GS2017-6

Evaluation of diamond-drill core from the Tower Cu-Zn-Ag-Au deposit, sub-Phanerozoic Thompson nickel belt, central Manitoba (part of NTS 63G14)

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In Brief:

- Drillcore re-evaluation places the Tower Cu-Zn-Ag-Au deposit within the Ospwagan group of the Thompson nickel belt
- Stratiform mineralization (T2 zone) may have formed by sub-seafloor hydrothermal replacement
- Potential exists for additional deposits along strike in the Ospwagan group

Citation:

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Summary

The Tower Cu-Zn-Ag-Au deposit is a pelitic-mafic or Besshi-type volcanogenic massive-sulphide (VMS) system located in the sub-Phanerozoic Thompson nickel belt (TNB). Preliminary findings suggest that the deposit is hosted in Pipe formation rocks of the Ospwagan group. A simplified stratigraphy for the deposit consists of impure chert and siliceous rock of the Pipe formation P1 member, overlain by pelite and local sulphide-facies iron formation of the P2 member. This is overlain by laminated calcareous rock and a thick sequence of impure chert with intercalations of calcsilicate, iron formation, siliceous rock and minor carbonate, constituting the P3 member. The upper portion of the P3 member hosts a heterogeneous chlorite schist unit that has not been identified in the Pipe formation from other parts of the TNB and appears to result from hydrothermal alteration, likely during sulphide deposition.

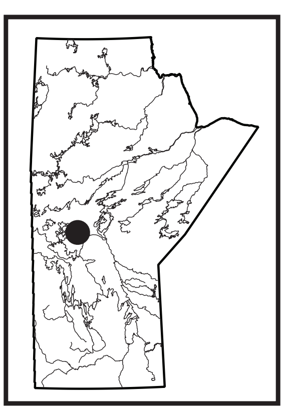
The T1 zone mineralization is discordant to stratigraphy and is hosted within the P2 member in the north and P3 member in the south. The T1 zone varies from a sulphidic schist to a sulphide breccia, the latter consisting of fragments of wall-rock, interpreted to result from mobilization along a late (D_3-D_4) structure. The T2 zone mineralization and associated chlorite schist may represent a stratiform zone of sub-seafloor hydrothermal replacement that is mostly in situ.

It is suggested that mafic to ultramafic magmatism, associated with either the Bah Lake assemblage or ca. 1883 Ma Molson-age intrusions, could have provided the heat source to drive the hydrothermal circulation system required to generate the Tower VMS deposit. Both of these magmatic events are of regional extent, suggesting there could be potential for VMS mineralization throughout the TNB. Volcanogenic massive-sulphide systems typically occur in clusters, suggesting that additional deposits could be found along strike from the Tower deposit.

Introduction

The Tower Cu-Zn-Ag-Au deposit is located along the eastern margin of the sub-Phanerozoic Thompson nickel belt (TNB; Figure GS2017-6-1). It was discovered by Falconbridge Ltd. (now Glencore plc) in September of 2000 during exploration for Ni-sulphide deposits in the William Lake area (Assessment Files 73953 and 73950, Manitoba Growth, Enterprise and Trade, Winnipeg). Follow-up drilling in the autumn of 2000 and winter of 2001 intersected significant Cu-Zn-Au mineralization in what has become known as the T1 zone of the Tower deposit (Assessment File 63G13256). The property was then acquired by Pure Nickel Inc. in August of 2007 (Beaudry, 2007). Rockcliff Resources Inc. (now Rockcliff Copper Corporation) entered an option and joint-venture agreement with Pure Nickel to explore the Tower deposit in February of 2008. Rockcliff Resources completed drill programs from 2010 to 2014 to further delineate the deposit, and additional mineralization was discovered in the T2 zone during the 2012 drill program (Assessment Files 63G13256, 63G14375). An indicated-resource estimate released in 2013 included 1 Mt at 3.7% Cu, 1.0% Zn, 17 g/t Ag and 0.5 g/t Au (Caracle Creek International Consulting Inc., 2013). In April of 2015, Rockcliff Resources agreed to sell its interest in the Tower property to Akuna Minerals Inc., and Pure Nickel sold its remaining interest to Akuna Minerals in June of the same year. Akuna Minerals is working toward developing the Tower deposit, and hopes to bring the deposit into production in the near future.

The significance of volcanogenic massive-sulphide (VMS) mineralization within a metallotect known for its magmatic Ni deposits is uncertain and brings into question the affinity of the hostrocks. Initial work by Falconbridge suggested that the mineralization is hosted by metasedimentary rocks of the Ospwagan group intruded by altered ultramafic rocks (Beaudry, 2007; Assessment Files 73953, 73950), implying that the Precambrian rocks in the area are part of the TNB; however, later work by Rockcliff Resources suggested that the metasedimentary rocks are subordinate to metavolcanic rocks and metamorphosed ultramafic intrusions (Assessment File 63G13256), possibly more in keeping with greenstone belts in the adjacent Superior province or Trans-Hudson



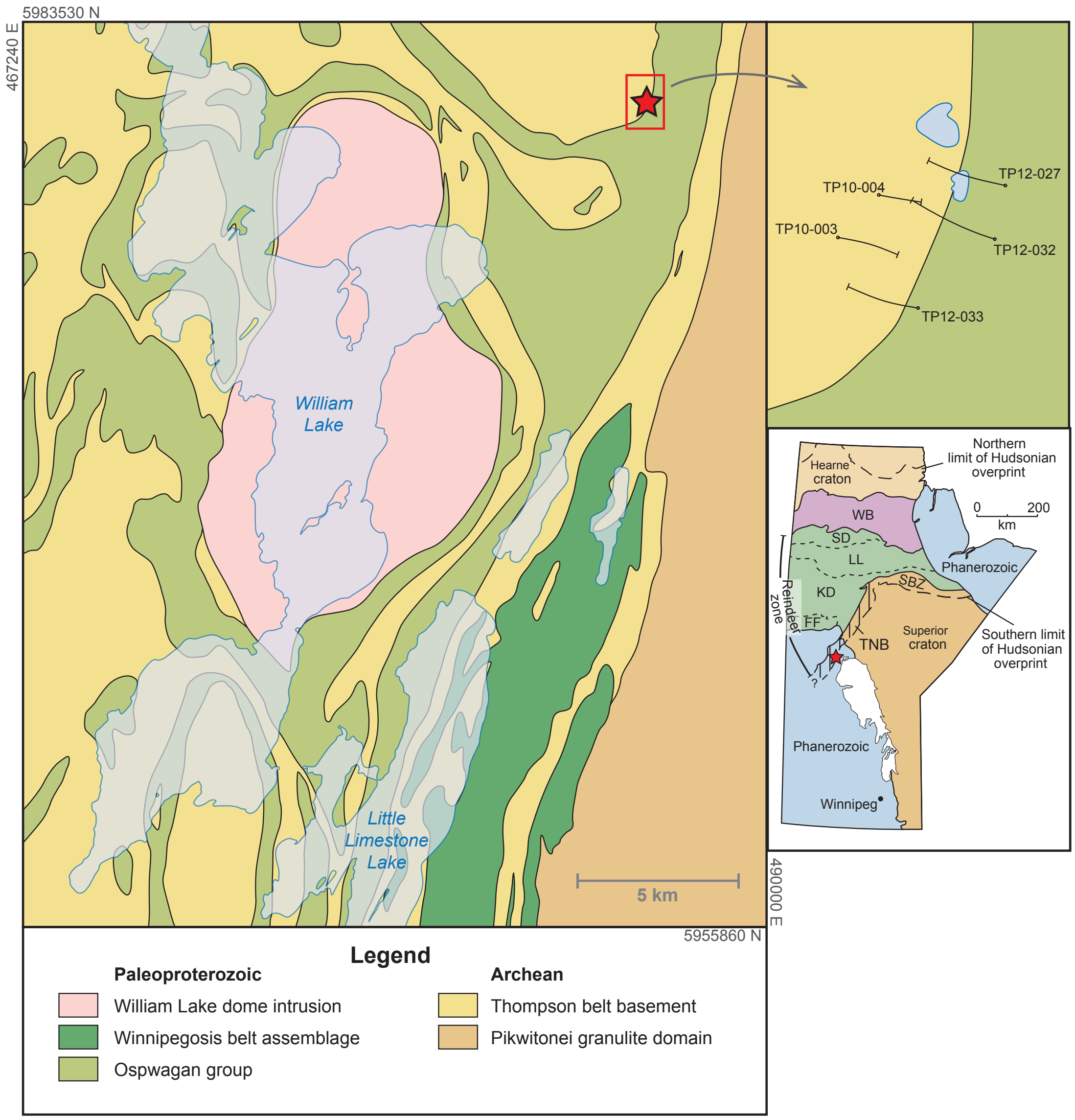


Figure GS2017-6-1: *Simplified lithology of the William Lake area (modified from Macek et al., 2006). The red star indicates the location of the Tower deposit. Abbreviations: FF, Flin Flon domain; KD, Kisseynew domain; LL, Lynn Lake domain; SBZ, Superior boundary zone; SD, Southern Indian domain; TNB, Thompson nickel belt; WB, Wathaman batholith.*

orogen (THO). Garnet, amphibole and biotite are described as common constituents in the ultramafic rocks: this association (garnet-bearing ultramafic rocks) is unknown in the Paleoproterozoic of the TNB but is described in the adjacent Pikwitonei granulite domain (PGD) of the Superior province (Böhm, 2005a, b; Couëslan, 2014). In contrast, VMS deposits have not been documented in the TNB or PGD but are widespread in the Flin Flon domain (FFD) of the THO: several deposits have been discovered to date in the eastern sub-Phanerozoic FFD, including the Talbot deposit, located roughly 32 km northwest of the Tower deposit (Simard et al., 2010). In this regard, it is possible that the rocks hosting the Tower deposit could represent a klippe of juvenile FFD rocks lying unconformably on TNB rocks.

The aim of this project is to evaluate the rocks hosting the Cu-Zn-Ag-Au mineralization at Akuna Minerals' Tower property, establish any genetic affiliation to the FFD, TNB or PGD, and identify possible VMS-related hydrothermal alteration. Identifying the provenance of the hostrocks could expand the mineral potential of the TNB or PGD, or expand the known area of the sub-Paleozoic FFD. Identifying VMS-related hydrothermal alteration could also provide exploration vectors to additional mineralization.

Regional geology

The TNB forms a segment of the Superior boundary zone, flanked to the northwest by the Kisseynew domain of the Trans-Hudson orogen and to the southeast by the PGD of the Superior craton. The TNB is underlain largely by reworked Archean gneiss of the Superior craton, which is typically quartzofeldspathic with enclaves of mafic to ultramafic rock. It is commonly migmatitic and characterized by complex internal structures that are the result of multiple generations of Archean and Paleoproterozoic deformation and metamorphism; clearly recognizable paragneiss is rare. The gneiss is interpreted to be derived from the adjacent PGD, which was subjected to amphibolite- to granulite-facies metamorphic conditions from ca. 2720 to 2640 Ma (Hubregtse, 1980; Mezger et al., 1990; Heaman et al., 2011; Guevara et al., 2016a, b). The granulites of the PGD were exhumed and unconformably overlain by the Paleoproterozoic supracrustal rocks of the Ospwagan group (TNB) prior to intrusion of the Molson dike swarm and associated ultramafic intrusions at ca. 1883 Ma (Bleeker, 1990; Zwanzig et al., 2007; Heaman et al., 2009; Scoates et al., 2017). The Archean basement gneiss and Ospwagan group were subjected to multiple generations of deformation and metamorphic conditions ranging from middle-amphibolite facies to lower-granulite facies during the Trans-Hudson orogeny (Bleeker, 1990; Burnham et al., 2009; Couëslan and Pattison, 2012).

The dominant phase of penetrative deformation is D_2, which affected the Ospwagan group and ca. 1883 Ma magmatic rocks. This deformation phase resulted in the formation of F_2 nappe structures, which incorporated the underlying Archean gneiss. The nappe structures have been interpreted as either east verging (Bleeker, 1990; White et al., 2002) or southwest verging (Zwanzig et al., 2007; Burnham et al., 2009). The recumbent folds are associated with regionally penetrative S_2 fabrics. The D_2 phase of deformation is interpreted to be the result of convergence between the Superior craton margin and the Reindeer zone of the Trans-Hudson orogen from ca. 1830 to 1800 Ma. The D_3 phase of deformation resulted in isoclinal folds with vertical to steeply southeast-dipping axial planes (Bleeker, 1990; Burnham et al., 2009). Mylonite zones with subvertical stretching lineations parallel many of the regional F_3 folds. Tightening of D_3 structures continued during D_4, marked by localized retrograde greenschist metamorphism along northeast-striking, mylonitic and cataclastic shear zones that commonly record southeast-side-up sinistral movement (Bleeker, 1990; Burnham et al., 2009). Although all rocks described in this report have been subjected to at least amphibolite-facies metamorphic conditions, the 'meta-' prefix has been omitted from rock names for brevity.

Review of the Ospwagan group stratigraphy

The following summary of the Ospwagan group is sourced largely from Bleeker (1990) and Zwanzig et al. (2007). The Paleoproterozoic Ospwagan group unconformably overlies Archean basement gneiss in the TNB. The lowermost unit of the Ospwagan group is the Manasan formation, which consists of two members: the lower M1 member, consisting of layered to laminated sandstone with local conglomerate layers near the base; and the overlying M2 member, consisting of semipelitic rock (Figure GS2017-6-2). The Manasan formation is interpreted as a transgressive, fining-upward sequence deposited along a passive margin. This siliciclastic system grades into the overlying calcareous sedimentary rocks of the Thompson formation.

The Thompson formation consists of three members: the T1 member comprises a variety of calcareous–siliceous rocks including chert, calcsilicate and impure marble; the T2 member is a semipelitic calcareous gneiss that is rarely present; and the T3 member consists of impure dolomitic marble with local horizons of calcsilicate (Figure GS2017-6-2). The Thompson formation represents a transition from a siliciclastic-dominated to a carbonate-dominated system.

The Pipe formation is subdivided into three members (Figure GS2017-6-2). The P1 member consists of a graphite-rich, sulphide-facies iron formation at the base (the locus of the Pipe II and Birchtree orebodies), overlain by a silicate-facies iron formation. The top of the P1 member consists of a reddish, laminated, siliceous rock. The P1 member grades into the overlying pelitic rocks of the P2 member, the top of which is marked by a sulphide-facies iron formation (the locus of the Thompson orebody). The overlying P3 member consists of a wide variety of rock types, including laminated, siliceous sedimentary rocks; silicate-, carbonate- and local oxide-facies iron formations; and semipelitic rocks, calcsilicate and a local horizon of relatively pure dolomitic marble. The Pipe formation represents a mix of chemical sediments and fine to very fine siliciclastics that were deposited in either an open-marine environment (Zwanzig et al., 2007) or during the development of a foredeep basin (Bleeker, 1990).

The Setting formation is divided into two members and is defined to include all siliciclastic rocks above the iron formation of the uppermost P3 member (Figure GS2017-6-2). The S1 member consists of rhythmically interbedded quartzite and pelitic schist with local calcareous concretions, which are very characteristic of the S1 member. The S2 member consists of thickly layered greywacke, with local horizons grading from conglomeratic at the base to pelitic at the top. No contact has been observed between the S1 and S2 members. It is possible that they represent a lateral facies change as opposed to a vertical succession. The S2 member appears to be missing altogether in the area of the Pipe mine, where contacts between the S1 member and the overlying Bah Lake assemblage are exposed. The Setting formation is interpreted to have been deposited by turbidity currents in a relatively deep-marine environment, possibly a foredeep basin (Bleeker, 1990). The coarse clastic material and thick turbidite bedding of the S2 member may record the onset of active tectonism, or a lateral sedimentary facies change, possibly to a submarine-channel or upper-fan environment.

At the top of the Ospwagan group is the Bah Lake assemblage, which consists of mafic to ultramafic volcanic rocks dominated by massive to pillowed basalt flows with local picrite and minor synvolcanic intrusions (Figure GS2017-6-2). The Bah Lake assemblage is dominated by a high-Mg suite (similar to normal mid-ocean ridge basalt; N-MORB) that occurs throughout much of the main TNB, and an incompatible-element–enriched suite (similar to enriched mid-ocean ridge basalt; E-MORB) that occurs in the northwestern Setting Lake area and along the margin of the Kisseynew domain (Zwanzig, 2005). The enriched suite is interpreted to overlie the high-Mg suite;

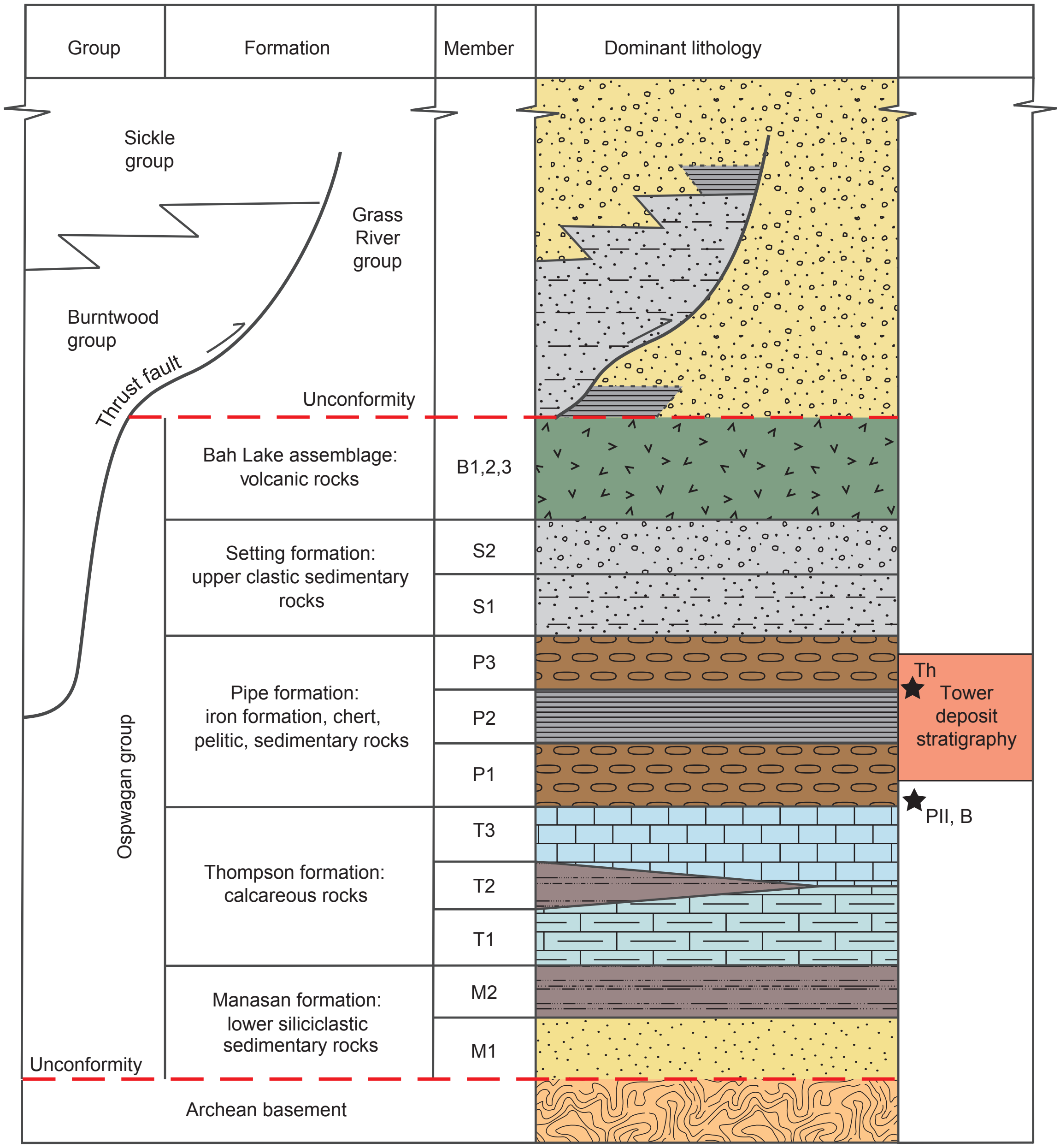


Figure GS2017-6-2: *Schematic tectonostratigraphic and lithostratigraphic section of the Ospwagan group of the Thompson nickel belt, and the Grass River and Burntwood groups of the adjoining Kisseynew domain (modified from Bleeker, 1990; Zwanzig et al., 2007). Black stars indicate the stratigraphic positions of the Birchtree (B), Pipe II (PII) and Thompson (Th) orebodies. A more detailed description of the Ospwagan group is found in the text.*

however, it is uncertain if this represents a stratigraphic or tectonic relationship. The Bah Lake assemblage may suggest the onset of active rifting in the TNB (Zwanzig, 2005; Zwanzig et al., 2007), or that the foredeep was magmatically active (Bleeker, 1990).

A minimum age for the Ospwagan group is provided by crosscutting amphibolitized dikes interpreted to be part of the Molson dike swarm, and the possibly comagmatic Ni-ore–bearing ultramafic sills, which intruded the Ospwagan group at all stratigraphic levels at ca. 1883 Ma (Bleeker, 1990; Zwanzig et al., 2007; Heaman et al., 2009; Scoates et al., 2017).

Geology of the Tower deposit

The Tower deposit occurs along the eastern margin of the sub-Phanerozoic TNB (Figure GS2017-6-1). The Precambrian rocks are overlain by 70–170 m of Paleozoic limestone and sandstone (Beaudry, 2007). Falconbridge geologists described the mineralization as being hosted in a thick package of Pipe formation pelitic and siliceous sedimentary rocks intruded by altered ultramafic rocks (Assessment Files 73953, 73950), consistent with the compilation map of Macek et al. (2006), which suggests that much of the Tower property is underlain by Pipe formation. Approaching the mineralization from the hangingwall, the sequence consists of magnetite-bearing silicate-facies iron formation, turbidite, and then a thick sequence of pelitic sedimentary rocks intruded by an altered ultramafic sill (Beaudry, 2007). The mineralization is hosted in the pelitic sedimentary rocks in proximity to the ultramafic body.

Conversely, work conducted by Rockcliff Resources determined the Tower property to be underlain by dominantly volcanic and ultramafic rocks with subordinate sedimentary rocks (Assessment File 63G13256). The volcanic rocks are dark green and fine to medium grained, contain possible pillow selvages, and have sharp contacts with adjacent units. The ultramafic rocks are described as peridotite, as well as dark green to black, fine- to coarse-grained rocks variably enriched in amphibole, garnet and biotite. In contrast to the Falconbridge work, the mineralization of the T1 zone was interpreted to occur within the ultramafic rocks, in proximity to the sedimentary rocks that consist of grey pelitic rocks containing fine-grained garnet and muscovite, interpreted to be part of the Pipe formation.

The mineralization of the T1 zone has a strike of approximately 013° with a dip of 75–85° to the east (Assessment File 73953). The mineralization transects the regional foliation and is characterized by rounded to subangular, millimetre- to centimetre-scale fragments of wallrock within a matrix of semisolid to solid sulphide (Assessment Files 73953, 63G13256 and 63G14375), interpreted to be mobilized from the source and forming the matrix of a fault breccia. Mineralization of the T2 zone is interpreted to be in situ, within pervasively altered volcanic rocks approximately 200 m northeast of the T1 zone (Rockcliff Resources Inc., 2013; Assessment File 63G13256).

The Cu- and Zn-rich nature of the mineralization is unusual for the TNB, as are the low concentrations of Ni and platinum-group elements (Beaudry, 2007). The mineralization is more akin to VMS deposits west of the TNB in the sub-Phanerozoic Flin Flon belt (Simard et al., 2010), and has been interpreted as a Besshi-type VMS system (Beaudry, 2007).

Methodology

Five drillholes were selected for re-examination. Drillholes TP10-003 and TP10-004 were both collared in the footwall and drilled toward the east, across strike through the T1 zone, and into the hangingwall (Figure GS2017-6-1). Drillholes TP12-027, TP12-032 and TP12-033 were collared in the hangingwall and drilled toward the west, across strike through the T1 zone, and into the footwall. Drillholes TP10-004 and TP12-032 were collared roughly across strike from each other, allowing for the construction of a continuous section from the footwall into the hangingwall of the deposit.

The core from each drillhole was laid out in its entirety, allowing for the entire sequence to be viewed and separated into petrographically distinct intervals. Each interval was then described petrographically and an interpretation of the protolith was made based on the mineral assemblages. Protolith interpretations were made within the context of the Ospwagan group stratigraphy; however, it is possible that the units represent varying types and intensities of hydrothermal alteration affecting volcanic rocks, which can be easily mistaken for sedimentary rocks (cf. Tinkham and Karlapalem, 2008). With this in mind, 27 samples were collected from key stratigraphic units for whole-rock lithogeochemical and Sm-Nd isotopic analyses. These analyses will be compared with the well-documented lithogeochemical and Sm-Nd isotopic compositions of the Ospwagan group (Zwanzig et al., 2007; Böhm et al., 2007) and published analyses from the sub-Phanerozoic Flin Flon belt (Simard et al., 2010). Thin sections were also made from each of the 27 samples to augment petrographic descriptions. The findings of this report should be considered preliminary and of an interpretive nature, pending the results of these analyses.

Stratigraphy of the Tower deposit

Iron formations are present throughout the footwall and hangingwall stratigraphy of the deposit (Figure GS2017-6-3). Because iron formations are only described from the Pipe formation of the Ospwagan group (Bleeker, 1990; Zwanzig et al., 2007), all sedimentary rocks intersected by the five drillholes are interpreted to be from the Pipe formation. The following stratigraphic description is from the footwall in the west to the hangingwall on the east side of the Tower deposit. The widths of stratigraphic units are reported as approximate true width, while thicknesses of potentially discordant intrusive units are reported as the length of the core intercept. Because of intense deformation, stratigraphic units commonly pinch and swell, and significant variations should be expected along strike. The stratigraphy presented below is idealized and based only on the five examined drillholes. All units were metamorphosed to amphibolite-facies conditions, and typically have moderate to strong foliations.

P1 member

The lowermost unit intersected in the footwall consists of grey, medium-grained and laminated to layered, impure chert (Figure GS2017-6-4a). The impure chert is garnet, biotite and amphibole bearing, and forms a package at least 22 m thick with intercalations, <20 cm thick, of garnet-bearing and biotite- and amphibole-rich silicate-facies iron formation. Upsection from the impure chert is a grey, fine- to coarse-grained, diffusely layered siliceous rock, 4.5–11 m thick (Figure GS2017-6-4b). The siliceous rock is garnet and biotite bearing, and could represent either quartz-rich siltstone or impure chert. The siliceous rock contains bands of, and is locally gradational into, biotite-bearing and amphibole-rich impure chert. Sparse sulphidic horizons, <25 cm thick, are present.

The siliceous rock has a mineral assemblage similar to the red laminated chert of the Pipe formation P1 member, exposed at the Pipe II mine near Thompson (Macek and Bleeker, 1989; Bleeker, 1990). The presence of this unit adjacent to pelitic schist (see P2 member) is in general agreement with this interpretation. Therefore, this package of rocks is tentatively interpreted as part of the Pipe formation P1 member.

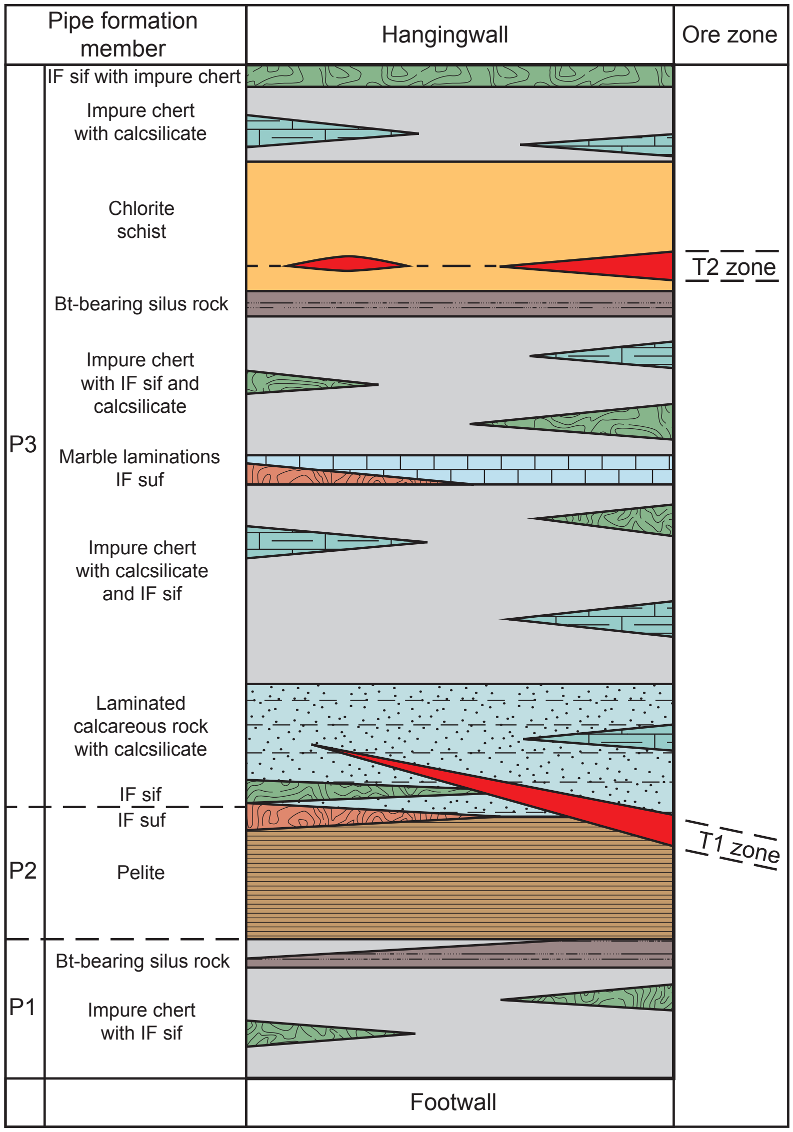


Figure GS2017-6-3: *Schematic lithostratigraphic section of the Pipe formation rocks hosting the Tower deposit. Abbreviations: Bt, biotite; IF sif, silicate-facies iron formation; IF suf, sulphide-facies iron formation; silus, siliceous.*

P2 member

Upsection from the siliceous rocks is a grey-brown, medium- to coarse-grained and diffusely layered pelitic schist (Figure GS2017-6-5a). The pelite can be 4–120 m thick, and is garnet bearing and enriched in muscovite and biotite. Staurolite occurs locally and the pelite locally becomes increasingly sulphidic in the uppermost part of the section. Spinel was identified in a thin section from the sulphidic portion of the pelite. Above the pelite is a 1.0–2.3 m thick sulphidic schist (Figure GS2017-6-5b) that is brown-grey, medium to coarse grained, gradationally banded and strongly magnetic in places. It is sulphide bearing and biotite and quartz rich, and is locally gradational into sulphidic chert.

The presence of iron formation on either side of the pelite (see P3 member) is a strong indicator that it is the Pipe formation P2 member. The sulphidic schist is likely correlative with a sulphide-facies iron formation of regional extent that marks the top of the P2 member (Bleeker, 1990; Zwanzig et al., 2007). This provides a good indicator that the sedimentary sequence is younging toward the east, from the footwall into the hangingwall of the deposit, and that the deposit remains structurally and stratigraphically upright.

P3 member

The sulphidic schist is locally overlain by silicate-facies iron formation (Figure GS2017-6-5b). The iron formation is grey to dark green, medium grained and strongly magnetic. It can be up to 3 m thick and is sulphide, grunerite, magnetite and biotite bearing, hornblende rich and siliceous. Magnetite forms discrete, equant porphyroblasts. The iron formation varies from relatively homogeneous to banded at a scale of <2 cm.

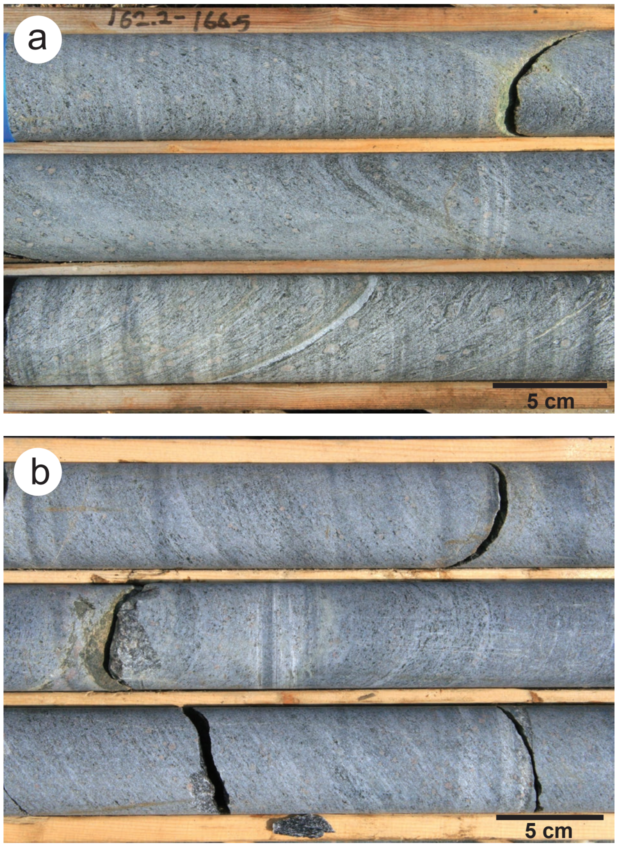


Figure GS2017-6-4: *P1 member rocks at the Tower deposit:* ***a)*** *garnet- and amphibole-bearing impure chert with incipient silicate-facies iron formation (bottom row; TP10-004, 162.2 m);* ***b)*** *garnet- and biotite-bearing siliceous rock (TP12-032, 578.6 m).*

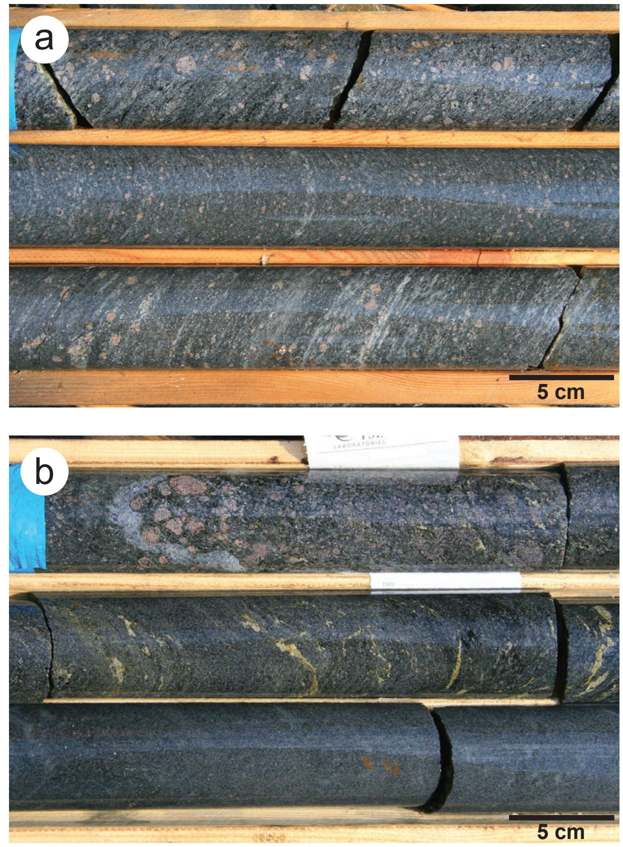


Figure GS2017-6-5: *P2 member rocks at the Tower deposit:* ***a)*** *garnet-bearing, muscovite- and biotite-rich pelitic schist (TP12-032, 573.6 m);* ***b)*** *sulphide- and garnet-bearing pelite (top row), sulphidic schist (middle row), and P3 member magnetite-bearing, silicate-facies iron formation (bottom row; TP10-004, 206.15 m).*

More typically, the sulphidic schist and pelite are overlain by an 8.9–38 m thick sequence of laminated calcareous rock (Figure GS2017-6-6a). The rock is grey-green, medium grained and weakly magnetic in places. It is typically laminated, but local zones can have diffuse bands <20 cm thick. The rock is siliceous, biotite bearing and hornblende rich. Local layers can be garnet bearing or clinozoisite bearing, and minor sulphide can be present. Intercalations of calcsilicate occur as clinozoisite-rich bands <10 cm thick. The calcsilicate varies from layered to mottled.

The calcareous rock is overlain by a thick (54–100 m) sequence of impure chert with intercalations of calcsilicate, silicate-facies iron formation and biotite-bearing siliceous rock (Figure GS2017-6-6b). The impure chert is typically grey to green-grey, medium to coarse grained, nonmagnetic and laminated to crudely layered on a scale <3 m. The chert is typically garnet bearing and siliceous with varying proportions of hornblende and biotite. The hornblende and biotite combined generally make up <30% of the rock. Local layers may be clinozoisite bearing. The chert locally grades into bands of silicate-facies iron formation <2.5 m thick. The iron formation is dark green-grey and contains a greater amount of garnet and hornblende (up to 40%) than the impure chert. Laminations of stringy chert are locally present within the iron formation (Figure GS2017-6-6b). The calcsilicate occurs as light green-grey, mottled to layered bands <20 cm thick. The calcsilicate is typically rich in clinozoisite±tremolite, with minor carbonate. The siliceous rock occurs as bands <40 cm thick that are garnet bearing and biotite rich with little to no hornblende. Sparse laminations of carbonate can be present within this sequence and become more abundant toward the upper contact.

The impure chert is locally overlain by a thin (25–85 cm) sulphide-facies iron formation (Figure GS2017-6-6c). The iron formation is brown-grey, fine to medium grained and weakly to moderately magnetic. The composition varies from a sulphide- and biotite-rich siliceous rock to a biotite- and hornblende-bearing, grunerite- and sulphide-rich rock.

Laminations and thin layers (<1 cm) of carbonate appear to form a local marker horizon at the Tower property (Figure GS2017-6-6d). These layers commonly occur in clusters within impure chert, just above the sulphide-facies iron formation when present. The carbonate layers are commonly interlayered

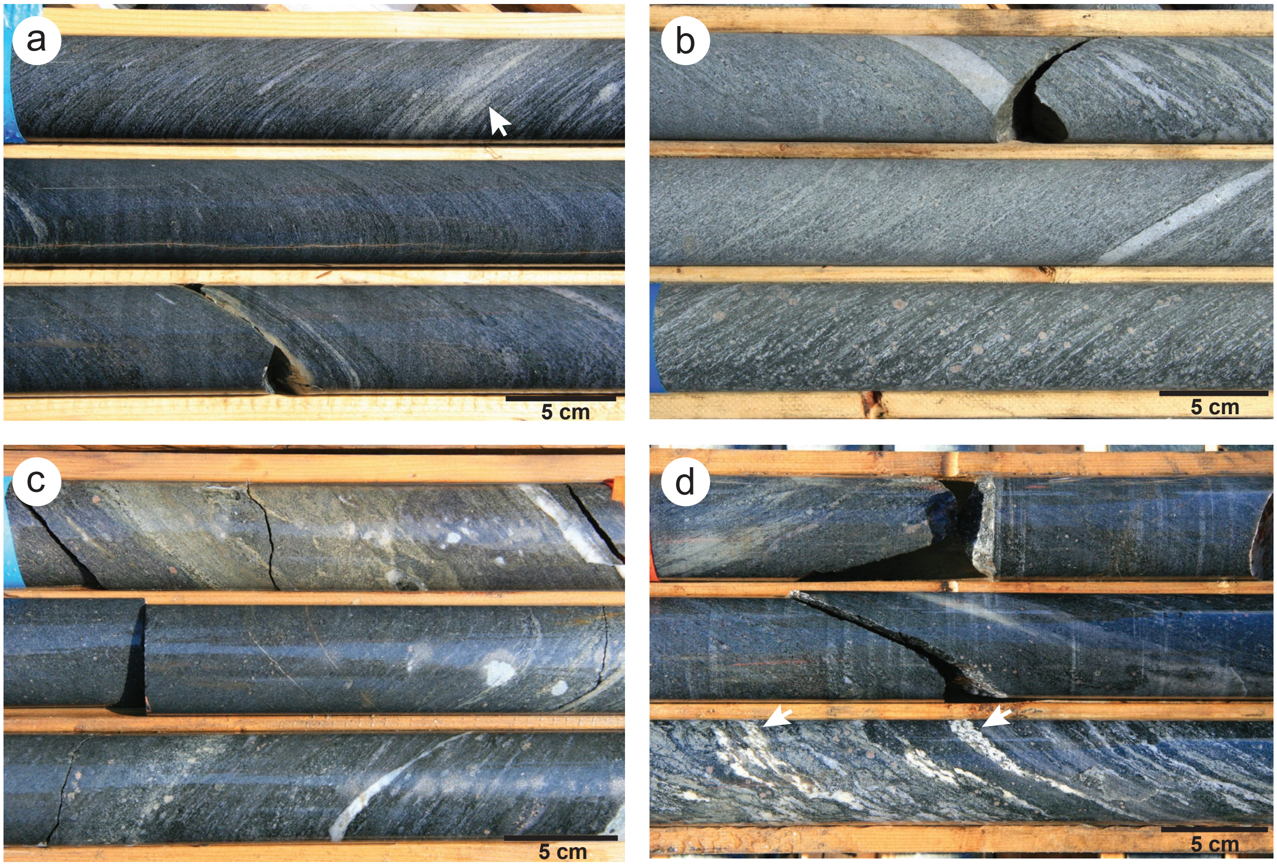


Figure GS2017-6-6: *P3 member rocks at the Tower deposit:* ***a)*** *amphibole-rich, laminated calcareous rock with diffuse clinozoisite-rich calcsilicate band (arrow; TP10-004, 227.1 m),* ***b)*** *garnet-bearing and amphibole-rich impure chert (top two rows) and iron formation with stringy chert laminations (bottom row; TP10-004, 274.5 m);* ***c)*** *sulphide-facies iron formation (top row) and impure chert (bottom two rows; TP12-032, 425.5 m),* ***d)*** *impure chert with diffuse bands of calcsilicate and thin carbonate layers (arrows; TP12-033, 350.5 m).*

with clinozoisite-rich and locally garnet-bearing calcsilicate. The impure chert sequence continues above the carbonate layers for an additional 64–150 m, and is similar to the sequence described below the sulphide-facies iron formation.

The impure chert is overlain by a 7–34 m thick sequence of garnet- and hornblende-bearing, biotite-rich siliceous rock (Figure GS2017-6-7a). The siliceous rock is brown-grey, medium grained and moderately magnetic in places. Chlorite and magnetite are present in minor amounts. The siliceous rock is interbedded with layers of impure chert, <2 m thick, which are generally enriched in clinozoisite. Local layers of silicate-facies iron formation, <1 m thick, can be present. The iron formation is strongly magnetic, garnet and magnetite bearing, and rich in grunerite. The magnetite occurs as discrete equant grains (Figure GS2017-6-7a). The siliceous rock grades over 3–5 m into the overlying chlorite schist.

The chlorite schist is the uppermost unit in two of the three drillholes collared in the hangingwall. Drillhole TP12-032 intersected the entire interval of schist, which is approximately 42 m thick. It is dark green to purplish green to brown, coarse grained, crudely banded on a scale of <50 cm, variably magnetic and compositionally heterogeneous (Figure GS2017-6-7b). The schist is typically quartz and chlorite rich with variable amounts of garnet, staurolite, magnetite and biotite, and minor sulphide±carbonate, hornblende and muscovite. Garnet porphyroblasts up to 10 cm across locally form up to 70% of the rock. Quartz occurs as discrete laminations, layers and pods <10 cm thick. A sulphidic horizon up to 4 m thick occurs near the base of the unit (Figure GS2017-6-7c). The contact between the chlorite schist and overlying impure chert is gradational over approximately 1 m (Figure GS2017-6-7d).

The units above the chlorite schist are described from drillhole TP12-032, and their continuity along strike is not known. The chlorite schist is overlain by a 14 m thick sequence of impure chert with intercalations of calcsilicate, <40 cm thick, that are most abundant toward the base of the unit. The impure chert is green-grey, fine to medium grained and magnetic in places. It varies from laminated to diffusely banded at a scale of <5 cm. The rock is siliceous, typically garnet bearing and hornblende rich with varying amounts of biotite and clinozoisite. The calcsilicate is light greenish yellow, fine grained and locally magnetic.

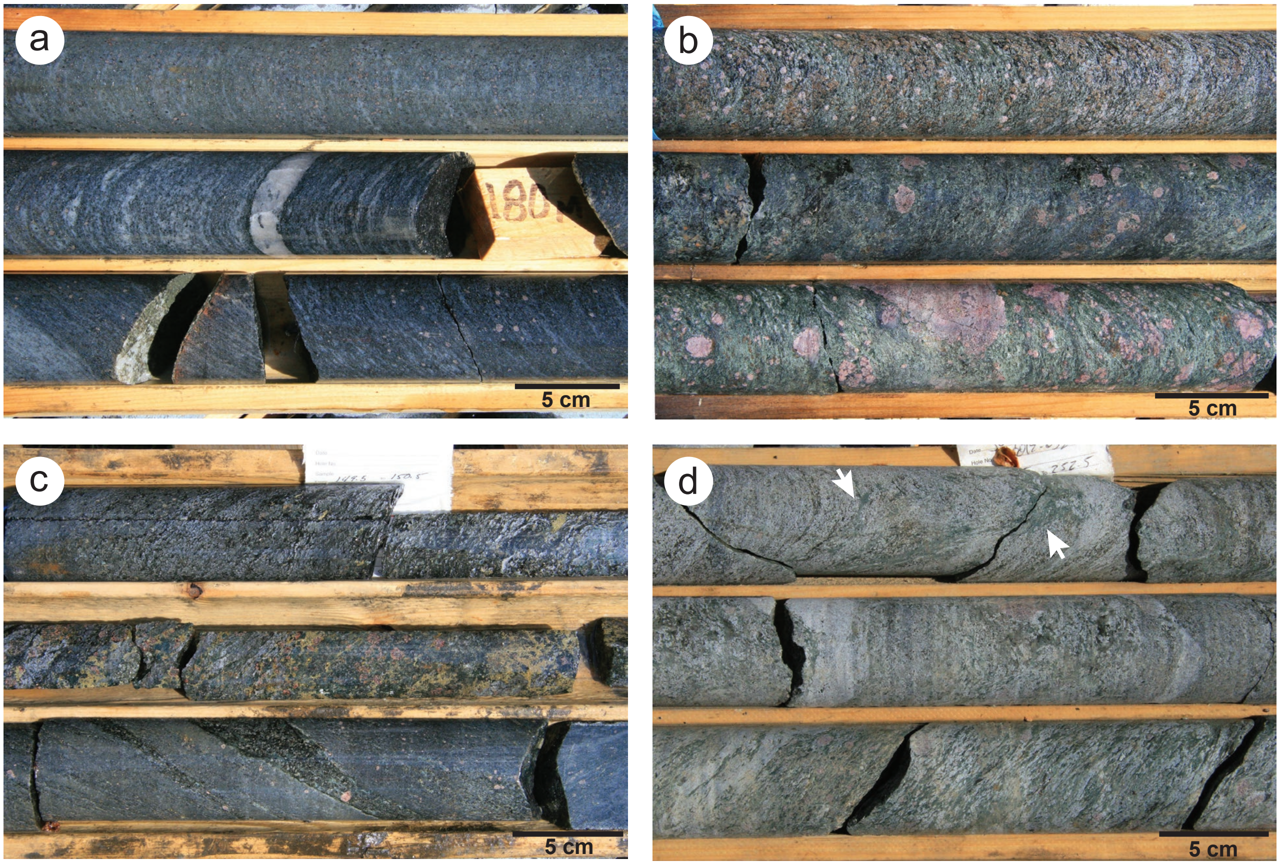


*Figure **GS2017-6-7:** P3 member rocks at the Tower deposit: **a)** magnetite-bearing, silicate-facies iron formation (top row) and garnet- and biotite-bearing siliceous rock (bottom two rows; TP12-027, 178.4 m); **b)** garnet-bearing and staurolite-rich chlorite schist (top row) grading into garnet-rich chlorite schist (bottom row; TP12-032, 285.85 m); **c)** chlorite schist with pods of semimassive sulphide from the T2 zone (middle row; TP12-033, 149.3 m); **d)** diffuse chlorite-rich bands in impure chert (arrows, top row) overlying chlorite schist (bottom two rows; TP12-032, 251.4 m).*

It is diffusely layered to massive, biotite bearing and clinozoisite rich with minor carbonate.

The impure chert is overlain by a sequence of silicate-facies iron formation and impure chert at least 11 m thick. The iron formation is dark brown-grey, medium grained and locally magnetic. It is laminated to layered with local stringy chert laminations; sulphide, biotite and garnet bearing; and hornblende rich. The iron formation is compositionally gradational into impure chert, which forms bands <2 m thick intercalated throughout the sequence. The chert is grey, fine to medium grained and nonmagnetic. It consists of a garnet- and hornblende-bearing siliceous rock with minor biotite.

The progression from pelite to sulphide-facies iron formation (see P2 member) to a sequence of predominantly chert, silicate-facies iron formation and calcsilicate suggests that the latter sequence represents the Pipe formation P3 member (Bleeker, 1990; Zwanzig et al., 2007). The silicate-facies iron formation with stringy chert laminations is similar to iron formation with stringy or 'ornamental' chert occurring in the lower part of the P3 member at the Pipe II mine (Macek and Bleeker, 1989). The carbonate laminations and layers that form a local marker horizon at the Tower deposit may represent incipient carbonate deposition correlative with the P3 member dolomite present along the Pipe mine–Birchtree mine trend in the exposed TNB. Although the carbonate layers and laminations appear to be conformable, they could represent carbonate veins that formed during diagenesis or during early prograde metamorphism and became transposed during the D_2 and D_3 deformation events.

The chlorite schist does not to correlate with any known part of the Ospwagan group sequence. No feldspar is present in thin sections of the schist, suggesting very low concentrations of Na. Calcium is likely present in only small amounts, contained in the minor amounts of carbonate. The K content is likely variable, contained within biotite and the minor amounts of muscovite. High concentrations of Mg-Fe and Al are suggested by the abundance of chlorite, garnet and staurolite. Overall, the mineral assemblage of the schist is that of a rock low in alkalis (especially Na and Ca) and rich in Al, Mg and Fe. Rocks of this bulk composition are characteristic of the proximal hydrothermal alteration zones associated with VMS deposits (Lydon, 1988; Franklin et al., 2005; Bailes et al., 2016; Buschette and Piercy, 2016). It should be noted that the sulphidic horizon toward the base of this unit is coincident with the T2 zone mineralization in drillhole TP12-027 (Assessment File 63G13256). The chlorite schist appears to be conformable with the Pipe formation stratigraphy; however, it is possible that it could be a discordant conduit that was transposed during the D_2 and D_3 phases of deformation. The gradational nature of the upper and lower contacts of this unit could indicate the waning influence of the hydrothermal fluids away from the alteration horizon or conduit (Figure GS2017-6-7d).

Intrusions

The Ospwagan group stratigraphy was intruded by various igneous bodies ranging in composition from ultramafic to felsic. Ultramafic schist, likely representing altered bodies of peridotite, occurs at various stratigraphic levels within the P3 member (Figure GS2017-6-8a) in intersections ranging from 10–100 m long. The ultramafic schist is pale grey-green, medium to coarse grained and weakly to strongly magnetic. It typically consists of varying amounts of talc, anthophyllite, chlorite, carbonate and serpentine, with minor magnetite and carbonate. The ultramafic schist commonly grades into ultramafic amphibolite at the margins, possibly representing a pyroxenitic envelope. It is possible that the ultramafic bodies intersected in holes TP10-003 and TP12-033 are correlative; however, it is not possible to correlate the host stratigraphy.

Bands of plagioclase amphibolite were intersected in all drillholes and occur at all stratigraphic levels (Figure GS2017-6-8b). The amphibolite is dark green-grey, medium to coarse grained, relatively homogeneous and nonmagnetic. Minor biotite, and rarely garnet, can be present. Although generally <6 m long, an intersection of amphibolite in hole TP10-003 was approximately 113 m long. The plagioclase amphibolite is interpreted as metamorphosed diabase and gabbro dikes, likely related to the Molson dike swarm (Heaman et al., 2009), or possibly one of the older Paleoproterozoic dike swarms of the northwestern Superior province (Heaman and Corkery, 1996; Halls and Heaman, 2000).

Small granitoid dikes <2 m thick occur sporadically in the core. An intersection of granodiorite >73 m long occurs directly below the Phanerozoic unconformity in hole TP12-032 (Figure GS2017-6-8c). The granodiorite is biotite and muscovite bearing and strongly foliated to protomylonitic. The intrusion could be related to the nearby William Lake dome (Layton-Matthews et al., 2007).

Mineralization

The main focus of this study is the rocks that host the Tower deposit; however, the mineralization of the T1 and T2 zones is discussed here briefly in relation to the stratigraphy. For more information regarding mineralization and resource estimates, the reader is referred to Assessment Files 63G1148 and 63G13256, and Caracle Creek International Consulting Inc. (2013).

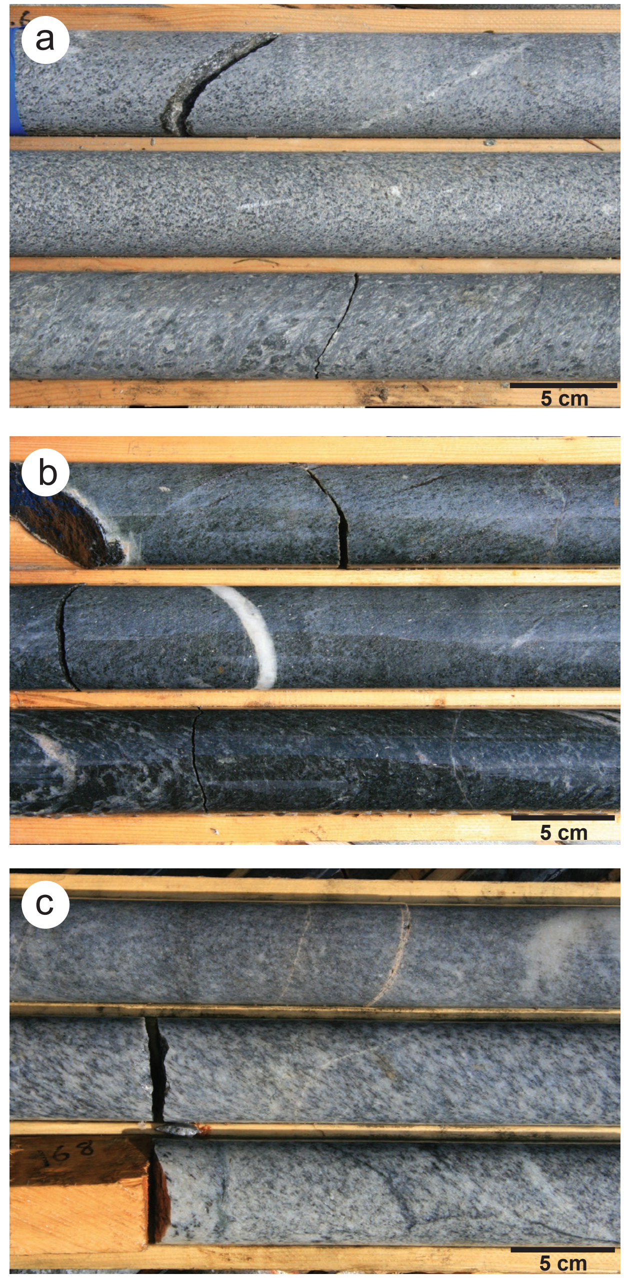


Figure GS2017-6-8: *Intrusive rocks at the Tower deposit:* ***a)*** *altered peridotite now represented by talc- and anthophyllite-rich ultramafic schist (TP10-003, 566.45 m);* ***b)*** *gabbro dike now represented by plagioclase amphibolite (TP10-003, 183.85 m;* ***c)*** *high-strain granodiorite (TP12-032, 167.6 m).*

T1 zone

The nature of the T1 zone mineralization varies along strike from that of a sulphidic schist with foliation-parallel sulphide stringers to solid sulphide. The sulphidic schist is typically biotite rich and contains up to 12% net-textured sulphide (Figure GS2017-6-9a). The solid sulphide occurs as the matrix to a breccia that contains rounded fragments of ultramafic amphibolite and rounded to angular fragments of white quartz and impure chert <7 cm across (Figure GS2017-6-9b).

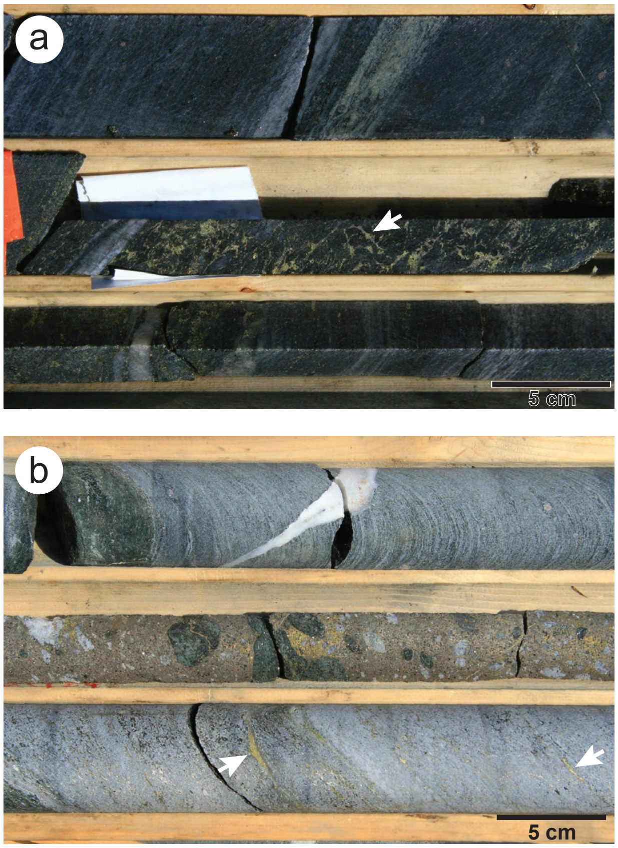


Figure GS2017-6-9: *Mineralization in the T1 zone of the Tower deposit:* ***a)*** *sulphidic schist with stringers of chalcopyrite and pyrrhotite (middle row, arrow indicates crosscutting veinlets of sulphide; TP12-027, 438.4 m);* ***b)*** *sulphide breccia with fragments of ultramafic amphibolite and white quartz (middle row); foliation-parallel stringers of chalcopyrite are visible in the laminated calcareous rock (arrows, bottom row; TP12-032, 565.7 m).*

The fragments are unsorted and matrix supported with rotated fabrics. The contacts of the breccia are locally marked by a seam of siliceous mylonite <5 mm thick. Local foliation-parallel stringers of chalcopyrite are present in the core up to 2 m away from the sulphide breccia. The T1 zone appears to be slightly discordant to the stratigraphy. In the northernmost intersection (TP12-027), the T1 zone occurs in place of the sulphide-facies iron formation, immediately above the P2 member pelite, whereas the mineralization in the southernmost intersection (TP10-003) occurs within the P3 member stratigraphy approximately 96 m above the upper contact of the P2 member pelite. The discordant nature of the mineralization and its occurrence as a sulphide breccia has led to the interpretation that the sulphide was mobilized from its source, forming a fault breccia (Assessment File 63G13256). The brittle-ductile nature of the structure suggests it likely belongs to the later D_3-D_4 phase of deformation.

T2 zone

Mineralization of the T2 zone occurs as disseminated sulphide blebs <1 cm across, and as stringers and pods of semisolid sulphide <5 cm across within the chlorite schist (Figure GS2017-6-7c). The stringers and pods typically have diffuse margins, parallel the foliation and appear to be concentrated toward the bottom of the chlorite schist unit. Of the selected drillcore, the T2 zone was only described from TP12-027 (Assessment File 63G13256); however, mineralization was also described from the same stratigraphic level in TP12-033, and weak Cu mineralization is present in TP12-032. This suggests that, although the T2 zone may pinch and swell, it appears to form a continuous horizon toward the bottom of the chlorite schist across the Tower property, suggesting that it may be stratiform and in situ.

Economic considerations

Volcanogenic massive-sulphide deposits of the pelitic-mafic lithofacies association, or Besshi type, are typically hosted in argillite and siltstone; and, to a lesser extent, in marl, chert and iron formation, intercalated with mafic flows, dikes or sills (Franklin et al., 2005; Piercey, 2011). The mafic magmas commonly have a mid-ocean ridge basalt (MORB) affinity, although magmas of alkali/ocean-island basalt affinity are also reported (Piercey, 2011). Deposits of this type are typically related to arc-rifting in oceanic environments (Franklin et al., 2005; Piercey, 2011). Regional semiconformable hydrothermal-alteration zones are poorly documented within the pelitic-mafic lithofacies association; however, footwall alteration zones with chlorite and quartz-sericite alteration are typically well developed and mineralized (Franklin et al., 2005). The association of potentially stratiform, in situ Cu-Zn mineralization with hydrothermally altered, low-alkali and high–Fe-Mg-Al rocks (T2 zone) strongly suggests that the Tower deposit is of VMS affinity. The pelite and chert-rich chemical sedimentary rocks that host the Tower deposit appear to be part of the Ospwagan group Pipe formation, intruded by mafic and ultramafic sills and dikes. Hostrocks of this type are suggestive of a pelitic-mafic lithofacies association, or Besshi-type, VMS deposit.

Volcanogenic massive-sulphide deposits form in areas of high heat flow, typically related to upwelling mafic magmas in extensional basins (Franklin et al., 2005; Galley et al., 2007; Piercey, 2011). The thinned crust and pooled mafic magmas provide the necessary heat to drive the hydrothermal circulation required to scavenge, transport and deposit base metals (Piercey, 2011). Mafic and ultramafic intrusive rocks are common in the stratigraphy that hosts the Tower deposit and are likely related to the ca. 1883 Ma Molson dike swarm and coeval ultramafic intrusions associated with magmatic Ni deposits in the TNB. Although not intersected in the drillcore at Tower, mafic to ultramafic volcanic rocks occur in the Bah Lake assemblage at the top of the Ospwagan group. Both of these suites are characterized by MORB-like trace-element signatures, which suggests relatively shallow partial melting of the mantle, likely in an extensional environment and accompanied by high heat flow. Hydrothermal activity sufficient to form VMS deposits could be associated with either suite of mafic-ultramafic rocks.

Cummingtonite-cordierite schist is reported from northern Setting Lake (Macek et al., 2006). Rocks of this composition are commonly interpreted to result from metamorphism of intense chlorite alteration (Baldwin, 1976; Pan and Fleet, 1995; Peck and Smith, 2005; Zheng et al., 2011). Although the schist is hosted in the Setting formation, it is in spatial proximity to the Bah Lake assemblage, which could suggest a paragenetic relationship. The Bah Lake assemblage is most abundant in the Setting, Ospwagan, Liz, and Mystery lakes areas but has been largely avoided by exploration programs because of its stratigraphic position upsection from the more prospective Pipe formation. These areas may warrant more detailed study if the VMS mineralization at the Tower property is found to be paragenetic with the Bah Lake assemblage. Examples of Molson-age mafic and ultramafic dikes are ubiquitous in the TNB; if the Tower deposit is found to be contemporaneous with Molson-age intrusions, the entire strike length of the TNB could have notional potential for VMS deposits.

The relationship between the Bah Lake assemblage and the Molson and ultramafic magmatism is disputed. Bleeker (1990) and Zwanzig (2005) called for discrete periods of magmatism, with the Bah Lake volcanism occurring prior to ca. 1890 Ma (Zwanzig, 2005). Conversely, Heaman et al. (2009) considered the Molson-age intrusions to be feeders to the Bah Lake assemblage. In either scenario, the mafic magmatism largely postdates the sedimentation of the Ospwagan group. This has implications for the hydrothermal alteration and mineralization of the T2 zone. If the VMS system was contemporaneous with magmatism, the T2 zone must have formed after deposition of the Pipe formation. This implies that the T2 zone formed either through sub-seafloor replacement, as hydrothermal fluids percolated laterally through the Pipe formation, or the T2 zone represents a transposed footwall stringer zone. Regardless, the alteration and variably mineralized T2 zone can be traced along strike at the deposit scale.

The T1 zone consists of mobilized sulphide, likely within a later D_3-D_4 (brittle-ductile) structure. What remains unclear is the source of this mineralized sulphide. One possibility is that it was sourced from the T2 zone. In this scenario, the T1 zone should continue to climb, up-stratigraphy, to the southwest until it intersects the T2 zone. Another possibility is that the T1 zone is sourced from elsewhere in the stratigraphy. Of note is the relative proximity of the T1 zone to the regionally extensive sulphide-facies iron formation at the top of the P2 member. If this sulphide-rich horizon is the source of the mineralization, it would imply that this horizon could be prospective for VMS mineralization as well as magmatic Ni. Another possibility is that the T1 zone sulphide was sourced from farther up the stratigraphy, where sulphide mounds formed on the paleoseafloor.

Volcanogenic massive-sulphide deposits typically occur in clusters, around calderas or along linear rifts (Galley et al., 2007). The MORB-like character of the mafic magmas in the TNB is suggestive of an extensional environment. This, combined with the apparent lack of a volcanic edifice in the Tower property area, suggests that the VMS system likely developed along an extensional fault. Hence, there may be potential for additional deposits along strike. Although not on trend, intersections of anomalous Cu and Zn have been encountered in drillholes on the adjacent William Lake property (Beaudry, 2007), suggesting that VMS mineralization may well be more widespread than is currently recognized in the sub-Phanerozoic portion of the TNB.

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