Thomas Frisch

CANADA

DEPARTMENT OF MINES AND TECHNICAL SURVEYS

GEOLOGICAL SURVEY OF CANADA
MEMOIR 290

BAY OF ISLANDS IGNEOUS COMPLEX WESTERN NEWFOUNDLAND

By
Charles H. Smith

EDMOND CLOUTIER, C.M.G., O.A., D.S.P.
QUEEN'S PRINTER AND CONTROLLER OF STATIONERY
OTTAWA, 1958

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PREFACE

Ultrabasic rocks are the source of several of the important minerals of commerce and new information on their nature and origin is of interest and value. In addition to their economic importance, they also supply information on the composition and differentiation of earth layers below the granitic crust.

The Bay of Islands Complex is well suited to detailed studies. It consists of a group of large, well exposed layered intrusions, with rock types ranging from peridotite, dunite and pyroxenite to gabbro and quartz diorite in composition. The author describes these rock types, their interrelationships, genesis and relation to the formation of chromite concentrations. Comparisons are made with other districts in which similar rock types are found.

The information gained from this study will be of value in the elucidation of petrological problems existing in less well exposed areas of ultrabasic and basic rocks.

GEORGE HANSON,
Director, Geological Survey of Canada

OTTAWA, April 25, 1956

BAY OF ISLANDS IGNEOUS COMPLEX WESTERN NEWFOUNDLAND

CHAPTER I

INTRODUCTION

General Statement

The ultrabasic and gabbroic rocks of western Newfoundland form layered plutons that show many of the characteristics of the Bushveld, Stillwater, and other classic gabbroic plutons upon which modern theories of petrology are built. The Newfoundland plutons also have distinct features that are significantly different from those of the classic areas. They are unique in that they combine features of typical gravity-stratified sheets (e.g., the Bushveld Complex) with those of the massive ultrabasic plutons characteristic of orogenic belts (e.g., the serpentinite masses of the Appalachian belt). The Bay of Islands rock suite—peridotite, dunite, pyroxenite, and banded and massive gabbroic rocks—is typical of the rock types found in other orogenic belts where post-emplacement deformation and intense serpentinization have blurred the nature and primary interrelationships of the rock types. Two features of the Bay of Islands Complex, rugged terrain and the property of ultrabasic rocks to inhibit the growth of vegetation, combine to create excellent exposures of the plutons. Furthermore, the well-developed primary banding and the presence of primary minerals undestroyed by serpentinization provide much information on the original nature of the ultrabasic rocks. It is therefore hoped that the conclusions drawn here may prove of value in those areas of ultrabasic and gabbroic rocks where field relationships are not as favourably exposed.

The Bay of Islands Igneous Complex, with its surrounding volcanic and sedimentary rocks, is the only part of a major belt of eugeosynclinal type rocks exposed along the west coast of Newfoundland (*see* Map 1). This part of the belt lies between Port au Port Peninsula and Bonne Bay. A submerged link possibly connects it with similar eugeosynclinal type rocks near Hare Bay in northern Newfoundland. The Complex is a north-northeasterly trending, discontinuous belt of layered ultrabasic and gabbroic rocks, 60 miles long and up to 10 miles wide. The principal members of the Complex, from south to north, are known as the Lewis

Hills, Blow Me Down Mountain, North Arm Mountain, and Table Mountain plutons. Each has a thick basal ultrabasic zone overlain by banded and massive gabbroic rocks. Thin ultrabasic layers, commonly feldspathic, are also found in separate banded gabbro bodies such as those that outcrop on Mount St. Gregory and along the seacoast. Other small isolated areas of ultrabasic rocks are interpreted as 'cold punches' along fault planes. The major ultrabasic plutons have westerly dipping floors that are generally conformable with the underlying Ordovician sedimentary and volcanic rocks. The latter rocks are metamorphosed to amphibolite and calcic hornfels near their contacts with the intrusive rocks.

This report is concerned mainly with the description and interpretation of the igneous rocks of the Complex. In the northern half of the Complex the surrounding sedimentary and volcanic rocks have been mapped in detail, but in the southern half of the Complex only the ultrabasic and interbanded gabbroic rocks of the principal plutons have been studied.

Accessibility

The Bay of Islands Complex may be reached by boat or car from Corner Brook. Corner Brook (population 8,635) is situated on Humber Arm, a branch of the Bay of Islands. Its principal industry is the manufacture of pulp and paper.

A three-hour drive north from Corner Brook takes one to Woody Point on scenic Bonne Bay. A newly completed branch road from Woody Point to Trout River passes along the flanks of Table Mountain. From Trout River the Trout River Ponds, occupying a fault-controlled valley of great beauty, may be travelled by boat.

The North Arm Mountain pluton is best reached by boat from Corner Brook, a distance of about 28 miles. From an anchorage in North Arm, numerous exposures of its basal contact and metamorphic aureole may be visited in a short time. Parts of the Mount St. Gregory highlands may also be examined near the north shore of the Bay of Islands. No roads or navigable waterways traverse the land-area between the Trout River Ponds and the Bay of Islands.

The northern flanks of Blow Me Down Mountain are accessible by footpath from Benoit's Cove, a village on the south shore of Humber Arm that is connected by road with Corner Brook.

Serpentine Lake and the upper part of Serpentine River form a convenient route of travel across the southern half of the Complex. A footpath connects Serpentine Lake with the main line of the Canadian National Railways. Chartered aircraft fly from Corner Brook to Serpentine Lake in fifteen minutes. The northern side of the Lewis Hills, and the southern part of Blow Me Down Mountain are easily reached from the shores of the lake.

The southern part of the Lewis Hills may be reached by dirt road to Point au Mal on Port au Port Bay. There a tractor road, impassable to cars, leads to the workings of the Newfoundland Asbestos Company on Lewis Brook. A permanent road to replace the tractor road is now being built. An alternate route is to travel by motor boat from the mouth of Fox Island River to Mine Cove, and then to walk inland.

Although travel by boat and highways now being built is making the flanks of the mountains increasingly accessible, travel in the mountains must be entirely on foot. Thick growths of low trees and high cliffs make climbing difficult in places, but the mountains are flat-topped and unwooded, and are easily traversed on foot.

Previous Geological Work

The geology of western Newfoundland was first described by J. B. Jukes (1842)[1], and later in more detail by A. Murray and J. P. Howley (1881, 1918). For many years the only published geological map of the area, however, was the geological map of Newfoundland compiled by J. P. Howley (1907). Ultrabasic and associated volcanic rocks are shown on this map as a single unit called "serpentines, diorites, dolerites, etc."

Except for brief descriptions of the asbestos and chromite occurrences, the ultrabasic and gabbroic rocks of the Bay of Islands Complex were little known prior to 1933. During the period 1933 to 1936, several investigators studied parts of the Complex and outlined the general features of the plutons. A. K. Snelgrove (1934, 1934a) published a short description of the Blow Me Down pluton and summarized the available information on the chromite deposits. Earl Ingerson (1935) described the northern half of the Complex. He interpreted Table Mountain and North Arm Mountain as parts of two separate, differentiated laccoliths. This view was disputed by A. F. Buddington and H. H. Hess (1937). John R. Cooper (1936) described the geology of the southern half of the Complex. Ingerson and Cooper conducted the only detailed petrographic studies of the Complex and their published results were of great help in the present study. In 1947, John Troelsen mapped part of Table Mountain and Mount St. Gregory areas during a study of the stratigraphy and structure of the Bonne Bay-Trout River area.

Field Work and Acknowledgments

The present report is based upon field work done during the summers of 1951, 1952, and 1953. The study was initiated as a strategic mineral investigation of the chromite deposits, and later expanded to include a detailed study of the intrusive rocks of the area. Geological data were

[1]Dates in parentheses are those of references cited in the bibliography.

first recorded on vertical aerial photographs and later transferred to topographic maps that became available in 1953.

In the field, the writer was assisted by W. A. Craig, J. Davy, W. G. Gates, and A. Warren in 1951; by S. H. Kranck, J. T. Miller, C. C. MacKenzie, P. Lee, and I. Snow in 1952; and by M. Genes, W. H. Hopper, P. Lee, T. Parsons and I. Snow in 1953. He is deeply indebted to these men for their cooperation and enthusiasm. Mr. and Mrs. George Lee, Port au Port, and Mr. and Mrs. Taylor Parsons, Trout River, extended hospitality to members of the field party on many occasions.

Laboratory studies were made at Yale University, where a study of the ultrabasic rocks of Table and North Arm Mountains was submitted in 1952 as partial fulfilment of the requirements for the degree of Doctor of Philosophy. Professors Alan M. Bateman, Matt. S. Walton Jr., and Horace Winchell were very helpful in this study. Further work was carried out at Geological Survey headquarters in 1952 and 1953.

Physiography

The Bay of Islands Igneous Complex forms a distinct physiographic unit along the west coast of Newfoundland. The unit may be described in general terms as an uplifted, deeply dissected peneplane bounded by steep slopes. On the east, it is separated from the Long Range by a lower, irregular surface whose summits may be remnants of another large erosion surface. On the west it is bounded by a coastal lowland with an average elevation of 100 feet. The uniformity of this lowland is broken along the coast by several ridges underlain by resistant intrusive and extrusive rocks that are almost as high as the main peneplane surface.

The upland has a gently undulating surface of low relief (*see* Plate I). It is over 2,000 feet in elevation, and has a maximum elevation of 2,673 feet in the Lewis Hills. Where underlain by ultrabasic rocks the peneplane surface has been converted into a felsenmeer. The toxic effect of magnesium silicates on plant growth has prevented the growth of trees except in swampy areas. Differential erosion on a fine scale has developed a ridge and valley pattern in the interbanded peridotite and dunite of North Arm Mountain, the more resistant peridotite layers forming ridges about 25 feet high parallel to the regional trend of the bedrock structures. In areas underlain by gabbroic rocks, local relief may be as much as 300 to 400 feet.

The peneplane has been correlated with the upland surface of the Long Range and named the Long Range peneplane by Twenhofel and MacClintock (1940), who suggest that it may be late Cretaceous or early Cenozoic in age. If the present rate of erosion, based upon the amount of post-glacial downcutting of streams, is projected back in time, it appears doubtful that the uplifted peneplane surface could have survived erosion throughout all of Cenozoic time, and a later Cenozoic time for the uplift of the surface appears probable.

The topography is largely controlled by lithology and structure. Igneous and metamorphic rocks underlie the upland areas whereas sandstone and shale underlie adjacent lowlands. Streams flow either northeasterly, parallel to the strike of the igneous and sedimentary rocks, or northwesterly, parallel to the trend of numerous transverse faults. The Trout River Ponds are structurally controlled by the Trout River transverse fault. Serpentine Lake occupies a valley probably controlled by another series of transverse faults.

The contacts of the resistant intrusive rocks with the surrounding sedimentary rocks are characterized by abrupt scarps. Northwesterly trending fault-line scarps up to 1,500 feet high provide excellent exposures of the ultrabasic rocks. The scarps are steep near the top, but their lower parts are buried in sliderock except where exposed by torrential streams. Northeasterly trending scarps follow the basal contacts of the plutons. These are generally wooded where underlain by sedimentary and metamorphic rocks, but the upper parts, underlain by ultrabasic rocks, are barren (*see* Plate II A).

Slump features are common on the mountain slopes and testify to their instability. Plate IX shows the typical concave outline of a large landslide on the southwest side of Table Mountain near the Narrows. It consists of large blocks of gabbro and ultrabasic rocks. Smaller landslides obstruct the entrance to Rope Cove Canyon, and a series of slump terraces form the western wall of the canyon. Much of the bedrock near the coast between Mine Cove and Lewis Brook is covered by landslides.

Numerous canyons notch the ultrabasic scarps (*see* Plate II B). They are 1 mile to 3 miles long and up to 1,500 feet deep. They vary from V-shaped forms typical of post-glacial stream erosion to theatre-headed valleys that have been called cirques (Twenhofel and MacClintock, 1940a, p. 1734). Stone Brook valley, on the southwest side of North Arm Mountain, is incised up to 1,000 feet into the upland surface and is typical of the youthful V-shaped valleys. Stone Brook follows an irregular course with many offsets and waterfalls, and completely fills the bottom of its valley, which extends over 2 miles from the north shore of the Bay of Islands.

The theatre-headed valleys are confined mainly to the barren ultrabasic rocks. One of the better developed of them is near the northeast corner of Table Mountain. They notch the southwest ultrabasic scarp of Table Mountain along Second Trout River Pond, but are absent from the gabbro scarp on the opposite side of the Pond. They are randomly located on the northern and southern sides of the mountains. Their long axes trend northeasterly, parallel to the regional structural trend, except in the Lewis Hills, where they are controlled by local northwest-striking structures.

The V-shaped valleys are obviously the result of stream action. It is difficult to evaluate the role of glaciation in developing the theatre-headed valleys, as any evidence of glacial scour there may have been, has been destroyed by frost action and post-glacial erosion. However, many of the cirque-like forms in ultrabasic rocks are apparently the result of the headward migration of knickpoints, which were initiated at the scarps, and the rapid lateral recession of the valley walls by slumping rather than by glacial scour. The migration of knickpoints up branches of streams (*see* Plate III) accompanied by rapid erosion of the interfluvial areas, aids in widening the heads of the theatres. Scarps underlain by ultrabasic rocks are unstable and it appears that slumping of the ultrabasic rocks combined with the headward migration of knickpoints are important factors in the production of cirque-like forms in the area. Many of these canyons are due to rapid erosion in post-glacial time.

For a more complete treatment of the surface features of Newfoundland, Wisconsin glaciation and post-glacial emergence, the reader is referred to papers by Twenhofel and MacClintock (1940, 1940a) and by Flint (1940).

Glaciation

Frost action has obliterated most minor glacial features. Where preserved, striæ merely show local topographic control of ice movement along the major valleys. The deep, steep-sided, U-shaped valleys of Bonne Bay, Trout River Pond, the inlets of the Bay of Islands, and the valleys of Serpentine River and Fox Island River apparently acted as channels through which glacier ice passed to the sea. In Second Trout River Pond, bottom has not been reached at 200 fathoms according to a local tradition based on soundings made many years ago and not necessarily reliable. This deepening is probably due to glacial scour. Erratics of pink granite, limestone, and quartzite occur on the upland surface. Similar rocks outcrop near the intrusions. There are no remnants of till sheets on the peneplane surface and only minor areas of till in the larger valleys.

The glaciation of western Newfoundland presents many unsolved problems, principally involving the extent, number, and ages of glaciations, and it is difficult to obtain indisputable evidence bearing on any of these problems. Coleman (1926) and Fernald (1930) state that part of the Long Range has never been glaciated, whereas Twenhofel (1912) and Twenhofel and MacClintock (1940a) claim that the entire Range has been glaciated. The presence of erratics on the mountains of the Bay of Islands area supports the idea that the entire area was glaciated, but erratics are not abundant and are found within a few miles of outcrops of their parent rock.

No glacial striæ were found on the upland surface, but they may have been destroyed by frost-wedging. The general scarcity of erosional and depositional glacial phenomena on the upland surfaces suggests that glaciation of the surface was not intensive, and certainly not comparable in intensity with that of the Canadian Shield.

An indirect method of interpreting the extent of glaciation was applied by Flint (1940). He described the wave-cut cliffs and emerged deltas along the coast between Bonne Bay and St. Georges Bay. The elevations of these features combine to form an inclined plane that he named the Bay of Islands surface. This surface rises at the rate of 3 feet per mile to the north-northwest, and has an elevation of 70 to 80 feet on the north shore of the Bay of Islands, 100 to 110 feet at Trout River, and 125 feet at the north end of Bonne Bay. Flint concluded that the tilt of this surface proves that the Labrador ice-sheet had invaded the island of Newfoundland. However, the possibility that either circumglacial uplift, related to an ice-sheet that never reached as far as the Bay of Islands area, or tectonic uplift unrelated to glaciation has tilted the strand lines must be considered. The importance of circumglacial uplift is difficult to evaluate. One line of reasoning that suggests the possibility of a recent tectonic uplift of the area should be considered before accepting uplifted shoreline features as an absolute measure of post-glacial recoil in western Newfoundland. Recent radio-carbon dating suggests that the last of the Wisconsin maximum was about 11,000 years ago (Flint and Deevey, Jr., 1951). If the V-shaped valleys of the area are of post-glacial age, the rate of erosion of the ultra-basic rocks appears to be so great that the uplifted peneplane surface would have been completely destroyed had it reached its present altitude in early Cenozoic time. As the elevation of the peneplane was unquestionably of tectonic origin it may be argued that this tectonic uplift continued into late Cenozoic time, later than the 11,000 years when the ice was last at a maximum, and that, therefore, it may have tilted the strand lines produced at that time.

CHAPTER II

GENERAL GEOLOGY

General Statement

Cambrian and Ordovician sedimentary and volcanic rocks occupy a broad re-entrant in the crystalline rocks of the Long Range between Port au Port peninsula and Bonne Bay (*see* Map 1). A section perpendicular to the coast at this latitude shows the gneisses and schists of the Long Range faulted against Cambrian schists, quartzites, and limestones. Westward, in stratigraphic succession, are Ordovician limestones and dolomites (St. George and Table Head groups), overlain by clastic sedimentary rocks of the Humber Arm group. In the upper part of the Humber Arm group, near the coast, volcanic rocks are abundant and are intruded by the ultrabasic and gabbroic intrusions of the Bay of Islands Igneous Complex.

The Humber Arm group includes clastic and volcanic rocks of middle to late Ordovician age whose occurrence along the west coast of Newfoundland is restricted mainly to this broad re-entrant. The upper part of the group contains impure sandstones (subgreywackes), shales, and volcanic rocks. These rocks formed during a period of relative uplift of the source area, accompanied by volcanism. They constitute a typical eugeosynclinal association (Stille, 1941), composed as they are of clastic sedimentary rocks and basic volcanic rocks that have been folded and cut by ultrabasic, gabbroic, and dioritic intrusions.

The sedimentary and volcanic rocks of the map-area have been folded along axes with a northeasterly trend. They have been intruded by gabbroic and ultrabasic intrusions, and then broken by transverse and longitudinal faults of great magnitude. The complex igneous and structural history of the area, and the scarcity of fossils in the sedimentary rocks, make correlation of individual sedimentary formations difficult. It is probable that the area has undergone several periods of folding, intrusion, faulting, uplift, and peneplanation. To decide whether or not these events are related to the Taconian, Acadian, Appalachian, or other periods of orogeny is, unfortunately, largely a matter of personal opinion.

The name Bay of Islands Complex was proposed by J. R. Cooper (1936, p. 7), "as a general term to include the following separate igneous masses (from north to south): 1) Table Mountain mass, 2) North Arm Mountain mass, 3) Blow Me Down mass, 4) Lewis Hills mass. .....The comprehensive name, Bay of Islands Complex, is justified because of unity of geology, especially with regard to the sequence of banded rocks, and because all the masses are thought to be parts of a single lopolith or thick sheet". In this report, the name Bay of Islands Igneous Complex is used to include all the ultrabasic and gabbroic intrusions of the Bay of Islands re-entrant, from Bluff Head to Bonne Bay. It thus includes the four principal areas of ultrabasic rocks mentioned by Cooper, and also the numerous areas of banded gabbros that contain minor amounts of ultrabasic material and are exposed along the shore of the Gulf of St. Lawrence. The ultrabasic and gabbroic rocks are grouped under this heading solely because of their similar lithology and genesis. No implication regarding their form, lopolithic or otherwise, is intended.

Two large bodies of ultrabasic rocks occur north of Bay of Islands. They are referred to as the North Arm Mountain pluton and the Table Mountain pluton. Both plutons are characterized by thick zones of interbanded peridotite and dunite. They are parts of an originally continuous mass now offset by the Trout River fault. Each is overlain by a zone of genetically related, banded gabbros. Much of the banded gabbro zone of the Table Mountain pluton has been eroded away, whereas that of the North Arm Mountain pluton is largely preserved. The banded gabbro in these upper zones grades into massive gabbro. The massive gabbro is intrusive into a 'roof' of diabasic metavolcanic rocks that is overlain by pillowed basic volcanic rocks along Gregory River. A third intrusive body, composed of banded gabbros with small feldspathic dunite layers, underlies Mount St. Gregory. Along the coast north and south of Trout River are other areas of banded and massive gabbros that are intensely sheared and faulted. Quartz diorite intrusions cut the North Arm Mountain gabbros at many places and also outcrop in a belt parallel to the gabbros along the coast.

Two large bodies of ultrabasic rocks outcrop south of the Bay of Islands; the Blow Me Down pluton on the north is separated by the Serpentine River valley from the Lewis Hills pluton on the south. The Blow Me Down pluton closely resembles the North Arm pluton in its lithology and morphology. It has a similar layered sequence of ultrabasic rocks overlain by banded and massive gabbros and metavolcanic rocks. The Lewis Hills pluton, however, differs from the other three by the almost erratic distribution of its banded gabbro zones throughout the ultrabasic rocks, and also by the excess of dunite over peridotite. The pseudo-complexity of the Lewis Hills pluton is probably due to the more nearly horizontal attitude of its main ultrabasic-gabbro contact over much of the area.

Metamorphic rocks, ranging in composition from biotite phyllite, through amphibolite, to calcic hornfels, form contact aureoles along the unfaulted basal margins of the plutons. The relatively high lime content of these rocks, characteristic of the contact aureole over great distances, is attributed to lime metasomatism during the early phases of the intrusion. Metavolcanic rocks form the roof of the North Arm Mountain intrusive body. They are the only roof rocks positively identified. They grade upward into unmetamorphosed pillowed volcanic rocks.

TABLE OF FORMATIONS

Period or epoch	Group	Lithology
Pleistocene and Recent		Delta deposits, alluvial fan deposits, sliderock, erratics.
Pre-Carboniferous	Bay of Islands Igneous Complex	Peridotite, dunite, pyroxenite, gabbro, anorthosite, troctolite, diorite, quartz diorite.
Ordovician (Middle and Late)	Humber Arm group	Volcanic rocks of basaltic to sodic composition. Metavolcanic equivalents. Feldspathic and chloritic sandstone; red, green, and black shale; minor clastic limestone; amphibolite equivalents along basal contacts of ultrabasic plutons.
Ordovician	Table Head group	Limestone, dolomite, shale.
	St. George group	Limestone, dolomite.

Sedimentary and Volcanic Rocks

ST. GEORGE AND TABLE HEAD GROUPS

Ordovician limestone and dolomite of the St. George and Table Head groups form cuesta-like hills east of the belt of ultrabasic rocks. They dip westerly and underlie the beds intruded by the Bay of Islands Complex. In the northern half of the Complex, they lie well to the east of the ultrabasic belt, but south of the Bay of Islands they outcrop closer to it. Along the strongly deformed, southern margin of the Lewis Hills intrusion, on Fox Island River, St. George limestone is in contact with ultrabasic rocks. This is interpreted as a secondary, faulted contact rather than a normal igneous contact because the ultrabasic rocks are deformed and have caused no visible alteration of the limestone. The fact that pre-Humber Arm rocks are more common at the southern end of the Complex than farther north suggests that the northern plutons were intruded into a slightly higher level of the stratigraphic sequence. However, post-intrusion deformation has been more intense along the southern margins of the Complex, as shown by folding and faulting of the contact of the Lewis Hills pluton, and this deformation is partly responsible for folding and faulting the lower formations closer to the ultrabasic belt. There is no indication that primary ultrabasic rocks were intruded into the pre-Humber Arm rocks, although later deformation has brought them into contact in places.

HUMBER ARM GROUP

Schuchert and Dunbar (1934, p. 86) proposed the term Humber Arm series for the sequence of clastic rocks exposed along the shores of Humber Arm in the Bay of Islands. They describe it as a thick unfossiliferous series of sandstones and shales that represents a change in conditions of deposition from those prevailing during the formation of the lower and middle Ordovician limestones and dolomites. The unit rests upon the Cow Head breccia and overlies rocks of the Table Head group. It is of Chazyan age in part, but its upper limit is unknown and it may include rocks of middle and late Ordovician, or possibly even of Silurian age. The strata are strongly folded and faulted and key beds can be followed for only short distances. The rocks are generally unfossiliferous, especially the upper beds, and constitute a lithological unit whose time range is only partly known. For these reasons the term 'group' is more appropriate than 'series'.

The Humber Arm group is largely restricted to a broad re-entrant in the crystalline rocks of the Long Range, between Port au Port Peninsula and Bonne Bay (Map 1). It does not outcrop south of Port au Port Peninsula, and only a few exposures of Humber Arm rocks are known along the northwest coast of Newfoundland north of Bonne Bay[1]. Rocks correlated with the Humber Arm group outcrop at the northern end of the Long Range (Cooper, 1937; Betz, 1939).

Whereas the limestone and dolomite of the St. George and Table Head groups reflect a quiescent environment, the Humber Arm group marks the inception of early Palæozoic orogenic conditions in western Newfoundland. The lower formations of the group suggest a rising source-area and the upper formations indicate the advent of submarine volcanism. The late Ordovician volcanism represents the generation of basaltic magma and reflects the building up of orogenic forces prior to mountain building. The ultrabasic and gabbroic rocks, which are closely associated in space with the volcanic rocks, may represent the later expression of this period of basic igneous activity.

Various subdivisions of the Humber Arm group have been made by workers in adjacent areas, outside the belt of igneous intrusion. Troelsen

[1] Helgi Johnson, personal communication.

(1947), working near the north end of the Complex, divided the group into the Gadd's Point slates, the Mackenzie Brook formation, the South Arm formation, and the Skinner Cove volcanic rocks. Weitz (1953), mapping east of the Bay of Islands, divided the group into the Summerside, McIvers, and North Arm formations and the Island volcanic rocks. In the early stages of the present study, the writer mapped only those sedimentary and volcanic rocks near the margins of the plutons, and did not attempt to subdivide the Humber Arm group. In later mapping of the regional setting of the northern half of the Complex, it was possible to divide the Humber Arm group into two broad lithologic units, one composed mainly of sedimentary rocks and the other composed of volcanic rocks. The sedimentary unit corresponds partly to the South Arm formation of Troelsen, or the North Arm formation of Weitz. It differs from the pre-Humber Arm rocks in the abundance of sandstone and shale, and the lesser amounts of limestone. The volcanic unit forms the upper part of the Humber Arm group. It includes the Skinner Cove volcanic rocks of Troelsen and the Island volcanic rocks of Weitz. It consists chiefly of basic volcanic rocks, with some red to black shale and impure sandstone. The ultrabasic and gabbroic rocks were intruded into the upper part of the Humber Arm group.

The Humber Arm sandstone is commonly light greenish grey and medium to coarse grained. In a few localities it grades into conglomerate, but conglomerates are of only minor importance. Graded bedding, the most common internal structure, was used to determine the tops of beds. In several thin sections, the sandstone was seen to contain 75 per cent angular to sub-rounded quartz grains, 5 per cent detrital plagioclase (sodic andesine) and lesser amounts of altered, yellowish brown biotite, pale green chlorite, zircon, garnet, and iron ore. Rock fragments are not common, except in local conglomeratic facies where quartz-bearing pebbles are most abundant. The Humber Arm sandstone is more quartzose than typical greywackes, and is best described as a greenish grey, feldspathic, chloritic sandstone. Its mineralogy suggests that it was derived from the erosion of granitic rocks, possibly the Precambrian rocks of the Long Range.

Humber Arm shales are commonly red to black, although green hues are developed along secondary fractures due to the reduction of ferric iron by organic solutions. Weathering of symmetrical marcasite nodules in the black shales has formed yellow and rust stains on the rocks.

Limestone members of the Humber Arm group are rare near the belt of intrusions. A small bed of light grey arenaceous limestone or calcarenite near Liverpool Brook consists of 60 per cent angular quartz. It also contains, in order of abundance, plagioclase, biotite, and zircon.

Volcanic rocks, typical of the western, upper parts of the Humber Arm group, outcrop along much of the coastal belt between Fox Island

River and Bonne Bay. A few flows are intercalated with the sedimentary rocks at the base of the plutons, but flows are most abundant between and to the west of the plutons. Metamorphosed equivalents of the volcanic rocks form extensive caps over parts of the gabbro masses, especially in the Mount St. Gregory highlands and to a lesser extent on Blow Me Down and Table Mountains. Troelsen (1947) has named the volcanic rocks north of Trout River the Skinner Cove volcanic rocks, and given them the status of a formation. Similarly, Weitz (1953) treats the volcanic rocks in the Bay of Islands as a formation, the Island volcanic rocks. Because of the complex structural and intrusive history of the area, it is not possible to correlate directly between these two areas. An attempt has been made to map the larger areas of volcanic rocks separately, especially in the northern part of the Complex. In places, however, the sedimentary units include some volcanic rocks, and the volcanic units, some sedimentary rocks.

Most of the volcanic rocks are basaltic andesites, although the composition of the plagioclase phenocrysts varies from albite to labradorite in different flows. They are dark greenish grey or, rarely, shades of red and are megascopically uniform in appearance. Flows with well-developed pillows and breccias are common. Both these and the more massive flows are commonly vesicular and, in places, porphyritic. In thin sections they are seen to have a felted, diabasic or ophitic fabric.

The least altered volcanic rocks are fairly simple in composition, being composed mainly of plagioclase, clinopyroxene, and some iron ore. Plagioclase forms euhedral to subhedral phenocrysts up to 4 mm. in length and is also a dominant component of the fine-grained groundmass. Some phenocrysts show oscillatory zoning, with twinning superimposed on the zonal structure. Others have altered cores and clear rims. The plagioclase ranges in composition from albite (An_5) to labradorite (An_{70}). Reassimilated margins and the alteration of phenocrysts along cracks suggest that the more sodic plagioclases have formed by alteration of calcic plagioclase. Diopsidic augite, green in hand specimen and slightly pinkish in thin section, occurs as euhedral phenocrysts and fills interstices between plagioclase laths of the groundmass. No fresh olivine grains were detected, and the presence of olivine is inferred only from pseudomorphs of green amphibole and colourless penninite. Secondary minerals in the volcanic rocks include green actinolitic hornblende, apparently derived from primary pyroxene, serpentine minerals, chlorite, iron ore, and sulphide minerals. Calcite replaces plagioclase either completely or along single zones in twinned crystals. Calcite also occurs as vesicular fillings together with chlorite or prehnite, and forms veins that contain native copper and hydrocarbons. Metamorphism of these volcanic rocks by intrusive gabbros has formed the amphibolites and granulites described below.

Metamorphic Rocks

Most of the metamorphic rocks of the map-area are members of the Humber Arm group that were altered by the ultrabasic and gabbroic rocks. They form a narrow contact-metamorphic aureole along the basal contacts of the ultrabasic rocks and flat-lying areas of metavolcanic rocks in the roof of the gabbroic rocks.

METAMORPHIC ROCKS ALONG THE BASE OF THE PLUTONS

The metamorphic rocks of the northwesterly dipping basal contact aureoles show evidence of moderate to high-grade thermal metamorphism and metasomatism related to intrusion, together with later retrograde effects produced by mechanical deformation. In some localities the retrograde effects have almost obliterated the evidence of thermal metamorphism. Contact-metamorphic rocks form narrow aureoles along the southwest margins of each of the four major plutons. They extend from south of Fox Island River in the Lewis Hills to the Gulch north of Table Mountain, and are offset along their strike by northwesterly striking transverse faults. Similar contact rocks border the eastern part of the Mount St. Gregory banded gabbro mass for 3 miles. The aggregate length of the metamorphic aureoles along the basal contacts of the major intrusions is about 34 miles. The visible aureoles average 200 to 500 feet in thickness, except in the southern part of the Lewis Hills where they have an outcrop width of almost a mile. There, inliers of ultrabasic rocks in the metamorphic belt suggest that much of the aureole may be underlain by ultrabasic rocks and the increased outcrop width is partly due to the flatter attitude of the contact between the overlying metamorphic rocks and the underlying ultrabasic rocks.

The basal contact of the North Arm Mountain pluton affords the best opportunity for detailed study of the metamorphic effects of the ultrabasic rocks (*see* Figure 1). This contact is well exposed on cliff faces along the northwest side of North Arm; there the metamorphic rocks have been less affected by the dynamic retrograde metamorphism that has elsewhere partly or completely destroyed the original metamorphic minerals of the aureole. Even in this locality, however, post-intrusion movements have imposed a gneissic to platy structure on the ultrabasic rocks adjacent to the contact and caused some mineralogical changes in the metamorphic rocks. Although the basal contact is better preserved there than elsewhere, it does not completely reflect conditions immediately following magmatic emplacement.

The metamorphic rocks grade from calcic hornfels at the contact, through diopside amphibolite and amphibolite, to phyllite in the outer part of the contact aureole. Pink pyrope garnet is present in all rock types. The

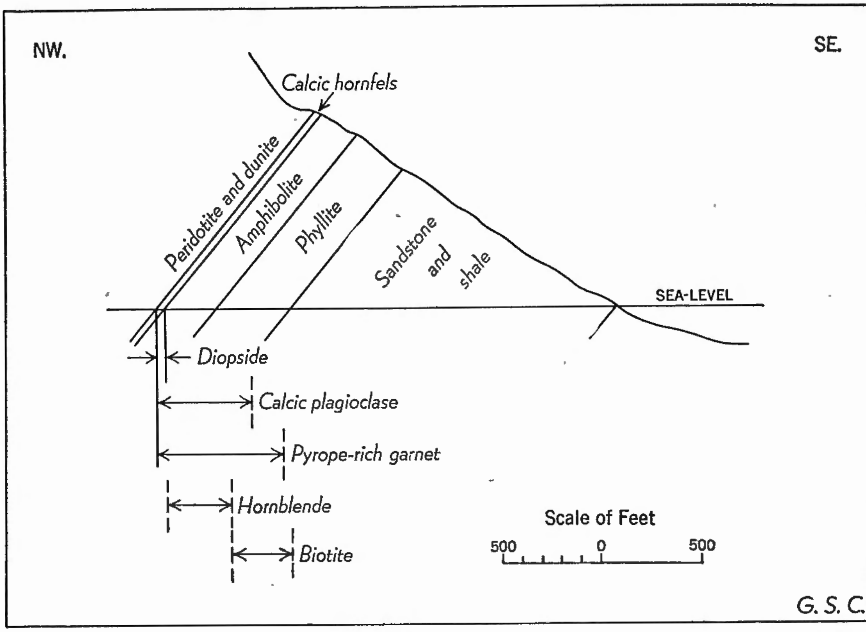


Figure 1. Section across contact aureole, North Arm Mountain pluton.

foliation of these rocks is conformable in strike with both the ultrabasic contact and the primary banding of the igneous rocks, except where disturbed by later faulting. Beyond the contact aureole, the country rocks are predominantly sandstones and shales with minor volcanic rocks.

Calcic hornfels occurs at the contact in a zone 1 foot to 10 feet wide. It is typically a fine-grained, compact, light greenish grey to purplish grey rock with fine mineralogical banding. Colourless to light green diopsidic pyroxene and calcic plagioclase are the characteristic minerals. The optical properties of the pyroxene show that it belongs, compositionally, at the diopside-salite boundary of the diopside-hedenbergite series according to the usage proposed by Hess (1941). One plagioclase was identified as anorthite, near An_{90}, although a similar specimen from the Table Mountain aureole contained labradorite near An_{60}. The plagioclase is altered to aggregates of an almost opaque mineral with low birefringence that resembles hibschite (hydrogarnet). Most specimens contain a high percentage of secondary calcium-rich minerals. These include prehnite, clinozoisite, xonotlite and calcite.

The amphibolite zone averages 200 feet in width. The amphibolite is a medium-grained, light to dark grey foliated rock composed essentially of hornblende and altered plagioclase. The hornblende crystals form distinct bands with their long axes roughly parallel to the banding. Thin sections show that the hornblende is poikiloblastic and pleochroic with X = golden yellow and Z = deep reddish brown. Pleochroism apparently decreases in intensity away from the ultrabasic contact, although further study is necessary to confirm this observation. The amphibole has the optical properties of a common hornblende, slightly richer in magnesium than iron. It is altered in places to actinolitic amphibole. The plagioclase of the amphibolites is almost completely altered to aggregates of very fine-grained minerals with low relief and low birefringence whose composition is indeterminate. Pink garnet porphyroblasts, colourless in thin section, have well-developed rims of vermicular amphibole and sericite, except in those specimens where the fabric has been modified by cataclasis. In one part of the aureole, a variation in the index of refraction of the garnet was noted, from $N = 1\cdot765$ to $N = 1\cdot737$ nearer the contact. A garnet grain from a diopside-bearing amphibolite was identified as a pyralspite garnet near pyrope in composition, but containing iron and calcium. This determination is based upon physical properties ($N = 1\cdot743 \pm 0\cdot003$; S.G. = $3\cdot57 \pm 0\cdot003$; $a = 11\cdot545 \pm 0\cdot005$ A). Zoisite, epidote, sphene, apatite, iron oxides, and secondary sericite and penninite are minor constituents.

The outer phyllite zone is characterized by several hundred feet of dark greenish grey to dark grey, fine-grained to aphanitic rocks with a coarse planar parting that grades into a poorly developed schistosity. Small-scale drag-folds, and porphyroblasts of garnet and quartz, are visible in hand specimens. In thin section, garnet, quartz, sodic plagioclase, zoisite, and muscovite grains are enclosed in a fine-grained matrix of biotite, chlorite, quartz, and iron oxides. The chloritized garnet porphyroblasts are characterized by a sieve structure, and in some specimens a poorly developed pinwheel structure.

The metamorphic aureole of the Table Mountain pluton, although not greatly destroyed by later deformations, is more variable in width, more commonly faulted, and less well exposed than the North Arm Mountain aureole. Streams flow across the aureole where it has been broken by faulting. In the less deformed interstream areas, vegetation conceals much of the aureole. Up to 500 feet of contact rocks, dominantly garnetiferous amphibolites, are exposed adjacent to the triangular-shaped area of ultrabasic rocks on the southwest shore of Second Trout River Pond. This amphibolite is similar to that previously described, except that the hornblende is strongly pleochroic in green hues instead of yellow to brown. The metamorphic foliation is parallel to the banding of the ultrabasic rocks, but both planar elements are oblique to their mutual contact, reflecting the structural deformation of this dragged extension of the Table Mountain pluton. The embayment of the aureole at Second

Trout River Pond is similar in form to the embayment of St. George limestone between the two southern plutons of the Complex, which was caused by post-intrusion deformation.

Along the basal contact of the Table Mountain pluton the peridotite banding is much deformed, and in places strikes at right angles to the contact. The dip of the aureole is variable, ranging from 40 to 80 degrees west. The light green calcic hornfels zone is locally up to 50 feet wide. Diopside is the dominant and in some specimens the only unaltered component of the hornfels. It was noted up to 400 feet from the contact. Fresh calcic plagioclase (An_{55} to An_{60}) was recognized in only a few thin sections. It appears to have been the mineral most susceptible to attack by later solutions, which have altered it to sericite and zoisite. Moderate-pink garnets form porphyroblasts up to 1 inch in diameter. The amphibolite zone varies from 200 to 500 feet in width. The hornblende of this zone is less pleochroic than that in the North Arm Mountain amphibolites. This feature, and slightly lower refractive indices, suggest it is a common hornblende possibly a bit richer in calcium and magnesium than those described from North Arm Mountain.

North of Sellars Brook contact rocks are cut by, and included in, the ultrabasic rocks. The small ultrabasic projections have a bluish colour on weathered surfaces in contrast to the yellowish brown weathered surfaces of the main peridotite mass. Their relationship to the surrounding amphibolites suggests that they were tectonically emplaced either by solid intrusion or by faulting. A similar serpentinite projection at the mouth of Shoal Brook is capped by unmetamorphosed sedimentary rocks. A narrow alteration zone composed of calcic zeolites is found at the contact, similar to those occurring along the Winter House Brook faulted contact described on pp. 48-50.

The metamorphic aureoles of the Blow Me Down and Lewis Hills plutons have been largely destroyed by post-intrusion deformation. Secondary flowage effects along the serpentinite-metamorphic rock boundary have imparted a gneissic and schistose structure to the serpentinites, and a schistose structure to the amphibolitic rocks. This has caused retrograde metamorphism of the contact rocks that could be mistakenly interpreted as a lower grade of primary metamorphism. Green schists are more abundant than amphibolites along the Blow Me Down Mountain contact. They have a cataclastic or mylonitic fabric superimposed upon their original metamorphic fabric and are composed of altered plagioclase, quartz, sericite, chlorite, apatite, and titanite. Specimens near the contact contain shattered garnet porphyroclasts that have been smeared out along the foliation planes and partly altered to penninite. Epidote and titanite also occur as porphyroclasts. The rocks are cut by veins of calcite and prehnite.

The Lewis Hills pluton appears to have borne the brunt of post-intrusion deformation, which has resulted in considerable blurring of the original contact-metamorphic effects. The contact aureole north of Hines Pond is narrow and consists of schistose rocks. The rocks contain minor amounts of diopside with slightly pleochroic green actinolitic amphibole. No fresh plagioclase was detected but areas of secondary alteration minerals (saussurite) may represent original plagioclase. Epidote, garnet, titanite, and chlorite occur as minor constituents. Southwest of Hines Pond the zone is up to a mile in width, but the grade of metamorphism is generally lower. Phyllite and actinolite schist form much of the zone. Small areas of serpentinite occur in parts of the aureole, and south of Chrome Point unmetamorphosed shales are faulted into the ultrabasic rocks. The ultrabasic rocks in this area are extremely deformed and the contact is offset by numerous faults and folds.

A narrow amphibolite zone 3 miles long, lies along the eastern side of the Mount St. Gregory gabbro mass. The amphibolite consists of brown pleochroic hornblende partly altered to actinolite, altered plagioclase, and epidote. It is almost identical with the amphibolite along the ultrabasic contact of the major plutons.

The characteristic features of the contact metamorphic aureoles are as follows:

(1) An outward gradation from minerals indicative of high rank thermal metamorphism (diopside and calcic plagioclase) to minerals characteristic of lower rank thermal metamorphism, indicating a genetic relation of the aureole to the intrusive bodies.

(2) A well-defined foliation indicating the role of stress during metamorphism.

(3) A higher lime content (in diopside, hornblende, epidote, plagioclase) and a higher magnesia content (in diopside, hornblende, pyrope) than is present in the sedimentary rocks beyond the aureole. The content of alumina increases with distance from the contact.

(4) A narrow width in spite of the evidence of high temperature metamorphism.

(5) A pyrope-rich garnet in all parts of the aureoles.

The original nature of the contact rocks is speculatory. The intrusions are commonly conformable with the country-rocks which include sandstone, shale, and minor limestone and volcanic rocks. Two contrasting theories of origin of the metamorphic rocks merit consideration:

(1) They were formed by thermal metamorphism of argillaceous dolomites, or of basic volcanic rocks.

(2) They were formed by the introduction of lime and magnesia into the country-rocks during intrusion.

Dolomitic rocks are exceedingly rare in those members of the Humber Arm group that have been intruded by the ultrabasic rocks. It is, therefore, extremely unlikely that they were continuously present at the metamorphic contacts. Basic volcanic rocks are more common than dolomite but less abundant than the sandstone and shale members of the Humber Arm group. In fact, volcanic members are practically absent near the metamorphic aureole of the North Arm Mountain pluton. Therefore, it is logical to assume that sandstone and shale, rocks low in lime and magnesia, were the chief rock types in the contact zone. It is therefore concluded that the presence of a continuous zone of metamorphic rocks richer in lime and magnesia than the normal sedimentary rocks, at unfaulted intrusive contacts throughout the Complex, indicates that lime and magnesia have been added to the country-rocks by contact metasomatism.

Harker (1939, p. 283) states that "diopside-bearing amphibolite is found in close association with intrusions of peridotite (now serpentine), but there is no reason to doubt that it belongs to the general regional system of metamorphism". Turner and Verhoogen (1951, p. 488) cite the metamorphism of initially noncalcareous sedimentary rocks to calc-silicate rocks at contacts with intrusions of peridotite. They state that the added lime is magmatic and conclude, "some writers regard it as having been concentrated in residual solutions left after crystallization of the ultrabasic 'magma', while others envisage it as having been set free during serpentinization of diopsidic pyroxene in the peridotite".

Although the metasomatic addition of lime and magnesia to the country-rocks seems probable, the fact remains that the ultrabasic rocks of the intrusions are notably poor in lime. The chemical analysis of peridotite near the floor of the North Arm Mountain intrusion (p. 30) shows a lime content of 0.31%. The primary minerals of the peridotite, forsterite and enstatite, are non-calcic and serpentinization of these minerals would not supply the lime for metasomatism. Only near Mount St. Gregory and in the metavolcanic roof of the pluton are similar amphibolites found associated with lime-bearing intrusive rocks (gabbros). The present zonal relation of the contact minerals to the intrusions implies that the lime has not been added by residual solutions after crystallization of the intrusions. If the addition of lime was a late phenomenon, crosscutting, lime-rich veins and a more random relationship of the metamorphic minerals to the contact would be expected. Neither the action of late residual solutions nor serpentinization of the peridotite is capable of explaining the lime metasomatism.

The banding of the igneous rocks and the parallel foliation of the metamorphic rocks indicate that the peridotite was emplaced under stress and, during emplacement, flowed parallel to the basal contact. The first magma to reach the horizons now exposed, probably consisted of low temperature components and was thus more fluid than the final ultrabasic mush containing those of higher temperature. Lime would be one of these low temperature components and it is reasonable to conclude that the lime metasomatism was caused by a more calcic, more fluid advance front of the peridotite magma. The idea of an advancing calcic (i.e. gabbroic) forerunner of peridotite intrusion is supported by the field observations of Bowen (1928, p. 158) and by the theoretical concept of the injection of a two-phase (gabbroic plus ultrabasic) magma discussed in later chapters. This use of contact metasomatism to infer the character of the early phases of intrusion is analogous to the use of chilled border zones for the same purpose.

METAVOLCANIC ROCKS OVERLYING THE NORTH ARM MOUNTAIN PLUTON

Fine-grained, massive, basic rocks in the northern half of the Complex are considered to represent the roof zones of the igneous intrusions, originally composed of basic volcanic rocks, but now metamorphosed to amphibolites and pyroxene-hornblende granulites. A small granulite area is restricted to the centre of the banded gabbro zone of Table Mountain. Similar metamorphic rocks form a flat-lying cap over many of the hills south of Trout River Pond, and grade into unmetamorphosed volcanic rocks near Gregory River. The change from fine-grained, massive basic rocks to unmetamorphosed volcanic rocks is so gradual that it cannot be recognized in the field unless pillows or other features diagnostic of lavas are evident. In the western parts of the Blow Me Down and Lewis Hills areas, fine-grained basic rocks of similar appearance are exposed, but have not been sufficiently studied to be separated from the closely associated plutonic gabbroic rocks.

Great difficulty is experienced in the field in attempting to subdivide and interpret the metamorphosed basic rocks. They consist of plagioclase (altered or fresh), amphibole (green or brown) and/or clinopyroxene (fresh or partly replaced by amphibole). Rocks of this composition may have various origins and, therefore, attempts to classify them genetically can lead to error. Thus, in the present study, rocks composed essentially of plagioclase and clinopyroxene are believed to be plutonic gabbros, volcanic basalts, or metamorphic granulites derived from basaltic rocks. Rocks composed of plagioclase and hornblende have formed by the regional metamorphism of plutonic gabbros or basaltic volcanic rocks, by the thermal metamorphism of basaltic rocks, or by the metasomatic alteration of sedimentary rocks. It is clear, therefore, that characteristics other than mineral composition alone must be used to attempt genetic subdivisions.

The gabbroic rocks of plutonic origin are distinguished by their coarse grain size, texture, banding, interbanding with ultrabasic rocks, crosscutting features, and the general absence of amphiboles. The volcanic rocks are distinguished by their fine grain size, ophitic and in places porphyritic textures, their association with sedimentary rocks, and their internal structures, pillows, breccias and sperulites. The fine- to medium-grained, massive, basic rocks of the area are the principal problem. They occur in places as an intermediate zone between the coarse-grained plutonic gabbros and the pillowed volcanic rocks. This suggests that they may be either plutonic in origin, representing chilled phases of the gabbro, or volcanic in origin, representing thermally metamorphosed volcanic rocks. Although they may have a composite origin, most are believed to be altered basic volcanic rocks because they grade into less altered, pillowed rocks, and in places, contain structures resembling deformed pillows and spherulites; some less altered phases have ophitic textures similar to those in the pillowed volcanic rocks.

The rock types included in the metavolcanic rock unit vary from massive, fine-grained, diabasic rocks, similar in microscopic aspect to the pillowed volcanic rocks of the Humber Arm group, to fine-grained, augite-hornblende-plagioclase rocks with a granulitic fabric. The transition from volcanic rocks to metavolcanic rocks was arbitrarily mapped where pillows or other megascopic volcanic structures are obliterated. Otherwise it was impossible to separate in the field the amphibolite and granulite facies from less altered, massive volcanic rocks. The extent of the metavolcanic rocks as shown on the geological map may thus be enlarged at the expense of less-altered volcanic rocks. It must be remembered that, whereas the gabbro-metavolcanic rock contact is commonly sharp, the contact between metavolcanic rocks and volcanic rocks is gradational and arbitrarily drawn. The contact between the coarse-grained plutonic gabbros of the North Arm pluton and the overlying metavolcanic rocks is flat over much of the North Arm pluton and individual metamorphic zones are not as well exposed as they are along the steeper-dipping basal contacts. However, there appears to be a general decrease in metamorphic grade from east to west, from granulitic and structureless amphibolitic rocks that cap the gabbros to pillowed volcanic rocks along the Gregory River valley.

The low-rank metavolcanic rocks are grey to greenish black in colour. The least altered varieties retain a primary ophitic texture and are commonly porphyritic. Flow structure is rare. White plagioclase phenocrysts and microlites are ragged in outline or clouded with clear rims. Secondary green amphibole has almost obliterated the primary augite and only small remnants of the latter are preserved. Chlorite and secondary iron ore minerals are common, and prehnite, calcite, sulphide minerals, and rarely quartz are minor constituents.

The more highly metamorphosed rocks have a fresher purplish black cast, but because of their fine grain size even this does not allow their separation from some of the less altered volcanic rocks in the field. In some outcrops a 'salt and pepper' appearance on the freshly broken surface is characteristic of the more metamorphosed rocks, the original ophitic texture being replaced by a granulitic fabric composed of partly aligned plagioclase and hornblende. Fresh, anhedral plagioclase, in places poorly twinned, varies within the andesine-labradorite range, with compositions of An_{50} to An_{55} predominating. Hornblende, pleochroic in shades of brown and poikiloblastic in habit, develops in place of green uralite. It resembles the brown amphibole found in the amphibolites along the basal contacts. Small amounts of epidote and iron ores are present. Other specimens of the more metamorphosed metavolcanic rocks contain light green to colourless diopsidic augite with an anhedral and in places a porphyroblastic habit. Thus the same sequence of metamorphic minerals occurs in the metavolcanic roof rocks of the intrusions as in the metamorphic rocks along the basal contacts of the plutons. Although the concept of lime metasomatism explains the formation of the basal metamorphic aureole, where volcanic rocks are not abundant enough to explain the great lateral continuity of the aureole, it cannot be applied directly to the origin of the metavolcanic rocks of the roof, for in these there is apparently no great change in the bulk composition of the rocks. Further study of the chemical composition of the metavolcanic rocks is required to evaluate the relative importance of thermal metamorphism and contact metasomatism.

Igneous Rocks

A generalized columnar section of the North Arm Mountain pluton is shown in Figure 2. This is typical in part of all the major plutons of the Complex and is, in a sense, the 'type section' of the Complex. It shows that the plutons are layered bodies with thick ultrabasic zones overlain by thick gabbro zones. Their basal contacts are in general conformable with the underlying sedimentary rocks. Their roof contacts are, to a minor extent, crosscutting. Interbanded contacts separate the main gabbro zones from the main ultrabasic zones. These interbanded zones contain various rock types, composed of olivine, clinopyroxene, and plagioclase in various proportions.

The Table Mountain pluton has a similar cross-section, but faulting and erosion have removed much of the gabbro zone. Although it now consists almost wholly of ultrabasic rocks, this was probably not the case at the time of emplacement. The Blow Me Down Mountain section is identical with that of North Arm Mountain, but the roof zone was not completely outlined. In the Lewis Hills, many of the smaller gabbro areas are composed entirely of interbanded gabbroic and ultrabasic rocks, typical

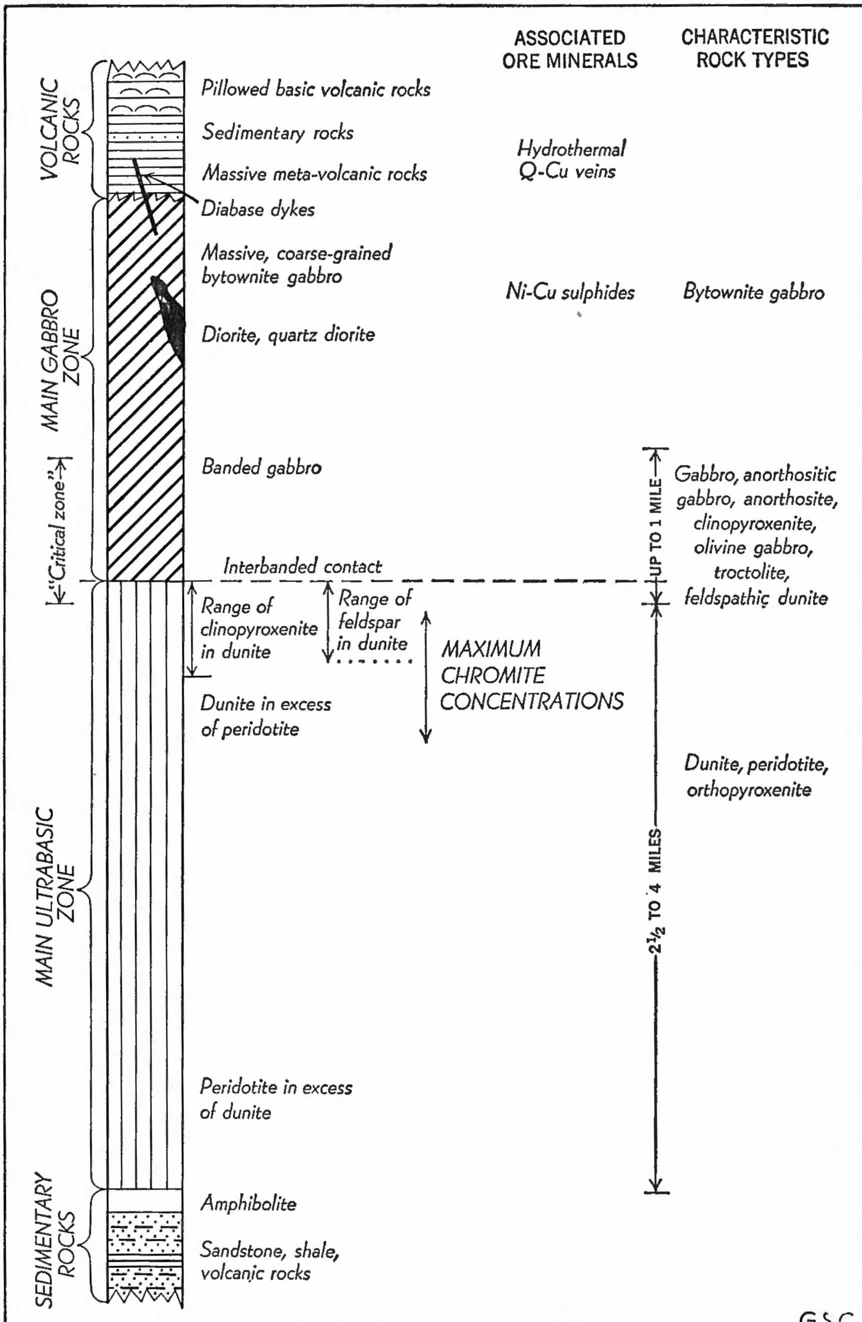


Figure 2. Generalized columnar section of North Arm Mountain pluton (not drawn to scale).

of the interbanded contact zone. It is only towards the western part of the Lewis Hills area that metavolcanic roof-type rocks are found, and the gabbro-metavolcanic relations in this area are still not solved.

ULTRABASIC ROCKS

Each pluton has a lower or basal zone composed of ultrabasic rocks. These zones are from 2½ to 4 miles thick. There are also numerous small detached masses of ultrabasic rocks or serpentinite. For instance, small ultrabasic layers, commonly feldspathic, are associated with the Mount St. Gregory gabbros, and sheared serpentinite bodies are found along faults marginal to the plutons and within the sedimentary rocks.

The major ultrabasic zones are underlain by narrow aureoles of metamorphosed sedimentary and volcanic rocks, and are overlain by banded and massive gabbroic rocks. In contrast to the Bushveld, Stillwater, and other more typical gravity-stratified sheets, there is no chill zone of gabbroic composition against the metamorphic rocks. Relic grains of olivine and enstatite show that the primary ultrabasic rocks near the contact were coarse grained and had the same composition as the ultrabasic rocks in the

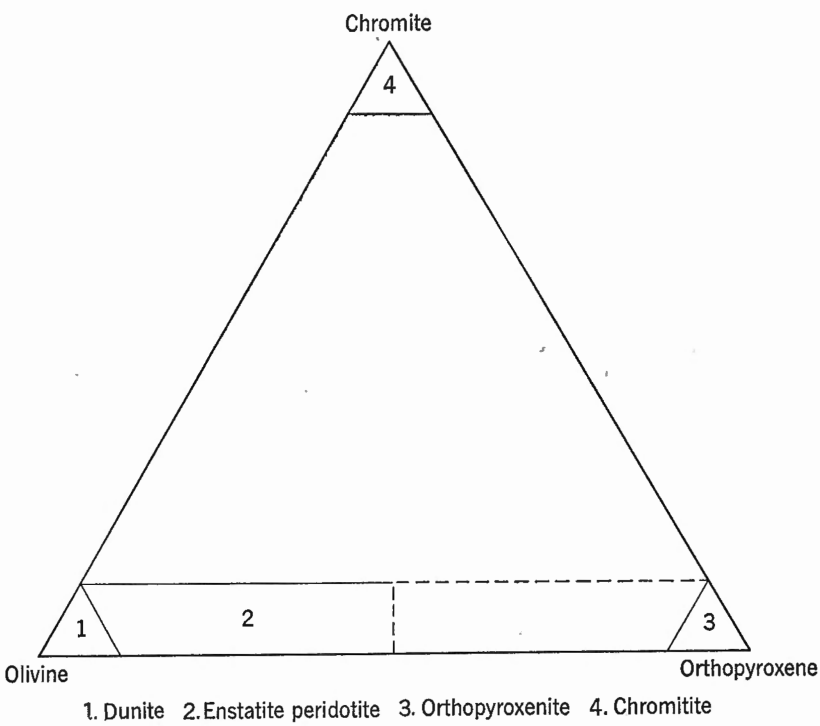


Figure 3. Principal rock types of the ultrabasic zone.

central parts of the masses. This olivine probably crystallized prior to its arrival at the contact. In places in the upper parts of the ultrabasic zones, hydrogarnet-bearing dunites form discrete but discontinuous layers that are most abundant near the gabbroic rocks. Ultrabasic layers interbanded with gabbroic rocks are less abundant in upper parts of the section.

It is not known if the basal zones of all the plutons were part of a single sheet because a 14-mile gap separates the Blow Me Down and North Arm Mountain plutons. However, it is probable that the Table Mountain and North Arm Mountain plutons are the faulted segments of a formerly continuous pluton (*see* Figure 9). The Blow Me Down and Lewis Hills plutons may also be the faulted parts of a single intrusive body but this is more difficult to prove.

The ultrabasic rocks have a distinct, banded structure that is fairly consistent in attitude over great distances. The banding consists of variations in the relative amounts of olivine, pyroxene, and chromian spinel. There are, however, no distinct variations in the chemical composition of the individual minerals. The banding is conformable with the basal contact and the bedding of the underlying sedimentary rocks. It generally follows the north-northeasterly regional trend of the area, and dips 35 to 70 degrees to the northwest.

The mineralogical composition of the rock types of the ultrabasic zones is shown in Figure 3. Only the fields represented by common rock types are labelled. These are defined as follows:

Dunite........................	over 90% olivine
Enstatite peridotite (harzburgite or saxonite)................	less than 90% olivine, the remainder dominantly orthopyroxene
Orthopyroxenite (enstatolite)......	over 90% orthopyroxene (enstatite)
Chromitite.....................	over 90% chromite (rare)

The compositional fields in Figure 3 do not correspond to any natural physical-chemical boundary between the rock types. All the dunite, for example, contains at least a little chromite and enstatite. When the content of enstatite rises to 10 per cent or more, it is convenient to signal this increase with another name and map unit, but the value of 10 per cent has no unique petrologic significance. Other investigators have used the arbitrary value of 5 per cent as the cut-off between dunite and peridotite.

Serpentine minerals derived from olivine and enstatite form 60 to 99 per cent of the ultrabasic rocks. Some secondary magnetite is a product of serpentinization. Clinopyroxenes occur mainly in the upper parts of the ultrabasic zones, and even there they are relatively uncommon.

Forsterite is the most abundant primary mineral in the ultrabasic rocks, forming dunite or, where the enstatite content is over 10 per cent, enstatite peridotite (harzburgite or saxonite). The distinction between dunite and enstatite peridotite is sometimes difficult to make in the field as the enstatite content commonly hovers about the 10 per cent dividing line and small variations cannot be distinguished with the naked eye. Otherwise the dunite can be distinguished from peridotite by differences on weathered surfaces. Dunite weathers to a light yellowish brown, smooth surface, whereas peridotite weathers to a slightly darker, stucco-like surface caused by resistant enstatite and chromite grains standing in relief. For detailed field work a celluloid plate on which 100 squares are ruled aids estimation of the enstatite content. The plate is placed on a weathered surface and the number of squares opposite enstatite grains are counted. This gives a minimum figure for the enstatite content.

With the increase in enstatite content, the rock may be classified as an orthopyroxenite (enstatolite). Orthopyroxenite is less common than peridotite and dunite. It occurs as bands and dykes rarely over several inches thick. Similarly chromitite, a rock composed dominantly of chromite, is a relatively unimportant rock type of the ultrabasic zone.

The well-exposed areas of ultrabasic rocks show no systematic chemical variations (i.e. cryptic layering) of primary mineral components. Although altered calcic plagioclase (hydrogarnet) and diopside occur near the gabbro zones, the main body of ultrabasic rocks is composed of the primary minerals; olivine (forsterite), orthopyroxene (enstatite) and brownish, translucent to opaque chromium-bearing spinel. In the classic areas of gravity-stratified sheets an evolution in the composition of mineral components is common, the olivine and orthopyroxene becoming more iron-rich towards the top of the pluton. In the Bay of Islands Complex, the ultrabasic rocks as a unit are chemically homogeneous. They consist mainly of forsterite and enstatite over large 'stratigraphic' thicknesses. The variation in rock types has the character of simple mechanical mixing, in varying proportions, of olivine and enstatite themselves of uniform composition. The mineral components of the ultrabasic zone are of uniform composition and fail to reflect the overall gravity stratification suggested by the presence of gabbroic rocks overlying ultrabasic rocks.

Dunite

Only general statements can be made regarding the relative restriction of dunite or peridotite to certain parts of the ultrabasic zone because dunite is interbanded with peridotite over wide areas. An unexplained feature is the predominance of dunite in the upper parts of the ultrabasic zone, near the banded gabbros. This is particularly evident in the Blow Me Down pluton, where the ultrabasic zone can be divided into a lower part composed dominantly of peridotite, a middle part composed of peridotite and dunite, and an upper part composed dominantly of dunite. This sequence is not as well shown in the North Arm pluton, where dunite occurs with peridotite in the lower part of the intrusion. Dunite is most abundant in the Lewis

Hills pluton, possibly because it has not been tilted sufficiently to reveal completely an underlying, more peridotitic zone as is found in the other plutons.

Dunite occurs in the form of bands and crosscutting dykes. Individual bands vary in thickness from a fraction of an inch to hundreds of feet. Correlation of separate bands is difficult even for short distances because of their tendency to pinch out or pass laterally into peridotite. The stratigraphic sections of Ingerson (1935, pp. 432-433) show dunite bands localized at the bottom and top of the ultrabasic zones of the northern plutons. No such continuous basal dunite layer is present because the proportions of olivine to enstatite varies laterally. Indeed, the basal layer is commonly a porphyroclastic peridotite. The true thickness of the upper dunite zone in any one of the plutons is not determinable because of the interbanded nature of the contacts, lateral variations between peridotite and dunite, and because the zone is offset in places by transverse faults.

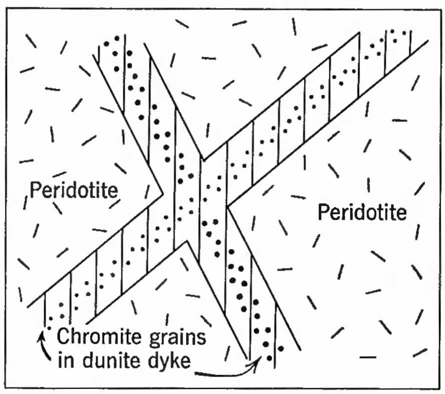


Figure 4. Sketch showing age relationships of crosscutting dunite dykes.

Dunite also occurs as thin crosscutting dykes. These dykes are approximately an inch wide. They cut enstatolite bands, and in places are cut by enstatolite dykes. There is no evidence of chilled contacts or bleaching of the wall-rocks near the dykes. The relative ages of small dunite dykes can be locally determined by such crosscutting relationships as those shown in Figure 4. This figure shows an earlier dunite dyke with a central parting of coarse chromite, cut by a later dunite dyke with a central parting of fine chromite. The chromite crystals show a preference for the centre of such dykes regardless of their attitude. They are analogous to the central magnetite partings found in serpentine veins, and their mode of origin may be similar. The dunite dykes are, in places, branches of the main dunite bands, and form connecting links between adjacent bands (*see* Figure 7).

Dunite with a low enstatite content weathers to a smooth, even-grained surface on which only a few grains of black, lustrous chromite stand out in positive relief. The absence of enstatite knobs makes the dunite easily recognizable in the field. Its weathering colour is generally a yellowish orange, slightly lighter than that of the peridotite. Sheared dunites adjacent to the basal contacts of the plutons weather to a dark greenish grey colour. Surface weathering extends to a maximum depth of 1 inch or 2 inches, and is commonly less than ½ inch in depth. Below the weathered zone the fresh dunite has a greyish olive-green to black colour and an aphanitic appearance.

The dunite is a monomineralic rock originally composed almost entirely of olivine, with very minor amounts of enstatite and chromite, but is now largely altered to serpentine minerals. Many of the rocks here classified as dunite would be called serpentinite in a more rigid, lithological classification. However, the texture of the serpentine aggregates derived from olivine is sufficiently distinctive to indicate the primary composition of the rock and to allow rock names to be applied on the basis of the pre-serpentinization composition of the rock.

The olivine grains are subrounded and fractured, in contrast to the large enstatite grains that are irregular and deformed but are less intensely fractured. Serpentinization has affected the olivine more than the enstatite. Large enstatite grains stand out in contrast to the smaller olivine relics and serpentine minerals, and impart a pseudoporphyritic fabric to the rock. Uniform optical extinction of numerous adjacent relics of olivine is characteristic of much of the fabric of the dunite and, where present, is indicative of the lack of post-serpentinization deformation. It can thus be seen that the size of the primary olivine grains was about 4 mm. The olivine is colourless, and has a large optic angle and positive sign. Its indices of refraction ($N_x = 1 \cdot 652 \pm 0 \cdot 003$; $N_z = 1 \cdot 668 \pm 0 \cdot 003$) indicate that it is forsterite with an average composition of Fo_{92}. There is little variation from this composition throughout the entire Complex. Olivine was clearly the first mineral to crystallize from the magma, for it occurs as small rounded inclusions in chromite and enstatite. Enstatite is only a minor component of the dunite and is described below in a section on peridotite.

Chromite is an accessory mineral that forms about 1 per cent of the average rock in the ultrabasic zone. Small concentrations tend to occur in dunite rather than peridotite, and the chromite there has the unusual habit of occurring in thin streaks along the centre line of thin dunite bands. It is rarely euhedral, more commonly occurring as irregular masses interstitial to, or enclosing, olivine. Megascopically it is black and shiny, but in thin section it is yellow to reddish brown and translucent, with opaque rims in places.

The chromite is neither of uniform composition throughout the ultrabasic zone nor homogeneous in single grains. Variations in composition are not related to location within the pluton, but are controlled more by serpentinization and other late and post-intrusion changes. The concentrates from heavy liquid separations contain grains with different specific gravities and magnetic properties, so that it is difficult to assign the spinel to a specific variety of chromite. The term chromite is used here to include minerals also described as picotite and chromium-bearing spinel.

Serpentine minerals, which form 60 to 99 per cent of the dunites, have a characteristic mesh structure indicative of their derivation from olivine. Different morphological varieties of serpentine occur, but all have the general optical properties of antigorite. Optically, the serpentine ranges from a colourless type with extremely low birefringence to a slightly pleochroic type with X = light greenish yellow and Z = light green. Fibrous varieties are oriented transverse to the fractures they occupy, with their long axes at right angles to the boundary of the olivine grain they replace. Where serpentinization has gone to completion, a core of unoriented serpentine minerals marks the site of the original olivine grain. The fibrous serpentine terminates in the middle of each veinlet, which is the outline of the original grain or fracture boundary from which replacement started. This forms a structural break or medial fracture which, in the more serpentinized peridotites, is commonly occupied by a ribbon of magnetite grains. Magnetite is not present in the medial fractures of the less serpentinized rocks. This is probably due to the low iron content of the olivine, the ability of the serpentine minerals to incorporate a certain amount of iron into their structures, and the secondary migration of iron. Secondary magnetite is commonly deposited as scattered blebs, along the cleavage planes in bastite, in later crosscutting veins, and along the margins of chromite grains.

Enstatite Peridotite

Enstatite peridotite, also known as harzburgite or saxonite, is as abundant as dunite in the basal zones of the three northernmost plutons. It occurs as thin bands that alternate with dunite (*see* Plate VI A) or as bands that are several hundreds of feet thick. It is composed of the essential minerals olivine (forsterite) and orthopyroxene (enstatite), with the olivine content commonly in excess of 75 per cent.

Weathered peridotite surfaces are dusky yellow to moderate yellow-brown. Light green to bronze, platy enstatite grains and black chromite grains form knobs on these weathered surfaces. On fresher surfaces the peridotite is pseudoporphyritic, and consists of green lustrous bastite or partly serpentinized enstatite set in a greenish black, aphanitic groundmass of serpentine.

The olivine of the peridotite is identical in composition to that of the dunite, i.e., it contains 5 to 8 per cent Fe_2SiO_4. It is commonly enclosed in enstatite or in chromite. Average enstatite grains are 2½ mm. in length and in places are as large as 5 mm. The mineral is partly or completely altered to serpentine with a bastite structure. The characteristic 010 parting of the enstatite is preserved in the bastite. Undulose extinction is commonly observed and may be related to pre-serpentinization phenomena because such grains are surrounded by olivine relics that maintain their optical continuity. The enstatite is colourless, and has a moderate axial angle and low birefringence. Its Z index, which is readily determined by oil immersion as it parallels the cleavage, varies between the limits of 1·664 to 1·672. According to the tables of Hess and Phillips (1940) this indicates a composition of $En_{93\text{-}97}$. The small variation in composition bears no relation to position in the intrusion.

Yellowish to reddish brown chromite occurs as anhedral grains (*see* Plate X A and B), fractured and cut by serpentine minerals. The chromite anhedra enclose olivine and are themselves enclosed by enstatite. The paragenetic sequence in the peridotite is:

Olivine.....................enclosed in chromite and enstatite
Chromite...................encloses olivine, interstitial to enstatite and enclosed in enstatite
Enstatite...................encloses olivine, and chromite

No chemical analyses of the ultrabasic rocks were made for this study, but the previously published analyses given below show their high magnesia and low lime and alumina content.

Table I

Chemical Analyses of Ultrabasic Rocks of the Basal Zones

	1	2	3	4
SiO_2	39·08	39·45	39·26	33·91
Al_2O_3	1·11	1·27	1·82	0·60
Fe_2O_3	4·81	4·06	4·27	4·79
FeO	2·87	3·88	3·02	3·23
MgO	37·80	37·01	38·53	41·61
CaO	0·31	0·16	1·25	0·05
Na_2O	0·00	0·04	0·23	0·13
K_2O	0·00		—	—
H_2O+	12·95	13·31	10·05	14·08
H_2O-	0·49	0·34	0·58	0·53
TiO_2	0·04	0·04	0·04	tr
P_2O_5	tr	tr	—	—
Cr_2O_3	0·23	0·42	0·39	0·60
MnO	0·10	0·21	0·11	0·11
CO_2			0·47	0·16
	99·79	100·19	100·02	99·80

(1) Basal margin of North Arm Mt. pluton. Analyst: A. H. Phillips (Hess, 1938).
(2) Serpentinite in Humber Arm sandstone, Liverpool Brook. Analyst: A. H. Phillips (Hess, 1938).
(3) Peridotite, one mile northeast of Hines Pond, Lewis Hills. Analyst: A. Willman (Cooper, 1936).
(4) Dunite, head of Rope Cove Canyon, Lewis Hills. Analyst: W. Willman (Cooper, 1936).

Orthopyroxenite

Both orthopyroxenite and clinopyroxenite are found in the Bay of Islands Complex. Clinopyroxenite occurs near the contact between the gabbro and ultrabasic zones (*see* Figure 15). It is interpreted as a reaction product rather than as a true primary member of the ultrabasic zone and is described below (p. 38) together with the related phenomena observed along the gabbro-ultrabasic contact.

Orthopyroxenite (enstatolite) bodies are small and not as abundant as peridotite or dunite. They form dykes and bands that rarely exceed a few inches in thickness and have no continuity along strike. The dykes may pinch out, join orthopyroxenite bands, or cut across orthopyroxenite bands. Enstatite is the essential mineral component, forming up to 98 per cent of the rock. Its average composition is $En_{94}(N_z = 1 \cdot 670)$, similar to that in the peridotite. Minor amounts of chromite and olivine are commonly enclosed in the enstatite.

The orthopyroxenite dykes are primary features formed during the consolidation of the ultrabasic rocks. Bowen and Tuttle (1949, p. 460) explain the formation of crosscutting pyroxenite dykes as due to SiO_2 — saturated water vapour streaming through cracks in peridotite and converting the peridotite to pyroxenite. It is probable that most of the orthopyroxenite formed in this way, during cooling of the ultrabasic rocks.

Ultrabasic Foliates

Post-serpentinization dynamic metamorphism has formed local areas of porphyroclastic peridotite and dunite with a well-developed foliation. These rock types occur near the basal contacts of the plutons, and among the altered ultrabasic rocks near the Big Level, Lewis Hills. They are also developed in places along the large transverse faults, although slickensided serpentinites are more common along major fault zones. Good exposures occur within 10 to 20 feet of the basal contact of the North Arm Mountain pluton, where dunite foliates have a distinct fissility and resemble black shales when viewed on an unweathered surface. Deformed peridotites resemble augen-gneisses (Plate IV A), with sinuous foliation and eyes of enstatite. Differential weathering accentuates this structure, and causes enstatite layers to stand in relief above serpentine layers.

A sequence of increasing mechanical deformation may be traced in thin sections of the porphyroclastic foliates along the basal contact of the North Arm Mountain pluton. In the first stage of deformation, serpentine is restricted to definite layers, between layers of granular olivine and pyroxene. Enstatite grains form porphyroclasts up to ½ inch in size, with undulose extinction and a well-developed 010 cleavage. In places, they are sheared along their cleavage planes, so that they resemble the traditional deck of cards used to illustrate shearing phenomena. The margins of the enstatite porphyroclasts are outlined by rims or mantles of granular olivine. The individual olivine grains do not show oriented extinction and are not separated by areas of serpentine. They form cataclastic aggregates concentrated at the ends of ellipsoidal pyroxene porphyroclasts. Apparently the serpentine has flowed plastically around large enstatite crystals which acted as buttresses during deformation. As the serpentine flowed, it carried the smaller olivine residuals with it, piling them up against the enstatite buttresses.

In more metamorphosed zones, denser rocks have formed due to the squeezing out of the serpentine. Well-foliated crystalline ultrabasic rocks are interbanded with hornblende gabbro in the Lewis Hills. Thin sections show that they consist of layers of granular olivine and pyroxene with knots of large enstatite grains and only minor amounts of serpentine, concentrated along a few distinct planes. They greatly resemble the cataclastic dunites from Anita Bay, Milford Sound, New Zealand, described by Turner (1942) and the dunite-mylonites of St. Paul's Rocks on the Mid-Atlantic Ridge (Tilley, 1947).

The fabric of the ultrabasic foliates is significant in two respects. First, it resembles the primary banding of the ultrabasic rocks, which is described in more detail below, but the secondary foliation is more sharply defined as shown by the fracturing and weathering of the rocks. The main distinction between the two fabrics is due to the conditions prevailing at their time of origin. The primary banded structure developed by magmatic processes at higher temperatures favourable to both flow and recrystallization. The secondary planar structure developed at lower temperatures after the ultrabasic rocks were serpentinized. The extreme plasticity of the serpentine allowed flow to take place at a temperature too low to allow recrystallization. Evidently the foliates of the basal contacts are related to shearing stresses localized along the contacts during tilting of the ultrabasic plutons.

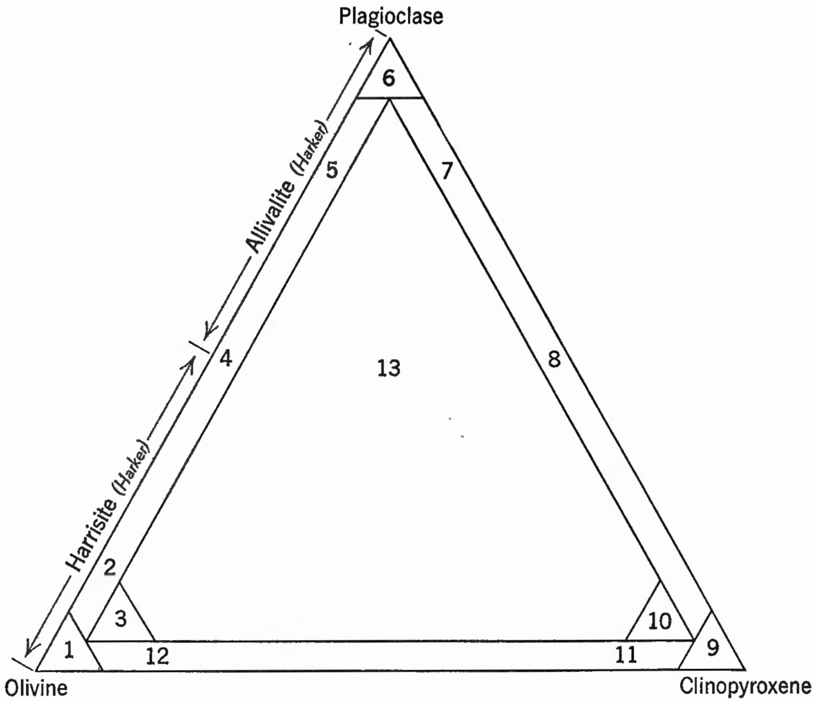
The second significant feature of the deformational fabric is the evidence it provides on the manner in which olivine crystals may accumulate during deformation, by the gradual elimination of serpentine. The end product is a mechanical differentiation which results in the formation of dunite mylonites on the one hand and serpentinites on the other. Under extreme deformation, such as that to which ultrabasic plutons are subjected during orogeny, this may lead to the formation of minor secondary ultrabasic apophyses intrusive into younger rocks, and thus make the primary ultrabasic bodies appear younger than they are.

INTERBANDED ROCKS OF THE GABBRO-ULTRABASIC CONTACT

A great variety of gabbroic and ultrabasic rock types are interbanded along the gabbro-ultrabasic contact of the major plutons. In the larger interbanded areas, such as those that border the North Arm ultrabasic rocks, the ultrabasic bands decrease in abundance and the gabbroic rocks become more massive away from the main ultrabasic body. Some of the smaller gabbroic areas shown in the Lewis Hills pluton are composed entirely of interbanded gabbroic and ultrabasic rocks. Along their strike the banded gabbro zones grade into feldspathic dunites and finally into normal dunite. In the Mount St. Gregory area, and along the coast north of Chimney Cove, are banded gabbros with small ultrabasic layers. The ultrabasic layers are more abundant in the western parts of the area. This suggests that they constitute a 'stratigraphic horizon' comparable to the gabbro-ultrabasic contacts of the major plutons, and that they have been separated from related, larger, ultrabasic zones by faulting. These interbanded zones are comparable in part with the critical zone of the Bushveld Complex (Hall, 1932), the banded zone of the Stillwater Complex (Peoples, 1933), and the ultrabasic complex of the Isle of Rhum (Harker, 1909).

Primary banding, from a fraction of an inch to hundreds of feet thick, is the outstanding feature in this zone. The abrupt alternations between white feldspar, black pyroxene, and brown olivine (serpentine) bands are very distinctive in the field. The dip of the bands tends to be steeper than that in the ultrabasic rocks and reversals of dip are common. Because of the lenticular nature of individual layers (*see* Figure 7), their lateral variations in mineral composition and thickness, their structural complexity, and the tendency for only the more resistant bands to outcrop in the inter-stream areas, it is not possible to set up a simple stratigraphic column, as attempted by Ingerson (1935). The bands are individual rock types of limited extent rather than continuous rock units.

Stated in simple terms, the rocks of the interbanded zone are composed of three principal primary minerals, whose abundance in different bands may vary from 0 to 100 per cent. These three essential minerals are plagioclase (bytownite), clinopyroxene (diopside to augite) and olivine (forsterite to chrysolite), and the various rock types that they may form are indicated graphically in Figure 5. In addition to the minerals listed above, orthopyroxene may locally become an important constituent, and small amounts of chromite, magnetite and sulphide minerals are present as primary components. From a limited amount of work done on the mineral components, no systematic or significant variation in the chemical composition of the mineral species was recognized. The variation of rock types appears to represent mainly a mechanical mixture of the mineral components in various proportion, with chemical variations being of lesser importance and randomly distributed.



1. Dunite
2. Feldspathic dunite
3. Feldspathic wehrlite
4. Troctolite
5. Anorthositic troctolite
6. Anorthosite
7. Anorthositic gabbro
8. Gabbro
9. Clinopyroxenite
10. Feldspathic pyroxenite
11. Olivine clinopyroxenite
12. Wehrlite
13. Olivine gabbro

Figure 5. Rock types of the interbanded zone.

The olivine-rich bands of the interbanded contacts are similar to the dunite bands of the main ultrabasic zone. Rock types representative of the series, feldspathic dunite-troctolite-anorthositic troctolite-anorthosite, occur where feldspar is more abundant. In the terminology of Harker (1908), these rock types are known as harrisite and allivalite. Rocks of this composition are confined to the interbanded contacts of the plutons.

Feldspathic Dunite

The presence of feldspathic dunite (a variety of harrisite) signifies an increase in the lime-alumina content of the ultrabasic rocks. It occurs interbanded with normal dunites and gabbro near the top of the ultrabasic zone. In the Lewis Hills pluton, banded gabbro zones grade along strike to feldspathic dunite, and then into dunite. As the compositional changes are not confined to vertical sections, they are not due to the influence of gravitative differentiation alone.

The term feldspathic dunite is used here in a genetic sense, and the strictly lithological term 'hydrogarnet dunite' is more correct. These rocks resemble normal dunites except for the presence of interstitial masses of white alteration products that form about 5 per cent of the rock. The secondary aggregates are composed principally of hydrogarnet (a member of the series $3CaO.Al_2O_3.3SiO_2$–$3CaO_2.Al_2O_3.6H_2O$) but, as a few remnants of plagioclase are associated with the hydrogarnet, it is logical to conclude that the hydrogarnet formed from plagioclase and that the rock was originally a feldspathic dunite. This interpretation is supported by the similarity in chemical composition between the hydrogarnet and the plagioclase, and by the results of experimental hydrothermal studies wherein hydrogarnets have been formed from plagioclase. C. O. Hutton (1943) pointed out that the mineral previously reported as grossularite in many descriptions of garnetiferous gabbros associated with basic and ultrabasic rocks was actually a hydrogarnet with a lower index of refraction than that of normal garnets. He proposed the term "hydrogrossular" for "all the members of the garnet-hydrogarnet series with a composition intermediate between plazolite and grossularite". Index determinations on the aggregates from the Bay of Islands dunites indicate that some of the hydrogarnets are beyond the hydrogrossular range and closer to the tri-calcium aluminate hexahydrate end-member in composition. For this reason only the general term hydrogarnet is used here, even though hydro-grossular may be the more common type.

Feldspathic dunites are distinguished in the field by the presence of small, white, resistant knobs on the yellowish brown weathered surfaces, and by small, white flecks of hydrogarnet on the greenish black fresh surfaces. In those rocks in which white hydrogarnet is abundant, fresh feldspar is also common and the rock type grades into troctolite. The surface of the more feldspathic facies is extremely rough and corrugated. Although fresh, brownish to colourless, isometric hydrogarnets may form large crystals, more typically they occur as aggregates of small crystals associated with a colourless, low birefringent mineral of low positive relief (chlorite?). The refractive index of the hydrogarnet ranges from $N = 1.59$ to 1.68, and the better developed crystals have the higher indices. All specimens, regardless of index, show hydrogarnet structures in X-ray powder patterns. The powder patterns are similar but not quite identical to that of hibschite. The cell dimension is $A_o = 12.12 \pm 0.25$. The hydro-garnet aggregates are interstitial to other silicate minerals and it is possible that the alteration is not in the form of simple replacement but that some solution, migration and redeposition of lime has taken place. Apparently the formation of hydrogarnet is due to the alteration of calcic plagioclase during serpentinization of the ultrabasic rocks.

The olivine in the feldspathic dunites is of the forsterite variety. Small amounts of clinopyroxene are also present.

Troctolite

The dividing line between feldspathic dunite and troctolite is not clearly defined in the literature and, as there was no necessity to differentiate them precisely on the scale of the present mapping, an exact dividing line was not adhered to in the field. In general, olivine-bearing rocks with up to 5 or 10 per cent plagioclase were called feldspathic dunite, and those with more than 10 per cent plagioclase and little or no clinopyroxene were called troctolite. The plagioclase of the troctolite is bytownite near An_{75} in average composition. The olivine is slightly higher in iron than that of the main ultrabasic zone and is a variety of chrysolite. It is interstitial to and, hence, later than the plagioclase. In the feldspathic dunite, on the other hand, the olivine appears to be earlier than the interstitial plagioclase. Where clinopyroxene is present, it occurs as either separate interstitial grains or as irregular rims between plagioclase and olivine. It is pinkish in colour and has a well-developed diallage parting. The clinopyroxene is in the diopside-augite compositional range. Irregular areas of a brown pleochroic amphibole rim the pyroxene grains in places and represent reaction products between the pyroxene and adjacent plagioclase grains.

Coronas, typical of the olivine-plagioclase rocks of many areas, are only partially developed in the Bay of Islands troctolite. Clinopyroxene occurs in places between olivine and plagioclase grains, suggesting that a limited reaction has taken place between these minerals. Commonly, however, no reaction zone exists, and instead an 'expansion fabric' is well developed (*see* Plate V). This consists of fractures that radiate from a serpentinized olivine nucleus into the surrounding plagioclase. These fractures show no relation to the crystallographic directions of the plagioclase, and either die out away from the olivine nucleus or join the fractures radiating from another olivine grain. Evidently as olivine altered to serpentine it expanded and caused fracturing of the adjacent plagioclase. The fabric indicates that serpentinization of the olivine has been an automorphic effect, unrelated to outside influences, such as younger intrusions, and is evidence favouring the idea that the serpentinization of the main ultrabasic bodies has also been an automorphic process. The age of the shattering relative to the development of secondary brown amphibole was not clearly determined. In some places, the fractures appear to cut the amphibole, but in others the amphibole appears to have crystallized along the fractures.

Osborn and Tait (1952) in their study of the system diopside-forsterite-anorthite have shown that "at some moderate temperature (below about 900 degrees) an accumulation of calcic plagioclase and magnesian olivine crystals becomes a metastable assemblage which will, if conditions are appropriate, transform to the low temperature stable assemblages of pyroxene, spinel and plagioclase, or pyroxene, spinel and olivine". The plagioclase-olivine rocks of the Bay of Islands Complex are metastable assemblages in which clinopyroxene has developed to only a limited extent. The preservation of this assemblage at lower temperatures is possibly due to rapid cooling under stress. In such an environment, reaction between olivine and plagioclase would be restricted because the interstitial fluid in which it takes place would be squeezed out of the crystal mush. Further evidence for the role of stress during crystallization is the development of banded structures in the rocks and the absence of well-developed coronas, such as are found in the troctolitic rocks of other areas. It is noteworthy that rocks of this composition in the Bay of Islands area are restricted to the zones in which banding is well developed and do not occur in the massive gabbroic rocks. Suggested courses of crystallization are outlined in Figure 6.

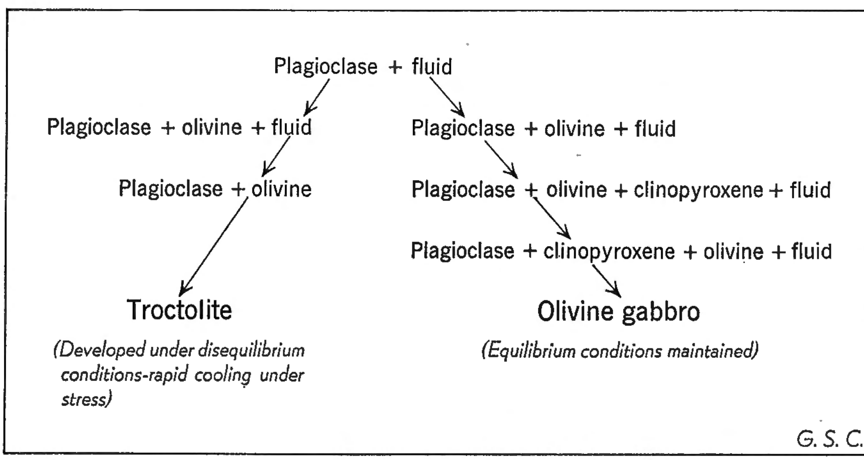


Figure 6. Alternate lines of descent in the origin of troctolite and olivine gabbro.

Anorthosite

Anorthosite layers are thin, small in lateral extent, and restricted to the banded zones. Anorthosite is rarely exposed in the areas of massive leucogabbro that outcrop farther away from the gabbro-ultrabasic contacts. It is composed wholly of bytownite similar in composition to that of the massive gabbroic rocks.

Clinopyroxenite

Small bands of clinopyroxenite are exposed in places near the base of the banded zone or in the upper parts of the ultrabasic zone. In the North Arm Mountain pluton clinopyroxenite forms a thick, lens-like member over 3 miles long, but is irregular in detail, as it consists of a group of small pyroxenite lenses in dunite and is discordant with the main gabbro contact (*see* Figure 15). At the southwest end of North Arm Mountain it projects into the ultrabasic zone and pinches out near the Stowbridge chromite deposit. Similar bodies occur here and there along the base of the Table Mountain gabbroic zone and among the banded rocks of the Lewis Hills pluton.

The clinopyroxenite is greyish olive, with a granular appearance on weathered surfaces. The stubbly, crystalline forms appearing on weathered surfaces stand out in marked contrast with the smooth surfaces of the dunite with which it is interbanded. It is a medium- to coarse-grained rock that consists almost wholly of clinopyroxene. The clinopyroxene has the following properties; colourless, well-developed 100 parting, extinction angle of 40 degrees, $2V = 57 \cdot 5$ degrees, $Ny = 1.675$. Its composition, following Hess (1949), is $Ca_{48}Mg_{49}Fe_3$, i.e. practically a pure diopside. A small amount of olivine is invariably present. It may be partly or completely altered to serpentine, whereas the pyroxene is relatively unserpentinized.

The distribution and origin of clinopyroxenites in layered basic-ultrabasic bodies is a problem of interest. In the Bay of Islands Complex clinopyroxenites occur only near the main gabbro-ultrabasic contacts, never in other parts of the main ultrabasic zones. However, they are not always present near the contacts for instead, in some places, extensive areas of feldspathic dunite occur along the contact zone. This locus of clinopyroxenite formation is even better demonstrated by other layered complexes. Zavaritsky (1932) describes a clinopyroxenite zone intermediate between ultrabasic and gabbroic rocks of the Rai-iz massif of Russia. A similar peridotite-dunite-pyroxenite-gabbro sequence is found among the ultrabasic rocks of the Eastern Townships of Quebec, although the structural relationships between the different rock types are not as clear. In the Unst ultrabasic body of the Shetland Islands, the sequence peridotite-dunite-clinopyroxenite occurs, but the gabbro fraction is not recognized (Amin, 1954). T. P. Thayer (1946) states that the Kenai, Alaska intrusions and those of eastern Oregon grade through pyroxenitic into gabbroic facies, whereas those of Cuba have an intermediate troctolite facies. Both of these facies are present along the contact zones in the Bay of Islands area. It is, therefore, established that where clinopyroxenites are found in layered ultrabasic bodies, they occupy a definite stratigraphic position intermediate between the ultrabasic and gabbroic rocks.

A possible explanation for this stratigraphic relationship is expressed in the equation:

$$CaO + MgO + 2SiO_2 \longrightarrow CaMg(SiO_3)_2 \text{ (diopside)}$$

It is probable that the formation of diopside is due to the reaction of lime from the gabbroic magma with magnesia from the ultrabasic mush. Another possible reaction in a similar chemical environment is

$$MgO + Al_2O_3 + Cr_2O_3 \longrightarrow MgO.\ (Al, Cr)_2O_3 \text{ (spinel)}$$

The relationship of this reaction to the development of spinel and, ultimately, chromite is discussed in Chapter V, where both field and experimental evidence are gathered to support the interpretation that these minerals originated by reaction between two zones of contrasting chemical composition during intrusion of a composite magma. It follows that both the clinopyroxene and the chromite can be interpreted as reaction products formed near the principal ultrabasic-gabbro contacts.

BANDING IN ULTRABASIC AND GABBROIC ROCKS

Physical Features

A banded structure characterizes many of the ultrabasic and gabbroic rocks of the area (*see* Plates VI and VII). The bands parallel the basal contacts of the plutons and the regional trends of the area, and are uniform in attitude over large areas. The banding is a primary, pre-serpentinization structure distinct from the secondary gneissic structure found along the margins of the ultrabasic bodies.

Rhythmic banding in the ultrabasic rocks is caused by a variation in enstatite content. Adjacent bands range in composition from dunite to peridotite and enstatolite (*see* Plate VI A). This is a purely physical relationship, as no chemical differences (cryptic layering) between the enstatite and forsterite of adjacent bands is apparent. A similar rhythmic banding occurs in the gabbroic rocks, where variations in the proportions of plagioclase, clinopyroxene, and olivine have resulted in the development of bands of contrasting composition and physical appearance (*see* Plate VII A). Variations in the plagioclase composition are small and very erratic and cannot be related to their position in the layered succession.

Individual bands may have either sharply defined walls (*see* Plate VI A) or gradational contacts. They range from a layer of single crystals to bands several hundred feet in thickness. Where banding consists of alternating layers of peridotite with different amounts of enstatite, it is not readily visible except where etched out by differential erosion. Elsewhere wisps of chromite grains may outline the banded structure. Folding of bands is found in places (*see* Plate VI B), especially in the enstatolite bands. These folds show the results of plastic flowage by their

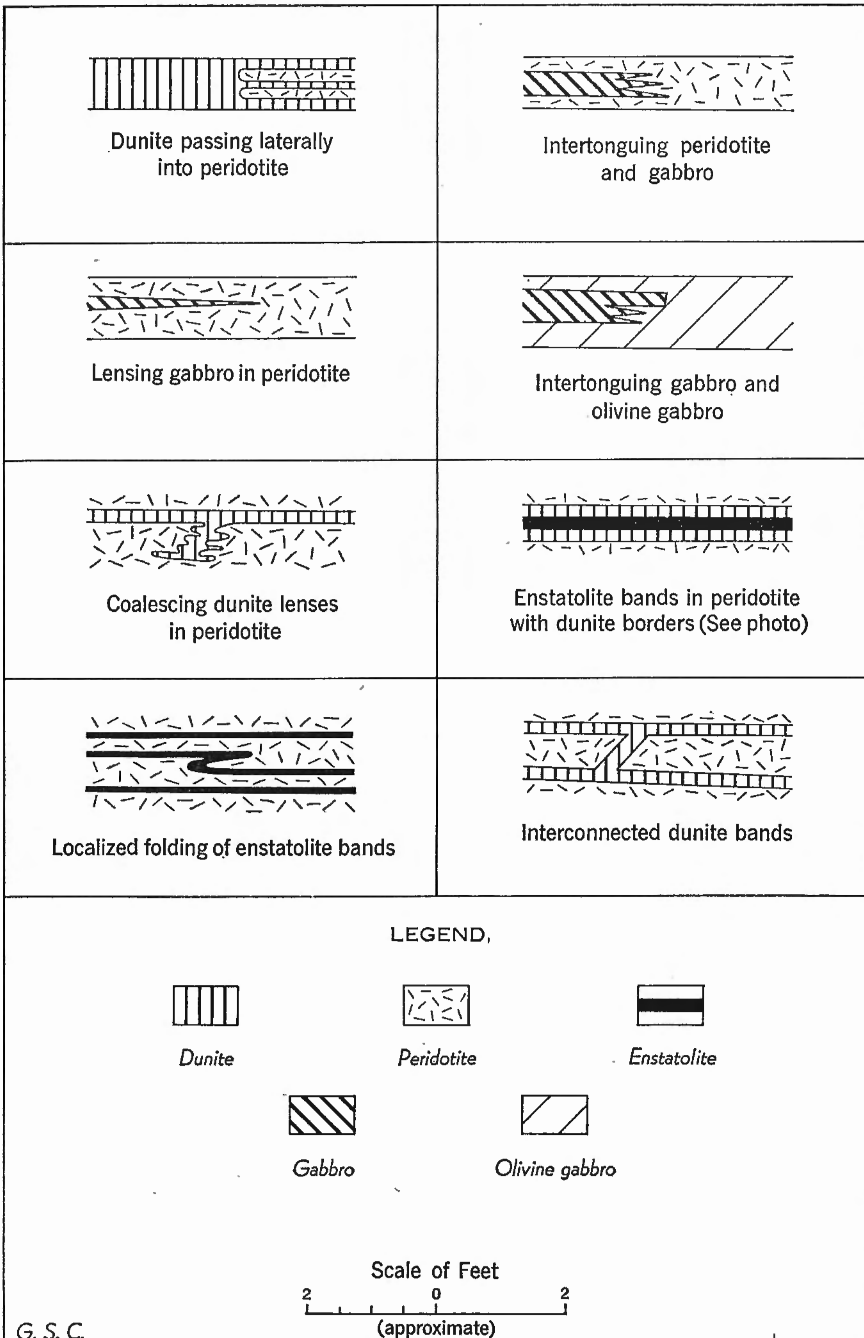


Figure 7. Small scale features of primary banding.

thickened crests and troughs, and are due to pre-consolidation movements of the magma. On vertical surfaces, isoclinally folded enstatolite bands may be seen paralleled by non-folded bands, which shows that folding was limited to small zones in the intrusion (*see* Figure 7) and probably related to different rates of laminar flow during emplacement.

The bands have an overall lenticular form although, in individual outcrops, some appear to have a uniform width. The smallest are merely lenses a few inches long; the largest are thousands of feet long. Some of the clinopyroxenite and chromite occurrences appear to have great lateral continuity but, actually they are aggregates of small lenses rather than single bands. Figure 7 illustrates common irregularities in the banding, showing the tendency for bands to pinch out and pass laterally into different rock types. This is present on a larger scale in the Lewis Hills pluton, where banded gabbroic zones can be traced laterally into feldspathic dunite and thence into dunite. Another feature illustrated in Figure 7 is the tendency for enstatolite bands to be bordered by dunite (*see also* Plate VI A). This 'segregation' effect is observed in both the parallel bands and in adjacent, irregular, pod-shaped bodies of small size. It is apparently due to the concentration of residual liquors along stress planes. The fluid removed from the walls leaves 'purified' dunite rims, and reacts with olivine in the centre to form enstatite along the site of the stress plane.

The orientation of the mineral components in the banded rocks of the area has not been determined, although in the gabbro a tendency toward the alignment of the 010 planes in plagioclase is apparent, but not pronounced.

Origin

There are two main theories for the origin of banded structures in basic igneous rocks:

(1) Origin through crystal settling or convection. This is based upon the similarity of banded structures with the bedding of sedimentary rocks.

(2) Origin by pre-consolidation flow. This is analogous to the formation of flow structures in volcanic rocks and along the margins of large granitic batholiths.

Grout (1918), Coats (1936), and Hess (1938) have explained banding in gabbroic rocks as the result of gravitative crystallization differentiation. Wager and Deer (1939) have developed a theory of origin based upon variations in the velocity of convection currents. Harker (1904), Zavaritsky (1932), Turner (1942) and Walton (1951) have explained certain types of banding in ultrabasic rocks as flow or shear structures formed in largely crystalline masses. Bowen (1928) has combined deformation of a crystallizing mass in which crystals are accumulating, auto-intrusion of liquid portions into rifts in the crystal mush, and gravitative differentiation in the smaller liquid layers to explain banded structures. In the Bay of Islands Complex there is no physical difference between the banding in the gabbroic rocks and that in the ultrabasic rocks. Therefore an explanation of the banding in the ultrabasic rocks should also explain the banding in the gabbros.

Despite the general acceptance of crystal settling or convection theories to explain the banding in gabbroic rocks, the general features of the Bay of Islands Complex appear to exclude this theory of origin. No known simple conditions of crystal settling explain the small scale inter-banding of early-formed olivine and late-formed enstatite without causing some variation in the chemical composition of the mineral components. An evolution in the chemical composition (cryptic layering) such as that found in the Bushveld and Skaergaard plutons, is the main argument for an origin of large scale banding by gravitative differentiation in situ. This chemical evolution does not occur in the Bay of Islands Complex. The chemical homogeneity of the thick ultrabasic zone implies that the banded structure was superimposed upon it as a unit and not by a systematic building up of crystal accumulations to a thickness of 4 miles.

Recent attempts to explain the banding of gabbros have appealed to the difference in specific gravity of calcic plagioclase (S.G. = 2.68–2.73) and augite (S.G. = 3.2–3.6). Coats (1936, p. 412) proposed the following mechanism which he called "rhythmic differential settling":

> "If two crystals of differing densities are settling in a liquid the density of which is but slightly less than that of the lighter sort of crystal, both varieties will settle toward the bottom. As these two kinds of crystal approach the bottom, the proportion of crystals to liquid will increase. When a certain limiting value is reached, since the sinking of the heavier crystals tends to displace the adjacent fluid upward, this liquid, because of its viscosity, and the slow rate of settling of the lighter crystals, carries them upward. There is thus produced a layer rich in the lighter crystalline constituent, over one rich in the heavier".

Hess (1938) presents a somewhat related theory of origin based upon short periods of turbulence in the magma and the effect of these periods on the differential settling of plagioclase and pyroxene. Wager and Deer (1939, p. 271) explain the layering of the Skaergaard intrusion as due to variations in velocity of convection currents. They state (p. 272) that:

> "Such variation combined with different rates of sinking of the crystals, because of their different specific gravities, should produce a winnowing effect on the crystals forming in the Skaergaard magma, and gravity stratified layering should result".

The fundamental basis for each of the above theories is that a difference in specific gravity exists between two mineral components, the plagioclase and the pyroxene. In attempting to apply these theories to the banding in the ultrabasic zone of the Bay of Islands plutons we are faced with the fact that virtually no difference in specific gravity exists between the two primary mineral components. Enstatite and forsterite have a specific gravity of 3.3, differing only in the second decimal place. Therefore theories dependent upon a difference in specific gravity between component minerals cannot explain the banding. As the banded gabbroic rocks occur in close association with the banded ultrabasic rocks, it is not logical to assume that they had a completely different mode of origin. Rather it is more probable that crystal settling did not play the principal role in the development of the primary rhythmic layering of either rock type.

Other theories of origin, based upon rhythmic cycles of crystallization of different components have been applied to the Bay of Islands Complex. Ingerson (1935, p. 437) attempted to explain the banded structure by the influence of the following factors on the normal course of crystal settling: (i) renewed injection of ultrabasic magma, (ii) dissolving of olivine from border facies, and (iii) concomitant extrusion causing relief of pressure. Cooper (1936) accepted this explanation, and in regard to variations of magma composition along the strike of the banding he stated (p. 46):

> "The problem is complex, involving the position of the feeding conduit, the depth of the chamber, tilting or other deformation during differentiation, possible volcanic action 'draining' parts of the magma chamber, etc."

No field evidence exists to confirm or deny the operation of these processes. Although they might explain several cycles of banding, it is highly improbable that they could explain the thousands of thin alternating ultrabasic bands of similar composition that are found through a thickness of 2 to 4 miles.

An alternative hypothesis, based upon pre-consolidation flow of the partly crystalline magma explains the banded phenomena more fully and is compatible with the location of these plutons in a tectonic belt. Flow layers are defined by Balk (1937, p. 15) as "tabular, disc-like rock bodies, composed essentially of those minerals that build up the surrounding rocks, but in different proportions". Flow structures have been described in volcanic rocks, and in the margins of plutonic rocks that range in composition from granitic to ultrabasic. They have also been produced experimentally by the deformation of plastic materials. The formation of banded structures in ice by glacier flow is particularly instructive (*see* Plate VII B). Meier, Rigsby, and Sharp (1954) describe flow foliation from the Saskatchewan glacier, Alberta, as alternating laminae of white bubbly ice and denser bluish ice that have formed parallel to the margins of the glacier. The distinction between sedimentary and flow structures is as much a problem of glacial geology as it is of petrology. Sharp (1954, p. 827) states that: "the arrangement, spacing, orientation, relation to flow direction, and relics of older structures strongly indicate that the foliation is developed in the ice by flowage independent of sedimentary layering". Detailed studies now being carried out on the mechanism of glacier flow should add to our meagre knowledge of the flowage of olivine crystal mushes. Preliminary studies (Meier, Rigsby and Sharp, 1954) show that recrystallization of the ice crystals may, to some extent, destroy the oriented pattern of crystals developed during flow.

A concise description of the role of stress in the formation of banded structure in ultrabasic rocks of the Rai-iz massif in the Arctic Urals is given by Zavaritsky (1932, p. 177): "the banded structure might most probably be ascribed to the stress which acted upon a nearly crystalline mass of olivine grains having contained a small residual acid solution. Under the influence of the stress it was irregularly distributed in bands in the mass of the crystallizing rock, and on its definitive consolidation this irregularity led to the formation of the banded structure. It is notable that the banding is developed also in the adjacent gabbro-amphibolites". Bowen and Tuttle (1949) have demonstrated that if water vapour, charged in SiO_2, streams through a crack in an olivine-bearing rock, pyroxenite is developed by reaction with the olivine. The presence of late-crystallized enstatite of uniform composition throughout the ultrabasic plutons, either in the interstices between olivine grains or in distinct bands outline the loci of late stage silica-rich solutions during the later cooling stages of the plutons. These solutions could have migrated along certain planes (flow planes) or may have been trapped between the olivine grains. In places they entered crosscutting fractures to form orthopyroxenite dykes similar to the orthopyroxenite bands. The concentration of late-stage fluids along definite parallel planes is due to the role of stress at the time of emplacement of the mass of olivine crystals. The attitude of the stress planes, parallel to the walls of the plutons, is similar to the attitude of the ultrabasic foliates. In these rocks the stress planes were developed at low temperatures and under conditions unfavourable to the crystallization of enstatite. The crystallization of enstatite along stress planes in the olivine mass may be compared to the development of laminae of white bubbly ice and denser bluish ice by recrystallization during glacier flow. Whereas, in the latter, only one mineral, ice, will crystallize, in an ultrabasic mush, falling temperature causes the formation of enstatite in the loci of the late-stage solutions. It is probable that these same late water-rich solutions would, with further decrease in temperature, produce partial serpentinization of the ultrabasic mass.

The origin of banding in gabbroic rocks has received more attention than the origin of banding in ultrabasic rocks and, the theory of origin by crystal settling has been most generally applied. However, as theories based upon specific gravity do not explain the banding in ultrabasic rocks, they probably do not explain the banding in the associated and conformable banded gabbroic rocks. An important feature of the gabbroic rocks is that they are well banded along their contacts with ultrabasic rocks, and massive away from these contacts. This is interpreted as an indication that the gabbroic phase of the intrusions was largely fluid at the time of emplacement with crystallization more advanced towards the bottom, or ultrabasic fraction, of the magma. Centres of crystallization of plagioclase and olivine, which were set up near the base of the gabbroic fraction, were streaked out into lens-shaped bodies by the flowing magma during the tectonic emplacement of the partly differentiated mass. Possibly some clinopyroxene also formed at this time. However, much of the clinopyroxene has developed through reaction of olivine and plagioclase and may have been generated by local reactions even after the banded structure was superimposed on the partly crystalline mass. Thus the clinopyroxene, in part at least, originated in a manner similar to that of the orthopyroxene of the ultrabasic rocks.

The above interpretation of the banded structure as due to flow is the most logical way of superimposing a banded structure on a thick, homogeneous ultrabasic unit. The homogeneity of the ultrabasic masses implies that the structure was imposed on the unit as a whole, and was not systematically built up by gravitative differentiation. The flow origin of the banding also explains the tapering of the ultrabasic zone (*see* Chapter III) as well as the interbanding of gabbroic and ultrabasic rocks and the lens-like anorthositic bodies in the ultrabasic rocks.

The distinction between flow-banding and true 'cryptic banding' is of great importance in the interpretation of the past history of a pluton. If the banded structure is interpreted as a gravity differentiation phenomenon, then the pluton was emplaced as a horizontal body. The Bay of Islands plutons, however, although originally gravity-differentiated masses at depth were later tectonically emplaced as wedge-shaped bodies (*see* pp. 84-86). The overall, gravity-stratified relationship (i.e. a gabbroic cap and an ultrabasic base) has been maintained, but the small-scale banded structure was developed during the final emplacement of the mass and, hence, is related to the movement of magma parallel to its confining walls, which were not necessarily horizontal. Although it is not definitely known if the plutons were horizontal, inclined, or vertical at the time of final emplacement, it is most probable, in view of the tectonic history of the masses, that they were finally emplaced in inclined strata during the deformation of the area. Whereas the attitude of the flow banding can be used to determine the attitude of the margins of the plutons, flow-banding in itself is not evidence that the plutons were originally horizontal bodies. Only the large scale relation of gabbro overlying ultrabasic rocks is evidence for gravity stratification. This latter relationship was developed early in the differentiation history of the plutons, however, and has only been preserved in a somewhat deformed state during their later emplacement history.

GABBRO

Rocks of gabbroic composition are as abundant as ultrabasic rocks in the Bay of Islands Complex. They occur as discrete layers in the interbanded zones or as large, massive to poorly banded bodies that form the western margins of, and partly overlie, the major ultrabasic zones. A relatively massive area of major dimensions underlies the Mount St. Gregory highlands, with its eastern margin interbanded with ultrabasic rocks. This gabbro is intrusive into the overlying metavolcanic rocks. A linear body of banded and massive gabbro with some ultrabasic layers forms Mount St. Gregory, and another strongly deformed body of similar composition forms the coastal region north of Chimney Cove. Coarse-grained gabbro and metagabbro also underlie large areas near Blow Me Down Mountain and outcrop along the western margin of the Lewis Hills.

The most common rock type is medium- to coarse-grained, light green leucogabbro, which locally grades to a melagabbro by increase in the clinopyroxene content, to anorthositic gabbro by increase in the plagioclase content, or to hornblende gabbro by secondary alteration. Except near the interbanded zones, the gabbro is commonly massive or poorly banded. It has rough, pitted weathered surfaces due to ridges of green and reddish brown pyrabole that stand out in relief against the white plagioclase.

The plagioclase of the gabbroic rocks occurs as well-twinned subhedral crystals that range from 2 to 4 mm. in greatest dimension. In the zone of interbanded rocks, the laths are imperfectly aligned, in contrast to their random orientation in the main gabbro zones. The plagioclase of the massive gabbro is typically bytownite, with no systematic difference from that of the interbanded zones. It varies from An_{70} to An_{80} in composition, with the majority of determinations falling between An_{70} and An_{75}. This variation is not related to distance from the main gabbro-ultrabasic contacts. It was not possible to relate mineralogical variations to position within the main areas of gabbro because banding is not well developed. In the few areas where banding is found, it is irregular and strongly divergent from that in the interbanded zones.

Augite commonly occurs as single grains or equigranular aggregates between plagioclase laths, or as poikilitic grains enclosing plagioclase. It is colourless or has a pinkish tinge, and a well-developed diallage parting.

It is biaxial positive with $2V = 49°$ to $51°$, and $Z \wedge c = 45°$ to $47°$. The average intermediate index of numerous crystals is $1{\cdot}695 \pm 0{\cdot}005$, which indicates a composition of $Ca_{40}Mg_{40}Fe_{20}$. This composition is quite different from that of the clinopyroxene of reaction origin found near the major ultrabasic contacts.

Olivine occurs in phases of the gabbroic rocks and contains more iron than that in the ultrabasic zone. It is an early mineral to form and is enclosed in plagioclase. It is locally serpentinized, but as a rule is better preserved than the olivine of the ultrabasic zone. In places it is as iron-rich as $Fo_{80}Fa_{20}$.

Magnetite is a common accessory mineral in the gabbro and is late in the crystallization sequence, but no concentrations of magnetite were found. It locally has reaction rims of brown hornblende against plagioclase. Similar brown hornblende has also formed by the alteration of augite. Prehnite occurs in thin late veinlets that cut the other minerals.

QUARTZ DIORITE

A northeasterly trending belt of hybrid rocks, with the approximate composition of quartz diorite, is exposed near the coast between Chimney Cove and Bonne Bay. South of Chimney Cove it is cut off by the Park fault. It may be correlated with a belt of granite described by Cooper (1936) at Lark Mountain on the south side of the Bay of Islands. This granitic belt occupies a similar structural position to the quartz diorite north of Chimney Cove. Other small, possibly related bodies of diorite and quartz diorite cut the coarse-grained gabbroic rocks of the Mount St. Gregory highlands. Dykes of quartz diorite composition are common near the Gregory River copper lodes, but no intrusions of this composition were found in the ultrabasic areas.

The rocks of the coastal belt are very impure. They are pink to grey, medium-grained rocks but are darker to the east where they contain numerous basic inclusions and are bounded by a zone of altered volcanic rocks. Altered plagioclase varies in composition from andesine to albite. Zoned, untwinned euhedral phenocrysts of albite are common. Interstitial quartz forms up to 40 per cent of the rock. Hornblende and epidote are abundant, and zircon, apatite, pyrite, and magnetite are accessory minerals.

At the northern end of the quartz diorite belt, near the Lookout Hills, the quartz diorite is intruded into flatly sheared, dark, fine-grained rocks that resemble the metavolcanic rocks of the Gregory highlands. These rocks are so intermingled that it was necessary to map them as a hybrid unit. The metavolcanic rocks increase in abundance to the north, toward the Lookout Hills.

BASIC DYKES

Porphyritic diabase dykes that cut the above intrusive and volcanic rocks are the youngest igneous rocks of the area. In the Mount St. Gregory highlands they form small and discontinuous bodies, rarely over two feet wide, that cut both the quartz diorite and the coarse-grained gabbros. They are most abundant in the Mine Cove area where a swarm of steeply dipping diabasic dykes cut both ultrabasic and volcanic rocks. This is the only place where diabase dykes cut the ultrabasic rocks. They appear to cross the flat contact between Humber Arm volcanic rocks and overlying ultrabasic rocks that has been interpreted as evidence for the Lewis Hills thrust in this area. Asbestos veins are formed in fractures in the ultrabasic rocks near the dykes. In thin section the dykes are seen to consist of large euhedral bytownite phenocrysts in a groundmass of augite, green hornblende, plagioclase and magnetite. Most are very altered, and contain secondary albite, epidote, penninite, prehnite, and magnetite, although their primary ophitic fabric is commonly retained.

Distinct from the porphyritic diabase dykes are coarser grained hornblende gabbro (bojite) dykes that cut the banded gabbroic and ultrabasic rocks of the Complex. One dyke of this composition occupies a northeasterly trending fault zone east of Rope Cove Canyon. Elsewhere in the Complex these dykes are small and discontinuous. They are coarse-grained to pegmatitic and consist of approximately equal amounts of black hornblende crystals up to 2 inches long, and white plagioclase. The hornblende is pleochroic in green and brown and has an intermediate index of 1·665. The plagioclase is bytownite.

Post-Intrusion Hydrothermal Action near Ultrabasic Rocks

The action of late-stage hydrothermal solutions near the margins of the ultrabasic bodies has modified the primary mineral composition of the rocks in the metamorphic aureole and created pseudo-metamorphic contacts along fault contacts. Prehnite, calcite, xonotlite and other secondary calcium-bearing minerals have formed in the metamorphic aureole. The recognition of pseudo-metamorphic contacts is of importance for their interpretation as true thermal metamorphic contacts may lead to erroneous conclusions regarding the age of intrusions. Colour changes in rocks adjacent to ultrabasic plutons and the presence of secondary low temperature silicate or sulphide minerals along contact zones have been interpreted by some investigators as due to contact metamorphism by the plutons. This interpretation may not be justified, and the changes observed may be due to hydrothermal action localized along the contact zone.

The development of a pseudo-metamorphic contact is best demonstrated at the fault contact between serpentinite and sedimentary rocks of the Humber Arm group in Winter House Brook (*see* Plate VIII A). The contact consists of a resistant, white-weathering layer 4 to 5 feet thick followed by green to black shale that grades through rust coloured shale into red shale. The colour changes suggest a 'bleaching effect' caused by the pluton. However, the contact differs from a normal igneous contact in the small size of the bleached zone and in the absence of amphibolite or other metamorphic rocks. A sandstone lens 15 feet from the contact shows no evidence of thermal metamorphism. The resistant contact layer is composed of brecciated fragments, wholly altered to secondary lime silicate minerals. The alteration is clearly due to hydrothermal solutions acting along the Gulch fault.

The presence of xonotlite ($5CaO.5SiO_2.H_2O$) among the secondary calcium-bearing minerals is of interest because of its unusual mode of occurrence and because this is apparently its first reported occurrence in Canada (Smith, 1954). Very few occurrences of xonotlite are known, probably because it is not easily identifiable in small amounts. Both massive, i.e. unoriented, xonotlite aggregates and fibrous forms occur in the Bay of Islands area. Fibrous xonotlite occurs in veins up to 3 inches wide and several feet long that cut altered sedimentary rocks near fault contacts at Winter House Brook, Shoal Brook, and near First Trout River Pond. Xonotlite also occurs with other lime-bearing minerals that fill joints in serpentinite near the basal contact of the North Arm Mountain pluton. Four separate bodies, composed of prehnite, pectolite, phlogophite, xonotlite, and other minerals, are exposed on the valley walls of a stream draining into North Arm. The largest body is 55 feet long and 10 feet wide. Xonotlite also occurs as a secondary mineral formed from the metamorphic minerals of the contact aureole. In the North Arm aureole, near the occurrences described above, a calcic hornfels is altered to a mass of prehnite, xonotlite, calcite, and clinozoisite, with only ragged relics of the original diopside remaining. About 30 miles farther south, in the Lewis Hills pluton, fragments of xonotlite were found in the alluvium, suggesting it also formed in other parts of the ultrabasic masses.

The xonotlite is a slightly pinkish, fibrous mineral that alters to chalky white on exposure. It is tough and hard ($H = 6 \cdot 5$), but easily scratched on chalky weathered surfaces. It has the following properties:

fibrous, positive elongation,
biaxial positive, very small 2V
specific gravity $2 \cdot 68 \pm 0 \cdot 04$
$N_x = 1 \cdot 592 \pm 0 \cdot 003$
$N_z = 1 \cdot 583 \pm 0 \cdot 003$
$N_z - N_x = 0 \cdot 009$

The specimens described above are deposited in the Brush Collection of Yale University (Specimens Nos. 6776 and 6777).

The xonotlite occurs as a vein-forming mineral, formed after solidification, serpentinization, and faulting of the ultrabasic plutons. It occurs in joints and faults in either the ultrabasic or metamorphic rocks. Xonotlite has been reported in limestone near igneous contacts at Tetala de Xonotla, Mexico (Winchell and Winchell, 1951) and at Goose Creek, Virginia (Shannon, 1925). No source of lime exists in the ultrabasic rocks now exposed at the contacts, and it is probable that the necessary lime was derived from the calcium-rich rocks of the contact metamorphic aureole by solutions circulating along fault planes. Whether or not these solutions were very late emanations from the ultrabasic rocks or hydrothermal solutions related to later igneous activity is not known.

CHAPTER III

STRUCTURAL GEOLOGY

General Statement

The Bay of Islands Igneous Complex forms a distinctive petrological and physiographic province along the west coast of Newfoundland, but its structural features conform to the general northeasterly structural trend of Newfoundland. The four major ultrabasic plutons outcrop along a north-northeasterly belt parallel to the Long Range on their east. The primary banding of the ultrabasic and gabbroic rocks and the strata of the Humber Arm group have been folded along northeasterly axes, now offset by later transverse faults. The primary bands of the major ultrabasic plutons dip generally to the northwest, although their attitude may vary locally from the regional attitude, particularly in the Lewis Hills pluton where eastward dips are recorded over wide areas.

The plutons have been offset by northwesterly directed stresses similar to those that have deformed much of the Appalachian belt. The southern half of the Complex has been displaced along the Serpentine Lake transverse fault, and the northern half along the Trout River transverse fault. Other transverse faults occur both in and along the margins of the plutons. Earlier discussions have centred around whether the plutons were emplaced as separate masses (Ingerson, 1937) or whether they are the faulted remnants of a single parent pluton (Cooper, 1936, Buddington and Hess, 1937). The bulk of the evidence, particularly from the northern half or the Complex, favours the latter view, and suggests that the plutons reached their present position by high-angle reverse faulting.

There is evidence, partly masked by post-intrusion deformation, that the plutons were emplaced in an active tectonic environment in which their original form was modified considerably. Thickening of the ultrabasic zones at depth, tectonic banding, intrusive contacts between gabbroic facies of the plutons and surrounding metavolcanic rocks, and other data suggest the forceful emplacement of a partially differentiated layered pluton. The various terms, laccolith, lopolith or sill do not adequately describe the structural features of the pluton.

Folds

Folding is of secondary importance to faulting in determining the structural grain of the area. The sedimentary and igneous rocks are repeated by a series of northeast-trending reverse faults, and folds are either confined to parts of separate fault blocks, or related to transverse faults. Igneous banding is as valuable as the layering of sedimentary rocks in solving minor structural details. In the three northernmost ultrabasic plutons, the attitude of the banded structure is uniform and has a general northeasterly strike and northwesterly dip. In the western parts of the area, near the Gulf of St. Lawrence, easterly dips are more common in both sedimentary and igneous rocks. The relation of this change in attitude to nearby faults is not clear. Fold structures are most common and very complex in the sedimentary rocks of the area. In the Trout River park area, small folds along north-northeast axes are related to the general shortening of the area. The sedimentary and volcanic rocks along the north shore of the Bay of Islands form a series of asymmetrical folds. An anticlinal structure near Liverpool Brook plunges to the north, whereas an adjacent syncline to the west plunges to the south. These structures are cut off to the north by a northwest-trending transverse fault.

Variations from the regional attitude are in many places related to the later transverse faults. Along the Trout River fault, southeast of the Trout River Ponds, the Humber Arm rocks have been flexed into a series of small folds striking northwest parallel to the trace of the fault plane. More pronounced, northwesterly striking trends occur in the northern part of the Lewis Hills pluton and along the Serpentine Lake–Serpentine River valley that separates it from the Blow Me Down pluton. The northern gabbro bands of the Lewis Hills pluton are folded into a structure that resembles a northwest-plunging syncline with a northwest-dipping axial plane. South of Serpentine Lake, the Humber Arm group and the underlying St. George group form a major northwesterly plunging anticline separating the two ultrabasic plutons. These divergent structures are related to one of the latest periods of deformation in the area. During this period, the plutons acted as buttresses against which the northwesterly directed forces pushed the sedimentary and volcanic rocks. The St. George and Humber Arm groups were forced between the two plutons and the banding of the northern part of the Lewis Hills intrusion was folded along northwesterly trending axes. Northwesterly trending structures also occur in the gabbro-amphibolite complex of the Big Level. The structural evolution of this part of the area is not known.

Faults

The northern half of the Complex can be divided into four main structural units that acted as discrete blocks during deformation (*see* Figure 8). These are the Coastal, Mount St. Gregory, North Arm Mountain, and Table Mountain masses. Each has been pushed to the northwest along

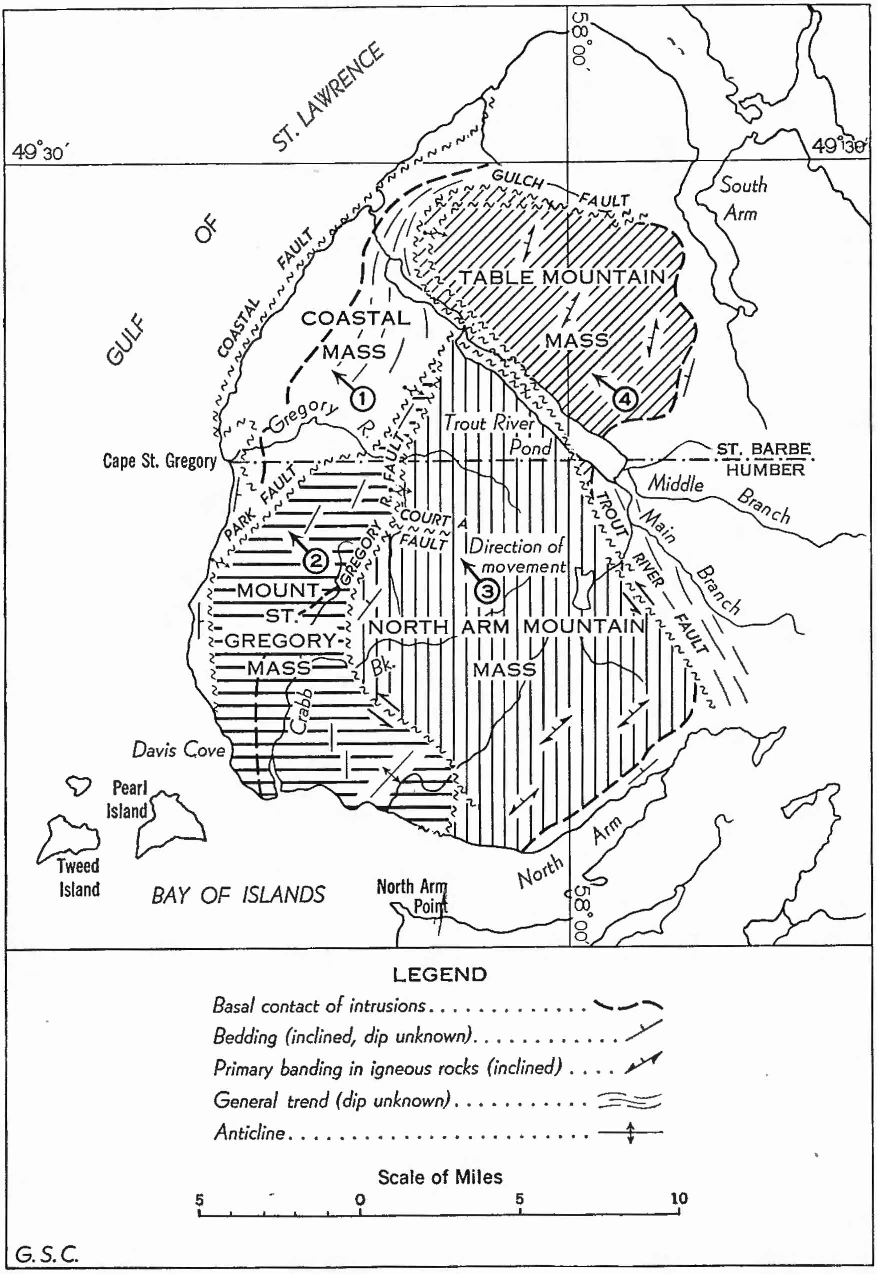


Figure 8. Structural units of northern half of the Bay of Islands Igneous Complex.

high-angle reverse faults that either pass into, or are truncated by, transverse faults along their strike. The sequence of fault movements is given below:

1. Coastal mass displaced to northwest along the Coastal fault.
2. Mount St. Gregory mass displaced toward the northwest along the Park fault, which appears to cut the Coastal fault at its southwest end.
3. North Arm Mountain mass displaced upward toward the northwest along the Gregory River fault.
4. Table Mountain mass displaced upward to the northwest along the Trout River and Gulch faults. The Trout River fault truncates the northeast end of the Park and Gregory River faults.

The younger faults brought successively deeper layers to the crustal level now exposed by erosion. Thus the Coastal and Mount St. Gregory masses represent the shallower or upper parts of the plutons and the Table Mountain mass the deepest part of the plutons exposed.

The structural pattern is somewhat similar in the southern half of the Complex (*see* Figure 11), although the mapped area is not large enough to furnish full details on the fault system. The Lewis Hills and Blow Me Down Mountain plutons have been displaced westward against the Coastal mass and acted as buttresses against and between which the weak Ordovician sedimentary rocks to the east have been thrust. The two plutons, and the related plutons to the north of the Bay of Islands, have been faulted into their present positions and, consequently, the original intrusive relationship has been considerably obscured.

Two principal types of faults are distinguished in the map-area. One comprises major northwesterly trending transverse faults whose net-slips have moderate horizontal components. These include the Gulch fault, the Trout River fault, the Serpentine Lake faults, and other similar tear faults that flank or cut through the ultrabasic masses. On most of these faults, the north side has moved west relative to the south side. Weitz (1953, p. 126) reports similar displacements along east-trending faults in the area east of the North Arm pluton, and this may be true of many of the transverse faults along the west coast of Newfoundland.

The second type of faulting comprises northeasterly trending longitudinal faults. These include both high-angle reverse faults and low-angle thrust faults. In places, they pass along strike into northwesterly striking transverse faults. The two types of faults are probably closely related in time and mode of origin.

FAULTS IN THE NORTH HALF OF THE COMPLEX

Gulch Fault

The Gulch fault is a transverse fault that forms the northern margin of Table Mountain. Its topographic expression is a steep-sided valley known locally as the Gulch. Exposures of the fault surface are rare because the lower parts of the valley are covered by sliderock, but the presence and attitude of the fault can be inferred from the following evidence:

Absence of a contact-metamorphic aureole. As noted in the section on metamorphic rocks, the contact aureole is a persistent feature along the base of each of the four ultrabasic plutons. At Shoal Brook its northerly trend swings abruptly westward along the line of the Gulch. It then thins rapidly and disappears within a mile, near Winter House Brook. This is the end of the contact aureole, and west of there the high resistant ridges that mark the metamorphic rocks give way to the lowland valley floor. Winter House Brook cuts across the projected strike of the amphibolite zone, but no amphibolite occurs in the continuous sequence of sedimentary rocks exposed in its valley. Near the western end of the Gulch is a small exposure of sheared sedimentary rocks that strikes north 35 degrees east and dips 40 degrees southeast. These shattered rocks are within 10 feet of sheared serpentinite at the foot of the scarp, but have not been thermally metamorphosed. The absence of a contact-metamorphic aureole is one of the chief criteria used to infer the presence of a fault in this area.

Shearing at the base of the Table Mountain scarp. The best exposures of ultrabasic rocks at the foot of the Table Mountain scarp occur where streams have cut through the sliderock mantle. There, zones of green and blue, highly deformed and polished serpentinite are exposed. Shear surfaces are irregular in the serpentinite but, in general, they dip approximately 60 degrees southward beneath Table Mountain. Although these shear surfaces are only found in the serpentinite, there being no sedimentary rocks exposed along most of the scarp, they probably reflect the attitude of the Gulch fault. The nature of the shearing is best seen at the western end of the valley, where shear zones over 300 feet wide at the foot of the scarp strike parallel to the valley and dip to the south. These zones consist of partly rounded greenish black serpentinite masses in a schistose serpentinite matrix. In places they have gouge partings.

Hydrothermal activity. Hydrothermal activity along the fault at Winter House Brook has already been described (page 49). Further evidence of hydrothermal activity in the fault zone occurs $\frac{1}{2}$ mile to the east, where the serpentinite has been brecciated and soaked by hematite-bearing solutions that also deposited quartz and carbonate minerals. Near the western end of the Gulch, 4 miles west of Winter House Brook, copper-nickel minerals are found in the strongly sheared serpentinite.

Truncation of structure. Primary banding in the ultrabasic rocks of Table Mountain strikes northeasterly at right angles to the strike of the fault and no change in trend is noticeable at the top of the Table Mountain scarp. Although exposures at the foot of the scarp are too sheared to show the attitude of the primary banding there, the regional relationships suggest that the banding is truncated by the fault.

Trout River Fault

The Trout River fault is a well exposed structure that can be traced by its topographic expression from the basal contact of the North Arm Mountain pluton northwesterly to Second Trout River Pond and thence along the lake, a distance of over 15 miles. It is one of the most prominent structural features of the northern half of the area. The nature of this fault has been the subject of controversy (Buddington and Hess, 1937; Ingerson, 1937), but the present field work indicates that it is a continuous transverse fault that has offset the Table Mountain pluton from its original position as a continuation of the North Arm Mountain pluton (*see* Figure 9). The fault surface exhibits the following features:

Topographic expression. Between North Arm and Second Trout River Pond, the fault trace is outlined by a fault-line scarp of ultrabasic and gabbroic rocks that in places is almost 1,000 feet higher than the lowland of sedimentary rocks on its northeast. Within 2 miles of the lake the height of the fault-line scarp decreases to less than 100 feet, due to the presence of resistant amphibolites on the northeast side of the fault. The valley of Second Trout River Pond is structurally controlled by this fault.

Truncation of structure and lithology. The Trout River fault separates rocks of different composition and attitudes. It truncates, almost at right angles, the primary banding of the ultrabasic rocks of the North Arm Mountain pluton, and separates these rocks from the sedimentary rocks of the Humber Arm group, which strike almost parallel to the fault trace. Similarly, it separates the gabbroic rocks of the Mount St. Gregory highlands on the southwest from the Humber Arm rocks on the northeast. At the southeast corner of Second Trout River Pond it separates, with angular discordance, banded gabbro from banded peridotite. Along Second Trout River Pond it truncates at right angles the primary banding of the ultrabasic rocks of the Table Mountain pluton, and separates them from the massive gabbro and metavolcanic rocks on the southern side of the lake.

Absence of contact-metamorphic aureole. The truncation of the contact aureole of the North Arm Mountain pluton and its absence from the northeast side of the pluton indicate the presence of a fault. Except for the dragged extensions of the Table Mountain contact zone, the fault trace is bounded on its northeast by a lowland of unmetamorphosed sedimentary rocks. Similarly there is no metamorphic aureole on the southwest side of Table Mountain, where Second Trout River Pond marks the line of the fault.

Drag features along the fault zone. Banded rocks have been dragged parallel to the trace of the fault in many places. Ultrabasic rocks and amphibolite at the basal contact of the North Arm Mountain pluton have been dragged through an angle of 60 degrees before being sheared off by the fault. On the opposite side of the fault, unmetamorphosed sedimentary rocks strike almost parallel to the fault trace. Along the fault scarp northeast of Sandy Pond is a small slice of deformed ultrabasic rocks with some amphibolite. This is part of the North Arm pluton contact aureole that has been dragged along the fault plane. The basal contact of the Table Mountain pluton also shows pronounced drag at the fault. This is shown by a triangular area of ultrabasic rocks at the southern end of Second Trout River Pond, where the primary igneous banding is, in places, almost parallel to the fault plane. Similar drag effects are found at the Narrows between First and Second Trout River Ponds, where the igneous banding is parallel to the fault. This drag shows that the Table Mountain block has moved to the northwest with respect to the North Arm Mountain block.

Exposures of the fault surface. The fault 'plane' is actually a zone of shearing at least several hundred feet wide. Near the head of Second Trout River Pond the shearing is highly irregular in attitude but commonly dips very steeply. At the Narrows the fault zone strikes north 30 degrees west and dips steeply to the southwest. The exposure consists of over 200 feet of sheared and polished serpentinite. This is only one of several branches of the transverse fault in this locality. Another branch to the southwest truncates the northeastern end of a banded gabbro-dunite contact, and yet another trends along the lake separating the banded gabbro and dunite on the northeast from massive gabbro on the southwest side of the lake.

Conclusions:

(1) There is a single continuous fault zone separating North Arm Mountain and Table Mountain.

(2) The fault zone has a nearly vertical dip.

(3) The nature of the drag at the fault zone indicates that the Table Mountain pluton has been displaced northwesterly relative to the North Arm Mountain pluton.

(4) The Table Mountain pluton is the faulted extension of the North Arm Mountain pluton.

Gregory River Fault

The Gregory River fault (*see* Figure 8) is irregular in trend due possibly to the combined effect of a series of transverse and longitudinal faults. It is older than the latest movements along the Trout River fault and these may have contributed to its irregular pattern. The fault extends northwestward from the shore of the Bay of Islands to the Gregory River valley and thence northerly along the valley to the Narrows. At its southern end, where it separates ultrabasic and gabbroic rocks on the northeast from sedimentary and volcanic rocks on the southwest, the fault is nearly vertical and is characterized by intense shearing and brecciation. Near the Stowbridge chromite deposit the gabbro-ultrabasic contact is deflected 60 degrees to the south, suggesting that the northeast side of the fault has moved northwest relative to the southwest side.

Along the Gregory River valley, the fault separates the unmetamorphosed volcanic rocks of the valley floor from the massive gabbros that underlie Mount St. Gregory although, west of the fault, amphibolite is found along the normal intrusive contact of the gabbroic rocks with the sedimentary and volcanic rocks. Along Gregory River, however, the gabbros are strongly sheared and gouge zones are present along the contact which apparently dips steeply towards the east. North of the stream draining Chimney Cove Pond, the fault is poorly exposed and its position and nature are assumed.

The Court A fault (*see* Figure 8) is the only subsidiary fault related to the Gregory River fault. It is a small tear fault of minor displacement that cuts the hanging-wall of the Gregory River fault. It is of interest as the locus of copper mineralization.

Exposures of the Gregory River fault surface suggest a single high-angle east-dipping reverse fault that has been subsequently folded, but it may be composed of two elements, a northeast-trending tear fault and a northerly trending high-angle reverse fault. The North Arm Mountain mass has been faulted over the Mount St. Gregory mass along this system. The Gregory River fault is older than the Trout River fault, but younger than the Park fault.

Park Fault

The Park fault (*see* Figure 8) forms the western margin of the Mount St. Gregory highlands, separating the igneous rocks of the highlands from the sedimentary rocks of Trout River Park. It is marked by a rugged fault-line scarp that extends from the shore of the Gulf of St. Lawrence to the Narrows. At its southern end the fault is almost vertical and marked by a wide schistose zone between volcanic rocks and deformed sedimentary rocks. East of North Head, large inclusions of red shale are found in the sheared gabbro. On the Feeder, a stream draining into Gregory River near

Chimney Cove, deformed serpentinite is found at the base of the gabbro scarp. Similar serpentinite masses occur at the north end of the fault, near the Narrows, and also along the Coastal fault east of Chimney Cove. These serpentinite bodies have probably been mechanically squeezed into the fault zone during deformation. The fault is not exposed on Gregory River, but unmetamorphosed sandstone outcrops in the stream floor below the gabbro hills. The sandstone there dips 60 degrees east, under the gabbro. Along the scarp overlooking Trout River Park, the gabbro is sheared and brecciated but the attitude of the fault is not certain.

The Park fault is cut by the Trout River fault at the Narrows and has been offset to the west so that it forms the western margin of the Table Mountain pluton. South of Chimney Cove it cuts across the projection of the Coastal fault. It is therefore younger than the Coastal fault and older than the Trout River fault. It is a high-angle, east-dipping reverse fault along which the Mount St. Gregory mass has been thrust westward over the sedimentary foreland.

Coast Fault

Extensive shearing along the sea-coast suggests the presence of a regional, partly submerged thrust fault. This fault separates the gabbroic rocks of the coastal belt from underlying volcanic rocks. Between Chimney Cove and Trout River, the gabbroic rocks reach the shore in most places, and the fault plane is not exposed. Its presence is, however, indicated by extensive schistosity developed in the gabbroic and ultrabasic rocks. Along the shore south of Trout River, small masses of volcanic rocks are faulted into the gabbroic rocks. North of Trout River these schistose volcanic and gabbroic rocks dip easterly at moderate angles. North of Chimney Cove, where the fault contact is exposed in the shore cliffs, it strikes north 25 degrees west and dips 70 degrees to the northeast. The fault is offset by a transverse fault at Chimney Cove, and then truncated by the Park fault.

The Coast fault was probably one of the oldest faults of the area, along which the underlying gabbroic rocks were thrust over the volcanic rocks of the eugeosyncline. Its attitude is not certain but some exposures of the fault suggest a moderate to steep easterly dip, i.e., it is more probably a high-angle reverse fault than a thrust fault.

STRUCTURAL INTERPRETATIONS OF THE NORTH HALF OF COMPLEX

Previous studies of the structure of the north half of the Complex have been chiefly concerned with the relationship of the two principal ultrabasic bodies to the Trout River fault. Ingerson (1935), shows two faults along the line of the Trout River fault, an older one that truncates the North Arm Mountain pluton and a younger one along the line of the Trout River Ponds. He concludes that the ultrabasic bodies are parts of two separate plutons because:

(1) The ultrabasic layer of the Table Mountain pluton is thicker than that of the North Arm Mountain pluton.

(2) The two ultrabasic bodies have their floors at different stratigraphic horizons in the Humber Arm group, although each pluton is, apparently, a concordant body.

Buddington and Hess (1937) interpret the ultrabasic zones as the faulted remnants of a single intrusive body, but they state that the displacement is due to a low-angle thrust fault that has moved the Table Mountain pluton more than 7 miles to the west along a flat floor. Their chief evidence is a small block of vertically banded gabbro that overlies a relatively flat floor of peridotite near the shore of First Trout River Pond. They state, "that the gabbro is an outlier of a low-angle thrust sheet can hardly be questioned, though the lower contact of the gabbro and the actual fault plane is (*sic*) hidden by talus slopes".

The writer favours the interpretation that the Table Mountain mass is derived from a deeper level in the parent pluton, and attained its present structural position by an upward movement of the north side of the Trout River fault. This suggests that the parent pluton had a wedge-shaped ultrabasic zone that thickened down dip rather than the parallel walls characteristics of a sill. This interpretation is shown in a schematic block diagram (*see* Figure 9).

The fact that one ultrabasic layer is thicker than the other does not eliminate the possibility that one pluton is the faulted extension of the other. Individual bands in the ultrabasic zone vary in thickness along their strike, and it is possible that the ultrabasic zone itself could vary in thickness either along strike or down dip. In view of the relation of the plutons to the Trout River fault it is logical to assume they represent different levels in a parent pluton, brought to their present position by differential movements along the Trout River fault.

The stratigraphy of the Humber Arm is not yet sufficiently known to permit the conclusion on which Ingerson's second argument is based. The sedimentary rocks at the base of each pluton are similar in lithology and, as the Humber Arm group is poorly dated by fossils, the stratigraphic position of the plutons cannot be determined. If Ingerson's assumption is correct and the two basal contacts are at different horizons within the same group, the interpretation favoured in this report is still valid for even a relatively concordant, intrusive body could be expected to cut across sedimentary structures to some degree along a strike length of over 16 miles.

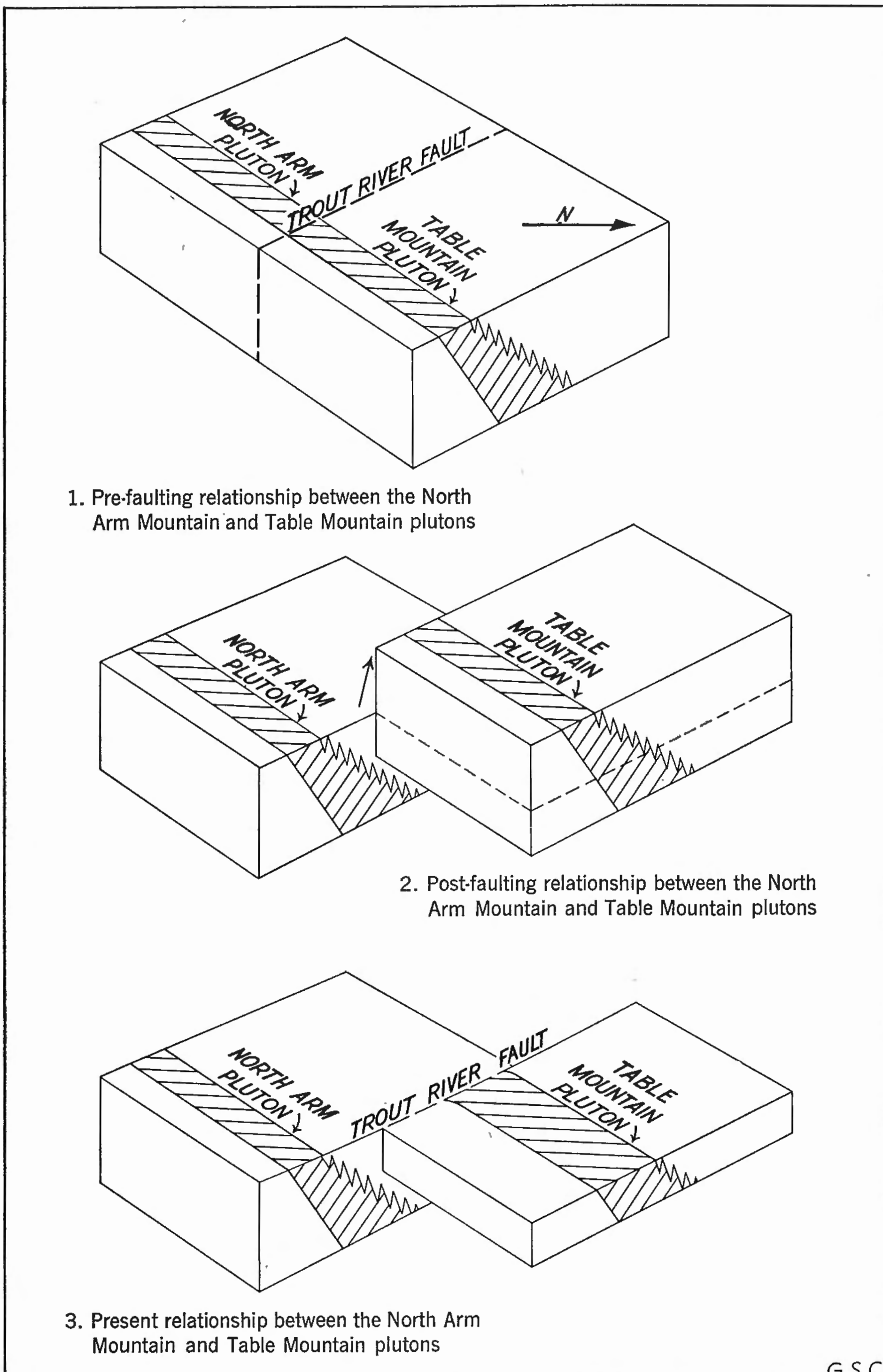


Figure 9. Block diagrams showing relationship between North Arm Mountain and Table Mountain plutons.

The block described by Buddington and Hess is approximately 600 feet by 1,300 feet in size. Its summit is 200 feet above the sliderock slopes of Table Mountain and 600 feet below the top of the gabbro cliffs of Table Mountain, which lie 1,000 feet to the northeast. The banding in the gabbro of the block has the same attitude as that of the Table Mountain gabbro. The block lies on a floor of serpentinite that dips 5 to 10 degrees toward Trout River Pond. There are two other possible interpretations for the block that merit consideration. The block may be a slice of gabbro in the Trout River fault zone because of its present position on the line of the fault. Other similar slices are found farther to the southeast along the Trout River fault and also along the south side of the Blow Me Down intrusion. This could also explain the outcrops of dunite and banded gabbro occurring along the lake shore at the Narrows. The banded gabbros at the Narrows, however, overlie dunite conformably, whereas, in the block under consideration, the vertical banding in the gabbro is truncated down dip by a flat floor of peridotite. The interpretation of the block as a fault slice does not satisfactorily explain the flat floor.

The presence of a flat floor beneath the block, and the fact that similar smaller gabbro blocks form knobs on the hill slope, suggest that the block is merely a very large piece of sliderock that has slumped from the side of Table Mountain. Plate VIII B shows large gabbro blocks in the process of spalling off the gabbro cliffs of Second Trout River Pond and sliding downhill to the lake. This illustrates the mechanism involved. Landslides with blocks of large size are common on the steep sides of the mountains. Plate IX shows such a slump, more than $\frac{1}{2}$ mile wide, composed of gabbro blocks. This slump is only $\frac{1}{2}$ mile from the block under discussion.

The interpretation of this block as a landslide feature removes the main evidence for low-angle thrust faulting in the area. As pointed out by Ingerson (1937), the great difference in thickness of the two plutons is not compatible with a low-angle thrust relationship. However, this variation in thickness can be explained by high-angle reverse faulting. The irregularity of fault traces in the area, suggestive of low-angle faulting, is apparently due to movements along combinations of transverse and longitudinal faults, and to the deformation of earlier-formed fault planes.

The absolute movement of the Trout River fault is not known. If the two ultrabasic bodies were once continuous, the only way of explaining their present structural relationship is for the Table Mountain side of the fault to have moved up relative to the North Arm Mountain side. Drag features at the fault surface indicate that the Table Mountain block did not move vertically upward, but that the direction of movement had a strike-slip component. Drag-folds and other field evidence, however, suggest that the net slip on the Trout River fault had a greater vertical than horizontal component.

The greater thickness of ultrabasic rocks in Table Mountain shows that the ultrabasic zone is wedge-shaped, and increases in thickness with depth. Most of the overlying gabbro cover has been eroded and faulted from the Table Mountain side of the fault, but much of it remains on the North Arm Mountain side of the fault.

The extension of the Trout River fault to the southeast is not known. To the northwest it ceases to be prominent, and is related to the northeasterly trending longitudinal faults as shown in Figure 10.

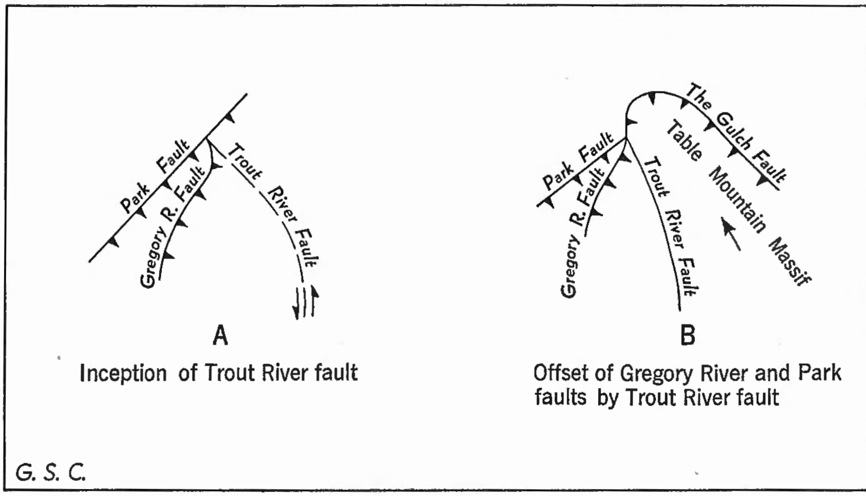


Figure 10. Evolution of Trout River fault.

The Trout River fault probably began as a tear fault related to the Park and Gregory River longitudinal faults. Increasing deformation caused the Trout River fault to assume a more active role, breach the Gregory River fault, and offset the northern part of the Gregory River fault to the west. Thus the Table Mountain block was pushed farther westward against the sedimentary foreland. The Trout River and Gulch faults do not extend past the offset part of the Park-Gregory River fault system and thus do not offset the granitic and gabbroic belts along the coast.

FAULTS IN THE SOUTH HALF OF THE COMPLEX

Blow Me Down Brook Fault

The northeast scarp of the Blow Me Down intrusion coincides with a major transverse fault that truncates the regional trend of the ultrabasic and gabbroic rocks (*see* Figure 11). The intensity of deformation is indicated by a $\frac{1}{2}$-mile-wide shear zone in the ultrabasic rocks along the base of the fault-line scarp. The ultrabasic rocks are completely serpentinized and there are asbestos stringers up to $\frac{1}{4}$ inch wide. In the tributary streams to Clarks Brook the limestone and shale of the Humber Arm group have been deflected from their normal trend so that they are parallel to the fault zone. They are relatively unmetamorphosed. Only on one ridge was an outcrop of albite-chlorite-titanite schist found. Its banding is transverse to that of the main fault zone, and it is apparently a slice in the fault zone dragged from the basal contact of the intrusion.

Near Blow Me Down Brook the fault trace is irregular. Some outcrops of bluish weathering serpentinite occur in the bush-covered hills bordering the intrusion. These may represent 'cold punches' of serpentinite into the sedimentary rocks during later deformation. Farther to the northwest the fault lies in the valley of Blow Me Down Brook, where its position is only approximately located.

The offset on this fault is similar to that on the transverse faults to the north, i.e., the north side of the fault has moved westerly relative to the south side.

Transverse Faults in the Blow Me Down Pluton

Faults within the ultrabasic masses are difficult to recognize because of the uniform lithology and the discontinuous nature of outcrops. Shearing in the ultrabasic rocks is common but, unless accompanied by changes in topography or pronounced alteration, only very detailed study will serve to outline the major shear zones. One of the best methods of recognizing major shear zones in the ultrabasic masses is to map deflections in the basal contacts of the plutons or in the gabbro-ultrabasic contacts. In the North Arm Mountain pluton, these contacts are remarkably straight for miles. In the Blow Me Down pluton, however, the basal contact is folded in two places, which may indicate the presence of two transverse faults. As no actual offset of the basal amphibolite zone was seen at either of these flexures, the displacement on these faults is probably small.

The northernmost flexure of the basal contact of the Blow Me Down pluton (*see* Figure 11) is connected by northwesterly trending shearing with a similar flexure of the gabbro-ultrabasic contact. This zone of shearing may represent a fault of minor displacement.

The southern flexure of the basal contact lies on a pronounced northwest-striking linear. Farther to the northwest, a branch of Simm's Brook has cut into the shear zone and there is well-developed, nearly vertical, sheeting on the valley walls. Still farther to the northwest, the gabbro-ultrabasic contact is offset and the gabbro and ultrabasic rocks are considerably deformed. Like the northern shear zone, this zone follows a northwesterly trend for over 5 miles. It is another fault zone with small displacement but great lateral continuity.

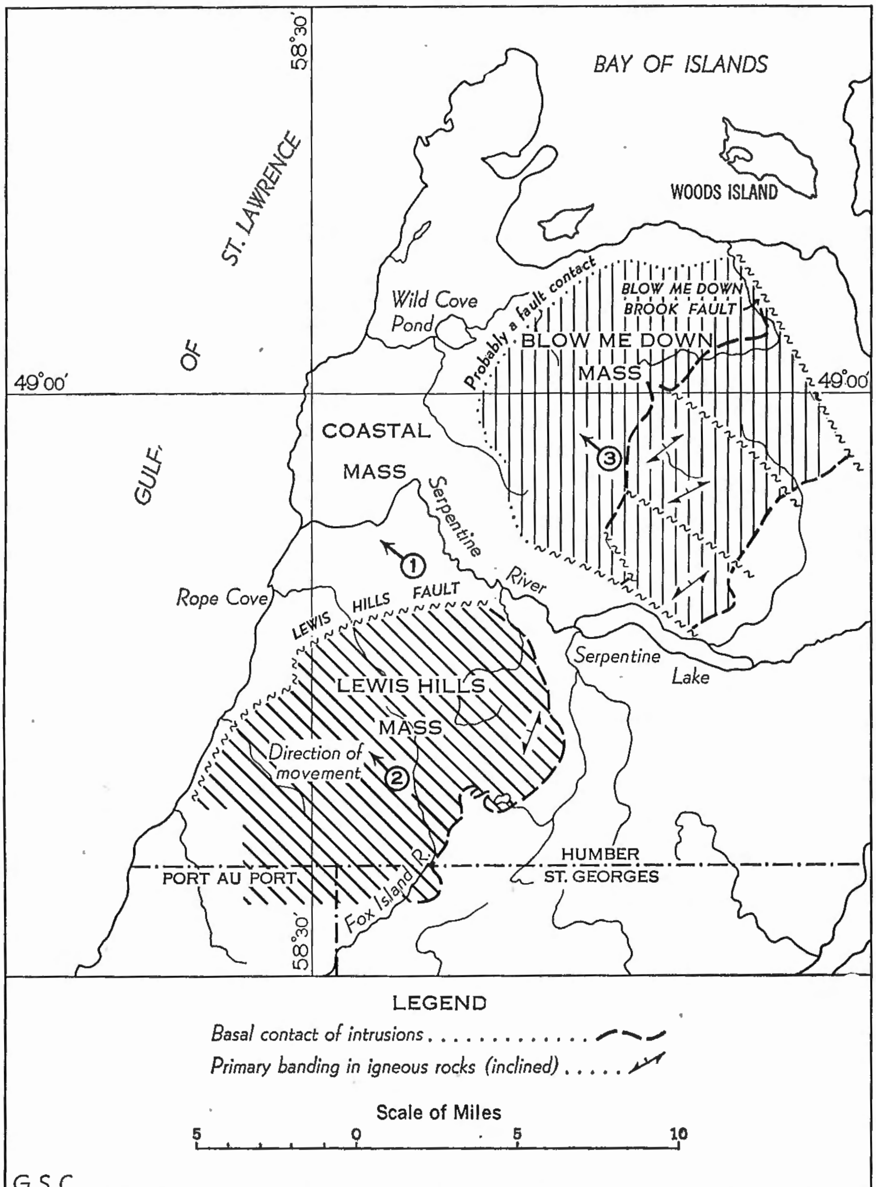


Figure 11. Structural units of the southern half of the Bay of Islands Igneous Complex.

Serpentine River Valley Faults

The Serpentine River valley probably occupies a wide zone of northwest-trending transverse faulting and folding between two igneous buttresses, the Blow Me Down and Lewis Hills plutons. The valley is largely drift covered, but the few outcrops found together with data from aeromagnetic maps indicate that it is underlain by shales and sandstone of the Humber Arm group. These are complexly folded, but commonly dip steeply and strike parallel to the trend of the valley. The underlying St. George group, which forms a steeply plunging anticline near the head of the valley, apparently has been folded into the gap between the two igneous masses. Metamorphic rocks, indicative of normal igneous contacts in other parts of the area, are not exposed along the valley and it must be concluded that the southern end of Blow Me Down Mountain and the northern end of the Lewis Hills are bordered by faults.

A well-defined branching fault forms the southern end of the Blow Me Down Mountain. A zone of shale and limestone up to half a mile wide separates it from Serpentine Lake. These sedimentary rocks are not metamorphosed. The shales have northwesterly strikes and moderate northeasterly dips. They strike at right angles to the primary igneous banding in the pluton. A small serpentinite body, apparently representing a 'cold intrusion' is found in the sedimentary rocks of the Red Gulch Brook. The northeast-trending amphibolites of the basal metamorphic aureole are also truncated by the shale and limestone near the north shore of Serpentine Lake. The ultrabasic rocks along the fault zone are deformed, and sheeting is locally well developed. The fault divides at Red Gulch Brook, and the branches extend both north and south of a block of poorly banded gabbro. The northern branch separates the gabbro from dunite. It is marked by a straight stream valley, on the south wall of which the dunite is very schistose. The schistosity strikes north 10 to 35 degrees west. Sliderock obscures much of the southeast end of the gabbro block, but troctolites that outcrop near the base of the block suggest that the main gabbro-ultrabasic contact is nearby. The offset of the gabbro-ultrabasic contact shows that the north side of this transverse fault has moved to the west relative to the south side. This relative movement is similar to that along the transverse faults to the north.

Faulting of the Lewis Hills Pluton

The Lewis Hills pluton is the most deformed member of the igneous complex. Its internal structure and marginal contacts are extremely irregular in comparison with those of the northern intrusions. Only the ultrabasic and their associated banded gabbroic rocks were mapped, and a more detailed study of the metagabbro and metavolcanic rocks to the west is required before the structure will be fully understood.

The north side of the Lewis Hills pluton is probably bounded by several faults. The contacts are sheared and the adjacent sedimentary rocks are relatively unmetamorphosed. On the east side of the Lewis Hills, the presence of contact metamorphic amphibolites suggest a normal intrusive contact. This contact is folded and offset along numerous small faults. Both the amphibolites and the adjacent ultrabasic rocks have a secondary foliation, and the serpentinites are locally so schistose that they resemble black shales. The deformation is most intense southeast of Fox Island River, where areas of unmetamorphosed Humber Arm rocks and amphibolites are incorporated into the ultrabasic masses and, conversely, ultrabasic 'tongues' intrude the amphibolites.

The Lewis Hills mass as a whole is a resistant core of igneous rocks that has been faulted to the northwest against the relatively incompetent sandstone and shale of the Humber Arm group. The latter underlie both the Serpentine River lowlands and a narrow fringe along the Gulf of St. Lawrence shore. Cooper (1936) states that the Lewis Hills have been thrust to the west along a nearly horizontal fault plane. The fault trace is characterized by a high scarp most of which is outside of the area mapped. A relatively flat fault contact for the ultrabasic-gabbro mass exposed in the Mine Cove-Lewis Brook area is suggested by exposures along the shore, by the re-entrant of Humber Arm rocks at Bluff Head Brook, and by the results of diamond drilling of the asbestos deposits. This body, however, appears to be a faulted extension of the main Lewis Hills ultrabasic mass, and has probably moved independently of the main massif. Thus its basal contact may not reflect the attitude of the entire Lewis Hills thrust.

STRUCTURAL INTERPRETATIONS OF THE SOUTH HALF OF THE COMPLEX

Successive stages of deformation have greatly confused the emplacement and early structural history of the southern half of the Complex in several ways. Unlike the simpler structural relationship between the North Arm and Table Mountain ultrabasic masses, where a single fault separates the two plutons, the Blow Me Down and Lewis Hills plutons are separated by a valley about 2 miles wide underlain by sandstone and shale of the Humber Arm group. An original structural continuity between the two bodies cannot be effectively proven, at least not by simple faulting. Furthermore, later deformation has superimposed new deflections on the basal metamorphic contacts of the plutons north and south of Serpentine Lake, so that the original drag features on the Serpentine Lake faults have been obscured. Thus the relative direction of movement between the two bodies cannot be satisfactorily resolved. Lastly, the internal structure of the Lewis Hills pluton, that is the spatial relationship between the gabbroic and ultrabasic rocks, is quite different from the structure of the Blow Me Down or other northern plutons.

A northwesterly striking fault zone borders the two plutons along the Serpentine River valley. Along the west side of the Lewis Hills, outside the area mapped for this report, Cooper (1936) shows fault contacts between igneous and metamorphic rocks on the east and unmetamorphosed Humber Arm rocks on the west. At Mine Cove and Bluff Head a relatively flat fault contact separates ultrabasic and gabbroic rocks from underlying Humber Arm sandstones.

Cooper suggested that the Lewis Hills mass was thrust 6 miles to the west along a flat thrust (the Lewis Hills overthrust) from its original position as the southern extension of the Blow Me Down mass. Walthier (1949) inferred a displacement of $4\frac{1}{2}$ to 5 miles for the southern part of the overthrust.

Cooper's assumption that the two plutons were part of a single intrusive body is probably valid, although not open to direct proof. Apparently deformation that post-dates the dislocation of the parent intrusion has forced the underlying Humber Arm and St. George sedimentary groups into the structural break between the two plutons. Considering the structural trends within the two plutons, the westward horizontal displacement of the Lewis Hills pluton relative to the Blow Me Down Mountain pluton is probably less than 1 mile and certainly not the 6 miles suggested by Cooper (1936). There are no drag features along the Serpentine River valley to support Cooper's assumed direction of movement and the offset gabbro block in the south side of Blow Me Down formatio suggests an opposite direction of movement. Probably the two plutons have been jostled many times along a series of transverse faults since they were originally faulted apart. The idea of a flat thrust is based on the fault at Bluff Head, but this underlies a separate fault block exposed south of Lewis Brook. North of there the fault is not exposed, but its presence is assumed on the basis of abrupt changes in lithology. Its attitude is not known. A simple flat thrust of small displacement does not explain the different internal structure of the two plutons. As discussed below, the distribution of gabbro, chromite, and dunite in the Lewis Hills pluton suggests a layered pluton similar to, but less inclined than, the Blow Me Down Mountain pluton. Tilting of the Blow Me Down pluton relative to the Lewis Hills pluton has been the major result of the faulting, and low-angle thrusting appears to be of lesser importance. It is not possible to determine the directions of absolute movements but, apparently, the Blow Me Down mass has been raised and rotated relative to the Lewis Hills mass, and the displaced gabbro block on the south side of the Blow Me Down mass was broken off during this tilting process. The Mine Cove igneous rocks are part of a separate fault block that has moved independently of the Lewis Hills mass. This fault block is separated from the Lewis Hills mass by a transverse fault along the Lewis Brook valley.

Diabase dykes that cut the flat fault contact of these igneous rocks in the Mine Cove area indicate post-faulting igneous activity, but it is not possible to date the faulting more closely.

Form of the Plutons

Ingerson (1935), who first studied the area in detail, suggested that the two northern plutons were the remnants of two tilted laccoliths. His interpretation was based upon the presence of a conformable 'floor' and on the supposed presence of the 'roof' of the Table Mountain pluton exposed in the Feeder, a stream draining Table Mountain. The internal structure of the plutons was interpreted as consisting of a thick ultrabasic zone at the base, overlain by a gabbroic 'critical zone', and capped by an 'upper border facies' of ultrabasic rocks identical in composition with the basal zone.

Buddington and Hess (1937) stated that the 'roof' contact described by Ingerson was actually a fault contact, and that the sedimentary rocks of the 'roof', instead of being overturned as the Ingerson hypothesis required, were right side up, as can be seen by graded bedding near the contact. In preference to Ingerson's hypothesis Buddington and Hess supported Cooper's hypothesis (1936), that the individual plutons of the Bay of Islands Complex were originally parts of a single lopolith.

Both the above hypotheses are based upon the presence of a conformable floor, but as no true roof rocks were identified, evidence was lacking by which the authors could distinguish a pluton with a domed roof (laccolith) from one with a sunken roof (lopolith). The mapping on which this report is based has outlined the roof of the North Arm Mountain pluton and shown it to consist of a flatly dipping contact between gabbro and metavolcanic rocks. It is possible that detailed mapping of the western parts of the Blow Me Down and Lewis Hills plutons will show the existence of similar features. Part of the roof of the Table Mountain pluton may be indicated by small areas of augite-andesine-hornblende granulite that occur in the central parts of the gabbro cap. The roof contact exposed in the Mount St. Gregory highlands is more or less horizontal, but steepens in dip to the west. Small apophyses of gabbro cut the overlying metavolcanic rocks.

The external form of the Bay of Islands plutons cannot be described by any rigorously defined term such as laccolith or lopolith. The North Arm pluton is wedge-shaped in cross-section, with a conformable floor dipping about 45 degrees to the northwest, and a relatively flat but irregular roof dipping slightly to the west. The gabbro area underlying Mount St. Gregory cannot be proven to represent the western side of a lopolith, as it occurs in a separate fault block and has a different lithologic sequence of rock types from the North Arm Mountain pluton. The bodies are wedge-shaped intrusions, best described by the non-committal term 'pluton'.

The 'upper border facies' of ultrabasic composition presumed by Ingerson does not occur in the North Arm Mountain pluton, nor is there evidence for it elsewhere in the complex. The small area of ultrabasic rocks on the west side of Table Mountain, which at first glance seems to be in this stratigraphic position, is really part of the basal ultrabasic zone raised to its present structural position of faulting.

The typical internal structure of the plutons is illustrated in Figure 2. The ultrabasic zone is not sill-like but wedge-shaped, and increases in thickness with depth (*see* Figure 9). The contact between gabbro and ultrabasic rocks is not sharp but, rather, an interbanded zone of variable width and there is no doubt that both rock types were emplaced and crystallized simultaneously. Although gradational by interbanding in detail, the contact is sharp in relation to the great thickness of the ultrabasic zone. The contact drawn on the geological map may thus be considered as a relatively abrupt chemical boundary that separates the non-aluminous, non-calcic ultrabasic rocks from lime and alumina-bearing gabbroic rocks.

The origin of the interbanded ultrabasic and gabbroic layers has been difficult to explain, especially by gravitative differentiation in situ, as numerous changes in magma composition are required to explain the sequence. Figure 7 illustrates relationships in the North Arm Mountain and Table Mountain plutons in which bands of gabbro pinch out along their strike, their place being taken by peridotite. Similar relationships occur in the Lewis Hills pluton, where interbanded gabbroic zones grade along strike through feldspathic dunite into dunite. The above features, as well as the wedge-shape of the main ultrabasic zones, suggest that the smaller ultrabasic layers in the banded gabbro areas near the main ultrabasic contact may be tongues from the ultrabasic zone rather than discrete layers parallel to it. In this case they need not be explained by reversals in the composition of the crystallizing magma, an unlikely event in these circumstances, but rather by the flowage of a partly crystalline magma as described in Chapter IV.

Figure 12 is a hypothetical reconstruction of the form of a parent pluton based on evidence from cross-sections of the various members of the Complex. All the members of the Complex have conformable floors, and the wedge-shaped form is suggested by the greater width at Table Mountain than North Arm Mountain. In the sketch, the intertonguing at the gabbro-ultrabasic contacts is exaggerated. The nature of the roof is inferred from the features of the North Arm Mountain pluton. The members of the Bay of Islands Complex may have been derived by faulting from separate wedge-shaped intrusions, each of which had an overall primary form as shown in this sketch, or they may be the faulted remnants of one parent pluton of this general form.

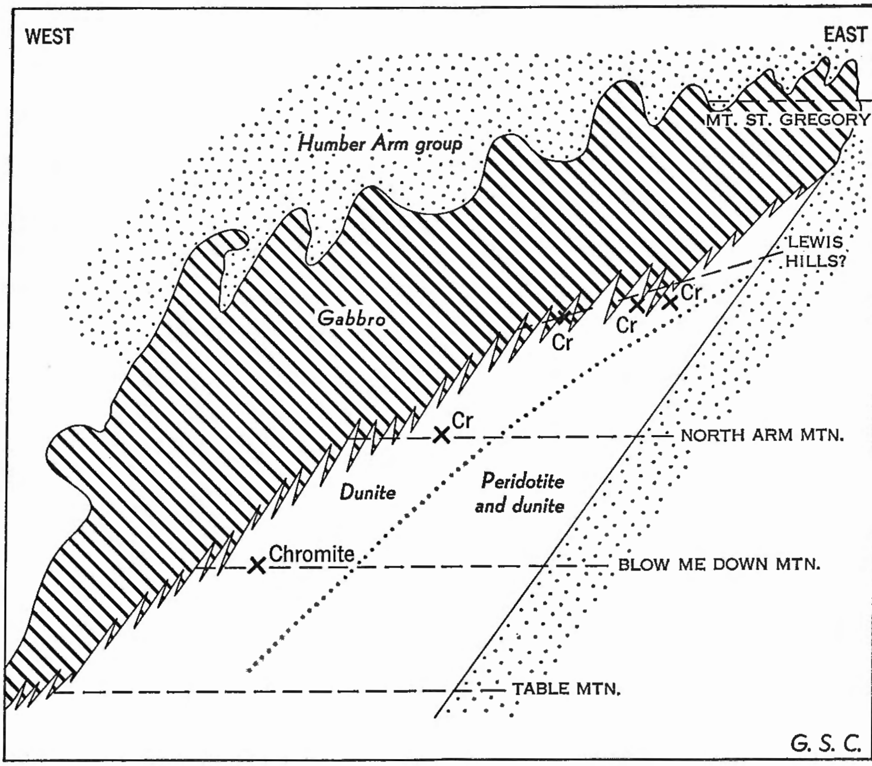


Figure 12. Hypothetical section of parent pluton.

Extensions of the Bay of Islands Complex

The Bay of Islands Complex is one of several northeasterly trending belts of ultrabasic and gabbroic rocks in Newfoundland. It is restricted to a belt of eugeosynclinal-type rocks that outcrop between Port au Port Peninsula and Bonne Bay (*see* Map 1). The south end of this belt is deflected westerly at Port au Port peninsula. Its continuation in the Appalachian belt is probably represented by Mount Albert and other ultrabasic plutons in northern Gaspé. The eastern ultrabasic belts of Newfoundland may belong to the same belt as the small and isolated ultrabasic occurrences reported from southern Gaspé and northern and southern New Brunswick.

At the northern end of the Bay of Islands Complex the termination of the eugeosynclinal belt coincides with a pronounced offset in the Long Range crystalline rocks. South of Bonne Bay they and Cambrian and Ordovician calcareous rocks outcrop well to the east of the principal ultrabasic plutons. North of Bonne Bay these rocks are offset to the northwest in line with the extension of the Complex. Mapping in the Bonne Bay area has not shown whether this offset is a flexure or a transverse fault resulting from the thrusting of the Long Range rocks to the northwest. The trend of the offset is however parallel to that of the major transverse faults within the Complex. Along the remainder of the west coast of Newfoundland Lower Ordovician rocks are most common and rocks of the Humber Arm group occur only as slices in thrust faults. Eugeosynclinal-type rocks are generally absent, and the nearest exposures of them are in the Pistolet Bay-Hare Bay area at the north tip of the island.

The relationship of the Bay of Islands Complex to the ultrabasic rocks of Hare Bay is open to speculation, as much of northern Newfoundland has not been geologically mapped, but the most logical interpretation is that the Hare Bay plutons are an extension of the Bay of Islands eugeosynclinal belt. It seems probable that the rocks of this belt, after being offset by the Bonne Bay flexure, are concealed beneath the waters of the Gulf of St. Lawrence to reappear in northern Newfoundland as shown on Map 1. The interpretation that the eugeosynclinal belt is not offset but extends beneath the Long Range thrust seems unlikely in view of the manner in which the Cambrian and Lower Ordovician rocks are warped around the Long Range north of Bonne Bay.

Age of the Orogeny

Basic and ultrabasic igneous rocks are the youngest rocks exposed in the Bay of Islands area. The only direct dating that can be obtained from this part of the Appalachian ultrabasic belt is that the plutons are post-Humber Arm group in age. The age of this group is not well established, but it is, in part, Middle Ordovician.

From a petrological viewpoint the abundance of basic volcanic rocks may indicate a time when basic and ultrabasic magma started to be generated in the area. The fact that the ultrabasic and gabbroic plutonic bodies are related in space with the basic volcanic rocks, suggests a common source for both and emplacement along related fractures in the earth's crust. Although it is not possible to state how long such a zone of weakness would be maintained, it is probable that the plutonic rocks were emplaced during the same period (Ordovician) as the volcanic rocks, rather than during a later orogenic period (Devonian).

Later ages have been assigned to ultrabasic bodies elsewhere in the Appalachian belt. In general, the younger intrusions are small serpentinite bodies, and the possibility has not been considered that they are derived from earlier ultrabasic plutons and emplaced during a later deformation by flowage and cold reintrusion. The effects of secondary reintrusion of serpentinite bodies are described in Chapter IV. They may considerably obscure the primary features and true age of ultrabasic plutons.

CHAPTER IV

PETROLOGICAL INTERPRETATION

General Statement

This chapter is devoted largely to speculation resulting from the study of the Bay of Islands Complex. It is hoped that this speculation will stimulate or irritate workers engaged in further detailed studies on the mode of occurrence and conditions of formation of ultrabasic rocks. First, a summary of the general types of plutons containing ultrabasic rocks is given, and the Bay of Islands Complex is discussed in relation to them. Next, the concept is developed of a continuous series of layered pluton-types, ranging in composition from gabbroic layered bodies with only minor ultrabasic layers (Bushveld Complex) to ultrabasic bodies with only minor gabbroic or feldspathic layers. Then the mode of origin of these various types of plutons is discussed with particular emphasis on possible modifications during contemporaneous and post-intrusion tectonism. Finally an attempt is made to point out the effect of tectonism in obscuring the nature of primary intrusions and erroneous interpretations that may result from studies of isolated, deformed ultrabasic plutons in orogenic belts.

Classification of Plutons Containing Ultrabasic Rocks

The contrasting features of ultrabasic bodies have been emphasized by the numerous classifications proposed by petrologists. Classifications have been based on:

(1) The relative amounts of ultrabasic and gabbroic rocks in the pluton.

- (a) *Gabbroic layered plutons with minor ultrabasic layers.* These are the typical gravity-stratified sheets (Buddington, 1936). Examples include the Bushveld and Stillwater Complexes and, in a sense, the Palisade sill with its olivine-enriched zone. Numerous examples occur in the Canadian Shield.
- (b) *Ultrabasic plutons with minor or no gabbroic layers.* Examples include the numerous small serpentinite bodies found along orogenic belts.

(2) The tectonic environment of the pluton.

- (a) *In non-orogenic (plateau) areas.* Bushveld and Stillwater Complexes.
- (b) *In orogenic belts.* These have been referred to as 'alpine-type' intrusions (Benson, 1926), or as 'injected peridotites' (Guild, 1947).

(3) The nature of the parent magma.

(a) *Derived from a gabbroic magma.* This type is referred to by Hess (1938) as a member of the basaltic magma series. Crystal settling in situ, possibly with convection or other modifications, is the dominant process.

(b) *Derived from a primary peridotite magma.* Referred to by Hess (1938) as a member of the ultramafic magma series. These supposedly formed by the intrusion of either a liquid, a crystal mush, a serpentine magma, or by solid flow.

Apart from these relatively simple and more common types of plutons, there are apparently more complex types, such as those composed of concentric rings of dunite surrounded by pyroxenite and gabbro. Their mode of origin is complex and is not satisfactorily explained as the result of gravitative differentiation in situ, as are the gravity-stratified sheets. Examples are the Blashke Island Complex (Walton, 1951) and the ring-like bodies of the Urals (Duparc and Grosset, 1916). The kimberlite and mica peridotite hypabyssal intrusions in non-orogenic zones are another group that is probably genetically distinct from the major groups listed above.

There have been attempts to develop two broad generalizations about the major types of plutons that contain ultrabasic rocks. Thus, Hess (1938) has stressed the idea that ultrabasic plutons containing little or no gabbroic material are typical of orogenic belts and have been derived from a primary peridotite magma. On the other hand, gabbroic layered plutons containing minor ultrabasic layers are considered to be typical of non-orogenic belts and formed by gravitative differentiation of a basaltic magma. The terms *gabbroic layered pluton* and *ultrabasic pluton* are used in this report to refer to the two types around which these two contrasting concepts have been built. The terms are used only to emphasize the essential difference between the two types of plutons, namely that one is composed dominantly of layered gabbroic rocks and the other is composed mainly of ultrabasic rocks. They are descriptive terms without genetic implications. Similar subdivisions have been used in recent geochemical studies of ultrabasic rocks. Davis and Hess (1949) have attempted to relate the abundance of radium in ultrabasic rocks to their assumed mode of origin. Holyk and Ahrens (1953) have made a similar study of potassium in ultrabasic rocks. The evidence from the Bay of Islands Complex suggests that the different modes of occurrence need not reflect different primary magmas. Instead, magmas that, under quiescent (non-orogenic) conditions, differentiate to form gabbroic layered plutons with thin ultrabasic layers, may give birth to typical ultrabasic plutons when emplaced in an orogenic belt. This is a secondary change accomplished by tectonism during and after the time of emplacement and need not indicate two distinctly different types of primary magma.

Features of the *gabbroic layered plutons* have been summarized by Buddington (1936, p. 348) as quoted below. They have

> "a relatively thin basal zone of a composition intermediate between the extreme variations occurring above. The material just above the basal zone is generally more mafic than that of the basal zone and passes upward into more feldspathic or felsic material. Accordingly, in the sheet as a whole, the specific gravity is intermediate at the base to a minimum in the upper part. This gradation may, however, be very irregular. In the lower portion of the thick sheets there may be ultrabasic segregations and small scale alternation of more mafic and more felsic material or bands of different mineral composition, and the uppermost part may be granite."

Where the mineral components show a compositional variation related to their height in the pluton, Wager and Deer (1953, p. 335) refer to the pluton as a 'Skaergaard-type' layered intrusion. These terms are applied mainly to gabbroic plutons that contain little or no ultrabasic rock.

Ultrabasic plutons form the bulk of the world's ultrabasic rocks and are used by Hess (1937, 1938) as evidence for a primary peridotite magma. Hess has called attention to the following features of such plutons:

(1) They occur in strongly deformed orogenic belts.

(2) They lack associated gabbroic rocks.

(3) They occur as concordant lenses, sheets, sills, and stock-like bodies. Larger bodies of batholithic proportions may occur at the crest of anticlinal structures, suggesting phacolithic forms.

(4) They are intruded during the first great deformation of the belt.

(5) Their border facies are of the same composition as the main rock mass.

(6) Effusive facies of the magma are almost entirely lacking.

(7) They produce remarkably little contact metamorphism.

(8) They show little variation that might be ascribed to magmatic differentiation. Feldspathic peridotites are only rarely associated with this type of pluton.

(9) They have few accessory minerals except chromite, picotite and magnetite.

(10) They contain less than 10 per cent total iron and commonly between 6 and 8 per cent. In contrast, those formed by the crystallization differentiation of basaltic magmas have an iron content slightly greater than 11 per cent. Chemical analyses indicate that the magnesia/iron ratio of ultrabasic plutons is greater than 7·5, whereas the ratio in basaltic differentiates is less than 7·5.

The essential differences between the two main types of plutons in which ultrabasic rocks occur, as suggested in current geological literature, are summarized in the following table:

Feature	Ultrabasic plutons	Gabbroic layered plutons
1. Relation to orogenic belts	Intruded into rocks of eugeosynclinal facies during the early stages of mountain building	Intruded into plateau areas
2. Associated igneous rocks	Gabbroic facies minor	Extensive sills, dykes, and related rocks of gabbroic composition
3. Layering	Rare, except for banding due to various proportions of olivine and pyroxene, without cryptic layering	Layered according to specific gravity and mineral composition (cryptic layering)
4. Nature of border zone	Has the same composition as main mass of pluton	Has chill zone of intermediate composition representing composition of primary magma
5. Extent of contact metamorphism	Minor effect on adjacent country rock	Greater metamorphic effects on adjacent country rock
6. Chemical composition	Ultrabasic Mg/Fe ratio greater than 7·5 Lime and alumina content low	Basaltic Mg/Fe ratio of ultrabasic layers less than 7·5 Lime and alumina content high

Relationship of Bay of Islands Complex to Gabbroic Layered Plutons and Ultrabasic Plutons

J. R. Cooper (1936, p. 40) first pointed out that "The Bay of Islands Complex is a connecting link between the lopoliths, and the ultrabasic rocks long recognized as characteristic of geosynclines". Thus, the Bay of Islands plutons combine features previously considered typical of gabbroic layered plutons with features typical of ultrabasic plutons.

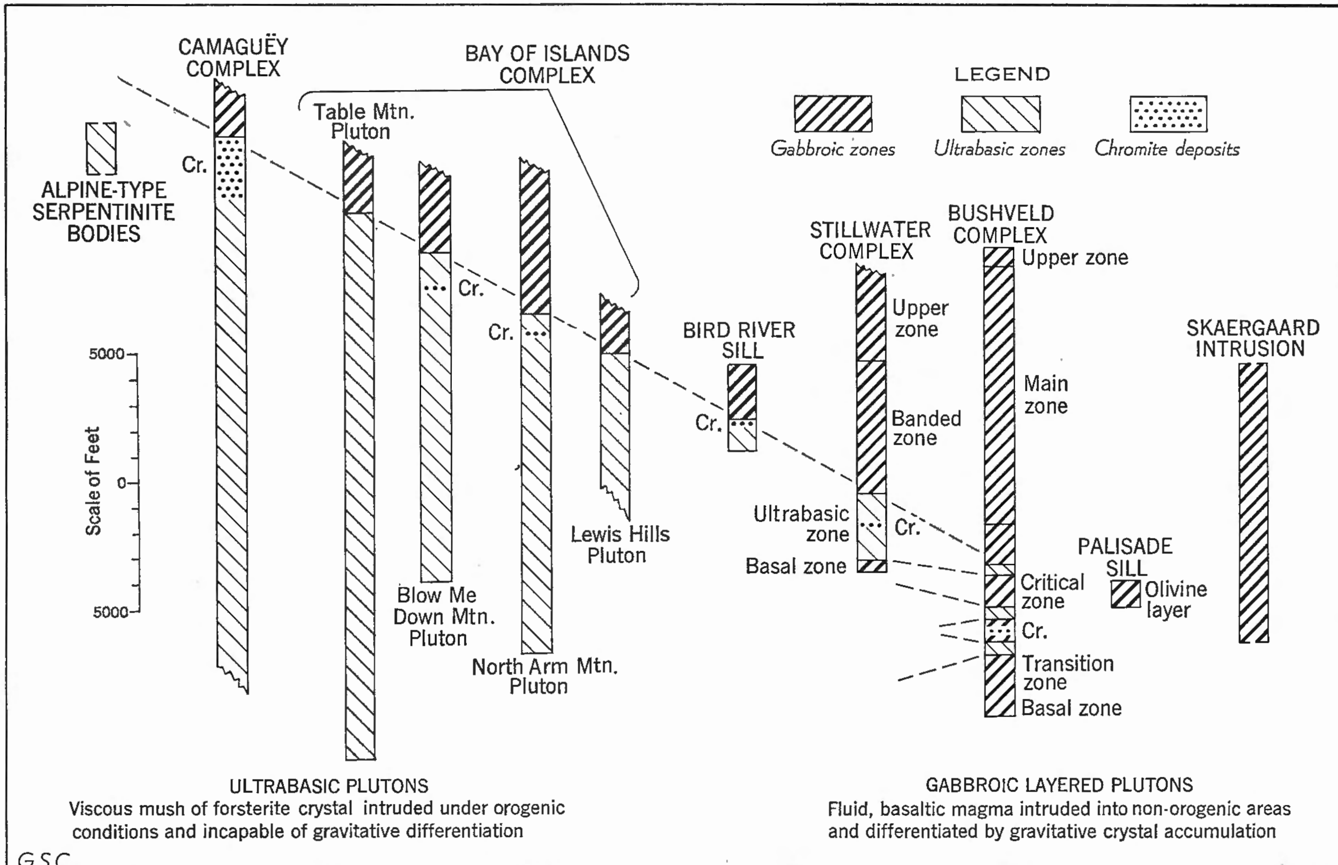


Figure 13. Ultrabasic-gabbroic layered pluton series.

It is possible to arrange many of the plutons containing ultrabasic rocks described in the literature into a series in which the abundance of ultrabasic rocks varies from 0 to 100 per cent (*see* Figure 13). The ultrabasic and gabbroic layered plutons are end-members of this series and the Bay of Islands Complex is an intermediate member. Certain of the contrasting hypotheses concerning the genesis of the end-members do not appear justified when considered in the light of this series.

The Bay of Islands Complex occurs in the Appalachian orogenic belt and is intrusive into a typical eugosynclinal suite of rocks. It has been deformed by later tectonism which has partly, but not completely, obscured the primary features of the plutons. Thus, whereas their location in an orogenic belt may suggest that the plutons are of the 'alpine-type', the large areas underlain by associated gabbroic plutonic rocks show that these plutons are also closely related to plutons of the gabbroic-layered type commonly considered typical of non-orogenic areas. The gabbroic rocks are genetically related to the ultrabasic rocks as shown by the interbanded nature of their mutual contacts and by the presence of rock types of intermediate composition (feldspathic dunite) near the principal contacts. As described in Chapter III, the Table Mountain and North Arm Mountain plutons are parts of the same parent body that has been dismembered by the Trout River fault. Although they were derived from the same magma, the Table Mountain pluton now consists dominantly of ultrabasic rocks with only a small roof zone of gabbroic rocks whereas the North Arm Mountain pluton contains a large thickness of gabbroic rocks that overlie the ultrabasic rocks. Thus, faulting alone has apparently developed an ultrabasic pluton from a gabbroic layered pluton!

The plutons of the Bay of Islands Complex show an overall gravity stratification, with gabbroic rocks overlying ultrabasic rocks. Yet, unlike gabbroic layered plutons of the Skaergaard-type, the rocks of the basal, ultrabasic zones show no significant variation in the composition of the essential minerals, forsterite and enstatite. Neither can small variations in plagioclase composition be related to stratigraphic position in the overlying gabbroic rocks. Because of the overall gravity-stratification of the bodies, this homogeneity of the ultrabasic zone and the random variation in the gabbroic rocks were quite unexpected. The ultrabasic zones resemble the typical ultrabasic plutons, which commonly have olivine and orthopyroxene of a similar uniform composition.

The nature of the primary border zone of the ultrabasic areas is not discernible in many exposures, as deformation localized along the contacts has caused flowage in the ultrabasic rocks and developed a secondary gneissic structure. Thus it is not possible to determine the nature of the original border zone directly. However, no gabbroic chill zone is found, and the composition of the present border zone is similar to that of the main ultrabasic zone. In this respect the border zone resembles that indicated by Hess to be characteristic of ultrabasic plutons.

The contact metamorphic aureole of the ultrabasic rocks is suggestive of medium- to high-grade thermal metamorphism. Although this aureole is not very thick, it is probable that post-emplacement deformation that caused flowage in the ultrabasic rocks near the contacts also modified the metamorphic aureole. Despite this later deformation the metamorphic aureole is much larger than that generally associated with ultrabasic plutons.

In the present study no new chemical analyses were made of the ultrabasic rocks. Hess (1938), however, gives an analysis of an ultrabasic rock from the North Arm Mountain pluton which has an Mg/Fe ratio of 9·5. This is considered by him as indicative of the ultrabasic magma series.

The Bay of Islands Complex, therefore, has features commonly considered to be characteristic of the two extreme types of ultrabasic-bearing plutons. Figure 13 is a graphic representation of these and other basic layered intrusions that seems to fit into an ultrabasic layered pluton series. The Bushveld Complex is at the gabbroic end of the series. It is a typical gabbroic layered pluton in which minor amounts of ultrabasic rocks occur. Of the Bushveld, Hall (1932, p. 320) states "Rocks characterized by the predominance of some member of the olivine group, whether fresh or altered, are restricted to the Critical Zone and are quantitatively subordinate within it".

The Palisade sill, with an olivine layer about 50 feet above its basal chill zone may also be considered an end-member of the series. It is particularly instructive as to differentiation process, and its origin by crystal settling has been clearly demonstrated (Walker, 1940).

Ultrabasic rocks are more abundant in the Stillwater Complex. They form a 2,500-foot layer immediately above the basal chilled zone. In the more gabbroic plutons, including the Stillwater Complex, basaltic chilled zones are used to infer the nature of the parent magma. Plutons in which ultrabasic rocks are abundant generally lack such chilled zones.

Numerous plutons of the Canadian Shield are characterized by layered structures and consist chiefly of either ultrabasic or gabbroic rocks (MacLaren, A. S., 1953). Where tops of flows are determined in the adjacent pillowed volcanic rocks, the gabbro zone is found to overlie the ultrabasic zone and the bodies resemble typical gravity-stratified sheets. Cryptic layering is rarely observed because of the extensive secondary alteration of these plutons. The ultrabasic belts of northwestern Quebec and Ontario include intrusive bodies characterized by layered structures together with other deformed bodies composed entirely of serpentinite. These Precambrian plutons cannot be placed in any one position in the sequence of Figure 13 because of their variability in composition. However, the Bird River sill of Manitoba is shown because of its important chromite deposits.

The plutons of the Bay of Islands Complex occupy a median position in the sequence for reasons discussed above. The different aspect of individual plutons in the Complex, due to the influence of later deformation in modifying their original form, is at once apparent. The Camagüey Complex of Cuba is a stratiform body similar to the Bay of Islands Complex (Flint, de Albear and Guild, 1948, p. 59). The thickness of the ultrabasic and gabbroic layers are unknown, but the ultrabasic zone is apparently very thick, whereas the gabbroic rocks form only 10 to 12 per cent of the exposures. It is, therefore, not plotted to scale, and its position in the sequence is determined only by the apparent relative abundance of gabbroic and ultrabasic rocks. The position of its important chromite deposits, stratigraphically below the gabbro zone, is also shown.

At the ultrabasic end of the series are the 'alpine type' of serpentinite bodies, devoid of gabbroic rocks. They are considered typical of orogenic belts and are interpreted by Hess (1938) to be evidence for a primary peridotite magma. However, the existence of a pluton series, such as that shown in Figure 13, indicates a closer genetic relationship between ultrabasic plutons and gabbroic layered plutons. It suggests that both types of plutons were derived from a similar parent magma, and that differences in cooling, emplacement history and post-emplacement deformation have caused the variations in the relative abundance of gabbroic and ultrabasic rocks in the sections of the plutons now exposed to view.

Relationship of Ultrabasic-Gabbroic Layered Pluton Series to Hypotheses of Origin

The most general and commonly accepted mode of differentiation of the gabbroic layered pluton is by the gravitative crystallization differentiation of a gabbroic magma in situ, with or without convection. Different hypotheses have been proposed to explain the formation of ultrabasic plutons. Hess (1938) has advocated a primary peridotite magma distinct from a gabbroic magma. Bowen (1915) has raised cogent objections to hypotheses based on the intrusion of liquid peridotite magmas. Hess attempted to explain ultrabasic plutons as due to the intrusion of magma of serpentine composition, but Bowen and Tuttle (1949) have shown that no liquid phase exists in systems of the composition of serpentine at geologically reasonable temperatures and pressures. Bowen, in numerous articles since 1917, has advocated the idea that dunites have formed by the intrusion of masses of olivine crystals and therefore that ultrabasic plutons were never intruded in a liquid form. He suggests (1949, p. 455): "It may be that an olivine aggregate is more capable of flow in the crystalline state than other common anhydrous rock-forming minerals, because its crystals are built up of SiO_4 groups without any chain, sheet, or space linkages—i.e., sharing of O atoms by Si atoms."

There are thus two distinctly different hypotheses for the formation of ultrabasic rocks. One is based on crystal settling from a basaltic magma and is applied to the ultrabasic rocks of the gabbroic-layered plutons; the other is based on intrusion of a primary ultrabasic magma and applied to the ultrabasic plutons. However, as mentioned above, the two end-members of the pluton series need not have distinctly different modes of origin, but rather share certain features of genesis. The Bay of Islands Complex, as an intermediate member of this series, may offer a solution to this problem. Study of this complex has shown that deformation, flowage, and faulting have developed ultrabasic plutons of batholithic dimensions from plutons that originally had a large component of gabbro. Although similar to ultrabasic plutons, these plutons have not formed from an ultrabasic magma, but by the mechanical dislocation of primary plutons composed partly of gabbro.

It is suggested that other ultrabasic and intermediate members of the pluton series that occur under similar orogenic conditions may have originated in the same way. It seems to the writer that the final elucidation of the problem will result, not from the examination of isolated and highly deformed 'alpine-type' plutons, but from a general consideration of the interrelationship of gabbroic, composite and ultrabasic bodies in time and space in an orogenic cycle. Furthermore, failure to consider the role of later deformation in blurring the primary features of intrusions has in the past led to misconceptions, and will continue to do so.

Mode of Emplacement of the Bay of Islands Complex

TYPES OF PLUTONS IN COMPLEX

The different members of the Bay of Islands Complex vary greatly in their relative content of gabbroic and ultrabasic rocks. The gabbroic belts along the Gulf of St. Lawrence contain minor, ultrabasic layers; the North Arm Mountain pluton contains a large mass of ultrabasic rocks and an extensive gabbro cap; the Table Mountain pluton is composed dominantly of ultrabasic rocks and only a small part of the gabbro cap remains; smaller, isolated bodies composed entirely of serpentinite occur along fault planes. It is obvious that some of these bodies cannot have been directly derived from primary peridotite magma, and the others from a primary gabbroic magma. The field relationships show that all, regardless of the amount of ultrabasic rock they contain, were derived from the same primary magma. Furthermore, it is apparent that those with the most ultrabasic material are the most highly deformed.

NATURE OF THE PRIMARY PLUTONS

The small serpentinite bodies localized along fault planes cannot be considered as primary plutons. Their sheared and altered aspect, smaller size, and their occurrence as lenses along faults younger in age than the principal igneous intrusions, show that they have been greatly deformed and emplaced as solid masses ('cold punches') during later tectonism. They have many of the features of typical ultrabasic plutons (ultrabasic composition, no chill zone, lack of contact metamorphic effects, etc.) but they are derivatives of primary plutons.

The Table Mountain pluton is not a typical primary pluton, for it has been modified considerably by faulting and is composed chiefly of ultrabasic rocks. It has been shown earlier to be the faulted extension of the North Arm Mountain pluton. The relative abundance of gabbro and ultrabasic rocks in the Lewis Hills pluton is imperfectly known, partly because its western gabbroic margins are still unmapped, partly because some of the gabbroic rocks are altered and their origin is uncertain, and partly because the banded gabbro bodies form irregular zones and cap-like bodies over much of the dunite zone of the pluton.

The North Arm Mountain and Blow Me Down plutons are 'type' examples of primary intrusions. They have, however, been modified to some extent by later deformation but less so than any of the other plutons. Nevertheless they approach most closely the typical primary pluton from which the nature of the parent magma can be judged.

The North Arm Mountain pluton has a narrow, moderately high temperature, metamorphic aureole along its basal contact which indicates that the ultrabasic or gabbroic material passing this contact during primary emplacement was at an elevated temperature and capable of altering the country rock. There is no evidence of a former chill zone. The primary features of the ultrabasic rocks near the basal contact are obscured by the post-serpentinization deformation of ultrabasic rocks parallel to the contact, but large grains of primary olivine suggest that the primary minerals crystallized elsewhere and were transported to their present position in the solid state.

The North Arm Mountain pluton has the overall structure of a gravity-stratified sheet. The layered ultrabasic zone, which is up to $2\frac{1}{2}$ miles thick (up to 4 miles in the Table Mountain pluton), shows no gravitative or cryptic layering. Banded structure is common, but is not related to differences in the specific gravity of the primary minerals nor to compositional variations in the primary minerals. Evidence suggesting that this banding is a flow structure has been presented above. The only general variation in the ultrabasic zone is that dunite is more common near the top than near the bottom, and this variation is not well defined. The presence of chromite concentrations near the top of the zone contradicts the expected specific gravity relationship and the oft-quoted hypothesis that the heavier ores (e.g. chromite) occur near the base of ultrabasic sills. Another variation is noticeable when the ultrabasic zone of the North Arm Mountain pluton is compared with that of Table Mountain (*see* Figure 9). The difference in thickness between the two zones, in view of their structural relationship (p. 60) shows that the ultrabasic zone of the North Arm Mountain pluton increases greatly in thickness with depth. It is not a simple, sill-like body, although the amount of thickening attributable to secondary deformation is not known.

The presence of disseminated hydrogarnet, altered feldspar, and layers of clinopyroxenite in the dunite near the top of the ultrabasic zone, shows a compositional change toward the overlying gabbro. These hybrid layers are not common, however, and, although the contact is interbanded, the change from an ultrabasic zone to a gabbroic zone, in view of the great thickness of the ultrabasic zone, is abrupt. The width of the 'critical zone' between the gabbroic and ultrabasic rocks varies. In numerous places the gabbroic rocks do not conformably overlie the ultrabasic rocks, and these local disconformities are due to deformation during and after emplacement.

The thickness of the gabbro zone is indeterminate because much of it is unbanded. The roof contact is generally flat, but local intrusive features are found. Banding is best developed near the ultrabasic zone, the upper parts of the gabbro zone being generally massive. The attitude of the banding is irregular in places. There is no systematic variation in the composition of the plagioclase. The genetic relationship of the gabbro to the ultrabasic zone is indisputable, despite local disconformable relations, because of the interbanded nature of the contact and the compositional variation within certain ultrabasic layers near the top of the ultrabasic zone.

EMPLACEMENT OF A TYPICAL PLUTON

The interlayering of gabbro with the underlying ultrabasic rocks is common everywhere, even in orogenic belts. It can be used in many places to determine the top of the enclosing sedimentary sequence. As gravitative differentiation is apparently the most logical explanation of this consistent layered relationship, it must be concluded that gravitative differentiation was important in the formation of the Bay of Islands Complex. Later deformation has modified the original form of the plutons and has even developed ultrabasic plutons, but it has not destroyed the evidence of the primary layered structure nor the relationship of the ultrabasic rocks to the gabbroic rocks.

The North Arm Mountain pluton is a pre-orogenic, relatively conformable intrusive body formed from a magma that differentiated at depth into an ultrabasic basal zone (olivine crystals plus interstitial fluid composed mainly of silica and water) and an overlying gabbroic zone (probably largely fluid but with some feldspar crystals). While the magma was in this state of incipient crystallization, but after the primary separation into gabbro and ultrabasic components, it was subjected to orogenic forces and intruded higher into the crust. The temperature at the time of this second emplacement was still high enough to metamorphose the wall-rocks and cause reactions among the components of the magma. The distance the magma travelled at this stage was probably not great. Homogenization of the ultrabasic layer was accomplished partly by mechanical mixing of the olivine mush during emplacement and partly by reactions with the interstitial fluid. Banded structure developed in the ultrabasic mush at this stage, and the reaction of the silica-bearing interstitial fluid with olivine during further cooling caused the crystallization of enstatite along stress planes. This formed the enstatolite layers, and the interstitial enstatite of the peridotites. The wedge-shape of the ultrabasic zone was also developed at this stage.

The gabbro fraction was probably more fluid and more easily injected into the crust. It advanced as a gabbroic forerunner of the ultrabasic mush, part of which was extruded to form basaltic flows, and also formed small gabbroic apophyses devoid of ultrabasic rocks in the roof of the intrusive body. Banding in the gabbroic rocks is best developed near the ultrabasic zone, suggesting a more advanced stage of crystallization in that part of the gabbro zone at the time of emplacement. Interbanding between the two zones is partly due to an incomplete separation of the gabbroic and ultrabasic components at depth, accentuated during intrusion by the development of planar flow structures in the hybrid mass. Reactions between the chemically dissimilar gabbroic and ultrabasic rocks prior to and during emplacement were localized in the 'critical zone' and caused the crystallization of spinel and diopside. Chromium, present in the late interstitial fluid, was able to enter the spinel structure and form local chromite enrichments. Although the layered nature of the pluton was generally maintained, it was not, at this later stage, emplaced as a horizontal body. Strong forces that caused active folding at the time of final emplacement also caused the ultrabasic mush to be emplaced somewhat passively as a tilted wedge possibly close to its present attitude.

After emplacement and serpentinization, the primary pluton was dismembered by faulting and folding. Tilting of the plutons caused post-serpentinization flowage along the basal contacts. Thrust faulting and associated transverse faulting destroyed the continuity of the complex, bringing bodies of different composition to their present structural positions. Thus, the nature of the parent pluton is obscured because each fault block represents only a part of the entire body.

The Bay of Islands Complex is a fossil remnant of a 'plateau-type' intrusive body that has been deformed in an orogenic belt. It illustrates one way in which secondary plutons that differ from their parent intrusive body and their parent magma may be developed by deformation. Deformation of the unsolidified magma has only partly destroyed its primary gravitative structure, and the relation of gabbro overlying ultrabasic rocks has been generally maintained. It is conceivable that in other areas the degree of autointrusion at this stage will be different, resulting in either a closer or more remote relation between the gabbro and ultrabasic phases of the magma. To carry the process of autointrusion to an extreme, the gabbroic fraction could be largely represented by basaltic flows in the eugeosynclinal belt. Post-emplacement deformation has further obscured the nature of the parent plutons, and left the complex as a series of non-typical plutons in fault blocks, some dominantly gabbroic, and others dominantly ultrabasic, in composition. It is clear therefore that major theories of petrogenesis cannot be based on the study of single, isolated igneous bodies in an orogenic belt. In the case of ultrabasic plutons, large areas of gabbro, composite, and ultrabasic plutons must be mapped in an orogenic belt, and their structural and time relationship determined, before their true magmatic history will be understood.

CHAPTER V

ECONOMIC GEOLOGY

General Statement

Minerals of possible economic interest associated with the igneous rocks of the Bay of Islands Igneous Complex include: asbestos, chromite, and copper-nickel sulphides in the ultrabasic rocks, and both copper-nickel and quartz-copper minerals in the gabbroic rocks.

In the Mount St. Gregory highlands, quartz veins containing pyrite, chalcopyrite, and arsenopyrite fill transverse faults cutting the roof of the pluton. Copper-nickel sulphide minerals form small concentrations near Winter House Brook, Bonne Bay, and the ultrabasic rocks of Table Mountain and the Lewis Hills. They also occur as minor accessory minerals in the gabbroic rocks. Quartz veins that contain pyrite, chalcopyrite, and sphalerite have been mined at the York Harbour mine on the northwest side of Blow Me Down Mountain. They were not examined in this study. Small wires of native copper are found in calcite veins in the Humber Arm volcanic rocks.

Chrysotile asbestos occurs only locally and in small amounts along the margins of the plutons. Other veins are found in the sheared ultrabasic bands associated with banded gabbro north and south of Trout River. Asbestos veins are best developed near the southern part of the Complex where the ultrabasic rocks are cut by later porphyritic diabase dykes.

Chromite and related minerals of the spinel group constitute on the average about 1 per cent of the average ultrabasic rocks. Massive lenses or layers of chromite are restricted to the upper parts of the ultrabasic zones just below the gabbroic layered zone. This spatial relationship is interpreted as due to a favourable chemical environment in the upper part of the ultrabasic zone.

Copper

Copper mineralization occurs as: (1) pyrite-chalcopyrite-arsenopyrite-sphalerite-quartz veins associated with quartz diorite intrusions; (2) chalcopyrite-nickeliferous pyrrhotite veins and disseminations associated with gabbro; and (3) native copper in calcite veins within the basic volcanic rocks.

The most important deposits are the quartz veins found near Gregory River and York Harbour. The York Harbour deposits have not been studied by the writer and the last published account of them is given by Cooper (1936). The other localities mentioned below indicate the extent of copper and nickel mineralization following the intrusion of the gabbroic rocks.

GREGORY RIVER COPPER LODES

Copper occurrences were discovered in the Mount St. Gregory area in 1921 by tracing float up Gregory River to its source, and several additional occurrences were discovered in 1922. In 1929, M. F. Bancroft studied the geology of these deposits and an equipotential survey of the Court A and Powell Brook areas was carried out by the Swedish American Prospecting Company of Canada. The mineralized areas were under lease by the Cape Copper Mines Limited in 1953, who drilled the Court A and Mitchell lodes and undertook some surface exploration at the Narrows. The other mineralized areas mentioned below have not been actively explored for over 20 years, and are so overgrown by bush and obscured by sliderock that a detailed examination was not possible.

The mineralized belt is about 6 miles long and extends northerly from Mitchell Brook to the Narrows, most of the veins being in the southern part. They fill fractures that cut the roof of the North Arm Mountain pluton. The contact between the medium-grained gabbros of the Bay of Islands Complex and the overlying, fine-grained, basic metavolcanic rocks is irregular in detail, but it has a regional westerly dip that steepens near the mineral deposits. The economic minerals occur near the roof contact on both sides of it. The Narrows, Ore Brook, and Court B lodes are in the gabbro, Number 9 lode is very near the contact, and the Court A and Mitchell lodes are in the metavolcanic rocks (*see* Figure 14). The mineralization is chiefly of the hydrothermal cavity-filling type, and only minor amounts of sulphide minerals are disseminated in the adjacent wall-rocks. Pyrite, chalcopyrite, and arsenopyrite are the principal primary minerals, and occur in a matrix of quartz and calcite. The mineralization may be related to quartz diorite intrusions that cut the gabbro near the mineral deposits. Except for the lack of sphalerite, the deposits resemble those of York Harbour, near the Blow Me Down pluton.

Structural controls are important in the localization of the copper deposits. The Court A lode is localized along a west-striking transverse fault that can be traced from the Court A lode to Gregory River. This fault is interpreted as a tear fault related to the major Gregory River fault. The Jumbo and Number 9 lodes are closely associated with the main fault. Quartz boulders several feet in diameter were noted in streams crossing the Gregory River fault. These are further evidence of hydrothermal activity along the Gregory River fault and its associated tear faults.

Possibly because it is inaccessible along the steep western wall of the Gregory River valley, this fault has not been adequately explored.

Figure 14 shows the locations of the principal mineral showings described below. The Narrows lode occurs north of the area shown in the Figure.

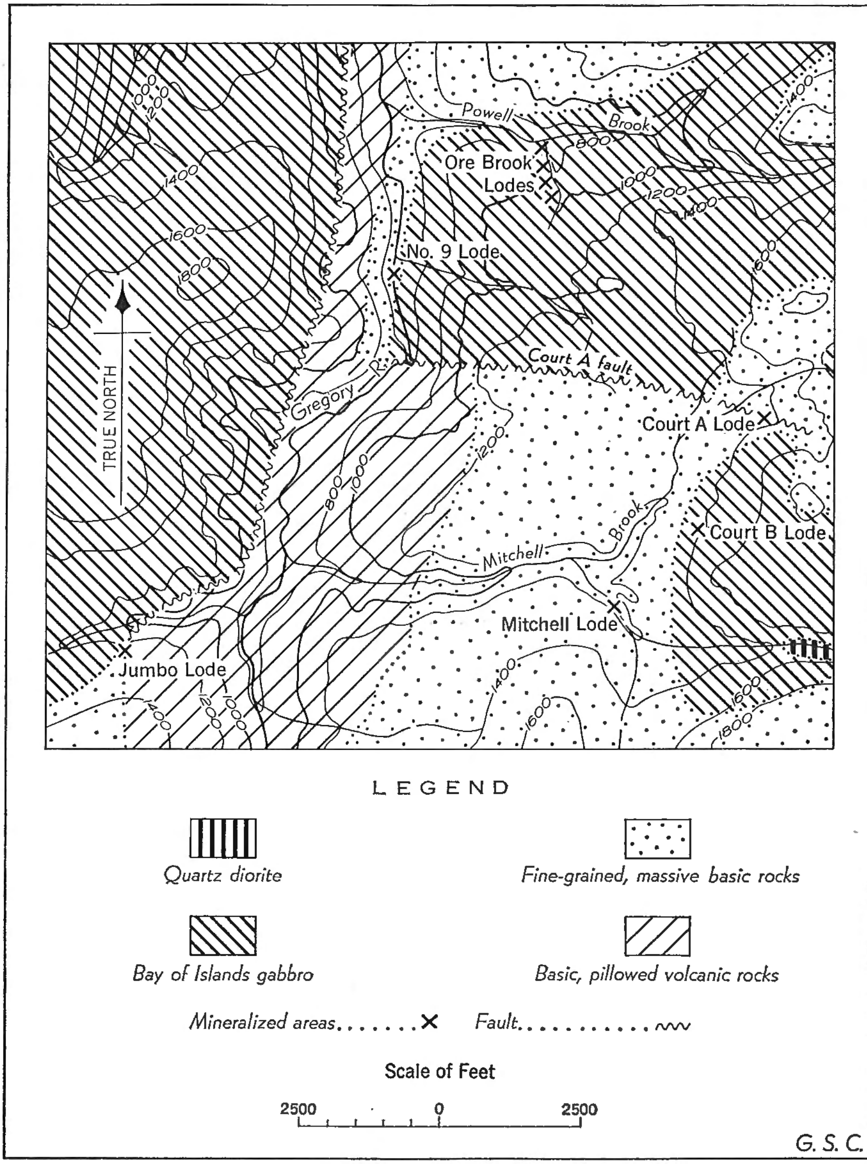


Figure 14. Geological map of area near Cape Copper Mines Ltd.

Court A Lode

The Court A lode is on a branch of Mitchell Brook. Sulphide minerals are exposed in the bed of the stream for over 100 feet, but the stream alluvium prevents detailed examination of the outcrops. Diamond drilling has extended the known limits of the mineralized zone for 1,000 feet along strike. The controlling structure dips steeply, but its direction of dip is not known. The sulphide minerals, pyrite, chalcopyrite, and arsenopyrite, occur in quartz veins and disseminated through the metavolcanic host rocks.

The Court A lode is localized along a west-trending transverse fault, the Court A fault, related to the Gregory River fault. This fault is only slightly mineralized where it crosses Gregory River. On the plateau between the Gregory River valley and the Court A lode it is expressed topographically by a slight linear depression occupied by a small pond. There are no outcrops along the depression but medium- to coarse-grained leucogabbros outcrop nearby on the northern side of the fault. Gossan float was found about 600 feet west of the pond, near the depression.

Jumbo Lode

The Jumbo lode is on the south wall of the valley of Jumbo Brook, a tributary of Gregory River. It is localized near the Gregory River fault, which there forms a highly sheared contact between the gabbroic and volcanic rocks. Two abandoned adits mark the site, but they were not entered. Specimens of massive ore from the dump consist of pyrite, chalcopyrite, and arsenopyrite.

Mitchell Lode

The Mitchell lode is on the southwest bank of Mitchell Brook, in the fine-grained, basic metavolcanic rocks that form the roof of the pluton. It has a general easterly trend, similar to the Court lodes. It consists of pyrite, chalcopyrite, arsenopyrite, and quartz. Diamond drilling by the Cape Copper Mines Limited failed to trace the structure across the brook, but there is a mineralized zone on the opposite bank, about 200 feet upstream from the Mitchell lode outcrops. It is up to $1\frac{1}{2}$ feet wide and contains quartz, pyrite, and chalcopyrite. Only further exploration will show whether or not it is the offset extension of the Mitchell lode or a separate vein.

Narrows Lode

The Narrows lode is on the Narrows Head, overlooking Trout River Pond. The country rock is a medium- to coarse-grained, fresh, massive leucogabbro. A cap of basic metavolcanic rocks overlies the gabbro and forms the top of the Head. The main showing occurs in a strong shear zone that strikes north 75 to 80 degrees east and dips 65 to 70 degrees to the south. The zone of shearing is irregular but well defined. It extends along the hillside for at least 400 feet. It pinches and swells, and encloses blocks of sheared gabbro. The hanging-wall consists of massive barren quartz, up to 8 feet thick. Below this are small discontinuous sulphide lenses containing pyrite, chalcopyrite, and minor arsenopyrite. The sulphide minerals have been leached locally and in places small coatings of malachite, azurite, cuprite, and native copper have formed. The extent of this secondary alteration is very minor.

Number 9 Lode

Number 9 lode is on the west bank of Gregory River just above river level, very close to the intrusive contact of coarse-grained gabbros and fine-grained metavolcanic rocks. It is now largely grown over, so that no information was obtained on the extent or trend of the mineralization. Ore specimens collected from the dump consist of massive pyrite, with quartz and carbonate minerals.

According to Douglas *et al.* (1940): "It is exposed for some 50 feet along strike, is probably not less than 10 feet thick, and dips northwest at 70 degrees". It thus dips toward the easterly dipping Gregory River fault. If the Gregory River fault is of pre-mineralization age, the mineralization of the Number 9 lode may be due to solutions that rose along the Gregory River fault.

Ore Brook Lodes

Ore Brook is a small tributary stream that drains north into Powell Brook. At least four lodes, the Hall, Palmer, Number 6, and Number 7 lodes, were exposed in the valley of this stream, according to descriptions in old reports, but it is not now possible to identify them separately. Numerous small mineralized quartz veins occur along the bed of the stream, but because of the abundant sliderock and vegetation on the valley walls their economic value is not known. The bedrock consists of greenish, medium- to coarse-grained leucogabbro that has been brecciated and intruded by diorite and quartz diorite. Later quartz veins, cutting both the gabbro and the quartz diorite, contain pyrite, chalcopyrite, and arsenopyrite.

Two well-developed gossans, 5 to 6 feet wide, occur in the upper part of the western valley wall near the mouth of Ore Brook, and near the head of a tributary stream west of Ore Brook. No attempt has been made to trace these mineralized zones under the overburden. Although the trend of the mineralized veins is rather variable, northeast strikes are dominant.

WINTER HOUSE BROOK, BONNE BAY

Veins of massive sulphide minerals outcrop on the southeast side of Winter House Brook about 3,000 feet upstream from the bridge on the Woody Point–Shoal Brook road. At this point the stream cuts across an area of coarse-grained gabbro that is in fault contact with shales and sandstones of the Humber Arm group both downstream and upstream. The gabbro forms a series of waterfalls at its downstream contact. Above the first fall, along the southeast bank, veins and irregularly shaped masses of sulphide minerals are exposed for 5 feet above water level, above which they are concealed by overburden. About four mineralized areas were noted over a distance of 40 feet along the bank. One of these is 6 inches wide and 2 feet long, and the others are of comparable size. Their inaccessibility and the lack of outcrop farther from the river bank make it impossible to assess the economic significance of the mineralization except as a guide to further prospecting.

Two assays of randomly chosen samples from the outcrop area and a third of float gave the following results:

Field Number	% Cu	% Ni	% Pt. group	Au (oz)
Float	2·50	1·25	Tr.	Tr.
B.B.1	1·48	1·33	Tr.	0·005
B.B.2	1·79	1·54	Tr.	0·003

Study of polished surfaces shows that the specimens consist dominantly of lamellar pyrrhotite that has been fractured and cut by veins of chalcopyrite and pyrite. No pentlandite was recognized and it is possible that much of the nickel is present in the pyrrhotite.

NORTH SIDE OF TABLE MOUNTAIN

Copper minerals are found at the northwest end of the Gulch, about 4 miles along the telephone line from Winter House Brook. They occur near the base of the Table Mountain scarp in highly sheared serpentinite. The shearing is part of the fault zone along the northern boundary of Table Mountain.

The copper mineralization is exposed at the top of a sliderock slope between two branches of a small stream that drains the Table Mountain scarp. Malachite stains cover 10 square feet of the bedrock, and mineralized float is found in the sliderock. The extension of the mineralized outcrop was not determined, but the ore is strongly magnetic and could be easily traced with a magnetometer. Under the microscope, the fresher specimens are seen to consist mainly of cubanite veined with bornite, and chalcocite. More oxidized specimens contain magnetite, cuprite, native copper and malachite.

An assay of one of the better mineralized specimens from the area yielded the following results:

Cu	8·0%
Ni	0·05%
Au	0·01%
Palladium	Faint trace

SOUTH SIDE OF SECOND TROUT RIVER POND

An area of limonite stain occurs on the bank of a stream that drains into Second Trout River Pond at the Rubbles. The sulphide minerals in this locality are non-magnetic pyrrhotite (troilite) with minor blebs of chalcopyrite. They occur as irregular masses interstitial to the silicate minerals. The deposit is probably too small to be of economic value.

ROPE COVE CANYON

Oxidized copper minerals occur in sheared dunite near the head of Rope Cove Canyon. A caved pit now marks the outcrop area and only a few mineralized fragments from the dump can be found on the sliderock slopes. According to Cooper (1936): "Most of the ore fragments on the dump are cuprite-native copper but polished sections show the following paragenesis: chalcopyrite, chalcocite and/or melaconite, native copper, cuprite, tenorite, and malachite". The ore fragments are strongly magnetic, probably due to the presence of magnetite. The mineralization resembles that found in sheared ultrabasic rocks on the northern side of Table Mountain.

HUMBER ARM VOLCANIC ROCKS

Small wires of native copper are found in calcite veins that cut the basic volcanic rocks of the Humber Arm group along the entire coastal area. They are of no economic value.

Asbestos Deposits

Asbestos deposits are scarce in the Bay of Islands Complex, in comparison with the abundance of exposures of ultrabasic rocks in the area. The larger deposits occur in the Mine Cove–Lewis Brook section at the southern end of the Complex, and in a belt of deformed, interbanded ultrabasic and gabbroic rocks that outcrop along the coast, north and south of Trout River. Asbestos occurrences discovered near Lark Harbour on the south side of the Bay of Islands may be extensions of this coastal belt. Surface and underground exploration has been concentrated in the small but complex area of ultrabasic and gabbroic rocks that outcrop between Lewis Brook and Mine Cove.

Asbestos was first reported from this locality by Alexander Murray in 1873, and since then there have been numerous attempts to exploit both the asbestos and chromite resources of this area. During the period 1946 to 1948, Asbestos Corporation carried out diamond drilling and underground exploration. The caved remains of at least nine prospect shafts were found on the seaward-facing slopes and two adits and numerous test pits have been driven into the west wall of the Lewis Brook valley. The latter location is currently being explored by Newfoundland Asbestos Limited.

LEWIS BROOK-MINE COVE AREA

The Lewis Brook-Mine Cove asbestos deposits occur in a wedge-shaped area of intrusive rocks bounded by Lewis Brook valley on the east and the Gulf of St. Lawrence on the west. Asbestos occurs on the east and west slopes of a flat-topped hill up to 1,400 feet high. The geology of this area is not well known. It is one of great structural and petrographic complexity and its interpretation will require a more detailed study. The metavolcanic rocks of Bluff Head and the metagabbroic rock types of the Big Level occur there, closely associated with fresher, banded gabbro and serpentinite. These rocks are deformed and cut by numerous diabasic dykes that in places cannot be distinguished from the other types of gabbroic rocks. Also, slump terraces and rubble of ultrabasic and gabbroic blocks that occur along the east and west slopes of the ridge in the vicinity of the asbestos occurrences prevent the correlation and mapping of individual units. The nature of the contact between ultrabasic rocks and underlying Humber Arm volcanic rocks at Mine Cove and the presence of sedimentary rocks in the valley of Lewis Brook downstream from the asbestos deposits, suggest that a relatively flat floor of Humber Arm rocks may underlie the entire area. Cooper (1936) interprets this contact as part of the Lewis Hills overthrust.

Asbestos occurrences in the western part of the area occur along the irregular slopes that face the Gulf of St. Lawrence. A coastal zone of Humber Arm sandstone, red shale and volcanic rocks outcrops at the mouth of Bluff Head Brook on the south and Lewis Brook on the north. Between these brooks, except for a few outcrops of Humber Arm rocks near or just below sea-level, the igneous rocks of the Complex extend to the sea-coast. They consist of alternating serpentinite, banded gabbro and diabase dykes with serpentinite forming less than half of the exposures seen. Cross-fibre asbestos up to 1 inch, but averaging $\frac{1}{16}$ to $\frac{1}{8}$ inch, in length, occurs in the boulders that form the beach near the serpentinite outcrops. Ribbon veins up to 1 foot in aggregate thickness are exposed on the cliff faces along the beach. Locally they occur near intrusive diabase contacts. Farther inland, the slopes are extremely rugged and the geology is obscured by the rubble of numerous slump blocks. Although good fibres are found in local exposures, individual serpentinite bands could not be traced because of their close association with different types of gabbroic rocks.

In 1952, rock outcrops were too few for mapping along the steep valley wall of Lewis Brook in the neighbourhood of the Newfoundland Asbestos Limited deposits. Two adits had been driven into the serpentinite of the valley wall, but these were largely timbered. According to local reports on the deposit, two 'troctolite' dykes were intersected by the adits and the major asbestos concentrations were found in the serpentinite near these dykes. A sample of this 'troctolite' collected from the dump contained colourless pyroxene partly altered to light green actinolitic amphibole, and colourless to dusty prehnite. The rock has a relict ophitic texture in which rectangular feldspar laths have been completely altered to prehnite. As there are no relict crystals of olivine, the dyke cannot be classified as a troctolite, but rather it resembles other diabase dykes that outcrop along the coast.

OTHER ASBESTOS AREAS

Other loci of asbestos mineralization that might be interesting to the prospector are as follows:

(1) Along the faulted margins of the major plutons. Along the northeast side of Blow Me Down Mountain, northwest and southeast of Blow Me Down Brook, there are numerous small cross-fibres in strongly deformed peridotite. The fibres, up to $\frac{1}{4}$ inch in length, are in ribbon bands up to 3 inches wide and occur for a considerable distance along the fault. Asbestos fibres are also found along the faulted margins of the Table Mountain pluton, from First Trout River Pond to Winter House and Shoal Brooks. Numerous asbestos veins up to $\frac{1}{16}$ inch long are found in the deformed ultrabasic rocks of this section.

(2) Asbestos concentrations occur in small ultrabasic layers in the gabbro that extends along the coast from Chimney Cove to north of Trout River. This is a zone of altered and deformed banded gabbroic rocks. Serpentinite layers form a minor part of this banded zone and asbestos occurs in places in the serpentinite. Along the northeast shore of Trout River Bay there are at least two serpentinite bands that contain small amounts of asbestos fibre up to $\frac{1}{2}$ inch long. Locally the fibres form ribbon bands up to 6 inches wide. Near Wallace Brook, northeast of Trout River, a serpentinized band in the banded gabbro complex contains fibres up to $\frac{3}{4}$ inch long. South of Trout River are several other similar asbestos occurrences in which fibres up to $\frac{1}{4}$ inch wide form stockworks in the serpentinite. At all the localities examined in this coastal belt, the serpentinite occurs as small bands, less than 100 feet thick, in the gabbroic rocks. The asbestos fibre generally forms only 1 per cent or less of the serpentinite.

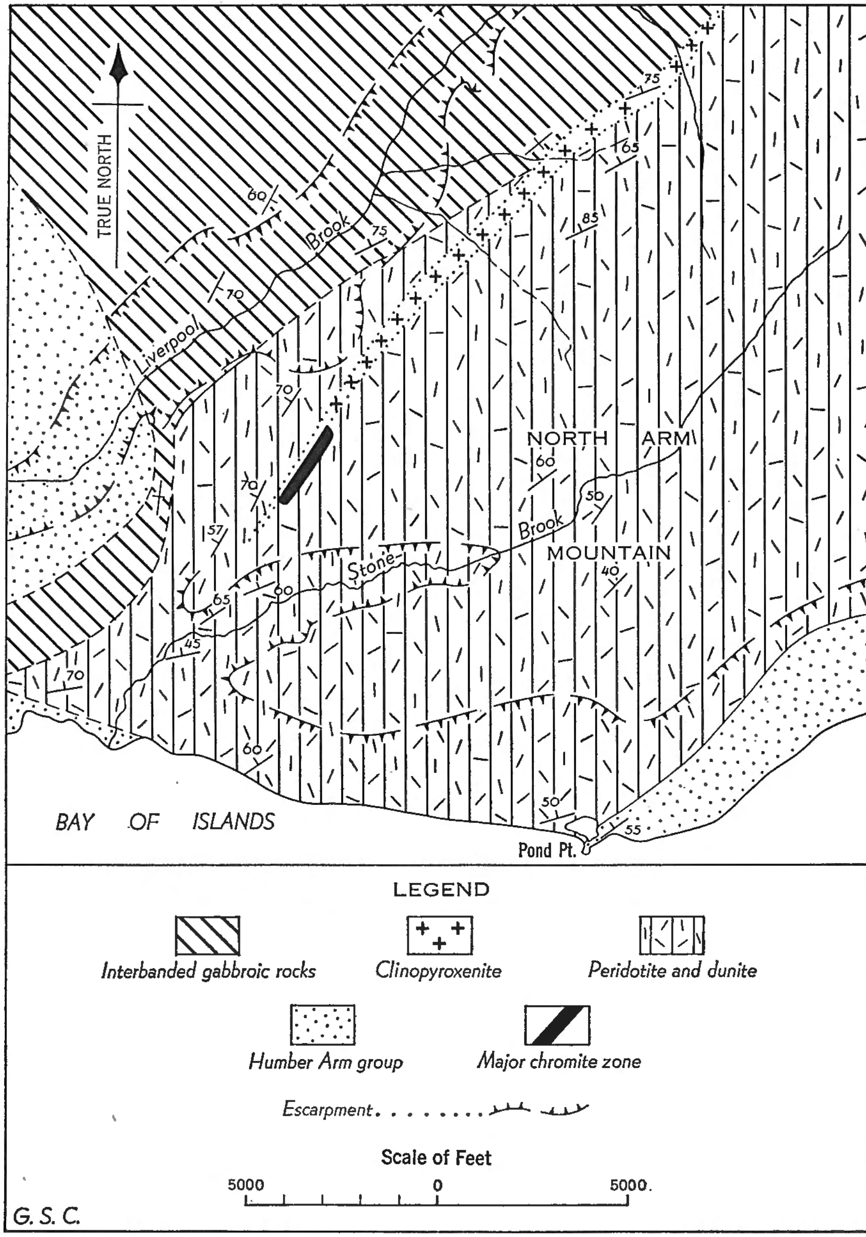


Figure 15. Geological map of Stowbridge chromite deposit.

Chromite Deposits

TABLE MOUNTAIN

No chromite concentrations are known in the ultrabasic rocks of the Table Mountain pluton.

NORTH ARM MOUNTAIN

Stowbridge Chromite Deposit

The Stowbridge deposit, which comprises the largest concentration of chromite in the North Arm Mountain pluton, is at the southwest end of North Arm Mountain. It is 1½ miles from the shore of the Bay of Islands, at a mean elevation of 1,500 feet. The deposit is accessible by boat from Corner Brook, a distance of 22 miles. North Arm and much of the Bay of Islands are frozen over during the winter months.

The area surrounding the deposit is a barren, undulating surface up to 2,100 feet in elevation. Outcrops are partly obscured by large areas of low grass and frost-riven boulders.

The chromite concentrations occur as lenses in the upper part of the ultrabasic zone (*see* Figure 15). They are over 2 miles stratigraphically above the base of the pluton and about 1,400 feet stratigraphically below the main contact of the ultrabasic zone with the overlying banded gabbro zone. The chromite lenses have a general north-northeasterly trend and dip 60 to 80 degrees to the north-northwest. This is parallel to the regional trend of the banding in the ultrabasic rocks, although the trend is modified in places by drag effects on faults that border the ultrabasic zone on the southwest.

The chromite lenses occur in dunite although minor concentrations are found in closely associated clinopyroxenite. Peridotite layers occur near, but not in, the chromite zone. Feldspathic dunite forms bands above and below the chromite zone, but no feldspar was found associated with the chromite. Enstatolite bands are also found in the adjacent ultrabasic rocks. A close associate of the chromite concentrations is a zone of coarse-grained, green clinopyroxenite that is locally parallel to the banding in the ultrabasic rocks. The clinopyroxenite thickens to the northeast, and cuts across the regional trend until it forms the basal layer of the banded gabbro zone. Still farther northeast, it thins and finally pinches out. Individual clinopyroxenite lenses occur both above and below the chromite zone. The clinopyroxene of these lenses is near pure diopside.

Eight exposures, separated by large areas of frost-riven boulders and containing small chromite bands, occur within a distance of ¾ mile. The smallest outcrops are 10 to 20 feet in diameter, the largest 180 by 80 feet. A typical outcrop is illustrated in Figure 16. The chromite zone is not

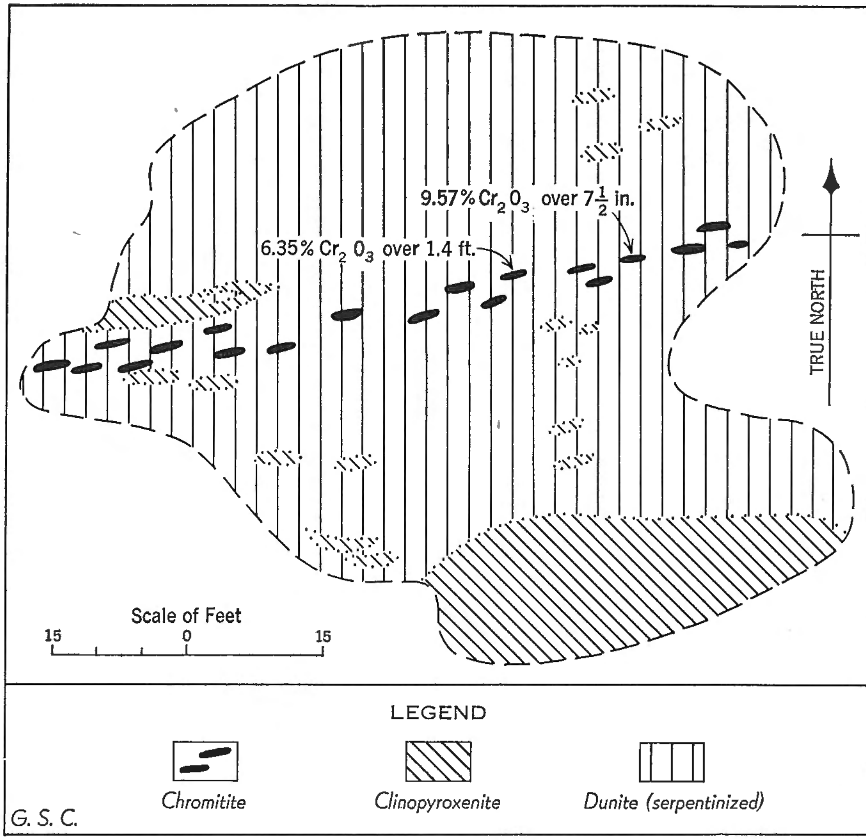


Figure 16. Sketch map illustrating typical chromite-bearing outcrop at Stowbridge deposit. Note that chromite occurs with clinopyroxenite as a series of discontinuous lenses in dunite.

composed of persistent chromite bands such as those that characterize the Bushveld and other stratiform deposits, but rather of an aggregate of small, closely spaced lenses in dunite. The zone of lenses averages 3 feet in width with a maximum width of 30 feet; the individual lenses of massive chromite vary in width from a single grain to 4 inches, and in length from a few inches to almost 10 feet. The chromite lenses, generally regular in trend, are in places warped into typical flowage folds with pronounced thickening of the crest and troughs. Some of the bands split into several smaller parallel bands before pinching out. Serpentine, derived from olivine, forms a large part of the chromite zone, a fact that accounts for the low assays obtained from chip samples cut across the entire chromite zone (*see* Table II).

Table II

Analyses of Channel Samples, Stowbridge Chromite Deposit

Sample No.	Width	% Cr_2O_3	Pt. group
H200	1·4 ft.	5·36	N.D.
H201	1·4 ft.	6·35	N.D.
H202	3·2 ft.	2·39	Tr.
H203	7½ in.	9·57	"
H204	1·9 ft.	3·11	"
H205	7·0 ft.	7·77	"

Chip samples cut across several of the better mineralized zones. Analyses by Division of Mineral Dressing and Metallurgy, Dept. Mines and Tech. Surveys, Ottawa.

The primary silicate minerals in the chromite zone have been largely destroyed by serpentinization, and the chromite is enclosed in serpentine or kaemmererite. In the main ultrabasic zone, where accessory chromite is found associated with primary silicate minerals, the chromite grains may be enclosed by enstatite or may themselves enclose olivine (*see* Plate X A). They are commonly anhedral, with a vermicular form (*see* Plate X B).

The chromite is a magnesian-rich variety. In thin section it is seen to vary in colour from yellowish or reddish brown to black. Opaque rims surround the grains, and their borders commonly have a pronounced serrate form. The lack of uniformity of the chromite grains is shown by differences in colour, and response to concentration tests. When concentrated on the Haultain superpanner, successive chromite fractions of different specific gravity are obtained. Each of these fractions, in turn, can be broken down into portions of different magnetic intensity. The chemical differences between fractions were not determined, but this inhomogeneity shows that analyses of chromite concentrates are not typical of the composition of the ideal chromite molecule. It is possible that these differences in the chromite grains are the result of alteration during serpentinization.

Because of the irregular distribution of chromite in the chromite zone, and the lack of good exposures, it is not practical to attempt to estimate the amount of chromite in this deposit. The low tenor of the chip samples given above suggests that there is no chromite of direct shipping grade, and only a small tonnage of concentrating ore could be expected. The probable grade of concentrates is about 41 per cent Cr_2O_3 with a maximum chrome-iron ratio of 1·70 to 1. The concentrates would have to be chemically beneficiated to raise them to metallurgical grade. The poor quality of chromite obtained from this deposit, and the small tonnage of low-grade chromite exposed indicate that it is of no economic value at the present time.

Table III

Analyses of Stowbridge Chromite Concentrates

—	No. 1 Sample	No. 2 Sample
Cr_2O_3	41·66	40·26
Al_2O_3	21·92	19·34
Fe*	16·78 {FeO–14·1, Fe_2O_3–7·8}	19·25 {FeO–13·5, Fe_2O_3–12·5}
MgO	13·53	13·90
CaO	—	Tr.
SiO_2	0·39	1·92
Mn	Strong trace	
Ti	" "	
Na	" "	
V	" "	
Ni	Trace	
Cu	Faint trace	
Ga	" "	
Pt	None detected	
Cell dimension (A_0)	8·249	
Cr/Fe	1·70	1.43

(1) Sample X3: Analyst: H. B. Wiik (Chemical); Geol. Surv., Canada, (Spectrographic).
(2) Analyst: E. J. Lavino & Co. (Snelgrove, 1934).
* Only total iron determined.

BLOW ME DOWN MOUNTAIN

The chromite lenses in the Blow Me Down Mountain pluton are only of scientific interest and indicate the parts of the ultrabasic body in which conditions were most favourable for the crystallization of chromite. No lenses or groups of lenses of economic value were found.

Figure 17 shows the parts of the pluton in which small, scattered, non-economic chromite lenses occur. In this zone lenses of massive chromite over an inch wide are rare. Some bands contain 10 to 20 per cent disseminated chromite over widths of less than 2 feet. These bands are rarely over 100 feet long and most pinch out or are obscured by the felsenmeer that covers the surface. None the less, compared with stratigraphically lower parts of the ultrabasic outcrop area to the east, the difference in abundance of chromite is pronounced. In this eastern part outcrops are abundant but not even small chromite concentrations were found.

Chromite lenses are limited to an outcrop area up to 2 miles wide adjacent to the Blow Me Down gabbro. The area is offset by faulting, and apparently is not of constant width across the pluton. This zone of

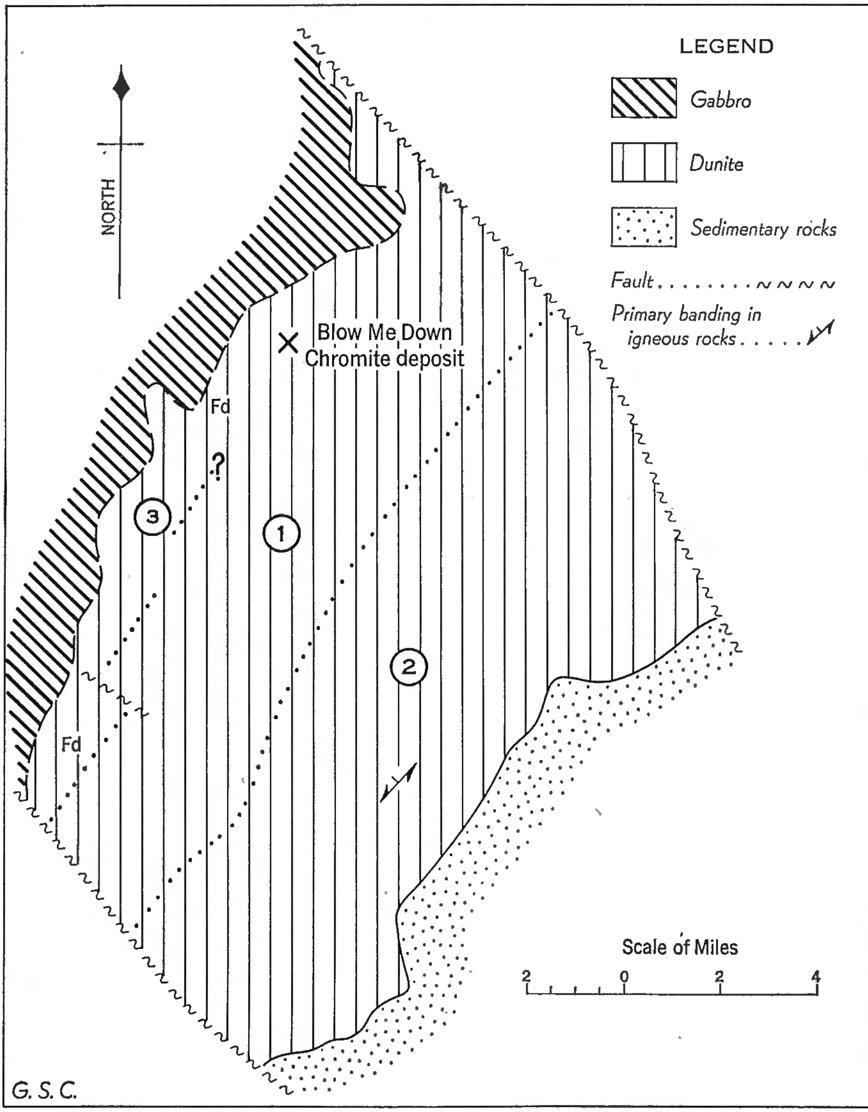


Figure 17. Sketch map illustrating spatial localization of chromite in Blow Me Down pluton.
1. Area in which small, non-economic chromite lenses occur.
2. Ultrabasic rocks barren of chromite.
3. Area in which feldspathic dunite (Fd) layers occur.

minor chromite concentrations is up to 6,000 feet thick and lies below the main gabbro contacts. The chromite occurs in dunite, which is far more abundant than peridotite in the upper part of the ultrabasic zone. Feldspathic dunite bands in the upper part of the chromium-enriched zone form a marginal facies underlying the gabbro along parts of the ultrabasic-gabbro contact. They may be as much as 2,000 feet below the gabbro contact or entirely absent, the lowest appearance of feldspar being in the gabbro proper.

The largest chromite lens in this zone, since removed by mining, was exposed on the peneplain surface near a scarp of the valley of Blow Me Down Brook. Snelgrove (1934, p. 13) states, "One hundred tons are said to have been mined from a lens of nearly solid chromite which was eight feet wide and twenty-one feet long at the surface". No chromite can be seen in outcrops at the pit, but a nearby dump contains several tons of chromite boulders up to a foot in diameter. A trench 900 feet southwest of the pit contains disseminated chromite, but the intervening rocks are barren. The trench is now largely filled by debris, but samples from it are reported by Snelgrove to have assayed 12% Cr_2O_3 over a 3- to 5-foot width. Nearby outcrops show that the concentrations have no great length along strike.

A few coarse-grained pyroxenite bands occur in the neighbourhood of the chromite concentrations. They are smaller than the Stowbridge pyroxenites and are composed of enstatite rather than diopside.

In thin section the chromite is yellowish orange, and enclosed in a matrix of kaemmererite and antigorite. Chemical analyses (*see* Table IV) indicate that it is a low-grade chromite. By mechanical concentration the grade can be raised to 32% Cr_2O_3, with a chrome-iron ratio of about 1·3/1 to 1·9/1. The high alumina content shown by analyses of samples 2 and 3 suggests the mineral should be called a chromian spinel rather than chromite.

LEWIS HILLS

Small chromite lenses are more abundant and widespread in the Lewis Hills pluton than in the other members of the Complex.

The two principal concentrations are located on the sides of the Fox Island River valley, about 13 miles from the mouth of the river. The deposit on the southeast side is known as Chrome Point, that on the northwest side is known as either the Compagnon deposit or the Springers Hill deposit. In addition to these, there are many small concentrations of chromite west of Hines Pond and southwest of Petleys Pond. The latter occur in dunite, close to bodies of feldspathic dunites and pyroxenites. The largest of these concentrations, south of Petleys Pond, is about 200 feet long and 2 feet wide. Other small concentrations, commonly composed of about 50% chromite over a width of 3 to 5 inches and a length of 10 to 25 feet, occur along the Fox Island River valley wall northeast and southwest of the Springers Hill deposit, and also on top of the mountain between the river and the overlying gabbroic rocks to the north. In no place are they spaced closely enough to be mined as an economical unit.

Table IV

Analyses of Blow Me Down Chromite Concentrates

	No. 1 Sample	No. 2 Sample	No. 3 Sample
Cr_2O_3	32·56	32·27	31·52
Al_2O_3	18·63	32·60	32·18
Fe*	17·06	11·43	11·74
MgO	22·53	17·91	17·97
CaO	—	—	—
SiO_2	0·90	1·77	2·05
Mn	Strong trace		
Na	" "		
Ti	Trace		
V	"		
Ni	"		
Cu	Faint trace		
Ga	" "		
Pt	None detected		
Cell dimension (A_0)	8·207		
Cr/Fe	1·31/1	1·94/1	1·84/1

(1) Sample X2, Analyst: H. B. Wiik (Chemical); Geol. Surv., Canada, (Spectrographic).
(2) Analyst: E. J. Lavino & Co. (Snelgrove, 1934).
(3) Analyst: A. E. LaRochelle, Mines Branch, Ottawa.
* Only total iron determined.

Springers Hill Deposit

The Springers Hill deposit occurs on the north side of Fox Island River, in an area of dunite with some peridotite layers. A road now under construction between Point au Mal and the Lewis Brook asbestos deposit forms the most convenient means of access. From this road at the Cache, the showings are 2 miles by trail to Fox Island River and thence 5 miles along the river to Springers Hill. The layering in the ultrabasic rocks there has a regional trend of north 5 to north 15 degrees east and a westerly dip of 30 to 45 degrees. Trenches dug in 1941 were almost obliterated by sliderock when examined in 1952 and it was impossible to make a quantitative evaluation of the deposit. Massive chromite fragments up to 5 inches in diameter were seen in the sliderock but they were not abundant enough to suggest the presence of large chromite concentrations. An analysis of one of these, given in Table V, shows that although the chrome content is good, the chrome-iron ratio is low and the ore would require chemical as well as mechanical beneficiation. Horning (1941) calculated a tonnage of 11,897 tons averaging 36·97% Cr_2O_3 based upon drilling and assay results from the two largest lenses.

Smaller chromite concentrations occur for several miles along the strike of the Springers Hill deposit. The chromite is present as small layers in the dunite of the valley wall, but no continuous chromite zone exists. In thin section, the chromite is deep red, slightly altered, and enclosed in a matrix of serpentine. No primary silicate minerals were found in the chromite zones.

Chrome Point Deposit

The Chrome Point deposit forms a small hill on the southeast side of Fox Island River, south of Springers Hill. It occurs near the margin of the pluton, in a zone of dunite and peridotite with a few bands of feldspathic dunite, pyroxenite, gabbro. These have been strongly deformed by stresses localized near the margin. The regional trend is northeasterly and, where undeformed, the primary banding generally dips northwest or vertically. A poorly defined zone of feldspathic dunite, pyroxenite and gabbro outcrops east of the deposit.

The Chrome Point deposit consists of small lenses of chromite in dunite. Bedrock is only exposed on the top of the hill, as the sides are covered with sliderock. Two bands of fairly massive chromite outcrop on

Table V

Analyses of Fox Island River Chromite Concentrates

	No. 1 Sample	No. 2 Sample
Cr_2O_3	43·39	36·85
Al_2O_3	13·40	15.24
Fe*	20·56	13·09
MgO	14·09	17·17
CaO	—	—
SiO_2	0·98	8·76
Mn	Strong trace	
Na	" "	
Ti	Trace	
V	"	
Ni	"	
Cu	Faint trace	
Ga	" "	
Pt	Not found	
Cell dimension (A_0)	8·272	
Cr/Fe	1.45	1·93

(1) Springers Hill concentrate. Analyst: H. B. Wiik (Chemical); Geol. Surv., Canada, (Spectrographic).
(2) Chrome Point: Analyst: Arnold (Meissner, 1901).
* Only total iron determined.

the hill and are about 4 inches wide and possibly 100 feet long. Snelgrove (1934, p. 16) reports forty masses of chromite that "vary from half an inch up to three feet in width, from a few inches up to eighteen feet in length, and from 10 per cent to 75 per cent in tenor of chromite". Thin sections of the ore show orange translucent chromite grains in a matrix of serpentine minerals.

MINE COVE-LEWIS BROOK AREA

Several chromite and asbestos occurrences outcrop in the area of ultrabasic rocks exposed near the coast, between Mine Cove and Lewis Brook. This is an area of great structural and petrographic complexity in which ultrabasic rocks, gabbro, and metavolcanic rocks are so closely associated that their separation on the present scale of mapping is not possible. Slump terraces and rubble composed of ultrabasic and gabbroic material extend several thousand feet inland and obscure the bedrock relationships. It is, therefore, impossible to trace chromite or asbestos occurrences for any distance in this area. The area is one of the most favourable parts of the entire Complex for the localization of chromite, but due to the deformation of the serpentinite it is unlikely that any of the pods are sufficiently continuous to form a large deposit.

The Mine Cove area is accessible from Point au Mal either by boat or by tractor road via the Lewis Brook asbestos deposit. There are no sheltered harbours along this part of the coast, and the boat trip can be made only in fine weather. The coast is rugged and at Bluff Head the abrupt sea-cliffs are approximately 1,600 feet high.

There are at least four chromite concentrations in the area. The chromite occurs in serpentinite interbanded with gabbro, a petrographic environment similar to that found to be most favourable for chromite concentration in the other members of the Complex. The attitude of banding in the gabbroic rocks of this area is extremely variable.

The Bluff Head chromite deposit is at the head of Mine Cove canyon, the cirque-like valley of Bluff Head Brook, at an elevation of approximately 1,000 feet. It was the first deposit in the area to attract attention, and it is reported (Murray and Howley, 1918) that 1,000 tons of ore of unstated grade were mined in 1895. The caved remains of the old workings may still be seen, partly covered by sliderock from the canyon wall and filled with water. They were investigated by R. Wiseman (1942) who reported "only two small lenses of low-grade chromite were seen throughout all the workings and the whole would not amount to more than ten or twenty tons. The bottoms of the winzes are, however, not accessible."

The canyon wall near the chromite deposit is composed of serpentinite cut by several altered feldspar porphyry dykes with chilled margins. The serpentinite is slickensided, and contains small asbestos veins. On top of the wall, gabbroic rocks are more abundant than serpentinite. In the debris-covered area about the mine workings, several small chromite bands are still exposed. These strike approximately north 35 degrees east and dip 50 degrees southeast. Massive chromite boulders up to 10 inches in diameter are common on a nearby dump.

Table VI

Analyses of Mine Cove Area Chromite Concentrates

	No. 1 sample	No. 2 sample	No. 3 sample
Cr_2O_3	49·51	46·68	49·23
Al_2O_3	13·70	6·79	7.50
Fe*	15·10	11·12	13·38
MgO	14·27	19·06	18·66
CaO	—	—	—
SiO_2	1·58	9·40	6·51
Mn	Strong trace		
Na	" "		
Ti	Trace		
V	"		
Ni	"		
Cu	Faint trace		
Ca	" "		
Pt	None found		
Cell dimension (A_0)	8·280		
Cr/Fe	2·24	2·87	2·52

(1) Sample A509 from Niger Campbell pit. Analyst: H. B. Wiik (Chemical); Geol. Surv., Canada, (Spectrographic).
(2) Bluff Head Mine. Analyst: A. E. LaRochelle, Mines Branch, Ottawa.
(3) Bluff Head Mine. Analyst: E. Waller. (G. W. Maynard, 1897.)
* Only total iron determined.

In thin section the chromite-bearing rock appears as red chromite grains in an antigorite matrix cut by asbestos veins. Chemical analyses show that the highest grade of chromite has been obtained from this part of the Complex, but there is no surface indication that a mineable tonnage of ore remains.

There are three trenches in a small zone of disseminated chromite 1,700 feet east-northeast of the Bluff Head deposit. A small pile of chromite ore lies by one trench, with pieces of massive chromite up to 4 inches in diameter. About $1\frac{1}{2}$ miles north of the Bluff Head deposit are two small chromite concentrations that were investigated by Niger Campbell. The southern one is marked by two small pits and several caved trenches sunk in deformed serpentinite that contains bands of olivine clinopyroxenite and gabbro. The only chromite observed is in a small pile near the pits, where fragments of massive chromite range up to 3 inches in diameter. The northern concentration occurs in a serpentinite zone less than 10 feet wide between two gabbro bands. It is composed of small bands 2 to 3 inches wide but has no great extent.

ORIGIN OF THE CHROMITE DEPOSITS

The following field facts must be taken into account by any theory of origin for the chromite deposits.

(1) The mode of occurrence of chromite does not differ from that of the rock-forming minerals, forsterite and enstatite, except for an apparent spatial localization of chromite concentrations. The chromite occurs in layers, similar to those formed by olivine and pyroxene. Chromium-bearing spinels are an accessory component of all the ultrabasic rocks of the Complex, and crystallized contemporaneously with the silicate minerals. The spinels enclose olivine, but are themselves enclosed by enstatite grains. An explanation of the mode of origin of the chromite concentrations must be based on the fact that the chromite is a primary mineral, formed during crystallization of the igneous rocks.

(2) Chromite concentrations are restricted to the upper parts of the ultrabasic zones, and the important deposits are less than $\frac{1}{2}$ to $\frac{3}{4}$ mile stratigraphically below the main gabbro contacts. In the Lewis Hills and Mine Cove areas, the stratigraphic relationships are not quite as clear, as the gabbro phase is irregularly distributed throughout the pluton. The spatial relationship of the chromite to the gabbro is, however, maintained.

A columnar section through a typical Bay of Islands pluton (*see* Figure 2) may be subdivided according to three chemical environments: (1) the main ultrabasic zone characterized by the minerals enstatite and forsterite, an environment of high magnesia and low iron and alumina content; (2) the main gabbro zone characterized by calcic plagioclase and clinopyroxene, an environment of high lime and alumina content and comparatively low magnesia content; and (3) intermediate between these, the zone of maximum chromite concentrations. Small amounts of calcic plagioclase and diopside form discrete bands in this zone and increase in abundance near the gabbro contact. This is also an environment of generally high magnesia content, but with lime and alumina concentrated along certain bands. It appears that in the Bay of Islands Complex, this distinctive chemical environment in which small amounts of alumina occur with the chromium and magnesia is most favourable for the formation of chromite concentrations.

Studies of the systems anorthite-forsterite-silica (Andersen 1915) and anorthite-forsterite-diopside (Osborn and Tait, 1952) have added to our knowledge of the formation of spinel (*see* Figure 18). They show that, at temperatures above 1,320 degrees as determined in a dry melt, liquids of a composition intermediate between anorthite and forsterite will, on cooling, form spinel as a transient phase. If equilibrium conditions are maintained, the spinel phase disappears on further cooling. Rapid cooling, or the removal of the liquid phase by reactions or movements during intrusion, may prevent the resorption of spinel at lower temperatures.

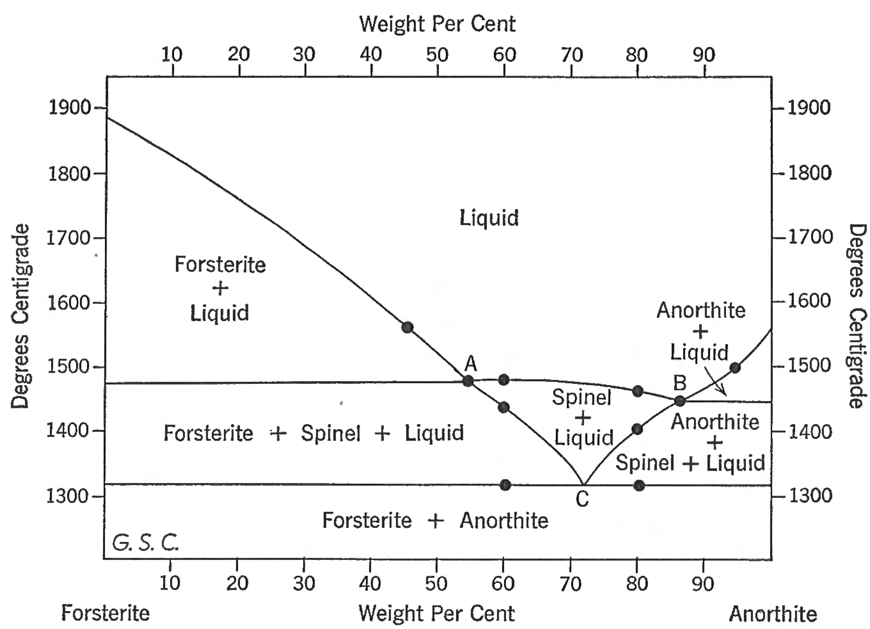


Figure 18. Equilibrium diagram of the system forsterite-anorthite (Osborn and Tait, 1952).

Bowen (1928) implied that chromium would readily enter the spinel structure to form chromite and thus explain the small amounts of accessory chromite common in ultrabasic rocks. This has been confirmed by the work of Wilde and Rees (1943) on the system MgO-Al_2O_3-Cr_2O_3. They show that there is complete solid solution between $MgO.Cr_2O_3$ and $MgO.Al_2O_3$. It is, therefore, reasonable to expect that any chromium present in the magma will enter the spinel structure to form chromite. In fact, spinel forms one of the major 'traps' for chromium during the cooling history of the ultrabasic rocks. As chromium will not enter the olivine or enstatite structures in appreciable amounts, the crystallization of spinel is one of the first events in the cooling history of the magma that allows chromium to enter the crystal phase.

Localized spinel-forming reactions that take place in the upper parts of the ultrabasic zones during crystallization of the magma, are apparently quantitatively capable of removing chromium from the surrounding crystal mush by local diffusion through the interstitial liquid, and hence could produce the chromite concentration of the Bay of Islands Complex. The chromium content of the normal ultrabasic rocks of the Complex is about 0.40 per cent Cr_2O_3. Enormous amounts of chromium, in an unconcentrated form, are thus locked up in the ultrabasic rocks. If the chromium in a cube of ultrabasic rock 500 feet on an edge was concentrated into a cube 110 feet on an edge, an orebody of 200,000 tons of chromite, containing 35 per cent Cr_2O_3 could be formed. These figures are illustrative only, as no chromite deposit this size is known in the Bay of Islands Complex. In view of the great size of the ultrabasic zones, however, the formation of chromite concentrations is evidently a minor and all too infrequent event in the history of the plutons as a whole. They could form without extreme prior concentrations of chromium in the liquid phase of the magma, that is from fluids not greatly enriched in chromium. And, according to this hypothesis, the lower horizons are barren because there was insufficient alumina to effect the segregation of spinel.

The deposits are late magmatic segregations, but whether the chromite crystallized during the pre-intrusion differentiation cycle of the magma, or during the time of tectonic emplacement cannot be satisfactorily determined. The position of the chromite concentrations in the upper parts of the ultrabasic zone in opposition to the usual specific gravity relationships, and the evidence for late crystallization of the chromite, suggest that the chromite may have formed during the later cooling history of the pluton, after it had attained its present structural position. Reactions between the gabbroic and ultrabasic magmas during and after intrusion, but prior to consolidation, apparently caused the formation of spinel and the segregation of chromite. The chromite was generated in situ because of the distinctive chemical environment prevailing in the upper parts of the intrusion, and the disequilibrium caused by the tectonic emplacement of a partly differentiated sill.

APPLICATION OF THEORY TO CHROMITE PROSPECTING

Hypotheses based upon an early magmatic origin for chromite generally assume that chromite concentrations should occur near the base of ultrabasic layers. As alumina is more abundant in the upper parts of such layers, nearer their gabbroic caps, a better place to look for chromite is in the upper parts of ultrabasic layers. This interpretation is applicable to primary layered bodies, but not directly to deformed serpentinite bodies whose primary petrochemical form has been destroyed by later deformation.

In the Bay of Islands area, prospecting for chromite should be restricted to that part of the ultrabasic body within a mile or less of the main gabbro contacts. The actual distance depends on the dip of the contact or of the layering. In terms of 'stratigraphic' thickness, the favourable zone is $\frac{1}{2}$ to $\frac{3}{4}$ mile below the gabbro. In the Lewis Hills pluton and the Mine Cove fault block, practically the whole area underlain by ultrabasic rocks is favourable because none of the ultrabasic rocks is far removed from gabbro or feldspathic dunite, in these areas. Chromite concentrations, as may be expected, are also numerous in these areas.

The areal relationship of chromite deposits to gabbro contacts is known in other areas, and has been used to advantage in prospecting. From a study of the Camagüey district of Cuba, Flint, de Albear, and Guild (1948) draw the following conclusions: "The most significant results of the detailed mapping of the Camagüey district has been to show that, although the economic deposits in the district occur in ultramafic rocks, there is a general relationship between the ore deposits and the feldspathic rocks and overlying volcanic rocks. A statistical study of their distribution reveals that 94 per cent of the known ore of the district, including both mined ore and reserves, lies within half a mile of the belts of feldspathic rocks . . . . The conclusion seems inescapable that the distribution of the ore deposits is in some way related to the localization of the feldspathic rocks, which seem to be near the top of a more or less stratiform complex." In the Bird River sill of Manitoba, according to Bateman (1943), chromite layers occur near the top of the peridotite zone, within 170 feet of the overlying gabbro. A similar stratigraphic relationship is mentioned by Thayer (1946) who states that in certain intrusions of eastern Oregon the larger deposits are relatively near the gabbro. In yet other areas, where gabbro forms the major part of the layered intrusions (Bushveld, Stillwater), plagioclase is common in the chromite zones, proving a higher lime and alumina concentration along these zones than in the normal ultrabasic rocks.

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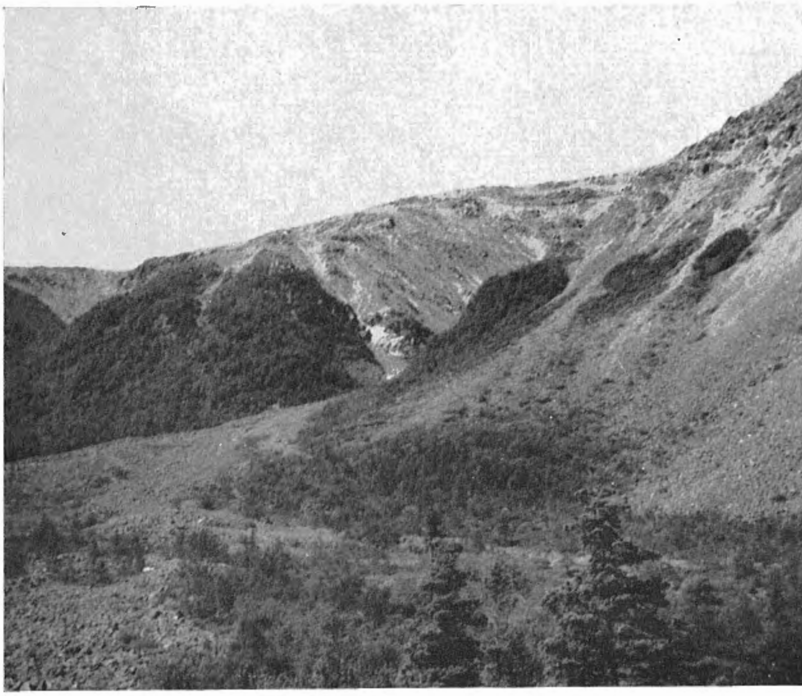
PLATES

PLATE I

106427

Table Mountain from the Bonne Bay road, showing the Long Range peneplane. The barren peneplane is underlain by ultrabasic rocks and the wooded hills are underlain by sedimentary and metamorphic rocks of the Humber Arm group. (Page 4.)

PLATE II



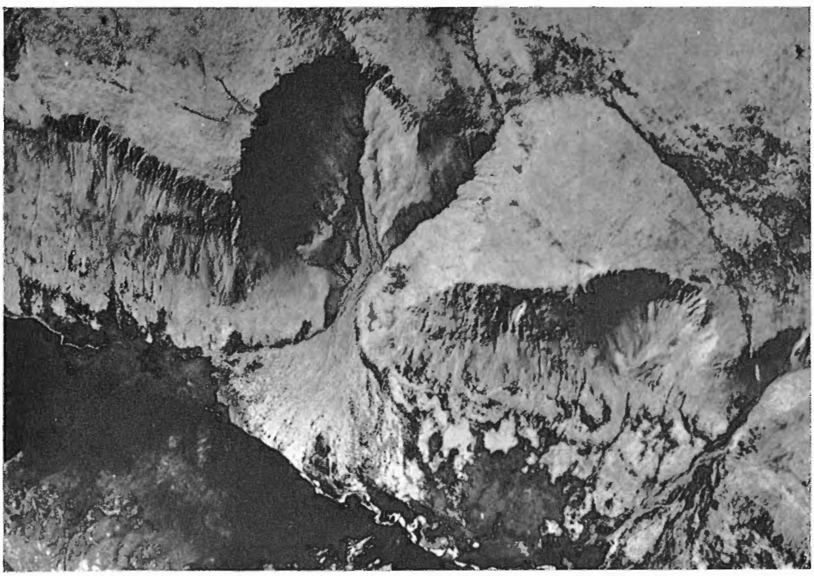
C.H.S. 5-3-51

A. Basal contact of North Arm Mountain pluton exposed along the north shore of North Arm. The wooded areas are underlain by sedimentary and contact metamorphic rocks of the Humber Arm group, and the barren areas at the top of the scarp by ultrabasic rocks. Alluvial fan in foreground. (Page 5.)

106428-21

B. Valley of Stone Brook, North Arm Mountain. Local relief is as much as 1,800 feet. The depth of this post-glacial valley demonstrates the rapid erosion of ultrabasic rocks in the area. Scale, 1 inch equals about $\frac{2}{3}$ mile. (Page 5.)

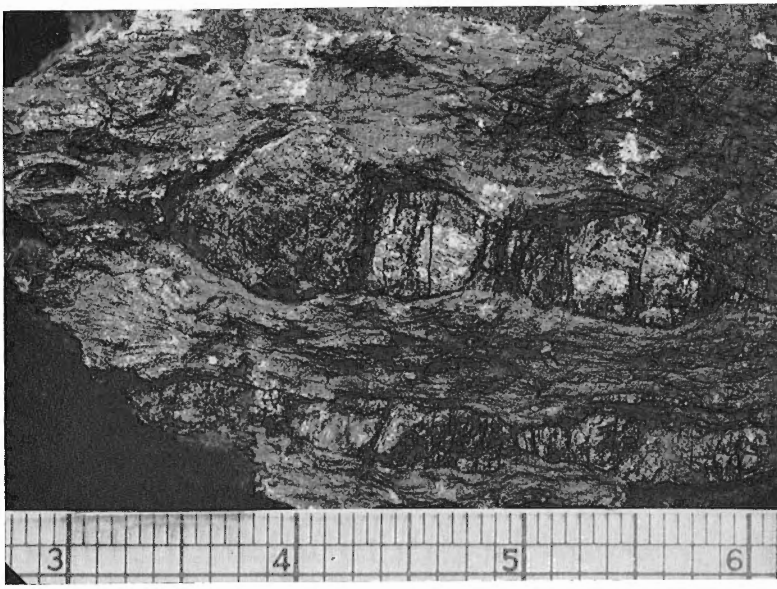
PLATE III



106428

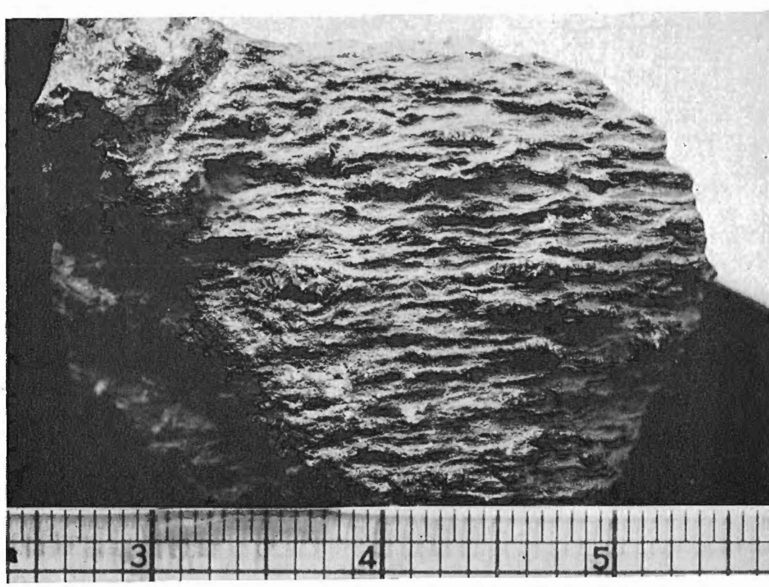
Northern fault-line scarp of Table Mountain showing notches cut by headward-eroding streams, and the development of cirque-like forms by active mass-wasting and breakdown of the interfluves between tributaries. Scale, 1 inch equals $\frac{3}{4}$ mile. (Page 6.)

PLATE IV



109910

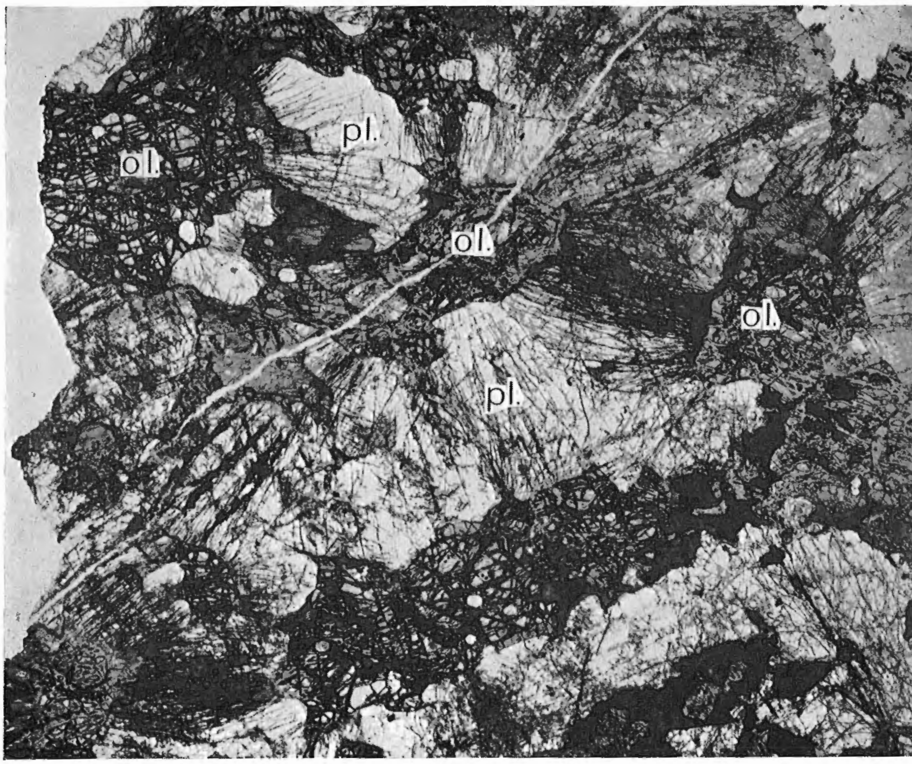
A. Peridotite augen-gneiss composed of enstatite 'eyes' in a groundmass of serpentine. Basal contact of North Arm Mountain pluton. (Page 31.)



109911

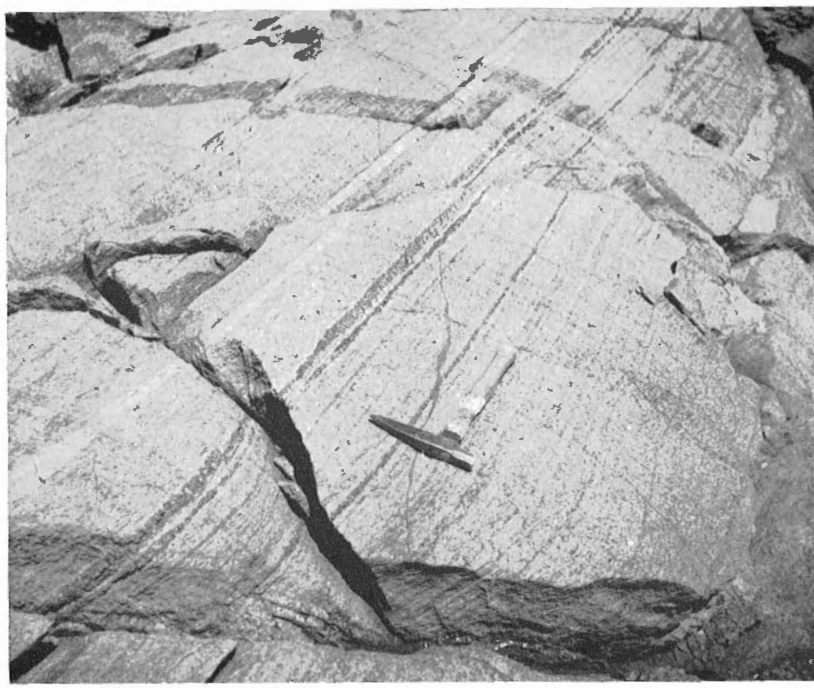
B. Peridotite augen-gneiss composed of enstatite 'eyes' in a groundmass of serpentine. Basal contact of Table Mountain pluton. (Page 31.)

PLATE V



Expansion fabric in troctolite. Radiating fractures extend from partly serpentinized olivine grains (ol.) into surrounding plagioclase (pl.). X 12.5 mag. (Page 36.)

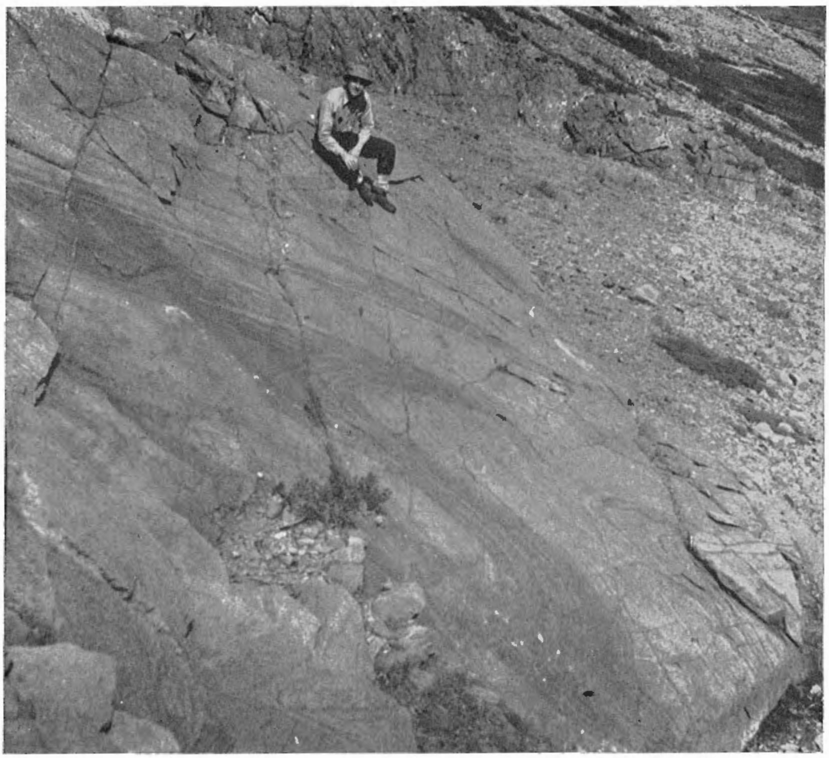
Plate VI



C.H.S. 1-7, 51

A. Interbanded enstatolite (dark), peridotite (mottled), and dunite (light). Note dunite borders along enstatolite bands. Table Mountain pluton, near the Narrows. (Page 39.)

Plate VI



C.H.S. 3-5, 51

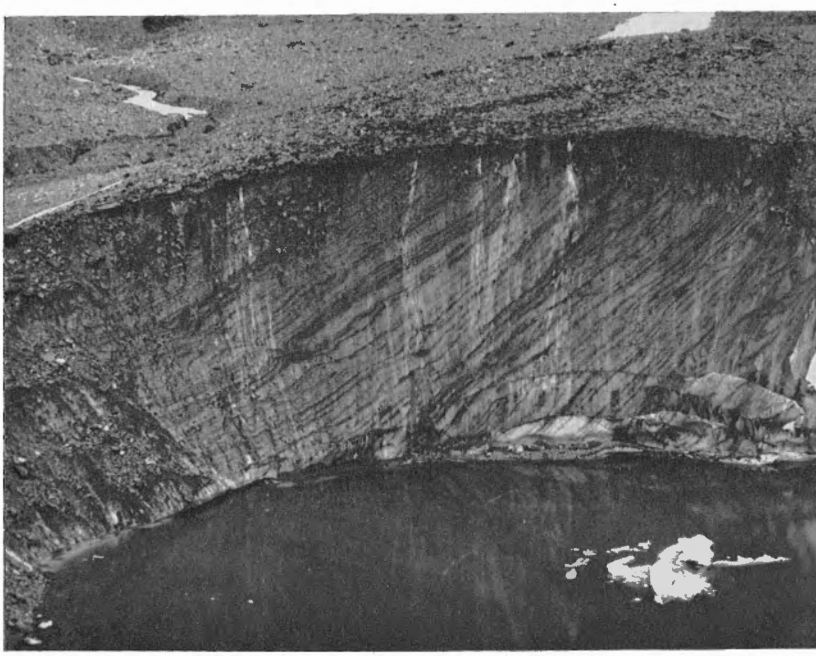
B. Folded enstatolite bands with thickened crests. Table Mountain pluton near the Gulch. (Page 39.)

PLATE VII

C.H.S. 1-9, 51

A. Detail of banded gabbro. Note splitting and pinching of anorthosite bands (white). Boulder in Trout River Pond. (Page 39.)

PLATE VII



J.A.E. 2-7, 52

B. Flow structures in ice. A near-vertical face in kettle hole, Wenkchemna glacier near Lake Louise, Alta. Cliff is 40 to 50 feet high. (Page 43.) *Photo and interpretation by J. A. Elson.*

PLATE VIII

C.H.S. 3-1-1951

A. Fault contact between serpentinite (right) and shales (left) of the Humber Arm group. Contact is marked by a 4- to 5-foot layer of calcium-bearing minerals of hydrothermal origin. Near headwaters of Winter House Brook, Bonne Bay. (Page 49.)

PLATE VIII

C.H.S. 4-2, 51

B. View looking northwest along Second Trout River Pond, from the southeast end of the lake. Note the large gabbro blocks spalling off the mountainside, and other blocks in the process of moving down-slope to the lake. The thrust sheet outlier described by Buddington and Hess (1937) may have originated in a similar manner. (Page 62.) *Photo by J. Davy.*

PLATE IX

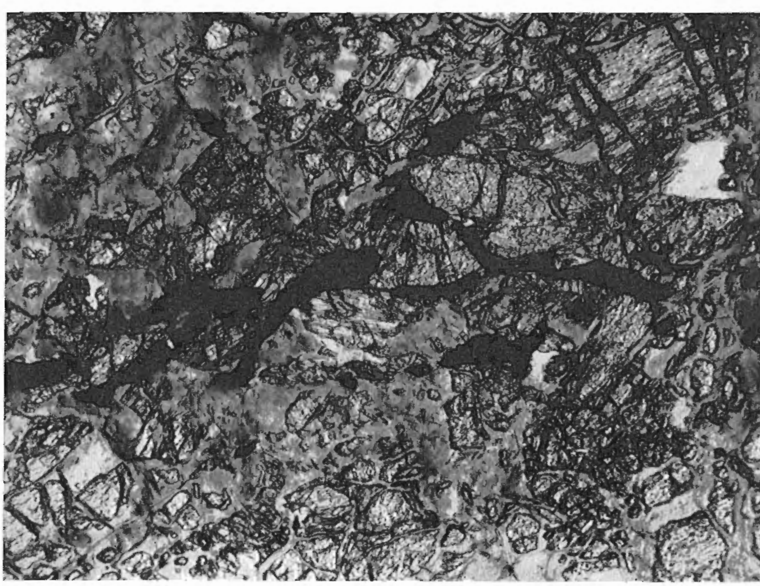
106587-B-144

The narrows between First and Second Trout River Ponds, looking north. Note large slump on side of Table Mountain etched out by streams, and location of block X under discussion (left). (Pages 5, 62.)

PLATE X

109912

A. Chromite (black) enclosing serpentinized olivine. Note irregular serrate edges of chromite grains. X 10 mag. (Pages 30, 99.)



109913

B. Vermicular late chromite (black) interstitial to silicate minerals. X 10 mag. (Pages 30, 99.)

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