CHAPTER V
CHAPTER – V

PETROCHEMISTRY AND GENESIS OF ULTRAMAFIC MAFIC - MAFIC ROCKS AND CHROMITE - V-Ti MAGNETITE ORES

5.1. INTRODUCTION

The study of Archaean ultramafic and mafic complexes has received a greater attention as these rocks form important components of several late Archaean greenstone belts (Windley and Smith, 1976). In Precambrian terrains the ultramafic-mafic complexes are usually found as lenses, which have been deformed and metamorphosed along with the adjacent supracrustal rocks. In spite of strong tectonic and metamorphic overprint, the original igneous stratigraphy, the process of generation and differentiation of magmas and their mode of emplacement may be elucidated in many cases by careful investigation of the field setting, petrography and geochemistry of these complexes.

Apart from the layered igneous complexes, a distinctive group of ultramafic to mafic rocks, komatiite-tholeiite series, occurs in many Archaean greenstone belts, either forming the basal members (Viljoen and Viljoen, 1969; Arndt et. al., 1977) or overlying the sedimentary units (Bickle et. al., 1975; Hendersen, 1981; Kroner et. al., 1981). Komatiites were thought to be restricted to Archaean greenstone belts, indicating unusually high geothermal gradients in the upper mantle before 2.5 Ga ago. The recognition of Proterozoic and Palaeogene komatiites (Gansser et. al., 1979), however, indicated that such conditions might not have been unique to the Archaean. In highly deformed and metamorphosed Archaean terranes, the distinction between komatiites and layered igneous complexes has major implications for models of
crustal evolution.

The mafic-ultramafic rocks occur with in the Dharwar Supergroup as well as with in the Peninsular Gneissic Complex (PGC) as enclaves of Sargur Schist Complex in Karnataka. Though the intrusive complexes occur with in the Dharwar as well as Sargur settings, the extrusive bodies occur mainly confined to the Sargur Schist Complex. Layered intrusive complexes with dunite +/- chromite, peridotite, pyroxenite, gabbro, anorthosite +/- chromite, titaniferous vanadiferous magnetite are seen in Nuggihalli and Sargur areas in older supracrustal belts and in parts of Shimoga schist belt (Masanikere-Taverekere-Magyathahalli-Hanumalapur), North Kanara (Kaiga-Mothimakki-Suryakalyanigudda) and Goa (Usgaon complex) area in the younger Dharwar belt.

Spinifex texture, pillowed, nodular are ocellar ultramafics are reported from several localities of Karnataka and Goa. The report came from Ghattihosahalli-Kummanaghatta area from the western margin of Chitradurga Schist Belt (Viswanatha et. al., 1977). Subsequently, extrusive komatitic rocks were reported from Mayasandra (Charn et. al., 1988), Banasandra (Srikantia and Bose, 1985; Srikantia and Rao, 1990), Nagamangala (Srikantia and Venkataramana, 1989; Devapriyan et. al., 1994), J.C. Pura (Venkatadasu et. al., 1991), Karighatta (Srikantappa at. al., 1992), Nuggihalli and Holenarasipura belts (Jaffri et. al., 1997) Mothimakki (Kariyanna and Asit Saha, 1997), Kalyadi (Subbarao and Naqvi, 1999), Sandur (Manikyamba and Naqvi, 1997 and 1998), and from Goa (Desai and Deshpande, 1979).

The belt north of J. C. Pura with the chain of enclaves with in the PGC hosting several chromite parts has been described by Ramachandra Rao (Sampath Iyengar, 1931) and
referred to as Shivani belt. In J.C. Pura area, ultramafic extrusive komatiites with periodic and basaltic komatiite variants with in the PGC are overlain by the Bababudan sequence with a quartz pebble conglomerate offering another spectacle of Sargur-Dharwar orogenic discrimination in Karnataka. (Venkata Dasu et. al., 1991). Though the intrusive ultramafics with chromites have been reported earlier for its PGE mineralogy from one of the locations from Chitradurga belt, the presence of vast areas occupied by the pillowed, nodular or ocellar ultramafites extending for over 50 km strike length and over 500 m width, as a major belt is being reported for the first time. Thus, the additional belt of extrusive ultramafic volcanism reported here, once again confirms the existence of a long linear and parallel komatiitic belt on the western side of the Chitradurga Schist belt signifying the extent of ensialic rifting (Goodwin, 1981), representing the zones of mantle tapping (Arndt, 1994) which took place during early Archaean.

Vishwanatha (1974), Naqvi and Hussain (1979), Hussain and Ahmed, (1982) and Jafri et. al., (1983), suggest that the ultramafic to mafic rocks in the Nuggihalli and Holenarasipur belts represent komatiitic volcanic rocks that form the basal members of the individual schist belts. Fine-grained anorthositic rocks associated with the ultramafic complexes were compared to lunar anorthosites and interpreted as primitive anorthositic basalt lavas (Drury et. al., 1978; Naqvi and Hussain, 1979).

Based on detailed geological and structural work in these belts, Swami Nath et. al., (1974), Janardhan and Srikantappa (1974) have shown that the ultramafic - mafic complexes are stratiform layered intrusives and are not komatiite metalavas. The fine-grained anorthosites of the Holenarasipur belt are now regarded as cumulates from
basic magmas, their fine-grained texture being attributed to intense shear deformation (Bhaskar Rao and Veerabadrappa, 1982).

The contrasting views regarding the origin of the ultramafic-mafic complexes led to the postulation of different models for the evolution of the Archaean crust in southern India. Formation of the greenstone belts by a process of initial rifting and eventual ocean opening was proposed by Naqvi (1976), in contrast to an intracrustal development implied by Swami Nath et al., (1974).

Paranthaman and Vidyadharan (2005) have identified and reported a new belt of Archaean ultramafites with a dominant extrusive and an equally dominant intrusive component (a part of the study area) in the northern strike extension of Nuggihalli belt and the J.C. Pura enclave (Venkata Dasu et al., 1991 and Ramakrishnan et al., 1994). This major schist belt extends in NNW-SSE for more than 50 km strike length and 500 m average width from north of Arasikere in the south, across the Vedavathi river up to the southern part of the Shimoga Schist belt in the north. This belt is disposed with an easterly shift with respect to Nuggihalli and a westerly shift with respect to J. C. Pura Schist belt and occur as major enclaves within the Pennissular Gneissic Complex, which are represented by medium grained foliated grey biotite gneiss, banded gneiss, granite gneiss, pink granites with quartz and pegmatite veins. This prominent belt of enclaves (Sargur) is located west of the Kibbanahalli arm and north of J. C. Pura Schist belt with Arasikere-Banavara granites towards south and Kadur granites in the west (east of Shimoga and Bababudan Schist belt and west of Chitradurga belt).

Granular chromite in the form of disseminations associated with olivine, besides the layers and pods, are recorded from east of Rangapura-Kallangavi within dunite. Thin
layers of chromite, varying is size from a few mm to 3 cm are recorded from southwest of 940 m hill. Within the serpentinised dunite, chromite is associated with kammerarite (80 % chromite, 15 % kammerarite and 5 % altered olivine). Floats of chromite are seen over 100 m x 5 m area. The chromite bearing ultamafics trend N 10° W to S 10° E with 70° northeasterly dips showing tight F2 fold structures.

Around Donnakaranahalli (few mm to >2 cm) with gaps of 5 to 15 cm between each layer are seen associated with the meta-pyroxinite band of 5 m width. The layers trend N10-20° W – S 10-20° E with 80° southwesterly dips. Within the major zone of dunites, 2 to 4 cm size crystals of altered chromite are seen. This dunite body exposed over 500 m x 150 m is carbonated and altered to talc at the margins and show tight F2 folds with NW-SE axes. Rich, grey coloured remobilized chromite veins (1 to 2 cm wide) cut across the original chromite layers in dunites. Other areas like Udugere, Vaddarahalli, Doddaannegere, Sankalpura, Doddapayyanahalli, Adigerahalli, Mallenahalli, Basavanahalli, north of Shivaganga gudi, Bhairagoundanahalli, Yeredegere and Rangayyanuru exposed, black greenish peridotite with excellent pillow, nodular or ocellar structure and polyhedral jointing pillows are ovoid, bulbous and deformed in shape (10-50 cm x 50-100 cm) showing characteristic marginal rinds. In between bigger pillows, smaller ones are seen with cuspate margins adjoining to pillows. Radial cracks are seen at places, though vesicles and amygdales are uncommon. However, pillowed ultramafics exhibit polyhedral jointing with 20-70 cm width and up to 1 m length. Margins of each polyhedral joint form a chilled zone of 2 to 6 cm width. Nodular or ocellar structures are well exposed at Doddapayyanahalli and Mallenahalli area. They consist of nodules or ocelli of 1 cm to 3 cm diameter with
concentric rings embedded in very fine-grained and serpentinised peridotite matrix. They are rounded and generally occur as protruding clusters on weathered outcrops. Meta-ultramafites show varying degrees of serpentinisation, steatitisation and carbonitisation. Veins of antigorite and chrysotile asbestos and talc are also recorded from several localities. The massive type grades into talc – tremolite – chlorite rock with development of octahedral magnetite crystals.

The intrusive ultramafic variants with chromite association occur in the northern part of the area and extend for 10 to 35 km strike length; whereas the extrusive ultramafics closely associated with the intrusive component and equally dominant rock type in the belt extends for 40 km strike length towards the middle and southern part of the area. Ultramafic bodies of the various dimensions occur as major enclaves within the PGC. Minor shiny amphibolite bands occur in both ultramafic associations and show development of garnet at gneiss/granite contacts.

In addition to ultramafic rocks, minor banded magnetite quartzites, fuchsite quartzites and chromiferous chert bands also occur in this belt. One such chrommiferous chert band occurs associated with the deformed and folded peridotitic komatiite near Udugere. The F2 folds plunge 50° towards NW and show a general NW-SE axis. Gabbro dykes trending ENE – WSW to EW (pre-tectonic) and dolerite dykes (post-tectonic) trending NW – SE directions are recorded from the area. The area has undergone amphibolite facies metamorphism and three episodes of deformation with F1 and F2 in NS to NNW - SSE as coaxial and F3 as broad warps and crenulation cleavages in NE – SW to EW direction parallel to the trend of the enclosing gneisses.
In the study area, though the co-existence of both the chromite bearing ultramafic intrusives and the pillowed and nodular extrusive komatiitic ultramafic rocks by Paranthaman and Vidyadharan (2005) only the chromite bearing ultramafic intrusives and the associated V-Ti magnetite bearing mafic rocks have been investigated. The discussion of their petrogenesis is based on field observations and on detailed petrographic and geochemical data.

In this section, the major and trace element data of representative samples of dunite, peridotite, chromitite, serpentine and gabbro-anorthosite suite of rocks along with their CIPW norms and critical elemental ratios are presented. The chemical data has been used to find out the overall chemical and genetic relationship between the members of the ultramafic - mafic complex and also to elucidate the geochemical feature of the parent magma using various discriminated diagrams.

5.2 MAJOR AND TRACE ELEMENTS IN ULTRAMAFIC ROCKS

5.2.1 Major elements

The chemical analysis data of two dunites, two peridotites, two chromitites and one serpentinite along with their normative mineral compositions are presented in the Table 5.1. From the above, it is evident that dunite, and peridotites are olivine normative of the study area. In dunites of the study area, SiO$_2$ varies from 37.43 to 39.12 wt. %, MgO varies from 32.48 to 34.25 wt. %, Fe$_2$O$_3$ varies from 8.93 to 9.50 wt. %. In peridotite SiO$_2$ varies from 34.21 to 37.34 wt. %, Fe$_2$O$_3$ 10.24 to 11.14 wt. %, MgO 38.02 to 40.25 wt. %. In dunite and peridotite TiO$_2$, MnO, CaO, Na$_2$O and P$_2$O$_5$ are present in small quantities and exhibit marginal variation. Chromitites show variation in SiO$_2$ from 11.73 to 12.81 wt. %, Al$_2$O$_3$ 5.89 to 6.96 wt. %, Fe$_2$O$_3$ 17.65 to
24.27 wt. %, MgO 20.37 to 22.34 wt. %, CaO 0.20 to 0.31 wt. %. In serpentinite, SiO₂ is 36.54 wt. %, Fe₂O₃ is 11.86 wt. %. MgO, 39.66 and CaO is 0.75 wt. %.

5.2.2 Trace elements

Higher charged cations such as Cr, V, Ti, Sr are rather immobile under most metamorphic conditions (e.g., Pearce and Cann 1973, Hamilton et. al., 1979) and hence can be used as petrogenetic indicators.

The trace elements like Cr, Ni and V were analysed in the ultramafic rocks of the study area and the data (in ppm) for two dunite, two peridotite two chromites and one serpentinite sample is presented in Table 5.1. The average trace element composition of ultramafic rocks from Sargur area (after Srikantappa et. al., 1984) is also presented in Table 5.2 for comparison.

In dunites of the study area Cr values are noticed up to 7300 ppm, Ni varies from 800 to 1200 ppm. In peridotite Cr varies from 6500 to 6800 ppm, Ni varies from 1200 to 1300 ppm. In serpentinite Cr values are present up to 4700 ppm and Ni values noticed up to 1000 ppm.

5.3 MAJOR AND TRACE ELEMENTS IN GABBRO–ANORTHOSITE ROCKS

5.3.1 Major elements

Chemical data of 14 samples of gabbro-anorthosite suite of rocks of the study area along with elemental ratios and CIPW norms are presented in Table 5.3. The major element data of gabbro - anorthosite suite of rocks of other similar layered complexes are also presented in Table 5.4 for comparison.

The gabbro - anorthosite suite of rocks of the study area show moderate variation in
SiO₂ content which ranges from 42.57 wt. % to 50.59 wt. %. Magnetite-gabbro shows little variation in Al₂O₃ from 13.50 wt. % to 18.00 wt. %, CaO from 4.01 wt. % to 9.99 wt. %. The lowest value of Al₂O₃ is found in magnetite-gabbro and highest value in gabbroic anorthosite. Highest value of CaO is found in anorthositic-gabbro, the lowest value of CaO is noticed in magnetite-gabbro. This variation in Al₂O₃ and CaO is due to variable amounts of plagioclase and clinopyroxene in these rocks (Table 5.3). The Fe₂O₃ ranges from 0.62 wt. % to 29.45 wt. %. Higher amount of Fe₂O₃ is noticed in magnetite-gabbro. The chemical composition of magnetite-gabbro of the study area can be broadly compared well with ferrogabbro of Ovre complex of Norway (Gjelsvik, 1957), ferro gabbro of Busheveld Igneous Complex of South Africa (Hall, 1932) and bleached magnetite-gabbro of Nuggihalli ultramafic complex (Radhakrishna et. al., 1973).

Gradual increase in TiO₂ content (0.34 to 4.50 wt. %) is noticed in these rocks from gabbroic-anorthosite through gabbro and to magnetite-gabbro. TiO₂ content also shows positive correlation with increase of iron content and this feature is due to the presence of variable amounts of ilmenite in these rocks.

MgO content of these rocks show variation from 0.92 to 12.88 wt. %. The lowest value (0.92 wt. %) is noticed in one sample of magnetite gabbro. These rocks show a marginal variation in Na₂O and K₂O content. Na₂O content ranges from 1.46 to 3.97 wt. %. The highest value (3.97 wt. %) is noticed in magnetite gabbro and the lowest value (1.46 wt. %) is seen in gabbro. The K₂O content in these rocks varies from 0.06 to 0.55 wt. %. The lowest value of K₂O is noticed in gabbro and the highest, in magnetite gabbro. The P₂O₅ content of these rocks varies from 0.03 to 0.11 wt. %.
An appreciable amount of sulphur (0.22 to 0.42 wt. %) is noticed in the magnetite gabbro. It is contributed by the sulphides like pyrite and chalcopyrite present in the rock.

The $\text{K}_2\text{O}/\text{CaO}$ ratio of these rocks ranges from 0.001 to 0.01 wt. %. This value is less than that of proterozoic anorthosites.

The $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratio of these rocks ranges from 7.22 to 56.30 wt. % and $\text{CaO}/\text{Na}_2\text{O}$ values range from 1.52 to 7.26 wt. %.

The $\text{CaO}/\text{Al}_2\text{O}_3$ ratio of these rocks is less than 1 and ranges from 0.29 to 0.99 wt. %. Such low ratios are characteristic of tholeiitic magmas.

5.3.2 Trace elements

The trace elements like Cr, Ni and V were analysed in the gabbro-anorthosite suite of rocks of the study area and the trace element data in ppm is given in Table 5.3. In these rocks Cr varies from <10 to 180 ppm, Ni varies from 20 to 180 ppm and V varies from 20 to 1200 ppm.

5.4 GEOCHEMICAL INTERPRETATION

In this section, the geochemical and genetic features of various rocks and ores are evaluated based on their major element content. AFM diagram (Fig. 5.1) of dunite, peridotite / serpentinite, chromitites and gabbro-anorthosite suite of rocks reveal an alkali poor tholeiitic trend with high degree iron enrichment. The magnetite -gabbro associated ores fall towards the FeO end whereas the dunite, peridotite/serpentinite, chromitites fall in the midway between FeO and MgO ends which may be attributed to higher FeO contents in the dunite, serpentinite/peridotite,chromitties.

Nesbitt et. al., (1979) suggests that the major elemental ratios such as $\text{CaO}/\text{TiO}_2$ and
Al₂O₃/TiO₂ could be used for recognition of peridotitic komatiite, since they should show chondritic abundances. At high degree of partial melting of peridotite, CaO, TiO₂ and Al₂O₃ enter the melt provided that olivine is the only residual phase.

The Al₂O₃-TiO₂ and CaO-TiO₂ (Fig. 5.3a & b) plots for the dunite, peridotite, chromitite, serpentinite and the gabbro-anorthosite suite of rocks of the study area clearly indicate that they are non-chondritic in nature.

Brookes and Hart (1974), have drawn attention to the fact that any mafic or ultramafic rocks which are rich in calcic pyroxene or calcic amphibole could be chemically similar to komatiites and therefore, emphasised the need to have spinifex texture to confirm its liquid composition. A metamorphic rock without primary texture indicative of liquid composition could have attained a komatiitic geochemistry either by its igneous progenetor fractionally crystallizing from a melt or by the metamorphic modification of its original chemistry. In the present study, the ultramafic rocks exhibit gradual relationship with gabbroic rocks, which in turn grade into magnetite gabbro and to V-Ti magnetite layers. Further, the gabbroic rocks, which are associated with ultramafic rocks, do not show any chemical features suggestive of their similarity with basaltic komatiites. Therefore, the overall geochemical features suggest that the ultramafic rocks of the study area form a part of layered complex and cannot be designated as komatiites. They are tholetites. The process of differentiation is well documented by systematic compositional trends connecting the ultramafic rocks with those of gabbro-anorthosite suites of rocks, on MgO vs. CaO, MgO vs. Al₂O₃ and MgO vs. FeO⁴ diagrams (Fig.5.3c, d & e), thus indicating that the ultramafic rocks and gabbro-anorthosite suite of rocks of the study area originated from a common parent
magma of tholeiitic composition which is rich in Cr-Fe-Ti-V by crystal-liquid fractionation and gravitational settling.

Major element variations (SiO$_2$, Al$_2$O$_3$, FeO$^t$, TiO$_2$, Na$_2$O, CaO, MgO and K$_2$O) and trace element variations with $100 \text{ FeO}^t/(\text{FeO}^t + \text{MgO})$ as an index of fractionation (F) for the ultramafic rocks and gabbro-anorthosite suite of rocks are shown in Fig.5.5a & b, 5.6a & b, 5.7a & b, 5.8a & b and 5.9a & b.

The ultramafic-gabbro-anorthosite suite of rocks exhibits a wide range of compositional variations as shown by the F $[100 \text{ FeO}^t/(\text{FeO}^t + \text{MgO})]$. The F ranges from 19.01 to 51.74 in ultramafic rocks, 15.50 to 95.50 in gabbro-anorthosite suite of rocks with a moderate to high degree of iron enrichment, representing the early, middle and the late stage of crystallization. These trends are due to removal/concentration of major elements during various stages of crystallization. Early stage crystallization is characterized by removal of MgO and enrichment of Al$_2$O$_3$, CaO, FeO$^t$ and TiO$_2$ with increase in F $[100 \text{ FeO}^t/(\text{FeO}^t + \text{MgO})]$.

A plot of FeO$^t$ vs. TiO$_2$ for the ultramafics and gabbro-anorthosite suite of rocks of the study area show the positive correlation between FeO$^t$ and TiO$_2$ (Fig. 5.4a) thus indicating a general increase of titanium with increase in total iron. In other words, there is an increase in ilmenite with increase in magnetite. This relationship suggests a solid solution relationship between two mineral phases, which is clearly evident in polished sections of ores. And the MgO vs. TiO$_2$ diagram (Fig. 5.4b) shows negative correlation between MgO and TiO$_2$ indicating an increase in TiO$_2$ from ultramafic rocks through gabbroic-anorthosite, anorthositic-gabbro to magnetite-gabbro. The FeO$^t$ and TiO$_2$ vs. F $[100 \text{ FeO}^t/(\text{FeO}^t + \text{MgO})]$ diagrams (Fig. 5.6a & b) show a
positive correlation. These diagrams indicate middle to late stage crystallization of Ti magnetite along with ilmenite.

TiO$_2$ content increase in the melt during the early fractionation of olivine and chromite. The Ti preferentially enters the lattice of magnetite and ilmenite during magmatic crystallization (Irvin, 1967). Since only accessory amounts of ilmenite and Ti magnetite are associated with ultramafics, anorthositic gabbro, gabbroic anorthosite and gabbro and since large amounts of Ti magnetites and ilmenite are noticed in magnetite-gabbro and V-Ti magnetite ores, it is suggested that much of Ti was incorporated in the ilmenite and Ti magnetite during the late middle stages of differentiation, i.e., the formation of magnetite-gabbro and V-Ti magnetite layers.

CaO and Al$_2$O$_3$ concentrations increase with F to a certain stage and then decrease with the crystallization of clinopyroxenes, plagioclase and Ti magnetite (Fig. 5.5a & b), thus indicating that the concentration of these elements is largely controlled by plagioclase formation. FeO$^t$ and TiO$_2$ increase till the late stage is reached when Ti magnetite is crystallized. Crystallization of clinopyroxenes and plagioclase mainly represent the early middle stage where crystallization of Fe-Ti-oxide increases and crystallization of Fe-Ti oxides is maximum giving rise to V-Ti magnetite ore layers, when F is 95.50. This relationship is indicated by the positive correlation between FeO$^t$ and F as well as TiO$_2$ and F (Fig. 5.6a & b).

Na$_2$O increases with F (Fig. 5.7a) from ultramafic to gabbro anorthosite suite of rocks and suddenly decrease at the crystallization stage of V-Ti magnetite.

K$_2$O exhibits a mixed trend with F in ultramafic to gabbro anorthosite suite of rocks of the study area. (Fig. 5.7b).
MgO vs. F (Fig. 5.8a) shows a negative correlation throughout the crystallization process. MgO decreases from 40.25 to 20.37 wt. % in ultramafic rocks and from 12.88 to 0.92 wt. % in gabbro-anorthosite suite of rocks indicating early crystallization of chromite, olivine, pyroxene, minor plagioclase followed by late crystallization of plagioclase clinopyroxene giving rise to gabbro-anorthosite suite of rocks. SiO$_2$ vs. F (Fig. 5.8b) exhibits a mixed trend during the crystallization process.

5.4.1 Geochemical interpretation of major vs. trace elements

At the time of fractional crystallization process of the magma, distribution of trace element is largely controlled by nature of bonding, ionic radius, electronegativity (Ringwood 1955) and their preference for major elements. However, the octahedral site preference energy influences the relative rate of removal of transition elements (Cr, Ti, Ni, V) through fractionating phase in basic magma (Walsh and Clarke, 1982). The MgO vs. minor elements like Cr, Ni and V diagrams are used to know the distribution of these trace elements.

It is well known that the Cr and Ni contents in the ultramafic - mafic series shows a gradual decrease in their content from ultramafic end to mafic end member. Cr and Ni concentrate in the early stage and decrease in the late stages in the igneous series as shown by Arth (1976). Ni substitutes for Mg or Fe$^{+2}$ in early-formed olivines, chromites and to a lesser extent in subsequently formed pyroxene and Fe-Ti oxides. It is observed that Cr and Ni contents gradually decrease (Fig. 5.10a & b) from ultramafic rocks to gabbro-anorthosite suite of rocks and these elements show a negative correlation with F[100 FeO$^+$(FeO$^+$ + MgO)] (Fig. 5.9a & b) for the whole range of ultramafic-mafic sequence of the study area. This relation suggests that Ni is
accommodated in early-formed chromite, olivine and pyroxene as well as in the late-formed pyroxene, and ilmenite rich Ti magnetite. This relation is also indicated by the plot of Ni vs. MgO exhibiting positive correlation (Fig. 5.10b).

Cr exhibits positive correlation with MgO (Fig. 5.10a). In addition, the slope of the trend is steep in ultramafics. It clearly suggests that lot of chromite is associated with dunite. This resulted in less chromium in the late differentiates such as gabbro-anorthosite suite of rocks.

Optical studies of ultramafic and gabbro-anorthosite suite of rocks reveals that the concentration of magnetite and ilmenite increase from anorthositic gabbro, through gabbro, magnetite-gabbro to V-Ti magnetite ores. Hence, part of the vanadium content may be present in pyroxenes of pyroxenites and gabbro-anorthosite suite of rocks of the study area. V vs. MgO plot shows negative correlation between V and MgO (Fig. 5.11) indicating that V and Mg behaved as incompatible elements during crystallization of ultramafic- gabbro-anorthosite suite of rocks.

5.5 GENESIS OF CHROMITE AND V-Ti MAGNETITE DEPOSITS: A REVIEW

5.5.1 Genesis of chromite deposits: A review

Major chromite deposits are classified into two geologic deposit types: Stratiform deposits (Bushveld Igneous Complex of South Africa, Kemi intrusions in Finland, Great Dyke, Zimbabwe, Stillwater complex USA, Sukinda ultramafic complex, Orissa, India, Sargur schist belt and Nuggihalli ultramafic-gabbro complex of Dharwar craton, India). Only layered complexes more than 1.9 Ga old are known to contain economic deposits of chromite. Podiform deposits (also called Alpine type) occur in
portions of oceanic crust and mantle that have been thrust up on to continents or Island arcs (e.g., Ural mountains of Soviet Union, Appalachian mountains of N-America and Tethian mountain chains of Albania, Greece and Turkey) and island arcs (Philippines, the Caribbean basin, Cuba). The host rock for Podiform type chromite deposits is peridotite. These deposits are seldom older than 1.1 Ga.

In the following section, a brief review of the genetic models proposed for the chromite deposits associated with ultramafic-mafic complexes of different areas are presented for easier understanding of the controversies in the proper perspective. Chromite deposits have long been considered unimpeachable illustrations of early magmatic segregations. Field and microscopic evidence testify that in many places this is the case. However, Diller, J. G Singewald, Junior, Sampson, C. S. Ross, E. Cameron and others cite evidence of some late magmatic or even hydrothermal chromite.

The Stillwater complex of Montana is considered to be an early magmatic segregation type complex. Some of the South African geologists consider that the chromite deposits in the Bushveld Igneous Complex are to be early magmatic segregations formed by sinking of early-formed crystals. Also, Cameron and Emerson (1959) come to similar conclusion for the eastern Bushveld chromite. Sampson (1929) considered the formation of chromite as late magmatic. Thayer (1956) points out that it is not correct to consider the formation of magmatic segregations with reference to the crystallization of olivine alone during the crystallization of magma and call them as “early magmatic” or “late magmatic”. He considered that in a broad sense all primary (magmatic) chromite deposits are early magmatic.
Jackson (1963), Irvine and Smith (1967), Cameron and Desborough (1969), Ghisler (1970) and Chakraborty and Chakraborty (1984) while explaining the mechanism of formation of chromite deposits of the Stillwater (Montana), Muskox (Canada), Bushveld (South Africa) and Fiskaenaset (Green Land) and Sukinda valley (Orissa) proposed that the cumulus texture and persistent compositional banding in the chromite layers presumably originated by magmatic differentiation and gravity settling of the early formed chromite crystals.

According to Cameron and Eugene (1959) the origin of the Steele port and Leader chromite seams in the pyroxenite series of the eastern part of Bushveld Igneous Complex, is as follows:

i) The chromitite seams are a consequence of magmatic differentiation. Available evidence is against formation by selective metasomatism of pre-existing sedimentary rocks ii) Structural features of the chromites and associated rocks indicate that magmatic currents played an important part in the development of contrasting rock types, producing differential deposition of crystals settling through the magma iii) Textural features of the pyroxenites associated with the chromitite are consistent with, settling and early crystallization of chromite and bronzite, crystallization of interstitial liquid to yield over growths on the settled crystals and interstitial plagioclase. iv) Biotite and clinopyroxene are end stage magmatic products. Their crystallization was unrelated to the formation of chromite. v) The order of crystallization of minerals in the pyroxenites, in the upper and lower members of the steel port, seam and in the basal transition and upper member of the Leader Seam indicated that the chromitite seams did not form by residual liquid accumulation. vi) Textual and compositional
features of the chromitite are consistent with the hypothesis that the main accumulation of chromite and bronzite was accomplished magmatic sedimentation, but they indicate that part of the chromite and bronzite crystallized in situ. Partial resolution of settled crystal seams required as a supplementary process. Diffusion related to thermal gradients may have contributed to enrichment in chromite. vii) Formation of the chromitite seams partly or wholly as a consequence of liquid immiscibility offers formidable difficulties, but liquid immiscibility may conceivably have played a supplementary role in chromite formation.

Cameron and Emerson (1959) based on the geochemical and mineralogical data concluded that the chromite deposits of Bushveld Igneous Complex are formed through progressive differentiation from mafic, magnesium-chromium enriched units near the base to intermediate, FeO-enriched, more sialic phases near the top, confirming the conventional tholeiitic differentiation. The layering is due to (i) magma mixing upon frequent multiple intrusions, (ii) reversals in the thermal or chemical regime resulting in shift in crystallization ratios near the magma chamber roof, so that different crops of crystals fell to the floor at successive times, (iii) variations in the oxygen fugacity ($fO_2$) in the melt and expansion - contraction of the chromite stability field (iv) differential magmatic sedimentation of minerals of different shapes and sizes, (v) sheeted, turbidite like spreadings of crystal mushes banked along the walls outward and downward across the floor (vi) the evolution and gravitative separation of immiscible silicate and oxide magmatic fluids (vii) in situ bottom crystallization controlled by interaction of melt composition pressure-temperature gradients (viii) differential foliation of lighter phase like plagioclase during sedimentation of denser
crystals like pyroxene and chromite and (ix) variations in total pressure and to combinations of all of these above factors. Differential magmatic sedimentation appears to explain many of the physical and chemical requirements.

Del Lama et al., (2000) have reported the textural features in the chromite – bearing ultramafic-mafic complex of the Jacurici River valley, Chromian district located in northeast Bahia state of Brazil. They opine that these chromite-bearing ultramafics exhibit a great geological, petrological complexity, intensely deformed, metamorphosed and transformed by metasomatic processes. The minerals of the spinel group show extensive solid solutions at high temperatures and fields of significant exsolution with temperature decrease, depending on the composite of the original phase.

Gornostayev et al., (2000) while describing the chromite deposits of Ukarine shield, concluded that they are associated with early proterozoic ultramafic complexes of presumably ophiolite nature. These chromite deposits contain base metal sulphides and PGE. The chromite chemistry of Ukarine deposits is comparable to typical ophiolite complexes.

Bao Peisheng and Wang Xibin (2000) while describing the podiform chromite deposits hosted in mantle peridotites, they identified two types of chromite deposits: High-Cr type chromite deposits (PPG series) and high-Al type chromite deposits (PTG series). And they concluded that these two types of podiform chromite deposits produced by varying degree of partial melting and reconstituted primary pyrolite. Spier et al., (2000) opines that the chromite deposits of Bacuri mafic-ultramafic complex representing the second largest reserves of chromite in Brazil. It is a major
stratiform complex overprinted by ductile deformation and associated regional amphibolite facies metamorphism, intrusive into gneiss-migmatite terrains of the Guyana shield. It consists of thick layer of chromitite (12 m) located at the base of ultramafic zone. Several thinner layers of massive chromitite are located within the ultramafic zone. The matrix of massive chromitite consists mainly of metamorphic silicates (serpentine, chlorite, tremolite). The compositional variation trends of chromite of this complex are similar to typical stratiform chromite deposits such as Bushveld and Great Dyke.

According to Verma (1964) the chromite mineralization in the Keonjhar district of Orissa is characterized by its intimate association with rocks of ultrabasic belt. The intrusive nature of the deposits, the presence of ore apophyses branching from the main masses into the surrounding rocks and the clear evidence of wall-rock replacement indicate that the ore bodies were formed by the introduction of an end phase chromite-rich differentiation fraction of the parent ultrabasic magma. Deep-seated slow crystallization of the magma resulted in the production of great series of highly differentiated rocks and in the final separation of the chromite-rich residuum. Instead of segregating with rocks of early magmatic period, the ore-bodies were deposited from mineralising solutions that extracted and transported the chromite and accompanying gangue mineral into shear zones, which had developed in the host rocks after their consolidation but prior to the onset of mineralization.

According to Banerjee (1972) the chromite deposits of Sukinda chromite field are formed by early magmatic segregation through fractional crystallization and gravitative accumulation of early magmatic chromite in pulses. This rhythmic
variation is probably due to fluctuations in the cooling gradient of the geotherms, advancing from the base inward and upward the ultrabasic mass, as visualised by Hess (1960) in the case of the Stillwater complex, Montana. This process was possibly aided by extremely sluggish diffusion and lack of strong convection currents. Under such conditions, the upper layer may remain molten at a time when the lower layer, due to the advancing gradient of lower temperature, precipitates chromite and/or silicates.

The Sukinda chromite field forms the eastern part of a 50 km long belt of basic-ultrabasic intrusives, extending discontinuously from Kantalsuan in Keonjhar district south-eastward to Katpal in Dhenkal district and then east and north-eastward up to Tungaisuni in Cuttak district. Further northeast chromiferous ultrabasic rocks are seen in Nausahi where the lithological association, time of emplacement and tectonic setting are similar to the Kantalsuan-Tungaisuni belt. The occurrences of chromiferous ultrabasic rocks in similar setup in the northeastern fringe of the Iron Ore stage block (Sarkar and Saha, 1963) is noted in south Singhbum, Bihar. A tabular comparison among all the above chromiferous bodies, given below, brings out their fundamental similarity.

Banarjee (1972) concluded that the chromiferous suites of south Bihar and Orissa were all emplaced after the deposition of iron ore stage sediments, but before the granite intrusion and folding movements. Some occurrences like those of the Sukinda and south Singhbum were given enough time to crystallise in situ, before folding movements started, others like the Katpal and Nausahi deposits, which were possibly a little later, got caught up in the great upheaval, before they could completely
consolidate. Hence the lenticular shape of the chromite deposits therein, and their orientation in diverse trends. This early orogenic ultrabasic magmatism, characterised by a super abundance of Cr, appears to mark out the iron ore orogeny from the later Satpura orogeny; in the latter, chromiferous ultrabasic rocks are very few in number.

Banerjee (1972) described the Sukinda Valley chromites as Alpine type whereas Chakraborty and Chakraborty (1984) opine that the chromiferous ultramafic rocks of Sukinda Valley of Orissa are intrusive into the Iron-ore supergroup (2950-3200 my) at the eastern periphery of the Indian Precambrian shield. Both laterally and vertically, chromite occurs as persistent layers, lenses or pockets in the serpentinised and silicified dunite-peridotite extending over a strike length of 25 km. The ultramafic rocks and the chromitite layers are co-folded with the Iron-ore supergroup into a plunging syncline. Primary layering, ball and pillow structures, cross laminations, graded bedding etc. can be detected and he concluded that the chromite deposits are predominantly stratiform in nature and were presumably formed in situ by crystal settling, the layering having been accentuated by the fluctuation of fO2. The geological features suggest a single magmatic cycle.

The dunite – peridotite underlying the Sukinda valley formed as olivine – rich cumulate from an ultramafic magma, which was presumably derived through partial fusion of subcrustal or mantle material at great depth and was subsequently emplaced into the Precambrian iron ore supergroup of metasediments as a concordant sheet. A hydrous magnesian melt could exist at a temperature of 1000° C and serpentinization took place below 500° C. The absence of gabbroic rocks, their intermediate and acid differentiates and the lack of rock association of true ophiolite complex in the near
vicinity of Sukinda valley all rule out the possibility of dunite – peridotite being formed by fractional crystallization of a basaltic magma.

Varadarajan and Narendra (1986) considered that the chromite deposits in the zoned ultramafic-mafic complexes along the Indus suture zone, Ladhak, Himalaya are formed by magmatic crystallization and gravity settling under condition of magmatic currents. They also opine that these chromite deposits are comparable to chromites from Alaska, British Columbia and Urals and they are distinct from either podiform or stratiform complexes.

Dessai et. al., (1992) considered that the chromite deposits in mantle peridotites from Indus ophiolite belt of Ladhak, Himalaya, that they are Podiform chromites.

According to Srikantappa et. al., (1980) the chromite and minor V-Ti magnetite-bearing ultramafic-mafic complex of the Sinduvalli area of Sargur schist belt extends for 22 km with an average width of 200 m. The ultramafic rocks represented by dunite, harzburgite peridotite and pyroxenite form a distinct but conformable unit within the surrounding basement gneisses, which have a regional strike of about 15°. The Sargur schist complex comprises a portion of the high-grade gneissic terrain of Karnataka and contains a variety of ultramafic and associated basic rocks and enclaves of metasediments is consisting of orthoquartzite, banded ironstones, carbonates as well as K-rich pelitic rocks.

The chromites also exhibit relict igneous structures viz., rhythmic layering, size graded layering, troughs due to erosion and textures such as cumulus, occluded silicate and chain structures. There is also some indication that there may be relict chemical features in that cryptic layering may be present. The igneous structures textures and
chemistry suggests that the ultramafic rocks are of stratiform origin. It follows that some of the gabbroic rocks (and the anorthositic rocks as well) belong to a now dismembered, layered, stratiform complex.

The chromitite layers of Sinduvalli show the imprint of several deformational episodes. This has resulted in the formation of westerly dipping planar fabric, which are superimposed upon the way-up structures demonstrated by the rocks. Preliminary olivine-spinel geothermometry suggests that the re-equilibration has proceeded during conditions of regional metamorphism and has been considerably affected by a late stage post-tectonic hydrothermal event connected with magnesite formation.

Naganna et. al., (1978) based on the relict primary textures such as occluded silicate texture and clot texture, concluded that the chromite deposits of Nuggihalli schist belt as well as Kadakola – Sinduvalli area are podiform deposits.

5.5.2 Genetic models on V-Ti magnetite deposits: A Review

A review of genetic models of V-Ti magnetite ores of different igneous complexes of the world is given in the subsequent pages. The mode of occurrence and genesis of V-Ti magnetite ores is one of the most important topics, which are debated since a long time in the field of economic geology. From the beginning of the 20th century, many workers have carried out detailed investigation on mineralogy, textures, relation of V-Ti magnetite ores with host rocks and their genesis by Singewald (1912), Ramdhor (1969), Osborne (1928), Kamiyama (1929), Hall (1932), Bateman (1951), Vincent and Philips (1954), Buddington and Lindsley (1964), Lister (1966), Willemse (1969), Haggerty (1976), Naganna et. al., (1977), Cawthorn and Mc Carthy (1981), Reynolds (1985), Chakraborty et. al., (1988), Govindaiah et. al., (1989), Mohanthy and Sahoo
All over the world the Ti-magnetite ores with varying amounts of vanadium content are associated with gabbro-norite-anorthosite complexes such as Bushveld igneous complex of South Africa, (Willemse 1969, and Reynolds 1985) and Skaergaard layered intrusion of Greenland (Vincent and Philips 1954; and Vincent 1960). Further these deposits have also been reported to occur as irregular veins and lenses, stringers and impersistent layers in non-layered massifs gabbro-anorthosite intrusives e.g., Adirondack, (Buddington et. al., 1963) and Duluth gabbro of Minesotta, USA; Allard lake and St. Charles anorthosite plutons of Canada (Lister, 1966).

In India, the V-Ti magnetite deposits associated with gabbro-anorthosite complexes have been studied from different areas by different workers. They include Dublabera-Katwar pahar areas of Singhbum district, Bihar (Dunn and Dey, 1937) south of Guru Mahisini pahar area in Mayurbhanj district of Orissa (Roy, 1955), Nausahahi in Keonjhar district, Orissa (Chakraborty, 1959a and 1988) and also from Salika area in the Bankura district of West Bengal (Bose and Roy, 1966).

In Karnataka, the V-Ti magnetite deposits have been reported and described from Nuggihalli greenstone belt (Chakraborty, 1961, Naganna et. al., 1975, Nijagunappa and Naganna, 1983, Vasudev 1984), Devanarasipura (Naganna et. al., 1977), and Masankere areas of Shimoga green stone belt (Channappa and Subramanya, 1973, Chayapathi 1976, Ramiengar et. al., 1978, Vasudev and Sreenivasan, 1979, Govindaiah et. al., 1986, Pathan and Govindaiah, 1986) and in many parts of North Kanara schist belt such as Mulemane, Ankola and Angadibail (Vasudev and Ranganathan, 1994).
Katwar Pahar deposits of Bihar, Nuggihalli, Devanarasipura and Masanikere deposits of Karnataka have been described as layered type, occurring in gabbro-anorthosite complexes spatially with ultramafic rocks (Vasudev and Sreenivasan, 1979, Nijagunappa and Naganna 1983, Pathan and Govindaiah, 1986). The genetic aspects of other occurrences in India are poorly documented.

Reynolds (1985) has classified the genetic models for the concordant V-Ti magnetite layers in the Bushveld complex and other gabbroic intrusions into (i) earlier genetic models and (ii) later genetic models

**5.5.2.1 Earlier genetic models:** The earlier genetic models acknowledge the existence of genetic links between concordant layered Fe-Ti oxide ores and their host rocks and invoked mechanism involving either the separation and accumulation of Ti magnetite crystals from which the ores crystallized.

Molengraaff (1904) opined that the concordant V-Ti magnetite layers in the Bushveld complex and other gabbroic intrusions represent magmatic segregation deposits.

Wagner (1928), in contrast to the views expressed by Osborne (1928) suggested that Ti magnetite crystals on the consolidated floor of magma chamber.

Hall (1932) opined that a dense ‘magnetite’ liquid might separate from crystallizing magma and accumulate on the floor of the magma chamber. Any unconsolidated plagioclase crystals on the floor of the chamber would float upward through the oxide liquid and became concentrated towards the top of such layers, thereby accounting for their plagioclase rich gradational upper contacts. Subsequent crystallization of these
magnetite liquids then gave rise to ore-rich layers.

Bateman (1951), elaborate on the ideas proposed by Hall (1932), concluded that concordant Fe-Ti ores were formed by late gravitative accumulations of oxide liquids on top of the consolidated pile. This liquid might also have been injected into the country rocks to form discordant bodies.

Lister (1956), while discussing origin of some selected Fe-Ti oxide ores concluded that the concordant Fe-Ti oxide ore layers are formed by separation and later accumulation of an immiscible Fe-Ti oxide liquid.

Coertz (1966), in contrast to the views proposed by Lister, (1966) suggested that the Bushveld Ti magnetite layers were derived from the crystallization of an oxide liquid that was intruded from below and spread laterally along certain lithological contacts.

Philpotts (1967), Kolker (1982) and Reynolds (1983a) opined that separation and accumulation of immiscible Fe-Ti oxide liquid remains an attractive hypothesis because it can explain many of the field and textural relationships of the Fe-Ti oxide ores.

5.5.2.2 Later genetic models: All later genetic models virtually invoke mechanisms whereby episodic increase in oxygen fugacity trigger the crystallization of copious quantities of Ti magnetite that are required for the development of ore-rich layers. The formulation of these models was proposed by experimental studies of synthetic Fe-bearing systems, which indicated that oxygen fugacity is an important controlling factor in the crystallization of magnetite and Fe-bearing phases (e.g., Muan and Osborne, 1956; Osborne, 1959; Roeder and Osborne, 1966). These studies indicate that the Fe$_2$O$_3$/FeO ratios of magmas are, to a larger extent, dependent on oxygen
fugacity and that magnetite precipitation would be enhanced by an increase in the amount of Fe$^{3+}$ in the melt.

Molyneux (1970a and 1970b), proposed that crystallization of Ti magnetite might take place near the floor of the magma chamber in response to increased oxygen fugacity. Hill and Roeder (1974), demonstrated experimentally that precipitation of Ti magnetite from basaltic liquid is largely controlled by oxygen fugacity and that it is always a late crystallization phase. They also confirmed Osborne (1959) earlier observation that marked Fe-enrichment of the residual liquid will occur during prolonged fractional crystallization under low or decreasing oxygen fugacity conditions.

Irvine (1975), suggested that Ti magnetite rich layers might develop in response to episodic oxidation events that were caused by extensive contamination of the basic parental magma by granitic liquid from melting of sialic roof rocks.

Klemm et al., (1982), have postulated that the formation of Ti magnetite layers might result from higher oxygen fugacity caused by the presence of volatiles derived from floor rocks that are undergoing thermal metamorphism below the intrusion.

Reynolds (1985), proposed that the expulsion of highly fractionated interstitial liquid from the semi-consolidated crystal pile into the base of the overlying magma column might represent an effective mechanism for the introduction of additional oxygen into the system. This liquid would be gradually expelled from the crystal pile during the upward displacement of part of the intercumulus phase during solidification in response to compaction, infiltration metamorphism (Irvine 1978, 1980a) and/or annealing (Reynolds, 1979a).
Reynolds (1985b), proposed a genetic model to explain the origin of Ti magnetite layers of BIC, based on the petrological, mineralogical, geochemical and field characteristics of these layers and the host rocks as (i) a lengthy period of fractional crystallization of basaltic magma, probably involving several influxes, finally resulting in the concentration of substantial amounts of Fe, Ti and V in the late stage residual magma, (ii) the crystallization of abundant plagioclase leads to an increase in the total Fe content and concomitant increase in the density of the residual liquid which may accumulate to form a stagnant layer on the floor of the magma chamber. This dense layer will not mix with the overlying magma and this providing a suitable environment for crystallization of Ti magnetite and (iii) The ore-rich layers develop largely by in situ bottom growth of Ti magnetite crystals. Densification and subsequent conversion of ore-rich layers into monomineralic rocks is accomplished by annealing at elevated subsolidus temperatures.

Govindaiah et. al., (1986) based on the field setting, mineralogical and geochemical features of ultramafic-mafic rocks and their associated V-Ti magnetite deposits of the Masanikere layered complex of Dharwar craton, concluded that the genesis of V-Ti magnetite layers/bands is intimately related to fractional crystallization process that were responsible for the formation of their gabbro-anorthosite host rocks. The distribution and geologic relationship of these ores clearly indicate that they are the integral components of the layered sequence of gabbro-anorthosite suite of the area. The chemical and elemental variations indicate that the members of the Masanikere complex are the result of fractional crystallization of iron-rich tholeiitic magma, indicating
gravitative settling of Ti magnetite crystals and the later injection into the metasediments of the Chitradurga group. The layered nature of the V-Ti magnetite ores, their cumulus and close association with ultramafic–mafic suite of rocks and their tectonic setting suggest that the Masanikere complex is not of Alpine type, but a part of stratified igneous complexes.

According to Kariyanna et al., (1996) the V-Ti magnetite deposits and associated sulphides in the Durgagudda Upper-Kaneri layered complex of Joida area of Dharwar craton, their genesis and evolution can be summarized as follows: -

(i) the V-Ti magnetite layers / bands of the study area were formed by fractional crystallization of V-Fe-Ti – rich magma of tholeiitic composition, probably involving several influxes leading to the concentration of substantial amount of Fe, Ti and V in the late stage residual magma. According to them, development of discrete layers and concentration of Fe-Ti oxides may be attributed to in situ bottom crystallization in response to episodic increase on fO₂.

(ii) the development of polygonal ilmenite-magnetite aggregates with gently curved grain boundaries and well defined triple junctions indicate that extensive annealing of crystalline assemblage took place at subsolidus temperature. However, true equilibrium was rarely attained in the present case.

(iii) the Ti magnetite exhibits a wide range of ulvospinel, pleonaste and ilmenite-bearing intergrowth. The development of these minerals can be related to ambient fO₂ at various stages of subsolidus cooling. Consistently, low fO₂ gave rise to well develop ulvospinel exsolution intergrowth of magnetite and ilmenite on cooling to below the magnetite – ulvospinel solvus. Higher fO₂ resulted in the formation of a wide range of ilmenite-bearing intergrowths that range from external granules through various
lamellar types (iv) disseminated sulphides represented by pyrite, chalcopyrite, cubanite sphalerite and covalite blebs/inclusions of chalcopyrite in magnetite may indicate their formation from immiscible liquids (v) late stage ductetic alteration resulted in variable degrees of oxidation of alteration of oxide bearing assemblages. The most commonly encountered feature is the extensive replacement of magnetite by chlorite in the study area.

Govindaiah et. al., (1997) concluded that the V-Ti magnetite deposits of the Durgagudda Upper-Kaneri layered complex are formed between 590° C and 875° C with $fO_2$ ranging between $\log_{10} = -22.00$ to $-13.50$.

Vinayaka et. al., (1999) opine that the chromite-V-Ti magnetite ore bodies are stratiform in nature and chromite is formed in situ by early magmatic segregation and the Ti magnetite with exsolved phases of ilmenite, ulvospinel, pleonaste is formed by late magmatic residual segregation through in situ bottom crystallization in response to episodic increase in $fO_2$.

5.6 GENESIS, EVOLUTION AND EMPLACEMENT OF ULTRAMAFIC-MAFIC ROCKS AND THEIR ASSOCIATED CHROMITE V-Ti MAGNETITE DEPOSITS OF THE STUDY AREA

Any genetic model for the genesis, evolution and the emplacement of ultramafic-mafic rocks and their associated chromite V-Ti magnetite deposits of the study area should satisfactorily explain a number of intriguing problems such as; (a) the origin and composition of parent magma and crystallization of the underlying ultramafic rocks and their associated chromites and the overlying gabbro-anorthosite suite of rocks and their associated V-Ti magnetite layers (b) nature of igneous intrusion weather alpine
or stratiform type and the rock association (c) mineralogical, textural and geochemical variation in chromites and V-Ti magnetites and their host rocks and (d) mode of emplacement of the chromite and V-Ti magnetite-bearing ultramafic-mafic rocks.

The field, petrological, mineralogical, textural and geochemical data for ultramafic rocks and their associated chromite ores, gabbro-anorthosite suite of rocks and their associated V-Ti magnetite ores, indicate the genesis of chromite–V-Ti magnetite layers is intimately related to fractional crystallization processes that were responsible for the formation of their host ultramafic–mafic rocks. The chromite–V-Ti magnetite ore bodies are the integral components of the layered ultramafic-mafic sequence as indicated by the occurrence of chromitite layers in rhythmic fashion, cumulus texture both in chromites and V-Ti magnetite ores and their associated ultramafic-mafic rocks, the chemistry of chromite, which is typical of the stratiform igneous bodies such as Bushveld Igneous Complex and Stillwater complex, Sinduvalli area and the chemistry of Ti magnetite-ilmenite pairs in the V-Ti magnetite ores and their similarity in temperatures of equilibration and fO₂ conditions with those of Bushveld Igneous Complex, Boula-Nausahi of Orissa, Joida area of Dharwar craton, Skaergaard intrusion of Greenland, layering in V-Ti magnetites, gradational contacts with the gabbros and gabbroic anorthosites and their associated ultramafics.

The mineralogical, textural and chemical data indicate that the members of the ultramafic–mafic complex of the study area are the result of fractional crystallization of Cr–Fe–Ti–V-rich tholeiitic magma, involving several influxes under low to moderate pressure in chromite–olivine–pyroxene–plagioclase–Ti magnetite order giving cumulus character of the ultramafic–mafic rocks.
The cumulus texture and compositional banding in chromitite ores indicate that chromite is formed *in situ* before or contemporaneously with olivine (as indicated by textural features) by early magmatic differentiation and gravity settling. The layering in the chromitite might have been developed by the fluctuation of fO$_2$ during fractional crystallization, a mechanism proposed for the formation of chromitite layers of by Cameron and Desborough (1969) as well as Chakraborty and Chakraborty (1984) for the Sukinda valley chromite deposits of Orissa. The temperature estimation based on the olivine-spinel thermometry in the area lie between 460º C to 500º C thus indicating re-equilibration of chromite-bearing ultramafic rocks at subsolidus temperatures during amphibolite facies metamorphism around 475 to 610º C at 2-3 K bars as indicated by the metamorphic minerals in the ultramafic rocks, mineral chemistry of plagioclase-hornblende in the metagabbros.

The primary chromite has been subjected to alteration and exhibit three combinations of phases such as (i) chromite – ferritchromit - magnetite, (ii) chromite - ferritchromit and (iii) chromite - magnetite. It is suggested that alteration of chromite occurred in two stages. Formation of ferritchromit represents the first stage and formation of magnetite is the last stage of alteration. Chemistry of the altered phases of chromite of the study area indicates that there is decrease in Cr, Al and Mg and increase in total Fe from parent chromite (core) to ferritchromit (inner rim) and to magnetite (outer rim). This zonation is a replacement process in which Fe is substituted for Mg, Al and to a lesser extent Cr. This replacement has taken place during serpentinization in the lower amphibolite grade metamorphism as suggested for similar chromite ores else where (Roeder, 1994). The chromium released from chromite during alteration, entered into
antigorite and chrome-chlorite, Cr-serpentine during prograde metamorphism. Another prograde metamorphic alteration of chrome is re-crystallization of exsolved ilmenite lamellae into coarser ilmenite inclusions and rounded segregated grains, which took place during lower amphibolite grade metamorphism.

The Ti magnetite ores as well as the magnetite-gabbro exhibit cumulus texture and compositional banding indicating the enrichment of Fe, Ti and V in the residual liquids. The plagioclase which is the chief constituent of gabbro-anorthosite suite of rocks and the in situ bottom crystallization of plagioclase might have lead to increase in the total Fe content and a concomitant increase in the density of the residual liquids which may have accumulated to form stagnant layers as proposed by Reynolds (1985).

Densification and subsequent conversion of ore-rich layers into essentially monomineralic rocks is accomplished largely by annealing at elevated subsolidus temperatures between 580°C to 650°C and augmentation of the expelled liquid from the underlying crystal pile during infiltration metasomatism. This lead to the expulsion of high proportion of interstitial liquid and formation of essentially of monomineralic Ti-magnetite layers that exhibit polygonal grain boundaries and triple junction. Development of ilmenite and pleonaste has taken place during subsolidus cooling.

The present investigation confirms the views of the earlier workers (Govindaiah and Pathan, 1996; Devaraju et. al., 2004 and Paranthaman and Vidyadharan, 2005) that the ultramafic - mafic complex of the study area is a part of the layered igneous complex.

The present study shows that both the chromite bearing ultramafic rocks confined to western portion and V-Ti magnetite bearing gabbro – anorthosite suite of rocks confined western portion of the study area are co-magmatic and
derived from a common parent magma of Cr – Fe – Ti – V rich tholeiitic composition involving several influxes under low to moderate pressure. The present study also indicates that this layered complex is similar to other layered igneous complexes like Bushveld Igneous Complex of South Africa, Nuggihalli ultramafic - gabbro complex and Sindhivalli ultramafic – mafic complex in the Sargur area of Dharwar craton and Mayurbunj basic intrusion of Orissa.