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FIELD TRIP GUIDEBOOK: STRATIGRAPHY AND ORE DEPOSITS IN THE THOMPSON NICKEL BELT, MANITOBA

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Field Trip Guidebook: Stratigraphy and ore deposits in the Thompson nickel belt, Manitoba

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Cover illustrations:

Left: Headframe overlooking the flooded open pit of the Pipe II mine.

Right: P2 member sulphide-facies iron formation exposed at the Pipe II mine. This horizon is the locus of the Ni sulphide orebody at the Thompson mine.

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Introduction

The Thompson nickel belt (TNB) occurs along the northwest margin of the Archean Superior craton and forms a segment of the Superior boundary zone of Manitoba (Bell, 1971; Bleeker, 1990). This zone was tectonically reworked during the ca. 1.9–1.7 Ga Trans-Hudson orogeny in response to convergence with the Paleoproterozoic Kisseynew domain and other arc-derived terranes of the Reindeer zone (Figure 1, Bleeker, 1990; Ansdell, 2005). The exposed northern segment of the TNB has a trend of 030°, a width of 30–40 km, and a strike length of approximately 200 km. The southern segment of the belt is covered by Paleozoic limestone and is believed to be continuous over a similar strike length (Burnham et al., 2009).

Nickel exploration began in the TNB in 1946 with a 10 year geophysical and geological exploration program and culminated with the discovery of the Thompson deposit in 1956 by the International Nickel Company of Canada Ltd. (Inco, now Vale Canada Ltd.; Fraser, 1985; Layton-Matthews et al., 2007). As of 2018, there have been approximately 18 deposits discovered in this prolific belt including four past producing mines, three mines currently on care and maintenance, and one mine currently in production (Figure 2, Layton-Matthews et al., 2007). A historical resource for Vale's mining operations in the Thompson area is listed to be 150 Mt at 2.32% nickel and 0.16% copper (Layton-Matthews et al., 2007), and total reserves for the Thompson operation as of 2017 was 27.5 Mt at 1.75% nickel (Lightfoot et al., 2017). Exploration in much of the TNB focuses on the Paleoproterozoic Ospwagan group cover sequence which arguably contains all of the nickel deposit-hosting ultramafic intrusions in the belt. Two regionally extensive horizons within the Ospwagan group are known to contain an abundance of sedimentary sulphide. It is these horizons that are believed to have been a source of crustal sulphur contamination for intruding ultramafic magmas, and host many of the belt's nickel deposits. This field trip will concentrate on the stratigraphy and the setting of nickel mineralization in the TNB.

Tectonic evolution of the Manitoba–Saskatchewan segment of the Trans-Hudson orogen

The Manitoba–Saskatchewan segment of the Trans-Hudson orogen (THO) can be subdivided into four main components, the tectonically reworked margins of the Archean Hearne, Superior, and Sask cratons, and the Paleoproterozoic Reindeer zone (Figure 1, Lewry and Collerson, 1990; Ansdell, 2005; Corrigan et al., 2009). The Sask craton is interpreted to underlie much of the Reindeer zone and Paleozoic rocks to the south; however, it is only exposed in three small windows in the Flin Flon–Glennie complex (Corrigan et al., 2009). The Reindeer zone consists of volcanic, intrusive, and sedimentary rocks that

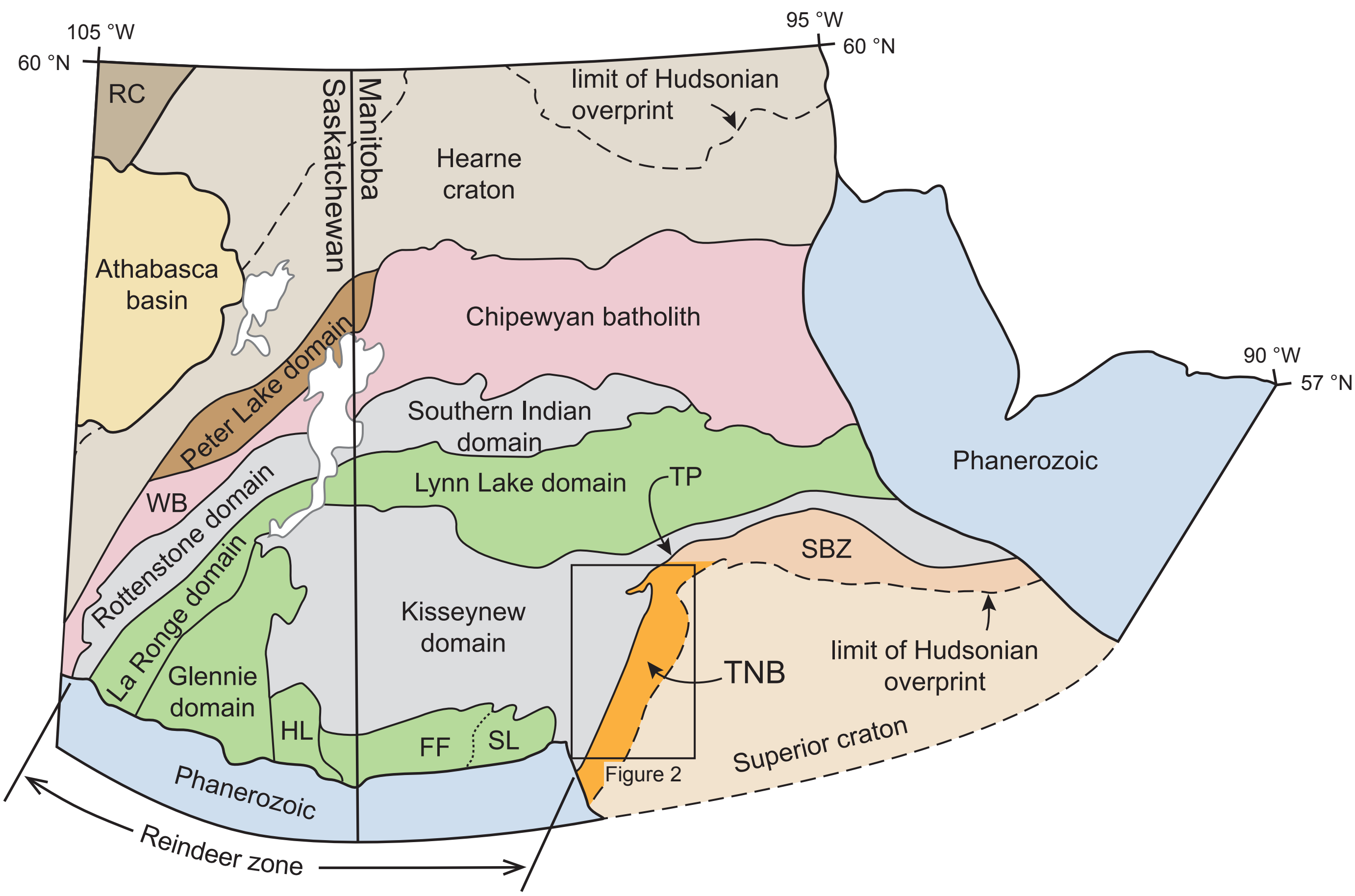

__Figure 1:__ Tectonic-elements map of the Manitoba–Saskatchewan segment of the Trans-Hudson orogen (modified from Lewry et al., 1990). Abbreviations: FF, Flin Flon domain; HL, Hanson Lake block; RC, Rae craton; SBZ, Superior boundary zone; SL, Snow Lake subdomain; TNB, Thompson nickel belt; TP, Thompson promontory; WB, Wathaman batholith.

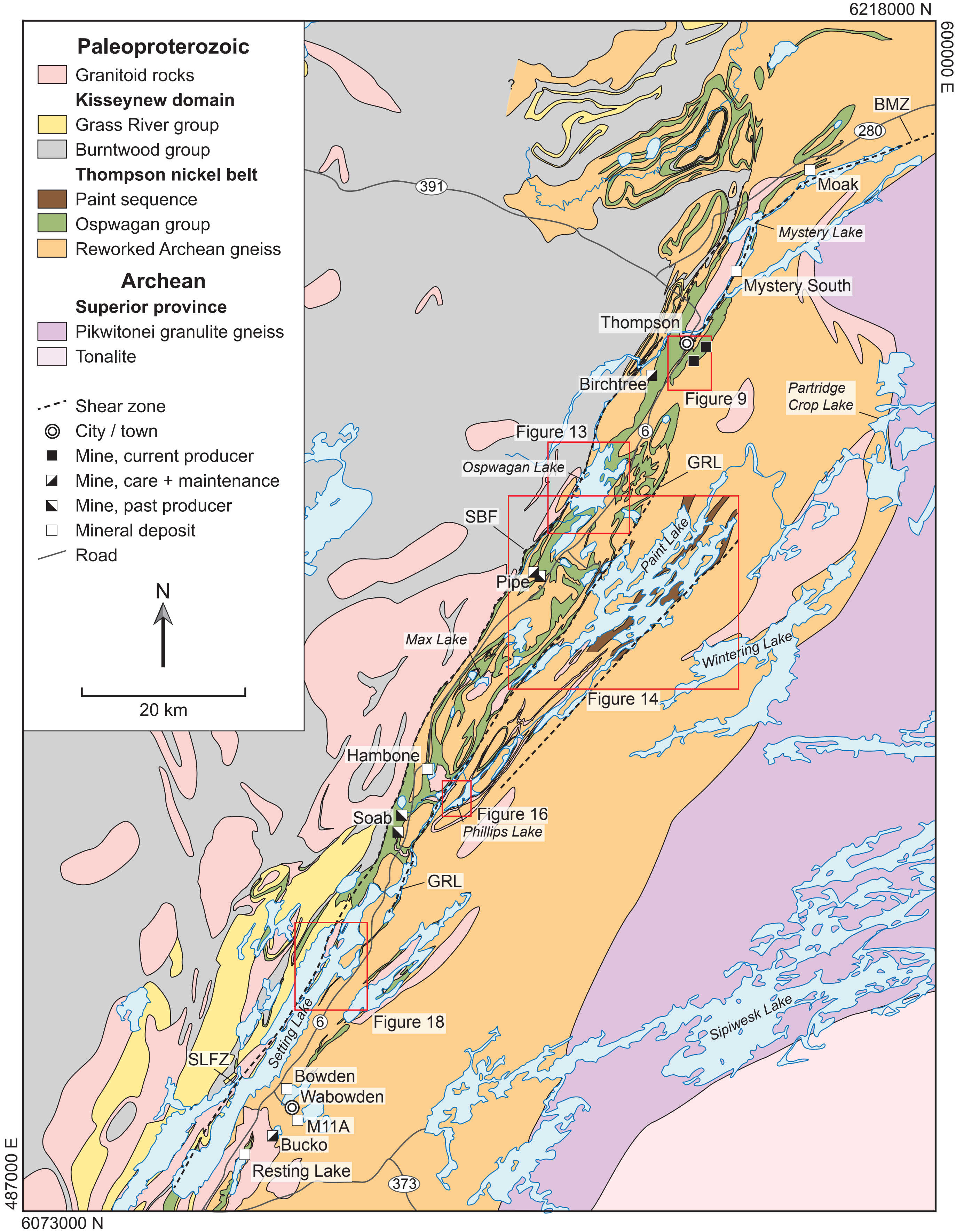

***Figure 2**: Geology of the Thompson nickel belt and adjacent portions of the Kisseynew domain and Superior craton (modified from Macek et al., 2006; Manitoba Mineral Resources, 2013). Abbreviations: BMZ, Burntwood mylonite zone; GRL, Grass River lineament; SBF, Superior boundary fault; SLFZ, Setting Lake fault zone.*

are considered the remnants of the Manikewan Ocean that once separated the three Archean cratons (Stauffer, 1984). These rocks developed in a variety of tectonic environments ranging from island-arc and back-arc to oceanic island and ocean floor settings (Ansdell, 2005). The timing of the initial rifting of the Manikewan Ocean is poorly constrained, but estimated at ca. 2.1 Ga (Halls and Heaman, 2000; Ansdell, 2005; Corrigan et al., 2009). After basin formation, passive margin sequences were deposited along the Hearne and Superior craton margins (Needle Falls and Ospwagan groups, respectively, Ansdell, 2005; Zwanzig et al., 2007; Corrigan et al., 2009).

The oldest preserved arc rocks in the Reindeer zone are dated at ca. 1.92 Ga and indicate active subduction in the Manikewan Ocean (Ansdell, 2005; Corrigan et al., 2009). Similar magmatic ages from the Lynn Lake, La Ronge, Glennie, and Flin Flon belts suggests that arc rocks within the Reindeer zone may have all been generated above subduction zones that were linked (Ansdell, 2005). Oceanic-arc magmatism continued until at least ca. 1.88 Ga, and the Flin Flon–Glennie complex developed with the juxtaposition of arc rocks along faults at ca. 1.87 Ga (Ansdell, 2005; Corrigan et al., 2009) or successively between 1.88 Ga and 1.85 Ga (Zwanzig and Bailes, 2010). Accretion of the La Ronge–Lynn Lake arcs to the southeast margin of the Hearne craton occurred by ca. 1.87 Ga, and was followed by the generation of the Chipewyan (Wathaman) batholith from ca. 1865 to 1845 Ma (Ansdell, 2005; Corrigan et al., 2009). The Chipewyan batholith represents an Andean-type magmatic arc that developed on the southern margin of the Hearne craton and implies an active subduction zone dipping under the craton margin and its accreted terrane at this time.

The Kisseynew domain (KD) is interpreted to be the remnants of a pre-1.89 Ga ocean basin that was trapped during continental collision (Zwanzig and Bailes, 2010). The basin was filled by turbidite deposits (Burntwood group) starting ca. 1.855 Ga, at the onset of terminal collision. Alternatively, the Kisseynew paleobasin might have formed after collision of the Flin Flon–Glennie complex with the southern margin of the La Ronge–Lynn Lake domain (Ansdell et al., 1995). In this scenario back-arc rifting between the Flin Flon–Glennie complex and the La Ronge–Lynn Lake domain resulted in the formation of the paleobasin which quickly filled with turbiditic deposits derived from the adjacent arcs from ca. 1.855 to 1.84 Ga (Ansdell et al., 1995). Correlative fluvial-alluvial deposits (Missi and Sickle groups) were generated on the flanks of the Flin Flon–Glennie complex and La Ronge–Lynn Lake domain, respectively. The correlative fluvial-alluvial Grass River group (Zwanzig et al., 2007) occurs along the flank of the TNB, but was shed from the arc terranes of the Reindeer zone, suggesting early (ca. 1.84 Ga, Percival et al., 2005) collision between these tectonic elements.

Successor-arc magmatism from ca. 1.86 to 1.83 Ga suggests continued subduction underneath the Flin Flon–Glennie complex (Ansdell, 2005). This culminated with collision between the arc accretionary complex and the Sask craton. Continued convergence led to underthrusting of the Sask craton below the Flin Flon–Glennie complex and overthrusting of the Kisseynew basin deposits on the complex. Terminal collision of the Superior craton with the Hearne craton and intervening Reindeer zone was underway by ca. 1.84–1.83 Ga, and represents the final closure of the Manikewan Ocean (Ansdell, 2005; Corrigan et al., 2009; Zwanzig and Bailes, 2010; Couëslan et al., 2013). During collision the northwestern part of the TNB or a similar pericontinental terrane fragment with Archean basement was thrust beneath the northeastern part of the KD (Zwanzig et al., 2006; Whalen et al., 2008). These rocks are exposed in several structural culminations in the KD. Terminal collision led to widespread tectonothermal reworking of elements in the THO.

Geology of the Thompson nickel belt

The TNB is largely underlain by reworked Archean gneiss of the Superior craton (Figure 2). The gneiss is typically quartzofeldspathic with enclaves of mafic to ultramafic rock. It is typically migmatitic and characterized by complex internal structures which 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 Pikwitonei granulite domain 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 Pikwitonei granulites were exhumed and unconformably overlain by the Paleoproterozoic supracrustal rocks of the Ospwagan group (Figure 3, Scoates et al., 1977; Bleeker, 1990; Zwanzig et al., 2007). The majority of rocks in the TNB have been subjected to at least amphibolite-facies metamorphic conditions; however, the 'meta-' prefix has been omitted from rock names for brevity. Local undeformed and unmetamorphosed gabbro dikes that are orthogonal to the regional trend of the TNB belong to the ca. 1270 Ma Mackenzie large igneous province (LeCheminant and Heaman, 1989; Mackie et al., 2009).

The Ospwagan group

The following description of the Ospwagan group is sourced largely from Bleeker (1990) and Zwanzig et al. (2007). The lowermost unit of the Ospwagan group is the Manasan formation which consists of two members: the lower M1 member consists of layered to laminated sandstone with local conglomeratic layers near the base; and the overlying M2 member consists of semipelitic rock. The mica-rich, quartzofeldspathic character of the M2 member makes it susceptible to partial melting at higher metamorphic grades. As a result, its appearance can vary considerably with metamorphic grade and degrees of partial melting. The Manasan formation is interpreted as a fining-upward, transgressive 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 consists of a variety of calcareous–siliceous rock, calcsilicate, and impure marble; the T2 member is a semipelitic, calcareous gneiss that is only rarely present; and the T3 member consists of impure dolomitic marble with local horizons of calcsilicate. Rocks of the Thompson formation, especially the T3 member, make an excellent stratigraphic marker that varies little with metamorphic grade and is easily recognized even in

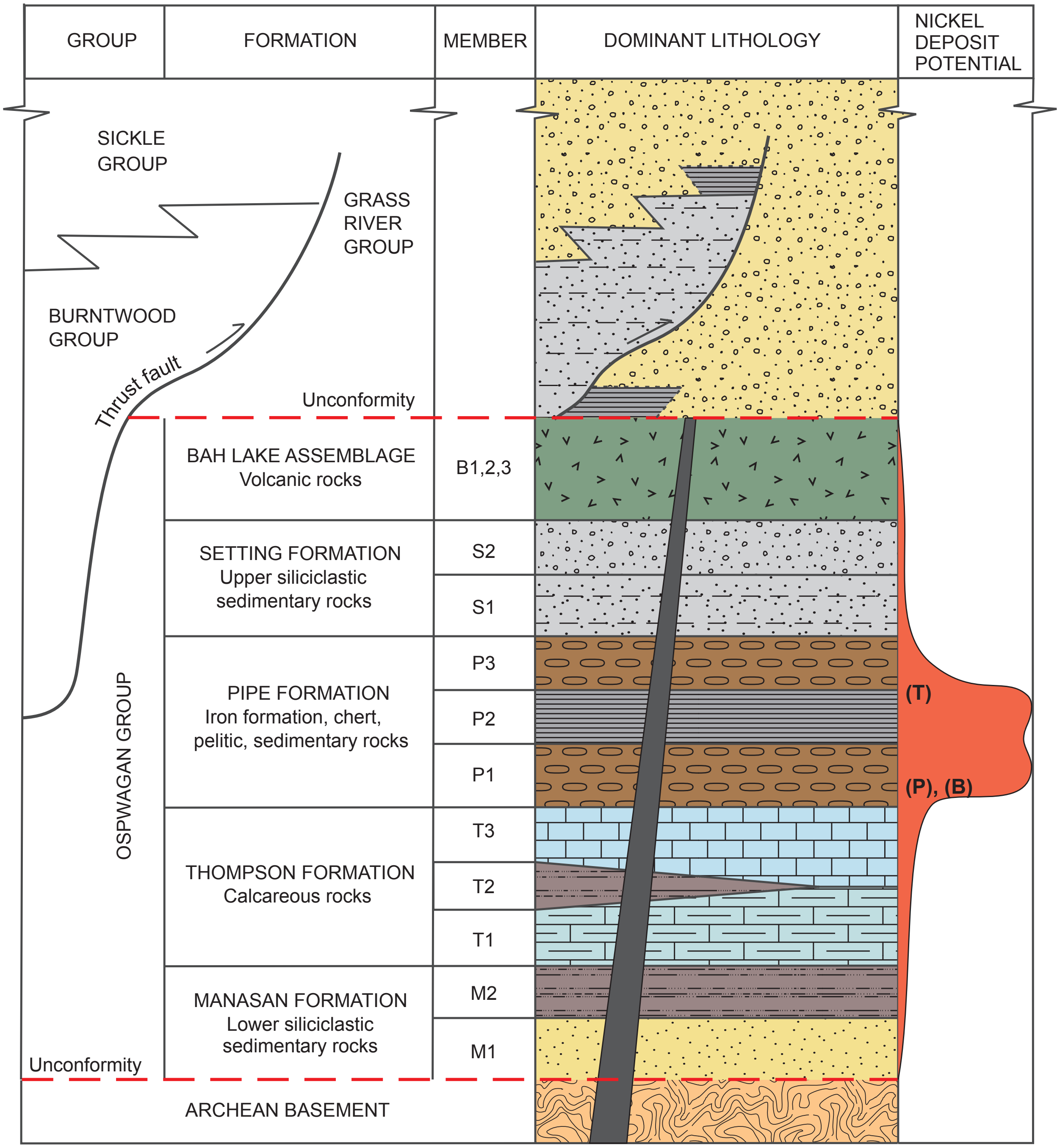

***Figure 3:** Schematic tectonostratigraphic and lithostratigraphic section of the Ospwagan group from Zwanzig et al. (2007; modified after Bleeker, 1990). (T), (P) and (B) indicate the lithostratigraphic positions of the mineralized ultramafic sills at the Thompson, Pipe II, and Birchtree mines, respectively.*

areas that attained lower granulite-facies metamorphic conditions. The Thompson formation represents a transition from a siliciclastic-dominated to a carbonate-dominated system.

The Pipe formation is subdivided into three members. The P1 member consists of a graphite-rich, sulphide-facies iron formation at the base (the locus of the Pipe II and Birchtree ore bodies) which is 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 the P2 member is marked by the presence of a sulphide-facies iron formation (the locus of the Thompson orebody). The P3 member consists of a wide variety of rock-types including laminated, siliceous sedimentary rocks; silicate-, carbonate-, and local oxide-facies iron formations; 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 consists of all siliciclastic rocks above the uppermost P3 member iron formation. The Setting formation is divided into two members. The S1 member consists of rhythmically interbedded quartzite and pelitic schist with local calcareous concretions. The concretions are characteristic of the S1 member. The S2 member consists of thickly layered metagreywacke, with local horizons grading from conglomeratic at the base to pelitic at the top. No contact between the S1 member and S2 member has been observed. The S2 member appears to be missing altogether in the Pipe mine area, where contacts between the S1 member and the overlying Bah Lake assemblage are clearly exposed (Zwanzig et al., 2007). It is possible that they represent a lateral facies change as opposed to a vertical succession. 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 shallowing of the basin, the onset of active tectonism or a lateral sedimentary facies change to a submarine-channel or upper-fan environment (Zwanzig et al., 2007).

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 mafic tuff, and minor synvolcanic intrusions. 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; 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).

The clastic material that formed the Ospwagan group was sourced almost entirely from erosion of the Superior craton. Whole rock neodymium-model ages for the Ospwagan group range from ca. 3.2 to 2.8 Ga. This is slightly younger than the ages determined for the underlying Archean gneiss (ca. 3.5–3.2 Ga; Böhm et al., 2007), but similar to model ages determined for the North Caribou terrane of the Superior craton to the southeast (ca. 3.32–2.84 Ga; Skulski et al., 2000; Rayner and Stott, 2005; Corkery, unpublished data, 2007; Couëslan, 2016). Hundreds of Archean detrital zircon have been dated from the Ospwagan group (Rayner et al., 2006; Burnham et al., 2009; Machado et al., 2011); however, a single ca. 1.97 Ga zircon was recovered from Setting formation greywacke (Bleeker and Hamilton, 2001), providing a maximum age for deposition. 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. 1880 Ma (Bleeker, 1990; Zwanzig et al., 2007; Heaman et al., 2009; Scoates et al., 2017).

The Ospwagan group rocks were subjected to multiple generations of deformation and metamorphic conditions as high as the lower granulite-facies during the THO. Granulite-facies assemblages in semipelitic and pelitic rocks of the Ospwagan group can become almost indistinguishable from the Archean basement gneiss; however, petrological end members such as marble, quartzite and iron formation remain recognizable at the highest metamorphic grades and can be used as marker horizons. Basement-like gneiss or migmatite successions between isolated but still recognizable marker horizons may represent 'ghost successions' of the Ospwagan group (Zwanzig et al., 2007). Distinguishing Archean from Ospwagan group rocks at high metamorphic grade may require the use of lithogeochemistry or Sm-Nd isotope geochemistry (Böhm et al., 2007; Zwanzig et al., 2007).

The Paint sequence

The 'Paint sequence' refers to three northeast-striking belts of metasedimentary rocks that occur in the Paint Lake area and appear to continue along strike to the Phillips Lake area (Couëslan, 2016, 2018). The stratigraphy of the Paint sequence is unconstrained but consists dominantly of arkosic wacke and arkosic arenite, with subordinate iron formation and pelite, and rare boudins of calcsilicate. Paint sequence rocks have only been recognized east of the Grass River lineament (Figure 2), in areas of granulite-facies metamorphism where primary textures and structures are all but obliterated save for centimetre-scale compositional layering. Wacke is the most abundant member of the sequence. It commonly contains centimetre- to decimetre-thick layers of arenite and iron formation. Pods of in situ to in-source leucosome are abundant. Outcrops of wacke are characterized by rusty weathered surfaces because of the presence of minor but ubiquitous pyrrhotite. The composition of the wacke and arenite are gradational into each other and they are commonly interbedded. The arenite is typically interlayered with centimetre- to metre-thick layers of wacke and rarely occurs interlayered with pelite. The iron formation occurs as discontinuous layers and lenses <3 m thick within the wacke. Iron formations are typically of the silicate facies; however, significant pyrrhotite and magnetite can be present.

A maximum age for the Paint sequence is provided by five ca. 2436 Ma detrital zircon grains obtained from a sample of wacke (Couëslan, 2016). The Paint sequence rocks are intruded by relatively straight-walled mafic dikes, which are tentatively interpreted to be part of the Paleoproterozoic Molson dike swarm, suggesting a minimum age of ca. 1880 Ma for the sequence. Whole rock neodymium-model ages for the Paint sequence range from ca. 3.6 to 3.0 Ga (Couëslan, 2016, unpublished data, 2019). This overlaps with model ages determined for the Ospwagan group (ca. 3.2–2.8 Ga) and the underlying Archean gneiss (ca. 3.5–3.2 Ga; Böhm et al., 2007). The Paint sequence rocks contrast with the Ospwagan group rocks in terms of their distinct detrital zircon population and unique trace-element compositions (Couëslan, 2016, 2018). Recent work suggests the Paint sequence may have potential to host mineralized ultramafic intrusions (Couëslan, 2018).

Ultramafic magmatism and nickel sulphide ores

Most ultramafic bodies in the TNB occur as discrete disjointed boudins characterized by tectonized contacts with the enclosing country rock. Larger, less deformed bodies commonly exhibit a thin pyroxenitic basal zone and a thicker lower zone of chromite-bearing peridotite or dunite that grades upward into a thinner pyroxenite upper zone. This asymmetric variation in cumulate olivine and whole rock MgO content is consistent with the younging direction when enclosed in Ospwagan group rocks (Bleeker, 1990; Layton-Matthews et al., 2007; Franchuk et al., 2016). This correlation between magmatic zonation and younging direction suggests the ultramafic intrusions formed concordant to semi-concordant sill-like bodies. Although ultramafic intrusions also occur in Archean gneiss, all economic deposits to date occur where the ultramafic sills are associated with the Ospwagan group cover sequence (Bleeker, 1990; Layton-Matthews et al., 2007; Zwanzig et al., 2007). These deposit-hosting ultramafic bodies generally occur along two horizons within the Pipe formation characterized by high concentrations of sedimentary sulphides (sulphide-facies iron formations; Bleeker, 1990; Zwanzig et al., 2007; Lightfoot et al., 2017). Uranium-lead zircon ages of ca. 1880–1877 Ma have been determined for ultramafic bodies at the Pipe II mine, Setting Lake, and Paint Lake (Hulbert et al., 2005; Heaman et al., 2009; Scoates et al., 2017). These ages are identical to the potentially comagmatic mafic to ultramafic intrusions of the Molson dike swarm (ca. 1880 Ma, Scoates and Macek, 1978; Heaman et al., 1986, 2009). The ultramafic bodies and mafic dikes provide a minimum age for the Ospwagan group which was intruded by these rocks at all stratigraphic levels (Bleeker, 1990; Burnham et al., 2009). A suite of large coeval granitoid plutons occur within the TNB (ca. 1.89–1.87 Ga), which are attributed to either continental arc-magmatism (Zwanzig et al., 2003; Percival et al., 2004, 2005) or crustal melting associated with the mafic magmatism (Heaman et al., 2009). A ca. 1880 Ma thermal pulse is suggested by metamorphic monazite in pelite at Paint Lake (Couëslan et al., 2013) irrespective of the petrogenesis of the felsic magmas.

Thompson nickel belt ores typically consist of pyrrhotite–pentlandite or pyrrhotite–pentlandite–pyrite with minor chalcopyrite; however, millerite, gersdorffite, cubanite, and nickeline have also been described (Eckstrand et al., 1989; Bleeker, 1990; Chen et al., 1993; Layton-Matthews et al., 2007). Sulphur isotopic data for Thompson ores are typically within mantle range values; however, Se/S ratios are indistinguishable from, or only marginally higher, than the sedimentary sulphides found within the Ospwagan group (Figure 4, Eckstrand et al., 1989; Bleeker, 1990). Mixing calculations based on the Se/S ratios indicate 70–100% of the sulphide mass was derived through the assimilation of country rock sulphide (Bleeker and Macek, 1996). Thompson nickel belt ores occur as interstitial, net-textured, brecciated, and solid magmatic sulphide intimately associated with the ultramafic bodies, and locally associated nickel-enriched sedimentary sulphide that formed through the redistribution of nickel and other metals during high-grade metamorphism (Eckstrand et al., 1989; Bleeker, 1990; Lightfoot et al., 2017). The nickel redistribution halo is typically in the range of tens of metres to possibly hundreds of metres at Thompson.

Deformation and metamorphism

The Ospwagan group was affected by four main generations of deformation during the Trans-Hudson orogeny (Bleeker, 1990; Burnham et al., 2009). Early deformation (D_1)

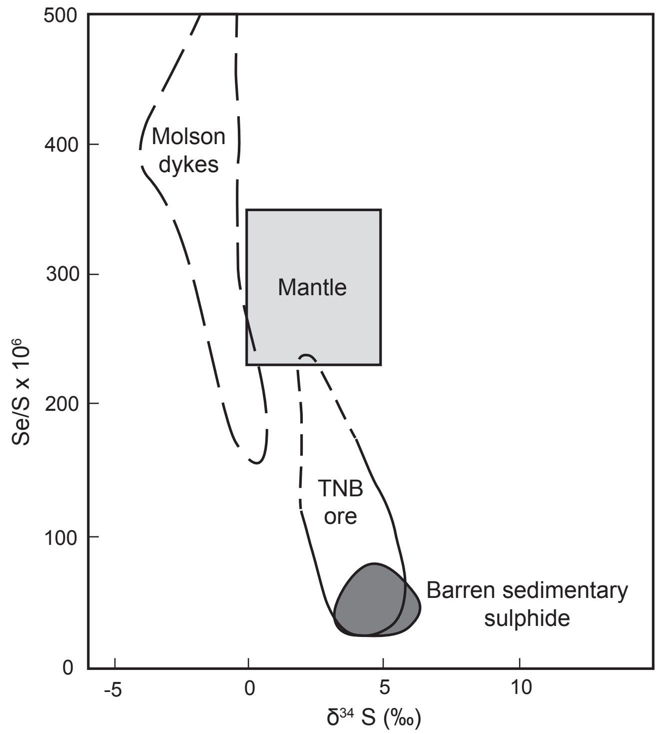

***Figure 4**: Comparison of δ^{34}S and Se/S x 10^6 ratios of sulphides from Thompson nickel ores, Ospwagan group sedimentary rocks, and Molson dikes (Eckstrand et al., 1989).*

predates the ca. 1880 Ma mafic to ultramafic magmatism; however, this generation is largely obscured by later deformation. F_1 folds are only well defined on the shoulder of the Pipe II open pit, where cross-cutting mafic dikes intruded folded Ospwagan group rocks, and where F_1 fold closures are deformed by F_2. Although Bleeker (1990) interpreted this deformation phase as the main nappe-forming event, two lines of evidence suggest otherwise. (1) The magmatic zonation of ca. 1880 Ma ultramafic sills is consistent with the younging direction in the host Ospwagan group stratigraphy. (2) It is now recognized that Kisseynew domain rocks belonging to the ca. 1860–1830 Ma Burntwood and Grass River groups are infolded into the western-most nappe structures (Percival et al., 2005; Zwanzig et al., 2007). Therefore, the nappe structures must post-date the ca. 1880 Ma magmatic event and the deposition of the ca. 1860–1830 Ma Kisseynew basin sedimentary rocks. The D_1 deformation phase is now tentatively interpreted as a relatively upright folding event (Burnham et al., 2009).

The dominant phase of penetrative deformation is D_2, which also affected the mafic intrusions. The F_2 fold generation refolded and tightened F_1 folds and resulted in east-verging (Bleeker, 1990; White et al., 2002), or southwest-verging (Zwanzig, 1998), isoclinal to recumbent F_2 folds (nappes) which incorporated the underlying Archean gneiss (Figure 5a). These are accompanied by regionally penetrative S_2 fabrics. Microstructural observations suggest that peak metamorphic conditions of middle amphibolite- to lower granulite-facies were attained during, and possibly outlasted D_2 (Coueslan and Pattison, 2012). This may have been followed by a period of metamorphic relaxation prior to the D_3 generation of deformation. The D_2 generation of deformation is estimated to have

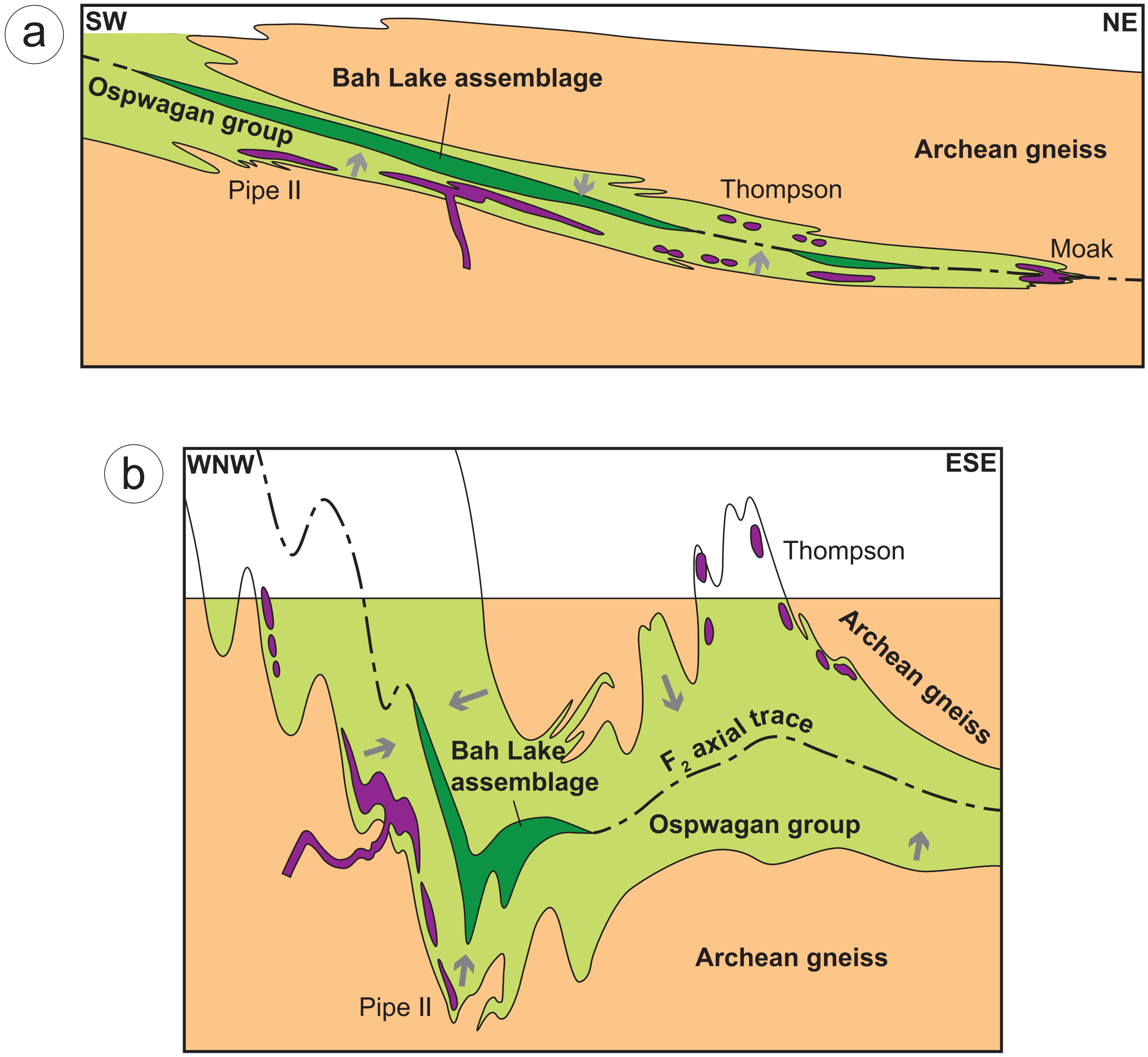

***Figure 5**: Cross-sections through regional fold structures: **a)** northeast longitudinal cross-section through a regional F_2 nappe structure. Upright F_3 folds are neglected in this schematic section to illustrate the original geometery of the F_2 nappe; **b)** schematic east-southeast cross-section through the refolded nappe structure of a) (adapted from Bleeker, 1990).*

begun by ca. 1840 Ma and may have continued until at least ca. 1780 Ma (Couëslan et al., 2013).

The D_3 generation of deformation resulted in isoclinal folds with vertical to steeply southeast-dipping axial planes (Figure 5b, Bleeker, 1990; Burnham et al., 2009). The isoclinal nature of both F_2 and F_3 results in a co-planar relationship between S_2 and S_3 along F_3 fold limbs. Mylonite zones with subvertical stretching lineations parallel many of the regional F_3 folds, and are characterized by retrograde lower amphibolite- to greenschist-facies mineral assemblages (Bleeker, 1990; Burnham et al., 2009). The D_3 generation of deformation was likely underway by ca. 1750 Ma (Couëslan et al., 2013). Tightening of F_3 structures continued during D_4, marked by localized retrograde greenschist-facies metamorphism along mylonitic and brittle cataclastic shear zones that commonly indicate southeast-side-up, sinistral movement (Bleeker, 1990; Burnham et al., 2009). The D_3–D_4 structures appear to exert a first order control on the present day distribution of metamorphic zones within the belt (Couëslan and Pattison, 2012). Four metamorphic zones subparallel the strike of the belt and the regional D_3–D_4 structures (Figure 6). This suggests that peak metamorphism and associated isograds were established prior to, or early during, D_3. The metamorphic isograds were then deformed by D_3–D_4 structures into their current configuration (Figure 7). Steep geothermal gradients occur across strike of the metamorphic-facies zones.

The TNB can be separated into four discrete metamorphic zones that parallel the regional strike of the belt. Three of the zones occur in a nested pattern with an inner middle amphibolite-facies zone, flanked on both sides by upper amphibolite-facies zones, flanked in turn by lower granulite-facies zones (Figure 6; Couëslan and Pattison, 2012). The zones are characterized by low pressure-high temperature metamorphic assemblages with pelitic rocks containing andalusite, sillimanite, and/or cordierite. Metamorphic conditions are estimated to

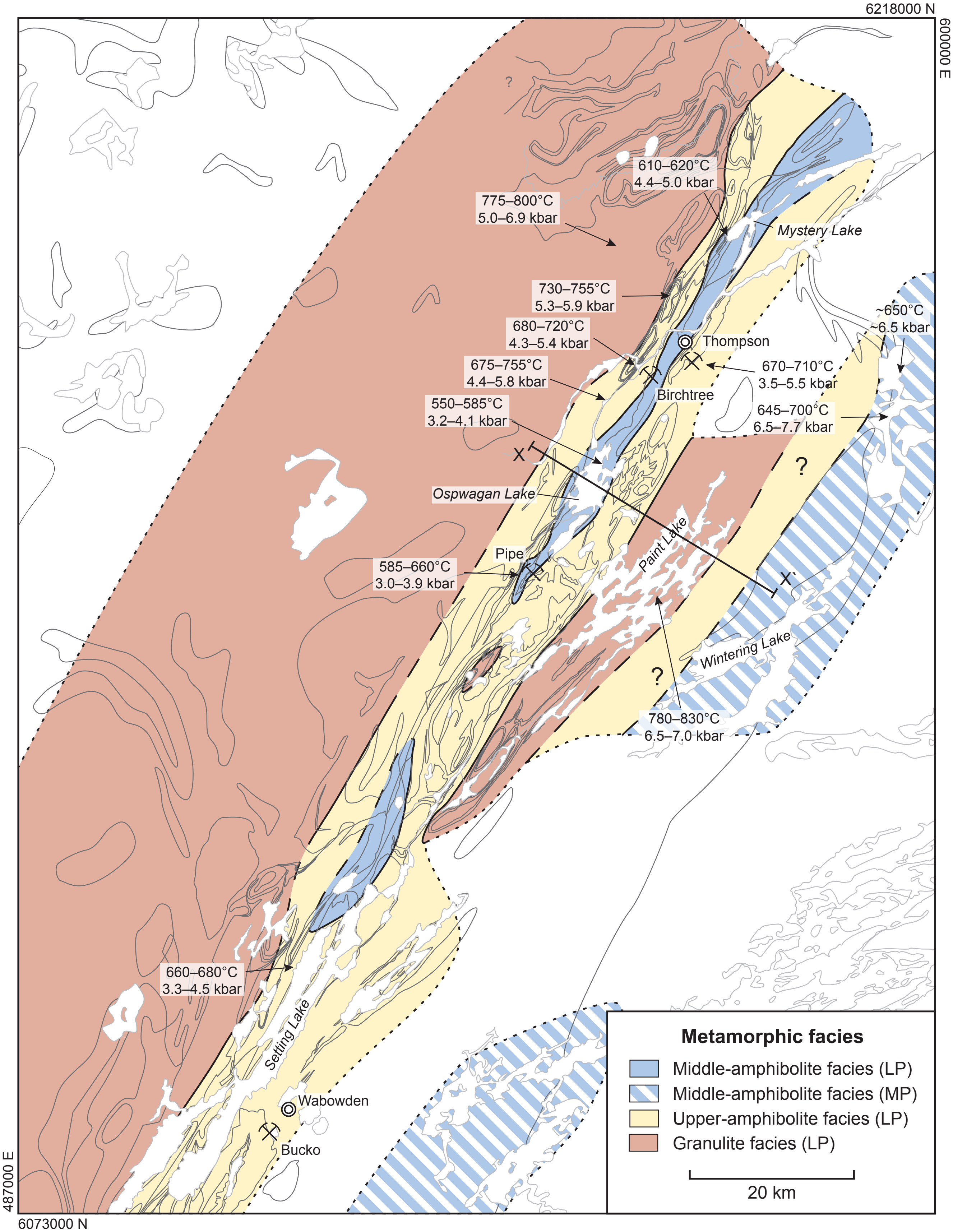

Figure 6*: Metamorphic-facies map of the Thompson nickel belt and adjacent Kisseynew domain (adapted from Couëslan and Pattison, 2012). Pressure–temperature estimates are from observed mineral assemblages and phase-equilibria models. Solid black lines indicate facies-zone boundaries, dashed lines define inferred boundaries, and dotted lines indicate the limits of the data. Solid dark grey lines indicate the the lithological boundaries in Figure 2. Line X–X` indicates the position of the schematic cross section in Figure 7e. Abbreviations: LP, low pressure; MP, medium pressure.*

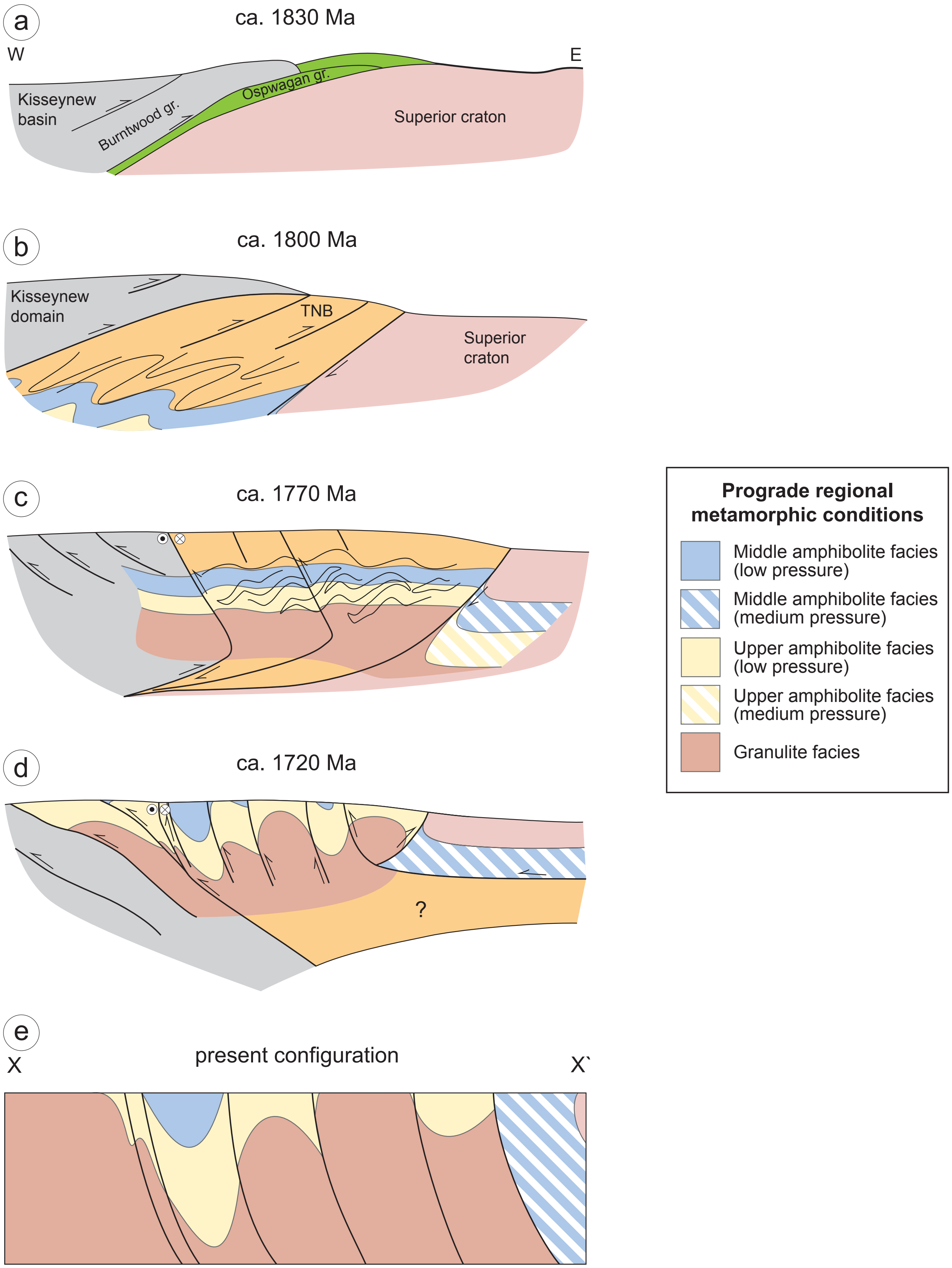

***Figure 7**: Schematic model of the major tectonic phases in the TNB (modified from Ellis and Beaumont, 1999; Couëslan et al., 2013): **a)** thrusting of Burntwood group rocks onto the Superior craton margin along early D_2 faults; **b)** continued convergence (D_2) results in the creation of a thrust and nappe stack, which consists of re-worked Archean gneiss and Ospwagan group rocks with intercalations of Burntwood group rocks; **c)** peak metamorphism was attained during metamorphic relaxation likely outlasting D_2 and possibly continuing into early D_3; **d)** the metamorphic-facies zones were deformed by upright F_3 folding accompanied by steeply east-dipping mylonite zones that record vertical stretching and sinistral movement; **e)** continued differential uplift along D_3–D_4 structures led to the present day distribution of metamorphic-facies zones as presented in this schematic cross-section line along line X–X` (Figure 6).*

be 550–620°C and 3–5 kbar in the inner middle amphibolite-facies zone, 640–710°C and 3–5.5 kbar in the upper amphibolite-facies zone, and 775–830°C and 5–7 kbar in the flanking granulite-facies zones (Couëslan and Pattison, 2012). Pressure–temperature–time paths in the TNB are broadly clockwise (Figure 8; Couëslan and Pattison, 2012; Couëslan et al., 2013). Although peak metamorphic conditions were likely attained diachronously in the three metamorphic-facies zones, prograde and peak metamorphism is broadly synchronous with ca. 1850–1750 Ma granitoid magmatism, which might have contributed to the low pressure–high temperature metamorphism characteristic of the belt (Couëslan et al., 2013).

Recent work has outlined a fourth metamorphic zone along the eastern margin of the TNB (Couëslan, 2014, unpublished data, 2016). Pelitic rocks from this zone are characterized by kyanite-bearing, middle amphibolite-facies assemblages. The kyanite occurs within Archean gneisses along the transition between the Pikwitonei granulite domain and the Superior boundary zone. Although originally interpreted as Archean (Hubregtse, 1977), the kyanite occurs as part of the overprinting Hudsonian metamorphic assemblage in rocks of the Partridge Crop and Duck lakes areas (Couëslan, 2014, unpublished data, 2016). This finding suggests a zone of medium pressure (Barrovian-style) metamorphism along the eastern margin of the TNB, with metamorphic conditions at Partridge Crop Lake estimated to be 640–700°C and 6.5–8.0 kbar (Couëslan, 2014).

An alternative tectonic model has been presented that promotes long-lived transpressional tectonics from ca. 1850 to 1750 Ma, and possibly as late as 1720 Ma (Gapais et al., 2005; Burnham et al., 2009; Machado et al., 2011). This transpressional model predicts non-systematic variations in metamorphic grade throughout the belt (Machado et al., 2011); however, the metamorphic study of Couëslan (2013) and the delineation of systematic, belt-parallel metamorphic-facies zones is in strong agreement with the nappe-tectonics model of Bleeker (1990).

Overthrust and onlapping stratigraphy of the Kisseynew domain

Geochemical data, Nd isotope ratios, and detrital zircon ages suggest that the detritus of the Grass River group and Burntwood group were eroded from the same juvenile Paleoproterozoic arc terrane (Percival et al., 2005; Böhm et al., 2007; Zwanzig et al., 2007; Burnham et al., 2009; Machado et al., 2011), and that their deposition was partly coeval. Locally preserved sedimentary structures suggest that they represent the deep-water and shallow-water facies, respectively, of the Kisseynew basin fill. Most of the following descriptions of Burntwood and Grass River group rocks are from Zwanzig et al. (2007).

The Burntwood group forms a monotonous succession of greywacke-derived, grey-weathering, weakly graphitic, garnet-biotite gneiss and migmatite. Locally preserved rhythmic layering, alternating between quartz-plagioclase–rich and more garnet-biotite–rich gneiss is interpreted to be derived from gradded greywacke-mudstone beds. Local calcsilicate lenses <5 cm thick were likely derived from calcareous concretions. Narrow belts of Burntwood group rocks in the southern Setting Lake area contain sporadic amphibole, magnetite, or pyrrhotite, and have been mapped as part of a Burntwood group–Grass River group conformable transition. Burntwood group rocks were brought into contact with Ospwagan group rocks only along major faults. On Setting Lake and farther west, the Burntwood group is faulted against Bah Lake assemblage amphibolite, whereas in the northern part of the TNB, it is commonly faulted against the P3 member of the Pipe formation (Figure 3). These faults are interpreted as flats and ramps on early D_2 thrusts.

The Grass River group occurs along the eastern margin of the Kisseynew domain and extends locally into the TNB in the Setting Lake area where the basal contact is sharp and overlies the Bah Lake assemblage. This contact has been interpreted as a strongly sheared unconformity (Zwanzig, 1998). The Grass River group is informally divided into four conformable stratigraphic units: the polymictic conglomerate at the base, followed by the lower, middle, and upper sandstones. The contacts between these units are gradational. The polymictic conglomerate is clast supported by pebbles and cobbles that range from mafic-ultramafic igneous to quartz-rich sedimentary. The conglomerate was likely deposited in an alluvial fan environment, possibly along a fault scarp, or within fault-bounded sub-basins. The lower sandstone is well layered and contains sparse pebbles and isolated, wedge-like conglomerate beds. The sandstone is arkosic with minor biotite, hornblende, epidote, and magnetite. The overall conglomerate to lower sandstone succession constitutes a fining-upward sedimentary wedge, likely deposited in a fluvial-alluvial environment. The middle sandstone is a well-bedded arkosic succession with biotite and magnetite the only mafic minerals. It contains distinct, multicolored calcareous layers and lenses 5–20 cm thick. The upper sandstone is a layered to laminated arksoic sandstone with sparse felsic pebbles and local felsic conglomerate beds. Cross-bedding is common. Mafic minerals consist of biotite and magnetite, and either sillimanite as discrete knots or muscovite depending on the local metamorphic grade. Crossbedding along with multicolored calcareous layers and lenses are common. The succession of middle to upper sandstone is interpreted as a coarsening-upward shallow-water deposit.

Day 1 stop descriptions: Thompson South pit and Pipe pit mines

Day 1 will be split between surface tours of the South pit at the Thompson mine and the Pipe pit at the Pipe II mine. In the morning, we will examine exposures of the lowermost stratigraphy of the Ospwagan group along the west shoulder of the South pit. In the afternoon, we will examine the outcrops of Ospwagan group rocks at the Pipe II mine. The Pipe II mine is the type locality for the Pipe formation which forms extensive outcrops along the east shoulder of the open pit.

Geology of the Thompson mine

The Thompson deposit was discovered in 1956 by Inco Ltd. (now Vale Canada Ltd.) and production began in 1961. The deposit has a 6 km long surface trace, is proven beyond a depth of 2000 m, and is still partly open (Bleeker and Macek, 1996; Vale Canada Limited, 2018). Underground operations

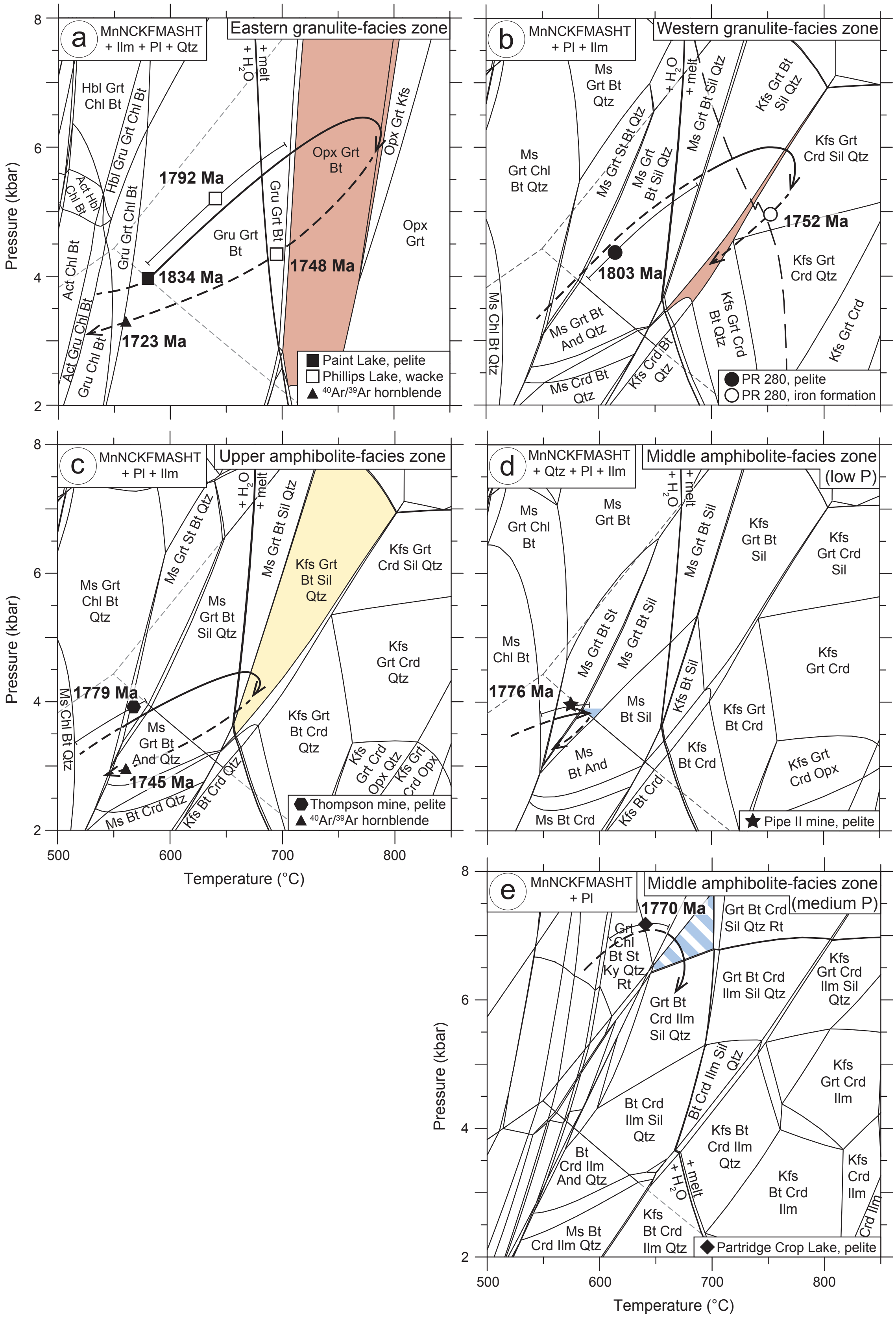

***Figure 8**: Metamorphic P-T-t paths for pelitic rocks from the Thompson nickel belt: **a)** Paint Lake–Phillips Lake, eastern granulite-facies zone; **b)** northwest of Thompson airport near PR 280, western granulite-facies zone; **c)** Thompson mine, upper amphibolite-facies zone; **d)** Pipe II mine, inner middle amphibolite-facies zone (low pressure; adapted from Couëslan et al., 2013); **e)** Partridge Crop Lake, marginal middle amphibolite-facies zone (medium pressure; modified from Couëslan, 2014; Couëslan, unpublished data, 2015). Abbreviations: Act: actinolite; And, andalusite; Bt, biotite; Chl, chlorite; Crd, cordierite; Grt, garnet; Gru, grunerite; Hbl, hornblende; Ilm, ilmenite; Kfs, K-feldspar; Ky, kyanite; Ms, muscovite; Opx, orthopyroxene; P, pressure; Pl, plagioclase; Qtz, quartz; Rt, rutile; Sil, sillimanite; St, staurolite.*

at Thompson are currently accessed from the T1 and T3 headframes, and open pits were operated from 1986 to 1992 to mine the crown pillars of the deposit. The T1 shaft (1350 m deep) is used to access the South End Development, with mining occurring down to the 4850 foot level (1480 m; Golder Associates, 2010; Vale Canada Limited, 2015, 2018; S. Kirby, pers. comm., 2019). The T3 shaft (795 m deep) is used to access the T1D Footwall zone (or Footwall Deep; Golder Associates, 2010; Vale Canada Limited, 2015, 2018). Bulk mining of the Footwall zone below the 4200 foot level (1280 m) began in 2012 (Vale Canada Limited, 2015; Lightfoot et al., 2017). A figure for the total tonnage of the Thompson deposit has not been released, but past production from the mine is estimated at 2.5 Mt Ni, and reserves are estimated at 27.54 Mt at 1.75% Ni (Lightfoot et al., 2017).

The Thompson structure consists of a nearly upright, variably plunging, high-amplitude F_3 antiform (Figure 9, Figure 10). Archean gneiss occurs along the outside limbs of the structure, while Ospwagan group rocks occur in the core. The area was subjected to upper amphibolite-facies metamorphic conditions (670–710°C, 3.5–5.5 kbar) during the Trans-Hudson orogeny which resulted in the partial melting of semipelitic and pelitic members of the Ospwagan group (Couëslan and Pattison, 2012). The main hinge at T1 folds the dominant S_2 foliation and plunges 50–60° toward the south. A considerable portion of the Thompson orebody is contained within this hinge zone, but mineralization also occurs along the east limb associated with parasitic F_3 fold structures (Bleeker, 1990; Lightfoot et al., 2017). The orebody is stratabound and occurs within the upper portion of the P2 member of the Pipe formation, at the same stratigraphic level as the uppermost sulphide-facies iron formation in the Ospwagan group (Bleeker, 1990; Zwanzig et al., 2007). Passive remobilization of ore along the mine horizon has occurred but does not extend beyond the range of the ultramafic boudins that represent the parent sill. Local remobilization of sulphide has also occurred along late faults and as infilling in late-stage tension gashes (Bleeker, 1990; Bleeker and Macek, 1996).

The South pit at Thompson mine is located at the main F_3 hinge, northeast of the T1 headframe. The west shoulder of the pit offers continuous outcrop of the Archean basement–lower Ospwagan group sequence, and provides the best known exposures for the Manasan formation and T1 member of the Thompson formation (Figure 11). The location of these outcrops in an area near the neutral surface of the F_3 antiform allowed for excellent preservation of the section, including the angular unconformity at its base (Macek et al., 2004).

Directions to stops 1-1 to 1-13: South pit, Thompson mine

0.0 km Turn south off of PTH 6 onto Burntwood Road (last traffic light heading south out of Thompson) and head to the Thompson mine and smelter complex. Follow all posted speed limits at the mine property and smelter complex.

2.2 km Stop at main office. The South pit is located immediately northeast of the mill complex and T1 headframe of the Thompson mine. The mine site is only accessible when accompanied by authorized personnel. The location for each stop is indicated in Figure 11.

Stop 1-1: Archean basement gneiss (UTM: 14U 572280 mE, 6175320 mN)

Archean-age gneiss underlies much of the TNB. It is interpreted to be derived from Pikwitonei granulitic gneiss that was tectonically reworked and hydrated during the THO. Gneiss in the Pikwitonei domain was subject 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). Mafic dikes with glassy chill margins intrude the Pikwitonei gneiss and have been dated as old as ca. 2.09 Ga, indicating that the area had been exhumed by this time (Halls and Heaman, 2000). The Archean gneiss was unconformably overlain by the Ospwagan group prior to tectonic reworking during the Trans-Hudson orogeny.

Stop 1-2: Archean gneiss–Manasan formation unconformity (UTM: 14U 572280 mE, 6175320 mN)

This stop highlights the locally angular relationship between the Archean gneissosity and the bedding of the overlying Ospwagan group. The thinly layered, monotonous quartzite of the M1 member of the Manasan formation forms the base of the Ospwagan group. The monotonous quartzite contrasts with the more heterogeneous granitic to granodioritic gneiss of the underlying Archean basement.

Stop 1-3: Garnet-bearing core of mafic dike (UTM: 14U 572305 mE, 6175367 mN)

This stop is located at the core of a composite mafic dike that cuts both the Archean gneiss and Manasan formation rocks. The core of the dike consists of a dioritic phase that contains amphibole, garnet, biotite, plagioclase, and quartz.

Stop 1-4: Mafic dike margin (UTM: 14U 572318 mE, 6175383 mN)

The margin of the composite mafic dike consists of layered gabbro with a fine-grained chill margin at the contact with the enclosing Manasan formation, M1 quartzite. This phase has been dated at ca. 1861 Ma (Bleeker and Hamilton, 2001; Scoates et al., 2017), which is considerably younger than the ca. 1885–1877 Ma Molson dike swarm (Heaman et al., 2009), and is closer in age to the ca. 1870 Ma mafic-ultramafic magmatism in the sub-Phanerozoic Winnipegosis komatiite belt to the south (Waterton et al., 2017).

Stop 1-5: Cross bedding in mafic dike (UTM: 14U 572300 mE, 6175410 mN)

What appears to be cross-bedding is preserved in the layering of the gabbro at this stop. The cross-bedding suggests that the layering formed in a dynamic magmatic environment and is not a product of metamorphic segregation (i.e. not metamorphic layering).

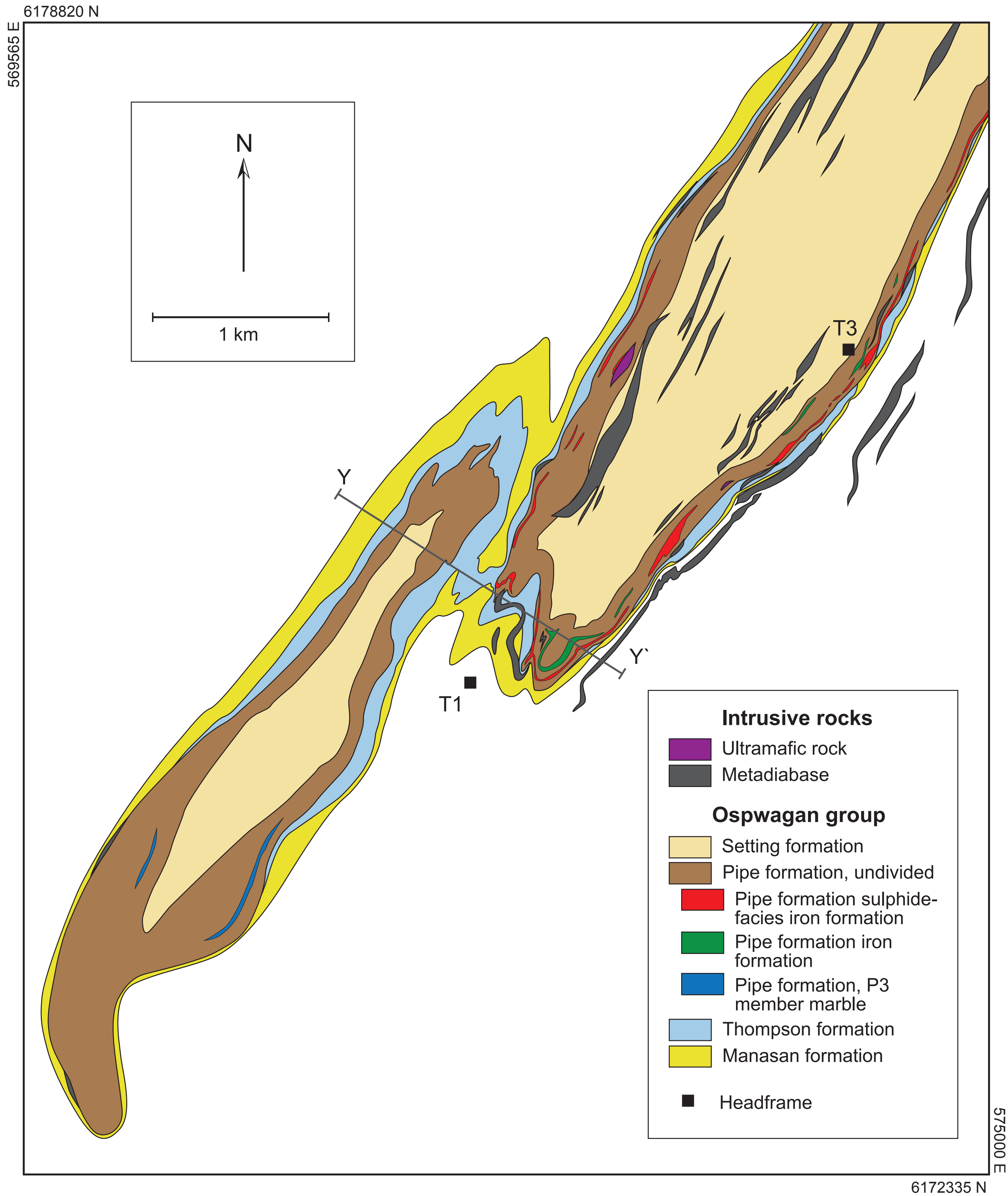

Figure 9*: Geology of the Thompson structure (Macek et al., 2006). Line Y–Y` indicates the location of the cross-section in Figure 10.*

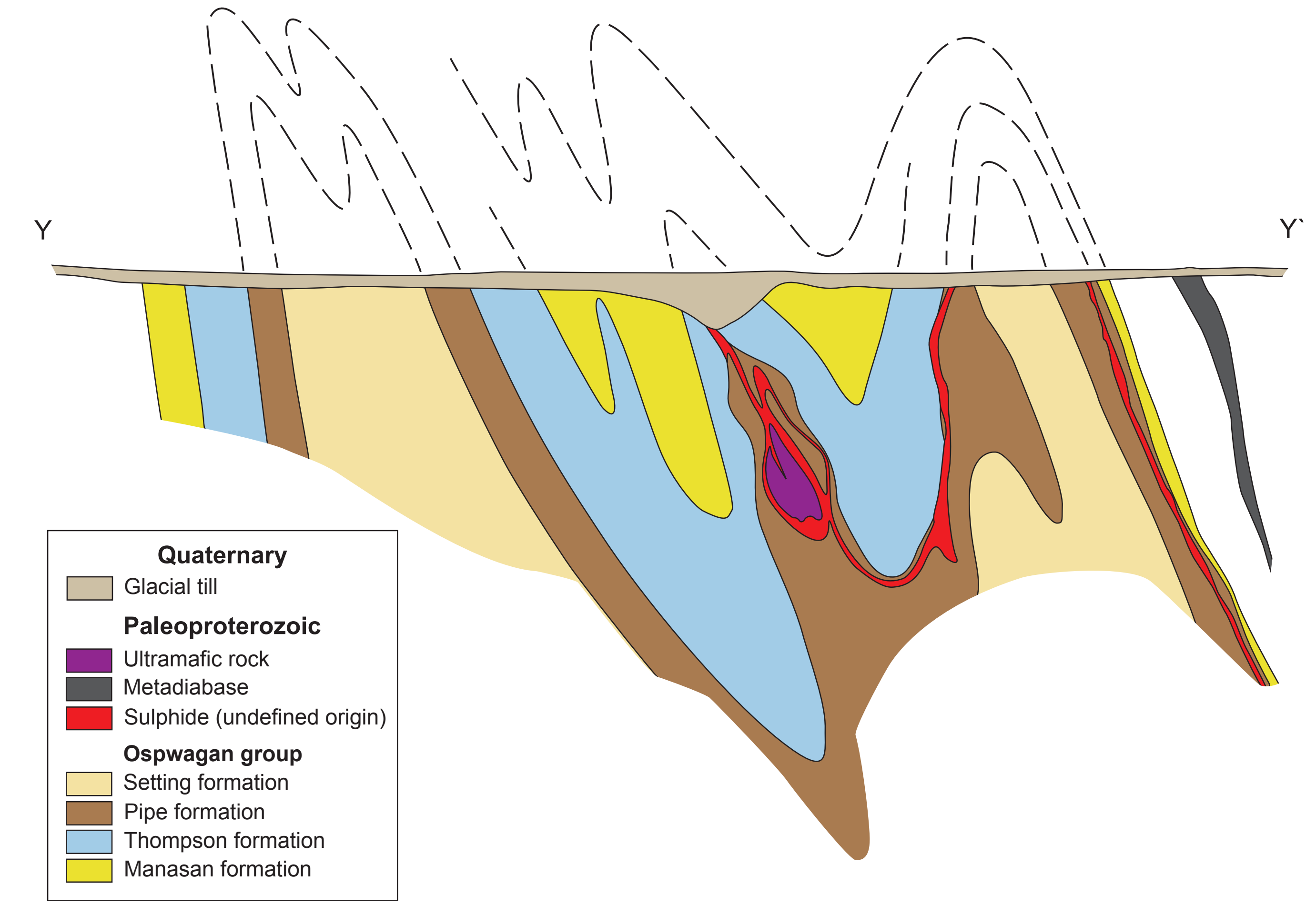

***Figure 10**: Cross-section through the Thompson structure facing northeast (line Y–Y`, figure 9; Bleeker, 1990). The high amplitude, steeply inclined folds belong to the F_3 generation, which refolded the overturned limb of an F_2 nappe.*

Stop 1-6: Composite mafic dike (UTM: 14U 572270 mE, 6175400 mN)

At this stop all three zones of the composite dike can be viewed: the dioritic core, layered gabbro, and chill margin. The layered zone to core margin is sheared and folded. From this vantage point, the orientation of the mafic intrusion can be seen to change from a dike cutting the Archean gneiss, to a climbing sill at the unconformity with the M1 quartzite.

Stop 1-7: Manasan formation, M1 conglomerate (UTM: 14U 572277 mE, 6175413 mN)

Lenses of M1 conglomerate can be seen at this stop, occurring approximately 2–3 m above the unconformity. The conglomerate consists of 1–3 cm long quartz pebbles in dark grey amphibole- and biotite-rich matrix. Conglomerate makes up a minor component of the M1 member of the Manasan formation, which is interpreted as the base of a passive margin sequence.

Stop 1-8: Archean gneiss–Manasan formation unconformity (UTM: 14U 572270 mE, 6175440 mN)

This stop is another example of the angular unconformity between the gneissic layering of the Archean basement and the overlying M1 quartzite. Because of the high strain that typifies much of the TNB, the unconformity at most localities displays a parallel relationship between the Archean gneissosity and the bedding of the overlying Ospwagan group.

The M1 member of the Manasan formation dominantly consists of arkosic quartzite. The quartzite contains quartz, K-feldspar, and plagioclase, along with minor biotite and muscovite. Local carbonate-bearing concretions are also present. The quartzite is well bedded with horizons of varying thickness. The well-defined bedding allows for the tracing of minor folds and “pseudo-boudins”. The pseudo-boudins are interpreted as the product of strain-related dissolution rather than strong ductility contrast. Development begins with the formation of a tension gash which is filled with siliceous material. As the siliceous material in-fills the tension gash, layers above and below are drawn into the boudin neck. The boudin-neck folds die out away from the tension gash toward nearly planar bedding (Macek et al., 2004).

Stop 1-9: Manasan formation, M1 quartzite (UTM: 14U 572324 mE, 6175421 mN)

The rocks at this stop consist of strongly deformed M1 quartzite. Looking toward the northeast from this vantage point, the quartzite of the M1 member can be seen grading into the

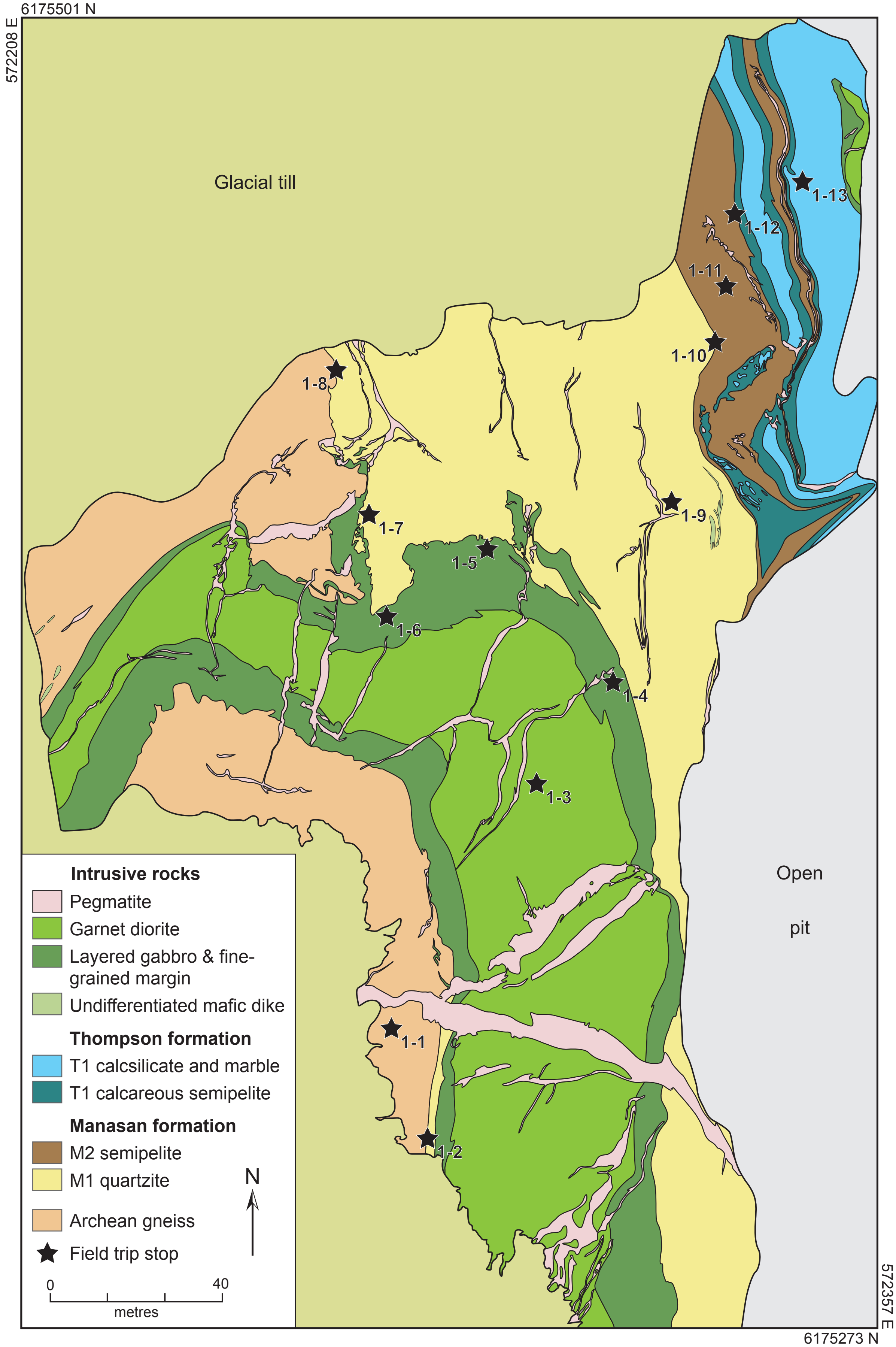

***Figure 11**: Geology of the west shoulder of the South pit at the Thompson mine (adapted from Macek et al., 2005). The location of field trip stops are indicated by stars.*

overlying semipelitic gneiss of the M2 member of the Manasan formation.

Stop 1-10: Contact between the M1 quartzite and M2 semipelite (UTM: 14U 572336 mE, 6175448 mN)

At this stop, we are standing on the contact between the M1 and M2 members of the Manasan formation. The more siliceous M1 member contrasts with the more micaceous composition of the M2 member. The Manasan formation is interpreted to represent a fining upward clastic sequence that grades from relatively mature sand at its base (M1 quartzite) to a more pelitic sediment, possibly a siltstone, at its top (M2 semipelite).

Stop 1-11: Manasan formation, M2 semipelite (UTM: 14U 572334 mE, 6175454 mN)

The semipelite of the M2 member of the Manasan formation contains quartz, feldspar, biotite, and muscovite. Local horizons contain fibrous sillimanite knots. The semipelite is characterized by pools of internally derived melt or leucosome. Although it has been partially melting, a crude primary compositional layering is still apparent. The migmatitic textures of the M2 semipelite can make it appear similar to the Archean basement gneiss; however, the underlying M1 quartzite and overlying calcareous rocks of the Thompson formation can be used as stratigraphic markers to help identify the M2 member.

Stop 1-12: Contact between the Manasan formation and Thompson formation (UTM: 14U 572350 mE, 6175465 mN)

The contact between the M2 semipelite of the Manasan formation and the overlying calcareous semipelite of the T1 member of the Thompson formation is marked by the disappearance of pods of internally derived melt and the appearance of greenish layers of calcsilicate. The T1 semipelite is locally more siliceous than the M2 member, and commonly contains calcic amphibole and epidote.

Stop 1-13: Thompson formation, T1 calcsilicate and marble (UTM: 14U 572358 mE, 6175474 mN)

The T1 semipelite is overlain by mottled diopside-bearing calcsilicate interbedded with variably micaceous marble. The overall trend appears to be toward increasingly carbonate-rich rocks. The T1 member of the Thompson formation is therefore interpreted to represent a switch from a siliciclastic dominated system (Manasan formation) to a carbonate-dominated system (Thompson formation).

Optional stop: Thompson formation, T2 semipelite and T3 marble (UTM: 14U 572419 mE, 6175565 mN)

Calcareous semipelite, possibly representing the T2 member, is in contact with impure marble of the T3 member. The T3 marble is locally sheared and partially retrogressed; however, coarse-grained diopside (commonly replaced by tremolite) and olivine (replaced by serpentine) are present. These coarse-grained minerals are characteristic of this unit, which is also referred to as the T3 'megacrystic' marble.

Geology of the Pipe II mine

The Pipe II mine is located approximately 33 km south of Thompson, Manitoba. After its discovery in 1957, the deposit was mined by Inco Ltd. as an open pit from 1969 to 1984. Clay and silt overburden was removed by dredge prior to mining and exposed extensive outcrops. An exploration shaft was sunk to 480 m and was later deepened to 935 m to facilitate the mining of deeper portions of the orebody; however, unfavourable market conditions pre-empted the start of underground production. The open pit operation halted at a depth of 245 m after producing approximately 18 million tonnes of ore. Historical information is from Galley et al. (1990) and Bleeker and Macek (1996).

The main structure at the mine site consists of a tight F_3 synform, plunging approximately 80° to the northeast (Figure 12, Fueten et al., 1986; Galley et al., 1990). The synform is cored by Ospwagan group rocks and bounded by reworked Archean gneiss. Bedding and foliations are near vertical as is a locally pronounced stretching lineation. Accessible outcrops along the east shoulder of the open pit provide one of the most complete successions of Ospwagan group rocks in the TNB, and make it the type locality of the Pipe formation. The surface trace of the ore zone is approximately 1 km long, most of this along the stratigraphic base of an ultramafic intrusion (Bleeker, 1990; Galley et al., 1990; Bleeker and Macek, 1996). The lens-like ultramafic body occurs on the west limb of the main mine structure and is more than 2 km long and up to 150 m thick. It is considered to be the boudinaged remnant of a more extensive sill, and is the southern-most member of a discontinuous array of ultramafic bodies that stretch from the Pipe II mine to the Birchtree mine along the same stratigraphic horizon (Galley et al., 1990; Bleeker and Macek, 1996). The Pipe II ultramafic body intruded along the base of the Pipe formation, immediately above the P1 member sulphide-facies iron formation (Figure 3, Bleeker, 1990; Galley et al., 1990). The sill is conformable with the enclosing Ospwagan group, and its differentiation profile of sulphide–dunite–peridotite–orthopyroxenite corresponds to the stratigraphic younging direction of the sedimentary succession (Galley et al., 1990; Bleeker and Macek, 1996). Intrusion of the ultramafic body occurred at 1880.2 ±1.4 Ma (Scoates et al., 2017). No evidence for a contact metamorphic aureole is preserved in the enclosing metasedimentary rocks; however, a metamorphic field gradient is recorded at the Pipe II mine that increases west to east, from middle amphibolite-facies assemblages (585–600°C, 3.7–4.0 kbar) adjacent to the ultramafic body to upper amphibolite-facies assemblages (665–710°C, 3.5–4.0 kbar) in eastern exposures (Couëslan et al., 2011).

Directions to stops 1-14 to 1-26: The Pipe II mine

0.0 km Drive southwest at the intersection of PTH 6 and Burntwood Road (last traffic light heading south out of Thompson)

33.5 km Turn left onto gravel road leading to the Pipe mine

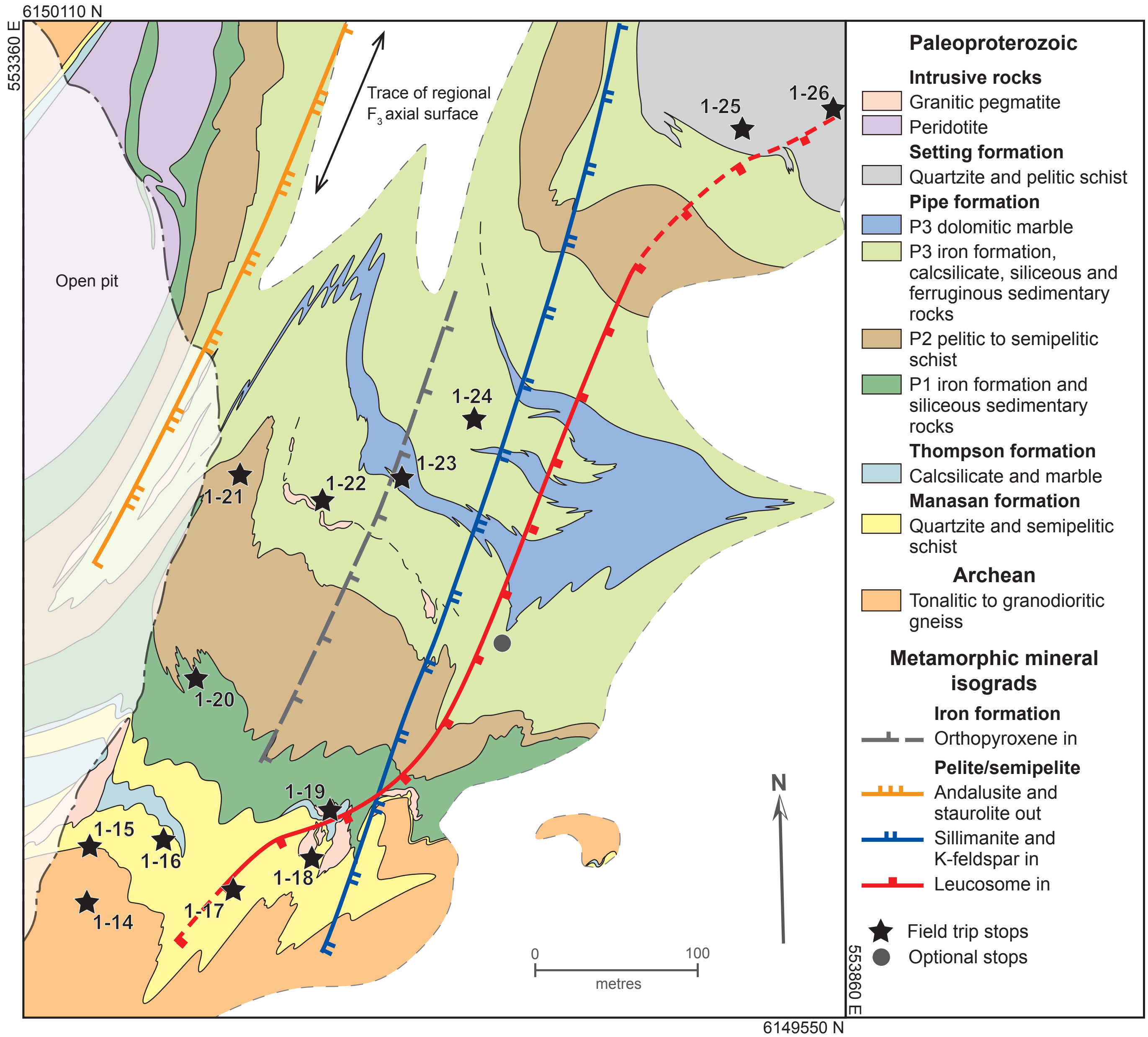

***Figure 12:** Simplified geology of the east shoulder of the open pit at the Pipe II mine (adapted from Bleeker, 1990; Couëslan et al., 2011).*

34.3 km Stop at mine gate. The mine site is only accessible when accompanied by authorized personnel. The location for each stop is indicated in Figure 12.

Stop 1-14: Archean basement gneiss (UTM: 14U 553388 mE, 6149605 mN)

This first stop occurs within the Archean basement gneiss. The gneiss is granodioritic, and contains biotite, hornblende, and rare garnet. The rock was strongly deformed during the Archean and again during the Paleoproterozoic. The gneiss contains bands of amphibolite which likely have multiple origins. Disrupted lenses of amphibolite with diffuse contacts oriented parallel to the Archean gneissosity may represent mafic rocks of Archean age; however, amphibolite lenses that are discordant to the Archean gneissosity, and display sharp contacts with locally preserved flame structures may represent Paleoproterozoic mafic dikes such as those of the Molson swarm.

Stop 1-15: Archean gneiss–Manasan formation unconformity (UTM: 14U 553386 mE, 6149624 mN)

As at many other locations in the belt, the contact between the Archean gneiss and the Manasan formation at Pipe mine appears conformable. This apparently conformable relationship may be the result of transposition due to high strain along the Archean–Ospwagan group unconformity. The Archean gneiss is typically light grey and heterogeneous in contrast to the monotonous beige M1 quartzite. The quartzite contains local pebbly lenses, and becomes characterized by graded bedding toward the contact with the overlying, dark grey M2 semipelite.

The contact between the M1 quartzite and the M2 semipelite is marked by an abundance of quartzofeldspathic segregations.

Stop 1-16: Manasan formation, M2 semipelite (UTM: 14U 553415 mE, 6149634 mN)

The M2 semipelite of the Manasan formation is characterized by monotonous rhythmic layering and an abundance of mm- to cm-scale quartzofeldspathic segregations. It has been suggested that the segregations represent incipient melting and leucosome formation; however, their unusually quartz-rich composition and sharp margins could be more suggestive of subsolidus segregations. The semipelite consists of a quartz-feldspar matrix with abundant muscovite, biotite, and minor tourmaline.

Stop 1-17: Manasan formation–Thompson formation contact (UTM: 14U 553458 mE, 6149614 mN)

As we move east, quartzofeldspathic segregations in the M2 semipelite become more abundant and more feldspathic. Transposition and shearing also become more intense. Progressing toward the top of the M2 member we see a dark layer that is free of quartzofeldspathic segregations, and disrupted layers of calcsilicate that mark the contact with overlying rocks of the Thompson formation. The dark horizon consists of more siliceous and calcareous semipelite of the T1 member of the Thompson formation.

Stop 1-18: Manasan formation, M2 semipelite with abundant pegmatite (UTM: 14U 553498 mE, 6149642 mN)

On a regional scale, the contact between the M2 semipelite of the Manasan formation and the calcareous rocks of the overlying Thompson formation is commonly marked by the presence of pegmatite. At some locations the M2 member may be entirely absent, with pegmatite occurring between the M1 member and Thompson formation rocks. It is not clear if the pegmatite is injected, or represents pools of locally derived melt. Rafts and schlieren of the M2 semipelite are commonly present within the pegmatite. Quartzofeldspathic segregations in nearby M2 semipelite are characterized by diffuse margins and compositions similar to granite, which provide strong evidence for partial melting and the formation of leucosome.

Stop 1-19: Thompson formation–Pipe formation contact, the Pipe II mine horizon (UTM: 14U 553509 mE, 6149660 mN)

At this stop we see beige dolomitic rocks belonging to the Thompson formation that grade into sulphidic chert of the overlying Pipe formation. The sulphidic chert horizon is known as the lower sulphide-facies iron formation, and is the host of the nickel-bearing ultramafic body at the Pipe II mine. It likely provided a source of external sulphur for the intruding ultramafic magma. An isolated boudin of ultramafic schist occurs nearby and could represent a fragment of the Pipe II ultramafic body. Intense deformation that is characteristic of many parts of the TNB is evident from the structural repetition of Manasan, Thompson, and Pipe formation rocks at this location.

Stop 1-20: Pipe formation, P1 silicate-facies iron formation (UTM: 14U 553456 mE, 6149731 mN)

The P1 member silicate-facies iron formation consists of interlayered chert and iron-rich horizons. The iron-rich layers vary in composition but typically contain varying proportions of biotite, grunerite, hornblende, garnet, magnetite, pyrrhotite, and quartz. The upper portion of the P1 member consists of reddish, laminated chert which grades upward into the overlying P2 member pelite. The transition to a depositional system dominated by iron formation, chert, and pelite suggests a deepening of the basin and possibly the development of a foredeep.

Stop 1-21: Pipe formation, P2 pelite (UTM: 14U 553458 mE, 6149828 mN)

The red chert at the base of the P2 member grades upward into the P2 pelitic schist. At the base, the P2 pelite contains zoned lenses of chert characterized by dark carbon-rich rims, pale centers, and local pink, garnet-bearing cores. In general, the P2 pelite becomes increasingly aluminous toward the top, and can contain up to 50% mica. The pelite may be thinly laminated, and contains local cherty and calcsilicate horizons. The top of the P2 member is marked by the upper sulphide-facies iron formation, which coincides with the mine horizon at Thompson.

The sulphide-facies iron formation transitions into siliceous sediment of the overlying P3 member of the Pipe formation. The siliceous sediment is very fine-grained and laminated to thinly bedded. It contains 32 boudinaged and folded layers of calcsilicate. The siliceous sediment is interpreted as a hybrid rock (as is much of the P3 member) formed by the deposition of fine-grained siliciclastic material, carbonate, and chert.

Stop 1-22: Pipe formation, lower P3 member (UTM: 14U 553503 mE, 6149820 mN)

The P3 siliceous sediment grades over several metres into iron formation with stringy chert laminations. This transition occurs as a rhythmic interfingering of the siliceous sediment and iron formation, which suggests a change in the depositional system.

Optional stop: tracing stratigraphy (UTM: 14U 553582 mE, 6149756 mN)

This stop takes another look at the base of the P3 member of the Pipe formation. Boudinaged calcsilicate layers occur within siliceous sediment that grades into the overlying iron formation. Continuing up-section, there is a marker horizon of dense chert followed by a layer of garnet-rich ironstone that grades into calcsilicate. The uppermost ~1.5 m of exposure contains the following transition: the calcsilicate grades upward to a thin, sharp black line, which is overlain by a horizon with a chert-rich base grading into a diopside-rich top, in turn capped by a chert-rich layer of iron formation, at which point the rocks disappear under cover.

Optional stop: tracing stratigraphy, continued (UTM: 14U 553533 mE, 6149809 mN)

At this stop we see garnet-rich ironstone grading into calcsilicate. The calcsilicate is bounded by a thin black line, and is overlain by a thin horizon with a chert-rich base and a diopside-rich top, followed by chert-rich iron formation. This transition, recognized from the previous stop, occurs over an interval of <30 cm at this location. Recognizing this marker horizon allows us to continue to trace the stratigraphy up section from the previous stop.

Stop 1-23: Pipe formation, P3 marble (UTM: 14U 553542 mE, 6149841 mN)

The P3 dolomitic marble is a regionally discontinuous horizon within the Pipe formation. It is beige, layered, and can contain vugs. The marble is relatively pure, although local olivine and other silicates can occur. The marble was intruded by a Molson dike which locally resulted in a thin contact aureole. The lower contact of the marble is transitional into the underlying silicate-facies iron formation.

Stop 1-24: Pipe formation, upper P3 member (UTM: 14U 553590 mE, 6149850 mN)

Overlying the P3 dolomite is a succession of iron formation, chert, and calcsilicate horizons. A thinly layered horizon of silicate-facies iron formation is overlain by a graded chert horizon that contains coarse fragments of chert at the base grading into fine fragments toward the top. The chert layer is overlain by a horizon of thickly layered iron formation. Above this lies a layer of carbonate-facies iron formation characterized by thick, brownish iron carbonate-rich bands, and a calcsilicate layer that can be used as a stratigraphic marker. Fragmented and massive chert horizons occur near the top of the P3 member.

Stop 1-25: Pipe formation–Setting formation contact (UTM: 14U 553706 mE, 6150060 mN)

The massive chert at this stop occurs near the top of the P3 member of the Pipe formation. The massive chert is overlain by a horizon of rusty chert and a thin, garnet-rich silicate-facies iron formation. The iron formation represents the top of the Pipe formation and is in relatively sharp contact with the overlying pelitic schist of the Setting formation. The Setting formation pelite is characterized by ubiquitous sillimanite knots which give the rock a speckled texture.

Stop 1-26: Setting formation (UTM: 14U 553826 mE, 6150062 mN)

At this stop we can see interlayered pelitic schist and quartzite of the Setting formation. This unit has been interpreted as turbidite deposits. Local concretions consist of quartz, clinozoisite, garnet, and carbonate, and are characteristic of the Setting formation. The transition from dominantly chemical and fine siliciclastic sedimentary rocks of the Pipe formation to the coarser siliciclastic rocks of the Setting formation could indicate the onset of active tectonism and shallowing of the sedimentary basin, or deposition of axial zone sediments in a foredeep basin.

Directions to optional stop: Bah Lake assemblage, pillowed basalt flow

0.0 km Intersection of Pipe mine road and PTH 6, turn north towards Thompson.

2.7 km Pull onto the right shoulder of the highway near the crest of a hill, cross the ditch on foot and walk eastward roughly 60 m through the gap in the tree line, continue into a clay quarry, turning south, and follow the east edge of the tree line for another 60 m.

Optional stop: Bah Lake assemblage, pillowed basalt flow (UTM: 14U 555810 mE, 6150855 mN)

The clay quarry contains isolated outcrops of Bah Lake assemblage rocks, including a pillowed basalt flow along the east tree line. The pillows are elongate suggesting substantial vertical stretching; however, way-up indicators are preserved and suggest tops are toward the south. The trace-element chemistry of the Bah Lake assemblage is MORB-like, indicated renewed extension along the Superior craton margin, or possibly a magmatically active foredeep. Exposures of Setting formation, S1 member quartzite and pelite occur farther along the quarry. The contact between the Setting formation and Bah Lake assemblage rocks appears to be sharp and indicative of a stratigraphic contact.

Day 2 stop descriptions: drillcore and Thompson mine underground

Day 2 will concentrate on mine stratigraphy and the ore environment. It will be split between core viewing and an underground tour at the Thompson mine led by Vale geologists.

Directions to optional stop: Burntwood group-like turbidite

0.0 km Intersection of Thompson Drive North and Provincial Road 391 (last traffic light heading north out of Thompson), drive north past the Thompson airport on Provincial Road 391.

18.2 km Turn right, crossing the ditch into a clay quarry (may require four wheel-drive vehicle depending on conditions). Follow the first right and park on the northeast side of the outcrop with the geodetic survey pillar. Proceed on foot through the tree line along the southeast side of the outcrop to additional outcrop area overlooking low ground (approximately 20 m).

Optional stop: Burntwood group-like turbidite (UTM: 14U 561225 mE, 6189435 mN)

This Burntwood group-like sequence consists of greywacke and pelite interbedded on a mm- to m-scale. The greywacke commonly occurs as lenses or boudins and can be internally layered. The more pelitic layers contain abundant garnet, sillimanite, and cordierite. Several varieties of pegmatitic segregations are present and range from conformable to discordant. Peritectic or restitic cordierite and garnet are present in the pegmatitic segregations.

Burntwood group rocks are in faulted contact with Pipe formation, P3 member rocks in the Mel zone of the northern TNB. Distinguishing between the potentially Ni-bearing pelites of the Pipe formation and the non-prospective pelites of the Burntwood group can be challenging. Cordierite is a characteristic amphibolite-facies mineral of Burntwood group pelite because of the rock's relatively low K content. Cordierite does not generally occur in amphibolite-facies pelite of the Ospwagan group, which contains higher concentrations of K. Unfortunately, much of the Mel zone is underlain by granulite-facies rocks, and both Ospwagan and Burntwood group pelites can contain cordierite (Couëslan and Pattison, 2012). Bulk-rock trace-element and Sm-Nd isotope geochemistry can be helpful for differentiating these two units (Böhm et al., 2007; Zwanzig et al., 2007); however, this outcrop demonstrates the difficulties that can arise when working in this area. These rocks yield a crustal residence model age of ca. 3.05 Ga, which is considered older than typical for Burntwood group rocks (ca. 2.15–2.62 Ga) and within the range for Ospwagan group rocks (ca. 2.82–3.22 Ga; Böhm et al., 2007); however, the bulk-rock major-element geochemistry of pelite in this outcrop is more similar to the low-K pelites of the Burntwood group than Ospwagan group rocks (Couëslan, 2013). It is possible that trace-element geochemistry could provide more clarity as to the affinity of these rocks.

Day 3 stop descriptions: Ospwagan Lake

Day 3 consists of an excursion on Ospwagan Lake, where portions of each of the Ospwagan group formations can be observed, including extensive outcrops of Bah Lake assemblage volcanic rocks. Ultramafic intrusions may be exposed on low-lying reefs during low-water years.

Geology of the Ospwagan Lake area

Nickel mineralization was first discovered on Lower Ospwagan Lake in 1929, making it one of the earliest Ni discoveries in the TNB (Fraser, 1985). Much of the lake is underlain by Ospwagan group rocks, part of a continuous belt that stretches from Pipe mine to Moak Lake as a synformal structure (Figure 2, Figure 13; Bleeker, 1990; Burnham et al., 2009). The regional synform is cored by Ospwagan group rocks and bounded by Archean gneiss. Bedding and foliations are typically steeply southeast dipping or near vertical. A basement inlier is present on Upper Ospwagan Lake, and likely represents a sheath-like F_3 dome that pierces through the synform. Ospwagan Lake hosts many of the best exposures of Ospwagan group rocks outside of the mine environment. The most continuous sequence occurs at Niven Point, where a section from the lower Pipe formation to Bah Lake assemblage is exposed. The supracrustal package at Ospwagan Lake is host to several large boudinaged ultramafic bodies. The largest ultramafic body might once have formed part of a continuous sill stretching from the Pipe II to Birchtree mines; however, no significant Ni-Cu mineralization has been discovered. Exposures on Ospwagan Lake include some of the lowest metamorphic grade rocks in the TNB with pressure-temperature estimates of 3.2–4.1 kbar and 550–585°C (Couëslan and Pattison, 2012).

Directions to stops 3-1 to 3-13: Ospwagan Lake (UTM: 14U 562150 mE, 6158175 mN)

0.0 km Head southwest out of Thompson on PTH 6 from Burntwood Road (last traffic light heading south out of Thompson).

20.9 km Turn right onto access road to Ospwagan Lake, continue on road to boat launch (approximately 300 m).

Stop 3-1: Archean gneiss–Manasan formation unconformity (UTM: 14U 559015 mE, 6159515 mN)

This outcrop includes the unconformity between the Archean basement gneiss and Manasan formation rocks. Although sheared, the contrast between the heterogeneous beige Archean gneiss and more monotonous dark Manasan formation M1 member is apparent. Shearing along the contact has resulted in an apparently conformable relationship between the Archean and Manasan formation rocks. Quartzofeldspathic pebbles are abundant in the M1 member.

Optional stop: Manasan formation, M2 member (UTM: 14U 559040 mE, 6159575 mN)

This outcrop contains strongly sheared Manasan formation M2 semipelite with quartzofeldspathic segregations. The semipelite is muscovite and biotite rich, and appears relatively siliceous. Rubbly-slatey material at the south end of the outcrop could represent the sheared top of the M1 member. Thompson formation rocks underlie the lake to the east.

Stop 3-2: Pipe formation, P1 member iron formation and P2 member pelite (UTM: 14U 558415 mE, 6158195 mN)

The P1 member iron formation is fine grained and sheared; however, it is strongly magnetic and chert layers are still visible. Relatively siliceous P2 is present at the north end of the island. Layers of two-colored chert are present, similar to the P2 at Pipe mine. A large ultramafic body underlies the lake to the north and west of this island and occurs at the same stratigraphic level as the ultramafic bodies at the Pipe and Birchtree mines. No significant mineralization has been reported.

Stop 3-3: Pipe formation, P2 member pelite (UTM: 14U 558380 mE, 6158140 mN)

These outcrops of P2 member pelite are sheared and retrogressed with local kink-bands. Quartz-rich segregations have been described as chert lenses; however, some of these segregations appear to be discordant to the sedimentary layering and likely represent quartz veins. Local pseudomorphs of staurolite and/or andalusite may be present.

Stop 3-4: Pipe formation, P3 member rocks – Niven Point (UTM: 14U 558455 mE, 6158065 mN)

A sequence of Ospwagan group rocks from the Pipe formation to the Bah Lake assemblage are exposed along Niven Point. The Pipe formation rocks consist of strongly magnetic and thickly bedded iron formations, and dolomitic marble, both

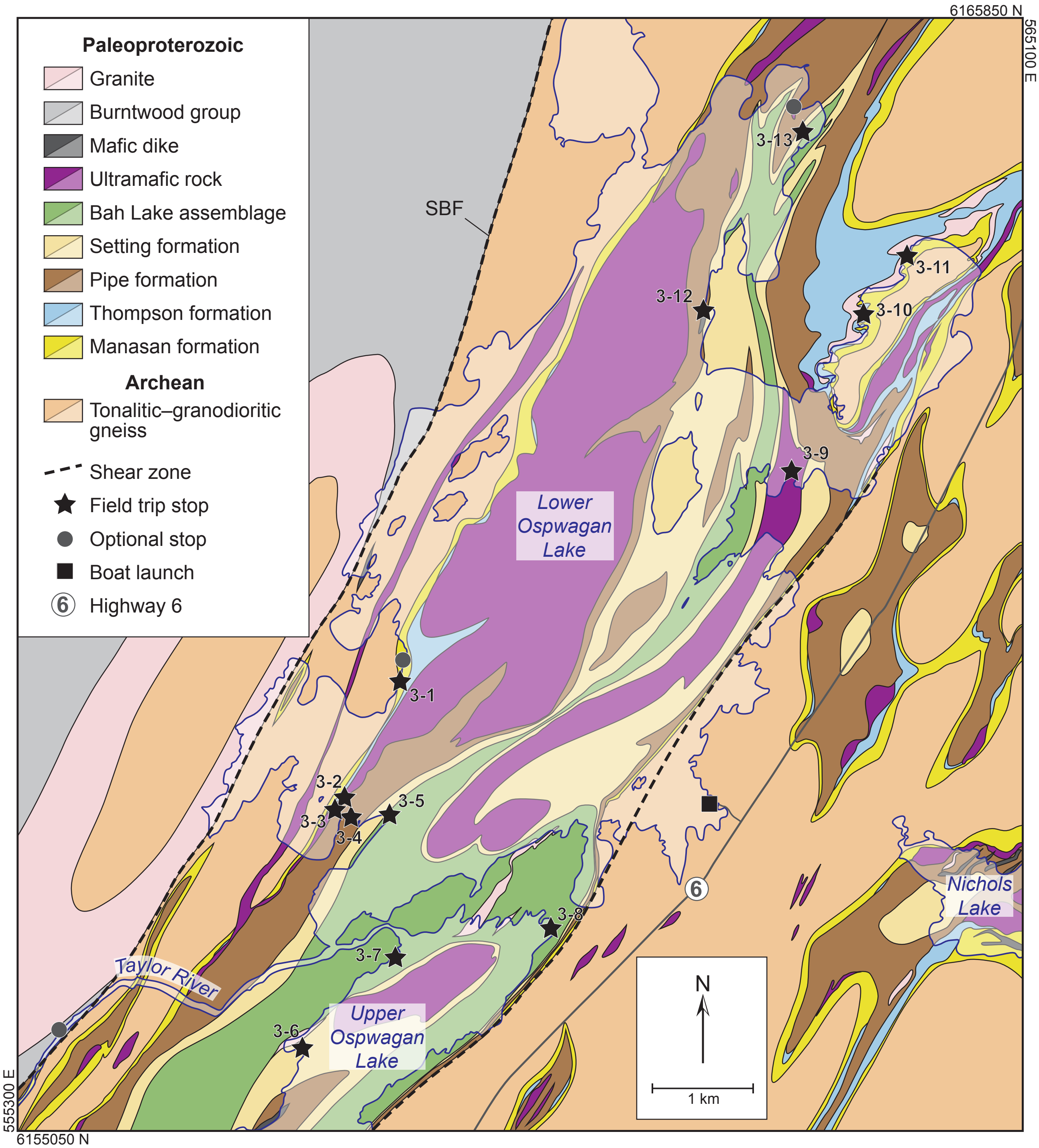

***Figure 13**: Geology of the Ospwagan Lake area (adapted from Macek et al., 2006). Abbreviations: SBF, Superior boundary fault.*

assigned to the P3 member. Rectangular garnet porphyroblasts are locally present within the iron formation. The regionally discontinuous marble horizon is also present at the Pipe II and Birchtree mines.

Stop 3-5: Setting formation and Bah Lake assemblage rocks – Niven Point (UTM: 14U 558885 mE, 6158085 mN)

The majority of this portion of Niven Point is underlain by Setting formation rocks, with Bah Lake assemblage basalt occurring toward the east end of the point. The Setting formation rocks consist of interbedded arenite, wacke, and pelite. Local pebbly beds are present. The Bah Lake assemblage basalt consists of either flow breccia, or sheared pillowed flows with local bedded tuff. Sparse lenses of garnet schist may represent interflow mafic pelite or possibly minor zones of hydrothermal alteration.

Stop 3-6: Setting formation – Upper Ospwagan Lake (UTM: 14U 558064 mE, 6155801 mN)

This outcrop contains Setting formation rocks consisting of m-scale beds of arenite with minor pelitic beds <15 cm thick. The abundance of arenite/quartzite at this location is more similar to Setting formation rocks at Mystery Lake and the Thompson 1C open pit, as opposed to the wacke-rich rocks at the Pipe II mine. The characteristic pink concretions of the Setting formation are also present.

Stop 3-7: Bah Lake assemblage, pillow basalt – Upper Ospwagan Lake (UTM: 14U 558985 mE, 6156750 mN)

This shoreline consists of Bah Lake assemblage basalt with well preserved pillows. Pillow structures indicate tops to the west. Minor subvolcanic intrusions are also present along this shoreline. The ca. 1890 Ma Favel Island complex on Setting Lake is interpreted to be in intrusive contact with the Bah Lake assemblage, suggesting that these volcanic rocks are older than the geochemically similar Molson dike magmatism.

Stop 3-8: Bah Lake assemblage basalt and 'picrite' – Upper Ospwagan Lake (UTM: 14U 560470 mE, 6156985 mN)

Massive flows of Bah Lake assemblage basalt and 'porphyroblastic picrite' are exposed along this narrow point. The 'picrite' flows are recognized by their pock-marked weathered surfaces and strongly magnetic property. The pock-marked surface is caused by the preferential weathering of serpentinized olivine porphyroblasts. Bulk rock geochemistry suggests these rocks are komatiites rather than 'picrites'. Although these flows represent ultramafic volcanic rocks, the Bah Lake assemblage is interpreted to be older (see Stop 3-7) and is isotopically and geochemically distinct from the deposit-forming ultramafic bodies of the Thompson nickel belt. A 30 cm thick layer of plagioclase-garnet stringers occurs within the basalt and may represent a flow top, or sheared pillowed horizon.

Optional stop: Burntwood group – Taylor River (UTM: 14U 555700 mE, 6155995 mN)

This exposure of Kisseynew domain rocks consists of interlayered pelite and wacke of the Burntwood group. Semiconformable leucosome forms 30–40% of the outcrop. Coarse-grained, pinitized cordierite porphyroblasts occur within the pelite and leucosome. Cordierite associated with the leucosome can be up to 10 cm across. Strong retrogression of the metamorphic assemblages is likely a function of close proximity to the Superior boundary fault.

Stop 3-9: Peridotite (UTM: 14U 562850 mE, 6161445 mN)

Serpentinized peridotite is present along this point. A small reef of peridotite is also exposed during periods of low water levels approximately 90 m north of the point. The reef is characterized by well preserved cumulate textures. This ultramafic body occurs relatively high in the Ospwagan group sequence and is hosted by Setting formation rocks.

Stop 3-10: Thompson formation calcsilicate (UTM: 14U 563535 mE, 6162955 mN)

Calcsilicate of the Thompson formation occurs as a laminated amphibole schist. While some laminations appear siliceous, the presence of carbonate is suggested by the vuggy weathered surface, and local laminations appear to contain up to 40% diopside. Small disrupted pegmatite segregations are common and contain coarse-grained quartz, feldspar, and green amphibole.

Stop 3-11: Pegmatite with xenolith of Thompson formation (UTM: 14U 563990 mE, 6163585 mN)

As noted at Pipe pit, pegmatitic intrusions are commonly associated with the Manasan formation M2 member semipelite, either as injections or locally derived melts. This semi-conformably pegmatite is several km long (stretching from the island at the entrance to this bay, to the north and east of this location) and has intruded along the contact between the M2 member and the Thompson formation. The M2–Thompson formation contact is preserved in a xenolith that is 10s of metres across.

Stop 3-12: Pipe formation, P2 member pelite (UTM: 14U 562015 mE, 6163045 mN)

These rocks represent some of the lowest metamorphic grade rocks exposed in the TNB. Fine-grained staurolite can be found at the end of the point, while coarse-grained andalusite occurs within quartz veins farther into the bay. Andalusite is not found in the groundmass of the pelite, suggesting it is unstable in the presence of muscovite.

Stop 3-13: Bah Lake assemblage mafic tuff (UTM: 14U 562965 mE, 6164810 mN)

Although the Bah Lake assemblage consists dominantly of massive to pillowed basalt flows, well-laminated mafic tuff is well exposed along the west side of this island.

Optional stop: Ultramafic intruded by Molson dike (UTM: 14U 562895 mE, 6165000 mN)

This low-lying reef is exposed during periods of low water. It consists of ultramafic rock intruded by a Molson dike. The intrusions are interpreted to be coeval and possibly co-genetic.

Directions to optional stops: Manasan formation semipelite and Thompson formation marble at Paint Lake marina

0.0 km Intersection of Ospwagan Lake access road and PTH 6, turn south toward Paint Lake.

7.7 km Turn left onto Provincial Road 375 which leads to the Paint Lake marina and campground. You are now entering a provincial park and a vehicle park pass may be required.

13.5 km Park vehicle in the marina parking lot.

Optional stop: Manasan formation, M2 member semipelite (UTM: 14U 561480 mE, 6150060 mN)

Manasan formation, M2 member semipelite is exposed in a road-cut along PR 375 approximately 60 m northwest of the marina parking lot. The semipelite contains abundant leucosome, and sparse garnet. The intensity of partial melting indicates significantly higher metamorphic grades than observed at either Pipe mine or Ospwagan Lake. Discontinuous, boudinaged layers of Thompson formation calcsilicate indicate stratigraphic tops to the southeast.

Optional stop: Thompson formation marble (UTM: 14U 561430 mE, 6149920 mN)

Thompson formation marble is exposed along the shoreline, south of the Paint Lake lodge. The marble is sheared and contains serpentinized olivine, tremolite, phlogopite, and chlorite.

Day 4 stop descriptions: Paint Lake

Day 4 consists of an excursion on Paint Lake. Paint Lake is the type locality for the Paint sequence rocks, a succession of dominantly arkosic wacke with subordinate arkosic arenite, and minor iron formation and pelite. Much of the Paint Lake area was subjected to granulite-facies metamorphic conditions. Many of the units observed here are also present in the Phillips Lake area to the south.

Geology of the Paint Lake area

The Paint Lake area was once considered to represent a basement dome of high-grade metamorphic rocks, possibly even preserved granulites of the Pikwitonei domain (Paktunç and Baer, 1986; Russell, 1981). Although the Paint Lake area is underlain by dominantly Archean gneissic rocks, three northeast-striking belts of Paleoproterozoic metasedimentary rocks are present in the area and have been informally termed the Paint sequence (Figure 14; Couëslan, 2016). A single northeast-striking belt of Ospwagan group rocks is recognized along the western shore of Paint Lake. In addition, numerous Paleoproterozoic felsic, mafic, and ultramafic intrusions occur in the area including a swarm of thin carbonatite dikes, which occur through the central islands of Paint Lake over a strike length

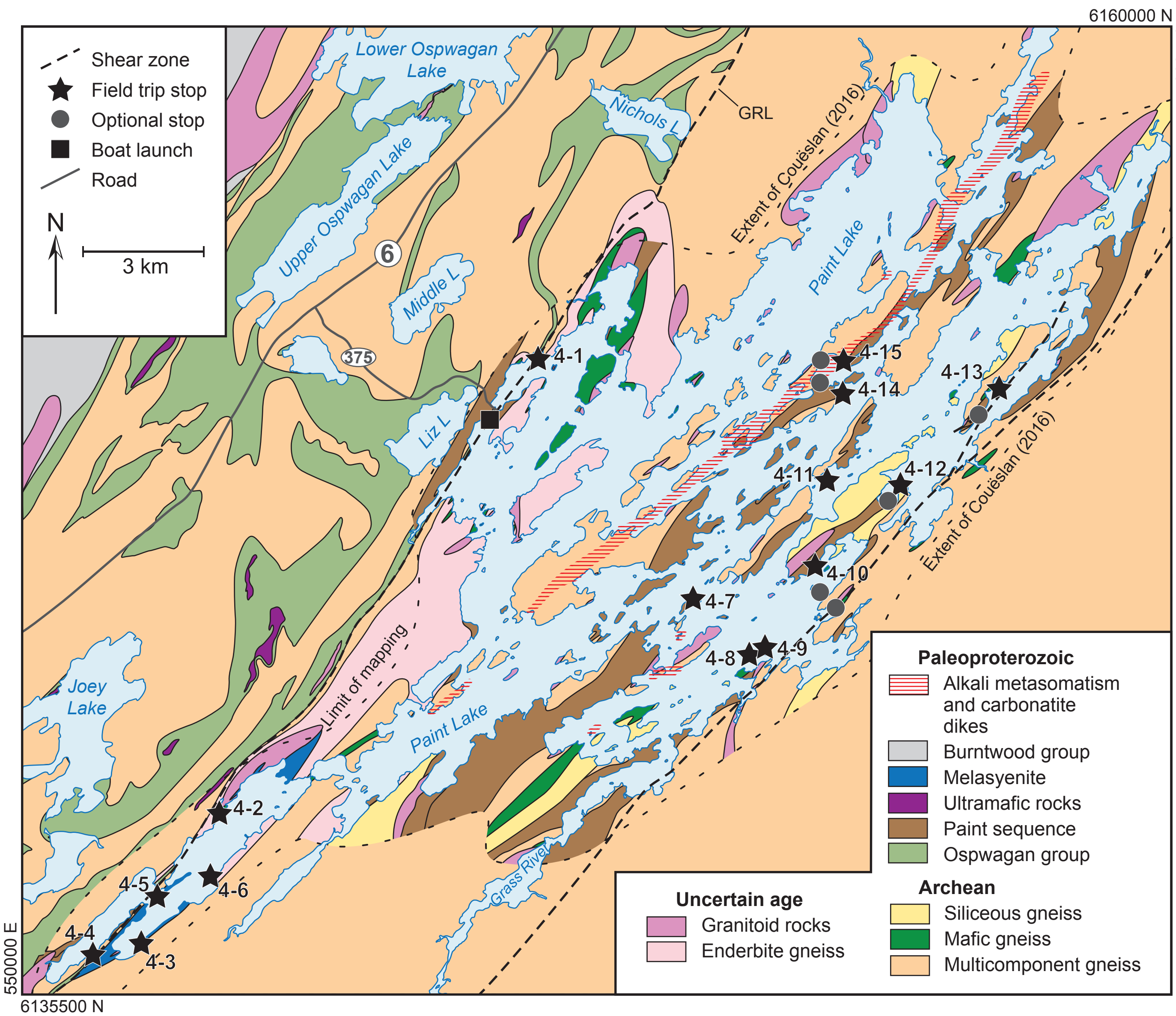

***Figure 14**: Geology of the Paint Lake area (modified from Macek et al., 2006; Couëslan, 2016). No attempt has been made to reconcile the geology of Macek et al. (2006) and Couëslan (2016). Abbreviations: GRL, Grass River lineament; L, Lake.*

of at least 23 km. Rocks in the area are typically high strain and minor isoclinal folds are abundant. The dominant regional structures in the Paint Lake area are upright, shallow to moderately plunging isoclinal F_3 folds. Regional trends for S_2 foliations and axial trends for F_3 folds are approximately 045°, but rotate from typical TNB trends of approximately 025° close to the marina, to approximately 075° in the southeast corner of the lake (Couëslan, 2009). Discrete D_3–D_4 mylonitic shear zones occur in several places trending subparallel to the regional structures. Mineral assemblages indicate lower granulite-facies metamorphism over most of the area (approximately 780–830°C and 6.5–7.0 kbar; Couëslan and Pattison, 2012) and are interpreted to be Hudsonian. A slightly lower metamorphic grade occurs along the western shore of Paint Lake in the vicinity of the Grass River lineament, which is characterized by upper amphibolite-facies metamorphic assemblages. Amphibolite-facies retrogression of granulite-facies assemblages is relatively common, and greenschist-facies retrogression is locally associated with the shear zones.

Directions to stops 4-1 to 4-14: Paint Lake (UTM: 14U 562150 mE, 6158175 mN)

0.0 km Head southwest out of Thompson on PTH 6 from Burntwood Road (last traffic light heading south out of Thompson).

28.6 km Turn left on PR 375 which leads to the Paint Lake marina and campground. You are now entering a provincial park and a vehicle park pass may be required.

39.3 km Turn right into the marina parking lot and boat launch area.

Stop 4-1: Archean gneiss–Ospwagan group unconformity (UTM: 14U 562570 mE, 6151340 mN)

The unconformity between the Archean gneiss and Manasan formation is exposed on a prominent reef in the bay north of the marina. The exposure indicates the sequence is younging to the east. Calcsilicate boudins in the M2 member indicate the transition into Thompson formation rocks.

Stop 4-2: Mafic gneiss (UTM: 14U 554865 mE, 6140255 mN)

Hornblende-rich, garnet-bearing mafic gneiss is common in the Paint Lake area. These rocks are assumed to be Archean and have traditionally been interpreted as part of a layered gabbro complex; however, cordierite-rich horizons along this shoreline suggest high-Mg, high-Al bulk compositions similar to the intense chlorite alteration associated with subseafloor hydrothermal activity (i.e., VMS-related alteration). This may suggest a volcanic origin for these rocks rather than intrusive. The rocks are geochemically similar to arc tholeiites.

Stop 4-3: High-Mg mafic gneiss (UTM: 14U 552885 mE, 6136945 mN)

Mafic gneiss at this outcrop is characterized by green amphibole, abundant orthopyroxene, and typically no garnet. Unlike the mafic gneiss at the previous stop, the bulk composition of this gneiss is similar to low-Ti tholeiite and boninite. These types of magmas are generally associated with rifted arc environments.

Stop 4-4: Manasan formation (UTM: 14U 551830 mE, 6136820 mN)

The contact between the Manasan formation, M1 member quartzite and M2 member semipelite is exposed on a small point. This succession from the M1 quartzite to the M2 semipelite indicates younging to the east, which is consistent with the younging direction of the Ospwagan group rocks near the marina, approximately 18 km along strike to the north-northeast.

The laminated quartzite is well preserved whereas the semipelite forms a diatexite. This outcrop demonstrates how certain units within the Ospwagan group (quartzite, iron formation, calcsilicate, and marble) are typically recognizable at all metamorphic grades, while units susceptible to partial melting (pelite and semipelite) become difficult to distinguish from the Archean gneiss. This gave rise to the concept of the Ospwagan group 'ghost succession' (Zwanzig et al., 2007), in which readily recognizable marker units are isolated by intervals of migmatitic gneiss.

Stop 4-5: Melasyenite (UTM: 14U 553335 mE, 6138195 mN)

This melasyenite is weakly peralkaline and meets the chemical criteria of an ultrapotassic rock as defined by Foley et al. (1987). It is characterized by dark brown to black K-feldspar phenocrysts <3 cm long, in a matrix of reddish-brown biotite, amphibole, and honey-brown K-feldspar. Dark brownish-green amphibole crystals <2 cm long are less common. Although geochemically enriched in LREE and LILE, it is also enriched in Ni and Cr, suggesting it is a primitive, mantle-derived melt. The age of this unit was determined at ca. 1883 Ma (Couëslan, 2016), contemporaneous with the intrusion of the Thompson-type ultramafic sills and Molson dikes. Similar rocks are known from the Max Lake area to the west.

Stop 4-6: Melasyenite and pyroxenite (UTM: 14U 554580 mE, 6138705 mN)

This exposure is more typical of the melasyenite, which has in most places been subjected to some degree of metasomatism and shearing. The K-feldspar are usually pink to brick-red, and rounded. Dikes of alkali granite are common. The melasyenite is in contact to the east with a pyroxenite intrusion. The pyroxenite is interpreted to be related to the Thompson-type ultramafic intrusions.

Stop 4-7: Paint sequence wacke (UTM: 14U 566435 mE, 6145445 mN)

The Paint sequence consists dominantly of arkosic wacke. In this outcrop the wacke is interbedded with arenite and local, discontinuous horizons of sulphide-bearing iron formation. The wacke is characterized by weak but pervasive gossan staining suggesting the presence of disseminated sulphide. Rare lenses of molybdenite are present; however, it is not associated with any visible alteration. Some horizons were boudinaged during D_2–D_3, and the boudin necks were filled by orthopyroxene-bearing leucosome. This demonstrates that the rocks attained granulite-facies metamorphic conditions during the Trans-Hudson orogeny. Similar rocks at southern Phillips Lake host a mineralized ultramafic body, suggesting the Paint sequence could potentially be host to Ni deposits (Couëslan, 2018).

Stop 4-8: Siliceous gneiss (UTM: 14U 567855 mE, 6144045 mN)

The siliceous gneiss at this outcrop is assumed to be Archean and its protolith is uncertain. It is common to both the Paint Lake and Phillips Lake areas. Relatively homogeneous textures at many locations suggested a plutonic protolith; however, it can also be well banded and local patches can be extremely quartz-rich. Depending on whether the banding represents primary layering or metamorphic segregations, it could suggest a sedimentary or possibly volcanic protolith.

Stop 4-9: Mafic gneiss (UTM: 14U 568160 mE, 6144190 mN)

This outcrop is similar to the rocks observed at stop 4-2. At this locality it consists of layered mafic gneiss, typically garnet-bearing, which grades toward the east into a layered leucocratic gneiss. The unit was previously interpreted as a metamorphosed mafic layered intrusion that grades from gabbro to anorthositic gabbro; however, the leucocratic gneiss is typically siliceous and can contain >40% quartz. This could imply that the rocks represent an Archean volcanic sequence that grades upward into either siliceous metavolcanic or metapsammitic rocks. Pegmatite dikes on the long northwestern point contain local accumulations of brown, metamict allanite.

Optional stop: Mylonite (UTM: 14U 569930 mE, 6145255 mN)

The mafic gneiss at this location is cut by northeast-trending mylonitic shear zones up to 3 m thick. The shear zones are likely related to sinistral transpression during D_3–D_4 deformation. The shear zones typically formed along the limbs of regional-scale, subvertical isoclinal F_3 folds. Movement along these shear zones appears to have exerted a first-order control on the distribution of metamorphic zones in the TNB.

Optional stop: Calcareous arenite (UTM: 14U 569490 mE, 6145595 mN)

Rafts of calcareous arenite occur within granodiorite at this outcrop. The arenite consists of alternating quartzofeldspathic and diopside-rich laminations. Its affinity is uncertain. Paint sequence arenite occurs along the shore to the west; however, calcareous sections are unknown. Similar calcareous arenite occurs in the Archean gneisses on Natawahunan Lake.

Stop 4-10: Paint sequence arenite (UTM: 14U 569520 mE, 6146335 mN)

Although less abundant than wacke, arkosic arenite is a significant component of the Paint sequence. Here, the arenite is interbedded with pelite and contains local garnet-bearing concretions. Both of these features make it similar to the Setting formation of the Ospwagan group. The pelite is characterized by a light gossan stain suggesting the presence of minor sulphide.

Stop 4-11: Paint sequence wacke (UTM: 14U 569650 mE, 6148365 mN)

This outcrop contains Paint sequence wacke with lenses of oxide-facies iron formation <15 cm thick. The wacke also contains boudinaged bands of amphibolite <50 cm thick, which are interpreted to be related to the ca. 1880 Ma Molson dike swarm.

Stop 4-12: Paint sequence pelite (UTM: 14U 571495 mE, 6148245 mN)

Pelite is a relatively minor component of the Paint sequence and is typically found interbedded with arenite. The pelite occurs as beds <20 cm thick and contains abundant garnet and sillimanite. Graphite is also present along with pyrrhotite resulting in a light gossan stain.

Optional stop: Paint sequence iron formation (UTM: 14U 571205 mE, 6147850 mN)

Paint sequence iron formation typically occurs as minor discontinuous layers in the wacke; however, it can form the dominant component of some outcrops. At this stop, Paint sequence iron formation is interbedded with pelite and arenite. The silicate-facies iron formation contains abundant orthopyroxene and garnet, along with thin chert laminations. The arenite forms beds <1 m thick with local intercalations of pelite.

Stop 4-13: Pelite of uncertain affinity (UTM: 14U 573880 mE, 6150630 mN)

This pelitic diatexite is characterized by very coarse-grained garnet, sillimanite, and cordierite. The affinity of the pelite is unknown, but it is texturally and mineralogically distinct from the Paint sequence pelite at stops 4-10 and 4-12. The prismatic habit and dark core of the sillimanite are reminiscent of andalusite (var. chiastolite), suggesting they may be pseudomorphous.

Optional stop: Garnet-rich mylonite (UTM: 14U 573350 mE, 6149850 mN)

This steep shoreline is underlain by a garnet-rich mylonitic shear zone. The garnet is coarse to very coarse grained and gives the weathered rock a knobby appearance. The mylonite is likely derived from pelitic rock similar to stop 4-13, and appears to be a splay of the shear zone exposed northeast of stop 4-9.

Stop 4-14: Dolomite carbonatite (UTM: 14U 570055 mE, 6150455 mN)

Carbonatite dikes are present through the central islands of Paint Lake as a swarm with a strike length of at least 23 km (Chakhmouradian et al., 2009; Couëslan, 2016). The individual dikes are typically <1 m wide and occur in a zone up to 500 m wide that is characterized by pervasive Na-Ca metasomatism. The metasomatized country rock is characterized by a bleached appearance. The dikes are commonly foliated and the swarm appears to parallel regional D_3–D_4 structures. Two varieties of carbonatite are present: a more primitive, grey dolomite carbonatite; and a more evolved, pink calcite carbonatite. The carbonatites intrude both Archean gneiss and Paint sequence wacke, and crosscut all but the latest, discordant and massive pegmatite dikes. Preliminary bulk-rock and phlogopite Rb-Sr isotope geochemistry, and U-Pb apatite geochronology suggest an age of ca. 1703 Ma for the carbonatite magmatism (Chakhmouradian, unpublished data, 2018). Similar dikes may also be present ~120 km along strike at Split Lake (Macdonald et al., 2017).

This shoreline exposes discontinuous dikes of dolomite carbonatite <60 cm thick that appear black on the weathered surface and light grey on fresh surfaces. The dolomite carbonatite contains serpentinized olivine, magnesiohornblende, magnetite, phlogopite, and local apatite. The dikes commonly have selvages of magnesiohornblende and/or diopside. The abundance of magnetite in the dolomite carbonatite makes it easy to differentiate from the non-magnetic, dolomitic marbles of the Ospwagan group.

Stop 4-15: Calcite carbonatite (UTM: 14U 570000 mE, 6151315 mN)

The calcite carbonatite exposed along this point is characterized by an earthy yellow-orange coating on weathered surfaces, and is pink on fresh surfaces. It contains magnesiohornblende and apatite, with minor magnetite, sulphide, and scapolite. The dikes are lined by selvages of amphibole and/or clinopyroxene±scapolite. At least two ages of pegmatite are present in this outcrop, with the youngest, relatively massive dike cross-cutting the carbonatite.

Optional stop: Calcite carbonatite (UTM: 14U 569570 mE, 6151240 mN)

Calcite carbonatite occurs along this shoreline as a series of dikes up to 1.5 m wide. The widest is well exposed along a steep rock face immediately south of the point. Pods of apatite-bearing clinopyroxenite are associated with the carbonatite.

Optional stop: Dolomite carbonatite (UTM: 14U 569590 mE, 6150755 mN)

Dolomite carbonatite is exposed across this shallow point during low water periods. The carbonatite occurs as dikes <2 m wide within metasomatized pegmatite.

Day 5 stop descriptions: Phillips Lake core and Setting Lake excursion

Day 5 consists of viewing core from southern Phillips Lake and an excursion on Setting Lake. The Phillips Lake core was drilled by Falconbridge Ltd. (now Glencore Canada Corporation) in 1996 and intersects some of the units observed during day 4 on Paint Lake, as well as a mineralized ultramafic body. The core is stored in Wabowden at the CaNickel Mining Limited core storage facility. Setting Lake is situated along the boundary between the TNB to the east and the Kisseynew domain of the Reindeer zone to the west, and provides an opportunity to see both Kisseynew domain and TNB rocks.

Geology of the Phillips Lake area

Phillips Lake is located 60 km south-southwest of Thompson in the eastern TNB (Figure 2). Inco Ltd. conducted airborne and ground geophysical surveys, diamond-drilling and outcrop mapping in the southern Phillips Lake area from 1952 to 1975 (Assessment File 92118, Manitoba Growth, Enterprise and Trade, Winnipeg). Their work led to the discovery of one large, and several smaller, ultramafic bodies in the area. Drillcore logging and mapping also resulted in the recognition of metasedimentary rocks, including 'skarn' and iron formation, that were restricted to narrow bands in the enclosing gneiss. Later work by Falconbridge Ltd. from 1980 to 1996 (Assessment Files 94497, 94506) delineated an ultramafic body with >1800 m strike length and significant Ni-mineralization along the footwall contact (Figure 15). Falconbridge geologists interpreted

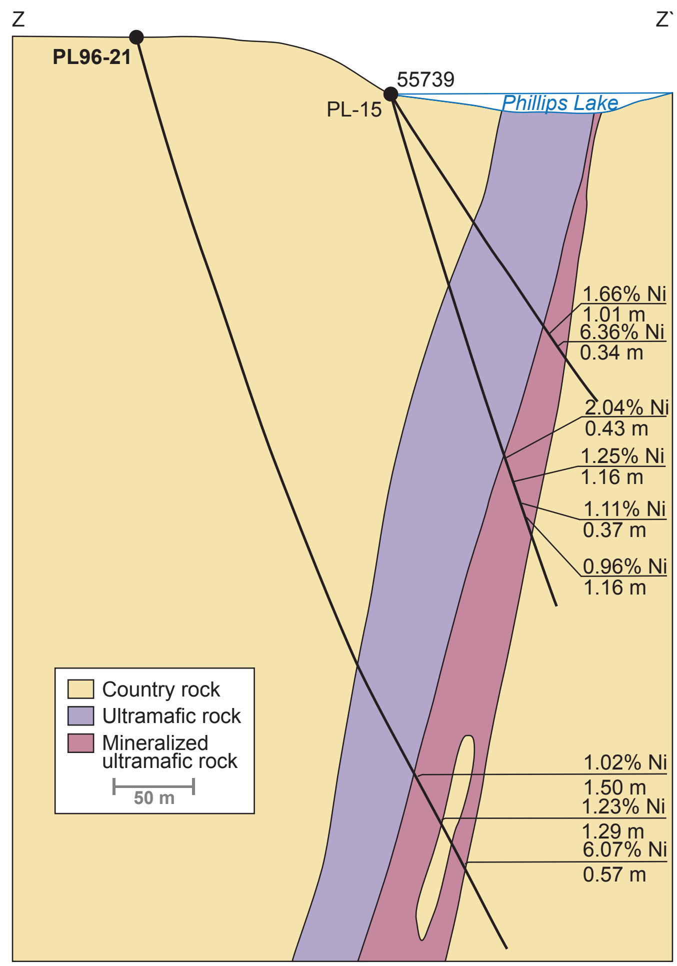

***Figure 15**: Cross-section Z–Z`, looking north through Falconbridge Ltd. drillholes PL-15 and PL96-21, and Inco Ltd. drillhole 55739 (modified from Assessment File 94497). The locations of drillholes and section Z–Z` are shown in Figure 16.*

the ultramafic body to be hosted in Archean gneiss; however, regional compilation mapping by the Manitoba Geological Survey (MGS) suggested that the mineralized ultramafic body was hosted by a 'ghost succession' of the Ospwagan group (Macek et al., 2006; McGregor et al., 2006; Zwanzig et al., 2007). Most recently, shoreline mapping in 2012 and relogging of drillcore in 2018 suggest that Paint sequence rocks host the mineralized ultramafic intrusion (Figure 16; Couëslan, 2016, 2018).

The Phillips Lake area is underlain by a highly attenuated basement dome (Burnham et al., 2009) consisting largely of Archean multicomponent gneiss, which comprises a mixture of tonalitic and granodioritic gneisses in varying proportions, with subordinate mafic and ultramafic phases (Macek et al., 2006; Couëslan, 2016). Other, more discrete Archean units in the area include a mafic gneiss suite, which could be derived from a layered gabbro complex (Macek, 1985) or possibly volcanic rocks (Couëslan, 2016); and a garnet- and magnetite-bearing siliceous gneiss unit. A thin sheet of Paint sequence rocks mantles the dome and is thickened in the fold closure at the south end of Phillips Lake (Couëslan, 2018), which is also the location of the mineralized ultramafic body. Rocks in the area attained granulite-facies metamorphic conditions during the Trans-Hudson orogeny, with orthopyroxene being a common constituent in rocks of felsic, intermediate and mafic bulk composition (Couëslan and Pattison, 2012; Couëslan, 2016).

Drillhole PL96-22 – Phillips Lake (collar UTM: 14U 542700 mE, 6121620 mN)

At the top of drillhole PL96-22 is a roughly 40 m thick interval of relatively monotonous biotite hornblende gneiss, which is of uncertain affinity (Figure 17). This is followed by a thick interval (300 m) of multicomponent gneiss. The multicomponent gneiss consists of a variety of phases including metaplutonic and metasedimentary phases, with zones of metasomatism/amphibolitization. Below the multicomponent gneiss is a relatively continuous, 95 m interval of wacke. The wacke contains local intercalations of iron formation and is likely correlative with the Paint sequence. The wacke is followed by a 15 m interval of calcsilicate and marble. The calcsilicate is crosscut by a 70 cm interval of sulphide breccia. The sulphide breccia likely represents remobilized sulphide along a fault/shear zone. The source of the sulphide and affinity of the calcsilicate is unknown. Below the calcsilicate is a 45 m interval of serpentinized peridotite. Relict cumulate texture is locally preserved with <2% interstitial sulphide. This is followed by a 45 m interval of wacke similar to that occurring above the ultramafic body. Below the wacke is a 140 m interval of mafic gneiss and highly disrupted wacke. The mafic gneiss is likely correlative with the high-Mg mafic gneiss at Paint Lake (stop 4-3). The wacke is disrupted by abundant leucosome and/or pegmatite injection. Iron formation associated with the disrupted wacke is well layered and texturally distinct from iron formation occurring in previous intervals.

Drillhole PL96-21 – Phillips Lake (collar UTM: 14U 542910 mE, 6121770 mN)

The core from drillhole PL96-21 presents a similar sequence as PL96-22 (Figure 17). A 55 m interval of biotite hornblende gneiss is followed by a 210 m interval of multicomponent gneiss. The multicomponent gneiss grades over 30 m into wacke that is likely correlative with the Paint sequence. The wacke forms a 105 m interval with local intercalations of iron formation. At the base of the wacke interval is a 1 m thick interval of sulphide breccia that likely correlates with the sulphide breccia in PL96-22. Below the wacke is a 135 m interval of serpentinized peridotite. The peridotite becomes increasingly sulphidic toward the base of the interval where a 60 cm interval of solid sulphide is present. Pentlandite mineralization is visible in the solid sulphide. The increasing sulphide content toward the structural footwall of the ultramafic body suggests that the stratigraphy in the drillcore remains upright. Toward the top of the peridotite interval is a 1.5 m band of biotite-tremolite gneiss, which may be correlative with the calcsilicate in PL96-22. Below the peridotite is a 50 m interval of wacke, which is similar to the interval present above the ultramafic unit. This is followed by 5 m of mafic gneiss at the end of the hole.

Interpretation and significance of drillholes PL96-21 and PL96-22

Mineralized ultramafic bodies in the TNB are typically associated with Ospwagan group rocks. The presence of mineralization in the peridotite at Phillips Lake is significant because the intrusion appears to be hosted in Paint sequence wacke (Figure 17). Weakly disseminated sulphide is ubiquitous in rocks of the Paint sequence and can be more abundant in iron formation and pelite horizons. This suggests that, in addition to the Ospwagan group, the Paint sequence is also capable of hosting mineralized ultramafic intrusions; however, the matter is complicated by the presence of the calcsilicate of uncertain affinity. The two most likely scenarios for the calcsilicate are that 1) it represents a tectonic sliver of Ospwagan group rocks spatially associated with the Paint sequence rocks, or 2) it is part of the Paint sequence. In the first scenario, the calcsilicate could be part of the Thompson formation, with the spatially associated sulphide breccia representing remobilized sulphide from the overlying P1 member sulphide-facies iron formation. In this scenario, the mineralized peridotite is in direct contact with Ospwagan group rocks at a similar stratigraphic level to that of the Pipe II and Birchtree ore bodies. The close spatial association of the Paint sequence rocks would in this case be coincidental and possibly of little consequence with respect to Ni mineralization. In the second scenario, the mineralized peridotite is hosted entirely in Paint sequence rocks, suggesting that ultramafic rocks hosted by the Paint sequence could be a new exploration target in the TNB.

Geology of the Setting Lake area

The Setting Lake area straddles the boundary between the Kisseynew domain and the Thompson nickel belt. This boundary is marked by the Setting Lake fault zone (SLFZ; Figure 18; Zwanzig, 1998; Burnham et al., 2009). Southeast of the SLFZ is the Soab structure, a homoclinal sequence of northwest-facing Ospwagan group rocks (Burnham et al., 2009). Much of this sequence lies under the lake with only the upper part of the sequence exposed on islands and reefs. This is the type locality

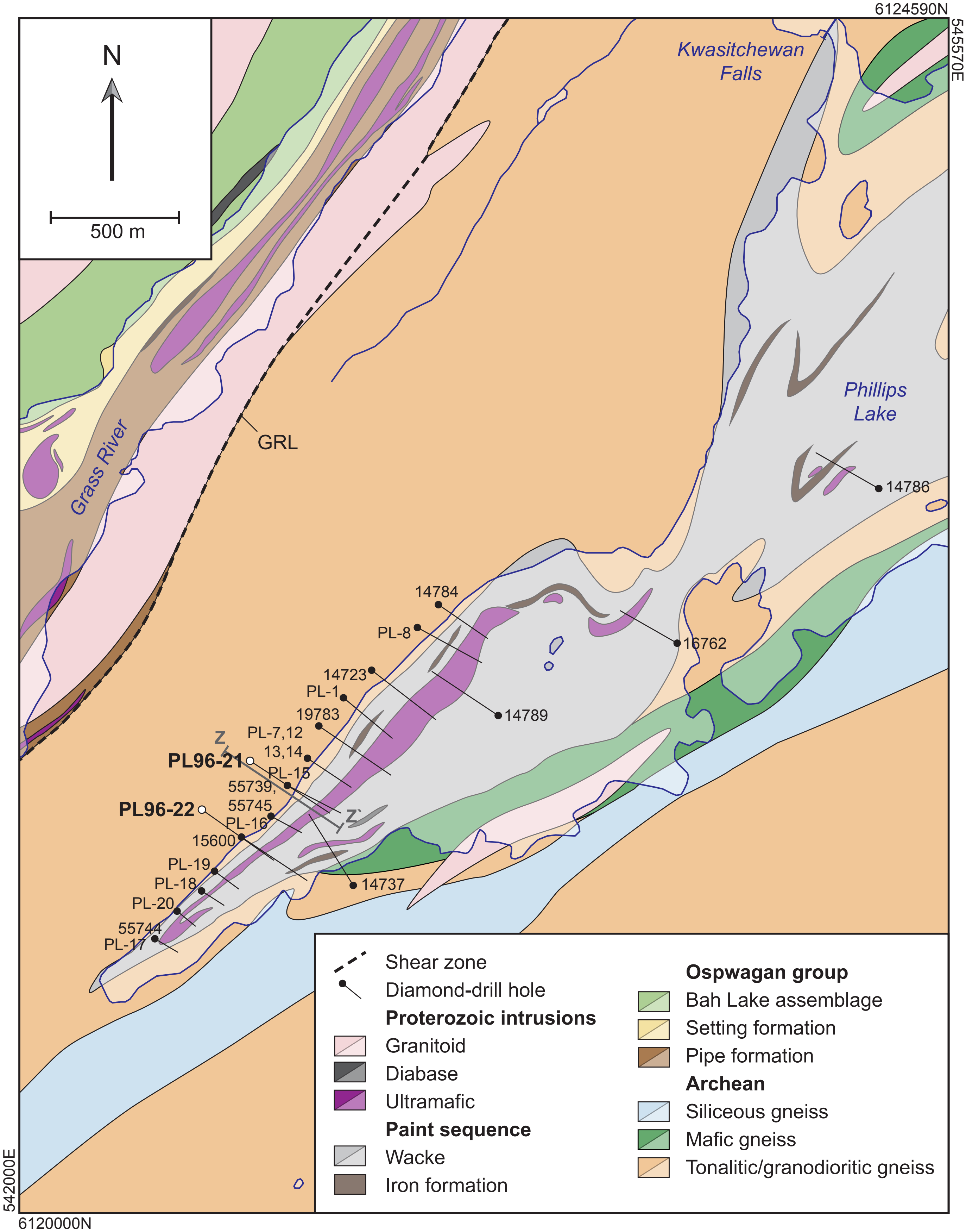

***Figure 16**: Geology of the southern Phillips Lake area, showing the locations of section Z–Z` (Figure 15), and diamond-drillholes by Inco Ltd. (five-digit numbers) and Falconbridge Ltd. (beginning with 'PL'; modified from Assessment File 94497; Macek et al., 2006; Couëslan, 2012). Diamond-drillholes completed by Callinan Mines Ltd. from 2005 to 2006 are not shown (Assessment Files 74292 and 74388). Abbreviations: GRL, Grass River lineament.*

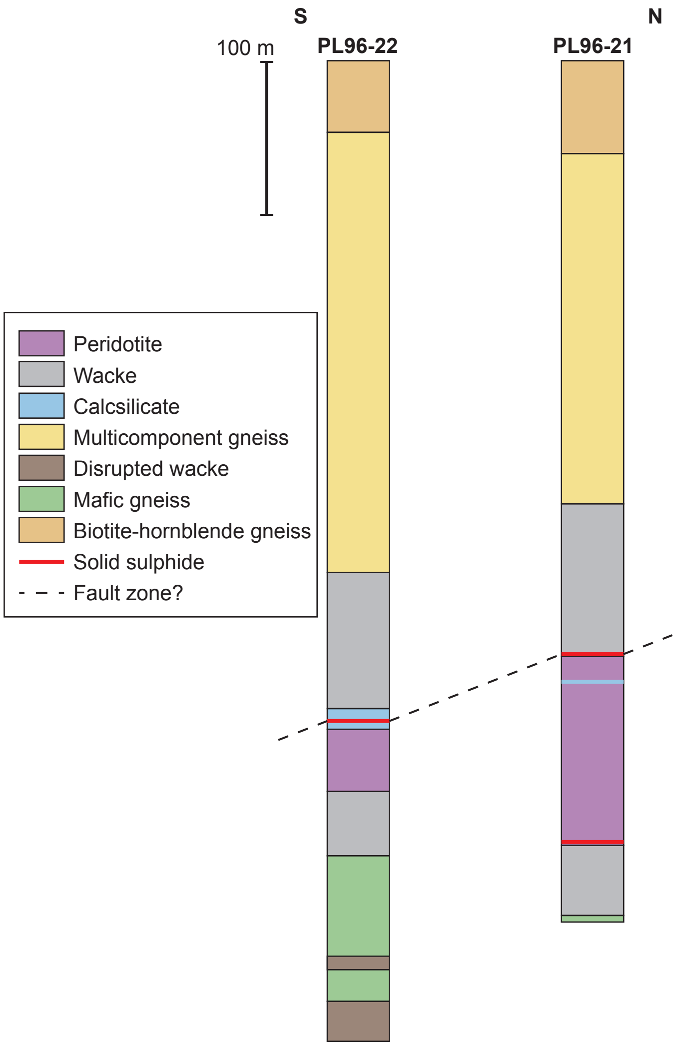

***Figure 17**: Schematic logs of drillcores PL96-21 and PL96-22 in the southern Phillips Lake area (Couëslan, 2018); see Figure 16 for drillhole locations.*

for the Setting formation which is locally well exposed, including the most extensive exposures of the S2 member. Most of the eastern shoreline of Setting Lake is underlain by Archean gneiss. The Soab structure is truncated to the northwest by the SLFZ. Northwest of the SLFZ is the Setting Lake antiform (Burnham et al., 2009). The doubly plunging antiform is cored by the ca. 1891–1878 Ma Favel Island complex, a gneissic quartz dioritic to granodioritic composite intrusion (Zwanzig et al., 2003; Percival et al., 2004). This intrusive complex is bounded by previously folded rocks of the Grass River group and Bah Lake volcanic assemblage. The southeast limb of the Setting Lake antiform is truncated by the SLFZ. The SLFZ manifests as greenschist-facies retrogression within mylonitic foliation to structureless cataclasite with abundant small faults and affects a zone up to 400 m wide (Burnham et al., 2009). It forms much of the contact between the Grass River group and Bah Lake volcanic assemblage, and is considered to be a segment of the Superior boundary fault (Figure 2), which separates the Superior craton (TNB) from the Reindeer zone (Kisseynew domain).

Directions to stops 5-1 to 5-8: Setting Lake (UTM: 14U 530640 mE, 6100650 mN)

0.0 km Head southwest out of Thompson on PTH 6 from Burntwood Road (last traffic light heading south out of Thompson).

89.6 km Turn right onto access road for the Setting Lake boat launch. The boat launch is at the end of the road (approximately 450 m).

Stop 5-1: Grass River group conglomerate (UTM: 14U 528370 mE, 6106720 mN)

This outcrop contains clast-supported, polymictic conglomerate at the base of the Grass River group. The majority of pebbles and cobbles are mafic, with subordinate felsic clasts. The arkosic matrix is dominated by feldspar and hornblende. The conglomerate was likely deposited in an alluvial fan environment along the margin of the Kisseynew basin. Detrital zircon from this unit are dominantly Paleoproterozoic with minor Archean grains.

Stop 5-2: Grass River group hornblende arkose (UTM: 14U 528165 mE, 6106650 mN)

The polymictic conglomerate is overlain by hornblende arkose (or lower sandstone), which is exposed along this point. Together the conglomerate and arkose represent a fining upward succession likely deposited in a fluvial-alluvial environment. The hornblende arkose is rather homogeneous with a crude layering suggested by variations in hornblende content. The unit can contain local pebbles and wedge-like lenses of conglomerate.

Stop 5-3: Grass River group biotite arkose (UTM: 14U 528025 mE, 6106855 mN)

The hornblende arkose grades into the overlying biotite arkose (or middle sandstone), which is exposed on this reef. The biotite arkose varies from crudely to well-bedded. It contains sparse quartz pebbles and local magnetite that can make this unit magnetic.

Stop 5-4: Grass River group sillimanite arkose with concretions (UTM: 14U 525810 mE, 6104165 mN)

This outcrop consists of sillimanite arkose (or upper sandstone), which is similar to the underlying biotite arkose, but contains sparse pebbles and beds with sillimanite knots. It also contains epidote-bearing 'concretions' or alteration zones. These zones appear to be concordant and are <20 cm thick. Discordant epidote veins are also present in this outcrop. This arkosic unit is muscovite-bearing at lower metamorphic grade. The middle and upper sandstone units (biotite arkose and sillimanite arkose) represent a coarsening upward, shallow-water succession.

Stop 5-5: Bah Lake assemblage–Grass River group unconformity? (UTM: 14U 525210 mE, 6097290 mN)

This reef exposes a sheared contact between amphibolite, interpreted as part of the Bah Lake assemblage, and the poly-

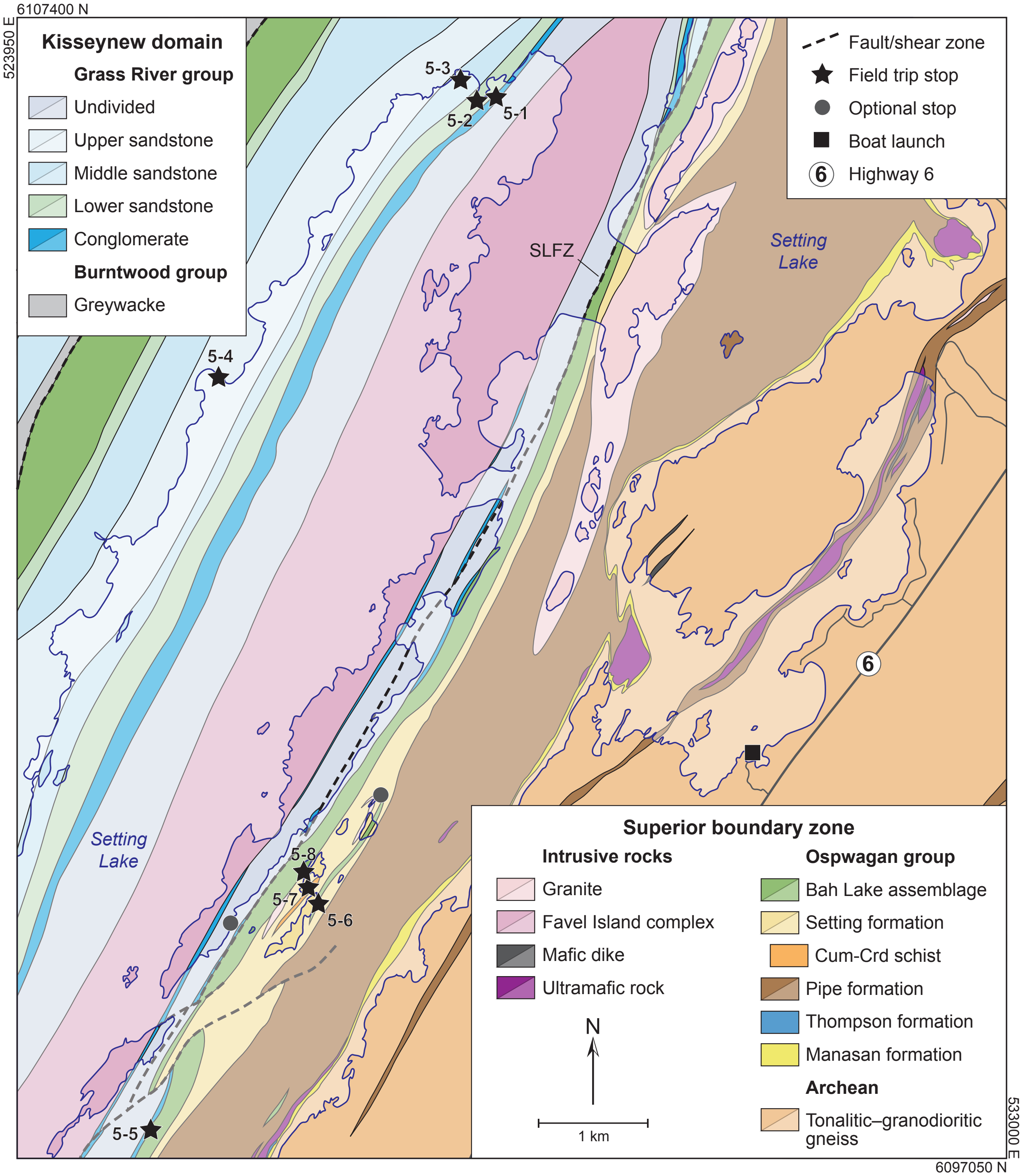

***Figure 18**: Geology of the Setting Lake area (adapted from Macek et al., 2006). Abbreviations: Crd, cordierite; Cum, Cummingtonite; SLFZ, Setting Lake fault zone.*

mictic conglomerate of the Grass River group. This contact has been interpreted as a sheared unconformity (Zwanzig, 1998). A late mafic dike intruded the conglomerate, indicating <1850 Ma mafic magmatism.

Optional stop: pseudotachylite along the Setting Lake fault zone (UTM: 14U 525940 mE, 6099215 mN)

Pseudotachylite in Grass River group rocks along this shoreline indicates late (D_4 or later) brittle faulting along the Setting Lake fault zone.

Stop 5-6: Setting formation, S2 member (UTM: 14U 526730 mE, 6099420 mN)

This outcrop contains thickly bedded S2 member wacke with local beds grading from conglomerate at the base to mudstone at the top. Grading indicates tops to the west. These rocks are considerably coarser grained than the S1 member arenite, wacke, and pelite observed at the Pipe mine and Ospwagan Lake. The relationship between the S1 and S2 members is uncertain and may represent a lateral facies change toward deposits proximal to a submarine channel environment.

Stop 5-7: Setting formation cummingtonite-cordierite schist (UTM: 14U 526650 mE, 6099525 mN)

An isolated band of cummingtonite-cordierite schist occurs within Setting formation rocks on this island. The unit varies from homogeneous to laminated, and its origin is uncertain. Trace-element geochemistry is suggestive of a contaminated MORB-affinity, and the petrography and major-element geochemistry are suggestive of a metamorphosed, Mg-rich chlorite-altered rock similar to rocks found in VMS-related systems.

Stop 5-8: Lamprophyre dike (UTM: 14U 526600 mE, 6099650 mN)

A narrow (<80 cm) lamprophyre dike intrudes relatively homogeneous amphibolite, which could be part of the Bah Lake assemblage, on the south end of this island. The lamprophyre is part of a suite of late-kinematic dikes that occur along the Setting Lake fault zone in the northern Setting Lake area (Ducharme and Zwanzig, 1999; Zwanzig, 1999). The lamprophyre is discordant to the regional foliation, but is itself internally foliated with local shear bands. The dike contains <5 cm felsic xenoliths. Hornblende and biotite occur as phenocrysts in a groundmass of dominantly hornblende with lesser amounts of biotite, feldspar, and carbonate. The weathered surface is characterized by pock-marked weathering.

Optional stop: Setting formation, S2 member (UTM: 14U 527295 mE, 6100355 mN)

This outcrop displays bedding in the Setting formation S2 member, which grades from conglomerate to mudstone. The rocks are isoclinally folded by F_3 with pebbles displaying varying intensities of stretching.

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