GEOLOGY AND MINERALIZATION
OF THE
CAPE HORN - LYELL COMSTOCK AREA, MT. LYELL

by

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University of Tasmania

February, 1971
"... there must be few places in the world where everything is as simple as at Mt. Lyell."

J. Bradley, 1957.
FRONTISPIECE

Southern slopes of the Comstock Valley looking east. Prominent outcrop at right is Comstock chert.
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ABSTRACT

Stratigraphically the disseminated chalcopyrite-pyrite ores of Cape Horn occur in ignimbrites (?) whereas altered lithic tuffs higher in the sequence of the Mt. Read Volcanics are the host rocks for the chalcopyrite-pyrite-bornite mineralization of Lyell Comstock. Chalcopyrite-sphalerite-galena deposits in replacement (?) chert provide a mineralogical transition between the copper ores and the banded exhalative pyrite-galena-sphalerite lenses of the Tasman Shaft.

On the basis of ore and gangue mineralogy and texture correlations are tentatively proposed between the West Lyell and Cape Horn, and between the Lyell Comstock and North Lyell-Crown Lyell deposits. This stratigraphic correlation is in sympathy with the distribution of trace elements in the sulphides.

Geochemical studies indicate that at least some of the Cape Horn ore represents an addition to, rather than a replacement of, the host rock. Copper mineralization occurs in areas of sericitization and silicification locally representing the most intense facies of alteration.

Clastic sediments overlying the ores contain feldspars in all stages of alteration and detrital sulphides. These are interbedded with Middle Cambrian limestones, providing the most convincing evidence to date of the age of the hydrothermal alteration and sulphide deposition.
1. INTRODUCTION

1.1 NATURE AND SCOPE OF THE STUDY

This thesis deals with the geology and genesis of the ores of an area of about one square mile at the northern section of the Mt. Lyell copper field, Western Tasmania (Fig. 1) encompassing the recently developed Cape Horn mine and the old Lyell Comstock and Tasman and Crown Lyell Extended workings.

The purpose of the study was to establish a stratigraphic succession in the volcanic host rocks for the mineralization and to relate this to the nature and genesis of the ore bodies. This has involved a study of the mineragraphy of the ores as well as hydrothermal alteration and chemical composition of the host rocks.

The investigation has also involved review, and sometimes reappraisal, of the descriptions and hypotheses of earlier workers, particularly those concerned with the genesis of the ores.

1.2 PROCEDURE

Geological information was recorded on 1:2400 base maps of the Mount Lyell Mining and Railway Company, occasionally supplemented with 1:1200 maps and aerial photographs. Additional information was gained from diamond drill core logging.

Some twelve weeks were spent in the field, time being approximately equally divided between field mapping and core logging.
Fig. 1  Locality Map.
1.3 LOCATION AND ACCESS

Cape Horn is located approximately 4 kilometres NNE of Queenstown, which lies on the Lyell Highway some 250 kilometres from Hobart. The area studied has excellent access with a road around the perimeter and roads to diamond drilling sites, particularly in the eastern section.

Outcrop in the area is generally sparse except in road cuttings.

1.4 PREVIOUS LITERATURE

Reports written prior to 1958 have been summarized in a paper by Wade (1958).

Early work on the geology and ores is remarkable in its diversification of ideas. Thus Thureau (1886) regarded the schists as volcanic muds, Peters (1893) saw the Iron Blow deposit as an upturned "pyritic swamp" and Allan (1893) linked mineralization with the emplacement of porphyries.

Gregory's (1905) book was the first comprehensive report of the area and by careful petrographical studies he recognized the schists as altered volcanic rocks. He also made an important contribution to the understanding of the structure of the area. Structural studies were extended by Loftus Hills (1927) who supported Gregory's interpretation of the mineralized rocks as lavas and tuffs.

Nye et al. (1934) initiated a new phase of geological thought for the deposits when they regarded the host rocks as intrusive porphyries, largely Devonian in age. This view was supported by Edwards (1939), who made the first detailed mineralogical survey of
the sulphides and concluded that the ores were of "mesothermal replacement" type. Conolly (1947) extended this viewpoint and his work initiated much fruitless exploration of Middle Owen Conglomerate horizons for ore.

Carey (1953) reinstated the concepts of earlier workers when he concluded that the Lyell schists were metamorphosed Cambrian geosynclinal sediments and volcanics. Carey (op. cit.) regarded the mineralization as being controlled by the strong zone of faulting along the Owen Conglomerate contact - the "Lyell Shear". Bradley (1954, 1956, 1957) supported the controlling influence of the Lyell Shear but regarded regional metasomatism and granitization as the critical features of ore genesis. This viewpoint was rejected by Wade and Solomon (1958) who emphasized the controlling effect of intersecting N-S and E-W Tabberabberan faults on ore distribution.

The possibility that some of the mineralization was Cambrian in age was an idea introduced by Hall and Solomon (1962), supported by Campana and King (1963) and elaborated on by Solomon (1964) and Solomon and Elms (1965). However these workers still regarded the Devonian phase as being predominant. The recognition of the importance of Cambrian processes in mineralization grew with the publication of Solomon's (1967) paper in which extremely haematitic zones in the Owen Conglomerate overlying ore bodies were regarded as fossil gossans. Trace element studies (Loftus Hills, 1967; Loftus Hills and Solomon, 1967; Loftus Hills, 1968; Loftus Hills et al., 1969) produced results more compatible with a Cambrian rather than a Devonian age for the ores, as did sulphur isotope studies (Solomon et al., 1969). Detailed mapping of the West Lyell Open Cut (McDonald, 1968) combined with
diamond drill core logging emphasized the stratigraphic nature of the mineralization, a pre-deformation origin for which was demonstrated by the mineralographic and petrological investigation of Markham (1968).

1.5 NOMENCLATURE

Following established local practice, the writer has occasionally adopted the colloquialisms of the field. It should be noted that throughout the text of this thesis the terms "Comstock" and "Lyell Comstock" are synonymous, the same applying to the terms "Tasman Shaft" and "Tasman and Crown Lyell Extended workings".

1.6 ACKNOWLEDGEMENTS

The writer expresses his gratitude to the staff and fellow graduate students of the Geology Department, University of Tasmania for stimulating and helpful discussion. Dr. M. Solomon, whose guidance, supervision and encouragement throughout the year were invaluable and Mr. R. J. Ford, for instruction on analytical techniques, are especially thanked.

The writer is indebted to the Mt. Lyell Mining and Railway Company Ltd. and to Mr. G. F. Hudspeth (General Manager) for providing accommodation in Queenstown.

The enthusiasm and interest of Mr. K. O. Reid (Chief Geologist) and other members of the geological staff at Mt. Lyell are gratefully acknowledged. The Tasmanian Department of Mines is also thanked for providing transport to the field area.

For her cheerful assistance throughout the project, I gratefully thank my wife Lee.
2. **STRATIGRAPHY**

From mapping in the Cape Horn-Lyell Comstock area the following succession has been established (Figs. 2 and 3):-

<table>
<thead>
<tr>
<th>Maximum thickness (m.)</th>
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<tbody>
<tr>
<td>PLEISTOCENE</td>
</tr>
<tr>
<td>Glacial Moraine</td>
</tr>
<tr>
<td><strong>ORDOVICIAN</strong></td>
</tr>
<tr>
<td>Gordon Limestone</td>
</tr>
<tr>
<td>Owen Conglomerate</td>
</tr>
<tr>
<td>Jukes Conglomerate</td>
</tr>
<tr>
<td><strong>CAMBRIAN</strong></td>
</tr>
<tr>
<td>Mt. Read Volcanics</td>
</tr>
<tr>
<td>Quartz Keratophyric Lapilli Tuff</td>
</tr>
<tr>
<td>Greywacke Conglomerate and Limestone</td>
</tr>
<tr>
<td>Chert (Stratified portion)</td>
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<tr>
<td>Keratophyric and Quartz Keratophyric Lavas</td>
</tr>
<tr>
<td>Lithic Tuffs</td>
</tr>
<tr>
<td>Felsitic Ignimbrites (?)</td>
</tr>
<tr>
<td>Ignimbrites (?) and minor siltstones</td>
</tr>
</tbody>
</table>

2.1 **CAMBRIAN SYSTEM**

**Mt. Read Volcanics**

1. Ignimbrites (?) and Siltstones

   This unit consists of volcanics and clastic sediments, the upper zone of which has been strongly hydrothermally altered and forms the host rock for the bulk of the Cape Horn mineralization. The series crops out distinctively west of Cape Horn as a deeply weathered reddish brown to white series of rocks.
Fig. 2: Geology of the Mt. Lyell Area (From Solomon, 1969).
Fig. 2: CAPE HORN-COMSTOCK AREA STRATIGRAPHY

QUATERNARY
- Pleistocene Moraine and Alluvium

ORDOVICIAN
- Upper Owen Conglomerate
- Middle Owen Conglomerate
- Lower Owen Conglomerate

CAMBRIAN
- Quartz Keratophyric Lapilli Tuff
- Greywacke Conglomerate, Limestone and Siltstone
- Chert and Massive Sulphides
- Keratophyric and Quartz Keratophyric Lava
- Intermediate Lithic Tuffs
- Felsitic Ignimbrites (?)
- Ignimbrites (?) and minor Siltstones

Geological Boundary
Inferred Geological Boundary
Bedding Strike and Dip
Cleavage Strike and Dip
Fault

Heights relative to sea level.
The volcanics show a streaky, fragmental texture in hand specimen (e.g. 37873) with some specimens appearing to have a very high proportion of clastic fragments (37874), this feature together with penetrations of silt into overlying volcanic layers somewhat suggestive of load casts, indicates that the "volcanics" may be epipyroclastics or clastically reworked tuffs.

In thin section (37873) the volcanics consist of albite crystals, often rounded, embayed and rounded quartz, tabular aggregates of chlorite showing an internal decussate texture, probably hornblende pseudomorphs, and pumiceous to hypocrystalline volcanic fragments, now completely devitrified in a quartz-sericite-chlorite-carbonate groundmass. Rounding of volcanic material is regarded by Fisher (1960) as one of the features common in laharic breccias, or volcanic mud flows, although rounding of feldspars is common in ignimbrites (Ross and Smith, 1960) and also occurs locally in lavas higher in the succession.

Vitric fragments are generally replaced either by carbonate and chlorite or by sericite. Many fragments are elongated, have irregular flame-like terminations and are bent around the phenocryst minerals (37873; Plates 1 and 2). The presence of deformed "fiamme" (Ross and Smith, op. cit.), together with the streaky lamination apparent in hand specimen, constitutes the strongest evidence in favour of an ignimbritic origin for the rocks, although some workers (e.g. Shirinian, 1966) note that flame textures may be developed in tufflavas.

The siltstone lenses and beds are commonly 0.3 to 3 metres thick and may be traced for some 100 metres along strike.
Plate 1 Bent "fiamme" in ignimbrite (?); vitric material replaced by sericite and quartz. Groundmass consists largely of carbonate, quartz, chlorite and albite.
37873 x 40 crossed nicols

Plate 2 Frayed termination to "fiamme".
37833 x 40 crossed nicols
The siltstones are finely laminated, often cross-bedded (37875a, b), with one specimen (37876) showing an elongated ellipsoidal quartz chlorite fragment in banded sericite quartz rock, the fragment probably being an altered mud pellet.

In thin section the laminated siltstones contain fresh to altered albite grains and quartz grains (37876) in a fine-grained sericite-quartz-carbonate-chlorite matrix, the mineralogy being largely a function of the hydrothermal alteration.

Considering the above evidence, it is probable that the siltstones were deposited in shallow ephemeral ponds following extrusion of each ash-flow tuff bed. The degree of clastic reworking of the ignimbrites is at this stage problematical.

2. Felsitic Ignimbrites (?)

The rocks of this unit are strongly porphyritic with abundant phenocrysts of pink albite (average diameter 2 mm.) and quartz in a felsic groundmass, sometimes pale green and very well laminated (37877) or grey and poorly layered (37878).

The groundmass (37877) consists of microcrystalline quartz, sericite, minor chlorite and potash feldspar (?), with tabular albite and rarely, embayed quartz phenocrysts. The lack of feldspar microcrystals is suggestive of an original vitric groundmass. Commonly the albite phenocrysts are fractured and bent (Plate 3), especially where the crystal elongation is at a high angle to the foliation, which also tends to wrap around the phenocrysts (Plate 4). The presence of unlaminated specimens (37878) is a strong argument against the foliation being tectonic; such rocks possibly being from the unwelded
Plate 3  Fractured and bent albite phenocryst elongated perpendicular to the foliation in felsitic ignimbrite (?). Groundmass has been replaced by quartz and sericite.

37877  x 40  crossed nicols

Plate 4  Slight warping of groundmass foliation around albite phenocrysts.

37877  x 40  crossed nicols
zone at the top of the flow. Similar rocks appear in the Queen Lyell area at the same stratigraphic horizon as sericite quartz schists, often pyritic.

Chemically the rock has a high silica content indicating it is an altered rhyolite (Table 1).

3. Intermediate Lithic (?) Tuffs

This horizon consists of fragmental and irregularly banded lithic tuffs with fragments and patches of a porphyritic material consisting of albite phenocrysts in a chloritic groundmass within a felsitic groundmass, also containing albite phenocrysts. As can be seen from Fig. 3 the unit crops out widely in the area.

Paucity of outcrop in the relatively unaltered zone near the East Queen River prevents rigorous subdivision of the unit, but from drill hole information in the Comstock area (e.g. drill hole NL 968) and surface mapping in the strongly altered schists, it is apparent that the tuff is coarsely fragmental at the base, grading up into finer tuffs or lavas which are mappable as chlorite flecked sericite-quartz schists.

In thin section (37879, 37890) the rock consists of euhedral phenocrysts of albite (An$_6$-7) up to 1.5 mm. in diameter with combined Carlsbad-albite and pericline twinning and rare embayed quartz phenocrysts. The albite phenocrysts tend to occur in clusters, but glomeroporphyritic texture is not developed. The groundmass consists of fine-grained recrystallized quartz, albite, potash feldspar, chlorite and minor rutile, with minor sericite, partially replacing feldspar. Sericite also is present as stringers with minor associated pyrite.
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<th>Specimen No.</th>
<th>Locality</th>
<th>Description</th>
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<tr>
<td>1</td>
<td>37877</td>
<td>6476W, 6858N Banded porphyritic felsite ignimbrite (?)</td>
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<tr>
<td>2</td>
<td>37891</td>
<td>3770W, 8220N Quartz keratophyric lava.</td>
</tr>
<tr>
<td>3</td>
<td>37892</td>
<td>4500W, 8150N Spilite of calcalkaline affinities. Same unit as 2.</td>
</tr>
<tr>
<td>4</td>
<td>37903</td>
<td>D.H. C50 656' Quartz keratophyric lapilli tuff.</td>
</tr>
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</table>
4. Keratophyric Lavas

A sequence of relatively unaltered porphyritic keratophyric lavas crop out along the Lyell Comstock road between 730 and 1400 metres west of the Lyell Comstock open cut. They thin rapidly towards the east probably lensing out near the Comstock chert body.

In thin section the rocks display all the typical textural features of calcalkaline lavas of the andesite-dacite-rhyolite association. Albite, quartz, chlorite, haematite and minor apatite are phenocryst minerals with albite, quartz, chlorite and K-feldspar (37891; identified by staining with sodium cobaltinitrite), forming the groundmass. Accessory rutile, zircon and monazite are also present. Late stage veinlets of carbonate, epidote and sericite, sometimes with traces of pyrite (37892) are rarely present.

Quartz occurs as strongly rounded and embayed phenocrysts, up to 2 cm. in diameter (37891-37893; Plate 5) in which a bipyramidal form is occasionally preserved. Albite (ca. An₄) occurs as glomerocrysts up to 1 cm. in diameter, swallowtail terminations and rounding of phenocrysts indicating partial resorption by the magma. Combined Carlsbad-albite and pericline twinning, forming a "mesh", is common although the former type in many cases occurs alone. Concentric alternating layers of clear albite and albite with minute inclusions of dusty opaques and chlorite (37893; Plate 6) may outline original oscillatory zoning.

Ferromagnesian phenocrysts are invariably altered to slightly pleochroic chlorite, often showing anomalous reddish-brown birefringence, and haematite. Replacement by iron oxides is only complete in the case of the coarser phenocrysts (up to 2 cm. long) which are invariably
Plate 5  Skeletal and embayed quartz phenocryst (right) and
euhedral poikilitic biotite (?) phenocryst, the
latter replaced by fine grained haematite (left).

37892 x 63 parallel nicols

Plate 6  Hornblende (?) phenocryst replaced by chlorite and
a haematite rim (upper left). A euhedral albite
crystal showing relict oscillatory zoning (?) is
at lower right. Both crystals in a hyalopilitic
groundmass with albite laths in a felsic feldspar-
quartz base.

37893 x 95 crossed nicols
poikiloblastic with albite inclusions (Plate 5). Smaller phenocrysts (Plate 6), up to 1 mm. long, normally have a chloritic core with a haematite rim while the finest phenocrysts contain no iron oxides (37892; Plate 7). Skeletal iron oxides are also present. The correlation between alteration and grainsize suggests that this phenomenon occurred prior to quenching of the groundmass and is probably due to resorption of biotite and hornblende due to their restricted stability field at low pressures and high temperatures as suggested by Green and Ringwood (1968).

Apatite is present both as hexagonal prismatic crystals up to 0.5 mm. long (Plate 7) and as minute needle-like crystals.

The groundmass has a hyalopilitic texture (Williams et al., 1954) and consists of minute laths of albite in a groundmass consisting largely of devitrified glass now represented by decussate chlorite, quartz and potash feldspar. From the base upwards there is a gradual transition from a predominantly chloritic to a felsitic groundmass, the latter having a pinkish colouration, probably due to submicroscopic haematite in quartz. This trend is also marked by a decrease in the number of albite laths in the groundmass and a pronounced increase in the acidity of the lava (Table 1).

5. Comstock Chert

The chert body crops out immediately west of the Lyell Comstock ore zone. From diamond drilling it is clear that locally the chert overlies a chlorite flecked schist of Unit 3 (Drill Hole 50) and that the chert is overlain by sediments containing detrital chert has been established both from field mapping and diamond drill
Plate 7  Euhedral apatite (left) and corroded ferromagnesian (right) phenocrysts; the latter is replaced by chlorite. The groundmass has a hyalopilitic texture as in Plate 6, but is more mafic and strongly chloritic.

37892  x 95  parallel nicols

Plate 8  Corroded barite grains in chert, with possibly two generations of barite. Note the relatively coarsely recrystallized chert adjacent to the barite.

37894  x 63  crossed nicols
In surfce outcrop the chert body is cut by innumerable haematite veins which, within the main mass of the chert, show a stockwork structure (Solomon, 1964; Plate 47, No.1) and are tightly crumpled (Solomon, 1964, 1967), but in thin section the haematite exhibits colloform textures (31088, 31093) identical to those of sulphides found in drill core material and thus a significant proportion of the haematite is probably derived from oxidation of sulphides, possibly via goethite.

In thin section the rock consists of a microcrystalline aggregate of quartz grains (averaging 0.02 mm. in diameter) with patches and veins of more coarsely recrystallized chert (37894, 31094) often adjacent to stylolites which are frequently outlined by irregular haematite veinlets. These often have a form suggesting the haematite formed as a film over fine euhedral quartz crystals indicating crystallization in an open cavity (31094). Barite occurs as patches and veinlets with corroded anhedral outline surrounded by patches of coarsely recrystallized chert (Plate 8). These features are possibly due in part to late stage "pressure solution" phenomena related to pore formation and filling by crystallization of the chert from a gel. It is likely that hydrothermal remobilization of barite occurred at this stage. Fluorite is occasionally present as an accessory mineral in late stage quartz-fluorite-sulphide veins in the main mass of the chert (ML 124; Plate 9). Quartz pressure shadows are common, adjacent to the quartz-sulphide intergrowths. These veins represent the fracture filling-replacement type ores of the chert.
Plate 9  Irregular mutual boundary texture between pyrite (black) and fluorite (dark grey) in coarsely recrystallized chert (white).

ML 124  x 63  parallel nicols

Plate 10  Fine, banded sulphides in exhalative (?) chert. Occasional pyrite grains are recrystallized.

37895  x 63  parallel nicols
breccia as discussed by Markham (1968), whose suggestion that the chert body represents complete silicification of volcanics in a fracture zone is tentatively accepted by the writer at this stage. At least the high degree of alteration is consistent with that in the surrounding schists as shall be discussed in a later section.

However some of the chert overlying the possible "replacement" chert previously described, as well as rocks of Unit 3, contains fine bands of sulphides with relict colloform and framboidal textures (37895, 101905) probably formed under open cast conditions. Some of these rocks (37895; Plate 10) contain minor fluorite and clear detrital quartz grains in a cream-coloured chert matrix. It is likely that the massive sulphide lenses of the Comstock open cut and Tasman Shaft occur at the same horizon.

6. Greywacke Conglomerate and Limestones

This unit consists of a complex of clastic and chemical sediments largely defying internal subdivision. The sediments are extremely important however, because they:

(i) provide information on the age of the Mt. Read Volcanics, and (ii) provide further evidence in favour of a Cambrian age for the mineralization.

In general the sediments include limestones, chert conglomerates and banded grey siltstones (37896A) west of Lyell Comstock (e.g. Drill Holes C50 and C58) and greywackes and conglomerates, minor limestones and opaline cherts dominate the succession south of the open cut.
The limestones (37896b, 91160) are pink to green in colour and are often brecciated and veined with calcite. From thin section examination they may be classed as bio-calcisparites containing sponge spicules and trilobite fragments. The fauna includes Dorapyge (Dr. P. Quilty, pers. comm.) indicating an Upper (?) Middle Cambrian age. Of particular interest is the fact that the limestone which directly overlies chert containing sulphide bands in drill hole C50 shows no sign of replacement by sulphide. As may be deduced from a subsequent section on hydrothermal alteration, this, in itself, is good evidence that sulphide deposition preceded limestone deposition.

A chert conglomerate which locally overlies the Comstock chert body on its north-western margin (37897) consists of subrounded pebbles of chert and quartz in a chlorite matrix. Sulphides are present in the matrix although some chert pebbles contain fine pyrite bands (Plate 11). The rock is veined sporadically by barite.

The major sediment type south of the Lyell Comstock open cut are conglomerates and greywackes. An important type contains volcanic rock and mineral fragments including pumice and shard fragments, albite, quartz, chlorite, jasper, chert and associated sulphide fragments (37898, 37899, 37900) in a matrix which may be either chert with diagenetic carbonate or chlorite sometimes with diagenetic (?) muscovite (38901). The sediments are poorly sorted, the angular grains generally being up to 1 cm. in diameter, with medium to low sphericity. Of particular significance is that within one section all stages of the alteration of plagioclase may be seen, from unaltered albite crystals to albite partially or completely altered to carbonate or sericite or, in extreme cases, by tabular aggregates of irregular
Plate 11  Portion of chert fragment with fine pyrite bands in chert conglomerate. Matrix (dark) consists largely of decussate chlorite.
37895 x 40 crossed nicols

Plate 12  Subhedral carbonate pseudomorph of plagioclase (white) and albite and quartz grains in sericitic matrix of greywacke breccia.
37899 x 63 crossed nicols
quartz grains occasionally with minor carbonate or chlorite (Plate 12). The significance of these mineralogical variations will be discussed more fully in a later section, but such an assemblage shows extreme disequilibrium with respect to hydrothermal alteration and could not have been produced by subsequent attack on a sediment containing fresh feldspars.

The pyrite grains (37898; Plate 13) are angular and are almost invariably surrounded by pressure shadows of quartz, carbonate and chlorite. Another interesting feature is the irregular "syneresis" (?) cracks confined to chert fragments.

7. Banded Quartz Keratophyric Lapilli Tuffs

The tuffs crop out prominently along the Lyell Comstock road immediately west of the open cut. The tuffs characteristically display planar bedding, the beds being between 10 and 30 cm. thick and consisting of coarsely fragmental albite, chlorite and quartz in a groundmass of similar composition.

The even bedding suggests the unit represents a submarine tuff flow (Fisher, 1960). The basal 10 metres of the tuff unit commonly contain fragments and lenses of the underlying sediments.

In thin section the tuff contains angular quartz phenocrysts, sometimes deeply embayed and often fractured (38902, 38903; Plate 14) slightly sericitized albite, chlorite, minor rutile and iron oxides and apatite needles in a microcrystalline groundmass of quartz, potash feldspar and secondary sericite and chlorite. In contrast to the other rocks, epidote is often abundant, in one specimen (38904) constituting some 20% of the rock.
Plate 13  Greywacke–chert breccias with clastic pyrite (high reflectivity), chert and volcanic grains in mudstone matrix.
37900 (upper)       37898 (lower)
x 2/3

Plate 14  Fine epidote crystals poikilitically enclosed in chlorite (dark) and fractured quartz phenocrysts in siliceous groundmass of quartz keratophyric lapilli tuff.
37902       x 63       crossed nicols
It is probable that the tuff is a product of a late stage violent phase of the volcanism.

The tuffs also are found at depth (38905; Plate 15) along the Jukes Conglomerate contact to the east of the Comstock open cut and extend south into the Queen Lyell area. From previous descriptions, there is little doubt that the tuff extends westwards, cropping out on the southern slopes of Crown Hill (Solomon, 1960) and is also the "quartz-feldspar porphyry" referred to by Edwards (1939) who mentioned the siltstone fragments in the basal section, although the latter author regarded them as roof pendants in a Devonian porphyry.

Chemically the tuff is of dacite composition (Table 1) and is notable in having a high CaO content (comparable with dacites quoted by Turner and Verhoogen, 1960, p.285) together with the absence of carbonate. This factor, together with the stratigraphic position of the tuff, suggests that eruption post-dated the alteration of the underlying volcanic suite. Also, the mineralogy of the tuff is identical with that of Smith's (1969) lowest grade of burial metamorphism.

It is also significant that in the case of the tuff abutting against the Jukes Conglomerate contact in the Lyell Comstock-Queen Lyell area (39805) there is no further sericitization of albite and no introduction of sulphide compared with the other specimens. This is consistent with a Cambrian age for the alteration and mineralization, but the tuff in this zone would have been expected to be strongly altered had the main phase of the mineralization been in the
Plate 15  Zoned, unsericitized plagioclase phenocryst in quartz keratophyric lapilli tuff. Core is oligoclase-albite (An$_{11}$), rim is albite (An$_{4}$). 37905 x 63 crossed nicols

Plate 16  Embayed and fractured quartz phenocrysts in volcanic fragments of Jukes Conglomerate. Texture is typical of quartz keratophyric lapilli tuff. 37906 x 63 crossed nicols
Devonian, with the Tabberabberan fracture zones providing channelways for ascending hydrothermal solutions.

2.2 ORDOVICIAN SYSTEM

Jukes Conglomerate

The Jukes Conglomerate overlies the banded tuffs in the Comstock area and consists of coarse rounded maroon tuff fragments, normally greater than 6 cm. in diameter in a greywacke matrix of essentially the same composition. Minor greenish siltstone bands also occur. Information on the formation locally is sketchy and is gained entirely from drill core, but most significantly the conglomerate is normally absent in the Cape Horn area and thickens towards the north being approximately 11 metres thick half-way between Cape Horn and Lyell Comstock (drill hole NL 962) and is at least 70 metres thick in the Comstock Valley (drill hole C50).

In thin section (38906; Plate 16) the fragments are recognizable as being derived from the underlying tuff, particularly in the characteristic angular, fractured and embayed nature of the quartz. Mineralogically the fragments show strong signs of surface weathering, the chlorite and feldspar having been altered to sericite and quartz with formation of "limonite" and iron oxides. Rutile appears to have been largely replaced by leucoxene. Thus there is little doubt that in the area the Jukes Conglomerate represents a "talus breccia" (Solomon and Elms, 1965) formed by "exfoliation and disintegration of the underlying volcanic and pyroclastic material with little or no transport of material" (Banks, 1962).
Banded greenish grey siltstones within the Jukes Conglomerate become more prominent south of Lyell Comstock, where the formation directly overlies mineralized volcanics. The siltstones in turn are overlain by haematite conglomerate similar to that previously described and are assigned to the Jukes Conglomerate because of thin (up to 0.5 cm.) layers of haematitic sandstone, similar to the conglomerate, within the silts (38907). In thin section the silts, contain quartz, sericite, carbonate, chlorite and minor fine pyrite grains in a quartz-sericite matrix. The sericite "grains" consist of a rectangular mesh of fine sericite flakes with a tabular form and are obviously feldspar pseudomorphs. It is thus evident that the siltstone, which was probably deposited in a reducing pond environment, is interbedded with "normal" Jukes Conglomerate and contains material derived from mineralized volcanics belonging to a horizon underlying the quartz-keratophyric lapilli tuff. This suggests a possible unconformity above the volcanics. This interpretation is supported by the thinning of the Jukes Conglomerate between Comstock and Cape Horn.

**Owen Conglomerate**

The Owen Conglomerate was not studied during the present investigation but has been well described by previous writers (e.g. Wade and Solomon, 1958; Solomon, 1964, 1967, 1969; Solomon and Elms, 1965; Banks, 1962). The Owen Conglomerate has been subdivided into Lower, Middle and Upper members marked by the upward passage from coarse-grained, pale coloured, quartzose paraconglomerates to haematitic sandstones and chert conglomerates with a pale grey
sandstone at the top of the succession. The maximum thickness of the formation in the Queenstown area is approximately 730 metres.

Of particular interest is the presence of haematite bodies and chert-haematite breccias within Middle Owen sediments overlying the three known occurrences of probable Cambrian surface-formed ore, namely in the Lyell Comstock, North Lyell and Iron Blow area. Solomon (1967) has interpreted these bodies as fossil gossans or limonitic screes formed during Ordovician weathering of the underlying sulphides.

**Gordon Limestone**

The presence of the Gordon Limestone in the Comstock Valley directly west of the mine has been confirmed by diamond drilling.

### 2.3 Pleistocene Moraine

Morainal sediments crop out in the Comstock Valley and sporadically on the hillside south of the valley. The presence of a rock flour matrix and dolerite fragments, probably derived from the Sedgewick Plateau to the north, make the glacial sediments easily distinguishable from the Owen Conglomerate screes in the area.
3. STRUCTURE

3.1 REGIONAL STRUCTURE

The major structure of the Mt. Lyell area is the sharp N-S trending upturn of the Owen Conglomerate-Mt. Read Volcanics contact, the Lyell Shear. The upturn marks the eastern limb of a major Devonian anticlinorium, but local disturbance dates from at least the upper Cambrian Jukesian Orogeny as indicated by rapid thinning of the Owen Conglomerate westwards from the mine area (Wade and Solomon, 1958).

Disturbance during Owen sedimentation is marked by the Haulage Unconformity within Upper Owen sandstones. Sediments underlying the unconformity are tightly crumpled, probably due to eastwards slumping of conglomerate and underlying volcanics resulting from continued uplift of the west wall of the Owen Basin (Solomon, 1964).

The most important phase of deformation occurred in the Middle Devonian Tabberabberan Orogeny. Solomon (1964) has classified the Tabberabberan structures into:-

1. Early N-S structures (TFL).
2. Late NW-NNW structures (TF2).

The TFL phase involved folding into major anticlinoria and synclinoria of which the major feature is the West Coast Range Anticlinorium, the eastern contact of which, as previously mentioned, is strongly overturned and thrust faulted in the mine area (Solomon, 1964) and which has been regarded by many authors (e.g. Carey, 1953) as a major controlling factor governing ore location.
**TF2 Structures.**

Solomon (op. cit.) classified these later structures into the following:

1. Faults with dip slip and strike slip components and the Linda Fault Zone.
2. Folds.
3. Cleavage related to the folds, with a steeply plunging lineation, T1_1.
4. Shallowly pitching cleavage lineation, T1_2.

### 3.2 STRUCTURE OF THE CAPE HORN AREA

Solomon showed that the NW trending cleavage was related to folding which tended to a similar style in the Owen Conglomerate. In the Cape Horn-Lyell Comstock area both cleavage and bedding of the volcanics have a similar E-W to NW-SE strike to within approximately 400 metres of the conglomerate-schist contact. This pseudo-parallelism of cleavage and bedding is marked by:

(i) Bedding in the quartz keratophyric lapilli tuffs and underlying sediments is broadly concordant with cleavage in the more altered lithic tuffs cropping out on the ridge to the north.

(ii) Cleavage generally closely parallels bedding near the East Queen River as shown by the thin siltstone beds and lenses within the volcanics of Unit 1.

Although there is little doubt that the cleavage is a transposition structure, in the sense of Whitten (1966), related to TF2
folding. However, fold closures, such as those noted near West Lyell by Solomon (1964, pp.304-306) are rare in the area, probably due to the general NW strike of the volcanics following the TF1 phase. A few fold closures have been noted within the schists near the Owen Conglomerate contact (e.g. at 3040W, 3700N) but these are probably the result of minor dextral faulting, related to a late stage of the TF2 deformation. This interpretation is supported by numerous kink bands in the vicinity. Tight folding also occurs within pyritic ores of the Tasman Shaft and the Comstock open cut.

Within 400 metres of the Owen Conglomerate contact the cleavage (and bedding) of the volcanics swings sharply from a NW trend to a N-S trend and, on occasions, approaches a SW azimuth as the conglomerate contact is approached, giving rise to a tight north-east trending, steeply plunging anticline, particularly prominent in the Lyell Comstock area. Strong shearing undoubtedly occurred along the contact as shown by the large number of tensional quartz veins in the vicinity, which often extend into the Owen Conglomerate.

The Cape Horn-Lyell Comstock area is bounded to the south by the E-W trending North Lyell Fault (Figs. 2 and 3), a structure forming part of a large scale structure which is part of a major E-W disturbance, the Linda Fault Zone (Solomon, 1962, 1965). Tabberabberan movement on the North Lyell fault involved some 750 metres of sinistral transcurrent displacement with several hundred metres of vertical movement, south side down (Solomon, 1964; Solomon and Elms, 1965). It is clear that the swing in the cleavage in the schists from a NW to a N-S trend may be entirely due to sinistral movement on the North Lyell fault provided that the bulk of transcurrent movement occurred
subsequent to cleavage development in the volcanics. This interpretation is in accord with Solomon's (1964) view that the sharp swing in the Lyell Shear to a NE trend between Cape Horn and Lyell Comstock is also related to the transcurrent movement.

Little is known about the Comstock Fault but mapping by the writer indicates that it should be placed farther north than by Wade and Solomon (1958) as shown in Fig. 4.

3.3 PRE-DEVONIAN MOVEMENT ON THE NORTH LYELL FAULT

From the geological map (Fig. 3) it can be seen that there is apparent unconformity between the Owen Conglomerate and the altered volcanics with rocks progressively lower in the Cambrian sequence abutting against the conglomerate contact between Comstock and Cape Horn (Figs. 5 and 6). The banded pyritic ore at the Iron Blow may be tentatively correlated with the Tasman Shaft mineralization and is hence probably towards the top of the volcanics.

This pattern is consistent with tilting of the block between Cape Horn and Lyell Comstock south side up with subsequent erosion prior to deposition of the Owen Conglomerate. A pre-Tabberabberan phase of movement on the North Lyell Fault is thus tentatively proposed.
N-S Section East of Comstock Ore Zone (500'W).

Pleistocene Moraine
Ordovician Gordon Limestone
Owen Conglomerate
Jukes Conglomerate
Cambrian Quartz Keratophyric Lapilli Tuff

Grey Shale
Limestone
Banded Pyritic Chert
Lithic Tuff
Fault
NW–SE Section South of Comstock.

Moraine
Owen Conglomerate
Jukes Conglomerate
Quartz Keratophyric Lapilli Tuff

Graywacke Conglomerate
Chert Conglomerate
Carbonate-chlorite-sericite-quartz schist
Pyritic quartz-sericite schist
Tuff.

Fig. 5
Section Through Cape Horn Mine Area Bearing 100° M.

Fig. 6
4. HYDROTHERMAL ALTERATION AND THE LYELL SCHISTS

Hydrothermal alteration has been recognized as an important control over ore deposition by proponents of a Devonian age for the ores (e.g. Edwards, 1939; Carey, 1953; Wade and Solomon, 1958) and by those favouring a Cambrian age for at least a major phase of the mineralization (e.g. Solomon, 1964; Rafter and Solomon, 1967; Markham, 1968). The former authors, however, believed that the role of alteration is subordinate with both the zonation of alteration and ore formation being structurally controlled. These writers believed that alteration post-dated cleavage development, whereas recent workers (Rafter and Solomon, 1967; Loftus-Hills et al., 1967; Markham, 1968) regard the alteration as being pre-cleavage and ascribe the stronger cleavage in the more altered ore-zone rocks to the presence of a higher proportion of phyllosilicates and carbonates in these rocks prior to the Tabberabberan Orogeny.

In the Cape Horn-Lyell Comstock area a range in alteration assemblages from intensely silicified and sericitized rocks, through carbonate-bearing assemblages (the "marginal schists" of Wade and Solomon, 1958 and Solomon, 1964) to albitized volcanics exists and because of their relevance to ore formation the assemblages shall be discussed in some detail.

4.1 CHLORITES AND QUARTZ CHLORITE SCHISTS

Optically the chlorites of the ore zone differ from those of the outer zones of hydrothermal alteration. All chlorites display anomalous birefringence, although those of the ore zone show stronger
pleochroism (37908, 37909) with green-blue to purple anomalous bi-refringence colours whereas those from the outer zones (37873, 37892) typically display green, reddish brown or maroon colours, or occasionally the "normal" first order grey. All the chlorites analysed from the Lyell area (Solomon, 1964) contain sufficient iron to be classed as ripidolites although those richest in iron occur in the more altered zone and the optical properties previously described are consistent with this trend (Deer et al., 1962, v.3, p.152).

Quartz chlorite schists occur only within zones of sulphide mineralization having been described from West Lyell as occurring in thin "beds" up to 6 metres thick within more sericitic schists (Solomon, 1964, p.245) and similarly at, and north of, Cape Horn quartz-chlorite "beds" are normally less than 2 metres thick, often interbedded with fragmental sericite-chlorite-quartz schists and quartz sericite schists. Quartz chlorite schists are also present at Lyell Comstock ore zone near the hanging wall of the chert body (37880).

Typically, haematite is present in abundance in these schists as clots and irregular veinlets intimately associated with chlorite. Towards the margin of a chlorite-quartz-haematite "bed" a zonation from haematite-chlorite intergrowths, frequently with a decussate chlorite core (37908; Plate 17) outwards through a haematite-quartz-chlorite assemblage to schist with large sericite blebs, was noted. The same pattern of alteration is commonly observed adjacent to haematite-chlorite clots in the lithic tuffs as shown by specimens 37881 and 37882. The haematite (37882; Plate 18) commonly forms a triangular network of intersecting lamellae, a form which indicates replacement of magnetite according to Uytenbogaardt (1951). This interpretation is
Plate 17  Zonation in haematitic "clot" in quartz chlorite schist. Light core of clot is decussate chlorite passing into a zone rich in haematite (black). Rim consists of fine grained quartz-chlorite-haematite.

37908  x 40  parallel nicols

Plate 18  Triangular network of haematite lamellae in chlorite (pale grey) and quartz (white).

37882  x 95  parallel nicols
supported by the occurrence of relict magnetite grains within sulphides adjacent to haematite clots near Cape Horn (37883, 37884; Plate 19) and the presence of magnetite adjacent to sulphides in the Lyell Comstock workings (101906; Plate 20).

Also prominent in the quartz chlorite schists are quartz porphyroblasts exhibiting regrowth, typified by specimen ML 97 (Plates 21 and 22) in which a clear quartz core is surrounded by a zone of radially regrown quartz at the inner margin of which there is a concentration of extremely fine dusty opaques, possibly haematite. The rim is characterized by a thin zone extremely rich in haematite lamellae. Pre-cleavage regrowth is indicated by both core and rim having similar strain extinction.

Some quartz chlorite schists (37885; Plate 23 and 24) contain abundant laths up to 6 mm. long of quartz aggregate in a fine decussate chlorite matrix. The laths are pseudomorphous after euhedral plagioclase, or after pre-cleavage sericite replacements of plagioclase. The quartz and chlorite are both unstrained and the former occasionally bears chlorite inclusions. Fine-grained haematite and subhedral to euhedral pyrite are also present. These features indicate post-deformation recrystallization; a process which probably also involved some "cleaning up" of the rock to the extent of removing some chlorite from the quartz "laths".

Occasional minute fluorite euhedra, up to 0.02 mm. in diameter, are enclosed within the quartz. Bothwell and Moss (1957) report 1.0% F in a sericite from Mt. Lyell. As fluorite is otherwise absent at Cape Horn and is extremely rare at West Lyell, the presence of fluorite possibly indicates that the quartz has replaced sericite.
Plate 19 Poikiloblastic chalcopyrite (faint grey) in quartz (dark grey). Haematite (white) occurs as concentrations of lamellae at upper margin of chalcopyrite grain. A few small equant magnetite (mid-grey) grains are present in chalcopyrite (near centre of plate).

37883 x 62 reflected light, 1 nicol

Plate 20 Magnetite (brown) partially replaced by haematite (light grey-brown) and associated chalcopyrite in quartz chlorite schist.

101906 x 350 reflected light, 1 nicol
Plate 21  Regrowth of quartz in quartz chlorite schist. Consists of inner clear primary quartz with a concentration of minute haematite inclusions at hub of regrowth zone.
ML 97 x 63 parallel nicols

Plate 22  Quartz grain of Plate 21 but taken under crossed polars. Note that the uneven extinction is partially due to radial regrowth, but has been modified by strain.
ML 97 x 63 crossed nicols
Plate 23  Elongate quartz aggregates pseudomorphing feldspar in fine grained chlorite matrix of quartz chlorite schist.
37885   x 63   parallel nicols

Plate 24  As for Plate 23, but with parallel polars, showing the microcrystalline nature of the quartz.
37885   x 63   crossed nicols
The above evidence suggests that the quartz-chlorite-haematite assemblage is secondary after sericitic assemblages containing magnetite, the magnetite probably being coeval with hydrothermal introduction of sulphides. A possible mechanism for the process is shown in the equation below, the chlorite formula being chosen to coincide with an analyzed chlorite from the Iron Blow (listed in Solomon, 1964, Table 21).

$$6Fe_3O_4 + 2KAl_2(AlSi_3O_{10})(OH)_2 + 3Mg^{2+} + 4H_2O + 4OH^- \rightarrow 6Fe_2O_3 + Fe_6Mg_3Al_3(Si_5Al_3)O_{20}(OH)_{16} + SiO_2 + 2K^+ + 4H^+ + 2H_2O$$

4.2 QUARTZ-SERICITE SCHISTS

Highly siliceous quartz-sericite schists are characteristic of the ore zones. These rocks have been well described from West Lyell by Solomon (1964) and Markham (1968) and shall only be treated briefly here. The schists are often strongly pyritic and consist of bands and patches of sericite (37886, 37887) in a groundmass of recrystallized quartz aggregate, the quartz grains frequently having lobate sutured boundaries as defined by Spry (1969, p.19), indicating some deformation subsequent to recrystallization (Plate 25). The quartz aggregate probably forms by further strain of quartz porphyroblasts similar to those discussed previously (Plates 21 and 22). The occurrence of discrete quartz and aluminosilicate-rich bands and patches is not a primary feature of the rocks and was not observed in the relatively unaltered volcanics cropping out near the East Queen River.
Plate 25  Strained quartz aggregate in quartz sericite schist, showing incipient formation of lobate sutured boundaries.

37886  x 660  crossed nicols

Plate 26  Embayed euhedral quartz phenocryst in quartz sericite schist. Minor secondary regrowth of quartz is present.

37888  x 170  crossed nicols
The banding is probably due to metamorphic differentiation in the sense of Spry (1969, pp.236-237).

The only mineral of primary origin in the quartz sericite schists is quartz, frequently showing the typical embayed form of quartz in the less altered volcanics (37888; Plate 26). The quartz shows secondary hydrothermal regrowth.

Pyrite porphyroblasts and chalcopyrite grains most commonly occur within phyllosilicate layers, a feature noted also by Markham (1968) which is possibly also due to metamorphic effects.

A pre-cleavage origin for the sulphides is indicated by the common quartz pressure shadows adjacent to pyrite porphyroblasts and chalcopyrite grains (Solomon, 1964; Markham, 1968) as shown in Plate 27.

It is tentatively suggested that the formation of the chlorite-free schists may be related to the origin of the quartz chlorite schists by a complementary process caused by local differences in the pH of the hydrothermal fluids. A possible equation for the process, derived in part from Meyer and Hemley (1967) is:

\[
\begin{align*}
\text{Fe}_6\text{Mg}_3\text{Al}_3(\text{Si}_5\text{Al}_3)\text{O}_{20}(\text{OH})_{16} + 3\text{Al}^{3+} + 4\text{SiO}_2 + 3\text{K}^+ + 6\text{H}^+ & \rightarrow \\
\text{chlorite} & \text{quartz} \\
3\text{KAl}_2\text{AlSi}_3\text{O}_{10}(\text{OH})_2 + 6\text{Fe}^{2+} + 3\text{Mg}^{2+} + 8\text{H}_2\text{O} & \text{sericite}
\end{align*}
\]

The writer suggests tentatively that the quartz chlorite and quartz sericite schist types need not imply different original rock types, but were formed by the metasomatic processes outlined.
Plate 27  Irregular pressure shadows of quartz adjacent to chalcopyrite grains.
37886 x 63 crossed nicols

Plate 28  Fractured apatite crystals associated with chlorite, quartz and sulphides. Portion of sericite band at lower left.
37909 x 63 nicols partially crossed
4.3 OTHER SCHIST TYPES AT CAPE HORN

The schists described in the previous sections represent the most extreme cases of alteration at Cape Horn. The ore zone is marked by the complete absence of albite and by the presence of carbonate generally only in minor quantities; the mineralogy being quartz-sericite-carbonate-haematite-sulphides with minor apatite and occasional traces of zircon and monazite.

In general, the schists of this group bear recognizable textural similarity to the parent ignimbrites (?). Relict fragmental texture is marked by discontinuous bands or stringers of sericite and chlorite in a fine-grained quartz mosaic or a quartz-sericite intergrowth (37889, 37910). Embayed quartz phenocrysts are common (e.g. 37911). Especially towards the margins of the ore body carbonate is present, most commonly enclosed in the quartzose mosaic, but never intergrown with sericite, possibly suggesting disequilibrium between these two minerals (37909, 37911). Apatite is a common accessory; in one specimen (37909; Plate 28) fractured apatite euhedra up to 1 mm. in diameter locally comprise 50% of the rock in a thin band associated with carbonate, chlorite, quartz, sulphides and haematite.

The sediments are represented in the ore zone by sericite schists, consisting of knots and kernels of quartz aggregate and quartz grains in a strongly sheared sericite matrix (37912). The quartz grains together with minor sulphides present are often fractured and strung out along the cleavage.

Within approximately 150 metres of the hanging wall of the ore body (37876) slightly altered albite and albite replaced by a
quartz-carbonate intergrowth are present in the sericitic matrix of the sediments, a similar effect being noted in the ignimbrites. Traces, but not appreciable amounts, of sulphides occur in this zone stratigraphically below the ore horizon at Cape Horn.

The same does not hold for the volcanics overlying the ore horizon where sericitic alteration persists and pyrite is present in the schists although chalcopyrite is scarce. A distinctive sericite quartz schist with coarse chlorite and sericite flecks (37913), possibly a lava, occurs at the top of Unit 1. In thin section (ML 54) the sericite and chlorite "phenocrysts" are strongly elongated along the TFL lineation with elongations of up to 7 in contrast to 2 for albite in the unsheared volcanics. In association with this process pyrite euhedra are frequently fractured and strung out along the cleavage (Plate 29).

The overlying porphyritic felsites of Unit 2 are present as sericite-flecked, sericite-quartz schists in the Queen Lyell area (37914, 37915) and in most cases contain only traces of chlorite although exceptions occur (37916). Pyrite may account for up to 15% of the rock.

4.4 ALTERATION IN THE LITHIC TUFFS

From Figs. 3 and 10 it can be seen that the trend of the lithic tuffs is markedly discordant with the hydrothermal alteration zonation.

Their mineralogy in the albite fringe zone, notably quartz-albite-chlorite has previously been discussed. To the south and east of the Lyell Comstock open cut the mineral assemblage is quartz-chlorite-calcite-sericite with accessory pyrite (37917).
ZONATION OF HYDROTHERMAL ALTERATION ASSEMBLAGES

- Chert and Silicified Volcanics
- Sericite Zone
- Carbonate Zone
- Albite Zone
- Post Alteration Assemblages
- Diamond Drill Hole

Fig. 10
Plate 29 Sheared sericite (light) and chlorite (dark) lenses in sericite-quartz groundmass. Note the pyrite grain (upper right) which has been fractured and drawn out along the schistosity.
ML 5½ x 63 parallel nicols

Plate 30 Carbonate (pale grey) sericite (white) intergrowth in probable replaced plagioclase phenocryst of carbonate-zone schist.
37917 x 95 crossed nicols
Calcite, often with chlorite, and quartz pseudomorphs the plagioclase and is occasionally intergrown with sericite (Plate 30). Significantly, the margins of the calcite "grains" are frequently frayed, or have thin quartz-sericite pressure shadows.

In the Queen Lyell area the carbonates become rarer and probably more sideritic (37918) and are associated with chlorite or groundmass quartz. Feldspar phenocrysts are completely altered to a sericite-quartz intergrowth occasionally with minor siderite.

Some 400 metres north-east of Cape Horn the alteration has proceeded to the stage where carbonate is rare, occurring sporadically in the groundmass. The rocks still display the characteristic patchy fragmental texture of the tuffs (Plates 31 and 32) and quartz-sericite-chlorite mineralogy (37919) with common haematite-chlorite patches (37881, 37882) and associated narrow zones of quartz-chlorite alteration. Chalcopyrite appears in this zone, being absent in those previously described and is often associated with the haematite.

4.5 THE COMSTOCK AREA

The main mass of the chert at Lyell Comstock has been tentatively suggested as representing a zone of complete silicification of volcanics by Markham (1968) and the writer (Chapter 1). The main evidence previously mentioned, is:

(i) the structural discordance of the chert,
and (ii) the presence of tabular grains of chert, presumably feldspar pseudomorphs in the overlying conglomerates.
Plate 31  Lithic type in sericite zone (2955W, 4120N) showing characteristic patchy fragmental texture. Greenish patch near lens cap is a haematite-chlorite "clot". Diameter of lens cap is approximately 4.5 cm.

Plate 32  Patchy fragmental texture in lithic tuffs from three alteration zones. Left: carbonate zone (37917) Centre: siderite rich specimen (37918) Right: sericite zone (37919) Dark area at bottom of 37919 is a haematite-chlorite clot. x 1/3
In addition the occurrence of fluorite and barite in the chert is also consistent with this hypothesis. As has been stated (section 4.1) fluorite may be precipitated by reaction between $F^-$ released during alteration of sericite and $Ca^{2+}$ present in hydrothermal solution. Similarly sufficient $Ba^{2+}$ may be released during destruction of sericite to enable the formation of barite. Preliminary analyses show that concentrations of Ba in excess of 1% occur in sericitic schists at West Lyell (Dr. M. Solomon, pers. comm.). Similar Ba enrichment in mineralized volcanics has been reported from Rosebery by Gee (1970) who demonstrated by electron microprobe studies that the Ba resided in sericite.

Within the lithic tuffs "cherty" quartz sericite schists are commonly present in drill holes immediately south of, and stratigraphically below, the mineralized zone at Comstock and probably represented feeder channels for the mineralization. These schists (e.g. 37920) containing minor sulphides, and minute crystals of rutile showing the characteristic knee-shaped twinning, are present within quartz, indicating that the schist formed by alteration of pre-existing volcanics rather than by surface exhalative processes. The sericite is a pale green "batchelorite", a "mineral" common in the schists and also veins (37921) near the Comstock chert. "Batchelorite" also occurs at North Lyell and the Iron Blow and pyrophyllite has been identified in this material by Solomon (1964). The occurrence of pyrophyllite indicates the presence of solutions too acid to co-exist with sericite and is thus an intermediate phase in the alteration between complete silicification and the normal muscovite-bearing assemblages.
4.6 SYNTHESIS

In the area two foci of the alteration occur coinciding with the economic mineralization at Cape Horn and Lyell Comstock. The zonation of hydrothermal assemblages is due to variations in pH of the ore fluids as may be expressed most succinctly by a series of equations:

\[ 4\text{SiO}_2 + 2\text{Al}^{3+} + 4\text{H}_2\text{O} \rightleftharpoons \text{Al}_2\text{Si}_4\text{O}_{10}(\text{OH})_2 + 6\text{H}^+ \quad \text{(III)} \]

quartz pyrophyllite

\[ 3\text{Al}_2\text{Si}_4\text{O}_{10}(\text{OH})_2 + 2\text{K}^+ \rightleftharpoons 2\text{KAl}_3\text{Si}_3\text{O}_{10}(\text{OH})_2 + 6\text{SiO}_2 + 2\text{H}^+ \quad \text{(IV)} \]

pyrophyllite muscovite

(from Hemley and Jones, 1964)

\[ \text{KAl}_3\text{Si}_3\text{O}_{10}(\text{OH})_2 + 2\text{Ca}^{2+} + 2\text{HCO}_3^- + 4\text{H}_2\text{O} \rightleftharpoons 2\text{CaCO}_3 + 3\text{SiO}_2 + \text{calcite quartz} \]

muscovite

\[ 3\text{Al(OH)}_4^- + 4\text{H}^+ + \text{K}^+ \quad \text{(V)} \]

(from Meyer and Hemley, 1967)

\[ \text{NaAlSi}_3\text{O}_8 + \text{Ca}^{2+} + \text{HCO}_3^- + 3\text{H}^+ \rightleftharpoons 3\text{SiO}_2 + \text{CaCO}_3 + \text{Al}^{3+} \]

\[ + \text{Na}^+ + 2\text{H}_2\text{O} \quad \text{(VI)} \]

albite quartz calcite

The step involving formation of carbonate may be absent as the deposition of carbonate is strongly pressure controlled (Holland, 1965, 1967). In such a case feldspars are directly sericitized.

\[ \text{KAlSi}_3\text{O}_8 + 2\text{NaAlSi}_3\text{O}_8 + 2\text{H}^+ \rightleftharpoons \text{KAl}_3\text{Si}_3\text{O}_{10}(\text{OH})_2 + 6\text{SiO}_2 + 2\text{Na}^+ \quad \text{(VII)} \]

K-feldspar albite muscovite quartz

(from Meyer and Hemley, 1967)
or, where the parent rock is deficient in potash feldspar:

\[ 3NaAlSi_3O_8 + K^+ + 2H^+ \rightleftharpoons KAl_2Si_3O_10(OH)_2 + 6SiO_2 + 3Na^+ \quad \ldots \text{VIII} \]

albite muscovite quartz

(from Helgeson, 1967).

The use of pyrophyllite as a geothermometer in the Mt. Lyell area is of dubious validity. Hemley and Jones (1964), in a series of experiments in which aluminosilicate assemblages were buffered with quartz found that pyrophyllite formed only above 350°C. However, on textural grounds, notably in the microcrystalline nature of the cherts, the colloform textures in the associated sulphides and the syneresis (?) cracks in chert clasts in the overlying sediments, the writer considers that the chert probably formed as a hydrous gel. Recent studies by Fournier (1967) on systems with activities of silica in solution sufficiently high to be compatible with the occurrence of amorphous silica indicate that pyrophyllite may persist metastably at temperatures below 200°C, the precise temperature depending on water fugacity. The formation of the thermodynamically stable kaolinite at these temperatures is inhibited by the mechanism:

\[ Al_2(Si_2O_5)(OH)_4 + 2SiO_2 \cdot nH_2O \rightleftharpoons Al_2(Si_4O_{10})(OH)_2 + (2n+1)H_2O \quad \ldots \text{IX} \]

Other reactions involving hydrogen metasomatism may further explain the observed assemblages. Chlorite occasionally occurs with calcite (?) in feldspar pseudomorphs in the carbonate zone probably due to the process (Meyer and Hemley, 1967):

\[ 2NaAlSi_3O_8 + 4(Mg,Fe)^{2+} + 2(Fe,Al)^{3+} + 10H_2O \rightleftharpoons \]

albite

\[ (Mg,Fe)_4(Fe,Al)_2Si_2O_{10}(OH)_8 + 4SiO_2 + 2Na^+ + 12H^+ \quad \ldots \text{X} \]

chlorite quartz
A siderite-bearing zone near the contact between the sericite and relatively siderite-poor carbonate zones is present but is probably thin and cannot be adequately mapped because of lack of adequate drill hole information and also the occurrence of widespread minor siderite in the sericite zone. The siderite is possibly equivalent in part to the haematite in quartz chlorite schists, forming under conditions of higher carbon dioxide and lower oxygen fugacity. This is supported by the intimate association of siderite and chlorite in some specimens (e.g. 37918).

The association between sulphide mineralization and hydrothermal alteration assemblages shall be discussed in more detail after the nature of the mineralization has been described.
5. ORE DEPOSITS

5.1 CAPE HORN

The Cape Horn ore body occupies a narrow zone west of the Owen Conglomerate contact and is currently in the early stages of mine development. Current ore reserves (above 800 feet above sea level) are 4 million tons of 1.6% Cu, the sulphur content of the ore being relatively low at 4.5% FeS₂, compared with between 6 and 20% FeS₂ at West Lyell (latter figure from Solomon and Elms, 1965). In addition the Cape Horn ore contains 0.45 oz. Ag and 0.032 oz. Au per ton (Mr. K.O. Reid, pers. comm.).

The mineralogy of the Cape Horn ores, as established by this study, is relatively simple with pyrite and chalcopyrite dominating the assemblage. Haematite, molybdenite, tennantite, rutile, galena, sphalerite and gold are accessories, all except the former two being rare. Supergene alteration of the ores is minor and restricted to the upper levels of the mine; chalcocite and native copper have been identified in hand specimen and in one specimen covellite-digenite intergrowths replace chalcopyrite (37882).

The Cape Horn ores display deformational textures identical to those described recently from West Lyell by Markham (1968).

Generally chalcopyrite appears to have behaved plastically under stress, whereas pyrite is extensively fractured. Textures developed due to this phenomenon include the migration of chalcopyrite into pressure shadow zones around pyrite porphyroblasts (37910, 101907; Plate 33), into minute fractures in the gangue, especially in the quartz-rich schists (101909, 101910, 37888, 37889; Plate 34).
Plate 33  Deformation textures associated with microfault in Cape Horn ore. Note pressure shadows of chalcopyrite adjacent to pyrite grain; fracturing of pyrite grain and shearing of haematite lamellae.

101907 x 62 reflected light, parallel polars

Plate 34  Chalcopyrite stringers filling fractures in quartz gangue, Cape Horn ore.

101909 x 95 reflected light, parallel polars
and into fractures in pyrite porphyroblasts (101908). In many cases chalcopyrite acts as a host for recrystallized pyrite euhedra (101909; Plate 35).

Etching with H$_2$O$_2$/NH$_4$OH reveals the texture in the chalcopyrite of the disseminated ores. The chalcopyrite (101908, 37889) is fine to coarse grained, individual grains showing a mutual boundary texture with sub-grains being developed in some cases. The formation of sub-grains in chalcopyrite appears to be a process similar to that involving the recrystallization of quartz to produce an aggregate of quartz grains. The chalcopyrite also shows polysynthetic deformation twinning. These features are similar to those described from Broken Hill by Richards (1966) and indicate a pre-cleavage origin for the chalcopyrite.

Pyrite crystals frequently show evidence of cataclastic deformation, frequently being fractured with the fragments occasionally strung out along the cleavage to form pressure tails (37910, 101911).

Recrystallization of pyrite is common, the pyrite often having euhedral cubic (101912; Plate 36) or pyritohedral outline (101913). Poikyloblastic pyrite is occasionally found (101914, 101915).

Haematite is almost ubiquitous accessory but is only concentrated to any extent in chloritic schists. The characteristic occurrence of haematite as a network of lamellae in the quartz chlorite schists was referred to in Section 4.1 and sulphides, particularly chalcopyrite are frequently closely associated with this haematite (37882, 37883, 37884). Evidence for a pre-deformation origin for the haematite is provided by the alignment of haematite lamellae shown in Plate 33.
Plate 35  Euhedral pyrite porphyroblasts in chalcopyrite.  
Note the post-recrystalline embayment of the 
pyrite crystals.
101909    x 95     reflected light, 
parallel polars

Plate 36  Recrystallized pyrite euhedra in quartz.
101915    x 62     reflected light, 
parallel polars
Of the other phases, molybdenite is present as lamellae in quartz sericite schist (101916); tennantite and galena occur as minute inclusions in poikyloblastic pyrite (101915), the latter mineral also occurring as isolated grains in 37882. Sphalerite is widely spread but is always present in trace amounts. Gold is very rare, and typically occurs in secondary quartz veins (101917; Plate 37), a situation similar to that reported from North Lyell by Edwards (1939).

Hence it is obvious that metamorphic textures are dominant in the Cape Horn ores, effectively obscuring any original texture. In particular the recrystallization of pyrite is a feature typical of greenschist facies rocks in Norway (Vokes, 1969) and it is apparent that the ores have undergone the same grade of metamorphism as is common in Western Tasmania.

The Cape Horn ores, with their simple mineralogy and metamorphic texture, are virtually identical to the West Lyell ores as documented by Markham (1968). The similarity extends to the host rocks, the occurrence of apatite rich bands in tuffaceous schists probably being the most striking parallel and it is likely that detailed mapping between West Lyell and Cape Horn might reveal a stratigraphic correlation.

5.2 **LYELL COMSTOCK**

Little is known about the Lyell Comstock ore bodies which were closed after the Second World War. The copper-rich ore occurred in the form of four discrete N to NE trending lenses dipping steeply to the west (Solomon, 1964).
Plate 37  Irregular gold grain in post-cleavage quartz vein, Cape Horn chalcopyrite grain at right.
101917 x 350 reflected light

Plate 38  Deformed molybdenite lamellae in siliceous gangue, Comstock. Whitish grains (high reflectivity) at centre and upper left of plate are pyrite.
101919 x 160 reflected light, parallel polars
A reconnaissance study of the ore was carried out on specimens held by the Mt. Lyell Mining and Railway Co. museum. Minerals identified were chalcopyrite, bornite, chalcocite, covellite, pyrite, molybdenite, galena, sphalerite, magnetite, haematite, digenite, tennantite, mawsonite, betekhtinite and stromeyerite. In addition Edwards (1939) reported gold. The main type of ore present was of the disseminated chalcopyrite, bornite, pyrite type. Texturally this type of ore displays the same deformation textures as the Cape Horn ore. In chloritic schists (101906; Plate 20) fine-grained chalcopyrite is closely associated with haematite which replaces magnetite. Very little pyrite is present. The chalcopyrite grains are frequently frayed with flame-like terminations strung out along the cleavage. This specimen is described in the Mt. Lyell collection as "typical schist ore". One specimen (101919; Plate 38) is rich in molybdenite, which occurs as lamellae strung out along the cleavage or as deformed plates.

In siliceous schists (101918) the character of the ore is markedly different. Fine-grained, often finely porous, pyrite with irregular, rounded outline is abundant. However, especially when in association with chalcopyrite and bornite, pyrite is frequently recrystallized to coarser, generally pyritohedral, poikiloblasts with inclusions of gangue, chalcopyrite and bornite. In one case (Plate 39) a zone of chalcopyrite, partially replacing pyrite is enclosed by a rounded growth of pyrite with a finely porous rim. The pyrite textures are typical of the "relict colloform textures" described by Markham (1968) from Crown Lyell ore.
Plate 39  Irregular chalcopyrite-pyrite intergrowth
(possibly "atoll" type replacement of pyrite by chalcopyrite) surrounded by pyrite with a porous outer rim (relict colloform texture ?), Comstock. 101918  x 160  reflected light, parallel polars

Plate 40  Intergrowth of betekhtinite (cream), bornite (brown) and chalcopyrite (yellow) with associated minor galena (white), mawsonite (orange-brown) and chalcocite (grey-blue, lower left), Comstock. 10192D  x 350  reflected light, parallel polars
Bornite and chalcopyrite occur as coarse, irregular "splashes" enclosing pyrite euhedra. In some cases bornite has partially replaced chalcopyrite and pyrite. Minor fine covellite veinlets, of probable supergene origin, are occasionally present within and at the margin of bornite grains. Chalcocite also appears to be a secondary sulphide. Mawsonite and galena are present in trace amounts.

Two specimens of massive copper-rich ore were also available to the author. One (101920) consisted dominantly of a myrmekitic-type intergrowth of chalcocite and chalcopyrite, although chalcocite consistently embays the chalcopyrite and it is clear that some replacement of chalcopyrite has taken place. Minor bornite is present as isolated grains and also rarely, due to partial replacement by chalcocite, in lattice intergrowths with chalcopyrite showing a "basket-weave texture" (Brett, 1964). Minor galena, betekhtinite, tennantite, mawsonite and stromeyerite are also present. Betekhtinite occurs in association with chalcocite, bornite, chalcopyrite and galena, most commonly at chalcocite or bornite-galena grain boundaries (Plates 40 and 41). Mawsonite is rare occurring as fine rounded grains at betekhtinite-bornite grain boundaries.

Stromeyerite, as is the case with the greater proportion of chalcocite, appears to be secondary and either occurs as a thin film moulded on boundaries of chalcocite against either galena or sphalerite (Plate 42) or as grains associated with chalcocite in chalcopyrite (Plate 41).

The other specimen (101921) consisted of coarse pyritohedral poikiloblasts or fine grains of pyrite similar to 101918 enclosed in bornite and a siliceous gangue. Minor irregular chalcopyrite grains
Plate 41 Chalcocite (blue-grey), stromeyerite (light brown), bornite (brown), galena (white) and betekhtinite (cream) in chalcopyrite, Comstock.

10192D x 350 reflected light, parallel polars

Plate 42 Rim of stromeyerite (dark brown) at galena (white) - chalcocite (green-brown) boundary with chalcopyrite (yellow), Comstock.

10192D x 660 reflected light, polars partly crossed
are also present, the chalcopyrite as well as both the pyrite and gangue are embayed by bornite. Small rounded grains of mawsonite are common in the bornite and gangue. Chalcocite is present as thin films at bornite-gangue grain boundaries and as grains in bornite. Rounded tennantite grains occur in bornite. Digenite is present as fine veins in bornite possibly outlining bornite grain boundaries. Concentrations of fine pyrite crystals and gangue occur at "triple points" where three veins meet (Plate 43).

It is interesting to note that the mineralogy and texture of the specimens described is identical with that of North Lyell ores as described by Markham. In particular betekhtinite has been identified previously from North Lyell (Markham, op. cit.; Markham and Ottemann, 1968). Mawsonite occurs at West Lyell but is only abundant at Crown Lyell and North Lyell (Markham, op. cit.; Markham and Lawrence, 1965).

5.3 ORES ASSOCIATED WITH THE COMSTOCK CHERT

5.3.1 Chert Breccia Ores

Comstock chert breccia ores were briefly described by Markham (1968). Specimens 101922 and 101923 are typical of this ore and consist of irregular networks of fine chalcopyrite veins (101922) or patches of sphalerite (101923) in a matrix of brecciated chert. The irregular margins of the grains heal fine fractures in the chert. Pyrite is rare but in 101922 occurs in the form of clusters of fine relict fremboids suggestive of formation as a gel. This is consistent with the very fine porous nature of the ore in this specimen. Associated minerals include galena and tennantite. Rutile and haematite appear to pre-date the sulphides and in 101922 chalcopyrite is moulded onto and embays a rutile grain.
Plate 43  Veins of digenite (blue) in bornite meeting at triple point at which there is a concentration of pyrite and gangue, Comstock.
101921  x 95  reflected light, parallel polars

Plate 44  Mutual boundary texture of sphalerite-chalcopyrite in chert-breccia ore, Comstock. Note the small rounded tennantite inclusion (grey) in the sphalerite.
101923  x 160  reflected light, parallel polars
Within the sphalerite patches mutual boundary texture between sphalerite and chalcopyrite (Plate 44) is present. The sphalerite has a pale honey coloured internal reflection indicating a low iron content. Markham (1968) described hexastannite from this ore, this mineral occurring as exsolution blebs in sphalerite.

Similar sulphides occur sporadically west of Comstock largely in "chert" at the contact of the keratophyric lavas and underlying lithic tuffs (101924).

5.3.2 Banded Chert Ores

This ore is distinguished from the chert breccia largely on the basis of texture. The ore is prominently banded due to layers of pyrite up to 0.5 cm. thick in chert (101905). Minor pyrite fragments also occur. In general the pyrite has been recrystallized to medium-grained euhedral to slightly rounded porphyroblasts which enclose the other sulphides, although some very fine pyrite showing a relict frambooidal texture is present especially in 101925.

The pyrite rich layers show markedly different compositions ranging from virtually massive pyrite with minor interstitial galena, to layers rich in sulphides, notably a lead-zinc assemblage with rounded sphalerite and tennantite grains in a galena matrix and a cupriferous assemblage in which tennantite, chalcopyrite and hexastannite are the dominant "matrix" phases. A "triple point" texture is apparent in hexastannite under crossed nicols indicating an approach to textural equilibrium. The banded ores are probably of exhalative origin.
An interesting feature of the mineralization is the presence of "sulphide breccias" overlying the lithic tuffs in drill hole NL 968. These consist of (101926a,b) coarse (up to 1.5 cm.) clastic fragments of chert and pyrite in a chert matrix with minor sulphides. The fragments are mainly pyrite, normally massive, but layers of chert with clusters of fine recrystallized pyrite grains (relict frambooids ?) also occur. Also in 101926a two chalcopyrite grains are present; these have been deeply corroded and largely replaced by secondary chalcocite. Associated minerals are galena and minor sphalerite, hexastannite and tennantite. This breccia ore is overlain by clastic breccias in which volcanic fragments in all stages of alteration, as well as chert and sulphide fragments, are present.

It is extremely probable that the breccias were formed by erosion of the banded ores followed by rapid deposition possibly into an actively forming structural trough. The occurrence of overlying breccias containing volcanic fragments is consistent with this hypothesis, the later appearance of these fragments being due to complete removal of chert somewhere along strike.

Similar breccias (101927) occur along strike, that is nearer the present surface, in drill hole NL 1000 and consist of fragments of chert and jasper in a cherty matrix. Associated sulphides include irregular chalcopyrite masses and veins with associated pyrite porphyroblasts and minor haematite.
5.4 TASMAN SHAFT ORE

In this discussion the massive ore present in two lenses in the Comstock open cut will also be included. These ores consist of massive banded pyrite-galena-sphalerite lenses with minor chert, which, in the Lyell Comstock open cut, are concordant with the cleavage in the surrounding schists (Plate 45). On the basis of texture and stratigraphic position the ores are tentatively correlated with the banded chert ores previously discussed, the latter having a higher silica content probably because of a greater proximity to the exhalative source.

Colloform textures are present with strong development of concentric structure with fine radial cracks in the pyrite of the open cut ore (Plate 46; 101928) and are slightly disrupted due to later deformation effects. Sphalerite occurs as irregular grains often rounded and enclosed in galena. Tennantite occurs as rounded grains also normally enclosed in galena. Traces of arsenopyrite are present, characteristically having a skeletal form having been partially replaced by galena although the residuals are occasionally recrystallized to laths or rhombs. A thin coating of bournonite is sometimes present on the arsenopyrite grains. In addition, Markham (1968) identified secondary chalcocite and covellite in the ore.

Deformation, however, masks most of the primary textural features of the ore. Hand specimens from the Tasman Shaft and also the more southerly lens in the Comstock open cut are tightly isoclinally folded (101929; Plate 47), chert layers within the ore are boudinaged and pyrite is strongly brecciated. In polished section
Plate 45 Massive pyrite-galena-sphalerite lens, Comstock open cut.

Plate 46 Colloform pyrite with associated galena and sphalerite from Comstock lens.

101928 x 62 reflected light, parallel polar
Plate 47  Banded pyrite-galena-sphalerite. Deformation phenomena included brecciation of pyrite, tight folding and boudinage of chert layer (left).
it can be seen that the brecciated pyrite grains form augen around which galena has flowed plastically (101930, 101931).

These deformation textures were also noted by Rafter and Solomon (1967) and Markham (1968) who made a comparison between the Tasman Shaft and massive Rosebery ores.

5.5 DEVONIAN REMOBILIZATION

Quartz "splashes" often carrying sulphides, were recognized as post-cleavage by Solomon (1964). These tensional quartz veins, representatives of a Devonian phase of mineralization, are common throughout the area and are undoubtedly hydrothermal in origin (Loftus-Hills et al., 1969) as indicated by notably lower Se (Loftus-Hills et al., op. cit.), Co and Ni in pyrite (Loftus-Hills and Solomon, 1967; Loftus-Hills, 1968) in the tension gashes at West Lyell compared with pyrite in the surrounding disseminated ore.

Texturally chalcopyrite from a similar quartz vein at Cape Horn (101932) reveals on etching with NH₄OH/H₂O little of the deformation twinning characteristic of chalcopyrite in the schists. This is consistent with Markham's (1968) observation of decussate intergrowths of chalcopyrite and calorite in West Lyell ore, a texture which is probably present in 37922, although this may also be due to non-hydrothermal recrystallization.

Further evidence of a post-deformation phase of hydrothermal activity is provided by the common occurrence of embayed recrystallized pyrite porphyroblasts both in chalcopyrite (101909, 101912) and in the siliceous gangue (101912) as shown in Plate 35.
A Devonian phase of sulphide reworking is only to be expected in view of the development of fractures during Tabberabberan deformation coupled with the fact that pressures and temperatures of approximately 2 kilobars and 300°C are consistent with lower greenschist facies metamorphism. Regional metamorphism as a possible means of generating a hydrothermal fluid has been recognized by many workers (e.g. Barnes and Czamanske, 1967; Krauskopf, 1967).

The major remaining problem is to attempt an evaluation of the relative importance of Cambrian and Devonian processes in ore concentration and localization, a problem which is largely outside the scope of the present investigation.

However, two pieces of independent evidence tend to suggest a minimal effect of a Devonian phase of mineralization:

(i) The minerals within the Devonian tension gashes reflect closely that of the surrounding altered and mineralized country rocks. In the pyritic sericite-quartz schists of the Queen Lyell area the tension gashes have a quartz-pyrite mineralogy (37923); in the carbonate zone quartz-carbonate veins are present (37924) and in the Cape Horn area a quartz-chlorite-chalcopyrite remobilization assemblage predominates (101932, 37925). On textural grounds (Solomon, 1964; Markham, 1968; this study) it is clear that the alteration is pre-cleavage and, on the basis of included fragments in Cambrian and younger sediments (Solomon, 1967; this study) probably Cambrian in entirety. The similarity between tension vein and enclosing schist mineralogy, particularly the lack of chalcopyrite-bearing veins outside areas of disseminated
copper mineralization, suggests that the veins were largely formed by lateral secretion in situ in the manner suggested by Boyle (1961) rather than by a process involving solution transport over long distances. However a detailed geochemical study would be required to test the hypothesis.

(ii) Devonian remobilization has not resulted in any replacement mineralization in limestones and calcareous sediments overlying mineralized cherts in the Lyell Comstock area. A possible exception is the chert conglomerate (32897) which overlies the main chert mass, but this is probably due to remobilization of sulphide due to hydrothermal solutions being forced outwards during crystallization of the chert.

5.6 RELATIONSHIP OF MINERALIZATION TO STRATIGRAPHY AND ALTERATION

The presence of a crude zonation in the ore bodies was mentioned by Markham (1962) and Solomon (1964), but was defined purely on geographical grounds.

Stratigraphically the host rocks for the Comstock mineralization are younger than those at Cape Horn, as established by the present study. A similar relationship probably exists between the host rocks of the North Lyell-Crown Lyell mineralization and those of the West Lyell ores, as was also suggested by Loftus-Hills (1968). The exhalative pyritic ore bodies of the Lyell Comstock area and probably the Iron Blow are the uppermost mineralized horizons in the Cambrian succession. (Fig. 7).

The West Lyell and Cape Horn disseminated chalcopyrite-pyrite ore bodies occur in sericitic schists and their formation is almost
Palaeoprofile of the Cape Horn-Lyell Comstock Area Prior to Extrusion of the Quartz Keratophyric Lapilli Tuff. (Not to Scale).

- Limestone, Siltstones & Minor Chert
- Massive Banded Sulphides
- Banded Chert-Sulphide Ore
- Chert-Sulphide Breccia grading up to Greywacke Breccia
- Silicified Volcanics (?)
certainly connected with the hydrothermal alteration. Similarly the bornite-digenite-chalcocite rich assemblages such as those at North Lyell and Lyell Comstock are deposited at high oxygen fugacity and are associated with advanced argillic alteration (Meyer and Hemley, 1967).

The wide range in alteration assemblages over a small distance towards the top of the volcanic pile contrasts markedly with the broader zones of alteration near Cape Horn and West Lyell. This may be attributed to rapid changes in the pH of the hydrothermal fluids over small areas, a feature exhibited by geothermal districts.

In a detailed discussion of natural geothermal waters, Ellis (1967) states that at temperatures between 250°C and 350°C the solubility of silica from volcanic rocks approaches that of amorphous silica; that is, it greatly exceeds the solubility of quartz. During steam loss at lower pressures amorphous silica is rapidly deposited, its solubility at 150°C being 620 ppm compared with 400 ppm at 100°C. This results in the common occurrence of opal and siliceous sinter at and near the surface in volcanic areas (Ellis, op. cit.).

In such sinters pyrophyllite and alunite are commonly associated (Meyer and Hemley, 1967); although the lack of alunite at North Lyell and Lyell Comstock may be related to the scavenging of sulphate from solution during barite formation.

Furthermore, it is significant to note that in the Salton Sea area, California, scales of copper-rich sulphides including digenite, bornite, chalcocite, stromeyerite, tetrahedrite, chalcopyrite and pyrite in an opaline matrix were formed at 130°C to 220°C in discharge pipes at the outlet of a metal-charged hydrothermal brine. It is also
interesting that despite the fact that the lead and zinc concentrations exceeded the copper content by factors of 10 and 100 respectively, the copper-rich phases were selectively precipitated (Skinner et al., 1967).

The similarity of the Mt. Lyell and Salton Sea cases suggests a similar mode of origin for the bornite-rich ores in the former case. An obvious objection lies in the fact that much of the chalcocite and all of the stromeyerite appears to be secondary. A primary origin for some of the digenite at least is indicated by lattice intergrowths of digenite and chalcocite, as noted by Edwards (1939) and Markham (1968) on North Lyell ore, although the former writer termed digenite "blue chalcocite". From the assemblages little can be gleaned about the temperature of ore deposition since kinetic studies by Brett (1964) indicate that phases in the systems Cu-S and Cu-Fe-S cannot be quenched in nature. However enargite has been identified at North Lyell by Edwards (1939) and Markham (op. cit.) and at the Iron Blow. Studies on the Cu-As-S system (Skinner, 1960) indicate that enargite forms at temperatures in excess of 320°C, although the effect of iron on the system is unknown.

The "chert" ores pose a problem as it is difficult to distinguish between sub-surface "fracture-filling" ores and exhalative ores on the basis of texture although this has been attempted. The texture and stratigraphic position of the banded massive and chert-rich ores is clearly indicative of an exhalative origin whereas structural evidence is probably the best in favour of a replacement origin for part of the chert.

The chert ores provide a transition between copper-rich and lead-zinc assemblages and are important for that reason alone.
Thus there is a correlation between alteration assemblages and ore types, but, equally important, a stratigraphic control is present. The lowest mineralized horizons (Cape Horn, West Lyell) are typified by a simple mineralogy with pyrite and chalcopyrite predominant, and a low Cu:S ratio (Table 2). The higher level bornite-rich ores (North Lyell and Lyell Comstock in part) have a low pyrite content and a high Cu:S ratio. In these ores Ag, Zn and Pb are enriched relative to the West Lyell type and phases bearing trace elements such as Sn (mawsonite, hexastannite) appear. In their high Pb and Zn contents the ores appear to be transitional to the Iron Blow and Tasman Shaft mineralization.

Significantly, the Salton Sea example shows that the composition of an ore formed is more strongly related to the environment of deposition than to the relative concentration of certain elements in the hydrothermal fluid.

<p>| TABLE 2 |
| Compositions of Lyell Ores |</p>
<table>
<thead>
<tr>
<th>% Cu</th>
<th>% Fe</th>
<th>% S</th>
<th>Au (oz/ton)</th>
<th>Ag (oz/ton)</th>
<th>% Zn</th>
<th>% Pb</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Cape Horn</td>
<td>1.60</td>
<td>2.1</td>
<td>2.4</td>
<td>0.032</td>
<td>0.045</td>
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<tr>
<td>2</td>
<td>Comstock</td>
<td>2.07</td>
<td>10.7</td>
<td>8.2</td>
<td>0.012</td>
<td>0.125</td>
</tr>
<tr>
<td>3</td>
<td>West Lyell</td>
<td>0.7</td>
<td>9.5</td>
<td>6.1</td>
<td>0.012</td>
<td>0.067</td>
</tr>
<tr>
<td>3</td>
<td>North Lyell</td>
<td>6.25</td>
<td>4.5</td>
<td>6.0</td>
<td>0.017</td>
<td>0.341</td>
</tr>
<tr>
<td>4</td>
<td>Iron Blow</td>
<td>1.0</td>
<td>43.8</td>
<td>50.0</td>
<td>0.04</td>
<td>1.50</td>
</tr>
</tbody>
</table>

1 Mr. K. O. Reid, pers. comm.
2 Edwards (1939)
3 Edwards (1939), and Wade and Solomon (1958)
4 Wade and Solomon (1958).
6. GEOCHEMISTRY OF ORE DEPOSITION

6.1 COBALT, NICKEL AND SELENIUM IN PYRITE

Studies of Co, Ni and Se contents of Mt. Lyell sulphides (Loftus-Hills, 1967; Loftus-Hills and Solomon, 1967; Loftus-Hills, 1968; Loftus-Hills et al., 1969) are extremely significant when viewed in the light of stratigraphy and mode of origin of the ore bodies. These results are summarized as averages in Table 3.

TABLE 3

Average Cobalt, Nickel and Selenium in Mt. Lyell Sulphides
(Results for Co and Ni from Loftus-Hills, 1968; for Se from Loftus-Hills et al., 1969. py = pyrite. cp = chalcopyrite.)

<table>
<thead>
<tr>
<th>Ore Body</th>
<th>Mineral</th>
<th>Co (ppm)</th>
<th>Ni (ppm)</th>
<th>Co/Ni</th>
<th>Se (ppm)</th>
<th>S/Se</th>
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<tbody>
<tr>
<td>WEST LYELL</td>
<td>py</td>
<td>881</td>
<td>163</td>
<td>5.40</td>
<td>69</td>
<td>7,740</td>
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<tr>
<td>(disseminated ore)</td>
<td>cp</td>
<td>306</td>
<td>68</td>
<td>4.50</td>
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<td></td>
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<tr>
<td>WEST LYELL</td>
<td>py</td>
<td>81</td>
<td>112</td>
<td>0.72</td>
<td>115</td>
<td>4,270</td>
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<tr>
<td>(quartz veins)</td>
<td>cp</td>
<td>19</td>
<td>19</td>
<td>1.00</td>
<td>50</td>
<td>7,000</td>
</tr>
<tr>
<td>CROWN LYELL</td>
<td>py</td>
<td>125</td>
<td>97</td>
<td>14.4</td>
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<td>5,800</td>
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<tr>
<td>(disseminated ore)</td>
<td>cp</td>
<td></td>
<td></td>
<td></td>
<td>114</td>
<td>3,070</td>
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<tr>
<td>NORTH LYELL</td>
<td>py</td>
<td>162</td>
<td>52</td>
<td>3.12</td>
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<td>2,040</td>
</tr>
<tr>
<td>(massive)</td>
<td>cp</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>IRON BLOW</td>
<td>py</td>
<td>54</td>
<td>52</td>
<td>1.04</td>
<td>10*</td>
<td>53,400</td>
</tr>
<tr>
<td>COMSTOCK</td>
<td>py</td>
<td>21</td>
<td>24</td>
<td>0.88</td>
<td>109</td>
<td>4,900</td>
</tr>
<tr>
<td>(massive py lens)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>TASMAN SHAFT</td>
<td></td>
<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
</tbody>
</table>

* one sample
The above figures indicate the following trends:

(i) There is a positive correlation between the Co content of pyrite and the Cu concentration in the mineralized areas (Loftus-Hills, 1967). The high Co content in pyrites in the Crown Lyell-North Lyell area is consistent with the identification of linnaeite by Edwards (1939) and Markham (1968).

(ii) There is a strong trend between the Ni content of pyrites and the stratigraphic level of mineralization, the Ni content being lowest for the exhalative bodies, namely the Iron Blow and Tasman Shaft ores and the massive pyritic lens at Comstock.

(iii) The selenium content closely follows Cu, being enriched in the North Lyell chalcopyrites; this being expected in view of the probable presence of berzelianite at North Lyell (Edwards, 1939; Loftus-Hills et al., 1969). However the Se content of the Iron Blow pyrites are the highest of the field, Se being concentrated in surface volcanic processes (Loftus-Hills and Solomon, 1967; Loftus-Hills et al., 1969). The average value of Se in the Tasman Shaft pyrite is higher than that of Crown Lyell pyrites and comparable with that of North Lyell chalcopyrites. The low Se content of the Comstock pyrite is probably a reflection of inadequate sampling at this locality.

(iv) Devonian remobilization results in a marked impoverishment in Co relative to Ni of pyrites compared with those of the disseminated ores. There is also a partitioning of Se, this element being enriched in chalcopyrite relative to pyrite in the quartz veins (Loftus-Hills, 1967; Loftus-Hills et al., 1969). This trend is the reverse of the extremely high Co/Ni
ratios in pyrites in the Crown Lyell area and is inconsistent with a major phase of Tabberabberan mineralization being responsible for the North Lyell and Crown Lyell ores.

In conclusion, the results show a strong correlation with the stratigraphic horizon of mineralization, and are consistent with results on trace element concentrations in other ore deposits of volcanic origin, for example Rio Tinto, Spain (Hegemann and Leybold, 1954) and Matagami and Noranda, Canada (Roscoe, 1965). The results from Mt. Lyell are also markedly different from all the Tasmanian Devonian Sn and Pb-Zn deposits (Loftus-Hills, 1967; Loftus-Hills et al., 1969).

6.2 SULPHUR ISOTOPE STUDIES

Two different types of sulphur are present in the Mt. Lyell ore bodies as shown by Solomon et al.,(1969). The first group includes ores from the Iron Blow, West Lyell, Comstock and Tasman Shaft which had a range of $\delta^{34}S$ values from $+5.2$ to $+10^\circ/oo$ with an average of $+7^\circ/oo$. By way of contrast sulphides from the North Lyell-Crown Lyell area had a $\delta^{34}S$ range of $-7.7$ to $+0.8^\circ/oo$, averaging $-2.6^\circ/oo$.

These results conflict with the Co, Ni and Se contents of sulphides which seemed to favour one phase of mineralization caused by stratigraphically ascending hydrothermal fluids. For the copper-rich ores of the North Lyell area Solomon et al. (op. cit.) suggested three possibilities:-

(i) The sulphur for these ores could have been derived by a preferential remobilization of sulphur enriched in the light $S^{32}$ isotope during movement on the North Lyell Fault or
possibly a late volcanic stage.

(ii) The ores may have been formed during an entirely separate Devonian phase of mineralization.

(iii) The ores may owe their distinct mineralogy and isotopic composition to supergene weathering during and subsequent to the Jukesian Orogeny. In support of this proposal the case of chalcocite at the "King Lyell" copper-clay deposit which probably formed during a period of supergene enrichment, possibly Tertiary (Solomon, 1969), was cited. This chalcocite has a $\delta^{34}S$ of -3.6‰.

The possibility that the North Lyell-Crown Lyell ores were generated during movement on the North Lyell Fault does not seem likely in view of the fact that:

(i) Markham (1968) noted the same deformational textures in Crown Lyell ore and West Lyell ore, the former presumably having undergone both TF1 and TF2 deformation. Any additional effects in the former ore due to movement on the North Lyell Fault were not noted.

(ii) Ores identical in texture and mineralogy to those at Crown Lyell and North Lyell occur at Lyell Comstock which has not been a zone of movement to the extent experienced at North Lyell.

Hypothesis (ii) is also not favoured since geochemically and mineralogically the North Lyell ores fit a definite stratigraphic zonation and can be adequately explained by a Cambrian mechanism (section 5.6). The "missing link" in the zonation is provided by the chert breccia ores at Comstock which are transitional in type between the copper-rich and pyritic exhalative ores.
At this stage hypothesis (iii) is favoured. The North Lyell ores were certainly exposed during Middle Owen sedimentation as shown by the chert-haematite breccia overlying the ore body (Solomon, 1967). They may also, as has been demonstrated in the case of the Comstock ores, have been exposed during Middle Cambrian erosion and near the surface during deposition of the Jukes Conglomerate. Thus supergene enrichment and oxidation of ore is to be expected and is supported by the observation that the bulk of the chalcocite at North Lyell (Edwards, 1939; Markham, 1968) and at Comstock is secondary. Isotopic exchange and re-equilibration probably occurred during these prolonged phases of weathering.

Barite occurs as irregular pre-cleavage veins at West Lyell, in the North Lyell area (Solomon et al., 1969) and both within the chert and schist in the Lyell Comstock area (37934). The West Lyell sulphate is probably related to a late-stage high oxygen fugacity phase. $\delta^{34}S$'s for West Lyell veins were in the range +23.5 to 26.2$^\circ$/oo with $\delta^{18}O$'s in the range +7.8 to +8.1$^\circ$/oo compared with $\delta^{34}S$'s between +20.9 and 25.7$^\circ$/oo and a $\delta^{18}O$ of 11.2 ± 0.2$^\circ$/oo for the North Lyell veins. Solomon et al. (1969) regarded a sea-water or connate origin for the barite as being unlikely considering the relatively low $\delta^{34}S$ values.

Residual barite in the North Lyell chert-haematite scree breccia (Solomon, 1967) has a $\delta^{34}S$ similar to that of the underlying sulphate (Solomon et al., 1959), but those from the Iron Blow haematite "gossan" have higher $\delta^{34}S$ values (+36.6 to +41.6$^\circ$/oo) indicating that the parent Cambrian barite may have been similarly enriched in the heavy isotope and thus a sea water origin for the sulphate is not precluded.
The relative "lightness" of the North Lyell barites may be due to exchange during weathering processes in which the nearby sulphides were also enriched in $^{32}S$.

It is thus probable that the ore sulphur of the Mt. Lyell area was derived, at least initially, from a single source, possibly seawater. Thus if a Cambrian origin is accepted for the "exhalative" ores of the Iron Blow, Tasman Shaft and Comstock Open Cut it is logical to extend the argument to the underlying disseminated ores, the only possible exception being the North Lyell ores. The $^{34}S$ values of the sulphides in post-cleavage quartz veins are identical to those in the surrounding disseminated ores, supporting the "lateral secretion" hypothesis for the origin of these veins and implying no further supply of sulphur was introduced.

6.3 MODE OF ORE FORMATION AT CAPE HORN

It may be deduced from the preceding sections that it is believed that the Cape Horn ores formed in a sub-surface environment, as has been suggested for the West Lyell ores by Rafter and Solomon (1967), Markham (1968) and Solomon et al. (1969) who envisaged a mechanism of mineralization involving pore filling and replacement of silicates.

Stanton (1966) has proposed a method by which these two processes may be distinguished. Stanton plots analyses of sulphide-bearing rocks on three component variation diagrams. Any consistent decrease in one component concurrent with a corresponding increase in sulphide content implies replacement of the component in question by sulphide. If the results show a random scatter, formation of sulphides by bulk addition to the rock is indicated.
The advantage of Stanton's method is that it may be used successfully on ores in which evidence of the depositional process has been obscured by later events, in this case metamorphism and deformation. A further advantage of the method is that small samples may be used.

A possible disadvantage of the method if applied to Mt. Lyell ores is that later hydrothermal remobilization may modify or destroy pre-existing patterns. At Cape Horn less than 3% (visual estimation from drill core logging and field mapping) of all sulphides occur in post-cleavage veins and thus remobilization should not substantially interfere with the application of the method.

A series of medium-grained fragmental quartz sericite-chlorite carbonate schists was chosen because all minerals of the hydrothermal assemblage were represented and nearly the widest possible range of sulphide concentrations for the Cape Horn ore was revealed (Table 4, nos. 1-11). In addition a chlorite-sericite-quartz schist with haematite-chlorite "blebs" (Table 4, no.12) and an almost sulphide-free schist from the carbonate zone (Table 4, no.13) were analysed. Only one specimen (Table 4, no.1) appeared to contain appreciable remobilized carbonate.

The results are adequately summarized in Figs. 8 and 9, and in both cases the points associated with progressively higher S contents in the range 0-10% show a wide scatter indicating the ores formed by a process of addition of S and metals to the host rocks. It is interesting to note that preliminary results on sulphide-rich quartz sericite schists indicate ore formation by replacement of sericite (Dr. M. Solomon, pers. comm.). This apparent anomaly requires further comment.
### TABLE 4

Composition of Hydrothermally Altered Rocks

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<tr>
<th>Specimen Number</th>
<th>Locality</th>
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</thead>
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<td>1</td>
<td>D.H. CH 2/329'</td>
</tr>
<tr>
<td>2</td>
<td>D.H. CH 2/331'</td>
</tr>
<tr>
<td>3</td>
<td>D.H. CH 2/343'</td>
</tr>
<tr>
<td>4</td>
<td>D.H. CH 2/349'</td>
</tr>
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<td>5</td>
<td>D.H. CH 2/359'</td>
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<td>6</td>
<td>D.H. CH 2/363'</td>
</tr>
<tr>
<td>7</td>
<td>D.H. CH 2/370(\frac{1}{2})'</td>
</tr>
<tr>
<td>8</td>
<td>D.H. CH 2/376(\frac{1}{2})'</td>
</tr>
<tr>
<td>9</td>
<td>D.H. CH 2/379'</td>
</tr>
<tr>
<td>10</td>
<td>D.H. CH 2/422'</td>
</tr>
<tr>
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<td>D.H. CH 2/444'</td>
</tr>
<tr>
<td>12</td>
<td>3492W, 3835N</td>
</tr>
<tr>
<td>13</td>
<td>D.H. KL 968/621'</td>
</tr>
</tbody>
</table>

1-9: Medium grained fragmental quartz, sericite, chlorite, "siderite" schists with varying amounts of pyrite and chalcopyrite.

10: Finer grained than 1-9; a dark green quartz-chlorite sericite siderite schist.

11: Similar to 10 but less chloritic.

12: Chlorite-sericite quartz schist with haematitic clots.

13: Carbonate rock schist belonging to the lithic tuffs.
### TABLE 4
Composition of Hydrothermally Altered Rocks

<table>
<thead>
<tr>
<th></th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
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</thead>
<tbody>
<tr>
<td>SiO₂</td>
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<tr>
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<td>1.53</td>
<td>1.61</td>
<td>0.98</td>
<td>1.18</td>
<td>0.30</td>
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<td>1.85</td>
<td>2.21</td>
<td>0.02</td>
<td>-</td>
</tr>
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<td>3.26</td>
<td>3.08</td>
<td>5.47</td>
<td>1.75</td>
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<td>1.20</td>
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<td>3.34</td>
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</tr>
<tr>
<td>S</td>
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<td>3.97</td>
<td>3.58</td>
<td>6.38</td>
<td>2.01</td>
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<td>1.12</td>
<td>0.51</td>
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<td>4.09</td>
<td>1.61</td>
<td>Tr.</td>
</tr>
<tr>
<td>BaO</td>
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<td>0.15</td>
<td>0.12</td>
<td>0.01</td>
<td>N.D.</td>
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<td>0.30</td>
<td>0.095</td>
<td>0.125</td>
<td>0.770</td>
<td>0.33</td>
<td>0.17</td>
<td>0.06</td>
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</tbody>
</table>

* CO₂ not determined probably very minor.

N.D. - not determined
Tr. - rock contains traces of pyrite; S not determined.
$\text{Al}_2\text{O}_3 - \text{CO}_2 - \text{K}_2\text{O} - \text{S}$ Variation Diagram

- $< 2\%$ S
- $2 - 4\%$ S
- $4 - 6\%$ S
- $> 6\%$ S

Fig. 8
SiO$_2$-CO$_2$-Al$_2$O$_3$-S Variation Diagram

Fig. 9
The series of rocks analysed in this study belong to the "Ignimbrites and Minor Siltstones" horizon and if not ignimbrites they were originally tuffs containing a high proportion of pumiceous material. Ross and Smith (1960) state that the porosity range for ash-flow tuffs is commonly between 20 and 70% although Marshall (1935) reported porosities in the range 3.4-28.6% for ignimbrites of the Lake Taupo area, New Zealand. Also, V. Starostin (quoted by Smirnov, 1970) found a linear relationship between the effective porosity of siliceous volcanics and the degree of alteration as expressed by the sum of the chlorite and sericite percentages.

A sudden drop in pressure of a hydrothermal fluid causes "flashing off" of steam and thus may cause ore deposition by supersaturation of the residual solution as has been suggested, for example, by Holland (1967). This mechanism is possibly significant for the Comstock and North Lyell ores which were probably deposited in shallow fracture zones (Markham, 1968) and similarly, porous zones in the volcanics could be favourable host rocks for the same reason. Certainly, a high proportion of the mineralization occurs in fragmental sericite-chlorite-quartz schists at West Lyell similar to those analysed at Cape Horn. Notably, within the Cape Horn ore zone the altered siltstones and conglomerates are markedly depleted in sulphides (e.g. 37912) compared with the surrounding fragmental rocks, thus supporting the hypothesis that the original porosity of the tuffs could be a factor favouring their role as ore hosts. The mineralization in the quartz-sericite schists at West Lyell presently being studied by Solomon could be due to the position of these rocks in a more widely mineralized area. Also there is the
possibility that a higher sulphur availability at West Lyell caused relatively non-selective mineralization involving replacement.

6.4 BARIUM AS A GUIDE TO ORE

The presence of 1.26% Ba, as a replacement for K in sericite of the Rosebery footwall schists (Gee, 1970) and similar concentrations in sericitic schists at West Lyell (Dr. M. Solomon, pers. comm.) has given rise to the suspicion that Ba may be an indicator of economic mineralization. The values for the Cape Horn rocks are much lower (Table 4), a reflection of:

(i) relatively low K$_2$O values in the schists analysed since Gee (op. cit.) found a strong positive correlation between K$_2$O and Ba content, and

(ii) the position of the Cape Horn ores near the edge of the zone of strong alteration, most rocks having similar Ba contents to "unaltered" Tasmanian quartz keratophyres, which average 1070 ppm BaO (8 analyses - Solomon, 1964).

The absence of barite as a first generation alteration mineral at Cape Horn and West Lyell enables some constraints to be placed on the chemical nature of the ore-forming environment. Barite is however fairly abundant in the Lyell Comstock and North Lyell areas which is compatible with probable relatively high oxygen fugacities near the surface.

The high barium concentrations in the altered zones could have been introduced by either of two mechanisms:

(i) The Ba could have been introduced by a juvenile fluid, its concentration declining away from the outlet(s) of the fluid.
The Be may have been derived by leaching of the surrounding volcanics, being concentrated in the most altered zones.

6.5 MECHANISM OF ORE DEPOSITION

To answer the above question is equivalent to determining whether the alteration was a prograde process, the albite, carbonate, sericite and silicification zones being formed in that order, or a retrograde process with the outer alteration zones overlapping the inner zones possibly due to mixing of the active hydrothermal fluid with relatively dilute and alkaline connate or meteoric waters. The question is obviously an important one, for if the former process is favoured it introduces the possibility that at least a part of the iron and copper in the deposits may have been derived during leaching and alteration of the volcanics.

Intergrowths of sericite-carbonate (37917) and carbonate-albite (37876) have already been noted near the boundaries of the alteration zones, but these merely reflect the increasing pH environment away from the mineralized zones and do not favour either mechanism.

The occurrence of separate patches and zones of sericite and carbonate (37909, 37911) is more significant. As discussed previously (4.3) the sericite occurs as bands in the schist whereas the carbonate is largely enclosed within the siliceous gangue.

The following relations may exist between the sericite and carbonate:

(i) Carbonate and sericite formed in equilibrium.

(ii) Replacement of carbonate by sericite, or of sericite by carbonate.
(iii) Formation of carbonate and sericite at different times; replacement not involved.

Hypotheses (i) and (ii) are not favoured as in both intergrowths or at least mutual boundaries between carbonate and sericite would be expected. The most probable explanation is that carbonate formation preceded introduction of sericite, the carbonate being protected from the later acidic solutions responsible for sericite formation by its protective mantle of quartz. Thus the most feasible, but certainly not established, mechanism of alteration involved a prograde series from albitization, in the least extreme, to silicification, in the most extreme case.

6.6 ESTIMATION OF CONDITIONS OF ORE DEPOSITION

From the preceding sections on hydrothermal alteration and ore deposits it may be seen that the dominant minerals present immediately after mineralization were quartz, sericite, chlorite, pyrite, chalcopyrite, carbonates and magnetite. The lack of a reliable geothermometer for the Lyell ores makes any speculation dangerous as to the quantitative estimate of ore-forming conditions, but the lack of feldspars in the assemblage indicates temperatures in excess of approximately 200°C (A. Folku, quoted by Smirnov, 1970) and according to Smirnov (op. cit.), quartz sericite rocks are commonly formed in the range 200 to 300°C whereas propylitized (chloritized) rocks form at about 250°C. The presence of enargite may be taken to indicate temperatures in excess of 320°C (Skinner, 1960) at North Lyell and the Blow, but this geothermometer has not yet undergone rigorous testing
and the effects of other components (e.g. iron) on the system have not been evaluated. As stated previously, it is believed that the occurrence of pyrophyllite at North Lyell and Comstock has little or no geothermometric significance.

A temperature of approximately $300^\circ$C thus appears reasonable for the formation of the Cape Horn ores, and is in accord with present-day geothermal systems, such as the Salton Sea system (Helgeson, 1967) and the WaioTapu area, New Zealand (Weissberg, 1969).

Further constraints may be now placed upon the system now that a temperature has been selected. 150 ppm total sulphur as sulphate in solution is a reasonable assumed value and is in accordance with reported concentrations in New Zealand geothermal systems at temperatures near $300^\circ$C (Weissberg, 1969) but is considerably in excess of that in the Salton Sea system at the same temperature (White et al., 1963).

A chloride concentration of 4 m is further assumed in accordance with the Salton Sea system (White, 1968). The K/Na ratios of hydrothermal waters is a function of temperature; K/Na of a fluid at $300^\circ$C is 0.12 (White op. cit.). Furthermore the activity ratio of potassium to hydrogen, $aK^+/aH^+$, for solutions in equilibrium with muscovite and quartz at $300^\circ$C is approximately 3 (Hemley and Jones, 1964). From these figures the pH of the fluid phase is $3.3 \pm 1$.

Assuming magnetite and not haematite was in equilibrium with the sulphide assemblage initially oxygen and sulphur fugacities of the order of $10^{-32}$ and $10^{-10}$ respectively would be expected (interpolation of results of Raymahashay and Holland, 1969), although the above
figures could vary somewhat since the presence of three phases in the system Fe-O-S is required to fix both fO₂ and fS₂ precisely.

An fO₂ of 10⁻³² implies a ratio of sulphate to sulphide species in solution of 0.1 (Helgeson, 1969) and thus a concentration of 15 to 20 ppm or 2 x 10⁻⁴ m sulphate would be expected in solution. Results by Malinin et al. (1969) on the solubility of barite in chloride solutions show that for 4 m alkali chloride solutions at 300°C the solubility product of barite is approximately 10⁻⁶. For the sulphate concentration shown above up to 650 ppm of Ba²⁺ could be carried in solution without barite being precipitated.

The above calculation is not intended to be quantitative but it shows how under reasonable conditions of alteration more than sufficient barium may be transported in solution to account for the observed enrichment in the sericite. In the North Lyell and Comstock ore bodies the oxygen fugacity was higher and thus there was a higher sulphate to sulphide ratio in solution. This is compatible with both the occurrence of barite as a gangue mineral and the relatively high metal to sulphur ratios present in these deposits.

A further note is necessary regarding the source of sulphur in the deposits. Studies on other disseminated sulphide deposits associated with volcanic rocks in geosynclinal environments have indicated low standard deviation of δS³⁴ values within individual deposits as at Mt. Lyell (Solomon et al., 1969), but variable mean S³⁴ enrichment between deposits. The New Zealand geothermal areas are characteristic of this type and show a similar narrow range of δS³⁴ values which indicates a magmatic origin for the sulphur, the
parent magma possibly being derived from the upper crust (Steiner and Rafter, 1966). However in a reappraisal of the systems Ellis (1967) considers that the hydrothermal waters are derived by circulation of meteoric waters in a convective cycle in which a few percent of the dissolved ions are incorporated from the magmatic source as has also been suggested by Smirnov (1970). This provides the heat for the alteration reactions that are responsible for most of the material in solution. Time for such processes in natural systems is ample as carbon isotope studies at Steamboat Springs, Nevada (Craig, 1962) showed that descending meteoric water spend between 30,000 and 300,000 years underground.

In such a long-lived system the problem of sulphide supply becomes trivial. By thermodynamic calculations Helgeson (1969) demonstrates that with increasing temperature the equilibrium ratio of sulphide to sulphate content in solution increases even at constant oxygen fugacity. Time is a critical factor in the process since the sulphate ion is notoriously sluggish in oxidation-reduction reactions (Barton, 1967).

Sufficient time is also important for another reason. For such reactions to be allowed to proceed, prolonged periods of relative quiescence of eruptive volcanic activity would be required. This is consistent with the observation that mineralization genetically and spatially related to volcanics took place either at the end of volcanism or at a pause during cycles of eruption. Examples of such a situation are worldwide and include such examples as the Rouyn-Noranda Area, Canada (Goodwin, 1965), Cyprus (Vokes, 1964), Rio Tinto,
Spain (Williams, 1966, 1964) and the Japanese "Kuroko" deposits (Horikoshi, 1969).

A surface origin for the magmatic fluid is also an attractive hypothesis for another reason. West of Cape Horn and thus with increasing depth in the volcanic pile, the intensity of alteration fades away, first carbonates and then albite appearing in abundance. No zones of sericitic alteration or silicification within these fringe zone assemblages are present indicating a lack of ore channels from depth. The same argument holds for the West Lyell ore body.

The mechanism outlined above in only tentatively proposed, but appears to be more acceptable than alternate hypotheses. Mineralizing juvenile waters being exsolved, or connate waters absorbed at depth and later being exsolved from a magma to produce an ore-forming solution (Kennedy, 1955) is a feasible process in relatively deep-seated granitic rocks but cannot occur in a near surface environment as Burnham (1967) has demonstrated that a granodioritic magma would boil off all its water content in rising from 10 kilometers to 3 kilometers in depth.

Syntaphral and lateral sliding of the shallower mineralized volcanics over the sub-strata (Schermerhorn, 1970) until the ore-bearing rocks overlie barren rocks is incompatible with the situation at Mt. Lyell because:

(i) A "root zone" of altered volcanics with transgressive epigenetic sulphide feeder channels has not been found.

(ii) No stratigraphic "break" occurs below the ore zone.
The conditions of ore formation in the Lyell Comstock and Tasman Shaft areas will not be discussed in the same detail as with Cape Horn but the following features deserve special mention:

(i) The local instances of attack on the wall rock by strongly acid solutions to produce silicification and advanced argillic alteration are probably connected with the local oxidation of relatively alkaline solutions, that is those initially in equilibrium with sericite. Such solutions would possibly be similar to those responsible for Cape Horn mineralization in having a sulphide to sulphate ratio of approximately 10 to 1. Oxidation in local fracture zones open to the surface would produce ionic H_2SO_4-rich solutions in contrast to the feebly dissociated H_2S-bearing solutions initially present. This hypothesis is consistent with both the rapid fluctuations in alteration assemblages in the Comstock area and the presence of barite in silicified rocks.

(ii) It has been argued, mainly on textural grounds, that deposition of these higher level ores largely occurred in gel form. The use of colloform textures as an indicator of this mechanism of formation, in the sense of Markham (1968), Edwards (1954) and earlier workers has been strongly criticized by Roedder (1968), for example. These textures are common in ores of volcanic affinities especially in those of probable shallow sub-surface or exhalative origin (e.g. Smirnov, 1970).

Regardless of whether or not a colloidal stage in ore development is implied, most workers, including Roedder (op. cit.) believe that the texture is an indicator of rapid sulphide deposition
which is also to be expected in a surface environment. It should be noted that the evidence of a gel state in the Mt. Lyell instance is not only based on colloform textures, but also on the presence of grains of quartz aggregate with relict syneresis (?)-type cracks and jasper fragments in breccias largely derived from underlying exhalative and replacement "cherts".

(iii) Massive banded pyritic deposits such as those of the Iron Blow, Comstock and the Tasman Shaft are believed to be products of submarine exhalative mineralization (e.g. Kinkel, 1966) and this is consistent with the presence of overlying limestones at Comstock.

The stratigraphically lower members of the volcanic succession are believed to be ignimbrites indicating subaerial extrusion. If this is correct then marine transgression must have occurred prior to formation of the Blow-type ores. This event cannot be precisely located in the stratigraphic sequence, but may have been a factor of some significance in sulphide development.
CONCLUSIONS

The host rocks for sulphide mineralization in the Cape Horn-Lyell Comstock area include ignimbrites (?) and minor siltstones overlain by lithic tuffs, spilitic to quartz keratophyric lavas and cherts.

Ore deposits formed at different stratigraphic levels are texturally, mineralogically and chemically distinct. The Cape Horn disseminated chalcopyrite-pyrite mineralization occurs in the ignimbrites towards the base of the established succession and is mineralogically simple. The Lyell Comstock ores are complex and phases such as bornite, chalcocite, betekhtinite, mawsonite and stromeyerite appear. Relict colloform textures are apparent in the pyrite in this ore. On the basis of texture and mineralogy the mineralized zones of Cape Horn and West Lyell are tentatively correlated as are the North Lyell-Crown Lyell and Comstock ores.

The sphalerite-chalcopyrite-galena mineralization in the chert at Comstock is transitional in type between the cupriferous mineralization and the overlying banded pyrite-galena-sphalerite ores of the Tasman Shaft.

A correlation exists between the stratigraphic level of the mineralization and the trace element content of the sulphides. Ni being progressively depleted and Se enriched towards the top of the volcanics. Also Zn, Pb and Ag are enriched in the higher level ores in accord with the classical paragenetic sequence.

A zonation of hydrothermal alteration exists with albite, carbonate, sericite and "chert" rich assemblages dominating, reflecting
a respective increase in the pH of the hydrothermal fluid. The ore zones of Cape Horn and Lyell Comstock represent two focii in intensity of the alteration. Geochemical studies indicate that at least some of the Cape Horn ores represent an addition of sulphide to the host rock. Mineralization probably was contemporaneous with ore deposition.

The problem of the mechanism of ore emplacement remains unsolved at this stage. It is tentatively suggested, however, that meteoric or sea-water rather than juvenile ascending solutions played an important role in ore deposition.

A Cambrian age for the bulk of the mineralization, is advocated on the following grounds:

(i) The ores show a well-defined zonation with respect to their stratigraphic position.
(ii) All ores show evidence of deformation.
(iii) Clastic sediments overlying the ores at Lyell Comstock contain albite as well as carbonate, sericite and chert pseudomorphs after plagioclase, and detrital sulphides, suggesting hydrothermal alteration and sulphide deposition preceded sedimentation.
(iv) In the same sequence as these sediments are fossiliferous limestones containing Dorsipyge.
(v) The limestone shows no sign of replacement and an overlying tuff, unlike the lower level volcanics, contains appreciable Ca but no carbonate. This is consistent with the interpretation that no major phase of hydrothermal alteration took place after the Cambrian.
Using a Phillips Vacuum X-ray Spectrograph (P.W.1540), seventeen rock samples were analysed for ten major oxides.

**Sample Preparation:** Each rock was crushed to minus one-eighth of an inch by a Sturtevant Roll-Type Jaw Crusher followed by grinding to 200 mesh powder with a Siebtechnik Laboratory Vibratory Disc-type Grinding Mill. Ignition losses were determined by heating approximately 0.5 gm. of sample to 1000°C for no less than half an hour. 0.28 gm. of sample was then used in the preparation of a fused disc with a borate mix using the method of Norrish and Hutton (1964).

**X-Ray Settings:** The settings are shown in Table 5.

**Absorption Corrections:** Absorption corrections were determined using Algol computer programme U727/3 which employs absorption factors determined by Norrish and Hutton (1969).

**Standards:** W1 and Tl were the standards used. Cu determinations were carried out using pressed, boracic acid-backed pills. Artificial standards were prepared using chalcopyrite-kaolinite mixtures to give copper concentrations of 300 ppm, 0.2%, 1% and 2%. Ba determinations were also performed using pressed pills. Standards were supplied by Mr. P. Robinson.

S was determined by a method developed by Mr. R.J. Ford, in which samples and standards were mixed with zinc in weight proportions of 1:3, powdered in the Siebtechnik Mill. Pressed pills were then prepared. Mr. P. Robinson supplied standards.
## Table 5

Instrument Settings for X.R.F. Analyses

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<th>Element</th>
<th>Tube</th>
<th>Crystal</th>
<th>Collimator</th>
<th>Voltage (kV)</th>
<th>Current (mA)</th>
<th>Counter E.H.T. Setting</th>
<th>Peak Angle (°)</th>
<th>Background (°)</th>
<th>Counter Channel Height (V)</th>
<th>Channel Width (V)</th>
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<td>20</td>
<td>C₆F₆</td>
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Water and Carbon Dioxide Analyses: H₂O and CO₂ were determined using Gee's (1970) modification of Riley's (1958) method. Samples were heated to 110°C for 30 minutes; the evolved H₂O and CO₂ being collected in tubes of magnesium perchlorite and "Ascarite" respectively.

Sources of Error: Of the 17 analyses, 10 lie within the acceptable range of 98%-102%. The possible sources of error in the analyses are numerous, due to the relatively large number of significant elements and the number of techniques employed.

Initially many analyses gave low totals. The main reason for this was a tacit assumption that all CO₂, S and H₂O were lost during ignition. It was found, however, that in most cases the sum of CO₂, S and H₂O contents exceeded the ignition loss. Hence correction factors had to be applied to all non-volatile elements determined using discs. The correction has the following form:

\[ \text{True } \% A = \frac{\text{Original } \% A \times \% \text{ sum of non-volatile elements} + \text{I.L.}}{\% \text{ sum of non-volatile elements} + H₂O + CO₂ + S} \]

where I.L. = originally determined ignition loss in \%.

\[ H₂O + CO₂ + S = \text{sum of percentages of } H₂O, CO₂ \text{ and } S. \]

The percentages of Fe were determined by a norm-type calculation and were not derived experimentally. From the determined Cu content the amount of S and Fe combined in chalcopyrite were calculated. The amount of Fe combined with the remainder of the S content to satisfy the requirements of pyrite was then determined. Hence the total amount of Fe in the sulphides was known. It was then a simple matter to determine the equivalent amount of Fe₂O₃ and subtract it from the amount determined initially.
APPENDIX B

ROCK SPECIMEN CATALOGUE

The majority of the following specimens are lodged in the collection of the Geology Department, University of Tasmania. A few specimens, particularly ores from the Lyell Comstock area, were borrowed from the Mt. Lyell Mining and Railway Company Museum, Queenstown. In these cases both the numbers in the museum catalogue and the University collection are given. The description given by the museum is reported (in quotation marks) together with any necessary amendment. In addition three thin sections belonging to the company were borrowed; these are distinguished in the text of this thesis by the prefix "ML".

All co-ordinates refer to the Mt. Lyell Mine Grid. Specimens from drill core are referred to in the following manner:—

NL 968/621 = D.D.H. NL 968, 621 feet.

R = hand specimen
T = thin section
P = polished section
Pt = polished thin section
D = X-ray disc and pressed pill
Z = pressed pill with zinc.
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<td>Ignimbrite (?)</td>
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<td>Sheared ignimbrite (?)</td>
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<td>37875</td>
<td>TR 5850W, 3405N</td>
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<td>37876</td>
<td>RT CH44/285</td>
<td>Altered siltstone</td>
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<td>37877</td>
<td>RTD 6475W, 6860N</td>
<td>Porphyritic felsite (ignimbrite ?)</td>
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<td>Unfoliated ignimbrite (?)</td>
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<td>) T W side Comstock</td>
<td>&quot;Chlorite feldspar schist&quot;</td>
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<td>quartz chlorite schist</td>
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<td>Pyritic quartz chlorite schist</td>
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<td>37883</td>
<td>RP NL 962/972</td>
<td>Haematite, magnetite, chalcopyrite and pyrite in quartz chlorite schist</td>
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<td>RP NL 962/980</td>
<td>Porphyritic quartz chlorite schist</td>
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<td>RT 3460W, 4205N</td>
<td>Quartz sericite schist with chalcopyrite</td>
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<td>37886</td>
<td>RT CH 46/285</td>
<td>Quartz sericite carbonate schist</td>
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<td>TDZ CH 2/444</td>
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<td>RPt 3880W, 2895N</td>
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<td>RT 6165W, 7500N</td>
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<td>Spilitic lava</td>
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<td>RT 3880W, 8230N</td>
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<td>37894</td>
<td>RT 1525W, 5845N</td>
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<td>RT C 50/1044½</td>
<td>Banded chert with pyrite</td>
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<td>R C 50/887½</td>
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<td>37896b</td>
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<td>37906</td>
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<td>Mudstone and sand stone (Jukes Conglomerate ?)</td>
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<td>Quartz chlorite haematite schist</td>
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<td>RT 2355W, 5446N</td>
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<td>37915</td>
<td>RT NL 975/286</td>
<td>Sericite quartz schist</td>
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<td>37916</td>
<td>RT 2778W, 5135N</td>
<td>Chlorite flecked sericite quartz schist.</td>
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<td>Specimen Number</td>
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<td>Rock Type</td>
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<td>37917</td>
<td>RTD NL 968/621</td>
<td>Calcite chlorite sericite schist (lithic tuff)</td>
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<td>37918</td>
<td>RT NL 969/379</td>
<td>Quartz sericite siderite chlorite schist</td>
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<td>RT NL 912/38</td>
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<td>37920</td>
<td>RT NL 968/1532</td>
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<td>37921</td>
<td>RT NL 968/1673</td>
<td>Green sericite</td>
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<td>RT CH 2/391</td>
<td>Quartz chlorite sericite schist with chalcopyrite</td>
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<td>37923</td>
<td>R 2520W, 5530N</td>
<td>Coarse pyrite in quartz</td>
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<td>37924</td>
<td>R NL 968/444</td>
<td>Calcite in quartz vein</td>
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<td>37925</td>
<td>R CH 44/1188</td>
<td>Quartz-chlorite-chalcopyrite in vein</td>
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<tr>
<td>37926</td>
<td>DZ Ch 2/329</td>
<td>Medium grained fragmental quartz, sericite, chlorite, carbonate schists with pyrite and chalcopyrite.</td>
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<td>37927</td>
<td>DZ CH 2/331</td>
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<td>37928</td>
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<td>37934</td>
<td>R 1185W, 5430N</td>
<td>Quartz sericite schist with barite.</td>
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**ORE SPECIMENS**

In describing ore specimens the following abbreviations have been used:

py = pyrite
cp = chalcopyritenb = bornite
dg = digenite
c = chalcoite
bnn = bournonite
cv = covellite
mt = magnetite
hm = haematite
rt = rutile
tn = tennantite
gn = galena
sl = sphalerite
apy = arsenopyrite
Au = gold
st = stromeyerite
mv = mawsonite
hx = hexastannite
mb = molybdenite
bt = betekhtinite
<table>
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<tr>
<th>Specimen Number</th>
<th>Locality</th>
<th>Description</th>
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<tbody>
<tr>
<td>101905</td>
<td>RP 2.5S C 56/707</td>
<td>Banded chert with py,cp,hx,tn, gn,sl</td>
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<tr>
<td>101906 (L180)</td>
<td>P 3/4 Comstock</td>
<td>&quot;Typical schist ore&quot; - cp,mt,hm in quartz chlorite schist</td>
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<tr>
<td>101907</td>
<td>× RP 1-29 CH 44/807</td>
<td>Banded hamt in schist with py and cpy</td>
</tr>
<tr>
<td>101908</td>
<td>RP 2.5S C 28/120</td>
<td>Quartz-sericite schist with cpy and py</td>
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<tr>
<td>101909</td>
<td>RP 7-41 CH 46/278</td>
<td>Cr,py</td>
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<td>101910</td>
<td>RP 7-42 CH 46/247</td>
<td>Euhedral py with apatite in quartz sericite schist</td>
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<tr>
<td>101911</td>
<td>RP 2.5S C 962/580</td>
<td>Recrystallized euhedral pyrite in quartz sericite schist</td>
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<tr>
<td>101912</td>
<td>RP 2.5S C 44/1016½</td>
<td>Euhedral py in quartz</td>
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<tr>
<td>101913</td>
<td>RP 2.5S NL 1011/104</td>
<td>Cplpy in quartz-sericite-chlorite-&quot;siderite&quot; schist</td>
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<td>101914</td>
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<td>Cplpy in quartz-sericite-chlorite-&quot;siderite&quot; schist</td>
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<tr>
<td>101915</td>
<td>RP 7-44 CH 44/1016½</td>
<td>Cplpy in quartz-sericite-chlorite-&quot;siderite&quot; schist</td>
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<tr>
<td>101916</td>
<td>RP 7-45 CH 15/397</td>
<td>Cp,py in quartz-sericite schist</td>
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<td>101917</td>
<td>RP 7-46 CH 2/417</td>
<td>Cp in sericite schist, Au and cp in quartz</td>
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<td>101918 (L33)</td>
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<td>&quot;Cp associated with mt&quot;. No mt - py,cp,py,mb in quartz sericite schist</td>
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<td>101919 (L184)</td>
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<td>&quot;Molybdenite coating in schist&quot;</td>
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<tr>
<td>101920 (L178)</td>
<td>P 3/4 No.4 stope, Comstock</td>
<td>&quot;Bornite&quot;-cc,dg,tn,sp,gn,p,rt in &quot;replacement&quot; (?) chert</td>
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<tr>
<td>101921 (L179)</td>
<td>P 1.2S No.1 open cut Comstock</td>
<td>&quot;Bornite&quot;-bn,cc,dg,tn,sp,gn,p,rt in &quot;replacement&quot; (?) chert</td>
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<td>101922</td>
<td>RP 3/4 C 58/557</td>
<td>Cp,tn,sp,gn,py,rt in &quot;replacement&quot; (?) chert</td>
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<td>101923</td>
<td>RP 3/4 C 58/558</td>
<td>Sl,gn,tn,cp,py in &quot;replacement&quot; (?) chert</td>
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<td>101924</td>
<td>P 3/4 5380W, 7475N</td>
<td>Py,tn in &quot;replacement&quot; (?) chert</td>
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<td>101925</td>
<td>RP 3/4 C 56/707</td>
<td>Mainly relict framboidal py in chert</td>
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<td>101926a,b</td>
<td>RP 3/4 NL 968/1732</td>
<td>Py,sp,gn,tn,cp,cc,sl,hx (?) in chert breccia</td>
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<tr>
<td>101927</td>
<td>RP 3/4 NL 1000/930</td>
<td>Cp,py, hm,tn in chert breccia</td>
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<td>101928</td>
<td>RP 3/4 860W, 5630N</td>
<td>Colloform py with gn,sp,tn,ap,sp,tn</td>
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<td>101929</td>
<td>R Taosan Shaft</td>
<td>Banded py,gn,sp,tn</td>
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<tr>
<td>101930</td>
<td>RP 3/4 904W, 430W, 5640N</td>
<td>Brecciated, banded py, gn,sp with chert</td>
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<td>101931</td>
<td>RP 3/4 dump; 430W, 5640N</td>
<td>Quartz-chlorite-cp in tension vein</td>
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<td>RP 3/4 CH 2/376</td>
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### THIN SECTIONS HELD BY THE MT. LYELL MINING AND RAILWAY CO. LTD.

<table>
<thead>
<tr>
<th>ML 54</th>
<th>3110W, 4625N</th>
<th>Flecked sericite chlorite quartz schist (similar to 37913).</th>
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<tr>
<td>ML 97</td>
<td>Cape Horn 1490 bench</td>
<td>Quartz chlorite schist.</td>
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### SPECIMENS REFERRED TO BY SOLOMON (1964)

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<tr>
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<td>31094</td>
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REFERENCES


As quoted by Loftus-Hills (1967).


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ADDITIONAL REFERENCES
