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THE GEOLOGY OF THE HAVDALSVATN

AND INGULSVATN AREA

A geological study of igneous and sedimentary rocks

and a discussion on the palaeo tectonic

setting of the Skorovas region.

by

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The dissertation is submitted in partial fulfillment of

the requirements for the Degree of Bachelor of

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London

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The author would like to express his gratitude to Dr.C.Halls for encouragement in the field. A special thanks to Dr.I.Gibbson and co-workers at Bedford College for their technical assistance.

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Abstract

The rocks of the Havdalsvatn and Ingulsvatn area form a minor art of the Gjersvik Nappe in the Central Norwegian Caledonides and onsist of a spilitised metavolcanic sequence which have been thrusted nder a thick sequence of volcanogenic clastic sediments.

The metavolcanic sequence consists of predominantly basaltic pillow lava, which are locally intruded by gabbroic, trondhjemite and keratophyric bodies and capped by acid pyroclastics and related lavas, vasskis and jasper in order of decreasing age. Separating the metavolcanic sequence from the overlying metaconglomerates is a thin band of marble which is believed to represent the last episode of sedimentation on the volcanic basement before tectonism. The volcanogenic clastic sediments consist of a basal 'fine felsic conglomerate' which underlies a 'coarse polymict conglomerate' that grades vertically upwards to a 'fine calc conglomerate'. Phyllites and arkoses represent the finer volcanogenic sediments.

Two episodes of folding can be recognised. The first, F1 has a fairly flat lying axial plane and has produced a penetrative schistosity. The second F2, is locally developed along a series of narrow thrust zones in which the schistosity has been coarsely crenulated. Both large scale F1 and F2 fold structures are absent.

Stable trace element assemblages, in addition to the mineral assemblage of the Skorovas orebody, indicate that the basaltic rocks have originated in an island arc environment. These rocks show characteristics of island arc tholeiites which are a product of incipient arc evolution. The presence of trondhjemite and locally granodiaritic intrusion in the Skorovas region, may reflect the initial stages of calc alkaline igneous activity which was interupted by the obduction of the arc environment onto the continental mass during the Caledonian orogen.

troduction

A nearly continuous Paleazoic geosynclinal belt occurs in Norway d adjacent parts of Sweden, stretching for some 1500 km. Regional ological studies demonstrate that the belt represents an allocthonous rip of metavolcanic rocks with associated metasediments and significant nounts of both basic and acidic intrusives. The belt has been thrusted rom the west, over the Pre-Cambrian crystalline basement and its mantle f Eocambrian sediments during the climax of the Caledonian Orogeny.

Associated with the metavolcanic rocks are stratiform massive pyritic ase metal deposits. The orebodies are characteristically lensoidal haped, variable in size, and usually penetratively deformed with their nclosing host rocks. These range from greenstones, greenschists and mphibolites, depending on the regional metamorphism. Generally the etamorphic facies increases northwards from lower greenschists in Sor-Trondelag to almandine - amphibolite characteristic of the Nordland region.

Skorovas is situated in the Grong district fig. 10 within the central Norwegian Caledonides. Lying 250 km north-east of Trondheim and 200 km south of the arctic circle at latitude 64^{o} 39' north and longitude 13^{o}17' east.

Geologically Skorovas lies in a segment of the greenstone belt which forms part of the Gjersvik nappe. Though regionally the area is characterised by a series of superimposed nappe structures, which have a sense of overthrusting towards the east.

The area is economically important since containing the Skorovas orebody which is the largest in the Grong district. Feil

The region was originally mapped by S. Foslie during the 1920's and he was the first to officially record the mineralisation at Skorovas. Since the advent of large scale mining operations much attention was given to mapping directly within the vicinity of the orebody. It has only been in recent times that mapping on a systematic scale has materialised, away from the orebody. A significant contribution to the understanding of the geology has been accomplished by Imperial College field parties under the supervision of Dr. C.Halls.

The area mapped by the author lies 10 km west east of Skorovas and corporates metalvolcanic and metasedimentary rocks, within a NW - SE NE - SW ending belt 6 km long and 3 km wide, between Havdal and Inggulsvatn g11 . Exploration operations were based at a small fishing hut on e northern shores of Havdalsvatn using the mine as a supply base.

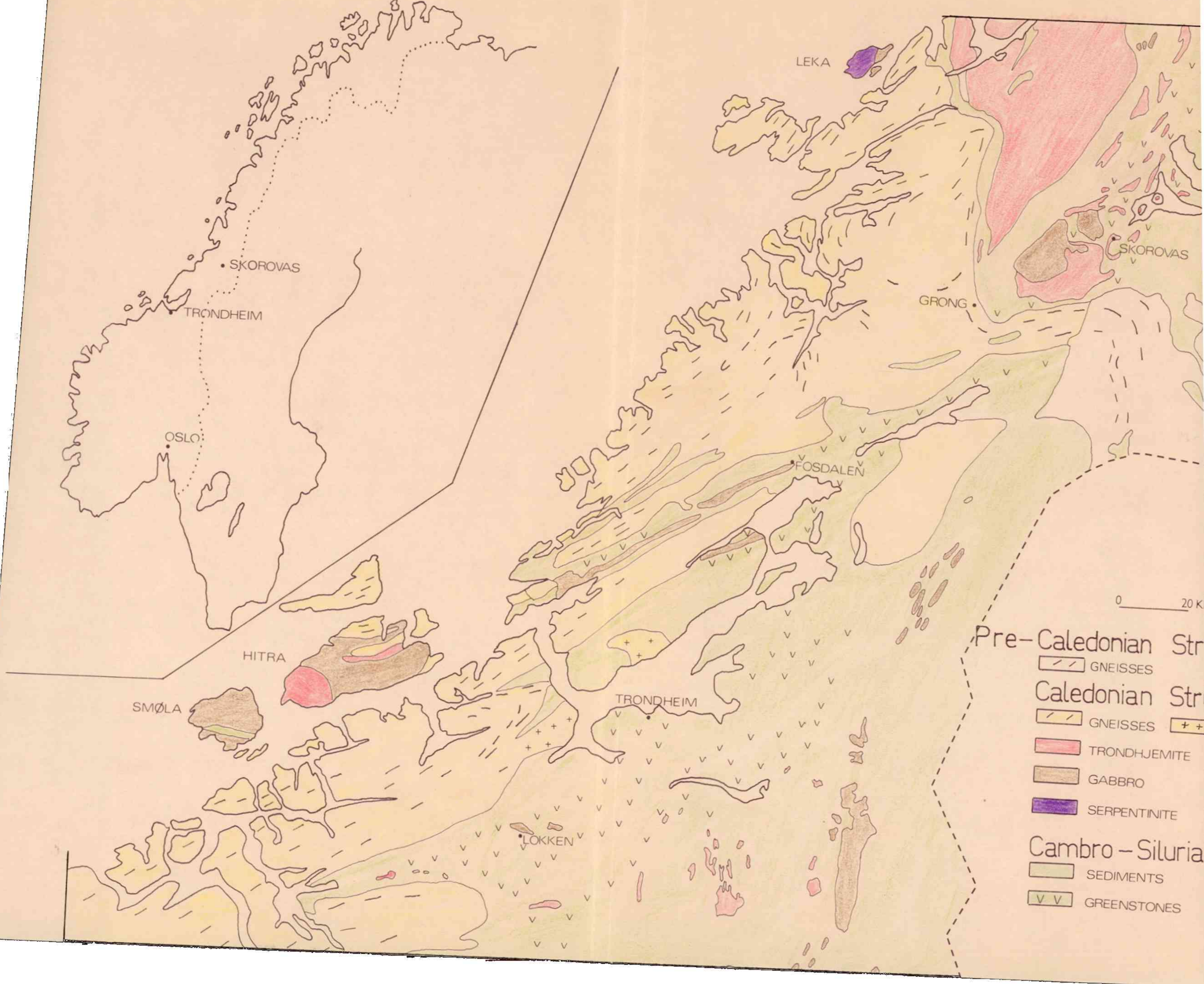
The terrain is typical of the landscape termed by the Norwegians s high fjell. A terrain of high altitude and steep slopes above the imber line. For the most part, the relief of the higher fjell lay etween 500 - 800 metres. Though towards the west, the high ground falls way down thickly forested slopes to the Tunnsjo depression. The linear idges and valley features which dominate the upland topography closely eflect the lithological and structural variation in units, parallel o the NW - SE tectonic grain. Roche moutonees litter the lower slopes, rotruding as elongated islands in a sea of glacial drift. Many small glacial rock basins support areas of marshy ground or small lakes. The larger ribbon lakes tend to be concordant to the tectonic grain of the region, thus indicating the trend of ice movement as NE - SW. Drainage in most areas is poorly developed, with only youthful streams gouging down through the rocks.

Rock exposure varied considerably from excellent on the upper fjell slopes, while the lower regions were blanketed by a variety of hardy vegetation. The diverse range of flora and fauna constituted an additional source of interest.

Precipitation is high and snow patches which were so numerous in the beginning of July gradually diminished towards the end of August. Though, by this time fresh snow was falling and the geological programme was in its final stages.

LOCATION MAP

Geological Map of Trondelag



SKETCH MAP TO LOCATE THE MAIN LITHOLOGICAL UNITS IN THE MAPPING AREA

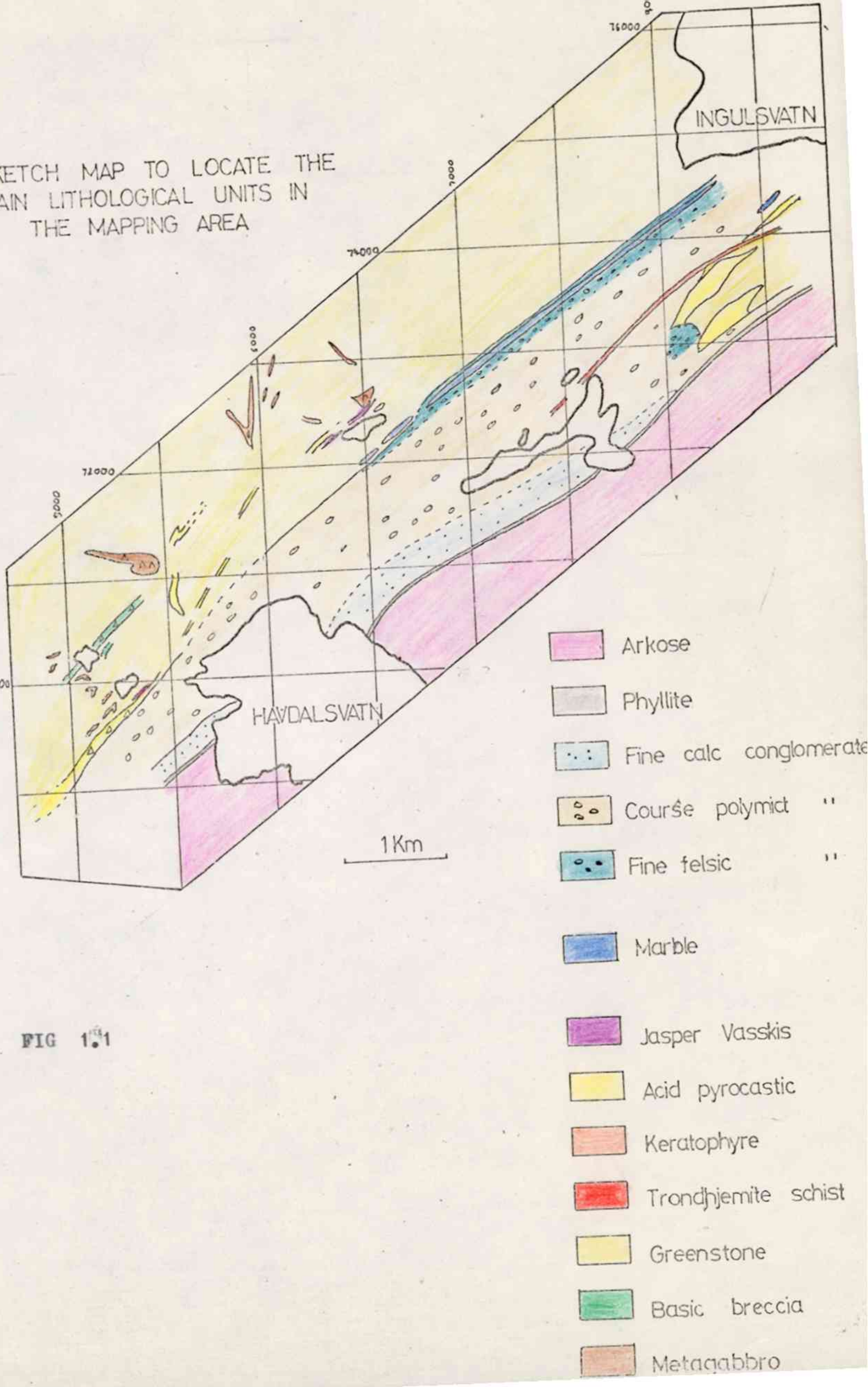


FIG 1.1

DIAGRAMMATIC STRATIGRAPHICAL SUMMARY OF THE MAPPING AREA

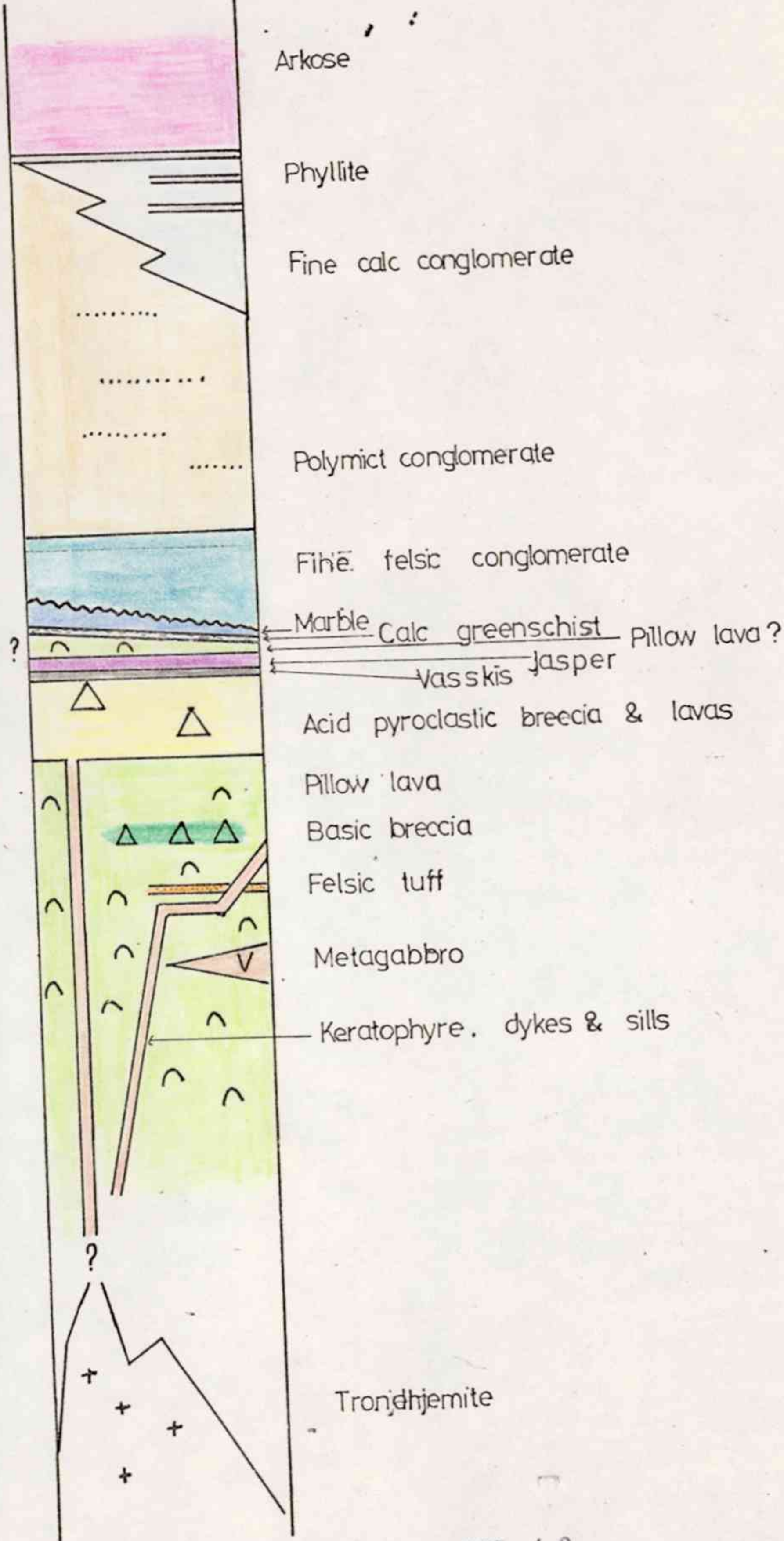


FIG 1.2

eenstone Introduction

The main lithological units are shown on the sketch map of the pping area fig. 11 accompanied by fig.12 the general stratigraphical ccession.

Greenstone is a descriptive rather than a genetic term describing altered basic igneous rock and generally applied on a regional scale describe extensive areas of such altered rocks. Here, the term will rigidly applied to basic metavolcanic rocks of greenschist facies ving a general composition of albite, epidote and chlorite.

Like the metasediments, the metavolcanics are a continuation of the orwegian greenstone belt. The metavolcanics at Skorovas are called the jersvik greenstones while those to the south of Norway, are termed the tøren greenstones. The greenstones are separated by an area called the rong Cumalation in which the Pre-Cambrian basement breaks through the reenstone belt.

The greenstones at Skorovas form a thick sequence of spilitised submarine and possibly sub aerial volcanics. They are of economic interest because the Skorovas orebody occurs within the sequence.

The Trones sheet compiled by the Norges Geologiske Undersokelse from Foslie's data show little lithological variation within the metavolcanic sequence in the mapping area. However, a number of lithological units are apparent in the field. The lack of any subdivision could be attributed to representation at the 1:100,000 scale and also, none of Foslie's field slips and notes have been published.

Geological investigations indicate two broad igneous groups within the metavolcanic sequence.

1) those of basic composition

2) those of acidic composition

The sequence also includes chemical products of volcanic exhalation in addition to the igneous rocks. These are the jaspers and varskis.

The basic igneous rocks include:

1) Pillow Lavas.
2) Basic Breccia.
3) Massive Lavas.

llow Lavas

This group forms the greater part of the metavolcanic sequence in 1e mapping area and are associated with intercalculation of more cidic lavas, pyroclastics, tuffs and dyke material. Clasical pillow tructures were observed in a number of localities, varying in size rom 1/2 metre to a metre in length. (photo 1). The pillow structures ended to be tectonically obliterated towards Skorovas and north-westwards way from the contact with the metasediments,(see fig.11). Invariably he pillows had a deformed ellipsoidal shape, which during progressive ectonic deformation, the primary fabric is destroyed and a penetrative schistosity imposed. This feature is illustrated in fig.13 . Though this was not always the case, as known pillow lavas near the mine, identified by the presence of jasper, had no penetrative schistosity imposed but appeared quite massive in texture.

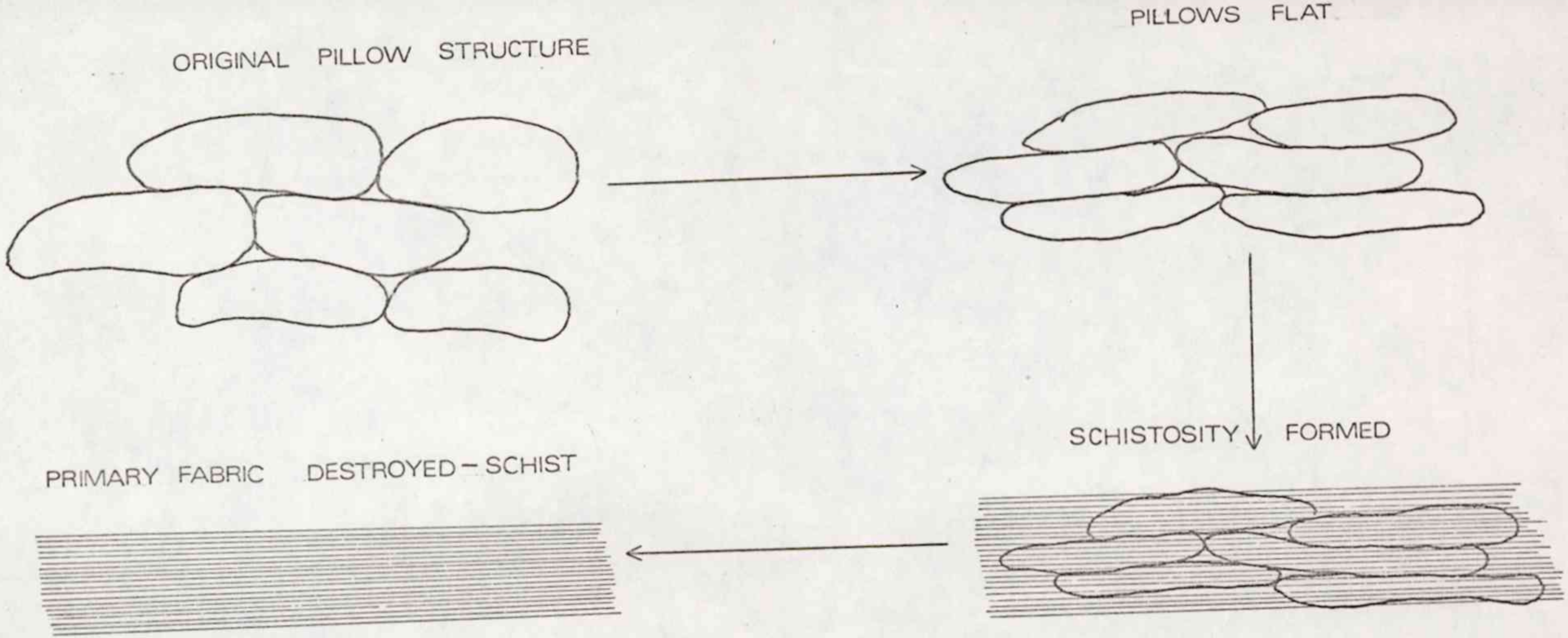
The pillows may show an internally zoned structure which comprises of an epidote core, a zone of vesicles and a chloritic rim. Fig.14 . The epidote cores are usually characterised by a disorientated fracture pattern, bearing no relationship to the regional schistosity. The fractures are exaggerated by the relative ease of weathering of the epidote. The epidote content of the pillow lavas was suggested as a criteria for sub-dividing the lavas into two units. This would be a simplification of the evidence, since the epidote has a rather eratic distribution. However, there is a general increase in epidote content, especially on a macroscopic scale, knots and veins, away from the contact with the metasediments. The general trend also corresponds with the increase in deformation of the pillow structures.

The epidote knots, (a field term used instead of epidote core when the pillow structure were not recognised), vary in size from a few centimetres to over a half metre in diameter. In the majority of cases the epidote knots have suffered progressive deformation and are flattened in the plane of schistosity. Excessive deformation has imposed a fracture cleavage on the knots, while the regional schistosity deflects around the knots. Field evidence suggests epidote genesis relates to progressive epidotization of the pillow core. Probably associated with diffusive activity of an aqueous epidote rich solution migrating towards the pillow core.

Above: Basaltic lava showing well developed pillow structures.

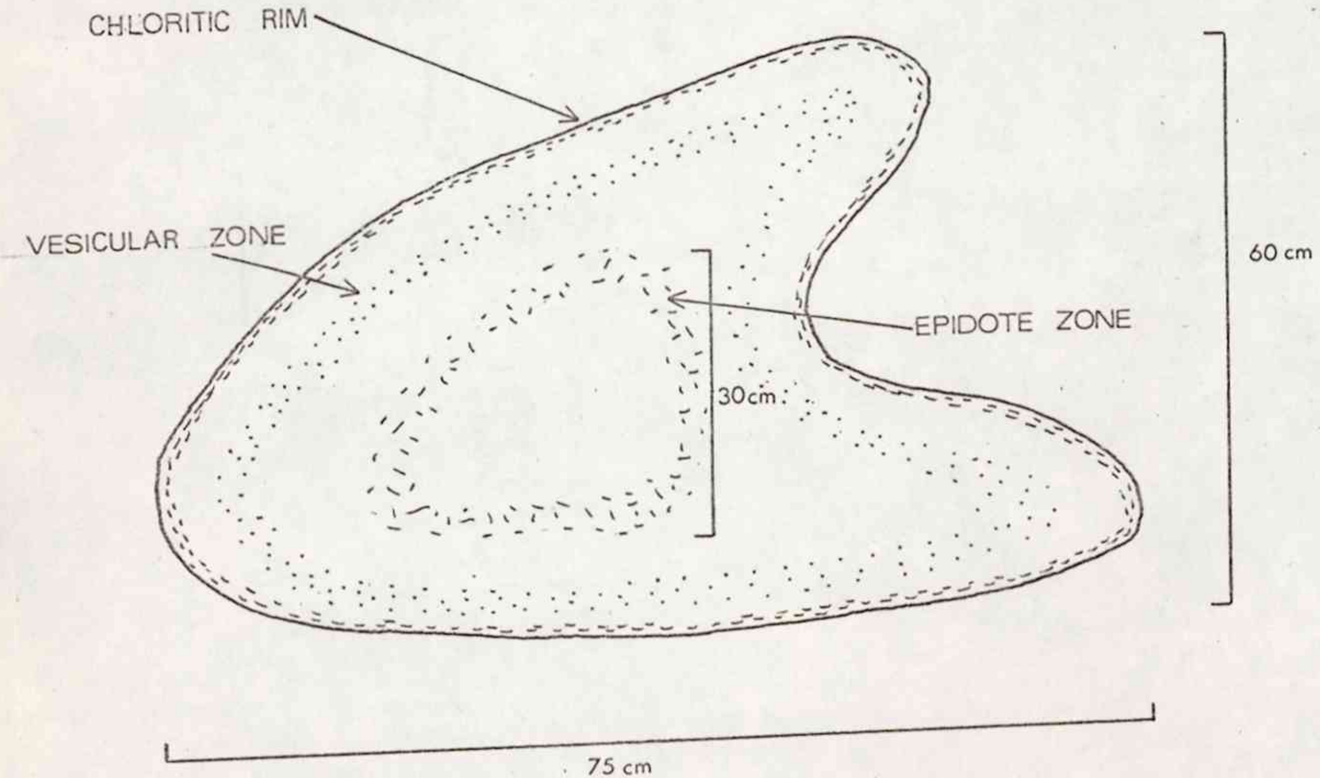
Below: The more typical appearance of the pillow lavas.

Photo 1.



PROGRESSIVE DEFORMATION IN PILLOW LAVAS SKOROVATN

FIG 1.5

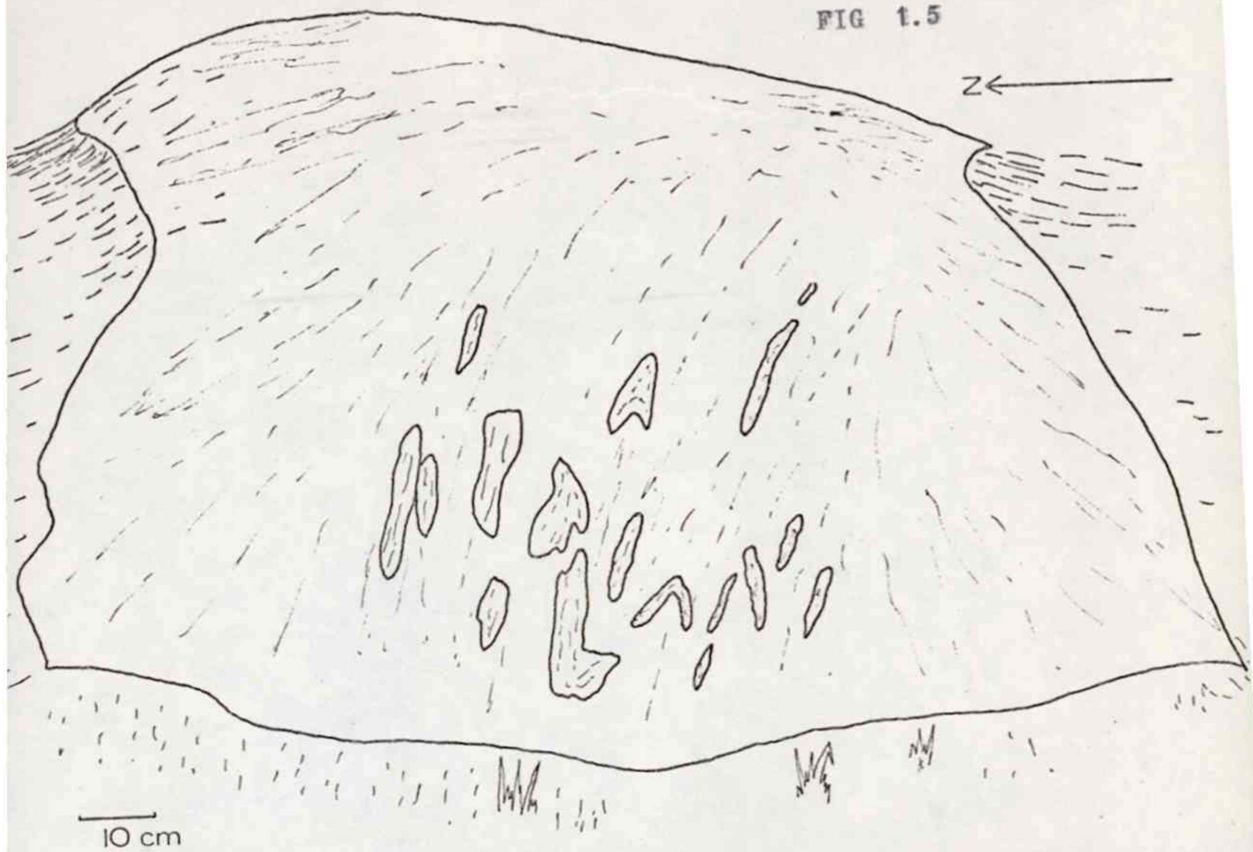


PILLOW STRUCTURE

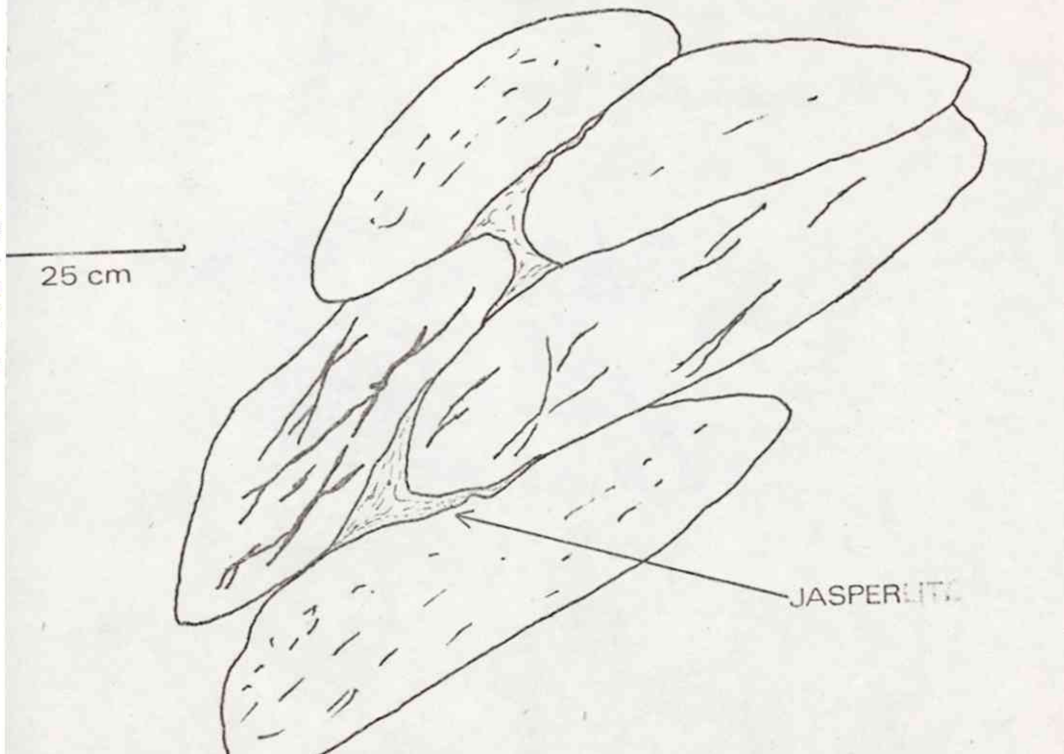
SHOWING ZONING

FIG 1.4

FIG 1.5



HIGHLY DEFORMED JASPERLITE IN PILLOW LAVAS 3440 70620



JASPERLITE IN THE CUSPS OF PILLOWS

A useful criteria was found, concerning the material which ccasionally fills the cusps between adjacent pillows, fig. 15 . n strongly deformed pillows jasper and chert occur as discordant rods. This appears to be the case within the lavas adjacent to the orebody at Skorovas. Thus, concluding the association of the orebody with a submarine environment. The criteria proved useful in the mapping area for recognising pillow lavas with an apparent lack of any primary structure.

Basic Breccia

The basic breccia forms a relatively minor constituent of the metavolcanic sequence. Their occurrance is confined to a few localities, notably 70900 3800, 70250 3050, where they are observed as discontinuous bands of variable width. In the mapping area they were seldom over 15 metres wide, but on the northern shores of Blahammeren a comparatively long exposure 1.5 km, attained a width greater than 120 metres. This might be consequential to deposition or structural control. Although it cannot be proven that the basic breccia constitutes a single horizon based on stratigraphical connotations, its field appearance does imply this.

The unit is fairly homogeneous, essentially consisting of chloritic greenstone fragments in a chloritic schistose matrix. The latter is composed of chlorite, epidote and albite. Mineralogically the fragments are a vesicular basic lava. A fine grained matrix consisting of dis-orientated albite laths, chlorite and epidote with partial replacement of feldspar phenocrysts by the latter two minerals. A few quartz grains occur in the matrix, but are usually confined to infillings of vesicles. The majority of vesicles have calcite or epidote infillings, though some are composite.

The fragments have a smooth rectangular shape and are between 3-5 cm in length. The bands are readily recognised in the field as the softer matrix is invariably weathered out, leaving the fragments protruding from the surface in a mozaic pattern.

The fragments are less deformed than one would expect, considering the relatively highly deformed nature of the adjacent pillow lavas. The random orientation of the albite laths, so characteristic of the fragments, indicate that most of the tectonic strain has been absorbed by their enclosing matrix. The absence of elongated amygdals also supports this idea. The major to minor axis of the fragments never exceeds three.

The characteristics of the breccia suggest a ~~submarine~~ pyroclastic origin, although this fails to explain the consistency in both, size range and distribution of the fragments. Another explanation involves expansion processes in situ. Similar to flow top brecciation in which the chilled surface is brecciated by movement of the underlying liquid magma. However, the thickness and only local development along a considerable strike length opposes this suggestion.

assive Lavas

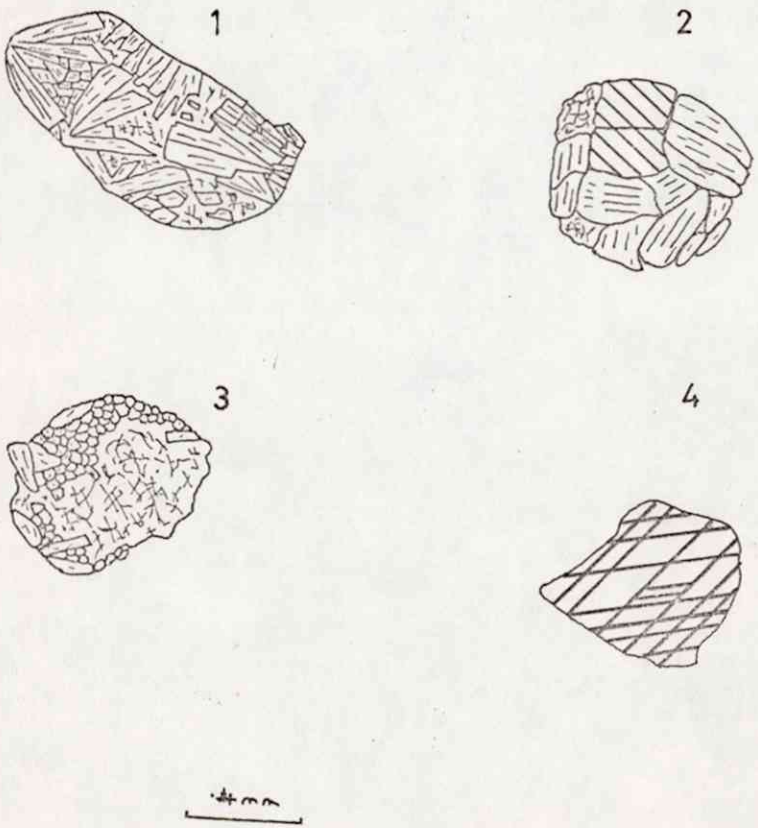
The massive lavas are the second most abundant lava type in the egion. In the field they appear to have a similar composition to the pillow lavas, but their physical appearance is different. They have a characteristically massive texture and are considerably harder than the pillow lavas. Mineralogically the two lava types are similar except that there is no preferred orientation of mineral grains within the massive lavas. There appears to be no definite contact between the two lavas except in the case where the occasional massive band occurs within the pillow lava sequence or vice versa. The lack of any recogniseable contact, relates to the problem of distinguishing a lava which shows no pillow structure and is not truely massive. However, the massive lava is devoid of jasper, vasskis and pyroclastic horizons which are characteristic of the pillow lava sequence. Also epidote knots are scarce in the massive lavas but the opposite is true for the pillow lavas.

Geochemical analysis indicates that the massive and pillow lavas are chemically similar, and it is likely that they were both derived from the same parent magma. The absence of chemical sediments and pyroclastic horizon, suggests that the massive lava results from an intrusive rather than extrusive phase of igneous activity. Possibly a near surface intrusive event.

mygdale Mineralogy

Amygdales are a common feature in both the pillows and massive .avas. They have a somewhat random distribution, occurring as small isolated patches within an exposure. The dimensions of individual amygdales varied considerably, with course 1.5 cm, and fine amygdales occurring together. Variation is also a feature of the amygdale infillings in which epidote, calcite, quartz and chlorite occur within a single handspecimen. Although all types of material constitute infillings within one exposure, there is a tendency for calcite to predominate in the upper parts of the greenstone sequence. In thin section the amygdales show a range of morphologies, from a single crystal to a composite aggregate of minerals. The latter is a common feature in the greenstones in which quartz fills the interstices of epidote. Occasionally epidote has a distinct fibrous texture (fig.1.6).

Apparently, both spherical and ellipsoidal shaped amygdales occur within a single exposure. The reason for this is hard to explain since one assumes the ellipsoidal shape is a product of deformation. The amount of strain shown by the amygdales was considerably greater in the pillow lavas than in the massive lavas.



1 Monocrystalline quartz including a radial growth of epidote.

2 Epidote aggregate with minor polycrystalline quartz.

3 Monocrystalline quartz including a fine aggregate of epidote.

4 Calcite infilling.

Sketch to show various amygdaloidal compositions

FIG1.6

Greenstone Mineralogy

It was evident that in the field that the basaltic greenstones were dominated by a mineral assemblage comprising of chlorite, albite, actinolite/tremolite, epidote and quartz. The proportion of each constituent varied considerably and no strict criteria could be used in the field for sub-dividing the greenstones. However, certain observations indicated that some of the balsaltic greenstones were trending towards andesitic composition. The presence of quartz phenocrysts coupled with a lighter matrix would suggest some acidic horizons within the greenstone stratigraphy. Also, the presence or absence of relict pyroxene phenocrysts may indicate some trends within the greenstones. However, considering the structural complexity of the region and the difficulty of making field distinctions among some of the greenstones, made the mapping of rigorous horizon a difficult matter.

The mineral assemblage of the greenstones is consistent with the greenschist facies of metamorphism. Assuming an original basaltic composition for the greenstones, a mineral assemblage consisting essentially of calcic phagioclase and pyroxene, it is apparent that a total breakdown of primary minerals has occurred.

An average or representative greenstone is a fine grained rock which may or may not be porphorytic or amygdaloidal. The groundmass consists of albite, chlorite, epidote, quartz and tremolite/actinolite in order of decreasing abundance. The chlorite, amphiboles and albite laths show a parallel to subparallel alignment which imparts a schistosity on the rock. But in the more massive greenstones there is a lack of a linear or planer fabric. A number of fresh albite laths cut the schistosity and the existence of albite crystals associated with unstrained quartz crystals, suggests a post tectonic growth. Also calcite, which occurs as a minor constituent of the groundmass bears no relationship to its surrounding minerals, showing neither alignment or deformation. Occasionally chloritic flakes have a sigmoidal nature, indicating the axis of incremental strain direction to have changed during progressive deformation. It also appears to segregate into distinct patches, unrelated to any primary mineral phase and bearing no relationship with the tectonic fabric of the rock. An obvious product of post tectonic growth.

The plagioclase phenocrysts are of albite composition, more rarely oligoclase, showing varying degrees of alteration. Very few fresh albite phenocrysts occur, but usually have corroded edges and numerous epidote inclusions. The degree of alteration is very much dependent on the size of the phenocrysts and amount of deformity suffered by the rock. At the corroded margins of the phenocrysts is invariably the mineral chlorite, while epidote seems to originate from growth within the albite crystals, because of its common occurrance as inclusions. Occasionally feldspars are completely pseudomorphed by epidote aggregates.

As previously mentioned, relict pyroxene phenocrysts are present in a number of the basaltic greenstone samples. They are invariably pseudomorphed by chlorite and/or amphiboles and the secondary amphibole is often replaced by chlorite. Some of the greenstones are associated with quartz phenocrysts and moderate quantities of quartz in the matrix. Rather than a secondary origin for quartz, it is suggested that these greenstones are in the compositioned range of an 'acidic andesite'.

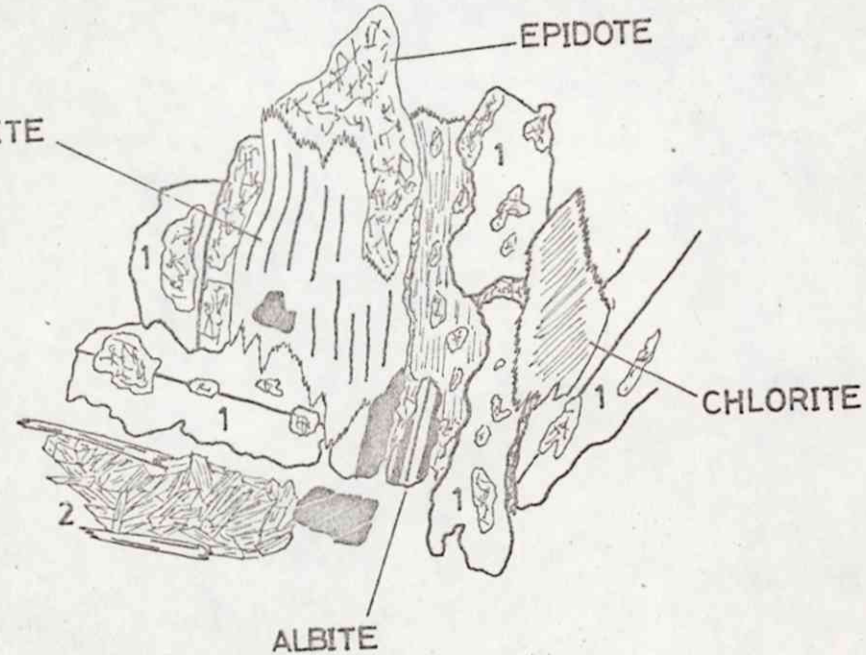
Opaque minerals are usually associated with the groundmass, showing varying degrees of alteration. Pyrite to limonite, and limonite to leucoxene. Veining is a common characteristic of the greenstones, with quartz and epidote predominating and minor volumes of chlorite and albite.

Metagabbro

The metagabbro occurs as small sheet and pod-like bodies intruded into the metavolcanic sequence. It is fine to medium grained rock, though tending to fine at the contact with the metavolcanics. Locally developed at the contact are patches of epidote enclosing small gabbroic xenoliths, thus suggesting epidotization of the metagabbro.

Mineralogically the metagabbro is completely converted to a metamorphic assemblage of minerals, though still retaining a primary gabbroic texture. The original calci plagioclase which is characteristic of gabbros, has been completely broken down to form albite and epidote (fig.1.7). Further alteration is manifested by the partial breakdown of albite which is represented by the formation of chlorite. The original pyroxenes show replacement by chlorite, actinolite, epidote and biotite. The relatively high percentage of opaque minerals is attributed to the breakdown of the primary ferromagnesium minerals and quartz, which occurs as irregular microcystalline patches, is also thought to be of secondary origin. The only macroscopic chemical and structural modifications appear as small, 5 cm, cracked epidote knots, which is rather surprising considering the large and numerous epidote knots, so characteristic of the adjacent pillow lavas.

The gabbroic sheets and pods occurring within the metavolcanic succession are interpreted as small high level intrusions closely related to the generation of the pillow sequence.



1 Partial replacement of albite by epidote

2 Chloritic aggregate pseudomorph of an amphibole

·4 mm

Sketch of Thin Section. METAGABBRO

fig 1.7

eenschist.

The term greenschist is used in the present context to describe schistose rock, lacking any primary characteristics.

The units occurs as relatively thin bands, 1-5 metres thick, through- ut the metavolcanic sequence. They are usually exposed in prominant carp features which occasionally span the whole mapping area. These eatures are clearly recognised on aerial photographs and are believed o have originated from tectonism.

Mineralogically the greenschist rock is rich in chlorite, epidote, alcite and albite. Chlorite is invariably aligned with the regional chistosity but the albite crystals only show subparallel alignment. he calcite content varies significantly throughout the greenschists orizon and those abundant in calcite have a characteristically pitted eathered surface.

A particularly interesting greenschist horizon occurs locally under- neath the marble band. Here, the horizon is dominated by the presence of carbonate rods which exceed a metre in length. Petrographically the horizon is identical to the calc greenschist bands which are seen within the metaconglomerates and therefore, indicates a sedimentary origin for this particular horizon.

A characteristic feature of the greenschist horizons is the local development of a coarse crenulated clevage and the abundance of disrupted quartz veins and stringers (photo 2). The quartz is believed to be a product of metamorphic segregation of silica during tectonism.

The tectonic features associated with the greenschist horizon suggest that the unit was formed by the accumulation of strain during progressive deformation. They are generally conformable with the lithological boundaries of the mapping area, but have an anastomosing nature which would seem to oppress a sedimentary origin.

Quartz associated with the greenschist lithology.

Photo 2.

elsic Rocks.

The felsic rocks form a small proportion of the metavolcanic sequence. Three types are described in the following discussion:

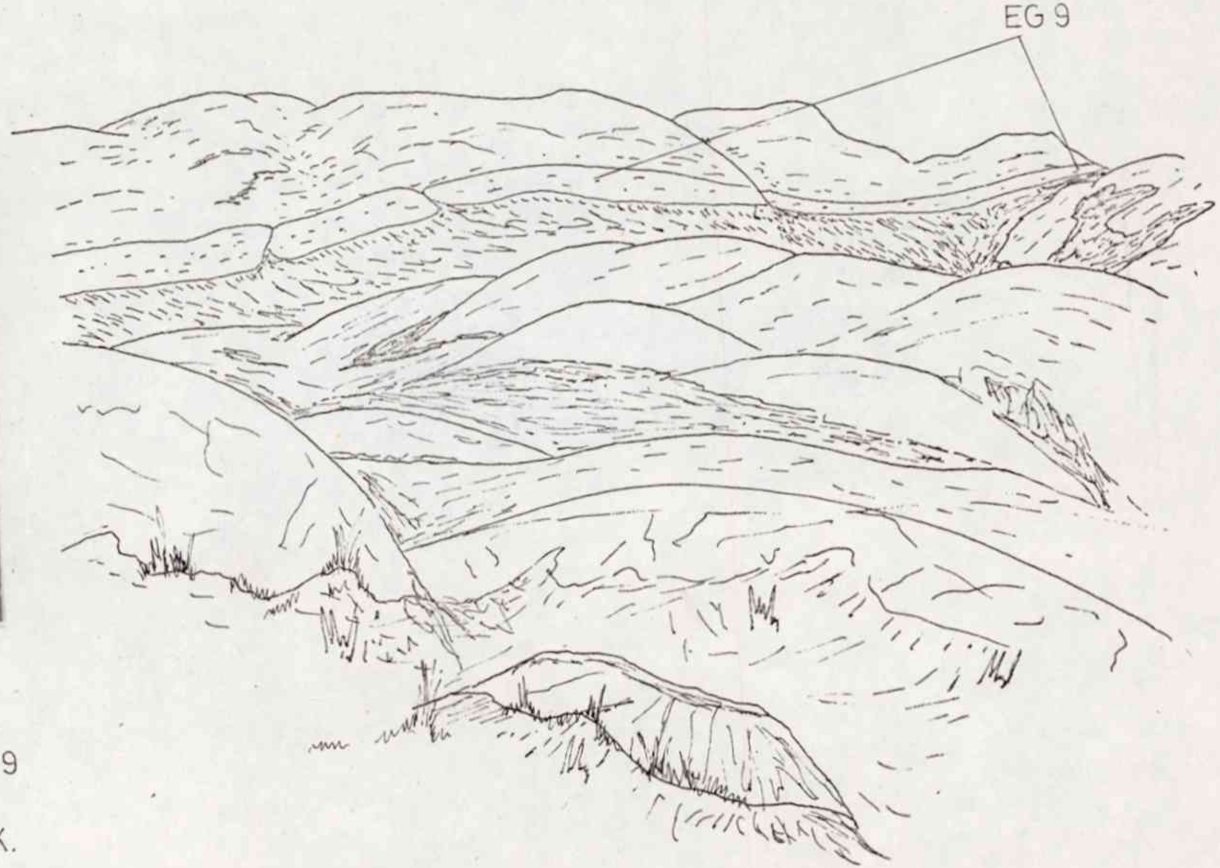
1) Kerotophyres
2) Coarse acid pyroclastic breccia
3) Fine grained felsic tuff.

However, to describe only three is rather an oversimplification of what constitutes one of the major complex problems associated with the region. A number of minor acid lavas and intrusives are present which do not fit into any of the above subdivision. A particularly interesting rock forms a significant physical feature (fig.18). Termed EG 9 in the field, it is a medium grained feldspathic intrusive which has been considerably chemically altered. The rock consists of corroded albite, highly deformed quartz, epidote and chlorite in order of abundance. R.W 17 is a quartz-feldspar porphyritic lava with anomously high proportions of sericite in a fine grained quartz and albite matrix. Geochemical analysis indicates the rock to be of rhyolitic composition.

BASAL CONTACT OF LITHOLOGY EG 9

AN ALTERED FELDSPATHIC ROCK.

Nb. HIGHLY SHISTOSE BASIC LAVA



FIELD SKETCH PANORAMIC VIEW NEAR HAVDALSFJELL I
72500 3700

FIG 1.8

eratophyres.

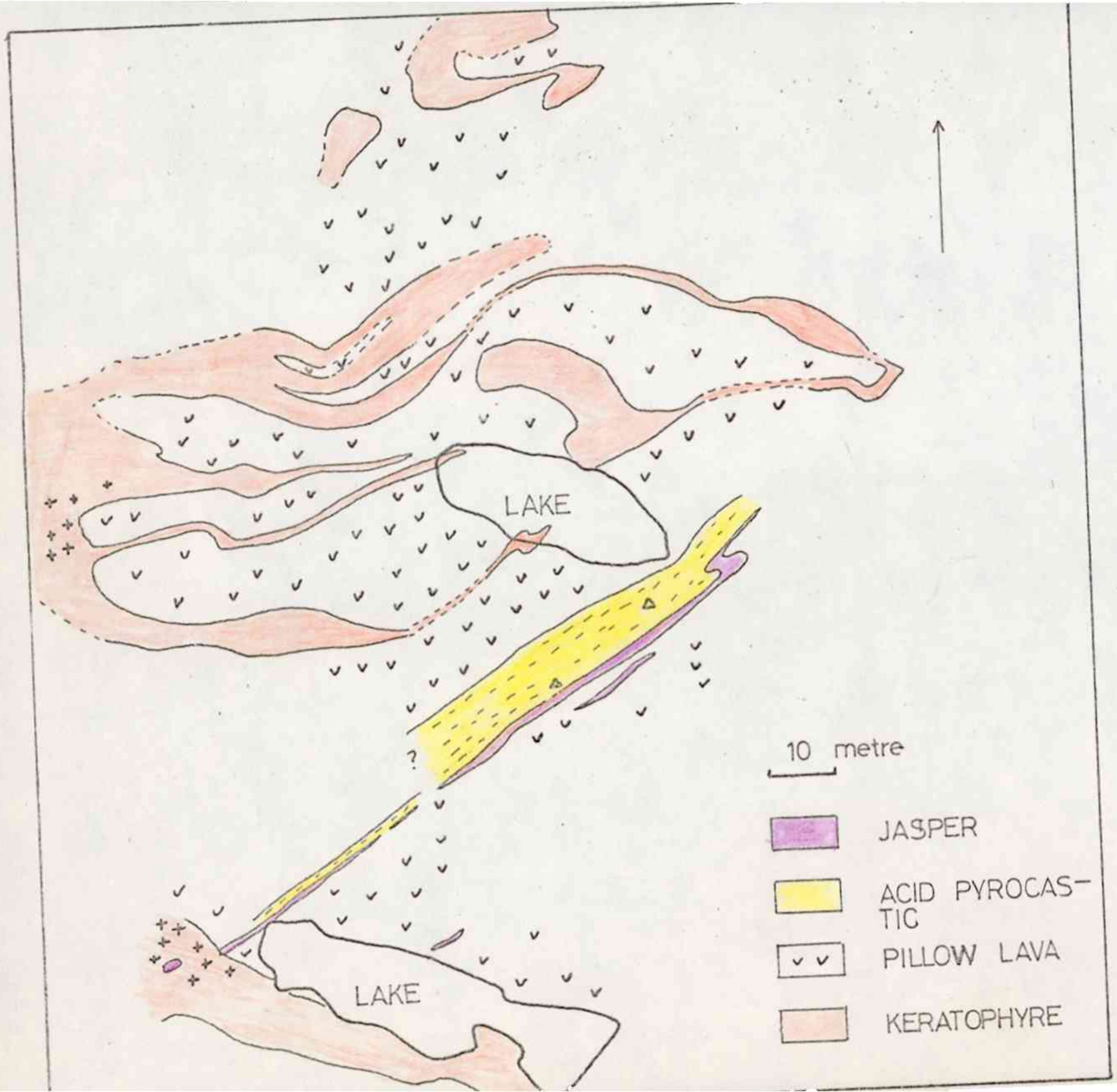
The term keratophyre has been used traditionally in the greenstone elts where sea floor metamorphism has led to the spilitic character of he volcanic sequence. It is used to describe felsic rocks with a ineralogy of sodic or sodicpotassic feldspar, quartz and minor amounts f altered ferromagnesium minerals.

The keratophyres occur as small dykes, sills, leves lenses and possibly lava throughout the metavolcanic sequence. They form only a minor constituent in the north east region, but greatly increase in abundance towards Skorvatn, in the south west.

In the field the keratophyres are a fine grained, white weathering rock with a characteristic flinty fracture. In hand specimens, their colour varies between a pinky white to a grey green. The latter is attributed to an increase in chloritic material in the matrix. Pyrite occurs as isolated grains and small veinlets in the keratophyric bodies and usually shows alteration to limonite. Fracturing is also widespread, usually oblique to the regional schistosity and invariably quartz filled.

The keratophyres show a definite orientation, parallel to the schistosity. They have been tectonically squeezed, dislocated and intricatly folded, though it was virtually impossible to determine any structural information because of the nature of the contacts. The rather irregular shaped appearance of many of the keratophic bodies is a surface expression of their sill like nature and thus not associated with folding. In some cases, these irregular shaped bodies occur adjacent to linear jasper bands. The development of a schistosity varied considerably from the more abundant massive bodies to a poorly developed keratophyric schist. The schistosity is attributed to the breakdown of potash feldspar and micas to form abundant sericite.

The keratophyres are thought to be genetically related to the trondhjemite as they have a similar geochemistry (Scott 1973). Also, the greater abundance of keratophyric bodies in the vicinity of the exposed trondhjemite would suggest a rather close relationship between the two lithlogies. However, no direct contact with the trondhjemite was observed.



DETAILED MAP. COMPASS AND TAPE. SHOWING SURFACE EXPRESSION OF KERATOPHYRE

GEOLOGICAL STRIKE OF REGION INDICATED BY JASPER AND ACID PYROCLASTIC BAND

JASPER XENOLITH WITHIN 'COURSE' KERATOPHYRE. VERY RARE OCCURANCE.

cid (Felsic) Pyroclastic Breccia.

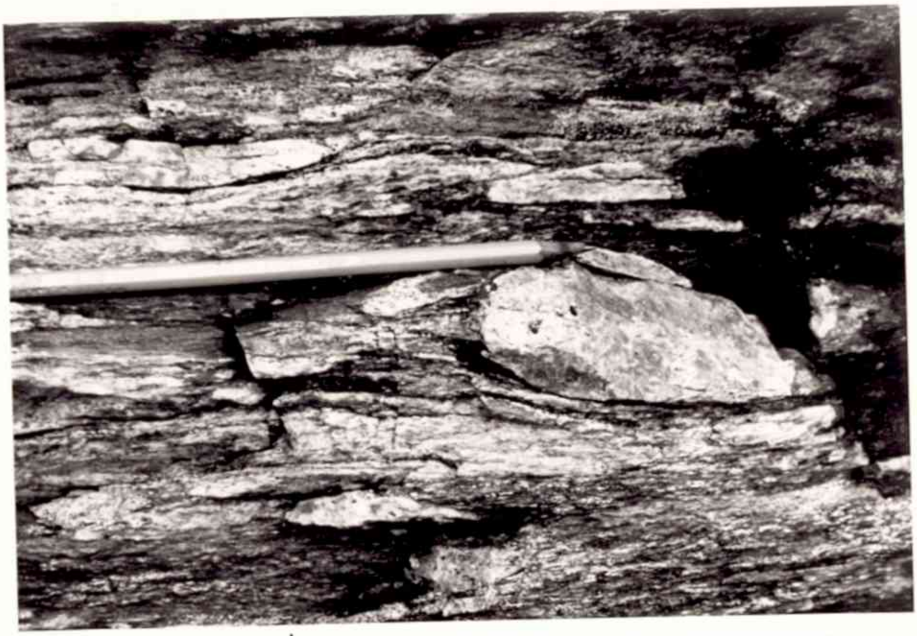
The acid pyroclastic breccia forms a particularly interesting ithology within the metavolcanic sequence, since its presence indicates xplosive acid volcanism, not known previously in the region. tratigraphically the unit underlies the jasper and vasskis bands, lthough this may not always be apparent. The unit is typically eterogeneous, with only the sporadoic occurrance of fragments.

The fragments are of keratophyric mineralogy, essentially albite phenocrysts in a fine grained mozaic matrix of quartz and albite. Although the composition of the fragments remains fairly constant throughout the lithology, their enclosing matrix varies considerably. In the extreme case, the matrix is exemplified by a chloritic schist, with small fragments of feldspar, quartz and secondary carbonate material. At the other end of the scale, the matrix consists of a high proportion of felsic material. In the instances where the fragments are absent along strike, which is the majority of cases, the unit appears, somewhat keratophyric in nature. A criterion used to distinguish the unit from the keratophyres is its stratigraphical relationship with the jasper and vasskis bands. Although this is limited to instances where they occur together, ~~but~~ and is an association which cannot always be demonstrated.

The fragments have been flattened in the plane of schistosity, though the amount of deformation is variable. Strain analysis is possible because the fragments show a major to minor axis ratio which are high and thus not appreciably influenced by the probable subspherical dimension of the original fragments. The ratio's are in the order of 5-10 which is considerably greater than the keratophyric ~~darts~~ clasts occurring in the overlying metaconglomerates.

The variable thickness of the unit, 3 m - 200 m, is believed to relate to deposition and structural control. The acid pyroclastic breccia is more predominant in the south west region, towards Skorovas, which also corresponds to a greater abundance of fragments. Locally the pyroclastic facies is transitional to a coarse schistose quartz porphyry. Clearly seen to the S.E. shores of Blahammoran Lake.

The pyroclastic breccia is clearly a product of a different phase of volcanism than the predominantly basaltic phase. Since the fragments are



Above and below:

Acid pyroclastic breccia within a particularly chloritic rich matrix.

Photo 3.

etrographically identical to keratophyres, regarded by Scott (1973) as hemically similar to trondhjemite, suggests that all three lithologies re genetically related. Primarily relating to an acid intrusive phase ith subsequent acid volcanism.

Considering the close relationship between the pyroclastic horizon nd jasper, it is concluded that the former has formed within a submarine nvironment. The apparent abundance of fragments in addition to the ncreasing thickness of the horizon towards the southwest, is possibly due to its distal relationship with an original volcanic centre. The centre is assumed to be located in the vicinity of Skorovas.

elsic Tuffs.

A third division of the felsic rocks was made in the field, that f the felsic tuff. It forms relatively thin lenticular bands interalated within the pillow lava sequence. They are discontineous, though pecific horizon were traced over 1 km along their strike length. The ands are concordant with the geological strike and schistosity of the egion.

The unit is distinguished from the keratophyres and the acid pyroclastic breccia, by its characteristic brown colour weathering and associated feldspars and quartz grains protruding from the surface. Mineralogically they show similarities with the keratophyres although having a coarser matrix and a higher percentage of sericite. They have a characterically granular and massive texture. However, there is no direct evidence indicating a tuffaceous, and their true mode of origin is not known exactly, but they predate the keratophyric bodies as in a few localities they are cut by keratophysic dykes.

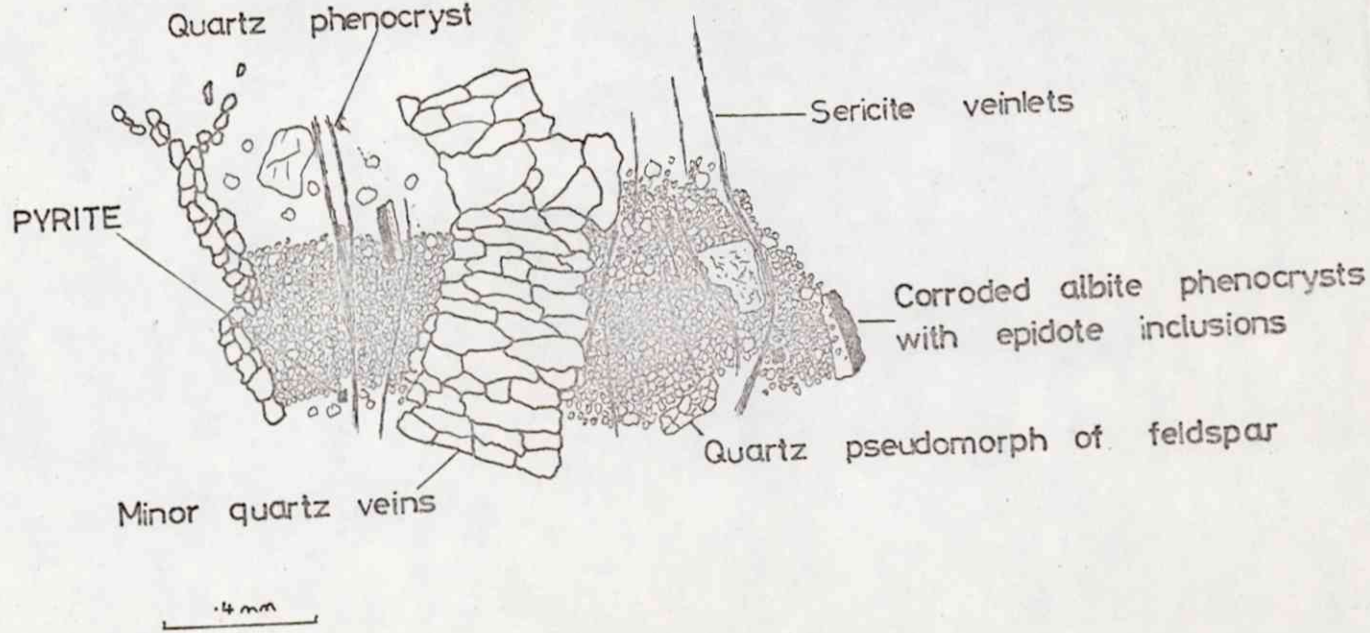
eratophyre Mineralogy

Contrary to the appearance of the keratophyric bodies in the field, his section reveals a remarkable homogeneity between individual pecimens from numerous localities.

The mineralogy is simple, consisting essentially of feldspar phenorysts in a fine grained mosaic of quartz and plagioclase. The composition f the plagioclase was impossible to determine by optical methods. Accessory inerals in the matrix are muscovite, chlorite and pyrite. The latter howing all stages of alteration to leucoxene. The feldspar phenocrysts ave a composition falling in the range of albite. They are invariably corroded and in certain cases pseudomorphed by a secondary quartz mosaic, but more often than not by chlorite. A few of the fresher albite phenocrysts show overgrowth along their grain boundaries. In some sections a synneusis texture, in which aggregates composed of albite crystals, with minor amounts of sphene, cubic pyrite and chlorite are observed.

The keratophyres show evidence of internal deformation, the quartz invariably exhibits strain extinction both in the matrix and secondary veins. The feldspars show twin deformation while in certain thin sections, chlorite has formed in pressure shadows around feldspar and pyrite crystals. In a few samples, the quartz and plagioclase matrix shows a preferred orientation which is usually cut by quartz veins and a network of fine sericite veinlets. The quartz crystals occurring in the veins show elongation in the direction of the preferred orientation.

A section out across an acid pyroclastic breccia fragment shows an increase in chlorite material in addition to a higher proportion of calcite in the matrix. The albite phenocrysts are much larger than seen in the keratophyres but show typical replacement by chlorite and quartz.



SKETCH OF THIN SECTION. KERATOPHYRE

FIG 1.9

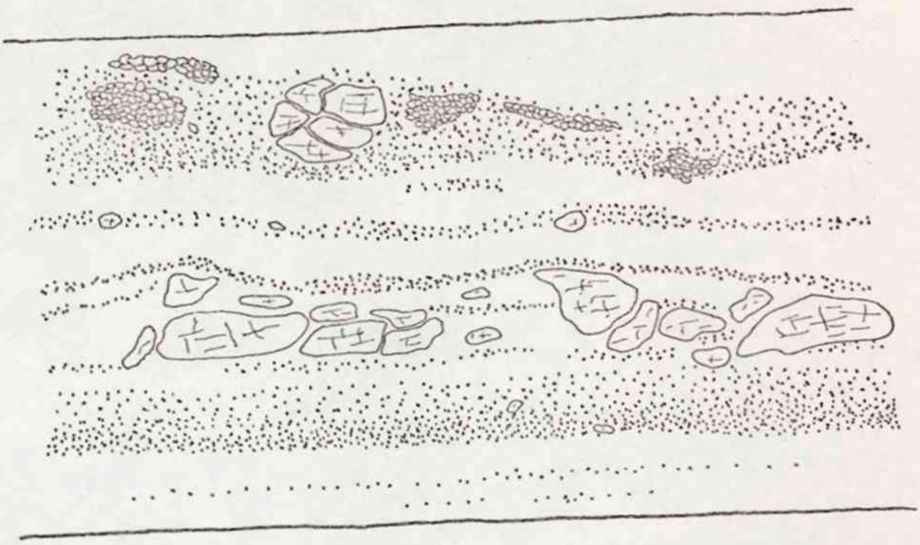
asskis

Stratigraphically the vasskis locally underlies the jasper bands nd overlies the acid pyroclastic breccia. However, the sequence is not lways apparent.

The vasskis has a variable mineral assemblage, consisting of pyrite, agnetite and a variety of gangue minerals. The most common variety of asskis consists of small finely dissemenated pyritic lenses, and euhedral rains of magnetite within a green chloritic schist. (fig.111). Fine banding (fig.111) is a significant feature of some hand specimen, and this suggest a syngenetic origin for the vasskis lithology. Pyrite occurs interbedded with a meta-shale rich in fine magnetite grains, and magnetite occurs pseudobanded with pink jasper and chert. However, the vasskis horizon incorporated massive pyrite, massive magnetite, and barren chloritic schists and cherts in any combination. Facies variation of the vasskis bands is developed both vertical and laterally along strike. Occasionally a vertical sequence is observed from massive pyrite, banded pyrite and magnetite, magnetite and jasper in order of decreasing age.

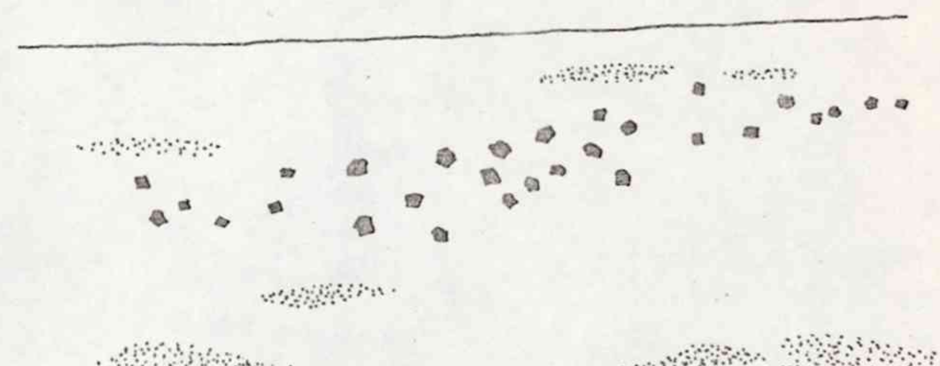
The vasskis is essentially a chemical sediment which was deposited on the volcanic basement under anaerobic conditions. Bacteria playing a significant part in the percipitation of magnetite and pyrite from hydrated iron oxides.

INTERBANDED FINE PYRITE AND SHALE
WITH MASSIVE PYRITE FRACTIONS



1 cm

MAGNETITE GRAINS AND FINELY DISSEMINATED
PYRITIC LENSES IN A CHLORITIC SCHISTOSE
MATRIX



SKETCHES ILLUSTRATING TWO TYPES OF VASSKIS

fig 1.11

Above: Massive magnetite band occurring within a thrusts zone.
Nb. The Greenschist at the top of the magnetite band.
Pillow lavas underneath.

Below: Close up of the magnetite band.

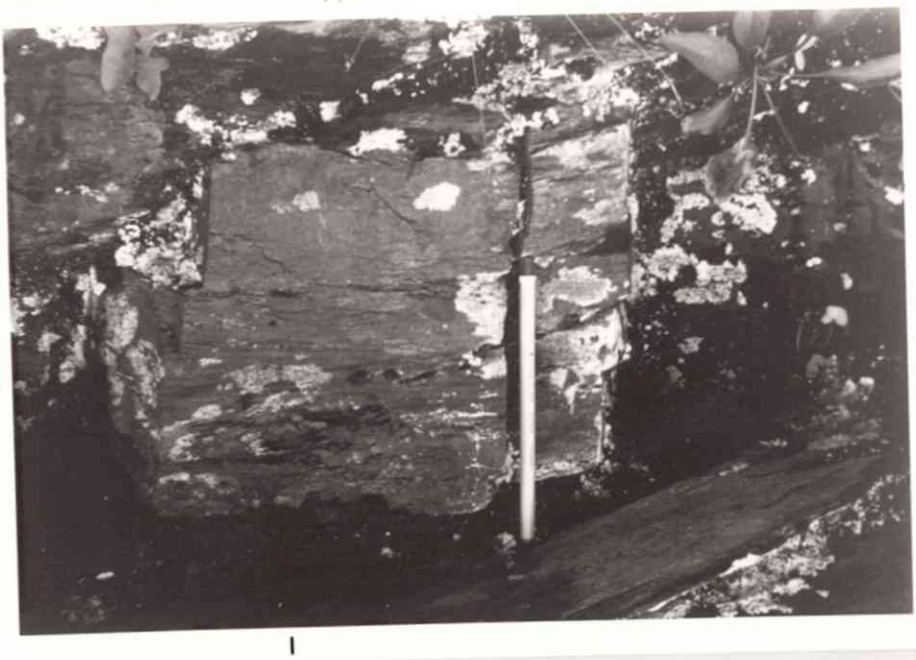


Photo 4.

asper

The jasper occurs as discontinuous bands which occur throughout he pillow lava sequence. The bands invariably follow the geological trike and are concordent with the regional schistosity. They are of aramount importance as marker horizon.

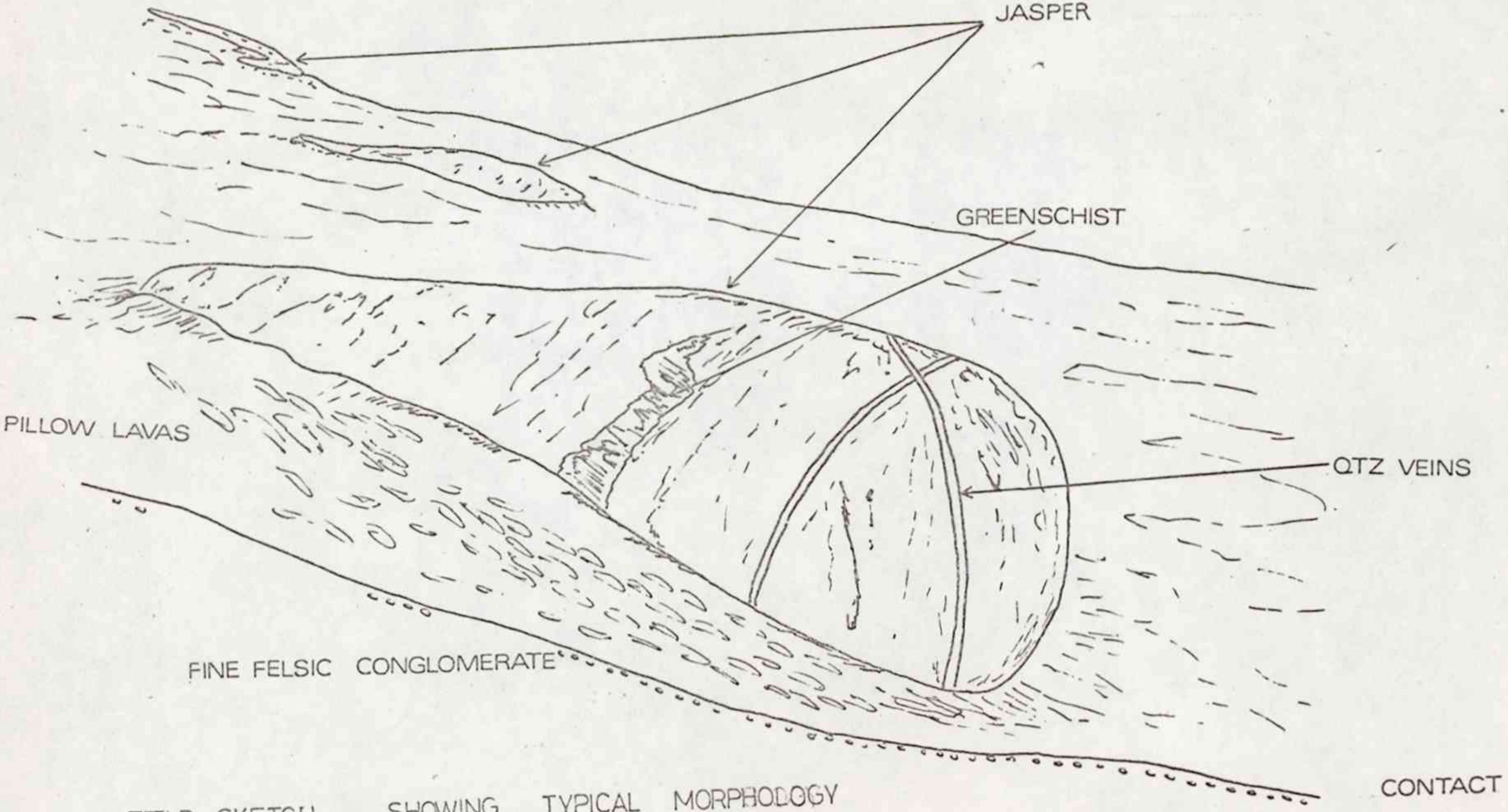
Mineralogically the jasper consists of crytocrystalline quartz and fine grains of platey heamatite and very finely dispersed red oxides. It is a fine oxide fraction which gives the unit its characteristic red colour. In the extreme case, in which the oxides are absent, the colour of the rock is a dark grey and termed a chert. Increasing heamatitie content produces a deep shade of red. In some instances it was possible to observe variations in colour from the base of an exposure to the top. Closely associated with the jasper bands are small lenses of magnetite and in particular quartz veins, which produces a striking colour contrast of white on red.

Stratigraphically the jasper overlies the vasskis which in turn overlies the acid pyroclastic breccia. This relationship is not clear in all sections because locally either the pyroclastic breccia or the vasskis may be absent. Further complication due to the structural complexity of the region, invert the stratigraphical sequence of these lithologies.

The variation in width 10 cm - 5 m, of the jasper bands could result from depositional, erosional or structural control, or more likely a combination of these factors. Structural control is evident since there has been a considerable thickening of fold cores and attenuation of limbs, associated with the folded jasper bands. However, during progressive deformation the jasper bands were at some stage very competent as indicated by bondinage structures (photo 45) and widespread fracturing. The fractures are commonly filled by quartz and minor amounts of pyrite.

The jasper horizons diminish towards Ingulsvatn in the north east region, which corresponds to the only sightings of chert. This is believed to reflect the distance from the volcanic centre which probably lay in the Skorovas area.

The origin of jasper is not fully understood although there is evidence indicating both organic and inorganic origins. Many authors propose the idea of siliceous organism migrating towards active volcanic centres.



FIELD SKETCH. SHOWING TYPICAL MORPHOLOGY AND DISTRIBUTION OF JASPER.

Nb. contact

70200 3850

FIG 1.12

Above: Jasper band showing boudinage structure. Nb. The inflow of chloritic rich vasskis.

Below: Jasper band with characteristic small magnetite lenses.

Photo 5.

wever, this does not appear to happen at the present time. In the
pping region, the close relationship between the jasper and underlying
id pyroclastic breccia suggests a common origin. The jasper probably
iginated from the dispersion of fine coloidal silica and hydrated iron
ides which were derived from the explosive phase of acid volcanism.
ie silica suspension circulating in the sea water while the pyroclastic
orizon was being deposited. The silica apparently fixes the heamatite
nder oxilising conditions, and preserves the oxidised character of the
asper while settling on the sea floor under anaerobic conditions.
educing conditions must be present since the vasskis which locally under-
ies the jasper, contains pyrite and magnetite.

Conclusion

It is apparent from the greenstone discussion that there have been ro distinct phases of volcanism in the Skorovas region.

The first phase and most predominant phase consisted of the extrusion f basaltic lavas on to a sea floor. The basaltic sequence was possibly nterupted by minor acid activity indicated by the presence of fine acidic uffs. Minor basic explosive volcanic activity also interupts the basaltic ava sequence. The lavas could be subdivided into events depending on heir chemical composition. The existence of basaltic andesites (Gjlelsvik 968) and tholeiitic basalts (Scott 1973) has already been proven in the icinity of the orebody.

The first phase was followed by a period of explosive acid volcanism. This period of igneous activity resulted in the intrusion of the trondhjemites and keratophyres and the subsequent extrusion of acid lavas and the deposition of the pyroclastics. Both probably deposited on the flanks of the volcanic centre. The vast quantity of silica and iron generated during this period resulted in the deposition of the jasper and vasskis horizon. The jasper originating from the fixation of hydrated iron oxides by silica in an oxidising environment. The anerobic conditions on the sea floor led to the percipitation of pyrite and magnetite, the two iron mineral phases of the vasskis horizon.

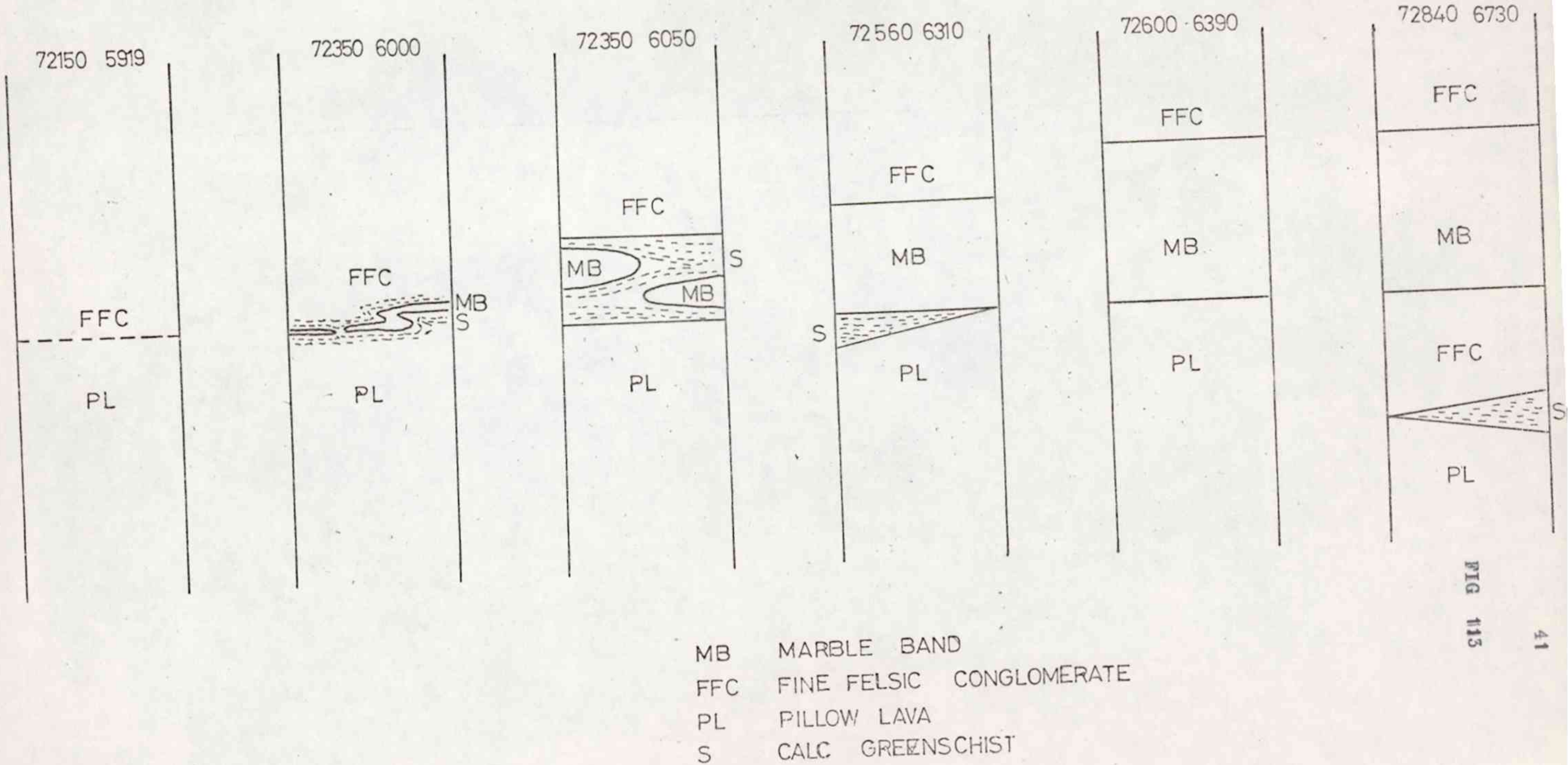
Marble

The marble in an important horizon as it occurs stratigraphically above the metavolcanics and underlies the metaconglomerates. It probably represents the last episode of sedimentation on the volcanic basement before uplift and the resultant rapid erosion of the volcanic pile.

The marble forms a single horizon adjacent or close to the meta-volcanic sequence. In the north east section of the mapping area, the marble forms a 50 metre high cliff section which thins out towards the south west into small lenses. The band follows the geological strike and is concordant with the regional schistosity.

The marble is relatively pure, massive, finely crystalline rock which is locally ferruginous giving rise to its weathered brown appearance. In a number of localities, weathering has produced solution holes. The fresh sample varies in colour from pure white to pinkish brown.

The contact relationships of the marble band with the surrounding lithologies is complex. Involving sliding across the stratigraphical boundaries, dislocation and folding. A summary of the various contacts is illustrated in fig. 1.13 . Ruptured tight F1 folds and boudinage structures are characteristic of the marble bands in the south west regions. Boundinage structures are confined to regions in which the marble is enveloped within a carbonate rich chloritic schist.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72840 6730
FFC
MB
FFC
S
PL
MB MARBLE BAND
FFC FINE FELSIC CONGLOMERATE
PL PILLOW LAVA
S CALC GREENSCHIST

FIG 1.15

he Metaconglomerates.

The metaconglomerates and related volcanogenic sediments are a ontinuation of the major greenstone belt of Norway. Although the ntire correlation of this group is not apparent within other regions of Norway, there is considerable similarities on a finer scale. This is best illustrated by what the author terms as a coarse polymict conglomerate which is petrographically similar to the Fjeldheim conglomerates and the basal conglomerates of the Lower Holvin series (Chadwick 1964).

On the Trones sheet the metaconglomerates are subdivided into the trondhjemite and calc conglomerate. The author however, recognises three distinct units:-

1) Fine Felsic Conglomerate
2) Coarse Polymict Conglomerate
3) Fine Calc Conglomerate

The metaconglomerates are overlain by a thick sequence of finer volcanogenic sediments which extend into Sweden. The sediments range from phyllites which directly overlie the fine calc conglomerate to arkoses.

ine felsic Conglomerate.

The fine felsic conglomerate is composed predominantly of ellipsoidal elsic clasts, 3 cm in length. The clasts range from a pale white olour to a dark grey and are keratophyric in composition with only a inor proportion of phaneritic types. Others include varying amounts f jasper, epidote and marble clasts in that order of abundance. The arble clasts are rare in this horizon. The clasts have a preferred rientation which reflects the tectonic flattening in the plane of schistosity.

The matrix is greyish green in colour, consisting of quartz, albite, and minor proportions of chlorite and epidote minerals. In the rare instances (photo 6) in which chlorite constitutes a significant component of the matrix, a schistosity is invariably present. However, the matrix is usually characterised by a massive grit like texture. Calcite forms a noticeable high proportion of the matrix near the contact with the marble band.

Stratigraphically, the unit directly overlies the metavolcanic sequence in the north east region, towards Ingulsvatn, where it also occurs as a tectonically upthrust wedge in the coarse polymict conglomerate. The unit pinches out towards the southwest and is entirely absent to the north of Haudalsvatn. In this region, the fine felsic conglomerate overlies the marble bound. The contact relationships of the unit with the metavolcanics, marble and coarse polymict conglomerate can be attributed to a combination of phenomena associated with, lateral facies variation in the conglomeratic units and the discordances introduced by dislocative movements along the stratigraphic horizons. A true assessment of the situation is hindered by the lack of quantitive structural evidence and vegetation masking the contacts. The leading unsolved question refers to the occurrance of the marble band within the fine felsic conglomerate. This aspect has already been mentioned in the discussion of the marble horizon.

In all probability, the fine felsic conglomerate represents the basal conglomerate, originally overlying the marble horizon and underlying the coarse polymict conglomerate. The apparent thinning of the unit towards the south west could have resulted from, non-deposition, erosion or thrusting underneath the coarse polymict conglomerate.

Above: Strongly deformed fine felsic conglomerate, with an unusually high proportion of chlorite in the matrix.

Below: Fine felsic conglomerate rich in greenstone and jasper clasts.

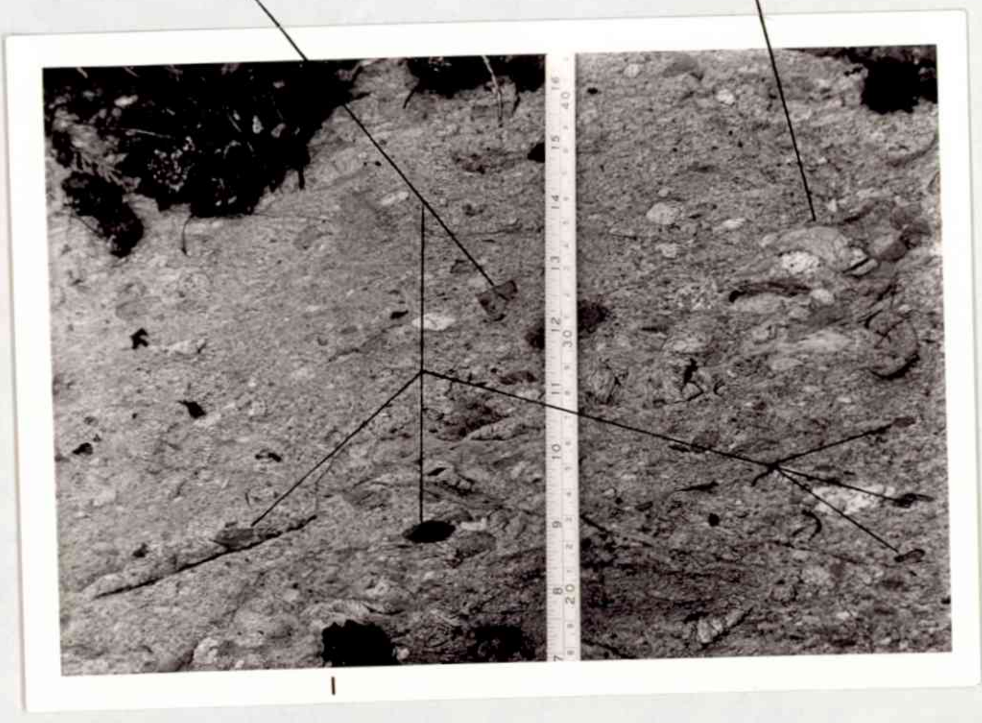


Photo 6.

The remarkable feature of the unit is the abundance of felsic lasts as opposed to jasper, marble or greenstone types. It appears hat the unit was derived from the erosion of felsic material and more pecifically, the felsic pyroclastics, lavas and keratophyres. Thus, t is reasonable to suggest that the volcanic sequence was capped by elsic material, prior to tectonic uplift. The near absence of jasper, particularly resistant rock and marble clasts, which are both significant omponents of the coarse polymict conglomerate indicate only minor tectonic plift. Apparently not effecting the relatively deeper region of the riginal submarine environment in which the jasper and 'limestone' were eposited. However, the fine felsic conglomerate may have been tectonically emplaced into the region from outside and thus not related to the original environment in which the other rocks were deposited.

The Coarse Polymict Conglomerate.

As the name implies, the coarse polymict conglomerate is composed of a variety of clast types. The most predominating are trondhjemite, marble, greenstone, jasper and locally metagabbro. Though, the horizon is called the coarse polymict conglomerate there is however, a noticeable lateral and vertical variation in facies. In close proximity of the meta-volcanic sequence, the horizon is dominated by a clast assemblage comprising of jasper, marble and greenstone types. Further away from the metavolcanic unconformity and also along strike of the unit towards Ingulsvatn, there is an increase in trondhjemite clasts to the extent that it becomes the predominant clast type. An arbitary trace could be followed in the field defining the limit of jasper clast occurrance. However, this was not possible for the greenstone or marble clasts which tended to persist throughout the horizon, though decreasing in size up the sequence.

The horizon is typically unsorted on a small scale, with clast sizes ranging between 1m to 5 cm in a single exposure (fig. 7). photo Regionally the clasts decrease in size away from the metavolcanic unconformity and therefore, up the sequence. The clasts are ellipsoidal in shape and flattened in the plane of schistosity. The extent of deformation is partly dependent on the clast material and to some extent the surrounding matrix. The jasper and trondhjemite clasts tended to be less flattened and more fractured than either the marble or greenstone types. The trondhjemite clasts associated with a silica rich matrix were less deformed than those occurring in a predominantly chloritic matrix. In the latter case, the greenstones and epidote clasts appear to have deformed homogenously with the matrix.

The matrix essentially consisted of four components, quartz, albite, chlorite and calcite, varying in grainsize and relative abundance. The quartz, albite and chlorite mineral components persisted throughout the horizon while calcite is confined to regions containing a high proportion of marble clasts. The amount of chloritic material controls the extent of imposing schistosity. In all cases, the schistosity is deflected around the more competent clasts. Where the matrix is composed of mostly quartz and albite, no schistosity is developed. Regionally the matrix becomes more acidic away from the metavolcanic unconformity, which corresponds to the increase in trondhjemite clasts.

ccessory minerals in the matrix are notably cubic grains of magnetite nd muscovite.

Intercalcated with the coarse polymict conglomerate are lithologies f both igneous and sedimentary origin. Four distinct types are ecognised:

1) The most persistant lithology forms a series of parallel liscontineous bands, never exceeding 3 metres in thickness. The unit is characterised by a gritty texture, which reflects the abundance of clastic quartz and albite grains over the flakey mineral constituents. These minerals are chlorite and sericite. In a few localities graded bedding was observed which confirms its sedimentary nature.

2) A very interesting lithology, which is described on the Trones sheet as a trondhjemite schist, strikes parallel to the tectonic grain of the area for 3 km. Texturally, the rock is medium to coarse grained with a poorly developed schistose fabric. Mineralogically it consists of corroded albite, deformed quartz, chlorite and sericite, showing no recogniseable igneous or sedimentary relationships. However, field evidence indicates the horizon to be of igneous origin, either a dyke or sill, which appears to have been intruded along a thrust plane. The intrusion appears to cut across a tectonically upthrusted wedge of metavolcanics within the coarse polymict conglomerate.

3) A number of chloritic carbonate schist bands occur within the coarse polymict conglomerate. The horizon is dominated by the presence of carbonate rods, flattened in the plane of schistosity. The horizon is exclusively associated with regions in which the coarse polymict conglomerate is characterised by a high proportion of marble clasts. The unit is of particular interest as occasionally occurring at the contact with the marble horizon.

4) Occasionally thin bands of the fine calc conglomerate occur in the coarse polymict conglomerate. This unit is discussed in detail in the next section.

The size of the clasts suggests rapid erosion of the volcanic sequence during a major tectonic uplift stage. The extent of uplift is illustrated by the presence of marble clasts within the coarse polymict conglomerate, As the original limestone horizon was deposited in the relatively deeper regions of the submarine environment. The abundance of trondhjemite clasts

Above and below:

Unsorted, strongly deformed coarse polymict conglomerate. Nb. The fracture pattern in the trondhjemite clasts (light coloured) in relationship to their major axes. Also the greater deformed greenstone clasts (pale grey colour).

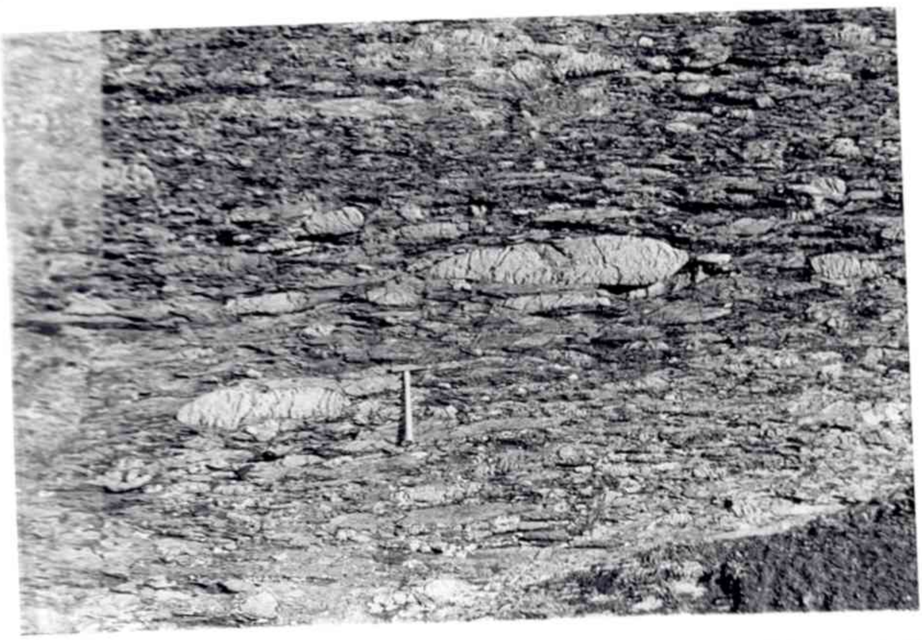


Photo 7.

ithin the unit may be a reflection of its high level intrusion into he volcanic sequence.

The trondhjemite schist within the coarse polymict conglomerate nust indicate local minor igneous activity during the initial stages of uplift.

ine Calc-Conglomerate.

The coarse polymict conglomerate is transitional to the fine calc onglomerate. The grading occurs with a specific pattern. Firstly, by n increase in carbonate material in the matrix and a subsequent decrease n both quartz and albite grains. Secondly, an increase in marble clasts t the expense of trondhjemite clasts.

The marble clasts are ellipsoidal in shape and between 2-5 cm in length. They have been flattened in the plane of schistosity. The whole unit weathers with a characteristic brown colour and is dominated by a pitted surface. The clast to matrix ratio is high and is more consistent throughout the unit, than seen previously in the other two types of conglomerates. The unit is typically unsorted, though the clast size is confined to relatively narrow limits. However, there are locally developed thin, < 5 cm, chloritic schist bands dispersed throughout the unit. The matrix has a schistose texture which deflects around the marble clasts. Chlorite and calcite are the major components of the matrix with only subordinant amounts of epidote, quartz and albite.

The width of the fine calc conglomerate is approximately 400 metres on the northern shore of Havdalsvatn, but pitches out in the north west region towards Ingulsvatn. Intercalated with the unit are thin phyllite bands which also separate the unit from the massive arkose. The phyllite is a calcareous-muscovite rich rock weathering with a sandy brown colour. Pyrite is present as small 2 mm cubes and in many cases altered to limonite. The phyllite horizon which occurs at the contact with the overlying arkose is characterised by a number of minor folds. Both F1 and F2 dislocation folds are recogniseable. These features are absent in the bands occurring within the fine calc conglomerate. F2 dislocation folds indicate tectonic movement along the arkose - phyllite contact. This is clearly supported by the topographical feature occurring at the contact. The arkose forms a 200 metre cliff section with the phyllite outcropping at the base.

The vertical facies changes, from the coarse polymict conglomerate, reflects the density contrast between the marble and the trondhjemite clasts. The former having a lower density and thus, able to be transported further under the prevailing condition. The direction of facies change suggests that material was transported in a south easterly direction. NB onclusion

Certain geometrical and morphological trends within the conglomerate equence correlate with the present day position of the eroded derivatives. oth the fine calc conglomerate and fine felsic conglomerate have an nverse thickness relationship to the marble band and felsic volcanics espectively. The maximum thickness of the fine felsic conglomerates owards the NE of the region coinciding with diminishing occurrance of 'elsic material in the metavolcanics. The thinning of the fine calc :onglomerate towards the NE coincides with the maximum thickness of the narble band.

The jasper bands which are abundant to the SW of the mapping region diminish rapidly towards the NE. The trend corresponds to a lateral facies change of the coarse polymict conglomerate from the jasper, marble and greenstone clast assemblage to one which is dominated by trondhjemite and greenstone in the NE. Thus, there appears to be a positive correlation between the absences of jasper within the metavolcanic sequence and clasts in the coarse polymict conglomerate.

From the above discussions it is believed that the present day spatial relationship between the metavolcanics and metaconglomerates is a mirror image of their position prior to nappe emplacement. Thus, there has been no substantial lateral tectonic movement between the metavolcanics and metaconglomerates sequences.

Regional Structural Setting

Skorovas is located within the Scandinavian Caledonides which, in Nord Trondelag, consists of a superimposed pile of nappe structures of varying lithology and metamorphic grades. The nappes trend approximately north-east and have a common west to east sence of overturning.

The Havdalvatn/Ingulsvatn area lies within part of the Gersvik nappe, and lying on a south-east margin of a greenstone belt which strikes northeast - southwest and is some 20 km wide and has a strike length of about a 100 km. The Gersvik nappe was defined in the summary published by Oftedahl (1956), on the basis of extensive mapping in the Grong region by Foslie (1922-43). The exact position of the Gersvik nappe is not clearly indicated on the map provided by Oftedahl, but lies approximately in the vicinity of the arkoses. The mapping area incorporates only a small section of the lower part of the Gersvik nappe.

In the area between Lake Turnsjoen and Lake Limmøngen, just to the NW of Ingulsvatn, the greenstones have been pushed over the volcanogenic clastic sediments. In the mapping area the greenstones dip under the sediments. This at least suggests a complex pattern of deformation, which has been superimposed on an intial system of nappes. The complexity of the problem is made apparent by the ambiguity in the structural interpretation, summarised in the general map presented with the overall description by Oftedahl (1965).

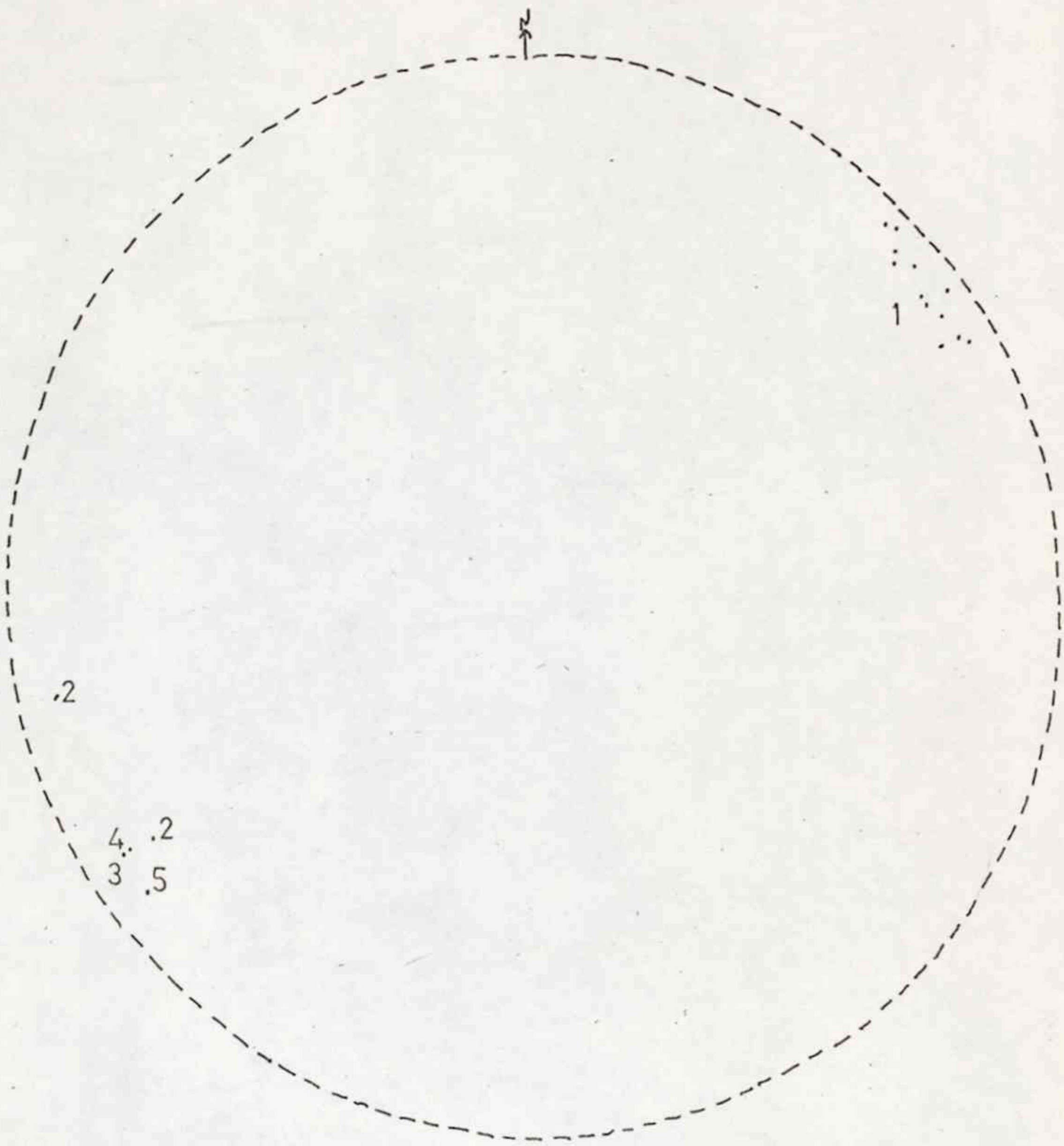
Structure

Two phases of deformation are recognised in the Havdalsvatn and Ingulsvatn region. The earlier phase is designated F1 and the latest F2.

F1 Folding Phase

The most obvious indication of an earlier deformation in the area is the occurrance of a schistosity designated S1. The schistosity is always axial plane to the F1 phase of folding. The F1 fold structures varied in style and wavelength, though generally tight, isodinal folds, with a wavelength between 2-3 metres and a near horizontal axial plane. Reference to fig.114 shows that the axial direction to be NE - SW and this is substantiated by the remarkable consistency in strike of the schistosity. Evidence for the tightness of the folds is provided by the jasper, vasskis and acid pyroclastic horizons which occur locally

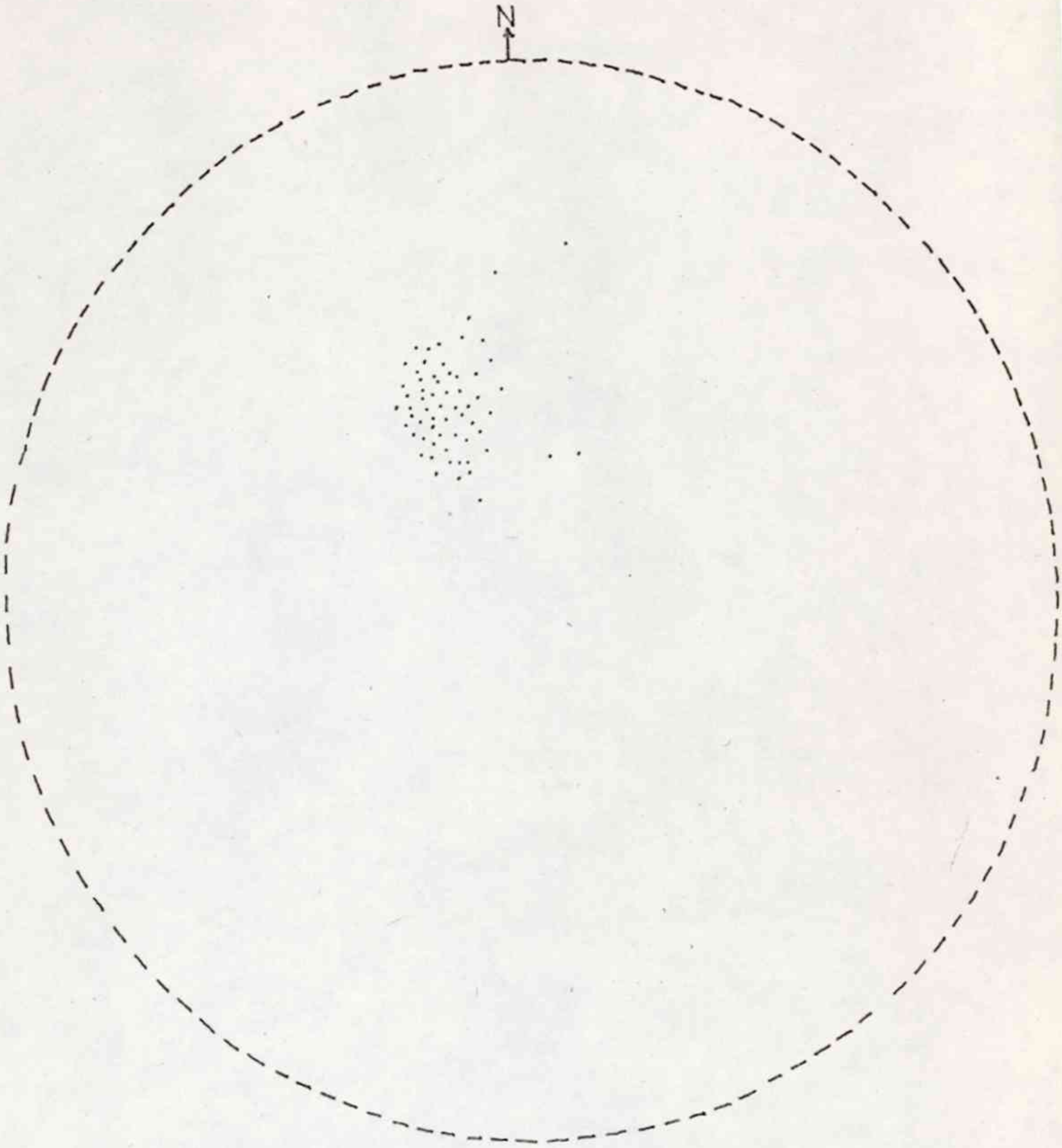
Hvilken projeksjon
Ikke lik Kolling



F1 MINOR FOLD AXIS

1 PHYLLITE BAND
2 GRIT BAND. CONGLOMERATE
3 SCHISTOSE KERATOPHYRE
4 PILLOW LAVA
5 CONGLOMERATE

FIG 1.14



POLES TO S1

FIG 1.15

throughout the metavolcanic sequence, as a series of bands in close proximity to each other. The jasper bands show considerable thickening of hinge zones while their limbs have been attenuated. Boudinage structures occur locally on the limbs which reflects the greater competence of the unit relative to its surrounding greenstone and vasskis matrix (photo 4).

Major F1 folds, those compariable to nappe structures which are responsible for the development of the schistosity are absent. The largest fold recognised in the mapping area, having a wavelength of 10 metres, was indicated by Z and S folds developed in a grit band in the metaconglomerates, on the lower and upper limbs. However, folds of this magnitude are rare and it is believed that the minor F1 folds recognised in the mapping area are parisitic folds lying at the hinge zone and the limbs of major F1 structures.

Field evidence indicates that only F1 folding appears to have produced any significant amount of internal deformation of the rocks. In all cases the deformed particles are flattened in the plane of the S1 schistosity. Potential indicators for determining the amount of strain are the pillow structures, amygdals, vesicles, conglomeratic clasts and acid pyroclastic breccia. Unfortunately one can only assume that the fine felsic conglomerate was originally spherically shaped. The assumption is based on the small size of the clasts.

Analytical data derived from the dimensions of the fine felsic clasts, reveals that the major to minor axis ratio for the jasper fragments is 2:1 and 3:1 for the felsic fragments. Thus, indicating the greater competence of jasper relative to the felsic clasts (fig.1.18). Measurements were compiled from the pyroclastic breccia since their major to minor axis is very high and thus the results are not appreciably effected by the primary subspherical shape. The ratios are considerably higher than for the felsic clasts in the fine felsic conglomerate and are in the range of 5 - 10:1.

The pillow lavas show a complete history of deformation, which was previously illustrated in fig.1.3 . The earliest deformation probably results from superincombent loading during cooling after erruption on the seafloor. During tectonic deformation the pillows are flattened in the plane of schistosity. Rotation of the pillow structures may have resulted, since it is unlikely that the plane of schistosity coincided ith the primary layering. Further tectonic strain produces internal eformation such as elongation of vesicles and amygdales, and the ubparallel alignment of minerals. Eventually the primary fabric is estroyed and a penetrative schistosity is imposed. Here, occurs the ompletion of the alignment of flakey minerals. A particularly intersting phenomena is the complete random orientation of albite laths n the fragments of the basic breccia unit. This suggests that the ectonic strain has been absorbed by the chloritic matrix enveloping he breccia fragments.

The extent of tectonic strain within the polymict conglomerates lepends on the clast type and the enclosing matrix. Where the matrix is particularly rich in chlorite and epidote, the greenstone clasts appear to have deformed homogeneously with the matrix. This producing a well developed linear and planar fabric. The more competent, trondhjemite, jasper and carbonate clasts appear only slightly deformed, with the schistosity of the matrix deflecting around the clasts. However, the extent of deformation varied regionally and partly relates to the matrix composition.

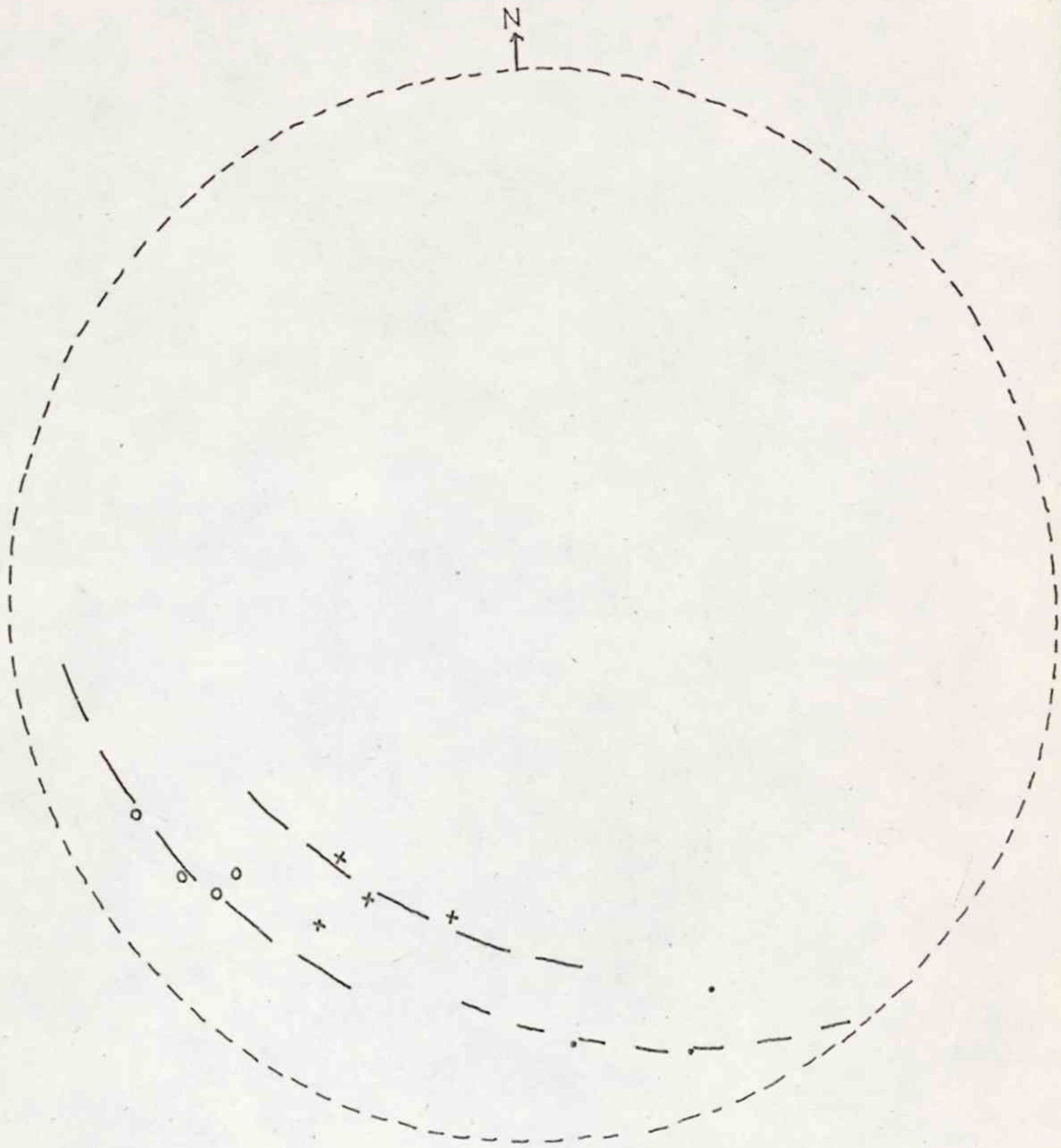
2 Folding Phase

The acculation of strain during progressive deformation appears to ave been confined to a series of narrow thrust horizons which are onformable to the geological strike and regional schistosity of the egion, but have an anastomosing effect. The features are clearly ecognised on the aerial photographs. Dislocation along these horizons as produced post schistosity folds, characterised by a coarse crenulation f the S1 schistosity and are designated F2 folds. The nature of these orizons may reflect the gross differencies in lithology and competence etween rock units in which they act as planes of weakness in which assive competent blocks are allowed to slide. The relative movement of the competent blocks is believed to have been propagated by their reaction to differential compressive stress. The phenomena is probably assisted by the presence of the massive trondhjemite bodies, which disrupt the homogenetity of the greenstone sequence and act as stable competent blocks relative to the host greenstones.

The axial direction of the minor F2 folds is more variable than the F1 (fig.117), though occasionally they are near coaxial. The variations, between SW and SE, is expected, since they are generated by dislocation of distinct horizon which is a local phenomena.

Though the minor F2 folds are developed in a number of lithologies they are particularly characteristic of the schistose units, the phyllites and greenschist horizons. Commonly associated with the greenschist horizons is quartz and calcite, believed to have been derived from metamorphic segregation of silica during tectonism.

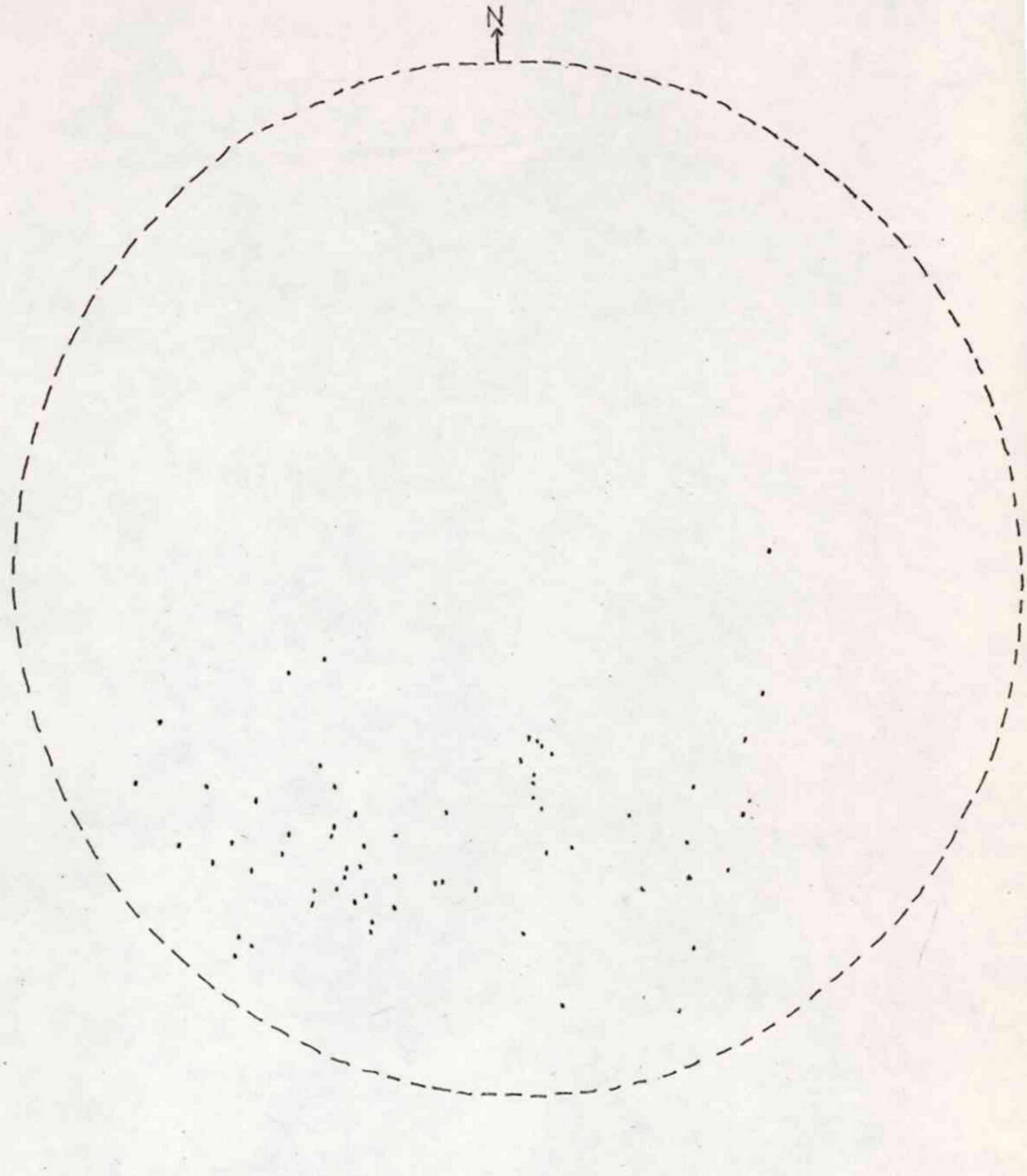
The consistency of the strike of the schistosity in the mapping area suggests that major F2 structures are absent. However, the occasional undulation of the metaconglomerate clasts may indicate major post S1 folding. In comparison to F1, the fold style is open and more or less vertical axial planar. For a fuller discussion on the F2 phase of folding the reader is referred to the report by Scott (1973).



POST SCHISTOSITY F2 FOLDS . GREENSCHIST HORIZON

	o	3850, 70850.
FOLD AXES	+	3950, 72230.
	•	6710, 73795.

FIG 1.16



POST SCHISTOSITY F2 FOLDS

VARIOUS LITHOLOGIES

• FOLD AXES

FIG 1.17

FIG 1.18

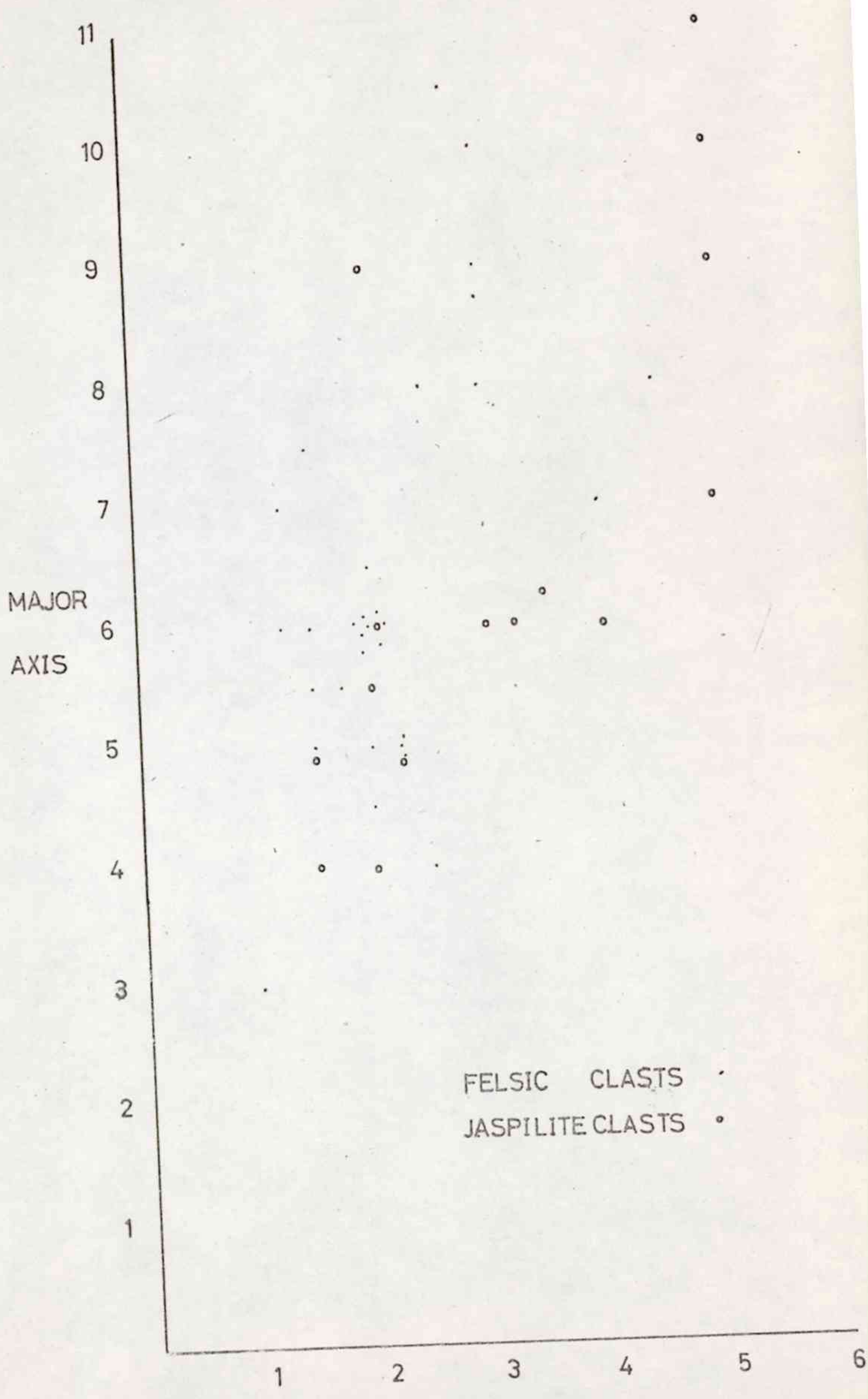
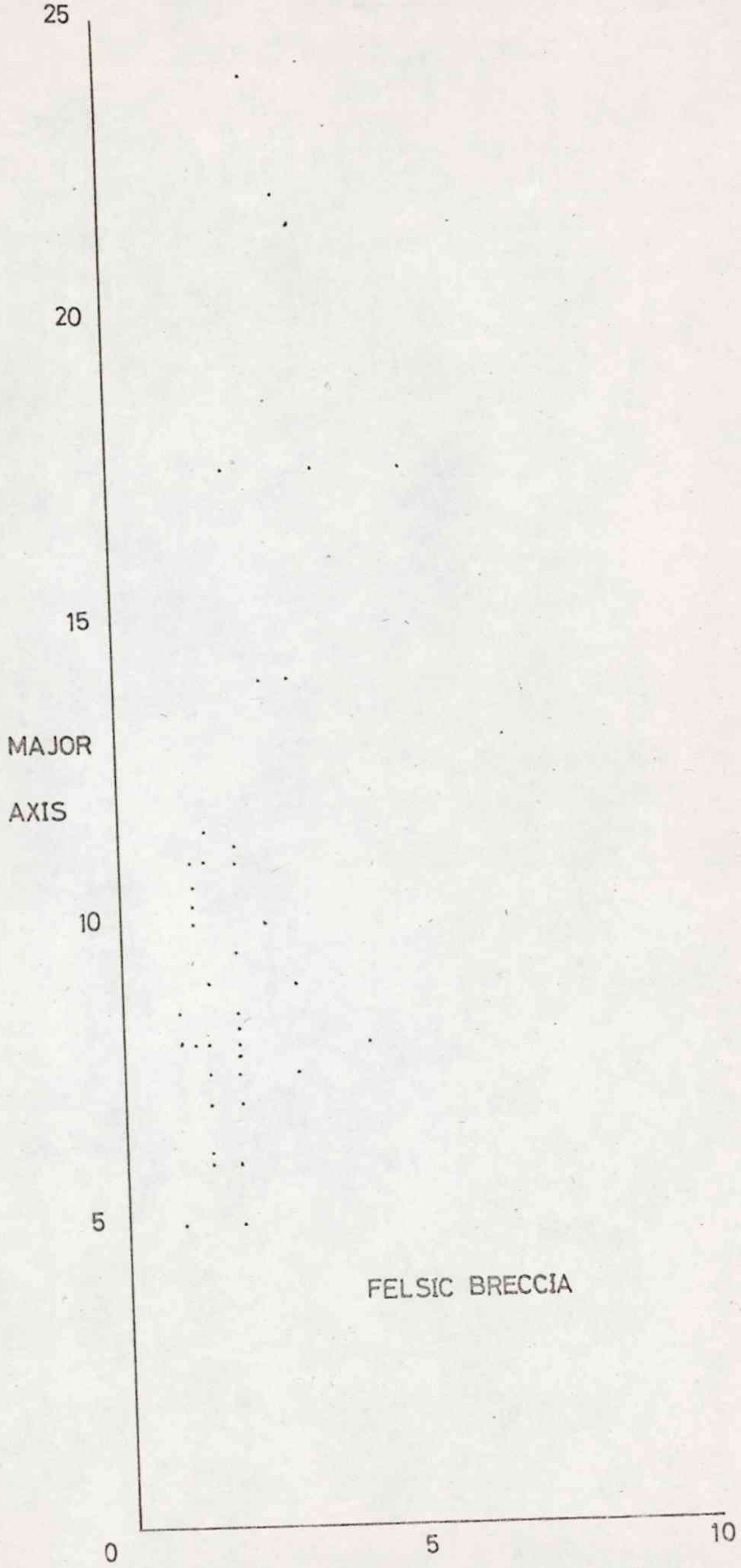

11
10
9
8
7
MAJOR AXIS
6
5
4
3
2
1
FELSIC CLASTS
JASPILITE CLASTS
1
2
3
4
5
6
MINOR AXIS

FIG 1.19



hrusts

Thrusts are a major feature of the mapping area. The thrusts run arallel to the tectonic grain and lithological strike of the region. hey are clearly recognised on aerial photographs since they form rominent scarp features which are of considerable strike length, often xtending right across the mapping area. The rocks adjacent to the more rominent thrusts show little brecciation and only where displacements of ypes or crenulation of the Si shistosity are recognised, can the thrust e clearly defined.

The majority of thrusts appear to dip south eastwards and are usually confined to schistose horizons or stratigraphical boundaries where dislocation of the units is a characteristic feature. Dislocation is thought to be the major factor contributing to the complex structural relationship at the contact between the metavolcanics and metaconglomerates.

Near the shores of Ingulsvatn, to the NE of the mapping area, pillows, acid volcanics and marble appears to have been thrusted as a wedge into the polymict conglomerate. The wedge is about 0.5 km from the metavolcanic/ metaconglomerate contact. However, the thrust contacts were not clearly defined.

The genetic aspects of thrusting have already been discussed in the F2 section.

PART II

GEOCHEMICAL ASPECTS OF THE

GREENSTONE ROCKS

ojectives of the Study.

The aim of this study is to determine the possible palaeo-tectonic nvironment in which the Skorovas region originated. To achieve this bjective, end samples of basic and acidic rocks were collected in the ield. Supported by samples obtained by previous Imperial College field parties, these ere analysed to determine their geochemistry.

The relevant literature to which much importance is attached in the nd discussion falls into two distinct groups:-

1) Major and minor element geochemical methods to discriminate the palaeo-tectonic setting, in which volcanic sequences, now obducted onto continental masses, may have originated.

2) The characterisics of ~~marine~~ massive sulphide ore deposits and their host rocks, associated with specific tectonic environments.

From the published data together with results of the geochemical analysis, it is hoped that a useful discussion will materialise concerning the original tectonic environment of the Skorovas region.

ochemical Methods.

In recent years there has been considerable interest in the use of ·ochemical methods to discriminate the palaeo-environments in which suites ' volcanic rocks, now tectonically incorporated in orogenic belts, may ive originated. When a statistically adequate sample of volcanic rocks rom a range of known enviroments have been thoroughly studied, in terms f specific major, minor and trace element contents an empirical system f ratios can be established. The ratios are only useful if the element oncentration is consistent within a single volcanic type and shows considerable ariation between volcanic types. The technique used is then to compare he analogous ratios and composition ranges in the ancient volcanic suites ith those of modern control and thereby determine the affinity of the ocks.

The way in which the geochemical studies are applied can be conveniently considered in terms of:-

1) Major element content.

2) Minor element content.

In both cases the chemical mobility of the various component elements must be considered in terms of the various factors in the post-consolidation history of the rocks.

Major element geochemistry of submarine volcanic rocks has various limitations resulting from interaction with circulating seawater during the deuteric stages of their thermal history. Also, because of bulk chemical changes operating during burial metamorphism, and subsequent deformation and regional metamorphism during nappe emplacement.

Metamorphism commonly causes major element redistribution within a volcanic sequence and to some extent, into and out of that system. Smith (1968) indicates that there is a tendency for Ca, Mg, Fe and Na to migrate over distances up to a metre, during low grade metamorphism. This phenomena is clearly seen in the mapping area with the presence of epidote knots and carbonate rods. Investigations of certain alkalie elements in oceanic basaltic material in situ indicate Rb and K enrichment during ocean floor weathering, but subsequent depletion during greenschist metamorphism (Cann 1970). The depletion of these two elements is attributed to a lack of suitable atomic sites in the crystal lattices of various minerals stable in this facies.

hemical interaction with circulating saline fluids affects a number of he major and minor element concentrations in submarine lavas. The illow margins show a progressive enrichment of K and related minor lements relative to their interior (Hart 1969). Calcium is less ffected by secondary processes, although during greenschist metamorphism here is often a depletion of available Ca sites which effectively decreases he Sr concentration. However, Na is strongly affected by both metamorphism nd saline waters, in which there is a substantial increase in the Na content of the lavas.

From the above considerations it is apparent that a certain amount of scepticism is required when using major element concentration to distinguish volcanic types. This has encouraged work in which attention is focused on minor elements in which some are stable during post-volcanic processes.

Pearce and Cann (1971, 1973) have compiled data demonstrating that the trace elements Y - Zr - Ti and Nb, apparently preserve their original abundance in ancient volcanic rocks. Thus, valid comparisons can be drawn between ancient and modern volcanic rocks using there element as a basis for study.

Pearce and Cann (1973) distinguish:

1) Basalts errupted within plates, (Ocean Island and Continental Basalts) using a Ti - Zr - Y diagram.

2) a) Altered plate margin basalts, (Ocean Floor Tholeiites, Low Potassium Tholeiites and Calc Alkaline Basalts) can be identified using a Ti - Zr diagram.

   b) Unaltered plate margin basalts can be identified by using a Ti - Zr - Sr diagram.

3) Y - Nb can be used for indicating the alkalic nature of a basalt.

Jakes and Gill (1970) have demonstrated that the rare earth elements show a significant variation between valcanic types, and are apparently stable with respect to low grade metamorphism.

The distribution of the rare earth elements in volcanic rocks, compared to the pattern found in chrondrite show :-

1) A flat distribution for oceanic tholeiites and island arc tholeiites, though the latter have a greater variation in the total range.

2) A pronounced enrichment of the lighter rare earth and a depletion of the heavier rare earth for the calc-alkaline volcanics.

Some of these trace and minor element criteria are used in this study to provide a classification of the basaltic greenstones and to suggest a possible environment for their origin. The major element distribution may prove useful in distinquishing between basalts and basaltic andesites and also providing information concerning bulk chemical modification during post volcanic processes.

lcanism and associated massive sulphide deposits in relation to
ate tectonics.

The great volume of geophysical and geological data accumulated uring the last two decades from the oceanic and continental environments, as led to the formulation of the theory of plate tectonics and its cceptance, as a working hypothesis in practice. The following section ummarises the main aspects, laying particular emphasis on the genesis f massive sulphide ore deposits with relationship to specific tectonic nvironments.

Modern day theories of plate tectonics require that oceanic crust e generated along mid oceanic ridges. The tholeiitic lavas, derived rom the partial melting of mantle periodotites moves laterally away rom the crust. This probably involves the influence of convection urrents within the underlying upper mantle. The oceanic crust has a stratified appearance, consisting of a lower layer of gabbroic intrusives, passing upwards into a dolerite sheet complex. These are overlain by pillow lavas, hyaloclastites and locally, thin pelagic sediments. During lateral movement down the ridge flanks intrusions of alkalie basalts are a common occurrance, but they are subordinate to the tholeiitic lavas in volume.

The massive sulphide deposits occurring in this type of tectonic environment are characterised by a simple mineralogy. This essentially consists of pyrite, chalcopyrite and minor amounts of sphalerite. The host volcanic rocks are mafic to ultramafic in composition and invariably spilitised. Felsic volcanic phases are rare or absent and there is also a lack of pyroclastic material. There are however, aquagene tuffs, hyloclastites and pillow breccias intercalated with the basaltic flows (Smitheringale 1972). Sediments only constitute a small proportion of the volcanic pile and are predominantly chemical deposits such as cherts, jasper, umber and ochres (Bear 1963, Constantinou 1972).

Examples of these ore deposits span the whole Phanerozoic, appearing first in the early Palaeozoic ophoilites, right up to present day metallic gels found on the floor of the Red Sea (Deyens and Ross 1969). The earlier Palaeozoic deposits include the Notre Dame Bay area (Smitheringale 1972) and Bay of Island (Duke 1971) in north central and western Newfoundland.

e best known example is provided by the massive orebodies of Cyprus which e of Mesozoic age (Moores and Vine 1971, Gass 1968, Constantinou 1972). tchell and Bell(1973) describe these deposits associated with ophiolite ites as 'Cyprus type Deposits'. Tertiary deposits of this affinity occur Ergani Maden in Turkey (Griffiths et al 1972).

Island arcs are found at destructive plate margins within oceanic isins or separated by marginal basins from the continental margins. On ne convex side of the arc is situated the subduction zone while on the oncave side or continental side occurs the marginal basin in which oceanic rust may or may not be being generated.

The lavas associated with the initial stages of island arc volcanism re invariably island arc tholeiites (Jakes and Gill 1970). These are ollowed at a later stage of island arc evolution, by the calc alkaline eries which range from basalts to rhyolites. The island arc tholeiites, hough chemically similar to the oceanic tholeiites can be distinquished y the concentration of trace elementsthey contain (Carr and Pearce 1971).

The subduction of oceanic crust at the Benioff Zone conduces lateral variation of magma types across a mature island arc (Kuno 1966). This explains the presence of the calc alkaline suite as a late stage phase of volcanism. The phenomena has been attributed to melting of the mantle wedge and selective partial melting of oceanic lithosphere and pelagic sediments, in or above the Benioff Zone. The exact mechanism is not fully understood however (Kuno 1960, Ringwood 1966). Kuno (1966) has used this phenomena to explain the gradual continuous change from tholeiitic lavas on the ocean side, through high alumina basalts and calc-alkaline rocks to olivine basalts on the continental side of Japan.

Mineral deposits formed at the early stages of island arc evolution are characterised by a simple, but more variable mineralogy than the 'Cyprus type Deposits'. Generally the massive pyrite deposits are rich in either one or both the metals zinc and copper, with only minor amounts of lead. The host rocks are variable, but are primarily island arc tholeiites with minor volumes of andesites. In some cases the whole suite from basalts to rhyolites are recognised (Hutchinson 1973). The thickness of the volcanic pile is large, in the order of 40,000 ft (Goodwin 1970). Sediments associated with the mineral deposits are mainly chemical cherts and iron formations, siliceous tuffs, volcanogenic grewackes and volcanoclastics (Hutchinson 1973).

ie sediments indicating intermittent volcanism, possible the final ;ages of a volcanic cycle. Examples of these deposits are Besshi in ıpan, Menstrask in Sweden (Grip 1951), and the Keewatin greenstone belt ıcorporating Noranda, Timmins and Matagami orebodies (Hutchinson 1973).

Further island arc development is signified by the predominance of he calc alkaline suite of lavas and associated acidic intrusive phases. 'olcanism is typically explosive with the generation of felsic lavas from .pparently domical centres. Endogenous rhyolite domes contaning both 'elsic lavas and the porphyritic subvolcanic equivalent appears to have ›layed a significant part of volcanism (Hutchinson 1973).

Massive pyrite deposits associated with the later stages of island ırc evolution are characteristically polymetallic, rich in both lead and zinc with only minor amounts of chalcopyrite. The gangue minerals vary considerably from carbonate rich, occurring in older deposits, to bedded gypsum - anhydrite and baryte in the younger deposits. The host volcanics are felsic rocks, with basalts at depth into the volcanic pile, if they are present. Intermediate and felsic rocks predominate adacent to the orebody, together with breccias and pyroclastic rocks. Sediments form an important constituent within the volcanic pile. Epiclastic sediments predominate over the chemical deposited cherts and ironstones and the volcanoclastic rocks.

These deposits first appear in Proterozoic rocks such as the Errington and Ver Lake deposits of Ontario (Martin 1957). The large Palaeozoic deposits of the Bathurst District in northern New Brunswick (McAllister 1960) and at Buchans Newfoundland (Swanson and Brown 1962). The economically important Tertiary Kuroko deposits of Japan (Horikoshi 1969).

The whole idea concerning the subdivision of volcanogenic massive sulphide deposits based on chemical and host rock criteria is only in its infancy. Comparing the work compiled by Mitchell (1973) and Hutchinson (1973) on the subject reveals major discrepancies between the massive sulphide deposits associated with island arc development. The theory is also limited when applied to Pre-Cambrian deposits, due to the lack of information concerning the tectonic regime in this period. However, the theory provides an excellent parameter in discriminating between the two end members, i.e. those deposits of oceanic crust affinity and those of calc alkaline affinity.

part from differing in volcanic rock composition, they differ considerably n mineralogy. The former having a simple mineralogy while the latter are olymetallic. The third type of deposit described, those associated with he initial stages of island arc evolution, have characteristics in between he other deposits. This is not surprising when one considers the magmas roduced are transitional between true oceanic tholeiites and the calc - lkaline suite. Thus, one would expect all three types to be present, lthough not necessarily coeval with each other. Further, as these zones ave been obducted during orogery onto continental material, one would nvisage a complex structural relationship to have developed between the hree magma types.

The Skorovas orebody consists of a simple assemblage of minerals. The predominant material is massive pyrite with smaller amounts of sphalerite and chalcopyrite. Lead is present only in trace amounts. The gangue is comprised of quartz, chlorite and carbonate minerals. The host rocks are tholeiites, possibly island arc and oceanic (Scott 1973), with minor volumes of basaltic andesites, felsic material and greenschist sediments. Calc alkaline volcanics are near absent.

From the brief description of the orebody it is justifiable to exclude its affinity to the calc-alkaline suite. However, the existance of trondjemite and quartz diorite intrusive phases in the Skorovas region, suggests the initial stages in the production of a calc-alkaline suite. The presence of these acid intrusive bodies cannot be used in excluding an oceanic ridge environment as they occur in Cyprus (Bear 1963). Cyprus is thought to be an excellent example of an original oceanic ridge environment. But Miyashiro (1972) has indicated the possible affiliation of the Cyprus ophidite complex as an original Mesozoic island arc environment.

The presence of island arc tholeiites and andesitic basalts indicate an early evolutionary stage in island arc development. This is substantiated by the occurrance of pyroclastic units in the close vicinity of the Skorovas orebody.

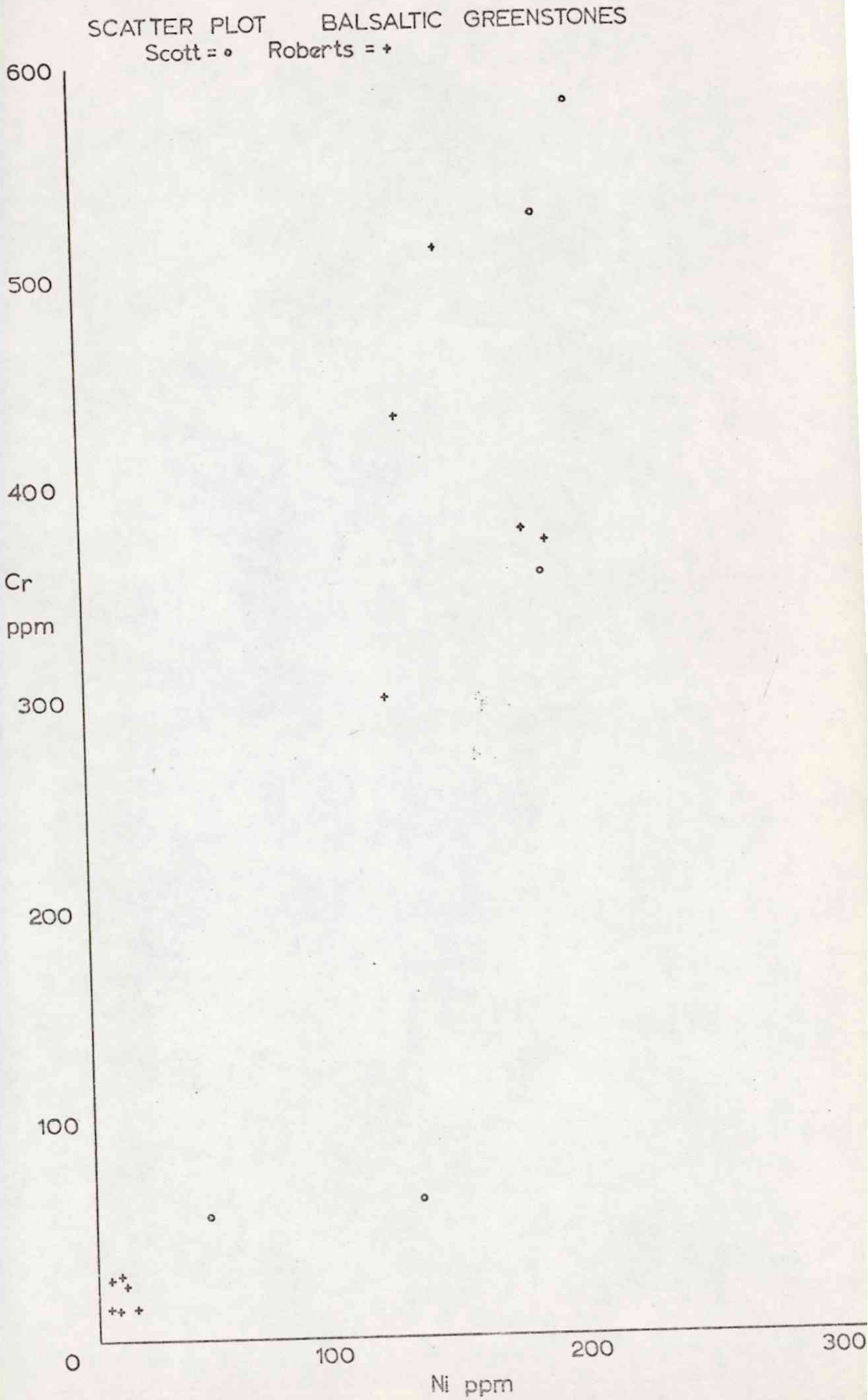
ief Summary of Previous Works.

One of the most comprehensive geochemical studies of the Skorovas egion was compiled by Gjelsvik (1968). He recognises three extrusive ocks in the close vicinity of the orebody; basaltic, andesitic and eratophyric. They are all of spilitic character having high Na, low g and Ca, and mostly very low K. Though these trends are modified djacent to the orebody where the extrusives have undergone pronounced eaching. Here there is a strong depletion of Na and Ca, some leaching f silica, whereas Mg and occasionally K show an increase in concentration. t is unfortunate that no trace elements data is included in his work, ince this could possibly indicate the nature of the basaltic greenstones.

Scott (1973) distinquishes two types of basaltic greenstones, from eochemical and mineralogical evidence in the Skorovas region. The first, called a tholeiitic greenstone is characterised by relict pyroxene phenocrysts, high MgO, Ni, Cr and low Na_2O content. The second type, a spilitic greenstone is more predominant in the region and has a relatively low MgO, Ni, Cr and high Na_2O content. He concludes that the greenstones show no convincing similarities with any one present day magma series but displaying characteristics of both island arc tholeiites and abyssal basalts. The latter statement should be treated with caution since statistically the data was inadequate.

Regionally the Grong district is described as a possible calc-alkaline suite (Gale and Roberts 1972). However, data collected by Roberts (1974) and interpreted by the writer, using the Ti-Zr and Ti-Zr-Y discrimination diagram, indicate that the basaltic greenstones are oceanic tholeiites. The results correspond to the abyssal basalts described by Scott (1973) found in the Skorovas region. Further comparison between Roberts (1974) and Scott's data reveals a considerable similarity between the basaltic greenstones. As previously mentioned, Scott recognised two types of basaltic greenstones. These two groups are also distinquished from Roberts data corresponding to one characterised by high MgO, Ni, Cr and low Na_2O and the other group by low MgO, Ni, Cr and high Na_2O. The former being the tholeiitic greenstone and the latter the spilitic greenstone. Thus, there appears to be a good correlation of basaltic types on a local and regional scale.

FIG 2.0


SCATTER PLOT BALSALTIC GREENSTONES
Scott = ∘ Roberts = +
600
500
400
Cr
ppm
300
200
100
0
100
200
300
Ni ppm

mparing the Ni and Cr content to those of standard present day volcanic ocks, the spilitic greenstones correspond to the island arc tholeiites. e tholeiitic greenstones appear to be considerably enriched in Ni and r, and therefore, cannot be justifiably compared with any standard olcanic rock. However, the volcanic rocks of high Ni and Cr content re exclusively oceanic tholeiites.

The most serious handicap of previous workers has been the lack of ield evidence indicating pillow structures and pyroclastic units. The atter being essential for indicating submarine volcanic activity. Thus, here has been a tendency by previous workers to regard Skorovas as solely oceanic floor origin. Though this possibility still must be considered, evidence provided by pyroclastic units suggest an island arc environment. The possible environments will be discussed at a later stage in the study.

st Volcanic Processes.

tamorphism.

The metavolcanics and metasediments of the Skorovas region have quired their present metamorphic character, not during a single event, it during a series of events. These events involved a period of residence 1 the seafloor and subsequent incorporation in the compressional fields f the Caledonian oregeny as a result of obduction onto the Pre Cambrian asement.

Miyrashio (1972) has given an account of ocean floor metamorphism in itu, in the vicinity of mid ocean ridges. Metamorphic basalts, dolerites nd gabbros which have been dredged from the ocean floor either belong to he zeolite, greenschist or amphibolite facies. A schistosity is lacking n nearly all cases. Associated with low crystallisation in such metamorphic terrains is the introduction of Na and removal of Ca. The outgoing fluids rich in Ca would be ultimately emitted into the sea to contribute to carbonate deposition.

Island arc metamorphism is usually represented by high and low pressure regional metamorphic belts, occurring parallel to each other. The high pressure belt corresponds to the trench zone and is believed to originate from the rapid transport of the lithospheric plate down the subduction zone. On the continental side of an island arc is situated the low pressure belt, accompanied by granitic and andesitic rocks. It is thought to result from the gross thermal and chemical affects caused by the rising calc-alkaline batholiths from the underlying subduction zone.

Mineralogical Alteration and Metamorphic Facies.

The metavolcanics and meta sediments have a mineral assemblage consistent with the greenschist facies of metamorphism. The latter is divided into three (Winkler 1967).

1) Quartz - albite - muscovite - chlorite.

2) Quartz - albite - epidote - biotite.

3) Quartz - albite - epidote - almandine.

Only the last subfacies is absent from the mapping region as almandine nor its basaltic equivalent, an aluminous amphibole were present. This subfacies represents the highest grade within the greenschist facies.

the other two subfacies, the quartz - albite - muscovite - chlorite edominates.

The most noticeable aspect of the greenstone mineralogy is the persistance albite as phenocrysts and in the matrix, though there is some deviation ere oligoclase is present as phenocrysts. Thus, there is a near total eakdown of the primary calci-plagioclose mineral phase to form albite and idote. Further, alteration is manifested by the partial breakdown of bite crystals with the formation of chlorite. The close relationships epidote as inclusions in albite may indicate partial segregation of e two mineral phases during the breakdown of calcic plagioclase. riginal ferromagnesium minerals recognised as phenocrysts have altered chlorite and actinolite. They show a strong tendency to be pseudomorphed ather than corroded as is the case for albite phenocrysts. The scacity f chlorotoid and stilpnomelane in the mineral assemblges suggests that he rocks are materially low in Al_2O_3. The great abundance of chlorite ndicates iron to predominate over magnesium.

oda Metasomatism.

The relatively high proportion of Na to K, characteristic of spilites as been attributed to:-

1) Chemical interaction of Na ions from the seawater with the submarine lavas.
2) Chemical interaction of Na ions from percolating saline ground waters.
3) Chemical segregation during metamorphism in a closed system.

Two schools of thought can be distinguished from the above theories. Both 1) and 2) involve Na ions being introduced into a volcanic system, while 3) concerns the redistributions of Na ions within a closed system. However, when considering the consistency of Na content throughout the entire Skorovas region and the existance of soda rich basalts on the sea-floor, excludes the possibility of the last theory. Miyashiro (1972) suggests the soda rich basalts result from saline sea water trapped between successive lava flows.

Chemical segragation is a noticeable feature of the greenstones, clearly indicated by the widespread occurrance of epidote knots. They are composed chiefly of light yellowish - green epidote.

heir genesis is considered pre-tectonic as epidote clasts occur in
he metaconglomerates and thus, their origin is probably linked to
hemical alteration in the submarine environment. The occurrance of
pidote or veins and amygdal infillings suggest precipitation from
irculating fluids. A more complex process such as remobilization and
elective precipitation, is needed to explain the occurance of epidote
nots in the pillow cores. The presence of small gabbroic xenoliths,
5cm, within epidote, locally developed at the contact of the gabbro
ith the metavolcanics, suggest deep seated aqueous fluids migrating
long the contact zone.

Tectonic and metamorphic redistribution of mineral phases is most
conspicuous in thrust and fault zones. Small scale metamorphic
redistribution of mineral components, particularly epidote, quartz and
calcite as veins, occurs in every horizon mapped and in fact in every
thin section studied.

alytical Methods.

Two analytical methods have been used in this study, the X.R.F. and e direct reading spectrometer.

The first method the X.R.F., due to circumstances beyond the authors ntrol, had to be done outside of Imperial College. For this reason, am deeply indeted to Dr. Ian Gibson and co-workers at Bedford College. here generous help at such short notice made possible the determination f the following minor elements, Ti, Zr, Y Nb, Sr and Rb. The elements f concentrations greater than 5 ppm are of a high degree of accuracy, ithin 2%. However, Nb and a number of rubium concentrations are so low s to be beyond the calibrated accuracy of the machine.

The precision of analysis by the direct reading spectrometer is not ufficient for the purposes of detailed petrographic comparisons. Of he 26 elements analysed, only 9 are within an accuracy of 5%. These elements include Cu, Co, Ni, Cr, Mg, K, Al, Ba and Sr. The most serious rawback to this method is the absence of Si, Na and Fe. The silicon content tended to fluctuate depending on the rock type, although fairly accurate for the acid end members, but showing values considerably below the average expected for basic rocks. Sodium analysis is not possible by this method since it is used as a flux in preparing the samples.

The main purpose of this study is to establish sample criteria for the environmental affinity of the rock suites and therefore, full chemical analysis is not needed. To this extent the analytical data obtained is sufficiently useful to provide a basis for discussion.

rief Sample Description.

W 15 Fine grained Metagabbro. Non-weathered. Epidote and quartz vein present.

Location 5450 72280

48 Metagabbro. Non-weathered. Mostly plagioclase and amphibole. Minor epidote and patches of secondard quartz present. Visible pyrite grain.

Location 3970-71230

49 Pillow Lava. Non-weathered. Lineated amphibole phenocrysts. Minor patches of secondard quartz in addition to epidote and quartz vein present.

Location 3290-69830

I M 17 Amygdaloidal massive basaltic greenstone. Non-weathered. Usually dark grey in colour. Quartz, collite and chlorite amygdal infillings.

Location

G 46 Altered basaltic greenstone. Non-weathered. Highly deformed. Secondary quartz visible.

Location 4140 71440

G 53 Amygdaloidal basaltic greenstone. Non-weathered. Quartz and chlorite anygdal infillings. Minor epidote vein and visible pyrite grains.

Location 6630-69820

G 51 Schistose basaltic greenstone. Non-weathered. amphibole phenocrysts.

Location 5130-68870

R W 32 Amygdaloidal massive basaltic greenstone. Quartz, carbonate and epidote amygdal infillings. Visible pyrite grains.

Location 6280-72880

G 57 Very fine grained epidote rich greenstone.

Location Lower Heisbahn, Gruberfjell.

G 6 Deformed medium grained acid dyke.

Location Grundersfjell.

19 Coarse grained Grondhjenite. Relatively high proportion of chlorite and epidote.

Location 8450 71650

41 Coarse grained Grondhjemite. A minor proportion of ferromagnesium minerals.

Location Pylon hole, N.W. of Stalvikfjell.

44 Acid (felsic) pyroclastic breccia. Rich acid matrix. Keratophyric 'type' fragments.

Location 0170 68200

G 43 Same lithology as G 44.

Location 3200 69750

G 45 Quartz feldspar porphyritic acid (felsic) lava.

Location 0200 67720

G 54A Dark grey quartz rich keratophyre.

Location 6600 69900

G 55B Same lithology and location as G 54A.

S4 144 Dark grey quartz keratophyre. Visible pyrite grains.

Location 3100 74650

R W 17 Quartz feldspar porphyritic lava. Sericite rich matrix.

Location 3950 71050.

R W 17B Same lithology and location as R W 17.

G 42 Leucocratic quartz rich keratophyre.

Location 2370 69590

G 56 Same lithology and location as G 57.

G 52 Schistose basaltic lava. Non-weathered. Amphibole phenocrysts. Minor epidote veins.

Location 5130 68870

G 50 Pillow core. Non-weathered. Lilac coloured carbonate amygdals.

Location 2960 69880

R W 49 Dark greenstone. Weathered appearance. Characterised by a very deep green colour. Rich in stilpnomelane and pyrite.

Location 6150 73600

R W 18 Basic breccia fragment. Weathered. Quartz, calcite and epidote amygdal infillings.

Location 3870 70900

W 48 Pillow lava. Non-weathered. Lineated amphibole phenocrysts. Minor epidote veins.

Location 5600 72000

W 50 Altered lava. Non-weathered. Mostly chlorite, amphibole, epidote and plagioclase. Minor patches of quartz present.

Location 5980 72920

50A Pillow Rim. Non-weathered. Minor quartz veins and amphibole phenocrysts.

Location 2960 69880

55 Massive greenstone. Non-weathered. Minor quartz and epidote veins. Secondary quartz in matrix.

Location Lower Heisbahn, Gruberfjell.

47 Fine grained intrusives. Rich in epidote quartz and feldspar. Possibly andesitic composition.

Location 3450 69940

R W 7 Some lithology as G 57.

Location 4280 71550

MAJOR AND MINOR ELEMENT DATA. GREENSTONES

	'ISLAND ARC THOLEIITES'												'OCEANIC THOLEIITES'		'CALC ALKALINE'			
	G51	G48	G50	RW 48	RW 18	G50A	G52	RW 15	G49	RW 50	G57	IM 17	RW 49	G46	G56	G55	G53	RW 32
SiO_2	32.9	33.8	34.9	41.3	40.9	40.0	48.8	50.9	30.8	49.4	43.7	50.1	56.7	38.1	50.3	48.8	64.6	46.8
K_2O	1.6	0.7	0.8	0.5	0.2	1.2	0.8	0.7	1.7	0.4	0.4	0.1	0.3	0.2	0.1	0.1	0.8	0.4
CaO	5.2	10.8	9.55	8.5	13.0	5.8	2.9	3.8	5.7	5.0	3.9	4.6	2.9	1.9	1.9	2.5	1.3	2.0
MgO	8.4	8.1	7.3	7.4	2.5	5.4	6.8	7.1	6.8	6.1	5.0	5.3	2.5	2.8	2.8	2.0	1.8	2.3
Al_2O_3	12.1	16.7	12.5	10.8	11.7	13.4	12.7	16.1	16.1	23.6	13.4	18.9	14.8	9.8	15.8	14.4	14.5	11.7
Co	31.7	33.8	40.3	33.0	26.0	32.7	33.5	33.9	32.2	28.6	25.0	26.1	8.3	11.0	7.4	12.5	8.0	15.8
Ni	178.2	61.6	205.9	118.3	171.2	106.9	134.7	60.2	73.8	20.2	14.4	9.7	2.2	0.5	0.9	1.0	2.9	2.2
Cr	482.3	158.4	721.0	382.0	428.4	491.9	431.9	250.0	261.7	20.8	11.0	15.7	1.8	2.8	1.4	1.7	24.3	66.0
Cu	31.2	86.9	55.9	43.5	85.4	60.8	61.8	86.1	89.1	91.4	47.7	52.6	16.8	12.7	8.3	10.8	10.9	23.7
Ba	272.7	205.4	119.9	100.3	82.4	100.0	122.7	89.8	292.7	98.1	97.7	42.5	78.6	120.9	45.3	14.9	132.0	107.4
Sr	91	156	108	266	332	143	165	169	279	241	201	229	99	70	125	164	54	226
Rb	20	10	5	2	1	13	5	7	39	1	1	0	2	4	1	0	11	3
Ti	1738	1558	2097	1918	2517	2577	3296	3896	4957	5514	7072	6593	8211	8511	7372	5994	6133	2457
Zr	11	11	15	18	21	27	20	28	27	32	54	62	59	86	124	114	165	81.8
Y	5	3	5	5	8	6	9	10	10	15	23	20	28	30	43	35	52	27
Nb	1	1	1	1	2	2	1	2	0	0	3	4	1	4	1	3	2	2
Fe_2O_3	6.43	4.7	5.8	6.1	3.3	7.4	6.0	7.1	5.9	7.4	6.3	7.4	6.28	7.4	5.1	5.1	5.6	3.0

FIG 2.1

MAJOR AND MINOR ELEMENT DATA ACID ROCKS														
	TRONDJE-MITE		PYRO-CLASTIC		KERATOPHYRE				Qtz. Feld. Porphyrite		Felsic Lava		Andesites	
G1	G19	G41	G44	G43	G54A	G55B	$S_4$144	G42	RW17	RW17B	G45	G6	RW7	G47
SiO_2	68.2	64.4	77.4	72.0	61.9	83.4	65.4	60.3	54.6	72.5	68.7	54.8	63.8	46.2
K_2O	0.5	1.1	1.4	0.9	0.1	0.1	0.1	0.2	1.0	1.1	2.5	1.8	1.8	2.0
CaO	2.7	0.7	0.7	1.0	0.6	1.5	1.0	0.8	0.1	0.3	1.7	1.5	2.4	2.2
MgO	0.7	0.5	1.5	0.5	0.5	0.7	0.3	0.7	0.3	0.5	0.8	0.3	1.5	1.2
Al_2O_3	12.5	10.0	15.5	10.2	14.5	15.3	12.5	10.0	10.4	13.0	11.5	11.3	14.2	11.9
Ba	75.5	84.2	212.8	59.2	33.7	572.9	17.3	18.4	295.0	214.7	95.8	347.5	522.9	445.7
Sr	148	82	46	125	74	985	65	94	40	40	64	630	885	315
Rb	10	9	43	22	0	42	0	3	22	20	34	93	42	52
Ti	1378	1078	659	599	1738	2577	1138	659	719	659	1798	1438	2677	2697
Zr	59	106	161	122	172	168	216	136	143	145	133	172	180	173
Y	18	26	79	50	59	48	59	46	37	40	33	8	18	19
Nb	2	1	1	0	2	2	4	0	0	2	2	9	6	7
Fe_2O_3	2.4	1.0	1.9	0.7	2.0	3.3	1.4	0.7	0.9	1.0	1.7	1.0	3.1	1.8

FIG 2.2

scussion of Results.

The analytical results have been discussed in terms of four asses of information.

1) Major oxide analysis.

2) Minor elements of calcic and alkalic affinity.

3) The group Ti, Zr, Y and Nb.

4) Minor elements of the transitional group.

ajor Oxides.

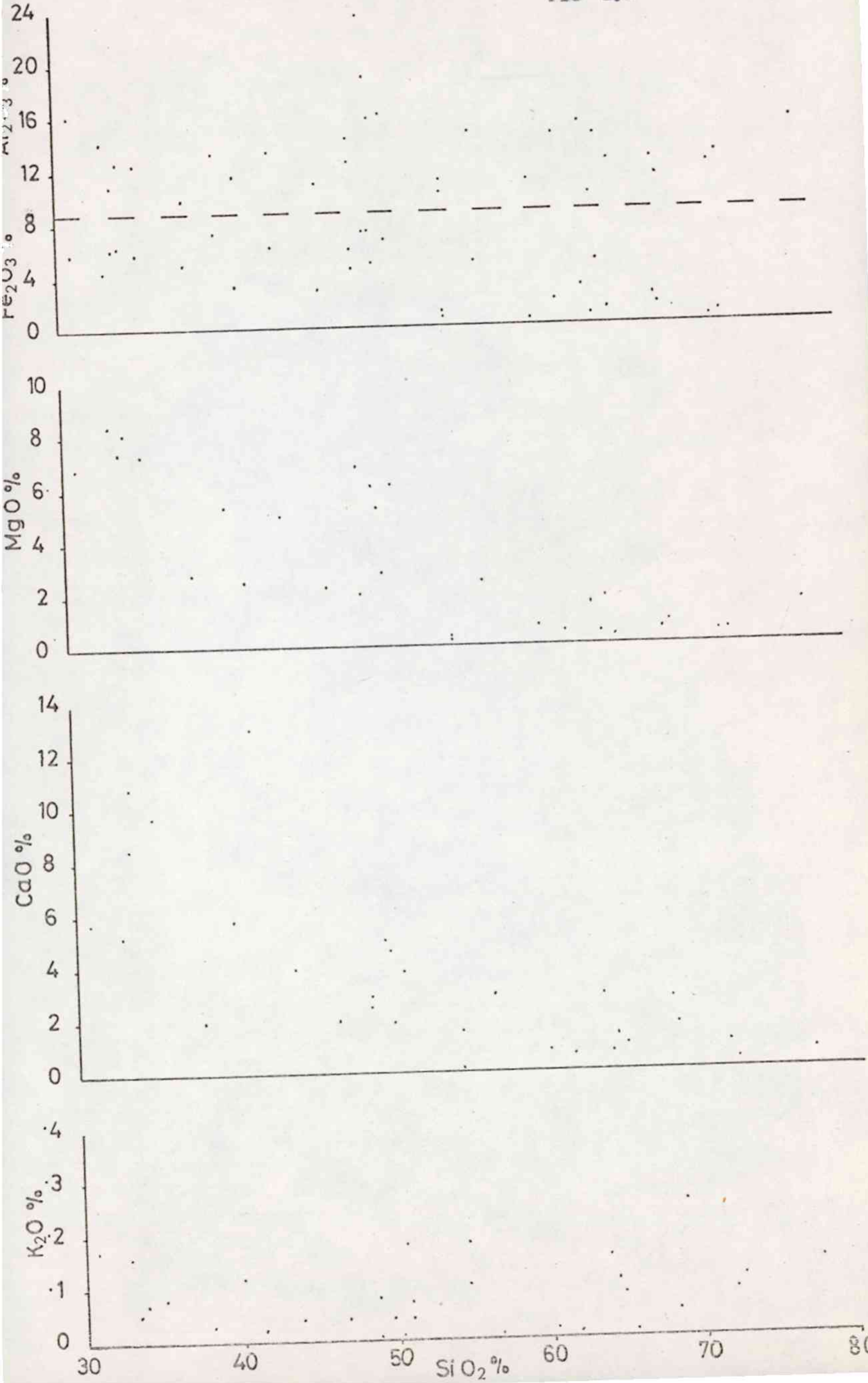
Any meaningful subdivision of the greenstones incorporating major xide analysis is severely restricted by the absence of Na, an essential kalic component, and the rather inaccurate silica values obtained. ecause of inaccurate analysis, coupled with the relative mobility haracteristics of certain major oxides, excludes conventional comparison echniques involving A.F.M. and alkalic-lime methods. Nevertheless, ome interesting patterns have emerged. A very interesting aspect of he major oxide analysis can be seen by comparing fig. 23 with fig. he former, a Harker variation diagram in which MgO and CaO show a general lepletion trend towards the acid rock members, though with a fairly wide scatter. A large proportion of the scatter can be attributed to inaccurate silica data. However, when the silica is replaced by the element zirconium (fig.24), as the abissica variation parameter, the trend is better defined and the plots considerably smoothed. The use of Zr as a silica indicator of the samples is substantiated by visual inspection of the rocks. However, the Zr content should not be used as a guide to the silica content of the tholeiitic greenstones.

The silica values for the acidic rocks clearly define samples with an andesitic character. Both G 53 and R W 7 have silica concentration in accordance with andesite. Certain discrepancies do arise since G 47 is very similar mineralogically and physically to R W 7, but has a very low silica content of 46.2%. Again in samples R W 17 and R W 17b, which are both quartz and feldspar porphyritic lavas, but differ in silica content of nearly 20%. Thus, even in the more acid members a certain amount of scepticism is necessary before interpreting the results.

Harker Variation Diagrams For The

Volcanic and Plutonic Rocks

FIG 2.3



Variation Diagrams for the Volcanic and Plutonic Rocks

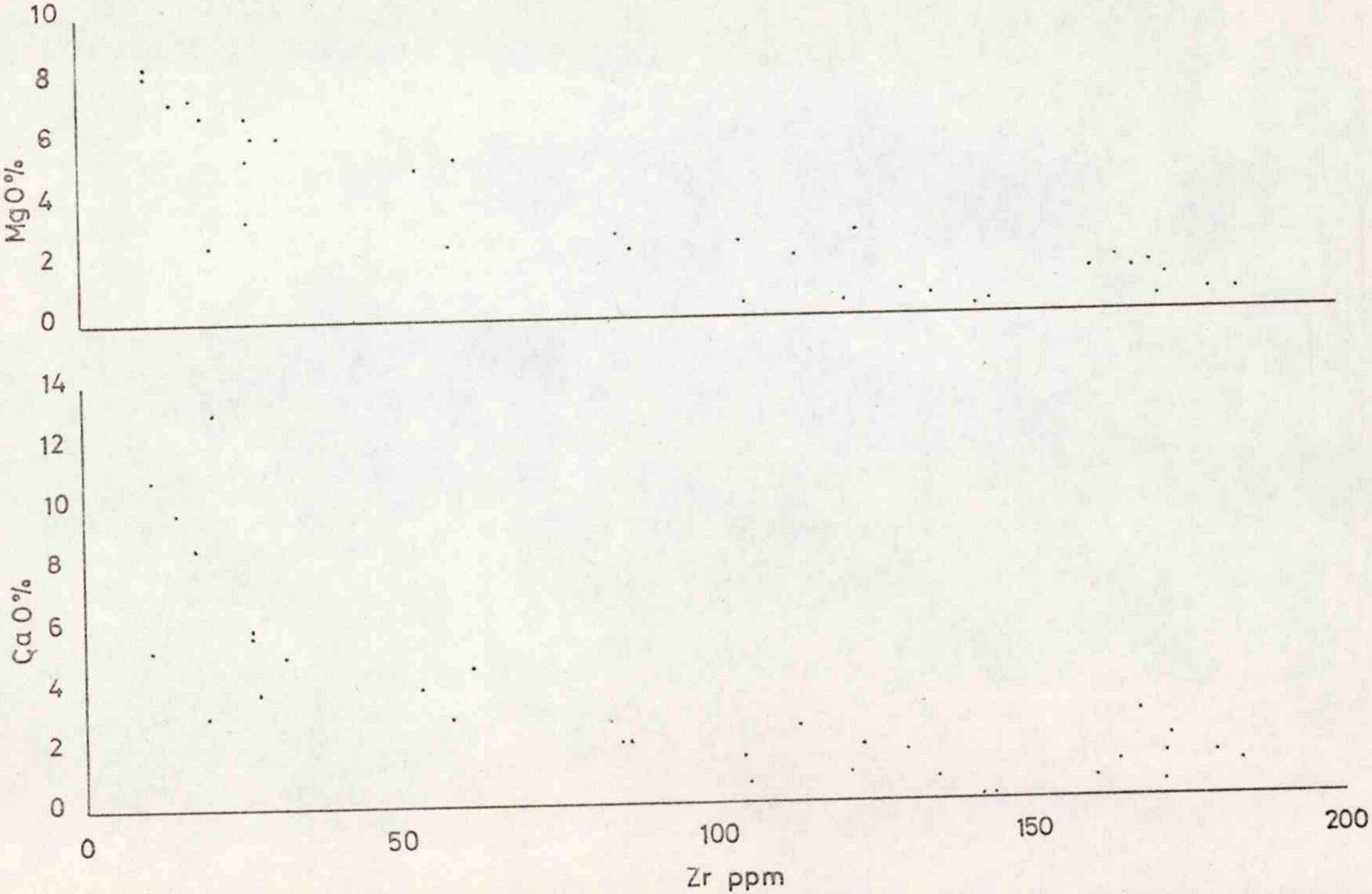


FIG 2.4

Mineralogical observations in addition to macroscopic field evidence suggest a considerable variation of major element concentrations. This appears to be the case from inspection of the data. The Harker variation diagram indicates a vague concentration of K_2O towards the acid end members, but there is little internal correlation within either the acid or basic fields.

A subdivision of the basaltic greenstones based solely on MgO and CaO concentration is not thought justifiable, although a certain group of greenstones have relatively high concentration of these major oxides. They are also characterised by high concentrations of transition elements and low Ti and Zr content. This data suggests an island arc site of origin for this group of greenstones as will be discussed presently. The rather erratic distribution of MgO and CaO within this particular group is attributed to the variable proportion to different ferromagnesium minerals present in each sample. Tremolite and actinolite contain Ca and a variable proportion of Mg and Fe elements, while epidote contains Ca and chlorite the element Mg.

The considerable fluctuation of Al_2O_3 results from the breakdown of primary calcic plagioclase and manifestation of various ferromagnesium minerals.

Sr, Ba and Rb.

These trace elements have alkalic or calcic affinities and are thus subject to fluctuation as a result of post volcanic alteration.

The Sr values show a considerable range of concentrations but show a vague depletion trend towards the acid end members. Although corresponding to the depletion trend of CaO, the Sr shows little correlation with the CaO concentrations for individual specimens. The apparent independence between Sr and CaO suggest the former occurs within other atomic sites rather than just a straight Ca substitution relationship. This is substantiated by Sr enrichment at the pillow margins relative to the core, which is the reverse of the CaO trend.

The existence of anomously high value such as occurs in R.W 7, an andesite, where the Sr concentration approaches 1000 ppm, suggests a genesis related to Sr bearing pore fluids. However, the concentrating mechanisms and the Sr relationship with the associated minerals is not known.

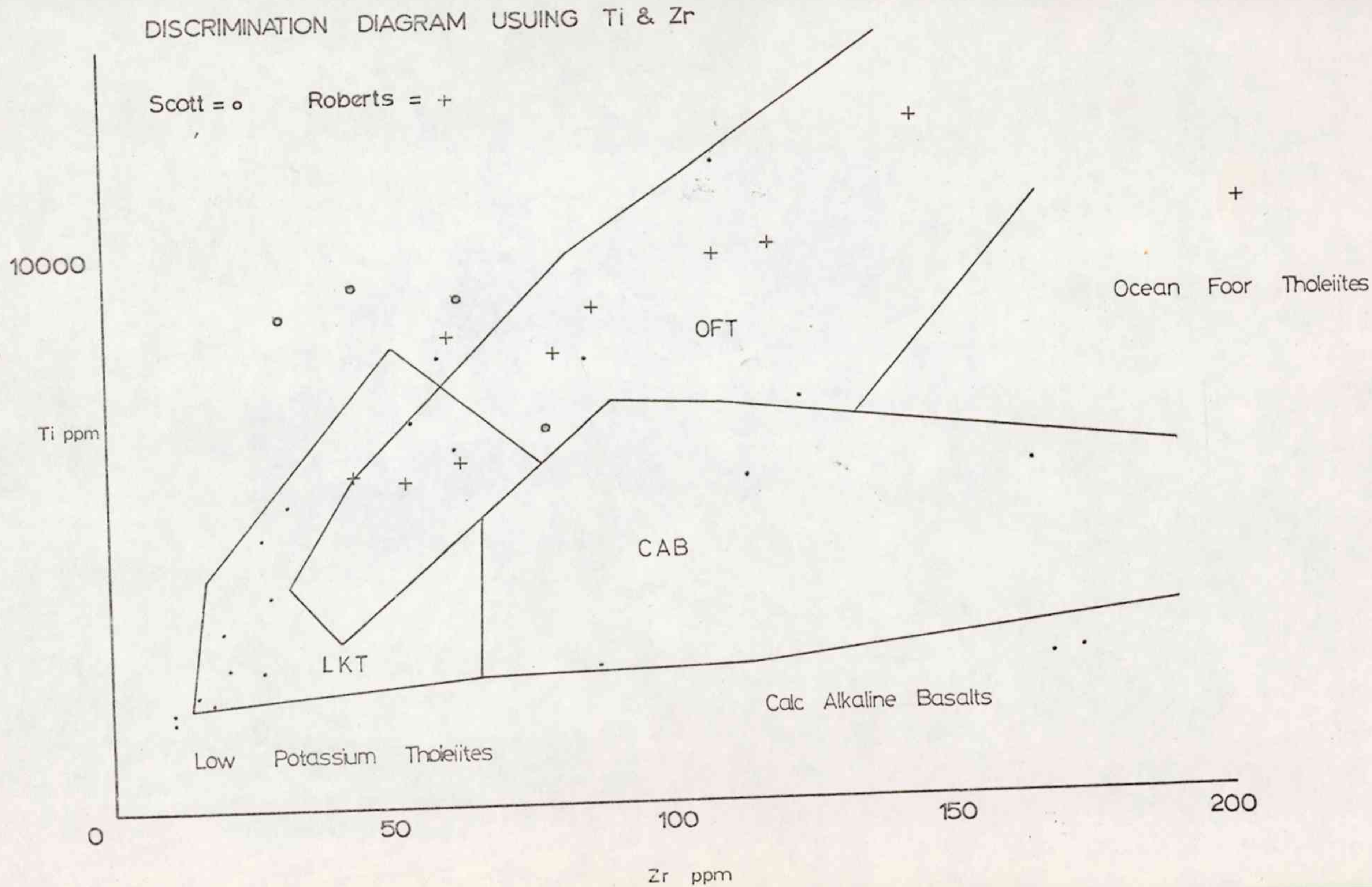
Barium shows a completely unsystematic distribution and appears
be unrelated to alkalic or calcic major elements. The greenstones
iow barium concentrations within the range or island arc tholeiites,
it occasionally values are considerably higher. However, the values
re well above the range for oceanic tholeiites. Pillow and core
amples show no significant deviation, suggesting a minimal redistribution
ie to ionic exchange with sea-water. Even the heavily deformed samples
how no marked depletion of barium.

The Rb values are generally low for all the rocks analysised,
ut concentrating towards the acid end members. This corresponds to the
enerally higher K_2O content of the acid members. There is moderate
orrelation between Rb and K_2O content for individual samples. Rb appears
o have been enriched by 60% at the pillow margins relative to the core,
hich correlate with the increase of K_2O content at the margins.

Y - Zr - Ti and Nb.

Pearce and Cann (1971) have demonstrated the use of trace elements Y - Zr - Ti and Nb in identifying ancient volcanic rocks and the paleo-tectonic setting. A more detailed summary of their conclusion has already been outlined.

The majority of basaltic greenstones analysed plot on the Ti - Zr discrimination diagram in the field of low Potassium tholeiites (fig. 25). This is analogous to island arc tholeiites. This group of greenstones analysed, show considerable variation in mineralogy and physical properties. They include massive, schistose, pillow, porphyritic, lavas and breccia fragments, showing varying degrees of deformation. The variation of samples confirm to a certain extent, Pearce and Cann's theory concerning the preservation of these elements, during post volcanic alteration. They are therefore, dependent on the primary chemical characterisics of their parent magma. There is however, some moderate variation between samples and more surprisingly within a single pillow structure. Here, analytical data indicates that the pillow rim is enriched by 25% Ti and 100% Zr relative to their core. The variation in concentrations is beyond the scope of the writer and either suggests a substantial amount of doubt concerning the stability of these elements during post volcanic alteration,



DISCRIMINATION DIAGRAM USUING Ti & Zr
Scott = o
Roberts = +
10000
Ti ppm
OFT
Ocean Foor Tholeiites
CAB
LKT
Calc Alkaline Basalts
Low Potassium Tholeiites
0
50
100
150
200
Zr ppm

FIG 2.5

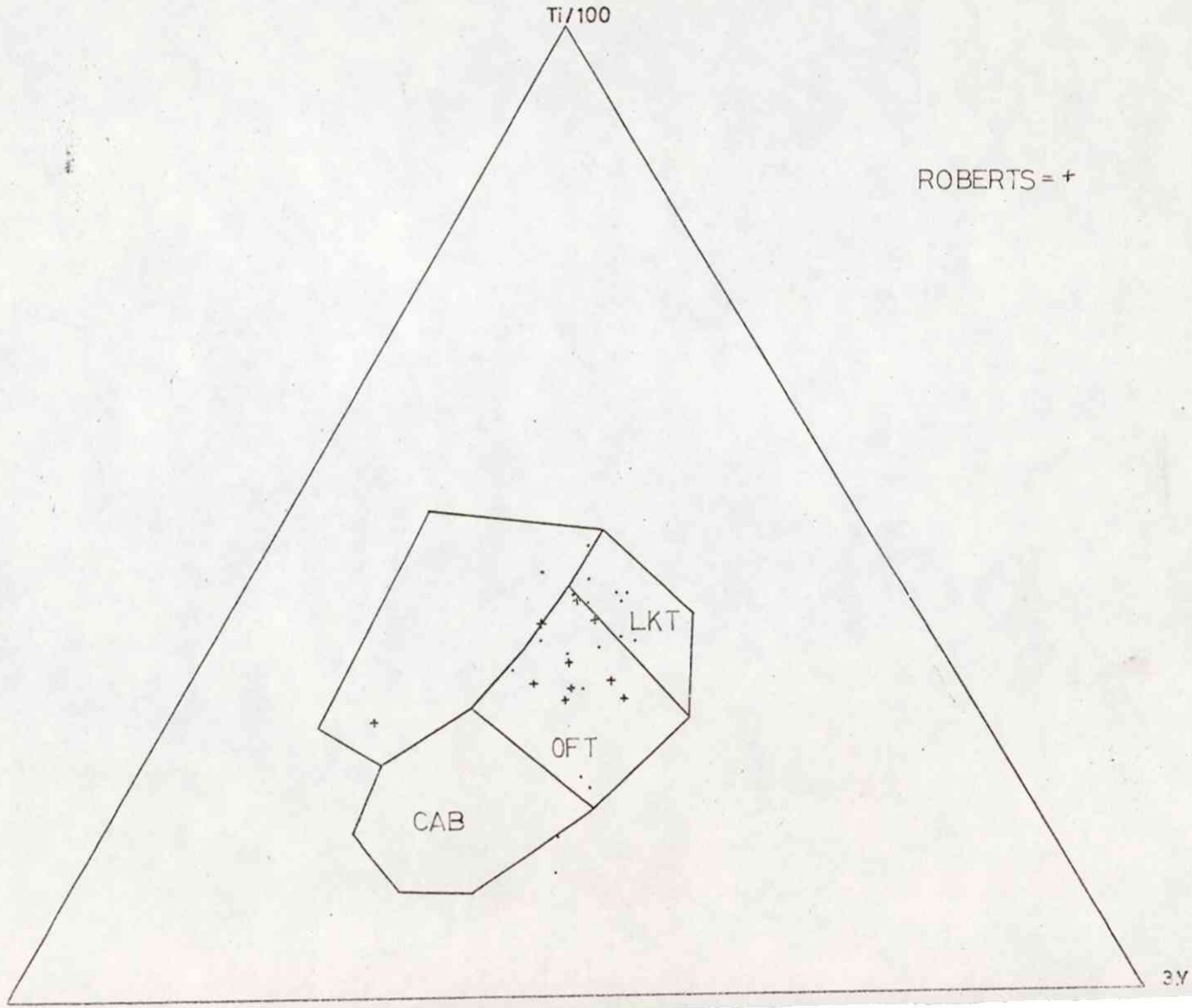


FIG 2.6

r fractionation occurs within a cooling pillow structure. But as Zr s a rather large incompatible ion, one would suppose enrichment occurs luring the last stages of solidification, i.e. at the pillow core. The converse appears to be true though.

Although a certain amount of variation occurs within this group of greenstones, they are bounded by relatively narrow limits. There are 35 ppm Zr and very low Ti values 6000 ppm. The latter being equivalent to about 1% TiO_2.

The predominance of an island arc tholeiitic character is supported by the low Y concentrations, although these tend to be considerably lower than the average for this type of rock. The low Y values tend to move the plots on the Ti - Zr - Y (fig. 26) towards the ocean floor basalts, but still 50% of the total greenstone samples plot in the field of low potassium tholeiite field. Two gabbros analysed also plot within this field.

Of the other greenstones analysed, two plot in the oceanic field and are characterised by substantially higher Ti and Zr content than those of island arc affinity. Four greenstone plots straddle the boundaries of the calc alkaline field. There does not appear to be any significant correlation between the plots, though tending to have high Zr and Y contents. The Zr concentration is in the order of 100 ppm.

From the trace element data it would appear that the greenstones are of island arc affinity, and more specifically, the earlier stages of evolution. This is evident, as they plot in the field of low potassium tholeiites. However, as previously mentioned, data collected by Robert's (1974) from the Grong region plot on the Ti - Zr and Ti - Zr - Y discrimination diagram in the field of oceanic tholeiites. Thus, assuming Pearce and Cann's (1973) method to be valid, there appears to occur two types of tholeiitic basalts in the region.

The felsic rocks are characterised by low Ti 3000 ppm, concentration and relatively high Zr 100-200 ppm content. The latter is more variable than Ti which is the converse for the greenstones. There is surprisingly only moderate variation of these elements considering the physical difference in character between individual samples.

The felsic pyroclastics, quartz porphyritic lava and keratophyres have very similar Ti and Zr content which strongly suggests a genetic relationship between these rocks.

Two two trondhjemites analysed show a considerable variation in r content. The lowest value is characterised by a high ferromagnesium ontent in the trondhjemite, Both however, have lower Zr concentration han the other felsic rocks. This is rather unusual since Zr concentrates s a late stage fractionation phase within a magma. Thus, one would xpect the Zr content of the trondhjemite to be higher than that of the 'elsic extrusives and keratophyre.

Transition elements.

Transition element variation within the greenstones provides a sensitive criterion for distinguishing two lava types, those of high and those of low transition element concentration.

From fig.26-7 there is a pronounced negative correlation between the Cr - Ni - Cu - Co concentrations and zirconium content. As all four of the transition elements correlate, suggests that they occur in a single mineral phase. This is rather an exception to the general rule, as Cr is preferentially concentrated in pyroxene and Ni in olivine. Considering the presence of relict pyroxene phenocrysts in many of the greenstones analysed, indicates that the two elements are contained within pyroxene and possibly olivine was subordinate or absent in the original rock. The pyroxenes are now pseudomorphed by amphibole and chlorite. The depletion of copper may relate to progressive separation of the copper as a hydrothermal fraction.

The enormously high Ni content 60 - 200 ppm and Cr 150 - 500 ppm correspond to the greenstones which plot in the field of low potassium tholeiites on the Ti - Zr discrimination diagram. Scott (1973) terms these rocks as tholeiitic greenstones. The Ni and Cr concentrations exceed the predicted average value for low potassium tholeiites, by a factor of 8. However, the values are compatible for those of oceanic tholeiites which range between 30 - 200 ppm Ni and 200 - 400 ppm Cr content. Thus, there appears to be a considerable discrepancy between conclusions derived from Ti - Zr content and the observed values for Ni and Cr. Two possible explanations are put forward to explain the apparant high Ni and Cr concentrations, assuming the rocks to be island arc affinity. The first considers that these particular greenstones are a manifestation of a very primitive and largely undifferentiated basic magma. This is purely speculative.

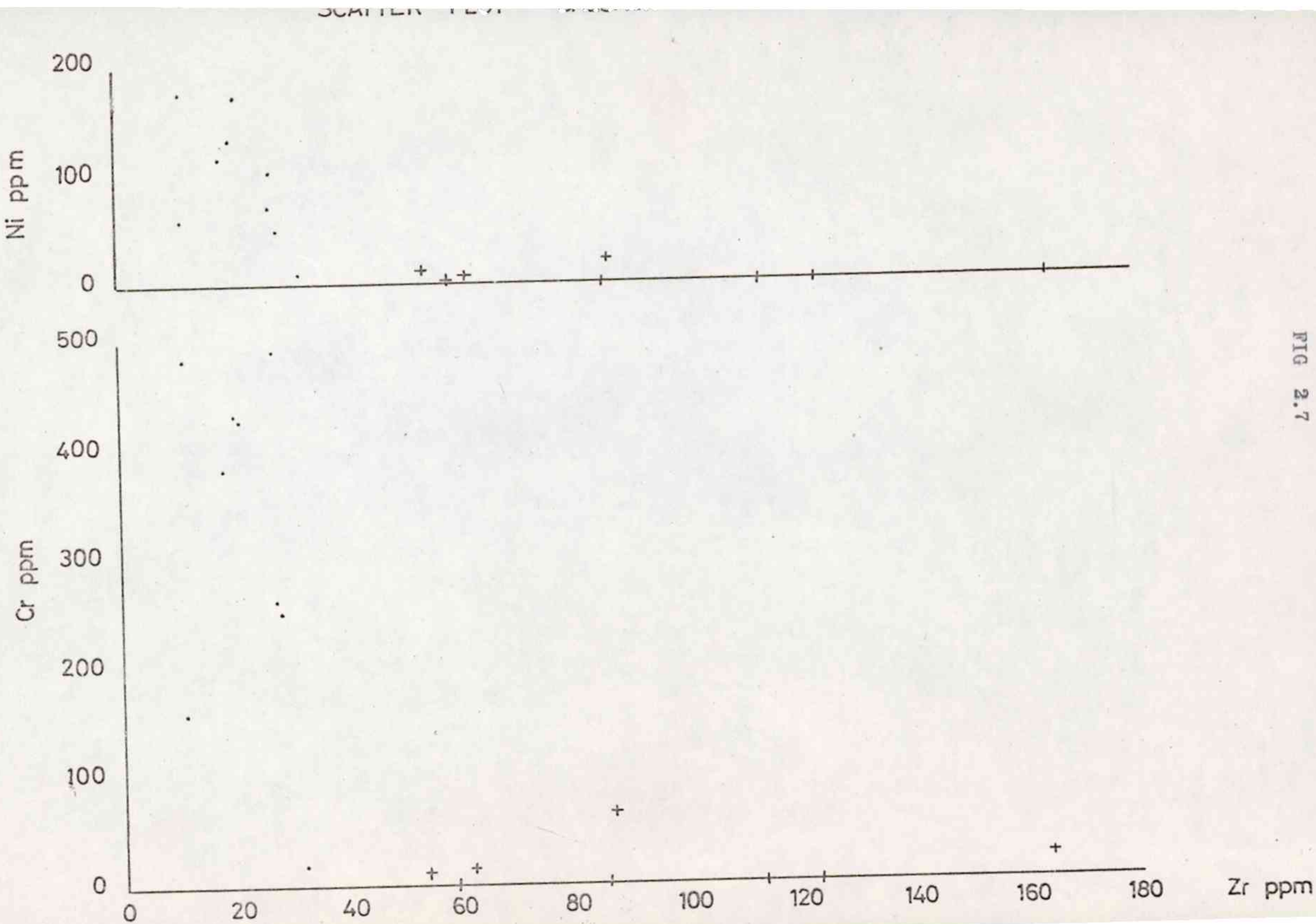

200
100
0
Ni ppm
500
400
300
200
100
0
Cr ppm
0
20
40
60
80
100
120
140
160
180
Zr ppm

FIG 2.7

SCATTER PLOT

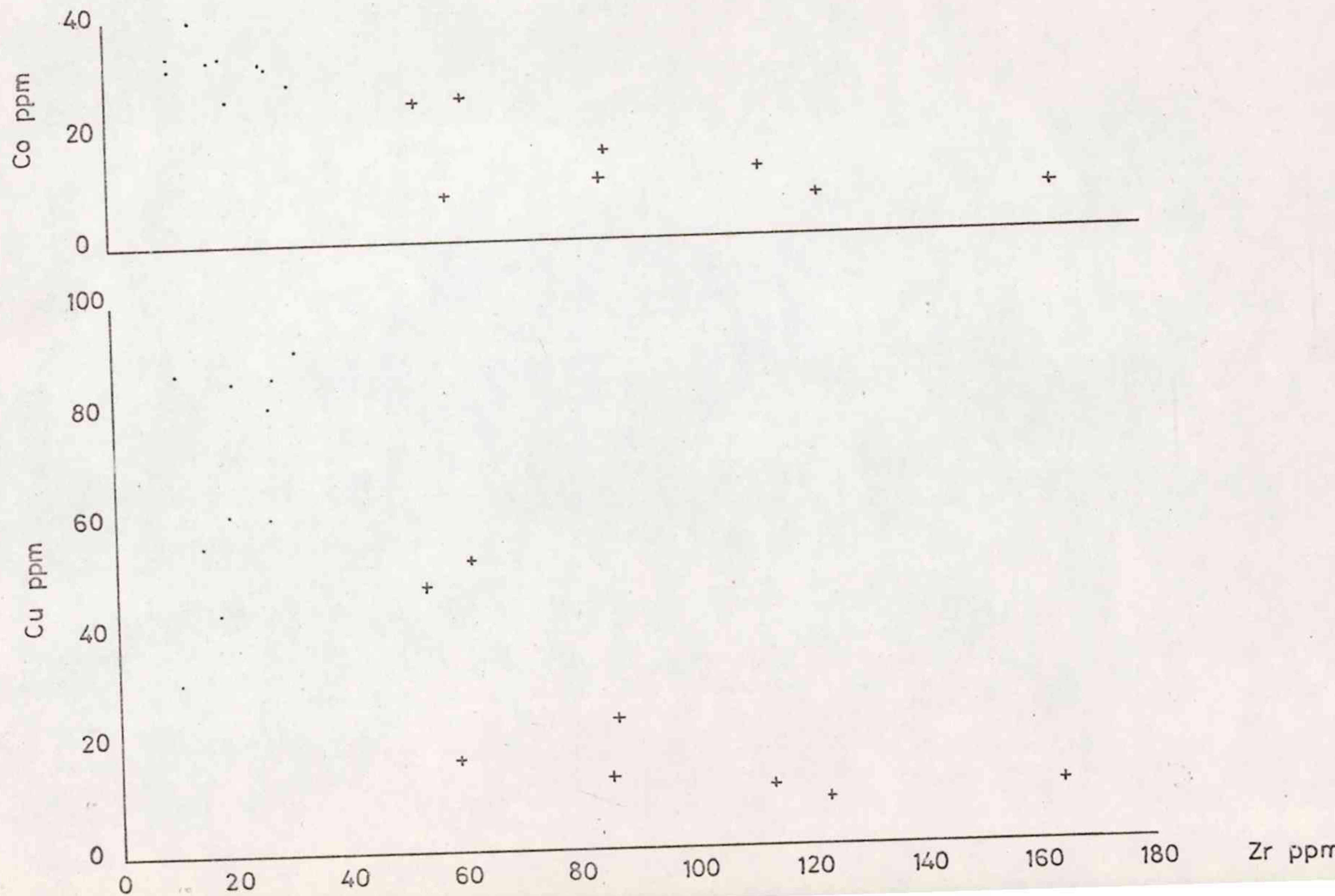


FIG 2.8

he second, and more plausible explanation suggests that high Ni and r contents are a local phenomena, perhaps associated with an oreforming nvironment such as that of Skorovas. Massive pyrite samples from the korovas region, analysed by Palmer (1972) give transition elements remarkably similar to those obtained in this group of greenstones. Thus, prior to the genesis of the orebody, the greenstones contained enormously nigh transition element concentrations.

The remaining greenstones are characterised by a low Ni and Cr content but there appears to be a systematic decrease in Cu and Co within the entire range of greenstone analysed. These rocks correspond to those termed by Scott (1973) as spilitic greenstones. On the Ti - Zr discrimination diagram, some plot in the oceanic tholeiitic field while the rest in the calc alkaline field. The latter having the lowest values for transition element concentration.

There appears to have occurred enrichment of certain transition elements within the core of a pillow structure relative to its rim. Ni shows enrichment of a 100%, Cr 45%, but Cu and Co are constant within a few percent. The Ni and Cr enrichment correlate with the increase in MgO and CaO and decrease of Zr at the pillow core. The variation may explain the rather poor internal correlation within the high Ni and Cr greenstones. The lower Zr values at the pillow core suggests migration during the initial formation of the pillow structure, as Zr is a relatively stable element during post volcanic alteration processes. Relating this explanation for the variation of Ni and Cr, would indicate that the difference in concentration are primary events and only partially due to secondary alteration.

The acid end members have very low Ni 5 ppm and Cr 5 ppm concentrations.

Conclusion

In the final section the author has attempted to correlate known eological evidence from a rather small section of the Skorovas region ith the geochemical data. The writer is fortunate in having mapped an rea, justifiably considered as 'a Geological sample' of the Skorovas egion. However, conclusions are limited to some extent by the inter-retation from the geochemical data. The theory in which this inter-retation depends, is based on empirical relationships between trace element ontents in volcanic rocks. The empirical approach is, perhaps, debatable ut since it has proved a useful working tool, it has gained acceptance n the recent geological literature.

A geological investigation provides a foundation for the present iscussion in determining the paleo-tectonic environment of the Skorovas egion. The major geological features and their interpretation are ummarised in synophis as follows.

Two aspects of igneous activity are recognised in the Skorovas egion. The predominant phase consists of submarine extrusion of basaltic lavas accompanied by gabbroic intrusion. The other phase of igneous activity resulted in the high level intrusion of trondhjemite and keratophyric bodies, which were accompanied by local explosive volcanism, forming the acid pyroclastics and felsic lavas. Within there extremes, though not necessarily in time, are minor andesites and possibly rhyolites.

The most significant aspect of the geological investigation is the presence of the acid pyroclastic breccia, thus proving volcanic explosive activity not known previously in the region. The volcanic centre is believed to be located at Skorovas as the frequency of the pyroclastic unit decreases along strike away from the orebody.

The nature of the greenstones excludes any possible clarification based on petrological evidence or major element analysis. Therefore, much importance is placed on the results of the trace element concentrations. Geochemical data based on Ti - Zr concentrations indicate that the majority of greenstones fall in the range of island arc tholeiites (low potassium tholeiites). However, considering the Ni and Cr contents alone, the greenstones would be clasified as oceanic tholeiites.

aring in mind the close similarities of the two types of tholeiites, th products of the upper mantle, suggest a certain amount of overlap etween chemical characteristics. The high Ni - Cr - Cu concentration f the greenstones are compatible with concentrations found in the massive yrites occurring in the Skorovas region. It is suggested by the author, hat prior or coevial to the information of the Skorovas orebody, normously high values of Ni - Cr - Cu characterised the island arc holeiites.

It is felt necessary to place the Skorovas area in perspective with ts immediate surroundings, the Grong district, and its equivalent further field, the major greenstone belt of Norway.

Regionally, Gale and Roberts (1972) have described the greenstones from he Grong district as lavas of calc alkaline affinity. They clasify much f the southern greenstone belt of Norway as calc alkaline with a western lank of tholeiitic greenstones. Extrapolating the boundary separating the two fields northwards, implies that the tholeiitic field lies some 100 km west of Skorovas. However, geochemical data interpreted by the writer from Roberts' (1974) recent work in the Grong destrict, indicates the greenstones are oceanic tholeiites. But in the immediate vicinity of Skorovas the greenstones are island arc tholeiites with only minor volumes of calc alkaline rocks. Further evidence excluding a predominant calc alkaline suite in the area is provided by the simple assemblage of minerals constituting the Skorovas orebody. This can only be interpreted as a product of early island arc evolution or alternatively an oceanic ridge environment. Deposits associated with the calc alkaline suite of rocks are characteristically polymetallic with a complex assemblage of gangue minerals.

The existence of two types of tholeiitic rocks occurring on the scale such as the Grong district can be explained by considering present day island arc environments. The island arc tholeiites are initially extruded onto an oceanic crust comprised of the oceanic tholeiites. Since the greenstone belt constitutes an allocthonous strip, one would expect both types of tholeiites to be present in close proximity.

Gale and Roberts (1972) also suggest that as calc alkaline igneous activity continues, a more acidic product is formed, namely the trondhjemite nd extruded felsic equivalents. Thus, the trondhjemite post date the alc akaline phase and represent the last stages of igneous activity rior to nappe emplacement. However, at Skorovas the felsic lava irectly overlie the tholeiitic greenstones. Therefore it is unlikely hat the trondhjemite is an end member of the calc alkaline suite, and onsidering the near absence of the latter, suggests that it was ntruded before calc alkaline activity.

The trondhjemite and associated extrusives may relate to special onditions within the magma chamber which oppresses, the development of a typical calc alkaline phase, or favours the former relative to the latter. However, the presence of granodioritic in the Skorovas region suggest at least the initiation of calc alkaline activity. The rising baccolith within the island arc environment.

The theory relating to the origins of trondhjemite indicated previously, contradict that proposed by Gale and Roberts (1972) based on evidence derived from the southern regions of the greenstone belt of Norway. Thus existence of trondhjemite bodies along the greenstone belt, though apparently intruded into different host rocks suggest, that the bodies are consanqueneous and to some extent, unrelated to the evolutionary pattern of island arc development. As the trondhjemite bodies are intruded prior to nappe emplacement, suggests that they are genetically related to underlying continental material, Sialic material caught up in the Benioff zone during the last stages of Proto-Atlantic contraction and subsequent manifestation of the Caledonian orogeny. To substantiate the age relationships between the trondhjemite intrusions and there affiliation with continental material is beyond the scope of the present study. Thus, the above theory is purely speculative and considerably lacking in factual evidence.

The predominance of the calc alkaline phase envisaged by Gale and Roberts (1972) in the southern regions of Norway suggests a lateral transition, from a fully developed arc system to an incipient arc system towards the central and northern parts of Norway. The lateral variation may relate to the sense of contraction of the Proto Atlantic Ocean. The mature island arc development in the south representing an advanced stage of contract while to the north and central parts of Norway, the initial stages of contraction. Thus, the sense of contraction is southwards.

The sequence of events as envisaged for the Skorovas region:

1) Initial stages of island arc evolution with the extrusion of island arc tholeiites on the oceanic tholeiitic crust. The development of a back arc marginal geosyncline.

2) The growth of volcanic pile and accompanied gabbroic intrusions. Spilitization of the tholeiitic lavas.

3) Last stages of tholeiitic generation and the formation of the Skorovas orebody.

4) High level intrusion of trondhjemite material and minor felsic extrusive activity, with the deposition of vasskis and jasper. Possibly the initial stages of calc alkaline activity. Deposition of limestone in the deeper parts of the marginal basin.

5) Sporadic parogenic movement resulting in the accumulation of volcanogenic clastic sediments (fine felsic conglomerate).

6) Major tectonic activity and the formation of the coarse conglomerate and finer (arkose) volcanogenic sediments in the deeper parts of the basin.

7) Nappe emplacement.

The conclusions summarised above depend to a large extent on logical deduction, based on field work, but supported by analytical studies, the basis of which has been the principle of uniformitarism as applied to the petrogenesis and geochemistry of Phanerozoic volcanic sequences.

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