Pyhäsalmi Volcanogenic Massive Sulfide Deposit, Central Finland

Jesse Hettula

UNIVERSITY OF OULU
GEOSCIENCES

October 2017
Table of Contents

1. SUMMARY ........................................................................................................................................... 1
2. INTRODUCTION .................................................................................................................................. 1
3. VOLCANOGENIC MASSIVE SULFIDES ................................................................................................. 2
   3.1 Definition ........................................................................................................................................ 2
   3.2 VMS model .................................................................................................................................... 2
   3.3 Alteration zones and their interpretation for mineral deposit exploration ................................. 4
   3.4 Metal zonation within mound-type massive sulphide ................................................................. 8
   3.5 Hydrothermal fluid plumes in seawater ......................................................................................... 10
   3.6 Classification of VMS deposits .................................................................................................... 11
4. REGIONAL GEOLOGY OF THE PYHÄSALMI MINE ......................................................................... 14
5. PYHÄSALMI MINE ............................................................................................................................ 17
6. PYHÄSALMI ORE DEPOSIT DEFORMATION PHASES AND STRATIGRAPHY ...................... 17
7. MODELS AND THEORIES ON HOW THE PYHÄSALMI DEPOSIT FORMED .............................. 20
8. DISCUSSION ..................................................................................................................................... 23
9. REFERENCES ....................................................................................................................................... 24
1. SUMMARY

Pyhäsalmi mine is located in central Finland, at the eastern side of the Pyhäjärvi lake. The Pyhäsalmi deposit is polymetallic Zn-Cu VMS ore body with total reserve, mined and yet to be mined, of 58.3 Mt @ Cu 0.9 %, Zn 2.4 %, S 37.8 %, Au 0.4 g/t and Ag 14 g/t. At the end of 2013, 51 Mt of ore has been mined. The mine will be in operation until August of 2019.

The Pyhäsalmi deposit is hosted in a felsic-dominated bimodal Proterozoic succession. Local hydrothermal alteration is composed of sericite-quartz alteration, and intensifies when it is in close proximity with the upper ore body. The deep ore body is thrusted into unaltered metamorphosed hangingwall volcanic rock, thus separated from the alteration zone.

The Pyhäsalmi district and deposit has been subjected to four different tectonic phases (D1 – D4) and intrusions accompanied by them. These tectonic processes have thrusted the deposit in upright position from the original position.

Basic theory of VMS formation processes can be used for modeling Pyhäsalmi deposit formation process, which in turn can benefit massive sulfide exploration.

2. INTRODUCTION

The purpose of this Bachelor’s thesis is to introduce basic concepts of volcanogenic massive sulfides (VMS) and Pyhäsalmi deposit, central Finland. This thesis explains what the VMS deposits are, how they form, how they may be classified and what methods are effective for VMS exploration. This subject was chosen as a preliminary study for later work on Pyhäsalmi deposits alteration zones.

The Pyhäsalmi deposit was discovered in 1958. Pyhäsalmi mine started to operate from year 1962 onwards. From the very beginning, study of the Pyhäsalmi mine area has taken place by intensive drilling and geochemical sampling. Mainly by these methods, there has been some success at explaining how Pyhäsalmi deposit formed.
3. VOLCANOGENIC MASSIVE SULFIDES

3.1 Definition

Volcanogenic massive sulfides (VMS) or volcanic-hosted massive sulfides (VHMS) are stratabound or stratiform massive bodies of sulfide ore and other accessory minerals. Massive sulfide deposits consist of >60% of sulfides. VMS deposits form at or just below the sea floor in any relatively deep submarine environments from seawater derived hydrothermal fluids that are driven by convective circulation above a deep heat source, which is most commonly a magma chamber (Hannington, 2014).

Most often submarine massive sulfide ore precipitates on top of volcanic footwall rocks, and more rarely into turbidites or other such deep – sea sedimentary rocks that are mixed with volcanic rocks. VMS deposits appear in several different marine volcanic settings, usually in zones of extension and active volcanism along mid-ocean ridges, but also at island arcs, back-arc basins and fore-arc troughs. VMS deposits are forming today on the seafloor around undersea volcanoes (Robb, 2005).

3.2 VMS model

VMS deposits form because of convective hydrothermal circulation within the footwall. Convective hydrothermal systems require both a heat source and high footwall permeability to allow effective fluid flow and depth for circulation. Heat enables convection; thus, a magma body is needed within a few kilometers of the seafloor. The temperature of the hydrothermal fluids can vary from 400 °C to just a few degrees higher than ambient seawater. In terms of ore forming processes, higher temperature is better since mineral and ligand solubility in aqueous solutions increases with temperature. High-temperature hydrothermal fluids (>300 °C) are generally saline (7 wt. % NaCl), acidic (pH 4 – 6), reducing, rich in H₂S and metals such as Fe, Mn, Pb, Zn and Cu. Low temperature hydrothermal fluids have salinities up to twice that of modern
seawater (3.2 wt. % NaCl) and are a mix of hydrothermal fluids and pore water. Seawater itself is cold (2 °C), alkaline, neutral (pH 7-8), oxidizing and rich in SO₄, but poor in metals (Robb, 2005).

In VMS hydrothermal systems, the hydrothermal fluids for the most part originate from seawater. Fluids seep down from the seafloor towards the magma chamber, close in to the chambers proximity and, when the fluid is hot enough, the fluid flow direction turns up towards the ocean floor. During this circulation, the fluids flow through the oceanic crust, scavenging metals while gradually increase the fluids metal ion content. The source of sulfur is essentially from the sulfate component of the seawater itself with reduction of sulfate to sulfide during fluid-rock interaction prior to venting. It is possible that some of the hydrothermal fluids originate from the magma, yet there are not any good general estimates on how much magmatic fluids can affect the overall ore forming process (Robb, 2005).

Fractures, faults or other anomalies which create permeability in the footwall rock act as a structural control to focus the fluid flow when high temperature fluids begin to rise to the seafloor. After effective fluid discharge control has formed, an efficient means of precipitating solids from the discharging fluids are needed. In general, precipitation happens by mixing of hot hydrothermal fluid with cold ambient seawater. This lowers the hydrothermal fluids temperature, makes it more oxidized and less saline. As a result, the hydrothermal fluid becomes less capable to solute metals. Because ore mineral precipitation happens rapidly and in a small area, it is possible that mineral deposit can form just above or at the fluid discharge site. When the solids have precipitated from the fluid and the fluid venting has stopped, a cover of volcanic or sedimentary material is needed to protect the ore deposit from oxidation and erosion (Large, 1992).

Hydrothermal circulation leaves behind large alteration zones around the hydrothermal venting system as a proof of activity. Hydrothermal alteration can be seen up to 20 km laterally around some deposit, and 2-4 km deep below the massive sulfide lens. It is most intense in and around the upflow zone through which hydrothermal fluids rise to surface, forming a characteristic assemblage of magnesian-chlorite-quartz-pyrite minerals. Sericite gradually replaces chlorite as the dominant alteration mineral with increasing distance. In the far-field alteration zone, alteration is much less intense, and is characterized by sericite, chlorite and epidote (Ridley, 2013). Usually, VMS related
regional alteration is hard to differentiate from diagenesis and/or regional metamorphism (Large, 1992).

When sulfides start to precipitate from the hydrothermal fluid, zoned structures of mounds, sheets and/or pipes will form. Pyrite is the most common sulphide mineral in the massive sulfide, followed by pyrrhotite, chalcopyrite, sphalerite, galena and, more rarely, sulphosalts and bornite. The most common non-sulphide minerals include magnetite, hematite and cassiterite. In VMS deposits, quartz, chlorite, barite, gypsum and carbonates are ganque minerals (Lydon, 1984).

The amount of the Cu, Zn, Pb and Fe in hydrothermal solution is strongly temperature-dependent. Rapid increases in concentration occur above 310°C for Zn and 320°C for Cu and Fe. The small difference (10°C) in this divide for Cu and Zn appears to be sufficient to cause the observed zoning from Cu to Zn sulfides outward from the high-temperature interior to the low-temperature exterior of a growing chimney or mound (Large, 1992).

Temperature differences effectively separate vigorous, metal sulfide-precipitating, high-temperature black smokers (300 – 400°C) from low-temperature anhydrite-precipitating, metal-poor, white or gray smokers (200 – 300°C). The color difference between white and black smokers is caused by the lack of metals in white smokers. White smokers can evolve into black smokers if the temperature rises above 300°C and vice versa (Robb, 2005).

Black smokers have the biggest contrast in fluid density when compared to seawater. Because of this contrast, black smokers exhibit high-velocity (1 – 2 m/s) and turbulent discharge and produces buoyant hydrothermal plumes that rise well above the seafloor (Shanks, 2012).

3.3 Alteration zones and their interpretation for mineral deposit exploration

Hydrothermal fluids in the VMS circulation systems breaks down feldspar and volcanic glass in the primary volcanic footwall host rocks. These minerals are replaced by sericite, chlorite, carbonate, pyrite, and quartz in varying propositions, depending on the
hydrothermal fluid chemistry. Equations 1 and 2 describe what happens during hydrothermal alteration:

\[ 3 \text{NaAlSi}_3\text{O}_8 + 2\text{K}^+ + 2\text{H}^+ = \text{KAl}_3\text{Si}_3\text{O}_{10}(\text{OH})_2 + 6\text{SiO}_2 + 3\text{Na}^+ \quad \text{Equation 1} \]

Equation 1 involves a loss of Na$_2$O from albite to the fluid, and gain of K$_2$O to the sericized albite. During sericite alteration, the hydrothermal fluids pH increases and silica is conserved by the deposition of quartz (Gifkins et al., 2005).

Equation 1 is typical reaction in the outer parts of the alteration system, whereas equation 2 is more important closer to stringer mineralization in footwall pipe zones, where chlorite rich assemblages becomes dominant over sericite-rich assemblages. These areas are referred to as the chlorite alteration zone.

\[ 2\text{KAl}_3\text{Si}_3\text{O}_{10}(\text{OH})_2 + 3\text{H}_4\text{SiO}_4 + 9\text{Fe}^{2+} + 6\text{Mg}^{2+} + 18\text{H}_2\text{O} = 3\text{Mg}_2\text{Fe}_3\text{Al}_2\text{Si}_3\text{O}_{10}(\text{OH})_8 + 2\text{K}^+ + 28\text{H}^+ \quad \text{Equation 2} \]

During equation 2, the fluid becomes more acidic. Precipitating chlorite gains FeO and MgO, and a relinquishes K$_2$O, provided that Al$_2$O$_3$ remains constant. Al$_2$O$_3$ content may, however, chance if carbonate alteration is present (Large et al., 2001a).

When carbonate alteration is present, it commonly develops in either chlorite- or sericite-bearing zones close to the ore. Large et al. (2001b) concluded that the proximal carbonate zones form by infill of porosity and selective replacement of various components of the volcanic rocks, rather than by direct precipitation on the sea floor.

Stockwork or stringer zone is a vein-type sulphide mineralization contained in a pipe of hydrothermally altered rock. The stringer zone contact with massive sulphide lens is gradual, unlike the sharp contact between the massive sulfide and footwall rock around the stringer zone (Lydon, 1984). Stringer zone mineralization can be economical, with high grades of Cu and Au (Large, 1992).

Chloritization of the central part of the hydrothermal alteration pipe, where the water/rock ratios are at maximum (Lydon 1988), is the most common form of alteration. High hydrothermal fluid temperatures (>250°C) and basic hydrothermal fluids (pH >5) are also beneficial factors for chloritization (Large et al., 2001b; Cooke et al., 2001).
Chlorite alteration zones are dominated by chlorite (50 wt% - 80 wt%) with subordinate quartz + pyrite + sericite ± carbonate. A slight Cu enrichment is common in chlorite alteration zone, when compared to sericite alteration zone (Gifkins et al., 2005).

The upper central part of the hydrothermal alteration pipe can also be dominated by silicification. These cores are composed of quartz + pyrite and quartz + pyrite + sericite ± chlorite assemblages (Gifkins et al., 2005). In siliceous core zones, hydrothermal alteration has been most intense, where hydrothermal fluid flow and temperature are at their highest, destroying all primary textures of the host rock. The core is commonly intersected by networks of pyrite + chalcopyrite stringer veins. When present, siliceous core is formed by supersaturation of Si in the hydrothermal fluid. It is surrounded by chlorite alteration.

Usually these chlorite alteration zones are surrounded by an outer zone of K₂O enrichment (sericite alteration zone) (Lydon, 1984). Sericite alteration forms in lower temperatures and slightly acidic fluids (Large et al., 2001b; Cooke et al., 2001), forming assemblages of sericite + chlorite + quartz + carbonate + pyrite. Minor disseminated sphalerite may be present in sericite alteration zone (Gifkins et al., 2005).

The alteration index (AI) is a tool for mapping VMS related alteration. Ishikawa et al. (1979) developed this index to quantify the intensity of sericite and chlorite alteration in the footwall volcanics proximal to Kuroko deposits in Japan. The AI is based on the amount of elements lost and gained during alteration:

\[
AI = \frac{100 \times (K_2O + MgO)}{(K_2O + MgO + Na_2O + CaO)}
\]  

Equation 3

Rocks with AI values in between 20 – 60 are referred to as a hydrothermally unaltered rock. AI values of 50 – 100 refers to a hydrothermally altered rock, where AI value of 100 represents complete replacement of feldspars and glass by sericite and/or chlorite.

The AI index has limitations: it is unable measure carbonate alteration, it cannot separate chlorite and sericite alteration from each other and it cannot detect the presence of pyrite. To overcome these limitations, Large et al. (2001a) introduced the chlorite-carbonate-pyrite index (CCPI):
\[
CCPI = \frac{100 (MgO + FeO)}{(MgO + FeO + Na_2O + K_2O)}
\]

Equation 4

In equation 4, the FeO represents the total iron (FeO + Fe_2O_3) content of the rock. The CCPI has been designed to measure the increase in MgO and FeO associated with the chlorite development, increasing pyrite grade towards the stringer zone and Mg-Fe carbonate alteration. This increase sets chlorite alteration apart from the sericite alteration, which is useful because chlorite alteration is commonly associated with central part of the hydrothermal fluid system. In addition, the CCPI is also affected positively by carbonate alteration and pyrite, magnetite or hematite mineralizations. These minerals also precipitate in the central part of the hydrothermal system, which makes this index a very useful tool for VMS exploration. The CCPI value ranges from 0 to 100. Large et al. (2001a) combined the Ishikawa AI with CCPI in a X-Y graph and named it the alteration box plot.

Unaltered rhyolite, dacite and andesite/basalt rock compositions are bound to a certain area of the alteration box plot. Unaltered rhyolitic volcanic rocks are bound to AI = 20 - 65 and CCPI = 15 - 45. If the rock composition is unaltered dacite, it would be bound to AI =20 - 65 and CCPI= 45 - 65. Unaltered andesite/basalt is bound to AI = 20 to 65 and CCPI = 65 - 85.

For example, if geochemical analysis of yet unidentified rock points to values of AI = 58 and CCPI = 76, the rock would be unaltered if it is indeed basalt. However, if the rock has chlorite + pyrite ± sericite alteration, then the rock has altered from a more felsic composition. The alteration box plot should be interpreted only after the footwall rock composition is known. Figure 1B demonstrates how the AI and CCPI behave around the Hellyer VMS deposit in Tasmania, Australia, where the altered rock protolith is andesite.
3.4 Metal zonation within mound-type massive sulphide

A particular metal zonation within the VMS ore body can form during mineralization. Mound-type massive sulfide ore bodies are typically zoned from footwall to hangingwall, respectively starting with basal massive pyrite mound, massive pyrite-chalcopyrite-gold zone, massive pyrite-chalcopyrite zone without gold, low-grade Pb-Zn-Ag zone, high-grade Pb-Zn-Ag-Au zone and the uppermost layer of barite and sediments, as is shown in Large (1992).

Large (1992) concluded, based on his studies on Australian VMS and references therein, which processes result in metal zonation within the mound-type massive sulfide. VMS mineralization starts with low-temperature hydrothermal fluid flux, which...
then gradually evolves into 300 - 350°C hydrothermal fluid. Eventually temperature begins to fall to the point when fluid flux stops completely. Large (1992) divided this process into four stages, based on differences in sulfide precipitation and solubility.

Temperature regime has a critical control on base and precious metal solubility and deposition. In the stage 1, when low-temperature hydrothermal fluid (150°-250°C) are discharged onto the seafloor, copper, lead and zinc are transported in ore-forming solutions as chloride or bisulfide complexes, depending on the solution chemistry (Large, 1977). These metals are released from the chloride and bisulfide complexes due to higher pH and lower temperature of the seawater, which decreases metal solubility. As a result, copper, iron, zinc and lead begin to precipitate as sulfides in significant amounts. At the same time, sulfates (barite and anhydrite) may form a layer or cap on top of the sulfide mound. The sulfates precipitate from the seawater on top of the sulfide mound as a result of increasing temperature.

During the stage 2, when fluid temperature rises to 250°C and towards 300°C, copper and gold solubilities have risen significantly, introducing chalcopyrite and native gold to the massive sulfide mound. At the same time, Pb-Zn-Ag solubility also rises so high that these metals will not immediately precipitate when temperatures begin to fall. The high-temperature fluid will solute the previously precipitated galena and sphalerite, and move them upwards in the sulfide mound, until they precipitate again due to falling temperatures. This results in a massive Cu ± Au sulfide base upon which lies massive low-grade Zn-Pb-Ag-Ba sulfide. The uppermost layer of the mineralization constitutes of sulfate.

Continued input of high-temperature fluids (300° to 350°C, stage 3) shapes the stringer zone to more clearly defined form. Chalcopyrite and pyrite precipitates in the stringer zone and in the basal part of the mound “pushing” the Zn-Pb zone upwards with the same principle that was explained before. The transition zone in between is low grade Zn-Pb-Ag massive sulfide, which probably represents an interval of lead and zinc leaching between the chalcopyrite front and the upper zone of lead-zinc deposition (Fig. 1A). If the hydrothermal fluid temperature rises above 350°C, all metal solubility will rise so high that the fluid will be grossly undersaturated and it will solute all but pyrite as it enters to the massive ore. Pyrite alone will continue to replace earlier precipitates on top of the stringer zone to form massive pyrite zone. All other zones will move
upwards, all the while enriching the metal grade of topmost Pb-Zn-Ag-Au zone even further. This process is referred to as zone refining.

At this temperature regime during stage 3, gold is at first carried by chloride (AuCl$_2^-$) complex to the stringer zone and massive pyrite-chalcopyrite zone. Gold solubility gradually decreases due to falling temperatures along the way, which forces gold to precipitate. Switchover from AuCl$_2^-$ to Au(HS)$_2^-$ transport happens at low grade Zn-Pb-Ag zone because of change in hydrothermal fluid chemistry. The bisulfide complexes will then leach gold from the middle zones of the porous massive sulfide. Gold is then carried from the middle zones to the uppermost Zn-Pb-Ag-Au-Ba zone by bisulfide complexes, where gold precipitates again due to mixing with seawater.

During stage 4 of VMS hydrothermal activity, the fluid temperature falls, and while doing so, the low temperature fluid will move through the massive sulfide. It precipitates late-stage sphalerite, galena, barite and carbonate veins which will somewhat overprint the earlier mineralizations in all zones.

In addition to the temperature control on the development of these zones, metal solubility is also affected by the hydrothermal fluids pH, salinity and $\delta$O$_2$. These values vary from deposit to deposit and are affected by footwall geochemistry and hydrothermal fluids origin. A complete zonation pattern is typically developed in mound-style VMS. Other VMS deposit types, such as sheets, pipes and stacks of bodies often exhibit distinctive characteristics.

### 3.5 Hydrothermal fluid plumes in seawater

Hydrothermal fluid behavior in seawater is mainly governed by the density of the hydrothermal fluid, and the density is affected by salinity and temperature variation of the hydrothermal fluid. Fluid salinity will increase the density, whereas higher temperatures will lower it. In a situation where high-temperature, low salinity hydrothermal fluid encounters seawater, the fluid will be less dense than the seawater. This difference causes the hydrothermal fluid to be buoyant. Thus, hydrothermal fluids will rise and spread as a column towards the ocean surface. It will not, however rise too
high because the “black smoke” will rapidly cool down and mix with seawater, causing the hydrothermal fluid density to increase. When hydrothermal fluids density is equal to the seawater, hydrothermal fluid forms a stratigraphic layer in the sea, where it will remain until it has assimilated into the seawater (Sato, 1972).

These hydrothermal fluids will appear as a black smoke – hence the name for black smokers. The fluid appears black because of sheer amount of precipitating tiny sulfide particles in the hydrothermal fluid. After the sulfides precipitate, they will sink down to the ocean floor and form a sheet-like massive sulphide proximal to the hydrothermal venting site and metalliferous sediments at more remote areas. Sato (1972) suggested this method as one of the two models for sheet-like VMS deposit formation.

For Kuroko VMS deposit where the ore body has a sheet-like structure but the hydrothermal fluids had higher density than seawater, Sato (1972) suggested that the hydrothermal fluid migrated downslope when encountering seawater. The fluid moved until it arrived at the lowermost point of local topography, where a brine pool accumulated. The brine pool remains until it is completely precipitated all its sulfides and forms a massive sulfide structure that is restricted by local topography. Uneven topography and density difference between hydrothermal fluid and seawater may explain why some deposits are not located at the top of central part of hydrothermal alteration zone. However, the brine pool model is inconsistent with observations from active seafloor hydrothermal system. The only modern analogies for the brine pool model are in the Red Sea (Shanks, 2012)

3.6 Classification of VMS deposits

VMS deposits have been classified by using criteria such as base metal content (Solomon, 1976; Franklin et al., 1981; Large, 1992), tectonic setting (Sawkins, 1976), host rock textures (Morton and Franklin, 1987), host rock lithology (Sangster and Scott, 1976) and host rock composition (Barrie and Hannington, 1997; Franklin et al., 2005; Galley et al., 2007). Since most of the metals in VMS deposits are derived from leaching of the footwall rock (Large, 1992), and the footwall composition is somewhat
controlled by tectonic setting, there tends to be some overlap and common agreement between these classification schemes.

Because all VMS deposits are polymetallic, Franklin et al. (1981) classified them into two kinds of base metal groups, the Zn-Pb-Cu group and Cu-Zn group. He observed unifying features within these groups, namely in the footwall compositions and alteration features.

The Zn-Pb-Cu group is hosted by felsic volcanic rocks and/or sedimentary rocks. This group shows alteration halos that are well defined and laterally extensive, usually with sericite-quartz core and Mg-rich chlorite halo. Cu-Zn group deposits are, on the other hand, hosted by mafic volcanic rocks, and subvolcanic intrusions are common. Alteration zones are well defined and vertically extensive, but they are the opposite to the Zn-Pb-Cu groups: The core is Mg-rich chlorite which is surrounded by sericite + quartz-rich halo.

Barrie and Hannington (1999) introduced a new classification scheme for VMS deposit. Their idea took Franklin et al. (1981) observations a step further, and divided VMS deposits into five groups. These groups differ from each other by their host rock compositions, with emphasis on the pre-alteration composition of volcanic host rocks. The names for these groups are mafic, bimodal-mafic, mafic-siliciclastic, bimodal-felsic and bimodal-siliciclastic. These groups host rock compositions change gradually from mafic dominated host rocks to felsic dominated host rocks and sediments. Barrie and Hannington (1999) concluded that by using this classification scheme, they also effectively grouped different VMS deposits by their time of deposition, volcanic setting, base metal content, tonnage and grade of the deposits. However, their scheme also had some inconsistencies when this scheme was used to classify VMS districts instead of individual VMS deposits (Franklin et al., 2005).

Franklin et al. (2005) modified Barrie and Hanningtons (1999) classification system by slightly changing the terminology and definition of different VMS settings to better reflect the relative amounts and compositions of volcanic and sedimentary strata. The updated group names and definitions are as follows:

1. Back-arc mafic type. The host rock composition is predominately mafic, which has formed in mature intra-oceanic back-arc setting. Stratigraphic succession contains ophiolite and ophiolite-like assemblages, dominated by pillowed and
massive tholeiitic basaltic flows. When compared to all other deposit types, back-arc mafic type is Cu rich and Zn poor, small in number of deposits and size. They are found almost exclusively in Phanerozoic rocks.

2. Bimodal-mafic type. The host rock succession is basalt dominated that can contain less than 25% felsic volcanic strata. However, deposits are often hosted by these felsic rocks. Bimodal mafic types tectonic settings are primitive arc or primitive volcanic arc. These deposits predominate in Late Archean and Early Proterozoic rocks, and they are the most common of VMS deposit types. They have the second highest Cu content of these five groups.

3. Pelitic-mafic type. The host rock is composed of sub-equal proportions of mafic volcanic or intrusive rocks and sedimentary rocks, while felsic volcanic rocks are minor or absent. Principally they are of Middle Proterozoic age and younger, and they can form in mature back-arc settings. This type of deposits are less numerous than the deposits of other types, but their average tonnage is second only to the felsic siliciclastic type deposits.

4. Bimodal-felsic. The host rock is dominated (35-70%) by felsic volcanic rocks and it is accompanied by basalts (20-50%) and terrigenous sedimentary rocks (~10%). Bimodal-felsic type VMS deposits form in continental margin arcs and related back-arc settings. They are most abundant in Phanerozoic rocks, but occurs also in Late Archean and Early Proterozoic rocks. These deposits are the second-most numerous, and on average contain the most Zn and Ag of the five deposit types. They also commonly contain barite.

5. Felsic-siliciclastic type. The host rock composition contains minor amounts of felsic and mafic volcanic rocks, of which felsic rocks are commonly more abundant, and they are dominated by siliciclastic felsic rocks. These deposits form in mature continental back-arc settings, of which the majority is dated to Phanerozoic age. Felsic-siliciclastic VMS deposits represent the greatest tonnage of the VMS types, and they have the largest average deposit size. They have on average the lowest Cu content and the highest Pb content of the five deposit types.

Galley et al. (2007) later added a sixth group:

6. Hybrid bimodal-felsic type. This group was added to represent a cross between VMS and shallow-water epithermal mineralization. They resemble bimodal-
felsic type deposits, but contain shallow-water epithermal features, such as aluminous alteration assemblages and precious metal enrichments.

4. REGIONAL GEOLOGY OF THE PYHÄSALMI MINE

The Pyhäsalmi VMS ore body is hosted by bimodal-felsic volcano-sedimentary rocks that are genetically related to a 1.93-1.92 Ga Paleoproterozoic island arc. The island arc formed on an oceanic plate, at the western side of the Archean basement complex. The oceanic plate later collided with the Archean basement (Korsman, 1988), forming a major thrust zone. The major thrust zone has reactivated several times during later orogenic movements. This zone is referred to as the Raahe-Ladoga zone (RLZ), and it hosts numerous VMS deposits, which can be seen in figure 2 (Mäki, 1986; Laine, 2015; Mäki et al., 2015). Mineralizations are typically found in the specific lithological associations (Laine et al., 2015).

The Pyhäsalmi deposit is the main deposit in its own stratigraphic group, which is called The Pyhäsalmi group. It has several satellite deposits, namely Mullikkoräme, Ruostesuo, Kangasjärvi, Hallaperä, Vuohtojoki, Kaskela, Konttikallio and Kokkopuro. Four of them have been mined out and the rest are uneconomic (Mäki et al., 2015). The Pyhäsalmi mine is still in operation, producing 1.38 Mt of ore in 2016 (First Quantum Minerals, 2017).

The other stratigraphic group that contains VMS deposits in RLZ is called the Vihanti group. It contains one major massive sulfide ore (the Vihanti deposit) and it is accompanied by three minor uneconomic satellite deposits. The Vihanti mine produced 28.1 Mt of sulfide ore (1.447 Mt Zn, 131 000 t Cu, 103 000 t Pb, 190 t Ag, 3038 kg Au) from 1954-1992 (Laine et al., 2015, Mäki et al., 2015). The host rock is mainly composed of thick layers of intermediate and felsic metavolcanic rocks and minor interstratified mafic metavolcanic rocks with calc-silicate and graphite tuffite interlayers.
These two stratigraphic units are separated by one major southwest-northeast and two major southeast-northwest trending faults, respectively the Oulujärvi shear zone (OjSZ), and the Revonneva (ReSZ) and Ruhaperä (RuSZ) shear zones.

Mäki et al. (2015) suggested separate lithostratigraphical classification schemes for Pyhäsalmi and Vihanti groups, indicating that there were at least two different volcanic environments in RLZ, despite their similar age of 1.93-1.92 Ga (Kousa et al., 2013). The Pyhäsalmi group represents a lower bimodal-felsic (mafic-felsic) metavolcanics sequence whereas the Vihanti groups bimodal-felsic (intermediate-felsic) metavolcanic host rock was deposited during continuous evolution of the arc.

Beyond the south-west border of the RLZ, the bedrock is composed of the Svecofennian Central Finland Granitoid Complex (1.89-1.87 Ga).
Figure 2. Regional geology of Raage-Ladoga zone (RLZ). The most important VMS mineralizations of the Vihanti and Pyhäsalmi area are marked with stars. The black dashed lines represent the main structural features of the RLZ (OjSZ = Oulujärvi shear zone, ReSZ = Revonneva shear zone, RuSZ = Ruhaperä shear zone) (based on Mäki et al., 2015; © Maanmittauslaitos, National Land Survey of Finland, 2013)
5. PYHÄSALMI MINE

The Pyhäsalmi deposit was discovered in August 1958 at the eastern side of the Pyhäjärvi lake by a local farmer, who was digging a well in his yard (Helovuori, 1979). After the first reserve estimations were done, the mine construction started in August 1959. Open pit mining commenced on March 1, 1962. Since 1976, all production has been from the underground mine (Mäki et al., 2015).

The underground mine reached the depth of 1000 m in 1996, and a new ore body was found underneath the 1000m mark in December of the same year. The decision to build a mine extension was made in 1998 and completed construction of a 1450 m deep automated hoisting shaft was done in 2001. Inmet Mining Corporation acquired the Pyhäsalmi mine in 2002, and in 2013 First Quantum Minerals took over the Inmet Mining Corporation and also the Pyhäsalmi mine (Mäki et al., 2015).

The total reserve, mined and yet to be mined, of Pyhäsalmi deposit are 58.3 Mt @ Cu 0.9 %, Zn 2.4 %, S 37.8 %, Au 0.4 g/t and Ag 14 g/t (Laine et al., 2015). At the end of 2013, the total production of the mine has been 51 Mt (Mäki et al., 2015) and the total ore production of 2016 has been 1.38 Mt (First Quantum Minerals, 2017). The estimated closing date of the mine is during August 2019 (personal comm. with Mikko Numminen, the acting mine geologist at Pyhäsalmi mine).

6. PYHÄSALMI ORE DEPOSIT DEFORMATION PHASES AND STRATIGRAPHY

The Pyhäsalmi deposit is hosted within deformed Proterozoic volcanics of the Svecovennian orogeny. The region has gone through a high degree of deformation, linked to four compressional phases and several intrusions of both mafic and felsic composition (Lickorish, 2012; Laine et al., 2015).

According to Laine et al. (2015), the structural evolution of the Vihanti-Pyhäsalmi area is divided into two major compressional stages, which are separated by extensional
phase. The first compressional phase is divided into three different compressional sequences (D1 – D3) and the second compressional phase has one compressional sequence (D4). The first compressional phase began approximately 1.91 Ga, by thrusting and flat lying folding of epicontinental sediment towards NE (D1). During the D1-D2 phase, Vihanti – Pyhäsalmi volcanic arc most probably collided against the Archean basement, causing major thrust fault zone between these units. This major thrust zone has reactivated several times during later orogenic movements and is known as the Raahe-Ladoga zone (RLZ). Continuous SW-NE compression (D3) resulted in tectonic thickening and intense folding, which refolded the earlier flat lying structures into upright position along the RLZ.

During 1.89-1.85 Ga, RLZ was subjected to voluminous intrusive activity along major faults, due to ceased first compressional phase and starting extensional phase. The progressive extensional phase caused uplift of granulite facies metamorphic blocks along the earlier shear zones. The highest peak for metamorphosis has been dated to 1952±2 Ma (U-Pb) at conditions of 600 °C/ 4-5 kb (Korja et al., 1994). After the extensional stage ended, the second compressional stage (D4) began (1.82-1.79 Ga). The earlier SW-NE compression was shifted to N-S. It overprinted all the earlier structures, causing the F3 trending folds to be refolded into large-scale F4 folds. The D4 stage created the crustal scale, sinistral, SW-NE trending Oulujärvi Shear Zone. The horsetail system of the shear zone caused intense shearing in the Pyhäsalmi area, resulting in remobilization of the massive sulphide ores and reactivation of the Archaean craton shear zones. This was accompanied by large amounts of late granites and pegmatite intrusions close to the major shear zones between 1.82 and 1.79 Ga. These intrusions are also present at close contact with the Pyhäsalmi deposit (Mäki et al., 2015).

The Pyhäsalmi deposit host rock structure is complex. In addition, the host rock protolith identification is hard and at times impossible (Helovuori, 1979) because of severe obliteration of primary textures. Whole-rock geochemical analysis is, therefore, the most reliable method to produce some divisions within the stratigraphy (Lickorish, 2013; Mäki et al., 2015).

The whole-rock geochemical data is used to identify different lithostratigraphical units by their immobile element ratios. High field strength elements (HFSE), such as Zr, Nb,
Y and rare-earth elements (REE), are used as an outline for systematic fractionation during magma cooling. Since HFSE elements are incompatible and immobile, they enrich in the magma body as it progressively fractionates to more felsic composition. If a compatible immobile component such as TiO$_2$ or Al$_2$O$_3$ is plotted against an incompatible monitor of fractionation such as Zr, the shape of the resulting curve depends on the fractionation behavior of the individual compatible component (Barrett & Maclean, 1994).

In a TiO$_2$-Zr plot the fractionation curve has a positive slope from basalt to basaltic-andesite compositions because of the Fe-Ti enrichment trend typical of tholeiitic rocks. From andesite to rhyolite tholeiitic compositions, the TiO$_2$-Zr fractionation curve has a negative (depletion) slope due to titanomagnetite crystallization, but Zr builds up in the magma. Thus, certain Al/Zr ratios, which are also called immobile element ratios, represent certain rock composition. This ratio does not change even when the rocks are subjected to hydrothermal alteration. Instead, the altered rock samples form an alteration line, which exhibits mass change (loss or gain) in the altered rock. Because the immobile element ratio does not change during alteration, hydrothermally altered rocks protolith composition is identifiable even if the primary textures are obliterated (Barrett & Maclean, 1994).

The Pyhäsalmi deposit is hosted in a volcanic sequence, which is primarily rhyolitic. Geochemical studies have identified several distinct rhyolitic units, which are referred to as Rhyolite A, Rhyolite B, and Rhyolite X. The alteration zone is predominately altered rhyolite, but it also includes altered mafic sills of the Lehto synvolcanic intrusion, which are also referred to as Mafic A-2 (Mäki et al., 2015). The alteration zone is composed of intensifying sericite alteration towards the upper ore body.

The hangingwall is typically composed of altered Rhyolite B, but chances above the mineralization into basic composition (Mafic A-1). The thin Rhyolite X unit is somewhat reliable indicator for the composition change, since it appears just before Mafic A-1. The Mafic A-1 is composed of coherent and brecciated tholeiitic mafic volcanic rocks. Figure 3 shows the local geology of the Pyhäsalmi deposit.
7. MODELS AND THEORIES ON HOW THE PYHÄSALMI DEPOSIT FORMED

The Pyhäsalmi deposit extends vertically from surface to 1420m at depth. It has been divided into two parts, upper ore body and “new” deep ore body. The upper ore body has a steeply dipping, elongated conical shape, which extends from surface to 1100m depth and is surrounded by tectonically foliated alteration zone, which is clearly more abundant in east- than in west-side of the deposit. The alteration zone extends equally in both south and north (Mäki et al., 2015).

The upper ore body shows no apparent metal zonation. It is inferred that this is due to intense deformation, which leaves very few chances for preservation of the fine-grained
sulphide assemblages (Imaña, 2003), and post-VMS tectonic stacking of different parts of the original massive sulphide lens (Mäki et al., 2015).

The deep ore body is comprised of a single piece of massive sulphide with a knob or a potato-shaped blob (350 x 200 x 300m) and extends between levels 1100 and 1420m. It is partially enclosed by weakly altered (high Na) volcanic sequence, which is composed of tectonically banded felsic and mafic schists. This unit is separated from the upper ore body by a shear zone that contains talc-sericite alteration (Imaña, 2003).

Imaña (2003) described and mapped each of the ore types in the deep ore body and concluded in his Master’s thesis that the deep ore body, unlike the upper ore body, shows a typical sulphide and gangue zonation, which consists of a massive inner part composed of massive pyrite followed by pyrite-chalcopyrite-sphalerite and pyrite-sphalerite. In terms of Fe, Cu and Zn, the metal zonation is clear. Similarly, the gangue mineralogy also shows gradual changes from calcite, calcite-dolomite adjacent to the massive pyrite core, and quartz-calcite-barite, barite-calcite and barite towards the upper part.

The uppermost pyrite-sphalerite zone does not appear at the top of deep ore body, but in the bottom and in all sides of the deep ore body. Instead, massive pyrite dominates at the top and in the center of the deep ore body, and pyrite-chalcopyrite-sphalerite zone is located in between these two units.

Imaña (2003) also concluded that the sulphide ores were formed by syngenetic sub-seafloor replacement and deposition near the palaeo-seafloor. Syngeneic subsea-floor replacement was interpreted to occur within the pyrite dominated zone based on a wide distribution of colloform and banded carbonate patches along with numerous inclusions of rapidly emplaced host facies (porphyritic rhyolites and mafic volcanics). These textures are interpreted as evidence for open space filling and replacement within an unconsolidated sulphide mound and its underlying volcanic strata at the sub-seafloor environment. Subseafloor deposition was interpreted to occur at the uppermost layer of the mineralization. This interpretation is based on the uppermost zinc-layers lack of altered relicts of its host facies, which suggest that its deposition took place above the mixing zone of carbonate-sulphate, probably near the paleo-seafloor.

Lickorich (2012) constructed a model of how the Pyhäsalmi deposit may have deformed from a vertically extensive stratabound structure into the current shape. He used a
simplified stratigraphy and drew these stratigraphic units into cross-sections which were based on the Pyhäsalmi deposit drill-core data, geochemistry data, and dip data derived from drill-core at the area of 61100N to 63800N. The spacing of the cross-sections is 200m. In these cross-section, Lickorich hypothesized several faults lines based on the observed lithostratigraphy.

Lickorich (2012) refers to the western margin of the Pyhäsalmi deposit as a hangingwall structure since the Zn-rich ore is located on the western margin of the deposit. And since the Zn-rich ore surrounds the deep ore body in all sides but the top, the deposits current structure may be a result of downward facing sheath-fold with the upper ore-body forming the upper limb of the fold.

The basic theory of VMS-sulfide formation dictates that the VMS deposit should be hosted in a zone of alteration. However, at Pyhäsalmi the deep ore body is enclosed by unaltered volcanic rock. The thrust/shear model for the tectonics of the Pyhäsalmi mine may provide explanation for this observation. The mineralization has been carried by a thrust into the overlaying hangingwall, and it also moved a small footwall section from underneath of the thrusted deep ore body into the deposit, introducing unaltered rock into the nose of the deposit (Lickorich, 2012).

Laitala (2015) concluded in his Master’s thesis that the Pyhäsalmi area has well defined pyrrhotite horizons. They are interpreted as metalliferous sediments that are commonly found both in modern and ancient submarine volcanic sequences that are associated with the VMS deposits. These pyrrhotite horizons formed in an anoxic and reducing environment, most likely on the seafloor of a Palaeoproterozoic ocean. The pyrrhotite horizons mark the tectonic hiatus, where VMS deposits could have had time to form, and may represent ore horizon equivalents in the stratigraphic succession.
8. DISCUSSION

Constructing a model for VMS deposit can be challenging since it is not uncommon for them to be subjected to deformation and metamorphism during orogeny. In this regard, the Pyhäsalmi deposit is not an exception.

However, by making correct observations of which lithostratigraphical units belong to the footwall or the hangingwall during VMS mineralization, it is easier to visualize how the mineralization has deformed from stratabound structure into its current form. However, identifying the stratigraphic successions may not always be possible due to intense deformation and metamorphism. In Pyhäsalmi case, lithostratigraphical unit identification has been possible due to whole rock geochemistry (immobile element ratios) and intensive drilling.

The Pyhäsalmi deposit behaves well in the bounds of VMS formation model:

- The footwall composition is bimodal-felsic (according to the classification of Franklin et al. 2005), and so the footwall alteration should be sericite-quartz dominated, which is then surrounded by chlorite alteration, which is true for the most part. However, the chlorite alteration zone appears to be absent.
- The upper ore body can be found by mapping alteration intensity, even though the deep ore body is not. As it happened, the deep ore body was found by accident during drilling (personal comm. with Mikko Numminen, the acting mine geologist at Pyhäsalmi mine).
- The VMS ore body has metal zonation at least in the deep ore body, where Zn and Cu rich zones are overlaying the massive pyrite.
- At a regional scale sulfide-enriched sediments may represent ore horizon equivalents in the stratigraphic succession.

By studying the Pyhäsalmi deposit, ore exploration methods in Pyhäsalmi district can be more effective by focusing on identifying those stratigraphic units that are associated with VMS formation. However, these methods would not be directly applicable in the Vihanti district, since the Vihanti districts VMS deposits formed in different volcanogenic environment.
9. REFERENCES


http://www.first-quantum.com/Our-Business/operating-mines/Pyhasalmi/Production-Stats/default.aspx


Large, R.R., Gemmell, J.B., Holger, P., (2001a). The Alteration Box Plot: A simple approach to understanding the relationship between alteration mineralogy and


