A re-appraisal of the Epoch NiS deposit, Filabusi greenstone belt, Zimbabwe: a case for hydrothermal Ni?

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ABSTRACT

The Epoch NiS deposit, discovered in 1970 and subsequently mined from 1972 to 1998, has been considered as a magmatic ore system. However, on the basis of field, petrographic and analytical data obtained during the initial exploration and evaluation stages, integrated with limited new analytical data (SEM and ore microscopy), we propose that the genesis of the Epoch NiS may have a hydrothermal origin. Key features that have prompted a re-appraisal of the genesis of the Epoch mineralisation, include: a) ore lenses are irregularly distributed along foliation fabrics of the host talc-carbonate rocks and as such no relationship to basal ultramafic/mafic units; b) the thickness of the ore lenses are highly variable from tens of centimetres to several metres; d) ore mineral association characterised by dominant millerite-pyrite-chalocpyrite; e) high Pd/Pt ratios. Although our hydrothermal model is somewhat speculative, we suggest that on present evidence there is little doubt that the Epoch NiS lacks the features of typical magmatic sulphide deposits. More geochemical and isotopic data is needed to test our model.

Keywords: Mineral Systems, Mineral Exploration, Zimbabwe, Hydrothermal Ni, NiS deposits, Filabusi Greenstone Belt

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1. Introduction

Zimbabwe has one of the highest metallic endowment Archaean cratons on Earth. It contains large abundances of Au, Ni, Cr and PGE deposits associated mainly with greenstone belts (e.g., Campbell and Pitfield, 1994). The Zimbabwe Craton is the result of the accretion of diverse tectono-stratigraphic terranes (Kusky and Kidd, 1992; Kusky 1998; Dirks and Jelsma, 2002; Begg, 2009).

These terrane boundaries are associated with long-living stable continental lithosphere and their internal suture contacts, represented by greenstone belts with significant ultramafic-mafic components. The latter are well known to be highly prospective for metal accumulations and in particular for NiS deposits, thereby becoming key targets for NiS exploration in recent years (e.g., Kerrich et al., 2005; Hoatson et al., 2006).

In addition, these terrane margins, where associated with lithospheric thickness changes and translithospheric faults coupled with active and voluminous magmatism, represent features that are largely present in the most important metallogenic provinces in the world, and responsible for world-class magmatic NiS deposits in Australia (Mt Keith), China (Jinchuan) and Russia (Pechenga; e.g., Begg et al., 2009).

Hydrothermal Ni deposits are generally rare and difficult to identify due to the effects of metamorphic, tectonic and weathering overprinting of the original mineralogy and/or textural features (González-Álvarez et al., 2010, and this issue). However, a holistic view on mineral systems and the increasing reporting of a hydrothermal component in many magmatic NiS deposits (e.g., Almeida et al., 2007) raises questions on the re-assessment of the main mechanisms of formation of previously described NiS deposits.

In 1970-71 the senior author was part of the mineral exploration team that led to the discovery of the Epoch NiS deposit and subsequent evaluation trenching and drilling program (Pirajno, 1971). During this time 45 drill holes with a total of about 12,500 m were drilled, leading to a preliminary definition of three main ore zones (Main, North and West). From 1972 onward underground mine development resulted in the production of about 18,776 tonnes of nickel metal between 1976 and 1986, with cobalt and copper as by-products (Baglow, 1998). The Epoch nickel mine closed down in 1998.
In this contribution we embark on a re-appraisal of the Epoch Ni mineral system for which a magmatic origin was initially assumed (Baglow, 1986). We have re-examined samples and field data from the original data set and on the basis of petrographic analyses, ore minerals-host rock association, we propose that the Ni mineralisation at Epoch is not of magmatic origin, but rather the result of hydrothermal activity associated with talc-carbonate alteration of ultramafic rocks.

2. Regional setting

The Zimbabwe Craton covers most of Zimbabwe and extends to the southwest into Botswana (see inset of Fig. 1; Wilson, 1990) with a sharp Moho at about 40 km (Gore et al., 2009). The accretion history of the Zimbabwe Craton encompassed the following stages: (1) the ~3.5 Ga Tokwe cratonic core (the oldest part of the Craton; Wilson et al., 1978), after which several stages of juvenile crustal growth resulted in several episodes of igneous intrusions and greenstone accretion (Dirks and Jelsma, 2002 and references therein); (2) subsequent episodes of crustal growth at 2.9-2.7 Ga added the Lower Greenstone belts of the Belingwean Supergroup (2.9 Ga) and lower Bulawayan Supergroup (2.8 Ga), and the Chingezi granites (2.9–2.8 Ga; Wilson et al., 1978; Dirks, 1998); (3) the next important episode of cratonic growth is represented by the Upper Greenstone belts within the Upper Bulawayan (2.7 Ga) and Shamvaian (2.65Ga) Supergroups associated with the Sesombi (2.7 Ga) and Wedza (2.65 Ga) magmatic events (e.g., Dirks, 1998); and (4) at a later stage, shortening of the crust caused strain and resulted in crustal melting and diapirism at ~2.6 which produced a large metamorphic signature in the Zimbabwe Craton throughout (Dirks and Jelsma, 1998). This event produced the Chilimanzi intrusions, which constitute the last major pre-Great Dyke igneous event prior to the cooling and stabilisation of the Craton (Wilson et al., 1978; Jelsma and Dirks, 2002).

The Upper Greenstones are widespread throughout the Zimbabwe Craton and contain the most important NiS deposits of Zimbabwe (e.g. Trojan, Shangani, Epoch; Kusky, 1998; Prendergast, 2003). Most of the greenstone lithologies in the Zimbabwe Craton were formed during the 2.75–2.58 Ga events (Dirks and Jelsma, 2002).

Following cratonization, the Great Dyke was emplaced at ~2.5 Ga (e.g., Wilson, 1990; Wingate, 2000). The Great Dyke is the most characteristic geological feature in Zimbabwe as a linear mafic-ultramafic north trending intrusive system that traverses the entire Zimbabwe
Craton from north to south (see inset of Fig. 1). It is ~550 km in length and 4-11 km in width, and comprises layered intrusions with an upper mafic succession of gabbro-norite and a lower ultramafic succession of pyroxenite-dunite-chromitite (Wingate, 2000). The Great Dyke contains major PGE and NiS deposits (Wilson et al., 1989; Wilson, 1992).

After the emplacement of the Great Dyke, several large geological units were formed or emplaced such as: the Limpopo Mobile and Magondi belts, and the Mashonaland, Karoo-Nuanetsi and the Umkondo Sills (e.g., Wilson, 1990; Wilson, 1992; Kusky, 1998; Wingate, 2000; Dirks and Jelsma, 2002; Prendergast, 2003).

2.1 The Filabusi Greenstone Belt

The Filabusi Greenstone Belt (FGB) is located in the central-south of Zimbabwe, West of the Great Dyke (inset of Fig. 1). The FGB features widespread ultramafic intrusions into the Lower sequence as Phurombuzi/Pangani serpentinite suites of Upper Greenstone age (Campbell et al., 1990). Metamorphism reached greenschist to locally amphibolite facies. Geodynamically the FGB is part of a discrete, structurally evolving basin with a common deep rooted volcanic source, based on the FGB’s fragmental distribution and variability (Campbell et al., 1990).

The FGB is part of the Bulawayan Group, comprising an older succession (Lower Greenstones) with ultramafic lavas, banded iron-formation and a succession of bimodal volcanic rocks. These rocks are unconformably overlain by thin komatiitic lava flows and massive to pillow tholeiitic basalts, pelitic sediments, banded iron-formation, limestone and conglomerate (Fig. 1).

The previous unconformable units are part of an Upper Greenstones or Filabusi Succession, which is where the host rocks of the Epoch NiS deposit occur (Baglow, 1998). The host rocks of the Epoch deposit form a protuberance that extends from the Phurombuzi Serpentinite (Fig. 1), which is part of the ca. 2.7 Ga Mashaba Ultramafic Suite (Baglow, 1986). The Epoch intrusion is emplaced into metabasaltic rocks of the Upper Greenstones and metagabbro and metabelites of the Lower Greenstones. The Epoch intrusion consists of talc, serpentinite, talc-carbonate and chlorite schist, described below.
3. Epoch NiS deposit

The Epoch Ni deposit was discovered in 1969-1970 during a stream sediment and soil sampling program and is manifested at the surface by a few gossan outcrops (Figs. 2 and 3). The Epoch NiS deposit forms a body that protrudes from a layered ultramafic complex in the Filabusi Greenstone Belt, in SW Zimbabwe (Baglow, 1986; 1998; Figs. 1 and 3). During mining operations, the Epoch ore bodies were delineated on the basis of 0.4% Ni cut off, which allowed the definition of SW-plunging three main ore bodies, namely a Main Zone, a North Ore Body and a West Lode (Baglow, 1998). The Main Zone strikes northwest and dips 50 to 70º to the southwest and near the surface was 240 m long and 10 to 50 m wide (Pirajno, 1971), narrowing at depth to 100 m long and 40 m wide (Baglow, 1998). The North Ore Body was situated in the footwall of the Main Zone, with an E-W trend and dips of 70-80º to the south and was 360 m and 5-20 m wide near the surface. The richest ore is in the West Lode, which occurs at the junction of the Main Zone and the North Ore Body, where ore grades reach 1.2% Ni. The total resource in Epoch was estimated in 5.61 million tonnes @ 0.67% nickel (Pirajno, 1971), which produced as byproducts ~100 tonnes of Co, ~800 tonnes of Cu and ~7 kg Au (Bartholomew, 1990).

The Epoch NiS deposit belongs to a series of NiS deposits in the Zimbabwe greenstone belts associated with the late Archaean Reliance komatiitic event: the Trojan and Shangani NiS deposits, two prospects at Hunters Road and Damba-Silwane, and several other occurrences (Prendergast, 2003). Sulphur in the Reliance nickel deposits was interpreted to have been sourced dominantly at upper crustal.

3.1 Host rocks

The Epoch Ni mineralisation is hosted predominantly in sheared talc-carbonate rocks that also contain dolomite, chlorite and magnesite (Fig. 2). Serpentinite, chlorite schist and remnant lenses of coarse-grained pyroxenite are locally present. Mineralogically, relic textures include chlorite pseudomorphs after pyroxene and olivine. Talc is more abundant than carbonate and generally occurs as flakes which make up the groundmass of the rock. Dolomite and magnesite occur in elongate form, mostly aligned along the sheared fabric of the rock, but also as porphyroblasts. Periclase and pyrope are locally abundant. Chlorite is present as porphyroblast, the above-mentioned pseudomorphs or as thin elongated plates parallel to the schistosity. Tremolite is locally present, whereas magnetite can be quite abundant and is usually found in the centre of the chlorite porphyroblast. Serpentinite forms
lenses and irregular zones and pseudomorphs after olivine and pyroxene and is represented by lamellae of antigorite and chrysotile fibres. The footwall rocks are massive to schistose diorite, thinly bedded argillite, banded chert, quartzite and quartz-sericite schist. The hangingwall comprises a succession of metasedimentary and metavolcanic rocks.

_talc-carbonate_ is the main host rock of the sulphide ore, grey-greenish in colour and with a variably trending foliation fabric (NW, E-W, NE, NNW). Carbonate forms porphyroblasts and pseudomorphs after olivine in a talcose groundmass. The carbonate minerals include elongate magnesite and dolomite grains aligned along the foliation fabric. Chlorite is present as lens-shaped porphyroblasts and as thin elongate plates also parallel with the foliation. Tremolite is locally present, whereas magnetite is quite abundant and occurs as clusters or disseminated grains and is also found in the middle of the chlorite porphyroblasts.

The composition of four samples of talc-carbonate is shown in Table 1.

_serpentinite_ is dark green and occurs as lenses and isolated zones up to 40 m long, within the talc-carbonate rocks, with which contacts are gradational. Serpentine minerals are mainly represented by fibro-lamellae of antigorite, locally with cross-fibres of chrysotile. Serpentine minerals locally pseudomorph olivine and pyroxene. Carbonate and talc replace the serpentine minerals. Magnetite occurs as dustings disseminated throughout and as small grains.

_pyroxenite_, only seen in drill core, forms lenses from 25 to 100 m long also within the talc-carbonate rocks. Pyroxenite is dark green to black in colour, coarse-grained with individual crystals up to 4 cm long. Pyroxenite is composed of augite (black in colour) and diopside (pale dull green). Serpentine minerals (antigorite and lizardite), carbonate and chlorite occur interstitially and along microfractures.

_chlorite schist_ occurs as lenses and dyke-like bodies within the talc-carbonate, possibly along shear zones and as nearly continuous bands along the contacts with the wall rocks. In the latter case the chlorite schist bands probably represent the alteration type known as “black wall” (Pirajno, 2012). The chloritic rocks show two varieties: a) monomineralic, pale-green to nearly colourless chlorite and b) chlorite associated with tremolite-actinolite acicular crystals. Locally, small bands of tourmaline are present and cut across the schistosity.
3.2 Relict textures

Outlines of primary crystals are commonly exhibited in the talc-carbonate and serpentinite rocks. In talc-carbonate, chlorite pseudomorphs with magnetite inclusions have faint to clear outlines of former pyroxene crystals, which look like “augen”, embedded in a talcose and fluidal matrix. In carbonated serpentinite, former olivine crystals recognisable by their shape are completely replaced by carbonate and rimmed by serpentine.

3.3 Mineralisation and ore minerals

The sulphide mineralisation occurs as disseminations, stringers and veins, forming a series of irregular lenses of variable thickness, aligned along the shear fabrics of the host rocks (Figs. 2 and 4). Locally, veins of massive sulphides (up to 80 vol.%) fill fractures in the host rocks. The sulphide ore lenses, are locally numerous and show variable size, from a few metres to several tens of metres and do not display a relation to a basal mafic-ultramafic unit. This coupled with the atypical sulphides association (see below), ore alignment along deformation fabrics, and replacement textures exhibited by ore minerals of mafic silicates, such as pyroxene, strongly suggest a hydrothermal genetic model.

The more common ore minerals identified by reflected light microscopy and SEM analysis include (Fig. 5): pyrite (FeS2, 60 vol%), millerite (NiS, 20 vol%), bravoite ((Fe,Ni,Co)S2, 12 vol%), chalcopyrite (CuFeS2, 5 vol%), pentlandite ((Fe,Ni)9S8, 3 vol%)1. Pyrite contains exsolution bodies of pyrrhotite. Trace amounts of other sulphide ore minerals include: pyrrhotite, vaesite, chalcocite, covellite, linnaeite, bornite, neodigenite and mackinaurite. Pyrrhotite is monoclinic with lamellae of hexagonal pyrrhotite. Covellite and chalcocite tend to replace chalcopyrite along grain boundaries, whereas mackinaurite occurs as thin veins in pentlandite grains. Linnaeite and neodigenite form discrete grains. The dominant oxide mineral is magnetite, which occurs in concentrations ranging from 1 to 48 vol%, with grain size ranging from 0.001 to 3 mm and is commonly associated with the serpentinisation of olivine. Magnetite is in places altered to martite and hematite. Droplets and specks of platinoids have also been identified by SEM and comprise sperrylite, cooperite, and micron-size inclusions in sulphides of chiluite (Bi5Te2Mo2O21) and native Pb. Electron-micro-probe analyses of five samples identified phases containing Pd, such as [S,Te (Fe, Pd)], [S, Sb (Fe, Cu, Pd)], Te, [Sb (Fe, Pd, Ru)], [S,Te,Sb (Fe, Ni, Ru, Pd)], [Te (Fe, Ni, Au, Pd)], but

1 Stochyometric formulae
reference to similar phases in the literature were not found. The Pd-bearing phases are invariably associated with millerite and chalcopyrite, whereas pentlandite does not appear associated with Pd-bearing phases.

4. A hydrothermal ore genesis model for the Epoch Ni mineral system

Baglow (1986, 1998), although advocating a magmatic origin for the Epoch Ni sulphides, did comment on the fact that the ore zones form lenses and veins, which do not occur as “basal massive sulphide accumulations”. Fig. 4 shows an example of the typical distribution of the ore lenses alignment along the dominant foliation fabric. The ore-hosting intrusion itself is a protrusion tectonically “extruded or injected” from the main ultramafic body of the Phurombuzi Serpentinite (Fig. 6), which in turn is part of a layered Mashaba Ultramafic Suite. The ore mineral association is dominated by millerite and chalcopyrite,

Despite geochemical proxies not being available at this stage for the Epoch deposit, the sequence of mineralogical, textural, deposit scale and structural features pose a rather intriguing question regarding the origin of the formation of the Epoch NiS deposit, or at least the final result of it. The Epoch Ni deposit could be the result of hydrothermal fluid-rock interaction, or a combination of magmatic Ni enrichment enhanced by hydrothermal concentration.

The main arguments for a magmatic origin of the Epoch NiS mineralisation were based on the model of a “cold intrusion” that was squeezed along the fault zone into the surroundings diorites and sedimentary units (Pirajno, 1971). This was based on: (1) The absence of a contact with a metamorphic aureole, (2) the deposits projects as an apophysis from the main ultramafic unit and is bounded by faults, (3) fragments of the country rock are scattered within the talc-carbonate zone near the ultramafic contacts.

Conversely, all the previous features could also be explained as the result of fluid flow through the volcano-sedimentary sequence and focussed along the shear zone that trapped an ultramafic unit. The abundance of water, could have led to the first phase of rock-fluid interaction from the pore fluids of the neighbouring shales and conglomerates. This, along with CO\textsubscript{2} release from the limestone could have been an excellent framework for the deeply altered ultramafic unit by widespread carbonatisation, converting it into talc-carbonate alteration. This combined with the availability of organic ligands from the shale beds, could
have enhanced the mobility of the Ni to redistributed and concentrated in lenses along the shear zone (Greenwood, 2012).

Millerite and pentlandite are nickel-rich sulphide minerals that often occur in weakly mineralised ultramafic units (e.g., Eckstrand, 1975; Donaldson, 1981). In a first serpentinisation alteration phase pyrite would be the result of the reducing conditions, whereas magnetite would be formed during carbonate alteration due to the released Fe and O2. The presence of lizardite as the main serpentine minerals suggests serpentinisation under low-temperatures (50-300°C; Evans, 2010).

The mineralisation is patchily distributed, and the main ore zones (four separate lenses; Fig. 2) are oriented along the foliation, and are constrained by two faults. As presented by Baglow (1980) the sulphide assemblage millerite-pyrite (with polydymite and pentlandite, and Cr-bearing magnetite; Fig. 5) is very rich in sulphur content; a feature that would be difficult to explain as the result of a magmatic process. However, an external source for the sulphur and the process of serpentinisation could explain this mineralogical association.

These serpentinising fluids would have flowed through the shear zone of the Epoch deposit and pervasively but differentially altered individual domains of the host rocks, concentrating as disseminating sulphides the Ni, and forming the magnetite (Fig. 6). Intense metasomatism is strongly supported by the presence of tourmaline bands and needles of black tourmaline up to 4 cm long parallel to the schistosity and within a chloritic matrix (Baglow, 1980). However, the youngest granitic intrusion described in the vicinity is 5 km away, and therefore unlikely to be related to the presence of tourmaline. All these features can be interpreted as supporting the Ni redistribution and accumulation by fluid-rock interaction.

Pirajno (1971) reported on samples analysed for PGE, which showed that Pd was the most abundant platinoid present, associated to chalcopyrite and millerite. This could be the result of Pt and Pd being more soluble than Ru, Rh and Ir under fluid transport (e.g., Mountain and Wood, 1988). The higher Pd/Pt ratio as the result of Pd being more mobile in a fluid system is a previously reported feature in many other identified hydrothermal Ni deposits (e.g., Mountain and Wood, 1988; Barnes and Liu, 2012; González-Álvarez et al., this issue).

The host rocks have undergone such a complex sequence of chemical alteration: (1) serpentinisation, (2) carbonatisation, and (3) talc-carbonatisation. This makes it very complex to reconstruct with accuracy the original fingerprint of the deposit. Therefore, the caveat to a
possible hydrothermal origin of the Epoch Ni deposit, is that could it be the result of a metasomatised magmatic NiS deposit? In other words, was Epoch the result of an upgrading on the Ni concentration due to serpentinisation processes associated with external sulphur addition? This question, as well as uncertainties in the rock-fluid mass transfers remain to be answered, and await more geochemical and isotopic (e. g. sulphur) analyses of the host rocks and the ore zones of the deposit.

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