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FIELD TRIP GUIDEBOOK

EVOLUTION OF THE THOMPSON NICKEL BELT, MANITOBA: SETTING OF NI-CU DEPOSITS IN THE WESTERN PART OF THE CIRCUM SUPERIOR BOUNDARY ZONE (FIELD TRIP A1)

by

W. Bleeker[1] and J. Macek[2]

1 Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario, K1A 0E8
2 Manitoba Energy and Mines, Geological Services Branch, 1395 Ellice Avenue, Suite 360, Winnipeg, Manitoba, R3G 3P2

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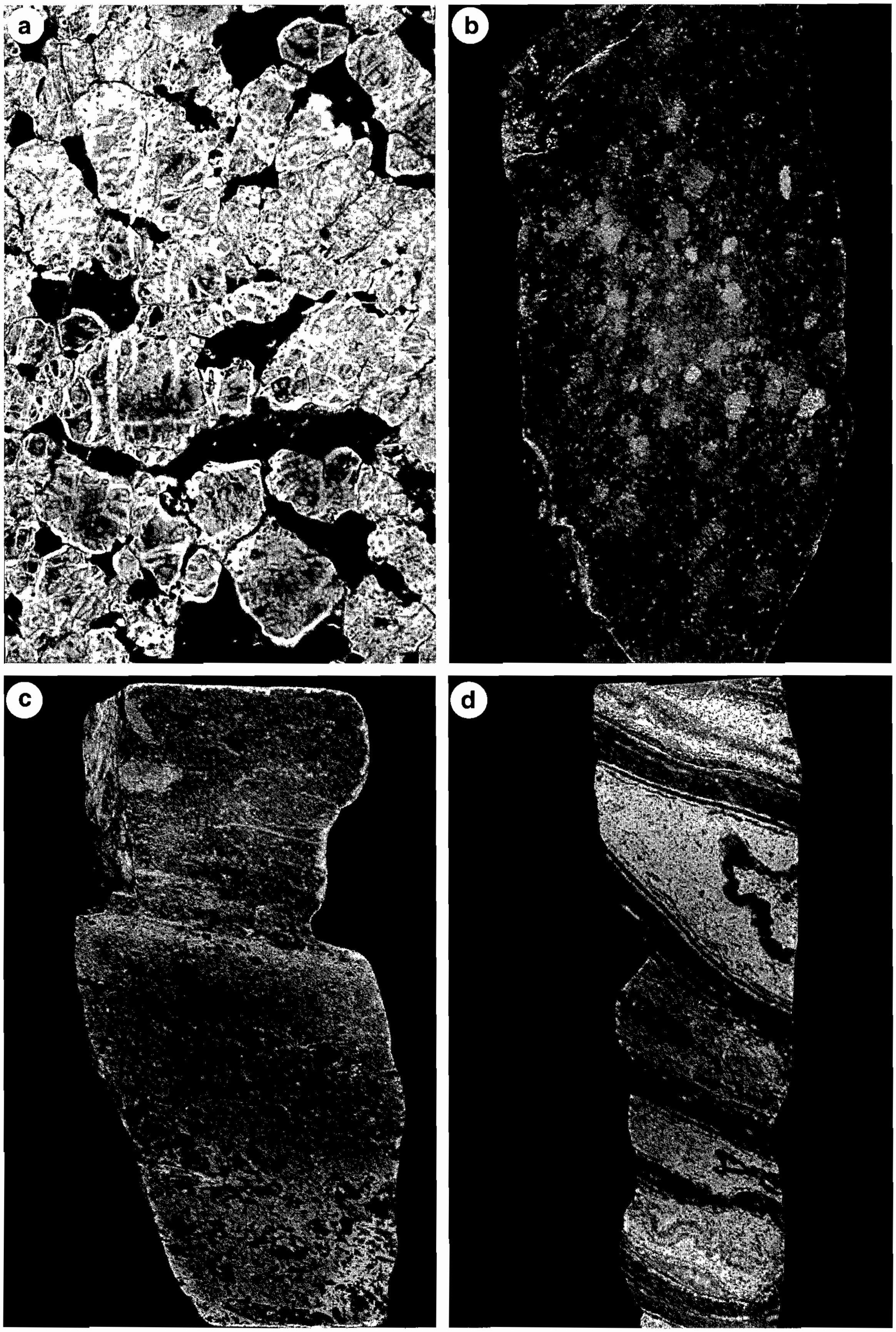
FRONTISPIECE CAPTIONS

a) *Net-textured interstitial sulphide ore (black) in serpentinized dunite host (various grey shades; serpentine pseudomorphs after original cumulate olivine). Sulphide assemblage consists of pyrrhotite and pentlandite with accessory chalcopyrite. The mineralized dunite occurs as augen-shaped boudins along the Thompson ore zone. Transmitted light, field of view approximately 3 mm wide.*

b) *Massive magmatic sulphides from the Thompson ore zone. Porphyroblasts of pentlandite (up to 1 cm in size) occur within a matrix of pyrrhotite, finer grained pentlandite and accessory chalcopyrite. The coarse-grained, porphroblastic habit of pentlandite is due to slow, late-metamorphic (retrograde) growth in response to exsolution from a peak-metamorphic, homogeneous monosulphide solid solution (mss). The retrograde porphyroblasts overgrow accessory biotite flakes that outline the layer-parallel S_2 fabric in the ore. Massive magmatic sulphides at Thompson contain approximately 10.5 wt% Ni on a 100% sulphide basis. Sample is approximately 20 cm tall.*

c) *Contact between massive magmatic sulphide ore (upper part) and Ni-enriched sedimentary sulphide ore (lower part) from the Pipe II orebody. The relatively clean magmatic sulphide ore contains approximately 3.5 wt% Ni and is further characterized by accessory grains of Zn-rich, magmatic chromite. Ni-enriched sedimentary sulphide ore is derived from originally barren sulphide facies iron formation and is devoid of chromite grains. Nickel and several other metals (e.g., Co, Cu, Pd; but excluding Ir and Cr) have completely equilibrated across the contact. Contacts like these are interpreted as evidence for arrested, in-situ bulk assimilation of sulpide iron formation into the overlying ultramafic sill. Sample is approximately 30 cm tall.*

d) *Ni-enriched sedimentary sulphide ore from sulphide facies iron formation adjacent to magmatic sulphides of the Thompson ore zone. Impure, graphitic pyrrhotite layers are interlayered with graphitic schist and chert and show tight to isoclinal folding. Although identical in general character to barren sulphide facies iron formation further away from the Thompson orebody, this sample contains 3.5 wt% nickel and about 900 ppb Pd. Background values in barren sulphide facies iron formation are less than 500 ppm nickel and 10 ppb Pd. Sample is approximately 40 cm tall.*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GEOLOGY AND ORE DEPOSITS OF THOMPSON NICKEL BELT

by

Wouter Bleeker, Geological Survey of Canada

INTRODUCTION

The geology of the Thompson Nickel Belt (hereafter referred to as the TNB) will be examined along an east-to-west transect. We will begin the transect in unreworked 2.7 Ga granulites of the Pikwitonei subprovince, which constitute the southeastern Archean "foreland" of the TNB. Following the effects of progressive overprinting by Early Proterozoic structures and metamorphism, we will cross into the TNB proper and complete the transect along the western bounding fault with the internal zone of the Trans-Hudson Orogen. The general transect will set the stage for a detailed examination of the Early Proterozoic autochthonous cover sequence in the belt—the Ospwagan group—and the structural-stratigraphic setting of the world-class Thompson Ni-Cu sulphide deposits.

GEOLOGY OF THE THOMPSON NICKEL BELT

Tectonic Setting

The TNB (Zurbrigg, 1963; Bell, 1971; Coats and Brummer, 1971; Coats *et al.*, 1972; Cranstone and Turek, 1976; Weber and Scoates, 1978; Peredery *et al.*, 1982; Green *et al.*, 1985; Bleeker, 1990a, b and c; Machado *et al.*, 1990; Lucas *et al.*, 1993; Bleeker *et al.*, 1995) forms a 10 to 35 km wide belt of variably reworked Archean basement gneisses and Early Proterozoic cover rocks along the northwestern margin of the Superior craton (Figs. 1, 2, and 3). The belt is interpreted as a remnant of one of the Early Proterozoic continental margins that formed in response to break-up of a Late Archean (super?) continent, which included the Superior province.

Between 1.9 and 1.8 Ga, several dispersed Archean (micro-) continental plates reassembled into a large continental mass encompassing much of the pre-Grenvillian core of Laurentia (Fig. 1). This reassembly, culminating in the Hudsonian orogeny, is manifested by a network of Early Proterozoic orogenic belts. The 3000 km long Trans-Hudson Orogen, which stretches from south of the Canada-U.S.A. border all the way to northern Quebec, is the most extensively developed member of the Early Proterozoic collisional belts, and is thought to have formed in response to closure of a wide ocean basin—the Manikewan ocean (Staufer, 1984; see also Lewry, 1981). The TNB forms the southeastern external margin of the Trans-Hudson Orogen, along which juvenile Early Proterozoic domains collided with Archean foreland. The eastern boundary of the TNB is defined by the limit of Proterozoic overprinting on Archean basement gneisses, whereas its western boundary is defined by the westernmost occurrence, at surface, of continental margin rocks such as reworked Superior province gneisses and (or) its

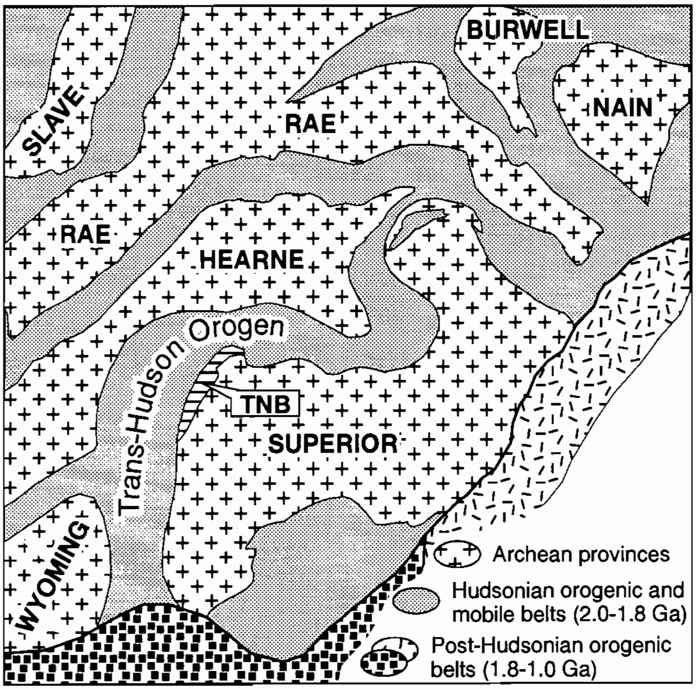


Figure 1: Schematic ("exploded") map of Laurentia showing Archean provinces separated by a "pan-Laurentian" system of Early Proterozoic orogenic and mobile belts (modified after Hoffman, 1989). Note location of the Thompson Nickel Belt (TNB).

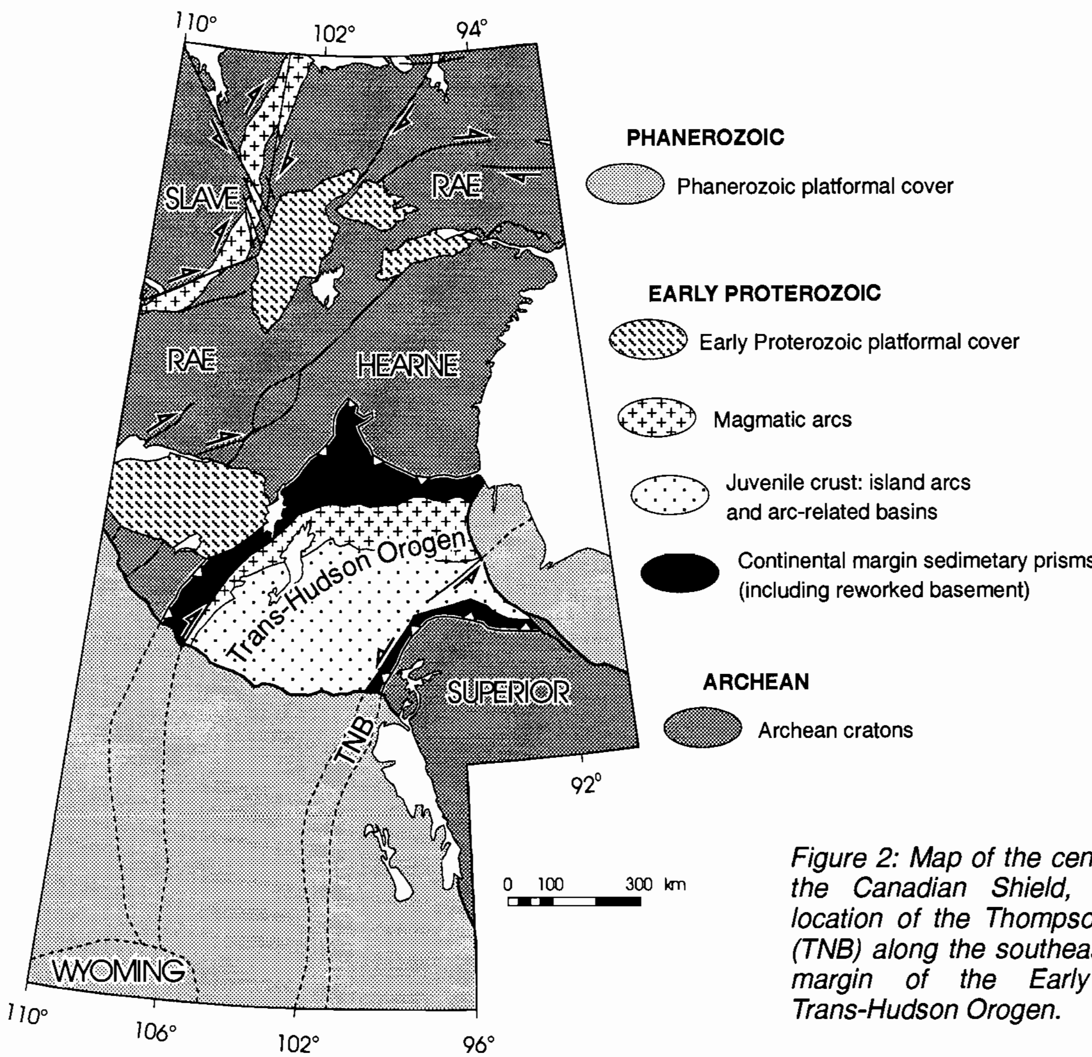


Figure 2: Map of the central portion of the Canadian Shield, showing the location of the Thompson Nickel Belt (TNB) along the southeastern external margin of the Early Proterozoic Trans-Hudson Orogen.

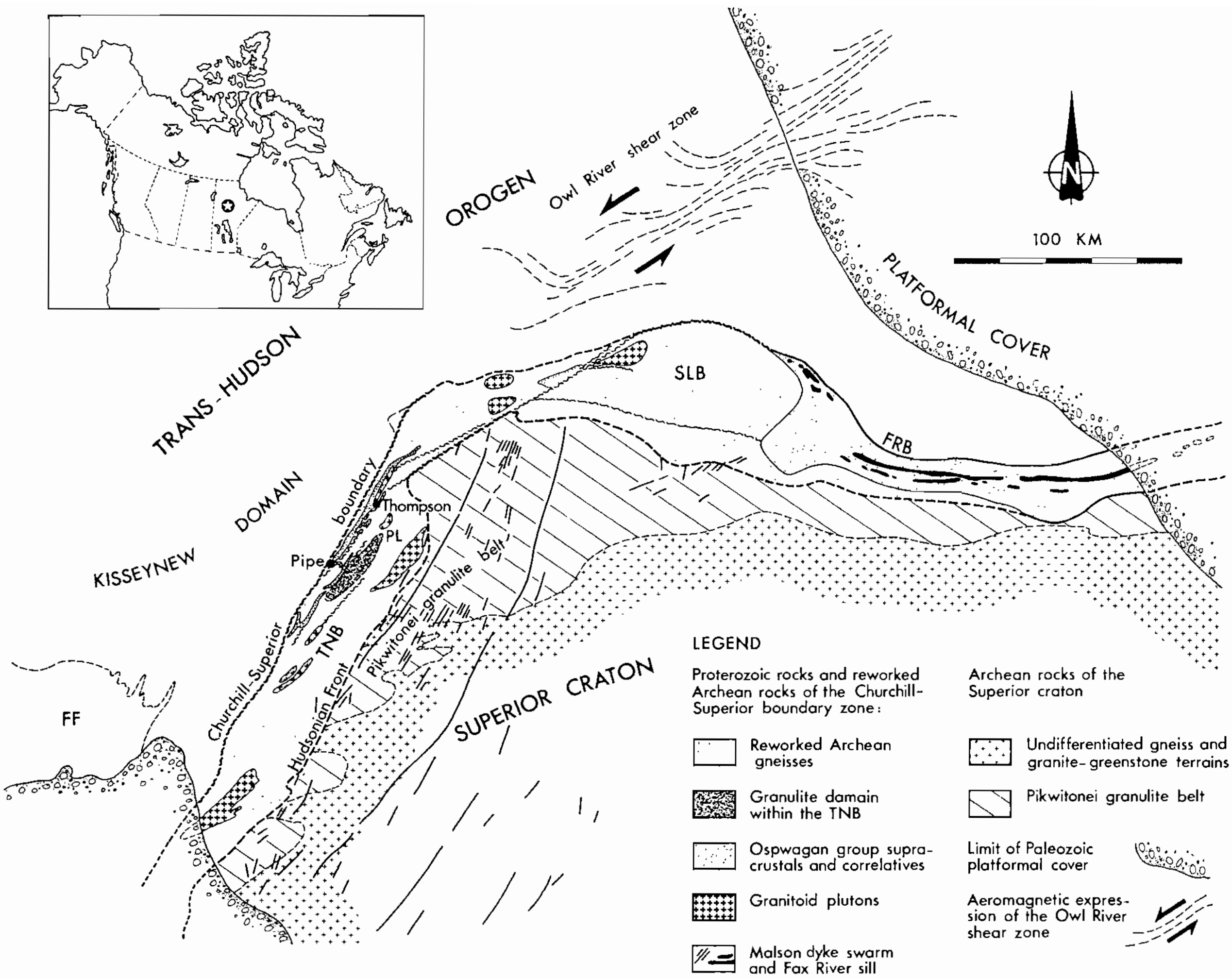


Figure 3: Simplified geological map of northern Manitoba, illustrating the location of the Thompson Nickel Belt (TNB) along the northwestern margin of the Archean Superior structural province. The TNB is part of the "Churchill-Superior Boundary Zone", which separates unreworked Archean crust in the southeast from the internal zone of the Early Proterozoic Trans-Hudson Orogen in the northwest. Abbreviations: FF for the 1.9-1.8 Ga Flin Flon-Snow Lake volcanic belt; SLB for Split Lake Block; and FRB for Fox River Belt.

autochthonous cover. Although poorly exposed, the TNB is well-defined by its strong gravity and magnetic expressions (Gibb, 1968; Kornik and MacLaren, 1966; Kornik, 1969). The stong geophysical signatures allow delineation of the southern extension of the TNB below Phanerozoic platformal cover, as far south as South Dakota (Green *et al.*, 1979; Thomas *et al.*, 1987).

To the north-northeast, the western boundary fault of the TNB appears structurally continuous with the Owl River shear zone (Bell, 1966), the aeromagnetic expression of which suggests a minimum sinistral displacement in the order of 100 kilometers (Bleeker, 1990a; see Fig. 3). Lithotectonically, however, the TNB as part of the Churchill-Superior Boundary Zone (Weber and Scoates, 1978) swings to the east and has its extension in the Split Lake Block and the Fox River Belt (Fig. 3). Like the TNB, the Split Lake Block is dominated by variably reworked basement gneisses intruded by Early Proterozoic dyke swarms, whereas the Fox River Belt consists of a homoclinal, steeply north-dipping sequence of Aphebian supracrustals and related intrusives (Baragar and Scoates, 1981; Scoates, 1981 and 1990).

Unreworked Archean crust to the southeast of the TNB includes low- to medium-grade granite-greenstone and gneiss terranes, and the high-grade Pikwitonei granulite belt (Fig. 3). Lithologically, the Pikwitonei granulites show sufficient similarities to the lower grade terranes to suggest that it represents deeper level exhumation of an overall gneiss and granite-greenstone crust (Roussel, 1965; Weber and Scoates, 1978; Hubregtse, 1980). The Pikwitonei granulite belt, which has an associated linear gravity high for much of its length (Gibb, 1968), parallels nearly the entire southeastern margin of the Churchill-Superior Boundary Zone.

Archean crust of the northwestern Superior craton is further characterized by mafic to ultramafic dykes of the Molson swarm (Ermanovics and Fahrig, 1975; Scoates and Macek, 1978; Paktunc, 1987). This dyke swarm, which has been dated at 1883 Ma (Heaman *et al.*, 1986), can be followed into the TNB (Cranstone and Turek, 1976; Bleeker, 1990c), but is lacking from younger allochthonous Aphebian crust to the northwest of the TNB constituting the internal zone of the orogen. Recent paleomagnetic studies have identified a dyke swarm that predates the Molson dykes (Zhai *et al.*, 1994).

The contact between the reworked external margin (the TNB) and the internal zone of the Trans-Hudson Orogen is marked by the Churchill-Superior boundary fault, a markedly linear, steeply dipping, mylonitic to cataclastic fault zone (the "Setting Lake lineament" of Rance, 1966; Coats *et al.*, 1972; Bleeker, 1990c). Across this fault zone, the TNB is juxtaposed against a collage of largely juvenile (Chauvel *et al.*, 1987; Stern *et al.*, 1995) Early Proterozoic terranes (Lewry, 1981; Lewry *et al.*, 1985 and 1987; Green *et al.*, 1985; Hoffman, 1988; Lucas *et al.*, 1993). From a TNB perspective, the Kisseynew Domain, comprising monotonous metaturbiditic gneisses (e.g., Zwanzig, 1990), is the most important component of this collage. The Kisseynew Domain probably represents the remnants of a back-arc-like basin (e.g., Hoffman, 1988; Ansdell *et al.*, 1995).

Local Geology

Variably reworked Archean basement gneisses are volumetrically the dominant rock type in the TNB and are at least partially derived from Pikwitonei granulite protoliths. Along the eastern boundary of the TNB—"the Hudsonian front"—the Pikwitonei granulites can be mapped into the TNB (Rance, 1966; Cranstone and Turek, 1976; Bleeker, 1990c) where they show progressive overprinting by Hudsonian ductile deformation and amphibolite facies metamorphism.

Various gneisses along the western margin of the TNB have been dated by the Rb-Sr whole rock method (Cranstone and Turek, 1976) and results indicated an Archean origin for these rocks. This has been confirmed by U-Pb zircon dating (Machado *et al.*, 1987 and 1990).

Locally, the TNB gneisses contain relics of granulite facies assemblages or pseudomorphs thereafter. At one locality, such pseudomorphic textures have been recognized in gneisses that occur structurally just below the contact between basement gneisses and rocks of the cover sequence. Since this contact can be shown to represent an Early Proterozoic angular unconformity (Bleeker, 1990c), granulitic basement comparable to the Pikwitonei granulites must have been exposed prior to deposition of the Early Proterozoic cover rocks. Differential uplift of the granulites may have resulted, in part, from asymmetrical extension during rifting along the TNB margin.

Remnants of the thin Early Proterozoic cover sequence, which is referred to as the Ospwagan group (Scoates *et al.*, 1977; Bleeker and Macek 1988a and b; Macek and Bleeker, 1989; Bleeker, 1990c), occur along the western margin of the TNB, in deeply dissected remnants of regional-scale fold interference patterns. Although an empirical stratigraphy had been recognized by INCO geologists (Peredery *et al.*, 1982), extreme deformation and poor exposure have long obscured the fundamental relationships between the supracrustal rocks and the gneisses. Detailed mapping of the Thompson open pit (Bleeker, 1989), the Pipe open pit (Bleeker and Macek, 1988a and b) and remapping of other key areas (Macek and Bleeker, 1989) has revealed the unconformable relationship and the existence of 1) a rare sillimanite-rich meta-regolith just below the unconformity, 2) a basal pebbly conglomerate, 3) a lower transgressive sequence which everywhere youngs away from the gneiss-metasediment interface, and 4) a detailed lithostratigraphy (Fig. 4), which can be correlated throughout much of the TNB, including the southern sub-Paleozoic extension of the belt (Nägerl and Bleeker, 1992).

A lower, fining-upward clastic sequence (Manasan formation) is overlain by a package of rocks dominated by chemical and pelitic sediments (Thompson and Pipe formations), after which there is a return to a coarse siliciclastic facies (Setting formation). Finally, the upper clastic rocks are overlain by mafic to ultramafic volcanics (Ospwagan formation). The volcanics are interpreted to be consanguineous with ultramafic sills which intruded the supracrustal sequence at various levels—most commonly near the base—and generated the nickel sulphide deposits.

Ospwagan group supracrustals, unlike the polymetamorphic basement gneisses, record a single, albeit complex Hudsonian P-T loop, with peak-metamorphic conditions ranging from lower to upper amphibolite facies (staurolite grade to garnet+sillimanite+K-feldspar grade, Fig. 5). Minor granitoid plutons, some of which show migmatitic envelopes, occur along the TNB. The granitoids are characterized by the occurrence of muscovite, and locally garnet, in addition to biotite. They appear anatectic in origin and are probably coeval with the thermal peak of Hudsonian metamorphism. U-Pb zircon dating has been hindered by Archean inheritance, but one pluton yielded a concordant monazite age of 1822±3 Ma (Machado *et al.*, 1987). Along the southern part of the TNB, a metadiorite intrusion has been dated at 1836^{+5}_{-3} Ma (U-Pb zircon; Bleeker *et al.*, 1994). To date, this metadiorite represents the oldest and only juvenile syn-orogenic intrusion in the belt. It is of interest that this age is similar to that of post-accretion igneous activity in the internal zone.

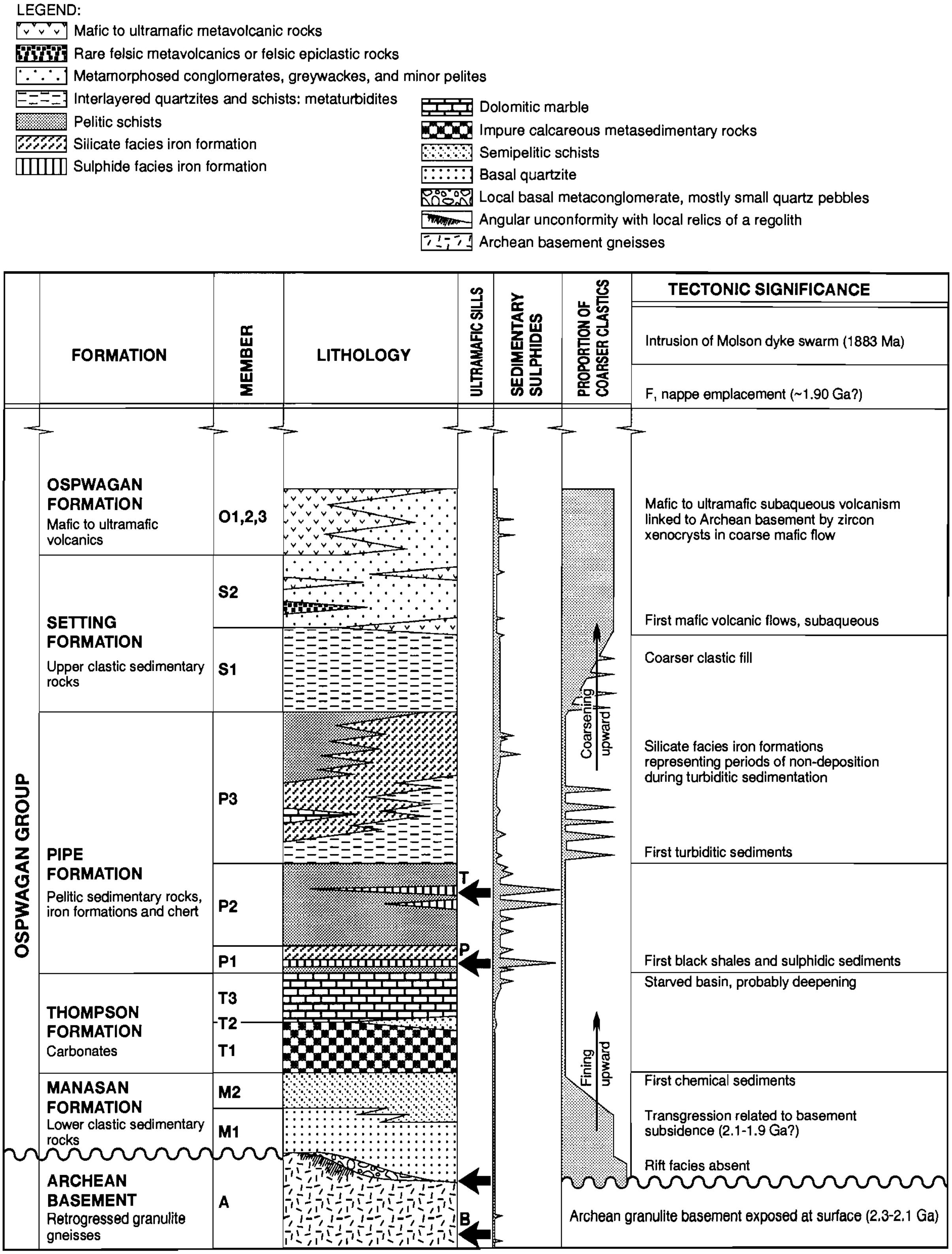


Figure 4: Generalized stratigraphy of the Ospwagan group cover sequence (after Bleeker, 1990c). Known intrusion levels of ultramafic sills are indicated in the central column (bold arrows: Thompson (T), Pipe II (P), Bucko (B), and others). The next column shows an approximate distribution of crustal sulphide material throughout the stratigraphy (full-scale peaks correspond to massive pyrrhotite layers in sulphide iron formations).

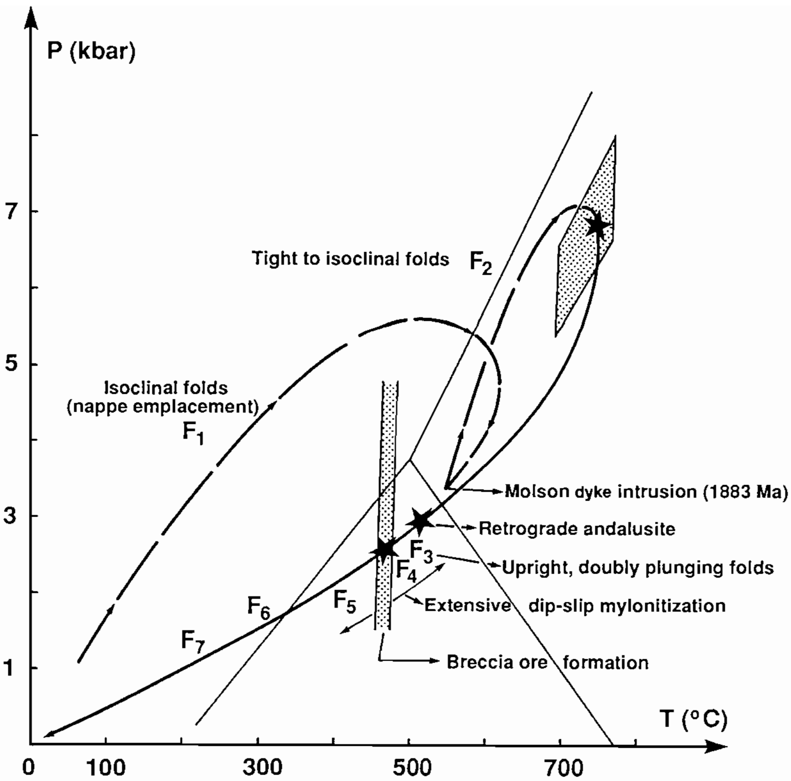


Figure 5: Generalized P-T-t loop for the Thompson Nickel Belt. Peak-metamorphic conditions are for Thompson.

Significance of the Molson Dyke Swarm

Various workers have interpreted emplacement of the Molson dyke swarm as an indication of initial rifting along the Thompson Nickel Belt segment of the Trans-Hudson orogen (Hubregtse, 1980; Green *et al.*, 1985). Although a causal relationship is indeed suggested by spatial association and parallelism of the dyke swarm with the Thompson Nickel Belt, the 1883 Ma age (Heaman *et al.*, 1986) of the swarm is clearly too young for the initial rifting event (Heaman *et al.*, 1986; Bleeker, 1990a). Recently, paleomagnetic work and preliminary U-Pb age dating has identified an older, ca. 2.1 Ga dyke swarm in the northwestern Superior Province, which is a more likely candidate for an initial rift swarm (Zhai *et al.*, 1994; L. Heaman, pers. comm.). Also relevant to the question of the age of initial rifting is a 2025±25 Ma age (Pb-Pb and U-Pb) on apatite cement from the Richmond Gulf Group, along the eastern Hudson Bay coast (Chandler and Parrish, 1989).

Along the western margin of the Thompson Nickel Belt, mafic dykes which Bleeker and Macek (1988a and b) correlate with the Molson swarm of the Archean foreland on the basis of their chemistry, orientation and general morphology, cross-cut not only basement gneisses but also the complete cover sequence and earliest Hudsonian fold structures (F_1). Rifting, subsidence, deposition of the Ospwagan group and F_1 deformation thus all preceded 1883 Ma, unless it can be demonstrated that the dykes in question are not part of the Molson swarm but belong to an as yet unrecognized younger swarm.

Significance of the Ospwagan Group Cover Sequence

The autochthonous nature and detailed lithostratigraphy of the Ospwagan group cover sequence allow inferences to be made about the early tectonic evolution of the Thompson Nickel Belt (Fig. 4). The lower clastic metasedimentary rocks of the Ospwagan group (Manasan formation) consists of a thin, locally developed quartz and feldspar pebble conglomerate, overlain by quartzites of subarkosic to quartz-arenitic composition. These arenites fine upwards into quartz-rich siltstones which are overlain by wackes. The rather mature nature and uniform thickness of this lower clastic sequence, over strike lengths of tens of kilometers, suggests its deposition was related to a transgression that swept throughout the area in response to passive margin subsidence, rather than to infill of localized rift basins. The Thompson Nickel Belt thus represents a domain somewhere on the subsided margin of the Superior plate, but it does not constrain the location of the rifted margin proper. Remnants of the Ospwagan group are truncated by the fault-bounded western margin of the Thompson Nickel Belt. The original extent of the Ospwagan group and, consequently, the original extent of the Superior plate must have reached further west. It thus appears that a marginal part of the Superior plate must occur below the internal zone of the orogen or was sliced off.

The clastic Manasan formation is overlain by the Thompson formation, which comprises chert and siliceous dolomite of variable thickness. The Thompson formation is overlain by graphitic sulphide facies iron formation, silicate facies iron formation, pelitic schists with intercalated sulphide facies iron formation and again silicate facies iron formation, collectively comprising the Pipe formation. The lower part of this chemical sediment-dominated package (the Thompson formation) suggests establishment of a stable and sediment starved platform which evolved into a more active tectonic environment with the recurrence of a immature, coarse clastic sedimentation (the Setting formation; Fig. 4). Turbiditic sedimentary rocks of the Setting formation are intercalated with minor mafic volcanic rocks and are overlain by the main sequence of mafic to ultramafic volcanics (Ospwagan formation) at the top of the Ospwagan group.

Age of the Ospwagan Group

Attempts at dating the Ospwagan group directly have been largely unsuccessful. Metasedimentary rocks are cut by probable Molson dykes, bracketing their age between 2.4 and 1.88 Ga. Sr-isotope systematics possibly suggest a narrower 2.1-1.88 Ga range (Brooks and Theyer, 1981).

A greywacke from the Setting formation, sampled at Ospwagan Lake, contains a variety of detrital zircons spanning a range from 2850 Ma down to 2675 Ma (Bleeker and Machado, unpubl. data). Although this result does not provide a tight constrain on the age of the Ospwagan group, it supports geological evidence that the Ospwagan group is post-Archean in age and that Superior Province basement was the major, if not the only source of the detritus. Two single, xenocrystic zircons from a coarse basaltic flow or sill in the Ospwagan formation have $^{207}Pb/^{206}Pb$ ages of ca. 2740 and 2640 Ma, respectively. This result clearly ties not only the Ospwagan group metasediments, but also the metavolcanics at the top of the sequence to the underlying Superior province basement (Bleeker and Machado, unpubl. data).

Recently, well-preserved komatiites were discovered by Cominco Limited beneath Lake Winnipegosis in the sub-Paleozoic extension of the Thompson Nickel Belt. Associated basalt flows have been dated at 1864^{+6}_{-4} Ma (Hulbert *et al.*, 1994). The preservation of these komatiitic volcanics, locally at sub-greenschist facies metamorphic grade, is exceptional and in stark contrast with multiply folded, faulted, and thoroughly metamorphosed Ospwagan group-like metasedimentary and ultramafic rocks that have been intersected by exploration drill holes in the same general sub-Paleozoic area (Nägerl and Bleeker, 1992) somewhat further to the northwest. It is not clear therefore how these post-Molson swarm komatiitic and basaltic volcanics relate to the Ospwagan group, if at all.

Structural-Metamorphic History

Earliest structures due to compressional tectonism are isoclinal F_1 folds, which may be of regional extent such as the nappe-like F_1 fold that dominates the Moak Lake-Pipe Lake region (Figs. 6, 7, and 8). Intensely reworked basement forms the core of this structure and thus locally overlies downward-facing supracrustals, such as at Thompson Mine (Bleeker, 1990c).

The metamorphic regime during F_1 is unknown, but basement involvement in the nappe structure suggests at least lower amphibolite facies conditions. Timing of F_1 is also uncertain, but appears to have predated 1883 Ma, since F_1 folds are cross-cut by mafic dykes correlated with the Molson dyke swarm. F_1 is overprinted by a second phase of tight to isoclinal folds (F_2), which developed under high-grade conditions. On a mesoscopic scale, F_2 folds are prominent and fold Molson dykes (Fig. 9). F_2 folds were probably recumbent and appear to represent a second phase of ductile thrusting.

The thermal peak of regional metamorphism outlasted F_2 and occurred between 1.82 and 1.78 Ga. The combined F_1-F_2 history has been interpreted as a phase of crustal thickening during which a southeast-tapering wedge of recumbent folds and thrusts was emplaced on the margin of the Superior plate (Bleeker, 1990a, b, and c). Thermal relaxation of the thickened crust culminated in generation of anatectic granites and the thermal peak of metamorphism.

It should be pointed out however that due to the complex polyphase history, the vergence of the early structures is ambiguous. An easterly vergence has been assumed mainly because a clear record of long-lived subduction beneath the Superior province is absent. Recent seismic reflection profiles (Lucas *et al.*, 1993) show, however, that the Thompson Nickel Belt and the boundary zone in general are dominated by a strongly reflective, southeasterly dipping imbrication structure, at least on the southern line extending from Cross Lake to Snow Lake (Fig. 10).

In an attempt to reconcile the above conflict, the easterly dipping imbrication structure is attributed to a late-stage flip in vergence direction, probably at the onset of the important D_3 event in the Thompson Nickel Belt.

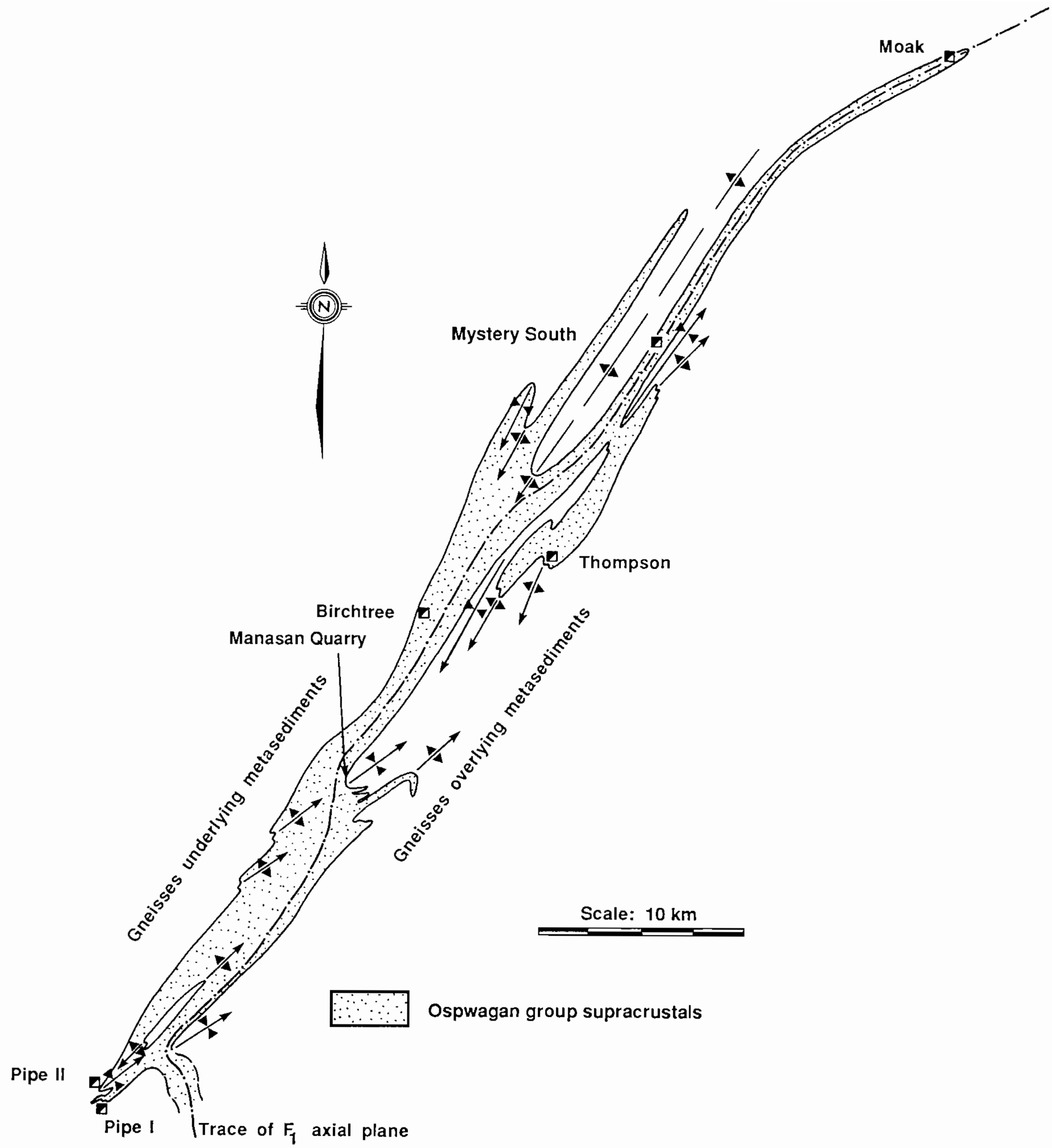


Figure 6: Simplified map of the Moak Lake-Pipe Lake area in the northern part of the Thompson Nickel Belt. The Ospwagan group cover sequence is preserved in a deeply dissected remnant of a refolded nappe structure. Metasedimentary rocks are completely enclosed by reworked basement gneisses, except at the southern end where the structure is poorly defined and possibly open; the supracrustals always young away from the basement/cover contact. The trace of the axial plane of the F_1 nappe and the axes of major F_3 folds are indicated. Also shown are the localities of major nickel sulphide deposits (Moak, Mystery South, Thompson, Birchtree, and Pipe I and II).

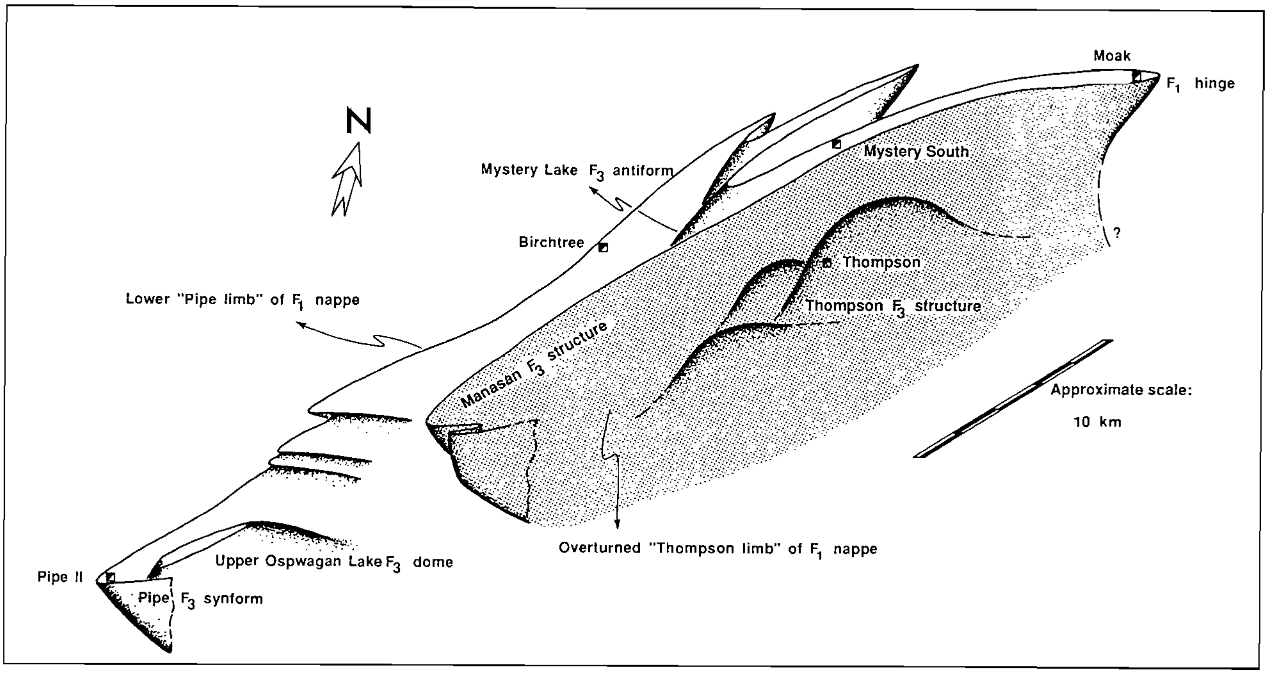
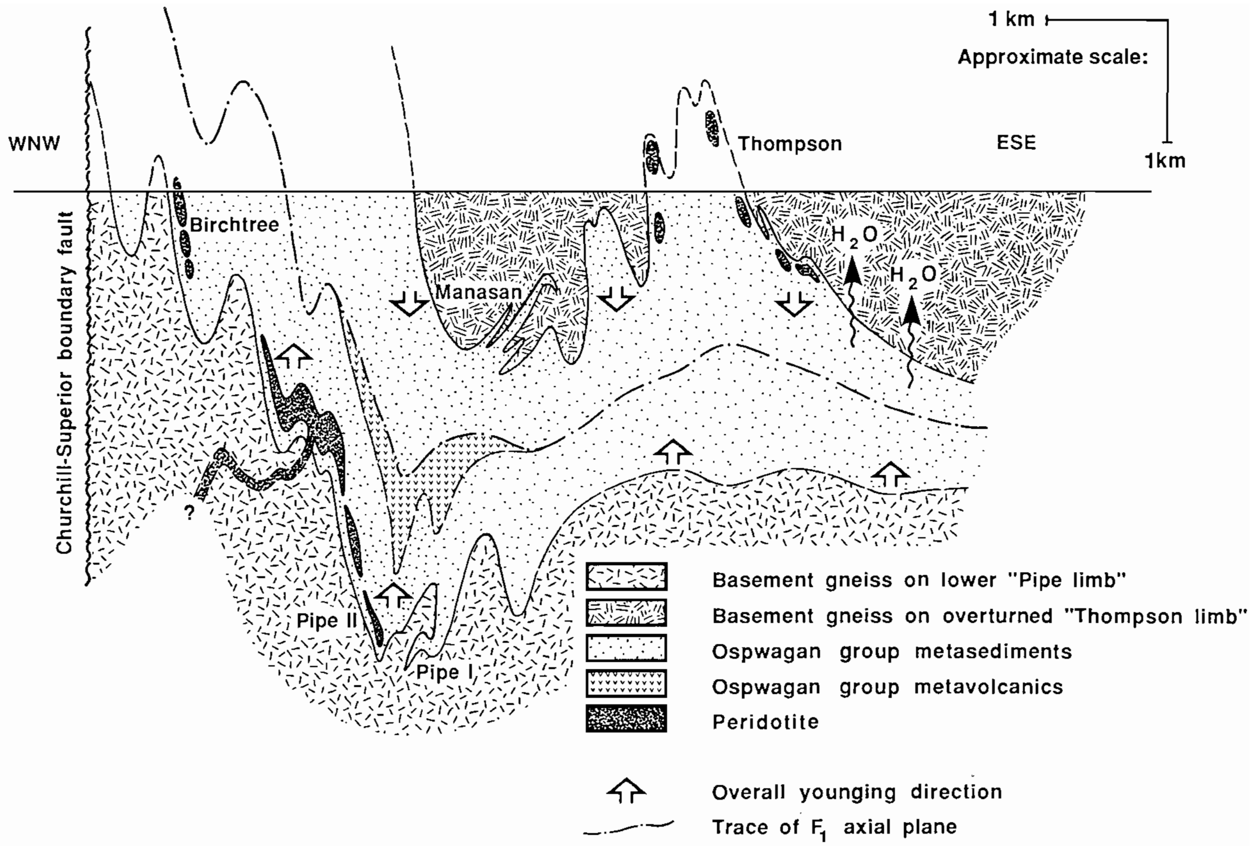


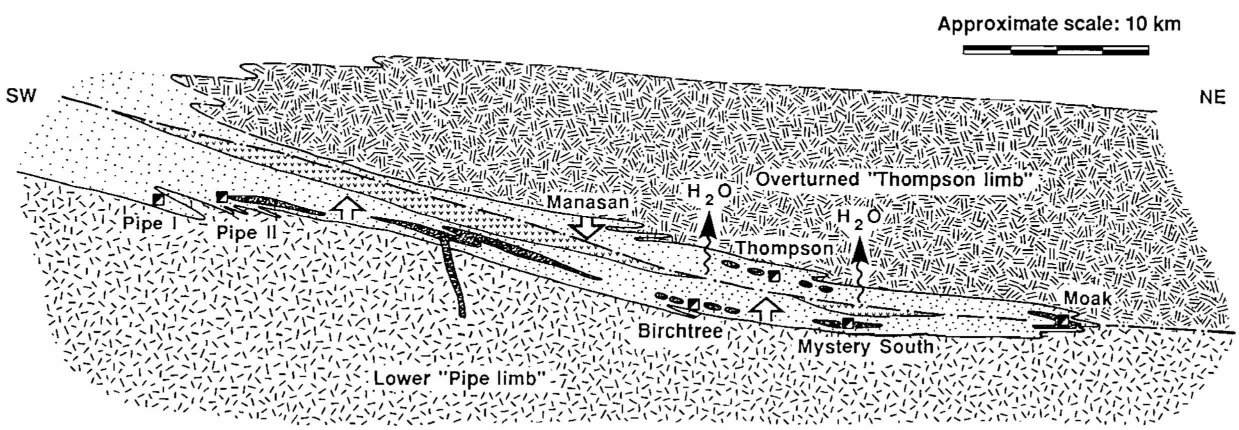
Figure 7: 3D sketch of the regional structure in the Moak Lake-Pipe Lake area, as viewed from the south. All upright structures are F_3 folds. The downward-facing Thompson and Manasan F_3 fold structures define the regionally overturned limb of the nappe (the "Thompson limb"). Localities such as Pipe II and Birchtree are located on the lower, upward-facing limb of the nappe (lower "Pipe limb").

Figure 8 (next page): Schematic structural sections throught the refolded nappe structure.

Upper diagram (a) shows a cross-section in which various structural elements are projected onto a common section. Upright, slightly west-vergent folds belong to the F_3 generation. Note the location of the Thompson deposit on the overturned limb of the nappe.

Lower diagram (b) shows a simplified, northeast-to-southwest longitudinal section through the nappe, which further illustrates the relative position of the various deposits.


1 km
Approximate scale:
1km
WNW
ESE
Thompson
Birchtree
Manasan
Churchill-Superior boundary fault
H_2O
H_2O
?
Pipe II
Pipe I
Basement gneiss on lower "Pipe limb"
Basement gneiss on overturned "Thompson limb"
Ospwagan group metasediments
Ospwagan group metavolcanics
Peridotite
Overall younging direction
Trace of F_1 axial plane


Approximate scale: 10 km
SW
NE
Overturned "Thompson limb"
Manasan
H_2O
H_2O
Thompson
Pipe I
Pipe II
Moak
Birchtree
Mystery South
Lower "Pipe limb"

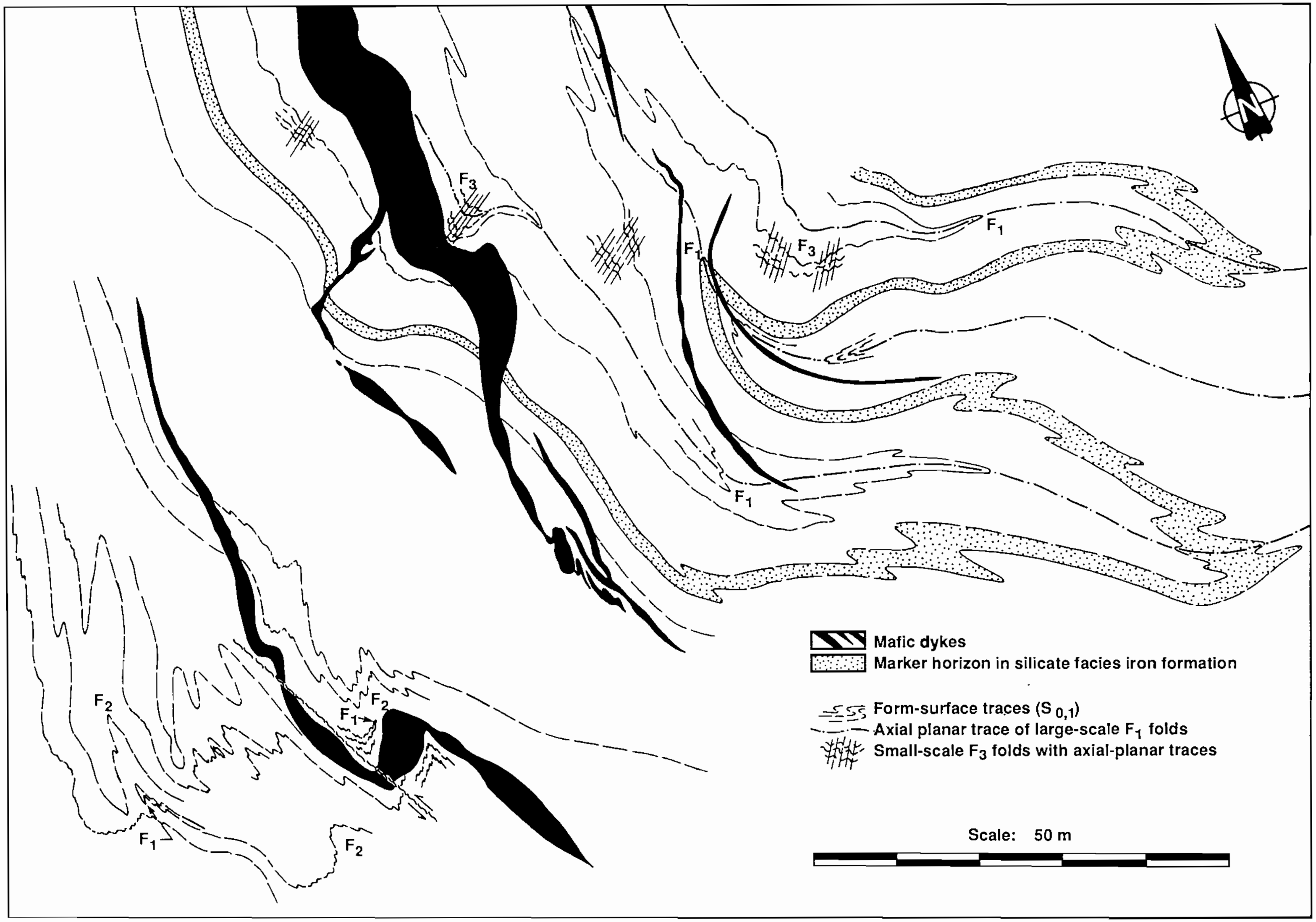


Figure 9: Form surface map of F1-F2-F3 overprinting relationships in the core of the Pipe II synform (F3), Pipe II open pit. The local structure is dominated by tight to isoclinal F1 folds. Mafic dykes, which tentatively are correlated with the Molson dyke swarm, cut these folds. F1 folds and dykes are locally refolded by asymmetric F2 folds. The two early fold generations and the dykes are then refolded into a relatively open northeast-trending synform (after Bleeker, 1990c).

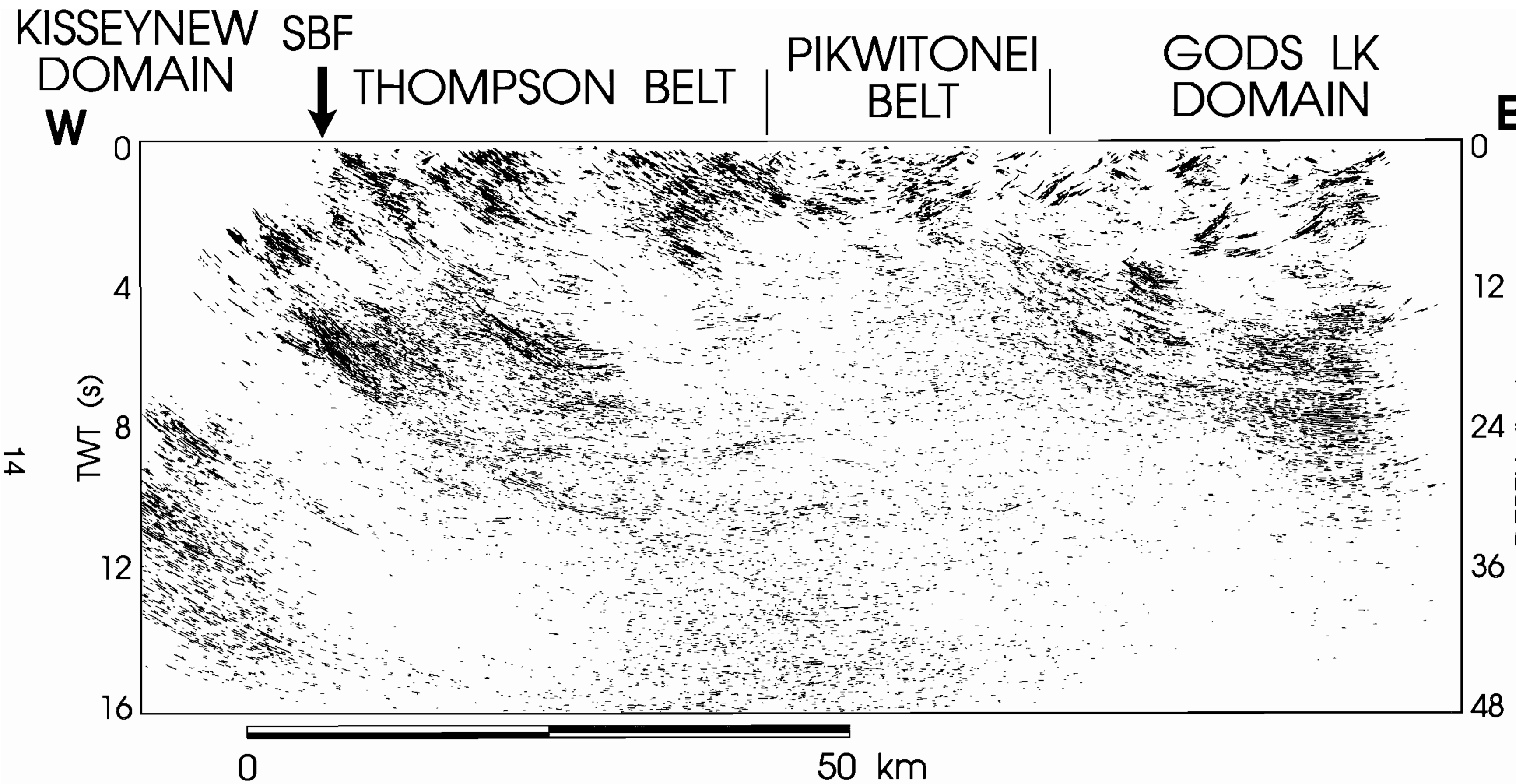


Figure 10: Lithoprobe seismic reflection profile (Line 2) extending from the Cross-Lake greenstone belt in the Superior province, across the southern portion of the TNB, into the the Kisseynew gneiss domain of the internal zone of the Trans-Hudson Orogen (after White and Lucas, 1994).

From 1786 Ma onwards, up to about 1765 Ma, the Thompson Nickel Belt experienced intrusion of a vast number of pegmatite dykes that are associated with hydrous, retrogressive, amphibolite facies metamorphism. At about 1770 Ma, during F_3 upright folding, metamorphic temperatures must have cooled rapidly because late-F_3 folded micas do not show any noticeable kink-band boundary migration. In terms of their asymmetry, F_3 folds, although nearly upright, show a weak but consistent westerly vergence. It is suggested that the sudden influx of pegmatites dykes and associated hydrous fluids marks the onset of west-vergent thrusting, during which the Superior plate started to override the internal collision zone. Syn-D_3 mylonites and sheath folds are probably the result of this deformation, which led to rapid uplift of the Thompson Nickel Belt, now being the overriding plate, around 1770 Ma.

At about that time, under falling metamorphic temperatures, intense sinistral transpression of the nappe/thrust pile produced the doubly-plunging F_3 folds, which are the most obvious structures throughout the Thompson Nickel Belt, transposing the preexisting recumbent fold/thrust pile into a steep gneiss and schist belt. Axial planes of macroscopic F_3 folds dip steeply to the southeast, trend 035° to 050°, and form a left-stepping *en échelon* pattern, with axial traces trending 0-15° clockwise of the 035° trend of the belt (Bleeker, 1990a). Extensive mylonitization occurred late- to post-kinematic with respect to F_3 folds and is confined to shear zones which tend to be parallel to steeply dipping limbs of the upright F_3 folds. Obvious kinematic indicators in these reverse shear zones reveal a dip-slip, dominantly east-side up displacement sense (Fueten and Robin, 1989; Bleeker, 1990a). Hence the reverse shear zones accommodated further shortening across the belt and further differential uplift of the Superior plate margin relative to the internal zone.

The ductile shear zones are overprinted by steep, pseudotachylite-generating brittle-ductile and brittle faults, which trend (sub) parallel to the belt. These faults are concentrated along the western margin of the belt and are especially numerous along the Churchill-Superior boundary fault. Early faults indicate sinistral strike-slip displacement, and are a late expression of the F_3 sinistral transpressive regime. Late movement on these faults is dextral. A late system of conjugate, transcurrent brittle faults is pervasive throughout the belt and developed in conjunction with dextral transpression on the main boundary fault.

NICKEL DEPOSITS IN THE MOAK LAKE-PIPE LAKE AREA

Introduction

Although several deposits are known from the southern part of the belt (Fig. 11), some of which were mined briefly such as the Falconbridge-owned Manibridge deposit (Coats and Brummer, 1971), the Moak Lake-Pipe Lake area in the northern part of the belt has been the most productive. All deposits in this area occur within the complexly folded Early Proterozoic cover sequence, the Ospwagan group. Within the cover sequence, the deposits are associated with ultramafic sills that show variable degrees of disruption or boudinage.

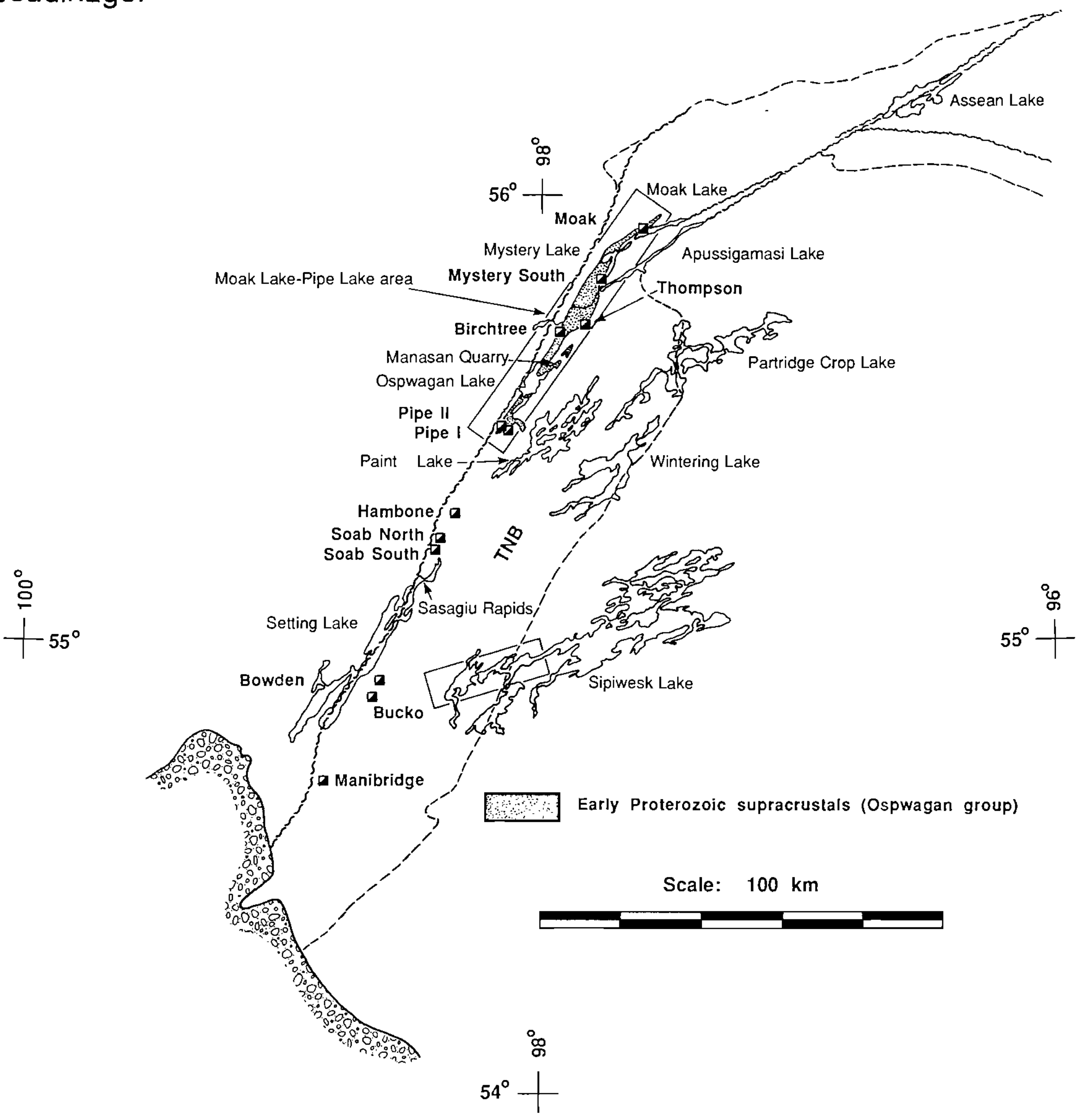


Figure 11: Location map showing the various nickel sulphide deposits known along the TNB. Also indicated is the Moak Lake-Pipe Lake area, where most of the productive deposits occur within complexly folded rocks of the Ospwagan group cover sequence.

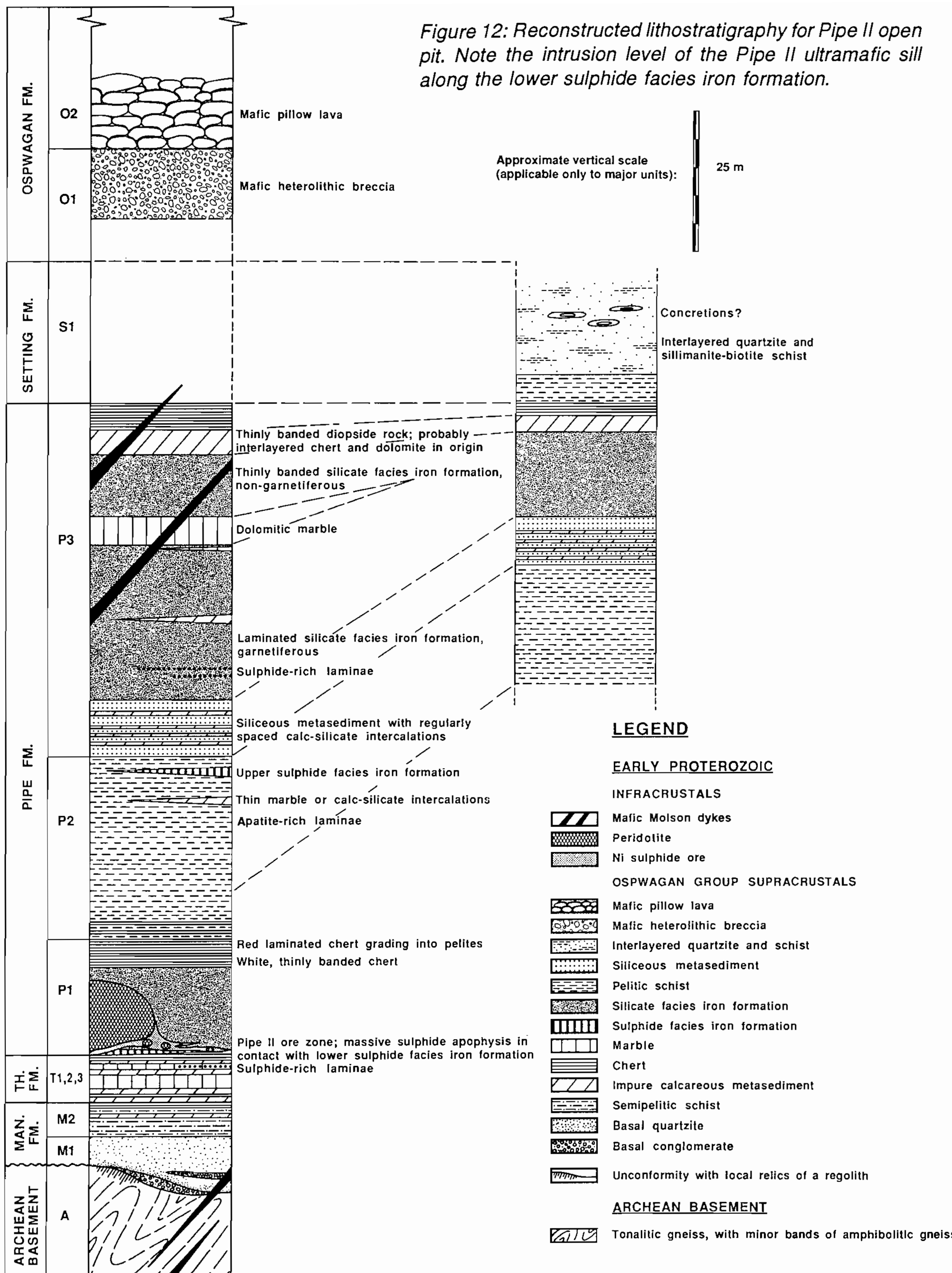
The strain magnitude is extreme on the overturned "Thompson limb" of the nappe, but is somewhat less on the lower upward facing "Pipe limb". Another interesting consequence of the F_1 nappe structure is the metamorphic-metasomatic contrast between the lower "Pipe limb" and the overturned "Thompson limb". Reworked grey gneisses on the lower limb are still largely tonalitic in character, comparable to unreworked equivalents of the Archean foreland, whereas reworked gneisses on the overturned "Thompson limb" are often pink in colour, have a considerable potassium feldspar content and show an abundance of pegmatitic sweats. The apparent addition of components such as potassium is attributed to upward migration of metamorphic fluids derived from the dehydrating metasediments in the synformal core of the nappe (Fig. 8).

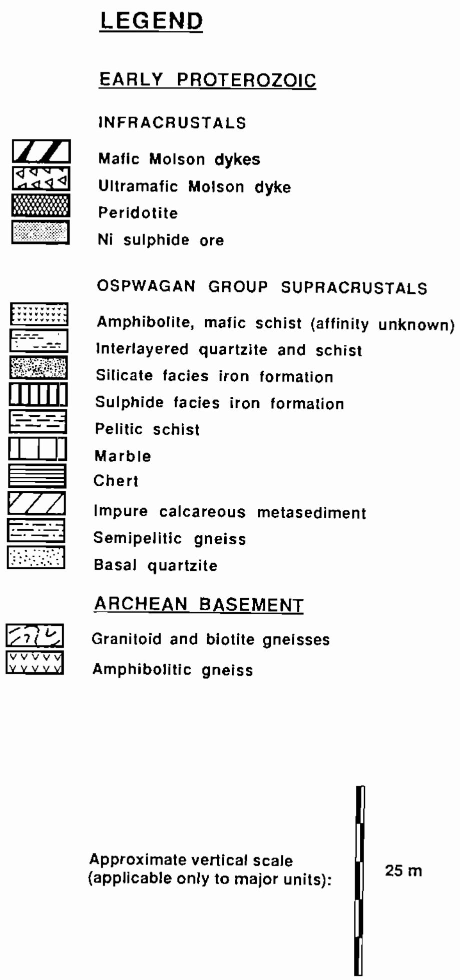
Lithostratigraphic analysis of the multiply transposed supracrustals shows that ultramafic sills intruded the cover sequence at various levels (e.g., Fig. 4). Sill or dyke-like bodies occur also within the underlying basement. The Pipe II sill intruded low in the cover sequence, below the pelitic schists, along a graphitic sulphide facies iron formation (Fig. 12). Other ultramafic bodies on the lower limb of the nappe, such as those at Birchtree Mine and on Ospwagan Lake, occur in similar lithostratigraphic position,. The Thompson sill intruded higher in the sequence, near the top of the pelitic schist unit (Fig. 13). This horizon is also characterized by large concentrations of sedimentary sulphides in the form of disseminated, banded or massive pyrrhotite in a host of extremely graphitic schist or interlayered with chert. This second level of sulphide facies iron formation can also be identified at Pipe II open pit, where it forms a 10 to 50 cm thick band of pyrrhotite with inclusions of graphitic schist and chert. Hence, all known deposits in the Moak Lake-Pipe Lake area are associated with major sedimentary sulphide concentrations. They combine the occurrence of ultramafic sills with either the lower (Pipe II deposit) or the upper (Thompson deposit) sulphide facies iron formation (Fig. 14).

Thompson Ni Sulphide Deposit

After 10 years of intense exploration in the area, during which the low grade Moak and Mystery deposits were outlined, INCO discovered the Thompson ore body in 1956. A shaft had been sunk at Moak the previous year, for underground exploration and to prepare this deposit for production, but these operations were discontinued as soon as the Thompson discovery proved to be of sufficient tonnage (Fraser, 1985). At Thompson, production commenced in 1961 at which time 23 million tonnes of ore with a combined Ni-Cu grade of 2.97 wt% had been outlined (Zurbrigg, 1963). A figure for the total tonnage has not been published but is estimated at 80 to 100 million tonnes at a similar grade. However, if dilution by wall rocks and redistribution of Ni to originally barren sedimentary sulphides (see below) is taken into account, the original tonnage of magmatic Ni sulphides must be considerably lower, possibly in the range of 25 to 50 million tonnes with approximately 10 wt% Ni on a 100% sulphide basis. Present production capacity of the Thompson metallurgical complex amounts to approximately 50,000 tonnes of refined Ni per year (Hopkins, 1986). Thompson underground operations provide approximately 6000 tonnes of ore per day.

Figure 12: Reconstructed lithostratigraphy for Pipe II open pit. Note the intrusion level of the Pipe II ultramafic sill along the lower sulphide facies iron formation.





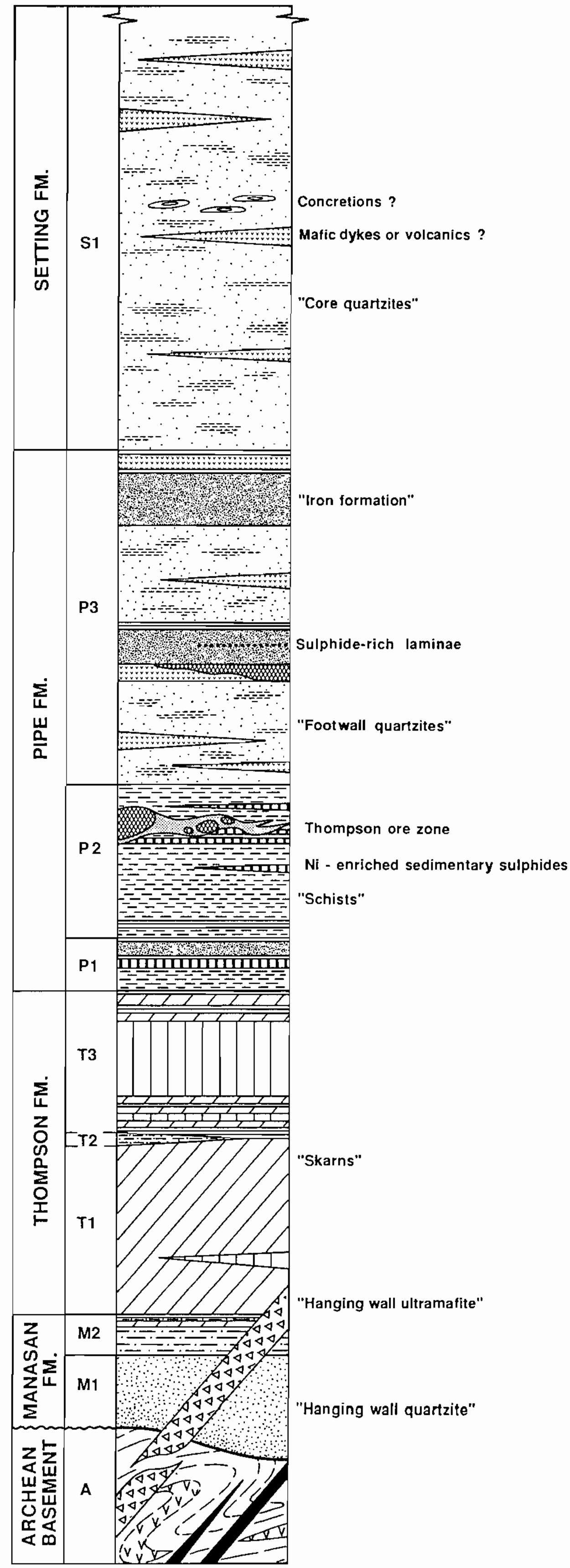


Figure 13: Reconstructed lithostratigraphy for Thompson mine. Note the intrusion level of the Thompson sill along the upper sulphide facies iron formation.

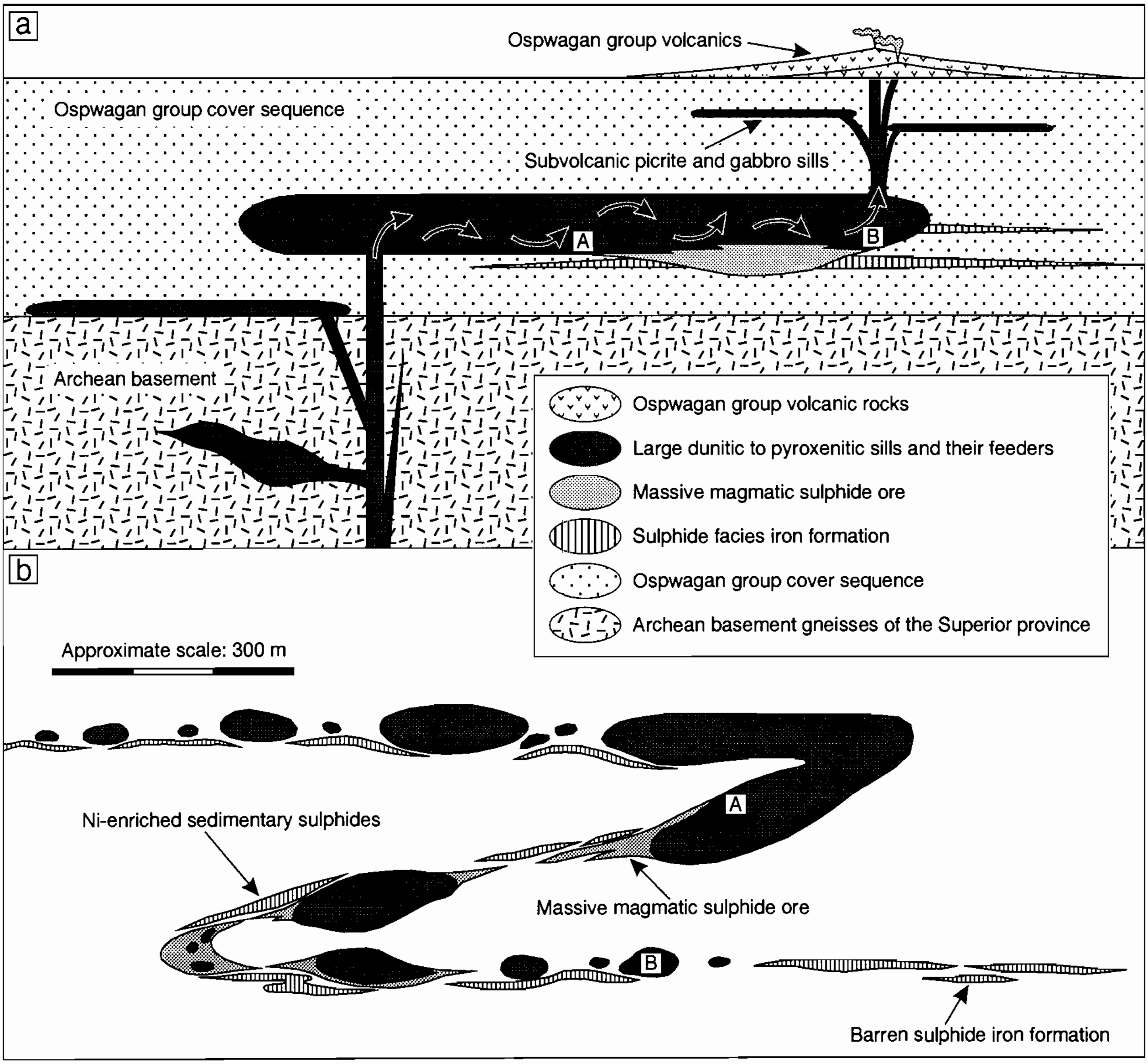


Figure 14: Schematic genetic section showing the overall stratigraphic relationships between the Ni sulphide deposits, the Ospwagan group, and Archean basement (after Bleeker, 1990c). Upper diagram (a) shows relationships prior to deformation. Ultramafic sills intruded within basement gneisses, along the basement/cover contact and, most commonly, within the Ospwagan group cover sequence. Although some deposits are hosted by dunitic to peridotitic sills within the basement environment, Thompson-type deposits formed where large sills intruded along sulphide facies iron formations within the cover sequence. The sills probably acted as conduits for turbulently flowing magmas that ascended further to feed sills and volcanics higher in the stratigraphy. Lower diagram (b) illustrates the present stuctural style of folded, stretched and boudinaged ultramafic sills, remobilized massive sulphides, and remnant sedimentary sulphides. Note that magmatic sulphides were remobilized "passively" along the intrusive horizon due to stretching of the enveloping, metasedimentary horizons. Massive magmatic ore lenses, after deformation and remobilization, are still confined between points A and B.

During the 1970's and early 1980's, additional ore supply came from the Pipe II open pit. From 1986 to 1992, additional supply has come from several large open pits mining the crown pillars of the Thompson orebody, while in 1989, the nearby Birchtree Mine was reopened.

The Thompson ore zone has a 6 km long surface trace, has been proven beyond a depth of 2000 m and is still partly open. The ore body is stratabound and occurs within and generally near the top of the pelitic schist unit (P2 member of the Pipe formation; Figs. 13 and 15).

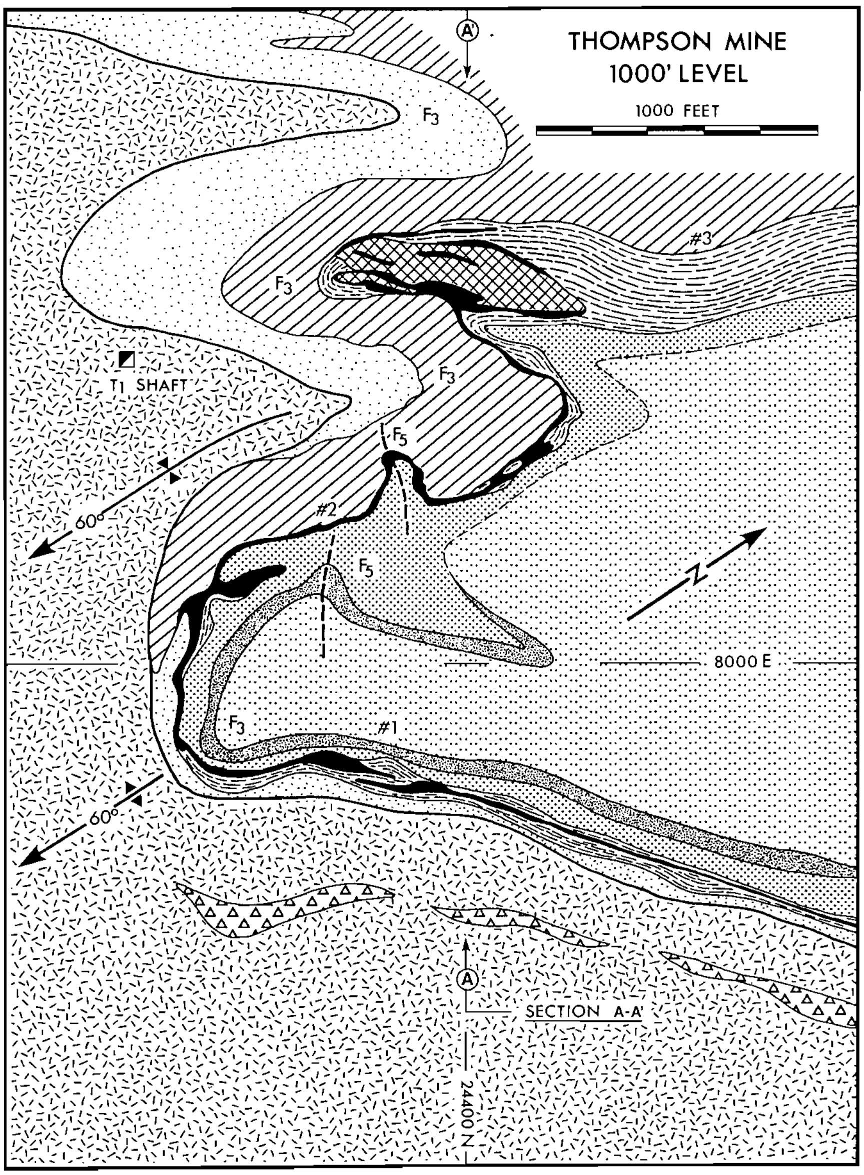


Figure 15a: Underground level map of Thompson mine ("1000 foot level"; modified after Zurbrigg, 1963). For legend and section A-A' refer to Figure 15b. Note the steeply south plunging F_3 folds. One of the limbs of the large-scale F_3 structure is refolded by subvertically plunging F_5 folds.

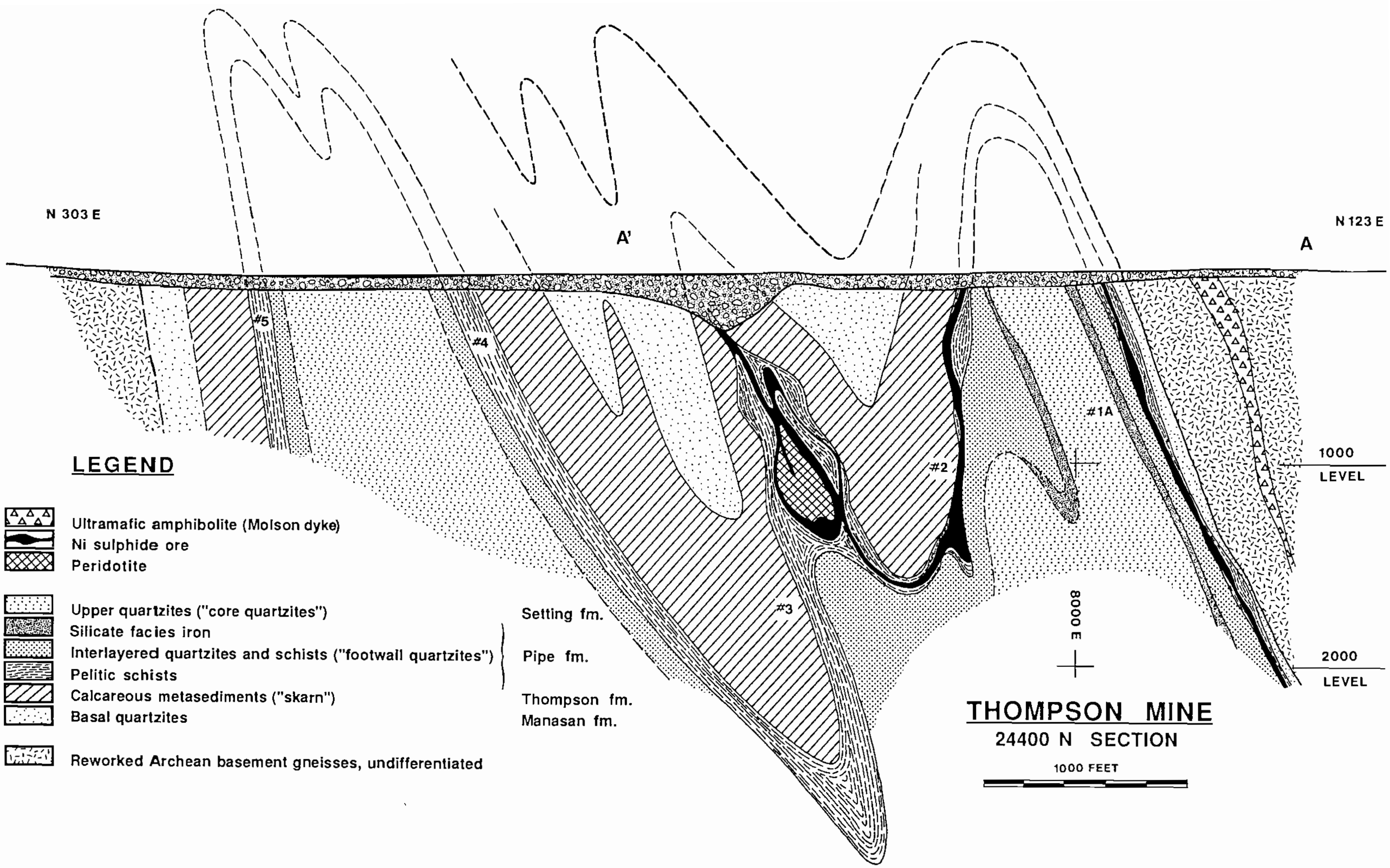


Figure 15b: Vertical cross-section through the downward-facing Thompson structure, looking towards the northeast. High amplitude, steeply southeasterly inclined folds all belong to the F3 generation, which refold the overturned "Thompson limb" of the F1 nappe structure.

Despite the complex and protracted structural-metamorphic history of the ore body, involving six folding phases and sillimanite+garnet+K-feldspar-grade metamorphism, no significant remobilization of sulphides has been observed across the original host horizon, except along late faults and as infilling of late stage tension gashes. Important remobilization occurred, however, along the original host horizon, but is described as "passive" in response to folding of and extension along the original horizon (Bleeker, 1990c). Consequently, magmatic Ni sulphides are still confined to a single lithostratigraphic horizon and, within this horizon, do not extend beyond the range of ultramafic boudins that represent remnants of the original parent sill (Fig. 14).

Pipe II Ni Sulphide Deposit

The Pipe II deposit was discovered in 1957. In 1961 INCO decided to mine the upper 220 m of the orebody by open pit method and to sink a 480 m deep shaft to further explore the orebody underground. Dredging of silt and clay overburden began in 1967 and production from the open pit started in 1969. To facilitate mining of deeper parts of the ore body, the shaft was deepened to a depth of 935 m. However, unfavourable market conditions prevented the underground production from starting up. The open pit was mined out in 1984 at a depth of 245 m, after producing approximately 18 million tonnes of relatively low grade interstitial, breccia and massive sulphide ore. Currently, INCO is reassessing the Pipe II underground mine and extensive exploration in the area has resulted in the discovery of a new ore lens, the "Pipe Deep" deposit.

The surface trace of the Pipe II ore zone is approximately 1 km long and occurs for much of its length along the stratigraphic base of a more than 2 km long, up to 150 m thick serpentinized ultramafic body. The lens-like ultramafic body, which is considered the boudinaged remnant of a larger sill (Bleeker, 1990c), occurs on the western limb of the main mine structure—a tight, steeply northeast-plunging F_3 synform with reworked basement gneisses on the limbs and highly deformed, staurolite-grade Ospwagan group cover rocks in the core. The large ultramafic boudin dips 75° to the southeast and is concordant with the enveloping metasedimentary rocks. Facing of the sill, based on its differentiation profile from sulphides, to dunite, peridotite, and orthopyroxenite, is towards the southeast and conforms to the overall younging direction of the enclosing metasedimentary rocks on the western limb.

On a regional scale, the Pipe II ultramafic boudin is the southernmost member of a discontinuous array of ultramafic bodies which stretches from Pipe II open pit, along the west shore of Ospwagan Lake, to Birchtree Mine. Where the lithostratigraphic position of these bodies could be checked, they occur at exactly the same horizon within the Ospwagan group cover sequence, above the carbonates and lower sulphide facies iron formation and below the first silicate facies iron formation. This intrusive level is below the lithostratigraphic horizon of the Thompson ore zone.

Sulphide ore is concentrated near the abrupt southern termination of the Pipe II ultramafic boudin and extends beyond the parent body onto the eastern limb of the macroscopic F_3 fold, in the form of a complexly shaped, tapering apophysis—the so-called "hanging-wall stringer". The more than 200 m long apophysis consists of extraparental massive sulphides (Bleeker, 1990c) carrying minor inclusions of serpentinized dunite right up to the point where it pinches out. Along its entire length it retains a similar lithostratigraphic position as the parent sill on the western limb. The tapering apophysis is overprinted by F_2 and F_3 folds and is thus an early structural feature. This strongly suggests it formed during F_1 boudinage of the sill, by flow of basal massive sulphides towards a boudin neck (e.g., Fig. 14). The local presence of sulphide mineralization in the sill possibly controlled location of the boudin neck.

Genesis of Ni Sulphide Mineralization in the Thompson Nickel Belt

The TNB hosts a large number of Ni sulphide deposits (Fig. 11), in a number of different environments (e.g., Figs. 4 and 14). The richest deposits, such as Thompson, Birchtree, and Pipe II, were generated by sills that intruded within the Ospwagan group ("high sulphur environment"), along either one of two sulphide facies iron formations. Common contacts between massive magmatic sulphides and sedimentary sulphides (Fig. 16) suggest a relationship of arrested *in-situ* assimilation of sulphidic country rocks. Magmatic sulphides in these deposits, distinguished by their relatively high Cr content (Bleeker, 1990c), show depleted Se/S ratios that are indistinguishable from or only marginally higher than those of the sedimentary sulphides (Fig. 17; Eckstrand *et al.*, 1989; Bleeker 1990c). Mixing calculations based on Se/S ratios indicate that 70-100% of the sulphide mass was derived, through assimilation, from the country rocks.

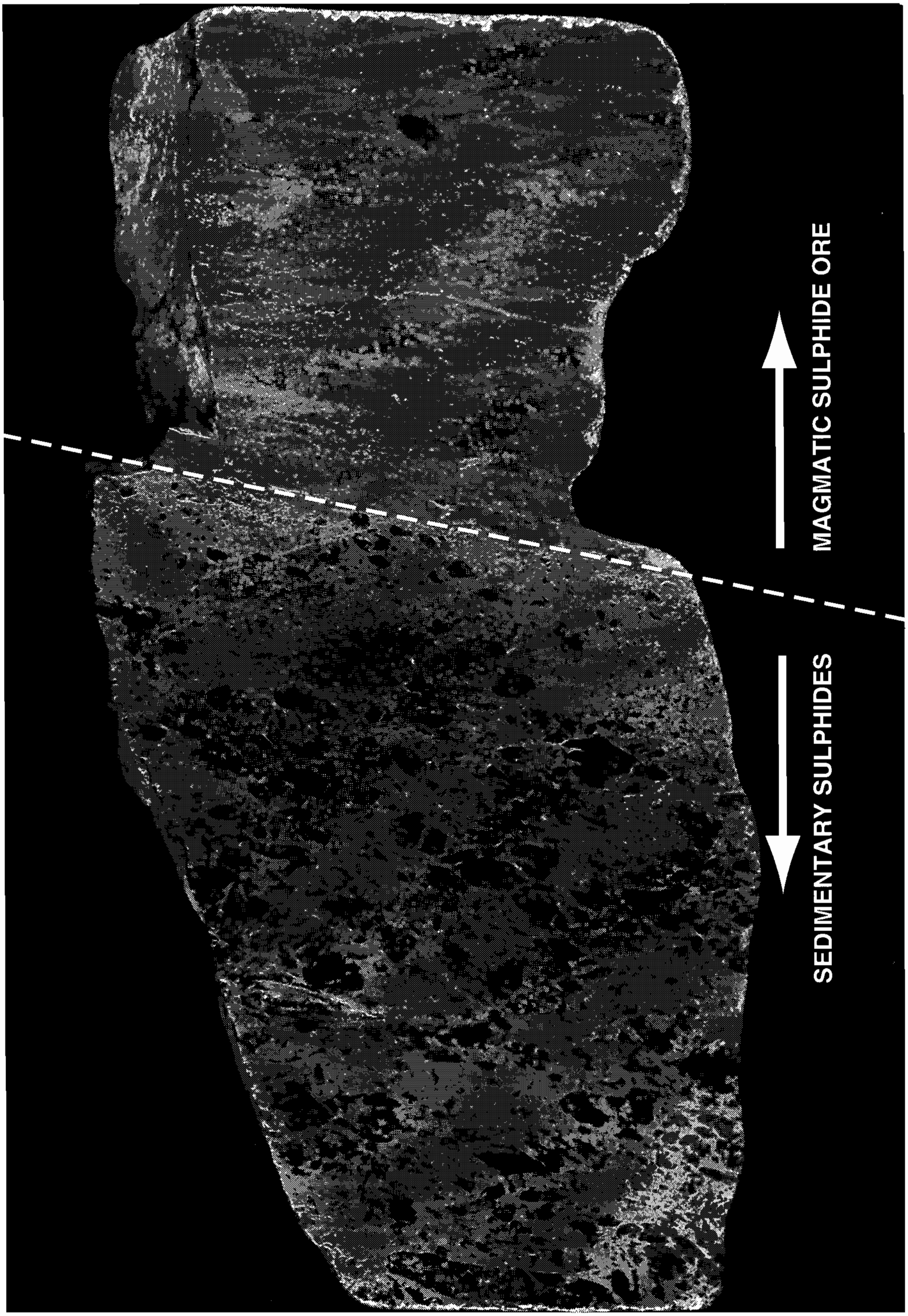
The Se/S ratios correlate with variable metal tenors that point at different R-factors, and hence somewhat different assimilation and sulphide separation histories for the various deposits (Fig. 18).

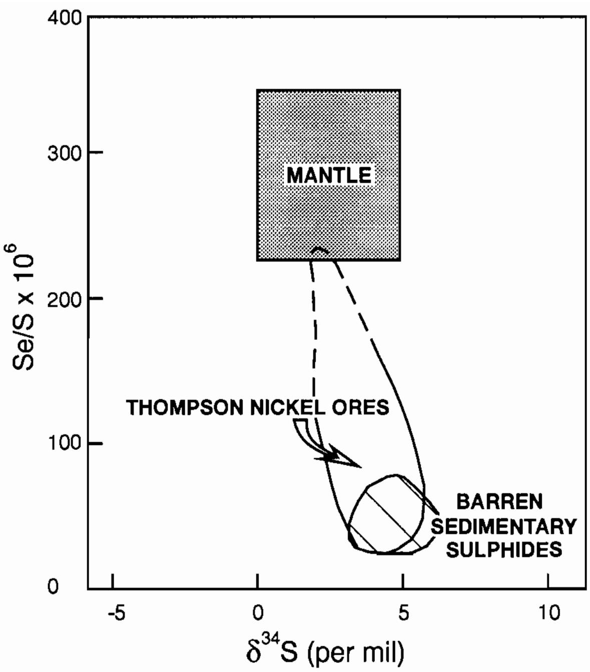
Massive magmatic sulphides are an important, but not the only ore type in the metasediment-hosted deposits. Ni-enriched sedimentary sulphides (Bleeker, 1990c) constitute a second important ore type (e.g., Fig. 16). Ni-enriched sedimentary sulphide ore formed by extensive redistribution of Ni and other metals, during high-grade metamorphism, from high-grade magmatic sulphides to previously barren sedimentary sulphides in the immediate country rocks (Fig. 19).

Figure 16 (next page): Contact between massive magmatic sulphides of the Pipe II orebody and sedimentary sulphides of the lower sulphide iron formation.

The highly graphitic sedimentary sulphides (below) consist of pyrrhotite and contain numerous inclusions of disrupted layers of graphitic schist and chert. The magmatic sulphides (above) are "clean" (i.e., few silicate and graphitic impurities; note however the small dark inclusion of dunite). Zn-rich chromite grains are a ubiquitous accessory in the magmatic sulphide ore, but are absent in the adjacent sedimentary sulphides.

Ni and Co concentrations have equilibrated across the contact during amphibolite facies metamorphism and, consequently, the originally barren sedimentary sulphides were transformed into Ni-enriched sedimentary sulphide ore. Contact relationships like these are interpreted as evidence for arrested assimilation (in situ) of sedimentary sulphides.


MAGMATIC SULPHIDE ORE
SEDIMENTARY SULPHIDES



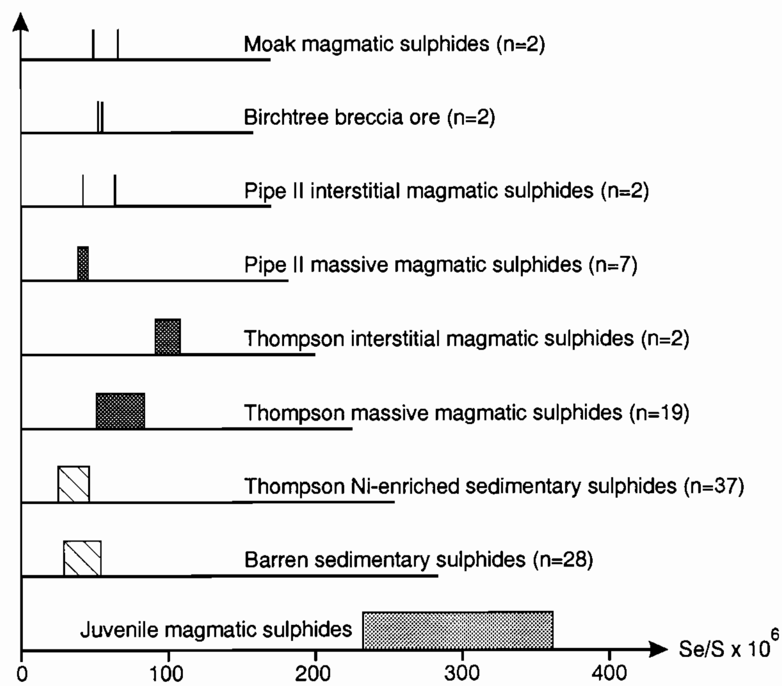


Figure 17: Se/S ratios for different sulphide types. Upper diagram is after Eckstrand et al. (1989) and shows the general "mixing trend" between the signature of barren sedimentary sulphides and values expected in uncontaminated mafic to ultramafic rocks ("mantle"). The lower diagram (from Bleeker, 1990c) shows Se/S ratios for a large number of samples from different orebodies as well as barren sedimentary sulphides. Note the general similarity in Se/S ratios between magmatic sulphides and barren sedimentary sulphides, suggesting bulk assimilation of country rocks was an important process. Also note the subtle differences in average Se/S ratios between different ore types at Thompson. Ni-enriched sedimentary sulphide ore shows an identical signature to its barren sedimentary precursor. Magmatic sulphide ores show slightly higher Se/S ratios, reflecting variable mixing ratios between mantle-derived magmatic sulphide and crustally derived assimilated sulphide.

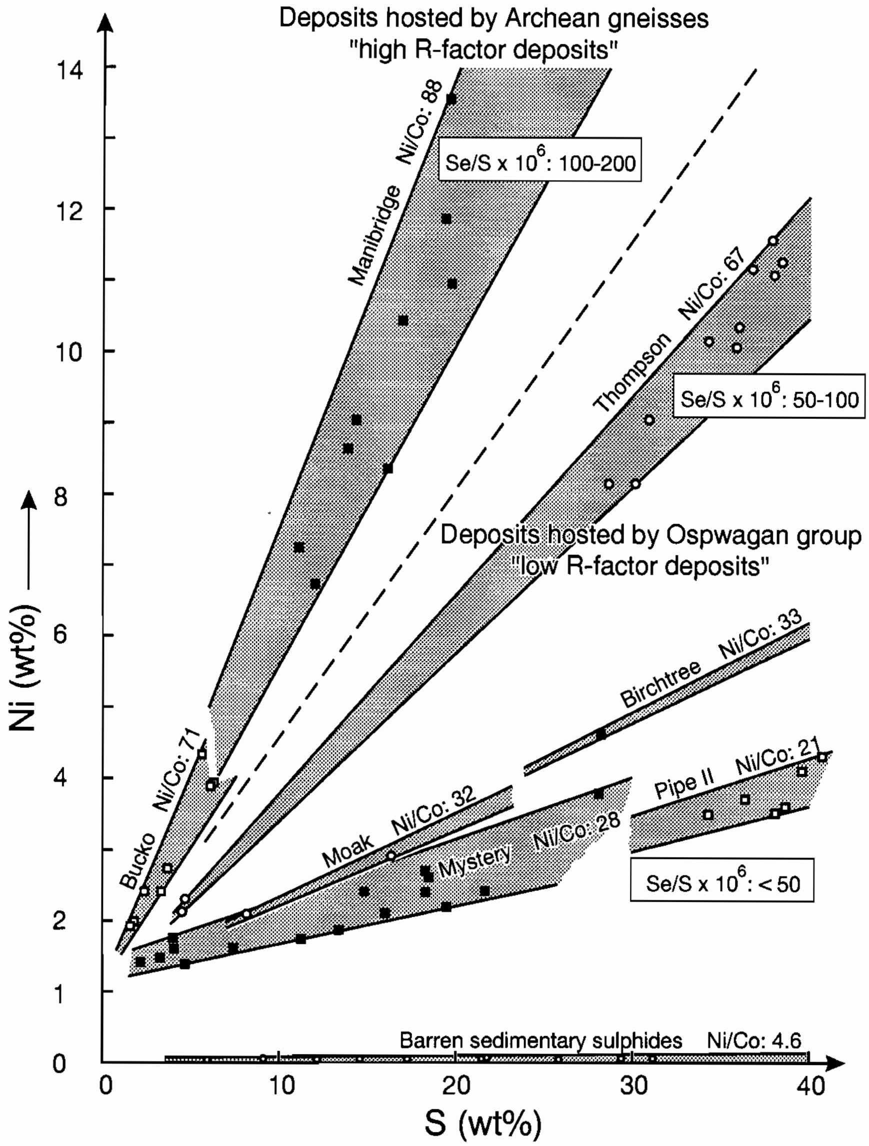


Figure 18: Nickel versus sulphur diagram for samples from barren sedimentary sulphides and a number of different deposits. Data points for magmatic sulphide ores from an individual deposit define a linear trend, the slope of which reflects the Ni tenor of the magmatic sulphides (weight percentage of Ni in 100% sulphides). Tenors vary strongly between the different deposits, reflecting different assimilation and sulphide separation histories and, thus, different R-factors. High tenor deposits (high R-factor), such as Manibridge and Bucko, occur within low-suphide environments such as the basement gneisses. Low tenor deposits (low R-factor) occur in the high-sulphide environment of the Ospwagan group. Variable tenors (and R-factors) correlate with a systematic variation in other geochemical parameters such as Ni/Co ratios and Se/S ratios.

Ni-enriched sedimentary sulphide ore is characterized by a general appearance that is identical to that of barren sedimentary sulphides, a low Cr content, variable Ni tenors, and anomalously low Ir/Pd ratios (Bleeker, 1990c; see also Paterson *et al.*, 1984).

PGE distribution patterns of this and other ore types are shown in Figure 20. Field evidence and mixing calculations based on Ni/Co ratios indicate that such Ni-enriched sedimentary sulphides constitute a considerable portion of the total ore tonnage, particularly at Thompson (Bleeker, 1990c). The driving force for the metasomatic Ni redistribution was the high concentration of Ni in magmatic sulphides relative to that in adjacent barren sedimentary sulphides, while both sulphide types transformed into the same monosulphide solid solution phase during high-grade metamorphism. Given the extent of the Ni redistribution halo, which ranges up to several ten's of meters to possibly hundred's of meters at Thompson, it is unlikely that solid state diffusion alone was a dominant transport mechanism. Only a metasomatic, fluid-assisted process appears capable of explaining the extent of the Ni halo. To some extent, the gain of Ni in Ni-enriched sedimentary sulphides is balanced by a corresponding loss in nearby magmatic sulphides. However, from a mass balance point of view, Ni loss in massive magmatic sulphides does not appear of a sufficient magnitude to explain all the Ni-enriched sedimentary sulphides. Although this Ni mass balance is difficult to quantify, it is suggested that additional Ni may have been supplied by serpentinization of the ultramafic rocks. Further work is needed on the exact details of the Ni metasomatism in environments such as Thompson.

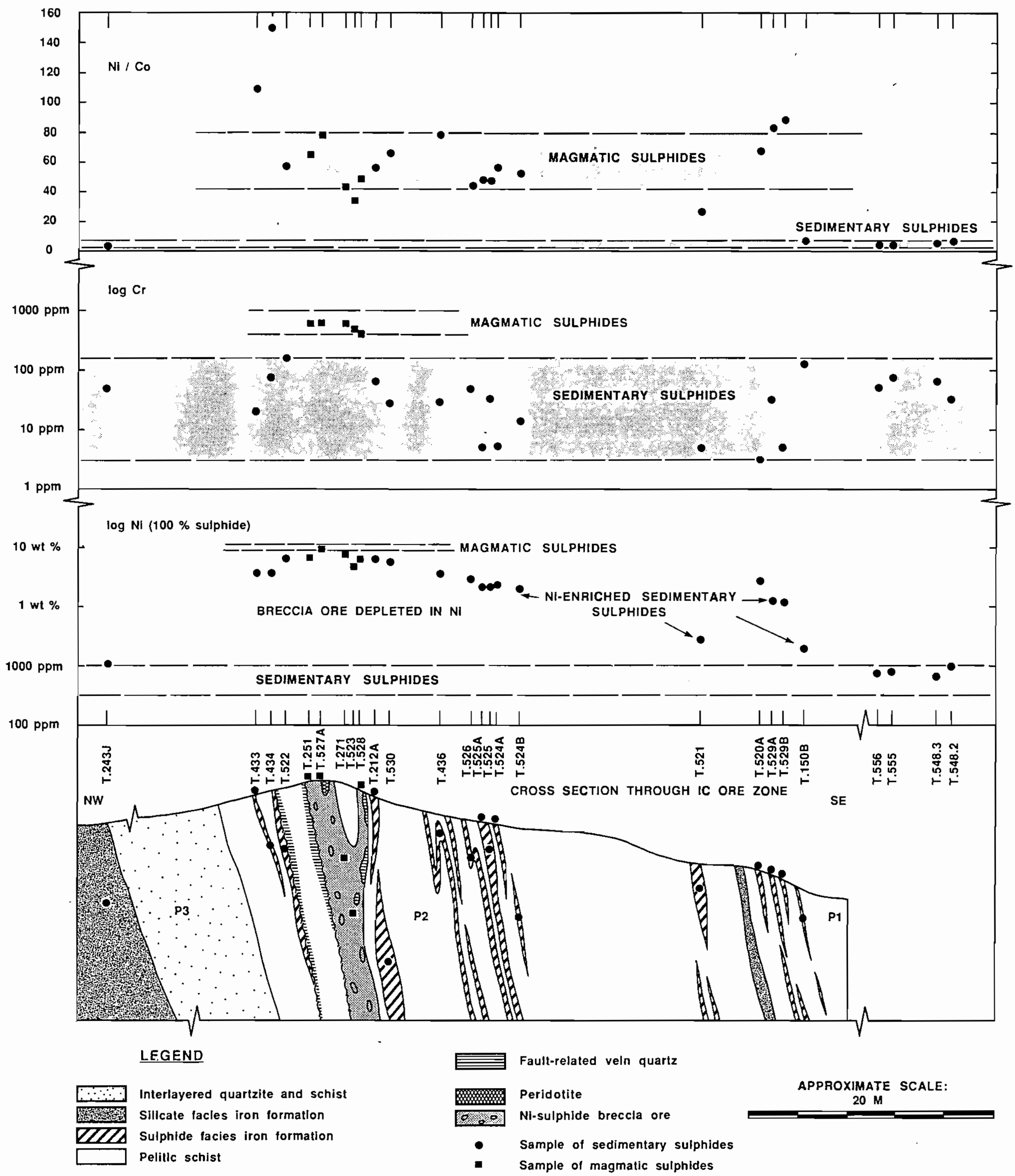
The relative importance of Ni-enriched sedimentary sulphide ores gives the Thompson deposit a unique character and explains the former controversy between strictly magmatic models and syn-sedimentary exhalative models. Depending on the location along the Thompson ore zone one can observe either the magmatic or the sedimentary end members of the complex spectrum of ore types (Bleeker, 1990c).

Deposits associated with ultramafic bodies hosted by Archean basement gneisses ("low-sulphur environment") are dominated by disseminated sulphides and show relatively high Se/S ratios. Although overall of low grade, due to their disseminated character, metal tenors in the interstitial sulphides are high (e.g., up to 20 wt% Ni on a 100% sulphide basis).

Table 1 lists representative analyses for some typical ore types as well as barren sedimentary sulphides.

Figure 19 (next page): Ni, Cr, and Ni/Co values in whole-rock sulphide samples (n=22) across the 1C ore zone at the north end of the 1C pit at Thompson. Sample locations of magmatic (solid squares) and sedimentary (solid dots) sulphide samples are indicated on the cross-section. Four barren sedimentary sulphide samples (on right side of diagram) were sampled at larger distances from the Thompson orebody.

Note (1) the extensive redistribution of Ni (and Co) from magmatic sulphides to adjacent sedimentary sulphides; (2) the inverse correlation of Ni-enrichment with distance from the main ore zone; (3) the minor Ni-depletion of magmatic sulphides relative to average magmatic sulphides at Thompson; (4) magmatic and sedimentary sulphides can still be distinguished on the basis for their Cr content; and (5) the peak in Ni/Co ratios on the edge of the Ni halo, suggesting greater mobility for Ni relative to Co.


Ni / Co
160
140
120
100
80
60
40
20
0
MAGMATIC SULPHIDES
SEDIMENTARY SULPHIDES
log Cr
1000 ppm
100 ppm
10 ppm
1 ppm
MAGMATIC SULPHIDES
SEDIMENTARY SULPHIDES
log Ni (100 % sulphide)
10 wt %
1 wt %
1000 ppm
100 ppm
MAGMATIC SULPHIDES
BRECCIA ORE DEPLETED IN NI
NI-ENRICHED SEDIMENTARY SULPHIDES
SEDIMENTARY SULPHIDES
T.243J
T.433
T.434
T.522
T.251
T.527A
T.271
T.523
T.528
T.212A
T.530
T.436
T.526
T.525A
T.525
T.524A
T.524B
T.521
T.520A
T.529A
T.529B
T.150B
T.556
T.555
T.548.3
T.548.2
CROSS SECTION THROUGH IC ORE ZONE
NW
SE
P3
P2
P1
LEGEND
Interlayered quartzite and schist
Silicate facies iron formation
Sulphide facies iron formation
Pelitic schist
Fault-related vein quartz
Peridotite
Ni-sulphide breccia ore
Sample of sedimentary sulphides
Sample of magmatic sulphides
APPROXIMATE SCALE:
20 M

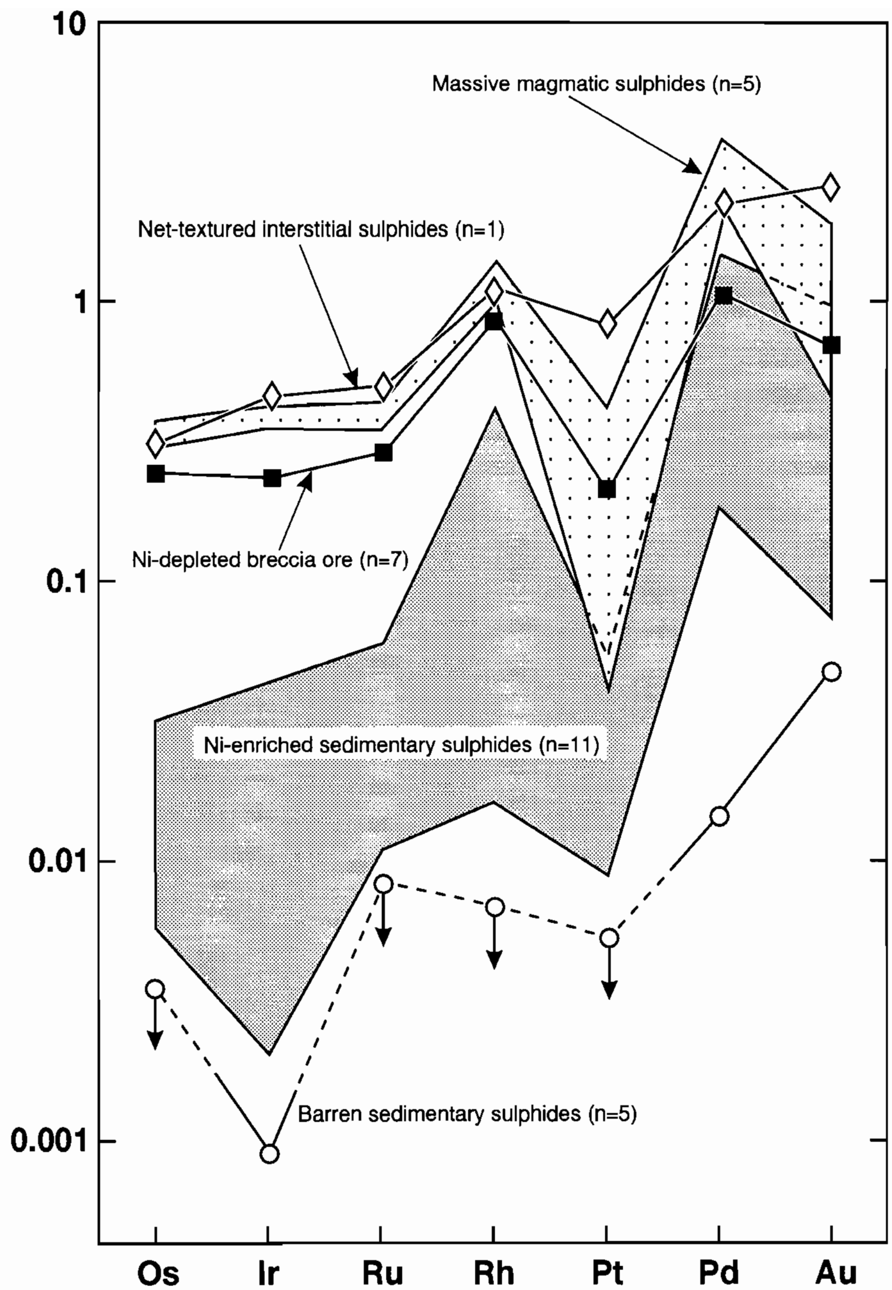


Figure 20: Chondrite-normalized PGE-Au patterns for various ore types at Thompson: net-textured interstitial magmatic sulphides (n=1); massive magmatic sulphides (n=5); Ni-depleted breccia ore (n=7); Ni-enriched sedimentary sulphides (n=11); and barren sedimentary sulphides (n=5). Ranges for massive magmatic sulphides and Ni-enriched sedimentary sulphides indicate average values plus or minus one standard deviation. In barren sedimentary sulphides, abundances for Os, Ru, Rh, and Pt are below detection limits; hence, indicated values are maxima. Note the low Ir abundances in sedimentary sulphides and the intermediate characteristics of Ni-enriched sedimentary sulphides. The high Pd/Ir ratio (steeper slope) is typical for Ni-enriched sedimentary sulphide ore and indicates relative immobility of Ir in association with the Ni redistribution (after Bleeker, 1990c).

Table 1: Representative analyses of barren sedimentary sulphide samples and ore samples[1]

		S wt%	C wt%	Ni ppm	Cu ppm	Co ppm	Zn ppm	Cr[2] ppm	Ni/S ($x10^4$)	Ni/Cu	Pd ppb	Ir ppb	Pd/Ir[3]	Se/S ($x10^6$)
Barren sedimentary sulphides:														
O.66A	Ospwagan Lake	29.6	4.2	370	340	53	980	<10	12.5	1.1	14	0.4	35	21
O.66B	Ospwagan Lake	30.8	3.6	360	400	62	580	<10	11.7	0.9	--	--	--	19
P.63	Pipe II open pit	19.6	3.9	200	330	40	1100	37	10.2	0.6	12	1.1	11	42
P.79	Pipe II open pit	6.0	2.0	120	140	32	160	85	20.0	0.9	--	--	--	42
T.150B	Thompson	6.8	3.7	330	240	48	88	130	48.5	1.4	--	--	--	37
T.243J	Thompson	9.2	1.1	260	370	74	91	51	28.3	0.7	--	--	--	25
T.548-3	Thompson	19.0	14.0	330	1000	63	910	33	17.4	0.3	4	0.2	20	71
Thompson, Ni-enriched sedimentary sulphide ore (samples ordered by increasing Ni enrichment):														
T.521		34.4	3.9	2500	900	93	63	<10	72.7	2.8	60	0.9	67	17
T.525		23.1	7.1	13000	1200	270	190	19	563	10.8	69	1.1	63	37
T.436[4]		19.9	4.0	18000	700	230	190	29	905	25.7	450	3.3	136	38
T.425A		32.9	1.9	52000	4000	660	280	<10	1581	13.0	990	5.1	194	53
T.512C		27.6	5.6	66000	6000	1200	310	140	2391	11.0	600	28	21	36
T.514		29.4	4.8	75000	3100	1200	340	28	2551	24.2	350	1.5	233	29
T.549		32.2	3.5	105000	8800	1100	420	44	3261	11.9	2070	19	109	21
T.426		31.6	3.1	109000	5600	870	470	<10	3449	19.5	1750	2.7	648	23

Footnotes:

[1] Data taken from Bleeker (1990c).

[2] Note uniformly low Cr content of barren and Ni-enriched sedimentary sulphides (<200 ppm).

[3] Note the high Pd/Ir ratios in many of the Ni-enriched sedimentary sulphide ore samples, indicating low initial Ir concentrations in the sedimentary sulphide material, and relative immobility of Ir. Pd behaved much more mobile and its abundance correlates strongly with that of Ni.

[4] This sample is shown in Figure d of the Frontipiece

Table 1: Continued

		S wt%	C wt%	Ni ppm	Cu ppm	Co ppm	Zn ppm	Cr ppm	Ni/S ($x10^4$)	Ni/Cu	Pd ppb	Ir ppb	Pd/Ir	Se/S[7] ($x10^6$)
Thompson, interstitial and massive magmatic sulphides:														
T.372[5]	interstitial	4.7	0.6	13000	270	270	280	590	2766	48.1	330	27	12	92
T.188B[6]	massive	36.8	0.1	101000	3800	1500	500	500	2744	26.6	4200	200	21	113
T.296	massive	35.9	0.0	93000	5000	1500	700	1200	2590	18.6	2900	160	18	68
T.534	massive	35.7	0.1	98000	3700	1300	560	930	2745	26.5	4770	160	30	48
T.512A	massive	38.5	0.1	102000	3900	1400	580	610	2649	26.2	3000	200	15	88
T.512B	massive	38.0	0.0	105000	5300	1500	550	560	2763	19.8	2520	170	15	87
Pipe II, interstitial and massive magmatic sulphide ore:														
P.32	interstitial	2.8	0.1	5100	230	230	87	3200	1821	22.2	26	6.5	4	65
P.29	massive	39.8	0.2	31000	2500	1300	150	690	779	12.4	140	30	5	43
P.37A	massive	38.8	0.7	26000	400	1200	160	810	670	65.0	52	28	2	43
P.89	massive	38.4	0.0	24000	2100	1200	200	870	625	11.4	38	33	1	39

Further footnotes:

[5] Photomicrograph of this sample is shown in Figure a of the Frontipiece.

[6] Sample is shown in Figure b of the Frontipiece.

[7] Note the low Se/S ratios of Pipe II magmatic sulphides, which are indistinguishable from those of barren sedimentary sulphides. This indicates that essentially 100% of the magmatic sulphide material was derived by bulk assimilation of adjacent sedimentary sulphides. Se/S ratios in magmatic sulphides at Thompson are slightly more elevated, indicating a higher magmatic (S and Se) input. Se/S ratios correlate with Ni tenors.

EXCURSION STOPS

Day 1: Sipiwesk Lake, Eastern Boundary of the TNB

Day 1 of the fieldtrip will be spend on Sipiwesk Lake, which straddles the eastern boundary of the TNB (Fig. 21). A number of stops will be made, along a roughly east-to-west transect across the boundary, to highlight the following points of interest:

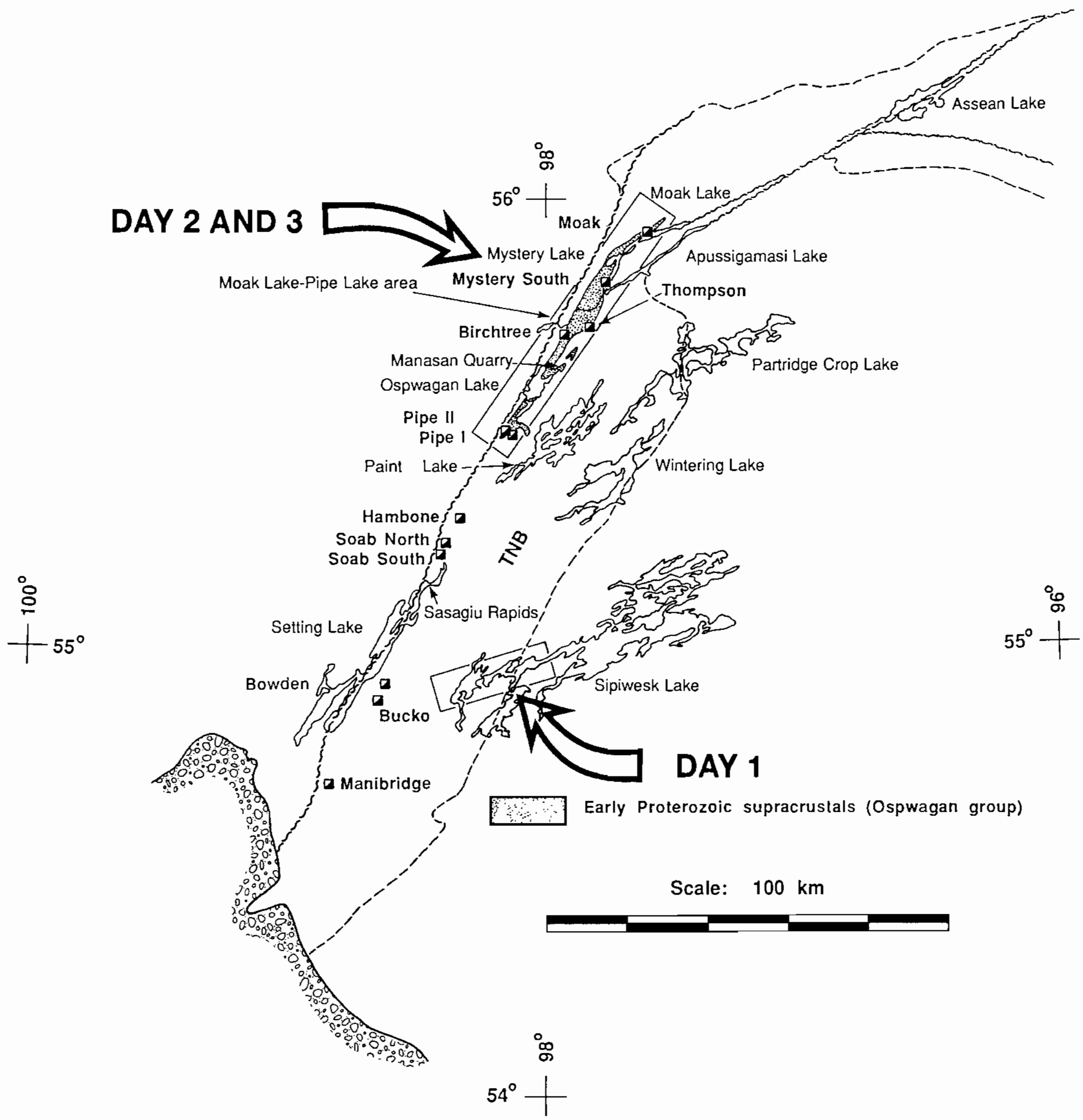


Figure 21: Location map. Day 1 of the field trip will be spend on Sipiwesk Lake, straddling the eastern boundary of the TNB. Several stops will be made across the "Hudsonian front". Day 2 and 3 will focus on the Moak Lake-Pipe Lake area, further north along the TNB, to examine the geological relationships between basement gneisses, the Ospwagan group cover sequence, and the Thompson and Pipe II nickel sulphide deposits.

1. Unreworked 2.69 Ga granulites of the Pikwitonei domain

Pikwitonei granulites consist of granulite facies tectonites of felsic to intermediate and mafic composition, with minor amounts of ultramafic rocks. Strongly stretched and disrupted mafic and ultramafic bands (enclaves) occur within more dioritic to tonalitic host gneisses. Some of the mafic granulites can be identified as (layered) gabbroic complexes, whereas the felsic to intermediate host gneisses are probably tonalitic to dioritic intrusives. Unequivocal supracrustal rocks (e.g., mafic volcanics) have not been identified on this part of Sipiwesk Lake. The strong foliation is variable in orientation but generally dips steeply towards the southeast. Since Pikwitonei granulites can be mapped far into the TNB, an interesting question is how much of the steeply east-dipping seismic reflection fabric, as observed along Lithoprobe Line #2 (Fig. 10), is actually Kenoran in age rather than Hudsonian.

2. The Molson dyke swarm, dated at 1883±2 Ma

Abundant Molson dykes can be seen on Sipiwesk Lake, varying in width from less than 10 cm to more than 100 m. The dykes are important because they provide time markers for deformational and metamorphic effects associated with the TNB boundary. Most dykes are approximately vertical and no systematic tilt of the swarm is observed across the boundary. Trends are mostly northeast although northwesterly trends are observed as well. Molson dykes are cut by Early Proterozoic pegmatites and both sets of intrusions are deformed in the frontal shear zones of the TNB.

3. First appearance of Early Proterozoic structural effects

Minor vein-filled fractures are seen east of the front, whereas the front itself is relatively abrupt and marked by an increasing number of amphibolite facies shear zones.

4. The appearance of coronitic garnets in two-pyroxene granulites

To the east of first frontal shear zones, small, pin-head size, coronitic garnets developed in intermediate to mafic two-pyroxene granulites. Garnet development is a static and anhydrous metamorphic response to the Early Proterozoic metamorphism of the Archean two-pyroxene granulites. That much of the pin-head garnets are due to Proterozoic overprinting, and not to Late Archean isobaric cooling, is shown by the occurrence of similar garnets in Molson dykes (e.g., Bleeker, 1990a; some coarser coronitic garnet that occurs in one of the metagabbros is possibly Late Archean in age). Garnet growth is due to the pressure sensitive reaction:

orthopyroxene + calcic plagioclase → garnet + clinopyroxene + quartz

Quartz produced during the reaction occurs as vermicular intergrowths with the garnets. Thermobarometric calculations on the garnet-bearing assemblage suggest the eastern boundary zone was subjected to conditions of about 550°C and about 5 kbar. This metamorphism is attributed to "loading" of the granulite facies foreland during the Early Proterozoic. The actual frontal shear zones and hydrous retrogression appear to overprint the garnet-bearing assemblages. Across the boundary, garnet growth is becoming more pronounced in domains that escaped influx by hydrous fluids. Locally, mafic granulites have been transformed into spectacular garnet-clinopyroxene (+plagioclase) rocks, which are called, informally, "pseudo-eclogites".

5. Static hydrous retrogression of the granulites
Where hydrous fluids were introduced into the granulites, the rocks are retrogressed to hornblende+biotite-bearing rocks. A pronounced colour change accompanies this retrogression from the rusty yellowish colour typical for granulites to grey or black for the amphibolite facies retrogressive products. Along the front, hydrous fluids were introduced along discrete fractures, along shear zones, and together with pegmatites. During the hydrous retrogression, magnetite tends to react to biotite with a correlating drop in rock magnetic susceptibility. The magnetic quiet zone which marks the eastern boundary of the TNB is attributed to this pervasive retrogression.

6. First appearance of subhorizontally intruding pegmatites
Near the frontal shear zones, 10-50 cm thick, pink pegmatites appear. They typically show a horizontal intrusive attitude, compatible with intrusion into a thrust regime (σ_3 is vertical). An attempt to date these pegmatites has been unsuccessful due to lack of zircon and monazite. The pegmatites are deformed in the frontal shear zones.

7. First appearance of somewhat younger, red aplite dykes
Somewhat further west, red aplite dykes are seen to intrude the granulites, Molson dykes, and the pink pegmatites. These aplites are probably a late expression of the same magmatism that produced the pegmatites. This magmatism possibly can be correlated with intrusion of the Wintering Lake granodiorite, a large pluton further north along the boundary zone. Machado et al. (1987) report an 1822±3 Ma monazite age for this pluton.

8. Dynamic retrogression of the granulites along Early Proterozoic shear zones
Significant shear zones can be mapped along the front and further west into the TNB. These shear zones show variable trends but are on average parallel to the eastern boundary of the TNB. Dips of the shear zones are variable, but lineations are invariably down-dip. Several of the shear zones dip shallowly towards the west and can be described as frontal thrust along which the TNB appears to have ramped up over its foreland. In the shear zones, granulites were retrogressed and dynamically recrystallized to strongly foliated hornblende+biotite+plagioclase gneisses.

Day 2: Thompson and Pipe II Open Pits

Day 2 will start with a visit to one of the mined out open pits at Thompson (the South Pit). The remainder of the day will be spend at the Pipe II open pit. At both localities, the main focus will be on the lithostratigraphy and deformation history of the Early Proterozoic cover sequence, the Ospwagan group.

Day 2, early morning: Thompson open pit

Stop 2-1: Shoulder of the South Pit of the Thompson Open Pit

The shoulder of the South Pit, if accessible, has one of the better exposures of the contact between basement gneisses and the Ospwagan group cover sequence. A weak angular relationship between the gneissic structure of the basement and the overlying metasediments is preserved locally. At the base of the Ospwagan group, small lenses of quartz pebble conglomerate are developed along the unconformity, overlain by the basal quartzite member of the Manasan formation (M1 member). Quartzites grade up into semipelitic gneisses (M2 member), which in turn are overlain by the calc-silicate gneisses of the Thompson formation. Note the disposition of the Ospwagan group around the hinge of the major south-plunging F_3 antiform (as outlined by the shape of the open pits) and the overall downward facing nature of the structure.

Day 2, morning and afternoon: Pipe II open pit

Pipe II open pit is unique in that it exposes the most complete section through basement and cover sequence of the Thompson Nickel Belt. In spite of intense polyphase deformation many primary features are preserved allowing reconstruction of the lithostratigraphy (Fig. 12) and the detailed setting of the Pipe II ore body. Better than average preservation of the Pipe II sequence is probably due to its location on, or close to, hinges of not only F_3, the main mine-scale fold, but also F_1 and F_2. Secondly, metamorphic conditions are at staurolite grade, hence 50-100° lower than at Thompson. After an introduction during which maps and polished hand samples will be examined, we will walk a traverse from basement gneisses to volcanic rocks at the top of the Ospwagan group.

Stop 2-2: Strongly reworked basement gneisses intruded by Molson dykes

The grey gneisses show a Hudsonian transposition foliation. Although grain size is greatly reduced relative to Archean protoliths, a coarse relict texture is still visible locally. Note the reasonable preservation of the mafic dykes in spite of the extreme strain in the basement gneisses. Mafic dykes like this occur throughout the western Thompson Nickel Belt and are chemically indistinguishable from Molson dykes.

Stop 2-3: Gneiss-basal quartzite contact and Manasan formation

Examination of the lower transgressive sequence comprising: (1) a thin and local pebbly conglomerate; (2) quartzite with beds that preserved graded bedding; (3) upward-fining laminated quartzites; and (4) semipelitic schists. Note the characteristic dark reddish colour and the microcline-rich sweats of the latter lithology. Exceptional preservation of the basal quartzite can be attributed to its location on a F_2 fold hinge.

Stop 2-4: Manasan to Thompson formation

Transition from semipelitic schist, via cherty layers, to impure calcareous metasediments, marble and sulphidic chert (Thompson formation and P1 member of Pipe formation). The sulphidic chert, rather thin in these outcrops, are laterally continuous with nearly massive pyrrhotite. It is this horizon that must have supplied the bulk of assimilated sulphur to the Pipe II sill. Note the strong difference between this carbonate-rich sequence and that of Thompson, reflecting lateral facies changes as well as the large original separation between the two localities.

Stop 2-5: First silicate facies iron formation
This unit (remainder of P1 member of Pipe Formation) directly overlies the intrusive horizon of the Pipe II sill. Some members of this iron formation are highly magnetic, some show regular chert banding, others are massive. It is distinct from silicate facies iron formations higher up in the sequence and grades via white and red cherty units into pelitic schists.

Stop 2-6: Pelitic schists and upper sulphide facies iron formation (P2 member of Pipe formation)
The upper sulphide facies iron formation occurs near the top of the pelitic schists and can be correlated with sedimentary sulphides near the top of the pelitic schists at Thompson.

Stop 2-7: Main silicate facies iron formation (P3 member of Pipe formation)
Siliceous metasediments and the main cycle of silicate facies iron formations (P3 member of Pipe formation). This package of rocks is extremely well preserved, as illustrated by the fact that many centimeter to decimeter-wide layers can be followed throughout the entire eastern shoulder of the pit. Subtle differences allow a succession of chert-banded iron formations to be recognized, which is interrupted by a clean dolomitic marble and capped by thinly-banded diopside rocks and chert. Note that the succession of iron formations shows a trend from laminated, garnetiferous varieties in the lower part to thinly banded, non-garnetiferous, pyroxene-bearing varieties above the dolomitic marble. The same compositional trend appears to be present in Thompson, and probably reflects a progressive decrease of a residual Al-rich pelitic component. The sequence of iron formations and dolomitic marble shows several large-scale, tight to isoclinal F_1 folds, which are cross-cut by a dense swarm of Molson dykes (see Fig. 9). The post-F_1 nature of these dykes will be demonstrated.

Stop 2-8: Further into the core of the Pipe F_3 synform
A large outcrop, isolated from previous localities, exposes a succession of clastic and chemical sediments, which structurally repeats previously observed lithostratigraphy. The structurally highest part of the outcrop adds however a section of stratigraphy which correlates with the "core quartzites" at Thompson (S1 member of Setting formation).

Stop 2-9: Ospwagan group volcanics (O1 and O2 members of the Ospwagan formation)
Further into the core (300 meters) of the Pipe II synform, the local stratigraphic top of the Ospwagan group comprises mafic, heterolithic volcanic breccias overlain by basaltic pillow lavas.

Stop 2-10: Pipe II ultramafic sill and its upper contact
Exposures on the western limb of the Pipe II synform show the upper part of the ultramafic sill and its contact with overlying silicate facies iron formation. In spite of serpentinization and tremolitization of the ultramafic rocks, primary textures are locally well preserved. A trend from dunite to orthopyroxenite can be observed. Relict textures in the orthopyroxenite show a decrease in grain size towards the upper contact, which is marked by a talc schist envelope. Metasediments overlying the sill will be examined, to demonstrate the stratigraphic position of the sill.

Stop 2-11: Examination of Pipe II Open Pit from observation booth
This locality offers an excellent view of the geology in the pit walls, especially the massive sulphide apophysis in the southeastern wall. Note the tapering and conformable nature of the apophysis.

Day 3: Ospwagan Lake, Manasan Quarry, and Drill Core Display

Day 3, early morning: basal unconformity on Ospwagan Lake

Stop 3-1: Western shore of Lower Ospwagan Lake
Across the lake, on its western shoreline, several exposures of the (deformed) basal unconformity exist. One of these shows an excellent section of highly reworked basement gneisses overlain by clastic rocks of the Manasan formation: a thin pebbly conglomerate, quartzite, and semipelitic schist. The unconformity is tightly folded by upward-facing F_3 folds.

Day 3, basal unconformity and lower part of the Ospwagan group, Manasan Quarry

Stop 3-2: Stratigraphy and fold structure at the Manasan Quarry, eastern limb
Manasan Quarry is situated 15 km south-southwest of Thompson, and like the Thompson open pit, on the overturned "Thompson limb" of the F_1 nappe (Fig. 5). The quarry exposes basement gneisses, an up to 50 m thick package of basal quartzites, and stratigraphically overlying metasediments in a downward facing, northeast-plunging F_3 synform. The rather mature basal quartzite, with up to 90 wt% SiO_2, is used as a flux in smelting and converting processes at the Thompson plant (e.g., Boldt and Queneau, 1967). The eastern limb of Manasan synform exposes the basal unconformity, along which pebbly conglomerate and quartzite overlie basement gneisses. The basal conglomerate is about 30 cm thick and consists of quartz pebbles in a dark, amphibole-rich matrix. Originally, this matrix was probably a calcareous cement. Concentrations of apatite have also been

observed in the matrix. A few pebbly conglomerate layers occur higher up within the quartzites. On the south side of the quarry, quartzites (M1 member) are overlain by semipelitic schists (M2 member). F_{1or2}-F_3 interference can be demonstrated in the semipelitic schists. Note that the S_3 axial plane trends 050°, clockwise from the trend of the Thompson Nickel Belt. This F_3 fold trend is typical and forms an *en échelon* pattern, which dominates the late structure of the belt.

Stop 3-3: Stratigraphy and fold structure at the Manasan Quarry, western limb
On the western limb of Manasan synform, at the northwest end of the quarry, a section will be examined that exposes from east to west: (1) reworked basement gneisses, (2) white, grey or pink basal quartzites with a few quartz pebble conglomerate layers near its base, (3) semipelitic schists, and (4) a thick package of banded calc-silicate gneisses and minor marble. Basement gneiss at this locality comprises an orthogneiss of intermediate composition with mafic xenoliths. The thin basal conglomerate layers are not easily recognized due to stretching of the pebbles and green amphibole growth in the matrix. In general, the basal quartzite may contain considerable amounts of feldspar and mica, but locally grades into orthoquartzite. Semipelitic schist, the uppermost unit of the lower clastic transgressive sequence, forms a thin band on this limb. Note the characteristic colour of this unit and the presence of highly stretched microcline-rich sweats. Banded calc-silicate gneisses or "skarns" at this locality are very similar to comparable rocks at Thompson, illustrating the original proximity of the two localities. On the lower "Pipe limb" of the nappe, the skarn and marble sequence is quite different in character or lacking.

Day 3, afternoon: drill core and sample display and wind up of field trip
Several drill cores will be displayed from some typical sections through both INCO Ltd. and Falconbridge Ltd. deposits.

Formal end of field trip.

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REFERENCES

Ansdell, K.M., Lucas, S.B., Connors, S.B., and Stern, R.A., 1995. Kisseynew metasedimentary gneiss belt, Trans-Hudson orogen (Canada): back-arc origin and collisional inversion. Geology, v. 23, p. 1039-1043.

Baragar, W.R.A., and Scoates, R.F.J., 1981. The Circum-Superior belt: a Proterozoic plate margin?. *In* Kröner, A., ed., Precambrian Plate Tectonics, Elsevier, Amsterdam, p. 297-330.

Bell, C.K., 1966. Churchill-Superior boundary in northeastern Manitoba. Geological Survey of Canada, Paper 66-1, p. 133-136.

Bell, C.K., 1971. Boundary geology, upper Nelson River area, Manitoba and northwestern Ontario. *In* Turnock, A.C., ed., Geoscience Studies in Manitoba, Geological Association of Canada, Special Paper, no. 9, p. 11-39.

Bleeker, W., 1989. Litho-structural map of the Thompson Open Pit, Thompson Nickel Belt, Manitoba, scale 1:1200, with accompanying map description. Geological Survey of Canada, Open File 2089.

Bleeker, W., 1990a. New structural-metamorphic constraints on Early Proterozoic oblique collision along the Thompson Nickel Belt, northern Manitoba, Canada. *In* Lewry, J.F., and Stauffer, M.R., eds., The Early Proterozoic Trans-Hudson Orogen of North America, Geological Association of Canada, Special Paper 37, p. 57-73.

Bleeker, W., 1990b. Thompson Area—General Geology and Ore Deposits. *In* Galley, A.G., Bailes, A.H., Syme, E.C., Bleeker, W., Macek, J.J., and Gordon, T.M., eds., Geology and Mineral Deposits of the Flin Flon and Thompson Belts, Manitoba, Field Trip Guidebook, 8th IAGOD Symposium, Geological Survey of Canada, Open File 2165, p. 93-136.

Bleeker, W., 1990c. Evolution of the Thompson Nickel Belt and its Nickel Deposits, Manitoba, Canada. Unpublished Ph.D. thesis, University of New Brunswick, Fredericton, New Brunswick, 400 p.

Bleeker, W., and Macek, J.J., 1988a. Thompson Nickel Belt Project: Pipe Pit Mine. *In* Report of Field Activities 1988, Manitoba Energy and Mines, p. 111-115.

Bleeker, W., and Macek, J.J., 1988b. Pipe Pit Mine (Sheets A,B,C,D,E,F,G and H), 1:400. Manitoba Energy and Mines, Geological Services, Preliminary Maps 1988T-1 to 1988T-8.

Bleeker, W., Nägerl, P., and Machado, N., 1995. The Thompson Nickel Belt, Manitoba: some new U-Pb ages. Geological Association of Canada, Program with Abstracts, v. 20, p. A8.

Boldt, J.R., and Queneau, P., 1967. The winning of Nickel. Longmans Canada Limited, Toronto, Ontario, 487 p.

Brooks, C., Theyer, P., 1981. Rb/Sr geochronology in the Thompson Nickel Belt, Manitoba: implication for Aphebian crustal development and metallogenesis. Canadian Journal of Earth Sciences, v. 18, p. 932-943.

Chandler, F.W., and Parrish, R.R., 1989. Age of the Richmond Gulf Group and implications for rifting in the Trans-Hudson Orogen, Canada. Precambrian Research, v. 44, p. 277-288.

Chauvel, C., Arndt, N.T., Kielinzcuk, S., and Thom, A., 1987. Formation of 1.9 Ga old continental crust. I: Nd isotopic data. Canadian Journal of Earth Sciences, v. 24, p. 396-406.

Coats, C.J.A., Brummer, J.J., 1971. Geology of the Manibridge nickel deposit, Wabowden, Manitoba. *In* Turnock, A.C., ed., Geoscience Studies in Manitoba, Geological Association of Canada, Special Paper no. 9, p. 155-165.

Coats, C.J.A., Quirke, T.T., Bell, C.K., Cranstone, D.A., and Campbell, F.H.A., 1972. Geology and mineral deposits of the Flin Flon, Lynn Lake and Thompson areas, Manitoba, and the Churchill-Superior front of the western Precambrian shield. Guidebook, Field Excursion A31-C31, International Geological Congress, XXIV Session, Canada.

Cranstone, D.A., and Turek, A., 1976. Geological and geochronological relationships of the Thompson Nickel Belt, Manitoba. Canadian Journal of Earth Sciences, v. 13, p. 1058-1069.

Eckstrand, O.R., Grinenko, L.N., Krouse, H.R., Paktunc, A.D., Schwann, P.L., and Scoates, R.F.J., 1989. Preliminary data on sulphur isotopes and Se/S ratios, and the source of sulphur in magmatic sulphides from the Fox River Sill, Molson Dykes and Thompson nickel deposits, northern Manitoba. *In* Current Research, Part C, Geological Survey of Canada, Paper 89-1C, p. 235-242.

Ermanovics, I., and Fahrig, W.F., 1975. The petrochemistry and paleomagnetism of the Molson dykes, Manitoba. Canadian Journal of Earth Sciences, v. 12, p. 1564-1575.

Fraser, H.S., 1985. The Great Thompson Nickel Discovery. Inco Limited, Manitoba, 388 p.

Fueten, F., and Robin, P.Y.F., 1989. Structural petrology along a transect across the Thompson Belt, Manitoba: dip-slip at the western Churchill-Superior boundary. Canadian Journal of Earth Sciences, v. 26, p. 1976-1989.

Gibb, R.A., 1968. A geological interpretation of the Bouguer anomalies adjacent to the Churchill-Superior boundary in northern Manitoba. Canadian Journal of Earth Sciences, v. 5, p. 439-453.

Green, A.G., Cumming, G.L., and Cedarwell, D., 1979. Extension of the Superior-Churchill boundary zone into southern Canada. Canadian Journal of Earth Sciences, v. 16, p. 1691-1701.

Green, A.G., Hajnal, Z., and Weber, W., 1985. An evolutionary model of the western Churchill province and western margin of the Superior province of Canada and the north-central United States. Tectonophysics, v. 116, p. 281-322.

Heaman, L.M., Machado, N., Krogh, T., and Weber, W., 1986. Precise U-Pb zircon ages for the Molson dyke swarm and the Fox River sill: implications for Early Proterozoic crustal evolution in NE Manitoba, Canada. Contributions to Mineralogy and Petrology, v. 94, p. 82-89.

Hoffman, P.F., 1988. United Plates of America, the birth of a craton: Early Proterozoic assembly and growth of Laurentia. Annual Review of Earth and Planetary Sciences, v. 16, p. 543-603.

Hoffman. P.F., 1989. Precambrian geology and tectonic history of North America. In Balley, A.W., and Palmer, A.R., eds., The geology of North America—an overview, The geology of North America, Volume A, Geological Society of America, Boulder, Colorado, p. 447-512.

Hopkins, J.L., 1986. Live at twenty-five. Canadian Mining Journal, July 1986, p. 29-31.

Hubregtse, J.J.M.W., 1980. The Archean Pikwitonei granulite domain and its position at the margin of the northwestern Superior Province, central Manitoba. Manitoba Mines Branch, Geological Paper 80-3, 16 p.

Hulbert, L., Stern, R., Kyser, T.K., Pearson, J., Lesher, M., and Grinenko, L., 1994. The Winnipegosis komatiite belt, central Manitoba. *In* Manitoba Mining, Minerals and Petroleum Convention '94, Manitoba Energy and Mines, p. 21.

Kornik, L.J., 1969. An aeromagnetic study of the Moak Lake-Setting Lake structure in northern Manitoba, Canadian Journal of Earth Sciences, v. 6, 373-381.

Kornik, L.J., and MacLaren, A.S., 1966. Aeromagnetic study of the Churchill-Superior boundary in northern Manitoba. Canadian Journal of Earth Sciences, v. 3, p. 547-557.

Lewry, J.F., 1981. Lower Proterozoic arc-microcontinent collisional tectonics in the western Churchill Province. Nature, v. 294, p. 69-72.

Lewry, J.F., Macdonald, R., Livesey, C., Meyer, M., Van Schmus, R., and Bickford, M.E., 1987. U-Pb geochronology of accreted terranes in the Trans-Hudson Orogen, northern Saskatchewan, Canada. *In* Pharaoh, T.C., R.D. Beckinsale and D. Rickard, eds., Geochemistry and Mineralization of Proterozoic Volcanic Suites, Geological Society Special Publication no. 33, p. 147-166.

Lewry, J.F., Sibbald, T.I.I., and Schledewitz, D.C.P., 1985. Variation in character of Archean rocks in the western Churchill Province and its significance. *In* Ayres, L.D., P.C. Thurston, K.D. Card and W. Weber, eds., Evolution of Archean Supracrustal Sequences, Geological Association of Canada, Special Paper 28, p. 239-261.

Lucas, S.B., Green, A., Hajnal, Z., White, D., Lewry, J., Ashton, K., Weber, W., and Clowes, R., 1993. Deep seismic profile of a Proterozoic collision zone: Surprises at depth. Nature, v. 363, p. 339-342.

Macek, J.J., Bleeker, W., 1989. Thompson Nickel Belt Project: Geological Investigations at Pipe Mine, Setting Lake and Ospwagan Lake. *In* Report of Field Activities 1989, Manitoba Energy and Mines, p. 73-87.

Machado, N., Heaman, L., Krogh, T.E., and Weber, W., 1987. U-Pb geochronology program: Thompson Belt Northern Superior Province. *In* Report of Field Activities 1987, Manitoba Energy and Mines, p. 145-146.

Machado, N., Krogh, T.E., and Weber, W., 1990. U-Pb geochronology of basement gneisses in the Thompson Belt (Manitoba): evidence for pre-Kenoran and Pikwitonei-type crust and Early Proterozoic basement reactivation in the western margin of the Archean Superior Province. Canadian Journal of Earth Sciences, v. 27, p. 794-802.

Nägerl, P., and Bleeker, W., 1992. Investigations of the sub-Paleozoic extension of the Thompson Nickel Belt. Geological Association of Canada, Program with Abstracts, v. 17, p. A82.

Naldrett, A.J., Hoffman, E.L., Green, A.H., Chou, C.L., and Naldrett, S.R., 1979. The composition of Ni-sulphide ores, with particular reference to their content of PGE and Au. Canadian Mineralogist, v. 17, p. 403-415.

Paktunc, A.D., 1987. Differentiation of the Cuthbert Lake ultramafic dykes and related mafic dykes. Contributions to Mineralogy and Petrology, v. 97, p. 405-416.

Paterson, H.L., Donaldson, M.J., Smith, R.N., Lenard, M.F., Gresham, J.J., Boyack, D.J., and Keays, R.R., 1984. Nickeliferous sediments and sediment-associated nickel ores at Kambalda, Western Australia. *In* Buchanan, D.L., and Jones, M.L., eds., Sulphide deposits in mafic and ultramafic rocks, The Institution of Mining and Metallurgy, p. 81-94.

Peredery, W.V., and Geological Staff, 1982. Geology and nickel sulphide deposits of the Thompson belt, Manitoba. *In* Hutchinson, R.W., Spence, C.D., and Franklin, J.M., eds., Precambrian Sulphide Deposits, H.S. Robinson Memorial Volume, Geological Association of Canada, Special Paper 25, p. 165-209.

Rance, H., 1966. Superior-Churchill structural boundary, Wabowden, Manitoba. Unpublished Ph.D. thesis, University of Western Ontario, London, Ontario.

Roussel, D.H., 1965. Geology of Cross Lake area. Department of Mines and Natural Resources, Province of Manitoba, Publication 62-4, 69 p.

Scoates, R.F.J., 1990. The Fox River sill, northeastern Manitoba—a major stratiform intrusion. Manitoba Department of Energy and Mines, Geological Report GR82-3, 192 p.

Scoates, R.F.J., 1981. Volcanic rocks of the Fox River Belt, northeastern Manitoba. Manitoba Department of Energy and Mines, Geological Report GR81-1, 109 p.

Scoates, R.F.J., and Macek, J.J., 1978. Molson Dyke Swarm, Manitoba Mines Branch, Geological Paper 78-1, 53 p.

Scoates, R.F.J., Macek, J.J., and Russel, J.K., 1977. Thompson Nickel Belt project. *In* Report of Field Activities, Manitoba Mineral Resources Division, p. 47-53.

Stauffer, M.R., 1984. Manikewan: an Early Proterozoic ocean in central Canada, its igneous history and orogenic closure. Precambrian Research, v. 25, p. 257-281.

Stern, R.A., Syme, E.C., Bailes, A.H., and Lucas, S.B., 1995. Paleoproterozoic (1.90-1.86 Ga) arc volcanism in the Flin Flon Belt, Trans-Hudson Orogen, Canada. Contributions to Mineralogy and Petrology, v. 119, p. 117-141.

Thomas M.D., Sharpton, V.L., and Grieve, R.A.F., 1987. Gravity patterns and Precambrian structure in the North American central plains. Geology, v. 15, p. 489-492.

Weber, W., and Scoates, R.F.J., 1978. Archean and Proterozoic metamorphism in the northwestern Superior province and along the Churchill-Superior boundary, Manitoba. *In* Fraser, J.A., and Heywood, W.W., eds., Metamorphism in the Canadian Shield, Geological Survey of Canada, Paper 78-10, p. 5-16.

White, D.J. and Lucas, S.B., 1994. A closer look at the Superior Boundary Zone. 1994 Lithoprobe Trans-Hudson Orogen Transect Workshop Report, No. 38, p. 35-41.

Zhai, Y., and Halls, H.C., 1994. Multiple episodes of dike emplacement along the northwestern margin of the Superior Province, Manitoba. Journal of Geophysical Research

Zurbrigg, H.F., 1963. Thompson Mine geology. Canadian Institution of Mining and Metallurgy Bulletin, v. 56, p. 451-460.

Zwanzig, H.V., 1990. Kisswynew Gneiss Belt in Manitoba: stratigraphy, structure and evolution. *In* Lewry, J.F., and Stauffer, M.R., eds., The Early Proterozoic Trans-Hudson Orogen of North-America, Geological Association of Canada Special Paper 37, p. 95-120.

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