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Petrology, mineral chemistry and geochemistry of Late Triassic Ni-Cu ore-bearing mafic-ultramafic intrusions, Hongqiling, northeastern China: petrogenesis and tectonic implications

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Abstract:

The Hongqiling magmatic Ni-Cu sulfide deposit, situated on the southern margin of the eastern Central Asian Orogenic Belt (CAOB), is composed of over 30 mafic-ultramafic intrusions. These ore-bearing intrusions composed mainly of harzburgite, lherzolite, websterite, orthopyroxenite and norite (gabbro). The constituent minerals are olivine, diopside, bronzite, calcic-hornblende, plagioclase and spinel with orthopyroxene as a dominant mineral in these intrusions. These ore-bearing intrusions are not Alaskan-type complexes. Spinel and clinopyroxene both exhibit different chemical compositions from those in the Alaskan-type complexes. The rocks that make up the intrusions have high contents of MgO (average value = 25.20 wt.%) and low TiO₂ (average value = 0.58 wt.%). The high MgO contents of the minerals and the high Mg[#] (71) of the calculated melt in equilibrium with olivine demonstrate that the parental magma of the Hongqiling mafic-ultramafic intrusions was a high-Mg tholeiitic magma. The Hongqiling ore-bearing mafic-ultramafic intrusions and the calculated “trapped liquids” for the olivine-orthopyroxene cumulate rocks are all enriched in large-ion lithophile elements and depleted in high field strength elements. The Ce/Pb, Ta/La, Th/Yb, (La/Sm)_{PM} values and the depletion of Nb and Ta suggest that the magma experienced crustal contamination. The Hongqiling ore-bearing intrusions display many similarities with mafic-ultramafic intrusion which formed in a post-collisional extensional environment in the western CAOB (e.g. Huangshanxi). Common features include their whole-rock compositions

and mineral chemistry. Combined with the evolutionary history of the eastern segment of the CAOB, we believe that the Late Triassic Hongqiling mafic-ultramafic intrusions formed in a post-collisional extensional environment.

Keywords: Mineral chemistry; High-Mg tholeiitic magma; Post-collisional extension; Mafic-ultramafic intrusion; Hongqiling

1. Introduction

A number of small Ni-Cu ore-bearing mafic-ultramafic intrusions occur in the southern part of the Central Asian Orogenic Belt (CAOB); examples include the Kalatongke (Song and Li 2009; Zhang et al. 2009a; Li et al. 2012b), Huangshandong (Han et al. 2004; Mao et al. 2015), Huangshanxi (Han et al. 2004; Mao et al. 2014), Erbutu (Peng et al. 2013) and Hongqiling (Wu et al. 2004; Lü et al. 2011; Wei et al. 2013; Hao et al. 2014a, b, 2015) intrusions (Fig. 1). The tectonic environment in which these intrusions formed is the subject of intense debate. Some researchers have suggested that these intrusions are related to a mantle plume (Zhou et al. 2004; Pirajno et al. 2008; Liu et al. 2016); others have proposed that these intrusions are typical Alaskan-type complexes, which form in subduction zones (Han et al. 2007; Xiao et al. 2009); and still others have suggested that these intrusions formed in a post-collisional extensional environment (Wu et al. 2004; Tang et al. 2011; Sun et al. 2013b, c; Hao et al. 2015).

The Hongqiling Ni-Cu ore-bearing mafic-ultramafic intrusions are located on the southeastern margin of the CAOB (Fig. 1). These intrusions provide important data that inform discussions of the genesis and tectonic environment of the mafic-ultramafic intrusions in the CAOB. Many studies of the Hongqiling ore-bearing intrusions have focused on determining the ages of the rocks; whole-rock elemental and isotopic compositions have recently been measured (Wu et al. 2004; Xi et al. 2006; Feng et al. 2011; Lü et al. 2011, 2012; Wei et al. 2013; Hao et al.

2014a, b, 2015). These studies have greatly improved our understanding of the origin of the Hongqiling ore-bearing intrusions and the formation of sulfide ores in the magmatic system. However, studies of mineralogy and mineral chemistry have not received much attention; in particular, no work has been conducted on the trace elements in the minerals to date.

Primary and accessory minerals (e.g. olivine, spinel and clinopyroxene) play an important role in helping to identify the tectonic setting and the physical and geochemical conditions for Ni-Cu ore-bearing mafic-ultramafic intrusions develop (Dick and Bullen 1984; Nimis and Ulmer 1998; Barnes and Roeder 2001; Downes et al. 2003; Driouch et al. 2010; Liu et al. 2015, 2017b). For example, spinel is a useful petrological and geodynamic indicator in mafic-ultramafic intrusions (Irvine 1965; Dick and Bullen 1984; Liipo et al. 1995; Barne sand Roeder 2001). Olivine is a predominant mineral, and its composition (e.g. MgO and FeO_T) can be used to estimate the compositions of parental magmas (Roeder and Emslie 1970). Clinopyroxene is another predominant mineral in mafic-ultramafic intrusions, and its contents can also indicate the characteristics of magmas (Kushiro 1960; Loucks 1990; Xue et al. 2015).

In this study, we combine new whole-rock and mineral composition data with previously published zircon U-Pb ages to evaluate the characteristics and composition of the parental magma of the Hongqiling Ni-Cu ore-bearing mafic-ultramafic intrusions. We also constrain the petrogenesis and tectonic setting of these intrusions.

2. Geological setting

The CAO B represents one of the most important sites of juvenile crustal growth during the Phanerozoic (Jahn 2004). The eastern part of the CAO B evolved from the amalgamation of several micro-continental blocks and/or terranes in the Paleozoic (Wu et al. 2011). From west to east, it is divided into the Erguna Block, the Xing'an Block, the Songnen-Zhangguangcai Range, the Jiamusi Massif, and the Khanka Massif (Fig. 2a). More than 30 mafic-ultramafic intrusions are located in the Hongqiling area (Fig. 2c). They trend northwest-southeast and are separated by a series of subparallel northwest-trending faults (Wu et al. 2007). These intrusions intruded into the Early Paleozoic Hulan Group, which contains gneiss, amphibolite, mica schist and marble (Shi and Lan 1985) and has a metamorphic grade that ranges from the greenschist facies to the amphibolite facies (Fig. 2b and 2c).

The Hongqiling metallogenic belt is divided into three mafic-ultramafic belts. These belts are Nos. I, II and III from southwest to northeast (Fig. 2c). However, only several intrusions that display Ni-Cu sulfide mineralization have been found to date in belt I (Fig. 2c). The ore-bearing intrusions, such as Hongqiling Nos. 1, 2, 3 and 7, are exposed over an area of approximately 0.5–0.7 km². The No. 1 and No. 7 intrusions contain medium and large Ni resources, respectively (Lü et al. 2011). In plan view, the Hongqiling No. 1 intrusion is an irregular lens in shape (Fig. 3a). The intrusion trends northwest, and it is 980 m in length and 150–280 m in width

([Yang et al. 2005](#)). The No. 1 intrusion intrudes into the Hulan Group and consists of olivine pyroxenite, ilherzolite, pyroxenite and gabbro ([Qin 1995](#); [Wu et al. 2004](#)). The boundaries between rock units are gradational. The average Ni and Cu grades of the ores are 0.65 wt.% and 0.13 wt.%, respectively. The Hongqiling No. 2 intrusion trends northwest and is 1700 m in length and 10–350 m in width ([Hao et al. 2012](#)). The intrusion consists of pyroxene peridotite, pyroxenite and gabbro, and the main rock type is gabbro. The boundaries between rock types are gradational ([Fig. 3b](#)). The average Ni grade of the ores ranges from 0.40 to 0.60 wt.%. The Hongqiling No. 3 intrusion trends north-south and is 2500 m long and 40–500 m wide. The intrusion consists of olivine pyroxenite, gabbro and amphibole pyroxenite. The boundaries between rock types are gradational ([Fig. 3c](#)). The average grades of its Ni and Cu ores are 0.39 wt.% and 0.11 wt.%, respectively. The Hongqiling No. 7 intrusion trends northwest; it is 700 m long and 30 m wide and has average Ni and Cu grades of 2.31 wt.% and 0.63 wt.%, respectively ([Qin 1995](#)). The main rock type of the intrusion is orthopyroxenite. A small harzburgite occurs within the orthopyroxenite unit. Minor norite is present in places at the margin of the intrusion ([Fig. 3d](#)). Most of the Hongqiling ore-bearing mafic-ultramafic intrusions have relatively confirmed U-Pb zircon ages based on studies in recent years. They were formed in the Late Triassic based on recent reports ([Table 1](#)).

3. Sample description

All the samples in this paper were collected from the Hongqiling Nos. 1 (42° 53' 55" N, 126° 25' 29 " E), 2 (42° 54' 6" N, 126° 24' 47" E), 3 (42° 53' 44" N, 126° 25' 6" E) and 7 (42° 52' 5" N, 126° 29' 35" E) intrusions. Representative photomicrographs of typical samples are shown in [Fig. 4](#).

Harzburgite is massive structure and mainly composed of olivine (40–60 vol.%), orthopyroxene (20–30 vol.%), clinopyroxene (5–10 vol.%) and minor amounts of plagioclase, and spinel ([Figs. 4a and 4b](#)). Olivine grains are generally anhedral granular, and orthopyroxene grains are subhedral to anhedral. Olivine grains are partially altered to serpentine ([Fig. 4b](#)). Some spinels and sulfide inclusions are hosted by olivine or pyroxene grains ([Figs. 4h and 4i](#)).

Lherzolite is massive structure and mainly composed of olivine (40–50 vol.%), orthopyroxene (15–20 vol.%), clinopyroxene (15–20 vol.%) and hornblende (5–10 vol.%). Hornblende occurs as overgrowth rim of pyroxene ([Figs. 4c and 4d](#)). It commonly shows a poikilitic texture characterized by the occurrence of multiple small rounded olivine inclusions within pyroxene ([Fig. 4c](#)).

Websterite is massive structure and mainly composed of orthopyroxene (40–45 vol.%), clinopyroxene (35–45 vol.%) and hornblende (5–10 vol.%) ([Fig. 4e](#)). It displays granular texture ([Fig. 4e](#)). Orthopyroxene and clinopyroxene grains are subhedral to euhedral granular.

Orthopyroxenite is massive structure and mainly composed of orthopyroxene (55–65 vol.%), clinopyroxene (15–25 vol.%) and plagioclase (5–10 vol.%). Some of the orthopyroxene has been altered to tremolite. Minor amounts of clinopyroxene and hornblende occur in the interstitial spaces (Fig. 4f). Orthopyroxene grains are subhedral to euhedral granular. Clinopyroxene occur as anhedral granular.

Norite is massive structure and mainly composed of orthopyroxene (25–40 vol.%), clinopyroxene (15–25 vol.%), plagioclase (20–30 vol.%) and minor amounts of hornblende (Fig. 4g). Orthopyroxene grains are subhedral to anhedral granular. Plagioclase grains are in anhedral granular.

4. Analytical methods

The major elements were measured by a Shimadzu XRF-1800 instrument at the Yanduzhongshi Geological Analysis Laboratories Ltd. The loss-on-ignition (LOI) was measured as the weight loss of the samples after 1h baking at a constant temperature at 1000 °C. The powdered samples, weighing approximately 1.2 g were fused with lithium tetraborate ($\text{Li}_2\text{B}_4\text{O}_7$, 6 g) at 1050 °C for 20 min. The accuracy and reproducibility were monitored by Chinese national standard sample GSR3. The analytical precision for most elements was better than 1%. Trace elements were analyzed using a Perkin–Elmer Sciex ELAN 6000 ICPMS at the

Yanduzhongshi Geological Analysis Laboratories Ltd, following procedures as described by Qi et al. (2000). The analytical precision for trace elements was better than 5%.

The major elements of minerals were determined by wavelength-dispersive X-ray emission spectrometry using a JEOL-JXA8230 electron microprobe at the Institute of Mineral Resources, Chinese Academy of Geological Sciences. The operating condition was 15 kV acceleration voltages with 10 nA beam current and a beam diameter of 1 μm . Trace elements in spinel, olivine, clinopyroxene and orthopyroxene were determined by laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) at the National Research Center for Geoanalysis. The analytical procedures, operating conditions, and data acquisition and reduction are given in Liu et al. (2008).

5. Results

5.1. Major element compositions of minerals

Representative compositions of spinel, olivine, clinopyroxene and orthopyroxene from the Hongqiling mafic-ultramafic intrusions are listed in Supplementary Table 1 (Table S1).

5.1.1. Spinel

Spinel grains from the Hongqiling ore-bearing mafic-ultramafic intrusions mainly occur in

harzburgite and lherzolite. They are small in size, usually hosted by other minerals (e.g. olivine, pyroxene and hornblende). Spinel contains 23.05–41.56 wt.% Al_2O_3 and 22.61–38.42 wt.% Cr_2O_3 . The $\text{Cr}^\#$ [$\text{Cr}^\# = 100 \text{ Cr} / (\text{Cr} + \text{Al})$] and $\text{Mg}^\#$ [$\text{Mg}^\# = 100 \text{ Mg} / (\text{Mg} + \text{Fe}^{2+})$] of spinel vary from 27 to 53 and from 28 to 57, respectively. The composition of spinels in these intrusions is similar to those of the Huangshandong and Huangshanxi intrusions (Mao et al. 2014, 2015), but is significantly different from the compositions of those from the Alaskan-type complexes, boninites and komatiites (Barnes and Roeder 2001) (Fig. 5).

5.1.2. Olivine

There are no significant variations within individual olivine grains. They are characterized by low SiO_2 (39.33–39.68 wt.%), FeO_T (10.63–13.94 wt.%), NiO (0.06–0.19 wt.%), CaO contents (0.00–0.10 wt.%) and high MgO (44.03–48.2 wt.%) contents (Table S1). The Fo values vary from 85 to 89. The olivine from the Hongqiling ore-bearing mafic-ultramafic intrusions is characterized by the depletion in Ca (<1000 ppm), similar to those from the coeval mafic-ultramafic intrusions in the orogenic belts (e.g. Huangshandong, Huangshanxi, Erhongwa and Poyi; Sun et al. 2013b; Sun et al. 2013c; Mao et al. 2014, 2015; Liu et al. 2015; Xue et al. 2016). The Ca contents in olivines from Hongqiling ore-bearing intrusions are much lower than those from OIB (ocean island basalts), CFB (continental flood basalts) and komatiite (Sobolev et al. 2007) (Fig. 6a).

5.1.3. Pyroxene

Orthopyroxenes are mainly bronzite and clinoenstatite (Fig. 6b), with Mg[#] from 83.2 to 88.0, containing 1.12–4.90 wt.% Al₂O₃, and 0.02–0.67 wt.% TiO₂ (Table S1). Clinopyroxene varies from diopside to augite in composition (Fig. 6b), with Mg[#] ranging from 83 to 86. It contains 1.96–5.85 wt.% Al₂O₃ and 0.18–0.94 wt.% TiO₂ (Table S1). The CaO contents of clinopyroxene from the Hongqiling intrusion (average 19.7 wt.%, <22 wt.%; Table S1) are very similar to those from the Huangshanxi intrusion (<22 wt.%, Mao et al. 2014), but are significantly lower than those from the Xiadong Alaskan-type ultramafic intrusion (~25 wt.%, Su et al. 2012).

5.1.4. Other minerals (hornblende and plagioclase)

According to the hornblendes classification scheme of Leake et al. (1978), hornblendes are compositionally classified as pargasite and edenite (unpublished data). Plagioclase occurs in all kinds of rocks, but is different in modal abundance. These plagioclases are albite, oligoclase, andesine and labradorite (unpublished data).

5.2. Trace element compositions of minerals

The abundances of trace elements in spinel, olivine and pyroxene from the Hongqiling ore-bearing mafic-ultramafic intrusions are listed in Table 2.

The abundances of the incompatible trace elements are higher in clinopyroxene than in the

other minerals. Many incompatible trace elements are below the detection limits in the spinel and olivine. The spinel is enriched in transition metals, such as V, Cr, and Zn. In Fig. 7a, the pattern defined by the spinel from Hongqiling differs from the pattern of komatiite. The olivine is enriched in transition metals, such as Sc, V, Cr, Co, Ni and Zn. The primitive mantle-normalized trace element patterns for clinopyroxene from the Hongqiling ore-bearing mafic-ultramafic intrusions are similar to those from the Huangshanxi ore-bearing mafic-ultramafic intrusion (Mao et al. 2014) and show Nb depletion (Fig. 7). Both the orthopyroxene and clinopyroxene are depleted in Nb and Ta (Fig. 7; Table 2).

5.3. Major and trace elements of whole rocks

Major and trace element analyses of the Hongqiling ore-bearing mafic-ultramafic intrusions are listed in Tables 3 and 4. The loss on ignition (LOI) values of these samples are variable. For comparison, we have recalculated the whole-rock raw data to 100% on an anhydrous basis. The following discussion employs the normalized values. The SiO₂ contents of these samples range from 42.95 wt.% to 53.74 wt.%, whereas the ranges of the TiO₂, Na₂O, K₂O, FeO_T and MgO contents are 0.35–0.83 wt.%, 0.37–2.11 wt.%, 0.22–1.48 wt.%, 6.12–11.40 wt.%, and 12.35–33.35 wt.%, respectively; the Mg[#] values fall within the range of 80.88–86.85. On a Harker variation diagram (Fig. 8), all of the samples show negative correlations among SiO₂,

CaO and MgO, whereas a positive correlation between MgO and FeO_T is observed.

On the chondrite-normalized rare earth element (REE) and the primitive mantle-normalized trace element diagrams, all of the samples show similar trace element patterns. They are enriched in large-ion lithophile elements (LILEs; Th and La) and depleted in high field strength elements (HFSEs; Nb, Ta, Zr and Hf) (Fig. 9a). These trace and REE element patterns are similar to those of the rocks from Huangshanxi (Mao et al., 2014), but are significantly different from OIB (Sun and McDonough 1989) (Fig. 9). These rocks have relatively low Σ REE concentrations (13.02–79.61 ppm) and slight enrichment in the light REEs relative to the heavy REEs (Fig. 9b). The trace element patterns of all of our samples show negative Nb and Ta anomalies (Fig. 9a), and weak negative Eu anomalies (Fig. 9b).

6. Discussion

6.1. Effect of alteration

The close spatial relationship and similar trace element patterns of the Hongqiling ore-bearing mafic-ultramafic intrusions suggest that they likely have a common mantle source and underwent similar magmatic processes in the same tectonic setting. All of these rocks suffered variable alterations (e.g. serpentinization and amphibolization), which is consistent with

the variable LOI values (1.51–6.65 wt. %). The REEs and the HFSEs are relatively immobile during low-grade metamorphism or hydrothermal alteration (e.g. [Staudigel and Hart 1983](#)), whereas the Cs, Rb, Ba, K, and Sr elements are relatively mobile. In the [Fig. 10](#), the “immobile” elements (La, Hf, Nb, Y, Th and Ta) experienced little change in correlation, whereas the “mobile” elements (Sr and Pb) exhibited negligible correlations. This trace element characteristics of the Hongqiling ore-bearing intrusions suggest that Hongqiling ore-bearing intrusions experienced the low-grade metamorphism or hydrothermal alteration. In addition, the Pd/Ir ratios of Hongqiling rocks are very low ([Wei et al. 2013](#); [Hao et al. 2014b](#)), also suggesting hydrothermal alteration had a very small impact on the Hongqiling ore-bearing intrusions. Based on the mobility of elements, we choose HFSEs and REEs (e.g. Th, Zr, Hf, Nb, Ta and Y) to highlight the magma evolution process.

6.2. Fractional crystallization and crustal contamination

Fractional crystallization is an important controlling factor in magmatic evolution ([Ripley and Li 2013](#); [Liu et al. 2017a](#)). Poikilitic texture is commonly considered to be evidence of fractional crystallization, and it is very commonly observed in the mafic-ultramafic rocks in the Hongqiling ore-bearing intrusions under the microscope ([Fig. 4](#)). Examples include spinel grains hosted by olivine or pyroxene ([Fig. 4h and 4i](#)), olivine hosted by orthopyroxene or clinopyroxene

(Fig. 4a, 4b and 4c), and orthopyroxene hosted by clinopyroxene (Fig. 4b). These features indicate that fractional crystallization of spinel and olivine occurred in the early stages of evolution of the magma, followed by orthopyroxene, clinopyroxene and other accessory minerals (e.g. plagioclase, hornblende and mica). The positive correlation between FeO_T and MgO in bulk rock samples suggests that they formed by fractionation crystallization of olivine and pyroxene (Fig. 8b). Al_2O_3 and CaO are negatively correlated with MgO (Fig. 8c and 8d), suggesting that pyroxene or plagioclase played an important role in the evolution of the magma. The weak negative Eu anomalies in REE patterns were likely caused by the crystallization of plagioclase. As shown in Fig. 8, the major element compositions of the Hongqiling mafic-ultramafic rocks are mainly controlled by the abundances of olivine and pyroxenes, indicating that the mafic-ultramafic rocks are olivine-pyroxene cumulates. Taken together, olivine, orthopyroxene and clinopyroxene were likely the principal crystallized phases in the magma of these mafic-ultramafic rocks.

Crustal contamination can introduce sulfide saturation in the mantle magmas that give rise to ore-bearing mafic-ultramafic intrusions (Ripley and Li 2013). The presence of abundant orthopyroxene indicates that the magma was oversaturated with SiO_2 , suggesting that SiO_2 -rich crustal material may have been introduced into the magma of the Hongqiling ore-bearing intrusions. The values of Ce/Pb do not change during partial melting and magmatic differentiation, and the mantle generally has high Ce/Pb ratios (25 ± 5), whereas the continental

crust exhibits relatively low values (<15) (Hofmann 1988). The Ce/Pb ratios of the Hongqiling ore-bearing intrusions range from 1.1 to 3.9, with an average of 3.2. These Ce/Pb values differ appreciably from the ratios of the mantle but are similar to the ratios of continental crust, suggesting that the Hongqiling ore-bearing mafic-ultramafic intrusions experienced crustal contamination. Similarly, the Ta/La ratios of the Hongqiling ore-bearing intrusions (average value = 0.02) are lower than the ratios of primitive mantle (0.06, Wood et al. 1979) indicating that the rocks experienced the crustal contamination during the rock-forming stages. Moreover, on the Th/Yb-Nb/Yb diagram (Pearce 2008), our samples fall into the magma-crust interaction zone (Fig. 11a), which also indicates crustal contamination. This conclusion is further supported by initial $^{187}\text{Os}/^{188}\text{Os}$ ratios (0.34 to 3.02) measured in the Hongqiling ore rocks and reported by previous workers (Wei et al. 2013; Sun 2013a; Hao et al. 2014b). The values of $^{187}\text{Os}/^{188}\text{Os}$ are much higher than those of Cu-Ni sulfide deposits free of crustal contamination (e.g. the Kambalda Cu-Ni sulfide deposit, which displays $^{187}\text{Os}/^{188}\text{Os}$ values of 0.10889 ± 0.00035 ; Foster et al. 1996). All of these evidences indicate that Hongqiling ore-bearing mafic-ultramafic intrusions have experienced crustal contamination.

The trace elements ratios are closer to those of the upper crust, indicating that the majority of the material came from the upper crust (Figs. 11c and 11d). However, this is inconsistent with the very positive and high $\epsilon\text{Nd}(t)$ values of the Hongqiling ore-bearing intrusions (1.63–10.34, Wu et al. 2004; Wei et al. 2013; Sun et al. 2013a). Because the old upper crust has very negative

$\epsilon\text{Nd}(t)$ values. It may have resulted from mixing with partial melts derived from the lower part of a juvenile crust. Partial melting of juvenile crust due to mafic magma underplating is widely believed to be responsible for the formation of abundant A-type granites that are characterized by positive $\epsilon\text{Nd}(t)$ values and highly negative Nb anomaly in the region (e.g. [Wu et al. 2002](#); [Sun et al. 2005](#)).

6.3. Geochemical characteristics of the parental magma

Because the fractional crystallization of olivine and spinel occurred in the early stage of the magma evolution, their crystallization temperature can be taken to approximate the lower limit of the temperature of the parental magma. Based on partition coefficients between olivine and melt ([Roder and Emslie 1970](#); [Weaver and Langmuir 1990](#)),

$$T_{\text{ol}} = 1066 + 12.06 \text{ Mg}^{\#} + 312.3 (\text{Mg}^{\#})^2, \quad (1)$$

The crystallization temperature of olivine is 1220–1324°C (average 1308°C).

The crystallization temperature of spinel can be calculated by the following equation ([Fabries 1979](#)):

$$T_{\text{sp}} (^{\circ}) = (4250 * Y_{\text{Cr}^{\text{sp}}} + 1343) / (\text{Ln } K_{\text{Cr}^{\text{o}}} + 1.825 * Y_{\text{Cr}^{\text{sp}}} + 0.571), \quad (2)$$

where $\text{Ln } K_{\text{Cr}^{\text{o}}} = 0.34 + 1.06 * (Y_{\text{Cr}^{\text{sp}}})^2$, and $Y_{\text{Cr}^{\text{sp}}} = \text{Cr}^{3+} / (\text{Cr}^{3+} + \text{Al}^{3+} + \text{Fe}^{3+})$

The crystallization temperature of spinel is 1391–1412°C (average 1405°C). The above

crystallization temperatures of olivine and spinel suggest that the temperature of the parental magma likely exceeded 1308°C (average).

The parental magma of Ni-Cu sulfide deposits is usually komatiitic or basaltic magma (Keays 1995; Naldrett 1999). The mineral composition is affected by the temperature, the pressure and their parental magma compositions. Komatiitic magma is derived from a higher degree of partial melting of mantle than basaltic magma. Komatiitic magma has a higher Mg content, temperature and pressure than basaltic magma. Generally, the $Cr^{\#}$ values of spinels in komatiites are higher than spinels formed from basaltic magma. Figs. 5 and 7 show that the geochemical signature of spinel from Hongqiling is distinctly different from spinel formed from komatiites (Barnes and Roeder 2001; Yao 1999; Pagé and Barnes 2009). Moreover, Ca contents in olivines from Hongqiling ore-bearing intrusions (<1000 ppm) are much lower than those from komatiite (Sobolev et al. 2007). These suggest that the parental magma of the Hongqiling ore-bearing intrusions was not komatiitic magma. Additionally, the Si and Al indices in clinopyroxene can be used as effective tracers to indicate the characteristics of magma (Kushiro 1960). The SiO_2 and Al_2O_3 contents of most of the clinopyroxenes from the Hongqiling ore-bearing intrusions reflect non-alkaline characteristics (Leterrier et al. 1982) (Fig. 12a), and most of their compositions plot in the field of tholeiitic magma in the diagram of Al^{IV} vs. TiO_2 (Fig. 12b). All of the above evidences imply that the parental magma of the Hongqiling ore-bearing mafic-ultramafic intrusions was tholeiitic magma.

6.4. Composition of the parental magma

6.4.1. Calculation of major elements contents in the parental magma

Chilled margins have not been identified in the Hongqiling ore-bearing mafic-ultramafic intrusions. Therefore, the composition of the parental magma can only be approximated. We estimate the major element compositions of the parental magma based on the compositions of spinel and olivine.

We assume that the spinels that have the highest Cr_2O_3 content and are enclosed in olivine are primary, which retains their initial compositions, following the empirical equation of [Maurel and Maurel \(1982\)](#):

$$(\text{Al}_2\text{O}_3)_{\text{sp}} \% = 0.035 \times (\text{Al}_2\text{O}_3 \%)_{\text{liq}}^{2.42} \quad (3)$$

In this way, we determine that the equilibrated Al_2O_3 content of the Hongqiling parental magma ranges from 14.61 to 18.64 wt.%.

On the other hand, the partitioning of Fe and Mg between olivine and its parental melt can be used to estimate the $\text{Mg}^\#$ molar ratio of the melt in equilibrium with olivine. [Roeder and Emslie \(1970\)](#) provided the following expression that describes the partition coefficient between olivine and melt (in mol):

$$K_d = (\text{Fe} / \text{Mg})^{\text{Ol}} / (\text{Fe} / \text{Mg})^{\text{magma}} = 0.3 \pm 0.03 \quad (4)$$

Using the most Mg-rich olivine composition from the Hongqiling mafic-ultramafic intrusions (Fo = 89) and the above partition coefficient, we estimate that the melt in equilibrium with the olivine in the Hongqiling ore-bearing mafic-ultramafic intrusions contains a $Mg^{\#}$ value of 71. Moreover, the $Mg^{\#}$ of mineral compositions from the Hongqiling mafic-ultramafic rocks reveals that they are Mg-rich (Tables S1). Thus, we think that the parental magma of the Hongqiling ore-bearing mafic-ultramafic intrusions was high-Mg tholeiitic magma.

6.4.2. Calculation of trace element contents in the “trapped liquids”

The trace element composition of parental magma can be estimated by the trace element partition coefficients between melt and clinopyroxene for olivine-clinopyroxene cumulate rocks (e.g. Tian et al. 2011; Li et al. 2013; Tang et al. 2013). However, this method is not applicable to olivine-orthopyroxene cumulate rocks, such as the harzburgite found in the Hongqiling ore-bearing intrusions.

The trace elements of the equilibrium melt with the Hongqiling olivine-orthopyroxene cumulate rocks can be estimated from whole-rock compositions using the method of Godel et al. (2011):

$$C^i_{WR} = F_{Opx\ (initial)} \times C^i_{Opx} + F_{Ol} \times C^i_{Ol} + F_{Liq} \times C^i_{Liq} \quad (5)$$

where C^i_{WR} , C^i_{Opx} and C^i_{Ol} are the concentrations of the element i in the whole-rock and the cumulus orthopyroxene and olivine, respectively. In our calculation, we assumed the concentrations of trace elements in olivine and orthopyroxene to be zero when they are below the

detection limits of LA-ICP-MS. The errors caused by this assumption are negligible because the concentrations of these elements in the minerals of interest are two orders lower than those in the whole-rocks. F_{Opx} (initial) is the fraction of cumulus orthopyroxene prior to post-cumulus overgrowth and F_{Ol} is the fraction of olivine (for the pyroxenite $F_{\text{Ol}} = 0$) (Godel et al. 2011). We use back-scattered electron images to measure the proportion of the minerals. As the exchange of trace elements between olivine with liquid is negligible, we choose the samples with a higher proportion of olivine and lower content of hornblende and plagioclase. This can ensure that there was a sufficient amount of “trapped liquids” in the samples before solidification and can minimize the effects of hornblende and plagioclase in the samples. The calculated trace elements of the “trapped liquids” are listed in Table 4 and plotted in Figs. 9a and 9b.

Th and Nb are incompatible elements in olivine and orthopyroxene, and they do not enter into the olivine and orthopyroxene during the fractional crystallization (Table 2). The contents of Yb in olivine and orthopyroxene are much lower than that in whole-rock (Tables 2 and 4). Therefore, the ratios of Th/Yb and Nb/Yb for olivine-orthopyroxene cumulates should be close to those of the “trapped liquids”. The similarity of the Th/Yb and Nb/Yb ratios of the whole-rock samples and the calculated ratios of “trapped liquids” by the equations of Godel et al. (2011) indicate that the method we used above is correct and feasible (Fig. 11a).

We use their mantle-normalized patterns for comparison. The estimated “trapped liquids” compositions are enriched in LILEs (Th) and depleted in HFSEs (Nb, Ta, Zr and Hf) relative to

primitive mantle (Fig. 9a). The estimated “trapped liquids” compositions of the Hongqiling olivine-orthopyroxene cumulate rocks are remarkably similar to the estimated compositions of “trapped liquids” from the Huangshanxi olivine-orthopyroxene cumulate rocks in the western CAOB (Mao et al. 2014), suggesting that Hongqiling and Huangshanxi rocks maybe formed in the same tectonic environment and have similar magma sources. All the trace element compositions of the mafic-ultramafic rocks, orthopyroxene, clinopyroxene and the “trapped liquids” display characteristic negative Nb and Ta anomalies, implying that the parental magma for the Hongqiling intrusion was depleted in Nb and Ta, and the initial magma might have experienced some degree of crustal contamination.

6.5. Tectonic setting and its evolution

The formation of the ore-bearing mafic-ultramafic intrusions in the CAOB has been suggested to have taken place in several different tectonic settings, including: (1) in association with a mantle plume (Zhou et al. 2004; Pirajno et al. 2008; Liu et al. 2016), (2) Alaskan-type complexes, which form in subduction zones (Han et al. 2007; Xiao et al. 2009), and (3) a post-collisional extensional environment (Wu et al. 2004; Tang et al. 2011; Sun et al. 2013b, c; Hao et al. 2015).

Most OIBs are considered to be the expression of plume tails, whereas continental and

oceanic large igneous provinces with chemistry similar to that of OIBs are generally interpreted to represent plume-head events (Campbell and Griffiths 1990; Condie 2001; Ernst and Buchan 2003; Saunders et al. 2005). The trace and REE element patterns of the Hongqiling ore-bearing intrusions are significantly different from OIBs (Sun and McDonough 1989) (Fig. 9). On the Nb/Yb-Th/Yb diagram and the Th/Nb-Ce/Nb diagram (Figs. 11a and 11b), the samples plot far from the OIB area. The above evidences suggest that the Hongqiling ore-bearing mafic-ultramafic intrusions have little connection with mantle plume. In addition, the magma derived from mantle plumes is always accompanied by large-scale radial dyke swarms and layered intrusions (Coffin and Eldholm 1994). However, this is not the case for the Hongqiling ore-bearing mafic-ultramafic intrusions. They occupy only small areas (0.5–0.7 km²) (Figs. 2 and 3), and the related magmatism is insufficiently intense. Moreover, no mantle plume has been found in northeast China at present. Thus, the mantle plume origin can be ruled out.

Alaskan-type complexes are generally thought to have formed in a subduction zone environment (Johan 2002; Pettigrew and Hattori 2006; Thakurta et al. 2008). Morphologically, these intrusions all form linear arrays along major tectonic sutures and intruded during terrane accretion, and these intrusions are usually among 12–40 km² (Johan 2002). They have a concentric zoning structure in a petrographic context. In terms of mineral assemblage, clinopyroxene is common, and orthopyroxene is either missing or occurs in very low abundance in the Alaskan-type complexes; magnetite is a common mineral, with local abundances as high

as 15~20% by volume (Taylor 1967; Himmelberg and Loney 1995). The mineral chemistry of the Alaskan-type complexes is characterized by high Fe-Cr, low Al chromite and Ca-rich diopsidic clinopyroxene with a wide range in composition (Irvine 1974; Rublee 1994; Helmy and Mahallawi 2003). In fact, the morphology, mineralogy and mineral chemistry of the Hongqiling ore-bearing mafic-ultramafic intrusions are distinct from those of Alaskan-type complexes. For example: (1) The Hongqiling ore-bearing intrusions are very small, and they occupy only small areas (0.5–0.7 km²) (Figs. 2 and 3). (2) The Hongqiling ore-bearing intrusions contain large amounts of orthopyroxene and plagioclase, but they contain a small amount of magnetite. (3) Spinel grains in the Hongqiling ore-bearing intrusions exhibit different Cr-Fe³⁺-Al compositions from those in the Alaskan-type complexes (Fig. 5). (4) The CaO contents of clinopyroxene from the Hongqiling intrusion (< 22 wt.%, Table S1) are significantly lower than those from the subduction-related Xiadong ultramafic intrusion (~25 wt.%, Su et al. 2012). All of the above evidences imply that the Hongqiling ore-bearing mafic-ultramafic intrusions have little connection with Alaskan-type complexes that was emplaced during subduction.

The whole-rock compositions and mineral chemistry (e.g. spinel and olivine) for the Hongqiling ore-bearing intrusions are similar to those of the Huangshanxi intrusion, which was considered to be formed in a post-collisional extensional environment. Similarly, the estimated “trapped liquids” compositions of the Hongqiling olivine-orthopyroxene cumulate rocks are remarkably similar to those of the Huangshanxi olivine-orthopyroxene cumulate rocks in the

western CAO (Mao et al. 2014), suggesting that both formed in post-collisional extensional environments. In fact, the disappearance of the marine strata occurred after the Early Triassic, indicating the end of Paleo-Asian Ocean subduction (Peng et al. 2012). The occurrence of molasse is an effective indicator of the beginning of an orogeny. Late Triassic molasse is widespread in the east of the CAO, and it rests unconformably on middle-Late Triassic granites (Peng et al. 2012; Zhao et al. 2012), suggesting that the study area entered the orogenic stage in the Late Triassic. Moreover, the A-type granite is generally considered to be formed in an extensional environment (Martin. 2006; Konopelko et al. 2007), and a number of Late Triassic A-type granites occurred in the study area (Wu et al. 2002; Zhang et al. 2018), suggesting that the study area was in post-collision extensional environment in the Late Triassic (Wu et al. 2002, 2011). Geochronologic data show that the Hongqiling mafic-ultramafic intrusions formed in the Late Triassic (237 Ma to 208 Ma) (Table 1). Therefore, we believe that the Hongqiling mafic-ultramafic rocks formed in a post-collisional extensional environment.

The $\epsilon\text{Nd}(t)$ values of the Hongqiling mafic-ultramafic rocks fall in the range of 1.63 to 10.34 (Wu et al. 2004; Wei et al. 2013; Sun et al. 2013a), which indicate that the source was depleted mantle. The generation of high-Mg magmas requires high mantle potential temperatures (e.g. Nisbet et al. 1993; Zhou et al. 2004; Zhao and Zhou 2013); our calculation suggests that these temperatures likely exceeded 1308°C. In contrast, the continental lithosphere is commonly cold. Thus, these high magma temperatures suggest that the source of the Hongqiling magma is

likely asthenosphere rather than continental lithosphere. Considering the source characteristics and tectonic setting of the Hongqiling ore-bearing mafic-ultramafic intrusions, the evolutionary history of the study area can be concluded the following scenario. At about 250 Ma, the sustained subduction caused the final closure of the Paleo-Asian Ocean basin (Wu et al. 2007; Chen et al. 2009; Zhang et al. 2009; Wei et al. 2013). Simultaneously, the lithosphere was gradually thickened by the effect of the initial collision of the Jiamusi Block and North China Craton, which also provide favorable conditions for delamination (Wu et al. 2007; Chen et al. 2009; Zhang et al. 2009b). However, the lithosphere became unstable with the increase of potential energy, resulting in the thinning of the lithosphere and upwelling of the asthenosphere during the Middle to Late Triassic. The Hongqiling primary magma was related to an interaction between upwelling depleted asthenospheric melts and was contaminated by the crust during their ascending process. Finally, the Hongqiling ore-bearing mafic-ultramafic intrusions formed during the Late Triassic in a post-collision extensional environment.

7. Conclusions

(1) The Hongqiling ore-bearing mafic-ultramafic intrusions have experienced fractional crystallization and crustal contamination.

(2) We calculate that the parental magma of the Hongqiling ore-bearing mafic-ultramafic intrusions was a high-Mg, Nb-Ta-deplete tholeiitic magma.

(3) The Hongqiling ore-bearing mafic-ultramafic intrusions formed in a post-collisional

extensional environment in the Late Triassic.

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Figure captions

Figure 1. Simplified geological map of the Central Asian Orogenic Belt (modified after [Jahn 2004](#)) and the distribution of important ore-bearing mafic-ultramafic intrusions in the region. Sources of U-Pb zircon ages: Kalatongke ([Han et al. 2004](#)), Huangshanxi ([Han et al. 2004](#)), Huangshandong ([Han et al. 2004](#)), Erbutu ([Peng et al. 2013](#)) and Hongqiling ([Wu et al. 2004](#)).

Figure 2. (a) Tectonic sketch map of NE China (modified after [Wu et al. 2007](#); [Zhou et al. 2013](#)); (b) Regional geologic map for the Hongqiling Cu-Ni deposit (modified from [Li et al. 2010](#)); (c) Detailed distribution of the mafic-ultramafic intrusions in the Hongqiling area. Age sources: the age of No.1 intrusion is from [Wu et al. \(2004\)](#), the age of No.2 intrusion is from [Liu et al. \(2010\)](#), the age of No.7 intrusion is from [Lü et al. \(2011\)](#).

Figure 3. Geological planes for the Hongqiling No. 1 intrusion (a), No. 2 intrusion (b), No. 3

intrusion (c), No.7 intrusion (d) (modified from [JBGMR, 1988](#); [Qin 1995](#); [Wu et al. 2004](#)).

Figure 4. Representative microphotographs of the Hongqiling ore-bearing mafic-ultramafic intrusions.

(a) Olivine inclusion in orthopyroxene, rock type: harzburgite; (b) The poikilitic olivine grains, rock type: harzburgite; (c) and (d) Rock type: lherzolite; (e) Poikilitic texture, rock type: websterite; (f) Small spinel inclusions within olivine and orthopyroxene crystals, rock type: orthopyroxenite; (g) Orthopyroxene crystals enclosed in plagioclase, rock type: norite; (h) and (i) Sulfide inclusions within olivine crystals. Ol—olivine; Opx—orthopyroxene; Cpx—clinopyroxene; Spl—spinel; Pl—plagioclase; Hbl—hornblende; Serp—serpentine; Sulf—sulfide.

Figure 5. Compositional comparison of spinels from the Hongqiling ore-bearing mafic-ultramafic intrusions and elsewhere in the world. Data sources: Huangshandong are from [Mao et al. \(2015\)](#), Huangshanxi are from [Mao et al. \(2014\)](#), other types of rocks are from [Barnes and Roeder. \(2001\)](#).

Figure 6. (a) Plot of Ca versus Fo contents in olivine and (b) The Wo–En–Fs diagram of pyroxenes (after [Morimoto et al. 1989](#)) in the rocks from the Hongqiling ore-bearing mafic-ultramafic intrusions. The data for continental flood basalts (CFB), ocean island basalts (OIB), and komatiites are from [Sobolev et al. \(2007\)](#). Data for the Annette and Duke Island Complexes are from [Li et al. \(2012\)](#).

Figure 7. (a) MORB-normalized trace and major elements for spinel and (b) primitive mantle-normalized incompatible element patterns for clinopyroxene in the rocks from the Hongqiling ore-bearing mafic-ultramafic intrusions. The spinel from the MORB values are from [Pagé and Barnes \(2009\)](#); spinel from the Huangshanxi are from [Mao et al. \(2014\)](#), spinel in

komatiites values are from Yao (1999); primitive mantle values are from Sun and McDonough (1989).

Figure 8. Comparison of major oxide contents in whole-rocks from the Hongqiling ore-bearing mafic-ultramafic intrusions. Mineral abbreviations are the same as in Figure 3. Previous studies data are from Wu et al. (2004), Wei et al. (2013) and Li et al. (2018).

Figure 9. (a) Primitive mantle normalized incompatible element and (b) chondrite normalized REE patterns for the Hongqiling ore-bearing mafic-ultramafic intrusions. Primitive mantle values and chondrite are from Sun and McDonough (1989); Data for Huangshanxi intrusion are from Mao et al. (2014).

Figure 10. Diagrams of trace elements versus Zr of the Hongqiling ore-bearing mafic-ultramafic intrusions.

Figure 11. (a) Diagram of Th/Yb versus Nb/Yb; (b) Diagram of Ce/Nb versus Th/Nb; (c) Diagram of $(Th/Nb)_{PM}$ versus $(La/Sm)_{PM}$; (d) Diagram of Th/Yb versus Nb/Th. OIB, N-MORB, arc volcanic rocks and subduction-related mafic intrusive rocks are from Pearce. (2008) and Peltonen, (1995). UCC and LCC are from Rudnick and Gao (2004). The legends and data sources are same as Fig. 8.

Figure 12. (a) SiO_2 versus Al_2O_3 and (b) Al^{IV} versus Ti diagrams for clinopyroxenes from the Hongqiling ore-bearing mafic-ultramafic intrusions. Original plot form Leterrier et al. (1982).

Table 1. Geochronological data for the Hongqiling ore-bearing mafic-ultramafic intrusions.

Intrusion	Dated rocks	Dating methods	Ages / (Ma)	Data sources
No.1 intrusion	Gabbro	SHRIMP zircon U–Pb	216 ± 5	Wu et al., 2004
	Gabbro	LA-ICP-MS zircon U-Pb	220 ± 3	Feng et al., 2011
	Olivine pyroxenite	Biotite ⁴⁰ Ar- ³⁹ Ar isochron	225 ± 1	Xi et al., 2006
	Disseminated Ore	Re–Os isotopic dating	237 ± 16	Sun, 2013a
No.2 intrusion	Gabbro	SHRIMP zircon U–Pb	212 ± 2	Hao et al., 2013
	Disseminated ore	Re–Os isotopic dating	215 ± 24	Sun , 2013a
No.3 intrusion	Amphibole pyroxenite	Amphibole ⁴⁰ Ar- ³⁹ Ar isochron	228 ± 3	Liu et al., 2010
No.7 intrusion	Massive Ore	Re–Os isotopic dating	208 ± 21	Lü et al., 2011

Table 2. Average concentrations of trace elements in olivine, orthopyroxene and clinopyroxene from the Hongqiling ore-bearing mafic-ultramafic intrusions (ppm).

Sample	P-II-3	P-II-13	P-II-14	P-II-15	P-III-6	P-I-7	P-I-10	P-II-3	P-II-13	P-II-14	P-III-6
Rock type	Lherzolite	Lherzolite	Harzburgite	Orthopyroxene	Harzburgite	Harzburgite	Harzburgite	Lherzolite	Lherzolite	Harzburgite	Harzburgite
Mineral	spinel					olivine					
n	2	4	2	2	3	3	2	3	2	2	2
Sc	—	—	—	—	—	4.83	5.57	4.47	3.28	1.44	4.62
V	1319	561	704	1010	1076	7.33	4.39	4.89	12.45	10.82	1.90
Cr	114788	120758	86340	103475	102896	151	191	53.87	163	153	75.93
Co	284	544	311	375	418	140	139	166	151	193	200
Ni	974	1045	850	925	1454	1807	1325	1361	2472	1235	1899
Cu	2.14	144	555	180	3.85	2.53	1.07	b.d	1.30	3.20	b.d
Zn	2376	5206	2487	3804	2656	79.38	81.94	44.94	67.07	73.78	76.94
Ga	123	218	157	232	163	b.d	2.26	0.51	4.87	3.82	b.d
Rb	—	—	—	—	—	1.37	1.05	b.d	7.77	13.26	1.63
Sr	—	—	—	—	—	1.07	7.66	0.21	21.07	27.63	0.28
Ba	—	—	—	—	—	1.99	28.73	0.40	71.73	64.23	2.31
Y	—	—	—	—	—	b.d	0.44	b.d	0.75	0.97	0.37
Zr	—	—	—	—	—	3.11	5.65	b.d	12.19	24.25	b.d
Nb	—	—	—	—	—	b.d	b.d	0.16	0.55	0.34	b.d
Ta	—	—	—	—	—	b.d	b.d	b.d	b.d	b.d	b.d
La	—	—	—	—	—	0.10	0.67	b.d	1.58	1.45	b.d
Ce	—	—	—	—	—	b.d	1.22	b.d	2.91	3.02	b.d
Pr	—	—	—	—	—	b.d	b.d	b.d	0.32	b.d	b.d
Nd	—	—	—	—	—	b.d	0.12	0.29	1.03	0.72	b.d
Sm	—	—	—	—	—	b.d	b.d	b.d	b.d	b.d	b.d
Eu	—	—	—	—	—	b.d	b.d	b.d	b.d	b.d	b.d
Gd	—	—	—	—	—	b.d	0.40	0.54	0.19	0.38	b.d
Tb	—	—	—	—	—	b.d	b.d	b.d	b.d	b.d	b.d
Dy	—	—	—	—	—	0.12	b.d	b.d	b.d	b.d	b.d
Ho	—	—	—	—	—	b.d	b.d	b.d	b.d	b.d	b.d
Er	—	—	—	—	—	0.14	0.27	b.d	b.d	b.d	b.d
Tm	—	—	—	—	—	b.d	b.d	b.d	b.d	b.d	b.d
Yb	—	—	—	—	—	0.32	b.d	0.30	b.d	b.d	b.d
Lu	—	—	—	—	—	b.d	b.d	b.d	b.d	b.d	b.d
Hf	—	—	—	—	—	b.d	0.13	0.18	0.23	0.29	b.d
Pb	—	—	—	—	—	0.48	0.73	0.13	0.58	0.68	b.d
Th	—	—	—	—	—	b.d	0.17	b.d	b.d	b.d	b.d
U	—	—	—	—	—	b.d	b.d	b.d	b.d	b.d	b.d

n: number of analysis; -: unmeasured ; b.d: below detection limit.

Continued **Table 2**

Sample	P-I-10	P-II-13	P-II-14	P-III-8	P-III-3	P-VII-11	P-I-7	P-I-10	P-II-13	P-III-3	P-III-8	P-VII-2	P-VII-11
Rock	Harzburgit	Lherzolite	Harzburgit	Websterit	Websterit	Norite	Harzburgit	Harzburgit	Lherzolite	Websterit	Websterit	orthopyroxenite	Norite
type	e		e	e	e		e	e		e	e		
Mineral	orthopyroxene						clinopyroxene						
n	3	2	2	5	3	3	3	2	4	5	5	2	2
Sc	10.22	10.96	10.73	19.89	16.58	27.17	47.96	53.22	51.28	49.32	45.15	27.43	40.05
V	34.03	43.75	50.82	95.97	76.33	124	221	247	235	203	191	141	175
Cr	1568	952	1032	2175	1827	1454	6012	5267	6407	1594	1303	3181	2736
Co	95.99	42.91	42.91	81.27	72.00	60.98	33.03	25.50	39.69	31.02	35.91	23.37	25.20
Ni	777	211	209	135	203	138	296	107	249	49	74	233	44
Cu	6.84	b.d	b.d	1.60	3.50	1.53	4.02	1.63	1.62	0.87	0.99	2.54	1.27
Zn	65.46	58.19	64.23	146	131	67.30	17.47	19.18	31.60	36.71	44.01	17.99	25.78
Ga	5.32	7.70	7.02	3.59	4.32	6.68	6.04	8.22	5.86	4.50	6.30	5.07	6.59
Rb	1.72	10.11	9.67	1.42	1.53	6.00	0.96	0.95	0.66	1.91	5.29	4.21	7.97
Sr	4.26	24.47	24.35	4.26	5.71	14.82	29.03	37.58	16.07	55.64	34.53	41.86	32.78
Ba	14.89	93.25	86.75	20.33	25.84	56.52	1.19	20.81	2.61	42.65	41.61	30.56	78.74
Y	1.81	3.15	2.46	1.93	2.05	2.53	10.98	26.95	10.63	12.58	11.64	12.11	9.54
Zr	10.58	25.00	26.00	4.24	5.97	13.04	8.72	37.35	8.26	24.95	27.65	35.97	23.48
Nb	b.d	0.24	0.31	b.d	b.d	0.22	b.d	b.d	b.d	0.15	1.30	0.19	0.31
Ta	b.d	b.d	b.d	b.d	b.d	b.d	b.d	b.d	b.d	b.d	b.d	b.d	b.d
La	0.37	2.18	2.09	0.32	0.41	1.36	0.81	2.45	0.44	3.22	6.63	2.42	2.46
Ce	0.93	4.35	4.28	0.76	1.37	2.41	3.14	9.66	1.69	10.74	16.75	7.18	5.69
Pr	0.16	0.38	0.29	b.d	b.d	0.17	0.71	2.29	0.39	1.74	2.02	1.19	0.73
Nd	0.25	1.70	1.80	0.58	0.65	0.99	4.07	12.15	2.90	9.06	9.37	6.78	3.54
Sm	0.12	0.15	0.15	0.10	b.d	0.44	1.30	4.14	1.11	2.83	2.28	2.07	1.09
Eu	b.d	b.d	b.d	b.d	b.d	b.d	0.44	1.42	0.51	0.64	0.65	0.76	0.43
Gd	b.d	0.49	0.49	0.35	0.37	b.d	2.38	5.21	1.49	2.94	2.42	2.70	1.41
Tb	b.d	b.d	b.d	b.d	b.d	b.d	0.30	0.84	0.29	0.43	0.38	0.40	0.26
Dy	0.20	0.30	0.31	0.25	0.22	0.12	1.73	6.13	1.96	2.62	2.43	2.64	1.78
Ho	0.11	0.15	0.12	b.d	b.d	b.d	0.40	0.97	0.40	0.47	0.45	0.52	0.40
Er	0.24	0.26	0.26	0.23	0.26	0.32	1.42	3.04	1.21	1.19	1.19	1.38	1.01
Tm	b.d	b.d	b.d	b.d	b.d	b.d	0.14	0.33	0.19	0.18	0.17	0.16	0.15
Yb	0.35	0.67	0.71	0.37	0.39	0.23	1.36	2.93	1.36	1.12	1.05	1.07	0.98
Lu	b.d	b.d	b.d	b.d	b.d	b.d	0.15	0.30	0.12	0.17	0.16	0.18	0.13
Hf	0.25	0.74	0.54	0.17	0.27	0.37	0.48	1.95	0.35	0.82	1.04	1.22	0.85
Pb	0.93	2.42	1.83	0.64	1.03	1.66	0.09	0.82	0.25	3.21	1.51	2.27	2.35

Th	b.d	b.d	b.d	b.d	b.d	b.d	b.d	1.79	b.d	0.32	0.33	0.51	0.42
U	b.d	b.d	b.d	b.d	b.d	b.d	b.d	0.11	b.d	0.14	0.12	0.21	0.15

n: number of analysis; -: unmeasured ; b.d: below detection limit.

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Table 3. Major oxides contents of the rocks from the Hongqiling ore-bearing mafic-ultramafic intrusions (%).

Sample	P-I-10	P-II-13	P-II-14	P-III-7	P-III-8	P-VII-11	P-VII-15
Rock type	Harzburgite	Lherzolite	Harzburgite	Harzburgite	Websterite	Norite	Orthopyroxenite
	e		e	e			e
SiO ₂	40.03	43.43	42.05	42.17	53.01	50.56	47.17
TiO ₂	0.53	0.79	0.61	0.33	0.35	0.5	0.79
Al ₂ O ₃	6.11	7.77	7.51	6.22	4.50	13.3	9.33
Fe ₂ O _{3T}	10.97	11.54	12.13	10.66	10.4	6.67	11.06
MnO	0.16	0.19	0.19	0.16	0.21	0.15	0.14
MgO	31.08	25.27	27.44	29.66	22.06	12.10	20.71
CaO	3.62	5.44	4.58	4.53	7.15	12.30	3.72
Na ₂ O	0.35	0.92	0.84	0.38	0.58	2.07	0.38
K ₂ O	0.30	0.49	0.32	0.22	0.34	0.22	1.41
P ₂ O ₅	0.05	0.14	0.07	0.06	0.05	0.04	0.11
LOI	6.65	3.59	3.90	5.68	1.51	1.53	4.59
Total	99.84	99.57	99.64	100.15	100.06	99.49	99.39
FeO _T	10.59	10.82	11.40	10.16	9.49	6.12	10.49

Table 4. Trace element contents of the rocks from the Hongqiling ore-bearing mafic-ultramafic intrusions (ppm).

Sample	P-I-10	P-II-13	P-II-14	P-III-7	P-III-8	P-VII-11	P-VII-15	Estimated trapped liquid	
Rock- type	Harzburgite	Lherzolite	Harzburgite	Harzburgite	Websterite	Norite	Orthopyroxenite	P-I-10	P-II-14
Li	13.07	13.07	13.05	9.89	8.73	7.13	41.75		
Be	0.25	0.37	0.28	0.25	0.46	0.21	0.80		
Sc	12.77	19.95	17.07	21.20	43.81	32.52	9.73		
V	67.87	127	105	102	170	173	86.83		
Cr	3297	2856	3028	2781	1525	1398	1218		
Co	112	95.30	105	140	98.90	38.84	66.66		
Ga	5.55	9.29	8.23	7.57	7.17	10.44	10.58		
Rb	9.57	15.02	9.83	8.62	18.25	7.08	39.27		
Sr	98.55	107	111	118	77.26	221	94.38	619	592
Y	9.71	13.81	10.46	8.10	10.41	10.89	14.77	59.98	61.28
Zr	31.85	60.61	43.35	31.08	51.22	24.15	72.73	172	148
Nb	0.84	1.27	1.01	0.75	1.87	0.53	2.38	5.56	4.84
Cs	1.70	4.99	4.50	3.34	1.92	9.88	6.00	9.07	30.01
Ba	64.00	85.62	59.74	69.00	81.35	44.60	149	287	225
La	2.76	5.48	3.28	4.19	6.00	1.98	7.77	15.08	12.34
Ce	7.51	13.47	8.83	8.70	13.30	5.69	18.48	43.61	39.25
Pr	1.05	1.79	1.27	1.28	1.99	0.88	2.91	6.74	7.87
Nd	5.11	8.14	6.12	6.11	8.66	4.58	12.15	33.20	34.55
Sm	1.49	2.29	1.74	1.52	2.05	1.56	3.20	9.74	11.30
Eu	0.44	0.64	0.50	0.34	0.36	0.54	0.79	2.94	3.34
Gd	1.41	2.22	1.69	1.25	1.74	1.62	2.80	7.79	8.87
Tb	0.29	0.44	0.33	0.27	0.37	0.35	0.52	1.94	2.22
Dy	1.88	2.87	2.18	1.72	2.18	2.32	2.96	12.22	13.90
Ho	0.39	0.59	0.46	0.35	0.47	0.50	0.64	2.43	2.81
Er	1.00	1.53	1.16	0.97	1.29	1.26	1.62	5.19	7.24
Tm	0.15	0.23	0.17	0.14	0.18	0.18	0.24	1.01	1.15
Yb	0.95	1.53	1.15	0.89	1.12	1.23	1.55	5.76	6.22
Lu	0.15	0.23	0.18	0.13	0.19	0.18	0.24	0.97	1.22
Hf	0.93	1.60	1.20	0.93	1.25	0.76	1.97	5.25	5.88
Ta	0.07	0.09	0.07	0.08	0.17	0.04	0.18	0.45	0.50
Pb	2.83	2.31	2.18	3.34	3.30	1.82	9.22	14.42	8.40
Bi	0.07	0.05	0.04	0.17	0.14	0.04	1.56	0.49	0.27
Th	0.69	1.19	0.84	0.88	2.34	0.40	1.53	3.95	5.59
U	0.12	0.22	0.12	0.18	0.55	0.07	0.35	0.81	0.77

Draft

Figure 1

161x107mm (300 x 300 DPI)

Figure 2

171x128mm (300 x 300 DPI)

Figure 3

178x183mm (300 x 300 DPI)

Figure 4

180x136mm (300 x 300 DPI)

Figure 5

176x131mm (300 x 300 DPI)

Figure 6

179x80mm (300 x 300 DPI)

Figure 7

161x59mm (300 x 300 DPI)

Figure 8

159x132mm (300 x 300 DPI)

Figure 9

179x66mm (300 x 300 DPI)

Figure 10

174x236mm (300 x 300 DPI)

Figure 11

179x168mm (300 x 300 DPI)

Figure 12

209x79mm (300 x 300 DPI)