Metalliferous Deposits within Newfoundland Ophiolite Complexes

by

John F. Copp

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Robert B. Scott

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Abstract

Because of their apparent metalliferous concentrations, six metamorphosed sedimentary rocks were collected from the island of Newfoundland. The samples were studied in order to determine the origins of the metals contained within. X-ray diffraction, atomic absorption, and thin section work revealed a hydrothermal origin. Through comparison to metalliferous sediments found in active spreading centers, a similar origin for the samples was concluded. The metals are the result of precipitation from fluids which were heated by, and reacted with, oceanic lithosphere which was still cooling at that time.

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Table of Contents

List of Tables

List of Figures

Introduction

In this study I worked with a collection of rocks gathered from the island of Newfoundland. The rocks, originally sediments, were collected because they showed evidence of being rich in various metals. My primary objectives in studying these rocks were: 1) to identify the mineral assemblages present, 2) to determine the types and abundances of the metals present, and 3) to identify the individual origins of these deposits.

In recent years, many researchers have been studying metalliferous deposits associated with igneous oceanic lithosphere. Deposits have been found on top of the igneous rocks of the sea floor and at the base of the sediment column which show evidence of having been precipitated by hydrothermal fluids (see discussion, page 3). Deposits of metal sulfides have been found within the pillow basalts of ophiolite complexes and have also been found to be the result of hydrothermal activity, while the associated igneous rocks were relatively fresh and still forming (Upadhyay and Strong, 1973). Little work has been done on the metalliferous sediments immediately overlying the igneous rocks of ophiolite complexes.

Newfoundland contains several ophiolite complexes which run through the full sequence of rock types, and which allow extensive sampling. The materials I worked with were within the sedimentary column either immediately overlying igneous oceanic crust, or overlying igneous materials known to have formed on the sea floor, the assumption being that they were subjected to similar conditions and processes, for they all showed signs of a concentration of metals. However, they showed little resemblance to the metal-rich sulfides which were known to occur in abundance within the underlying pillow basalts. Through studying these rocks, I hoped to find evidence of hydrothermal activity enriching the sediments in metals.

Sea Floor Metal-Rich Deposits

Metalliferous deposits rich in iron and manganese are widespread on the sea floor, and have been extensively studied. Two primary views as to the sources of these metals have been expressed, one being the slow precipitation of metals by "normal" seawater, the other being precipitation from hydrothermal waters.

Hydrogenous deposits are formed by the slow oxidation and precipitation of metals taken into solution through weathering of the continents. Iron, manganese, and other metallic elements have low solubilities in alkaline and oxidized liquids like seawater, so their extraction is reltively simple. The deposits form only in areas of the sea receiving minimal amounts of material from terrigenous sources, otherwise the nonmetalliferous sediments would dilute the deposits and prevent the concentration of metals. This tends to restrict the formation of such deposits to abyssal plains and seamounts. These slowly accumulating deposits form sea floor concretions known as manganese nodules.

In areas of the sea floor that are undergoing relatively high rates of deposition of organic matter, diagenetic deposits may form. Hydrogenous metals are buried in the sediments along with the organic matter. The gradual decomposition of the organic matter creates reducing conditions by depleting the oxygen contained within the pore waters of the sediment (Bonatti, 1978).

This leads to re-solution of metals such as iron and manganese, and their concentration within the pore waters of the sediments. With compaction of the sediments, these pore fluids migrate upwards until they reach areas within the sediments that are still oxidized, where the metals re-precipitate and are deposited as oxides, and the cycle begins again. If the oxidized zone lies near the boundary between the sediments and overlying seawater, metal crusts and concretions may form on the sea floor (Bonatti, 1978). Iron may be extracted from this cycle by combining with sulfur, which is less mobile than iron oxides, so that once buried, the iron sulfides remain within the sediments.

Hydrothermal Processes

In the mid-1960's, several investigators working in the vicinity of the East Pacific Rise reported geochemical anomalies centering along the axis of the ridge. Boström and Peterson (1966), of the Institute of Marine Sciences at the University of Miami, and Skornyakova (1964), of the Institute of Oceanography of the U.S.S.R., detected unusually high concentrations of iron and manganese from sediments in this area. Joensuu and Bonatti (1966), also of the University of Miami, obtained deposits from the crest of the East Pacific Ridge containing more than 30% iron by weight.

The formation of metal deposits similar to those described above (see Table 1) has been directly observed in areas characterized by volcanism, such as near the island of Thera in the Mediterranean (Bonatti, Honnorez, Joensuu, Rydell, 1972) and near Matupi Harbor in the western Pacific (Ferguson and Lambert, 1972). In both places, the deposits were seen to form by the precipitation of metals from warm solutions, rich in silica and metal (see Table 1) that were discharged through the sea floor by hydrothermal springs. In addition, active hydrothermal springs have been observed in the vicinity of the Galapagos spreading center (Ballard and Corliss, 1977). These deposits are chemically distinct from hydrogenous deposits, which show Fe to Mn ratios near unity (see Figure 1).

One accomplishment of the Deep Sea Drilling Project (DSDP), aboard the Glomar Challenger, was the discovery of metalliferous deposits buried at the bottom of the sedimentary column, immediately overlying basaltic basement. These deposits were found in older sea floor than at the East Pacific Ridge, away from modern spreading centers, across the Atlantic, Pacific, and Indian Oceans (Bonatti, 1975). These deposits, several meters thick, probably formed close to a spreading center, and moved away with the lateral spreading of the oceanic crust, with the overlying sediments deposited during this movement. This indicates that processes leading to the creation of metal-rich deposits at spreading centers have been active in the past.

Estimations of the conductive heat flow from recently formed lithospheric plates have been made, using the temperature of the material when emplaced and the thickness and composition of the plate. When compared to actual sea floor measurements, values match well except for those taken near active spreading centers. Here, the measurements show a wide scatter of values, but such that not even the highest values match the estimations of conductive heat flow. This indicates that a substantial amount of the cooling is done by convection rather than by conduction. Williams (1974) used this evidence to conclude that nearly 80% of the geothermal heat released near ridge crests is by convective circulation.

Researchers have found recently formed oceanic crust to be quite permeable through a number of routes, including surface faults and fracture systems, cracks between dikes, and deep tensional fractures. This combination suggests penetration by sea water to depths of several kilometers beneath the sea floor. On the Reykjanes Peninsula of Iceland, this has been confirmed by Palmason and Semundsson (1974). Due to the height of the water column within today's oceans and seas, ample pressure exists to make sea water penetrate the young, highly fractured oceanic crust, and to prevent the water from becoming gaseous. If heated to temperatures of 200-500^{o}C, the sea water reaches pressures ranging from 1000-3000 bars (Honnorez and Kirst, 1975). Such pressures are high enough to expel the same fluids out of the crust in even the deepest parts of the ocean.

Sea water within the crust, and at a few hundred degrees centigrade, can react considerably with basalt. This has been proven experimentally by Hajash (1975) who has shown that the sea water changes from an originally oxidized and slightly alkaline solution to an acidic and reduced solution, highly enriched in Fe, Mn, Si, and Ca, in addition to other elements, as a result of reaction and leaching with the basalt. In addition, he found considerable amounts of magnesium were transferred from the seawater into the basalt.

Additional evidence of hydrothermal processes has been obtained from basalts of the sea floor. DSDP cores from Hole 319A and Site 321 contain secondary, vein-filling phases containing sulfides, iron-rich clays and oxides, and carbonates. Petrochemical and chemical studies have identified them as hydrothermal deposits (Scott and Swanson, 1976). Similar minerals have been formed by reacting tholeitic ocean basalt with sea water at 500°C and 0.6 kbars (Hajash, 1975).

Ophiolite Complexes

Large deposits of copper, iron, lead and zinc have long been known to occur in the Apennines, Cyprus, and Newfoundland. The deposits consist of lens-shaped mineralizations, up to 100m across, in between pillow basalts and dispersed in stock work fashion within dikes immediately underlying the basalts.

Studies of the rocks revealed a sequence, running from bottom to top, of ultramafics, gabbroic cumulates, gabbroic with dikes gradually increasing, pillow basalts, and sediments (see Figure 2). Further work into their seismic properties, chemistry, and structure revealed that they were segments of oceanic lithosphere that had been transported above sea level. The recognition of these sequences as ophiolite complexes, when coupled with knowledge of hydrothermal activity within the sea floor, explained the origin of these bodies. Hydrothermal fluids circulating through the crust possess large quantities of sulfides, partly through reduction of sulfates contained within "normal" sea water, and partly through leaching of sulfur from the basalt. These ions tend to combine with the metals in solution forming less mobile metal sulfides which are subsequently precipitated upon discharge from the vents. Apparently, this sulfide mobilization occurs while the pillow basalts are still forming during the early stages of volcanism, for such mineralizations ofter show a stratigraphic distribution within the pillow basalts (Upadhyay and Strong, 1973). The stockwork mineralizations are thought to be the result of later remobilization and emplacement along fault zones (Upadhyay and Strong, 1975).

Therefore, two forms of mineralization are associated with hydrothermal activity in oceanic lithosphere, one type being sulfide metals within and immediately below pillow basalts, the other being metal oxides deposited at the base of the sedimentary column, immediately overlying pillow basalts. The oxygen content of the seawater and the rate of burial probably are major factors controlling the types of deposits.

The Geologic Setting of Newfoundland

The structural province consisting of the island of Newfoundland has been divided into 9 tectonic zones (see Figure 3), designated A to H, with the boundaries of each zone determined by major faults and structural discontinuities (Williams, Kennedy, and Neale, 1972). Two of the zones, A and H, on the west and east sides of the island, respectively, contain rocks considered "continental" in origin, that is, they are composed of Precambrian granitic and gneissic materials, overlain by Cambrian and Ordovician platformal sequences. The Western Platform (A) and the Avalon Platform (H) are separated by a mountain belt (B-G). In addition to metamorphosic sediments, silicic volcanic rocks, and granitic intrusions, segments of this belt contain ultramafic to mafic sequences now recognized as ophiolite complexes (Norman and Strong, 1975, Upadhyay, Dewey, and Neale, 1971).

The materials I worked with come from zones C, D, and E, on the western side of the mountain belt. Specifically, they come from the Burlington Peninsula and the southwest side of Notre Dame Bay (Figure 3). Of concern are three ophiolite complexes recognized in this area, named after the areas in which they occur: Baie Verte, Ming's Bight, and Bett's Cove (see Figure 4), (Upadhyay and Strong, 1973). Two other ophiolites occur within the zones studied, but they were not directly associated with any of the samples I worked with. The Ming's Bight ophiolite is usually considered to be an extension of the Baie Verte.

Church and Stevens (1971) interpreted the Bett's Cove-Baie Verte ophiolites as representing one continuous sheet of oceanic crust prior to Devonian folding and faulting. Their view is that the ophiolites were moved from the east onto a sequence of late Proterozoic to Lower Paleozoic clastic sediments (Fleur de Lys Supergroup). They believe this occurred during the closing of an ocean basin due to underthrusting of the North American continent along a subduction zone dipping to the southeast.

Dewey and Bird (1971) studied the same ophiolites and proposed that western Newfoundland began to evolve through continental rifting and generation of a proto-Atlantic crust in the Precambrian. They believe that in the Cambrian or early Ordovician, a marginal basin formed to the west of the Fleur de Lys terrain (this material is being used as a "fixed" reference point). In this area the Bett's Cove and Baie Verte materials formed in inter-arc basins. The marginal basin began to contract in the early Ordovician, and segments of the crust were tectonically emplaced as ophiolites during a continental collision in middle Ordovician times.

Kennedy (1973) studied the Burlington Peninsula and concluded that the ophiolites present represented oceanic lithosphere of at least two different ages. He distinguished between the two ophiolites on the sides of Baie Verte by naming the one on the west side the "Advocate Group". In this group, he recognized what he interpreted as pre-Ordovician structures and concluded that they formed in a small ocean basin in pre-Ordovician times. Westward subduction of oceanic lithosphere (to the east of the Burlington Peninsula) resulted in island arc volcanism. In the early Ordovician, continued sea floor spreading accompanied obduction of the oceanic lithosphere, emplacing the Bett's Cove complex. Kennedy agrees with Church and Stevens (1971) in believing that the Baie Verte group originated to the east of the Burlington Peninsula, and is allochthonous.

The different stratigraphic interpretations of the above authors are shown in Figure 5. Disagreements still exist as to the modes of formation, the methods of emplacement, and the timing of events.

Methods

The rocks that I worked with on this project were collected on the island of Newfoundland by Robert B. Scott in 1974. The samples were taken because they were at or/near the base of the sediment column, within ophiolite complexes or highly similar surroundings. In addition, colorations or recognizable minerals suggestive of metalliferous concentrations were used as sampling criteria. The six rocks which I studied were chosen from this collection.

My first work with the samples was to study the samples macroscopically, trying to get an idea of the minerals present and their relationships within the samples. I then studied the rocks extensively with an X-Ray Diffractometer in order to learn the types and locations of the major minerals present within the rock. Slabs of rock were then cut from the most promising samples and sent off to be ground down into thin sections.

While the thin sections were being processed, I worked on sampling specific locations within the rocks which showed concentrations of metal-rich materials. For example, some of the rocks were stratified and I wasn't interested in materials which suggested histories unrelated to the metalliferous deposits. These samples I then proceeded to put into solution for elemental analysis by Atomic Absorption Spectroscopy. Four samples from the same location within one rock were used for statistical purposes. For calibration, I used three USGS standards: BCR-1, a basalt; DTS-1, a dunite, and MAG-1, a marine mud from the Gulf of Maine. Although the overall chemical compositions of these standards were unlike those of my samples, they did have the high metal concentrations that I needed.

To get these rocks into solution, one has to break the atomic bonds. This was done by reacting 0.5 gram samples with hydrofluoric acid, hydrazine sulfate (for organic matter), and nitric acid. Two standards and one sample didn't go completely into solution after this process, but did go sufficiently into solution after being treated with aqua regia (hydrochloric and nitric acid). Later, during the analysis, one of the standards showed values inconsistent with the other two. Since this standard, a dunite, possessed several of the high metal values I needed, I substituted it with a standard already in solution, W-1 (diabase), which correlated with the values of the other standards. Dilutions for the standards were 1:200, 1:2000 (2M), 5M, 10M, and 20M. Sample dilutions were 1:200 and 10M. A few additional dilutions of the samples were necessary to successfully bracket values with those of the standards. Results of the AA work are given in Table 2.

The thin sections, when returned, were then studied with a petrographic microscope in an attempt to identify all of the minerals present within each sample which underwent analysis by AA. Also, I sought to determine the histories of the rocks by studying textures and relationships among the minerals present.

It might be added that extensive reading was also necessary, concerning the history of Newfoundland, ore deposits, hydrothermal and other forms of metal enrichment, and on the origin and identification of minerals, in addition to other topics.

Metal Deposits From Tilt Cove

Two of the rocks that I worked with came from Tilt Cove, situated on the east side of the Burlington Peninsula (Figure 4). Tilt Cove represents the northern end of an ophiolite assemblage often referred to as the Betts Cove Ophiolite Complex (Upadhyay, Dewey and Neale, 1971). The Snooks Arm Group, considered to be lower Ordovician in age, contains the Betts Cove Ophiolite within its base, and consists (from bottom to top) of ultramafics, gabbro, sheeted dikes, pillow lava, volcano-clastic sediments, pillow lava, volcano-clastic sediments, and a final layer of pillow lava (Upadhyay and Strong, 1973).

In the area of the contact between the sheeted dikes and pillow lavas, numerous types of sulfides are found, including pyrite, chalcopyrite, and sphaelerite. The assemblage evidences stratigraphic control and sedimentary features which, along with the mineralogy, reveals a strong resemblance to Troodos-type sulfide mineralizations. On the origin of this body, strong and Upadhyay (1974), say "the mineralization resulted primarily from syngenetic precipitation of sulfides from upwelling geothermal brines during hiatuses of volcanism in an ocean ridge-type environment." So at least while the pillow basalts were forming, metallic deposits were being laid down.

The two samples 20A and 20B were collected at the contact between the lower-most pillow basalts and overlying sediments. They were collected because of their bright red color (similar to that of jasper), which turned out to be due to abundant iron oxide in the rock. This layer of material was quite distinct from the overlying volcanoclastic sediments.

X-ray analysis of the two revealed calcite, ankerite, hematite, dolomite, quartz, and chlorite (not in order of abundance) within both of the rocks. Atomic absorption data (Table 2) revealed that both contained substantial amounts of iron, with lesser amounts of manganese. The concentrations indicated that the rocks had undergone some form of metal enrichment. When the distributions were plotted on a triangle diagram similar to that of Figure 1, the points from these samples (and the rest of the samples analyzed) all fell within the range designated as "hyrdothermal."

Thin section study of sample 20B indicated that the rock was a coarsely crystalline (.75mm-1mm) limestone containing abundant hematite. As the photograph in Figure 7 indicates, the hematite is concentrated in an arrangement suggestive of the directions of cleavage of a calcite crystal. In patches resembling filled voids, fine-grained quartz surrounds euhedral crystals of dolomite. This relationship can be seen in the photograph in Figure 8. The patches are dissected by numerous veins containing calcite and hematite. In places, penninite (a chlorite) occurs in vein-like form.

The concentration of hematite within the calcite varies, some areas have only minor amounts, and other areas have so much that the hematite forms short irregular bands. Convoluted layering can be seen within the limestone, but these layers rarely contain bands of hematite. The irregular distribution of hematite and its lack of correspondence with apparent sedimentary layers indicates that the hematite formed within the rock after the calcite was in place. Apparently the hematite entered by way of solutions passing through the rock. A possible mode for this is described by Krauskopf (1957) who describes the precipitation of iron oxide from originally slightly acidic iron-rich fluids after their reaction with limestone, which converts the fluids into alkaline ones (with lower solubilities for iron and manganese). Creation of such acidic metal-rich solutions by reaction of hot sea water with basalt has been done experimentally (Hajash, 1975). Oxidizing conditions could also have been responsible for the enrichment, in the absence of the limestone dissolving and altering the fluid chemistry.

Further indication of hydrothermal solutions is indicated by what appears to be filled voids, which contain quartz and dolomite. The quartz is very fine-grained (.01-.03mm) which indicates it was deposited, with the dolomite, by fluids. I wasn't able to distinguish between dolomite and ankerite, but the X-ray data indicated both were present within the rock.

Ankerite occurs within sedimentary rocks as a result of both hydrothermal and low temperature metasomatism (Deer, Howie, and Zossman, 1962). Dolomite is known to occur in even more types of environments. Fractures containing hematite and calcite run through the filled voids, and can be seen cutting through the large crystals of dolomite. The fracturing and remobilization probably occurred during emplacement of the ophiolite, at this time the layers seen within the limestone probably underwent their convolution.

Penninite is a silica and magnesium-rich chlorite, and was distinguished from other possible chlorites on the basis of its "anomalous-blue" interference colors in thin section, in addition to its other optical properties. The thin section showed the penninite occurring in vein-like form cutting through the limestone in isolated areas of the rock. The differences in form and distribution between the chlorite and hematite suggest that the two minerals formed at different times, for the two are often unassociated within the rock. Chlorites are found in low-grade regionally metamorphosed rocks, and characterize the greenschist facies. This form of metamorphism would have occurred during the emplacement of the ophiolite suite. This evidence suggests that the hydrothermal activity leading to the deposition of iron occurred while the igneous rocks were cooling by convection.

It might be mentioned that rocks which host hydrothermal processes are affected by metamorphism which ranges from the zeolite to greenschist facies (Bonatti, 1975). However the differences in form and distribution between the hematite and penninite suggests that this was not the way in which the penninite formed.

Sample 20A is also a hematite-rich limestone. It possesses a higher iron content than does 20B (see Table 2). In hand specimen, no convoluted layers are present, unlike 20B. Further differences appeared in thin-section study. The hematite is dispersed throughout the limestone and is not concentrated into forms within the crystals of calcite, whose outline is masked by the abundant hematite. The arrangement of the hematite is rather mottled, some areas are quite dark, while the surrounding rock is a strong red. This uneven, yet thorough, arrangement indicates penetration by hydrothermal fluids depositing iron by the same methods outlined for 20B. If the hydrothermal fluids (which appear to be the only possible source of the iron) were discharging the metals into sea water where they subsequently precipitated, I would expect to see more evidence of lamination within the rock.

This specimen contains two types of what appear to be filled voids. One of these forms is exactly identical to the filled voids in 20B. The other type contains no silica, and is shown in the photograph in Figure 9. It contains euhedral dolomite crystals (again it could be ankerite) separated by fine-grained hematite within a matrix of calcite or dolomite (very fine-grained). The hematite may have entered the voids at the same time as it was deposited within the rest of the rock.

Numerous fractures, often at right angles to one another, cut across the rock. By studying the cross-cutting relationships, the oldest veins consist of hematite and calcite. Cutting this are veins containing chlorite (penninite) and calcite, with no hematite. Cutting this is a vein containing dolomite on the walls, with quartz and chlorite lying in between, the quartz crystals oriented at right angles to the vein. From this, one might infer that the hematite entered the rock before the silica and chlorite. I can't explain why the voids containing quartz possess no chlorite, unless both types of void filling, and the veins, occurred before metamorphism formed the chlorite. The penninite may have formed from materials contained within the vein, with magnesium coming from the dolomite.

In a brief summary, these two rocks evidence hydrothermal enrichment taking place prior to the low-grade regional metamorphism which accompanied emplacement of the ophiolite. In addition, their location immediately above the pillow basalts indicates that they formed while the lithospheric material was still relatively young, for no overlying sediments were observed to show similar mineralizations.

Metal Deposits from Ming's Bight

Three of the rocks that I worked with are from Ming's Bight, a small peninsula on the northern end of the Burlington Peninsula. This smaller peninsula lies between two bays, Baie Verte and Ming's Bight. Norman and Strong (1975) have interpreted the Ming's Bight area as an ophiolite assemblage which ranges from ultramafics at the bottom to sediments overlying pillow basalts at the top. The assemblage is exposed by five distinct structural blocks within the peninsula, separated by fault zones containing either serpentinized peridotite or talc-carbonate materials. Further differences between the blocks occur in the number of units present, their orientations, and physical or chemical differences between corresponding units. Materials exposed on this peninsula are referred to as the Baie Verte Group (Norman and Strong, 1975).

Workers studying these rocks tend to agree that the Baie Verte ophiolite sheet was emplaced during the early Ordovician, and later folded into a tight syncline during the Acadian Orogeny (Middle Devonian) (Church and Stevens, 1971). Kennedy (1973) also interpreted deformed and metamorphosed ultramafic rocks to the west of the peninsula as pre-Baie Verte ophiolites. Church and Stevens (1971) and Dewey and Bird (1971) studied igneous and metamorphic lithologic similarities between the east and west sides of the bay, and looked at islands in the bay between the two (undeformed mafic and gabbroic rocks), before concluding that the western ultramafic belt represents the base of the Baie Verte Group. This belt of ophiolites overlies Fleur de Lys metasediments. Garnetiferous amphibolites of this area have been interpreted as a dynamothermal aureole resulting from thrusting (Church and Riccio, 1974). The rocks of Ming's Bight are considered to have been sufficiently distant from the thrusting zones so as to have escaped much of the deformation during emplacement (Norman and Strong, 1975).

Ming's Block

Sample 28 came from Ming's Block (Figure 10) along the shore of Ming's Bight. This block contains several elements of an ophiolite, except that the gabbro unit is in fault contact with pillow basalts. Sediments of the area are described as predominantly water-lain tuffs. Sample 28 was distinguished from these materials because in hand specimen it was seen to be predominantly composed of magnetite. X-ray analysis revealed magnetite, quartz, calcite, prochlorite, epidote, and hematite. AA data indicated that the rock was very rich in iron, about 38% by weight. Figure 6 suggests a hydrothermal source for the iron enrichment. In hand specimen two types of apparent cherts were seen, one red and one green, and there were indications of strong remobilization of calcite while the rock was being folded.

By thin-section study, the red chert turned out to be a mixture of fine-grained quartz and hematite, with minor amounts of magnetite. The quartz and the hematite seem to have been laid down simultaneously on the sea floor, for the iron doesn't show the mottled distribution observed in the samples from Tilt Cove, but instead shows signs of lamination.

As can be seen in the photograph in Figure 11, the rest of the rock is very rich in magnetite. This magnetite lies within crudely parallel inter-connecting layers of calcite and very fine-grained quartz. The magnetite is largely within the layers of quartz, or on the borders of these layers, whereas the calcite contains lesser amounts. Since the rock shows signs of concentrations of calcite resulting from folding during regional metamorphism, it is possible that the iron and quartz were precipitated by hydrothermal solutions, and later introduction of calcite followed planes of weakness within the rock created by the magnetite.

Laboratory experiments have indicated that magnetite can be formed as a precipitate from aqueous solutions at low temperatures and pressures. By adding alkali to a solution containing both ferrous and ferric iron, precipitates have been obtained that contain brownish-black, magnetic materials with X-ray patterns similar to those of magnetite (Weiser, 1953). Hydrous Fe_3O_4, rather than hydrous Fe_2O_3, can be formed by oxidation of ferrous hydroxide in an alkaline medium if the rate of supply of oxygen is slow (Huber, 1958), a larger supply of oxygen (higher Eh) would allow little or no magnetite to form. Differences in environment of deposition of the chert and surrounding rock may account for the two distinct types of iron oxide.

The formation of magnetite from hematite only calls for heat and a reducing atmosphere (Deer, Howie, and Zussman, 1962). If the calcite within the rock is primary, conditions may have occurred that recrystallized hematite into magnetite, with the chert representing a system already closed to such conditions so that its hematite remained unaltered.

A greenish-colored area occurred as a pod-like inclusion within the rock. A portion of it was included in the thin section, and it revealed prochlorite, epidote, calcite, and quartz. Epidote, calcite, and quartz are all known to occur within the overlying tuffaceous sediments. Prochlorite is an iron and magnesium-rich member of the greenschist facies and is often found associated with magnetite. Apparently this pod consists of metamorphosed tuffaceous sediments, which could easily have been transformed during emplacement of the ophiolite. Both this chlorite pod and the hematite-rich chert seem to have been folded into the magnetite body.

Deer Cove Block

Two samples, 37 and 40, come from the southern end of the Deer Cove Block (see Figure 10). The block consists of a thick (900m) layered gabbroic unit in fault contact with intensely deformed dikes, pillow lavas, and sediments (an additional 700m). A rough transition from dikes and lavas in the northern end of the block, to sediments in the southern end, can be discerned. Sediments in the southern end include volcanogenic tuffs, ferruginous chert, pyrite, magnetite, and chalcopyrite.

Sample 37 was a distinct green-colored schist, very distinct from overlying and underlying tuffs. At the time of sampling it was suspected that the rock contained fuchsite, a chromium-rich muscovite. X-ray study of the rock revealed quartz, dolomite, muscovite, and plagioclase. Thin-section work showed that the quartz and dolomite were present in nearly equal amounts, forming crudely parallel and interconnecting layers. Subophitic textures varied from quartz enclosing dolomite to dolomite enclosing quartz. Muscovite was optically verified and seen to form veins running through the rock in bands parallel to the other layers. The green color was quite unlike muscovite, and as the AA data in Table 2 shows, this is due to the high level of chromium within the mineral. The quartz grain sizes were quite small (.03mm), again indicating an origin by precipitation. Plagioclase occurred in medium sized crystals within a seam running through the rock, and showed a compostion of about An_{50} by the Michel-Levy law, as applied to albite twins (Kerr, 1959). Calcite (recrystallized into dolomite) may have been the original constituent of the rock.

An investigation by Whitmore (1946) into the paragenesis of chromium-bearing muscovites has shown that they occur most commonly in hydrothermal replacement zones, associated with quartz and gold deposits. I have yellow opaques within the rock, but I wasn't able to identify them as gold (further work will be done). The chromium is probably ultimately derived from leaching of chromite-rich pods which occur in the ultramafics of this area (Strong, 1974).

Albite is known as a relatively common, low-grade metamorphic product, but intermediate compositions such as An_{50} are rarer. Possibly the calcium present within the dolomite affected the crystallization of the plagioclase, which was the result of the regional metamorphism of the area. This metamorphism was probably also responsible for the crude layering of materials within the rock.

Sample 40 was also collected in the southern end of the Deer Cove Block. It was seen within 10' long boudins lying within tuffaceous sediments, 10-15' from underlying pillow basalts. In hand specimen, one could see pyrite, chalcopyrite, magnetite, and dense black patches. X-ray study showed stilpnomelane (a brittle mica), calcite, quartz, and magnetite within the dark patches. AA revealed about 20% iron and a bit more copper than in the other samples (Table 2).

Thin-section study showed that stilpnomelane was the dominant mineral, forming laths oriented perpendicular to the direction of the veins they were within. The veins passed between blocks of fine-grained materials, which upon inspection appeared to be additional stilpnomelane associated with epidote and small grains of quartz. Some of these blocks contained opaques outlining ellipsoidal bodies enclosing numerous grains of stilpnomelane. As if this wasn't enough, large .75mm non-oriented laths of stilpnomelane occurred with calcite and clinozoisite in areas unassociated with either blocks or veins. Apparently, favorable conditions existed for the development of stilpnomelane.

Stilpnomelane is a rather widespread mineral which often occurs in iron-rich materials, and is considered to be a product of low-grade regional metamorphism. In the lake Superior region, James (1954) describes very fine-grained stilpnomelane with rocks that still retain their clastic texture. It is possible that the outlined ellipsoidal bodies are the same sort of feature. Also, stilpnomelane has been described as an accessory mineral in some rocks which indicate that the original constitution of the host rock is not the main factor governing the occurrence of the mineral during periods of regional metamorphism (Deer et al, 1962). Hutton (1938) has described stilpnomelane which formed after the main phase of metamorphism during a period of post-tectonic recrystallization. Such behavior would help to explain why there are so many forms of the same mineral within one rock.

Grain sizes of magnetite are uniform throughout the rock, irregardless of surroundings. Perhaps it formed with pyrite and chalcopyrite before the stilpnomelane began to form. Question does exist as to how and when the iron, 20% of the rock by weight, entered the rock. The processes of recrystallization have pretty well removed any grounds for speculation.

The occurrences of clinozoisite and epidote may also be attributed to regional metamorphism.

In a brief summary, all three rocks from this peninsula possess evidence of metal enrichment, two of them by sources directly attributable to hydrothermal activity. It also appears that this activity took place prior to the low-grade regional metamorphism of the area (which is not known for metal enrichment).

Metal Deposits From Loon Bay

The last of the six samples I studied comes from Loon Bay, located on the southwest side of Notre Dame Bay. Although the sediments do not overlie materials presently recognized as ophiolitic, they do overlie pillow basalts, which indicates sea floor deposition and possibly similar alteration processes. Overlying the pillow basalts are volcano elastic materials, which underlie shales (some pyritiferous), banded cherts, pink cherts, and calcareous layers. Sample 44 resembled a pink chert, but it was weathering to a black color which suggested it might be rich in manganese.

Thin-section work revealed a boudinized layer of rhodonite, (Mn, Fe, Ca) SiO_3, surrounded on both sides by fine-grained layers of quartz. The rhodonite layer showed a poikilitic texture around quartz crystals (.02mm), which at times reached proportions of 50-60%. Some calcite was also seen within the rock, either as vein-filling or as large crystals within the rhodonite layer. The photograph in Figure 12 shows a vein passing through the rhodonite layer, which apparently enabled the rhodonite to recrystallize and get rid of its content of quartz. The quartz layers surrounding the central layer suggest an origin as precipitates because of their .03 mm grain size. These layers contain abundant epidote and scattered actinolite needles (Figure 13).

One possibility for the rhodonite is that it formed as a radiolarian chert, which has a tendency to recrystallize. If the material were to encounter waters rich in manganese, this could cause recrystallization as rhodonite. An implication of hydrothermal activity must first be qualified. The leaching of metals from basalt is known to result in the solution of considerable amounts of iron and manganese. Iron is known for its tendency to precipitate before manganese, yet the rhodonite has a ratio of Mn:Fe of 20:1 (Table 2). There may have been an Eh gradient in the waters such that by the time the Hydrothermal fluids had reached the layer of concern, the iron had already been oxidized and precipitated. Or increasing alkalinity as the fluids moved towards the sea floor (since sea water is alkaline) may have had a similar effect. Either way, if the hydrothermal fluids were derived from basic, igneous rocks, considerable amounts of iron must have been precipitated. Why the surrounding layers of quartz didn't alter into rhodonite makes the matter more complicated. Actionolite and epidote have been described as results of hydrothermal activity in experimental reactions (Hajash, 1975). They're also known as products of low-grade regional metamorphism. Subtle chemical differences between the materials involved may also have been responsible for the difference in alteration.

Rhodonite has also been described as a metasomatic replacement of rhodochrosite by circulating hydrothermal solutions which bring in SiO_2 (Deer et al, 1962). Such a mechanism would also explain the abundant SiO_2 within the layer. This brings up the question of where the rhodochrosite ($MnCO_3$) came from. Rhodochrosite-bearing ore deposits forming replacement bodies have been described in limestone, and attributed to warm meteoric waters (Ham and Oakes, 1944). Syngenetic rhodochrosite deposits have also been described. If either of these methods were associated with hydrothermal activity, the same arguments for the absence of iron would apply.

Summary

Thin section work performed on the samples indicates that all six rocks were originally sediments before undergoing metamorphism. As described in this paper and shown in Table 2, the rocks also showed signs of having been enriched in metals. Both the metal distribution (Figure 6) and the mineralogy indicate a hydrothermal source for the enrichment. Two of the rocks (20A and 20B) offer evidence that the hydrothermal metamorphism did not occur simultaneously with the low-grade regional metamorphism which accompanied the emplacement of the ophiolites.

Instead, it seems likely that the enrichment occurred by processes identical to those responsible for the metalliferous deposits found in active spreading centers. Such a hypothesis is supported by the similarities in metal distribution (compare Figure 6 with Figure 1). Also, five of the rocks occur in known ophiolite complexes and the sixth rock occurs in a highly similar geologic environment. In addition, all six rocks lie at the base of the sediment column, similar to the distribution indicated by DSDP cores.

Although these deposits are not presently economically feasible as a source of metals for industry, their occurrence appears to be widespread both in the present and in the past. Therefore, such deposits should be considered in the study of other ophiolites.

One item worth considering is that the samples I studied are in areas where deposits of economically important metal sulfides are also known. They are located in a different stratigraphic position, but are considered to be of identical origin. This implies that where one finds deposits similar to the samples I studied, one might also expect to find valuable metal sulfide deposits in a stratigraphically lower position. This may someday be confirmed by DSDP cores.

Table 1 Composition of samples from metalliferous hydrothermal deposits. The first two are discharges from springs in volcanic areas, the rest are from the vicinity of spreading centers. Sources: Matupi Harbor, Ferguson and Lambert 1972; Thera, Bonatti and Joensuu 1972; EPR, Bostrom et al 1969; EPR (amph D2), Bonatti et al 1966; Bauer Deep, Sayles and Bischoff 1973; MAR 26°N, Scott et al 1974; Afar, Bonatti et al 1972.

	PER CENT				PPM			
	Si	Al	Fe	Mn	Ni	Co	Cu	Zn
THERA	11.6	1.2	35.0	0.6	<5	<5	30	—
MATUPI HARBOR	—	—	44.0	0.034	—	—	47	52
E. PACIFIC RIDGE	6.1	0.5	18.0	6.0	430	105	730	380
E. PAC. RIDGE (AMPH D2)	8.2	0.5	32.5	1.94	400	35	74	—
BAUER DEEP	18.0	1.4	18.2	5.7	950	90	1,100	600
MAR. 26°N	—	—	.01	39.2	100	18	12	—
D.S.D.P. (E. PACIFIC)	—	—	17.5	4.5	535	83	917	358
AFAR (Fe-RICH)	14.0	3.7	29.0	0.15	18	17	11	—
AFAR (Mn-RICH)	<0.2	0.0	0.15	54.2	<10	<5	<5	—

	%		PPM						
	FeO	MnO	Fe	Mn	Cu	Ni	Cr	Co	Zn
20A	13.0	1.25	101,000	9,670	88	94	44	25	132
20B	10.1	1.33	78,470	10,300	8	58	14	18	162
28	48.5	0.57	377,000	4,410	34	45	35	34	45
37	2.1	.13	16,300	1,010	5	144	1,767	28	15
40	25.1	.73	195,000	5,650	400	22	31	25	36
44	.95	21.2	7,380	164,100	44	17	0	36	52

Table 2 Metal abundances of the samples as determined by Atomic Absorption.

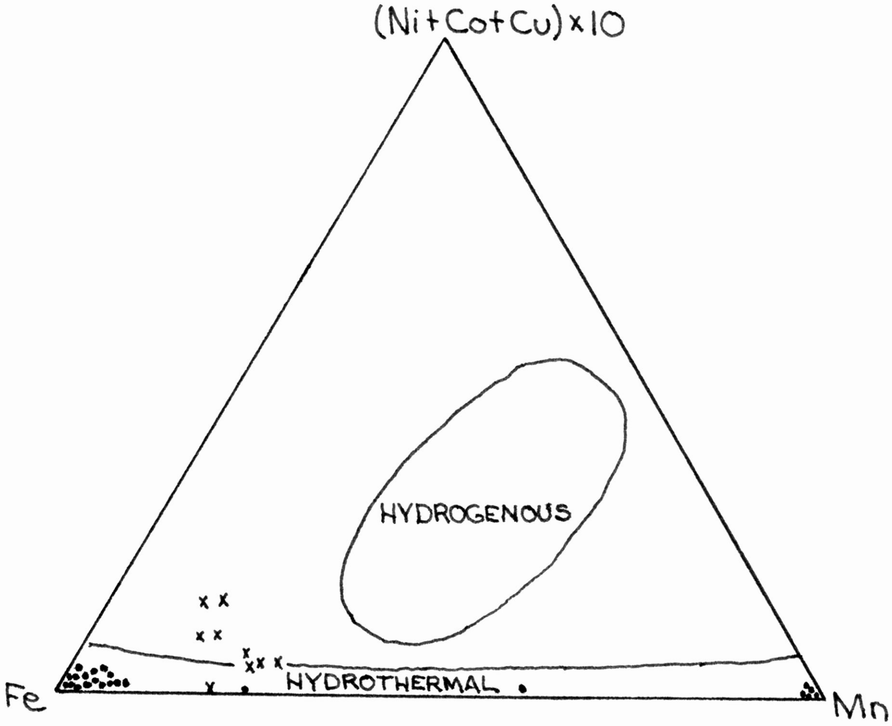

Figure 1. Ratio Fe : Mn : (Cu + Ni + Co) in metalliferous sediments from the sea floor, modified from Bonatti, 1975.

● Hydrothermal deposits from Thera, Stromboli, Afar, and seamount Amph D2.

× Fe-rich sediments from the East Pacific Rise.

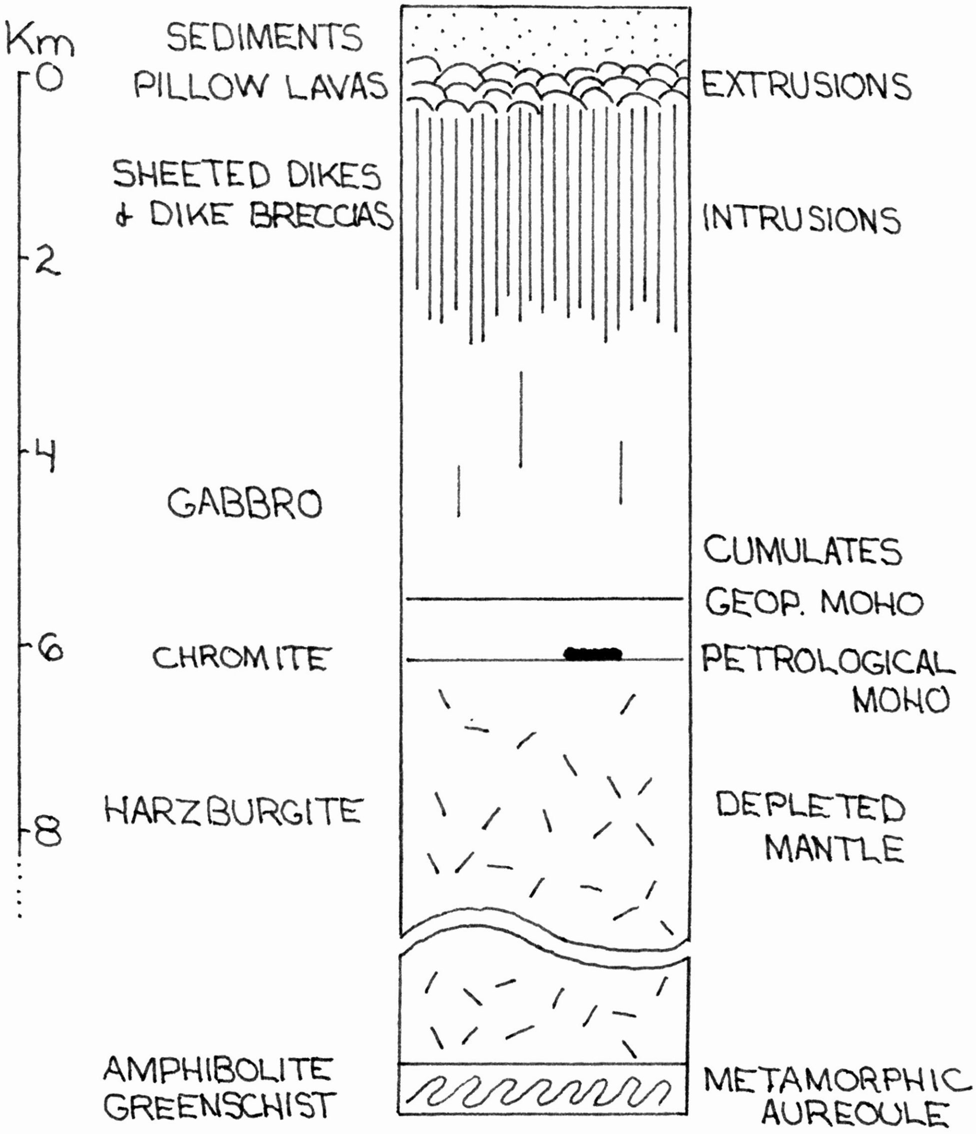

Figure 2. Vertical cross-section of an ophiolite complex from Newfoundland.

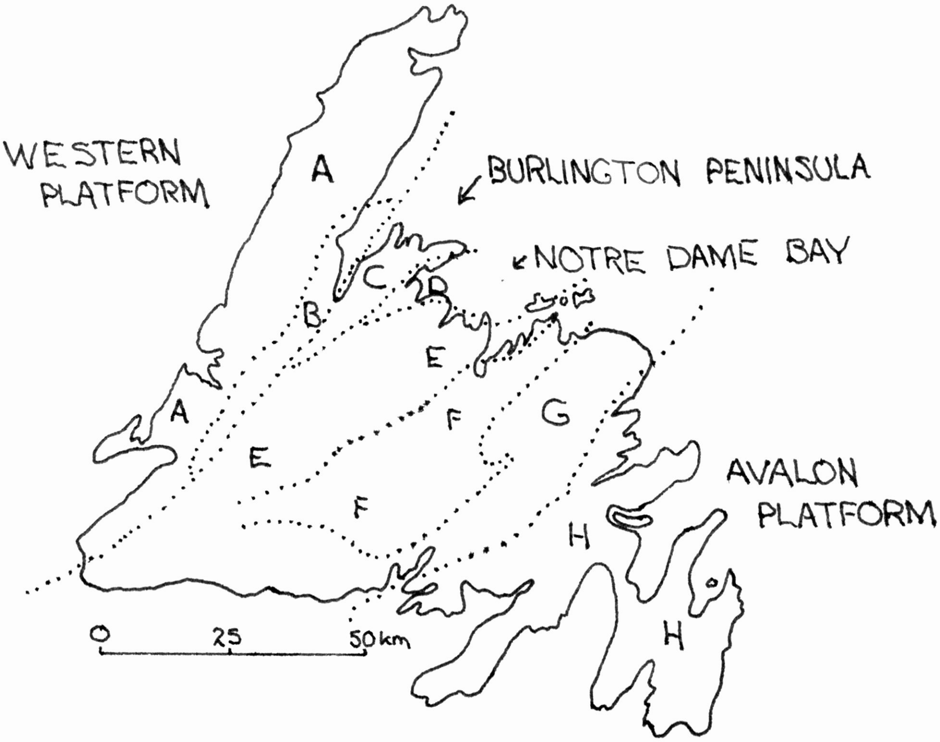

Figure 3. Newfoundland, showing tectono-stratigraphic zones.

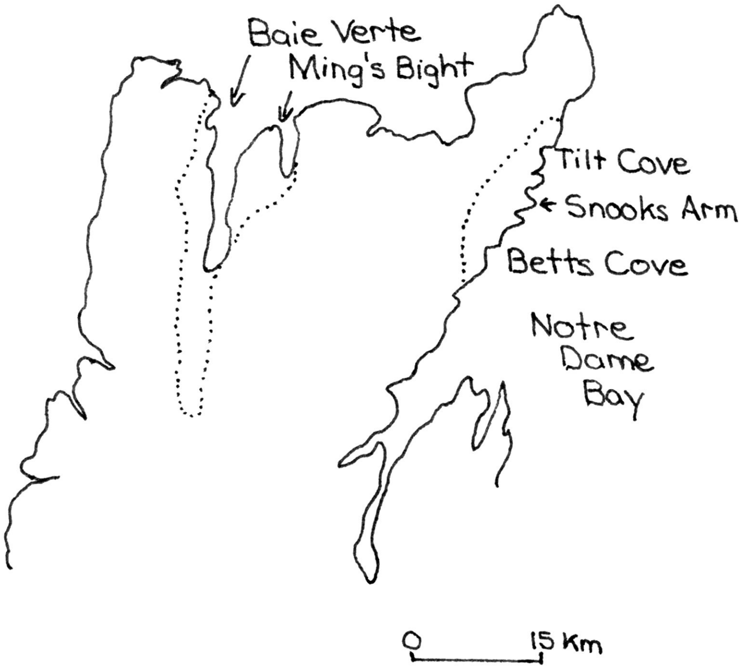

Figure 4. Burlington Peninsula, showing the Betts Cove, Baie Verte, and Ming's Bight ophiolites.

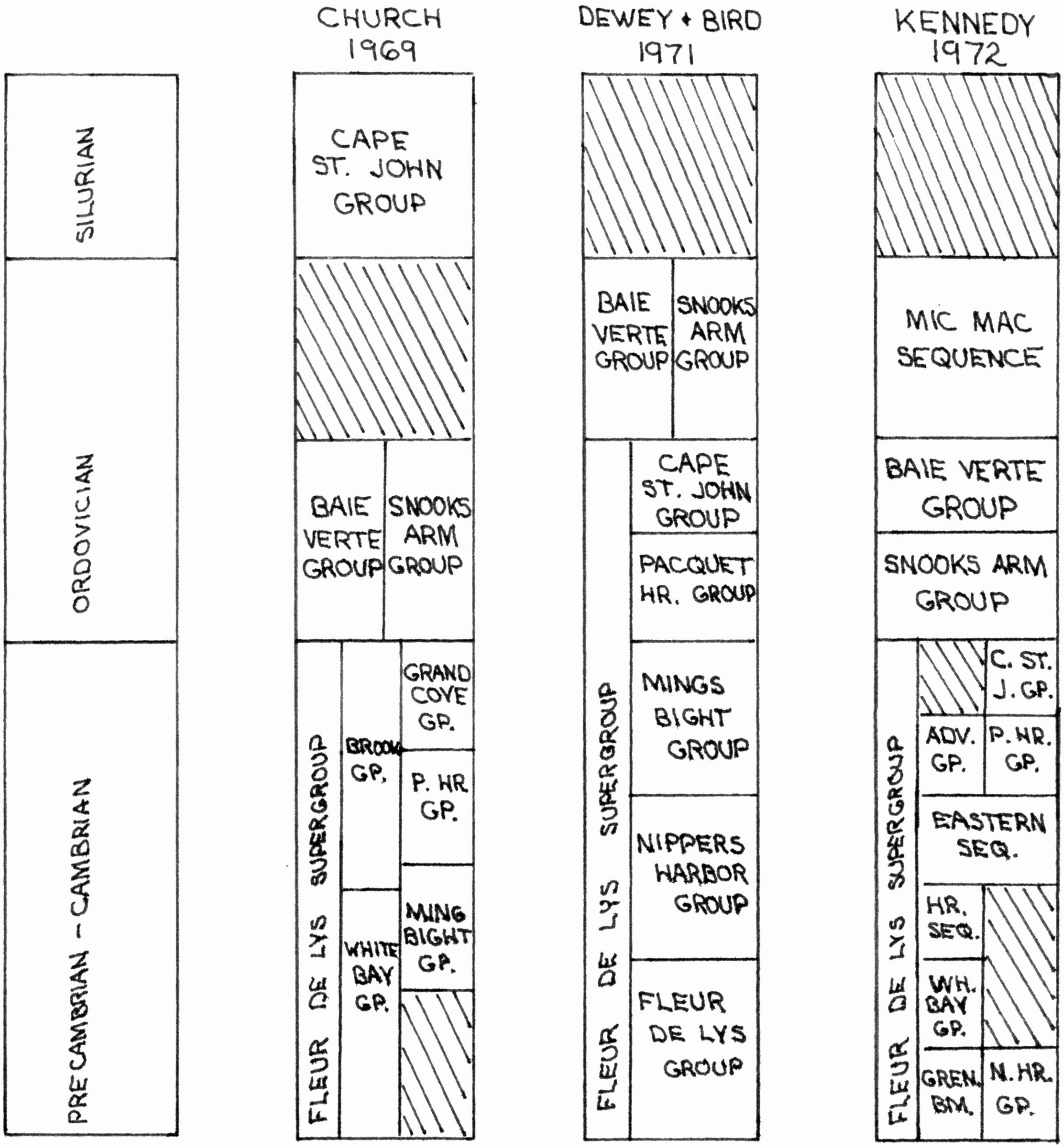

Figure 5 Stratigraphic interpretations of the Burlington Peninsula, after Norman and Strong, 1975.

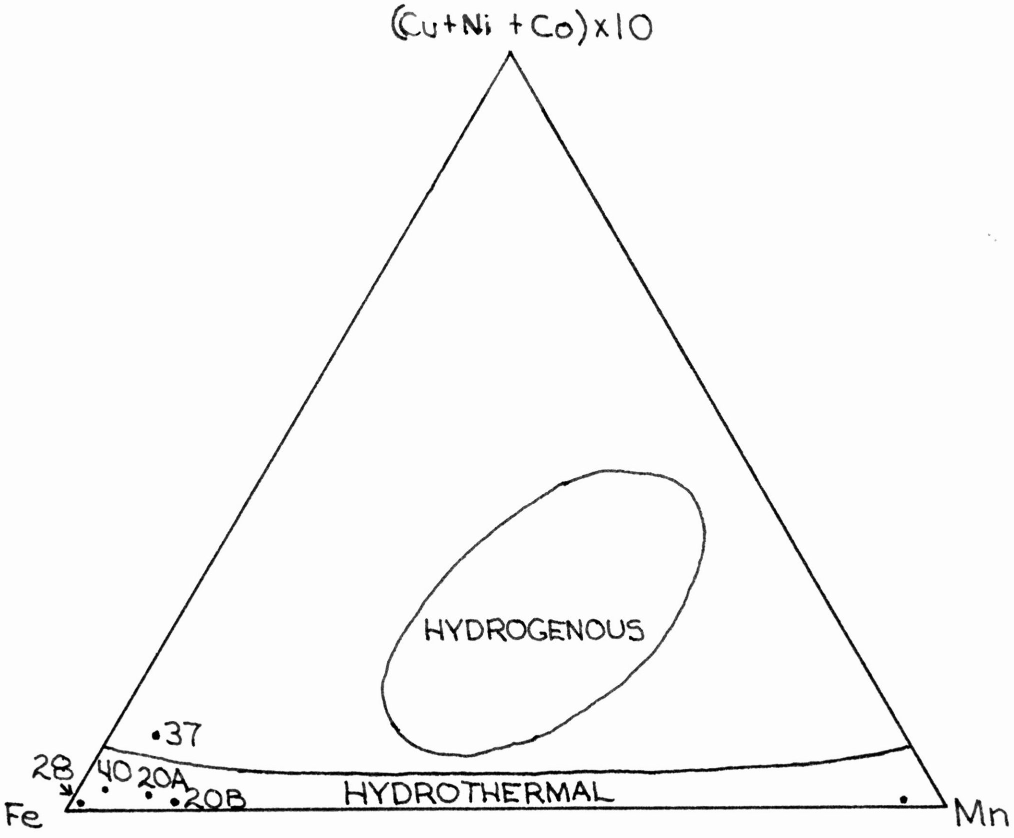

Figure 6. Ratio Fe : Mn : (Cu + Ni + Co) from the samples analyzed in my study. Data from AA analysis. Note the resemblance to the hydrothermal deposits shown in Figure 1. Hydrogenous and hydrothermal ranges are from Bonatti, 1975.

Figure 7. Photograph showing the distribution of hematite within calcite, from sample 20B. Black material is hematite, white is calcite, as is the gray material (crossed Nicols). Note how the hematite in the center of the photograph appears to be distributed within cleavage planes of calcite.

Figure 8. Photograph within an apparent filled void contained within sample 20B. The euhedral crystal is dolomite (or ankerite). The fine-grained material is quartz (crossed Nicols). Dense bands on the margins of the picture are veins containing calcite and hematite, which cross-cut the materials filling the void.

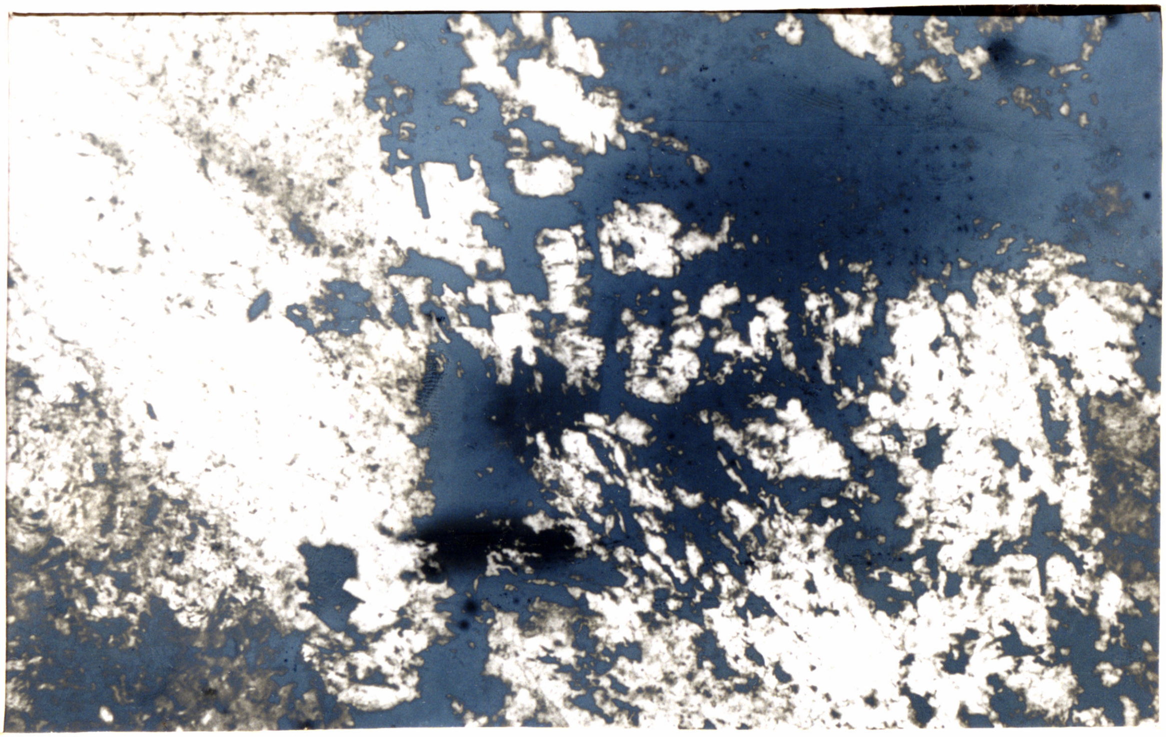

Fig 7

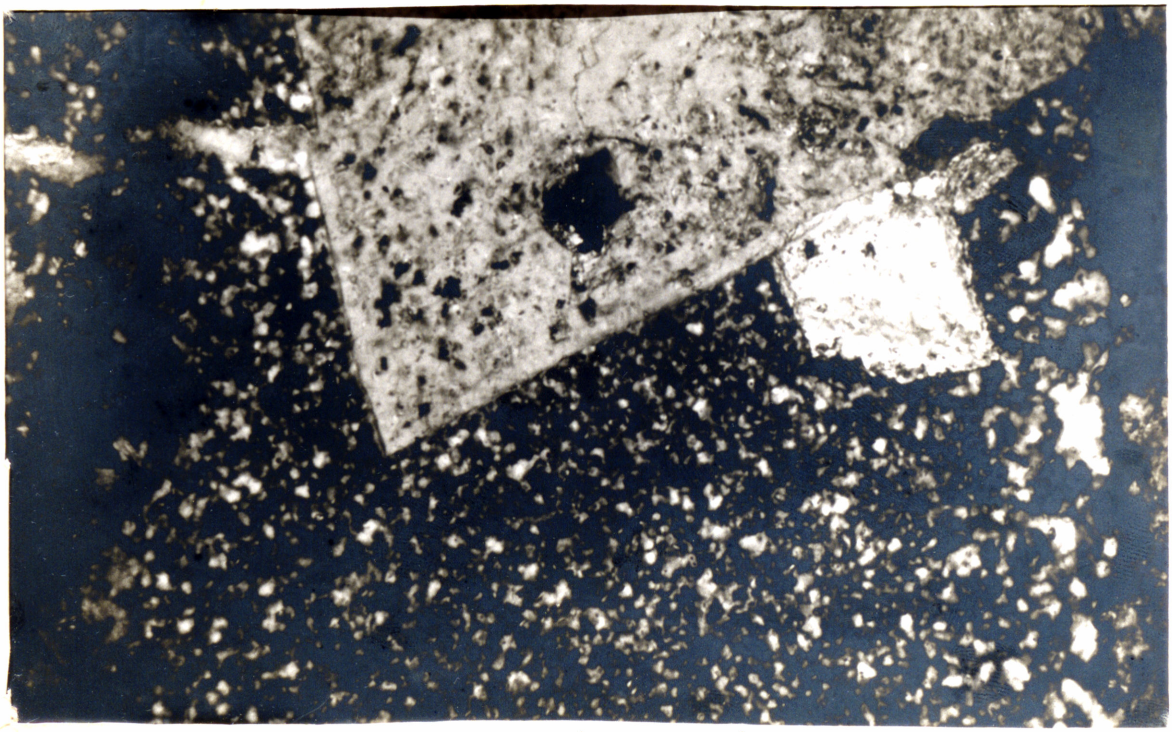

Fig 8

Figure 9. Photograph of an apparent filled void within sample 20A. White material is dolomite (or ankerite). Black material is hematite. The black material surrounding the void is hematite contained within calcite, and is typical of the distribution of hematite within the entire sample. Rounded white patch is another filled void containing calcite only. Veins seen on the right side contain calcite. Note the differences in the distribution of hematite between this sample and 20B (Figure 7).

Figure 11. Photograph of the distribution of magnetite within sample 28. Black material is magnetite. Background material is quartz and calcite. Unfortunately, this photograph doesn't clearly show the layering I described as a characteristic of the rock.

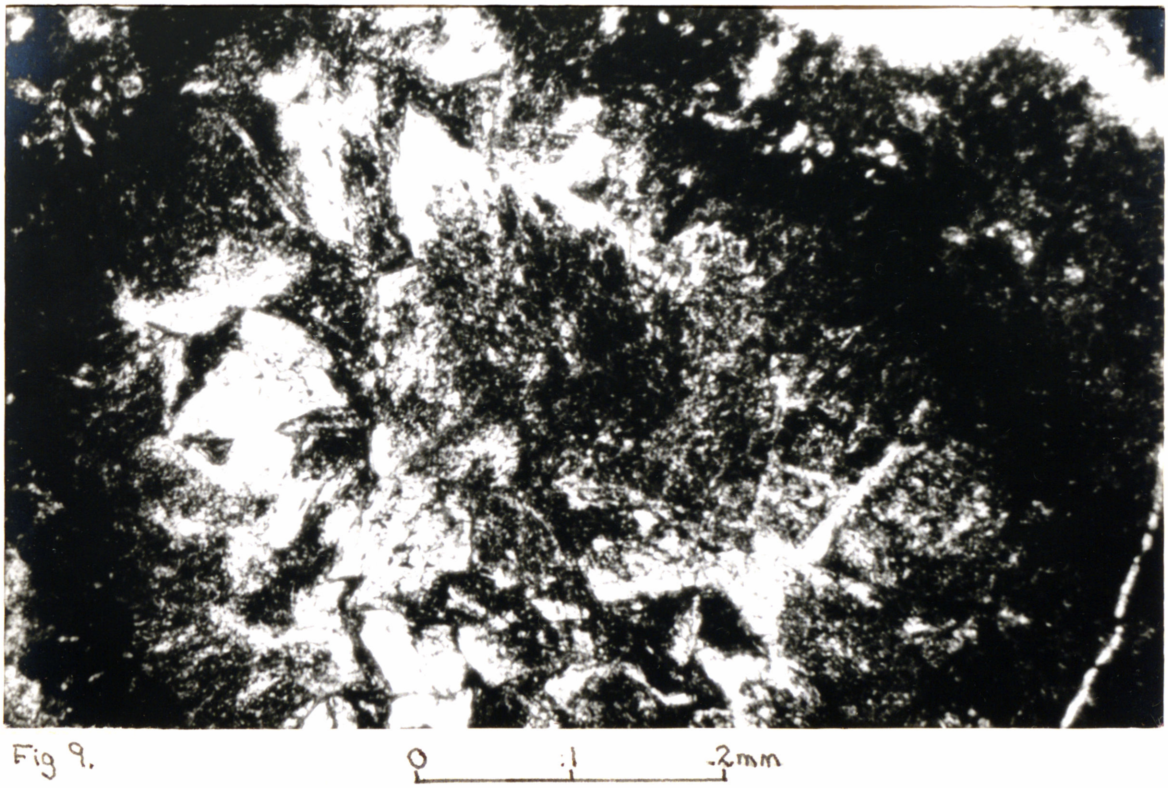

Fig 9.

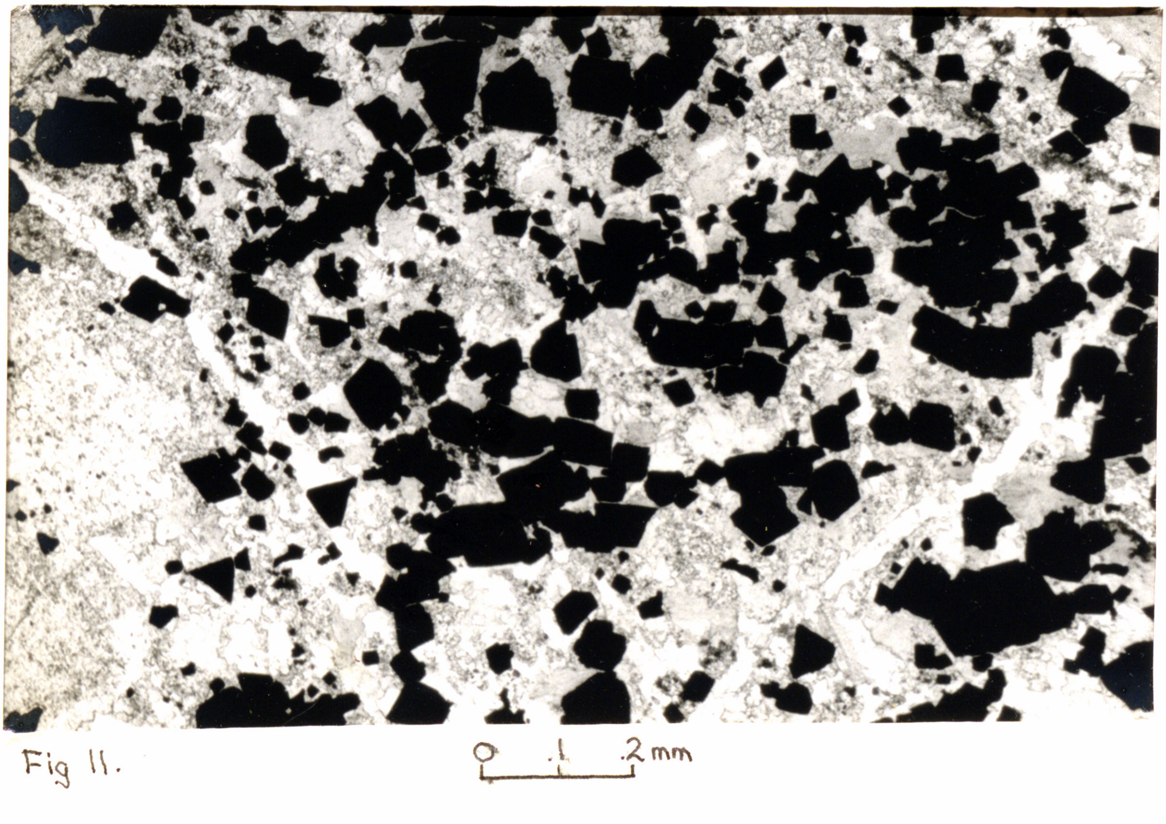

Fig 11.

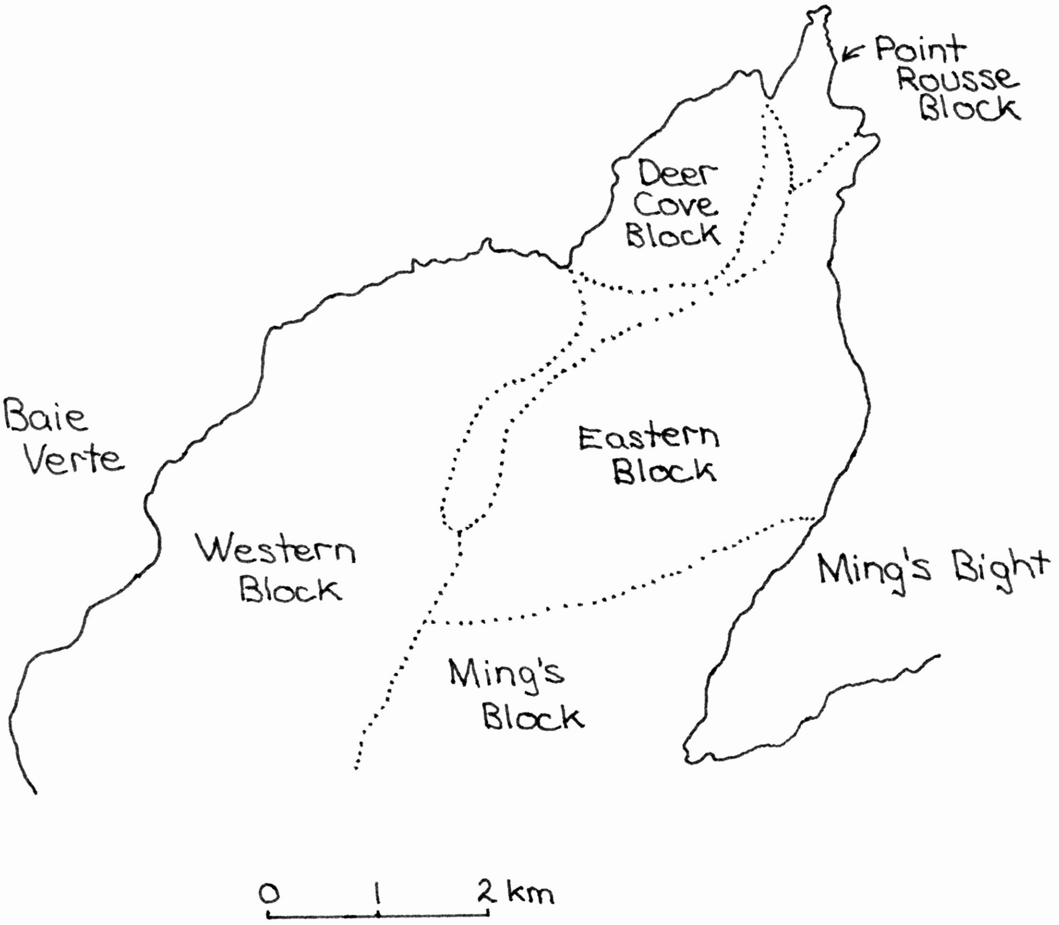

Figure 10. Peninsula occurring between Ming's Bight and Baie Verte, showing separation into five structural blocks.

Figure 12. Sample 44. Materials in the center of the photograph represent a vein passing through the layer of rhodonite. Large crystals of medium gray material represent rhodonite, which appears to have recrystallized. The other white, black, and gray materials within the vein are quartz crystals (crossed Nicols). Materials on the right and left sides of vein show rhodonite (gray) containing quartz crystals (white and black).

Figure 13. Photograph of one of the two layers occurring above and below the layer of rhodonite within sample 44. Granular, high-relief crystals are epidote. Acicular needles are actinolite. Both are contained within a matrix of quartz (plane polarized light).

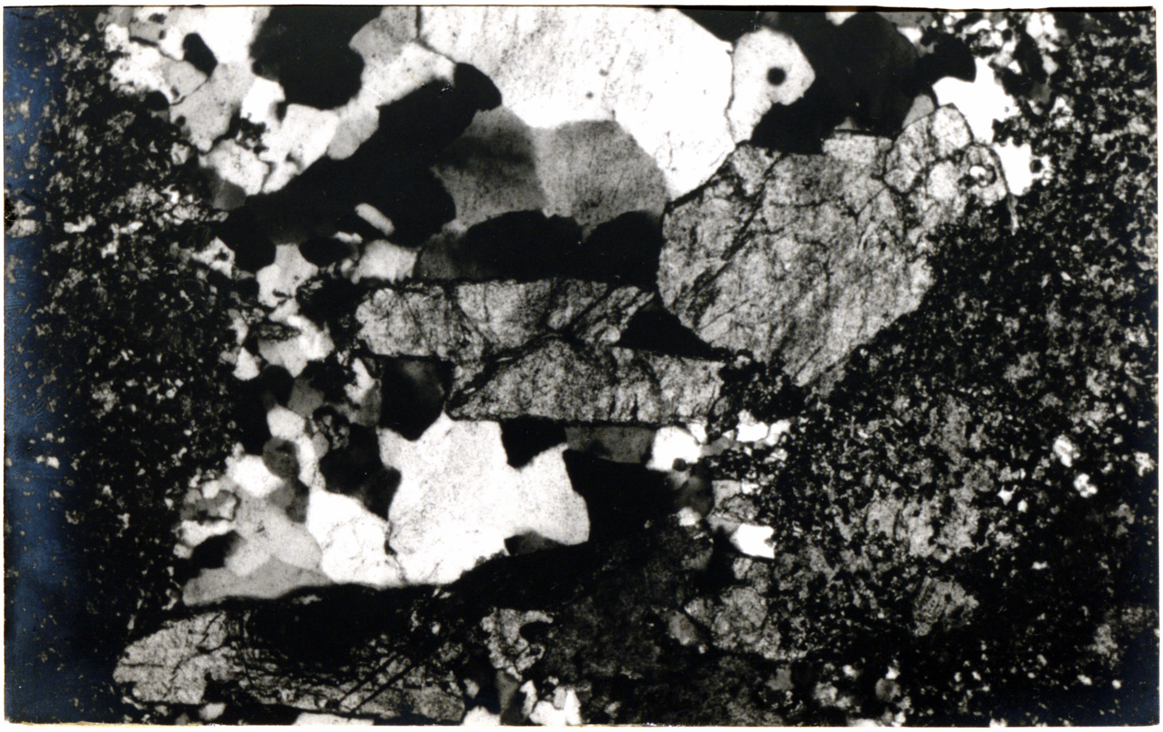

Fig 12

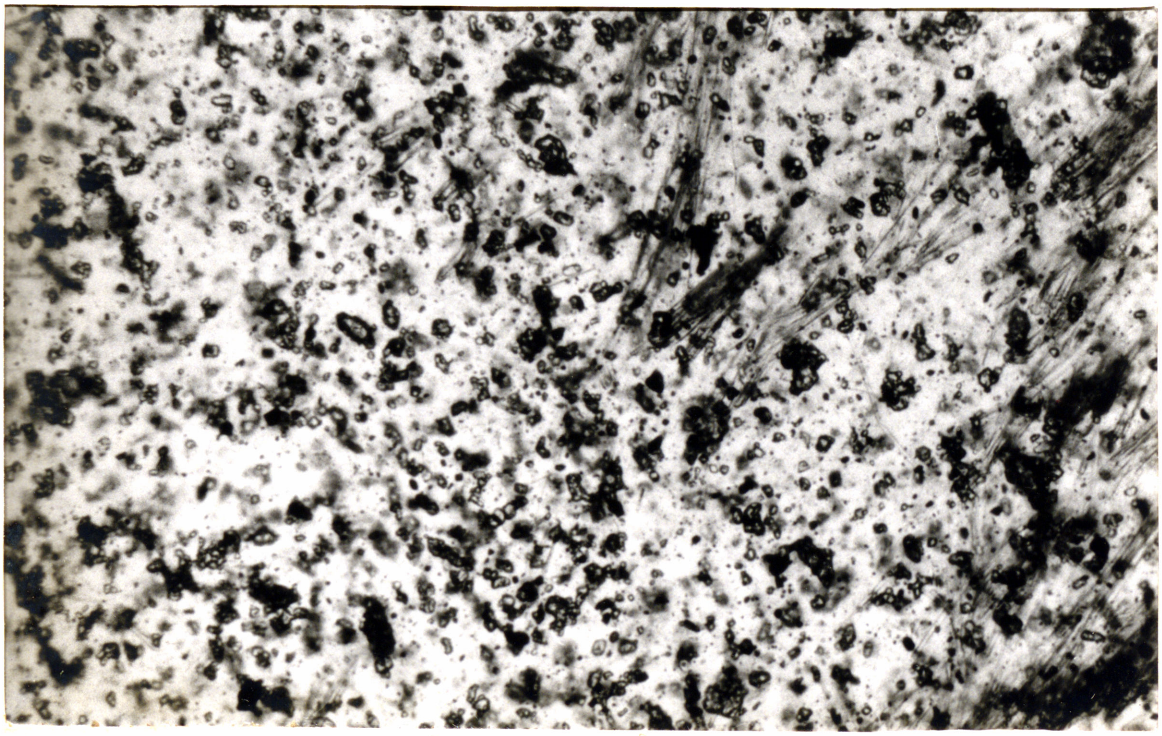

Fig 13.

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