Dedikasi untuk ayah dan bundaku

# THE PETROGENESIS OF HOST ROCKS AND THE ASSOCIATED HYDROTHERMAL MINERALISATIONS OF THE EAST ARM, GREAT SLAVE LAKE, CANADA

by

Muhammad/Mohd Zuhudi Muda
B.Sc. (Malaya)

A thesis submitted for the degree of
Master of Philosophy

University of Southampton
1979

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UNIVERSITY OF SOUTHAMPTON

ABSTRACT

FACULTY OF SCIENCE

GEOLOGY

Master of Philosophy

# THE PETROGENESIS OF HOST ROCKS AND THE ASSOCIATED HYDROTHERMAL MINERALISATIONS OF THE EAST ARM, GREAT SLAVE LAKE, CANADA

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The East Arm of Great Slave Lake contains a great thickness of lower Proterozoic (Aphebian: 2460 - 1750 m.y.) sedimentary and igneous rocks that are only mildly deformed or metamorphosed. Three separate periods of intrusive magmatic activity have been recognised: 1) Early Aphebian (2200 - 2100 m.y.) alkaline complexes and probably related breccia pipes; 2) Middle Aphebian alkaline volcanism ($\sim$1870 m.y.) with hypabyssal intrusions; and 3) Late Aphebian ($\sim$1790 m.y.) calcalkaline diorite stocks and laccoliths.

All the intrusive rocks in the East Arm contain examples of silver, bismuth, nickel-cobalt arsenide vein-type mineralisation. The later ones contain earlier magnetite-apatite-actinolite with uranium and rare earth element minerals. The mafic margins of the early intrusions contain disseminated copper-iron-nickel sulphides and iron-titanium oxides which also pre-date the vein mineralisations. Some of the ore minerals identified are new to the East Arm - these include stromeyerite, parkerite, hauchecornite, electrum and two Pb-Bi sulphosalts (aikinite? and betekhtinite?). Arsenide mineralisation was also found at a locality where none had been previously reported. Bismuth and nickel do not show the antipathy reported from other arsenide occurrences in the world.

A model is postulated from the detailed study of one of the early Aphebian Complexes for the genesis of the arsenide ore-fluid. It involves hydrothermal leaching principally of mafic minerals. Models for various modes of deposition are derived from study of the late Aphebian Complexes. They involve fluid migration up early joints and deposition in open spaces. The magnetite-apatite-actinolite with uranium and rare earth element mineral deposits were emplaced earlier in joints, sometimes passively and sometimes forcefully, and were probably generated as immiscible liquids during cooling of their host plutons. The similarity of the ore deposits in the rocks of different ages suggests a common petrochemical process for their formation.

It is proposed that the metasomatic and magmatic hydrothermal mineralisations are the inevitable products of magmatic activity, which, in turn, is the inevitable product of the geotectonic evolution of the area. With one notable exception, the Ni-Co arsenide deposits of the world may have been formed as a consequence of similar geotectonic processes.

# P A R T  I

# P E T R O G E N E S I S

# 1. INTRODUCTION

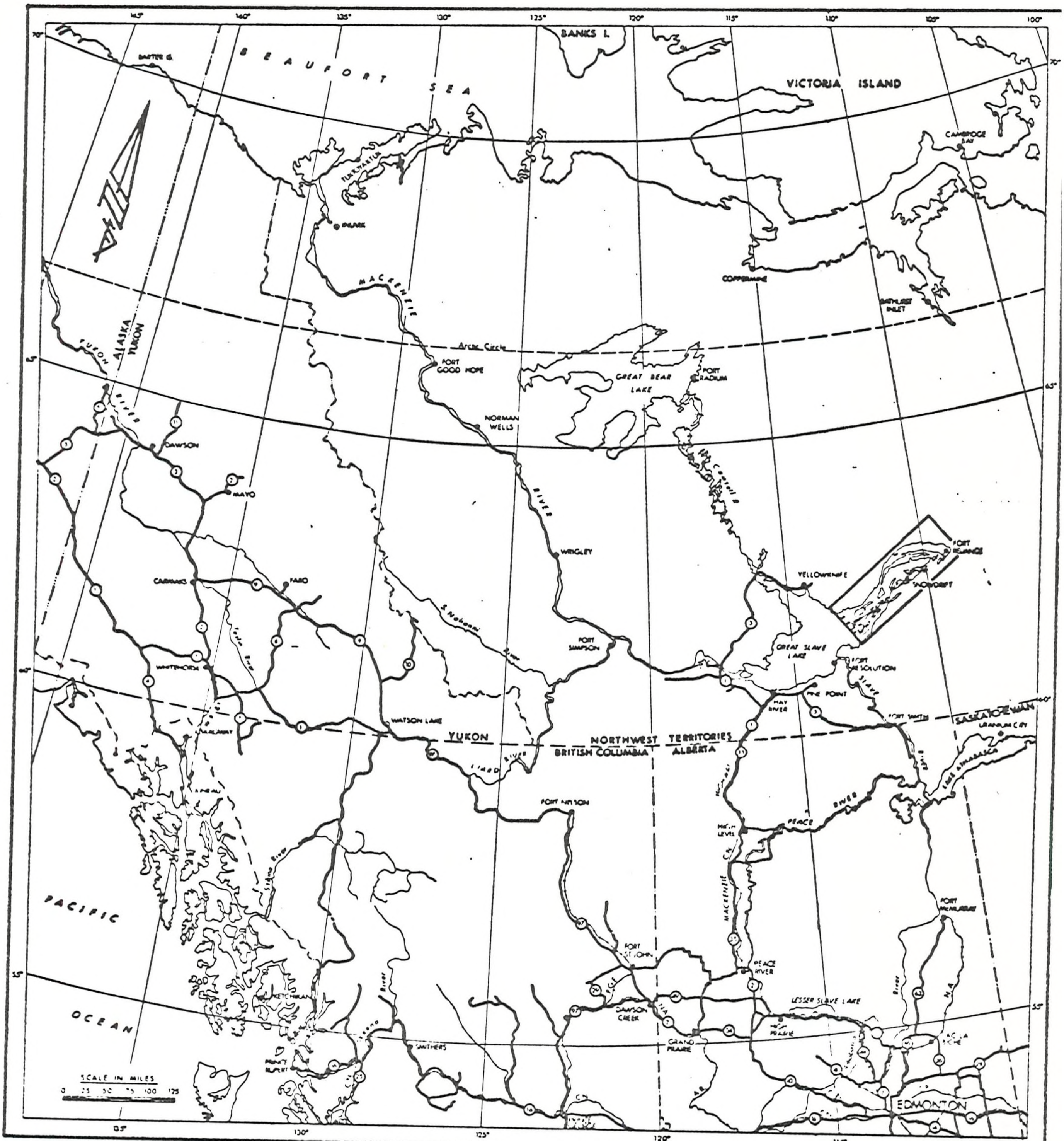
Great Slave Lake is situated in between two other great lakes that is Great Bear Lake to the north and Lake Athabasca to the south. It straddles the presently exposed margins of the Canadian craton Most of the lake is underlain by Paleozoic strata, but Aphebian (2200-1700 m.y.) and younger Precambrian rocks outcrop on the eastern shore and islands particularly in the East Arm. It is the Aphebian rocks that this thesis is concerned with.

In the early 1900's not much interest was concentrated around Great Bear Lake neither around Great Slave Lake. The pioneer workers, especially prospectors, were returning from the Klondike and looking for gold although only silver ore was discovered. In the 1930's gold was found at Yellowknife and interest shifted to Great Slave Lake. The first reconnaissance survey in the Great Slave Lake region was by Bell (1902). Since then not much work was done, until in late twenties; in the thirties the East Arm region was investigated in some detail particularly by Stockwell (1933 and 1936). This was the result of a gold strike which brought many of the prospectors from Great Bear Lake.

The regional geology of the East Arm was only very generally understood and the only mapping that covered most of the area was that of Stockwell (1936) which was carried out from 1929 to 1931. The first systematic report of the geology of the East Arm was made by Stockwell in 1933, after Lausen (1929) had made the first attempt at systematic stratigraphy of the area. The stratigraphic nomenclature used now is basically that of Stockwell revised and amplified by Hoffman (1968b). The stratigraphic work of Hoffman (1968, 1969, 1970, 1973) and Hoffman et al. (1970, 1973) has contributed greatly to the understanding of the geologic development of the East Arm as a whole.

Contributions primarily concerned with the geology of the East Arm are listed below:

1) Bell (1902): Reconnaissance regional geology (report on exploration in the Great Slave Lake region).
2) Rutherford (1929): Description of stromatolitic dolomites in the East Arm.
3) Lausen (1929): First attempt at systematic stratigraphy.




Map 1: Location of the study area.

4) Stockwell (1933): First systematic report of the geology of the East Arm.

5) Stockwell (1936): First geological map of the East Arm (G.S.C. maps 377A and 378A; 1 inch to 4 miles).

6) Brown (1950a): Placed Stockwell's geologic mapping of the western half of the East Arm on an adequate topographic base.

7) Brown (1950b): Modified Stockwell's geology of the Reliance area.

8) Brown (1950c): Modified Stockwell's geology of the Christie Bay area.

9) Wright (1951): Further modified the geology of the Christie Bay map area.

10) Wright (1952): Further modified the geology of the Reliance map area.

11) Barnes (1951): Mapped Snowdrift area at 1 inch to the mile.

12) Barnes (1952): Mapped McLean Bay area at 1 inch to the mile.

13) Stockwell et al. (1968): Compiled final maps of the Christie Bay and Reliance areas at 1 inch to 4 miles.

14) Hoffman (1968): Detailed stratigraphic analysis and reclassification of the Great Slave Supergroup.

15) Hoffman (1969): Sedimentological analysis and paleogeographic reconstruction of the Great Slave Supergroup and Et-Then Group.

16) Reinhardt (1969b): Map modifications and general structure in the Wilson Island - Petitot Islands area at 1 inch to 4 miles.

17) Hoffman et al. (1979): Made stratigraphic and paleogeographic correlations among the Great Slave Supergroup - Et-Then Group and other areas of proterozoic sediments around the Slave Province.

18) Reinhardt (1972): General description of exotic breccia occurrences in the area of Simpson Islands and Wilson Islands.

19) Hoffman (1973): Synthesis of the early Proterozoic stratigraphic and tectonic evolution of the East Arm area and the western margin of the Slave Craton.

20) Hoffman et al. (1973): Descriptions of almost complete geology of the East Arm area.

The object of this thesis is the study of the hydrothermal mineralisations and the associated host rocks (mainly the Aphebian magmatic rocks) in the East Arm. Unlike other fields of geology little has been written in much detail of the economic geology of the area. Although some of the deposits in the East Arm area were discovered in the forties, (e.g. those at Sachowia Lake and Regina Bay), no detailed mineralogy was documented. However, a brief report on the occurrence of the Ni-Co-Bi mineralisation at Blanchet Island was made by Mason (1969) after its discovery in 1968. Walker (1970) has reported on the mineralisation along Vestor Channel for Vestor Explorations Ltd. In 1972, following the study of granite in the Slave Province, Davidson reported on some U-Th-REE mineralisations in the Blachford Lake Complex. The sandstone-type uranium deposits in the Proterozoic strata were described by Morton (1974). In 1975, Stanworth and Badham carried out investigations of the mineral deposits in the East Arm for Rio Algom Ltd. Subsequently Badham (1976 and In Press) has documented the relationship between the diorites and their associated Mnt-Ap-Act-U, Ag, Ni-Co-As mineralisations in the East Arm.

It is important to note that this study is based on the fieldwork of Badham and Stanworth in 1971 and 1975, and that the author has not visited the area. The work on the petrography and geochemistry of the host rocks (Part I) and on the petrography of the ore deposits (Part II) was carried out on a comprehensive suite of samples collected by Badham. Here a unified concept for the geologic and metallogenic evolution of the East Arm area is proposed.

## 2. PHYSIOGRAPHY AND SURFICIAL GEOLOGY

The main body of Great Slave Lake crosses the boundary between the Canadian Shield and the bordering area of paleozoic rocks, and the east arm of the lake extends at right angles to the contact for 175 miles into the Shield (Map 1). The East Arm basin owes its existence to deep erosion of a belt of mixed hard and soft rocks that is bordered on three sides by uniformly more resistant formations, generally granitic. The topography of the East Arm as a whole contrasts with the surrounding areas of the Canadian Shield by varying widely, commonly with local reliefs of several hundred feet with cliffs or steep slopes. Ridges, elongate in a northeast - southwest direction, parallel to the structural grain, are prominent. Diorite intrusions are usually cliff-forming with relief of about 400 feet. Although the East Arm is generally a low lying area, numerous fault-line scarps are present, the highest of which rises up to 350 feet above the lake level.

The surrounding granitic upland of the Canadian Shield presents a rather monotonous plateau on which low, rounded hills rarely exceed 100 feet. Within the western half of the East Arm, the general level of the bordering lands and of numerous islands and peninsulas in the lake rises gradually from near lake level at the entrance to the arm to 450 to 550 feet above the lake in the east part of the area, with the result that the country becomes progressively more rugged and picturesque towards the east. Within the eastern half of the East Arm, the general level rises gradually from 450 feet above the lake in the southwest corner of the area to 700 feet in the southeast part and 900 feet a few miles inland from the north shore of the lake. The main lake itself is generally about 500 feet above sea-level and its water is rather shallow, about 100 to 300 feet deep over the paleozoic unconformity. The East Arm is generally deeper but variable: at Christie Bay the water is over 2000 feet deep.

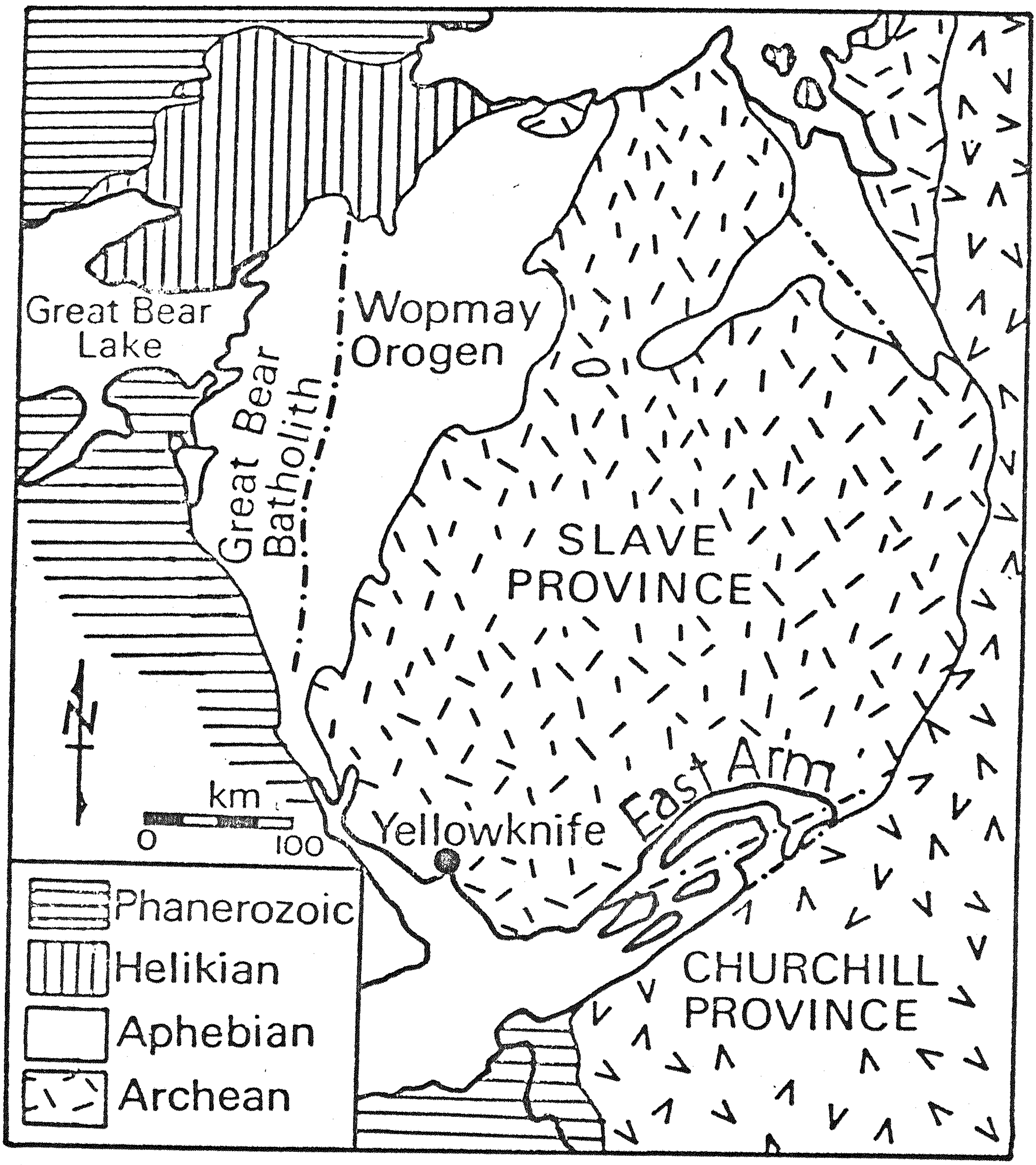
The East Arm area is also covered with pleistocene and recent deposits with outcrop along ridges (along faults) and in lake shore areas. The cover is predominantly sandy till and muskeg which in most instances likely developed on till. Glacial boulders are scattered over much of the East Arm but thick deposits of moraine are rare. Since the retreat of the continental ice sheet, large deltas have formed at the mouths of rivers flowing into the southwest corner of the arm.

The deltas, sheltered lowlands and some of the islands are locally well-timbered but many of the rocky slopes and hill-tops are sparsely wooded. The drainage is disrupted and run off is limited to the spring due to a dry climate of the area. Roche moutonnée topography is prevalent and, together with glacial strive, indicates pleistocene ice movement was towards the southwest, down the axis of the East Arm. However all rocks are generally well-exposed, particularly on shores where they are glacially-polished and kept free of lichen and vegetation by water and winter-ice.

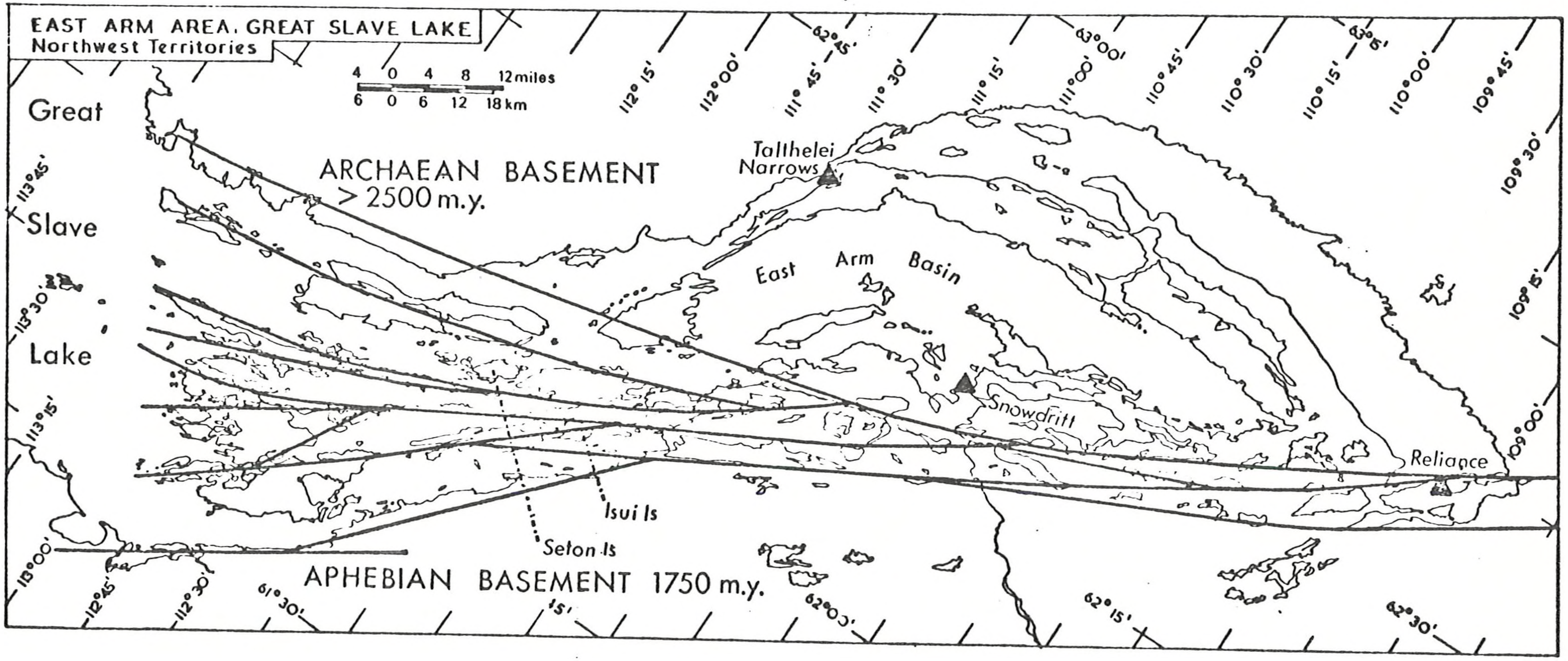
## 3. REGIONAL GEOLOGY

The East Arm of Great Slave Lake contains a great thickness of Lower Proterozoic (Aphebian: 2.5 to 1.7 b.y.) and Middle Proterozoic (Helikian: 1.7 to 1.2 b.y.) sedimentary and igneous rocks that are well-preserved and unmetamorphosed although mildly deformed. These rocks were deposited and emplaced in a tectonic basin produced by the MacDonald fault system (Map 3). It is a complex graben which is now defined by the ramifying network of faults of this system. It lies between the Archean ( 2.5 b.y.) Slave craton and the Churchill province and represents a complex area of Aphebian sedimentary and igneous rocks and Archean basement stabilised by the Hudsonian (about 1.7 b.y.) orogenic event (Map 2). It has been argued that this faulting was originally typical of a rifted graben (aulacogen) and subsequently became transcurrent (Hoffman, 1973). On the other hand Badham (1978) argued that the system has always been essentially transcurrent (a right-lateral strike slip system) and active intermittently throughout the Aphebian. However the fault movements controlled both the deposition and preservation of sedimentary rocks and became the loci of igneous activity. The nature and history of the sedimentary rocks have been well-documented by Hoffman (1968, 1969, 1973) and his stratigraphy is summarised in Table 1 after some minor modification by Badham (1975).

Within the various sedimentary sequences there is a wide chemical and temporal range of magmatic rocks. Volcanic rocks are found in all groups of the East Arm but the Seton Volcanics represent a period of particularly widespread and active magmatism. Intrusive rocks are less common, but there were two important periods of intrusions, that is at the beginning and end of Great Slave Supergroup times. The earliest intrusions consist of two alkaline complexes (the Maple-Blachford Lake complex, of which Odin Dyke is a part, and the Easter Island Dyke) and three probably related lines of diatreme breccia pipes in Inconnu, Vestor and Preble Channels (Reinhardt, 1972) (Map 5). The petrochemistry and field relationship of these rocks are reported in more detail elsewhere but are summarised here in Table 2. These intrusions appear to be between 2200 m.y. and 2050 m.y. old and are confined to the western end of the East Arm (Map 4). The igneous complexes are alkaline




**Map 2:** Tectonic subdivisions of the NW Canadian Shield – after Fraser et al., 1972.




Map 3: The MacDonald Fault System.

and locally peralkaline (Davidson,1972) and rare igneous matrices to the breccia pipes have affinities with late differentiates of the complexes. The Easter Island Dyke has been dated at 2200 m.y. and 2170 m.y. (G.S.C. 62-93, and Burwash and Baadsgaard, 1962). The dyke intrudes the Archean basement. Its relationships to the Sosan Group and to the breccia pipes are less clear and at present there exists some controversy. Badham (1977 and In Press) suggested that the dyke intrudes the lower Sosan Group and is cogenetic with the Vestor Channel breccia pipes. The peralkaline granite in the Blachford Lake Complex has been dated at 2057 ± 56 m.y. (Wanless in Davidson, 1978). The rocks of this complex cut Archean basement and the more mafic, alkaline rocks of the closely related Maple Complex (now part of the Blachford Lake Complex and named as Caribou Lake Gabbro in Davidson, 1978) which also cuts the Archean basement. Therefore it is probable that the Easter Island Dyke and the Maple-Blachford Lake Complex are effectively contemporaneous.

The younger intrusions are stocks and laccoliths of calc-alkaline diorites and were emplaced throughout the East Arm (Map 4), at the end of Great Slave Supergroup times. They outcrop intermittently from the Îles Basses in the west to Meridian Lake in the east (a distance of some 200 km) and are essentially confined to the line that acted as a boundary between the shelf and basin facies in the Upper Great Slave Supergroup times (Hoffman, 1969, 1973). They have been dated variously at 1845, 1795, 1785 and 1630 m.y. (Hoffman, 1969). While these are K-Ar ages and are not thoroughly satisfactory, an age of 1790 m.y. is generally accepted and best agrees with paleomagnetic data (Irving and McGlynn,1976).

Finally, hypabyssal magmatic rocks are common but formed a minor constituent of the areas of greatest volcanism.

Closely related to these various magmatic rocks are the mineralisations, firstly of magnetite-apatite-amphibole (actinolite) in pipes and pegmatites, secondly of uranium-thorium-rare earth element minerals in pipes, veins and pegmatites, and thirdly of Ag, Bi, Ni-Co-Fe arsenides, sulpharsenides and sulphosalts in veins and skarns. It has been observed that the older intrusions have in places been tilted and eroded, such that both top and bottom are exposed. The younger intrusions were emplaced at shallow depths and the tops are preserved. The model for genesis and deposition of the mineralisations can be derived from the study of these intrusions. Two occurrences of arsenide mineralisations are apparently not associated with any intrusion. However the mineralisation is so closely similar to that clearly related to the intrusions that we can deduce the presence of unexposed igneous rocks. Volcanic pipes and hypabyssal plutons are closely related to the alkaline Seton Volcanic Formation. Two such pipes have been found to be mineralised and the mineralisation has close affinities with that related to the major intrusions. The mineralisation in the volcanic rocks is thought to be a different manifestation of the same petrochemical process that operated on the major intrusions.

## 4. DETAILED GEOLOGY OF THE EAST ARM AREA

### a) GENERAL

As already mentioned, the East Arm is a westward-opening graben controlled by the splaying branches of the MacDonald fault system. The parallel basement mylonite zone, at least 300 miles (480 km) long, indicates an earlier history of transcurrent movements. A thick sequence of Aphebian and Helikian sediments overlie the basement. The evolution of the East Arm graben can be divided into three stages (Stockwell, 1936 a, b; Hoffman, 1969, 1973) in which three groups of sedimentary and igneous rocks were deposited and partially eroded. All three groups line unconformably not only on Archean basement but also on older groups. Badham (1978) has suggested that the deposition and preservation of all these strata were essentially controlled by strike slip (with or without concomitant vertical) movements on various splays of the fault system.

The earliest deposited group, the Wilson Island Group, was deformed and partially eroded before the emplacement of a series of alkaline igneous complexes (Davidson, 1972, 1978; Badham, 1978). The main thickness of strata, the Great Slave Supergroup, was mildly deformed, faulted, partially eroded and intruded by alkaline intrusions in the early Aphebian and by dioritic rocks towards the end of the Aphebian. The youngest group, the Et-Then Group, is the post-Aphebian proximal erosional debris of the Churchill Province and of horst-blocks of earlier East Arm strata, and is essentially undeformed. The Et-Then Group unconformably overlies the dioritic intrusions.

b. LITHOLOGIES

In this section a rather brief description of sedimentary lithologies of the East Arm is presented. The nature and history of the sedimentary rocks have been described in considerable detail by Hoffman (1968, 1969, 1973). Some minor modifications have been made by Badham (1975) and are summarised in Table 1. However the detailed petrography, petrochemistry and petrogenesis of the magmatic rocks are presented in the later part of this chapter.

Basement rocks to the north of East Arm are Archean, predominantly at greenschist facies, and these dip beneath the East Arm where inliers can be recognised, generally at slightly higher grade and often weathered, in horst-blocks. The mylonites on NE-trending faults on the south shores of Great Slave Lake imply that the Archean and the Wilson Island Group predate the Great Slave Supergroup. On the other hand the initiations of the MacDonald fault system predated the Wilson Island Group by the evidence of pre-Wilson Island Group deformation in the Archean which post-dates diabase dykes which are, in the Slave Craton, younger than the Kenoran orogeny (about 2.5 b.y.).

The Wilson Island Group is now preserved mainly on Wilson Island, around Basile Bay and on the south shore between MacDonald Lake and Preble Island. It consists of a minimum of 4835 feet and a maximum of 10,000 feet of coarse clastic sediments, dolomites, volcanics and argillites. The coarse clastics include conglomerates, fluviatiles and beach sands while dolomites are stromatolitic with indications of intertidal deposition. The volcanics include tuffs, flows, pillow lavas and iron formation, and are predominantly mafic-intermediate. The argillites are fine, green, banded and monotonous and seem to be a distal turbidites succession. The paleocurrent data indicate NE to SW transport in the Wilson Island area. There is no clear direction indicated from elsewhere. On Basile Peninsula the basal white quartzite with local conglomerates, lies unconformably on a rotted gneissose granite. Elsewhere the base is not seen. Before the deposition of the Union Island Group, the Wilson Island Group was metamorphosed to greenshist facies and fairly tightly folded. The age of the Wilson Island Group is not known. It must post-date 2390 m.y.

| ERA | GROUP | FORMATIONS | IGNEOUS ROCKS | AGE m.y. | REFERENCES |
|---|---|---|---|---|---|
| HELIKIAN | | | Diabase Dikes (mcKenzie) | 1315<br>1295,1250 | Reinhardt, 1969<br>Hoffman, 1968 |
| | | | Diabase Sills | - | - |
| | Et-Then | Preble<br>Murky | | | |
| APHEBIAN | | | C.Calc-alkaline diorites | 1630,1785,<br>1795,1845 | Hoffman, 1968 |
| GREAT SLAVE SUPERGROUP | Christie Bay | Pearson<br>Portage Inlet<br>Tochatwi<br>Stark<br>Hearne<br>Wildbread<br>Pekanatui Point | Basalt. | - | - |
| | Pethei | Blanchet<br>McLean<br>Utsingi<br>Talthelei<br>Douglas Peninsula<br>Charlton Bay | | | |
| | Kahochella | McLeod Bay<br>Gibralter<br>Seton | B.Seton Volcanics<br>Sodic basalt<br>and andesites | 1872 | Baadsgaard et al., 1973<br>Olade and<br>Morton, 1972 |
| | Upper Sosan | Akaitcho River<br>Kluziai<br>Duhamel<br>Hornby Channel | A. Alkaline<br>intrusions and<br>breccia pipes | 2170<br>2200<br><br>2057 | G.S.C. 62-93<br>Burwash and Baadsgaard, 1962<br>Baadsgaard, 1962<br>Wanless, in Davidson, 1978 |
| | Lower Sosan | Susanne | Tuffs | | |
| | Union Island | | Alkali-Continental tholeiite basalts and sills | | Goff and Scarfe, 1976 |
| | | | Rhyolite and tuff | | |
| | Wilson Island | | Basalt,trachyte,rhyolite. | | Stockwell, 1976 |
| ARCHEAN BASEMENT | | | | 2460 and older | Reinhardt, 1969 |

Table 1: Stratigraphic Column after Hoffman, 1968, and from Stanworth and Badham, 1975.

and must predate approximately 2000 m.y., the minimum age at which the Union Island and Susanne (Lower Sosan) Groups were deposited.

The Union Island Group outcrops in only a small area on Union Island and adjacent islands. It has nowhere been found unconformable on Wilson Island Group, yet it is everywhere unconformable on basement. Thus the Wilson Island Group rocks must have been eroded (from horst-blocks) from those areas where the Union Island Group is now preserved (as graben). The Union Island Group is older than the Great Slave Supergroup and separated from it by an unconformity. It consists of black and grey shale/slate, pillowed basalt, dolomite and greywacke. The details of these units at MacDonald Lake have been described by Hoffman (1968). Cherts are common in all the dolomites and many of them are volcanogenic rather than sedimentary. The Union Island Group is nowhere seen to underlie the Sosan Group unconformably, but always in fault contact. Nevertheless it must do so firstly as a geometric necessity; secondly as the Sosan Group clastics contain clasts of Union Island lithologies, and thirdly because a large clast of Union Island black shale occurs in a breccia pipe cutting the Sosan Group at Inconnu Channel. Before the deposition of the Great Slave Supergroup, the Union Island Group was gently folded, uplifted and partially eroded.

The Great Slave Supergroup has been divided by Stockwell (1936) into six formations, the basal unit of which rests on a deeply eroded surface. It is about 20,000 feet thick and the strata occur in an easterly trending synclinorium. Along the north flank the strata dip very gently south but are folded in the central and southern parts. The rocks are cut by sills, laccoliths and possibly also stocks of hornblende-biotite diorite or quartz diorite. A biotite from one of these intrusions yielded a K-Ar age of 1845 m.y. which indicates that the rocks are Aphebian and were emplaced during an early phase of the Hudsonian orogeny. The intermittent vertical movements on a pre-existing fault system are the fundamental control of all East Arm sedimentation and deformation. The faults that were clearly in existence in pre-Sosan times splayed out to the west from an area around Snowdrift. Most of the major splay-faults were active at the end of the Union Island Group times.

The basal Sosan Group is from 3000 to 5000 feet thick and composed predominantly of feldspathic quartzite: it includes a thin basal stromatolitic dolomite member, and, near the top, red or pink siltstone and shale. By the paleogeographic interpretations deduced from sedimentary facies and paleocurrent directions in the various 'Hornby Channel' Formations, the 'Hornby Channel' Formation in the Simpson Islands area is thought not to be equivalent in age to that in the Snowdrift and Reliance areas. Badham and Stanworth (1975) concluded that the Sosan Group west of Basile Bay is older and have renamed it the Lower Sosan Group (Stanworth, In Prep.). The Hornby Channel Formation overlies basement unconformably in the Snowdrift and Reliance areas. At Snowdrift the Hornby Channel Formation is overlain by Duhamel dolomites. This thins to the NE and also change facies into clastic sediments. The Kluziai sandstones overlie Duhamel dolomites conformably at Snowdrift and thin to the NE where the facies become less mature. They overlie fluviatile sandstones at Reliance where no dolomite was deposited.

In all areas east of Basile Bay the lower Sosan Group formations are overlain conformably by the Akaitcho River Formation (a homogeneous red siltstone and fine sandstone). These lithologies are generally porous and permeable.

The overlying Kahochella Group is as much as 7000 feet thick and consists of red shale, siltstone, arkose, carbonate, and a few beds of oolitic iron formation. In the central part of the basin, andesite flows and pyroclastics with inter-calated red sandstone and shale occur within the sequence. The Pethei Formation comprises about 1,500 feet of limestone and dolomite with abundant stromatolitic zones along the north limb of the synclinorium. Lateral equivalents in the southern part of the basin are fine-grained limestone and interbedded limestone and argillite and in the southwest are graded beds of greywacke and shale with interbeds of chert and limestone. Conformably overlying the Pethei are red mudstone, dolomite and limestone forming the Stark Formation. Breccias, probably of different origins, are common in the carbonate rocks. The Tochatwi Formation comprises about 3,000 feet of non-marine, mollasse-type red sandstone and shale. The youngest rocks of the Great Slave Supergroup form the Pearson Formation which consists of 500 feet of basalt, commonly vesicular and amygdaloidal, with related tuff and thin beds of argillite. Orientations of cross-bedding and other current structures suggest a source area to the west and north of the present basin for most clastic sediments. The sequence includes near-shore sediments deposited in fluvial non-marine, shallow marine and deep water marine environments. The youngest reddish sediments were deposited in a shallow water, near shore, partly marine environment and were subject to subaerial exposure.

The Great Slave Supergroup is overlain unconformably by the Et-Then Group, which rests on truncated folds in Great Slave strata and is later than the diorites that intrude the older rocks.

The Et-Then Group is divided into two formations, the oldest of which, the Murky, is composed of several thousands of feet of conglomerate and minor arkose and shale. The sequence is made up of cyclical alternations of course boulder-conglomerate that grade upwards, with decreasing pebble-size into thin, commonly lenticular arkose and rare shale beds. The cycles vary in thickness up to 50 feet. Near MacDonald fault, granitic rocks and granite-gneiss predominate, whereas to the north, clasts of the Great Slave Supergroup are most abundant. The conglomerate may thin to the north and the clasts appear to decrease in size in the same direction suggesting a source to the south. These strata pass through a transition zone in which the amount of sandstone increases into the Preble Formation which consists of at least 4,000 feet thick of buff, pink, or red lithic sandstone, and feldspathic quartzites with intercalated thin beds of pebble-conglomerate, shale-chip conglomerate, and red shale. Cross-bedding and thin laminations are common and mudcracks are found throughout the sequence. The quartzites consist of rectangular to subrounded, well-sorted grains of quartz, rock fragments, and feldspar in a carbonate cement.

Et-Then Group rocks are cut by diabase sills and dykes. The dykes are part of the north-northwesterly striking Mackenzie swarm that yield K-Ar whole rock dates of between 1215 and 1555 m.y. (Fahrig et al., 1963), but the sills which are cut by these dykes are not dated. Et-Then Group rocks are gently folded about easterly trending axes and, along with older rocks, are cut by the northeast trending MacDonald fault and related parallel faults. Most displacement of rocks appears to have been completed before the intrusion of the Mackenzie diabase dykes and most probably began during the late stages of the Hudsonian orogeny.

# 5. GEOLOGY AND PETROGENESIS OF THE MAGMATIC ROCKS

## General

Intrusive igneous rocks of various Aphebian ages have been recognised throughout the East Arm and their distribution is shown on Map 4. These alkaline rocks were formed at three separate periods, not two as was previously thought. Chemical and geochronologic data on these rocks are now available and substantiate the division. A new complex (The Maple-Blachford Lake Complex) has been located and documented (Davidson, 1978). A series of diatreme breccia pipes of hitherto unknown age and origin has been correlated with the earliest magmatic phase. The older intrusions have suffered greater tilting and uplift and are exposed at different levels but generally at deeper levels. Nevertheless a close genetic relationship between all these intrusions is obvious. The Maple-Blachford Lake intrusion is thought to be about the same age as the Easter Island Dyke and the closely related breccia pipes. The younger calc-alkaline diorites were probably emplaced just after the final deposition of Great Slave Supergroup rocks (Pearson Formation) but clearly long before the Et-Then Group.

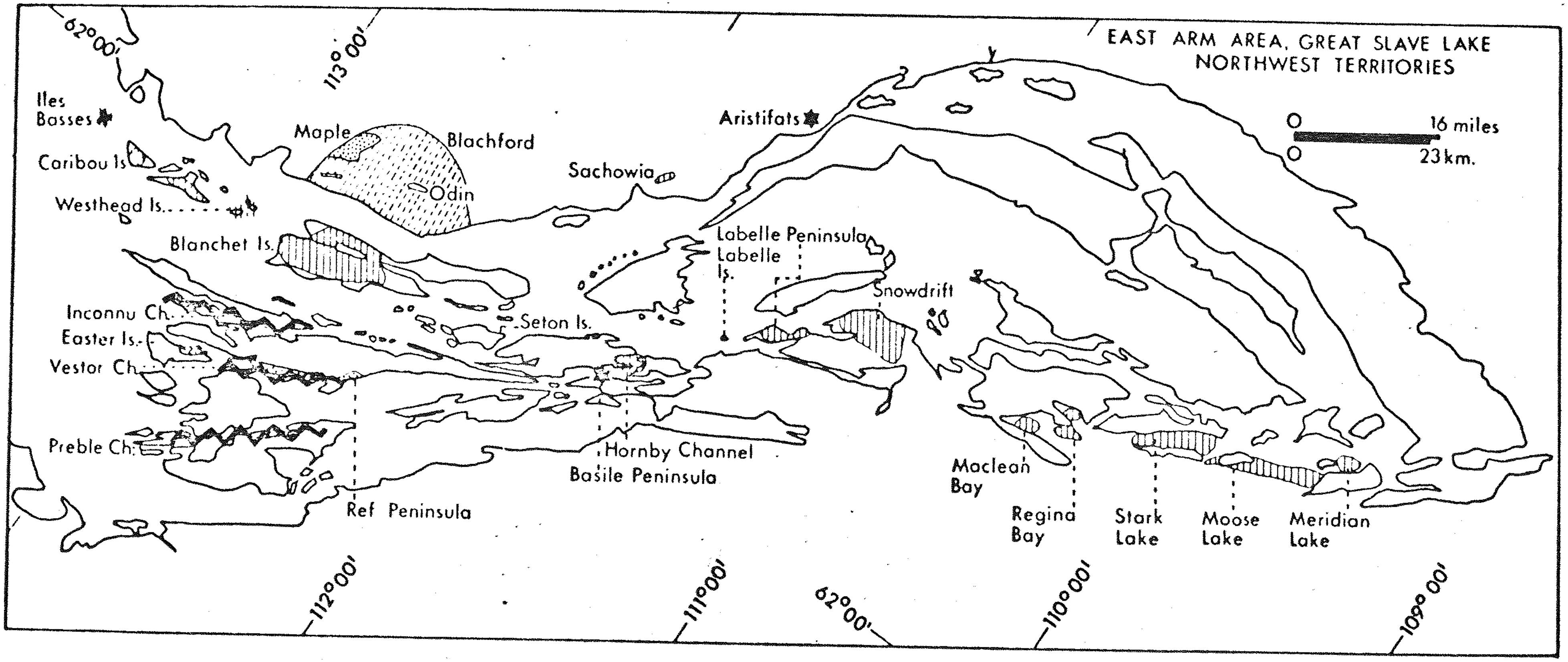
The three groups of igneous rocks, labelled A, B and C in Table 1 are summarised as the following:

Group A: This Lower Aphebian group includes the Easter Island Dyke, the newly discovered Maple-Blachford Lake complex, and lines of diatreme breccia pipes of Vestor Channel, Preble Channel and Inconnu Channel.

Group B: This middle Aphebian group consists of the Seton Volcanic Formation which is made up of flows, tuffs and hypabyssal necks of alkali basalt with minor felsic differentiates.

Group C: This group includes all the late Aphebian diorites such as the Îles Basses, Caribou Islands, Westhead Island and Blanchet Island, Stark Lake, Moose Lake and Meridian Lake laccoliths and the Hornby Channel, Labelle, Snowdrift, MacLean Bay, Regina Bay stocks or plugs.

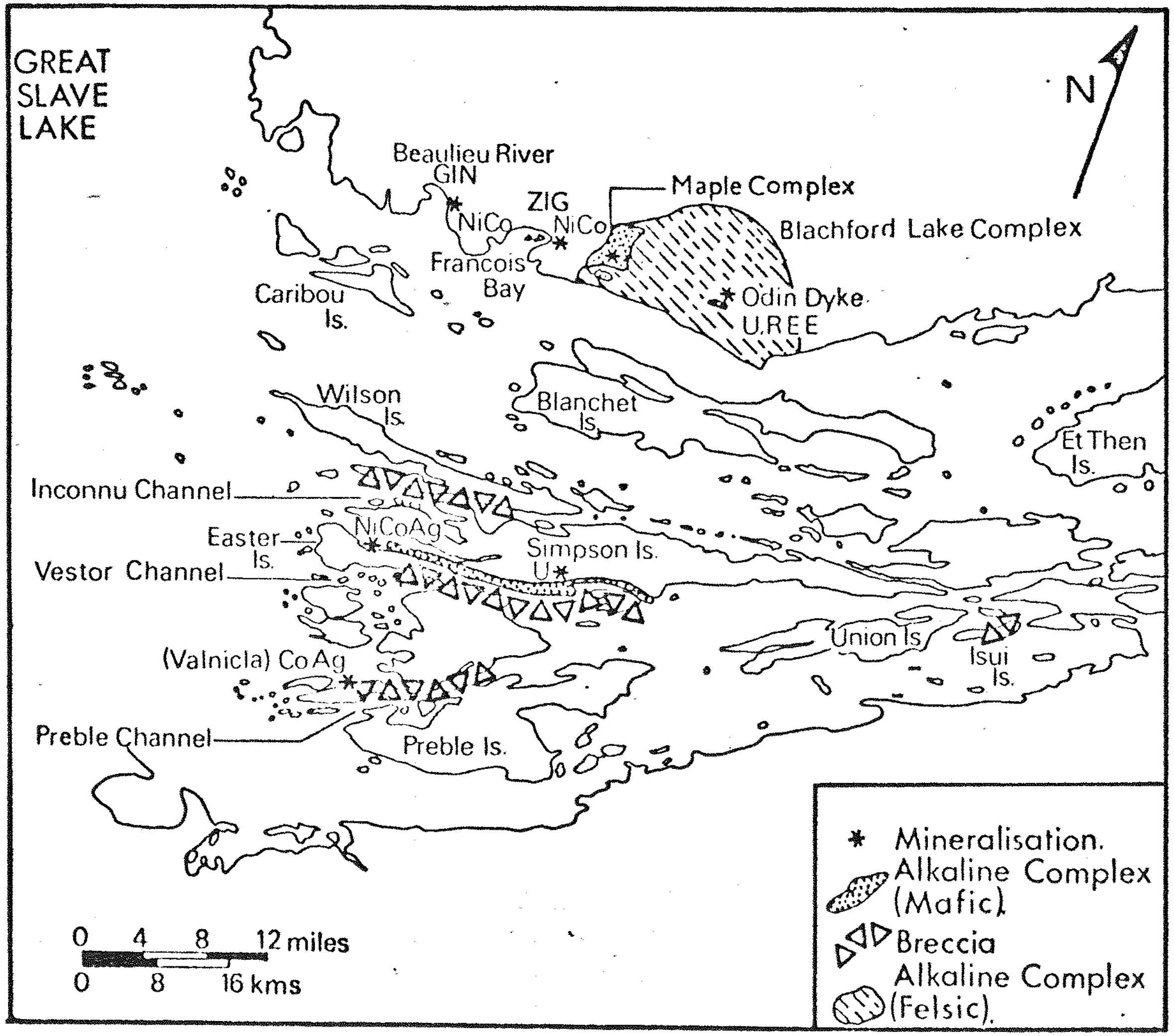
All these magmatic phenomena lie in major splays of the MacDonald fault system. These faults were clearly active before, during and after the emplacement of the magmatic rocks. As already mentioned these faults not only controlled the timing and nature of magmatism, but also




Map 4: The East Arm area - Geology.

the deposition, deformation and preservation of the various lower Aphebian sedimentary sequences. It is understood that alkaline magmatism is characteristic of deep continental fracture zones, whether they be graben (such as the East Arm), faulted monoclines, belts of block faulting or transcurrent faults (Bailey, 1974). In such zones the location of magmatism is controlled by faults and the timing is dependant on activity on those faults. Thus temporal and spatial data from alkaline rocks may contribute significantly towards an understanding of the geotectonic nature and history of the fracture zone.

Finally all these intrusions are closely associated with uranium-thorium-rare earth elements or Ag, Bi, Ni-Co-Fe arsenide mineralisations. Similar mineralisations are also found elsewhere on the same major lineaments that control the intrusions. While some of these mineralisations are clearly related to the late Aphebian magmatism (Badham, In Press) it is felt that the others indicate as yet unexposed lower Aphebian alkaline complexes. The details of the mineralisations are reported in Part II of this thesis.




Map 5: The western part of East Arm area - Geology

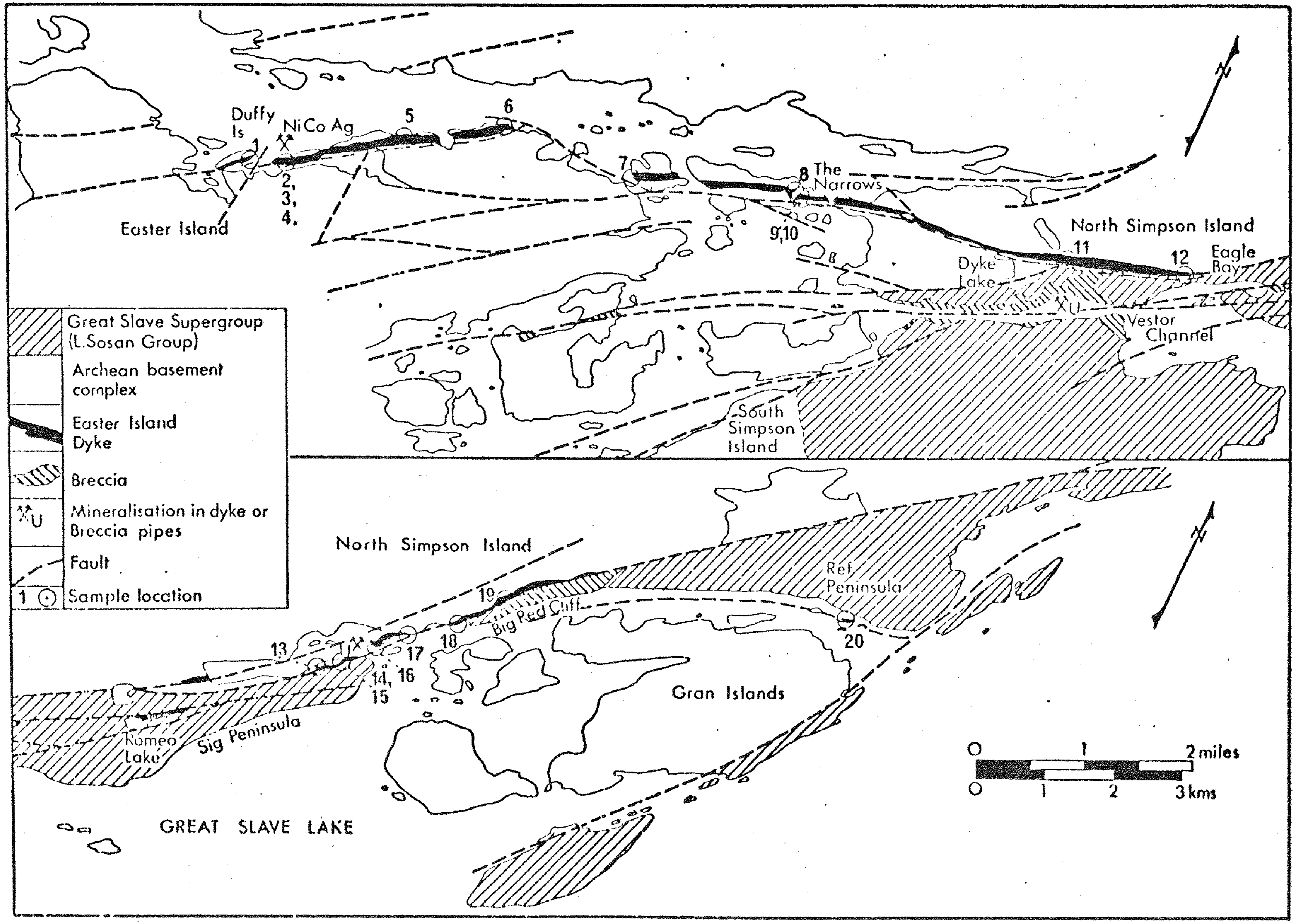
## Group A. THE LOWER APHEBIAN MAGMATIC ROCKS

### a. The Easter Island Dyke

#### i. General

This remarkable dyke is exposed from Easter Island through Vestor Channel and out to Ref Peninsula. It has now been mapped over a distance of 29 km with an average width of 50 metres (Map 6). The dyke is vertically differentiated and has been rotated by faulting, approximately 10°, such that successively deeper levels are generally exposed to the southwest. It is thought that a vertical section of between 1 and 2 km is now exposed. The roof is exposed intermittently over the easternmost 12 km and can be seen in a number of places in Sig Peninsula, and as far as Ref Peninsula.

The intrusion is differentiated from an ultramafic rock through gabbro to an alkali syenite main phase. In its intermediate portions it is polyphase and contain chilled margins of a 'brick-red trachytic phase' of bostonites. The dyke contains no xenoliths in its lower portions, but has sandstone and gneissic xenoliths in a multiply-intruded matrix of chilled albite syenite in the top. The westernmost outcrops consist of ultramafic cumulates, but to the northeast the rock varies from pyroxenic alkali syenite with olivine-sulphide-rich margins on Easter Island to predominantly albite syenite with subsidiary biotite and amphibole (predominantly hornblende and some actinolite) at Dyke Lake. The olivine-sulphide-rich margins, now altered to a sulphide-talc (serpentine) - carbonate assemblage, thin out eastward from a maximum of 3 metres; that on the north disappears 3 km before that on the south. Alteration and gas-brecciation is common and the degree of alteration decreases downward. It must be emphasised that in its western half (i.e. lower part of the dyke) only the margins of the dyke are altered. In contrast the eastern (upper) parts are pervasively hydrothermally altered. The rock originally consisted of variable syenite which was emplaced in progressively more differentiated pulses. It was proposed by Badham (1975) that the brick-red vesicular albite trachytes (bostonites) were formed from the latest magma. Small gas-brecciation pipes are common, especially near Dyke Lake, Big Red Cliff and Ref Peninsula. They are




Map 6: The Easter Island Dyke - Geology and Mineralisation.

intimately related to the roof zone of the dyke and intrude Sosan Group sandstone. These pipes with their bostonite matrices, are similar to the diatreme breccia pipes in the Vestor Channel area. However near Vestor Channel the dyke appears to have been the parent to the breccia pipes and bostonites (Badham, 1975). The dyke is intermittently spatially coincident with the line of breccia pipes between Vestor Channel and Ref Peninsula.

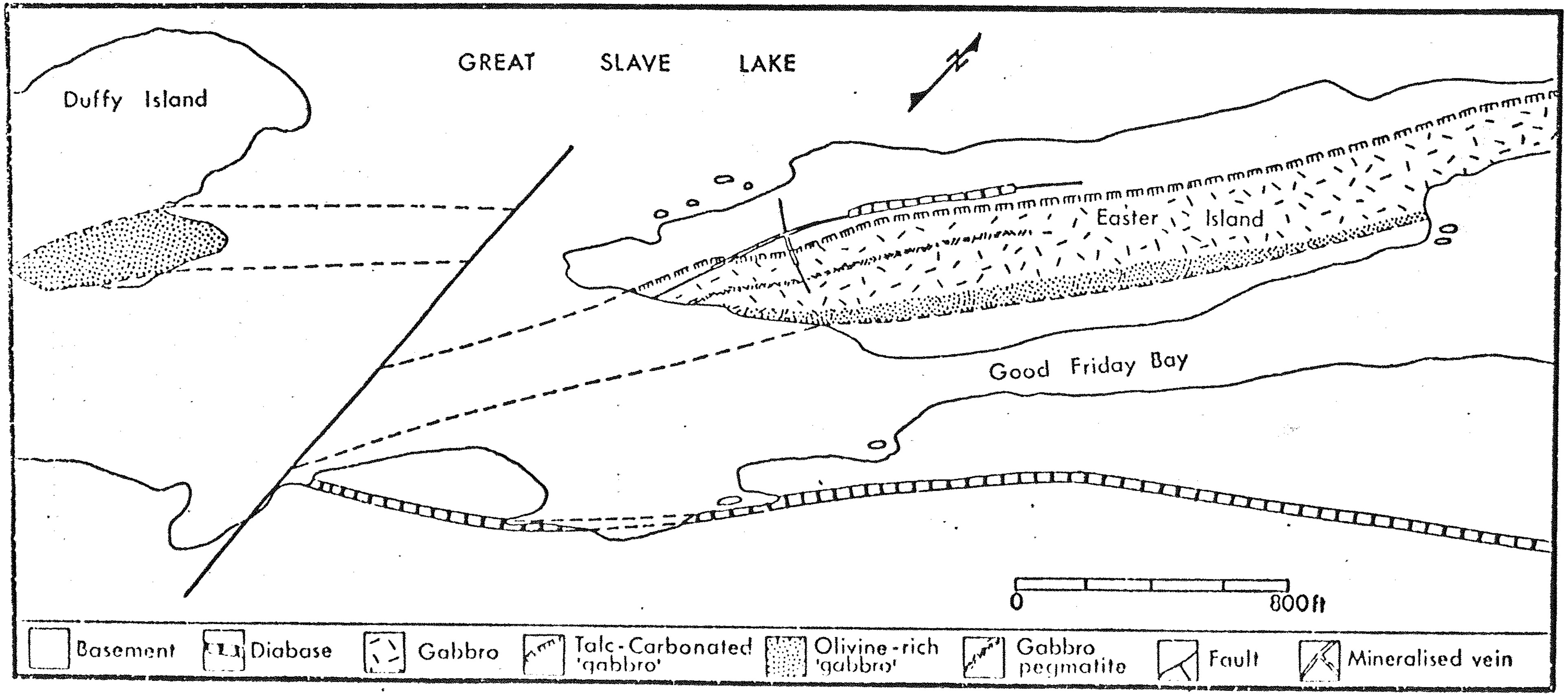
The alteration of the top is thought to be autohydrothermal and the breccia pipes probably represent the escape zones of spent fluids and gases along with the last remnants of silicate liquids. The roof of the dyke is close to the unconformity of Sosan Group on Archean basement and most of brecciation and veining occur at this unconformity.

It is important that the dyke intrudes not only Archean basement but also the Lower Sosan (Susanne) Group, which thus provides a minimum age for the intrusion. As was mentioned earlier the K-Ar ages of 2200 m.y. and 2170 m.y. age may be spurious because of the prevalent alkali metasomatism.

ii. Petrography/Petrology

The samples studied were collected from various places along the entire exposed length of the dyke and in places across its width. This includes the samples from Duffy Island, Easter Island, The Narrows, Sig Peninsula, Romeo Lake, Eagle Bay, Dyke Lake, Big Red Cliff and Ref Peninsula (Map 6).

On Duffy Island (Map 7) the dyke is some 80 feet thick but is not well exposed. It appears to consist entirely of homogeneous ultramafic rock (Sample 1 in Table 2) cutting Archean garnet-biotite gneiss. The rock consists of about 40% cumulate olivine enclosed by some 25% titanangite with minor pigeonite and small amounts of albite-microcline antiperthite, interstitial amphibole (hornblende) and biotite. Blebs of ilmenite, magnetite with hematite-rutile-anatase, and Fe-Ni-Cu sulphides comprise up to 10% of the rock.




Map 7: Duffy Island and Easter Island areas - Geology and Mineralisation.

The oxide and sulphide blebs show liquid immiscibility textures. The primary pyroxene and amphibole are both rimmed by green, sodic varieties (aegirine and riebeckite). The olivine has been slightly serpentinised in places and secondary Fe-Ni-Cu sulphides and hematite-rutile are abundant in these places.

On Easter Island (Map 7) the dyke consists of ultramafic and gabbroic rocks on the margins and mafic syenitic rock as the main phase. The marginal facies, which are separated from the country rocks by a very narrow chill zone, thin to the east from a maximum of about 3 metres. While the northern ones thin out by locality 5 and the southern by locality 8, at the Narrows (Map 6). These marginal facies (Samples 2 and 4) are quite similar to the dyke at Duffy Island but have suffered strong alteration to a talc-dolomite-siderite assemblage which contains up to 5% Fe-Ni-Cu sulphides (Plate 3 F & G). The rock still retains the cumulate and liquid immiscibility textures with a considerable amount of oxide blebs-hematite, rutile and anatase altered from ilmenite and magnetite. Serpentinisation of olivine is not uncommon (Plate 1 G). Between these marginal facies the dyke consists of 35 to 40% antiperthitic alkali feldspars; 25 to 30% titanaugite rimmed by aegirine and a blue amphibole, probably riebeckite; and about 10% each of an amphibole-hornblende with blue rims (riebeckite), biotite and skeletal magnetite-ilmenite-hematite and rutile-anatase (Plates 1 G and 3 F, G). The feldspars are predominantly intergrown anhedral albite and microcline but some relict oligoclase and secondary albite have been observed. Nowhere was quartz observed. In the centre of the dyke there is a thin pegmatitic facies consisting of coarse intergrown alkali feldspar and the blue amphibole, riebeckite. Euhedral apatite which is an ubiquitous minor constituent of the dyke, is abundant in the pegmatite.

Eastwards along Easter Island and into North Simpson Island the dyke becomes progressively more feldspathic (Samples 5 to 10 in Map 6 and Table 2). The ultramafic cumulate border phase disappears and a narrow chill zone of red-stained, oriented laths of albite becomes more noticeable. The red staining of albite is due to fine hematite dusting. From Sample 5 to Sample 8, the rock composition

PLATE 1

A) Odin Dyke pegmatite sample showing coarse, euhedral albite with minor microcline and orthoclase. The interstices are filled with dark amphiboles (riebeckite and hornblende) and pyroxenes (augite/aegirine and hypersthene?).

B) Maple (Caribou Lake) pegmatite containing riebeckite, hornblende, augite, aegirine and hypersthene? and lesser quartz and albite with some microcline.

C) Euhedral albite and microcline antiperthites replacing pyroxenes (augite and aegirine). Odin Dyke pegmatite. P.P.L.

D) Euhedral apatite (ap.) intergrown with aegirine (aeg.) and hornblende (hbl.), some replacing altered and hematitised albite and microcline (ab./micr.). Maple (Caribou Lake) pegmatite. P.P.L.

E) Euhedral plagioclase (labr.) intergrown with augite (aug.). Some minor quartz (qtz.), later apatite (ap.) and opaques (op.) are present. Maple (Caribou Lake) gabbro. P.P.L.

F) Quartz (qtz.) partly replacing and overgrowing altered and hematitised albite and microcline (ab./micr.) with inclusions of aegirine (aeg.) and augite (aug.). Maple (Caribou Lake) pegmatite. P.P.L.

G) Plagioclase (olig.) intergrown with olivine (ol.) and some minor biotite (bi.). Olivine is partly altered to serpentine (serp.) and also replaced by opaques (op.). Easter Island Dyke olivine gabbro. P.P.L.

H) Oolites partly replaced by late calcite (calc.), dolomite (dol.), siderite (sid.), chlorite (chl.) and opaques (op.) in a matrix of earlier calcite, dolomite and siderite. Fragment in Aristifats Lake agglomerate. P.P.L.

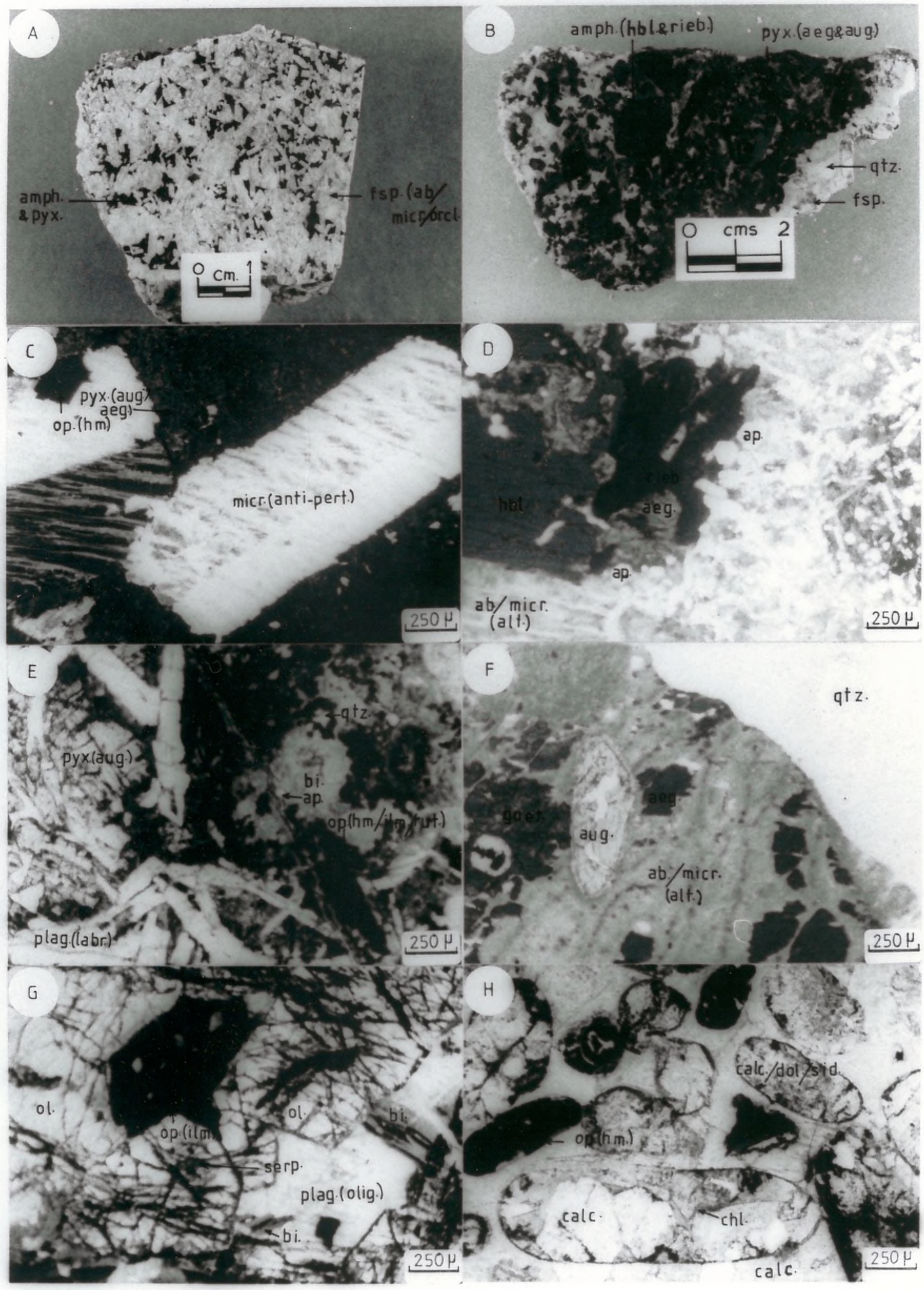

A
amph. & pyx.
fsp. (ab/micr/orcl.
0 cm. 1
B
amph. (hbl & rieb.)
pyx. (aeg & aug.)
qtz.
fsp.
0 cms 2
C
pyx. (aug/aeg.
op. (hm)
micr. (anti-pert.)
250 µ
D
ap.
rieb.
hbl
aeg.
ap.
ab/micr.
(alt.)
250 µ
E
qtz.
pyx (aug.)
bi.
ap.
op (hm/ilm/rut.)
plag. (labr.)
250 µ
F
qtz.
goet.
aeg.
aug.
ab/micr.
(alt.)
250 µ
G
ol.
op. (ilm)
ol.
bi.
serp.
plag. (olig.)
bi.
250 µ
H
calc./dol./sid.
op. (hm.)
calc.
chl.
calc.
250 µ

PLATE 1

varies slightly by which albite and microcline content increases progressively. By Dyke Lake the proportion of feldspars (albite and microcline) in the main part of the dyke has increased to about 60%. Here the pyroxene becomes progressively less important and consists increasingly of aegirine, although some titanaugite is found. Riebeckite, hornblende, biotite and hematite-rutile are minor constituents. Some alterations of these minerals especially feldspars are quite clearly observed. The feldspars are partially altered to carbonate (mainly dolomite with some siderite) and sericite, while the mafics are chloritised.

From Dyke Lake eastwards (Samples 11 and upward) the feldspars are increasingly altered to carbonates and sericite, and the mafic minerals present are altered to chlorite, hematite and probably epidote. The whole rock becomes dusted with secondary hematite and thus shows a reddish-brownish coloration. Small veinlets of albite and bostonite, often heavily carbonated and hematitised, are common in the dyke and at the contact zone with the country rocks. Here the bulk of the dyke consists predominantly of alkali feldspars (mostly antiperthitic albite and microcline with minor orthoclase), riebeckite, hornblende acicular apatite and also strongly pleochroic biotite (Samples 11 and 12). The amount of feldspar is about 80 to 90%. The rock, coarse hypidiomorphic red albite-syenite as it has been called, is quite strongly altered. The alteration products include carbonates (dolomite and siderite), sericite, chlorite, epidote (?) and hematite-rutile. From Eagle Bay eastward to Ref Peninsula only the roof zones are exposed (Samples 13 to 20). The roof zones are made up of successive pulses of progressively more feldspathic rocks until the latest phases are pure albitites and bostonites (equivalent to Samples D and E in Table 3). Albitites and bostonites are generally reddish or brownish in appearance due to fine dusting by secondary hematite. Often veinlets of bostonites and albitites cut the surrounding rocks and are particularly common in the Sosan Group sandstones which lie unconformably on the Archean basement . These albitites and bostonites are vesicular in places and associated with small brecciated zones interpreted as gas breccia pipes (Samples D and E in Table 3). It is important to note that the albitite and bostonite veinlets are closely similar to the igneous matrices of the breccia pipe at Vestor Channel and Big Red Cliff located on Map 6.

Before alteration, the red-brown bostonite was composed of about up to 95% albite, trace to minor quartz and in places up to 10% each of apatite and mafics. The texture is idiomorphic to hypidiomorphic with an imperfect trachoidal arrangement of albite laths. These are mostly less than 2mm in length with equidimensional albite phenocrysts rarely present. Alterations include hematitisation, sericitisation, carbonatisation, and chloritisation. Although locally a dark grey bostonite is present, it is petrographically identical to the reddish bostonite. Sericitisation has affected bostonite to varying degrees from only a minor alteration of albite to complete replacement and the rock can contain up to 10% secondary sericite. In drill cores available, certain contact zones and thin dykes are completely altered to pale green sericite with dolomite over width of up to several feet. Dolomitization of the bostonite is ubiquitous with replacive dolomite often constituting 20 to 60% of the rock. The chlorite content in these varies from trace to about 25%, the chlorite either replacing the mafics or albite. Albitite on the other hand consists essentially of pure albite, although some alterations do persist as in the case of the bostonites. It is observed that some massive bostonites are composed of bostonite clasts in a comminuted and recrystallised albite matrix. The bostonite has been extensively brecciated in many areas during diatreme activity.

iii. Petrochemistry

The chemically and petrographically analysed samples were collected from the entire exposed length of the dyke and in places, across its width (Map 6). These samples are numbered 1 to 20 on the map and also in Table 2. The analysed bostonites from Preble Channel and Vestor Channel diatremes are presented in Table 3 as Sample D and E respectively.

The central phases provide a more representative phases of the dyke than do the marginal facies. This is because of the cumulate nature of the margins in the west and their altered state throughout. The samples were selected on the basis of lack of weathering and secondary veining and analysed by X-ray fluorescence spectrometer (see Appendix for the procedure). The results have been corrected

| Sample | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 | 14 | 15 | 16 | 17 | 18 | 19 | 20 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| D miles: | 0 | 0.3 | 0.3 | 0.3 | 1.4 | 2.3 | 3.5 | 5.3 | 5.5 | 5.5 | 7.8 | 8.6 | 12.3 | 12.5 | 12.5 | 12.9 | 13.1 | 13.7 | 14.2 | 17.1 |
| Central or marginal % | C | NM | C | SM | C | C | C | C | C | SM | C | C | NM | NM | C | C | C | C | C | C |
| $SiO_2$ | 37.0 | 35.6 | 44.2 | 42.9 | 46.2 | 42.7 | 45.5 | 48.5 | 42.1 | 45.7 | 49.8 | 47.8 | 50.8 | 46.0 | 46.4 | 53.8 | 47.0 | 53.6 | 46.0 | 47.3 |
| $TiO_2$ | 2.6 | 11.4 | 5.6 | 4.2 | 4.5 | 3.9 | 4.0 | 3.4 | 5.4 | 8.6 | 2.3 | 2.2 | 1.9 | 3.4 | 2.9 | 1.4 | 1.8 | 2.4 | 2.6 | 2.1 |
| $Al_2O_3$ | 4.9 | 7.3 | 11.5 | 13.4 | 13.5 | 11.7 | 11.9 | 14.0 | 12.6 | 10.2 | 15.2 | 14.6 | 17.4 | 10.9 | 16.1 | 16.6 | 16.7 | 15.3 | 14.4 | 16.0 |
| FeO | 18.5 | 23.0 | 16.0 | 12.2 | 14.4 | 15.1 | 15.7 | 13.3 | 14.0 | 18.2 | 11.2 | 10.4 | 5.8 | 13.6 | 10.8 | 7.8 | 9.4 | 9.6 | 12.4 | 11.2 |
| MgO | 18.9 | 6.4 | 5.4 | 4.9 | 4.6 | 9.1 | 4.8 | 3.9 | 6.6 | 8.9 | 3.5 | 3.6 | 4.0 | 4.1 | 2.2 | 2.7 | 5.2 | 3.2 | 4.5 | 4.8 |
| CaO | 4.2 | 8.1 | 7.3 | 7.1 | 6.1 | 2.4 | 5.4 | 5.0 | 3.9 | 1.5 | 4.3 | 4.3 | 5.6 | 4.8 | 5.2 | 2.7 | 2.7 | 2.8 | 4.8 | 2.8 |
| $Na_2O$ | 1.3 | 2.2 | 3.6 | 1.7 | 4.1 | 2.1 | 3.8 | 4.4 | 2.1 | 1.6 | 4.9 | 3.8 | 4.1 | 4.6 | 1.9 | 5.9 | 4.3 | 4.9 | 4.4 | 3.9 |
| $K_2O$ | 0.7 | 1.1 | 1.4 | 2.6 | 1.8 | 1.8 | 1.3 | 2.5 | 2.2 | 1.8 | 2.3 | 2.6 | 2.6 | 2.5 | 4.2 | 3.0 | 2.5 | 2.6 | 1.3 | 2.2 |
| $H_2O$ | 6.1 | 2.1 | 2.6 | 4.0 | 2.4 | 4.4 | 3.3 | 2.2 | 4.0 | 4.6 | 2.4 | 3.8 | 2.1 | 1.6 | 2.8 | 1.7 | 3.3 | 2.2 | 2.7 | 2.8 |
| $CO_2$ | 2.2 | 0.3 | 0.4 | 8.0 | nd | 3.5 | 0.4 | nd | 4.0 | 0.3 | 1.5 | 3.6 | 6.0 | 11.8 | 8.0 | 4.4 | 3.6 | 3.2 | 3.6 | 3.7 |
| Total | 96.4 | 98.5 | 98.0 | 101.0 | 97.6 | 96.6 | 96.1 | 97.2 | 96.9 | 101.4 | 97.4 | 96.8 | 100.3 | 103.3 | 100.5 | 100.0 | 96.5 | 99.8 | 96.7 | 96.8 |
| p.p.m. | | | | | | | | | | | | | | | | | | | | |
| Zr | 199 | 213 | 307 | 304 | 407 | 397 | 362 | 471 | 353 | 300 | 583 | 641 | 909 | 321 | 516 | 758 | 669 | 843 | 441 | 526 |
| y | 63 | 59 | 72 | 92 | 80 | 88 | 123 | 91 | 87 | 72 | 102 | 107 | 150 | 94 | 117 | 124 | 113 | 120 | 93 | 106 |
| Nb | 100 | 95 | 113 | 130 | 125 | 131 | 34 | 138 | 131 | 111 | 153 | 171 | 215 | 125 | 151 | 202 | 177 | 185 | 152 | 156 |
| P | 1382 | 1394 | 1877 | 2089 | 2164 | 2289 | 2910 | 2931 | 1509 | 1695 | 2942 | 2641 | 2642 | 1562 | 3528 | 1645 | 2267 | 2769 | 4143 | 3512 |
| Rb | 19 | 33 | 31 | 75 | 34 | 50 | 37 | 47 | 44 | 33 | 61 | 31 | 99 | 100 | 166 | 64 | 54 | 61 | 23 | 61 |

Table 2: Partial chemical analyses for samples of the Easter Island Dyke. The distance of each sample east of Sample 1 is shown. The sample location within the dyke is shown by NM (North Margin), SM (South Margin) or C (Central). Total iron is shown as FeO.

Continued overleaf.

| Sample | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 | 14 | 15 | 16 | 17 | 18 | 19 | 20 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| p.p.m. continued | | | | | | | | | | | | | | | | | | | | |
| Sr | 285 | 242 | 303 | 192 | 471 | 366 | 366 | 572 | 566 | 144 | 432 | 338 | 163 | 240 | 258 | 325 | 326 | 186 | 574 | 551 |
| Ba | 292 | 1262 | 775 | 356 | 770 | 610 | 585 | 768 | 804 | 965 | 602 | 652 | 171 | 789 | 3956 | 688 | 591 | 578 | 601 | 796 |
| La | 21 | nd | 30 | 32 | 40 | 39 | 31 | 55 | 23 | nd | 72 | 75 | 242 | 34 | 38 | 59 | 50 | 74 | 73 | 38 |
| Ce | 109 | 78 | 104 | 78 | 110 | 102 | 127 | 138 | 83 | 95 | 161 | 125 | 156 | 84 | 117 | 72 | 128 | 173 | 137 | 111 |
| V | 252 | 1410 | 578 | 363 | 391 | 314 | 295 | 237 | 454 | 1020 | 100 | 94 | 66 | 316 | 224 | 49 | 14 | 126 | 166 | 151 |
| Cr | 528 | 87 | 16 | 15 | nd | nd | 24 | nd | nd | 72 | nd | nd | nd | 89 | nd | nd | nd | nd | nd | nd |
| Mn | 1612 | 1671 | 1450 | 2184 | 1322 | 1210 | 1600 | 1464 | 1154 | 928 | 1461 | 1041 | 1242 | 3925 | 3192 | 1066 | 756 | 828 | 991 | 856 |
| Ni | 1018 | 236 | 56 | 81 | 32 | 28 | 24 | 19 | 36 | 290 | 15 | 16 | 16 | 104 | 37 | 11 | 10 | 22 | 7 | 10 |
| Cu | 122 | 296 | 56 | 65 | 41 | 44 | 19 | 32 | 37 | 600 | 24 | 14 | 9 | 79 | 14 | 11 | 13 | 14 | 22 | 16 |
| Zn | 167 | 206 | 197 | 19 | 118 | 108 | 206 | 259 | 57 | 89 | 130 | 26 | 16 | 8 | nd | 11 | 57 | 11 | 15 | 58 |
| S | 1071 | 5228 | 1648 | 1327 | 1096 | 960 | 1113 | 1068 | 1322 | 3942 | 426 | 417 | 735 | 1648 | 1214 | 564 | 386 | 298 | 646 | 587 |
| Cl | 789 | 286 | 282 | 416 | 387 | 476 | 393 | 292 | 350 | 412 | 453 | 427 | 470 | 459 | 443 | 738 | 551 | 683 | 552 | 572 |
| Ga | 18 | 35 | 33 | 28 | 39 | 39 | 36 | 40 | 36 | 34 | 40 | 40 | 37 | 26 | 36 | 41 | 36 | 42 | 37 | 38 |
| As | 4 | nd | 1 | 3 | 1 | 2 | 14 | nd | 13 | nd | 4 | 10 | 3 | 10 | 8 | 3 | nd | 3 | 3 | 2 |
| Mo | 32 | 28 | 32 | 44 | 37 | 37 | 34 | 38 | 40 | 34 | 44 | 50 | 54 | 42 | 43 | 52 | 50 | 49 | 42 | 48 |

Table 2: Continued from previous page. Partial chemical analyses for samples of the Easter Island Dyke. The distance of each sample east of Sample 1 is shown. The sample location within the dyke is shown by NM (North Margin), SM (South Margin) or C (Central). Total iron is shown as FeO.

against the calibration curves derived from international standards. Water and carbondioxide were analysed gravimetrically.

In Table 2 the data are not rounded up to 100% nor are analyses presented on water-carbondioxide free basis. It is thought that much of the hydration and carbonation were autometasomatic and hence the values given in the table are a meaningful representation of the original chemistry of the samples. The data are then plotted on standard AFM and alkalis-silica diagrams (Fig. 1), after Irvine and Baragar (1971). The strongly alkalic nature and relative lack of iron enrichment are well shown. So too are the obviously anomalous chemistries of the marginal samples compared with the consistency of the central phase samples. The variations in chemistry are examined as a function of distance along the dyke, that is from Sample 1 eastward to Sample 20. The assumption here is that the samples taken increasingly further to the east represent successively higher levels in the dyke as was postulated from the field and petrographic studies. The chemical variations can be separated into three parts as generalised below:

1. An initial anomalous value for the cumulates (Samples 1 and 2)
2. A progressive differentiation of the dyke with distance (Samples 3 to 12).
3. Irregular but relatively constant values eastwards from Sample 12 to Sample 20.

However the exact chemistry of the parent magma of the dyke is not easy to estimate. The average composition of the parent magma is quite similar to that of Brogger's (1894) analysis of the classic camptonite from Maena in the Oslo Graben. The Easter Island Dyke is petrographically and chemically equivalent or quite similar to camptonite. It was Brogger who recognised the intimate relationship of both bostonites and breccias with camptonites and maenaites, and argued their derivation by differentiation from an alkalic olivine-gabbro parent. In the classic Gran area of the Oslo Graben, large intrusions of such olivine gabbro outcrop close to the camptonites. Hence it is quite reasonable that the bostonites and the breccia pipes of East Arm were derived by differentiation from the same parental magma as the Easter Island Dyke.

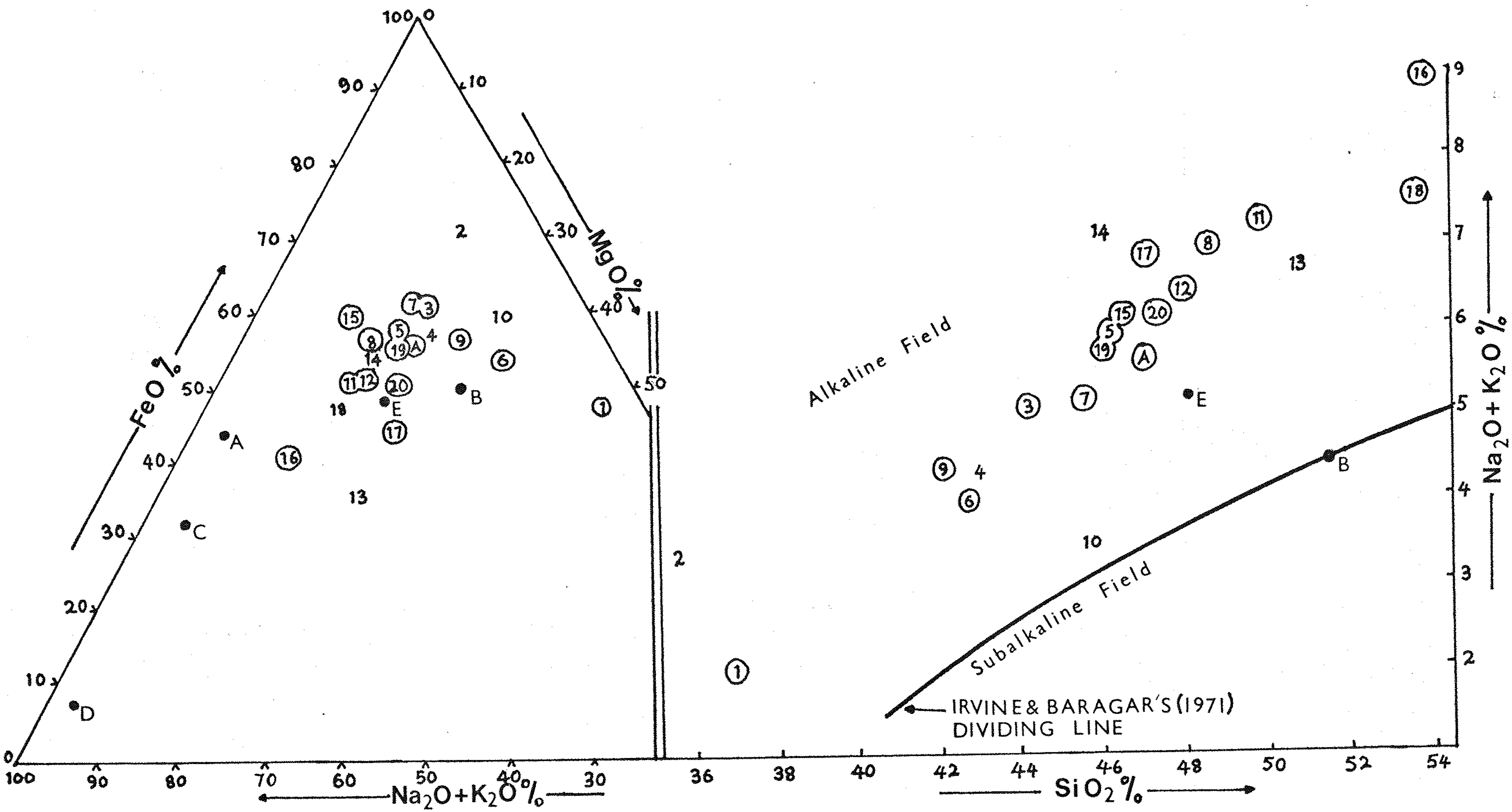



**Fig. 1:** AFM and Alkalis: Silicadiagrams for the Easter Island Dyke (ringed numbers represent central and open numbers, marginal samples, and Ⓐ is the calculated average composition of the Dyke); the Maple (Caribou Lake) gabbro (Sample B), fine-grained syenite (Sample A); the fine-grained Odin Dyke (Sample C); and the Preble Channel (Sample D) and Vestor Channel (Sample E) bostonites.

Besides the Fe-Ni-Cu sulphide mineralisation in the mafic and ultramafic cumulates, in some places within the Easter Island Dyke, U, Ag, Bi, Ni-Co-Fe arsenide mineralisations are found. This hydrothermal mineralisation is clearly related to the early Aphebian alkaline magmatism in the area.

iv. Evolution of Easter Island Dyke

Petrographically the dyke is a marginally saturated sodic orthosyenite. It is proposed that an originally volatile, alkaline magma rose up a tight fracture in Archean basement (Badham, In Press). As the magma approached its final levels, olivine crystals accumulated at the base and volatile fractions began to migrate upwards. Initially the volatiles rose through the whole of the crystallising dyke causing ubiquitous soda metasomatism. Later, and especially in the lower half of the intrusion, the volatiles were able to escape only up the margins which were subsequently further altered. The upper portions of the dyke reached the unconformity with lower Proterozoic sandstones whose lesser competence and probable saturation with water permitted the escape of late differentiates with the hydrothermal fluids and gases. Hydrothermal fluids in the upper reaches of the dyke probably caused the pervasive alterations mentioned. However it is not clear to what extent the hydration and carbonation processes can be considered to have been autometasomatised. With regard to the mineralisation in the dyke, a schematic vertical section model of the dyke is shown in Fig. 2 from Badham (In Press).

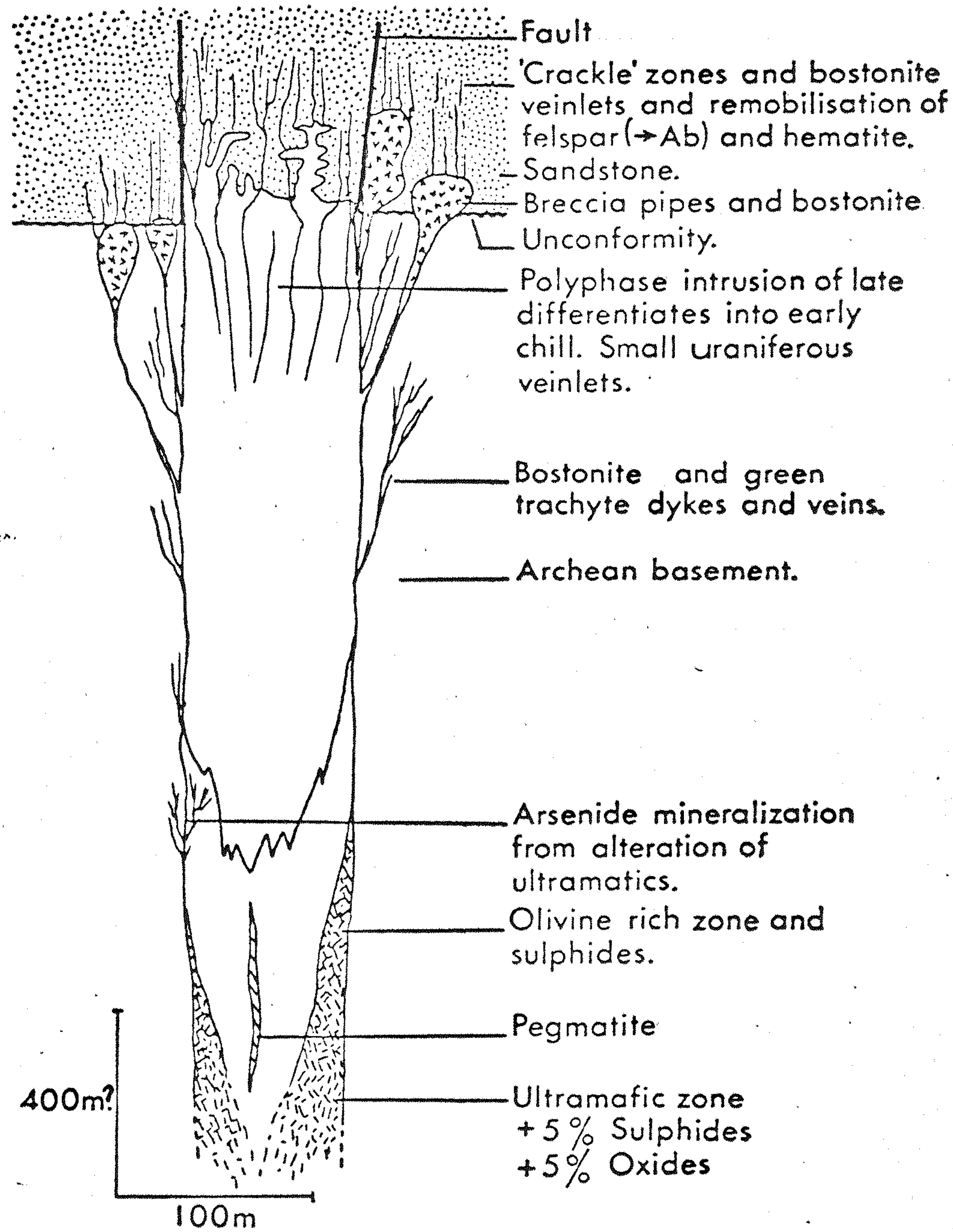


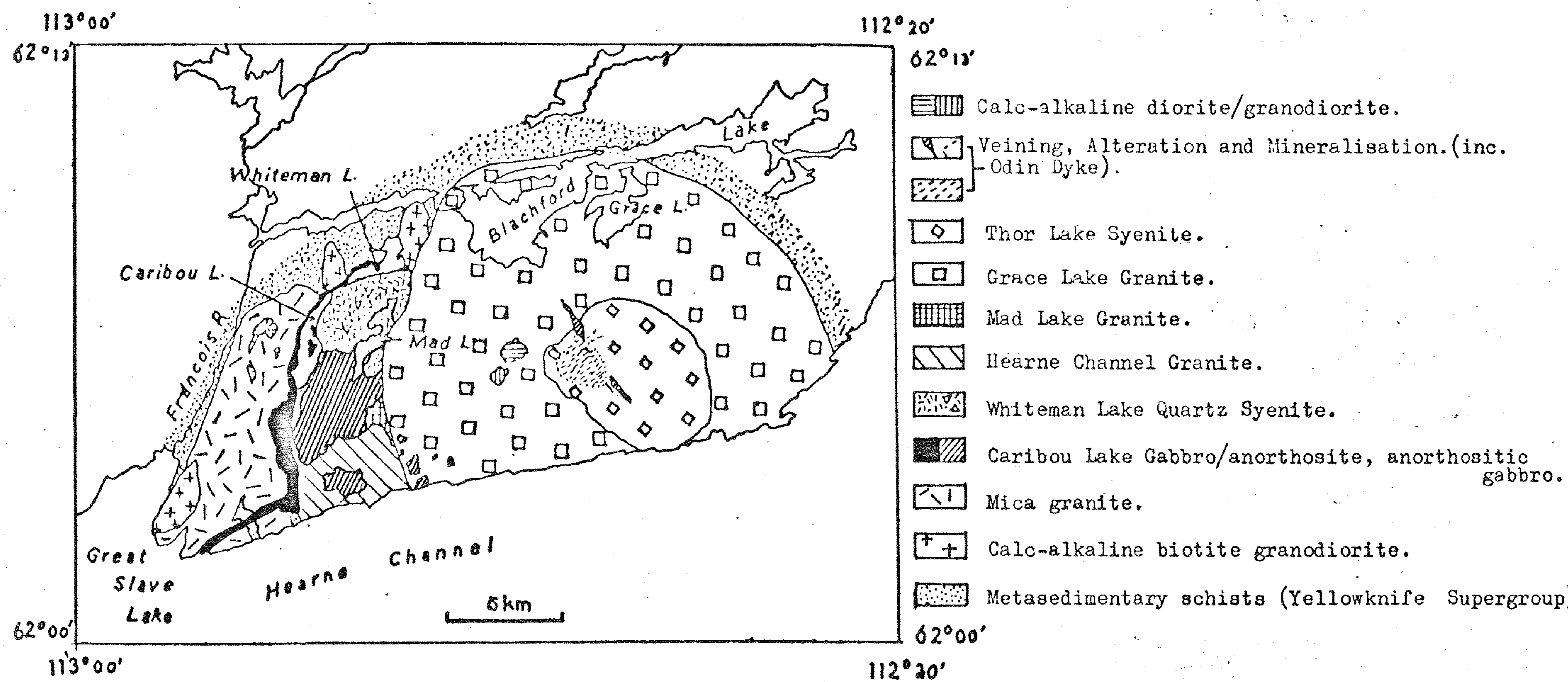

**Fig. 2: Vertical section of the Easter Island dyke.**

b. The Maple-Blachford Lake Complex.

i. General Geology

This alkaline complex intrusion lies on the north shore of Great Slave Lake (Maps 4 and 8) and underlies an area close to 235 $km^2$. The Odin Dyke is a part of the complex. The initial Maple Complex (named here after the Maple Claim Group on Ni, Co, Ag mineralisation) was originally described by Henderson and Joliffe (1941) as consisting of gabbro, syenite and anorthosite cut by Archean granite. Now the Maple Complex is considered as a part of the Blachford Lake Complex and known as Caribou Lake Gabbro (Davidson, 1978). In 1975, Badham had observed unequivocal evidence along the western margins of the 'Maple Complex' that the gabbroic rocks postdated the Archean rocks and predated the deposition of the mid-Aphebian Great Slave Supergroup in the East Arm. This is supported by petrographic works by Badham (In Press), Davidson (1978) and the author. The detailed geology of this complex intrusion is well documented by Davidson (1972, 1978). The gabbros are chilled against the granites, which are reddened and recrystallised. Archean pegmatites do not cut the contact, whereas late differentiate phases (both pegmatitic and fine-grained) of the gabbros do cut the granites. These gabbros, containing xenoliths of granite and metasedimentary materials, are undeformed whereas the granites are slightly foliated. Therefore the complex is post-Archean.

On the east side the 'Maple Complex' (Caribou Lake Gabbro and its associated rocks) is cut by the main Blachford Lake Complex. This Blachford Lake Complex consists of peralkaline granite and alkali syenite with late pegmatitic phases (Davidson, 1972, 1978). In this study the author did not examine the rocks of the Blachford Lake Complex (because no samples were available) except for one such pegmatite, that is the Odin Dyke, at its centre. A small mass of diorite is apparently included in the main body of the alkaline granite (Davidson, 1972). A K-Ar radiometric age of 2057 ± 56 m.y. has been obtained on an amphibole (riebeckite) from the alkaline granite of this complex. (Wanless, in Davidson, 1978). The data suggest that the Maple-Blachford Lake Complex is related and similar in age to the Easter




Map 8 : Revised geology of the plutonic rocks in the Blachford Lake area, from Davidson, 1978.

Island Dyke. On the other hand the diorite plutons within the Blachford Lake alkaline granite are texturally and compositionally very similar to the younger diorite laccolith on Blanchet Island, 10 km to the south (Davidson, 1978). The Maple-Blachford Lake Complex was intruded by a swarm of diabase dykes. Dykes of a similar trend (ENE) are overlain by the Sosan Group in the East Arm.

ii. Petrography/Petrology

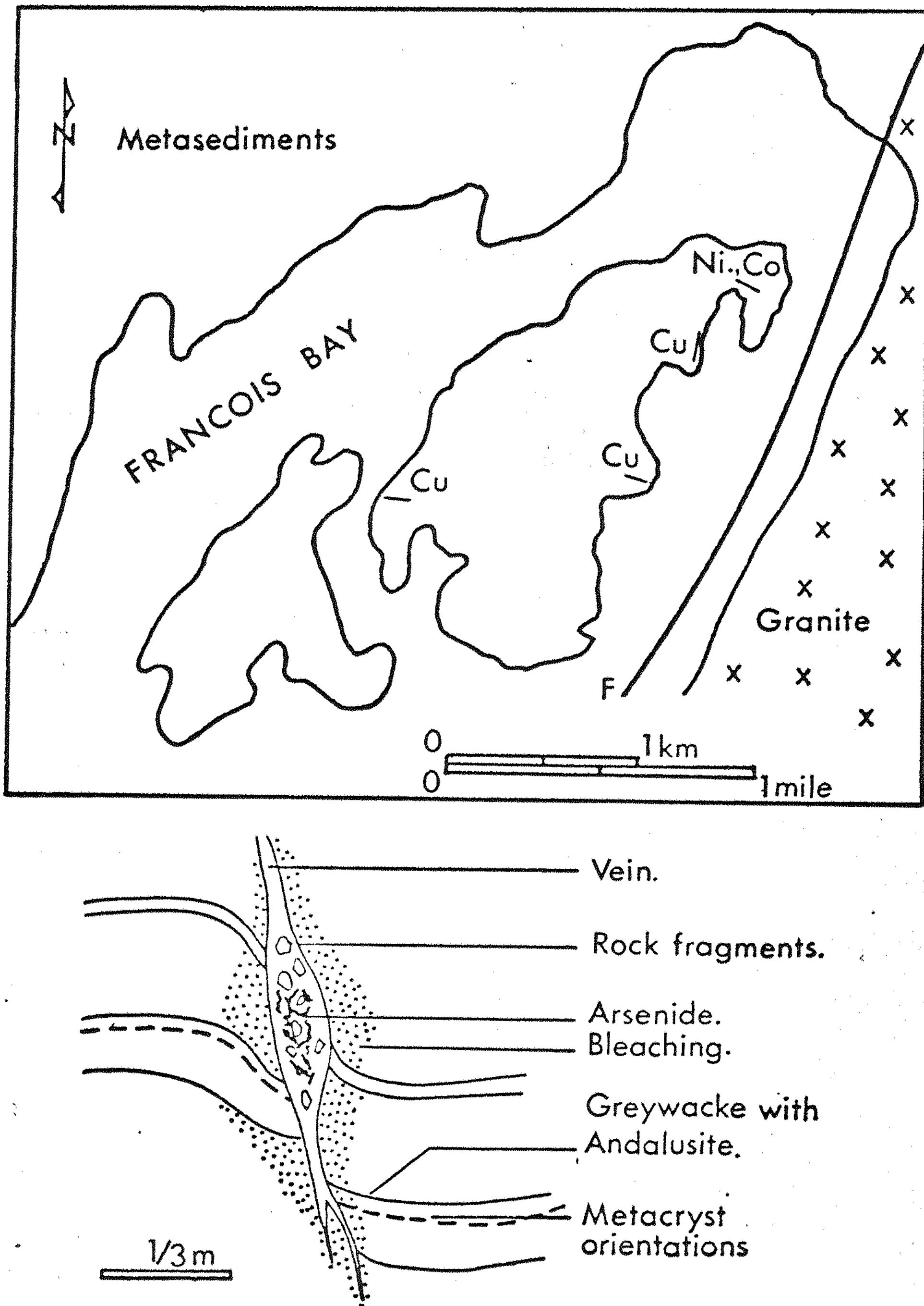
Davidson (1978) has documented in detail the petrology of the Blachford Lake Complex. Only samples of small parts of the Complex, that is the Maple (Caribou Lake) gabbros and related rocks, and also the Odin pegmatitic dyke were examined by the author.

The 'Maple Complex' consists predominantly of homogeneous gabbro, albite syenite (locally with quartz) and anorthosite (and anorthositic gabbro). The gabbro, varies from fine to coarse-grained, contains some 45 to 50% feldspars; 25 to 30% colourless augite and some aegirine; about 20% very strongly coloured biotite; and about 5% apatite, which is ubiquitous as late crystallised needles but often occurs in clusters and lenses too (Plate 1 E). Anhedral quartz was observed especially in the samples taken from the western part of the gabbro complex. The commonly anhedral feldspars consist predominantly of zoned sodic plagioclases (ranged up to labradorite) rimmed and replaced by albite antipertites. Some microcline and minor orthoclase were also observed. Some perthites were distinguished. The subhedral pyroxenes are often altered to green and brown amphiboles (hornblende and riebeckite), biotite and iron-titanium oxides. The rock appears to be a mildly alkaline gabbro which has suffered a certain degree of soda metasomatism. Bands of more mafic gabbro, which contain up to 5% disseminated Cu-Fe sulphides are seen in places. These sulphides and oxides occur as blebs, showing liquid immiscibility textures against the silicates. Skeletal ilmenite, magnetite, hematite and rutile were also observed in the bands of virtually pure plagioclase ($An_{20}$ - $An_{40}$). However these petrographic data seem generally to support an origin by accumulation for these rocks.

In the same intrusion, some multiply-intruded pegmatitic phases are not uncommon. These pegmatites (Plate 1 B & F) contain coarse euhedral feldspars (albite and microcline antiperthites); hornblende-riebeckite amphiboles; red-brown biotite; pyroxene (augite, aegirine and hypersthene ); apatite; and overgrown quartz. There are progressively more feldspars found in later phases. Individual pegmatitic areas are complex, consisting of an initial crystallisation of albite antiperthite with some microcline, and aegirine overgrown by large crystals of riebeckite. The latest phases consist of acicular apatite with some sphene, zircon, monazite and quartz. The anhedral overgrown quartz, resulting from a silicification process, contains inclusions of feldspars, amphiboles and pyroxenes. Alteration of feldspars, pyroxenes and amphiboles is common with indications of secondary minerals such as chlorite, sericite, carbonate (dolomite and siderite) and perhaps epidote. However some secondary oxide of iron and titanium (such as hematite, rutile and anatase) were observed replacing the primary minerals. From the petrographic data, these pegmatites would represent alkaline differentiates from the parent gabbro.

In association with the aforementioned gabbro, is a quartz syenite phase which is well exposed along the south shore of Whiteman Lake (Map 8). Although quartz syenite is the predominant rock type, it varies systematically in composition and appearance. In places this syenite is devoid of quartz. This medium-grained rock generally contains 85 to 90% feldspars; about 5% rhombic and fine-grained dolomite and siderite; up to 2% quartz and some hornblende. Other minor accessories include acicular apatite, chlorite, zircon, sphene, monazite, sericite-muscovite, iron and titanium oxides and perhaps epidote. The feldspars are made up of albite, microcline, some orthoclase and some plagioclases (andesine and oligoclase). They are moderately sericitised, chloritised and carbonatised and dusted with fine-grained hematite. Most of the primary mafics, especially amphibole, are altered to secondary chlorite, carbonates, hematite and epidote. Fine microcrystalline quartz occurs either as interlocking crystals with feldspars or filling the interstices like the carbonates. However this quartz syenite is closely associated with fine-grained pink dykes which consist predominantly of red-stained albite and microcline with accessory apatite, sphene, zircon and monazite. The dykes are often strained and somewhat cataclastic and are cut by veinlets of carbonates and hematite. It is important to note that these syenitic rocks are quite similar to the late bostonitic phases of the Easter Island Dyke. Around the contact with the Archean rocks, there is no sign of even incipient breccia pipes but thin veinlets of syenitic material are common. It is presumed that there was a concentration of volatiles late in the differention history of the 'Maple Complex'. Where this was trapped, the late pegmatitic phases grew. Thus the fine-grained dykes represent areas of partial escape of these volatiles and late differentiates.

The Odin Dyke, a pegmatite in the alkali syenite portion of the Blachford Lake Complex, lies along 1 km of an east-west striking photolinear. This pegmatitic dyke is highly variable. A fine-grained, red, feldspathic marginal phase is usually present but the bulk of the dyke, which is some 100 metres thick, consists of irregular patches of pure feldspars (mainly albite with some microcline), feldspars and amphiboles, and pure amphiboles (hornblende and riebeckite). Other minor accessories include trace pyroxenes (augite and aegirine), microcrystalline quartz, apatite, sericite-muscovite, chlorite, rare earth element minerals and iron-titanium oxides. Different areas are of different grain-size and show different intermineral relationships. The reddish brown feldspar is albite antiperthite with small patches of microcline. Some orthoclase was determined. Relict plagioclase inclusions are often found in the coarse sodic feldspar. The amphiboles are predominantly riebeckite but some hornblende and actinolite often found. Some augite and aegirine are enclosed in the large riebeckite crystals. Beside apatite, other rare earth element minerals such as zircon and monazite are abundant in places. In the brecciated areas the fractures are dusted with hematite and usually contain fluorite, albite, sparse analcime, acmite (as prisms) and rare earth element minerals. Some of these rare earths have not yet been identified. Bastnaesite has been identified by Davidson (1978). Larger areas of brecciation contain fragments of early carbonates and comminuted rocks. A small fracture in the northern margin of the dyke contains fragments cemented by apatite and riebeckite, with subordinate magnetite, carbonate and fluorite.




**Map 9: Zig area** - Geology and Mineralisation.

iii. Petrochemistry and Evolution

For the partial chemical analyses, three rock samples from the Maple-Blachford Lake Complex were taken. The results are shown in Table 3. The analyses have not been rounded up to 100% again, in order to facilitate comparisons with others presented here. Sample A is an augite-plagioclase-biotite gabbro typical of the Maple (Caribou Lake) gabbro - on its western edge. Sample B is a fine-grained, pink, syenitic dyke within the 'Maple Complex'. The samples are quite closely similar to the Easter Island Dyke gabbroic rock and albite-syenitic rock respectively. The only difference is that there is no quartz in the albite syenite of the Easter Island Dyke. The 'Maple' gabbro is marginally (mildly) alkaline and the syenitic dyke more so: the data are not inconsistent with the assumption that the latter is the differentiate of the former. The potash content of the syenite is rather high for the proportions of albite to microcline and orthoclase observed in thin sections. Sample C, from a relatively fine-grained and unbrecciated portion of the Odin Dyke, shows the dyke to be strongly differentiated and strongly alkaline. However, the results can also be compared to the analyses of bostonites (Samples D and E) from Preble Channel and Vestor Channel diatremes respectively. Finally all these rocks are chemically alkaline in nature, whatever their appearances. The evolution of the Maple-Blachford Lake Complex is in ways similar to that of the Easter Island Dyke but the only essential difference being their mode of emplacement. The Odin Dyke is also similar to the later phases of the 'Maple' (Caribou Lake) Complex and of the Easter Island Dyke. It was emplaced by a series of gas brecciation events which introduced successively later differentiates of the parent magma. Ni-Co-arsenide mineralisation found (but not sampled) in the gabbro and in the Gin and Zig areas may have a genetic relationship to the Maple-Blachford Lake Complex. This mineralisation is so similar to that of the Easter Island Dyke that a close genetic and possibly temporal relationship is suggested.

| Sample | A | B | C | D | E |
|---|---|---|---|---|---|
| $SiO_2$ | 68.6 | 51.5 | 64.0 | 72.0 | 48.0 |
| $TiO_2$ | 0.6 | 2.8 | 0.3 | 0.2 | 1.4 |
| $Al_2O_3$ | 17.1 | 15.7 | 17.7 | 18.2 | 16.5 |
| FeO | 6.2 | 8.5 | 5.0 | 0.6 | 8.3 |
| MgO | 0.4 | 3.4 | 0.8 | 0.4 | 3.6 |
| CaO | 0.4 | 4.8 | 1.0 | 0.1 | 7.6 |
| $Na_2O$ | 2.6 | 2.7 | 4.2 | 5.5 | 2.2 |
| $K_2O$ | 5.0 | 1.8 | 5.6 | 2.0 | 2.9 |
| $H_2O$ | 1.1 | 1.3 | 0.9 | 1.0 | 1.9 |
| Total | 102.0 | 92.5 | 99.5 | 100.0 | 92.4 |

Table 3: Partial chemical analyses for:

A - Maple-Blachford Lake Complex syenitic dyke;

B - Maple (Caribou Lake) Gabbro;

C - Odin Dyke fine-grained phase;

D - Preble Channel bostonite with sandstone fragments;

E - Vestor Channel bostonite.

c. The Diatreme Breccia Pipes

i. General Geology

In the East Arm area, there are lines of diatreme breccia pipes along Vestor, Preble and Inconnu Channels and a few isolated pipes occur to the east (Maps 4 and 5). Most of the Preble Channel system and the western part of the Vestor and Inconnu Channels systems have been described by Reinhardt (1972). All the three systems were mapped by Walker and Badham for Vestor Explorations Ltd. in 1970 and again by Badham in 1975. Badham (1975 and In Press) suggested that the Vestor Channel diatremes are cogenetic with the Easter Island Dyke from the field relationships cited above. All three systems can be seen to be post-Lower Sosan (Susanne) Group and pre-Et-Then Group. Hoffman (1978) suggested that they post-date the stark Formation of the Chritie Bay Group, due to the supposed presence of blocks of stark limestone characterised by its stromatolitic and rippled structures in the breccias. To the author this controversy still remains unresolved. The breccia pipes do contain blocks of some stromatolitic and rippled limestone, but these may have been derived from the Union Island Group limestones. It is very likely that the Vestor Channel breccia pipes were derived from the Easter Island Dyke parent magma due to their very close association.

ii. Petrography/Petrology

The Vestor Channel pipes, which outcrop intermittently from the west end of Vestor Channel up to Big Red Cliff, (Map 4 and 5), are highly variable and complex in detail, though they are similar in general. Various sizes of assorted blocks, sometimes angular, sometimes abraded and rounded, are found in a matrix that in places consists of comminuted rock fragments and elsewhere consists of red trachytic bostonite (Sample E in Table 3). The blocks include basement rocks, Sosan Group sandstones, and numerous fragments of dolomite and argillite that are thought to be from restricted fault-bounded deposits of the lowermost Sosan age (Badham, In Press) No blocks of the Easter Island Dyke are found. The pipes appear to have been initially gas breccia pipes that were constricted in the basement but that expanded and died out in the sandstones. Later eruptions up the same pipes set the emplacement of the bostonite magma. It is felt that the first pulse was highly gas-charged and often fragmented but the second was rather less volatile and cemented the first. Petrographically, these bostonites are indistinguishable from those in the upper parts of the Easter Island Dyke and two appear to be cogenetic in the Dyke Lake area. The pipes consist of about 90% albite and 10% carbonates (mainly dolomite with some siderite). Traces of apatite, quartz and mafics were observed. Besides carbonates other alteration products include chlorite, sericite and hematite. These breccia pipes are clearly characterised by albite, iron and magnesium carbonates and hematite alteration haloes. The crackle zones in the sediments where no definite pipe has been recognised are often filled with albite, carbonates and hematite, and surrounded by similar alteration haloes as surround the pipes.

From these observations, Badham (In Press) has proposed that these pipes represent the escape paths of volatiles and late differentiates from the Easter Island Dyke. In places they rupture the top of the dyke itself. The above proposition is remarkably reasonable. Elsewhere they intrude other faults in the complex fault zone (Maps 4 and 5 superimposed on Map 3). According to Reinhardt (1972) these pipes may be related to some uranium and copper mineralisations in the sandstones. However the pipes disrupted and partially destroyed these earlier mineralisations which are seen in fragments and there is no mineralisation genetically associated with the breccia pipes in this area.

The Inconnu and Preble Channels diatremes are closely similar to those in Vestor Channel except that an igneous matrix has only been identified in one small exposure in Preble Channel (Reinhardt, 1972). These pipes contain blocks of Archean basement lithologies, Lower Sosan Group sandstones, argillites and carbonates and rarely black shales, that may be from a Union Island Group lithology not exposed in these areas. Carbonate (dolomite and minor siderite) and hematite veins, crackle-zones and alterations are ubiquitous. Although albitisation is less abundant than in Vestor Channel, it has been recognised in all the pipes. From these data, it is not unreasonable to suppose that both the Preble and Inconnu Channels diatreme systems are underlain by alkaline igneous rock somewhat similar to the Easter Island Dyke (which is thought to be the parent for the Vestor Channel breccia pipes). This is further supported by the presence of Co-Ni-Cu arsenide mineralisation in a vein at the west end of Preble Channel System (that is the claims of Valnicla Copper Mine Ltd. shown in Map 5). This mineralisation is closely akin to that in the Easter Island Dyke.

Albite-carbonate crackle-zones have been recognised in Sosan Group sandstones, both on Isui Island (Map 5) and on Fairchild Peninsula near Fort Reliance (Map 4). Both occurrences are on the master fault of the system.

iii. Petrochemistry and Evolution

Two samples of the bostonites from these diatremes were chemically analysed and the results are given in Table 3. Sample D is a bostonite from Preble Channel. The rock consists predominantly of albite with some microcline and minor carbonate (ferroan dolomite). The sample is unavoidably contaminated with small fragments of quartzite for it proved impossible to select a sample entirely free of these. It was found that the rock is similar to the bostonites from, for example the Oslo Graben (Brogger, 1894). Sample E is a bostonite from Vestor Channel diatreme pipe, consiting of about 90% albite and about 10% carbonates (mostly ferroan dolomite and some siderite). Again it is found similar to the bostonite of the Oslo Graben, and also to the bostonite facies of the Easter Island Dyke. Chemically both of these bostonites (from Preble and Vestor Channels) are alkaline in nature (Fig. 1).

d. Summary and Discussion

It is proposed that the two major magmatic complexes and three lines of diatreme breccia pipes found in the East Arm are all clearly related both genetically and temporally. Each intrusion is morphologically different and hence represents different levels and modes of emplacement of a similar alkaline and highly volatile parent magma. These intrusions are characterised by gas-brecciation in their upper levels, progressively more severe soda metasomatism in later phases and an enrichment in rare earth elements and uranium in the later differentiates. As well as the volatiles themselves which escaped from the upper levels, three characteristic volatile magmas were able to escape from the parent intrusions. One gave rise to feldspar-amphibole pegmatites, the second to amphibole-apatite-magnetite pegmatites rich in rare earth elements and the third to fine-grained albite dykes and veins here called bostonites. The breccia pipes are thought to overlie magmatic complexes similar to the Easter Island dyke.

The variation along the Easter Island Dyke has been shown to be an originally vertical feature (Fig.2). The dyke varies from melanocratic syenogabbro at its base through albite syenite and virtually pure albitites and bostonites at its roof. This variation has been ascribed to the process of differentiation during crystallisation and is generally similar to that in the very much larger Maple-Blachford Lake Complex. The main intrusive phase of the 'Maple' (Caribou Lake) Complex is best defined chemically as an alkaline gabbro and petrologically as syenogabbro. The late differentiates are more alkalic and syenitic in nature. Although it is cut by more felsic alkaline rocks of the main Blachford Lake Complex it is thought that all these rocks are related and represent intrusion of progressively more differentiated magmas in what should perhaps be considered as a sub-circular alkaline complex. Thus, the main intrusions are best classified as alkali syenites and are clearly similar to the camptonite and gabbroic associations of the Oslo Graben (Brogger, 1894) or to the alkaline complexes of the St. Lawrence Graben (Phillpotts, 1974).

The alkaline magma was clearly emplaced up splaying branches of the MacDonald fault system. Later movements have probably destroyed other alkaline complexes but the presence of breccia pipes and mineralisations indicate that many of the main splay systems contained, or still contain at depth, alkaline igneous rocks. The time of emplacement was therefore clearly one of major fault movements. The 2200 m.y. ages for the Easter Island Dyke are K-Ar ages and may not be thoroughly reliable because of the pervasive metasomatism of the rocks. The 2057 ± 56 m.y. age for the Blachford Lake Complex granite is also a K-Ar age and thus can only provide a minimum age for the complex. The 'Maple' (Caribou Lake) Complex must therefore be older than this. It is reasonable to suppose that these rocks were emplaced in the 2200 to 2100 m.y. interval. Furthermore the Big Spruce Lake alkalic complex, north of Yellowknife is a complex whose parent magma seems to have been remarkably similar to that of the East Arm rocks and has an age of 2111 ± 40 m.y. (Martineau and Lambert, 1974). It is proposed here that these rocks were all formed at the same time and define a major period of faulting and alkaline magmatism in and around the East Arm in the Lower Great Slave Supergroup times (Table 1). While the magmatism is petrochemically similar to that in major rift-valleys of the world, this fault pattern and history are quite distinct from those bounding such rifts. It is interesting to note further the proposal of Irving and McGlynn (1976) that there was a major period of extension throughout Canada at about 2100 m.y. following the deposition of Huronion Supergroup.

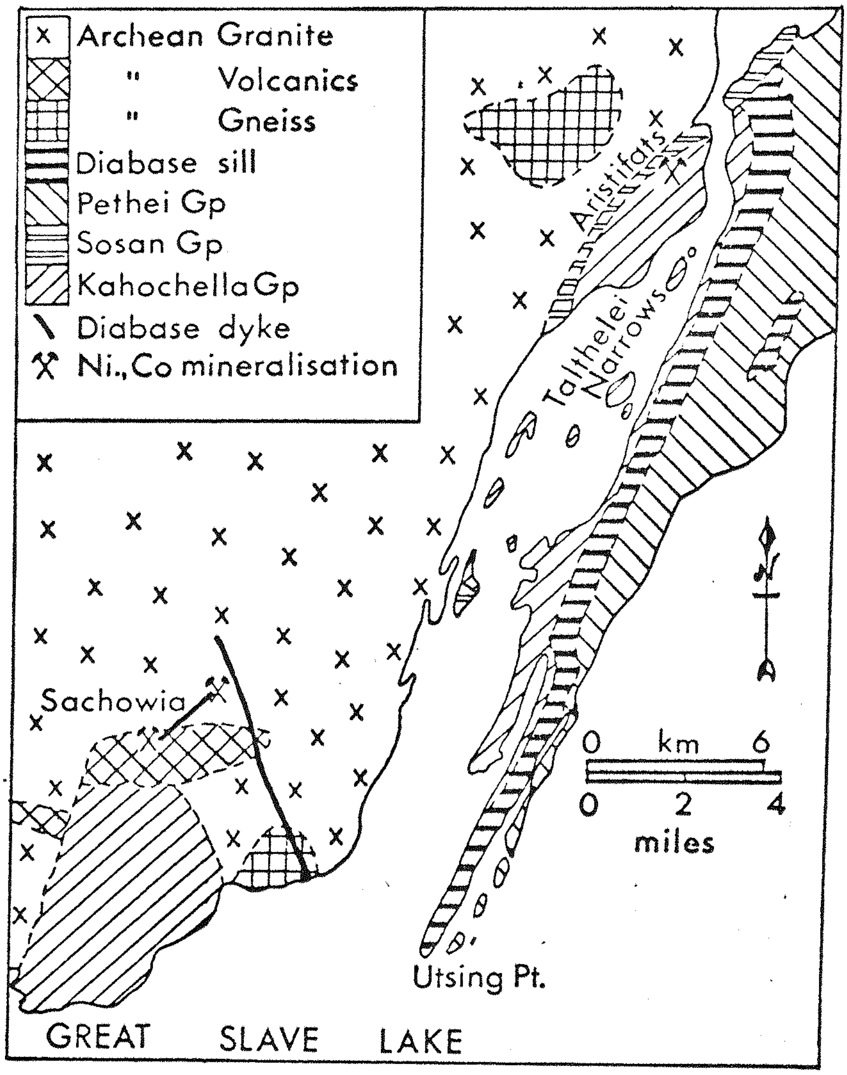
## Group B. THE MIDDLE APHEBIAN MAGMATIC ROCKS

### i. General Geology

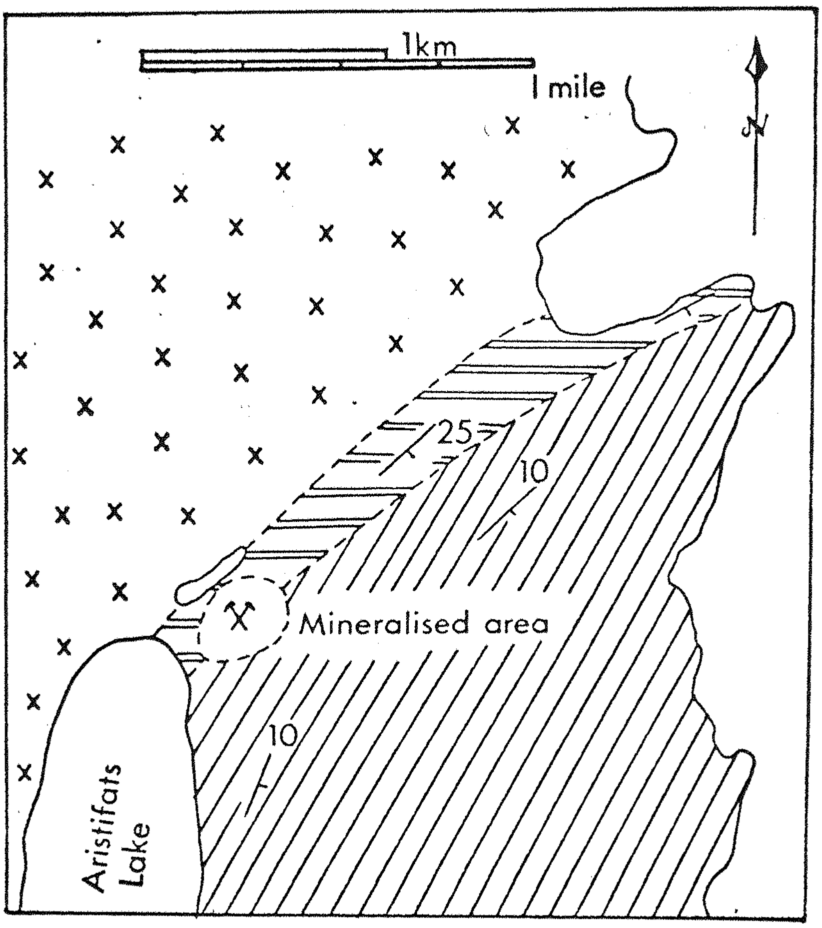
Although igneous activity occurred intermittently throughout Great Slave Supergroup times (Table 1), the volcanics of the Seton Formation represent a period of particularly widespread and active magmatism. The volcanic rocks occur in all formations between the Duhamel and McLeod Bay but are particularly concentrated at the junction between the Sosan and Kahochella Groups. The Seton Formation should not be considered a stratigraphic division within the Great Slave Supergroup but should now refer collectively to all volcanic rocks in the Sosan and Kahochella. Since Hoffman's classification could lead to confusion, it is felt that his term 'Seton Formation' as such is now superfluous. Volcanic rocks on Seton Island the type section have been dated at 1872 ± 10 m.y. (Baadsgaard et al., 1973), although Olade (1975) suggests that this is the age of metasomatism and that the rocks might be slightly older.

### ii. Petrography/Petrology

The volcanic and associated hypabyssal rocks are predominantly mafic, but silicic centres are locally important (Hoffman et al., 1977). The rocks are chemically alkaline (Olade and Morton, 1972) but Olade (1975) speculated that they were originally subalkaline and have suffered soda metasomatism. Some of the hypabyssal rocks are albite granophyres, albite trachytes (Olade and Morton, 1972) and albite porphyries (Hoffman et al., 1977). This suggests that the soda metasomatism is genetically and intimately related to the magmatic process and is not fortuitous. It is important to remember that soda-metasomatism and albite-rich differentiates characteristic of the older intrusive rock. Hoffman et al., (1977) have documented a large number of plugs, pipes and agglomerate filled vents around Seton Island, Snowdrift and Aristifats Lake (Map 4). It has been noted that the pipes on the northwest side of Talthelei Narrows are mineralised and that more southerly pipes are preserved at higher levels and might carry mineralisation at depth.




**Map 10: Aristifats Lake and Sachowia Lake areas -**
Geology and Mineralisation.




**Map 11: Aristifats Lake area** - Geology and Mineralisation.

A small vent on Basile Peninsula (Map 4) cuts shales and tuffs of the Gibralter Formation and is associated with pillow lavas and breccia. The vent-agglomerate consists of rounded blocks of various hydrothermally cemented volcanic rocks and tuffs with Fe-Cu sulphides. The centre of the vent has been intensely altered to a chloride-carbonate assemblage with up to 5% sulphides (pyrite and chalcopyrite). These carbonates are made up of dolomite with some calcite and siderite. In Hornby Channel, near Basile Peninsula (Map 4), a plug of albite porphyry contains specks and veinlets of apatite and fluorite. This plug and its host rocks of Kahochella Group age are cut by an ENE-trending diabase dyke. At Aristifats Lake (Map 11) there is a more deeply eroded vent cutting Archean rocks and the Lower Akaitcho River Formation. The vent contains fragments of granite, greenstone, red calcareous and hematitic shales and mafic volcanic rock in a matrix of rock flour and yellow carbonates. Towards the centre the whole assemblage is progressively and concentrically altered to chlorite, then sericite, and finally to carbonates (Plate 1 H). Most of these fine to coarse-grained carbonates are made up of dolomite with some calcite and siderite. Some of the rock fragments and carbonates were cut by quartz and later carbonate veins. Oolites filled with carbonates, chlorite, sericite, iron oxide (hematite) and siliceous materials (chalcedony and chert) are common. It was observed that the feldspars (mostly albite) of the rock fragments were severely altered. In the central zones there are significant showings of Ni-Co arsenide mineralisation in addition to the more common Cu-Fe sulphides and oxides.

### iii. Summary

These brief descriptions of the petrology of the Seton volcanics indicate that they are also quite closely related to the earlier intrusions, but their mode of emplacement is different. This similarity is chiefly due to the presence of alkaline albite-rich differentiates and soda metasomatism which are characteristics of the early Aphebian intrusions. This is supported by the presence of Ni-Co arsenide mineralisation which closely akin to that in the Easter Island Dyke and will be discussed in detail in Part II of this thesis.

## Group C. THE LATE APHEBIAN MAGMATIC ROCKS

### i. General Geology

A number of plugs (or stocks) and laccoliths of calc-alkaline diorites were intruded at shallow levels into the Great Slave Supergroup times (Map 4). They outcrop intermittently from Îles Basses in the west to Meridian Lake in the east (a distance of some 220 km) and are essentially confined to the line that acted as a boundary between shelf and basin facies in Upper Great Slave Supergroup times (Hoffman, 1969, 1973, 1977). These intrusions were emplaced up the principal faults of the East Arm and have been dated at 1845, 1795, 1785 and 1630 m.y. (Hoffman, 1969). These are K-Ar ages and thus not thoroughly satisfactory. However an age of 1790 m.y. is accepted by most workers as reasonable and consistent with other data.

The intrusions were restrained in small plugs or stocks in older rocks but expanded into large laccoliths where they were emplaced at the boundary between the Pethei and Christie Bay Groups, particularly in the Stark Formation (Table 1). The uppermost formations of the Pethei Group are predominantly well-bedded carbonates with minor argillaceous and calc-argillaceous intercalations. The Stark Formation consists of red mudstone and stromatolitic carbonates but these are frequently chaotically mixed in breccias that are thought to be principally the result of a collapse following dissolution of a thick evaporite sequence (Badham, 1975; Badham and Stanworth, 1977; Hoffman et al., 1977). It was proposed by Badham (1975) and Badham and Stanworth (1977) that the intrusions expanded easily into the consequent zone of solution collapse and they may even have enhanced the rate and extent of dissolution. However no intrusion is seen in rocks younger than the Stark Formation. Intrusions which failed to reach the top of the Pethei Group are usually stocks, not laccoliths, although some expansion into other formations, particularly of the Kahochella Group, can be seen, for example in Northern Hornby Channel. Badham (In Press) proposed that the diorites must post-date the Christie Bay Group. In addition Hoffman et al., (1977) consider that the intrusions post-date nappes which in turn post-date the Christie Bay Group.

The tops and margins of these intrusions are most complicated. The tops consist of a series of progressively more differentiated rocks that successively intruded earlier phases. Contacts are usually nebulous and gradational, although the latest differentiates (finer-grained trachytic red felsite dykes and veinlets) often cut earlier rocks sharply, but with no chill. There are also breccias of early-crystallised material cemented by red felsite. It is clearly observed that the tops are pervasively altered to a carbonate-sericite-hematite assemblage, which alteration extends out into the country rocks. However the actual contacts consist of a zone of igneous and sedimentary rock fragments in a red felsite matrix. These zones vary between 10 cm and 30 metres in width. The proportion of igneous matrix decreases outwards and carbonate becomes the predominant cement. The presence of plutonic rocks in the breccias is thought to be due to the rising up of the main body of igneous rock through the explosion-breccias which were modified and cemented by late differentiates and altered by hydrothermal solutions (Badham, 1975, and In Press). Contact metamorphic aureoles are extremely narrow and hornblende-hornfels facies is only seen directly beneath the laccoliths.

The laccoliths have flat bottoms and steep sides. The bases are often concordant over a few tens of metres but are broadly slightly discordant. There has been little disturbance of underlying strata and contact metamorphism is restricted to a few metres or tens of metres. The actual roofs of the intrusions are generally eroded but may be present on the Îles Basses and on West Caribou Island (Map 4).

ii. Petrography and Petrology

The intrusions are mineralogically and texturally highly variable. They consist predominantly of medium-grained, porphyritic diorite (Plate 2). The phenocrysts are predominantly plagioclases and hornblende in a matrix of alkali feldspars and minor quartz. The plagioclases are usually oligoclase to andesine, but are commonly overgrown by albite and by microcline with patch perthites of albite. The rocks are usually hydrothermally altered such that feldspars are replaced by sericite and carbonate, and hornblende by chlorite, hematite and carbonate. In the west andesine-hornblende diorite is the most common phase. The quartz content is generally about 10% or less but there are a few areas, with gradational contacts with the main phase, where quartz and biotite are abundant. To the east, diorites still constitute the bulk of the intrusions but the amount of alkali feldspar replacement and the number of quartz-biotite bearing areas increase progressively. The latter are later pulses of the diorites, as can be seen particularly at Hornby Channel and MacLean Bay where the emplacement of these later phases was concentrated towards the centre and tops of the intrusions. These later phases are mineralogically different, but seem to be chemically similar (see Table 4) suggesting that they are not differentiates but are products of crystallisation under differing conditions (presumably of P Total and $PH_2O$).

For the partial chemical analyses, samples of all the main diorites were taken (see the accompanying maps for all the locations). Sample 1 (Table 4) is from East Caribou Island. It is the south-western contact of laccolith with Pethei Group carbonates contact breccia. This brown-stained breccia consists of fine fragments of coarse and fine diorite and carbonate (dolomite and calcite) in comminuted carbonate impregnated matrix. The diorite fragments consist of intergrowths of plagioclases (An 20-30) overgrown by albite with minor chlorite and some microcrystalline quartz (up to 15%).

Sample 2 is the contact phase from Blanchet Island. It is the southern contact of laccolith with Pethei Group calcargillites. The diorite consists of fine-grained laths of plagioclase (An 20-30) almost completely overgrown by albite with interstitial chlorite, quartz, carbonates, hematite and magnetite. In some samples of the contact phase the amount of carbonates, mostly dolomite and calcite, increased up to about 90% of the whole rock. The main phase of the intrusion (Sample 3) which is 200 metres of southern contact consists of plagioclases (An 25-30) overgrown by albite, microcline and some orthoclase. Total feldspar is 50-60%, but partially replaced by carbonates and sericite. Relict phenocrysts (about 10%) of hornblende, actinolite, pyroxene and biotite are altered to chlorite, hematite and carbonates (mostly dolomite or calcite with some siderite). Minor sphene, epidote, apatite, zircon and monazite are observed. The matrix (30%) is made up of alkali feldspar, probably albite, and quartz. However in some samples the microcrystalline and overgrown quartz constitutes about 15-20% of the rock.

The sample of the north end of Hornby Channel (Sample 4) is rather dark, fine-grained phase from near the contact. It is highly hematitised and contains phenocrysts of plagioclase overgrown by albite microcline and some orthoclase (a total of about 60% feldspars). Most of the hornblende, actinolite and biotite are altered to chlorite hematite and carbonates, while the feldspars are altered to chlorite and carbonates. Interstitial quartz (up to 10%) and minor apatite are observed with traces of monazite and zircon. Another sample of the north end of Hornby Channel (Sample 5) is the red trachytic phase which consists predominantly of laths of plagioclases (An 20-30) corroded and overgrown by alkali feldspars (albite and microcline) and some 15% euhedral quartz. Total feldspar is about 75%. Other minor minerals include chlorite, carbonates and apatite. The reddish colour of this trachytic phase is essentially due to hematite dusting. Sample 6 is taken from the northernmost island at north end of Hornby Channel. It is made up of phenocrysts of plagioclases (An 20-30) overgrown by albite and microcline in a matrix of alkali feldspars and quartz. These feldspars constitute about 60% of the rock. Chlorite, hematite, carbonates, sparse biotite after hornblende and minor apatite made up the rest of the rock.

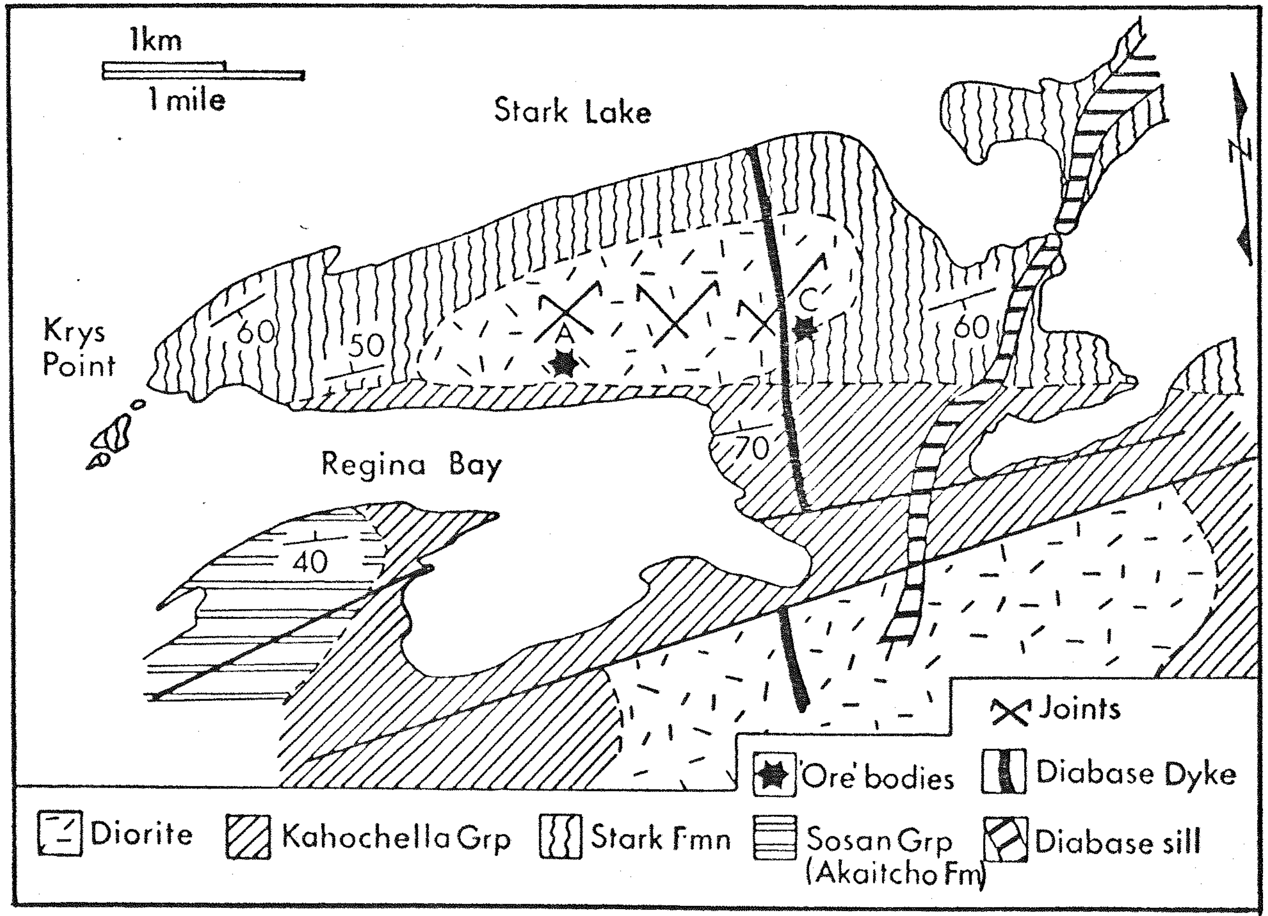
From the first island west of Western Labelle Peninsula the diorite (Sample 7) consists of the main phase. It is made up of 60% phenocrysts of plagioclases (An 20-30) overgrown by albite and microcline. The mafics (up to 15%) essentially consist of actinolite and biotite after hornblende and are strongly altered to chlorite, carbonates and hematite. However most of the feldspars are altered to sericite and carbonates. The matrix of alkali feldspar and minor quratz is badly altered. Small, rounded crystals of quartz constitute about 15-20% of the rock. Sample 8, which is from the westernmost island in the chain west of Labelle Peninsula, is petrographically similar to that of Sample 7. The Eastern Labelle Peninsula sample (Sample 9) is taken from the main phase of the intrusion and consists of broken phenocrysts of zoned plagioclase (An 30) overgrown by albite and microcline (45-50%). Interstitial subhedral quartz constitutes 30-35% of the rock. The 10% relicts of clinoamphibole (actinolite and hornblende) are altered to chlorite, carbonates biotite and hematite. The fine-grained feldspathic matrix has hematitic stain.

However minor apatite, monazite and zircon are observed. Locally in some other samples from the Eastern Labelle Peninsula, allanite is also found. Nevertheless the diorite of the Western Labelle Peninsula is similar to that of the Eastern Labelle.

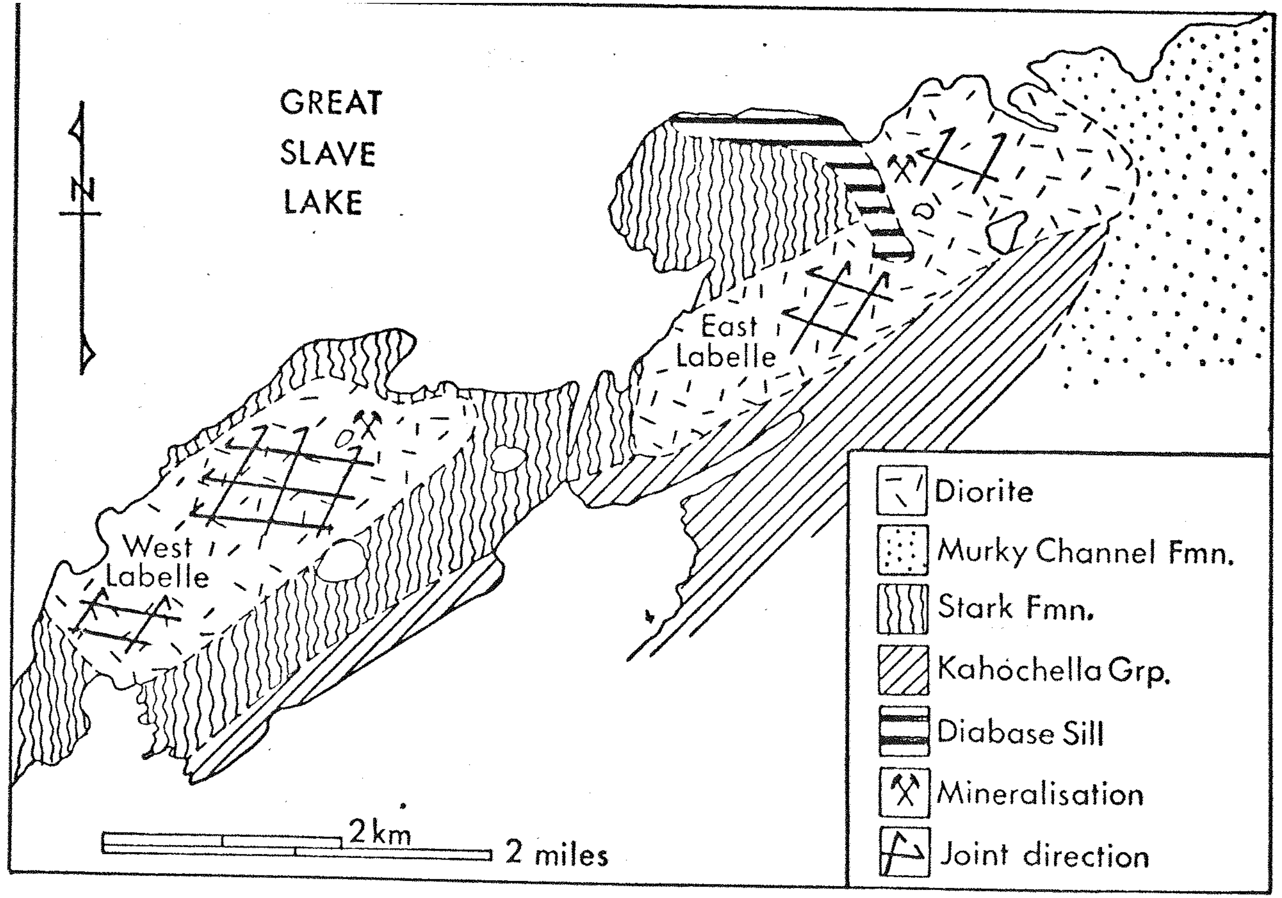
From the shoreline north of Murky Lake and west of Snowdrift the diorite (Sample 10) is petrographically similar to that of Sample 9. Sample 11 and sample 12 were taken progressively eastwards from Sample 10. Both of these are similar to Sample 9 but Sample 12 contains less quartz (about 20-25%) and has better preserved clinoamphibole (especially actinolite), with biotite as an important alteration product.

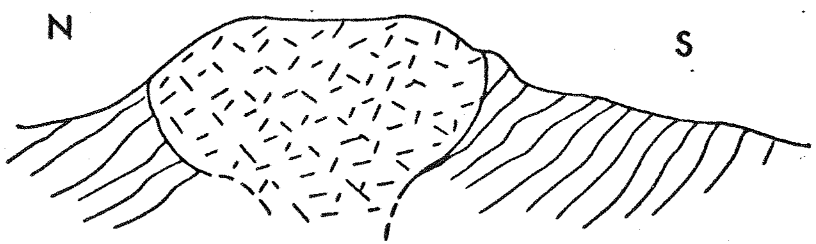
At MacLean Bay there are two phases observed, that is a more mafic phase (Sample 13) and a more felsic central phase (Sample 14). The mafic phase contains about 60% feldspars, mostly microcline and albite with few relict sodic plagioclase laths. Some of these coarse feldspars are zoned. The mafics (15-20%) are made up of relict hornblende and actinolite. Quartz is quite abundant (up to 20%) and this includes large overgrown crystals poikilitically enclosing feldspars and amphiboles. The alteration products include carbonates, chlorite and hematite. Minor apatite, monazite and zircon are also observed. However, the felsic phase is similar to the mafic phase, except that the feldspars are larger and quartz smaller. Nevertheless their proportions are the same. Biotite is the dominant mineral (about 10%). Here the small fractures are healed with carbonates and perhaps minor analcime.

Sample 15 is the main phase near southern contact from Regina Bay diorite stock. It contains small phenocrysts of plagioclases (An 20-30) overgrown by microcline, orthoclase, and then albite (50%), in a matrix of fine-grained quartz and alkali feldspar. Quartz (15-20%) is often found as small phenocrysts. Relict, altered clinoamphiboles, mainly actinolite with some hornblende constitute 15% of the rock. The alteration products such as carbonates (mainly calcite and dolomite), chlorite and hematite are not uncommon. Here apatite (carbonate - and chloroapatite), sphene, zircon and monazite are quite abundant compared

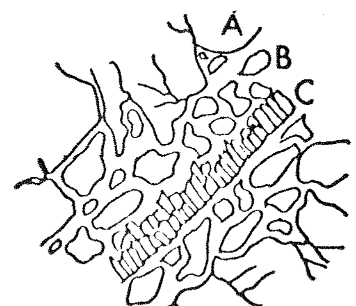



Map 12: Regina Bay area - Geology and Mineralisation.



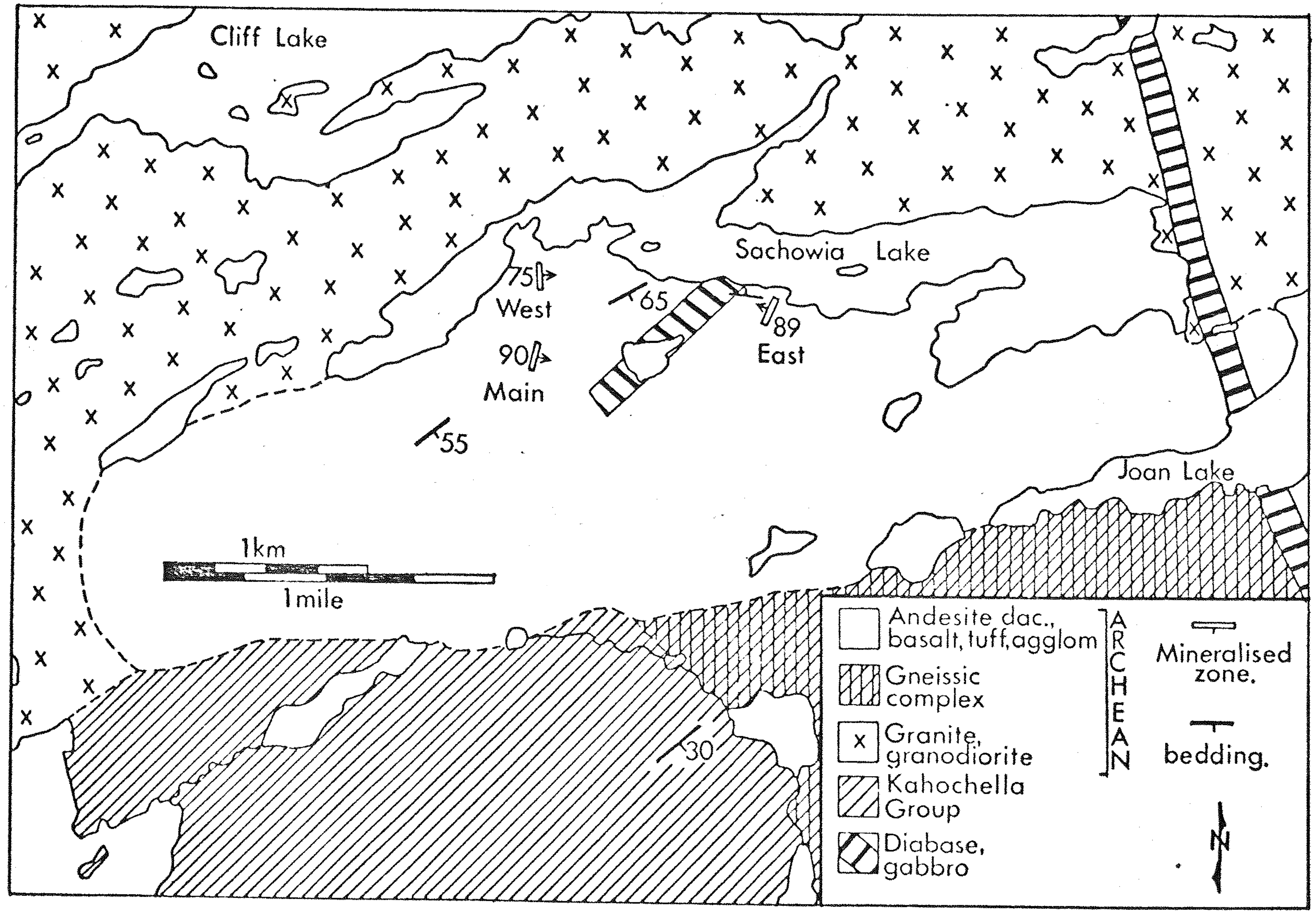

Sketch Section through Western Labelle Peninsula



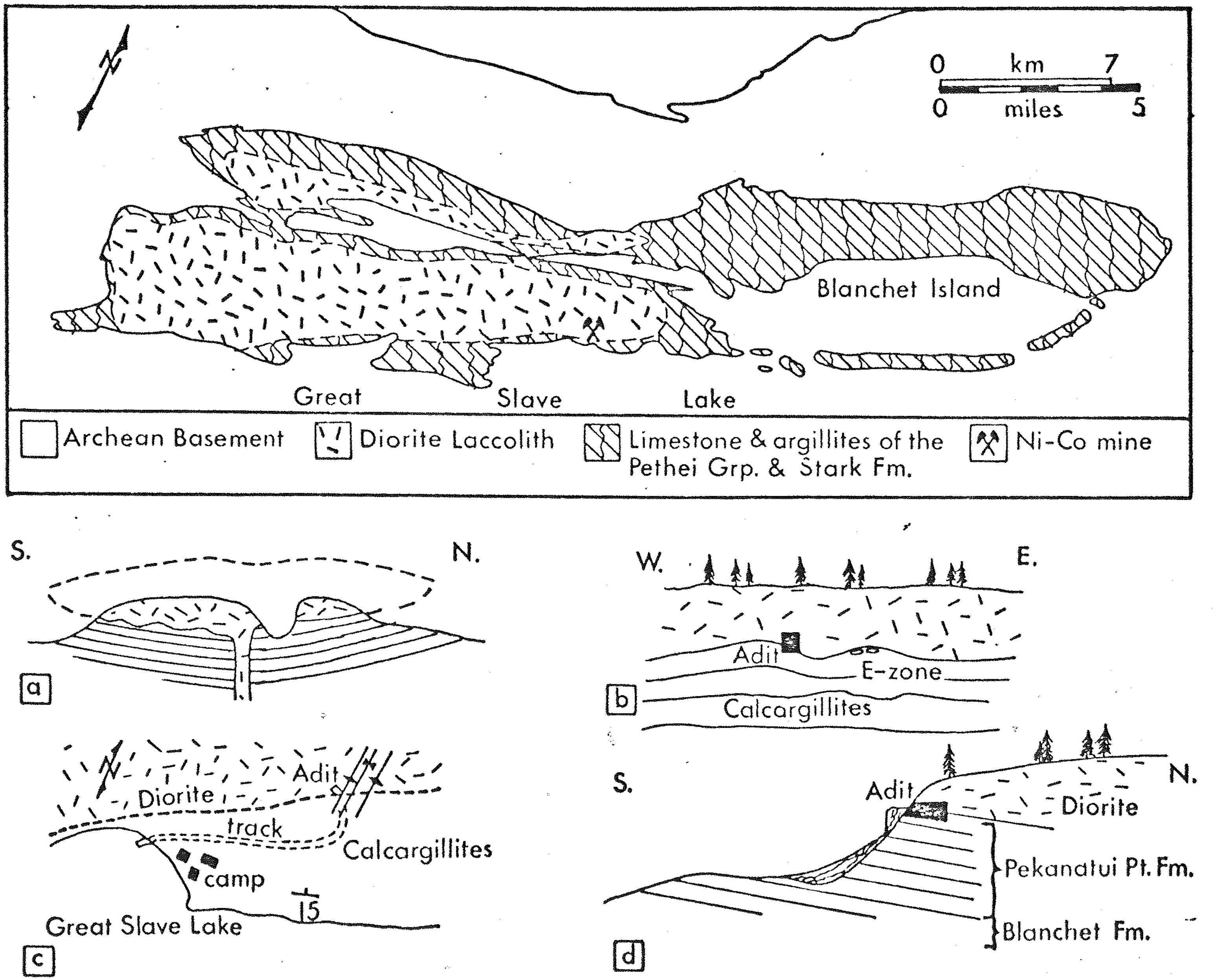

Plan of the E. Labelle vein

2 m.

**Map 13: West Labelle and East Labelle areas** - Geology and Mineralisation.




**Map 14: Sachowia Lake area** - Geology and Mineralisation.




**Map 15: Blanchet Island area** - Geology and Mineralisation

PLATE 2

A) Feldspars replaced by actinolite (act.), later altered to chlorite (chl.). Some biotite (bi.) and opaques (op.) have replaced the above minerals. West Labelle diorite. P.P.L.

B) Late monazite (mon.), zircon(zr.), opaques (op.), biotite (bi.) and chlorite (chl.) after intergrown quartz (qtz.) and albite (ab.). Albite is also altered to sericite (ser.). Regina Bay diorite. P.P.L.

C) Quartz (qtz.) with relief inclusions of orthoclase (orcl.), calcite (calc.), chlorite (chl.) and minor opaques. At the quartz rims actinolite (act.) is strongly altered to chlorite (chl.) and opaques (op.). Regina Bay diorite. P.P.L.

D) Late calcite, opaques (uraninite, coffinite, hematite and goethite) and chlorite (penninite) replacing quartz (qtz.) - feldspars (fsp.) intergrowth. Regina Bay diorite. P.P.L.

E) Large euhedral remnant of biotite (bi.) altered to chlorite (chl.) and epidote, and calcite in a matrix of quartz (qtz.) - feldspars (ab. and orcl.) intergrowth. Most of the feldspars are sericitised and also replaced by secondary calcite (calc.). East Labelle diorite. P.P.L.

F) Coarse, cleaved, acicular actinolite (act.) intergrown with and also partly replacing albite (ab.). Secondary chlorite (chl.) and opaques (op.) have replaced all the above minerals. East Labelle diorite. P.P.L.

G) Remnant of coarse, euhedral biotite (bi.) with feldspar inclusions (ab.) intergrown with albite and anhedral quartz (qtz.). East Labelle diorite. P.P.L.

H) Zoned euhedral albite (ab.) - highly sericitised, chloritised and carbonatised, intergrown with euhedral quartz (qtz.). East Labelle diorite. P.P.L.

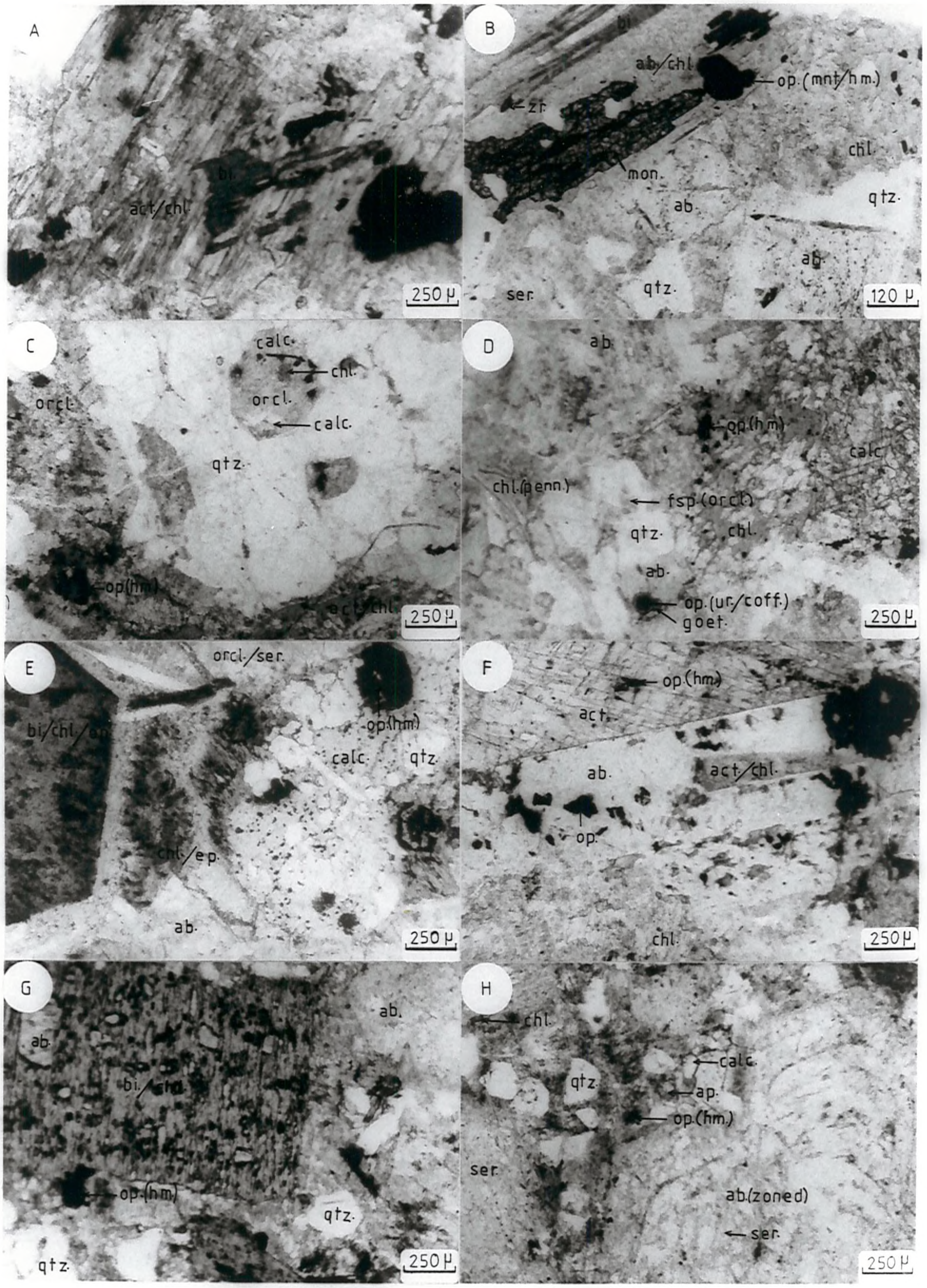

A
act/chl
bi
250 μ
B
bi
zr
ab/chl
op.(mnt/hm.)
mon.
chl.
ab.
qtz.
ab.
ser.
qtz.
120 μ
C
calc.
chl.
orcl.
orcl.
calc.
qtz.
op(hm)
act/chl
250 μ
D
ab
op.(hm)
calc
chl.(penn.)
fsp.(orcl.)
qtz.
chl.
ab.
op.(ur/coff.)
goet.
250 μ
E
orcl./ser.
op.(hm)
calc.
qtz.
chl./ep.
ab.
250 μ
F
op.(hm)
act.
ab.
act/chl.
op.
chl.
250 μ
G
ab.
ab.
op.(hm)
qtz.
qtz.
250 μ
H
chl.
calc.
qtz.
ap.
op.(hm.)
ser.
ab.(zoned)
ser.
250 μ

PLATE 2

to other diorites in the East Arm. However, the main phase from the centre of the intrusion (Sample 16) contains strongly hematitised intergrown plates of albite and microcline with very rare relict plagioclase laths (60%). Large relict phenocrysts of clinoamphiboles (up to 20%), mainly actinolite and some hornblende, are quite common. Quartz crystals are sparse. Apatite, sphene, zircon and monazite are also common. Alterations of the primary minerals are however similar to those of other diorites but are not really severe.

The eastern Stark Lake diorite sample (Sample 17) is the main phase with biotite (nearby main phase without biotite was not analysed). The diorite consists of large laths of plagioclase, An 20-30 with rims of An 10, partially overgrown by microcline and albite (65%). In addition, small phenocrysts of biotite (15%) with some actinolite and hornblende are altered to carbonates, chlorites and hematite, and surrounded by small crystals of zircon, apatite, monazite and also allanite. Interstitial, subhedral quartz (15%) often found with some fine matrix alkali feldspars. Closely related to these diorites is the felsite dyke in the Sachowia area. Sample 18 is the centre of this felsite dyke. It is predominantly consisting of fine-grained intergrowths of approximately equal amounts of quartz, plagioclase (An 30) and alkali feldspars (albite and microcline) which are strongly hematitised and altered to carbonates (dolomite, calcite and siderite). Here there are no mafic minerals observed. However it is quite similar to the contact phases of other dioritic intrusions elsewhere in the East Arm. On the other hand Sample 19 is the fine, red felsite of the same area. It contains fine-grained, interlocking mass of quartz and hematite-dusted feldspar (mostly albite) with numerous small apatite crystals. This red felsite is also quite similar to the contact phases of other intrusions.

### iii. Petrochemistry

A total of 19 samples (including two samples of Sachowia felsite) were analysed on the basis of homogeneity and lack of weathering. These include most of the diorite stocks and laccoliths which outcrop in the East Arm area. The analyses are presented in Table 4 in such a way that the distance (D) of each sample is eastward from locality 1

| Sample | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 | 14 | 15 | 16 | 17 | 18 | 19 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| D miles | 0 | 26.8 | 26.8 | 54.8 | 54.8 | 56.8 | 64.8 | 64.0 | 71.2 | 75.2 | 78.8 | 84.4 | 88.4 | 88.4 | 99.2 | 99.6 | 113.2 | - | - |
| % | | | | | | | | | | | | | | | | | | | |
| $SiO_2$ | 48.0 | 65.9 | 54.5 | 52.2 | 58.3 | 63.0 | 60.2 | 63.5 | 65.1 | 64.0 | 65.6 | 65.9 | 64.0 | 62.6 | 61.6 | 64.9 | 64.2 | 71.0 | 61.7 |
| $TiO_2$ | 0.4 | 0.4 | 0.7 | 0.9 | 0.8 | 0.6 | 0.6 | 0.7 | 0.7 | 0.6 | 0.6 | 0.6 | 0.6 | 0.7 | 0.6 | 0.6 | 0.6 | 0.2 | 1.9 |
| $Al_2O_3$ | 8.6 | 19.5 | 18.1 | 17.8 | 19.4 | 18.4 | 17.7 | 17.2 | 17.7 | 17.9 | 17.6 | 17.6 | 17.3 | 16.9 | 16.6 | 17.2 | 17.2 | 21.0 | 17.2 |
| FeO | 1.8 | 0.7 | 6.8 | 9.4 | 4.8 | 2.0 | 4.2 | 5.4 | 3.7 | 4.4 | 4.2 | 2.1 | 3.0 | 3.0 | 3.5 | 5.2 | 4.8 | 1.1 | 2.6 |
| MgO | 0.9 | 1.0 | 4.5 | 4.5 | 2.4 | 1.4 | 1.6 | 3.4 | 3.9 | 2.6 | 2.6 | 2.7 | 2.6 | 3.2 | 3.0 | 2.9 | 3.0 | 0.6 | 1.7 |
| CaO | 26.2 | 2.4 | 3.1 | 1.2 | 3.7 | 2.8 | 5.8 | 1.5 | 2.1 | 1.6 | 3.1 | 2.8 | 5.5 | 5.1 | 3.4 | 2.4 | 2.7 | 0.8 | 4.0 |
| $Na_2O$ | 2.6 | 8.5 | 4.5 | 2.1 | 3.1 | 5.3 | 3.5 | 2.0 | 2.6 | 2.9 | 2.5 | 5.4 | 3.7 | 3.2 | 4.7 | 2.8 | 2.4 | 4.2 | 4.7 |
| $K_2O$ | 0.6 | 0.3 | 1.1 | 3.7 | 3.0 | 2.5 | 2.0 | 5.3 | 4.7 | 5.8 | 4.5 | 2.4 | 1.2 | 1.5 | 0.6 | 4.4 | 4.5 | 2.2 | 1.4 |
| $H_2O$ | 0.6 | 0.6 | 2.4 | 3.2 | 1.8 | 0.5 | 1.9 | 0.9 | 0.9 | 1.1 | 2.2 | 0.8 | 1.5 | 1.6 | 2.0 | 1.1 | 1.4 | 1.6 | 1.5 |
| $CO_2$ | 16.3 | 1.5 | 0.6 | 1.0 | 4.7 | 3.5 | 2.8 | 0.3 | 0.4 | 1.5 | 0.8 | 1.2 | 2.2 | 3.0 | 2.1 | 0.4 | 1.2 | 0.6 | 4.5 |
| Total | 106.0 | 100.8 | 96.3 | 96.0 | 102.0 | 100.0 | 100.3 | 101.2 | 101.8 | 102.4 | 103.7 | 101.5 | 101.6 | 100.8 | 98.1 | 101.8 | 102.0 | 103.3 | 102.2 |
| p.p.m. | | | | | | | | | | | | | | | | | | | |
| Rb | 21 | 8 | 20 | 146 | 96 | 49 | 77 | 238 | 122 | 160 | 130 | 52 | 38 | 53 | 12 | 191 | 158 | 69 | 44 |
| Sr | 112 | 30 | 142 | 60 | 53 | 58 | 45 | 185 | 237 | 113 | 203 | 126 | 240 | 130 | 123 | 248 | 198 | 117 | 35 |
| Ba | 160 | 95 | 331 | 694 | 144 | 216 | 203 | 973 | 1171 | 1468 | 1457 | 488 | 325 | 195 | 247 | 847 | 877 | nd | nd |

Table 4: Partial chemical analyses for samples of the East Arm diorites (1–17) and the Sachowia Lake felsites (18,19). The distance of each diorite sample east of sample 1 is shown. Total iron is presented as FeO. The high total for sample 1 arises from its high carbonate content which was not allowed for in the correction procedure.

Continued overleaf.

| Sample | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 | 14 | 15 | 16 | 17 | 18 | 19 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| p.p.m. continued | | | | | | | | | | | | | | | | | | | |
| Ce | 24 | 11 | 63 | 56 | 84 | 80 | 77 | 88 | 63 | 59 | 94 | 66 | 99 | 76 | 96 | 71 | 79 | 207 | 150 |
| La | 9 | nd | 42 | 43 | 55 | 53 | 63 | 65 | 38 | 27 | 49 | 7 | 60 | 59 | 56 | 56 | 59 | nd | nd |
| Zr | 63 | 333 | 140 | 165 | 215 | 324 | 117 | 186 | 196 | 197 | 184 | 212 | 190 | 199 | 214 | 177 | 185 | 121 | 171 |
| Y | 25 | 22 | 13 | 28 | 29 | 45 | 25 | 28 | 40 | 36 | 34 | 54 | 78 | 80 | 36 | 27 | 40 | 3 | 27 |
| Nb | 8 | 17 | 11 | 13 | 17 | 17 | 13 | 12 | 16 | 14 | 11 | 17 | 12 | 14 | 27 | 16 | 12 | 9 | 11 |
| V | 48 | 20 | 105 | 154 | 109 | 57 | 80 | 84 | 95 | 97 | 95 | 92 | 78 | 76 | 66 | 82 | 82 | 16 | 195 |
| Cr | 26 | 2 | 17 | 35 | 16 | 6 | 47 | 49 | 47 | 44 | 49 | 48 | 45 | 50 | 46 | 44 | 46 | nd | 9 |
| Mn | 2339 | 51 | 609 | 1006 | 633 | 617 | 594 | 631 | 316 | 466 | 367 | 143 | 199 | 274 | 308 | 624 | 404 | 139 | 666 |
| Ni | 4 | 5 | 16 | 22 | 19 | 4 | 17 | 32 | 27 | 22 | 20 | 16 | 22 | 43 | 16 | 25 | 25 | 150 | 95 |
| Cu | 2 | nd | 3 | 2 | 1 | 6 | 11 | 6 | 9 | nd | 4 | 15 | nd | 6 | 6 | 17 | 25 | 2 | 5 |
| Zn | nd | nd | 81 | 60 | nd | nd | 32 | 75 | 38 | 132 | 49 | 3 | 9 | 12 | 30 | 36 | 12 | nd | nd |
| Ga | 8 | 18 | 16 | 21 | 21 | 20 | 18 | 17 | 18 | 20 | 21 | 19 | 19 | 15 | 19 | 19 | 16 | 18 | 16 |
| Mo | nd | nd | nd | 4 | nd | nd | nd | nd | nd | nd | 4 | 4 | 7 | nd | 5 | nd | nd | 9 | nd |
| Pb | 1 | 2 | 6 | 2 | 3 | 3 | 10 | 12 | 12 | 11 | 13 | 26 | 7 | 9 | 8 | 17 | 8 | 5 | 4 |
| As | 5 | 3 | 2 | 3 | 1 | 9 | 5 | 4 | 2 | 6 | 2 | 3 | 3 | 2 | 1 | 5 | 3 | 480 | 414 |

Table 4: Continued from previous page. Partial chemical analyses for samples of the East Arm diorites (1-17) and the Sachowia Lake felsites (18,19). The distance of each diorite sample east of sample 1 is shown. Total iron is presented as FeO. The high total for sample 1 arises from its high carbonate content which was not allowed for in the correction procedure.

(Sample 1) on the Caribou Islands. It was hoped that Sample 1 would represent the border phase. However, it is immediately apparent that it is chemically aberrant and is highly contaminated by carbonates (mainly calcite and dolomite), so the sample is not considered further. The analyses for Samples 2 - 17 show a remarkable consistency and the general lack of variation in the analyses is clear. Sample 2 is anomalous for soda and alumina of the major oxides and 7 out of 28 elements in total are aberrant. This sample is a contact phase and was taken near the Blanchet Island arsenide mineralisation. Sample 4 is rather anomalous in having low silica and high iron, titanium, vanadium and manganese. These anomalies reflect the rather mafic nature of the sample. Sample 5 from the more normal parts of the same intrusion is not anomalous. Of other samples, none has any major element anomaly and five have no anomalies at all. The number of anomalies is however small for the components. MgO, $K_2O$, $CO_2$, $H_2O$, Pb, Sr, Ba, Cu, Zn, Ga, Mo and Pb value are scattered and all these elements are characteristically mobile during alteration. That alteration is the prime cause of these variations as shown by the $CO_2$ and $H_2O$ variations and also by the close association in variability of $K_2O$, Rb and Ba. However, the most homogeneous components are $SiO_2$, $TiO_2$, $Na_2O$, La, Zn, Y, Nb, V, Mn and Ni. It has been suggested that Ti, Zr, Y, Nb and to a lesser extend La and V are immobile during alteration (Floyd and Winchester, 1975) and these data support the contention. The homogeneity of $Na_2O$ is surprising in the light of the normal mobility of sodium in hydrothermal systems and of the petrographic evidence for late magmatic alkali feldspar growth. Nevertheless, all the samples have been similarly affected.

Samples 13 and 14 from the MacLean Bay intrusion are closely chemically similar, though they are plagioclase-hornblende and plagioclase-quartz-biotite bearing respectively. This suggests that the differences are the result of crystallization under differing conditions (of P Total and $P_{H_2O}$) and not a consequence of differentiation. However, the quartz-biotite bearing phase (Sample 14) is the younger and was probably emplaced into the cooling earlier phase (Sample 13) after some uplift had taken place.

The diorites are shown to be calcalkaline and there is a range of composition. This is well explained by the plotted data on the standard AFM and alkalis-silica diagrams from Badham (In Press) (Fig. 3). From the plotted Harker diagrams (Badham, In Press) among the major elements there is a general decrease of FeO, MgO and $TiO_2$ with increasing silica, but neither $Al_2O_3$ nor the alkalis show any concomitant increase. Of the trace elements, Zr and Y vary directly with silica, but there are no other clear trends. Thus, in general there is little evidence that the variations are due to differentiation from a single source. Therefore, it is concluded that the individual intrusions are each slightly different, mainly due to the result of crystallisation. On the other hand, the diorites show remarkably little change over the entire 220 km sampled. This is shown by the data plotted against distance eastward from an arbitrary point (site of Sample 1) (Badham, In Press). However, there are slight increases in Zr and Y and a decrease in $TiO_2$ suggesting very small increases in alkalinity to the east. In conclusion, the only general changes up the length of the East Arm are mineralogical and this is observed to be the result of exposure at slightly higher levels in the intrusions in the east. There is no evidence for increasing alkalinity or for increasing differentiation to the east. To the author there is no such fundamental division of diorites only to the west of latitude $111^{o}30'$ and monzonites only to the east of this line, as was suggested by Hoffman et al. (1977).

Regarding the Sachowia felsites, Badham (1978) has proposed that they were probably the same age as the diorites because of their petrographic similarity to the contact phases of the diorites. The presence of quartz distinguishes them from bostonites and albitites that are closely related to the early Aphebian alkaline intrusions, such as the Easter Island Dyke. Analyses of two of these felsites (Samples 18 and 19) are consistent with the lack of mafics and the high feldspars and quartz contents. Sample 18 shows some evidence of having been differentiated from the main diorite composition and both samples clearly show the effects of nearby mineralisation in their high Ni and As values. Sample 18 is similar in many ways to the other felsitic contact phase such as that of Sample 2. On the other hand Sample 19 is generally similar to the other coarser grained diorites. Both felsites are calcalkaline. Therefore it is concluded

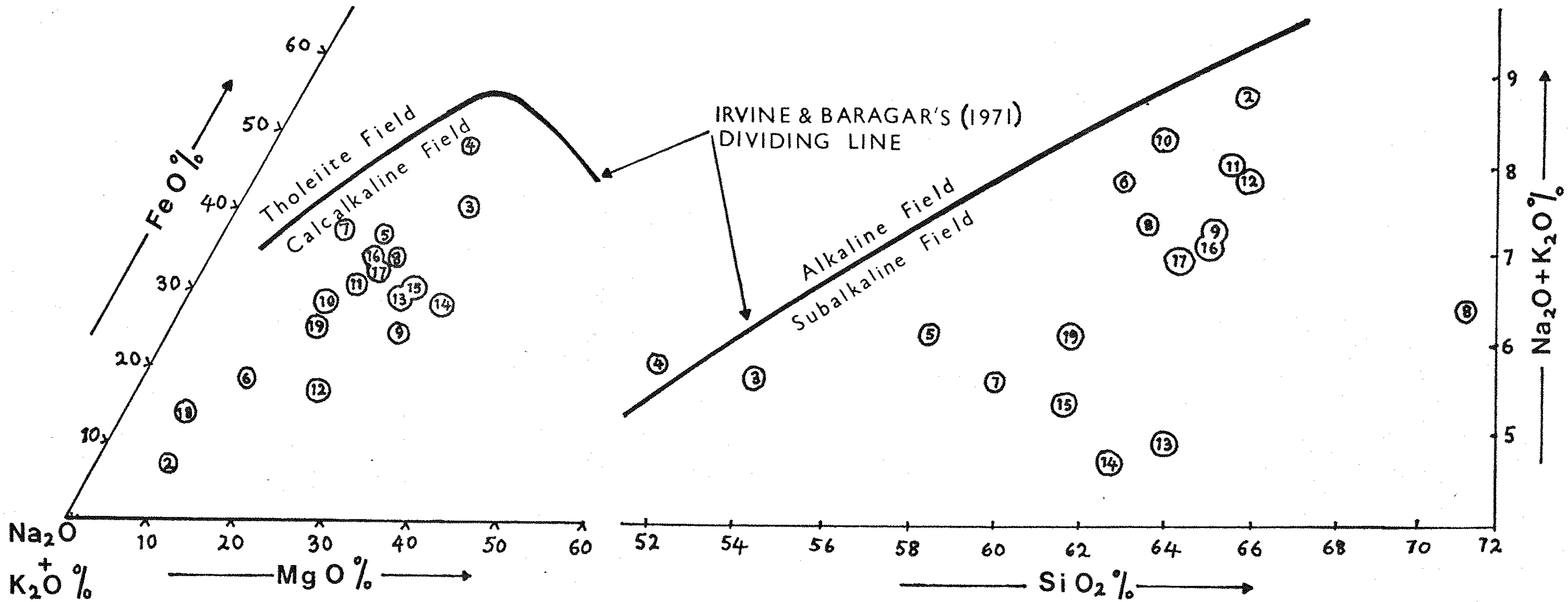



Fig. 3: AFM and Alkalis-Silica diagrams for the East Arm diorites (Samples 2 to 17); and Sachowia felsites (Samples 18 and 19).

that these felsites are chemically related to the diorites and are probably contemporaneous. It is thus not unreasonable to suggest that the Sachowia area is underlain by a diorite intrusion.

All the intrusions are mineralised with early veins, pegmatites and hydrothermal breccias of actinolite with apatite and subordinate magnetite and uranium minerals. A number of intrusions also have younger, but related mineralisations of Ni-Co-Fe-arsenides, sulpharsenides, silver, bismuth and sulphosalts. The details of these mineralisations will be presented in Part II of this thesis.

iv. Evolution of the Diorites

From the petrographic study and chemical analyes, it is suggested that the diorites emplaced in the East Arm area are evolved from the late differentiates of gabbro. The initial diorites essentially contain plagioclases (with some alkali feldspars), pyroxenes (augite ), hornblende, some biotite and are with or without quartz. As the cooling dioritic magma rises, most of the pyroxenes are altered to amphibole (mostly actinolite) and the plagioclases are altered to sodic feldspars, presumably albite. Thus, the amount of sodium increases while calcium decreases spontaneously. The diorites then became more sodic in nature. However at this stage, remnants of pyroxene and hornblende still existed. On rising up the late magma seems to have increased its water content and volatiles and as a result, hydrothermal mineralisations were emplaced in fractures and structural traps of the diorites. At this stage the diorites are altered to actinolite and alkali feldspars with albite, quartz and biotite, though some hornblende, plagioclases and remnants of pyroxenes still remained. This alteration is autohydrothermal. Chloritization, carbonatisation, sericitisation, silification, epidotisation and hematitisation are common. These alterations predate the actinolite veining but post-date the alkali feldspars replacements of plagioclases. Most of the feldspars are altered to carbonates and sericite, while the mafics are altered to chlorite, carbonates and hematite. Apatite, sphene, zircon, monazite and locally allanite are the late minerals to crystallize. The only differentiation from the main diorite composition is the evolution of the felsites. To summarise the evolution of these diorites, a model (Fig. 4) is presented here.

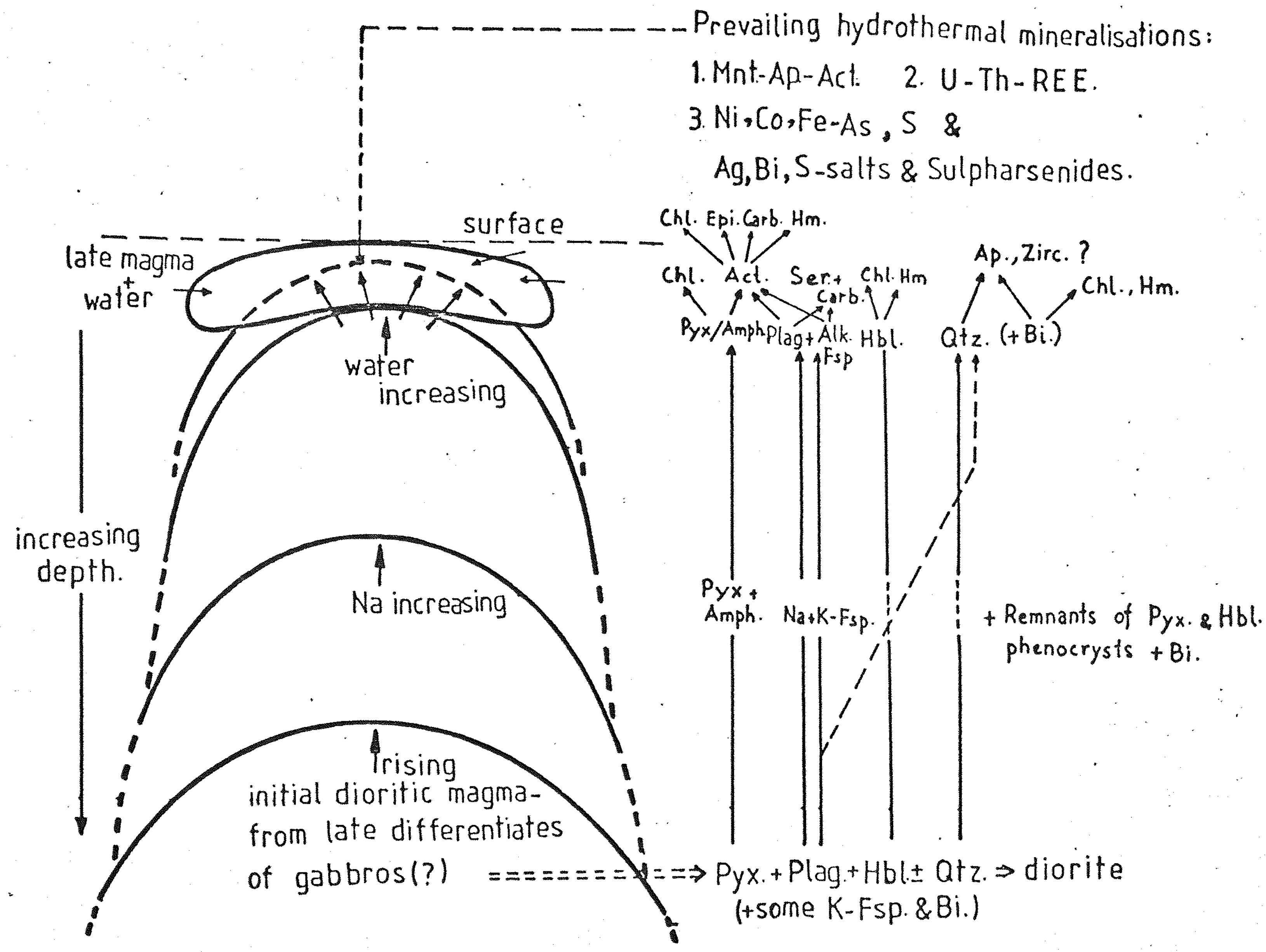



Fig. 4: A model of the evolution of calc-alkaline diorites.

## v. Summary and Discussion

Hoffman et al. (1977) pointed out the similarity between the East Arm diorites and the bulk of the plutonic rocks of the Great Bear Batholith, a similarity that is enhanced by comparison of the chemistries (Badham 1973 b and in this thesis). The comparison is important because both groups of intrusions are of closely similar age and may both have been generated above an east-dipping subduction zone (Badham, 1973 a, b.; Hoffman and McGlynn, 1978). The diorites are calcalkaline and they do resemble those of the Great Bear Batholith both chemically and temporally. Thus, if the subduction model is correct, then it must be concluded that these East Arm diorites are related in some way to subduction beneath the graben. However that the generation of these diorites was directly related to subduction is highly unlikely, both from consideration of the geometry of the process and the absence of potash zonation. From geometrical considerations it is improbable that a subducting plate of hydrous oceanic crust could have extended from the proposed trench (Badham, 1973 b) to the east end of the East Arm. It is found that similar igneous rocks are absent from other aulacogens that joined marginal oceans that were closed by subduction, for example Anadarko Trough, Montreal graben and Belt Trough (Burke and Dewey, 1973). This would suggest that continental margin subduction was not able to affect the aulacogens. However, intermediate calcalkaline melts may be derived by partial melting of hydrous basalt or upper mantle or could be derived by crustal anatexis. Here the possibility of crustal anatexis is simply ruled out owing to the evidence for a thin crust under the East Arm (Barr 1971) coupled with the clear lack of high regional heat flow in the area during the Aphebian times. Therefore it can only be concluded that the melts were mantle-derived and that their generation was related only indirectly to the subduction process in manner not yet well understood.

## 6. GENERAL SUMMARY

From the field and petrographic studies of the East Arm magmatic rocks, several generalisations can be made:

1) Three groups of Aphebian magmatic rocks are recognised. The early Aphebian intrusions are made up of two intrusive complexes (The Easter Island Dyke and Maple-Blachford Lake Complex) and three lines of diatreme breccia pipes (Vestor, Preble and Inconnu Channels). The middle Aphebian rocks are made up of Seton volcanics. The late Aphebian intrusions are dioritic stocks and laccoliths.
2) The early and middle Aphebian intrusions have been shown to be alkaline. The late Aphebian ones are calcalkaline. Locally the older intrusions are peralkaline. All the magmatic rocks are characterised by soda metasomatism and late-stage differentiation.
3) The igneous rocks and related rocks were emplaced in splay faults of the MacDonald transcurrent system, as a result of Aphebian magmatism throughout the East Arm. Therefore the emplacements are related to periods of transcurrent movements on the fault system and not to rifting sensu stricto.
4) The igneous rocks appear to be in three distinct age groups: between 2000 and 2200 m.y. old for the early Aphebian intrusions; 1872 m.y. old for the middle Aphebian Seton volcanic; and about 1790 m.y. old for the late Aphebian diorite intrusions. The older intrusions are of the same age as the similar Big Spruce Lake Complex and the younger ones are probably of the same age as the Great Bear Batholith.
5) The older rocks provide a minimum age for the lowest units of the Great Slave Supergroup, and imply that the Wilson Island Group is similar in age to the Lower Huronian. On the other hand, the emplacement of younger intrusions provides an age for the end of the Great Slave Supergroup times.
6) All these intrusions are closely associated with hydrothermal mineralisations in the East Arm which will be described in detail in Part II of this thesis.

Regarding the lines of diatreme breccia pipes several generalisations can be made:

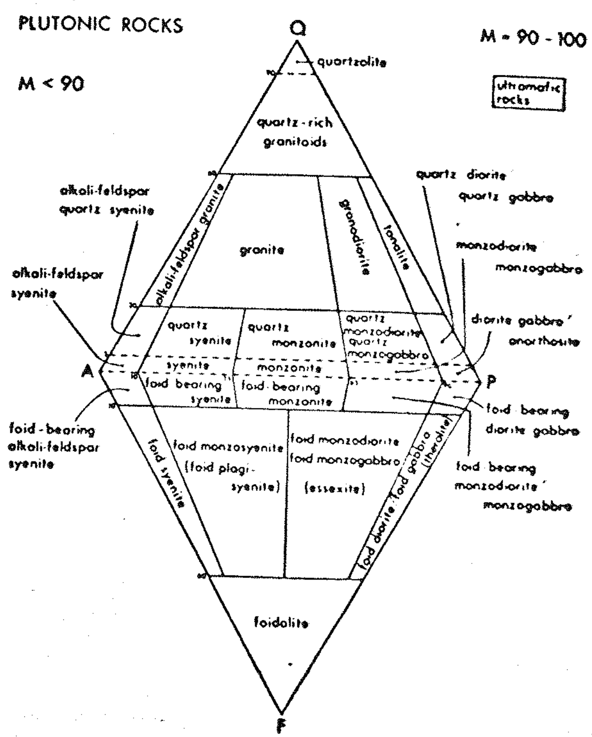
1) The pipes lie in zones along major faults.
2) The pipes have formed principally at the basement - Sosan Group unconformity.

3) Pipe fragments include basement and lower Sosan (Susanne) Group debris. No other sedimentary rocks were recognised.

4) The pipes are clearly the result of high-pressure gas escape. Much of the matrix consists of comminuted basement and sedimentary rocks.

5) Many of the pipes show two phases of emplacement and red felsite ( bostonite ) and green trachyte magmas were commonly introduced during the second phase. In some places breccia fragments include felsite. The felsite is closely similar to the marginal and upper level phases of the Easter Island intrusion and hence a genetic relationship is proposed.

From the detailed petrography of the older and younger intrusions of the East Arm, a summarised classification of these rocks presented in Fig. 5. It is a classification in the form of a QAPF double-triangle, from Streckeisen (1967). The diagram is based on modal composition cited beside the diagram, and it is the classification for plutonic rocks having less than 90% dark minerals. Alkali syenites (i.e. of the Easter Island Dyke and Blachford Lake Complex) are situated at the extreme end towards the alkali feldspar apex. On the other hand the calcalkaline diorites and gabbros are situated at the opposite end. This classification is made clear for the East Arm magmatic rocks.




*Q* quartz (or: silica minerals, mainly quartz);
*A* alkali feldspars: orthoclase, microcline, perthite, anorthoclase, albite ($An_{00-05}$);
*P* plagioclase ($An_{05-100}$), scapolite;
*F* feldspathoids (foids): leucite and pseudoleucite, nepheline, sodalite, noseane, hauyne, analcime, cancrinite, etc.;
*M* mafic minerals (micas, amphiboles, pyroxenes, olivines, opaque minerals, accessories (zircon, apatite, titanite, etc.), epidote, allanite, garnets, melilites, monticellite, primary carbonates, etc.).

**Fig. 5:** The QAPF double-triangle for plutonic rocks having less than 90% dark minerals, from Sorensen (1974) after Streckeisen (1967). The diagram is based on model oompositions as above.

# 7. DISCUSSION OF THE GEOTECTONIC EVOLUTION OF THE EAST ARM SUBPROVINCE

From mapping of the East Arm area, Hoffman (1968, 1969) postulated that the Aphebian Goulburn, Epworth, Snare and Echo Bay Groups and the Great Slave Supergroups were a part of a single geotectonic unit i.e. the Coronation Geosyncline. His later mapping shows that the Great Slave Supergroup was not actually part of the geosyncline but was set back into the Archean craton (compare Hoffman, 1969, 1970 and 1972). Hoffman (1973) proposed that the East Arm and Bathurst Inlet were aulacogens, projecting from the geosyncline into the craton.

However, Hoffman (1973) and Hoffman et al. (1973) proposed lower Aphebian rifting of the western margins of the Slave craton to produce new ocean crust. They further proposed the concomitant production of 'failed arm' of triple junctions i.e. the East Arm and Bathurst 'graben' and the subsequent closure of the ocean by subduction and collision in the late Aphebian to produce the Wopmay orogen. Initially Hoffman proposed that the units involved in the Aphebian orogeny be ascribed to the 'Wopmay Orogen'. On the other hand Badham (1973a, b) argued that there is no evidence for closure by collision in the Wopmay Orogen and consequently there is no requirement for initial rifting. Furthermore Badham (1973b) has proposed that there may always have been oceanic crust to the west of the North American craton. Hence there is no clear evidence for lower Aphebian rifting of the western margin of the Slave craton. Recently Badham (1978) discussed the early evolution of the East Arm. His descriptions concerned the nature of faulting, the basement, the Wilson Island Group, the Union Island Group, and the lower Great Slave Supergroup, and he proposed an alternative model to the failed arm hypothesis. The alternative model is rationalised by the following facts:

1) The East Arm was clearly connected to the Wopmay Orogen, as was the Bathurst graben. The orogen resulted from 'Andean' orogeny along the western margins of Slave craton (Badham, 1973b; Hoffman and McGlynn, 1976). There is no sign that the craton was rifted in earliest Aphebian times to produce a new ocean and furthermore the

Slave craton was always bounded to the west by ocean (and therefore oceanic crust?).

2) The Slave craton was stabilised at approximately its present thickness at around 2500 m.y. It is now bounded by the Wopmay orogen and the two fault systems on three sides and by a sharp 1750 m.y. metamorphic front (Thelon Front) on the fourth.

3) The Churchill province was stabilised at approximately its present thickness at around 1750 m.y. The province consists of remnant keels of Archean greenstone, highly deformed troughs of Aphebian rocks and vast intervening areas of updated Archean and various anatectic Aphebian granites and gneisses. The tectonic grain of the province strikes NE-SW (Davidson, 1972a).

The Archean cratons must have been as thick and stable at around 2500 m.y. as they are now. Badham (1978) proposed that the Churchill Province was matured and thickened by northwesterly directed compression between the Superior and Slave cratons in the period of 2500-1750 m.y. and the East Arm and Bathurst grabens are the result of deflection of this compression along the sides of the rigid Slave craton. The above proposition is summarised in Fig. 6 (adopted from Badham, 1978).

Regarding the 'failed arm' hypothesis several points appear to be at variance with it:

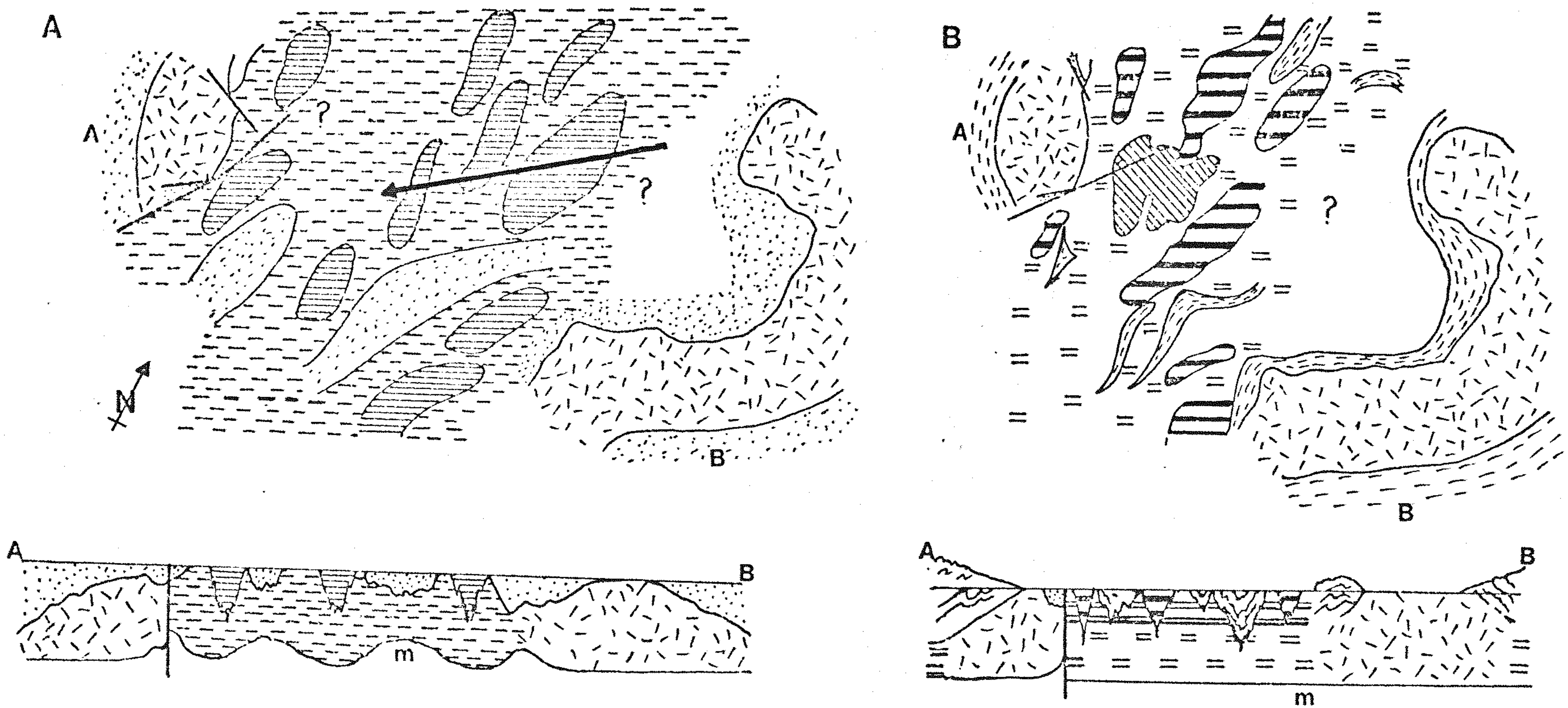
1) The fault system has always been essentially right-lateral and strike slip, and totally contrasted to those around rifted grabens.

2) Faulting was probably initiated in the late Archean and certainly predated the supposed initial rifting. Faulting continued in the same style long after the supposed termination of rifting.

3) The Archean basement north of the East Arm shows no sign of lower Aphebian uplift and erosion adjacent to the East Arm.

4) The Wilson Island Group was formed and preserved by essentially the same controls that operated on younger groups and is clearly not part of a 'pre-aulacogen basement'.

5) The magmatism, which is predominantly alkaline and calcalkaline, cannot be related to rifting either in time or in space, but was localised in the major strike-slip faults and emplaced intermittently throughout the Aphebian.




Undeformed Aphebian sediments, both in basins on metastable crust and in marginal belt.

Deformed Aphebian basins showing fold trends.

Archean crust, stable since 2500 m.y.

Remnants of Archean greenstone belts in immature crust.

Compressed Archean remnants in mature crust.

Dubawnt Group, Middle Proterozoic cover.

Mature late Aphebian crust.

Immature early Aphebian crust.

**Fig. 6:** Model for the evolution of the East Arm in relation to Aphebian tectonics of the Canadian Shield, from Badham, 1978. Diagram A represents the state in the Lower Aphebian (~2400 m.y.) and diagram B, the state at the end of the Aphebian (~1750 m.y.). The large black arrow shows the direction of compression in the Churchill province and resultant shear on the MacDonald system.

6) The basement on the south side must have been updated and eroded throughout the Aphebian times and somewhere to the east of the East Arm.

Thus the above points refute the concept of a 'failed arm' for the evolution of the East Arm graben. The East Arm graben was in existence long before its supposed genesis at a triple junction, and the main stages of its evolution have been due to the continuous activities along the fundamental strike-slip fault.

# P A R T   I I

# M E T A L L O G E N E S I S

# 1. INTRODUCTION

The geological evolution of the East Arm area has been documented in some detail, as have the magmatic host rocks related to the various mineralisations. Some of these mineral deposits could be of economic importance, especially those with Ni-Co-Fe arsenides, uranium and probably Bi.

In Part II each of these deposits is described and attempts are made to relate them to a unified concept of the total evolution of the area. As part of such studies analytical data have been obtained via optical, diffractometric and electron microprobe techniques. The details of the analytical procedures and corrections applied are given in the Appendix.

## 2. TYPES OF MINERALISATIONS

The East Arm intrusions contain hydrothermal mineralisations, firstly of magnetite-apatite-actinolite in pipes and pegmatites, secondly of uranium-thorium-rare earth element minerals in pipes, veins and pegmatites and thirdly of Ag, Bi, Ni-Co-Fe arsenides, sulpharsenides, sulphides and sulphosalts in veins and skarns. The older intrusions contain two separate types of mineralisations; firstly sulphides and oxides occur in the ultramafic and mafic portions; and secondly Ni-Co-Fe arsenides with some bismuth and silver (reported?) occur in veins. On the other hand the younger intrusions, particularly at Western Labelle Peninsula, contain magnetite-apatite-actinolite associated with uranium and rare earth elements as well as Ag, Bi, Ni-Co-Fe arsenides. It is found that the arsenides occur intimately with sulpharsenides and some sulphides and sulfosalts. However the magnetite-apatite-actinolite deposits were emplaced earlier in joints, sometimes passively and sometimes forcefully, and were probably generated as immiscible liquids. The arsenide mineralisations essentially involved fluid migration up early joints and deposition in open spaces. In Blanchet Island, deposition in structural traps at the base of the laccolith is also observed.

Some of the mineralisations, particularly at Aristifats Lake, are related to the extrusive rocks of the Seton Volcanics. In addition, two occurrences of arsenide mineralisations are not apparently related to the intrusions. As was mentioned earlier, it is not unreasonable to suppose a hidden igneous body underneath yet unexposed.

darker uraninite often replaces the lighter type and hence some workers have called it secondary uraninite (Ramdohr, 1969).

Uraninite observed in this study is mostly in cubic forms, although some small, botryoidal crystals are found. However uraninite is often altered to coffinite (Plate 8). In the West Labelle area, uraninite is closely associated with Ni-Co-Fe arsenides and molybdenite, while in Regina Bay area it is associated with magnetite-apatite-actinolite. It is found that the uranium mineralisation is just after the deposition of molybdenite and magnetite, but pre-Ni-Co-Fe arsenides. In addition, uraninite seems to be replacing allanite, particularly at West Labelle. Hematite is ubiquitous in all the uranium mineralisations and mainly replacive. (Plates 7 and 8). The association of uraninite and magnetite-hematite is not uncommon. Uranium here is thought to be leached out and remobilised from older strata and then hydrothermally deposited in veins. It has been shown that for pitchblende to be taken into solution the uranium must be oxidised to the hexavalent state ($U^{6+}$). Various writers such as Lovering, Tolmachev and Frederickson (in Barton, 1956) proposed that in an acid, oxidising environment uranium is present in solution as the uranyl ions, $UO_2^{+2}$. Rodden and Warf (in Barton, 1956) have mentioned the possibility of uranium forming complexes with carbonate, sulphate, chloride, fluoride, and sulphite ions.

It is concluded that the uranium mineralisation was introduced into the veins at the same time as hematitic alteration. Subsequent brecciation and the introduction of new minerals served to recrystallise, and sometimes remobilise the uranium minerals. This however true for the uranium mineralisation in association with Ni-Co-Fe arsenides of West Labelle. Much of the uranium minerals have been remobilised, and replaced by arsenides. Some of these uranium minerals could not be identified with certainty, although they have been detected by Geiger-counter. The author just managed to identify some fine grains of uraninite and coffinite in the arsenides using electron microprobe technique, autoradiography and X-ray diffraction methods. It is interesting to note that no thorium minerals were identified in these mineralised veins. Galena is common in these uranium veins. Usually it fills the cracks in uraninite crystals and also surrounds the grains.

## 3. DETAILED MINERALOGY

### a. Oxides

Oxides of iron and titanium are very common in all the deposits. Hematite is the most ubiquitous of the vein minerals, being present in one form or another. Usually it is manifested by a red dusting of carbonates. Needles in rosettes are not uncommon, particularly in the mineralised veins at Regina Bay (Plate 7G). Often it forms martitization texture with magnetite and ilmenite. It is usually formed from the oxidation of magnetite or the breakdown of ilmenite and mafic minerals and occurs in almost all the stages in the paragenetic sequence.

Magnetite is one of the earliest minerals to be deposited in the veins. On the other hand it is also usually found as blebs together with ilmenite, particularly in the ultramafic and mafic rocks of the older intrusions as a result of breakdown of olivine by hydrothermal fluids. As it was mentioned above, magnetite is normally oxidised to hematite and perhaps maghemite. However ilmenite usually altered to rutile and anatase (Plate 3 G). Petrographically rutile and anatase are quite similar, both grey in colour, and their identification is based on their different anisotropic colours and internal reflections.

In West Labelle area, a copper oxide mineral, cuprite ($Cu_2O$) was identified. It is observed to fill the fractures in chalcopyrite. Elsewhere in the East Arm no cuprite was identified.

### b. Uranium Minerals

Most of the uranium minerals are in the form of uraninite ($UO_2$) or the silicate coffinite ($U(SiO_4)_{1-x}(OH)_{4x}$) as they were confirmed by electron microprobe. Uraninite itself is made up of three distinct varieties, $\alpha$-uraninite-$UO_2$ (uraninite in the strict sense), $\beta$-uraninite - $U_3O_8$ (pitchblende in the strict sense) and $\gamma$-uraninite-$U_4O_7$ (Wasserstein, 1954). The name pitchblende usually refers to colloform, botryoidal or massive uraninite. In this study, the lighter and darker varieties are clearly distinguishable. The darker variety has lower reflectivity and hardness than the 'normal' lighter uraninite. In addition, the

Regarding the uranium secondaries, a few are identified using X-ray diffraction method. They include the yellow, greenish yellow soddyite, autunite, schoepite, dumontite(?) and the orange, orange red clarkeite (gummite). Others could not be identified with certainty due to their mixture with the above minerals. These uranium secondaries are usually deposited along fractures and joints in the vein.

c. Sulphides

Small amounts of pyrite and chalcopyrite are associated with almost every stage of the mineralisations. However, sulphides are only common in veins in the third stage of the general paragenetic sequence of the mineralisations in the East Arm. In the ultramafic and mafic rocks of the early intrusions, the sulphides and oxides formed immiscible blebs within the silicates. These early sulphide blebs are disseminated in the rocks and those identified are pyrite, chalcopyrite, pyrrhotite, pentlandite, bravoite, and millerite. The blebs are thought to have recrystallised during the cooling of the intrusions (the Easter Island Dyke in particular). However they contain no arsenic, silver or bismuth. The sulphide content is rather high in the altered margins of the intrusions. Much of the nickel here was probably leached from the altered nickel-riched olivine. However some of chalcopyrite crystals are altered to secondary copper minerals, such as chalcocite, digenite and covellite, by late supergene processes.

In the mineralised veins pyrite and chalcopyrite are common. Other sulphides include bravoite, marcasite, millerite, pentlandite, valeriite(?), heazlewoodite, bornite, idaite(?), chalcocite, digenite, covellite, galena, linnaeite(?), argentite, bismuthinite and molybdenite. All these minerals were identified petrographically and some by X-ray diffraction and electron microprobe techniques. It is possible some minerals, especially in the linnaeite group (Co-sulphides) are present but have not been identified with certainty. However it is surprising to see that no zinc minerals have been identified, although some workers have reported the presence of sphalerite in the East Arm area. Most of the above minerals are shown in the accompanying plates.

Specks of these sulphides, especially bismuthinite, millerite and chalcopyrite-digenite are seen to replace Ni-Co-Fe arsenides. There is no doubt that some of the sulphides such as molybdenite, pyrite and chalcopyrite are earlier in the general paragenetic sequence of the vein mineralisations. Marcasite was seen to replace pyrite in West Labelle. Bismuthinite was usually deposited after native bismuth and the arsenides. This is also true for argentite and other sulfosalts which were deposited after native silver.

d. Arsenides

Studies of the arsenide minerals, especially the di- and tri-arsenides, have been hampered for many years by a profusion of structural and chemical data that not only seen to be inconsistent, but which were frequently in error. This confusion is due to the fact that most arsenide minerals occur in highly complex intergrowths, and that analyses and X-ray diffraction studies were made on mixtures, rather than on single phases. Although some detailed papers by Holmes (1947) and Roseboom (1962, 1963) sorted out much of the confusion, errors were still common, even in the sixties. A history of the evolution of ideas on arsenides is given by Holmes, and the result of his classification is adopted here (Fig. 7) together with that of Radcliffe and Berry (1968).

The higher arsenides consist of two distinct series. The first, an isometric triarsenides series, is grouped under the general name of 'skutterudite'. There appears to be complete solid solution between Ni-Co- and Fe- end members, but triarsenides with only single element in the metal sites are confined, apparently, to the cobaltian member. In Ni- and Fe-rich members the triarsenides break down to the second series which is a group of monoclinic or orthorhombic diarsenides. This can be subdivided further into three mineral groups - the safflorites, monoclinic Co-Fe diarsenides; löellingite, an orthorhombic Fe-rich diarsenide; and rammelsbergite, and its polymorph pararammelsbergite (Radcliffe, 1966), a Ni-rich orthorhombic diarsenide. At high temperatures, experimental data (Ramdohr, 1969) indicate complete solid-solution between cobalt and nickel in the diarsenides and the skutterudite field is rather small. At lower

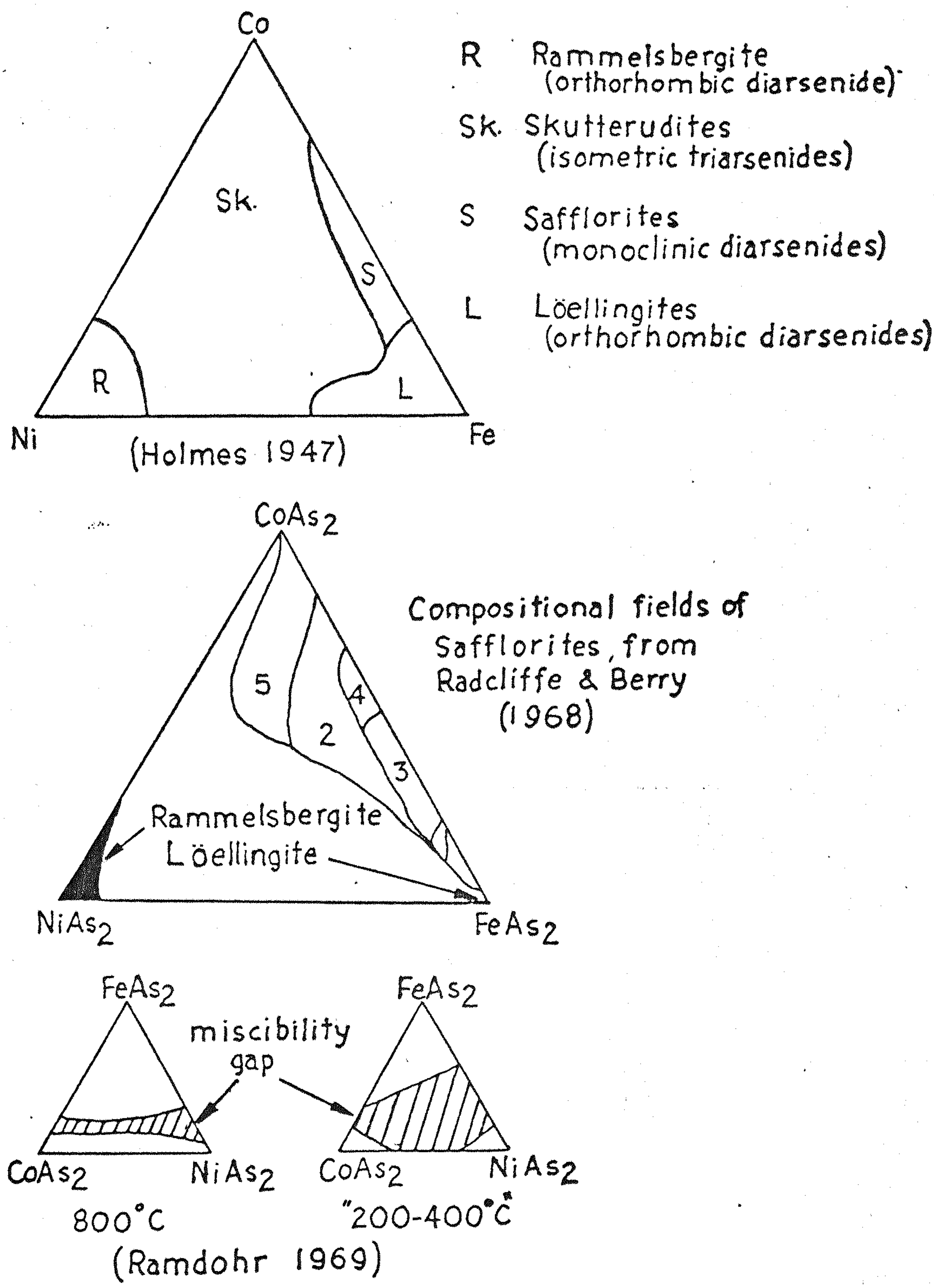



**Fig. 7 : Stability fields and compositional ranges of Di- and Tri-Arsenides.**

temperatures this solid-solution breaks down, and two separate diarsenides result (Fig. 7). However, the existence of a pure cobalt diarsenide has not been documented in the literature, and was not found in this study.

The advent of the electron microprobe made it possible to prove the validity of Holmes work in the cryptically zoned and banded samples. Studies by Radcliffe (1966, 1968) and Radcliffe and Berry (1968) have been especially important. It was found that the diarsenides are essentially stoichiometric and that previous postulations of metal - or arsenic-dificient members are erroneous. There is no documented substitution for nickel, cobalt and iron, and only sulphur has been shown to replace arsenic up to a maximum of about 5 wt.%. Radcliffe and Berry (1968) proposed slightly different limits to the Fe-Co diarsenides and distinguished five different safflorites which, because of differing Fe : Co ratios, have slightly different X-ray diffraction patterns. In addition, Radcliffe and Berry also note that the safflorites are predominantly orthorhombic, although monoclinic examples have been recorded by Petruk (1971, 1972). Skutterudites have frequently been reported as being arsenic-deficient, and this is well-documented by Petruk (1971) who gives an average Fe : As ratio of 1 : 2.8.

The studies of the relations between coexisting phases and the trends in mineral zoning are of greater interest, than the exact chemical nature of the mineral phases. In this study, the higher arsenides were identified by the X-ray diffraction method, and by electron microprobe technique , as well as optically. These arsenides were separated into the isometric and non-isometric groups. The isometric series were identified as the skutterudites, and the non-isometric series as either rammelsbergite and safflorite-löellingite. This is well confirmed by the electron microprobe analyses. Furthermore the optical data, including the use of microhardness and reflectance measurements, allowed clear separation of the rammelsbergite and safflorite löellingite groups. Optically safflorite is hard to distinguish from löellingite. Safflorite and loëllingite have characteristic zoning textures (Plates 10, 15 and 17) which rammelsbergite does not. Star-shaped safflorite twins are common, particularly from the West Labelle mineralised vein. Normally rammelsbergite is optically distinguished from safflorite or löellingite by its higher reflectance and whiter colour. On the other hand skutterudite is distinctively isotropic, but its intergrowth with isotropic gersdorffite (Ni-Co sulpharsenide) and to some extent, cobaltite, often made distinction rather difficult. In this study, the above difficulty was solved by the use of electron microprobe technique.

During the investigation two lower arsenides, namely niccolite and maucherite, were identified. Niccolite usually forms the bulk of the arsenides in the veins and maucherite is formed at the expense of the niccolite. Pararammelsbergite was not definitely identified, although some of the 'rammelsbergite' grains seem to have higher reflectance (61-62%) than the 'normal' rammelsbergite (59-60%).

The arsenides occur as single crystals in the wall rock or gangue, as dendrites, rosettes and tubercules, and as massive patches formed by the coalescence of finer structures. The rosettes are not the cross-section of dendrites, but are clearly spherical. The tubercules are morphologically similar to the dendrites, but are straight-stemmed and unbranched.

All three structures are made up of very finely-banded, mixed arsenides. Relationships in the banded areas are variable, but the following generalisation can be made:

1. Cores are more coarsely crystalline, and rims most finely-banded. (Plates 3, 15, 17 and 18).
2. Banding may parallel and reflect the shape of the outer walls of the structure, may parallel inclusions in the core and may be highly variable (Plates 15, 17 and 18).
3. Niccolite generally occurs in the centre of structures.
4. Rammelsbergite and skutterudite are most common on the outer rims of niccolite cores. (Plate 18C).
5. Many of the structures examined in hand specimens and polished-sections show that niccolite with rammelsbergite rims formed initially, and that this recrystallised later and was rimmed further by other arsenides, especially skutterudites and safflorite-löellingite at the outermost rims. (Plates 15, 17 and 18).
6. In banded di- and triarsenide sequences, there are no distinct preferential replacement by bismuth and carbonates or sulphosalts.

(Plate 15, 17 and 18). All arsenides contain bismuth inclusions.

7. In polished sections of Sachowia Lake samples, bismuth was seen in niccolite (Plate 15), although rammelsbergite, and skutterudite do contain this metal.

8. Only trace amounts of silver are associated with the arsenides, usually as disseminated fine-grains.

9. There are Ni-rich areas in the veins. Co- and Fe-As-rich areas are more widespread, generally occur on either sides of the nickel rich areas, and are rich in bismuth inclusions.

The electron microprobe data substantiate many of these generalisations and permit further ones:

1. There is no Co-rich diarsenide, but all the skutterudites are Co-rich.

2. Fe-rich diarsenides are only seen in thin, discontinuous bands in diarsenide sequences (Plate 17).

3. Co- and Fe-rich arsenides are more common in the outer margins of the structures (Plate 17 and 18).

4. Co- and Fe-rich diarsenide sequences are often intimately interbanded with triarsenides (Plate 18).

Spot analyses probably integrate a number of cryptic compositional bands, but the variations between Ni, Co and Fe are of interest. Nickel and cobalt vary antipathetically. Iron shows some sympathetic variations with cobalt but generally decreases inwards from the margins.

One of the most intriguing facets of the arsenides is the banding. From the descriptions above it is suggested that many of the bands, and certainly the well-crystallised cores, are not original, but are post-depositional features. It has been shown that some dendrites can grow by rim replacement and gradual infilling. Others have clearly grown outwards. However, Kidd and Haycock (1936), and Stanton (1972) have assumed that the arsenides accreted in successive bands around silver, bismuth, or carbonate cores. Both mention the possibility of colloidal precipitations. In the mineralised veins of the East Arm, the accretion hypothesis can be applied. Since Ni-rich arsenides generally forms at the cores, it is assumed that the Co- and Fe-rich arsenides grew around them. However, the exact mechanism of banding is not known. It is thought that the bands may have developed with their present chemistry by direct precipitation from a gel; or a mixed arsenides, stable at higher temperatures, may have formed first and broken down into the banded di- and triarsenide sequences as it became unstable.

Phase relation studies of the arsenide minerals are not common, perhaps the best being that of Yund (1961) on the Ni-As system. Yund implies that rammelsbergite is unstable and reverts to pararammelsbergite at 590 ± 10°C under the vapour pressure of the assemblage. Evidence from fluid inclusion and isotope work (Robinson, 1971) and from the preservation of pre-arsenide assemblages, indicates that temperatures in veins containing pararammelsbergite cannot have been in excess of 300°C. This anomaly still remains to be corrected.

e. Sulpharsenides

Sulpharsenides are quite common in the East Arm mineralised veins. They are closely associated with the arsenides already mentioned. Gersdorffite and cobaltite are the two sulpharsenides most commonly encountered in the study. It is not unreasonable to expect arsenopyrite or glaucodot but these have yet to be found by the author.

These sulpharsenides clearly replace the earlier arsenides, especially in fractures and around the outer rims of the arsenides (Plates 15 and 18). In addition they are often found intergrown with banded di- and triarsenides. As was mentioned earlier, these sulpharsenides are rather difficult to distinguish from skutterudites, due to the fact that they are isotropic or only weakly anisotropic. Gersdorffite is usually pitted and fractured and often contains tiny blebs of bismuth (Plate 15). On the other hand cobaltite is smooth in appearance and has distinct crystal-shape (Plate 13).

In the general paragenetic sequence, the sulpharsenides were deposited just after the deposition of the arsenides. It is found that gersdorffite and cobaltite are closely associated with native metals, especially bismuth and some electrum (Au, Ag) (Plate 15 and 13). Blebs of the native metals are observed to be replacive, usually in fractures.

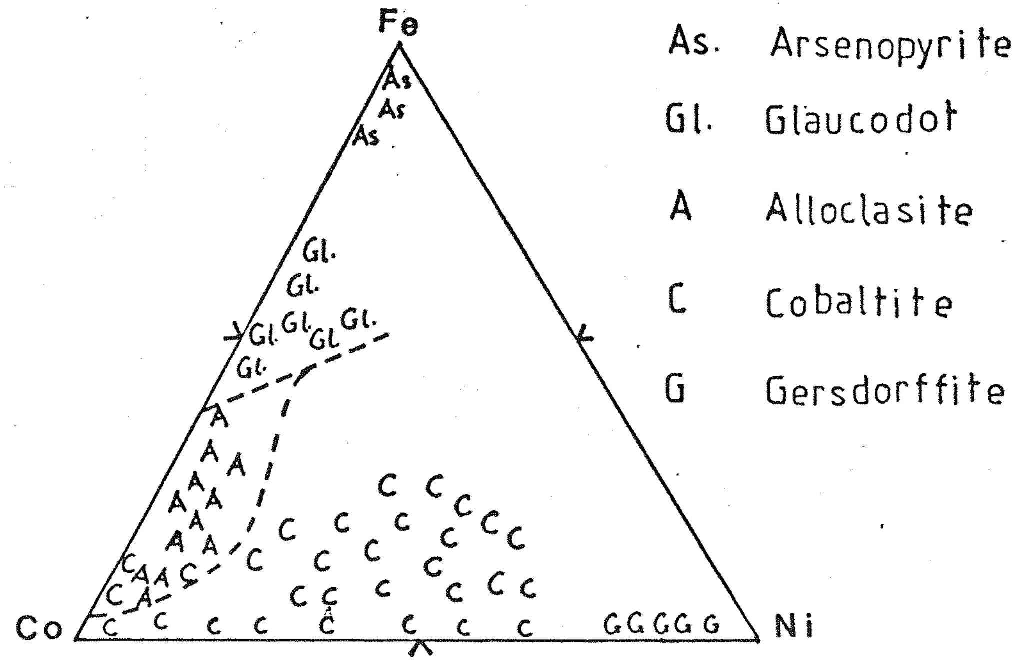



Fig. 8: Compositional fields of Sulpharsenides– from Petruk, 1972.

Gersdorffite and cobaltite often resemble each other. This is probably due to the fact that at high temperature a solid-solution member between cobaltite and gersdorffite does really exist (Klemm and Weiser, 1965). From the electron microprobe data of the East Arm mineralisation the phase 'cobaltite', which is thought to be a solid solution member, contains high nickel. From the compositional fields of sulpharsenides (Fig. 8) taken from Petruk (1972), the high nickel cobaltite is equivalent to gersdorffite. Petruk (1972) has pointed out that the cobaltite of Cobalt-Gowganda area also contains high nickel values.

f. Sulphosalts

The sulphosalts occur as small, massive and complex replacements of arsenides, as intergrowths with sulphides and as small disseminated grains in the gangue. In polished sections they are all closely similar, and different phases are recognised only by subtle nuances of colour. The complexity of intergrowths often makes X-ray diffraction work impossible, and certain identification can only be made after microprobe analyses.

Stromeyerite ($Cu_2S.\ Ag_2S$) was seen in only one polished section (Sample from West Labelle), as blebs in carbonate gangue along with native silver, argentite, chalcocite, digenite and covellite (Plate 12A). It is the first occurrence of this mineral reported in the East Arm. Proustite or ruby silver ($3Ag_2S.As_2S_3$) is found in several polished sections of West Labelle veins and also occurs along with silver and other sulphides mentioned above. The red-internal reflection is typical of proustite.

Most abundant sulphosalts are the Bi-sulphosalts. Hauchecornite ($Ni_4(Bi,Sb)\ S_4$) and parkerite ($Ni_3\ Bi_2S_2{}^2$) were identified. These Bi-sulphosalts are found replacing the arsenides, and also closely associated with sulphides such as chalcopyrite, pyrite, bravoite, bismuthinite and in one polished-section with galena (Plates 4, 15, 16, 18, 19). Native silver is clearly associated with hauchecornite and parkerite. An unknown Bi-Pb-sulphosalt was identified in a polished-section of a West Labelle sample. Optically it has the characteristics of aikinite ($2PbS.\ Cu_2S.\ Bi_2S_3$) and each of those elements was identified by electron microprobe. However no certain analytical data are available. In addition a Pb-sulphosalt was also identified, as intergrowths with this 'aikinite'. It contains lead, copper and iron in addition. Optically this phase has some of the characteristics of betekhtinite ($Pb_2(Cu, Fe)_{21}S_{15}$). These last two sulphosalts are closely associated with Cu-Fe-Sulphide and uraninite mineralisation. They are found to be replacing galena and chalcopyrite.

g. Native Metals

Bismuth is the most abundant of the native metals found in the East Arm vein mineralisations. Some silver are found as disseminated grains and stringers in the carbonate gangue (Plate 12). In addition, a polished section of West Labelle sample contains electrum (Au, Ag) as it was confirmed by electron microprobe (Plate 13).

It is surprising not to see much silver here compared to other Ni-Co-arsenide mineralisations elsewhere. On the other hand, bismuth is abundant, especially in the mineralised veins in Blanchet Island, Sachowia and Zig areas. (Plates 14, 15, 16, 17 and 18). Bismuth is free of silver and vice versa. This was confirmed by microprobe analyses (Plates 12 and 16). No native animony or arsenic or gold free of silver were identified.

h. Gangue Minerals

Carbonates, apatite and actinolite are ubiquitous gangue minerals in the magnetite-uranium minerals association (Plates 6 and 9). Apatite and actinolite generally predate the uranium deposition, although some late apatite and actinolite do replace uraninite (Plate 8A). In West Labelle area, allanite is locally deposited in the veins prior to the uranium mineralisation. Detailed mineralogy shows that there are several generations of carbonates which are mostly made up of calcite and ferroan-dolomite. Only chloroapatite was identified by electron microprobe.

In association with the arsenide mineralisations, carbonates are again the predominant gangue mineral. They are mostly ferroan-dolomite with some calcite. These carbonates are made up of several generations in the paragenetic sequences. The early carbonates are replaced by the arsenides, and the later ones often filled the fractures in the arsenides.

i. Alteration Minerals

Annabergite, erythrite, malachite, azurite, goethite and uranium secondaries were all identified in surface outcrops of the veins which are extensively weathered. Annabergite and erythrite form in a matter of weeks on massive vein arsenides exposed on the surface. These alteration minerals are readily recognised by their characteristic colours.

## 4. DETAILED MINERALISATION AREAS

### a. Easter Island

#### History

The mineralisation was discovered in 1950 and has been investigated by drilling, trenching and a ground magnetometer survey. A few bags of high grade silver ore were shipped early on, but the property has remained idle since 1965.

#### Host Rock

The gabbroic and syenitic rocks of the dyke are considered as the host rocks for the mineralisation. The dyke intruded the Archean basement of pegmatitic granite. The Archean basement, mostly garnet-biotite gneiss here, contains a few specks of pyrite and chalcopyrite. In Part I of this thesis, rocks have been described in considerable detail. On the Easter Island, the dyke is usually vertical, but dips steeply to the southeast at Good Friday Bay (Map 7). Along the shores of Good Friday Bay it is a zoned gabbro with olivine and sulphide-rich margins. The margins are cut by small faults which contain both mineralised veins and diabase. Although the sulphide-rich margin persists for 5 miles to the east, no further arsenide veins are known.

#### Dyke Mineralisation

In the Good Friday Bay area both margins of the dyke are altered. The alteration zones consist of talc-carbonate-chlorite-sulphides. They are interpreted as the product of carbonation of olivine-pyroxene-sulphide rich margins. The altered zones contain primary sulphides, but most of the sulphide has been remobilised into talc-carbonate stringers during alteration. Pyrite, pyrrhotite and chalcopyrite have all been noted, together with supergene bornite, digenite, covellite and hematite. Magnetite and ilmenite are the primary oxides identified but their alterations to hematite-rutile and anatase are common. Other sulphides identified are bravoite, pendlandite and millerite blebs. The blebs are free of arsenic and silver. In general

PLATE 3

Easter Island Dyke mineralised gabbro and veins:

A) Hand specimen showing the arsenides replacing and replaced by ferroan dolomite (Fe-dol.). Mineralised vein.

B) Niccolite (nicc.) core rimmed by skutterudite (sk.) in ferroan dolomite gangue (dark). Later millerite (mill.) crystals have replaced the skutterudite. Mineralised vein. P.P.L.

C) Same as B - crossed nicols.

D) A large grain of skutterudite (sk.) - rammelsbergite (ramm.) intergrowth, in ferroan dolomite. Remnants of earlier niccolite (nicc.) are found within the core of the structure. Later chalcopyrite (cp.) and bravoite have replaced the above ore minerals. P.P.L.

E) Aggregates of pyrite (py.) crystals with cubic bravoite cores in dark grey ferroan dolomite gangue (Fe-dol.). Mineralised vein. P.P.L.

F) Intergrown pyrrhotite (po.) and pentlandite (pent.) replaced by chalcopyrite (cp.) and ilmenite. These minerals occur with liquid immiscibility textures against the silicates, mostly olivine (ol.), which is partly altered to serpentine (serp.). Mineralised gabbro. P.P.L.

G) Aggregates of pyrite (py.) crystals replacing earlier ilmenite (ilm.) in olivine (ol.). Some crystals of ilmenite are altered to rutile (rut.). Mineralised gabbro. P.P.L.

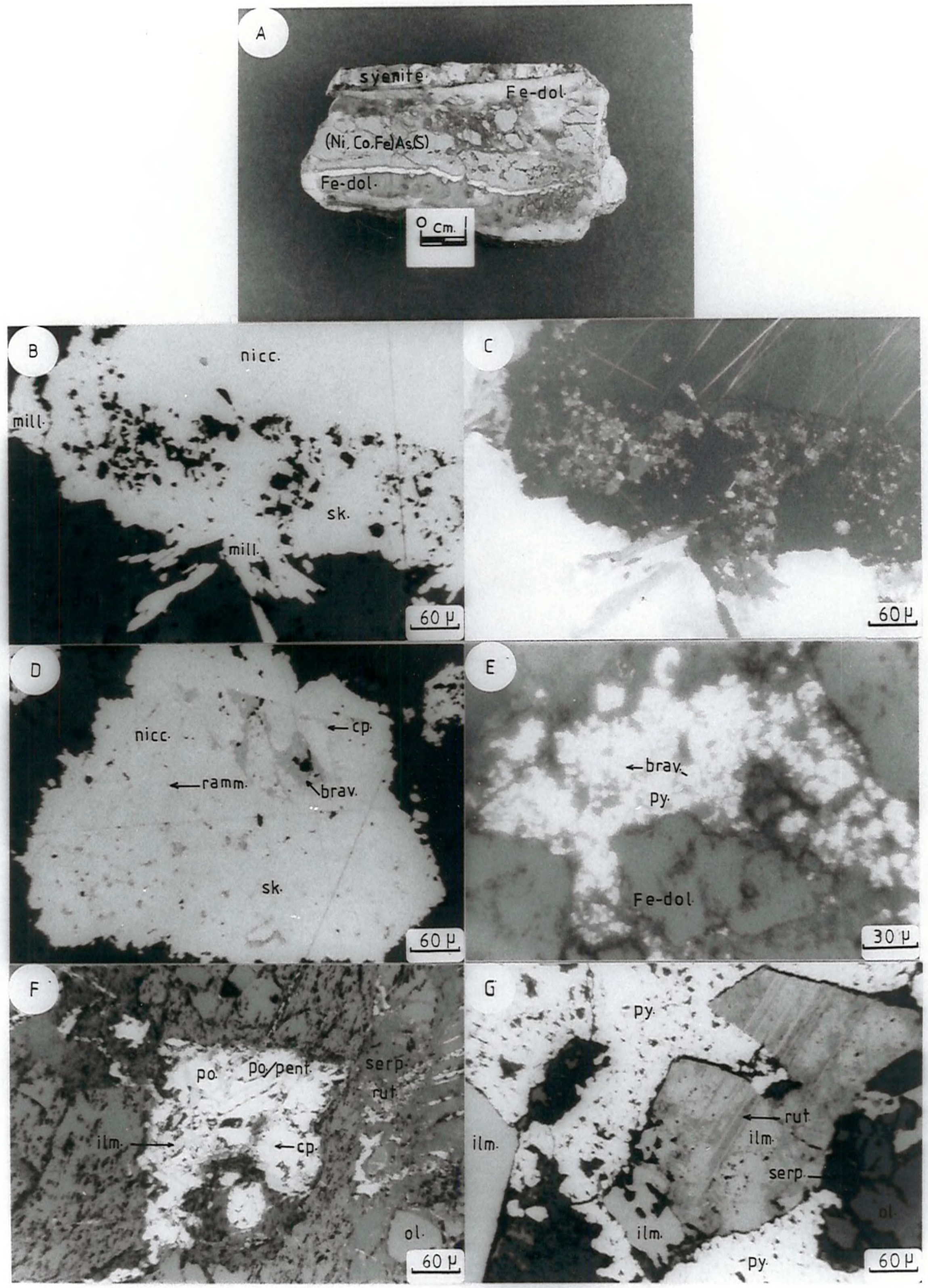

A
syenite
Fe-dol
(Ni, Co,Fe)As(S)
Fe-dol
cm.
B
nicc
mill
sk
mill
60 μ
C
60 μ
D
nicc
cp
ramm
brav
sk
60 μ
E
brav
py
Fe-dol
30 μ
F
po
po/pent
serp
rut
ilm
cp
ol
60 μ
G
py
rut
ilm
ilm
serp
ol
ilm
py
60 μ

PLATE 3

PLATE 4

Easter Island Dyke mineralised vein:

A) Niccolite (nicc.) replaced by intergrown group of minerals including millerite (mill.), gersdorffite (gersd.), hauchecornite (hc.) and bravoite along fractures. P.P.L.

Ac) Atomic number distribution of A - electron probe microphotograph.

(Ni+Co) Nickel and cobalt contents. High Co is restricted to the area marked by the dotted line. Ni $K\alpha$ and Co $K\alpha$.

(Fe) Iron content. Fe $K\alpha$.

(Bi) Bismuth content. Bi $K\alpha$.

(Sb) Antimony content. Sb $K\alpha$.

(As) Arsenic content. As $K\alpha$.

(S) Sulphur content. S $K\alpha$.

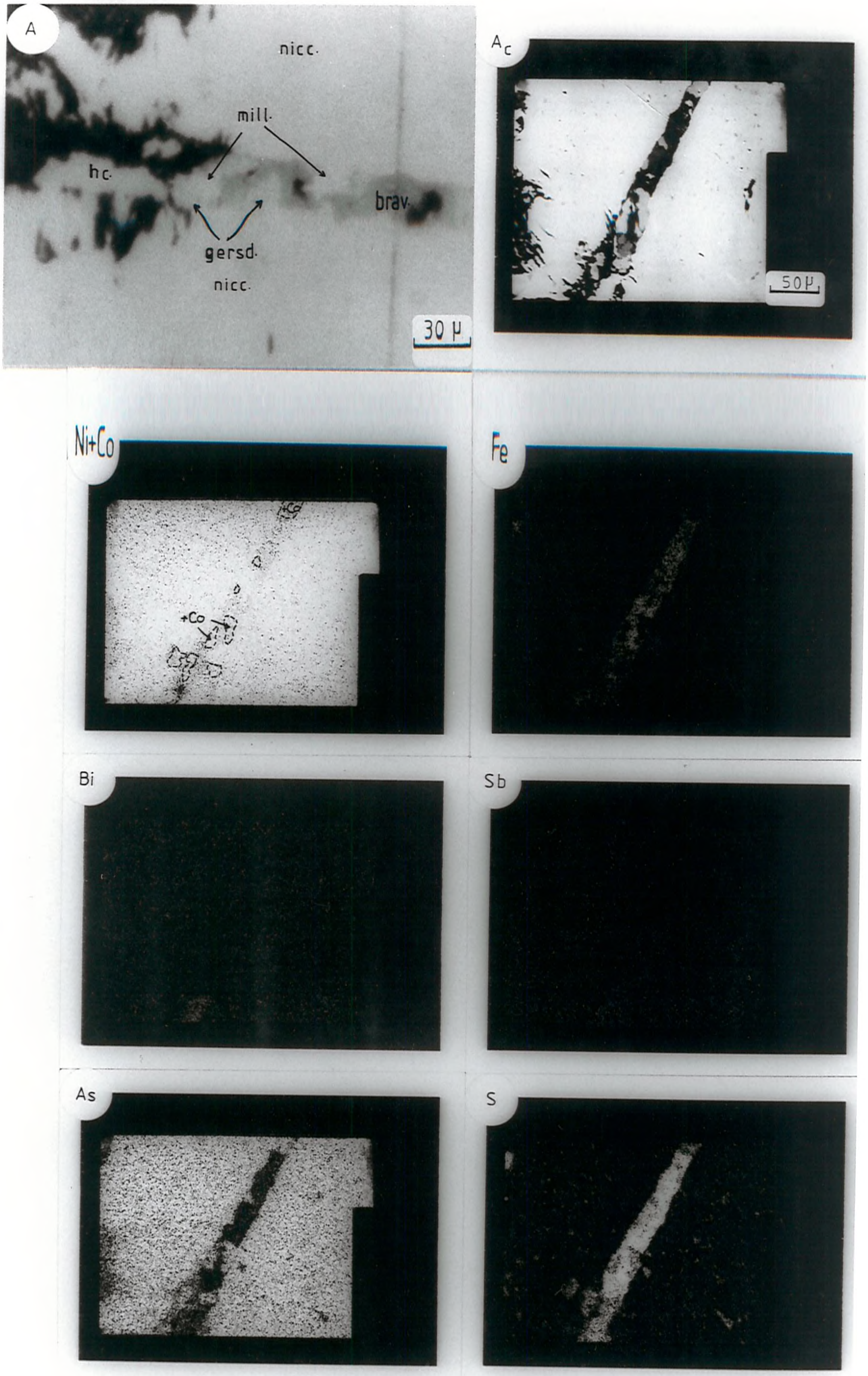



PLATE 4

nickel and copper contents of the dyke are less than 0.1% but in the altered margins values are variable up to 0.3% (Badham, In Press). This mineralisation was tested by drilling in 1961 and 1967 but no significant tonnages were outlined. There are central areas where the olivine has been serpentinised and small blebs of nickel sulphides, pentlandite and millerite in the alteration zones offset to the original high nickel content of the olivine.

## Vein Mineralisation

Vein mineralisation occurs on the margins of the dyke at its western end. Veins and shear zones run out from the intrusion and into the Archean gneisses (Map 7). The veins are at low angles to the margins of the intrusion and are most abundant on the north side. In the interior of the dyke the veins contain only ferroan-dolomite and some calcite or siderite, but in the altered margins and for a few metres beyond the contact the veins contain spectacular arsenide mineralisation in a ferroan dolomite-calcite-siderite-quartz gangue. The gneisses are entirely unaltered around these veins. This vein mineralisation consists predominantly of botryoidal niccolite which is usually rimmed by gersdorffite and skutterudite. Rammelsbergite, cobaltite, millerite, parkerite, hauchecornite, bravoite, pyrite and chalcopyrite were also identified. In this study the author could not find any native silver or other silver minerals in association with the arsenides, although silver was reported before. However small blebs of native bismuth were identified.

## Paragenetic Sequence and Mode of Deposition

The paragenetic sequence of the mineralisation is separated into two, firstly the dyke mineralisation and then the vein mineralisation. In the mafic and ultramafic cumulates of the dyke, the oxides are thought to have been deposited very early in the paragenetic sequence as a result of the breakdown of the primary silicates. Almost immediately the sulphides were deposited as a result of the same phenomenon. It is suggested that during cooling of the intrusion autometasomatic hydrothermal activity on the margins caused the olivine to alter to talc, and carbonate was introduced. The sulphides were recrystallised, and perhaps further sulphides were engendered by exsolution of Cu-Fe-Ni from olivine. Such reactions are typical of hydrothermal alteration of ultramafic rocks under oxidising conditions. Some of the sulphides, especially chalcopyrite then altered into secondary copper minerals such as digenite and covellite.

However, the earliest arsenides to be deposited are niccolite and rammelsbergite. Then they were rimmed by skutterudite and other sulpharsenides. All the sulphides were finally deposited, although some pyrite, chalcopyrite and bravoite are prior to the arsenide deposition. Native silver and bismuth were probably deposited during the early part of the late sulphide stage. Both paragenetic tables are presented here (Tables 5 and 6). Stage I of the host rock mineralisation (Table 5) consists principally of early disseminated copper-nickel-iron sulphides and iron-titanium oxides. This is followed by the deposition of intermediate Fe-Ti oxides in Stage 2. Time spent for Stage 3 (a and b) is much longer than the above stages and predominantly contains intermediate Cu-Ni-Fe sulphides. Stage 4 is an oxidation stage and represents the final deposition of secondary sulphides and oxides. For the vein mineralisation (Table 6) the first stage is also represented by early sulphides and oxides. Stage 2 contains Ag?, Bi, Ni-Co-Fe arsenides and sulpharsenides. Stage 3 is made up of intermediate sulphides and the later part of the stage (Stage 3b) contains bismuth sulphosalts. The time of bismuth deposition was prolonged throughout Stage 3. Finally, oxidation of primary minerals has taken place in Stage 4.

Regarding the vein mineralisation, it is thought that at greater depth the alteration of olivine would involve the release of metals (Ni, Co, Fe, Bi, Ag?, As) but not sulphur, since the conditions would have been reducing for the hydrothermal alteration. It is proposed that these metals were carried up the dyke margins and into fractures in higher levels of the dyke which were already cooled. Carbonates and quartz were remobilised from the dyke along with silver?, bismuth, arsenides and sparse sulphides.

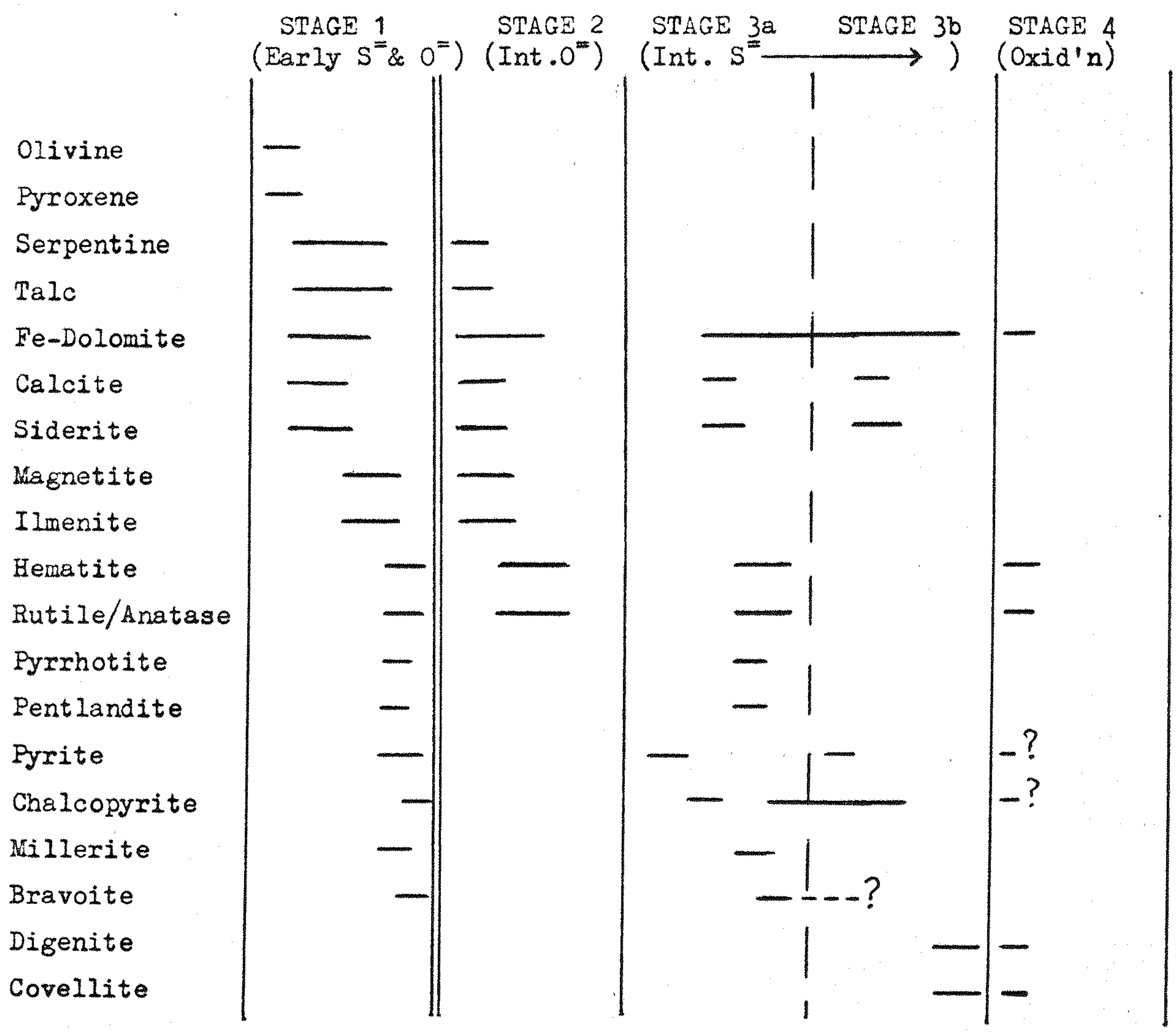



Table 5: Paragenesis of the Easter Island Dyke.
Double line represents main period of fracturing and brecciation.

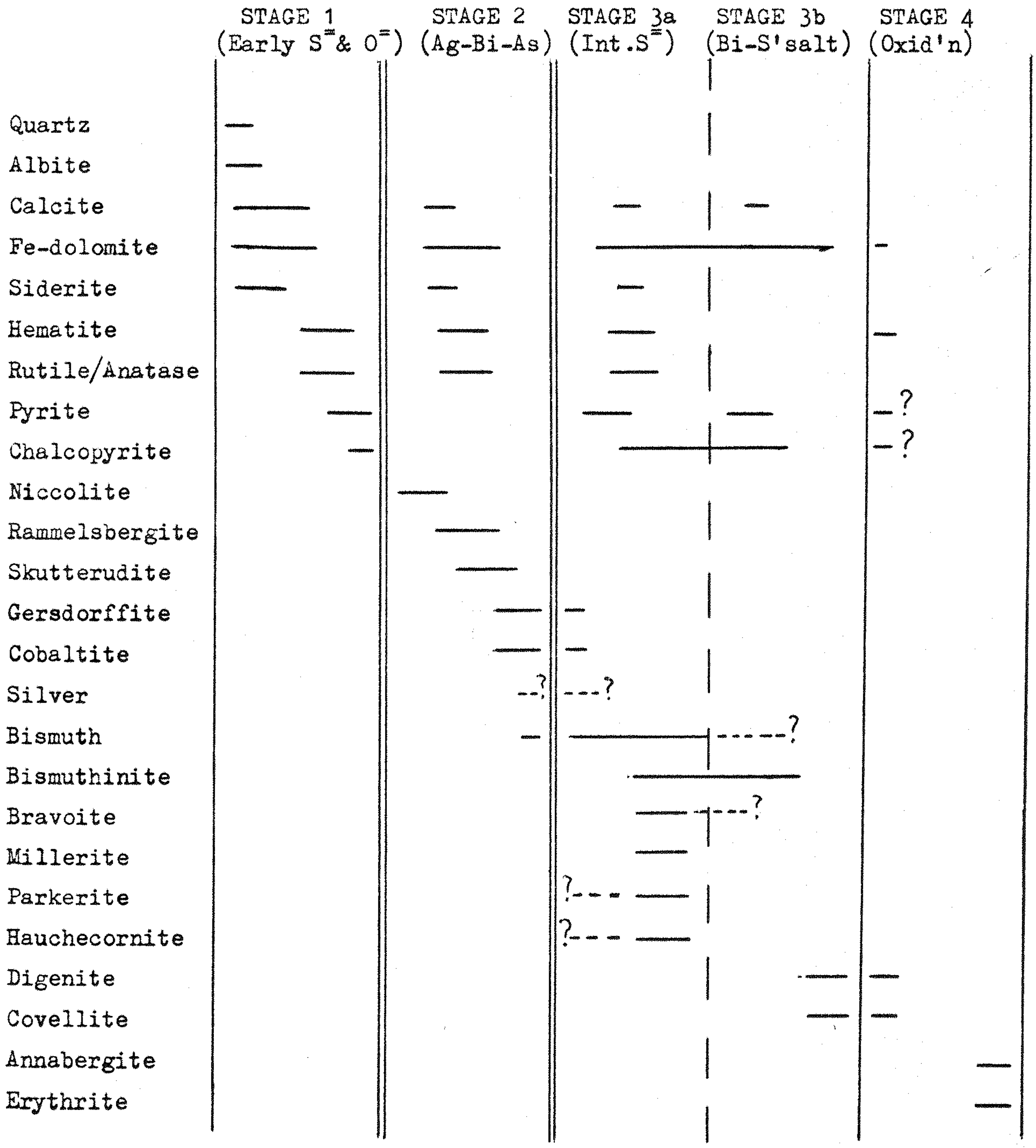



Table 6: Paragenesis of Easter Island Dyke's mineralised veins.
Double lines represent main periods of fracturing and brecciation.

b. The Maple-Blachford Lake Complex

In this study, the author did not examine much of the detail mineralisations in the complex. However some details about this complex was documented in Part I of this thesis. A brief description of the mineralisation is reported here.

In general the Maple-Blachford Lake complex is so much similar to the Easter Island dyke and their mineralisations are almost equivalent to each other. Although some arsenide mineralisation are found directly associated with this complex no samples were available in this study. Two arsenide mineralisations have been discovered in Gin and Zig areas (Maps 5 and 9). These mineralisations are thought to be closely related to the Maple-Blachford Lake Complex. The detailed mineralisations at Zig will be discussed in a later part of this thesis.

The Maple-Blachford complex as a whole consists essentially of alkaline rocks described earlier. The Maple Gabbro was later named as Caribou Lake Gabbro by Davidson (1978). It essentially consists of marginally alkaline gabbro with zones of more mafic and sulphide-rich cumulate gabbro and more oxide rich anorthosite. The sulphide-rich gabbro contains blebs of chalcopyrite and nickeliferous pyrite which show immiscibility textures against the silicates. The gabbro is generally quite fresh and there is no sign of recrystallisation of the sulphides. The anorthosite layers contain interstitial skeletal ilmenite.

The Odin Dyke is part of the Blachford Lake Complex. This dyke, which is an alkali syenite pegmatite (albite-riebeckite-biotite) dyke, and other similar dykes cut the Blachford Lake Complex. The dyke is extremely complex, containing areas of greatly varying grain-size and proportion of its principle constituents riebeckite and alkali feldspar. The dyke is commonly brecciated with later phases cementing fragments of earlier ones. Apatite and rare-earth element minerals are common accessories and in places are concentrated in small pockets, overgrowing the primary silicates. There are small breccia-pipes and 'crackle-zones' cutting all phases. One such fracture zone contains magnetite, uraninite, monazite and unidentified rare-earth element minerals. However a grab sample of this vein was reported to contain 0.3% $U_3O_8$, 0.2% $ThO_2$ and 20.5% total rare-earth element minerals. (D.I.A.N.D. files, Yellowknife from Badham, In Press). Davidson (1972) describes zones of weak radioactivity in fluorite-cemented syenite breccias about 1 km northwest of the Odin Dyke which may be similar.

However it is thought that the genesis of the sulphide mineralisation in the Maple gabbro is closely similar to that of the Easter Island Dyke and their paragenetic sequence is more or less the same.

c. Aristifats Lake

## History

Copper and cobalt mineralisation has been known for some time in carbonate veins in Kahochella (Seton volcanics) agglomerates. Early trenching yielded high grades, but no tonnage potential. However it was realised more recently that the agglomerates here are not a thin skin cover on Archean granite, but a Seton volcanic vent. Recent drilling has proven this contention and the potential for greater tonnage of metals has been increased. Drilling of the vent in 1974 found that high concentrations of mineralisation did not continue at depth. However more drilling is going on.

## Host Rocks

The Archean granitic basement is overlain unfonformably by a variable thickness (100 - 400 feet) of red mud-chip conglomerate, red siltstone, red shale, grey shale and hematitic shale, belonging to the Akaitcho Formation of the Sosan Group. These are overlain conformably by globular, oolitic hematite beds, tuffs, red shale and agglomerates of the Seton Formation of the Kahochella Group. In the mineralised area a vent pipe of agglomerate extends over 200 feet in depth. These rocks outcrop on both sides of Talthelei Narrows, and are cut by a thick diabase sill on Utsingi Point.

## Mineralisation

The detailed petrography of the agglomerate has been documented in Part I of this thesis. The drill samples taken have apparently copper mineralisation but with the cobalt stain, due to the presence of erythrite. The mineralised samples were taken from the central zone of the vent, which is made up of completely carbonated rock fragments in a carbonate matrix, all riddled with carbonate veins. The ore minerals usually crystallised on the margins of rock fragments. The minerals observed are mainly chalcopyrite replacing fractured and brecciated pyrite. Supergene alteration of the primary copper and iron sulphides is clearly prominent especially along the fractures.

PLATE 5

Aristifats Lake mineralised agglomerate:

A) Hand specimen showing pyrite (py.) and chalcopyrite (cp.) with traces of arsenides in interstices of altered feldspars (fsp.). Secondary malachite (mal.) and erythrite (eryt.) are shown on the weathered surfaces.

B) Fractured and brecciated pyrite (py.) partially replaced by chalcopyrite (cp.) and digenite (dig.) - goethite (goet.) along fractures. P.P.L.

C) Euhedral pyrite (py.) and traces of cobaltite (cob.), gersdorffite (gersd.) and löellingite (loel.) partially replaced by chalcopyrite (cp.) in the altered feldspar and carbonate gangue (dark). Digenite (dig.) and goethite (goet.) have replaced chalcopyrite along fractures. P.P.L.

Cc) Atomic number distribution of a part of C - electron probe microphotograph.

(Ni) Nickel content. Ni $K\alpha$.

(Co) Cobalt content. Co $K\alpha$.

(As) arsenic content. As $K\alpha$.

(S) Sulphur content. S $K\alpha$.

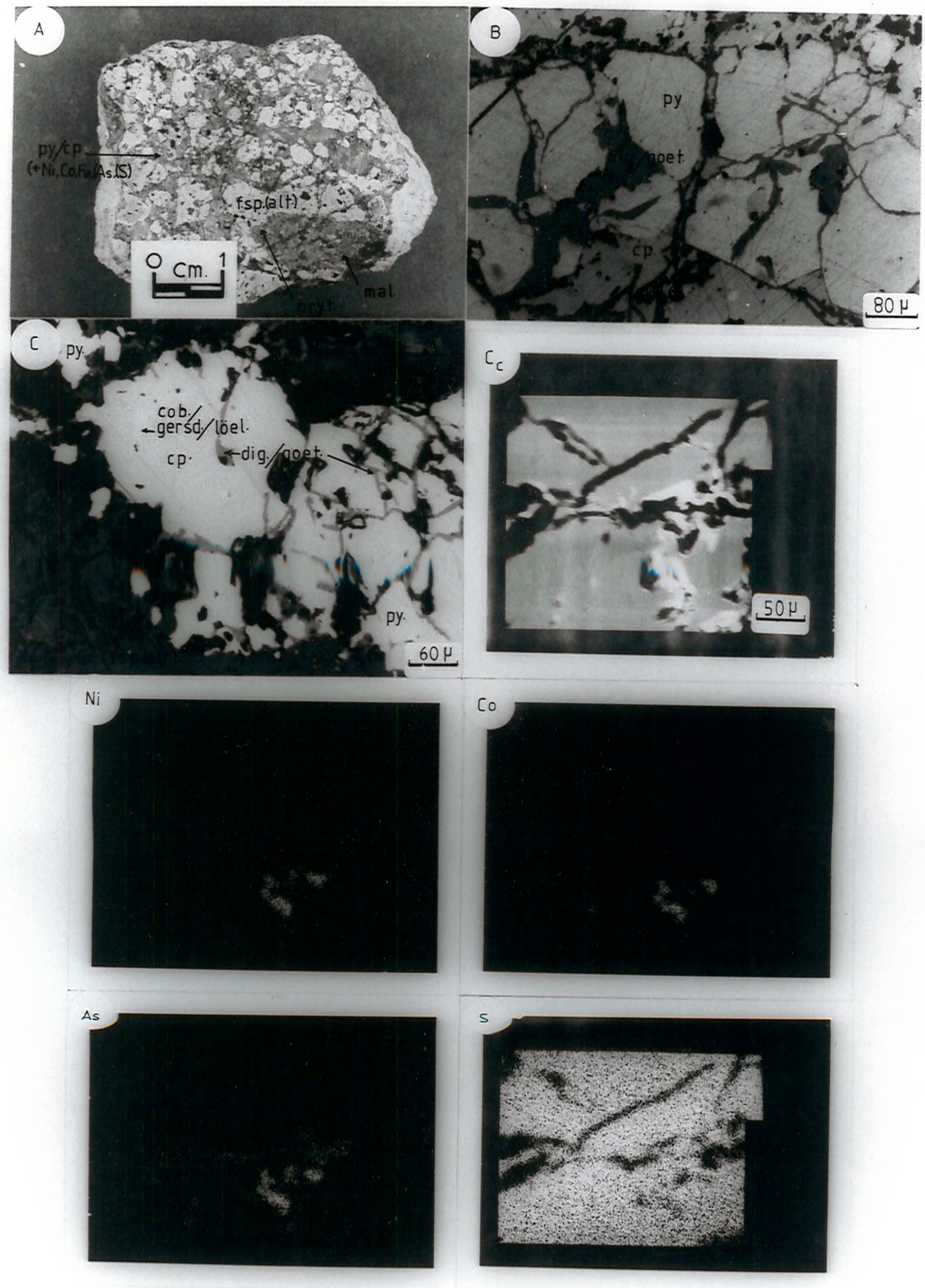

A
py/cp
(+Ni,Co,Fe(As,S)
fsp.(alt)
O Cm. 1
mal
B
py
cp
80 μ
C
py
cob.
gersd.
löel.
cp.
dig.
goet.
py.
60 μ
Cc
50 μ
Ni
Co
As
S

PLATE 5

Digenite, covellite and goethite are common supergene minerals here. However early arsenides and sulpharsenides are observed as fine white blebs in chalcopyrite, and were confirmed by electron microprobe analyses (Plate 5). They are cobaltite and skutterudite. Preliminary drilling has indicated up to 500 tons averaging 6% copper, 2% cobalt and 3 ounces per ton silver. Grades of 2% copper and 0.5% cobalt are more widespread. In this study the author could not identify any native silver. Argentite was identified optically intergrown with the supergene copper and iron sulphides.

Paragenetic Sequence and Mode of Deposition

The mineralisation is clearly the final product of hydrothermal alteration in a vent. It is proposed that the alteration is concentric from fresh to argillic to propylitic to carbonated, and that the mineralisation was deposited in the most altered rock. Altered and weakly mineralised Seton vents have been seen elsewhere.

In the paragenetic sequence it was observed that the arsenides and sulpharsenides have been replaced by the later pyrite and chalcopyrite. Supergene alteration minerals will be the latest in the sequence. The detail paragenetic table is shown in Table 7. Stage 1 is predominantly made up of early Cu-Fe sulphides and Cu-Fe-Ti oxides. Ni-Co-Fe arsenides and sulpharsenides were deposited in the second stage. Intermediate Cu-Fe sulphides occurred in the relatively longer Stage 3. Stage 4 is an oxidation stage which contains secondary Cu-Co-Fe-Ti minerals.

It is important to note the presence of cobalt in the mineralisation since cobalt is not a common element in mineralised volcanic stockworks. This suggests that the Seton volcanics may be more closely related to other East Arm magmatic rocks than has hitherto been suspected. The similarities of the mineralisation to that at Valnicla (Badham, In Press) are also noted.

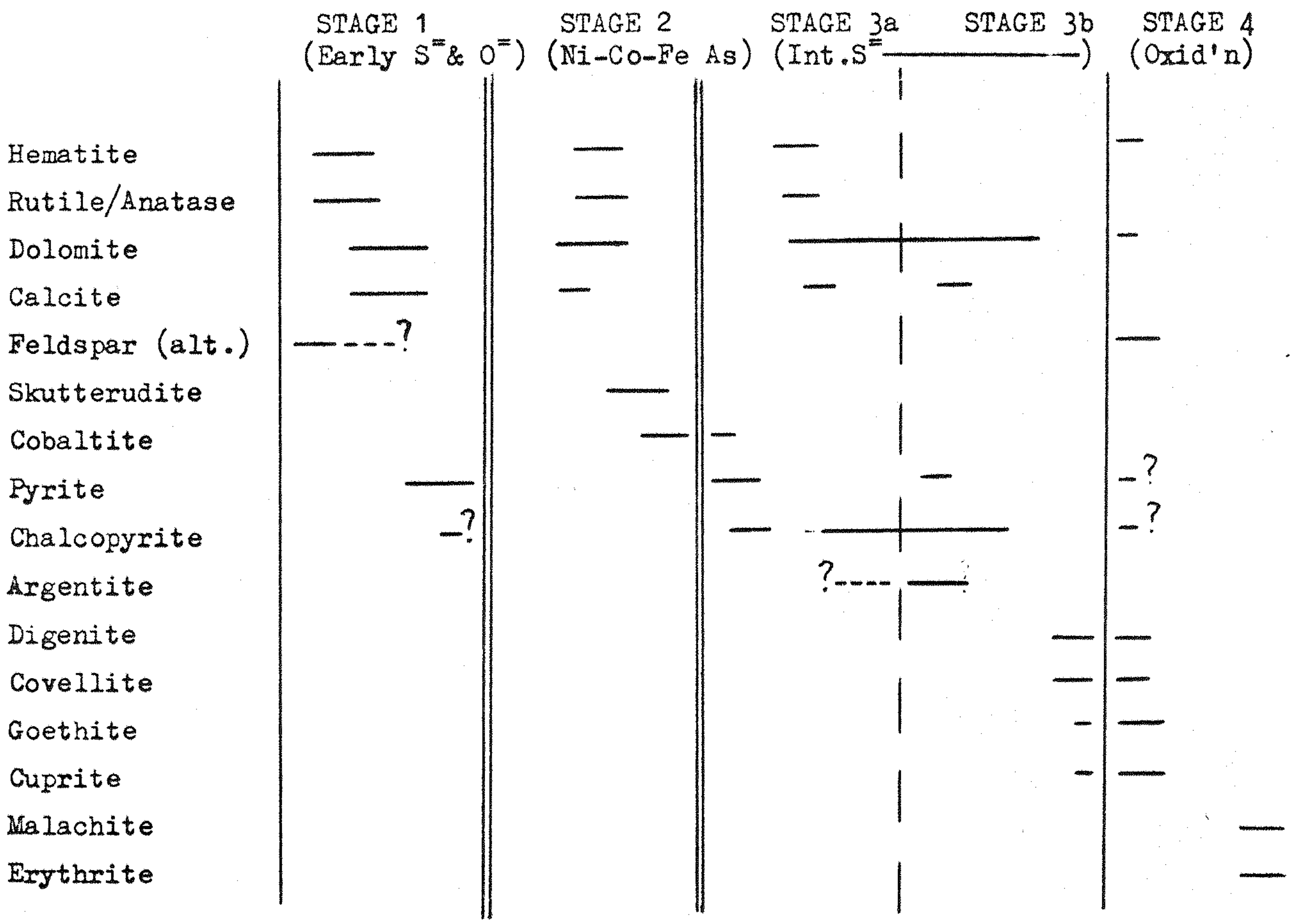



Table 7: Paragenesis of the mineralised vent at Aristifats.
Double lines represent main periods of fracturing and brecciation.

d. Regina Bay

## History

The Regina Bay diorite was staked in 1949 as the Rex Group following discovery of radiometric anomaly of the C-zone. Further prospecting uncovered 5 radiometric zones of which only A and C were of any significance. These two zones were trenched and drilled in the early fifties and a short adit was advanced on the C zone. However the property has been inactive since 1954.

## Host Rocks

The host rock is the calc-alkaline diorite stock which has been documented earlier on in Part I. This hornblende-plagioclase porphyritic diorite plug (or stock) intrudes the contact between red mudstones and buff limestones of the Stark Formation and similar rocks of the Upper Kahochella Group. All these rocks are cut by diabase sills and dykes which post-date the mineralisation.

## Mineralisation

The five radioactive veins all lie in joints relatively near to the contact, but none were seen to extend beyond the contact, or even through the contact breccia. The A, B, E, and Stevens veins were not examined by the author. However company reports indicate that they lie in NE trending vertical joints. (Badham, In Press). They vary between 0 and 6 feet wide and 200 feet long, and fade out into stringers at either end. They contain actinolite and magnetite and radioactive minerals are erratically distributed. They are economically unimportant.

The C zone has been explored in detail by Badham (1975). The vein is 570 feet long, 18 to 50 inches wide, vertical and NW-striking. Drilling indicates that it pinches out into a stringer zone at either end and at a depth of some 150 feet. Actinolite and apatite filled much of the vein. Actinolite is oriented perpendicular to the vein wall (Plate 6) and crystals over 1 foot long were seen. Intra-crystallisation movements in the vein have resulted in bent crystals.

PLATE 6

Regina Bay mineralised veins:

A) Hand specimen of magnetite (mnt.) - apatite (ap.) - actinolite (act.) + uraninite (ur.) pegmatite. Early white calcite (calc.) cross-cut by needles of actinolite (dark).

B) Hand specimen showing curved acicular actinolite (act.) crystals cross-cut by apatite (ap.) crystals. Magnetite (mnt.) and hematite (hm.) found in the central part of the pegmatitic vein in diorite.

C) Hand specimen showing intergrowth of magnetite (mnt.) - apatite (ap.) - actinolite (act.) and late white calcite (calc.) veinlets cutting all the minerals. Massive magnetite/hematite concentrated at the centre of the pegmatitic vein.

D) Hand specimen showing acicular actinolite cross-cut by apatite crystals (brown). Calcite (white) is both earlier and later than apatite-actinolite.

E) An autoradiograph of one of the polished-sections containing uranium. Scale: 3x

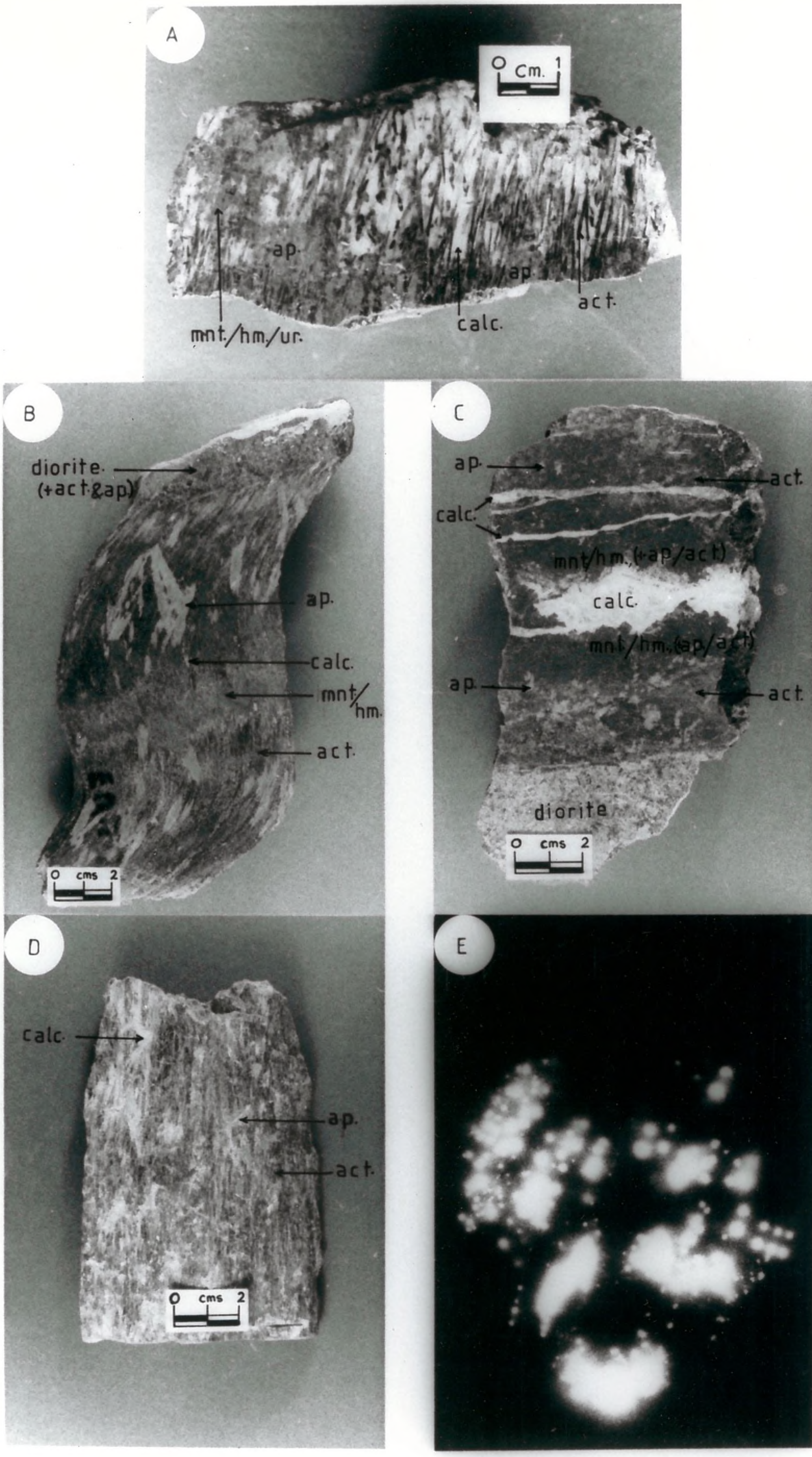

A
0 Cm. 1
ap.
ap.
mnt./hm./ur.
calc.
act.
B
diorite
(+act.&ap.)
ap.
calc.
mnt/hm.
act.
0 cms 2
C
ap.
calc.
act.
mnt/hm.(+ap/act)
calc.
mnt/hm.(ap/act)
ap.
act.
diorite
0 cms 2
D
calc.
ap.
act.
0 cms 2
E

PLATE 6

PLATE 7

Regina Bay mineralised veins:

A) Pyrite (py.) being partly replaced by magnetite (mnt.) - hematite (hm.). Later galena (gn.) replaces all the above minerals. Late apatite (ap.) and calcite have corroded the ore minerals. P.P.L.

B) Fractured and brecciated pyrite (py.) replaced by later galena (gn.). P.P.L.

C) Fractures in pyrite (py.) healed by late chalcopyrite (cp.) and galena (gn.). Some late crystals of apatite (ap.) replaced the ore minerals. P.P.L.

D) Late pyrite (py.) and chalcopyrite (cp.) corroding the earlier uraninite (ur.) along the rims and fractures in earlier apatite (ap.) gangue. P.P.L.

E) Euhedral and zoned uraninite (ur.) replaced by hematite (hm.). Some of the hematite is formed after actinolite (act.). Late euhedral apatite and needles of actinolite often replaced both uraninite and hematite. P.P.L.

F) Similar to E. P.P.L.

G) Rosette needles of hematite (hm.) replaced apatite (ap.) and actinolite (act.).

H) Massive rounded crystals of uraninite (ur.) have replaced by needles of actinolite (act.) and corroded by later hematite (hm.). P.P.L.

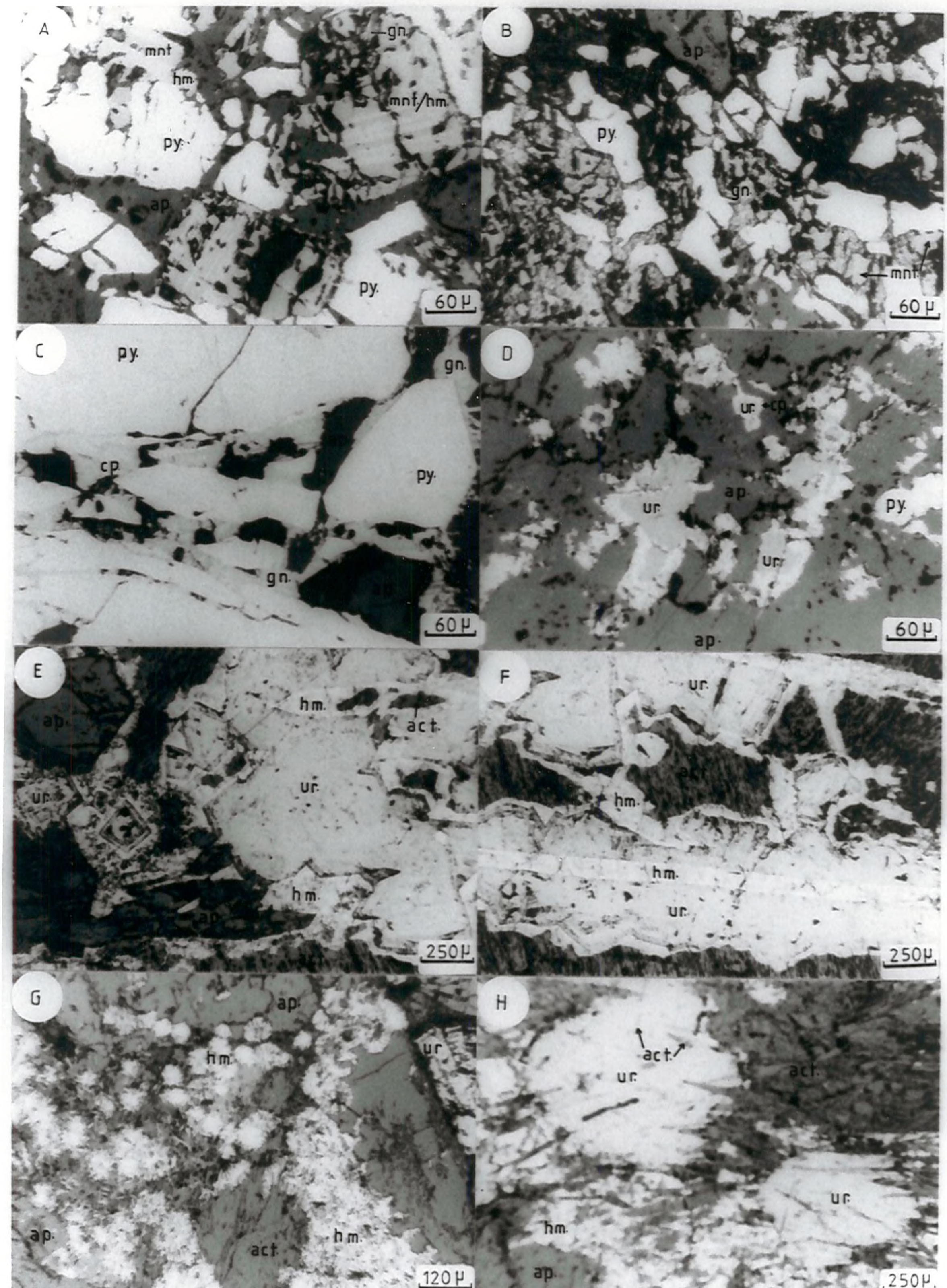

A
mnt
hm
gn
mnt/hm
py
ap
py
60µ
B
ap
py
gn
mnt
60µ
C
py
gn
cp
py
gn
60µ
D
ur
cp
ap
ur
py
ur
ap
60µ
E
ap
hm
act
ur
ur
hm
250µ
F
ur
act
hm
hm
ur
250µ
G
ap
ur
hm
ap
act
hm
120µ
H
act
ur
act
ur
hm
ap
250µ

PLATE 7

PLATE 8

Regina Bay mineralised veins:

A) Cubic and zoned uraninite (ur.) altered to coffinite (coff.), and partly replaced by apatite (ap.) and actinolite (act.). Magnetite (mnt.) - hematite (hm.) pre-date the uraninite. P.P.L.

Ac) Atomic number distribution of A - electron probe microphotograph.

(U) Uranium content. U $M\alpha$.

(Si) Silicon content. Si $K\alpha$.

B) Intergrowths of gersdorffite (gersd.), loellingite (loel.) and safflorite (saff.) in calcite (dark). P.P.L.

Bc) Atomic number distribution of the smaller grain in B - electron probe microphotograph.

(Ni) Nickel content. Ni $K\alpha$.

(Co) Cobalt content. Co $K\alpha$.

(Fe) Iron content. Fe $K\alpha$.

(As+S) Arsenic content. As $K\alpha$. S-rich areas superimposed.

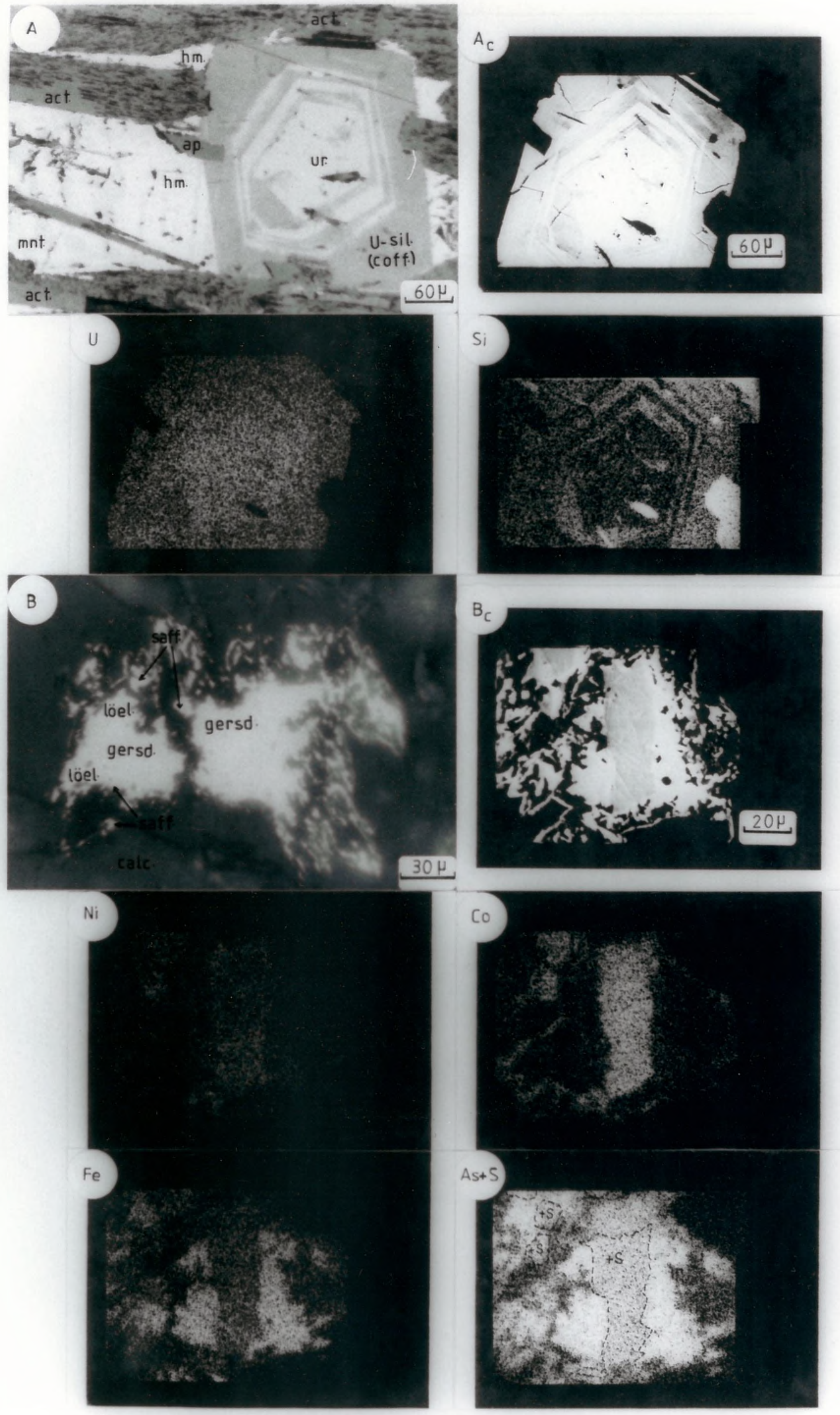

A
act
hm.
act
ap
ur
hm.
mnt
U-sil.
(coff.)
act
60μ
Ac
60μ
U
Si
B
saff
löel
gersd
gersd
löel
saff
calc
30μ
Bc
20μ
Ni
Co
Fe
As+S
+S

PLATE 8

The apatite has grown coevally with amphibole as horizontal pencils, parallel with the vein wall (Plate 6). The outer margins of this zone are marked by numerous fractures in the diorite coated with stellate actinolite and this may extend four or five feet on either side of the vein (Badham, In Press). The inner margins of this zone are sheared and the shears are covered with stellate actinolite and, locally, uranium secondaries. The central part of the vein is filled with irregular pegmatitic growths of actinolite, apatite and magnetite with interstial carbonate, some sulphides and local concentrations of uranium secondaries. It is reported that the veins contain 0.29% $U_3O_8$ over 4 feet in the adit area. In this area the author has identified uraninite and some coffinite in polished section (Plates 7 and 8). An autoradiograph of one of the samples is shown in Plate 7. Other sulphides include pyrite, chalcopyrite and galena which generally post-date the uraninite deposition. Uraninite is usually altered to its silicate variety coffinite (Plate 8A).

In one of the polished sections, two small blebs of arsenides and sulpharsenides are identified (Plate 8B). They are in the form of safflorite, löellingite, skutterudite and cobaltite-gersdorffite. This is an important showing in the Regina Bay area.

## Paragenetic Sequence and Mode of Deposition

This C zone vein shows a complex paragenesis with at least three phases of growth. Magnetite, apatite and actinolite do seem to have crystallised simultaneously. Then uranium is deposited and replaced all the above minerals. Some late actinolite and apatite post-date the uranium mineralisation. It is thought that uraninite and hematite (mostly after magnetite) are contemporaneous. The arsenide and sulpharsenide mineralisation is suggested to be later than uraninite, but prior to the sulphide mineralisations. Most of the galena is probably radiogenic and replacing pyrite and chalcopyrite in fractures. The detailed paragenesis is shown in Table 8. The first stage of the paragenesis contains early Cu-Fe sulphides and Fe-Ti oxides. Mnt-Ap-Act with uranium were deposited early in the second stage (Stage 2a) and followed by the deposition of small amounts of Ni-Co-Fe arsenides and sulpharsenides in Stage 2b. Stage 3 is characterised by the deposition of intermediate Cu-Fe-Pb sulphides. The oxidation stage (Stage 4) mainly contains uranium secondaries.

The Regina Bay mineralisation was suggested to have been produced in an open fissure system which was filled by late differentiates and later by hydrothermal fluids driven up from the hotter core of the intrusion. The open fissure system usually permitted the complete spectrum of mineralisation but here the deposition of the arsenides was for some reason limited.

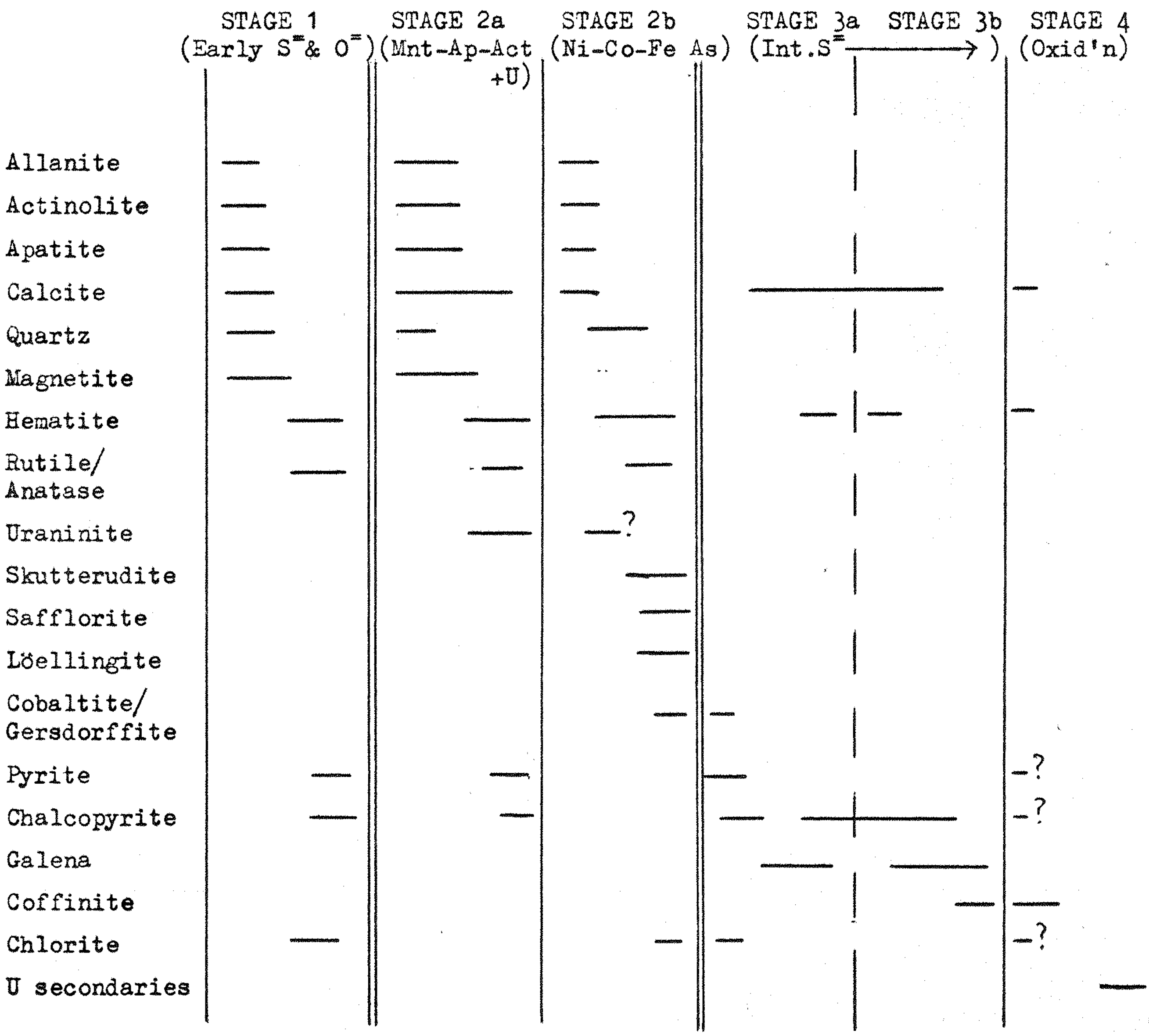



Table 8: Paragenesis of vein mineralisation at Regina Bay.
Double lines represent the main periods of fracturing and brecciation.

e. Labelle Peninsula

## History

The two radioactive zones on Labelle Peninsula were certainly known in the early fifties, but no record of their actual discovery could be found. The zones occur in the central parts of high level diorite plugs. The western zone has been trenched and drilled extensively. There was no sign of evaluation on the eastern zone.

## Host Rocks

The diorite plugs of Labelle Peninsula do appear to be slightly separate bodies. They intrude the Kahochella-Stark Formation boundary (an unconformity with the Pethei Formation missing). Both are similar in being hornblende-plagioclase porphyritic diorites, with rare biotite, but little quartz. The contacts are near vertical, often dipping steeply beneath the plug. Some 20 feet away from the contact the intrusion becomes finer-grained and impregnated with polyphase veinlets of finer igneous rock of which the youngest and most prolific is red felsite. Nevertheless the contact is sharp and the sediments are unbrecciated and are only rarely cut by red felsite stringers. The Stark Formation at the contact consists of laminated argillites and calc-argillites, in which salt pseudomorphs are common. There are also intraformational breccias of calcargillite in the Stark, and these clearly predate the intrusion. These breccias look remarkably like 'broken beds' resulting from collapse after evaporite solution.

The Eastern diorite is overlain unconformably by Murky Formation conglomerate, and is intruded by a diabase sill. The offset produced by this intrusion indicates that the plug does not have vertical sides, but that it is mushroom-shaped. The surrounding rocks were folded prior to intrusion.

## Mineralisation - West Labelle

There is a small area of radioactive anomalies in the NE corner of the western plug. While much of the plug is coarse-grained and homogeneous, in the mineralised area it is cut by numerous unoriented

PLATE 9

West Labelle mineralised veins:

A) Hand specimen containing allanite (allan.), uraninite (ur.) and molybdenite (mo.) in calcite (white).

B) Actinolite (act.) vein in diorite host rock. P.P.L. (polished thin section).

C) Same as B, except with ilmenite (ilm.) - rutile (rut.) replacing actinolite. The contact between actinolite vein and diorite is strongly hematitised. P.P.L. (polished thin section).

D) Allanite (allan.) - calcite (calc.) vein in diorite host rock. P.P.L. (polished thin section).

E) Same as D, except the vein also contains opaques - pyrite (py.), marcasite (marc.) and uraninite (ur.). Uraninite is also found in diorite as well as actinolite (act.) - chlorite (chl.). P.P.L. (polished thin section).

F) Euhedral allanite (allan.) crystals replacing calcite (calc.) and formed along the border of carbonate apatite (carb-ap.) crystals. Some uraninite replaced the allanite and also formed along the grain-borders of carbonate apatite. P.P.L. (polished thin section).

G) Mineralised vein containing calcite (calc.), carbonate apatite (carb-ap.) and opaques (rutile, hematite, pyrite and digenite) in fine-grained felsitic host rock. P.P.L. (polished thin section).

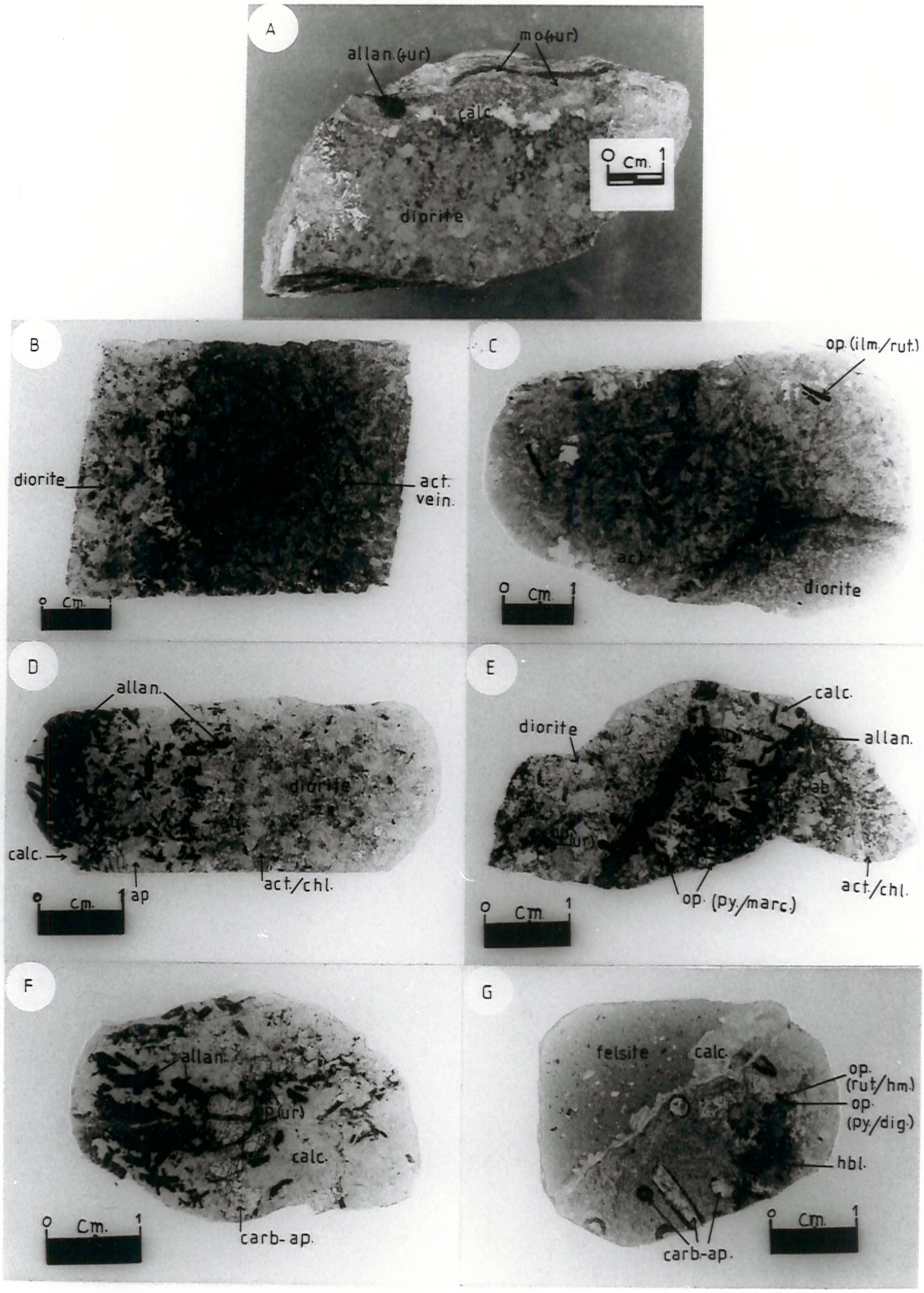

A
allan.(+ur)
mo(+ur)
calc.
diorite
0 cm. 1
B
diorite
act.
vein.
0 cm. 1
C
op. (ilm./rut.)
act.
diorite
0 cm. 1
D
allan.
diorite
calc.
ap
act./chl.
0 cm. 1
E
diorite
calc.
allan.
ab
act./chl.
op. (py./marc.)
0 cm. 1
F
allan.
calc.
carb-ap.
0 cm. 1
G
felsite
calc.
op. (rut/hm.)
op. (py./dig.)
hbl.
carb-ap.
0 cm. 1

PLATE 9

PLATE 10

West Labelle mineralised veins:

A) Fractured pyrite (py.) replaced by digenite (dig.) and hematite (hm.) needles along fractures and outer rims, in calcite (dark) gangue. P.P.L.

B) Ilmenite (ilm.) altered to hematite (hm.) and rutile (rut.) inclusions of earlier calcite (dark) are found within the aggregates of above minerals. P.P.L.

C) Cubic pyrite (py.) and quartz crystals are enclosed within later chalcopyrite (cp.). P.P.L.

D) Intergrown skutterudite (sk.) - rammelsbergite (ramm.) - löellingite (loel.), coffinite (coff.) and goethite (goet.) are after uraninite. Calcite and ferroan dolomite (darker grey) are stained with erythrite (eryt.). P.P.L.

E) Zoned, twinned safflorite starlets in calcite (calc.) and ferroan dolomite (Fe-dol.) gangue. P.P.L.

F) As E in crossed nicols, showing zoning and mimetic star-shaped twinning.

G) Chalcocite (cc.) - digenite (dig.) - covellite in actinolite (act.) replacing earlier calcite and have replaced by late euhedral quartz (qtz.). Blebs of chalcocite-covellite are found as inclusions in quartz. P.P.L.

H) Bornite (bn.) exsolving idaite?(id.) and chalcopyrite (cp.) and later digenite (dig.). Note the aggregates of hematite needles overgrowing bornite. P.P.L.

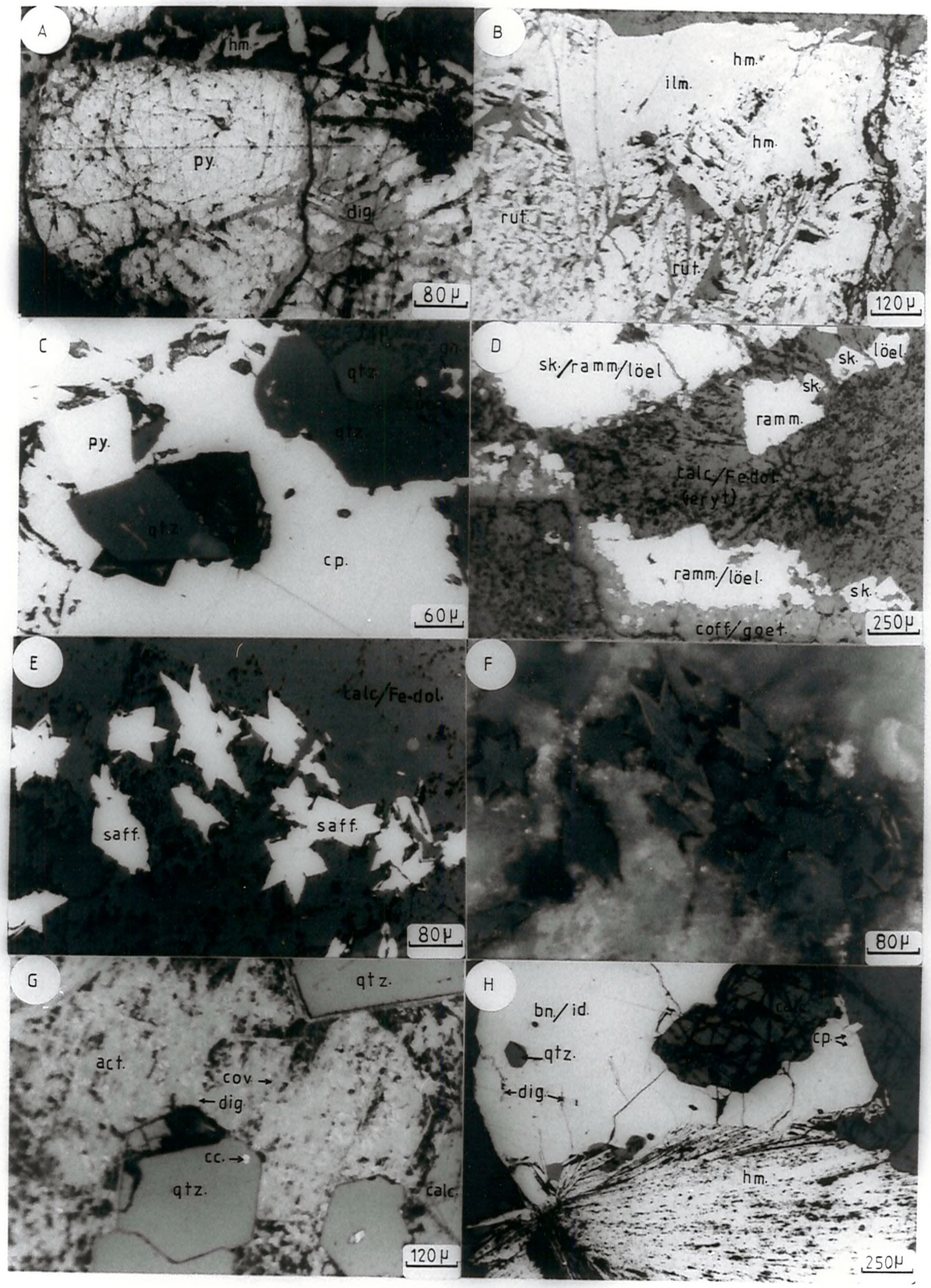

A
hm.
py.
dig.
80μ
B
hm.
ilm.
hm.
rut.
rut.
120μ
C
py.
qtz.
qtz.
qtz.
cp.
60μ
D
sk./ramm/löel
sk.
löel
sk.
ramm.
calc/Fedol (eryt)
ramm./löel
sk.
coff/goet
250μ
E
talc/Fe-dol
saff.
saff.
80μ
F
80μ
G
qtz.
act.
cov
dig
cc.
qtz.
calc
120μ
H
bn/ id.
qtz.
dig
calc
cp
hm.
250μ

**PLATE 10**

PLATE 11

West Labelle mineralised veins:

A) Uraninite (ur.) and galena (gn.) enclosed by spheroids of coffinite (coff.) - goethite (goet.), in zoned allanite. P.P.L.

B) Molybdenite (mo.) and uraninite (ur.) disseminated in calcite (calc.) and apatite (ap.) gangue. P.P.L.

C) Rounded and cubic uraninite (ur.) crystals overgrown on massive molybdenite (mo.), in calcite gangue (dark). P.P.L.

D) Fractured uraninite (ur.) consisting of lighter and darker phases and partially altered to coffinite (coff.). Radiogenic galena (gn.) is enclosed in uraninite. P.P.L.

E) Corroded uraninite (ur.) in altered allanite (allan.). Uraninite is again altered to coffinite and galena. P.P.L.

F) Zoned and altered allanite (allan.) enclosing chalcopyrite, bornite, covellite, digenite, hematite and rutile. P.P.L.

G) Uraninite (ur.) rims around apatite (ap.). P.P.L.

H) Altered allanite (allan.), again replaced by uraninite (ur.) intergrown with apatite (ap.). P.P.L.

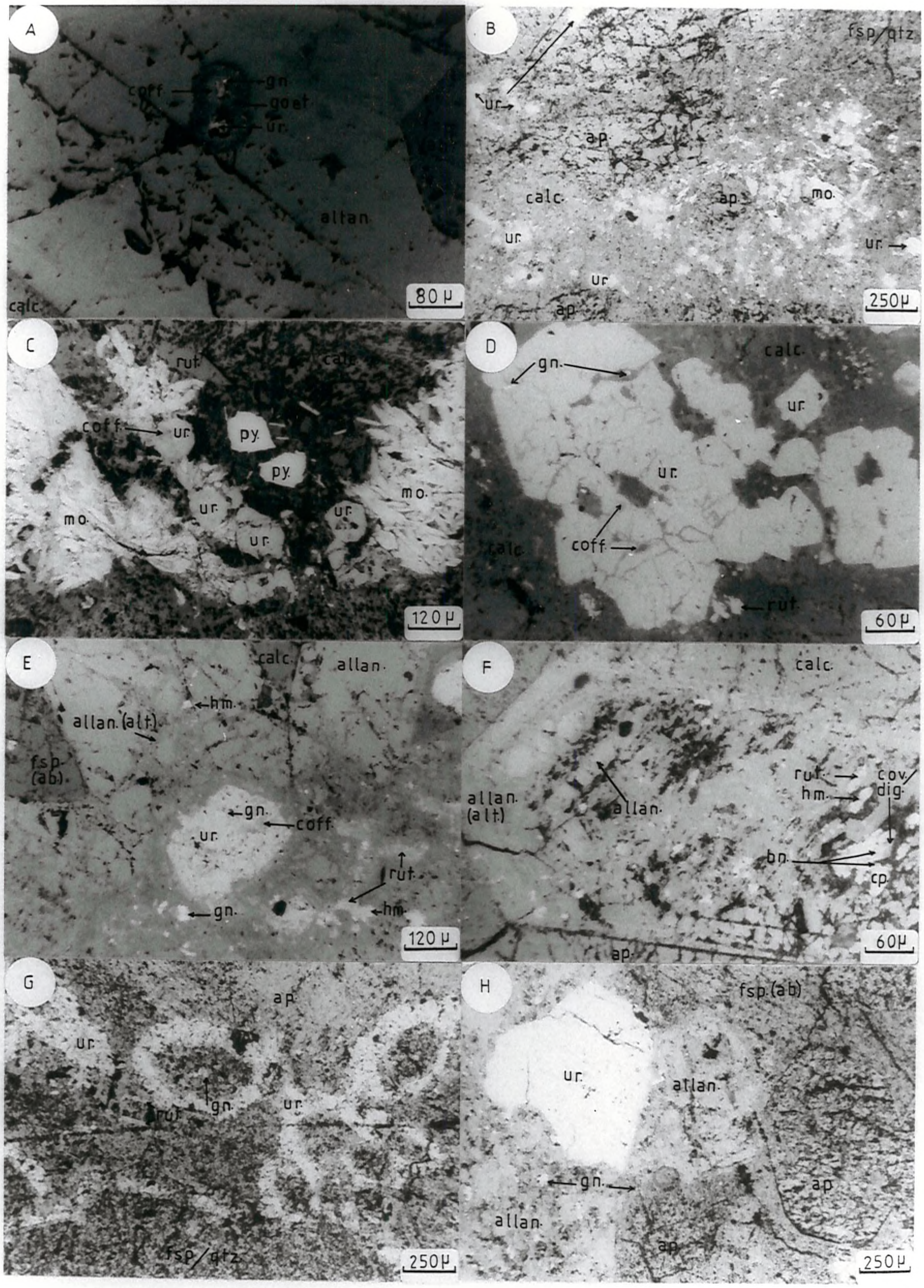

A
coff
gn
goet
ur
allan
calc
80 µ
B
fsp/qtz
ur
ap
calc
ap
mo
ur
ur
ur
ap
250µ
C
rut
calc
coff
ur
py
py
mo
mo
ur
ur
ur
120 µ
D
gn
calc
ur
ur
calc
coff
rut
60µ
E
calc
allan
hm
allan (alt)
fsp (ab)
gn
coff
ur
rut
gn
hm
120 µ
F
calc
rut
cov
hm
dig
allan (alt)
allan
bn
cp
ap
60µ
G
ap
ur
gn
rut
ur
fsp/qtz
250µ
H
fsp (ab)
ur
allan
gn
ap
allan
ap
250µ

PLATE 11

PLATE 12

West Labelle mineralised veins:

A) Stringers of native silver (Ag) in calcite (calc.) and earlier feldspar (ab.). Note the close association of silver with chalcocite (cc.). Chalcocite is intergrown with stromeyerite (str.) and contains fine inclusions of native silver. Argentite (arg.) has replaced native silver. P.P.L.

Ac) Atomic number distribution of chalcocite-stromeyerite + silver grain in A - electron probe microphotograph.

(Ag) Silver content. Ag $K\alpha$.

(Cu) Copper content. Cu $K\alpha$.

(S) Sulphur content. S $K\alpha$.

B) Native silver (Ag) bleb in chalcocite (cc.). Note the euhedral quartz crystals overgrowing chalcocite. P.P.L.

Bc) Atomic number distribution of the same sample as B but of different silver grain - electron probe microphotograph.

(Ag) Silver content. Ag $K\alpha$.

(Cu) Copper content. Cu $K\alpha$.

(S) Sulphur content. S $K\alpha$.

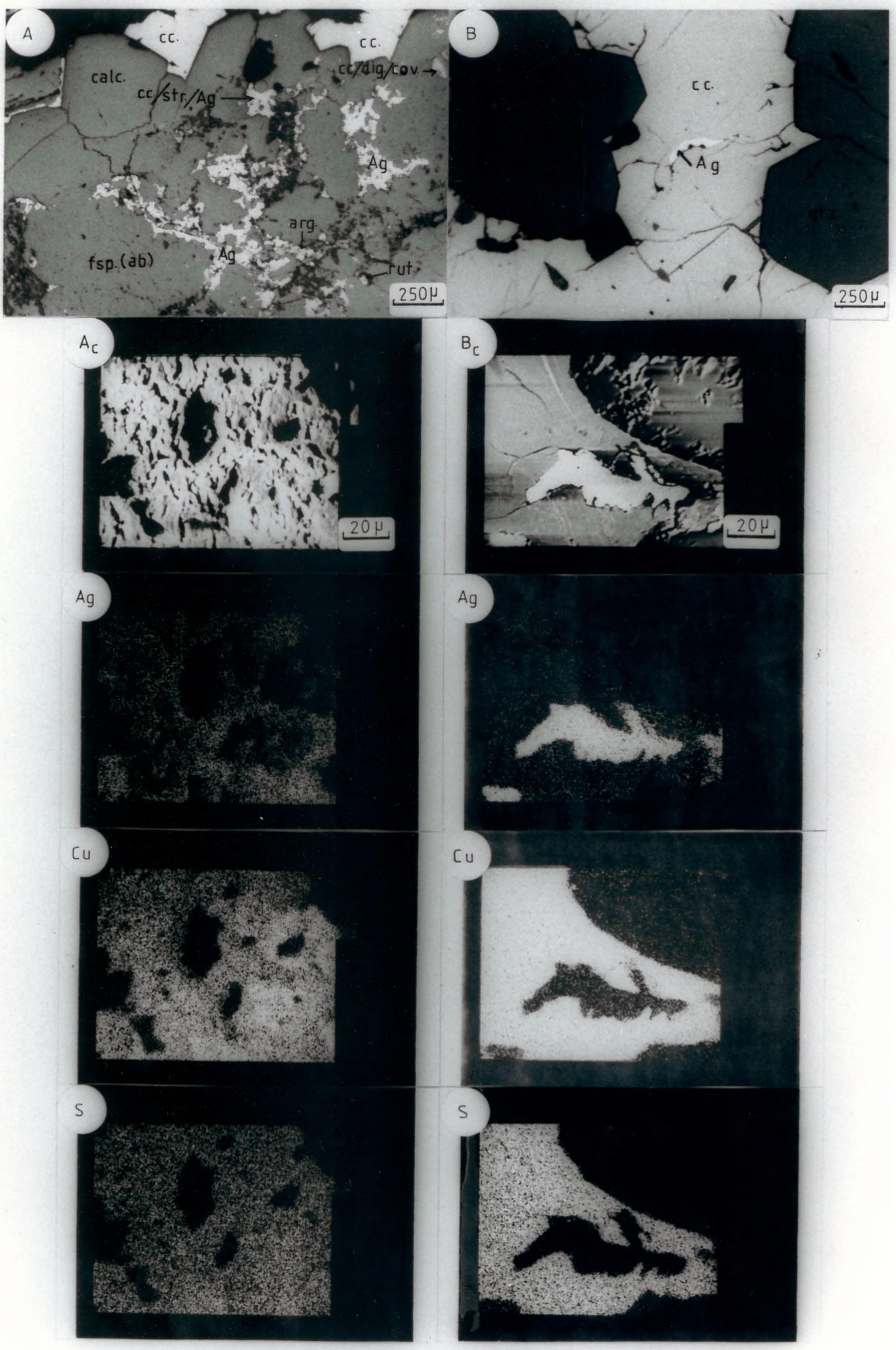

A
cc.
cc.
calc.
cc/str/Ag
cc/dig/cov
Ag
Ag
arg.
fsp.(ab)
Ag
rut.
250µ
B
cc.
Ag
qtz
250µ
Ac
20µ
Bc
20µ
Ag
Ag
Cu
Cu
S
S

PLATE 12

PLATE 13

West Labelle mineralised vein:

A) Inclusions of later electrum (Au, Ag), bismuthinite (bism.), chalcopyrite (cp.), bornite (bn.) and galena (gn.) in euhedral cobaltite (cob.). Some earlier calcite grains are also enclosed. P.P.L.

Ac) Atomic number distribution of A - electron probe microphotograph.

(Au) Gold content (the high background is due to gold-coating during analysis). Au K$\alpha$.

(Ag) Silver content. Ag K$\alpha$.

(Bi) Bismuth content. Bi K$\alpha$.

(Ni) Nickel content. Ni K$\alpha$.

(Fe+Co+Cu) Iron, cobalt and copper contents. Cu only present where indicated. Fe K$\alpha$, Co K$\alpha$ and Cu K$\alpha$.

(As+S) Arsenic and sulphur contents. As K$\alpha$ and S K$\alpha$ superimposed.

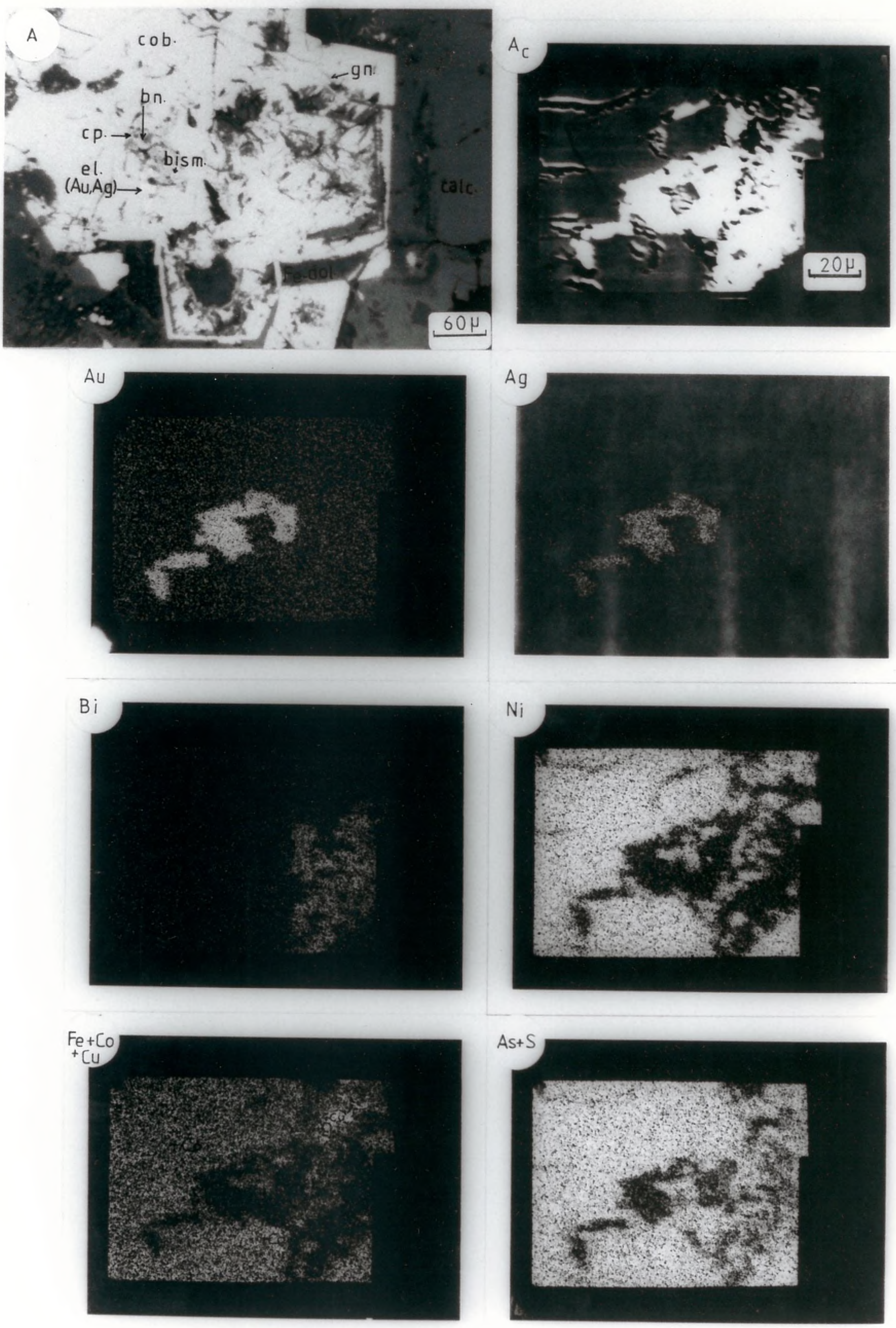

A
cob
gn
bn
cp
bism
el.
(Au,Ag)
talc
Fe-dol
60μ
Ac
20μ
Au
Ag
Bi
Ni
Fe+Co
+Cu
As+S

PLATE 13

stringers and patches of fine-grained diorite and felsite. These in turn are cut by pegmatitic veins of actinolite-apatite-hematite/ilmenite/rutile. The centres of the veins often contain a little carbonate (mostly cacite and some dolomite) and chalcopyrite. In one area similar veins are frequently filled with carbonate-quartz (locally with allanite) veins with Ni-Co-Fe arsenides, native silver, stromeyerite, argentite, proustite (ruby silver), molybdenite, magnetite-hematite-ilmenite-rutile, uraninite-coffinite and uranium secondaries, bornite-idaite, chalcocite, cuprite, and other supergene copper sulphides, cobaltite, pyrite, marcasite, electrum (Au, Ag), bismuthinite, galena and some Bi- and Pb-sulphosalts (Plates 9, 10, 11, 12, and 13). Uranium minerals are also found in some other veins nearby. The arsenides are generally in stringers instead of having botryoidal texture. Niccolite and maucherite are rare. Rammelsbergite and skutterudite are the predominant arsenides. Star-shaped safflorite are however common and often intergrown with loellingite. Here gersdorffite is not as common as cobaltite. In one of the polished sections, marcasite is found to be replacing pyrite along the rims. This is the only sample of marcasite found during investigation.

Mineralisation- East Labelle

A small radioactive anomaly was found near the northern contact of the eastern diorite. The diorite is similar to the western one. The anomaly occurs where a breccia vein fills one of these joints. The breccia is up to 10 metres wide and 50 metres long, and the average width is about 1 metre. The breccia consists of a jumble of angular fragments of ordinary diorite and of a fine-grained variety cemented by magnetite, apatite and actinolite, and some carbonate. Most of the zone consists about 90% of breccia fragments which have been torn off the walls by the mineralising event, and the walls are permeated by stringers of magnetite and actinolite. The central parts of the zone consist of apatite-magnetite actinolite pegmatite with interstitial carbonate. The main radiometric anomaly is over the centre of the vein, but no uranium minerals were identified.

Paragenetic Sequence and Mode of Deposition

It can clearly be seen that there is a complete paragenesis in West Labelle mineralised veins. Diorite → felsites and other differentiates → Ap-Act-Mnt pegmatites (+uranium) → carbonate + arsenides → carbonate, quartz, sulphides and sulphosalts. During or just after the transition to the latest stage of the sequence, native metals were deposited. The complete paragenesis is shown in Tables 9 and 10 for the West Labelle and East Labelle mineralisations respectively. The first stage of West Labelle mineralisation contains considerable amounts of various early sulphides and oxides. Stage 2 contains the earliest true vein mineralisation, with Mnt-Ap-Act + U in the earlier part of the stage (Stage 2a) and Ag, Ni-Co-Fe arsenides in the later part of the stage (Stage 2b). Intermediate sulphides and silver-bismuth sulphosalts were deposited in Stage 3a and b respectively. Stage 4 (oxidation stage) is characterised by Cu, Ni, Co, Fe and U secondary minerals. In East Labelle, the paragenetic sequence (Table 10) contains no uranium in Stage 2 and the absence of arsenide mineralisation. Otherwise the stages of mineralisation are similar to those at West Labelle.

The two showings (East and West Labelle) demonstrate different manifestations of the same phenomenon. As was mentioned earlier, the West Labelle mineralisation is similar to that of Regina Bay, which was produced in an open fissure system. West Labelle, however, provides the complete spectrum of mineralisation. The eastern showing (East Labelle) is an area where there was no open fracture system during cooling and the volatile pressure in the core of the intrusion exceeded the strength of the cool upper margin. Hydrothermal streaming through the fracture produced mineralisation and immediately sealed the system, preventing from further mineralisation.

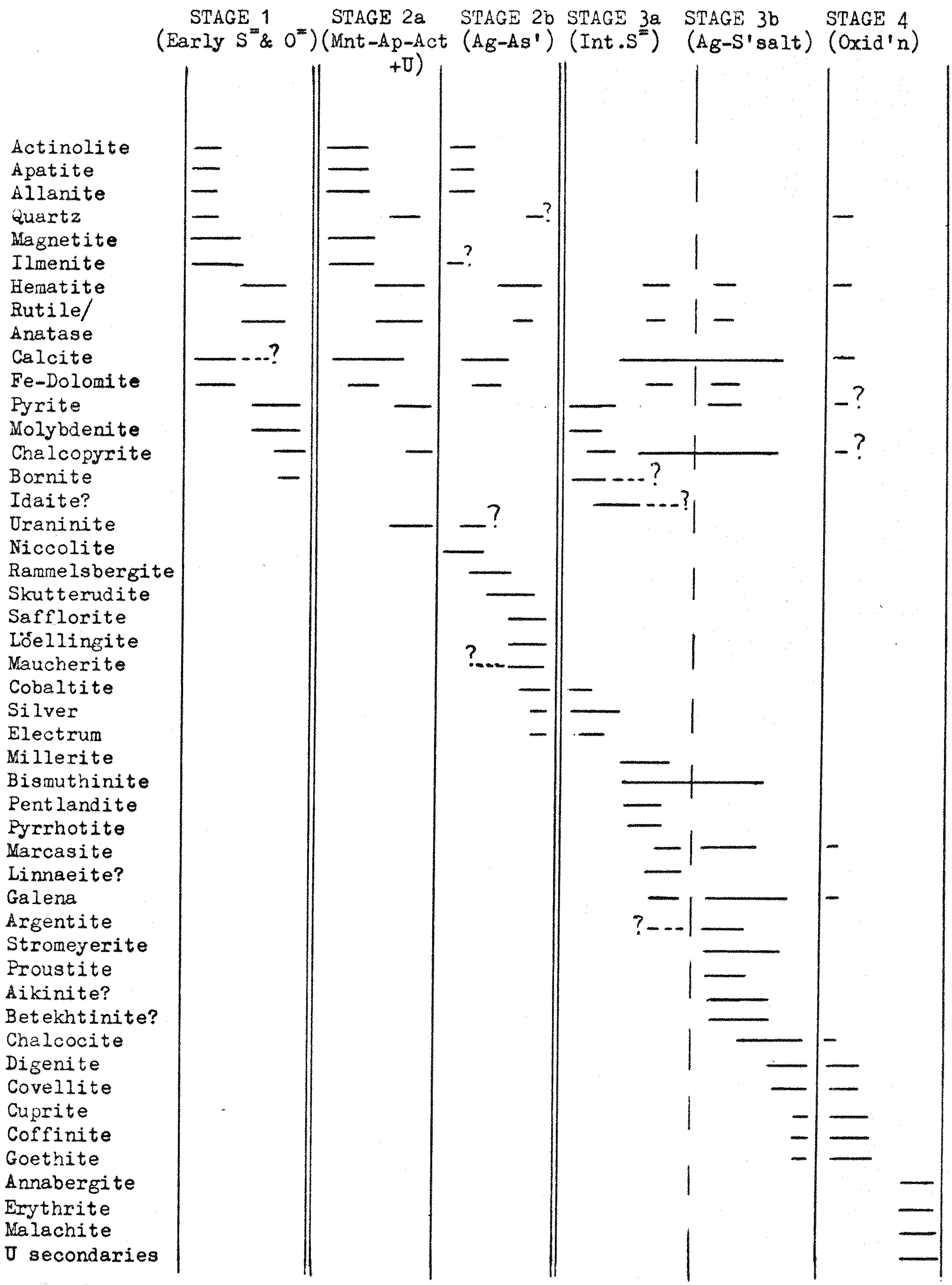



Table 9: Paragenesis of mineralised veins at West Labelle.

Double lines represent main periods of fracturing and brecciation.

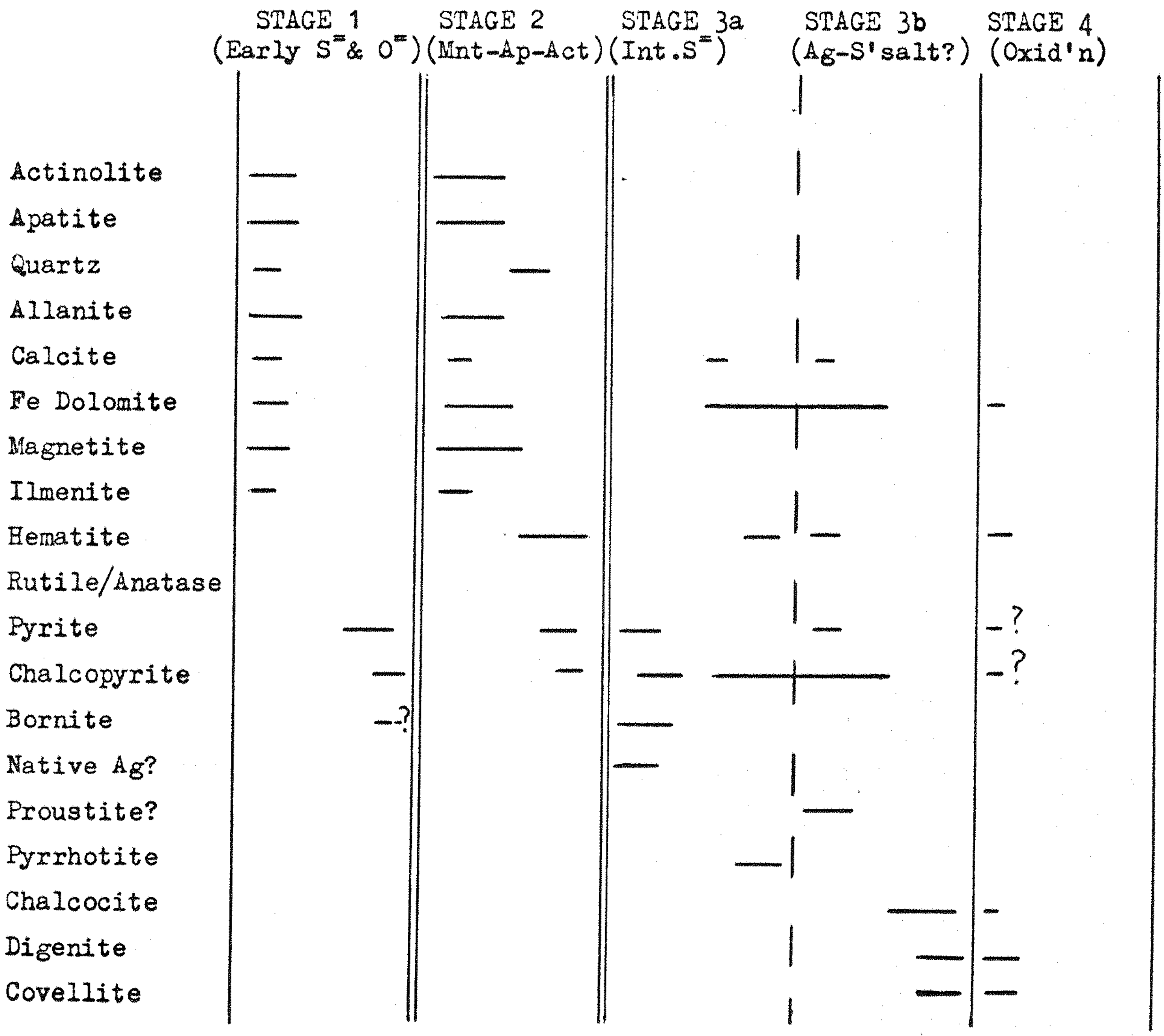



**Table 10:** Paragenesis of mineralised veins at East Labelle.
Double lines represent main periods of fracturing and brecciation.

f. Sachowia Lake

## History

Niccolite mineralisation in shear zones in Archean Yellowknife Group metavolcanics was discovered in 1940, but although considerable drilling and trenching was carried out in the fifties, it was not until 1969 to 1971 that any ore was mined. Over this period a few tons of high grade nickel ore was shipped from a reserve calculated at some 5000 tons. The property has remained unworked since 1971.

## Host Rocks

Archean granite, gneiss and metavolcanics all outcrop on the shores of Sachowia Lake (Maps 10 and 14). These were cut by a NW trending diabase dyke and subsequently a NE trending dyke. Granites, gneiss, volcanics and dykes are all altered and faulted, whereas the Kahochella sediments are relatively fresh and undisturbed. The Kahochella Group here consists of interbedded shales, tuffs and agglomerates, with rare intermediate flows, overlain by oolitic hematite ironstone.

Most of the rocks in the area are cut by $20^{\circ}$-striking fracture zones which, in three areas contain dykes of red felsite. The margins of the felsites have sheared, but the felsites themselves are fresh and undeformed.

## Mineralisation

There are three zones of mineralisation corresponding to the three red felsite dykes, an East, a Main and a West zone. In each of these zones the red felsite has been brecciated on one margin by further movement on the fracture that permitted the initial intrusion. The breccias have been cemented by yellow and white carbonate (mostly ferroan dolomite with some calcite) and quartz. Open spaces in the breccia cement and dilatancies, the later engendered by further small faults movements, have been filled by ore minerals, of which niccolite is predominant. In addition skutterudite, rammelsbergite, gersdorffite, cobaltite, löellingite, safflorite, pyrite, chalcopyrite, parkerite, hauchecornite, bravoite, native bismuth, bismuthinite and some galena

PLATE 14

Sachowia Lake mineralised veins:

A) Hand specimen of breccia containing nickel-cobalt-iron arsenide mineralisation in the matrix. Erythrite (eryt.) is shown on the weathered surface.

B) Hand specimen showing arsenide mineralisation in ferroan dolomite and calcite gangue.

C) Hand specimens showing arsenide mineralisation in ferroan dolomite gangue within felsite (felst.) host rock.

D) Hand specimens showing arsenide and bismuth mineralisation in ferroan dolomite-calcite gangue.

E) Hand specimen showing banded arsenides with bismuth in felsite host rock.

F) Hand specimen showing fractured and brecciated arsenides.

G) Inclusions of native bismuth (Bi) in earlier rammelsbergite (ramm.). Note the alteration of native bismuth to bismuthinite (bism.) along the edges. Other inclusions include chalcopyrite (cp.). P.P.L.

H) Fine inclusions of native bismuth (Bi) in both safflorite (saff.) and rammelsbergite. Much of the bismuth has been altered to bismuthinite (bism.). P.P.L.

deposition (e.g. at the base of Blanchet Island laccolith). In the volcanic system, with the bulk of magmatism being at higher levels, it is probable that the mineralising fluids were generated within hypabyssal magma chambers and circulated (and precipitated) within the vents (e.g. Aristifats Lake).

The similarities of all these arsenide deposits justifies the assumption of a common petrochemical process for the generation of the mineralising fluids. It has been argued that such fluids were generated in the roots of the Easter Island Dyke by breakdown of olivine and possibly also clinopyroxene and primary sulphide phases. The relatively high level of emplacement presumably permitted an effective convective circulation of water which was able to leach at least Ag, Bi, Ni, Co, Fe, As and Sb to form a metalliferous fluid depleted in reduced sulphur species. This fluid was free to rise up fractures within or around the intrusion. In the small Easter Island intrusion the rise was not great and precipitation occured close to the site of generation. However in the larger Maple (Caribou Lake) gabbroic complex, the mineralising fluid may have been driven up rather further from more mafic cumulate rocks in the roots of the complex. It is important to note that the arsenide mineralisations completely or partially replaced the earlier Mnt-Ap-Act + U, REE (especially at West Labelle).

If a common petrochemical derivation is to be accepted then it is implicit from the above that the middle and late Aphebian igneous bodies are also underlain by more mafic rocks. For the middle Aphebian volcanic rocks there is evidence of such mafic rocks in the abundant 'spilites' (Olade and Morton, 1972; Olade, 1975). It is therefore not unreasonable to propose the existence of mafic complexes (or complexes just like that of the Easter Island Dyke) among the feeder systems of this volcanism. There is no evidence for the presence of mafic roots to the late Aphebian diorites. However they are mostly high-level laccoliths and plugs which must have been derived from some magmatic feeder-system and their evolution has been discussed in Part I of this thesis. On the other hand one might argue that because of the presence of the arsenide mineralisation these feeder-systems must have contained more mafic rocks at depth.

c) Time and Duration of Mineralisations.

Considering the processes around the upper parts of the intrusions, a continuum of events can be predicted of the general type: metamorphism to metasomatism to progressively lower temperature hydrothermal events. This is a continuum both in time and space. Metamorphism in the East Arm is not very severe with regard to the Aphebian intrusions, but the Archean basement was highly metamorphosed. Burial metamorphism is rather weak and has been discussed by Walker (1977).

A time-distance plot of a single intrusion (Fig. 10) allows predictions concerning vein mineralisations to be made. At any distance (D) from an intrusive contact a certain sequence of events will occur and represented by the vertical line through D1. Some of these events may be unable to occur until a certain time - e.g. hydrothermal veins cannot develop until fractures and joints have opened (the fractures and joints may be opened by an increasing hydrostatic head or by external mechanisms such as isostatic adjustments of the intrusion). Thus there may be no hydrothermal veins until time T1, even though a hydrothermal fluid existed prior to this time. From the time T1 onwards, a regular sequence of hydrothermal events occurs in the fractures and joints until they become sealed. Thus high temperature hydrothermal events (resulting the sulphide and oxide mineralisation in mafic and ultramafic cumulates in early intrusions) can be clearly distinguished from the medium and lower temperature hydrothermal ones (i.e. the vein mineralisations) which would start at time T2

The relative ages of all the events around all the intrusions (either early or late) are almost similar. The polyphase magmatisms often made the above prediction quite complicated. The mineralised veins fit the latter sequence of events and also in the general paragenetic sequence of mineralisations. It is proposed that the mineralising fluids were derived via the 'felsic' intrusions regardless of their ages from the more 'mafic' intrusions.

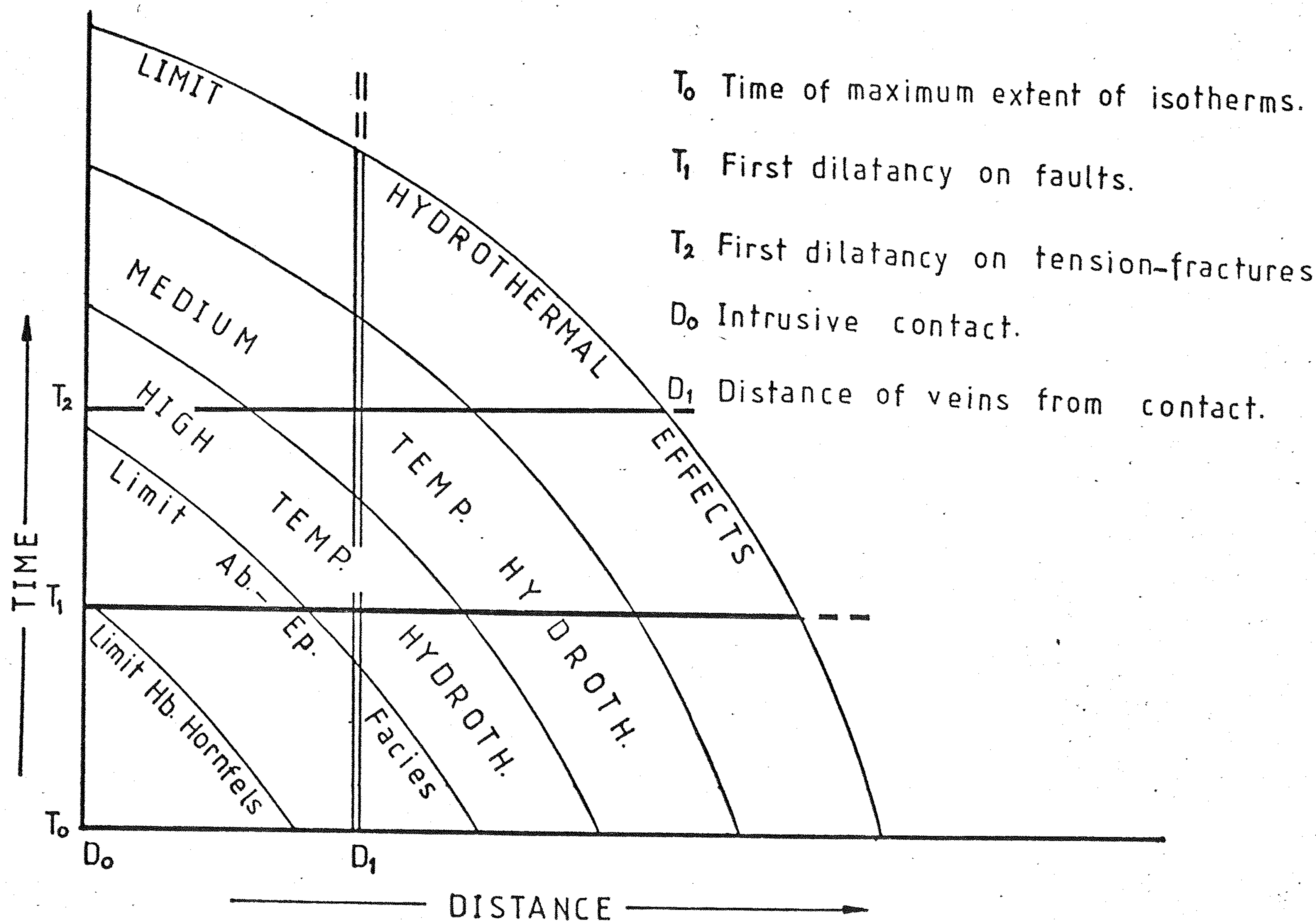



**Fig. 10: Idealised Time-Distance plot for development of veins around a cooling intrusion.**

| Gangue | Zone | Ore-Minerals (Latest Minerals) | | Economically Important Elements | Composition of Wolframite and Sphalerite | Trace Element Data* | Wall-Rock Alteration |
|---|---|---|---|---|---|---|---|
| Quartz; Chalcedony; Barite, Dolomite, Calcite; (Chlorite, dashed) | 7. | Barren: (pyrite). | *Mesothermal and Epithermal Lodes* Generally at right-angles to granite ridges. | | ↑ | ↑ | Silicified granite. Silicified slate. Skarn-type rocks derived from greenstones and calc. sediments. |
| Quartz; Chalcedony; Barite, Dolomite, Calcite | 6. | Hematite. Stibnite. Jamesonite. Tetrahedrite. Bournonite. Pyragyrite? Siderite. Pyrite: (marcasite). | | Fe. Sb. | | | Chloritized granite: peach. Chloritized slate: peach. Hematitized granite. Hematitized slate. |
| Quartz; Chalcedony; Barite, Dolomite, Calcite; Flourite | 5b. | Argentite. Galena. Sphalerite. | | Ag. Pb. Zn. U. Ni. Co. Bi. | | Bi and Sn decrease in galena. | |
| Quartz; Hematite; Flourite | 5a. | Pitchblende. Niccolite. Smaltite. Cobaltite. (Native bismuth: bismuthinite?) | | | | | |
| Quartz; Chlorite; Hematite; Flourite | 4. | Chalcopyrite. Sphalerite. Wolframite: (scheelite). Arsenopyrite. Pyrite. | *Hypothermal Lodes* Generally parallel to granite ridges and dykes. | Cu | Fe increases in wolframite and decreases in sphalerite. | In, Mn, and Sn decrease, and Ge and Ga increase, in sphalerite. | |
| Quartz; Felspar; Mica; Tourmaline; Chlorite; Hematite; Flourite | 3. | Chalcopyrite: (stannite). Wolframite: (scheelite). Arsenopyrite. Cassiterite: (wood tin). | | Sn. Cu. W. As. | | | Tourmalinized granite: capel. Tourmaline slate hornfels: capel. |
| Quartz; Felspar; Mica; Tourmaline | 2. | Wolframite (scheelite). Arsenopyrite: (molybdenite?) Cassiterite. | | Sn. W. As. | | | Greisenized granite. Quartz-sericite slate hornfels. Kaolinized granite (?). |
| Quartz; Felspar; Mica; Tourmaline | 1. | Cassiterite. Speculanite. | | Sn. W. As. | | | |
| Quartz; Felspar; Mica; Tourmaline; Flourite | Greisen bordered veins | Arsenopyrite. Stannite. Wolframite. Cassiterite. Molybdenite. | Veins often in granite cusps. | Sn. W. As. | | | |
| Quartz; Felspar; Mica; Tourmaline; Flourite | Pegmatites | Arsenopyrite, Wolframite. Cassiterite. Molybdenite. (Earliest Minerals) | | Sn. W. As. | | | |

Table 14: A Generalised Mineral Paragenesis, etc., of the Mineral Deposits of Southwest England, from Park and McDiarmid, 1964.

The general paragenetic sequence of mineralisations in the East Arm intrusions is quite close to the general paragenesis of the mineral deposits of Southwest England (Table 14, from Park and MacDiarmid, 1964) although the host rocks are different. The ore deposits lie in similarly controlled structural sites. The principal differences are that there are no tin or wolfram (tungsten) minerals found in the East Arm and furthermore the arsenide mineralisation is zoned with successively younger and perhaps lower temperature minerals deposited outwards from the core. Consequently, it is proposed that the multi-element assemblage of the East Arm vein deposits is the result of the constriction of separate stages of an evolving hydrothermal system into generally one depositional location. The assemblage is retained in the source (i.e. 'mafic' intrusions) and recirculated in 'felsic' rocks until all the elements are contained in a complex polymetallic ore-fluid and able to escape later in the sites of their deposition.

d) Comparisons with Other 'Arsenide' Deposits.

From the study of the mineralisations in the East Arm area, it is found that the arsenide mineralisation is remarkably quite similar to the other similar ore deposits in the world (Table 15).

The deposits at Great Bear Lake and Cobalt-Gowganda, Ontario, are remarkably very similar to those in the East Arm with respect to the arsenides. The above deposits have been documented in detail by Badham (1972), and Jambor and Petruk (1971) and Petruk (1972) respectively. They are similar in mineralogy, in the controls and mode of deposition and in their zoning, although the host rocks are different. The deposits at Great Bear Lake are associated with the Great Bear Batholith (late Aphebian granite pluton); at Cobalt-Gowganda the ores were suggested to have been derived from the diabase host rock (Jambor and Petruk, 1971). Halls and Stumpfl (1969) and Boyle and Dass (1971) stressed the relationships of Archean sulphide and ore-veins and proposed a leaching model. The other principal differences are that in the East Arm, not much silver is associated with the arsenides and in contrast Bi is quite abundant. Furthermore, in this special case Bi is not at all antipathetic with the nickel-rich arsenides and hence contributes a distinct difference from the association documented at Great Bear Lake and also Cobalt-Gowganda regions. But generally the preference for Bi is more towards the iron-rich arsenides.

The Jachymov, SW England and Kongsberg deposits also show a similar paragenesis and mineralogy. The almost identical deposits of Jachymov and SW England are preceded by a period of molybdenum, wolfram, tin and arsenopyrite mineralisation but at Jachymov the products of each stage coincide in space. Similarly, the deposits at Kongsberg are preceded by veins containing pyrite, pyrrhotite, sphalerite, galena and molybdenum. In the East Arm, the arsenide mineralisation is preceded by veins with molybdenite, pyrite, chalcopyrite, pyrrhotite, bornite, millerite, bravoite, pentlandite and oxides such as magnetite, hematite, rutile and uraninite. However all these mineralisations show some sort of telescoping. The deposits at Kongsberg are localised where veins cross pre-existing sulphide-rich fahlbands (Vokes, 1967).

Kroutov (1972), in describing the veins of the Khouvu-Axy district relates them to hydrothermal solutions percolating from the 'foyer abyssal' up regional faults. The paragenesis, mineralogy and morphology of the veins is similar to some deposits of the East Arm veins, and the minerals were emplaced after a period of skarning and metasomatism around intermediate plutons.

The deposits in the Chalanches, France, area post-date sulphide-carbonate veins. Their mineralogy and paragenesis are similar to that of the East Arm. Ypma (1972) implies that these veins may have their source in diabase (same as the Cobalt-Gowganda deposits), in which they lie. He notes that depositional temperatures range between 280°C and 150°C, perhaps equivalent to those of the East Arm.

In many of the mining provinces in Mexico and the Andes, at low temperature deposits of silver, copper and sulphosalts are typical of the mines. Gillerman (1968) relates the very similar New Mexican veins to Early Tertiary monzonite intrusions and has shown that silver and sulphosalt mineralisation overgrows older uranium and arsenide mineralisation at depth in the Bullard Peak veins. Ramdohr (1969) has noted the supraplutonic-subvolcanic nature of the Sorpresa (Bolivia) and Cusco (Peru) deposits. These two deposits show the typical Ni-Co etc. assemblage in areas otherwise characterised by the later evolved Ag-Bi-sulphosalt veins.

| Age and Deposit | Main elements other than Ni-Co-Fe-As-Ag | Host Rock | Associated Intrusions | Theories of Origin | Reference |
|---|---|---|---|---|---|
| 1700-1400 m.y. | | | | | |
| East Arm, Great Slave Lake, Canada. | Mo-Cu-Pb-Bi-U-S | Gneisses, limestone and 'diorites'. | Intermediate dykes, stocks and laccoliths. | Hydrothermal from intrusions. | This work. |
| Great Bear Lake, Canada. | Cu-Pb-Zn-Bi-Sb-U-S | Intermediate volcanics. | Calcalkaline plutons. | As above | Badham, 1973 |
| Thunder Bay, Canada. | Cu-Pb-Zn-Sb-S-C | Clastic sediments. | 'Granite' | As above | Bastin, 1939 |
| Cobalt-Gowganda, Canada. | Mo-Cu-Pb-Zn-Sb-S | As above | Nipissing Diabase. | 1. From diabase.<br>2. From hidden granite.<br>3. From leaching of sulphides. | Jambor, 1971<br>Bastin, 1939<br>Boyle and Dass, 1971 |
| 400-250 m.y. | | | | | |
| Kongsberg and Modum, Norway. | Pb-Sb-S-Ti | S-rich schists? and gneisses. | Vinor Amphibolite. | 1. From parent magma of amphibolite.<br>2. From sulphides Fahlbands. | Gammon, 1966<br>Vokes, 1967 |
| Erzegebirge, Germany. | Cu-Pb-Zn-Sb-Bi-S-Sn-U | Mica schists | Calcalkaline plutons. | Hydrothermal from intrusions. | Stanton, 1972 |
| Jachymov, Czechoslovakia. | As above | As above | 'Granitoids' | As above | Pavlu, 1971 |
| Lainijaure, Sweden. | No Ag | ? | ? | ? | Ramdohr, 1969 |

Table 15: Age, Chemistry, Host Rock and possible sources of Ni-Co-Fe-As-Ag Deposits in the world.

Continued overleaf.

| Age and Deposit | Main elements other than Ni-Co-Fe-As-Ag | Host Rock | Associated Intrusions | Theories of Origin | Reference |
|---|---|---|---|---|---|
| <u>400-250 m.y. continued</u>. | | | | | |
| Khouvu Axy, U.S.S.R. | Bi-S | Sediments and volcanics. | Intermediate plutons and basic sills. | Hydrothermal up faults from "Foyer abyssal". | Droutov, 1972 |
| Cornwall, U.K. | Sn-W-Cu-Pb-Zn-U-S | Clastic | Calcalkaline batholiths | Hydrothermal from batholiths | Park and McDiarmid, 1970 Stanton, 1972 |
| Bou Azzer, Morocco. 'Age Uncertain' | Au-Cu-U | ? | ? | ? | Ramdohr, 1969 |
| Pozoblanco, Spain. | Pb-Zn-Bi-Sb-S | ? | ? | ? | Ramdohr, 1969 |
| Sarrabus, Sardinia. | Cu-Pb-Zn-Bi-Sb-S-Ba | Black Schists | Granite/Aplite porphyry. | Hydrothermal from intrusions | Stanton, 1972 |
| Dobschau, Hungary. | Cu-S-B | Diorite | Diorite | As above | Stanton, 1972 |
| <u>100-10 m.y</u>. | | | | | |
| Chalanches, France. | Cu-Pb-Zn-Bi-Sb-S-Au-Hg | Diabase | Diabase | From diabase magma? | Ypma, 1972 |
| Kalterberg and Val d'Annivers, Switzerland. | Cu-Pb-Zn-Sb-S | ? | ? | ? | Ramdohr, 1969 |
| Schladming, Austria. | No Ag | ? | ? | ? | Ramdohr, 1969 |
| Talmessi, Iran. | Cu-Sb-S | ? | ? | ? | Ramdohr, 1969 |

Table 15: Continued from previous page. Age, Chemistry, Host Rock and possible sources of Ni-Co-Fe-As-Ag Deposits in the world.

Continued overleaf.

| Age and Deposit | Main elements other than Ni-Co-Fe-As-Ag | Host Rock | Associated Intrusions | Theories of Origin | Reference |
|---|---|---|---|---|---|
| 100-10 m.y. continued. | | | | | |
| Chanarcillo and Japonesa, Chile. | Cu-Bi-Sb-S-C | ? | Calcalkaline intrusions. | Hydrothermal from intrusions | Ramdohr, 1969 |
| Cusco, Peru. | No Ag | ? | As above | As above | Ramdohr, 1969 |
| Sorpressa, Peru. | Cu-Pb-Zn-Bi-S-C | ? | As above | As above | Ramdohr, 1969 |
| Bullard Peak, New Mexico, U.S.A. | Cu-Pb-Zn-Bi-Sb-S-U-Cl | Gneisses | Monzonite | As above | Gillerman, 1968 |
| Wickenberg, Arizona, U.S.A. | As above | As above | Granitic pegmatites | As above | Stanton, 1969 |
| Batipolas, Mexico. | Pb-Zn-Sb-S-Ba | Andesite/ Diorite | Granite | As above | Stanton, 1972 |
| Sabinal, Mexico. | Cu-Pb-Zn-Sb-Bi-S | Limestone | Alaskaite | As above | Stanton, 1972 |
| Age Unknown | | | | | |
| Shashani, Rhodesia | None | ? | ? | ? | Ramdohr, 1969 |

Table 15: Continued from previous page. Age, Chemistry, Host Rock and possible sources of Ni-Co-Fe-As-Ag Deposits in the world.

Finally it has been shown that many of the Ni-Co etc. deposits have been, and most may be, related to hydrothermal activity around large intermediate intrusions. The deposits at Cobalt-Gowganda remain a unique and special case, since they generally occur in a restricted structural site in diabase and evolved in many periods of mineralisation. Otherwise most of these Ni-Co etc. deposits in the world are more or less similar with regard to their mineralogy, paragenesis, structural controls and mode of deposition. In the East Arm the Ni-Co-Arsenide deposits are very closely related to the intrusions regardless of their age (lower, middle and late Aphebian).

## 7. SUMMARY AND THE ORIGIN OF THE SILVER, BISMUTH NICKEL-COBALT ARSENIDE ORE TYPE

It has been shown that the nickel-cobalt arsenide deposits throughout the world occur typically in podiform hydrothermal veins. The ore-association is rather complex. This is because of polymetallic assemblages and also because of extended sequences of deposition and a resultant host of accretionary and replacement textures. Earlier, it has been shown that the aresenide mineralisation in the East Arm is a continuum of events in and also around the alkaline and calc-alkaline intrusions. In Part I it has been argued that these intrusions are the products of continent-margin tectonics and perhaps partial melting beneath the edge of this continent. Furthermore it was argued that the metals in the ore-veins were derived from the intrusions mentioned above. The validity of this hypothesis may be demonstrated by the comparisons with other similar deposits in the world.

The East Arm deposits and some other deposits in the world are summarised in Table 15. The table demonstrates a number of factors common to these deposits. However it is clear that the Ni-Co etc. deposits of the world fall into three distinct age groups, i.e.: firstly, 1700 - 1400 m.y. (associated with Aphebian orogenic events); secondly, 400 - 250 m.y. (associated with Hercynian orogenic events); and thirdly, 100 - 10 m.y. (associated with Alpine/Andean/Laramide orogenic events). These orogenic periods tend to be metallogenic eras for the deposits, not exactly due to anomalous metal content in these areas, but to the coincidence of a number of geotectonic processes. Consequently it is proposed that the deposits form an intimate part of each of these orogenic events, regardless of their origin. It is found that each of these orogens, apart from the Alpine, is of continent-margin (i.e. Andean) type. In each case these ore-deposits are emplaced in the orogenic hinterland.

The ore-deposits are usually part of a polymetallic association that typically includes the following elements: Cu, Pb, Zn, S, Bi, Sb, As, U, and less commonly, Hg, Mo, Sn, W, as well as Ni, Co, Fe and Ag. Generally the Ni-Co etc. veins are pre-dated by sulphide and sulphosalt mineralisation. This 'telescopic' manner of deposition demands not only a common source for the ore-fluids but also a common environment of deposition. This environment must relate in some ways to the history of continent-margin orogens.

In many cases the veins have been ascribed to deposition from hydrothermal fluids that originated in various types of intermediate to acidic igneous rocks. The environment of deposition is controlled by the structures caused by intrusion of these igneous bodies. It is important to note that these igneous rocks, wherever analysed, are calcalkaline to alkaline in nature i.e. typical of the inner parts of continent-margin orogens.

Other propositions for the origin of the Ni-Co-arsenide ore-deposits include lateral secretion from pre-existing sulphide bodies (Cobalt-Gowganda and Kongsberg), and also from diabase magma (Cobalt-Gowganda, Kongsberg and Chalanches). It is found that the deposits at Cobalt-Gowganda are unique. They are anomalous with respect to their larger magnitude and also the lack of bismuth and uranium. Among other theories for the Cobalt-Gowganda deposits is the origin in hidden or distant granitic rocks. However there is no granite of the right age exposed in the area and furthermore there is no temporal relationship between the deposits and any form of orogenic event. On the other hand the lateral secretion hypothesis is not satisfactory since it has continually been demonstrated that the sulphide acted purely as a suitable structural site for deposition. (Gammon, 1966). Therefore the origin of the deposits at Cobalt-Gowganda region is not well understood.

T A M M A T

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# APPENDIX.

## A. Mineral Identification.

Mineral identification was made by the following techniques:

### 1) Thin sections.

Thin sections were prepared for the identification of non-opaque (translucent) minerals under transmitted light microscope. From this study the nature of the host rocks is documented. Some of the thin sections were stained to determine the nature of the carbonates using the method of Dickson (1965).

### 2) Polished sections.

The preparation technique was adopted from Cameron (1961) and improved by the author. Polished sections were studied under reflected light microscope, with reflectance and Vickers Hardness Number tester.

### 3) Polished thin sections.

This is to study the relationships between opaque (ore) and translucent (gangue) minerals.

### 4) Chemical etching.

This is to determine the nature of zoning and certain chemical compositions.

### 5) Electron microprobe (semiquantitative analysis).

The polished sections were coated either with carbon or gold. Marked (ringed) areas were examined using a JEOL JXA-50A Electron Probe Microanalyser. The operating conditions were at 25 kV for all the elements determined except oxygen (10kV) and at $3 \times 10^{-8}$ amp beam current. The specific lines used for analysis of each element are given in the captions to the various plates.

6) Autoradiographs.

This is to determine the presence of radioactive minerals especially uranium minerals. The films were exposed to the $\alpha$-radiations from the polished sections, polished thin sections and also hand specimens for several days in the dark using ordinary film or in daylight using Kodak CA 80-15 (alpha-sensitive film) using the method of Basham and Easterbrook (1978). Hence the radioactive minerals, especially that of uranium, can be exactly located when the autoradiograph is superimposed on the polished sections or polished thin sections.

7) Geiger-counter detection.

This is a simple method of detecting any radioactive minerals especially in the hand specimens.

8) X-ray diffractometry.

For the X-ray diffractometric analyses, the specimen is powdered. The powder is mounted on either a metal-glass mount or on glass slides using glue or durofix and analysed using a Phillips PW 1965/30 X-ray diffractometer. The operating conditions were at 36 kV and 24 mA and using Cu $K\alpha$ radiation. The peaks for each mineral obtained, as recorded on the chart, were corrected using some standard pure minerals such as quartz.

B. X-ray fluorescence analysis.

Analyses for the elements expressed as oxides and all other elements presented in Tables 2,3 and 4 were performed by Dr. J.P.N. Badham in conjunction with his work, using the Phillips PW 1212 X-ray fluorescence spectrometer in the Geology Department of Southampton University. Water and carbondioxide analyses were done using the Thermogravimetric Balance. Results were corrected using the method of Cosgrove (1972).

ACKNOWLEDGEMENTS

This thesis was undertaken at the suggestion and under the supervision of Dr. J.P.N. Badham. His encouragement has contributed as much as his advice and help throughout the two-year work.

I am very grateful to the Federal Government of Malaysia who sponsored me to make this work possible.

In the Geology Department of Southampton University, I am most grateful to the members of the staff, especially Dr. M.E. Gosgrove and Mr. T. Clayton, who taught some of the analytical techniques. I have benefitted from informal discussions with other research students here, especially C.W. Stanworth, and Dr. Y.A. Kettaneh. I must thank them for their ideas and their assistance for the better understanding of my work.

During the preparation of my samples and analyses I was consistently helped by the technicians here, especially K. Gosden, R. Jones, J. Merefield, R. Saunders and G. Jack. The electron probe analysis was helped by Mr. E. Heath of the Materials Laboratory at the Mechanical Engineering Department of Southampton University. The photography for this thesis was made possible with the help of B. Marsh. I acknowledge them all with gratitude and appreciation.

The typing of this thesis was done by Mrs. M. Sillence. I am grateful for her work.

Finally, affectionate thanks to all the members of my family and also my friends who inspired and encouraged me to do this work.