by Paul T. Robinson ^{1,2}, Mei-Fu Zhou ², John Malpas ² and Wen-Ji Bai ³

Podiform Chromitites: Their composition, origin and environment of formation

1. Centre for Marine Geology, Dalhousie University, Halifax, Nova Scotia B3H 3J5 Canada

2. Department of Earth Sciences, The University of Hong Kong, Pokfulam Rd., Hong Kong, China

3. Institute of Geology, Chinese Academy of Geological Sciences, 26 Baiwanzhuang Rd, Beijing 100037, China

Podiform chromitites are normally either high-Cr or high-Al in composition and typically only one type occurs in any given ophiolite. Despite some superficial similarities, high-Cr and high-Al chromitites differ significantly in mineral and bulk-rock geochemistry. Many podiform bodies have been deformed and recrystallized by continued mantle flow but their magmatic origin is widely accepted. We suggest that podiform chromitites occur in the upper parts of ophiolitic or island arc mantle and are formed by reaction between rising melts and the host peridotites. High-Cr and high-Al ores are slightly different in their environment of formation.

Introduction

Chromitite orebodies occur in two distinct forms: stratiform and podiform. Stratiform bodies are areally extensive layers or bands in mafic layered intrusions. Their mineralogies and textures leave little doubt that they crystallized from mafic magmas probably by mixing of evolved and more primitive melts (Irvine, 1977). Classic examples of stratiform chromitites include those in the Muskox, Stillwater and Bushveld layered intrusions.

Podiform chromitites, on the other hand, typically occur as lenticular or pod-shaped bodies, a few meters or tens of meters in length, which are hosted in depleted peridotites of presumed mantle origin. These bodies are individually small but, where abundant, they form large-scale chromitite deposits. The origin and tectonic environment of formation of these deposits have been topics of investigation for many years. Although some features of podiform chromitites remain enigmatic, recent work suggests that they form by melt-rock reaction as mafic magmas pass through, or are ponded in, segments of the upper mantle and that their compositions reflect varying degrees of partial melting of the parental peridotites.

Petrography of podiform chromitites

Podiform chromitites are typically lens-shaped although many occur as tabular or pencil-like bodies. Many bodies are concordant or sub-concordant to the foliation in the surrounding peridotites and exhibit fabrics interpreted to be the result of solid-state flow and recrystallization at relatively high temperatures (Cassard et al., 1981). Most podiform chromitites are hosted in depleted harzburgite, although some occur in Cpx-bearing harzburgite or lherzolite.

Most podiform chromitites, whether hosted in harzburgite or lherzolite, are enclosed in dunite pods or have dunite envelopes several centimeters to a meter thick (Thayer, 1964; Thayer, 1969; Leblanc, 1987; Zhou et al., 1996). These dunite envelopes grade outward into the surrounding harzburgite or lherzolite with increasing

modal proportions of pyroxene. Where lherzolite is the host, a band of harzburgite typically lies between the lherzolite and the dunite.

Massive, disseminated, nodular and brecciated chromitites are the most common textural types (Figure 1A-D, see inside back cover) and these commonly grade into each other. Massive chromitites typically have sharp contacts with the enclosing dunite whereas disseminated bodies grade outward into the surrounding rock, in some cases passing into interlayered chromitite and dunite.

Massive chromitites are typically coarse-grained rocks composed of more than 80 volume % chromite. Individual chromite grains are mostly 1 to 5 mm across (some are up to 30 mm), anhedral in shape, and show the effects of corrosion or reaction in addition to mechanical disruption. Textural relationships suggest that the coarse size of most massive chromitites is a primary magmatic feature and not due to recrystallization. The chromite grains are closely packed, leaving very narrow interstices filled with silicate minerals (Thayer, 1964; Greenbaum, 1977).

Disseminated chromitites consist of 25-80 volume % chromite grains, more or less uniformly scattered in the host silicates (mainly olivine and orthopyroxene) (Figure 1C-D). Individual chromite grains are generally smaller (0.5-2 mm) and more regular than in the massive varieties. Many disseminated chromitites grade into massive bodies with increasing chromite contents and larger crystal sizes.

Nodular varieties consist of small (0.3 to 2.5 cm), spherical or elliptical aggregates of chromite in an olivine-rich matrix that may itself have disseminated chromite (Figure 1C-D). In some cases, individual nodules may be crudely oriented parallel to their long axis. Closely associated with nodular chromitites are orbicular types which have a solid, nodule-like nucleus of chromite surrounded by a mantle of olivine and an outermost rim of fine-grained chromite plus olivine. Both types can grade into massive chromitites with increasing amounts of chromite.

Banded chromitites, with rhythmically alternating chromite-rich and olivine-rich layers, are present in many chromitite pods. Such banding usually parallels the long axes of the chromitite pods or lenses, which in turn may parallel the foliation in the host peridotites. Many bands show small-scale folding, faulting and cataclasis. In brecciated chromitites disseminated chromite typically fills spaces between dunite fragments (Figure 1D).

Olivine (Fo 92-97) is the most common matrix mineral, although many chromitites have small amounts of orthopyroxene (En 94-96) and clinopyroxene. In some cases, olivine and orthopyroxene are partly to completely replaced by serpentine, and cracks and veinlets are filled with carbonate and garnet.

Inclusions of olivine, clinopyroxene, orthopyroxene, albite, Fe-Ni alloy and sulfides are present in many chromite grains (e.g., Lorand and Ceuleneer, 1989). In addition, chromites in several bodies, such as those in Al Ays of Saudi Arabia, Troodos in Cyprus, New Caledonia, and Oman (Johan et al., 1983; McEllduff and Stumpfl, 1991), contain volatile- and alkali-rich mineral and fluid inclusions. The presence of volatiles can also be deduced from the occurrence of pegmatitic chromitites and associated pegmatitic gabbro and troctolite. The abundance of fluid inclusions and hydrous

Figure 2 Cr# versus Mg# for podiform chromites. High-Al-chromitites have Cr#s of 60 or less whereas high Cr-varieties have Cr#s of 70 or higher. Data are from Luobusa, Dongqiao, Sartohay, Hegenshan Solon, Yushi and Hongguleleng ophiolites of China.

minerals in chromite grains indicates relatively high water content in the parental magma which, in turn, suggests formation in a supra-subduction zone environment. In such an environment H₂O, K₂O-rich fluids, and possibly silicate melts derived from the subducted slab can be added to the overlying mantle wedge (Ringwood, 1974; Pearce et al., 1984; Roberts, 1988).

Typically, many of the silicate minerals, particularly olivine and pyroxene, are too magnesian to have crystallized from a melt and probably owe their composition to Mg-Fe subsolidus exchange between silicates and chromite (Irvine, 1967; Roeder et al., 1979; Lehmann, 1983).

Table 1 A comparison of high-Cr and high-Al chromitites from the Luobusa and Sartohay ophiolites of China (after Zhou and Robinson, 1994)

	High-Cr	High-Al
Chromitite	Cr#>70	Cr#<60
	TiO ₂ <0.2wt.%	TiO ₂ >0.2 wt.%
Peridotite	Harzburgite	Lherzolite
	TiO ₂ <0.05 wt.%	TiO ₂ >0.8 wt.%
	MgO=42-51 wt.%	MgO=36-48 wt.%
	Ca=0-1.5 wt.%	Ca=0-3.4 wt.%
	Al=0.01-1.20 wt.%	Al=0.3-2.20 wt.%
Mineral chemistry		
Olivine (Fo)	90-93	86-91
Opx (En)	91-92	87-88
(Al ₂ O ₃)	0.71-4.13 wt.%	2.29-5.98 wt.%
(CaO)	0.48-1.54 wt.%	0.92-4.34 wt.%
Cpx (En)	47-53	43-51
(Al ₂ O ₃)	2.75-4.72 wt.%	2.21-7.68 wt.%

Chromite compositions and parental magmas

Podiform chromitites are normally either high-Cr or high-Al in composition (Figure 2) and typically only one type occurs in any given ophiolite. In a few cases, such as the Zambales ophiolite of the Philippines, both high-Cr and high-Al varieties are present but they occur in two distinct lithologic associations and are believed to have crystallized from two compositionally distinct parental magmas (Hock et al., 1986; Yumul, 1992; Yumul and Balce, 1994).

Despite some superficial similarities, high-Cr and high-Al chromitites differ significantly in mineral and bulk-rock geochemistry (Table 1). High-Cr chromitites are characterized by Cr#s (100Cr/(Cr+Al)) greater than 70, by very low bulk rock TiO₂ and by highly magnesian olivine and orthopyroxene. These chromitites are typically hosted in highly depleted harzburgites (Table 1).

High-Al varieties, on the other hand, have Cr#s less than 60, have higher bulk-rock TiO₂ and less magnesian silicate minerals. The high-Al chromitites are typically hosted in less depleted Cpx-bearing harzburgites or lherzolites. Both types have similar shaped chondrite-normalized PGE patterns, enriched in Os, Ir, and Ru relative to Rh, Pt, and Pd (Zhou and Robinson, 1994). However, high-Cr varieties typically have somewhat higher Ir, Ru, and Rh contents than high-Al varieties (e.g. Zhou and Robinson, 1994).

Figure 3 TiO₂ versus Cr# of high-Cr and high-Al podiform chromitites. MORB and boninite fields are from Arai (1992). Data sources are the same as in Figure 2.

Experimental studies have shown that in peridotites, the degree of depletion can be related to the degree of partial melting that the rock has undergone. As the degree of melting increases, the residual minerals become depleted in Fe, Al, Ca, and Ti, the Mg#s of residual olivine and orthopyroxene increase, and the Al₂O₃ content of orthopyroxene decreases (Mysen and Kushiro, 1977; Jaques and Green, 1980; Michael and Bonatti, 1985).

Variations in the chrome content of accessory and ore-forming chromite also are believed to be related to the degree of partial melting with Cr#s increasing with increasing degrees of melting (Irvine, 1967; Arai, 1980; Dick and Bullen, 1984). Typically, high-Al chromitites plot in the MORB field whereas high-Cr varieties plot in the boninitic field (Figure 3) (Arai, 1992; Zhou and Robinson, 1994). Thus, high-Cr chromitites, hosted in highly depleted harzburgites, are believed to have crystallized from boninitic melts formed by relatively high degrees of partial melting whereas high-Al varieties, hosted in less depleted Cpx-bearing harzburgite and lherzolite, are believed to crystallize from tholeiitic melts formed by lower degrees of melting (Roberts, 1992). This does not mean that the mantle sections from which the melts were derived were the same composition

originally or that the differences in mineral chemistry are caused entirely by the most recent episode of partial melting. The same degree of depletion and the same chromite compositions could be achieved by recycling mantle sections that had undergone several episodes of partial melting (Zhou and Robinson, 1994).

Origin of podiform chromitites

Textural and mineralogical evidence indicate that podiform chromitites form by crystallization of mafic magmas in the upper mantle (Brown, 1980; Cassard, et. al., 1981; Lago et al., 1982). Cumulate textures, cyclical layering, and silicate gangue minerals are all evidence of a magmatic origin and current models assume formation in magmatic conduits cutting the mantle peridotites (e.g., Leblanc and Ceuleneer, 1992). Many podiform bodies have been deformed and recrystallized by continued mantle flow but their magmatic origin is widely accepted.

What is not so well understood is the mechanism by which chromite is precipitated. Upper mantle melts are generally basaltic in composition and it is not clear how chromite, as a minor component of such melts, can be concentrated to form massive or disseminated deposits. To form monomineralic layers or pods, chromite must have crystallized alone for a period of time and/or have been concentrated by physical processes. Likewise, it is not clear why chromitites are abundant in some mantle sections and totally absent in others (Roberts, 1988).

Johan et al. (1983) and Auge (1987) suggested that chromium can be extracted from magmas by a fluid phase but there is no relevant experimental evidence to support this interpretation. Contamination by wall rock assimilation was proposed for chromitite formation in the Muskox (Irvine, 1975) and Kemi (Alapieti et al., 1989) layered intrusions because alkali-rich inclusions occur in chromites from these bodies. Other models focus on migration of melts from the silicate-chromite cotectic into the primary phase volume of chromite. Ulmer (1969) and Murck and Campbell (1986) emphasized the importance of oxygen fugacity in the crystallization of chromite, and Cameron and Desborough (1969) suggested that contamination of the magma by gaseous emanations from the country rock could increase its oxygen fugacity, leading to the formation of monomineralic chromite horizons. Cameron (1980) subsequently abandoned this hypothesis in favor of chromite crystallization in

response to changes in total pressure, a mechanism also invoked to explain chromite layers in the Stillwater complex (Lipin, 1993). However, the magnitude of the pressure change required to form a 1-m-thick chromitite layer would be unrealistically large (>1 kbar) and, in general, the effect of pressure on mineral stability is small (Hatton and von Gruenewaldt, 1987; Campbell and Murck, 1993). The presence of chromitite layers in stratiform mafic intrusions strongly supports the interpretation that the parental magmas are basaltic in composition. Irvine (1977) and Irvine and Sharpe (1986) suggested that chromite precipitation from mafic melts in stratiform intrusions could be triggered by mixing of evolved and primitive magmas. This model is supported by textural evidence for common cotectic crystallization of chromite and olivine and is compatible with the observed sequence of layers in stratiform complexes (Sharpe and Irvine, 1983; Irvine and Sharpe, 1986; Murck and Campbell, 1986; Roeder and Reynolds, 1991; Campbell and Murck, 1993). We have used a modified version of Irvine's model to explain the precipitation of podiform bodies (Zhou and Robinson, 1994; Zhou et al., 1996).

Most podiform chromitites have dunite envelopes which grade outward into harzburgite or lherzolite. These envelopes are believed to have formed by melt-rock reaction in the same manner as dunites around many dikes in peridotite (Dick, 1977; Quick, 1981; Bodinier et al., 1990; Kelemen, 1990; Kelemen et al., 1992; Kelemen and Dick, 1995; Dick and Natland, 1995; Edwards and Malpas, 1995). As batches of relatively primitive, mafic magma rise through the host peridotite, pyroxenes are dissolved from the wall rock, leaving behind a dunite residue a few centimeters to a few meters thick.

The characteristic occurrence of dunite envelopes around podiform chromitites suggests a genetic relationship between the two. Removal of pyroxenes from the host peridotite by melt rock reaction would increase the silica content of the magma as demonstrated by experimental results (Fisk, 1986; Kelemen et al., 1990). Such an increase in silica would move the melts from the olivine-chromite cotectic into the primary phase field of chromite (Figure 4).

Once the melt has moved into the chromite stability field and chromite starts to precipitate, the melt would migrate toward the olivine-chromite cotectic, where olivine and chromite would again precipitate together (Figure 4). Batches of new magma rising through the conduit would mix with the fractionated melts, moving them back into the chromite stability field.

The interplay of melt-rock reaction, chromite fractionation and magma mixing would lead to many fluctuations in melt composition, producing both massive and disseminated chromitites and phase layering within a single podiform body. Continuous melt-rock interaction would result in the precipitation of chromite alone to form massive ores. If the magmas and host rock reached equilibrium and interaction ceased, chromite and olivine would precipitate together to form disseminated ores. Nodular ores probably form by a combination of chromite precipitation and convective overturn of the magma pocket.

Figure 4 Phase relations in the system olivine-quartz-chromite as determined by Irvine (1977). A primitive magma (A) on the olivine-chromite cotectic would react with the peridotite host to form a range of more silicic compositions along a line such as A-X, all of which lie in the chromite field. Precipitation of chromite would drive the melts along lines parallel to XY, moving them again toward the olivine-chromite cotectic. Batches of new mafic melt with a composition similar to A could mix with the fractionated magma along a line such as AY, thus moving it again into the chromite field. (after Zhou, et al., 1996).

Tectonic setting of podiform chromitites

The abundance of volatile- and alkali-rich mineral and fluid inclusions in chromite grains (Johan et al., 1983; McElduff and Stumpfl, 1991) indicates relatively high water contents in the parental magmas which, in turn, suggests formation in a supra-subduction zone environment. In such an environment H₂O, K₂O-rich fluids, and possibly silicate melts derived from the subducted slab can be added to the overlying mantle wedge (Ringwood, 1974; Pearce et al., 1984; Roberts, 1988). The melts from which chromitites precipitate are typically boninitic or tholeiitic with island arc characteristics (e.g., Malpas and Robinson, 1987; Roberts, 1988; Yumul and Balce, 1994; Zhou et al., 1996).

We suggest that podiform chromitites occur primarily in supra-subduction zone mantle sections and that their chemical compositions can be correlated with formation in different tectonic settings,

specifically island arcs and nascent spreading centers, such as back-arc basins (Zhou and Robinson, 1994).

Beneath island arcs, there is a high input of volatiles from the subducted slab into the overlying mantle wedge, leading to high degrees of partial melting and the formation of boninitic or highly depleted tholeiitic melts. In these environments, the old lithospheric mantle immediately beneath the crust would be invaded by large volumes of melts generated at depth by melting of depleted harzburgites (Figure 5). The new magmas would rise through and react with the old mantle, producing pods of high-Cr chromitite with dunite envelopes. Little spreading would be expected in such environments and the melts would tend to pond in the upper mantle, thus enhancing melt-rock interaction. If spreading did occur, the old lithospheric mantle would be thinned and eventually removed.

However, some podiform chromitites could still form by melt/rock reaction because melt compositions in these environments may also change with time. For example, in the Troodos ophiolite of Cyprus, the melts became increasingly depleted with time, changing in composition from relatively enriched andesites and dacites, to depleted tholeiites, to highly depleted boninites (Robinson et al., 1983). Such depleted melts presumably form at depth either by increased degrees of partial melting or by remelting of a progressively depleted source (Crawford et al., 1989 and references therein). As the melts separate from their source rocks and move upward through the mantle they are no longer in equilibrium with their host rocks and would react with them.

Ophiolites formed in such environments would be characterized by depleted mantle peridotites produced by high degrees of partial melting. The magmas from which the chromitites crystallized would be more refractory than the host peridotites and would react strongly with them, producing high-Cr chromitites and cumulate sequences of pyroxenite and gabbro. Such bodies would be similar to Papua-type ophiolites that contain low-Al tholeiites, Opx-cumulates and Cpx-free harzburgites (Ishiwatari, 1985).

In nascent spreading centers, such as those in back-arc basin, melts would still be generated in mantle wedges above subduction zones but less volatiles are added to the wedge and the degree of par-

tial melting is lower. The lower degrees of partial melting will produce less depleted magmas from which high-Al chromitites crystallize (Zhou and Robinson, 1994; Yumul and Balce, 1994). Melts generated by partial melting will rise into pre-existing, lithospheric mantle which is thinned during spreading (Figure 5). As the melts pass through the upper asthenospheric and lithospheric mantle they will react with the wall rocks, and precipitate high-Al chromitites. However, these melts are not as far out of equilibrium with their host rocks as those beneath island arcs and less reaction will take place.

The residual peridotites formed in this environment would be moderately depleted Cpx-bearing harzburgites or lherzolites, similar to those of the Yakuno-type ophiolites which are characterized by high-Al tholeiite, and Cpx-type cumulates (Ishiwatari, 1985). Such bodies are similar to MORB ophiolites as defined by Pearce et al. (1984).

High-Al chromitites are associated with troctolite dikes (Figure 1A) that show a crystallization sequence of plagioclase before clinopyroxene (Thayer, 1969; Zhou and Robinson, 1994). Gabbros dredged from the Garret Fracture Zone and troctolites in the mantle of the East Pacific Rise have a similar crystallization sequence which has been explained by formation at relatively low pressures compared to pyroxenite dikes (Elthon, 1979; Hebert et al., 1983; Edwards and Malpas, 1995). Pyroxenite dikes associated with high-Cr chromitites indicate high pressure crystallization from a high-Mg basaltic magma (c.f., Casey et al., 1983; Elthon, 1989). Thus, the associated mafic dikes are consistent with the proposed tectonic scenario for high-Cr chromitites beneath island arcs and high-Al equivalents in back arc basins (Figure 5).

In New Caledonia high-Cr and high-Al chromitites, hosted in harzburgite and lherzolite, respectively, are closely associated along a paleotransform fault (Leblanc, 1995). We suggest that such occurrences reflect formation in a fore-arc environment where hot asthenosphere rising along transform faults as subduction is initiated (Stern and Bloomer, 1992; Malpas et al., 1997).

Beneath mature spreading centers, such as those of mid-ocean ridges, pre-existing lithospheric mantle is thinned and removed during spreading (Figure 5) and MORB magmas remain remarkably

Figure 5 A diagram illustrating proposed relationships between tectonic environments and the formation of podiform chromitites of different composition (after Zhou and Robinson, 1997). Beneath island arcs there are thick sections of old, stagnant lithospheric mantle in which rising melts can pond. The addition of water and other volatiles to the mantle wedge beneath the arc will enhance melting, leading to the production of highly depleted melts. These melts will react with the host peridotites to form abundant podiform chromitites rich in chromium. In back arc basins lower degrees of partial melting produce tholeiitic magmas from which high-Al chromitites crystallize. In this environment, the lithosphere is thinned and eventually removed by spreading so melt-rock interaction is most pronounced in the early stages of spreading. Beneath mature spreading centers, such as mid-ocean ridges, there is no old lithospheric mantle and the MORB melts are equilibrated with the peridotites through which they pass. Thus, no reaction takes place and no chromitites are produced.

constant through time. Thus, magmas formed at depth would be relatively equilibrated with the wall rocks through which they pass and little reaction would take place. Because the magma compositions would not be modified significantly by melt/rock reaction, no large chromitite bodies would form.

Conclusions

Podiform chromitites form in the upper parts of ophiolitic or island arc mantle by reaction between rising melts and the host peridotites. High-Cr chromitites form beneath island arcs where refractory melts, formed by relatively high degrees of partial melting, react with thick sections of old lithospheric mantle. High-Al varieties form in nascent spreading centers, such as back-arc basins, where less refractory melts react with thinned lithospheric mantle. No chromitite bodies would be expected beneath mature spreading centers, such as mid-ocean ridges, because there is no old lithospheric mantle and the rising melts are equilibrated with the depleted peridotites through which they pass.

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Paul T. Robinson is Professor of Earth Sciences, at Dalhousie University, Halifax, Nova Scotia, Canada (currently Visiting Professor in the Department of Earth Sciences, The University of Hong Kong). He is also a member and former Director of the Centre For Marine Geology at Dalhousie University. He received his B.S. degree from the University of Michigan (1959) and his Ph.D. from the University of California, Berkeley (1964). His main research interests are ophiolites and ocean lithosphere and current projects include studies of the lower ocean crust in the Indian Ocean and of ophiolites in China.

Mei-Fu Zhou received his B. Sc. from Nanjing University, China (1983), M.Sc. degree from University of Saskatchewan (1992) and Ph. D. degree from Dalhousie University, Canada (1995). His main research interests focus on the geochemistry and metallogeny of mafic/ultramafic rocks. He is currently in the Department of Earth Sciences, University of Hong Kong, managing the departmental ICP-MS laboratory.

John Malpas is currently Professor and Head, Department of Earth Sciences, The University of Hong Kong. He received his M.A. from Oxford University (1971) and PhD from Memorial University of Newfoundland (1976). He has an extensive research programme investigating the genesis and tectonic setting of ophiolites, ocean island volcanics and aspects of environmental geochemistry. He has been involved for several years with the Ocean Drilling Program.

Wen-Ji Bai, graduated from Changchuan College of Geology, China (1958), has been a research professor in the Institute of Geology, Chinese Academy of Geological Sciences, Beijing. His main research fields are ophiolites and their ore deposits and he has published over 80 papers and 5 monographs in Chinese.