



The Archean Komatiite-Related Inyala Chromitite, Southern Zimbabwe

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Abstract

The chromitites at the Inyala mine, Zimbabwe, form part of a layered intrusion enclosed within the Archean granulite facies gneisses of the north marginal zone (NMZ) of the Limpopo belt. Chromitites occur within an olivine-orthopyroxene-chromite cumulate sequence which was derived from a komatiitic magma.

An electron microprobe study and a comparison with the mineralogy of unaltered komatiites from the Belingwe greenstone belt, Zimbabwe, strongly suggest that some of the original olivine, chromite, and orthopyroxene mineral compositions are preserved. Olivine-chromite mineral pairs continued to equilibrate down-temperature until about 680°C. Calculations based on the diffusion of Fe and Mg in olivine suggest that the cooling was rapid (several 1,000°C/Ma), hence it is reasonable to expect igneous mineral compositions to be preserved. Calculations based on mineral-melt, Fe-Mg partitioning for olivine and orthopyroxene indicate that the parent komatiitic liquids varied in composition from 26 to 16 wt percent MgO. Thus some of the chromites crystallized from melts which are among the most magnesian komatiites known and which would have had an eruption temperature of ca. 1,520°C.

Orthopyroxene cumulates, which are unusual in komatiitic sequences probably formed in response to assimilation of adjacent supracrustal rocks, possibly banded iron-formation. Oxygen activity calculations suggest that assimilation may have led to an increase in the oxygen activity in the melt which may be responsible for the crystallization of both chromite and orthopyroxene.

Introduction

CHROMITE deposits of Archean age are known from a variety of layered mafic-ultramafic intrusions worldwide (e.g., Stowe, 1987a). The chemistry of Archean chromite deposits is diverse (Rollinson, 1995a), suggesting that they have originated from a variety of parental magma types in a range of tectonic settings. A particular difficulty in using mineral chemistry to decipher the origin of chromitites is that chromite grains frequently experience substantial chemical change during cooling (e.g., Eales and Reynolds, 1986) and subsequent metamorphism (Evans and Frost, 1975). This means that their compositions reflect subsolidus processes rather than the composition of the parent magma (Rollinson, 1995a, b). The effects of subsolidus exchange can be acute in Archean chromite deposits because they have frequently been metamorphosed, sometimes to granulite grade, further obscuring primary igneous compositions.

This study describes the geologic setting and mineral chemistry of a chromitite body from the Archean Limpopo belt in southern Africa in which it is possible to recognize original igneous mineral compositions. The purpose of this paper is to show from the mineral chemistry that the chromitites were derived from a magnesian komatiite. The preservation of original igneous compositions can be understood in terms of mass balance within the cumulates and the rapid cooling of the igneous intrusion. As will be shown below, the chromitites are found in a medium-pressure granulite terrane, nevertheless the emphasis of this paper is on the processes prior to the granulite facies metamorphism.

Geologic Setting

The Inyala chromitite is located toward the western end of the north marginal zone of the Limpopo belt (NMZ) in Zimbabwe (Fig. 1). It is one of several chromitites associated

with ultramafic bodies enclosed within the 2.6 to 2.7 Ga tonalite-trondhjemite-granodiorite (TTG) gneisses of the north marginal zone.

The north marginal zone is the most northerly of three subzones of the Limpopo belt and crops out principally in southern Zimbabwe. It represents a zone of granulite facies gneisses (Robertson and du Toit, 1981) which have been thrust onto the amphibolite facies to greenschist facies granites and greenstones of the Zimbabwe craton to the north (Rollinson and Blenkinsop, 1995). The precise relationship between the north marginal zone and the Zimbabwe craton has been the subject of some debate, centering around the estimated displacement along the bounding thrust. Rollinson (1993) argued that the thrust represented a crustal-scale shear zone which formed the boundary between two distinct terranes. More recent evidence (Mkweli et al., 1995) suggests that this is not the case and it is likely that the north marginal zone represents typical Zimbabwe craton lower crust and that a section from the north marginal zone into the Zimbabwe craton marks a near-complete crustal cross section.

The geology of the north marginal zone of the Limpopo belt has recently been described by Rollinson and Blenkinsop (1995) and Kamber and Biino (1995). Five key stages have been identified:

Stage 1. An early suite (>2.7 Ga) of supracrustal rocks, comprising banded iron-formation, basaltic lavas, and ultramafic rocks. These lithologies are similar in composition to those found in the greenstone belts of the Zimbabwe craton.

Stage 2. Intrusion of charnockitic magmas of tonalitic composition, between 2.6 and 2.7 Ga (Berger et al., 1995). This event dominated the geology of the north marginal zone and accounts for 90 percent of the present-day outcrop.

Stage 3. The intrusion of the porphyritic granites of the Razi granite suite between 2.62 and 2.58 Ga (Mkweli et al.,

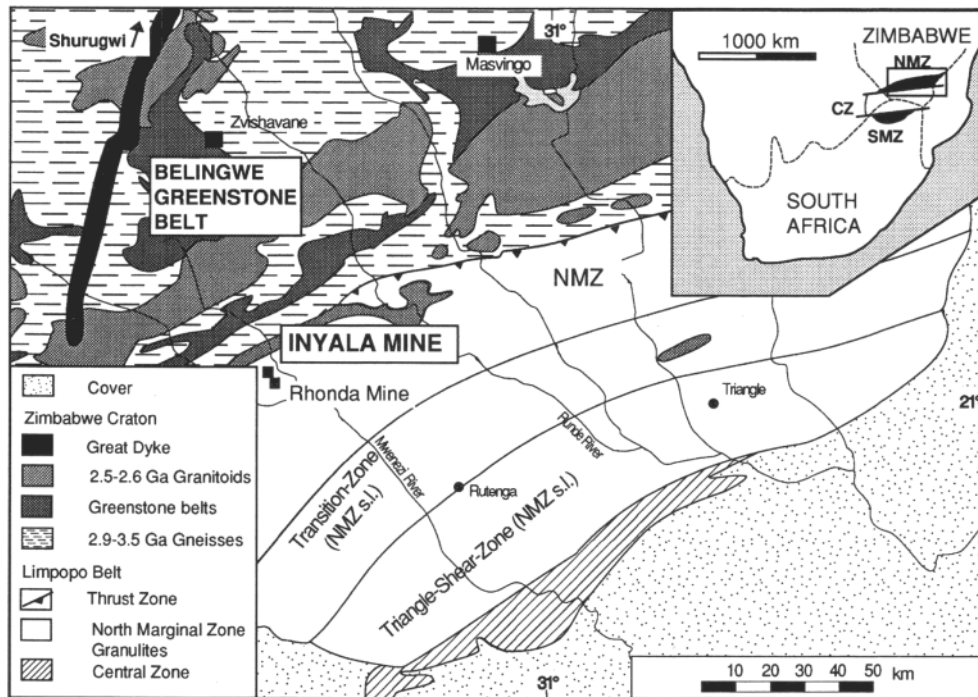


FIG. 1. Geologic map of the north marginal zone of the Limpopo belt showing the location of the Inyala mine. Note that the north marginal zone is subdivided into the north marginal zone proper (NMZ), a transition zone, and the triangle shear zone (NMZ s.l.). The inset map shows the location of the north marginal (NMZ), central (CZ), and south marginal (SMZ) zones of the Limpopo belt in southern Africa.

1995) along the thrust boundary between the north marginal zone and the Zimbabwe craton.

Stage 4. Coeval with the final stages of granitoid emplacement was a low- to medium-pressure granulite facies event which affected all lithologies except the very last intrusions. The P-T conditions of the granulite facies metamorphism have been calculated from the data of Sherlock (1996) for the north marginal zone in the particular area of this study, based on a felsite collected 8 km to the southeast of the Inyala mine. Using the garnet-orthopyroxene thermobarometric methods described in Rollinson (1989), an early garnet-orthopyroxene assemblage records metamorphic equilibration at about 5 kbars and 800°C. A later generation of garnet equilibrated at temperatures of around 730°C and at slightly higher pressures (ca. 6 kbars). These results confirm the anticlockwise P-T path described for the north marginal zone by Sherlock (pers. commun.).

Stage 5. Final exhumation of the north marginal zone took place during the Proterozoic at 1.97 Ga (Kamber et al., 1996) implying prolonged residence of the north marginal zone at depth.

Field relationships at the Inyala mine

The Inyala chromitite belongs to the early supracrustal suite of the north marginal zone (stage 1 above) and is located within a sequence of dunites and pyroxenites. Inyala is one of a number of ultramafic bodies in the western part of the north marginal zone many of which contain chromite. Normally these ultramafic bodies are associated with banded iron-formation and although this is not the case at the Inyala mine,

chromitiferous ultramafic rocks associated with banded iron-formation are located within 2 km of the mine (Prendergast, 1995). At the Inyala mine itself, hornblendites (probable metabasalts of the early supracrustal suite) are found as lenses in the pyroxenites. Although intrusive contacts have not been directly observed at the Inyala mine itself, geologic relationships elsewhere in the north marginal zone indicate that the dunites and pyroxenites predate the intrusion of the tonalite-trondhjemite-granodiorite suite. Zircon ages for tonalite-trondhjemite-granodiorite in this part of the north marginal zone are between 2603 ± 64 and 2637 ± 19 Ma (Berger et al., 1995), indicating that the early supracrustal suite dates to at least 2.64 Ga.

Detailed geologic mapping at Inyala has shown that there are two ultramafic bodies each comprising dunites, chromitites, orthopyroxenites, and amphibolites enclosed in tonalite-trondhjemite-granodiorite gneisses (Prendergast, 1995). They represent fragments of a layered intrusion which now form parallel lenses 300 to 400 m apart, each 800 by 170 m and about 150 m thick, which follow the regional strike (ENE) and dip to the south. Facing directions determined underground suggest that the succession faces northwest (Prendergast, 1995) and is inverted. The relationship between the dunites and orthopyroxenites is complex. In both lenses the orthopyroxenites form an incomplete sheath (0–80 m thick) around the dunite, although in places there are also serpentinized dunite layers within the orthopyroxenite. Contacts between the pyroxenite and dunite are sharp. The chromitites form irregular pods within serpentinized dunite. Some chromitite lenses are discordant to the orthopyroxenites, although some

TABLE 1. Representative Mineral Analyses from the Inyala Mine, Zimbabwe

	Chromite												
	Massive chromitite			Impure chromitite			Pyroxenite			Dumite			
SiO ₂	0.05	0.03	0.05	0.05	0.08	0.03	0.03	0.05	0.82	0.05	0.05	0.05	0.10
TiO ₂	0.19	0.18	0.19	0.19	0.21	0.24	0.24	0.51	0.53	0.06	0.08	0.19	0.18
Al ₂ O ₃	11.74	11.92	11.74	11.82	11.73	3.28	3.28	4.38	5.53	14.15	14.29	8.99	8.57
FeO	15.85	15.60	16.08	14.46	14.16	72.04	70.76	68.07	65.58	26.96	27.66	34.95	34.51
MnO	0.38	0.33	0.39	0.29	0.38	0.54	0.50	0.40	0.38	0.30	0.41	0.35	0.46
MgO	13.96	13.97	13.82	14.62	14.72	2.17	2.51	2.37	2.58	9.36	8.76	6.17	6.23
Cr ₂ O ₃	57.50	57.73	57.52	58.67	58.09	12.82	14.99	18.82	20.87	47.88	47.59	47.22	48.39
Total	99.67	99.76	99.79	100.10	99.37	94.82	93.59	94.60	96.29	98.76	98.84	97.92	98.44
Cations including Fe ³⁺ to 24 oxygens													
Si	0.013	0.007	0.013	0.013	0.020	0.008	0.008	0.016	0.236	0.014	0.014	0.014	0.029
Ti	0.037	0.034	0.037	0.037	0.040	0.054	0.054	0.114	0.115	0.012	0.015	0.040	0.037
Al	3.537	3.586	3.536	3.532	3.527	1.156	1.156	1.525	1.879	4.379	4.437	2.930	2.782
Fe ³⁺	0.739	0.678	0.737	0.607	0.641	11.856	11.177	10.610	8.661	1.628	1.596	2.645	2.549
Fe ²⁺	2.650	2.654	2.700	2.444	2.380	6.990	6.819	6.988	7.149	4.294	4.498	5.431	5.400
Mn	0.082	0.071	0.084	0.062	0.082	0.135	0.127	0.101	0.093	0.067	0.092	0.081	0.107
Mg	5.319	5.316	5.266	5.526	5.597	1.118	1.505	1.042	1.110	3.664	3.438	2.541	2.558
Cr	11.623	11.655	11.626	11.763	11.714	3.000	3.541	4.394	4.757	9.943	9.911	10.317	10.538
Total	24.000	24.000	24.000	24.000	24.000	24.000	24.000	24.000	24.000	24.000	24.000	24.000	24.000
Fe#	0.333	0.333	0.339	0.307	0.298	0.880	0.859	0.870	0.866	0.540	0.567	0.681	0.679
Cr#	0.767	0.765	0.767	0.769	0.769	0.753	0.754	0.742	0.717	0.694	0.691	0.779	0.791
Pyroxene													
Pyroxenite													
SiO ₂	57.62	57.60	57.60	57.20	56.02	57.72	57.89	57.73	57.43	58.40	58.09	58.04	58.04
TiO ₂	0.02	0.02	0.02	0.02	0.02	0.04	0.02	0.02	0.02	0.02	0.02	0.02	0.02
Al ₂ O ₃	0.57	0.65	0.65	0.58	0.65	1.10	0.88	0.87	0.85	0.82	0.82	0.90	0.90
FeO	7.79	7.84	7.74	7.74	7.75	7.72	7.64	7.67	7.44	5.02	5.12	5.17	5.17
MnO	0.38	0.27	0.34	0.34	0.37	0.22	0.16	0.14	0.25	0.11	0.08	0.11	0.11
MgO	33.36	33.63	33.96	33.96	33.86	33.52	34.50	34.01	33.88	35.41	35.57	35.28	35.28
CaO	0.25	0.17	0.17	0.17	0.16	0.20	0.09	0.07	0.16	0.22	0.23	0.20	0.20
Na ₂ O	0.21	0.00	0.02	0.02	0.02	0.01	0.04	0.01	0.03	0.03	0.01	0.03	0.03
Cr ₂ O ₃	0.02	0.07	0.07	0.07	0.05	100.53	101.26	100.54	100.10	100.03	99.94	99.75	99.75
Total	100.22	100.25	100.30	100.10	98.90	100.53	101.26	100.54	100.10	100.03	99.94	99.75	99.75
Cations to 6 oxygens													
Si	1.994	1.991	1.991	1.982	1.969	1.987	1.979	1.986	1.985	1.995	1.988	1.990	1.990
Ti	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Al	0.023	0.026	0.026	0.024	0.027	0.044	0.035	0.035	0.035	0.033	0.033	0.036	0.036
Fe	0.226	0.227	0.227	0.224	0.228	0.222	0.218	0.221	0.215	0.143	0.147	0.148	0.148
Mn	0.011	0.008	0.010	0.010	0.011	0.006	0.005	0.004	0.007	0.003	0.002	0.003	0.003
Mg	1.722	1.733	1.733	1.755	1.774	1.720	1.759	1.745	1.746	1.803	1.815	1.804	1.804
Ca	0.009	0.006	0.006	0.006	0.006	0.007	0.003	0.003	0.006	0.008	0.009	0.007	0.007
Na	0.014	0.000	0.001	0.001	0.001	0.001	0.003	0.001	0.002	0.002	0.001	0.002	0.002
Cr	0.000	0.002	0.001	0.002	0.001	3.990	4.003	3.995	3.997	3.988	3.995	3.991	3.991
Total	4.000	3.994	4.005	4.005	4.017	88.56	88.95	88.77	89.04	92.64	92.53	92.40	92.40
Mg#	88.42	88.42	88.34	88.67	88.62	88.56	88.95	88.77	89.04	92.64	92.53	92.40	92.40

Olivine

	Dunite				Impure chromitite				Pyroxenite				
SiO ₂	41.61	41.47	41.43	41.07	41.40	41.97	42.10	42.39	40.75	40.65	40.32		
FeO	7.28	6.41	8.98	8.58	8.72	3.20	3.19	3.03	10.12	10.21	10.08		
MnO	0.11	0.00	0.12	0.11	0.11	0.04	0.04	0.05	0.24	0.29	0.19		
MgO	51.05	51.99	50.06	49.67	49.51	54.10	53.78	54.35	48.47	49.57	47.96		
CaO	0.01	0.01	0.00	0.01	0.07	0.01	0.00	0.01	0.03	0.00	0.04		
NiO	0.54	0.48	0.39	0.36	0.47	0.86	0.83	0.86	0.20	0.17	0.15		
Total	100.60	100.36	100.98	99.80	100.28	100.18	99.94	100.69	99.81	100.89	98.74		
Cations to 4 oxygens													
Si	1.003	0.998	1.006	1.003	1.007	0.999	1.004	1.003	1.002	0.991	1.002		
Fe	0.147	0.129	0.182	0.175	0.177	0.064	0.064	0.060	0.208	0.208	0.210		
Mn	0.002	0.000	0.003	0.002	0.002	0.001	0.001	0.001	0.005	0.006	0.004		
Mg	1.834	1.865	1.827	1.809	1.795	1.820	1.912	1.917	1.777	1.801	1.777		
Ca	0.000	0.000	0.000	0.000	0.002	0.000	0.000	0.000	0.001	0.000	0.001		
Ni	0.011	0.009	0.008	0.007	0.009	0.017	0.016	0.016	0.004	0.003	0.003		
Total	2.997	3.002	2.994	2.996	2.993	3.001	2.996	2.997	2.997	3.009	2.997		
Mg#	92.60	93.53	90.86	91.17	91.01	96.79	96.78	96.97	89.52	89.65	89.46		

Analyses were made by a Cambridge instruments, Microscan 9 electron microprobe, at the University of Oxford; operating conditions were 20 kV accelerating voltage, a specimen current of ca. 4 × 10⁻⁸ amps, spot size of 2–3 microns, and a counting of 10–30 s; silicate and oxide standards were used and a standard ZAF correction applied

of the contacts within the ultramafic bodies are tectonic. There is some dispute over the significance of the pyroxenites, for the field relationships suggest that they may represent a reaction zone between the serpentized dunite and its host. Alternatively, the presence of dunitic pods within the pyroxenites suggests that the two are integral to the layered intrusion. This issue will be addressed later in the paper.

Underground exposures show that, in places, primary igneous layering is preserved in the chromitites and dunites. Chromitite-dunite contacts are sharp, with the chromitite above the dunite. The lower part of chromitite layers shows grain size layering with centimeter-wide laminae of fine-grained chromitite in a matrix of coarse-grained massive chromitite. In the upper part of the layer, close to the contact with the overlying dunite, are laminae of dunite within the chromitite. Within chromitite layers there are also bands containing large (1 cm) olivine crystals in a chromitite matrix. These are described below as "impure chromitites." There are also crosscutting veins of dunite in the chromitite and irregular fingerlike structures of dunite penetrating into chromitite producing isolated xenoliths of chromitite, a few centimeters long in a dunitic host. These are thought to have developed due to gravitational instability in the partially solidified magma chamber in which dunite intruded the chromitite as a crystal mush.

Petrography and Mineral Chemistry

Samples were collected from the Inyala orebody 2 on the 350 m level between crosscuts 15N and 35S. Representative mineral compositions are given in Table 1. For chromite compositions, quoted Fe/(Fe + Mg) ratios reflect Fe²⁺/(Fe²⁺ + Mg) ratios. Fe²⁺ and Fe³⁺ were determined from stoichiometry using the equation of Droop (1987). For the coexisting silicate minerals, the Mg number = Mg/(Fe_{total} + Mg).

Chromitites

Massive chromitites are, for the most part, coarse grained (>1 mm) and the chromite grains are uniform in composition. Within individual grains Cr₂O₃ contents vary between 57.0 and 58.0 wt percent, Cr/(Cr + Al) ratios are high (between 0.76–0.77), and Fe/(Fe + Mg) ratios low (between 0.33–0.35).

Impure chromitites

Inhomogeneous impure chromitites contain layers of dunite up to 1 cm wide in a chromitite host. The chromite-rich layers contain individual chromite grains up to 3 mm across, but generally they are much finer grained. Both large and small grains show within the grain variations in the Cr₂O₃ content between 57.5 and 58.8 wt percent, Cr/(Cr + Al) ratios between 0.76 to 0.77, and Fe/(Fe + Mg) ratios between 0.29 and 0.33. The olivine-rich layers contain very large (up to 1 cm) elliptical olivine grains. These grains are rich in inclusion trails. The olivine is very magnesian (Fo₉₇), with a high NiO content (0.8 wt %), and even the very large grains show no zonation. Throughout, in both chromite-rich and olivine-rich bands, the olivine is extensively altered to a mixture of serpentine and dolomite. Phlogopite postdates the serpentinization.

Dunites

The dunites are characterized by coarse-grained olivine with grains typically between 2 and 10 mm in length. Accessory phases include chromite (ca. 1%) and in some samples orthopyroxene (0–5%). Up to 40 percent of the olivine is serpentinized and a small amount of phlogopite overgrows the serpentine. Olivine contains abundant trails of fine non-opaque inclusions (cf. olivine in komatiite described by Renner et al., 1994). Olivine compositions are uniform within individual dunite layers but between layers range from $\text{Fo}_{90.9}$ to $\text{Fo}_{93.5}$. Individual grains are very homogeneous, the greatest variation being adjacent to chromite grains. Where present, orthopyroxene ($\text{Mg}/(\text{Fe}_{\text{total}} + \text{Mg}) = 0.924\text{--}0.926$) forms large (1–4 mm) grains which are randomly distributed and partially serpentinized. Chromite is subhedral and fine grained (0.5 mm), homogeneous within layers, but between layers varies in Cr content from 46.8 to 48.4 wt percent Cr_2O_3 ; $\text{Cr}/(\text{Cr} + \text{Al})$ ratios vary from 0.69 to 0.79 and $\text{Fe}/(\text{Fe} + \text{Mg})$ ratios from 0.54 to 0.58. Phlogopite has an $\text{Mg}/(\text{Fe}_{\text{total}} + \text{Mg})$ of 0.94. The dunites are interpreted as olivine cumulates.

Pyroxenites

The pyroxenites contain the mineral assemblage orthopyroxene, Cr magnetite (magnetite with up to 21 wt % Cr_2O_3) \pm olivine \pm amphibole \pm phlogopite. Some samples are massive, others are banded and contain Cr magnetite-olivine and olivine-orthopyroxene laminae. The proportion of orthopyroxene varies considerably. In the massive pyroxenites there is up to 95 percent orthopyroxene and grains vary in size from 0.5 to 6 mm across. The large grains ($\text{Mg}/(\text{Fe}_{\text{total}} + \text{Mg}) = 0.88$) are irregular in shape and are internally strained; they show exsolution textures and contain abundant small inclusions of an opaque mineral in ill-defined trails. Smaller grains have straight grain boundaries and an interlocking texture. The Cr_2O_3 content of the orthopyroxene is low (0.02–0.04 wt %) and the A_2O_3 content varies from 1.1 to 1.84 wt percent. In a few places the orthopyroxene is altered to a magnesian hornblende ($\text{Mg}/(\text{Fe}_{\text{total}} + \text{Mg}) = 0.89\text{--}0.90$) particularly poor in Al.

In the banded pyroxenites, olivine-orthopyroxene laminae become harzburgitic with ca. 30 percent olivine and ca. 70 percent orthopyroxene. Grains are typically up to 1 mm in length, but larger orthopyroxene grains, up to 6 mm across, are poikilitic and enclose smaller rounded olivine grains. Rounded, irregular olivine grains are also interstitial to orthopyroxene grains. Magnetite-olivine-rich bands contain 65 percent olivine, 30 percent Cr magnetite, and 5 percent orthopyroxene. Olivine grains ($\text{Fo}_{88.2\text{--}89.5}$) normally vary from 0.5 to 2 mm across and are slightly serpentinized. The Cr magnetite varies in grain size from 1.5 mm to the micron scale. The larger grains are irregular in shape and are interstitial to olivine grains; the smaller grains are rounded. Their Cr content varies with grain size, indicative of a diffusion relationship in which the chromites reequilibrated with silicate phases during cooling and in which there is a relationship between blocking temperature and grain size. They vary in their chrome content from 12.3 to 20.9 wt percent Cr_2O_3 but have $\text{Cr}/(\text{Cr} + \text{Al})$ ratios similar those of the chromites.

The banded nature of the orthopyroxenites is interpreted

as a relict igneous banding. This observation is supported by the textural relationships between orthopyroxene, olivine, and Cr magnetite. The differences in grain size, the observed grain shape, and the presence of a poikilitic texture all suggest that the orthopyroxenites preserve relict igneous textures. This was, in part, modified during later deformation and granulite facies recrystallization as is evidenced by some fine-grained, granular orthopyroxene surrounding larger (igneous) grains. These observations indicate that the orthopyroxenites are primary igneous cumulates and not the product of the later granulite facies metamorphism. As such they form an integral part of the layered dunite-chromitite succession.

Petrogenesis

Chromitites

Chromite compositions are plotted on a $\text{Cr}/(\text{Cr} + \text{Al})$ vs. $\text{Fe}^{2+}/(\text{Fe}^{2+} + \text{Mg})$ projection of the spinel prism (Fig. 2a). Mass balance arguments dictate that, in rocks where the dominant mineral is chromite, original igneous compositions are preserved because there are few other phases available for element exchange (Eales and Reynolds, 1986). Chromites in chromitites have high $\text{Cr}/(\text{Cr} + \text{Al})$ and low $\text{Fe}^{2+}/(\text{Fe}^{2+} + \text{Mg})$. Chromite in impure chromitites coexists with a small amount of olivine (Fo_{97}). The highly magnesian nature of the olivine suggests that there has been Fe-Mg exchange between olivine and chromite and that the chromite may have originally been more magnesian than its present composition.

The measured compositions of Inyala massive and impure chromites plot in Figure 2b on the same trend as skeletal chromite phenocrysts in komatiites from the 2.7 Ga Reliance Formation of the Belingwe greenstone belt (data from Scholey, 1992; Zhou and Kerrich, 1992; and Renner et al., 1994; recalculated using the charge balance equation of Droop, 1987). The Inyala and Belingwe chromites are also similar in their Cr_2O_3 contents which vary between 57.0 to 58.7 wt percent at Inyala and between 51.4 to 58.9 wt percent at Belingwe. This comparison is instructive because the Belingwe komatiites are the freshest and best-documented komatiites in southern Africa and offer the best opportunity to study primary magmatic mineral compositions. Rollinson (1995a) showed that chromites in komatiites tend to have a distinctive trend on a $\text{Cr}/(\text{Cr} + \text{Al})$ vs. $\text{Fe}/(\text{Fe} + \text{Mg})$ diagram and can be distinguished from chromites in other ultramafic rocks. This is apparent in Figure 2c, where the Inyala chromitites are compared with seam chromitites of komatiitic origin from Shurugwi, Zimbabwe. There is a close correspondence between the Inyala and Shurugwi chromitite compositions confirming the likely komatiitic parentage of the Inyala chromitites. In contrast, chromitites from the Bushveld intrusion (data from the Upper Critical Zone, Eales and Reynolds, 1986) plot in a very different field. It was also shown by Rollinson (1995a) that, in komatiitic chromites, the most magnesian compositions are likely to approximate the original liquidus compositions and that the iron-enrichment trend is due to reequilibration with the melt on cooling. Thus, Figure 2b demonstrates that chromites from Inyala have compositions which are close to the original liquidus compositions of the chromites from Belingwe and could approximate the primary magmatic chromite compositions.

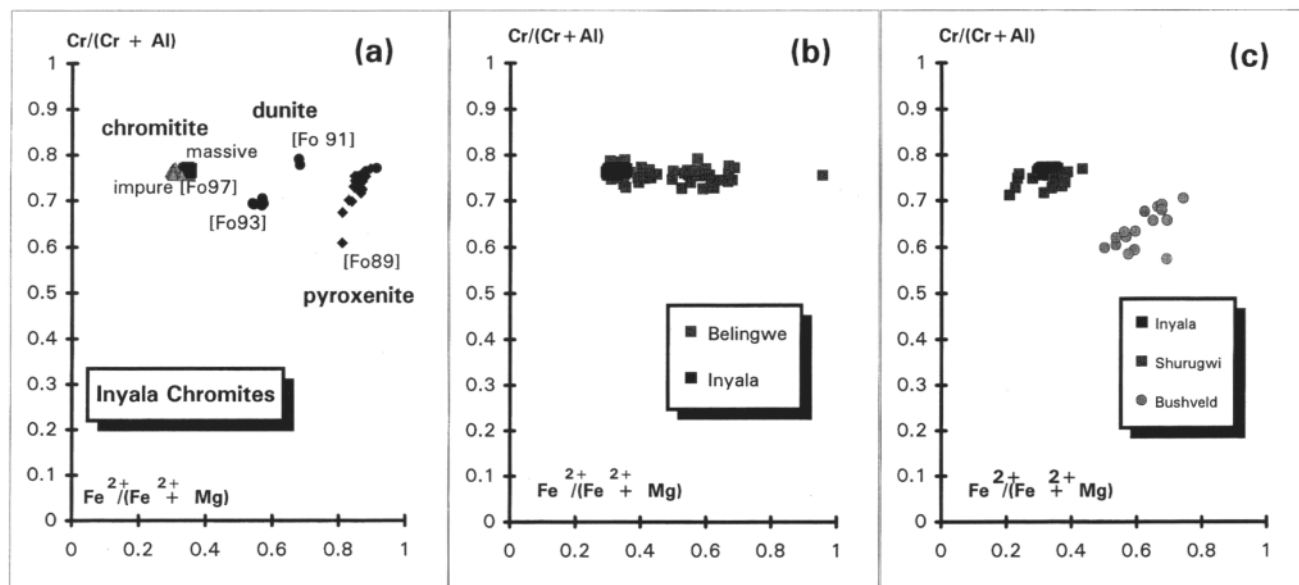


FIG. 2. (a). Chromite compositions at Inyala: massive chromitites (black squares), impure chromitites (gray triangles), chromite in dunite layers (black circles), Cr magnetite in pyroxenite layers (black diamonds). The compositions of the coexisting olivines are shown in brackets. (b). A comparison between chromite compositions at Inyala (black squares) and those in komatiite lavas from the Belingwe greenstone belt, Zimbabwe (gray squares; data from Scholey, 1992; Zhou and Kerrich, 1992). (c). A comparison between chromite compositions at Inyala (black squares) and seam chromitites from komatiites at Shurugwi, Zimbabwe (gray squares; data from Cotterill 1979; Fernandes, 1987; Stowe, 1987b), and from the Bushveld intrusion, South Africa (gray circles; data from the Upper Critical Zone, Eales and Reynolds, 1986).

Dunites

A plot of Fo vs. NiO for olivine from the Inyala chromitites, dunites, and pyroxenites is given in Figure 3. From mass balance considerations olivines in the purer dunites are thought to have retained their original compositions whereas this is not the case for olivine in the chromitites and pyroxenites. There is a strong correlation between Fo and the NiO content of the olivine. Figure 3 also shows the Fo-NiO trend for the Belingwe komatiites (Renner et al., 1994). The two trends are slightly different in slope and in extent but the following observations can be made:

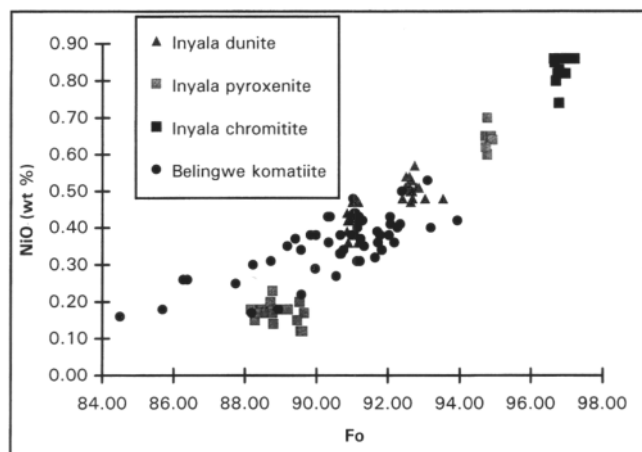


FIG. 3. Plot of Fo (fosterite) content vs. wt percent NiO in olivines from the Inyala mine and in olivines from the komatiites of the Belingwe greenstone belt (data from Renner et al., 1994).

1. There is a significant overlap in compositional range between olivine from the Belingwe komatiites and the Inyala ultramafic rocks. The most iron-rich olivines on the Belingwe trend represent rim compositions and reflect down-temperature reequilibration between olivine and melt.

2. The olivine of one Inyala dunite in the range $\text{Fo}_{92.6-93.5}$ ($\text{NiO} = 0.47-0.54$ wt %) is very similar in composition to the most magnesian olivine in the Belingwe komatiites (Renner et al., 1994). These Belingwe olivines form comparatively rare early phenocrysts (core compositions: $\text{Fo}_{92.3-93.9}$, $\text{NiO} = 0.40-0.53$ wt %). Such phenocrysts could have separated from the melt early in its crystallization history to form olivine-rich cumulates. Nisbet et al. (1993) have argued that these olivine phenocrysts crystallized in equilibrium with a komatiite with ca. 26 wt percent MgO. Such a liquid had a liquidus temperature of 1,520°C.

3. A second Inyala dunite contains olivine in the compositional range $\text{Fo}_{90.9-91.2}$ ($\text{NiO} = 0.39-0.47$ wt %) which is very similar in composition to the ubiquitous microphenocrysts in the Belingwe komatiites with core compositions in the range $\text{Fo}_{91-92.4}$ ($\text{NiO} = 0.34-0.48$ wt %). At Belingwe, olivine with the composition Fo_{91} is derived from a komatiite magma with ca. 19 wt percent MgO (Nisbet et al., 1993).

4. At Inyala, the most magnesian olivine has a composition of Fo_{97} . This occurs as isolated grains in impure chromitite. These olivines are considerably more magnesian than those in komatiite magmas suggesting that they are the result of reequilibration with chromite. Results presented below suggest that this took place during the cooling of the melt.

These observations suggest that the Inyala dunites are derived

from several different komatiitic magmas with compositions in the range 19 to 26 wt percent MgO.

Olivine-chromite thermometry

The coexistence of olivine and spinel permits the calculation of equilibration temperatures based on Fe^{2+} -Mg and Cr-Al- Fe^{3+} exchange reactions (Sack and Ghiorso, 1991). Olivine-spinel pairs in dunites yield temperatures in the range 610° to 680°C and in the chromitites between 490° to 640°C. Olivine-Cr magnetite pairs in the pyroxenites equilibrated between 620° to 750°C. If the original (igneous) chromite compositions from the chromitites are combined with original olivine compositions from the dunites ($F_{\text{O}_{91}}$), then temperatures of between 1,100 to 1,200°C are indicated. These temperatures are in the same range as those calculated by Zhou and Kerrich (1992) for the Belingwe komatiites, although the application of the new chromite Ni thermometer (Griffin et al., 1994) suggests temperatures in the range 1,100 to 1,400°C. An experimental study by Kinzler and Grove (1985) of a komatiite with 17.2 wt percent MgO at 1 atm crystallized olivine on the liquidus at ca. 1,360°C and spinel at ca. 1,330°C. Thus the calculated temperatures for the Inyala samples are typical komatiite melt temperatures, although they are probably below liquidus temperatures. The significance of the lower equilibration temperatures is uncertain and could reflect either the rapid cooling of the komatiite or reequilibration during the granulite facies metamorphism.

Diffusion in olivines

The olivine in this study shows very little evidence of Fe-Mg zoning even though some grains are up to 10 mm in length, suggesting that these grains are well equilibrated. What is not clear is whether this took place during the igneous or the metamorphic stage of their evolution. Since there is a large difference in cooling rate between the two events, the blocking temperature equation is used here to estimate the cooling rate of the dunites. For the dunites the blocking temperature for Fe^{2+} -Mg exchange between olivine and spinel is known and so the blocking temperature equation can be inverted to calculate the cooling rate. The diffusion data for olivine are taken from Jaoul et al. (1995), a blocking temperature of 700°C is assumed for Fe^{2+} -Mg exchange and the grain radius is estimated from petrography. An "A" value (a function of grain shape) of 55 was adopted for olivine.

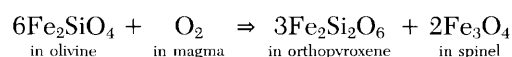
Diffusion calculations for the dunites show that olivine grains with a radius of 2 mm suggest cooling rates of 2,200°C/Ma, whereas grains 1 mm in radius suggest cooling rates of 8,900°C/Ma. While the measured cooling rates are clearly imprecise and may contain an element of both igneous and metamorphic processes, the high values indicate that diffusion took place predominantly in a rapid cooling igneous regime rather than in the slower cooling environment of metamorphic belts (3–30°C/Ma). This result further confirms the view that these rocks preserve original igneous olivine and chromite compositions, despite the later granulite facies metamorphism.

Pyroxenites

Orthopyroxene in komatiite lavas is not commonly reported in the literature. Chemically unaltered komatiites at Belin-

gwe, Zimbabwe, and komatiites from the Alexo flow, Ontario, contain rare pigeonite mantled with augite in the groundmass (Arndt and Fleet, 1979; Renner et al., 1994). One-atm experimental studies have also reported pigeonite in a synthetic komatiitic melt with 17 wt percent MgO (Kinzler and Grove, 1985). Bronzite and enstatite-rich cumulates of possible komatiitic origin are described from layered intrusions in the Barberton greenstone belt, South Africa (de Wit et al., 1987), but these are not well studied. Arndt and Fleet (1979) report a bronzite-rich cumulate in the Alexo komatiite flow, Ontario. This 10-m thick flow contains a thin orthopyroxene-rich band between an olivine cumulate layer and an upper zone containing phenocrystal olivine. The bronzite cumulate comprises 70 to 80 percent prismatic bronzite together with olivine and chromite in a fine-grained groundmass. The bronzite has crenulate grain boundaries, thought to be indicative of rapid cooling. The bronzite has an $\text{Mg}/(\text{Fe}_{\text{total}} + \text{Mg}) = 88$ (identical to that of the mineral compositions of pyroxenites described here) and coexists with olivine of the same $\text{Mg}/(\text{Fe}_{\text{total}} + \text{Mg})$. Kinzler and Grove (1985) noted that, in the Alexo flow, the pyroxenes in the groundmass of the komatiite were different from those observed in the cumulate layer. This led them to suggest that the bronzite cumulate layer was produced either by a process of magma mixing, during the differentiation of the flow or by the assimilation of a silicic component into the komatiite magma. They showed that mixing between komatiite and a silica-rich component would shift the melt composition from the field of primary olivine crystallization into the orthopyroxene field.

An alternative mechanism for orthopyroxene crystallization is demonstrated in an experimental study by Barnes (1986) who showed that the chromite-pyroxene peritectic is controlled by oxygen activity in the melt and that a small change in oxygen activity can shift the melt from the field of chromite precipitation to that of orthopyroxene precipitation. The effect of increasing the oxygen activity in an olivine-bearing magma is illustrated by the reaction:



and predicts that an increase in the activity of oxygen will result in both orthopyroxene precipitation and an increase in the magnetite component of spinel. Both these features are observed in the pyroxenites described here and support the hypothesis that crystallization of pyroxene was accompanied by a change in oxygen activity in the melt. A change in oxygen activity could be brought about either by magma mixing or by assimilation as discussed above.

The hypothesis that the orthopyroxene at Inyala was derived from a komatiitic liquid may be further tested by calculating the composition of the assumed parental melt using the distribution coefficient for FeO and MgO between orthopyroxene and the parental liquid (Barnes, 1986):

$$\log \left[\frac{(\text{FeO}/\text{MgO})_{\text{orthopyroxene}}}{(\text{FeO}/\text{MgO})_{\text{melt}}} \right] = 0.2029 - 1150/T(^{\circ}\text{K}).$$

An iterative calculation using the FeO content of the very fresh komatiite samples from the Belingwe greenstone belt, Zimbabwe, reported by Bickle et al. (1993) and orthopyrox-

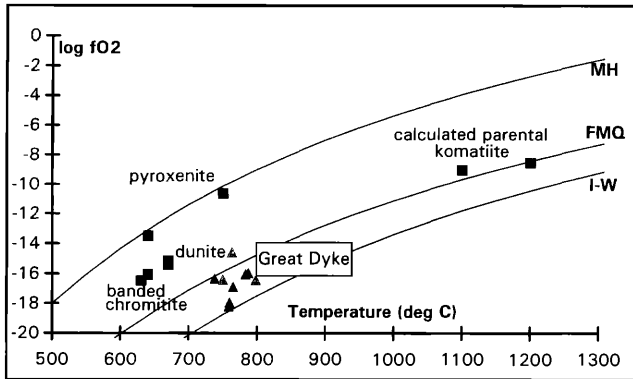


FIG. 4. Oxygen activity vs. temperature plot, showing the calculated compositions for olivine-orthopyroxene-spinel assemblages from the Inyala pyroxenites, dunites, and parental komatiite magma. Also shown are two data points for banded chromitite from the Rhonda mine, a comparable geologic setting, located 8 km to the southwest. The buffer curves for the reaction iron-wustite (IW), quartz-fayalite-magnetite (FMQ), and magnetite-hematite (M-H) are shown for comparison. Olivine from bronzitites (gray triangles) and dunites (black triangles) from the Great Dyke (data from Wilson, 1982) are also shown for comparison.

ene from the pyroxenites with $Mg/(Fe_{total} + Mg) = ca. 0.88$ suggests a parent liquid with 16 wt percent MgO at temperatures of between 1,300° to 1,330°C. These melt compositions are lower than those calculated from olivine compositions (19–26 wt % MgO) but are still consistent with a komatiitic parent liquid. It is important to note, however, that the thickness of the orthopyroxene-rich layers at Inyala is much greater than those reported elsewhere from komatiites.

Oxygen activity calculations

To investigate a possible change in oxygen activity during the differentiation of the komatiite, oxygen activities were calculated for the mineral assemblage olivine-orthopyroxene-spinel using the equation of Ballhaus et al. (1991). Figure 4 shows a plot of oxygen activity versus temperature in the dunites and pyroxenites calculated relative to QFM. The relative oxygen activities are calculated for the equilibration temperatures given above using the olivine-spinel thermometer of Sack and Ghiorso (1991). Two important observations arise from these calculations:

1. There is an important difference in oxygen activity between the dunites and the pyroxenites. Pyroxenites have an oxygen activity 2.5 log units higher than that of dunites and of banded chromitites from the Rhonda mine, located 8 km to the southwest of Inyala (Fig. 1). This suggests that there was a major difference in oxygen activity between the olivine-rich and orthopyroxene-rich compositions. When bronzitites and dunites from the Great Dyke (Wilson, 1982) are studied a similar difference is not observed (see Fig. 4).

2. The oxygen activity for the komatiites was calculated by using the original liquidus olivine and spinel compositions, from the dunites and chromitites, respectively, and the calculated olivine-spinel temperatures. Strictly, the calculated oxygen activity is only valid at the point at which the melt became saturated in orthopyroxene. At 1,200°C this falls very close to the QFM buffer and is in close agreement with that pre-

dicted for the Belingwe komatiites by Nisbet et al. (1993), although the oxygen activity prior to the orthopyroxene saturation cannot be known.

It is important to note that the measured difference in oxygen activity is for the temperature interval 650° to 750°C. Thus it cannot be proven that the measured difference reflects a real difference in oxygen activity between the dunites and pyroxenites during the igneous history of the rock. However, the form of the equation of Ballhaus et al. (1991) means that the difference in calculated oxygen activity is almost independent of temperature. Further, it has already been argued from the rapid cooling of these rocks that original compositions tend to be preserved. What is not known is the extent to which the Cr magnetites in the pyroxenites have changed in Fe^{3+} during cooling. Thus, while it is thought to be likely, it cannot be proved that the measured differences in oxygen activity reflect changes in magma composition.

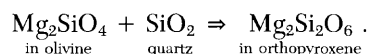
Discussion

The origin of monomineralic chromite layers in layered igneous intrusions is a problem which has taxed petrologists for several decades. In the last ten years opinion has shifted from models based on gravitational setting (Irvine, 1980a) toward models in which in situ crystal growth is favored (Irvine, 1980b). Many recent models favor a process of magma mixing to trigger chromite precipitation. The range of models includes the mixing of primitive and evolved magmas of the same parentage (Irvine, 1977; Wilson, 1982) and the mixing of two distinct magmas (Irvine et al., 1983). In these cases the effect of magma mixing is to replace the crystallization of the silicate phase(s) with the crystallization of chromite. The same effect can also be achieved through the assimilation of silicic country rock by an ultramafic melt (Irvine, 1975). A change in the oxygen activity of the melt is also seen as a means of triggering chromite precipitation (Ulmer, 1969). This latter mechanism could be related to either assimilation or magma mixing.

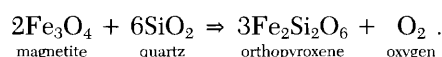
The results presented here show that there are strong similarities in the mineral chemistry of chromites and olivines from Inyala with those found in fresh komatiitic lavas from the Belingwe greenstone belt, Zimbabwe (Fig. 2b). Inyala chromite compositions are also very similar to those of the komatiite-hosted chromitites from the Shurugwi greenstone belt, Zimbabwe (Fig. 2c), although these chromitites are less well analyzed than those from the Belingwe komatiites and their compositions have been partly disturbed by later shearing. These observations suggest that the Inyala chromitites were derived from a komatiitic magma, hence in this study, the origin of monomineralic chromite layers has to be addressed in magmas of a komatiitic composition.

At Inyala the chromitites occur within the dunites, nevertheless the presence of orthopyroxenites may have an indirect bearing on the problem of chromite precipitation from a komatiite magma. It has already been noted that there are very few well-documented examples of orthopyroxenites associated with a komatiitic parent magma. Where orthopyroxenites have been described as part of a komatiite flow, it has been suggested that the orthopyroxene crystallization was triggered by either magma mixing, assimilation of siliceous

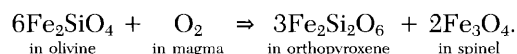
country rock and/or a change in the oxygen activity of the melt. Although both magma mixing and assimilation are theoretically possible (Nisbet and Chinner, 1981; Huppert et al., 1984), the very high melt temperatures estimated for the parent komatiite at Inyala (ca. 1,520°C) suggests that contamination is the more likely possibility. Banded iron-formation is frequently the host to dunites in this region and could be the possible contaminant. The assimilation of banded iron-formation by a komatiitic melt could trigger the precipitation of orthopyroxene through the addition of silica by the reaction:



It is possible that orthopyroxene would also be produced through the reaction:



The increased oxygen activity produced by the above reaction could lead to the further precipitation of orthopyroxene accompanied by a magnetite rich-chromium poor spinel by means of the reaction:



Petrological evidence presented in this paper, namely the coexistence of orthopyroxene and chrome-magnetite in the orthopyroxenites, and the evidence for an increase in oxygen activity in the pyroxenites support the hypothesis that the pyroxenites formed from the assimilation of banded iron-formation by komatiitic magma. Support is also found in the field relationships at Inyala because the orthopyroxenites form, for the most part, at the boundary between the dunites of the main orebody and the country rock. It has been demonstrated that the assimilation of banded iron-formation by a komatiitic melt will trigger orthopyroxene crystallization and will raise the oxygen activity of the melt. These two mechanisms are also regarded as possible causes of chromite precipitation in ultramafic melts. Thus, although it cannot be unequivocally demonstrated for the chromitites, it is proposed from the evidence of the pyroxenites that country-rock assimilation is an important mechanism of chromite precipitation in komatiite melts.

The recognition of the importance of assimilation and changes in oxygen activity in komatiites may explain why some komatiite magmas produce chromite-rich cumulates whereas others produce sulfide-rich melts rich in Ni and the platinum-group elements. The difference in patterns of mineralization may lie in the differing compositions of assimilated material and the resultant changes in oxygen and sulfur activities in the melt.

Conclusions

1. The chromitites at the Inyala mine form part of an cumulate sequence which was derived from a komatiitic parent liquid.
2. The komatiitic liquids varied in composition from 26 to 16 wt percent MgO.
3. The most magnesian komatiites would have had a liq-

uidus temperature of ca. 1,520°C and had the capacity to assimilate the country rock.

4. Pyroxenites (bronzite cumulates) formed in response to assimilation of adjacent supracrustal rocks. There is some indication that the mixing gave rise to an increase in oxygen activity in the melt.

5. Olivine-chromite mineral pairs continued to equilibrate down-temperature until about 680°C. Diffusion of Fe and Mg in olivine suggests that the cooling was rapid and that igneous compositions and blocking temperatures are preserved.

Acknowledgments

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