CHAPTER IV

CHROMITE MINERALISATION
Chapter IV

Chromite Mineralisation

The chapter deals with field characters, structures and textures, ore petrography and mineral chemistry of chromites associated with the mafic-ultramafic rocks. On the basis of these characteristics, the chromites have been classified.

The chromite ores are confined to the Lower Zone which consists of dunite-peridotite association (olivine-chromite cumulate and olivine-pyroxene-chromite). Mineralisation is well exposed at two localities namely Nanus-NW of Bondla and at Bondla, mainly between Bondla and Poikul. Chromite at Nanus occurs as thin layers 1-3 cm thick which alternate with layers of dunite. The latter vary in thickness from 4-10 cm. The chromite layers strike N40° and dip 30°-35° towards N130°. The peridotites are sheared and fractured. The fractures trend (i) N65° and dip 50° towards N335° and (ii) N110° and dip 25°towards N350°. Along these fractures, the layered chromites are remobilised and the layers appear to form a
criss-cross pattern of veins. The original layering is still discernible at places. Along the shear fractures fibrous actinolite and chlorite have developed perpendicular to the layering. These veins vary in width from 2.0 cm to 1.0 m.

The chromites at Bondla can be traced intermittently over a distance of about one kilometre. In the lower part of the peridotites they occur as thin layers 1 cm to approximately 10 cm thick. Such layers alternate with thicker layers of peridotite. The chromite layers invariably exhibit a sharp lower contact with a gradational top. Individual layers at places pinch out along the strike. At places, the layers bifurcate, giving the appearance of veins. Such a pattern of veins could be the result of 'unconformities' formed during deposition (Wager, 1968). Stratigraphically higher in the section, the chromite layers become thicker. In the vicinity of Poikul, a massive chromitite layer is exposed which extends over an outcrop width of more than 300 m. The layer is of chromitite and consists of 75% fine grained chromite with about 25% silicates. The chromitite grades into a chromite-olivine cumulate with progressive increase
of olivine. In some chromite-olivine layers the chromite to silicate ratio is low, thus making the chromite appear as dissemination in silicates.

**Ore Petrography**

The section on ore petrography describes the structures and textures exhibited by chromites in hand specimen and in thin- and polished- sections.

**Structures of Ores**

The term structure is here meant to include the forms and shapes assumed by ore minerals and their aggregates as seen by unaided eye in hand specimen.

Some of the important structures observed include layered and disseminated structure, occluded-silicate structure, chromite-net (reticulate) structure, compact and massive ores and schlieren of chromite.
Layered and Disseminated Structure

The term 'layer' is here used in preference to 'band' because the former has a genetic connotation. The layered structure is made up of alternating dark and light coloured layers. The dark layers consist of chromite. These alternate with light coloured layers made up of serpentinised olivine. The layers are repetitive in nature. Individual chromite layers vary in thickness from 1 cm to as much as 3 cm. Invariably, the olivine-rich layers are many times thicker than the chromite layers. At places, the chromite layers pinch out laterally. Individual layers may be composed entirely of chromite giving rise to massive chromite layers. Invariably, however, little interstitial silicate is always present. The chromite layers exhibit a sharp lower contact and a gradational upper contact with the olivine-rich layers. At places, chromite is interstitial in olivine-rich layers giving rise to disseminated structure.

Occluded Silicate Structure

The structure is exhibited by massive ore layers
which contain little silicates in them. It is observed that olivine either singly or as an aggregate of crystals is enclosed within chromite forming spherical, rounded or elliptical islands of olivine in chromite. At places, olivine is elongated forming lenticles and streaks within chromite.

**Chromite-Net (Reticulate) Structure**

The structure is generally shown by chromite layers which contain considerable proportion of interstitial silicates. In this case subhedral grains or grain aggregates of chromite have interstitial olivine so that chromite forms a network with olivine grains in between (Plate X Photo 1).

**Massive Structure**

Massive structure is shown by thicker layers of chromite. The layer in this case is made up of fine grained chromite with very little interstitial silicates. Majority of the chromite grains are subhedral and tightly packed.
Schlieren of Chromite

The thin veinlets of chromite 1-2 cm thick are seen at places at an angle to the general layering of the rock. Some of them are a couple of cm long with tapering ends. Such ore forms are included under streaks and are also referred to as schlieren.

Chromite Petrography

Chromite has been studied both in transmitted light and incident light. In microsections the chromite from chromitite is typically dark brown in colour. Individual grains are subhedral to anhedral. The grains join each other and form aggregates with irregular, cuneiform interstices in between filled by serpentinised olivine (Plate X Photo 2) forming myrmekitic intergrowth with gangue. Chromites from peridotite are dark brown in colour, typically subhedral and join each other forming chains of chromite.

Under incident light, the chromite is greyish-white in colour and has pitted surface (Plate XI Photo 1).
The reflectivity of a grain in general is moderate. Between cross nicols it is faintly anisotropic. However, careful examination shows that the borders of grains have higher reflectivity than the central part. Such high reflecting borders are made up of ferrit-chromit (Spangenberg, 1943). It is the result of alteration of chromite during which magnetite replaces the former. Well developed grains of chromite show exsolutions of ilmenite parallel to (111) (Plate XI Photo 2). Similar exsolutions have been described from certain Fe-rich chromites from Bushveld. Exsolution bodies of ilmenites in Fe-poor chromites have also been described from the Bastetal and from the Monche Tundra, Cola (Ramdohr, 1980). In oil, the chromite shows brown-internal reflection. At places, especially along thin layers and veins, it is replaced by fibrous chlorite which has developed perpendicular to the layer or vein in question (Plate XII Photo 1).

Textures of Chromites

Under textures, are described, the inter-relationship between chromite and silicate minerals as seen under the microscope. Some of the important textures observed
are layered texture, micro-reticulate texture, chain texture, occluded silicate texture, disseminated texture and myrmekitic texture.

**Layered Texture**

The most commonly observed texture is layered texture. It is made up of alternate layers of chromite and serpentinised olivine. The chromites are subhedral and are arranged into layers that alternate with olivine rich layers. In chromite layers serpentinised olivine occurs interstitially.

**Micro-reticulate Texture**

The texture is commonly exhibited by layered chromites. In this case subhedral to rounded grains of chromite form a framework. The interstices between chromites are occupied by olivine giving the appearance of a chromite network. At places, rounded, elliptical or irregular islands of serpentinised olivine are found within a ring of chromite grains (Plate XII Photo 2). Such a texture is referred to as ring texture (Chang Chung, 1969).
Chain Texture

The texture is common in layered chromites and chromite veins. Polygonal crystals of chromite join each other to form continuous chains.

Occluded Silicate Texture

The texture is exhibited both by massive chromites and layered chromites. It is made up of rounded, elliptical or irregular grains of silicates that are poikilitically enclosed within chromite (Plate XII Photo 2).

Disseminated Texture

The texture is exhibited by chromites from olivine-rich layers. Euhedral to subhedral layers of chromite occur in between olivine crystals.

Myrmekitic Texture

The texture is typically observed in massive chromite ores. Elongated or cuniform silicates are surround-
ed by anhedral aggregates of chromites. Such a texture is described as myrmekitic texture by Ramdohr (1980). The silicates are poikilitically enclosed in a cumulate of chromite forming a heteradccumulate.

Mineral Chemistry

Mineral chemistry of chromite has been investigated with an electronprobe microanalyser. Representative polished thin sections of chromite from chromitite, peridotite and troctolite were analysed for major element composition. Three samples of chromite bearing rocks were selected for study. A minimum of three spot analyses were carried out per grain and average composition was determined. At least four grains were analysed in each sample to know the variation amongst the grains. The chromites were analysed from core to rim to know the variation in composition within a grain. In all fourteen micro-chemical analyses are presented in Table No. 4.1. Total iron was determined as FeO. Fe₂O₃ was calculated from total FeO on the basis of spinel stoichiometry. The major element oxide percentages were recalculated to cations per formula unit for 24 cations.
**Table No. 4.1 Electronprobe Microanalysis of Chromites from Bondla**

<table>
<thead>
<tr>
<th>Oxides</th>
<th>B19A</th>
<th>B19A</th>
<th>B19A</th>
<th>B19A</th>
<th>B15</th>
<th>B15</th>
<th>B15</th>
<th>B15</th>
<th>B15</th>
<th>BT1</th>
<th>BT1</th>
<th>BT1</th>
<th>BT1</th>
<th>BT1</th>
</tr>
</thead>
<tbody>
<tr>
<td>TiO₂</td>
<td>0.72</td>
<td>0.72</td>
<td>0.75</td>
<td>0.78</td>
<td>0.81</td>
<td>0.51</td>
<td>0.70</td>
<td>0.72</td>
<td>0.71</td>
<td>0.70</td>
<td>3.35</td>
<td>3.41</td>
<td>3.37</td>
<td>3.45</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>17.02</td>
<td>17.30</td>
<td>17.15</td>
<td>17.17</td>
<td>7.63</td>
<td>1.58</td>
<td>7.97</td>
<td>1.36</td>
<td>7.01</td>
<td>1.38</td>
<td>13.26</td>
<td>13.12</td>
<td>13.25</td>
<td>12.87</td>
</tr>
<tr>
<td>Cr₂O₃</td>
<td>45.17</td>
<td>45.26</td>
<td>45.19</td>
<td>45.19</td>
<td>51.65</td>
<td>54.92</td>
<td>51.25</td>
<td>52.05</td>
<td>51.49</td>
<td>54.25</td>
<td>41.10</td>
<td>38.16</td>
<td>39.59</td>
<td>38.01</td>
</tr>
<tr>
<td>FeO</td>
<td>17.32</td>
<td>17.35</td>
<td>17.12</td>
<td>17.46</td>
<td>27.10</td>
<td>27.09</td>
<td>27.33</td>
<td>27.20</td>
<td>28.21</td>
<td>27.45</td>
<td>20.10</td>
<td>19.77</td>
<td>19.20</td>
<td>20.74</td>
</tr>
<tr>
<td>MgO</td>
<td>10.41</td>
<td>10.51</td>
<td>10.46</td>
<td>10.52</td>
<td>1.99</td>
<td>1.43</td>
<td>1.97</td>
<td>1.45</td>
<td>1.85</td>
<td>1.09</td>
<td>7.25</td>
<td>7.21</td>
<td>7.89</td>
<td>6.66</td>
</tr>
<tr>
<td>MnO</td>
<td>0.33</td>
<td>0.28</td>
<td>0.30</td>
<td>0.30</td>
<td>1.78</td>
<td>1.98</td>
<td>1.75</td>
<td>1.91</td>
<td>1.70</td>
<td>2.06</td>
<td>0.51</td>
<td>0.50</td>
<td>0.47</td>
<td>0.44</td>
</tr>
</tbody>
</table>

Cations recalculated on the basis of 32 oxygens

<table>
<thead>
<tr>
<th>Ti</th>
<th>0.1367</th>
<th>0.1365</th>
<th>0.1426</th>
<th>0.1489</th>
<th>0.1715</th>
<th>0.1109</th>
<th>0.1476</th>
<th>0.1571</th>
<th>0.1504</th>
<th>0.1536</th>
<th>0.6618</th>
<th>0.6861</th>
<th>0.6627</th>
<th>0.6852</th>
</tr>
</thead>
<tbody>
<tr>
<td>Al</td>
<td>5.1176</td>
<td>5.1931</td>
<td>5.1489</td>
<td>5.1595</td>
<td>2.534</td>
<td>0.5444</td>
<td>2.6444</td>
<td>0.4696</td>
<td>2.3444</td>
<td>0.4769</td>
<td>4.1017</td>
<td>4.144</td>
<td>4.0901</td>
<td>4.0127</td>
</tr>
<tr>
<td>Mg</td>
<td>3.9665</td>
<td>3.9992</td>
<td>3.9798</td>
<td>4.0063</td>
<td>0.8373</td>
<td>0.6235</td>
<td>0.8291</td>
<td>0.6338</td>
<td>0.7831</td>
<td>0.4769</td>
<td>2.847</td>
<td>2.8864</td>
<td>3.0865</td>
<td>2.631</td>
</tr>
<tr>
<td>Mn</td>
<td>0.0706</td>
<td>0.0597</td>
<td>0.0644</td>
<td>0.0644</td>
<td>0.4280</td>
<td>0.4914</td>
<td>0.4187</td>
<td>0.4572</td>
<td>0.4086</td>
<td>0.5122</td>
<td>0.1124</td>
<td>0.1130</td>
<td>0.1041</td>
<td>0.0988</td>
</tr>
<tr>
<td>Cr #</td>
<td>0.64</td>
<td>0.637</td>
<td>0.639</td>
<td>0.638</td>
<td>0.819</td>
<td>0.958</td>
<td>0.812</td>
<td>0.962</td>
<td>0.831</td>
<td>0.963</td>
<td>0.676</td>
<td>0.69</td>
<td>0.667</td>
<td>0.666</td>
</tr>
<tr>
<td>Fe³⁺ #</td>
<td>0.108</td>
<td>0.102</td>
<td>0.107</td>
<td>0.103</td>
<td>0.119</td>
<td>0.167</td>
<td>0.119</td>
<td>0.212</td>
<td>0.121</td>
<td>0.18</td>
<td>0.183</td>
<td>0.225</td>
<td>0.206</td>
<td>0.228</td>
</tr>
<tr>
<td>Al #</td>
<td>0.320</td>
<td>0.325</td>
<td>0.321</td>
<td>0.323</td>
<td>0.158</td>
<td>0.034</td>
<td>0.165</td>
<td>0.029</td>
<td>0.147</td>
<td>0.0298</td>
<td>0.264</td>
<td>0.27</td>
<td>0.263</td>
<td>0.258</td>
</tr>
<tr>
<td>Mg #</td>
<td>0.517</td>
<td>0.519</td>
<td>0.521</td>
<td>0.517</td>
<td>0.115</td>
<td>0.085</td>
<td>0.113</td>
<td>0.086</td>
<td>0.104</td>
<td>0.065</td>
<td>0.391</td>
<td>0.393</td>
<td>0.422</td>
<td>0.363</td>
</tr>
</tbody>
</table>

B19A - Peridotite  B15 - Chromitite  BT1 - Troctolite
From the microprobe chemical data (Table No. 4.1) it is seen that majority of the grains show chemical zoning. There is an increase in Cr$_2$O$_3$ from core to rim as seen from analysis 5-10 (Table No. 4.1). The Cr # (Cr/Cr+Al) varies between 0.81 to 0.83 for the core and it is between 0.95 and 0.96 for the rims. Fe$^{+3}$ 
$[\text{Fe}^{+3}/\text{Cr+Al+Fe}^{+3}]$ also shows an increase from core to rim whereas, Mg # $[\text{Mg/Mg+Fe}^{+2}]$ shows a progressive decrease rimward. This compositional variation can be attributed to secondary alteration and development of ferrit-chromit (Spangenberg, 1943) along the borders of chromite crystals. The chromites from the chromitites in general (chromite cumulate - B15) are very rich in Cr$_2$O$_3$ and have high Cr # $[\text{Cr}/(\text{Cr+Al})]$. Cr$_2$O$_3$ varies between 54-51 wt.% and Cr # between 0.81 and 0.96. Cr$_2$O$_3$ content (45 wt.%), and Cr # (0.63-0.64) decrease in peridotite chromite (olivine-chromite cumulate - B19A). Cr$_2$O$_3$ decreases further to about 39 wt.% with Cr # 0.66 in disseminated chromite from troctolite (BT1). Thus overall there is a progressive decrease in Cr$_2$O$_3$ from chromitite chromite (B15) to peridotite chromite (B19A) to troctolite chromite (BT1). It may be noted though that the composition of accessory chromites (from BT1) may show considerable
variation due to sub-solidus re-equilibration with associated phases (Irvine, 1967; Dickey and Yoder, 1972) and due to alteration (Jan et al., 1984; 1985). The accessory chromites are impoverished in Mg # \[\text{Mg}/(\text{Mg}+\text{Fe}^2+)\] and Cr # compared to unaltered chromites, possibly due to substantial Mg-Fe exchange with silicates-olivines and pyroxenes during sub-solidus re-equilibration. The altered chromites generally show enrichment in Cr and \(\text{Fe}^2+\) compared to unaltered parts of the grain. \(\text{Cr}_2\text{O}_3\) content varies with modal mineralogy. Higher wt.% \(\text{Cr}_2\text{O}_3\) is observed with increasing modal percentage of chromite. Such an increase is common (Cameron, 1975; Wilson, 1982).

\(\text{Cr}_2\text{O}_3\) and \(\text{Al}_2\text{O}_3\) show an inverse relationship. Total Fe (as FeO) in chromites is very high varying between 24 and 40 wt.%. \(\text{Cr}_2\text{O}_3\) exhibits a negative correlation with FeO_(t) within samples specially of chromitites. The latter is almost constant within a sample whereas, amongst samples it shows considerable variation between 24 and 40 wt.%. A typical feature of chromites from chromitite (B15) is their high \(\text{Cr}_2\text{O}_3\) which is more than 51 wt.% and more than 35 wt.% FeO_(t). Whereas, in peridotite chromites (B19A) \(\text{Cr}_2\text{O}_3\) decreases to about 45 wt.% and FeO
to 25 wt.%. This behaviour of Cr₂O₃ and FeO is characteristic of layered chromites.

In general TiO₂ in all samples is more than 0.5 wt.% - a feature typical of stratiform chromites. Dickey (1975) suggests 0.3 wt.% TiO₂ as the dividing line between podiform and stratiform chromites. The TiO₂ in accessory chromites (BT1) in particular is much higher, more than 3.0 wt.%. Herbert (1982) showed that in cumulate rocks, TiO₂ varies from 0.1 to 2.0 wt.% This is so because of geochemical behaviour of Ti⁺⁴ which is enriched in the melt relative to solid phases during crystallization (Burns, 1973). Cameron (1979) reported 0.3-2.5 wt.% TiO₂ in chromites from Eastern Bushveld. Wilson (1982) noted 0.23-2.24 wt.% TiO₂ in the chromites from Great Dyke. Chromites from other layered - and SE Alaskan-type complexes generally contain more than 0.3 wt.% TiO₂ (Dickey, 1975; Bird and Clark, 1976). Evans and Wright (1972) and Hill and Roeder (1974) showed that in more fractionated basalts the chrome-spinels are richer in TiO₂. TiO₂ from Bondla chromites do not show any significant correlation with any of the major elements (Thayer 1964; Dickey, 1975; Greenbaum; 1977).
All chromite analyses contains very low wt.% MgO unlike podiform chromites. Similarly they contain higher wt.% Al₂O₃ over MgO unlike the podiform chromites which show a reverse relationship. The (Cr+Al) variation in chromites is very small (11.5-14.5). The Cr/Al ratio however, shows a wide variation from 1.5 to 25 which according to Dickey (1975) is the result of either bulk chemistry or temperature of crystallisation. A positive correlation is observed between Cr/(Cr+ Al) and Fe+2/(Mg +Fe+2). The Mg # shows a much broader range. It varies from 0.06 (in chromitite chromite - B15) to 0.52 (in peridotite chromite - B19A). The variation in the oxidation ratio [2Fe₂O₃/(2Fe₂O₃+FeO)] is very small (0.39-1.11). The Fe+2/Mg ratios are higher and have a broad range from 0.9 to 14 (in comparison with podiform chromites) which according to Dickey (1975) is indicative of fractional crystallization.

The MnO content of the chromites show a positive correlation with FeO. It is lower in peridotite chromites (about 0.3 wt%) and it increases in chromitite chromites (more than 1.7 wt %). The chromites from chromitite and peridotite are more ferriferous and contain more than 16...
wt.% total iron (FeO) when Cr$_2$O$_3$ percentage is about 45 wt.% - a feature typically observed in some of the stratiform chromites from Bushveld complex. Various mechanisms have been suggested for this variation: (i) reaction with the magma (Henderson, 1975), (ii) post cumulus changes under variable oxygen fugacity (Cameron, 1975), (iii) exchange of Mg-Fe with the silicate minerals during subsolidus re-equilibration (Irvine, 1967), (iv) alteration attendant to changing fO$_2$ and fH$_2$O during chloritization and serpentinization (Ahmed and Hall, 1981; Jan et al., 1985) and (v) increase in Fe$^{+2}$ during metamorphism (Evans and Frost, 1974; Bliss and MacLean, 1975; Hoffman and Walker, 1978).

Cr # exhibits a reciprocal behaviour with Al # [$\frac{Al}{(Al+Cr+Fe^+3)}$] implying that Cr is preferentially incorporated in Al-spinel lattice. A comparison of this relationship with Fe$^{+3}$ # [$\frac{Fe^{+3}}{(Cr+Al+Fe^{+3})}$] indicates that Al is incorporated preferentially into the chromite lattice and that the gradual upward increase in the Fe$^{+3}$ content (increasing Fe$^{+3}$ from peridotite chromite upwards in the section) seems to be at the expense of both Al and Cr. The chromites from the chromitite (B15) and perido-
tite (B19A) have Fe$^{3+}$ # in the range of 0.10–0.18. These values are higher than those observed in podiform chromites (Jan and Windley, 1990). The accessory chromites (BT1) show still higher values. Jan and Windley (1990) suggest that ferric iron shows a considerable increase in the altered and/or equilibrated grains. Several workers have also shown that composition and stability of chrome-spinels depend not only on temperature, pressure and bulk chemistry but also depends on oxygen fugacity ($fO_2$) (Rivalenti et al., 1981). With increasing $fO_2$ chrome-spinel becomes enriched in Fe$^{3+}$ and impoverished in Mg.

The Mg # of accessory chromites (BT1) varies sympathetically with Mg # of co-existing olivine and Ca-poor pyroxenes. A slight decrease in Mg # (except for one analysis) is shown by accessory chromites. The associated Ca-poor pyroxenes also show a slight decrease whereas, the olivines have almost constant composition. The fluctuations in these ratios may be affected by several factors including changing temperature and $fO_2$ conditions with differentiation. Fisk and Bence (1980) ascribed the decrease of Mg # in chromites to decreasing temperature.
The decrease in Mg # observed in the present case may probably reflect a decrease in Mg content in the melt due to crystallization of Mg rich silicates, however, in this case the decrease could be due to loss of Mg and enrichment of Fe³⁺ during metamorphic alteration and serpentinitisation (Stowe, 1994).

In order to bring out the characteristic features of the chromites, the chemical data have been plotted in different variation diagrams. In the Cr-Al-F³⁺+2Ti ternary diagram (Irvine, 1967; Dickey, 1975; Bird and Clark, 1976; Eales and Marsh, 1983). The chromites from peridotite (B19A) occupy the overlapping fields of residual peridotites (ophiolite peridotites) and stratiform chromites. The accessory chromites from troctolite (BT1) occupy overlapping fields of stratiform -and SE Alaskan-type complexes. The chromitite chromites (B15) however, due to their high Cr and low Al content plot far removed from fields normally occupied by chromites from the three main type of complexes - namely, podiform -, stratiform - and SE Alaskan-type (Fig. 4.1a).
**Fig. 4.1a** Cr-(Fe$^{3+}$+2Ti)-Al plot for chromites from Bondla [based on Irvine (1967), Dickey (1975), Bird and Clark (1976) and Eales and Marsh (1988)].

**Fig. 4.1b** Mg # vs Cr # plot for chromites from Bondla. [based on Irvine (1967), Leblanc et al., (1980). Dick and Bullen (1984)].
In the Mg # vs Cr # face of the spinel prism (Irvine, 1967; Leblanc et al., 1980) the peridotite chromites (B19A) and the accessory chromites (BT1) straddle the fields of stratiform and SE Alaskan-type complexes, whereas, the chromitite chromites (B15) plot outside these fields due to their high Cr # and low Mg # (Fig. 4.1b). Such chromites have been reported from Kenticha Hill (Bonavia, et al., 1993) and from the ultramafic rocks of the Jijal complex, NW Pakistan (Jan and Windley, 1990). However, the behaviour of Mg # is opposite to that shown by Jijal chromites. The chromites from chromitite (B15) have low Mg # than the chromites from peridotite (B19A) and troctolite (BT1).

The variation in Cr # are similar to Cr/(Fe+2+Fe+3) ratio which largely depends on the availability of Cr+3 in the magma. Fisk and Bence (1980) demonstrated that Cr # of chromite co-precipitating with olivine decreases with falling temperature and that this trend may be related to a decrease in Cr+3 in the magma with continued spinel crystallisation. The Cr # of Bondla chromites increases from peridotite chromites (B19A) to chromitite chromites (B15) and then decreases further upwards. This
could be related to enrichment of Cr$^+$ in the magma brought about by replenishment in the magma chamber.

In the Mg # vs Fe$^{+3}$ # face of the spinel prism, the peridotite chromites (B19A) plot in the field of stratiform chromites, the accessory chromites (BT1) occupy the overlapping fields of stratiform - and SE Alaskan-type complexes and chromitite chromites (B15) due to their low Mg # plot outside these fields. The peridotite chromites (B19A) have low Fe$^{+3}$ # (about 0.10), the accessory chromites (BT1) have between 0.18- 0.22 whereas, the chromitite chromites (B15) show Fe$^{+3}$ # varying between 0.12-0.21. Fe$^{+3}$ # of chromitites varies inversely with Mg #. Such a relationship could be ascribed to a drop in temperature at constant fo2. Overall increase in Fe$^{+3}$ content of the spinels from the chromitite layer upwards could be caused by gradual depletion of chromium in the crystallizing magma (Fig.4.2a).

A plot of Cr$_2$O$_3$ vs FeO + Fe$_2$O$_3$ is depicted in Fig.4.2b. In this diagram the peridotite (B19A) and accessory chromites (BT1) plot in the field of stratiform chromites. The chromitite chromites (B15), however, plot
Fig. 4.2a Fe\(^{3+}\) # vs Mg # plot of chromites from Bondla [based on Irvine (1967) and Dick and Bullen (1984)].

outside the field of stratiform chromites due to their higher Cr$_2$O$_3$ and FeO$_{tot}$.

In the TiO$_2$ vs Cr$_2$O$_3$ variation diagram (Fig. 4.3a) the peridotite chromites (B19A) occupy the field of oceanic cumulates near the boundary of the field of SE Alaskan-type complexes. The accessory chromites (BT1) and the chromitite chromites (B15) lie outside these fields due to their higher TiO$_2$ and Cr$_2$O$_3$ respectively.

In the TiO$_2$ vs Al$_2$O$_3$ diagram (Fig. 4.3b) the peridotite (B19A) and accessory chromites (BT1) lie in the field of stratiform chromites forming two separate clusters due to their variation in TiO$_2$ content. The chromitite chromites (B15), however, fall far removed from any of the fields due to their low content of Al$_2$O$_3$.

The TiO$_2$ content of chromites can provide information on the degree of differentiation. The TiO$_2$ content of the Bondla chromites in general increases from chromitite chromites (B15) into troctolite chromites (BT1). Evans and Wright (1972) and Hill and Roeder (1974) have shown that TiO$_2$ content of chrome spinels increases with
4.0
1.0
SE ALASKAN-TYPE
COMPLEXES

OCEANIC CUMULATES

OCEANIC TECTONITES
(Depleted)

OCEANIC TECTONITES
(Highly Depleted)

10 20 30 40 50 60
Cr2O3 Wt.%

0.1

0.01

0

TiO2 Wt.%

Fig. 4.3a TiO2 wt.% vs Cr2O3 wt.% plot of chromites from Bondla [based on Herbert (1982), Irvine (1967), Bird and Clark (1976)].

Fig. 4.3b TiO2 wt.% vs Al2O3 wt.% plot for chromites from Bondla.
progressive fractionation of the basaltic magma. Ti by virtue of its lower distribution coefficient gets enriched in the melt relative to solid phases during fractionation (Eales and Reynolds, 1981). It is natural therefore, that stratigraphically higher chromites are derived from more fractionated liquid.

A number of workers have shown that the stability of chrome-spinels depends not only on the bulk composition, temperature and pressure but also on oxygen fugacity (Irvine, 1965; 1967; Ulmer 1969; Hill and Roeder, 1975; Fisk and Bence, 1980; Rivatenti et al., 1981). Hill and Roeder (1974) and Fisk and Bence (1980) observed that with increasing $f_{O_2}$ the chrome-spinel is enriched in Fe$^{3+}$ and is impoverished in Mg. In order to evaluate the effect of $f_{O_2}$ on the composition of chromites, the analytical data were plotted in the Cr- and Al-free face of the spinel prism (Fig. 4.4) contoured with a set of surfaces representing the possible compositions for different $f_{O_2}$ at constant pressure and temperature (Irvine, 1965). The intersection of three surfaces ($f_{O_2}$ isobars) with the Cr and Al free faces of the spinel compositional prism are shown in (Fig.4.4) alongwith the
Fig. 4.4 $\text{Fe}^{3+}$ $\#$ vs $\text{Mg} \#$ diagram contoured with theoretical constant $fO_2$ isobars from Irvine (1965) showing plots of Bondla chromites. Dashed lines-isobars for Al-free system, dotted lines isobars for Cr-free system.
plots of Bondla chromites. The chromitite chromites (B15) show low fo2 than those in peridotite (B19A) and troctolite (BT1). The chromites from stratiform complexes are normally formed under low fo2 (Jan and Windley, 1990). Cr and Fe+3 contents are related to the percentage of cumulus chromite to cumulus olivine. The chromitite chromite is bound to contain more Cr and less Fe+3 than peridotite chromite (Olivine-chromite cumulate) (Jackson, 1969). The trend of increasing Fe+3 shown by chromitite chromites is largely the result of an increase in fo2 due to crystallization (Roeder and Reynolds, 1991) and replenishment, however, in this case the increase could be related to alteration. It is seen from the diagram that Fe+3 is highest in the accessory chromites (BT1) and decreases in peridotite chromites (B19A). Chromitite chromites (B15) show a range of Fe+3 spanning from peridotite chromites (B19A) to troctolite chromites (BT1). An opposite behaviour is observed in chromites from Alpine-peridotites wherein Fe+3 is higher in spinels from chromitites and spinel-rich dunites than in accessory spinels (Dick, 1977). In the present case however, the increase in Fe+3 may be ascribed in part to secondary alteration, given the altered nature of the rocks. This is also supported
by the trend shown by the Mg # which is reverse to that normally encountered.

Classification of Chromites

The continental geological setting of the Bondla complex, the repetitive nature and way up stratigraphy of the chromite layers alternating with those of olivines, the cumulate textures shown by chromites and associated silicates are all consistent with the stratiform nature of chromites. Mineralogically, the chromites show features which are typical of layered chromites. Mineral chemical features such as high Cr$_2$O$_3$ (more than 45 wt%), TiO$_2$ more than 0.51 wt.% and FeO more than 17 wt.% in chromites containing over 45 wt.% Cr$_2$O$_3$, Al$_2$O$_3$ more than MgO, Cr/Fe ratio (between 1 and 2) which decreases upwards, are features which are characteristics of stratiform chromites. However, in order to distinguish them from segregated layered chromites in ophiolites, the Bondla chromites could be classified as "Bushveld type" following Stowe (1994). The high Cr # ranging from 0.63 to 0.96 and low Al$_2$O$_3$ content between 7.0 and 17.0 wt.% (excluding altered rim compositions) permit them to be classified as
Archaean greenstone-hosted chromites (LeBlanc, 1984; Stowe, 1994).

As pointed out by Irvine (1967) spinels can be used as petrogenetic indicators. Using spinel compositions, Dick and Bullen (1984) divided the Alpine-type peridotites into three groups: Type-I peridotites and associated volcanics having spinels with Cr # less than 0.6, Type-III peridotites and associated volcanics containing spinels with Cr # more than 0.6 and Type-II peridotites and volcanics, those spanning the range of spinels in Type-I and Type-III peridotites. The chromites from Bondla show Cr # more than 0.60 and therefore, they show similarities to spinels in Type-III peridotites of Dick and Bullen (1984). Such high chrome spinels analogous in Type-III peridotites are reported from arc-related lavas from SE Alaska and from continental layered intrusions (Dick and Bullen 1984).

The rock association chromitite-dunite-peridotite (Thayer, 1969) permits the Bondla chromites to be classified as early magmatic segregation deposits (Park and Mac Diarmid, 1975; Jensen and Bateman, 1979). The chromites
can also be categorised as high grade ores (more than 45 wt.% Cr2O3) suitable for the metallurgical industry however, their Cr:Fe(t) ratio is slightly lower than the recommended value of 2.5.

The associated mafic-ultramafic complex can be classified as a medium - and small - sized sheet-like or funnel shaped intrusion emplaced in non-orogenic area (Jensen and Bateman, 1979; Edwards and Atkins, 1986) similar in form to the Skaergaard and Rhum intrusions (Wager and Brown, 1968). Considering the zonal distribution of the rocks - ultramafics forming the Lower Zone and the mafics forming the Upper Zone, the complex shows similarities to the Great Dyke Complex, Zimbabwe where chromites are confined to the lower ultramafic zone (Page, 1977; Wilson, 1982).

**Sulphide Mineralisation**

The sulphide mineralisation is confined to the NW-SE trending zone of mylonite. Along this zone the rocks are converted to quartz-sericite-schist. They are intruded by three sets of quartz-calcite veins which are described earlier.
The sulphides are confined to the quartz-calcite veins. The intersections of veins are the favourable locales for the sulphides. The sulphides occur as lenticles, segregations, clots, stringers, disseminations and also as thin veinlets.

Ore Petrography

The sulphide ore assemblage was studied under incident light. The sulphides are made up of pyrite, pentlandite, chalcopyrite, pyrrhotite with subordinate galena, millerite and Ni- and Co-sulphoarsenides.

Pyrite is the predominant phase in the mineral assemblage. It belongs to two generations. Pyrite-I is euhedral and occurs as megacrysts showing compositional zoning (Plate XIII Photo 1). It is invariably fractured and shows intergrowth with chalcopyrite. Pyrite-II is fine grained, interstitial and spongy in appearance. At places, it replaces pyrite-I. Pyrite-II shows intergrowth with pentlandite (Plate XIII Photo 2). Pentlandite also occurs as discrete grains and as exsolution lamellae in pyrrhotite (Plate XIV Photo 1). The latter is replac-
ing pyrite. The sulphoarsenides are represented by Ni- and Co-gersdorffite. Ni-gersdorffite either occurs as discrete grains or is seen to vein pyrite. Co-gersdorffite often occurs as euhedral grains (Plate XIV Photo 2) often zoned and is intergrown with pyrite. Galena and millerite have restricted occurrence and are generally associated with pentlandite.

Mineral Chemistry

The sulphide minerals were analysed by standard electronprobe technique using an energy dispersive system at C.S.I.R.O., Australia. Between 5 and 15 analyses were performed on each mineral phase from each ore sample. Fifteen representative chemical analyses from about 50 analyses performed in total, are presented in Table No 4.2. These are the first mineral analyses published on the sulphides from Bondla (Dessai, et al., 1994). It is seen from the table that intergrain compositional variation in pyrite within a sample is not very significant. Considerable variation in element concentrations is however observed between pyrites of two generations. Pyrite-I contains between 0.78 and 1.23 wt.% Co whereas, in pyrite-II
Table No. 4.2 Electronprobe Microanalysis of Sulphides from Bondla

<table>
<thead>
<tr>
<th></th>
<th>Pyrite I</th>
<th>Pyrite II</th>
<th>Chalcopyrite</th>
<th>Pentlandite</th>
<th>Pyrrhotite</th>
<th>Millerite</th>
<th>Ni-Gersdorffite</th>
<th>Co-Gersdorffite</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fe</td>
<td>44.46</td>
<td>44.82</td>
<td>45.10</td>
<td>45.16</td>
<td>29.47</td>
<td>29.88</td>
<td>24.87</td>
<td>32.81</td>
</tr>
<tr>
<td>Co</td>
<td>1.23</td>
<td>0.78</td>
<td>0.88</td>
<td>0.01</td>
<td>0.02</td>
<td>0.02</td>
<td>6.29</td>
<td>0.09</td>
</tr>
<tr>
<td>Ni</td>
<td>0.22</td>
<td>0.39</td>
<td>0.05</td>
<td>0.81</td>
<td>0.00</td>
<td>0.02</td>
<td>34.25</td>
<td>31.77</td>
</tr>
<tr>
<td>Cu</td>
<td>0.01</td>
<td>0.02</td>
<td>0.01</td>
<td>0.01</td>
<td>33.08</td>
<td>33.23</td>
<td>0.02</td>
<td>0.01</td>
</tr>
<tr>
<td>S</td>
<td>53.19</td>
<td>52.67</td>
<td>53.68</td>
<td>52.78</td>
<td>35.02</td>
<td>34.99</td>
<td>33.35</td>
<td>34.56</td>
</tr>
<tr>
<td>bal</td>
<td>99.86</td>
<td>99.09</td>
<td>99.39</td>
<td>98.83</td>
<td>97.65</td>
<td>98.24</td>
<td>98.76</td>
<td>99.28</td>
</tr>
</tbody>
</table>
Co percentage varies from 0.01 to -0.88 wt.%. Ni in pyrite-I varies between 0.22 and 0.39 wt.%, whereas in pyrite-II it is between 0.05 to 0.81 wt.% The percentage of Cu in both pyrites is insignificant. Pyrite-I shows higher percentage of arsenic (0.41 to 0.75 wt.%) than pyrite II (0.06 to 0.17 wt.%). Thus it is seen that pyrite is significantly richer in Co and As and poorer in Ni whereas, pyrite-II is richer in Ni and has insignificant concentrations of Co and As.

Chalcopyrite contains 29 to 30 wt.% Fe and between 32 and 33 wt.% Cu. The concentrations of other elements such as Ni, Co and As are insignificant.

Pyrrhotite contains about 57.74 wt.% Fe and 0.72 wt.% Ni. In Pentlandite, the Fe percentage varies from 24.8 to 32.8 wt.% and Ni from 31.7 to 39.6 wt.%. Some samples contain significant amount of Co which may be up to 6 wt.%. The Ni percentage in Ni-gerdsorffite varies from 21.2 to 24.8 wt.%. As is between 41.5 and 42 wt.% and Fe is between 3.4 and 4.9 wt.% Co-gerdsorffite contains about 22 wt.% Co, 40 wt.% As and 8 wt.% Ni.
Discussion

With the limited data available it is not intended here to go into the details of the genesis of the sulphides. However, pending detailed work certain broad deductions can be made within the realm of the available information.

It is widely believed that base metal mineralisation in Archaean greenstones is the result of deposition from hydrothermal solutions. However, there seems to be no consensus of opinion among different workers on the nature and origin of ore fluids. Hutchison and Burlington (1984) and Viljon (1984) considers sea-water derived exhalative fluids as carriers of the base and precious metals. Philips et al. (1984) Fyfe and Kerrisch (1984) advocate metamorphically derived fluids as the transporter of metals whereas, Burrows (1986) and Rock and Groys (1988) suggest magmatically derived fluids as the prime ore carriers.

The sulphide mineralisation at Bondla is controlled along the weak zone within a network of veins and veinlets
of quartz-calcite and sulphides indicating that the mineralized area could represent a focus of hydrothermal solutions which were guided towards the surface along the fracture system. The latter may have served as a feeder system for the mineralizing solutions. The vein mineralogy, the presence of significant amount of calcite, the abundance of Ni and Co in the ores are all consistent with Co2 rich hydrothermal fluids. The presence of hydrothermal epidote in the alteration haloe may indicate as a first approximation a temperature of about 250 - 270°C (Browne, 1978). This temperature range is also corroborated by the experimental phase relations of the assemblage pyrrhotite - pentlandite which indicates temperature less than 310°C. The low temperature of ore formation is also supported by the presence of millerite in the ore assemblage. In most documented exhalative deposits, precipitation temperatures are commonly below 300°C (Sato, 1972). At temperatures of 300°C and below there is considerable uncertainty regarding phase relations in the Cu - Fe - S system (Vaughan and Craig, 1978). Nevertheless a H2S dominant ore fluid should successively precipitate pyrite - pyrrhotite - pentlandite with decreasing Fe activity and temperature. Absence of magnetite and haematite with
quartz may indicate low oxygen fugacity conditions. High Co/Ni ratios and high Co and Ni content of pyrite may indicate volcanogenic derivation. The mineralisation bears some similarities with the fracture controlled stringer ores of Kuchiganehal of Ingaldhal sulphide zone from Chitradurga schist belt. The latter mineralisation has been suggested as of volcanogenic, hydrothermal derivation (Ananth-Iyer and Vasudev, 1985). However, further detailed work on the sulphide mineralisation of Bondla is essential before any definite conclusions on the nature of mineralisation could be drawn.