THE ULTRAMAFIC COMPLEX AND RELATED ROCKS
OF
DUKE ISLAND, SOUTHEASTERN ALASKA

Thesis by

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ABSTRACT

Duke Island, 59 square miles in area, is at the southern end of southeastern Alaska. Sedimentary and volcanic rocks, possibly Mesozoic in age, are metamorphosed and intruded by gabbroic, ultramafic and granitic plutons, in that order. The granitic rocks may be of Cretaceous age.

Primary gabbroic rocks are dominantly two-pyroxene gabbro and norite. Their plagioclase is An_{50}-An_{70}.

Ultramafic rocks crop out as two main areas and more than a dozen minor ones. The rocks in the main areas probably are continuous at depth forming the Duke Island ultramafic complex. Constituent minerals are olivine, clinopyroxene, and hornblende; orthopyroxene and plagioclase characteristically are absent. Rock units are classified as dunite, peridotite, olivine pyroxenite, and hornblende pyroxenite. Hornblende pyroxenite contains 10-20 per cent magnetite and typically occurs as a border zone. The olivine-bearing units have remarkable layering which developed by gravitational settling of crystals from a body of circulating magma. Most of the olivine pyroxenite is cut by an intrusion represented at the present surface by dunite and peridotite.

Hornblende-anorthite (An_{95}) pegmatite, an ultramafic derivative, occurs in an aureolé around the complex. In the aureole, pyroxene gabbro is altered to hornblende gabbro with plagioclase intermediate between those of pegmatite and primary gabbro.

The relationship of ultramafic and primary gabbroic rocks indicates that they formed from ultramafic magma and normal gabbroic magma respectively.

Mechanisms of crystallization differentiation, multiple intrusions, solid intrusion, and vapor transfer are examined as possible explanations of the distribution of rock types within the ultramafic complex. No one is sufficient, but all have applicability. The border zone is accounted for by transfer of water, silica, lime, and iron from the ultramafic magma to olivine-bearing rocks initially solidified from the magma body onto its walls. The required reactions are demonstrable in other parts of the complex, and the process is related to the development of the surrounding aureole. Evidence is given of late magmatic recrystallization in the complex and of local replacement of olivine pyroxenite by dunite. Disequilibrium, largely arising from multiple intrusion, and transfer of materials by an aqueous-rich vapor phase are probable causes.

A sequence of events is summarized.

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PART I. INTRODUCTION

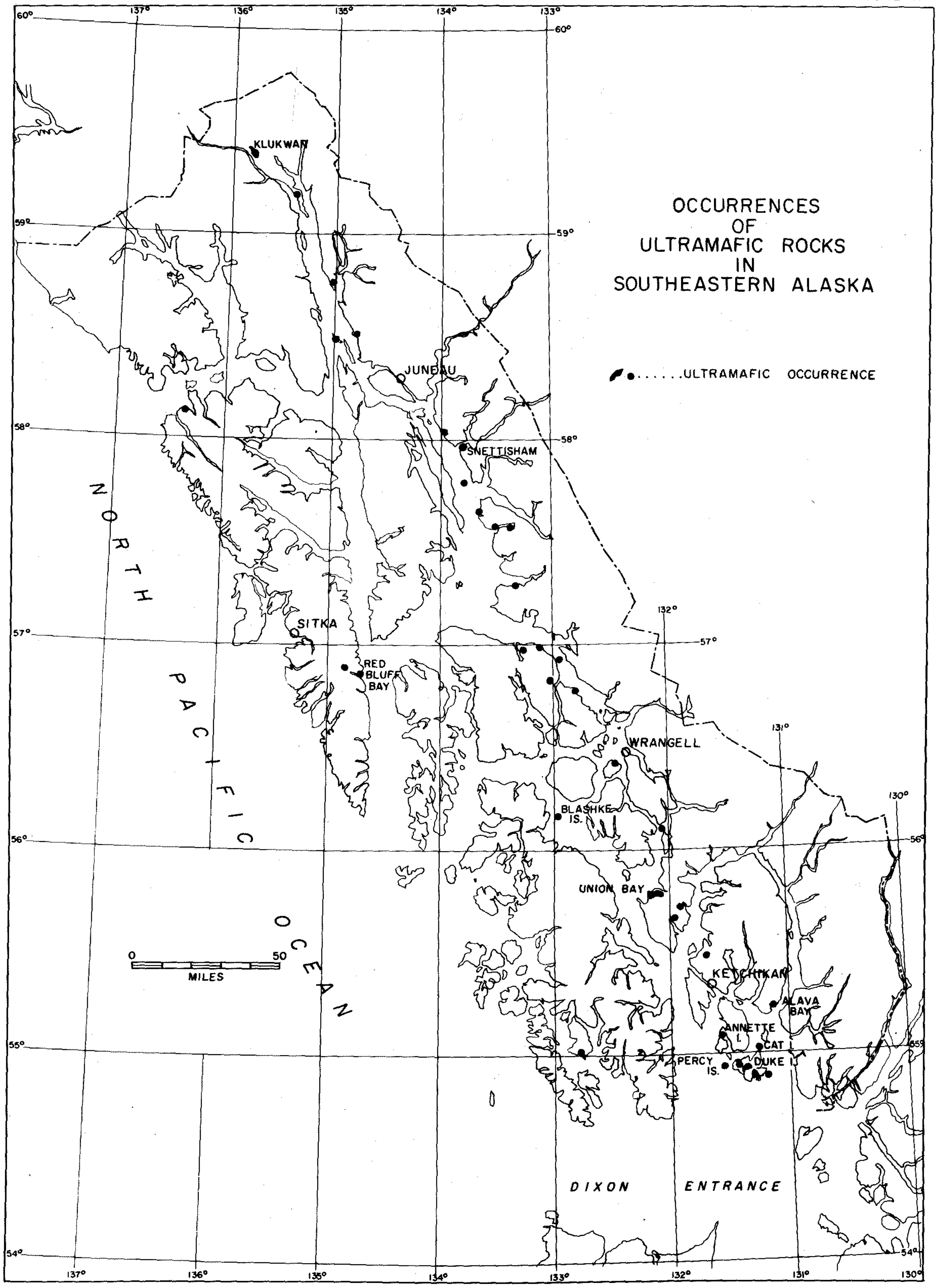
Ultramafic rocks in the visible parts of the crust of the earth characteristically occur in relatively small bodies in narrow belts of moderate to intense orogeny. One of these belts extends the length of southeastern Alaska, a distance over 300 miles (fig. 1), and apparently continues in an arc to Kenai Pennisula in southern Alaska (Hess, 1954). In addition to their common tectonic environment, these ultramafic rocks have many similar features in mineralogy, petrology, and internal structure, and therefore they probably have a similar origin. At the southern end of the belt, at Duke Island, an ultramafic complex contains remarkable layered structures, adding even more interest to these distinctive rocks.

LOCATION AND ACCESS

Duke Island is centered at approximately 54°55'N. and 131°20'W. and is situated on the north edge of Dixon Entrance (fig. 1). The main island covers about 59 square miles, and close by and included in the descriptions herein are several smaller islands with a total area slightly under 3 square miles. The larger of these satellitic islands are Dog Island, Kelp Island, and East Island.

The area is readily accessible from Ketchikan, 40 miles to the north, by either boat or seaplane. A landing

FIGURE I.


OCCURRENCES
OF
ULTRAMAFIC ROCKS
IN
SOUTHEASTERN ALASKA
......ULTRAMAFIC OCCURRENCE
KLUKWAN
JUNEAU
SNETTISHAM
SITKA
RED BLUFF BAY
WRANGELL
BLASHKE IS.
UNION BAY
KETCHIKAN
ALAVA BAY
ANNETTE I.
CAT I.
DUKE I.
PERCY IS.
NORTH PACIFIC OCEAN
DIXON ENTRANCE
0
50
MILES
60°
59°
58°
57°
56°
55°
54°
137°
136°
135°
134°
133°
132°
131°
130°

strip for large planes and a U. S. Coast Guard station are located on Annette Island, 8 miles northwest of Duke Island. Duke Island has no permanent habitation.

TOPOGRAPHY

The topography of Duke Island is subdued compared to that of most of southeastern Alaska. One prominent peak, Mount Lazaro, with an elevation of 1785 feet, rises in the southwest corner of the island. Elsewhere, only one point is above 600 feet, and local relief does not exceed 300 feet. The hills are streamlined by Pleistocene ice that moved approximately S.30°W. The hummocky topography of glacial terrain is especially well developed in the central and south-central parts of the island where ultramafic rocks crop out. Numerous small fresh-water lakes dot the surface of the island, and two long coves, Hall Cove and Morse Cove, carry salt water well into the interior. The shore line is irregular and rocky, swept clean by storm waves and the strong currents of 15- to 20-foot tides. Broad wave-cut benches occur in the poorly resistant metamorphic rocks along the southwest shore.

VEGETATION

The principal trees on the island are hemlock, cedar, shorepine and alder. The forest is appreciably less dense than is typical of the region, and its growth shows close control by drainage and the type and availability of soil. Everywhere along the well-drained shoreline is a fringe of dense growth about a hundred yards wide. Inland the steeper slopes have relatively heavy timber, particularly southern slopes as they were lee of ice movement and are covered by glacial debris. Areas underlain by metamorphic and granitic rocks are generally low and wet, and stands of trees are patchy and stunted. The ultramafic rocks provide only small quantities of soil of poor grade and, with the exception of the hornblende-bearing varieties, are largely barren. Only the gabbroic rocks support uniformly heavy forest.

PREVIOUS WORK

All recorded previous geologic work on Duke Island has been confined to the examination of shoreline exposures. The first known reference to the ultramafic rocks was made by Chapin (Buddington and Chapin, 1929, p. 189), who reported the presence of peridotite and pyroxenite. In 1934, A. H. Koschmann spent 15 days mapping the shoreline, and his unpublished map and report are available on open file at the U. S. Geological Survey offices. He later described and discussed some of the hornblende-rich rocks (Koschmann, 1935).

Locations and brief descriptions of comparable ultramafic rocks in southeastern Alaska are given by Buddington and Chapin (1929, pp. 188-198). A more recent unpublished summary covering Alaska and British Columbia was compiled by Walton (1951a). Papers giving detailed descriptions of individual bodies of ultramafic rock are: Guild and Balsely (1942); Kennedy and Walton (1946); Walton (1951a) and (1951b); Robertson (1955); Stebbins (1957); Ruckmick (1957); Ruckmick and Noble (1959).

PRESENT WORK

The field work on which the present report is based was done during the summers of 1955, 1956, and 1957. Operations were carried out from temporary camps set up in three parts of the island, Pond Bay, Hall Cove, and Judd Harbor. The whole island was first mapped on a reconnaissance basis, and then the ultramafic areas were remapped in more detail. Most of the field mapping was done on vertical aerial photographs. A preliminary copy of the U. S. Geological Survey map for the area (Prince Rupert sheet) was enlarged, and some of the details of shore-line features have been modified by the author from the air photographs. The geology of the whole island, plotted on this base, appears in plate 1 at a scale of 1" = 2640'. The two main ultramafic areas that were mapped in more detail appear in plates 2 and 3 at 1" = 1320' Two small parts of the ultramafic rocks with considerable complication, originally mapped at 1" = 20' and 1" = 10' using tape and compass control, and are shown in plates 4 and 5 respectively.

To facilitate making reference to specific localities, the writer has added two topographic place names, Knob Hill and North Hill, to the map.

The laboratory work has largely been a petrographic study. Thin sections and polished sections were prepared by R. von Huene at the California Institute of Technology. Indices of refraction of minerals were determined on sized materials by oil immersion in sodium light. The indices of pyroxenes and olivine are reliable to about $\pm$ 0.002, and those of plagioclase to about $\pm$0.003. In the metamorphic rocks, plagioclase composition is determined only by the maximum $\lambda+$ extinction angle of albite twins observed in thin section. The universal stage was used to measure the optic angles of pyroxene grains in thin section. Only grains were measured in which direct rotation from one axis to the other was possible, and all tabulated values but one are averages of two or more measurements. Corrections of angles were made for differences between the indices of the pyroxenes and the stage hemispheres. Optic angles are reproducible to within $\pm$ 2 degrees. Modal analyses were done by the point counting method using a mechanical stage. Each mode is based on a minimum of 1500 counts.

Materials for chemical work were crushed in a large diamond mortar, and all sizing was done with bolting cloth screens. The mineral separates were purified with heavy liquids and a Franz magnetic separator. Analyses for major elements were done by W. H. Herdsman, Glasgow, Scotland.

A preliminary study has been made of the trace element **distribution** in the ultramafic and gabbroic rocks. The analyses were made by the author in the spectrographic laboratory of the California Institute of Technology under the direction of A. A. Chodos and E. Godijn. Each sample was mixed with 4 times its own weight of spectrographically pure graphite and exposed using the following equipment and methods.

Spectrograph: Jarrel-Ash 3.4 m grating instrument, Wadsworth mount, dispersion, 5.2 A/mm in the first order.

Excitation: 19-ampere short-circuit D.C. arc from a Jarrel-Ash Varisource. Sample as the anode. Analytical gap, 4 mm magnified 5X and focused on the slit. Central 2 mm used with a slit width of 25 microns; 25-mg samples were burned to completion (90-120 sec.). Total energy method with no internal standardization.

Electrodes: High purity one-quarter inch graphite rod as the anode. U.C.C. #3417. Pointed one-eight inch cathode.

Wave-length Range: 2300-4800 A in the first order.

Plates: Eastman Kodak III-O.

Processing: 4 minutes in DK-50 developer at 20°C, 20 sec. short stop, 10 min. in acid fix, 20 min. wash.

Plate Calibration: Selected iron lines after the method of Dieke and Crosswhite (1943). Each plate is calibrated.

Densitometer: Applied Research Laboratories model #2250.

Spectral line intensities were converted to concentrations of elements using the standard curves that are used regularly in the spectrographic laboratory.

The isotopic composition of some quartz samples was determined by the author under the supervision of S. Epstein. The experimental method involves the reduction of the quartz with graphite at high temperatures according to the equation:

$$SiO_2 + 3C \longrightarrow SiC + 2CO$$

The CO gas is then converted to CO_2 in the presence of a catalyst at appropriate temperatures.

$$2CO \longrightarrow C + CO_2$$

Measurement of the isotopic composition of the oxygen in the mass spectrometer was done on the CO_2 gas. The technique and apparatus were essentially the same as that used by Clayton and Epstein (1958). Results are reported in terms of δ:

$$\delta = \left[\frac{R_{sample}}{R_{standard}} - 1\right] 1000$$

where R = ratio O^{18}/O^{16}. The standard used here is mean sea water, that used by Clayton and Epstein.

In the text to follow, the general geology of the island is first briefly described. The gabbroic and ultramafic rocks are then described in detail with a minimum of interpretation. The origin of these rocks and the causes of inferred processes are then discussed, and finally, a sequence of events that is considered to best account for the observed features is given.

PART II. GENERAL GEOLOGY

METAMORPHIC ROCKS

The oldest rocks exposed on Duke Island form a complex assemblage of metamorphosed sedimentary and volcanic rocks. They include chloritic, micaceous, and amphibolitic schists and gneisses, apparently derived largely from clastic sediments, and volcanic rocks of intermediate composition, primarily dacitic lavas and agglomerates. Buddington and Chapin (1929) have grouped this assemblage with a broad area of comparable lithologic types adjacent to the Coast Range batholith, which they call the Wrangell-Revillagigedo belt of metamorphic rocks. They state (p. 74) that this belt is made up mainly of Carboniferous and Triassic formations, although beds as young as Cretaceous and as old as Ordovician may be included. The most specific reference to Duke Island is made by Chapin (Buddington and Chapin, 1929, p. 134):

> "Greenstone and slate made up of an interbedded series of altered tuffs, flows, and black slate, with some intrusive rocks, occur on Gravina, Duke, Annette, and Revillagigedo Islands, in the Ketchikan district. The greenstone and slate overlie the Upper Triassic rocks with apparent conformity. On the evidence of a few fossils found in the intercalated sediments and on the grounds of structure and analogy with rocks of known age, these rocks are regarded as Upper Triassic or Jurassic."

The conformable relationship was apparently observed on Gravina Island, where Chapin did his most detailed work, and not on Duke Island (cf. Chapin, 1919, p. 90).

The exposure of the metamorphic rocks on Duke Island is poor, except along the shore line, and only a rapid reconnaissance has been made over them. The rocks are structurally and petrologically complex, and the different types are difficult to distinguish. Consequently, no attempt has been made to separate them into units on the final map.

The main occurrences of metamorphic rocks are along the southwest shore between Point White and Hall Cove, and on the north shore in the vicinity of Niquette Harbor and Pond Bay. As shown on the map, minor bodies occur elsewhere along the shore. Evidently the metamorphic rocks fringe the igneous core of the island, but this may be in part the result of few outcrops inland or a failure to distinguish the metamorphic amphibolites from the hornblendic gabbros.

The original sediments were apparently largely fine-grained clastics, as they are commonly associated with conglomerates, and their mineralogy indicates that the chemical composition probably corresponds to that of graywacke. Foliation is generally well developed, and intense folding occurs locally. In some places two or more sets of lineations suggest refolding, a definite possibility as the metamorphic rocks are cut by major intrusions of several ages. Much of the bedding is either obliterated or parallels the foliation, but marked local divergences between the directions of bedding and foliation have been noted.

The lavas are more massive than the metasedimentary rocks and appear to have acted as competent buttresses during the shearing that accompanied metamorphism. They are generally fine-grained, and primary igneous textures are still evident locally. Most of the lavas are equigranular, but some are porphyritic. In places near the granitic rocks, porphyroblasts of albitic plagioclase are extensively developed and stand out distinctively on the light brown to medium gray-green weathered surface. Plagioclase is the dominant mineral and is partially altered to clinozoisite or epidote and albite. Biotite, and hornblende partially altered tochlorite, are the main mafic minerals but are generally present in small percentages. Quartz is rare, and potash feldspar is absent. The agglomerates are everywhere closely associated with the lavas and weather to much the same color. Both fragments and matrix are near to the flow rocks in composition. The fragments are a few inches in diameter and are angular to subangular in outline.

The metamorphic rocks are intruded by major masses of both granitic and gabbroic rocks, but they have not been observed in contact with a large body of ultramafic rock. It has not been possible to study the changes in any one unit as it approaches the contacts with the igneous rocks, but a general summary relating mineral assemblages of all metamorphic rocks to distance from the igneous contacts is given in table 1. The distances are taken from the map and possibly the contact is closer in the third dimension. The specimens collected within 1000 feet of a contact have been separated into two groups, depending on whether the igneous rock is granitic or gabbroic. The following features are noted in the table.

(1) The mineral assemblages do not change significantly as the granitic rocks are approached.

(2) As gabbros are approached, the following changes take place at about 1300 feet from the contact.

   (a) The number of mineral phases decrease from 6 or 7 to only 4 or 5. This probably indicates a closer approach to equilibrium with fewer relict minerals, and the more complex chemical composition of the minerals of higher metamorphic rank.

   (b) Sphene, calcite, chlorite, muscovite, and generally, epidote disappear. On the basis of its mineralogy specimen I-9-2 (entry 29) seems out of place in the table, but the reason is not known.

   (c) Magnetite and apatite become common phases.

TABLE I. VARIATIONS IN THE MINERAL ASSEMBLAGES OF THE METAMORPHIC ROCKS WITH DISTANCE FROM THE CONTACTS OF IGNEOUS BODIES

Entry	Specimen	Environment		Mineralogy														
		Igneous Type	Distance From Contact Feet	Sphene	Calcite	Chlorite	Muscovite	Epidote	Quartz	Plagioclase	Biotite	Magnetite	Hornblende	Garnet	Pyroxene	Cordierite	Sulphides	Apatite
1	N-23-9	Gabbro	3500		X	X	?	X	X	X	(X)		A					
2	I-1-1	Granite	3000	X	X	(X)	X		X	5							X	
3	I-1-2	Granite	3000	X		X	X	X	X	5			A	X				
4	I-20-3	Granite	3000	X	X	X	X	X	X	X	X		A					
5	S-15-3	Granite	2500		X	X	X	X	X	5								
6	S-15-5	Granite	2300		X	X	X	X		6	X							
7	I-1-4	Granite	2200			X		X		7	X							
8	N-23-7	Granite	2000		X		X	X		4	X	X	g					
9	I-4-1	Gabbro	2000	X	X	X		X	X	X							X	
10	I-4-2	Gabbro	2000			X		X		37	(X)	X	A				X	X
11	I-4-7	Gabbro	1400	X		X	(X)	X	X	8								
12	N-42-14	Gabbro	1300					X		47	X		g					
13	N-42-13	Gabbro	1000	X				X	X	53			g					
14	N-23-4	Granite	800			X	X	X	X	X	X							
15	S-15-2	Granite	200				(X)	X	X	23	X	X	g					X
16	N-23-3	Granite	100		X	X	X	X	X	X			g					
17	N-22-1	Granite	0				X	X	X	X			g					
18	S-17-4	Gabbro	700-						X	40	X	X		X				
19	S-17-5	Gabbro	700-						(X)	46	X	X		X			X	
20	S-17-6	Gabbro	700-							60		X	b					X
21	N-42-12	Gabbro	600						X	53	(X)	X	bg					
22	I-10-2	Gabbro	500?						X	55	X		bg					X
23	I-10-6	Gabbro	500						X	52	X		(X)					X
24	I-5-5	Gabbro	400							46	X		gb					X
25	N-46-9	Gabbro	300							62		X	b					
26	N-42-8	Gabbro	250					X	X	46	X	X		X		X	X	
27	I-5-4	Gabbro	200						X	48	X	X	b					X
28	H-6-1	Gabbro	200							68		X	g	X				X
29	I-9-2	Gabbro	200			X	X	X	X	27	X	X	g				X	X
30	I-10-1	Gabbro	200							40	X	X	gb				X	X
31	R-9-6	Gabbro	100							46	X	X						
32	N-26-4	Gabbro	100						X	40	X	(X)	b					X
33	H-6-2	Gabbro	100						X	62	X		g				X	
34	N-41-8	Gabbro	100						X	47	X	X		X	X			
35	N-46-7	Gabbro	50						X	42	X		b	X				
36	N-45-6	Gabbro	50						X	55	X				X		X	X

Notes: (1) The anorthite content of plagioclase is given where determined.

(2) X indicates the presence of a phase. (X) indicates the presence of a phase in small quantities.

(3) For hornblende: "A" indicates actinolitic character; "g" indicates green color; and "b" indicates brown color.

Biotite also seems to become more common, although it is present in several specimens from beyond the 1300 foot cutoff.

(d) The anorthite content of the plagioclase shows a marked increase. This undoubtedly accompanies the disappearance of epidote.

(3) Within 1000 feet of the gabbro:

(a) Brown or brownish green hornblende is relatively common. Biotite may be more reddish near the gabbro, but this does not seem to be as general as the color change in hornblende.

(b) Garnet appears in some rocks of appropriate composition. Specimen I-1-2 (entry 3) contains a small amount of garnet, but the rock is clearly among those of lower metamorphic grade, and probably the garnet is manganiferous (Harker, 1939, p. 217). The garnets within 1000 feet of the gabbro are usually large red euhedral metacrysts and probably belong to the almandine-rich series.

(4) Two specimens within 100 feet of the gabbro contain pyroxene: one contains hypersthene, and the other augite ($n_y = 1.717$). Cordierite is present in one specimen (N-42-8, entry 26)

collected about 250 feet from gabbro.

The textures show a roughly parallel change. Away from the gabbro, the minerals commonly have poorly-developed, shreddy forms, and relict minerals are common. Near the gabbro, the rocks generally are well-crystallized schists and gneisses, with slightly coarser grain sizes, and the mineral grains are relatively free of inclusions.

Thus the grade of metamorphism characteristic of areas away from gabbro is either the same as that imposed by the granitic rocks, or is determined by the granitic rocks. The mineral assemblages are representative of the albite-epidote-amphibolite facies and the greenschist facies (Turner and Verhoogen, 1952). Some of the apparent variations in grade in these rocks are due to retrograde effects or, perhaps, repeated metamorphism. Near the gabbro, the mineral assemblages are largely indicative of the amphibolite facies, and the hypersthene-bearing rock may belong to the pyroxene granulite facies. These higher grades of metamorphism clearly are related to the gabbro.

GABBROIC ROCKS

Gabbro is the most common rock type on Duke Island. Two main species, pyroxene gabbro and hornblende gabbro, have been distinguished on the map. The beginning of a further subdivision of the pyroxene gabbro on the basis of its content of pyroxene and olivine has been made but to be completed would require extremely detailed petrographic study.

The gabbroic rocks are younger than the metamorphic rocks, definite intrusive contacts being displayed wherever their contact is exposed. The pyroxene gabbro is believed to be older than the ultramafic rocks, whereas the development of the hornblende gabbro probably was caused by the ultramafic and granitic rocks. Evidence for these relationships will be given in a later section.

ULTRAMAFIC ROCKS

In the vicinity of Duke Island occurs the greatest concentration of ultramafic rocks in southeastern Alaska. Within the area shown on plate 1 are two major areas of ultramafic rocks and more than a dozen minor ones. Four miles to the west, another large ultramafic complex is exposed on the Percy Islands. Ten miles to the northwest, a large dunite body crops out on Annette Island, just north of the airport. A smaller body of pyroxenite is exposed on Cat Island, about 2 miles north of the northeast corner of Duke Island.

The largest area of ultramafic rocks at Duke Island is about 5.4 square miles. It lies in the center of the island and is well exposed along Hall Cove. Hereafter, it will be referred to as the Hall Cove ultramafic area. It is shown on plate 2. The second largest area occurs at the south end of the island and is about 3.6 square miles. It is exposed in the vicinity of Judd Harbor and will be called the Judd Harbor ultramafic area. Plate 3 is a detailed map of this area. There is reason to believe that the two main areas are the outcrop of one body continuous at depth. This body is the Duke Island ultramafic complex.

A feature common to several of the ultramafic complexes in southeastern Alaska is a concentric zoning. In its extreme development, the zoning has a dunite core surrounded by successive rings of peridotite, olivine pyroxenite, pyroxenite, and gabbro. This structure was first described by Buddington (Buddington and Chapin, 1929, pp. 190-191) in the Blashke Island complex, where it has the most symmetrical development. Kennedy and Walton (1946) recognized zoning in ultramafic bodies at Kane Peak and Mount Burnett (Union Bay). More recently, Ruckmick (1957) has confirmed the existence of zoning at Union Bay, and Stebbins (1957) has demonstrated it in the Percy Islands complex. An additional aspect of the Blashke Island zoning, discovered by Kennedy and Walton (1946), and later corroborated in detail by Walton (1951), is a progressive increase in the Fe:Mg ratio of olivine and pyroxene from the central zone outward, a feature which Walton calls "cryptic zoning".

Concentric zoning is only crudely developed in the

Duke Island ultramafic complex. It is best shown in the Hall Cove ultramafic area, where peridotitic rock is more or less surrounded by olivine pyroxenite which, in turn, has an almost complete rim of hornblende pyroxenite. In the Judd Harbor ultramafic area, the same rock types predominate, but as shown in plate 3, they are disposed side by side rather than in a concentric fashion. Imperfect as the zoning may be, its explanation is, nevertheless, one of the major problems in the evolution of this ultramafic complex.

The most outstanding feature of the Duke Island ultramafic complex is a remarkable layering, closely resembling graded bedding in sedimentary rocks. An example is shown in figure 2. This phenomenon is extensively developed, and an understanding of its origin is fundamental to explaining the evolution of the complex and may well have a bearing on the interpretations of the other ultramafic bodies in southeastern Alaska.

Evidence on the age of the Duke Island ultramafic rocks is meager. Like the gabbro, they are undoubtedly younger than the metamorphic rocks. In several localities, ultramafic rock is cut by younger granitic rocks. Notable examples are the quartz-feldspar dikes in the hornblende pyroxenite on the north shore of Kelp Island (plate 3), and granitic dikes in the olivine pyroxenite at the northern end of the Judd Harbor ultramafic area. Along

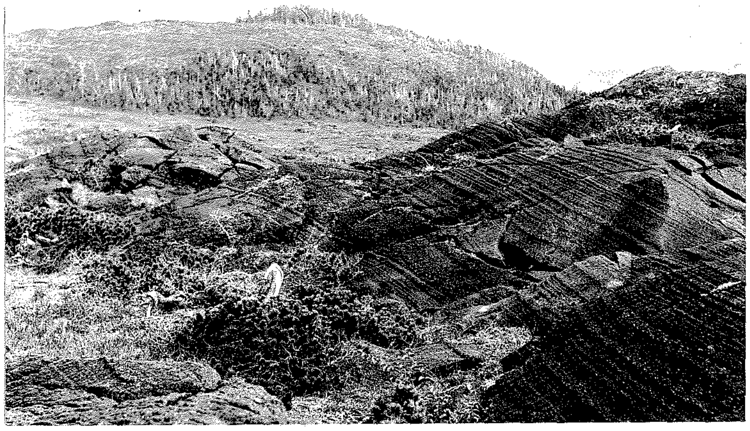


Figure 2. Graded layering in the Hall Cove ultramafic area. The rock is peridotite. Pyroxene stands out in relief on the weathered surface, olivine is depressed. Pyroxene is the coarser mineral and is concentrated in the lower part of each layer. In the background is Knob Hill. The view is to the east.

Form Point, on the north shore of Duke Island, probable inclusions of hornblendite occur in granodiorite.

The ultramafic rocks of southeastern Alaska are so similar in lithology, structure, and tectonic environment, and constitute such a distinctive petrologic suite, that it is reasonable to believe that they are of the same age. Walton (1951a) has made a comprehensive summary of the evidence for the ages of these masses and for ultramafic rocks in British Columbia, as gathered by various investigators, and concludes that they are best considered "to be the oldest plutonic rocks of the Jurassic-Cretaceous petrogenetic cycle." The available evidence on Duke Island does not disagree with this conclusion.

BASIC PEGMATITE

The next youngest rock type after the ultramafic rocks is a remarkable pegmatite of calcic plagioclase (An_{96}) and hornblende. This material is particularly common in the southern part of Duke Island, where it occurs in dikes cutting the ultramafic and gabbroic rocks. Some of it has been described by Koschmann (1935). Identical pegmatite occurs at the Blashke Islands (Walton, 1951) and the Percy Islands (Stebbins, 1957), and somewhat comparable dike rock is present in the ultramafic bodies at Union Bay (Ruckmick, 1957) and Klukwan (H. P. Taylor, Jr., personal communication). Thus the pegmatite seems closely related to the ultramafic rocks in distribution and composition. Similarities in composition will be described in more detail in a later section. On the north side of Judd Harbor, the pegmatite is in contact with and apparently is cut by a small mass of granitic rock. No basic pegmatite was seen to cut granite, although many pegmatite dikes occur in gabbro immediately adjacent to granitic rocks at the mouth of Hall Cove. Therefore it is probable that the pegmatite is pre-granite.

GRANITIC ROCKS

A major body of granitic rocks crops out in the northwest part of Duke Island, and 7 smaller masses and numerous dikes occur in the gabbro that underlies most of the eastern half of the island. As discussed above, some of these rocks are later than the ultramafic bodies. Evidence that granitic rocks cut gabbro is abundant wherever the contact is exposed, and it is probable all the granitic rocks are younger. The metamorphic rocks are intruded and, to some extent, metasomatized by granitic materials as shown along the shore north from Hall Cove. A major granite body, presumably an outlier of the Coast Range batholith, occurs on Annette Island, immediately to the north of Duke Island (Buddington and Chapin, 1929), and probably the acid plutons underlying parts of Duke Island are also related to the batholith.

If so, they would be "Jurassic to Cretaceous" in age (Buddington and Chapin, 1929, p. 253).

The first age determinations based on radioactivity methods for minerals from Alaskan igneous rocks have recently been published by Matzo, Jaffe, and Waring (1958). Lead-alpha ages for zircons from two granitic rocks from southeastern Alaska are given. One, a quartz diorite from Tolstoi Point, Prince of Wales Island, showed an age of 103 million years, and the other, a granodiorite from the Coast Range batholith at Turner Lake, near Taku Inlet, showed an age of 93 million years. These figures are considered by the authors to be compatible with a Cretaceous age assignment (p. 537).

The Duke Island granitic rocks are medium to light gray on both fresh and weathered surfaces. They are commonly foliated by parallelism of biotite, and in places moderate gneissic banding is developed by segregation of quartz into lenses and bands. The trend of foliation tends to parallel the contacts with the older rocks and to wrap around inclusions of foreign material, suggesting that it is a flow pattern.

The term "granitic" here covers the compositional range from quartz diorite to quartz monzonite. A small amount of diorite has been included, but no high potash granites have been found on Duke Island. Modal analyses of thin sections have been plotted on a quartz-plagioclase-potash feldspar diagram in figure 3, and the volume percentage of mafic minerals has been related by means of tie lines. The triangular plot suggests the existence of two groups of rocks. This may well be the result of the limited sampling, although most of the specimens richer in potash feldspar do come from small bodies in the gabbro. The position of the more potash-rich group is a close approximation of the liquidus minimum in the system $NaAlSi_3O_8$-$KAlSi_3O_8$-SiO_2 (Schairer, 1950), and this has been used as evidence of a magmatic origin (e.g. Tuttle and Bowen, 1958). The position of the other rocks is not incompatible with variations which might be expected to develop by fractional crystallization effects. The tie lines show a general inverse relationship between the quantities of potash feldspar and femic minerals, a feature normal to the calc-alkaline trend of igneous rock variation.

In thin section, most quartz shows marked undulatory extinction due to strain. Its texture is one of mutual interference with the alkali feldspars, but it is interstitial to plagioclase in the quartz diorites. Potash feldspar is either untwinned or shows the cross-hatch twinning of microcline. Oriented microperthitic blebs of albite are common in the coarser potash feldspar, and in some thin sections their distribution shows that the host has oscillatory zoning. Rims of both albite on microcline and microcline on albite have been observed. As shown in

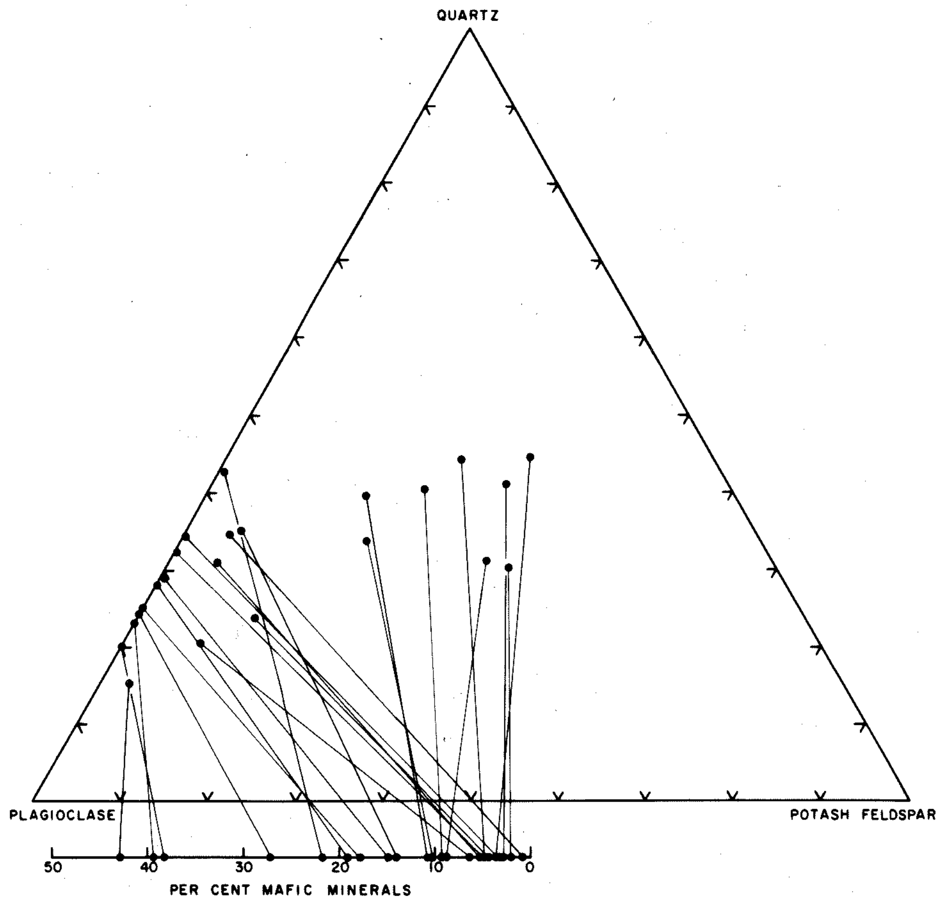


Figure 3.- Volume per cent relationship of quartz, plagioclase, potash feldspar, and mafic minerals in the Duke Island granitic rocks.

table 2, the plagioclase ranges from An_{50} to An_{8}, An_{15} to An_{25} being the common type. Small amounts of myrmekite are present in the more sodic plagioclase. The anorthite-rich plagioclase occurs in rocks with a high content of mafic minerals, mostly from the region just northwest of the Hall Cove ultramafic area where gabbro and granitic rocks are adjacent. The exposure here is poor, and the actual contact was not observed. An almost complete range of types from granodiorite to gabbro is present. The presence of quartz was used as the principal field criterion for drawing the contact, but this may have been too arbitrary, and some of the rock classed as granitic may actually be quartz gabbro. Biotite is ubiquitous, but where mafic minerals are common, hornblende predominates. Chlorite is locally present as an alteration of biotite, and muscovite is relatively common in the more alkalic rocks. Hypersthene was observed in thin sections from four specimens, and in two of these, its indices of refraction indicate compositions of approximately Of_{33} and Of_{36}. This is not particularly iron-rich as compared to the gabbro pyroxenes.

The characteristic texture of the granitic rocks is xenomorphic granular, all mineral grains being anhedral with sutured boundaries. Only the quartz diorite is an exception, as it contains subhedral to euhedral plagioclase. Medium grain size is most typical, the average quartz and feldspar grains being 2-5 mm. In some specimens, quartz

18	I-3-7	1.530-4	1.538	8-16	42.29	25.36	27.43	4.02	--	--	--	0.06	0.78	0.06 zr
19	N-37-3	1.540	1.548	24	39.95	32.45	24.52	0.61	1.82	--	0.30	--	--	0.36 sph
20	I-1-11	1.536	1.540	16	28.32	32.86	29.75	8.88	--	--	--	--	--	0.19 zr
21	S-21-7	1.536	1.540	16	43.09	33.16	20.11	2.99	--	--	--	--	0.32	0.32 mu
22	R-16-1	1.532-6	1.536-9	10-22	20.42	34.23	39.27	5.57	0.36	--	--	--	--	--
23	N-37-8	1.539	1.545	20	29.59	38.36	29.82	1.62	--	--	0.17	--	--	0.45 chl

		Orthopyroxene Composition n_x	n_z	% Of
7	N-18-5	1.692	1.710	33
8	R-23-3	1.698	1.712	36

qtz - quartz	px - orthopyroxene	mu - muscovite
K-spar - microcline	mte - magnetitic oxides	sph - sphene
plag - plagioclase	ap - apatite	gar - garnet
bi - biotite	cz - clinozoisite	zr - zircon
hb - hornblende	chl - chlorite	

and feldspar are segregated, and the quartz may reach grain sizes of 10-15 mm whereas the two feldspars form a relatively fine intergrowth. Commonly, an extremely fine, granular intergrowth of quartz and two feldspars is present in veins which are apparently due to replacement or recrystallization of the coarser material. The coarse material appears to be primary, as the feldspars are perthitic, with euhedral zoning. The fine material is not perthitic. Some cataclasis is probably related to the development of the two grain sizes, as the coarse material is strained whereas the fine is not, but the main change seems to be one of recrystallization.

DIABASE DIKES

The youngest known igneous rocks in the area are small, fine-grained, dark gray diabase or basalt dikes. These intrude the granitic rocks in the vicinity of Point White and cut the ultramafic rocks exposed at East Island. Nothing definite can be said as to their age. Possibly they are equivalent to the Tertiary basalt dikes common in many parts of southeastern Alaska (Buddington and Chapin, 1929, pp. 221-273).

QUATERNARY DEPOSITS

Quaternary deposits are minor and almost entirely of glacial origin. Thin, poorly-sorted accumulations of silty clay, sand, and gravel cover much of the lower areas, particularly those underlain by metamorphic and granitic rocks. Most of this material is locally derived, but a few large boulders are exotic granitoid rocks and probably come from the Coast Range batholith to the northeast. Relatively extensive alluvium covers the lower southern slope of Mount Lazaro. Recent marine sand beaches are rare and small; the largest ones occur in Hall Cove near its mouth, and along the southern shore of the island in the vicinity of Cape Northumberland.

PART III. DETAILED DESCRIPTIONS OF ULTRAMAFIC AND GABBROIC ROCKS

FELDSPATHIC ROCKS

CLASSIFICATION AND MAPPING

Two units of gabbroic rocks have been distinguished on the map: pyroxene gabbro; and hornblende gabbro. This distinction is to some extent arbitrary, because most of the pyroxene gabbro contains some hornblende, but most of the hornblende gabbro is devoid of pyroxene and distinctive in lithology and association.

The mapping of the gabbroic rocks was only a reconnaissance in the early part of the field work. The major divisions were not fully recognized until the work was well along, and the limits shown on the map are largely drawn from field notes and specimens. In spite of these shortcomings, the general distribution of types is believed to be known with sufficient accuracy to justify the conclusions drawn.

A third unit of feldspathic rock is the basic pegmatite. In the ultramafic rocks, it forms distinct dikes, and the larger of these have been mapped with fair accuracy (see plate 3 in particular). In the gabbroic rocks, the pegmatite does not form extensive, well-defined bodies, and because of the limited time spent mapping these areas only the distribution has been shown. Thus on plate 1 any outcrop containing basic pegmatite has been given the color allotted to this unit regardless of whether it is predominant or present in only minor amounts. Pegmatite and hornblende gabbro are generally associated, and outcrops containing both rocks are denoted by the symbol "hgb,hp." If pegmatite is exceptionally abundant, the symbol is "hp,hgb."

PYROXENE GABBRO

Pyroxene gabbro underlies most of the eastern half of Duke Island. It is uniform and generally massive, and is dark gray on both fresh and weathered surfaces. The grain size is surprisingly fine considering the size of the body and is mostly 1-3 mm and rarely exceeds 5-6 mm. Little of the rock is porphyritic.

The principal minerals in thin sections of the pyroxene gabbro are clinopyroxene, orthopyroxene, olivine, plagioclase, hornblende, and ilmenomagnetite. Apatite and sulphides are the main accessories. Biotite is rarely present. Mineralogical and modal data are given in table 3. Chemical analyses of one of the rock specimens (N-36-8) and an orthopyroxene separate (spec. N-25-1) are given in table 8.

Orthopyroxene (hypersthene) and clinopyroxene (augite) commonly are both present in the rock, the latter predominating. However, facies of the gabbro with almost

TABLE 3. OPTICAL DATA ON THE PYROXENE GABBRO

Section 1. Properties of Minerals from Pyroxene Gabbro

Entry	Specimen	Olivine n_y	Olivine % Fa	Clinopyroxene n_y	Clinopyroxene 2V	Orthopyroxene n_x	Orthopyroxene n_z	Orthopyroxene % Of	Plagioclase n_x'	Plagioclase n_z'	Plagioclase % An
1	I-33-1			1.708					1.577	1.583	94
2	N-12-2			1.684		1.690	1.705	29	1.571	1.577	82
3	I-1-9			1.689		1.685	1.698	24	1.572	1.576	81
4	I-11-3	1.690	18.5	1.679	47°15'				1.586	1.572	72
5	I-1-8			1.691		1.685	1.700	25	1.567	1.571	71
6	N-25-1C					1.693	1.708	32	1.566	1.571	70
7	R-21-1			1.701					1.566	1.571	70
8	N-24-13			1.687	48°30'	1.691	1.707	31	1.563	1.570	66
9	I-18-1	1.742	45	1.687		1.696	1.710	34	1.563-6	1.568-72	64-72
10	N-12-8					1.700	1.716	39	1.564	1.568	65
11	N-43-6	1.736	40	1.692		1.688	1.701	27	1.562	1.567	62
12	N-36-8			1.694	46°30'	1.694	1.710	33	1.561	1.567	61
13	R-20-3			1.703		1.697	1.715	36	1.556-64	1.562-70	55-67
14	I-31-9					1.710	1.735	47	1.560	1.567	60
15	N-37-2			1.689					1.561	1.566	60
16	N-15-4			1.697	48°30'	1.707	1.722	45	1.560	1.565	58
17	N-36-9			1.697					1.560	1.565	58
18	R-1-5			1.699					1.559	1.565	58
19	T-13-2					1.692	1.707	31	1.559	1.564	57
20	N-36-5			1.697		1.699	1.717	38	1.559	1.564	56
21	R-5-3			1.697					1.558	1.564	56
22	N-39-5			1.700		1.712	1.727	49	1.556	1.561	51
23	N-36-6			1.693	47°45'				1.556	1.561	51
24	N-14-4	1.805	73	1.706	49°0'	1.715	1.731	52	1.556	1.561	51
25	I-7-3			1.707	49°40'	1.713	1.728	50	1.556	1.562	51
26	I-11-2			1.696		1.705	1.720	43	1.554	1.561	50
27	T-11-4					1.707	1.725	46	1.555	1.560	50
28	N-43-1			1.696		1.709	1.722	45	1.555	1.560	49
29	T-3-1					1.717	1.728	51	1.548	1.555	41
30	R-13-1			1.698		1.718	1.732	54	1.547	1.552	37

Section 2. Modes of Pyroxene Gabbro

Entry	Specimen	ol	clpx	rhpx	hb	plag	mte	ap	Others
2	N-12-2	0.39	18.32	--	33.60	46.90	0.10	0.19	0.48 idd
3	I-1-9	2.32	26.52	14.26	6.53	48.46	1.35	--	0.54 idd
4	I-11-3	37.16	25.56	--	3.91	27.17	6.20	--	--
6	N-25-1C	--	3.63	28.29	24.24	43.64	0.12	0.06	--
8	N-24-13	--	20.20	19.56	13.71	46.47	0.06	--	--
10	N-12-8	--	--	10.92	38.52	41.89	4.94	3.60	0.12 serp
11	N-43-6	8.82	21.95	6.82	19.48	41.63	0.32	0.11	0.74 serp
12	N-36-8	--	22.38	1.53	13.34	48.96	13.68	0.05	--
14	N-31-9	--	--	8.22	56.71	32.00	2.52	0.54	--
15	N-37-2	0.31	24.56	0.87	12.62	59.75	1.74	--	0.12 idd
20	N-36-5	0.06	23.00	6.91	17.80	45.01	7.08	0.06	0.06 serp
22	N-39-5	--	9.76	1.66	15.10	69.20	4.25	--	--
23	N-36-6	0.06	20.53	2.07	28.04	45.64	3.63	--	--
24	N-14-4	9.80	6.10	15.23	22.68	23.52	16.35	6.10	0.06 bi
31	N-14-5	--	9.08	31.35	11.76	47.05	0.63	tr	--
32	N-36-4	--	3.31	0.76	42.23	40.70	12.47	0.38	--
33	N-21-3	0.30	--	35.16	3.63	48.03	8.48	3.73	0.56 bi
34	N-39-1	0.17	9.80	--	20.01	48.14	12.82	--	8.93 cz

ol - olivine
clpx - clinopyroxene
rhpx - orthopyroxene
hb - hornblende
plag - plagioclase
mte - magnetic oxides
ap - apatite
idd - iddingsite (?)
serp - serpentine
bi - biotite
cz - clinozoisite

exclusively one or the other do occur, and where petrographic information is available, the orthopyroxene gabbro or norite has been indicated on plate 1 with the symbol "ngb." Norite may be more common in the western part of the body, but data are sparse. Both pyroxenes occur as subhedral prisms 0.5-3 mm in length. The orthopyroxene is weakly pleochroic in shades of pink and green, and the clinopyroxene is pale green to colorless. Textural indications of relative ages are commonly indefinite or contradictory, but it is probably more characteristic for orthopyroxene to rim or be interstitial to clinopyroxene. This suggests that orthopyroxene appeared later but that the two minerals overlapped considerably in their crystallization periods. Thin sheetlike inclusions of orthopyroxene of the type usually attributed to exsolution are rarely present in some of the clinopyroxene, but the reverse relation has not been observed. According to the indices of refraction, the content of orthoferrosilite equivalent in the orthopyroxenes ranges from 24-52 per cent. The composition of the one analysed pyroxene is Of_{34} and the optical determination (Of_{32}) is in agreement. A chemical analysis has not been obtained for the gabbro clinopyroxene. The charts prepared by Hess (1949, p. 634) correlating chemical composition with optical properties have not proved reliable in the ultramafic rocks but may be more applicable here because they are primarily based on pyroxenes from gabbros. They indicate the range of composition of the clinopyroxene in the Duke Island gabbros to be about $Ca_{37}Mg_{53}Fe_{10}$ to $Ca_{38}Mg_{29}Fe_{33}$. Coexisting pyroxenes and plagioclase have been related graphically in figure 4. The plot shows that both pyroxenes become more iron-rich as the plagioclase becomes more sodic. The variation does not appear to be related to the relative quantities of the two pyroxenes or to have a systematic areal distribution.

Olivine has been recognized in quantity in only 4 of 24 thin sections taken from the pyroxene gabbro and appears to be sporadic. Its occurrence has been indicated on the map with the symbol "pxogb." The range in composition of the olivine, as determined from indices of refraction, is Fa_{18} to Fa_{73} (table 3, section 1). The variation shows the same relation to plagioclase as does that in the pyroxene. The olivine crystals are equant and subhedral and range in size from a small fraction of a millimeter to 2-3 mm. Locally they have undergone partial to complete pseudomorphic replacement by iddingsite (?) and magnetite. As expected, textural relationships show that olivine probably is the first mineral to have crystallized. It is included in the plagioclase and pyroxenes, and generally it has a corona of hypersthene or is bordered by numerous small hypersthene crystals. Kelyphitic rims of pale-green fibrous hornblende surround some of the olivine. These are

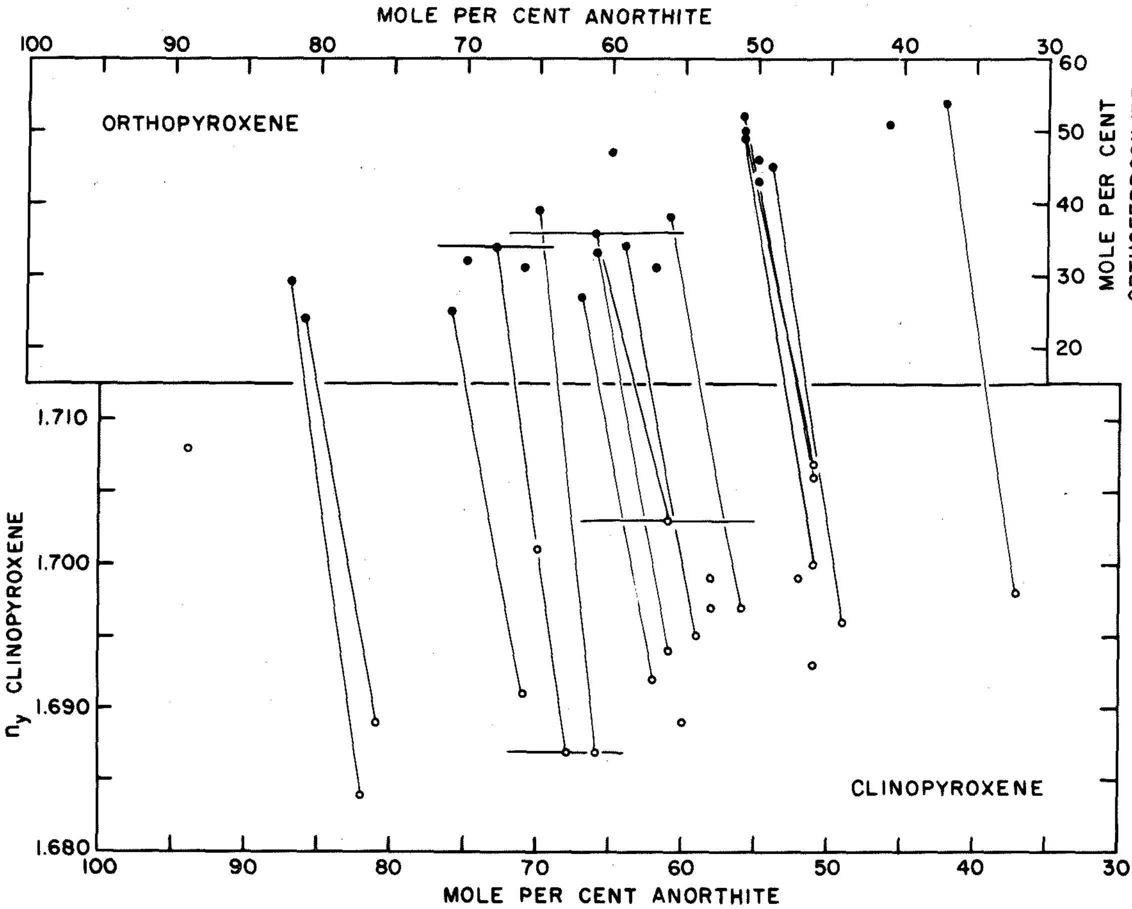


Figure 4.- Comparison of the variations in coexisting pyroxenes and plagioclase in the Duke Island pyroxene gabbro (Based on optical data.).

thin or absent where hypersthene is present.

Plagioclase occurs as subhedral crystals which are usually lathlike but may be equant. Its grain size is 1-4 mm and uniform; rarely does it form phenocrysts. Twinning is primarily of the albite type. The composition of the plagioclase determined from indices of 001 cleavage flakes and the curves of Tsuboi (Rogers and Kerr, 1942, p. 244), ranges from An_{40} to An_{95} but is most commonly An_{50} to An_{75}. Zoning is sparse and weakly developed. Age relations with pyroxene are commonly indefinite, and probably the two mineral species have crystallized simultaneously over most of their range. In many thin sections the plagioclase is slightly clouded by incipient saussuritization, but coarse clinozoisite is rare.

Magnetic oxides (magnetite and ilmenite) are present in quantities that range from a fraction of one per cent to 15 per cent and are characteristically associated with pyroxene and hornblende. They generally form anhedral masses, 0.5-1.0 mm in size. In some specimens they occur as remarkable graphic intergrowths in hypersthene. This texture seems to have developed during the replacement of olivine. Ilmenite is presumed to be present because of the high percentage of TiO_2 in the chemical analysis of the clinopyroxene gabbro (table 8, section 1). As no polished sections of gabbro have been made, the relative amounts and textural relationships of ilmenite and magnetite are unknown.

Hornblende, in appearance, is the most variable mineral in the gabbro. Its color ranges widely in shades of green and brown, even in one thin section. It is commonly brown against the magnetic oxides. According to C. G. Engel (personal communication) a brown color in hornblende commonly indicates the presence of appreciable amounts of TiO_2. The hornblende may be well crystallized or shreddy but is always anhedral. By texture, it is the latest mineral, as it typically rims pyroxene, olivine, and magnetite and commonly veins plagioclase. Pseudomorphous replacement of pyroxene by hornblende is widespread.

Biotite has been found in only one or two specimens collected near granitic bodies and may be an alteration imposed by these younger intrusions.

Apatite is typically associated with magnetite, in places to as much as 6-8 per cent of the rock.

Little is known about the internal structural relationships of the pyroxene gabbro. Parallelism of plagioclase laths is developed only locally. Distinctive layering is known at two places, one just north of the Hall Cove ultramafic area, 1200 feet west of the linear valley marking the continuation of the cove, and the other 4000 feet east of Judd Harbor. The Hall Cove example is the better (figs. 5 and 6). The layers are regular and continu-

Figure 5. Rhythmic layering in norite. Tops of layers are to the upper right. The locality is just north of the Hall Cove ultramafic area, 1200 feet west of the Hall Cove fault.

Figure 6. Rhythmic layering in norite. The seams standing in relief are almost pure pyroxene and mark the base of each layer. The top of the layer is slightly more feldspathic than the normal gabbro. Tops of layers are to the left. The locality is the same as figure 5.

ous for several tens of feet and range in thickness from 2 or 3 inches to 1 foot. The lower boundary of each layer is sharply defined by a marked concentration of pyroxene. Proceeding upward through the layer, the percentage of mafic minerals decreases in a rapid gradation, and the gabbro becomes normal or even slightly feldspathic. Plagioclase laths are well oriented parallel to the plane of the layering. These characteristics are similar to those of rhythmic or gravitational layering of the Skaergaard intrusion (Wager and Deer, 1939, pp. 36-45) and other gabbroic complexes. In both Duke Island localities, the layers strike about N.70°W. and dip south at moderate to steep angles. Tops of layers, determined by grading (Peoples, 1933), face south and indicate that the layers are in normal position.

HORNBLENDE GABBRO

The hornblende gabbro is divisible into two groups by environment and mineralogy. One group occurs in a large area between and partially surrounding the two main ultramafic areas. The other occurs mainly in the eastern parts of Duke Island and is apparently related to the granitic intrusions.

Hornblende Gabbro Associated with Ultramafic Rocks

The hornblende gabbro near the ultramafic rocks is transitional between pyroxene gabbro and basic pegmatite.

It ranges from a relatively fine-grained, uniform rock to a streaky, uneven-textured rock in which porphyroblastic and pegmatitic facies are prevalent. The area of its occurrence has no well-defined limit but grades into pyroxene gabbro and has patches of pyroxene gabbro throughout on both macroscopic and microscopic scales. Hornblende gabbro is almost everywhere cut and permeated by basic pegmatite. These features indicate that the rock is an altered and recrystallized pyroxene gabbro.

Plagioclase and hornblende in about equal amounts are the dominant minerals. The plagioclase is generally as calcic as that in the pyroxene gabbro and grades upward in anorthite content to the high values (An_{95}) typical of the basic pegmatite (table 4). The distributions of the determined plagioclase compositions in the three types of feldspathic basic rocks are shown in histograms in figure 7. The gaps in the histogram for hornblende gabbro may only reflect the limited sampling but may have some real meaning. One of the first stages in the transition from pyroxene gabbro to hornblende gabbro is the appearance in the plagioclase of seams or veins of fine, granular plagioclase of distinctly higher index. At a more advanced stage the original plagioclase is completely reduced to the finer grain sizes, and pyroxene is altered to hornblende. About where pyroxene disappears, the plagioclase becomes coarser again, first as small porphyro-

TABLE 4. COMPOSITION OF PLAGIOCLASE IN HORNBLENDE GABBRO

Entry	Specimen	Plagioclase n_x'	n_z'	% An
1	N-12-6	1.579	1.584	98
2	N-36-3	1.578	1.582	94
3	N-15-9	1.577	1.582	93
4	N-15-10	1.576	1.580-1.583	91
5	N-40-5	1.576	1.581	91
6	T-11-3	1.574	1.579	88
7	R-37-2	1.574	1.580	88
8	N-36-4	1.574	1.579	87
9	N-40-2	1.573	1.578	84
10	S-24-3	1.571	1.576	81
11	N-39-2	1.571	1.576	80
12	N-40-4	1.571	1.576	80
13	R-37-4	1.568	1.574	76
14	R-37-5	1.568	1.573	75
15	I-14-1	1.565	1.570	68
16	R-13-3	1.560	1.566	59
17	R-37-3	1.559	1.565	58
18	N-39-6	1.558	1.564	57
19	N-41-5	1.558	1.564	57
20	R-19-7	1.558	1.562	57
21	N-39-4	1.558	1.563	54
22	R-1-2	1.557	1.562	53
23	N-39-3	1.556	1.561	51
24	I-14-2	1.554	1.560	49
25	N-41-6	1.554	1.560	48
26	I-7-1	1.553	1.559	47
27	N-40-3	1.549	1.554	41
28	R-19-2	1.545	1.550	33
29	R-19-11	1.539	1.545	20

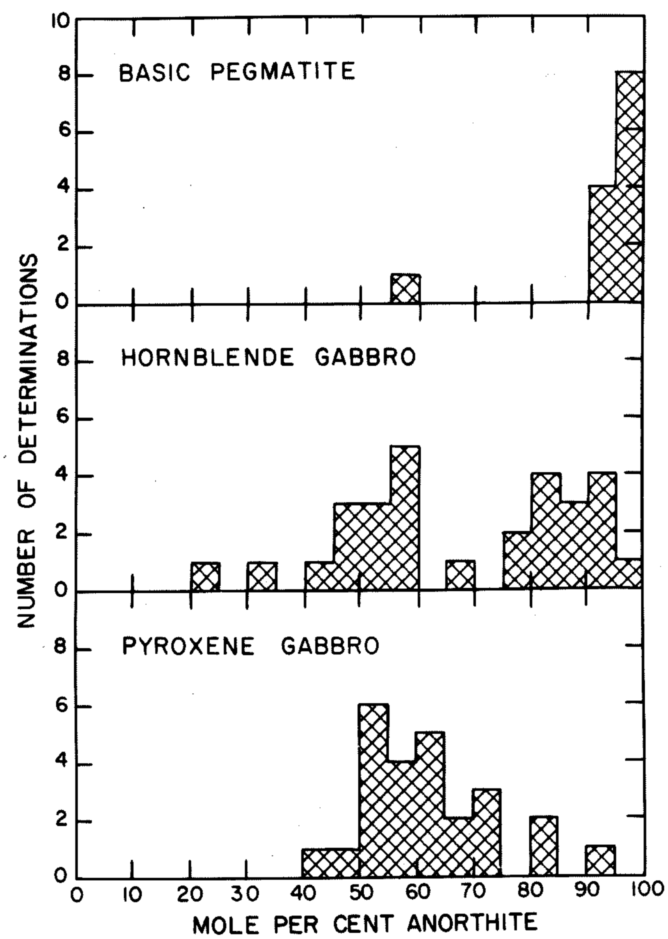


Figure 7.- Histograms for the composition of plagioclase in the different types of mafic rocks occurring at Duke Island, Alaska.

blasts, then along veins or dikelets.

The hornblende is a green variety, moderately to weakly pleochroic, and forms anhedral prisms but is not shreddy or fibrous. Except for the grain size, it is identical in appearance with that in the basic pegmatite.

Clinozoisite as an alteration of plagioclase is the most common of the other minerals. Zones of intense saussuritization of the type common in the Union Bay gabbro (Ruckmick and Noble, 1959) have not been recognized at Duke Island.

Magnetite is erratic in occurrence, to 20 per cent in some places and almost absent in others. On the aeromagnetic map, the level of magnetic intensity is appreciably higher over this type of hornblende gabbro than over pyroxene gabbro, suggesting that magnetite is more common in the hornblende gabbro. This would, however, be difficult to demonstrate by modal analyses because of the erratic distribution of the mineral.

Pyroxene, where present, is enclosed in hornblende. Biotite is rarely present and apatite, pyrrhotite and pyrite are sparse.

A chemical analysis, mode, and norm of a specimen of hornblende gabbro is given in table 8 . However, the specimen is not entirely representative, as it has a lower magnetite content and less calcic plagioclase (An_{57}) than average.

Hornblende Gabbro Associated with Granitic Rocks

The contact of granitic and gabbroic rocks, where exposed, generally is easily delineated and shows clearly that the gabbro is the older rock. For several hundred feet away from the contact, the pyroxene gabbro is brecciated, veined, and permeated by granitic materials and is altered to hornblende gabbro. Feldspathization is common and gives the rock a non-uniform, slightly porphyroblastic texture. Under the microscope these rocks show a characteristic xenomorphic granular texture. The plagioclase is strained, poorly twinned, and more commonly zoned than is typical of the gabbros. In three specimens of the more altered material, the plagioclase was found to be An_{20}, An_{33}, and An_{47}, all more sodic than normal, apparently indicating the addition of alkalies from the granitic intrusions. Replacement of plagioclase by epidote and biotite is common. Pyroxene is rare and occurs only as tiny relict grains after extensive alteration to hornblende. The hornblende is bright green to bluish green with moderate pleochroism and absorption. The color also may indicate addition of alkalies. Biotite is exceptionally abundant for the gabbros and ranges in color from rich brown to reddish-brown.

BASIC PEGMATITE

Dikes of basic pegmatite are widely distributed in the southwestern part of Duke Island and in Kelp Island. They range in width from a fraction of an inch to 200 feet, but most of them are 3 inches to 10 feet. Some of the larger dikes have been traced for 1200 feet along the strike. Essential minerals are hornblende and plagioclase in about equal amounts.

The hornblende in thin section is medium green and weakly pleochroic. Tiny inclusions of opaque material, either magnetite or ilmenite, are common. A chemical analysis of a separate of the hornblende is given in table 8, specimen I-27-1. The calcium and aluminum contents of the hornblende are higher than those of most igneous hornblendes, and the water content given in the analysis is so low that the analysis is probably in error. The analysis closely resembles that of hornblende from the ultramafic hornblendite (table 8, specimen I-31-4). The latter has a lower Fe:Mg ratio, but whether this is a characteristic relationship is not known.

The plagioclase has, according to its indices of refraction, an average composition of about An_{96}. Zoning is generally absent. The exceptionally high calcium content of the plagioclase was first recognized by Koschmann (1935). Walton (1951) has described similar plagioclase at the Blashke Islands, and Kennedy and Walton (1946, p. 83) reported it at Mount Burnett (Union Bay). The association of the calcic plagioclase with ultramafic rocks seems well

TABLE 5. COMPOSITION OF PLAGIOCLASE IN BASIC PEGMATITE

Entry	Specimen	Plagioclase n_x'	n_z'	% An
1	N-39-1	1.580	1.585	99
2	N-15-7	1.579	1.584	98
3	S-24-4	1.579	1.585	98
4	I-28-2A	1.578	1.584	98
5	N-41-3	1.578	1.584	98
6	N-40-1	1.578	1.584	98
7	S-24-6	1.578	1.584	96
8	S-24-1	1.578	1.584	96
9	S-8-2	1.577	1.581	94
10	T-3-2	1.577	1.582	92
11	R-37-1A	1.577	1.582	92
12	I-27-1	1.575	1.581	90
13	R-13-2	1.560	1.566	59

established. The one sample of basic pegmatite with a relatively low anorthite content (fig. 7) comes from along Morse Cove at the east side of Duke Island, well away from the ultramafic rocks. Ilmenomagnetite is a prevalent accessory mineral although somewhat erratic in distribution. In the dikes it occurs as relatively coarse clots, some an inch or two in diameter. Commonly it is present in concentrated seams just outside the dike walls. In the Judd Harbor ultramafic area, extensive alteration zones enriched in magnetite accompany the pegmatite dikes. Clinozoisite is abundant as an alteration of plagioclase along fractures and grain boundaries. This alteration gives the plagioclase a chalky appearance that commonly acquires a purplish surficial stain. Sphene has been observed in one or two thin sections. Biotite is rare.

The grain size of the pegmatite ranges enormously, from a fraction of an inch to almost four feet (fig. 8). Usually it is one quarter to six inches, the coarser material in the larger dikes.

Many of the dikes show a well-developed comb-structure of tapered hornblende crystals oriented normal to the dike walls (fig. 9). Systematic zoning does not seem to occur in the dikes, either in relative abundances of minerals or in the composition of plagioclase.

In the southern part of Duke Island are extensive outcrops of swarms of innumerable dikes stacked side by

Figure 8. Basic pegmatite showing the large size of the hornblende crystals. The plagioclase has the composition Ab_4An_{96}. The locality is the north side of Judd Harbor.

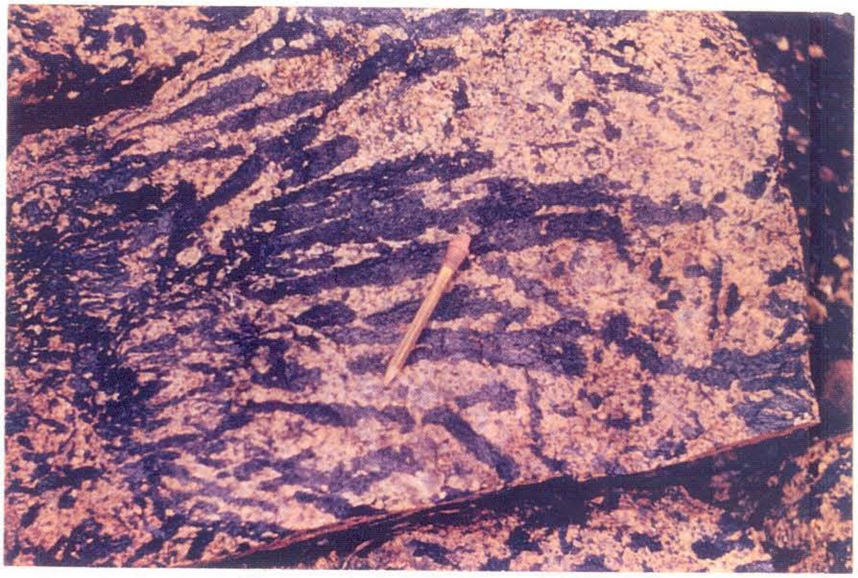


Figure 9. Basic pegmatite showing tapered hornblende crystals in a well-developed comb structure. One of dike walls is just to the left of the photograph, and the crystals become fewer and larger toward the center of the dike. The locality is the small cove about 2000 feet east of Judd Harbor.

side. Many are not separated by septa of country rock, yet they apparently do not intersect one another. Where this phenomenon has its most extreme development the dikes fade into one another, and if comb structure is present, the rock appears to be streaked with zones of perpendicular hornblende prisms. This feature is illustrated in figure 10, and several good examples are shown in Koschmann (1935, figs. 3-6).

ULTRAMAFIC ROCKS

CLASSIFICATION AND RELATIVE ABUNDANCE OF TYPES

The classification of ultramafic rocks is based on the relative abundances of olivine, clinopyroxene, and hornblende. Orthopyroxene is not present in any of the rocks of the main complex. The units were chosen in the field as the most natural for mapping purposes. Although the contacts are gradational, much of the gradation between the principal groups takes place over only a few feet, and once the main compositional breaks were recognized mapping was comparatively straightforward. The classification is given in table 6.

The classification is approximately the same as that used by Ruckmick (1957). The term "pyroxene dunite" is new and applies to small patches of dunite with sporadic clusters and clots of coarse pyroxene and small patches of olivine pyroxenite. The rock differs from peridotite

Figure 10. Basic pegmatite. The rock is streaked with hazy parallel bands of perpendicular hornblende crystals. The bands parallel the swarms of pegmatite dikes in the vicinity. The locality is the southwest corner of Kelp Island.

TABLE 6. CLASSIFICATION OF THE ULTRAMAFIC ROCKS

Rock name and symbol	Approximate volume percentage of mineral			
	Olivine	Clino-pyroxene	Hornblende	Magnetite other opaques
dunite, du	85–100	0–15	––	0–5
pyroxene dunite, pdu	65–85	15–35	––	0–5
peridotite, pd	45–85	15–55	––	0–5
olivine pyroxenite, opx	5–45	55–95	0–10	0–10
hornblende-olivine pyroxenite, hopx	5–10	35–75	10–45	5–15
hornblende pyroxenite, hpx	––	20–80	20–80	5–25
hornblendite, hb	––	0–5	85–95	5–10

in that it is non-uniform.

The basis of the classification can be further illustrated by microscopic data. In figure 11, the results of modal analyses have been plotted on a triangular diagram with coordinates of volume percentages of olivine plus serpentine, pyroxene plus hornblende, and magnetite plus ilmenite and other spinels. The extent of variation of the principal rock groups is shown.

Histograms of the approximate areal abundances of the ultramafic rocks classed according to their relative proportions of olivine to pyroxene plus hornblende are given for the Hall Cove and Judd Harbor ultramafic areas in figure 12. The plots were made by weighting individual modal analyses according to the map areas they are believed to represent. Assumptions in allotting areas and shortcomings in using modes as a method of sampling are, of course, major sources of error. However, the histograms are believed to be improvements over ones which could be obtained by plotting numbers of modes because the thin sections are not systematically distributed. The principal features shown by the histograms are two peaks: one for the rocks essentially devoid of olivine; the other for those containing 15-30 per cent olivine. These correspond to hornblende pyroxenite and olivine pyroxenite respectively. The author believes from field observations that the olivine pyroxenite peak should be sharper, and that the positions

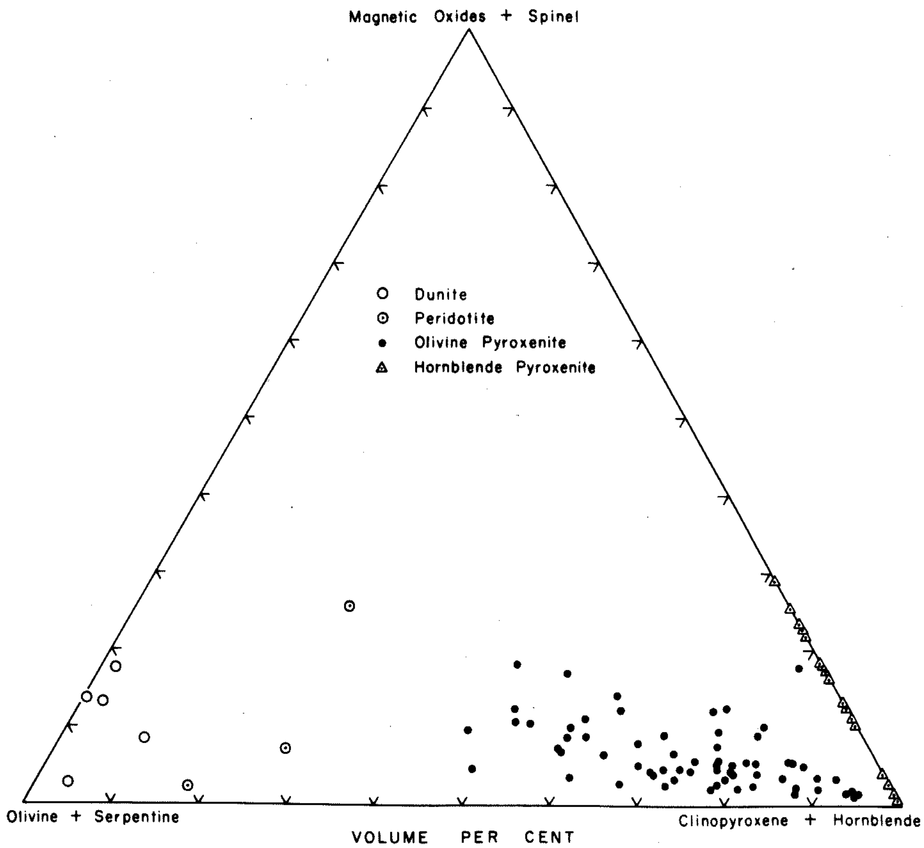


Figure 11. - Modal data on the Duke Island ultramafic rocks.

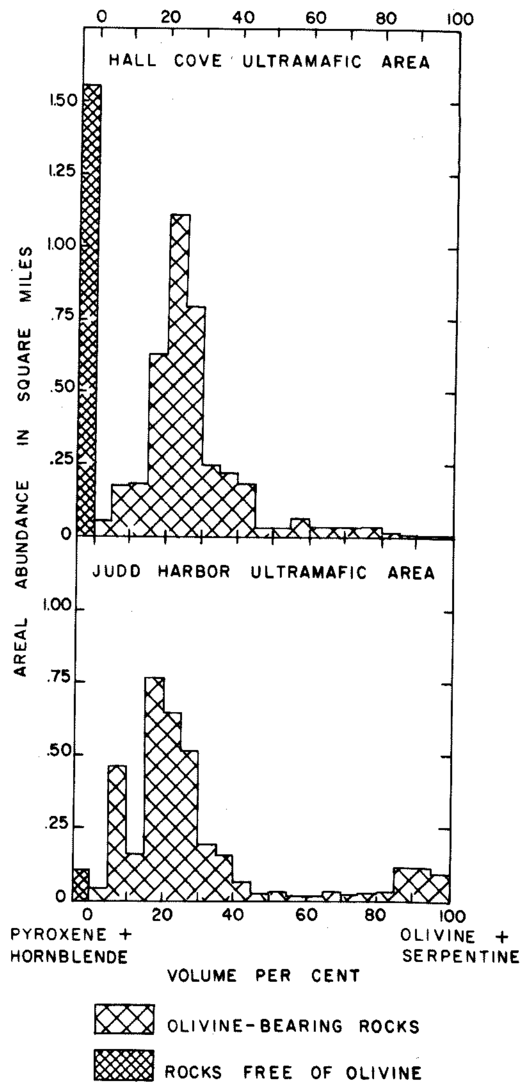


Figure 12.– Histograms showing the areal abundance of the Duke Island ultramafic rocks.

of this peak in the two areas should be interchanged because the olivine pyroxenite in the Judd Harbor ultramafic area is almost certainly the richer in olivine. A third less distinct peak in the histogram for the Judd Harbor ultramafic area is for rocks in the range 80-100 per cent olivine because of the dunite in this area. Similar frequency distributions exist in several other ultramafic bodies in southeastern Alaska. The Union Bay complex would show all three peaks, the high olivine one being much more pronounced because of the larger proportion of dunite present. Most of the "structural peridotite" mapped by Ruckmick (Ruckmick and Noble, 1959) at Union Bay would be two rock types, dunite and olivine pyroxenite, by most methods of sampling, and therefore the range 50-80 per cent olivine would be low on the histogram. The Annette Island complex would give only the dunite peak, whereas the Percy Islands complex has the two peaks in the low olivine region (Stebbins, 1957). In every example peridotite is relatively sparse. In the field common relationships are olivine pyroxenite in contact with ultramafic rocks containing either several times as much olivine or essentially none.

PETROLOGY OF AVERAGE TYPES

Dunite and Peridotite

Dunite and peridotite are close in appearance and can be grouped as a fairly natural unit by mineralogy and distribution. They grade together over considerable distances and are intermingled. Rather than draw an arbitrary boundary between them, they have been distinguished on the map only by different shades of the same color. They will be described together.

Fresh unaltered dunite is uncommon at Duke Island. In the few localities where it occurs, it is greenish gray and weathers to the reddish brown color typical of unaltered dunite occurring in other parts of the world. The more prevalent serpentinized material has a distinctive chamois-like weathered surface and is dull black on the fresh surface because of fine-grained magnetite in the serpentine. The rock is massive and granular in appearance where fresh, but serpentinization tends to obscure grain size so that only rarely can the olivine crystals be seen and then usually only the larger ones by the flash of light from newly broken surfaces. In thin section, the rock is a granular mosaic of equant olivine crystals with simple unsutured boundaries. Grain size is 1-12 mm and tends to be seriate. Generally the rock is laced with tiny veinlets of serpentine, each with fine-grained magnetite either as a medial seam or in a uniform intergrowth. Straining of olivine grains is rare. Where clinopyroxene is present in small amounts, it is interstitial to the olivine, but it may form subhedral prisms where more common. Chromite is the one mineral apparently unique to the dunite and peridotite, as it has not been recognized in the pyroxene-rich rocks. It occurs either as tiny octahedra disseminated widely through the rock and coating fractures, or as massive veins and clots with dimensions of several inches. The latter are weakly to moderately magnetic, probably because of magnetite as intergrowths or in solid solution.

The more peridotitic rocks have greenish gray clinopyroxene, which contrasts sharply with serpentinized olivine. On the weathered surface, the pyroxene stands out in marked relief so that extremely small amounts are detected with ease. It may be interstitial to olivine, or be early in appearance, or may show mutual interference with olivine. It is everywhere coarser than the olivine and commonly forms large poikilitic crystals, 1-2 inches in length, that may contain 30 per cent or more of olivine crystals. Examples of these pyroxenes are shown in figures 13 and 14. Veinlike masses of pegmatitic pyroxene (figs. 15 and 16) are locally common in both dunite and peridotite. They range in width from less than 1 inch to 2 feet, and pyroxene crystals more than 6 inches long have been noted. Coarse olivine occurs in some of the more podlike masses of this material and reaches grain sizes of more than 5 inches.

Modes of thin sections of dunite and peridotite and optical data on their pyroxene and olivine are given in table 7, sections 1 and 2. Chemical analyses of dunite,

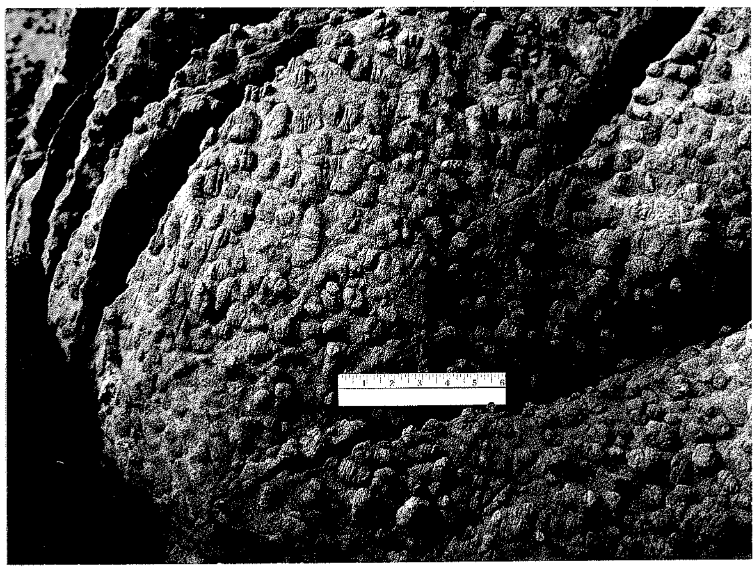


Figure 13. Poikilitic pyroxene crystals in peridotite. The crystals are more than one third olivine inclusions. The locality is the peridotite zone of the Hall Cove ultramafic area.

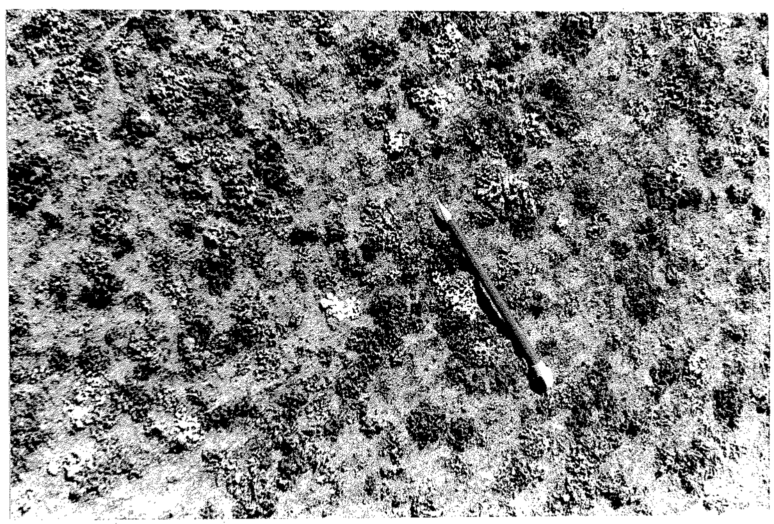


Figure 14. Poikilitic pyroxene crystals in peridotite. The olivine inclusions are coarser than in figure 13. The locality is a small patch of peridotite in the northern olivine pyroxenite zone of the Hall Cove ultramafic area.

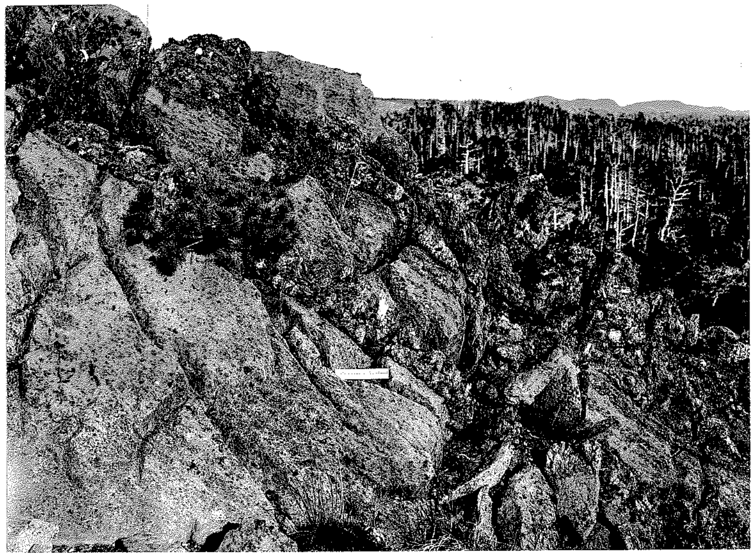


Figure 15. Veins of coarse-grained pyroxene in the peridotite zone of the Hall Cove ultramafic area. The scale is 6 inches.

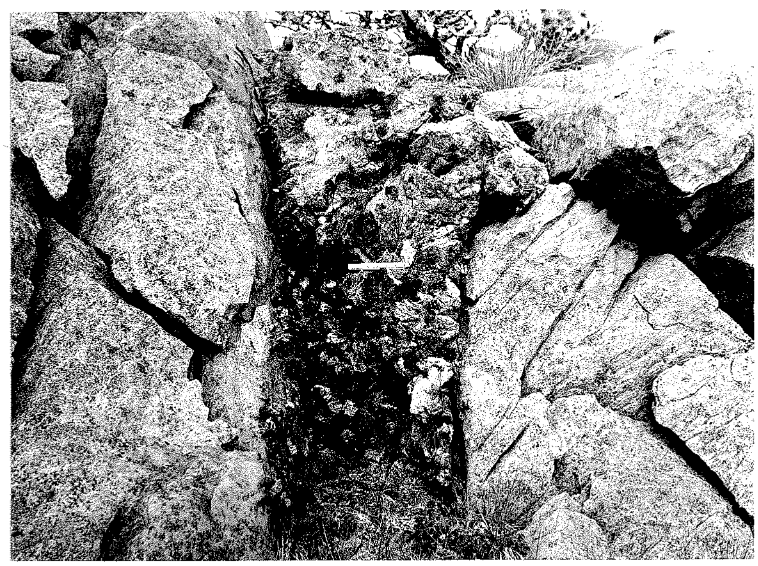


Figure 16. Vein of coarse pyroxene in the peridotite zone of the Hall Cove ultramafic area.

TABLE 7. OPTICAL DATA ON THE ULTRAMAFIC ROCKS

Entry	Specimen	Olivine n_y	Olivine % Fa	Clinopyroxene n_y	Clinopyroxene 2V	Modes ol	Modes clpx	Modes hb	Modes serp	Modes mte	Modes cz	Modes sp
Section 1. Dunite												
1	I-36A-1			1.684		57.04	3.79	--	36.32	2.84	--	--
2	H-4-4	1.684	15.5			37.20	0.15	--	48.70	13.92	--	--
3	H-1-3	1.685	16.0	1.684		17.50	1.55	--	63.19	17.78	--	--
4	I-36-7	1.692	19.5			36.32	2.46	--	47.94	13.28	--	--
Section 2. Peridotite												
5	T-10-3	1.686	16.5									
6	T-10-10	1.686	16.5	1.680								
7	T-10-9	1.689	18.0	1.683								
8	T-10-5	1.688	17.5	1.683								
9	T-10-6	1.689	18.0	1.684								
10	T-10-7	1.688	17.5	1.684								
11	T-10-11	1.688	17.5									
12	T-10-12	1.689	18.0	1.684								
13	T-10-13	1.688	17.5									
14	N-24-15	1.685	16.0	1.685	53°20'							
15	I-40-2	1.687	17.0	1.685								
16	T-10-2	1.689	18.0									
17	T-15-1	1.690	18.5	1.686								
18	T-11-2	1.684	15.5	1.686								
19	I-44-3	1.684	20.5	1.686								
20	H-2-2			1.686								
21	H-1-8	1.689	18.0	1.686								
22	N-5-7	1.690	18.5	1.686								
23	N-24-17	1.690	18.5	1.684	54°50'							
24	N-42-4	1.691	19.0			21.06	2.52	--	62.84	13.58	--	--
25	T-13-3	1.691	19.0									
26	N-24-16	1.692	19.5									
27	N-24-8	1.693	20.0			60.79	9.46	--	21.13	8.53	--	--
28	I-29-1	1.693	20.0	1.689		57.20	5.86	1.95	19.47	15.51	--	--
29	N-24-10	1.694	20.5	1.690	53°20'	79.78	17.04	0.48	0.26	2.43	--	--
30	I-30-2			1.695	54°20'	28.51	23.87	0.26	21.68	25.66	--	--
Section 3. Olivine pyroxenite												
31	T-18-2			1.682								
32	I-38-4			1.682								
33	I-38-3			1.683								
34	T-16-1			1.683								
35	I-36A-2			1.684	56°0'	7.32	77.79	0.29	10.92	3.66	--	--
36	N-24-17	1.685	16.0	1.684	54°35'							
37	I-38-2			1.685								
38	I-46-1	1.686	16.5	1.685								
39	I-37-5	1.689	18.0	1.685		6.96	85.96	2.79	0.87	3.42	--	--
40	T-16-4	1.690	18.5	1.685								
41	T-16-2	1.692	19.5	1.685								
42	T-16-7	1.693	20.0	1.685								
43	T-16-5	1.690	18.5									
44	H-1-1	1.686	16.5	1.686		7.99	80.02	1.40	8.45	2.13	--	--
45	R-38-4	1.688	17.5	1.686								
46	H-1-9	1.689	18.0	1.686		11.44	55.22	5.37	13.72	14.24	--	--
47	T-16-3	1.692	19.5	1.686								
48	T-16-6	1.692	19.5	1.686								
49	T-15-2	1.694	20.5									
50	R-36-3			1.687		38.57	48.36	0.55	7.85	4.83	--	--
51	R-36-4			1.687		11.53	72.39	--	12.53	3.53	--	--

TABLE 7. (Continued)

Entry	Specimen	Olivine n_y	Olivine % Fa	Clinopyroxene n_y	Clinopyroxene 2V	Modes ol	clpx	hb	serp	mte	cz	sp
Section 3. Olivine Pyroxenite (continued)												
52	R-36-5			1.687		7.89	72.75	1.65	8.02	9.67	--	--
53	R-27-3			1.687		3.99	71.48	19.49	1.63	3.40	--	--
54	N-12-3			1.687	54°10'	26.78	51.76	0.48	10.22	10.76	--	--
55	N-24-18	1.689	18.0	1.687	55°0'	16.78	76.75	0.11	1.64	4.71	--	--
56	H-11-9	1.689	18.5	1.687								
57	N-12-1	1.692	19.5	1.687	52°20'	19.77	49.34	0.24	18.10	12.54	--	--
58	I-45-1	1.694	20.5	1.687								
59	I-44-3	1.694	20.5	1.687								
60	T-10-4	1.695	21.0	1.687								
61	T-10-8	1.696	21.5	1.687								
62	R-36-2			1.688		9.32	86.98	1.93	2.13	3.53	--	--
63	R-32-9			1.688		43.69	26.02	--	22.99	7.29	--	--
64	R-41-1			1.688		24.47	61.17	0.66	1.44	12.24	--	--
65	R-32-8			1.688								
66	I-37-2			1.688		23.17	71.18	0.81	2.29	2.54	--	--
67	I-37-3			1.688		2.84	93.52	0.99	1.32	1.32	--	--
68	T-9-3			1.688								
69	R-32-3			1.688		27.13	42.39	15.81	2.85	11.02	0.79	--
70	R-32-1			1.688		9.97	70.26	3.39	11.80	4.56	--	--
71	R-36-1			1.688		9.52	69.26	8.46	9.79	2.86	--	--
72	R-27-2	1.687	17.0	1.688		9.39	78.68	--	7.66	4.26	--	--
73	I-36-2	1.687	17.0	1.688	55°15'	9.34	75.74	2.78	7.83	4.33	--	--
74	I-36-1	1.689	18.0	1.688	54°55'	18.62	67.32	0.06	8.68	5.24	--	--
75	I-36-12	1.689	18.0	1.688		8.43	77.36	0.69	8.31	5.19	--	--
76	N-24-14	1.693	20.0	1.688	53°10'	33.83	48.08	8.82	1.82	7.15	0.30	--
77	N-24-7			1.689	53°10'	24.59	69.62	0.13	1.58	4.08	--	--
78	N-34-1			1.689	55°10'	3.88	63.54	29.54	1.49	1.55	--	--
79	N-30-1			1.689		20.49	30.57	3.34	36.82	8.71	0.05	--
80	H-1-2			1.689		8.96	75.01	9.52	0.97	5.53	--	--
81	I-36-3	1.687	17.0	1.689	55°5'	22.09	59.24	0.42	9.24	9.00	--	--
82	H-1-4	1.692	19.5	1.689		15.37	58.08	4.84	15.09	6.61	--	--
83	R-38-1	1.695	21.0	1.689		31.10	55.58	1.43	1.82	10.06	--	--
84	N-42-1			1.690	54°0'	8.39	81.89	5.04	2.55	2.11	--	--
85	R-32-4			1.690		16.09	70.18	2.03	6.74	4.94	--	--
86	R-32-7			1.690		12.32	73.52	8.21	1.89	4.05	--	--
87	I-38-5			1.690								
88	R-28-1			1.690		3.22	92.35	1.07	1.39	1.90		
89	I-39-2			1.690								
90	I-30-1			1.690		13.11	61.75	9.06	9.32	6.75		
91	I-36-4	1.693	20.0	1.690	56°10'	25.67	41.14	19.05	10.32	3.81		
92	I-36-10	1.693	20.0	1.690	52°35'	18.37	55.98	1.86	16.74	7.03		
93	I-39-3	1.695	21.0	1.690								
94	I-37-2	1.696	21.5	1.690		23.17	71.18	0.81	2.29	2.54		
95	I-31-2	1.696	21.5	1.690		5.52	69.21	10.18	3.91	11.17		
96	N-30-3			1.691	53°0'	6.21	76.08	0.25	11.81	5.66		
97	I-36-11			1.691		29.12	48.82	1.67	9.63	10.74		
98	R-38-5			1.691		14.26	76.13	0.22	4.00	5.38		
99	H-11-7	1.693	20.0	1.691		8.92	82.56	2.51	0.79	5.23		
100	R-38-3	1.694	20.5	1.691		20.20	69.88	0.58	4.49	4.48	--	--
101	I-36-8	1.695	21.0	1.691	55°0'	7.42	53.10	27.30	10.13	2.05	--	--
102	R-41-3	1.695	21.0	1.691								
103	T-9-1	1.696	21.5	1.691								
104	N-24-1			1.692	51°45'	20.70	68.41	0.08	1.57	9.23	--	--

TABLE 7. (Continued)

Entry	Specimen	Olivine n_y	Olivine % Fa	Clinopyroxene n_y	Clinopyroxene 2V	Modes ol	clpx	hb	serp	mte	cz	sp
Section 3. Olivine Pyroxenite (continued)												
105	I-36-5			1.692	55°45'	2.74	77.09	17.15	1.91	1.08	--	--
106	I-36-6			1.692	55°40'	12.22	69.30	7.97	8.38	2.12	--	--
107	I-36-9			1.692	55°5'	6.38	86.16	3.60	1.83	2.02	--	--
108	R-38-2			1.692		11.51	79.52	1.27	2.12	5.57	--	--
109	R-44-2			1.692		13.33	72.94	0.06	1.84	12.32	--	--
110	H-18-5			1.692		14.06	76.24	3.22	0.79	5.69	--	--
111	I-29-2	1.694	20.5	1.692	54°50'	14.84	68.33	6.75	2.17	7.90	--	--
112	H-18-1	1.696	21.5	1.692		24.87	65.81	0.19	1.08	8.04	--	--
113	R-44-1	1.698	22.5	1.692		19.21	73.28	0.10	1.57	5.83	--	--
114	I-31-1	1.694	20.5	1.693	55°30'	7.90	41.34	12.19	20.82	16.78	0.54	0.42
115	H-18-2	1.695	21.0	1.693		29.40	46.11	0.92	5.53	18.02	--	--
116	N-14-1			1.694	53°0'	13.39	66.43	7.50	6.73	5.95	--	--
117	I-32-2	1.690	18.5	1.694	54°10'	10.87	56.51	17.30	2.79	12.46	0.06	--
118	H-11-1					30.40	44.56	0.91	14.16	9.97	--	--
119	I-37-4					20.35	69.05	--	6.02	4.58	--	--
120	N-12-7					10.91	66.46	0.24	19.51	2.87	--	--
121	S-19-1					6.71	74.86	11.60	1.77	5.05	--	--
122	S-29-5					2.97	78.08	0.99	8.97	9.03		
Section 4. Hornblende-Olivine Pyroxenite												
123	T-18-4			1.681								
124	T-1-2	1.697	22.0	1.689								
125	H-1-7			1.693								
126	T-7-3			1.694								
127	T-6-1			1.695								
128	T-2-6			1.695								
129	T-7-2			1.695								
130	H-19-4			1.695		2.90	62.87	16.26	--	17.95		
131	N-34-3			1.697	54°40'							
132	T-12-2			1.697								
Section 5. Hornblende Pyroxenite												
133	T-13-1			1.690								
134	N-2-1			1.693		--	68.07	20.59	--	11.18	0.16	--
135	T-12-3			1.695								
136	T-7-1			1.695								
137	I-31-3			1.696	57°0'	--	68.58	27.40	--	4.01	--	--
138	H-19-2			1.697		--	50.73	31.12	--	17.52	--	0.63
139	H-19-3			1.698		--	58.29	19.31	--	21.99	--	0.41
140	N-4-1			1.699	53°35'	--	64.48	14.13	--	8.33	4.23	8.83
141	T-20-1			1.700								
142	R-43-4			1.700		--	12.63	64.79	--	20.60	--	1.98
143	I-32-5			1.698-1.707								
144	N-3-1			1.701		--	31.56	49.88	--	18.31	0.23	tr
145	N-42-5			1.708	52°10'	--	48.67	38.04	--	12.74	0.18	0.36
146	N-30-4			1.709								
147	N-30-2			1.709	53°10'	--	18.40	68.99	--	12.54	--	0.06
148	I-36-13					0.06	59.76	21.45	--	18.13	0.60	--
149	I-37-1					--	30.24	68.88	--	0.88	--	--
150	N-4-3					--	38.24	38.11		8.01	0.32	15.31
151	N-5-8C					--	1.25	69.13	--	29.00	--	0.63
152	N-14-2					--	1.81	71.95	--	25.43	--	0.80
153	N-15-1					--	--	88.76	--	10.42	0.81	--
154	N-24-12					--	--	97.21	--	2.78	--	--
155	N-30-5					--	--	95.98	--	1.36	2.65	--
156	N-50-1					--	--	80.42	--	16.53	0.31	2.55

TABLE 7. (Continued)

Entry	Specimen	Olivine n_y	Olivine % Fa	Clinopyroxene n_y	Clinopyroxene 2V	
Section 6. Miscellaneous						
157	H-21-1			1.684		Pyroxene vein in peridotite
158	T-11-1	1.692	19.5	1.685		Pegmatitic segregation
159	T-40-1			1.689		Pyroxene vein in peridotite
160	T-18-1			1.682		Light green pyroxenite, shore of Hall Cove
161	T-25-1			1.685		Coarse pyroxene, East Island
162	I-46-1			1.686		Fine-grained pyroxene vein
163	N-24-4			1.690	52°30'	Coarse-grained pyroxene vein
164	T-9-2			1.695		Pyroxene vein

and olivine and pyroxene from peridotite appear in table 8.

The chemical analysis of the olivine from specimen I-40-2 shows that its composition is Fa_{17}. The same composition is given by optical data and a curve published by Poldervaart (1950, fig. 2) to relate composition to n_y index of refraction. Indices indicate the range of composition in olivine in dunite and peridotite to be Fa_{16}-Fa_{21}.

The analysed clinopyroxene (Specimen I-40-1) is from a coarse vein similar to that shown in figures 15 and 16. Its composition in terms of atomic ratios is $Ca_{48}Mg_{42.5}Fe_{9.5}$, or about $Di_{82}He_{18}$. The n_y index of this pyroxene is 1.689 and is near the high end of the range for n_y indices (1.684-1.691) of pyroxenes from dunites and peridotites. The indices of clinopyroxenes generally increase with increasing Fe:Mg ratio, but the charts prepared by Hess (1949) to relate chemical composition of clinopyroxenes to their optical properties have not proved reliable for the pyroxenes in the ultramafic rocks. This point is discussed on page 77.

The analysed dunite specimen (No. H-4-4) is considerably serpentinized, as indicated by 8.85 per cent H_2O+. Serpentinization results in the oxidation of iron and the appearance of hypersthene in the norm. If the analysis is recalculated to eliminate the normative hypersthene, the result given in table 9 is obtained. The composition of the olivine in the norm of the recalculated analysis is

TABLE 8. CHEMICAL ANALYSIS AND RELATED DATA

Section 1. Rocks

Entry	A	C	D	H	G
Rock	du	opx	hpx	pxgb	hgb
Specimen	H-4-4	R-38-2	I-31-3	N-36-8	N-39-6
SiO_2	35.39	47.55	46.63	43.83	48.96
Al_2O_3	1.34	4.77	7.08	17.06	19.34
Fe_2O_3	6.57	2.93	3.91	6.63	2.56
FeO	7.72	5.58	5.01	10.07	6.21
MgO	38.16	18.46	13.74	6.18	6.79
CaO	0.80	18.66	21.66	10.65	11.38
Na_2O	0.27	0.45	0.65	2.05	2.67
K_2O	0.02	0.03	0.08	0.03	0.14
H_2O+	8.85	0.70	0.29	0.75	1.20
H_2O-	0.55	0.20	0.20	0.25	0.25
TiO_2	0.12	0.52	0.74	2.37	0.41
Cr_2O_3	0.25	0.23	0.15	N.D.	N.D.
MnO	0.11	0.03	tr	0.12	0.15
P_2O_5	tr	0.04	tr	0.11	0.02
CO_2	tr	nil	nil	nil	nil
	100.15	100.15	100.14	100.10	100.08

C.I.P.W. Norms

	A	C	D	H	G
magnetite	10.39	4.22	5.83	9.60	3.75
ilmenite	0.16	0.92	1.38	4.60	0.77
chromite	0.49	0.22	0.22	--	--
leucite	--	--	0.43	--	--
nepheline	--	2.01	2.85	--	--
orthoclase	--	--	--	--	1.11
albite	2.30	--	--	17.51	22.88
anorthite	2.74	11.25	16.21	37.71	40.92
heden-bergite	--	7.52	7.47	4.52	4.02
diopside	1.18	57.60	55.45	8.32	9.22
ferro-silite	0.72	--	--	5.60	3.88
enstatite	8.46	--	--	9.35	7.33
fayalite	6.69	2.37	0.51	1.13	2.27
forsterite	66.86	13.88	6.00	1.64	3.84
larnite	--	--	3.64	--	--
	99.99	99.99	99.99	99.98	99.99

Modes

	A	C	D	H	G
olivine	37.20	11.51	--	--	--
serpentine	48.70	2.12	--	--	--
clinopyroxene	0.15	79.52	68.58	22.38	0.16
orthopyroxene	--	--	--	1.53	--
hornblende	--	1.27	27.40	13.34	63.74
plagioclase	--	--	--	48.96	27.06
magnetic oxides	13.92	5.57	4.01	13.68	0.21
clinozoisite	--	--	--	--	8.82
apatite	--	--	--	0.05	--
	99.97	99.99	99.99	99.94	99.99

Analyst, W. H. Herdsman, Glasgow, Scotland

TABLE 8. (Continued)

Section 2. Minerals

Mineral	olivine		clinopyroxene						orthopyroxene		hornblende	
Entry	A'		B'	C	C'	D	D'		I		E	J
Rock	pd		pd	opx	opx	hpx	hpxm		ngb		hb	hp
Specimen	I-40-2		I-40-1	I-37-2	R-38-2	I-31-3	H-19-3		N-25-1		I-31-4	I-27-1
SiO_2	38.29		50.80	51.50	49.71	49.45	48.42		49.80		41.20	42.48
Al_2O_3	0.49		4.04	4.31	3.58	7.03	6.38		4.14		16.16	14.56
Fe_2O_3	0.68		1.39	1.17	1.63	2.44	2.31		1.51		4.12	3.92
FeO	15.52		4.53	4.36	3.66	4.01	3.83		19.36		7.04	9.51
MgO	42.24		14.79	15.20	15.96	11.84	14.33		21.28		13.96	11.63
CaO	1.69		23.26	22.42	23.98	24.19	22.86		2.80		12.63	12.26
Na_2O	0.34		0.37	0.47	0.61	0.22	0.66		0.36		2.27	2.53
K_2O	0.04		0.03	0.04	0.09	0.05	0.09		0.08		0.62	0.41
H_2O+	0.10		0.30	0.20	0.30	0.10	0.25		0.21		0.38	0.91
H_2O-	0.20		0.09	nil	nil	0.20	0.25		0.11		0.29	0.22
TiO_2	0.07		0.36	0.32	0.56	0.56	0.68		0.12		1.42	1.38
Cr_2O_3	0.15		N.D.	N.D.	N.D.	N.D.	N.D.		N.D.		N.D.	N.D.
MnO	0.15		0.10	0.06	0.05	0.05	0.08		0.37		0.04	0.33
P_2O_5	tr		0.08	0.06	tr	0.03	tr		tr		0.03	tr
CO_2	nil		nil	nil	nil	nil	nil		nil		nil	nil
	99.96		100.14	100.11	100.13	100.17	100.14		100.14		100.16	100.14
x + y[1]	2.055	w+x+y[2]	1.984	2.013	2.129	1.958	2.074	x+y[2]	2.074	w+x+y[4]	8.39	7.99
z	2.000	z	2.028	2.000	1.938	2.020	1.962	z	1.962	z	8.00	8.00
%Al in z	2.2	%Al in z	5.4	5.6	4.7	9.5	8.3	%Al in z	5.0	%Al in z	24.9	21.3
		Ca[2]	48.0	46.8	47.7	53.0	48.2	Ca	5.7			
		Mg	42.4	44.1	44.1	36.2	42.0	Mg	60.7			
		Fe	9.4	8.9	8.0	10.7	9.7	Fe	33.6			
		Ca[3]	46.0	44.3	46.1	49.9	45.0					
		Mg	44.1	46.4	45.6	38.9	44.7					
		Fe	9.9	9.3	8.3	11.4	10.3					
$\frac{Fe+Mn}{Fe+Mn+Mg}$	17.7	$\frac{Fe+Mn}{Fe+Mn+Mg}$	18.2	16.8	15.3	22.8	18.8	$\frac{Fe+Mn}{Fe+Mn+Mg}$	35.6		30.4	49.6
n_y	1.687	n_y	1.689	1.688	1.692	1.696	1.698	n_x	1.693			
		2V	--	--	--	57°	--	n_z	1.708			

Analyst, W. H. Herdsman, Glasgow, Scotland

[1] Calculated on the basis of 4 oxygen and 2.000 z atoms

[2] Calculated following Hess, 1949

[3] Atomic ratios of Ca, Mg and Fe after subtraction of Ca as $CaAl(Al,Si)_2O_6$

[4] Calculated on the basis of 24 oxygen and 8.00 z atoms

Symbols:

du	- dunite	hb	- hornblendite
pd	- peridotite	hp	- basic pegmatite
opx	- olivine pyroxenite	pxgb	- pyroxene gabbro
hpx	- hornblende pyroxenite	ngb	- norite
hpxm	- hornblende pyroxenite with magnetite	hgb	- hornblende gabbro

Fa_{16}, the same as that of the olivine in the rock as determined from its index of refraction. This suggests that the recalculated analysis is a fairly good approximation of the composition of the unserpentinized material, and that serpentinization did not involve appreciable chemical change beyond the addition of water and oxygen. The recalculated analysis agrees reasonably well with the olivine analysis (table 8, section 2, entry A') and with the analysis of unserpentinized dunite in the Union Bay complex (Ruckmick, 1957). Anorthite and nepheline appear in the norm but not the mode.

TABLE 9. CHEMICAL ANALYSIS OF DUNITE SPECIMEN H-4-4 RECALCULATED TO ELIMINATE NORMATIVE HYPERSTHENE

Oxide	Percentage	Norm	
SiO_2	38.92	Mt....................2.57	
Al_2O_3	1.47	Cr....................0.50	
Fe_2O_3	1.76	Il....................0.17	
FeO	14.14		
MgO	41.97	Ne....................1.26	
CaO	0.88	An....................2.76	
Na_2O	0.30		
K_2O	0.02	He.........0.28	
TiO_2	0.13	Di.........0.96	.....1.24
MnO	0.12	Fa........17.80	
Cr_2O_3	0.27	Fo........73.61	....91.41
	99.98		

Olivine Pyroxenite

The fresh surface of typical olivine pyroxenite is medium greenish gray, mottled with dark gray where the olivine grains are extensively serpentinized. The weathered surface has a brownish cast with small yellowish brown pits marking the sites of olivine crystals (fig. 17). In the shoreline exposures, the reactions of pyroxene and olivine to weathering are commonly reversed, and the olivine stands out slightly in relief.

The common grain size in olivine pyroxenite is 1-8 mm for pyroxene and 0.5-4.0 mm for olivine. The olivine pyroxenite in the Judd Harbor ultramafic area is commonly coarser than average, the pyroxene being 2-15 mm and the olivine 1-8 mm. Seriate grain sizes are the general rule, and a considerable range is present everywhere. Abnormally coarse-textured variants are common (fig. 18) and some of these underlie remarkably large areas.

Modal analyses of thin sections of olivine pyroxenite are given in table 7, sections 3 and 4. As shown in figures 11 and 12, the olivine content of the rock ranges from 5 to 45 per cent, the most common rock having about 25 per cent. Hornblende is sparse in the vicinity of major dunite and peridotite bodies and only reaches appreciable quantities adjacent to the hornblende pyroxenite zone and some of the basic pegmatite dikes. In localities where it is especially abundant, the map designation for olivine pyroxenite, "opx,"

Figure 17. The typical surficial appearance of olivine pyroxenite is shown. The pits result from the weathering of olivine.

Figure 18. Coarse-grained olivine pyroxenite. The locality is East Island.

has been modified to "hopx." Olivine pyroxenite is devoid of plagioclase except in a few places adjacent to basic pegmatite, where the plagioclase probably has been introduced metasomatically.

Under the microscope, the olivine pyroxenite shows a simple packing of subhedral grains of pyroxene and olivine. Most commonly, olivine is interstitial, but the rounded forms of primary olivine crystals may adjoin or be surrounded by pyroxene. Examples of each mineral included in the other have been observed and may even be seen in one thin section. These textural relationships and the very constant proportion of olivine to pyroxene throughout most of the unit strongly suggest that these two minerals have overlapped considerably in their periods of formation. Magnetite as a primary mineral occurs as tiny subhedral to anhedral grains that are both interstitial to and included in the pyroxene and olivine. Weakly pleochroic green hornblende is everywhere the latest primary mineral and is characteristically interstitial to pyroxene. Yellowish-green serpentine with fine magnetite dust generally is present as an alteration of olivine along grain boundaries and in tiny veinlets. Reddish iddingsite (?) is a rare alteration product of olivine.

A chemical analysis, mode, and norm of a typical specimen (R-38-2) of olivine pyroxenite are given in table 8, section 1. Analyses of pyroxene separates from this specimen and from another (I-37-2) appear in table 8, section 2.

Their compositions are $Ca_{48}Mg_{44}Fe_8$ and $Ca_{47}Mg_{44}Fe_9$ respectively, and their n_y refractive indices are 1.692 and 1.694 respectively. Optical data on pyroxene and olivine in olivine pyroxenite are listed in table 7, sections 3 and 4. The range of n_y for pyroxene in olivine pyroxenite is 1.682-1.694 and in hornblende-olivine pyroxenite is mostly 1.689-1.698. The range of n_y for olivine is 1.685-1.698, which indicates a compositional range of Fa_{17}-$Fa_{22.5}$. In any rock the refractive indices indicate that olivine has a slightly higher Fe:Mg ratio than the coexisting pyroxene. Figure 19 is a plot of the n_y indices of coexisting pyroxene and olivine. A graphical comparison of the indices of pyroxene and olivine from the three major ultramafic units is made in figure 20. The histograms show that the indices of the olivine proxenite minerals overlap those from peridotite and dunite but have a significantly higher mean value. Almost all the pyroxenes analysed from hornblende pyroxenite have indices greater than those from olivine pyroxenite. Figure 21 is a plot of 2V against n_y for the clinopyroxenes of the various ultramafic rocks and the pyroxene gabbro.

Hornblende Pyroxenite

The typical Duke Island hornblende pyroxenite is massive, medium-grained, and dark greenish gray on both fresh and weathered surfaces. As a rule it is slightly finer-grained than the olivine pyroxenite, although

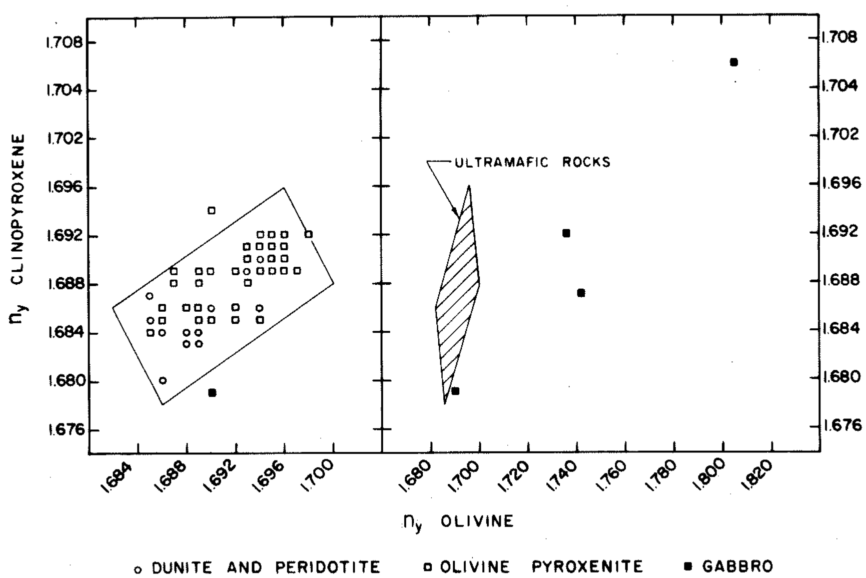


Figure 19.- Comparison of the refractive indices of clinopyroxene and olivine in the ultramafic and gabbroic rocks.

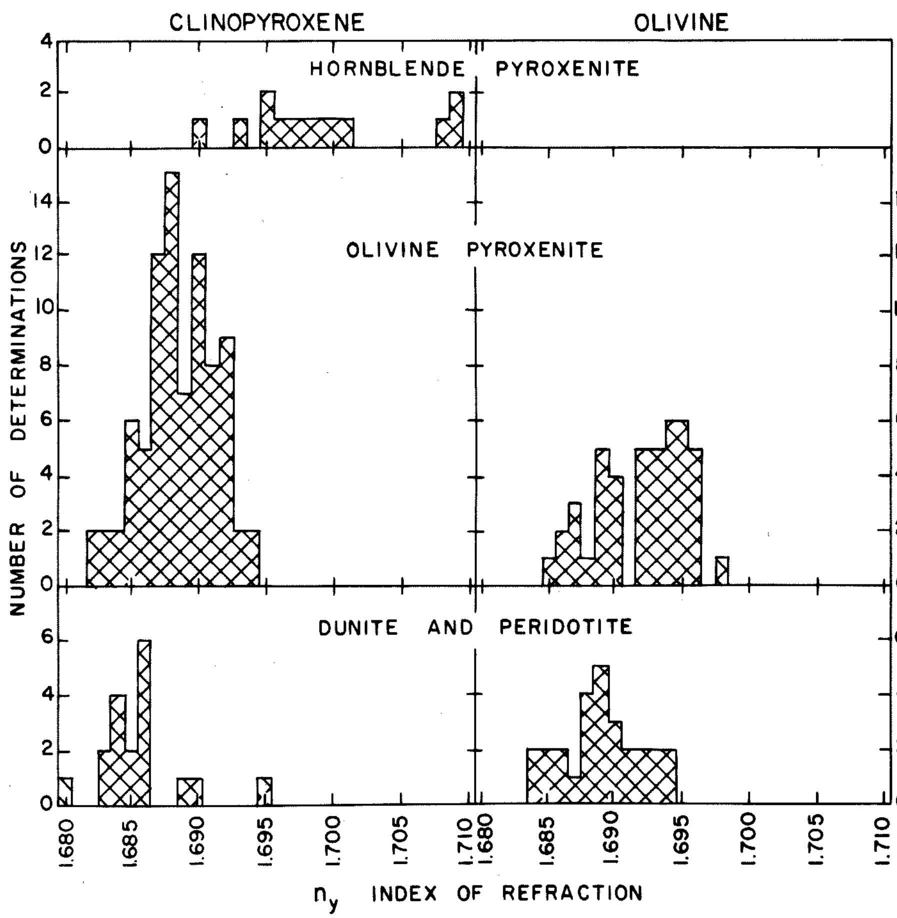


Figure 20.- Comparison of the refractive indices of clinopyroxene and olivine in the major ultramafic rock units.

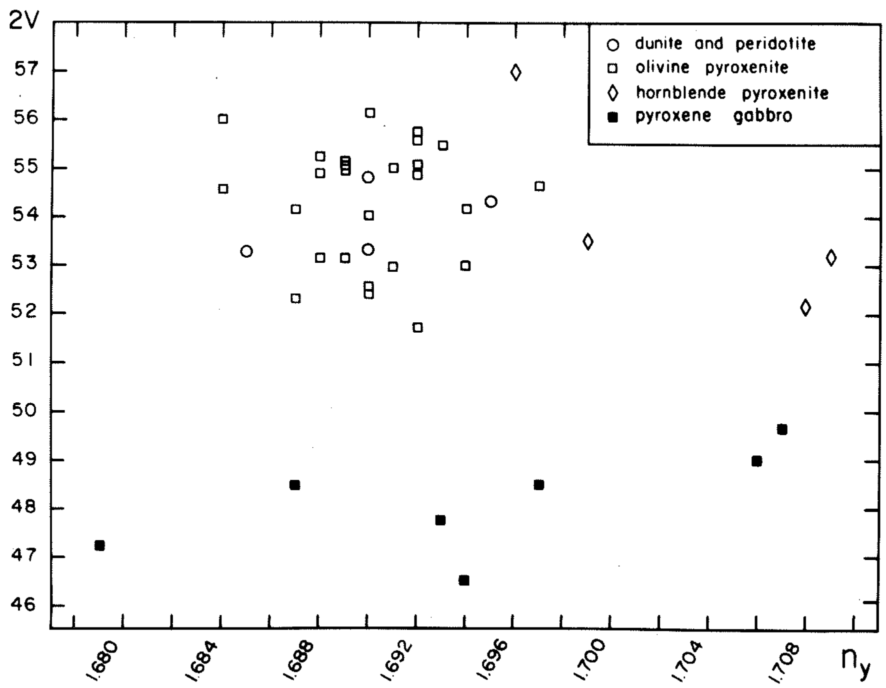


Figure 21.- Plot of n_y against 2V for clinopyroxene from Duke Island rocks.

hornblende may form large pegmatitic or veinlike crystals. The rock commonly is easily disintegrated, hence good specimens are difficult to obtain, and the exposures are not as extensive as those of other ultramafic units. Diamond drill holes show that this poorly consolidated condition persists to depths of at least 500 feet and is proportional to the amount of hornblende.

Modes of the hornblende pyroxenite and optical data on the pyroxene from this unit are given in table 7, section 5. A chemical analysis of a specimen (I-31-3) is given in table 8, section 1, and chemical analyses of two pyroxene separates (Specimens I-31-3 and H-19-3) are given in table 8, section 2.

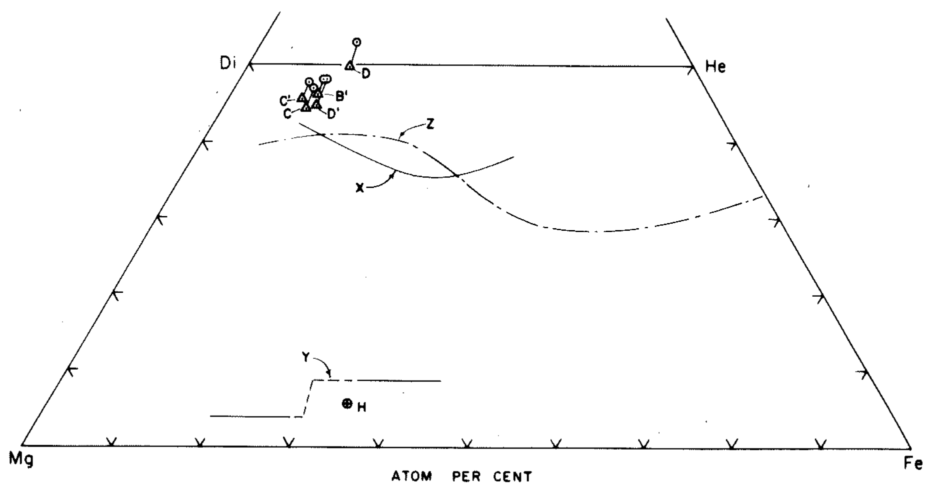
The clinopyroxene in the hornblende pyroxenite is greenish gray but is slightly darker than that in the olivine-bearing ultramafic rocks. In thin section, its color is pale green. The crystals are subhedral prisms, 2-10 mm in length. Pyroxene is probably the earliest mineral to crystallize, as all the others surround it or are interstitial to it. Some of the pyroxene crystals have central zones charged with numerous tiny oriented inclusions of opaque material. The inclusions have not been identified but are probably magnetite or ilmenite and may be the result of exsolution. Not all the pyroxenes show these inclusions, but a systematic geologic or areal distribution has not been recognized. The range of the intermediate index of refraction is 1.690-1.709, and as shown in figure 20, this is significantly higher than that of pyroxene from the olivine-bearing units. The analysed pyroxenes have the following properties:

Specimen	Composition	n_y	2V
I-31-3A	$Ca_{53}Mg_{36}Fe_{11}$	1.696	57°
H-19-3	$Ca_{48}Mg_{42}Fe_{10}$	1.698	--

For comparison, all the analysed pyroxenes have been plotted on the standard Di-He-Mg-Fe quadrilateral diagram for pyroxenes in figure 22. The increase in Fe content of the pyroxene in the hornblende pyroxenite over the other types is only 1 or 2 per cent, and is about 8 per cent less than is expected from optical properties by the charts of Hess (1949). Evidently some other element or elements are affecting the optical properties. The Al_2O_3 content of the hornblende pyroxenite pyroxene is almost twice that of the other pyroxenes and of the pyroxenes used by Hess in compiling the optical charts, and this may be responsible for the high indices. However, according to Segnit (1953, p. 219), alumina decreases the indices of diopside. The reason for the discrepancy is not known.

One of the analyses of pyroxene from hornblende pyroxenite plots on the Ca side of the Di-He join. This cannot be due to contamination because hornblende, the only other silicate mineral in the rock, has a much lower ratio of Ca to Mg + Fe than clinopyroxene. The analysis may be in error because only two of the six analyses of Duke Island pyroxenes meet the requirement recommended by Hess (1949, p. 626) that the two groups of ions, Z and W + X + Y (table 8), should each sum to 2.00 ± 0.02, and the Ca:Mg ratio of pyroxene I-31-3 is suspiciously high compared to the other clinopyroxene analyses. However, the total rock analysis for specimen I-31-3 also is rich in calcium and has larnite in the norm. If the amount of Ca in the hypothetical pyroxene molecule $CaAl(Al,Si)_2O_6$, is subtracted and the atomic proportions of Ca, Mg and Fe recalculated, the pyroxene falls inside the quadrilateral. These recalculated proportions have been plotted in figure 22 for all the clinopyroxene analyses.

It is noteworthy that the range of variation in composition of the ultramafic pyroxenes is small in figure 22 and that the increase in calcium is as marked as the increase in iron. This same trend is shown by clinopyroxene in the Union Bay complex (Ruckmick, 1957). It is unusual and, for comparison, the trends of variation of clinopyroxenes are shown for the early and middle stages of the Skaergaard intrusion as determined by Brown (1957, fig. 3) and for common mafic magmas as determined by Hess (1941, fig. 10). The curves of Brown and Hess are for clinopyroxenes coexisting with orthopyroxenes and should represent the approximate limit of solid solution of orthopyroxene in



⊙ Clinopyroxene: total Ca:Mg:(Fe''+Fe'''+Mn).

▲ Clinopyroxene: as above, less Ca as hypothetical $CaAl(Si,Al)_2O_6$.

⊕ Orthopyroxene.

B' Pyroxene vein in peridotite.
C Olivine pyroxenite.
C' Olivine pyroxenite.
D Hornblende pyroxenite.
D' Hornblende pyroxenite.
H Norite

X Trend of Ca-rich pyroxene in the Skaergaard intrusion. After Brown, 1957.
Y Trend of Ca-poor pyroxene in the Skaergaard intrusion. After Brown, 1957.
Z Trend of clinopyroxenes in common mafic magmas. After Hess, 1941.

Figure 22.- Plot of Duke Island pyroxenes on a Di-He-Mg-Fe diagram.

clinopyroxene for the conditions existing during the crystallization of the magmas involved. That the ultramafic pyroxenes lie on the calcium side of these curves is compatible with the absence of orthopyroxene, indicating that orthopyroxene had not reached its saturation limit in either the clinopyroxene or the magma from which the rock formed.

Hornblende in the hornblende pyroxenites appears black in hand specimen and under the microscope is medium gray-green to yellow-green with weak pleochoism and weak to moderate absorption. It is not uralitic in habit but occurs as well-formed crystals interstitial to or including pyroxene. In some of the rocks where it is exceptionally abundant, it forms early-looking prisms. A chemical analysis is not available for this mineral, but the composition probably is similar to that of hornblende from the hornblendite (table 8, section 2) because the hornblendite is transitional to hornblende pyroxenite, and the two minerals are very similar optically.

Magnetite is prevalent in the hornblende pyroxenite, almost to the point of being an essential mineral. It shows remarkably constant concentration at about 10-20 per cent although this is not adequately illustrated by the modes tabulated in table 7, section 5. Characteristically the magnetite is disseminated through the rock as 0.5-3.0 mm masses which stand out in distinct relief on the weathered surface. In thin section, magnetite commonly is interstitial to pyroxene but may show subhedral crystal form against hornblende. Ilmenite is invariably present as discrete grains and in places occurs as tiny exsolution blades along the octahedral planes of the magnetite.

A dark green spinel is commonly associated with magnetite, and although not identifiable in polished sections, is believed to be the material forming the most prevalent exsolution lamallae in magnetite. Most of it contains a considerable percentage of magnetite dust, and in a few thin sections it appears to have slight zoning in the amount of included magnetite, the outer rim being clear. The index of refraction of the spinel varies appreciably, even in one specimen, but is about 1.75-1.76. According to data given in Winchell and Winchell (1951, p. 82), this indicates a composition intermediate between true spinel and hercynite. Chromium, which causes a rapid increase of index to values in excess of 1.80, probably is not present in an appreciable concentration. This conclusion is supported by a spectrochemical analysis of a typical magnetite-bearing hornblende pyroxenite which contains about ½ per cent modal spinel and shows only 80 ppm chromium (table 11, entry D').

Plagioclase generally is absent from hornblende pyroxenite. The only locality where it occurs over an appreciable area is the triangular bulge in the hornblende pyroxenite zone on the southeast side of the Hall Cove ultramafic area. Even here, it does not make up much more than 5 per cent of the rock and, actually, what was called plagioclase in the field has proved under the microscope to be entirely clinozoisite and prehnite (?). However, these probably are alteration products of plagioclase.

Rare specks of sulphides are widely disseminated through the average hornblende pyroxenite and commonly impart a rusty appearance to the weathered rock. Pyrrhotite is the most prevalent mineral, but pyrite, pentlandite, and chalcopyrite have been observed.

Specimen I-31-3, for which a chemical analysis is given in table 8, section 1, was selected because it was fresher than most of the specimens collected from the hornblende pyroxenite unit. However, it comes from near the contact with olivine pyroxenite and has a lower than average magnetite content. Several thousand feet of drill core, broken in 10-foot sections, from the hornblende pyroxenite have been analysed for "soluble iron" (i.e. iron in oxide minerals soluble in hydrochloric acid) and TiO_2. Histograms of the results of these analyses are plotted in figure 23. They show a marked concentration of soluble iron at 11-12 per cent and of TiO_2 around 1.4 per cent. If it is assumed that about half the TiO_2 is in pyroxene and hornblende as indicated by the analyses of these minerals, then the common hornblende pyroxenite has about

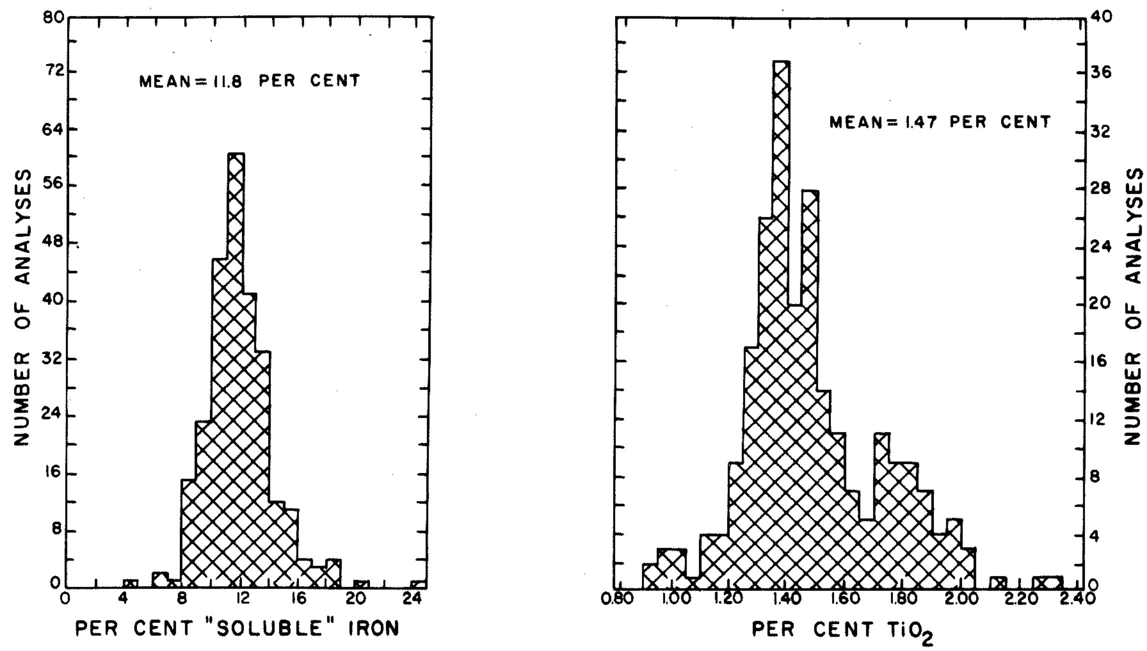


Figure 23.- Results of analyses of drill core from the hornblende pyroxenite zone.

$1\frac{1}{2}$ per cent ilmenite and 15-16 per cent magnetite. A rough calculated analysis of hornblende pyroxenite is given in table 10 together with its norm. It was assumed in the calculation that the rock was composed of 16 per cent magnetite, 1.5 per cent ilmenite, 0.5 per cent hercynite, 20 per cent hornblende (sp. I-31-4), and 62 per cent pyroxene (sp. H-19-3). An analysis of magnetite-bearing pyroxenite from Union Bay (Ruckmick, 1957) is given for comparison.

Hornblendite

Hornblendite occurs only in small bodies or as small zones in the larger rock units. Probably much of it is a late magmatic alteration of hornblende pyroxenite and olivine pyroxenite, but some appears to form primary segregations. Hornblende is the only major mineral. Magnetite and ilmenite are common accessories, but the amounts are much less than in the hornblende pyroxenite. Hornblendite in the south part of the Judd Harbor ultramafic area is commonly associated with basic pegmatite and in places contains small percentages of calcic plagioclase and clinozoisite. Elsewhere, it is essentially free of leucocratic minerals. Pyroxene is rarely present and then only as ragged relict grains.

The grain size of the hornblendite differs greatly. The most prevalent material has 5-20 mm crystals, but

TABLE 10. COMPOSITION OF MAGNETITE-BEARING HORNBLENDE PYROXENITE

	I	II
SiO_2	38.26	37.54
Al_2O_3	7.19	5.35
Fe_2O_3	13.29	15.52
FeO	9.66	9.64
MgO	11.67	11.85
CaO	16.70	17.40
Na_2O	0.86	tr
K_2O	0.18	tr
H_2O+	0.24	0.28
H_2O-	--	0.32
TiO_2	1.78	2.23
MnO	0.06	0.16
	99.89	100.29

Norm of I		
Mt		19.29
Il		3.35
Lc		0.85
Ne		3.99
An		15.36
He	4.73	46.25
Di	41.52	
Fa	1.12	8.05
Fo	6.93	
Larnite		2.85

I Calculated analysis for typical magnetite-bearing hornblende pyroxenite in the Duke Island ultramafic rocks. The methods of calculations are given in the text.

II Analysis of magnetite-bearing pyroxenite from the Union Bay ultramafic complex. The analysis is from Ruckmick, 1957, specimen 32a.

hornblendite facies of the hornblende pyroxenite zone may be as fine as 1-2 mm. Coarse pegmatitic veins or segregations with hornblende crystals more than 6 inches long occur locally (fig. 24), and crystals 4-7 feet in length were observed in a pod in the ultramafic body cropping out on the north shore of Duke Island, opposite Vegas Islands.

An analysis of hornblende from hornblendite is given in table 8, section 2, specimen I-31-4. As this is almost the total rock a norm has been calculated for the mineral and is as follows:

Mt		6.07
Il		2.75
Lc		3.00
Ne		10.02
An		31.99
He	3.50	23.16
Di	19.66	
Fa	4.22	22.38
Fo	18.16	
Larnite		0.61
		99.98

Thus hornblendite is a very much undersaturated rock in terms of its normative minerals.

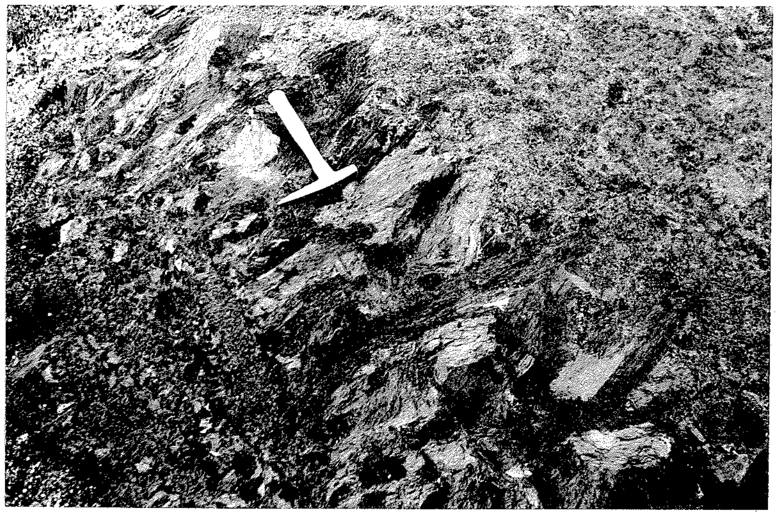


Figure 24. Pegmatitic vein or segregation of coarse hornblende in finer grained hornblendite. The locality is the north shore of Kelp Island.

TRACE ELEMENT DISTRIBUTION

The method of analysis used in making a preliminary investigation of the trace element distribution in the ultramafic and gabbroic rocks is described on page 7. The results are given in table 11. The sensitivity values listed in the table are the average figures for the spectrographic laboratory of the California Institute of Technology. Each sample was exposed in duplicate, and the average is regarded as one analysis. Absolute values are probably only good to about 50 per cent of the amount indicated, but all results are from 2 plates, exposed and developed on the same day, and the relative values should be reliable.

The data are plotted in figure 25. The diagram is like that used by Nockholds and Mitchell (1946, fig. 9) and Wager and Mitchell (1950, fig. 5). Every reason for the sequence in which the analyses are plotted cannot be given at this point, and in fact, a completely logical sequence probably does not exist. Hence "trends" or apparent inconsistencies therein do not necessarily have genetic significance. The diagram is primarily a means of comparison. The basic subdivision is into ultramafic and feldspathic types. The ultramafic rocks are arranged in an order paralleling Bowen's reaction series. The second olivine pyroxenite is from higher in the layered rocks, but as will be seen later this may not have any significance. Basic pegmatite is placed next because it has much in common

TABLE 11. TRACE ELEMENT DATA. CONCENTRATIONS ARE IN PARTS PER MILLION.

Section 1. Rocks

Entry		A	B	C	C'	D	D'	E	F	G	H	I
Specimen		H-4-4	I-40-2	I-37-2	R-38-2	I-31-3	H-19-3	I-31-4	R-37-1A	N-39-6	N-36-8	N-25-1
Rock		du	pd	opx	opx	hpx	hpxm	hb	hp	hgb	pxgb	ngb
Element	Sensitivity											
Ag	0.5	0.1	0.5	0.4	0.4	0.4	1.0	0.4	0.2	0.4	0.5	0.4
Ba	20	tr	tr	tr	tr	(10)	21	105	77	88	65	(g)
Co	2	125	240	120	70	60	53	88	46	138	89	59
Cr	5	1350	820	1500	925	670	77	60	43	258	135	140
Cu	1	8	4	45	245	300	115	24	2	42	120	33
Ga	5	--	--	--	--	--	8	tr	7	11	12	7
Mn	2	1175	1650	1300	860	940	1175	940	800	1375	895	1225
Ni	2	535	810	460	280	115	45	235	19	34	100	60
Sc	2	4	8	64	140	180	93	100	12	40	74	17
Sr	2	3	--	35	32	42	85	135	510	525	195	300
Ti	2	75	90	1150	2475	3150	>10000	4700	1925	2525	8300	785
V	2	44	38	180	360	615	1140	760	435	368	1180	280
Y	10	--	--	--	tr	tr	10	10	tr	10	tr	tr
Yb	1	1.5	2.0	1.5	1.5	2.0	4.0	3.5	2.5	2.0	4.5	1.5
Zn	80	(38)	(65)	tr	--	tr	tr	(12)	--	tr	(57)	tr
Zr	2	tr	6	8	12	18	28	23	13	12	43	10

The following elements were sought but are not present in amounts greater than the sensitivity values shown in parentheses: As(200); B(10); Be(1); Cd(10); Ge(10); La(80); Mo(3); Nb(15); Pb(10); Pt(100); Sb(200); Sn(5); Ta(200); Th(200); and U(500).

Section 2. Olivine

Entry		A	B	C	C'
Specimen		H-4-4	I-40-2	I-37-2	R-38-2
Rock		du	pd	opx	opx
Element	Sensitivity				
Ag	0.5	0.5	0.5	0.5	0.5
Ba	20	tr	tr	tr	tr
Co	2	300	285	260	270
Cr	5	140	215	35	45
Cu	1	175	5	200	80
Ga	5	--	--	--	--
Mn	2	1700	1850	1825	1575
Ni	2	1000	835	840	745
Sc	2	4	4	2?	5
Sr	2	7	7	9	8
Ti	2	15	8	--	30
V	2	18	25	15	25
Y	10	tr?	tr	--	tr?
Yb	1	4	4	4	4
Zn	80	145	95	115	110
Zr	2	5	5	tr	<5

TABLE 11. Continued

Section 3. Pyroxene

Entry		Clinopyroxene						Orthopyroxene
Entry		B'	C	C'	D	D'	H	I
Specimen		I-40-1	I-37-2	R-38-2	I-31-3	H-19-3	N-36-8	N-25-1
Rock symbol		vn in pd	opx	opx	hpx	hpxm	pxgb	hgb
Element	Sensitivity							
Ag	0.5	(0.3)	(0.3)	(0.2)	(0.3)	(0.3)	0.8	1.2
Ba	20	tr	tr	tr	tr	tr	tr	tr
Co	2	33	32	33	40	34	92	125
Cr	5	1125	2625	1200	580	14	33	123
Cu	1	6	32	85	270	70	7450	27
Ga	5	--	--	--	tr?	70	--	--
Mn	2	1030	835	790	975	895	1750	3200
Ni	2	113	140	130	62	13	110	110
Sc	2	140	130	160	175	210	135	17
Sr	2	35	40	40	25	55	14	--
Ti	2	1425	1450	2325	2425	3600	2675	275
V	2	378	355	440	515	560	805	200
Y	10	tr?	tr?	tr	tr	tr	26	--
Yb	1	1.5	1.2	1.5	1.8	1.9	5	4
Zn	80	--	--	--	--	--	--	143
Zr	2	13	10	15	15	20	37	16

Section 4. Hornblende

Entry		D	D'	E	F	F'	G
Specimen		I-31-3	H-19-3	I-31-4	R-37-1A	I-27-1	N-39-6
Rock		hpx	hpxm	hb	hp	hp	hgb
Element	Sensitivity						
Ag	0.5	(0.3)	(0.4)	(0.4)	0.8	0.8	0.6
Ba	20	115	180	105	tr	153	55
Co	2	58	56	88	69	50	48
Cr	5	525	35	60	225	3	48
Cu	1	205	200	24	8	5	10
Ga	5	2	3.5	tr	2	7.5	6
Mn	2	770	625	940	1600	1925	1625
Ni	2	205	24	235	63	24	43
Sc	2	180	180	100	86	62	62
Sr	2	165	315	135	33	130	75
Ti	2	3900	5900	4700	4000	6000	3050
V	2	695	835	760	695	630	425
Y	10	10	10	10	16	35	13
Yb	1	2.5	3.5	3.5	4	5	3
Zn	80	tr	tr	tr	tr	(64)	tr
Zr	2	15	23	23	32	62	18

TABLE 11. Continued

Section 5. Plagioclase

Entry		F	G	H	I
Specimen		R-37-1A	N-39-6	N-36-8	N-25-1
Rock		hp	hgb	pxgb	ngb
Element	Sensitivity				
Ag	5	--	--	--	--
Ba	20	84	17	153	15
Co	2	--	4	3	--
Cr	5	--	11	--	--
Cu	1	2	7	7	5
Ga	5	10	14	19	16
Mn	2	16	54	22	20
Ni	2	tr	2	4	1
Sc	2	--	--	--	--
Sr	2	1025	1300	495	540
Ti	2	tr	105	150	--
V	2	42	40	50	44
Y	10	--	--	--	--
Yb	1	--	--	--	--
Zn	80	--	--	--	--
Zr	2	--	--	--	--

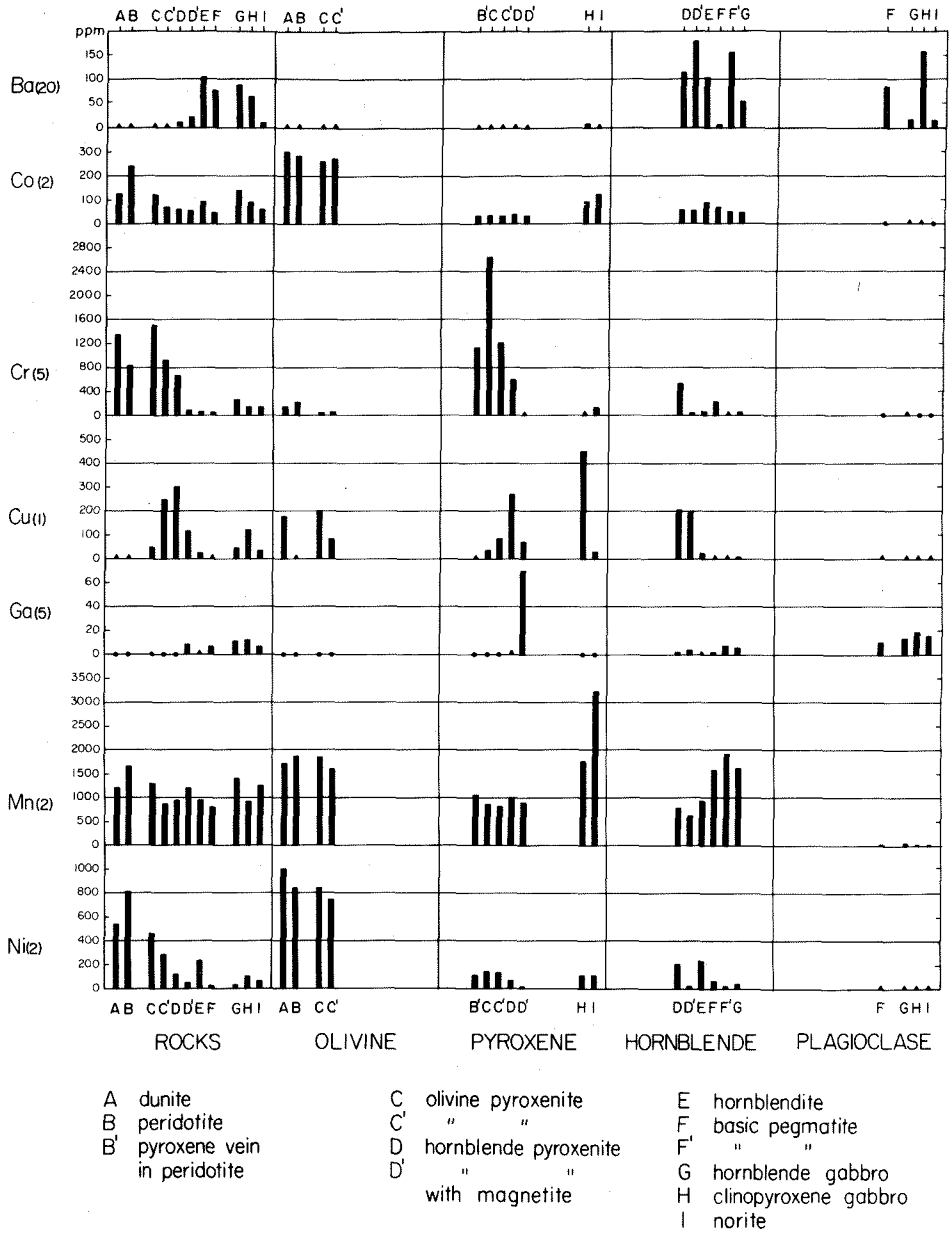


Figure 25. Plot of the trace element data for the mafic and ultramafic rocks of Duke Island and their constituent minerals. Concentrations are given in parts per million. The numbers in parentheses are the sensitivity values of the analytical method for that particular element. A solid triangle indicates that the element was detected in trace amounts. A dot indicates that the element was not detected.

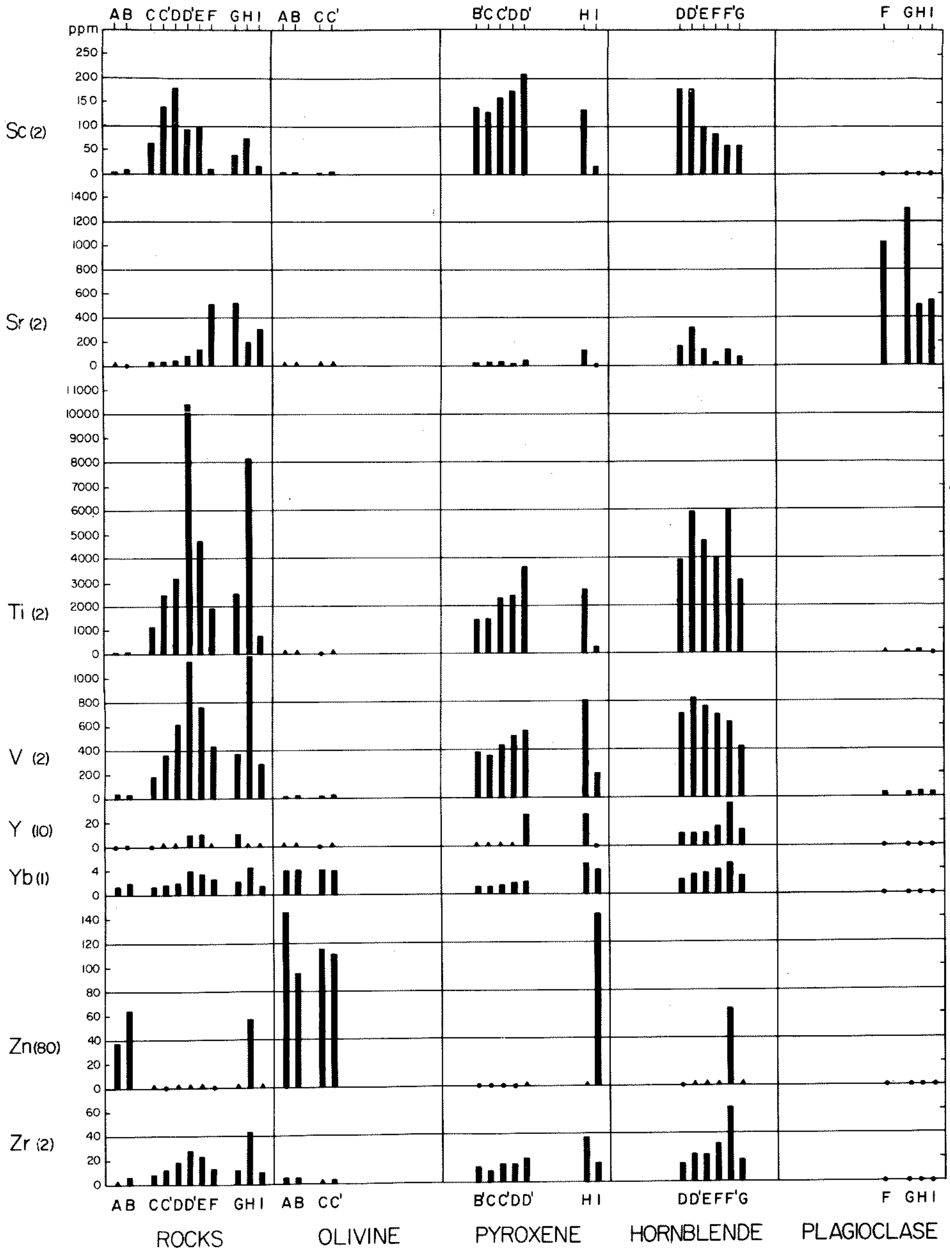


Figure 25. Continued.

with hornblendite. Hornblende gabbro follows because of its association with basic pegmatite. Chemically, the clinopyroxene gabbro is more like hornblende gabbro than is the norite, hence norite is placed last.

The trace element content of the various rocks and minerals is similar in general to that found in most previous studies of mafic and ultramafic igneous rocks (e.g. Wager and Mitchell, 1950; Ross, Foster, and Myers, 1954). Extensive discussions of the principles determining minor element distribution are given in the references cited, hence only brief comment will be made on the major features shown by the analyses.

Silver is uniformly distributed in all rocks and minerals in small, constant quantities.

Barium is concentrated in hornblende relative to the other mafic materials, and in two of the four analysed plagioclase samples.

Cobalt is markedly concentrated in olivine and shows a slight preference for hornblende as compared to pyroxene. Materials balance requires that appreciable cobalt be in magnetite and ilmenite, a commonly observed feature (Wager and Mitchell, 1950, p. 158).

Chromium is relatively abundant in the olivine-bearing rocks but does not concentrate in olivine. In dunite and peridotite, it must occur primarily in chromite, substantiating the petrographic observations (p. 57), but in the pyroxenic rocks, it is almost entirely in pyroxene. The low amount of chromium in the magnetite-rich hornblende pyroxenite is notable.

Copper, scandium, titanium, vanadium, ytterbium, yttrium, and zirconium show remarkable parallelism in their distributions. In the column of rocks in figure 25, each shows two "peaks," one at the magnetite-bearing hornblende pyroxenite, and the other at the clinopyroxene gabbro. All the elements except copper are rare in olivine and show parallel distributions in pyroxene. In hornblende they seem to form two groups.

Parallel distributions for these elements are not apparent in the data of Nockholds and Mitchell (1946, fig. 9) or Wager and Mitchell (1950, fig. 5). As the rocks studied by these authors come from well-substantiated examples of series developed by crystallization differentiation, the data presented here might be construed to mean that the Duke Island rocks are not a differentiation series. However the rocks richest in this group of elements also have the most magnetite and, probably, the highest ratio of ferric to ferrous iron. It is well established that the oxidation state of iron greatly influences the course of igneous differentiation. Thus, the minor element data may be a reflection of a fundamental control in igneous rock variation that was different for the Duke Island rocks as compared to the previously studied rock series.

A reason undoubtedly important for the parallelism of these elements is the simplicity of mineralogy in the ultramafic rocks. In dunite and olivine pyroxenite, sites into which the trace elements can enter are few. With the appearance of more phases, the possibility of diversity increases. This is probably the reason for the development of two groups in the hornblende. Again, the data need not be inconsistent with differentiation.

<u>Manganese</u> is relatively constant throughout the rocks. It shows some concentration in olivine and in the gabbro pyroxene compared to the ultramafic pyroxene.

<u>Nickel</u>, like cobalt, is strongly concentrated in olivine and is slightly more abundant in hornblende than in clinopyroxene.

<u>Strontium</u> shows marked preference for plagioclase

and a lesser preference for hornblende.

Zinc is most abundant in olivine and the orthopyroxene.

LAYERING IN THE ULTRAMAFIC ROCKS

GENERAL REMARKS AND NOMENCLATURE

The feature which distinguishes the Duke Island ultramafic rocks from those in other parts of southeastern Alaska is the abundance of stratification closely resembling the bedding of sandstones. An example is shown in figure 26. Igneous stratification of this sort has been given many names in the geological literature, but none of the described examples includes all the variants occurring in the Duke Island rocks. Here all stratification will be called layering, and appropriate modifiers will be applied where necessary. The principal types of layering are:

(1) Graded layering

(2) Non-graded layering

(3) Fragmental layers

They are all transitional. The term layered series is applied to those units containing layering.

Essentially all the layering occurs in the olivine-bearing rocks and involves only two minerals, clinopyroxene and olivine. At one locality on Kelp Island, magnetite bands or layers occur in hornblende pyroxenite, but it is questionable whether these are of the same origin. Not all the ultramafic rocks are layered; the distribution and attitude of the layering will be described for the individual areas.

Plates 1, 2, and 3 show the distribution of layers by trend lines. The spacing of lines roughly shows the amount of layering. The length of lines gives the lateral persistence of sections of layered rocks but does not necessarily show the continuity of individual strata. Topography influences the mapped trend of layering only in the vicinity of Knob Hill in the Hall Cove ultramafic area. Elsewhere, relief is low, and layers dip at relatively steep angles, hence topographic effects are negligible.

GRADED LAYERING

The most impressive graded layers are 2-10 inches thick (figures 26-29), and these make up the bulk of the volume of layered rocks. The thinnest layers are essentially one crystal diameter thick and these characteristically occur in groups to form a finely laminated rock that is interstratified with the thicker layers. The greatest thickness recognized for a continuously graded layer is about 25 feet, but the common maximum thickness is 2-4 feet.

The grain size of the typical graded layer is 4-10 mm at the base and becomes progressively finer upward until

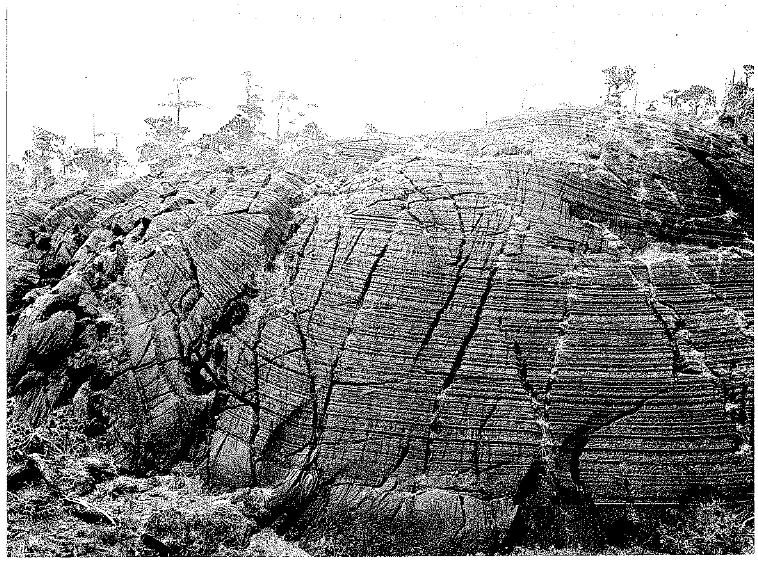


Figure 26. Well-developed graded layering in olivine-rich olivine pyroxenite. The thick layer is about 3 feet through. To the left the layers are draped over a large olivine pyroxenite block. Note that the irregularity due to the block disappears in about 10 feet because of thinning of the layers over it. The locality is close to the line of measurement of lithologic section B, near its base.

it is 0.2-2.0 mm at the top. The thickest layers generally are the coarsest. Ascending through a series of layers, the change from coarse to fine is gradational, whereas that from fine to coarse is abrupt. Layers without distinct grading commonly alternate with the graded layers, but outright reversals in the direction of grading are rare. On cursory examination, the layers appear to be extremely well sorted. Actually, a considerable spectrum of grain size is present at any level, although sorting does improve in the upper part of a layer. It has not been possible to obtain quantitative data on the sorting so that comparison with sandstones can be made. The rock cannot be disaggregated along grain boundaries for seiving, and measurements of grain diameters in thin section or on the etched weathered surfaces are unreliable. Thin sections, by their nature, do not give a representative picture of the sorting, and the measurements on weathered surfaces are biased towards the larger grain sizes. Probably the crystals underwent further growth from interstitial magma after deposition, and this would modify the initial grain size distribution.

Both pyroxene and olivine show sorting according to grain size. Inasmuch as they have almost the same densities but the average pyroxene is coarser, olivine is concentrated in the upper part of a layer. In classifying the layered rocks, the limits given in table 6 are applied to the total composition of a sequence of layers rather than to thin bands that have been enriched in one of the minerals because of their grain size. Obviously this procedure has many practical advantages in mapping, and it probably is more important for most theoretical considerations to know the composition of the total crystalline accumulate in any given locality than to know the degree of minor variation superimposed by grain-size sorting.

The character of the layers naturally differs somewhat with the rock type. Figure 27 shows layering in dunite almost free of pyroxene. The grading is slight and is evident only because of etching due to weathering. Examples of dunite layering are rare, but this may in part stem from the difficulty of recognizing grading in the commonly serpentinized dunite. Figure 28 shows the layering typical of peridotitic rock in which sorting by grain size commonly results in almost perfect mineralogical differentiation, the upper part of each stratum containing only a very minor amount of pyroxene. The olivine pyroxenite layers generally contain more pyroxene than olivine, even in their upper portions (fig. 29).

A surprising feature of the graded layers is the virtual absence of planar or linear orientation of pyroxene crystals. At only a few localities in peridotite has any suggestion of preferred orientation been observed, and the pyroxenes of the other rocks are invariably non-oriented.

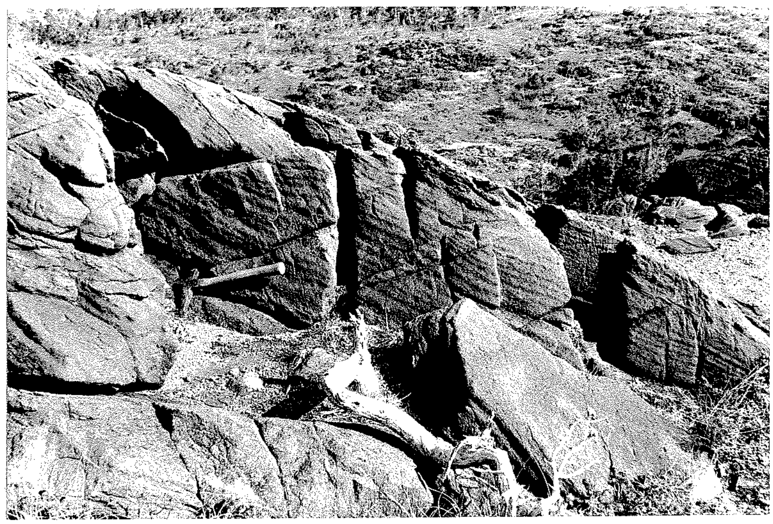


Figure 27. Layering in dunite. The variation in grain size is very slight. Pyroxene is present only as an interstitial phase. The locality is on the east side of the top of Knob Hill.

Figure 28. Graded layering in peridotite. A marked mineralogical sorting is shown, the upper part of the layers being almost entirely of olivine. The locality is the main zone of dunite and peridotite in the Judd Harbor ultramafic area.

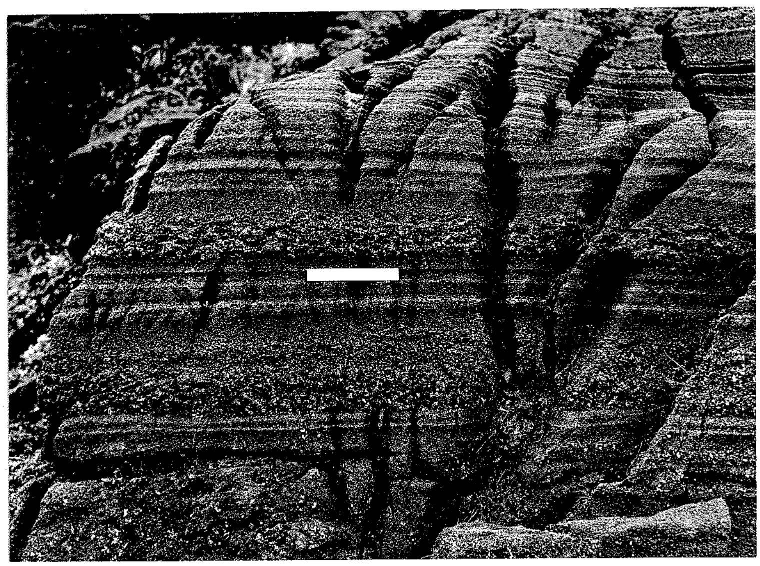


Figure 29. Graded layering in olivine pyroxenite. The coarser crystals standing out in relief are clinopyroxene. Olivine is more concentrated in the finer grain sizes. The scale is 6 inches. The locality is 600 feet east of Knob Hill.

This is a distinct contrast to the stratification in gabbros described of other areas, where plane-parallelism of plagioclase laths is the rule, and linear alignment is relatively common (Wager and Deer, 1939, pp. 38-50; Grout, 1918, p. 446). Evidently, the combination of stubby pyroxene prisms and spheroidal olivine crystals in the ultramafic rocks arranges best as a massive rock.

The lateral continuity of layers differs considerably. Sequences of 4 or 5 layers extending only a few feet are not uncommon, but more characteristic are strata traceable for several tens of feet. The maximum distance that a single stratum has been followed is about 250-300 feet, but undoubtedly many could be traced even farther if it were not for breaks in exposure, structural complications, and lack of individuality among layers. It is very unusual to see a single horizon pinch out, fade out, or terminate for reasons other than structural complications. Generally, not one stratum but a whole section of layers ends. Some of the sections are laterally continuous for 900 feet and one might deduce that some individual layers also persist this far.

The manner in which layers terminate is of interest. Figure 30 shows a single layer splitting into two layers with a total thickness equal to that of the original. Insofar as the author knows, this example is unique for thick, well-developed layers in an otherwise laterally continuous

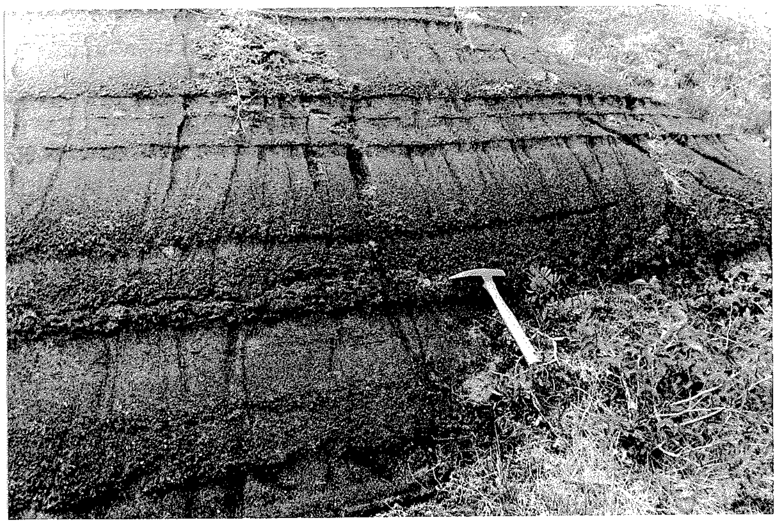


Figure 30. Section of coarse, thick layering showing one graded layer splitting into two. The locality is the peridotite zone of the Hall Cove ultramafic area, about 1000 feet west of Knob Hill.

section. More typically, an appreciable section of layers becomes less and less distinct over a distance of a few tens of feet and finally disappears. Where it is fading out, the layering closely resembles the non-graded layering.

Doubtless the grading by grain size in the layers is a result of gravitational sorting of crystals settling from a magma. It has therefore been used during mapping to determine direction of tops of layers just as graded bedding is used in sedimentary rocks. Comparable layering in gabbroic rocks has been put to similar use (Peoples, 1936, p. 358), but the grading in the gabbro layers is one of distribution of minerals according to their density rather than grain size, the pyroxenes being concentrated in the lower part of each layer and the feldspar at the top. Stoke's law, which gives the terminal settling velocity of spherical particles in a viscous medium, states:

$$v = \frac{2}{9} r^2 g \frac{(\rho_1 - \rho_2)}{\eta}$$

where

ρ_1 and ρ_2 are the densities of particles and liquid, respectively

η is the viscosity of the fluid

g is the acceleration due to gravity

r is the radius of the spheres.

Thus it is not surprising that both density and particle size should be effective in the production of grading by gravity sorting. For determining tops the difference in the grading of the gabbro and ultramafic layers is unimportant.

In many of the photographs of layered rocks, the strata have steep to vertical dips. It will be shown that considerable post-depositional folding and deformation of the rocks have occurred; hence horizontality cannot be used as a criterion of gravity accumulation. At present, there seems to be no way in which to estimate the original attitude of the layering. Wager and Deer (1939, fig. 14) indicate that the rhythmic layering of the Skaergaard intrusion had original dips as steep as 20°, and Carr (1954) considers dips of 35°-50° in the banded gabbros of the Isle of Skye to be primary.

NON-GRADED LAYERING

Non-graded layering is the name applied to appreciable thicknesses of layered rocks in which graded layers are absent. It is not meant to refer to the thin laminations that commonly alternate with the better developed graded layers. The non-graded layering is not nearly as spectacular as the graded layering, nor is it as clearly the result of gravity settling of crystals, although this is believed to be its origin. Examples are shown in figures 31 and 32. The layering is primarily by thin alternations of pyroxene-rich and olivine-rich materials, or by discontinuous parallel bands of almost pure olivine. Pyroxene

Figure 31. Non-graded layering in the olivine pyroxenite of the Judd Harbor ultramafic area. The layering forms part of the large fold pattern mapped in the area and for that reason is nearly vertical. The locality is near the top of the most southerly of the main hills within the area.

Figure 32. Non-graded layering in olivine pyroxenite in the Judd Harbor ultramafic area. Note the segregation of coarse pyroxene in the dunite band beneath the geological hammer.

is everywhere the coarser, but grain-size grading is rare. The thickness of the layers seldom exceeds 4-5 inches. Individual bands persist for distances ranging from a foot or two to several tens of feet, but seldom farther. However, sections formed by the layering are remarkably continuous; in the Judd Harbor ultramafic area, non-graded layering forms a section several thousand feet in thickness that extends laterally for over a mile. Where the layers are deformed, they are folded in a systematic fashion, small folds being in harmony with the large ones.

The graded and non-graded layers fade into one another both laterally and vertically, and in a few places they form alternate sections. Mapping shows that the two are everywhere structurally conformable, and the same fundamental origin is implied. It does not seem unreasonable to expect that, just as graded and non-graded beds occur in sediments, graded and non-graded layers could exist in igneous rocks and be formed by essentially the same process.

FRAGMENTAL LAYERS

The fragmental layers are analogous to conglomerate beds in much the same way that graded layering is analogous to graded bedding in sandstones. The fragmental layers consist of fragments of olivine pyroxenite set in a matrix that is, for the most part, peridotitic. Most of the fragments are 2-10 inches in length, but blocks more than 6 feet long occur in some layers (figs. 33 and 34). The fragments are not appreciably rounded, and although corrasion or attrition by transport may have occurred, it cannot be proved. Fragments typically show planar orientation (fig. 35), but linear orientation, although sought for, has not been found.

The fragmental layers alternate with normal graded layers, grade into them, and commonly show vertical grading in particle size themselves (fig. 33). They are really only a variant of the normal graded layering, and the boundary between the two types is arbitrary. In the graded fragmental layers, a reliable visual appraisal of sorting can be made. Even in the better graded strata sorting is poor near the base and improves upward as the average particle size becomes finer. In the lower parts, blocks of olivine pyroxenite several feet in length occur together with large quantities of discrete crystals only a small fraction of an inch in diameter. The upper parts of many fragmental layers closely resemble normal graded layers.

Most of the fragmental layers are 1-10 feet thick. They are everywhere thicker than the normal graded layers in the same section, and generally,the layers with the larger fragments are the thicker. The lateral continuity of the layers is not great, and they thin conspicuously and undergo a change in facies to finer grain sizes. As will be shown, they occur in sections characterized by

Figure 33. Graded fragmental layers. The fragments are olivine pyroxenite, the finer-grained and lighter-colored material in the top parts of the layers and forming the matrix for the fragments is peridotitic. The scale is 6 inches. The locality is along lithologic section A, near its middle.

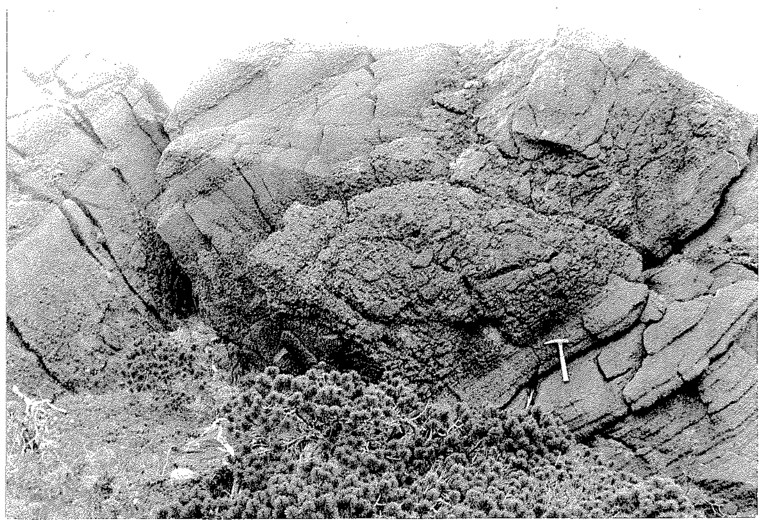


Figure 34. Graded fragmental layer. The locality is the same as figure 33. The olivine pyroxenite fragments are much larger and more poorly sorted than in figure 33. Note the 6-foot block in the upper part of the photograph.

Figure 35. Fragmental layer showing planar orientation of fragments. Peridotite zone, Hall Cove ultramafic area.

considerable depositional complication, and probably the conditions which provided the fragments were local and did not permit the layers to have extensive development. A few, however, persist for more than 200 feet, and clearly the fragmental layers are layers and not local piles of rubbly olivine pyroxenite blocks.

INTERPRECIPITATE MAGMA

Layering in the ultramafic rocks is attributed fundamentally to the accumulation of crystals by gravitational settling from a magma because many of the layers are graded in such a way that coarse grain sizes are consistently below and finer ones above. Wager and Deer (1939, p. 120) point out that crystals accumulating from a magma under the influence of gravity should resemble a loosely packed sandstone and would retain 10-30 per cent pore liquid or interprecipitate magma. They were unable to make an accurate measurement of the amount of interprecipitate liquid in the Skaergaard intrusion, on the basis of textural relationships, but succeeded in justifying an assumed value of 20 per cent as a first approximation of an average figure.

Interprecipitate magma must have existed in the Duke Island layered rocks, but evidence bearing on its quantity and composition is limited. Two major phases, clinopyroxene and olivine, persist from bottom to top in the exposed layered series. Generally, their primary and interprecipitate fractions cannot be distinguished by textural relationships, certainly not to the point of making reliable quantitative estimates. The mafic silicate minerals show so little variation in composition that zonal overgrowths, expectable products of interprecipitate magma (Wager and Deer, 1939, p. 121), are not discernible. Primary magnetite is a common but quantitatively minor accessory and probably formed after the primary minerals accumulated in layers. Hornblende, though generally present as an interstitial phase, does not appear as a primary precipitate at any definite horizon, and its concentration is not obviously related to elevation in the layered series. However, the widespread occurrence of hornblende probably is good evidence of a water-bearing magma, and interprecipitate magnetite indicates an appreciable iron content, but a more definite statement cannot be made at the present time.

Vertical grain-size grading apparently is not present in the rhythmic layering at the Skaergaard intrusion. Presumably, each mineral in the rock has a limited spectrum of sizes. At Duke Island, crystal size varies considerably. The massive rocks are like poorly sorted sediments, and even the graded layers are not perfectly sorted. Because of the poor sorting, the percentage of interprecipitate magma might be expected to be appreciably less than in the

Skaergaard intrusion. On the other hand, plagioclase in the Skaergaard rocks is tabular, and the rhythmic layering commonly shows igneous lamination by planar and linear alignment of minerals. This packing system might allow less interstitial liquid than the non-oriented accumulations of Duke Island. Thus effects of sorting and orientation probably cancel, and major differences in the pore-liquid content of the two types of deposit are not expected.

In discussions to follow, frequent mention will be made of interprecipitate magma and to features attributable to its effects.

HALL COVE ULTRAMAFIC AREA

GENERAL FEATURES

The Hall Cove ultramafic area has a distinctive lobate configuration (plate 2). The predominant rock type is olivine pyroxenite. Hornblende pyroxenite forms an almost complete peripheral rim, and olivine pyroxenite surrounds a small east-trending zone of peridotite thus giving a crude development of the concentric zoning typical of several of the ultramafic bodies in southeastern Alaska.

The rocks in the area apparently are cut by a major northerly trending fault whose trace is in part occupied by Hall Cove.

The main layered structures occur in the olivine-bearing rocks west of the fault. For the most part, they trend east and dip moderately or steeply south. Tops of layers are to the south. Much of the layering appears to be at a considerable angle to the hornblende pyroxenite border zone and the outer boundary of the ultramafic area. Proceeding up through the layered series, or proceeding from north to south, the sequence of rocks is: olivine pyroxenite; peridotite; and olivine pyroxenite. The northern or lower boundary of the peridotite zone is profoundly disconformable with the layering and apparently is an intrusive contact. Many blocks of olivine pyroxenite are included in the peridotite. The southern or upper contact of the peridotite zone is indefinite and may be either intrusive or gradational and conformable with the layering. If it is conformable, the southern olivine pyroxenite zone probably is younger than the peridotite whereas the northern olivine pyroxenite zone definitely is older. Therefore although the area is zoned in a physical sense, the olivine pyroxenite parts may be of two distinct ages.

Large-scale folding occurs in the southern olivine pyroxenite zone, and several faults have been mapped.

A small isolated area of ultramafic rocks in the vicinity of North Hill shows features which suggest that it is part of the outcrop of the main body.

The small amount of layering east of the major fault is erratically folded and bears no obvious relation to the main layered series.

HALL COVE FAULT

The major lineament on the aerial photographs of Duke Island is marked by Hall Cove and the shallow valley that projects on from the northern end of the cove. The lineament trends about N.30°E. The contours of the aeromagnetic map (plate 7) are deflected notably along this line, and structures in the layered ultramafic rocks on opposite sides fail to match. A major fault is postulated, hereafter referred to as the Hall Cove fault. The dip of the fault is unknown but probably steep. The east block is believed to be relatively depressed by several thousand feet of dip-slip movement, although the boundaries of the ultramafic area apparently are not offset appreciably. Evidence for this displacement will be given in a later section.

HORNBLENDE PYROXENITE ZONE

The hornblende pyroxenite border zone in the Hall Cove ultramafic area is continuous, and the few breaks indicated on the map (plate 2) probably are due to lack of exposure. The width of the zone ranges from only a few feet to about 1500 feet. In one or two places at the northwest corner of the area, the border rock is hornblende-olivine pyroxenite, but the olivine is partially replaced, chiefly by magnetite, and the hornblende and pyroxene are optically identical with those in hornblende pyroxenite.

Two apparent divergences of hornblende pyroxenite from its characteristic position as a peripheral rim are noted. West of the Hall Cove fault hornblende pyroxenite crops out as a small patch in the middle of the southern olivine pyroxenite zone. A strip of hornblende pyroxenite transects the main lobe of ultramafic rock east of the Hall Cove fault. Both the occurrences may reflect the position of the ultramafic boundary in the vertical dimension, the first from below and the second from above.

The hornblende pyroxenite has massive, uniform, medium-grained texture and relatively uniform composition. The rock disintegrates readily, as is typical of this unit, and as a result, the topographic elevation of the border zone around the south and west sides of the area is 50-75 feet lower than the adjacent olivine pyroxenite. The unit is virtually devoid of internal structure. In only one outcrop, the most southerly in the area, was any suggestion of layering recognized, a faint magnetite banding of unknown origin.

Along its outer boundary, the hornblende pyroxenite zone is adjacent to either gabbro or quartz diorite. The contact is exposed in only about a half-dozen outcrops, for but a few feet in each. In several places, it can be located within 10 feet but is not actually visible. There are no fine-grained facies of hornblende pyroxenite attributable to chilling.

The contact of hornblende pyroxenite with gabbro where exposed is abrupt and, locally, even knife-edge sharp. Generally, the rocks show slight effects of mutual alteration or hybridization. For a distance of a few feet from the contact the hornblende pyroxenite in places contains sparse plagioclase and clinozoisite, and in places small dikes or veins of gabbroic material are present. All the plagioclase is anorthite-rich, probably related to basic pegmatite; hence the dikes do not give relative ages of the major rock masses. The gabbro near the contact is commonly the hornblendic type with anorthitic plagioclase. Along the northwest boundary of the area, the only detectible alteration of the pyroxene gabbro is a clouding of the plagioclase by incipient saussuritization. This, however, may not be a contact effect.

The contact of the border zone with the quartz diorite is sharp where exposed, but the age relationships are indefinite. In some places adjacent to the contact the granitic rock contains more than typical amounts of hornblende and is cut by numerous epidote stringers. The hornblende pyroxenite appears to be normal. As discussed previously, evidence from other parts of Duke Island indicates that the granitic rock is the younger.

Along its inner boundary the hornblende pyroxenite zone is everywhere against olivine pyroxenite. The contact is gradational but can be defined to within a few feet. Figure 36 is a diagrammatic representation of the nature of the contact. As shown, in going from olivine pyroxenite to hornblende pyroxenite, the first indication of the approach to the contact is the sporadic appearance of hornblende in amounts up to 15 per cent. In the short distance in which the main transition takes place, hornblende and magnetite appear in abundance, and olivine disappears except for scattered grains. Accompanying this change is an increase in the indices of refraction of the pyroxene. Commonly the olivine pyroxenite has olivine-rich patches a few inches in size, and these patches persist into the hornblende pyroxenite, apparently as relicts, but become less and less distinct as they are gradually replaced by magnetite, spinel, hornblende, and possibly clinopyroxene. On plate 2, three hornblende-olivine pyroxenite areas of several thousand square feet within the hornblende pyroxenite zone appear to be unreplaced remnants of olivine pyroxenite. The contact of the hornblende pyroxenite and olivine pyroxenite seems to be petrologic rather than structural, as one mineral assemblage takes the place of another. The textural relations suggest that the hornblende pyroxenite mineral assemblage is the younger.

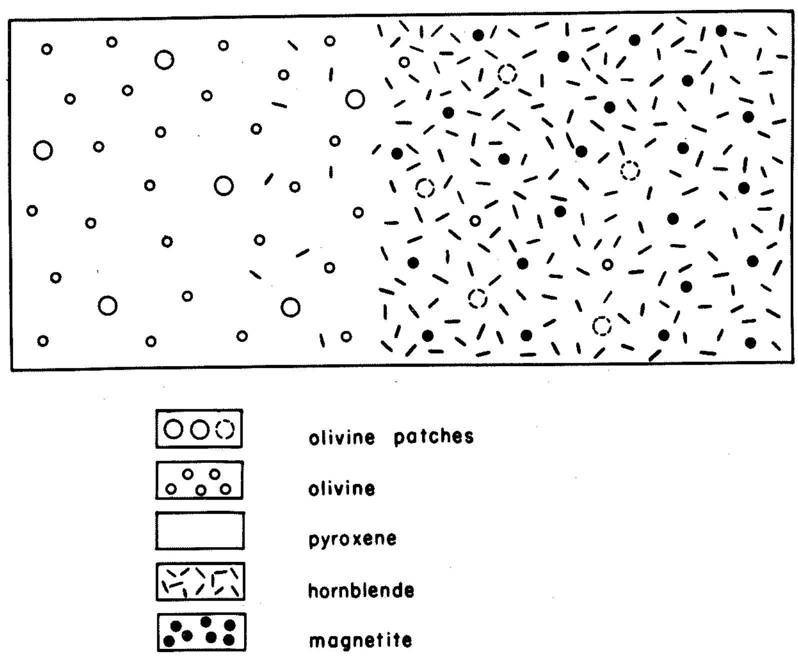


Figure 36.- Diagrammatic representation of a common relation along the contact of olivine pyroxenite with hornblende pyroxenite. The length of the diagram is a few tens of feet.

OLIVINE-BEARING ROCKS WEST OF THE HALL COVE FAULT

Northern Olivine Pyroxenite Zone

In the olivine-bearing ultramafic rocks west of the Hall Cove fault the layering dips consistently south or southwest, and tops of layers, as indicated by grading, face south. Thus the stratigraphically lowest rock unit is to the north; it consists of olivine pyroxenite. The outcrop of this unit is almost separated into two parts by a north-trending fault having 1200 feet of apparent right lateral movement.

Layering in the northern olivine pyroxenite zone strikes N.60°-70°E., approximately parallel to the boundary of the ultramafic area, and generally dips 60°-70°S. The contact of ultramafic and gabbroic rocks is exposed in only one outcrop where it can be measured and dips 60°S. However, the relief on the contact is only a few inches, and thus the layering and contact may or may not be parallel at this point.

Graded layers are present, but the most extensively developed layering is the type shown in figure 37. It has sharply defined alternate bands of pyroxene and olivine without obvious grading and is a fairly distinctive variant of the ultramafic layering. This layering is present on both sides of the fault and is a further indication that the blocks are correlative.

Fragmental layers have not been observed in the

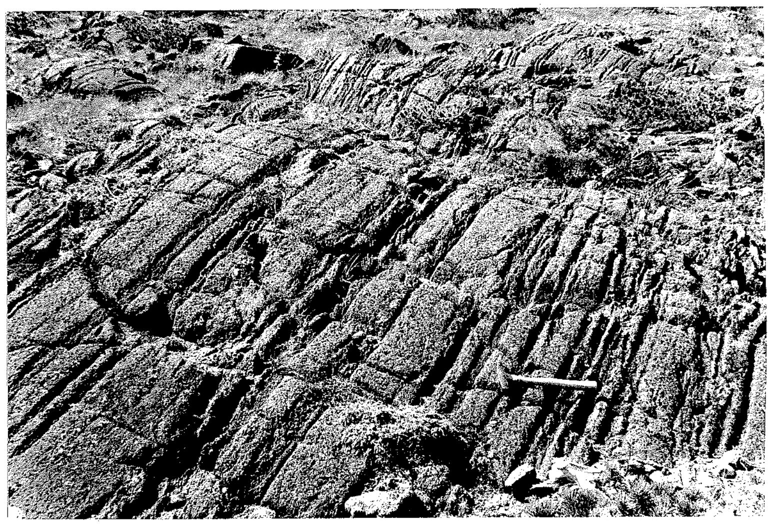


Figure 37. Layering in the northern olivine pyroxenite zone of the Hall Cove ultramafic area, east of Knob Hill. Olivine and pyroxene are largely segregated into discrete bands without appreciable grading.

zone, but in the upper parts, just east of Knob Hill, individual blocks of olivine pyroxenite are included in the layered series. Some of these blocks are layered (fig. 38) and are fragments of an earlier crystalline accumulate included in the later layered rocks. Figure 39 shows a structure occurring a few tens of feet south of or stratigraphically above the locality of figure 38. The highly deformed layering is best interpreted as a slump structure whose development was penecontemporaneous with deposition. These features are the lowermost indications of disturbed conditions of sedimentation in the main layered series of the Hall Cove ultramafic area. An abundance of comparable features occur in the peridotite zone.

Peridotite Zone

Northern Boundary

The contact of the peridotite with the northern olivine pyroxenite zone is a highly irregular surface that cuts sharply across the olivine pyroxenite layering. The peridotite layering strikes about east and dips 40^{o}-50^{o}S. or about 20^{o} flatter than the layering in the olivine pyroxenite. The contact is thus comparable to an angular unconformity in sedimentary rocks. The analogy to sediments can be carried one step farther, because innumerable blocks and fragments of different sizes of olivine pyroxenite are included in the lower peridotite

Figure 38. In the left part of the photograph is an angular block of layered olivine pyroxenite which has been overlain by a later generation of olivine pyroxenite layers. Subsequently, there has been tilting to the right. The scale is 6 inches.

Figure 39. Highly deformed layering in olivine pyroxenite probably resulting from slumping during accumulation. Note the alternation of deformed and relatively normal layers.

layers like a basal conglomerate. These structural relationships suggest that the northern boundary of the peridotite zone is an intrusive contact and that the peridotite is the crystalline precipitate of a younger magma deposited on or against the northern olivine pyroxenite zone. No chill zone is present along the contact. Many small bodies of dunitic material occur in the olivine pyroxenite immediately north of the contact; some of these are dikelike but most have highly irregular outlines. Their character and origin will be discussed later.

Blocks and Fragments

The peridotite zone of the Hall Cove ultramafic area forms an east-west strip about 6400 feet long and 3400 feet wide. It has extensively developed and highly regular layering but also has the most intricately complex structures of any of the Duke Island rocks. Two features are common: layering; and included blocks of olivine pyroxenite. The fragmental layers are a combination of these two features in which the blocks and fragments form layers. Essentially all the fragmental layers occur in the western part of the peridotite zone.

The blocks and fragments are irregular to angular and may be equant, rectangular, or elongated slabs (figs. 40-45). With the exception of about a dozen pieces of non-ultramafic material they consist solely of olivine pyroxenite. Most of the blocks are structureless, but many contain well-developed layering. Commonly they are disposed so that their layering is completely discordant with that of the surrounding peridotite (figs. 43 and 44).

The blocks shown in figures 40 to 45 are in the smaller size classes and range from aggregates of only a few crystals to bodies a couple of tens of feet in length. Truly large blocks are shown in generalized form in plate 2. They commonly are 100-200 feet to a side, and one may be 600 feet in length. Proof that such large masses are single blocks is given in plate 4, where several have been shown in detail. Outcrop areas have been mapped to show the extent of interpretation. In some places where contacts are covered, soil and broken rock have indicated their position, but much of the shape given to blocks in covered areas has simply been patterned after their exposed forms. It is clear that many of the blocks are angular; the "coarse-grained and uneven-textured" materials are the result of secondary processes to be discussed. The size range of the blocks is great, and several larger than 100 feet are shown. The peridotite layering is much deformed but, where most regular, has the general dip of the peridotite zone as a whole (40^{o}-50^{o}S.S.E.). In contrast, the layered olivine pyroxenite blocks, both large and small, have random orientation (note in particular the large block at locality A). Layering within the blocks is not significantly deformed, and the random attitudes cannot be explained by folding. The blocks must be separate fragments.

The relationship of peridotite layering to small fragments is well illustrated in figures 40, 41 and 43. The tendency for layers to rise over blocks is typical, regardless of their size, and hereafter when this relationship is referred to, the layers will be said to drape over the blocks. The layers underneath small blocks commonly are slightly depressed; this is probably due to both the weight of the block and to the differential compaction of the layers around the block. Intense deformation of layering occurs beneath and around the lower edges of the large blocks. This deformation is local, and the folding is not systematic (figs. 45-47, and plate 4). The deformation is interpreted to be the result of impact and subsequent settling of the blocks in the loosely consolidated peridotite layers.

The small individual blocks and fragments in the fragmental layers are pieces of olivine pyroxenite that were included in the magma at the time the layers were being deposited. A similar origin is implied for the large blocks because of their distribution (plate 2) and because they are both overlain and underlain by peridotitic layers. The angular shape of the blocks and the difference in their composition from the peridotite indicate that they

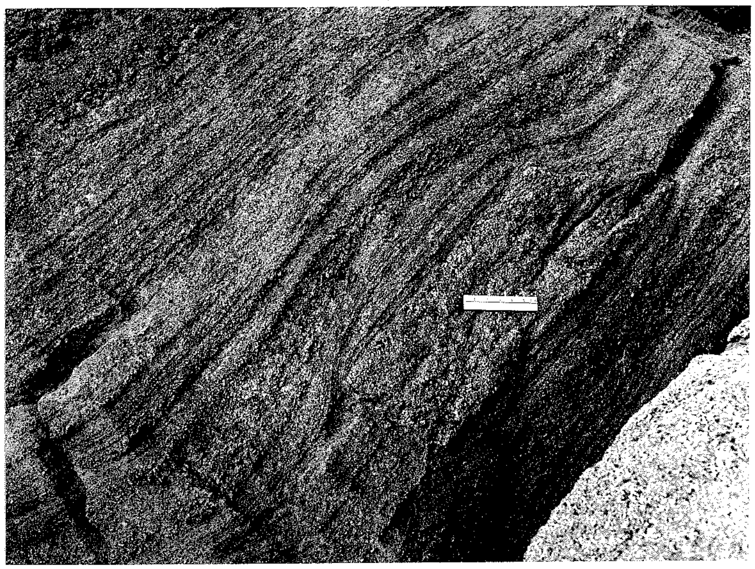


Figure 40. A small block of olivine pyroxenite in layered peridotite. The block is angular indicating that it was well-solidified when broken loose from its original position. The manner in which the layers drape over the block is typical. Note that part of the draped layers have been removed, probably by scour. The locality is in the peridotite zone of the Hall Cove ultramafic area, near its west end.

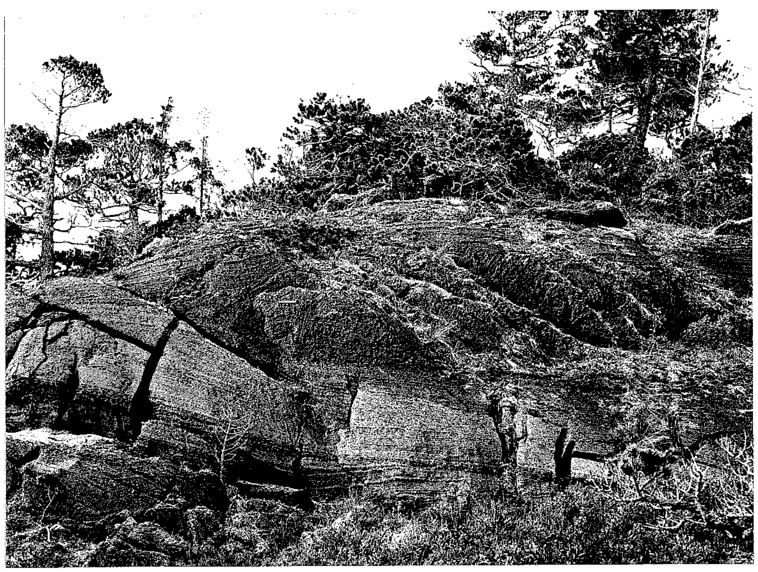


Figure 41. A medium-sized block of olivine pyroxenite in layered peridotite. A 6-inch scale is lying near the left end of the block. The layers beneath the block have been depressed by its weight.

Figure 42. A small fragment of layered olivine pyroxenite in layered peridotite. The locality is near the west end of the peridotite zone in the Hall Cove ultramafic area.

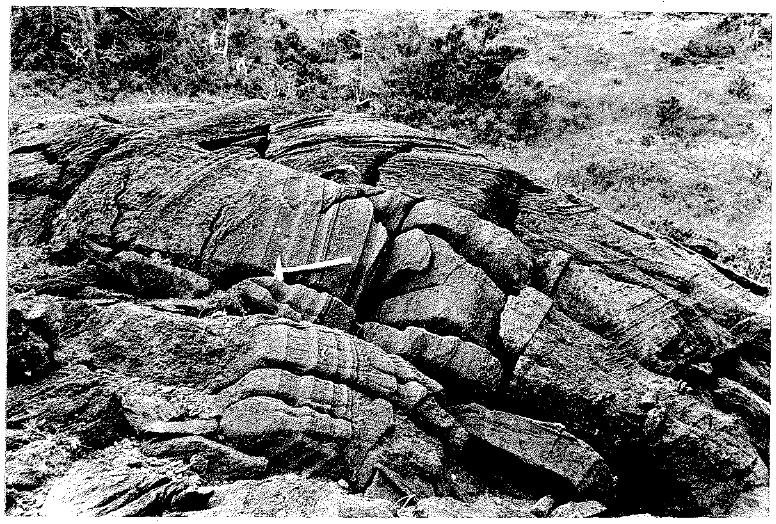


Figure 43. A block of layered olivine pyroxenite over which are draped layers of peridotite. The grading in the layering in the block showsthat it is slightly overturned. Peridotite zone of the Hall Cove ultramafic area.

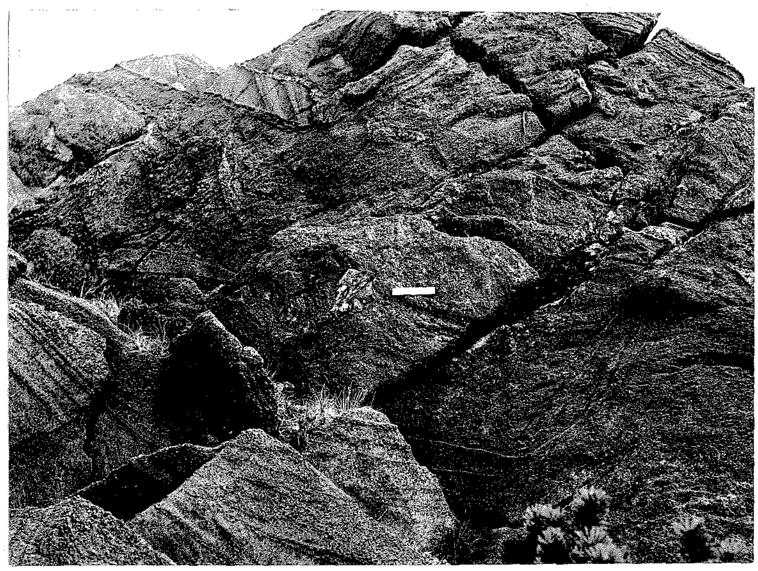


Figure 44. A rubbly mass of slabs and fragments of olivine pyroxenite in peridotite. Some of the fragments are pieces of coarse pyroxene crystals. The locality is the south part of the area mapped in plate 5. The scale is 6 inches long.

Figure 45. Deformation of layering apparently due to the impact of the angular block of olivine pyroxenite in the left part of the photograph. Part of the disturbed layering seems to have been redeposited as new layers, suggesting current action with a component of movement to the right. The locality is close to the line of lithologic section A, near its center.

Figure 46. Deformation of layered peridotite attributed to the influx of olivine pyroxenite blocks. A small fragment is visible a few feet to the right and above the 6-inch scale. A larger mass of olivine pyroxenite occurs in the foreground and numerous other blocks are present in the immediate vicinity. Knob Hill is in the background. The view is to the east.

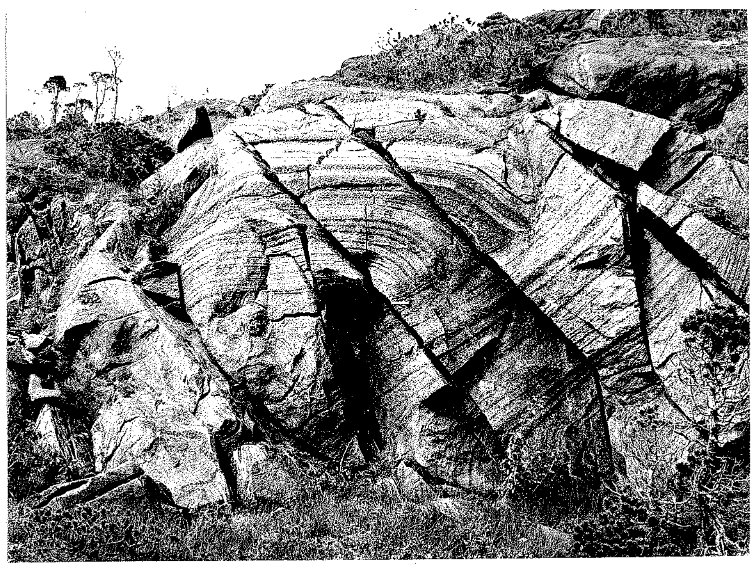


Figure 47. Major deformation of peridotite. Blocks of olivine pyroxenite are not visible in the photograph, but this outcrop is between two of the largest blocks mapped in plate 2. A 6-inch scale rests on the outcrop near the center of the photograph.

represent a distinctly earlier generation of material. Probably none of the fragments would stay in suspension very long in view of the abundant evidence that even individual crystals were settling out. This inference, and the layering in many blocks, suggest that the olivine pyroxenite was derived locally while the layers were accumulating and that it was not carried up by the magma from great depth. A possible source is the northern olivine pyroxenite where transgressed by peridotite. The material at the present level of erosion could not itself be the source because it is stratigraphically beneath the level of the blocks, but possibly at one time the olivine pyroxenite zone projected up the side and over the top of the position now occupied by peridotite.

Some of the olivine pyroxenite blocks actually contain blocks themselves, both as individuals and in fragmental layers. Examples are shown in plates 4 (loc. A) and 5 (loc. C) and in figure 48. Layering has not been observed in fragments definitely established as belonging to the earlier generation, but it may be present (e.g. plate 4, loc. B). Such layering is possible if the northern olivine pyroxenite zone is a source of fragments, because layered blocks have been observed in this zone (fig. 38). Each generation of fragments may not be evidence of a new intrusion, but it does indicate that conditions of sedimentation were unstable and that earlier deposits were repeatedly disrupted.

Variations in the Zone

The character of the peridotite zone at its western end and the changes with elevation in the layered series are shown in lithologic sections A, B, and C (plate 8). Layering is better developed in this part of the zone than to the east, and the sections include more than 90 per cent of the fragmental layers. Otherwise, this part of the zone is fairly typical.

Two poorly-defined, major zones of large blocks are apparent in plate 2 within the rocks shown in section A (although they are not obvious in section A) and these zones persist to the east paralleling the general trend of layering. In sections B and C, fragments are smaller and fragmental layers are fewer and thinner. This undoubtedly indicates that the availability of inclusions was less as accumulation continued, but the sections were measured progressing away from the northern olivine pyroxenite zone and, because of the dip of the layers, may also show a lateral change. Unfortunately, the low relief does not permit a determination of the significance of the lateral change in a north-south direction.

Pyroxene becomes appreciably more abundant in the rocks measured in sections B and C, and the rock is actually olivine-rich olivine pyroxenite grading to normal olivine

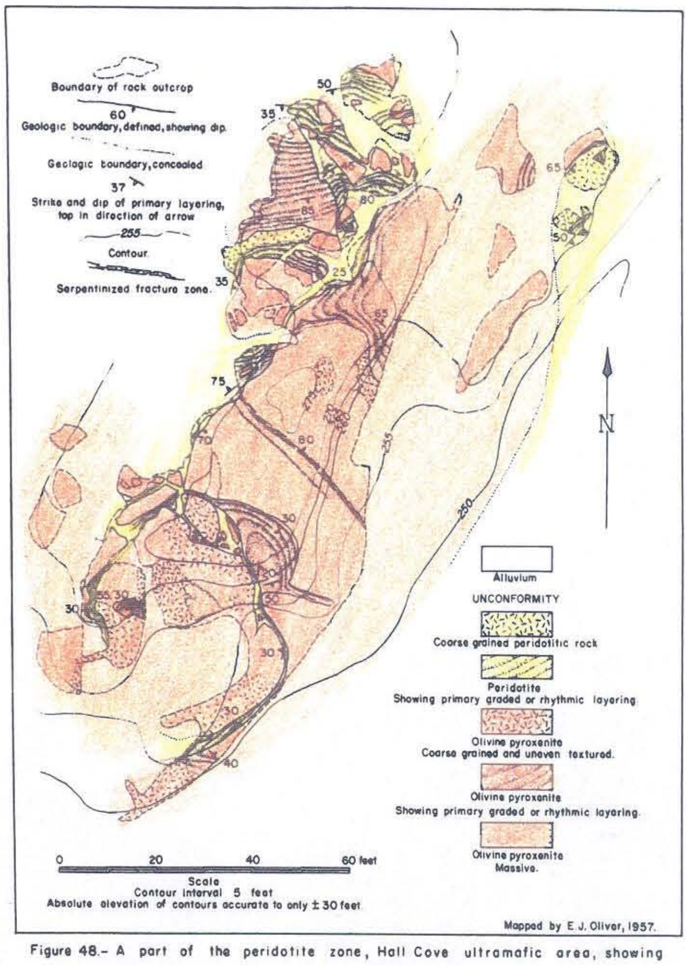


Figure 48.- A part of the peridotite zone, Hall Cove ultramafic area, showing two ages of olivine pyroxenite blocks.

pyroxenite. Despite its composition, however, this material is included in the peridotite zone. It almost certainly grades down into normal peridotite and has the well-developed layering and fragmental layers typical of the peridotitic rocks. Thus, although there is a change in the composition of the layered rocks described in lithologic sections A, B, and C, a natural, sharply-defined break is not apparent, and this part of the layered series is not considered to be a logical place in which to define an arbitrary contact.

In the eastern third of the peridotite zone layering is not prevalent in either blocks or matrix, and coarse-grained and uneven-textured facies are extensively developed. As a result, it is not completely clear that the structures observed even represent a fragmental zone, but this explanation was considered the most satisfactory during mapping and is consistent with relationships observed to the west. The map representation of this part of the zone is semi-diagrammatic at best.

Peridotite Dikes

Dikelike bodies of peridotite occur in olivine pyroxenite at several localities in the northern or lower half of the peridotite zone. They have sharp, subparallel contacts and range in width from 1 inch to 6 feet, and in length from 3 to 50 or more feet. Examples are shown in plates 4 and 5 and in figures 49 and 50. Some apparent dikes probably are septa of peridotite separating unrelated blocks and may be the remnant of material that has been squeezed out from between blocks rather than an intrusion into olivine pyroxenite (e.g., in plate 4, localities D and E are suggestive in this respect). Proof that some are intrusive is best seen in figure 50, where layers in the olivine pyroxenite can be correlated across the dike and are visibly displaced. This is locality B in plate 5.

The dikes are unusual in the following respects:

(a) They intrude only olivine pyroxenite blocks. None of the dikes cuts peridotite, and no comparable bodies have been definitely recognized outside the peridotite zone of the Hall Cove ultramafic area.

(b) The peridotite in the dikes can be traced without discontinuity into the peridotite surrounding the blocks. The peridotite is everywhere identical in composition and textural features, and there is no reason to suspect different generations.

(c) The dike rock is commonly layered. The layering is identical with that in the layered peridotite matrix enclosing the blocks and generally is sufficiently well-graded that direction of tops of layers can be determined. It is everywhere

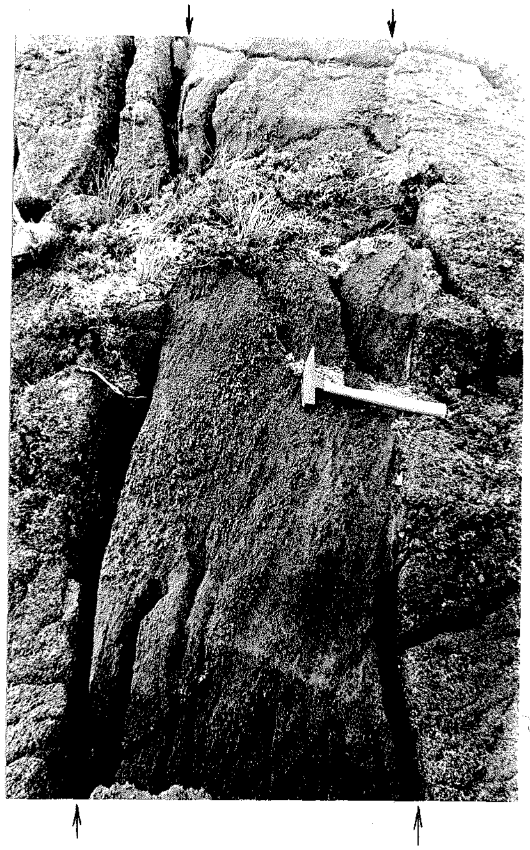


Figure 49. A 2-foot dike of layered peridotite cutting olivine pyroxenite. The relationship of layering to dike contacts (arrows) is notable. The olivine pyroxenite apparently was originally one large block. The locality is at the west end and near the base of the peridotite zone in the Hall Cove ultramafic area. The regional dip of the peridotite-zone layering is towards the observer at about 40°-50°, appreciably different from that in the dike. Joint swarms (foreground) approximately parallel the dike.

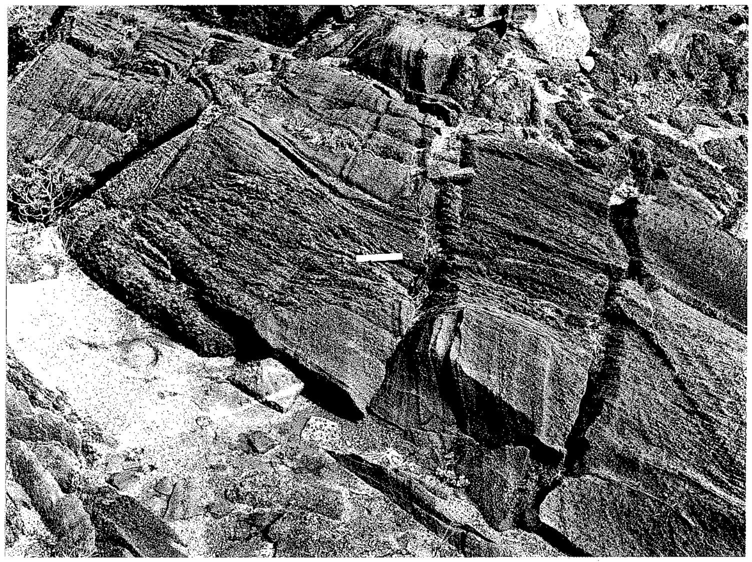


Figure 50. Peridotite dike cutting olivine pyroxenite. The layering in the olivine pyroxenite is notably offset due to dilation caused by the dike; the layer that is groovelike is most easily correlated across the dike. A deformed pyroxene-rich layer in the peridotite is visible just to the right of the 6-inch scale. Joint swarms cut across the dike from left to right. This is locality B in plate 5.

at least slightly irregular and rarely coincides with the regional attitude of layering in the peridotite zone.

Notable examples showing these features are the 5-foot dike at the south end of plate 4 (loc. G), a 3-foot dike at the northwest corner of plate 5 (loc. A), and figure 9. Discussion of the evolution of the dikes will be postponed until the origin of the layering has been considered.

In figure 51 is shown a feature that resembles a dike in some respects. Layers in olivine pyroxenite are correlated across intervening peridotite, which has transverse layering. The main body of peridotite does not have the form of a dike, but the southern olivine pyroxenite body is partially transected by a small peridotite dike. The relationships indicate two blocks of olivine pyroxenite, the northern one underlying draped peridotite layering, and the southern one resting on top. As the blocks probably fell several hundred feet to reach their present position in the layered series, it is unlikely that two originally-adjacent inclusions would come to rest at exactly the same spot but at times sufficiently different that peridotite layers could be deposited between them. On the other hand, the layered peridotite does not appear to have been squeezed between the blocks. A nearly identical relationship occurs at locality C, plate 4, between two blocks of olivine pyroxenite and layered peridotite. No good explanation can

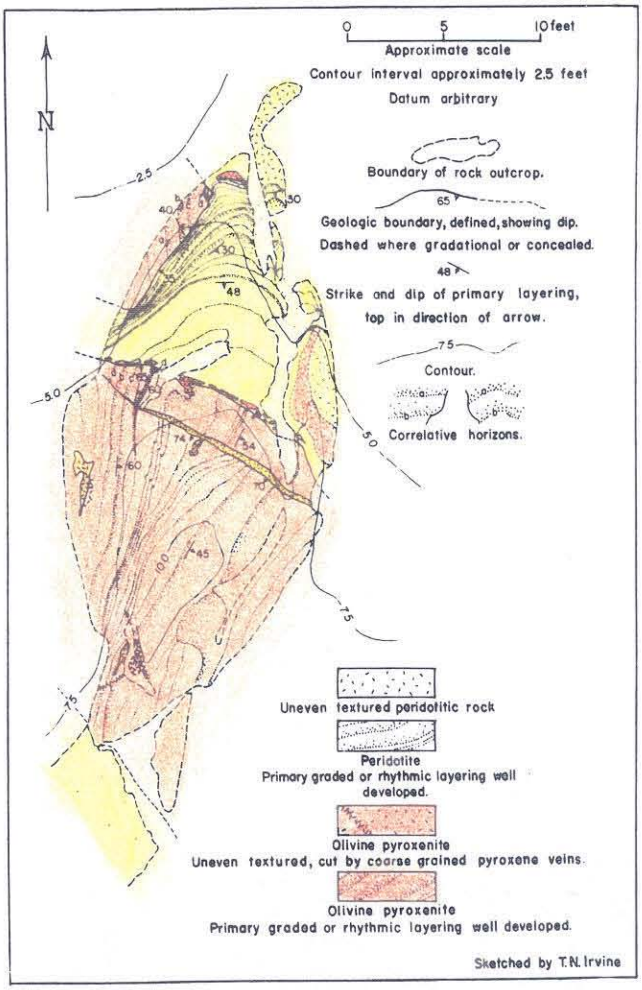


Figure 51.- Layered olivine pyroxenite transgressed by layered peridotite. Hall Cove ultramafic area.

be offered for either example.

Quartz inclusions

Other than olivine pyroxenite blocks, the only definite inclusions observed in any of the ultramafic rocks are composed of quartz. The quartz fragments range in size from 2 inches to 2½ feet (figs. 52-54). They are generally more or less equant, but one is markedly elongate in two dimensions and is probably slablike. All are angular to subrounded. In places the layers drape over the inclusions as they do over the olivine pyroxenite blocks (fig. 53), showing definitely that the pieces of quartz are inclusions, not veins.

All the observed quartz inclusions are in the western and southern parts of the peridotite zone of the Hall Cove ultramafic area. Ten diffent localities are known over a stratigraphic distance of about 1500 feet in the layered series; 7 are indicated in lithologic sections A, B, and C. Several localities have 2 or 3 fragments, and in one layer 3 occurrences of fragments are spread over a distance of 130 feet.

The quartz is coarse-grained, massive, and white, and resembles vein or pegmatite quartz. Some of the inclusions are stained along fractures and grain boundaries by hematite, but no other contaminant has been noted.

In the ultramafic rock surrounding each inclusion is a reaction rim ranging from one quarter of an inch to

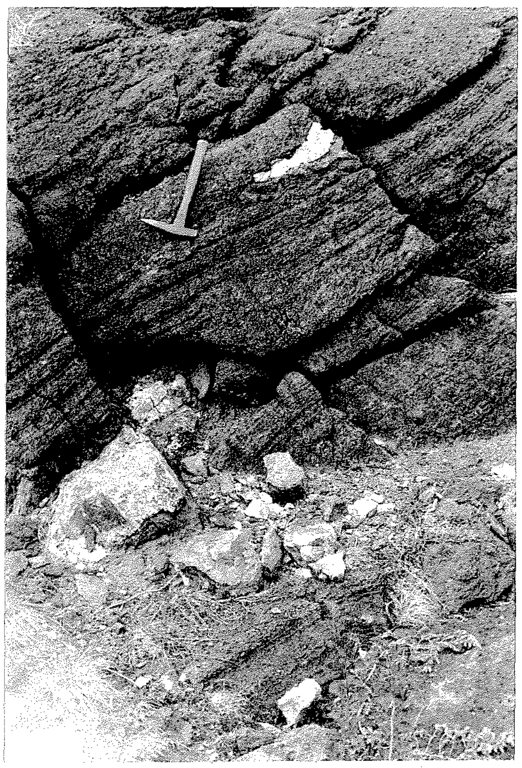


Figure 52. Two quartz inclusions in the peridotite zone of the Hall Cove ultramafic area. The block in the lower part of the photograph is the largest known example. The locality is the 518-foot level of lithologic section A.

Figure 53. Quartz inclusion in the upper part of the peridotite zone of the Hall Cove ultramafic area. The olivine pyroxenite layers are notably draped over the inclusion. The locality is the 210-foot level of lithologic section C.

Figure 54. Slablike quartz inclusion in the lower part of the peridotite zone in the Hall Cove ultramafic area.

3 or 4 inches in thickness. This is composed essentially of clinopyroxene of brighter green color than is typical of that in the normal olivine pyroxenite and peridotite. Olivine is absent from the rim, and no orthopyroxene has been recognized in the specimens examined microscopically. A specimen (No. T-17-1) from one of the thicker rims contains coarse hornblende that is almost colorless in thin section. It also has fine-grained quartz interstitial to the pyroxene and a small amount of fine, chalky white material that could not be identified completely by either optics or X-ray powder pattern, but which is in part plagioclase. Optical properties of minerals from two of the specimens of the reaction rim are as follows:

Specimen	Material	n_y	2V
I-46-2	Olivine in the rock	1.690	
	Clinopyroxene in the rock	1.685	52°30'
	Clinopyroxene in the rim	1.686	55°30'
T-17-1	Clinopyroxene in the rim	1.686	56°10'

It was expected that the reaction of olivine and quartz had produced enstatite and that the latter had been taken up in solid solution in the clinopyroxene. However, the 2V measurements indicate that, if anything, the clinopyroxene in the rim has a lower enstatite content than that in the rock. A separate of the pyroxene from specimen T-17-1 was prepared, and Mr. Henry Schwarcz kindly compared its composition with that of one of the chemically analysed pyroxenes (specimen I-37-2) by X-ray fluorescence methods. Using specimen I-37-2 as a standard, and assuming that the rim pyroxene has the same ratio of ferrous to ferric iron and the same content of Na_2O and K_2O, the analysis given in table 12 was obtained.

TABLE 12. X-RAY FLUORESCENCE ANALYSIS OF CLINOPYROXENE FROM A REACTION RIM AROUND A QUARTZ INCLUSION

	I (standard)	II
SiO_2	51.50	54.93
Al_2O_3	4.31	0.86
Fe_2O_3	1.17	(1.42)
FeO	4.36	(6.52)
MgO	15.20	13.73
CaO	22.42	21.84
Na_2O	0.47	(0.47)
K_2O	0.04	(0.04)
H_2O+	0.20	N.D.
H_2O-	nil	N.D.
TiO_2	0.32	0.10
McO	0.06	0.07
P_2O_5	0.06	N.D.
CO_2	nil	N.D.
	100.11	99.98
Atomic ratios		
Ca	46.8	46.3
Mg	44.1	40.6
Fe	8.9	13.1

I. Clinopyroxene from typical olivine pyroxenite. Specimen I-37-2. W. H. Herdsman, Analyst.

II. Clinopyroxene from reaction rim. Specimen T-17-1. H. P. Schwarcz, analyst. Values in parentheses are explained in the text.

There is no increase in the ratio Fe+ Mg:Ca in the rim pyroxene, and actually, the proportion of Mg is lower. The principal differences are that the rim pyroxene has:

(1) A smaller content of TiO_2. This could be the major factor in determining the color of the mineral.

(2) A lower content of Al_2O_3 and a higher content of SiO_2. Evidently, in the presence of abundant silica, fewer Al atoms occupy tetrahedral positions in the pyroxene structure than is typical in the ultramafic rocks. The Al may have been taken up by either hornblende, or plagioclase, or both.

It seems clear that the development of the reaction rim involves more than is expressed simply by the equation

$$\underset{\text{olivine}}{(Mg,Fe)_2SiO_4} + \underset{\text{quartz}}{SiO_2} \longrightarrow \underset{\text{enstatite}}{2(Mg,Fe)SiO_3}$$

but all the details are not understood, particularly the behavior of magnesium.

The derivation of the quartz is unknown. The only near sources are quartz veins and small granitic pegmatite dikes with parts of coarse quartz occurring in the gabbro around the northwest corner of the Hall Cove ultramafic area. Quartzites and cherts are not known in the vicinity of Duke Island. The metamorphic rocks along the shore north from Hall Cove are cut by quartz veins, but these are a considerable distance from the area in which the inclusions occur. Any of the veins and pegmatitic dikes related to the granitic rocks probably are younger then the ultramafic rocks, as this is the only definite age relationship observed for the two rock groups. If the veins and dikes in the gabbro are the source of the inclusions, a time period between the solidification of the gabbro and the emplacement of the ultramafic rocks is required during which highly silicic material was introduced. No other indication of more than one age of silicic igneous material has been recognized at Duke Island, but the author has observed two ages of granitic rocks at Alava Bay, Revillagigedo Island, 40 miles northeast, one older than a hornblendite body and the other younger. Thus all quartzose igneous material in the region does not belong to the same age, and perhaps there are two ages at Duke Island.

An attempt was made to correlate the quartz inclusions with various types of dikes and veins on the basis of the isotopic composition of their oxygen. The analytical procedure and method of expressing the results are described on pages 8-9. The data on quartz are given in table 13. The results are reproducible to about $\pm$ 0.2 per mil. The inclusions match best with the quartz veins in the gabbro, the correlation that seems most likely on geologic grounds. However, it must be concluded that correlation is neither substantiated nor disproved by the isotopic data. Sampling is not complete enough to know the isotopic variability of either the inclusions or the veins and dikes. Furthermore, although the inclusions are obviously not in chemical equilibrium with the ultramafic rocks, some isotopic exchange between them may have occurred. Exchange might, for example, be effected via water diffusing between the inclusion and the surrounding rock. Studies of the isotopic composition of igneous minerals suggest that quartz in isotopic equilibrium with ultramafic minerals should have a low O^{18}/O^{16} ratio (H. P. Taylor Jr., personal communication). The range in the isotopic composition of oxygen in natural quartz as known at the present time is about 6.5 to 34 per mil as compared to mean sea water (S. Epstein, personal communication). The inclusions have one of the lowest ratios. This is not proof of isotopic exchange between the inclusions and the ultramafic material, but it is suggestive.

A puzzling problem related to the quartz inclusions is how they persisted in a magma that contained forsteritic olivine and thus was clearly undersaturated in silica; one might expect immediate solution or assimilation of the quartz. Another problem is that other inclusions have not been observed. Factors which may have been important in determining these relationships are:

(a) The rate of solution of an inclusion in a melt may be dependent in part on the temperature of fusion of the inclusion relative to the temperature of the melt. The fusion temperature of pure quartz probably would be higher than that of almost any other inclusion the magma is likely to acquire. Thus quartz might dissolve much more slowly than other inclusions. This does not mean that the quartz has to melt in order to be assimilated into the magma, the explanation is one of kinetics. Melting of other inclusions would, of course, facilitate their solution, and the coarse grain size of the quartz might reduce its rate of solution.

(b) A separation of quartz from other types of inclusions might have been effected because of relative densities. Most of the exposed country rock that could have contributed inclusions is considerably denser than quartz and would settle more rapidly in the magma. Possibly, quartz even floated and remained suspended until it was in some way carried down to the floor. Separation of the inclusions could then have been made almost as soon as they entered the magma body.

(c) Possibly, while the quartz was in suspension in the magma the extraction of heat from the melt for the solution of the quartz caused the precipitation of an armoring rim of pyroxene. Once armored, pure quartz unlike chemically-complex inclusions would not melt, and hence the rim could persist. If armoring really was effective, the quartz might be pieces of rocks at depth which were included in the magma and which remained in suspension while other types of inclusions dissolved. The rim pyroxene would then have been in equilibrium with large amounts of magma and would not be expected to have a greater magnesium content than other pyroxene separating from the magma at the same time. However, by this explanation, some of the inclusions probably had to remain in suspension during the time required to accumulate at least 1500 feet of layered rocks.

(d) Even though the deposition of quartz fragments may have been delayed because of their low densities, burial is undoubtedly an important reason why they still exist. By burial in the layered rocks, they have been insulated from reaction with all but the immediately adjacent ultramafic material. Much of the extent of this reaction is probably represented by the rim of bright green pyroxene.

TABLE 13. ISOTOPIC COMPOSITION OF OXYGEN IN QUARTZ FROM DUKE ISLAND ROCKS

Entry	Specimen	Occurrence	δ (‰)
1	T-2-1	Inclusion in peridotite	7.96
2	I-46-2	Inclusion in peridotite	8.12
3	T-2-2	Vein in norite	8.95
4	T-2-4	Granitic pegmatite in gabbro	9.10
5	T-23-1A	Vein in metamorphic rocks	10.25
6	T-23-3	Vein in metamorphic rocks	10.16
7	N-18-3	From typical quartz diorite	10.70

<u>Southern Boundary</u>

The southern limit of the peridotite zone is a gradational change into olivine pyroxenite without definite angular discordance in the layering of the two rock types. The change may be a progressive one due to differentiation of the magma by fractional crystallization. Regardless of the reason, a break comparable to the intrusive contact forming the northern boundary of the peridotite zone is not obvious; hence a contact has not been drawn on the map, the change being shown only in the coloring.

On the other hand, the exposure is not so complete that a contact could not have been overlooked or misinterpreted. If the contact does exist, the most likely place for it is just beyond the upper limit of lithologic section C. On the map, this corresponds to the tip of the narrow gabbro lobe projecting into the west side of the ultramafic area. From this point, a contact could trend slightly north of east following the upper limit of the more peridotitic rock as far as the Hall Cove fault, where it would, presumably, be displaced out of sight. Neither fragmental layers nor quartz inclusions have been observed south of this line. It will be noted that the layering at the tip of the gabbro lobe flattens from 40°-50° to as low as 10° and then steepens again farther south. This may be the result of folding or even primary deposition, but it might indicate that a contact has been crossed. A small amount of hornblendized olivine pyroxenite with magnetite is present near by and may be related to a contact.

Uncertainty with regard to the nature of the contact gives two possible interpretations of the structural history of the area. If the contact is a conformable gradation, then the southern olivine pyroxenite probably has been deposited on top of the peridotite and is therefore younger than the peridotite. If the contact is intrusive, then it may correspond to the northern boundary of the peridotite zone; the peridotite would then be younger.

Southern Olivine Pyroxenite Zone

South and west from the peridotite zone, or proceeding upward through the olivine pyroxenite, layering is much less common, occurring sporadically in thin sequences of limited lateral extent. One notable feature of the layering, particularly in the upper part of the series, is that the layered rock is much finer-grained than the adjacent massive rock, but the reason is not known. Most of the layering shows at least poorly developed grading. Fragmental layers have not been observed.

Directions of dips and tops of layers are consistently to the south or southwest, and unless numerous unrecognized faults are present, the structure is comparatively simple. The main feature is a large open syncline whose axis trends N.65°E and plunges 40°-60°S.W. The north limb of the fold trends west and dips 35°-50°S. The south limb trends N.20°E. and dips 30°-45°N.W.

Moderately coarse, uneven textures are common in olivine pyroxenite throughout much of the zone, and pods or segregations of pegmatitic pyroxene and olivine occur locally. Small patches of dunite are concentrated in several areas (plate 2); figure 55 is a photograph of one of these patches.

North Hill Ultramafic Area

An area of $\frac{1}{4}$-square mile of ultramafic rocks crops out at North Hill. This is within 1500 feet of the Hall Cove ultramafic area, and the rocks in the two localities are very similar and probably are parts of one original body. Hornblende pyroxenite forms a partial peripheral rim, and again the rim may be more extensive than shown on the map simply because of the nature of exposure. Most of the area is underlain by olivine pyroxenite with graded layering trending N.60°W. and dipping 30°-40°S.W., with tops of layers facing south. In the northern half of the area, the layering is irregular and discontinuous. In contrast, the southern or upper third of the exposed mass has approximately 250 feet of highly regular layering in rock slightly more rich in olivine than normal olivine pyroxenite. The position and attitude of this section is such that its base lines up reasonably well with the northern boundary of the peridotite zone in the Hall Cove ultramafic

Figure 55. One of the numerous small, irregular-shaped bodies of dunite (light-colored material) occurring in the southern olivine pyroxenite zone of the Hall Cove ultramafic area. A slight suggestion of planar character is evident. This approximately parallels the layering in the vicinity. For a discussion of the origin of this type of dunite, see pages 247-264.

area, and the increase in olivine suggests that the two may in fact correspond.

OLIVINE-BEARING ROCKS EAST OF THE HALL COVE FAULT

The part of the Hall Cove ultramafic area east of the Hall Cove fault makes two tonguelike areas projecting southeast.

The smaller tongue has olivine pyroxenite at its base and hornblende pyroxenite at its tip. The olivine pyroxenite is apparently devoid of layered structures. Several small outcrops of dunite or pyroxene dunite are present within the area of olivine pyroxenite, and one of these, occurring along the east side of the peninsula in Hall Cove, is about 20 by 6 feet and consists of 50-55 per cent olivine, 10-15 per cent clinopyroxene, and 35-40 per cent magnetite. The concentration of magnetite is the largest known to the author in any Duke Island rock. This is surprising, because hornblende pyroxenite is the common magnetite-rich unit, whereas dunite is generally magnetite-free.

The principal olivine-bearing rock exposed in the large tonguelike area is olivine pyroxenite, with only a few small patches of dunite. Layering in the olivine pyroxenite is both graded and non-graded, and this is the best locality in which to see the transitional relationship between the two. Plate 2 shows the structure in the layering.

This structure has no simple relationship to that on the west side of the Hall Cove fault. The layers strike about the same, but directions of dip and tops are opposite. East of the fault the layers are folded; west of it, they dip steeply but are comparatively regular. This disharmony is not only one of the major reasons for believing the fault exists but is also one of the more problematical structural features of the area.

FAULTS AND JOINTS

Three faults are shown in the western part of the Hall Cove ultramafic area. Two are not visible in outcrop but have been postulated to explain irregularities in contacts. The third is a sheared zone rich in serpentine.

One of the faults offsets the northern olivine pyroxenite zone (p. 123) at a point where a sharp bend occurs in the northern boundary of the ultramafic area. Its direction, N.30°E., is parallel to the general lineation in the area but is also the direction of glaciation; hence the topographic expression may be deceiving. The fault does not appear to extend south into the peridotite. Perhaps displacement took place prior to the accumulation of the peridotite and in advance of the emplacement of the magma from which the peridotite formed.

In the southern part of the area, an apparent right-lateral displacement of about 1500 feet in the hornblende pyroxenite border zone is attributed to a fault trending N.60°-70°E. The trace assigned to the fault is a prominent lineament marked by several cliffs and one lake. The northeastern continuation of the fault into the ultramafic rocks is indefinite, but the lineament converges with the axis of the major syncline in this area, and the fault may terminate in the fold; the fold becomes appreciably tighter at the point of convergence. Air photographs show that the direction N.65°E. is common to many lineaments in the area. However, neither folding nor definite displacement due to faulting has been related to any of them.

The serpentinized zone cuts the southern olivine pyroxenite zone and is about 1000 feet west of Hall Cove. The strike is about N.15°E. and differs from that of the Hall Cove fault by 10°-20°. The attitude is approximately vertical. The zone differs in width and definition from place to place but is about 30 feet wide where best exposed. Definite displacement along the zone has not been noted.

Joint swarms are impressively developed in the western part of the Hall Cove ultramafic area, An example is shown in figure 79, and many others can be recognized in the photographs of the layered structures. The swarms are numerous parallel, but discontinuous fractures that controlled serpentinization and, therefore, show prominently on the weathered surface. They are developed throughout the entire area west of the Hall Cove fault and everywhere have nearly the same strike and dip (N.10°E. $\pm$ 10°, 80° to vertical). This attitude is the same as that of the serpentinized shear zone described before. In the west-central part of the area is a second, less prominent set of joint swarms that trends about N.65°E., the same direction as the topographic lineaments in this vicinity. All the north-trending joints are clearly later than the consolidation of all the ultramafic rocks and are completely unrelated to any of the deformational structures in the layered rocks. They cut gabbro and the quartz veins in the gabbro, and they are believed to cut the granitic rocks, although this has not been shown conclusively. Joint swarms are rare east of Hall Cove and, for this reason, are believed to be related in some way to the Hall Cove fault, even though they do not parallel it. Most of the straining of mineral grains in the ultramafic rocks seen in thin section appears to be related to the joints. This and the constancy of direction suggests that they are due to shearing. However, significant displacement has not been recognized on any one fracture plane.

THICKNESS OF THE LAYERED SERIES

The stratigraphically lowest layering in the northern olivine pyroxenite zone roughly parallels the adjacent boundary of the Hall Cove ultramafic area, hence the boundary may be the original bottom of the ultramafic complex now turned on end. However, evidence in other layered intrusions (Wager and Deer, 1939; Wilson, 1956; Rossman, 1954) indicates that igneous stratification commonly has dips flatter than the contacts. This relationship is suggested at several places in the Duke Island ultramafic rocks and may exist along the northern boundary of the Hall Cove ultramafic area because the dip of the contact was measurable at but one locality, and the total relief on the contact here was only about six inches. Layered rocks may, therefore, be present beneath the lowest now exposed.

The top of the layering in the northern olivine pyroxenite zone is truncated at the contact with peridotite and, hence, is also unknown. The maximum exposed thickness for the zone, as determined from the map, is about 3900 feet. A total thickness of 1700 feet has been measured in the peridotite zone and is shown in lithologic sections A, B, and C, The three sections do not appear to repeat and in fact, they may be separated by considerable unexposed thicknesses of layered rocks. The peridotitic layering may be flatter than the northern boundary of this zone, and therefore the lowest layers again may not be exposed. Uncertainty over the nature of the southern boundary of the peridotite zone has been discussed (p. 154).

The layering in the southern olivine pyroxenite zone is not continuously developed, and the structure is complicated by the syncline and at least one fault. Therefore a thickness estimate for the zone is conjectural. Sections taken from the map indicate an approximate thickness of 6500 feet. This probably is not large by more than a factor of two, and actually nothing indicates that the top of the section has been reached.

In summary, the total visible thickness of apparently non-repeated layered series exposed in the Hall Cove ultramafic area is about 12,000 feet.

JUDD HARBOR ULTRAMAFIC AREA

DISTRIBUTION OF ROCK TYPES

The Judd Harbor ultramafic area has the characteristic rock units (dunite, peridotite, olivine pyroxenite, and hornblende pyroxenite), but the pattern of their distribution is somewhat atypical (plate 3). The main unit of dunite and peridotite is along the north side of the area in contact with surrounding gabbro instead of being disposed centrally in the more common zonal arrangement of the ultramafic rocks in southeastern Alaska. Unlike the Hall Cove ultramafic area, the only occurrence of hornblende pyroxenite along the border is confined to two small masses on the southern fringe, one exposed on Kelp Island, the other at Judd Harbor. An appreciable amount of hornblende pyroxenite is developed through the middle of the olivine pyroxenite.

Part of the reason for this atypical distribution may be a lack of exposure along most of the outer boundary of the northeastern half of the ultramafic rocks. However, the same parts of the boundary that lack a hornblende pyroxenite border zone are marked by a series of relatively steep-walled, linear gullies occupied by streams and lakes; this suggests that they are largely fault contacts.

The southern part of the boundary, unless faulted where covered by water, shows the same relationships of hornblende pyroxenite to gabbro as were observed in the Hall Cove ultramafic area (pp. 119-120). The contact is exposed in five outcrops on Kelp Island, and in each it is comparatively sharp but still so hazy and irregular that the dip could not be measured.

OLIVINE PYROXENITE

Olivine pyroxenite underlies about three-fourths of the Judd Harbor ultramafic area and is the most extensively layered part of the Duke Island ultramafic complex. Most of the layering is the non-graded type, and in only a few places has evidence of direction of tops been found. The olivine pyroxenite is exceptionally coarse grained (10-20 mm) over large areas, and commonly the pyroxene crystals are oriented with their C-axes normal to the layering (see pp. 246-247).

The principal structural feature is a large, inverted S-pattern in the layering due to a steeply plunging anticline and syncline. The fold axes trend N.70°-80°E. and plunge 60°-70°S.W. This is approximately the same attitude as that of the syncline in the southern olivine pyroxenite zone in the Hall Cove ultramafic area. In the few localities where the layers are sufficiently well graded that direction of tops can be determined, the upper surfaces face south and west along the fold axes. No evidence of reversals due to isoclinal folding have been recognized. The southern flank of the anticline and the northern flank of the syncline both trend about N.65°W. and generally dip about 50°-70°S. The limb common to both folds trends N.50°W. and dips 85°S., indicating that the folds are slightly overturned to the north. A few minor folds occur on the limbs of the major ones and have approximately the same disposition, including being slightly overturned to the north. The nature of the folds suggests that the rock was relatively plastic during folding, and it is probable that the deformation took place shortly after accumulation of the layers and before the interprecipitate magma had completely solidified.

Numerous irregular patches of dunite or pyroxene dunite with dimensions ranging from a few feet to a few hundred feet occur in the olivine pyroxenite and seem to be concentrated along the axial parts of the folds.

DUNITE AND PERIDOTITE

In the main zone of dunite and peridotite, the two rocks are intermixed and transitional without a natural compositional break. Consequently no contact has been drawn between them on the map, and the principal areas of each have only been distinguished with different shades of the same color. On the other hand, the contact of these two units with olivine pyroxenite is easily defined and, at the most, grades over 10 to 20 feet. Dunite predominates over peridotite and the zone on the whole has much less pyroxene than the peridotite zone in the Hall Cove ultramafic area. Partial serpentinization of olivine is pervasive. Veins of coarse pyroxene are common, and a few tiny veinlets of chromite or chromiferous magnetite are present in the pyroxene-free dunite.

Relatively good graded layering is developed in the peridotitic facies (fig. 28), but layering has not been recognized in the dunite, probably because of the absence of contrasting pyroxene. The principal structure formed by the layering is a basin elongated in an easterly direction and having steep inward dips (70°-80°) on the south side and low inward dips (15°-25°) on the north. This structure bears no relation to the folded pattern in the olivine pyroxenite. In fact, the contact of these two major units cuts across the olivine pyroxenite folds in a way to suggest that dunite is part of a mass which has been intruded into the olivine pyroxenite and whose emplacement may have caused the folding in the latter. One local exception to this relationship occurs near the southwest end of the contact. Here, a trace of the direction of the olivine pyroxenite layering appears to persist into the dunite for a few tens of feet.

Several comparatively small bodies of olivine pyroxenite are isolated along the north side of the dunite zone. They contain layering that appears to be structurally unrelated to that occurring in the dunite, but which might be fitted into the folded pattern of the main olivine pyroxenite zone. In a gross way, these bodies resemble the large blocks of layered olivine pyroxenite that occur in the peridotite zone of the Hall Cove ultramafic area and thus suggest parallel histories for the olivine-rich parts of the two areas. The similarity may be even greater than can be shown, because at the time the Judd Harbor ultramafic area was mapped the large dimensions of the blocks in the Hall Cove ultramafic area had not been recognized, and there has been no subsequent opportunity in which to make a careful comparison.

HORNBLENDE PYROXENITE, HORNBLENDITE, AND BASIC PEGMATITE

The masses of hornblende pyroxenite underlying the southern part of the Judd Harbor ultramafic area are generally typical of this rock unit, except for some magnetite banding or layering in the Kelp Island occurrence.

This layering has a slight suggestion of mineralogical sorting, with magnetite concentrated at the supposed bottom and grading sharply upward into pyroxene and hornblende. The Kelp Island hornblende pyroxenite is one of the few occurrences of this rock type that seems to be stratigraphically above olivine pyroxenite. If the magnetite banding is gravitational layering, then it is possible that the rock is a late-stage differentiate of the same magma from which the olivine pyroxenite formed. The layering would have to have been folded into at least one anticline with steep or vertical limbs. However, this is possible because the fold fits in reasonably well with the pattern of the larger folds in the olivine pyroxenite.

Several areas of hornblendite have been shown on the map. Most of this rock is coarse-grained and massive, but part has pegmatitic veins or segregations with large perpendicularly-oriented crystals (fig. 24). The distribution, mineralogy, and textural relationships of this material suggest that it has formed by the concentration and reaction of late magmatic fluids along limited zones in olivine pyroxenite.

Some of the rock mapped as hornblende pyroxenite is actually fine-grained hornblendite and probably indicates that pyroxene, which is characteristically rimmed by hornblende in this zone, has reacted completely with the magma from which it crystallized. It also is possible, but not certain, that some of the hornblende has crystallized directly from the magma and accumulated by gravity.

Basic pegmatite dikes are common in the Judd Harbor ultramafic area and cut all the rock types, although they are very rare in dunite. They are commonly associated with hornblendite and hornblendic alteration in the olivine pyroxenite. Some contain appreciable amounts of coarse magnetite, and marked concentrations of magnetite are present in the rock adjacent to the walls of many dikes. A relatively concentrated zone of near-vertical dikes trends N.60°W. through the middle of the main olivine pyroxenite area and is accompanied by extensive alteration of olivine pyroxenite to magnetite-bearing hornblende-olivine pyroxenite and hornblende pyroxenite. The alteration zone is definitely younger than the olivine pyroxenite, as it crosses the layering and obliterates the structure. The prominence of magnetite is well shown by the large aeromagnetic anomaly that occurs over the zone. Textural relations indicate that magnetite is the principal mineral taking the place of olivine, whereas hornblende replaces pyroxene. The alteration and the dikes must be genetically related because of their close spatial relationship and because hornblende and magnetite, primary minerals in the dikes, are the principal minerals produced by the alteration. However, every dike does not have an alteration zone, and every patch of alteration does not have a visible dike. In many places, the amount of alteration seems far too great to have been caused solely by the quantity of dike exposed. The zone must have been invaded by late magmatic materials which produced either dikes, or alteration, or both.

Where alteration is extreme, the rock produced is hornblende pyroxenite. This material is mineralogically the same as that which occurs in the border zone, containing optically identical hornblende, pyroxene, and hercynitic spinel, plus magnetite with the same proportion of associated ilmenite. The only differences are that, in the alteration zone, magnetite is slightly more sporadic and grain size is coarser. The grain size seems to reflect the coarse-grained character of the olivine pyroxenite in the vicinity of the alteration zone.

THICKNESS OF LAYERED ROCKS

Because of low dips and structural complications, no estimate of thickness will be attempted for the zone of dunite and peridotite.

A thickness of 5300 feet has been obtained for layered rocks in the visible part of the olivine pyroxenite zone by using sections taken from the map. The structure is obviously complicated, and this estimate is doubtful, but probably minimal. It is not possible to tell whether either top or bottom of the zone is exposed.

FAULTS AND JOINTS

The possibility that much of the northern boundary of the Judd Harbor ultramafic area is marked by faults has been considered (p. 163). Two minor faults have been indicated on the map in plate 2. Both may be related to folding and, possibly, to the emplacement of the dunite as they apparently do not extend beyond the ultramafic area. Joint swarms similar to those so prevalent in the western part of the Hall Cove ultramafic area are not present in the Judd Harbor ultramafic area. The pegmatite dikes commonly trend N.60°W., as does the contact of the main zones of olivine pyroxenite and dunite. This apparently is the major fracture direction in the area.

OTHER ULTRAMAFIC AREAS

The next largest ultramafic body underlying Duke Island, ¼-square mile in area, is in the northwest part of the island about one mile south of Form Point. It is olivine pyroxenite with a partial border zone of hornblende pyroxenite on the east side. Small amounts of gabbro crop out at several places along the boundary, and in two outcrops, sharp, nearly vertical contacts of gabbro with hornblende pyroxenite are exposed. Difficulty was encountered in distinguishing the gabbro from the mafic-rich diorite or quartz diorite in the area, and the amount of gabbro may actually be larger than is shown on the map.

Otherwise, the ultramafic body is completely surrounded by granitic rocks and possibly represents a portion of the body cropping out as the Hall Cove ultramafic area that has foundered in the younger granitic intrusion. A small amount of graded layering is exposed in one outcrop of olivine pyroxenite and strikes N.70°W., dips 55°N.E., and has tops facing northeast. Vertical joint swarms identical to those in the Hall Cove ultramafic area and having the same strike (N.10°-20°E.) are extensively developed. That they do have the same direction is a further suggestion that their development post-dated the solidification of the granitic rocks.

A body of olivine pyroxenite and hornblendite 500 feet by 40 feet in size crops out at the shore about 2000 feet northeast of Point White. It is engulfed in and cut by several dikes of gneissose granitic rock. The hornblendite is concentrated in the marginal parts of the body.

A strip of hornblendite 4000 feet long and 1000 feet wide strikes N.70°W. along the north shore of Duke Island just opposite Vegas Islands. The hornblendite is coarse-grained and massive and has a minor dissemination of ilmeno-magnetite and sparse clots of epidote or clinozoisite. Two small exposures of olivine pyroxenite occur near the west end but are considerably hornblendized. Crystals of hornblende with maximum dimensions of 4-7 feet occur in pegmatitic segregations; some of these have been described by Koschmann (1935). Granitic rocks occur on all sides of the body. The contact on north and east, where exposed in the tidal zone, is knife-edge sharp and dips 35°S. At the west end of the body, the contact is hazy and transitional over a distance of 20 feet. The southern boundary is not exposed. Age relations are not clear, although the granitic rock probably is younger as it contains abundant inclusions of gabbro and has foliation that swings around the ultramafic body in a flowlike pattern.

At the tip of Form Point, several small masses of hornblendite, the largest 20-30 feet in length, are included in granodiorite. Some gabbro inclusions also are present. Foliation in the granodiorite generally wraps around the inclusions.

A body of hornblendite about 1200 feet in diameter crops out at Ryus Bay. It is bordered by gabbro on the south and west, but the contact is not exposed. A small mass of quartz diorite apparently intrudes the contact on the southeast and, where in contact with hornblendite, appears to be contaminated with mafic material.

A 300-foot by 900-foot mass of hornblendite is exposed on the south side of the head of Niquette Harbor. The body is bounded by granitic rock on the south and east, but the contacts are not exposed. The northwest boundary may be marked by a fault beneath and parallel to Niquette Harbor.

A small body of olivine pyroxenite occurs in a narrow septum of gabbro intruded by granitic rock 4000 feet west of the southern extremity of Pond Bay.

Two small outcrops of ultramafic material occur along the shoreline of the southwest apex of Dog Island, one on the northwest side of the apex, the other on the south. Both are surrounded by gabbro that has been intruded, feldspathized, and hornblendized by granitic material. The ultramafic rocks have also been metamorphosed, being sheared and altered to a serpentine (?). The original rock apparently was fine-grained olivine pyroxenite or peridotite.

An interesting ultramafic rock occurs in a 400-foot exposure along the north shore of Duke Island, about one mile west of Grave Point. The rock is bounded at both ends by gabbro, but the contacts are not visible. It is exceptionally fine-grained (0.5 mm) and comes closer to being a chilled ultramafic rock than anything known to the author in this area or in southeastern Alaska. Modes of three specimens of the rock are as follows:

	I-3-3	I-3-4	I-3-5
Olivine	27.53	43.97	12.90
Serpentine	7.44	2.55	.44
Orthopyroxene	.12	12.13	38.03
Magnetitic oxides	4.69	2.92	5.92
Hornblende	41.24	33.63	30.55
Hercynitic spinel	18.98	4.79	12.16
	100.00	99.99	100.00

They show that the rock is non-uniform. This is one of the two bodies of ultramafic rocks known on Duke Island to contain orthopyroxene and the only one to contain coexisting olivine and hercynitic spinel. The occurrence of such large amounts of hornblende and olivine together is atypical, as is the large amount of hercynitic spinel. In specimen I-3-5, the olivine has an n_y refractive index of 1.688, indicating a composition of $Fa_{17.5}$; the orthopyroxene has $n_x = 1.674$, $n_y = 1.685$, indicating a composition of Of_{13}; and the hercynitic spinel has an index of about 1.745. In all specimens, the hornblende is pale brown and primary in appearance.

Just north of Flag Point on the east side of Duke Island, olivine pyroxenite occurs as a small lens in sharp contact with metasedimentary rock essentially composed of hornblende, plagioclase, and biotite. The ultramafic rock is slightly sheared and serpentinized, probably indicating mild metamorphism by granitic intrusions occurring near by.

An isolated outcrop of olivine pyroxenite is present in gabbro about ¼-mile south of the half-way point along the length of Morse Cove. The rock is composed of olivine, orthopyroxene, clinopyroxene, and magnetite (?), and is extensively altered to coarse serpentine.

Olivine pyroxenite with minor patches of dunite and peridotite underlies East Island. A large dike of basic pegmatite cuts olivine pyroxenite in one of the nearby rock islands. An almost vertical planar structure, apparently non-graded layering, striking N.60°E. was noted on another of the small islands. Joint swarms strike N.50°-60°E. and 65°-70°N.W. The largest positive anomaly on the aeromagnetic map (plate 6) occurs between East Island and Kelp Island and probably indicates that the East Island ultramafic body is of major dimensions and has a large zone of hornblende pyroxenite.

PART IV. GENERAL THEORY

DERIVATION OF THE MAJOR ROCK UNITS

BASIC PEGMATITE

The basic pegmatite has plagioclase more calcic than that in the gabbroic rocks and therefore probably is not a late derivative of the gabbro. Experimental studies on synthetic systems indicate that, in normal magmatic processes, the latest plagioclase should be the most sodic. According to R. H. Jahns (personal communication), plagioclase in magmatic pegmatite bodies is at least as sodic as that in the major plutonic masses from which the pegmatite is most likely to have been derived.

The basic pegmatite is believed to be a late member of the ultramafic sequence because:

(1) It is a very much undersaturated rock in terms of its normative minerals.

(2) The pyroxenic ultramafic rocks are exceptionally rich in calcium and, thus, a plausible source for the pegmatite minerals.

(3) Hornblende in the pegmatite is essentially the same as that in the hornblendite, with the same distinguishing features (high Ca and Al).

(4) At Duke Island, the distribution of pegmatite shows a distinct spatial association with hornblende pyroxenite. In the gabbro, the pegmatite

is concentrated near the greatest development of the hornblende pyroxenite zone.

(5) Similar pegmatite occurs with most of the ultramafic bodies in southeastern Alaska (fig. 58).

PYROXENE GABBRO

The Duke Island pyroxene gabbro is a typical gabbro in terms of its chemical composition, mineralogical variations, and textural features. The relationships of mafic minerals and plagioclase shown in figure 4 are consistent with normal magmatic fractionation. Textural relationships show the sequence of mafic minerals in the discontinuous reaction series of Bowen (1928). Norite and two-pyroxene gabbro seem to be sufficiently intermixed to be considered differentiates of the same magma type. Textural relationships suggest that the noritic facies was the later to develop, but the reason for the absence of clinopyroxene is not understood. This, however, is a problem common to many gabbroic masses and requiring more study in both the field and laboratory.

HORNBLENDE GABBRO

The hornblende gabbro occurring near the granitic rocks is, on the basis of its areal distribution, mineralogy, and texture, fairly definitely the result of metasomatic alteration of pyroxene gabbro by the acid intrusions.

The hornblende gabbro associated with the ultramafic complex may be divisible into two groups by plagioclase composition (fig. 7). The type with the intermediate plagioclase could be a late product of the normal evolution of the same magma that crystallized pyroxene gabbro. However, this hornblende gabbro cannot be distinguished from that with the calcic plagioclase on the basis of areal distribution. As with the basic pegmatite, the type of hornblende gabbro with plagioclase more calcic than that in the pyroxene gabbro probably cannot be a late derivative of the magma from which the pyroxene gabbro crystallized. The hornblende gabbro is believed to be a metasomatic alteration of pyroxene gabbro by material derived from the ultramafic rocks because:

(1) Its textural features (streaky unevenness; porphyroblastic to pegmatitic facies) are that of a recrystallized, altered rock.

(2) It is spatially concentrated around the ultramafic rocks.

(3) It is intimately associated with basic pegmatite.

(4) It has plagioclase of composition intermediate between those of pyroxene gabbro and basic pegmatite.

Presumably, the alteration accompanied the introduction of the basic pegmatite.

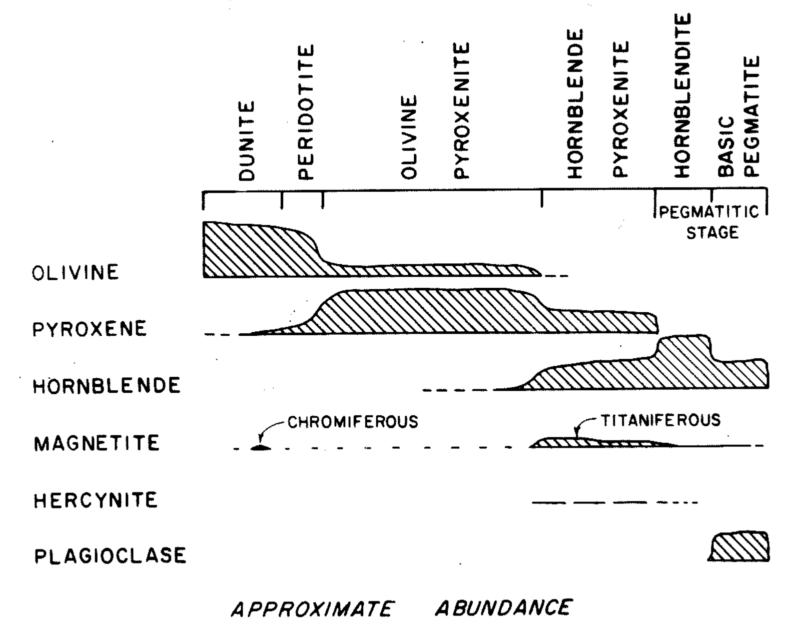
ULTRAMAFIC ROCKS

Most of the variations in complex bodies of igneous rocks probably arise because in the chemical systems involved, coexisting fluid and crystalline phases generally differ in composition when in equilibrium. In order to examine the origin of the major variations in the Duke Island ultramafic rocks in the light of this principle, the structural problems will be momentarily neglected, and the major rock units will be considered only as petrologic types. If their evolution has followed some systematic course, then it should be possible to arrange the units in a logical sequence.

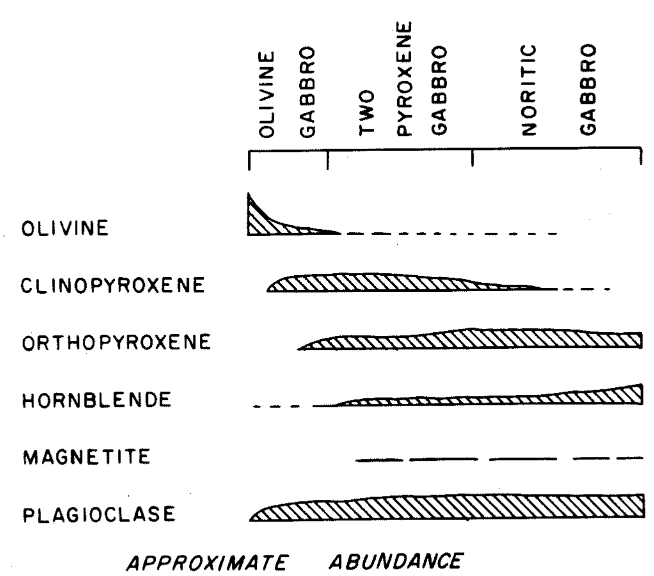
One possible sequence is:

> dunite-peridotite-olivine pyroxenite-hornblende pyroxenite-hornblendite-basic pegmatite

This will be recognized as a probable sequence in temperature of formation and might be expected to develop through fractional crystallization. However, other explanations are possible; for example, the same sequence in reverse might be accounted for by fractional fusion. Figure 56-A is a comparison of the mineralogy of the ultramafic rocks arranged in this sequence, and although the diagram is designed only to facilitate discussion and is not meant to imply age relations for the rocks, it is approximately an order of crystallization of the minerals as indicated by textures. For purposes of pointing out interesting



A. Mineralogical comparison of the Duke Island ultramafic rocks.



B. Mineralogical comparison of the Duke Island pyroxene gabbros.

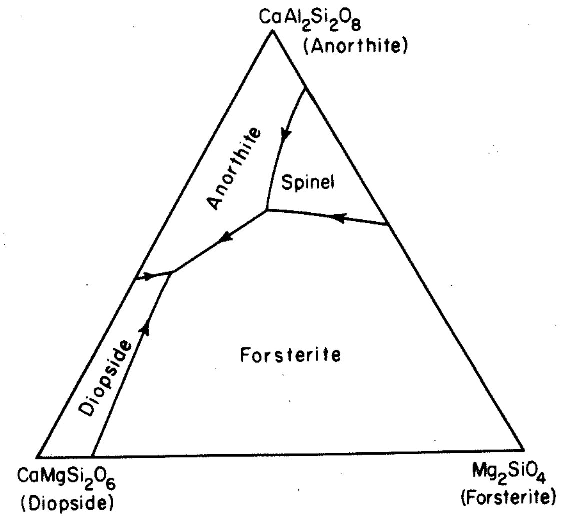
FIGURE 56

relationships and problems, the diagram will be considered an order of crystallization as might be developed because of crystallization differentiation.

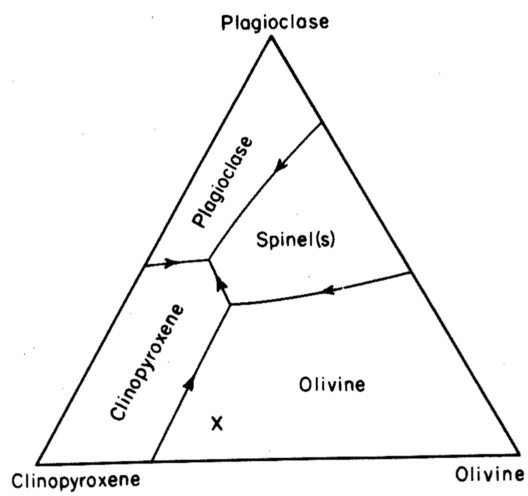
Olivine is shown to be the first mineral to crystallize and is then joined by clinopyroxene. As these minerals are very close to their respective magnesian end members, forsterite and diopside, one might expect the system Mg_2SiO_4-$CaMgSi_2O_6$ (Bowen, 1914) to give a fair approximation to the crystallization history of the natural rocks. If, in the synthetic system, a melt on the olivine side of the eutectic composition is subjected to crystallization differentiation, the crystalline precipitates that are formed are equivalent to dunite and an olivine pyroxenite with 87 per cent pyroxene. In the ultramafic rocks of southeastern Alaska the most common olivine-bearing units are dunite and a uniform olivine pyroxenite with 70-85 per cent pyroxene (H. P. Taylor, Jr., personal communication; and the author's observations). The similarity of the synthetic products and the natural rocks is impressive and suggests to the author that these two rock types are crystalline accumulates of a common magma undergoing crystallization differentiation.

Chromite, or chromiferous magnetite is indicated to appear in small amounts near the end of the crystallization of dunite. Chromite has not been recognized in the pyroxenic rocks, although chromium is just as abundant in olivine pyroxenite as in dunite (fig. 25). Probably chromium enters the pyroxene structure to such an extent that the magma from which the pyroxene-bearing rocks formed did not become saturated with the oxide mineral. Thus, clinopyroxene could be the reaction product of chromite.

The transition from olivine pyroxenite to hornblende pyroxenite is marked by the disappearance of olivine and the appearance of ilmenomagnetite, hercynitic spinel, and hornblende. This is a conceivable reaction relation during crystallization and can be illustrated as follows. Although the stable existence of magnesian spinel and diopside at liquidus temperatures has not been demonstrated in synthetic systems, Osborne and Tait (1952, p. 430), in discussing the system diopside-forsterite-anorthite (fig. 57-A), suggest that at some temperature below 1145°C calcic plagioclase and magnesian olivine become unstable together and that spinel and diopside can coexist. In more complex systems such as represented by the magma from which the ultramafic rocks formed and containing ferrous and ferric iron, the spinel would be hercynitic and along with magnetite probably persists in equilibrium with clinopyroxene to liquidus temperatures. If this is true, a projection of the liquidus fields of plagioclase, olivine, and spinel(s) might be as shown schematically in figure 57-B. Thus a magma starting from point X could by fractional crystalli-



A. The system diopside-forsterite-anorthite. After Osborne and Tait (1952)



B. Schematic representation of possible relations in a more complex system.

FIGURE 57

zation, give the sequence: olivine; olivine + clinopyroxene; clinopyroxene + spinel(s); clinopyroxene + spinel(s) + plagioclase. Clinopyroxene and spinel(s) are reaction products of olivine. If sufficient water was present, hornblende might be expected to appear in reaction relation to pyroxene.

Yoder and Tilley (1956) show that the curve for the disappearance of olivine in the crystallization of a tholeiitic basalt magma saturated with water also marks the appearance of hornblende. Thus hornblende could be the reaction product of olivine in the Duke Island ultramafic rocks although this is not entirely substantiated by textural relationships. These authors also show that pyroxene and hornblende can crystallize together, another phenomenon that is possible in some of the Duke Island ultramafic rocks.

One of the problems in the origin of parts of the pyroxenite and hornblende pyroxenite zones of several ultramafic bodies in southeastern Alaska is their relatively high, but very constant content (15-20 per cent) of magnetite. A possible explanation for the constancy provided by a hypothesis of crystallization differentiation is that this is the proportion of magnetite that crystallizes simultaneously with mafic silicates in the absence of feldspar. The quantity is not out of line with the limited experimental data available on systems involving magnetite and mafic minerals (Muan and Osborne, 1956). Magnetite layering in the hornblende pyroxenite (p.167) could be supporting geological evidence. Textural relationships are not altogether in agreement with this postulate, because the opaque oxide minerals are commonly interstitial to pyroxene. However, the extremely low temperatures indicated by the titanium content of the magnetite (pp. 202-204), if true, suggest that much of the ilmenite has unmixed from magnetite. This process might have begun during late magmatic stages so that, effectively, the oxide minerals were entirely recrystallized while the other silicates were still forming. Consequently, the textural relationships need not be completely diagnostic of the relative ages of the minerals.

As hornblende becomes more common at the expense of pyroxene, the magnetite content of the rock also decreases. This could reflect both the decrease of iron in the magma because of abundant crystallization of magnetite in the hornblende pyroxenite, and the greater entry of ferric iron into hornblende as compared to pyroxene (table 8, section 2).

Plagioclase is the last mineral to appear. As might be expected from the high calcium content of the pyroxenic rocks, the plagioclase is anorthite. It is notable that the hornblende in the basic pegmatite probably has a higher Na_2O content and Na:Ca ratio than the plagioclase. Apparently albite could not form because of low silica, a further indication of the low silica content of the fluid materials from which the pegmatite crystallized.

Thus most of the petrologic features of the Duke Island ultramafic rocks can be accounted for by crystallization differentiation following well-established principles that minerals begin to crystallize from a cooling magma when the magma is saturated with them and cease forming only when replaced by a reaction product. There has been more than one intrusion of magma involved in the evolution of the Duke Island ultramafic complex, but probably each one had at least part of this sequence of crystallization.

It has been stated that the ultramafic rocks of southeastern Alaska probably are all of the same origin (plate 1), and therefore the applicability of the sequence of differentiation given above should be tested in the other complexes. Figure 58 is a tabulation of the rock types occurring in the principle ultramafic bodies. Sources of information besides the author's own observations are: Walton (1951); Stebbins, (1957); Ruckmick (1957); and Taylor (personal communication). A hypothetical order of crystallization is given in the lower part of the diagram. The Union Bay complex is the only one with a complete suite of known rock types. Several bodies have only the last part of the sequence, suggesting that they may be products of "derivative" magmas.(As fractional melting can probably produce much the same variation in the composition of the liquid magma as fractional crystallization, "derivative" as used here is not meant to imply that the magma is the residual after the crystallization of the earlier rocks in the sequence. In some places, fusion during the formation of the magma may never have progressed to the point that olivine, for example, was melted.) The three bodies with only the first part of the sequence might be explained as not having undergone extreme fractionation. Moreover, all rock types in a complex may not be visible at the present erosion level.

A pyroxenite in which clinopyroxene is the only mineral in significant quantities is a common rock type in some of the complexes but is absent at Duke Island. The cessation in the crystallization of olivine in the Duke Island ultramafic rocks was explained by having any, or all, of magnetite, hercynitic spinel, and hornblende serve as reaction products. The absence of olivine from the pure pyroxenites cannot be accounted for in this way, and this may mean that the interpretation based on the Duke Island rocks is not correct. Some other explanation must be sought. Orthopyroxene is common in basic igneous rocks in reaction relationship to olivine but generally is absent from the ultramafic rocks of southeastern Alaska. Considerable solid solution of orthopyroxene and clinopyroxene can occur, particularly at elevated temperatures (Poldervaart and Hess, 1951; Atlas, 1952; and Boyd and

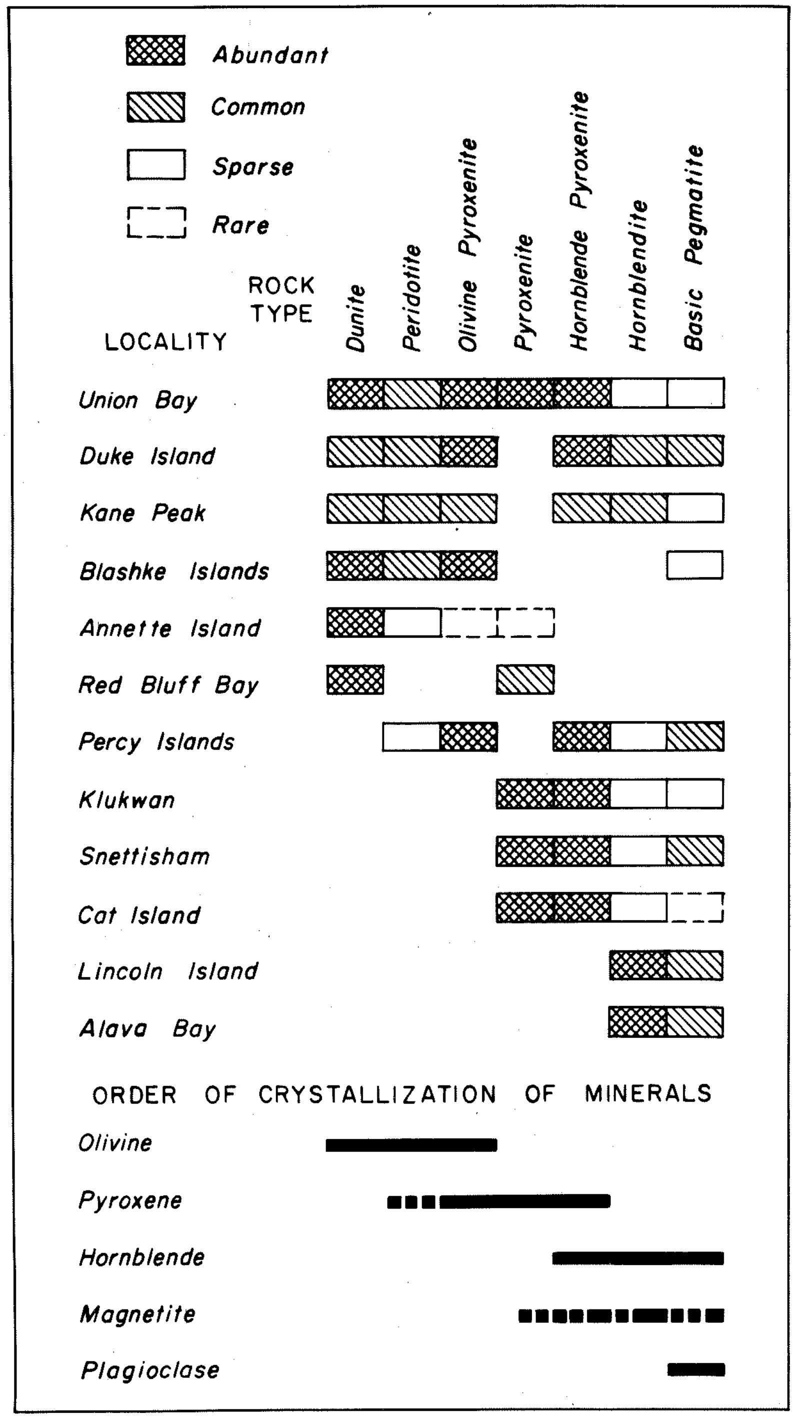


Figure 58.- Comparison of the relative abundance of ultramafic rock types at different localities in southeastern Alaska.

Schairer, 1957), and thus reaction enstatite might be entirely absorbed into the clinopyroxene. If this happened, a marked increase in the Mg + Fe:Ca ratio of clinopyroxene in pyroxenite as compared to that in olivine pyroxenite might be expected. However, chemical analyses of clinopyroxenes from both Duke Island (fig. 22) and Union Bay (Ruckmick, 1957, fig. 13) show that, if anything, this ratio decreases. The only other possible explanation that the author can suggest is that the boundary of the liquidus fields of clinopyroxene and olivine in the natural magma resembles that shown schematically in figure 59. At point A, the boundary changes from a cotectic to a peritectic, and diopside with little or no orthopyroxene in solid solution becomes the reaction product of olivine. The tendency of the more Fe-rich pyroxene to be richer in Ca may indicate that the Ca:Mg ratio of the magma increased proportionately faster than the SiO_2 content so that, eventually, all the MgO and Al_2O_3 had to go into diopside rather than CaO and Al_2O_3 going into plagioclase and MgO into enstatite. This explanation is, however, without experimental support, and for the present at least, there seems to be no substantiated way to account for the disappearance of olivine in all the ultramafic bodies of southeastern Alaska by processes related to fractional crystallization in one magma type. The same problem exists for derivation of a pyroxenite magma by fractional melting

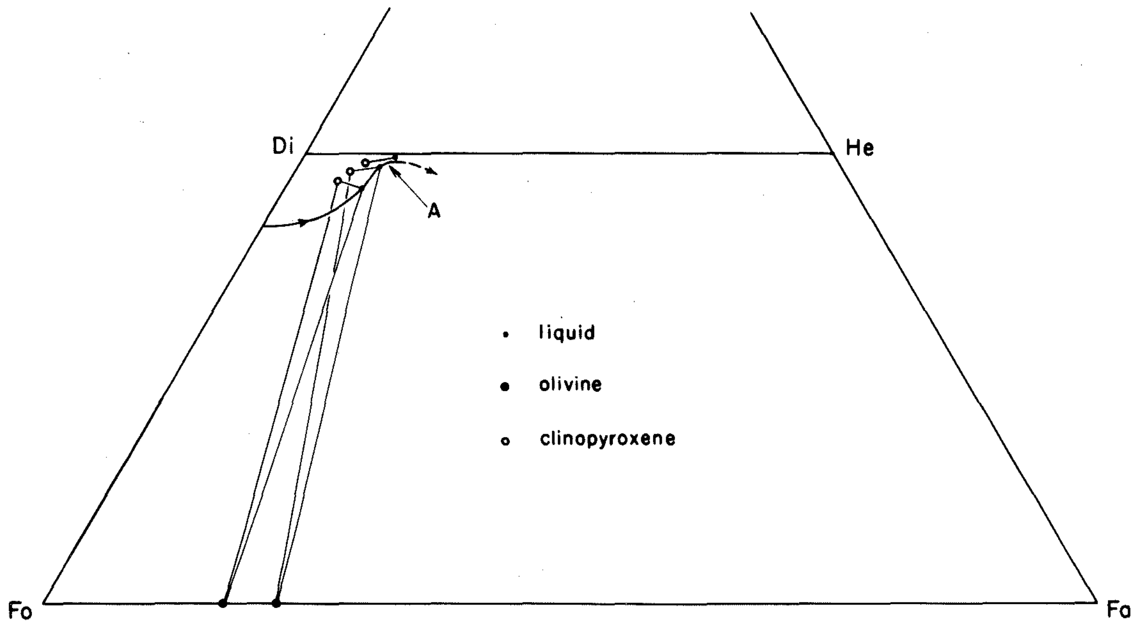


Figure 59.- Schematic projection of possible liquid-solid relations in the magma from which the ultramafic rocks formed.

of an olivine-bearing source, if the known phase relations have applicability at the depths where magmas form.

In summary, most of the variations in the ultramafic rocks of southeastern Alaska can be accounted for by simple processes of crystal-liquid equilibrium. If this type of process did not always apply, then those which did operate must have been closely kindred.

EVIDENCE AND NATURE OF AN ULTRAMAFIC MAGMA

RELATION OF ULTRAMAFIC AND GABBROIC ROCKS

Figure 56-B is a comparison of the mineralogy of the various types of pyroxene gabbro and, like figure 56-A, is approximately an order of crystallization of the minerals as indicated by textures. A complete representation of the crystallization history of a body of gabbro as large and varied as that occurring at Duke Island cannot possibly be made on so simple a diagram because the evolution of the body has undoubtedly been influenced by many local factors. The figure is therefore meant to apply in only a very general way.

If the gabbroic and ultramafic rocks were derivatives of the same magma, then conceivably, it should be possible to fit their respective crystallization histories together as a continuous sequence. In the ultramafic suite, this join would, presumably, be made at or near the end of the crystallization of hornblende pyroxenite as this is the "latest" ultramafic unit and the principal ultramafic rock to come in contact with gabbro. By the time this stage is reached, olivine has ceased to form, pyroxene is on the way out, and hornblende and magnetite are prominent phases. The early minerals in the gabbro are olivine, pyroxene, and plagioclase; hornblende and magnetite are late. To fit these crystallization histories together requires a complete repetition of sequence. This can probably be explained in several ways, but nevertheless, the facts remain; the sequence of appearance of minerals in the two rock groups were different.

The ultramafic rocks are extremely undersaturated in silica as compared to the pyroxene gabbro. This is illustrated particularly well by a comparison of their normative minerals. The norms for the analyses of the ultramafic rocks show not only olivine, but leucite, nepheline, and larnite. Anorthite is present in the ultramafic norms but, except in the basic pegmatite, apparently was not allowed to form as a modal mineral, in part at least because of low silica. Alumina and alkalies are primarily taken up by the mafic minerals.

Chemical data on the gabbroic minerals are not complete enough to compare with those for the ultramafic minerals. Although optical data are not fully reliable indicators of chemical composition, refractive indices suggest considerable overlap in the Fe:Mg ratios of the mafic minerals from the two rock groups and perhaps even lower Fe:Mg ratios in some of the gabbro minerals. This, if true, is an unusual condition for later differentiates. In figure 19, the intermediate indices of coexisting clinopyroxene and olivine have been plotted, and the trends of gabbroic and ultramafic rocks are compared. The data on gabbro are few, but the trends appear independent, and the gabbro shows much greater variation even under limited sampling. Figure 21 shows that the clinopyroxene in the gabbro has a different optical trend from that in the ultramafic rocks and gives no indication that the trends could ever join in a continuous sequence.

The olivine-bearing gabbros do not occur near the ultramafic rocks, as one might expect if the rocks were differentiates of the same magma. On the other hand, although norite is commonly in contact with hornblende pyroxenite, hypersthene does not occur in any of the ultramafic rocks of the main complex, even as an interstitial phase, and it is not present in the norms of the ultramafic rocks.

Plagioclase is abundant in the gabbro, and it might be expected to occur interstitially in the ultramafic rocks as a product of the crystallization of the interprecipitate magma if the rocks were derived from the same magma type. One would expect differentiates of the same magma to show gradational contacts. The ultramafic rocks are, however, virtually devoid of plagioclase, except in the pegmatitic stage, and the contact of the ultramafic and gabbroic rocks is consistently sharp and intrusive in appearance. This relationship might be explained in some places as the result of removal of interprecipitate magma, for example by filter pressing, but it seems fortuitous that the mechanism, whatever it might be, should always work, not only in the Duke Island complex, but in ultramafic bodies at Union Bay, Klukwan, Percy Islands, and in many of the other parts of southeastern Alaska.

Basic pegmatite probably is a member of the ultramafic suite. It cuts pyroxene gabbro. Some of the hornblende gabbro is believed to be a metasomatic alteration of pyroxene gabbro imposed by the ultramafic complex; this suggests that the pyroxene gabbro is older than the ultramafic rocks. Dikes of gabbroic rocks have not been observed in any of the ultramafic rocks. Some of the small bodies of non-feldspathic ultramafic material in the gabbro could be dikes, although they also could be inclusions.

The only known geologically-possible sources for the quartz inclusions in the peridotite zone of the Hall Cove ultramafic area are quartz veins and granitic pegmatite dikes in the pyroxene gabbro. This could indicate that a period of granitic intrusion separated the formation of the gabbroic and the ultramafic rocks.

The structural relationship of the gabbroic and ultramafic rocks is difficult to define because of the paucity of internal structure in the gabbro. However, the graded layering suggests that much of the ultramafic rock is stratigraphically above the gabbro, and hence, if the rocks are all differentiates of one magma, gravitational settling of crystals cannot be the only factor causing differentiation. At one significant locality along the northern boundary of the Hall Cove ultramafic area, about 1200 feet west of the Hall Cove fault, graded or rhythmic layering occurs in gabbro to within three feet of the contact (p. 37). The contact is knife-edge sharp and sufficiently irregular to rule out a fault. A small body of gabbro occurs in the hornblende pyroxenite border zone, and is believed to be an inclusion, indicating that the gabbro is the older rock. The hornblende pyroxenite has clinopyroxene, but no orthopyroxene or plagioclase. The gabbro is norite. The contact trends roughly parallel to the layering in both ultramafic and gabbroic rocks, and the graded layering in each type shows that the gabbro is underneath. It is inconceivable that these rocks could be products of one magma body differentiated in place, even though both types are accumulative. Except for the closeness of layering, the contact is not much different from that observed at other places along the boundaries of the ultramafic areas. Consequently, even though gabbro may be stratigraphically or physically on top of the ultramafic rocks at other places, it need not be a later differentiate of the same magma from which the ultramafic rock formed.

Thus the textural, mineralogical, petrological, and structural relationships of the gabbroic and ultramafic rocks do not support crystallization differentiation of one magma, or even one magma type, as a mechanism of formation of the rocks. Rather, they suggest that two very different types of magma have been involved, an early one of relatively normal gabbroic composition, and a later one of ultramafic composition.

PROBLEMS RELATED TO ULTRAMAFIC MAGMAS

There are, however, some unanswered problems regarding ultramafic magmas.

(1) The apparent absence of ultramafic lavas is sometimes cited as evidence against the existence of ultramafic magmas. Further evidence in the Duke Island area is the absence of ultramafic dikes outside the main complex. If the magma is as fluid as both experimental and field relationships indicate, it should enter any available fracture. Dikes or veins of pegmatitic hornblendite occur

in the gabbro in the vicinity of Kelp Island and Judd Harbor, but the small widely-scattered occurrences of olivine-bearing ultramafic rocks are not dikelike and are subject to other interpretations.

(2) The common association of ultramafic and gabbroic rocks suggests a close relationship. Almost all the ultramafic bodies in southeastern Alaska and British Columbia are accompanied by gabbro. If an ultramafic magma has existed, it must have come from a part of the earth's interior different from and presumably deeper than the source area of the gabbro magma. As the ultramafic magma would probably have the higher temperature, and as gabbroic and basaltic rocks are so common in the crust of the earth, one might speculate that temperature conditions severe enough to form an ultramafic magma would almost inevitably result in gabbro melts as well. Structural control for emplacement would then be the major factor in the association. However, this is undoubtedly casting the problem aside lightly.

(3) The high temperatures of ultramafic magmas should produce extensive metamorphic aureoles in many of the country rocks that surround ultramafic intrusions. Lack of metamorphism is not a problem at Duke Island. The metamorphic rocks against gabbro show mineral assemblages consistent with their environment, and the main areas of ultramafic rocks are completely enveloped by gabbro. At

their nearest approach, the ultramafic and metasedimentary rocks are hornblende pyroxenite and amphibolite, respectively, and this does not seem incongruous. The principal effect of the ultramafic magma on the gabbro is metasomatic and has resulted in the production of large amounts of hornblende, anorthite-rich plagioclase, and some clinozoisite and magnetite. This represents a major enrichment in Ca, H_2O, and probably Fe. The same mineralogy is developed to some extent in the metamorphic rocks nearest the ultramafic areas. The metasomatism is primarily localized around the probable upper part of the ultramafic body, as would be expected.

EVIDENCE BEARING ON THE COMPOSITION OF THE MAGMA

The chemical analyses and mineralogy of the ultramafic rocks show that the magma from which they crystallized was exceptionally low in silica and alkalies as compared to common magmas, and rich in magnesia and lime. The most abundant ultramafic rock at Duke Island is olivine pyroxenite, and this rock forms the major part of the layered series. It is believed to have formed by cotectic crystallization of olivine and clinopyroxene with continuous accumulation of crystals by gravity settling. The presence of graded layering containing primary pyroxene in the dunitic zones indicates that, even during the crystallization of dunite, the composition of parts of the magma periodically touched the cotectic boundary and, probably, that the entire liquid fraction of the magma was never far from that boundary. Hence it would seem that even the most dunitic rocks formed from a magma with appreciable calcium, and thus a pure dunite melt is not believed to have existed at Duke Island. Alumina is low compared to feldspathic rocks but shows notably large concentrations in pyroxene and hornblende. The low Fe:Mg ratio of the mafic silicates and the relatively large quantity of magnetite suggest that much if not most of the iron was in the ferric state. This might also be deduced from the trace element data (p. 95).

The concentration of water in the magma is important but, unfortunately, cannot be estimated reliably. Pervasive interstitial hornblende in the rocks of the layered series indicates that the magma did contain significant water throughout its crystallization. The basic pegmatite and calcic hornblende gabbro are believed to be products of material expelled from the ultramafic magma, and if this interpretation is correct, probably large amounts of water have been lost from the ultramafic complex. It is noted that the interpretations that the ultramafic magma contained considerable water and that it lost water are both contrary to interpretations made by Ruckmick (1957) for the Union Bay complex. That such different conclusions could be reached points out a major difference between the rocks in the two localities; at Duke Island, all rocks generally are more hydrous.

TEMPERATURE OF THE ULTRAMAFIC MAGMA

Temperature estimates for a magma whose composition is not definitely known and for which good geothermometers are lacking must be tenuous, to say the least. However, for the ultramafic rocks, maximum temperatures are of most concern, and for this some indirect evidence is available.

Even the dunite is believed to have crystallized from a magma close in composition to the cotectic boundary of clinopyroxene and olivine. The clinopyroxene and olivine are close to the ideal compositions of diopside and forsterite, respectively, and as the latter two are the high temperature end members in their respective solid solution series, their eutectic temperature should represent a fair approximation of the maximum temperature of the magma. The temperature of the eutectic is 1387°C (Bowen, 1914). This is 500°C lower than the melting temperature of fosterite and, probably, at least 400°C lower than the melting point of the typical olivine (Fa_{17}) in the ultramafic rocks of southeastern Alaska. This lowering is a major step toward explaining the ultramafic rocks at temperatures normally considered to be geologically reasonable, and is a second point in favor of the derivation of dunite from a magma in which the melt fraction had a composition close to the liquidus boundary between the fields of olivine and pyroxene. Yoder (1955) has shown that the melting temperature of diopside is lowered from 1391°C to 1282°C at 5000 bars water pressure. Although the ultramafic magma may not have contained an amount of water equivalent to this pressure, it did contain some. Furthermore, it must have been significantly different from a pure diopside-forsterite melt, because the rocks have other constiuents, in particular FeO, Fe_2O_3, and Al_2O_3, all of which would reduce the melting temperature. Thus, the author believes that 1387°C is the absolute maximum temperature of the magma and that the actual maximum may have been lower than this by 100°C or more.

Buddington, Fahey, and Vlisidis (1955) give an empirical curve for the solubility of TiO_2 in magnetite as a function of temperature, where TiO_2 is present as ilmenite both in solid solution in the magnetite and as separate grains in the rock. As with the application of all solid solution geothermometers, attainment of chemical equilibrium must be assumed. The presence of separate grains of ilmenite is required to prove that magnetite was saturated with ilmenite for the temperature at which equilibrium was established. If the assumptions and the curve are correct, then the temperature obtained is that at which the rock was quenched. For igneous rocks this probably is not the temperature at which they crystallized initially, but one can only assume the difference is not great.

As described previously (pp. 80-81), the Duke Island hornblende pyroxenite contains magnetite and ilmenite as discrete grains, and hence the magnetite should be useful as a thermometer. Unfortunately, none has been separated and analysed, and only available analyses are from drill core and are for acid soluble iron and TiO_2 in the total rock. These analyses, however, do illustrate a point. The average content of soluble Fe is 11.8 per cent and of TiO_2 is 1.47. If it is assumed that approximately half the TiO_2 is in hornblende and pyroxene, as shown by the analyses of these minerals, and the remainder is in ilmenite and magnetite, then the rock has 17.1 per cent ilmenite+magnetite with 4.4 per cent TiO_2. (Buddington, et al, list TiO_2 contents of 3.91 and 3.6 per cent for magnetite in comparable ultramafic rocks in southeastern Alaska occurring at Haines and Klukwan respectively.) Even if all the ilmenite was in solid solution in the magnetite, the maximum temperature that the TiO_2 content could indicate is 750°C according to the curve of Buddington, et al. As much of the ilmenite is in separate grains, the temperature of quenching probably is several hundred degrees lower. One caution must be emphasized in addition to the obvious shortcomings in the data. The solution of hercynitic spinel in magnetite as indicated by its presence in exsolution lamellae (p.81) might affect the solubility of ilmenite sufficiently that the curve of Buddington, et al, is not applicable to these rocks. However, the effect would have to be great to negate the indication that the temperature at which crystallization, or recrystallization, ceased was relatively low for igneous rocks.

A possible indication of the minimum temperature at which magmatic activity still occurred in the ultramafic rocks has been observed in hornblendite at Alava Bay, Revillagigedo Island. In this body, disseminated magnetite occurs in tiny octahedra that are exceptionally well-formed as compared to magnetite in most of the ultramafic rocks. Accompanying it in some places is a light dissemination of tiny pyrite cubes. It seems possible, although it cannot be proved, that these minerals could have formed together during late magmatic stages. Kullerud and Yoder (1957) have shown that the dissociation curve of pyrite begins at 743°C and about 180 psi total pressure and increases to 815°C at 75,000 psi. The pyrite in the ultramafic rock could not have formed at temperatures above this curve. Pyrite and magnetite do not coexist in equilibrium above a temperature of about 675°C (Kullerud, 1957), and if these minerals have formed in equilibrium from late fluids of the ultramafic magma, then magmatic activity was still going on at this temperature.

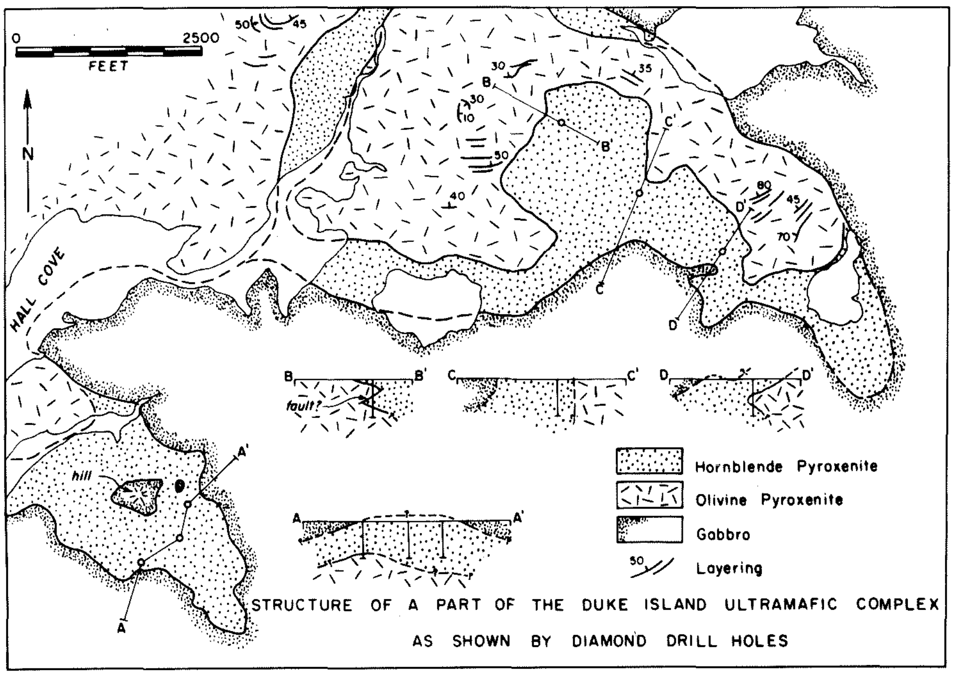
In conclusion, limited evidence indicates that the temperature of the ultramafic magma was not unreasonably high by commonly accepted geological standards and that magmatic activity and other processes effective in determining the final mineralogical features of the rocks persisted to relatively low temperatures. The broad range in the temperature of formation of the rocks may be of major significance.

EVIDENCE OF THREE-DIMENSIONAL STRUCTURE

DRILL HOLE DATA

The low relief of Duke Island affords little opportunity for observation of the structure of the ultramafic rocks in three dimensions, and less direct evidence has to be relied upon. The magnetite-bearing rocks have been investigated with nine diamond drill holes, each to a depth of 500 feet. Data obtained from six holes in the hornblende pyroxenite border zone of the Hall Cove ultramafic area are summarized in figure 60.

One of the holes in section AA' passed out of the border zone rock into olivine pyroxenite at a depth of 435 feet. This fact and an apparent capping of gabbro on a nearby hill suggest that the contact of the ultramafic and gabbroic rocks is relatively flat in this area. This is part of the evidence on the displacement of the Hall Cove fault, indicating that the east block of the fault is down relative to the west block.



STRUCTURE OF A PART OF THE DUKE ISLAND ULTRAMAFIC COMPLEX AS SHOWN BY DIAMOND DRILL HOLES

FIGURE 60

The holes in sections BB', CC', and DD' show that the contact of hornblende pyroxenite and olivine pyroxenite is irregular. Relationships are not consistent but may suggest that hornblende pyroxenite is the upper unit. No concordance between contact and adjacent layering is evident.

The three other holes were drilled in the central part of the main olivine pyroxenite zone in the Judd Harbor ultramafic area to investigate strong magnetic anomalies. The main rock type in the core from all the holes is olivine pyroxenite. One hole intersected a large dike of basic pegmatite and several small ones. Much of the olivine pyroxenite in this hole was altered to magnetite-bearing hornblende pyroxenite or hornblende-olivine pyroxenite. The other two holes encountered only minor amounts of pegmatite and hornblende pyroxenite. The amount of magnetite in the rocks is surprisingly small considering the magnitude of the aeromagnetic anomaly. Drilling failed to show an appreciable body of magnetite-rich rock within 500 feet of the surface, even though the holes were located on prominent anomalies of ground surveys. Polarization of the magnetite may be important. It was noted that the behavior of compass and dip needle was much more erratic in this area than in the magnetite-rich border zone.

AEROMAGNETIC MAP

The aeromagnetic map shows a general northwest grain. Magnetic lows occur over dunite and peridotite, and as expected, marked positive anomalies are present over hornblende pyroxenite. One exception to the rule is the absence of a marked anomaly over the broad bulge in the hornblende pyroxenite zone on the west side of the Hall Cove ultramafic area, about three fourths of a mile south of North Hill, even though the magnetite content of the rock here is fairly normal. However, the north contact of the hornblende pyroxenite makes a pronounced V-shaped bend to the south where it is crossed by a stream, and this may indicate that the zone is shallow in this area.

The southwest quarter of Duke Island, including the two main ultramafic areas, has a much higher than average total magnetic intensity. Outside the ultramafic areas, this high matches almost exactly the distribution of hornblende gabbro and basic pegmatite. It may mean any or all of the following:

(1) A higher content of magnetite in hornblende gabbro and basic pegmatite than in pyroxene gabbro (or, perhaps, more strongly polarized magnetite).

(2) Hornblende pyroxenite is present beneath the hornblende gabbro and basic pegmatite.

(3) The entire ultramafic suite extends beneath the

hornblende gabbro and basic pegmatite, joining the rocks exposed in two main areas into one body.

(4) The topographic effect of Mount Lazaro on the magnetic intensity is appreciable. However, this effect presumably is largely eliminated from the map.

GEOLOGICAL INFERENCES AND SUMMARY

The Judd Harbor and Hall Cove ultramafic areas are believed to be the outcrop of one large ultramafic body at depth because:

(1) The basic pegmatite is believed to be a member of the ultramafic suite and the hornblende gabbro to be the product of metasomatic alteration of the pyroxene gabbro accompanying the introduction of the basic pegmatite. Basic pegmatite and hornblende gabbro are so abundant between the two ultramafic areas that the presence of considerably more ultramafic material at depth is suggested. This accords with both aeromagnetic and drill hole data.

(2) The rocks in both areas show extensive layering. This is a sufficiently uncommon feature to indicate that it may all have developed in one magma chamber.

The concept that the ultramafic rocks exposed in the two areas are joined at depth is the main reason for believing that the east block of the Hall Cove fault has been dropped relative to the west one.

Other inferences which might be made and which will be discussed in more detail later are:

(1) The intrusion has a floor or bottom on which the layering originally accumulated.

(2) Hornblende pyroxenite commonly forms a continuous border zone. Where visible the zone is widest around the upper part of the layered rocks, and judging from the distribution of aeromagnetic anomalies, this same relationship could exist at depth if the assumption that one large body is represented is correct.

(3) The plane of the layering is commonly at a high angle to the hornblende pyroxenite zone and, hence, to the outer boundary of the ultramafic complex. Evidence in other layered igneous complexes suggests that layering generally has flatter dips than adjacent contacts, being banked against them (Wager and Deer, 1939; Rossman, 1954; and Wilson, 1956). Where evidence of other relationship is lacking, this one will be assumed in the Duke Island ultramafic complex. This requires a conical or ellipsoidal shape for the body. Certainly,

considering the thickness of exposed layered rocks compared to the apparent lateral extent of the ultramafic rocks, it is difficult to imagine the body as being very sill-like.

In the cross sections and block diagrams presented and discussed later, no pretense is made to defend the details of configurations shown at depth beyond the assumptions given above and those fundamental to the interpretation. The cross sections can only be diagrammatic representations of hypothetical interpretations. The depths shown for the ultramafic body are the minimum required by the estimated thickness of layered rocks. That the body may extend to greater depths is entirely possible, but lesser ones are unlikely. The feeders to the ultramafic mass are shown to be dikelike because the Duke Island ultramafic complex is apparently elongated and the main ultramafic rocks in the vicinity of Duke Island occur in a line between East Island and the Percy Islands. Fracture control is suggested.

The East Island ultramafic body off the southeast corner of Duke Island may also be joined to the main complex. However, data to support this conjecture are so limited that the possibility will not be considered further.

EVOLUTION OF THE MAJOR ZONING IN THE ULTRAMAFIC ROCKS

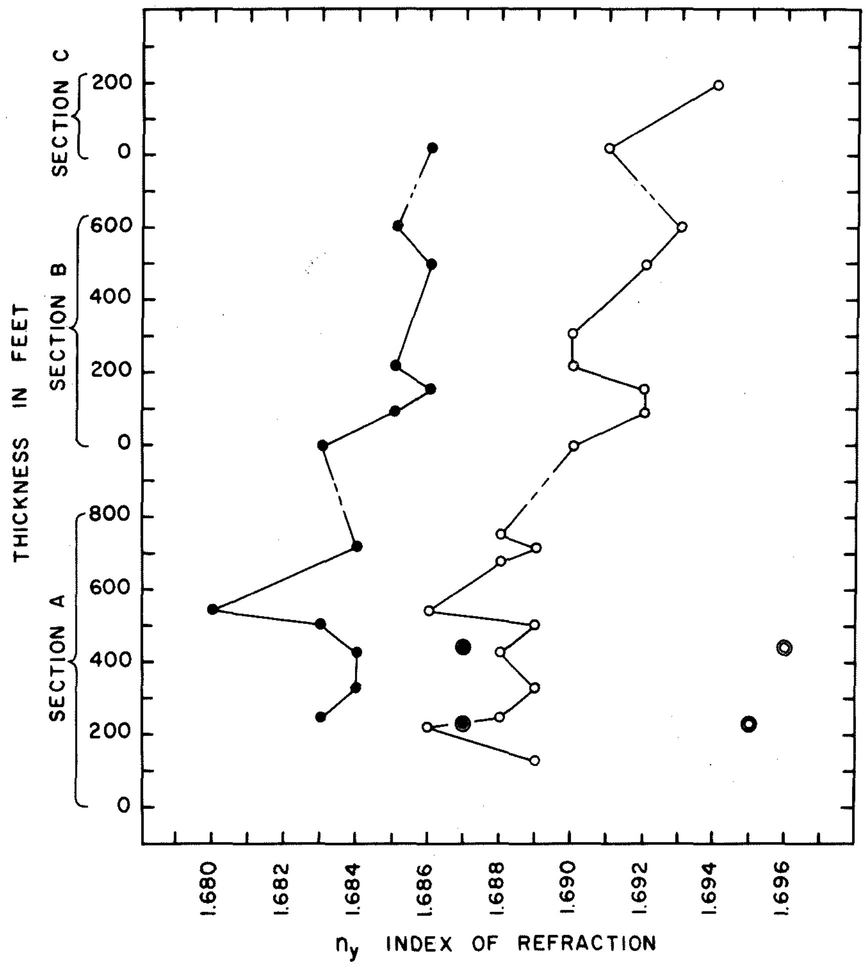
POSSIBLE MECHANISMS

Differentiation by Fractional Crystallization

Crystallization differentiation is probably the best substantiated and most widely accepted process for the production of variations in igneous complexes. One mechanism whereby this process can be effected is the removal of crystals from the melt by gravitational settling. Gravitational layering is extensively developed in the Duke Island ultramafic complex, and it has been shown that most of the petrographic features of the major rock types can be accounted for by crystallization differentiation. This process therefore deserves prime consideration in the examination of the major zoning.

The Duke Island ultramafic complex clearly is not a simple differentiated sill or stratiform sheet, nor is it probable that the complex could have had such a form but was subsequently deformed. All the rocks might have formed from one magma, but more than one injection of this magma or its derivatives (p. 187) is required. The features along the northern boundary of the peridotite zone in the Hall Cove ultramafic area must be the result of an addition of magma. On the other hand, the southern olivine pyroxenite zone in this area may be conformably above the peridotite and could, therefore, be a later differentiate of the same intrusion from which the peridotite formed. The hornblende pyroxenite is stratigraphically above the olivine pyroxenite in only a few places, and if it is a differentiate, the distribution must largely be controlled by some factor other than gravitational accumulation.

An expectible result of differentiation, judging from the Skaergaard intrusion (Wager and Deer, 1939) and from general theoretical considerations, is a progressive change from early to late stages in the composition of minerals in solid solution series. At Duke Island, a slight increase in the Fe:Mg ratio of pyroxene and olivine is indicated in the sequence dunite and peridotite-olivine pyroxenite-hornblende pyroxenite, and other regular variations in both major and trace elements are evident in these rocks (table 8, fig. 25). However, indices of refraction suggest considerable overlap for the different units (fig. 20). An optical study was made of changes in the composition of minerals belonging to solid solution series and occurring in the layered series ("cryptic layering" by the terminology of Wager and Deer, 1939, p. 37). Figure 61 shows the intermediate indices of refraction of pyroxene and olivine in the Hall Cove peridotite zone for specimens collected along the lines of measurement of lithologic sections A, B, and C. The increase of indices



- ● Clinopyroxene
- ● *Clinopyroxene in olivine pyroxenite fragment.*
- ○ Olivine
- ◎ Olivine in olivine pyroxenite fragment.

Figure 61. - Variation in the n_y refractive index of clinopyroxene and olivine with elevation in the peridotite zone of the Hall Cove ultramafic area as sampled along the lines of lithologic sections A, B, and C.

with elevation are small, and reversals are frequent. Figure 62 is a map showing the areal variation in the intermediate index of clinopyroxene in both the Judd Harbor and Hall Cove ultramafic areas. Comparison of the plot with the pattern of the zoning and layering (plate 1) shows that the indices generally are higher in olivine pyroxenite than peridotite, but if any change takes place with elevation in the layered series, it is not regular or not detectable by the method used.

The absence of cryptic layering in layered igneous complexes is a major problem not confined to the Duke Island ultramafic complex. Rossman (1954) reports an absence of cryptic layering for a layered basic intrusion in the Fairweather Range (Mount Crillon), southeastern Alaska, in which a 32,000-foot thickness of gabbro shows no significant change in the composition of either pyroxenes or plagioclase. Brown (1956) found no change in the minerals from 2600 feet of layered feldspathic peridotite and allivalite exposed on the Island of Rhum. The explanation may be that the composition of the crystallizing magma has been maintained by the introduction of new magma; this is essentially the explanation chosen by Brown. Many of the small fluctuations in the Duke Island layered series could be accounted for by a convection hypothesis, which calls for continual introduction of different parts of the magma body into the zone of

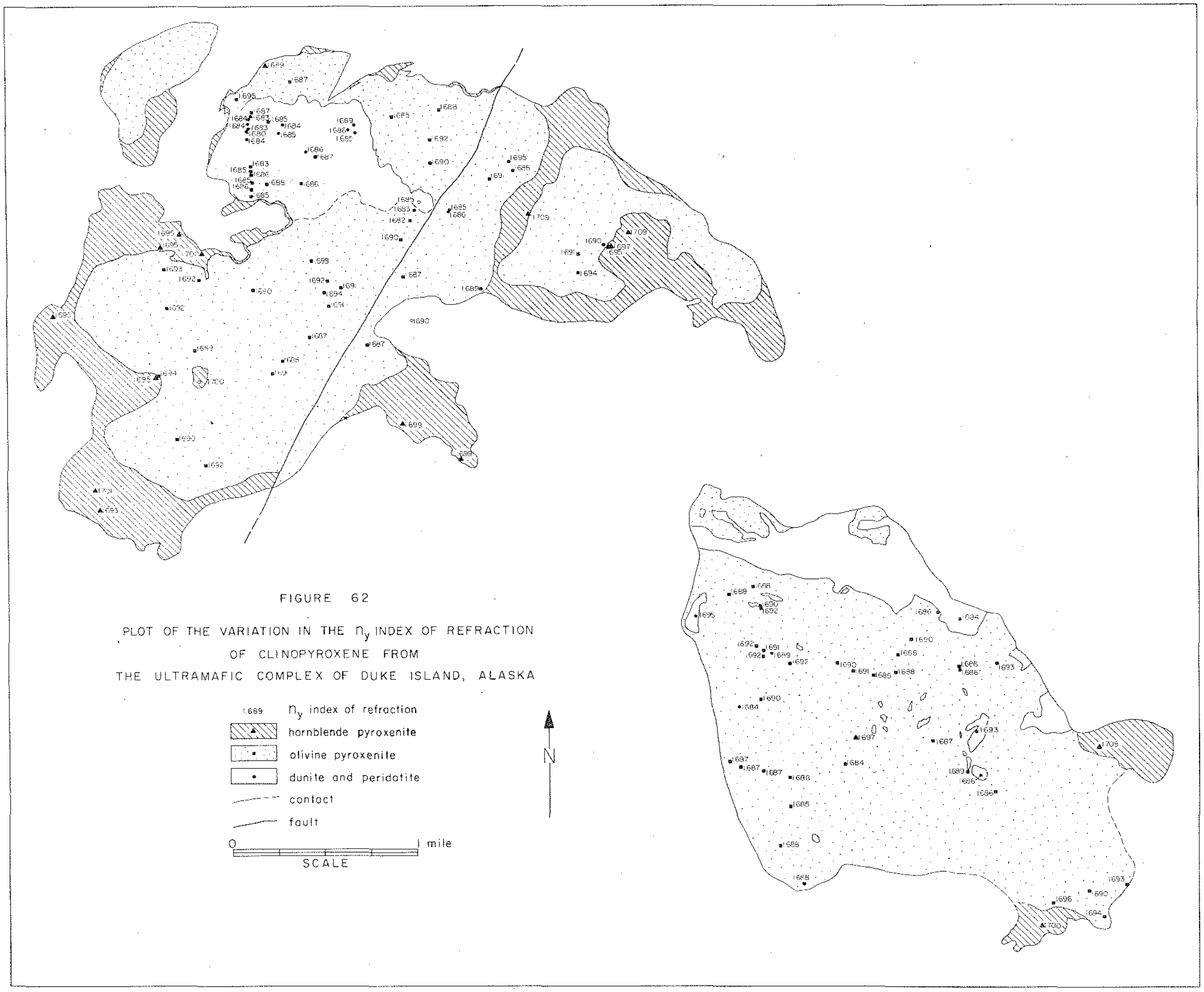


FIGURE 62

PLOT OF THE VARIATION IN THE n_y INDEX OF REFRACTION OF CLINOPYROXENE FROM THE ULTRAMAFIC COMPLEX OF DUKE ISLAND, ALASKA crystallization. Thus inhomogeneities in the magma might be reflected in the rocks. One major addition of magma in the Hall Cove part of the Duke Island ultramafic complex is certain, having resulted in the formation of the peridotite zone, and a few minor additions might be justified by structural evidence, but for the main part of the layered series evidence of repeated intrusions is lacking. Rossman (personal communication, 1956) makes a similar observation on the Mount Crillon gabbro.

From theoretical considerations of the distribution of water in magma bodies, Kennedy (1954, p. 496) has suggested that cryptic layering may not develop in layered intrusions because the crystals formed near the floor and were always in contact with the liquid from which they formed. Evidence will be presented to show that the crystals of the Duke Island complex probably did not form near the floor, and that the magma underwent repeated stirring or convection. Convection would not permit the distribution of water required by Kennedy's hypothesis.

Progressive oxidation of the magma during crystallization might prevent an increase in the Fe:Mg ratio of pyroxene and olivine. The Fe^{++}:Mg ratio of the magma could be held constant and the ferric iron produced would be largely unacceptable to the mafic silicates. However, this mechanism probably requires the production of enstatite (at least in solid solution in clinopyroxene) at the expense of olivine, a change that is not observed. Such a process could not account for a lack of variation in plagioclase. Possibly, if much of the iron was in the ferric state initially, little variation in the Fe^{++}:Mg ratio would develop. This would be compatible with the low iron content in the mafic silicates of all of the Duke Island ultramafic rocks. Variation should be shown by other elements, and a detailed study of the trace elements in the rocks of the layered series might be informative in this respect. By this explanation, magnetite crystallizing from the interprecipitate magma should become more common with stratigraphic elevation. The modal analyses cannot be used to test this possibility because they include magnetite that has resulted from serpentinization. The aeromagnetic map shows an increase in total magnetic intensity over the upper part of the layered series exposed in the Hall Cove ultramafic area, but this could be due to several reasons.

The iron-rich hornblende pyroxenite border zone has its greatest development near the upper parts of the layered series, even though it does not conform with the layering. This suggests that loss of iron and, perhaps, other components to the border rocks may have occurred during the crystallization of the magma and prevented the development of marked cryptic layering. This possibility will be considered further in the discussion of the origin of border zone.

Multiple Intrusions

A mechanism based on separate intrusions of magmas of different composition can, in theory, account for almost any structural situation that one might encounter in an igneous complex. For this reason, it is difficult to criticize--it is seldom contradicted by field evidence. Multiple intrusion has undoubtedly played an important role in the evolution of many igneous complexes, and the field relationships are incontrovertibly in its favor in the Duke Island ultramafic complex (p.126). However, unless positively demonstrated with structural criteria, appealing to such a mechanism is, in a sense, a last resort because it simply shoves the major problems of rock variation down into the realms of speculation, the regions where magmas are formed. To some extent, the author has already appealed to the last resort to explain the relationships of the gabbroic and ultramafic rocks. The ultramafic bodies of southeastern Alaska are all so similar in composition, mineralogy, and structure that, if magmas of different composition are involved one might reasonably expect that the magmas could be derived from a single parent material by some simple process of chemical fractionation. The most probable processes are two;

fractional crystallization and fractional fusion.

Ruckmick (1957) has chosen a mechanism of multiple intrusion to explain the zonal structure of the Union Bay ultramafic complex. Melts of three different compositions are called upon which are, in order of intrusion, equivalent to: (1) pyroxenite, or pyroxenite with magnetite; (2) olivine pyroxenite; and (3) dunite. Each magma is intruded into the central part of the preceding one, which has, by that time, largely crystallized. The control for the sequence of intrusion is, supposedly, the order of melting, the later magmas having the higher temperatures. If gabbro is considered part of the complex, then it represents another magma earlier than the pyroxenite.

The magmas that Ruckmick has chosen are not types generally accepted by igneous petrologists and are extremes in composition with respect to one another. Ruckmick recognizes the problems connected with the derivation of such magmas and attempts to explain their formation by fractional melting in simple eutectic or cotectic systems. However, the problems still exist. It is difficult to imagine how a liquid equivalent to dunite and containing so little calcium could be derived from the same source area as the calcium-rich pyroxenic magmas; a dunite melt probably cannot be derived from anything but dunite. Certainly, a dunite magma cannot be parent to the pyroxenic magmas. A pyroxenite magma that does not crystallize olivine is almost as problematical. Thus, it would seem that Ruckmick must postulate not only three or four magmas but also three or more source areas. This requires a great deal of a process supposedly capable of explaining a similar distribution of similar rocks in many ultramafic complexes.

The phase diagram for the system Mg_2SiO_4-$CaMgSi_2O_6$ (Bowen, 1914) shows a continuous liquidus between the eutectic point and pure forsterite; there is no reason to suspect liquid immiscibility in this range. Ruckmick postulates liquid magmas equivalent to olivine pyroxenite and dunite. Judging from the Mg_2SiO_4-$CaMgSi_2O_6$ phase diagram, these magmas may differ in temperature by more than 400°C, which is an unlikely gap. If these magmas did exist, then one might expect a complete suite of intermediate (peridotitic) melts whose abundance would be even greater than that of the dunite melt because they have lower temperatures. However, peridotite is relatively rare among the ultramafic rocks of southeastern Alaska. In the author's opinion, the best explanation for this is that peridotite liquid, if it did exist, almost invariably underwent crystallization differentiation so that dunite formed while it crystallized on the olivine liquidus, and olivine pyroxenite formed while it was on the liquidus boundary between the fields of olivine and pyroxene. Crystallization differentiation because of gravity settling of crystals is known to occur in shallow, rapidly cooled intrusions (e.g. the Palisades sill and the Shonkin Sag), and calculations of rates of crystallization and of gravitational settling velocities of crystals (p. 291) suggest that this is an almost inevitable process in any slowly-cooled, deep-seated ultramafic melt. Thus it seems probable that, even if a dunite melt did exist, a considerable part of the dunite occurring in the ultramafic rocks has formed by differentiation of a pyroxenic magma. If some of it has, which part? Perhaps it all has, and a dunite liquid never did exist.

Formation of a border zone by multiple intrusions presents further problems, but before these can be discussed a definition of the border zone has to be formulated because it differs somewhat in different complexes. In the discussions to follow, the border zone will, by the author's definition, include those ultramafic rocks occurring in the peripheral parts of a complex and showing the following features.

(1) They are for the most part macroscopically free of olivine.

(2) They generally contain pyroxene with distinctly higher refractive indices than those of pyroxene in the interior zones. Judging from the available chemical analyses, this means a slightly higher Fe:Mg ratio, Ca:Mg+ Fe ratio, and Al_2O_3

content.

(3) They commonly contain 10-20 per cent ilmeno-magnetite in a uniform dissemination.

(4) They generally contain more hornblende than the interior zones.

With the possible exception of (2), no one feature is a unique characteristic, but at least two of them are invariably present. They are the features which make the border zone problematical. At Duke Island the border zone is almost entirely hornblende pyroxenite. By this definition, the border zone in the Union Bay complex is the combined zones of pyroxenite and hornblende pyroxenite. It is clear in Ruckmick's discussions that much, although not all, of these units constitute most of the solid products of the first intrusion (pyroxenite magma) in his hypothesis. However, the composition of this intrusion was chosen to explain features (1) to (3) whereas the production of hornblende is attributed with good reason to contamination of the magma with water and other materials from the wall rocks. The size and location of the main magnetite-bearing part of the pyroxenite and hornblende pyroxenite zones in the Union Bay complex makes this explanation a definite possibility. However, if a hypothesis of multiple intrusions of this sort is to have general applicability, it should in this author's opinion explain as separate intrusions those parts of border zones showing features (1) to (3) in other ultramafic complexes.

The principal examples in southeastern Alaska of ultramafic bodies with border zones are the complexes at Union Bay, Duke Island (particularly as exposed in the Hall Cove ultramafic area), and the Percy Islands (Stebbins, 1957). Similar border zones of pyroxenite, or hornblende pyroxenite, or both occur in two ultramafic bodies in southern British Columbia (Aho, 1956; Camsell, 1913). In every example the zone is remarkably continuous, and even considering them individually, the author has difficulty imagining them formed simply as the earliest in a series of intrusions. Furthermore, no one has reported a definite transection of the border zone by, for example, dikes of either olivine pyroxenite or dunite. This is very strange if multiple intrusion is the cause of the zoning, because the emplacement of the later magmas must have caused some expansion of the magma chamber; this expansion is well shown in those parts of the Duke Island complex where definite evidence of multiple intrusion can be demonstrated. Such continuous border zones seem to imply some physicochemical control in the differentiation of the complex by the contact. Another interesting feature apparent on the maps of each of Duke Island, Union Bay, and the Percy Islands is that the border zone is narrowest and has the least magnetite where dunite is most common. The Blashke Island complex, although well zoned both in terms of its major rock units and the composition of its mineral phases, has a major proportion of dunite and lacks a border zone of olivine-free pyroxenite (Walton, 1951). These relations seem too consistent to be accounted for by multiple intrusion. They are much more in accord with some mechanism of differentiation in situ.

The available evidence at Duke Island indicates that hornblende pyroxenite is younger than olivine pyroxenite, and in some places, it is even a reaction or alteration product of olivine pyroxenite (p. 121 and p. 169). If these age relations are valid and the border zone does represent a separate magma, then this magma would be the latest, and its emplacement would have to be controlled by the contact. Although the intrusion might resemble a cone sheet or ring dike, such precise control seems unlikely, and the mechanism would not account for the thinning and change in character of the zone near dunite. Furthermore, the contact of hornblende pyroxenite and olivine pyroxenite is not a structural break (p.121) and does not look like most intrusive contacts.

Solid Intrusion

Intrusion of ultramafic rocks in a largely solid condition has been repeatedly advocated by Bowen (Bowen, 1928, p. 167; Bowen and Schairer, 1933; Bowen and Tuttle, 1949) and is a mechanism widely accepted for some ultramafic types (Hess, 1954, p. 401; Ross, 1954, pp. 723-728).

At Duke Island, the layering indicates that the ultramafic rocks have crystallized from a magma. If there has been any solid intrusion, the amount of internal flow could not have been great because most of the layering is relatively undeformed. The magma must have crystallized at or near the present position of the complex relative to its country rock.

In the rocks of the Hall Cove ultramafic area, the only place where intrusion of predominantly solid material may have occurred is in the peridotitic dikes (pp. 137-143) Considerable doubt exists even here, and certainly, if the dikes are solid intrusions, the conditions leading to their emplacement probably were very different from those visualized by Bowen and others.

The rocks of the Judd Harbor ultramafic area are folded and may be largely bounded by faults. If faulting and deformation occurred simultaneously--a good possibility--then the movements could constitute a kind of solid intrusion. The relationship of dunite to olivine pyroxenite suggests that dunite is a discrete intrusion whose emplacement caused the folding in the olivine pyroxenite. The small areas of olivine pyroxenite occurring along the north side of the main zone of dunite have not definitely been shown to be the outcrop of blocks buried during the accumulation of the olivine-rich layered series as in the Hall Cove ultramafic area. Thus the dunite could have been the original bottom part of a differential layered complex and could have been squeezed upward into olivine pyroxenite shortly after the main accumulation of crystals in the complex. More internal deformation in the dunite than is evident might be expected, but its junction with olivine pyroxenite could essentially be a fault that has healed through subsequent recrystallization. It is perhaps notable that the principal difference between the dunite zone in the Judd Harbor ultramafic area and the peridotite dikes in the Hall Cove ultramafic area is one of scale; the relationship of layers to contact is much the same. However, if both are solid intrusions, they must have been emplaced under different structural conditions, none of which is particularly easy to visualize.

Vapor Transfer

To explain the concentric rock-type zoning and cryptic zoning in the Blashke Islands complex, Walton (1951a) has developed a hypothesis based largely on vapor transfer in an ultramafic magma. The magma involved is of wehrlitic (peridotitic) composition and contains liquid, crystals, and aqueous-rich vapor. Because of thermal gradients, water in solution in the magma is more concentrated in the peripheral regions of the magma body, depressing the freezing temperature there so that the body crystallizes from the center outward. Fe, Si, and Ca are supposed to diffuse outward through the vapor phase so that dunite crystallizes in the core and the Fe:Mg ratio of the mafic silicates increases outward. As cooling continues crystallization begins at the border as well and forms olivine augitite (olivine pyroxenite). The two fronts of crystallization join in the wehrlite (peridotite) zone.

The occurrence of such a precise combination of thermal and chemical gradients is difficult to imagine, even in one complex. Moreover, preservation of the zoning produced is unlikely in a magma where gravitational settling of crystals probably is appreciable. The vapor phase is postulated primarily to facilitate diffusion, and the magma must be saturated with water at an early stage. If water was as abundant as seems to be implied, one would expect to see signs of it. However, the rocks in and around the Blashke Islands complex are relatively anhydrous, and miarolitic cavities in the ultramafic rocks are not reported.

A mechanism of vapor transfer will be used to explain secondary features in the Duke Island ultramafic complex, and the possibility of vapor transfer is given some consideration in the discussion of the evolution of the hornblende pyroxenite border zone. However, convection probably occurred in the main body of magma and would not allow the thermal and chemical gradients required by Walton's hypothesis. The mechanism could operate only in the interprecipitate magma where stirring would not be possible. The interprecipitate magma is the most likely place for a vapor phase to form, but diffusion would be restricted to tortuous channelways through the crystalline mesh and, thus, probably would not be as effective as Walton requires. The results could only be secondary, as the major variations would already have been established by primary crystallization.

In summary, none of the above mechanisms (crystallization differentiation, multiple intrusion, solid intrusion, and vapor transfer) is adequate in itself to account for all the features of the Duke Island ultramafic rocks. On the other hand, each one may have some applicability. If so many mechanisms are involved, it is perhaps strange that several complexes would show many similar features. However, crystallization differentiation is a possibility in any magma, and a high water content (or a low saturation limit) may be a characteristic feature of ultramafic magmas, so that vapor transfer could be of general importance. Multiple intrusion is an unpredictable phenomenon but always a possibility. Ultramafic bodies are commonly interpreted as solid intrusions, and although the mechanics of intrusion are conjectural, they might be largely dependent on high water vapor pressure as a driving force.

ORIGIN OF THE HORNBLENDE PYROXENITE BORDER ZONE

The Problem

The size and character of hornblende pyroxenite border zone preclude the possibility that it is a chilled rim of original magma. The zone probably has not developed by assimilation of pyroxene gabbro into the ultramafic magma because, compared to the gabbro and any of the other ultramafic units, hornblende pyroxenite has more iron, a higher $CaO:SiO_2$ ratio, and probably more water (although this latter is not obvious from the chemical analyses). The zone bears some resemblance to the configuration of the border group in the Skaergaard intrusion (Wager and Deer, 1939), but differs in that it is comprised entirely of rocks that are late differentiates rather than having a large proportion of early differentiates.

The parts of the hornblende pyroxenite that cap the layered olivine pyroxenite could be late differentiates resulting from gravitational accumulation, but this mechanism cannot account for hornblende pyroxenite occurring beneath or alongside layered olivine pyroxenite. As hornblende pyroxenite is the lowest temperature major rock type in the ultramafic suite, it might be close in composition to the lowest temperature derivative (p. 187) magma of truly ultramafic composition. This also is indicated by the presence of numerous, relatively small ultramafic bodies in southeastern Alaska in which pyroxenite and hornblende pyroxenite are the only abundant units. Thus the border zone rock might represent a separate injection of magma as postulated by Ruckmick (1957), but the difficulties facing this mechanism (pp. 222-225) make it unlikely.

Two hypotheses are presented here which might account for the normal features of the border zone. In some respects the hypotheses are similar to that preferred by Walton (1951) as an explanation of the concentric zoning in the ultramafic bodies, and undoubtedly, awareness of Walton's discussions has influenced their formulation. The hypotheses primarily are attempts to account for the features of the border zone using essentially only one intrusion of ultramafic magma and having processes based on fractional crystallization as a major control.

The features of the border zone which are, in the author's opinion, consistently developed under normal circumstances and for which explanation is required are:

(1) A continuous zone (but narrower near dunite).

(2) A relatively uniform rock that is richer in Fe, Ca, Al and H_2O than the other major ultramafic units. Magnetite is more common in the thicker parts of the zone but has a well-defined upper level of concentration at about 20 per cent and, except for small veins or bands, does not form massive bodies.

At Duke Island, the mineral assemblage of the hornblende pyroxenite seems to have developed at the expense of that of the olivine pyroxenite. Thus the ultramafic rock originally in contact with gabbro or other country rock is believed to have been olivine pyroxenite.

Postulates for which some justification has been made previously and which will be assumed as fundamental here are:

(1) The order of crystallization in the ultramafic magma is approximately that outlined on pages 180-192.

(2) The Fe:Mg ratio of the residual magma is progressively increased by fractional crystallization, but the iron was largely in the ferric state and, therefore not available to the silicate minerals.

(3) The water content of the residual magma was progressively increased because of the crystallization of anhydrous minerals.

(4) The basic pegmatite was derived from the ultramafic magma, and the hornblende gabbro around the ultramafic areas is, in large part, a metasomatic alteration of pyroxene gabbro that accompanied the introduction of the basic pegmatite. These two represent a major transfer of lime, water, and probably iron from the

ultramafic complex to the gabbro masses.

An idealized model of a layered ultramafic body showing the normal features of the border zone is shown in figure 63.

Hypothesis I

It is proposed that during the late stages of crystallization of the ultramafic magma the fluid materials (whatever their nature) that eventually deposited the basic pegmatite were generated and, while passing through the slightly cooler zone just inside the boundary of the body, reacted with the olivine pyroxenite there to produce hornblende pyroxenite. The reaction would be essentially the same as that which is indicated to have occurred along the dike zone in the main olivine pyroxenite zone of the Judd Harbor ultramafic area (pp.169-170).

Under conditions of fractional crystallization, the residual magma should become more capable of producing the pegmatitic facies with increasing elevation in the layered series and as it became more different from a magma that would precipitate dunite. Thus, the size of the reaction zone should be larger near the upper part of the layered olivine pyroxenite and away from dunite. The probable distribution of the magnetite-bearing hornblende pyroxenite, as indicated by surface exposures and by the aeromagnetic map, correlates well with the distribution of basic peg-

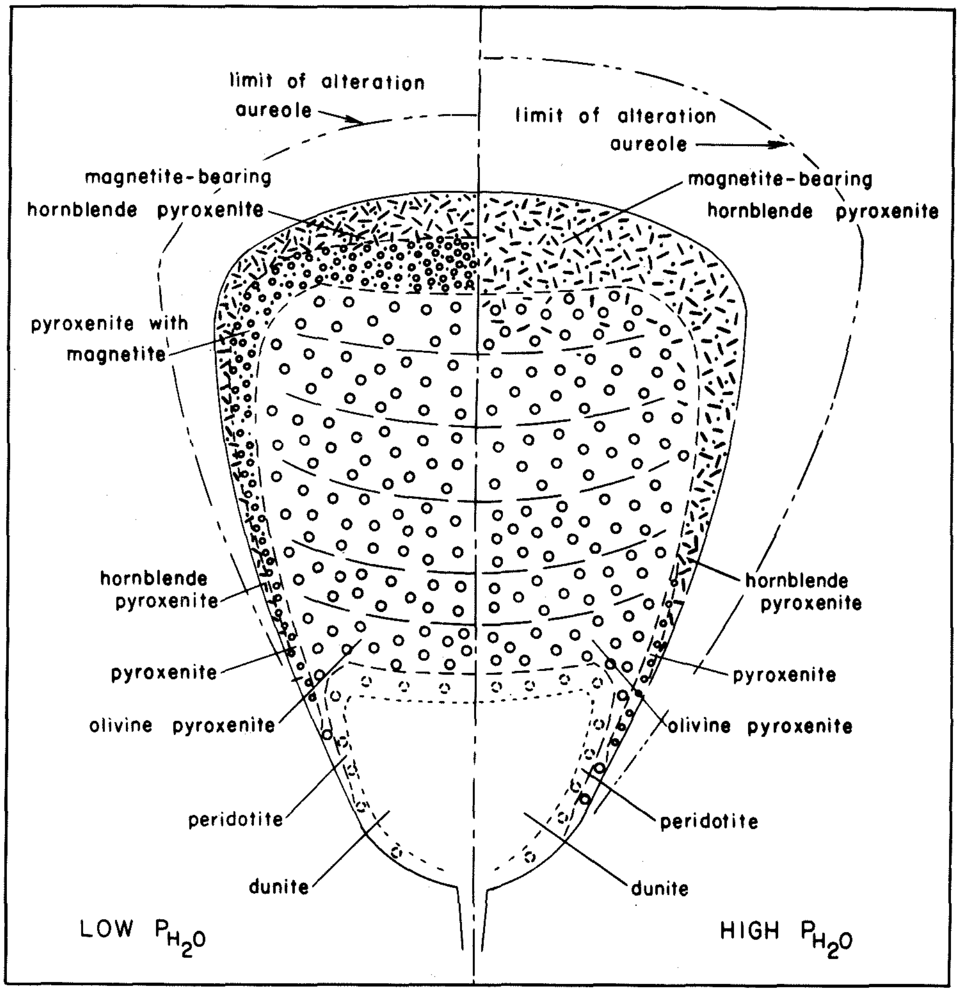


Figure 63.- Idealized configuration in a differentiated layered ultramafic complex. A contrast in the character of the zoning is shown that might develop because of different amounts of water in different complexes. The diagram is drawn as a section but is a fair approximation of the plan of several complexes if the layering is removed.

matite and calcic hornblende gabbro.

The iron content of the residual magma might be expected to vary somewhat depending upon the part of the complex from which it was derived. This could account for the smaller percentage of magnetite in the border zone near dunite. The altering fluid material is a derivative of a magma that has, at some time in its history, crystallized olivine pyroxenite. Hence the alteration of olivine pyroxenite to more ferriferous hornblende pyroxenite is only a slight modification of the processes occurring in the normal magmatic reaction series. The rate of temperature drop in the intrusion is small by this stage of its solidification, and therefore the fluid material might be expected to react only until it had modified the rock in the cooler border zone to the point that they were in equilibrium with each other. Any fluid not used in the reaction would be free to move on and would not be expected to deposit magnetite in the concentrations typical of contact metasomatic magnetite deposits.

The reaction would primarily result in addition of material to the border zone if the analogy to the magmatic reaction series is valid. As olivine is the principal mineral to disappear, one might assume that Mg is the one element not added in appreciable quantity. Holding MgO constant, comparison of the analyses of typical olivine pyroxenite (table 8, section 2, specimen R-38-2) and magnetite-bearing hornblende pyroxenite (table 10, No. I) shows that about one third of the material in the present border zone would have to be added during the late magmatic stages. This quantity becomes rapidly less if any Mg is removed. The increase in volume in the border zone would, presumably, be compensated for by decrease in volume in the interior of the ultramafic body due to losses of material, and to cooling and crystallization.

It might be argued that the basic pegmatite is the residual liquid of only hornblende pyroxenite magma. However, the quantity of basic pegmatite is exceedingly large; the area of its occurrence is greater than the outcrop of hornblende pyroxenite and, in extensive parts, is underlain by more than 50 per cent dike rock. Although it is not possible to tell how much hornblende pyroxenite exists beneath the basic pegmatite and hornblende gabbro, it would seem that, in order to produce so much highly calcic material, the residual fluid of the entire ultramafic body must be drawn on, not just the hornblende pyroxenite.

This points up one of the hitherto neglected problems in the derivation of the basic pegmatite from the ultramafic magma, a problem that has, in fact, served as one of the lines of evidence that the pyroxene gabbro and ultramafic rocks were not derived from the same magma. Basic pegmatite is abundant and contains about 40-50 per cent plagioclase. Except for one small occurrence of hornblende pyroxenite, none of the ultramafic rocks contain plagioclase. As with the relationship of pyroxene gabbro, it is perhaps strange that so much feldspathic material could be derived from the same source as material so barren of feldspar. The probable explanation is that the pegmatite fluid is rich in water, and as long as it was interprecipitate magma and in intimate contact with pyroxene the Al_2O_3 and other components that generally form feldspar were used to make pyroxene into hornblende. Plagioclase began to form only after the fluid entered fractures to form dikes. It also is possible that chemical fractionation accompanied the separation of the pegmatite fluid(s) from the ultramafic mass. However, this cannot be proved at the present time.

How or why the pegmatite left the ultramafic body is a matter of speculation. The hypothesis is based on the assumption that the basic pegmatite is a direct derivation of the ultramafic magma and, therefore, has moved outward. If it is not a direct derivative, then because of its composition it must represent a separate intrusion, a possibility which is not particularly attractive and which would not account for the distribution of the pegmatite around the upper part of the complex except by coincidence. Considerable evidence of late magmatic recrystallization, autoreplacement, and segregation will be cited (pp. 241-265) and serves to emphasize the importance of processes operating at this stage in causing redistribution of material. It is important to remember that by this stage the rate of cooling and crystallization in the complex is at its lowest point. The time available for the process proposed here to go on must represent a major fraction of the total period of crystallization in the complex. The temperature range during which it may have occurred is indicated to be large (pp. 201-205). Water has concentrated and is undoubtedly important as a catalyst. Separation of an aqueous-rich vapor phase could be an important factor in that:

(1) It might force the pegmatite magma out of the intrusion.

(2) It could act as a medium of transport by both flow and diffusion.

(3) It could assure uniform replacement in the hornblende pyroxenite zone.

Positive evidence of a gas phase has not been recognized in any of the Duke Island ultramafic rocks. The author has observed vuglike cavities in a hornblendite body at Alava Bay, Revillagigedo Island, and one might infer from these that a gas phase did form in some ultramafic bodies. The cavities give no clue as to the stage in the crystallization that a gas phase might have begun to form; they are late in appearance, but this is to be expected. Early vapor phase could be lost without leaving comparable trace of its existence. Zavaritsky (1937, p. 72) reports gas-filled miarolitic cavities in the Nizhny Tagil dunite massif in the Ural Mountains of Russia. This would seem to indicate that even the interprecipitate magma of dunite can contain enough volatile material to separate a gas phase.

The hypothesis requires that the fluid material which reacts with the border rocks move outward while the front of reaction advances inward. The only apparent control for the position of the front is thermal. Although the contact of hornblende pyroxenite and olivine pyroxenite is gradational, it still seems too sharp to be accounted for solely by the low thermal gradients that must have existed at the time. The contact relationship would be much easier to explain if at least one of the components required in the reaction was migrating inward. Perhaps the fluids concentrated in the border zone and moved laterally in it to some extent before escaping into the country rock.

Hypothesis II

An alternative hypothesis is suggested here that requires only slight modification of much of the process suggested above. A detailed analysis will not be attempted.

Although most of the crystals in the Duke Island ultramafic body have settled and accumulated on the floor of the magma body, it is reasonable to assume that a shell of ultramafic material also formed around the periphery.

Probably most of the initial shell was olivine pyroxenite. Goranson (1937) has pointed out that, theoretically, water can be lost through the walls of an intrusion throughout much of its crystallization period by an "osmotic pressure" effect. Possibly there was a temporally comparable loss of other more mobile components from the Duke Island magma, perhaps even related to a loss of water. Judging from other geological environments involving metasomatic processes and considering the composition of the magma involved here, the quantitatively most significant of these components might well be oxides of Ca, Fe, and Si. This material, upon moving out into the slightly cooler shell of ultramafic rock surrounding the magma, could react and result in magnetite-bearing pyroxenite or hornblende pyroxenite. The mobile materials would have to move largely by intergranular diffusion, but the shell would probably thicken as crystallization progressed, and if each increment was replaced only shortly after it formed, the distances required for movement of material would not have to be large. Eventual accumulation of the layered series against the shell would blanket out the process except to the extent that it might occur in the interprecipitate magma. The blanketing effect plus the change in composition of the magma as crystallization progressed could account for the thickening and changes in the border zone with elevation in the layered series and away from dunite. Continual loss of iron from the magma might account for the absence of cryptic layering.

As they stand, hypothesis I begins at about the stage that hypothesis II ends. They should perhaps be integrated to form a continuous sequence of events but have been left separate in order to emphasize as much as possible the factual evidence as compared to speculative interpretations.

A hypothesis along the lines of the two given above will be assumed as the origin of the hornblende pyroxenite in the final analysis. It is not an essential part of the explanations of the relationships of dunite and peridotite to olivine pyroxenite and can be readily replaced by some alternative. Therefore, that such a hypothesis might prove invalid need not detract from the explanations of other features.

SECONDARY FEATURES IN THE ULTRAMAFIC ROCKS

RECRYSTALLIZATION

Parts of the olivine pyroxenite and peridotite have exceptionally coarse grain size and blotchy, uneven textures. The average grain size of this material may be 2-10 times that in the more normal rocks. In the peridotite, coarse poikilitic crystals grade to large crystals identical to those forming late veins and pegmatitic segregations. This coarse-grained facies shows the following features.

(1) Sporadic occurrence in irregular, patchy areas

which may either transgress or parallel the trend of the primary layering. The layering can commonly be traced through these areas, but the primary textures are obscured and obliterated. The long axes of many of the coarse pyroxene crystals are normal to the layering. These features are illustrated in figures 64 to 67.

(2) A bulk composition not much different from that of the more normal-looking rock in the same area. The layering may be somewhat wavy and irregular, suggesting a slight volume increase. Segregation of pyroxene and olivine occurs, but it is not obvious that either is consistently more abundant than average. Petrography has not shown a consistent mineralogical difference between the coarse rock and the adjacent finer material.

(3) A distribution with no obvious limits. Coarse textures have been observed almost everywhere in the olivine-bearing ultramafic rocks. In the Hall Cove ultramafic area, they are relatively common in and around the large olivine pyroxenite blocks in the peridotite zone. The nature of their occurrence is illustrated in plate 4, designated "coarse-grained and uneven-textured."

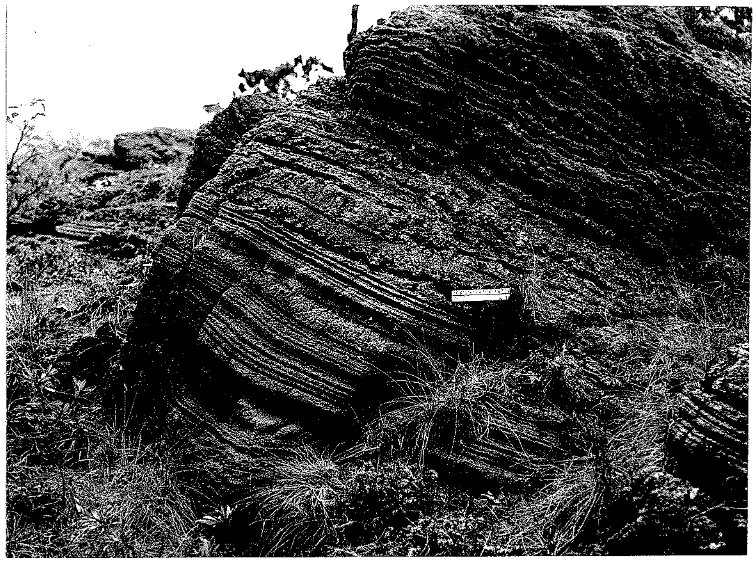


Figure 64. Layered olivine pyroxenite showing the contrast between normal primary textures (below) and the coarse-grained, recrystallized rock. The recrystallized layers are somewhat wavy, suggesting a slight volume increase. In outcrop, a weakly developed orientation of pyroxene crystals normal to the layering is evident in the coarse rock. The locality is the northern olivine pyroxenite zone of the Hall Cove ultramafic area.

Figure 65. Coarse, recrystallized zone cutting across primary layering in olivine pyroxenite. The locality is near that shown in figure 64.

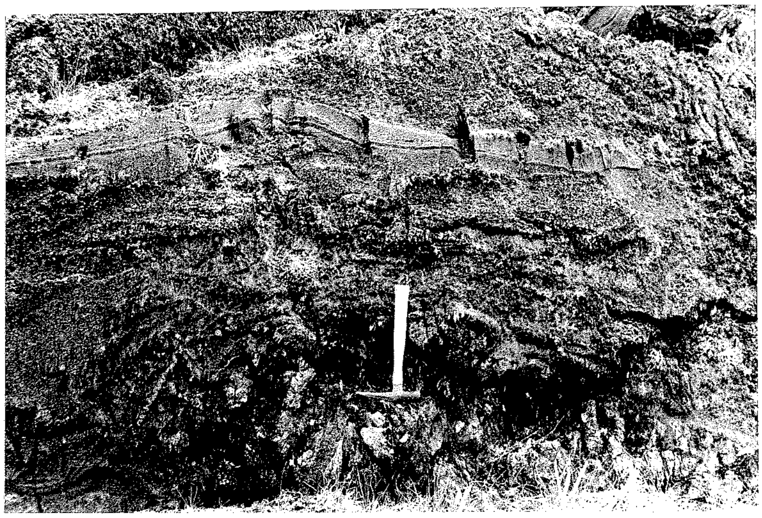


Figure 66. Recrystallized olivine pyroxenite. The fine-grained layered rock has primary textures. The coarse, uneven textured material is believed to be secondary. Note that the very coarse pyroxene near the head of the geologic hammer tends to be oriented normal to the layering. The olivine pyroxenite is part of a large block in the peridotite zone of the Hall Cove ultramafic area. The locality is the top of Knob Hill.

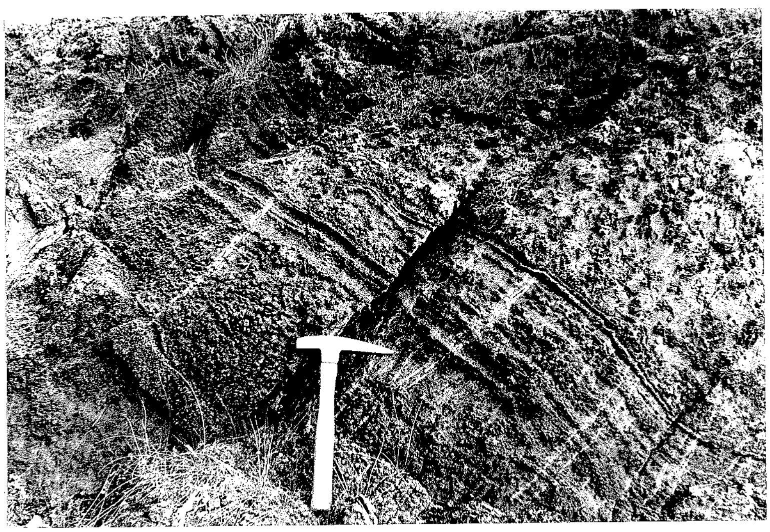


Figure 67. Partially recrystallized layered peridotite. The pyroxene crystals near the geologic hammer are highly poikilitic, being charged with olivine inclusions. The locality is the Hall Cove ultramafic area, east of Knob Hill.

Patches of coarse texture are prevalent in the northern olivine pyroxenite zone just east of Knob Hill, and they are widespread throughout the southern olivine pyroxenite zone. In the latter, they are most common around the small patches of dunite and pyroxene dunite shown on the map. The greatest development of coarse-grained material is in the Judd Harbor ultramafic area where, although not mapped, it is estimated to make up 40 or more per cent of the main olivine pyroxenite zone and is especially common along the axial regions of the two large folds.

The transgressive relationship of the coarse material to the primary layering and the partial obliteration of the layering are regarded as conclusive evidence that much of the coarsening is a secondary effect superimposed on the original textural and structural features of the primary crystalline accumulates. The widespread occurrence, the close association with veins and segregations, and the common poikilitic pyroxene suggest that the coarse, patchy material is a late magmatic product resulting from local concentration of residual magmatic fluids. The lack of evidence of prominent changes in composition or volume indicates that the process is primarily one of recrystallization.

Recrystallization with control by the layering probably is the best explanation of the orientation of the long axes of pyroxene crystals normal to the layering. The only comparable structures known to the author to occur in other layered mafic or ultramafic intrusions are the "perpendicular feldspar rock" and the "wavy pyroxene rock" in the border group of the Skaergaard intrusion (Wager and Deer, 1939, pp. 26-27) and the "Harrisite structure" in the layered peridotite of the Island of Rhum (Wager and Brown, 1951). These phenomena are explained as primary features due to crystal growth normal to the interface of liquid and previously precipitated crystalline material. Such a process does not seem capable of accounting for the distribution of coarse, perpendicularly-oriented pyroxene crystals in zones transverse to the layering as occur at Duke Island. On the other hand, it was not in mind during the mapping of the Judd Harbor ultramafic area and may have more applicability in the origin of the non-graded layering there than is realized. One feature suggestive of this sort of process was noted in the eastern half of the peridotite zone of the Hall Cove ultramafic area. Along the upper surface of many of the olivine pyroxenite blocks is a rim of coarse pyroxene crystals oriented normal to the surface. The upper part of some of the blocks seems to be recrystallized, but the rim on others has the appearance of a vein along the contact with overlying peridotite. Not every block has a rim, and the rim is rarely complete, but where developed it invariably is on the upper side of the block. This relationship was first pointed out by Mr. E. J. Oliver. It is shown at several places in plate 4. In figure 68 is a structure that could be interpreted as the draping of layering over crystals that grew at the interface of earlier accumulates with the overlying magma. However, the author believes that much of the bulging of the layers resulted after their deposition because of the growth of the coarse crystals. Veins of identical coarse pyroxene cutting across layering are abundant in the same area, as is some of the most definite evidence of recrystallization.

SECONDARY DUNITE

In the descriptions on the various areas of ultramafic rocks, frequent mention has been made of small, patchy bodies of dunite. These bodies have the following characteristics:

(1) They have small dimensions. Maximum dimensions for individual bodies range from one foot to 200-300 feet.

(2) Smaller bodies may be pure dunite, but the larger ones typically contain small masses of olivine pyroxenite and appreciable coarse-grained pyroxene, either as sporadic disseminations or in clusters. Poikilitic pyroxene and

Figure 68. Peridotite layering deformed or draped over a large pyroxene crystal oriented normal to the layering. The pyroxene is typical of that in late stage veins cutting the peridotite in the same area (fig. 15).

veins of coarse-grained pyroxene are common. The texture and grain size of the dunite itself is relatively normal, as are the refractive indices of the olivine.

(3) They have extremely irregular forms. All gradations from narrow bands to more or less equant bodies occur. The olivine pyroxenite against the dunite has a ragged, "moth-eaten" appearance, and a slight coarsening of the pyroxene at the contact is common. The boundaries of the patches are a distinct contrast to the sharp, straight-walled boundaries of the peridotite dikes and olivine pyroxenite blocks which occur in the main peridotite zone of the Hall Cove ultramafic area. However, many of the patches occur within angular olivine pyroxenite blocks (plate 4). Some blocks show an abrupt change from straight-walled boundaries to ragged, irregular forms against peridotite, but relict angular outlines can commonly be traced in the peridotite by higher than normal pyroxene content.

(4) The patchy bodies transect layered structures in the olivine pyroxenite without displacing or deforming them in any way. The layers do not drape over the dunite as they do over

included olivine pyroxenite blocks. The patches are commonly elongated parallel to the layering, but this is not a general rule. In some places, traces of layering can be followed continuously from olivine pyroxenite into and through the dunitic areas. Included pieces of olivine pyroxenite may contain poorly developed layering, and this invariably has the same orientation as that in the surrounding olivine pyroxenite.

(5) The dunitic bodies are associated with recrystallized rock, and the two are transitional in many places.

(6) The dunitic patches have wide distribution but tend to occur in zones. The principal zones in the Hall Cove ultramafic area are (a) in the blocks in the eastern half of the peridotite zone, (b) along the northeastern fringe of the peridotite zone, and (c) two or three main areas in the southern olivine pyroxenite zone. In the Judd Harbor ultramafic area, the patches are concentrated along the axes of the two main folds. These zones do not parallel the layering although the patches themselves commonly do.

The above features are illustrated in figures 69 to 74.

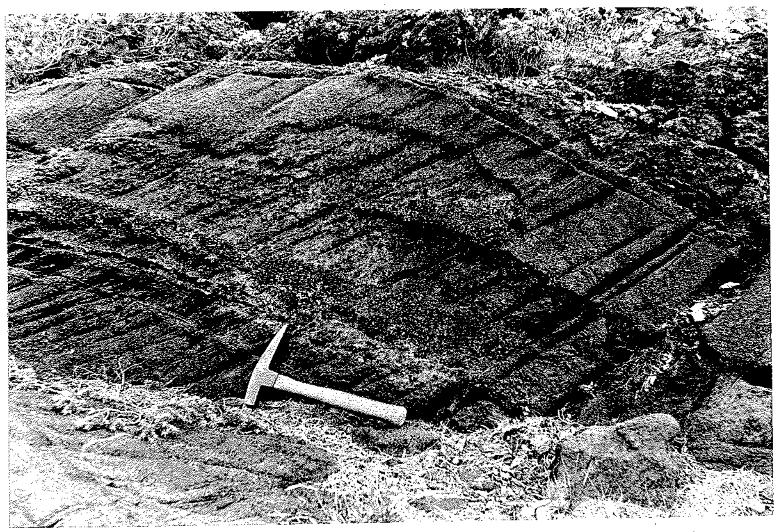


Figure 69. Layered olivine pyroxenite (fine-grained, even-textured rock) replaced by an irregular body of peridotite. An apparent trace of relict layering is evident in the peridotite due to differential concentration of poikilitic pyroxene, and slight coarsening of olivine pyroxene right at the contact is evident. The olivine pyroxenite is part of a relatively large block in the peridotite zone of the Hall Cove ultramafic area. Serpentinized joint swarms are the cause of the linear pattern going diagonally upward to the right.

Figure 70. Layered olivine pyroxenite showing a ragged, "moth-eaten" appearance against peridotite. The locality is the peridotite zone of the Hall Cove ultramafic area.

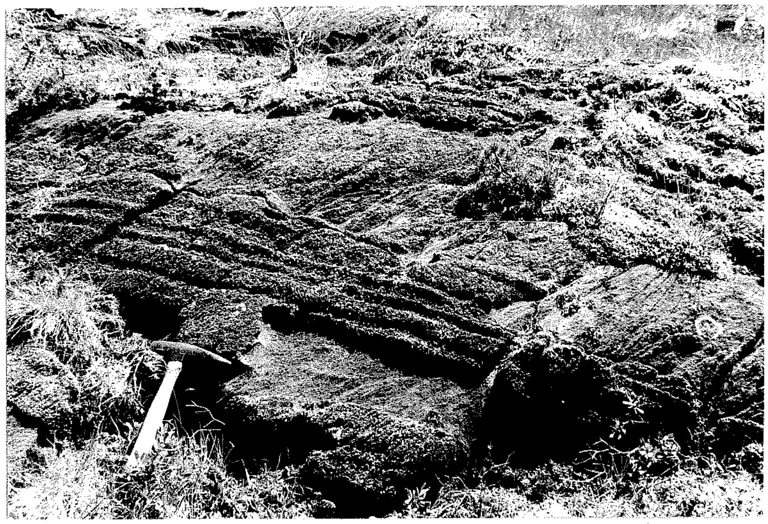


Figure 71. Layered olivine pyroxenite irregularly transgressed by dunite but showing no deformation. The layering probably is partially recrystallized.

Figure 72. Olivine pyroxenite, transected by dunite and recrystallized. Note the abundance of coarse-grained pyroxene in veins or segregations.

Figure 73. Irregular body of dunite (yellow) in layered olivine pyroxenite. The layering is regular despite the presence of dunite, and no evidence of bulging or displacement could be detected in the field. This is one of the many patchy bodies of dunite that are mapped in the main olivine pyroxenite zone of the Judd Harbor ultramafic area. Most of the olivine pyroxenite is believed to be recrystallized, but a small amount of smooth-textured primary material is visible in the photograph.

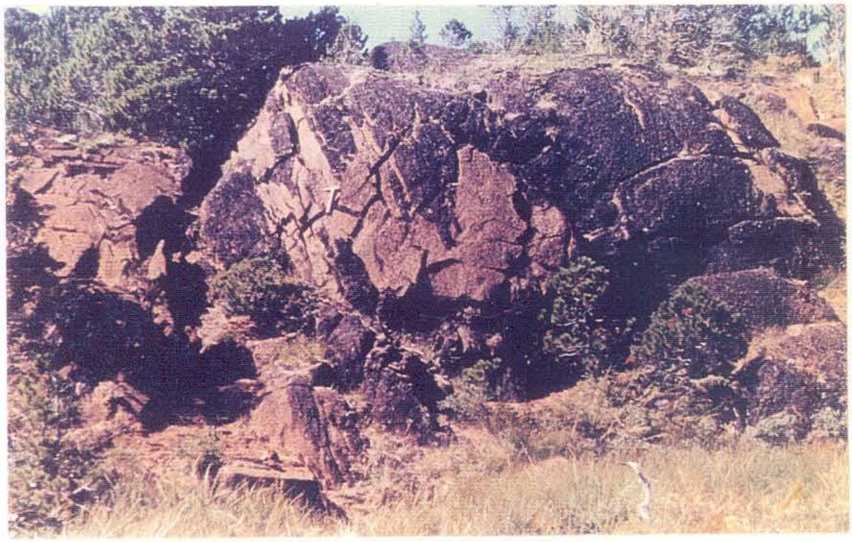


Figure 74. Small dunite body in olivine pyroxenite. The features and locality are essentially the same as in figure 73.

The patchy dunitic bodies do not represent inclusions in olivine pyroxenite, because the layers do not drape over them and because they occur in zones transverse to the layering. They are not intrusive because of their relationship to the layering. That the trace of olivine pyroxenite layering persists undisturbed through some of the patches indicates that the dunite has formed by volume for volume replacement. The relict contacts of angular blocks in peridotite also suggest replacement. The small included pieces of olivine pyroxenite may be remnants, because the layered examples are never misoriented. None of the other characteristics refutes a replacement origin. The commonly associated coarse-grained pyroxene veins may well represent pyroxene that has been displaced from the rock but not completely removed from the environment. A similar interpretation can be applied to much of the poikilitic pyroxene, except that only slight redistribution is required. Deposition of some coarse pyroxene seems to have caused a local volume increase (fig. 68). Control of replacement by layering is indicated by the tendency of some patches to parallel the trace of layering. Obviously, one cannot hope to prove a replacement origin for each individual body, but the similarity in form and occurrence implies a common genesis.

Some indication of replacement of pyroxene by olivine has been observed in peridotite. Two examples are shown in figures 75 and 76. In figure 75, it is clear that the layering has been obliterated by removal of pyroxene. It is not possible to tell whether olivine has been added or not. In figure 76, a pyroxene-rich stratum shows a prominent gap, but traces of the layer indicate that it was once continuous. Immediately beneath the gap, partial obliteration of several layers show that the discontinuity probably was not produced by primary sedimentation but, rather, represents a zone of replacement transverse to the layering. In this example, the replacement was apparently volume for volume.

PROPOSED EXPLANATION

The phenomena ascribed to replacement and recrystallization must be genetically related because of their close spatial association and more or less transitional relationship. Their widespread distribution throughout the ultramafic complex suggests that they are a part of the evolution of the complex and not due to processes operating from the outside. A considerable amount of redistribution of material is implied. How this distribution was effected is one of the major problems. The other main problem is why redistribution occurred. Replacement and recrystallization occur in a rock because an environment develops in which the original crystalline material is no longer in equilibrium. A plausible expla-

Figure 75. Peridotite layering partially obliterated due to apparent leaching of pyroxene and, possibly, replacement by olivine.

Figure 76. Pyroxene-rich layering in peridotite partially replaced by olivine. Several of the smaller layers in the lower part of the photograph show the same discontinuity as occurs in the main layer, indicating that the gap was not formed during primary deposition. The locality is on the top of Knob Hill, near its north end.

nation of these phenomena in the ultramafic rocks requires that some reason for disequilibrium be found.

Transfer of Materials

The erratic way in which the patchy areas of replacement dunite and recrystallization transect and obliterate layering, and their close association with veins and segregations indicate that the magma was at or approaching a vein-forming stage during their development. Clearly, it is not possible to know exactly what this means with regard to the composition and condition of the magmatic residuum, but the following deductions can be made.

(1) The fluid material was mobile or of low viscosity as evidenced by:

(a) Its tendency to concentrate in local zones and veins.

(b) The general absence of obvious feeders to the bodies of replacement dunite. It seems probable that much of the necessary migration of material took place through intergranular channelways.

(c) The coarse grain size of the pyroxene. Grain size is known to be dependent on many things, but one of the reasons for the coarse textures characteristic of the recrystallized zones may be greater freedom of movement of the various chemical components so that fewer centers of crystallization are required.

(2) Water was probably concentrated, because the only minerals to form previously in the olivine-bearing rocks were anhydrous.

Bowen and Tuttle (1949, p. 459) observed in work on the system $MgO-SiO_2-H_2O$ that enstatite was converted to forsterite by extraction of silica by water vapor at temperatures above 600°C. From this, they suggest that the streaming of water vapor undersaturated in silica and at appropriate temperatures through cracks in enstatite pyroxenites could produce veins of dunite and, conversely, that water vapor saturated in silica could produce enstatite veins in dunite. Barth (1952, p. 226) points out, in regard to dunite veins, that such a process is not logically limited to veinlike or dikelike bodies but that any form is conceivable. The process, insofar as it has been developed, is not adequate to account for replacement dunite and recrystallization in the Duke Island ultramafic rocks because the pyroxene is diopsidic rather than enstatite and the replacement is apparently volume for volume. Not only must SiO_2 be transferred, but also CaO, MgO, and to some extent FeO.

Since the original observation of Bowen and Tuttle, transfer of other materials by a water-rich gaseous phase has been noted in hydrothermal experimentation. Contamination of charges with iron from the walls of the apparatus is a common problem. Transfer of alumina, silica, and alkalies is reported by Jahns and Burnham (1958). Moreover, large amounts of material are transferred by diffusion through the gaseous phase, and movement of fluid itself is not essential (Howe and Burnham, 1957). To the author's knowledge, no one has yet reported transfer of CaO and MgO, but little work has been done on the proper systems under appropriate conditions. There is no reason why some movement of these materials would not occur, the question is one of rates. Certainly, if transfer by diffusion through a gas phase occurred in the ultramafic rocks, CaO and MgO would be involved because of their exceptional abundance.

Definite evidence of a water-rich gaseous phase has not been recognized in the Duke Island ultramafic rocks. However, transfer by diffusion through such a medium is considered to be the most plausible mechanism for the redistribution of material in the zones of recrystallization and dunite replacement.

Problem of Disequilibrium

Two points are significant with regard to the problem of disequilibrium:

(a) The phenomena attributed to replacement and recrystallization are localized in zones, suggesting that the residuum of the interprecipitate magma was concentrated by being moved out of its original environment into a new one. This in itself is probably the main reason for disequilibrium. It may cause disequilibrium in several ways and also be a factor in promoting the formation of a vapor phase. For example, mechanical deformation may have caused the localization of the interprecipitate magma by squeezing some parts of the pile of crystals and liquid and dilating others. Such a process could result in some chemical fractionation, the more mobile components of the magma being enriched in the dilated zones. A mechanical concentration of water could depress the crystallization temperature of the localized magma and create what effectively is a superheated liquid. This liquid could not be in equilibrium with the earlier formed crystals, and considerable amounts of solution and subsequent redeposition might ensue. Continued crystallization in the parts of the magma where water was concentrated could give rise to resurgent boiling.

(b) In the Hall Cove ultramafic area, much of the recrystallization and replacement is near the peridotite zone, either in included blocks or along the northeast margin of the zone. In these areas, the refractive indices of pyroxene and olivine in primary olivine pyroxenite generally are significantly higher than for the same minerals in the primary peridotite (see figs. 61 and 62). Here is an environment of definite disequilibrium. The peridotite at these places is known to be the product of a younger magma. Its interprecipitate liquid must have been more or less in equilibrium with its crystals and should, therefore, have had higher Mg:Fe and Mg:Ca ratios than would be compatible with the olivine pyroxenite, even under the same conditions of temperature and pressure. If this liquid was tapped at the proper time, for example by an intergranular channel of water-rich gaseous fluid, it might be expected to supply material necessary to replace olivine pyroxenite and to create the conditions necessary for recrystallization. In other words, the recrystallization and replacement probably are metasomatic effects imposed on the olivine pyroxenite by the interprecipitate magma of the peridotite. It is notable that these effects are erratic in the included blocks. If an early magmatic process had been involved, a more uniform product might be expected. Evidently the blocks persisted out of equilibrium, even though in relatively close contact with the interprecipitate liquid until the late stages of crystallization when localization of the magmatic residuum and, probably, the formation of a gas phase promoted reaction.

In the Judd Harbor ultramafic area, the emplacement of dunite (or the magma from which it crystallized) followed the crystallization of the main zone of olivine pyroxenite. The interprecipitate magma of this material probably provided the magnesia for the replacement dunite occurring within the area. The locus of replacement and recrystallization apparently was the dilated zone in the axial regions of the two folds.

SIGNIFICANCE OF RECRYSTALLIZATION AND REPLACEMENT

The recrystallization and replacement demonstrate the importance of late magmatic processes in effecting redistribution of material and in determining textural features. Although these features are most prominent locally, they may occur to some degree in all the rocks. Most of the Duke Island ultramafic rocks show some uneven textures with a recrystallized appearance.

The formation of dunite by autoreplacement in the ultramafic rocks has not been recognized in previous studies in southeastern Alaska and, in fact, is a process rarely mentioned in the literature on ultramafic bodies. Its identification at Duke Island is primarily dependent on the presence of layering, and there is no reason why it might not occur at other localities. Certainly, all dunite has not formed by replacement, even at Duke Island, nor need every ultramafic complex with dunite show some replacement. However replacement does occur and must be kept in mind when considering the origin of this rock type. The ragged outlines of bodies of replacement dunite at Duke Island are distinctive and, if observed in other areas, might be indicative of replacement. On the other hand, this need not be the only form of replacement bodies.

A notable feature of some of the dunite bodies in southeastern Alaska is that they have large parts virtually devoid of silicate minerals other than olivine. If dunites are crystalline accumulates from more pyroxenic magmas, one might expect them to show at least some interstitial pyroxene. The possibility that dunite can be produced by replacement of olivine pyroxenite suggests that a somewhat comparable process may occur which is capable of leaching interstitial pyroxene. Figures 75, 76, and 77 are notable in this respect. Pyroxene veins are common in and around dunite bodies, and a leaching process could perhaps be the source of pyroxene in thin pyroxenite rims surrounding some of the dunite bodies in southeastern Alaska, notably the Annette Island body (unpublished map by H. P. Taylor, Jr.).

ORIGIN OF THE LAYERING

POSSIBLE MECHANISMS

The layering in the ultramafic rocks is fundamentally due to gravity settling of crystals and rock fragments from a body of magma and accumulation on the floor of the magma chamber. Coates (1936) has proposed a mechanism based

Figure 77. Irregular dunite body (light-colored material) with a segregation of coarse-grained pyroxene in uniform, primary-looking peridotite. The relationships suggest partial leaching of the pyroxene. Note that the fine-grained, fracture-controlled pyroxene veins fade in the dunite. The locality is in the northern part of the Hall Cove ultramafic area, just east of the Hall Cove fault.

on simple experiments whereby layering is produced when two sorts of crystals of differing density settle simultaneously. By this mechanism, a stage is reached at which sinking of the denser crystals causes a current in the liquid that carries the less dense crystals upward. Thus a low density layer develops over a high density one. Dense crystals falling into the upper layer slip through until the mesh becomes too tight, and then the process repeats. Similar results might be produced with crystals at the same density but of different sizes. However, Coates' experiments were done in small vials, and the layers produced were hazy and discontinuous. It seems improbable that the mechanism is the cause of the continuous, sharply defined layers that occur in the Duke Island ultramafic rocks. Hess (1938, p. 266) gives further criticism of Coates' mechanism. Some factor in addition to gravity settling must cause the supply of crystals to be intermittent or fluctuating.

Several mechanisms have been proposed which might result in periodicity in the supply of crystals (Ussing, 1913, p. 361; Grout, 1918b; Bowen, 1928, p. 168; Hess, 1938, p. 266; Wager and Deer, 1939, pp. 266-270; Yoder, 1955; and Wager, 1953, 1959). They fall into two general classes:

(1) Gravity settling accompanied by currents in the magma.

(2) Gravity settling accompanied by fluctuations in the rate of crystallization of the magma due to variations in external pressure, water vapor pressure (or the pressure of other volatiles), or to control by rates of diffusion of ions.

SITE OF FORMATION OF CRYSTALS

The problem of the place in the magma chamber where crystals form must be dealt with before any hypothesis can be set up to explain the layering. Crystallization is generally assumed to take place at or near the top and sides of the chamber because these are the main cooling surfaces. However, Kennedy (1954, p. 496) has postulated that crystals might form near the floor because water and other volatiles, concentrated at the top of the chamber, depress the liquidus temperature so that, as cooling progresses, the magma in the lower regions solidifies first. The fragments in the fragmental layers in the Duke Island ultramafic rocks could not have formed near the bottom of the chamber because they are different in composition and mineralogy from the matrix material that encloses them and from the adjacent graded layers of discrete crystals. The fragments must be pieces of an earlier generation of ultramafic rock that fell into the magma at some point higher in the chamber. As they form well-developed graded layers contiguous with layers in which individual crystals are graded, it can be inferred that individual crystals also came from higher in the chamber and formed near the contact of magma and earlier rocks.

Two alternative hypotheses are visualized by the writer:

(1) The crystals form along the roof and walls of the chamber and are carried downward by bodily movement of magma. Deposition on the floor is much like the deposition of sedimentary beds by water currents.

(2) The crystals form intermittently at the top of a quiescent magma body, and then, because of densities greater than the magma, settle to the floor. Sorting takes place in each batch because of the different settling velocities of crystals of different sizes.

A difference between the two hypotheses is that by the first crystals settle only a short distance, whereas in the second, they settle thousands of feet, the thickness of layered rocks (Wager and Deer, 1939, p. 276).

EVIDENCE OF CURRENTS

Scouring and Cross-layering

The best evidence of the operation of currents during the accumulation of sedimentary rocks probably is scouring and cross bedding. Figures 78 to 86 show structures in the layered ultramafic rocks considered by the writer to be the

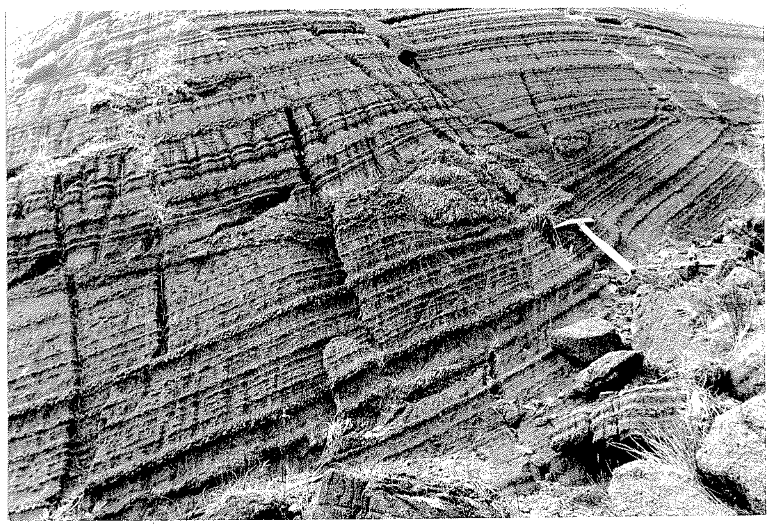


Figure 78. Layered peridotite with olivine pyroxenite fragments. The layers are truncated beneath the block and accompanying gravel-like material. Cross-layering is visible in the upper right. The locality is the Hall Cove peridotite zone, near its west end.

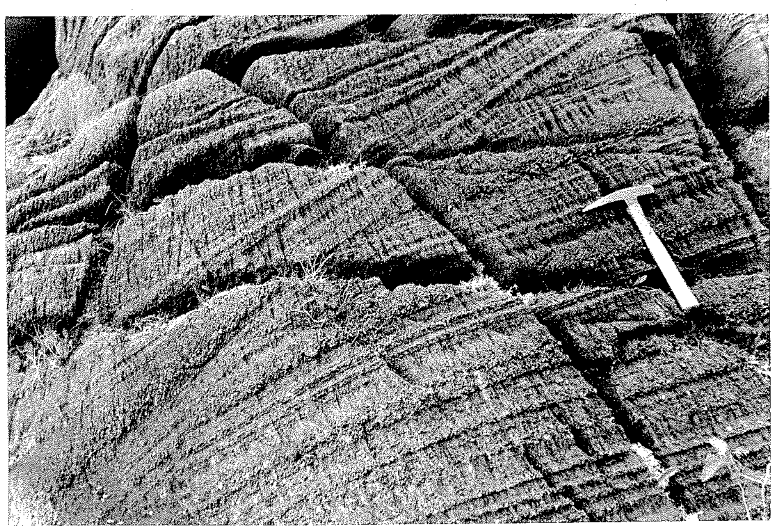


Figure 79. Cross-layering in peridotite. The occurrence is a few feet to the left of that shown in figure 78. Joint swarms are extensively developed, their trend paralleling the handle of the geological hammer.

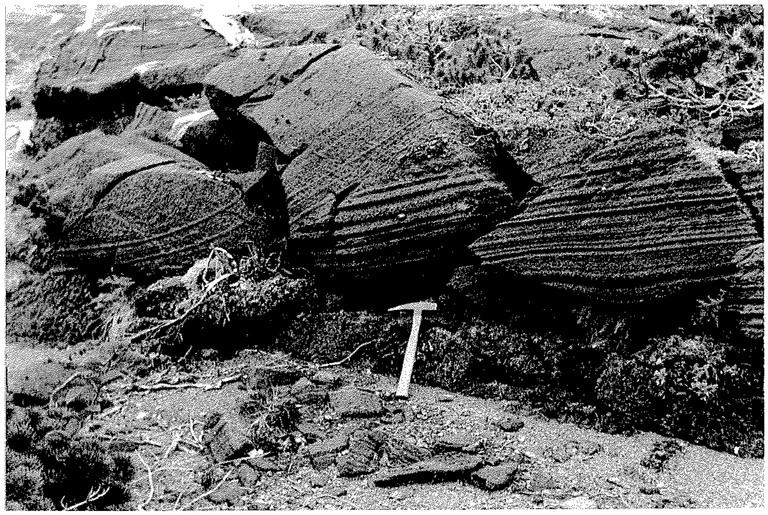


Figure 80. Cross-layering in peridotite. The geological hammer rests on an olivine pyroxenite block. The locality is the east end of the peridotite zone in the Hall Cove ultramafic area.

Figure 81. Cross-layering in peridotite. A large olivine pyroxenite block occurs just below the geological hammer. The locality is the peridotite zone in the Hall Cove ultramafic area.

Figure 82. A rumpled layer in peridotite which has been planed before the deposition of the overlying layer is shown. A small fragment of olivine pyroxenite is visible in the rumpled layer near the upper left corner of the photograph, and the next layer is draped over the block. The locality is the peridotite zone of the Hall Cove ultramafic area, near its west end.

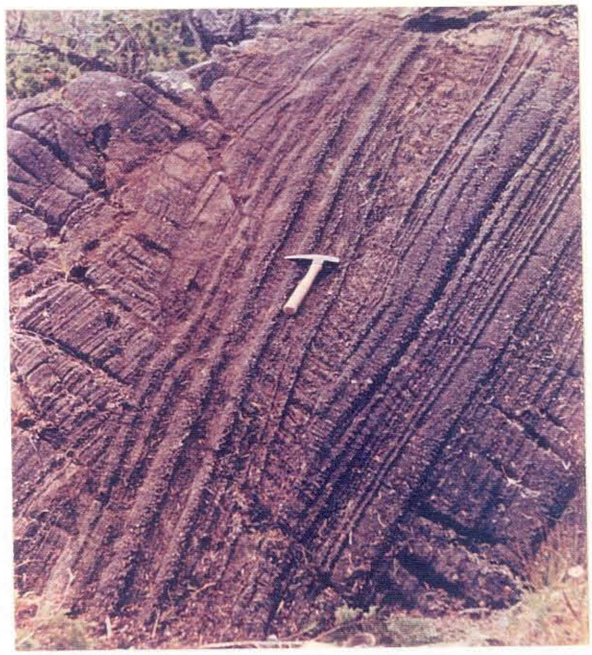


Figure 83. Cross-layering in olivine pyroxenite. The crossed layers are slightly richer in olivine than the others. Figure 84 is a sketch of this outcrop. The locality is just north of the central part of the peridotite zone of the Hall Cove ultramafic area.

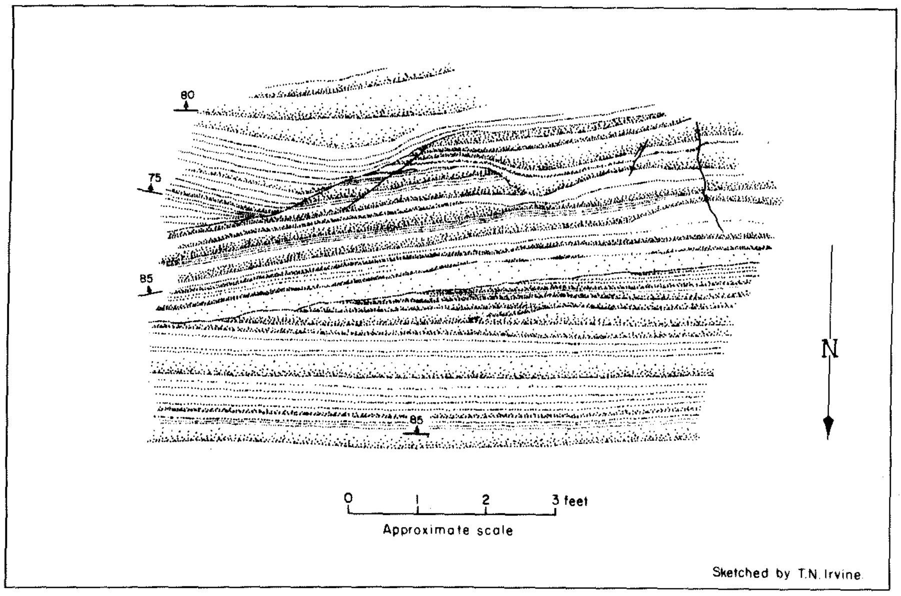


Figure 84. Cross-layering in olivine pyroxenite, Hall Cove ultramafic area.

result of current scour.

The areas in figures 78 and 79 are only a few feet apart. In figure 78, a block of olivine pyroxenite and a gravel-like deposit of smaller fragments truncate layering in peridotite. The layers are not only depressed by the fragmental deposits, but parts of them have been removed. Cross layering is evident in the upper right corner of the photograph and in figure 79. The different sets of layers are identical in composition and character, and as with the fragments, some material must have been removed from the lower layers and transported laterally. No evidence of slumping is apparent.

Figure 80 is another example of cross layering which probably requires currents for an explanation. The bottom-set and fore-set layers differ slightly in composition, but they are both peridotitic. This indicates that neither one is a layered block, because all definitely-recognized blocks and fragments are olivine pyroxenite. The geological hammer serving as scale in the photograph rests on a block that shows the contrast between olivine pyroxenite and peridotite.

Figure 81 is another example of cross layering in peridotite. Olivine pyroxenite blocks are common in the area, and a large one occurs just below the geological hammer.

Figure 82 shows a rumpled peridotitic layer whose top was apparently planed or scoured flat before the next layer was deposited. A small block of olivine pyroxenite is visible in the layer near the upper left corner of the photograph. This and similar fragments probably have caused the disturbance of the rumpled layer. The current which is supposed to have caused the scouring apparently was too weak to move the block, but draped the next layer over its upper surface.

Figure 83 shows cross layering in which all of bottom-set, fore-set, and top-set layers are present. On first inspection, the fore-set layers seem to constitute a layered dike, and they are slightly richer in olivine than the others. However, the layers of the other two sets do not match. Figure 84, a sketch of the same outcrop, shows how the pattern of the top-set layers swings parallel to their contact with the crossed layers. Scour is visible in the fore-set layering, and a tiny feature suggestive of cross layering occurs in the bottom-set layers. Nothing indicates that the disconformable relation is due to slumping localized in one or more of the sets. Current erosion is the best explanation of the structure.

Figures 85 and 86 are sketches made by E. J. Oliver. Figure 85 shows repeated cross-layering in peridotite. Fragments and a relatively large block of olivine pyroxenite occur in the outcrop just to the right of the layering. Figure 86 shows an impressive example of truncation and current erosion of deformed layers in peridotite. Three

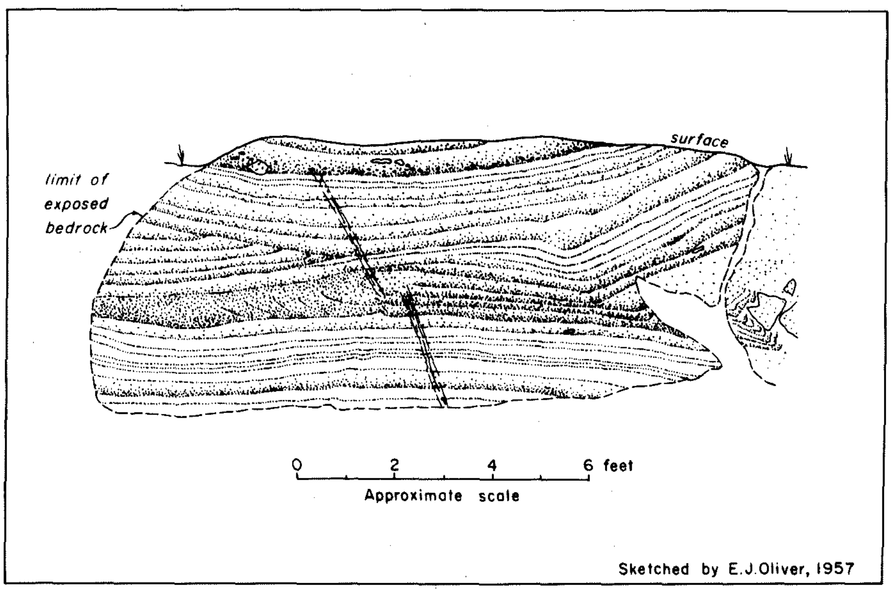


Figure 85. Cross-layering in peridotite, Hall Cove ultramafic area.

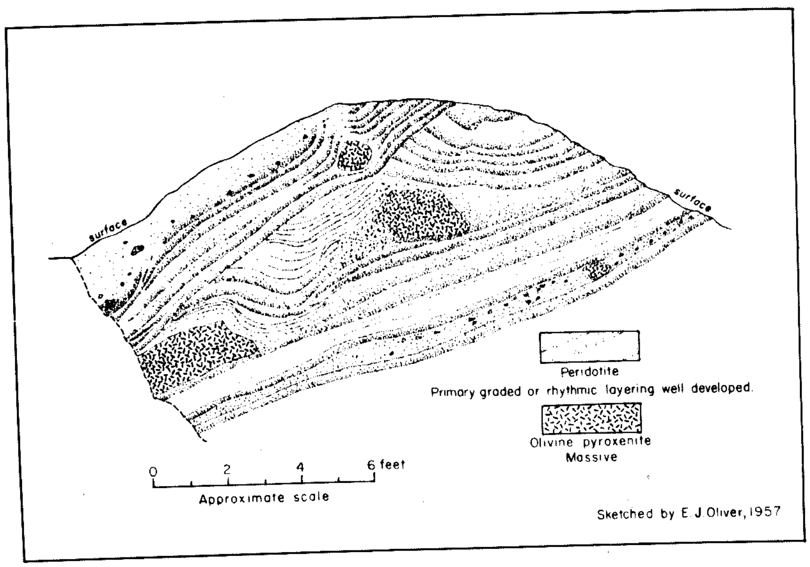


Figure 86. Layered structures in peridotite related to olivine pyroxenite blocks.

moderately large fragments of olivine pyroxenite are present in the exposure shown in the sketch, and near this, but not shown, are some very large blocks of this material. These very large blocks may have caused most of the deformation. The sketch shows that the layers drape over one of the blocks but are partly eroded. The thick graded layer on one side of the middle block degenerates into many thin layers on the other side. This could, perhaps, indicate that the block has caused an eddy in the main current.

On the whole, evidence of current scouring is sparse in the ultramafic layering, and although the writer believes the above described features are examples, it is only fair to say that some at least are subject to other interpretation. If currents did operate to form the layering, they have been primarily a means of transport and deposition, not erosion. In most of the examples of scour described above, blocks of olivine pyroxenite are present, and their position in the section commonly corresponds to the surface of scour. Three possible reasons for this relationship are:

(1) The currents which accompanied introduction of blocks were stronger than normal and were capable of picking up fine crystals of earlier layers.

(2) The blocks deformed the layers on which they fell, and the currents scoured the irregular

surface. Disturbed layers probably would be less resistent to erosion.

(3) The blocks were barriers and upset the pattern of current flow. Erosion of layers occurred during the process of establishing a surface in equilibrium with the new pattern.

Current Transport

Fragmental Layers, Evidence of Lateral Transport.

The fragments and blocks would, by a direct settling hypothesis, be pieces of olivine pyroxenite that became dislodged from the roof of the magma chamber and settled to the floor. By any simple process of dislodgement, the fragments would be expected to accumulate in irregular piles. However, many of the fragments occur in well-defined layers with relatively constant thickness and considerable lateral extent. In the 1700 feet of layered rocks described in lithologic sections A, B, and C, there are more than 50 of these layers. Their formation is no coincidence. They must have been spread over the floor, and the best spreading mechanism is one of magmatic currents moving laterally.

Quartz Inclusions, Evidence of Vertical Transport.

Wager and Deer (1939, pp. 264-266) describe xenoliths of gneiss in the Skaergaard intrusion which were partially fused and now appear as granophyre. Considering the relative densities of the xenoliths and the Skaergaard magma and other geological evidence, they conclude that both the original and the fuxed xenoliths should have floated, but the xenoliths occur in the layered series. On these occurrences, the authors state (p. 265):

> "Since the layered rocks accumulated from the bottom upwards as a solid mass of sediment, the granophyre patches cannot have floated up from below, but must have been deposited like the crystal precipitate which forms the layered rocks. On the evidence of the relative specific gravities given above, it is most unlikely that the granophyre could have sunk through the Skaergaard magma, especially at the time the dense hortonolite ferrogabbro was forming. We consider that the granophyre must have been carried down by the magmatic currents, and held there by the simultaneous deposition of the cloud of crystals which formed the layered series and which had a mean density greater than the magma."

The quartz inclusions in the layered peridotite of the Hall Cove ultramafic area present exactly the same problem. According to Dane (1942, p. 36), the density of liquid basalt at 1200-1250°C is 2.60-2.65 gm/cm^3, and the density of liquid diopside at 1391°C is 2.67. The density of quartz at 1000°C is slightly under 2.55 and would probably be even less at 1200-1300°C. (The phase diagram for SiO_2 shows that tridymite would not form at the pressures involved. Tuttle and Bowen, 1958, p. 29.) Thus the quartz should have floated. The author agrees with Wager and Deer and believes that the quartz inclusions show downward transport of material by magmatic currents.

The quartz inclusions are generally in or at the top of a fragmental layer. It has been suggested that the currents accompanying the introduction of the fragments were stronger than normal, and this may have been the condition required to bring the quartz down to the floor of the magma chamber.

By the direct settling hypothesis, the concentration of settling crystals is not likely to have been great enough to carry the quartz down, nor is it probable that the mafic minerals would make a stable pile on small, equant pieces of quartz floating freely in a magma of low viscosity. The only way the quartz could settle hundreds of feet in a denser magma is by having the mafic minerals attach themselves so as to increase the total density of the fragment. The quartz now occurs as discrete fragments which part readily from the enclosing rock, and the reaction rim around most of the inclusions is thin. Furthermore if the settling process took very long the quartz would probably have dissolved as the magma was clearly under-saturated with silica. Rapid transport and burial, insulating the quartz from all but the interprecipitate magma, is a more plausible explanation.

Other Evidence

A structure suggesting lateral transport of ultra-mafic material by currents is shown in figure 45. The angular block of olivine pyroxenite has dropped on layered peridotite and deformed it. Part of the layered material appears to have been swept to the right and redeposited. It does not seem possible to account for the structure by slumping; the redeposited material forms a layer which is too regular over too great a distance. A current, having a component of motion to the right with respect to the direction in which the photograph was taken, is indicated.

The streamlined manner in which the layers drape across individual blocks is suggestive of current flow (fig. 87). The layers generally thin over the top of blocks and thicken off their edges. This is not always, if ever, due to slumping from the top because the upper surfaces of many blocks conform with the plane of the layering (fig. 88). The layering tends to change thickness so that any irregularity is removed in as little vertical distance as possible. If the layers are thin relative to the size of the irregularity, then the effect of the latter may persist for a considerable vertical distance, but becomes progressively less pronounced (fig. 26). If the layers are comparatively thick or the irregularity small, then a planar surface may be re-established after the deposition of only one or two strata. For example, the block in figure 45 has no effect on the configuration of the layer above it. By a direct settling hypothesis without currents, there seems to be no obvious reason why

Figure 87. Peridotite layers deposited or draped over an olivine pyroxenite block. Peridotite zone, Hall Cove ultramafic area.

Figure 88. Draping of peridotite layers over an olivine pyroxenite block. The layers are thinnest on top of the block, and some are absent.

the layers should consistently thin over the blocks.

Blocks of olivine pyroxenite with dimensions measured in tens and hundreds of feet have caused considerable deformation of the layers on which they fell. They must at the same time have caused disturbance and stirring of the magma above the floor. Several dozen of these blocks are shown in plate 2, and if their distribution in three dimensions is considered, they must number in the hundreds. One can deduce that currents were common, if not intense. Stirring of the magma would probably destroy the sorting pattern of crystals in a direct settling hypothesis, hence graded layers would not be expected. Yet some of the best developed graded layers occur against blocks or in sections alternating with them. Thus, it can be inferred that disturbance of the magma helps, not hinders, the formation of layers.

Analogy with Sedimentary Deposits

In recent years a rather distinctive type of sedimentary series has been recognized by students of sedimentation and has been generally attributed to turbidity current deposition. Probably the outstanding single feature of the series is the abundance of graded bedding. As graded layering is so prominent in the Duke Island ultramafic rocks, it is interesting to make a further comparison of the two types of deposit. Kuenan and Carozzi (1953, pp. 364-365) list 21 characteristics of sedimentary series typical of the action of turbidity currents. Three of these pertain to fossils and shale inclusions and are not pertinent to this comparison, and five others are criteria for determining direction of current flow. The other 13 are summarized below.

All or most of the series show:

1. Interstratification of fine-grained deposits with coarse-grained beds.
2. Regular bedding. In the few cases in which a bed pinches out, the next one re-establishes the original bottom slope.
3. Absence of wave ripple mark, channel scour, coarse or mutually opposed cross-bedding, and other sedimentary features indicating small depths.
4. Slump structures and pull-aparts.

In the coarse-grained beds of the series, one finds:

5. Grading. (In the sediments, grading is commonly restricted to particles finer than 10 cm; the presence of a limit is the important point in comparison with igneous grading.) Grading is commonly combined with lamination, especially in the finer grain-sizes.
6. The sorting in each horizon of the graded bed is poor. (The sorting becomes progressively better upward. Pettijohn, 1956, pp. 171-172.)
7. The lower contacts are always sharp, the upper contacts generally so, but some coarse beds grade into fine ones.
8. Load casts formed by coarse beds sinking into the underlying "pelagic" beds.
9. Convolute bedding.
10. Absence or insignificance of scour at the base.
11. Generally speaking, the beds with abnormally coarse grain for any one locality are also the thicker ones in the section.
12. Each bed maintains its own characteristic features over its entire length in an exposure.

A final characteristic:

13. Slide conglomerates or breccias occur among the normal rocks. The conglomerates can be entirely isolated, or they may be present in groups with or without graded beds or "pelagic" beds in between them. Although generally emplaced

> without erosion of the underlying formation, local stripping of the underlying deposits may occasionally be noted. Although imbrication is possibly present in some places, it is inconspicuous. Cross bedding never goes right through such conglomeratic beds. They tend to have streaky internal structure and to contain cobbles or even boulders.

Almost every one of the above characteristics has a relatively good analogue in the layered ultramafic rocks of Duke Island. The interstratification of coarse and fine strata, the marked regularity and continuity of strata, the tendency of strata to re-establish original slope, the presence of slump structures, the association of grading and lamination, the nature of the sorting, and the direct relation of thickness to particle size are all typical of the igneous layering. The slide conglomerates are virtually identical to the fragmental layers.

Most of the features for which analogy is absent or tenuous are not surprising because they are even uncommon in the sediments. Pull-aparts have not been recognized at Duke Island, but it is interesting that features described by Hess (1938, p. 265) in the Stillwater complex and by Carr (1954) in the Cuillon complex seem to be reasonable facsimiles of this structure. Load casts exactly like those in sedimentary rocks are not present in the ultramafic rocks, although the depressions in layers beneath blocks are a kind of load cast.

The similarity of sedimentary and igneous features is a strong point in favor of the current hypothesis.

Currents are capable of producing every characteristic feature and, probably, most of the structures observed in the ultramafic layers, even though the precise mechanics may not be completely understood.

The criteria given by Kuenan and Carozzi for determining the direction of current flow apply to coarse beds and are as follows:

(1) Flow marks.
(2) Current ripple marks.
(3) Small scale current bedding.
(4) Tendency of pebbles to lie with longest axes parallel to the direction of flow.
(5) Horizontal grading, with dimunition of grain size away from the source. This happens too gradually to be observed in an exposure of normal size.

These features have not shown up well in the ultramafic rocks. The rocks do not split along layering planes, and seldom is there opportunity to look for flow marks. Features (2) and (3) apparently do not occur. Orientation of crystals is rare, but evidently orientation of pebbles in the type of sediment described by Kuenan and Carozzi is not common either because they have to call upon observations of others. Horizontal grading has been noted in the ultramafic layers, particularly the fragmental layers, but generally in at least two directions. This probably is because the fragments are introduced locally, and as a consequence, the continuity in the character of the layers is limited both away from the source and laterally. It is noted that the fragmental layers are almost entirely confined to a fan-shaped area in the peridotite zone of the Hall Cove ultramafic area. The fan has its apex in the northwest corner of the zone and spreads to south and east to the southern boundary of the zone. As shown in lithologic sections A, B, and C (plate 8), fragmental layers are fewer and thinner to the south and east, and the fragments are smaller. This may indicate a general southeast current flow for this part of the complex.

MECHANICS OF CURRENT GENERATION

Wager and Deer (1939, pp. 267-268) propose the following mechanism to account for the regularity and persistence of the layering in the Skaergaard intrusion.

> "The hypothesis which we wish to put forward to account for the circulation in the Skaergaard intrusion is based on the idea of convection, due to the increase in density of the magma on cooling. Without the separation of a solid phase, slight increase in density would take place on slight cooling, but if some crystallization also took place there would be a much more marked increase in the mean density of the liquid plus crystals...Grout (1918, A and B) has extended the term, two phase convection, to cover the case of convection due to formation of solid phases in the liquid; it is a two phase convection rather like that visualized by him which we postulate. Cooling of the Skaergaard magma must have taken place from the top and sides of the intrusion, but mainly from the former. The cooled and partly crystallised magma along the top and walls would have a higher specific gravity than the rest and would sink. Thus a convective circulation would be established in which currents descended along the walls, crossed the floor, to about the centre, then rose to the top where they could spread out

> toward the walls again. While passing along the top of the intrusion the magma would be sufficiently cooled and increased in density to sink again and repeat the process. Such a circulation must tend to be established in any liquid which increases specific gravity on cooling, and which is cooled from the upper surface."

Production of repeated layering is attributed to variation in the velocity of the currents thus affecting their capacity to carry particles of different densities. This also applies to particles of different sizes, although Wager and Deer were not confronted with the problem of vertical grain size gradation in the layers of the Skaergaard intrusion. Factors which they believe to favor convection in the Skaergaard magma are:

(1) The shape of the intrusion. Convection cells tend to have diameters four times their thickness, which is approximately the ratio for the Skaergaard magma during much of its crystallization.

(2) Low viscosity in the magma. They quote Becker who stated that basaltic magma has a viscosity only 60 times that of water.

The complete three dimensional configuration of the Duke Island ultramafic rocks is not known, but in view of the thickness of layered rocks (ca 12,000 feet), dimensions favorable to convection may have existed. The viscosity quoted by Wager and Deer seems inordinately low in comparison with more recent figures. For example, Birch and Dane (1942, pp. 133-137) give viscosities of basalt glass at

1200°C ranging from 700-3200 poises, and MacDonald has calculated figures of 3×10^3 to 2×10^4 poises for Mauna Loa lavas (1954, pp. 170-173) and 2.2×10^4 to 3.8×10^4 poises for Halemaumau lavas (1955, pp. 90-91). Birch and Dane list the viscosity of diopside glass to be 106 poises at 1200°C, and viscosities measured in the system CaO-MgO-Al_2O_3-SiO_2 (Machin, Yee, and Hanna, 1952) suggest that the magma from which the Duke Island ultramafic rocks crystallized was even more fluid than basaltic magma. Thus two-phase convection is just as plausible an explanation of the Duke Island layering as it is of the Skaergaard layering.

A mathematical treatment of convection is extremely complex, but it is of interest to examine the hypothesis of two-phase convection by some simple calculations of the rate of crystallization of the magma and the settling velocities of crystals. If the magma body is considered to be a semi-infinite sheet cooling through its upper surface, the heat flux through the upper contact is:

$$dQ/dt = - K\, dV/dX$$

where Q is the quantity of heat, t is time, K is the coefficient of thermal conductivity of the rock at the contact, and dV/dX is the thermal gradient at the contact. An average value of K is about 5.0×10^{-3} cal/sec.cm°C (Birch, 1942, pp. 251-258). It will be assumed that the thermal gradient at the contact is constant during the crystallization of one-foot (ca. 30 cm) layer and, for purposes of illustration which will be obvious below, equivalent to the fantastically large value of 1000°C/100 meters, or 0.1°C/cm. The heat of fusion of a diopside-forsterite mix is about 100 cal/gm, and the density of the layer is about 3.0 gm/cm^3. The time of crystallization of the layer is, then,

$$\frac{30 \times 3.0 \times 100}{5.0 \times 10^{-3} \times 0.1} = 1.8 \times 10^{7} \text{ seconds, or about 7 months.}$$

Using Stokes' law and assuming a viscosity of 400 poises, it can be shown that a 2-millimeter sphere of diopside in a melt of its own composition could settle about 1500 feet in 1.8 x 10^7 seconds. Thus, in the time it takes to crystallize a one-foot layer, one of the smaller crystals could settle a distance equivalent to a major fraction of the total exposed thickness of layered rocks. It is difficult to imagine how a density contrast large enough to induce convection would be established by this model. As soon as the smaller crystals reached the floor, steady state settling should prevail. A more realistic thermal gradient only makes matters worse.

If convection currents did operate to form the layering, then cooling through the walls of the magma chamber must have been appreciable. Depending on the shape assumed for the magma body, the area of cooling surface may be increased three to four times over that assumed in the above model, and crystals along the walls would be in a position more favorable to convection. Steep, inward dipping walls would also help in that the crystals settling to them would roll and tumble downward in a slow, partially hindered descent. In this way, they might concentrate to the point at which slumping would occur. One slump could instigate others in a sort of cascade effect. Thus, intermittent turnover of the magma could be accomplished, and the currents produced would be of fluctuating velocity and capable of depositing graded layers. Evidence of slumping is present at Duke Island (fig. 39), and slump structures have been reported in other layered igneous complexes (Carr, 1954).

The heat flow considerations suggest that the process should work best early in the crystallization period when cooling was greatest, and layering is notably better developed in the stratigraphically lower parts of the rocks of the Hall Cove ultramafic area. Shallow intrusion would have the same effect and is a condition observed in the Skaergaard intrusion but probably not satisfied at Duke Island.

EVOLUTION OF THE PERIDOTITE DIKES

It is believed that the explanation of the peridotite dikes occurring in the Hall Cove ultramafic area must satisfy the following conditions.

(1) The olivine pyroxenite blocks in which the dikes are emplaced are large inclusions that have fallen to the floor of the magma body and have thus been incorporated into the layered series.

(2) The peridotite dikes and their layering are of the same age as the layered peridotite in which the blocks are embedded.

(3) Graded layering is produced by currents.

The possible mechanisms for the formation of the dikes are divisible into the following classes:

A. Once intruded, the dikes are essentially closed systems with no addition or loss of material. In order to explain the layering as a progressive accumulate, it probably is required that either:

(1) The material injected into the dikes was largely liquid, and the layering developed in situ during crystallization.

(2) The material was largely solid, the layering having formed prior to injection.

B. The dikes (particularly those with layering) were open systems; material was added and removed after the space now occupied by the dike was established.

A(1). If the dikes were injections of liquid and if they existed as closed systems while crystallizing, a highly olivine-rich liquid is required. The dikes might then be evidence of a liquid peridotite magma. However, such an origin is considered improbable for the following reasons:

(a) Most of the dikes are open at one end, and many are open at both ends. The dike probably could not exist in this form during crystallization without crystals of the accumulating layered series being added or liquid escaping. Thus, the dike cannot be considered to have the same composition as the liquid magma which existed at the time.

(b) The graded layering shows that the magma must have been undergoing fractional crystallization. Reason has been given to indicate that olivine pyroxenite is the product of simultaneous crystallization of olivine and pyroxene. Thus a peridotite magma undergoing fractional crystallization in a closed dike would be expected to give a differentiated product of dunite and olivine pyroxenite. No indication of such differentiation is evident.

(c) It is unlikely that a system of currents capable of producing layering could operate within the restricted confines of a narrow dike. This might be construed as evidence against the current hypothesis, but is only one weak argument against much favorable evidence.

It would seem better to seek some other explanation for the dikes to which the current hypothesis can be adapted.

A(2). The similarity of the peridotite in the dikes to that in the surrounding layered series suggests that the dikes could be intrusions of the layered series that still contained appreciable interprecipitate liquid. Deformation of the peridotite layering attributable to the impact of large blocks is common. As the dikes intrude only blocks, it is proposed that their intrusion into a block could have accompanied its impact on the layered series. If the blocks were fractured, which seems possible in view of their angular habit, then it is conceivable that they would break apart upon touching down, and as they settled into the peridotite because of their momentum and large weight, the loosely consolidated layered material might be squeezed up into the fractures. Some of the relationships shown on plate 4, particularly locality F, are suggestive of such a mechanism.

It might seem that the mechanism would more than account for the irregularity in the layering of many dikes. Much internal flow and deformation is expectable. However, layers with little or no deformation are abruptly cut off by blocks in many places (e.g. plate 4, loc. A; plate 5, loc. D and H). These contacts are essentially faults. The same relationship might characterize the dike contacts because the distance that layered peridotite must be intruded is small, and interprecipitate magma could lubricate the contact.

The intrusion-upon-impact mechanism probably could not account for the dike at locality B in plate 5 (fig. 50). The relationship requires that the dike be squeezed down rather than up. That this could be caused by the impact of the large overlying (southern) block is unlikely.

An alternative mechanism whereby even the dikes in plate 5 are accounted for by semi-solid intrusion is as follows. The presence of well-consolidated blocks in the layered series probably resulted in considerable differential compaction as accumulation continued. Shifting and fracturing of the blocks could have occurred as part of the adjustments. This could also result from any other movements which might occur around a magma body. Thus, fractures might open in the blocks and be filled with the still-soft layered peridotite. Expulsion of interprecipitate liquid would carry crystalline material into the long, narrow dikes (e.g. plate 5, loc. F and G). The origin of the dikes would thus be essentially that generally proposed for sandstone dikes (Pettijohn, 1957, pp. 191-192).

B. If the dikes were open systems, liquid may have formed the initial intrusion and been subsequently displaced by crystalline material introduced along with the accumulation of the main layered series. Some of the dikes may well be the filling of crevices in shattered blocks resting on the floor of the magma body. Figure 89 is particularly suggestive of this type of process. In this example, the three bodies of olivine pyroxenite may be unrelated blocks, but the manner in which the peridotite layers fill in between the blocks is the feature important to the present discussion.

In plate 5, the whole east-trending (horizontal) strip of peridotite as well as the definitely transgressive material might be accounted for as dikes if somehow the space which they now occupy could have been opened and maintained until the liquid was displaced by crystals. It must be remembered that the dike is emplaced in a large block and the magma in the dike would be continuous with that surrounding the block. Consequently the dike needed solid support to hold it open, or it would have closed off and expelled its contents. Presumably the space for the dike could have been established when the olivine pyroxenite block landed on the floor, and the solid support to hold the dike open need not be completely visible at the present erosion level. Currents in the main body of magma could have resulted in periodic introduction of crystalline material into the dike, or perhaps even continuous flow of magma laden with crystals through the dike. Small olivine pyroxenite blocks might be carried in, or simply be dislodged from the dike walls, to be included in the layered peridotite accumulating in the dike. Eventually, shifting of the blocks could expell residual liquid, deform the

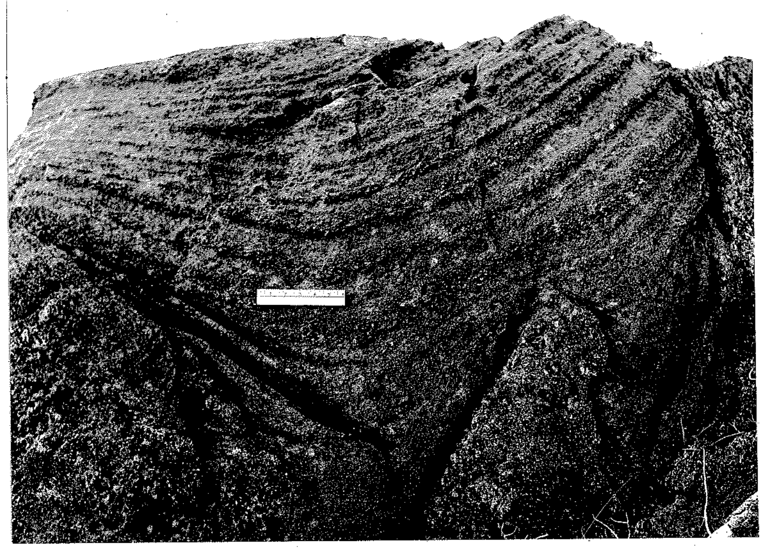


Figure 89. Layered peridotite filling the space between olivine pyroxenite blocks. Compare with figure 49.

layering, and result in the apparent impression that the large upper (southern) block has fallen on the peridotite layers.

In conclusion, none of the above mechanisms is considered to be completely satisfactory. Certainly, it is unlikely that the dikes represent the exact composition of the intruded liquid; the environment in which the dikes formed is one of accumulation of crystals from a large body of magma, and the dikes are clearly open to the influx of this material. The writer favors the solid intrusion mechanisms, particularly the one in which the blocks are shifted subsequent to burial by compaction or deformation. However, the lack of extreme deformation in the dyke layering is difficult to rationalize, as are some of the relationships shown in plate 5. None of the mechanisms proposed here seems to adequately explain the structure shown in figure 51.

PART V. SUMMARY OF THEORY AND SEQUENCE OF EVENTS

STRUCTURAL FEATURES

The Hall Cove and Judd Harbor ultramafic areas show a rather striking parallelism of structural features in the olivine-bearing rocks. These features are summarized in table 14. It would probably be a point in favor of a hypothesis if it could account for the development of each feature in one area simultaneously with that of the comparable feature in the other area. If the two areas are the outcrop of the same body, they might both be expected to reflect in their rocks any major event in the history of the complex.

PREFERRED HYPOTHESIS

The most simple and straightforward sequence of events the author can suggest to account for the structural features of the various rocks in the Hall Cove and Judd Harbor ultramafic areas is the following:

(1) A large body of ultramafic magma was emplaced into pre-existing gabbroic and metasedimentary rocks. The minimum dimensions of the body were about 7 miles long, $3\frac{1}{2}$ miles wide, and 2 miles thick. Most of the magma was probably injected in a relatively short period, but minor additions may have been made to the chamber at various times during its crystallization. The major part of the

TABLE 14. COMPARISON OF THE STRUCTURAL FEATURES IN THE OLIVINE-BEARING ROCKS OF THE HALL COVE AND JUDD HARBOR ULTRAMAFIC AREAS

Hall Cove ultramafic area	Judd Harbor ultramafic area
1. At least two major periods of intrusion. (a) The intrusion from which the main zone of peridotite formed cuts at least part of the olivine pyroxenite. (b) Blocks of olivine pyroxenite in peridotite are inclusions.	1. At least two major periods of intrusion. (a) The intrusion of the main body of dunite and peridotite (or the magma from which it formed) cuts the olivine pyroxenite. (b) Areas of olivine pyroxenite on the north side of the dunite zone apparently are the outcrop of inclusions or roof pendants.
2. All olivine-bearing units contain layering, regardless of their age. Layers in dunite and peridotite are discordant with respect to at least part of the olivine pyroxenite layering.	2. All olivine-bearing units contain layering, regardless of age. Layers in dunite and peridotite are discordant with respect to the olivine pyroxenite layering.
3. A major syncline occurs in the olivine pyroxenite. Its axis trends N.65°E., plunges 40°-60°S.W.	3. A major anticline and syncline occur in the olivine pyroxenite. Their axes trend N.70°-85°E., plung 60°-70°S.W.

olivine pyroxenite and hornblende pyroxenite now visible in the two areas are crystallization products of this magma. The northern and southern olivine pyroxenite zones in the Hall Cove ultramafic area are assumed to be of the same generation. Rocks richer in olivine, not visible at the present level of erosion, may have formed during the early stages of crystallization of the magma. Circulation of the magma, accompanied by gravitational settling of crystals, is believed to have caused the graded layering. Crystallization differentiation could have resulted in a late magma capable of precipitating the parts of the hornblende pyroxenite which now appear to cap the layered olivine pyroxenite (e.g. the Kelp Island occurrence, and the hornblende pyroxenite pierced by the drill hole along section AA', figure 60). For the most part, however, the explanation chosen for the border zone is one of transfer of lime, iron, silica, and other components to the periphery of the body, and reaction of these materials with olivine pyroxenite (pp.230-241). Most of the basic pegmatite was derived from this intrusion, as were the materials which caused the alteration of pyroxene gabbro to hornblende gabbro.

(2) A second major intrusion occurred. The primary rocks of the main zones of peridotite and dunite formed from this body. The intrusion was emplaced both into the central part of the previous olivine pyroxenite and

along the boundary between olivine pyroxenite and gabbro. The intrusion must have come in contact with pre-ultramafic rocks in the Hall Cove area, or enroute to this site, in order to acquire the quartz inclusions. The size of the intrusion, according to exposures, was only a small fraction of the size of the first intrusion. It is suggested that the second intrusion, though continuous, had two principal loci and raised the rocks at these points so that they are now exposed in the Judd Harbor ultramafic area and the eastern part of the Hall Cove ultramafic area. Evidently the intrusion was forcibly emplaced only shortly after the crystallization of the olivine pyroxenite in the first intrusion because it had the following effects on the earlier rock:

A. In the Hall Cove ultramafic area

(1) Tilting, fracturing, and minor faulting of the northern olivine pyroxenite zone.

(2) Folding of the southern olivine pyroxenite zone to produce a syncline, and possibly, faulting.

(3) Folding and irregular deformation of the olivine pyroxenite now visible east of the Hall Cove fault. (The fault had not yet developed.)

(4) It is possible that a small amount of hornblende pyroxenite magma was still present

as a residuum of the first intrusion. Squeezing of this material by the second intrusion might account for some of the irregularities in the distribution of hornblende pyroxenite.

B. In the Judd Harbor ultramafic area

(1) Folding of the olivine pyroxenite to produce the anticline and syncline, plus minor faulting.

(2) If faults actually exist along the northern boundary of the rocks in the area (p. 164), the intrusion might have been the cause, by punching the olivine pyroxenite upward into the overlying gabbro. By this process, the normally continuous hornblende pyroxenite border zone would be removed.

(3) The folding accompanying the second intrusion might also have upset the normal development of a continuous border zone of hornblende pyroxenite. The late magmatic fluids, instead of migrating outward, may have been forced into areas of low pressure or dilation. This is a possible explanation of the occurrence of a zone of basic pegmatite dikes and hornblende pyroxenite through the center of the olivine pyroxenite zone. Some of the

late magmatic fluids of the first intrusion may have been incorporated into the magma of the second intrusion so that a hornblende pyroxenite zone is absent from the dunite areas.

The second intrusion must have been considerably richer in the components of olivine than the first. However, for reasons given on p. 199, it is not believed to have been dunitic or even peridotitic liquid. It is postulated, following the suggestions of Bowen (1928, p. 158), that the magma contained a major proportion of crystals; a figure of about 50 per cent will be assumed. This cannot be proved, but the character of the layered deposits in the peridotite zone of the Hall Cove ultramafic area is suggestive of rapid continuous accumulation and may indicate the availability of larger amounts of crystalline material than could be accounted for by crystallization of a liquid magma during the time of their accumulation. The liquid fraction probably had a composition on the olivine side of the cotectic boundary between olivine and clinopyroxene, and 35 per cent olivine equivalent is considered a plausible figure. As such, the composition of the liquid may be very close to that in the first intrusion. Most of the crystalline material is believed to have settled immediately, accumulating as dunite. Toward the end of this period, as crystallization proceeded, circulatory movement was established in the liquid portion, resulting in deposition of layered material on top of the dunite. Because of the composition of the liquid, the early formed crystals would be primarily of olivine. However, under conditions of convective overturn, parts of the liquid magma in the zone of cooling along the walls and roof of the chamber are believed to have crystallized to the extent that their composition touched the cotectic boundary. Consequently, some pyroxene formed until continued circulation of the magma brought new magma to the zone of cooling. Thus, a peridotitic rock rather than pure dunite could develop in the layered series. Burial would prevent resorption of the pyroxene in another part of the liquid. Eventually, continued crystallization caused the entire magma to reach the cotectic boundary, and olivine pyroxenite then formed. Most of this has been removed by erosion from the exposed parts of the intrusion, but some remains in the Hall Cove ultramafic area, making up the layered rocks described in lithologic sections B and C.

The olivine pyroxenite of the first intrusion was considerably fractured and broken by the second intrusion. Blocks and fragments of it repeatedly broke loose from the roof and walls of the second magma body while the latter was crystallizing, particularly in the early stages, and became incorporated into the new layered rocks. Slumping of the blocks from the walls may have been a major factor in promoting convective circulation for the second time. Once started, the currents would loosen other blocks and the process would be self-perpetuating. The peridotite dikes developed sometime during or shortly after this stage.

Replacement dunite in both areas is considered to be largely a metasomatic effect of the second intrusion on the rocks of the first. The replacing fluids apparently invaded zones of dilation in the older rocks, such as the fold axes in the Judd Harbor ultramafic area, and the deformed parts of the Hall Cove ultramafic area. They also attacked parts of the olivine pyroxenite near the margin of the second intrusion.

(3) Emplacement of the granitic rocks is believed to be the cause of southward tilting of layered rocks in the peridotite zone of the Hall Cove ultramafic area; the original angle of dip of the layering is not known, but tilting of at least 20° probably is involved. Some of the outlying ultramafic masses surrounded by granitic rocks may represent portions of the main complex that have foundered in the granitic magma.

(4) Faulting occurred in the rocks of the area. The main break is the Hall Cove fault. Joint swarms represent shearing accompanying this faulting and affect gabbroic, ultramafic, and granitic rocks.

(5) Serpentinization occurred along the joint swarms.

Serpentine might also have formed during some of the earlier events but cannot be distinguished.

(6) Diabase dikes were intruded.

(7) Final erosion resulted in the present land forms. No explanation can be offered for the persistence of Mount Lazaro above the otherwise-flat island. Plate 7 is an exploded isometric block diagram interpreting the shape of the main ultramafic body at depth after the completion of this process.

ALTERNATIVE HYPOTHESIS

It has been indicated under the description of the Hall Cove ultramafic area (p. 155) that two interpretations are possible with regard to the relationship of the peridotite zone and the southern olivine pyroxenite zone. The above sequence of events assumes that the contact of these two units is intrusive, peridotite being the younger rock. An alternative interpretation puts the southern olivine pyroxenite zone stratigraphically above the peridotite as a later differentiate. A possible sequence of events is shown in figure 90. The line of section is approximately from south to north through the middle of the Hall Cove ultramafic area. The hornblende pyroxenite zone has not been shown on the diagrams. The events are:

(1) Intrusion of magma which crystallizes olivine and pyroxene under conditions of convective circulation, to give the northern zone of layered olivine pyroxenite.

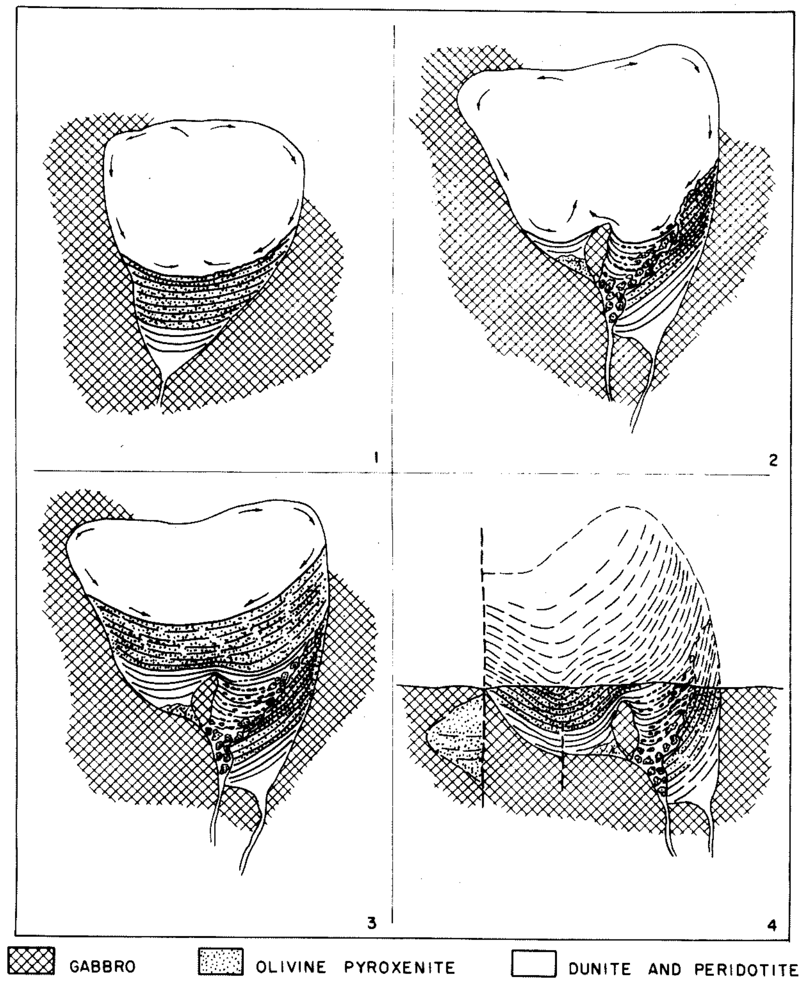


Figure 90. Diagrammatic representation of a possible sequence of events in the evolution of the rocks of the Hall Cove ultramafic area.

The magma may first have crystallized dunite as shown in figure 90-1, but these rocks have not been exposed. Minor slumping and break-up of early layers (figs. 38 and 39) may be a forewarning of event (2) below.

(2) A second injection of magma into the chamber (fig. 90-2). This intrusion caused tilting and disruption of the earlier layering. The mass of gabbro shown within the ultramafic body is a postulated extension of the lobe of gabbro that projects into the ultramafic body (plate 1) and is presumed to have been pried from the wall by the second intrusion. The precise interpretation of the behavior of the gabbro is not critical to the explanation of the relationship of the olivine-bearing rocks but is, nevertheless, a problem both here and in the preferred hypothesis.

The second magma was richer in olivine than that which crystallized olivine pyroxenite earlier but it does not have to be dunitic or even largely crystalline. It may have been of essentially the same composition as the initial intrusion. For a while, olivine is the principal phase to crystallize, and dunite and peridotite are deposited. The explanation for the formation of peridotite rather than pure dunite is the same as that used above (p. 307).

The earlier olivine pyroxenite was shattered and oversteepened by the second intrusion. As convective circulation proceeded, blocks of olivine pyroxenite repeatedly slumped and were included in the layered peridotite. Introduction from only the north side could explain the apparent decrease in the size and abundance of blocks away from the northern contact of the peridotite zone.

(3) Crystallization differentiation eventually resulted in the deposition of olivine pyroxenite again (fig. 90-3). Most of the flattening of the layering near the gabbro lobe would be a primary depositional feature.

Much of the difficulty with this particular interpretation is that other parts of the ultramafic rocks do not have an obvious parallel history, and in order to account for all the structural features, additional events must be postulated. By this mechanism, the folding of the syncline in the southern olivine pyroxenite zone of the Hall Cove ultramafic area must be attributed to some unknown outside force, whereas the folds in Judd Harbor olivine pyroxenite seem definitely related to the emplacement of the main body of dunite and peridotite. Similar difficulty is encountered in explaining the deformation in the olivine pyroxenite east of the Hall Cove fault. If it is assumed that the two main zones of dunite and peridotite are of equivalent age, then the Judd Harbor ultramafic area does not have a younger olivine pyroxenite zone to match the southern olivine pyroxenite zone in the Hall Cove ultramafic area. If the southern olivine pyroxenite zone at Hall Cove is considered to be equivalent to Judd Harbor olivine pyroxenite, the former area has no dunite to match the dunitic zone in the latter, and the problem of olivine-rich intrusion at Judd Harbor remains. The dunite zone in the Judd Harbor ultramafic area might be a solid intrusion of an early differentiate which originally underlay the olivine pyroxenite and which was equivalent in age to the peridotite zone in the Hall Cove ultramafic area, but the mechanics are unlikely, and a force adequate to explain the intrusion is not known. It is therefore concluded that this hypothesis is only a remote possibility. Perhaps the main thing it illustrates is the difficulty in interpreting the southern boundary of the Hall Cove peridotite zone as anything but an intrusive contact.

In conclusion, the problems encountered in explaining the internal structural features of the ultramafic rocks are many, even after an ultramafic magma is accepted. To solve them assuming the ultramafic and gabbroic rocks are derived from a common magma seems impossible. Although a negative argument, this, more than any other factor, has lead to the decision that Duke Island ultramafic rocks were formed from an ultramafic magma.

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APPENDIX I. PHOTO CREDITS

The photographs in the following figures were taken by James A. Noble.

2	38	50
13	39	54
15	40	64
16	41	72
26	44	87
29	46	88
33	47	89

Other photographs were taken by the author.