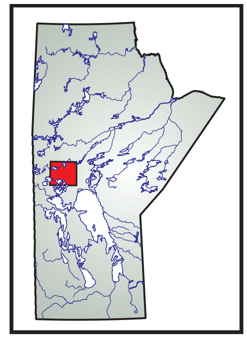
GS-6 New geological mapping, geochemical, Sm-Nd isotopic and U-Pb age data for the eastern sub-Phanerozoic Flin Flon Belt, west-central Manitoba (parts of NTS 63J3–6, 11, 12, 14, 63K1–2, 7–10)

by R-L. Simard, C.R. McGregor, N. Rayner[1] and R.A. Creaser[2]



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Summary

All volcanogenic massive sulphide (VMS) deposits found to date in the exposed portion of the Flin Flon Belt are located in juvenile-arc volcanic rocks. Recent exploration activity, however, has focused on the southern extension of the Flin Flon Belt beneath the Phanerozoic cover, leading to the discovery of several VMS deposits in poorly documented volcano-sedimentary successions.

Preliminary observations of the stratigraphy, whole-rock geochemistry, Sm-Nd isotope and geochronological data for 10 VMS-hosting successions, mainly from the sub-Phanerozoic portion of the Flin Flon Belt, lead to one main conclusion. Volcanogenic massive sulphide deposits west of Hargrave Lake (Moose, Limestone, Sylvia, Kofman, Reed Lake, Rail) are hosted in 'lower metamorphic-grade volcanic and sedimentary rocks' that are derived from juvenile, bimodal, tholeiitic to transitional oceanic-arc rocks of presently unknown age. Deposits east of Hargrave Lake (Watt's River, Fenton, Harmin, Talbot Lake), in comparison, are hosted in 'higher metamorphic-grade volcanic and sedimentary rocks' that formed in a rifting arc and/or at the onset of back-arc magmatism ca. 1865 Ma. The maturity and/or location of these deposits within the 'rifted basin' may have varied between deposits, as suggested by varying amounts of subduction-related contamination in the mantle source.

Results of this ongoing study have demonstrated that combined stratigraphic, geochemical, isotopic and geochronological data can be successfully applied, even in highly deformed and/or metamorphosed sub-Phanerozoic strata, to unravel the tectonostratigraphic affinity and, in some cases, to identify protoliths of ore-hosting successions, as shown for the Limestone, Sylvia, Rail, Watt's River, Fenton, Harmin and Talbot Lake deposits. This integrated approach provides a new framework for modern mineral exploration for hidden VMS deposits in the various exposed and covered 'domains' of the belt.

Introduction

The Flin Flon Belt is one of the largest Paleoproterozoic volcanogenic massive sulphide (VMS) districts in the world. It is composed of a series of tectonostratigraphic assemblages (juvenile arc, juvenile ocean-floor back arc, ocean plateau, ocean-island basalt, evolved plutonic arc) flanked to the north by gneisses of the Kisseynew Domain and extending south under the Phanerozoic rocks of the Western Canadian Sedimentary Basin (Figures GS-6-1, -2). All VMS deposits found to date in the exposed portion of the Flin Flon Belt are located within the juvenile-arc volcanic rocks.

Extensive mineral exploration spanning almost a century in the exposed parts of the Flin Flon Belt has led to the discovery of at least 25 ore deposits (as well as 40 subeconomic deposits; Syme et al., 1999b), the most recent being the Lalor Lake deposit (2007) in the Snow Lake area. Recent exploration activity, however, has also focused on the southern extension of the Flin Flon Belt beneath the Phanerozoic cover. Advances in geophysics during the last 20 years have rendered the Phanerozoic cover virtually transparent, thus leading to the discovery of several VMS deposits (e.g., Moose, Sylvia, Kofman, Reed Lake, Talbot, Fenton, Harmin, Watt's River), as well as other base-metal deposits (e.g., Namew Lake Ni-Cu deposit), in the sub-Phanerozoic Flin Flon Belt. A number of these new sub-Phanerozoic VMS deposits, when plotted on the latest geological map of the sub-Phanerozoic Flin Flon Belt published in the late 1990s, do not coincide with volcanic rocks but fall into sedimentary gneiss domains, such as the eastern Kisseynew Domain (Figure GS-6-2). This raises questions regarding the nature of these deposits (e.g., are all of them VMS deposits?), the nature of the rocks that host them and/or the quality of available sub-Phanerozoic bedrock maps.

In order to address these questions, a new mapping program for the sub-Phanerozoic portion of the Flin Flon Belt was initiated in 2009; this includes a major data-compilation component (see details in Simard and McGregor, 2009). One of the major objectives of this project is to better characterize the various geophysical domains in the sub-Phanerozoic portions of the Flin Flon Belt by using existing (data compilation) and new geochronological, isotopic and geochemical data. The results will facilitate comparison/correlation between the sub-Phanerozoic domains and the well-documented tectonostratigraphic

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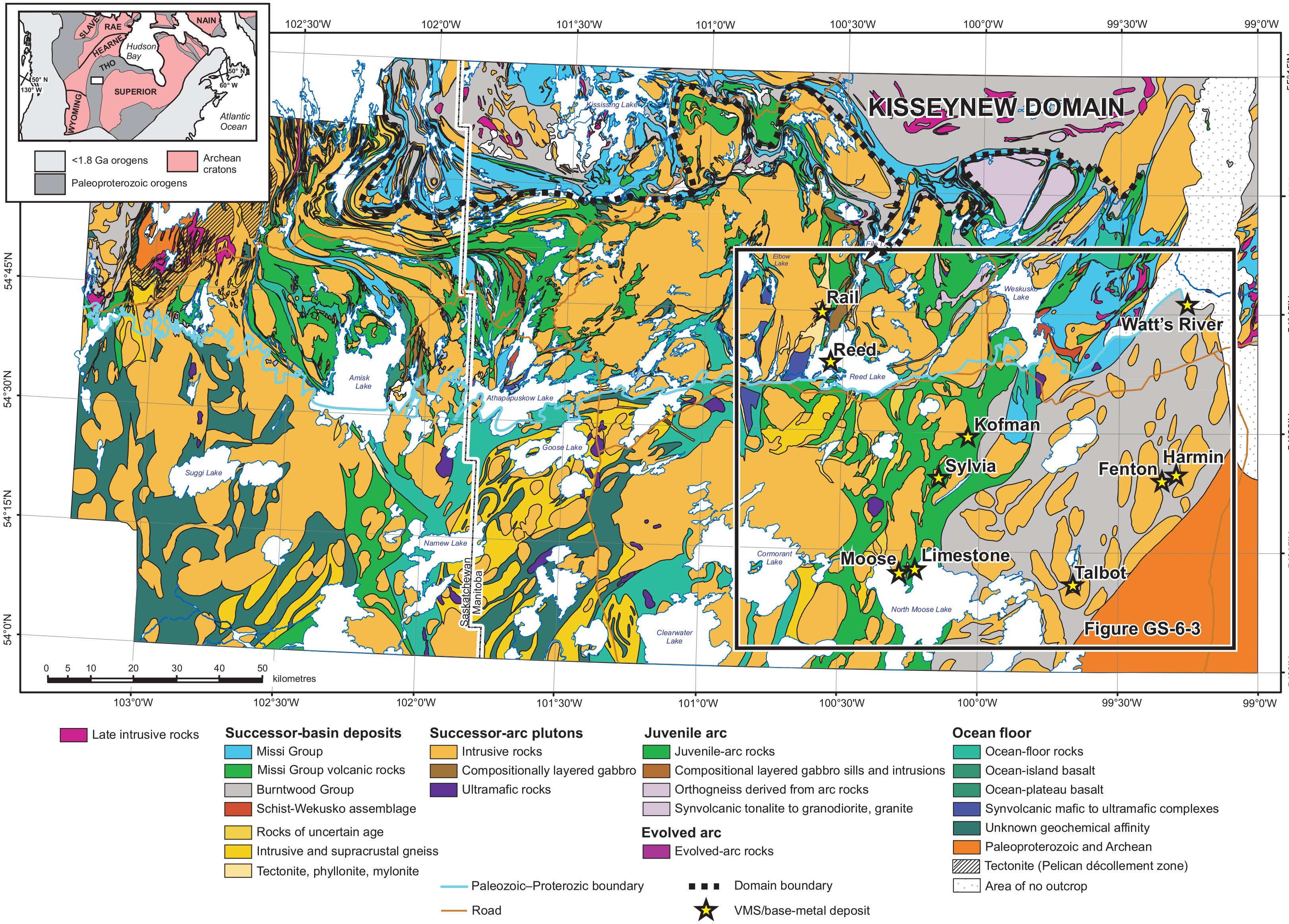

***Figure GS-6-1:** Generalized geology of the Flin Flon Belt, including its sub-Phanerozoic southern extension (modified from NATMAP Shield Margin Project Working Group, 1998); inset map shows the location of the Flin Flon Belt within the Trans-Hudson Orogen (THO).*

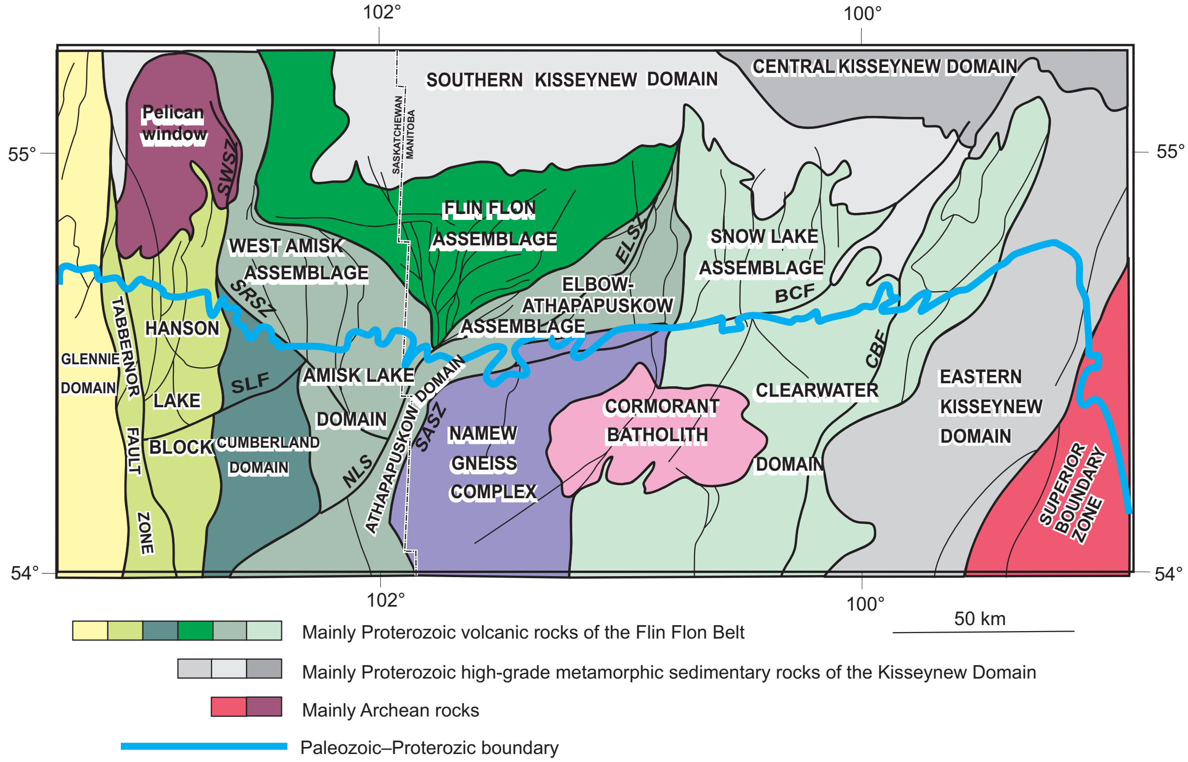

Figure GS-6-2: *Interpreted lithotectonic domains of the Flin Flon Belt (after NATMAP Shield Margin Project Working Group, 1998). Abbreviations of major faults and shear zones: BCF, Berry Creek Fault; CBF, Crowduck Bay Fault; ELSZ, Elbow Lake Shear Zone; NLS, Namew Lake structure; SASZ, South Athapapuskow Shear Zone; SLF, Suggi Lake Fault; SRSZ, Spruce Rapids Shear Zone; SWSZ, Sturgeon Weir Shear Zone. The heavy blue line is the contact between the exposed Precambrian rocks to the north and Paleozoic-covered Precambrian rocks to the south.*

assemblages of the exposed portion of the belt, resulting in revised geological maps for the sub-Phanerozoic geology of the Flin Flon Belt in Manitoba.

This paper provides a summary of the geology, whole-rock and Sm-Nd isotope geochemical, and U-Pb age results for samples collected from the Talbot, Harmin, Fenton and Moose VMS deposits in 2009, along with a summary of the geology and whole-rock geochemistry of the Limestone, Watt's River, Sylvia, Kofman, Rail and Reed Lake VMS deposits investigated in the summer of 2010. Samples collected in 2010 for Sm-Nd isotopic analysis and U-Pb radiometric dating were submitted for analyses, and results are pending. Geochemical, isotope and geochronological data are available in Data Repository Item DRI2010006.[3]

Regional setting

The Paleoproterozoic Flin Flon Belt is part of the juvenile Reindeer Zone of the Trans-Hudson Orogen (Figure GS-6-1), a collision zone formed during the 2.0–1.8 Ga amalgamation of several Archean cratons into Laurentia (Hoffman, 1988). The Flin Flon Belt consists of a series of tectonostratigraphic assemblages (juvenile arc, juvenile ocean-floor back arc, ocean plateau, oceanic-island basalt and evolved plutonic arc) that range in age from 1.92 to 1.87 Ga (Syme et al., 1999a). All of the VMS deposits mined to date in the Flin Flon Belt are hosted within the ca. 1.9 Ga juvenile-arc assemblages (Syme et al., 1999a).

The Flin Flon Belt was assembled during a period of intraoceanic accretion ca. 1.88–1.87 Ga, when part of the arcs and oceans were amalgamated to form the Amisk Collage (Lucas et al., 1996). This accretionary complex was subsequently 'stitched' together by calcalkaline plutons related to the 1.87–1.84 Ga successor-arc formation with coeval subaerial volcanism (ca. 1.87–1.85 Ga; Lucas et al., 1996). During the waning stage of successor-arc magmatism, uplift unroofed this collage, which led to deposition of marine to subaerial volcaniclastic and epiclastic units (Schist-Wekusko assemblage; Ansdell et al., 1999), development of a paleosol and deposition of voluminous continental alluvial-fluvial deposits (Missi Group) and broadly coeval marine turbidites (Burntwood Group) into the Kisseynew paleobasin to the north (ca. 1.85–1.84 Ga; Zwanzig, 1990; Lucas et al., 1996).

[3] MGS Data Repository Item DRI2010006, containing the data or other information sources used to compile this report, is available on-line to download free of charge at http://www2.gov.mb.ca/itm-cat/freedownloads.html, or on request from minesinfo@gov.mb.ca or Mineral Resources Library, Manitoba Innovation, Energy and Mines, 360–1395 Ellice Avenue, Winnipeg, MB R3G 3P2, Canada.

Arc and successor-arc processes were followed by a protracted interval of deformation and metamorphism in the Trans-Hudson Orogen (1.84–1.69 Ga) related to collision between the juvenile Paleoproterozoic Reindeer Zone and Archean cratons (Gordon et al., 1990; Lewry, 1990; Ansdell et al., 1995; Fedorowich et al., 1995), during which the Flin Flon Belt was overthrust by nappes of metasedimentary gneiss of the Kisseynew Domain to the north (Zwanzig, 1990).

Exposed Flin Flon Belt and Kisseynew Domain

The exposed Flin Flon Belt contains six geographically separated juvenile-arc assemblages (Hanson Lake, West Amisk, Birch Lake, Flin Flon, Fourmile Island and Snow Lake), which are separated by major faults or intervening ocean-floor, Burntwood Group sedimentary or plutonic rocks, or a combination of these rock types (Figure GS-6-1). The arc assemblages are internally complex, comprising numerous fault-bounded and folded volcanic suites (e.g., Bailes and Syme, 1989). They comprise a wide range of arc-related volcanic, volcaniclastic and intrusive rocks (Bailes and Syme, 1989; Syme and Bailes, 1993; Stern et al., 1995a; Lucas et al., 1996; Bailes and Galley, 2007). These assemblages are bimodal with dominantly andesite/basalt and rhyolite/dacite. Volcaniclastic sedimentary rocks, arc-rift basalt and shoshonite form subordinate but stratigraphically important members. Synvolcanic intrusions range widely in composition and vary in intrusive style from granitoid plutons to dikes and sills. Tholeiitic (1.90–1.89 Ga), calcalkaline (1.89–1.88 Ga) and shoshonitic (1.885 Ga) suites have been widely recognized, whereas boninite (>1.892 Ga) occurs only in the Snow Lake arc assemblage (Stern et al., 1993; Stern et al., 1995a; David et al., 1996).

The juvenile ocean-floor assemblages are composed mainly of mid-ocean-ridge basalt (MORB)–like basalt and related kilometre-scale, layered, mafic–ultramafic plutonic complexes (Figure GS-6-1; Syme and Bailes, 1993; Stern et al., 1995b). The volcanic sequences comprise thick units of pillowed and massive basalt. Uranium-lead zircon ages for these ocean-floor assemblages in the exposed portion of the Flin Flon Belt indicate that the ocean-floor magmatism was coeval with tholeiitic-arc volcanism at ca. 1.9 Ga (David et al., 1993; Stern et al., 1995b).

Voluminous plutons and coeval volcanism and deposition of sedimentary rocks of the successor arc took place between 1.88 Ga and 1.83 Ga. Large plutons, emplaced during three distinct magmatic stages, are highly variable lithologically and geochemically throughout the belt. Early (1.88–1.86 Ga) and middle (1.86–1.84 Ga) successor-arc plutons show both normal and reverse compositional zonation, from diorite to granodiorite, with moderate K content, steep rare earth element (REE) patterns and no Eu anomalies. In contrast, late (1.843–1.826 Ga) successor-arc plutons have high K content (granite in places), greater enrichment in high-field-strength elements (HFSE), and both variable REE pattern slopes and Eu anomalies (Whalen et al., 1999). Mafic hornblende–rich gabbro and diorite inclusions are present, and locally abundant, in plutons of all ages.

Volcanic, volcaniclastic, and sedimentary rocks with dates of ca. 1.88–1.83 Ga have been documented across the central and eastern parts of the exposed Flin Flon Belt, and are termed successor-basin deposits. They include the Schist-Wekusko assemblage, the Missi Group and the Burntwood Group. The Schist-Wekusko assemblage is of limited geographic extent and is characterized by marine trachyandesite conglomerate and felsic volcanic rocks with successor-arc ages (1864–1856 Ma; Stern et al., 1995a; Ansdell et al., 1999).

The fluvial-alluvial Missi Group sedimentary deposits are characterized by thick packages of conglomerate, pebbly sandstone and massive sandstone. Uranium-lead isotopic analysis of detrital zircon populations in Missi sandstone, combined with crosscutting intrusive age constraints, brackets deposition between 1854 and 1842 Ma (Ansdell et al., 1992). Detrital zircon ages in all Missi Group rocks indicate provenance from dominantly Flin Flon–Glennie Domain sources (1.92–1.85 Ga), as well as older (2.6–2.2 Ga) crustal sources (Ansdell et al., 1992; Ansdell, 1993). Volcanic rocks of the Missi Group include basalt, andesite, trachyandesite, dacite, rhyolite and equivalent intrusive rocks. Mafic volcanic rocks of the Missi Group are in places enriched in Zr, Y and TiO_2 relative to juvenile-arc volcanic rocks that predate 1.88 Ga (Connors and Ansdell, 1993, 1994). Gneissic equivalents of the felsic volcanic rocks of the Missi Group have been reported throughout the southern Kisseynew Domain (Figure GS-6-2; Bailes, 1975, 1980c, b; Zwanzig et al., 1996).

The marine turbidites of the Burntwood Group include greywacke, siltstone, mudstone and rare conglomerate. In the low-grade metamorphic Flin Flon Belt, these sedimentary rocks are generally in fault contact with other units. In the central Kisseynew Domain, in comparison, Burntwood Group psammitic to pelitic gneisses with upper almandine-amphibolite–facies mineral assemblages predominate, together with derived migmatitic paragneiss (Figure GS-6-2; Bailes and McRitchie, 1978; Bailes, 1980a), whereas the flanking zones/domains are composed of structurally interlayered gneisses derived from the basinal sedimentary rocks and the older arc, back-arc and successor-arc units (Zwanzig, 1990).

Sub-Phanerozoic Flin Flon Belt and Kisseynew Domain

The latest regional mapping of the sub-Phanerozoic portion of the Flin Flon Belt, conducted in the early 1990s by A. Leclair (Geological Survey of Canada), involved integration of high-resolution aeromagnetic and gravity data with drillcore information (Leclair et al., 1997). Leclair's compilation resulted in the recognition of several major domains (Figure GS-6-2), each with a distinct lithotectonic character and potential field-anomaly pattern. Some rock assemblages similar to those found in the exposed portion of the Flin Flon Belt and Kisseynew Domain were recognized in the sub-Phanerozoic, whereas others appear to be unique to the sub-Phanerozoic (Leclair et al., 1997). As a result, only three lithotectonic domains beneath the Phanerozoic cover were correlated with the 1.92–1.83 Ga volcanic and plutonic rocks of the exposed Flin Flon Belt: the Clearwater Domain with the Snow Lake arc assemblage, the Athapapuskow Domain with the Elbow-Athapapuskow ocean-floor assemblage and the Amisk Lake Domain with the West Amisk arc assemblage (Figure GS-6-2). The Namew Gneiss Complex, located just east of the Saskatchewan-Manitoba border, appears to be unique to the sub-Phanerozoic. It consists of upper-amphibolite–facies orthogneiss containing 1.88–1.83 Ga alkaline intrusive rocks with enclaves derived from older supracrustal rocks. Southwest of Reed Lake, the Namew Gneiss Complex and the Clearwater Domain are intruded by a large granodioritic pluton, the 1.83 Ga Cormorant Batholith (Leclair et al., 1997). The Clearwater Domain is bounded to the east by the 'eastern Kisseynew Domain', which includes mainly migmatitic quartzofeldspathic, pelitic and calcsilicate paragneisses that are intimately interlayered with metaplutonic veins and sheets (Leclair et al., 1997).

Geology of the sub-Phanerozoic eastern Flin Flon Belt and Kisseynew Domain

During the last two field seasons, several VMS deposits of the sub-Phanerozoic were studied in order to document the main geophysical/geological domains observed in the eastern portion of the sub-Phanerozoic Flin Flon Belt and Kisseynew Domain (Figure GS-6-3). Despite some similarities with boundaries defined by Leclair et al. (1997), lithotectonic domains defined in this report use a more generalized nomenclature for all but the 'Cormorant Batholith Domain'. Further work is required to more firmly delineate and subdivide the tentative domains discussed here. Based on the presently available geological

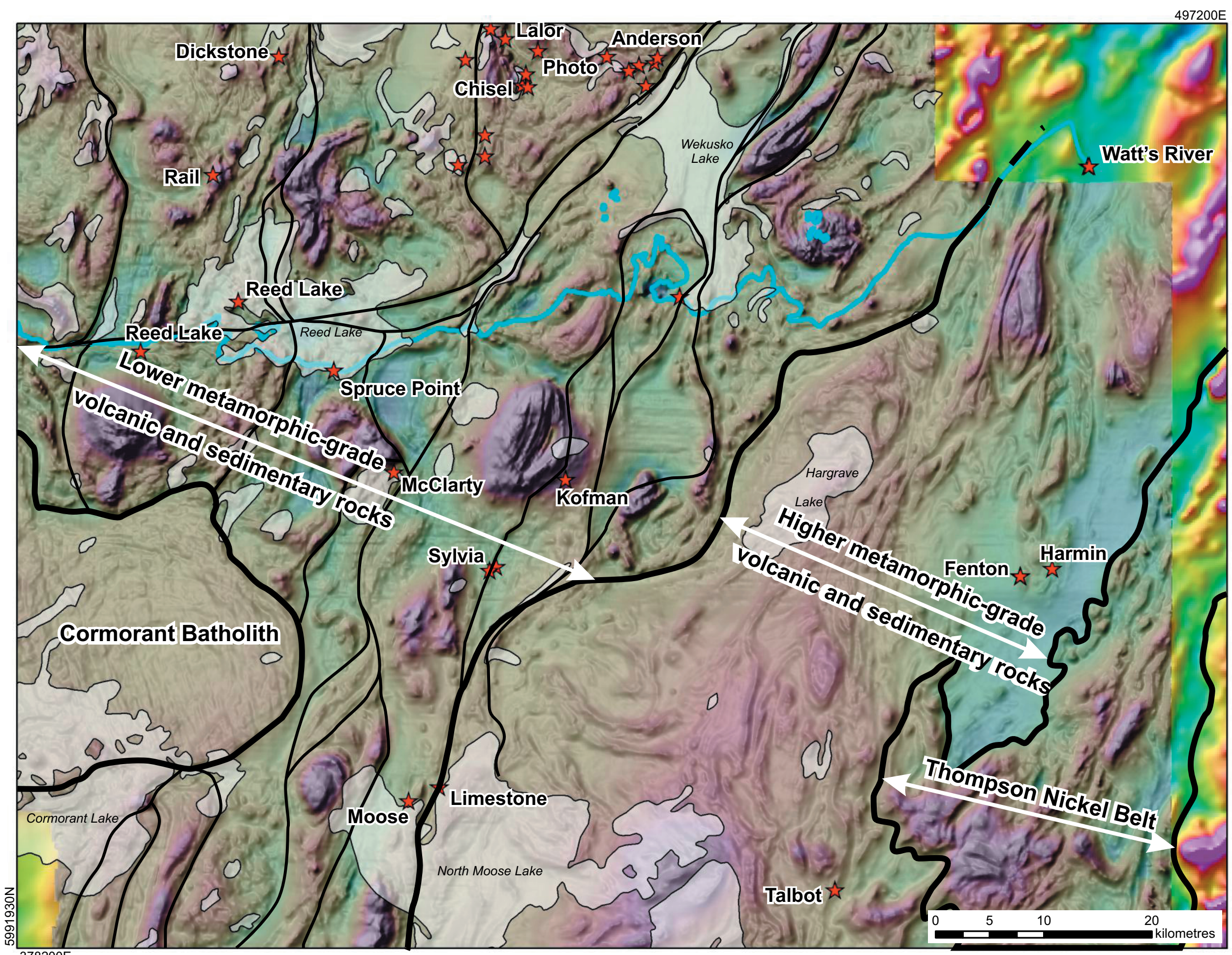

Figure GS-6-3: *Main geophysical domains of the eastern Flin Flon Belt superimposed on the residual total magnetic field; thick blue line traces limit of Phanerozoic cover; thick black lines are domain boundaries; thin black lines are faults and/or subdomain boundaries; red stars indicate known base-metal deposits.*

and geophysical data, three major domains are recognized, as outlined in Figure GS-6-3.

The 'Cormorant Batholith Domain' forms a roughly oval-shaped, 60 by 25 km, moderate-intensity aeromagnetic anomaly with a weak internal magnetic fabric transected by a series of linear north-northwest-striking structures southwest of Reed Lake that possibly represent faults or fractures (Figure GS-6-3; Leclair et al., 1997).

The Cormorant Batholith intrudes a package of 'lower metamorphic-grade volcanic and sedimentary rocks' with an overall north-northeast-trending magnetic pattern that extends from west of Reed and Cormorant lakes to east of Hargrave and North Moose lakes (Figure GS-6-3). The magnetic fabric of this 'lower metamorphic' domain is characterized by alternating north-northeast-trending curvilinear highs (mainly volcanic rocks) and arcuate lows (mainly sedimentary rocks or intrusions) disrupted by strong positive aeromagnetic anomalies (mainly mafic and ultramafic plutons). The metamorphic grade of these rocks varies gradually along strike, whereas abrupt across-strike transitions separate lower greenschist from middle-amphibolite facies domains. The lower metamorphic-grade rocks host several VMS deposits in the sub-Phanerozoic, such as Moose, Limestone, Sylvia, Kofman, McClarty Lake and Reed Lake (VMS Ventures Inc.), along with the past-producing Spruce Point mine. The lower metamorphic-grade rocks appear to extend north of the Berry Creek Fault into the exposed portion of the Flin Flon Belt where several other VMS deposits are located (e.g., Rail, Reed Lake [Rockcliff Resources Inc.], Chisel, Lalor, Anderson).

East of the lower metamorphic-grade rocks, a north-northeast-trending belt of 'higher metamorphic-grade volcanic and sedimentary rocks' is characterized by curvilinear to tightly folded magnetic patterns and extends from east of Hargrave and North Moose lakes to the Thompson Nickel Belt (Figure GS-6-3; boundaries modified from Macek et al., 2006). These higher grade rocks, which underwent peak metamorphism at upper-amphibolite to lower-granulite facies, are typically migmatitic gneiss with varying proportions of younger plutonic veins and sheets. Although most of these rocks were logged as metasedimentary gneiss, they host the Talbot, Harmin, Fenton and Watt's River VMS deposits.

VMS deposits in the 'lower metamorphic-grade volcanic and sedimentary rocks'

Six VMS deposits were studied in the 'lower metamorphic-grade volcanic and sedimentary rocks' west of Hargrave and North Moose lakes, namely the Moose, Limestone, Sylvia and Kofman deposits, along with two potentially correlative deposits in the exposed shield to the north: the Reed Lake (Rockcliff Resources Inc.) and Rail deposits. This section provides a summary of the geology and whole-rock[4] and isotope[5] geochemistry of these deposits.

Moose deposit

Location, cover, drillhole depth: Approximately 40 km south of Reed Lake near the north shore of North Moose Lake; ~180 m of Pleistocene glacial overburden and Phanerozoic sedimentary rocks; >500 m.

Stratigraphy: Well-preserved aphyric basalt (Figure GS-6-4a) is 'underlain' by massive to brecciated (lapilli tuff), aphyric, plagioclase-phyric and quartz-phyric rhyolite (Figure GS-6-4b), and crosscut by several plagioclase-phyric mafic dikes <3 m thick. No clear stratigraphic top indicators were found in the logged core samples.

Mineralization: Disseminated to near-solid sulphide zones (pyrite, pyrrhotite, sphalerite, chalcopyrite) are associated with felsic volcanic rocks. They commonly occur at or near the contact between basalt and rhyolite, at a depth of ~200 m, and consist of semidisseminated to near-solid sulphides that average ~15% pyrite, 5–7% pyrrhotite, 5–7% sphalerite and ~3% chalcopyrite (Figure GS-6-4c), as well as rare sulphide stringers.

Metamorphic facies: Upper greenschist (chlorite-biotite-epidote).

Alteration: Basalt is weakly to moderately chloritized and rhyolite is commonly sericitized and/or silicified. Late brittle faults that commonly display strong carbonatization were observed both above and below the mineralization.

Limestone deposit

Location, cover, drillhole depth: Approximately 4 km east-northeast of the Moose deposit; ~100 m of Pleistocene glacial overburden and Phanerozoic sedimentary rocks; ~500 m.

Stratigraphy: Highly deformed mafic (chlorite-biotite-garnet-quartz±staurolite; Figure GS-6-4d) and felsic

[4] One hundred and forty (140) representative rock samples were selected for geochemical analyses from a suite of more than 350 specimens collected during the mapping of the various VMS deposits of the eastern sub-Phanerozoic Flin Flon Belt and Kisseynew Domain. All samples were analyzed at Activation Laboratories Ltd., Ancaster, Ontario, with major-element concentrations determined by x-ray fluorescence (XRF), trace elements and rare earth elements (REE) by inductively coupled plasma–mass spectrometry (ICP-MS; using fusion for sample preparation), and Sc by instrumental neutron activation analysis (INAA). Analytical error for the XRF method was <1% for major elements. For trace elements, precision was better than 6% and analytical error was better than 5% (Young, 2002). Major- and trace-element compositions of all the analyzed samples are available in Data Repository Item DRI2010006.

[5] Fourteen (14) samples were analyzed for Sm and Nd concentrations and Nd isotopic compositions at the University of Alberta Radiogenic Isotope Facility, University of Alberta, Edmonton, Alberta. Results are presented in Table GS-6-1.

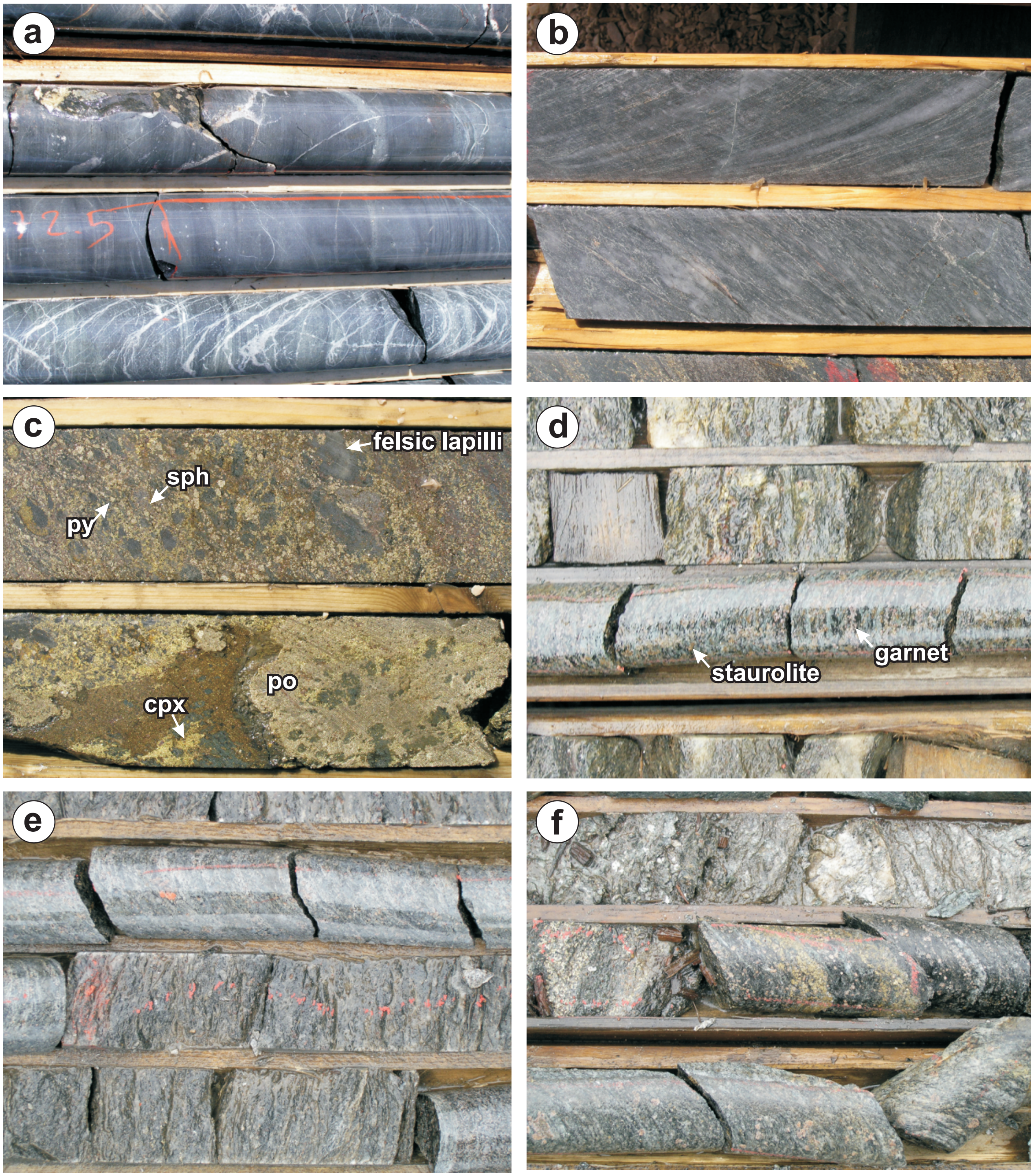

Figure GS-6-4: *Main rock types in drillcore from the Moose (a–c) and Limestone (d–f) deposits:* ***a)*** *massive aphyric basalt;* ***b)*** *monolithic felsic lapilli tuff;* ***c)*** *near-solid to solid sulphide mineralization in felsic lapilli tuff (py, pyrite; po, pyrrhotite; cpy, chalcopyrite; sph, sphalerite);* ***d)*** *chlorite-biotite-garnet-staurolite-quartz schist (mafic rock);* ***e)*** *sericite-quartz-biotite-chlorite-staurolite-garnet schist (felsic rock); and* ***f)*** *mineralized (pyrite-pyrrhotite) schist.*

(sericite-quartz-biotite-chlorite±staurolite±garnet; Figure GS-6-4e) schists are crosscut by less deformed, 3–15 m thick, fine-grained, hornblende-rich mafic rock interpreted as younger mafic dikes. The overall stratigraphy of the deposit consists of mafic schist (no staurolite) 'underlain' by intercalated, variably mineralized, mafic and felsic schist (with staurolite), in turn 'underlain' by mafic schist (no staurolite). No top indicators were observed during core logging. The premetamorphic nature of these rocks is unclear. A chlorite-biotite-garnet-staurolite-quartz metamorphic mineral assemblage is commonly observed in pelitic rocks (Spear and Cheney, 1989), but a similar assemblage has also been documented in altered mafic volcanic rocks in the footwall of VMS deposits of the Snow Lake area (Froese et al., 1989; Galley et al., 1993).

Mineralization: Minor (<10%) disseminated and stringer sulphide zones (pyrrhotite-pyrite- chalcopyrite±sphalerite) are associated with felsic schist intervals (Assessment File 71835, Manitoba Innovation, Energy and Mines, Winnipeg). The mineralization, observed mainly in felsic schist between 100 and 350 m in depth, consists of disseminated to near-stringer pyrite (<10%), pyrrhotite (<30%) and chalcopyrite (<3%), with sphalerite and chalcopyrite in fractures (Figure GS-6-4f).

Metamorphic facies: Lower amphibolite.

Sylvia deposit

Location, cover, drillhole depth: Approximately 20 km south-southeast of Reed Lake; ~100 m of Pleistocene glacial overburden and Phanerozoic sedimentary rocks; >500 m.

Stratigraphy: Intercalated, strongly foliated mafic (basalt) and felsic (rhyolite) rocks are 'underlain' by graphitic argillite. The volcanic stratigraphy consists of a thick succession of aphyric and plagioclase-phyric basalt and mafic volcaniclastic rocks intercalated with generally thinner, aphyric and quartz±plagioclase–phyric rhyolite. No top indicators were observed in the core.

Mineralization: This is classified as a stratabound VMS deposit consisting of several solid bands of pyrite, pyrrhotite, chalcopyrite, sphalerite and galena, flanked to the east by graphitic argillite (Rockcliff Resources Inc., 2008b). The mineralization is commonly hosted by, or associated with, the felsic volcanic rocks (quartz-sericite schist) and consists of near-solid to solid, thinly layered, very fine grained pyrite±chalcopyrite (<5%) with some stringer mineralization.

Metamorphic facies: Upper greenschist (chlorite-biotite±hornblende).

Alteration: The entire volcanic sequence that hosts the deposit has been extensively sericitized and/or chloritized. Late brittle faults are commonly associated with strong carbonatization.

Kofman (Kof) deposit

Location, cover, drillhole depth: Approximately 20 km southeast of Reed Lake, ~10 km northeast of the Sylvia deposit; ~100 m of Pleistocene glacial overburden and Phanerozoic sedimentary rocks; >300 m.

Stratigraphy: Well-preserved aphyric basalt is 'underlain' by a thick package of massive to brecciated aphyric rhyolite and massive silicified (light grey) amygdaloidal basalt, which is in turn 'underlain' by aphyric basalt.

Mineralization: This was described by Ferreira et al. (1996) as a stratabound VMS deposit consisting of disseminated to solid bands of pyrite, pyrrhotite, chalcopyrite and sphalerite contained in quartz-sericite schist derived from rhyodacite and rhyolitic tuff (see below). The mineralization is commonly found near the margins (top and bottom) of a very distinct, >100 m thick package of massive, variably chloritized and sericitized, amygdaloidal silicified (light grey) basalt (5% quartz-filled amygdules, 1–6 mm in diameter) associated with the aphyric rhyolite. It consists mainly of solid to near-solid medium-grained sulphides (>90% pyrite, <5% chalcopyrite, with minor sphalerite and pyrrhotite) and minor stringer mineralization (>75% pyrite, ~25% chalcopyrite). Portions of the mineralization show tectonic milling texture (*Durchbewegung* ore), comprising rounded clasts of coarser grained sulphide within the finer grained sulphide intervals.

Metamorphic facies: Middle greenschist (chlorite-sericite-epidote-actinolite).

Alteration: Silicification related to the mineralization is very intense and concentrated in the massive, light grey, amygdaloidal host basalt. Without whole-rock geochemistry, this altered basalt resembles, and was logged and reported as, quartz-phyric rhyolite/rhyodacite in previous publications (Ferreira et al., 1996; Rockcliff Resources Inc., 2007, 2008a).

Reed Lake deposit (Rockcliff Resources Inc.)

Location, cover, drillhole depth: Beneath the western portion of Reed Lake; 2 km north of the Paleozoic limestone cover; ~1000 m.

Stratigraphy: Intercalated well-preserved aphyric basalt and quartz- and quartz-plagioclase–phyric rhyolites. Basalt forms thick successions (>150 m) of variably altered, massive to pillowed flows with minor mafic breccia intervals (flow-top breccia?). The rhyolites form thick (>300 m), massive to brecciated units.

Mineralization: A series of en échelon near-solid sulphide lenses is separated by chlorite schist and chloritized andesite (Hudson Bay Exploration and Development Co. Ltd., 1998).

Metamorphic facies: Lower to middle greenschist (epidote-chlorite).

Alteration: Quartz-epidote patches are common throughout the mafic rocks. The rhyolites are variably altered (purplish hematization, silicification, sericitization).

Rail deposit

Location, cover, drillhole depth: 5 km northwest of Reed Lake; 15 km north of the Paleozoic limestone cover; >500 m.

Stratigraphy: The Rail property occurs within the West Reed–North Star shear zone that separates the Elbow-Athapapuskow ocean-floor assemblage on the west from the Fourmile Island island-arc assemblage to the east (Figure GS-6-1). The observed stratigraphy consists of two faulted successions of variably altered, intercalated aphyric basalt and quartz-plagioclase–phyric rhyolite, with a 2–7 m wide fault zone separating the two. The lower succession is intruded by a medium-grained, pinkish grey granodiorite sheet, just below the main mineralized interval.

Mineralization: The main mineralized interval is observed in the lower succession where it is associated with aphyric rhyolite. It consists of 3–10 m of near-solid, fine- to medium-grained sulphides (15–30% chalcopyrite, 5–10% sphalerite, 30–70% pyrrhotite, 10–15% pyrite) that commonly display *Durchbewegung* texture. Sulphide stringers (chalcopyrite) and thinner bands of solid sulphide (mainly pyrrhotite) are observed downward in the stratigraphy in decreasing abundance. In the upper succession, disseminated pyrite and pyrrhotite are observed in both mafic and felsic rocks, whereas near-solid to solid sulphide (pyrrhotite and pyrite) is restricted to the felsic rocks and some quartz veins. The near-solid to solid sulphides commonly show tectonic milling texture (*Durchbewegung*), comprising rounded clasts of rhyolite and quartz within sulphide.

Metamorphic facies: The upper sequence is amphibolite (biotite-hornblende-garnet±epidote); the lower sequence is middle greenschist (chlorite+epidote).

Geochemistry of the 'lower metamorphic-grade volcanic and sedimentary rocks'

Mafic (basalt to andesite/basalt) and felsic (rhyolite/dacite) rocks from the Moose, Limestone, Kofman and Reed Lake deposits, as well as from the lower succession of the Rail deposit, have low Nb/Y ratios (~0.1), suggesting subalkaline affinity (Figure GS-6-5a). The REE patterns of both the mafic and felsic rocks on the chondrite-normalized REE diagram are flat, with no enrichment in either light REE (LREE) or heavy REE (HREE; Figure GS-6-6a). The pronounced negative Nb and Ti anomalies on the mantle-normalized trace-element diagram (Figure GS-6-6a) and distribution of the samples on the Th-(Zr/117)-Nb/16 diagram (Figure GS-6-5b) indicate that these rocks are the products of subduction-related processes in an oceanic-arc environment. Their low $(La/Yb)_N$ ratios clearly demonstrate a tholeiitic magmatic affinity (Figure GS-6-6a).

Rocks of the Sylvia deposit differ from the other deposits in that they have slightly higher Nb/Y ratios (~0.3; Figure GS-6-5a). The REE patterns for both the mafic and felsic rocks are moderately enriched in LREE compared to HREE, the HREE pattern is flat and $(La/Yb)_N$ ratios are moderate (Figure GS-6-6c). These characteristics suggest a more transitional affinity than the clearly tholeiitic affinity of the previously described rocks. All other geochemical characteristics (Figures GS-6-5a, b, -6c) suggest that rocks of the Sylvia deposit were also the products of subduction-related processes in an oceanic-arc environment.

Basalt to andesite/basalt and rhyolite/dacite of the upper stratigraphic succession of the Rail deposit are completely different than any other succession in terms of their geochemistry. Although the mafic rocks have low Nb/Y values (subalkaline affinity; Figure GS-6-5a), their trace-element and REE patterns are flat, depleted in LREE relative to HREE (Figure GS-6-7c) and show depletion in the strongly incompatible elements with no significant Nb or Ti anomalies, typical of N-MORB (Figure GS-6-7c). They plot within the N-MORB field on the Th-(Zr/117)-Nb/16 diagram (Figure GS-6-5b). This, along with all the other geochemical characteristics outlined above, suggests that they were the product of mid-ocean-ridge magmatism.

Neodymium isotopic compositions of whole-rock samples from the Moose deposit yielded juvenile $\varepsilon_{Nd\ (1850\ Ma)}$ values of +4.0 to +4.5 (Table GS-6-1), suggesting a juvenile magma source with minimum to no involvement of continental crust in the magma genesis. Geochronological and Sm-Nd isotopic analyses for hostrock samples from the other deposits are pending.

Base-metal/VMS deposits in the 'higher metamorphic-grade volcanic and sedimentary rocks'

Four base-metal deposits were studied in the 'higher metamorphic-grade volcanic and sedimentary rocks' west of Hargrave and North Moose lakes, namely the Watt's River, Fenton, Harmin and Talbot Lake deposits. This section provides a summary of the geology, whole-rock and isotope geochemistry, and geochronology[6] for these deposits.

'Higher metamorphic-grade volcanic and sedimentary rocks' have been metamorphosed at upper-amphibolite to

[6] Uranium-lead geochronology was carried out using the sensitive high-resolution ion microprobe (SHRIMP) at the Geological Survey of Canada in Ottawa. Analytical procedures followed those described by Stern (1997), with standards and U-Pb calibration methods following Stern and Amelin (2003). The internal features of the zircons (such as zoning, structures, alteration) were characterized in back-scattered electron mode (BSE) using a Zeiss Evo® 50 scanning electron microscope. The 'Isoplot version 3.00' application of Ludwig (2003) was used to generate concordia diagrams and calculate weighted means. All errors reported in the text are given at the 2σ uncertainty level. The SHRIMP U/Pb zircon results are available in Data Repository Item DRI2010006.

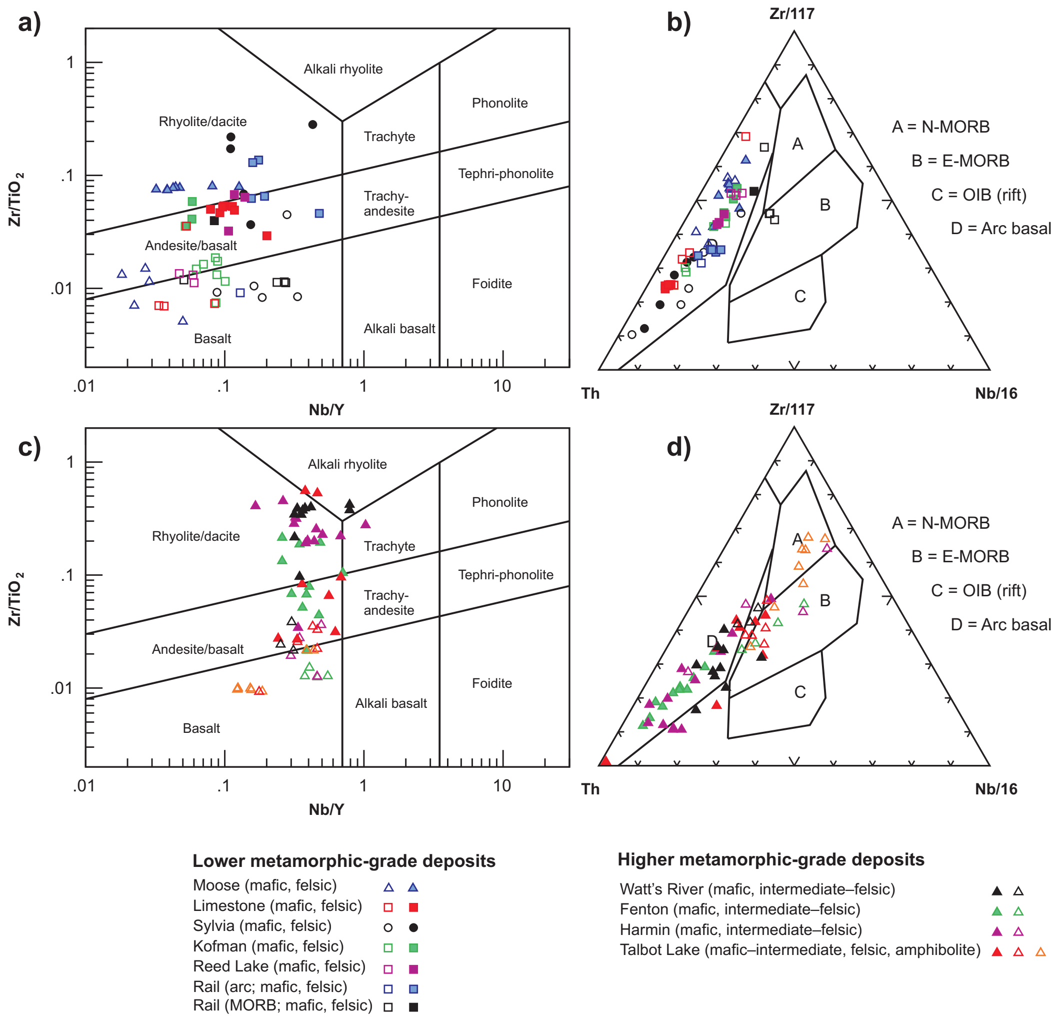

***Figure GS-6-5:** Geochemical characteristics of the hostrocks from VMS deposits in the sub-Phanerozoic Flin Flon Belt and Kisseynew Domain: **a, c)** Zr/TiO_2 vs. Nb/Y classification diagram (modified from Winchester and Floyd, 1977); **b, d)** Th-(Zr/117)-(Nb/16) discriminant diagram of Wood (1980). Abbreviations: E-MORB, enriched mid-ocean-ridge basalt; N-MORB, normal mid-ocean-ridge basalt; OIB, ocean-island basalt.*

lower-granulite facies, resulting in variable amounts of melt (anatectic texture and felsic intrusive veins/sheets); thus, no primary textures are preserved and protoliths are difficult to ascertain. Historically, the high-grade rocks were logged as metasedimentary rocks such as pelite (>50% biotite), greywacke (<30% biotite), pebbly greywacke and siltstone/quartzite (<10% biotite). More recently, they have been logged according to their mineral composition and fabric as 'mafic' biotite-rich quartzofeldspathic gneiss (>35% biotite), 'intermediate' quartzofeldspathic gneiss (15–35% biotite) and 'felsic' quartzofeldspathic gneiss (<15% biotite). However, in view of the recent discoveries of stratabound VMS-like mineralization in these rocks, the possibility has arisen that they are of volcanic origin (quartzite / felsic quartzofeldspathic gneiss = metamorphosed rhyolite and felsic tuff; biotite-rich quartzofeldspathic gneiss/greywacke = metamorphosed basalt or andesite), but this hypothesis remains to be tested.

Although each deposit has a unique stratigraphy, they all share some common characteristics. Each stratigraphic section consists of 30–300 m thick successions of layered quartzofeldspathic gneiss, grading from mafic (>30% biotite) to intermediate (15–30 % biotite) to felsic (<15% biotite). Each of these successions displays variable degrees of purplish green calcsilicate alteration (actinolite, diopside, calcite) that either completely replaced the primary mineralogy over several metres or developed in 2–25 cm thick bands. Sillimanite±cordierite are common in the intermediate and felsic quartzofeldspathic gneisses.

Watt's River deposit

Location, cover, drillhole depth: Approximately 52 km east-southeast of Snow Lake; 15–65 m of Pleistocene glacial overburden and locally a few metres of Phanerozoic sedimentary rocks; >1000 m.

Stratigraphy: Several thin (30–70 m), mafic to felsic

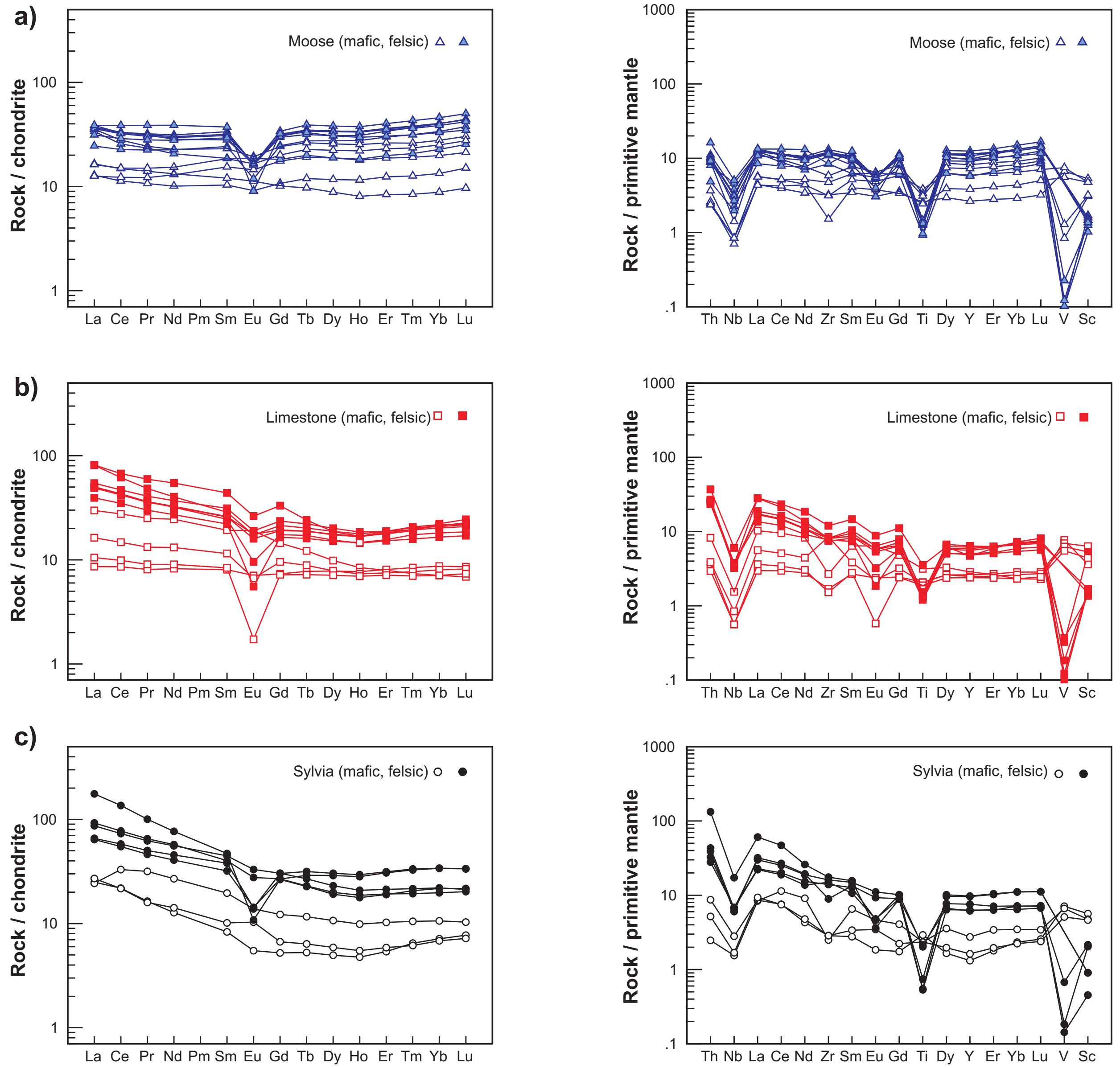

Figure GS-6-6: *Chondrite-normalized rare earth element (REE) patterns (normalizing values from McDonough and Sun, 1995) and mantle-normalized incompatible trace-element patterns (normalization values from Sun and McDonough, 1989) for rocks of the* ***a)*** *Moose deposit,* ***b)*** *Limestone deposit, and* ***c)*** *Sylvia deposit.*

quartzofeldspathic gneissic successions are locally crosscut by younger, fine-grained, massive amphibolite dikes. Minor brittle fault zones, which do not seem to affect the stratigraphy, commonly display carbonatization and sericitization.

Mineralization: Three main zones of sulphides with associated precious metals (gold and silver; HudBay Minerals Inc., 2007) are hosted in felsic quartzofeldspathic gneiss intervals, where they form narrow solid-sulphide bands, disseminations, blebs, splashes and stringers of coarse-grained pyrite, pyrrhotite, chalcopyrite, lesser sphalerite and minor galena. The mineralization commonly displays *Durchbewegung* texture.

Fenton deposit

Location, cover, drillhole depth: Approximately 70 km southeast of Snow Lake; 20 m of Pleistocene glacial overburden and ~50 m of Phanerozoic sedimentary rocks; >600 m.

Stratigraphy: The upper portion of the observed stratigraphy is dominated by intermediate to mafic gneisses, whereas the lower portion consists mainly of felsic quartzofeldspathic gneiss. Abundant aplite, pegmatite, granodiorite and leucogranite randomly crosscut these gneisses.

Mineralization: Hosted in the upper portion of the felsic quartzofeldspathic gneiss, it consists of disseminated pyrrhotite, sphalerite and chalcopyrite, and near-solid

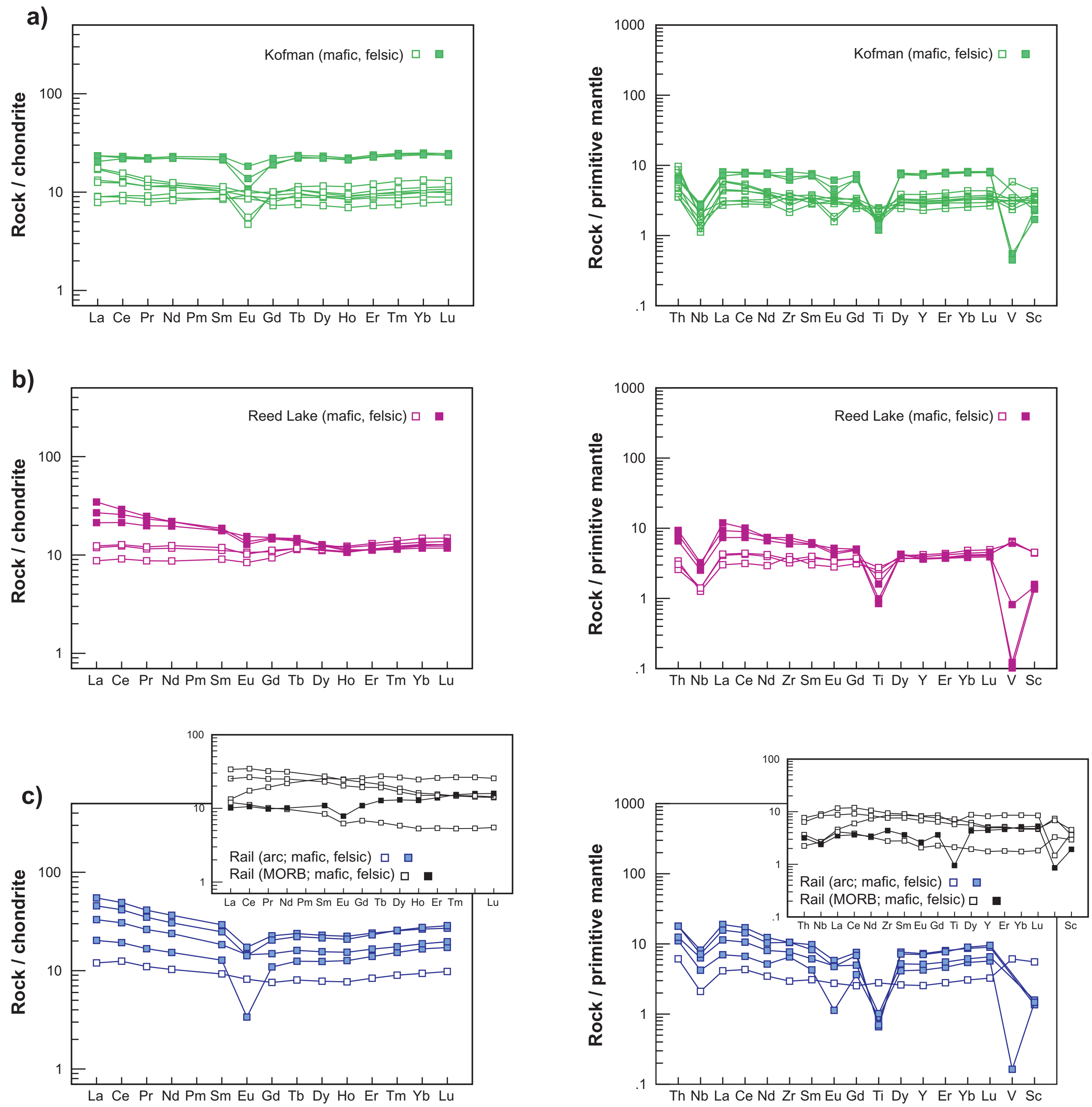

Figure GS-6-7: *Chondrite-normalized rare earth element (REE) patterns (normalizing values from McDonough and Sun, 1995) and mantle-normalized incompatible trace-element patterns (normalization values from Sun and McDonough, 1989) for rocks of the* ***a)*** *Kofman deposit,* ***b)*** *Reed Lake deposit, and* ***c)*** *Rail deposit.*

sulphide intervals of medium- to coarse-grained pyrrhotite, sphalerite and chalcopyrite.

Metamorphic facies: Upper amphibolite to lower granulite (hornblende-biotite-garnet±hypersthene).

Geochronology: A sample of quartzofeldspathic gneiss (105-09-HAR 042 402.3–404.5) from immediately below the mineralization was collected for U-Pb geochronology. Abundant, small zircon grains were recovered from the sample. Most consist of small, clear colourless cores surrounded by pale brown, slightly turbid rims. In back-scattered electron (BSE) images, the cores are characterized by oscillatory zoning, whereas the rims are typically unzoned. Fifty-five zircon grains were analyzed by SHRIMP and yielded ages ranging from 1919 to 1790 Ma. Analysis of eight high U, low Th/U rims yielded a weighted mean $^{207}Pb/^{206}Pb$ age of 1804 ±7 Ma, which is interpreted as the age of metamorphic overprint. The remaining analyses yielded a unimodal age population centred at 1865 Ma, which is interpreted as the approximate age of a single detrital zircon source (Figure GS-6-8). The strong metamorphism at ~1804 Ma obliterated all primary textures in the quartzofeldspathic gneiss, which makes it difficult to confidently interpret a volcanic or sedimentary protolith.

Table GS-6-1: Samarium-neodymium isotopic results for hostrocks of VMS deposits in the sub-Phanerozoic Flin Flon Belt and Kisseynew Domain.

Sample		Sm (ppm)	Nd (ppm)	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	2σ	T_{DM}	~T (ma)	ε_{Nd}^{T}
Moose deposit									
105-09-MAW 067 172.5	Aphyric basalt	1.44	4.42	0.1977	0.512856	0.000008	N/A	1850	4.0
105-09-MAW 067 223.6	Plagioclase-phyric rhyolite*	4.48	13.57	0.1998	0.512907	0.000008	N/A	1850	4.5
105-09-MAW 071 147.0	Aphyric basalt	3.32	9.26	0.2167	0.513101	0.000008	N/A	1850	4.3
105-09-MAW 071 250.8	Plagioclase-phyric rhyolite	4.28	12.59	0.2056	0.512965	0.000006	N/A	1850	4.3
Fenton deposit									
105-09-HAR 042 120.0	Sillimanite-rich intermediate quartzo-feldspathic gneiss	4.43	24.12	0.1110	0.511752	0.000007	2.09	1850	3.1
105-09-HAR 042 247.8	Amphibolite-gabbro	3.41	12.35	0.1671	0.512441	0.000006	N/A	1850	3.2
105-09-HAR 042 402.0	Intermediate to felsic quartzofeld-spathic gneiss*	5.77	25.69	0.1359	0.512063	0.000007	2.15	1850	3.2
Harmin deposit									
105-09-NIM 062 225.7	Mafic gneiss (~50% biotite)	5.48	28.94	0.1145	0.511776	0.000006	2.12	1850	2.7
105-09-NIM 062 470.0	Felsic quartzofeldspathic gneiss*	5.55	23.81	0.1410	0.512105	0.000007	N/A	1850	2.8
105-09-NIM 062 558.8	Intermediate to mafic gneiss	5.79	25.34	0.1382	0.512089	0.000007	2.16	1850	3.2
Talbot Lake deposit									
105-09-TLS 010 204.5	Intermediate quartzofeldspathic gneiss (~20% biotite)	5.40	20.18	0.1619	0.512363	0.000005	N/A	1850	2.9
105-09-TLS 010 345.8	Felsic quartzofeldspathic gneiss (~10% biotite)	12.06	55.18	0.1321	0.511999	0.000008	2.17	1850	2.9
105-09-TLS 010 645.0	Intermediate to felsic quartzofeld-spathic gneiss*	4.05	19.13	0.1279	0.511942	0.000008	2.16	1850	2.8
105-09-TLS 010 679.5	Garnetiferous amphibolite	2.45	7.50	0.1978	0.51284	0.000006	N/A	1850	3.7

* host mineralization or closely associated with mineralization

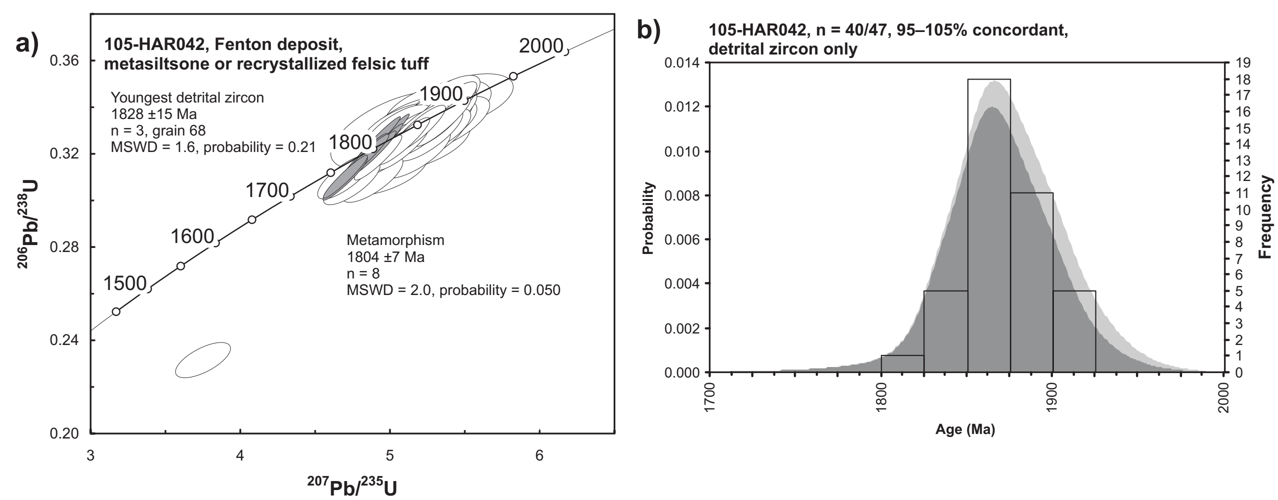

***Figure GS-6-8:** U-Pb data from the sub-Phanerozoic Flin Flon Belt and Kisseynew Domain: **a)** concordia diagram for Fenton deposit sample, with ellipses plotted at the 2σ uncertainty level; light grey ellipses represent analyses of high U metamorphic rims; white ellipses represent detrital zircon cores; **b)** probability-density diagram for Fenton deposit sample, the dark grey curve and histograms representing data that are within 5% of concordia; light grey curve represents all data regardless of discordance; replicate analyses and results from metamorphic rims are not included.*

In an attempt to constrain the maximum age of sediment deposition, replicate analyses on the youngest analyzed detrital zircon yielded a mean $^{207}Pb/^{206}Pb$ age of 1824 ±15 Ma. This relatively imprecise age, however, may have been affected by the strong ~1800 Ma metamorphic overprint. Assuming that the unimodal ~1870 Ma zircon population is from a single igneous/volcanic source, it could represent an age estimate for volcanism and VMS mineralization in the Fenton deposit area.

Harmin deposit

Location, cover, drillhole depth: Approximately 3 km northeast of the Fenton deposit, ~70 km southeast of Snow Lake; 20 m of Pleistocene glacial overburden and ~50 m of Phanerozoic sedimentary rocks; ~700 m.

Stratigraphy: This is very similar in every aspect to the Fenton deposit except for increased presence of late felsic intrusive rocks (up to 50%) crosscutting the stratigraphy. The stratigraphy is dominantly an upper succession of mafic to intermediate gneisses, and felsic gneiss at greater depths.

Mineralization: Hosted in the upper portion of the felsic gneiss, it forms disseminated (pyrrhotite, pyrite, sphalerite and chalcopyrite) to near-solid sulphide intervals of medium- to coarse-grained pyrrhotite, pyrite, sphalerite and chalcopyrite.

Metamorphic facies: Upper amphibolite to lower granulite.

Talbot Lake deposit

Location, cover, drillhole depth: Approximately 80 km south-southeast of Snow Lake; 5 m of Pleistocene glacial overburden and ~100 m of Phanerozoic sedimentary rocks; >850 m.

Stratigraphy: Comprises several 100–300 m thick, mainly mafic and intermediate gneiss sequences with well-foliated garnetiferous amphibolite intervals. Thinner (<15 m) intervals of felsic quartzofeldspathic gneiss are found toward the top of the intermediate gneiss intervals. The mafic gneiss is much more abundant in this deposit compared to other 'higher metamorphic-grade' deposits described here. The mafic gneiss is very biotite rich (60–70%) and usually garnetiferous. Ductile to brittle fault zones occur throughout the stratigraphic sequence.

Mineralization: The main mineralization, which is hosted in the most felsic gneiss, consists of disseminated to near-solid sulphides such as pyrite, chalcopyrite, sphalerite and pyrrhotite. A 'hangingwall brittle fault zone' occurs in almost all of the logged holes, located approximately 10–30 m into the hangingwall of the mineralization.

Metamorphic facies: Middle to upper amphibolite (hornblende, biotite, garnet).

Alteration: Extensive calcsilicate alteration zones occur throughout the stratigraphic sequence.

Geochronology: A sample of quartzofeldspathic gneiss (105-TLS010) from immediately below the mineralization was collected for U-Pb geochronology; only metamorphic zircons were recovered (high U, low Th/U; see Data Repository Item DRI2010006 for details).

Geochemistry of the 'higher metamorphic-grade volcanic and sedimentary rocks'

All gneisses of the 'higher metamorphic-grade volcanic and sedimentary rocks' have relatively low Nb/Y ratios (~0.3), suggesting a subalkaline affinity (Figure GS-6-5c). Based on their 'silica content' (Zr/TiO_2; Figure GS-6-5c) and other geochemical characteristics, they show three distinct geochemical signatures, as detailed in the following paragraphs.

The intermediate to felsic (rhyolite/dacite; Figure GS-6-5c) gneisses of all the studied deposits have REE patterns characterized by slight LREE enrichment, flat HREE patterns and pronounced negative Nb and Ti anomalies (Figure GS-6-9a). On a Th-(Zr/117)-Nb/16 diagram, they appear to be either products of subduction-related processes in an oceanic-arc environment (felsic volcanic/volcaniclastic rocks) or siliciclastic rocks derived from an evolved continental source (Figure GS-6-5). Their moderate $(La/Yb)_N$ ratios suggest a tholeiitic to transitional magmatic affinity (Figure GS-6-9).

The mafic gneisses (andesite/basalt; Figure GS-6-5c) of the Watt's River and Talbot deposits, when compared to primitive mantle, have trace-element and REE patterns enriched in incompatible elements, which is typical of E-MORB, along with slightly negative Nb and Ti anomalies commonly associated with subduction-related processes (Figure GS-6-9a). Their distribution away from the Nb pole in the Th-(Zr/117)-Nb/16 diagram (Figure GS-6-5d) suggests more of an E-MORB signature for these rocks, but with some contamination from subduction processes (contaminated mantle). The mafic gneisses of the Fenton and Harmin deposits are basaltic in composition (Figure GS-6-5c) and have trace-element and REE patterns typical of E-MORB (Figures GS-6-5d, -9b) without Nb and Ti anomalies. They most likely represent E-MORB basalt derived from an 'uncontaminated' mantle.

The well-foliated amphibolite intervals (basalt; Figure GS-6-5c) of the Talbot Lake deposit have flat to very slightly LREE-depleted REE patterns and are depleted in strongly incompatible trace elements with no significant Nb or Ti anomalies, typical of N-MORB (Figure GS-6-9d). They also plot within the N-MORB field on the Th-(Zr/117)-Nb/16 diagram (Figure GS-6-5d), which suggests, along with the other geochemical characteristics indicated above, that they were the product of mid-ocean-ridge magmatism. A Sm-Nd isotopic analysis from one amphibolite sample yielded an $\varepsilon_{Nd\,(1850\,Ma)}$ value of +3.8 (Table GS-6-1), which suggests that the amphibolite was derived from a juvenile magma source with minimum to

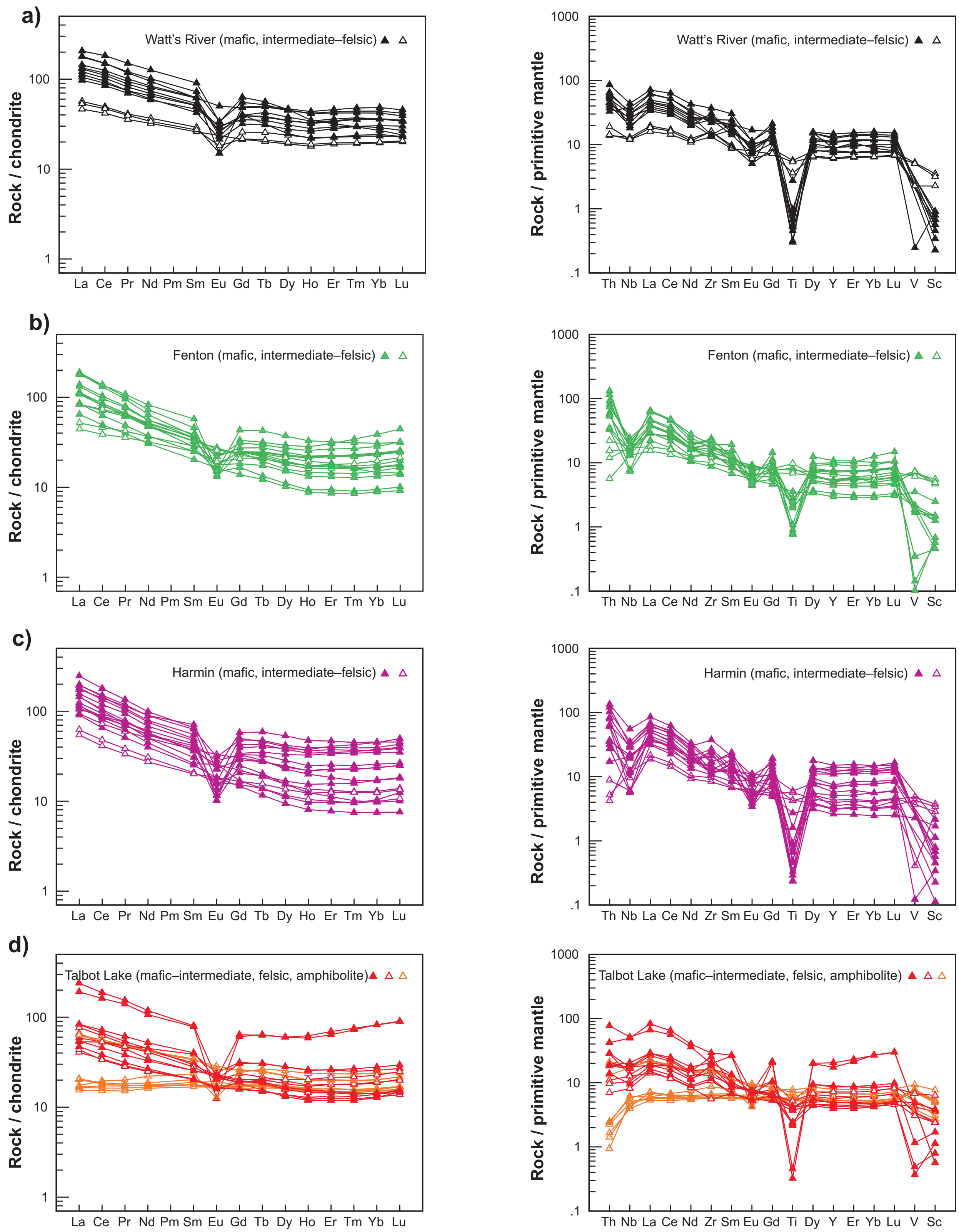

***Figure GS-6-9:** Chondrite-normalized rare earth element (REE) patterns (normalizing values from McDonough and Sun, 1995) and mantle-normalized incompatible trace-element patterns (normalization values from Sun and McDonough, 1989) for the **a)** Watt's River deposit, **b)** Fenton deposit, **c)** Harmin deposit, and **d)** Talbot Lake deposit.*

no involvement of continental crust in the magma genesis.

Analyses of Sm-Nd isotopes from both mafic and intermediate-felsic rocks from the Fenton, Harmin and Talbot Lake deposits yielded $\varepsilon_{Nd\,(1850\,Ma)}$ values of +2.7 to +3.2 (Table GS-6-1), suggesting a juvenile magma source with minimum to no involvement of continental crust in the magma genesis.

The felsic rocks hosting the base-metal deposits may be felsic siliciclastic rocks derived from an evolved continental source. Alternatively, the association of 'arc-derived' felsic rocks and contaminated E-MORB mafic rocks, similar to that observed for the Watt's River deposit, may suggest that they formed in a rifted arc and/or at the onset of back-arc magmatism. Samarium-neodymium isotope and geochronological analyses of these rocks are pending.

The lack of older inherited zircons, as well as the positive $\varepsilon_{Nd\,(1850\,Ma)}$ values for the rocks of the Fenton deposit, suggest that they were derived from a juvenile magmatic source (volcanic or plutonic) and were not proximal to or sourced by older, evolved continental crust. The lack of arc contamination in the E-MORB geochemical signature suggests that the rifted basin was slightly more mature (wider) when the Fenton and Harmin deposits formed, compared to the Watt's River deposit.

The association of juvenile 'arc-derived' felsic rocks and E-MORB mafic rocks suggests that the Talbot Lake deposit, like the Watt's River deposit, formed in a rifting arc and/or at the onset of back-arc magmatism. The great abundance of N-MORB amphibolite within the succession suggests either that these are late dikes of different affinity or that the rifted basin in which they were emplaced was wide enough for normal mid-ocean-ridge magmatism (ocean-floor spreading) concurrent with formation of the Talbot Lake deposit.

Discussion and economic considerations

Preliminary observations of the stratigraphy, whole-rock geochemistry and Sm-Nd isotope data of the six documented VMS-hosting successions from the 'lower metamorphic-grade volcanic and sedimentary rocks' domain (Figure GS-6-3) lead to one main conclusion: all of these VMS deposits are hosted in rocks that formed in a bimodal tholeiitic to transitional oceanic-arc setting, and were derived from a juvenile magma source with little to no involvement of continental crust in the magma genesis. The recognition of massive flows and breccias with very little volcaniclastic material in the least deformed of these strata (Moose, Kofman, Reed) suggests that they represent proximal volcanic deposits. Subsequent tectonometamorphism has obliterated part or all of the primary textures in some of these volcanic successions (Limestone, Sylvia, Rail), in some cases making it difficult to identify protoliths. Whole-rock geochemistry in these 'lower metamorphic-grade rocks', however, has generally not been 'altered' to the same extent as in the 'higher metamorphic-grade rocks' and can commonly be used to assess source rocks and protoliths. At this stage of the project, collection of additional geochronological data is needed/underway to more fully assess whether the volcanic successions that were later juxtaposed in multiple distinct fault panels (Figure GS-6-3) represent one or multiple dismembered tholeiitic to transitional arc system(s). Additional geochronological data will also allow for more informed correlations of the various sub-Phanerozoic successions with those of the exposed portion of the belt.

Compared to 'the lower metamorphic-grade rocks' domain, tentative data for the four base-metal (VMS?)–hosting successions studied from the 'higher metamorphic-grade volcanic and sedimentary rocks' domain (Figure GS-6-3) suggest they formed in a different tectonic setting. Although these rocks have been metamorphosed at upper-amphibolite to lower-granulite facies and none of the primary textures remain, protolith recognition (igneous versus sedimentary) may be possible by applying a combination of whole-rock geochemistry and Sm-Nd isotope analyses. All data collected so far suggest that these deposits formed in a rifting arc and/or at the onset of back-arc magmatism. The maturity or width of the 'rifted basin' in which these deposits formed may have varied slightly from one to the other, as suggested by the varying amounts of subduction-related contamination in the source mantle. Zircon U-Pb age data for the Fenton deposit hostrock suggest that part of the rifting arc formed ca. 1865 Ma, which could represent an age estimate for volcanism and VMS mineralization in this area. If it does, these VMS deposits would therefore be younger than the ones found in the Flin Flon or Snow Lake arc assemblages (~1.90–1.89 Ga) and would be coeval with the 'successor-arc' magmatism (~1.88–1.83 Ga). Similar ages have been obtained in recent geochronological studies of sub-Phanerozoic VMS deposits in the Saskatchewan portion of the Flin Flon–Glennie Domain (Morelli et al., work in progress).

All VMS deposits found to date in the exposed portion of the Flin Flon Belt are solely located within the juvenile, 1.90–1.89 Ga arc-volcanic rocks. Tentative data presented here, however, provide evidence for a younger, prospective, ~1.87 Ga spreading back-arc basin of 'successor-arc' age in the eastern portion of the belt, as well as prospective tholeiitic to transitional oceanic-arc environments farther west. This clearly demonstrates that combined stratigraphic, geochemical, isotopic and geochronological data have the potential to unravel the tectonostratigraphic affinity of most sub-Phanerozoic strata, even those that are highly deformed and/or metamorphosed, providing integrated and modern exploration models for the hidden VMS deposits in the various 'domains' of the belt.

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References

Ansdell, K.M. 1993: U-Pb zircon constraints on the timing and provenance of fluvial sedimentary rocks in the Flin Flon and Athapapuskow basins, Flin Flon Domain, Trans-Hudson Orogen; *in* Radiogenic Age and Isotopic Studies: Report 7, Geological Survey of Canada, Paper 93-02, p. 49–57.

Ansdell, K.M., Connors, K.A., Stern, R.A. and Lucas, S B. 1999: Coeval sedimentation, magmatism, and fold-thrust belt development in the Trans-Hudson Orogen: geochronological evidence from the Wekusko Lake area, Manitoba, Canada; Canadian Journal of Earth Sciences, v. 36, p. 312–293.

Ansdell, K.M., Kyser, T.K., Stauffer, M.R. and Edwards, G. 1992: Age and source of detrital zircon from the Missi Formation: a Proterozoic molasse deposit, Trans-Hudson Orogen, Canada; Canadian Journal of Earth Sciences, v. 29, p. 2583–2594.

Ansdell, K.M., Lucas, S.B., Connors, K.A. and Stern, R.A. 1995: Kisseynew metasedimentary gneiss belt, Trans-Hudson Orogen (Canada): back-arc origin and collisional inversion; Geology, v. 23, no. 11, p. 1039–1043.

Bailes, A.H. 1975: Geology of the Guay-Wimapedi lakes area; Manitoba Mines, Resources and Environmental Management, Mineral Resources Division, Publication 75-2, 104 p.

Bailes, A.H. 1980a: Geology of the File Lake area; Manitoba Energy and Mines, Mineral Resources Division, Geological Report GR78-1, 134 p.

Bailes, A.H. 1980b: Geology of the File Lake area; Manitoba Energy and Mines, Mineral Resources Division, Geological Report GR78-1, 134 p.

Bailes, A.H. 1980c: Origin of early Proterozoic volcaniclastic turbidites, south margin of the Kisseynew sedimentary gneiss belt, File Lake, Manitoba; *in* Early Precambrian Volcanology and Sedimentology in the Light of the Recent, Precambrian Research, v. 12, no. 1–4, p. 197–225.

Bailes, A.H. and Galley, A.G. 2007: Geology of the Chisel–Anderson lakes area, Snow Lake, Manitoba (NTS areas 63K16SW and west half of 63J13SE); Manitoba Science, Technology, Energy and Mines, Manitoba Geological Survey, Geoscientific Map MAP2007-1, scale 1:20 000.

Bailes, A.H. and McRitchie, W.D. 1978: The transition from low to high grade metamorphism in the Kisseynew sedimentary gneiss belt, Manitoba; *in* Metamorphism in the Canadian Shield, J.A. Fraser and W.W. Heywood (ed.), Geological Survey of Canada, Paper 78-10, p. 155–178.

Bailes, A.H. and Syme, E.C. 1989: Geology of the Flin Flon–White Lake area; Manitoba Energy and Mines, Mineral Resources Division, Geological Report GR87-1, 313 p.

Connors, K.A. and Ansdell, K.M. 1993: Transition between the Flin Flon and Kisseynew domains, File Lake–Limestone Point Lake area, northern Manitoba; *in* Report of Activities 1993, Manitoba Energy and Mines, Geological Services, p. 42–46.

Connors, K.A. and Ansdell, K.M. 1994: Revision of stratigraphy and structural history in the Wekusko Lake area, eastern Trans-Hudson Orogen; *in* Report of Activities 1994, Manitoba Energy and Mines, Geological Services, p. 104–107.

David, J., Bailes, A.H. and Machado, N. 1996: Evolution of the Snow Lake portion of the Palaeoproterozoic Flin Flon and Kisseynew belts, Trans-Hudson Orogen, Manitoba, Canada; Precambrian Research, v. 80, p. 124–107.

David, J., Machado, N. and Bailes, A.H. 1993: U-Pb geochronology of the Proterozoic Flin Flon Belt, Snow Lake area, Manitoba; Geological Association of Canada–Mineralogical Association of Canada, Joint Annual Meeting, Program with Abstracts, p. 22.

Fedorowich, J.S., Kerrich, R. and Stauffer, M. 1995: Geodynamic evolution and thermal history of the central Flin Flon Domain, Trans-Hudson Orogen: constraints from structural development, Ar/Ar and stable isotope geothermometry; Tectonics, v. 14, p. 472–503.

Ferreira, K.J., Mitchell, J. and Gula, D. 1996: Mineral deposits and occurrences in the south half of the Cormorant Lake area, (NTS 63K/1 to 63K/8); Manitoba Energy and Mines, Geological Services, Mineral Deposit Series Report No. 33, 150 p.

Froese, E., Lemkow, D.R. and Fedikow, M.A.F. 1989: Metamorphism of the host rocks to the North Cook Lake massive sulphide-type copper deposit, Snow Lake area (NTS 63K/16); *in* Report of Field Activities 1989, Manitoba Energy and Mines, Minerals Division, p. 139–144.

Galley, A.G., Bailes, A.H. and Kitzler, G. 1993: Geological setting and hydrothermal evolution of the Chisel Lake and North Chisel Zn-Pb-Ag-Au massive sulphide deposit, Snow Lake, Manitoba; Exploration and Mining Geology, v. 2, p. 271–295.

Gordon, T.M., Hunt, P.A., Bailes, A.H. and Syme, E.C. 1990: U-Pb ages from the Flin Flon and Kisseynew belts, Manitoba: chronology of crust formation at an Early Proterozoic accretionary margin; *in* The Early Proterozoic Trans-Hudson Orogen of North America, J.F. Lewry and M.R. Stauffer (ed.), Geological Association of Canada, Special Paper 37, p. 177–199.

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.

HudBay Minerals Inc. 2007: Technical report, Watt's River deposit, Snow Lake, Manitoba; HudBay Minerals Inc., 102 p.

Hudson Bay Exploration and Development Co. Ltd. 1998: Private company reports on the Rail deposit, Reed Lake deposit and Sylvia deposit; Hudson Bay Exploration and Development Co. Ltd.

Leclair, A.D., Lucas, S.B., Broome, H.J., Viljoen, D.W. and Weber, W. 1997: Regional mapping of Precambrian basement beneath Phanerozoic cover in southeastern Trans-Hudson Orogen, Manitoba and Saskatchewan; Canadian Journal of Earth Sciences, v. 34, p. 634–618.

Lewry, J.F. 1990: Lithoprobe Trans-Hudson Orogen transect: general geology of the transect area and an overview of currently planned and needed research; *in* LITHOPROBE: Trans-Hudson Orogen Transect, LITHOPROBE Secretariat, University of British Columbia, Report 12, p. 5–6.

Lucas, S.B., Stern, R.A., Syme, E.C., Reilly, B.A. and Thomas, D.J. 1996: Intraoceanic tectonics and the development of continental crust; 1.92–1.84 Ga evolution of the Flin Flon Belt, Canada; Geological Society of America Bulletin, v. 108, p. 602–629.

Ludwig, K.R. 2003: User's manual for Isoplot 3.00: a geochronological toolkit for Microsoft Excel; Berkeley Geochronology Center, Special Publication 4, 74 p.

Macek, J.J., Zwanzig, H.V. and Pacey, M. 2006: Thompson Nickel Belt geological compilation map, Manitoba (parts of NTS 63G, J, O, P and 64A and B); Manitoba Science, Technology, Energy and Mines, Manitoba Geological Survey, Open File OF2006-33, 1 CD-ROM.

McDonough, W.F. and Sun, S.-s. 1995: The composition of the Earth; *in* Chemical Evolution of the Mantle, W.F. McDonough, N.T. Arndt and S. Shirey (ed.), Chemical Geology, v. 120, p. 223–253.

NATMAP Shield Margin Project Working Group 1998: Geology, NATMAP Shield Margin Project area, Flin Flon Belt, Manitoba/Saskatchewan; Geological Survey of Canada, Map 1968A, 1:100 000 scale.

Rockcliff Resources Inc. 2007: Independent technical report: Snow Lake Project, Manitoba, Canada; Rockcliff Resources Inc., 91 p.

Rockcliff Resources Inc. 2008a: Private company report of the Kof property, Snow Lake, Manitoba; Rockcliff Resources Inc.

Rockcliff Resources Inc. 2008b: Private company report on the Sylvia property, Snow Lake, Manitoba; Rockcliff Resources Inc., 117 p.

Simard, R-L. and McGregor, C.R. 2009: Toward a new sub-Phanerozoic Precambrian basement map of the Flin Flon Belt, Manitoba (parts of NTS 63J, K, L); *in* Report of Activities 2009, Manitoba Innovation, Energy and Mines, Manitoba Geological Survey, p. 15–21.

Spear, F.S. and Cheney, J.T. 1989: A petrogenetic grid for pelitic schists in the system SiO_2-Al_2O_3-FeO-MgO-K_2O-H_2O; Contributions to Mineralogy and Petrology, v. 134, p. 17–32.

Stern, R.A. 1997: The GSC sensitive high resolution ion microprobe (SHRIMP): analytical techniques of zircon U-Th-Pb age determinations and performance evaluation; *in* Radiogenic Age and Isotopic Studies, Report 10, Geological Survey of Canada, Current Research 1997-F, p. 1–31.

Stern, R.A. and Amelin, Y. 2003: Assessment of errors in SIMS measurement of zircon U-Pb isotopic ages using a natural zircon standard (6266) and NIST SRM 610 glass; Chemical Geology, v. 197, p. 111–142.

Stern, R.A., Lucas, S.B., Syme, E.C., Bailes, A.H., Thomas, D.J., Leclair, A.D. and Hulbert, L. 1993: Geochronological studies in the Flin Flon Domain, Manitoba-Saskatchewan, NATMAP Shield Margin Project area: results for 1992–1993; *in* Radiogenic Age and Isotopic Studies: Report 7, Geological Survey of Canada, Paper 93-02, p. 59–70.

Stern, R.A., Syme, E.C., Bailes, A.H. and Lucas, S.B. 1995a: 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.

Stern, R.A., Syme, E.C. and Lucas, S.B. 1995b: Geochemistry of 1.9 Ga MORB- and OIB-like basalts from the Amisk collage, Flin Flon Belt, Canada; evidence for an intra-oceanic origin; Geochimica et Cosmochimica Acta, v. 59, p. 3131–3154.

Sun, S.-s. and McDonough, W.F. 1989: Chemical and isotopic systematics of oceanic basalts: implication for mantle composition and processes; Geological Society, Special Publication 42, p. 313–345.

Syme, E.C. and Bailes, A H. 1993: Stratigraphy and tectonic setting of Early Proterozoic volcanogenic massive sulphide deposits, Flin Flon, Manitoba; Economic Geology, v. 88, p. 566–589.

Syme, E.C., Lucas, S.B., Bailes, A.H. and Stern, R.A. 1999a: Contrasting arc and MORB-like assemblages in the Paleoproterozoic Flin Flon Belt, Manitoba, and the role of intra-arc extension in localizing volcanic-hosted massive sulphide deposits; Canadian Journal of Earth Sciences, v. 36, p. 1788–1767.

Syme, E.C., Lucas, S.B., Bailes, A.H. and Stern, R.A. 1999b: Contrasting arc and MORB-like assemblages in the Paleoproterozoic Flin Flon Belt, Manitoba, and the role of intra-arc extension in localizing volcanic-hosted massive sulphide deposits; Canadian Journal of Earth Sciences, v. 36, p. 1767–1788.

Whalen, J.B., Syme, E.C. and Stern, R.A. 1999: Geochemical and Nd isotopic evolution of Paleoproterozoic arc-type granitoid magmatism in the Flin Flon Belt, Trans-Hudson Orogen, Canada; Canadian Journal of Earth Sciences, v. 36, p. 250–227.

Winchester, J.A. and Floyd, P.A. 1977: Geochemical discrimination of different magma series and their differentiation products using immobile elements; Chemical Geology, v. 20, no. 4, p. 325–343.

Wood, D.A. 1980: The application of a Th-Hf-Ta diagram to problems of tectonomagmatic classification and to establishing the nature of crustal contamination of basaltic lavas of the British Tertiary volcanic province; Earth and Planetary Science Letters, v. 50, p. 11–30.

Young, G.M. 2002: Geochemical investigation of a Neoproterozoic glacial unit: the Mineral Fork Formation in the Wasatch Range, Utah; Geological Society of America Bulletin, v. 114, no. 4, p. 387–399.

Zwanzig, H.V. 1990: Kisseynew gneiss belt in Manitoba: stratigraphy, structure, and tectonic evolution; *in* The Early Proterozoic Trans-Hudson Orogen of North America, J.F. Lewry and M.R. Stauffer (ed.), Geological Association of Canada, Special Paper 37, p. 95–120.

Zwanzig, H.V., Parent, M. and Machado, N. 1996: Kisseynew Belt in Manitoba: 1.84 to 1.79 Ga staging of progressive continental collision and tectonic wedging; Geological Association of Canada–Mineralogical Association of Canada, Joint Annual Meeting, Program with Abstracts, v. 21, p. A105.