

ALPINE-TYPE CHROMITE IN NORTH BORNEO, WITH SPECIAL REFERENCE TO DARVEL BAY¹

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ABSTRACT

Chromite occurs as layers and pods in dunite and serpentinite lenses within most of the peridotite outcrops of Sabah. The crystals are anhedral, rounded, cataclastic, and unaltered.

In Darvel Bay, only the Saddle group of islands and Mount Silam contain chromite ore, yet all the ultramafic bodies have high 2770 ppm Cr and 1530 ppm Ni averages. The residual character of the ultramafites is indicated by 74 ppm K and 412 ppm Ti averages. The ultramafites occur within the Silumpat Gneiss, considered to be metamorphosed tholeiitic basalt, characterised by 230 ppm Cr, 96 ppm Ni, 653 ppm K, and 5778 ppm Ti averages.

The Sabah chromites have a wide range of chromium content, and aluminium increases reciprocally with chromium decrease. An X-ray fluorescence powder dilution method is given for chromite analysis.

Unit cell edge (a , Å) varies approximately with composition thus: Al cations/unit cell = $407.3 - 48.6 \times a$; Cr cations/unit cell = $58.46 \times a - 473.88$. Reflectivity at 590 nm varies linearly with composition thus: Al cations/unit cell = $27.93 - 1.82 \times \text{reflectivity} (\%)$. Hardness ranges from 1200 to 1500 VH.

The chromite, together with the ultramafite envelope, is presumed to have originated in the mantle and been emplaced as tabular masses into the crust.

INTRODUCTION

The main chromite occurrences in Sabah (North Borneo) are in the northern islands of Banggi and Malawali, the Labuk Valley of central Sabah, and the Darvel Bay area of the south-east (Fig. 1).

Approximately 65 percent of the ultramafic outcrops were intensively prospected from 1959 to 1963 by the Borneo Mining Co. Ltd. (Naylor Benzon and Co. of London) and found to contain insufficient volume of chromite for economic exploitation. In all the investigated areas the chromite occurs in dunite and serpentinitized dunite bodies and lenses (forming about 5% of the ultramafite total) which occur always within larger peridotite outcrops.

In the Darvel Bay area, the ultramafic bodies were emplaced into a basement of metamorphosed oceanic tholeiitic basalt (The Silumpat Gneiss) and are structurally concordant to their host-rocks (Fig. 2). Hutchison and Dhonau (1969) have shown that the ultramafic body of the Saddle island group is of a younger age and of a higher level of

¹ Presented at the Twelfth Pacific Science Congress, Canberra, Australia: August 24th, 1971. [Hutchison, 1971, (abstract)].

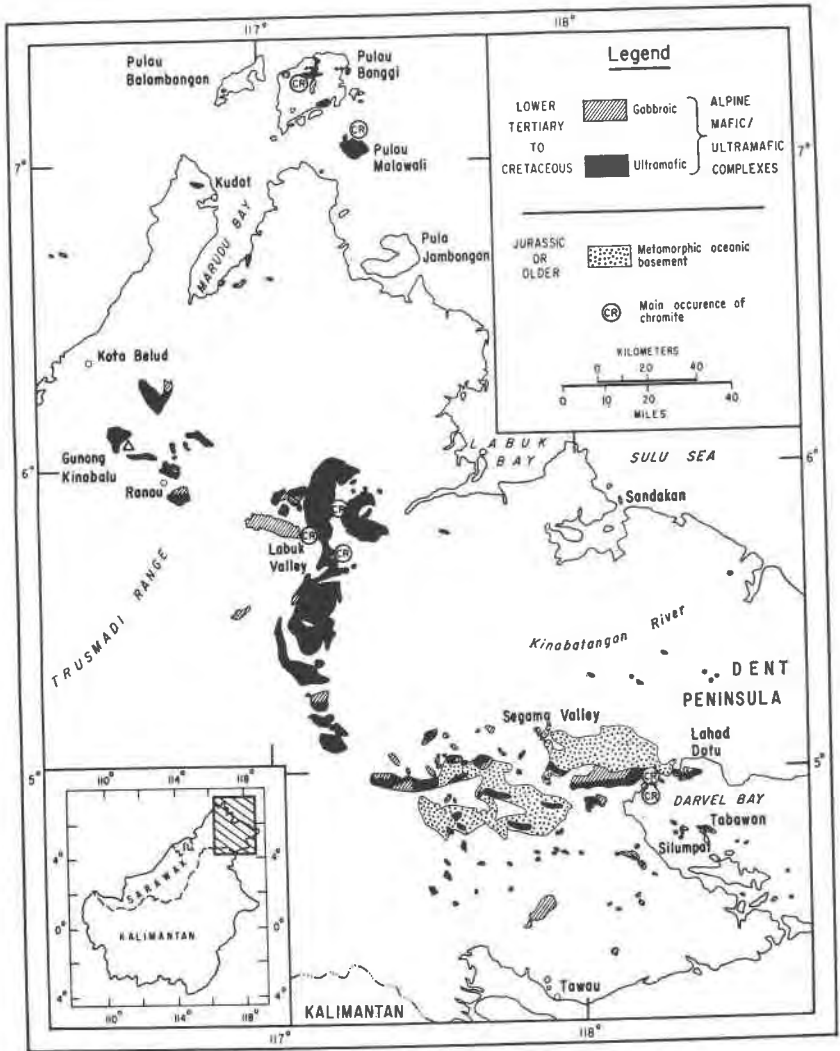


FIG. 1. Outline geologic map of Sabah showing the main occurrence of chromite in relation to the crystalline basement and the alpine ultramafic/mafic complexes (modified after Kirk, 1968). Sedimentary formations are left unshaded.

emplacement than the other ultramafites of the region (Fig. 2). They have also shown (Hutchison and Dhonau, 1969 and 1971) that the peridotite of Tabawan island was of a higher emplacement temperature and has caused a thermal aureole in the Silumpat Gneiss from the regional greenschist, through almandine amphibolite, to hornblende

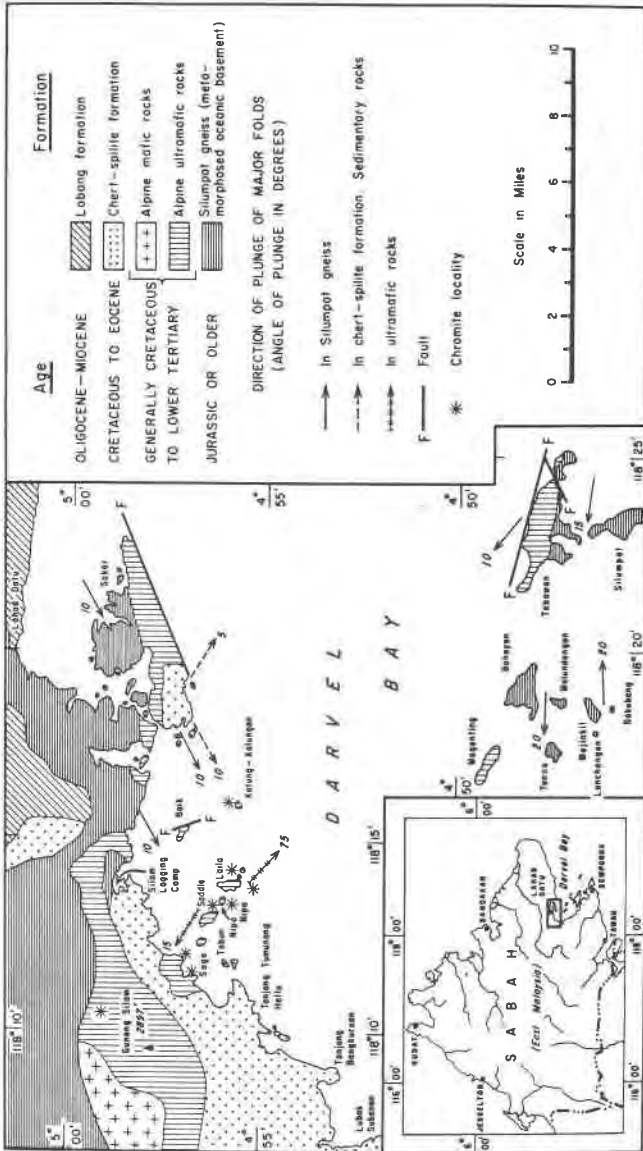


Fig. 2. Geologic map of the Darvel Bay area of Sabah after Hutchison and Dhonau (1969 and 1971).

granulite facies. By contrast, the younger Saddle island group ultramafite was emplaced in equilibrium with the greenschist facies conditions, and it is this emplacement alone in the Darvel Bay area that contains chromite ore.

TEXTURE

The chromite occurs as layers and bands, ranging in thickness from a few millimeters to several meters, and as lenses, pods, and layer-zones of disseminated crystals. The layers have been folded and boudinaged, but in several instances original cumulate textures are preserved. On Katung-Kalungan island (Fig. 2), for example, the chromite bands are sharply folded, but even after folding they preserve excellent graded layering reminiscent of an originally stratiform origin (Fig. 3). In the Labuk Valley, 800 meters east of Katai on Bukit Lumisir, the distribution of narrow bands of fine grained chromite in several dunite bodies suggests crystal settling (Collenette, 1964).

Generally, however, the distribution of chromite is tectonically

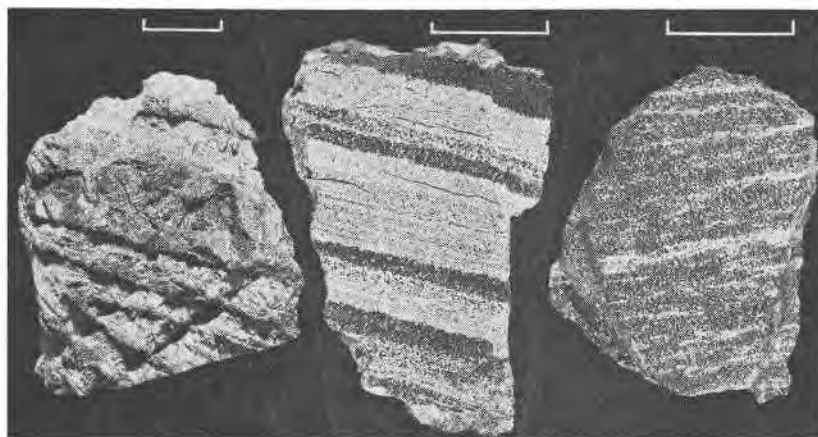


FIG. 3. Left: sharp fold in chromite-serpentinite layers. The chromite layers show high relief on the natural outcrop surface. (Specimen J 1173 from Katung-Kalungan island).

Center: End view of the same specimen on a polished surface showing graded layering of the chromite, inverted by folding.

Right: Chromite disseminations and segregations in serpentinite. The distribution probably resulted from tectonic emplacement, and the original layering has been largely obliterated. (Polished surface of specimen J 1187 from Laila island). The white bars represent 3 cm on each adjacent specimen.

controlled, and original stratiform structures largely obliterated (Fig. 3). In some areas, for example at Bidu-Bidu in the Labuk Valley area, the ultramafic foliation is not parallel to, and is superimposed on an original stratiform layering. Newton-Smith (1967) took this to indicate that the chromite had a magmatic-settling origin before tectonic emplacement.

Individual localities are listed under Table 1.

The chromite crystals from all localities in Sabah have rounded anhedral outlines (Fig. 4). All are cataclastic to some degree. The massive podiform bodies are also brecciated and healed by the serpentinite matrix (Fig. 4). Sabah chromite is remarkably free from hydrothermally altered rims as have been described by Mihálik and Saager (1968). The absence of compositional zoning is indicated by similar reflectivity values across crystals and by sharp and single X-ray diffraction peaks. Irregular patches and zones of ferritchromite as described by Engin and Aucott (1971) have been observed very rarely in Sabah. From the whole collection, only two examples of higher reflectivity ferritchromite have been found (A and B in Fig. 4).

TRACE ELEMENTS IN COUNTRY ROCKS

Method

Analyses were by comparison with U. S. Geol. Survey PCC-1, taking Ti = 44, Cr = 2600, and Ni = 2300 ppm and W-1 taking K = 5310, Ti = 6400 ppm.

Whole rock specimens, pulverised in a tungsten carbide-lined Tema mill vessel to avoid contamination, were made into pressed pellets backed with boric acid after the manner of Norrish and Chappel (1967) and analysed on a Philips X-ray vacuum spectrometer. $K\alpha$ lines were used throughout and the pulse height analyser carefully set to ensure maximum peak to background ratios.

Operating conditions

Element	Tube	Crystal	kV	mA	Counting		Path	Detector	Sensitivity per 10 ppm
					time sec.				
K	Cr	PET	44	26	40	vacuum	flow c.	2.5 counts/ 4 sec.	
Ti	Cr	LiF	44	26	10	vacuum	flow c.	49 counts/ 1 sec.	
Cr	W	LiF	44	26	10	air	flow c.	11 counts/ 1 sec.	
Ni	W	LiF	50	20	10	air	scintillation counter	31 counts/ 1 sec.	

Background measurements were made by averaging the two readings made at $\pm 1.7^\circ 2\theta$

SPECIMEN NUMBER	WEIGHT PERCENT										ATOMS PER UNIT CELL						Reference	Locality			
	Cr ₂ O ₃	Al ₂ O ₃	Fe ₂ O ₃	FeO	MgO	SiO ₂	TiO ₂	Tr.	Al	Cr	Fe ³⁺	Fe ²⁺	8 R ²	ea							
86 120	55.76	15.29	14.65	14.49	1.30	11.4			4.6	11.4		4.2	3.6		Collenette (1964)	Baregi Island					
NW 6	56.56	15.35	NL	13.22	14.14	2.60	0.25	4.7	11.3			3.1	4.9		Collenette (1964)	Melawai Island					
NW 12 (32)	54.99	11.05	0.87	12.32	16.65	4.25	0.23	3.7	12.3			5.7	2.3		Collenette (1964)	Melawai Island					
NB 5452	54.5	8.9	16.0	15.0	4.5	0.03	3.0	12.5	0.5	3.0	5.0		8.333		Willson (1951)	Melawai Island					
J 1187	53.5	12.7	15.7	15.5	3.5	0.10	4.1	11.7	0.2	3.1	4.9		6.288		This Paper	Lala Island Darvei Bay					
EW 105	51.57	18.04	6.51	8.95	13.60	1.25	0.25	5.5	10.5			3.2	4.8		Collenette (1964)	Labuk Valley (Bakit Lumair)					
LM 2	50.48	10.47	0.22	12.61	18.76	4.77		3.5	11.4	1.1	2.0	6.0			Collenette (1964)	Soodie Islands Darvei Bay					
NB 1	50.42	16.09	15.79	15.39	1.75			5.0	10.5	0.5	2.5	5.4			Newton-Smith (1957)	Labuk Valley (Porog)					
LD 6	48.48	10.48	1.76	11.41	19.15	6.33		3.7	11.4	0.9	2.3	5.7			Collenette (1964)	Mount Silan Darvei Bay					
NB 2	48.30	16.69	15.57	15.55	3.40			5.5	10.5			3.3	4.7		Newton-Smith (1957)	Labuk Valley (Porog)					
LD 4	47.37	9.25	5.05	9.31	18.40	8.61		3.5	12.1	0.4	3.3	4.7			Collenette (1964)	Kalong Island Darvei Bay					
NB 3	45.93	16.57	14.49	17.60	4.92			5.5	10.2	0.3	2.7	5.3			Newton-Smith (1957)	Labuk Valley (Porog)					
J 224	45.5	16.5	15.5	15.9	3.1	0.15		5.3	9.9	0.6	2.4	5.6			This Paper	Labuk Valley Siidu-Siidu					
J 227	45.3	16.4	15.7	15.9	1.8	0.20		5.7	9.4	0.9	2.1	5.9			This Paper	Labuk Valley Siidu-Siidu					
EW 108	43.79	24.41	2.82	11.20	19.00	1.75	0.25	7.3	8.7			2.5	5.2		Collenette (1964)	Labuk Valley (Bakit Lumair)					
J 8080	39.6	27.4	14.1	17.4	2.5	0.19		8.1	7.8	0.1	2.5	5.5			This Paper	Mount Silan Darvei Bay					
EW 68	36.69	12.53	39.29	7.99	2.19	0.79		4.6	9.4	2.0	6.9	4.1			Collenette (1964)	Baregi Island					
H 134	35.59	24.46	13.54	17.37	5.09	0.50		8.1	7.9			2.9	5.1		Collenette (1964)	Labuk Valley (Siku-Puku)					
LS 49	32.9	24.05	13.8	15.5	7.05			8.3	7.7			3.2	4.8		Collenette (1964)	Mount Silan Darvei Bay					
LS 50	32.55	25.25	13.75	15.05	5.65			8.5	7.1	0.4	2.4	5.6			Collenette (1964)	Mount Silan Darvei Bay					
NB 4169	31.4	28.4	14.2	17.2	4.4	0.19		9.1	6.5	0.1	2.6	5.2			Collenette (1966)	Kinabalu Area Pataharung					
										Totals		0.51		0.63		0.59		0.74		0.65	

TABLE 1. CHEMICAL ANALYSES AND UNIT CELL CAPTURED ON THE BASIS OF 32 OXYGENS, OF SOME CHROMITES FROM SARAH. A FEW UNIT CELL EDGES DETERMINED BY POWDER DIFFRACTION ARE INCLUDED. LOCALITIES ARE DESCRIBED IN APPENDIX.

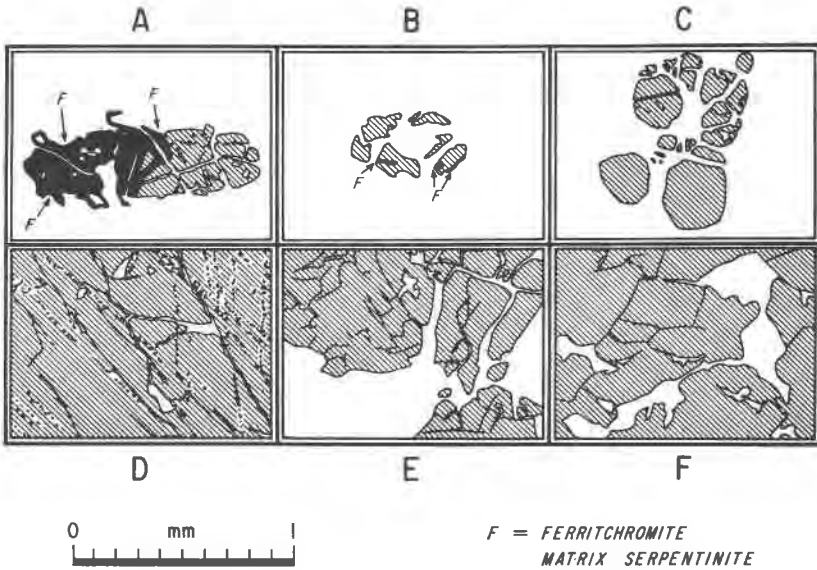


FIG. 4. Drawings from polished sections of chromite occurrences in Sabah. The matrix is generally serpentinite. F = ferritchromite with considerably higher reflectivity than chromite, which is shown hatched. A = specimen 7182, accessory hydrothermally altered chromite from the Ranau Luhan road, Kinabalu area. B = J 1186, accessory chromite with small areas of ferritchromite from Laila island. C = J 227, rounded chromite grains from Bidu-Bidu in the Labuk Valley. D = BB 120, massive fractured chromite from Banggi island. E and F = NB 4169, massive fractured chromite from the Kinabalu area, E from the edge and F from the interior of a layer.

In view of the apparent absence of chromite, except as a very minor accessory, in the ultramafic bodies of Sakar and the outer islands of Tabawan and Maganting (Fig. 2), a selection of rock specimens from all the Darvel Bay ultramafic outcrops was analysed for chromium and related elements. The results are shown in Figure 5. Specimens 1 through 17, and 19 and 27 are ultramafites. At the same time a number of *Silumpat Gneiss* specimens, 18 and 20 through 26 were analysed.

The rock contents of chromium, nickel, and titanium, when compared in relation to the potassium contents, show the sharp distinction between the ultramafic rocks and the *Silumpat Gneiss* (Fig. 5). Excluding number 27 from this study because it is abnormally enriched in potassium, we find that the Darvel Bay ultramafic rocks have the following average contents:

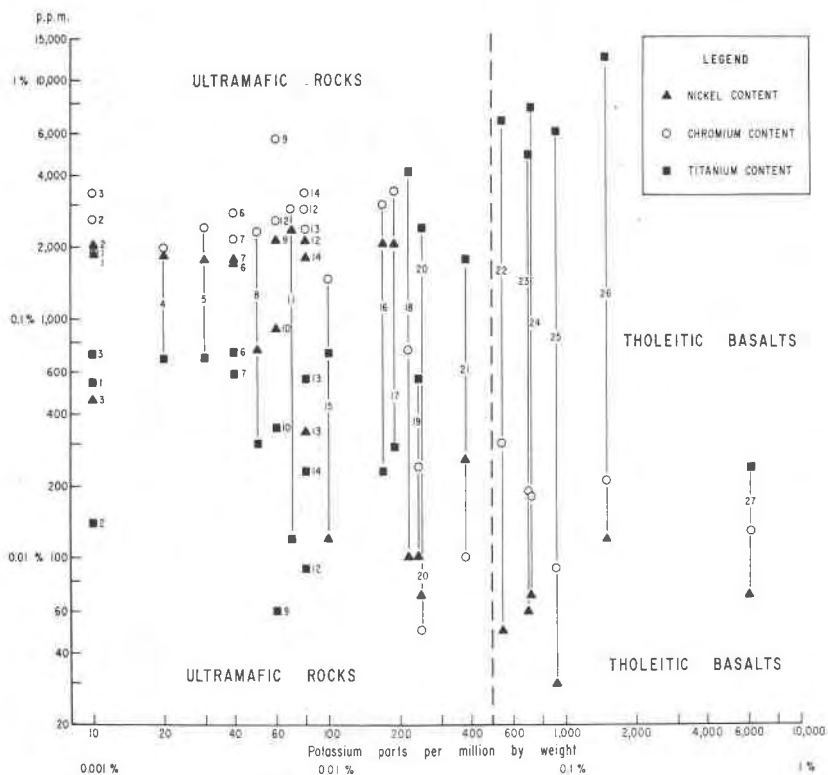


FIG. 5. Plot on logarithmic scales of the Cr, Ni, Ti, and K contents of some Darvel Bay rocks. The dashed vertical line is an arbitrary division between the potassium-based fields of ultramafic rocks and tholeiitic basalts. The Silumpat Gneiss extends to the right from number 18, but 27 is an ultramafic rock anomalously rich in potassium. The Silumpat Gneiss is characterised by its richness in potassium and titanium; the ultramafites by their richness in chromium and nickel. 1 = J 1080, Dunite from Tabawan. 2 = J 1095, Serpentinite from Tabawan. 3 = J 1191, Pyroxenite from Nipa-Nipa. 4 = J 1168, Peridotite from Maganting. 5 = J 1103, Peridotite from Tabawan. 6 = J 1083, Peridotite from Tabawan. 7 = J 1101, Peridotite from Tabawan. 8 = J 1175, Peridotite from Katung-Kalungan. 9 = J 1220, Serpentinite from Saddle. 10 = J 1245, ultrabasite from Silam. 11 = J 1329 Peridotite from Sakar. 12 = J 1182, Serpentinite from Baik. 13 = J 1200, Pyroxenite from Giffard. 14 = J 1319, Peridotite from Sakar. 15 = J 1196, Pyroxenite from Enam. 16 = J 1088, Dunite from Tabawan. 17 = J 1265, ultrabasite from Sakar. 18 = J 1344, Silumpat Gneiss from Silam. 19 = J 1185, Pyroxenite from Laila. 20 = J 1075, Silumpat Gneiss from Tabawan. 21 = J 1324, garnet amphibolite from Sakar. 22 = J 1050, Silumpat Gneiss from Silumpat. 23 = J 1060, Silumpat Gneiss from Silumpat. 24 = J 1261, Silumpat Gneiss from Silam. 25 = J 1127, Silumpat Gneiss from Tabawan. 26 = J 1337, Silumpat Gneiss from Sakar. 27 = J 1215, Pyroxenite from Enam.

K 74 ppm, ranging from 10 to 240 (excluding 27)

Ti 412 ppm, ranging from 60 to 730

Cr 2770 ppm, ranging from 1450 to 5690 (excluding 19 and 27)

Ni 1531 ppm, ranging from 120 to 2350 (excluding 19 and 27).

When compared with the worldwide average ultramafic values of Goles (1967): K = 200 ±, Ti = 300, Cr = 2400, and Ni = 1500 ± ppm, the Darvel Bay ultramafic rocks seem remarkably low in potassium and high in chromium. This might be taken to indicate that they represent a suite of rocks strongly residual in character and to have come accordingly from a Mantle region from which much basaltic magma has already been withdrawn by partial melting.

It is interesting to note that although no segregations of ore occur on the islands of Sakar, Tabawan, and Maganting, yet the ultramafic rocks from these islands are often as rich in chromium as the rocks from the Saddle islands. Saddle island Group specimens in Figure 5 are 3, 8, 9, 10, 12, 13, 15, 19, and 27. These do not differ significantly in their distribution of Cr, Ni, Ti, and K from the ultramafic masses of other Darvel Bay areas. Of course, the specimens that were selected for analysis did not contain layers and segregations of chromite. Such layers and segregations occur only in the Saddle island Group and on Mount Silam. Hutchison and Dhonau (1969) showed by structural data (summarised on Fig. 2) that the deformation of the Saddle island Group ultramafic body is later than elsewhere in Darvel Bay. Accordingly it must be concluded that not all ultramafic emplacements are of the same age and presumably not all from the same part of the Mantle.

The Saddle island Group ultramafites have come from a Mantle area that had a considerable development of chromite layering, and Hutchison and Dhonau (1969 and 1971) showed that this was a cold emplacement. On the other hand the ultramafite of Tabawan was of sufficient temperature to have raised the Silumpat Gneiss in a thermal aureole to as high as hornblende granulite facies.

Specimens 18 and 20 through 26 are of Silumpat Gneiss, which Hutchison and Dhonau (1969 and 1971) believe to be metamorphosed oceanic basaltic crust. For the specimens analysed, the following average contents have been found:

K	653	ppm, ranging from 220 to 1500
Ti	5778	ppm, ranging from 1760 to 12690
Cr	232	ppm, ranging from 50 to 740
Ni	96	ppm, ranging from 30 to 260

These values compare reasonably well with the average worldwide

tholeiitic basalt: K 5400 ppm with a lower limit of about 500 ppm (Manson, 1967); Ti 9900 ppm (Manson, 1967); Cr 218 ppm (Prinz, 1967); Ni 85 ppm (Prinz, 1967). From a comparison of these values, the Silumpat Gneiss can well be regarded as typically tholeiitic in character. Metamorphism to amphibolite and granulite facies has caused a partial depletion in potassium and perhaps titanium. One of the specimens J1324, is of a *Silumpat Gneiss* inclusion in the ultramafic mass of Pulau Sakar. It is now a garnet amphibolite, and although it must have been considerably modified by the enclosing ultramafic mass, its characters as shown in Figure 5 (No. 21) are still that of the Silumpat Gneiss. It is hard therefore to escape the conclusion that the Silumpat Gneiss is distinct both chemically and mineralogically from the ultramafic bodies of the Darvel Bay area. Dr T. P. Thayer (personal commun.) has questioned our interpretation that the Silumpat Gneiss is divorced in age and consanguinity from the ultramafic bodies because as he has noted elsewhere, Thayer (1967), he regards "the ultramafic and gabbroic parts of ophiolite complexes as being identical with, and part of, the alpine igneous rock suite." However the problem has been to a large extent resolved by Leong (1971) who has recently re-mapped large areas of the hinterland of Darvel Bay. He has found that there is a distinct Crystalline Basement which includes the Silumpat Gneiss and is, by radiometric and stratigraphic evidence, of pre-Lower Triassic age. In addition there are distinct occurrences of banded gabbros associated with the ultramafic rocks and these form the peridotite-gabbro complexes which post-date the Crystalline Basement. The problem in Sabah has arisen from the fact that the banded gabbros of the alpine complexes have in the past been confused, especially by Fitch (1955, and 1970), with the banded amphibolite gneisses which constitute the Silumpat Gneiss of the Crystalline Basement.

CHROMITE CHEMISTRY

A number of analyses of chromites separated from ultramafic rocks of Sabah have been found in the literature. These are given in Table 1 together with a few analyses by the present author by X-ray fluorescence. The conversion to unit cell cationic proportions is best illustrated by an example: NB5452.

<u>Oxide</u>	<u>Weight percent</u>	<u>Molecular weight</u>	<u>Cationic proportions</u>
CrO _{1 1/2}	54.5	75.01	0.7170
AlO _{1 1/2}	8.9	50.98	0.1745

Total iron expressed as

FeO _{1/2}	16.0	79.85	.2003 less .0003 for ilmenite impurity = .2000
MgO	16.0	40.32	.3968 less $1\frac{1}{2} \times .0748$ for serpentine impurity = .3968 - .1122 = .2846
Total cations			1.3761
SiO ₂	4.5	60.09	.0748
TiO ₂	0.03	79.90	.0003

For the basic formula $R^{3+}_{16}R^{2+}_8O_{32}$, then multiply by 24/1.3761, giving $(Cr_{12.5}Al_{3.0}Fe_{0.5})_{16}(Fe_{3.0}Mg_{5.0})_8O_{32}$. The total iron of 3.5 is first used to make R^{3+} up to a total 16, the remainder going to R^{2+} . In all the samples, the impurities were considered for consistency to be ilmenite and serpentine and their amounts calculated, as shown in the above example, by allocating all Ti to ilmenite and all Si to serpentine.

Purification Before Analysis

Samples were crushed and sieved -120 + 230 U. S. standard, washed as often as necessary with water to remove fine powder, then dried. The very magnetic fraction was collected in a Frantz Ferrofilter and discarded. The powder was now slowly passed through a Frantz isodynamic separator with a side tilt of 15° and a forward tilt of 25° (Rosenblum, 1958) and the chromite was collected within the range + 0.1 A, and -0.5 A. The non magnetic fraction was usually serpentine or olivine. If the chromite fraction visibly contained some green material, it was re-run at 0.4 or 0.45 A.

X-ray Fluorescence Analytical Method

Because of the severe matrix problems to be expected between chromium and iron, it was decided to dilute the specimens in a neutral matrix. Only in this way are matrix effects between chromium and iron reduced to a useful degree (Mitchell and Kellam, 1968). Specimens were prepared by mixing thoroughly in a 2 × 1 inch plastic vial with plastic ball pestle on a Wig-L-Bug¹ for 10 min, 0.4 g of the chromite specimen finely ground in an agate mortar with 2.0 g of Analar manganous carbonate. This homogenized mixture was made into a pressed pellet by the method of Norrish and Chappell (1967) and used directly in the spectrometer. Two chromite standards were used, NB 5452 and NB 4169 to obtain calibration curves. At first I tried to make up and use an artificially prepared standard using weighed amounts of Analar Cr₂O₃·(NH₄)₂, Fe₂O₃, Al₂O₃, MgO, quartz, and TiO₂ and mixing them with manganous carbonate in the same ratio. The problem of particle size effect made a nonsense of the results for Mg and Al. Therefore it must be concluded that artificial standards using powder pellets are useless for light element analysis. Provided two widely different chromite specimens can be obtained which have been analysed by wet chemical methods,

¹ Obtainable from Crescent Dental Manufacturing Co., 7750 West 47th St., Lyons, Illinois, 60534, USA

then the simple X-ray dilution procedure is capable of giving satisfactory results. This is especially important because of the difficulty of putting chromite into solution. One important point should be noted and that is that the Fe_2O_3 calibration curve between the two standards had a negative slope, showing the overall importance of matrix effects (Mitchell and Kellam, 1968). Accordingly it is possible that iron determinations by this method are less than perfect. But it appears that Cr, Al, Mg, Si, and Ti determinations are good.

Instrumental Settings for the Elemental Determinations

Element	Tube	Crystal	kV	mA	Counting time sec.	Average sensitivity for 1% in chromite
Cr_2O_3	W	LiF	44	8	10	603 counts per sec.
Al_2O_3	Cr	PET	44	26	20	20 counts per 2 sec.
MgO	Cr	KAP	44	26	40	6 counts per 4 sec.
Fe_2O_3	W	LiF	50	8	10	790 counts per sec.
TiO_2	Cr	LiF	44	16	10	6700 counts per sec.
SiO_2	Cr	PET	44	26	20	88 counts per 2 sec.

All determinations were with a flow counter and vacuum path. The background of each peak was determined at $\pm 1.6^\circ 2\theta$ from the peak position for each element except in the case of iron where it had to be at $\pm 2.3^\circ$ because of the close position of the manganese K β peak.

Discussion

The analysed chromites of Table 1 are shown on Figure 6. The majority of Sabah chromites fall in the field of aluminian chromite but a few are in the chromian spinel compositional range of Stevens (1944), and a few can be considered as of metallurgical grade. Figure 7 shows that with only three exceptions (BW 68, MW 12(32), and BB 120) the Mg:Fe ratio in R^{2+} lies between the small range of 4.7:3.3 and 6.0:2.0. This remarkably small range in R^{2+} while Cr:Al ranges from 12.5:3.0 to 6.8:9.1 (Table 1) is remarkable. Thayer (1970) has brought attention to this very fact that as alpine-type chromites become less and less rich in chromium, they become richer in aluminium and *not* in iron; indeed the iron characteristically remains approximately constant. In contrast, stratiform chromites become progressively richer in total iron as their chromium content falls. The Sabah chromites are in this respect typically Alpine. Another interesting difference is that stratiform chromites characteristically have a unimodal and restricted chromium range. Alpine type, or podiform chromites, have a much wider range of chromium content which is characteristically bimodal (Thayer, 1970). The range of chromium in the Sabah chromites (Fig. 6) is considerably wider than in any stratiform body and there is a suggestion of a bimodal chromium con-

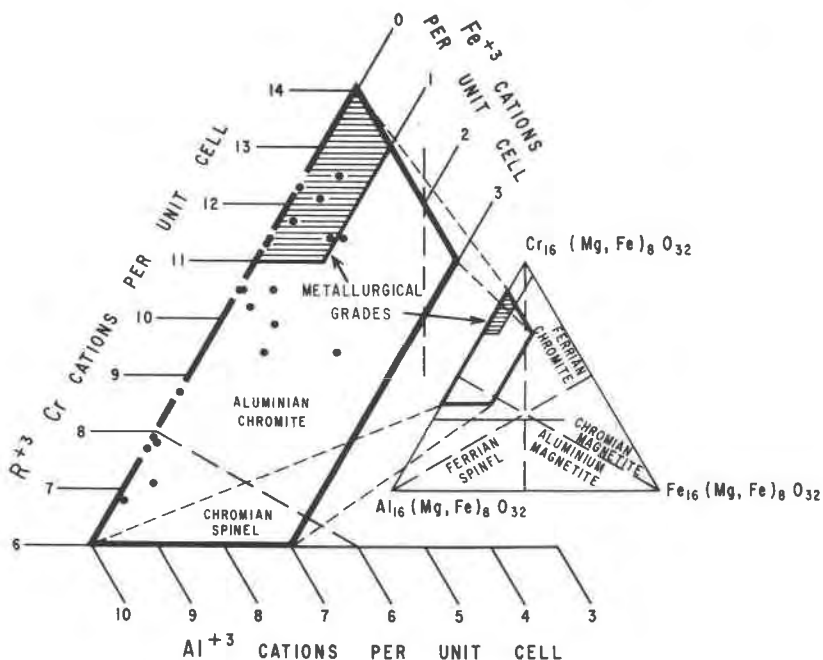


FIG. 6. Distribution of Sabah chromites in the $Cr_{16}-Al_{16}-Fe_{16}R^{3+}_{32}$ triangular classification. The actual cationic proportions are plotted in only a small portion (left) of the complete triangle (shown right).

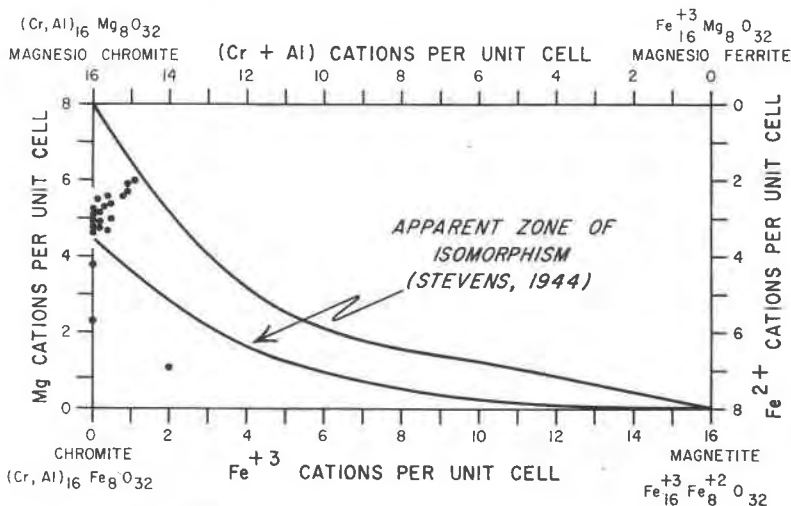


FIG. 7. Diagram to show the remarkable lack of variation in $Mg:Fe^{2+}$ ratio for the same chromites. Stratiform chromites show a much wider range.

tent with one mode around 7.9 and the other around 11.3 chromium cations per unit cell.

PHYSICAL PROPERTIES

Unit Cell Edge

Stevens (1944, Fig. 2) plotted percent Cr_2O_3 against cell edge a in Å and found, with a few significant exceptions, a good linear fit. His data can be recalculated to give the following regression equation:

$$\text{percent Cr}_2\text{O}_3 \text{ in chromite} = 301.85 (a\text{\AA}) - 2445.66$$

His data consisted of 25 measurements of chromites ranging from Cr_2O_3 0.0 to 59.5 percent but excluding two samples which had high ferric iron contents. But I do not feel that this was a valid justification for their exclusion, for the difficulty of putting chromite into solution for analysis must cast doubt on chromite $\text{Fe}^{3+}:\text{Fe}^{2+}$ determinations.

Stevens (1944) must have been fortunate in his choice of data, for the Sabah chromites do not fit well on his curve. Indeed it is surprising that a linear relationship between Cr_2O_3 and a exists, for in the chromite unit cell $(\text{Cr}, \text{Al}, \text{Fe}^{3+})_{16} (\text{Fe}^{2+}, \text{Mg})_8 \text{O}_{32}$ there are five variables with Cr, Fe^{3+} , and Mg of roughly equivalent ionic radius (0.63, 0.64, and 0.66 Å respectively) and Al (0.51 Å) and Fe^{2+} (0.74 Å). Hence it should be expected that variation in Al and in Fe^{2+} would have much more control on the unit cell edge than Cr. Again we would expect to find a different relationship in stratiform chromites, where total iron increases as Cr decreases, than in podiform chromites, in which aluminium increases as chromium decreases. In the present study several chromites listed in Table 1 were measured for unit cell edge (a) on a diffractometer.

Chromite mixed with an almost equal amount of potassium bromate was smeared on a glass slide and run with $\text{CuK}\alpha$ radiation at $1/8^\circ$ per minute from 61.5 to $63.5^\circ 2\theta$. The 2θ difference between KBrO_3 K α_{1220} at 61.61 and the chromite K α_{044} at approximately 63.16° is added to 61.61 to give the precise chromite 2θ K α_{044} . From this, $a^2 = 32 d_{044}^2$. The results are tabulated in Table 1. Chromite peaks for all Sabah specimens are sharp and well resolved indicating the homogeneous nature of each specimen and the general absence of hydrothermal alteration to ferritchromite.

The variation of a in Å with Al and Cr cations per unit cell is shown on Figure 8. The Al content in R^{+3} is considered to be the most important variant because of the small size of the Al cation. The ratio of Mg:Fe in R^{+2} is likely to have a pronounced effect also on a and the value given against each point on Figure 8 is the number of Fe^{+2}

computed in the unit cell. Since this has a large ionic radius it is likely to affect the size of a . The points in solid circles are from this study and those in open circles are from Stevens (1944). A considerable spread of values has been obtained (Fig. 8) and the spread from the computed best-fit linear correlations is not systematically related to Fe^{2+} contents as hoped. The regression equations for these data, bearing in mind that the spread is considerable, are:

Number of aluminium cations per unit cell = $407.30 - 48.60 (a \text{ in } \text{\AA})$

Number of chromium cations per unit cell = $58.46 (a \text{ in } \text{\AA}) - 473.88$

But these equations can be used only for very approximate determinations. Perhaps the reason for the imperfect nature of the data of Figure 8 is the oxidation state of iron. Many chromites may have some of the iron in R^{2+} oxidised to ferric iron. The change in ionic size would be expected to distort the lattice. Be this as it may, the results of Figure 8 show that at this time X-ray diffraction determination of chromite unit cell edge (a) cannot be with certainty related to the composition of the chromite and accordingly doubt must be cast on the validity of Steven's (1944) relationship between Cr_2O_3 and a .

Reflectivity

Polished specimens of 11 of the chromites listed on Table 1 were compared with SiC No. 88 (reflectivity given as 20.16 percent at a 590 nm) on a Vickers microscope fitted with an EEL digital-readout microphotometer. The results of several determinations on different crystals within each polished specimen are given in Table 2. An excellent correlation was obtained of reflectivity with the number of aluminium cations per unit cell (Fig. 9) giving a computed regression equation:

$$\begin{aligned} \text{Number of aluminium cations per unit cell} \\ = 27.93 - 1.82 \times \text{reflectivity (at 590 nm)} \end{aligned}$$

The relationship given in Figure 9 appears to be a precise and quick method of determining Al content in chromites. However this relationship has been determined on alpine-type chromites in which, as chromium content drops, aluminium rises progressively. This same relationship is not expected to apply exactly to stratiform chromites in which the overall iron content rises with chromium fall-off. For such chromites the slope would be expected to be less steep or even reversed.

A wavelength of 590 nm was used, so that a comparison could be

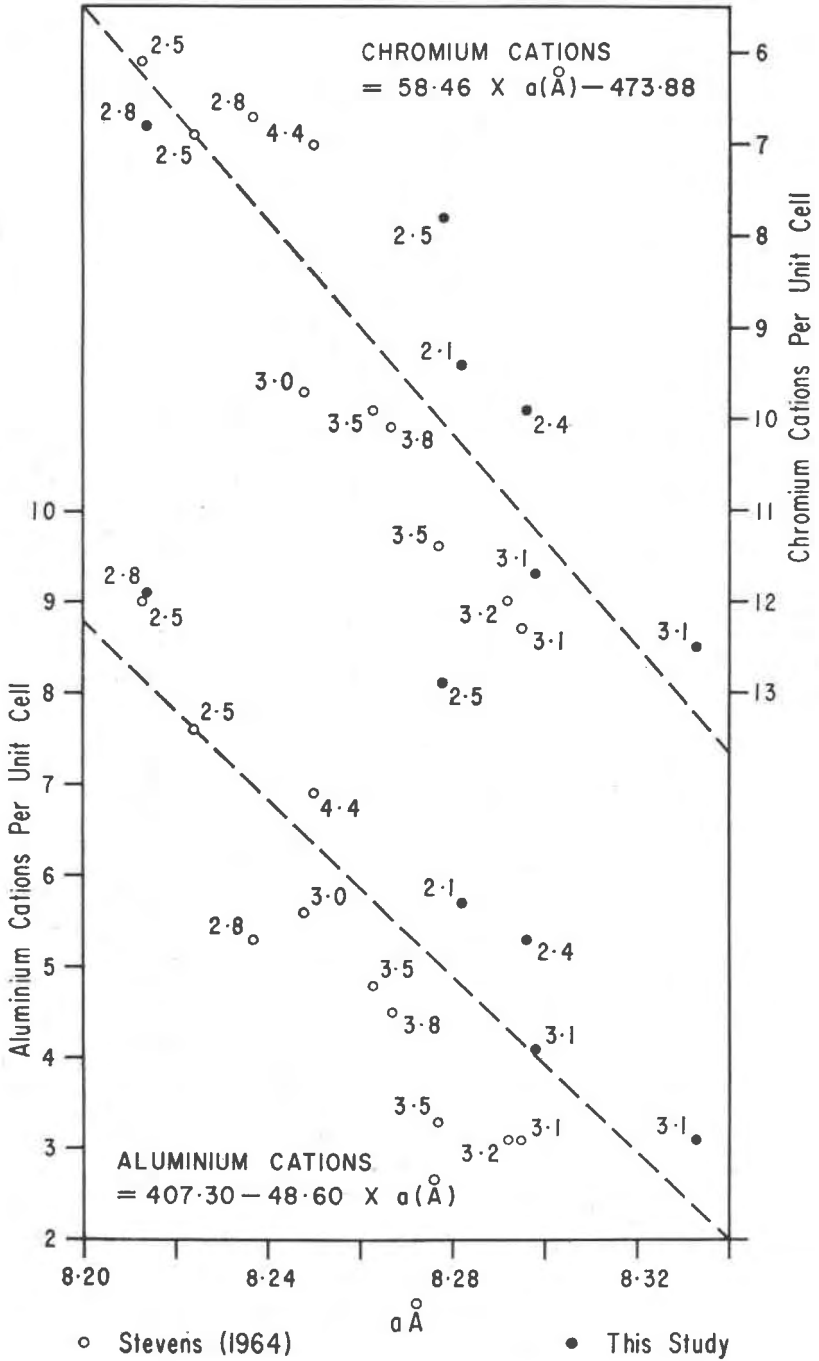


TABLE 2. REFLECTIVITY AT WAVELENGTH 590 NM (COMPARED WITH SIC NO. 88) AND VH NUMBERS FOR SOME SABAH CHROMITES.

SPECIMEN NUMBER	REFLECTIVITY % 590NM	VICKERS HARDNESS
C H R O M I T E		
NB 5452	13.7 ± .1	1380 ± 90
J 1186	13.5 ± .3	1400 ± 20
J 1187	13.4 ± .4	1510 ± 60
J 224	12.6 ± .3	1430 ± 40
BB 120	12.5 ± .1	1370 ± 60
J 227	12.4 ± .3	1290 ± 240
EW 105	11.9 ± .1	1470 ± 120
J 8090	11.4 ± .3	1320 ± 80
H 134	10.7 ± .1	1400 ± 90
NE 4169	10.2 ± .3	1450 ± 170
J 1182	9.6 ± .1	1470 ± 60
F E R R I T C H R O M I T E		
J 1186	19.7 ± .9	1070 ± 50

made with the values of Mihálik and Saager (1968), who quoted a similar range of values but did not try to relate the reflectivity to composition. They noted that chromite grains in the Basal Reef of the Witwatersrand System often show higher reflecting borders as a result of alteration *in situ*. The alteration results in an increase in iron and chromium, and a decrease in aluminium and magnesium. This alteration is now regarded by most workers as being hydrothermal (see Engin and Aucott, 1971) and it gives rise to higher reflectivity values. Indeed it resembles magnetite in polished section, and if we follow the scheme of Stevens (1944) as shown in Figure 6, the hydrothermal alteration of the Sabah chromites would be expected to put them into the field of ferrian chromite, which is now called by most people ferritchromite.

Reflectivity values of the Sabah ferritchromite are higher than those given by Mihálik and Saager (1968). They given 13.6 percent



FIG. 8. Relationship between the unit cell edge (a , Å) and the number of Al and Cr cations per unit cell for 6 Sabah chromites and 10 from Stevens (1944). The number beside each point is the number of iron cations in R^{2+} .

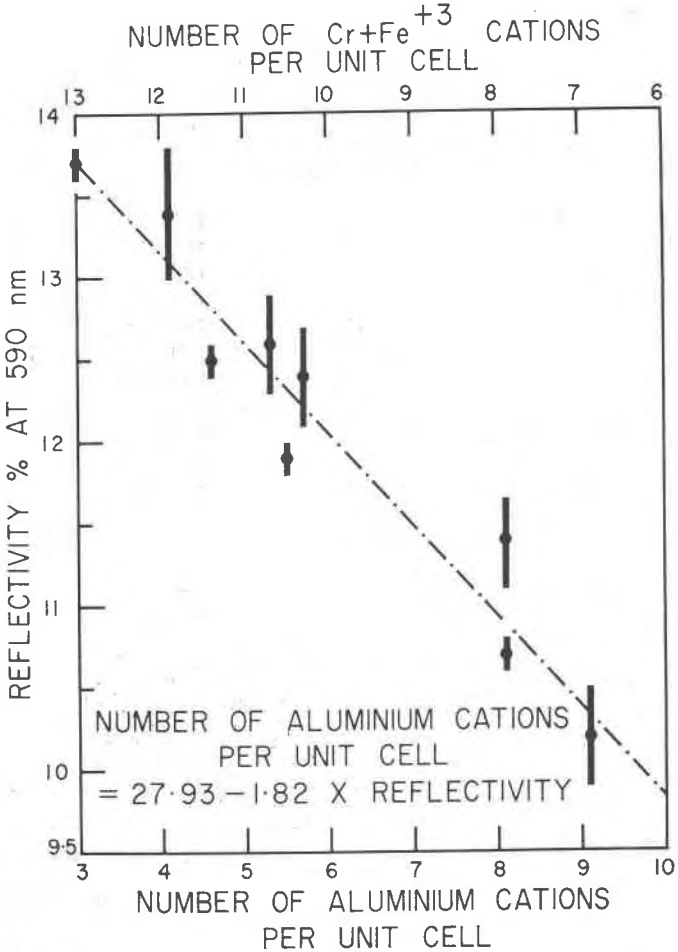


FIG. 9. Variation of chromite reflectivity at 590 nm wavelength (compared with SiC No. 88 of given 20.16 percent reflectivity) with change in aluminium content.

as their highest determination, but this study indicates a value of the order of 20 percent (Table 2). This reflectivity is more characteristic of magnetite, and indeed Ramdohr (1968) shows an alpine chromite (p. 929) rimmed progressively by ferritchromite and then by magnetite.

Hardness

Hardness was determined using a 200 g load and 25 sec. contact time on a Leitz miniload hardness tester calibrated against a hardened

steel standard. The values obtained are listed in Table 2. The figures are considerably higher than those quoted by Mihálik and Saager (1968), but hardness can be of little determinative use for indicating chromite composition because at these high values the margin of error is greater than the range of variability sought.

ULTRAMAFITE—ORIGIN AND EMPLACEMENT

The wide range of chromium content of Sabah chromites in common with all podiform chromites, the tendency to a bimodal chromium content, and the absence of iron enrichment as chromium falls, all point to major and essential differences between Alpine-type and stratiform ultramafites. These differences must reflect a radically different ultimate origin. It seems reasonable therefore to presume that Alpine-type ultramafites have evolved in the Mantle just as stratiform ultramafites have differentiated from a gabbroic magma in the crust. For these reasons the hypothesis of McTaggart (1971) that Alpine-type ultramafites were originally identical to stratiform types as cumulates in basic magma chambers high within the crust, subsequently subsiding during tectonism to form cold, fault-controlled intrusions downward into lower tectonic levels, cannot be accepted.

The layering of chromite within dunite, showing excellent grading (Fig. 3) indicates that the chromite layers owe their origin to "igneous sedimentation" just as in stratiform deposits, but this differentiation must have taken place in the Upper Mantle. Partial melting of Mantle material to give basaltic magma, which eventually extrudes at oceanic ridges, must leave behind a strongly ultramafic residuum. The very low potassium contents of the ultramafic rocks of Darvel Bay (Fig. 5) indicate this residual character.

The association of the Sabah chromite-bearing ultramafic rocks with gabbro bodies and highly metamorphosed (generally in almandine amphibolite but occasionally in hornblende granulite facies) tholeiitic metabasalts could be accommodated within an oceanic spreading zone with rising Mantle current and accompanying high heat-flow. The work of Karig (1971), further elaborated by Dewey and Bird (1971), shows that such ocean spreading zones not only occur in the main ocean ridges, but also in ridges behind the island arcs of the western Pacific in marginal or inter-arc basins. Indeed Karig (1971) has designated the Sulu basin, which lies in the top right corner of Figure 1, as an inactive or fossil marginal basin with high heat-flow.

The curvilinear outcrop pattern of the ultramafite bodies of North Borneo (Fig. 1) suggest sheet-like emplacements from the Mantle into the crust, and the convex westwards pattern, from Labuk Valley to

Darvel Bay, must be taken to indicate that the ultramafite sheet dips eastwards, implying a westwards and upwards thrusting of Mantle material into the Crust.

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Appendix. Description of chromite localities of Table 1

Bangi Island (Kapitangan)—as thin lenses of 5 to 10 cm thickness in a 1 m band of serpentinized dunite within peridotite. (Sungei Tengah)—as bands and pods up to 10 cm width in strongly faulted and sheared serpentinized dunite (Collenette, 1964).

Malawali Island: strongly brecciated chromite-bearing serpentinite.

Labuk Valley: (Porog)—as veins and disseminations, striking $40^\circ/\sim 90^\circ$ in a dunite belt 61 m wide. One of the larger veins is banded and 4.6 m thick, passing into solid chromite 3.7 m thick. It is locally brecciated. Individual bands vary up to 23 cm thick and crystals seldom exceed 3 mm (Collenette, 1964). (Bukit Lumisir)—as sheared and mylonitized veins. One ore-body is of 4.6×1.8 m of coarse grained chromite. Dunite lenses contain layers, veins, and accessory chromite (Collenette, 1964). (Sungei Taguuk)—as pods and bands up to 15 cm thick in dunite lenses. (Tangulap and Kuun-Kuun)—as a 6 mm layer and disseminations in dunite lenses. (Ruku-Ruku)—as a pod, 80×50 cm, and as thin stringers in dunite within peridotite.

Saddle Island group: The thickest dunites and serpentinites of Sabah occur here. As tabular layers grading into disseminations, and sometimes as irregular bodies with sharp contacts against dunite (Collenette, 1964). (Katung-Kalungan island)—as numerous discontinuous bands a few mm to 6 cm thick in serpentinite. (Laila island)—as common discontinuous bands, 3 to 13 mm thick, locally swelling to 60 cm. (Saddle island)—as a fractured band, 30 cm thick, increasing to 1 m, outcropping over a 3 m distance. (Nipa-Nipa)—as bands 13 mm to 5 cm thick in an ore zone 50 cm wide.

Mount Silam: An ore body, 13×7 m \times 60 cm maximum thickness, dips irregularly N.E. at 15° . It has a skin (after sheared dunite) of schistose serpentinite 2 to 30 cm thick, weathering to a lateritic clay. The ore body and serpentinite envelope is thought to have been detached from its parent dunite and intruded upwards into the overlying peridotite (Collenette, 1964).

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