Geodynamic implications of ophiolitic chromitites in the La Cabaña ultramafic bodies, Central Chile

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Abstract

Chromitites (> 80% volume chromite) hosted in two ultramafic bodies (Lavanderos and Centinela Bajo) from the Palaeozoic metamorphic basement of the Chilean Coastal Cordillera were studied in terms of their chromite composition, platinum-group element (PGE) abundances and Re-Os isotopic systematics. Primary chromite (Cr#=0.64-0.66; Mg#=48.71-51.81) is only preserved in some massive chromitites from the Centinela Bajo ultramafic body. This chemical fingerprint is similar to other high-Cr chromitites from ophiolite complexes, suggesting that they crystallised from arc-type melt similar to high-Mg island-arc tholeiites (IAT) and boninites in supra-subduction mantle. The chromitites display enrichment in IPGE (Os, Ir, Ru) over PPGE (Rh, Pt, Pd), with PGE concentrations between 180 and 347 ppb, as is typical of chromitites hosted in the mantle of supra-subduction zone (SSZ) ophiolites. Laurite (RuS_2)-erlichmanite (OsS_2) phases are the most abundant inclusions of platinum-group minerals (PGM) in chromite, indicating crystallisation from S-undersaturated melts in the sub-arc mantle. The metamorphism associated with the emplacement of the ultramafic bodies in the La Cabaña has been determined to be ca. 300 Ma, based on K-Ar dating of fuchsite. Initial $^{187}Os/^{188}Os$ ratios for four chromitite samples, calculated for this age, range from 0.1248 to 0.1271. These isotopic compositions are well within the range of chromitites hosted in the mantle section of other Phanaerozoic ophiolites. Collectively, these mineralogical and geochemical features are interpreted in terms of chromite crystallisation in dunite channels beneath a spreading center that opened a marginal basin above a supra-subduction zone. This implies that chromitite-bearing serpentinites in the metamorphic basement of the Coastal Cordillera are of oceanic-mantle origin and not oceanic crust as previously suggested. We suggest that old subcontinental mantle underlying the hypothetical Chilenia micro-continent was unroofed and later altered during the opening of the marginal basin. This defined the compositional and structural framework in which the protoliths of the meta-igneous and meta-sedimentary rocks of the Eastern and Western Series of the Chilean Coastal Cordillera basement were formed.

Keywords: Chromite; ophiolite; marginal basin; Chile; Chilenia

1. Introduction

The Palaeozoic metamorphic basement of the Chilean south-central Coastal Cordillera contains scattered small bodies of serpentinised ultramafic rocks associated with a heterogeneous assemblage of meta-sedimentary and meta-volcanic rocks (Aguirre *et al.* 1972; Hervé 1974, 1977; Godoy and Kato1990; Willner *et al.* 2005; Glodny *et al.* 2008). These ultramafic bodies and their host rocks have been suggested to be slices of fossil oceanic lithosphere, which were accreted to the Pacific margin of Gondwana (Hervé *et al.* 1981; Forshythe 1982; Glodny *et al.* 2005; Willner *et al.* 2005). The origin of the ultramafic rocks has been the subject of many studies and remains a hotly debated topic with two main hypotheses. An initial survey by Barra *et al.* (1998) of chromitite samples collected from streams draining the largest ultramafic bodies in the area of La Cabaña, about 60 km west of the city of Temuco (Fig. 1), revealed that these chromitites have an island arc geochemical signature. Additionally, Höfer *et al.* (2001) integrated the petrology and ages of deposition and metamorphism of the meta-sedimentary rocks with the mechanism of emplacement of the meta-volcanic and ultramafic rocks from La Cabaña, concluding that these rocks represent portions of oceanic lithosphere formed in a marginal basin developed behind an island arc, sited on the Paleo-Pacific margin of Gondwana. In contrast, Glodny *et al.* (2005) and Willner *et al.* (2005), based on studies of the northern part of the Coastal Cordillera (34°-36° S), have suggested that these are oceanic crustal rocks that were incorporated into the Coastal Cordillera by basal accretion.

These observations and contrasting models raise questions about the true origin of the chromitites and their host ultramafic rocks at La Cabaña, as well as their relationships to the surrounding meta-sedimentary and meta-volcanic rocks. In an attempt to bridge this gap we examined the geochemistry of platinum-group elements (PGE), Re-Os isotopes and *in situ* chemical analyses of chromite in chromitites from the La Cabaña ultramafic bodies. Our study contributes important data to the ongoing debate regarding interpretations of the origin and evolution of the metamorphic basement of the Chilean Coastal Cordillera, and provides new insights into the evolution of the Palaeozoic eastern margin of the Gondwana supercontinent.

Geology of the chromitites

Geological overview of the metamorphic basement of central Chile

The La Cabaña ultramafic bodies (after Vergara 1970) are situated near the town of Carahue in central Chile, approximately 60 km west of Temuco (Fig. 1). Geologically, they are part of the Palaeozoic metamorphic basement of the Chilean Coastal Cordillera, a paired metamorphic belt of Palaeozoic age that extends for more than 1,000 km along the south-central part of the Pacific coast of Chile. The westernmost unit of this metamorphic belt is known as the Western Series, whereas the easternmost belt is termed the Eastern Series. The rocks of these two units record distinct metamorphic grades: high-pressure/low-temperature (HP-LT) in the Western Series and low-pressure/high-temperature (LP-HT) in the Eastern Series (e.g., Aguirre *et al*. 1972; Godoy and Kato 1990; Willner *et al.* 2005). The boundary between the two units is the sinistral NW-SE-striking Lanalhue Fault Zone (Glodny *et al.* 2008) in the south, whereas it is marked by a late reverse structure, the Pichilemu-Vichuquén Fault in the north (Willner *et al.* 2005).

The Western Series comprises meta-greywackes containing lenses of greenschists and blueschists and associated Fe, Mn-rich meta-sediments, meta-cherts and bodies of mafic-ultramafic rocks (upper mantle dunite, cumulate dunite-pyroxenites and gabbros), with some of them deformed and metamorphosed together with the surrounding host rocks (Aguirre *et al.* 1972; Hervé 1977; Frutos and Alfaro 1987; Barra *et al.* 1998; Höfer *et al.* 2001). Höfer *et al.* (2001) suggested an island arc position for the chromitites, which they interpreted as having a back-arc setting on the basis of data from the neighbouring rocks. Although limited isotopic data exist for the metamorphic host rocks (e.g., Hervé *et al.* 2013), no isotopic data have been obtained so far for the mafic-ultramafic rocks.

The Eastern Series consists of alternating psammites and metapelites characterised by very low grades of metamorphism; these rocks are weakly deformed and mostly preserve the original bedding, except in the easternmost part of the belt where higher metamorphic grade has obliterated the primary fabric of the rocks (Willner *et al.* 2000; Hervé *et al.* 2013).

The rocks of the Coastal Cordillera have been interpreted as either an accretionary complex (Hervé *et al.* 1981; Glodny *et al.* 2005, 2008; Willner 2005; Willner *et al.* 2005), or a backarc basin (Frutos and Alfaro 1987; Rabbia *et al.* 1994; Hufmann and Massonne 2000; Höfer *et al.* 2001) formed during the Late Carboniferous.

The La Cabaña Ultramafic Bodies

The La Cabaña ultramafic bodies consist of two small outcrops (Barra *et al.* 1998): the Centinela Alto (or Lavanderos) to the east and Centinela Bajo to the west (Fig. 2).

Lavanderos is a lens-like ultramafic body 200 × 35 m in size, trending NE-SW. It is hosted in Palaeozoic mica-schists enclosing meta-volcanic rocks (Fig. 2; Vergara 1970; Alfaro 1980; Höfer *et al.* 2001). The body largely consists of antigorite, with accessory clinochlore and chromite mostly transformed to Fe^{2+}-rich porous chromite (Alfaro 1980; Barra *et al.* 1998). Talc locally replaces both antigorite and chlorite. The antigoritite shows a penetrative mylonitic foliation towards the boundaries of the body, which was most likely produced during its fault-controlled emplacement (Barra *et al.* 1998), or during greenschist-facies metamorphism (Höfer *et al.* 2001).

Centinela Bajo is a larger oblate body of 6 × 3 km in size that protrudes into the mica-schists. The rocks are dunites made up of coarse-grained porphyroclastic olivine (1-2 mm) replaced by pseudomorphic (mesh texture) lizardite, with minor amounts of primary pyroxenes. The degree of alteration of the primary silicates may vary from 20 to 80% volume of the rock. The secondary silicates are accompanied by secondary chlorite and amphibole, and a late generation of chrysotile and carbonate veins cross-cutting all of the described minerals. Höfer *et al.* (2001) described an outcrop of meta-gabbros in contact with dunites in the northern part of the ultramafic body. However, in our field studies we did not find a continuous section from the mantle dunite to gabbros, owing to poor exposure in the rainforest. Despite the poor exposure Höfer *et al.* (2001) reported that the gabbros show clear evidence of metamorphism (uralitisation of clinopyroxene, secondary chlorite and minor zoisite and titanite), unlike the peridotites supposedly in contact with them. Detailed structural mapping reveals that shear zones filled by mylonitised serpentinite surround and isolate blocks of undeformed serpentinised dunite; in the latter, mantle foliation (defined by trains of Cr-spinel in the porphyroclastic rocks) and random distributions of Cr-spinel grains are obliterated by the late mylonitic foliation near the contacts with the shear zones. The foliations in the serpentinite rocks have a strike and dip identical to the foliation of the host schists (Höfer *et al.* 2001; Fig. 2).

Chromitites

Chromitites have been reported in both ultramafic bodies at La Cabaña (Fig. 2 and 3; Barra *et al.* 1998; Höfer *et al.* 2001), although chromitites have only been observed *in situ* in Lavanderos (Fig. 3a-e). The Lavanderos chromitites are lenses and veins of a few centimetres thick of massive chromite (>80% chromite grains) hosted in heavily serpentinised dunite (Fig. 3a-c). The contact between the massive chromitite and the host rock is sharp (Fig. 3c) but can locally grade to disseminated chromite ore through a zone of schlieren chromite (Fig. 3a-b). Frequently, late veins of chrysotile transgress the chromitite veins, disrupting their continuity (Fig. 3b-c). The metamorphic alteration that affected the rocks of the body has also completely transformed the magmatic chromite to secondary Fe^{2+}-rich porous chromite (Fig 4a; Barra *et al.* in press).

At Centinela Bajo, the poor exposure in the forest does not allow for direct observation of the chromitite bodies. We only identified a few boulders of massive chromitite on the slopes of the vegetation-covered hills (Fig. 2 and 3d-e). The features of these boulders, such as the massive texture of the ore, their irregular to angular shape and relatively large size, and their distribution along the drainage network suggest the presence of hidden large chromitite bodies (Fig. 2; Höfer *et al.* 2001). The studied chromitite blocks consist of >80% former euhedral chromite grains, which are fractured and complexly zoned (Fig. 4b). The zoning consists of crystallographically-ordered rims (or patches) of Fe^{2+}-rich chromite (Barra *et al.* in press) surrounding the unaltered (magmatic) cores. In the altered rims there are abundant inclusions of Cr-clinochlore, which may also fill one of the generations of fractures cutting the chromite grains. Cr-clinochlore is also the main constituent of the interstitial matrix between chromite grains (Barra *et al.* in press).

Methods

The compositions of major and minor elements in primary igneous cores of chromian spinels were obtained using a SX-100 CAMECA electron microprobe analyser (EMPA) at the Centro de Instrumentación Científica of the University of Granada (Spain), using the procedures described by Gervilla *et al.* (2012). In the former, calibration for chromite was performed using natural and synthetic standards: MgO for Mg, Fe_2O_3 for Fe, Al_2O_3 for Al, Cr_2O_3 for Cr, SiO_2 for Si, TiO_2 for Ti, $MnTiO_3$ for Mn, NiO_2 for Ni and $Pb_5(VO_4)_3Cl$ for V.

Six chromitite samples were analysed for bulk platinum-group elements (PGE) abundances. The analyses were performed by Actlabs, Canada using nickel sulfide fire assay and instrumental neutron activation analysis (INAA) technique. Detection limits were 5 ppb for Ru and Pt, 2 ppb for Os and Pd, 0.2 ppb for Rh, and 0.1 for Ir.

Whole-rock Re-Os isotopic analyses were performed on mineral separates of chromite from four selected chromitite samples. Chromite grains were analysed at the Isotope Geochemistry Laboratory, University of Maryland, USA, following the procedure described by Walker *et al.* (2002a). Briefly, about 0.5g of finely-ground chromite was introduced into a Pyrex™ Carius tube with a 2:1 mixture of concentrated nitric and hydrochloric acids, and appropriate amounts of a mixed Re-Os spike. The tube was then sealed and heated at 270 °C for >24 hours. Upon opening the tube, Os was separated by solvent extraction techniques (Cohen and Waters 1996) and Re by standard column-chemistry procedure (Morgan and Walker 1989). Osmium was further purified by microdistillation (Birck *et al.* 1997). Osmium isotopic ratios were determined using the UMd *VG Sector* 54 thermal ionisation mass-spectrometry (TIMS) using Faraday cups, whereas Re abundances were analysed using a Cetac Aridus nebuliser coupled to a Nu Plasma multi-collector inductively-coupled plasma-mass spectrometer (MC-ICP-MS). External precision for $^{187}Os/^{188}Os$, and Os and Re concentrations for the quantities measured was better than 0.1%, 0.2% and 2% respectively. Procedure blanks averaged 0.8 picograms for Os and 1.2 picograms for Re. Blank corrections for Os were negligible (<0.1%). Blank corrections for Re comprised as much as 1% for some samples. Initial $^{187}Os/^{188}Os$ ratios are calculated as per Shirey and Walker (1998).

Results

Chemistry of primary chromite

As the La Cabaña chromitites are variably altered, interpretations of their petrogenesis in terms of primary magmatic processes in the mantle require the analysis of the unaltered cores of chromite grains. Primary cores were only preserved in the massive chromitite of Centinela Bajo (Fig 4b and 5a-d; Barra *et al.* 1998). The assumption of an igneous origin for these cores is supported by EMPA analyses showing that they contain very low Fe_2O_3 (<2.60 wt.%; *i.e.*, [$Fe^{3+}\# = Fe^{3+}/(Fe^{3+}+Al^{3+}+Cr^{3+})$] <0.03), high MgO (>12.09 wt.%), and low TiO_2 (<0.23 wt.%), values that are typical for primary chromite from mantle-hosted ophiolitic chromitites (Fig. 5a-b; Table 1).

The primary chromite of the Centinela Bajo chromitites is highly homogeneous in terms of Cr# [Cr/(Cr+Al) atomic ratio] and Mg# [$Mg/(Mg+Fe^{2+})$ atomic ratio] (Fig. 5c). Thus, the Cr# values range between 0.64 and 0.66, which correspond to 48.71-51.81 wt% Cr_2O_3 and 17.33-18.52 wt.% Al_2O_3, whereas the Mg# varies from 0.58 to 0.67. These Cr# and Mg# values overlap with the compositional field of high-Cr podiform ophiolitic chromitite (Fig. 5a-c; Leblanc and Nicolas 1992). The contents of MnO, V_2O_3 and ZnO are lower than 0.48 wt.%, 0.29 wt.% and 0.20 wt.%, respectively.

Platinum-group elements (PGE)

The La Cabaña chromitites have relatively low PGE concentrations, with total PGE contents ranging from 180 to 347 ppb for each of the samples (Table 2), with slightly higher concentrations in samples from Centinela Bajo (294-347 ppb; mean 320 ppb) relative to those from Lavanderos (180-235; mean 213 ppb). Figure 6 shows the chondrite-normalised distribution pattern of the La Cabaña chromitites compared with chromitites hosted in the mantle sections of other ophiolites, and from other geodynamic settings. The general compositional trend of the La Cabaña chromitites fits very well with the distribution of the PGEs shown by other Type I chromitites (González-Jiménez *et al.* 2014a) that are hosted in the mantle sequence of ophiolite complexes and show an enrichment in the IPGE (Os, Ir and Ru =305-346.9 ppb) relative to the PPGE (Rh, Pt and Pd=13.9-42.9 ppb).

The La Cabaña chromitites exhibit steep positive patterns from Os to Ru (with positive Ir anomalies relative to Os and Ru in the samples CAB-7B and PODOB), followed by steep negative slopes from Rh to Pd (with Pd positive anomalies in the samples CAB-7B, LAC-2). This distribution of the PGEs in the La Cabaña chromitites also overlaps the fields of other Latin American chromitites hosted in the mantle sections of supra-subduction zone (SSZ) ophiolites (Fig. 6a-c), such as the "unmetamorphosed" high-Cr chromitites in Sagua de Tánamo, eastern Cuba (González-Jiménez *et al.* 2011), in Santa Elena in the Dominican Republic (high-Cr and high-Al; Fig. 6b; Zaccarini *et al.* 2011), and in the "metamorphosed" high-Al chromitites of Tehuitzingo, SW Mexico (Proenza *et al.* 2004). In contrast, the La Cabaña chromitites have lesser amounts of IPGE and a less pronounced negative segment from IPGE to PPGE than the chromitites (high-Cr) of the Vizcaino Peninsula, Mexico (Fig. 6b; Vatin-Perignon *et al.* 2000). These chromitites are interpreted to have crystallised in equilibrium with melts of boninitic affinity, generated by hydrous melting in suprasubduction environments. Finally, the La Cabaña chromitites are significantly different from the high-Al chromitites of the Pampean Ranges, Argentina, in which metamorphism has disturbed the original distribution of PGEs (Fig. 6c; Proenza *et al.* 2008).

The abundances of IPGE in the La Cabaña chromitites were controlled by abundant small inclusions (<25 µm) of Os-Ir-Ru alloys, laurite (RuS_2)-erlichmanite (OsS_2), irarsite (IrAsS), and minor omeiite ($OsAs_2$), and Ru-Ni alloys (Galdames *et al.* 2011). The PPGE-rich assemblage mainly includes Rh-rich minerals, such as hollingworthite (RhAsS), and an unidentified antimonide (Rh-Cu-Sb). Laurite-erlichmanite is found as inclusions in the cores of chromite grains but is more frequently replaced by Os-Ir-Ru alloys in pores of Fe^{2+}-rich chromite, suggesting its desulfurisation under reducing conditions. This process could be also responsible for the formation of Ru-Ni alloys after breakdown of Ru-bearing Ni-rich sulfides. In contrast, irarsite, omeiite and the unnamed Rh-Cu antimonide are most likely associated with oxidising fluids, which supplied As and Sb to the system, and redistributed the PGE. Abundant secondary inclusions of Ni-rich S-poor sulfides (heazlewoodite and gersdorfite) and Ni-arsenides accompany these PGM in the alteration rims of chromite.

Whole-rock Re-Os isotopes

Total Re contents in La Cabaña chromitites are very low (<0.35 ppb), with the exception of sample CEN-07, which shows a slightly higher Re content (2.2 ppb). Osmium concentrations vary from 38 to 202 ppb (Table 3). Measured $^{187}Os/^{188}Os$ ratios of chromite separates from three chromitite blocks of Centinela Bajo vary from 0.1252 to 0.1271, and a chromite separate from one chromitite pod from Lavanderos has a ratio of 0.1268 (Table 3). The $^{187}Os/^{188}Os$ for all four chromitite samples averages of 0.12652 ± 0.0008 (2σ). Estimates for the present $^{187}Os/^{188}Os$ ratio of the bulk oceanic mantle range from a low of ~0.125, based on the average of abyssal peridotites (e.g., Snow and Reisberg 1995; Liu *et al.* 2009), to a high of ~0.128, based on averages for chromitites and peridotites of Phanerozoic ophiolites extrapolated to the present (e.g., Walker *et al.* 2002b). Thus, ratios for all of our chromitite samples fall well within the range of estimates for the modern oceanic mantle.

Initial $^{187}Os/^{188}Os$ ratios have been calculated for 300 Ma, which is the estimated age of emplacement of the peridotites onto the continental crust inferred from the age of metamorphism of the meta-sedimentary rocks hosting the peridotites (Höfer *et al.* 2001; Willner *et al.* 2005). These ratios are almost identical to the measured ratios because of the very low Re/Os ratios of the chromitites. At La Cabaña, the initial $^{187}Os/^{188}Os$ ratios span from 0.1248 to 0.1271, with an average of 0.1264 ± 0.0011 (2σ) (Fig. 7). Although the total number of samples analysed for Os isotopes is too small to be statistically robust, there appear to be no major differences in Os isotopic compositions between chromitites of Centinela Bajo and Lavanderos.

Discussion

Parental melts of the chromitites

The composition of primary magmatic cores of chromite from Centinela Bajo chromitites with high Cr, and low Fe_2O_3 and TiO_2 contents is typical of chromite from chromitites hosted in the upper mantle sequence of ophiolite complexes (Fig. 5a-b). In the diagram Al_2O_3 versus TiO_2 defined by Kamenetsky *et al.* (2001) for chromites hosted in lavas of different geotectonic settings (Fig. 5d), some of these cores fall within the field of MORB and within the field of spinels with high Ti from basalts of modern back-arc basins; another set of data with lower TiO_2 plots out of these fields. This scattering in the plot is a common feature of many mantle-hosted chromitites worldwide (e.g., Pagé and Barnes 2009; Zaccarini *et al.* 2011; Escayola *et al.* 2011), which may reflect the increase of TiO_2 produced in melts crystallising chromitite as they migrate through the SSZ mantle while reacting with peridotite wall-rock (Graham *et al.* 1996; Zaccarini *et al.* 2011; González-Jiménez *et al.* 2011, 2014b).

Experimental and empirical studies show that the contents of Al_2O_3, TiO_2, FeO and MgO in chromite reflect those of the melt from which the chromite has crystallised (Maurel and Maurel 1982; Wasylenki *et al.* 2003; Kamenetsky *et al.* 2001; Rollinson 2008; Pagé and Barnes 2009). Considering that our chromites are high-Cr and low-Ti (Fig. 5a-d), similar to chromite from low-Ti high-Cr arc-lavas of Kamenetsky *et al.* (2001), we have estimated the Al_2O_3 and TiO_2 contents of the melt(s) in equilibrium with our chromite using the revised equations defined by Zaccarini *et al.* (2011) for those chromites (eqs. 1 and 2):

$$ln\,[\text{wt\% } Al_2O_{3\,(melt)}] = 5.2253\times\, ln\,[Al_2O_{3\,(spinel)}] - 1.1232\ (R^2{=}0.9723) \qquad \text{(eq. 1)}$$

$$ln\,[\text{wt\% } TiO_{2\,(melt)}] = 1.0897\ ln[\text{wt\% } TiO_{2\,(spinel)}] + 0.0892 \qquad \text{(eq. 2)}$$

The FeO/MgO ratios of the melts in equilibrium with Centinela Bajo chromites were estimated using the following empirical expression (Maurel and Maurel 1982; equation 3):

$$ln(FeO/MgO)_{spinel} = 0.47 - 1.07Al\#_{spinel} + 0.64Fe^{3+}\#_{spinel} + ln\,(FeO/MgO)_{melt} \qquad \text{(eq. 3)}$$

where FeO and MgO in wt%, $Al\# = Al/(Cr+Al+Fe^{3+})$ and $Fe^{3+}\# = Fe^{3+}/(Cr+Al+Fe^{3+})$.

These calculations suggest that the melts that have precipitated the chromite of the Centinela Bajo chromitites contained from 13.7 to 14.2 wt% Al_2O_3, and from 0.22 to 0.34 wt% TiO_2, with the FeO/MgO ratio varying between 0.8 and 1.2. The Al_2O_3 contents of the parental melt in equilibrium with the igneous chromite of the Centinela Bajo are similar to the melts that precipitated the high-Cr chromites in boninite lavas at Bonin Island in Japan (10.6-14.4 wt.% Al_2O_3; Hicky and Frey 1982;

Crawford *et al.* 1989), but also overlap those of high-Mg island-arc tholeiites (IAT) (11.4-16.4 wt.% Al_2O_3; Augé 1987).

Similar arc-type melts, but with lower Al_2O_3, TiO_2 and FeO/MgO ratios, have been suggested to be responsible for the formation of high-Cr chromitites in the mantle section of many SSZ ophiolites. Examples include Kempirsai in Kazakhstan (Melcher *et al.* 1997), Oman (Rollinson 2008), Muğla in Turkey (Uysal *et al.* 2009), Rutland Island in the Bay of Bengal (Ghosh *et al.* 2009), Santa Elena in Costa Rica (Zaccarini *et al.* 2011), and Mayarí-Cristal, in eastern Cuba (Proenza *et al.* 1999; González-Jiménez *et al.* 2011). These chromitites were interpreted to have crystallised near the Moho Transition Zone (MTZ) of oceanic lithosphere overlying a subduction zone in either a fore-arc or back-arc basin.

On the other hand, the La Cabaña chromitites show the typical chondrite-normalised PGE patterns of ophiolitic chromitites characterised by the enrichment in IPGE (i.e., Os, Ir and Ru), which differ significant from chromitites found in layered mafic intrusions and Ural/Alaskan complexes (Fig. 6a). The inclusions of primary Os-rich PGM like laurite, as a mineralogical expression of the abundances of IPGEs, indicate the sulfur-undersaturated nature of the melts that crystallised the chromitites (Garuti *et al.* 1999; Brenan and Andrews 2001; Andrews and Brenan 2002; Bockrath *et al.* 2004). Basaltic melts with low enough fS_2 to precipitate Os-rich PGM are produced during the partial melting of depleted peridotites (Jugo 2009), or secondarily by the reaction of IAT melts with mantle peridotite at increasing melt/rock ratio during the formation of chromitites, in SSZ mantle (Zhou *et al.* 1998; Büch *et al.* 2002; Shi *et al.* 2007; Marchesi *et al.* 2011).

Tectonic setting of the La Cabaña ultramafic bodies

According to Prichard *et al.* (2008), chromitites with appreciable amounts (~100 ppb) of total PGEs are precipitated only from melts that have been extracted after ≥20% partial melting of mantle sources. The fact that at La Cabaña the average PGE content of chromitites is around 250 ppb suggests that the melts were derived from a mantle source that had experienced degrees of partial melting above 20%. The primary olivine preserved in the dunites at Centinela Bajo have Mg# = 0.90-0.92 and NiO contents (0.16-0.45 wt.%), similar to the olivine in the ultra-depleted peridotites of the mantle sections of ophiolites of Eastern Cuba (Fig. 8). This leads us to suggest that the dunites hosting the chromitites at Centinela Bajo are residues resulting from high degrees of partial melting.

In the ophiolite environment, peridotites that have undergone multiple episodes of melt extraction until nearly barren in their major- and minor-elements signatures are common in the mantle beneath arcs developed above supra-subduction zones (e.g., Ohara *et al*. 2002; Marchesi *et al*. 2006; Ishimaru *et al*. 2007). In these settings, dehydration of the subducting slab may release a large volume of volatiles contributing to a decrease in the melting temperature, thus favouring the release of PGEs from their host peridotitic mantle at lower degrees of partial melting than relatively fertile, "dry" and less oxidised settings such as MOR (Jugo 2009). Chromitites that have crystallised in a MOR setting are significantly more depleted in PGEs than those from the La Cabaña (frequently <100 ppb; Leblanc and Nicolas 1992; Zhou *et al.* 1996, 1998; Proenza *et al.* 1999; Ahmed and Arai 2002; Prichard *et al.* 2008; Uysal *et al.* 2009; González-Jiménez *et al.* 2014a).

The similar geochemistry of the studied chromitites to "mantle hosted" chromitites from SSZ ophiolites provide the first evidence that the La Cabaña ultramafic bodies are portions of an oceanic mantle overlying a supra-subduction zone. The formation of chromitites in the suprasubduction zone environment is commonly associated with the crystallisation of basaltic melts at their site of extraction during the opening of marginal basins in either fore-arc or back-arc basins (Roberts 1998; Leblanc and Nicolas 1992; Malpas 1997; Melcher *et al.* 1997; Proenza et al. 1999; Rollinson, 2005, 2008; Gervilla *et al.* 2005; Robinson and Adetunji 2013; González-Jiménez *et al.* 2014b). This is consistent with the fact that the widespread meta-volcanic rocks and metabasites in the Western Series show E-MORB signatures (Rabbia *et al.* 1994; Hufmann and Massone 2000), suggesting that these rocks crystallised from basaltic melts derived from an upwelling mantle beneath a spreading center. During the formation and evolution of the marginal basin above a supra-subduction zone, the extraction of melts from the mantle takes place preferentially through dunite conduits sited beneath the axis of the spreading ridge. Numerical modelling (Braun and Kelemen 2002) shows that in these settings the dunite conduits form an interconnected network of channels filled with basaltic melts that might be extracted from different mantle sources. Where two or more melt-filled conduits intersect, batches of melt with different degrees of fractionation (i.e., different $a\mathrm{SiO_2}$) may mix to produce hybrid melts saturated with chromite, and thus, are capable of crystallising sizeable chromitite bodies (González-Jiménez *et al.* 2014b).

Under the proposed scenario, the upwelling arc-type melts might become saturated in chromite, precipitating chromitites prior to their emplacement at crustal levels as mafic rocks or extruded lavas. A similar interpretation has been proposed for other ophiolites (Leblanc 1986; Graham *et al.* 1996; González-Jiménez *et al.* 2011), and demonstrated by analysis of minor- and trace-element in chromite from a suite of mantle and crustal rocks of Oman (Dare *et al.* 2009) and Thetford Mines in Canada (Pagé and Barnes 2009). Thus, our observations and data provide significant evidence against previous interpretations that the metabasites and their associated ultramafic rocks (i.e., serpentinites and peridotites) found in the metamorphic basement of the Coastal Cordillera represent the segments of *the upper part of an oceanic crust* (Willner 2005). In contrast, we propose that these rocks correspond to different fragments of oceanic lithosphere (including mantle and overlying crust) that originally formed in a marginal basin developed above a supra-subduction zone. Most likely these portions of upper mantle were later incorporated into the continental crust by basal accretion within the accretion prism represented by the metamorphic rocks of the Coastal Cordillera.

Interpretation of the Re-Os isotopic data

As is typical of ophiolitic chromitites worldwide, the four La Cabaña chromites are characterised by very low $^{187}\mathrm{Re}/^{188}\mathrm{Os}$ ratios, hence calculated initial ratios are not very sensitive to age correction. Here, we have used the estimated age of the host rocks metamorphism (ca. 300 Ma) to calculate the initial $^{187}\mathrm{Os}/^{188}\mathrm{Os}$ ratio. However, it is reasonable to think that this is a minimum age because the chromitites and their PGM and base-metal sulfide (BMS) inclusions could have formed several million years before their emplacement onto the continental crust (e.g., González-Jiménez *et al.* 2014a). Of note, the initial $^{187}\mathrm{Os}/^{188}\mathrm{Os}$ ratios of the La Cabaña chromites fit well with the average evolution curve for chromitites of SSZ ophiolites worldwide (Fig. 7; Walker *et al.* 2002b).

The initial $^{187}Os/^{188}Os$ ratios of La Cabaña chromites, however, vary little compared with much larger variations noted for most other chromitites from Phanerozoic ophiolites (Fig. 7a). Büchl *et al.* (2004) also observed small variations of Os in chromitite pods hosted in the mantle section of the Troodos ophiolite. In these chromitites the range of variation of $^{187}Os/^{188}Os$ ratios of the chromite separates (0.1265-0.1301) is much lower than those of the associated peridotites (0.1235-0.1546). This observation is in agreement with more recent data by O'Driscoll et al. (2012) for the Shetland Ophiolite chromitites, where whole rock $^{187}Os/^{188}Os$ for discrete chromitite pods is remarkably homogenous, albeit with significant heterogeneities observed in chromitites from different localities. These authors suggest that during the formation of the chromitites under an open-system, potentially with significant input of numerous melts with variable composition, the mixing process involved in the formation of chromitites tend to homogenize the Os signatures. This is surprising, as the Os isotopic compositions are likely governed by the presence of discrete inclusions of PGM and BMS formed from melt/fluids with diverse magmatic and possibly post-magmatic contributions (Fig. 7; Malitch *et al.* 2003; Malitch 2004; Ahmed *et al.* 2006; Shi *et al.* 2007; Marchesi *et al.* 2011; González-Jiménez *et al.* 2012a,b, 2013a, 2014a).

Additionally, recent studies of ophiolite chromitite have shown that they often contain several populations of PGM and BMS, with different Os-isotope signatures at the sub-grain scale (e.g., Marchesi *et al.* 2011; González-Jiménez *et al.* 2012a,b, 2013a, 2014a). The *in-situ* analysis, using LA-MC-ICPMS, of large numbers of individual PGM and BMS from chromitites and host peridotites in the Ojén massif in Spain suggest that the formation of these chromitites, which involved partial melting, melt transport/pooling and melt–rock reactions, did not erase the original $^{187}Os/^{188}Os$ heterogeneity of the peridotite source (González-Jiménez *et al.* 2013a). Indeed, the results of this study highlights the complexity of the mechanism of formation of the chromitites in the oceanic mantle, which may require the fractional extraction and/or stepped pooling of melts for the preservation of the Os signatures during the mingling process necessary to crystallise chromite (González-Jiménez *et al.* 2013a).

A new geodynamic framework for the mafic-ultramafic rocks of the Chilean Coastal Cordillera

The structure of the present western margin of the South American continent is seen as the result of the amalgamation of several tectonostratigraphic terranes to the southern Andean margin during the Palaeozoic (Ramos *et al.* 1986; Bahlburgh *et al.* 1994; Ramos 2008, 2009; Bahlburgh *et al.* 2009). During the early to middle Devonian the hypothetical exotic terrane known as Chilenia (Ramos *et al.* 1984; Ramos and Basei 1997; Keppie and Ramos 1999) was separated from Laurentia and later accreted to the proto-Andean margin in the south-central part of Chile (Ramos *et al.* 1986; Davis *et al.* 2000; Ramos 2008). The continental crust of the Chilenia micro-continent was later intruded by the Southern Coastal Batholith ca. 295-307 Ma ago (Hervé *et al.* 1988, 2013; Lucassen *et al.* 2004; Glodny *et al.* 2008). The geochemical and isotopic studies indicate that the parental melts of these intrusive rocks were derived from a portion of sub-arc mantle overlying a supra-subduction zone (Lucassen *et al.* 2004). These observations lead us to suggest that this section of the paleo-margin of Gondwana remained as a passive margin until the early Late Carboniferous when a paleo-subduction zone was initiated (Fig. 9). This interpretation is consistent with previous hypotheses that the paleomargin of Gondwana was a passive margin from Ordovician to Early Carboniferous time (Hervé *et al.* 1987; Bahlaburg and Hervé 1997; Augustsson and Balhburg 2003). The formation of sedimentary rocks that now constitute the Western Series is related to the erosion of amalgamated terranes, as is indicated by the U-Pb ages of detrital zircons in these units (Hervé *et al.* 2013).

Rapalini and Vilas (1991) suggested that between the Late Carboniferous and the Early Permian, the vector of subduction of the proto-Pacific changed to a northward oblique direction, which resulted in the rotation of several continental blocks with strike-slip displacement parallel to the western continental margin of South America. This interpretation, based on paleomagnetic studies, is consistent with the fact that the Lanalhue Faut zone acted, in its initial stages of development, as a ductile deformation zone with divergent character (Glodny *et al*. 2008). We suggest that this change of the tectonic regime, which very likely also increased the angle of the subducting slab, coupled with the associated rollback of the subducting plate, enhanced the break-up of a portion of Chilenia by opening of a marginal basin in the fore-arc region (Fig. 9).

The emplacement of the La Cabaña peridotites into the continental crust is poorly constrained at 282 ± 6 Ma, based on K-Ar dating of fuchsite, whereas granitoids intruded the continental crust of Chilenia ca. 295-307 Ma. This suggests that this marginal basin was active for a very short period (~10 Ma). Interestingly, the Os isotopic values of the studied chromites are also within the range of values determined for the subcontinental lithospheric mantle near the studied area, determined by the analyses of mantle xenoliths located in the back-arc region of the Andes (Schilling *et al*. 2008). This suggest that peridotites that are representative of ancient SCLM beneath Chilenia could have become part of an oceanic basin being altered (i.e., oceanised), and later served as the basement for the accumulation of oceanic basalts and sediments, similar to that observed in other oceanic basins (Tsuru *et al.* 2000; Shi *et al.* 2007; 2013; Griffin *et al.* 2009; O'Reilly *et al.* 2009; Tang *et al.* 2013). This marginal basin very likely developed in a fore-arc position and must have been limited to the west by a fragment of the former Chilenia terrain (Fig. 9). Additional evidence for the formation of this marginal basin is provided by the geochemistry of the chromitites presented here, and the E-MORB signature of the metabasites widespread through the Western Series (Díaz *et al.* 1988; Vivallo *et al.* 1988; Rabbia *et al.* 1994; Hufmann and Massone 2000). This interpretation is also consistent with the presence of several Besshi-type massive sulfide deposits interpreted as formed in a proximal setting in a marginal basin (Alfaro and Collao 1990; Schira *et al.* 1990). In this new model, the protolithic sedimentary rocks of the Western Series resulted from erosion and later sedimentation within the marine basin of Carboniferous granitic plutons and possibly some parts of the previously-formed Eastern Series (Fig. 9).

A compressive regime can be related with the San Rafael orogeny, as described by Mpodozis and Kay (1990, 1992) for the northern part of Chile (28°-31°S), in which a hypothetical terrane (Terrane X) collided with the Chilean paleo-Pacific margin during the Early Permian. In an analogous way, the proposed separated fragment of Chilenia docked with the Gondwana margin, producing the Early Permian metamorphism preserved in the rocks of both Western and Eastern Series (Höfer *et al.* 2001; Glodny *et al.* 2008). We currently lack evidence for this fragment of Chilenia, which may have been lost by tectonic erosion associated with subduction.

Conclusions

(1) The geochemistry signature of the La Cabaña chromitites suggests that they precipitated from arc-type melts originated within a suprasubduction zone environment. This new interpretation rules out previous models that have suggested that the mafic and ultramafic rocks of the Coastal Cordillera were formed in an abyssal setting.

(2) The initial $^{187}Os/^{188}Os$ ratios for four chromitite samples, calculated for the estimated age of emplacement, range from 0.1248 to 0.1271. These isotopic compositions are within the worldwide range of reported Re-Os data for chromitites from other Phanerozoic ophiolites.

(3) The formation of the chromitites was associated with the formation and evolution of a marginal basin above the supra-subduction zone. However, unlike most chromitites reported worldwide, this basin did not develop behind an island arc but in a fore-arc setting. Available regional data suggest that they originated in a pull-apart basin that separated a portion of the Chilenia micro-continent. The opening of this marginal basin was associated with the change in the direction of subduction in the paleo-Pacific subduction zone, and involved the “oceanisation” of the old subcontinental lithospheric mantle.

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Figure captions

Figure 1. Geological sketch map of the Coastal Cordillera of south-central Chile with the location of the La Cabaña area (modified from SERNAGEOMIN, 2003)

Figure 2. Simplified geological map of the La Cabaña area (modified from Höfer *et al.* 2001) showing the location of chromitite bodies observed *in situ* and collected from the rainforest drainage network. Legend is inset in the figures.

Figure 3. Photographs showing morphologies and structures of the studied chromitites in the La Cabaña ultramafic bodies. (a), (b) and (c) are chromitites observed *in situ* at Lavanderos whereas (d) and (e) are boulders of chromitite collected from the rainforest drainage network. In photographs (b) and (c) late chrysotile (Ctl) veins in disrupt in the chromitite. The size of the different objects shown in the figures for scale are: pen in (a) 15 cm long, coins in (b) and (c) 2 cm and 2.3 cm in diameter respectively, and hammer in (d) and (e) 35 cm.

Figure 4. Back-scattered electron (BSE) images showing textures of chromitites of the La Cabaña area. (a) chromite grain completely transformed to secondary porous chromite at Lavanderos, (b) partly altered chromite with porous chromite rim and homogenous unaltered core at Centinela Bajo. We use the definition of porous chromite and partly altered chromite micro-structures as defined Gervilla *et al.* (2012).

Figure 5. Chemistry of primary chromite (black circles) of the Centinela Bajo chromitite as comparison with chromian spinel of various tectonic settings in terms of (a) Al-Cr-Fe^{3+} compositions, (b) TiO_2 versus Cr_2O_3. (c) Cr# [Cr/(Cr+Al)] versus Mg# [Mg/(Mg+Fe)]. (d) TiO_2 versus Al_2O_3. Data sources for chromian spinel of different tectonic settings taken from Kamenetsky *et al.* (2001) and Proenza *et al.* (2007).

Figure 6. C1-chondrite (Naldrett and Duke 1980) normalised patterns of the La Cabaña chromitite and comparison with the chromitites hosted in the ophiolitic mantle (Proenza *et al.* 2007 and references therein) and in Alaskan-type complexes (Garuti *et al.* 2003, 2005) of the Urals and Bushveld (UG2) Layered Complex (Naldrett *et al.* 2011). Other selected chromitites (with high-Cr and high-Al chromite) of Latin America have been included for comparison: Vizcaino, Baja California Sur, Mexico (Vatin-Perigon *et al.* 2000); Tehuitzingo, Acatlan Complex, Mexico (Proenza *et al.* 2004); Pampean Ranges, Argentina (Escayola *et al.* 2004); Santa Elena, Costa Rica (Zaccarini *et al.* 2011); Sagua de Tánamo, Mayarí-Baracoa ophiolite, Cuba (González-Jiménez *et al.* 2011).

Figure 7. Initial $^{187}Os/^{188}Os$ versus presumed age (Ga) of ophiolite formation for the La Cabaña chromitites and other chromitites hosted in the mantle section of Phanerozoic SSZ ophiolites. Here the term "age of ophiolite formation" is considered as either timing of the incorporation of the peridotite substrate into an "oceanic" column, with an overlay of basaltic and gabbroic crust and sediments, or the time when the crust-mantle sequence was emplaced onto the continental crust for preservation. The gray line represents the evolution trajectory for the primitive upper mantle (PUM) as reported by Meisel *et al.* (2001) whereas black one corresponds to estimates for chromitites worldwide (Walker et al. 2002b). The plot (a) for whole-rock and chromite separates includes data reported by Walker et al. (2002b) for chromitites worldwide ophiolite as well as data for Albania (Kocks *et al*. 2007); Kraubath, Austria (Melcher and Meisel 2004); Shetland, Scotland (Walker *et al.* 2002b; O'Driscoll *et al*. 2012); Mayarí-Baracoa, Cuba (Gervilla *et al*. 2005; Frei *et al.* 2006); Mugla, Turkey (Uysal *et al.* 2009); and Sartohay, China (Shi *et al.* 2012). The plot (b) includes data for *in-situ* PGM and BMS analysed in chromitites from the Eastern Desert, Egypt (Ahmed *et al.* 2006); Dobromirtsi, Bulgaria (González-Jiménez *et al*. 2013b); Kraubath, Austria (Malitch 2004), Loma Baya, Mexico (González-Jiménez *et al.* 2014a); Luobusa and Donqiao, Tibet (Shi *et al*. 2007); Mayarí and

Sagua de Tánamo, Cuba (Marchesi *et al.* 2011; González-Jiménez *et al.* 2012a), Oman (Ahmed *et al*. 2006); and the Ojén Lherzolite Massif, Spain (González-Jiménez *et al.* 2013a). Note the smoothing effect of bulk-chromite data compared to the *in-situ* analyses of individual PGM and BMS, as several generations of the latter control the budget of Os-isotopes in the chromitite.

Figure 8. NiO versus Mg# in olivine relicts in dunites of Centinela Bajo; these are very similar to those in ultra-depleted peridotites of the Cuban ophiolites (Proenza *et al.* 1999).

Figure 9. Geodynamic model proposed for the evolution of the Chilean Coastal Cordillera.

Tables

Table 1. Selected analyses of unaltered chromite cores of Centinela Bajo chromitites

Table 2. Platinum-group elements of chromitite samples from the La Cabaña ultramafic bodies.

Table 3. Re and Os isotopes and abundances of selected chromitite samples from the La Cabaña ultramafic bodies.

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FIGURE 1.

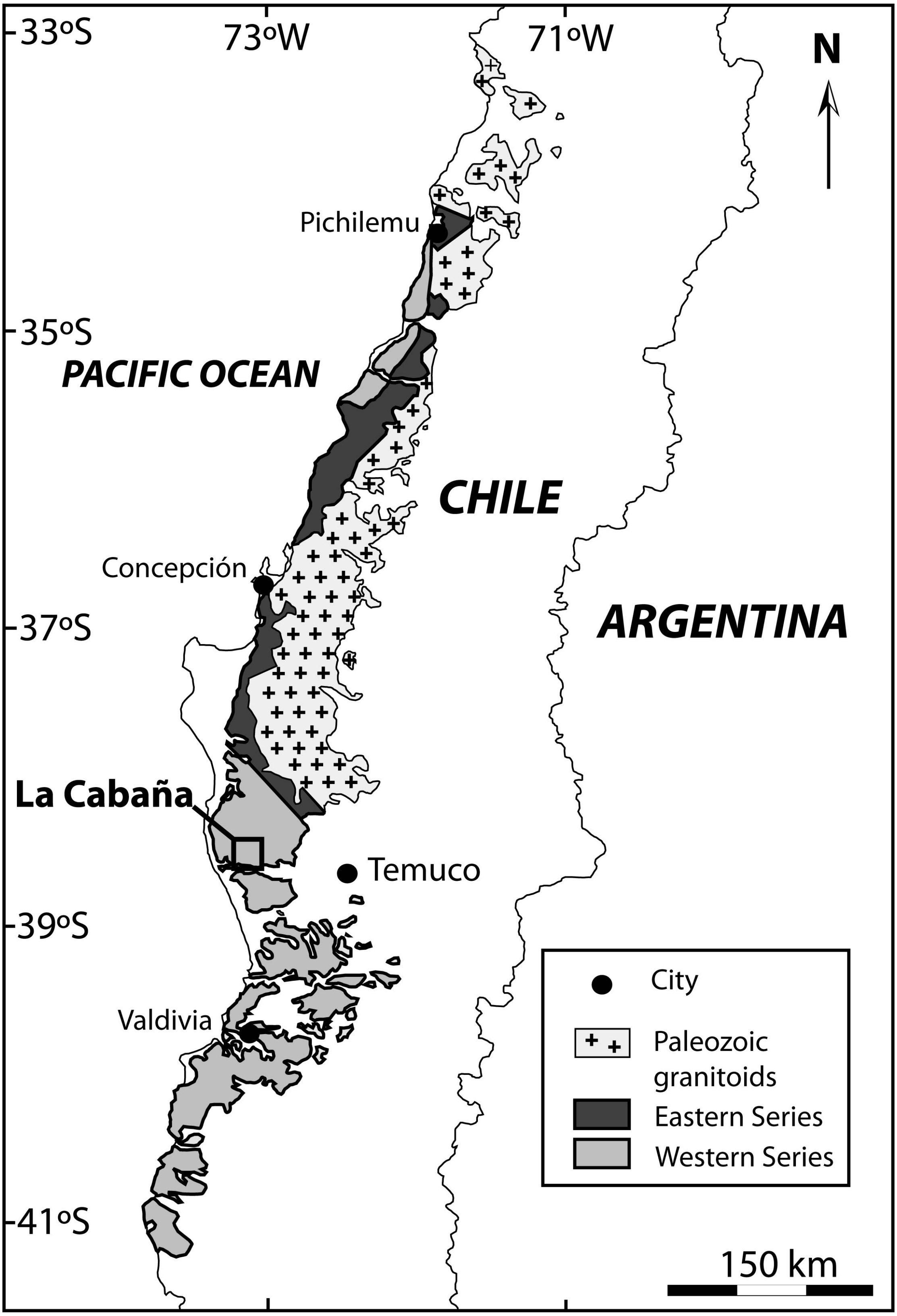
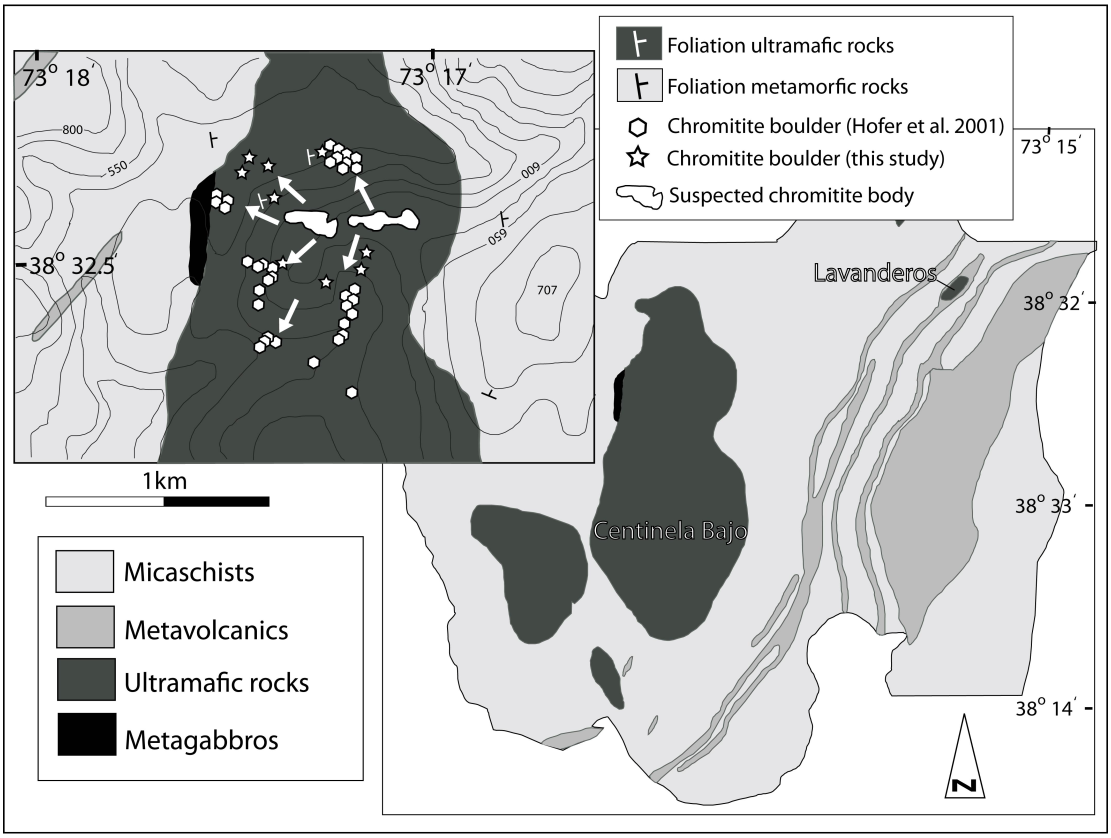

33ºS
73ºW
71ºW
N
Pichilemu
35ºS
PACIFIC OCEAN
CHILE
Concepción
37ºS
ARGENTINA
La Cabaña
Temuco
39ºS
City
Paleozoic granitoids
Eastern Series
Western Series
Valdivia
41ºS
150 km

FIGURE 2.



Foliation ultramafic rocks
Foliation metamorfic rocks
Chromitite boulder (Hofer et al. 2001)
Chromitite boulder (this study)
Suspected chromitite body
73° 18´
73° 17´
73° 15´
38° 32.5´
38° 32´
38° 33´
38° 14´
800
550
600
650
707
1km
Lavanderos
Centinela Bajo
N
Micaschists
Metavolcanics
Ultramafic rocks
Metagabbros

FIGURE 3.

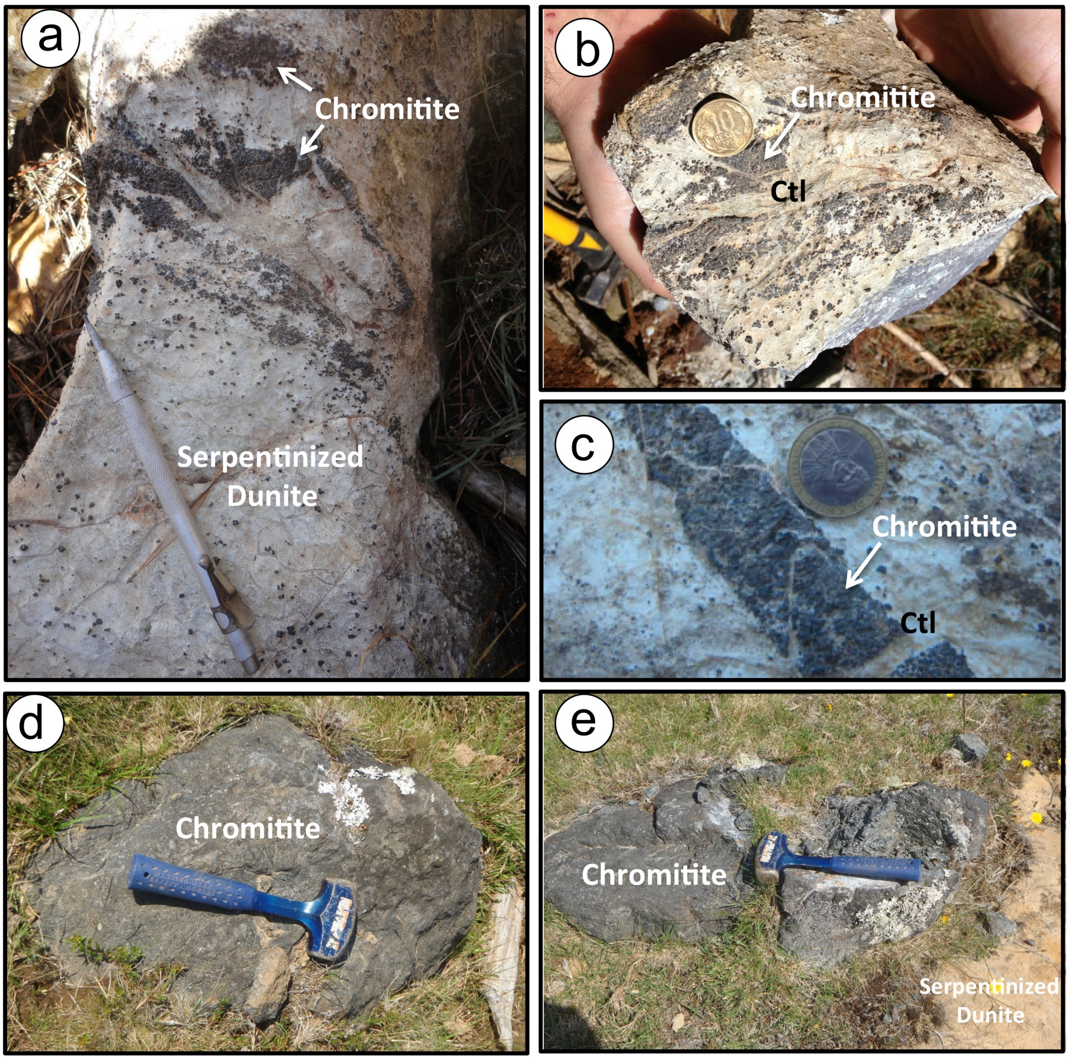

a
Chromitite
Serpentinized
Dunite
b
Chromitite
Ctl
c
Chromitite
Ctl
d
Chromitite
e
Chromitite
Serpentinized
Dunite

FIGURE 4.

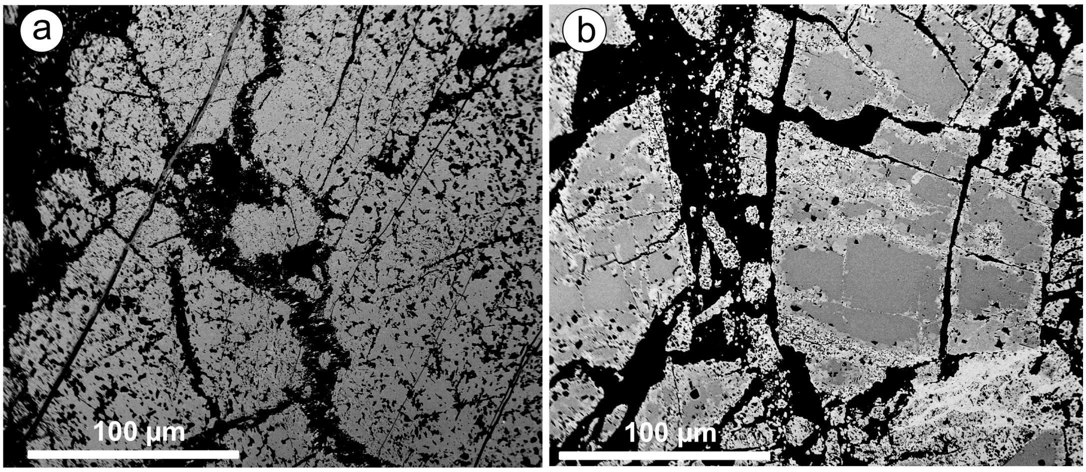


FIGURE 5.

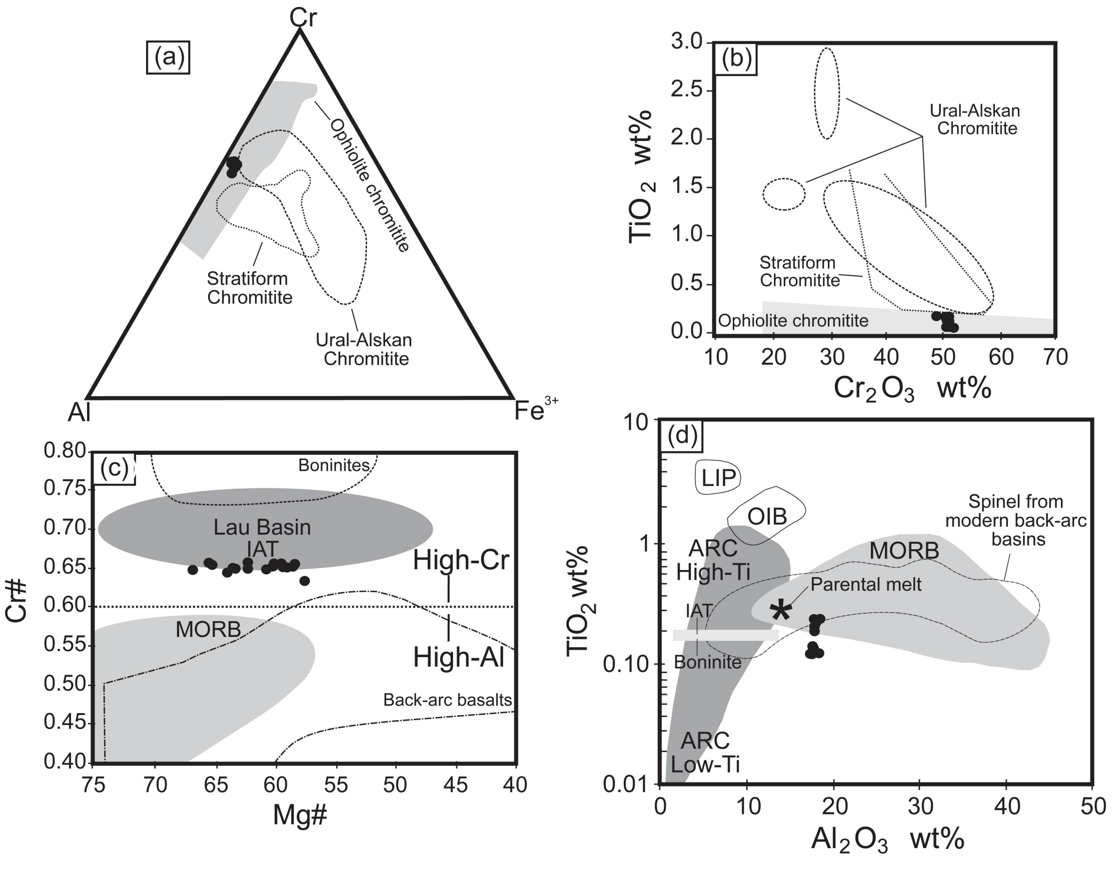
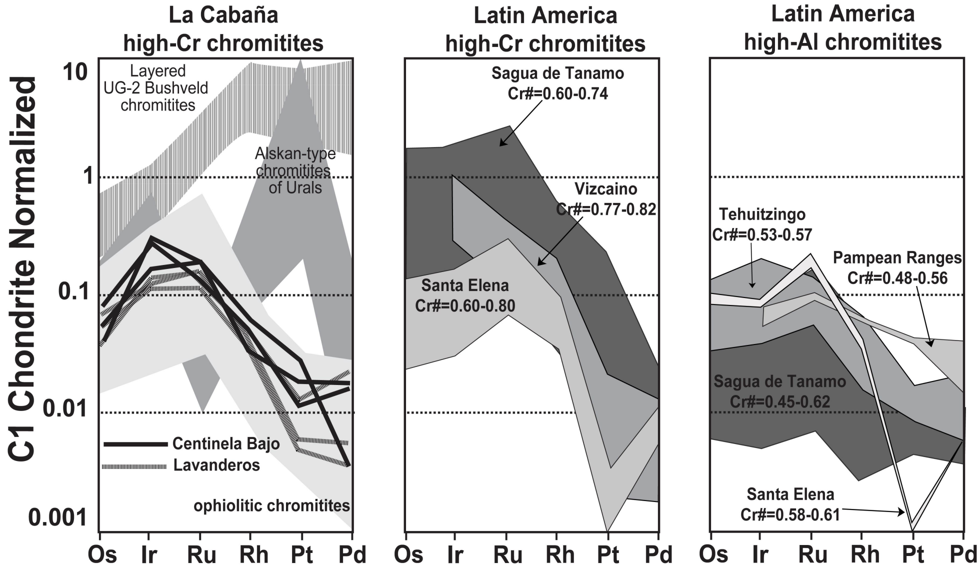


FIGURE 6



La Cabaña
high-Cr chromitites
Latin America
high-Cr chromitites
Latin America
high-Al chromitites
C1 Chondrite Normalized
10
1
0.1
0.01
0.001
Layered
UG-2 Bushveld
chromitites
Alskan-type
chromitites
of Urals
Centinela Bajo
Lavanderos
ophiolitic chromitites
Sagua de Tanamo
Cr#=0.60-0.74
Vizcaino
Cr#=0.77-0.82
Santa Elena
Cr#=0.60-0.80
Tehuitzingo
Cr#=0.53-0.57
Pampean Ranges
Cr#=0.48-0.56
Sagua de Tanamo
Cr#=0.45-0.62
Santa Elena
Cr#=0.58-0.61
Os
Ir
Ru
Rh
Pt
Pd

FIGURE 7.

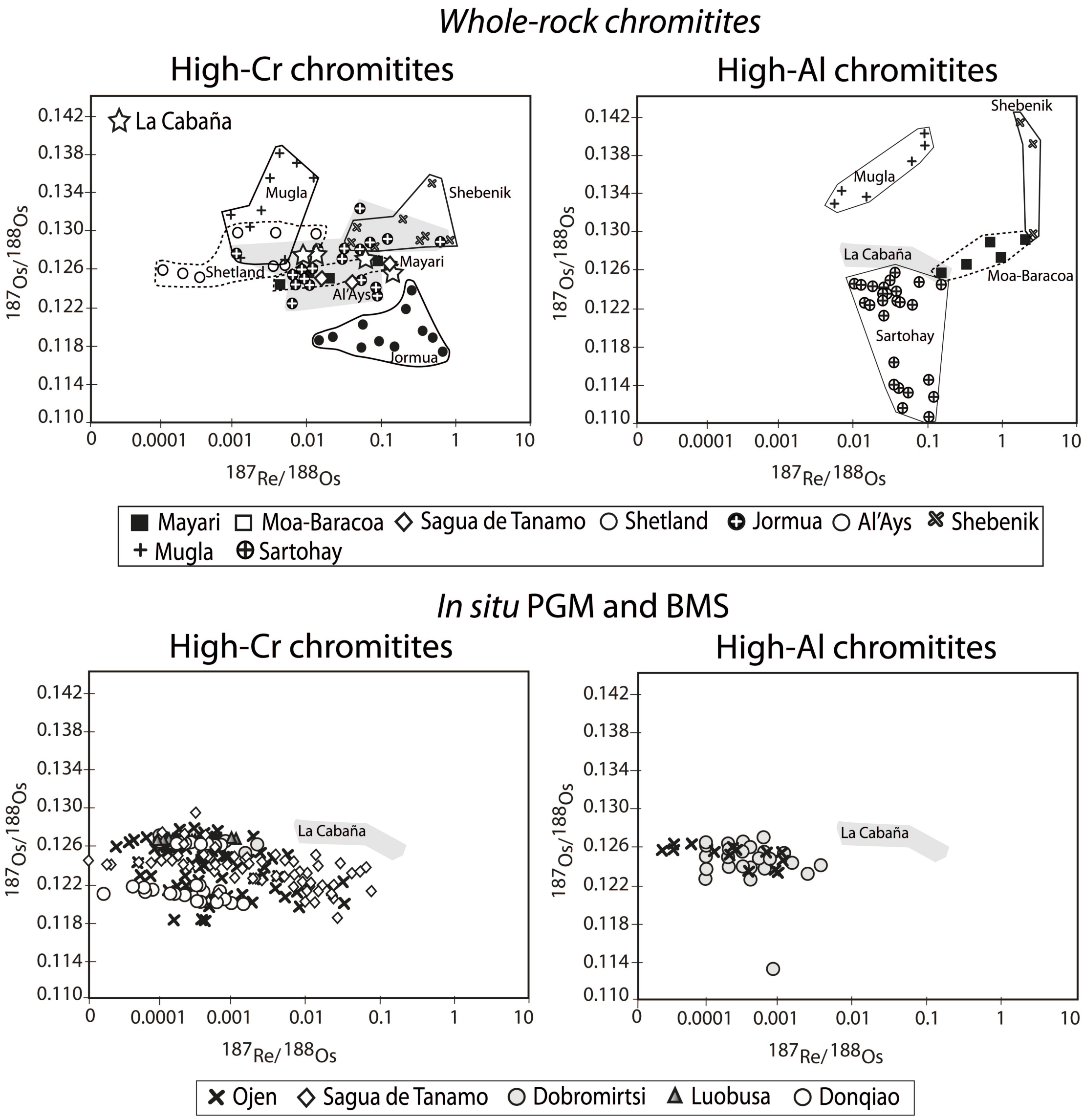
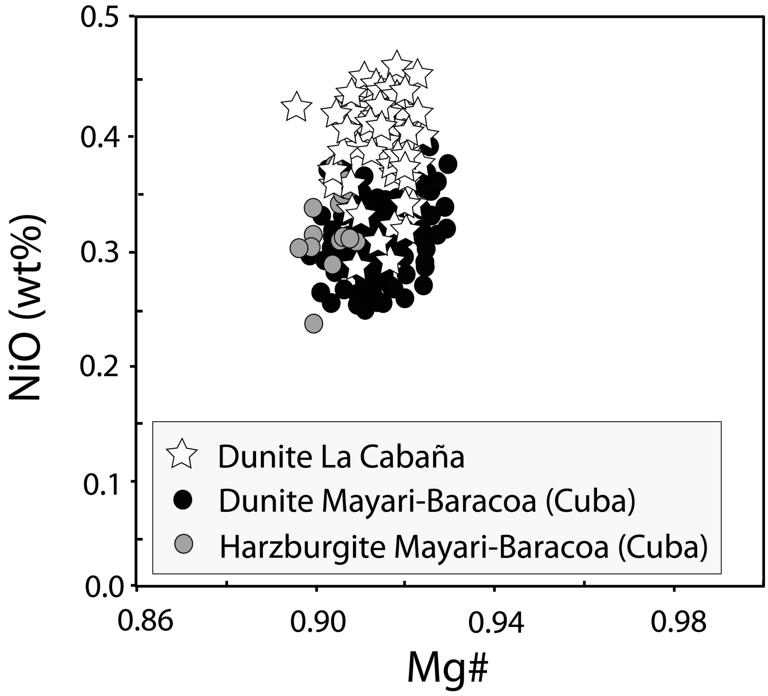

Whole-rock chromitites
High-Cr chromitites
High-Al chromitites
La Cabaña
Mugla
Shebenik
Shetland
Mayari
Al'Ays
Jormua
Sartohay
Moa-Baracoa
$^{187}Os/^{188}Os$
$^{187}Re/^{188}Os$
Mayari
Moa-Baracoa
Sagua de Tanamo
Shetland
Jormua
Al'Ays
Shebenik
Mugla
Sartohay
In situ PGM and BMS
High-Cr chromitites
High-Al chromitites
Ojen
Sagua de Tanamo
Dobromirtsi
Luobusa
Donqiao

FIGURE 8.



0.5
0.4
0.3
0.2
0.1
0.0
NiO (wt%)
0.86
0.90
0.94
0.98
Mg#
Dunite La Cabaña
Dunite Mayari-Baracoa (Cuba)
Harzburgite Mayari-Baracoa (Cuba)

FIGURE 9.

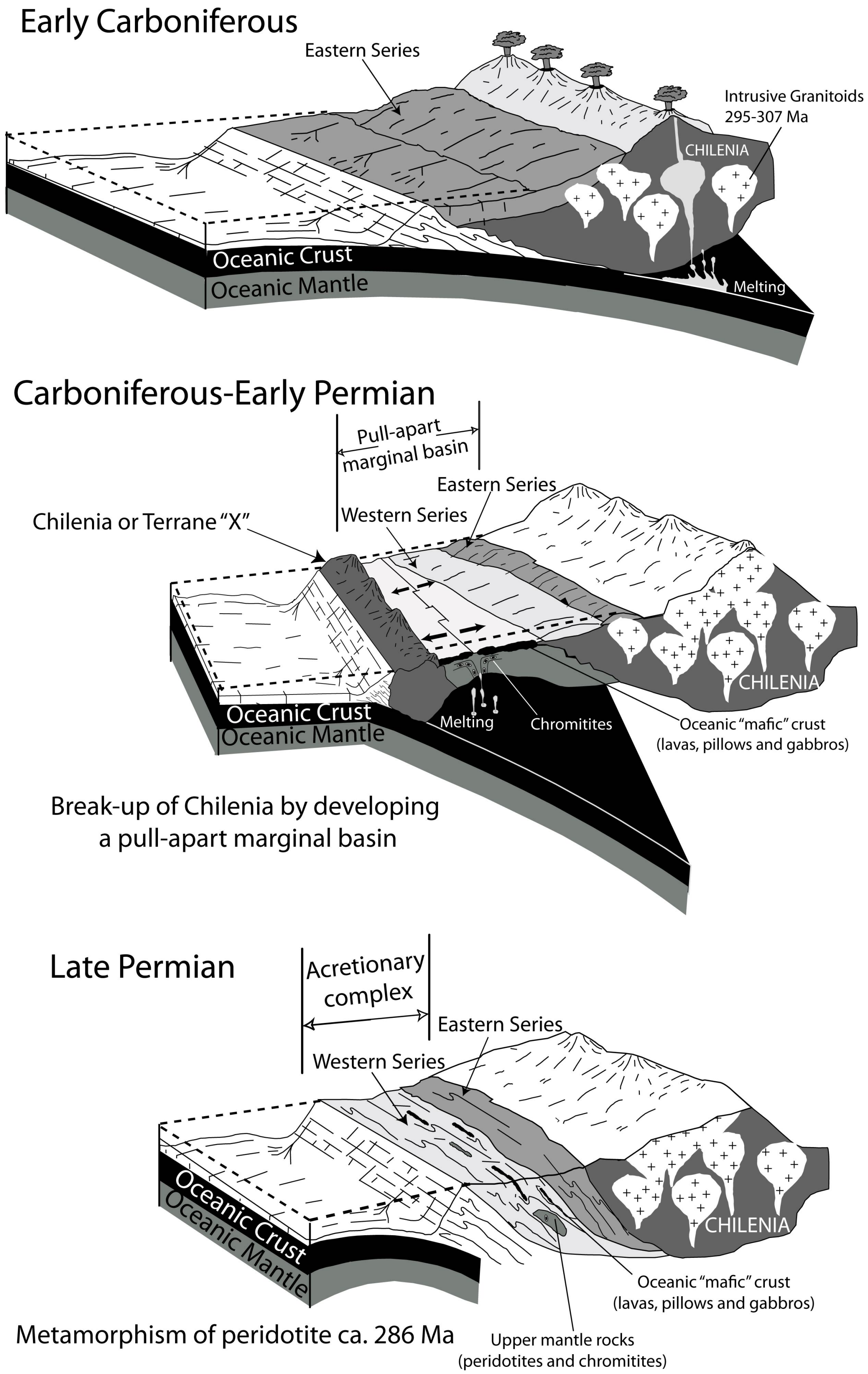

Early Carboniferous
Eastern Series
Intrusive Granitoids
295-307 Ma
CHILENIA
Oceanic Crust
Oceanic Mantle
Melting
Carboniferous-Early Permian
Pull-apart
marginal basin
Eastern Series
Chilenia or Terrane "X"
Western Series
CHILENIA
Oceanic Crust
Oceanic Mantle
Melting
Chromitites
Oceanic "mafic" crust
(lavas, pillows and gabbros)
Break-up of Chilenia by developing
a pull-apart marginal basin
Late Permian
Acretionary
complex
Eastern Series
Western Series
Oceanic Crust
Oceanic Mantle
CHILENIA
Oceanic "mafic" crust
(lavas, pillows and gabbros)
Metamorphism of peridotite ca. 286 Ma
Upper mantle rocks
(peridotites and chromitites)

Table 1. Selected representative analyses of unaltered chromite at Centinela Bajo.

	011-P2-1	011-P2-9	5BF-f1-1	5BF-f1-2	5BF-f1-4	5BF-f1-5	5BF-f1-11	5BF-f1-12	5BF-f1-13	5BF-f2-31
TiO_2	0.23	0.23	0.14	0.13	0.13	0.12	0.13	0.14	0.12	0.13
Al_2O_3	17.96	18.52	17.69	17.71	17.70	17.71	17.65	17.87	18.50	17.78
V_2O_3	0.24	0.29	0.21	0.21	0.26	0.20	0.18	0.23	0.28	0.18
Cr_2O_3	50.80	48.71	51.55	51.42	51.21	51.81	51.46	51.07	51.58	51.41
Fe_2O_3	2.13	2.56	2.39	2.38	2.57	2.60	2.32	2.24	1.59	1.91
MgO	13.54	12.09	12.79	12.66	12.82	13.34	12.64	12.78	13.80	12.42
MnO	0.32	0.48	0.39	0.41	0.36	0.40	0.37	0.41	0.41	0.36
FeO	13.65	15.66	14.98	15.12	14.90	14.16	15.12	14.88	13.54	15.43
NiO	0.17	0.14	0.16	0.10	0.21	0.05	0.17	0.13	0.08	0.09
Total	99.03	98.68	100.32	100.14	100.15	100.38	100.02	99.74	99.90	99.71
Ti	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Cr	1.26	1.22	1.28	1.28	1.27	1.28	1.28	1.27	1.27	1.28
Al	0.67	0.69	0.65	0.66	0.65	0.65	0.65	0.66	0.68	0.66
V	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe3+	0.05	0.06	0.06	0.06	0.06	0.06	0.05	0.05	0.04	0.05
Mg	0.64	0.57	0.60	0.59	0.60	0.62	0.59	0.60	0.64	0.58
Mn	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Fe2+	0.36	0.42	0.39	0.40	0.39	0.37	0.40	0.39	0.35	0.41
Ni	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00
Zn	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Cr#	0.65	0.64	0.66	0.66	0.66	0.66	0.66	0.66	0.65	0.66
Mg#	0.64	0.58	0.60	0.60	0.61	0.63	0.60	0.60	0.64	0.59
Fe^{3+}#	0.12	0.13	0.13	0.12	0.13	0.14	0.12	0.12	0.10	0.10

Table 2. Platinum-group elements of chromitite samples from La Cabaña (in ppb).

	Centinela Bajo			Lavanderos		
	PODOB	**CAB-7**	**CAB-7B**	**LA-C1**	**LA-C2**	**LA-C3**
Os	22	29	43	27	20	34
Ir	157	93.3	153	60.5	74.1	66
Ru	126	136	92	79	108	106
Rh	12.9	7	9.8	6.9	8.8	8.5
Pt	28	19	12	< 5	13	6
Pd	< 2	10	9	< 2	12	3

* Detection limits are 5 ppb for Ru and Pt, 2 ppb for Os and Pd, 0.2 ppb for Rh and 0.1 ppb for Ir.

Table 3. Re, Os concentration (in ppb) and Os isotopic data of selected chromite samples from the La Cabaña ultramafic bodies

	Re	Os	$^{187}Os/^{188}Os$	$^{187}Re/^{188}Os$	$^{187}Os/^{188}O_i$	γOs	T_{MA}	T_{RD}
CEN-07	2.230	128.0	0.12522	0.08	0.12482	-0.93	505	466
CEN-08	0.3514	202.4	0.12710	0.008	0.12706	0.85	146	149
CEN-09	0.2049	78.69	0.12703	0.01	0.12698	0.78	156	160
LAC-1008	0.3225	37.66,	0.12675	0.04	0.12655	0.44	213	221

* $^{187}Os/^{188}Os$ initial γOs, T_{MA} and T_{RD} and model ages were calculated by comparison with the Os-isotope evolution of the enstatite Chondrite (present-day ECR $^{187}Os/^{188}Os$= 0.1281±0.0004, $^{187}Re/^{188}Os$= 0.421;Walker et al., 2002a) using the parameter λ for ^{187}Re=1.555 ×10^{-11} a^{-1} from Shirey and Walker (1998). Note that T_{MA} and T_{RD} model ages are expressed in Ma.