncline in Tasmania. The more important phases of Tasmanian history are given in Table 1 and are compared to a reduced version of Bilibin's sequence.
References


### TABLE I

**Major Mineral Production in Tasmania**

<table>
<thead>
<tr>
<th>Metal</th>
<th>1963 Production</th>
<th>Total Production (to end of 1963)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Copper</td>
<td>15,602 tons</td>
<td>593,600 tons</td>
</tr>
<tr>
<td>Zinc</td>
<td>38,589 tons</td>
<td>717,380 tons</td>
</tr>
<tr>
<td>Lead</td>
<td>11,889 tons</td>
<td>approx. 670,900 tons</td>
</tr>
<tr>
<td>Gold</td>
<td>31,989 oz.</td>
<td>2,600,494 oz.</td>
</tr>
<tr>
<td>Silver</td>
<td>1,438,858 oz.</td>
<td>approx. 84,500,000 oz.</td>
</tr>
<tr>
<td>Tin</td>
<td>1,005 tons</td>
<td>148,000 tons</td>
</tr>
<tr>
<td>Wolfram</td>
<td>382 tons</td>
<td>approx. 28,750 tons</td>
</tr>
<tr>
<td>Osmiridium</td>
<td>-</td>
<td>31,008 oz.</td>
</tr>
</tbody>
</table>
Captions for Figures

Fig. 1. Locality Map of Tasmania.

2. Palaeogeography of the Younger Precambrian.

3. Palaeoprofile to illustrate the geological history of the Zeehan-Mt. Tyndall area up to the close of Owen Conglomerate deposition; see Campana and King (1963) for further sections.


5. Arcuate folds of the early phases of the Tabberabberan Orogeny.

6. Geological map of West Tasmania, compiled from Campana and King and other sources.

7. Stress and Strain Diagrams for the later part of the Tabberabberan Orogeny as typified by the Linda Fault Zone.
THE COPPER ORE DEPOSITS OF MT. LYELL

by

1
M. Solomon

and

2
R.G. Elms

INTRODUCTION

The copper deposits of Mount Lyell (145° 35' E; 42° S) occur on a narrow zone of intermittent copper sulphide mineralisation extending for 20 miles from Mount Darwin in the south to Lake Dora, north of Queenstown (Fig. 1). Further to the north along the line of this zone are the Rosebery-Hercules and Mount Farrell lead-zinc-copper zones.

Mining in the Queenstown area began in 1883 when the Iron Blow was worked for gold. A year later the adjacent sulphide ores were exposed and the field recognised as a potential copper producer. By 1898 forty-two companies had been formed to exploit the copper resources of Mount Lyell. In 1903 the two major companies, the Mount Lyell Mining and Railway Company Limited and the North Mount Lyell Copper Company, successfully amalgamated in an effort to make

1. Geological Consultant, Department of Geology, University of Tasmani
2. Chief Geologist, Mt. Lyell Mining and Railway Co. Ltd.
a profitable operation out of two uncertain ones. Most of the small companies were short-lived and by 1933 the last of them had sold out to the Mount Lyell Mining and Railway Company, which from that time has been the field’s sole producer (Blainey, 1956).

The field has produced 567,890 tons of copper from ore varying between 0.7 and 8.0% Cu, as well as 16,018,623 ozs. of silver and 594,521 ozs. of gold to the end of 1963. Production for 1962 was 14,445 tons of electrolytic copper from 2,169,377 tons of ore.

Mt. Lyell is one of a series of peaks varying from 2,600 to 4,800 feet above sea level that form the north-trending West Coast Range. The peaks rise sharply on the western flank from a deeply dissected peneplain that slopes gently to the coast. Most of the Mt. Lyell mines lie on the west side of a narrow divide joining Mt. Lyell and Mt. Owen and the workings extend between 700 ft. and 2,000 ft. above sea level. The divide is exposed to the prevailing westerly winds and receives an annual average rainfall of over 100 inches.

GEOLOGICAL HISTORY

The geological succession is as follows:

**Pleistocene:** Till and varved clays in the Linda and Comstock Valleys

**Jurassic:** Dolerite sill on Mt. Sedgwick

**Permian:** Tillite on Mt. Sedgwick, beneath dolerite

Tabberabberan Orogeny-Middle Devonian
Lower Devonian to Silurian: Eldon Group:

+ 7,000 ft: alternating sandstones and mudstones; rare limestones.

Ordovician: Junee Group: 1,000 feet: Gordon Limestone.

0-4,500 ft: Owen Conglomerate:

Lenticular siliceous conglomerates and sandstones, sub-divided as follows:-

Upper: Haematitic sandstones and fine grained conglomerates, largely marine.

Middle: Red sandstone and medium grained conglomerate, part marine.

Lower: Coarse grained, pale coloured conglomerate.

0-500 ft: Jukes Breccia-conglomerate:

talus breccia composed largely of Cambrian volcanics.

Jukesian Movement - intrusion of Darwin Granite

Cambrian: Mt. Read Volcanics:

10,000 (?) ft: Acid to basic, sodic and potassic lavas; pyroclastics and laminated siltstones.

The Mt. Read Volcanics, in which the orebodies occur, occupy an extensive arcuate belt which passes between Rosebery and Queenstown
with a north-south trend. The volcanics are largely acid to intermediate keratophyres and pyroclastics with rare spilitic flows and siltstone lenses; they are considerably altered, poor in ferromagnesian minerals and rich in soda and potash.

The volcanics have undergone deuteric (?) chloritisation and albitisation and were cleaved and slightly metamorphosed (low grade greenschist facies) in the Devonian. Their origin is probably part terrestrial and part marine.

The Mt. Read Volcanics have been interpreted as sheared intrusives (Twelvetrees, 1902; Nye, Blake and Henderson, 1934; Edwards, 1939; Conolly, 1947; and Alexander, 1953) but were recognised as altered volcanics by Gregory (1903); Hills (1927); Hills and Carey (1949); Scott (1954); Bradley (1954); and Wade and Solomon (1958). Scott and Bradley considered them to be basic lavas and greywackes that had been metasomatised in the Cambrian (Scott) or the Devonian (Bradley), to acid and intermediate porphyries.

Their age has been much discussed and it is not yet clear whether they are equivalent to the fossiliferous Dundas Group (lower Upper Cambrian to upper Middle Cambrian) or the Crimson Creek formation or older. Ten miles south of Queenstown, the volcanic sequence was intruded by tabular sheets of Cambrian granite (the Darwin granite) of similar composition to the enclosing rhyolites. The granite, of sub-volcanic origin, was exposed to erosion during Owen deposition as also were dykes of magnetite-hematite intruded into certain potash
rhyolites along the West Coast Range.

The Mt. Read volcanic arc developed off-shore from a N-S trending geanticline of Precambrian rocks (the Tyennan Geanticline) that had developed in the centre of Tasmania and which was only shallowly submerged during the Cambrian. It rose rapidly along marginal faults in the early Ordovician and shallow basins developed on its perimeter and within the Mt. Read volcanic arc. The Owen Basin, the largest, formed along the west margin of the Geanticline and was filled by detritus of largely Precambrian rocks. As the Tyennan Geanticline was eroded, the sea transgressed over most of western Tasmania and deposition of the Upper Owen marine sands was followed by widespread deposition of the Gordon Limestone and succeeding miogeosynclinal sediments of Silurian to mid-Devonian age. Deposition ceased in the mid-Devonian as a result of Tabberabberan movements.

STRUCTURE

The principal tectonic structures (Fig. 3) are Tabberabberan in age; the earliest were major synclines and anticline trending N-S to NNW with wavelengths of 5-6 miles, their trend probably controlled by the physical discontinuity against the Precambrian at the western margin of the Tyennan Geanticline. The dominating structure is the West Coast Range Anticlinorium, the axis of which passes through
Queenstown and continues north to Mt. Dundas (Fig. 1). The site of this structure was probably elevated as a horst in immediate pre-Owen time, to form the western wall to the Owen basin; at least in the mine area, this wall shed volcanic detritus into the Owen sediments. One of us (M. S.) believes that slumping from this wall in the mine area may have led to upturning of the Owen sediments during the Upper Owen and that further upturning and upthrusting took place in the Tabberabberan Orogeny. This line of upturning, known as the Great Lyell Fault Zone, extends several miles north of Queenstown and for 3-4 miles to the south (Figs. 3 and 4). In the Lyell mine area the base of the upturned Owen beds forms a west-dipping wall which has been proved by drilling over a vertical range of nearly 3,000 ft. (Fig. 6).

Smaller-scale folds trending WNW to NW were superimposed on these early N-S features and are expressed in the Owen Conglomerate as steep reversed faults and in the Cambrian and Silurian rocks as tightly appressed folds with axial surface cleavage. Cleavage generally strikes NW or WNW and dips steeply to the SW but in the Lyell area the cleavage in the Lyell Schists is 'moulded' around the Owen Conglomerate margin, the direction being controlled by the contact. Thus at Comstock a strong NE cleavage is developed against the NE trending contact. At Queenstown the structure is complicated by the WNW faults of the Linda Fault Zone which produce ridge and basin structures; the principal basins coincide with the
Linda and Comstock Valleys (Figs. 3 and 4). These faults are pronounced in the Owen Conglomerate and Precambrian rocks but fade out on entering the Cambrian or Silurian beds, where the displacement is dissipated in folds. Late in the WNW to NW faulting and folding, sinistral transcurrent movements took place along the WNW faults and on deavage faces; prior to this horizontal phase, the tectonic movements had been essentially vertical. The most important faults are the WNW trending Sedgwick and North Lyell faults, both of which have vertical displacements (north side up) of several thousand feet and transcurrent movements of 5,000 ft. and 2,500 ft. respectively (Figs. 2 and 3).

A few NE faults, probably tensional in origin, occur in the region but do not seem to be important controls over ore deposition, as was suggested by Conolly (in Alexander, 1953) and Bradley (1956).

Slumping (?) in the Upper Owen led not only to local up-turning and crumpling of the Upper Owen beds but also resulted in volcanic material being carried into the Owen basin. This Cambrian material, apparently forming a wedge in the Owen sequence, now forms a schist belt (the Tharsis schist zone) within upturned Owen beds and is the host for several important orebodies (Figs. 2 and 6).

ORE DEPOSITS

Mineralisation of the Lyell field took place towards the close of the tectonic activity of the Tabberabberan Orogeny. Ore deposition
was controlled by structural channels and the sulphides are generally post-cleavage although on a few discrete cleavage surfaces post-ore movement has taken place.

All the economic deposits, with the exception of Comstock, occur in a 1½ mile strip on the divide between Mt. Lyell and Mt. Owen (Fig. 3). However, sulphide mineralisation is distributed along a narrow, straight, N-S zone extending from South Darwin to Comstock and reappearing again at Lake Dora. Of these uneconomic prospects, the largest, from south to north, are (Fig. 1):

East Darwin (pyrite-chalcopyrite); Lake Jukes (bornite); Jukes Proprietary (chalcopyrite); Great Lyell (pyrite-chalcopyrite); and Lake Dora (pyrite, chalcopyrite and some sphalerite and galena). These orebodies occur in altered quartz keratophyres.

The orebodies of the Mt. Lyell area are largely in altered Mt. Read Volcanics adjacent to the steeply upturned base of the Owen formation and they form a series of echeloned lenses extending up to 2,000 ft. from the Owen-Mt. Read contact. Throughout the mineralised area the volcanics are altered to sericitic and chloritic "schists" by recrystallisation and hydrothermal alteration of the cleaved volcanics. These schists, which vary in texture and composition of the rock and the intensity of alteration, are known as the Lyell Schists; their formation pre-dates sulphide deposition.

The economic deposits of the Mt. Lyell area consist of iron
and copper sulphide and may be divided in general into the following types:

(a) Massive pyrite-chalcopyrite  
(b) Disseminated pyrite-chalcopyrite  
(c) Chalcopyrite-bornite  

In addition, there are occurrences of galena and sphalerite at Comstock and native copper (in the "copper clays") on the east flank of the Mt. Lyell-Mt. Owen divide.  
(a) The massive ores consist of up to 75% pyrite and the Cu content ranges from trace quantities to about 1%. The principal examples are the orebodies at South Lyell and the Iron Blow (previously described as the Big Blow, the Mt. Lyell orebody, or the Blow). The South Lyell orebody is very low in copper and was mined to provide flux for the siliceous ores from North Lyell. It is steeply inclined, elongated along cleavage and steeply pitching. The Iron Blow, 400 ft. away, differs mainly in that it contains more chalcopyrite (70-5%Cu).  

The Iron Blow was so named because a massive hematite body occupied the footwall of the ore and this hematite was thought to be the source of the detrital gold that initiated interest in the Lyell area. The hematite body has been removed as overburden but a similar mass at North Lyell appears to be a particularly hematitic section of a ferruginous conglomerate that occurs along part of the base of the Owen beds. Both the Iron Blow and the North Lyell masses have previously
been regarded as formed by magmatic (?) segregation (Power, 1892); deposition from hypogene solutions leaching pyrite (Gregory, 1905); oxidation of pyrite in situ (Peters, 1893; Montgomery, 1893); deposition from hydrothermal solutions, the iron being magmatic in intrusive porphyries (Edwards, 1939); deposition in a 'basic front' related to Devonian mineralisation and granitisation (Bradley, 1954); and transfer by hydrothermal solutions of iron in adjacent volcanics (Wade and Solomon, 1958). It is suggested by one of us (M. S.) that the hematite masses are dehydrated limonitic scree deposits formed during Owen deposition from weathering of Cambrian magnetite-hematite veins in the Mt. Read Volcanics, and possibly even from weathering of pre-Owen sulphide deposits. The Blow orebody is banana-shaped in section (Fig. 5) and replaces the Lyell Schists right up to the conglomerate wall which is, at least in part, a reverse fault. The orebody was worked by both underground and open-cut methods from 1897 to 1922. The ore consisted mainly of pyrite (up to 85%) but chalcopyrite was important in the upper levels, along with enargite, tetrahedrite, bornite, chalcocite and a little galena and sphalerite. The first 1,500,000 tons of ore mined averaged 2.85% Cu but grade declined with depth; total production was 5,497,000 tons averaging 1.28% Cu, 2.67 oz. Ag per ton, and 0.065 oz. Au per ton. The remnant of the orebody averages only 0.5% Cu. Of the small rich shoots within the orebody, the most important was the Mt. Lyell bonanza, which yielded 850 tons assaying 21.3% Cu and 1023 ozs. Ag
per ton; it contained stromeyerite with bornite, chalcopyrite and gold in quartz (Petterd, 1910). The £106,312 obtained from this ore did much to keep Mt. Lyell in production in the early, uncertain years.

The Iron Blow and South Lyell are the only pyrite-chalcopyrite orebodies with fairly well defined walls; the remainder are outlined by assay values only and the orebodies represent ill-defined zones of copper enrichment.

(b) Disseminated pyrite-chalcopyrite ores differ from type (a) in that the mineralisation is weaker, with total sulphide content varying between about 35% and 5%. In general, the copper content is inversely proportional to the distance from the Owen-Schist contact.

The chief orebodies of this type are those in the West Lyell Open Cut, and the Royal Tharsis and the Comstock. They are elongated along cleavage but generally dip slightly more steeply than the cleavage. Some disseminated orebodies (Crown Lyell No. 1 and Crown Lyell No. 3) occur in the North Lyell fault and one, the Lyell Tharsis orebody, occurs in the Tharsis schist zone.

The West Lyell Open Cut includes an echeloned series of ore lenses that strike about NW, parallel to the foliation, dip steeply (slightly steeper than the foliation which dips SW at about 75°) and pitch at about 90°. They vary in grade from over 1% to less than 0.5% Cu, contain from 6 to 20% FeS₂, and occur in both quartz-sericite and quartz-chlorite schists. Some of the orebodies contain molybdenite, and bornite and covellite have been reported in drill holes. The schists
show coarse compositional banding, and, in places, remnants of volcanic breccia (?), shale, and feldspar porphyry are recognisable.

Annual production from the cut is approximately 2,000,000 tons of ore carrying about 0.7% Cu and reserves for open-cut mining are estimated at 18,000,000 tons. The main ore body, the Prince Lyell, extends downwards to at least R.L. -800 ft. (a proved vertical extent of 2,300 ft.) and in depth improves in grade.

Other similar pyrite-chalcopyrite bodies occur in the vicinity but have proved too low in copper. With the Royal Tharsis and West Lyell orebodies, they form a strip of echelon lenses parallel to the Owen-Schist contact.

The Royal Tharsis orebody extends over a vertical distance of at least 1,800 ft., similar to Prince Lyell. Its strike and dip is approximately that of the foliation and the mineralogy is similar to the West Lyell orebodies. Like these, it gradually approaches the conglomerate wall in depth. It has been largely mined out above R.L. 750 ft., the ore from this section averaging 1.6% Cu. Approximately one million tons of 2.0% Cu ore are indicated below R.L. 750 ft.

The Comstock ore-bodies comprise four lenses, arranged in vertical echelon and dipping north-westerly. The top lens strikes NE to ENE and the bottom lens N, a swing that parallels the conglomerate

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1 Mine datum is 100 ft. below local mean sea level.
contact. The top lens dips steeply NW and pitches westerly at about 60°, the lower lenses dip west at about 60° and have a 90° pitch.(139,574),(949,612)

The ore lenses extend from the surface at R.L. 1,700 ft. to below R.L. 700 ft. The top two are mined out, the third partly mined and the fourth is virtually unmined. To the north of, and alongside, the orebodies is a pipe-like mass of brecciated chert veined by hematite. Veining and brecciation are pre-Ordovician in age, judging by pebbles gathered from the Owen Conglomerate, and the chert is apparently a product or associate of Cambrian volcanism, and possibly of pre-Ordovician mineralisation. A similar mass of chert occurs at North Lyell, in close association with the rich North Lyell orebodies.

Comstock ore minerals include chalcopyrite, pyrite and a little bornite, and Edwards (1939) records magnetite, galena and free gold. When mining ceased in 1944 about 400,000 tons of developed ore (2.2% Cu) remained.

The No. 2 Crown Lyell orebody is a relatively small lens that strikes N-S, dips west and plunges 60° northwesterly. It extends from its surface outcrop at R.L. 1,860 to R.L. 1,250 ft. and up to 1955 it had yielded 436,000 tons of ore assaying 1.65% Cu. It lies in schist within the North Lyell Fault Zone.

(c) Chalcopyrite-bornite ores are concentrated in the North Lyell area, lying partly in the North Lyell Fault zone and partly in the Tharsis schist zone (Fig. 3 and 5). Discovered in 1897 during
the building of a road to the Eastern Orebody, the North Lyell ores proved to be the richest on the field, averaging 5.4% Cu and yielding 4,642,860 tons of ore up to 1953, when production ceased. Mining recommenced in 1959 and reserves approximate 100,000 tons of 6% Cu ore. Smelting difficulties defeated the original North Lyell Company, for despite the presence of rich ore, it was compelled to amalgamate with the Mt. Lyell Company, which possessed poorer copper ore but an abundance of the pyritic ore necessary for successful smelting. The orebodies strike approximately NW but are most irregular in shape and therefore difficult to prospect. The original discoveries, which were in the Tharsis Schist zone, proved to be steep pipes but the orebodies in the North Lyell Fault zone were found to be very irregular.

The bornite-chalcopyrite ores are very low in pyrite and generally occur in a siliceous gangue. Sulphides other than pyrite, chalcopyrite and bornite are rare; Edwards (1939) records chalcocite, sphalerite, galena, tetrahedrite, enargite and gold. Native copper and copper carbonates are occasionally seen in fault zones below the surface.

A feature of the North Lyell area is the presence of large masses of brecciated and fractured hematitic chert (Fig. 5), similar to the Comstock chert, and the presence of a hematite bed at the base of the Owen Conglomerate. The chert, and the enclosing chloritic and sericitic schists, are hosts to the sulphides, which occur as
(a) bornite-chalcopyrite (b) chalcopyrite or (c) pyrite-chalcopyrite ores.

The Lyell Tharsis orebody occurs south of North Lyell and in the Tharsis schist zone. The high grade ore is worked out and only a low grade halo (1% Cu) remains. The ore was largely chalcopyrite with some bornite.

Galena-sphalerite ore has been mined by the Tasman and Crown Lyell Extended Co. from a small lode 800 ft. east of the Lyell Comstock mine. Typical grade was 28% Pb, 20% Zn, 0.5% Cu, and 17 ozs. of Ag per ton. The lode is steeply dipping and occurs in schist adjacent to the Owen Conglomerate.

The Copper Clays occur on the eastern side of Mt. Lyell-Mt. Owen divide in the narrow gullies draining the Linda Valley. All three major deposits, the Blocks, Consols and King Lyell, were worked between 1895 and 1910 by mining and sluicing. The ore consists mainly of nodular goethite ramified and replaced by veinlets of native copper and cuprite; clayey material and siderite occur in varying amounts. Old reports refer to veins of chalcocite, chalcopyrite and pyrite in the goethite. The ore overlies the Upper Owen beds conformably and underlies dark grey siliceous shales of the Gordon Limestone. Particularly at King Lyell the ore appears to have replaced a bed of limestone (or siderite). It is suggested by one of us (M.S.) that the goethite and copper minerals probably have a dual origin, being derived partly by indigenous oxidation of pre-existing sulphides formed in the Tabberabberan mineralisation and partly by deposition
from acid meteoric waters draining the outcropping sulphides on
the divide. The copper clays probably developed during the
prolonged sub-tropical erosion phases that characterised the
Mesozoic and early Cainozoic Eras in Tasmania. The steep slopes
and inert gangue generally allowed only limited gossan development
over the sulphide bodies and much of the copper removed by
weathering was trapped in the copper clays.

MINERALISATION

The important primary copper minerals in the Lyell area
are chalcopyrite and bornite, with minor quantities of chalcocite,
digenite, tetrahedrite, tennantite and enargite. Among the secondary
minerals, native copper and cuprite are locally important. Minor
amounts of galena, sphalerite, stannite, gold and other minerals
also occur (Edwards, 1939).

Markham (1963) suggests a general order of crystallisation
as follows:

Pyrite (earliest)
Pyrite + chalcopyrite and pyrite + bornite
Chalcopyrite + bornite
Bornite + chalcocite
Bornite + digenite

This sequence is characterised by an increase in both Cu : Fe
and metal : sulphur ratios.

ORIGIN

The ores are of the mid-temperature, 'hydrothermal' type, deposited after alteration of the country rocks, along an old fault zone that was strongly reactivated in the Tabberabberan Orogeny. No intrusive igneous source for the mineralisation is known either for the Lyell field or the sulphide mineralisation of the northerly extensions - Rosebery and Mt. Farrell. The nearest Devonian granite to Mt. Lyell is 20 miles to the north-west and the only post-Cambrian igneous rocks in the area are lamprophyre dykes. Three of these dykes have been found in the mine area; they are up to 6 ft. wide and several thousand feet in length, parallel to the cleavage and unmineralised. They are composed of biotite, olivine and aegirite, similar to lamprophyre dykes elsewhere in western Tasmania. They are possibly an indication of a granitic mass at some depth below Mt. Lyell.

The obvious association of the Mt. Read Volcanics with the major sulphide ore deposits in western Tasmania leads to speculation on the possibility of a syngenetic origin for the ores. Some indication of possible ore deposition associated with Cambrian volcanism has already been given and Devonian hydrothermal activity might well have regenerated some pre-existing metallic ore.

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References


**Figure Captions**

**Fig. 1.** The Mt. Lyell district.

**Fig. 2.** Geological map of the Mt. Lyell mine area.

**Fig. 3.** Principal structures in the Mt. Lyell district.

**Fig. 4.** North-Section through the Mt. Lyell area.

**Fig. 5.** East-West section through the Mt. Lyell area.

**Fig. 6.** Cross sections (generalised) of:

(a) The Blow orebody.

(b) The North Lyell orebodies.
QUEENSTOWN

+ AREA OF FIG. 3

MT. SEDGWICK

LINDA

MT. OWEN

MT. LYELL

QUEENSTOWN

W.C.R. ANTICLINORIUM

KING

RIVER

WEST

JUKES PROPRIETARY

LAKE JUKES

EAST DARWIN

MT. DARWIN

SCALE

0 1 2 MILES

FIG. 1
FIG. 3
FIG. 5

- **DEVONIAN SHALE**
- **SILURIAN SANDSTONES ETC.**
- **GORDON LIMESTONE**
- **OWEN CONGLOMERATE**

- **JUKES CONGLOMERATE**
- **MT. READ VOLCANICS**
- **PRECAMBRIAN**

0 1 2 MILES
Tin Ore Deposits of Mt. Bischoff

by

M. Solomon

INTRODUCTION

Mt. Bischoff (145° 31', 5'E/41° 25'S) is a quarter of a mile north of Waratah in north-west Tasmania. Discovered in 1871, the mine flourished until 1900, after which production declined to the present-day figure of 7-8 tons of tin metal p.a., recovered largely from alluvial deposits and dumps. The grade has averaged 1% tin and a little over 5,500,000 tons of ore have yielded 54,100 tons of tin metal.

GEOLOGY

Flat-lying Tertiary basalt capping sands and gravels forms a dissected plateau at about 2,000 ft. above sea level above which Mt. Bischoff rises some 600 ft. (Figs. 1 and 2). The Tertiary rocks are up to 200 ft. thick and unconformably overlie the following succession:

Jurassic: Dolerite

Devonian: Granite and quartz porphyry dykes

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Cambrian (?): Crimson Creek Formation: ? 10,000 ft:

Mudstones greywackes, breccias, spilites.

Precambrian (?): approximately 1,000 ft: grey quartzites and shales

0-250 ft: dolomite

1,000 ft+: grey quartzites and shales.

In the immediate vicinity of the mine workings the dolomite, which is an important host for ore, reaches its maximum thickness and is outcropping or overlain by 1-200 feet of quartzites and shales. Knight (1953) describes an altered dolomitic shale within the dolomite. The Cambrian (?) and Precambrian (?) sediments are essentially conformable but there is evidence of pre-Cambrian deformation. Syngenetic, "sedimentary" brecciation and crumpling is common in the Precambrian (?) quartzites, particularly in those above the dolomite.

The Precambrian (?) beds crop out in the axial zone of the E-trending Bischoff anticlinorium which plunges gently to the west (Fig. 1). The subsidiary E-W folds of this structure are well exposed in the mine area (Fig. 2) and are disrupted by longitudinal and oblique folding to produce "keel" structures. The structure is further complicated by minor NE folds, and by major tension (?) faults that are important in controlling distribution of the dolomite. The major structures were produced during the Tabberabberan
(Middle Devonian) Orogeny.

Quartz-feldspar porphyry dykes were intruded late in this Orogeny as a radial swarm centered on Mt. Bischoff and the hinge of the Bischoff anticlinorium. Several sills extend out from the dykes, particularly near the top and bottom of the dolomite bed. The porphyries are similar to the elvan dykes of Cornwall, England, and have been replaced in the mine area by topaz, tourmaline and to a lesser degree by sericite, fluorite and cassiterite (Groves and Solomon, 1964). Breccias along the dyke walls suggest the dykes were introduced into radial tension faults (rather than fractures as in Knight, 1953) which were possibly related to the rise of a granitic cupola from an underlying granite mass. It is suspected that the Meredith Granite, which crops out two miles south of Mt. Bischoff, continues north under Mt. Bischoff.

MINERALIZATION

The tin mineralization, mainly as cassiterite, was centred near the focus of the radial dyke swarm and formed five types of lodes.

(a) Pyrrhotite-talc-carbonate bodies

Most tin was produced from pyrrhotite-talc-carbonate bodies replacing dolomite. The carbonate is a coarse-grained Ca-Mg-Fe-Mn carbonate similar to the pistomacite of the Magnet Mine and
distinct from the fine-grained, almost pure sedimentary dolomite. The talc is largely metasomatised dolomite but some appears to replace tremolite. Chondrodite occurs in the Brown Face. Pyrrhotite almost completely replaces the dolomite in the Brown Face, Slaughteryard Face and Greisen Face orebodies (Fig. 2) near the axis of the anticlinorium but south of Greisen Face pyrrhotite replacement is less and concentrated near the base of the dolomite.

Stillwell (1945) described three types of ore from the Greisen Face: carbonate-sulphide, massive pyrrhotite, and talc-pyrrhotite. In the first type pyrite, arsenopyrite, and pyrrhotite are cut by later chalcopyrite and galena. Cassiterite occurs as fine grains in pyrite, arsenopyrite and carbonate, and is generally earlier than the sulphides. Colloform pyrite is cut by later sulphides. Cassiterite is rare in the massive pyrrhotite type but relatively common in the talc of the talc-pyrrhotite ore. Tin distribution in the ores is sporadic and can only be determined by assay.

The dolomite and sulphide ore probably cropped out during erosional phases between the late Mesozoic and Recent times; limonite gossans developed in places but elsewhere the weathering products were completely removed, leaving residual and detrital cassiterite at the surface. This chemically and mechanically concentrated cassiterite sand was some of the richest tin ore in the
world and enabled the mining company to declare huge profits early in its history. Primary ores were preserved only in synclinal hinges or where protected by a thin quartzite cover.

(b) **Pyritic Lodes**

Zones of pyrite and quartz-pyrite ore occur within the mineralised dolomite. Much of the pyrite may be secondary, derived from pyrrhotite by high temperature alteration but some of the pyrite appears to be primary and confined to dyke margins. Leaching of this ore leaves a crumbly quartz-pyrite rock, generally rich in cassiterite (e.g. White Face ore). Small quantities of secondary pyrite have been produced during oxidation of the pyrrhotite.

(c) **Fissure Lodes**

Several fissure lodes extend beyond the area of dolomite alteration. They vary from a few inches to 20 ft. in width and from steeply dipping to horizontal. They consist of quartz and sulphides with only minor pyrrhotite. Most of the lodes have approximately NW trend, parallel to a major fault direction, and probably are fault fillings. The largest such orebody, the Giblin Lode, which was worked by the Mt. Bischoff Extended Company, had an average width of 2 ft. and was worked for a vertical depth of 1,000 ft. It averaged 1% Sn and consisted of pyrite, sphalerite, arsenopyrite and cassiterite in a gangue of quartz, tourmaline, topaz, fluorite, etc. Production amounted to over 150,000 tons. Other important bodies were the North Valley and Queen lodes, which contained similar
mineral assemblages to the Giblin Lode. Stillwell (1943) reported stannite replacing cassiterite in the North Valley Lode.

(d) Disseminated and joint fillings in dykes

Portions of the dykes assay 0.3-4% Sn, the cassiterite occurring as a replacement of feldspar or with quartz as joint fillings. Together with cassiterite in numerous joints in sediments adjacent to the dykes, this ore type provides most of the visible cassiterite found in the mine.

(e) Alluvial and Lacustrine

Late Tertiary lacustrine (?) sands and gravels and Recent soils have been worked for detrital tin south of Mt. Bischoff and there have been extensive alluvial workings along the Waratah River.

Mt. Bischoff is ringed by a number of small silver-lead-zinc deposits that form an aureole around the tin mineralisation.

The tin mineralisation is apparently confined to a high-temperature vertical funnel that followed the path of an acid dyke swarm, both dykes and mineralisation probably stemming from a local, cupola-like bulge in the Meredith Granite. The similarity between the Mt. Bischoff and Renison Bell mines is striking, particularly with regard to stratigraphy, structure, mineralogy and igneous activity, and a common mode of origin is indicated.
REFERENCES


FIG. 2

- NORTH TERTIARY BASALT
- WARATAH RIVER
- NORTH VALLEY LODE
- MT BISCHOFF
- SLAUGHTERYARD SYNCLINE
- BROWN FACE SYNCLINE
- GREISEN FACE
- SOUTH SCALE 600 1200 FT

TERTIARY
- BASALT
- QUARTZ PORPHYRY DYKES
- CAMBRIAN
- MUDSTONE & GREYWACKE
- PRECAMBRIAN
- SANDSTONE & SHALES
- DOLOMITE

FIG. 2
LEAD-SILVER-ZINC ORE DEPOSITS AT MT. FARRELL

By

M. Solomon

INTRODUCTION

Several small sulphide deposits occur along a five mile belt between Mt. Murchison and Tullah (long. 143° 37' E, lat. 41° 43' S). Discovered in 1897, the important mines (North Mt. Farrell and New North Mt. Farrell) have produced 84,000 tons of lead and 10,173,400 oz. of silver from small scale intermittent workings to the present day, the ore averaging 13.5 per cent Pb, 16 oz/ton Ag, 2.5 per cent Zn and 0.5 per cent Cu. The New North Mt. Farrell workings are 900 ft. deep.

GEOLOGY

The deposits occur in or close to a half mile wide belt of laminated mudstones, tuffs and greywackes (the "Farrell Slates" of earlier workers) which strike NNE and dips steeply W (Fig. 1, and Hall and Solomon, 1962, Fig. 59) and is overlain and underlain

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by Cambrian (?) lavas and pyroclastics of the Mt. Read Volcanics. The sediments and tuffs form one limb of a pre-Ordovician fold which was steepened (overturned ?) and crumpled and sheared on NNW and NNE trends in the Middle Devonian Tabberabberan Qrogeny. Eastwards these rocks are unconformably overlain by Ordovician Owen Conglomerate (Fig. 1).

ORE DEPOSITS

The ore deposits occur in two lode zones (the Murchison and Farrell) which contain several subparallel lenticular fissure lodes in Tabberabberan shears. The lodes swing in strike from NNW to NNE, bifurcate locally, and dip steeply west. They are up to 300 ft wide, several hundred feet long, and contain ore shoots with a southerly pitch that are richest at fracture intersections and where they cut tuffaceous beds. In the Murchison Mine ore is confined to that part of a NNE tensional fault that cuts a 40 ft tuff bed. The ore is mainly unsheared but schistose galena in the New North Mt. Farrell Mine indicates local post-ore movement.

The Farrell lodes consist of early quartz and late siderite with mainly pyrite, sphalerite, galena, chalcopyrite, jamesonite and tetrahedrite. Some shoots are abnormally rich in copper and
zinc minerals. The Murchison lodes contain more arsenopyrite and pyrite and less galena and siderite. Wall rock alteration is slight; some disseminated pyrite occurs near the lodes but is distinct from syngenetic pyrite in the sediments (Hall et al., 1953).

The ores appear to be of medium temperature, hydrothermal origin and were probably deposited during the Tabberabberan Orogeny. The nearest igneous rock, the Murchison Granite, three miles south-east of Tullah, is of Cambrian age.

ACKNOWLEDGEMENT

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REFERENCES AND BIBLIOGRAPHY


FIG. 1
THE GEOLOGY OF THE MT. BISCHOFF DISTRICT

by

D. I. GROVES AND M. SOLOMON

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Hobart, 7th May, 1964.

Publications from the Department of Geology, University of Tasmania


27. CAREY, S. W., 1955a.—Wegener’s South America-Africa Assembly, Pit or Misfit? *Geol. Mag.*, Vol. 92 No. 4, pp. 196-200.


44. Teachers' Edition of No. 42.


64. KING, L.—The Origin and Significance of the Great Sub-Oceanic Ridges. *Univ. of Tas. Cont. Drill Sym.,* pp. 82-105.


Continued inside back cover.
ERRATA

PAPERS AND PROCEEDINGS OF THE ROYAL SOCIETY OF TASMANIA, VOLUME 98.
University of Tasmania Geology Department Publication no. 142.

GROVES, D.I. AND SOLOMON, M. THE GEOLOGY OF THE MT. BISCHOFF DISTRICT.

Page 8, Line 41, for correlated read correlated
" 9, " 123, " δ " γ
" 11, " 51, " pyroxene " pyroxene
" 12, " 18, " intergrowths " intergrowths
" 12, " 77, " marging " merging
" 12, " 82, " particularly " particularly
" 12, " 97, " far " for
" 12, " 110, " exhibit " exhibit
" 13, " 13, " an analyses " and analyses
" 13, " 20, " α " γ
" 13, " 20, " δ " α
" 14, " 35, " Aanalyses " Analyses
" 14, " 42, " amay " away
" 14, " 62, " mere " were
" 15, " 2, " α " ω
" 15, " 2, " δ " ε
" 15, " 51, " altred " altered
" 15, " 107, " Waartah " Waratah
" 15, " 124, " boht " both
" 18, " 20, " Mineraliaztion " Mineralization
" 18, " 54, " gren " green
" 19, " 50, " gass " gas
" 20, " 65, " geothemometer " geothermometer
" 21, " 131, " Ther-momter " Ther-mometer

Plate 1, Fig. 2, for 1/16 read 1/25
" 1, " 4, " x 35 " x 20
" 2, " 1, " x 68 " x 40
" 2, " 2, " x 68 " x 40
" 2, " 3, " x 68 " x 40
" 2, " 4, " x 35 " x 20


Bull. Geol. Surv. Tasm., 34
THE GEOLOGY OF THE MT. BISCHOFF DISTRICT

By

D. I. Groves
Department of Mines, Tasmania

and

M. Solomon
Department of Geology, University of Tasmania

(With two plates and seven text figures)

ABSTRACT

The basement rocks of the area are the Upper Precambrian (?) Bischoff quartzites and slates, which comprise a 2000 ft. succession of quartzites and sheared mudstones with a bed of dolomite in the upper part of the succession. These sediments have undergone considerable deformation during the pre-Cambrian and/or early Cambrian but show apparent conformity to the overlying mudstones, sandstones, greywackes, volcanic breccias and spilitic lavas of Cambrian (?) age. Late in the Cambrian, ultrabasic and basic dykes were intruded.

The Precambrian (?) and Cambrian (?) rocks were folded into a westerly plunging anticlinorium, with smaller subparallel folds on the limbs. These are distorted by small ENE folds and NNW faults. A granitic body underlies the area and radiating quartz porphyry dykes probably associated with a local, narrow cupola intrude the core of the anticlinorium at Mt. Bischoff. Greisenization of the porphyries is extreme, with the formation of topaz, tourmaline, muscovite, and cassiterite pseudomorphing primary feldspar. Mineralization associated with igneous activity has resulted in selective replacement of the dolomite bed at Mt. Bischoff by iron sulphides, t alc, quartz, carbonate and cassiterite, and the filling of tension fissures throughout the area. A zonation of mineral deposition around Mt. Bischoff is suggested by a nucleus of tin mineralisation with an outer rim of lead-zinc mineralisation. By analogy with other areas in Tasmania the folding, granitic intrusion and associated mineralisation are considered to have occurred during the Tabberabberan Orogeny (Middle Devonian).

Tertiary terrestrial sediments and basalts unconformably overlie older rocks and form a widespread plateau.

PHYSIOGRAPHY

The dominant topographic feature of the Waratah area is an extensive dissected plateau lying at an altitude of 2000 to 2100 feet. This gently undulating plateau is underlain by flat-lying, non-marine, Tertiary sediments and basalt to a depth not exceeding 150 feet, indicating an original Tertiary land surface at an elevation of about 1900 to 2000 feet. The plateau extends down to an altitude of about 1800 feet to the west of the Mt. Cleveland Range and rises to 2200 feet in the northeastern and central section of the Mackintosh area. Inland from there, a series of accordant summits lie between 3000 and 3500 feet, this level corresponding in elevation to the Lower Plateau Surface of Davies (1959).

Rising above this plateau area are several residual mountains including Mt. Bischoff (2598 feet), Mt. Cleveland (3200 feet), Mt. Meredith (2500 feet), Mt. Ramsay (3890 feet), St. Valentine Peak (3640 feet) and Mt. Pearse (3000 feet). These mountains owe their prominence and shape to the rock types of which they are composed; for example the shape of Mt. Bischoff is conditioned by the distribution of the porphyry dykes. Mt. Cleveland has a resistant backbone of Cambrian chert, and the arcuate form of Mt. Pearse reflects a plunging syncline in Owen Conglomerate.

Two main drainage systems have developed since the uplift of the land surface that followed pre-basalt peneplanation; these are the Pieman and Arthur River systems. The Pieman follows a westerly course some 30 miles to the south of Waratah while the Arthur rises in the Waratah District, flows northerly for 35 miles and then westerly, emerging at the coast about 50 miles north of the Pieman River. The Magnet Range forms the watershed between the tributaries of these two rivers, the Whyte River flowing westerly to the Pieman while Magnet Creek and its tributaries join the Arthur River to the east (Fig. 1). The streams generally meander slowly to the margins of the plateau, from which they fall in a series of rapids and small waterfalls to the valley floors some 800 to 900 feet below. Some structural control is suggested for the course of the Waratah River, the river trending just east of north and then swinging sharply to just north of west (Fig. 2); these directions may represent Tertiary structural lines.

In general, the Waratah area can be described topographically as comprising an extensive, deeply dissected plateau area, with sporadic monadnocks rising some 1000 feet above plateau level.

R.S.—2.
Fig. 1.—Sketch map of the Mt. Bischoff district.
STRATIGRAPHY

The basal sedimentary rocks in the area are Upper Precambrian (?) quartzites, sheared shales and dolomite which were previously called the Mt. Bischoff Series by Reid (1923) and are here called the Bischoff quartzites and slates. Overlying these rocks is a thick sequence of shales, mudstones, greywackes, subgreywackes, cherts and altered lavas which were previously called the Dundas Series and are here called Cambrian (?) sediments as their correlation is somewhat obscure. These sediments show apparent structural conformity with the Bischoff quartzites and shales. Unconformably overlying these sediments are conglomerates, sandstones and lignites of Tertiary age, which are generally covered by basalt (Fig. 2).

The succession is summarised below:—

Quaternary—

River gravels and alluvium.

Tertiary—

Gravels, conglomerates, siltstones and lignite: 50-100 ft.

—Unconformity—

Cambrian (?)—

Cambrian (?) argillites, greywackes, breccias, cherts and altered laves:

—Disconformity—

Precambrian (?)—

Bischoff quartzites, slates and dolomite: +2,000 ft.

PRECAMBRIAN

Bischoff quartzites and slates

These crop out in a relatively narrow, E-W trending inlier that extends from the head of Deep Creek (east of Mt. Bischoff) to the Magnet Mine. They consist of alternating quartzites and sheared shales and siltstones, with a bed of dolomite and associated dolomitic shales. The sequence shows strong lithological similarities to the Rocky Cape Group on the North West Coast of Tasmania (Spry, 1957) and the Zeehan Quartzite and Slate in the Zeehan area (Blissett, 1962), both of which are considered as Upper Precambrian. A correlation of these rocks with similar sediments of Upper Precambrian age was also suggested by Carey (1953) and Knight (1953), the sequence previously being regarded as Ordovician.

The succession in the vicinity of Mt. Bischoff is:—

Hangingwall Shales and Quartzites: +1000 ft.

Dolomite, including dolomitic shales: 0-200 ft.

Footwall shales: 0-30 ft.

Footwall quartzites, shales and siltstones: +1000 ft.

This succession differs from that given by Knight (1953), mainly in the addition of the Hangingwall sediments.

The Footwall quartzites and shales are dominantly thinly bedded although massive sequences of 15 feet in a single rock type have been observed. The quartzites contain a high proportion of quartz which has been quite strongly sheared, and thus gives no indication of original angularity and sphericity of grains. The quartz grains exhibit sub-parallel grain and optical elongation and marked undulose extinction, and are in places crushed. The grains vary from 0.1 mm. to 0.5 mm. in diameter. Muscovite occurs in most sections and is quite abundant in several, generally occurring as elongate flakes, 0.1 mm. to 0.4 mm. in length, which are aligned parallel to the elongate quartz grains. Quartz and muscovite comprise the only large grains present, these being set in a matrix of fine sericite and chlorite with sporadic occurrences of interstitial microquartz cement. Minor accessory minerals include well rounded grains of rutile, zircon and tourmaline with graphitic bands and flecks of limonite. Hypogene pyrite is common throughout these sediments.

The sheared shales or mudstones are well laminated and extremely fine grained (0.01 to 0.02 mm. diameter), and consist of quartz, muscovite, sericite and microquartz as in the quartzites. The laminations appear to be produced by variations in the proportion of muscovite which is also found abundantly on the bedding planes. An analysis of a shale from the mine area at Mt. Bischoff (Table 1) indicates that the sheared shales at Mt. Bischoff are apparently more mature than the average shale, having a higher K, O to Na, O ratio, and are more siliceous. However, they do not contain as much silica as the siliceous shale and are probably an intermediate type.

Some thirty thin sections of quartzite and sheared shales from throughout the exposed sequence were examined, and it was found that the sediments in the upper portion of the sequence contained up to 70% quartz grains while the sediments exposed in the Waratah River contained only 50 to 60% quartz. It is also apparent that the sediments become distinctly cleaner towards the upper beds, microquartz cement dominating the normal sericite matrix. Statistical examination of fairly continuous sections of these sediments in the Mt. Bischoff area and in the Waratah River east of Mt. Bischoff (see Fig. 4) revealed that the Mt. Bischoff section comprised 58% quartzites and 42% shales while the Waratah River section comprised 72.5% quartzites and 27.5% shales. Thus there appears to be an increase in shales towards the top of the sequence, accompanied by an increase in the silica content of the sediments.

**Table 1**

<table>
<thead>
<tr>
<th>Analysis of shale from Mt. Bischoff compared with other shales.</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
</tr>
<tr>
<td>SiO₂</td>
</tr>
<tr>
<td>TiO₂</td>
</tr>
<tr>
<td>Al₂O₃</td>
</tr>
<tr>
<td>Fe₂O₃</td>
</tr>
<tr>
<td>FeO</td>
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<tr>
<td>MnO</td>
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<td>Na₂O</td>
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</tr>
<tr>
<td>H₂O₂</td>
</tr>
<tr>
<td>H₂O</td>
</tr>
<tr>
<td>CaCO₃</td>
</tr>
<tr>
<td>Fe₂O₃</td>
</tr>
<tr>
<td>Fe₂O₃</td>
</tr>
<tr>
<td>CO₂</td>
</tr>
</tbody>
</table>

II.—Average shale (Petijohn, 1956).
III.—Siliceous shale (Petijohn, 1956).
The Footwall shales are dark grey in colour, lenticular, and represent a local transition to dolomite.

The dolomite is creamy or pale grey when fresh and weathers to pale brown or fawn. Thin sections of the dolomite (30649, 30649 (a), and 36.R.2 of the Mines Department indicate a fine-grained rock consisting almost exclusively of crystalline dolomite with minor interstitial quartz grains. Irregular patches of coarsely crystalline carbonate were developed during recrystallization related to mineralisation. The dolomite also exhibits a fine macroscopic banding which may represent bedding.

Analyses of the dolomite (Table 2) indicate that it is composed of almost pure dolomite mineral, Knight (1953) also indicating the presence of less than 1% MnO in the dolomite. The thin banding (bedding?) and fine grained texture of the dolomite, and its concordant relations to contiguous rocks, indicate a sedimentary origin (see also Knight, 1953). Thin bedded dolomitic shales occur within the dolomite at the Greisen Face and at the portal of the Main Tunnel.

Table 2
Analyses of Dolomite from Mt. Bischoff

<table>
<thead>
<tr>
<th></th>
<th>Weight %</th>
<th>Mole %</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>CaCO₃</td>
<td>MgCO₃</td>
</tr>
<tr>
<td>I</td>
<td>54.6</td>
<td>43.5</td>
</tr>
<tr>
<td>II</td>
<td>53.9</td>
<td>44.1</td>
</tr>
<tr>
<td></td>
<td>CaCO₃</td>
<td>MgCO₃</td>
</tr>
<tr>
<td></td>
<td>50.4</td>
<td>47.8</td>
</tr>
<tr>
<td></td>
<td>49.7</td>
<td>48.3</td>
</tr>
</tbody>
</table>


The dolomite reaches a maximum thickness of 250 ft. at Mt. Bischoff but is only about 100 ft. thick two miles to the north of the mountain. It has not been found elsewhere, either because it is lenticular or has poor outcrops.

The sediments above the dolomite, the Hanging-wall quartzites and shales, are similar to those below, and are dominantly sheared shales.

Pre-consolidation structures occur within both the Footwall and Hangingwall sediments. Somewhat deformed flow casts are exposed on a bedding surface on the North Valley Road, Mt. Bischoff and deformed examples occur in the Waratah River, 1/2 mile below the mill (Plate I, Fig. 1). As the structure is difficult to determine in these areas no attempt has been made to determine current directions.

Small-scale cross-bedding is visible in some of the coarse siltstones and quartzites but no graded bedding has been observed.

The most widespread feature is the crumpling and brecciation of the sediments: contortion may be confined to a layer a few inches thick between undisturbed material, or may involve several tens of feet of sediments. Plate I, Fig. 2, illustrates a type in which the sand matrix has been liquefied and has incorporated fragments of adjacent mud layers, which retained a shearing strength during the deformation.

The sandstones (Hanging Wall?) above the Greisen Face (Mt. Bischoff) are almost entirely very contorted and/or brecciated, so that the orientation of the rock is obscure. The deformation has a pre-consolidation appearance and involves fine-grained sandstones with little or no shale; in this case several tens of feet of sandy material appear to have been disturbed.
RECENT
ALUVIUM

TERTIARY

DEVONIAN
GREENES FACE
@ WHITE

CAMBRIAN
QUARTZ PORPHYRY
GREYWACKES & MUDSTONES

PRECAMBRIAN
DOLOMITE
QUARTZITES & SHALES

--- FAULT
--- LODE
--- ROAD
--- BEDDING
--- FOOT TRACK

Fig. 4.—Geological map of the Mt. Bischoff mine area. For Section A-B see Fig. 5. Base map by courtesy Lands and Surveys Dept.
Another type exhibits bands of crumpled and brecciated sandstone within a dark shale matrix: the sandstone bands vary from a few mm. to 1½ cm. thick and the shale matrix is generally the dominant lithology. The mechanism in this type is probably the "quickening" of clays containing plastic sandy layers.

CAMBRIAN

Cambrian (?) sediments

These sediments occupy the major portion of the Waratah area and are mainly chocolate-coloured, finely laminated to massive mudstones with minor yellow-brown to grey mudstones, greywackes, sub-greywackes, cherts, chert breccias, sandstones and altered lavas. The boundary between these beds and the Bischoff quartzites and slates is not exposed although to the south of Don Hill and on the Magnet Tram they crop out within several feet of each other. At these localities they appear conformable, although the change in lithology from siliceous shale to greywacke is abrupt and the Bischoff quartzites and slates are much more severely deformed than the overlying sediments.

Chocolate coloured mudstones are the dominant sediments in the area. These are extremely fine grained rocks in which even in thin section the constituent minerals are unrecognizable. Haematite and limonite are abundant, producing the red colouration, and thin carbonate veins are also present. In the Waratah River section these mudstones are rare and the finer sediments are grey, black and yellow-brown laminated mudstones which consist of extremely small, angular fragments of quartz, plagioclase, sericite, and calcite. The laminations are apparently caused by thin bands of limonite and graphitic material. The predominant rock types exposed in the Waratah River are subgreywackes, which also occur sparsely throughout the remainder of the area. The rocks are poorly sorted with an open framework and consist dominantly of angular quartz grains, several of which are elongate and exhibit undulose extinction. Other clastic grains include rounded albite, hornblende, augite, chlorite, magnetite and rare rock fragments. They range in grain size from 0.2 to 0.8 mm. diameter and generally the grains are not quite in contact. The matrix consists of chlorite, iron ores and small fragments of other minerals.

Specimen 30659, from the old Magnet tram-line, consists mainly of isolated grains, up to 0.2 mm. diameter, of angular to sub-angular quartz and rounded feldspar, chlorite and magnetite (in order of abundance). The feldspars are interesting in that while a few grains of fairly fresh albite are present, the majority appear to be similar to microcline. Most of these show a very fine, spindly like normal albite twinning and some show pericline cross-twinning; they are optically negative and 2V measurements range from 79° to 87°. Several contain blebs and irregular patches of clear quartz and some crystals appear to be almost entirely replaced by quartz. The significance of these feldspars will be discussed in the section on igneous activity. Somewhat deformed amygdulike bodies filled with a fine quartz aggregate are also present, with a few flakes of muscovite and small fragments of fine grained basaltic lava.

Other rocks are similar (30655, 30657, 30658, 30660) but vary in quartz (up to 40%) and iron oxide content. Specimen 30657b is unusual in that the matrix has several percent of small (0.05 mm. diameter) hornblende crystals, apparently as a detrital component.

Coarser grained sediments (mainly fine to coarse breccias) are common throughout the area. Many of these rocks contain a high percentage of albite laths and angular volcanic rock fragments with or without clastic muscovite. Fragments of shale are also common, occurring with minor quartz, biotite, carbonate and limonite in a fine matrix of chlorite, albite and a little quartz. Some of the breccias have a much lower volcanic component,
e.g., the breccia outcropping in the Waratah River, 
 mile below the Bischoff mill, which has indistinct 
bedding and is composed dominantly of angular 
rock fragments of the Bischoff quartzites and slates 
and Cambrian (?) sediments; several lava frag­
ments are also present. The fragments are several 
inches in diameter (to a maximum of 9 inches) 
and angular to subangular.

Some of these rocks may be volcanic but the 
presence of quartz and sedimentary rock fragments 
in quantity suggests they are not tuffs and may 
be the result of weathering of volcanic and Pre-
Cambrian rocks.

Nye (1923) classified these greywackes and 
breccias as micaceous or feldspathic “breccias” 
but the number of sediments in which mica was 
an important constituent proved to be very small 
while almost all of them contained feldspar.

Lavas and pyroclastics are rare in the Mt. 
Bischoff area but become more common towards 
the Magnet Mine where they are associated with 
a massive bed of banded chert and chert breccia. 
The volcanic rocks are described in detail in a 
later section.

An interesting member of the Cambrian (?) 
sediments in the Arthur River to the north of 
Waratah is a thin bed of limestone interbedded 
with yellow-brown laminated mudstones. The 
limestone is dark grey in colour, fine grained 
and contains small bands of coarsely crystalline calcite. 
The thickness of the bed is unknown due to 
limited exposure.

The Cambrian (?) sediments are apparently 
unfossiliferous throughout the area although Chap­
man (1929) records annelid remains (described as 
arthropod tracks—_Tasmanadia twelvetreesi_—by 
Glaessner, 1957) in slates on the Arthur River a 
few miles north-west of Waratah.

On the grounds of lithologic similarity, the 
Cambrian (?) sediments in the Waratah district 
have been correlated with the Dundas Group of 
Middle to Upper Cambrian age (e.g., Knight, 1953) 
and also beds lower in the Cambrian (Banks, 1962).

At the present time no direct correlation with 
other sequences is possible and attention can only 
be drawn to the similarities of the Waratah rocks 
to those of the Huskisson River and Dundas areas.

**TERTIARY**

The beds belonging to the Tertiary System con­
sist of conglomerates, gravels, siltstones, sandy 
clays and lignites, forming a sequence generally 
some 50 feet in thickness and reaching 100 feet in 
places.

The basal members consist of conglomerates and 
gravels containing boulders up to 12 inches in 
diameter. Small quantities of casseriterite and gold 
have been found in these basal beds. Above these 
are fine grained, poorly sorted siltstones with water 
borne pebbles and unevenly distributed boulders 
that distort the bedding. The boulders have been 
locally derived, consisting of quartz, quartzite, 
shale, greywacke, breccia, quartz porphyry and 
granodiorite. Boulders of Permian mudstones have 
also been found (Nye, 1923). Small outcrops of 
lignite and ligneous clays occur in the vicinity of 
Waratah and contain numerous leaf impressions 
which have been described by Johnston (1888) as 
belonging to the genera _Eucalyptus, Quercus, 
Laurus, Cycadites_ and _Ulmus_. This and the 
lithology indicate a terrestrial origin for these sedi­
ments and they are thought to have been deposited 
in streams and small lakes. At Don Hill (Fig. 4) 
Tertiary sands may be seen deposited against low 
bedrock cliffs. Differential compaction of the 
younger sediments has produced folds and high 
dip angles near these cliffs (Plate I, Fig. 3).

**IGNEOUS ACTIVITY**

Five major periods of igneous activity are repre­
sented within the Waratah area:

(a) Volcanic emission during deposition of the 
Cambrian (?) sediments, with the pro­
duction of dominantly basaltic lavas.

(b) Intrusion of ultrabasic and basic igneous 
rocks, probably soon after cessation of 
deposition of the Cambrian (?)

(c) Intrusion of acidic rocks including gra­no­
diorites and quartz-feldspar porphyries 
into the Bischoff Series and Cambrian (?) 
sediments during the Devonian.

(d) Intrusion of dolerite during the Jurassic.

(e) Extrusion of basalt over a vast area during 
the Tertiary.

(a) **CAMBRIAN VOLCANIC ACTIVITY**

Lavas are generally rare in the Cambrian (?) 
sediments but become more prolific towards the 
Magnet Mine. Individual flows appear to have 
covered considerable areas, an example in the 
Arthur River (near the confluence with the 
Waratah River) outcropping at fairly close intervals 
over about ¼ mile; the maximum exposed thickness 
is 100 feet. The lavas examined are mainly spilitic 
(e.g. 636, 642, 30658a, 30662a and 35G9 of 
the Mines Department). Many are porphyritic, 
with phenocrysts of albite, augite and chlorite in 
the ground matrix of basaltic sands, laths, chlorite, 
calcare, epidote, magnetite and ilmenite. Some have 
interstitial texture and consist of an interlocking 
aggregate of albite, augite and chlorite with inter­
stitial chlorite, calcite, &c. The albite phenocrysts 
are almost invariably clotted by sericite and in 
some appear to be completely altered to sericite 
and in some appear to be completely altered to a fairly 
interchange aggregate of sericite plates and 
shales. They reach a maximum size of about 
2 x 0.5 mm. As in the spilites elsewhere in Tas­
mania the albite displays low-temperature optics 
and varies in composition between Ab$_2$ and Ab$_3$. 
Augite occurs in only a few of the rocks, and is 
alterated to varying degrees to chlorite, or a very 
fine grained dusty aggregate of chlorite and possibly 
sericite (e.g. 30662a). The augite is similar to that 
described by Scott (1954) in the Lynch Creek 
basalts at Queenstown, this proved to be a diopsidic 
variety. Chlorite may be present, and in some of the rocks it is clearly pseudomorphing earlier minerals, 
probably augite (e.g. 642) and in some cases 
possibly olivine (e.g. 30662b). Most of the chlorite 
shows anomalous “Berlin Blue” interference 
colours and is very pale green in colour and non­
pleochroic. Refractive index measurements gave 
$\beta=1.615$ and $1.633$. 


The ground mass is very variable; it may have a typical basaltic form dominated by thin albite laths or it may be granular. Chlorite occurs interstitially to the feldspar laths, along with calcite, epidote, magnetite and ilmenite.

Some of the rocks are amygdaloidal, the amygdules being of ovoid shape and usually filled with combinations of chlorite, calcite and quartz. In slide No. 642, an amygdule nearly 1 cm. in diameter is filled with quartz, calcite and iron oxide, and rimmed by epidote crystals arranged radially. Calcite and chlorite veinlets are common and prehnite occurs rarely.

A specimen from a lava on the Magnet tram (30062b) consists entirely of euhedral chlorite pseudomorphs up to 2 x 2 mm., in a matrix of chlorite and quartz spherulites up to 0.2 mm. diameter. The chlorite appears to be pseudomorphing pyroxene and/or olivine crystals. Amygdule-like bodies in the ground mass are composed of aggregates of fine quartz spherulites (Plate II, Figs. 1 and 2). One of the "phenocrysts" which has a rim of quartz spherulites consists of chlorite and a fine grained quartz aggregate which appears to be partially replaced by a single skeletal quartz crystal.

Chlorite-quartz spherulite rocks occur in the Magnet Dyke and quartz spherulites line the rims of amygdules consisting of chlorite with cores of calcite in a basalt (636) from the West Magnet Mine. Rather similar spherulitic quartz rocks at Zeehan are assumed by Scott (1954) to be the result of silicification of basalt but the quartz-chlorite development appears to be related to a late stage of lava solidification rather than the regional metasomatism suggested by Scott (1954).

In all specimens collected in the field, albite is the only feldspar present, an observation similar to that made by Scott. However, slide No. 636 from the West Magnet Mine, has a coarse doleritic texture with almost unaltered labradorite (An 58) laths up to 2 mm. long. Mineralogical areas being occupied by pale green chlorite, calcite and magnetite (?). As already described, the rock is prominently amygdaloidal with chlorite, calcite and quartz spherulites. It appears to be a basalt, or a shallow intrusive, has many textural and mineralogical features in common with the spilites, and is unlike any Tertiary basalt. If it is Cambrian it gives an indication of a parental, more calcic melt.

Volcanic rocks with fragmental texture are not common, though some of those described as sandstones and breccias may in fact be pyroclastic. An interesting rock occurs near the head of the Arthur River (646). It consists largely of feldspar crystals, quartz spherulites rock, and basaltic fragments in a variable matrix of chlorite, iron oxides, and calcite. Part of the matrix consists of irregular veins and concretionary forms of a pale reddish-brown, isotropic mineral that is probably hydrossillimanite. Scott (1951a, 1954) has described a similar mineral in the King Island spilites, in which it appears to be due to hydrothermal alteration. This rock is of uncertain origin but may be the fine-grained equivalent of an autotlastic breccia (Wright and Bowes, 1963), formed by gas explosions within lava during solidification.

The dominance of K-feldspar in the greywackes interbedded with spilites is surprising in view of the lack of any feldspar other than albite in the spilites or the keratophyres. The reason is not known but it is suggested that the microcline in the sediments is derived from potassic rhyolite or quartz keratophyre flows and necks that have been largely disintegrated by explosive activity during eruption.

(b) CAMBRIAN INTRUSIVES

Basics and Ultrabasics

Peridotites, pyroxenites and serpentinites occur extensively in a wide belt to the southwest of the Waratah District and small masses of these rocks are also present in the Waratah District. Numerous boulders of weathered ultrabasic rocks occur north and south of the Arthur River Dam on the Waratah-Corinna Road, and two small masses occur at Mt. Bischoff. These rocks are similar to other ultrabasic masses in Tasmania, e.g., at Adaminaby, Anderson Creek and Argent Tunnel, and probably belong to the same tectonic phase.

With the spilites and albite gabbros these form a typical ophiolite association.

A number of gabbroic bodies intrude the Cambrian sediments west of the Magnet Mine. A typical example crops out ¼ mile east of the Whyte River bridge on the Corinna Road. It is composed almost entirely of plagioclase, hornblende and chlorite (in order of abundance) in an even grained granitic texture, the average diameter being 0.5 mm. The feldspar is loaded with inclusions (?) of yellowish-brown chlorite and sericite (?) and many of the laths possess a thin clear rim in optical continuity with the cloudy core. The plagioclase is albite or albite-oligoclase. The hornblende is somewhat fibrous and the crystals tend to be ragged. It appears to replace both augite (?) and yellowish chlorite. Its optical properties are α = 1.643, β = 1.660, 2V = 73° ± ve.

Nye (1923) refers to similar rocks west of the Whyte River some of which contain augite. Specimen 586 shows pyroxene (diopside ?) in all stages of alteration to hornblende and pale green chlorite.

Magnet Dyke

A long, narrow strip of igneous rock occurs on the contact of the Bischoff and Cambrian (?) sediments west of the Magnet Mine to the
Persic Mine, a total distance of some 5 miles. This mass of igneous rock has occasioned some controversy, Twelvetrees (1900) and Nye (1923) considering it to be a complex dyke, and Scott (1954) suggesting on petrological evidence that it represents a strip of volcanic rocks.

The "Dyke" is generally some 200 to 300 feet in width, reaching its maximum thickness at the Magnet Mine where it is some 1000 feet wide. At the mine it has a west dip and is more or less conformable. It consists over most of its length of a single, rather variable rock type which has been previously named a diabase porphyrite. At the Magnet Mine, however, the section is more complex. Twelvetrees (1900) describing the "Dyke" from east to west as websterite porphyrite, diabase porphyrite, and orbicular or spheroidal websterite (Fig. 6).

The diabase porphyrite, which extends for the whole length of the "Dyke", is extremely variable, with albite, pyroxene, chlorite, siderite, quartz, hornblende, pyrite and ilmenite occurring in varying proportions. Commonly the rock is porphyritic with phenocrysts of altered albite, and also chlorite completely or partially pseudomorphing pyroxene. Albite porphyrite is probably a more apt term than diabase porphyrite. The albite, which occurs as multiply twinned and untwinned crystals, has been extensively sericitized. The ground mass is generally very fine and consists dominantly of albite laths and chlorite with irregular masses of quartz and carbonate, the latter also commonly occurring as veins throughout the rock. Quartz occurs rarely as large anhedral crystals associated with chlorite. Small crystals of ilmenite, sphene, columnar epidote and cubes of pyrite also occur throughout the ground mass. Circular masses of carbonate, chlorite and quartz spherulites are also common in some sections, and are similar to amygdale fillings of the Cambrian (?) basalts. The circular bodies also show similar zonation to those in the basalts, with a carbonate nucleus lined by sheaves of chlorite, and quartz spherulites forming an outer margin.

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Fig. 6.—Section across the Magnet Dyke at the Magnet Mine (after Nye, 1923).
Although the albite porphyrite is dominantly porphyritic, some sections exhibit a doleritic texture. Specimen 35 N 14 (Mines Department), from the Waratah-Pleman Track, consists of interlocking laths of albite and augite with a very minor proportion of ground mass, and appears to be an altered dolerite. Specimen 712 has a similar texture and composition but also contains some hornblende and actinolite. Several of the albite crystals show myrmekitic texture.

In places the albite porphyrite consists entirely of quartz and chlorite, e.g., 708 and 35 F 11 (Mines Department). This rock consists of large spherulitic masses of quartz with interstitial sheaves of chlorite, and has been called a silicified variolite (Rosenbusch in Nye, 1923). Analyses of these "variolites" are given in Nye (1923) and quoted by Scott (1954) as intermediate phases of the silification of splitite, the end product being a spherulitic quartz rock (99.1% SiO₂). It is suggested here that the siliceous rock is a primary, or deuteric, feature confined to lenses and channels in the splittie body.

The pyroxenite on the east of the "Dyke" at the Magnet Mine is reported to occur as a lenticular body with a maximum width of 360 feet and thinning out to the north and south. The rock is extremely weathered, but resembles a decomposed basic rock with large pyroxene phenocrysts. A section of this rock described by Nye (1923) contained dominantly serpentine with a little chlorite. The name websterite porphyrite was applied by Rosenbusch who considered that in its unaltered state, the rock would contain phenocrysts of hornblende or enstatite in a ground mass of pyroxenes. Specimen 637 from North Magnet contains large phenoecysts of enstatite that is largely altered to talc (?); the ground mass is composed of talc (?), sericite, quartz and chlorite. A similar rock (2183—allotted to be from the northern slopes of Mt. Bischoff) consists of enstatite crystals surrounded and in part pseudomorphed by talc (?). Fine anhedral quartz crystals and small fibres of chlorite are also present. This specimen probably represents part of the pyroxenite mass on the southern margin of the Magnet Dyke at the Persic Mine.

On the western side of the albite porphyrite at the Magnet Mine, there lies a decomposed pyroxenite. It is weathered to a brownish-yellow rock which contains decomposed crystals of pyroxene, similar to those of the eastern mass. In places this rock has an orbicular structure, containing abundant spheres with the same mineralogical composition as the general mass of the rock. The occurrences of the pyroxenites bounding the albite porphyrite appears to be limited to the Magnet and Persic Mine areas, the only areas of mineralisation along the "Dyke". Several of the specimens labelled as "Dyke" at the Magnet Mine is the inclusion of a large block of the Bischoff quartzites between the easterly pyroxenite and the albite porphyrite and a smaller block within the pyroxenite itself (Nye, 1923).

The origin of the "Dyke" is uncertain. Several features indicate an intrusive origin: (a) the complex nature of the "Dyke" at the Magnet and Persic mines, (b) the distribution of the "Dyke" along 5 miles of the Cambrian (?)-Precambrian contact and along what appears to be a discordant structure at the mine, (c) the composition of the pyroxenite, and (d) the inclusions of Bischoff quartzites within the "Dyke".

However, as Scott (1954) has demonstrated, there is a strong petrographical similarity of a large portion of the albite porphyrite with the splittie in the Magnet area and she concluded that the rock was volcanic. Also, Cambrian (?) volcanic breccias are apparently found adjacent to the "Dyke", specimens 35 N 8 and 35 N 22 from North Magnet and the Persic Section (labelled websterite and diabase porphyrite respectively) being breccias composed almost exclusively of volcanic fragments. The writers agree with Scott on the nature of the albite porphyrite but suggest the pyroxenites are intrusive.

Alteration of the Cambrian Lavas and the Magnet Dyke

These rocks have undergone a pattern of alteration that is characteristic of Cambrian volcanics elsewhere in Tasmania, generally involving the processes of albitization, chloritization, carbonitisation and silification. The albite in amygdules and veinlets appears to be related to a late phase of the crystallisation of the flows, see also Solomon, 1960) and probably the chlorite, quartz and carbonate associated with the albite are also of similar (deuteric ?) origin. The cloudy albite forming phenocrysts shows no textures indicating its development by replacement of calcic feldspar and it appears to have crystallised at an early stage of lava solidification. The labradorite from the West Magnet mine shows no replacement by sodic feldspar but it does indicate the presence of at least some low-soda magma in the area.

The presence of albite and augite in ophitic texture in the King Island lavas (Scott, 1951b) is further support for a primary origin and a strong case can be made for the presence of a Na-rich magma during Cambrian volcanic activity.

Bradley (1954) considered the alteration of the Cambrian volcanics to be of Devonian age but it is likely to be Cambrian because similar intrusions in Tasmania it is likely to be mid-Devonian to Carboniferous in age. To the north of this granitic mass are dykes of quartzfeldspar porphyry which intrude the Cambrian (?) sediments and appear to be related to the granite. Quartz-feldspar porphyries and topazised quartz...
porphyries occur at Mt. Bischoff as a series of anastomosing dykes and sills, one of which apparently extends 3 miles north-east into Deep Creek (the granite mass indicated on the State Geological Map in Deep Creek appears to be a quartz porphyry dyke).

**Adammellite**

Two types of adamellite occur in the area examined; an even, medium-grained adamellite and a porphyritic adamellite, the porphyritic variety previously being considered as a late stage intrusive into the even grained granodiorite (Reid, 1923).

The even grained variety (30651, (e), (f) and (g)) consists of orthoclase, oligoclase (approx. An95), quartz and biotite as intergrown, generally euhedral crystals from 1 mm. to 3 mm. in length. Graphic interwroths of quartz and orthoclase are common. Other minor components of the adamellite include hornblende, zircon, topaz and tourmaline. Modal analyses of three thin sections of this rock averaged 54.5% total feldspar, 36.1% quartz, 9.0% biotite and 0.5% accessory minerals. Some 20% to 30% of the total feldspar is oligoclase and the rest orthoclase.

The porphyritic variety is similar but contains large phenocrysts of oligoclase up to 2 cm. in length which in places show slight zoning. Modal analyses on four thin sections of this rock averaged 55% total feldspar (including 9% oligoclase phenocrysts), 34% quartz, 11% biotite and accessories. Thus the porphyritic and even-grained adammelites are compositionally similar and are probably variations within the granitic mass rather than distinctive bodies. This is supported by the field distribution which appears to be irregular and gradational.

Aplitic phases, consisting almost exclusively of quartz with minor orthoclase, contain large irregular clots of black fibrous tourmaline which constitute the greater proportion of the rock in places. Contact effects of the granite were not seen for lack of exposure.

**Associated Quartz-Feldspar Porphyries**

A dyke of quartz-feldspar porphyry extends north from the adammellite and outcrops between the junction of Seven-mile Creek and the Arthur Falls. It consists of large phenocrysts of orthoclase and quartz in a fine-grained ground mass of orthoclase, quartz, a little plagioclase and a ferromagnesian mineral which is probably hornblende. Associated with this rock is a more basic type which macroscopically appears to consist of phenocrysts of feldspar in a ground mass of feldspar and probably hornblende. The latter rock appears to occur mainly along the margins of the main quartz-feldspar porphyry dyke.

A further acid dyke probably occurs some 200 yards north of the Magnet Tram bridge over the Arthur River. The rock is extremely weathered, feldspar being the only recognizable mineral in a dominantly limonitic ground mass. Both dykes intrude Cambrian (?) sediments and contact effects are limited to slight baking for a few feet from the contact.

**Quartz Porphyries at Mt. Bischoff**

The most prominent feature of the geology of Mt. Bischoff is the profusion of altered porphyry intrusives within the Precambrian sediments. The majority of the porphyries occur as dykes from fifteen to one hundred feet in width, forming an overall radial pattern with preferential trends east-west and north-south, and steep dips to the north and west respectively. Large irregular masses also occur, for example on the summit, and to the north-east of Mt. Bischoff and also at Don Hill, the first two being marginal points of several large dykes. Smaller, discordant intrusions of porphyry are common throughout the area, the majority apparently being offshoots from the main dyke system. Relatively thin sills of porphyry are also common, particularly at the junction of dolomite and Footwall shales (e.g., in the Brown Face and Slaughteryard Face). Alteration of the country rock by the porphyry causes thinning and slight baking for several feet from the contact.

Breccias composed of country rock fragments are consistently present on the walls of the porphyry dykes, the breccias varying from 2 to 26 feet in width. The fragments are angular or subangular, up to 2 inches in diameter, and are mainly in contact with each other. The matrix is largely crushed sedimentary material and fine grained porphyry. The contact between porphyry and breccia is irregular with small lenticular "splashes" of porphyry (both microscopic and macroscopic) extending into the breccia far several feet from the contact. Xenoliths of baked country rock are also common within the porphyry adjacent to the contact. In several cases, the structure as defined by the dolomite horizon is difficult to correlate across the dykes and in the case of the Western Dyke it is evident that there has been displacement along the line of the intrusion. This evidence, together with the occurrence of contact breccias of quite large dimensions, indicates that the porphyries were intruded into tensional openings containing fault breccias, and discounts the possibility of extreme brecciation during emplacement of the porphyries.

The matrix of the breccias also exhibits strong macroscopic bedding, cross-bedding structures and small swirls which probably reflect flow movements in the partly crystalline porphyries during intrusion.

Within the general category of quartz porphyry there are several gradational varieties which depend on the subsequent alteration of a primary porphyry type rather than variations within the original parent magma. Remnants of unaltered porphyry occur in the southern portion of the White Face Dyke and Falls Creek, and are recorded by Reid (1923) from the western extremity of the Brown Face Dyke. Analyses of unaltered porphyries are shown in Table 3 (Nos. 1 and 7).

The porphyry from the southern portion of the White Face Dyke (e.g., 30617) consists of bipyramidal quartz phenocrysts up to 5 mm. in diameter with lath-shaped phenocrysts of a soft white substance in a fine ground mass. The laths are orthoclase or albite, and are commonly altered to clay minerals (mainly kaolinite (?); X-ray powder photographs show strong similarities to orthoclase pattern but most of the lines are offset equal amounts in the
same direction, presumably due to alteration. D.T.A. curves confirmed the presence of kaolinite. The groundmass is largely quartz and sericite. Muscovite occurs sporadically throughout the rock as flakes up to 0.5 mm. in length. A modal analysis of a thin section indicated 59% groundmass, 25% quartz phenocrysts, 14% orthoclase phenocrysts and 2% muscovite.

The majority of the porphyries are topazised and contain only an insignificant proportion of original feldspar. These rocks are typified by specimens 30619 to 30629, 2173, 692, 693, 698, 700 and 702, an analyses of typical samples are given in Table 3 (Nos. 2 to 6). The porphyries consist of a fine-grained ground mass of dominantly quartz and topaz with sporadic occurrences of sericite and fine threads of talc. Small crystals of fluorite and tourmaline are rarely present. The prismatic topaz has been referred to as pyctime (Reid, 1923). The topaz has refractive indices $\alpha = 1.620$ $\beta = 1.610$ and $2V = 62$ indicating it has a molecular percentage $OH/(OH+F)$ between 6 and 4 (Deer et al., 1962). High fluorine content is indicated in the partial analysis given by Reid, 1923. Quartz phenocrysts are abundant, generally ranging in diameter from 0.5 mm. to 2.0 mm. and in some instances reaching 5.0 mm. These phenocrysts are perfectly formed hexagonal crystals near the margins of the dykes but within the dykes are generally slightly corroded or possess a narrow rim of quartz in optical continuity with the phenocrysts, the outlines of which are still visible. Fluid inclusions are present, and in some instances common, in the quartz which also contains minute inclusions of zircon, muscovite and topaz. The portions of the dykes with non-corroded crystals represent the chilled marginal material. Lath-shaped cavities are also evident, these generally being lined with groups of radiating acicular crystals of topaz growing inwards from the cavity walls; they were probably feldspar phenocrysts (Plate II, Fig. 3). Topaz is also evident as acicular crystals associated with irregular aggregates of anhedral quartz crystals in optical discontinuity, and as partial rims around corroded quartz phenocrysts. Sulphides are generally abundant, occurring as finely crystalline particles; many occurrences form lath-shaped aggregates, probably replacing feldspar (Plate II, Fig. 4). Pyrite is the dominant sulphide mineral although arsenopyrite occurs sporadically throughout the porphyries, and pyrrhotite is common in porphyry sills on the contact of replaced dolomite and slate. Finely crystalline sulphides are also found associated with the quartz-topaz ground mass. Oxidation of these sulphides has occurred near the surface and accounts for the high proportion of $Fe_2O_3$ in several of the analyses. Cassiterite is also present in the porphyries as fine granular crystals replacing feldspar laths and rarely as well-formed zoned crystals up to 0.5 mm. in diameter.

Table 3

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Total .... 100.20 99.77 99.94 100.39 99.86 99.94 99.64

Localities—
1, 2, 3—White Face Dyke.
4, 5, 6—Western Dyke.
7—White Face Dyke.
Analyst for 7.—A. D. Mackay (in Reid, 1923).
Unusual varieties of porphyry occur along the North Valley road (30630 a and b). These are quartz-topaz porphyries but contain abundant rosettes of muscovite in lath-shaped aggregates. A further variety (30631) contains a higher proportion of pyrite cubes, several of which enclose quartz phenocrysts, thus indicating crystallization after the quartz. Large phenocrysts of muscovite are also common in this rock. Muscovite is also common in a porphyry dyke on the contact of the Bischoff sediments with Cambrian (?) sediments to the west of Waratah (30633); this porphyry consists dominantly of sheared quartz phenocrysts with elongate aggregates of sericite in a fine quartz-sericite ground mass. It is unusual in that it has been sheared.

The White Face Dyke (see Fig. 4) was examined in some detail as it provides an opportunity of studying the gradation from a quartz-feldspar porphyry to a quartz-topaz porphyry. It was sampled closely from just south of the open cut northward into White Face, i.e., from unaltered to intensely altered material. The unaltered quartz-feldspar porphyry grades into a rock with a felsitic ground mass and idiomorphic quartz phenocrysts and lath-shaped masses of siderite, pyrite and fluorite with some remnants of orthoclase. Biotite and muscovite are also present, generally being associated with talc as minor constituents of the ground mass. This rock then grades into a quartz-topaz porphyry. The intermediate zone described above is only a few feet wide. Analysis 1 (Table 4) is of unaltered porphyry and analyses 2 and 3 are of quartz-topaz porphyry. Analyses 4, 5, and 6 are from the Western Dyke, taken at intervals of 2-300 ft., No. 4 being from the old Haulage, No. 5 near Slaughter-yard Face and No. 6 just south of Mt. Bischoff peak. Little variation is noted although potassium is more abundant in No. 6 which is the most westly sample, and may indicate the occurrence of minor original feldspar away from the main centre of mineralization.

The occurrence of a quartz-feldspar porphyry towards the limit of mineralization at Mt. Bischoff and the consistent occurrence of topaz, fluorite, sulphides and cassiterite in lath-shaped cavities and in the ground mass is strong evidence that the latter minerals are present as pseudomorphs after feldspar during alteration. This was also suggested by Dunn (1922), who considered that the alteration took place homogeneously in a crystal mush prior to emplacement. If this hypothesis were correct, it would be expected that the proportion of topaz and other replacement minerals would be constant throughout the dyke. Sections of porphyry taken across the Western Dyke were examined, 30625a and d being marginal samples and 30625 b and c samples from near the centre of the dyke. Modal analyses of these sections (as percentages) are given below. One thousand counts were made on 4 sq. cm. to determine the phenocryst-ground mass ratio. 1,000 counts for the same area for the composition of the ground mass and 1,000 counts for the composition of the phenocrysts.

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<tr>
<th></th>
<th>30625a</th>
<th>30625b</th>
<th>30625c</th>
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<tr>
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<td>1.0</td>
<td>1.0</td>
<td>1.0</td>
</tr>
<tr>
<td>Total phenocrysts</td>
<td>28.6</td>
<td>17.6</td>
<td>31.0</td>
<td>27.4</td>
</tr>
</tbody>
</table>

Phenocrysts—

The occurrence of muscovite, fluorite, siderite and talc pseudomorphing feldspar in place of topaz at the extremities of the mine area, and the occurrence of unaltered quartz-feldspar porphyry near Don Hill and in Falls Creek, suggest that a temperature gradient may have existed in the area with extensive topazisation in the areas of highest temperature towards the centre of Mt. Bischoff.

In places, alteration has been so extensive that the original texture of the porphyry no longer exists and the rock can only be distinguished as an alteration product by a gradation into recognizable porphyry. In the White Face, along the walls of the dykes, the topaz has replaced even the quartz, producing spherulitic aggregates of topaz in association with euhedral to anhedral crystals of casseriterite, many of which show marked zoning (e.g. 694 and 698). The spherulitic topaz comprises a central, granular aggregate surrounded by radiating prisms (Plate II, Fig. 4) and needles of topaz, and is rarely associated with radial blue tourmaline crystals.

The porphyries have also been totally replaced in the White Face and near the Main Tunnel entrance by tourmaline, fluorite and siderite. These rocks (30634 a to f) generally comprise large crystals of fluorite several mm. in length, which contain inclusions of casseriterite and siderite, these being bounded in places by sheaf-like masses of muscovite, producing a halo effect. The fluorite is completely surrounded and "intruded" by aligned acicular crystals of tourmaline which exhibit marked pleochroism from blue to colourless. Coarse crystals of siderite also occur and themselves riddled with a profusion of acicular crystals of tourmaline. In specimen 30634 d moderately thick veins of sheaf-like muscovite occur in association with the tourmaline.
The tourmaline is dark green in hand specimen and has refractive indices $\alpha=1.6580$, $\beta=1.6325$ ($\delta=0.0255$).

An analysis given by Reid (1923, p. 55) is as follows: SiO$_2$: 36.86, Al$_2$O$_3$: 36.72, Fe$_2$O$_3$: 5.66, MgO: 3.92, MnO: 0.66, CaO: 0.34, Na$_2$O: 3.57, K$_2$O: 1.11, B$_2$O$_3$: 10.56, F: 0.61; total: 100.01. Though no H$_2$O is given the analysis indicates a composition somewhere between a schorl and dravite type and this matches the optical properties.

The occurrence of these totally altered porphyries is limited and is probably due to very local influxes of pneumatolytic vapours or solutions, particularly on the margins of the porphyry dykes.

In summary, the sequence of events at Mt. Bischoff is considered to have commenced with the diapiric injection of a narrow cupula of granite material and to the hinge of the Bischoff Anticlinorium, the cupula probably being connected to underground extensions of the Meredith Granite (Fig. 3). The diapiric upthrusts produced tensile stress in the roof sediments with the formation of tension faults splaying outwards from the focus of injection. These tensional openings were then intruded by quartz porphyry dykes, several of which protracted along selective stratigraphic horizons as sill-like extensions. Associated with the emplacement of the porphyries, pneumatolytic vapours or solutions rose up from the cupula along the open channel-ways of the marginal brecciated walls of the porphyries, producing widespread replacements of primary feldspar by topaz, tourmaline, fluorite, sericite and cassiterite and finally, by pyrite, arsenopyrite, pyrrhotite and cassiterite.

As recognized by Twelvetrees and Petterd (1897) the quartz porphyry dykes are similar to the elavns of Cornwall, England. In this area, folded Palaeozoic slates and sandstones, and later granite stocks, are intruded by greisenized quartz porphyry dykes (MacAllister, 1908; Lindgren, 1933; Llewellyn, 1946; Dewey, 1948; and Dunham, 1952). The tin deposits are limited areally to the foci of intrusion with copper and lead-zinc lodes in the surrounding area. Greisenization has been extreme in the Cornwall deposits, the granite along the lode walls being altered to quartz, muscovite and topaz as aggregates pseudomorphing partly altered feldspars. Fluorite is present in places and the secondary quartz is filled with liquid inclusions. The quartz porphyries have also been altered to quartz, tourmaline, topaz and fluorite with kaolin locally. The alteration is limited to areas of mineralization and is not widespread through the granite.

Kaolinization of the granite has occurred over quite extensive areas and is considered a relatively low temperature effect, as in the marginal zones at Mt. Bischoff. In the porphyries there has been an addition of iron, titanium and bornite at Mt. Bischoff but potassium remains high, probably due to the formation of muscovite. It is probable that similar hydrothermal and pneumatolytic alteration processes occurred both at Mt. Bischoff and Cornwall, indicating a similar granitic source.

**D. I. GROVES AND M. SOLOMON**

**DOLERITE**

Two bodies of Jurassic dolerite intrude Cambrian (?) sediments in the area. The largest occurs north of the Magnet Mine and is a dyke-like mass extending for about 1500 feet in a north-easterly direction. The second mass is a small intrusive just west of the Magnet Tram, to the north of the contact of basalt and Cambrian (?) sediments. Thin sections of these rocks (715 and 35 G.2-Mines Dept.) show augite and labradorite exhibiting an ophitic to doleritic texture, the individual crystals occurring up to 1.5 mm. in length. Between the larger crystals is a fine ground mass of quartz and feldspar with some fine interstitial magnetite.

The dolerite is identical to phases of the dolerite sills intruding Permian and Triassic rocks in south-eastern Tasmania.

**TERTIARY ACTIVITY**

**BASALT**

A basalt sheet, some 50 to 150 feet thick, occurs extensively over the plateau area between Waratah and Guildford, on the Magnet Range, and as small remnants elsewhere in the area. It generally overlies Tertiary terrestrial sediments, the base being at an elevation of 1900 to 2000 feet.

The basalt (30652 and 35 G.6 Mines Department) generally consists of a fine intergrowth of labradorite and augite, up to 0.3 mm. in length, with interstitial calcite and magnetite. Olivine is generally present, occurring as larger crystals up to 2 mm. in diameter that have been slightly serpentinitized. Tachylite also occurs in places, specimen 35 G.7 (Mines Department) consisting of large particles of green glass, up to 6 mm. in length, with a ground mass of labradorite, calcite and augite. In places the basalt is extremely vesicular and in some cases is scoriaceous.

Three thin basaltic dykes occur in the Arthur River intruding Cambrian (?) sediments; one about 1½ miles upstream and one 2 miles downstream from the confluence with the Waratah River, and one about 1½ miles upstream from the old Magnet Tram bridge. These may represent ‘feeders’ for the main basaltic sheet. A specimen from the basaltic dyke near the Magnet Tram (30652b) contains phenocrysts of labradorite in a ground mass of labradorite, purplish-brown augite and magnetite. Magnetite also occurs abundantly as elongate blebs in the cleavage cracks of the labradorite phenocrysts. Amygdules of chlorite and a fibrous zeolite are also present. Olivine is apparently absent.

**STRUCTURE**

The dominating structural feature of the district is the E-W trending Bischoff Anticlinorium within which the Precambrian rocks outcrop. This structural trend is unusual in NW Tasmania, where the majority of both large and small scale structures trend NW or NNE (Carey, 1953; Solomon, 1962).

The Waratah district is supposed to lie between early Palaeozoic geanticlinal ridges, the margins of which trend about NNE in NW Tasmania. Early Tabberabberan folding is thought to have been
controlled in trend by these pre-existing surfaces, and later folding, on NW trends, is thought to be due to some "external" cause (Solomon, 1962).

The E-W structures may reflect some much earlier deformation, the trends of which have locally altered the dominant Tabberabberan directions. A pre-Cambrian deformation is indicated by the microscopic fabric in the Precambrian (?) sandstones and by other features to be described.

**Devonian Structures**

**Folding**

The form of the Bischoff Anticlinorium is indicated by the map of the Precambrian inlier (Figs. 1, 2 and 3). Its wavelength is approximately 5 miles and its amplitude approximately 2 miles. It plunges gently to the west and probably also to the east, to form an elongate dome. The contact between Bischoff quartzites and slates and the Cambrian (?) sediments on the southern flank shows apparent concordance, with minor faulting, while the northern flank is largely occupied by the Magnet Dyke which extends along a possible fault contact between the two sequences. The main zone of porphyry dykes is concentrated in the axial surface of the anticlinorium and it is suspected that a granite cupola exists beneath Mt. Bischoff.

Smaller folds with wavelengths of 100 to 1,000 feet occur on the anticlinorium; the majority have been mapped by using the base of the dolomite as a marker horizon. These folds have been outlined fairly accurately in the Bischoff mine workings and here they are typically associated with sub-longitudinal tensional faults that tend to obliterate limbs of the folds. In the Brown Face syncline the faulting is slightly oblique to the fold axis and as a result the dolomite base has a "keel" form. The folds of the Bischoff mine have very shallow plunges, dominantly to the west. Previous interpretations of the plunge of these folds (Knight, 1953, Hall & Solomon, 1962) have not recognized the importance of slightly oblique faulting producing keel-like structures. The exposed extent of these folds is limited by NNW faults, almost transverse to the fold axes (Fig. 4).

The Main Tunnel of the Mt. Bischoff mine extends from the north to the south side of the mountain and, although partly filled in, it provides a sub-surface cross section through a zone of minor folds on the anticlinorium. Mapping of the tunnel shows fairly gentle undulations below structures which at the surface have considerable amplitude, indicating that the minor folds on the anticlinorium die out within depths approximating to the wavelengths of the folds.

**North East Folds**

Distorting this general W trend of folding are smaller folds which show a variable trend, with a distribution of axial surfaces around 60°M. These NE folds and also the larger W structures are highlighted by the plot of axial surfaces shown in Fig. 7.
Fig. 7a, also by the plot of several hundred bedding readings taken between Don Hill and the Arthur-Waratah River confluence (Fig. 7b). These folds which vary in wavelength from 2 to 100 feet, vary in plunge from northerly to southerly. The axial surfaces generally dip towards the SSE on the southern limb of the anticlinorium and towards the NNW on the northern limb, although there are minor variations. The folds are restricted in exposure to areas of thin-bedded shales and quartzites and are abundant in the Mt. Bischoff mine where they are so intense as to almost obliterate the W folding. Small folds belonging to this group are also evident in thin-bedded mudstones of the Cambrian (?) sequence. The folds display all variations from a concentric to a similar style of folding, although a concentric style is dominant.

A small group of folds, of similar wavelength and style to the previously mentioned folds, occur with a NNW to NW trend. A large proportion of the folds are restricted to areas adjacent to NNW to NW faults, and were probably produced by local fault movements.

Generation of Folds

There is no reliable indication of the order of super-imposition of the two main fold generations, though it is suspected that the small wavelength folds are later than the larger.

Faulting and Jointing

The majority of known Devonian faults appear to be tensional and either of W or NNW trend. A plot of the poles of faults other than of W trend indicates a variation in strike from between 330° and 350°, with the majority at 340°. These faults dip fairly consistently to the west at high angles, though some are vertical and a few dip very steeply east. An important fault of this group displaces the minor W folds on the Bischoff anticlinorium. Other parallel faults cut the porphyry dykes and are mineralized, indicating a post-intrusive age. In a few, shearing of the one indicates later movement on the pre-existing faults.

Well developed joints, with 6 inches to 2 feet spacing, strike consistently at 340° and these joints in many cases carry quartz and cassiterite. They occur in porphyry dykes and in both Cambrian (?) and Precambrian sediments. They appear to be later than the folding and dyke intrusion and related to the mineralization phase.

The normal faults accompanying folding, with longitudinal or slightly transverse trend, have already been described.

As previously mentioned the porphyry dykes were probably injected into a set of radial tension fractures formed by rise of a cupola-like mass into the Bischoff Anticlinorium. Some movement along these fractures is indicated by the presence of brecias and by displacements of marker horizons across the dykes. The displacements are estimated to be less than 100 feet. Fissure veins several feet in width occupy tension fractures, some of which appear to be related to the radial porphyry pattern and others to the 340° trend.

The displacements developed by the north-south folds inferred by Knight (1953) appear to be due to movement on the 340°M faults.

Pre-Devonian (?) Structures

A number of small scale folds and thrusts are confined to the Precambrian sediments and may be related to some pre-Tabberabberan tectonic phase. Isocinal folds with wavelengths of several feet are confined to relatively thin-bedded quartzites of the Bischoff quartzites and slates. The folds trend 80°M to 90°M and occur within the limbs of the minor, E-W trending folds. The folds plunge shallowly both to the east and west, no steeply plunging folds being observed.

Chevron folds, with a wavelength of 1 to 10 feet are limited to the marginal zones of the Precambrian (?) core, occurring commonly in the Arthur River and Magnet Creek near the confluence of the two streams, and more rarely on the southern contact of the south-east slopes of Don Hill. Up to twenty chevron folds with subparallel vertical axial surfaces have been observed in a single outcrop in the Arthur River area. The folds, which trend between 40°M and 50°M and plunge steeply to the south-west and north-east, occur exclusively in thinly bedded quartzites and shales.

Small recumbent folds, generally with a wavelength of approximately one foot, occur sporadically throughout the area. They are locally dependent on the oriention of the Precambrian sediments and may represent some pre-Tabberabberan tectonic phase. Isoclinal folds with wavelengths of several hundred bedding surfaces generally dip towards the south-west and north-east, occur exclusive in thinly bedded quartzites and shales.

Small monoclinal ‘wars’ are a distinctive feature of the deformation of the Bischoff quartzites and slates. They occur on a small scale throughout the area, generally in shales and rarely in quartzites. The axial surfaces of the monoclinic bends are locally dependent on the orientation of larger folds and are variable throughout the area.

At this stage it is not possible to be certain of distinguishing between pre- and post-consolidation structures, particularly in rocks with a lack of cleavage associated with the folds.

Insufficient structural work has been done to firmly date the age and sequence of development of this group of folds. The quartz microfabric in the Bischoff quartzites indicates an early movement phase and this is further indicated by the presence of isolated blocks of quartzite within shales in parts of the Bischoff succession. These blocks are generally elongate parallel to bedding and taper at their extremities with shales “flowing” around and enclosing the blocks. In most cases it has been impossible to determine whether these blocks represent fold mudlons or boudins, but in some cases they appear to be part of a fragmented bed of quartzite.

Tertiary Faulting

The Tertiary terrestrial sediments are generally horizontal throughout the area although on the south-western flank of Don Hill they dip to the south-west at angles up to 50°; these steep dips,
and associated folds, may be due to late Tertiary faulting but are more probably due to compaction of the Tertiary sediments against the bedrock cliffs.

The linear nature of the Waratah River, as described in a previous section, is suggestive of a possible Tertiary fault trend.

**MINERALIZATION**

Mt. Bischoff is situated in a strip of dominantly low grade silver-lead-zinc mineralization which trends north-north-easterly and lies on the northwest flank of the large granodiorite mass centred at Mt. Meredith. Tin mineralization occurs within the granitic mass to the south of the Waratah-Corinna Road, in association with quartz porphyry dykes at Mt. Bischoff, and adjacent to reported quartz porphyry dykes at the Cleveland Mine.

**Tin Mineralization**

Mineralization is confined to aplite and pegmatitic bodies within the main granodiorite mass. The ore bodies are small and largely lenticular in shape and they comprise arsenopyrite, molybdenite, pyrite, chalcopyrite, sphalerite and cassiterite in a gangue of green, fluorine-bearing mica. Quartz and tourmaline are commonly present in association with the ore. The lodes do not appear to have any distinct structural control.

**Mt. Bischoff**

At Mt Bischoff, tin mineralization has been more localised and more intensive than in the granodiorite. The tin occurs in a number of widely differing types of ore body, which are limited to a circular area of 2000 feet radius from the centre of the mountain. The ore bodies comprise (a) a large replacement body in dolomite, (b) numerous vein deposits, (c) replacements of porphyry and (d) incrustations on joint faces, in and adjacent to porphyry dykes.

(a) The main ore body was formed by replacement of the dolomite bed which occurs towards the top of the Bischoff quartzites and slates. This bed, which was probably folded prior to mineralization, has been largely replaced by pyrrhotite, pyrite, talc, quartz and an iron-manganese magnesium carbonate. Typical analyses of this secondary carbonate are given in Table 4. Within the replaced dolomite are semi-continuous layers of apparently sheared talc which are possibly secondary after tremolite. This alteration probably results from metasomatic alteration of the dolomite along shear zones.

At the eastern end of the Brown Face open cut is an outcrop of dark green and yellowish rock that is composed of chondrodite partially replaced by serpentine (chrysotile?) and pyrrhotite. It is colourless and non-pleochroic, has 2V=72+ve, κ=1.615, ω=1.693, and is twinned on {001} and also on other composition planes (not yet determined). The identification of chondrodite was confirmed by X-ray diffraction.

The distribution of mineralization is extremely irregular within the body, some sections comprising massive sulphides while others are virtually sulphide free. The distribution of the sulphides themselves is more regular, pyrrhotite being the dominant sulphide towards the centre of mineralization, while pyrite and sphalerite are more common to the north and south. Mineralization was apparently confined to a relatively small area, as unmineralized dolomite crops out to the south in Happy Valley and to the north, on the northern slopes of the Waratah River valley.

The tin, dominantly as cassiterite, occurs throughout the replacement body and although microscopically associated with gangue minerals is most abundant in zones of high sulphide concentration. These zones are commonly slightly oblique to the margins of the dolomite bed and occur in different structural environments throughout the mine area. In general they occur within south-dipping limbs of the dolomite with local concentrations adjacent to fault walls. The zones of highest concentration may represent lode channels for the mineralization.

**Table 4**

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<td>SiO₂</td>
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<td>FeS</td>
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<td>SO₃</td>
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---

1. From Happy Valley Face (Reid, 1923, p. 86).
2. From Mt. Bischoff (Reid, 1923, p. 86).

Stillwell (1945) has divided the ore into three dominant types: (i) carbonate-sulphide ore, (ii) massive pyrrhotite ore and (iii) pyrrhotite-talc ore.

(i) The carbonate-sulphide ore consists essentially of coarse grained carbonates, grain diameter up to 5 mm., containing irregular splashes and veinlets of sulphides, dominantly pyrrhotite, pyrite and arsenopyrite. Small blebs of chalcopyrite and galena and crystals of cassiterite occur rarely in the sulphides. Small clusters of cassiterite up to 0.25 mm. in diameter occur more commonly in the coarse carbonate, and are partially coated with a thin film of chalcopyrite and stannite.

(ii) The massive pyrrhotite ore is dominantly pyrrhotite with colloform masses of pyrite and a little marcasite. Irregular threads of chalcopyrite, bismuthinite and galena are also present, and Stillwell reports minute grains of stannite (0.01 x 0.005 mm.) occurring in chalcopyrite veins. Minor proportions of quartz, fine platey talc, fine carbonate, fluorite and zeolite are embedded in the pyrrhotite.
(iii) The talc-pyrrhotite ore comprises alternate bands of pyrrhotite and talc up to 2 mm. wide. The pyrrhotite bands are crossed at irregular intervals by carbonate and pyrite veins, and contain rare cassiterite crystals. Small chalcopyrite veinlets occur on the boundary of pyrrhotite and talc masses.

(b) Vein deposits carrying cassiterite occur throughout the mountain. These lodes fill fractures which in places displace porphyry dykes and cut the pyrrhotite replacement body; hence they are considered a later phase of mineralization than the dolomite replacement. The veins have strike lengths up to 2,500 feet and dip length up to 1,000 feet and they pinch and swell along both dip and strike. The veins also commonly branch and converge forming a complex system of subsidiary ore bodies. The veins comprise cassiterite, in association with pyrite, pyrrhotite, arsenopyrite, sphalerite, chalcopyrite, galena, jamesonite, bis-muthinite, wolframite, stannite, quartz, siderite, tourmaline, fluorite and topaz, the relative proportions of these constituents varying considerably from one vein to another (see Stillwell, 1943). Apart from the major veins with an average width of 2 feet, there are subparallel sets of minor veins, some 1 to 2 inches in width, which commonly consist of quartz and cassiterite. These are generally too small to be economic.

The veins vary in strike from NNW to WNW and generally dip steeply to the west. This is also true of the minor vein deposits.

(c) The quartz porphyry intrusives have been mineralized throughout most of the mine area by topaz, tourmaline, fluorite, pyrite, pyrrhotite and cassiterite; these minerals generally pseudomorph feldspar. In places the proportion of this disseminated cassiterite ore was high enough for quarrying of the porphyry to prove economic.

(d) Well crystallized crystals of cassiterite, with a little quartz and tourmaline, also occur as in-crustations on joint faces within the porphyry and in the quartzites and shales, for up to 15 feet on either side of the intrusives. Where the joint frequency is considerable, the rocks have been mined by open cut methods.

Replacement of the dolomite and quartz porphyries probably involved a protracted period of alteration dependent to a large extent on temperature gradients and variations in gas and water pressure. However, the mineral assemblages formed at Mt. Bischoff can be satisfactorily divided into two temperature-dependent groups which are partially gradational. The higher temperature mineral assemblage consists of topaz, tourmaline, cassiterite, muscovite and some fluorite which largely replaces the quartz porphyries. This phase represents the introduction of fluorine and boron-rich vapours which were largely confined to high temperature zones delineated by the quartz porphyries.

Equivalent alteration of the dolomite horizon appears limited, although it is probably represented in the formation of chondrodite, minor topaz and tourmaline and possibly tremolite. Replacement of the dolomite horizon involves widespread magnesium metasomatism with alteration of virtually pure dolomite mineral (Table 2) to magnesium-rich carbonates and talc (Table 5). The dominant assemblage is sulphides, talc, carbonates, "serpentine", sericite, fluorite and cassiterite probably representing a slightly late more widespread, aqueous phase of mineralization under lower temperature conditions. Replacement of chondrodite by "serpentine" and possible replacement of tremolite by talc confirm this sequence.

Conformable pyrrhotite-pyrite-cassiterite bodies similar to that at Mt. Bischoff occur at Renison Bell and Mt. Cleveland (25 miles south, and 8 miles south-west, of Mt. Bischoff respectively).

The deposits at Renison Bell (Fisher, 1953; and Hall and Solomon, 1962) are gently dipping pyrrhotite sheets, up to 200 feet thick, or steeply dipping fissure lodes comprising pyrite and pyrrhotite. They occur in a Cambrian perhaps partly late Precambrian, succession of mudstones, greywackes, sandstones and dolomite, the pyrrhotite replacement bodies occurring mainly in a succession containing dolomites. The sediments have been folded into NW-trending anticline with normal faulting both parallel and oblique to the fold axis. Greisenized quartz porphyry dykes cut the sediments near the anticlinal hinge. The associations of greisenized quartz porphyry dykes with conformable pyrrhotite replacement bodies in Cambrian-Precambrian successions containing dolomite at both Mt. Bischoff and Renison Bell suggest a similar origin for the two deposits.

At the Mt. Cleveland Mine (Reid, 1923; Hall and Solomon 1962) pyrrhotite sheets replace certain beds in a sequence of Cambrian (?) sediments. As at Mt. Bischoff and Renison Bell, both replacements and fissure veins carrying cassiterite occur. The ore bodies comprise pyrrhotite, pyrite, chalcopyrite, arsenopyrite, quartz and cassiterite. Reid (1923) records that there are quartz porphyry dykes intruding the sediments in this area though later workers have failed to find the dykes. The Cleveland ores are basically similar to those at Mt. Bischoff but occur in a higher stratigraphic horizon and have a less obvious relationship to porphyry dykes.

Silver-Lead-Zinc Mineralization

Isolated, small silver-lead-zinc veins are common throughout the Waratah District, intersecting both the Bischoff quartzites and slates and the Cambrian (?) sediments. The largest deposit occurs at the Magnet Mine as a vein, some 10 to 15 feet in width, on the contact of albite porphyry and websterite porphyrite of the Magnet Dyke (Fig. 6). In depth this lode has a bifurcated structure with the main body of ore occurring at the junction of the two lodes, which were considered as intersecting steeply by Nye (1923), Cottle (1953) and Edwards (1960).

The lode is dominantly ankerite with bunches and veinlets of galena and sphalerite and lesser amounts of arsenopyrite, pyrite, boulangerite, pyrrargyrite, tetrahedrite and traces of chalcopyrite in a gangue of manganosiderite and ankerite (Edwards, 1960). Chalcopyrite occurs dominantly as segregation bodies in sphalerite, and pyrrargyrite and tetrahedrite as drop-like inclusions in galena. Crustification textures are extremely common,
with all sulphides interlayered with manganese-siderite. Cockade textures, developed about fragments of brecciated and serpentinized pyroxenite, are also common. Strong shearing and brecciation preceded the close of ore deposition with the production of elongate bent and brecciated fragments of galena and sphalerite which were later cemented by ankerite. This shearing and brecciation supports the theory of ore control by intersecting shears. Edwards (1960) also suggests that the mode of occurrence of the Magnet ore indicates an epigenetic origin and that the crustification and cockade textures indicate that the ore was, at least in part, deposited in open spaces.

The southern margin of the Magnet Dyke has also been a host for ore deposition, to the north-east of Magnet. A small lead deposit occurs at the Persic Section on the contact of Bischoff quartzites and slates with the Magnet Dyke. It may be significant that pyroxenite occurs on the southern margin of the dyke at this locality and also at Magnet. The Persic lodes are small, irregular and uneconomic, generally occurring as small splashes of galena and siderite within the country rock and not as a distinct vein as at the Magnet Mine. Small veins of quartz and carbonate which contain traces of copper and silver, also occur on the contact of Bischoff quartzites and slates and the Magnet Dyke at Fawkner's Show (North-east of the confluence of the Arthur River and Magnet Creek).

Small pockets of silver-lead-zinc mineralization are also common throughout the Waratah District away from the Magnet Dyke. The majority of these deposits are small, irregular and uneconomic veins, the most extensive deposit occurring at the Silver Cliffs Mine to the north-west of Mt. Bischoff. This well banded lode consists of galena, jamesonite, sphalerite, pyrite and minor boulangerite in a gangue of quartz and siderite. Similar small lodes occur on the northern slopes of the Arthur River (Fig. 1). Small lodes of sulphantimonides, jamesonite, stibnite, berthierite and boulangerite occur in Tinstone Creek to the south-west of Mt. Bischoff and are probably related to the silver-lead-zinc deposits, which commonly contain a high proportion of jamesonite. This group of prospects forms a "halo" of lead-zinc mineralization around the Mt. Bischoff tin deposit.

A genetic relationship between the silver-lead-zinc deposits and the Mt. Meredith granodiorite mass is suggested by the occurrence of the deposits in a belt on the north-west flank of the mass and their general lack of areal conformity with other igneous masses in the area. Silver-lead-zinc mineralization probably occurred during the Devonian metallogenic epoch along fractures in the roof sediments over a large granodiorite batholith which is now partially exposed. Tin mineralization was confined to aplite and pegmatitic bodies within the granodiorite and spread out by numerous porphyry dykes at Mt. Bischoff.

Determination of formation temperatures of the lodes in the Waratah District using the pyrrhotite and sphalerite geothermometer (Arnold and Reichen, 1962 and Kullerud, 1953) indicate a local temperature "high" at Mt. Bischoff with zonation outwards from the centre of intrusion. A lower, irregular temperature gradient existed over the remainder of the area, probably locally controlled by the granodiorite batholith.

The spatial association of sulphide-cassiterite and later stage silver-lead-zinc ores is a characteristic also of the Rimson Bell and Cleveland mines and is in fact, typical of many tin-rich provinces. Bilibin (1955) suggests, from a study of similar deposits in the U.S.S.R., that these sulphide-tin ores generally are related to the "late stages" of geosynclinal development, involving the intrusion of small granitic porphyries into fractured zones. He finds the majority of these deposits are Mesozoic or Tertiary.

Oxidation of the Orebodies

Gossans developed during the Tertiary over several outcropping sulphide orebodies. Indigenous gossan at the Magnet Mine is dominantly limonitic with bands of secondary minerals such as cerussite and pyromorphite. Gossans derived from pyrrhotite formed extensively over Mt. Bischoff. Though largely removed by mining, old reports indicate that they were a mixture of indigenous and exotic types. Pyrrhotite is very unstable in temperate conditions and breakdown resulted either in development of a friable limonitic crust with cassiterite concentrated at the base, or in complete removal of iron and sulphur, leaving a cassiterite or cassiterite-quartz sand on the surface. Where the orebodies cropped out on hillsides the cassiterite-quartz sand travelled down slope to form extensive eluvial and alluvial deposits. The spread of the exotic limonite was retarded to some extent by the presence of the dolomite bed but much of the dolomite was removed by the abundant sulphuric acid, resulting in further cassiterite concentration.

GEOLOGICAL HISTORY

Shallow water deposition of well sorted sands, silts and muds began during the Upper Precambrian. Chemical precipitation of dolomite with some contemporaneous deposition of muds and silts occurred towards the end of sedimentation with subsequent deposition of silts. Preconsolidation tectonics occurred in the sedimentary pile with the formation of pre-consolidation breccias and folds. A period of non-deposition, or deposition and erosion, ensued with minor tectonic activity.

Sedimentation again occurred during the Cambrian with the deposition of poorly sorted muds, silts and sands on a sinking sea floor. These sediments were in part derived from local uprisen areas of Precambrian rocks, as indicated by the common occurrence of detrital strained quartz grains, and from volcanic activity contemporaneous with deposition. Volcanic activity resulted in the formation of tuffs, volcanic breccias and spilitic rocks with associated near-surface basic intrusions. The origin of the albitic in the lavas is controversial but it is thought to be primary or deuteric. The final phase of the Cambrian igneous activity resulted in the intrusion of transgressive tabular sheets of ultrabasic and basic rocks.
The depositional and tectonic history of the Ordovician and Silurian periods cannot be determined in this area. It is probable from the occurrence of widespread marine sedimentary sequences of both ages in adjacent areas that they were also deposited near Waratah and have been subsequently eroded.

A major orogenic period, the Tabberabberan Orogeny, disrupted sedimentation in the Devonian. Evidence from Eugenana and Point Hibbs indicates that the tectonic movements associated with this period took place between the early Lower Devonian and the late Middle Devonian (Burns and Banks in Solomon, 1962). The first stage of tectonic activity was characterised by arcuate, long wavelength NNE-SSW-trending structures. The major activity occurred in the Meewee with the exception of small intrusions of dolerite during the Lower Tertiary, resulting in the formation of gossans and residual eluvial and alluvial deposits of cassiterite at Mt. Bischoff.

Recent sedimentation has occurred in the major rivers of the area, tin-bearing alluvium and gravels being deposited in the more mature sections of their course.

References

Acknowledgments

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———, 1951b.—A Note on the Occurrence of Intergrowth between Diopsidic Augite and Albite and of Hydrogrossular from King Island, Tasmania. Geol. Mag. 88, 429-431.


Fig. 1.—Slightly deformed flow casts in Precambrian quartzite, Waratah River.

Fig. 2.—Pre-consolidated breccia in Bischoff quartzites and slates, West Mt. Bischoff. 1/16 natural size.

Fig. 3.—Tertiary sediments showing differential compaction (?) against quartz porphyry bedrock, Don Hill, Mt. Bischoff. Plate by courtesy of J. Wilson.

Fig. 4.—Radiating topaz replacing quartz porphyry, White Face, Mt. Bischoff; x 35.
FIG. 1.—Amygdule containing quartz spherulites in Cambrian lava, Arthur River; polarised light, x 68.

FIG. 2.—Identical field of view, crossed nicols, x 68.

FIG. 3.—Radiating topaz pseudomorphing feldspar in quartz porphyry dyke, Mt. Bischoff; x 68.

FIG. 4.—Pyrite and topaz pseudomorphing feldspar in quartz porphyry dyke, Mt. Bischoff; x 55.
Publications from the Department of Geology, University of Tasmania


608. LOVELOCK, J. K.—Differentiation Problems in Basic Complexes in Relation to an Area in the Sierra Nevada, California. *Univ. of Tas. Dol. Sym.*, p. 61.


645. TANZANIA, University Geology Science Net., 1959.


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Publications from the Department of Geology, University of Tasmania


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RIB AND HACKLE MARKS ON JOINT FACES AT RENISON BELL, TASMANIA: A PRELIMINARY NOTE

MICHAEL SOLOMON and PATRICK ARTHUR HILL

ABSTRACT

Structures which resemble those previously described by glass technologists have recently been found on joint surfaces in Tasmania. The structures consist of concentric “ripples” (rib marks) crossed by arcuate, minor fractures (hackle marks); they are probably of tectonic origin but may have been formed by artificial impact.

INTRODUCTION

In recent years, attention has been directed to some spectacular structures thought to be due to impact (Deitz, 1959, 1960), and from time to time peculiar structures, obviously tectonic, have been described on joint faces (Woodworth, 1896; Parker, 1942; Raggatt, 1954; Hodgson, 1961a, 1961b; Roberts, 1961). Recently, structures either of tectonic or impact origin—the impact in this case man-made—have been discovered in Tasmania. They are on vertical joint faces in a road cutting 1 mile south of Renison Bell, in thick-bedded, fine-grained, homogeneous graywacke sandstone, dipping 10° northwest (pl. 1, A).

The structures, on normally planar joint surfaces, consist of concentric “ripples” which describe no more than 150° of arc and which are convex upward (pl. 1, B). The ripples are wavelike in profile and are slightly asymmetrical (fig. 1, a); their wavelengths, vary from a few millimeters to 3 cm., their amplitudes average 1–2 mm. and in a few cases exceed 2–3 mm. The structures are not quite concentric; wavelengths measured vertically are greater than those measured horizontally as if the ripples have “traveled” farther and faster upward than sideways. The ripples are crossed at right angles by approximately radial cracks, kinks, or steps.

The only inhomogeneities visible in the rock are planar fractures parallel or subparallel to the joint faces. Both the fractures and the joints are filled with thin films (0.5 mm. average) of finely striated quartz and chlorite. On the several joint faces of plate 1, A the striations have an approximately constant plunge of 50° and this plunge is maintained in the scaly remnants of chlorite that are preserved in the troughs of the ripples (pl. 1, B).

The relative ages of the ripples and the striations are unknown. On one structure the ripples appear to be earlier: their tops are planed off and they bear a fine grooving which is parallel to the striations. On other structures, however, the only indication that ripples are earlier is a possible smoothing of the ripple crests. If the striations formed before the ripples, then the striations should be deflected by the ripples. Such deflection was not observed but this may be because the amplitudes of most of the ripples are so small; unfortunately, where the amplitude is considerable and could be useful the striated quartz film does not extend over the crests (e.g., pl. 1, B).

ORIGIN OF THE STRUCTURES

Organic or sedimentary.—The Renison Bell structures are similar in appearance to cross sections through algal structures depicted by Bradley (1929) and Shrock (1948, p. 287). However, they are confined to the joint plane, whereas algal structures almost invariably extend laterally along the bedding.

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The Renison Bell structures are similar in part to those produced by the swishing to and fro of plants or seaweed by changing currents (e.g., tidal) and, in part, to the concentric rings formed by gas blisters in mud (Cloud, 1960). Again, however, such features generally have their greatest dimension along bedding surfaces.

**Tectonic.**—The probability that the ripples are tectonic depends upon their age relationship to the strated joint filling. If the ripples are pre-filling then they are probably tectonic, if they are post-filling then they may be tectonic but they could also have some other origin (see section, "Artificial or Man-made").

Approximately radial fractures or joints—like the percussion rays of Woodworth (1896)—are illustrated by Raggatt (1954), Parker (1942), Hodgson (1961a) and Roberts (1961) show radial features that form part of plumose “feather fracture” systems, of tectonic origin. Hodgson shows “plumose” cracks crossed by crudely concentric “curvilinear ridges” and suggests that these structures develop synchronously with jointing. Perfectly concentric structures, believed to be tectonic, occur on cleats in coal (Hofmann, 1910, Stutzer and Noé, 1940) and the coal containing these features is described as “augen kohl.” Stutzer and Noé (p. 253) agree with Hofmann that the “eyes” formed simultaneously with the cleats under the same stress system and that they developed because of minor inhomogeneities in the coal. Stutzer and Noé note that similar structures occur in the Zechstein shales.

Concentric and radial markings have been discussed in several analyses of fractures in glass. Preston (1926), Murgatroyd (1942), Oughton (1945), and McKenna

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**Fig. 1**—a, Generalized section at right angles to “ripples” (rib marks) on joint face at Renison Bell (Note the positions of the cracks are apparently unrelated to those of the crests and troughs. The top of the joint face in to the left.) b, Section through rib marks formed in glass by the repeated change of direction of a fracture during propagation. Based on Murgatroyd (1942) Scale immaterial, wavelength dependent on fracture velocity and rock type.

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**PLATE 1**

A. Concentric “ripples.” Area outlined in black is shown in pl. 1, B

B. Area outlined in fig. 2, showing concentric “ripples” (rib marks) and remnants of quartz-chlorite film preserved in troughs of ripples. Arcuate cross fractures (hackle marks) shown by arrows. Scale Length of matchbox is 2 inches.
Concentric "ripples" and Huckle Marks
Hackle marks and rib marks
describe the arcuate ripples as “rib marks” and the radial features as “hackle marks” (pl. 2, A). Rib marks represent pauses in the advance of the main fracture; as the advance slows, the fracture deflects through 90° until it is at right angles to the main fracture surface; on re-starting it curves back to its original direction (fig. 1, b). The ribs are normally sharp—unlike the rounded crests shown in figure 1, a—and the rib spacing is inversely proportional to the fracture velocity. Rib marks form in both slowly and rapidly propagating cracks from a variety of causes whereas hackle marks are high-velocity cracks formed by impact or explosion. Both rib and hackle marks develop under tensional stress.

The similarity between the structures in glass and those on the Renison Bell joints indicates a similar origin. The glass structures probably developed under local tensional stress following point impact; if the Renison Bell structures are tectonic they probably developed during joint growth at discrete points of tensional stress build-up caused by local inhomogeneities in the sediments. These irregularities may have caused momentary delays in joint propagation and on recommencement the joint fractures apparently extended more rapidly upward (toward the surface?) than sideways (pl. 1, B). The rounded ribs may be the result of a later (?) “planing-off” by movement along the joints at the time the striated quartz-chlorite film developed.

Most of these conclusions, developed by comparison with fractures in glass, are supported by Roberts (1961) who concludes that “feather fractures” in joints are similar to the cleavage fractures in steel, which develop suddenly at certain stress levels and propagate rapidly. Roberts cites Nadai (1950, p. 182) as stating that conchoidal fractures with concentric ripples develop under tensile stresses caused by concentrated impact.

Roberts (1961) agrees with Parker (1942) that feather markings only appear on shear joints, but his criteria for recognizing shear joints are not necessarily valid; the symmetrical disposition of joints about a fold axis (p. 486) does not prove that the joints are non-tensional and Muehlberger (1961) had demonstrated that the shear joints with plumose markings described by Parker are essentially tension fractures.

Artificial or man-made.—The Renison Bell structures may, however, be directly related to man-made impacts, particularly as the structures occur in a road-cutting excavated by drill-hole blasting. Similar structures (though with subdued hackle marking) have been observed in Silurian-Devonian siltstones in road cuttings in eastern Tasmania; these structures overlap one another on irregular fracture faces, are an inch or two in diameter and, like the Renison Bell examples, are confined to relatively homogeneous beds. They almost certainly formed during rock blasting. Coarse rib marks (pl. 2, C) were frequently developed in tough siliceous Permian mudstones during blasting for the Poatina Power Station, Tasmania. Somewhat smaller structures, with both radial and concentric elements, have also been found in Permian mudstones at Hobart (pl. 2, B), but these were formed as the rocks broke under jack-hammer impact. Crude percussion rays may also develop when “dimension stone” is artificially split or trimmed (Hill, 1961) and conchoidal fractures—comparable to rib marks—develop by knapping flints and cherts.

PLATE 2

A. Hackle marks (H) and rib marks (unmarked) produced in glass by impact of hammer. After Oughton (1945); scale not stated.

B. Hackle marks (H) and rib marks (arrows) in Permian mudstone from Hobart, Tasmania.

C. Symmetrical rib marks produced in Permian mudstone by explosion-impact. Compare with fig. 1, B. Specimen collected by H. L. Paxton.
Thus the Renison Bell structures could possibly have resulted from detonation during road construction. Pre-existing joints and inhomogeneities in the sediments may have distorted the stress field and acted as discrete foci for stress concentration. If due to blasting, the rib marks must be younger than the striated film; this, however, is difficult to reconcile with the planing-off of the rib marks mentioned above, unless there was some movement between joint blocks during or just after the explosion.

CONCLUSIONS

The Renison Bell structures have several features in common with joint markings described elsewhere and to fractures observed in glass. They are probably tensitional in origin and formed during the development of jointing, but may be a result of explosion impact.

REFERENCES CITED

STUTZER, O., and NOé, A. C., 1940, Geology of coal: Chicago, Univ. of Chicago Press.
Cambrian Succession in West Tasmania

Campana, King and McKenna (1960) have suggested in This Journal that the Mt. Read Volcanics (or porphyroids of Campana et alii and many earlier writers) are older than the Dundas Group and separated from it by an angular unconformity. They support their view by contrasting the deformation and metamorphism of the Volcanics and the overlying rocks.

The critical exposure (in a railway cutting north of Bulgo b a c) shows that there was erosion of the Mt. Read Volcanics during deposition of the overlying sediments but there is no evidence of angular discordance; while admitting local disconformity we question the "angular unconformity". Moreover, although there may be a difference in tectonic style we suggest that there is no marked difference in degree of deformation or metamorphism.

The correlation of the overlying sediments with the Dundas Group is apparently made on lithological similarity and the presence of (?) Hurdia davidi, as was done by Banks (1956); however, recent work on the Cambrian rocks of Tasmania indicates that the only valid basis for stratigraphical correlation is palaeontological and that the specimen named (?) Hurdia davidi by Chapman (1926) is too poor to be diagnostic. The age and stratigraphical position of the sediments above the Volcanics at Bulgobac Siding are at present unknown.

However, even if the correlation is correct, the disconformity (?) and the felspar grains in the Dundas rocks may be satisfactorily explained by assuming a volcanic accumulation near the margins of the sedimentary basin during part or all of Dundas deposition and erosion of volcanic material penecontemporaneously with volcanism and sedimentation (see Carey, 1953; Campana et alii, 1958). Crystal tuffs and lavas, both in the Dundas Group at Dundas and interbedded with fossiliferous Middle and Upper Cambrian rocks near Beaconsfield and Ulverstone demonstrate volcanic activity during Dundas deposition.

Although Campana et alii have put forward an interesting thesis, in our opinion the evidence is insufficient to show an orogeny between the formation of the Mt. Read Volcanics and the deposition of the Dundas Group.

Campana et alii explain the different structural relations at the base of the Owen Conglomerate at Red Hills and at Zeehan (Mt. Misery?) in terms of an orogeny between the Volcanics and the Dundas Group, but we suggest that the contrasting degrees of discordance at these localities can best be accounted for by assuming lateral differences in intensity of deformation of the pre-Owen beds during Jukesian movement (see Banks 1956, Solomon 1960).

References


The Proterozoic-Upper Cambrian Succession in West Tasmania

Having mapped for two years the regional geology of the Zeehan Quadrangle, we welcome the recent contribution by Campana, King and McKenna (1960) as an attempt to unravel the complicated Cambrian succession of the West Coast. Although our work is incomplete, we must question one of their major conclusions.

We agree that the name 'Dundas Group' should be restricted to the Middle-Upper Cambrian sequence (Psychagnostus gibbus-Glyptagnostus reticulatus Zones) and to rocks which can be correlated, for example the 'Huskisson Group' of Taylor (1954). However, we consider that in the Pieman-Zeehan-Dundas area the Group follows conformably on a sequence which ranges from Lower to Middle Cambrian, and not unconformably on Lower Cambrian as postulated by Campana et alii.

(1) Stratigraphical Sequence

Elliston (1954) defined the Carbine Group and the Dundas Group at Dundas, while between 1951 and 1954 B. L. Taylor and D. Burger mapped a considerable area round the Pieman River and Renison Bell. We
have continued this work, studying the Proterozoic-Cambrian succession over 350 square miles, and our mapping shows that overlying the Older Proterozoic there are two different series. The older sequence, probably ranging from Upper Proterozoic to Lower Cambrian, comprises pale grey saccharoidal quartzites, muscovite-bearing grey siltstones and fine quartzites, with grey or greenish-grey Cambrian, comprises pale grey saccharoidal quartzites, with quartz veins, and schistosity or cleavage is locally well developed. The sequence includes the Oonah Quartzite and Slate (Spry, 1958) which apparently passes up into the finer Carbine Group, though the latter may be only a facies variant. We suggest that this sequence is at least 7,000 feet thick.

The Oonah or Carbine quartzites and slates are overlain by a thick series of argillites, greywacke and conglomerates, conspicuous by their purple, red, green and grey colours, which have long been referred to the Dundas Group. However, Taylor (1954) showed that below the fossiliferous Dundas Group in the Pieman River-Huskisson River region there is an important thickness of argillites, slates and greywacke included by Ward (1909), Waterhouse (1914) and Conder (1918) in the Dundas Group, which he named the Crimson Creek Argillite. This sequence, which may be up to 10,000 feet thick, and which we propose to term a Formation because subdivision may be possible later, is composed of generally finer sediments than the overlying Dundas Group.

We agree with Campana et alii that the Success Creek Group is equivalent to the Carbine Group, and that it is of Lower Cambrian or possibly Upper Proterozoic age. However, we support Taylor's conclusion that the Crimson Creek Formation is Lower to Middle Cambrian, and that the Dundas Group, as defined, succeeds it conformably.

The Dundas Group has been described in detail by Elliston (1954). The beds closely resemble those in the Crimson Creek Formation, but also include a number of massive greywacke-conglomerates separated by fossiliferous shales or siltstones. The Glyptagnostus reticulatus Zone has been found only on the Huskisson River.

(2) Localities

The succession in different districts is briefly described below:

(i) Pieman River-Huskisson River

The Oonah Quartzite and Slate is probably unconformable on Older Proterozoic and passes up into the finer Carbine Group. Taylor commented on the incoming of 'pyroclastic' material (greywacke) in the upper part of the Carbine Group. Such bands are common in the Crimson Creek Formation and the Dundas Group. On structural and lithological evidence we support Taylor's opinion that the Carbine Group passes up into the Crimson Creek Formation near the mouth of the Wilson River, and that on the Huskisson River the fossiliferous Dundas Group follows conformably.

(ii) Remison Bell

Micaceous saccharoidal quartzites, siltstones and shales resembling those in the Carbine Group form a structural high and pass up into greywacke, conglomerate and argillite typical of the Crimson Creek Formation with no evidence of unconformity. In this area there are many sills of gabbro and serpentine. The Dundas Group as defined has not yet been identified and may be absent.

(iii) Dundas

The Carbine Group probably rests unconformably on Older Proterozoic schists. Elliston's conclusion that the Dundas Group is unconformable on the Carbine Group was based on doubtful evidence and our work so far indicates the presence of the Crimson Creek Formation beneath the Dundas Group. For example, north of Mt. Dundas the Carbine Group passes eastwards up into purple slates and greywacke resembling those in the Crimson Creek Formation elsewhere. Again, in Mariposa Creek on the west flank of Mt. Dundas west-dipping Carbine Group quartzites and slates are overlain by greywacke and conglomerate which we assign to the Crimson Creek Formation. In each locality strikes and dips are compatible with a passage up from the Carbine Group.

(iv) Zeehan District

The Oonah Quartzite and Slate can be traced for many miles north and north-west of Mt. Zeehan. The Carbine facies occurs near Queen Hill, Zeehan and on the Trial Harbour road. We consider that the Crimson Creek Formation is represented by part of the 'Keratophyric Tuff' of Twelvetrees and Ward (1910) which appears to follow the Carbine Group conformably east of Queen Hill and in the Austral Valley. Although the contact is frequently obscured by complex faulting and deep weathering, structural trends do not indicate a major unconformity. Part of the 'Keratophyric Tuff' which yielded Diplagnostus sp. was correlated by Opik (1951) with the Hodge Slate (Dundas Group).

(v) East of Trial Harbour

The quartzites and siltstones resemble those of the Carbine Group. Southwards across the Little Henty River they are succeeded by dark purplish cherts, argillites and greywacke which, on lithological grounds, we correlate with the Crimson Creek Formation. Once again comparable structures
indicate that there is no unconformity. About two miles south-east of Trial Harbour part of a trilobite was found by D. Groves; it has not yet been identified but is probably a species from the Dundas Group.

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16 December 1960.

References

Comment on the Note of Banks and Solomon

The statement of Banks and Solomon that ‘the only valid basis of stratigraphic correlations is palaeontological’, appears a little dogmatic. Significantly it has been disregarded by them in treating the Tasmanian Cambrian stratigraphy. Indeed, nine out of 13 sections have been assigned by Banks (1957, pp. 178-187) to the Dundas Group ‘on lithological and structural grounds’. Should these be regarded as of unknown age and stratigraphic position?

The sediments of the Queenstown area that have been correlated with the Dundas Group by Wade and Solomon (1958, p. 374) are, in the author’s words ‘unfossiliferous, but similar lithologies’ in the type area… contain trilobites and dendroids of Middle and Upper Cambrian age’. Is this correlation invalid? I do not know whether Banks and Solomon would now apply such a rigorous principle to their own stratigraphic conclusions, but surely all over the world stratigraphers are legitimately using other criteria in correlating unfossiliferous formations.

Questions of method apart, the correlation of the Dundas Group of the sequence overlying the Mt. Read Volcanics in the Bulgobac-Que River area has not been based merely on lithological grounds. Besides the presence of Hardia dawidi, whose specific determination was not questioned by Banks in his previous publications, six distinctive members of the Bulgobac-Que River sequence are also found, in identical order of succession, in the fossiliferous Dundas beds of the Huskisson River, 12 miles south. Among these members there is a horizon of light-coloured siliceous conglomerates, very similar in facies to the Owen Conglomerate, which appears to form an unmistakable marker bed.

That the general presence of felspathic fragments and lava pebbles in the Dundas Group may be satisfactorily explained by assuming erosion of a volcanic accumulation near the margin of the basin, is precisely our conclusion. It follows that the Mt. Read Volcanics are older than the Dundas sediments. We have postulated an angular unconformity between these two groups of rocks not only ‘by contrasting the deformation and metamorphism’, but also on specific stratigraphic evidence.

Thus, in the Mt. Murchison-Red Hill area the Owen Conglomerate-Jukes Breccia formations rest on the Mt. Read Volcanics with a right-angle unconformity, and contain abundant pebbles and boulders of metamorphosed Volcanics. In the Mt. Misery area this conglomerate shows not only conformable relations but also lithological gradations with the Mt. Misery Conglomerate forming the top of the Dundas Group. This would prove that the Mt. Read Volcanics were already metamorphosed and steeply folded at the time of deposition of the Jukes Breccia-Lower Owen Conglomerate, while undisturbed sedimentary conditions prevailed in the adjoining Dundas Trough (Mt. Misery area, Huskisson Syncline). The inference that the orogenic movements and related metamorphism that affected the Mt. Read Volcanics pre-date the Dundas sedimentation appears, therefore, stratigraphically justified.

Banks and Solomon suggest (but do not prove) ‘that there is no marked difference in degree of deformation or metamorphism between the Mt. Read Volcanics and the overlying sediments of the Bulgobac-Que River area’. But McKenna, who mapped this zone for the first time, has shown that the sedimentary succession is folded in a regular syncline, the limbs of which have a general dip of 47°. Shearing, schistosity or other metamorphic effects in the sediments are entirely absent, the succession being, in fact, largely formed of shales, greywacke and felspathic sandstones. By contrast, the position of the underlying Volcanics is subvertical throughout in spite of their more competent nature, and their metamorphism is so widespread that one can not but agree with the view of Carey that ‘the abundant igneous rocks of the district… all fall into that varied group of sheared acid and subacid porphyries which have gone under the name of porphyroids… or schistose quartz-felspar porphyries, and where the alteration was most extreme, sericite schists’ (Carey, 1945, p. 22; 1953, p. 1118). Similar alterations have not been described so far in the sediments of the Dundas Group of the area, nor have any been observed by us.
Comment on the Note of Bisset and Gulline

The unconformity between the Dundas Group and the Carbine-Success Creek Group questioned by Bisset and Gulline, was accepted by us on the basis of regional mapping in the Dundas-Renison Bell area, in agreement with previous authors. But our work was not then completed, and the problem can hardly be fully discussed before the various new maps and reports are published.

As for the stratigraphic position of the Crimson Creek argillites, there is indeed evidence that they conformably underlie the Ptychagnostus gibbus-Glyptagnostus reticulatus zones, as suggested by Blissett and Gulline. But our observations were not quite conclusive in this respect, so we preferred to consider the argillites as part of the Dundas Group, following a personal communication by Öpik who is inclined to regard them as a lateral facies of the fossiliferous Dundas beds. In addition, their local developments, paucity of outcrops and vertical gradation to the Dundas beds make them an ill-defined mapping unit. However, in sections and palaeoprofiles illustrating a work now in press we represent the argillites (and interbedded black shales and cherts) as conformably underlying the Ptychagnostus gibbus-Glyptagnostus reticulatus zones. They would represent the initial euxinic facies of the geosynclinal cycle, followed by the greywacke-conglomeratic sequence of the Dundas Group (Flysch facies).

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References
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THE DUNDA S GROUP IN THE QUEENSTOWN AREA

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(With 1 Plate and 4 Text Figures)

ABSTRACT

Reconnaissance geological surveys of the Dundas Group in the Queenstown area reveal that the sequence contains a suite of volcanic rocks including potassic rhyolites, quartz keratophyres, andesites, basalts and associated pyroclastics in addition to the sedimentary rocks, which are mainly paraconglomerates, greywacke sandstones and slates. The potassic rhyolites are confined to a narrow zone between Mt. Sedgwick and South Darwin and are flanked by lavas which vary from rhyolites to basalts but which generally contain a relatively high soda content. Viewing the West Coast as a whole this variable zone has a western limit beyond which the lavas are mainly of basic type and commonly spilitic.

The rock types which make up the Darwin Granite body are briefly described and it is noted that chemically they resemble members of the volcanic suite. The age of the granite is verified as pre-Junee Group.

It is suggested that the formation names previously proposed for the Dundas Group in this area are invalid.

INTRODUCTION

Geologists have been attracted to the Queenstown area ever since gold was first discovered there in 1883 by a prospecting syndicate. The clearing of the thick rain forest and the development of the associated copper deposits that followed the initial discovery soon revealed the complexity of the local geology, which became the subject of numerous technical reports. Of the scientists who visited the field in its early years, Professor J. W. Gregory was the first to extend his survey beyond the mine leases and his geological map published in 1905 remained the principal reference for many years. The Tasmanian Mines Department gradually developed its activities on the Lyell and the nearby Jukes and Darwin fields and by the middle 1930's an approximate picture of the geology had been obtained. However, not until Bradley began his work in 1950-51 had any attempt been made to produce a general geological map of this part of the West Coast, despite its obvious economic importance.

This pioneer work was followed in 1954-56 by a more detailed survey of the area by the writer on behalf of the Mt. Lyell Mining and Railway Co., as part of a new approach to the solution of the problems of ore deposition at Mt Lyell. The area then investigated included that part of the West Coast Range between Mt. Sedgwick and South Darwin, involving about 200 square miles of rugged terrain (Fig. 1). Aerial photographs of scale 1 inch = 1320 feet were used for mapping so that the survey was of a reconnaissance nature and hence by no means a complete study; similarly the petrographic work must be regarded as essentially preliminary. The specimen numbers quoted refer to the rock and slide catalogue of the Mt. Lyell Mining and Railway Co., Queenstown.

Discussion in this paper is confined to the Dundas Group which is of particular interest in that it is host to the Mt. Lyell copper deposits and contains a variety of rock-types which have puzzled geologists for many years. The Group was defined by Elliston (1954) as the sequence of sediments and volcanics exposed near Dundas township, underlying the Junee Group and overlying the Carbine Group (Precambrian). As a result of regional mapping programmes by private enterprise, Government departments, and the University of Tasmania, the extent of the Dundas Group outcrop on the West Coast has been determined approximately, correlations being made largely by lithology and continuity of outcrop but occasionally on palaeontological evidence. Those fossils collected to date show a range from the Ptychagnostus gibbus Zone to the Glyptagnostus reticulatus Zone, or approximately Upper Middle Cambrian to basal Upper Cambrian (Banks, 1956). No fossils have been found near Queenstown and correlation is made by continuity of outcrop from areas to the north, by lithology, and by relationships to overlying beds. The base of the Group is not exposed in this area.

The sediments of the Dundas Group are thought to have been deposited in a geosynclinal basin which extended over at least western Tasmania and part of Victoria (represented there by the Heathcotean Series) and the abundance of volcanics has led to the use of the term eugeosyncline (Kay, 1947) for the basin of deposition.

A feature of the Dundas Group on the West Coast is the development of a considerable thickness of dominantly volcanic material along the West Coast Range. This was recognised by Carey (1950, 1953) and further emphasised by Campana et al. (1958) in their subdivision of the Group into a "bedded series" and a "volcanic assemblage". The latter is particularly well exposed in the Bulugbac-Pinnacles-Rosebery-Mt. Tyndall area and forms a sharply defined N-S zone of dominantly volcanic rock, while the Dundas Group to the west is essentially a sedimentary sequence.
Elliston described considerable thicknesses of agglomerate and tuff in the succession at Dundas but these beds are now recognised as non-pyroclastic and it is clear that the type section belongs to the "bedded series".

South of Mt. Tyndall the volcanic assemblage of Campana et al. is difficult to distinguish as its percentage of sedimentary rocks increases and lavas become ubiquitous. As a general rule all Dundas Group sections south of the latitude of Mt. Sedgwick contain roughly equal proportions of volcanics and sediments.

The volcanic zone between Bullogbac and Mt. Tyndall is interpreted (Carey, 1950) as marking a line of volcanoes in the Cambrian eugeosyncline and Campana et al. have related this zone of volcanic activity to the early phases of rift valley formation.

No fossils have been reported in the volcanic zone and its stratigraphic relationship to the bedded series is unknown.

**PREVIOUS STUDIES AND PRESENT VIEWS**

Gregory's (1905) descriptions of the Dundas Group rocks in the Queenstown area, including chemical analyses and petrographic data, are still referred to. Of the earlier contributions from the Tasmanian Mines Department the most useful accounts of the Cambrian rocks are those by Twelvetrees (1900) and Hills (1914) on the Jukes-Darwin mining district and Hills (1927) on the Lyell field. Nye, Blake and Henderson (1934) completed a generalised survey of the area north of Mt. Jukes and their views on the origin of the Cambrian rocks were opposed to those expressed by Hills. Although later workers on the Mt. Lyell leases (e.g. Edwards, 1939; Conolly, 1947; Alexander, 1953) followed Nye et al., recent work confirms Hills' conclusions that the Cambrian rocks are largely of volcanic origin.

Hills and Carey (1949) and Carey (1953) discussed problems of West Coast geology and Banks (1956) included local information in his summary of the Dundas Group in Tasmania. The regional geology has been dealt with at length by Bradley (1954, 56, 57) and summarized by Wade and Solomon (1958).

The earlier geological opinions on the nature of the Dundas Group in this area varied from intrusive complexes with sedimentary roof-pendants (e.g. Nye, Blake and Henderson; Edwards; Conolly), to volcanics with minor sediments (e.g. Hills 1914, 1927) while others have suggested combinations of these extremes (e.g. Gregory, 1905). In 1953, Carey obviously considered that many of the so-called pyroclastics were greywackes and he stressed the sedimentary nature of the Group as a whole: he also heralded and supported the work of Bradley, who introduced a new approach when he postulated that many of the igneous rocks were a result of metasomatism of greywacke sediments and basic lavas, the metamorphic processes being a Tabberabberan age and related to sulphide mineralization. He envisaged the extensively developed quartz and feldspar porphyries (the term porphry is used in this text to describe a rock showing porphyritic texture) as originating by "porphyriticization", involving growth of feldspar and quartz in situ, whereas all earlier workers had considered these rocks to be igneous. He was supported by Scott (1954) though she suggested the metasomatism only affected basic lavas and was of late Cambrian age. These views have been criticised by Banks (1956) and Wade and Solomon (1956) and it is now suggested that the Dundas Group in the Queenstown area comprises a suite of only slightly metamorphosed lava flows (varying from rhyolites to basalts), agglomerates, tuffs, conglomerates, sandstones and slates. This is very similar to the description given by Hills 45 years ago.

Bradley's views were accepted by other workers (e.g. Carey 1953, p. 1109) probably largely because severe alteration of the Dundas rocks was obvious near the Lyell deposits and it seemed reasonable to assume a more "regional" metasomatism involving porphyrritization and granitization. However, Cambrian sequences in areas further afield, where there is no obvious metamorphism, contain identical porphyries and there is no reason to believe these are metasomatic. Examples include the Cambrian rocks in the D'Aguilar Range (south of Macquarie Harbour) on the High Rocky Point-Point Hibbs coastline, and in the Tullah-Mt. Farrell area. Those on the Southern Ocean coast are somewhat lenticular bodies interbedded with sediments (and pyroclastics?) and varying from a few inches to several hundred feet thick. They are fine grained, porphyritic, locally vesicular and a few exhibit scoriaceous tops. The evidence is overwhelming in favour of these rocks being lava flows. The Tullah and D'Aguilar Range bodies are less well exposed but they appear to be concordant igneous porphyries.

Further and more direct evidence that the Queenstown porphyries are igneous is provided by the discovery in them of high-temperature quartz crystals (see p. 4). Several of these porphyry occurrences (e.g. the Lynch Creek area) are lenticular, concordant, vesicular and amygdaloidal, fine grained, associated with pyroclastics and similar to lavas of other areas, so that they are almost certainly volcanic. However, some of them (e.g. in the West Queen River and near Little Owen) have elliptical, almost circular outcrops and show discordant contacts; these are either intrusions, or volcanic necks, or steep-sided Felican-type domes of viscous lava, burial of which by ash and sediment has produced pseudo-discordant relationships at their margins.

The majority of the porphyries are very probably lavas, judging by field relationships, similarities to other flows, and both macro- and microscopic textures, but there may be instances of intrusive bodies.

**VOLCANIC ROCKS**

*Lavas*

The principal lava-types observed in the Queenstown-Darwin area are as follows:

**Spherulitic potash rhyolites**

On Mt. Darwin, Intercolonial Spur, Whip Spur and Mt. Sedgwick (Fig. 1) are found distinctive massive outcrops of closely jointed, pink haematitic
FIGURE 1.—The Queenstown Area.
feldspar porphyry (D 32, 33). The outcrops of these porphyries, in plan, vary from those which are almost circular (e.g. Whip Spur and Mt. Sedgwick) to those with one dimension much greater than the other (e.g. Intercolonial Spur). Possibly the former represent volcanic necks of Peléan cores while the latter are flows. The groundmass of these porphyries is aphanitic and the pinkish feldspar laths are up to 3 mm. long and very sparsely distributed. In places, and particularly south-east of Mt. Sedgwick, the rock is laced with haematite magnetite veins which are up to two feet wide, generally lenticular and of random orientation.

Microscopic study of typical porphyries on Mt. Darwin and Intercolonial Spur show the phenocrysts to be subhedral laths of plagioclase feldspar (albite ?) many of which show slightly corroded margins. Alteration of some crystals is slight, but in others it is so intense that the crystal is merely an outline in a microcrystalline “felt” of sericite, calcite, &c. The groundmass, which may form more than 90% of the rock, is composed mainly of rather crudely developed spherulites generally less than 0.5 mm. diameter. These usually possess a core of clear quartz and often a rim of quartz or dark brown haematitic (?) "dust" (Fig. 2). Notable features of the chemical composition of a typical sample of this rock are the high silica and potash percentages (73.4% SiO₂, 8.0% K₂O—see Table 1, No. 1). Sodium-cobalt-nitrite stain tests (Chayes, 1952) gave negative reactions on the phenocrysts and strong indications of positive reactions throughout the groundmass. Though the fine grain of the base reduces the certainty of the latter observation, the stain results and the chemical analysis together suggest that much of the groundmass material is potash-feldspar.

![Figure 2.—Spherulitic forms in potash rhyolite, Mt. Darwin. Clear material in the spherulites is quartz.](image-url)

Similar rock (Q 75) forms a bold outcrop at the head of Whip Spur (see Figs. 1 and 3). Albite (?) phenocrysts up to 2 x 1 mm. in size occur in a microcrystalline feldspathic ground mass in which spherulites abound; these are less than 3 mm. diameter, consist of radiating fibres and have a thin dark circumference like many of those described from Mt. Darwin. Once again the base reacts positively to potash stain tests. A fine banding which shows on some weathered faces is interpreted as flow banding.

An interesting variant (D25) acts as host to the Lake Jukes copper orebody. It is a dense, mottled, pink and grey “felsite” similar to the porphyries just described but differing in locally showing micrographic texture between quartz and potash (?)-feldspar. Hills (1914) describes the rock as a granophyre. The texture may be primary but could also result from reconstitution of a spherulitic base during mineralization. Brecias and banded rocks associated with this and other potash rhyolites are described under the heading “Pyroclastics” on p.

The outcrops of the pink rhyolites are confined to a narrow zone roughly coincident with the axis of the West Coast Range between South Darwin and Mt. Sedgwick, and the only other occurrences known to the writer are at Red Hills, South Mt. Farrell and near Lake Rolleston, all within the West Coast Range. This localization suggests the existence of a narrow zone of distinctive volcanic centres in Cambrian times from which acid, potash-rich lavas were erupted. It is significant that a major part of the only granite of proved Cambrian age in the Queenstown area is of similar composition to these rhyolites and is intruded into them.

**Quartz keratophyres**

Quartz-feldspar porphyries outcrop in the West Queen River, at Harris’ Reward pack bridge (over the King River), in the Garfield River, west of Mt. Sedgwick, north of Darwin and at many other places within the area under discussion. Outside the area, similar rocks occur near South Mt. Farrell, Lake Dora and over a wide area north of Boko Siding on the Emu Bay Railway.

The boundaries and precise shape of these quartz porphyry outcrops are always difficult to define though generally the bodies conform to the local structural trend. The porphyry exposed in the West Queen River has an elliptical outcrop roughly concordant with the regional strike but its southern margin appears to cut across steeply dipping sandstone beds. This porphyry (Q31) is pink or grey in colour with phenocrysts of both clear quartz and feldspar set in an aphanitic groundmass. Thin section study reveals that the quartz phenocrysts show some well-defined crystal faces but most are corroded and embayed and frequently contain inclusions of the ground mass. A few of the crystals show fuzzy, poorly defined boundaries.

Bradley (1957) has discussed similar observations on this and other quartz porphyries and he considers they indicate that the phenocrysts have grown in situ by a system of processes involving solid diffusion. Actually all the features described are typical of present-day and ancient acid eruptives, the corrosion, embayments, &c., being attributed to reaction between developing quartz crystals and the residual liquids. Proof of their igneous origin has been provided by Dr. E. Williams who has shown me etched basal sections of quartz crystals from West Coast porphyries that
clearly display the cracking and heterogeneous twinning typical of the high temperature form. The feldspar of the typical West Queen River porphyry occur as hypidiomorphic crystals in approximately equal quantity to the quartz, the phenocrysts of the two minerals forming 60% of the typical rock. Few feldspar crystals show embayments or inclusions but many are markedly zoned, a feature which is often highlighted by alteration of the core to chlorite. In the rim to sericite (?) the crystals are clear, unaltered albite but the majority are clouded by alteration products and appear to have a composition of about Ab9. Ferromagnesian minerals are very scarce though chlorite and haematite confined in idiomorphic lath outlines suggest that some ferromagnesian mineral was once present in the rock.

The groundmass is a microcrystalline aggregate of feldspar, chlorite, and haematite of which the texture is obscured by alteration products.

A feature of these quartz porphyries and also the feldspar porphyries (keratophyres) is the presence in the microcrystalline chlorite-chlorite-albite (Ab9) aggregates some of which occur isolated in the groundmass but many of which are clearly replacing altered feldspar laths. Many are only visible under the microscope but others are seen in outcrop as vughs partially filled with albite, chlorite and epidote. The albite is characteristically confined to the rim while the ferromagnesian minerals fill, or partially fill, the core. A specimen in the Mt. Lyell Mining and Railway Company’s collection, presented by Professor J. W. Gregory in 1903, is a feldspar porphyry from the East Queen River with a vugh several centimetres across lined with albite, epidote and quartz. There is every gradation between typical zoned amygdales and irregular “clots” derived from feldspar alteration and they are all regarded as expressions of late-stage gas action (i.e. deuteric phenomena). The albite of this phase is usually pink or brown in colour and appears fresh and unaltered under the microscope.

A chemical analysis of a large specimen of typical West Queen River porphyry is given in Table 1, No. 3; in conjunction with the petrographic data it suggests the rock should be identified as a quartz keratophyre. The latter term is used to denote a rock with the chemical composition of a sodic rhyolite but with abundant albite or oligoclase and a very small percentage of primary ferromagnesian mineral, which is altered to chlorite. It is distinguished from keratophyre by abundant free quartz and consequent higher silica percentage. Gregory (1905, p. 57) described a quartz-feldspar porphyry from the Lyell Comstock tram line as a diabase porphyrite. However, his petrographic description and the chemical analysis he provided (see Table 1, No. 4) suggest he had examined a rhyolitic flow.

A striking quartz keratophyre (Q22) outcrops in a small quarry west of the Zeehan road three miles from Queenstown. It is a pink-brown porphyry containing prominent amygdales lined with albite and filled with chlorite and occasional specks of galena. The majority of the phenocrysts are of altered plagioclase feldspar in laths up to 4 mm. long, the remainder being of corroded quartz crystals and irregular, rounded masses of chlorite with dark rims which may represent altered feldspar crystals. The chlorite shows the dark blue or rich brown interference colours typical of pennine. The groundmass is a microcrystalline aggregate of quartz and feldspar, speckled with pyrite grains and containing rareapatite crystals.

A slightly different variety of quartz porphyry outcrops on the tram line between the upper and lower power houses at Lake Margaret. It varies in colour from grey-blue to yellowish grey and contains phenocrysts of quartz, feldspar, and chlorite up to 4-5 mm. diameter. The idiomorphism of the quartz crystals and zoning in the feldspars are visible in hand specimen.

**Sodi-potassic rhyolite**

Half a mile south of the upper zig-zag on the Lyell-Comstock tram line and on the west side of the East Queen River, there are bold outcrops of grey feldspar porphyry (Q23), parts of which show banding. Colours of individual bands vary from pale to medium grey and hence they are not particularly distinct; they vary from a few millimetres to a few centimetres in thickness and are persistent.

Many of the phenocrysts, which average 2 mm. across show vestiges of multiple twinning but they are so clouded by alteration products that identification is difficult; extinction angle measurements indicate a composition of about Ab9. They are enclosed by a quartzose (?) microcrystalline groundmass.

This rock was described by Bradley (1954, p. 223) as a soda trachyte but chemical analysis (Table 1, No. 5) of a specimen reveals a high silica content (72.9%) and roughly equal amounts of soda and potash, so that the term sodi-potassic rhyolite seems more appropriate until more definite petrographic data are available.

**Keratophyres**

Sodic feldspar porphyries of various types occur throughout the Dundas Group and are the most common of the lava-types in the Queenstown area. They are particularly well exposed at several places along the Lyell-Comstock tram line, on the Zeehan road, east of Mt. Sorell, and north and east of Mt. Jukes. Similar rocks make up the bulk of the "volcanic assemblage" (of Campana et al., 1958) which outcrops between Rosebery and the Sterling River Valley, and between Farrell Siding and Tullah.

Deeply weathered lavas of this type occur at the northern end of the Queen River gorge where they are interbedded with beds of slate several feet thick; at their base the flows have an uneven surface and incorporated fragments of the underlying sediments. Along the Lyell-Comstock tram line the lavas are interbedded with conglomerates, pyroclastics and thin beds of banded slate.

These porphyries (Q15, 26, 27, 29, &c.) are characterised by subhedral feldspar laths up to 1 cm. long, very fine or microcrystalline matrices, and “clots” of albite-chlorite crystals which are
taken as evidence of deuteric action. The feldspar phenocrysts are generally intensely altered, either to albite-chlorite, or to a dense brown "felt", but examination of fresher remnants indicates a composition near Ab₅₂. Small amounts of calcite have been observed as an alteration product of feldspar. Augite and hornblende laths are generally considerably chloritized, and embayed quartz crystals rich in inclusions are rare constituents. The groundmass in some thin sections is speckled with haematite and apatite crystals are observed occasionally.

Most of the feldspar porphyries can be classed as keratophyres, assuming that term to include trachytic rocks characterised by a relatively high soda content, the presence of abundant albite or oligoclase and intense alteration of what little ferromagnesian mineral may have been present in the rock. In this area there are all gradations to sodic rhyolites, on the one hand and to albite andesites on the other, with the result that there are considerable ranges in the chemical composition of the keratophyres. The analysis of the feldspar porphyry given in Table 1, No. 6 is typical of a rather more basic variety and on chemical composition alone would be described as a trachyandesite. It is transitional between the true keratophyre and the hornblende andesite in that it contains laths of hornblende and augite up to 1 × 4 mm. in size and partly altered to chlorite.

**Augite Trachyte (?)**

East of the Queenstown-Lynchford road (see Fig. 3) the grass-covered hills display tors of feldspar-pyroxene rocks and on some of the tor faces can be seen haphazardly distributed fragments of sedimentary rocks and basic lavas.

This rock has a rudely equigranular, more or less reticulated texture, with a grain size of about 1 mm. and sparse interstitial material. The feldspar content varies from 30% to 90%, the remainder of the rock being composed of augite (diopsidite?), quartz and the microcrystalline base. The feldspars occur as stumpy subhedral laths, only slightly altered and with a composition of Ab₃₀. The augite crystals are anhedral, fractured, and partially altered to chlorite and locally form 30% of the rock. Clear albite veins traverse the rock and albite-chlorite aggregates (amygdales?) are common.

From petrographic data, Twelvetrees (1902, p. 282) considered the rock to be extrusive and described it as a syenite porphyry. Bradley (1954) included it in his "Lynch Conglomerate" formation and considered it to be a sediment derived from weathering of underlying basalts. He inferred an unconformity at its base.

The outcrop (Fig 3) is nearly two miles long and relatively narrow and it is clearly concordant with contiguous beds. From the evidence outlined it is impossible to differentiate between suggestions that the rock is a crystal tuff, a lava, or a sill, but it is most unlikely to be a sediment. Its texture is rather similar to that seen in tuffs near Lyell Comstock described on p. 43.

South-east of Little Owen there outcrops a rock of similar composition but with a more pronounced porphyritic texture; on its northern margin it is interbedded with sandstones and tuffs but the major part of the outcrop shows discordant relationships to the neighbouring rocks. This occurrence therefore has more the features of an intrusion and may represent a section through a volcanic cone.

The only other example of this rock-type with which I am acquainted occurs as boulders on the side of the Zeehan-Comstock road nine miles from Zeehan. The boulders are in Pleistocene moraine and are derived from the plateau south of Mt. Dundas. One of them contains randomly oriented fragments of finely banded siltstone that appear to have been hornfelsed at their margins by baking. The inference is that the host rock was hot and probably fluid.

In summary then, the available evidence suggests that this rock-type may occur as an intrusive, a pyroclastic or as a lava flow. In all probability all these modes of occurrences exist and they are clearly interconnected. Uncertainty as to its exact nature makes it difficult to define the rock but tentatively it is named after the lava-type of similar composition; several occurrences of other lava-types may similarly be actually intrusive or in part pyroclastic.

The chemical composition of a sample gathered 500 feet east of Lynch Creek bridge is given in Table 1, No. 7. Although clearly related to the keratophyres and andesites of the area it is a distinctive rock type and warrants a particular term. Despite the abundance of albite and high soda content, the presence of augite invalidates the term keratophyre and augite trachyte is preferred.

**Andesites**

Many of the hill-tops in the Comstock-Crown Hill area are capped by tors of hornblende and augite porphyry. These rocks are typically grey or pinkish-brown in colour and composed of phenocrysts of feldspar and ferromagnesian minerals (up to 3 × 1½ cm.) set in an aphantic groundmass. Similar rocks occur as boulders in the Lake Margaret moraine and outcrop west of Mt. Tyndall.

The texture of these porphyries (Q1, 7a, 7b, 8, 9, &c.) varies from seriate to porphyritic and the dominant mineral is feldspar occurring as phenocrysts and in the very fine grained groundmass. It is generally clouded and partly chloritised but remnants of multiple twinning are discernible and extinction angle measurements indicate a composition of Ab₃₀. Zoning is common, many crystals showing a clear albite fringe enclosing a core of altered Ab₅₂ feldspar.

Hornblende occurs as pale green crystals, usually deeply embayed and containing "inclusions" of the groundmass; in some cases the crystal is more of a skeletal framework. Some have a dark rim and a core of chlorite, others are represented by chlorite-haematite aggregates.
SOLOMON

Figure 3.—Geological Map of the Lynch Creek Area.
There is every variation from hornblende porphyry without augite to augite porphyry with little hornblende; the total percentage of ferromagnesian minerals is constant and thus where there is much of one mineral there is little of the other. The augite (diopside?) occurs as cracked and poorly formed crystals, generally partly chlorkitized.

Quartz occurs sparsely in the form of shards or in clusters of fragments showing unit extinction (presumably representing a fragmented crystal). The remainder of these rocks is usually made up of a dark, partly feldspathic matrix, fragments of feldspar-pyroxene lava, and ragged crystals of calcite. Albition of these rocks varies from slight to intense, a more severe case being illustrated in Plate 1, Fig. 1. Here the albite replacement has proceeded to such a degree that a brecciated appearance has resulted. The isolated blocks of hornblende porphyry average 3 dm. across and are separated by pale albite rock containing some hornblende crystals. Some of the thin sections show irregular replacement zones of albite and one alteration product of hornblende is considered to be hornblende-pyroxene lava. A specimen taken on the west slope of Crown Hill contained in the groundmass tiny 0·1 mm. spherulites of radiating albite (?) fibres; the spherulites gave the typical "cross-figure" when rotated between crossed nicsols. Near the summit of Crown Hill the porphyries show banding, "pebble" outlines and inclusions of sediments or tuffs.

Chemical analysis of typical slightly albitized hornblende porphyry from Crown Hill (Table 1, No. 8), taken in conjunction with petrographic data suggests a suitable name for the rock is a hornblende (or augite) andesite.

Scott (1954) regards these andesites as originating by metasomatism of basic lavas, the hornblende crystals growing out of chlorite which has developed through breakdown of pyroxene. I was unable to establish definitely the relationships between hornblende, augite and chlorite but gained the impression that chlorite formed bom both augite and hornblende. Probably the chloritization and albization are deuteric or secondary phenomena and are obscuring the magmatic hornblende-augite relationships. Scott (1954, p. 141) discusses at length the parallelism of hornblende lavas in a boulder of andesite observed by Banks in the Margaret moraine. The orientation of the lavas is at an angle to the cleavage and she is unable to explain the feature by her metasomatic theory for the development of hornblende. In all probability it is a primary flow phenomenon.

Basalts

Basalts outcrop along Lynch Creek, between 1000 and 4500 feet east of the Lynch Creek bridge (see Fig. 3). Individual flows are limited in extent, in both horizontal and vertical direction, and they are associated with tuffs, siltstones and volcanic breccias. Red and brown clays are exposed along Lynch Creek in the walls of open cuts made during the pursuit of quartz-gold veins. These clays are made up of tuffs and perhaps lavas that have been altered as a result of the mineralization, and since deeply weathered.

The lavas (Q82, 83, 88) are grey-green in colour and porphyritic, with phenocrysts of dark green pyroxenes up to 2 cm. long and smaller pale grey feldspar laths, set in an aphanitic ground mass. The pyroxene occurs in euhedral crystals, is only slightly chloritised and has been identified by Scott (1954) as diopside. Although Scott describes the feldspar laths as albite I found they were so altered that identification was difficult and that what few measurements could be made indicated a composition of about Ab₃. The relatively low soda percentage in these rocks (Table 1) fails to support Scott's determination.

Some specimens gathered from the volcanic sequence contain hornblende laths and are not unlike the hornblende andesites of the Crown Hill area.

The Lynch Creek lavas are associated with lenses of breccia; this is usually very coarse grained and consists of fragments of basalt embedded in a basaltic matrix which usually shows evidence of having "attacked" the fragments. These lava fragments differ from the matrix and also the majority of the flows in being markedly vesicular and amygdaloidal. The amygdales are spherical, up to 5 mm. in diameter, and are made up in the following ways:

(a) Lined with albite enclosing a zone of haematite and a core of chlorite.
(b) Lined with calcite enclosing a chlorite core.
(c) Filled completely by calcite.
(d) Lined with chlorite enclosing a calcite core.

The breccias are thought to have originated by the process of "autobrecciation" whereby the chilled gas-laden crust on a mobile lava becomes broken up by, and incorporated into, the still molten rock beneath.

A variation (Q66) of the normal porphyritic texture is seen in a flow outcropping on the west flank of Miners Ridge; it is about 50 metres thick and extends for several hundred metres. The lava is composed of diopside and feldspar but has a roughly equigranular texture with augite granules set in a lattice-work of fine feldspar laths. A chemical analysis is given in Table 1, No. 12, along with others of the Lynch Creek basalts (Nos. 10 and 11).

Another suite of basalts with which I have a brief acquaintance occur outside the area now under discussion but are of some interest as regards distribution of lava types. They outcrop on the Southern Ocean coast between Point Hibbs and High Rocky Point and consist mainly of spilitic basalts and associated tuffs and breccias, very well exposed in steep cliffs. The flows vary from a few inches to over 100 feet thick, are fine grained apart from local coarser zones, and are dark or medium grey in colour. While some flows are homogeneous others are severely brecciated, resulting in a "conglomerate" of lava fragments in a lava base, similar to those described from Lynch Creek.

The commonest type is a grey porphyry with feldspar phenocrysts and dark euhedral augites set in an aphanitic, dark ground mass. The field-
spars are slightly chloritized, euhedral, albite crystals generally less than 1-2 mm long; similar feldspar is discernible in the ground mass either lattice of laths of King Island, Penguin, and Zeehan. Some of these have "pillow" structures and are in all probability of submarine origin.

These basalts have strong spilitic affinities and are similar to spilites described by Scott from King Island, Penguin, and Zeehan. Some of these have "pillow" structures and are in all probability of submarine origin.

Summary of Lava Descriptions

The Queenstown area exposes a continuously variable series of volcanic rocks ranging from basalts through andesites to rhyolites. The alkali content is equally variable, potash-rich rhyolites being associated with soda-rich keratophyres and relatively alkali-poor basalts. Apart from the potash-rhyolites the dominant primary feldspar type varies from Ab₆ₗ to Ab₂ₘ in composition and the rocks have undergone varying degrees of albition. Though these rocks were predominantly extrusive the discordant contacts mentioned in a few cases suggest some of the occurrences are actually intrusive, in the form of small sills, plugs, or vent-fillings.

Association of the lavas with slates and sandstones suggests the environment was at least temporarily aqueous and it probably ranged in place and time from terrestrial to marine.

Distribution of Lava Types

A notable feature of lava distribution in the Queenstown-Darwin area is the concentration of the potash rhyolites along a narrow N-S belt which coincides with the axis of the West Coast Range and locally with the Great Lyell Fault Zone. (see Solomon, 1959, p. 36). The Darwin Granite, of very similar composition, also occurs along this fault zone between Mt. Darwin and South Darwin (Fig. 4).

Away from this belt the lavas vary from basalts to rhyolites but are characterized by dominance of soda over potash.

Viewing the West Coast as a whole, a tripartite division of lava types can be recognised:—firstly the narrow, relatively insignificant potash rhyolites occurring within a wider zone of variable but soda-rich volcanics, which in turn is flanked on the west by a still wider area in which the majority (say 80%) of the lavas are basic and sodic (see Fig. 4).

The central acid to basic zone in the Queenstown area has a poorly defined western boundary partly due to lack of exposure but to the north, and particularly in the Rosebery district, the western edge is sharp and coincides with the bedded series—volcanic assemblage interface. The volcanics here and to the north are made up of similar lavas to those seen near Queenstown. West of Rosebery and Williamsford, the lavas of the basic zone are subordinate to sediments and are largely basic in character (e.g. the Curtin Davis Volcanics and the Montana Melaphyre, Elliston, 1954, and Scott, 1954) and only rare acid lavas have been recorded (e.g. at Montana, Twelvetrees and Ward, 1910, p. 19). In the Point Hibbs-High Rocky Point area the percentage of lavas in most of the Cambrian sequence is greater than that of the sediments but the dominance of basic types is maintained.

Cause of the Distribution

Again insufficient is known of the Cambrian palaeogeography and tectonic environment to be certain of the cause of the lava distribution described.

An important feature, however, is that the lava zones are roughly parallel to the western margin of the Precambrian outcrop, a relationship which appears to be continued north-east of Bulgobac (Fig. 4). The present Precambrian cherty zone coincides with the Cambrian margin of the Tyennan Block, a relatively stable area which Carey (1953) considers to have had a considerable influence on Palaeozoic sedimentation and tectonics and which was very likely exposed during the Cambrian. The narrow zone of volcanoes that are assumed to have built up the "volcanic assemblage", and another related to the potash rhyolite distribution, are both parallel to the Tyennan Block margin and may be imagined as lines of off-shore volcanoes in the Cambrian depositional area. These volcanic lines roughly coincide with major Devonian structures such as the Great Lyell Fault Zone and probably are expressions of Cambrian movement along these features, which also appear to be influenced by the Tyennan Block.

Albite in the Lavas

Scott (1950) has described albite in ophitic texture with diopside in Cambrian basalts from King Island and this is strong evidence that albite occurs as a primary mineral. The presence of fresh albite in tuffs north of Queenstown (see page 43) suggests the mineral formed at an early stage in crystallization and the fresh albite crystals in the augite trachytes, and also in some of the quartz keratophyres, has the appearance of a primary mineral.

The existence of rims off clear albite on more calcic feldspar indicates a phase of albite formation related to soda-enrichment of the residual material of the cooling flow. Albition of an even later phase is seen in the veinlets and the chlorite-albite aggregates; these are observed mainly in lava flows but the presence of similar features in pyroclastic rocks indicates the mobile material circulated beyond the limits of the flows. The continuity between the chlorite-albite "clots" and amygdales suggests this late albition is related to the release of volatiles from the lava.

It is accompanied by chloritization of both primary feldspars and ferromagnesian silicates.
Scott (1954) considers the albitization to be part of a widespread late Cambrian metasomatic involving not only albitite introduction but also chloritization, silicification and carbonation. Actually chlorite and albitite are the only replacement minerals of Cambrian age in the Queenstown area and their mode of occurrence suggests they are of igneous origin. With regard to the source of the soda in the porphyries she concludes (p. 144) that it is derived from conate waters trapped in the lower levels of the eugeosynclinal sediments. I would prefer to leave the problem with Turner and Verhoogen (1951, p. 210), who state when discussing olivine basalt magma: “Differentiation of the magma, assimilative re-action with rocks situated in the basal levels of the geosyncline, concentration of magmatic water rich in soda, and chemical activity induced by entrapped sea water and rising conate waters squeezed up from deeply buried sediments, are all factors of possible significance in evolution of spilites and keratophyre”.

The albitite described above is a pinkish brown in colour and easily recognizable. Albitite with a reddish tinge is often seen in quartz veins in the schists at Mt. Lyell and in the vicinity of other sulphide deposits; the veins are clearly post-schistosity and related to Devonian mineralization and this albitite is perhaps derived by solution from soda-rich Dundas Group lavas and tuffs.

**PYROCLASTIC ROCKS**

**Agglomerates** (fragments over 32 mm. diameter.)

Conglomeratic and breccia lenses are a feature of the Dundas Group in the Queenstown area. When the constituents are entirely of igneous origin, then the rock is probably the result of auto-brecciation or may be a true agglomerate, particularly if associated with lavas. Many of the lenses show no stratification or sorting and could be products of the nue ardente type of eruption. For instance, coarse breccias occur in the potash rhyolite assemblage on Mt. Darwin and also in the granophytic host to the bornite veins at Lake Jukes; generally the fragments vary from a few cm. to a few decimetres across, are angular to sub-angular, and of similar composition to the matrix. There is little or no stratification and the fragments are randomly orientated.

Apart from these breccias, which are fairly certain of volcanic origin, there are many which contain varying percentages of sedimentary material such as siltstone and sandstone fragments. These could be volcanic, or pyroclastic rocks deposited in water or sedimentary rocks and proof of volcanic origin must depend on the recognition of volcanic glass; this, of course, is completely devitrified and in most cases it is impossible to decide whether the deposit is pyroclastic or whether it is reworked volcanic material. In the environment envisaged by the writer both types might be expected for the frequency of volcanic material suggests the Queenstown district in Cambrian times was one of considerable volcanic activity with eruption taking place from numerous centres of ephemerical nature. The constantly changing conditions would result in re-working of both land- and water-deposited pyroclastic and igneous material and would produce a complex, rapidly varying suite of sedimentary and volcanic rocks.

A typical conglomeratic rock (Q78) that is probably of mixed origin outcrops near the head of Whip spur some hundred metres west of the pink rhyolite. It occurs as lenses associated with finely banded mudstones and lavas and consists of coarse fragments of feldspar porphyry and sandstone and smaller quartz pebbles in a confused matrix of albite crystals, chlorite, and microcrystalline quartz (?) and feldspar (?). There is little stratification and the framework is disrupted. Veinlets and replacement zones of clear albitite are common.

**Tuffs**

Again it is difficult to differentiate between sedimentary deposits and volcanic deposits but there are a number of rock types that can be described as tuffs with reasonable certainty. For instance, at 8263/3584, on the Lake Margaret tram line west of Crown Hill (Fig. 1) and the north Margaret tram line west of Crown Hill (Fig. 1) is a distinctly banded rock (Q11) with an average grain size $\frac{1}{2}$ mm. and a texture typical of orthoquartzites (equidimensional grains with an interstitial cement forming say 8-10% of the rock). It is composed of subangular grains of albitite, quartz and hornblende (in order of abundance) and microcrystalline interstitial matter. The presence of banding, the texture and the anomalous composition (a high percentage of quartz compared with the local lavas) suggest the rock may a tuff.

Similar remarks apply to a rock (Q17) forming tars above the northern zig-zag on the Lyell-Comstock tram line. It is mottled or rudely banded in pinks and dark greens, the bands being several inches thick, and it is interbedded with sandstones and paraconglomerates (i.e. conglomerates with disrupted framework and high matrix-pebble ratios). The rock has a granular texture, an average grain size about 1 mm. and is composed of quartz and fresh albitite with sparse, fine grained, interstitial material which is locally chloritic (giving the dark green colour).

**SEDIMENTARY ROCKS**

The relative proportions of sedimentary and volcanic rocks in the Cambrian sequence near Queenstown vary considerably and can only be determined approximately due to difficulties in identification. General field observations indicate that the proportion of sediments varies from a little above, to a little below, 50% in any square mile.

**Conglomerates**

The problems of identifying sedimentary conglomerates have already been discussed and rocks which are probably volcanic have been described. Those with more obvious sedimentary affinities are typified by the following examples:—at 8218/3588 there are outcrops of a deeply weathered, grey, feldspathic breccia-conglomerate containing fragments, sometimes “shard”-like, of slates and sandstone. The framework is disrupted and the distribution of fragments is chaotic, indicating rapid sedimentation.
Lenses of conglomerate occur in fine, grey sandstone in the Queen River gorge (south of Lynchford); an interesting type consists of small (up to 5 x 15 mm.) ovate pebbles of grey chert or very fine quartzite lying with long axes on the bedding planes, the plentiful interstitial material ranging from pebble size to microscopic and comprising "chert", quartz and rounded fragments of keratophyre. The source of the siliceous pebbles could be Cambrian chert beds, but none are known in this area and it is more likely that they are derived from Precambrian material.

A paraconglomerate containing sandstone lenses outcrops half a mile north-west of the Lyell Comstock open cut and immediately north of the mottled tuff described on p. 43. It is composed of fragments of volcanic and sedimentary material of Cambrian origin but in addition contains rounded pebbles of vein quartz identical with those which form the bulk of the Owen Conglomerate. It is thus very similar to typical Jukes Conglomerate but is almost certainly a member of the Cambrian sequence. Similar rocks occur throughout the Dundas Group (e.g. Elliston's Razorback Formation at Dundas) and grade, with increasing quartz pebble content, to siliceous breccias and conglomerates almost identical with beds in the Owen Formation. Thus there is good evidence that a Precambrian quartz-rich source was intermittently feeding material in limited amounts into the Cambrian eugeosyncline. Density currents sweeping off the Precambrian shoreline (to the east?) and also off the unstable volcanic belt probably carried much of the material into the "bedded series" depositional area.

The similarity of many of these Cambrian paraconglomerates to the Jukes Conglomerate renders impossible the positive identification of the latter formation by lithology alone.

Sandstones

Several greywacke sandstones outcrop in the Queenstown-Darwin area. They are variable in character, lenticular, "dirty" or "muddy", and again it is often difficult to differentiate between greywacke sandstone and tuffaceous sandstone. Thin beds outcrop north-east of Lynchford and a considerable thickness may be seen east of Little Owen (see p. 46).

The one exception to the greywacke-type sandstone is the clear, quartz sandstone (Q101) which outcrops along the crest of Miners Ridge east of Lynchford (Fig. 3). It is about 100 feet thick and can be traced for four miles; its grain size varies between ½ mm. and 1/30 mm. and it consists entirely of subangular quartz grains apart from a small number of biotite flakes aligned parallel to the poorly developed cleavage. It is interbedded with shale beds and is of some importance in that its prominent outcrop makes it a useful marker horizon for determining structure.

Mudstones and Slates

These beds consist of finely, irregularly alternating bands of coarse siltstone (or very fine sandstone) and claystone. The claystone bands are thickest and consist of argillaceous material studded with quartz grains of silt grade; the narrower siltstone bands are usually composed of angular and subangular quartz grains of silt grade. Locally a tendency to grading is seen and the bands look similar to varves. Very fine ripple markings and erosion of tops of layers are seen occasionally and good examples of slump structures on a small scale outcrop on Whip Spur (see also Bradley 1954, p. 223). An indication of the frequency of banding is given by measurements made on the polished face of a specimen in the Mt. Lyell Mining and Railway Company's geological museum; in a nine-inch section the frequency of bands varies from one band per inch to 35 bands per inch.

Bradley (1954) has described these beds as exposed in Lynch Creek and termed them the "Miners Slates". As a cleavage is developed in the rock the lithology is correctly named but the Miners Ridge, from which the term is derived, is actually due to the quartz sandstone already described. "Miners Slates" (see p. 46) outcrop over a wide area as isolated lenses which merge along strike to other Dundas Group rocks, these facies variations render suspect the use of the slates as marker beds and suggest that mud and silt developed in local, ephemeral basins through a considerable range of the Cambrian.

The principal accessible exposures are along Lynch Creek, in the Queen River gorge, and along South Owen Creek (east of Queenstown). The most continuous outcrop mapped so far is in South Owen Creek, where the slates have been traced for 1¼ miles with a fairly constant thickness of 300 feet. They outcrop either in low rises or in stream beds depending on the relative resistance of the propinquent beds. Rather similar finely banded rocks outcrop over a wide area around the junction of the Garfield and Currie rivers (Fig. 1) but the stratigraphic position of these beds is not known (they may, in fact, be of Silurian age).

The slates of the Tullah area and of the Rosebery and Hercules mines are very similar in lithology to the Lynch Creek slates.

STRUCTURE

Determination of structures in the Cambrian rocks is hampered by rapid lithological changes along strike and the consequent lack of "marker horizons", and also by the fact that few of the rock types display identifiable bedding planes.

However, the presence of relatively persistent units east of Lynchford has allowed at least a partial portrayal of the fold styles in that area. The folds are only slightly asymmetrical, have steeply dipping limbs, and wave lengths averaging about 4000 feet. The axial trend in this district is 180° and there are fairly frequent changes in the direction of plunge. East of Queenstown the folds have NW trending axes and they plunge NW (Fig. 3). The north point shown on Fig. 3 is magnetic with declination 10° E.

Folds on these trends affect the Junee and Eldon Groups and clearly belongs to a Devonian orogeny, yet there is good evidence that faulting, uplift and
erosion affected the Cambrian sequence prior to de­position of the Junes Group. The presence of pebbles of Darwin Granite and Dundas lavas in the Jukes Conglomerate and the history of the development of the Jukes Trough (or Owen Rift valley), in which the Owen Conglomerate was later de­posited, clearly indicate major faulting on N-S trends followed by erosion of elevated areas (see Bradley, 1954, 56; Wade and Solomon, 1958; Campana et al., 1958). A good example is the uplift and erosion of the Darwin Granite late in the Cambrian and prior to deposition of the Owen Conglomerate. This tectonic phase has been termed the Jukesian Movement by Carey and Banks (1954). Clearly defined and indisputable unconformity between the Junes and Dundas Groups has not been proved on the West Coast, to the writer's knowledge, but it has been inferred at several places. A very strong indication of unconformable relationships in the Queenstown district is found near Lynchford, on grounds of discordance of lithology trends and observed dips (Fig. 3) and also at Mt. Sedgwick, where the base of the Jukes Conglomerate appears to cut through part of the Cambrian sequence. Yet at Mt. Misery (three miles east of Zeehan) the Gordon Limestone-Owen Conglomerate-Dundas Group succession is apparently conformable and Carey and Banks (1954) describe other areas showing a similar relationship.

As Bradley (1954, p. 205) has already pointed out the unconformity at Mt. Jukes described by Hills (1914) is actually a contact between gently dipping Junes Group conglomerates and Dundas lavas with a steeply dipping cleavage but in which no stratification is visible. A similar relationship is seen at the north end of Mt. Huxley where the basal beds of the Jukes Conglomerate and the underlying rocks of the Dundas Group are so sheared and altered that it is impossible to investigate the Dundas-Junes interface. A complicating factor in determining Cambrian-Ordovician relationships is the likelihood of the interface acting as a décollement surface during Devonian folding, thus possibly producing "pseudo­ unconformities". The chances of such a process taking place are increased in the areas where there are thick wedges of Owen Conglomerate that locally deflect and modify the Devonian stresses. Banks (1956, p. 204), by "unfolding" the unconformities he has observed and studying the residual pre-Junes dips, has concluded that the intensity of Jukesian folding varied from place to place. The force of his argument is reduced if it is accepted that décollement development is a real possibility.

In summary, there is evidence that faulting and upheaval preceded Junes Group deposition but the exact and precise nature of the movements is unknown. The variation in relationship between the Junes and Dundas Groups tends to support Banks' view of localised zones of tectonic activity but study of this tectonism is possibly hampered by décollement surfaces.

METAMORPHISM

As a general rule the Cambrian sequence has been only slightly metamorphosed, for if the effects of weathering and albite metasomatism are omitted, the rocks have undergone little change. Locally, however, the Cambrian beds have undergone severe alteration as a result of the embayment of sulphides, mainly pyrite and chalcopyrite. This alteration involved sericitization, chloritization, propylitization and hydration and the resulting rocks are described as sericite and chlorite schists. At Mt. Lyell the process reached a peak and the West Lyell open cut area is the centre of an aureole of alteration extending outwards for three­quarters of a mile. As described in Wade and Solomon (1958) the aureole is very crudely zoned, the central sericite zone passing out to chlorite-rich areas and then to rocks that have undergone milder alteration which is included in the term propylitization.

Typical of the early, propylitic stage of alteration is a sheared feldspar porphyry (Q64) from the upper reaches of South Owen Creek. The feldspar phenocrysts are roughly aligned with the cleavage and are also slightly rounded, giving a poorly developed augen structure. The ground mass is a quartzose (?) microcrystalline aggregate almost obscured by thick sheaths of sericite (?) which lie parallel to the cleavage. Alteration of phenocrysts is in some cases slight but in others the feldspar is almost totally replaced by caliche growing from the crystal centres in reniform masses with dark rims. The initial phase of this alteration process is seen as the development of tiny caliche rhombs with parallel orientation throughout the feldspar crystals. Many other cases could be described but the complete treatment of the various types of hydrothermal alteration is the subject of another study.

The mineralization with which this metamorphic aureole is associated is of Devonian or Carboniferous age, as the sulphides are later than the Devonian cleavage and locally replace Junes Group rocks, and the remnants of basal Permi.an beds in the area are relatively undisturbed, have no cleavage and are unmineralized.

Other important areas of hydrothermal altera­tion apart from West Lyell are concentrated on the N-S line of copper deposits that mark the Great Lyell Fault Zone between Comstock and South Darwin.

Kaolinization of quartz porphyries is described by Bradley (1954, p. 233) near the Hospital at Queenstown and has been noted by the writer both south and north of Queenstown in small irregular zones. The alteration affects the entire rock apart from the quartz crystals and the result is a whitish or greenish clay that is studded with the residual quartz phenocrysts. The process appears to have no relation to mineralization and might in fact be a deuteric effect. It is selective in that it affects some porphyries but not others, yet no differences in environment, either structural or sedimentary, are discernible. Bradley, however, relates the kaolinisation to NE faults which he believes to be major controls over ore deposition.
STRATIGRAPHY

The frequent lithological variations, the difficulties in elucidating structures and lack of fossils have prevented the establishment of a stratigraphic succession for the Group in this area. However, Bradley (1954, p. 221) has suggested the following sequence in the Lynch Creek area (youngest at the top):

<table>
<thead>
<tr>
<th>Sequence</th>
<th>Thickness (feet)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lynch Conglomerate</td>
<td>3,000</td>
</tr>
<tr>
<td>Battery Volcanics</td>
<td>4,000</td>
</tr>
<tr>
<td>Miners Slate</td>
<td>3,000</td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td><strong>10,000</strong></td>
</tr>
</tbody>
</table>

The recent field work suggests that this succession has been measured across an anticline in which the Lynch Conglomerate and the “Miners Slate” occupy opposing limbs of the fold and are thus roughly equivalent (Fig. 3). The mapping has also shown that the lithologies change rapidly along strike and any sequence is therefore of only very local significance; this is at variance with the data on Bradley’s maps which show the beds extending north and south of the type section for considerable distances. Part of Bradley’s metamorphic hypothesis (1954, p. 224) is based on his belief that he could observe gradational metamorphism of the Lynch Conglomerate and Battery Volcanics along strike from Lynch Creek to Roaring Meg Creek. Actually the beds are not continuously exposed as his maps indicate, due to structural contortions (Fig 3); and in any case I would attribute the rock-type variations to original facies changes.

With regard to the terms Lynch Conglomerate, Battery Volcanics and “Miners Slates” the new structural picture renders suspect the proposed age relationships between these units, and other field evidence shows that whatever relationship may exist is likely to be a really restricted one. The rock types of the Dundas Group in this area, both volcanic and sedimentary, were almost certainly laid down in localized and ephemeral depositional environments and it is difficult if contemporaneous deposition of any one rock type or particular association of rock types over a wide area ever took place. This being so, it would seem unwise at this stage to assign formal names to individual members of particular sequences that have been measured in this area and it is suggested that Bradley’s terms be discarded. The terms he used are also unsatisfactory for other reasons, viz.:

Lynch Conglomerate: I have been unable to match the description given for this formation with field observations and suspect that a considerable proportion of it is occupied by augite trachyte or trachytic tuff, which Bradley considered to be a conglomerate. This formation has been correlated with the Sorell Conglomerate and the Dora Conglomerate but the former is clearly part of the Jukes Formation and similar remarks probably apply to the Dora Conglomerate. The structural interpretation shown on Fig. 3 suggests the Lynch Conglomerate is equivalent in part to the Miners Slate and also that it is overlain unconformably by the Junee Group, both observations being at variance with Bradley’s description.

Battery Volcanics: The age relationships are suspect and the locality name is unfortunate in that the “Battery” (of the King River Gold Mining Co.) is now non-existent and its position is seldom shown on published maps, new or old.

Miners Slates: Again the proposed stratigraphic relationship to the basalts is questionable and again the locality name could be improved; Bradley used the prefix “Miners” under the impression that the slates outcropped on Miners Ridge but as has already been mentioned (p. 44) this feature is due to the presence of a sandstone bed. In Wade and Solomon (1958, pp. 374-375) the terms Lynch siltstones and Miners sandstone (the capital ‘S’ in the 1958 text is a typographical error) were used as strictly lithological terms but this form of nomenclature is not now considered to be satisfactory, mainly because these terms are so similar to formation names and these carry so many implications. If it is desired to refer to the lithologies described under the terms Battery Volcanics and Miners Slates, and it is often convenient to do so during field work, then such terms as Lynch Creek-type volcanics and Lynch Creek-type slates could be used; similarly Crown Hill-type andesite, Whip Spur-type rhyolite, &c.

THICKNESS OF THE DUNDA GROUP

The only reliable measurements available have been made near Little Owen and on Lynch Creek. The former area was mapped in conjunction with work on the Mt. Lyell Mine leases on a scale of 1” = 100 ft. and the measurements obtained are as follows:—

<table>
<thead>
<tr>
<th>Lithology</th>
<th>Thickness (feet)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Conglomerates (?)</td>
<td>1,000</td>
</tr>
<tr>
<td>pebbly sandstones</td>
<td></td>
</tr>
<tr>
<td>and sandstones, probably partly volcanic, outcropping near Little Owen summit</td>
<td>700</td>
</tr>
<tr>
<td>Greywacke sandstones and tuffs, and augite trachyte (?)</td>
<td>300</td>
</tr>
<tr>
<td>Lynch Creek-type slates, exposed along South Owen Creek</td>
<td>500</td>
</tr>
<tr>
<td>“Dirty”, fine grained sandstone</td>
<td>2,500</td>
</tr>
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</table>

The rocks east of the Lynch Creek bridge are relatively undisturbed and they give the following sequence, measurements being by tape and level:—

<table>
<thead>
<tr>
<th>Lithology</th>
<th>Thickness (feet)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gordon Limestone</td>
<td>600</td>
</tr>
<tr>
<td>Owen Conglomerate (?)</td>
<td>Grey-brown quartzite</td>
</tr>
<tr>
<td>Jukes Conglomerate (?)</td>
<td>Grey to purple sheared paraconglomerate with pebbles of porphyry and sediments, and a few of quartz</td>
</tr>
</tbody>
</table>
M. SOLOMON

Dundas Group: Poorly exposed section; mainly finely and regularly banded pale grey shaly sandstone with lenses of para-conglomerate. Also lenticular quartz porphyry lavas, part kaolinised. The thickness is estimated at 300 feet.

Augite trachyte (?) 250 feet.
Basalts, tuffs and breccias plus 1000 feet.

Total of Dundas Group exposed—1550 feet.

While drawing horizontal sections through the area it was found necessary to assume a thickness of at least 4000 feet for the Group and generally a figure of twice this was used.

Hills (1914) suggests a thickness of 21,000 feet in the Jukes-Darwin area but he points out that the beds may be repeated by folding.

THE DARWIN GRANITE

The Darwin Granite is the only large intrusive body in the area and it is of Cambrian age. Despite Bradley’s assertions that it is Devonian, the occurrence of granite pebbles in the Jukes Conglomerate at South Darwin and at Mt. Sorell is clear proof that it is older than the Junee Group. The basal part of the Jukes Formation at South Darwin contains numerous and distinct sub-angular fragments (up to 24 x 30 cm.) of pink granite, white granite and haematitic material, together with smaller pebbles of quartz and chert (Plate 1, Fig. 2). The pebbles and fragments are clearly defined in a matrix of granitic detritus. In the Jukes Conglomerate exposed on the east face of Mt. Sorell the pebbles of granite and haematitic material are fewer in number and smaller in size.

The granite occurs as a vertical tabular body aligned north-south and extending from Mt. Darwin to South Darwin. Its outcrop is roughly three miles long and half a mile wide. The adjacent rocks to south and east contain pebbles of granite and are therefore younger but to the west and north consist of the pink potash rhyolites and tuffs, pre-granite in age.

The granite body is complex, being composed of parallel sheets of differing composition with a predominance of granitic types, chief of which are a pink orthoclase-quartz rock and a white plagioclase-quartz rock.

The pink granite (D42) forms a long line of tors on the eastern side of the Mt. Darwin-South Darwin Spur. It is coarse-grained and has a texture that is typically granitic, being not seriate yet not quite equigranular; intercrystal boundaries are commonly sutured. The mineralogy is simple, the only constituents being pink orthoclase, colourless quartz and pale green altered plagioclase feldspar and generally the mineralogical composition is orthoclase > quartz > plagioclase. However, in places the quartz content exceeds that of the feldspars and they occur in approximately equal quantity so that these varieties should be termed adamellites.

The orthoclase is fresh and most sections under the microscope show a fine, wispy cleavage; they often exhibit relatively coarse (“vein”) perthitic texture, the orthoclase enclosing parallel tongues of plagioclase (albite ?) in which multiple twinning is perpendicular to the length of the tongues. Quartz crystals are finely fractured and show undulose extinction. The plagioclase is considerably altered to chlorite, kaolin and other minerals and its composition is difficult to determine but the few reliable measurements indicate a composition of Ab60. Ferromagnesian minerals are rare and consist of ragged crystals of biotite and chlorite. The two available chemical analyses (Table 2, Nos. 1 and 2) resemble those of potassic granites.

The white granite (D40) occurs on the west flank of the pink variety; it is coarse-grained, locally almost pegmatitic, and is composed entirely of equal percentages of altered, cloudy plagioclase feldspar and colourless quartz. All that felspar hampers identification but it appears to be oligoclase; its alteration products are whitish in colour in contrast to the greens of the plagioclase in the pink granite. The crystals are loaded with inclusions of quartz, suggesting it crystallized after that mineral. As with the pink granite, sutured crystal boundaries are characteristic. A chemical analysis of a specimen of this granite type, given in Table 2, No. 3, shows a reversal of the K2O/Na2O ratio of the pink variety.

Workings for haematite and copper minerals in the granite prove that the alteration of the feldspars in both types extends many feet below the surface, indicating that it is not a weathering effect.

Within the granite complex are thin sheet-like tongues of sedimentary material trending parallel to the length of the granite. Between the pink and white granites is a long zone of schists showing patchy mineralization and in the pink granite occur dark-grey, sharply defined, hornfelsic lenses not unlike Lynch Creek-type slates; Bradley (1954, p. 325) emphasises the alternation of granite and hornfels by giving details of a traverse across the complex.

The outer margins of the granite are unfortunately not now exposed but Hills (1914) states that the “boundary-line between granite and the felsites which it has intruded is sharp and well-defined. This line of contact has been opened up in several places by trenching, and the granite can there be seen in contact with the felsite, being quite as coarse-grained as in the interior of the mass, no transition into finer grained varieties being observable at the margins”. Certainly the felsites on the western side of the granite show little alteration that can be ascribed to granite intrusion, unless the unduly heavy haematite-magnetite veining is related to the granite.

A feature of the Darwin Granite is the complete lack of apophyses and aplite or pegmatite veins. This and other points such as the lack of contact metamorphism, complex composition and the tabular shape serve to distinguish it from the massive, irregularly shaped and monotonous stocks and batholiths of Devonian age such as occur at Mt. Heemskirk and Mt. Meredith. The Murchison Granite, which outcrops in the gorge south of Mt. Farrell, is similar in form to the Darwin Granite and even more variable in composition.
Though it has not been studied in detail, sufficient is known to indicate that it contrasts too with the Devonian granites and may well be of Cambrian age.

Bradley (1954) has suggested that the Darwin Granite is a replacement of sedimentary rocks but the evidence he presents, and the evidence since gathered by the writer, is insufficient to provide a conclusion on the mode of emplacement. However, the similarity in composition between the pink Darwin Granite and the potassic rhyolites, and between the white granite and the keratophyres suggests a genetic relationship. It is highly probable that the granites and the lavas represent the intrusive and effusive phases of a single late (?) Cambrian unweiling of acid magma, and that the granite core of the volcanic pile rose up at a late stage to become emplaced among the products of the earlier eruptions.

ACKNOWLEDGEMENTS

I am indebted to the Mt. Lyell Mining and Railway Co. for permission to publish this paper, and also to Mr. M. L. Wade (Chief Geologist of the Mt. Lyell Mining and Railway Co. until 1958) for guidance and assistance both in the field and the office.

Thanks are also due to Messrs M. R. Banks and A. Spry (University of Tasmania) for helpful discussion and critical reading of the manuscript.

REFERENCES.


Carey, S. W., 1950.—Report to North Broken Hill Ltd. Unpublished.


TABLE 1

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9. Hornblende andesite, Mount Shasta, California (H. N. Stokes), as quoted in Hatch, Wells, and Wells, 1949, p. 271.
### TABLE 2

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Total | 100.06 | 98.90 | 100.41 |

Fig. I.—Albitised hornblende andesite from Crown Hill. The pale grey rock is largely fresh albite, the darker grey is hornblende-feldspar porphyry.

Fig. II.—Jukes Conglomerate, South Darwin Peak. The boulders to the right of the hammer are pink granite; the large one to the left is haematitic schist. The lack of contrast between boulders and the matrix is due to inadequate lighting of the subject at the time of photography.
The Mineralized Rift Valleys of Tasmania

By

B. Campana, S. B. Dickinson, D. King and
R. S. Matheson

(Proceedings No. 192 December, 1959)
DISCUSSION AND CONTRIBUTIONS

THE MINERALIZED RIFT VALLEYS OF TASMANIA

By B. Campana, S. B. Dickinson, D. King and R. S. Matheson

(Stillwell Anniversary Volume, December, 1958.)

Mr. M. Solomon (University of Tasmania): Campana, Dickinson, King and Matheson have made an important and stimulating contribution to the geology of the West Coast of Tasmania. However, certain of the hypotheses submitted require discussion as they appear to be founded on evidence which has not been subjected to critical examination.

THE OWEN CONGLOMERATE AND ITS ENVIRONMENT OF DEPOSITION

In presenting the suggestion that the Owen Conglomerate is a terrestrial formation, its striking red colours are brought in as supporting evidence with the remark that there is little need to dwell on the significance of 'red beds'” (p. 50). Now the authors have stated (p. 49) that the red colouration is most prevalent in the uppermost beds of the Owen Sequence (the Upper Owen of Wade and Solomon, 1958) and I would confirm this from my own observations, yet the following points strongly suggest that these beds were deposited in a marine environment:

(i) Worm tracks and borings are frequent, and trails and related features probably made by some merostome creature are found on bedding planes in the Upper Owen Sediments. If these are not marine then they are a record of the earliest known terrestrial forms, for the Owen Conglomerate is older than the Gordon Limestone and contains Lower Ordovician fossils (M. R. Banks in Hughes, 1957).

(ii) I have found brachiopods and other fossils on this horizon approximately one mile south of the King River ridge, east of Queenstown, and
(iii) Gregory (1905, p. 71) has identified crinoid stems and brachiopods in sandstones both to the north and south of Gormanston. His locality map indicates that the sediments belong to the Upper Owen.

If this evidence is accepted then the conclusion follows that the red colouration is not necessarily significant with regard to the depositional environment. The red colour is due to the presence of hematite which was apparently being deposited as detrital material during formation of the Owen Conglomerate. It is found as pebbles in the Jukes Conglomerate (just below the Owen Formation) and was probably derived in part from the hematite-magnetite veining seen in some of the Cambrian lavas and in Precambrian rocks.

A number of other significant features of the Owen Formation have also to be accounted for by the exponents of the terrestrial theory:—

(a) The coarse conglomerates which make up the Lower Owen member are usually yellowish grey in colour and very seldom show the red and pink tinges that might be expected in the piedmont environment postulated by the authors.

(b) I have mapped a number of pink and grey siliceous conglomerates and breccias in the Dundas Group, similar and locally identical to members of the Owen Formation but clearly in the Cambrian sequence. They occur along the west flank of the Sticht Range and north of Rosebery, the latter occurrences being of particular interest in that they are interbedded with monotonous sequences of mudstones and slates. These mudstones are very similar to, and are continuous with, sediments of the Dundas Group west of Rosebery which contain autochthonous marine fossils (Öpik, 1951).

This close association of marine sediments and siliceous conglomerates strongly suggests that the latter were deposited in a marine environment (perhaps by density currents). Therefore any proposal that the often identical Owen sediments are non-marine must be completely proved
The occurrence of Cambrian siliceous breccias and conglomerates near the Sticht Range, 5 miles north and along strike from Lake Spicer, leads one to suspect that the talus breccia illustrated in Fig. 5 (a) of the authors’ paper may actually be of Cambrian age. If this is so, then another item of evidence in support of the terrestrial theory (p. 49) is disqualified.

(c) A feature of the Owen Conglomerate is the almost complete absence of Cambrian pebbles and fragments. Mr. Wade has described to me a small area on Mt. Lyell where the Owen contains a number of Cambrian pebbles but this is so far unique and during examination of many square miles of Owen outcrop, from the D’Aguilar Ranges to Valentines Peak, I have seen very few pebbles of Cambrian origin. In attempting to envisage the authors’ piedmont environment I find it difficult to see why this should be so, but exclusion of the Cambrian rocks from the depositional area is easier to understand under the marine conditions suggested by Carey and Banks (1954, p. 267), in which the relatively softer rocks fail to survive prolonged re-working near shore lines. Such an environment would also account for the fairly high degree of sorting and lack of clay matrix in the Owen sediments.

In summary, I feel that although Campana et al. have provided some support for a terrestrial origin their hypothesis is by no means proved and must be backed by more critical evidence. It may well be that the zone of deposition lies close to the dividing line between the terrestrial and marine but the problem will probably remain unsolved until the sediments in question are examined in detail.

FAULTING AND MINERALIZATION

The new significance of the rift valley theory with respect to mineral exploration appears to have been over-emphasised by the authors, as pointed out by Hall and Cottle (1959).

The distribution of the known orebodies of the West Coast Range is related to certain structural features and
particular lithological boundaries. These controls have been recognised for some time and are still being studied in current mineral exploration programmes, e.g. in 1905 Gregory described the Great Lyell Fault Zone\(^1\) between Mt. Jukes and Mt. Read and he pointed out the tendency for ore to occur where the Fault Zone was intersected by transverse structures. This fault zone has since been extended north to Mt. Farrell by Carey (1953), who preferred to use the term Lyell Shear, and its approximate position is shown by Hall and Cottle in Fig. 1 (1959). The initial phases of the L.E.E. exploration work in South-West Tasmania in 1956 were mainly concerned with mapping regional geology and, in particular, studying the location and economic significance of the continuation of the Great Lyell Fault Zone south of Mt. Darwin. It has been suggested by many writers (e.g. Carey, 1953; Wade and Solomon, 1958) that movement took place along this feature at the close of the Cambrian and that it formed the western margin of the basin of deposition of the lower Owen beds.

Another prominent ore control that had been examined in detail by exploration geologists prior to the recent activity is the western boundary of the “volcanic assemblage” (Campana et al. 1958) which Hall et al. (1953) have suggested is in part faulted. The orebodies at Williamsford, Rosebery and the Pinnacles occur along, or close to, this feature.

Now Campana et al. have proposed, for the first time, that the Owen Conglomerate and the volcanic assemblage were deposited in a rift valley and also that the western edge, the Owen Rift fault, is the most important control over mineralization in the West Coast Range. The maps accompanying the authors’ 1958 paper show this fault as a linear feature linking Elliott Bay and Mt. Bischoff but in a later publication Campana and Dickinson (1959, p. 137)

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\(^1\) I support Hall and Cottle in their plea for retention of Gregory’s term, the “Great Lyell Fault”, though I suggest it would be improved by describing it as a fault zone (defined by Willis in “Geologic Structures” as a complex break in which the fissures occupy a wide strip). Clearly the long zone stretching north from Mt. Darwin and containing major structural elements and mineral deposits was recognised by Gregory and his term should therefore receive priority over Carey’s “Lyell Shear”, with which it is synonymous.
state that its exact location is doubtful and that "step faulting and/or satellite structures" are to be expected. They also say on the same page that it is coincident with the Great Lyell Fault Zone and it seems clear that the Owen Rift fault is actually a re-interpretation of previously known features. Thus the rift valley hypothesis, while of considerable academic importance, has not yet lightened the burden of the exploration geologist.

The postulated northerly continuation of the Owen Rift fault through Mt. Bischoff is not supported by the results of recent field work; the N-S faults of the area do not appear important and the tin deposits are probably related to a local and independent "high" in the Meredith Granite. The volcanic assemblage seen in the Pinnacles-Rosebery area does not occur near Waratah. If the tin deposits are independent of this fault, then the authors' hypothesis of stratigraphic zoning of mineral deposits along its length (p. 55) is considerably weakened. Certainly there is a geographical distribution of the copper and silver-lead mineralization on the West Coast but it seems premature to discuss it in terms of little known Cambrian stratigraphy. It would be equally reasonable, on available evidence, to suggest that the Pinnacles, Rosebery, Tullah and Mt. Lyell deposits were all at the same stratigraphic horizon.

AGE OF THE MINERALIZATION

The authors suggest on p. 53 that the age of the mineralization on the Great Lyell Fault Zone is doubtful and in succeeding discussion they consider a Cambrian age to be very probable.

The presence of pebbles of hematite and granite in the Jukes Conglomerate, and of chromite in the Owen Conglomerate (see Wade and Solomon, 1958) is evidence of Cambrian granite emplacement, hematization and chromite mineralization, the latter probably associated with serpentinite intrusion and perhaps nickel sulphides (Williams, 1958). The granite involved in this phase is the Darwin Granite, which occurs 13 miles south of Queenstown and
shows no evidence of having any associated copper mineralization.

It is equally clear that the Mt. Lyell deposits are Devonian in age, the main evidence being as follows:

(a) The ore is unsheared and post-schistosity.
(b) The schistosity is related to a structural pattern affecting rocks up to Lower Devonian in age.
(c) The evidence at hand suggests these Devonian movements were far more widespread and severe than those of Cambrian age.
(d) In the North Lyell stopes, bornite penetrated and probably replaced Owen Conglomerate.

Points (a), (b), and (c) also apply to the deposits at Rosebery, Williamsford, Pinnacles and others in the West Coast Range. The only other major deposits that can be dated are those of the Zeehan field where the silver-lead-zinc veins cut rocks ranging from Cambrian to Lower Devonian (Florence Quartzite) and clearly belong to a late Palaeozoic mineralization. Their zonal relationship to the Heemskirk granite is a strong indication of the age of that intrusion (Editorial Contribution, 1953).

Nye (1932) has described silver-lead-zinc mineralization cutting fossiliferous Silurian strata eight miles west of Waratah. This mineralization appears to be related to the Magnet orebody, which also, therefore, is probably post-Silurian.

To summarise the available evidence, it seems clear that the close of the Cambrian (Jukesian movement) was accompanied by intrusions of granite and serpentine (the latter containing chromite and possible nickel sulphides) and by hematisation. The Devonian orogeny equally clearly included widespread sulphide mineralization and granite intrusion and hence it seems an unnecessary complication to postulate important Cambrian copper mineralization followed by Devonian recirculation and concentration by "juvenile hydrothermal solutions".

I am grateful to M. R. Banks and K. L. Burns (University of Tasmania) for assistance in preparing this note.
REFERENCES


APPENDIX B

Supporting Papers

1. The Age of the Gawler Range Porphyry.
3. The Geochemistry of Darwin Glass.
THE AGE OF THE GAWLER RANGE PORPHYRY

By R. K. Johns and M. Solomon
THE AGE OF THE GAWLER RANGE PORPHYRY (1)

By R. K. JOHNS and M. SOLOMON (2)

[Read 14 August 1952]

SUMMARY

Recent field work has disclosed that the Gawler Range porphyry, hitherto regarded as a pre-Sturtian extrusive, intrudes what are considered to be basal Cambrian sediments in the Corunna area on Upper Eyre Peninsula. The intrusions of porphyry and the unstressed granites of the Middle-back Ranges area are thought to have taken place in the final stages of the orogeny following upon deposition of Cambrian sediments on terraces of the Adelaide miogeosyncline.

INTRODUCTION AND PREVIOUS WORK

While engaged on a reconnaissance geological survey of the Iron Knob-Port Augusta district (Corunna Military Sheet) the authors found evidence to indicate that the Gawler Range porphyry is intrusive into sediments which are probably of Cambrian age.

Jack (1922) first reported pebbles of Gawler Range porphyry in the Corunna Range conglomerate. He believed the conglomerate represented a lower Cambrian overlap onto the Pre-Cambrian foreland of Eyre Peninsula. He believed the porphyry to be extrusive onto a Pre-Cambrian surface (1917). In 1947, Mawson also reported occasional porphyry pebbles in the conglomerates and concluded that the latter and the equivalent Tent Hill formation were younger than the porphyry. On lithological grounds he correlated these sediments with basal Adelaidean of the Emeroo Range (east of Port Augusta) and therefore dated the intrusion as pre-Adelaidean. Segnit (1939) suggested that both the Tent Hill formation and the Corunna conglomerate were equivalent to the Pound quartzite horizon which he placed in the Upper Pre-Cambrian. Ward (1949) also favoured an upper pre-Cambrian or basal Cambrian age for the conglomerate.

GENERAL GEOLOGY

The Gawler Range porphyry extends into the north-west corner of the Corunna military sheet and eastwards to the longitude of Myall Creek H.S. (see plan). The bordering sediments in this area are well-worked conglomerates, quartzites, sandstones and sandy shales. Highly-folded quartzites and shales are exposed over much of the area between the Corunna Range and Pandurra H.S. and these are overlain unconformably by conglomerates, sandstones, shales and flagstones outcropping in the Corunna Range and in the Roopena H.S.-Lincoln Gap area. They have been squeezed into a broad synclinal fold in the Corunna Range but at Lincoln Gap the beds are horizontal or show a gentle easterly dip. The age of these sediments is uncertain though it is believed that they represent the terrace or shelf facies of the lower Cambrian. The Corunna Range sediments are thought to be a coarse facies of the sandstone outcropping near Roopena H.S.

The underlying contorted quartzites and shales may be basal Adelaidean or equivalent to beds of the iron ore series in the Middleback Range, i.e., Middle Pre-Cambrian.

(1) Published by permission of Director of Mines.
(2) Department of Mines, Geological Survey of South Australia.

Trans. Roy. Soc. S. Aust., 76, December, 1953
The general succession is summarized below:

**CORUNNA RANGE**

- Cambrian
  - Corunna sandstone, conglomerate and Quartzite

**LINCOLN GAP**

- Lincoln Gap Flagstones
- Tregalana Shales
- Corunna sandstone

--- Unconformity ---

Adelaidean or
Middleback Group: Quartzites and Shales.

This sequence may be matched at Whyalla, Mt. Laura and in the Moonabie Range (Miles, Bulletin 33 of Geol. Surv. of South. Aust., unpublished).

--- Fig. 1 ---

**THE AGE OF THE PORPHYRY**

The following are the more important facts relating to the age of the Gawler Range porphyry and its mode of emplacement:

1. A small body of Gawler Range porphyry intrudes the Corunna conglomerate three miles south of Wartaka H.S. The outcrop is of irregular shape and cross-cuts the bedding and at its margins shows chilling and partial digestion of the sediment.

2. Two miles south-south-west of Roopena H.S. a plug of similar porphyry about 20 acres in area is flanked by Corunna sandstone. Near the contact with the igneous rock the sandstone is locally recrystallised and a dark hornfelsic facies is developed.

3. The Corunna Range bifurcates north of the latitude of Tassie Creek, the two arms representing the limbs of a broad north-pitching synclinal fold, though the eastern limb is disturbed by faulting.
The low area between the ranges, the axis of the syncline, is now occupied by porphyry which, judging from the steepness of the limb of the fold and the relative elevations of the ranges and the valley between them, apparently intruded and digested the sediments in the axis of the syncline (see horizontal section).

Two other porphyry stocks were mapped in Corunna sandstone about four miles west of Pandurra H.S., but no contact phenomena were observed. The porphyry of these small bodies is generally of finer grain than in the main axis.

The evidence outlined above proves that the Gawler Range porphyry in this area is intrusive into the Corunna sediments. The form of the intrusion is indicated in the accompanying horizontal section.

Both Jack (1922) and Mawson (1947) reported the presence of porphyry pebbles in the Corunna conglomerate and concluded that the intrusion was older than the sediment. The authors, however, failed to find any such pebbles anywhere in the Range.

Howchin (Trans. Roy. Soc. S.A. 1928) recorded red and grey porphyry pebbles in the Sturt Tillite in the Crystal Brook area, and various geologists (Mawson, Sprigg, etc.) have suggested that they are similar to the Gawler Range porphyry. It is possible, however, that they were derived from the Wallaroo-Moonta or some other porphyry. Jack (1917) states that the Wallaroo-Moonta and the Gawler Range porphyries are similar in appearance.

RELATED IGNEOUS BODIES

Several other unstressed magmatic granites occur in the Middleback Range area, the largest of which is the Charleston granite. This outcrops near Moonabie at the southern end of the Range and is a pink, coarse-grained porphyritic variety. It intrudes a dense grey porphyry which is itself intrusive into the Adelaidean grit of the Moonabie Range succession. No pebbles of either grey porphyry or granite are to be found in the conglomerate which unconformably overlies this grit. The conglomerate is thought to be equivalent to that in the Corunna Range.

Evidence therefore indicates that the granite post-dates the Corunna sedimentation.

A uniform-grained tor granite outcrops in the Burkett Hill area, west of Iron Knob, and appears to felspathise and isolate outcrops of Corunna conglomerate. The time relationship between the granite and the Gawler Range porphyry is not known.

The intrusion of these unstressed granites and the porphyry is thought to have taken place in the final stages of the orogeny which followed upon the deposition of the Cambrian sediments on terraces of the Adelaide mio-geosyncline.

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REFERENCES

Jack, R. L. 1917 The Geology of the Moonta and Wallaroo Mining District. Geol. Surv. of S. Aust., Bull. 6
Jack, R. L. 1922 The Iron Ore Resources of South Australia. Geol. Surv. of S. Aust., Bull. 9
HOWCHIN, W. 1928 The Sturtian Tillite and Associated Beds in the Western Scarps of the Southern Flinders Ranges. Trans. Roy. Soc. S. Aust., 52

MAWSON, D. 1947 The Adelaide Series as developed along the Western Margin of the Flinders Ranges. Trans. Roy. Soc. S. Aust., 71

SEGNI, R. W. 1939 The Precambrian-Cambrian Succession in South Australia. Geol. Surv. of S. Aust., Bull. 18

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GEOCHEMICAL AND GEOLOGICAL EVIDENCE FOR THE ORIGIN OF DARWIN GLASS

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