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HIGH TEMPERATURE ACID ROCKS ASSOCIATED WITH SERPENTINITE IN EASTERN QUEBEC

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ABSTRACT. Highly serpentinized alpine-type peridotites frequently have small, relatively acid rock lenses associated with them. Weakly serpentinized or unserpentinized peridotites are usually associated with segregations of olivine gabbro. A study of the acid rocks found in the highly serpentinized peridotite at Asbestos in Eastern Quebec indicates that they are metasomatically altered gabbro bodies. The initial alteration was to a biotite-saussurite which was further altered to a microcline-diopside-albite-quartz rock. The degree of alteration was dependent on structural attitude. These rocks represent high temperature alteration in the amphibolite facies which indicates that the serpentinizaion of the surrounding peridotite has taken place at a higher temperature than is represented by the general level of regional metamorphism, the epidote-amphibolite facies. Chemical evidence suggests that the components which were added to the gabbros may have been derived in part from the peridotite undergoing serpentinization.

INTRODUCTION

The association of small, relatively acid rock bodies with alpine-type serpentinized peridotites has been noted frequently by geologists working on serpentinites. The term "acid" is used in a relative sense in that actual silica contents may range from 40 wt.% to 80 wt.%. As this implies, the variations in mineralogy are large when the numerous world occurrences are considered and may even be large in a small area. Thus, "acid rocks" is here meant to distinguish those small, usually light-colored, lenslike and dikelike bodies that are found in highly serpentinized alpine-type peridotites and never make up more than 10 vol.% of the serpentinite mass.

The acid rock types usually found in this association include albitites, corundum albitites, alaskites, hornblende, and biotite granites, edenite amphibolites and calc-silicate rocks. Commonly these acid bodies have been considered late-crystallizing silica and water-rich residues of the peridotite magma and, as such, have been used as a source for the silica and water required to serpentinize ultramafic masses (DuToit, 1919; Gordon, 1921). Larsen (1928) has shown that this process is quantitatively impossible. At the present time, they are considered metasomatic and hydrothermal in origin (Turner, 1960; Larsen, 1928; Bowen, 1956, p. 132).

The work undertaken here traces the chemical and mineralogical changes that have affected a particular suite of such rocks in the serpentinite sill at the open pit of the Jeffrey Mine of the Canadian Johns-Manville Company, Ltd. at Asbestos, Quebec (fig. 1).

ACKNOWLEDGMENTS

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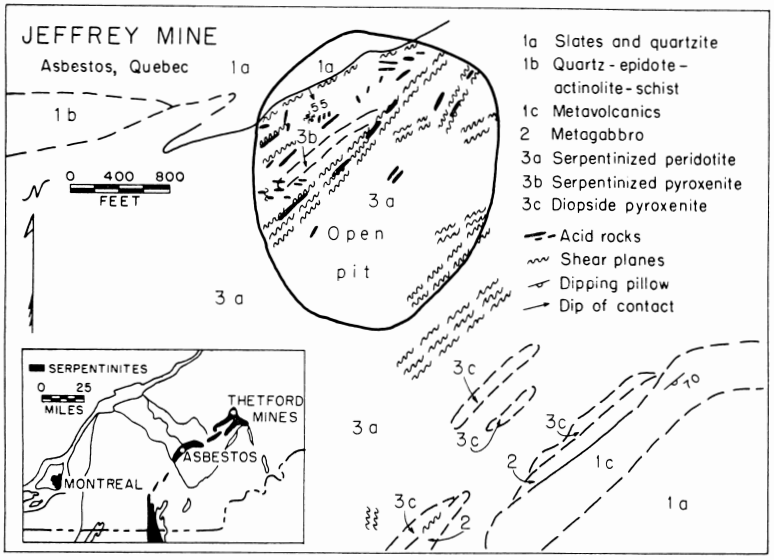


Fig. 1. Geologic sketch map of the open pit and immediate surroundings, Jeffrey Mine, Canadian Johns-Manville Co., Ltd., Asbestos, Quebec.

THE JEFFREY SERPENTINITE AND IMMEDIATE SURROUNDINGS

In this locality the serpentinite sill is 4000 ft thick and was emplaced between underlying siliceous slates, quartzites, and actinolite schists, and overlying metagabbro and metavolcanics (metabasalt and metaandesite) of the Caldwell Series which is of possible Grenville Age (Riordon, 1957). The upper contact is nowhere exposed so that it is impossible to ascertain the genetic relation, if any, between the gabbro and the ultramafic sill. These overlying rocks exhibit retrograde effects characteristic of the epidote-amphibolite facies. Basic plagioclase is altered to clinozoisite and oligoclase, and clinopyroxene is altered to actinolite and chlorite. The underlying metasediments are altered to minerals characteristic of the same facies. The lowermost Caldwell (Bennett Schists, not shown in fig. 1) locally carry hornblende, garnet, kyanite, and staurolite. The rocks immediately underlying the serpentinite consist of quartz, magnetite, and graphite with accessory muscovite, carbonate, oligoclase, and (corroded, detrital) microcline. Within this rock (1a, fig. 1), contact thermal effects are visible, in thin section only, for a distance of at most 4 inches away from the contact. Here minute sillimanite needles are sparsely developed and are associated with clouded Ab80 and fine grains of fresher-appearing Ab65. Some microlites of diopside are associated with corroded grains of primary carbonate. This development of andesine-diopside-sillimanite indicates hornfels development in the amphibolite facies. Beyond a few inches no hornfels minerals are found. A portion of the lower contact consists of actinolite-epidote-albite-quartz-magnetite schist (1b, fig. 1). The actual contact of this rock with serpentinite is not visible so that nothing is known of a hornfels development in it also.

The peridotites of the eastern Quebec "belt" average over 50 vol.% serpentinite (Cooke, 1937, p. 67). At the Jeffrey Mine the mass is virtually 100% serpentinite. The present mineralogy consists of three serpentine minerals (antigorite, chrysotile, and bastite), magnetite, and accessory amounts of presumably primary chromite. In some thin sections small amounts of talc and/or chlorite and/or brucite are found. Nickel (1958) has found traces of iron-nickel alloy ($FeNi_3$) in samples from Jeffrey. Minor amounts of millerite, pyrite, and diamond have been found in the past.

Because of the high degree of alteration, primary silicates are rare. They consist of Fo 90 (analysis 1, table 1) and En 93 (analyses 2, 3, table 1). The latter always contain exsolution lamellae of diopside (ext. angle = 37°) which results in about 2% CaO in the analyses. The upper third of the sill contains lenses of very coarse-grained diopside pyroxenites (analysis 4, table 1) which have been only superficially serpentinized along fractures and cleavages.

On the basis of abundance of bastite pseudomorphs after orthopyroxene, it was possible to distinguish one band in the lower third of the sill, which is dominantly a serpentinized orthopyroxenite (3b, fig. 1). However, it was found that the absence of such pseudomorphology is not a good criterion for designating a rock as serpentinized dunite. Many hand specimens show no bastite whereas their thin sections carry large amounts of fine-grained bastite crystals and orthopyroxene remnants that are usually brecciated and bent (pl. 1-E). This feature was also noted by Guild (1947), and Cooke (1937). Cooke (p. 70) suggests that the retention of bastite pseudomorphs depends on the original grain size of orthopyroxene. Tectonic activity that accompanies serpentinization results in a cataclastic texture whereby pyroxene and bastite are reduced in size and only the coarsest original grains are still evident in hand specimens. In this regard it must be noted that the upper third of the sill never carries any bastite even in thin sections. Company geologists have speculated that it may represent a serpentinized dunite. However, it is not designated as such in figure 1 because of the lack of corroborative evidence of any kind.

It was noted earlier that the serpentinite contact with the overlying rock types is not exposed, and that the partially exposed lower contact shows a weak hornfels development for a few inches into the metasediments. It must also be noted that this lower contact is quite irregular, and serpentinite in places is found mechanically squeezed into joints and fractures for several inches into the metasediments. Immediately to the west of the pit an arm of serpentinite protrudes into the metasediments for about 700 ft (fig. 1).

Within the Jeffrey serpentinite relatively acid rocks are found occurring as lensoid masses rarely over 60 ft in diameter and 15 ft thick. They are confined to the lower half of the sill and comprise about 5 vol.% of the whole mass in this locality. For purposes of description, biotite is used as a guide mineral. The masses consist of two distinct types: nonfoliated brown biotite-rich rocks and white biotite-poor rocks. In addition some intermediate varieties occur.

TABLE 1

	1	2	3	4	5	6	7	8
SiO_2	40.4	50.0	52.1	51.0	39.6	51.1	59.5	62.4
TiO_2	0.010	0.023	0.010	0.084	1.10	1.10	0.81	0.26
Al_2O_3	0.30	1.85	1.85	1.45	23.8	20.5	17.6	14.0
Fe_2O_3	—	1.34	—	—	1.06	—	—	—
FeO	9.6*	3.48	4.86*	2.64*	8.57	6.36*	4.95*	4.89*
MnO	0.14	0.17	0.16	0.078	0.10	0.041	0.038	0.13
MgO	48.8	36.0	36.0	19.0	5.85	4.50	2.85	4.00
CaO	0.20	1.75	2.35	26.0	8.90	6.20	4.20	9.40
Na_2O	0.06	0.10	0.10	0.14	1.70	3.06	4.04	5.60
K_2O	0.04	0.10	0.08	nd	4.85	4.00	4.25	2.70
H_2O^+	—	—	—	—	2.30	—	—	—
H_2O^-	—	—	—	—	0.14	—	—	—
ZrO_2	trace	nd	nd	nd	0.074	0.084	0.059	0.028
V_2O_3	nd	0.023	0.017	0.025	0.055	0.033	0.019	0.010
Cr_2O_3	0.030	0.86	0.72	1.10	nd	nd	nd	nd
NiO	0.29	0.12	0.09	0.032	nd	nd	nd	nd
SrO	nd	nd	nd	nd	0.115	0.042	0.031	trace
BaO	trace	0.003	0.001	trace	0.32	0.185	0.195	0.071
Mg/Mg^+Fe	0.90	0.93	0.93	0.93	0.53	0.56	0.51	0.59
Ca/Ca^+Mg^+Fe	—	—	—	—	0.36	0.36	0.35	0.50
Total	100.77	96.75	99.26	102.47	99.42	98.12	99.40	104.58

* FeO calculated from total Fe

** All iron calculated to total Fe for comparison purposes. No correction made for magnetite in rock.

BIOTITE-RICH TYPE

Analysis 5 (table 1) is the average of three analyses of this rock type. The rock consists of about 47 vol.% pleochroic reddish brown biotite (analysis 10, table 1), 2-3 mm across, occurring as clumps of sheaves surrounding gray areas, 3-6 mm across, made up of saussurite consisting of an intimate mixture of untwinned Ab85 ± 2 and low-iron clinozoisite. The oligoclase als ois found as larger, clearer grains unmixed with clinozoisite. The calculated mode for the rock is given as no. 5, table 2. Accessories in the rock consist of less than 1 percent zircon, apatite, zoisite, and grossularite with only rare remnants of corroded, twinned calcic plagioclase and augite. Only two grains of calcic plagioclase were found clear enough for U-stage work and, because of poor

TABLE 1

9	10	11	12	13	14	15	16	17	18	19
67.6	32.0	36.4	33.2	65.1	49.0	50.0	35.8	28.9	39.6	41.9
0.23	2.9	3.1	1.34	nd	0.061	0.18	0.004	0.94	trace	0.04
11.2	16.0	16.8	16.8	17.3	0.72	1.46	0.148	15.8	5.58	2.78
—	1.07	—	0.76	—	—	—				
							5.61**	7.63**	2.42**	6.49**
4.83*	16.52	16.60*	11.14	0.039*	14.60*	12.60*				
0.65	0.16	0.19	0.23	trace	1.42	0.34	0.095	0.37	0.29	0.27
3.85	13.2	11.7	21.9	0.10	8.8	9.4	39.4	28.9	37.1	34.8
9.40	0.39	0.88	0.20	0.29	25.5	25.0	0.074	0.68	0.23	0.23
5.93	0.43	0.50	0.20	0.32	0.45	0.89	nd	nd	0.01	nd
2.70	8.80	9.05	9.10	16.10	0.24	0.20	trace	0.02	0.02	0.13
—	2.92	3.00	3.51	—	—	—	12.30	12.60	—	—
—	0.18	0.06	0.09	—	—	—	0.42	0.22	—	—
0.059	0.014	0.052	nd	nd	trace	trace	nd	0.16	nd	nd
trace	0.079	0.066	0.031	nd	trace	trace	nd	0.033	nd	nd
nd	nd	trace	nd	trace	nd	nd	0.40	nd	nd	nd
nd	0.020	nd	0.076	nd	nd	trace	0.28	nd	nd	0.086
trace	trace	nd	trace	nd	trace	trace	nd	nd	nd	trace
0.050	0.80	0.27	0.13	0.09	nd	nd	trace	0.004	0.003	trace
0.59	0.57	0.56	0.77	—	0.52	0.57	0.91	0.84	0.95	0.88
0.51	—	—	—	—	—	—	—	—	—	—
107.60	96.05	99.23	99.48	99.34	101.31	100.64	99.44	97.09	86.20	87.60

TABLE 1

Analysis No.	Description
1	Remnant olivine from serpentinized peridotite.
2	Remnant orthopyroxene from serpentinized peridotite.
3	" " " " "
4	Clinopyroxene from pyroxenite lens in serpentinite.
5	Biotite-rich acid rock (average of 3).
6	Biotite-poor acid rock with about 30 vol.% biotite and porphyroblasts of albite.
7	Biotite-poor acid rock with about 20 vol.% biotite.
8	Biotite-poor acid rock with only remnant biotite.
9	" " " " " "
10	Biotite from biotite-rich acid rock.
11	Biotite from biotite-poor acid rock.
12	Biotite from dark serpentinite in contact with biotite-rich rock.
13	Microcline from biotite-poor acid rock (average of 2).
14	Diopside from biotite-poor acid rock.
15	" " " " "
16	Average of nine serpentinite analyses.
17	Dark serpentinite from contact with biotite-rich rock. Sample was biotite-free.
18	Dark serpentinite from contact with biotite-poor rock. Sample was biotite-free.
19	Dark serpentinite from contact with biotite-poor rock. Sample was biotite-free.

orientations, the determinations only indicated the composition was more calcic than Ab40. The rare grains of augite were found contained in biotite. They showed maximum extinction angles of 43° and weak, pinkish pleochroism. They are apparently aluminous augite and are optically quite distinct from the diopside found in lenses in the upper third of the sill (noted earlier). Plate 1-A shows the typical texture of this rock type. It will be noted that the gray saussurite areas surrounded by biotite have a subhedral shape suggesting the morphology of earlier calcic plagioclase. It is in these areas that the remnants of calcic plagioclase are found.

In each biotite-rich mass is found a network of thin veins, usually less than 5 mm wide, principally made up of iron-free zoisite ($\alpha = 1.697$). These veins pinch and swell and are discontinuous. Some portions swell to 1-2 cm and possess lenticular cavities that are filled with anhedral masses of pale pink to colorless anhydrous grossularite grains. Some lenticles are open cavities, the walls of which are lined with large (5-10 mm) dodecahedral crystals of grossularite. Phehnite and margarite are accessories in these veins.

The contacts of the biotite-rich masses with the surrounding serpentinite are always quite sharp. However, biotite flakes are always present in the serpentinite for 3 to 10 inches away from the contact. In addition the serpentinite in this zone is black and dense in contrast to the normal green serpentinite of the sill. Analysis 17, table 1, was made of (biotite-free) serpentinite from the contact. In thin sections it is seen to consist of biotite, stilpnomelane, and some penninite in a matrix of antigorite. The stilpnomelane replaces antigorite and is, in turn, partially to completely replaced by biotite. Composition of biotite from the contact serpentinite is given as analysis 12 (table 1).

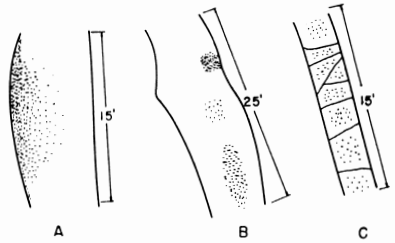


Fig. 2. Sketches of three types of acid rocks classed as intermediate and biotite-poor. A is a composite rock with a high biotite portion grading into a low biotite to biotite-free portion. B has disseminated biotite and "ghost" patches high in biotite. C has only disseminated biotite flakes.

INTERMEDIATE AND BIOTITE-POOR ACID ROCKS

Although the typical biotite-rich rocks consist of between 45 and 50 vol.% biotite, a large number of acid rock lenses in the Jeffrey serpentinite carry from 30 vol.% biotite to practically none at all. These types are illustrated diagrammatically in figure 2. Figure 2A is a composite type with a high biotite portion; 2B carries "ghost"-like high biotite patches; 2C carries only disseminated flakes.

Those which have the larger amounts of biotite appear very much like the biotite-rich bodies except for the presence of white patches, up to 15 mm across, homogeneously scattered throughout the mass. In thin sections the rock

is mineralogically and texturally similar to biotite-rich rocks; however, the white patches consist of porphyroblastic aggregate clusters of clear, twinned, anhedral albite (Ab 91±2). Plate 1-B shows one of these clusters. In addition, close examination of biotite shows that some grains are partially replaced by minute blebs and grains of microcline and diopside. Analysis 6 (table 1) was made of this rock type.

The majority of biotite-poor masses carry less than 10 vol.% biotite as scattered flakes or as patches that appear to be "ghosts" of biotite-rich rocks (fig. 2B). In thin section these rocks are found to consist of microcline (analysis 13, table 1), Ab 95±2, diopside (analyses 14, 15, table 1), and quartz, with accessory sillimanite and muscovite in rare cases. These rocks are usually highly fractured. Biotite flakes are found as accessories only in unfractured portions. Any thin fractures that cut through a biotite patch are bordered by a biotite-free zone of several inches' width on either side (fig. 2C).

These rocks show a tendency to segregate some phases. Pale green diopside is usually mixed with microcline, albite, and quartz but may occur as thin (2 mm) veins. Microcline and/or myrmekitic albite-quartz often occur with quartz and accessory sillimanite as coarse pegmatite patches.

The diopside and microcline are the products of the breakdown of biotite. Plate 1-C shows diopside and microcline replacing biotite as thin slivers or blebs parallel to the mica cleavage. Plate 1-D shows the typical texture of diopside, microcline, and albite in a portion of rock that carries no biotite accessory. In this figure the central diopside grain encloses a small crystal of microcline. The diopside is, in turn, partially enclosed in a larger microcline crystal. This texture is interpreted to be the result of simultaneous crystallization.

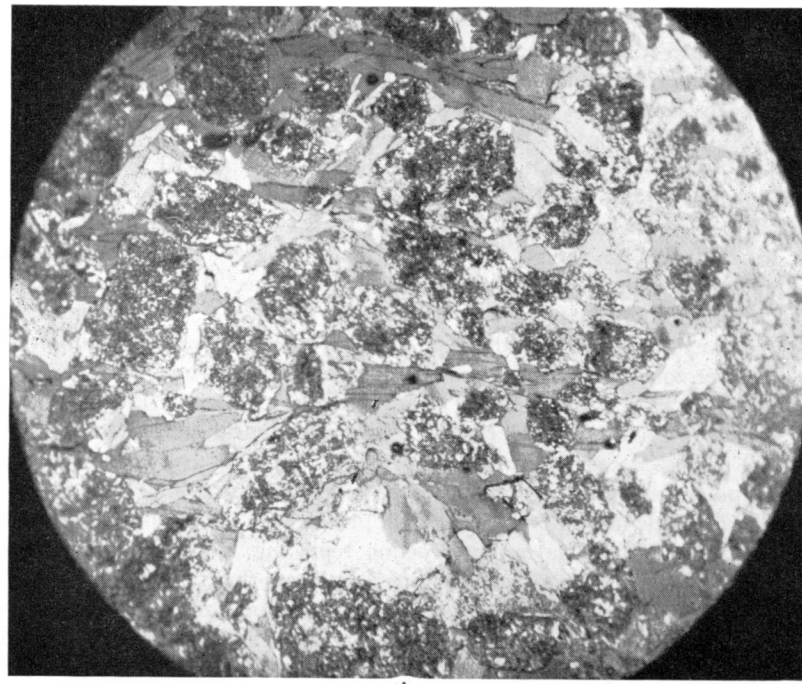
In contrast to the even contacts of the biotite-rich masses against serpentinite, the biotite-poor contacts are ragged and uneven. Feldspar grains with diopside are found contained in the black, dense serpentinite several inches away from the acid rock, or feldspar grains are found jutting out into the contact serpentinite. In addition, serpentinite is found squeezed for several inches into fractures in the acid rock. This ragged contact is interpreted largely as the result of the conversion of biotite (which occurs in the contact zone of serpentinite) to microcline and diopside. In thin sections the contact serpentinite zone for 6-8 inches away from the acid rock is found to consist of biotite flakes, in all stages of replacement by microcline-diopside, and some penninite in a matrix of antigorite. The antigorite is often partially replaced by shreds of talc. Rare biotite flakes show conversion to talc and a dust of magnetite grains. Analyses 18 and 19 (table 1) were made of dark serpentinite from contacts with biotite-poor acid rocks.

Finally it should be noted that in rare instances veins are found in the biotite-poor rocks that consist of fibrous parawollastonite. In a few instances vugs were found that were lined with albite and brushlike fibers of pink pectolite.

STRUCTURAL FEATURES

Structurally the Jeffrey serpentinite does not possess any consistent mappable elements such as numerous persistent compositional bands or primary S-planes. The lower contact with metasediments strikes approximately N60E and dips 55° SE, while the upper contact strikes about N45E and dips about

PLATE 1



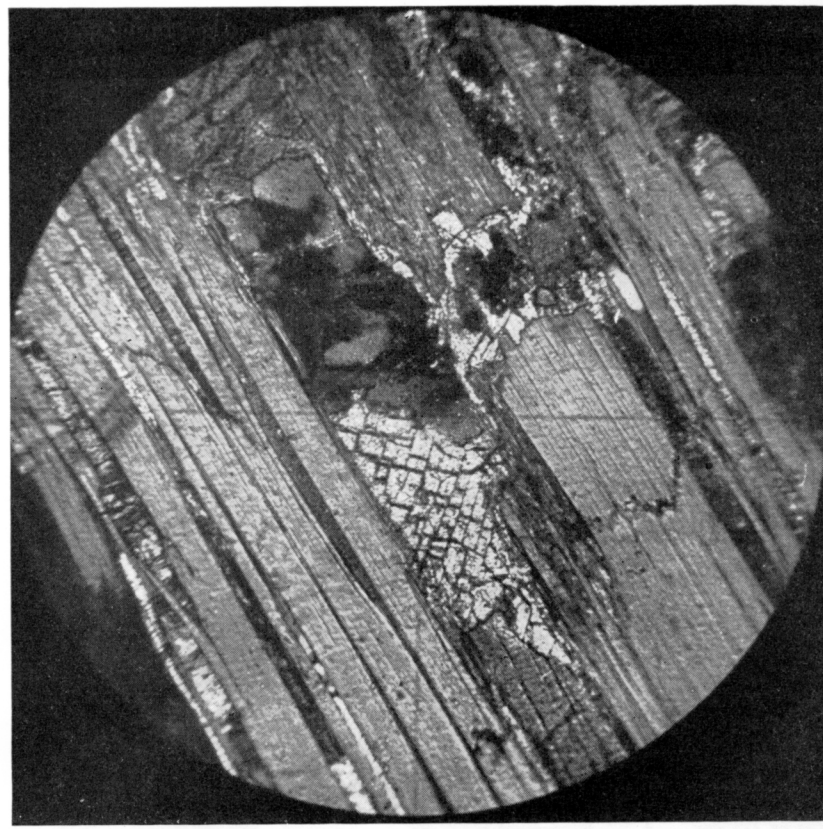
A.

B.

70° SE. The actual map-view distribution of acid rock lenses fortuitously depends on the mining operation and changes from month to month. However, by taking advantage of the lensoid habit of these bodies, it was possible to measure an average dip and strike for each body exposed at a given time. Care was taken not to measure any one body more than once.

The poles to each lens measured were plotted on the *upper hemisphere* of a Schmidt net. To avoid working at the edges of the net it was found to be

PLATE 1



C.

A. Biotite-rich acid rock. Biotite sheaves and oligoclase grains surround areas of saussurite (oligoclase and clinozoisite) that retain the subhedral shapes of original calcic plagioclase crystals. 50x. X Nicols.

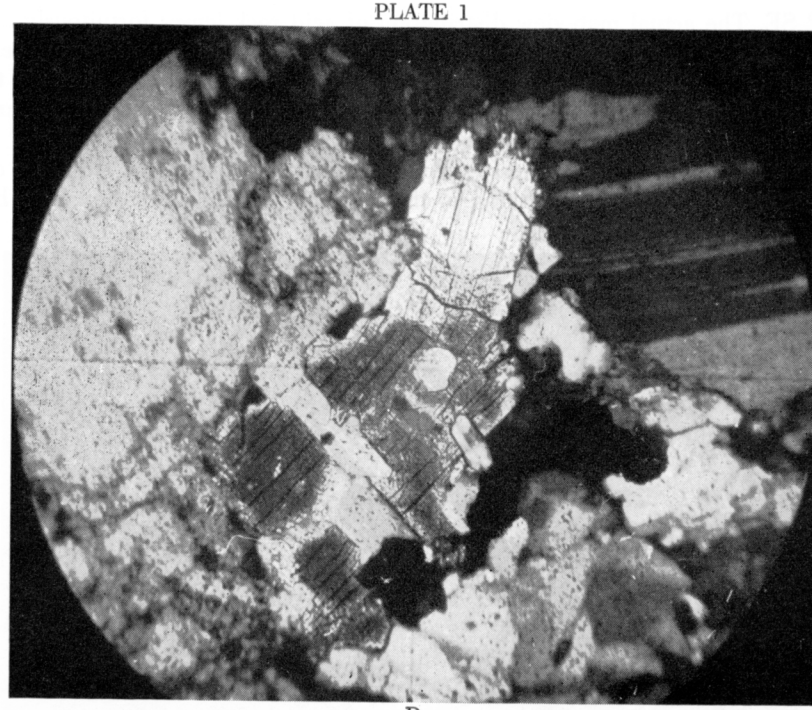
B. Porphyroblastic aggregate cluster of albite crystals in a biotite-rich matrix. The cluster here is about 10 mm. in diameter. The rock is classed as an intermediate type of acid rock. 25x. X Nicols.

C. Diopside (cross-hatched) and microcline (dark gray) replacing biotite. Small blebs of microcline and diopside are developed along other cleavages. 25x. X Nicols.

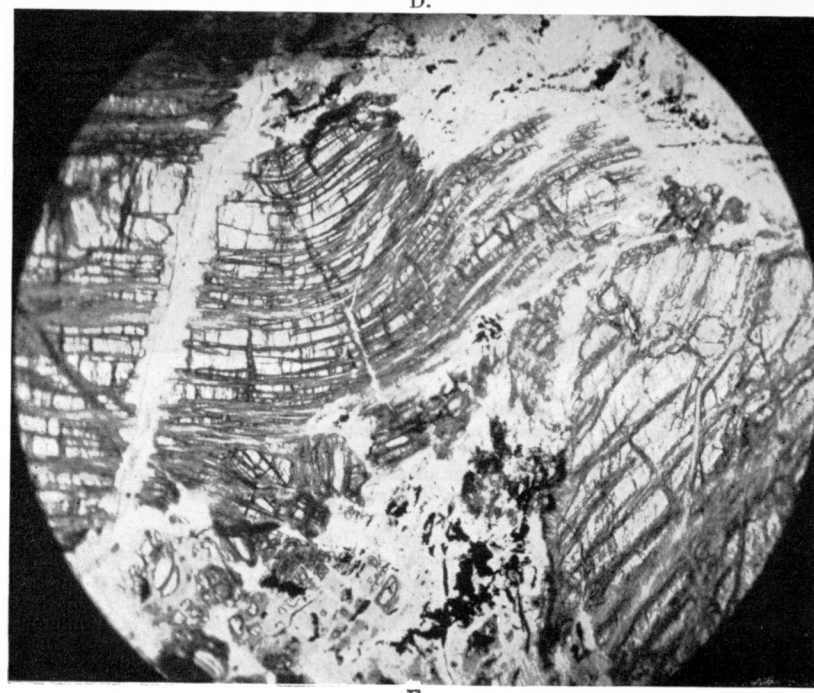
D. Diopside (in center) encloses a small crystal of microcline. Larger microcline crystal partially encloses this diopside grain and another grain of diopside. This texture is interpreted as the result of simultaneous crystallization of diopside and microcline. Twinned albite crystal in upper part of field. 25x. X Nicols.

E. Cataclastic texture of remnant enstatite in serpentinite. 25x. X Nicols.

PLATE 1



D.



E.

much clearer if the sill were "rotated" to a "horizontal" position by the number of degrees of dip of the closest contact. Figures 3A-3E are the plots of all the acid rocks exposed in May 1957, with the sill "rotated" 55° in a direction N30W.

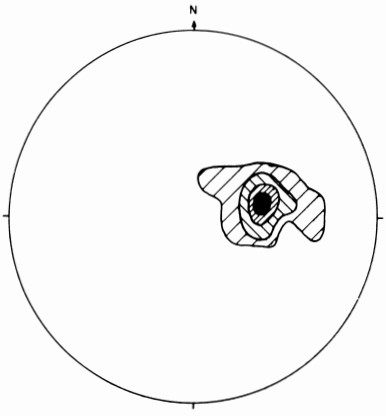
These plots show a strong anisotropy. The twenty-six biotite-rich bodies plot at a strong single peak (fig. 3A), whereas forty-nine biotite-poor bodies plot at two major peaks with a weak partial girdle between them (fig. 3B). When all seventy-five points are plotted together, one weak and one very strong peak result due to the coincidence of the biotite-rich maximum with one of the biotite-poor maxima (fig. 3C). In figure 3D are plotted the poles to thirty-three shear planes in the serpentinite. These show a major peak close to the same maximum of figure 3C.

Figure 3E is a composite drawing of the peak points only, with the planes normal to these peaks drawn in. A and B represent the two acid rock maxima. X and Y are two shear plane maxima. Point Z represents the maximum point of nineteen lineation element measurements. These lie in the plane of shear plane maximum X.

From these data it will be noted that there is a relationship between mineralogy and structural attitude as well as a remarkable symmetry. The acid rocks strike at almost right angles to the sill, and their dips lie symmetrically about 66° apart and about 33° to the pole to the sill (center point of net). Only one of these maxima (A) is approximately coincident with a major shear direction. In the field it was observed that biotite-rich bodies and some biotite-poor bodies were subparallel to shears which sometimes warped around them and sometimes cut them at low angles. However, many biotite-poor bodies were cross-cut by shears which offset the blocks of acid rock. Serpentinite is squeezed between the blocks thus separated. Within such acid bodies numerous fractures are developed parallel to the shear planes. As it was noted earlier, these fractures are bordered by biotite-free zones on either side.

INTERPRETATION

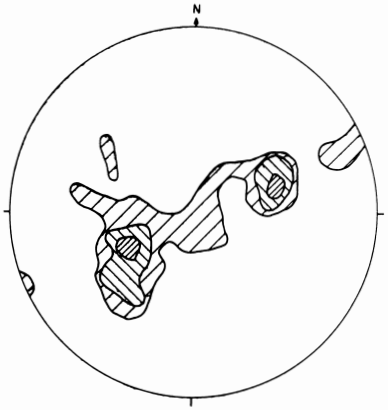
Although small acid rock masses are associated with highly serpentinized alpine-type peridotites, similar acid rocks are only rarely associated with unaltered peridotites (MacKenzie, 1960). On the other hand, the association of small olivine-augite gabbro or troctolite masses with unaltered alpine peridotites is common. Guild (1947) described olivine-augite-enstatite-calcic plagioclase lenses in the peridotite of Oriente Province, Cuba. Hadley (1949) in his work on the peridotite of Clay County, North Carolina, found that the body was partially altered to serpentinite and talc schist, and troctolite lenses in it were also partially altered. Olivine is rimmed by enstatite, and bytownite is rimmed by hornblende. These rims are apparently due to deuteric reactions and require the addition of at least water and silica. More highly altered lenses are converted to edenite amphibolites which are associated with veins of sodic plagioclase, corundum, biotite, actinolite, chlorite, and talc. All these minerals cannot be the result of the isochemical alteration of troctolite. They require the metasomatic addition of at least soda, potash, silica, and water.



BIOTITE-RICH ACID ROCKS

Sill horizontal
Contour: 36% 28% 20% 12%
Plot of 26 points

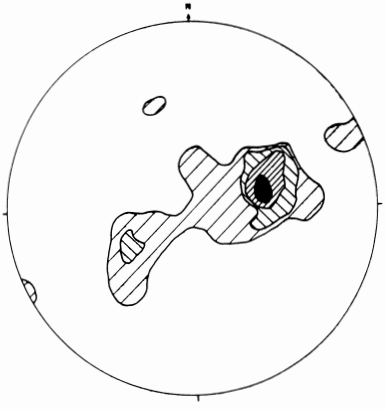
A



BIOTITE-POOR ACID ROCKS

Sill horizontal
Contours: 14% 10% 6%
Plot of 49 points

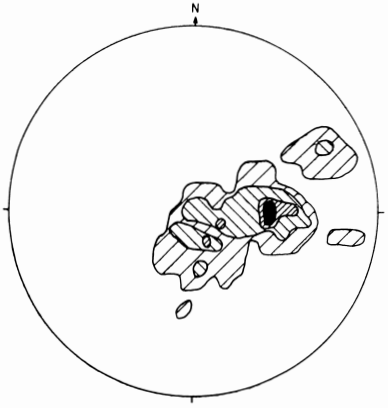
B



ALL ACID ROCKS

Sill horizontal
Contours: 18% 13% 9% 5%
Plot of 75 points

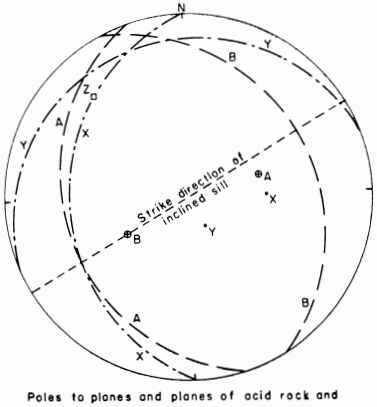
C



SHEAR PLANES

Sill horizontal
Contours: 24% 18% 12% 6%
Plot of 33 points

D



Poles to planes and planes of acid rock and shear plane maxima.
Sill horizontal.

E

Fig. 3.

In the case of the acid rocks in the Jeffrey sill the remnant minerals, calcic plagioclase and augite, found in the biotite-rich masses strongly suggest that these bodies are metasomatically altered gabbros. The absence of olivine or enstatite remnants does not necessarily mean that they were never present. The original bodies may have been olivine-augite gabbros or simply augite gabbros. The attempt to determine the calcic plagioclase composition optically was unsatisfactory. However, if it is assumed that the present clinozoisite-Ab85 content (no. 5, table 2) is totally the result of the saussuritization of the original calcic plagioclase, the implied composition of the original plagioclase is about Ab30, calcic labradorite. This composition is consistent with the plagioclase compositional range observed in unaltered gabbroic rocks in unaltered peridotites (Guild, 1947). The metasomatic conversion of an original gabbro, of specifically unknown composition, to the biotite-saussurite bodies, now observed, requires, as a conservative estimate, the addition of at least potash and water. It may also have required added silica and alumina. If gabbroic material were removed in great quantity there is no evidence of it in the field. However, it is probable that the accessory veins of zoisite (and grossularite) represent removal and segregation of lime, alumina, and silica from

TABLE 2

Calculated modes from rock analyses

	Mole %	Wt.%	Vol.%
	5		
Biotite	42.6	47.6	47.1
Ab 85	26.6	17.6	20.0
Clinozoisite	30.8	34.8	32.9
	8		
Microcline	14.3	15.9	17.7
Ab 95	47.0	49.5	53.0
Diopside	37.0	34.2	28.6
Quartz	1.7	0.4	0.4
	9		
Microcline	12.5	14.8	16.5
Ab 95	43.6	48.9	52.7
Diopside	34.4	33.9	28.2
Quartz	9.5	2.4	2.4

Fig. 3. A-C represent the plot of poles to lensoid acid rock bodies in the Jeffrey serpentinite. D is the plot of poles to shear planes in the serpentinite. E is a composite plot of maxima points from A-D; the planes normal to each maximum are drawn in. Point Z in E is the maximum of lineation elements found in shear plane X.

All plots are on the upper hemisphere of a Schmidt net. For clarity the plots have been moved 55° in a direction N30W to place the sill in a horizontal attitude and avoid plotting at the edge of the net.

the original gabbro. Analysis 17 (table 1) of the surrounding contact serpentinite shows an enrichment of titanium, iron, manganese, calcium, potassium, zirconium, vanadium, and barium when contrasted with the average of nine normal serpentinite analyses from the Jeffrey pit (analysis 16, table 1). The titanium, iron, manganese, and calcium may have been extracted from the original gabbro. However, the other elements must represent addition from an outside source. Cooke (1937, p. 83) presents a similar analysis of dark serpentinite in contact with an acid rock from Thetford Mines. He contrasts it with an analysis of normal serpentinite, and the same compositional differences are noted as at Jeffrey.

BaO which geochemically follows K_2O is enriched in the biotite of the acid rock (analysis 11), in the contact serpentinite zone, and in the biotite found in the contact serpentinite (analysis 12). This latter biotite has a Mg/Mg+Fe ratio of 0.77 in contrast to the biotite from the adjacent biotite-rich rock where the ratio is 0.57 (analysis 10). This indicates that the two biotites grew in adjustment with two different environments. The Mg to Fe ratio observed in the biotite-rich rock (0.53, analysis 5) indicates the more iron-rich composition of the original gabbro in contrast to the surrounding, highly magnesian peridotite (where the present ratio in serpentinite is 0.91, analysis 16). This difference is to be expected if the gabbros represent a late crystallizing fraction of the original ultramafic mass. Incidentally, it should be noted here that the ratio difference between the biotite of the biotite-rich rock and the whole rock itself is due to a small amount of ferric iron in the clinozoisite of the rock. An approximate calculation indicates less than 5 percent epidote in the clinozoisite.

From the petrographic work described earlier, it was possible to follow a sequence of changes in these acid bodies whereby more sodic plagioclase appeared, clinozoisite disappeared, biotite showed decomposition to microcline and diopside, and silica appeared in excess as quartz. In addition some fairly rare specimens showed a small excess of alumina as sillimanite. Analyses 5 through 9 (table 1) represent a sequence of acid bodies with decreasing visible biotite content. From optical estimates, analyses of rocks, and from analyses of microcline, biotite, and diopside it was possible to calculate approximate modes for rocks 5, 8, and 9 (table 2).

On the basis of the petrographic work it is concluded that the intermediate and biotite-poor bodies represent further alteration of the biotite-rich rock type. Figures 4 and 5 are simple variation diagrams on the basis of increasing SiO_2 going from rock 5 to rock 9 (left to right). The ratio of Mg/Mg+Fe in figure 3 varies from 0.53 to 0.59, rising to the latter value in almost biotite-free rocks 8 and 9. The variation in this ratio indicates a small enrichment in Mg or the removal of some Fe under the conditions of final alteration. In figure 4, Al_2O_3, FeO and TiO_2 drop through the sequence while Na_2O rises. K_2O shows only a small variation in higher biotite rocks 5, 6, and 7, and then drops in the low biotite rocks 8 and 9. MgO drops slightly through 5-6-7 and then rises slightly in 8 and 9. CaO drops steadily in the higher biotite rocks and rises sharply in low biotite rocks. Through the sequence of rocks which carry significant amounts of biotite (5-7), the CaO content is independent of

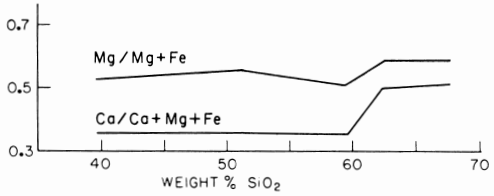


Fig. 4. Variation of ratios Mg/Mg + Fe and Ca/Ca + Mg + Fe against $\%SiO_2$. The variation is taken from acid rocks showing decrease in visible biotite content (going from left to right, analyses 5-9, table 1.).

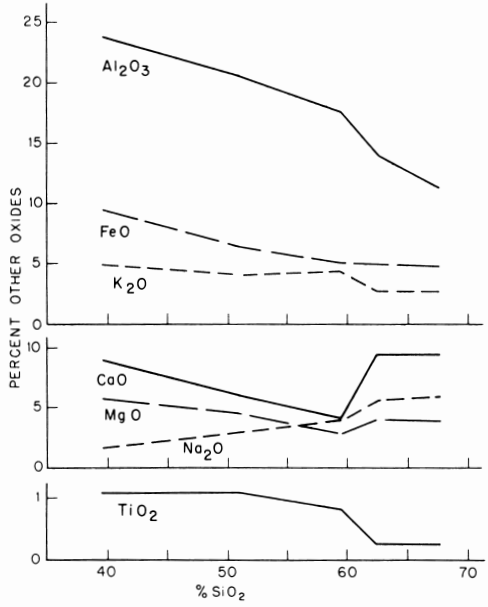


Fig. 5. Oxide variation plotted against $\%SiO_2$ for acid rocks with decreasing visible biotite content (going from left to right, analyses 5-9, table 1.).

Mg. However, there is a slight relationship between calcium and iron in the small epidote content of the clinozoisite of these rocks. This is small enough that CaO may be considered relatively independent of both Mg and Fe in the rocks in which biotite is stable. However, in the low-biotite to biotite-free rocks, 8 and 9, Ca is largely dependent on Mg and Fe in the diopside which replaces biotite as the major ferromagnesian mineral. The variation of the ratio Ca/Ca+Mg+Fe exhibits a more regular trend through the sequence of rock types than does the simple variation of CaO. In figure 4 this ratio shows little variation through the sequence of higher-biotite rocks 5, 6, and 7, and then jumps to a steady higher value in biotite-poor rocks 8 and 9. The jump apparently reflects the dependence of Ca on Mg and Fe and also suggests an introduction of Ca into these rocks thereby making biotite unstable (discussed below).

The most conservative interpretation that can be made of the data is that the alteration of the biotite-saussurite assemblage to the final diopside-micro-

cline-albite-quartz assemblage is the result of major additions of silica, lime, and soda, and obviously the loss of water. The dropping percentages of other oxides are due in part to relative dilution. Iron may have been removed in small amounts and small amounts of magnesium may have been added. Evidence is given below which indicates that potash remained fairly constant and that its drop-off in the variation diagram is due almost entirely to the dilution effect of added material.

The breakdown of biotite to microcline and diopside is of particular interest in that diopside is not the usual ferromagnesian mineral found in serpentinite-acid rock associations. The only other place in the region where diopside is significant is in the acid bodies associated with serpentinite at the Montreal Chrome Pit about seven miles southeast of Thetford Mines (Cooke, 1937, p. 79). At Thetford Mines itself, actinolite is the common mafic mineral (Cooke, 1937, p. 81), and in other occurrences around the world hornblende is often found in this type of rock. The absence of hornblende at Jeffrey is attributed to low available alumina. On the other hand, actinolite is not stable in the higher temperature amphibolite facies which the biotite-poor acid rocks represent (discussed below). This is apparently also the reason why diopside pyroxenites in the upper third of the sill have not been altered to actinolite amphibolites.

The conversion of biotite to diopside-microcline is a critical reaction in going from the epidote-amphibolite to amphibolite facies (Ramberg, 1952). The reaction requires both available lime and silica. Biotite (phologopite) is often stabilized in amphibolite facies marbles because of low available silica in a highly calcic environment (Ramberg, 1952, p. 152). An idealized reaction would be:

$$11.8\ SiO_2 + 5.9\ CaO + K_2Mg_{3.36}Fe_{2.64}Al_2Si_6O_{20}(OH)_4 \rightleftarrows$$
$$2\ KAlSi_3O_8 + 5.9\ CaMg_{0.57}Fe_{0.43}Si_2O_6 + 2\ H_2O + 0.1\ FeO$$

In this reaction a simplified biotite with a Mg to Fe ratio of 0.56 is converted to a diopside with a ratio of 0.57. These ratios correspond to the ratios determined in unreacted biotite (analysis 11, table 1) which was found coexisting with diopside (analysis 15) in a biotite-poor acid rock. A small release of iron occurs on the right-hand side. Ramberg (1952, p. 152) pointed out that in amphibolite facies carbonate rocks the Mg to Fe ratio is usually somewhat higher in diopside than in coexisting biotite. It is interesting to note that the idealized reaction predicts a diopside to microcline molar ratio of 2.95:1 in the product. Calculated modes for rocks 8 and 9 (table 2) give molar ratios of 2.6 and 2.8:1, respectively. This suggests that the K_2O content of the rock remained fairly constant during the alteration, and the drop observed in the variation diagram is due largely to relative dilution.

The association of diopside with sodic plagioclase led Tilley (1927) to assume that diopside was stable in the low epidote-amphibolite facies. Ramberg (1952) pointed out that the plagioclase composition is a response to bulk composition and that diopside is always indicative of the amphibolite facies. The presence of accessory sillimanite in some of the biotite-poor acid rocks further confirms amphibolite facies conditions.

The structural data presented earlier suggest that the original gabbro masses lay in two major orientations 66° apart. One group may have been in an attitude which corresponded with an original strong joint system whereas the other set lay in a complementary set of weaker joints. In this regard Guild (1947) noted the presence of two dominant joint sets in the peridotite of Oriente Province, Cuba. The poles to these joints plotted at two strong maxima about 80° apart. A plot of the poles to lenticular olivine gabbro bodies showed two maxima, about 25° apart, one of them corresponding to one of the joint maxima. The similarity of the two cases is striking.

During subsequent deformation, possibly accompanying serpentinization, the strongest set of joints would most likely be utilized as a shear direction, and the shear planes would act as zones of entry or localization of introduced material. Under the initial P-T conditions (possibly high epidote-amphibolite facies) the gabbros were altered to the biotite-saussurite assemblage by reaction with introduced potash and water. Increasing metamorphism moved into the amphibolite facies field and the introduced material became largely soda, lime, and silica rich. The acid bodies which lay transverse to the major shear direction were most thoroughly altered by the new material. They were cross-faulted and cross-jointed, and biotite persisted as unreacted remnants only in portions that were unfractured.

CONCLUSIONS

Ramberg (1952, p. 156) places the upper limit of the amphibolite facies at about 400°C. A pure magnesian serpentinite is stable to 500°C (Bowen and Tuttle, 1949). Thus, serpentinites are not diagnostic of the particular facies of metamorphism under which they formed. The mineralogy of the hydrothermally altered gabbros of the Jeffrey sill indicates amphibolite facies conditions were attained at some time in the history of the whole mass. The very weak contact metamorphic effect along the footwall indicates hornfels development also in the amphibolite facies. On the other hand, the surrounding country rocks represent adjustment only to epidote-amphibolite facies conditions. Thus, the serpentinite and its associated acid rocks are regionally out of grade with the country rocks. This disparity may be interpreted in several ways.

(1) The peridotite sill was intruded as a relatively cool mass which resulted in only a very weak hornfels development at the contact. During subsequent deformation, the sill, lying in a plane of weakness, acted as a locus for deformation and an avenue for the introduction of added components and water.

(2) The peridotite sill was first intruded at greater depth than is represented by the country rocks which now surround it. It was serpentinized (and the gabbros altered) at a depth commensurate with an amphibolite facies level of metamorphism. The plastic serpentinite, with its acid bodies, was subsequently moved as a solid into the higher level (and lower grade) country rocks in which it is now found. The heat loss was lower than the rate of movement so that the serpentinite was slightly hotter than these rocks and a small contact hornfels zone developed.

MacKenzie (1960) found a strong amphibolite facies hornfels development associated with the intrusion of the virtually unserpentinized Tinaquillo peridotite in Venezuela. Here again the country rocks are in the epidote-amphibolite facies. On the basis of this single case one is inclined to favor the second alternative suggested above.

On a broader regional basis, if one assumes that the serpentinites of the Eastern Quebec belt are more or less contemporaneous within short distances, then the Jeffrey body must have a history similar to the well-studied body at Thetford Mines (Cooke, 1937). At Thetford Mines there is no contact hornfels development, and the acid rocks carry actinolite, tremolite, and epidote which represent the epidote-amphibolite facies. The country rocks are also in the same facies. However, at the Montreal Chrome Pit, just seven miles southeast of Thetford Mines, diopside occurs in acid bodies and the level of metamorphism is higher. A detailed study of the region as a whole was not made by the writer but one significant observation can be made. According to Cooke's maps (1937), the serpentinite at Thetford Mines and the belt to the southwest of it lie on the crest of a regional anticlinal structure. But, at the Montreal Chrome Pit and at the Jeffrey Mine the serpentinites lie on the flanks of the same fold. It is possible that the flanking position suffered more intense deformation and development of higher temperatures than the crestal position, and thus was created a sharp P-T gradient across the structure which resulted in facies changes in relatively short distances. This would also account for the fact that the supposed original attitudes of the gabbro bodies at Jeffrey were only partially destroyed by subsequent deformation. That is, a weak partial girdle was developed during deformation by the rotation of some bodies, but the strong original maxima were not totally destroyed as they would have been had the whole serpentinite mass been remobilized and moved to shallower depths as is implied by the second alternative presented above.

Little can be said concerning the source of the components which effected the alteration of the gabbroic rocks in the Jeffrey sill. For that matter the alteration cannot be dated definitely as pre- or post-serpentinization, or synchronous with it. However, by comparing analyses 1, 2, and 3 of primary olivine and enstatite with analysis 16 (the average of 9 serpentinites), it will be noted that the small amounts of potash, soda, and lime found in the primary minerals are virtually absent in the serpentinite. For example, an approximate calculation shows that if the original peridotite consisted of 90 vol.% olivine (analysis 1) and 10 vol.% enstatite (analyses 2 and 3 with an average of 2% CaO) and was thoroughly converted to serpentine and magnetite, the serpentinite rock should carry about 0.3% CaO in the analysis. Analysis 16, of serpentinite, shows only 0.07% CaO, or less than one-fourth of the expected amount. This discrepancy is larger than would result from an analytical error of as much as 15%. On this basis it would appear that the alteration of gabbros accompanied serpentinization and resulted, at least in part, from oxides released during the process.

CHEMICAL ANALYSES

All chemical analyses were made by the writer at the Spectrochemical Laboratory of the Department of Geology, University of Chicago. Metals were determined spectrographically. Alkalies were determined (flame) photometrically against N.B.S. standards. Ferrous iron was determined by titration against standardized cerric solution. Consituent water was determined by the Penfield method.

In table 1 a dash (—) indicates that the oxide was not sought. The letters "nd" indicate the oxide was not detected within detection limits (usually 10 ppm ±). The observed analytical error is 2% for alkalies and 1% for ferrous iron. For spectrographically determined elements the expected error is 10% at least. In the work performed here, unknowns were arced alternately with knowns (such as G-1, W-1) which were diluted to bracket the percentage of the oxide sought in the unknown. Thus a working curve was re-established on each plate in the percentage range sought. Replicate arcings of knowns yielded observed variations within 7.5%. Thus, 10% may be safely used as the maximum error.

REFERENCES

Bowen, N. L., 1956, The evolution of the igneous rocks: New York, Dover Publications, 334 p.

Bowen, N. L., and Tuttle, O. F., 1949, The system MgO–SiO_2–H_2O: Geol. Soc. America Bull., v. 60, p. 439-460.

Cooke, H. C., 1937, Thetford, Disraeli, and eastern half of Warwick map areas, Quebec: Canada Geol. Survey Mem. 211, 160 p.

Du Toit, A. L., 1919, Plumasite and titaniferous magneite rocks from Natal: Geol. Soc. South Africa Trans., v. 21, p. 53-73.

Gordon, S. G., 1921, Desilicated granite pegmatites: Acad. Nat. Sci. Philadelphia Proc., pt. 1, v. 73, p. 169-192.

Guild, P. W., 1947, Petrology and structure of the Moa chromite district, Oriente Province, Cuba: Am. Geophys. Union Trans., v. 28, p. 218-246.

Hadley, J. B., 1949, Preliminary report on corundum deposits in the Buck Creek peridotite, Clay County, N. C.: U. S. Geol. Survey Bull. 948, p. 103-128.

Larsen, E. S., 1928, A hydrothermal origin of corundum and albitite bodies: Econ. Geology, v. 23, p. 398-433.

MacKenzie, D. B., 1960, High temperature alpine-type peridotite from Venezuela: Geol. Soc. America Bull., v. 71 p. 303-318.

Nickel, E. H., 1958, The occurrence of native nickel-iron in the serpentine rock of the eastern townships of Quebec: Special paper prepared for presentation to Mineralogical Soc. America Meeting, St. Louis, Mo., Canada Dept. Mines and Tech. Survey.

Ramberg, Hans, 1952, The origin of metamorphic and metasomatic rocks: Chicago, University of Chicago Press, 317 p.

Riordon, P. H., 1957, Evidence of pre-Taconic orogeny in southeastern Quebec: Geol. Soc. America Bull., v. 68, p. 389-394.

Tilley, C. E., 1927, Vesuvianite and grossularite as products of regional metamorphism: Geol. Mag. [Great Britain], v. 64, p. 372-376.

Turner, F. J., and Verhoogen, Jean, 1960, Igneous and metamorphic petrology: New York, McGraw-Hill Book Company, 694 p.