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MINERALISATION AT

MOUNT RANGITOTO AND

MOUNT GREENLAND,

WESTLAND
ABSTRACT

Gold- and silver-bearing veins located in Mine Creek, Mount Rangitoto and Cedar Creek, Mount Greenland have been worked sporadically since 1875, for small returns. The Mine Creek workings (1875 - 1904) yielded unknown but presumably small quantities of silver, and possibly gold, before operations ceased due primarily to poor recovery of the lower than expected silver grades encountered. The Cedar Creek workings yielded some 112kg (3,969oz) of gold from mineralised quartz veins, worked intermittently between 1885 and 1941, before operations finally ceased due to the poor financial position of the last company to work the ground, Mount Greenland Gold Limited, and their purchasing of an over-estimated ore reserve.

The Mine Creek veins consist of a quartz-pyrite-sphalerite-galena-minor chalcopyrite – electrum ± tourmaline ± calcite assemblage hosted in narrow shear zones in hornfelsed Greenland Group country rock. The mineralised shears which strike north-east and dip at 20 - 40° NW, are located near the crest of the Mine Creek Anticline, a horizontal or gently plunging upright close fold, but cross-cut both bedding and the axial surface of the fold at high angles.

Mineralisation of the shears probably occurred during the latest stages of emplacement of the strongly differentiated Triassic (c.214 Ma) aged Kakapotahi Granite. Magmatic mineralising fluids rich in Fe, Pb, Zn, Sb, As, Ag and Au, but poor in Cu, derived from the Kakapotahi Granite, passed upwards along the Mine Creek Fault propylitically altering the Rangitoto Granite adjacent to the fault, and into the mineralised veins. Sub-economic disseminated gold-silver mineralisation accompanied this alteration. The mineralising fluids were probably CO₂-rich with a salinity of approximately 18% NaCl equivalent, as indicated by fluid inclusion study.

Deposition of the ore minerals in the veins occurred as a fissure infilling process contemporaneous with fault movement, at estimated temperatures of 350 ° - 500°C and P H₂O pressures...
of 1-2 kbars. The observed paragenetic sequence pyrite - sphalerite - galena - chalcopyrite - electrum, in order of decreasing age, reflects deposition of the ore minerals under conditions of decreasing activities of sulphur and FeS, probably accompanied by decreasing temperature.

The major gold- and silver-bearing minerals, electrum and galena respectively, occur as replacements of earlier formed pyrite, a chemically favourable host for these minerals.

The Cedar Creek veins consist of up to a few per cent in total, of pyrite - chalcopyrite - arsenopyrite - sphalerite - galena - boulangerite - (jamesonite) - bournonite - tetrahedrite - gold hosted in a quartz ± calcite vein system up to 4m wide. Although the vein system located in the normally faulted crest of the William Tell Anticline is traceable for over a kilometre, known economic gold mineralisation is restricted to the ore shoot worked in the William Tell levels. The ore shoot which plunges steeply to the northwest along the strike of the vein and reputedly thins with depth, is probably located in a former zone of high permeability within the vein system, possibly due to increased fracturing.

Vein formation and mineralisation occurred penecontemporaneously with emplacement of the Rangitoto Granite into the previously folded overlying Greenland Group sediments, approximately 295 - 314 Ma ago. Magmatically derived mineralising fluids, metals and sulphur migrated upwards, probably along bedding planes, into the crest of the William Tell Anticline and thence upwards along the William Tell Fault, parallel to the axial plane of the anticline. Deposition of the ore minerals from the fluids occurred as fissure infillings associated with fault movement, at estimated temperatures of 230 - 400°C and maximum \( P_{H_2O} \) pressures of less than 2 kbars.

The observed paragenesis pyrite - arsenopyrite - chalcopyrite - sphalerite - galena - boulangerite - bournonite - tetrahedrite reflects ore deposition at successively lower
temperatures, the late stage sulphosalts being deposited at estimated temperatures of $230^\circ$ - $300^\circ$C. Changes in the composition of the mineralising fluids accompanied the decreasing temperature. The activity of sulphur decreased from approximately $10^{-12}$ atm - $10^{-4}$ atm during deposition, and was accompanied by the decreasing activity of PbS. Increases in the activity of Sb$_2$S$_3$ and Cu concentrations accompanied the above decreases resulting in the late stage paragenesis (galena - boulangerite - (jamesonite) - bournonite - tetrahedrite), observed.

Gold, the youngest of the ore minerals present, was probably deposited as a response to cooling of the mineralising fluids resulting in the decreased solubility of gold in the aqueous solutions.
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Surviving records of the Mount Rangitoto Silver Mining Company were generously made available by Mr I. Jacobs of Christchurch, and provided invaluable information on the early prospecting and mining ventures.

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CHAPTER I
INTRODUCTION

1.1 GENERAL INTRODUCTION

The Mount Greenland massif and the Rangitoto Range, separated by the Mikonui River, are situated some ten kilometres southeast of the former goldmining township of Ross (see Fig 1.1). The Mount Greenland massif hosts known gold-bearing quartz veins, whereas the Rangitoto Range to the south, contains both gold- and silver-bearing base metal quartz veins, and sub-economic disseminated gold-silver mineralisation.

Prospecting of the Rangitoto Range in the 1870's resulted in the discovery of gold- and silver-bearing base metal veins in Mine Creek. Prospecting and mining operations were carried out sporadically from 1875 to 1904, during which time a small amount of vein material was worked for its silver content.

Auriferous quartz was discovered in Cedar Creek, Mt Greenland, in 1885. Sporadic mining ensued and continued until 1941, although the quantity of stone extracted during this period was small.

Minor prospecting has been undertaken in both localities since the cessation of the mining operations.

No adequate investigation of the mineralisation in either Mine Creek or Cedar Creek has been made to date, and apart from brief observations made during the mining operations, and by Geological Survey staff, few details are known.

a) Aims of this Investigation

This investigation was undertaken with the following aims in mind:

i) To recount the mining history of these two deposits, and ascertain the likely reasons for the cessation of the mining operations.

ii) To describe the mineralogy and geochemistry of the vein assemblages, areas of disseminated mineralisation, and the host rocks surrounding the deposits.
Fig 1.1 Locality map. The distribution of exposed Paleozoic basement rocks is also shown.
iii) To determine the processes responsible for, and controls on, the mineralisation, particularly the sources of the economic minerals, and the conditions under which they were deposited.

It is intended that the rather general outline presented in this study will provide a base for the understanding of the geology surrounding, and mineralisation within, the study area. Such a base it is hoped, will be useful for both future studies of detailed aspects of the mineralisation, and in the possible exploration for similar deposits in nearby areas.

1.2 LOCATION

The study area, containing both the Mine Creek and Cedar Creek workings, covers an area of approximately 30km² situated some 10km southeast of Ross (see Fig 1.2).

The Cedar Creek workings on the southeastern slopes of Mount Greenland are accessible by four wheel drive vehicle from Ross - a distance of about 13km over a steep and rough track - or alternatively foot access may be readily gained from the Totara Valley Road.

The Mine Creek workings, some 6km to the southeast of the Cedar Creek workings, are accessible only by foot or helicopter.

1.3 GEOLOGICAL SETTING

The Westland foreland province consists of a thick sequence of alternating greywackes and argillites, intercepted as a turbite succession (Laird and Shelley, 1974). The sediments, of Cambrian-Ordovician age, were deposited in the Tasman Geosyncline, and have since been intruded by numerous granite bosses, ranging from upper Ordovician to Permian (Aronson, 1968; Adams, 1973), and perhaps Triassic (see chapter 6), in age.

A regional unconformity occurs above the Greenland Group, which was overlain by Cretaceous and Paleocene coal measures.
Fig 1.2. Map showing the location of the thesis area and the location of the Cedar Creek and Mine Creek workings.
followed by Eocene to Oligocene marine sandstones, mudstones and limestones. Uplift of the Greenland Group and intrusive basement rocks during the Rangitata and Kaikoura Orogenies has, however, resulted in the denudation of these deposits in many areas, their preservation being confined almost entirely to structural depressions.

Sediments of the Miocene to Pliocene Blue Bottom Group unconformably overly the Eocene to Oligocene marine deposits, and in places the Greenland Group, throughout the foreland province. Deformation of these sediments, the uppermost unit of which is the tectonically derived auriferous Old Man Gravels, occurred during the Kaikoura Orogeny, producing a series of NE-SW trending synclines and anticlines.

A third regional unconformity occurs above the Old Man Gravels, which are overlain by auriferous glacial and fluvioglacial deposits.

The Westland foreland province is separated from the allochthonous Torlesse Group in the east by the transcurrent Alpine Fault. Mylonites of the Fraser Formation are developed locally, to the west of the Alpine Fault, where they occur in fault contact with the less deformed sediments and intrusives of the foreland province. To the east of the Alpine Fault, a broad belt of schists is developed. These reach garnet-oligoclase zone adjacent to the fault, but decline in grade eastward to biotite and chlorite schists, finally giving way to the prehnite-pumpellyite metasediments of the Torlesse Group.

The proximity of the foreland province to an active plate boundary, possibly since Cambrian-Ordovician times, has resulted in a geological history dominated by tectonic activity. Several orogenic episodes are recognised, these being the Devonian-Carboniferous Tuhua Orogeny, the Cretaceous Rangitata Orogeny, and the late Tertiary
Kaikoura Orogeny. High temperature – low pressure metamorphism associated with granite intrusion is characteristic of the two older orogenies in Westland. Fault controlled uplift of the Greenland Group and intrusive basement rocks, and the overlying sediments, occurred during the Kaikoura Orogeny, and was accompanied by folding of the Cretaceous-Tertiary sediments into a series of broad synclines and anticlines.

1.4 PHYSIOGRAPHY AND CLIMATE

The Cedar Creek and Mine Creek workings (see Fig 1.2) are located at altitudes of 490m (1600 ft) and 600m (2,000 ft) respectively. In both areas slopes are steep, streams deeply incised and waterfalls common.

The lower slopes of both Mt Greenland (904m) and Mt Rangitoto (1125m) are covered with thick podocarp forest, which gives way above the workings in both areas, to sub-alpine scrub species. The dense vegetation cover makes travel away from formed tracks difficult.

Rainfall in the area is high, probably approaching 3,000mm per annum, being distributed fairly evenly throughout the year. Temperatures are generally low, averaging about 5°C in winter and 15°C in summer. During the winter frosts are common, and snow often lies thinly about the higher elevations.

1.5 PREVIOUS WORK

No detailed investigations of the geology or the mineralisation within the study area, have been previously undertaken. Descriptions of the mineralisation in Mine Creek, Mt Rangitoto, relating to the discovery and mining of the veins during the 1870's may be found in Reid (1886), Hector (1877, 1879), von Haast (1879), Cox (1877), and also in Reports on the Colonial Museum and Laboratory (1876, 1878), and in Reports on the Goldfields of New Zealand, between 1887 – c.1900. Many of these descriptions are brief, and devoted to discussion of some reportedly high assays of ore obtained during the early stages of prospecting.
Further descriptions of the vein mineralisation and the workings of the Mt Rangitoto Silver Mining Company, formed to work the ground, may be found in the Handbook of N.Z. Mines (1887, 1906), Mines Statements (1892, 1893), and Downey (1928).

Morgan (1908), in preparing the Geology of the Mikonui Subdivision (N.Z.G.S. Bull. No.6), visited the area briefly, but did not undertake any detailed geological studies.

Valuable information has been obtained from the surviving records of the Mt Rangitoto Silver Mining Company, which worked the ground in the 1870's. The accounts and records of the company are almost entirely devoted to describing the vein occurrences in a general way, the grade of the ore in the veins, and the operations of the Mt Rangitoto Silver Mining Company.

Newman (1977) gives a brief geological outline of the area.

The veins worked in Cedar Creek, Mt Greenland have also received very little detailed investigation. Numerous references to the veins and the workings in the area can be found in Mines Statements (1893 - ), Reports on the Goldfields of New Zealand (1885 - ), and also in Downey (1928). Details of the more recent mining operations and plans of the workings in the area were obtained from files held in the Mines Department, Greymouth. Morgan (1908) also briefly describes the geology and vein mineralisation in the Cedar Creek area.

In more recent times, Young (1964) recognised an assymetrical syncline (Cedar Syncline) - anticline (William Tell Anticline) pair in Cedar Creek, and noted that the main workings were in the crest of the anticline. Adamson (1966) further noted that the veins in the old workings were emplaced in a steeply dipping shear zone, probably by fissure infilling processes. Apart from making
these observations, neither Young nor Adamson attempted to describe further either the geology or the mineralisation of the area.

Mason (1955) noted the presence of bournonite in some samples of vein material, carrying free gold, from Cedar Creek.

The auriferous occurrences in the more celebrated Reef ton area, also in Greenland Group sediments, have been rather better, although still inadequately, investigated. Henderson (1917) provided the first accounts of the geology of the area, and this work was subsequently expanded upon by Gage (1948) who investigated the structure of the country rocks enclosing the auriferous quartz veins. Gage noted that the auriferous ore shoots were located in both the troughs of synclines, and less commonly but just as importantly, in the crests of anticlines. The veins were interpreted as being shear zone infillings, the dominant minerals present being quartz, calcite, gold, arsenopyrite, pyrite, stibnite, molybdenite, chalcopyrite and minor galena. Both Henderson (1917) and Gage (1948) considered the gold and associated minerals to be magmatically derived.

The sedimentology of the Greenland Group sediments has been described by Laird (1972), Laird and Shelley (1974) and to a lesser extent by Morgan (1908), Henderson (1917) and Suggate (1957). The sequence of alternating greywackes and argillites is considered to be a turbidite sequence, the clastic components being derived from the polycyclic weathering of an acid igneous terrain (Nathan, 1976). The sediments are quartz rich, containing only a few percent feldspar and rock fragments in a clay matrix. The mineralogy and petrology of the sediments has been described by workers such as Morgan (1908), Henderson (1917), Suggate (1957), Reed (1957), Laird (1972), Laird and Shelley (1974), Nathan (1976) and Mason (1978).

The generally unfossiliferous nature of the Greenland Group, has prompted several workers to undertake isotopic
dating studies on the group, the most notable being those of Aronson (1968), Adams (1973, 1975), and Adams et. al. (1975, 1979). These studies indicate that the Greenland Group was derived from source rocks at least 1200 Ma old. Deposition of the sediments occurred prior to 440 Ma ago, (Adams et. al. 1975) and possibly as much as 495 Ma ago (Adams, 1975), indicating Cambrian-Ordovician deposition. This age is supported by the discovery of graptolites in the Waitahu River area by Cooper (1975).

Granitoids of the Tuhua Group intrude the Greenland Group throughout much of its length. The age of these granitoids has also been investigated isotopically by the above workers, as well as by Mason (1961), Devereux et. al. (1968) and Tulloch (1979). These studies indicate that granite emplacement into the Greenland Group sediments of the Westland foreland occurred as several pulses. Late Ordovician - early Silvrian (420-440 Ma) emplacement of Tuhua Group granites has been noted by Adams (1973). A second pulse of intrusive activity, known as the Tuhua Orogeny, occurred during the Devonian-Carboniferous (300-370 Ma) (Aronson, 1968, Adams et. al. 1975) and was succeeded by a further pulse of intrusive activity, particularly in northern Westland and Nelson, of Jurassic-Cretaceous age, known as the Rangitata Orogeny. A significant number of Triassic aged intrusive rocks have also been located in Westland: Charleston 212 Ma (Aronson, 1968); Buller River 222 Ma (Eggers and Adams, 1979); Victoria Range (Tulloch, 1979). Further south similar aged intrusives have been located on Bounty Island 188 Ma, Ruapuke Island 211-217 Ma and at Pahia Point 188 Ma by Devereux et. al. (1968), suggesting the possible existence of widespread plutonism during this period.

Very little detailed mapping has been undertaken in the Tuhua Group granitoids, and consequently field relationships are obscure, and petrological and geochemical data are scarce, particularly in south Westland. Tulloch (1973)
described the petrology of the Barrytown stock, Hamill (1972) described the petrology of the western Hohonu Range, and Tulloch (1979) the petrology and geochemistry of the Victoria Range granites. These studies indicate that the granitoids are of calc-alkaline affinity and were emplaced, often at relatively high levels (2-4km), to the west of a subduction zone. Many of the granites show S-type characteristics (following White and Chappell, 1974) and may have originated by partial melting of sialic crust above the subduction zone.

Although these studies of both the Greenland and Tuhua Groups have helped to elucidate considerably the Paleozoic history of the western foreland, very little attention has been given to relating this history to ore genesis. This is rather surprising in view of the considerable quantities of gold won from the Greenland Group in New Zealand, and the related sediments of the Tasman Geosyncline in eastern Australia (e.g. the famous Bendigo reefs).
2.1 MOUNT RANGITOTO

Gold and silver-bearing base metal veins were discovered in Mine Creek by three prospectors - Messrs James Bevan, James Palmer and Edwin Kenway - in about 1875. Samples of ore, almost undoubtedly hand picked, were collected and forwarded to Melbourne for assay, two samples yielding 735oz (22.86kg) and 392oz (12.15kg) of silver, per ton of galena.

These results encouraged the prospectors greatly and on July 16, 1877, the Mount Rangitoto Silver Mining Company, with a nominal capital of £30,000 in 6000 shares of £5 each was formed, to prospect and work the ground. Drives and shafts were put in on the lode and stone extracted and stockpiled awaiting the construction of a packhorse or dray track to the workings. Some details of the work undertaken and the nature of the lodes may be found in the Manager's Report and the Director's Report for 1877 (see overleaf).

During this period a good deal of surface prospecting was undertaken on the discovered lodes, however, this was generally in a haphazard and ill-planned manner. This lack of systematic prospecting and development of the mine resulted in the Mine Manager, Frederick Manton's services being dispensed with.

The new manager, one Carl Schonfelter, was appointed in September 1877, and immediately abandoned the existing workings. A new tunnel, approximately 100 feet (30 metres) below the existing workings was begun, and according to Schonfelter's annual report for 1879 was continued for at least 690 feet (211m), with adjoining cross-cuts, winzes and stopes, and tunnels at higher and lower levels.

Schonfelter was rewarded with the discovery of stronger and richer veins in the new workings. This, together with the expectation of the imminent discovery of a larger,
MANAGER'S REPORT.

Mount Ragitoto Silver Mine.

To the Directors of the Mount Ragitoto Silver Mining Company (Limited).

Gentlemen,—

I beg to furnish you with the following report of the Mount Ragitoto Silver Mine:—

The workings upon the Main Lode comprise four tunnels or adits, and eight cross drives, as shown in plan hereunto, measuring 640 feet; also, the lower drive, 400 feet in length, to cut the Lode at a lower level, drain the upper workings, and facilitate the extraction of ore.

The Lode varies in thickness from one inch to two feet, underlying to the North-east, 20 degrees, but rising at a much greater angle to the South-west, the ore in that direction being more solid. The vein is composed of Ferruginous Quarts, Auriferous Pyrites or Bismuthiferous Iron, and Argenticiferous Galena and Blende, taking its course through Granite, Trap and Slate.

The Prince of Wales Lode is situated about half a mile north-west from the Main Workings, the vein being about 6 feet in width, composed of Conglomerated Quartz, intermixed with Specular Iron and small veins of Galena and Iron Pyrites. The work done upon this part of the Lode consists of two drives, one being 600 feet in length intersecting the vein, and the other 80 feet, to strike the Lode at a lower level. The ore had greatly improved at the time the work was discontinued, but is worthy of further attention, owing to the auriferous and argentiferous character of the ore, which I have assayed from time to time.

Haune's Lode is only a few yards from the mouth of the Lower Main Tunnel, being a small vein of quartz intermixed with very rich Argenticiferous Galena and Auriferous Pyrites, cropping out a few feet on the surface; very little work has been done upon it, and no time should be lost in ascertaining its value.

Austin's Lode (about 50 yards east of the main workings), is two feet thick, composed of micaceous slate with patches of Mende, Galena, and Blende. I think it would be desirable to work upon it, as there is every appearance of the existence of good, payable lodes.

The Copper Lode lies about a quarter of a mile North-east of the main workings, and is composed of a number of veins of quartz; floor spar with alternate bands of slate; stains of green carbonate; black oxide and yellow sulphurates of Copper, of very low percentage. As this Lode has not been opened, the ore may prove of better quality when properly exposed.

Five Clay of a very superior quality exists at no great distance from here, which will be an essential benefit, should you determine to erect smelting furnaces at, or near the mine.

The Surface Work performed consists of clearing timber, cutting rock from land-slips, cutting ore-paddocks, cutting foundation for houses and tents, erecting blacksmith's shop, charcoal house, powder magazine, assay houses, store house, manager's house, cutting and forming pack tracks, &c.

I have made several assays of the ore, and find the results to vary considerably, the yields being as follows:—Samples of ore taken from No. 7 drive and separated from gangue, yielded 10/ozs. 18lbs. wt. of Silver and 30 per cent. of lead, to the ten of ore.

Ore, dwt. grs.

| Samples similarly treated and taken from Churches' shaft or winze, yielded | 340 | 9 | 9 |
| No. 2 drive, yielded | ... | ... | 128 | 0 | 0 |
| No. 9 drive, yielded | ... | ... | 12 | 10 | 0 |
| No. 1 drive, yielded | ... | ... | 28 | 17 | 0 |
| Prince of Wales drive, yielded | ... | ... | 28 | 0 | 0 |

And during the year I have obtained results varying from only a trace of Silver to as high as 600oos to the ton; but as the ore attains a greater depth it improves in its Argenticiferous nature. At times the Lode is considerably narrowed by broken and unsettled country, but I have reason to believe that when the drives are extended into more settled country, the ore will improve in quantity and no doubt in quality,

As the new Real now approaches completion, I await your instructions to prepare for stopping and driving the ore ready for sending to Hikurangi, and, in order to expedite this work, I recommend that tenders be called for supplying the necessary timber to be used in the stopes.

I notice that the Ferruginous Quartz, met with in very large quantities in the various drives, contains Gold, and I think it desirable when the ore is being cut away, that a few tons of the above be also forwarded for treatment, as I believe it to be of a payable nature.

I have the honour to be, Gentlemen,
Your obedient Servant,

FREDERICK J. MANTON.
Mount Rangitoto Silver Mining Company, (Limited).

TO THE SHAREHOLDERS IN THE ABOVE COMPANY.

The following Reports of the Mount Rangitoto Silver Mining Company (Limited), were adopted at the Ordinary General Meeting of Shareholders, held on Monday, July 30th, 1877.

DIRECTORS' REPORT.

GENTLEMEN—

The Directors in submitting this, their first report on the operations of the Company, desire to express their satisfaction with the progress that has been made in developing the mineral resources of the Company.

Although the works at the mine have been carried on under more than ordinary difficulties, owing to the necessity of the country, and the expense incidental to conveying material over a rough mountain track, still the utmost economy has been exercised throughout—contract labor having been retained to an extent, when practicable.

It is, however, gratifying to report that, the Government road up the valley of the "Wanganui" now approaches completion. The last section, of about a mile in length, to connect the Government road with the mine, has been undertaken by the Company, and tenders have been accepted for the work in question, which it is intended should be finished simultaneously with that of the Government.

It is confidently expected that uninterupted communication, by the new road, will be established by the end of September, which will enable the Company to transport the ore without further difficulty or delay. The distance from the mine to the Port of Hokitika by this route will be about 32 miles.

Your Directors recommend that shipments of ore should be made to Melbourne, or Sydney, and also to London, in order that the mineral may receive the most skilful treatment in its reduction, and subsequent despatch; and, further, that the results of one shipment may be a check upon the other.

This course is considered the more important owing to the fact of the discovery of the presence of gold in the country. The results of several scientific assays proving it to exist in an already payable degree, and generally in association with the argentiferous products of the mineral.

The best opinion as to the machinery necessary for the proper and skilful treatment of the ore, will, thus be obtained, so as to enable the Company to decide upon this important matter should it be deemed desirable to erect suitable reduction furnaces at the mine.

With regard to the prospects of the mine and the work done therein, you are referred to the Mining Manager's Report, which contains full particulars relating thereto.

For the Board of Directors,

JOSEPH CHURCHES,
Chairman.
payable ore-body, resulted in the installation of a three stamp, water driven crushing battery, amalgamating barrels and two small buddles for treating the ore. This plant however, proved unsatisfactory for the type of ore, and in an attempt to improve recovery a wrought iron roasting pan, approximately 4.9m long and 2.5m wide, was erected, and used to calcine the ore before crushing. Recovery, however, remained poor. Gordon (1886) in the Annual Report on the Goldfields of New Zealand states

"... it appears as though very little gold or silver could be got by this process."

Mr Gordon "... examined the tailings lying around the battery and was surprised to find them full of quicksilver; and on washing some in an old fry-pan I (Mr Gordon) collected about 20 lbs of quicksilver and from 6dwt to 8dwt of gold and silver."

During the period 1877 to 1882 a considerable number of assays of the ore were made, both for the company, and by individuals. The results of these assays are summarised in Table 2.1. Many of these assays, in spite of some of the large sample sizes quoted, probably reflect hand picking of the ore. Indeed, the initial values obtained by Kirkland, which led to the formation of the company to work the ground, were never approached in any of the subsequent analyses. Furthermore, it seems as though no special effort was made by the company, to save the gold present in the ore.

Prospecting and mining operations continued at the mine until 1882, by which time the capital of the Mount Rangitoto Silver Mining Co. had been expended. According to Reid (1886):

"the original lode was lost, although several other lodes were met with in the dirves carrying more lead and less silver."

Attempts to raise capital, in England, to enable the continued working of the ground were unsuccessful, and operations were discontinued.
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Table 2.1 Assays of Mt Rangitoto Ore

N.D. = not determined
* = values given are Ag per ton of galena

Sources
1. Company Records
3. Hector (1877), Cox (1877), von Haast (1879).
The 500 acres of freehold land surrounding the mine, that had been purchased from the Crown by the Company, was sold in 1889 to a syndicate of six men, most of whom had been shareholders in the original Mount Rangitoto Silver Mining Company. Work was continued at the mine, and stone was extracted from Schonfelter's workings and treated on site.

Prospecting of the area was also undertaken by the syndicate, known as the Mount Rangitoto Freehold, and according to reports an area of mineralised granite was located. Gordon (1893) in the Annual Reports on the Goldfields of New Zealand quotes the following extract from a letter from the prospectors (Messrs Pollock and Bevan) to the Minister of Mines:

"You may be aware that for some time past we have been prospecting a huge belt of conglomerate, or a sort of bastard granite, which crumbles away on exposure, and that we have found to be auriferous. We have had many trials from the outcrop, taken indiscriminately from an extensive surface. One trial of a ton by chlorination in Melbourne gave us 6dwt 12 gr gold, another by the Cassel process yielded 6dwt 14 gr. Our own fire assays from the face of the formation have yielded a variety of results - some poor, some exceedingly good, but encouraging from a prospectors point of view, owing to the fact that we seldom or never failed to get gold, although nothing is visible to the naked eye. On another occasion we sent five small lumps to Reefton, and Mr Fenton found gold in all of them, and recommended us to further prospect the discovery. In February last a tunnel was driven right into the formation. The tunnel was 4ft 6ins by 6ft 6ins. At every 3ft of driving 12 lb of stone was taken out and tried. The tunnel was driven for a distance of 40ft, and nine of these trials of 12 lb each were made, and in every case gold was obtained. In order to get a good all round average the result of the nine trials were cupelled and run into one button of gold with the gratifying result that it yielded 4dwt 4gr of gold, equal to 6dwt 16gr per ton. The process adopted was by quick-silver amalgamation of the ordinary character, with but very primitive appliances at command. This mode of dealing with the stone proves that the gold is free and does not require
either the Cassel process or chlorination to extract it; hence the probability of making such a huge mass pay well, for it must be borne in mind there are simply millions of tons of stuff to operate on, which could be broken down for a mere nominal cost, probably less than 1s. per ton. The gold is very fine and will require careful treatment, but recent amalgamation pans and intelligent battery management ought to be sufficiently understood to make the undertaking a success."

Subsequent assays, reported in a 1907 prospectus, yielded 5dwt 13gr (9g) and 6dwt 12gr (10g) of gold per ton, and 40z 3dwt (129g) and 10dwt 14gr (17g) of silver per ton.

The shareholders of the Mount Rangitoto Freehold attempted to float a company to provide the capital to further develop the existing mine, and to mine the mineralised granite. These attempts, however, met with no success, and although several options were exercised on the property, no mining of the granite was undertaken.

The ground was apparently abandoned by 1907, and subsequent attempts by the remaining shareholders, and later on their descendants, to induce capital to be invested in the property were unsuccessful. Since this time only brief examinations of the property have been made.

2.2 PROSPECTING AND MINING HISTORY - CEDAR CREEK, MOUNT GREENLAND

Auriferous quartz was first discovered in Cedar Creek in 1885 by John Gagliardi. Numerous leases were quickly taken up along the strike of the reef, the major claims being the William Tell, Swiss Republic, All Nations and Lanarch properties (see Fig 2.1). Very little prospecting was done before the properties were abandoned. However, in 1887 the William Tell Claim had some further work done on it, and according to Downey (1928):
Fig 2.1. Plan of the Cedar Creek workings showing the location of the drives and the thickness (in feet) and location of the veins encountered in the various crosscuts (after Young, 1964). Two alternative positions for the Swiss Republic claim are shown.
"... quartz showing free gold was met with in an adit level driven 57 feet below the outcrop of a reef exposed in the creek. This adit was driven for 225 feet on the course of the reef, about 200 feet of the distance being on solid stone said to have been 4 feet wide with gold fairly distributed through it. A winze was then sunk on the reef at 76 feet from the western end of the adit, but the stone became broken and finally disappeared a few feet down. Another adit was driven 85 feet lower than No.1, for a distance of 567 feet, where a rise was put up to connect with the bottom of the winze. Although the prospects were not very promising, the company that held the claim erected a battery at considerable expense, and started to work out the quartz above the upper adit."

The quantity of stone worked and the gold returned are, however, unknown.

About 200 feet below the No.2 adit another adit was constructed for a distance of 1,000 feet and according to the Annual Report on the Goldfields of New Zealand (1888):

"... a block of stone or reef was cut through at about 800 feet from the mouth, but there is very little gold to be found in it. Several small leaders have been cut through further in, which contain a little gold."

The reefs are all stated to be in the form of segregated veins.

The property was worked until 1890, when the company collapsed, and the battery was sold and transported to the Minerva Mine at Blackball.

To the west of the William Tell workings, on the Swiss Republic Claim, a reef of 3 feet width running parallel to the reef in the William Tell workings was discovered and stated to contain "a little gold" (Report on the Goldfields of New Zealand, 1888). An adit was driven on the reef for about 400 feet and a crosscut put out in a southerly direction for 240 feet. In the crosscut two small leaders and a reef 4 feet in width were encountered, all of which were said to carry gold. No further work was ever done on the claim.
To the east of the ground held by the William Tell Company, and simultaneously with the abovementioned operations, the All Nations and Lanarch Companies put out crosscuts of 610 feet and 1,000 feet respectively. Numerous small veins and leaders were intersected and a reef formation 6 feet wide in the All Nations crosscut and 14 feet wide in the Lanarch crosscut was encountered. No payable values were found in any of the veins or leaders encountered.

The efforts of all of these companies were paralysed by the collapse of the William Tell Company and the sale of the battery in 1890, and the ground lay idle until 1896 when the Alpha Special Claim Syndicate took up all the old claims.

A little prospecting was carried out and a drive of 223 feet was put in on an outcrop of quartz carrying gold. One ton of the stone was crushed for a return of 4oz 2dwt gold. However, as the reef was driven on the stone became very poor, the syndicate became discouraged and the ground once again abandoned.

In 1913 the ground was once again taken up, by the Mount Greenland Gold-Quartz-Mining Company, and the old William Tell No.2 adit extended to the northeast and to the west. Nothing of value was found in the northeast extension but in the western extension a body of quartz about 8 feet wide was met and found to carry "very fair" gold, particularly on the hanging wall side.

This discovery provided sufficient impetus for the installation of a five stamp battery, and crushing commenced in 1917 and was continued sporadically until 1921. During this time 1,939 tons of quartz were crushed for a yield of 2,030ozs (62915g) of gold, a recovered grade of just over 1oz/ton (31 grams/tonne).

The lode was worked for a distance of about 60 ft when it was cut by a fault beyond which it could not be located. The payable portion of the lode was also found to quickly disappear on stoping upwards, although it did
continue underfoot. The company considered driving a low level adit a distance of 2,200 feet along the reef line to enable the remaining payable stone to be worked, and also in the hopes of finding new ore shoots. A subsidy from the Mines Department for this work was applied for, but rejected. A subsidy was however granted for -

"... continuing a cross-cut from the fault in our present reef to intercept another reef which is believed to be a continuation of the fault - the length required to drive in the cross-cut is approximately 150 to 200 feet."

Apparently no payable stone was located in the cross-cut, for the Company ran out of funds by the end of 1921 and operations ceased.

The Mount Greenland Extended Syndicate prospected ground adjoining the western edge of the Mount Greenland Gold-Quartz-Mining Company's ground from about 1919 to 1921. The prospectors state that they -

"... have found what we believe to be the continuation of their (the Mount Greenland Quartz-Gold-Mining Company) main reef and traced it practically right through the area from east to west. Surface prospects give colours of gold in the dish. The width varies from 5 to 8 or 9 feet. We have also traced a second reef parallel to the main reef, about 40 feet separating the two. The latter also varies from about 2 feet 6 inches to 4 feet."

A drive was commenced 200 feet vertically below the outcrop (the location of the drive is unknown to the author) and continued for over 350 feet. Reefs were encountered, but not found to contain payable values. The funds of the syndicate became exhausted by the end of 1921, and work ceased.

In 1929 the former William Tell property was taken up by Mr W.O. Bierwirth who formed a syndicate to prospect and work the ground. Bierwirth discovered that the reef stoped out by the Mount Greenland Gold-Quartz-Mining Company, and which varied from 4 to 6 feet in width, was only the
smaller portion of the whole reef. The larger portion, 8 to 9 feet wide, was parallel to, and separated from, the smaller by a narrow band of country rock.

Bierwirth also discovered that the reef which was apparently severed by a fault in the No.2 level was in fact "dragged" by the fault, and was present beyond the fault. The prospecting cross-cuts apparently pierced a portion of the reef track in which the reef existed above them, and probably below them also. Auriferous quartz from above the No.2 level was extracted by stoping. Some 2,735 tons of quartz was crushed by the company for a yield of 1,903ozs (59183g) a recovered grade of 0.70oz/ton (22 grams/tonne).

Bierwirths Gold Reefs Ltd worked the ground until May 1937.

The property was subsequently sold to a company called Greenland Gold Limited for £20,000. This new company considered that an ore block of approximately 25,000 tons with a grade greater than 0.50oz/ton (15 grams/tonne), remained above the No.2 level.

At considerable effort and expense a five stamp battery and two small cyanide tanks were hauled over the nine mile track from Ross and erected on the property. About 30 tons of concentrate and possibly a small amount of quartz was worked by the company before it got into financial difficulties.

Furthermore, it appears that the company was floated on misleading reports:

"It is now definitely proved that the thousands of tons of payable ore above No.2 level of William Tell does not exist ..."

and

"The only prospect worth considering in the mine is a lense of ore approximately 60 feet in length and an average width of four feet, situated at the west end of the No.2 William Tell. Values from samples taken by Mr E.J. Scoble (Inspector of Mines), if they are maintained at a lower level, should work out at one ounce per ton, with most of the gold recoverable over plates or strakes."

A request for assistance from the Mines Department to construct a drive to intersect the lens of ore below the William Tell No.2 level was declined due primarily to the poor financial position of the company.

The company, which had apparently been left with an almost worked out mine on its hands, could not raise sufficient capital to undertake the expensive prospecting operation and went into liquidation in 1941.

About 1970 the United Mining Company investigated the old workings and constructed a four wheel drive road to the site from Ross. The company's funds apparently became exhausted at this stage, for no work appears to have been done in developing the mine further.

2.3 REASONS FOR CEASING MINING OPERATIONS

2.3.1 Mount Rangitoto

The establishment of mining operations at Mount Rangitoto appears to have been based on a few exceptionally high assays (up to 735oz/ton) of silver from a galena-bearing vein. The early assays undoubtedly reflect the effects of hand-picking the ore, for subsequent assays undertaken by a variety of analysts, rarely exceeded 8-10ozs/ton of ore.

Selective mining of the galena-rich portions of the veins was continued in a badly planned and haphazard manner for a few years with generally disappointing results, the hopes of the company being based on the expectation of the veins to strengthen with depth. This proved to be the case, however the silver recovered by the company was apparently insufficient to meet expenses and operations ceased.

The grade and thickness of the veins exploited in Schonfelter's workings are unknown to the writer, the workings being at present inaccessible due to a fall in the access drive.
The method of ore treatment adopted by the company, viz. crushing followed by roasting and barrel amalgamation of the concentrates, was unsuitable for this type of ore, and resulted in poor recovery. This factor, possibly coupled with a low grade of ore, is thought to be primarily responsible for the cessation of the mining operation.

It must be stressed that Mount Rangitoto was only ever promoted as a silver mining venture, in spite of gold assays of up to several ounces per ton being reported. (A 90 ton shipment of ore is reported to have contained gold at the concentration of 2ozs/ton (62 grams/tonne). Why no attempt was made to recover gold from the mine is uncertain.

Subsequent reports of an auriferous 'granite formation' in the locality, reputedly containing 5-6dwt/ton (8-9 grams/tonne) and up to 4ozs/ton (124 grams/tonne) of silver, if true, should have been further investigated. Why this was, apparently, never done is unknown.

2.3.2 Cedar Creek, Mount Greenland

Mining operations ceased at Mount Greenland in about 1941, when Mount Greenland Gold Ltd went into liquidation. The company which initially had quite a large capital, expended most of this improving the track to the mine and installing plant. Records from the period 1938-41 suggest, however, that the amount of payably auriferous stone remaining in the mine, when purchased by the company, was very small, being restricted to a lens 60ft x 4ft (depth unknown) at the western end of the No.2 level. A report indicating the existence of 25,000 tons of then payable stone above the No.2 level appears to be exaggerated.

Attempts to raise capital to undertake development below the No.2 level and extract the remaining lens of payable ore known to exist, were unsuccessful, and operations ceased.
Several factors seem to have contributed significantly to the failure of the mining operation. Firstly, it appears that the success of the early operations, particularly those of the Mount Greenland Gold-Quartz-Mining Company (1913-21) and Bierwirths Gold Reefs Ltd (1929-37) was achieved largely by 'picking the eyes' out of the mine. These companies respectively crushed 1,939 and 2,735 tons of quartz for gold returns of 2,030ozs (62915g) and 1,903ozs (59183g), giving recovered grades of 1.0 (31 grams/tonne) and 0.7oz/ton (22 grams/tonne).

A contributing factor, with regard to the failure of the mining operation, appears to have been the (deliberately?) misleading report which resulted in the formation of Mount Greenland Gold Ltd, and their purchase of an over-estimated ore reserve. This undoubtedly put the company in a very bad financial position which ultimately caused its collapse. It also appears as though the directors of this company were not fully acquainted with the problems of quartz mining in such difficult country, and bad mine development by the previous company proved to be a severe disadvantage.

According to the records a lens of ore 60ft x 4ft (depth unknown) remains in and below the western end of the No.2 level. The possible existence of payable stone between the No.1 and No.2 levels should also be noted. According to the records the orebody which was 100 feet long in the No.1 level is only 60 feet long in the No.2 level. Whether or not the orebody continues to decrease in length below the No.2 level is unknown.
CHAPTER III

REGIONAL GEOLOGY

3.1 INTRODUCTION

Greywackes and argillites of the Greenland Group (including the Waiuta Group) are exposed discontinuously in a NE-SW trending belt west of the Alpine Fault, over 400km in length, extending from Milford Sound in the south to Karamea in the north. Granitoids of the Tuhua Group, intrude the Greenland Group sediments in many places, producing local and usually narrow contact metamorphic aureoles.

The Greenland Group sediments show little variation throughout their length, consisting of a uniform deep water turbidite sequence of greenish-grey greywacke and argillite, lacking calcareous and metavolcanic beds (Nathan, 1976). The thickness of the sequence is unknown, however, a minimum thickness of 5,200m was estimated in the Reefton area by Gage (1948).

Compositionally, the sediments are generally quartz-rich and are unfossiliferous, although a recent graptolite discovery (Cooper, 1975) indicates an early Ordovician age. This is supported by a Rb-Sr age of $495 \pm 11$Ma (Adams, 1975), inferred to represent the sedimentation of the group. Accordingly, the Greenland Group is considered to be contemporaneous with the Webb, Aorangi Mine and Patriach Formations, and the Mount Patriarch Limestone (Cooper, 1974) of Nelson area.

The similarity of the Greenland Group to the Ordovician sediments of south-eastern Australia has been noted by numerous workers (for example Benson and Keble, 1936; Cooper, 1975), and it is probable that the rocks of both areas were originally part of the same depositional basin, the Tasman Geosyncline, prior to rifting apart by the opening of the Tasman Sea in post-Cretaceous times.
Granitic intrusives of the Tuhua Group, which together with the Greenland Group make up the basement of the Westland foreland province, range in age from about 430Ma (Ordovician-Silurian) to about 280Ma (Carboniferous-Permian), although granites as young as 212Ma (Triassic) may also be present (for further discussion see Chapter 6). Radiometric dates obtained thus far suggest that a pulse of intrusive activity, known as the Tuhua Orogeny, took place between 280 and 370Ma ago.

Folding of the Greenland Group, and associated cleavage development, possibly by dewatering (Laird and Shelley, 1974; Shelley, 1975a), is thought to have occurred during the Ordovician (Shelley, 1975b), and is possibly related to intrusive activity. Folds are well developed throughout the Greenland Group, with gently plunging or horizontal upright folds with moderately to steeply dipping limbs, common. Bedding shows remarkable uniformity in strike, striking either NW-SE, or NNE-SSW in rocks of the redundant Waiuta Group (Suggate, 1957) included here in the Greenland Group.

3.2 SEDIMENTOLOGY, MINERALOGY AND PETROLOGY OF THE GREENLAND GROUP

3.2.1 Sedimentology

Laird (1972) and Laird and Shelley (1974) have described in some detail the sedimentology of the Greenland Group. The sequence described by them generally consists of alternating beds of sandstone and indurated mudstone, interpreted to be a turbidite succession. Sandstone beds are commonly graded with sharp basal contacts, show directional sole marks, load casts, flame structures, internal cross- and parallel-laminations, and top surface ripple marks. The turbidite sandstones are considered to have been derived from the east or south-east and deposited on a submarine fan complex.
3.2.2 Mineralogy and Petrology

The mineralogy and petrology of the Greenland Group has been described in varying detail by such workers as Morgan (1908), Henderson (1917), Reed (1957), Laird (1972), Laird and Shelley (1974), Nathan (1976) and Mason (1978). These studies indicate that the Greenland Group has a relatively uniform and distinctive modal composition. The sandstones are quartz-rich (35-50%) with very minor amounts (>6%) of sodic plagioclase and fine grained sedimentary and volcanic rock fragments (Nathan, 1976). Laird (1972) states that up to 90% of the sand-sized particles consist of quartz, with minor quantities of sodic plagioclase and rock fragments, the latter consisting of sandstone and mudstone, quartzite and representatives of a spilific volcanic suite.

Morgan (1908) states that "the greywackes consist mainly of angular fragments of quartz and various feldspars. Biolite is always present and although much is secondary some may be original." Epidote and minor calcite are also noted as being present in the groundmass.

Laird (1972) also notes the presence of micas (<1%), pyrite, and rare ferromagnesians in the sandstones of the Greenland Group. The composition of twelve samples selected for point counting by Laird (1972) are given in Table 3.1.

When plotted on a QFR diagram (see Fig 3.1) all of the samples plot near the quartz apex with most falling in the sublitharenite or subfeldsarenite fields of Folk et al. (1970).

The finer grained argillites or mudstones of the Greenland Group have been less well described than the sandstones. Morgan (1908) notes that they are often greenish-grey in colour with a well developed cleavage parallel to bedding.
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<td>4.9</td>
<td>3.5</td>
<td>3.2</td>
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</tbody>
</table>

**GREENLAND GROUP**
- 1 Fourteen Mile Bluff
- 2 Seventeen Mile Bluff
- 3 Mt Ryell
- 4 Headwaters of Punakaiki River
- 5 Headwaters of Moonlight Creek
- 6 Middle reaches of Moonlight Creek

**WAIUTA GROUP**
- 7 Slaty Creek, lower Buller River
- 8 Rahui, lower Buller River
- 9 Black's Point, Reefton
- 10 Black's Point, Reefton

**GREENLAND GROUP**
- 11 Mt Greenland, mine road
- 12 Mt Greenland, mine road

Sheet Numbers refer to N.Z. Geological Survey petrological record forms held in the Christchurch District Office.

Table 3.1 Point Counts of Samples of Greenland Group (including Waiuta Group) sandstones (from Laird (1972)).
Fig 3.1. Thin section analyses of Greenland and Waiuta Group samples plotted on a QFR diagram.

Q = quartz; F = feldspar; R = rock fragments.

All samples are quartz rich and are in or close to the subfeldsarenite and sublitharenite fields of Folk et al. (1970).

After Laird (1972).

Under the microscope he notes that "the argillites appear to be much altered, and show a fine-grained mosaic, consisting principally of amphibole, possible serpentinous material, and biotite. In the coarser bands, which approach grauwacke, quartz grains are plentiful, and a few fragments of feldspar may be seen." This description suggests a different composition from the sandstone beds, a view not held by Nathan (1976) who states that "the argillite beds are similar (to the sandstone) but contain a much higher proportion of recrystallised clay matrix."

It is assumed that quartz is the dominant clastic component.

3.3 GEOCHEMISTRY OF THE GREENLAND GROUP

Morgan (1908), Henderson (1917), and Nathan (1976) have all published geochemical analyses of Greenland Group sediments. Analyses from these authors are given in Tables 3.2 and 3.3.
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**TOTALS:** 100.03 100.33 99.80 100.49 100.32 100.23 100.54

1 Greywacke near junction of Boatman and Topfer Creeks
2 Micaceous greywacke (highly altered) from Waitaha sugarloaf
3 Micaceous greywacke from Flat Creek, Mount Rangitoto
4 Country rock, Blackwater Mine, altered by metasomatism
5 Argillite from Cedar Creek track
6 Argillite near junction of Boatman and Topfer Creeks
7 Country rock, Garden Gully, altered by metasomatism

No's. 2, 3 and 5 from Morgan (1908), p.101.

Table 3.2 Geochemistry of the Greenland Group (from Henderson (1917))
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<td>G</td>
<td>A</td>
<td>G</td>
<td>A</td>
<td>G</td>
<td>A</td>
<td>G</td>
<td>C1</td>
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Samples prefixed P held in the Petrology Collection, MGS, Lower Hutt
Sample prefixed UC is held in the Geology Dept, University of Canterbury
Analysis: S. Nathan.

G = greywacke
A = argillite
L1 = loss on ignition
G1 = average Greenland greywacke
G2 = average Greenland argillite
G3 = average bulk composition of Greenland Group
G/A = 1.9/1 (after Love, 1972)
W = Vederso's average worldwide slate (AWS)
Sample localities given in APPENDIX 1.

TABLE 3.3 Major and trace element composition of Greenland Group greywackes and argillites (from Nathan, 1976).
As Nathan (1976) concludes, the analyses show a uniform quartz or silica-rich sequence with $K_2O > Na_2O$, $CaO < 1.5\%$ and $Rb > Sr$. Mason (1978) has reported a Greenland Group sample from south Westland which does not, however, fit these constraints. This sample is apparently atypical, or alternatively not from the Greenland Group.

Nathan (1976) further notes the following trends:

1. An inverse linear relationship between $SiO_2$ and many elements e.g. $Al$, $Mg$, $Fe$ and transition trace elements, $K$ and alkali trace elements. This is interpreted as representing the diluent effect of quartz in an otherwise uniform matrix.

2. No clear relation exists between $SiO_2$ and $Na$, $Ca$, $Sr$, $Ta$, $Hf$ and the REE.

The petrological and geochemical data suggests that the Greenland Group is comprised largely of the two end products of weathering - quartz and clay minerals - of a terrane of granitic composition. More than one cycle of weathering of the source area appears likely to have been necessary to produce the quartz-rich, feldspar poor character noted.

### 3.4 GRANITOIDS OF THE TUHUA GROUP

Remarkably few detailed examinations have been made of the Tuhua Group intrusives since the pioneering work of Bell and Fraser (1906), although interest in the group appears to have been revived somewhat over the last couple of decades, possibly as a response to the availability of radiometric dating techniques. Since Mason (1961) published the first K-Ar dates for Tuhua Group granites, several other studies have been undertaken, notably those of Aronson (1965, 1968) and Adams (1973, 1975), the results of which are more appropriately discussed in Chapter 6. In summary, however, these works have shown that the Tuhua
Group ranges in age from 430 Ma - the Maybelle Bay and Windy Point Granites (Adams, 1973) - to possibly as young as 212 Ma (Aronson, 1968) for the Redjacket Granite (Nathan, 1975). A pulse of intrusive activity between about 280 Ma and 370 Ma was also noted by Aronson (1968) and correlated with the mid-Paleozoic Tuhua Orogeny.

3.4.1 Petrology and Geochemistry

Detailed petrological and geochemical studies of the Tuhua Group granites are insufficient to permit anything other than broad generalisations concerning the group. The granites are generally "alkali or calc-alkali, pale grey or less commonly pink, massive or porphyritic biotite-granites which may in places pass into granodiorite or syenite" (Reed, 1958).

Hamill (1972) investigated the granites of the western Hohonu Range where he recognised two varieties of granite, a pale grey sometimes feldspar porphyritic biotite-granite composed primarily of oligoclase, quartz, microcline and biotite (the Deutgam Granite), and a red, highly perthitic equigranular granite (the Brunner Granite). These varieties are interpreted as being of subsolvus and hypersolvus origin, respectively, and reflect the differing magmatic histories of the granite.

Tulloch (1973) studied the petrology of the Barrytown stock, north of Greymouth, and noted a similar mineralogy to that observed by Hamill (1972) for the Deutgam Granite in the Hohonu Range. According to Tulloch, the Barrytown stock was derived from partial melting of sialic material and emplaced as a mobile magma to a shallow depth (2-4km) at a temperature of about 750-800°C. Similar conditions of emplacement are inferred for Tuhua Group granites in the Hokitika District by Bell and Fraser (1906) who regarded the granites as "high level". Laird and Shelley (1974) suggest that, in the Reefton area at least, the granites were emplaced into wet, unconsolidated Greenland Group sediments.
A large number of geochemical analyses of the Tuhua Group granites have been made over the years, by numerous workers. Many of the early analyses, however, were for major elements only, thus limiting their petrogenetic usefulness. Hoggarth (1977) analysed both major and selected trace elements in a large number of samples, but drew few conclusions regarding the origin and conditions of emplacement of the granites analysed.

Tulloch (1979) combined petrographic and geochemical techniques with K-Ar age dating in a study of the Victoria Range granites. He concluded that the Paleozoic granites in this area were derived from the ultrametamorphism of the sialic crust above a subduction zone resulting in partial melting of the crust. The granite bodies mapped were generally S-type granites (after Chappell and White, 1974), although I-type granites, apparently derived by the partial melting of pre-existing igneous rocks with an intermediate SiO₂ content, were also noted.

3.5 LITHOLOGIC DESCRIPTIONS AND MAP UNITS USED IN THIS THESIS

During mapping of the Tuhua and Greenland Group rocks in the thesis area, it became apparent that further subdivision of both groups was desirable. Two distinct and easily recognisable varieties of granite, probably of differing ages (see Chapter 6), were observed on the southern flanks of Mount Rangitoto, and were mapped as such, whereas the degree of hornfelsing of the Greenland Group provided the basis for subdivision of that group. The following lithologic units are recognised.

1. Greenland Group Metasediments (gm)

Metasedimentary greywackes and argillites of the Greenland Group crop out extensively over the Mount Greenland massif. In the field the beds are generally well indurated and often cleaved, and are usually grey
to grey green in colour. The greywackes, which are dominant over the argillites, occur in beds generally less than 1.5 metres thick. The beds, which are commonly graded, always have sharp basal contacts, and usually overlie laminated or structureless argillite. Sedimentary muscovite is common in hand specimen.

The argillites, as noted, often contain either parallel or cross laminations and usually have gradational basal contacts with underlying greywacke beds, and sharp upper contacts with overlying beds.

Cleavage is pervasive in the muddy layers, but less well developed in the sandier greywackes.

2. Greenland Group Hornfelses (gh)

Hornfelses of the Greenland Group have only been located on the southern flanks of Mount Rangitoto where they are preserved as a roof pendant overlying the Rangitoto and Kakapotahi Granites. The pendant is comprised of knotted slates, spotted slates and greywackes, and locally schistose rocks. Recrystallisation is ubiquitous and has resulted in destruction of the slaty cleavage evident in the metasediments of Mount Greenland. Sedimentary structures are also commonly destroyed by recrystallisation, and bedding planes generally obscured.

In hand specimen the dominant textural features are the presence of spots of biotite, often up to a centimetre across, and commonly a metamorphic layering defined by the preferred orientation of the metamorphic biotite. These features are lacking in the metasediments of Mount Greenland.

Porphyroblasts are also present in some of the finer grained beds, giving rise to knotted slates, and locally, where crystallisation of biotite has produced a strong preferred orientation, sub-schists or linear schists may result.

3. Rangitoto Granite (rg)

The Rangitoto Granite crops out extensively in the thesis area, particularly on the northern slopes of the Rangitoto Range, and also above and west of the Mine Creek workings on the southern slopes.
The Rangitoto Granite is a grey, equigranular or slightly porphyritic biotite-granite, containing between 10 and 15% biotite as decussate aggregates, several millimetres across. Plagioclase (approx 40%) is the dominant felsic mineral with quartz and alkali feldspar accounting for 20 - 25% each of the rock. Partially assimilated xenoliths of Greenland Group hornfels are common.

In places the Rangitoto Granite has undergone significant hydrothermal alteration. Veins and veinlets of quartz and calcite are common in such cases, and the altered rocks are generally less resistant to erosion than their unaltered counterparts. Where this occurs the term altered granite, or altered Rangitoto Granite, is applied to the rocks in question. A fuller description of the alteration present is contained in Chapter 8.

4. Kakapotahi Granite (kg)

The Kakapotahi Granite crops out in the vicinity of Flat Creek, at the southern foot of Mount Rangitoto and extends southwards to the Purcell Range. The granite is a pale-grey or white leucocratic equigranular granite containing approximately equal amounts of quartz, plagioclase and K-feldspar. In contrast to the Rangitoto Granite, the Kakapotahi is biotite-poor consisting of less than 5% biotite, and in addition contains small amounts of muscovite and garnet, both visible in hand specimen. Xenoliths of Greenland Group sediments are absent.

The structural geology of the area is discussed in the following chapter, and the petrology and geochemistry of the rock types is discussed in Chapter 5.

5. Unmineralised Vein and Dyke Rocks

A variety of unmineralised vein and dyke rocks cut the hornfelsed pendant of Mount Rangitoto, and to a lesser extent the Rangitoto and Kakapotahi granites. Detailed investigation of these rocks was considered beyond the scope of this thesis, however, the following types were identified in the field, and mapped where appropriate:
a) Aplite Dykes

Aplite dykes, ranging from a few centimetres up to about 4m in thickness, are widespread throughout the hornfelsed pendant, and are occasionally seen cutting the Rangitoto Granite. The dykes, generally less than 50cm thick, appear in hand specimen to consist of a fine grained (<2mm) mixture of quartz and feldspar, and commonly contain up to a few per cent of muscovite and/or tourmaline, and less commonly traces of chlorite or biotite. Contacts with the surrounding country rocks are always sharp and often planar and wall rock alteration is generally absent. Where the aplite dykes contain quartz-tourmaline veinlets, however, tourmalinisation of the hornfelsed pendant adjacent to the aplite veins is common. Pegmatitic nodules are occasionally present in some of the larger dykes.

b) Pegmatite Dykes

Pegmatite dykes, consisting of a coarse grained assemblage of subhedral quartz, alkali feldspar and muscovite crystals, up to 1cm across, are considerably less common than their aplitic counterparts, and are confined almost entirely to the hornfelsed pendant. The dykes range in thickness from about 15cm to an estimated 20m but are generally less than 1m thick. Accessory tourmaline is present in many dykes, and wall rock contacts are always sharp, with little evidence of alteration.

c) Granitic Dykes

Fine grained granitic or micro-granitic dykes consisting almost entirely of quartz, alkali feldspar and muscovite, between 1mm and 5mm across, are also common in the hornfelsed pendant. The dykes generally average about 30cm in thickness, although one large dyke up to an estimated 50 metres across was observed.
Coarsely pegmatitic nodules, up to 45cm across, and commonly containing tourmaline, are present in this dyke. Again, wall rock contacts are sharp, and show few signs of alteration.

Compositionally, the above three classes of dyke appear identical, the major difference being in the grain-size of the dykes, which presumably reflects their cooling history.

d) Quartz Veins

i) Mount Rangitoto

Pneumatolytic and/or hydrothermal quartz veins are the most abundant of the vein and dyke rocks present. The veins are usually composed of pure quartz, although occasionally traces of muscovite, tourmaline and chlorite may be present. The veins range in thickness from less than 1cm to over 1m, but are usually less than 30cm thick. Wall rock contacts are sharp, and show little evidence of alteration. The quartz veins cut both the hornfelsed pendant and the Rangitoto Granite, but are best developed in the former. No visible sulphides or gold occur in any of these veins, the observed mineralisation being confined to the old mine workings in Mine Creek. A float boulder containing mineralised quartz veinlets was, however, located in the east branch of Mine Creek.

ii) Mount Greenland

Quartz veins exposed in the vicinity of the Cedar Creek workings show considerably more variation in thickness and form, than those of Mount Rangitoto. Descriptions of the Cedar Creek occurrences are given in section 10.1.2. Elsewhere on the Mount Greenland massif, quartz veinning is generally less well developed than in Cedar Creek, although Veronica Creek, Hope Creek and Montieth Creek (see Map) all contain significant vein systems. Bedded veins, occasionally massive
but most commonly consisting of stringers of quartz, are the dominant vein type, usually averaging between 50cm and 1 metre in thickness. These veins are usually best developed in the argillite beds. Veinlets a few centimetres thick cross-cutting bedding are also common throughout the area, as are quartz filled joints. Veinlets and irregular 'patches' of quartz occur ubiquitously in crush or fault zones, but only reach thicknesses of greater than about 1m in Cedar Creek, near the William Tell workings.

e) Alkaline Basic Dykes

Two fine grained analcimite dykes (U.C.9450) intruding the Rangitoto Granite were located in Churches Creek (G.R.564/363185). The two dykes are parallel and sub-adjacent, with thicknesses of 4cm and 1.5m. In thin section, the dykes are seen to consist of altered clinopyroxene and pseudomorphs of olivine. These phenocrysts are set in a groundmass of clinopyroxene prisms, in a fine grained mesostasis with analcime. Feldspars, micas and amphiboles are absent, reflecting a high degree of undersaturation and a lack of volatiles.

In Cedar Creek (G.R.557/440244) a 20cm analcime basalt dyke (U.C.9451), cross cutting bedding and segregated quartz veins, was located. Clinopyroxene (titanomagnesioaugite) and plagioclase (=An_{60}) occur as phenocrysts, and are set in a fine-grained groundmass of clinopyroxene, plagioclase, analcime and zeolites. These dykes, although not lamprophyres, almost certainly belong to the so-called Westland lamprophyre dyke swarms, and probably have a common origin, relating to Cretaceous rifting of the Tasman Sea.
CHAPTER IV

STRUCTURAL GEOLOGY

4.1 REGIONAL HISTORY

The structural geology of the Greenland Group sediments of the Westland foreland is, at present, poorly understood. The greywackes and argillites, of Cambrian-Ordovician age were deposited in the Tasman Geosyncline and are interpreted as turbidites (Laird, 1972; Laird and Shelley, 1974). Prior to lithification (Laird and Shelley, 1974; Shelley, 1975a) the wet, unconsolidated sediments underwent a regional metamorphism during which temperatures of about 400°C (Shelley, 1975a) were attained, at least locally. During this greenschist facies metamorphism, which was accompanied by extensive dewatering of the sediments and the development of a penetrative axial plane cleavage, the sediments were folded into a series of gently plunging upright folds (after Fleuty, 1964) with moderate to steeply dipping limbs.

The strike of both bedding and the axial surfaces of the folds is remarkably uniform, being predominantly NW-SE, although rocks of the now redundant Waiuta Group strike NNE-SSW, and have a similar structural trend. The reasons for this variation in strike are at present unknown, although Laird (1972), working in the Paparoa Range, attributed the variation to the effects of the Paparoa Tectonic Zone, a fault zone active since at least Cretaceous times. While this indeed may be the case in the Paparoa Range, it is difficult to envisage the movement along the Paparoa Tectonic Zone that would be required to produce a 60°-90° rotation in strike, in rocks some 30 kilometres or more (e.g. in the Reefton area) to the east of the Paparoa Tectonic Zone.

Shelley (1972; 1975a) and Laird and Shelley (1974), attributed the folding of the Greenland Group, north of Greymouth and in the Reefton area, to the gravitational
slumping of wet unconsolidated sediments off the flanks of rising granite bodies, probably during the Ordovician (Shelley, 1975b). In order to produce the simple structures and the uniformity in strike present, over a distance of 200 kilometres, or more, widespread granite intrusion would be required. No evidence for such widespread intrusive activity, during the Ordovician has been reported to date, although Adams (1973) obtained K-Ar dates for the Maybelle Bay, Windy Point and Barrytown stocks, ranging from 390 Ma to 425 Ma, indicating probable Ordovician-Silvrian emplacement of these stocks in the Westland foreland province. It seems likely, however, that folding of the Greenland Group occurred prior to emplacement of these granitoids. Adjacent to the Barrytown stock, for example, NW - SE trending folds approach the intrusive contact at right angles, and although the folds cannot be traced to the contact, they appear undisturbed by the intrusion, indicating that emplacement of the stock into an already folded sequence of sediments probably occurred.

Emplacement of the Tuhua Group granitoids into the regionally metamorphosed and folded Greenland Group probably began in the Late Ordovician (Adams, 1973) and continued sporadically throughout the Paleozoic, and possibly into the Triassic (see Section 6.3). Narrow contact aureoles surround most of the granitoids, which were probably emplaced at relatively high structural levels, at depths of about 2 - 7km (see for example Tulloch 1973, 1979; Hamill, 1972). A pulse of intrusive activity, the Tuhua Orogeny, occurred during the Carboniferous-Permian, but like earlier intrusions did not apparently disrupt the pre-existing structure of the Greenland Group, although contact aureoles were developed adjacent to the granitoids.

Following the Tuhua Orogeny, probably during the Cretaceous, gradual uplift of the Westland foreland occurred, and was almost certainly accompanied by block faulting. Overlying marine deposits, such as the Reefton Devonian, were almost completely destroyed by denudation. Non-marine sediments of the Pororari Group (Early Cretaceous),
including the Hawks Crag Breccia which contains abundant Greenland Group boulders, were deposited in fault controlled grabens on the Greenland Group basement, indicating that moderate relief, at least, was attained.

This regional uplift, which Sheppard et. al. (1975) suggest was New Zealand wide, was accompanied by the emplacement of Cretaceous intrusives, north of Greymouth, during the Rangitata Orogeny. No folding of the Greenland Group associated with the Rangitata Orogeny has been noted.

During the Late Cretaceous, c. 80 Ma ago, the New Zealand continent was rifted apart from Australia and Antarctica by the initiation of sea-floor spreading in the Tasman Sea and the South-West Pacific (Molnar et. al. 1975). Further block faulting may have accompanied this rifting in the foreland, however, there is little evidence for significant deformation of the foreland during this period. Subsidence of the foreland occurred during the early Tertiary with coal measures, overlain by transgressive marine sediments, deposited unconformably on the Paleozoic, or locally, the Cretaceous basement. Marine transgression continued until the end of the Oligocene, when regression apparently began, resulting in deposition of the unconformably overlying Miocene-Pliocene Blue Bottom Group. This regression is regarded by Warren (1967) as marking the beginning of the Kaikoura Orogeny in the foreland.

During the Kaikoura Orogeny, extensive block faulting of the Greenland Group and granitic basement rocks occurred along NE - SW trending faults, and resulted in the folding of the overlying Tertiary sediments into broad parallel trending synclines and anticlines (e.g. the Grey Valley Syncline and the Brunner Anticline). Uplift of the faulted Greenland Group horsts was considerable, giving rise to the present day Paparoa Range, the Mount Greenland massif, and probably many of the basement 'highs' now exposed west of the Alpine Fault.
In spite of the abundance of tectonic events affecting the foreland since Ordovician times, the Greenland Group has remained surprisingly undeformed. Since the pre-Tuhuan regional metamorphism and folding, probably during the late Ordovician, deformation of the group in the foreland has been almost entirely restricted to block faulting episodes, probably accompanied by gradual uplift, during the Cretaceous and the late Tertiary. The emplacement of the Tuhua Group and Cretaceous granitoids into the Greenland Group has had little or no effect on the earlier formed folds in the group, although hornfelsing of the sediments adjacent to the granitoids is ubiquitous.

4.2 MOUNT GREENLAND - STRUCTURAL GEOLOGY

4.2.1 Folding (see accompanying map and Cross-sections, map pocket)

Within the thesis area, north of the Mikonui River, the Greenland Group metasediments are folded into a series of gently plunging or horizontal upright folds (after Fleuty, 1964). A well developed axial plane cleavage (see Fig. 4.1) is associated with the folding, and is best developed in the argillite beds, where it is commonly so penetrative that it has destroyed the primary sedimentary structures. The cleavage is refracted between the argillite and greywacke beds, and combined with bedding and facing relationships, provides a useful tool for mapping the fold structures.

In Cedar Creek (see Map) Young (1964) recognised an assymmetrical syncline-anticline pair, and named the structures the Cedar syncline and the William Tell Anticline. These structures were also located by the writer, and mapped northwestwards along strike to Carley Creek. The Cedar Syncline lies c. 300m southwest of the William Tell Anticline, and strikes parallel to it.

In Cedar Creek the limbs of the Cedar Syncline dip, with minor variation, between 60° and 80°. The southern limb of the syncline in Cedar Creek dips at 50 - 60°, in
Fig 4.1. Greenland Group metasediments showing well developed cleavage in the argillite beds. Cleavage cross-cuts bedding at a high angle. Note also the presence of graded sand beds with parallel laminated mudstones at the top (right) of the beds.

Cedar Creek, Mount Greenland (G.R. 557/413243).
contrast to the steeper dip, 70 - 85°, of the northern limb. The hinge of the syncline is not exposed in Cedar Creek, but may, like the hinge of the adjacent William Tell Anticline, be faulted.

The William Tell Anticline, in Cedar Creek, has a more symmetrical form than the Cedar Syncline, both limbs in the hinge region, dipping between 65° and 85°. The dip of the northern limb of the Anticline, near the head of Cedar Creek, decreases northwards, flattening to about 40° as the Long Creek Syncline is approached. This decrease in dip reflects the more open nature of the fold structures to the north of the William Tell Anticline.

The forms of both the William Tell Anticline and the Cedar Syncline change northwestward along strike. In Swiss Creek, northwest of Cedar Creek, the William Tell Anticline tightens, becoming a tight fold (after Fleuty, 1964), the southern limb of which is overturned, dipping steeply (=80°) to the northeast. Although the hinge of the fold is not exposed, it seems likely that it is faulted, particularly in view of the faulting observed in the less tightly folded hinge of the anticline in the William Tell No. 2 Level. It is also apparent that the axial plane of the anticline, which is vertically dipping in the William Tell No. 2 Level, dips steeply northeastwards, probably at an angle of about 70°. The increased tightness of the anticline to the northwest, and the change in dip of the axial surface, is possibly due to either local variations in the intensity of the principal stresses during folding and/or to variations in the response of the sediments to these stresses.

The Cedar Syncline, northwest of Cedar Creek, remains assymmetrical in form, with the southwestern limb dipping at 50 - 70°. The northeastern limb of the syncline is overturned, however, and dips at about 80° northeastswards. The axial plane of the syncline, like the Mine Creek Anticline, must also dip steeply to the northwest. Tightening of the syncline appears to be less than that of the William Tell Anticline, the interlimb angle of about
30 - 40° present in Cedar Creek, persisting in Swiss Creek, although becoming tighter further northwest in Hope and Carley Creeks.

Northwestwards, along strike from Swiss Creek, the William Tell Anticline retains its tightened form with a steeply northeast dipping axial surface and overturned southwestern limb.

In Cedar Creek in the vicinity of the old mine workings, considerable variation in the dip of bedding was observed over distances of c.100m or less. These variations are thought to be caused by small folds developed between the Cedar Syncline and the William Tell Anticline, and for a distance of a few hundred metres north of the crest of the William Tell Anticline, before they apparently die out. These small folds cannot be traced for any great distance, and do not appear to be present northwest of Swiss Creek, or where the Cedar Syncline and the William Tell Anticline cross the lower reaches of Cedar Creek. Why these small folds cannot be traced for any great distance along strike is unclear, however, their close spatial relationship to the William Tell Anticline and the Cedar Syncline, and their apparent absence from the more open folds to the northeast, suggests that they have developed as a result of the more intense folding associated with the former.

Approximately 1200m to the northeast of the William Tell Anticline, and sub-parallel to it, the Long Creek Syncline is encountered. The fold is symmetrical or near symmetrical in form, both limbs dipping between 50° and 60°, defining a close fold (after Fleuty, 1964). The dip of the NW - SE striking axial surface must be steep (>65°), while the constancy of the strike of bedding throughout the area suggests a horizontal or shallowly plunging fold axis.

The Long Creek Syncline is paralleled approximately 1000m to the northeast by the symmetrical Eraser Creek anticline. Again neither the axial plane nor the fold axis of the fold have been observed, however, an upright axial surface and shallowly plunging or horizontal hinge, similar to that envisaged for the Long Creek Syncline seems most likely.
The folds developed on the Mount Greenland massif show a similar structural style to those reported elsewhere in the Greenland Group (see, for example Gage, 1948) and are interpreted as having formed prior to the Tuhua Orogeny, which saw the emplacement of the Rangitoto Granite immediately to the south of the Mikonui River. The Mine Creek Anticline in the hornfelsed pendant on the southern slopes of Mount Rangitoto (see Section 4.3.1) is considered to be of similar origin.

The fold style developed in the Greenland Group metasediments is simple, with gently plunging upright synclines and anticlines developed. Regionally, the folds have a wavelength of c.1100m, although in the vicinity of the former mine workings, the William Tell Anticline and the Cedar Syncline have a wavelength of c.300m, with smaller folds developed on the flanks. Why the wavelength of these two folds is so much less than the adjacent folds is unclear, however, local variations in the stress regime operating during folding, or variations in the response to this stress may have been important.

Folding apparently occurred by flexure slip processes, probably accompanied by extensive shearing along argillite beds at higher strain levels accompanying the development of tighter folds.

4.2.2 Faulting

Faults located in the Greenland Group metasediments can conveniently be divided into two sets; a well developed set striking parallel to bedding or the axial surfaces of the folds, and a less well developed set striking across bedding. Many of the faults cannot be traced for any distance, owing to the vegetated nature of the country, and the absence of suitable marker horizons makes determination of the sense and amount of movement on many of the faults impossible.
a) **Faults Striking Parallel or Sub-Parallel to Bedding or Axial Planes.**

Two major faults, both striking NW - SE were located in the Greenland Group metasediments of Mount Greenland. In Monteith Creek, near the northeastern boundary of the thesis area, a large fault - the Monteith Fault - was located. The fault is marked by a crush zone, which has in places been healed by recementation of the sediments, some 20 - 30m wide. The fault, which is upthrown to the northeast, dips steeply southwest (60-70°) and is normal in type. The direction of horizontal displacement, if any, is unknown.

The age of the fault is also unknown, however, its topographic expression and location at the top of the Totara River gorge suggest that recent movement on the fault related to the Kaikoura uplift of the Mount Greenland massif has occurred. Formation of the fault may, of course, have considerably pre-dated this latest movement. The William Tell Fault (see Geological Map) runs parallel to the crest and axial surface of the William Tell Anticline. The fault is well exposed in the William Tell No. 2 Level where a friable black pug, carrying numerous stringers of quartz containing pyrite, up to 1.5m thick occurs. This fault pug forms the southern wall of the vein system worked in the William Tell No. 2 Level (see Section 10.1).

Movement on the William Tell Fault was dominantly vertical, as indicated by faint slickensides in the pug (Adamson, 1966), and may have occurred as early as the Ordovician, during the regional folding episode. Sulphur isotope data (see Section 11.5) from sulphides in the adjacent quartz body suggest a magmatic origin for the sulphur in the veins, which coupled with the brecciated nature of the wallrock included in the vein quartz, indicates that movement also occurred during emplacement of the vein system, which is though to be related to intrusion of the Tuhuan-aged Rangitoto Granite in the late Carboniferous-Permian.

There is no good evidence for more recent movement on the William Tell Fault, and it thus seems likely that the
fault may have been inactive since the end of the Paleozoic.

Numerous small faults or shears parallel to and contained within the argillite beds are evident, however, no attempt has been made to systematically map these. These shears are usually best developed where the slaty cleavage is parallel or sub-parallel to bedding, although may also be present where the cleavage cuts bedding at higher angles. The age of these shears is unknown but the fractured nature of the argillite beds indicates that the movement occurred after lithification of the sediments. Occasional quartz stringers and veinlets, up to 15cm thick, commonly occur in the sheared argillite beds, although no evidence of mineralisation of these veinlets have been observed.

b) Faults Striking at an Angle to Bedding.

Several faults striking at an angle to bedding were located (see Geological Map). Most of these faults strike north or northeastwards and are steeply (60°-90°) dipping. Small crush or pug zones less than 75cm wide are characteristic, but occasionally these may reach thicknesses of 3 - 4m. Although the amount and sense of displacement on these faults cannot be determined, horizontal displacements, in general, are unlikely to be larger than some tens of metres. This is indicated by the existence of a fault in Swiss Creek which cuts the William Tell Anticline between Swiss Creek and Hope Creek, without noticeably displacing the Anticline.

A fault striking at c.290° and dipping vertically was also located in Swiss Creek, and strikes southeastwards towards the William Tell No. 2 Level. A similarly orientated fault striking 290°, and dipping 25° SW is also shown in the western end of the William Tell No. 2 Level where it cuts and displaces the vein system which is, according to records of the times, "dragged" over the fault. The fault located in Swiss Creek may be a continuation of this fault which displaces the vein system, or alternatively a genetically related fault. Both faults
are considered to post-date the mineralisation, and are therefore considered to be post-Paleozoic in age.

The Mount Greenland massif is probably separated from Rangitoto Range to the south by a large fault. Although not observed, the existence of the fault is indicated by the lack of hornfelsing of the Greenland Group metasediments, marked by the absence of biotite, immediately to the north of the Mikonui River. If the contact between the metasediments and the Rangitoto Granite were intrusive, rather than faulted, biotite would be present in the metasediments. Calculations by Jaeger (1957) indicate that for a granite body with a diameter of c.4km, estimated to be the minimum diameter of the Rangitoto Granite, and a temperature of 750°C emplaced at a depth of c.5km, temperatures in the country rock would exceed c.430°C within 2km of the pluton, and c.500°C within 1km. Assuming that biotite would form by reaction of chlorite and sericite in the matrix of the metasediments at temperatures above 400°C, biotite should occur in the metasediments, within about 2km from the Rangitoto Granite.

No biotite has been observed in the Greenland Group metasediments in lower Swiss, Cedar and Hope Creeks, at distances of approximately 1km from the Rangitoto Granite, and accordingly a fault contact between the Rangitoto Granite and the Greenland Group metasediments is indicated. The position of the contact (see Geological Map) is taken from Morgan (1908), with minor modification based on air photo interpretation, and has not been observed by the writer.

The fault, called here the Mikonui Fault, is downthrown probably by 1km or more, to the north. The amount and sense of horizontal displacement, if any, is however, unknown. Development of the fault probably occurred in pre-Cretaceous or Cretaceous times, and may have been related to a tensional tectonic regime during the Late Cretaceous.

Uplift of the Mount Greenland massif, north of the Mikonui Fault, since Pliocene times has been considerable and has probably also been accompanied by a similar amount of uplift of the Rangitoto Range. Movement on the Mikonui Fault may also have occurred at this time.
The relationship of the structures observed on the Mount Greenland massif, north of the Mikonui Fault, to those of the Rangitoto Range to the south are summarised in section 4.4.

4.3 MOUNT RANGITOTO

The structural geology of the Mine Creek area has not been investigated in any detail, the most comprehensive account to date being that of Morgan (1908), who described the general geology of the Mine Creek area.

Greenland Group hornfelses, exposed on the southern flanks of Mount Rangitoto, have been deformed into a single upright horizontal or gently plunging anticline (after Fleuty, 1964). Both bedding and the axial surface strike NW – SE, and bedding generally dips at angles greater than 40°. Numerous faults are evident in the field, although many cannot be traced for any great distance, owing to the nature of the country. Two predominant sets can, however, be recognised, one striking NNE – SSW, and another less well developed set striking NW – SE.

4.3.1 Folding (see Geological Map and Cross sections)

A single anticline, the Mine Creek Anticline was located on the southern slopes of Mount Rangitoto. The hinge of the fold has not been observed, however, a marked reversal in dip occurring at about the level of the old workings in Mine Creek, is sufficiently pronounced throughout the area to define the structure, with the aid of stereographic analysis.

A plot of 65 poles to bedding (see Fig 4.2) taken from the Greenland Group hornfelses, defines two maxima at about 030° and 210°, approximately equidistant from the centre. The pattern indicates a symmetrical or near symmetrical close fold (after Fleuty, 1964), with a steeply dipping or vertical axial plane, and a shallowly plunging, or horizontal, fold axis. The axial surface strikes NW – SE (c.300° – 120°), parallel to bedding.
Fig 4.2.  (a) Poles to bedding, Greenland Group hornfelses, Mount Rangitoto. Poles define a gently plunging or horizontal upright fold (see text for discussion).

(b) Poles to cleavage ($S_1$), Greenland Group hornfelses, Mount Rangitoto. Cleavage is sub-parallel to bedding (see text for discussion).

(c) Poles to joints, Mount Rangitoto. Joints parallel to bedding, and a steeply dipping NE-SW striking joint set are evident (see text for discussion).

(d) Poles to veins and dykes, Greenland Group hornfelses, Mount Rangitoto. Veins and dykes are generally sub-parallel or parallel to bedding or jointing (see text for discussion).
Both the fold style and orientation of the Mine Creek Anticline are identical to the observed structures on Mount Greenland (see section 4.2.1). This similarity suggests that the hornfelsed pendant preserved on Mount Rangitoto, has not been significantly disrupted by granite intrusion. Furthermore, the existence of the Mine Creek Anticline in the pendant suggests that the folds of both Mount Greenland and Mount Rangitoto pre-dated, perhaps by a considerable period of time, granite intrusion in the area.

Shelley (1972, 1975a) and Laird and Shelley (1974) have attributed folding of the Greenland Group north of Greymouth and in the Reefton area, to the gravitational slumping of wet unconsolidated sediments off the flanks of rising granite bodies. The hypothesis, however, does not readily explain observed structures in the thesis area.

Firstly, the constancy of strike of the Greenland Group sediments and fold axes both regionally and in the thesis area, suggests that the folds are related to a regional tectonic event rather than the intrusion of one or more granite bodies producing slump folds. Considerable variation in the strike of bedding and axial surfaces would be expected in slump folds surrounding intrusive bodies, however, such a variation is lacking in the Greenland Group sediments of the foreland.

Secondly, considerable variation in fold style with both vertical and horizontal proximity to intrusive pluton(s) would be expected if folding were due to gravitational slumping of wet sediments off the rising intrusive bodies. Such a variation is not evident in the Greenland Group, even though thicknesses of greater than 5,000m (Gage, 1948) are exposed in some localities. The striking similarity of the Mine Creek Anticline, preserved in a hornfelsed pendant, to the William Tell Anticline of Mount Greenland, which is at a higher structural level, as suggested by the lack of hornfelsing of the Greenland Group metasediments, provides further evidence for the uniformity of fold style over
both large areas and considerable thicknesses of sediment. This would not be predicted by models involving gravitational slumping.

Thirdly, the style of folding in the thesis area i.e. upright, horizontal or gently plunging, near symmetrical folds with moderate to steeply dipping limbs, is not particularly compatible with an origin by slumping associated with granite intrusion. Slump folds could reasonably be expected to have variably inclined axial surfaces, and a greater degree of asymmetry particularly at lower structural levels, than the folds observed.

Accordingly, it is considered that folding of the Greenland Group, in the thesis area at least, was not a consequence of gravitational slumping related to granite intrusion, but rather a part of a regional deformation prior to intrusion of the Rangitoto Granite, approximately 305 Ma ago (see Chapter 6). Although granite intrusion may have accompanied this deformation - the Maybelle Bay, Windy Point and Barrytown stocks give K-Ar ages of approximately 400 Ma (Adams, 1973) - no evidence for such intrusions producing folds in the Greenland Group, exists in the thesis area. The folds present, and the associated axial cleavages, are therefore considered to pre-date the intrusion of the Rangitoto Granite, and probably formed as a response to a regional compressive regime during the Ordovician.

4.3.2 Faulting

Although the traces of numerous faults were located in outcrop the nature of the country prevented most of these being traced any distance. Furthermore, the absence of suitable marker horizons makes determination of the amount and sense of displacement on most of the faults impossible.

The largest and most important fault in the area, the Mine Creek Fault (see Geological Map), controls the course
of the western branch of Mine Creek, and for much of its length marks the contact between the Rangitoto Granite and the hornfelsed pendant (see Fig 4.3). The fault, which is normal in character, strikes NNE - SSW, dips vertically or steeply to the east, and is upthrown to the west. The absence of the hornfelsed pendant overlying the Rangitoto Granite, up to the ridge crest west of the fault, suggests a minimum vertical displacement of about 300m to the west of the fault.

In the vicinity of the workings in Mine Creek, several sub-parallel splay faults are evident, and have resulted in considerable fracturing of the hornfelsed pendant between Schonfelter's and Manton's workings. Again, the displacement on the splays is unknown, but the faults are assumed to be normal in character.

The traces of numerous small faults were also located during mapping, however, none could be followed beyond the individual outcrops. Apart from the Mine Creek Fault and associated splays, two sets of faults could be distinguished.

a) Faults Parallel or Sub-parallel to Bedding

A group of seven fault traces striking NW - SE, parallel or sub-parallel to bedding, was located in the hornfelsed pendant. These faults, all marked by crush or pug zones less than 0.5m wide, are most commonly developed in the muddier beds, and have dips between 58° and 90°. The sense and amount of displacement is again unknown, however, the steep dips suggest that the faults are probably normal rather than reversed in character.

In Manton's Workings, one such fault cuts and displaces a mineralised vein, providing clear evidence that some, and possibly all, of these faults post date the vein mineralisation.

b) Faults Striking at a High Angle to Bedding

A number of faults striking NE - SW, at a high angle to bedding, were also located particularly in the eastern branch of Mine Creek. These faults are steeply dipping,
Fig 4.3. Mine Creek Fault, marking the contact between the Rangitoto Granite (left) and the Greenland Group hornfelses. Sense of movement is as indicated. Mine Creek, near Schonfelter's workings (G.R. S64/365197).

Fig 4.4. Small fault cross-cutting bedding in the surrounding Greenland Group hornfelses. Drag structures in the adjacent hornfelses indicate that the fault is a normal fault. Mine Creek, Mount Rangitoto. (G.R. S57/372202).
and drag structures in the adjacent sediments suggest, normal in character (see Fig 4.4), although the amount of displacement is unknown.

The age of the faults is difficult to determine, however, the brecciated nature of the crush zones suggests that the pendant rocks were cold and brittle during faulting. It seems probable, therefore, that these faults post-date granite intrusion in the area, and may be of the same age as those striking parallel to bedding.

None of the faults discussed above, excluding the Mine Creek Fault and some of the splays, shows any evidence of hydrothermal alteration or mineralisation. This suggests that if the faults pre-dated intrusion of the Rangitoto Granite they must have been impermeable to hydrothermal fluids. This situation would seem rather unlikely, and it is accordingly suggested that the faults mapped post-date both intrusion of the Rangitoto and Kakapotahei Granites, and the mineralisation. They are therefore considered to be post-Triassic in age and probably related to movement during either the Rangitata or Kaikoura Orogenies.

Rocks adjacent to the Mine Creek Fault, in contrast, show evidence of alteration and mineralisation. Accordingly, it is considered that this fault, and its splays, existed prior to cooling of the Kakapotahei Granite, and may have originated during that phase of intrusion. No evidence to suggest that significant movement on the Mine Creek Fault and splays has occurred since emplacement of the Kakapotahei Granite probably during the Triassic has been noted.

4.3.3 Development of Metamorphic Foliation

Well developed slaty cleavage and poorly developed schistosity, defined by the preferred orientation of metamorphic biotite, and less commonly muscovite, are widespread throughout the hornfelsed pendant. The foliation is developed parallel or sub-parallel to bedding, although may occasionally cut bedding at acute angles, up to about
20° (see Fig 4.5). Some shearing has occurred parallel to schistosity, disrupting quartz veins either along discrete planes or pervasively, producing small folds (see Fig 4.6).

The foliation, which strikes subparallel to both bedding and the axial plane of the Mine Creek Anticline (see Fig 4.2(b)) is considered to be an axial plane foliation. Metamorphic biotite formed during contact metamorphism of the Greenland Group defines the foliation. The biotite has mimetically crystallised and grown parallel to an earlier formed slaty cleavage, such as occurs at Mount Greenland, defined by the preferred orientation of chlorite, sericite, and probably other platy minerals forming the matrix of the sediments. The original cleavage may have had an origin by de-watering, as suggested by Shelley (1975a).

4.3.4 Jointing

Jointing is well developed throughout the hornfelsed pendant, but only poorly developed in the more massive granites. Field observations indicate the existence of two main joint sets, a dominant set developed parallel to bedding (NW - SE) and a less well developed set striking NE - SW (see Fig 4.3(c)). This latter set is developed sup-parallel to a recognisable fault set (see section 4.3.2), to which it is probably genetically related. No evidence of mineralisation along either joint set has been observed, and there is no apparent change in joint attitudes between the hornfelsed pendant and the Rangitoto and Kakapotahi granites.

4.3.5 Unmineralised Vein and Dyke Rocks

The unmineralised vein and dyke rocks recognised in the area have been described in section 3.4.5. Field relationships indicate that many of the vein and dyke rocks are intruded parallel to bedding, and less commonly parallel to joint directions (see Fig 4.2(d)). These
Fig 4.5 Upper. Axial plane foliation, defined by the preferred orientation of biotite, developed parallel to bedding in the Greenland Group hornfelses.
Mine Creek (G.R. S64/370199)

lower. Poorly developed axial plane cleavage ($S_1$) cross-cutting bedding ($S_0$) at c.15°, in sandy Greenland Group hornfelses.
Mine Creek (G.R. S64/357189)
Fig 4.6 Quartz veinlet in hornfelsed Greenland Group, folded and fractured by shearing parallel to cleavage. Veinlet is 6cm in width.
Float boulder, Mine Creek (G.R. S64/370200)
latter veins and dykes generally dip near vertically and cross-cut bedding at high angles. There are no apparent differences between the orientation of different vein or dyke types, and likewise no changes in orientation with differing host rocks are evident, either in the field or when plotted on a stereograph.

The dykes and quartz veins are inferred to be related to intrusion of the Rangitoto Granite, which underlies the hornfelsed pendant. In Manton’s Workings the mineralised veins are seen to cross-cut a micro-granitic dyke, indicating that the mineralisation probably post-dates the dykes, and by inference the unmineralised quartz veins also.

The origin of the alkaline basic dykes is uncertain, but, it seems most likely that they are either related to intrusion of the Kakapotahi Granite, or alternatively to rifting associated with the opening of the Tasman during the Cretaceous.

4.4 SUMMARY

Structures observed on both Mount Greenland and Mount Rangitoto indicate that folding of the Greenland Group sediments into a series of gently or horizontally plunging upright close folds (after Fleuty, 1964) occurred as a consequence of a regional metamorphism probably during, or prior to, the Ordovician. Faulting of the hinge regions of these folds e.g. the Mine Creek Anticline, may have occurred as a response to tightening of the folds at this time, or alternatively as a response to emplacement of the Rangitoto Granite during the Tuhua Orogeny, c.305 Ma ago (see Chapter 6). The faulted hinges of these folds, on Mount Greenland, proved to be favourable sites for the migration of magmatic fluids during the Tuhua Orogeny, and the formation of quartz vein systems, which are in places, payable auriferous.

Emplacement of the Rangitoto Granite, at the end of the Carboniferous, had little effect on the structure of the already
folded Greenland Group, although hornfelsing of the sediments occurred adjacent to the granite. The hornfelsed pendant preserved on the southern flanks of Mount Rangitoto, was however, cut by numerous granitic vein and dyke rocks, generally intruded parallel to bedding or joint directions. No mineralisation has been observed in these veins and dykes.

Emplacement of the Kakapotahi Granite, to the south of the Rangitoto Granite, occurred during the Late Triassic (see Chapter 6), and also had no discernable effect on the fold pattern produced during the Ordovician regional metamorphism. The Mine Creek Fault, however, was active during the emplacement of the Kakapotahi Granite, and acted as a channelway for the escape of magmatically derived, hydrothermal ore fluids. Ore deposits, such as those worked in Mine Creek, formed in fault or shear zones connected to the Mine Creek Fault, that were probably also active during this time.

Since the Late Mesozoic, faulting of the Greenland Group sediments has occurred, probably during the Rangitata and Kaikoura Orogenies. The faults either strike NW - SE sub-parallel to bedding or NE - SW, cross-cutting bedding at a high angle. The faults are generally normal in character, and although the sense and amount of displacement on them cannot often be determined, no evidence for large horizontal displacements has been observed. The mineralised veins in both Mine Creek and the William Tell No. 2 Level, have been cut by these post-mineralisation faults, although offsets observed are not large.

Uplift of the Mount Greenland massif, totalling at least 500m, has occurred since the late Pliocene, during the Kaikoura Orogeny. Although no structures definitely attributable to this uplift have been located, it is quite likely that many of the faults located in the Mount Greenland area are related to this uplift episode. The effects of the Kaikoura Orogeny on the Rangitoto Range are unknown, but presumably significant uplift and associated faulting could also have occurred in this area, since the Pliocene.
Fig 5.1 Section through the Greenland Group metasediments, Cedar Creek (G.R.S57/413243). This section appears typical of the Greenland Group throughout the Mount Greenland area, although perhaps contains fewer cross-bedded units than in many other localities. See text for discussion.
CHAPTER V

PETROLOGY AND GEOCHEMISTRY

INTRODUCTION

Petrographic features of the Greenland Group metasediments and hornfelses, and the Rangitoto and Kakapotahi granites are described. The sedimentology of the Greenland Group metasediments has been described only briefly, owing to the similarity of the metasediments in the thesis area to those described by Laird (1972) and Laird and Shelley (1974).

Quantitative geochemical analysis of the major rock types was also undertaken, by X-ray fluorescence spectrometry for major and trace elements, and by atomic absorption spectroscopy for gold and silver. Details of the methods used are given in Appendix III. Normative compositions were also determined for both the Rangitoto and Kakapotahi granites.

PETROLOGY

5.1 GREENLAND GROUP METASEDIMENTS

A detailed investigation into the sedimentology and petrology of the Greenland Group metasediments in the thesis area was considered beyond the scope of this work, however, some limited observations were made during general mapping and qualitative examination of a small number of thin sections. The sedimentology and petrology of the Greenland Group sediments in other localities has been described by Reed (1951), Laird (1972) and Laird and Shelley (1974).

5.1.1 Sedimentology

The Greenland Group metasediments of Mount Greenland consist of a well bedded sequence of greywacke and argillite (see Fig 5.1). The greywackes usually occur as medium-fine sands, although coarser sands are also present. The beds, which rarely exceed 1 - 2m, in thickness, commonly grade upwards from medium-coarse sands at the base, into parallel
metres

29
NG
Ungraded fine sandstone
Medium sandstone at base grading into black mudstone
over top 15cm

28 G
Fine sandstone (15cm) grading up into laminated siltstone and mudstone

28 G
Fine sandstone at base grading into laminated siltstone (2cm) and mudstone (2cm)
Interbedded siltstone and mudstone (1cm beds) grading to laminated mudstone

27 G
Fine sandstone at base grading upward into massive grey siltstone
Massive silty grey mudstone
Fine sandstone grading into muddy grey siltstone over top 5cm

26 G
Poorly laminated grey siltstone, grades upward into structureless black mudstone

26 G
Well laminated grey-black mudstone

25 NG
Ungraded, silty very fine sandstone
Parallel laminated siltstone

25 NG
Ungraded very fine sandstone
Structureless grey mudstone overlain by 5cm of laminated siltstone

24 G
Medium fine sandstone at base grading upward into very fine sand and parallel laminated siltstone.

24 G
Parallel laminated mudstone, and siltstone interbedded with structureless grey-black mudstone
Micaeous silty very fine sand, ungraded

23 G
As below

22 G
As below

21 G
Medium-coarse sandstone at base grading upward into fine sand, and silt which grades over a few cm into laminated siltstone.
laminated and/or dark grey-black structureless mudstones. Graded sandstone beds are far more frequent than ungraded beds, and may account for as much as 70 - 80% of the sand sized fraction. Ungraded beds are usually fine or very fine sands, and are often micaceous. The base of the sandstone beds is always sharp, the beds generally overlying laminated siltstones or mudstones. Sole marks are not widespread but may be well developed locally. Load casts and ripples are the dominant sole marks present, and occasionally flutes are observed. An excellent exposure of broadly spatulate flutes (c.f. Conybeare and Crook, 1968) was observed in the base of a graded sandstone bed in Swiss Creek (G.R. 557/393220).

A limited number of paleocurrent estimates was made using both sole marks and cross-beds. These, when corrected for structure, indicate a possible southwestern source for the sediments, however, a considerable amount of systematic investigation would be required to confirm this.

Laminated siltstones and mudstones less than 50cm thick are commonly developed at the top of graded sand beds. The base of these beds grades downwards into the underlying sands. In contrast the upper contacts with overlying sandstones or mudstones are always sharp. As well as occurring at the top of graded sandstone beds, laminated siltstones also occur interbedded with structureless dark-grey mudstones.

Cross-bedded siltstones and mudstones are less common than their parallel laminated counterparts, and are commonly absent in many sections of a few tens of metres thickness. Where present, however, the cross-bedded siltstones usually occur either at the top of graded sandstone beds, or more commonly, overlying parallel laminated siltstones.

Structureless grey or dark-grey mudstones occur most commonly as beds less than 75cm thick, although may on occasion attain thicknesses of several metres. The basal contacts of the beds are either sharp or gradational with the underlying siltstones or sandstones, whereas the upper contacts are always sharp.
Throughout the thesis area sandstones are dominant over siltstones and mudstones, the estimated sand : mud ratio being approximately 2 : 1.

5.1.2 Petrology

The petrology of the Greenland Group metasediments in the thesis area is similar to that described by Laird (1972) and Laird and Shelley (1974). The sediments consist of angular-subrounded fragments of quartz (35 - 50%), feldspar (2 - 4%), sedimentary and volcanic rock fragments (1 - 2%), and minor chert, contained in a matrix of sericite and minor chlorite.

The sandstones are very poorly sorted, particularly at the base of graded beds, with angular fragments (0.2 - 0.8mm) of quartz, and feldspar (0.2 - 0.4mm) in a fine grained matrix (see Fig 5.2). Sub-rounded volcanic and sedimentary rock fragments (0.2 - 0.7mm) are also present.

Quartz is the dominant sand-sized detrital mineral, most commonly occurring as strained grains with undulose extinction, although strain free grains are also present. Polycrystalline aggregates of quartz, sometimes containing small sericite flakes, are also present but are regarded as rock fragments. Fine grained quartz (< 0.05mm) is common in the matrix of the rock.

The feldspar present (2 - 4%), occurs as angular-subrounded anhedia or subhedia generally between 0.2 and 0.4mm across. Albite twinning is ubiquitous, and Michel-Levy extinction angle methods indicate a sodic oligoclase or albite composition. Occasionally the plagioclase occurs with an aggregate of quartz crystals (c.f. Laird, 1972, p.287), suggesting the presence of granitic rock fragments.

Rock fragments, other than those already noted, are minor constituent (1 - 2%) of the sandstones. The fragments (0.2 - 0.7mm) consisting largely of chlorite and minor sericite, are rounded, and although their origin is unknown, are thought to be volcanic.
Fig 5.2(a) Photomicrograph of typical Greenland Group greywacke, consisting dominantly of poorly sorted, angular quartz and feldspar in a fine grained matrix consisting of micas and clay minerals. Photograph is from the base of a graded bed. U.C. 9453 Thin section, XPL.
Cedar Creek, Mt Greenland (G.R. S57/413243)

Fig 5.2(b) Photomicrograph of Greenland Group greywacke showing alignment of inequigranular quartz grains parallel to incipient cleavage (S₁). Bedding is parallel to the sides of the photograph (N-S). U.C. 9453. Thin section, XPL
Cedar Creek, Mt Greenland (G.R. S57/413243)
Detrital muscovite (0.4 - 1mm) flakes occur in the sandstones, but rarely account for more than one percent of the rock.

The matrix (30 - 50%) of the sandstones consists of fine grained flakes of sericite, and minor amounts of chlorite. The matrix is recrystallised, reflecting the effects of low grade regional metamorphism, and is crudely aligned forming an incipient axial plane cleavage, which is well developed in the argillites. No recrystallisation of detrital quartz grains is evident.

Parallel laminated siltstones consist of 1 - 3mm thick laminae of silt interbedded with mudstone laminae, of similar thickness. The silty laminae are comprised of angular quartz fragments and minor plagioclase feldspar grains up to 0.04mm across. Compositionally the siltstones are similar to the sandstones, but do not contain easily identifiable rock fragments. These fragments may, like the sericitic matrix, have been recrystallised or alternatively may be less abundant in the finer grained beds. The muddy laminae consist almost entirely of very fine grained micas and clay minerals (see Fig 5.3(a)).

A well developed cleavage is common in the muddy laminae, but less well developed in the coarser silty laminae. Where the cleavage is at an acute angle or sub-parallel to bedding, the cleavage is defined by a well developed mica orientation. In some specimens, however, where cleavage cuts bedding at a higher angle, the mica preferred orientation is less intense (see Fig 5.3(b)) and appears buckled, the cleavage taking on the appearance of a strain slip cleavage passing through the hinges of the buckles. Alignment of elongate quartz grains parallel to incipient cleavage in the sandstones is present in some specimens, and may be due to dewatering during cleavage formation, accompanying fold development, as suggested by Laird and Shelley (1974) (see Fig 5.2(b)).

The dark-grey mudstones consist almost entirely of very fine grained micas and clay minerals with occasional very fine silt sized fragments of quartz. A penetrative cleavage defined by the preferred orientation of the micas is common.
Fig 5.3(a) Parallel laminated argillite, with cleavage ($S_1$), which is well developed in the finer grained laminae, cross-cutting bedding and being refracted by the coarser laminae. Thin section, cross polarised light. U.C. 9454 Cedar Creek, Mt Greenland (G.R. S57/413243)

Fig 5.3(b) Magnified view of cleavage ($S_1$) from Fig 5.3(a) above. The cleavage is defined by 'kinks' in the groundmass micas which have only a weak preferred orientation, and appears to be an incipient strain slip cleavage. Thin section, cross-polarised light.
The sedimentary structures observed, and the petrology of the Greenland Group metasediments of the thesis area suggests that the sediments, like those described by Laird (1972) and Laird and Shelley (1974) are probably turbidites deposited on the margins of a submarine slope, possibly a fan. The source of the sediments is apparently from the south-west. Post-depositional deformation of the sediments has resulted in the formation of a penetrative cleavage in the mudstone beds, and an incipient cleavage in the sandstone beds. Recrystallisation of the matrix of the sandy beds and the muddy beds accompanied cleavage formation, probably a consequence of regional folding during the Ordovician.

5.2 GREENLAND GROUP HORNFELSES

The Greenland Group hornfelses, preserved as a pendant on the southern flanks of Mount Rangitoto, are distinguished from the Greenland Group metasediments (gm) of Mount Greenland, by the ubiquitous occurrence of metamorphic biotite, and to a lesser degree muscovite. Characteristically the hornfelses are better indurated and darker in colour than the metasediments, and sedimentary structures, apart from bedding, are rare. 'Spots' of biotite up to 1cm across are common, and smaller spots (< 0.5cm) of sericite, and/or muscovite, are sometimes present, particularly in the finer beds. A well developed penetrative cleavage, defined by the preferred orientation of biotite, usually at an acute angle or sub-parallel to bedding, is evident in the field.

5.2.1 Mineralogy

a) Quartz: Equigranular quartz (0.03 - 3mm) occurs with biotite, throughout the hornfelsed pendant. The quartz shows widespread evidence of recrystallisation, although in some rocks detrital quartz grains remain. In the more metamorphosed rocks triple junctions are commonly approached, resulting in a typical polygonal-granoblastic texture (see Fig 5.4(a)).
Fig 5.4(a) Typical polygonal-granoblastic texture developed in the Greenland Group hornfelses. The biotite preferred orientation is parallel to cleavage. Thin section, cross polarised light. U.C.9455
Mine Creek, Mt Rangitoto (G.R. S64/347186)

Fig 5.4(b) Muscovite developed between sericitised andalusite porphyroblasts. The strong muscovite preferred orientation is oblique to cleavage, defined by the preferred orientation of biotite (as above). See text for discussion. Thin section, cross polarised light. U.C. 9456.
Mine Creek, Mt Rangitoto (G.R. S64/367200)
b) **Biotite**: Pale-dark brown pleochroic biotite (0.04 - 0.3mm) occurs throughout the rocks of the hornfelsed pendant, commonly forming a polygonal-granoblastic texture with quartz. A well developed preferred orientation, defining a cleavage direction is discernable in most specimens, and is probably due to mimetic crystallisation and growth of biotite, formed from the breakdown products of sericite and chlorite in the groundmass of the metasediments. Decussate aggreages of biotite may form 'spots' up to 1cm across, in a polygonal-granoblastic groundmass.

c) **Muscovite**: Muscovite is a common constituent of the pendant rocks, usually occurring as flakes similar in size to accompanying biotite (0.04 - 0.3mm) or as larger crystals up to 1.5mm across. The finer grained muscovite commonly occurs intergrown with biotite, in such cases the two minerals having a common orientation.

Larger flakes of muscovite, up to 1.5mm across are commonly poikiloblastic, including rounded quartz grains and less commonly biotite, indicating a late development during the metamorphic event. The larger flakes, which may form decussate aggreages several millimetres across, have no obvious preferred orientation. Twinning of the muscovite is rare, but has been observed in some sections.

Muscovite flakes, 0.1 - 0.3mm long are also common around the margins of sericitised porphyroblasts, and show a well developed preferred orientation (see Fig 5.4(b)) oblique to the cleavage defined by biotite. This re-orientation may be due either to re-orientation of the rocks with respect to stress prior to crystallisation of the muscovite, or alternatively, to a change in the stress regime existing during crystallisation of the two phases.

Very fine grained muscovite or sericite is also widespread, most commonly occurring as aggregates with quartz and occasionally biotite, as replacements of andalusite porphyroblasts. Larger flakes of randomly oriented muscovite may also occur in these porphyroblasts, which may account for up to 70 - 80% of the rock in some cases.
d) **Feldspar:** Relict albite, as determined by the Michel-Levy method, is present in some of the less metamorphosed hornfelses in the area, and may account for up to 5% of such samples. The crystals are commonly subhedral, are albite twinned, and are sedimentary in origin. As the effects of hornfelsing become more intense oligoclase appears at the expense of albite. Metamorphic oligoclase never exceeds c.1% of the hornfelses, and occurs as small anhedral crystals (≤ 0.2mm) intergrown with quartz in a polygonal-granoblastic texture.

e) **Andalusite:** Andalusite porphyroblasts, now completely altered to fine grained aggregates of sericite + quartz, are widespread throughout the pendant, particularly in the finer grained hornfelses, where knotted slates may result. The porphyroblasts (3 - 7mm) are generally rounded or elliptical in shape, and many posses 'core' structures (see Fig 5.5(a)) a common feature of andalusite. Less commonly the porphyroblasts are rectangular in shape, reflecting the presence of (100) or (010) sections. The alteration of andalusite to quartz + sericite requires the addition of potassium and water to the hornfelses.

Diamond shaped (001) sections of andalusite, now completely altered to quartz, are also present in some sections, although are not particularly common. One such section of andalusite, now replaced by quartz, was observed within a quartz-sericite porphyroblast, thought to have originally been cordierite (see Fig 5.5(b)), suggesting that extensive retrograde metamorphism has occurred, either as a consequence of slow cooling related to intrusion of the Rangitoto Granite, or, alternatively to subsequent intrusion of the Kakapotahi Granite.

f) **Cordierite:** No cordierite has been observed in thin section however, large generally rounded quartz-sericite porphyroblasts (up to 8mm) may have formed due to alteration of cordierite, as well as andalusite.

g) **Sillimanite:** Sillimanite is present, but not common in the pendant. Fine fibrous needles, often curved, occur cross-cutting quartz veinlets in the hornfels (see
Fig 5.5(a) Andalusite porphyroblast, now altered to quartz + sericite, showing relict core structure. Thin section, cross polarised light. U.C.9456.
Mine Creek, Mt Rangitoto (G.R. S64/367200)

Fig 5.5(b) Rectangular cordierite (?) porphyroblast, altered to quartz + sericite, overgrowing an andalusite porphyroblast replaced by quartz. See text for discussion. Thin section, cross polarised light. U.C.9456
Fig 5.6) and are thought to have formed as a consequence of contact metamorphism of the pendant, associated with intrusion of the Rangitoto Granite.

h) Tourmaline: Tourmaline occurs as an accessory mineral throughout the rocks of the pendant, but may become so abundant locally, as to result in the almost complete tourmalinisation of the hornfelses. The euhedral tourmaline crystals (up to 4mm long) occur either singly in the hornfelses, or where they are more abundant, as randomly oriented or poorly developed radiating aggregates in a dominantly quartz matrix. The crystals are commonly colour zoned, being green at the core and brown near the margins and have a low birefringence, indicating that they are of the schorl variety.

i) Pyrrhotite: Pyrrhotite occurs disseminated in the hornfelsed pendant at one locality near the old workings in the east branch of Mine Creek (G.R.664/366195) and accounts for up to 1% of the sediment. A fuller description of this occurrence is given in section 8.3.

j) Accessory Minerals: Magnetite occurs as an accessory mineral, usually as small (0.2mm) anhedral crystals disseminated throughout the hornfelses.

5.2.2 Textures

Macroscopeically, the Greenland Group hornfelses appear as spotted slates, with a well developed cleavage defined by the preferred orientation of biotite and fine grained muscovite. Locally sub-schists accompanied by incipient mineral segregation occur.

Microscopeically, the dominant texture is a granoblastic-polygonal intergrowth of quartz (+ minor feldspar), biotite and muscovite. In some sections two cleavage directions are evident. The earliest formed cleavage, which is observable in the field, is defined by the preferred orientation of biotite, and is usually parallel or sub-parallel to bedding. A subsequent cleavage, defined by the preferred orientation of muscovite around sericitised porphyroblasts, occurs at an angle to the earlier formed cleavage, and is only observable in thin section.
Fig 5.6 Sillimanite needles cross-cutting a quartz veinlet in hornfelsed Greenland Group. See text for discussion. Thin section, cross polarised light.
U.C.9457.

Mine Creek, Mt Rangitoto (G.R. S64/370199)
Retrogressive metamorphic textures such as the sericitisation and silicification of porphyroblasts and the growth of muscovite, are widespread and well developed throughout the pendant. These textures suggest, either slow cooling of the granite intrusive, or the existence of a subsequent thermal event. K-Ar dating suggests two phases of granite emplacement (see Chapter 6) in the Rangitoto Range area, and accordingly it seems most likely that the abundant retrogressive features observed are related to the latest phase of intrusive activity, viz emplacement of the Kakapotahi Granite.

Discussion of the P-T conditions of metamorphism implied for the aureole assemblage is contained in section 5.5, following.

5.3 RANGITOTO GRANITE

The Rangitoto Granite, forming the bulk of the Rangitoto Range, generally occurs as a pale grey equigranular biotite-granite (after Streckeisen, 1976). Examination of stained samples and qualitative thin section work indicates an approximate composition, plagioclase (40%), quartz (25%), K-feldspar (20%), and biotite (15%) commonly partly replaced by chlorite. Accessory minerals include rutile and zircon, and epidote, zoisite, calcite, chlorite and sericite are also present as alteration products of the feldspars. Partially assimilated Greenland Group xenoliths are common in the granite.

5.3.1 Mineralogy

a) Plagioclase: Plagioclase (0.1 - 5mm) occurs most commonly as subhedral to anhedral crystals, intergrown with quartz and K-feldspar. Undulatory extinction of Albite and Carlsbad twins indicates that normal zoning is common. Oscillatory zoning is also common. Measurement of extinction angles indicates that the composition of the plagioclase ranges from andesine (An_{40}) at the core to albite (An_{8-10}) or sodic oligoclase (An_{10-15}) at the margin (see Fig 5.7). Unzoned crystals have refractive
Fig 5.7 Plagioclase showing oscillatory zoning (An\textsubscript{40-}
An\textsubscript{10}) in Rangitoto Granite. Note partial replace-
ment by mimetic biotite. Thin section, cross polarised light. U.C. 9458.
Mine Creek, Mt Rangitoto (G.R. S64/365195)

Fig 5.8 Plagioclase, now completely altered to a
fine grained intergrowth of zoisite + epidote, and
secondary sericite and muscovite. Rangitoto Granite.
Thin section, cross polarised light. U.C. 9458.
(G.R. S64/365195)
indices lower than quartz, indicating that they are either sodic oligoclase or albite. The grain boundaries are commonly irregular in detail, and may have undergone some resorption. The widespread occurrence of plagioclase zoning suggests that fractional crystallisation processes were operative during crystallisation of the magma.

Alteration of plagioclase is extensive, and is generally either of sericitic or propylitic form.

Sericitisation is widespread, usually being best developed in small crystals as well as the cores of larger crystals, many of which have unaltered rims. The secondary sericite flakes are commonly aligned parallel to cleavage directions in the host plagioclase (Fig 5.8), although where alteration is more complete clusters of flakes with more random orientations predominate.

Propylitic alteration commonly accompanies sericitic alteration of the plagioclase feldspars, and is marked by the development of very fine grained aggregates of epidote and zoisite, often accompanied by sericite, or larger flakes of muscovite, calcite and chlorite. The alteration, which results in intense clouding of the feldspars, is usually confined to patches within the core of plagioclase crystals, and less commonly may mimic the zoning of the host crystal.

Both the sericitic and propylitic alteration are probably a consequence of late stage autoalteration processes operating during cooling of the magma.

b) Quartz: Quartz (0.01 - 4mm) usually occurs as anhedral crystals intergrown with the feldspars and biotite. The crystals, which are often highly fractured, are commonly strained and exhibit undulose extinction. The crystals are only occasionally poikilitic including both plagioclase and potassium feldspars, and biotite in minor amounts. Rutile occurs as fine needles within the quartz.

c) Potassium Feldspar: The potassium feldspar present is microcline, which almost always shows well developed cross-hatched twinning. The crystals (0.01 - 8mm) are
usually anhedral, occurring intergrown with quartz, plagioclase and biotite in the groundmass, no phenocrysts being present. The larger crystals are sometimes markedly poikilitic, including crystals of all the major minerals present. Perthitic textures, consisting of sodic plagioclase in a microcline host, are common.

No attempt has been made to determine the composition of the microcline.

d) **Biotite**: Pale brown, strongly pleochroic biotite occurs as single idiomorphic crystals (0.05 - 4mm), and as deccussate aggregates up to 8mm long. Only rarely does the biotite define a foliation in the granite, the biotite usually showing no obvious preferred orientation.

Chloritic alteration of the biotite is ubiquitous (see Fig 5.9(a)) chlorite being developed parallel to the basal biotite cleavage, and often occurring interlaminated with biotite. The secondary chlorite is biaxial negative, with strong purple or less commonly blue interference colours indicating that it is an Fe-Mg Chlorite (Albee, 1962). Fine grained epidote also occurs as an alteration product of biotite, sometimes together with secondary chlorite (see Fig 5.9(b)).

e) **Accessory Minerals**: Apart from the minerals already noted (viz. calcite, chlorite, epidote, zoisite, rutile) minor amounts of zircon were observed, generally as euhedral crystals with slightly rounded ends.

5.3.2 **Textures**

In the field the Rangitoto Granite is remarkably homogeneous, lacking obvious phenocrysts and almost always lacking foliation. Partially assimilated xenoliths of Greenland Group material usually 4 - 7cm long, but occasionally exceeding 10 - 15cm in length, are common, usually having a spherical or elliptical shape. The xenoliths show a complete range of assimilation, ranging from unaltered hornfelsed xenoliths with sharp contacts with the granite, to barely recognisable spots of residual biotite.
Fig 5.9(a) Magmatic biotite being replaced along cleavage by Fe-Mg chlorite (Albee, 1962). Rangitoto Granite. Thin section, partly cross polarised light. U.C. 9458.
Mine Creek, Mt Rangitoto (G.R. S64/365195)

Fig 5.9(b) Secondary epidote formed by the alteration of biotite, and possibly plagioclase. Rangitoto Granite. U.C.9473. Thin section, cross polarised light.
Mine Creek, Mt Rangitoto (G.R. S64/366197)
The margins of the xenoliths show clear evidence of assimilation. Euhedral to anhedral microcline, and less commonly plagioclase, crystals up to 1mm across are commonly developed along the margins of the xenoliths. The feldspars which are commonly poikilitic, including relict biotite from the hornfelsed xenoliths, sometimes have identical orientations to adjacent crystals in the granite, probably due to mimetic crystallisation and growth.

The least altered xenoliths appear identical in thin section to many of the Greenland Group hornfelses, consisting almost entirely of an equigranular mosaic of quartz (0.01 - 0.1mm) and biotite (0.01 - 0.6mm). With increasing assimilation, sodic oligoclase and albite subhedra-euhedra (0.05 - 0.15mm) become more abundant in the xenoliths. The plagioclase is commonly zoned, albite twins are often present, and very fine grained biotite is commonly included in the crystals.

As the degree of assimilation increases, the number and size of the plagioclase crystals in the xenoliths increases, until crystals approaching the size of those in the granite (5 - 8mm) are reached. The margins of the xenoliths, too, become more diffuse owing to encroachment by the feldspars of the granite, both microcline and plagioclase, accompanied by recrystallisation and loss of quartz from the xenolith. Continued granitisation of the margins of the xenoliths, accompanied by growth of the plagioclase feldspars in the xenoliths results in the almost complete destruction of the xenolith, only spots of biotite remaining to indicate its former existence. Presumably recrystallisation of the relict biotite completes the granitisation process.

The occasional occurrence of unoriented clusters of angular xenoliths within the granite, resulting in a migmatitic texture, indicates that the included xenoliths were probably originally fractured fragments of Greenland Group country rock, subsequently included in the magma and broken down by assimilation or granitisation processes.
Within the granite, myrmekitic intergrowths of quartz and sodic oligoclase are commonly developed along plagioclase-microcline contacts (see Fig 5.10). These intergrowths, up to 0.8mm across, have formed by the replacement of microcline by sodic oligoclase or albite, with the subsequent separation of excess quartz. The plagioclase in the myrmekite is commonly in optical and crystallographic continuity with the adjacent plagioclase, reflecting the effects of mimetic crystallisation.

Contacts between the feldspars, quartz and biotite are generally sharp but irregular in outline. Plagioclase crystals adjacent to biotite, are sometimes aligned parallel to the biotite flakes providing further evidence of mimetic crystallisation. Contiguous plagioclase crystals are also commonly oriented on occasion, probably due to mimetic crystallisation.

The poikilitic character of much of the microcline present, particularly the larger anhedral crystals (see section 5.3.1), suggests that these crystals probably formed during the later stages of crystallisation of the magma.

The genetic significance of the mineralogy and geochemistry (section 5.8) of the Rangitoto Granite is considered in section 5.10, where likely conditions of origin and emplacement of the granite are discussed.

5.4 KAKAPOTAHI GRANITE

The Kakapotahi Granite crops out along the lower southern slopes of the Rangitoto Range, and extends southwards across Flat Creek and the Kakapotahi River, to the Purcell Range. The granite is an equigranular leucocratic granite (after Streckeisen, 1976). Qualitative thin section and stained slab examination indicates an approximate modal composition of K-feldspar (35%), plagioclase (30%), quartz (30%), biotite (3 - 4%), and muscovite (≈ 1%). Accessory minerals include garnet and traces of apatite. Sericite, zoisite, and minor epidote occur as alteration products.
Fig 5.10 Myrmekite developed along plagioclase-microcline boundary due to the replacement of microcline by albite or Na-oligoclase, accompanied by the separation of excess quartz. Note the optical continuity of host and myrmekite plagioclase (black). Thin section, cross polarised light. U.C.9458. Rangitoto Granite.

Mine Creek, Mt Rangitoto. (G.R. S64/365195)
The normative composition of the rock is discussed in section 5.9, together with the geochemistry.

5.4.1 Mineralogy

a) Potassium Feldspar: Microcline (0.1 - 7mm) of unknown composition, is the potassium feldspar present. The crystals which average 2 - 3mm in size are generally anhedral, or less commonly subhedral, and cross-hatched twinning is ubiquitous. Many of the larger crystals are poikilitic, most commonly including rounded quartz crystals (up to 0.3mm), and less commonly plagioclase (up to 0.2mm). Perthite, consisting of sodic oligoclase in a microcline host, is common in the larger crystals.

b) Plagioclase: Plagioclase (0.1 - 6mm) is usually of anhedral or subhedral habit. Normal zoning is common, but oscillatory zoning is generally absent, in contrast to the Rangitoto Granite. The composition of the plagioclase, as determined by the Michel-Levy extinction angle method, ranges from andesine (An_{35}) in the cores of zoned crystals, to albite (An_{6-10}) at the margins. Most of the unzoned crystals appear to be either albite or sodic oligoclase although more calcic and probably earlier formed, andesine occurs as inclusions in poikilitic microcline.

Alteration of the plagioclase is ubiquitous and almost identical in nature to that observed in the Rangitoto Granite. Sericitisation is widespread, and best developed in the cores of plagioclase crystals, some of which have completely unaltered rims of albite or sodic oligoclase, possibly late stage overgrowths. The secondary sericite occurs as small randomly oriented flakes and less commonly as flakes parallel to cleavage. The alteration is not intense, most of the plagioclase being flecked with small flakes of sericite (up to 0.2mm). Occasionally growth of this sericite to form muscovite is evident, particularly in the smaller plagioclase crystals, which may be almost completely altered to secondary muscovite.

Propylitic alteration, of plagioclase to zoisite, sometimes associated with minor chlorite and epidote, is also widespread, but less intense than in the Rangitoto Granite.
Small plagioclase crystals are occasionally completely propylitised. In the larger crystals, however, propylitisation is generally confined to patches within the cores and is often accompanied by more intense sericitisation.

Pink and red varieties of the Kakapotahi Granite crop out in Galena Gulch (G.R.$64/340192). Thin section examination of these reddened granites indicates that their mineralogy is almost identical to the typical, pale-grey leucocratic granite, the difference in colour between the two varieties being due to a different alteration assemblage in the plagioclase. The pink and red granites have undergone only slight sericitic alteration, the plagioclase being flecked sparingly with sericite. Propylitic alteration is, however, absent.

The pink or red colour of the granites is attributed to the presence of very fine disseminations of an opaque mineral, probably hematite, in the plagioclase. The very fine grain-size of the hematite makes positive identification in thin section impossible, however, its occurrence in the plagioclase of reddened granites has been noted by numerous workers, and attributed to pneumatolytic processes operating during the latest stages of granite intrusion. Similar processes may have been responsible for the reddening of the Kakapotahi Granite.

c) Quartz: Quartz (0.02 - 4mm) occurs commonly as large, strained anhedral crystals, commonly containing inclusion trails, and less commonly as small anhedral crystals along the contacts between feldspars. Rounded inclusions of quartz are common in both poikilitic microcline and plagioclase.

d) Biotite: Biotite (0.1 - 6mm) occurs as single flakes or small decussate aggregates scattered sparsely throughout the rock. The brown-pale brown pleochroic flakes are commonly partially replaced by chlorite, along the margins and parallel to cleavage. The alteration or replacement of biotite by chlorite is less well developed than in the Rangitoto Granite. The anomalous blue interference colours of the chlorite indicate that it is an
Fe-Mg variety (Albee, 1962). Garnets often occur with the biotite, and in places distort the biotite flakes, bending the cleavage about the garnets. This is probably due to the growth of biotite about the earlier-formed garnets.

e) Muscovite: Muscovite (0.1 - 5mm) accounts for up to 1% of the rock, usually occurring as discrete flakes or small aggregates of crystals often associated with biotite. These flakes (see Fig 5.11(a)) are of primary magmatic origin. Small flakes also occur as alteration products of plagioclase, and occur scattered throughout the plagioclase host.

f) Accessory Minerals: Garnet (~0.2mm) is the most abundant accessory mineral, occurring as sub-rounded euhedra (see Fig 5.11(b)) either interstitial to or included in the felsic minerals. The pale pink colour of the euhedra, which are commonly fractured, suggests that the garnets are almandine, the most common variety of garnet in igneous rocks.

Apatite is also present in trace amounts, and minor epidote and chlorite occur as alteration products.

The mineralogy of both the Rangitoto and Kakapotahi granites is summarised by way of comparison in Table 5.6.

5.4.2 Textures

Apart from the absence of Greenland Group xenoliths in the Kakapotahi Granite, the textures observed are almost identical to those in the Rangitoto Granite (see section 5.3.2). Myrmekites up to 0.4mm across are common along plagioclase-microcline boundaries. Microcline, plagioclase and quartz are mutually poikilitic to varying degrees. Inclusions of microcline and quartz in plagioclase, with an orientation identical to crystals adjacent to the plagioclase, indicates that quite extensive replacement of both microline and quartz by plagioclase has occurred. This is further indicated by the embayed nature of many of the quartz and microline crystals adjacent to plagioclase. This replacement is thought to be a late stage crystallisation feature.
Fig 5.11(a) Magmatic muscovite flake, intergrown with quartz and feldspar. Kakapotahi Granite. Thin section cross polarised light. U.C.9460.
Galena Gulch, Mt Rangitoto (G.R. S64/340192)

Fig 5.11(b) Sub-rounded almandine garnet (black) deforming adjacent biotite. Note also the zoned, slightly altered highly poikilitic plagioclase. Kakapotahi Granite. Thin section, cross polarised light. U.C.9461.
Kakapotahi River (G.R. S64/342182)
Garnet is commonly included in all the felsic minerals, indicating its early crystallisation in the magma.

Normal zoning of the plagioclase is common, however, oscillatory zoning is considerably less common than in the Rangitoto Granite. Why this should be so is unclear, however, it is likely that the difference is due to the different crystallisation histories of the magmas.

The genetic significance of the mineralogy and textures observed is considered, together with the geochemistry of the granite (section 5.9) in section 5.11.

GEOCHEMISTRY

5.5 GREENLAND GROUP METASEDMENTS

Geochemical analyses of Greenland Group metasediments and hornfelses are given in Table 5.1.

Gold and silver were analysed by atomic absorption spectroscopy, the remaining elements being analysed by x-ray fluorescence spectrometry. Details of the analytical methods used are contained in Appendix III. Sample 10198 is an argillite, the remaining metasedimentary samples being greywackes.

a) Major Elements:

The geochemistry of the Greenland Group north of Greymouth has been described by Nathan (1976), and is characterised by the following features: \( \text{K}_2\text{O} > \text{Na}_2\text{O} \), \( \text{CaO} < 1.5\% \), and \( \text{Rb} > \text{Sr} \), which reflects the lack of feldspar and clastic volcanic detritus in the sediments.

The four samples analysed in this study all show the above characteristics, and are regarded as being geochemically typical of the Greenland Group elsewhere. The samples contain 60 - 71% \( \text{SiO}_2 \), the \( \text{SiO}_2 \) content increasing with increased grain size as noted by Nathan (1976). An inverse correlation with \( \text{Al}_2\text{O}_3 \) also exists, and is regarded by Nathan (1976) as reflecting the diluent
Geochemical analyses of major oxides (wt%) and selected trace elements (ppm, unless otherwise stated) from Greenland Group metasediments and hornfelses.

<table>
<thead>
<tr>
<th>No.</th>
<th>GREENLAND GROUP METASEDIMENTS</th>
<th>GREENLAND GROUP HORNFELSES</th>
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<tr>
<td></td>
<td>SI0₂</td>
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</tr>
<tr>
<td>10196</td>
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<td>10191</td>
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<tr>
<td>10192</td>
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<td>0.88</td>
</tr>
<tr>
<td>10193</td>
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<td>0.70</td>
</tr>
<tr>
<td>10194</td>
<td>51.5</td>
<td>0.92</td>
</tr>
</tbody>
</table>

**TOTAL** 98.39 97.41 99.34 99.74 99.53 100.55 98.53 97.42 99.27 97.83 94.00

**TABLE 5.1**: Geochemical analyses of major oxides (wt%) and selected trace elements (ppm, unless otherwise stated) from Greenland Group metasediments and hornfelses.

Sample numbers are XRF No's from University of Canterbury.

Descriptions and locations are contained in Appendix II.

Samples 10192 and 10193 have undergone sulphur metasomatism, Sample 10194 is highly tourmalinised.

Sample 10198 is an argillite, the remaining metasediment samples are greyvacke.

Where S is abundant, iron as Fe has been assigned to S to form FeS₂.

See text (section 5.8) for discussion.
effects of a clay mineral matrix, on quartz, with decreasing grain size.

Other characteristics of the Greenland Group have been discussed by Nathan (1976) in relation to the sedimentology and also hold for the samples analysed in this study. Perhaps worthy of note is a slightly higher MgO content 3.9 - 5.0%, in the Mount Greenland greywackes than those reported by Nathan (2.8 - 4.5%). This probably reflects an increased mafic content in the samples analysed, and may be due to a higher proportion of chlorite in the matrix, or alternatively, the presence of more Mg-rich chlorites in the Mount Greenland samples.

b) Trace Elements:
   i) Rubidum, Strontium and Barium:

   Characteristically the Greenland Group contains high concentrations of Rb (120 - 243ppm) and low levels of Sr (29 - 87ppm) relative to many crustal rocks (Nathan, 1976), the low Sr content reflecting the lack of calcic plagioclase in the sediments. The abundances of Rb (124 - 232ppm) and Sr (31 - 38ppm) in the Mount Greenland metasediments analysed in this study are typical of those analysed by Nathan (1976).

   Ba (622 - 1034ppm) shows considerable variation in the samples analysed, a trend noted by Nathan (1976), who also observed a linear correlation between K and Ba, indicating that most of the Ba is held in the clay matrix. A positive correlation was also noted in the metasediments analysed in this study, although the linear trend of Nathan (1976) was not so pronounced (see Fig 5.12(a)).

   The relationship of the trace and major elements discussed above, to the sedimentology of the Greenland Group has been discussed by Nathan (1976), to whom those interested are referred.

   ii) Yttrium, Niobium, Zirconium:

   Zr concentrations in the samples analysed range from 141 - 502ppm, the lowest values being in the argillite (10198) and highest values in the coarser grained greywackes. This increased concentration with grain size and SiO₂ content is probably due to the greater abundance of detrital zircons,
Fig 5.12 (a) $K_2O$ vs Ba. Note the linear trends for both metasediments and hornfelses, and the higher $K_2O$ contents of the hornfelses (see text for discussion).

(b) $K_2O$ vs Zr. A slight loss of Zr occurs during hornfelsing possibly due to resorption of detrital zircons.
the main Zr bearing mineral, in the coarser greywackes. The higher Zr content of the greywackes, as opposed to the argillites, is not universal (see Table 5.1), and is thought to reflect the provenance of the sediments, deeply eroded acid igneous terranes supplying abundant erosion resistant zircons, which were deposited in sandy beds along with hydraulically equivalent quartz grains. Y and Nb range in concentration from 32 - 47 ppm and 15 - 18 ppm respectively. These elements are not significantly enriched in the metasediments, and show no correlation with either grainsize or other elements. The values obtained are likely to be typical background values in the metasediments.

iii) Copper and Zinc:

Cu and Zn occur in the metasediments in narrow concentration ranges, Cu from 19 - 49 ppm and Zn from 81 - 101 ppm. Cu shows a positive correlation with Al\textsubscript{2}O\textsubscript{3} (see Fig 5.13(a)), and a negative correlation with SiO\textsubscript{2}. (The trends are possibly linear, however, the number of samples analysed is too small to define the trends with certainty.) The correlations indicate that Cu is most abundant in the finer grained (low SiO\textsubscript{2}, high Al\textsubscript{2}O\textsubscript{3}) argillites than the greywackes, and is probably adsorbed onto micas or clay minerals in the matrix of the sediments.

Zn in the metasediments shows a greater abundance and variation than Cu and also has a positive correlation with Al\textsubscript{2}O\textsubscript{3} (see Fig 5.13(b)), and a negative correlation with SiO\textsubscript{2} reflecting its presence in the matrix, where it is assumed to occur adsorbed onto the clay minerals and micas, along with Cu and the other transition metals.

iv) Lead, Tin and Antimony:

Pb (10 - 13 ppm) occurs in the greywackes in typical concentrations for such rocks, but is enriched in the argillite which contains 79 ppm Pb. This is some three times higher than the typical background for shales (see Table 5.2). It is not known why the argillite sample (10194) should show such enrichment in Pb, or if this value is typical of Greenland Group argillites elsewhere. The enrichment of Pb in the argillite again suggests that Pb is contained in the mica and clay mineral matrix of the metasediments.
Fig 5.13. (a) $\text{Al}_2\text{O}_3$ vs Cu. Cu increases with $\text{Al}_2\text{O}_3$ in the metasediments reflecting its presence in the micaceous groundmass. A slight enrichment of Cu probably accompanies hornfelsing.

(b) $\text{Al}_2\text{O}_3$ vs Zn. Zn increases with $\text{Al}_2\text{O}_3$ in the metasediments reflecting its presence in the groundmass also. Zn shows an increase due to hornfelsing.
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<tbody>
<tr>
<td>Ba</td>
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<td>220 - 510</td>
<td>472</td>
<td>510</td>
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<tr>
<td>Nb</td>
<td>8</td>
<td>7</td>
<td>17</td>
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</tr>
<tr>
<td>Rb</td>
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<td>-</td>
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<td>159</td>
</tr>
<tr>
<td>Sr</td>
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<td>90 - 200</td>
<td>122</td>
<td>61</td>
</tr>
<tr>
<td>Y</td>
<td>9 - 22</td>
<td>25 - 38</td>
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<td>As</td>
<td>1 - 4</td>
<td>4 - 9</td>
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<td>-</td>
</tr>
<tr>
<td>Sn</td>
<td>-</td>
<td>-</td>
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</table>

**TABLE 5.2**: Selected average trace element compositions of sandstones and shales from the United States, and of the Greenland Group.

Sn and Sb show no noticeable enrichment in the metasediments and no trends with regard to grain size or other elements are discernable.

v) Sulphur and Arsenic:

S and As are present in the metasediments in very low concentrations. No depletion or enrichment trends are evident for either element, and although the location of these elements within the metasediments is unknown, they are almost certainly held in the fine grained matrix.

c) Summary and Conclusions:

The characteristics of the Greenland Group noted by Nathan (1976) i.e. $K_2O > Na_2O$, $CaO < 1.5\%$ and $Rb > Sr$ hold for all samples analysed. A negative correlation between $SiO_2$ and $Al_2O_3$ also holds, and reflects the variation in abundance of clay minerals and quartz between argillites and greywackes.

The metasediments show concentration of Zr in the greywackes, whereas Cu, Pb and Zn are preferentially enriched in the argillites, indicating their presence in the matrix of the metasediments. Y, Nb, Sn, Sb, As and S show no significant enrichment in the metasediments, nor any depletion or enrichment trends with either grain size or other elements.

5.6 GREENLAND GROUP HORNFELSES

a) Major Elements:

The hornfelses of the pendant preserved on the southern slopes of Mount Rangitoto have undergone significant chemical changes during contact metamorphism. The hornfelses are characterised by a lower $SiO_2$ content (55 - 64\%) than the metasediments, and $CaO$ may be greater than 1.5\% (c.f. the Greenland Group metasediments). Like the metasediments $K_2O > Na_2O$, however, the differences between $K_2O$ and $Na_2O$ are generally greater in the hornfelses. Sulphur may show significant enrichment, particularly where sulphides have formed in the hornfelses (e.g. 10193).
Accompanying the loss in $\text{SiO}_2$ (5 - 6%) during hornfelsing of the metasediments, is a gain in $\text{Al}_2\text{O}_3$ ($\approx 3 - 4\%$) and $\text{K}_2\text{O}$ ($\approx 2\%$) and lesser gains in Fe-Oxide ($\approx 1.3\%$), MgO ($\approx 1.2\%$), MnO and CaO.

$\text{Na}_2\text{O}$, TiO$_2$ and $\text{P}_2\text{O}_5$ show no significant changes during hornfelsing.

These changes in major element geochemistry are difficult to interpret in view of the widespread effects of retrogressive metamorphism, particularly the widespread development of secondary sericite. Prograde metamorphism of the metasediments, consisting of a fine grained chlorite-sericite matrix supporting quartz and minor sodic feldspar, probably occurred by some or all of the following reactions (from Winkler, 1967):

1. $3 \text{Muscovite} + 5 \text{prochlorite} \rightarrow 3 \text{biotite} + 4 \text{Al-chlorite} + 7 \text{quartz} + 4 \text{H}_2\text{O}$
2. $\text{Clorite} + \text{muscovite} + \text{quartz} \rightarrow \text{cordierite} + \text{biotite} + \text{H}_2\text{O}$
   (If excess muscovite and quartz occur chlorite will disappear completely, and muscovite will remain as a residual mineral phase.)
3. $\text{Muscovite} + \text{quartz} \rightarrow \text{K-feldspar} + \text{andalusite} + \text{H}_2\text{O}$
4. $6 \text{Muscovite} + 2 \text{biotite} + 5 \text{quartz} \rightarrow 3 \text{cordierite} + 8 \text{K-feldspar} + 8 \text{H}_2\text{O}$

The reaction of equation (1) occurs early in the prograde metamorphism, prior to the onset of hornblende-hornfels facies conditions, the subsequent reactions occurring at or near the onset of these conditions (i.e. 500 - 600°C at 1 - 2 kbars). As can be seen from equation (1) quartz is produced by the reaction of chlorite and muscovite and could be lost to the rock along with metamorphic fluids, thus accounting for the lower SiO$_2$ content of the hornfelses.

Reactions (2) - (4) as outlined, also occur during prograde metamorphism, probably with little gain or loss of components, resulting in the disappearance of chlorite and the formation of biotite, cordierite and andalusite, and possibly K-feldspar.

The retrograde reactions (i.e. from RHS to LHS), however, all evolve quartz + muscovite (or sericite) at the expense of...
biotite and the Al$_2$SiO$_5$ polymorphs, andalusite and cordierite. The evolved quartz could readily be lost from the hornfelses during this retrogressive metamorphism. Accordingly it is considered that the lower SiO$_2$ content of the hornfelses, relative to the metasediments, is probably due to widespread effects of retrogressive metamorphism, preceded by a possible SiO$_2$ loss accompanying biotite formation, relatively early in the prograde metamorphic cycle.

The addition of Al$_2$O$_3$, K$_2$O, Fe-Oxides, MgO, MnO and CaO to the hornfelses cannot be accounted for by increased relative abundances due to a net silica loss of 5 - 6% from the hornfelses, the increases being too great to be explained by this mechanism alone. It is therefore apparent both from the geochemical data and the occurrence of both pyrrhotite-bearing and tourmalinised hornfelses in the pendant, that metasomatic alteration has accompanied hornfelsing of the metasediments. The nature of these metasomatic processes is obscured by the widespread retrogressive effects in the pendant. It seems unlikely, however, that appreciable addition of any of the above components occurred during the retrograde event, and therefore, it is considered that the pendant rocks have remained isochemical, with regard to addition of elements, since emplacement of the underlying Rangitoto Granite.

Alkali metasomatism penecontemporaneous with emplacement of the granite, has resulted in a widespread enrichment of K$_2$O and CaO, the geochemically similar MgO and Al$_2$O$_3$, and the transition element oxides Fe$_2$O$_3$ and MnO. These elements are all relatively mobile in hydrothermal fluids and their metasomatic addition to the hornfelses is not unexpected. These elements would have combined with SiO$_2$, either excess quartz present in the hornfelses or metasomatically added SiO$_2$, to produce biotite (K (Mg, Fe) (AlSi$_3$) O$_{10}$ (OH, F)$_2$), andalusite and cordierite (Al$_2$SiO$_5$) and in the absence of chlorite, muscovite (K Al$_2$ (AlSi$_3$) O$_{10}$ (OH, F)$_2$). Ca, Al, Fe and Mg are all constituents of tourmaline which also has a metasomatic origin.
Minor tourmalinisation of the hornfelsed pendant is widespread, tourmaline occurring in trace amounts in many of the thin sections examined. Locally, however, the pendant is almost completely tourmalinised, the hornfelses being altered to tourmaline (schorl) and quartz. Addition of Fe, Mg, Ca, Al, B and the trace elements F, Cu, Zn, Sn and As occurred as a result of tourmalinisation and was accompanied by a loss in K (see Table 5.1, sample 10194). This lost K may have been added to the more generally alkali metasomatised rocks of the pendant, which show an increase in K due to metasomatism.

Tourmalinisation associated with granite emplacement is usually regarded as a late stage pneumatolytic alteration, probably post-dating the thermal peak associated with granite emplacement. Volatile enriched fluids, or gases, residual from the crystallising or solid magma, resulted in this case, in both local tourmalinisation, and the more widespread 'alkali' metasomatism noted previously.

Retrogressive metamorphism of the metasomatically altered hornfelses is characterised by the breakdown of andalusite, biotite, cordierite, and possibly K-feldspar if sufficiently high temperatures were attained, resulting in the quartz-muscovite (or sericite)-biotite assemblages observed (see petrology, section 5.2). This alteration requires an addition of water, and was accompanied by a loss of SiO₂. In contrast the concentrations of the metasomatically introduced K₂O, CaO, Al₂O₃, MgO, FeO and MnO have been negligibly affected by this retrogression, no loss of these elements occurring during cooling, or subsequent but less intense re-heating due to intrusion of the Kakapotahi Granite, being evident.

The occurrence of disseminated pyrrhotite in the hornfelses in the east branch of Mine Creek (G.R.S64/366195) is also attributed to high temperature metasomatic processes. This alteration is described and discussed in Chapter 8.

b) Trace Elements:
   i) Rubidium, Strontium and Barium:

   Rb and Sr both occur in greater concentrations in the
In the hornfelses than in the metasediments, although the trend $\text{Rb} \gg \text{Sr}$ noted by Nathan (1976) for the Greenland Group still holds. In contrast Ba shows no significant enrichment.

A $\text{K}_2\text{O}$ vs Rb plot (see Fig 5.14(a)) shows a linear trend of increasing Rb with increasing $\text{K}_2\text{O}$ suggesting that Rb replaces K in both biotite and muscovite, and was probably also introduced into the hornfelses along with K, and the other mica forming components Al, Fe and Mg during metasomatism.

Sr, shows no significant enrichment in the hornfelses compared with the metasediments from Mount Greenland. A plot of $\text{K}_2\text{O}$ vs Sr (see Fig 5.14(b)), including the data of Nathan (1976), suggests that little or no Sr enrichment occurs during metasomatism. Tourmalinisation and sulphur metasomatism accompanied by a loss in K from the rocks, is accompanied by a significant enrichment in Sr, however.

Ba shows no increase with $\text{K}_2\text{O}$, (see Fig 5.12(a)), indicating a general lack of Ba addition during metasomatism. Redistribution of existing Ba from the matrix of the metasediments into metamorphic micas formed during hornfelsing and metasomatism has probably occurred.

ii) Yttrium, Niobium and Zirconium:

Y and Nb show no evidence of enrichment or depletion during the metamorphic event.

Zr shows a weak depletion trend (see Fig 5.12(b)) in the hornfelses, probably due to resorption of detrital zircon during metamorphism and metasomatism.

iii) Copper and Zinc:

Both Cu and Zn are enriched in the hornfelses, probably by addition during metasomatism. Plots of $\text{Al}_2\text{O}_3$ vs Cu and Zn (see Figs 5.13(a) and (b)) show a linear trend for $\text{Al}_2\text{O}_3$ vs Cu in the metasediments, and a somewhat more diffuse trend in the hornfelses. The relationship between $\text{Al}_2\text{O}_3$ and Cu, indicates that Cu, like Zn is contained within the micas of the hornfelses. Tourmalinisation and sulphur metasomation are accompanied by additional Cu enrichment.
Fig 5.14 (a) K2O vs Rb. A linear trend of increasing Rb with increasing K2O is evident for both the metasediments and hornfelses. A slight enrichment in both K2O and Rb is evident due to hornfelsing.

(b) K2O vs Sr. Sr shows no significant enrichment during hornfelsing and metasomatism.
Zn as would be expected follows Cu and is enriched in the hornfelses, although the correlation with Al₂O₃ is more diffuse. Tourmalinisation is accompanied by Cu and Zn enrichment, however, sulphur addition during metasomatism, appears to be accompanied by a loss of Zn, and a gain in Cu.

iv) Lead, Tin and Antimony:

Pb is neither depleted nor enriched during metamorphism and metasomatism, but is probably redistributed among the mica minerals. Sulphur metasomatism producing pyrrhotite (sample 10193) is, however, accompanied by an increase in Pb.

Sb, in contrast, is depleted during metamorphism and metasomatism, probably a consequence of the relatively high temperatures attained during metamorphism coupled with the high mobility of Sb.

Enrichment of Sn is only slight in the alkali- and sulphur-metasomatised hornfelses, but significant enrichment accompanies tourmalinisation, as would be expected. This enrichment is due to concentration of Sn in volatile rich, fluorine and boron bearing solutions, Sn having a strong affinity for these elements. The Sn is combined with silicates viz. tourmaline, no discrete Sn-bearing minerals having been observed.

v) Sulphur and Arsenic:

Enrichment or depletion trends of S are difficult to determine due to the generally low abundance of S in the rocks, and the relatively high limits of detection (20ppm). A slight enrichment of S, however, may accompany the widespread alkali metasomatism. Sulphur metasomatism, in contradistinction, adds significant quantities of S (up to 0.75%) to the hornfelses, the S combining with Fe to form pyrrhotite and pyrite. No S addition accompanies tourmalinisation.

As was only detectable in two hornfelsed samples (10193 and 10194), these samples having undergone sulphur metasomatism and tourmalinisation, respectively. No significant enrichment of As has occurred as a response to the more widespread alkali metasomatism. The high As content of sample 10199 (metasediment) cannot be readily explained.
c) **Summary and Conclusions:**

The enrichment and depletion trends discussed above are summarised in Table 5.3. Alkali metasomatism probably a late stage pneumatolytic process related to intrusion of the Rangitoto Granite, accompanied hornfelsing of the Greenland Group metasediments. Addition of $\text{Al}_2\text{O}_3$, $\text{K}_2\text{O}$, $\text{Fe}$-oxides, $\text{MgO}$, $\text{MnO}$, and $\text{CaO}$ occurred at this time, the elements reacting with excess quartz in the metasediments to form biotite, muscovite and the $\text{Al}_2\text{SiO}_5$ polymorphs, andalusite and cordierite. The trace elements Cu, Zn, Sn and Rb, and possibly S, were also introduced into the hornfelses at this time, these elements probably being located in the metamorphic micas. Loss of Zr, Sb and As also occurred due to metasomatism, probably reflecting resorption of detrital zircons, and the high mobility of Sb and As in hydrothermal fluids.

Retrogressive metamorphic effects are widespread and probably reflect a re-heating of the hornfelses associated with intrusion of the Kakapotahi Granite. Addition of $\text{H}_2\text{O}$ and loss of $\text{SiO}_2$ (5 - 6%) were the dominant geochemical changes associated with the retrogression, resulting in the observed mineral assemblage quartz-muscovite (or sericite)-biotite.

Locally, tourmalination is superimposed on the hornfelsing and alkali metasomatism of the roof pendant. Tourmalination is characterised by a loss of $\text{K}_2\text{O}$, a large gain in Sn and Fe-oxides, a slight gain in $\text{Na}_2\text{O}$ and loss of Rb. In spite of the large increase in Sn no discrete Sn-bearing minerals have been observed. Increases in the transition metals Cu and Zn, as well as As, also occur due to tourmalisation.

Discussion of the geochemical changes during sulphur metasomatism associated with disseminated pyrrhotite mineralisation is reserved for Chapter 8.

Of the economically interesting elements analysed (viz. Cu, Zn, Sn, Pb, As, S), Cu, Zn and to a lesser degree Sn and S, show enrichment during hornfelsing and metasomatism, Sb and As show depletion and Pb is relatively unaffected.
<table>
<thead>
<tr>
<th>HORNFELSEN + ALKALI METASOMATISM</th>
<th>TOURMALINISATION</th>
<th>SULPHUR METASOMATISM</th>
<th>RETROGRESSIVE METAMORPHISM</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>GAINED</strong></td>
<td><strong>LOST</strong></td>
<td><strong>STABLE</strong></td>
<td></td>
</tr>
</tbody>
</table>
| \( \text{Al}_2\text{O}_3 \) (3-4\%, \( \text{K}_2\text{O} \) (\approx 2\%)
| Fe-oxide (\approx 1.3\%)
| \( \text{MgO} \) (\approx 1.2\%)
| \( \text{MnO}, \text{CaO} \) | \( \text{SiO}_2 \)
| \( \text{K}_2\text{O} \) | \( \text{SiO}_2 \)
| \( \text{K}_2\text{O} \) | \( \text{H}_2\text{O} \) | \( \text{SiO}_2 \) (5-6\%)
| **MAJOR OXIDES**                 | **MAJOR OXIDES** | **MAJOR OXIDES**     |                          |
| \( \text{Na}_2\text{O} \)       | \( \text{Na}_2\text{O} \) | \( \text{Na}_2\text{O} \) |                          |
| \( \text{TiO}_2 \)              | \( \text{TiO}_2 \) | \( \text{TiO}_2 \) | No significant changes in major elements (excl. \( \text{SiO}_2 \))
| \( \text{P}_2\text{O}_5 \)      | \( \text{P}_2\text{O}_5 \) | \( \text{P}_2\text{O}_5 \) |                          |
| **GAINED**                       | **LOST**       | **STABLE**           |                          |
| \( \text{Rb} \)                 | \( \text{Rb} \) |                           |                          |
| \( \text{Cu}, \text{Zn} \)      | \( \text{Sr} \)
| \( \text{Sn} \) (slight)        | \( \text{Cu}, \text{Zn} \)
| \( \text{S} \) (slight)         | \( \text{Sn} \) (high)
| \( \text{As} \) (slight)        | \( \text{As} \) (high)
| **MINOR ELEMENTS**              | **MINOR ELEMENTS** | **MINOR ELEMENTS**     |                          |
| \( \text{Zr} \) (slight)        | \( \text{Rb} \) | \( \text{Zn} \) |                          |
| \( \text{As} \) ?               | \( \text{Sb} \) |                           |                          |
| **STABLE**                       | **STABLE**     | **STABLE**           |                          |
| \( \text{Ba} \)                 | \( \text{Ba} \) |                           |                          |
| \( \text{Pb} \)                 | \( \text{Pb} \) |                           |                          |
| \( \text{Y}, \text{Nb} \)       | \( \text{S} \) |                           |                          |

Table 5.3: Geochemical changes occurring during hornfelsing, metasomatism and retrogressive metamorphism of the Greenland Group metasediments. Question marks indicate possible gains or losses. See text for discussion.
These trends suggest that the magmatic metasomatic fluids were slightly enriched in both Cu and Zn, and that these elements could be expected in vein deposits associated with the Rangitoto Granite. Furthermore, Sb and As could be expected in vein deposits some distance from the pendant, where deposition occurs as a response to cooling. The Sb and As in the vein deposits of Cedar Creek, Mount Greenland, related to the Rangitoto Granite, is possibly derived, from metasomatism of the pendant rocks above the underlying granite, as well as from the granite body itself.

The origin of the metals in the vein deposits of both Mine Creek and Cedar Creek is discussed in terms of genetic models for the deposits in Chapter 13.

Tourmalinisation of the metasediments, although resulting in a significant Sn enrichment is not accompanied by the formation of Sn-oxides or sulphides, the tin ores.

No significant loss or gain of gold or silver has accompanied the hornfelsing and metasomatism of the metasediments (see Chapter 7).

5.7 CONTACT METAMORPHISM OF THE GREENLAND GROUP

P-T IMPLICATIONS OF THE AUREOLE ASSEMBLAGE

The characteristic mineral assemblage of the Greenland Group hornfelses of the pendant on the southern slopes of Mount Rangitoto, excluding retrogressive minerals, is as follows:

quartz-biotite-muscovite-oligoclase-andalusite
+ cordierite

Locally, sillimanite is present in the assemblage. Xenoliths of Greenland Group hornfels included in the Rangitoto Granite, show varying degrees of alteration, the highest grade metamorphic assemblage observed being quartz-biotite-oligoclase-microcline, in partially granitised xenoliths. No sillimanite or andalusite was identified in the xenoliths.
Subsequent retrogressive metamorphic effects have modified the mineral assemblages present, particularly in terms of sericitisation of the aluminium silicates.

The prograde mineral assemblage of the pendant is characteristic of the hornblende-hornfels facies of contact metamorphism as described by Turner and Verhoogen (1960). They record the following mineral assemblages in this facies:

A. Pelitic Rocks
   - Quartz-muscovite-biotite-cordierite (-plagioclase)
   - Quartz-muscovite-andalusite-cordierite (-plagioclase)

B. Quartzo-feldspathic
   - Quartz-plagioclase-microcline-muscovite-biotite
   - Quartz-plagioclase-microcline-biotite-hornblende

The P-T conditions of hornblende-hornfels facies metamorphism defined by Turner and Verhoogen (1960, p.519) range from 450°C at 600 bars $P_{H_2O}$, to 750°C at 2.9 kbars $P_{H_2O}$.

The prograde mineral assemblage present in the hornfelses of the pendant has been plotted on Fig 5.15. The abundance of muscovite in the hornfelses and the general absence of sillimanite provides an upper estimate of temperature during metamorphism of c.575° - 625°C. The presence of probable cordierite porphyroblasts now altered to sericite + quartz, suggests a minimum likely temperature of hornfelsing of C.520°C, although Turner (1968) extends the stability field of cordierite as low as 450°C at 1 kbar.

The pressures of metamorphism are not well known, due primarily to the uncertainty in location of the triple point of the $A_2SiO_3$ polymorphs (see Fig 5.15). The data of Fyfe (1967) indicate that andalusite is the stable polymorph below pressures of c.2.5 kbars, although the data of Holdaway (1971) suggest that andalusite is stable at pressures of up to 3.5 kbars. Taking the perhaps more generally accepted data of Fyfe (1967), the aureole assemblage i.e. quartz-muscovite- biotite-andalusite-cordierite suggests that pressures during the contact metamorphism were less than 1.6 kbars. Assuming a pressure gradient of 250 bars per kilometre a maximum depth of burial of c.6km is indicated during hornfelsing.
Phase diagram showing the stability fields of the metamorphic minerals in the aureole assemblage. The stability field of the pendant assemblage, (Mt Rangitoto) viz. Mu + Cd + Q + A is shaded, using the data of Fyfe (1967). Temperatures and pressures of metamorphism of 525-575°C and 0.5-1.8 kbars are indicated.

See text for further discussion.
Geochemical analyses of the Rangitoto and Kakapotahi granites, the petrology of which is discussed in sections 5.3 and 5.4, are given in Table 5.4, together with CIPW normative compositions. The geochemical characteristics of the granites are summarised in Table 5.5. Compared with the Kakapotahi Granite, the Rangitoto Granite is SiO₂ (69 - 72%), Na and K poor, and is enriched in the transition metals Fe, Mn, Ti and Cu, as well as Ca and Mg. Ba and Sr are both present in higher concentrations in the Rangitoto Granite than in the Kakapotahi Granite, with Ba > Sr > Rb. In contrast, due to lower Ba and Sr contents, Ba < Rb ≫ Sr in the Kakapotahi Granite. Pb and Ag also show significant enrichment in the Kakapotahi Granite, with Ag being enriched both relative to the Rangitoto Granite and the average for acid intrusive rocks (see section 7.4). The differences in geochemistry between the granites is readily explainable in terms of the observed mineralogy and the normative compositions of the granites.

Calculation of the CIPW norm:

Before discussing the geochemistry of the granites it should be noted that the norm calculations based on the geochemical analyses, do not include biotite, even though this is a significant component of both the Rangitoto (10 - 15%) and the Kakapotahi (3 - 4%) granites. Fe, Mg and the transition metals assigned to magnetite, ilmenite and hypersthene in the norm, in fact reside in the granites almost entirely in biotite and chlorite. The assignment of the transition metals to Fe-oxides and hypersthene results in a surplus of Al₂O₃ in the rock, which appears in the norm as corundum. K associated with the transition metals in the micas is included as orthoclase in the norm, and the silica required to form micas is included as normative quartz.

This method of calculating the norms, therefore, will give slightly high values of normative quartz and orthoclase, and corundum, hypersthene, magnetite and ilmenite will appear in the norm as significant components although they
TABLE 5.4: Geochemical analyses and normative compositions of the Kakapotahi and Rangitoto Granites.

Sample numbers are Canterbury University XRF No’s. Sample locations and descriptions are given in Appendix II. Major oxides quoted in wt %. All trace elements quoted in ppm. Au and Ag analysed by A.A. spectroscopy.

For the purpose of calculating the CIPW norms, an Fe₂O₃/FeO ratio of 0.4 was arbitrarily selected for all granitoid rocks.
<table>
<thead>
<tr>
<th></th>
<th>Rangitoto Granite</th>
<th>Kakapotahi Granite</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>69 - 72%</td>
<td>74 - 75%</td>
</tr>
<tr>
<td>TiO₂</td>
<td>&gt;0.2</td>
<td>&lt;0.1</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>&gt;14.5</td>
<td>&lt;14.5</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>&gt;1.6</td>
<td>&lt;0.8</td>
</tr>
<tr>
<td>MnO</td>
<td>0.05 - 0.08</td>
<td>0.04 - 0.07</td>
</tr>
<tr>
<td>MgO</td>
<td>&gt;0.8</td>
<td>&lt;0.3</td>
</tr>
<tr>
<td>CaO</td>
<td>&gt;1.9</td>
<td>&lt;1.1</td>
</tr>
<tr>
<td>Na₂O</td>
<td>&lt;3.6</td>
<td>&gt;3.7</td>
</tr>
<tr>
<td>K₂O</td>
<td>&lt;4.6</td>
<td>&gt;4.6</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.10</td>
<td>0.02</td>
</tr>
</tbody>
</table>

**Trace Elements**

<table>
<thead>
<tr>
<th></th>
<th>Rangitoto Granite</th>
<th>Kakapotahi Granite</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ba&gt; Rb&gt; Sr</td>
<td>&lt;40ppm</td>
<td>&gt;40ppm</td>
</tr>
<tr>
<td>Y</td>
<td>&gt;90ppm</td>
<td>&lt;70ppm</td>
</tr>
<tr>
<td>Zr</td>
<td>10-20ppm</td>
<td>&lt;1-2ppm</td>
</tr>
<tr>
<td>Pb</td>
<td>40-70ppm</td>
<td>80-100ppm</td>
</tr>
<tr>
<td>Ag</td>
<td>&lt;0.07ppm</td>
<td>&gt;0.3ppm</td>
</tr>
</tbody>
</table>

**TABLE 5.5**: Geochemical Characteristics of the Rangitoto and Kakapotahi Granites. (See text for discussion)
occur in the granites only in negligible amounts. Furthermore, the greater the abundance of biotite in the sample the greater will these components be in the norm. Accordingly, over-estimation of the abundance of quartz and orthoclase will be greater for the biotite-rich Rangitoto Granite than the Kakapotahi Granite.

The comparative mineralogy of the Rangitoto and Kakapotahi granites, determined from the normative compositions, qualitative thin section examination, and staining of cut slabs is given in Table 5.6. The differences in mineralogy observed between the two granites is reflected in the geochemistry, which can in large part be explained by the mineralogical differences.

a) Major Elements:

Relative to the Kakapotahi Granite, Rangitoto Granite is SiO₂ (69 - 72%), K₂O (< 4.6%) and Na₂O (< 3.6%) poor, reflecting the lower abundance of modal quartz, microcline and sodic plagioclase in the latter. As would be expected CaO (> 1.9%) is higher in the Rangitoto Granite than the Kakapotahi Granite, and reflects both the greater abundance and more calcic nature of the plagioclase present (see Table 5.4). The Ab/An ratio of the Rangitoto Granite (< 4.1) is accordingly lower than that of the Kakapotahi Granite (> 6.3).

On a K₂O - Na₂O - CaO diagram, the Rangitoto, and Kakapotahi granites lie in separate fields, the Rangitoto Granite being more calcic and less potassic than the Kakapotahi Granite (see Fig 5.16). Although this separation could be attributed to crystal fractionation processes, the probable difference in age between the granites (see section 6.5) and the field relationships suggest that this is unlikely. It therefore seems most probable, that the granites have crystallised from melts of differing composition, the Rangitoto Granite from a melt with a lower Ab/An ratio than that from which the Kakapotahi Granite crystallised. This suggestion is supported by the presence of andesine (An₄₀) in the
### Comparative modal mineralogy of the Rangitoto and Kakapotahi Granites based on qualitative thin section examination, normative composition and examination of cut slabs.

Only the most common minerals are included, and minerals present as alteration products are omitted.

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Rangitoto Granite</th>
<th>Kakapotahi Granite</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quartz</td>
<td>25 - 27%</td>
<td>30 - 33%</td>
</tr>
<tr>
<td>Microcline</td>
<td>20%</td>
<td>30%</td>
</tr>
<tr>
<td>Plagioclase</td>
<td>(An&lt;sub&gt;40-5&lt;/sub&gt;) 40%</td>
<td>(An&lt;sub&gt;35-5&lt;/sub&gt;) 30 - 35%</td>
</tr>
<tr>
<td>Biotite</td>
<td>10 - 15%</td>
<td>3 - 4%</td>
</tr>
<tr>
<td>Muscovite</td>
<td>-</td>
<td>1 - 2%</td>
</tr>
<tr>
<td>Chlorite</td>
<td>Tr (alteration of biotite)</td>
<td>Tr (alteration of biotite)</td>
</tr>
<tr>
<td>Garnet</td>
<td>-</td>
<td>Tr</td>
</tr>
</tbody>
</table>

**TABLE 5.6**
Fig 5.16 $K_2O$ - $Na_2O$ - $CaO$ diagram. The Rangitoto Granite shows a more calcic and less potassic character than the Kakapotahi Granite, due to the greater abundance and more calcic nature of the plagioclase present in the Rangitoto Granite. See text for further discussion.
Rangitoto Granite, whereas the most calcic feldspars in the Kakapotahi Granite are oligoclase (An<35).

Inclusion and assimilation of Greenland Group xenoliths in the Rangitoto Granite could not account for the more calcic nature of the Rangitoto Granite, the CaO content of the xenoliths (<1.7%) being less than that of the Rangitoto Granite (>1.9%). Assimilation of such low Ca-bearing xenoliths would tend to dilute the concentration of CaO in the Rangitoto Granite rather than increase it. The existence of abundant partially and completely assimilated xenoliths in the Rangitoto Granite therefore suggests that the parent magma may have been more calcic in nature than the geochemical analyses indicate. Possible An-Ab compositions of the parent magmas of the two granites, based on phase relations from Bowen (1928) for the system albite-anorthite are as follows:

\[
\begin{align*}
\text{Rangitoto Granite} & : \text{An}_8 \text{Ab}_{82} \\
\text{Kakapotahi Granite} & : \text{An}_5 \text{Ab}_{95}
\end{align*}
\]

The oxides of Ti, Fe, Mg and Mn are all more abundant in the Rangitoto Granite than the Kakapotahi Granite. These elements are all contained in biotite in significant concentrations, although undoubtedly some Ti occurs as rutile in both granites, and some Mg, Fe and Mn occurs in garnet in the Kakapotahi Granite. The enrichment of these transition metals in the Rangitoto Granite reflects the abundance of biotite (10 - 15%) in the granite, relative to the Kakapotahi Granite, which contains less than 5% biotite. These transition elements are abundant in the Greenland Group hornfelses where they occur in concentrations greater than in the Rangitoto Granite (see Table 5.1). Assimilation of xenoliths of hornfelsed Greenland Group within the Rangitoto Granite could, therefore, have enriched the Rangitoto Granite in Fe, Ti, Mg and Mn. It is unlikely, however, that sufficient assimilation of the Greenland Group occurred to account for all of the differences between the Kakapotahi and Rangitoto granites, and it is accordingly suggested that the parent
Rangitoto magma was enriched in the transition metals, relative to the Kakapotahi Granite, prior to assimilation of the Greenland Group country rocks.

b) Trace Elements:

Trace element analyses of 143 Westland granitoids were undertaken by Hoggarth (1977) whose data is summarised in Table 5.7. Although this data includes granites of both Rangitoto (Cretaceous) and Tuhuan (Paleozoic) age, it provides a useful measure of the background concentrations of many of the elements analysed in this study, viz. Ba, Sr, Rb, Zr, Y, Nb, Pb, Cu, Zn, Sn. Comparison of Hoggarth's data with that obtained in this study indicates that the Kakapotahi Granite contains below average concentrations of Ba, Sr, Zr, and Cu; average concentrations of Rb, Y, Nb and Zn; and higher than average concentrations of Pb and Sn. The Rangitoto Granite, in contrast, contains below average concentrations of Zr; slightly below average concentrations of Ba, Sr, Y, Nb and Cu; average concentrations of Rb and Zn; and above average concentrations of Pb and Sn.

Like the major element geochemistry, the trace element geochemistry indicates that the Rangitoto and Kakapotahi granites almost certainly formed from different magmas with distinct geochemical characteristics.

i) Barium, Strontium and Rubidium:

Ba (265 - 367ppm) occurs in greater concentrations in the Rangitoto Granite than the Kakapotahi Granite (1196ppm), due either partly or wholly to the greater abundance of calcic plagioclase, the main Ba-bearing mineral, in the Rangitoto Granite. Biotite, which may also contain Ba substituting for K is also more abundant in the Rangitoto Granite than the Kakapotahi Granite, and may also account, in part, for the higher Ba content of the Rangitoto Granite.

In contrast to Ba, Rb is more abundant in the Kakapotahi Granite (250 - 329ppm), than the Rangitoto Granite (170 - 232ppm). This undoubtedly reflects the greater abundance of K-feldspar, which commonly contains significant Rb, in the Kakapotahi Granite. Rb could also
### TABLE 5.7: Geochemical analyses of selected trace elements (ppm) from Westland granites and the Greenland Group.

Columns 1 - 8 inclusive from Hoggarth (1977)
Columns 9 and 10 this study.

1. All (143) granitic samples (Hoggarth, 1977)
2. Greenland Group (11 samples)
3. Buller Valley granites (57 samples)
4. Separation Point Batholith (9 samples)
5. Barrytown and Maybelle Bay granites (20 samples)
6. Mt Mantell granites (15 samples)
7. Deutgam Granite, Hohonu Range (9 samples)
8. Brunner Granite, Hohonu Range (6 samples)
9. Kakapotahi Granite (3 samples)
10. Rangitoto Granite (5 samples)

<table>
<thead>
<tr>
<th></th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
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<th>7</th>
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</thead>
<tbody>
<tr>
<td>Ba</td>
<td>429</td>
<td>472</td>
<td>437</td>
<td>495</td>
<td>329</td>
<td>466</td>
<td>631</td>
<td>336</td>
<td>42</td>
<td>306</td>
</tr>
<tr>
<td>Sr</td>
<td>271</td>
<td>122</td>
<td>283</td>
<td>621</td>
<td>81</td>
<td>344</td>
<td>462</td>
<td>43</td>
<td>22</td>
<td>163</td>
</tr>
<tr>
<td>Rb</td>
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<td>199</td>
<td>105</td>
<td>287</td>
<td>144</td>
<td>136</td>
<td>218</td>
<td>278</td>
<td>204</td>
</tr>
<tr>
<td>Zr</td>
<td>175</td>
<td>235</td>
<td>154</td>
<td>151</td>
<td>175</td>
<td>220</td>
<td>251</td>
<td>266</td>
<td>61</td>
<td></td>
</tr>
<tr>
<td>Y</td>
<td>53</td>
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<td>51</td>
<td>20</td>
<td>78</td>
<td>49</td>
<td>44</td>
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<tr>
<td>Nb</td>
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<td>16</td>
<td>8</td>
<td>18</td>
<td>22</td>
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<tr>
<td>Sn</td>
<td>12</td>
<td>5</td>
<td>13</td>
<td>12</td>
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<td>5</td>
<td>13</td>
<td>11</td>
<td>65</td>
<td>65</td>
</tr>
</tbody>
</table>

1. All (143) granitic samples (Hoggarth, 1977)
2. Greenland Group (11 samples)
3. Buller Valley granites (57 samples)
4. Separation Point Batholith (9 samples)
5. Barrytown and Maybelle Bay granites (20 samples)
6. Mt Mantell granites (15 samples)
7. Deutgam Granite, Hohonu Range (9 samples)
8. Brunner Granite, Hohonu Range (6 samples)
9. Kakapotahi Granite (3 samples)
10. Rangitoto Granite (5 samples)
Sr occurs in lower concentrations in the Kakapotahi Granite (9 - 43ppm) than in the Rangitoto Granite (130 - 175ppm), and almost certainly correlates with the lower Ca content of the Kakapotahi Granite.

These differences give rise to characteristic ratios of Ba, Sr and Rb in the two granites, the Rangitoto Granite having Ba > Rb > Sr, whereas in the Kakapotahi Granite the relative abundances are Ba < Rb > Sr. When plotted on the ternary diagram derived by Bouseily and Sokkary (1975) the Rangitoto and Kakapotahi granites plot as two distinct fields (see Fig 5.17). The Kakapotahi Granite has Ba : Rb : Sr ratios characteristic of a strongly differentiated granite, whereas the Rangitoto Granite in comparison is less strongly differentiated. The Rangitoto Granite plots outside the 'granite fields' of Bouseily and Sokkary, reflecting its slightly lower than average Ba content.

ii) Yttrium, Niobium and Zirconium:

The Kakapotahi Granite (45 - 86ppm) is enriched in Y relative to the Rangitoto Granite (27 - 35ppm). The presence of garnet within the Kakapotahi Granite almost certainly accounts for the increased Y abundance. Zr in contrast is most abundant in the Rangitoto Granite. The Y/Zr ratios of the two granites accordingly show marked differences being in the range 0.8 - 1.5 for the Kakapotahi Granite and 0.24 - 0.34 for the Rangitoto Granite.

The difference in the Y/Zr ratios of the two granites is unlikely to be due to crystal fractionation processes in the two magmas, neither of the elements being significantly concentrated in the crystallising silicate phases, excluding the Y-garnet association already noted. The difference in the ratios is therefore regarded as indicating probable crystallisation of the granites from two distinct magmas with differing Y and Zr concentrations.

There is no significant difference in Nb concentrations between the two granites.

iii) Copper and Zinc:

Cu occurs in the Kakapotahi Granite in barely detectable amounts (< 2ppm), however, the Rangitoto Granite...
Fig 5.17 Rb - Ba - Sr diagram (after Bouseily and Sokkary 1975), showing postulated differentiation trends and Ba : Rb : Sr ratios of different rock types.

The Kakapotahi Granite has Ba : Rb : Sr ratios characteristic of strongly differentiated granites. The Rangitoto Granite in contrast is less strongly differentiated with Ba : Rb : Sr ratios more typical of 'normal' or 'anomalous' granites.
in contrast, contains c.15ppm Cu, slightly below the average concentration of 23ppm obtained by Hoggarth (1977) for Westland granites. The extremely low levels of Cu encountered in the Kakapotahi Granite are difficult to explain. Cu generally occurs as a replacement for Ca and Na in plagioclase, and Fe in biotite. If most of the Cu present is in fact held in biotite, the biotite-poor character of the Kakapotahi Granite would explain the low Cu content. It seems most likely, however, that the Kakapotahi Granite crystallised from a Cu-poor melt, and that the Cu present occurs in both plagioclase and biotite.

Unlike Cu, Zn shows no enrichment or depletion in either of the granites, both having average Zn contents of c.49ppm, close to Hoggarth's (1977) average. Considerably more variation occurs in the Kakapotahi Granite than the Rangitoto Granite, although why this should be so is unclear.

iv) Lead, Tin and Antimony:

Both the Rangitoto and Kakapotahi granites contain Pb in concentrations above the average for Westland granites (see Table 5.7). This enrichment is attributed to high abundances of Pb in the magmas from which the granites crystallised. The Kakapotahi Granite shows the greatest enrichment (Pb = 75 - 97ppm), whereas the Rangitoto Granite has Pb contents in the range 45 - 63ppm.

The reasons for enrichment of Pb in these granites are unknown. It is possible, however, that both of the granites are part of a regional suite of Pb-enriched granites emplaced in the Westland foreland during the Paleozoic. Hoggarth (1977) determined the average Pb contents of Westland granites of Ordovician-Cretaceous age to be 38ppm, compared to the world average of 20ppm for felsic igneous rocks (Krauskopf, 1967, p.588) which indicates a possible regional enrichment.

The relative enrichment of Pb in the Kakapotahi Granite probably correlates with its more K<sub>2</sub>O and SiO<sub>2</sub> rich nature, Pb having a strong affinity to occupy K lattice sites in feldspars and micas. The low abundance of K<sub>2</sub>O and SiO<sub>2</sub> in the Rangitoto Granite correlates with a lower Pb concentration (see Fig 5.18).
Plot of Pb vs K$_2$O for the Kakapotahi and Rangitoto granites. The Kakapotahi Granite is enriched in both K$_2$O and Pb relative to the Rangitoto Granite, reflecting the more potassic nature of the Kakapotahi Granite, and the tendency for Pb to occupy K lattice sites in the K-feldspars and mica minerals.

Both granites are Pb-enriched compared to the world average for felsic igneous rocks (Pb = 20 ppm, Krauskopf 1967, p.588).
Sn occurs in identical concentrations in both the Rangitoto and Kakapotahi granites (c.65ppm) which is some 5 - 6 times higher than the values reported by Hoggart (1977) for Westland Granites. Under the classification scheme proposed by Flinter (1971) both granites are classified as 'stanniferous granitoids' i.e. contain greater than 10ppm Sn, the Sn being present either as discrete Sn-bearing minerals or as impurities in silicates.

The identical Sn contents of the Kakapotahi and Rangitoto granites is somewhat surprising at first glance, particularly in view of the differences in biotite abundances noted for the two granites. Biotite commonly contains significant quantities of Sn, probably as a replacement of Fe or Mg. In the case of the Rangitoto Granite the high Sn values are attributed to the abundance of biotite, which accounts for 10 - 25% of the rock.

The enrichment of Sn in the Kakapotahi Granite, is regarded as being due to the high degree of differentiation (see Fig 5.17) that the granite has undergone. Late stage residual melts, from which the Kakapotahi Granite has apparently crystallised, are commonly enriched in Sn, along with Y and Rb.

No search for discrete Sn-bearing minerals in either of the granites has been made.

Both the Rangitoto and Kakapotahi granites contain Sb concentrations of less than 10ppm. No significant enrichment or depletion trends have been noted, either within or between the granites.

v) Sulphur and Arsenic:

S occurs in both the Rangitoto and Kakapotahi granites in concentrations below the lower limit of detection (20ppm). A slight enrichment of S is, however, apparent in samples of the Rangitoto Granite collected from near the Mine Creek Fault (see section 8.2).

As, like S, occurs in both granites in concentrations below the lower limits of detection (5ppm). Accordingly, enrichment or depletion trends for both elements are unknown.
ORIGIN AND EMPLACEMENT OF THE GRANITES

The petrological and geochemical characteristics of the Rangitoto and Kakapotahi granites, discussed in the preceding sections, suggests that the two granites were derived from genetically unrelated magmas. This is supported by K-Ar dating (see Chapter 6) which indicates that the Rangitoto Granite was emplaced c.314 Ma ago (Carboniferous-Permian) and the younger Kakapotahi Granite was emplaced C.214 Ma ago, during the Triassic.

Mineralogically, the Rangitoto Granite differs from the Kakapotahi Granite, in that it contains a greater proportion of plagioclase (c.40%) which may be as calcic as An$_{40}$, less quartz (25 - 27%) and less microcline (c.20%) than the Kakapotahi Granite. Biotite also accounts for a significant proportion (10 - 15%) of the Rangitoto Granite. In contrast the Kakapotahi Granite contains plagioclase (30 - 35%), only as calcic as An$_{35}$, microcline (c.30%) and quartz (30 - 33%). Biotite occurs in greatly reduced amounts (3 - 4%) and may be accompanied by muscovite, and garnet is a common although minor component. These mineralogic differences are reflected in the major element geochemistry of the two granites (see Table 5.5), the Rangitoto Granite containing less SiO$_2$, Na$_2$O and K$_2$O, and more Fe-oxides, MgO, MnO, CaO, P$_2$O$_5$ and Al$_2$O$_3$ than the Rangitoto Granite.

Because of the presence of biotite in the granites, both contain hypersthene and corundum in the CIPW norm, and as would be expected from the mineralogy have differing abundances of Q-Ab-An-Or (see Table 5.4).

Trace element variations between the two granites perhaps provide the best geochemical evidence for the crystallisation of the granites from different magmas. The concentrations of Ba, Sr and Rb in the two granites (see Fig 5.17) indicates that the Kakapotahi Granite is a strongly differentiated granite, whereas the Rangitoto Granite is considerably less differentiated. Sn, Y, Rb and Pb, which commonly concentrate in residual melts during crystallisation, are all enriched in the Kakapotahi
Granite and provide further evidence for strong differentiation of the magma and crystallisation of the Kakapotahi Granite from a residual liquid. In contrast, the Rangitoto Granite is considerably less differentiated with only a slight depletion in Ba and Sr. Sn is enriched in the granite however, due almost certainly to the abundance of biotite, the main Sn-bearing mineral in granites, rather than to a high degree of differentiation. Pb as would be expected, is less abundant than in the Rangitoto Granite.

The markedly different Ag abundances of the two granites also suggests derivation from different magmas. Ag, which is considerably more mobile than Au, is enriched in the Kakapotahi Granite relative to the Rangitoto Granite (see Chapter 7), although the Au content of both granites is almost identical, Boyle (1966, 1979) notes that there is little evidence for concentration of either Au or Ag during differentiation in calc-alkali rocks. The high Ag content of the Kakapotahi Granite is, therefore, regarded as being indicative of crystallisation of a high-Ag magma. The Rangitoto Granite is similarly assumed to have crystallised from a relatively low-Ag magma.

From the data available, the origin of the two magmas is difficult to discern with any certainty. Work by Chappell and White (1974) on eastern Australian granitoids however resulted in erection of a simple classification scheme for granites, depending on their origin. Granites analysed by Chappell and White were classified either as I-type, derived from the partial melting of pre-existing igneous rocks, or S-type, derived from the partial melting of pre-existing sedimentary rocks. The criteria on which the classification was based are summarised in Table 5.8. The Rangitoto and Kakapotahi granites have characteristics of both I-type and S-type granites (see Table 5.9). Both granites have high Na₂O contents, considered by Chappell and White to indicate I-type granites, however, the ratio of \( \text{Mol Al}_2\text{O}_3 / (\text{Na}_2\text{O} + \text{K}_2\text{O} + \text{CaO}) \) for both granites is greater than 1.1 indicating that both are S-type granites.
### Table 5.8: Characteristics of S-type and I-type granites (after Chappell and White, 1974).

<table>
<thead>
<tr>
<th>RANGITOTO GRANITE</th>
<th>KAKAPOTAHI GRANITE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Na₂O from 3.20 - 3.55% for rocks with K₂O content of 3.91 - 4.53%</td>
<td>Na₂O from 3.75 - 4.38% for rocks with K₂O content of 4.63 - 5.11%</td>
</tr>
<tr>
<td>Mol. Al₂O₃/(Na₂O+K₂O+CaO) = 2.06 - 2.14</td>
<td>Mol. Al₂O₃/(Na₂O+K₂O+CaO) = 1.9 - 2.0</td>
</tr>
<tr>
<td>CIPW Normative corundum = 0.85 - 1.73 (ave = 1.2%)</td>
<td>CIPW Normative corundum = 0.73 - 1.15% (ave = 0.9%)</td>
</tr>
<tr>
<td>Garnet absent</td>
<td>Garnet present</td>
</tr>
<tr>
<td>Biotite 10 - 15%</td>
<td>Biotite 3 - 4% + muscovite</td>
</tr>
<tr>
<td>Average Cu, high Sn although neither types of mineralisation known</td>
<td>Low Cu, high Sn although neither types of mineralisation known</td>
</tr>
</tbody>
</table>

**I- TYPE**

Relatively high Na; Na₂O 3.2% in felsic varieties, decreasing to 2.2% in more mafic types

- Mol. Al₂O₃/(Na₂O+K₂O+CaO) < 1.1
- CIPW normative diopside or <1% normative corundum
- Hornblende common in mafic I-types
- Garnet generally absent
- Associated with W and porphyry Cu-Mo mineralisation

**S- TYPE**

Relatively low Na; Na₂O < 3.2% in rocks with c.5% K₂O, decreasing to 2.2% in rocks with c.2% K₂O.

- Mol. Al₂O₃/(Na₂O+K₂O+CaO) > 1%
- CIPW normative corundum
- Muscovite in felsic S-types, biotite in mafic S-types
- Garnet may be present
- Associated with Sn mineralisation

**TABLE 5.9**: Characteristics of the Rangitoto and Kakapotahi Granite considered to reflect either S-type or I-type origins (based on Chappell and White, 1974). See text for discussion.
The Rangitoto Granite contains a greater abundance of normative corundum (average of 5 samples = 1.2%) than the Kakapotahi Granite (average of 3 samples = 0.9%), the former indicating an S-type origin, the latter an I-type, although one value for the Kakapotahi Granite exceeds 1%. Accordingly either an S- or I-type origin may be inferred for the Kakapotahi Granite.

The occurrence of biotite in both granites indicates possible S-type origins, and the existence of muscovite and garnet in the Kakapotahi Granite further supports an S-type origin for the Kakapotahi Granite.

Consideration of all of Chappell and White's criteria applied to the Rangitoto and Kakapotahi granites, suggests that both granites are S-type, in spite of the high Na$_2$O contents and low contents of normative corundum in some samples. Accordingly, they are inferred to have formed by partial melting of sediments, possibly the Greenland Group, at depth. Melting is considered to have occurred as a response to a high heat flow above a subduction zone to the east of the Tasman Geosyncline.

Plots of the normative compositions of Q-Ab-Or for both granites are shown on the ternary diagram of Tuttle and Bowen (1958), for the system Q-Ab-Or-H$_2$O at P$_{H_2O} = 2$ kbars (see Fig 5.19). As can be seen from the figure the eutectic minimum is dependent on the Ab/An ratio of the rock, being deflected away from the Ab apex with decreasing Ab/An ratios. The eutectic minima for the Rangitoto Granite (Ab/An= 4.0) and the Kakapotahi Granite (Ab/An= 6.5) at P$_{H_2O} = 2$ kbars are designated, on the diagram, by points R and K respectively.

The composition of the eutectic minimum is also dependent on P$_{H_2O}$, being deflected towards the Ab apex with increasing pressure (see Fig 5.20).

The Kakapotahi Granite samples plotted lie close to the 'thermal valley' separating the Ab and Or fields, and indicate emplacement at temperatures approaching the eutectic minimum i.e. 700 - 750°C. A minimum pressure of crystallisation of P$_{H_2O} = 2$ kbars is estimated from Fig 5.19.
System Q – Ab – Or – H₂O projected onto the anhydrous base of the tetrahedron. The 2 kbar cotectic is plotted for Ab/An ratios of $\infty$, 7.8, 5.2, 3.8 and 1.8.

The Kakapotahi Granite samples plot close to the thermal valley separating the Ab and Or fields indicating emplacement at temperatures of 700° - 750°C, close to the eutectic minimum.

The Rangitoto Granite plots in the Ab field indicating possible crystallisation of the granite at PH₂O conditions exceeding 2 kbars.

(Diagram after Tuttle and Bowen, 1958).
Effects of pressure on the eutectic minimum in the system Q - Ab - Or - H₂O for Ab/An = ∞ (after Tuttle and Bowen, 1958). The trace of the eutectic minimum (E) with increasing pressure (dashed arrow) is shown on the diagram. The postulated trace of the eutectic minimum of the Rangitoto Granite (Ab/An = 4.0) with increasing pressure is shown as a dotted line. The Rangitoto Granite samples plot near the eutectic minimum at pressures P H₂O = 3 - 4 kbars, pressures which intuitively seem too high. See text for further discussion.
The Rangitoto Granite samples plot some distance away from the eutectic minimum predicted by the Ab/An ratio of the granite (see Fig 5.19) indicating probable crystallisation of the granite at \( P_{H_2O} \) conditions exceeding 2 kbars. The postulated trace of the eutectic minimum \( R \), for the Rangitoto Granite is shown in Fig 5.20 and indicates probable minimum crystallisation pressures of \( P_{H_2O} = 3.5 - 4 \) kbars. Temperatures of crystallisation at this \( P_{H_2O} \) would also have been close to the eutectic minimum i.e. c.700 - 750\(^{0}\).

The relationship of \( P_{H_2O} \) to \( P_{\text{total}} \) in the granites during crystallisation is unknown, and accordingly it is difficult to estimate likely depths of emplacement. As \( P_{H_2O} \) is unlikely to exceed \( P_{\text{total}} \) in a crystallising magma however, the above \( P_{H_2O} \) estimates for the Rangitoto and Kakapotahi granites may be regarded as minimum values of \( P_{\text{total}} \). Accordingly, minimum depths of emplacement, assuming a pressure gradient of 4km/kbar, based on the system Q-Ab-Or-H\(_2O\) are as follows:

- Rangitoto Granite - 14 - 16km
- Kakapotahi Granite - 8km

The 14 - 16km emplacement depth indicated for the Rangitoto Granite is almost certainly too great. The widespread occurrence of andalusite, the low pressure polymorph of \( Al_2SiO_5 \), in the contact aureole, and its generally low metamorphic grade, probably hornblend-hornfels facies, suggests that emplacement of the Rangitoto Granite probably occurred at shallower depths of 4 - 8km. Accordingly, 8km is regarded as a likely maximum depth of emplacement of the Rangitoto Granite.

The widespread retrogressive metamorphic effects noted in the contact aureole, and thought to be associated emplacement of the Kakapotahi Granite, similarly suggests that emplacement of this granite occurred at depths of less than 8km, rather than greater than 8km as suggested by the Q-Ab-Or diagrams. It is therefore considered most likely that emplacement of both the Rangitoto and Kakapotahi granites occurred at depths of 4 - 8km and
temperatures of 700 - 750°C. It is possible, however, that both granites reached higher levels before crystallising, perhaps as high as 2km depth.
6.1 INTRODUCTION

A small number of samples were selected for potassium-argon dating with the following objectives in mind:

a) To determine the age of the vein mineralisation at Mount Rangitoto, and its relationship to intrusive and hydrothermal activity in the area.

b) To determine the ages of the Rangitoto and Kakapotahi granites.

Nine samples were selected and analysed at the Institute of Nuclear Sciences, Lower Hutt.

6.2 PREVIOUS GEOCHRONOLOGICAL WORK

K-Ar and Rb-Sr dating studies, including samples of Greenland Group and/or Tuhuan granite, have been undertaken by Aronson (1965, 1968), Adams (1973, 1975), Adams et. al. (1974), Adams and Nathan (1977), and Mason (1961). Most of these studies have been regional in nature, with one or more of the following aims:

a) To determine the age of sedimentation of the Greenland Group.

b) To determine the age of emplacement of the Tuhuan granites in Nelson and Westland.

c) To determine the age of formation of the Constant Gneiss.

Aronson's studies, the most comprehensive to date, attempted to establish using mainly Rb-Sr isochron ages, a regional geochronology for the Paleozoic in Nelson and Westland.

6.2.1 Summary of Results

Greenland Group and Tuhua Group K-Ar and Rb-Sr ages
determined by previous workers, notably Adams et. al. (1975) and Aronson (1965, 1968) are summarised in Table 6.1 and Fig 6.1 respectively.

The granite ages determined by Aronson fall into two groups, interpreted by Aronson to-

"... indicate two major periods of plutonic and metamorphic activity in the Foreland Province, one during the late Mesozoic and one in the mid Paleozoic."

The younger dates, generally in the range 100 - 200 Ma have been correlated with the Rangitata Orogeny, and undoubtedly include samples of both granites intruded during the Rangitata Orogeny, for example the Separation Point batholith and the Buckland Granite stock (Adams and Nathan, 1978), whereas the less concordant dates possibly reflect thermal overprinting of older granites during the Rangitata Orogeny.

The older group of dates, in the range 280 - 380 Ma represent a period of mid-Paleozoic granite intrusion correlated with the Tuhua Orogeny.

Subsequent K-Ar dating by Adams (unpubl. report, 1973) revealed the existence of granite bodies pre-dating those reported by Aronson (1965). Three bodies were recognised, these being the Barrytown, Maybelle Bay and Windy Point stocks, the first with an age of about 390 Ma, and the last two with ages between 420 and 425 Ma.

These dates suggest the existence of three phases of granitic plutonism on the West Coast of the South Island, these being late Ordovician, Devonian-Carboniferous, and Jurassic-Cretaceous in age.

Adams et. al. (1975) noted that the ages obtained from the Greenland Group (see Table 6.1) fell into two groups:

a) Ages from 300 - 370 Ma.

These ages were interpreted as representing a thermal overprint associated with Tuhuan intrusive activity, particularly at about 300 Ma ago.
<table>
<thead>
<tr>
<th>SAMPLE NO.</th>
<th>ROCK TYPE, DESCRIPTION</th>
<th>LOCALITY</th>
<th>Kx</th>
<th>$^{40}\text{Ar}_n$/g</th>
<th>$^{40}\text{Ar}_r$</th>
<th>Age (Ma)</th>
</tr>
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<tbody>
<tr>
<td>NZKA 42</td>
<td>Slate</td>
<td>S37</td>
<td>3.08</td>
<td>39.93</td>
<td>96.9</td>
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<td></td>
<td>Metaseds. above Constant Gneiss</td>
<td>S37</td>
<td>3.57</td>
<td>63.94</td>
<td>95.8</td>
<td>404 ± 5</td>
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<td></td>
<td>Slate</td>
<td>S37</td>
<td>3.51</td>
<td>58.82</td>
<td>73.2</td>
<td>379 ± 5</td>
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<td></td>
<td>Slate</td>
<td>S37</td>
<td>3.01</td>
<td>55.23</td>
<td>60.6</td>
<td>412 ± 5</td>
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<td></td>
<td>Qtz-sericite-chlorite slate</td>
<td>S37</td>
<td>4.08</td>
<td>68.77</td>
<td>98.2</td>
<td>382 ± 6</td>
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<td></td>
<td></td>
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<td>49.07</td>
<td>97.6</td>
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<td></td>
<td></td>
<td>S44</td>
<td>3.33</td>
<td>52.60</td>
<td>91.0</td>
<td>360 ± 7</td>
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<td></td>
<td></td>
<td>S44</td>
<td>4.06</td>
<td>79.81</td>
<td>96.5</td>
<td>438 ± 7</td>
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<td></td>
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<td>S44</td>
<td>3.64</td>
<td>47.08</td>
<td>96.8</td>
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<td></td>
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<td>S44</td>
<td>3.29</td>
<td>42.36</td>
<td>97.4</td>
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<td></td>
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<td>24.29</td>
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<td></td>
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<td>S38</td>
<td>3.76</td>
<td>61.57</td>
<td>95.8</td>
<td>410 ± 8</td>
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<td>82*</td>
<td>Hornfelsed greywacke</td>
<td>S57</td>
<td>2.28</td>
<td>25.46</td>
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<td>261 ± 5</td>
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<tr>
<td>83*</td>
<td>Qtz-sericite-chlorite slate</td>
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<td>4.47</td>
<td>88.07</td>
<td>98.9</td>
<td>438 ± 7</td>
</tr>
<tr>
<td>84*</td>
<td></td>
<td>S57</td>
<td>4.49</td>
<td>81.09</td>
<td>99.1</td>
<td>407 ± 9</td>
</tr>
</tbody>
</table>

**TABLE 6.1**: Previously published K-Ar dates for Greenland Group (including Waiuta Group) sediments. * Indicates sample from Mt. Greenland massif.

Notes: 1. Grid References refer to NZMS1 1:63,360 series maps, Sample No's from Adams (1975)
2. Sources: Adams et. al. (1975), Adams (1975). $\lambda_f = 0.472 \times 10^{-3} \, \text{yr}^{-1}$, $\lambda_e = 0.584 \times 10^{-16} \, \text{yr}^{-1}$
Fig 6.1 Aronson's map showing Rb-Sr and K-Ar ages obtained by him for the rocks of the Western foreland (from Aronson, 1968).
b) Ages from 370 - 440 Ma.

This older group of ages is interpreted as representing metamorphism of the Greenland Group at least 440 Ma ago (Upper Ordovician) and suggests that the Greenland Group must be, at least, older than Upper Ordovician.

Aronson (1968, p.679) also noted periods of plutonism which correspond closely to the ages of Greenland Group metamorphism noted by Adams et. al. (1975).

6.3 PRINCIPLES OF THE K-Ar DATING METHOD

The principles of the K-Ar dating method have been described in some detail by Schaeffer and Zahringer (1966) and Dalrymple and Lamphere (1969), to whom the more interested reader is referred.

Briefly, the K-Ar dating method depends upon the decay of $^{40}$K in rocks to $^{40}$Ar, and the retention of this $^{40}$Ar through geological time.

Three naturally occurring isotopes of potassium exist, these being $^{39}$K, $^{40}$K and $^{41}$K, having abundances of 93.08%, 0.0119% and 6.91% respectively. Of the three only $^{40}$K is radioactive and has a half life of $1.31 \times 10^9$ years. The decay scheme is shown in Fig 6.2.

The fractions of $^{40}$K decaying along the $\beta$ and EC (electron capture) paths per unit time are given by the respective decay constants and:

$$\lambda_\beta = 0.581 \times 10^{-10} \text{ yr}^{-1}$$

$$\lambda_{ec} = 4.962 \times 10^{-10} \text{ yr}^{-1}$$

From the decay constants the branching ratio (R) can be calculated ($R = \frac{\lambda_{ec}}{\lambda_\beta}$) enabling the amount of $^{40}$K decaying to form $^{40}$Ca (89%) and $^{40}$Ar (11%) to be determined.

To make use of this method for age determination purposes, the original potassium content of the rock or mineral ($K_0$) must be known. This is related to the present $^{40}$K content ($K_t$) as follows:

$$K_0 = \frac{e^{-\lambda t}}{K_t}$$

$$\lambda = \lambda_{ec} + \lambda_\beta$$

$$t = \text{age of mineral or rock.}$$
The present argon content ($Ar_t$) in moles per mole of potassium is given by the following:

$$Ar_t = \frac{\lambda_{Ar}}{\lambda_{K} + \lambda_{\beta}} (K_0 - K_t)$$

The age of the rock or mineral ($t$) can be calculated, using the present abundances of argon and potassium in the mineral or rock, by the following formula:

$$t = \frac{1}{\lambda} \ln \left( 1 + \frac{R}{R} \cdot \frac{Ar_t}{K_t} \right)$$

Thus, by accurately determining the potassium and argon content of a rock or mineral sample, the age of that sample can be calculated.

Two conditions must, however, apply before the age of a mineral or rock can be determined:

1. No argon is present in the rock or mineral at the time of formation ($t_o$).
2. No argon has been lost from the sample since the time of formation ($t_o$).

These conditions in effect limit the minerals or rocks that can be reliably used for age determinations to micas (muscovite and biotite), hornblende, fine grained igneous
and volcanic rocks, and low grade metamorphic rocks.
feldspars, with the exception of the high temperature
K-feldspar sanidine, generally have low argon retentivities.
pyroxenes, as well as having low potassium concentrations
often contain excess argon. Accordingly neither yield
reliable dates, feldspars commonly giving young ages and
pyroxenes old ages.

6.4 METHOD

Nine samples - four granites, three Greenland Group
hornfelses, and two of mineralised vein material - were
collected for dating. Weathered surfaces were removed
from the samples and thin sections prepared of each to
check their suitability for dating. Suitable samples
were broken into chips, ground in a ring mill for a few
seconds, sieved - the 211\(\mu\) - 422\(\mu\) fraction retained and
then washed in distilled water. Mineral separates, where
required, were obtained using a Franz magnetic separator.

The samples or separates, thus prepared, were used
for both potassium and argon analysis. In both cases
sub-samples of 0.25 - 1.0g were removed using a mini
sample splitter, and then treated as follows:

a) Potassium Analysis:
The samples were dissolved in hydrofluoric and
nitric acid, mixed with boric acid and distilled water,
and then transferred into ion exchange columns where
they were eluted using nitric acid. The solutions
obtained were weighed and then analysed on a flame
spectrophotometer to determine the potassium content.

b) Argon Extraction and Analysis:
The procedure used for argon extraction and analysis
is similar to that described in some detail by Hulston
and McCabe (1972) and Adams (1975), although slight
modifications have been made to the equipment since
these publications.
Argon is extracted from the previously oven dried and weighed samples, in a bakeable glass system under an ultra-high vacuum. The sample, loaded into a quartz wool-lined molybdenum crucible is heated to 1600°C for 8 - 10 minutes, using an R.F. heating coil. Reactive gases evolved during fusion of the sample are removed from the system and the remaining argon sample, plus a $^{38}$Ar spike are then leaked into the mass spectrometer for analysis.

Peaks produced by $^{36}$Ar, $^{38}$Ar and $^{40}$Ar are scanned five times, using a chart recorder. The heights of the respective peaks are measured and the amount of radiogenic $^{40}$Ar present calculated. Having previously determined the potassium content of the sample, the age can be calculated.

**6.5 RESULTS AND INTERPRETATION**

The results obtained in this study are summarised in Table 6.2. The analytical errors quoted in the table were estimated using the equations of Cox and Dalrymple (1967), with the typical percentage errors (standard deviation) for the $^{38}$Ar spike quantity (1%), $^{40}$Ar/$^{38}$Ar and $^{36}$Ar/$^{38}$Ar ratio measurements (0.3% and 0.7% respectively) and the potassium content (0.5%), as determined by Adams (1975).

As can be seen from Table 6.2 age determinations based only on single minerals have been made for most samples, although for samples 9471 and 9458 two mineral fractions were dated. In both cases, however, the ages obtained for the two fractions were discordant.

**6.5.1 Age of Granite Intrusion**

a) Rangitoto Granite:

The results obtained suggest that emplacement of the Rangitoto Granite (U.C. 9473 and U.C. 9458), forming the upper part of the Rangitoto Range, occurred between 295 ± 2 and 314 ± 2 Ma ago. This age agrees well with those obtained elsewhere for 'Tuhuan Granites', on the West Coast, and the Rangitoto Granite is accordingly assigned a
## MINERALISATION, MT RANGITOTO

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<th>Grid ref.</th>
<th>% K</th>
<th>(40^{\text{Ar}}/39^{\text{Ar}}) x 10^3</th>
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### TABLE 6.2: K-Ar dates for the major rock types, and mineralised Mine Creek vein material, Mount Rangitoto.

Notes: 1) TR = total rock, Bi = biotite, Mu = muscovite, P = feldspar

2) Decay constants \(\lambda_p = 4.962 \times 10^{-10} \text{ yr}^{-1}\)

\(\lambda_{40} = 0.581 \times 10^{-10} \text{ yr}^{-1}\)

Isotopic abundance \(40^{\text{K}} = 0.01167 \text{ atomic}\%\)
Carboniferous-Permian age. Emplacement of the Rangitoto Granite is accordingly considered to have occurred during the late stages of the Tuhua Orogeny, during which time granite emplacement was widespread.

From one of the samples (U.C. 9458), a date was obtained for the feldspar fraction as well as the biotite fraction. The age obtained, $147 \pm 1$ Ma, is considerably younger than the biotite age ($295 \pm 2$ Ma) from the same sample. This discordance is interpreted as reflecting a significant argon loss from the feldspar fraction. The high potassium content of the feldspar separate from U.C. 9458 reflects the presence of abundant alkali feldspar, mainly microcline, in the dated material. The ease of argon loss, particularly from slow cooling alkali feldspars, has been noted by numerous workers, and as noted above, seems a likely reason for the discordant age obtained. The date is accordingly considered unreliable, and regarded as insignificant.

b) Kakapotahi Granite:

Biotite from the Kakapotahi Granite (U.C. 9461 and U.C. 9471) yielded significantly younger ages - $144 \pm 1$ and $175 \pm 1$ Ma - than the nearby Rangitoto Granite. These younger ages are interpreted as representing the minimum possible age of emplacement of the Kakapotahi Granite.

Magmatic muscovite from the Kakapotahi Granite (U.C. 9471), was also dated and yielded an age of $213 \pm 4$ Ma. Because of the greater argon retentivity of muscovite relative to biotite, and its lower susceptibility to thermal alteration, the age obtained is considered to reflect the likely age of emplacement of the Kakapotahi Granite.

This upper Triassic age ($213 \pm 4$ Ma) is in marked contrast with the Tuhuan (Carboniferous-Permian) age obtained for the Rangitoto Granite nearby, and implies that two phases of intrusive activity have probably affected the area. A Rb-Sr isochron age would, however, be required to determine with certainty the age of emplacement of the Kakapotahi Granite.
The mineralogical and geochemical differences between the two granites noted in Chapter 5, further suggest that the Rangitoto and Kakapotahi Granites are possibly not genetically related, supporting the K-Ar dates obtained.

6.5.2 Hornfelsing of the Greenland Group

Three samples of Greenland Group hornfelses from Mount Rangitoto were dated. Two of the samples (U.C. 9477 and U.C. 9456) yielded almost identical ages of 275 ± 2 and 277 ± 2 Ma, whereas the third, (U.C. 9469) yielded a considerably older age of 314 ± 2 Ma.

Three samples of Greenland Group metasediments from Mount Greenland, were dated by Adams et al. (1975) and yielded ages of 261 ± 5 Ma, 438 ± 7 Ma and 407 ± 9 Ma (see Table 6.1).

The dates obtained in this study indicate that hornfelsing of the Greenland Group sediments on the southern flanks of Mount Rangitoto occurred as a response to emplacement of the Rangitoto Granite between 295 ± 2 and 314 ± 2 Ma ago. The slightly younger ages obtained from U.C. 9477 and U.C. 9456 (275 ± 2 and 277 ± 2 Ma respectively) suggests that the hornfelses have undergone some argon loss either during, or since, hornfelsing hence yielding younger ages than the Rangitoto Granite. This argon loss may have been a response to re-heating of the hornfelses accompanying subsequent emplacement of the Kakapotahi Granite.

Partial argon loss is adopted by Adams et al. (1975) as an explanation for a young, 261 ± 5 Ma, age obtained from Mount Greenland, close to samples yielding considerably older ages - 407 ± 9 Ma and 438 ± 7 Ma. According to Adams et al. these ages reflect the effects of metamorphism at least 438 Ma ago. If this interpretation is correct, emplacement of the Rangitoto Granite between 295 Ma and 314 Ma ago has not affected the Greenland Group metasediments of Mount Greenland, at the present exposure level, sufficiently to interrupt potassium decay processes.
Accordingly, it seems that the exposed metasediments of Mount Greenland have not been heated above 250 – 300°C, the generally accepted upper temperature limit for argon retention in micas, since the late Ordovician, although the 261 Ma age obtained by Adams et. al., may reflect a Tuhuan thermal overprint related to intrusion of the Rangitoto Granite.

The lack of contact metamorphic textures and minerals (see section 5.1) in the Greenland Group metasediments, further suggests that thermal metamorphism accompanying granite intrusion during the Tuhua Orogeny, did not affect the Greenland Group metasediments of Mount Greenland at the present level of exposure. Contact metamorphism of the metasediments at depth below the Mount Greenland massif, is however, probable.

6.5.3 Age of Mineralisation, Mine Creek

Two samples of mineralised vein material from Mine Creek were analysed in an attempt to determine the age of mineralisation. The ages obtained, 374 ± 3 and 532 ± 4 Ma, showed a huge discrepancy, and as both samples came from closely spaced localities within the same vein, are considered unreliable.

The reasons for the anomalies, and old ages obtained, are unknown. It seems possible, however, that argon may be present in the numerous fluid inclusions within the vein quartz. The presence of argon in the late stage magmatically derived fluids contained within the inclusions, would not be unexpected and probably accounts for the variable and anomalously old ages obtained. It is also possible that the vein system has not remained closed with respect to potassium and argon since its formation, a requirement necessary for the use of the K-Ar method.

Although the age of the mineralisation in Mine Creek could not be determined isotopically, field and laboratory work have enabled some estimates to be made. Sulphur isotope data (section 11.5) indicate that the sulphide sulphur of both the Mine Creek and Cedar Creek veins is magmatic in origin.
The Mine Creek veins are seen to cut aplite dykes, almost certainly formed during emplacement of the Rangitoto Granite. Furthermore, the veins are associated with the Mine Creek Fault, along which significant alteration and sub-economic mineralisation occurs. This alteration is almost certainly due to emplacement of the Kakapotahi Granite, which has an anomalously high silver content and is regarded as the source of silver in the vein deposits (see section 7.4), probably during the Triassic.

The age of the mineralisation in the Cedar Creek veins is unknown however, the magmatic source of the sulphide sulphur present, coupled with the presence of the nearby Rangitoto Granite to the south of the Mikonui River which probably also occurs at depth below Mount Greenland, suggests that the mineralisation in Cedar Creek probably accompanied the later stages of emplacement of the Rangitoto Granite. Accordingly, the veins are regarded as being Late Carboniferous-Early Permian in age, and significantly older than the Late Triassic age inferred for the Mine Creek veins.

6.5.4 Summary and Significance of Results

The dates obtained by Laird et al. (1975) and those obtained as a part of this study, indicate that deposition of the Greenland Group at Mount Greenland occurred prior to 438 Ma (Late Ordovician-Silurian), at which time the sediments were regionally metamorphosed. Emplacement of the Rangitoto Granite, to the south of the Mikonui River between 295 and 314 Ma ago (Carboniferous-Permian), occurred during the Tuhua Orogeny, and was accompanied by hornfelsing of the Greenland Group pendant preserved on the southern flanks of Mount Rangitoto.

To the north of the Mikonui River, in contrast, the metasediments show little evidence of thermal metamorphism at the present exposure level. A 261 Ma age for a slate from Mount Greenland (Adams et. al., 1975), however, probably reflects the thermal effects of granite emplacement during the Tuhua Orogeny. This age, coupled with the sulphur isotope
data obtained for the veins, indicates the probable presence of a Tuhuan aged granite, possibly the Rangitoto Granite, at depth below the Mount Greenland massif. This granite, is considered to be the most likely source of the elements present in the Cedar Creek veins.

Emplacement of the Kakapotahi Granite probably occurred during the Late Triassic c.213 Ma ago, and was accompanied by the formation of the mineralised Mine Creek veins.

The surprisingly young age obtained for the Kakapotahi Granite combined with recent dating studies suggests that Triassic intrusive activity in the Westland foreland may have been more widespread than previously suspected. Aronson (1968) obtained a Triassic (213 Ma) Rb - Sr age from muscovite within the Redjacket Granite (terminology after Nathan, 1975), and more recently Tulloch (1979) has obtained a Triassic (232 - 213 Ma) age for the Tobin Epidiorite. Other Triassic ages from New Zealand include the Mistake Diorite from the Hollyford region 208 - 180 Ma (Williams and Harper, 1978), Bounty Island 188 Ma, Ruapuke Island 217 - 211 Ma, Pahia Point 188 Ma, (Devereux et al. 1968) and Buller River intrusives (Eggers and Adams, 1979).

Many of these dates, obtained from rocks to the east of the Alpine Fault, are from plutonic rocks, remnants of an island arc, within the Brook Street Terrane (Permian-Jurassic), extending eastward from the southeast of the South Island to the Bounty Islands and westwards to the Alpine Fault along the southern margin of the Maitai Terrane.

The presence of Triassic intrusives to the west of the Alpine Fault, suggests that the Brook Street island arc possibly extended west of the present position of the Alpine Fault, cutting across northern Fiordland and into the Westland foreland, which at that time may have been overlain by Carboniferous-Permian and possibly younger sediments, now removed by erosion.

The existence of Triassic plutonic activity in the Westland foreland, probably as a westward extension of the
recognised Brook Street island arc, is further supported by recent pollen dating of the volcaniclastic Topfer Formation, which indicates a Triassic age. Both basaltic and rhyolitic components are present in this formation, further indicating the existence of volcanic activity in the foreland.

No attempt to date the alkaline basic dykes observed in the thesis area was made, due to their high degree of alteration. They are considered, however, to have been emplaced during rifting of the Tasman Sea in the Late Cretaceous.

The identification of probable Triassic granite emplacement in the thesis area, has significant implications in terms of ore genesis. Polyphase intrusive systems, such as seem to exist in the Rangitoto Range, are commonly favourable environments for the formation of large scale disseminated ore deposits e.g. porphyry copper and molybdenum deposits. The recognition of such a system in the Rangitoto Range must, therefore, be regarded as an indicator of a potentially favourable environment for the formation of disseminated ore deposits, particularly porphyry copper-molybdenum type deposits, and disseminated gold-silver deposits.

Disseminated gold-silver mineralisation accompanied by hydrothermal alteration, does occur within the Rangitoto Granite, however, investigations within the thesis area suggest that this mineralisation is local in extent, and sub-economic (see Chapter 8 for description and discussion).
CHAPTER VII

GOLD AND SILVER IN ROCKS

7.1 INTRODUCTION

The problem of determining the origin of the gold and silver contained in many precious metal vein and stockwork deposits, particularly those associated with intrusive activity, has never been satisfactorily resolved. Some workers believe that the intrusive rocks, themselves are the source of the gold and silver, others believe that the intruded sediments in many cases supplied the gold and silver in the deposits.

As the source of these elements is important both in interpreting observed mineralisation, and also in defining likely exploration targets, it was considered worthwhile making a preliminary investigation into the gold and silver content of the major rock types encountered. By determining the local background contents of the major rock types it was hoped to determine the likely source(s) of gold and silver contained in both the vein deposits and the altered Rangitoto Granite (see Chapter 8), and if possible, to determine the migration pattern of gold and silver during the intrusive phase.

The gold and silver content of the altered or mineralised Rangitoto Granite was also investigated, as early reports suggested the likely presence of economic mineralisation in the granite.

7.2 METHOD

Due to the unavailability of both neutron activation and isotope dilution analytical facilities, analysis for gold and silver was undertaken by flameless atomic absorption using a Varian Techtron - AA5 spectrometer. The method used was based on those outlined by Tindall (1965) and Hildon and Sully (1971), with minor modifications made to the
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<td></td>
<td></td>
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<tr>
<td>10191</td>
<td>&lt;.002</td>
<td>.0133</td>
<td></td>
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<tr>
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<td>.0048±.0014</td>
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<td>.0042±.0005</td>
<td>&lt;.002</td>
<td></td>
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<td></td>
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<tr>
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<td>.0069</td>
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<td>.0438</td>
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<tr>
<td>10214</td>
<td>.0301±.0029</td>
<td>.7313±.0103</td>
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<td></td>
<td>24.29</td>
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</table>

**TABLE 7.1**: Gold and Silver contents of the major rock types of Mt. Greenland and Mt. Rangitoto.

**Note**: 1. Errors quoted are analytical errors based on replicate analyses.
2. Sample numbers are University of Canterbury XRF No's except for U.C.9484, which is a University of Canterbury sample number.
digestion process (after Newman, 1979) as outlined in Appendix III.

Sample preparation consisted of fine grinding of the samples, followed by digestion in HCl and aqua regia, addition of HBr to stabilize the gold and silver in solution, and finally concentration of the gold and silver into MIBK.

Because of the low levels of gold and silver being analysed, large samples (~20g) were used, with concentration into a final volume of 3.70ml of MIBK. Spiked duplicates were prepared for each sample, in the case of gold, providing an internal calibration standard for each sample. For silver, several standards were prepared for calibration purposes, and the concentration in the unknowns was calculated using the derived calibration.

Analysis of 5μl samples of MIBK was undertaken using a carbon cup atomizer. Replicate analyses of gold and silver suggest a precision of ± .0002ppm at concentrations of .003ppm, with a lower detection limit for both gold and silver of .002ppm.

7.3 RESULTS

Twenty-six rock samples were analysed by the above method. The results obtained are summarised in Table 7.1. Average background values for similar rock types have been calculated by Boyle (1979) and are given in Table 7.2 below:

<table>
<thead>
<tr>
<th>Rock Type</th>
<th>Gold</th>
<th>Silver</th>
<th>Ag</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>X</td>
<td>σ</td>
<td>X</td>
</tr>
<tr>
<td>Acid Intrusives</td>
<td>.0114</td>
<td>.0796</td>
<td>.05</td>
</tr>
<tr>
<td>Greywacke</td>
<td>.0132</td>
<td>.0192</td>
<td>0.2</td>
</tr>
<tr>
<td>Shale, Mudstone, siltstone, argillite</td>
<td>.0080</td>
<td>.0266</td>
<td>0.2</td>
</tr>
<tr>
<td>Quartzites and indurated sandstones</td>
<td>.0317</td>
<td>.0836</td>
<td>-</td>
</tr>
<tr>
<td>Slates, phyllites, meta-argillites</td>
<td>.0022</td>
<td>.0022</td>
<td>0.2</td>
</tr>
<tr>
<td>Hornfels, meta-pelites</td>
<td>.0064</td>
<td>.0262</td>
<td>-</td>
</tr>
</tbody>
</table>

Table 7.2 : Average gold and silver contents of selected rock types (from Boyle, 1979).
The results obtained in this study show the following:

1. The Kakapotahi Granite is significantly enriched in silver (\( \overline{Ag} = 0.400 \text{ppm} \)) compared with both the world average for acid intrusives (\( \overline{Ag} = 0.05 \text{ppm} \)), and the Rangitoto Granite (\( \overline{Ag} = 0.055 \text{ppm} \)).

2. The Greenland Group metasediments of Mount Greenland contain silver in concentrations significantly lower (\( \overline{Ag} < 0.008 \text{ppm} \)) than the world average for similar rocks (\( \overline{Ag} \approx 0.2 \text{ppm} \)). The Greenland Group hornfelses of Mount Rangitoto have a similar silver content (\( \overline{Ag} < 0.005 \text{ppm} \)) to the less metamorphosed metasediments.

3. The gold background for the four major rock types is approximately 0.003ppm (neglecting the high values obtained for samples 10183 and 10210). For the metasediments and hornfelses this is similar to Boyle's average (\( \overline{Au} \approx 0.004 \text{ppm} \)) but the gold content of the Rangitoto and Kakapotahi granites is significantly lower than Boyle's average for acid intrusives (\( \overline{Au} \approx 0.0114 \text{ppm} \)).

4. The Kakapotahi Granite gives Ag : Au ratios greater than 90, compared with less than 21 for the Rangitoto and altered Rangitoto granites. The Greenland Group metasediments and hornfelses have ratios of less than 7.

7.4 INTERPRETATION

The results obtained provide some insight both into the behaviour of gold and silver during intrusion of the granite bodies, and also into the source of these elements in the vein deposits.

7.4.1 Distribution of Gold and Silver in the Major Rock Types

a) Source of Silver in the Kakapotahi Granite:

The anomalously high silver content of the Kakapotahi Granite almost certainly reflects the existence of higher than normal concentrations of the element in the parent magma. Although slight sericitic and propylitic alteration of the granite has occurred (see section 5.4), this alteration is not intense and is certainly no more pervasive than in the Rangitoto Granite, which has a silver content some 6 - 8 times lower than the Kakapotahi Granite.
Accordingly, it is considered unlikely that secondary alteration processes have significantly altered the silver content of the Kakapotahi Granite. Furthermore, as mineralisation processes do not appear to have affected the granite either—no silver bearing ore minerals have been detected—it seems improbable that silver has been added to the granite subsequent to crystallisation, and a primary magmatic origin is indicated.

No attempt has been made to investigate the silver content of the silicates comprising the granite, although work by Boyle (1968) suggests that garnet, muscovite, biotite, chlorite and albite may all contain significant quantities of silver. The similarity of the ionic radii, and other chemical properties of silver, to potassium and sodium, suggests that silver may replace these elements in the silicate lattices. It is perhaps worth noting that the silver enriched Kakapotahi Granite is slightly richer in both sodium and potassium than the Rangitoto Granite which also has a lower silver content.

b) Silver Content of the Other Major Rock Types:

i) Rangitoto Granite: The Rangitoto Granite has a silver content typical of acid intrusives (Ag = 0.055ppm). The values obtained are somewhat more erratic than for the other rock types, ranging from 0.098ppm to 0.029ppm, probably reflecting the effects of local alteration and mineralisation of the granite (see section 8.2).

The Rangitoto Granite has undergone a similar degree of propylitic alteration to the Kakapotahi Granite, evident by the alteration of plagioclase feldspars. Superimposed on this alteration, particularly near the Mine Creek Fault, is a somewhat irregular and more intense propylitic alteration (see section 8.2), in places accompanied by higher than normal silver and gold values (e.g. samples 10183 and 10210).

Samples 10213 and U.C. 9484 are more highly altered being cut by a stockwork of quartz-calcite-K-feldspar veinlets (see section 8.2). Both samples show lower than average silver contents (0.013 and 0.028ppm respectively), indicating a possible loss of silver from the granite during
this more intense alteration. Gold, in contrast to silver seems to be little affected by the degree of alteration.

A fuller description of the alteration and gold-silver mineralisation of the Rangitoto Granite is given in the following Chapter.

ii) Greenland Group Metasediments and Hornfelses:

Both the Greenland Group metasediments and hornfelses contain unusually low concentrations of silver (<0.008ppm). This is probably best explained in terms of the evolution of the Greenland Group sediments. Nathan (1976) concludes that the Greenland Group was derived by polycyclic erosion and deposition of an acid plutonic source, and that the source rock for the final cycle was a quartzose metasedimentary unit. Quartz from plutonic rocks is almost always poor in silver (Boyle, 1968, p.47), most of the silver present in such rocks being contained in albite or the micas. These minerals have been almost entirely destroyed during the erosive cycle(s), resulting in a quartz-rich, silver-poor sediment. Subsequent regional metamorphism and recrystallisation of the clay matrix may have resulted in further depletion of silver from the metasediments.

Alternatively, and perhaps more unlikely, the low silver content of the metasediments and hornfelses may reflect derivation of the sediments from silver-poor source rocks.

The almost identical silver content of the metasediments and the hornfelses (see Table 7.1) suggests that no significant enrichment or depletion of silver has occurred in the sediments during granite emplacement and the associated contact metamorphism.

c) Gold Content of the Major Rock Types:

The average gold content of the major rock types ranges from 0.0026ppm in the Greenland Group metasediments to 0.0039ppm (excluding the obviously high values in samples 10183 and 10210) in the Rangitoto Granite.
Gold, unlike silver, shows no enrichment in either the Kakapotahi or Rangitoto granites, being present in both granites in concentrations below the average for acid intrusives (see Table 7.2). The reason for the low gold contents (0.0030 and 0.0039 ppm respectively) of the granites is unknown, however, it is possible that both granites may be a part of a regional suite of low gold-bearing granites, such as have been noted by Boyle (1979, p. 41).

A general positive correlation between gold and silver in igneous rocks has been noted by numerous workers (see for example Boyle 1979, p. 42). The granites in the thesis area show no such correlation, the Kakapotahi Granite having an Ag : Au ratio greater than 90, compared with an Ag : Au ratio of less than 21 for the Rangitoto Granite. The reason(s) for the lack of correlation of gold and silver between the two granites are unclear, but may be related to variations in the Ag : Au ratios of the source rocks prior to melting, or alternatively, reflect variation in the mobility of gold, relative to silver, in the intrusive systems.

The Greenland Group harnfelses and metasediments have almost identical gold contents (0.0026 and 0.0028 ppmrespectively), and have Ag : Au ratios in the range 0.5 to 5.7. These ratios are close to the average for harnfelses and metasediments (see Table 7.2), although as noted previously silver concentrations in these rocks are considerably below the average.

The data available suggest that no significant enrichment or depletion of gold has occurred in the rocks of the harnfelsed pendant during contact metamorphism, in spite of the fact that a redistribution of the gold present, associated with the growth of metamorphic micas, almost certainly accompanied the intrusive activity. The apparently inert behaviour of gold and silver during the thermal metamorphism is a little surprising in view of the chalco-phile tendencies of the elements.
7.4.2 Sources of Gold and Silver in the Mineralised Rocks and Veins:

a) Mine Creek Vein Deposits: Sulphur isotope data obtained as a part of this study (see section 11.5.1) clearly indicate a magmatic origin for the sulphur contained in the fluids that formed the vein deposits. Several lines of evidence indicate that these fluids were derived from the silver-rich Kakapotahi Granite.

The high silver content of the Kakapotahi Granite, coupled with the chalcophile tendencies of silver, suggests that magmatic fluids derived from this granite would probably be silver rich also. The silver rich nature of the ore fluids is reflected in the argentiferous nature of the galena present, and also by the presence of electrum in the vein assemblage. Furthermore, the Ag : Au ratios of the mineralised vein samples analysed, are always greater than 1 (see Table 9.2), again reflecting the silver rich nature of the ore-forming fluids. Silver rich fluids such as apparently formed the vein deposits are, unlikely to have been derived from the Rangitoto Granite, which has a considerably lower silver content than the Kakapotahi Granite.

The association of both the vein mineralisation and the disseminated mineralisation in the Rangitoto Granite with the Mine Creek fault, indicates that this fault has probably acted as a major feeder of magmatic fluids to the vein system. The fault post-dates the intrusion of the Rangitoto Granite, and therefore the magmatic fluids responsible for the vein mineralisation must also post-date the intrusion of the Rangitoto Granite. The most obvious source of these fluids is the younger (see section 6.5.1), silver-rich, Kakapotahi Granite.

The Greenland Group hornfelses of the roof pendant cannot be regarded as a significant source of either gold or silver in the Mine Creek vein deposits. Analysis of a single wall-rock sample (10214), collected 0.3m from the footwall side of a mineralised vein, shows that both gold and silver have been added to the wall rock, rather than depleted from it, suggesting that the wall-rocks are probably not the source of the silver and gold in the veins.
Furthermore, the low silver content of both the metasediments and the hornfelsed pendant, and the apparently inert behaviour of both gold and silver during hornfelsing of the sediments, indicates that the metasediments and hornfelses were not a significant source of either the silver or the gold in the vein deposits.

b) Mineralised Rangitoto Granite: Local enrichment of gold and silver associated with hydrothermal alteration occurs in the Rangitoto Granite. The source of the gold and silver present is uncertain, however, the association of the alteration with the Mine Creek Fault suggests that, like the vein deposits, both the silver and gold present were probably derived from the Kakapotahi Granite. It is possible, however, that the Rangitoto Granite itself is the source of the gold and silver in the mineralised portion, being derived from the bulk of the granite, as a result of late stage auto-alteration by hydrothermal fluids.

c) Cedar Creek Reefs: Sulphur isotope evidence (see section 11.5.1) indicates that the sulphur in the sulphides of the Cedar Creek reefs is of dominantly magmatic origin. Very little direct evidence exists to indicate the source of the gold and silver present in veins or reefs, however, several factors apart from the sulphur isotope data combine to suggest a magmatic origin for both the gold and silver.

The Greenland Group cannot be regarded as a likely source of either gold or silver. The wall rocks adjacent to the veins, where exposed, show little or no sign of alteration. If the 3,933oz (122,316g) of gold won from the William Tell levels were derived from the Greenland Group, with an average gold content of 0.0026ppm (see Table 7.1), a minimum volume of $46.88 \times 10^6$ tonnes ($\approx 1.77km^3$) of rock would be required to provide the gold won from the veins (assuming complete removal of the gold from the meta-sediments). This is equivalent to a volume of wall rock 1.77m wide x 5km long x 2km deep. No evidence for such extensive wall rock alteration exists in the field, and accordingly, the wall rocks adjacent to the veins are not considered to be the source of the gold present in the veins.
Regional metamorphism of the metasediments during the Ordovician may have liberated significant quantities of gold from the metasediments. Formation of the vein deposits, however, post-dates this metamorphism, and it is therefore considered that the metasediments are almost certainly not the source of the gold and silver present in the Cedar Creek reefs.

The most likely source of both the gold and silver in the Cedar Creek reefs, is considered to be a granite body, probably of Tuhuan age and possibly the Rangitoto Granite, underlying the Mount Greenland massif at depth. This is indicated both by the sulphur isotope data (section 11.5) and the structural relationships observed in the thesis area.
8.1 INTRODUCTION

Between 1889 and 1893 the granite surrounding the Mine Creek workings was prospected by a syndicate known as the Mount Rangitoto Freehold, and according to records of the times (see section 2.1) an area of mineralised granite was located. Trials of the stone are reported to have yielded up to 6dwt 14gr (10.1g) of gold and 10dwt 14gr (16.5g) of silver per ton. No mining of the granite was, however, undertaken.

A brief search for the mineralised granite seemed appropriate in view of the current interest in low grade large tonnage precious metal deposits, particularly gold, and was undertaken as part of this study. A small outcrop of sulphide (predominantly pyrrhotite and marcasite) bearing hornfels was located near the old mine workings during the search.

8.2 'MINERALISED' OR 'ALTERED' RANGITOTO GRANITE

Traces of disseminated mineralisation were located in the Rangitoto Granite in Mine Creek and its tributaries, particularly in the vicinity of the old mine workings (see Geological Map). The mineralisation, which is associated with propylitically altered Rangitoto Granite, was only observed near the Mine Creek Fault (see Fig 8.1) to which it is almost certainly genetically related.

Five samples of granite, altered to varying degrees, were analysed for both gold and silver by atomic absorption spectroscopy, and although all the values obtained were economically uninteresting, maximum concentrations of gold
Fig 8.1 Highly fractured and propylitically altered Rangitoto Granite adjacent to the Mine Creek Fault (dashed). Darker granite to the right of the fault is less fractured and altered. Note the partially blocked entrance to the prospecting drive in the 'altered' granite (centre left).

Mine Creek, west branch (G.R. S64/365197)
and silver approximately 11x (0.041ppm) and 5x (0.187ppm) the respective background values (see Table 7.1) were detected.

8.2.1 Mineralogy

The mineralogy of the Rangitoto Granite has been described previously (see section 5.3). Thin section examination of the 'altered or mineralised' portions of the Rangitoto Granite reveals a similar pre-alteration mineralogy. Plagioclase feldspar subhedral-anhedral up to 5mm across make up about 40% of the rock. Many of the crystals are normally zoned, ranging in composition from andesine - albite (An$_{40}$ - An$_{10}$). Alteration of the plagioclase is ubiquitous and well developed, the dominant alteration products being zoisite, with lesser amounts of sericite, epidote and calcite. Alteration of many crystals is pervasive and has resulted in the complete destruction of the former feldspar, including the more sodic rims (see Fig 8.2).

Quartz (0.01 - 4mm) occurs as anhedral crystals intergrown with the feldspars and biotite. Apart from being extensively fractured where associated with calcite stockworks, no other alteration is evident.

Quartz of secondary origin formed by the breakdown of plagioclase is also present in the altered Rangitoto Granite, forming small anhedral crystals adjacent to calcite veinlets.

Microcline (0.01 - 8mm) also occurs unaltered as intergrowths with quartz and plagioclase, as previously noted for the Rangitoto Granite.

Biotite flakes, up to 4mm across make up 10 - 15% of the rock and are commonly partly replaced by Fe - Mg chlorite, with strongly anomalous purple interference colours.

Calcite is ubiquitous in 'altered' samples of Rangitoto Granite collected from near the Mine Creek Fault, forming veinlets and grain boundary infillings between the feldspars and quartz. The abundance of calcite varies from 1% in the least altered samples to 15 - 20% of highly altered hand specimen sized samples. Where present in greater abundance,
Fig 8.2 Plagioclase poikilitically included in microcline, pervasively altered to zoisite and epidote. Sodic rims of the plagioclase are less altered than the more calcic cores, and the host microcline is unaltered. Thin section, cross polarised light. U.C.9484.

Near Mine Creek Fault. (.G.R. S64/395197)
the calcite forms pervasive stockworks in the granite, which has undergone fracturing (see Fig 8.3) associated with movement on the Mine Creek Fault. The stockwork veinlets commonly contain angular brecciated fragments of quartz and rare K-feldspar. Plagioclase is absent in the calcite of the veinlets, reflecting its high degree of instability in a CO₂-rich hydrothermal environment.

The high degree of alteration of plagioclase to zoisite and epidote followed by the replacement of these minerals, particularly in the stockworks, has liberated calcium which has combined with CO₂ to form the secondary calcite present.

Pyrite is the only sulphide mineral observed in the altered Rangitoto Granite occurring as small (< 0.01mm) anhedral to euhedral, generally along grain boundaries, where it may be associated with calcite. The pyrite which is absent in the unaltered granite occurs only in trace amounts and there is no obvious increase in pyrite abundance with increasing alteration and stockwork development. A small nodule of pyrite 2 - 3cm long was located in a small shear adjacent to the Mine Creek Fault, however, such nodules are absent in the body of the rock.

8.2.2 Geochemistry

Three samples of 'altered' Rangitoto Granite were collected and analysed, as well as one sample containing abundant calcite stockworks. The results of the analyses are given in Table 8.1. Samples 10210, 10211 and 10212 were collected at distances of 5m, 10m and 15m, respectively, from the Mine Creek Fault, and sample 10213 which contains abundant calcite stockworks, was collected 11m from the fault. All of the samples come from within the fracture zone associated with the Mine Creek Fault.

The results obtained for samples 10210 - 10212, when compared with the average composition of the unaltered Rangitoto Granite (see Table 8.1) indicate slight gains in the transition metals, Fe, Mg and Ti due to hydrothermal alteration, attributed to the development of secondary chlorite both by reaction with Al₂O₃ and SiO₂ released by
Fig 8.3(a) Calcite veinlets infilling fractures in brecciated quartz. 'Altered' Rangitoto Granite. Thin section, cross polarised light. U.C.9481. (G.R. S64/365197)

Fig 8.3(b) Calcite stockwork infilling fractures in and partially replacing plagioclase. 'Altered' Rangitoto Granite. Thin section, cross polarised light U.C. 9481. (G.R. S64/365197)
TABLE 8.1: Geochemical analyses of altered Rangitoto Granite.

Au and Ag by AA spectroscopy, remaining trace elements and major element oxides by XRF.
Samples 10210 - 10213 from 5m, 10, 15m and 11m from Mine Ck. Fault respectively.
See text for discussion.

* Probably too low due to analytical error.

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<tr>
<th>SAMPLE NO:</th>
<th>10210</th>
<th>10211</th>
<th>10212</th>
<th>10213</th>
<th>10183</th>
<th>Average Unaltered Granite</th>
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<tr>
<td>SiO₂</td>
<td>67.4*</td>
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<td>TOTAL</td>
<td>95.65</td>
<td>98.90</td>
<td>98.98</td>
<td>98.93</td>
<td>99.57</td>
<td>99.60</td>
</tr>
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</table>

|  |  |  |  |  |  |  |  |
| Ba | 265 | 301 | 265 | 286 | 367 | 307 |
| Nb | 10 | 11 | 10 | 11 | 11 | 12 |
| Rb | 204 | 187 | 186 | 168 | 198 | 204 |
| Sr | 128 | 140 | 155 | 133 | 172 | 163 |
| Y | 32 | 30 | 29 | 35 | 28 | 29 |
| Zr | 105 | 105 | 107 | 100 | 104 | 105 |
| S | 997 | 22 | <20 | 50 | <20 | <20 |
| Cu | 73 | 17 | 13 | 17 | 21 | 15 |
| Pb | 131 | 67 | 57 | 39 | 60 | 55 |
| Zn | 70 | 51 | 50 | 50 | 52 | 49 |
| Sn | 128 | 70 | 69 | 35 | 63 | 65 |
| Sb | 49 | 5 | 3 | <2 | <2 | <2-10 |
| As | <5 | <5 | <5 | <5 | <5 | <5 |
| Ag | .1873 | .0642 | .0449 | .0275 | .0983 | .0546 |
| Au | .0411 | .0055 | .0024 | .0037 | .0198 | .0071 |
the breakdown of plagioclase, and as a replacement of biotite. Alteration of plagioclase has resulted in slight losses of Ca and Na. $\text{Al}_2\text{O}_3$ shows no change in concentration due to alteration apparently being recombined with no net gains or losses. $\text{SiO}_2$ shows a slight loss due to alteration, probably a result of the replacement of biotite by chlorite, and the replacement of plagioclase by zoisite and epidote, all reactions which result in a nett $\text{SiO}_2$ loss.

Propylitisation of the Rangitoto Granite adjacent to the Mine Creek Fault is not intense, and accordingly changes in the major element geochemistry are only slight.

Of the minor elements Ba, Rb and Sr are all lost during hydrothermal alteration, due to the breakdown of the feldspars, particularly plagioclase. Y, Nb and Zr show no changes in concentration, reflecting their general stability in the hydrothermal environment.

The economically interesting elements viz. Cu, Pb, Zn, Sn, Sb, As, S, Au and Ag show no significant enrichment in samples 10211 and 10212, generally lying close to, or just above the average concentrations detected in the unaltered Rangitoto Granite (see Table 5.4). Sample 10210, in contrast shows significant enrichment in S, Cu, Pb, Zn, Sn, Sb, Ag and Au. The high S content (997ppm) is probably due to increased concentrations of the alteration product pyrite in the sample. No discrete Cu, Pb, Zn, Sn or Sb bearing minerals have been observed in this sample, these elements presumably occurring as lattice constituents in the alteration assemblage, particularly pyrite.

Enrichment of Au and Ag in sample 10210 is also attributed to the increased abundance of pyrite, which may contain up to c.200ppm Au and Ag (Boyle, 1979), substituting for Fe in the pyrite lattice. None of the silicates of the alteration assemblage are known to contain significant quantities of either gold or silver.

Sample 10183, although apparently unaltered, also contains high Au and Ag values. Why this should be so is unclear, however, the high values may be related in some unknown way to the higher than normal CaO content (2.44%) of the sample.
Sample 10213 contains abundant calcite, both in the form of pervasive stringers forming a stockwork, and as a replacement of plagioclase. In thin section the veinlets of the stockwork are seen to contain abundant angular quartz and feldspar fragments (see Fig 8.3), probably due to fracturing contemporaneous with fault movement and hydrothermal alteration accompanying emplacement of the Kakapotahi Granite.

Alteration in this sample is accompanied by losses in Si, Na, K, and Al, the latter due to the development of secondary calcite at the expense of plagioclase. Accompanying the losses is a big gain in Ca, and lesser gains in Ti, Fe and Mg.

Trace elements show similar trends to those noted for samples 10210 - 10212 (inclusive), although Ba, Rb and Sr as would be expected show less depletion than in the calcite poor samples. Of the economically interesting elements only S and As show enrichment. Cu, Zn, Au and Ag show no significant changes in concentration, and Pb and Sn show nett losses.

S and As being amongst the most mobile of elements in the hydrothermal environment, are commonly concentrated, with respect to the less mobile metals Cu, Zn, Pb, Sn, Au and Ag, in late stage hydrothermal fluids. The increased concentrations of S and As and lack of metal enrichment associated with the calcite stockworks in the altered granite, may reflect a late stage alteration by metal-poor, S- and As-rich fluids.

Although it is tempting to correlate the degree of alteration and mineralisation with distance from the Mine Creek Fault, insufficient data exists to do this with any confidence. The results in Table 8.1, however, suggests the possibility of both increasing alteration and mineralisation with proximity to the fault. If this is indeed the case, the geochemistry of samples 10210, 10211 and 10212 collected 5m, 10m and 15m respectively from the Mine Creek Fault, suggests that S, Cu, Pb, Zn, Sn, Sb, Au and Ag all increase in concentration as the fault is approached.
K, Ca and Na apparently decrease as the fault is approached, in contrast to Ti and Fe which also show an apparently increasing trend.

Although the trends noted could reasonably be expected on geological grounds, they may in fact be more apparent than real, being based on only three samples. Analysis of considerably more samples would be required to establish the nature of the relationship between alteration/mineralisation and distance from the Mine Creek Fault with any certainty.

8.3 SULPHIDE-BEARING HORNFELS

A narrow band of sulphur-metasomatised hornfels was located near Manton's Workings in Mine Creek (G.R.S64/367197), exposed in a recent slip. The altered zone, up to 1.5m thick, is parallel or sub-parallel to bedding, but can only be traced a short distance before being cut by a small fault in one direction, and disappearing under vegetation in the other. Accordingly, the extent of the alteration present and its relationship to other rock types is only poorly understood.

8.3.1 Mineralogy

The mineralogy of the sulphide bearing hornfels reflects the effects of thermal metamorphism, related to intrusion of the Rangitoto Granite, overprinted by sulphur metasomatism. Muscovite (c.40%), quartz (c.30%) and biotite (c.25%) with lesser chlorite (up to 5%) are the dominant silicate minerals present in the rock, with pyrrhotite, which may be altered to marcasite, accounting for up to 5% of the rock.

Muscovite (0.05 - 1mm) occurs intergrown with biotite laths, which may be partly replaced by muscovite, and also with groundmass quartz. Relict spots up to 7 - 8mm across are common, and consist of fine grained muscovite (or sericite), quartz and chlorite. The presence of chlorite
in these spots, which have probably formed by the break-
down of andalusite and possibly cordierite porphyroblasts,
is attributed to the effects of metasomatism, chlorite
being absent in such spots elsewhere in the pendant.

Quartz and biotite make up the groundmass of the rock,
 occurring as equigranular polygonal intergrowths similar
to those noted in the unaltered hornfelses elsewhere in
the pendant.

Pyrrhotite shows a variety of occurrences in the
 metasomatised hornfels, occurring most commonly as
irregular interstitial infillings, between quartz grains
and micas, up to 1mm across (see Fig 8.4). Less commonly
irregular or lensoidal aggregates of grains up to 3 - 4mm
long (see Fig 8.5) are present, and are commonly elongate
parallel to the foliation of the hornfels, defined by the
preferred biotite orientation. Thin films of pyrrhotite
are also present in the hornfels, being developed along
quartz-biotite contacts, presumably due to the reaction
of iron, released by the breakdown of biotite, with sulphur
in the metasomatic fluids.

In the groundmass of the rock pyrrhotite usually occurs
in association with chlorite in muscovite rich areas. This
association is presumably due to the almost complete
destruction of biotite to form muscovite, minor chlorite and
iron, which has reacted with sulphur in the metasomatic
fluids, forming pyrrhotite. Elsewhere, pyrrhotite occurs
as the incipient breakdown of biotite, occurring in decussate
clots within the hornfels. Veinlets of pyrrhotite, generally
less than 1mm across are also sparingly present within the
rock.

Inversion of pyrrhotite to marcasite, although not
extensive, occurs sparingly throughout the rock, the
marcasite occurring as anhedral crystals up to 0.06mm
across, and as fine grained aggregates replacing pyrrhotite.
Twinning is common, particularly in the larger crystals.

The metastability of marcasite, even at low temperatures,
suggests that inversion of pyrrhotite to form marcasite
almost certainly accompanied intrusion of the Kakapotahi
Fig 8.4 Pyrrhotite (black) occurring as an interstitial infilling to quartz (white) in S-metasomatised Greenland Group hornfels. Thin section, plane polarised light. U.C.9469. (G.R. S64/367196)

Fig 8.5 Irregular aggregates of pyrrhotite (po) elongate parallel to cleavage, developed in S-metasomatised Greenland Group hornfels. Polished section, plane polarised light, oil immersion. U.C.9469. (G.R. S64/367196)
Granite, or alternatively occurred subsequent to intrusion of this granite. Marcasite, if present prior to this intrusive stage, would almost certainly have become unstable under the influence of heat, inverting to another of the iron sulphide species, either pyrrhotite or pyrite depending on the activity of sulphur (As).

Rare small pyrite euhedra are present in the hornfels, particularly in the biotite-poor areas of the rock. In these areas the As may have been sufficiently high to combine with the low levels of iron present, thereby forming pyrite rather than pyrrhotite. Alternatively, the pyrite may be due to the inversion of marcasite, as a response to heating associated with emplacement of the Kakapotahi Granite.

8.3.2 Geochemistry

One sample of sulphur-metasomatised hornfels was analysed (sample 10193, Table 5.1), along with a slightly sulphur-metasomatised sample (sample 10192, Table 5.1). The analyses suggest that sulphur metasomatism is characterized by significant Ca, Pb, As and Cu addition, as well as significant S addition. As and Cu undoubtedly occur in the pyrrhotite and other sulphides formed by sulphur metasomatism, and accompanying inversion to marcasite. No obvious host minerals for either Ca or Pb have been detected in the hornfels. Presumably these elements are hosted in the micas or silicates of the metasomatised hornfels.

Losses in Fe due to metasomatism are apparent, and due to the breakdown of biotite and subsequent loss of Fe, some of which has recombined with S to form pyrrhotite. Accompanying this loss in Fe are slight apparent losses in the other transition metals Ti, Mg, Mn and Zn, and a loss in Na. The loss in Na is attributed to the destruction of sodic feldspar in the hornfels as a consequence of sulphur metasomatism, whereas the transition metals are presumably lost as a result of biotite breakdown.

No changes in either Ag or Au concentrations due to sulphur metasomatism have been observed.
CONCLUSIONS

8.4 Alteration and Mineralisation of the Rangitoto Granite

Propylitic alteration of the Rangitoto Granite although ubiquitous, is locally intensified adjacent to the Mine Creek Fault, and is apparently due to hydrothermal alteration of the fault zone accompanying emplacement of the Kakapotahi Granite, during the Triassic.

The alteration is characterised by the replacement of plagioclase by epidote, zoisite, calcite and minor sericite, and occasionally by slight sericitisation of alkali feldspars, mainly microcline. Biotite is replaced by Fe - Mg chlorite (Albee, 1962) which is also present in small amounts in the groundmass of the granite. Locally, pervasive calcite stockworks are developed in the granite.

Geochemically these mineralogical changes are accompanied by slight losses in Si, Na, K, and Al, due to the breakdown of feldspars, and slight gains in the transition elements Fe, Mg and Ti, presumably due to the development of secondary chlorite. Ca shows higher losses, than the transition metals reflecting the higher degree of alteration of plagioclase.

Of the trace elements, losses in Ba, Rb and Sr accompany feldspar destruction. Only in sample 10213, the most highly altered and mineralised sample collected, do the economically interesting elements show any changes in concentration. In this sample S, Cu, Pb, Zn, Sn, Sb, Ag and Au all show significant enrichment and, with the exception of Sn, accompany the development of minor amounts (<1%) of pyrite. Sn, derived from the hydrothermal fluids is presumably hosted in biotite, and possibly other phyllosilicate minerals.

A possible trend of increasing alteration and mineralisation approaching the Mine Creek Fault is indicated in this study, but further systematic sample analysis is required to either prove or disprove the existence of this trend.
No gold or silver enrichment accompanying the development of calcite stockworks has been detected, the high values obtained apparently being related to minor pyritisation accompanying the more widespread propylitic alteration of the granite.

Gold and silver concentrations, reported to be present in the Rangitoto Granite during early prospecting operations have not been detected in this study, the highest gold and silver concentrations obtained being 0.187ppm (2.9gr/tonne) silver, and 0.04ppm (0.62gr/tonne) gold. These concentrations although sub-economic are, however, significantly above the background values in the Rangitoto Granite (0.007ppm gold and 0.055ppm silver).

8.4.2 Sulphide-Bearing Hornfels

Sulphur metasomatism of the hornfelsed pendant accompanied intrusion of the Rangitoto Granite and is characterised by the formation of fine grained pyrrhotite (<5%) in the hornfelses. The extent of sulphur metasomatism located is, however, very small being confined to the vicinity of Manton's workings.

Hornfelsing and alkali metasomatism of the pendant (described in section 5.2) has been overprinted by sulphur metasomatism resulting in the replacement of biotite by muscovite and minor chlorite, and the liberation of Fe, which has reacted with S in the metasomatic fluids to form pyrrhotite. Accompanying the formation of pyrrhotite is an addition of Ca, Pb, As and Cu to the hornfels, As and Cu almost certainly occurring in the pyrrhotite formed. The host minerals for Ca and Pb are unknown. Losses in Fe, and slight losses in the other transition metals Ti, Mg, Mn and Zn are attributed to nett losses accompanying biotite destruction. Na is also lost during sulphur metasomatism.

No significant enrichment or depletion of either gold or silver accompanies the sulphur metasomatism.
Subsequent inversion of pyrhotite to marcasite is attributed to thermal effects related to emplacement of the younger Kakapotahi Granite.

The sulphur metasomatism observed is regarded as being a high temperature (c.550°C or more) effect, occurring during the waning stages of emplacement of the Rangitoto Granite. The metasomatic fluids apparently had only moderate to low sulphur activities ($A_s^2$) resulting in the formation of pyrrhotite rather than the more sulphur-rich pyrite.

This sulphur-metasomatism is economically uninteresting, no significant metal values (including gold and silver) being detected in the metasomatised hornfels.
CHAPTER IX

VEIN MINERALISATION: MOUNT RANGITOTO

9.1 EMPLACEMENT

9.1.1 Relationship to Structural Features

The gold- and silver-bearing base metal veins exposed in Mine Creek are located near the crest of the Mine Creek Anticline, a symmetrical, horizontal or gently plunging upright fold with an interlimb angle of about 60°. The axial plane strikes NW - SE (c.300°), and is cut off to the west by the Mine Creek Fault, which marks the western contact of the Greenland Group metasediments with the Rangitoto Granite (see Geological Map). The Mine Creek Fault and associated splays, are normal in character, and post date intrusion of the Rangitoto Granite.

The mineralised veins strike to the northeast (c.055°), and dip between 10° and 30° to the northwest, cutting both bedding and the axial plane of the Mine Creek Anticline at a high angle.

The veins exposed in Manton's workings in Mine Creek, are located in shear zones, with crushed Greenland Group sediments 20 - 30cm thick usually present on the footwall of the veins, and less commonly on the hanging wall also. Unfortunately, the sense and amount of movement on the mineralised shears cannot be determined.

The vein system in the old workings, is cross-cut by a small northwest striking, vertically dipping fault, with an offset of c. 1 metre. The mineralised veins in turn cross-cut earlier formed aplite dykes indicating that the vein system:

a) post-dates dyke emplacement, thought to be associated with intrusion of the Rangitoto Granite;

b) has been disrupted by post-formational faulting and formation of crush zones.
The lack of significant offset of the aplite dykes cut by the mineralised veins shows that although post formational faulting has disrupted the veins, no significant shearing along the veins has occurred since the emplacement of the aplite dykes.

9.1.2 Vein Morphology

The morphological characteristics of mineralised veins may be used to determine which of the processes - replacement or fissure infilling - was most important during formation of the veins (see, for example, Park and MacDiarmid, 1975). Unfortunately, many of the criteria listed by Park and MacDiarmid cannot be readily applied to the Mine Creek veins, due to the development of crush zones along the vein walls.

The best evidence regarding the mode of formation, of the mineralised veins is provided by the petrography of the veins, and the nature of the included wall rock fragments.

Lenticular fragments of wall rock, from less than 1 cm up to 15 cm in length, are common in the veins and are usually aligned parallel to the vein walls. The outline of the included fragments only rarely matches that of the wall rock enclosing the veins, indicating relocation of many fragments due to shearing accompanying vein formation processes.

The included wall rock fragments are commonly fractured, the fractures being healed by quartz veinlets which sometimes have matching walls.

In thin section the wall rock fragments, typically consist of fine grained quartz and sericite, and show sharp contacts with vein quartz. Occasionally, however, altered wall rock inclusions are observed, apparently undergoing replacement by the vein minerals. Similarly, around the margins of such inclusions, fine grained sericite, relict from the wall rock inclusions, is sometimes present in the vein quartz.

The above features indicate that the mineralised veins formed dominantly by fissure infilling processes, accompanied by minor replacement of wall rock fragments. The location
of the veins in crush zones further supports fissure infilling as the dominant mode of vein formation. Small veinlets of quartz occasionally carrying pyrite, in the wall rock of the vein, are often irregular in outline and contain traces of sericite and less commonly biotite. These veins have apparently formed by replacement of the sedimentary host rock.

Small stringers extending from the vein into the country rock occur where the crush zone adjacent to the vein is absent. These stringers often contain small vugs lined with quartz crystals and are thus considered to have formed by fissure filling processes.

Similar vugs, although not common, are present in the veins themselves, and provide further evidence of the dominance of fissure filling processes in the formation of the vein system.

9.2 MINERALOGY

Specimens of the mineralised veins exposed in Mine Creek and from Manton's workings were examined in both polished and thin section. The dominant sulphide species present in approximate order of abundance are as follows:

pyrite > sphalerite > galena > arsenopyrite > chalcopyrite

Electrum visible under the microscope, is a minor constituent of the veins also.

The most abundant gangue mineral is quartz, associated with minor amounts of tourmaline and calcite. Sericitised fragments of wall rock are commonly present in the veins also.

The distribution of the gangue and sulphide minerals in the veins is highly variable, ranging from massive pyrite with traces of sphalerite and quartz on the one hand, to barren quartz on the other. Galena and sphalerite are both considerably less abundant than pyrite and generally account for less than 10 - 15% of hand specimen sized samples of mineralised material.
2.1 Sulphide Mineralogy

a) Pyrite:

Pyrite crystals up to 2 - 3mm across occur throughout most of the mineralised veins. The abundance of square sections under the microscope suggests that the cube is the dominant form, however pyritohedron, as evidenced by six sided sections, are present also. The crystals, which illustrate well the idiomorphic character of pyrite, are commonly fractured, the fractures being healed by quartz, which on occasion partially replaces the pyrite. Sphalerite and galena, where present, commonly replace pyrite along both fractures and grain boundaries.

As well as the massive form, noted earlier, pyrite crystals occur as subhedral to euhedral aggregates up to 1cm long, in quartz. Such aggregates are commonly partially replaced by galena.

Pyrite euhedra up to 1mm across are common in the wall rock, particularly of the footwall, to distances of up to .75m from the veins, and are also common in and along the margins of wall rock fragments within the veins.

In the coarser grained, and generally less fractured pyrite, small inclusions of foreign minerals, up to .02mm long, are quite abundant. The most common species indentified are chalcopyrite, galena, an unidentified grey opaque phase, and electrum. The inclusions are usually monomineralic, although galena-chalcopyrite and galena-electrum inclusions are sometimes present (see Fig 9.1). Where pyrite is replaced by sphalerite small inclusions of sphalerite are common in the pyrite crystals.

The origin of the inclusions in the pyrite is unclear although two possibilities must be considered as likely:

a) Replacement of pyrite by the included minerals.
b) Entrapment of the included minerals during the growth of the pyrite crystals

Little evidence exists to support the idea of entrapment of the included sulphides and electrum. Textural relationships suggest that the included species dominantly post-date the deposition of pyrite. Accordingly, an origin by
Fig 9.1 Galena (grey) and composite electrum (yellow) -galena inclusions in pyrite. See text for discussion. Polished section, plane polarised light, oil immersion. U.C.9498.
Manton's Workings (G.R. S64/366197)

Fig 9.2 Sphalerite (grey) filling fractures in and partially replacing earlier formed pyrite (yellow) euhedra and subhedra. Polished section, plane polarised light. U.C.9508.
Stamping battery, Mine Creek. (G.R. S64/365195)
replacement is inferred for the included minerals.

Some of the more rounded and spherical inclusions of chalcopyrite, could have formed by entrapment during crystallisation and cooling of the pyrite, or by exsolution, however, these phenomenon, particularly exsolution, can hardly be regarded as having resulted in the formation of the more irregularly shaped chalcopyrite inclusions, and of the galena, galena-chalcopyrite and galena-electrum inclusions.

b) Sphalerite:
Sphalerite is confined to the mineralised veins where it occurs as a replacement of pyrite. Mosaics of crystals up to several centimetres, but usually less than 1cm, across are present, often containing incompletely replaced pyrite either as subhedral crystals or relict veinlets. Fractures within and between pyrite crystals are in places infilled with sphalerite (see Fig 9.2) and replacement of pyrite by sphalerite to produce 'caries texture' is not uncommon.

Galena is occasionally associated with sphalerite, particularly along sphalerite-pyrite contacts. The sphalerite-galena boundaries are commonly scalloped being concave with respect to sphalerite. Such boundaries are common in lead-zinc deposits, and have been interpreted as both replacement and equilibrium features. The occurrence of galena along the margins of sphalerite aggregates, where these are in contact with pyrite, suggests that, in this case, replacement of galena by sphalerite is probable.

Numerous minute lath shaped or sub-spherical inclusions up to 0.01mm long are abundant in the Mine Creek sphalerite. (see Fig 9.3), and have been tentatively identified as chalcopyrite. The inclusions are considered to be exsolution products of sphalerite. Many of the inclusions show common orientations, probably parallel to cleavage, further supporting an origin by exsolution.

c) Galena:
Galena commonly replaces the other sulphide minerals, reflecting its relatively late position in the paragenetic sequence. As well as the small inclusions of galena replacing
Fig 9.3 Sub-spherical and elongate inclusions of chalcopyrite (yellow) exsolved from sphalerite (grey): polished section, plane polarised light, oil immersion U.C.9508.
Ore from stamping battery site, Mine Creek. (G.R. S64/365195)

Fig 9.4 Galena (grey) filling fractures between pyrite (yellow) euhedra and partially replacing the earlier formed pyrite. Polished section, plane polarised light. U.C.9508.
Ore from battery site, Mine Creek. (G.R. S64/365195)
pyrite crystals discussed previously, galena also occurs as interstitial infillings between pyrite euhedra (see Fig 9.4) and as veinlets infilling fractures in pyrite. In such instances galena clearly post-dates the pyrite.

Galena rarely occurs in juxtaposition with sphalerite, but where this does occur, textural evidence suggests replacement of galena by sphalerite.

Small veinlets of galena can also be observed to cross-cut quartz, and in turn be cut by quartz veinlets suggesting brecciation and quartz deposition during and after the deposition of galena.

Deformation of galena crystals is generally absent, although occasional crystals interstitial to pyrite show curved polishing pits (developed parallel to cleavage) indicating some post-depositional deformation of the mineralised veins.

Galena and sphalerite both show a close association with pyrite, which they commonly replace. This association is probably due to the relative ease of replacement of pyrite by sphalerite and galena compared with the vein silicates in which galena and sphalerite are generally absent.

d) Arsenopyrite:

Euhedral crystals of arsenopyrite up to 0.8mm, but generally 0.2 - 0.4mm across, sometimes partially replaced by silicate minerals, form a minor component of the vein assemblage. Examination of polished sections suggests that the distribution of arsenopyrite is related to the presence of altered wall rock fragments in the vein assemblage. (Polished sections in which no wall rock fragments are present lack arsenopyrite.) The arsenopyrite euhedra occur disseminated most commonly in the wall rock fragments and less commonly in the cementing quartz.

In one specimen (10202) arsenopyrite euhedra, commonly fractured and cemented with quartz forming a cataclastic texture, occur in thin veinlets parallel to the vein walls (see Fig 9.5). These veinlets are apparently shear planes, probably formed during early movement along the fracture zone, and are separated by bands of quartz up to a few
Fig 9.5 Veinlets of fine-grained fractured arsenopyrite euhedra (white) developed along a quartz (q) wall rock (w) contact. Py = pyrite. See text for discussion. Polished section, plane polarised light. U.C.9496 Manton's Workings, Mine Creek (G.R. S64/366197)
millimeters wide. Aggregates of pyrite crystals in the quartz are considerably less fractured but may show a cataclastic texture where they are in contact with arsenopyrite.

The paragenetic position of arsenopyrite is uncertain, however, its association with wall rock fragments and its proximity to the vein walls, suggests that it occupies an early position in the paragenetic sequence.

e) Chalcopyrite:

Chalcopyrite occurs in the vein assemblage in minor amounts only. As well as the previously noted modes of occurrence, chalcopyrite occurs as fracture infillings and minor replacements of pyrite euhedra, and also as interstitial infillings, with or without quartz, between pyrite euhedra. The grains are small, up to 1 - 2 mm across and irregular in outline, the shape being controlled by the surrounding minerals. In contrast to pyrite, the grains are unfractured.

Polished section examination indicates that chalcopyrite is most abundant in sections containing galena, and to a lesser extent those containing sphalerite. Galena-chalcopyrite inclusions in pyrite suggest that both these minerals probably occupy a similarly late position in the paragenetic sequence.

9.2.2 Noble Metals

a) Electrum:

Pale yellow electrum shows a very close association with pyrite, always occurring within or adjacent to it (see Fig 9.6(a)). The electrum, consisting of approximately 50 - 55% silver and 45 - 50% gold (see section 12.3.1(b)) occurs most commonly as small (up to .04 mm) irregular shaped grains filling fractures in, and partially replacing pyrite. Some grains also occur as partial replacements of pyrite along the margins of small quartz filled fractures within pyrite aggregates.

Inclusions of electrum within pyrite are also common, (see Fig 9.6(b)) and may be due to replacement of the pyrite host. An origin by exsolution cannot, however, be ruled out, although the large size of the electrum grains and their generally irregular shape suggests that exsolution is unlikely. Within the pyrite, electrum may also occur with
Fig 9.6(a) Electrum (white) and quartz (grey) infilling a fracture in pyrite (py). Polished section, plane polarised light, oil immersion. U.C.9501.
Mine Creek, near Manton's Workings. (G.R. S64/366197)

Fig 9.6(b) Rounded inclusions of electrum (white) and galena (grey) in pyrite crystal. See text for discussion. Polished section, plane polarised light, oil immersion. U.C.9501.
Mine Creek, near Manton's Workings. (G.R. S64/366197)
galena, as minute inclusions up to 0.02mm long. These composite inclusions are also interpreted as replacement features.

No visible electrum has been detected in the quartz of the vein assemblage and there is no apparent affinity for the electrum to be associated with arsenopyrite, a common association in many hydrothermal ore deposits, including the Cedar Creek reefs. Clusters of pyrite within quartz and massive pyritic material are both associated with electrum. There is also an apparent increase in the electrum content of the veins with the abundance of sphalerite. This increase may, however, be apparent rather than real, being based on qualitative polished section observation. If the increase is indeed real, it seems that deposition of the electrum, like sphalerite, occurred at a late stage in the paragenetic sequence.

9.2.3 Gangue Mineralogy

The dominant gangue mineral in the vein assemblage is quartz, with minor amounts of calcite, tourmaline and chlorite. Sericite is also present in included wall rock fragments and as small relict flakes in the vein quartz where replacement of the wall rock is incomplete.

a) Quartz:

Two varieties of quartz can be distinguished in thin section, largely on the basis of grain size, and probably represent two different generations.

i) 1st Generation:

Fine grained anhedral-subhedral quartz, 0.01 - 1mm across, occurs associated with included and partially replaced wall rock fragments, adjacent to the vein margins (see Fig 9.7). The quartz is commonly highly strained, although deformation lamellae are rarely visible, and contains numerous minute fluid inclusions. The contacts with the coarser second generation quartz are sharp, although irregular in form. Pyrite and arsenopyrite euhedra occur in the first generation quartz.
Fig 9.7 Fine grained 1st generation vein quartz containing pyrite (black) and fine micaceous wall rock selvages, mainly muscovite. (Black circular spots are bubbles in the mounting resin.) Thin section, cross polarised light. U.C.9500.
Manton's Workings, Mine Creek. (G.R. S64/366197)
ii) 2nd Generation:

Coarsely crystalline anhedral quartz generally 2 - 8 mm across is characteristic of the second generation. This quartz is almost always highly deformed with deformation lamellae being well developed (see Fig 9.8(a)) and in places almost pervasive. The orientation of the lamellae is controlled by the quartz orientation, and accordingly the lamellae are not continuous across grain boundaries. The lamellae in some grains appear curved, and are refracted by fractures within the quartz.

Fluid inclusions, generally in poorly developed trails, are almost always present, and give rise to a milky appearance in hand specimen. Crystals adjacent to pyrite aggregates, in contrast to most of the second generation quartz, contain relatively few inclusions (see Fig 9.8(b)) and often do not have well developed deformation lamellae. This may be due either to re-crystallisation of this quartz subsequent to deposition by a pressure shadow effect, or alternatively may reflect the primary character of this quartz.

There is no marked difference in inclusion density between first and second generation quartz.

b) Calcite:

Calcite is a minor component of the vein assemblage. Crystals in the 0.5 - 2 mm range, occur as anhedral to subhedral mosaics associated with quartz and pyrite. Calcite is most commonly observed in quartz veins or veinlets closely associated with wall rock or wall rock fragments, occurring as crystals interstitial to the vein quartz. Monomineralic calcite veinlets up to 1 mm across also cut both country rock and quartz veinlets.

Most of the calcite present shows an absence of deformation lamellae, although these are occasionally developed. In view of the highly strained and deformed condition of the quartz in the veins, this lack of deformation in the much more easily deformed calcite is somewhat surprising and suggests that calcite formed late in the mineralisation sequence, probably after most of the movement along the mineralised shears had ceased.
Fig 9.8(a) Highly strained and fractured coarsely crystalline second generation quartz with well-developed deformation lamellae. Less strained quartz occurs interstitial to fractured pyrite (black). Thin section (slightly thick), cross polarised light. U.C.9496 Manton's Workings, Mine Creek. (.G.R. S64/366197)

Fig 9.8(b) Pyrite (black) surrounded by clear quartz with a low density of fluid inclusions. The surrounding second generation vein quartz, in contrast, has a high density of small fluid inclusions, giving a milky appearance in hand specimen. Thin section, plane polarised light. U.C.9496.
Manton's Workings.
The origin of the calcite in the vein system is unclear, however, two alternatives seem likely. Calcium released by wall rock alteration may have reacted with CO₂ rich fluids to form calcite, or alternatively, the calcite may have been deposited from entirely magmatically derived fluids.

Ankerite, the characteristic carbonate of hypothermal vein deposits, has not been identified in the vein system.

c) Tourmaline:
Tourmaline is a rare constituent of the vein assemblage generally occurring as irregular crystals 1 - 2mm across sparsely distributed throughout the vein quartz. The crystals are generally dark brown in colour suggesting that they are of the schorl variety.

Tourmaline is characteristic of hypothermal deposits (300°C - 500°C) and its presence in the Mine Creek veins suggests that depositional temperatures were probably within this range.

d) Chlorite:
Chlorite is rare in the vein assemblage being restricted to thin veinlets associated with pyrite and arsenopyrite. Fine grained aggregates of chlorite surround the sulphide minerals, and are present near the margins of the veins. The chlorite aggregates cross-cut both wall rock inclusions and quartz veinlets. Accordingly, it seems likely that chlorite, like calcite, formed during the later stages of mineralisation.

3.2.4 Mineralogical Zonation
Mineralogical zonation is apparent, although relatively poorly developed, within the mineralised vein system. Its effects are further obscured by the patchy distribution of the sulphide minerals within the veins.

Qualitative examination shows that pyrite, arsenopyrite, quartz, calcite, tourmaline and micas (excluding those in wall rock fragments) are most abundant near the vein walls, and are often associated with wall rock inclusions. Pyrite and arsenopyrite are common along the margins of such wall
rock inclusions. Sphalerite and galena, in contrast, are absent along the margins of wall rock inclusions.

Galena and sphalerite, the late stage sulphides, occur more abundantly near the centre of the veins, generally associated with pyrite aggregates in quartz. Where the vein consists predominantly of massive pyrite with minor quartz, sphalerite is common, occurring as thin bands 1 - 2cm wide parallel to the vein walls. Galena is rarely present in the massive pyritic vein material.

Electrum is apparently associated with late stage pyrite and sulphides. Earlier formed arsenopyrite, which usually has a strong affinity for gold, shows no tendency to occur with electrum in the veins.

The mineral zonation noted above, reflects either changing composition of the ore fluid and/or, changes in the physio-chemical conditions of ore deposition.

9.2.5 Mineral Paragenesis

The textural relationships between the minerals in the vein assemblage indicate a relatively simple paragenesis, with probably only two phases of deposition occurring contemporaneously with movement on the shears now hosting the mineralised veins. It seems most likely that the mineralising episodes were quite short-lived and occurred during the post-intrusive cooling phase, related to intrusion of the Kakapotahi Granite.

The paragenetic sequence deduced by observation of textural relationships between the ore and gangue minerals is summarised in Fig 9.9. Quartz, pyrite, arsenopyrite and tourmaline were deposited during the earliest mineralising episode related to the development of the shear system. Deposition of quartz and pyrite probably continued at a reduced rate after this initial pulse of mineralisation, and may have been accompanied by minor movements on the shears and possibly the deposition of minor chalcopyrite.

This mineralising episode was apparently followed by the deposition of galena and sphalerite, and the continued deposition of minor quartz and pyrite. Deposition of sphalerite and galena may have been due to a renewed episode
<table>
<thead>
<tr>
<th>Mineral</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quartz</td>
</tr>
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<td>Pyrite</td>
</tr>
<tr>
<td>Arsenopyrite</td>
</tr>
<tr>
<td>Pyrite</td>
</tr>
<tr>
<td>Sphalerite</td>
</tr>
<tr>
<td>Galena</td>
</tr>
<tr>
<td>Electrum</td>
</tr>
<tr>
<td>Chalcopyrite</td>
</tr>
<tr>
<td>Tourmaline</td>
</tr>
<tr>
<td>Calcite</td>
</tr>
<tr>
<td>Tourmaline</td>
</tr>
<tr>
<td>Calcite</td>
</tr>
</tbody>
</table>

**Fig 9.9** Paragenesis of the Mine Creek veins, showing the estimated temperatures of deposition of the vein minerals, and their relationship to tectonic events.

See text for discussion.
of fault activity and accompanying brecciation, or have occurred near the end of a continued period of movement on the shears. In either case, galena and sphalerite deposition apparently continued until most of the movement on the shears had occurred.

Electrum, the last of the ore minerals deposited in the shear system, was apparently deposited in fractures formed during the latest phases of movement, and is therefore considered to post-date this movement. Contemporaneous deposition with late stage galena and sphalerite is indicated by the existence of composite galena-electrum inclusions in pyrite, and the general association of sphalerite and electrum noted previously.

Minor chalcopyrite was also deposited during the latest stages of fracturing.

Calcite, which forms cross-cutting veinlets within the shears, was probably the last mineral of hypogene origin to form in the vein system. Calcite deposition may have been in part synchronous with electrum deposition, however, there is no good textural evidence to confirm this.

9.3 COMPARISON OF SULPHIDE MINERALOGY WITH EXPERIMENTAL DATA

1. Fe-S System:

The high temperature portion of the binary Fe-S system has been extensively investigated, mainly as a consequence of geothermometric investigations of pyrrhotite structure and composition. Early work (Kellurud and Yoder, 1959) showed that pyrite melts incongruently to pyrrhotite and sulphur at $742^\circ + 1^\circ$C at 10 bars pressure, when excess sulphur is present. The melting point of pyrite was also shown to be pressure dependent, being raised by about $14^\circ$C per kbar, in the absence of vapour pressure.

Evidence for the pyrite $\rightarrow$ phrrhotite + sulphur transition occurring at lower temperatures has been supplied by Lambert (1973), who depressed the transition point to $355^\circ$C by lowering the sulphur fugacity of the system, and adding water and coal to the experimental charge.
Phase relations in the system have been described by Kissin (1974), and are illustrated in Fig 9.10. In spite of much research into phase relations in the low temperature fields, many enigmas still remain (for discussion see Vaughan and Craig, 1978) particularly in regard to the stability fields and compositions of the pyrrhotite species.

The only mineral phase belonging to the Fe-S system present in the Mine Creek veins is pyrite. Accordingly, assuming deposition under conditions approaching the estimated load pressure of 2 kbars (see sections 5.7 and 5.9) a maximum temperature of formation of c.770°C can be inferred for the pyrite in the vein system. The presence of co-existing arsenopyrite in the vein system, the maximum thermal stability of which is c.702°C (Barton, 1969), however, indicates that the pyrite present formed at temperatures below c.700°C.

Pyrrhotite, although absent in the vein assemblage, occurs in the hornfelsed pendant some 50m from Manton's Workings. This occurrence is described in section 8.2 The structural state of the pyrrhotite present is unknown, however, the pneumatolytic origin inferred for the formation of this pyrrhotite suggests that it is most likely to be the high temperature hexagonal form, stable over wide compositional ranges above 250°C (see Fig 9.10).

Although both pyrite and hexagonal pyrrhotite are stable up to c.700°C, hexagonal pyrrhotite is the most common Fe-sulphide in high temperature hypothermal ore deposits, possibly due to its stability at lower sulphur activities (A\textsubscript{S\textsubscript{2}}) than pyrite (see Fig 9.11). From Fig 9.11 it is evident that at a given sulphur activity (A\textsubscript{S\textsubscript{2}}) pyrite will be the stable Fe-sulphide at lower temperatures, and pyrrhotite the stable phase at higher temperatures. Furthermore it is evident that the stability field of pyrite above temperatures of 500°C - 550°C is very small, and only slight increases in temperature, or decreases in A\textsubscript{S\textsubscript{1}}, would be required for pyrrhotite to become the stable phase.

It is accordingly suggested, that although maximum temperatures for pyrite and arsenopyrite deposition of c.700°C are possible it is more likely that deposition of these
Fig 9.10 High and low temperature phase relations in the system Fe-S. Pyrite is the only mineral of the Fe-S system present in the Mine Creek veins indicating deposition at temperatures below c.750°C at moderate to high sulphur activities. See text for further discussion. (Phase relations from Kissin, 1974)
Fig 9.11  Plot of sulphur activity ($A_{S_2}$) vs temperature for the system Fe-S (from Scott, 1973). Note the small stability field of pyrite above c. 500°C (shaded) compared to pyrrhotite, the common high temperature mineral in the Fe-S system. The region of principal ore-forming environments is shown by crosses. See text for discussion.
minerals in the vein system, occurred at temperatures below 400° - 500°C.

2. Fe-Zn-S System:

The Fe-Zn-S system has also been extensively investigated, particularly with regard to the composition of sphalerite, for use as a potential geothermometer and/or geobarometer.

Barton and Toulmin (1966) identified phase relations above 400° and these are now well established. The lower temperature relationships have been investigated by Scott and Kissin (1973) and are summarised in Fig 9.12.

The original geothermometer proposed by Kellurud (1953) based on the FeS content of sphalerite has been subsequently invalidated, however, by recognition of the dependence of sphalerite composition on the activity of FeS, which may be widely varying in the pyrrhotite series (Toulmin and Barton, 1964), and on the pressure of formation (Scott, 1973).

The composition of sphalerite, may however, be used as a geobarometer within certain temperature limits (Scott and Barnes, 1971). As can be seen from the phase diagram (Fig 9.13) the composition of sphalerite in equilibrium with co-existing pyrite and pyrrhotite is nearly constant between 300° and 550°C up to pressures of about 5 kbars, thus allowing geobarometric estimates within these limits.

The presence of co-existing pyrite and sphalerite in the vein system at Mount Rangitoto allows for application of the geobarometer to estimate pressures of formation. Sphalerites from the Mine Creek veins were analysed using a JEOL JXA-58 electron microprobe (see Appendix III) and the following results were obtained:

<table>
<thead>
<tr>
<th>Sample 10203:</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wt % Fe</td>
<td>2.92</td>
<td>3.10</td>
<td>2.96</td>
<td>2.90</td>
<td>3.49</td>
<td>3.30</td>
</tr>
<tr>
<td>Wt % FeS</td>
<td>4.60</td>
<td>4.88</td>
<td>4.66</td>
<td>4.57</td>
<td>5.49</td>
<td>5.19</td>
</tr>
<tr>
<td>Mole % FeS</td>
<td>5.04</td>
<td>5.33</td>
<td>5.14</td>
<td>5.04</td>
<td>6.01</td>
<td>5.72</td>
</tr>
</tbody>
</table>

The sphalerites analysed all contain low amounts of FeS, falling in the range 4.6 - 5.2 wt %. The composition of the sphalerites falls well to the right of the univariant boundary of Scott and Kissin (1973) shown in Fig 9.12. Furthermore,
Fig 9.12 T-X projection of the low temperature portion of the system Fe-Zn-S (after Scott and Kissin, 1973) showing the composition of sphalerite co-existing with minerals of the Fe-S system. Mine Creek sphalerites lie at the extreme right of the sphalerite + pyrite field (Mole % FeS = 4.5 - 5.5), and accordingly cannot be used to provide temperature estimates, for the deposition of sphalerite in the vein system. The FeS content of the sphalerite is probably largely controlled by the activity of FeS rather than temperature.
Figure 9.13 T-X projection of the sphalerite + pyrite + hexagonal pyrrhotite solvus isobars (after Scott, 1973). Mine Creek sphalerites (4.5 - 5.5 Mole % FeS) plot in the high pressure area to the extreme right of the diagram indicating pressures of formation above 5 kbars (= 20km depth assuming normal lithostatic loads). Such pressures are unrealistically high, probably due to the lack of buffering of the effects of the activity of FeS by the presence of pyrrhotite, in the Mine Creek assemblage.
the low concentrations of FeS in the Mine Creek sphalerites do not enable application of the sphalerite geobarometer. As can be seen from Fig 9.13, the Mine Creek sphalerites plot beyond the high pressure end of the calibrated geobarometer, their low FeS content indicating likely pressures of formation in excess of 6.5 – 7 kbars. Even allowing for the unlikely possibility of $P_{H_2O}$ in the vein system exceeding $P_{total}$, crystallisation pressures in excess of 6.5 – 7 kbars seen unrealistically high. Accordingly, pressures existing in the vein system at the time of sphalerite deposition are regarded as being unknown.

The predominance of fissure filling textures, and the inferred 6 – 8km depth of burial during granite emplacement suggests that pressures in the vein system were probably close to lithostatic load pressures. It seems likely therefore that $P_{H_2O}$ in the vein system was probably less than or equal to $P_{load}$ and therefore maximum pressures of 1.5 – 2 kbars are inferred, during the formation of the Mine Creek veins.

The reason(s) for the low FeS content of the Mine Creek sphalerites is unclear, but may be due to low sulphur activities ($A_{S2}$) in the mineralising fluids. Scott (1973) notes that the composition of sphalerites in equilibrium with only one mineral of the Fe-S system i.e. pyrrhotite or pyrite may be quite variable, presumably due to the lack of buffering of the effects of the activity of FeS, which Barton and Toulmin (1966) showed to be important in determining sphalerite compositions.
§.4 GEOCHEMISTRY

A detailed investigation into the geochemistry of the mineralised veins, based on systematic channel sampling and analysis, was considered to be beyond the scope of this thesis. A limited number (eight) of selected samples from the veins exposed in Manton's Workings were, however, analysed for major elements and selected trace elements in order to determine the relationships between the noble metals, the sulphide-forming metals and the gangue minerals, particularly quartz. It was hoped that these relationships would both reveal the distribution of the gold and silver present in the veins, and provide some insight into genesis of the vein deposits. Furthermore, it was hoped to obtain some idea of the concentrations of gold and silver present in the veins, particularly in view of the speculation that surrounded the values obtained during the early days of the prospecting venture.

In order to obtain the maximum possible range in concentration of the elements analysed, samples ranging from almost pure quartz, through to massive sulphide (mainly pyrite) with minor quartz were selected and analysed. The modal compositions of the samples estimated from hand specimen examination, are given in Table 9.1.

Analysis of the samples by atomic absorption for Au and Ag and XRF (see Appendix III) for the remaining elements, was undertaken on sub-samples taken from 150g of powdered rock, which had in turn been prepared from c.300g of chips, the large samples being used in an attempt to obtain representative gold and silver values.

It had also been hoped to investigate the composition of the sulphide minerals and electrum present in the veins to determine the distribution of gold and silver in the sulphides and the composition of the electrum. A preliminary examination of some polished sections from both Mount Rangitoto and Mount Greenland was undertaken using the Otago University microprobe (see Appendix III), however, insufficient time was available to undertake further investigation. The semi-
TABLE 9.1: Modal composition of mineralised vein samples, Mine Creek, Mt Rangitoto.
Estimates are based on visual examination of hand specimens used for geochemical analysis.

<table>
<thead>
<tr>
<th></th>
<th>10200</th>
<th>10201</th>
<th>10202</th>
<th>10203</th>
<th>10204</th>
<th>10205</th>
<th>10206</th>
<th>10207</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quartz</td>
<td>80</td>
<td>85</td>
<td>85</td>
<td>100</td>
<td>65</td>
<td>70</td>
<td>27-30</td>
<td>8</td>
</tr>
<tr>
<td>Pyrite</td>
<td>15</td>
<td>5-10</td>
<td>-</td>
<td>-</td>
<td>15</td>
<td>10</td>
<td>70</td>
<td>85</td>
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<tr>
<td>Sphalerite</td>
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<td>-</td>
<td>10</td>
<td>-</td>
<td>15</td>
<td>Tr</td>
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<td>Galena</td>
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<td>-</td>
<td>5</td>
<td>-</td>
<td>5</td>
<td>20</td>
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<td>5-10</td>
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<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>2</td>
</tr>
</tbody>
</table>
quantitative results obtained in the preliminary study are included in the following sections, as appropriate.

9.4.1 Results

The results obtained in this study are summarised in Table 9.2. As would be expected, the concentrations of Si, Pb, Zn and Fe, the major gangue and sulphide forming elements, correspond closely to the estimated modal compositions of the respective samples, with Si showing an inverse correlation with Pb, Zn and Fe. Maximum concentrations of both Pb and Zn of C.10% were recorded in samples 10205 and 10207 respectively. Although only weak correlations between Fe and Pb (corr. co-efficient = + 0.13) and Fe and Zn (corr. co-efficient = + 0.50) are evident, both sphalerite and galena have an intimate association with pyrite, never occurring where pyrite is absent. The presence of pyrite is therefore considered necessary for the deposition of sphalerite and galena, the concentrations of which, however, are not closely related to pyrite abundance.

Although usually regarded as a trace element, As is present in all the samples analysed in concentrations greater than 0.01%, excluding sample 10203, and therefore is more appropriately regarded as a major element. As occurs in the samples both in arsenopyrite, and probably also as an impurity in pyrite, replacing sulphur in the crystal lattice. Accordingly, a strong correlation between Fe and As (corr. co-efficient = + 0.97 (see Fig 9.14(a)), is evident. The other major elements i.e. Pb, Zn, and Si show negligibly weak correlations with As, indicating its general absence in galena, sphalerite and quartz. Sample 10203, which appears as barren quartz in hand specimen, contains expectedly low concentrations of As (172ppm), further reflecting its general absence in the vein quartz.

a) Trace Elements:

The trace elements viz. Cu, Sn, Sb, Ag and Au (and in some samples Pb and Zn), show some interesting correlations, both with the major elements and other trace elements.
<table>
<thead>
<tr>
<th>MAJOR ELEMENTS</th>
<th>Mineralised Quartz Veins, Mine Creek</th>
<th>Wall Rock</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>10200</td>
<td>10201</td>
</tr>
<tr>
<td>SiO₂</td>
<td>83.5</td>
<td>87.0</td>
</tr>
<tr>
<td>Al₃O₇</td>
<td>0.54</td>
<td>2.02</td>
</tr>
<tr>
<td>MgO</td>
<td>0.00</td>
<td>1.56</td>
</tr>
<tr>
<td>CaO</td>
<td>0.01</td>
<td>0.84</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.10</td>
<td>0.00</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.01</td>
<td>0.02</td>
</tr>
<tr>
<td>Ti</td>
<td>0.01</td>
<td>0.04</td>
</tr>
<tr>
<td>Mn</td>
<td>0.00</td>
<td>0.02</td>
</tr>
<tr>
<td>Fe</td>
<td>7.28</td>
<td>4.20</td>
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<tr>
<td>S</td>
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<tr>
<td>Pb</td>
<td>0.19</td>
<td>2.40</td>
</tr>
<tr>
<td>Zn</td>
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<td>0.80</td>
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<tr>
<td>Cu</td>
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<td>-</td>
</tr>
<tr>
<td>As</td>
<td>0.26</td>
<td>0.21</td>
</tr>
<tr>
<td>Total</td>
<td>99.01</td>
<td>98.90</td>
</tr>
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</table>

<table>
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<th>TRACE ELEMENTS</th>
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<th>10203</th>
<th>10204</th>
<th>10205</th>
<th>10206*</th>
<th>10207*</th>
<th>10214</th>
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<tbody>
<tr>
<td>Cu</td>
<td>41</td>
<td>31</td>
<td>45</td>
<td>&lt; 1</td>
<td>82</td>
<td>35</td>
<td>53</td>
<td>46</td>
<td>51</td>
</tr>
<tr>
<td>Pb</td>
<td>470</td>
<td>-</td>
<td>-</td>
<td>72</td>
<td>-</td>
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<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Zn</td>
<td>227</td>
<td>-</td>
<td>-</td>
<td>46</td>
<td>-</td>
<td>365</td>
<td>140</td>
<td>-</td>
<td>935</td>
</tr>
<tr>
<td>Sn</td>
<td>&lt; 15</td>
<td>&lt; 15</td>
<td>&lt; 15</td>
<td>&lt; 15</td>
<td>&lt; 15</td>
<td>&lt; 15</td>
<td>&lt; 15</td>
<td>&lt; 15</td>
<td>&lt; 15</td>
</tr>
<tr>
<td>Sb</td>
<td>26</td>
<td>9</td>
<td>67</td>
<td>5</td>
<td>110</td>
<td>98</td>
<td>38</td>
<td>66</td>
<td>24</td>
</tr>
<tr>
<td>As</td>
<td>-</td>
<td>-</td>
<td>172</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>&lt; 5</td>
<td></td>
</tr>
<tr>
<td>Ag</td>
<td>36.8</td>
<td>14.0</td>
<td>128</td>
<td>&lt; 0.2</td>
<td>173</td>
<td>272</td>
<td>130</td>
<td>118</td>
<td>0.7313</td>
</tr>
<tr>
<td>Au</td>
<td>11.5</td>
<td>12.4</td>
<td>17.6</td>
<td>0.38</td>
<td>19.6</td>
<td>11.5</td>
<td>30</td>
<td>97.4</td>
<td>0.0301</td>
</tr>
<tr>
<td>Ag/Au</td>
<td>3.20</td>
<td>1.13</td>
<td>7.27</td>
<td>&lt; 0.5</td>
<td>8.83</td>
<td>23.6</td>
<td>4.33</td>
<td>1.22</td>
<td>24.3</td>
</tr>
</tbody>
</table>

TABLE 9.2: Geochemical analyses of 8 samples of mineralised vein material, and 1 sample of wall rock, Mine Creek. Major elements are considered to be all elements present in concentrations > 0.01%, plus the 'normal' major element oxides. Major elements given in weight percent, trace elements in ppm. Au, Ag analysed by atomic absorption (Lime and Marble Ltd), all other elements by XRF. * Indicates S by addition, assuming FeAss, FeS₂, PbS and ZnS. SiO₂ calculated by difference.
Fig 9.14 Variation diagrams for the Mine Creek vein assemblage.
(a) Fe vs As; (b) Ag vs Pb; (c) Au vs Ag; (d) Ag vs Zn.
Note the good correlations between Fe vs As, and Ag vs Pb, reflecting the presence of As in pyrite and the occurrence of arsenopyrite with pyrite, and the occurrence of Ag in galena. Weak correlations between Au vs Ag and Ag vs Zn are also evident. See text for further discussion.
Cu is contained in the vein assemblage both as chalcopyrite and as impurities in other sulphide minerals, particularly sphalerite and pyrite. Although not particularly strong Cu shows a positive correlation with Zn (corr. co-efficient = +0.42) and Fe (corr. co-efficient = +0.38), and a negligible correlation with Pb. The tendency for Cu to follow Zn and Fe is attributed to the similar geochemical properties of the transition metals.

A good correlation between Cu and Sb (corr. co-efficient = +0.70) is also evident. This correlation, although not surprising, is a little difficult to explain the absence of and discrete Sb-bearing minerals in the veins. Sb commonly occurs associated with both Cu and Pb in hydrothermal ore deposits, and it therefore seems most probable, that the Sb present in the Mine Creek veins occurs as impurities in chalcopyrite and galena, the major Cu- and Pb-bearing minerals. This hypothesis is further supported by the good correlation between Sb and Pb (corr. co-efficient = +0.67), and the poor correlations between Sb and Fe (corr. co-efficient = +0.12), and Sb and Zn (corr. co-efficient = +0.09) suggesting that if indeed Sb is contained in pyrite and sphalerite, which is unlikely, its distribution must be quite erratic.

Concentrations of Sn in the vein system are below the limits of detection (15ppm) for all of the samples analysed, and no Sn-bearing minerals have been observed in the vein assemblages. The low concentrations of Sn present are typical of many base-metal bearing veins around granitic bodies, and are probably due in large part to the tendency of Sn to concentrate with alkalis and rare-earth elements in the petmatitic phases of the granites, rather than becoming incorporated in the associated hydrothermal fluids.

b) Distribution of Gold and Silver:

Gold and silver, combined to form electrum, occur in the vein assemblage, associated with pyrite (see section 12.2.2). A semi-quantitative microprobe analysis of four grains of electrum from sample 10202 revealed the following compositions:
Cu, Ni and Fe, the most common impurities in native gold, are all present in detectable concentrations in the Mine Creek electrum. Cu and Ni show similar abundances ranging from 0.00 - 0.07%, whereas Fe shows higher concentrations, ranging from 0.08 - 0.50%. Comparison with the compositional data of Chisholm (1979) suggests that the Mine Creek electrum contains Cu, Ni and Fe in concentrations typical for hypogene unoxidised gold from elsewhere in the world.

It is not clear why the Mine Creek electrum should have such a high Ag content. It is possible, however, that the high Ag content may be due to deposition from Ag-enriched hydrothermal fluids derived from the Ag-rich Kakapotahi Granite, believed to be the source of the ore forming elements.

The Ag:Au ratios of the Mine Creek electrum analysed range from 1.03 to 1.15, a marked contrast to the Ag:Au ratios of 1.13 to 23.6 (see Table 9.2) obtained for the mineralised vein samples analysed. The higher Ag:Au ratios of the vein samples c.f. the electrum indicates clearly that significant quantities of Ag occur in other forms, almost certainly as impurities in the sulphides, particularly galena.

A strong correlation between Ag and Pb (corr. co-efficient = + 0.92) is evident from Table 9.2 (see Fig 9.14(b)), clearly indicating that galena, rather than electrum is the dominant Ag-bearing phase in the vein assemblage, the Ag substituting for Pb in the galena lattice. Ag and Au in contrast show only a weak correlation (corr. co-efficient = + 0.15, see Fig 9.14(c)) reflecting both the low abundance
of electrom relative to galena in the veins, and the argentiferous nature of the galena present.

Only weak correlations between Ag and Zn (corr. co-efficient = + 0.18, see Fig 9.14(d)), and Ag and Fe (corr. co-efficient = 0.04, see Fig 9.15(a)) exist indicating that neither sphalerite nor pyrite host significant quantities of Ag, although undoubtedly Ag is present in trace amounts in both minerals.

As would be expected a strong correlation between Ag and Sb (corr. co-efficient = + 0.90) exists (see Fig 9.15(b)) reflecting the similar chemical properties of these elements, and their consequent similar behaviour in ore forming environments.

A moderate correlation between Cu and Ag (corr. co-efficient = + 0.51 see Fig 9.15(c)) also exists, the Ag content of most of the analysed samples, particularly those containing galena, being greater than the concentrations of Cu present. The correlation between Cu, and Ag is almost certainly due to the similar paragenetic positions occupied by chalcopyrite, sphalerite (containing exsolved chalcopyrite) and galena, as well as the occurrence of Ag, a common impurity, in chalcopyrite.

Like Ag, Au has a distribution predicted by its chemical characteristics, showing best correlations with Fe, As and Zn, and worst correlations with Cu, Pb and Sb which have stronger affinities for Ag.

A strong negative correlation exists between SiO₂ (mainly quartz) and Au (corr. co-efficient = - 0.83, see Fig 9.16(a)) clearly indicating that very little Au occurs in the quartz, most of it being associated with sulphides in the vein assemblage. Sample 10203, which appears as barren quartz in hand specimen, contains less than 0.2ppm Au, further supporting the association of Au with the sulphides.

Strong correlations between Au and Zn (corr. co-efficient = + 0.88) and Au and Fe (corr. co-efficient = + 0.76) (see Figs 9.16(b) and (c)) indicate that gold has associations with both of these elements. According to Boyle (1978, p.25) sphalerite does not normally contain significant quantities
Fig 9.15 Variation diagrams for the Mine Creek veins. (a) Ag vs Fe; (b) Ag vs Sb; (c) Ag vs Cu. Note the poor correlation between Ag and Fe, in contrast to the good positive correlations between Ag and Sb, and Ag and Cu. See text for discussion.
Fig 9.16 Variation diagrams for the Mine Creek veins. (a) Au vs SiO₂; (b) Au vs Zn; (c) Au vs Fe; (d) Au vs As. The strong negative correlation between Au and SiO₂ reflects the absence of gold in the vein quartz, Au being contained in and associated with the sulphide minerals, particularly pyrite and arsenopyrite. This is reflected in the strong correlations between Au and Fe, and Au and As. The good correlation between Au and Zn is due to the similar position in the paragenetic sequence of electrum and sphalerite. See text for further discussion.
of Au, although in many deposits sphalerite (and galena) are indicators of high Au values. No gold or electrum has been observed in, or closely associated with sphalerite, in the Mine Creek veins, although electrum is common in pyrite in polished sections containing sphalerite. It seems most probable therefore, that the correlation between Au and Zn reflects the association of electrum with sphalerite in the vein assemblage. This is supported by the similarly late position of both minerals in the paragenetic sequence.

By contrast, the association between Au and Fe is attributed to both the observed association of electrum and pyrite, which also contains small inclusions of electrum, and to the occurrence of minor amounts of Au in the lattice of pyrite.

Au and As (corr. co-efficient = + 0.65) also show quite a strong correlation (see Fig 9.16(d)) which is attributed to the observed association of electrum with pyrite, which occurs with arsenopyrite, the dominant As-bearing mineral. The inclusion of Au in the lattice (or as minute inclusions in arsenopyrite) is also probable.

The low degree of correlation of Au with Sb, Pb and Cu (corr. co-efficients = + 0.23, -0.02 and +0.28 respectively) in the Mine Creek veins (see Figs 9.17(a) - (c)), is typical of many hypothermal precious metal deposits, these elements having much greater affinities for Ag than Au.

The Ag:Au ratio of the Mine Creek vein samples analysed ranges from 0.5 to 23.6. The Ag:Au ratio shows a strong correlation with Pb (corr. co-efficient = +0.94), clearly illustrating the dependence of the Ag:Au ratio on the galena content of the sample. In hand specimens lacking visible galena (see Table 9.1), the Ag:Au ratio of the samples ranges from 1.13 to 4.33, whereas in those samples containing galena the Ag:Au ratio is greater than 7.27, reflecting the argentiferous nature of the galena.

The tendency for the Ag:Au ratio in galena-free samples to approach that of the electrum (c. 1.05) indicates that electrum is the main Au-and Ag-bearing phase in these samples.
Fig 9.17 Variation diagrams for the Mine Creek veins. 
(a) Au vs Sb; (b) Au vs Pb; (c) Au vs Cu.
Au shows negligibly weak correlations with both Pb and Cu indicating its general absence in galena and chalcopyrite the main Pb- and Cu-bearing minerals. 
Sb, probably contained as impurities in the sulphides, also shows no significant correlation with Au.
Ag, in contrast to Au, shows strong correlations with Pb, Sb and Cu, Au having strongest affinities for Fe, As, Zn.
See text for further discussion.
The Ag content of the samples analysed ranges from 0.2ppm in the barren quartz (sample 10202) to 272ppm in the most galeniferous sample (10205). By comparison Au ranges from 0.38ppm in the barren quartz (sample 10203) to 97.4ppm in the most pyritic sample (10207). The average Ag content of the samples analysed is 109ppm, and the average Au content 25ppm. It should be noted, however, that the samples analysed were not selected with the object of calculating grades for the vein system, but rather to investigate geochemical variations.

Maximum values obtained by Newman (1977) from nine samples were 44ppm Ag, and 22ppm Au.
10.1 EMPLACEMENT

10.1.1 Relationship to Structural Features

Observations regarding the structural setting of the Cedar Creek workings on Mount Greenland are few, consisting mainly of infrequent references in the company records and Mines Statements of the times, and more recently a brief paper by Young (1964), and an unpublished report by Adamson (1966). No systematic geological survey of the area has been undertaken.

The structural geology of the area surrounding the workings has been described in section 4.1.

The auriferous veins worked in the William Tell levels strike parallel to the regional bedding, dip vertically or near vertically, and as noted by Young (1964), are emplaced along the crest of the William Tell Anticline. Veins exposed in the All Nations and Lanarch cross-cuts to the southeast, are also emplaced along the crest of the William Tell Anticline and even though the veins encountered in the latter workings were unpayably auriferous, there can be little doubt that they are a continuation of the veins worked in the William Tell levels.

Where exposed in the William Tell No. 2 level the vein system has a well developed crush zone, marked by a friable black pug carrying faint slickensides, on the southwestern wall. The pug contains numerous stringers and veinlets of quartz up to a few millimetres across, and small amounts of pyrite throughout. The presence of this pug, together with the abundance of brecciated wall rock fragments in the quartz of the vein proper, indicates that the vein was emplaced along a well defined fault in the crest of the William Tell Anticline. Adamson (1966) suggests that the slickenside orientation indicates dominantly vertical movement, although no field observations were made by the writer to verify this.
Mineralisation of the vein system, which has been intersected by cross-cuts over a length of about 250m, is apparently restricted to the ore shoot worked in the William Tell levels. Unfortunately access could not be gained to these workings, and accordingly no observations regarding vein and host rock structures in the area of the main workings were possible. Plans and records of the times, however, provide a little information.

The ore shoot worked in the William Tell levels apparently plunges steeply to the northwest and according to the records thins with depth between the No. 1 and No. 2 levels.

A fault at the northwestern end of the No. 2 level has been reported, by Fry (1936), to dislocate the ore shoot. Rather than being severed, however, it seems that the ore shoot was 'dragged' across the fault, which, according to a plan drawn by Scoble (1934), strikes at 295° and dips at 25° southwards. The displacement on this post-mineralisation fault is not known, but is unlikely to be large if the ore shoot has in fact been 'dragged' across the fault, and perseveres, as reported, in the northwestern end of the No. 2 level.

10.1.2 Vein Morphology

The quartz body exposed in the William Tell No. 2 Level contains abundant fragments of brecciated greywacke country rock. These fragments which are up to 15cm across, are commonly fractured and healed with quartz, with very little displacement evident.

The vein-wall rock contacts are sharp, although irregular in detail. To the north-east of the quartz body, exposed in the prospecting drive, veins up to 10cm across and dipping at low angles into the country rock are evident. These veins are apparently offshoots of the main quartz body and are not obviously mineralised.

Small stringers of quartz up to 15cm across, running parallel to the main quartz body were observed in the No. 2 adit near the main quartz body.
Adamson (1966) notes the presence of vugs lined with drusy quartz, in the main quartz body in the No. 2 level and states that these indicate hydrothermal emplacement of the quartz body as a cavity filling. Although this is possible, the more generally massive nature of the quartz and the brecciation of included fragments suggests that emplacement of the veins accompanied brecciation and faulting and would be more appropriately described as the infilling of an active shear zone.

Although access could not be gained to the No. 2 Level, available plans and records indicate the existence of two parallel veins, 1.5 - 2 and 2 - 3 metres thick, at the northwestern end of the No. 2 Level. These veins possibly represent a splitting of the quartz body exposed in the No. 2 adit. A plan by O'Reilly (1934) notes that in several places the "reef consists of stringers", and provides further evidence for the splitting of the quartz body, into smaller veinlets.

The sharp vein-wall rock contacts noted, together with the splitting of the quartz body into parallel veins and stringers, the brecciated nature of the exposed quartz body, and the associated pug, combine to indicate that emplacement of the quartz body occurred as an infilling process along the faulted and brecciated crest of the William Tell Anticline.

Features indicating an origin by replacement processes have not been observed.

Surface exposures in the vicinity of the Cedar Creek workings show a variety of vein forms, and relationships to bedding and structure. Veins parallel to bedding are commonly developed in argillite beds and reach a thickness of 0.5 metres or more. In such cases contacts with adjacent argillites are generally planar, whereas those with grey-wacke are irregular and often sheared (see Fig 10.1), with veinlets extending into the sandy beds along fractures. In some cases veinlets cross-cutting sandy beds in this manner are truncated by shearing in adjacent argillite beds (see Fig 10.2).
Fig 10.1 Fractured quartz vein developed parallel to bedding. Note the sharp planar contact with the argillite (right) and the irregular contact with highly fractured greywacke (left). Stringers and veinlets also extend along fractures into the greywacke.

Cedar Creek  (G.R. S57/408250)

Fig 10.2 Quartz veinlets cross-cutting greywacke, but truncated by a highly sheared argillite bed.

Swiss Creek  (G.R. S57/396254)
Sheeted veins are also present in places, forming masses 1 - 2 metres thick, with individual veins usually less than 10cm thick (see Fig 10.3). Such veins have only been observed parallel to bedding.

Lensoidal veins are less common but also strike parallel to bedding. In this case the vein shape is probably controlled largely by the competency of the adjacent greywacke beds.

Angular, brecciated fragments of wall rock are common in some veins (see Fig 10.4) and are indicative of shear zone infilling. Selvages and elongate fragments of wall rock aligned parallel to the vein walls are also common, (see Fig 10.5) and are again indicative of shearing or fault movement, during vein formation.

10.2 VEIN MINERALOGY

Due to the present condition of the William Tell workings it was not possible to obtain in situ samples of mineralised quartz. However, samples from the floor of the No. 2 adit, the battery site, the Canterbury Museum, the Canterbury University Geology Department, and float boulders in Cedar Creek were obtained, and enable some observations to be made regarding the mineralogy of the vein system.

The samples studied can conveniently be divided into three groups:

a) 'Barren' quartz; either massive or containing brecciated wall rock fragments, with or without traces of pyrite, lacking visible gold.

b) 'Auriferous' quartz; as above but containing visible gold.

c) 'Auriferous sulphide bearing quartz'; auriferous quartz carrying sulphide and/or sulphosalt minerals including pyrite, arsenopyrite, chalcopyrite, sphalerite, galena, bournonite (Cu\(\text{Pb}\)\(\text{Sb}\)\(\text{S}_3\)), boulangerite (Pb\(\text{Sb}_4\)\(\text{S}_11\)), and tetrahedrite (Cu\(\text{Sb}_4\)\(\text{S}_13\)).
Fig 10.3  Sheeted vein consisting of stringers developed parallel to bedding in the Greenland Group. The vein is cut by an analcime basalt dyke (dark grey).
Cedar Creek, Mt Greenland  (G.R. S57/440244)
Fig 10.4 Angular, brecciated wall rock fragments included in vein quartz, indicating an origin by shear zone infilling.
Float boulder, Cedar Creek. (G.R. S57/406251)

Fig 10.5 Quartz vein containing numerous wall rock selvages aligned parallel to the vein walls, indicating a probable origin by fracture infilling.
Mt Greenland track (G.R. S57/400250)
Unfortunately, the spatial and structural relationships between these types of quartz cannot be determined.

10.2.1 **Auriferous Sulphide-Bearing Quartz**

1. Sulphide Mineralogy:

Samples of 'auriferous sulphide bearing quartz', thought to be from the ore shoot worked in the William Tell No. 2 Level, were examined in polished section. The following minerals and textural relationships were noted in the limited number of samples available:

a) Chalcopyrite:

Chalcopyrite is a common constituent of the 'auriferous sulphide bearing quartz' and although never comprising more than a few percent of any specimen, is the most abundant sulphide mineral. Anhedral aggregates of chalcopyrite up to a few millimetres across are common, and frequently occur in association with small wall rock selvages within the vein quartz, however, finer irregular shaped grains up to 0.25mm across, are also present as disseminations in the wall rock (see Fig 10.6). The anhedral chalcopyrite aggregates commonly occur in close association with pyrite. Individual pyrite euhedra or subhedra up to 0.4mm across, and less commonly clusters of euhedra up to 1mm across are frequently enclosed in the chalcopyrite aggregates which have overgrown, and in some places partly replaced, the earlier pyrite. This association occurs along contacts between wall rock inclusions and the vein quartz, and in fractures in the quartz. These fractures commonly contain the sulphosalts bournonite, boulangerite and tetrahedrite, and arsenopyrite, gold, sphalerite and galena may also be present.

b) Pyrite:

Pyrite is a ubiquitous, but not abundant, constituent of the quartz. Apart from its association with chalcopyrite, as noted above, pyrite occurs as euhedral crystals up to 1.5mm across. The cubic form is dominant, although occasionally well developed hexagonal sections are observed, reflecting the presence of some pyritohedra.
Fig 10.6 Anhedral chalcopyrite (cpy) surrounding fractured pyrite (py) euhedra, and as fine grains in the wall rock of the vein. Q = quartz. Polished section, plane polarised light. U.C. 9509.
William Tell No. 2 level. (G.R. S57/408247)

Fig 10.7 Arsenopyrite crystal (aspy) partially replaced by gold (Au) developed adjacent to calcite (ca). Q = quartz, cpy = chalcopyrite. Polished section, plane polarised light. U.C. 9510.
Floor of William Tell No. 2 level. (G.R. S57/408247)
Many of the pyrite euhedra present are fractured, sometimes intensely, reflecting their brittle nature and their relatively early position in the paragenetic sequence.

Inclusions of quartz and chalcopyrite are present in some of the pyrite euhedra, and both of these minerals, in places, heal fractures in pyrite.

c) Arsenopyrite:
Arsenopyrite is subordinate to both pyrite and chalcopyrite, most commonly occurring as euhedral to subhedral cyrstals up to 0.8mm across. Clusters of fine grained arsenopyrite are less commonly present, and usually occur close to quartz-wallrock contacts (see Fig 10.7), where they may be associated with both pyrite and chalcopyrite.

Like pyrite, arsenopyrite crystals are often fractured, however, in the case of arsenopyrite the fractures are most commonly infilled by either gold or quartz. As well as infilling fractures in, and partially replacing arsenopyrite, gold commonly occurs in fractures containing arsenopyrite, or as irregular shaped grains adjacent to the arsenopyrite. These relationships, apart from illustrating the strong affinities gold has for arsenopyrite, indicate that deposition of arsenopyrite pre-dates at least some fracturing of the vein quartz, and subsequent gold-sulphide-sulphosalt mineralisation.

d) Sphalerite:
Sphalerite occurs in the vein assemblage in only minor amounts, almost always occurring in close association with chalcopyrite. The grains are generally small (0.1 - 0.2mm across) and anhedral, and usually adjacent to chalcopyrite with which it has curved mutual boundaries. The boundaries are commonly both concave and convex with respect to sphalerite, and although replacement of chalcopyrite by sphalerite cannot be ruled out, it seems more likely that the curved boundaries are equilibrium boundaries formed during contemporaneous deposition of the two phase. Rarely, planar boundaries between the two minerals are
observed, and are also considered to be due to co-precipitation, rather than replacement of chalcopyrite by sphalerite.

Less commonly anhedral sphalerite crystals up to 0.7mm across occur in fractures in the vein quartz, commonly associated with the sulphosalts. Again, curved boundaries predominate, their general form suggesting probable replacement of the sphalerite by the sulphosalts (see Fig 10.8).

Chalcopyrite blebs, generally with curved, irregular outlines are commonly present but never abundant, in most sphalerite crystals. These blebs may be due to exsolution of chalcopyrite during the cooling of the sphalerite, inclusion of chalcopyrite during crystallisation of the sphalerite, or possibly due to incipient replacement of sphalerite by chalcopyrite. Although insufficient evidence exists to rule out any of these possibilities, entrapment of chalcopyrite by sphalerite seems most likely, particularly in the general absence of unambiguous replacement textures between the two phases, and the lack of widespread evidence for exsolution.

Sphalerite commonly replaces pyrite euhedra forming classic 'caries' textures, and clearly post-dates much of the pyrite present.

e) Galena:

Galena occurs only in minute amounts in the vein assemblage, and is the rarest of the sulphide minerals present. Irregular grains up to 0.4mm across occur as interstitial infillings between quartz grains, probably healing fractures. Similar shaped grains also occur adjacent to, and within, included fragments of wall rock within the vein quartz. Partial replacement of galena by quartz and calcite is evident.

Galena also occurs with the sulphosalts (see Fig 10.9), being partially replaced by both bournonite and boulangerite. Small grains (up to 0.1mm) also occur enclosed in anhedral chalcopyrite, and have almost
Fig 10.8 Sphalerite (sp) enclosed and partially replaced by younger tetrahedrite (tt). Pyrite (py) euhedra and subhedra and irregular gold (Au) grains are also enclosed in the tetrahedrite. Polished section, plane polarised light. U.C.9458.

Fig 10.9 Irregular aggregate of galena (gn) incompletely replaced by boulangerite (bl), in quartz (Q). Ca = calcite, Au = gold. Replacement of galena probably reflects increasing activity of Sb$_2$S$_3$ in late stage fluids. See text for discussion. Polished section, partly cross polarised light. U.C.9509.
certainly been trapped by contemporaneous or subsequent crystallisation of the chalcopyrite. Galena also occurs along the margins of some chalcopyrite grains, and again gently curved low-energy boundaries are present suggesting co-precipitation of chalcopyrite and galena at conditions approaching equilibrium.

Galena has not been observed in sections containing significant quantities of tetrahedrite (Cu$_{12}$ Sb$_4$ S$_{13}$), tending instead to be associated with boulangerite (Pb$_5$ Sb$_4$ S$_{11}$) and bournonite (Cu Pb Sb S$_3$).

2. Sulphosalt Mineralogy:

The auriferous sulphide bearing quartz of the Cedar Creek veins, apart from containing the sulphides described above, contains a suite of antimony-bearing sulphosalts including bournonite (Cu Pb Sb S$_3$), boulangerite (Pb$_5$ Sb$_4$ S$_{11}$) and another variety tentatively identified as tetrahedrite (Cu$_{12}$ Sb$_4$ S$_{13}$). Unfortunately the individual sulphosalt species are extremely difficult to distinguish in polished section, having almost identical optical properties, and accordingly the identifications made in this study should be regarded as tentative until detailed microprobe analysis of the specimens are available. Unfortunately, insufficient time was available to undertake such a study of the ores, however, the presence of both bournonite and boulangerite in the vein assemblage was confirmed by X-ray diffraction analysis of powdered sulphosalts. Unfortunately the scarcity of sulphosalt bearing quartz available for study prevented the more widespread use of X-ray diffraction as an identification tool.

A semi-quantitative microprobe analysis of some of the sulphosalts present revealed three species of differing copper content. These species are interpreted as being bournonite (Cu = 13.5 - 14.7 wt%), boulangerite (Fe = 0 wt%, Cu = 0 wt%) and jamesonite (Cu = 0 wt%, Fe = 2.8 wt%), bournonite being most abundant of the sulphosalts probed.

Although jamesonite is reputedly present in sample 10217, microscopic identification indicates that the jamesonite reputedly present, is in fact boulangerite. Nevertheless,
preliminary microprobe analysis suggests that small amounts of jamesonite are present in some samples at least. Furthermore, some of the microscopically identified boulangerite may in fact be jamesonite, which has almost identical optical properties. As jamesonite has not been identified petrographically, its probable presence can only be considered in general genetic and petrogenetic terms.

a) Boulangerite \( (\text{Pb}_5 \text{Sb}_4 \text{S}_{11}) \)

Boulangerite occurs both as equigranular unoriented aggregates, and as distinct needles or fibres (see (Fig 10.11)) within the vein quartz. Unoriented aggregates of boulangerite crystals up to 2mm across occur in fractures in the vein quartz, and are commonly associated with the sulphide minerals arsenopyrite, chalcopyrite and galena. Arsenopyrite occurs in boulangerite aggregates as an euhedral-subhedral crystals commonly slightly rounded, up to 0.2mm across. The arsenopyrite crystals are often fractured, the fractures being healed by boulangerite.

Irregular intergrowths are common with both bournonite (see Fig 10.12) and galena. Where intergrown with bournonite, boulangerite generally has an elongate or platy habit in contrast to bournonite which occurs as unoriented equigranular mosaics. The boundaries of the intergrowths are generally sharp, and follow margins of the boulangerite crystals, reflecting the more idiomorphic character of boulangerite. Intergrowths of boulangerite and bournonite are commonly fractured, the fractures generally being healed by quartz.

Intergrowths with galena are also common (see Fig 10.9) where galena is present, and are probably due to replacement of galena by boulangerite, a common phenomenon in many hypogene, sulphosalts bearing ore deposits.

Chalcopyrite commonly occurs with boulangerite, as well as the other sulphosalts, in mineralised fractures up to a few millimetres across in the vein quartz. Gently curved boundaries suggest cogenetic deposition of both minerals, although the presence of occasional
Fig 10.11 Radiating needles of boulangerite (bl) in vein quartz. Polished section, cross polarised light, oil immersion. U.C.9491.

Sample held by Canterbury Museum, probably from William Tell No. 2 level.

Fig 10.12 Intergrowth of boulangerite (bl) and bouyoumite (bo). See text for discussion. Polished section, cross polarised light, oil immersion. U.C. 9491.

Sample, probably from William Tell No. 2 level, held in Canterbury Museum.
angular inclusions of chalcopyrite in boulangerite, may reflect partial replacement of chalcopyrite by boulangerite.

Gold is commonly associated with boulangerite, and the other sulphosalts and sulphides in mineralised fractures in the vein quartz.

Elongate or fibrous needles of boulangerite occur disseminated in the vein quartz. The needles, occurring either singly or in radiating clusters, are orthorhombic in cross section and lack terminations. A strongly elongate or fibrous habit is common, with the needles which are occasionally curved attaining lengths of up to 5mm. Striations parallel to the long direction of the needles (001), are well developed.

b) Bournonite (Cu Pb Sb S₃)

Bournonite has an almost identical occurrence to boulangerite in the vein assemblage. Equigranular anhedral mosaics of crystals up to 2mm across are common in fractures in the quartz, and as noted, may be intergrown with boulangerite (see Fig 10.12). Bournonite shows a stronger tendency to be associated with chalcopyrite than does boulangerite, which has a closer association with galena, commonly partly replacing earlier formed chalcopyrite. Small irregular shaped blebs (up to 0.02mm) of chalcopyrite, commonly with curved boundaries, are often present in bournonite crystals. Their origin is uncertain, however, it seems most probable that the chalcopyrite has been entrapped during contemporaneous crystallisation of bournonite. Origins due to exsolution or incomplete replacement cannot, however, be ruled out.

Bournonite and chalcopyrite commonly have gently but irregularly curved mutual boundaries (see Fig 10.13) interpreted to be equilibrium boundaries produced by cogenetic crystallisation. Although partial replacement of chalcopyrite by bournonite, or vice versa, is possible, the boundaries are neither consistently concave nor convex with respect to either mineral suggesting probable equilibrium boundaries. Gold shows a tendency to occur as irregular grains along these boundaries (see Fig 10.14)
Fig 10.13 Intergrowth of bournonite (bo), chalcopyrite (cpy), partially replaced by gold (Au). See text for discussion. Polished section, cross polarised light, oil immersion. U.C.9458.

Sample from University of Canterbury, Geology Department collection.

Fig 10.14 Gold (yellow) forming veinlet along the boundary between chalcopyrite (cpy) and bournonite (bo) by replacement of both chalcopyrite and bournonite. Py = pyrite. (Black areas are polishing pits.) Polished section, cross polarised light, oil immersion. U.C.9458.

Geology Department collection.
often protruding into the chalcopyrite. In such instances the gold is regarded as being a late stage mineral, partially replacing chalcopyrite and to a lesser degree bournonite.

In some bournonite grains twinning is evident, usually being of simple lamellar form, although occasionally polysynthetic twins which intersect at almost 90°, are observed. Ramdohr (1969) attributes both types of twinning in bournonite to growth processes rather than the effects of post-depositional crystallisation.

As with boulangerite, bournonite also has a close association with gold, both minerals commonly infilling the same fractures in the vein quartz. Cogenetic textures are common, although the presence of veinlets of gold cross-cutting bournonite indicates that gold deposition, may, in part at least, post-date bournonite deposition.

c) Tetrahedrite (Cu$_{12}$ Sb$_4$ S$_{13}$)

The third sulphosalt species present, tentatively identified as tetrahedrite, also has a close association with gold, with which it shows a variety of replacement and equilibrium textures. Tetrahedrite most commonly occurs as anhedral aggregates of crystals up to a few millimetres across, usually associated with gold and arsenopyrite in fractures in the vein quartz. Gold commonly forms veinlets cross-cutting tetrahedrite aggregates (see Fig 10.15), and may also partially replace tetrahedrite.

Angular fragments of tetrahedrite are common in gold grains, and indicate considerable fracturing occurred prior to deposition of gold and after deposition of tetrahedrite.

Arsenopyrite euhedra (0.01 - 0.05mm) are also common in tetrahedrite grains, and may also be fractured, the fractures most commonly being healed by gold. Pyrite euhedra of similar size to the arsenopyrite euhedra, are also present in tetrahedrite which may partially replace the pyrite, to produce 'caries' textures.

No close associations between tetrahedrite and either bournonite or boulangerite have been observed,
Fig 10.15  Tetrahedrite (grey) fractured and extensively replaced by gold (yellow). See text for discussion. Polished section, plane polarised light. U.C.9458.

Geology Department (University of Canterbury) collection.
tetrahedrite showing a tendency to occur with gold and to a lesser degree chalcopyrite, rather than bournonite or boulangerite. Although all these minerals have been observed in some polished sections, tetrahedrite has not been observed adjacent to the other sulphasals.

An unidentified pale-gray mineral, translucent in cross polarised light, also occurs with tetrahedrite, particularly along gold-tetrahedrite boundaries. The origin of this phase is uncertain, however, the limited observations made suggest that this phase is possibly a reaction rim, or alternatively an early replacement phase forming at the expense of tetrahedrite.

3. Gold:

Gold occurs in the sulphide bearing auriferous quartz most commonly as fracture infillings (see Fig 10.16), often associated with sulphide or sulphasalt minerals. Grains are commonly irregular in shape, forming stringers generally less than 1mm wide, although occasionally attaining thicknesses of 4 - 5mm. Being a late stage mineralisation product gold exhibits a variety of textural relationships with the sulphides and sulphasals.

Gold exhibits a strong affinity for arsenopyrite, a common association in many gold bearing ore deposits, infilling fractures in and partly replacing arsenopyrite (see Fig 10.7). In such cases gold clearly post-dates the earlier formed higher temperature arsenopyrite. In contrast to arsenopyrite, early formed pyrite has no close association with gold, although both minerals may occur together in fractures in the vein quartz.

Gold may also partially replace chalcopyrite forming small crystals protruding into the latter. This is particularly common where chalcopyrite has a mutual boundary with bournonite, in which case partial replacement of the bournonite may also occur. Replacement of chalcopyrite by gold is also apparent in the absence of contiguous sulphasals, and reflects golds late paragenetic position.

Rounded inclusions of gold, up to 0.4mm across are present in both tetrahedrite (see Fig 10.17) and bournonite with which
Fig 10.16(a) Irregular veinlets of gold (yellow) occurring as fracture infillings in vein quartz. Traces of sulphosalts minerals (grey-white) are also present. Polished section, partly cross polarised light. U.C.9491.

Fig 10.16(b) Veinlets of gold (yellow) surrounding a wall rock fragment (w), and extending along fractures into the vein quartz (Q). Polished section, plane polarised light. U.C. 9458.

Geology Department (University of Canterbury) collection.
Fig 10.17  Gold (Au) and pyrite (py) included in tetrahedrite (tt). Pyrite is replaced by tetrahedrite and gold. Polished section, plane polarised light. U.C. 9458.

Geology Department (University of Canterbury) collection.

Fig 10.18  Gold (yellow) replacing earlier formed pyrite (py) both from the core and from the margins producing 'caries' textures. Chalcopyrite (cpy) is also replaced by gold. Polished section, plane polarised light, oil immersion. U.C.9458.

Geology Department (University of Canterbury) collection.
the included gold is generally regarded as being cogenetic. Veinlets of gold, up to 0.6mm across, are also present in bournonite and tetrahedrite, and to a lesser degree boulangerite, where they most commonly occur as fracture infillings. In such cases gold deposition clearly post-dates the formation of the sulphosalts. Replacement of tetrahedrite by gold is also apparent providing further evidence of the late paragenetic position of gold.

Commonly, particularly where gold is abundant, curved equilibrium boundaries with bournonite, and to a greater degree chalcopyrite and tetrahedrite, are common reflecting cogenetic deposition of chalcopyrite and these sulphosalts with gold. Equilibrium boundaries with boulangerite have not been observed, gold usually occurring as veinlets in or partial marginal replacements of boulangerite. This is probably due to the earlier position, compared with the other sulphosalts and gold, that boulangerite occupies in the paragenetic sequence.

Gold also occurs as small rounded inclusions in pyrite which may in turn be enclosed in tetrahedrite. The origin of this gold is unclear, however, in view of the generally late position of gold in the paragenetic sequence it seems most likely that these inclusions are due to the incipient replacement of earlier formed pyrite by gold (see Fig 10.18).

In contrast to the electrum in the Mount Rangitoto veins, the gold present in the Cedar Creek veins is a dark yellow colour, reflecting its high fineness, which microprobe analysis indicates is greater than 95%.

4. Gangue Mineralogy:

a) Quartz:

Quartz is easily the most abundant gangue mineral present in the Cedar Creek reefs, accounting for over 95% of the vein material. Thin section examination reveals the presence of two types of quartz. Massive, coarsely crystalline milky white quartz is easily the most abundant variety present making up the bulk of the vein material. The quartz consists of anhedral mosaics of grains 1 - 2mm across. Most of the grains are slightly to moderately strained, and exhibit
undulose extinction, although occasionally strain free grains are observed. Occasional small cavities up to a few millimetres across occur within this quartz, and are commonly lined with small quartz euhedra.

Angular brecciated quartz fragments, generally less than 1mm across, cemented by very fine grained crushed quartz, are present in fractures in the more coarsely crystalline variety of quartz. The fragments are generally only slightly to moderately strained. Contacts between the white massive crystalline quartz and the brecciated quartz are sharp, further indicating that the brecciated quartz is a fracture infilling.

Unfortunately, the relationship of the two types of quartz to the sulphide and gold mineralisation is unknown, primarily due to an insufficiency of mineralised material from which to prepare both polished and thin sections. It seems reasonable to suppose, however, that the brecciated quartz, which like the mineralisation is a fracture infilling, may be associated with the gold and sulphide mineralisation.

Polished section examination shows that the coarsely crystalline massive quartz is mineralised also, commonly containing needles of boulangerite disseminated within it. Gold also occurs in this quartz infilling fractures and forming stringers between quartz crystals.

b) Calcite:
Calcite occurs as fracture infillings in the vein quartz, forming irregular discontinuous veinlets up to several centimetres long by a few millimetres wide. Occasionally calcite is associated with gold in these fractures, suggesting a similarly late origin for the calcite. No association between calcite and the sulphosalts has been noted.

c) Wall Rock Fragments:
Wall rock fragments are ubiquitous in the vein quartz, commonly occurring as irregular shaped selvages up to a few millimetres across, along fractures in the quartz. These selvages are commonly associated with pyrite euhedra, and less commonly with chalcopyrite and arsenopyrite. Stringers of gold are sometimes developed along the contacts of such selvages with the vein quartz.
The wall rock material, which due to its crushed nature is always very fine grained, consists almost entirely of chloritic material, the quartz originally present in the metasediments having been removed by hydrothermal processes operating during ore deposition.

10.2.2 'Auriferous' Quartz

One sample (10209) of auriferous quartz, lacking any sulphide minerals apart from pyrite, was obtained from the Canterbury Museum. The sample consisted of massive vein quartz with slightly waxy or oily lustre, containing numerous wall rock fragments ranging in size from less than a millimetre to 7-8mm across. The larger fragments are almost always angular, and some have undergone fracturing and subsequent healing by quartz, although in many cases no significant displacement of the fracture fragments is evident. This quartz, apart from containing visible specks of free gold, is almost identical in appearance to the barren quartz at present exposed in the No. 2 level adit, and it is considered likely that the quartz is from near this locality.

Pyrite which accounts for less than one percent of the sample occurs as small euhedra 1 - 2mm across, commonly developed along wall rock quartz contacts, and also as euhedra disseminated throughout the quartz. Selvages of wall rock, ranging in size from less than 1mm to several millimetres also occur disseminated throughout the vein quartz, and have almost certainly formed by crushing and grinding of the wall rock due to fault movement contemporaneous with hydrothermal activity and mineralisation.

Gold occurs as irregular shaped grains scattered sparsely throughout the quartz. The grains, which range in size from barely visible to a few millimetres across, show no tendency to occur along the contacts between quartz and the larger wall rock fragments, preferring instead the vein quartz with its minute chloritic selvages. Sulphides (excluding pyrite) and sulphosalts are absent in the auriferous vein material, and there is no noticeable tendency for gold to be associated with the pyrite present.
Due to the inaccessibility of the main workings and the paucity of material available for examination few definite conclusions can be drawn regarding the relationship(s) of the auriferous quartz to the auriferous sulphide and sulphosalt bearing quartz. The absence of sulphides viz. galena, arsenopyrite, sphalerite and chalcopyrite, and sulphosalts in the auriferous quartz suggests that the mineralisation of the auriferous quartz may have occurred after deposition of these minerals, and therefore be a latest phase in the mineralising process. Alternatively, the gold present could be of a different generation to that present in the auriferous sulphide bearing quartz, either predating or post-dating deposition of the sulphides and sulphosalts.

Although it is not possible to discern why the two styles of mineralisation viz. sulphide-sulphosalt-gold and gold-quartz apparently occur in such close proximity in the same vein system, it is obvious that different mineral assemblages do occur in the veins, and reflect at least minor differences in the mineralising processes operating during vein formation, and also possibly the effects of structural control on mineralisation style within the vein system.

10.2.3 Barren Quartz:

Although the Cedar Creek reefs are traceable along strike for several kilometres, known economic gold mineralisation as described in the two preceding sections (10.2.1 and 10.2.2) is confined to the William Tell workings. The remainder of the reef system consists of barren quartz, composed almost entirely of milky white, massive, coarsely crystalline quartz, generally lacking wall rock fragments and calcite. Pyrite is occasionally present as euhedra up to a few millimetres across, along wall rock-vein contacts. Sulphide (excluding pyrite) and sulphosalt minerals, and gold, are all absent or extremely rare in the barren quartz.

The relationship(s) of the barren quartz to the sulphide-bearing auriferous quartz (section 10.2.1) and the auriferous quartz (section 10.2.2) cannot be determined by observation
due to the inaccessibility of the William Tell workings. Some inferences may, however, be drawn. The existence of a mineralised ore shoot in the William Tell workings in a generally barren or uneconomically auriferous and extensive vein system, suggests that the mineralisation in the William Tell workings is structurally controlled. The nature of the structure(s) is unknown, however, the magmatic origin for the metals and sulphur in the deposit, indicated by the sulphur isotope data available (see Chapter II), suggests that faults or shears in the vein system extending to depth are the most likely structures controlling the mineralisation. Suitable mineralising structures, such as the faults or shears envisaged, have not been located elsewhere in the vein system, consequently the veins generally consist of barren quartz with minor pyrite, formed by wall rock alteration processes, gold and sulphide-sulphosalt mineralisation being absent.

10.3 PARAGENESIS

The textural relationships of the minerals in the vein assemblages, and the inferred distribution of the quartz types recognised, indicates a relatively simple paragenesis for the Cedar Creek veins. The inferred paragenetic sequence is shown in Fig 10.19.

Barren, milky white massive quartz was deposited in a normal fault zone developed in the axial region of the William Tell Anticline, forming an extensive vein system which hosts the mineralisation worked in the William Tell workings. Pyrite, formed by wall rock alteration processes accompanied this quartz deposition.

Brecciation of the early formed quartz possibly accompanied by penecontemporaneous granite emplacement was followed by a period of sulphide mineralisation characterised by the deposition of the relatively high temperature sulphides pyrite and arsenopyrite, followed by chalcopyrite probably in part contemporaneous with pyrite and arsenopyrite. Minor sphalerite may also have been
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<td>CONTINUING BRECCIATION ACCOMPANIED BY MOVEMENT ON THE WILLIAM TELL FAULT</td>
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Fig 10.19 Paragenesis of the Cedar Creek veins, showing the inferred temperatures of deposition, and tectonic events occurring before and during mineralisation. The paragenetic sequence reflects ore deposition occurring at successively lower temperatures, accompanied by changes in the composition of the mineralising fluids. See text for discussion.
deposited synchronously with chalcopyrite.

Brecciated quartz and sulphide minerals indicate that brecciation of the vein assemblage occurred prior to the last stage of mineralisation. Deposition of sphalerite, chalcopyrite and galena accompanied and followed this brecciation. As deposition of the sulphide minerals waned sulphosalt species began to be deposited in the order boulangerite, bournonite, (jamesonite?) and tetrahedrite. Equilibrium boundaries and intergrowths between the sulphosalt species, particularly boulangerite-bournonite and bournonite-tetrahedrite, indicates considerable depositional overlap of these species.

Gold deposition occurred during the latest stages of mineralisation, being the last of the ore minerals to be deposited.

The paragenetic position of gold contained in the 'auriferous quartz', lacking sulphides or sulphosalts, is unclear. It is possible that this gold either pre-dates or post-dates deposition of the gold associated with the sulphosalts.

Calcite was the last of the vein minerals to be deposited.

The simple paragenesis observed in the Cedar Creek veins is typical of that encountered in many mesothermal sulphantimonide-bearing gold deposits, and reflects the combined effects of progressive deposition with decreasing temperatures accompanied by compositional changes in the mineralising fluids.

10.4 COMPARISON OF ORE MINERALOGY WITH EXPERIMENTAL DATA

The systems Fe-S, Fe-Zn-S, of which the minerals pyrite and sphalerite are members, have been described in the preceding Chapter (section 9.3) and will be considered here by way of summary only.
The presence of co-existing pyrite and arsenopyrite in the Cedar Creek veins indicates that deposition of these species occurred at temperatures below c.700°C, the maximum temperature at which arsenopyrite is stable (Barton, 1969). The absence of pyrrhotite in the vein assemblages suggests however, that deposition of pyrite and arsenopyrite probably occurred at temperatures of 500°C or less, pyrite only being stable over a narrow range of temperature and $A_{S_2}$ conditions above this temperature (see Fig 9.11).

The composition of the Cedar Creek sphalerites has not been investigated, and accordingly yield no useful information regarding temperatures and pressures of formation. The occurrence of chalcopyrite inclusions in some of the sphalerites observed is considered to be due to entrapment of chalcopyrite during sphalerite crystallisation, rather than exsolution of chalcopyrite during cooling of the host sphalerite. The absence of demonstrably exsolved chalcopyrite in the Cedar Creek sphalerites means, the chalcopyrite-sphalerite exsolution geothermometer cannot be applied to the Cedar Creek sphalerites, and accordingly no pressure or temperature estimates using these minerals can be made.

In contrast to the lack of information obtained from the sulphides in the Cedar Creek veins, the sulphosalts present yield some information regarding the physical and chemical conditions of ore deposition.

10.4.1 The Lead-Antimony Sulphosalts

a) Pb-Sb-S System:

The lead sulphantimonides lie on the binary join PbS (galena) - $Sb_2S_3$ (stibnite) in the system Pb-Sb-S. Three naturally occurring phases, as well as several synthetic phases, lie on this join, the natural phases being boulangerite (5PbS : 2$Sb_2S_3$) stable below 638°C, robinsonite (6PbS : 5$Sb_2S_3$) stable between 582°C and 318°C and zinckenite (PbS : $Sb_2S_3$) stable below 545°C. Phase relations in the system are shown in Fig 10.20.
Fig 10.20 Phase relations in the system PbS-Sb₂S₃ (from Craig et al. 1973). The stability fields of the naturally occurring phases boulangerite, robinsonite and zinckenite are shown. See text for discussion.

Fig 10.21 Log A vs temperature plot of the sulphidation curves of the lead sulphantimonides (from Craig et al. 1973). The sulphidation curves, combined with the observed paragenesis, indicate deposition of the sulphosalt species at successively decreasing temperatures and activities of sulphur. See text for further discussion.
Experimental work by Craig et. al. (1973) indicates that boulangerite has a maximum thermal stability of 638°C, breaking down above this temperature to form a PbS enriched phase (Phase I), which has not been recorded in nature, and a Sb$_2$S$_3$ enriched liquid. Below 638°C boulangerite is stable for the entire range of geologically interesting temperatures and therefore of little use as a geothermometer.

The presence of boulangerite in the Cedar Creek veins indicates maximum temperatures at the time of deposition of 638°C, and a relatively low activity of Sb$_2$S$_3$. High activities of Sb$_2$S$_3$ would have resulted in deposition of robinsonite or zinckenite below c.550°C both of which contain higher Sb$_2$S$_3$: PbS ratios than boulangerite).

Jamesonite (Pb$_4$ FeSb$_6$ S$_{14}$) contains constituent iron and plots to the Fe side of the PbS - Sb$_2$S$_3$ binary. If iron is present at the time of lead sulphantimonide formation, and the activity of sulphur (A$_{S_2}$) is within the field shown in Fig 10.21, jamesonite will be deposited as the Pb-Sb-bearing phase. If however, A$_{S_2}$ lies outside this field, for a given temperature of deposition, iron sulphides would form and co-exist with Fe-free lead sulphantimonides.

The Cedar Creek assemblage, containing both boulangerite which co-exists with pyrite, and jamesonite (tentatively identified by microprobe), probably reflects the effects of decreasing A$_{S_2}$ during sulphosalt deposition. Although the paragenetic position of jamesonite is unknown, it seems probable that deposition occurred as a response to decreasing A$_{S_2}$ in the hydrothermal fluids. Pyrite and boulangerite, stable at higher activities of sulphur would become unstable as A$_{S_2}$ decreased, giving way to deposition of jamesonite.

Unfortunately, jamesonite is stable over a wide range of temperatures and pressures, and cannot, therefore, be used to indicate likely P-T conditions of formation. If, however, the temperature of jamesonite deposition is known,
its presence provides a reasonable estimate of $A_S^2$ values during deposition.

b) $\text{Cu}_2\text{S-Sb}_2\text{S}_3$ System:

Tetrahedrite is the most common antimonian copper compound, and indeed sulphosalt, in the Earth's crust. In spite of considerable research the exact composition of tetrahedrite remains unknown, but probably lies between $\text{Cu}_3\text{SbS}_3$ and $\text{Cu}_{12}\text{Sb}_4\text{S}_{13}$ (Skinner et. al. 1972). (The latter, generally accepted formulae, is adopted for tetrahedrite in this study, in the absence of any compositional data, from the Cedar Creek tetrahedrite.) Solid solution with tennantite ($\text{Cu}_{12}\text{As}_4\text{S}_{13}$) occurs, and most naturally occurring tetrahedrites contain at least some arsenic replacing antimony in the tetrahedrite lattice.

Skinner et. al. (1972) investigated phase relations in the $\text{Cu}_2\text{S-Sb}_2\text{S}_3$ system, which is dominated by a join between chalcocite ($\text{Cu}_2\text{S}$) and stibnite $\text{Sb}_2\text{S}_3$, over the complete range of geologically interesting temperatures, and suggested that tetrahedrite had an extensive composition range which was dependent on temperature, but at all temperatures was less sulphur-rich than $\text{Cu}_{12}\text{Sb}_4\text{S}_{13}$. Accordingly, they consider that tetrahedrite compositions lie to the sulphur rich side of the $\text{Cu}_2\text{S-Sb}_2\text{S}_3$ join.

Heating experiments undertaken by Skinner et. al., indicate that tetrahedrite breaks down by reaction to form famatinite ($\text{Cu}_3\text{SbS}_4$) and a sulphur poor phase (Phase B) unknown in nature, at temperatures of $543^\circ + 2^\circ\text{C}$. Tetrahedrite remains a stable phase below this temperature, which is accordingly considered to be the maximum temperature of deposition of the Cedar Creek tetrahedrite.

c) $\text{Cu}_2\text{S-PbS-Sb}_2\text{S}_3$ System:

Goodell (1975) reviewed phase relations in the ternary system $\text{Cu}_2\text{S-PbS-Sb}_2\text{S}_3$, and noted that the system contains a number of conflicting assemblages. Bournonite has been reported with most other phases in the system, although is prevented from co-existing with chalcocite by the common galena-tetrahedrite association. According to Goodell (1975)
the three phase assemblage galena-tetrahedrite-bournonite is common, often occurring as equilibrium intergrowths. This assemblage, however, conflicts with the assemblage tetrahedrite-boulangerite which occurs in the more antimony rich portion of the phase diagram. This conflict may, according to Goodell, be resolved if bournonite is preserved metastably in the galena-tetrahedrite-bournonite assemblage, or alternatively if the tetrahedrite in these assemblages contains arsenic.

Many paragenetic sequences take a path through galena to boulangerite-galena to tetrahedrite-bournonite-galena to tetrahedrite-bournonite to tetrahedrite e.g. Chang and Bever (1973), Goodell (1975), reflecting both the differentiation of copper and lead, and the increasing activity of \( \text{Sb}_2\text{S}_3 \) with time.

Although the sulphantimonides in the Cedar Creek assemblage do not provide good information regarding temperatures and pressures of deposition, consideration of the paragenetic sequence and experimental relationships provides some useful information regarding changes in the activity and coposition of the mineralising fluids with time. The observed paragenesis of the sulphosalts in the Cedar Creek reefs is as follows:

   galena - boulangerite - (jamesonite?) - bournonite - tetrahedrite.

This trend reflects a decreasing activity of sulphur and PbS in the mineralising fluids, accompanied by increasing activity of \( \text{Sb}_2\text{S}_3 \) in the late stage mineralising fluids, as well as the differentiation of copper and lead noted previously. The late paragenetic position of tetrahedrite suggests that the activity of arsenic may also have been increasing in the late stage fluids, a commonly reported phenomenon often resulting in the late stage formation of sulpharsenides.

Activities of sulphur indicated by the sulphosalts assemblages in the inferred temperature range of deposition \((400^\circ - 250^\circ \text{C})\) probably ranged from approx. \(10^{-4}\text{ atm}\) during the earlier higher temperature stages of sulphide and boulangerite deposition, to approx. \(10^{-12}\text{ atm}\) during the latest cooler stages of deposition.
Maximum temperatures of sulphosalts deposition of approximately 530° - 540°C are indicated by the presence of bournonite and tetrahedrite in the vein assemblage. These temperatures, however, intuitively seem too high. The absence of biotite in the wall rock adjacent to the veins, and the low degree of metamorphism of the metasediments in general, indicate that mineralisation temperatures probably did not exceed 350° - 400°C which is accordingly regarded as a more likely maximum mineralisation temperature.

The absence of bornite, which is stable at similar temperatures and sulphur activities to the lead sulphantimonides in the vein assemblage, suggests that deposition of the sulphantimonides may have occurred above 230°C, the maximum thermal stability of bornite (Scott, 1974). Accordingly, 230°C is regarded as a likely minimum temperature of sulphantimonide deposition, although gold deposition could have occurred at temperatures significantly lower than this.

Attempts to derive more accurate mineralisation temperatures by fluid inclusion studies were unsuccessful due to the small size of inclusions present in the quartz.

Pressures within the vein system during mineralisation are unknown, however, the fissure infilling nature of the veins suggests the pressures within the veins during mineralisation were probably considerably less than lithostatic load pressures.

10.5 GEOCHEMISTRY

A detailed investigation of the geochemistry of the Cedar Creek veins was impossible due to the inaccessibility of the main workings of the William Tell No. 2 level. Four samples of vein quartz (samples 10209 - 10217, see Table 10.1) were, however, analysed, and one sample of the fault pug adjacent to the vein exposed in the accessible portion of the No. 2 level (sample 10208) was also analysed.
TABLE 10.1 : Geochemical analyses of Cedar Creek quartz vein samples. Major elements, including oxides are given in weight percent, trace elements in ppm unless otherwise indicated. Au and Ag analysed by AA spectroscopy (Lime and Marble Ltd), all other elements by XRF.

* Fe calculated by assuming FeS₂ and subtracted from total Fe₂O₃.
Sample localities are given in Appendix II.
The results obtained reflect the mineralogy of the vein quartz. Sample 10209 consisting of auriferous quartz lacking sulphides and containing abundant wall rock selvages, contains a lower SiO$_2$ content and higher Al$_2$O$_3$, Fe$_2$O$_3$, MnO, MgO, CaO, Na$_2$O, and K$_2$O contents than the other vein samples analysed, which only contain traces of included wall rock. Concentrations of the economically interesting trace elements are generally similar to those obtained in sample 10215 which consists of barren quartz. S and As, however, show a slight enrichment in sample 10209, almost certainly due to the presence of trace amounts of pyrite in this sample.

Low Ag values (~0.2ppm) occur in the auriferous quartz, reflecting the absence of silver-bearing sulphides or sulphosalts. Au (11.7ppm) is present as fine grained free gold, and probably contains up to 4 - 5% Ag, as indicated by microprobe analysis.

Samples 10216 and 10217 contain sulphide minerals, sulphantimonides, and free gold, totalling 4 - 5% of the rock, the remainder being vein quartz with only minor wall rock selvages present. This mineralisation has resulted in addition of the sulphide and sulphosalt forming elements Cu, Fe, Pb, Sb, S, Zn and As. Sn also shows a slight enrichment in these samples (21 - 35ppm) compared to the barren quartz (<15ppm). This increase in Sn content accompanying sulphide-sulphosalts-gold mineralisation suggests a possible magmatic origin for the ore forming fluids, a suggestion supported by sulphur isotope data (see Chapter 11).

Maximum concentrations of Cu (1.5%), Pb (0.7%) and Sb (0.3%) occur in sulphantimonide bearing samples (1216 and 10217). These samples also contain the highest Ag (22ppm) and Au (1070ppm) values, the Au and Ag occurring in both the sulphides and sulphosalts, as well as in free gold. No silver sulphides or sulphosalts have been identified in the Cedar Creek assemblage.

Semi-quantitative microprobe analysis of chalcopyrite, pyrite and bournonite from sample 10217 (see Table 10.2) indicates that Au and Ag occur in these minerals in the following concentration ranges:
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**Pyrite**

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<td>98.11</td>
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</tr>
</tbody>
</table>

**Bournonite**

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<td>Fe</td>
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<td>&lt;0.01</td>
<td>0.10</td>
<td>&lt;0.01</td>
<td>0.03</td>
<td>&lt;0.01</td>
<td>0.17</td>
<td>&lt;0.01</td>
<td>&lt;0.01 - 0.17</td>
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<td>&lt;0.01</td>
<td>&lt;0.01</td>
<td>0.03</td>
<td>0.04</td>
<td>&lt;0.01</td>
<td>0.06</td>
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<tr>
<td>Ag</td>
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<td>0.06</td>
<td>&lt;0.01</td>
<td>0.09</td>
<td>&lt;0.01</td>
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<td>0.01</td>
<td>&lt;0.01</td>
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<tr>
<td>Au</td>
<td>0.28</td>
<td>1.40</td>
<td>&lt;0.01</td>
<td>&lt;0.01</td>
<td>0.07</td>
<td>1.05</td>
<td>&lt;0.01</td>
<td>0.91</td>
<td>&lt;0.01 - 1.40</td>
</tr>
</tbody>
</table>

**TABLE 10.2**: Semi-quantitative microprobe analysis of chalcopyrite, pyrite and bournonite from Cedar Creek (sample 10217).

Compositions in weight percent, for pyrite and chalcopyrite, were calculated using Colby's (1972) MAGIC IV correction programme. Bournonite compositions were obtained directly from K values i.e. $K = \frac{I_{\text{unk}} \times C_{\text{std}}}{I_{\text{std}}}$ and are drift, dead-time and background corrected only.

The maximum estimated error using K+ values only is $\pm 10\%$ of the stated value.

Analyses undertaken on the Otago University JEOL JXA-58 microprobe.
Au Ag

Pyrite: 0.01 - 0.61 wt% 0.01 - 0.14 wt%
Chalcopyrite: 0.01 - 0.78 " 0.01 - 0.07 "
Bournonite: 0.01 - 0.40 " 0.01 - 0.09 "

These values are considerably higher than the maximum values noted by Boyle (1979) for pyrite and chalcopyrite:

<table>
<thead>
<tr>
<th></th>
<th>Au (wt%)</th>
<th>Ag (wt%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pyrite</td>
<td>0.02</td>
<td>0.02</td>
</tr>
<tr>
<td>Chalcopyrite</td>
<td>0.002 (20ppm)</td>
<td>0.13</td>
</tr>
</tbody>
</table>

Comparison of the Cedar Creek data with that of Boyle, indicates clearly that both Au and Ag, as well as occurring as lattice constituents in the sulphides, and sulphosalts, also occur as submicroscopic impurities within these minerals. Such occurrences are common in auriferous sulphide-sulphosalt bearing deposits.

Although not analysed, tetrahedrite in the vein assemblage could be expected to contain less than 0.01% Au and possibly up to 27% Ag, the maximum concentrations noted by Boyle (1979 p.28). Jamesonite and boulangerite could similarly be expected to contain up to 0.08% and 0.02% Ag respectively. These minerals can accommodate only trace amounts (<1ppm) of gold, however.

Semi-quantitative microprobe analysis of gold grains from three quartz samples indicate Ag contents between 0.043 and 4.47 wt% (see Table 10.3). Cu (0.05 - 0.13 wt%), Ni (<0.01 - 0.03 wt%) and Fe (<0.01 - 0.07 wt%) also occur in trace amounts in the gold grains analysed. These results are all within the range reported by Chisholm (1979) for West Australian gold, and are typical of gold compositions reported by other authors from elsewhere in the world.

Compared with the Mine Creek electrum, the Cedar Creek gold contains more Cu similar or slightly less Ni and considerably less Fe. These differences in composition are attributed to different conditions of deposition and mineral associations. The abundant pyrite present in the Mine Creek veins possibly reflects high activities of Fe ionic species,
### Table 10.3: Semi-quantitative microprobe analysis of Cedar Creek Gold.

Analytical conditions as described in Table 10.2 and Appendix III.

Estimated analytical error for all elements is less than ± 10% of the stated value.

N.A. = not analysed.
in the hydrothermal fluids. The incorporation of Fe into gold or electrum would be expected under such conditions, particularly in the apparent presence of low activities of the Cu ionic species. The enrichment of Cu in the Cedar Creek gold reflects its late paragenetic position and deposition with Cu-bearing sulphosalts, suggesting higher activities of Cu ionic species relative to Fe and Ni, in the Cedar Creek ore fluids.

Although no attempt has been made to describe in detail the wall rock alteration accompanying the mineralisation, analysis of the friable fault pug associated with the vein in the William Tell No.2 level (sample 10208) indicates addition of Ca, S, Sb and As to the fault pug accompanied by a loss in Na. K shows no apparent change.

10.6 COMPARISON WITH THE REEFTON QUARTZ LODES

Although the Cedar Creek gold-quartz veins have a similar geological setting to the more celebrated Reef ton quartz lodes to the north, and the Bendigo gold-quartz lodes in Victoria, also in Ordovician sediments of the Tasman Geosyncline) some differences in detail occur. A brief comparison with the quartz lodes of the Reef ton district, as described by Henderson (1917) and Gage (1948), accordingly seems warranted.

Both the Cedar Creek and Reef ton quartz lodes are structurally controlled, occurring as fissure infillings along normal faults in the crests of anticlines, and in the case of the Reef ton district, synclines also. Vein-wall rock contacts in both areas are sharp, with little or no alteration evident. The Cedar Creek reefs apparently consist of parallel veins and stringers, whereas the reefs of the Reef ton district are described by Henderson as lenticular bodies separated by reef tracks. Post formational faulting has disrupted the quartz lodes of both districts, in some cases resulting in the loss of the reef system beyond such faults. Mineralised ore shoots in both areas dip steeply along the strike of the reef, northeast in the case of the...
Reefton lodes, and northwest in the Cedar Creek veins.

These structural similarities clearly suggest a common origin for the gold-quartz lodes, related to regional tectonic events, probably regional metamorphism followed by widespread granite emplacement during the Tuhua Orogeny in Westland foreland. In spite of these structural similarities, differences in the mineralogy of the vein assemblages occur between the areas and are summarised in Table 10.4.

In both districts the gangue assemblage consists of quartz, sulphides and minor calcite, although lead sulphantimonides are present in the Cedar Creek veins, in contrast to the Reefton occurrences. The Reefton quartz lodes generally contain the assemblage:

Pyrite - arsenopyrite - stibnite - chalcopyrite - molybdenite - minor galena - gold.

Sulphides are generally developed along quartz-wall rock contacts. Pyrite and arsenopyrite are the most common sulphides, although locally the veins may contain up to 20% stibnite.

Gold occurs associated with "bluish oily quartz" generally being absent in massive white coarsely crystalline quartz. Low gold values are associated with chalcopyrite - molybdenite, and stibnite mineralisation. Enrichment of gold is reported to occur on wall rock selvages and vein walls.

The fineness of the gold ranges from 90 - 97%.

The Cedar Creek reefs are characterised by the following assemblage:

Pyrite - arsenopyrite - chalcopyrite - sphalerite - galena - bournonite - boulangerite - (jamesonite) - tetrahedrite - gold.

No stibnite has been detected, and the chalcopyrite-molybdenite association is also absent. Pyrite and chalcopyrite are the most abundant sulphides, but as with the Reefton lodes account only for a few percent of the vein material.

Gold is present only sub-economically in the massive white coarsely crystalline quartz making up the bulk of the
REEFTON QUARTZ LODES (Henderson, 1917)  
(Gage, 1948)

Gangue Minerals
Quartz + sulphides + minor calcite
  a) Quartz of 2 varieties
     i) White coarsely crystalline quartz, poorly auriferous quartz.
     ii) Bluish, strained quartz with oily lustre, highly auriferous.

Sulphides
Pyrite - arsenopyrite - stibnite - chalcopyrite - molybdenite - minor galena.
No sulphantimonides or sphalerite.

Gold
Free and in sulphides
Has negative correlation with stibnite and chalcopyrite - molybdenite
Fineness 90 - 97%
Enrichment on vein walls and wall-rock selvages.

CEDAR CREEK REEPS (this study)

Gangue Minerals
Quartz + sulphides + sulphantimonides + calcite.
  a) Quartz of 2 varieties
     i) White, coarsely crystalline poorly auriferous quartz.
     ii) Mineralised quartz with oily or waxy lustre, brecciated.

Sulphides and Sulphantimonides
Pyrite - arsenopyrite - chalcopyrite - galena - sphalerite - boulangerite - bournonite - (Jamesonite) - tetrahedrite.
No stibnite, molybdenite - chalcopyrite association.

Gold
Free and in sulphur-bearing species.
Positive correlation with sulphantimonides, and associated with arsenopyrite.
Fineness probably greater than 95%
Minor enrichment adjacent to country rock.

TABLE 10.4: Comparative mineralogy of the Reefton quartz lodes and the Cedar Creek reefs.  
Reefton data from Henderson, 1917; Gage, 1948.
vein system, mineralisation being confined to ore shoots, which in places contain mineralised brecciated quartz with an oily or waxy lustre. Gold shows a close association with the sulphides and lead sulphantimonides present, occurring both in the lattices of these minerals and in free form. Slight enrichment along quartz-wall rock contacts is evident.

The fineness of the Cedar Creek gold is probably greater than 95%.

The most significant differences noted between the Reef ton lodes and the Cedar Creek reefs are:

a) the absence of the chalcopyrite-molybdenite association, indicative of hypo-mesothermal deposition, in the Cedar Creek reefs;

b) the occurrence of stibnite as the main antimony-bearing phase in the Reef ton field compared with the lead sulphantimonides and tetrahedrite in the Cedar Creek Reefs;

c) the negative correlation between gold and antimony in the Reef ton lodes is not evident in the Cedar Creek reefs, where gold is associated with the late-stage sulphantimonides.

The absence of the chalcopyrite-molybdenite association in the Cedar Creek reefs is attributed to lower meso-epithermal conditions of deposition, (or possibly deposition from molybdenum poor fluids). The presence of the lead sulphantimonides in the Cedar Creek reefs may reflect deposition under conditions of higher activity of PbS relative to Sb$_2$S$_3$ than in the Reef ton lodes, possibly at slightly higher temperatures earlier in the paragenetic sequence. The general absence of galena in the Reef ton lodes further suggests lower activities of PbS in the ore-forming fluids in this area.

Inferred conditions of formation and mineralisation in the Cedar Creek reefs are summarised by way of a genetic model in Chapter 13. Economic considerations relating to vein formation and mineralogy, are discussed in Chapter 14.
CHAPTER XI

SULPHUR ISOTOPE ANALYSIS

11.1 INTRODUCTION

Sulphur isotope studies of a variety of ore deposits have during recent years added considerably to the understanding of ore forming processes, particularly in terms of:

a) Determining the source of sulphur in the deposits, and
b) Determining the temperature of deposition of sulphide minerals.

With the above aims in mind it was decided to undertake a brief sulphur isotope study of the sulphides from the Mount Rangitoto and Mount Greenland vein deposits. The technique was also seen as a potential means of investigating any possible genetic relationship between the two deposits, and their relationships to nearby granite intrusives.

The results obtained have helped to elucidate these problems somewhat, however, the usefulness of the exercise was limited because of the scarcity of Mount Greenland vein material available, bearing cogenetic sulphides suitable for geothermometry.

It had also been hoped to obtain samples from both Schonfelter's workings and Manton's workings at Mount Rangitoto to investigate vertical variations in the sulphur isotopic composition of the sulphides. Unfortunately it was not possible to obtain samples from Schonfelter's workings which lie beyond a fall blocking the access drive and accordingly nothing is known of the vertical variations in the isotopic composition of the vein deposits.

11.2 PRINCIPLES OF THE METHOD

Four isotopes of sulphur occur in nature, these being:

\[
\begin{align*}
\text{32}^\text{S} & \quad 95.10\% \\
\text{33}^\text{S} & \quad 0.74\% \\
\text{34}^\text{S} & \quad 4.20\% \\
\text{36}^\text{S} & \quad 0.02\% 
\end{align*}
\]
with relative abundances as indicated (Stanton, 1972). The two most abundant isotopes $^{32}$S and $^{34}$S are of particular geological interest, as they occur in a wide variety of rock types in sufficient quantity to be easily measured.

The isotopic composition of sulphur depends upon its geochemical history, which in turn reflects the geological processes leading to the formation of sulphur bearing minerals. This property of isotopic fractionation, together with the abundance of sulphur-bearing minerals in many ore deposits, can be particularly useful in helping to solve problems of ore genesis.

11.2.1 Fractionation and Abundance of $^{32}$S and $^{34}$S

Although the ratio $^{32}$S/$^{34}$S in nature is approximately 22.21, natural sulphides from a variety of geological environments show a range from about 20.8 - 23.3 (Stanton, 1972). The isotopic composition of sulphur from a particular sample is expressed in terms of $\delta^{34}$S which is defined as follows:

$$
\delta^{34}S = \frac{^{34}S_{sample}}{^{32}S_{std}} \times \frac{^{34}S}{^{32}S} \times 10^3 \%
$$

The standard used is the sulphur in the troilite (FeS) phase of the Canyon Diablo Meteorite the $^{32}$S/$^{34}$S ratio of which is 22.22 (Faure, 1977). This ratio is similar to that found in mafic igneous rocks, and consequently the $\delta^{34}$S value of a particular sample of terrestrial sulphur is a measure of the change in isotopic composition since introduction into the earth's crust.

The pioneering work of Thode et al. (1949), who reported the first variations in the isotopic composition of sulphur, suggested that fractionation of sulphur was caused by two geologically important processes:

1. Bacterial reduction of sulphate ($SO_4^{2-}$) ions to hydrogen sulphide ($H_2S$) with enrichment of $^{32}$S in the hydrogen sulphide.
2. Isotopic exchange reactions between sulphur-bearing ions, molecules and solids.

Subsequent workers have shown that in the case of isotopic exchange reactions:

1. The heavier isotope ($^{34}S$) is enriched in molecules of higher oxidation state, or greatest bond strength, and isotopic fractionation increases with increased differences in the oxidation states of the molecules concerned.

2. Isotopic fractionation is temperature dependent and decreases with increasing temperature.

3. The amount of isotopic fractionation occurring between two mineral species ($\alpha$ and $\gamma$) is given by the fractionation factor ($\alpha$), as follows:

$$\ln \alpha = \delta^{34}S_\alpha - \delta^{34}S_\gamma$$

The greater the fractionation factor the greater is the temperature dependence of the mineral pairs.

Although the above principles hold in general, isotopic fractionation may be influenced considerably by the physico-chemical environment of mineral deposition, particularly in the case of the formation of hydrothermal sulphides precipitated from aqueous solutions. In such cases the following have been shown to be the most important factors influencing fractionation:

1. The pH of the Ore Forming Fluid:

In an ore forming fluid the abundance of the sulphur bearing species $H_2S$, $HS^-$, $S^-$ and $SO_4^{2-}$, all of which fractionate sulphur isotopes, is controlled by the equilibrium positions of the following reactions:

$$H_2S_{(\alpha)} \rightleftharpoons H^+ + HS^-$$
$$HS^- \rightleftharpoons H^+ + S^-$$
$$2H^+ + SO_4^{2-} \rightleftharpoons H_2S_{(\alpha)} + 2O_2$$

The equilibrium positions of these reactions are controlled by the chemical equilibrium of the hydrogen ion, which is a measure of the pH of the ore forming solution.
Sakai (1968), in ascertaining the above, determined that \( \text{SO}_4^{2-} \) was strongly enriched in \( ^{34}\text{S} \) relative to \( \text{S}^- \), and that the enrichment increased with decreasing temperature. \( \text{H}_2\text{S(aq)} \) and \( \text{H}^- \) were found to concentrate \( ^{34}\text{S} \) less efficiently.

2. Oxygen Fugacity:

Oxygen fugacity has been shown by Ohmoto (1972) to be important in controlling the abundance of \( \text{SO}_4^{2-} \) relative to \( \text{H}_2\text{S(aq)} \). This is most apparent for the sulphur-bearing species \( \text{HSO}_4^- \), \( \text{KSO}_4^- \), \( \text{NaSO}_4^- \) for which the following equilibrium reactions hold:

\[
\begin{align*}
\text{HSO}_4^- & \rightleftharpoons \text{H}^+ + \text{SO}_4^- \\
\text{KSO}_4^- & \rightleftharpoons \text{K}^+ + \text{SO}_4^- \\
\text{NaSO}_4^- & \rightleftharpoons \text{Na}^+ + \text{SO}_4^- 
\end{align*}
\]

A high oxygen fugacity favours development of the \( \text{SO}_4^{2-} \) ion, which in turn enriches the sulphide ion (\( \text{S}^- \)) in \( ^{32}\text{S} \). Accordingly, minerals formed at high oxygen fugacity will be enriched in \( ^{32}\text{S} \).

Both low pH (high activity of the \( \text{H}^+ \) ion) and high oxygen fugacity, therefore, favour the development of sulphide minerals enriched in \( ^{32}\text{S} \), whereas a high pH and a low oxygen fugacity result in sulphide minerals enriched in \( ^{34}\text{S} \).

The effects of both pH differences and oxygen fugacity are greatest at temperatures below about 500\(^\circ\)C (Ohmoto, 1972).

3. Total Sulphur Content:

Ohmoto (1972) studied the system Fe-Ba-O\(_2\)-S\(_2\) and found that the total sulphur content of this system affected the size of the stability fields of the system (see Fig 11.1). The extent of the stability field for the sulphur-bearing minerals in this system will determine the possible ranges of \( \delta^{34}\text{S} \) for the mineral species concerned.

Thus in the system Fe-Ba-O\(_2\)-S\(_2\), under the conditions as given in the explanation accompanying Fig 14.1, pyrite forming from a solution with \( \delta^{34}\text{S} = 0\% \), can have a range of \( \delta^{34}\text{S} = +4 \) to \( -26.7\% \), depending on the fugacity of oxygen and the pH of the system.
Fig 11.1 Isotopic composition of sulphur in pyrite and barite in the system Fe-S-O as a function of pH and the fugacity of oxygen. Square brackets indicate $\delta^{34}S$ of pyrite; round brackets denote $\delta^{34}S$ of barite. Note that the $\delta^{34}S$ of pyrite varies widely depending on the fugacity of oxygen and pH of the system.

The total $^{34}S$ of the system is equal to zero per mil. The Fe-S-O mineral boundaries represent a total sulphur content of 0.1 moles S/kg H$_2$O; T = 250°C, ionic strength = 1.0. The dash-dot line is the barite soluble/insoluble boundary.

A decrease in the sulphur content of the system would decrease the size of the stability fields of the sulphur bearing species, and accordingly reduce the range of isotopic fractionation in these species.

4. Kinetic Isotope Effects:

The effects of isotopic diffusion, preferential mobilisation of one isotope with respect to the other, and differences in reaction rates of the two isotopes have been summarised by Jensen (1967), with regard to hydrothermal deposits. These effects can produce variations in the $\delta^{34}S$ values of sulphide minerals. However, at hydrothermal temperatures the effects are not marked, producing $\delta^{34}S$ differences of less than 10 per mil. Furthermore, only slight variations in $\delta^{34}S$ of sulphide minerals occur as a result of preferential mobilisation of $^{32}S$.

11.2.2 Application of the Theory to Hydrothermal Ore Deposits

The dependence of the sulphur isotope composition of sulphide species on the parameters discussed in the preceding section, makes interpretation of the absolute values of $\delta^{34}S$ for sulphide minerals difficult. This is a problem that must be considered in attempting to determine the source of sulphide sulphur.

Typically, igneous rocks have a narrow range of $\delta^{34}S$ values close to 0 per mil., reflecting a crustal or mantle source and homogenisation of the isotopes during igneous processes. Crystal fractionation may however result in increasing values of $\delta^{34}S$ with increasing fractionation, particularly during the late stages of crystallisation. Extreme enrichment, however, is generally less than $+11$ per mil (Jensen, 1967).

By comparison, sulphides formed by the bacteriogenic reduction of sea water sulphate, tend to show a wide spread in $\delta^{34}S$ values, generally in the range $-20$ per mil to $+45$ per mil (Ault and Kulp, 1959). A similarly large spread of values has been shown to exist in sedimentary rocks (excluding limestones and evaporites) by Ault and Kulp (1959).

Although different rock types show considerable, and often
overlapping ranges in $\delta^{34}S$ values, inferences may be made regarding the source of sulphur in sulphide minerals. In the case of hydrothermal vein deposits, particularly those formed at higher temperatures, the isotopic composition of the sulphides, can generally be used as an indicator of the source of the sulphur present.

Although the absolute values of $\delta^{34}S$ for sulphide minerals are affected by the factors considered previously, these effects do not affect the interpretation of differences in $\delta^{34}S$ values of cogenetic sulphide mineral pairs in terms of equilibrium temperatures.

During crystallisation, or precipitation, small differences in $\delta^{34}S$ occur in cogenetic sulphides. These differences reflect the temperature of formation of the sulphides provided the minerals are in equilibrium. The $\delta^{34}S$ value of the sulphide pair is related to temperature by the following equation:

$$\Delta^{34}S = \frac{A \times 10^6}{T^2}$$

where $\Delta^{34}S = \delta^{34}S_A - \delta^{34}S_B$ (A, B = sulfide minerals)
$T$ = absolute temperature (°K)
$A$ = constant.

Differing values of $A$ have been determined by numerous workers for different mineral pairs, over wide temperature ranges (see Table 11.1). $\Delta^{34}S$ is equivalent to the fractionation factor ($\ln\alpha$) and as mentioned in the preceding section, the greater the fractionation factor, the greater is the temperature dependence of the cogenetic mineral pair.

Consideration of bond strengths predicts that $^{34}S$ should be enriched in the order -

pyrite > sphalerite > chalcopyrite > galena.

The mineral pair pyrite-galena is thus the most sensitive geothermometer, however, as these minerals rarely occur both cogenetically and in equilibrium, the pair sphalerite-galena is most commonly used as a geothermometer.

Furthermore, it can be seen from equation 2 that isotopic fractionation decreases with increasing temperature and
<table>
<thead>
<tr>
<th>MINERAL PAIR</th>
<th>A</th>
<th>TEMPERATURE RANGE, °C</th>
<th>REFERENCE</th>
</tr>
</thead>
<tbody>
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<td>Pyrite-galena</td>
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<tr>
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<td>340-690</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>1.10</td>
<td>250-600</td>
<td>3</td>
</tr>
<tr>
<td>Sphalerite-galena</td>
<td>0.963</td>
<td>27-527</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td>0.62</td>
<td>340-690</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>0.66 x 10^6 T^{-2} - 0.1</td>
<td>250-600</td>
<td>2 by 4</td>
</tr>
<tr>
<td></td>
<td>0.80</td>
<td>250-600</td>
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</tr>
<tr>
<td></td>
<td>0.70</td>
<td>275-600</td>
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</tr>
<tr>
<td></td>
<td>0.78</td>
<td>0-1000</td>
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<td>Pyrite-sphalerite</td>
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<td>0.26</td>
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</tr>
<tr>
<td></td>
<td>0.30</td>
<td>250-600</td>
<td>3</td>
</tr>
<tr>
<td>Chalcopyrite-galena</td>
<td>0.65</td>
<td>250-600</td>
<td>3</td>
</tr>
<tr>
<td>Pyrite-chalcopyrite</td>
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<td>3</td>
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</tr>
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</tr>
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<tr>
<td>Sphalerite-pyrrhotite</td>
<td>0</td>
<td>250-600</td>
<td>3</td>
</tr>
</tbody>
</table>

TABLE 11.1 : Isotopic Fractionation of Sulfur Among Coexisting Sulfide Minerals in Isotopic Equilibrium with an External Sulfur Reservoir*

1. Sakai, 1968
2. Grootenboer and Schwarcz, 1969
4. Czamanske and Rye, 1974
5. Groves et. al. 1970, quoting Hulston

* The temperature dependence of $\Delta^{34}S$ values of mineral pairs is expressed as:

$$\Delta^{34}S = A \times 10^6 / T^2$$

where $\Delta^{34}S$ is the difference between $\delta^{34}S$ values of two coexisting minerals, $A$ is a constant and $T$ is the absolute temperature.

(From Faure (1977) p.414)
because of the inverse square relationship the geothermometer is most sensitive at low temperatures, where higher degrees of fractionation are observed.

11.3 THE METHOD

11.3.1 Sample Collection

Samples of vein material from both Mount Rangitoto and Mount Greenland were collected in the field. Care was taken to select the freshest samples possible, and also where possible those containing several sulphide species. Descriptions of the hand specimens were made, and relevant geological data recorded for each sample.

Polished sections were made from most samples to confirm mineral paragenesis and identifications, and to check for post formational alteration of the sulphides.

11.3.2 Sample Preparation

Pure separates of sulphide minerals were prepared by removing fragments of these minerals from the sample with a metal probe, and then by hand picking under a binocular microscope. Approximately 50 - 100mg of each mineral were prepared, although occasionally it was necessary, due to the scarcity of some sulphides, to use a smaller sample.

From the mineral separates 20 - 30mg each of pyrite, sphalerite and galena, and approximately 50mg of sulphosalt were removed, weighed, mixed with an excess of cuprous oxide (Cu₂O)-approximately 100mg - and ground with mortar and pestle to a uniform mixture. This mixture was then placed in a platinum reaction vessel, ready for analysis.

11.3.3 Sample Analysis

a) Conversion to SO₂

The reaction vessel containing the sample was placed in an evacuated pyrex system (see Fig 11.2) and heated to 1000°C for 5 - 7 minutes by passing an electric current through the reaction vessel. Sulphur dioxide (SO₂) is
Fig 11.2 Vacuum system, as used in this study, for the evolution and collection of SO₂ from sulphides, and measurement of the S-isotopic composition of the SO₂. The system is housed in the Institute of Nuclear Sciences, D.S.I.R., Lower Hutt.
formed by direct reduction, with carbon dioxide (CO₂) and water vapour being the other main gases given off.

b) Removal of CO₂ and Non-condensible gases:

The sample was passed through a cold trap, filled with alcohol and dry ice to remove water vapour and then frozen onto n-pentane chilled with liquid nitrogen. The non-condensible gases were then pumped away, and the n-pentane partially warmed to give off CO₂, which was collected in a second liquid nitrogen cooled trap. Further warming of the n-pentane results in SO₂ being given off.

The SO₂ sample was then passed via the cold finger into a variable reservoir from which it was leaked into the mass spectrometer for measurement.

c) Sample Measurement:

Sample measurement was undertaken on a VG Micromass 1202E mass spectrometer, coupled to a Hewlett-Packard programmable calculator with printout, and chart recorder.

At the beginning of each day's operation a reference sample of SO₂ is leaked into a second variable reservoir and used as a reference for the measurement of each sample (see Fig 11.2). A standard sample of Ag₂S was also run at the start of each day, as a check on the machine's operation, and also to provide a means of comparison with Canon Diablo Troilite δ³⁴S.

Pressures in the reference and sample variable reservoirs were equalised to within 1%, and measurement begun. The reference gas and then the sample gas, were alternately passed into the mass spectrometer via a two way valve and the ratio ³²S/³⁴S of each was measured, and recorded by the chart recorder and calculator printout.

After twelve and again after eighteen measurements of each, the δ³⁴S value and the standard deviation of the sample were printed out by the calculator. If agreement was satisfactory and the standard deviation sufficiently low (less than 0.1 per mil) the measurement was ended and the sample gas pumped out of the mass spectrometer, ready for the next sample.
266.

11.4 RESULTS

Twenty-one analyses of sulphides from Mine Creek, Mount Rangitoto (seventeen) and Cedar Creek, Mount Greenland (four) were made during this study. The $\delta^{34}S$ values obtained are given in Table 11.2 and the average values and ranges for the sulphides are given in Table 11.3.

The $\delta^{34}S$ values obtained from the Mount Rangitoto sulphides range from $+1.60 \pm 0.1\%$ to $-0.44 \pm 0.1\%$. Individual sulphides had the following averages:

- Pyrite: $\delta^{34}S = +0.95 \pm 0.2\%$
- Sphalerite: $\delta^{34}S = +0.52 \pm 0.2\%$
- Galena: $\delta^{34}S = -0.10 \pm 0.2\%$

The values obtained show a slight enrichment (approx 0.6%) in $^{32}S$, relative to Canon Diablo Troilite, and have a narrow range in values (see Table 11.3).

The experimentally predicted trend of $\delta^{34}S$ decreasing in the order pyrite > sphalerite > galena holds for all samples analysed (see Table 11.2).

The four analyses from Cedar Creek produced somewhat more varied results. The three pyrite samples analysed gave an average $\delta^{34}S$ value of $-0.96 \pm 0.2\%$, and a total range of $4.13 \pm 0.2\%$, considerably greater than the range of $1.38 \pm 0.2\%$, obtained for Mine Creek pyrites. The negative average $\delta^{34}S$ values from Cedar Creek indicates that the pyrite, and probably the other sulphides are enriched in $^{34}S$, relative to those from Mine Creek.

No mineral pairs were analysed from Cedar Creek and accordingly no geothermometric calculations for the deposit based on sulphur isotope abundances have been possible.
<table>
<thead>
<tr>
<th>INS 'R' NO.</th>
<th>CONTR. NO.</th>
<th>DESCRIPTION</th>
<th>δ^{34}S %o (CDT)</th>
</tr>
</thead>
<tbody>
<tr>
<td>9233/3</td>
<td>U.C.9502</td>
<td>Pyrite</td>
<td>1.60</td>
</tr>
<tr>
<td>/4</td>
<td>U.C.9500</td>
<td>Pyrite</td>
<td>1.05</td>
</tr>
<tr>
<td>/5</td>
<td>&quot;</td>
<td>Sphalerite</td>
<td>0.50</td>
</tr>
<tr>
<td>/6</td>
<td>&quot;</td>
<td>Galena</td>
<td>-0.16</td>
</tr>
<tr>
<td>/7</td>
<td>U.C.9479</td>
<td>Pyrite</td>
<td>0.22</td>
</tr>
<tr>
<td>/8</td>
<td>U.C.9502</td>
<td>Sphalerite</td>
<td>0.17</td>
</tr>
<tr>
<td>/9</td>
<td>U.C.9503</td>
<td>Pyrite</td>
<td>1.15</td>
</tr>
<tr>
<td>/10</td>
<td>&quot;</td>
<td>Sphalerite</td>
<td>0.79</td>
</tr>
<tr>
<td>/13</td>
<td>U.C.9496</td>
<td>Pyrite</td>
<td>0.78</td>
</tr>
<tr>
<td>/25</td>
<td>&quot;</td>
<td>Sphalerite</td>
<td>0.66</td>
</tr>
<tr>
<td>/15</td>
<td>&quot;</td>
<td>Galena</td>
<td>0.06</td>
</tr>
<tr>
<td>/17</td>
<td>U.C.9495/1</td>
<td>Pyrite</td>
<td>0.86</td>
</tr>
<tr>
<td>/18</td>
<td>U.C.9495/2</td>
<td>Pyrite</td>
<td>1.23</td>
</tr>
<tr>
<td>/19</td>
<td>U.C.9495</td>
<td>Galena</td>
<td>-0.44</td>
</tr>
<tr>
<td>/20</td>
<td>U.C.9498</td>
<td>Pyrite</td>
<td>0.74</td>
</tr>
<tr>
<td>/21</td>
<td>&quot;</td>
<td>Sphalerite</td>
<td>0.47</td>
</tr>
<tr>
<td>/22</td>
<td>&quot;</td>
<td>Galena</td>
<td>0.15</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>CEDAR CREEK, MOUNT GREENLAND</th>
</tr>
</thead>
<tbody>
<tr>
<td>9233/1</td>
</tr>
<tr>
<td>/2</td>
</tr>
<tr>
<td>/11</td>
</tr>
<tr>
<td>/16</td>
</tr>
</tbody>
</table>

Estimated error = ± 0.1%.

TABLE 11.2: Table of δ^{34}S values relative to Canon Diablo Troilite (CDT) from Mine Creek, Mt Rangitoto and Cedar Creek, Mt Greenland.
### Average sulphur isotopic compositions of sulphide minerals from Mine Creek (Mt Rangitoto) and Cedar Creek (Mt Greenland) expressed as $\delta^{34}S$ (‰) relative to Canon Diablo troilite.

<table>
<thead>
<tr>
<th></th>
<th>Mine Creek</th>
<th>Cedar Creek</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pyrite</td>
<td>+0.95</td>
<td>-0.96</td>
</tr>
<tr>
<td>Sphalerite</td>
<td>+0.52</td>
<td>+0.15</td>
</tr>
<tr>
<td>Galena</td>
<td>-0.10</td>
<td>-3.00</td>
</tr>
<tr>
<td>Total</td>
<td>+0.58</td>
<td>-0.68</td>
</tr>
</tbody>
</table>

|                | +0.22      | -3.00       |
| Max value      | +1.60      | +1.13       |
| $\delta^{34}S$ | +1.38      | 4.13        |
| Range          | 0.62       | 0.15        |
| No. of analyses| 8          | 3           |

|                | +0.17      | +0.15       |
| $\delta^{34}S$ | +0.79      | +0.15       |
| Total          | -0.44      | +1.13       |
| Range          | 0.59       | 4.13        |
| No. of analyses| 5          | 1           |

|                | +0.79      | +0.15       |
| $\delta^{34}S$ | +0.15      | +0.15       |
| Total          | +1.54      | +1.13       |
| Range          | 0.62       | 4.13        |
| No. of analyses| 4          | 4           |

Estimated error for average $\delta^{34}S$ and range $\delta^{34}S$ = ± 0.2‰
Estimated error for max. and min. $\delta^{34}S$ values = ± 0.1‰

**TABLE 11.3**

---

Average sulphur isotopic compositions of sulphide minerals from Mine Creek (Mt Rangitoto) and Cedar Creek (Mt Greenland) expressed as $\delta^{34}S$ (‰) relative to Canon Diablo troilite.
11.5 INTERPRETATION OF RESULTS

11.5.1 Source of Sulphur

a) Mine Creek, Mount Rangitoto:

The narrow range of $\delta^{34}$S values obtained from the Mine Creek sulphides, and the proximity of all the values to zero indicates a homogenous sulphur source with only slight enrichment of $\delta^{34}$S relative to Canon Diablo Troilite ($\delta^{34}$S$_{ave} = +0.6 \%$). The values obtained are typical of magmatic hydrothermal deposits (see Fig 11.3) and indicate a magmatic source for the sulphur, with little or no contamination by sulphur derived from the country rocks.

The possible effects of the pH of the ore forming fluids, the oxygen fugacity, total sulphur content, preferential mobilisation of isotopes and differences of reaction rates of the isotopes, on the absolute $^{34}$S values obtained for the sulphides is unknown. Although these effects are probably insignificant with regards to inferring a source of sulphur, as evidenced by the narrow range of values obtained and their proximity to zero, the following points may be noted, particularly with regard to the ore forming fluid:

1. pH of the Ore Forming Fluid:

The absence of sulphate minerals in the veins and wall rock at Mount Rangitoto, suggests that the $SO_4^{2-}$ ion probably was not a major component of the ore forming fluid. As the $SO_4^{2-}$ ion is most abundant in acidic solutions it seems likely that the ore forming solutions were not strongly acid.

The apparently low concentrations of $SO_4^{2-}$, which concentrates $^{34}$S, in the fluids, suggests that the sulphides forming from the hydrothermal fluids are only slightly, if at all, enriched in $^{32}$S relative to the fluid.

2. Oxygen Fugacity:

A high oxygen fugacity also favours the development of the $SO_4^{2-}$ ion in solution. The absence of sulphate minerals accompanying the mineralisation, suggests that
Fig 11.3 Sulphur isotopic ratios of ore deposits of magmatic hydrothermal, sandstone uranium and red-bed copper types. The length of the short vertical lines indicates the number of samples with a particular ratio, the shortest lines representing a single sample.

Both the Mine Creek and Cedar Creek sulphides have \( ^{34}S \) values typical of magmatic hydrothermal ore deposits. (from Jensen, 1967)
a low oxygen fugacity was likely, and that the sulphides in the veins are not enriched significantly in $^{32}\text{S}$, with respect to the ore forming fluid.

3. Total Sulphur Content:
The sulphur content of the solutions from which the sulphide minerals formed is unknown, and accordingly its effect on the $\delta^{34}\text{S}$ values obtained is unknown also.

4. Preferential Mobilisation and Reaction Rate Differences:
Jensen (1967) states that at hydrothermal temperatures the above effects are likely to produce only slight variations in the $\delta^{34}\text{S}$ values of sulphide minerals. There is no evidence that such effects have significantly affected the isotopic composition of the Mount Rangitoto sulphides.

From the above considerations it seems likely that the $\delta^{34}\text{S}$ values obtained for the sulphide minerals of Mine Creek are very similar to those of the fluid from which they precipitated, and reflect a magmatic origin, for the sulphur and the fluid.

It should be noted, however, that some of the sulphur present could have been derived initially by the assimilation of the Greenland Group sediments by the intruding biotite granite, rather than being of primary magmatic origin. Sulphur thus derived would be impossible to distinguish isotopically from primary magmatic sulphur.

The geological evidence, notably the proximity of the vein system to a granite body and the reported strengthening of the vein system with depth (i.e. towards the granite) provides further support for a magmatic origin for the sulphur in the sulphides in the Mine Creek vein system.

b) Cedar Creek, Mount Greenland:
The pyrites, and sulphosalt, analysed from Cedar Creek give $\delta^{34}\text{S}$ values compatible with a magmatic hydrothermal origin (see Fig 11.3). The $\delta^{34}\text{S}$ values give an average of $-0.7\%$ and range from $-3.0\%$ to $+1.1\%$. These values indicate a slight enrichment of $^{34}\text{S}$ in the sulphides from Cedar Creek, relative to those from Mine Creek. The slight negative average $\delta^{34}\text{S}$ value for the Cedar Creek sulphides suggests a dominantly magmatic origin for the sulphur.
Ohmoto (1972) showed that the absolute value of $\delta^{34}S$ in sulphide minerals was largely dependent on the pH, the oxygen fugacity and the total sulphur content of the hydrothermal fluids. The enrichment of $^{34}S$ in the Cedar Creek sulphides, compared with those from Mine Creek, may reflect deposition from a fluid with a lower pH and oxygen fugacity (reducing the amount of $^{34}S$ enriched $SO_4^{2-}$ ions in solution), and possibly a lower sulphur content also. This possible difference in the composition of the hydrothermal fluid would be expected with increasing distance from the magmatic source, and may even reflect an increasing component of meteoric water in the fluids from which the Mount Greenland sulphides were deposited.

The greater variation in $\delta^{34}S$ values obtained from the Cedar Creek sulphides reflects either a less homogeneous source for the contained sulphur, and/or changes in the composition of the hydrothermal fluids from which the sulphides were deposited. Some of the sulphur present in the sulphides may have been derived from the host rocks, although the proximity to zero of the $\delta^{34}S$ values suggests that only a small amount of sulphur, if any, was thus derived.

Accordingly, most of the sulphur present in the Cedar Creek sulphides is considered to be magmatically derived, and therefore it seems probable that a large portion of the hydrothermal fluids were magmatic in origin. A significant component of meteoric and/or metamorphically derived fluids may, however, have also entered the vein system.

11.6 INTERPRETATION OF RESULTS - GEOTHERMOMETRY

11.6.1 Determination of Temperatures of Formation

Three triplets and three sulphide pairs were analysed from the vein system at Mount Rangitoto, with the aim of determining temperatures of mineral formation.

A number of workers including Sakai (1968), Grootenboer and Schwarz (1969), Kajiwara and Krouse (1971), Czamanske and Rye (1974) and Smith et. al. (1979) have experimentally
calibrated sulphur isotope geothermometers for the common sulphides (pyrite, sphalerite, pyrrhotite, chalcopyrite, galena), over temperatures ranging from 0°C to 1000°C. These works enable a comparison of natural sulphide assemblages with the experimentally calibrated geothermometers, to be made.

The $\delta^{34}S$ values for the Mine Creek sulphides and the $\Delta^{34}S$ values for sulphide pairs, are given in Table 11.4. As can be seen from the table, isotopic fractionation between coexisting sulphides is only slight, ranging from $0.59 \pm 0.2$ to $1.67 \pm 0.2\%$ for the pair pyrite-galena, $0.32 \pm 0.2$ to $0.66 \pm 0.2\%$ for the pair sphalerite-galena, and $0.11 \pm 0.2$ to $1.43 \pm 0.2\%$ for the pair pyrite-sphalerite.

The values of the triplets - pyrite, sphalerite, galena - can be plotted on the experimental trisulphide geothermometer of Smith et. al. (1979) in the form of a triangular diagram (see Fig 11.4). Unfortunately the three triplets all plot above 600°C, the upper limit for which the geothermometer was calibrated. Extrapolation of the geothermometer above 600°C suggests that the sulphides from one sample U.C.9500 formed at about 670°C, whereas those of samples U.C.9496 and U.C.9498 formed at about 900°C. These temperatures are geologically unreasonable for base metal-bearing hydrothermal veins.

The trisulphide geothermometer, as well as determining temperatures of formation, can also be used as a check on the attainment of equilibrium between the sulphide minerals. If the sulphides are in equilibrium, plots of the $\Delta^{34}S$ values of the pairs combinations should plot on or close to the thermometer line, and as individual points or small areas. The three triplets analysed all plotted as single points rather than areas, and although not on the geothermometer line, were sufficiently close to suggest that equilibrium is probably being closely approached in the sample sulphides.

It should also be noted that calibration of the geothermometer was achieved using sulphides synthesised under non-hydrothermal conditions. This may result in some error in temperature estimates when the geothermometer is applied to natural hydrothermal conditions.
<table>
<thead>
<tr>
<th>SAMPLE</th>
<th>$\delta^{34}\text{S} (%)$ (CDT)</th>
<th>$\Delta^{34}\text{S} (%)$ Py/Sp</th>
<th>$\Delta^{34}\text{S} (%)$ Sp/Gn</th>
<th>$\Delta^{34}\text{S} (%)$ Py/Gn</th>
</tr>
</thead>
<tbody>
<tr>
<td>U.C.9502</td>
<td>Pyrite 1.60 ± 0.1</td>
<td>1.43 ± 0.2</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td></td>
<td>Sphalerite 0.17 ± 0.1</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>U.C.9503</td>
<td>Pyrite 1.15 ± 0.1</td>
<td>0.36 ± 0.2</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td></td>
<td>Sphalerite 0.79 ± 0.1</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>U.C.9495</td>
<td>Pyrite 0.86 ± 0.1</td>
<td>-</td>
<td>7/1. 1.30 ± 0.2</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Pyrite 1.23 ± 0.1</td>
<td>-</td>
<td>7/2. 1.67 ± 0.2</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Galena -0.44 ± 0.1</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>U.C.9496</td>
<td>Pyrite 0.77 ± 0.1</td>
<td>0.11 ± 0.2</td>
<td>0.60 ± 0.2</td>
<td>0.71 ± 0.2</td>
</tr>
<tr>
<td></td>
<td>Sphalerite 0.66 ± 0.1</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td></td>
<td>Galena 0.06 ± 0.1</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>U.C.9498</td>
<td>Pyrite 0.74 ± 0.1</td>
<td>0.27 ± 0.2</td>
<td>0.32 ± 0.2</td>
<td>0.59 ± 0.2</td>
</tr>
<tr>
<td></td>
<td>Sphalerite 0.47 ± 0.1</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td></td>
<td>Galena 0.15 ± 0.1</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>U.C.9400</td>
<td>Pyrite 1.05 ± 0.1</td>
<td>0.55 ± 0.2</td>
<td>0.66 ± 0.2</td>
<td>1.21 ± 0.2</td>
</tr>
<tr>
<td></td>
<td>Sphalerite 0.50 ± 0.1</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td></td>
<td>Galena -0.16 ± 0.1</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>U.C.9479</td>
<td>Pyrite 0.22 ± 0.1</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

**TABLE 11.4**: $\Delta^{34}\text{S}$ values of sulphide pairs from Mine Creek, Mt. Rangitoto.
The Mine Creek samples plot beyond 600°C, the upper temperature for which the geothermometer was calibrated. Depositional temperatures of 600° - 900°C are indicated. These temperatures are, however, unrealistically high.

All three samples plot as single points near the geothermometer line AA' indicating the near attainment of equilibrium in all samples.

See text for further discussion.
Smith et al. (1979) also determined curves of $\Delta^{34}\text{S}$ against $1/T$ ($T$ = temperature) for the sulphide pairs pyrite-sphalerite, sphalerite-galena, and pyrite-galena. Unfortunately, again the curves were not calibrated above 600°C. However, measurement of the slope of their calibration curves enables calculation of the constant $A$, which may then be used to calculate temperatures of formation beyond the limits of calibration (600°C). The values thus obtained are given in Table 11.5, together with temperature estimates similarly derived from the data of Czamanske and Rye (1974), Kajiwara and Krouse (1971) and Sakai (1968).

Several significant features are evident from Table 11.5.

1. The temperatures of formation for coexisting sulphide pairs show a huge range, from $185 \pm 13$°C to $1526 \pm 1388$°C. Many of the higher temperatures are obviously in error, being beyond the stability field of pyrite (see section 9.3).

2. The temperature agreement between pairs of minerals from samples containing three sulphides is poor, with the following temperatures being calculated (from Smith et al. (1979)), for respective pairs:

<table>
<thead>
<tr>
<th>Sample</th>
<th>Py-Sp</th>
<th>Sp-Gn</th>
<th>Py-Gn</th>
</tr>
</thead>
<tbody>
<tr>
<td>U.C.9496</td>
<td>$1378 \pm 1252$</td>
<td>$882 \pm 147$</td>
<td>$983 \pm 138$</td>
</tr>
<tr>
<td>U.C.9498</td>
<td>$781 \pm 289$</td>
<td>$1308 \pm 409$</td>
<td>$1104 \pm 187$</td>
</tr>
<tr>
<td>U.C.9500</td>
<td>$465 \pm 85$</td>
<td>$828 \pm 125$</td>
<td>$689 \pm 57$</td>
</tr>
</tbody>
</table>

Large variations both within and between samples are evident, with temperature differences of up to 530°C being recorded for different sulphide pairs from the same sample, and differences of up to 920°C occurring between samples.

Such varied data, from samples collected within close proximity from the same vein system, emphasises clearly that the sulphur isotope geothermometer is not a valid means of estimating temperatures of formation for the Mine Creek sulphides.

3. The large errors accompanying the calculated temperatures of formation for the sulphide pairs from Mine Creek reflects both the lack of sensitivity of the sulphur isotope geothermometer at high temperatures, and the low degree of
<table>
<thead>
<tr>
<th>Sample</th>
<th>$\Delta^{34}S_{Py-Sp}$</th>
<th>$\Delta^{34}S_{Sp-Gn}$</th>
<th>$\Delta^{34}S_{Py-Gn}$</th>
<th>A</th>
<th>Source</th>
<th>Temp FM ($^\circ$C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>U.C.9502</td>
<td>1.43 ± 0.2</td>
<td>0.30</td>
<td>2, 4</td>
<td>185 ± 13</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.356</td>
<td>3</td>
<td>226 ± 16</td>
<td></td>
<td></td>
</tr>
<tr>
<td>U.C.9503</td>
<td>0.36 ± 0.2</td>
<td>0.30</td>
<td>2, 4</td>
<td>640 ± 178</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.356</td>
<td>3</td>
<td>721 ± 200</td>
<td></td>
<td></td>
</tr>
<tr>
<td>U.C.9495/1</td>
<td>1.30 ± 0.2</td>
<td>1.10</td>
<td>2</td>
<td>647 ± 49</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>1.319</td>
<td>3</td>
<td>734 ± 57</td>
<td></td>
<td></td>
</tr>
<tr>
<td>U.C.9495/2</td>
<td>1.67 ± 0.2</td>
<td>1.10</td>
<td>2</td>
<td>538 ± 32</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>1.12</td>
<td>4</td>
<td>546 ± 33</td>
<td></td>
<td></td>
</tr>
<tr>
<td>U.C.9496</td>
<td>0.11 ± 0.2</td>
<td>0.30</td>
<td>2, 4</td>
<td>1378 ± 1252</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.356</td>
<td>3</td>
<td>1526 ± 1388</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.70</td>
<td>1</td>
<td>807 ± 135</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.80</td>
<td>2, 4</td>
<td>882 ± 147</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.71 ± 0.2</td>
<td>1.10</td>
<td>972 ± 137</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>4</td>
<td>983 ± 138</td>
<td></td>
<td></td>
</tr>
<tr>
<td>U.C.9498</td>
<td>0.27 ± 0.2</td>
<td>0.30</td>
<td>2, 4</td>
<td>781 ± 289</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td>0.356</td>
<td>3</td>
<td>875 ± 324</td>
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<tr>
<td></td>
<td></td>
<td>0.70</td>
<td>1</td>
<td>1206 ± 377</td>
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<tr>
<td></td>
<td></td>
<td>0.80</td>
<td>2, 4</td>
<td>1308 ± 409</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td>0.59 ± 0.2</td>
<td>1.10</td>
<td>1092 ± 185</td>
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<td></td>
<td></td>
<td>4</td>
<td>1104 ± 187</td>
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<tr>
<td>U.C.9500</td>
<td>0.55 ± 0.2</td>
<td>0.30</td>
<td>2, 4</td>
<td>465 ± 85</td>
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<td></td>
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<tr>
<td></td>
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<td>0.356</td>
<td>3</td>
<td>531 ± 87</td>
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<td></td>
<td></td>
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<td>1</td>
<td>757 ± 115</td>
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<td>0.80</td>
<td>2, 4</td>
<td>828 ± 125</td>
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<td>1.21 ± 0.2</td>
<td>1.10</td>
<td>680 ± 56</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td>4</td>
<td>689 ± 57</td>
<td></td>
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</tbody>
</table>

\[ T (^\circC) = \sqrt{\frac{A \times 10^6}{\Delta^{34}S}} - 273 \]

2. Kajiwara and Krouse (1971)
3. Sakai (1968)
4. Smith et. al. (1977) (values of A from slope of graph (Fig 1) in paper)

TABLE 11.5: Calculated Temperatures of Formation of Mine Creek Sulphides.
fractionation between the sulphide pairs compared with the accuracy of the analytical technique. This latter effect results in large percentage errors reflected in the errors quoted in Table 11.5.

11.6.2 Sources of Error

Clearly, the temperatures of formation calculated from the $\Delta^{34}S$ values of the Mine Creek sulphides (see Table 11.5), do not reflect the temperatures of formation. Many of the calculated temperatures exceed the stability fields of the sulphide minerals in the vein system and accordingly, the temperature estimates based on sulphur isotope data must be regarded as being in error.

Although the source(s) of the error are difficult to determine, the following factors seem most likely:

a) Non-cogenicity of Sulphide pairs:

For the sulphur isotope geothermometer to be valid the $\Delta^{34}S$ values calculated must be from cogenetic mineral pairs. Unfortunately, it is not always possible to determine the paragenetic relationship of mineral pairs with certainty, and commonly it is the lack of textures indicating non-cogenicity, such as replacement of one mineral by another, that must be taken as evidence of possible cogenicity of mineral pairs. Obviously, such mineral pairs may, in fact, be non-cogenetic, and if so the $\Delta^{34}S$ values obtained from such pairs, would only fortuitously reflect the temperature of formation.

Textures indicative of cogenetic deposition of sulphides are lacking from the vein samples studied from Mine Creek, and although mineral pairs showing obviously non-cogenetic relationships were rejected for analysis, those selected may also have been non-cogenetic. If this were indeed the case, the $\Delta^{34}S$ values obtained would not be a measure of the degree of isotopic fractionation between cogenetic mineral pairs, which as noted previously is temperature dependent, and accordingly the geothermometer would be invalidated.

b) Non-equilibrium of Sulphide pairs:

Cogenetic sulphide minerals may exist in disequilibrium due to a variety of causes, including post-depositional
re-equilibration due to preferential migration of isotopes, and/or rapid crystallisation during formation of the sulphides preventing the attainment of equilibrium.

Smith et al. (1977) in developing their experimental tri-sulphide geothermometer provided a means for investigating possible disequilibrium between the sulphide triplet pyrite-sphalerite-galena. If these phases are in equilibrium the $\Delta^{34}S$ values for the mineral pairs should plot as a point or small area on the triangular sulphide geothermometer of Smith et al., and also lie close to the calibrated thermometer line AA'.

The three pyrite-sphalerite-galena triplets analysed from Mine Creek when plotted on the geothermometer of Smith et al., plot as single points, although none of the triplets plot particularly close to the line AA'. This suggests that although equilibrium is approached in the co-existing sulphides, it is not attained, and accordingly the temperatures of formation calculated from the sulphide pairs, assuming the existence of equilibrium, will be in error. Furthermore, as the fractionation factor $\Delta^{34}S$ is inversely proportional to the square of the temperature of formation, the effects of disequilibrium will be most marked at high temperatures of formation. Such a disequilibrium may be responsible, in part at least, for the unacceptable temperature estimates obtained from the Mine Creek sulphides.

The possible reason(s) for the existence of isotopic disequilibrium between the Mine Creek sulphides is unknown.

c) Sample Impurity:

Another possible source of error is sample impurity, although every care was taken to avoid this during preparation and analysis. The narrow range of $\delta^{34}S$ values obtained suggests that the effect of impurities is small or negligible.

The lack of attainment or preservation of isotopic equilibrium, and the possible non-cogenicity of the sulphide pairs are probably the most important factors in invalidating the application of the sulphur isotope geothermometer to the Mine Creek sulphides. These conditions do not, however,
affect the absolute $\delta^{34}S$ values obtained, and therefore do not affect inferences regarding the source of the sulphur in the sulphide minerals of the vein assemblages.
CHAPTER XII

FLUID INCLUSION DATA

12.1 INTRODUCTION

Twelve samples, seven from Cedar Creek and five from Mine Creek, were collected and prepared for fluid inclusion examination. Unfortunately, from the twelve only two samples yielded inclusions of workable size, and one of these was destroyed during the final stages of preparation.

The lack of suitable inclusions for study severely limited the usefulness of the technique. Quantitative data on temperatures of quartz deposition based on fluid inclusion studies, are non-existent for Cedar Creek quartz and only one formation temperature was obtained from mineralised Mine Creek quartz. Some information regarding the nature of the ore-forming solution also emerged as a consequence of the study.

12.2 PRINCIPLES OF THE METHOD

During the growth of hydrothermal minerals, in this case quartz, portions of the mineralising fluid may become trapped in crystal defects. Sorby (1858) recognised that these minute samples of hydrothermal fluid could be used as a geothermometer, using as a basis, the thermal expansion and contraction of the entrapped fluid.

On cooling, the entrapped fluid contracts so that usually, at room temperature, a vapour phase exists within the inclusion. On re-heating of the inclusion this phase will revert to liquid at the homogenisation temperature, which reflects the temperature of entrapment of the fluid.

The use of fluid inclusions as a geothermometer demands that the entrapped fluids be small samples of the mineralising fluid which have not undergone any modification since entrapment. The major causes of post formational modification are:
1. Leakage of the inclusion fluid.
2. Reaction of the fluid with the host mineral and deposition or absorption of the fluid components.
3. Changes in volume of the inclusion.

It has been shown by numerous workers (see, for example Roedder, 1967) that these effects, are, in the case of quartz, generally negligible.

As well as satisfying the above criteria, it has also been found that the composition of the inclusion fluid and the pressure of entrapment will influence the temperature of homogenisation.

1. Composition of the Fluid.

The composition of the fluid may affect the homogenisation temperature of the inclusion, particularly as the presence of solutes will influence both internal vapour pressures and fluid compressibility within the inclusion.

2. Effect of Pressure.

The pressure of entrapment will affect the degree of filling of the fluid inclusion (see Roedder, 1967) as well as the density of the liquid in the inclusion. Both will affect the homogenisation temperature, which may need to be corrected for pressure to determine the true filling temperature of the inclusion.

A more detailed description of the principles of the method is given by Roedder (1967, 1972), and is summarised by Newman (1978).

12.3 METHOD

Samples of quartz were cut with a diamond saw. The cut face was ground, using successively finer grades of carborundum, until smooth and flat. The sample was then polished using an alumina-water mix on a Struers lapping machine. The polished face was mounted on a glass microscope slide, the sample ground down to approximately 0.5mm thickness, and the upper face of the sample polished as above.
A preliminary inspection of the slide was made under an ordinary petrographic microscope to locate potentially workable inclusions, and areas surrounding these marked with a diamond pencil. The sample was then removed from the slide by dissolving the Lakeside adhesive in warm alcohol. The polished chip thus obtained was broken, and the fragment containing the desired inclusion retained for study.

Freezing and heating runs were carried out using a Zeiss Jena microscope fitted with a Chaix Meca heating-freezing stage. Cooling runs were made using a 12.5x occular lens and a 25x objective. For heating runs a 25x occular lens with a 16x (16/0.20 Zeiss) long focal objective was used.

Liquid nitrogen was used as a coolant during freezing runs.

12.4 RESULTS

The single inclusion studied was from sample U.C.9498 being located in a quartz crystal lining a small cavity. No textural relationships between this quartz and the sulphide minerals were discernable in thin section.

Two phases were observed in the inclusion at room temperature (see Fig 12.1a) these being a water rich liquid and a vapour rich phase. The sample was cooled rapidly to -110°C, to begin the "freezing run", slow warming from a low temperature rather than gradual cooling being employed to overcome metastability of some phases.

During this initial cooling an intermediate phase was observed to appear on the margin of the vapour bubble, and on further cooling disappear. The formation of a solid phase on the margin of the vapour phase was observed at c. -98°C (see Fig 12.1b).

Gradual warming was begun at -110°C, at a rate of about 1°C per minute. The solid phase in Fig 12.1b remained stable from -110°C to -57°C when it disappeared suddenly into the
Two phases only present, a water-rich liquid (w) and a CO₂-rich vapour (v).

Rapid cooling begun
During cooling a solid phase (Sc) formed suddenly at 98°C, and remained stable to 110°C.

Rapid cooling to -110°C then gradual warming
On warming to -57°C the solid CO₂-rich phase (Sc) disappeared suddenly into the vapour phase (v), leaving a CO₂-rich vapour (v) and H₂O-rich solid or liquid phase (W or I).

Further warming to -49°C resulted in the appearance of a CO₂-rich liquid phase (Lc) on the margin of the CO₂-rich vapour bubble (V).

Liquid CO₂ phase (Lc) remained stable to -17°C before disappearing into the CO₂-rich vapour phase (V). An unrecognised clathrate (CO₂-hydrate) probably formed at this time (see text).

A slight shape change occurred at -3.0°C due to the melting of the H₂O-rich solid (I) to form H₂O-rich liquid (W).

CO₂-rich liquid (Lc) began to appear at -0.9°C.

CO₂-rich liquid (Lc) increased in volume until +13°C was reached.

CO₂-rich liquid (Lc) became unstable at +14°C disappearing suddenly to leaving a CO₂-rich vapour (V) and H₂O-rich liquid (W) room temperature.

Fig 12.1 Phase changes observed fluid inclusion (sample U.C.9498) during a "freezing" run. Gradual warming rather than cooling was used to overcome possible metastability of some phases during cooling. V = CO₂-rich vapour; W = H₂O-rich liquid; Sc = solid CO₂; Lc = liquid CO₂; I = ice.

See text for discussion.
vapour phase, (Fig 12.1c) strongly suggesting that the solid phase was CO$_2$ (melting point 56.6°C), and that the vapour phase (V) was CO$_2$ rich.

Further warming of the inclusion to -49°C resulted in a new phase appearing marginal to the vapour phase (Fig 12.1d). This new phase remained present until a temperature of -17°C was reached, when the vapour phase (V) expanded to occupy the space filled by the intermediate phase. This intermediate phase is interpreted as being liquid CO$_2$. As the liquid CO$_2$ coexists with CO$_2$ vapour below 17°C, it is apparent that warming from -49°C to -17°C followed the CO$_2$ phase boundary in Fig 12.2. If indeed the warming path was along this boundary an unrecognised clathrate (CO$_2$ hydrate), must also have existed in the inclusion.

Further warming to -3°C produced a slight change in shape of the CO$_2$ vapour bubble, interpreted as melting of the aqueous phase, which must have existed as ice between -3°C and -17°C. Below -17°C the ice may have reverted to liquid H$_2$O, as a consequence of the formation of liquid CO$_2$, and probably remained liquid over the stability range of the liquid CO$_2$ (-17°C to -49°C). Indeed, the existence of solid CO$_2$ apparently extending into the aqueous phase of the inclusion (W) (see Fig 12.1b), suggests that water may have metastably existed in the inclusion to at least -57°C not becoming frozen until further cooled.

Double freezing of the aqueous phase, has been reported in CO$_2$ rich inclusions by Collins (1979), and it seems probable that this also occurred in the inclusion studied.

Melting of the aqueous solution, evidenced by a slight shape change of the aqueous phase, probably occurred at -3.0°C. Further warming to -1.0°C saw the appearance of a liquid phase (liquid CO$_2$) (see Fig 12.1f) at the margin of the vapour phase. From the experimental work of Chen (1972) and Bozzo et. al. (1975), this suggests that the salinity of the aqueous phase in the inclusion is approx. 18% (see Fig 12.2).

The liquid CO$_2$ phase remained stable until 14°C (Fig 12.1h) then disappeared, leaving at room temperature; an aqueous
Fig 12.2 Phase relations in the system CO$_2$-H$_2$O (from Hollister et al., 1976). The dotted line indicates the path followed by the inclusion studied during the freezing run. The freezing point depressions (~3.0°C) indicates a salinity of c.18% NaCl equivalent.

CO$_2^V$ = CO$_2$-rich vapour; CO$_2^L$ = CO$_2$-rich liquid; C = CO$_2$ clathrate; H$_2$O$^L$ = H$_2$O-rich liquid; I = H$_2$O solid, ice.
phase (W) and a CO$_2$ rich vapour phase.

During the heating run, the temperature was raised at approximately 1°C per minute from room temperature to 400°C and changes noted. An estimate of the proportion of CO$_2$ rich vapour present was made at 31°C, the critical point of CO$_2$, the estimate being 30%. Further heating resulted in homogenisation of the inclusion at 290°C ± 5°C and decrepitation of the inclusion at 305°C ± 5°C.

Several small, secondary inclusions, were observed to homogenise at 330°C ± 15°C, and decrepitate at about 355°C ± 15°C.

It was also observed that the degree of filling of the secondary inclusions was quite variable, ranging from almost entirely vapour to almost entirely liquid.

12.5 INTERPRETATION

The freezing data suggest that the ore forming fluids;
a) were CO$_2$ rich;
b) had a salinity of about 18% (NaCl equivalent).

The likely pressure of sealing of the inclusion cannot be determined from study of the inclusion.

The homogenisation temperature of the primary inclusion studied suggests sealing of the inclusion at a temperature of about 290°C ± 5°C, and a maximum temperature of formation, the decrepitation temperature is likely to be about 305°C ± 5°C.

This implies that, if the quartz containing the inclusion were co-genetic with sulphide mineralisation a formation temperature of about 300°C ± 10°C is likely.

The location of the quartz in the lining of a drusy cavity, however, suggests that the quartz may have been deposited during the latest stages of vein formation, at considerably lower temperature than those of formation of the sulphide minerals.

Insufficient data from fluid inclusions exist to determine, if this is in fact the case, or if the homogenisation
temperature obtained is accurate, and likely to be representative of the vein quartz as a whole.

Allowing for the effects of pressure on the homogenisation temperature, a maximum temperature of formation of $450^\circ C \pm 20^\circ C$ is indicated, at geologically reasonable pressures. This indicated temperature of formation is in quite good agreement with temperatures predicted by study of both the vein assemblage, (section 9.3) and the country rock assemblage (section 5.7).
CHAPTER XIII

INTERPRETATION SUMMARY AND CONCLUSIONS
- GENETIC MODEL FOR THE MINERALISATION

13.1 GEOLOGIC SETTING PRIOR TO MINERALISATION

The Greenland Group, which hosts the known mineralisation in the thesis area, consists of a thick alternating sequence of greywackes and argillites, interpreted as being a turbidite succession. Deposition of the sediments occurred during Cambrian - Ordovician times in the Tasman Geosyncline, formerly an extensive depositional basin now fragmented by Cretaceous rifting, forming the Tasman Sea, and by post-Cretaceous movement on the Alpine Fault.

The Greenland Group sediments, in spite of their age and proximity to the Alpine Fault, have undergone relatively little deformation. Prior to lithification, the wet unconsolidated sediments underwent lower greenschist facies regional metamorphism during which temperatures of about 400°C (Shelley, 1975a) were attained. This metamorphism was accompanied by dewatering of the unconsolidated sediments. In spite of the relatively high metamorphic grade attained the Greenland Group shows only slight evidence of recrystallisation, predominantly detrital quartz occurring in a recrystallised mica and clay mineral matrix.

Deformation of the Greenland Group accompanying the regional metamorphism resulted in the formation of a series of horizontal or gently plunging upright folds (after Fleuty, 1964) with moderately to steeply dipping limbs. The strike of both bedding and the axial surfaces of the folds is remarkably uniform, being predominantly NW-SE, although rocks of the now redundant Waipata Group strike NNE-SSW. The reasons for this regional variation in strike are unknown. The development of a penetrative axial plane cleavage also accompanied the folding episode.
Intrusion of Tuhua Group granitoids into the thick, folded sequence of Greenland Group greywackes and argillites probably began during the late Ordovician (Adams, 1973) and continued sporadically throughout much of the Paleozoic. A pulse of intrusive activity, known as the Tuhua Orogeny occurred in Devonian - Carboniferous (300 - 370 Ma) times, during which emplacement of the Rangitoto Granite, the oldest recognised intrusive in the thesis area, occurred.

13.2 THE INTRUSIVE PHASES

Two phases of granite intrusion are recognised in the thesis area.

1) Intrusion of the Rangitoto Granite

Intrusion of the Rangitoto Granite occurred between c.295 and 314 Ma ago (Carboniferous - Permian) during the late stages of the Tuhua Orogeny. Petrological and geochemical data suggest that the granite is probably an S-type granite (after Chappell and White, 1974) derived by partial melting of pre-existing sedimentary rocks, probably in a region of high heat flow above a subduction zone.

The conditions of emplacement of the granite are not particularly well established. When plotted on the Q-Ab-Or-H₂O diagram of Tuttle and Bowen (1958), crystallisation pressures of P_H₂O = 3.5 - 4 kbars are indicated. These pressures, assuming a pressure gradient of 4km/kbar, and P_H₂O < P_total in the crystallising magma, indicate a minimum depth of emplacement of 14 - 16km. This depth is almost certainly too great, however. The generally widespread occurrence of andalusite, the low pressure Al₂SiO₅ polymorph, in the contact aureole, and the generally low metamorphic grade (hornblende-hornfels facies) of the aureole, suggests that emplacement of the granite probably occurred at shallower depths of c.4 - 8km. Accordingly 8km is regarded as a realistic maximum depth of emplacement.

Emplacement temperatures of 700⁰ - 750⁰C were probable, these being close to the eutectic minimum of Tuttle and Bowen (1958).
Contact metamorphism of the Greenland Group roof pendant hosting the Mine Creek vein mineralisation has resulted in widespread hornfelsing and alkali metasomatism, resulting in addition of $\text{Al}_2\text{O}_3$, $\text{K}_2\text{O}$, $\text{CaO}$, $\text{Fe}_2\text{O}_3$, $\text{MgO}$ and $\text{MnO}$ to the hornfelses. Andalusite and cordierite porphyroblasts, now altered to quartz + sericite, are widespread in the pendant along with metamorphic biotite and muscovite. Rarely sillimanite needles, may also be present.

Locally tourmalinisation and sulphur-metasomatism have also affected the hornfelses, the latter resulting in the formation of disseminated pyrrhotite within the hornfelses.

2) Intrusion of the Kakapotahi Granite

The strongly differentiated Kakapotahi Granite was probably emplaced during the Triassic, c.214 Ma ago (a Rb-Sr isochron age would be required to confirm this age), and like the older Rangitoto Granite, is regarded as being an S-type granite (after Chappell and White, 1974). When plotted on the $\text{Q-Ab-Or-H}_2\text{O}$ diagram minimum pressures of crystallisation of 2 kbars are indicated, suggesting emplacement of the Kakapotahi Granite at depths greater than 8km, assuming a pressure gradient of 4km/kbar. This estimate, seems intuitively too great however, particularly in view of the widespread retrogressive metamorphic effects noted in the contact aureole, thought to be associated with emplacement of the Kakapotahi Granite. Accordingly, a maximum depth of emplacement of c.8km is considered probable. Both the Kakapotahi and Rangitoto granites could, however, have been emplaced at higher levels than this, possibly to depths as shallow as 2km.

Temperatures of emplacement of the Kakapotahi Granite were also probably close to the eutectic minimum i.e. $\text{c.700 - 750°C}$. Retrogressive metamorphic effects accompanying intrusion of the Kakapotahi Granite are widespread in Greenland Group hornfelses. The replacement of andalusite and cordierite by sericite and quartz, and the development of muscovite, in places defining a secondary cleavage, are the most obvious retrogressive effects. During this metamorphism $\text{SiO}_2$ was lost from the hornfelses and $\text{H}_2\text{O}$ added. No significant changes in concentration of the other elements analysed have been detected.
Thermal effects accompanying both phases of granite emplacement are confined to rocks south of the Mikonui River, at the present exposure level, although it is probable that the Rangitoto Granite underlies the Mount Greenland massif at depth.

The mineralised occurrences of Cedar Creek and Mine Creek are considered to be related to emplacement of the Rangitoto and Kakapotahi granites, respectively.

13.3 POST INTRUSIVE MODIFICATIONS

Post intrusive tectonic activity in the Westland foreland has been dominated by the effects of the Rangitata and Kaikoura orogenies. The Rangitata Orogeny was marked by gradual uplift and block faulting of the Greenland and Tuhua Group basement (Sheppard et. al., 1975) accompanied by granite intrusion north of Greymouth. During the late Cretaceous, c. 80 Ma ago, the New Zealand continent was rifted apart from Australia and Antarctica by the initiation of sea-floor spreading in the Tasman Sea (Molnar et. al., 1975). Lamprophyres alkali basalts and related dyke rocks intruded the foreland during this time.

Following early Tertiary subsidence and marine transgression, further block faulting of the Greenland and Tuhua Group basement occurred as part of the Kaikoura Orogeny. Uplift of faulted basement blocks was considerable, resulting in the formation of the Paparoa Range and probably many of the basement 'highs' including the Mount Greenland massif and Rangitoto Range, now exposed west of the Alpine Fault.

Within the thesis area, numerous post-intrusive faults have been located, and generally strike either NW-SE parallel to bedding, or NE-SW cutting bedding at high angles. The faults are generally normal in character, although the sense and amount of displacement cannot generally be readily determined. Some of these faults cross-cut and displace the mineralised veins of both Cedar Creek and Mine Creek.
Of the major faults recognised, only the Mine Creek and William Tell faults are considered to be of pre-Cretaceous age. The Monteith Fault is probably of Kaikouran age, and the age of the inferred Mikonui Fault is unknown. In spite of considerable post-intrusive faulting, no associated folding of the Greenland Group has been detected.

Two alkaline basalt dykes located in the thesis area were probably emplaced during the Cretaceous Rangitata Orogeny and are considered to be related to rifting of the Tasman Sea.

THE MINERALISATION - CEDAR CREEK, MOUNT GREENLAND

13.4 SOURCE OF THE MINERALISATION

13.4.1 Heat and Fluids

The sulphur isotopic composition of pyrite from the Cedar Creek veins (\( \delta^{34}\text{S} = -3.00 \pm 1.13\% \)) indicates a predominantly magmatic origin for the sulphide sulphur present, and by inference a magmatic origin for the mineralising fluids.

The nearest known intrusive body to the Cedar Creek veins is the Rangitoto Granite immediately to the south of the Mikonui River. This granite is inferred to be separated from the Greenland Group metasediments of the Mount Greenland massif by the Mikonui Fault (see Geological Map - map pocket), a large normal fault with considerable upthrow to the southwest. The lack of hornfelsing of the Greenland Group metasediments of Mount Greenland combined with the structural setting of the mineralised veins and the sulphur isotope data, combine to suggest that an intrusive body, probably the Rangitoto Granite, underlies the Mount Greenland massif at depths of 2km or more below the present erosion level. Accordingly the Rangitoto Granite is regarded as being the most likely source of both the heat and fluids necessary for the mineralisation of the Cedar Creek veins.
The Rangitoto Granite was probably emplaced to depths of less than 8km at a temperature of 700 - 750°C before crystallising completely.

Migration of the magmatic mineralising fluids has been controlled by the structure of the overlying Greenland Group sediments (see Fig. 13.1), particularly the attitude of folds and faults. The location of the mineralised veins in the crest of the William Tell Anticline indicates that the fluids probably migrated upwards from the underlying granite, along bedding planes into the crest of the anticline, and thence upwards along the William Tell Fault.

The fractured and faulted hinge of the anticline has provided both a favourable route for the escape of hydrothermal fluids, and a site for accompanying ore deposition.

Local variations in the permeability of the William Tell Fault, perhaps caused by intersecting faults active during mineralisation, or by local variations in the degree of fracturing along the fault, may have given rise to both variations in the mineral assemblages present, and to the formation of payable ore shoots, such as that worked in the William Tell No. 2 level.

13.4.2 Gangue Minerals

Quartz and minor calcite are the only significant gangue minerals in the Cedar Creek veins. No direct evidence exists to indicate the origin of these minerals. The magmatic origin of the sulphide sulphur of the veins, and by inference the mineralising fluids, suggests that the quartz of the veins is probably of magmatic origin also. Some of the quartz present in the veins could, however, have been derived by wall rock alteration processes, and deposited as fissure infillings in the vein system. Although no geochemical study of the wall rock has been attempted, there is no visible evidence of extensive wall rock alteration adjacent to the veins, and accordingly wall rock alteration processes are not regarded as having supplied significant silica to the veins for deposition as quartz.

Calcite is regarded as having been either deposited directly from late stage magmatically derived fluids or alternatively
Fig 13.1 Schematic model for the genesis of the Cedar Creek vein mineralisation. Arrows indicate the direction of fluid migration. Py = pyrite; aspy = arsenopyrite; cpy = chalcopyrite; sp = sphalerite; gn = galena; bl = boulangerite; bo = bournonite; tt = tetrahedrite; Au = gold; sb = stibnite.

See text for discussion.
deposited by the reaction of calcium, released from the wall rock by alteration processes, with the mineralising fluids. A geochemical investigation of the wall rock would be required to determine which of these alternatives is most likely.

A supergene origin for the vein calcite is unlikely, in view of the association of calcite with the hypogene sulphide and sulphantimonide minerals, and the general absence of supergene alteration of these minerals.

13.4.3 The Ore Minerals

The ore minerals, pyrite, chalcopyrite, arsenopyrite, sphalerite, galena, boulangerite, bournonite, tetrahedrite and gold, are all of primary magmatic origin, both the metals and sulphide sulphur being derived from a granitoid, probably the Rangitoto Granite, underlying the Mount Greenland massif at depth. The envisaged source of the metals and their migration paths is shown in Fig. 13.1.

The sulphur isotopic composition of the Cedar Creek pyrite ($\delta^{34}S = -3.00$ to $+1.13\%$) indicates a predominantly magmatic origin for the sulphide sulphur in the veins.

The origin of the sulphide forming metals is less well established however, although several lines of evidence indicate a magmatic origin for the metals. The structural setting of the veins indicates derivation of the hydrothermal fluids from a granite body at depth. Late stage magmatic fluids such as those envisaged are commonly enriched in the sulphide forming metals viz. Cu, Pb, Zn, Fe and Sb, as well as As, all of which are present as ore minerals in the Cedar Creek veins. Furthermore, there is no evidence for derivation of the metals from either regional alteration of the Greenland Group metasediments accompanying granite emplacement, or from extensive wall rock alteration. This lack of alteration, coupled with the generally low levels of Cu, Pb, Zn, Sb and As in the Greenland Group suggest that the Greenland Group is unlikely to be the source of these elements in the vein deposits.

The presence of sulphantimonide minerals in the vein system further indicates a magmatic origin for the sulphide forming metals. Chang and Bever (1973) state that "... creditable
descriptions of mineral deposits that contain lead sulphosalts shows that essentially all of these are deposits of magmatic affiliation ..., most of the deposits being "cool-hypothermal and mesothermal" vein deposits. Accordingly, it seems probable that the Cedar Creek sulphides and sulphantimonides are also of magmatic origin.

The above features, coupled with the sulphur isotope data, clearly indicate a magmatic origin for the sulphur-bearing ore minerals.

Gold and silver, present either as free gold or in the sulphur-bearing minerals, are also of magmatic origin. Wall rock alteration could not have provided the quantity of gold (122,316g) won from the veins - the minimum quantity of metasediments required to supply this amount of gold to the veins would be approximately $47 \times 10^6$ tonnes ($\approx 0.18\text{km}^3$). No evidence for alteration of this amount of wall rock has been observed. Accordingly, the bulk of the gold present in the Cedar Creek veins is considered to be of magmatic origin. The geochemical similarity of silver to gold, and its association with the sulphosalt minerals, almost certainly reflects a magmatic origin for the silver present in the veins also.

No evidence for supergene gold enrichment has been noted.

13.5 TIMING OF THE MINERALISATION

The Cedar Creek vein mineralisation is genetically related to emplacement of a granitoid body into the overlying Greenland Group sediments, which host the mineralisation. The granitoid is not exposed at the present erosion level, and accordingly its age is uncertain, however, the lack of hornfelsing of the Greenland Group metasediments at horizontal distances of c.1km from the Rangitoto Granite, to the south of the Mikonui River (see Geological Map and cross sections, map pocket), suggests that the underlying granitoid may be, in fact, a down-faulted block of the Rangitoto Granite (see section 4.2.2 for discussion). If this is indeed the case, as seems most likely, the Cedar Creek mineralisation is penecontemporaneous with
emplacement the Rangitoto Granite i.e. 295 - 314 Ma old. Accordingly, a Carboniferous - Permian age is assigned to the mineralisation.

Mineralisation probably occurred during the waning stages of granite emplacement. Late stage magmatic-hydrothermal fluids are commonly enriched in Cu, Pb, Zn, Sb, Au, As and Ag, all of which are present in the Cedar Creek veins, suggesting late stage mineralisation, probably during the latest stages of granite crystallisation. Such late stage origins are typical of many gold-bearing magmatic-hydrothermal vein deposits.

13.6 CONDITIONS OF MINERALISATION

13.6.1 Temperature-Pressure

The temperature-pressure regime existing in the Cedar Creek veins during mineralisation remains virtually unknown, due to a lack of reliable geothermometric and geobarometric data. Observations based on the mineral assemblages present in the country rock and the mineralised veins indicate that mineralisation occurred below c.700°C the maximum thermal stability of arsenopyrite. Maximum temperatures of late stage boulangerite and tetrahedrite deposition of 648°C and 543°C are indicated by the respective thermal stabilities of these minerals.

These temperatures of deposition intuitively seem too high, however, particularly in view of the general absence of biotite and other thermal metamorphic minerals in the wall rock of the vein. Accordingly maximum temperatures of 400°C are considered likely during mineralisation.

No reliable measure of minimum mineralisation temperatures has been obtained, all of the ore minerals being stable to low temperatures, however, the absence of bornite in the vein assemblage suggests possible late stage deposition of the
sulphantimonides above c.230°C, the maximum thermal stability of bornite. Accordingly, 230°C is regarded as a possible minimum temperature of sulphantimonide deposition, although gold deposition could have persisted to lower temperatures.

The paragenetic sequence determined for the vein assemblage reflects deposition of the ore minerals with decreasing temperature.

Pressures in the vein system during mineralisation are unknown, but were probably less than lithostatic load pressures (c.2kbars). This seems particularly likely in view of the infilling nature of the vein system.

13.6.2 Composition of the Ore Fluids

The sulphantimonide assemblage in the Cedar Creek veins reflects changing composition of the mineralising fluids with decreasing temperature and time (see Fig.13.2). The observed paragenesis, galena-boulangerite-(jamasonite?)-bournonite-tetrahedrite, reflects a decreasing activity of sulphur and PbS in the mineralising fluids accompanied by an increased activity of Sb₂S₃ in the late stage fluids. Activities of sulphur in the inferred temperature range of deposition (400°C - 250°C) probably decreased from approx. 10⁻⁴ atm to 10⁻¹² atm during the latest stages of deposition. Arsenic, also apparently had a low activity in the late stage mineralising fluids, as evidenced by the lack of late stage sulpharsenide minerals, although higher arsenic activities may have existed during the early stage higher temperature arsenophrite deposition.

An increasing copper content of the mineralising fluids with time is indicated by the late appearance of the Cu-bearing sulphosalts, bournonite and tetrahedrite. This trend reflects the differing differentiation trends of copper and lead in the late stage mineralising fluids.

The acidity and salinity of the mineralising fluids are unknown. It seems probable, however, that the fluids were not
Fig 13.2 Diagram showing compositional changes in the ore fluids during mineralisation.

- Decreasing; - Increasing.

See text for discussion.
strongly acidic, and were at least weakly saline. Such fluids have been recognised by Fyfe and Henley (1973), and others, as being efficient transporters of gold, and transition and base metals.

Other compositional characteristics of the ore fluids are unknown.

13.7 CONTROLS ON THE MINERALISATION

13.7.1 Structural Controls

As noted previously, the Cedar Creek veins are located in a normal fault zone in the crest of the William Tell Anticline. Although the vein system is traceable for several hundred metres known economic mineralisation is confined to ore shoot in the William Tell workings. Unfortunately access could not be gained to these workings and accordingly controls on the configuration of the ore shoot are unknown. It is possible, however, that the ore shoot represents a zone of higher permeability within the vein system, possibly due to the effects of locally increased fracturing either associated with movement on the William Tell Fault or the intersection of secondary faults with the vein system.

13.7.2 Gold Distribution

The distribution of gold within the veins is not well known and accordingly controls on the distribution of gold are only poorly understood. Gold occupies a late position in the paragenetic sequence, commonly occurring with the lead sulphosalts where these are present. In such instances gold occurs both as a replacement of the sulphosalts, and as intergrowths with the sulphosalts.

Gold also shows a tendency to occur as a replacement of early formed arsenopyrite, and as fracture infillings around wall rock selvages. The gold-arsenopyrite association is common in many gold deposits and is probably due to the similar
covalent radius of arsenic (1.39Å) and gold (1.40Å) as well as the low but significant affinity of gold for sulphur (Boyle, 1979, p.32). Deposition of such gold probably occurs by replacement of both arsenic and iron in the arsenopyrite lattice.

The association of gold with wall rock selvages is not readily explained. Presumably, however, the selvages are chemically favourable sites for gold deposition.

The deposition of gold probably occurred as a result of decreasing temperatures of the hydrothermal fluids. Fyfe and Henley (1973) have shown that the solubility of gold as chloride complexes decreases markedly with decreasing temperature below c.250°C. Such a mechanism readily explains the late paragenetic position of gold in the Cedar Creek veins, and is in good agreement with the inferred temperatures of gold deposition in the Cedar Creek veins.

Compositional changes in the ore fluids, while controlling the sulphide and sulphosalt species deposited would probably have had a significantly smaller effect on gold deposition than the effects of decreasing temperature. Gold has only low solubilities in the sulphosalt species and would probably be little affected by changes in the sulphosalt species deposited. Silver, in contrast can occur in significant concentrations in the sulphosalt minerals, particularly tetrahedrite, and should show a strong tendency to occur with the sulphosalts, as well as the gold, of the Cedar Creek veins.

No evidence for early deposition of either gold or silver has been observed.

13.7.3 Sulphide and Sulphosalt Deposition

Sulphide and sulphosalt deposition in the Cedar Creek veins has been controlled by both decreasing temperature, and changes in the composition of the ore forming fluids with time. Sulphide minerals viz. pyrite, arsenopyrite, chalcopyrite, sphalerite and galena were deposited relatively early in the
paragenetic sequence, probably at temperatures of $300^\circ - 400^\circ C$ under conditions of relatively high sulphur activity ($10^{-4}$ atm.). Declining sulphur and possibly FeS activities with time, accompanied by increasing activity of $\text{Sb}_2\text{S}_3$ and decreasing temperatures, resulted in the subsequent deposition of galena, boulangerite, bournonite, tetrahedrite and possibly jamesonite.

The relationships of these minerals to compositional changes in the ore fluids has been discussed in section 13.6.2.

The genesis of the Cedar Creek vein deposits is summarised schematically in Fig. 13.1.

THE MINERALISATION - MINE CREEK, MOUNT RANGITOTO

13.8 SOURCE OF THE MINERALISATION

13.8.1 Heat and Fluids

Emplacement of the Kakapotahi Granite during the Triassic, c.214 Ma ago, is considered to have provided the heat necessary for the mineralisation of the Mine Creek veins. The granite was probably emplaced to depths of less than 8km before crystallising completely. Deposition of the mineralised veins occurred from hydrothermal fluids derived from the Kakapotahi Granite.

The sulphur isotopic composition of the Mine Creek sulphides clearly reflects a magmatic origin for the sulphide sulphur. $\delta^{34}\text{S}$ values for the main sulphide species are as follows:

\[
\begin{align*}
\delta^{34}\text{S} (\%) \\
\text{Pyrite} & : +0.22 - +1.60 \\
\text{Sphalerite} & : +0.17 - +0.79 \\
\text{Galena} & : -0.44 - +0.15
\end{align*}
\]

The $\delta^{34}\text{S}$ values have a narrow range and lie close to zero, reflecting derivation from an homogeneous source. The values are typical of those of sulphides of magmatic origin, and
accordingly a magmatic origin for the sulphide sulphur, and by inference, the mineralising fluids, seems probable. $^{34}\text{S}$ values from sedimentary or mixed sources generally show a greater range than those obtained in this study.

Propylitic alteration of the Rangitoto Granite adjacent to the Mine Creek Fault further indicates that the mineralising fluids were derived from the nearby Kakapotahi Granite. The Mine Creek Fault, which was probably active during emplacement of the Kakapotahi Granite, has acted as a channelway for the escape of the hydrothermal fluids, which migrated upwards along the fault, and through the now mineralised shears worked in both Manton's and Schonfelter's workings. Although the mineralised shears have not been traced to their intersection with the Mine Creek Fault, it seems almost certain that the shears and the Mine Creek Fault are interconnected, together providing a channelway for the escape of the mineralising fluids derived from the Kakapotahi Granite.

Metamorphic fluids derived from the propylitic alteration of the Rangitoto Granite adjacent to the Mine Creek Fault may have contributed to the mineralising fluids. The likely contribution of such fluids is, however, regarded as insignificant in terms of both volume and mineralising potential.

The widespread retrogressive metamorphism of the hornfelsed pendant, which probably accompanied intrusion of the Kakapotahi Granite, indicates that $\text{H}_2\text{O}$ was added to rather than released from the hornfelses during intrusion of the Kakapotahi Granite, further indicating the probable absence of a significant metamorphic component in the mineralising fluids.

No evidence for the presence of significant quantities of meteoric waters in the mineralising fluids has been detected. The sulphur isotope data and the probable depth at which mineralisation occurred (4 - 8km), suggest that most of the mineralising fluids were magmatic in origin.

The envisaged migration of fluids accompanying intrusion of the Kakapotahi Granite and the Mine Creek vein mineralisation is shown in Fig. 13.3.
13.8.2 Gangue Minerals

Quartz, easily the most abundant gangue mineral in the Mine Creek veins may have been derived from two sources. Si contained in the magmatic fluids derived from the Kakapotahi Granite has almost certainly been deposited as quartz in the vein system during mineralisation. Magmatically derived hydrothermal or pneumatolytic fluids are commonly enriched in Si and give rise to quartz veins.

Silica, derived from the retrogressive metamorphism of the hornfelsed pendant accompanying intrusion of the Rangitoto Granite, may also have been deposited in the vein system as quartz. An SiO₂ loss from the hornfelses of 5 - 6% is indicated by the geochemistry of the hornfelses (see Table 5.1). This SiO₂ was possibly removed from the hornfelses by solution in magmatically derived hydrothermal fluids, and deposited in the Mine Creek veins along with the magmatically derived SiO₂. Accordingly, a dual origin for the vein quartz seems likely.

Tourmaline, a minor constituent of the vein assemblages, is regarded as being magmatic in origin. Hypothermal magmatic vein deposits commonly contain minor amounts of tourmaline, probably due to enrichment of boron, lithium and fluorine in late stage magmatic fluids from which the deposits formed.

Calcite, which formed late in the paragenetic sequence is also considered to be of magmatic origin. The limited fluid inclusion data obtained, indicates that the mineralising fluids were CO₂ rich, and accordingly the deposition of calcite in such circumstances could reasonably be expected. Calcite, formed as a result of the alteration of plagioclase in the Rangitoto Granite near the Mine Creek Fault, further supports the likelihood of the vein calcite being magmatic in origin.

Chlorite, present in the veins in trace amounts, is either of latest stage hydrothermal origin, or secondary origin forming as a result of the breakdown of pyrite.
13.8.3 The Ore Minerals

The ore minerals, pyrite, sphalerite, galena, arsenopyrite, chalcopyrite and electrum are regarded as being of primary magmatic origin, both the metals and the sulphide sulphur being derived from the Kakapotahi Granite and deposited from magmatic hydrothermal fluids.

As noted in section 13.8.1, the $^{34}\text{S}$ values of the sulphides reflect a magmatic origin for the sulphide sulphur. Field relationships indicate that the Kakapotahi Granite is the most likely source of the sulphur in the vein sulphides.

The origin of the metals in the vein sulphides can be deduced from several lines of evidence. The general absence of wall rock alteration accompanying vein formation suggests that it is unlikely that the metals present in the veins were leached from the adjacent wall rock. The almost identical Cu, Pb, Zn, Au and Ag contents of the hornfelsed country rocks and the Greenland Group metasediments of Mount Greenland (see Table 5.1) further suggests that these elements have been neither significantly depleted nor enriched during the intrusive and mineralising phases. Accordingly, the hornfelses cannot be regarded as a likely source of the metals present in the vein deposits. The general lack of hydrothermal alteration of the hornfelses provides further evidence that these rocks are not the source of the metals in the veins.

The Kakapotahi Granite, when compared with the Rangitoto Granite and the Greenland Group hornfelses shows a significant enrichment in both Pb (83ppm) and Ag (0.4ppm), but is depleted in Cu (1-2ppm). The high Ag and Pb, and low Cu contents of the Kakapotahi Granite indicate that the Kakapotahi Granite was probably the source of the Pb and Ag present in the Mine Creek veins. Furthermore, the low Cu content of the granite probably explains the presence of only trace amounts of chalcopyrite in the veins. The argentiferous nature of the galena present in the Mine Creek veins, coupled with the presence of electrum in the vein assemblages further suggests that deposition of these minerals occurred from Ag-rich solutions.
Although the source of the Fe, Zn, Au and As present in the veins cannot be determined geochemically, all the major rock types containing similar concentrations of these elements, it seems probable that they were derived from the Kakapotahi Granite also.

Sub-economic mineralisation accompanying propylitisation of the Rangitoto Granite, adjacent to the Mine Creek Fault has resulted in enrichment of Cu, Pb, Zn, Sb, Ag and Au, accompanied by the formation of minor amounts of pyrite. An associated enrichment in Sn suggests that the mineralising fluids were magmatic in origin and probably derived from the Kakapotahi Granite. Accordingly, the Kakapotahi Granite is considered to be the source of the sulphur and metals present in both the ore minerals of the vein deposits, and the sub-economic disseminated mineralisation adjacent to the Mine Creek Fault.

13.9 TIMING OF THE MINERALISATION

Although attempts to date the mineralised veins in Mine Creek, using K-Ar methods were unsuccessful, dates obtained from the major rock units combined with field relationships enable the timing of the mineralisation to be determined within reasonable limits.

Intrusion of the Kakapotahi Granite occurred c.214 Ma ago, as indicated by K-Ar dating, and was accompanied by the formation of the mineralised veins in Mine Creek, which are accordingly regarded as being Triassic in age.

Vein formation could have occurred at any stage of granite emplacement, however, it seems most probable that the veins were deposited from late stage pneumatolytic or hydrothermal fluids. A late origin is suggested by the CO₂ rich nature of the ore forming fluids, and their apparent enrichment in Pb, Ag, Au, Sb, Cu and As, all elements which are commonly concentrated in late stage hydrothermal fluids.
Accordingly, the Mine Creek mineralisation is regarded as being of Triassic (c.214 Ma) age, and as having formed during the latest stages of emplacement of the Kakapotahi Granite. Propylitisation and sub-economic mineralisation adjacent to the Mine Creek Fault also occurred at this time.

13.10 CONDITIONS OF MINERALISATION

13.10.1 Temperature-Pressure

Temperatures of mineralisation in the Mine Creek veins are not well defined, due to an absence of suitable fluid inclusions for geothermometry, and the non-validity of temperature estimates based on both sulphur isotope and sphalerite compositional data.

The presence of arsenopyrite in the vein assemblage indicates that deposition occurred at temperatures below c.700°C, the maximum thermal stability for arsenopyrite. The presence of pyrite rather than pyrrhotite in the veins, however, suggests that deposition probably occurred at temperatures considerably below 700°C, pyrite being stable only over very narrow ranges of $\lambda_{S_2}$ above about 500°C - 550°C. Accordingly 550°C is regarded as a realistic maximum temperature of mineralisation. The presence of tourmaline in the vein assemblage further indicates hypothermal temperatures of mineralisation.

No good minimum estimates of mineralisation temperatures have been obtained from the vein assemblage. The general absence of low temperature minerals in the vein assemblage suggests that mineralisation temperatures were probably greater than c.350°C.

The single fluid inclusion studied from the vein system indicates a maximum temperature of quartz deposition of 450°C, at geologically reasonable pressures. The paragenetic position of the host quartz is uncertain, however, and until further
data is available this figure must be taken as an indication of possible mineralisation temperatures only.

Consideration of these data combine to indicate likely temperatures of mineralisation ranging from $350^\circ \text{C} - 500^\circ \text{C}$. Retrogressive metamorphic features in the hornfelsed country rock are compatible with this temperature range.

Pressures in the vein system at the time of mineralisation are unknown, estimates based on sphalerite geobarometry being geologically unreasonable. The inferred maximum depth of emplacement of the Kakapotahi Granite is 8km ($P_{\text{load}} = 2\text{ kbars}$). Assuming that $P_{H_2O}$ in the vein system was less than $P_{\text{load}}$, which seems probable in the case of a shear zone infilling such as this, maximum pressures ($P_{H_2O}$) of 2 kbars are indicated during mineralisation. Mineralisation pressures ($P_{H_2O}$) of 1 - 2 kbars are accordingly inferred for the Mine Creek veins.

### 13.10.2 Composition of the Ore Fluids

Fluid inclusion data indicates that the mineralising fluids were $CO_2$ rich, and had a salinity of c.18% (NaCl equivalent). The high $CO_2$ content indicated readily explains the presence of calcite as stockworks adjacent to the Mine Creek Fault, and minor amounts of late stage calcite in the vein assemblage.

The mineralisation temperatures envisaged suggest that the mineralising fluids had a high sulphur activity ($\log A_{S_2} \approx -5 \text{ atm}$, or more). Although the acidity and composition of the mineralising fluids are otherwise unknown, consideration of the sulphur isotope data obtained coupled with the observed mineralogy suggests that the fluids probably had a low oxygen fugacity and were only weakly acidic or alkaline.

Hydrothermal fluids such as those envisaged, viz. $CO_2$-rich, saline, weakly acidic or alkaline fluids with a high $A_{S_2}$, have been identified as the mineralising fluids in many precious metal vein deposits, and are efficient transporters of both Au and Ag, as well as the more common sulphide forming metals.
13.11 CONTROLS ON THE MINERALISATION

13.11.1 Structural Controls

Known vein mineralisation at Mine Creek is confined to three sub-parallel shears striking NE-SW, and dipping at 20° - 40° NW, located near the crest of the Mine Creek Anticline. The veins cut both bedding and the axial surface of the anticline obliquely. The shears are probably connected to the Mine Creek Fault at depth, and were probably active contemporaneously with both mineralisation and movement on the Mine Creek Fault.

Propylitisation and sub-economic gold-silver mineralisation of the Rangitoto Granite occurs on the footwall of the Mine Creek Fault, which has apparently been a major channelway for magmatic fluids passing into the now mineralised shear system exposed in Manton's and Schonfelter's workings.

13.11.2 Gold Distribution

Gold occurs predominantly as electrum associated with pyrite in the Mine Creek veins, although undoubtedly both pyrite and arsenopyrite, and possibly sphalerite, all contain minor amounts of gold. In contrast to the sulphides, the vein quartz does not contain significant gold.

The major controls on gold deposition are not well known. Electrum, the major gold-bearing phase, occurs only as fracture infillings or inclusions in earlier formed pyrite. This probably reflects the strong affinity that gold has for pyrite. Arsenopyrite, which also has a strong affinity for gold, shows no tendency to occur with electrum, probably because of the early paragenetic position of arsenopyrite.

The paragenesis of the Mine Creek sulphides indicates that electrum deposition occurred as a response to decreasing temperatures and changes in composition of the ore fluids. The limited data available suggest that the activities of $S_2$, FeS and possibly As decreased during mineralisation,
whereas concentrations of Pb, Zn, Au, Ag and Cu, and possibly CO$_2$, in the mineralising fluids increased during mineralisation. These compositional changes, coupled with decreasing temperature, were probably responsible for the deposition of electrum in the Mine Creek veins. Possible changes in the acidity and salinity of the mineralising fluids may also have accompanied late stage electrum deposition.

No evidence for significant changes in pressure during mineralisation processes has been noted.

13.11.3 Silver Distribution

Silver occurs predominantly in galena, as a substitution for Pb in the crystal lattice, and to a lesser extent in solid solution with gold in electrum. Trace amounts of silver probably occur in the sulphide minerals also, particularly chalcopyrite.

Argentiferous galena occurs as a replacement of pyrite throughout the veins, the pyrite apparently acting as a nucleation site for the late stage galena. Why this should be so is unclear. It is possible, however, that reduced stability of pyrite, induced by compositional changes in the mineralising fluids, may have resulted in the incipient breakdown of pyrite and release of S$^-$. This S may have reacted with Pb$^+$ in the mineralising fluids to form galena, as a replacement of pyrite.

Deposition of late stage galena from fluids enriched in both Pb and Ag, coupled with the similar geochemical properties of the elements, has resulted in the deposition of argentiferous galena, as a replacement of earlier formed pyrite.

Controls on the deposition of electrum, the other major silver-bearing phase, have been described in the preceding section.
13.11.4 Base Metal Distribution

The base metal sulphides (galena and sphalerite) occur as late stage replacements of pyrite in the Mine Creek veins. Deposition of these sulphides probably occurred as a response to changes in composition of the ore fluids, coupled with decreasing temperature. Early mineralising fluids with high activities of FeS and S$_2^-$, and containing significant quantities of As, became relatively depleted in these components during mineralisation. This depletion was accompanied by an enrichment in Pb, Zn, Ag and Au in the fluids resulting in base metal deposition, as a replacement of earlier formed pyrite.

The low concentrations of chalcopyrite present in the veins is probably due to the low concentrations of Cu in the Kakapotahi Granite, which is the source of the metals present, coupled with the low acidity of the mineralising fluids envisaged.

No evidence for significant pressure changes during mineralisation have been noted.

The genesis of the mineralised occurrences of the Mine Creek area is summarised schematically in Fig. 13.3.
Fig 13.3 Schematic model for the genesis of the Mine Creek vein and disseminated ore deposits. Py = pyrite; aspy = arsenopyrite; sp = sphalerite; gn = galena; cpy = chalcopyrite. See text for discussion.
CHAPTER XIV

ECONOMIC CONSIDERATIONS

14.1 IMPLICATIONS OF THE GENETIC MODEL

The genetic models outlined in the previous chapter have significant economic implications. The recognition of two probable phases of granite emplacement in the thesis area, the importance of structural control on the mineralisation present, and the recognition of the source of both the metals and sulphur present in the deposits and the mineralising fluids, are in particular economically significant.

Polyphase intrusive systems, particularly those of monzonitic or granitic composition, are particularly favourable systems for the formation of large volume - low grade disseminated copper-molybdenum and gold-silver 'porphyry' type ore deposits. Chappell and White (1974), note that tungsten and porphyry copper-molybdenum mineralisation tend to be associated with I-type granites, whereas S-type granites such as the Rangitoto and Kakapotahi granites tend to be associated with Sn mineralisation. This general trend combined with the granitic rather than monzonitic character of both granites and the equigranular rather than porphyritic nature of the older Rangitoto Granite, together with the low Cu content of the Younger Kakapotahi Granite combine to reduce the prospectivity of the Rangitoto Granite as a potential host to disseminated porphyry copper-molybdenum mineralisation.

A thin quartz vein containing both chalcopyrite and stibnite, was however located in a hornfelsed Greenland Group float boulder in the eastern branch of Mine Creek (G.R.S64/368198) suggesting the possibility of disseminated copper-molybdenum mineralisation, at depth in the Rangitoto Granite. No other evidence of the possible existence of such mineralisation has been noted at the present exposure level.
Although both the Rangitoto and Kakapotahi granites both contain higher than normal concentrations of Sn (≈ 65ppm), no evidence for tin mineralisation has been noted. Although the possibility of tin mineralisation in either the Rangitoto or Kakapotahi granites cannot be ruled out, it should be noted that high concentrations of tin in granite bodies often occur in unmineralised granitoids. The S-type origin of the granites in the thesis area, and their similar geological setting to granitoids containing tin mineralisation elsewhere in the world e.g. Bolivia, suggest a favourable geological environment for tin mineralisation. It should be noted, however, that tin mineralised granitoids have not been reported from elsewhere in the Westland foreland to date.

Sub-economic disseminated gold-silver mineralisation accompanying propylitic alteration is present in the Rangitoto Granite along the footwall of the Mine Creek Fault. The higher permeability of the fractured Rangitoto Granite adjacent to the fault has allowed hydrothermal fluids derived from the Kakapotahi Granite to alter and mineralise the adjacent Rangitoto Granite. Similar faults, or zones of high permeability, have not been located elsewhere in the Rangitoto Granite, in the thesis area. Such permeable zones, particularly faults, if located elsewhere in the Rangitoto Granite could, however, be altered and mineralised also, particularly if they are in close proximity to the Kakapotahi Granite which is the source of both the mineralising fluids and ore forming elements.

The vein deposits of Mine Creek almost certainly continue at depth beyond Schonfelter's workings to intersect the Mine Creek Fault, which was the major feeder of the ore fluids to the vein system. The possible existence of a significant orebody at the intersection of the vein system with the Mine Creek Fault is indicated by the genetic model, and supported by the reported increasing thickness and silver content of the veins in Schonfelter's workings. If such an ore shoot exists, it would probably plunge northeastwards along the line of the Mine Creek Fault. No such orebody has been
observed southwestward along the Mine Creek Fault where it would be expected to intersect the present ground surface, however, exposure of the Mine Creek Fault in the likely area of intersection is poor. The possible existence of an ore shoot along the vein-fault intersection, as predicted by the genetic model, accordingly, cannot be discounted.

The genetic model developed for the Cedar Creek mineralisation also has several significant economic implications. The magmatic origin of the ore forming elements suggests derivation from a granite body, probably the Rangitoto Granite, underlying the Mount Greenland massif at depth. Gold, and the sulphide-forming elements are derived almost entirely from this source, no supergene enrichment being evident.

Deposition of the vein system occurred in structurally prepared sites i.e. the faulted crest of the William Tell Anticline, although economic gold mineralisation has been located in only a small portion of the vein system. The recognition of the importance of structural control on vein formation and ore deposition, as noted by Gage (1948) in the Reefton quartz veins, obviously is of significance in terms of prospecting for other mineralised veins in the area.

The existence of sulphantimonide minerals in the Cedar Creek veins, and their common occurrence around the margins of Pb-Zn base metal deposits (Ramdohr, 1969) is also economically interesting, suggesting the possibility of base metal mineralisation at depth in the vein system, or possibly along the Greenland Group - granite contact.

Clearly, the general implications of the genetic model discussed above, are of considerable importance in planning a prospecting strategy both for evaluation of the known mineralisation, and in the search for as yet undiscovered mineralisation.
14.2 FUTURE PROSPECTING

14.2.1 Disseminated Ore Deposits

The probability of locating disseminated ore deposits of economic grade in the thesis area is regarded as being low. Available evidence suggests that if disseminated deposits occur in the Rangitoto Range area they are likely to occur in the Rangitoto Granite adjacent to fault zones of Triassic age or older age (e.g. the Mine Creek Fault), or in similarly permeable areas. Gold, silver and possibly minor base metals are the most likely ore minerals to be found in such areas.

Stream sediment sampling undertaken by the Carpentaria Exploration Co. between 1970 and 1972 failed to reveal significant base metal anomalies over the Rangitoto Range. Accordingly, if disseminated copper-molybdenum mineralisation occurs in the Rangitoto Granite in or near the thesis area, it is almost certainly below the present exposure level, and would be detected only by geophysical techniques or diamond drilling. The likelihood of such mineralisation occurring at depth in the Rangitoto Granite is low, but cannot be discounted.

Although no tin mineralisation has been detected in the thesis area, geological conditions for such mineralisation appear favourable. Stream sediment sampling over the thesis area and surrounding areas, would possibly reveal the presence of any existing tin mineralisation exposed at the present erosion level.

14.2.2 Vein Deposits - Mount Rangitoto

Precious and base metal-bearing veins have not been reported from elsewhere on the Rangitoto Range to date. If such veins do exist it seems probable that they will also be located along faults or shears active contemporaneously with emplacement of the Kakapotahi Granite. Accordingly, attempts to locate such structures should be made in prospecting for further mineralised veins, and be accompanied in the first instance by stream sediment and soil sampling of the hornfelsed pendant area.
The likelihood of discovering significant vein systems in the hornfelsed pendant is, however, considered to be extremely low. Further evaluation of the mineralised veins in Schonfelter's workings as a potential small volume - high grade gold and silver deposit is considered worthwhile. Attempts to clear the fall in the access drive should be made in the first instance, and if successful be accompanied by detailed mapping and sampling of the exposed veins. If results are encouraging, diamond drilling from within the drives could be undertaken to further evaluate the veins.

Attempts to determine the possible existence of a larger orebody developed along the intersection of the Mine Creek Fault with the Mine Creek vein system could be undertaken by geophysical methods, e.g. ground magnetometer or resistivity surveys across the Mine Creek Fault in the west branch of Mine Creek.

14.2.3 Vein Deposits - Mount Greenland

Auriferous vein deposits, apart from those in the crest of the William Tell Anticline, could be discovered within or around the thesis area, however thorough searches for such veins in the past have failed to locate payably auriferous occurrences. Accordingly, the chances of making new discoveries are considered to be low. Any search for such veins should, however, be concentrated along the axial regions of synclines and anticlines, particularly where faulting is detected in the hinge regions of the folds. Shortening of the regional fold wavelength may also be associated with vein formation, and mineralisation. Initially mapping and stream sediment sampling should be undertaken, possibly using associated elements e.g. antimony, copper, arsenic and lead as pathfinders for gold.

Further evaluation of the workings in the William Tell No.1 and No.2 levels appears warranted and should be made initially by gaining access to the No.2 level followed by mapping and sampling of the vein system. If results are
encouraging diamond drilling of the vein from within the drives could be undertaken to further evaluate the extent and grade of the veins.

The possibility of base- and precious-metal mineralisation occurring in the vein system at depth is indicated by the presence of sulphantimonide minerals in quartz of the William Tell levels should also be noted. No prospecting for such veins is considered to be warranted, however.
15.1 INTRODUCTION

The rather general approach adopted in this thesis, although providing a good framework for the establishment of a genetic model for the Mine Creek and Cedar Creek mineralisation, has resulted in only a rather cursory examination of some aspects of the ore deposits. Such an approach was considered desirable, however, in view of the absence of background data on both the mineralised occurrences and details of the surrounding geology.

Unfortunately, because access could not be gained to the William Tell No.2 level and Schonfelter's workings, as had originally been hoped, some aspects of both the Cedar Creek and Mine Creek vein deposits remain uninvestigated. The resulting lack of detailed data on the vein deposits is regarded as one of the most serious shortcomings of this study, however, it is hoped that data obtained from the accessible portions of the orebodies, and examination of old records and samples, has helped in part to overcome this problem. It is acknowledged, however, that revision of some aspects of the genetic model developed for the mineralisation would probably be required if access were gained to the William Tell No.2 level and/or Schonfelter's workings.

15.2 RECOMMENDATIONS FOR FUTURE RESEARCH

15.2.1 The Mineralisation

The following aspects of the mineralised occurrences are considered to be worthy of future research:
a) **Vein Mineralogy, Cedar Creek.** The late stage sulphantimonide minerals identified in the Cedar Creek veins should be further investigated. A quantitative electron microprobe study, coupled with x-ray diffraction work is required to positively identify all phases present. Combination of such a study with reflected light microscopy, would yield more reliable data on both the mineralogy of the sulphantimonides, and their paragenetic relationships, than that contained in this study. Furthermore, such a study would help considerably to determine conditions of sulphantimonide and associated gold mineralisation, thus being of both academic and economic interest.

The absence of sulphantimonide minerals in auriferous veins in the Greenland Group of the Reefton area (stibnite is the Sb-bearing phase in this area) suggests differing conditions of mineralisation in the two areas. Detailed studies of the vein assemblages, particularly the Sb-bearing phases could help to determine the conditions of mineralisation in the two areas.

b) **Geothermometry.** Temperatures of deposition of the vein minerals of both Cedar Creek and Mine Creek are poorly known. Although a preliminary attempt to obtain such data by fluid inclusion studies was abandoned, due to a lack of time and the general difficulty of obtaining suitable inclusions to work with from the limited material available, careful sample selection and preparation and thorough examination at greater leisure would probably overcome these difficulties. Such a study would, as well as enabling better estimation of the temperatures of mineralisation to be obtained, also provide valuable information on the nature of the ore-forming fluids, particularly their CO₂ content and salinity.

c) **Wall Rock Alteration.** No attempt has been made in this thesis to study the wall rock alteration accompanying vein formation and mineralisation. Such a study, by petrological and geochemical methods, could further elucidate the origin of the mineralised veins, particularly temperatures of formation. Systematic sampling and sample analysis would be required as part of such a study.
15.2.2 The Geological Setting

a) Isotopic Age Dating. The K-Ar dates obtained in this study provide reasonable age estimates for the Rangitoto Granite and the Greenland Group hornfelses analysed. The young age of the Kakapotahi Granite obtained, c. 212 Ma, should, however, be regarded as tentative until a Rb-Sr age is determined for the granite. In view of the importance of this granite in terms of the genetic model proposed for the mineralisation such a study is considered to be worthwhile. Furthermore, the recognition of Triassic aged granitoids from elsewhere in the western foreland, over recent years, suggests that Triassic intrusive activity in the foreland may have been more widespread than formerly believed. Attempts should also be made to map the extent of the Kakapotahi Granite beyond the thesis area, particularly if the granite does in fact yield a Triassic Rb-Sr age.

b) Petrology and Geochemistry. Some aspects of the preliminary petrological and geochemical investigations undertaken in this study are worthy of further investigation. Investigations of the nature and extent of alteration of the Rangitoto and Kakapotahi Granites both petrologically and geochemically, within and beyond the thesis area, could result in the identification of potentially mineralised areas, as well as better defining the origin and conditions of granite emplacement. Similar studies of the Greenland Group metasediments and hornfelsed equivalents would also provide more data on the nature and extent of the chemical and mineralogical changes accompanying contact metamorphism and metasomatism.

Additional studies of both the mineralisation and the surrounding rocks, such as those proposed, would help to both refine the genetic model developed for the mineralisation, and to increase our understanding of the Paleozoic geology of the western foreland.
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APPENDIX I

PLACE, STRUCTURE AND LITHOLOGICAL NAMES USED IN THIS THESIS

During the course of this study it was considered necessary to devise some locality names, to enable ease of description. In the case of Mount Rangitoto, localities have been named after people involved in the early prospecting and mining ventures, and structures and rock types mapped have been named after localities in which they occur, commonly creeks. Structures located on the Mount Greenland massif have also been named after the localities in which they occur.

The following names have been erected for use in this thesis:

1. Mine Workings

a) Manton's Workings: Two drives and stopes connected by shafts, inclines and winzes to lower levels, constructed under the direction of Frederick J. Manton, the first Mine Manager of the Mount Rangitoto Silver Mining Company. The drives are located approximately 100m from the confluence of the east and west branches of Mine Creek (G.R. 564 366197), in the east branch.

b) Schonfelter's Workings: Schonfelters workings are located 20m from the confluence of the east and west branches of Mine Creek (G.R. 564 365196) approximately 30m vertically and 100m horizontally from Manton's workings. These were the main mine workings constructed by Carl Schonfelter, the second Mine Manager of the Mount Rangitoto Silver Mining Company. The access drive is blocked c.35m from the entrance.

2. Localities

a) Churches Creek: The un-named tributary of Flat Creek immediately southeast of Mine Creek, joining Flat Creek at G.R. 564 347177. The creek is named after Joseph Churches, the first Chairman of the Mount Rangitoto Silver Mining Company.
b) Cedar Creek: The upper reaches of Cedar Creek are incorrectly labelled Langdon's Creek on the NZMS1 S57 Topographic Map (2nd edition, 1970). The name Cedar Creek, appearing on earlier plans and maps, has been used in this thesis. The upper reaches of the creek host the Cedar Creek reefs worked in the William Tell Workings.

3. Geological Structures

Geological structures recognised in this study have been given names of the creeks or localities in which they were either discovered or are best exposed. The structures are named on the accompanying geological map, the following names being erected for use in this thesis:

Faults:
- Mine Creek Fault
- Mikonui Fault
- William Tell Fault
- Monteith Fault

Folds:
- Mine Creek Anticline
- Long Creek Syncline
- Eraser Creek Anticline

Young (1964), identified the William Tell Anticline and the Cedar Syncline.

4. Rock Types

a) Rangitoto Granite: The Rangitoto Granite (described in Chapters 3 and 5) has been named after Mount Rangitoto and makes up the bulk of the Rangitoto Range, extending from the mid-southern slopes northwards to the Mikonui River.

b) Kakapotahi Granite: The Kakapotahi Granite (described in Chapters 3 and 5) crops out in the Kakapotahi River, after which it is named, and on the southern flanks of the Rangitoto Range extending southwards to the Purcell Range. The Kakapotahi Granite is probably younger than the Rangitoto Granite.

The subdivision of the granites is for the sake of clarity and ease of description only. It is not proposed that these
rock types be regarded as anything other than informal units for use in this thesis only.

APPENDIX II

SAMPLE DETAILS

Sample numbers used in the text, together with locations and rock type are given in Table A. All grid references are from NZMS1 1:63,360 topographic maps.

The locations of samples collected from Manton's workings are shown in Figure 1.
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**py** = pyrite; **sp** = sphalerite; **gn** = galena.

Grid references are from NZMS1, 1:63,360 topographic maps

Unknown = sample from Museum or Geology Department collections.

ZM prefix = Canterbury Museum

MG747 = Geology Department, Canterbury University.
Fig 1. Plan of Manton's workings, Mine Creek, showing sample locations, vein thicknesses and attitudes of the veins.
APPENDIX III

ANALYTICAL METHODS

1. X-Ray Fluorescence

X-ray fluorescence analysis was performed on the Philips PW1400 automatic spectrometer in the Department of Geology at Canterbury University. All samples were prepared as compressed powder pellets bonded with a 7% aqueous solution of polyvinyl alcohol. For most elements background intensities were used to correct for variations in mass absorption. International laboratory and synthetic standards were used to construct calibrations (which were the responsibility of Dr S.D. Weaver). The Canada Centre for Mineral and Energy Technology reference materials CD-1 (antimony ore), CCU-1 (copper concentrate), KC-1 (zinc-lead-tin-silver ore) and MP-1 (zinc-tin-copper-lead ore) were used for analysis of high concentrations of S, Fe, Cu, Zn, As, Sn, Sb and Pb.

Scatter line calibrations were erected for the elements Zn, Rb, Sr, Y, Zr and Nb using variations in the intensity of the Au $L_\beta$ tube line to correct for mass absorption differences.

Spectrometer parameters for each element analysed are given in Table B. The precision of the analyses has yet to be assessed, suffice to say that the PW1400 is the highest precision commercially available wavelength-dispersive spectrometer. The accuracy of the major oxide and trace element analyses of the silicate rocks may be judged from Table C in which results obtained for six international standard rocks are compared with recommended values. For other than trace components powder pellet calibrations are less accurate than methods involving dilution and fusion. Unfortunately fusion techniques had not been developed in the laboratory at the time of analysis. The method chosen however has the advantage that it is rapid, and it is believed that the accuracy obtained is acceptable.
### TABLE B: X.R.F. SPECTROMETER PARAMETERS

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<td>Rb</td>
<td>KA</td>
<td>F</td>
<td>S</td>
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<td>75</td>
<td>40</td>
<td>26.595</td>
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<td>KA</td>
<td>F</td>
<td>S</td>
<td>LiF 200</td>
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<td>25.110</td>
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<td>S</td>
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<td>KA</td>
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<td>S</td>
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<td>40</td>
<td>22.525</td>
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<tr>
<td>Nb</td>
<td>KA</td>
<td>F</td>
<td>S</td>
<td>LiF 200</td>
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<td>40</td>
<td>21.345</td>
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<td>0.90</td>
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<td>KA</td>
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<td>S</td>
<td>LiF 200</td>
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<td>13.995</td>
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<td>1.04</td>
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<td>F</td>
<td>S</td>
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<td>Ba</td>
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<td>28.255</td>
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</tbody>
</table>

**Notes:**
1. * Al filter
2. Y KA, Zr KA and Nb KA corrected for interferences from Rb KB, Sr KB and Y KB respectively.
3. Cr tube used with Flow Counter
   Au tube used with Scintillation Counter
<table>
<thead>
<tr>
<th></th>
<th>wt%</th>
<th>ppm</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>AGV-1 (AGV-1)</td>
<td>GSP-1 (GSP-1)</td>
</tr>
<tr>
<td>SiO₂</td>
<td>59.60  59.72</td>
<td>68.10  67.31</td>
</tr>
<tr>
<td>TiO₂</td>
<td>1.05  1.05</td>
<td>0.62  0.66</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>17.40  17.22</td>
<td>15.10  15.19</td>
</tr>
<tr>
<td>TFe₂O₃</td>
<td>6.80  6.84</td>
<td>4.10  4.33</td>
</tr>
<tr>
<td>MnO</td>
<td>0.10  0.10</td>
<td>0.04  0.04</td>
</tr>
<tr>
<td>MgO</td>
<td>1.35  1.55</td>
<td>0.93  0.96</td>
</tr>
<tr>
<td>CaO</td>
<td>5.30  5.00</td>
<td>2.03  2.02</td>
</tr>
<tr>
<td>Na₂O</td>
<td>4.25  4.31</td>
<td>2.80  2.80</td>
</tr>
<tr>
<td>K₂O</td>
<td>2.92  2.93</td>
<td>5.53  5.53</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.48  0.50</td>
<td>0.28  0.28</td>
</tr>
</tbody>
</table>

Notes: 1. Trace concentrations of S, As, Sn and Sb determined by spiking, no recommended values available.
2. Recommended values, in brackets, are those quoted by S. Abbey, 1978, X-ray Spectrometry, 7, 99-121.
2. Atomic Absorption Spectrometry

Gold and silver contents of the silicate rocks were determined at the University of Canterbury, by the methods outlined below. Vein rock samples were analysed for gold and silver by Lime and Marble Ltd, Nelson.

1. Sample Preparation:

Samples of approximately 300g were selected, and broken into small chips. 100-150g of chips were ground for 2 - 8 minutes in a ring mill using a tungsten carbide head, producing a fine powder. Approximately 20g of this powder was used for analysis. The methods of sample preparation are based on those outlined by Hildon and Sully (1971), Tindall (1965) and Newman (1977), with minor modifications.

2. Sample Digestion:

Digestion of the weighed samples was carried out over a steam bath by acid leaching, using concentrated analytical reagent grade acids. 25ml of HCl was added to the sample and heated for 10 - 15 minutes. After this time 20ml of aqua regia was added to the sample, which was then evaporated to dryness over the steam bath. Once dry, a further 20ml of HCl was added to the samples to return the soluble products into solution. 70ml of de-ionised H2O was then added to the sample, which was removed from the steam bath and allowed to cool. When cool approximately 7ml of HBr was added to the solutions to stabilize the gold and silver in solution.

The solution obtained was then filtered to obtain a clear filtrate with was used for concentration.

3. Concentration:

Because of the low gold and silver concentrations in the filtrate, concentration of the gold and silver into a smaller volume is required for analysis. 3ml of the organic solvent MIBK (methyl iso-butyl ketone) was added to each sample, which was shaken vigourously for a few minutes and then allowed to stand until the MIBK was separated from the filtrate. The two solutions were then separated using a separating funnel, the MIBK concentrate being made up to a standard volume of 3.70ml by addition of MIBK.
Duplicate samples were prepared as above, the duplicates being spiked with either 2\( \mu \text{g} \) or 4\( \mu \text{g} \) of gold, for calibration purposes.

Acid filtrates, retained from the separation process were also used to prepare silver calibration standards. 3 sets of 3 filtrates were spiked with 0.1, 1.0 and 10\( \mu \text{g} \) of silver, which was then concentrated into MIBK, as above. These spiked filtrates were used to erect calibration curves for silver.

4. Analysis:

Analysis of the samples and standard solutions was performed on the Varian Techtron AA5 atomic absorption spectrometer in the Chemistry Department, Canterbury University, using settings recommended in the Varian Techtron handbook. Because of the small volumes of analyte prepared (a requirement of the high degree of gold and silver concentration needed to analyse such low sample concentrations) a carbon rod atomiser rather than a flame, was used to atomise the samples. 5\( \mu \text{l} \) samples were placed in the carbon cup using a micro-pippete and the carbon cup was pre-heated using the ramp setting to volatalise the MIBK, prior to atomising the sample remaining in the cup. Ramp and atomising temperatures used were those recommended in the Varian Techtron handbook.

Replicate analyses of gold and silver samples suggest an analytical precision of \( \pm 0.0002 \text{ppm} \) at concentrations of 0.003ppm, and a lower limit of detection for both gold and silver of 0.002ppm.

MIBK standard solutions and blanks were intermixed with sample analysis to check the stability of operating conditions and for possible contamination of the carbon cup.

The gold and silver values obtained in this study are considered to be acceptably accurate in view of the analytical method used.

The digestion method and analytical conditions used by Lime and Marble Ltd, to analyse the vein rocks are unknown.
3. Microprobe Analysis

A preliminary microprobe analysis of mineralised samples was undertaken using the JEOL JXA-58 electron probe x-ray microanalyser in the Geology Department of the University of Otago. Polished sections of mineralised vein material were carbon coated under vacuum, and mounted in brass sample holders, using conducting paint to attach the carbon coated surface to the sample holder. Standards (plates 9 and 10) supplied by the Geology Department, University of Otago were used for analysis of all elements analysed. The materials from which the standards were prepared were unknown to the writer at the time of completing this thesis.

Spectrometer parameters for the elements analysed are given in Table D. Peak scans were used to check and correct for drift during analysis.

The data obtained was processed by an on line computer which provided K values corrected for drift, dead time and background. The K values obtained were determined as follows:

<table>
<thead>
<tr>
<th>Element</th>
<th>Crystal</th>
<th>Radiation</th>
<th>Setting</th>
<th>kV</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cu</td>
<td>LiF</td>
<td>KA</td>
<td>107.20</td>
<td>15</td>
</tr>
<tr>
<td>Ni</td>
<td>LiF</td>
<td>KA</td>
<td>115.35</td>
<td>15</td>
</tr>
<tr>
<td>Fe</td>
<td>LiF</td>
<td>KA</td>
<td>134.70</td>
<td>15</td>
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<tr>
<td>Ag</td>
<td>Pat</td>
<td>LA</td>
<td>133.06</td>
<td>15</td>
</tr>
<tr>
<td>Au</td>
<td>LiF</td>
<td>LA</td>
<td>88.74</td>
<td>15</td>
</tr>
</tbody>
</table>

Notes: 1. Au was analysed at 15kV, rather than the 35kV optimum setting recommended, to enable more rapid analysis. Estimated errors due to this procedure are considered to be less than ± 10% of the stated Au values. (Y. Kawachi, pers. comm.)
\[
K \text{ values} = \frac{I_{\text{unknown}}}{I_{\text{standard}}} \times C_{\text{standard}}
\]

where

- \( I_{\text{unknown}} \) = X-ray intensity of unknown
- \( I_{\text{standard}} \) = X-ray intensity of standard
- \( C_{\text{standard}} \) = Concentration of standard

Where all the major components of a mineral being analysed, e.g. pyrite, chalcopyrite, gold and electrum were determined the K values obtained were fed into an additional correction program to obtain better data. The correction program used was that of Colby (1972) based on the empirical method of Sweatman and Long, the program being held in the computing facility of the Department of Engineering, Canterbury University. Using this program corrections for generation, absorption and characteristic fluorescence were made, however, no correction for continuous fluorescence was possible due to limitations of the computer facility.

In the sulphide minerals S was determined by stoichiometry assuming ideal formulas, of:

- Pyrite \( \text{FeS}_2 \)
- Chalcopyrite \( \text{CuFeS}_2 \)

Where insufficient elements had been analysed to enable Colby's correction programme to be employed, the weight percents of the analysed elements stated are the uncorrected K values obtained. Consequent errors introduced by the lack of correction of the K values are considered to be less than \( \pm 10\% \) of the stated values (Y. Kawachi, pers. comm.), and are therefore significant only when concentrations of the elements being analysed exceed 2-3wt%. The lower limit of detection for all elements is 0.01%.

Because of the uncorrected data obtained for some minerals, and the likely errors in Au analysis the data presented in this thesis is best regarded as being semi-quantitative.
Analytical errors for all elements are considered to be less than $\pm 10\%$ of the quoted values, an acceptable degree of precision for this preliminary type of study. The accuracy of the analyses has not been quantitatively assessed, however, analysis of some standards run as unknowns during the general analysis, suggests that an accuracy of c.$\pm 2-3\%$ was attained at concentrations (of elements other than gold) of 100%.