ERRATUM

Geology and geochemical evolution of the northern Flin Flon and southern Kisseynew domains, Kississing–File lakes area, Manitoba (parts of NTS 63K, N)

by H.V. Zwanzig and A.H. Bailes
Manitoba Geological Survey
Manitoba Innovation, Energy and Mines

GR2010-1.pdf reprinted with revisions, 2011

The following items have been revised:

Page 66, Table 15: The data in the fifth, sixth, seventh, and eleventh columns were revised to have six decimal places.

Page 94: “Pegmatitic granitoid rocks emplaced during protracted deformation have yielded U-Pb ages from monazite and zircon of ca. 1812–1789 Ma (Parent et al., 1999).”

GR2010-1

GEOSCIENTIFIC REPORT

Geology and geochemical evolution of the northern Flin Flon and southern Kisseynew domains, Kississing–File lakes area, Manitoba (parts of NTS 63K, N)

By
H.V. Zwanzig and A.H. Bailes

Manitoba

Geoscientific Report GR2010-1

Geology and geochemical evolution of the northern Flin Flon and southern Kisseynew domains, Kississing–File lakes area, Manitoba (parts of NTS 63K, N)

by H.V. Zwanzig and A.H. Bailes
Winnipeg, 2010

Innovation, Energy and Mines

Hon. Dave Chomiak
Minister

John Clarkson
Deputy Minister

Mineral Resources Division

John Fox
Assistant Deputy Minister

Manitoba Geological Survey

E.C. Syme
Director

When using information from this publication in other publications or presentations, due acknowledgment should be given to the Manitoba Geological Survey. The following reference format is recommended:

Zwanzig, H.V. and Bailes, A.H. 2010: Geology and geochemical evolution of the northern Flin Flon and southern Kisseynew domains, Kississing–File lakes area, Manitoba (parts of NTS 63K, N); Manitoba Innovation, Energy and Mines, Manitoba Geological Survey, Geoscientific Report GR2010-1, 135 p.

NTS grid: parts of NTS 63K13–16, 63N1–4

Keywords: File Lake; Flin Flon Domain; geochemistry; igneous rocks; intrusions; Kisseynew Domain; Kississing Lake; Manitoba; metamorphism; mineral deposits, genesis; structural geology; tectonics; tectonostratigraphic units; volcanic rocks

Published by:
Manitoba Innovation, Energy and Mines
Manitoba Geological Survey
360–1395 Ellice Avenue
Winnipeg, Manitoba
R3G 3P2 Canada
Telephone: (800) 223-5215 (General Enquiry)
(204) 945-4154 (Publication Sales)
Fax: (204) 945-8427
E-mail: minesinfo@gov.mb.ca
Website: manitoba.ca/minerals

Cover illustration: Moss- and lichen-free outcrops of Missi Group volcanic rocks (dark) within metasediments after the forest fire of 1989.

TABLE OF CONTENTS

Page

APPENDICES

TABLES

FIGURES

MAPS

PRINT/DIGITAL COMPONENTS AND DISTRIBUTION FORMATS

The print format of GR2010-1 includes an accompanying DVD in back pocket to accommodate the digital only components of the report. The digital only format of GR2010-1 is a stand-alone DVD product.

	Print components	**Digital components on accompanying/ stand-alone DVD**
Report GR2010-1	√	√ PDF
Oversize Figures 8 and 43	√ (folded in back pocket)	√ PDF
Appendices 1–3	√	√ PDF
Appendix 4		√ Excel
Appendix 5		√ PDF
Accompanying Map GR2010-1-1	√ (folded in back pocket)	√ PDF
GIS Map GR2010-1-2		√ ArcReader project

Summary

This report focuses on adjacent and structurally overlapping parts of two lithotectonic domains (Flin Flon and Kisseynew) in the Paleoproterozoic Trans-Hudson Orogen (THO) in Manitoba (Figure 1). The supracrustal rocks of the Flin Flon Domain to the south are dominantly metavolcanic, whereas those of the Kisseynew Domain are mainly metasedimentary. The change is also from medium-grade metamorphic rocks preserving early structures in the south to highly metamorphosed rocks with deeper level structures in the north (Figure 2). The metavolcanic rocks have been newly subdivided in detail, based partly on their variably preserved field characteristics and U-Pb zircon ages but mostly on their surprisingly well-preserved geochemistry. These units and the revised structural geology provide a new control on the presence and distribution of volcanogenic massive sulphide (VMS) and gold deposits in the Kississing–File lakes area.

In previous reports (Zwanzig, 1990, 1999; Zwanzig and Schledewitz, 1992), the change in metamorphic grade and structural style was used to define the domain boundary because of structural complexity and an abundance of intrusive rocks. Units were defined mainly on mineral contents. In this report, the mainly volcanic rocks and early intrusions are recognized as structurally contiguous with the lower grade rocks to the south. They have the same high mineral potential and are therefore separated in detail from the younger sedimentary and intrusive rocks, a procedure started by Bailes and McRitchie (1978) and Bailes (1980a) before U-Pb zircon dating. Age and inferred primary lithology now serve as a better subdivision for describing the origins and metallogeny of these rocks (Ashton and Froese, 1988). A new regional division into tectonic subdomains (Figure 2) helps to elucidate their evolution and the potential for future discoveries of mineral deposits in different areas.

The area considered in this report extends 30 km south across Kississing Lake, near the Saskatchewan border, to Kisseynew Lake, and 80 km southeast to File Lake (Figure 2). It includes the main area of crustal-scale overlap and interleaving of the various sedimentary and volcanic units that are newly defined and described in this report. Although more difficult to identify at higher grade and greater strain than in the main part

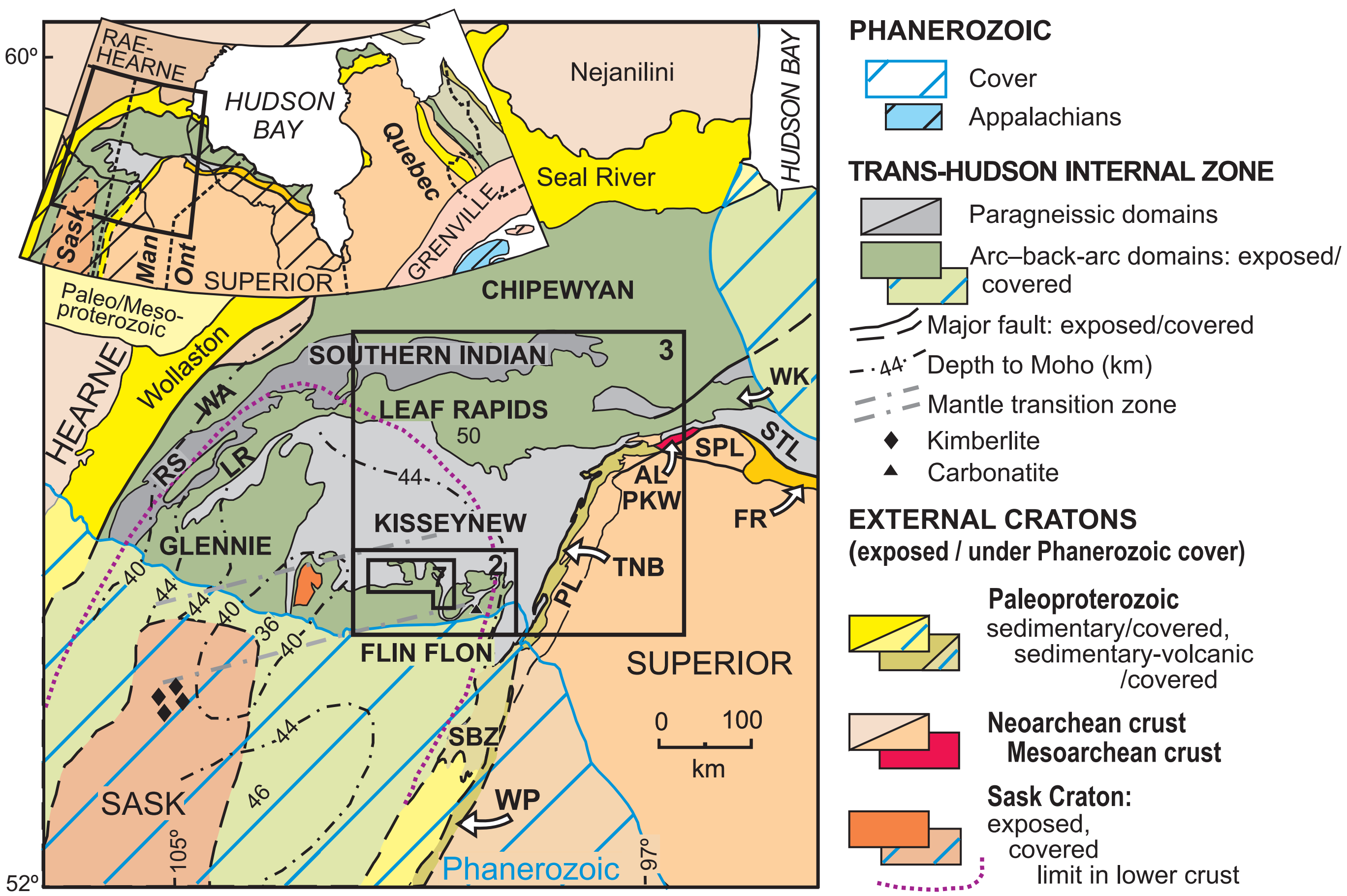


Figure 1: *Simplified geology of the Trans-Hudson Orogen (THO) in Manitoba and northwestern Saskatchewan, showing the major tectonic domains as exposed and beneath Phanerozoic cover. Distinguished are the internal zone (Reindeer Zone), Archean cratons and their Paleoproterozoic cover (external zones). Extent of the Sask craton is from Kreis et al. (2000) and its maximum extent in the lower crust is from Hajnal et al. (2005). Depth to mantle is contoured from seismic refraction profiles of Németh et al. (2005); mantle transition zone and approximate location of kimberlites are also after Németh et al. (2005). Domain names (west to east): WA, Wathaman; RS, Rottenstone; LR, La Ronge; SBZ, Superior Boundary Zone; WP, Winnipegosis; TNB, Thompson Nickel Belt; PL, Paint Lake; PKW, Pikwitonei granulite; AL, Assean Lake (Mesoarchean); SPL, Split Lake; FR, Fox River; STL, Stephens Lake; WK, Waskaiowaka. Shear zone: OSZ, Owl River Shear Zone. Locations of Figures 2, 3 and 7 are shown by the polygon with the appropriate number. Location is on the inset map, which shows the wider THO (eastern and western parts with respect to Hudson Bay).*

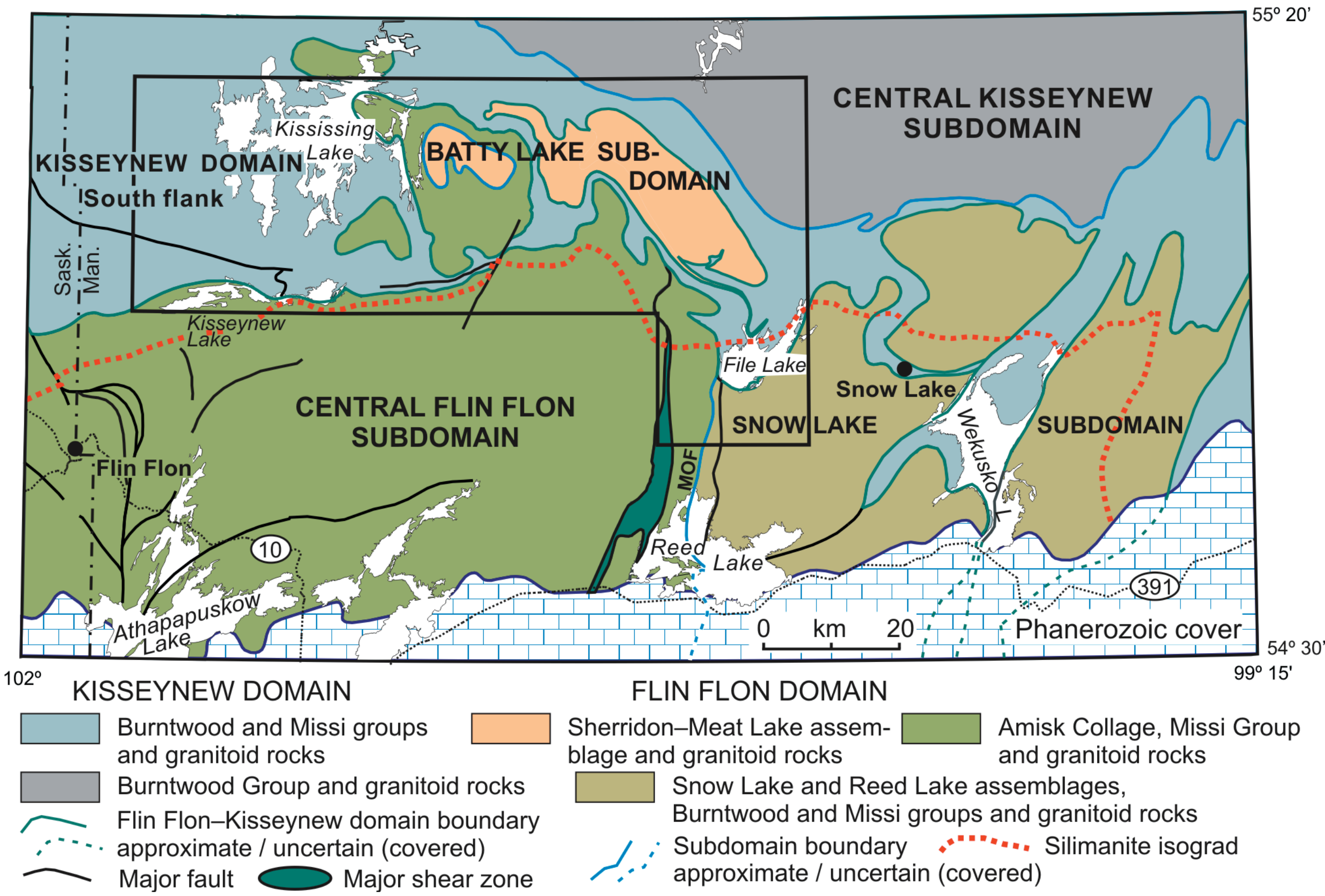


Figure 2: *Subdomains of the Flin Flon Domain and southern Kisseynew Domain in Manitoba, also showing the outline of the Kississing–File lakes area and the sillimanite-biotite-garnet isograd.*

of the Flin Flon Domain, units are traced north using a combination of 1) field criteria, 2) petrography, 3) structural and stratigraphic continuity, 4) careful study of their geochemistry and Nd isotopes, and 5) application of the existing U-Pb zircon ages.

The regional tectonostratigraphy defines several packages of mappable units regardless of whether contacts are faulted or depositional and merely sheared. Tracing the packages is assisted by previous studies (Bailes, 1980a; Zwanzig, 1990, 1999; Zwanzig and Schledewitz, 1992; Norman et al., 1995; Connors, 1996; Connors et al., 1999; Kraus and Williams 1999) and by the unique trace-element fingerprints of the various volcanic units in the Kississing–File lakes area. The oldest, predominantly volcanic rocks (Amisk Collage, Northeast Reed assemblage and Snow Lake assemblage) are recognized because they are intruded by the widespread layered sills of the 1886 Ma Josland Lake gabbro (Zwanzig et al., 2001). The geochemistry and exceptionally strong fractionation trends of these sills leads to their wide correlation and helps to determine tops at their early time of intrusion. The origins of the older volcanic assemblages within these packages are modelled on the lower grade rocks to the south and on more modern rocks with similar geochemistry. The geochemistry compares well to arc, arc-rift and back-arc assemblages (Stern et al., 1995a, b). An important result of this study is that the high-grade metavolcanic rocks have retained their immobile trace-element pattern sufficiently to enable recognition of their origin. Thus, the conclusion of Syme et al. (1999) that the arc and arc-rift assemblages in the Flin Flon Domain are most favourable for hosting VMS deposits can be applied up to the northernmost part of the Kississing–File lakes area. An important arc assemblage is the Fourmile Island assemblage, which is interpreted to include arc-rift volcanic rocks in the north that host the Dickstone VMS deposit. This assemblage is traced the full length of the Kississing–File lakes area and provides a good exploration target.

The field relationships, petrography and geochemistry of the abundant plutonic rocks that formed throughout the evolution of the area compare closely with those of coeval volcanic rocks, and early, middle and late successor-arc intrusions in the south (Whalen et al., 1999). The U-Pb ages of the plutons help to provide a time frame for the regional tectonic evolution of units that separate into those with high or low mineral potential. The intrusive suites are presented in this report as 1) coeval with the early volcanic rocks; 2) early successor arc, predating the sedimentary rocks typical of the Kisseynew Domain; and 3) late successor arc, coeval with and postdating these sedimentary rocks. An important new conclusion, based on a preliminary U-Pb zircon age (N. Rayner and D. Tinkham, pers. comm., 2009), is that the highly metamorphosed felsic rocks in the Sherridon–Batty Lake area (Zwanzig and Schledewitz, 1992) represent VMS-hosting, early successor-arc volcanic rocks (Sherridon–Meat Lake assemblage). Approximately coeval intrusions in the Batty Lake subdomain (Figure 2) may have acted as hydrothermal heat engines. This opens the possibility of new base-metal exploration where marine arc rocks of this age are present in other areas.

The younger metasedimentary rocks, most typical of the Kisseynew Domain, comprise the Burntwood Group of greywacke-mudstone turbidite (File Lake formation of Bailes, 1980a) and the Missi Group (Bailes, 1971; Stauffer, 1990), a syntectonic overlap assemblage of shallow-water sediments with locally intercalated volcanic rocks. These are readily recognized at all metamorphic grades. Their stratigraphy and continuity help to define structurally intact sequences. Typical columns constructed in the supracrustal rocks show that these types of sequences can be traced with their underlying older volcanic and intrusive rocks, and correlated for up to 100 km. A major unconformity clearly defines the stratigraphic base of the Missi Group and therefore way-up during its deposition. The relative abundance of the sedimentary rocks and their proposed sedimentary facies help to define the boundary of the paleobasin that formed the Kisseynew Domain at the margin of the Flin Flon Domain. The younger supracrustal units and coeval intrusions also elucidate the character of the latest arc magmatism that occurred during the onset of collision tectonics in the THO. This stage of evolution has not yielded VMS deposits to date, probably because it was subaerial. The distinctive geochemistry of these rocks helps to eliminate barren geology from mineral exploration. Similarly, the youngest intrusive rocks, which are concentrated in the Kisseynew Domain, are distinguished by their own distinctive geochemical patterns and Nd isotope ratios. These, too, can be eliminated at present from VMS and gold exploration.

This report recognizes that the structural style and northerly trend in the central part of the Flin Flon Domain previously existed also on its northern margin (Zwanzig, 1999). This conclusion has a profound influence on recognizing how the tectonostratigraphic units discussed in this report were juxtaposed. The structural style changes along the strike of the assembled packages, from subvertical, northerly striking fault panels with low internal strain in the south to isoclinally folded, gently northeast-dipping (nearly recumbent), highly attenuated panels in the north. The change is concomitant with the transition to the highest metamorphic grade in the north, as previously described by, among others, Bailes (1980a) and Zwanzig and Schledewitz (1992). The trend of units and the shallow dips are clearly a late structural feature (Zwanzig, 1999). In general, a gneissic volcanoplutonic infrastructure in the north is overlain by recumbent folds of para-autochthonous volcanic and partly allochthonous sedimentary units. In the lower part of the northern suprastructure, the metasedimentary rocks are interleaved with the volcanic rocks of the infrastructure. This already complex geometry is further complicated by refolding into dome-and-basin structures that are locally transposed into large sheath folds in the north. Late block faulting and tilting in the southern part of the Flin Flon Domain is thought to have been responsible for preserving the high-level rocks with their abundant mineral deposits.

Geochemistry, concentrated on elements that are least mobile during alteration and metamorphism, has been used to deduce the paleotectonic environments during the 1.89–1.87 Ga evolution of the Kississing–File lakes area. These deductions are supported by systematic changes in Nd isotope ratios. The geochemistry is consistent with the depositional environments suggested by the remnant primary features of the various assemblages, and leads to a new possible tectonic model. This model involved early (pre–1885 Ma) tectonic amalgamation of the volcanic rocks in the Flin Flon Domain with final extensive arc rifting and domain-wide re-amalgamation during the onset of terminal continental collision in the THO. Each type of tectonic environment is associated with a different suite of mineral deposits. Volcanogenic massive sulphide (VMS) deposits occur in oceanic-arc and arc-rift volcanic rocks. Some precious-metal deposits may occur in mid-ocean-ridge basalt (MORB)–like units that are concluded to have formed in back-arc and arc-rift environments in mafic sills (Josland Lake gabbro) that may have been the direct source of small gold deposits and platinum-group elements (Zwanzig, 1994; Peck et al., 2000; Bailes and Theyer, 2006). Another gold deposit in the Kississing–File lakes area (Puffy Lake) is synorogenic and associated with a major shear zone, like deposits elsewhere in the THO.

Other contributions of this report, which result from the comprehensive treatment of a large critical area, include an updated, most probable tectonic history of the western THO, new tectonic domain boundaries and a proposed subdivision of north-central Manitoba into tectonic subdomains (Figure 3). The tectonic history supports early (ca. 1890–1880 Ma) multistage accretion, deformation and rifting, followed by 1876–1845 Ma early successor-arc plutonism with newly confirmed coeval volcanism (Whalen and Hunt, 1994; N. Rayner and D. Tinkham, pers. comm., 2009). This was followed by ca. 1845–1830 Ma reassembly of the oldest volcanic-arc and back-arc crustal fragments during the youngest successor-arc magmatism, as suggested by Zwanzig et al. (2001). This process of arc rifting and reassembly may be the key to understanding the tectonics and metallogeny of the Flin Flon and Kisseynew domains. Arc rifting may have had a direct influence on VMS deposition, whereas the reassembly influenced orogenic gold deposition.

The work also leads to a new plate-tectonic model of the western part of the THO, consistent with the history of the wider THO extending to northern Quebec and Labrador (Figure 1) and beyond into Greenland and Scandinavia (St-Onge et al., 2006, 2009). The model suggests that the Kisseynew Domain formed early as a Japan-type back-arc sea that was filled with arc-derived sediments only during final mature-arc magmatism and terminal continental collision. This explains the newly found peri-Superior Archean crustal fragments in the Kisseynew Domain (Percival et al., 2005; Murphy and Zwanzig, work in progress, 2010) and the Archean fragments that remained attached to the Flin Flon Domain and had an early isotopic influence there. A paleogeographic reconstruction based on this model suggests that greenstone belts surrounding the Kisseynew Domain formed in a single volcanic arc that was locally rotated during terminal collision. Other belts in this assembly may be as highly prospective as the Flin Flon Domain and still contain undiscovered mineral deposits.

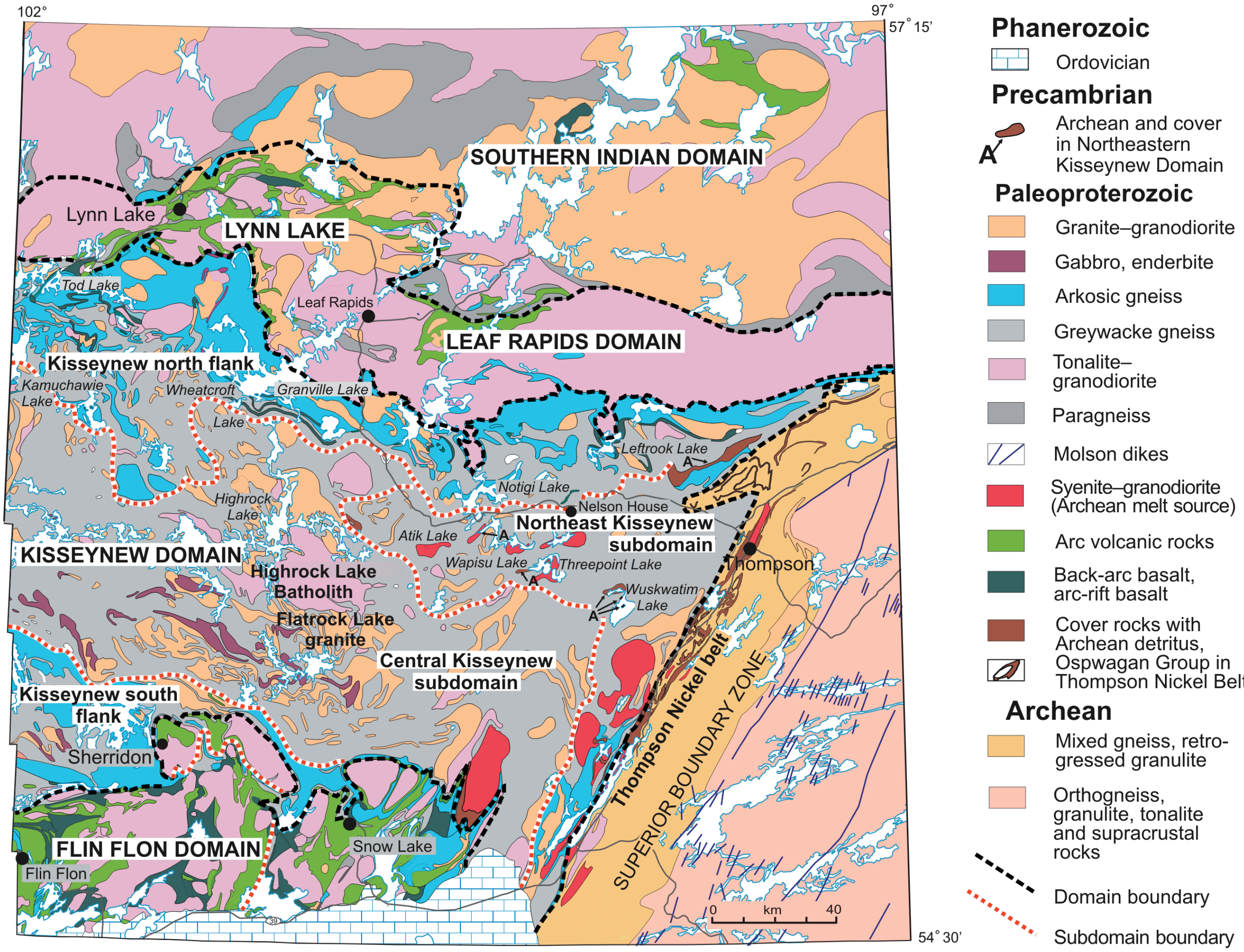


***Figure 3:** Geology of north-central Manitoba, showing the Kisseynew Domain with its subdomains and surrounding tectonic elements of the Trans-Hudson Orogen.*

1 Introduction

1.1 Report structure and terminology

The geology of the Kississing–File lakes area is presented in seven sections: 1) Introduction, 2) Regional geology, 3) Tectonostratigraphy, 4) Structure and metamorphism, 5) Geochemistry and chemostratigraphy, 6) Tectonic evolution, and 7) Economic geology. These sections are presented together in this report to allow a more integrated interpretation of the premetamorphic protoliths, as well as the tectonic history and metallogeny of the area. Important details in each part are cross-referenced in the following parts so that an individual topic can be read alone and further explored at will. The geochemical characteristics of all units are summarized in tables and illustrated in a tectonostratigraphic chart, as well as on the accompanying 1:100 000 scale geological map (Map GR2010-1-1, print format in back pocket, PDF on accompanying DVD; also in GIS format as Map GR2010-1-2 on accompanying DVD[1]). The geochemistry of the map units is illustrated in figures presented in Appendices 1–3, as adjacent diagrams of various element contents and element ratios for easy comparison. The units are listed and illustrated with digital photographs in Appendix 4[2], Table 1 on the accompanying DVD.

Section 2 includes an overview of the regional tectonic setting of the Flin Flon Domain and the adjacent Kisseynew Domain in the internal zone (Reindeer Zone) of the Trans-Hudson Orogen (THO). This section also summarizes the crustal architecture of the THO, as determined from published structural studies and seismic profiles. A short review of the geology of the Flin Flon and Kisseynew domains is included, primarily as determined during a NATMAP compilation program (NATMAP Shield Margin Project Working Group, 1998) and during mapping carried out in the Kisseynew Domain (Bailes, 1975, 1980a; Baldwin et al., 1979; Zwanzig and Schledewitz, 1992) and for this report.

In **Section 3**, the tectonostratigraphic and intrusive assemblages in the Kississing–File lakes area are defined by their tectonic affinity and introduced in broad chronological groups (oldest to youngest where known). The informal units and assemblages are listed and illustrated in Appendix 4, Table 1

[1] Map GR2010-1-2, the GS version of Map GR2010-1-1, also shows the locations of mineral deposits and geochemical samples including sample analyses.
[2] The complete geochemical dataset, unit averages and important ratios, and UTMs are also given in Appendix 4.

and shown on Maps GR2010-1-1 and -2). Where uncertainties exist in the age and affinity of units, they have been grouped with units of known age and similar geochemistry. Thin units were not all mapped but are presented in generalized tectonostratigraphic columns. The abundant intrusive rocks and their derived units of orthogneiss are summarized from existing reports.

In **Section 4**, the current understanding of large-scale structures that led to the repetition of major units is reviewed and updated. Structural evolution, which extended from 1886 to 1780 Ma, is divided into six major phases of deformation. The resulting complex geometry is described so that units with high mineral potential to be traced for exploration. The deformation phases provide an important relative time frame for the tectonic evolution of the THO in Manitoba.

Section 5, a major focus of this report, presents the geochemistry of the rocks and describes individual analyzed volcanic units in more detail. Metavolcanic rocks are divided into tectonic-assemblage types with arc or back-arc affinity, based on field data and geochemistry. Arc assemblages are further subdivided into those with strong and weak arc-like geochemical signatures. For some units of amphibolite and felsic gneiss, the geochemistry is the only unique distinguishing feature. An attempt is made to provide a subdivision for the intrusive rocks that succeed the early volcanic rocks into early and final 'successor-arc' stages of magmatism and subsequent synkinematic magmatism that followed the onset of collision tectonics.

In **Section 6,** the evolution of the Flin Flon–Kisseynew transition area is summarized in tectonic stages. The new tectonic analysis and new detrital zircon ages appear to lead to a fundamental rethinking of the evolution of the Flin Flon and Kisseynew domains.

In **Section 7**, descriptions of mineral deposits and large showings in the area are summarized from available published files and reports. The origin of some deposits is reinterpreted, and relationships are made to 1) the stages of tectonic evolution of the area, 2) local metallogeny, and 3) regional tectonostratigraphic controls of volcanogenic-massive-sulphide (VMS) and precious-metal deposits.

The prefix 'meta' is generally omitted for recognizable rock types, but the terminology of layered gneissic and migmatitic rocks is usually retained. Volcanic successions with similar ages and tectonic affinities that are structurally juxtaposed or have unknown contact relationships are termed 'assemblages' (Table 2). Assemblages generally consist of a number of mappable units at the scale of a formation that are not formally defined. Units may be in stratigraphic or fault contact. Nevertheless, units are named after their type locality and their dominant rock type. Although units are at the formation level, this stratigraphic term is avoided because of the generally uncertain contact relationships. Unit numbers and letters are assigned as on the geological legend, which uses a combination of age and geochemical affinity (Appendix 4, Table 1; Maps GR2010-1-1 and -2). Groups of assemblages are described from oldest to youngest, but individual units are defined by composition and tectonic affinity where their geochemistry is known. The order of units is stratigraphic only where depositional contacts and tops are known.

The definition of volcanic units and assemblages at the highest metamorphic grade relies partly on the signatures of their least mobile elements (Section 5.4). The major successions of sedimentary rocks and their derived migmatite are given formal group names because the base of the Missi Group is well defined and the top of the Burntwood Group is locally defined. Assemblages of units occur in fault-bounded structural blocks in the southern or main part of the Flin Flon Domain (Bailes and Syme, 1989) but are typically highly deformed and metamorphosed adjacent to the Kisseynew Domain in the north (Zwanzig and Schledewitz, 1992). For example, repetitions across thrust faults may be folded into large recumbent structures and, finally, refolded into basins and domes or large incipient sheath folds (Zwanzig, 1999). This complex geometry is resolved partly by using the lower part of the Missi Group, which retained a clearly defined base during the deformation. An understanding of the complex geometry is required to define and trace the lithotectonic assemblages across the Kississing–File lakes area. Successions that make up structural entities of higher order are called 'packages'. These may contain autochthonous sequences as well as para-autochthonous or allochthonous units or assemblages. A partial reconstruction of the main rock packages was undertaken in order to restore by retrodeformation the tectonostratigraphy of original fault blocks.

1.2 History of study and data acquisition

This study is based on the ongoing program of bedrock mapping, sampling, and stratigraphic, petrological and structural analysis done by the Manitoba Geological Survey. The Kississing–File lakes area was covered in four projects carried out intermittently between 1970 and 1996. Geochemical data were acquired from Manitoba Mines Branch Geochemical Laboratory, University of Saskatchewan and Activation Laboratories Ltd. between 1993 and 1997. Neodymium-samarium isotope data were acquired from the University of Alberta between 2001 and 2008. The latest acquired geochemical data, an isotopic age and structural inferences, were provided by Halo Resources Ltd. and their consultants during exploration in the Sherridon area between 2006 and 2008.

Table 2: Tectonostratigraphic assemblage types in the Kississing–File lakes area.

Assemblage type	Components	Age (Ga)
Early volcanic	Oceanic-arc, back-arc, ocean-floor (ophiolite) supracrustal and intrusive rocks	1.92–1.88
Early successor-arc	Volcanic-arc rocks, supracrustal and intrusive	1.88–1.85
Late successor-arc and basins	Emerged (continental)–arc volcanic and intrusive rocks, successor-basin deposits (fluvial-alluvial and marine)	1.85–1.83
Postcollisional	Intrusive rocks	1.82–1.77

The earliest of these projects was 1:15 840 scale mapping of the File Lake area between 1970 and 1972 (Bailes, 1980a). This work provided the first stratigraphic framework and an early structural analysis for the area that included structural cross-sections. Metamorphic isograds were mapped between rocks at upper-greenschist facies in the southern part of the area and upper-amphibolite facies and migmatite in the north. This provided an important understanding of the connection between the relatively well preserved rocks in the south and the equivalent higher grade metamorphic rocks in the north. Archived pulps remaining from early geochemical analyses were reanalyzed for major and trace elements during this study.

Regional 1:50 000 scale mapping was also carried out during this early period in the central and northern parts of the Kisseynew Domain, adjoining the Kississing–File lakes area (Baldwin et. al., 1979; Zwanzig, 1990). Archived samples from this project were provided to the University of Saskatchewan for a wide-ranging study of the geochemistry of granitoid rocks of the Kisseynew Domain in Manitoba and Saskatchewan (Hollings and Ansdell, 2002). A small selection of their data is used in this report for comparison with the intrusive rocks on the South flank of the Kisseynew Domain, including part of the Kississing–File lakes area.

Geological mapping at 1:50 000 scale in the Kississing–Batty lakes area and 1:20 000 scale at Kisseynew Lake was carried out between 1983 and 1988 (Zwanzig and Schledewitz, 1992). This work has provided an early tectonostratigraphic and structural framework for the gneissic rocks, together with an overview of their petrography and mineral deposits. An important result of this work was the separation of gneissic units derived from volcanic, sedimentary and intrusive rocks. Metavolcanic assemblages and intrusive units similar to those in the central part of the Flin Flon Domain were recognized. Basinal and basin-margin facies of the metasedimentary rocks were defined. Some of the complex structures of the Kisseynew Domain South flank were described. Refolded recumbent structures involving the high-grade metasedimentary rocks of the Kisseynew Domain that are interleaved with the meta-igneous rocks of the Flin Flon Domain were recognized. Geochemical data acquired during the 1:50 000 mapping formed a limited set of widely spaced samples that remained unpublished but was helpful in establishing the structural continuity of some units.

The excellent exposure between Kississing and Dow lakes resulting from the forest fire that swept through the area in 1989 permitted more detailed remapping, at 1:20 000 and 1:10 000 scales, of areas with economic interest or critical geological relationships (Zwanzig, 1992a, b; Schledewitz, 1993a–c; Zwanzig and Parent, 1993; Zwanzig et al., 1993; Zwanzig and Shwetz, 1994a; Zwanzig, 1996a).[3] The detailed preliminary maps, along with the earlier 1:50 000 scale maps (Zwanzig and Schledewitz, 1992) and maps from the Flin Flon–Snow Lake area and parts of Saskatchewan, were revised and compiled at 1:100 000 scale (NATMAP Shield Margin Project Working Group, 1998). Major early thrust faults at File Lake were recognized by Connors (1996) and shown in the compilation map, but reinterpreted by Zwanzig (1999). The tectonic origins of the main volcanic assemblages in the File Lake–Reed Lake area were recognized after reconnaissance-scale remapping, resampling and geochemical work there (Syme et al., 1995; Syme and Bailes, 1996). A tectonostratigraphic and structural analysis of the Kississing–File lakes area was presented by Zwanzig (1999). That work provided the structural and tectonostratigraphic framework for this report.

The detailed mapping along the northern margin of the Flin Flon Domain, plus the resampling and recovery of archived samples in the File Lake area, provided most of the material for the geochemistry of this report. The most intensive geochemical sampling was carried out on the meta-igneous units from Kisseynew Lake to Dow Lake and at Jungle Lake. Early versions of the compilation maps were used to guide the sampling. This allowed collection of samples from known tectonostratigraphic positions. The results of more widely scattered sampling along strike or across the typically complex folds in the area were used to help delineate the full extent of units. Recent mineral exploration and mapping by Halo Resources Ltd. and work by D. Tinkham (Laurentian University) have provided geochemical, structural and metamorphic data on the Sherridon–Meat Lake area.

2 Regional geology

2.1 Tectonic setting

The Flin Flon and Kisseynew domains form the southern half of the predominantly isotopically juvenile Paleoproterozoic internides (Reindeer Zone) of the Trans-Hudson Orogen (THO) in Manitoba and Saskatchewan (Ashton, 1999). They are, respectively, a metamorphosed oceanic volcanoplutonic domain and a migmatitic sedimentary-basin domain. Other important tectonic elements of the THO in Manitoba are shown in Figure 1 and defined in Figure 4. These include the Stephens Lake metasedimentary domain in the east and, successively northward, the Leaf Rapids volcanoplutonic domain (including the Lynn Lake and Rusty Lake greenstone belts), the Southern Indian metasedimentary and magmatic domain and the Chipewyan continental-arc–back-arc granitoid domain. The two latter domains are progressively more contaminated with Archean crust to the northwest (Beaumont-Smith and Böhm, 2002; Corrigan and Rayner, 2002), in part from the adjoining external zone (Cree Lake Zone; Lewry et al., 1986) that forms the margin of the Archean Rae-Hearne craton. The various domains represent crustal slices, remnants of composite terranes that formed in different tectonic environments, some in neighbouring terranes and others at an unknown distance. The internal domains are interpreted to have formed during plate-tectonic activity between 1.9 and 1.83 Ga (Lucas et al., 1996). They were amalgamated and built up in successive stages in an oceanic to pericontinental environment, as well as at the active continental margin of the Rae-Hearne craton. Final continent-continent collision, possibly lasting from 1.85 to 1.82 Ga, involved the external zones trapping the juvenile

[3] Available in original black-line format from Publication Sales, Manitoba Innovation, Energy and Mines, 360–1395 Ellice Avenue, Winnipeg MB R3G 3P2, Telephone 1-800-223-5215, 204-945-4154, Fax 204-945-8427, minesinfo@gov.mb.ca.

Orogen	Craton	Zone	Domain	Definition
Kenoran	Craton	**Cree Lake**		Archean rocks, reworked during the Hudsonian Orogeny with Paleoproterozoic supracrustal and/or intrusive rocks
			Nejanilini	Archean granulite, Archean and Paleoproterozoic orthogneiss, metasedimentary rocks and migmatite
Increased Archean supracrustal components	Rae-Hearne	? ? (Northwestern Trans-Hudson externides)	Wollaston	Gneissic plutonic rocks with Archean melt component or crystalization age; Paleoproterozoic foreland to retro arc cover; strong north-northeast late Hudsonian overprint
			Seal River –Great Island	Archean basement gneiss and greenstones with Archean and Paleoproterozoic metasedimentary cover sequences; east-west structural grain with north and northeast overprint
		Reindeer (Trans-Hudson internides)		Paleoproterozoic supracrustal and intrusive rocks, juvenile to highly contaminated with Archean crust; structurally emplaced partly over Archean Sask craton; rare Archean inliers
		Wathaman-Chipewyan subzone	Chipewyan	Mainly 1865-1850 Ma granodiorite (continental-arc and back-arc), juvenile to Archean-contaminated; plutons stitch Southern Indian and Seal River–Great Island domains
	?	Rottenstone–Southern Indian subzone	Southern Indian	Metasedimentary (greywacke, conglomerates) and arc–back-arc magmatic rocks, juvenile in the south, Archean crustal component in the north; local inlier of Archean gneiss
	?		Waskaiowaka	Paleoproterozoic orthogneiss with significant Archean melt component; metasedimentary rocks and migmatite
		La Ronge–Lynn Lake subzone	Leaf Rapids	**Eastern subdomain:** mainly juvenile-arc plutons **Lynn Lake–Rusty Lake belts:** Juvenile and locally Archean-contaminated Paleoproterozoic arc and back-arc assemblages intruded by coeval and younger arc plutons; local Sickle Group
Trans-Hudson	Sask	Flin Flon– Glennie Complex	Flin Flon	Mainly juvenile Paleoproterozoic arc and back-arc assemblages intruded by coeval and younger arc plutons. **Central Flin Flon subdomain:** Amisk collage with local Missi Group outliers and rare Archean fault slices. **Snow Lake subdomain:** Amisk- and Missi-age volcanic rocks with Burntwood Group **Batty Lake subdomain**: early successor-arc igneous rocks
			Kisseynew	Mainly juvenile arc–derived metasedimentary rocks–migmatite. **Central subdomain:** Burntwood Group, younger **Flanking subdomains**: Burntwood and Missi/Sickle sediments, local volcanics and arc–back-arc basement. **Northeastern subdomain:** Burntwood Group with Archean inliers having Paleoroterozoic cover and pre-Burntwood intrusions
Increased Archean supracrustal components			Stephens Lake	Predominantly metasedimentary rocks (greywacke, iron formation, arenite), mainly juvenile arc–derived
		Superior Boundary (Southeastern Trans–Hudson externides)		Archean rocks, variably reworked during the Trans-Hudson Orogeny ± Paleoproterozoic supracrustal/intrusive rocks
			Fox River	Neoarchean basement, slightly reworked; overlain by weakly metamorphosed/deformed Paleoproterozoic rocks
			Thompson Nickel Belt	Archean basement, strongly reworked; unconformably overlain by highly deformed and metamorphosed Paleoproterozoic platform to continental-margin succession; intruded by Paleoproterozoic back-arc and collisional granite
			Winnipegosis	Sub-Phanerozoic Archean basement, reworked and overlain by weakly metamorphosed/deformed Paleoproterozoic rocks
All Archean			Paint Lake	Neoarchean rocks, weak to strong Trans-Hudson reworking, include retrogressed granulite; Paleoroterozoic granite
Kenoran	Superior		Assean L.	Mainly Mesoarchean rocks, moderate reworking
			Split Lake	Neoarchean rocks, weak Trans-Hudson reworking
		Northwest Superior		Neoarchean orthogneiss, volcanoplutonic belts
			Pikwitonei	Granulite

Archean-contaminated plutons

***Figure 4:** Preliminary subdivision of the Trans-Hudson Orogen and the adjoining (Archean) cratons in Manitoba into lithotectonic zones and domains (definitions and relationships under review). The relative importance of Archean components in sedimentary, volcanic and intrusive rocks is illustrated in the left column. Location of domains is shown on Figure 1.*

domains, the small Archean Sask craton and various Archean crustal slivers in front of the advancing Superior craton. Thus, the THO is a collage of Paleoproterozoic and Archean domains. The Flin Flon, Glennie, La Ronge, Lynn Lake–Leaf Rapids and Kisseynew domains are the largest remnants of oceanic-arc, back-arc, pericontinental and probable fore-arc terranes in the THO.

The larger scale subzone that encompasses the Flin Flon Domain and the Glennie Domain (Figure 1) has been called the Flin Flon–Glennie Complex (NATMAP Shield Margin Project Working Group, 1998). It is interpreted to be allochthonous on Archean crustal rocks (Sask craton). The Sask craton is exposed in several structural culminations surrounded by the younger rocks of the Flin Flon–Glennie Complex and Kisseynew Domain in Saskatchewan (Lewry et al., 1986). The Pelican window forms the core of the largest of these structures. It is mantled by a wide mylonite zone with dikes that indicate a tectonic emplacement of the Flin Flon–Glennie Complex starting before 1826 Ma (Ashton et al., 1999, 2005). Seismic profiles (Lewry et al., 1994) suggest that the Sask craton structurally underlies much of the southwestern part of the Reindeer Zone (Figure 1). However, its full extent in the lower crust is uncertain and the eastern limit may lie farther west than shown in Figure 1 (White et al., 2005).

The Cree Lake Zone in Saskatchewan and northwestern Manitoba has an Archean basement and Paleoproterozoic supracrustal rocks deposited during and following continental rifting. Proterozoic granitoid plutons, some with significant Archean components, are common. The cover in the Wollaston Domain contains a mixture of Archean and Paleoproterozoic detritus and has been considered a foreland-basin to retro-arc succession (Tran et al., 2003; Corrigan et al., 2005).

Metasedimentary rocks in the northern half of the Reindeer Zone also contain a mixture of Archean and isotopically juvenile Proterozoic detritus, and are intruded by melts that were contaminated by Archean crust. This has been ascribed to early collision between parts of the Reindeer Zone and the Rae-Hearne craton margin, followed by northwest-directed subduction of oceanic tracts (Corrigan et al., 2005). The cover was thrust first to the northwest (e.g., Annesley et al., 2005) and then, together with its basement and the northwestern part of the Reindeer Zone, it was carried southeast toward the Kisseynew Domain (Corrigan et al., 2005) to form the uppermost crustal slices in the orogen. The final stage of overriding from the northwest may have lasted from the 1845 Ma deposition of the Missiage siliciclastic rocks (McLennan, Sickle, Grass River, Missi and Ouram groups) to peak metamorphism at 1810–1800 Ma (Machado et al., 1999; Zwanzig, 1999). A final northeast compression and sinistral transpression extended across the entire orogen.

In the southeast, the Reindeer Zone forms an apparently younger collisional boundary (Green et al., 1985; Weber, 1990) with its southeastern external zone (Superior Boundary Zone). Along this boundary and in the northeastern part of the Kisseynew Domain, the predominantly juvenile Proterozoic metasedimentary rocks and migmatite are underlain by, or structurally interleaved with, Archean basement gneiss that has a thin cover of Archean-derived metasedimentary rocks (Percival et al., 2006; Zwanzig et al., 2006). These are intruded by Paleoproterozoic rocks with melt components of Archean crust and mantle. A quartzite unit with Archean-derived detritus occurs on the northeastern end of the Flin Flon Domain.

Partly Archean-derived melts are also associated with unexposed exotic sources in the northern and eastern Reindeer Zone, indicating that some areas were pericontinental. Despite structurally underlying Archean rocks (Sask craton), the exposed central part of the Reindeer Zone is predominantly juvenile at surface. The rocks in the Kississing–File lakes area are minimally contaminated, consistent with the inferred structural superposition of the Paleoproterozoic rocks on the Sask craton after the arc magmatism (Figure 4). The early oceanic origin of volcanic rocks is associated with their metal-rich character, and this is addressed in the report. A zone of seismically anomalous mantle underlies the Kississing–File lakes area and extends southwest into Saskatchewan (Németh et al., 2005), where it flanks the Fort à La Corne diamond-bearing kimberlite field (Figure 1). This is interpreted as a late feature.

Along the para-autochthonous western margin of the Superior craton (i.e., in the Thompson Nickel Belt [TNB]), the cover rocks belong to the Ospwagan Group (Bleeker, 1990; Macek et al., 2006). These early Paleoproterozoic sedimentary rocks have an almost purely Archean provenance and were deposited on a continental shelf (Zwanzig, 2005a; Böhm et al., 2007; Zwanzig et al., 2007). They are capped by juvenile Paleoproterozoic mafic–ultramafic volcanic rocks and host ca. 1880 Ma ultramafic bodies (Hulbert et al., 2005) and abundant associated nickel deposits. The ages and relationships of the various mafic–ultramafic rocks along the Superior Boundary Zone in Manitoba indicate repeated episodes of crustal extension with abundant magmatism, apparently after the main rifting (Zwanzig, 2005a). The supracrustal rocks are intruded by mafic dikes that are interpreted as belonging to the prominent 1883 Ma Molson dike swarm (Heaman et al., 1986). The western edge of the TNB is intruded also by ca. 1890–1870 Ma calcalkaline plutons with large Archean melt components that suggest an origin in a possible continental-margin arc (Percival et al., 2006) or back-arc (Zwanzig et al., 2003). The inlying Archean gneiss and thin cover in the Northeast Kisseynew subdomain are possibly part of the Ospwagan Group and its basement (Percival et al., 2006; Zwanzig et al., 2006). They are intruded by ca. 1885 Ma, transitional to alkaline and calcalkaline rocks derived mainly from Archean subcontinental lithospheric mantle (Whalen et al., 2008).

The TNB had an extended history of deformation and metamorphism that lasted as late as 1770 Ma (Bleeker, 1990; Burnham et al., 2009). During terminal continental collision, the Superior Boundary Zone rocks were inserted laterally into the Kisseynew Domain, locally overriding it (White et al., 2002). A group of ca. 1835 Ma, juvenile, alkaline to sanukitoid plutons that extends across the boundary between the TNB and the Kisseynew Domain (Zwanzig et al., 2003) implies an earlier collision between these terranes than previously thought (e.g., Lucas et al., 1999). It also implies local underthrusting of Proterozoic juvenile-arc mantle.

The cover in the Fox River Domain (Scoates, 1981, 1990), along the northern margin of the Superior craton, probably also comprises Archean detritus but contains juvenile volcanic rocks and mafic–ultramafic sills that are the same age as the ca. 1883 Ma Molson dikes and the sills in the TNB (Heaman et al., 1986). Similar undated sedimentary rocks in the Winnipegosis Domain, beneath the Phanerozoic rocks, contain younger (ca. 1864 Ma) mafic–ultramafic rocks (Hulbert et al., 2005; Burnham at al., 2009). The cover rocks in these domains are nearly undeformed and are at the sub-greenschist facies of metamorphism.

Finally, the southerly overriding of the sub-greenschist–facies rocks of the Fox River Domain, on the northern margin of the Superior craton, by the amphibolite-facies siliciclastic rocks of the Stephens Lake Domain must have occurred after the waning of metamorphism in the north. A sharp break in metamorphic grade and no inverted isograds at this north-dipping contact support this interpretation. The orthogneiss and migmatite of the Waskaiowaka Domain probably overthrust the Fox River Domain at a similar time. These events may have occurred during an important tectonometamorphic phase in the TNB at about 1780 Ma. Late- to postmetamorphic north-north-east-trending mylonite zones with sinistral east-side-up displacement occur along the western margin of the TNB (Fueten and Robin, 1989; Bleeker, 1990; Zwanzig, 2000a). These may have been active during the thrusting in the north.

2.2 Geology of the Flin Flon and Kisseynew domains

The well-preserved, low-grade metamorphic rocks in the Flin Flon Domain have been recognized as being fault-bounded assemblages of volcanic and lesser sedimentary rocks formed in an oceanic island-arc environment (Bailes and Syme, 1989; Bailes and Galley, 1999). Geochemical work, U-Pb zircon dating and Sm-Nd isotope work have shown that they comprise mainly 1.90–1.88 Ga juvenile assemblages (Manitoba Geological Survey, 2006) formed in a variety of oceanic environments. These include arcs, back arcs, arc rifts, an ocean plateau and an ocean island (Stern et al., 1995a, b; Syme et al., 1999). Parts of these terranes were structurally juxtaposed to form an accretionary collage, the Amisk Collage, in the central Flin Flon area (Lucas et al., 1996), and units of similar age in the eastern part of the Flin Flon Domain (Snow Lake allochthon of Kraus and Williams, 1999) that comprise the Northeast Reed and Snow Lake assemblages (Syme et al., 1995). All of these assemblages appear to be present also in the Kississing–File lakes area but with the addition, on the northern margin of the Flin Flon Domain, of ca. 1.85 Ga metavolcanic rocks. A relatively young volcanic suite is preserved only on the eastern shore of Wekusko Lake at the eastern margin of the Flin Flon Domain, and as a narrow sliver southeast of Flin Flon.

Work in the central Flin Flon Domain had suggested that accretion took place during a pause in magmatism, shortly after 1.88 Ga (Lucas et al., 1996). However, differentiated mafic sills (Josland Lake gabbro) intruding the axial surfaces of major folds southwest of File Lake (Bailes, 1980a) indicate that early deformation occurred before ca. 1886 Ma, the crystallization age of the sills (Zwanzig et al., 2001). The intrusive relationship also indicates that these structures were recumbent during differentiation of the sills. Deformation fabrics in adjacent areas also suggest the presence of early, shallow-dipping structures (Ryan and Williams, 1999). These structures are consistent with the amalgamation and stacking of some arc and back-arc fragments between early volcanic episodes, as well as during and following successor-basin development.

The early volcanic assemblages are cut by 1876–1830 Ma 'successor-arc' intrusions (Lucas et al., 1996; Whalen et al., 1999) that constitute over 50% of the outcrops in the Flin Flon Domain. They belong to early, middle and late arc plutons formed during successive stages of continued subduction, involving the destruction of early back-arc basins. In this report, the plutons have been combined as 1876–1845 Ma 'early' ('early' and 'middle' of Whalen et al., 1999) and 1845–1830 Ma 'late' successor-arc plutons. Early and late successor-arc volcanic rocks are restricted to the northern and eastern margins of the Flin Flon Domain. The progressive development of steeply dipping, northeast- to northwest-trending foliation in successively older plutons suggests that the main upright structures in the Flin Flon Domain formed during the successor-arc magmatism (Ryan and Williams, 1999). Current structural work also shows that there were two phases of northerly trending folding and faulting in the Flin Flon area before the deposition of the Missi Group (R-L. Simard et al., work in progress, 2010). The late successor-arc plutons, on the other hand, probably formed during the onset of continental-collision tectonics (e.g., Zwanzig, 1999).

The Amisk Collage and early (1876–1845 Ma) successor-arc plutons are unconformably overlain by 1848–1830 Ma (Machado et al., 1999; Ansdell et al., 1999) terrestrial metasedimentary and local volcanic rocks (Missi Group; e.g., Gordon and Lemkow, 1987; Stauffer, 1990). The Snow Lake and Northeast Reed assemblages, on the other hand, are generally in structural contact with the greywacke-mudstone of the Burntwood Group (Zwanzig, 1990), originally the File Lake Formation of Bailes (1980a). These metasedimentary rocks are coeval with to slightly older than (about 1855–1840 Ma; David et al., 1996; Machado et al., 1999) the Missi Group. The Burntwood Group was deposited in turbidite fans extending into the Kisseynew basin (Bailes, 1980b) and was faulted against the volcanic rocks of the Snow Lake area. The terrestrial shallow-water sandstone and marine turbidite were probably deposited during the onset of collision tectonics, during sedimentary filling and closure of the Kisseynew basin (Zwanzig, 1999).

Rapid erosion of the volcanic and plutonic rocks supplied the immature, mainly juvenile detritus that constitutes the Missi and Burntwood groups (Bailes, 1980b; Zwanzig, 1990; Syme et al., 1995). In the main part of the Flin Flon Domain, Missi Group conglomerate, pebbly sandstone and sandstone–siltstone probably formed in isolated intermontane basins. Along the transition to the Kisseynew Domain, the finer grained siliciclastic rocks of the Missi Group likely formed in a long basin between exposed volcanoplutonic rocks and a bordering sea. The favoured interpretation (Zwanzig, 1999) is that the Burntwood Group was fed by detritus provided by the same fluvial-alluvial systems that formed the Missi Group. As the Missi Group prograded into the Kisseynew basin, it apparently formed unstable deposits on a steep slope; these were redeposited into deeper water by turbidity currents, forming greywacke-mudstone of the Burntwood Group. Although sedimentary facies changes are difficult to recognize in metamorphic rocks, analysis of Burntwood and Missi group rocks in this report reinforces reports of sedimentary facies changes by Zwanzig (1990) and Bailes (1980b). They support an interpretation of the Missi and Burntwood groups as being sourced in the Flin Flon Domain and part of a sedimentary system prograding into a basin represented by the adjacent Kisseynew Domain.

Intercalated volcanic rocks are restricted to the Missi Group along the northern and eastern margins of the Flin Flon Domain, with only two occurrences of possible felsic reworked tuff recognized in the Burntwood Group. One possible interpretation of this difference in abundance of intercalated volcanic rocks is that much of the Burntwood Group may predate the Missi-age magmatism and distal-facies Missi sedimentation. A range of depositional ages for different Missi basins has been suggested by Ansdell et al. (1999), and various tectonic origins have been proposed for the Kisseynew Domain (Ansdell et al., 1995; Zwanzig, 1997). The common occurrence of early (ca. 1.90–1.89 Ga) arc-rift basalt and back-arc–basin basalt in contact with the Burntwood Group in the Kississing–File lakes area (this report) and along the North flank of the Kisseynew Domain (Zwanzig et al., 1999) suggests that the Kisseynew Domain may have had a tectonic history as long as that of the Flin Flon Domain but one dominated by early back-arc volcanism and much later, rapid sedimentation during collision tectonics and final arc magmatism (ca. 1.84–1.83 Ga). Sedimentation may have taken place in a fore-arc and shallow trench or collisional-basin environment. These and other possible tectonic models are discussed later.

The Flin Flon Domain can be divided into three subdomains (Figure 2). The Central Flin Flon and Snow Lake subdomains contain rocks of similar age but with variations in chemical and isotopic characteristics, and ages of intrusions. The Batty Lake subdomain apparently contains younger metavolcanic rocks. This and differences in field relationships indicate different tectonic histories of these subdomains. The Central Flin Flon subdomain is underlain by the Amisk Collage, intruded by successor-arc plutons that are dominated by 1876–1845 Ma ages, with all of these rocks unconformably overlain by Missi Group meta-arenite and conglomerate. The Amisk Collage is dominated in the west by the Flin Flon arc assemblage, in the centre by the Elbow-Athapapuskow ocean-floor assemblage and in the east by the Fourmile Island arc assemblage (Figure 5; Syme et al., 1995).

The Snow Lake subdomain (Snow Lake allochthon of Kraus and Williams, 1999) includes the Northeast Reed (ocean-floor) assemblage and, farther east, the Snow Lake (arc) assemblage (Bailes and Galley, 1999, 2007). These are intruded by coeval (subvolcanic) granitoid plutons and late successor-arc plutons, both of which are rare in the Central Flin Flon subdomain. The Batty Lake subdomain and the area east beyond Wekusko Lake include, respectively, the Sherridon–Meat Lake and the Schist-Wekusko assemblages of volcaniclastic and volcanic rocks that have an early successor-arc age (1850–1855 Ma Sherridon gneiss, dated by N. Rayner and D. Tinkham, pers. comm., 2009; and the 1876 Ma McCaffrey-Liftover package of Ansdell et al., 1999). The early and successor-arc volcanic rocks in the latter subdomains are fault bounded along their contacts with the younger (1855–1840 Ma; David et al., 1996; Machado et al., 1999) Burntwood Group metagreywacke-mudstone. Plutons in the Snow Lake subdomain are dominated by the younger (1840–1830 Ma) successor-arc intrusions.

The West Reed Lake Fault Zone is interpreted as a major structural break that separated the subdomains, forming the eastern limit of the Amisk Collage and western limit of the Burntwood Group. However, both subdomains contain ca. 1890–1880 Ma igneous rocks and narrow fault slivers of the Burntwood-age sedimentary rocks that occur also in the Central Flin Flon subdomain. The distinctive 1886–1881 Ma layered sills (Josland and Mikanagan gabbros) in both domains may represent a rifting event that disrupted the previously accreted complex, leading to a period of separate evolution of the subdomains. Rifting probably formed at least part of the early Kisseynew back-arc–basin floor at that time (Zwanzig, et al., 2001).

The Kisseynew Domain forms a large part of the Reindeer Zone in Manitoba (Figure 3; Zwanzig, 1990, 1999; Gordon et al., 1990). The Kisseynew Domain features a central subdomain of migmatite derived from greywacke-mudstone, and abundant late granitoid rocks at transitional upper-amphibolite to granulite facies of metamorphism (Bailes and McRitchie, 1978; Growdon et al., 2006). Flanking subdomains have structurally interlayered gneissic units derived from Burntwood Group greywacke, arkosic sedimentary rocks: Missi Group in the south; and back-arc–basin basalt and gabbro, and plume-related volcanic rocks in the north (Zwanzig, 1990; Zwanzig et al., 1999). These mafic rocks and their intrusions are interpreted as basement to the Missi-age arkosic rocks (Sickle Group) on the Kisseynew North flank. Intercalations with similar ocean-floor affinity within the Burntwood Group may also represent the disrupted basement to the Burntwood Group (White et al., 2000). A northeastern subdomain has interleaved or structurally underlying Archean gneiss that has a thin metasedimentary cover with an Archean provenance (Percival et al., 2006). This formed in the pericontinental terrains in the THO that border the Superior craton.

The origin of the Kisseynew Domain has been described as a back-arc basin (Ansdell et al., 1995; Lucas et al., 1996), but a fore-arc or collisional origin cannot be ruled out, at least for the later stages of evolution (Zwanzig, 1999). In this report, we will argue that the Kisseynew Domain had a protracted history involving 1) development of early basalt in one or more back-arc basins; 2) later turbidite deposition, probably during the onset of continental collision; 3) Missi Group terrestrial sedimentation and volcanism at the basin margin; and 4) postcollisional magmatism and thermotectonism. The close similarity to Neoarchean gneissic sedimentary belts, interpreted as fore-arc and trench-fill deposits (e.g., Fralick et al., 2006), is discussed in the Section 6.1.6.1.

The boundary between the Flin Flon and Kisseynew domains is herein defined as the change from dominantly pre-Burntwood (>1.85 Ga) metavolcanic and intrusive rocks to younger, dominantly metasedimentary basinal and syntectonic intrusive rocks. The sedimentary facies in the basinal rocks change across the boundary from proximal fluvial-alluvial, mainly coarse sandy and conglomeratic facies (Missi Group; Bailes, 1971; Stauffer, 1990) to distal marine turbidite

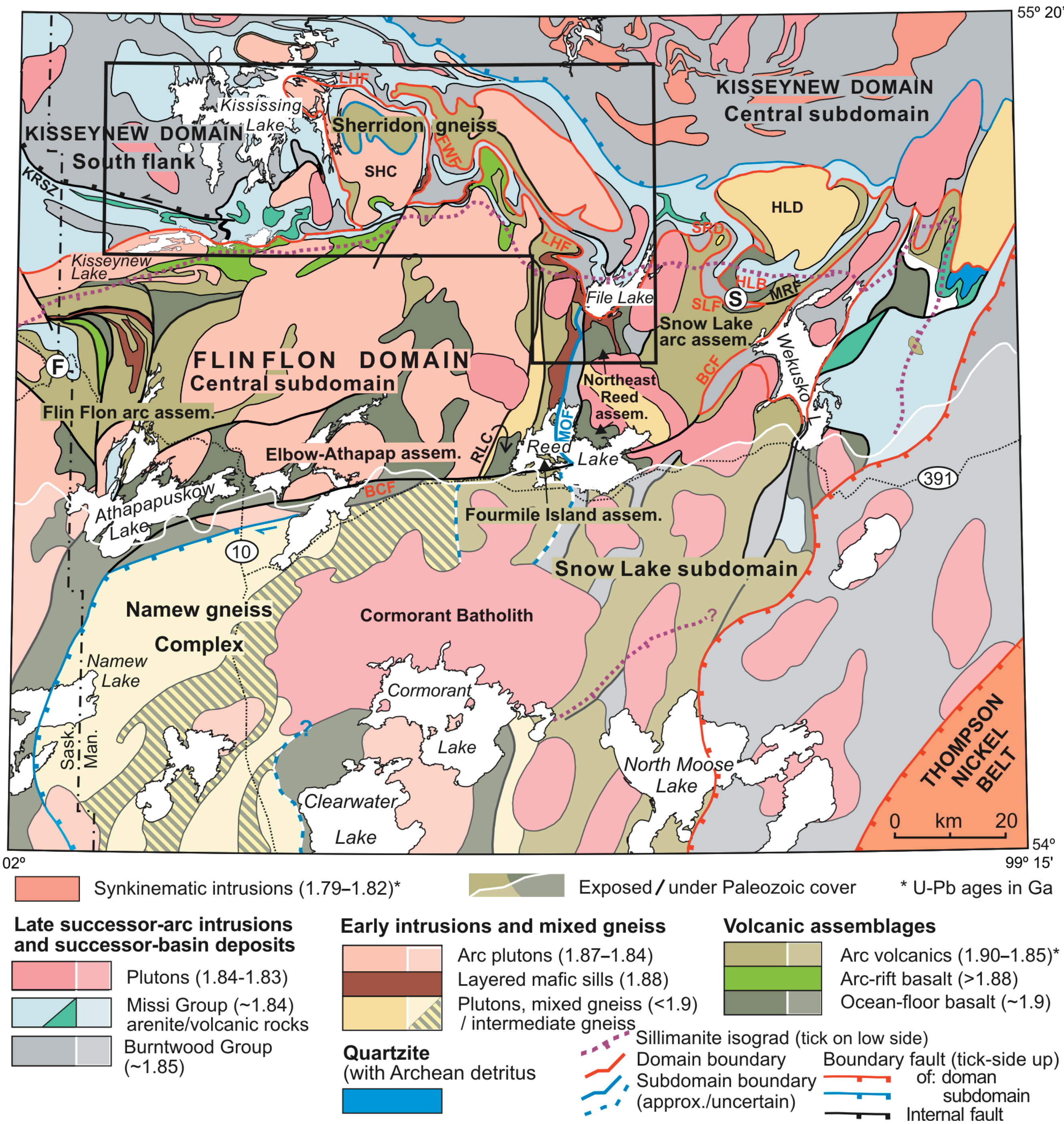


***Figure 5:** Major tectonic assemblages and sedimentary and intrusive rocks of the Flin Flon Domain, both exposed and beneath Paleozoic cover. Also shown are the structurally complex domain boundaries; the subdomains, including the southern parts of the Kisseynew Domain; and the sillimanite-biotite-garnet isograd. Abbreviations: BCF, Berry Creek Fault; F, Flin Flon; FWF, Fairwind Fault; HLB, Herblet Lake basin; HLD, Herblet Lake dome; KRSZ, Kississing River Shear Zone; LHF, Loonhead Lake Fault; MOF, Morton Lake Fault; MRF, McLeod Road Fault; RLC, Reed Lake mafic–ultramafic complex; S, Snow Lake; SHC, Sherridon-Hutchinson Lake Complex; SLF, Snow Lake Fault. Outline indicates the Kississing–File lakes area.*

(Burntwood Group, previous File Lake Formation), overlain by more distal Missi Group sandstone with only thin basal conglomerate at the transition (Bailes, 1980a, b; Zwanzig, 1990, 1999). In the most distal facies, the Burntwood Group metaturbidite grades upward through a thin unit of quartzite and pelite to nonmarine meta-arenite and subaerial volcanic rocks of the Missi Group. Thus, the boundary marks the regressive (prograding) depositional margin of the marine Kisseynew basin.

The area that encompassed the transitions in lithology and sedimentary facies was previously termed the 'Kisseynew South flank' (Zwanzig and Schledewitz, 1992). This subdomain name is retained in this report only for the area of mainly metasedimentary rocks and younger intrusive rocks in the vicinity of Duval Lake and the western part of Kississing Lake. The older igneous rocks extending from the central part of Kississing Lake to Batty Lake are herein described as part of the Flin Flon

Domain. The Missi Group in the Flin Flon Domain (as defined in this report) was deposited unconformably on the early volcanic and intrusive rocks, probably in isolated intermontane basins that drained into the Kisseynew basin. The Missi Group volcanic rocks occur along the paleobasin margin. The present irregular domain boundary was probably shaped by original embayments in the Flin Flon–Glennie Complex and by polyphase deformation.

The crustal architecture inferred from seismic-reflection profiles (Lucas et al., 1994) and parallel structural cross-sections (Zwanzig, 1993, 1999) indicates that the southeastern THO forms a gently northeast-dipping crustal stack. The interpretation of the main seismic reflection line (S1a in White et al., 2005) is consistent with east-over-west crustal stacking followed by northeast-over-southwest lower-crustal flow (Hajnal et al., 1996). The volcanoplutonic rocks of the Flin Flon Domain may be structurally underlain by exotic Archean crust (Sask craton in Figure 1; Lewry et al., 1994, Ashton et al., 1999, 2005; White et al., 2005). They are structurally overlain by the gneisses of the Central Kisseynew subdomain in the north and east. West of a large structural culmination in the Glennie Domain, the crustal structure dips northwest (Lewry et al., 1994). This crustal geometry formed between about 1850 and 1770 Ma by progressive arc and continental collision and continued convergence, followed by intracontinental transpression (Figure 6). Relatively young (about 1830–1790 Ma) intrusions are classified as synkinematic and probably formed during postcollisional convergence of the Archean cratons that bound the THO. However, arc magmatism of this late stage cannot be ruled out in the Kisseynew Domain (e.g., Hollings and Ansdell, 2002; Ansdell, 2005). The synkinematic intrusions formed during metamorphism that accompanied the development of the crustal-scale structural stack that forms much of the Reindeer Zone (Zwanzig, 1999). The Flin Flon Domain has been proposed to form a 10–15 km thick imbricate crustal slice emplaced from the northeast (Lucas et al., 1996), east (White et al., 2005) or both (Zwanzig, 1999). The more complex syn- to postcollisional structural history proposed by Zwanzig (1999) involves overturned early thrust faults rooted in the south, and subsequent repeated changes in fold vergence, first toward the west and then, during peak thermal deformation, toward the southwest and south, followed by northwest compression during uplift and slow cooling. This latter history is adopted in this report.

3 Tectonostratigraphy

The structural continuity of rocks in the Flin Flon–Kisseynew transition with lower grade, better preserved rocks to the south provides the basis for understanding the area's geology. Major volcanic assemblages are defined in the south and traced into the Kississing–File lakes area, with metamorphic grade increasing to the north. Assemblages were defined after extensive study (e.g., Bailes, 1980a; Syme, 1988, 1991; Bailes and Syme, 1989; Syme et al., 1995; NATMAP Shield Margin, Working Group, 1998; Bailes and Galley, 1999) and

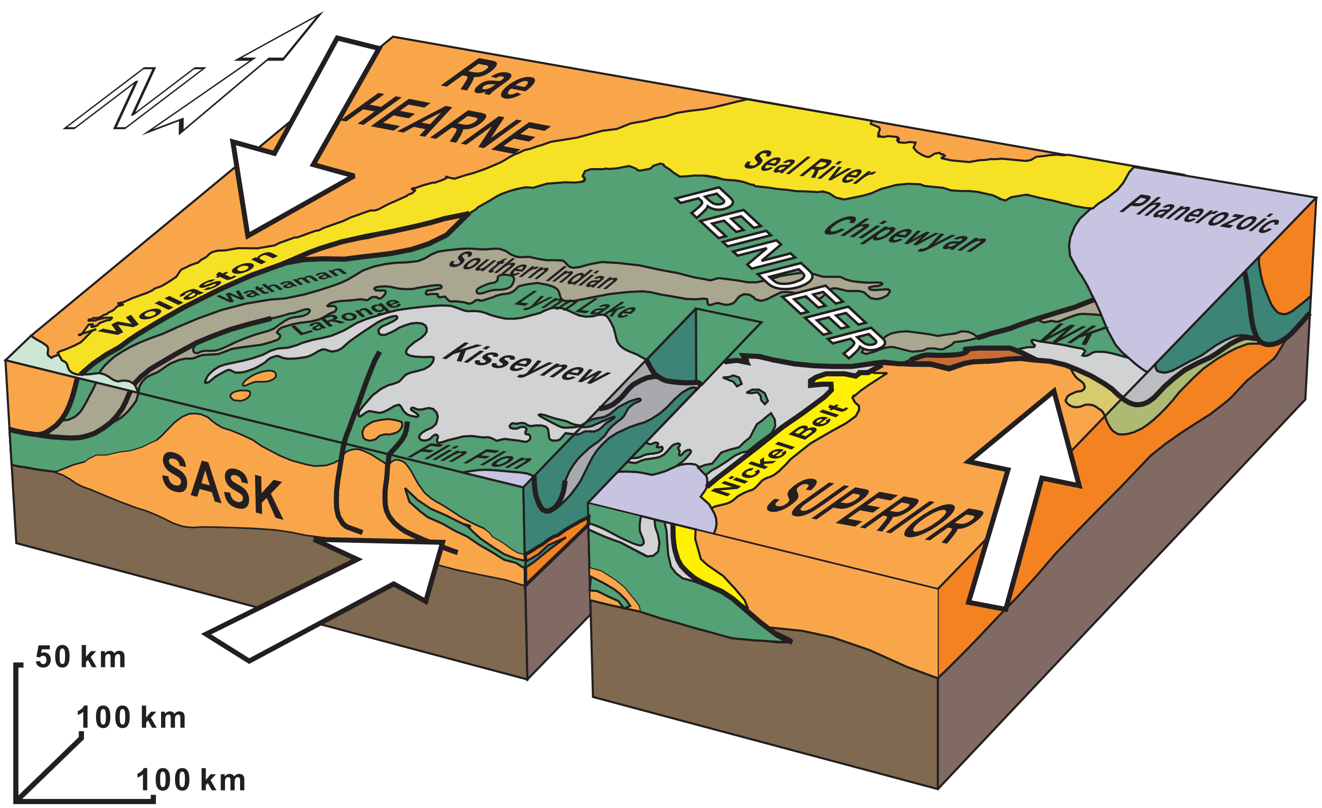


__Figure 6:__ Schematic diagram of interactive crustal sheets in the Trans-Hudson Orogen (after Zwanzig, 1999). Sequence of events: 1) the arc–back-arc and marginal-basin domains collided with the Sask craton to form the middle and lower sheets, 2) the Superior craton was inserted from the southeast to partly overlie the middle sheets, and 3) the northern internal and external domains overrode the crustal stack from the northwest as an upper sheet. The Kisseynew and Waskaiowaka (WK) domains were forced obliquely out of the collision zone.

can generally still be recognized in the more highly metamorphosed and deformed rocks in the north (Zwanzig and Schledewitz, 1992; Zwanzig, 1999). They are considered to be tectonic rather than stratigraphic elements because their outer contacts and some internal contacts are structural. Individual volcanic units that make up the assemblages are defined by their field characteristics and geochemistry. Depositional contacts between the various units are only locally identified in the north, whereas the stratigraphic sequences in the Flin Flon and Snow Lake areas are clearly defined within the same major fault block (Bailes and Galley, 1999; R-L. Simard et al., work in progress, 2010). In many areas, metavolcanic units are thin, partly because they have been attenuated during deep-seated deformation. These are shown in numerous tectonostratigraphic columns, generalized from subareas where clusters of detailed information were collected. Some of the thin units are depicted on 1:10 000 or 1:20 000 scale preliminary maps (Schledewitz, 1988; Zwanzig, 1992a, b; Schledewitz, 1993a, b; Zwanzig and Parent, 1993; Zwanzig et al., 1993; Zwanzig and Shwetz, 1994a; Zwanzig, 1996a).

The columns are recognized as representing a coherent regional tectonostratigraphy because individual units are traced from one column to another and correlations are supported by the regional mapping. The columns are organized in structural packages that occur in a northeast-dipping to vertical structural stack (Bailes, 1980a; Lucas et al., 1999). Depositional contacts in the Kississing–File lakes area are more common for the sedimentary rocks, which are mapped as groups and informal units, very locally as formations. The unconformable contact of the Missi Group with the underlying Amisk Collage and plutonic rocks is generally intact and traced for a total of 150 km along the various structures. The disconformable to conformable contact with the Burntwood Group extends for a similar distance. For the first time, volcanic units intercalated with the Missi Group metasedimentary rocks, which were mostly unrecognized until recently, are described on the Flin Flon northern margin. These details have important implications for the regional tectonic evolution included in Section 6.1.5. Rock packages containing a volcanic assemblage and a sedimentary sequence extend across the area and make up the mappable tectonostratigraphy.

The discussion of intrusive rocks is restricted to short petrographic comments with average modal mineral content summarized from descriptions given in Zwanzig and Schledewitz (1992), and some based on full descriptions by Bailes (1980a). Tracing intrusive belts from the more weakly metamorphosed plutons in the south into equivalent, highly metamorphosed and deformed bodies in the north provides an understanding of the abundant orthogneiss in the Kississing–File lakes area.

Five tectonic subdomains are defined by the ages of contained rocks, their tectonic origin and contact relationships. These subdomains are 1) the Central Flin Flon subdomain, 2) the Snow Lake subdomain, 3) the Batty Lake subdomain, 4) the Kisseynew South flank, and 5) the Central Kisseynew subdomain (Figure 2). They are defined, respectively, by the presence of 1) the Amisk Collage, abundant 1876–1845 Ma plutons and unconformably overlying Missi Group; 2) the Snow Lake, Northeast Reed and Schist-Wekusko assemblages with abundant 1.84–1.83 Ga intrusions and fault slices of the Burntwood Group or Missi Group; 3) Sherridon–Meat Lake successor-arc assemblages and coeval intrusive rocks; 4) the Burntwood Group with conformable to disconformable overlying Missi Group and 1.83–1.79 Ga intrusions; and 5) the Burntwood Group and 1.83–1.79 Ga intrusions. The presence, in the Central Flin Flon subdomain, of narrow fault slices of early successor-arc volcaniclastic rocks at Schist Lake and Burntwood-age or slightly older (1860–1856 Ma) greywacke and mudstone at the Trout Lake mine (Rayner, pers. comm., 2009) indicates a somewhat similar tectonic history throughout the Flin Flon domain, but one that developed at different structural levels and might have been originally located along strike (Section 4.1).

Early metavolcanic assemblages are recognized by the presence of layered Josland Lake sills, which yielded a U-Pb zircon isotope dilution–thermal ionization mass spectrometry (ID-TIMS) age of 1886 ±3 Ma from a sample taken southwest of File Lake (Zwanzig et al., 2001). This provides an important constraint: some of the high-grade metavolcanic rocks in the Kississing–File lakes area can be considered to be equivalent in age to the early (>1.88 Ga) volcanic rocks of the Amisk Collage and the Snow Lake assemblage in the south. The presence of progressively metamorphosed Josland Lake sills from south to north supports the conclusion that their hostrocks are also structurally continuous. The Flin Flon (arc) assemblage, Elbow-Athapapuskow (ocean-floor) assemblage and Fourmile Island (arc) assemblage, all belonging to the Amisk Collage, were defined in the southern part of Central Flin Flon subdomain but are herein interpreted to underlie, as well, the southern margin of the Kississing–File lakes area. Amphibolite derived from the Fourmile Island assemblage and Josland Lake sills continues to the northwest to Kississing Lake in narrow fault slices, thus defining a northern extension of the Central Flin Flon subdomain. Although discontinuous, such narrow belts of amphibolite extend around the Sherridon–Hutchinson Lake Complex and Adamson Lake and Big Island domes, probably as far northwest as Yakushavich Island on Kississing Lake (Figure 7).

The 1876–1845 Ma successor-arc intrusions of the Central Flin Flon subdomain also extend north into the Kississing–File lakes area, where they are exposed in a series of domes or sheath-like culminations, commonly with remnants of recrystallized volcanic rocks of the Amisk Collage and mantles of the overlying Missi Group. The culminations form structural inliers in the north, including the Big Island dome, Adamson Lake dome, Sherridon–Hutchinson Lake Complex and Batty Lake Complex in Figure 7a. They are interpreted as equivalent to the composite arc massif of the Flin Flon area, comprising the Amisk Collage and stitching successor-arc plutons that are unconformably overlain by the 1848–1830 Ma meta-arenite and conglomerate of the Missi Group. The granitoid culminations with their associated supracrustal rocks are the most northerly inliers of the Central Flin Flon subdomain (Figure 2). The Missi Group meta-arenite or quartzofeldspathic gneiss contains intercalated metavolcanic rocks that are not present in the main part of the Central Flin Flon subdomain.

The Sherridon–Meat Lake assemblage, with its preliminary age of 1850–1855 Ma (N. Rayner and D. Tinkham, pers. comm.,

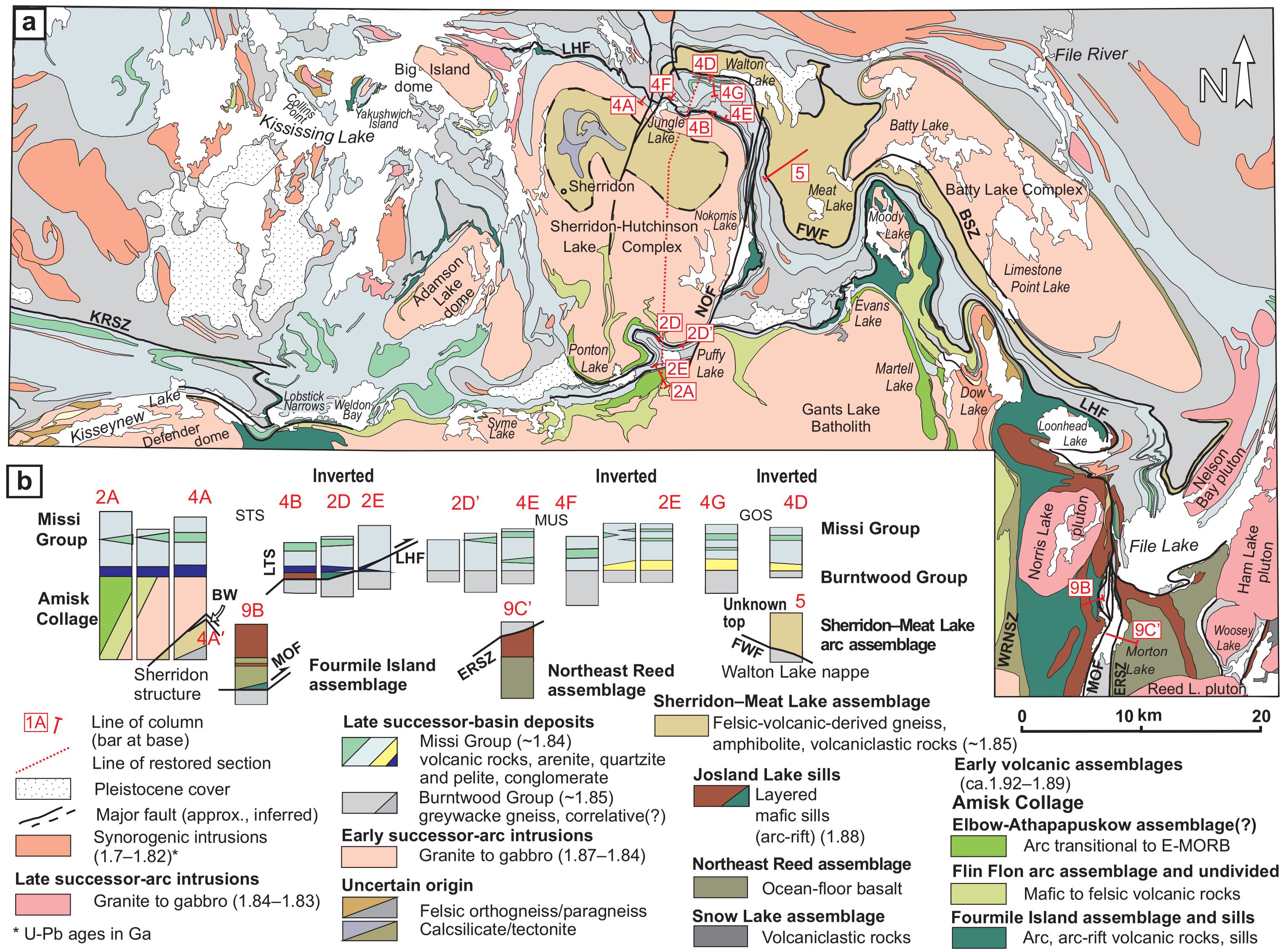


***Figure 7:** Geology of the Kississing–File lakes area. a) Map showing volcanic assemblages, major sedimentary units and intrusive suites. Locations of simplified tectonostratigraphic columns used in (b) are shown as numbered squares. b) Simplified restored tectonostratigraphic section derived from the structural interpretation of Zwanzig (1999), illustrating the key stratigraphic and structural relations. Spaces between columns represent folds. Wide labelled spaces are early isoclinal recumbent folds separating upright and inverted limbs: GOS, Gohl Lake syncline; MUS, Mud Lake syncline; STS, Star Lake synform. Fault/shear zones: BW, possible Burntwood Group in shear zone; BSZ, Batty Lake Shear Zone; ERSZ, East Reed Shear Zone; FWF, Fairwind Lake Fault; KRSZ, Kississing River Shear Zone; LHF, Loonhead Lake Fault; LTS, Loonhead Lake thrust sheet; MOF, Morton Lake Fault; WRNSZ, West Reed–Northstar Shear Zone. The folds and faults are discussed in Section 4.1. The columns are not to scale; estimated thicknesses are given in the text and Figure 8 (back pocket).*

2009), is defined herein as the lithotectonic package comprising the felsic to intermediate gneiss and amphibolite derived from predominantly juvenile volcanic-arc rocks exposed in the Sherridon–Meat Lake area and extending southeast to File Lake. The rocks are exposed in a structural basin at Meat Lake and in the structural culmination at Sherridon (Figure 7a). Whereas the Walton Lake nappe (Figure 8, back pocket) is probably a volcanic outlier, the Sherridon structure can be interpreted as an inlier at a deeper structural level. These volcanoplutonic areas, which were previously considered as part of the South flank of the Kisseynew Domain (Zwanzig and Schledewitz, 1992; Zwanzig, 1999), are now interpreted to be high-grade equivalents of the volcanoplutonic rocks of the Flin Flon Domain, and possibly part of the Snow Lake subdomain east of Wekusko Lake; in this report, they are named the Batty Lake subdomain. The Sherridon structure (between Sherridon and Jungle Lake) is interpreted as a window into a low structural level that contains the same rocks as the basin at Meat Lake but with the addition of more metasedimentary rocks (Zwanzig and Schledewitz, 1992). The Batty Lake Complex is interpreted as an extension of the Batty Lake subdomain. This is consistent with the age of the main intrusive phase of the Batty Lake Complex (ca. 1.86 Ga), which makes it part of the early successor-arc plutons and similar in age to the Sherridon–Meat Lake assemblage. The Batty Lake Complex is also closely associated with the Sherridon–Meat Lake assemblage by geochemistry and hydrothermal alteration. This association may have important economic ramifications (Section 7.1.3.10).

The Sherridon–Meat Lake assemblage in the Meat Lake area is surrounded by rocks of the Burntwood Group. Contacts are interpreted to be faults or shear zones, but some may be depositional. The predominantly felsic volcanic gneiss that occurs in the structural basin at Meat Lake, and the more mafic gneiss along its southeastern extension to File Lake are jointly bounded by the Fairwind Lake Fault in the southwest and the Batty Lake Shear Zone in the northeast (FWF and BSZ on Figure 7; Zwanzig, 1999). This structurally isolated body of predominantly metavolcanic gneiss, extending across Meat Lake and Walton Lake, has been interpreted as a small nappe (Walton Lake nappe of Zwanzig, 1999). The Batty Lake Shear Zone is also similar to the outer contact of the Sherridon structure, as it too contains a narrow sliver of Burntwood Group gneiss. In the Sherridon structure, the Sherridon–Meat Lake assemblage is separated from the surrounding intrusive rocks by a high-strain zone that contains local metagreywacke tentatively interpreted as Burntwood Group. Zones with high contents of graphite and Fe-sulphides, including massive graphite, may be the expression of a fault that surrounds the Sherridon–Meat Lake assemblage. The relationship suggests that the contacts of the Sherridon gneiss are similar to the Morton Lake Fault and have therefore been interpreted as a tightly folded subdomain boundary. The core of the structure contains uniform gneiss of a composition similar to greywacke. This unit is clearly not a turbidite but may be a correlative of the Burntwood Group, possibly deposited in an environment more proximal to its sediment source.

On Reed Lake, the west boundary of the Snow Lake subdomain is defined by the Morton Lake Fault (MOF), a major structural break that forms the eastern boundary of the Amisk Collage and western boundary of rocks containing major structural outliers (fault slices) of Burntwood Group rocks (Figure 5; Syme et al., 1995). These sedimentary rocks surround volcanic blocks that include the Northeast Reed (ocean-floor) assemblage south of File Lake and flank a unit of the Snow Lake (arc) assemblage in the extreme southeastern corner of the map area (Figure 7a). Well-preserved Burntwood Group rocks with turbidite structures, exposed at Reed, Morton and Wekusko lakes, extend north into equivalent staurolite-bearing greywacke-mudstone at File Lake and Snow Lake, and farther northwest into migmatitic paragneiss at Kississing Lake. These metasedimentary rocks and equivalent paragneiss units dominate the Kisseynew Domain.

Younger (1840–1830 Ma) successor-arc intrusions, which are common in the southern part of the Snow Lake subdomain, also extend northwest into orthogneiss and foliated intrusions west of Kississing Lake. They include the dated Reed Lake, Ham Lake and Nelson Bay plutons in the File Lake–Woosey Lake area. The undated, relatively massive Norris Lake pluton, west of File Lake, is also interpreted as part of the younger successor arc. Their importance is that they stitch together the earlier deformed structural sheets of volcanic rocks and the structurally interleaved Burntwood Group. Smaller intrusions in the northern and western parts of the Kississing–File lakes area are considered to be ≤1830 Ma by correlation with compositionally similar rocks dated in the Central Kisseynew subdomain.

The Kisseynew Domain includes the Central subdomain, in the northeast corner of the study area, and the South flank, which extends from Kississing Lake east beyond Batty Lake. The South flank is underlain mainly by the Burntwood Group and the disconformably to conformably overlying Missi Group (distal sedimentary facies and local volcanic rocks). The southern boundary of the Kisseynew Domain with the Flin Flon Domain is a series of faults and shear zones, including the Loonhead Lake thrust fault (LHF in Figure 5). This fault is interpreted to have carried the Amisk Collage, Josland Lake sills and Missi Group (basin-margin facies) over (or against) the Burntwood Group (columns 4B, 2D and 2E on Figure 7b; Zwanzig, 1999). The Loonhead Lake Fault truncates or merges with the Morton Lake Fault at File Lake and is refolded in a complex pattern northwest of the lake. East of the File Lake area, the Kisseynew Domain is bounded by the Snow Lake Fault (SLF in Figure 5) and major faults at Wekusko Lake. Where the bounding faults leave only a sheared sliver of the Missi Group above the Amisk Collage, the unconformity at the base of the Missi Group is taken as the regional domain boundary. The southern salients of the Burntwood and Missi groups in the Snow Lake subdomain have the mineral assemblages that record the increasing regional metamorphic grade from greenschist facies in the south to amphibolite facies in the north (Bailes and McRitchie, 1978; Bailes, 1980a).

Geological transects across the region provide structural sections rather than the early tectonostratigraphy. The structural sections, therefore, have been partly restored to upward facing from steeply dipping and structurally inverted segments to form a model of the early tectonostratigraphy pieced together from the Kississing–File lakes area (Figure 7b) by Zwanzig (1999). In these sections, the base of the Missi Group is used as datum for retrodeformation and to take advantage of systematic facies changes in the lower part of the Missi Group. The sections illustrate the north–south continuity of Missi Group units and the originally underlying volcanic thrust sheet. The lower part of the sections illustrates the abrupt break from the early volcanic and intrusive rocks to the Burntwood Group, which is interpreted to have originally underthrust the arc massif rather than being overthrust as interpreted elsewhere (e.g., Ansdell et al., 1999).

3.1 Early volcanic to successor-arc assemblages (ca. 1.90–1.85 Ga)

The metavolcanic rocks, subvolcanic intrusions and minor sedimentary rocks that are intruded by the oldest suite of plutons in the Kississing–File lakes area are considered to be part of the early (ca. 1.9 Ga) juvenile-arc and back-arc assemblages defined by Stern et al. (1995a, b) in the Flin Flon Domain. Such rocks have not been dated in the Kississing–File lakes area. Nevertheless, regionally extensive volcanic units (Moody Lake basalt and Storozuk volcanics) in the Fourmile Island assemblage, and the Fussey Lake basalt in the Northeast Reed assemblage are intruded by the 1886 ±3 Ma (Zwanzig et al., 2001) Josland Lake gabbro. The date was obtained on ferroan quartz diorite southwest of File Lake from the upper part of one of the thickest, most extensive Josland Lake differentiated sills (Bailes, 1980a). Other units interpreted to belong to early assemblages are intruded by slightly younger plutons. One unit (Puffy Mine andesite) is intruded by granite (Ragged Lake pluton) that yielded 1873 ±4 Ma zircons (Hunt and Zwanzig, 1990). Another unit (Ponton Lake basalt) is unconformably overlain by the ca. 1.85 Ga Missi Group. A narrow belt of fine-grained amphibolite west of Dow Lake (interpreted as part of Ponton Lake basalt) contains tonalite dikes that form a margin phase of the ca. 1890 Ma (J. Whalen, pers. comm., 2007) to 1876 +7/–6 Ma (Whalen and Hunt, 1994) Gants Lake Batholith.

Based on these field relationships the early assemblages in the study area are considered to have a minimum age of 1886 Ma. This includes the high-grade part of the Amisk Collage in the northern extension of the Central Flin Flon subdomain and the Northeast Reed assemblage in the Snow Lake subdomain. Only the Sherridon–Meat Lake assemblage has poorly constrained ages. It was tentatively correlated with the early volcanic rocks at Snow Lake because it lies along strike and has a similar composition, trace-element geochemistry, hydrothermal alteration and metallogeny (Zwanzig, 2000c). A recently determined ID-TIMS age of 1850–1855 Ma (N. Rayner and D. Tinkham, pers. comm., 2009), however, indicates that this assemblage of predominantly felsic rocks appears to belong to the early successor-arc deposits. It is described herein with the earlier volcanic rocks because of its similar geochemistry and metallogeny (Sections 5.7.5 and 7.1.3). Several undated units, moreover, have been broadly grouped with the Sherridon–Meat Lake assemblage (Maps GR2010-1-1 and -2, back pocket/DVD; Appendix 4, Table 1, DVD). Undated early dikes and sills, which are locally closely associated in the field with the early volcanic rocks although not generally distinguished from early successor-arc intrusions, are usually treated as coeval and included with the host unit where they are interlayered and share geochemical characteristics with the host flows. The term 'juvenile' has been dropped in this report for the early assemblages in the Kississing–File lakes area because all of these, as well as the successor-arc deposits, are generally isotopically juvenile. Those few that are not juvenile appear to be closely related to the rest in geochemistry.

The original stratigraphic relationships among most of the early assemblages are unknown because they have structural or intruded contacts. Faults or long-lived shear zones commonly occur between assemblages, similar to relationships observed in better preserved equivalents to the south (e.g., Lucas et al., 1996; Ryan and Williams, 1999). In the Kississing–File lakes area, as in the lower grade rocks to the south, units within some assemblages have depositional or interlayered contacts that are clearly stratigraphic. In addition, other contacts lie between units with very similar geochemical characteristics, suggesting a stratigraphic affiliation and accumulation within an evolving volcanic pile. A model that helps to establish the structural relations among the early assemblages and their stratigraphic relation with the successor-arc deposits was provided by Zwanzig (1999). The tectonostratigraphic units, their contact relationships and present (deformed) thicknesses are compiled in Table 3.

The tectonic affinity of the various units in the Kississing–File lakes area has been inferred from the petrology and physical volcanology (where preserved) but mainly from geochemistry. These data show that the rocks mapped as Amisk Collage comprise an arc-volcanic assemblage (included with the Flin Flon assemblage) and an ocean-floor assemblage (tentatively included with the Elbow-Athapapuskow assemblage). The arc-like rock consists partly of andesite, locally associated with rhyolite or intermediate breccia. The ocean-floor basalt is distinguished by thick uniform successions of pillows with abundant sills or dikes of the same composition. Similar characteristics define, respectively, the high-grade Sherridon–Meat Lake (arc) assemblage and the Northeast Reed (ocean-floor) assemblage. The former includes a particularly high proportion of felsic gneiss derived from rhyolite and probably volcaniclastic rocks and felsic subvolcanic intrusions.

3.1.1 Northeast Reed assemblage

A volcanic assemblage extending from Reed Lake to File Lake consists of 2.5–3 km of uniform pillowed to massive basalt that has been shown to be ocean floor (Syme and Bailes, 1996) and is therefore devoid of interlayered sedimentary rocks or felsic volcanic rocks. The assemblage is structurally isolated: its western boundary is the East Reed Shear Zone (ERSZ in Figure 7) and the northern limit is the eastern arm of the Loonhead Lake Fault (Connors, 1996); and the eastern limit was interpreted as a fault by Syme et al. (1998). All of these structures juxtapose basalt with the (younger) Burntwood Group turbidites. The metagreywacke structurally envelops the Northeast Reed assemblage but also occurs in the roof and core of the invading (underlying) plutons. The Reed Lake assemblage probably represents the remnant of a klippe that has been emplaced over the Burntwood Group (Zwanzig, 1999).

Table 3: Field characteristics of early volcanic to successor-arc assemblages in the Kississing–File lakes area.

Thickness (m)	Unit	Name (analyzed units)	↑Upper contact ↓lower contact	Lithology
AMISK COLLAGE (intruded by early-successor arc plutons, ca. 1.885 Ga amalgamation)				
Elbow-Athapapuskow assemblage (?)				
0–15	1b	Reed Lake complex	Stuctural lenses	Untramafic intrusive
1400	2a	Ponton Lake basalt	↑Unconformity (Missi) ↓unexposed	Mafic to ultramafic flows, dIkes and sills, foliated amphibolite (northwest), mafic tectonite (east)
Fourmile Island assemblage				
150	6a	Yakymiw Formation	↑Unexposed	Mudstone, siltstone, volcanic wacke
1200	2e	Storozuk volcanics	↑Conformable	Basalt and andesite, flows, breccia
700	2d	Dickstone rhyolite	↑Conformable	Rhyolite flows and breccia
>600	2c	Preaston andesite	↑Conformable	Flows and fragmental andesite
400	2b	Moody Lake basalt	↑Unconformity (Missi)	Pillow basalt, gabbro, basaltic Fe-andesite
150	3c[1]	Martell Lake basalt	↑Unexposed ↓tectonite	Fine-grained amphibolite
250	3d	Dow Lake andesite	↑Unexposed	Fine-grained amphibolite +garnet
130	3e[1]	SE Dow Lake porphyritic basalt	Interlayered	Plagioclase-phyric massive and pillowed basalt, breccia
Flin Flon Arc assemblage				
300	3a	Puffy Mine andesite	↑Unconformity (Missi) ↓intrusive	Amphibolite ± garnet-biotite, local pillows, calc-silicate alteration, veined
			Overlain by Ponton Lake basalt	
450	3b	Koscielny andesite	↑Unexposed / intrusive	
ASSEMBLAGES (not intruded by early successor-arc plutons but with 1.84 Ga re-amalgamation)				
Northeast Reed assemblage				
2500	1a	Fussey Lake basalt	↑Faulted / intrusive	Uniform pillow basalt, gabbro
Sherridon-Meat Lake assemblage				
?1400	5–5d	Sherridon gneiss	↑Faulted / faulted	Felsic volcanic gneiss, breccia, quartz porphyry and alteration
~100 aggregate	4c	Meat Lake amphibolite	Interlayered with Sherridon gneiss	Amphibolite, pillow basalt, breccia
~150 aggregate	4d	Walton Lake amphibolite	Interlayered with Sherridon gneiss	Fine-grained amphibolite (volcanic), gabbro
~300	4f	Cree Lake calcsilicate	Interlayered with Sherridon gneiss	Medium-grained mafic calcsilicate rock (presumed volcanic origin)
Snow Lake assemblage				
0–600	6a	Parisian Formation	↑Intruded ↓conformable	Volcanogenic paraconglomerate

[1] Units 3c and 2f may lie on opposing limbs of an anticline with unit 3d in the core.

Alternatively, it formed as a thrust imbricate within the Burntwood Group (Connors, 1996).

The northern part of the Northeast Reed assemblage (Butler Lake–Fussey Lake sequence of Bailes, 1980a) forms an important unit in the Kississing–File lakes area and has been named the Fussey Lake basalt (unit 1a in this report; Figure 9a). The unit is intruded by the 1.89–1.84 Ga Reed Lake pluton phase and Josland Lake sills, which have been dated as 1886 Ma northwest of Reed Lake (column 9C–9C' on Figure 8c).

3.1.2 Amisk Collage

In the Kississing–File lakes area, the Amisk Collage and prominent sills of Josland Lake gabbro extend from the west side of the Morton Lake Fault north to Moody Lake and continue west to Kisseynew Lake in roof pendants of a successor-arc pluton (Figure 8a, back pocket). The Loonhead Lake thrust sheet, which has Josland Lake gabbro and thin Amisk Collage at the base, extends from Evans Lake northwest to Kississing Lake, forming mantles on the Sherridon–Hutchinson Lake Complex and the Big Island and Adamson Lake domes.

The Amisk Collage in the Kississing–File lakes area includes, from southwest to northeast, 1) the Flin Flon arc assemblage, 2) the Elbow-Athapapuskow (ocean-floor or back-arc) assemblage, and 3) the Fourmile Island assemblage. These three assemblages were distinguished and mapped on field characteristics, which was aided by their distribution in continuous belts separated by structural discontinuities. Units containing thick uniform successions of pillow basalt and gabbro (units 1a–2b) with no sedimentary or felsic volcanic intercalations are interpreted as back-arc–basin basalt or arc-rift basalt. Coeval basalt to andesite and rhyolite (units 2c–e, 3a–d) belong to arc assemblages. Amphibolite derived from andesite, indicative of

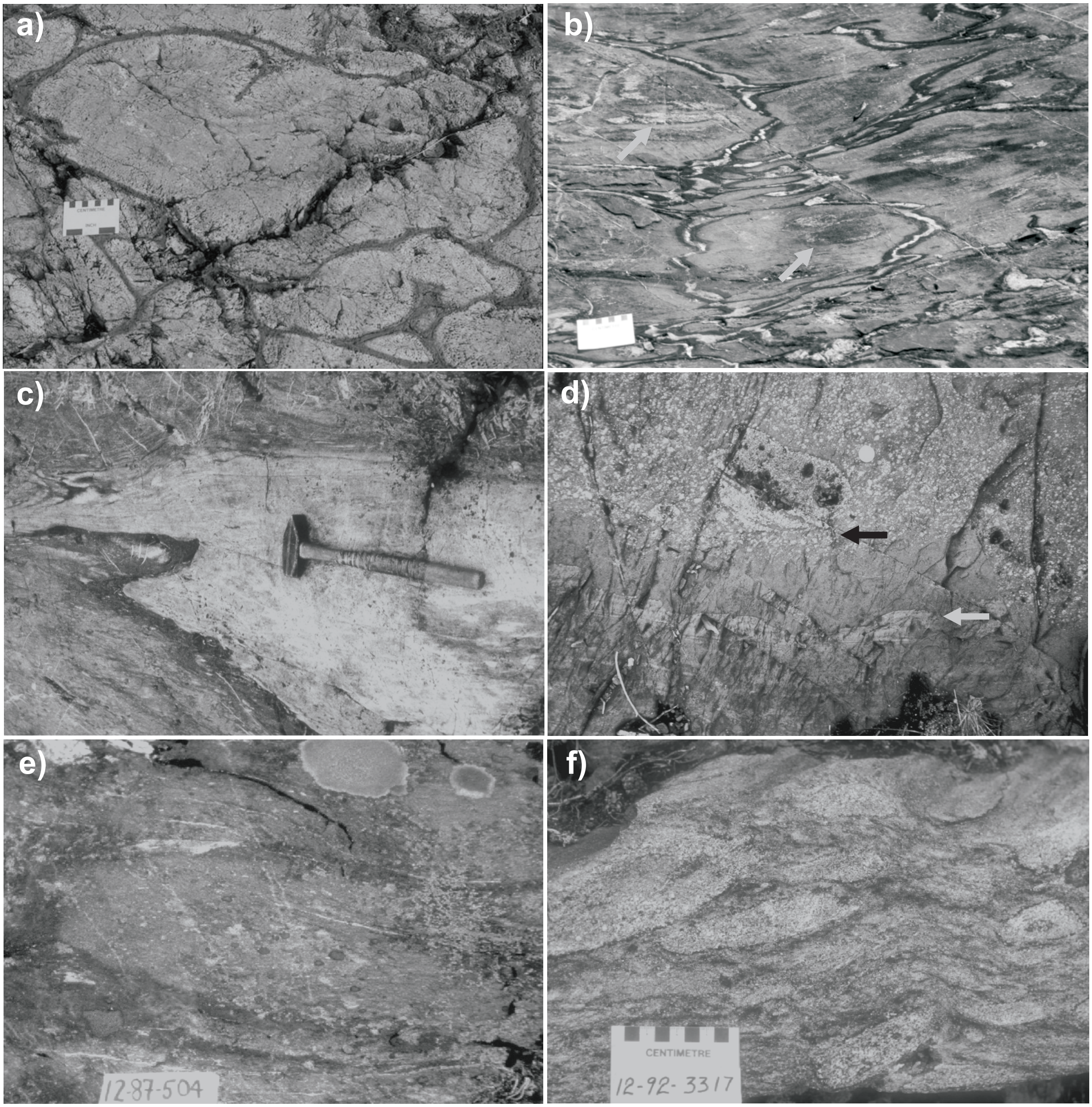


***Figure 9:** Primary structures of early volcanic assemblages (see Appendix 4 for these and additional photographs in colour): **a)** Well-preserved pillow structure and composition of Northeast Reed (ocean-floor) assemblage, Reed Lake, southwest of the study area. **b)** Ponton Lake basalt pillows, probable Elbow-Athapapuskow ocean-floor assemblage, south of Puffy Lake, near the sillimanite isograd; deformed pillow rinds have a plagioclase core and hornblende rims; epidosite seafloor-alteration domains are metamorphosed to plagioclase-diopside lenses (arrows). **c)** Dickstone rhyolite lobe (white) in finely fragmental rhyolite (dark), although most of the unit is massive; in combination with reaston andesite southwest of File Lake, this is a typical arc or arc-rift assemblage. **d)** Yakymiw volcanogenic sediments, showing a relatively felsic, fine-grained laminated bed (white arrow) and the coarse-grained base of an overlying bed (black arrow) with a large fragment, southwest of File Lake. **e)** Pillow structure in generally poorly preserved Moody Lake basalt, north of the sillimanite isograd, southwest of Moody Lake; tape is 10 cm long. **f)** Locally preserved fragmental texture of andesite in Meat Lake amphibolite (Sherridon–Meat Lake assemblage), 3 km northwest of Batty Lake.*

an arc origin, was distinguished in the field from amphibolite of basaltic origin by the presence of small, scattered garnet porphyroblasts. Where primary volcanic structures are preserved, breccia can also serve as a guide to arc-derived amphibolite. Ocean-floor units are recognized by abundant alteration to calc-silicate rock, found as plagioclase-diopside–rich alteration and interpreted as sea-floor alteration. Individual units were also distinguished by their consistent association with other rock types, such as the Josland Lake sills. Consequently, most of the individual units that were found to have a unique geochemistry during this study had already been recognized and traced on the basis of rock associations and macroscopic field characteristics. A notable exception is the rocks in the Dow Lake area, where preliminary geochemical plots were used to distinguish some of the highly strained amphibolite units (Zwanzig, 1996b).

3.1.2.1 Possible Elbow-Athapapuskow assemblage

Two small tectonic lenses of ultramafic rock (unit 1b), one west of the northern extension of the West Reed–North Star Shear Zone (WRNSZ in Figure 7) and another included in an adjacent intrusion, are interpreted as fragments of the ca. 1.90 Ga Reed Lake mafic–ultramafic complex (Syme et al., 1995). They are coarse-grained pyroxene-rich rocks ±olivine±spinel and considered to be part of a dismembered ophiolite in the Elbow-Athapapuskow assemblage (Stern et al., 1995a). The distinctive ultramafic lenses, lying consistently on the west side of the WRNSZ, form an important lithological and structural link between the Reed Lake area in the south and the high-grade rocks near Dow Lake in the north.

A monotonous succession of pillow basalt (Figure 9b) with subvolcanic gabbro sills makes up the Ponton Lake basalt (unit 2a). It lies structurally above andesite units (column 2A–2A' in Figure 8b, back pocket), below the WRNSZ, and is tentatively included in the Elbow-Athapapuskow assemblage, which has similarly high-Mg flows and sills, and lies on the other side (west) of the Gants Lake Batholith. The contact between Ponton Lake basalt and the underlying andesite units is not directly exposed but partly intruded by granitic rocks. The contact may be structural because there is no interlayering to suggest an overlap of the basalt and andesite volcanism. The Ponton Lake basalt is repeated along the southern margin of the Sherridon–Hutchinson Lake Complex. Another narrow belt of amphibolite correlated with the Ponton Lake basalt forms the northeastern sidewall of the Gants Lake Batholith west of Martell Lake (column 8A). The combined strike length of these discontinuous slivers of Ponton Lake basalt is 30 km. The basalt is unconformably overlain by the Missi Group in an early synclinal structure involving all of these belts.

3.1.2.2 Fourmile Island assemblage

The volcanic units between the West Reed–North Star Shear Zone and the Morton Lake Fault (WRNSZ and MOF in Figure 8a, back pocket) belong to the Fourmile Island assemblage (Syme et al., 1995). This assemblage is herein interpreted to extend from the Reed Lake area northwest to Kississing Lake, where it occurs in the Loonhead Lake thrust sheet. The Fourmile Island assemblage includes the compositionally and texturally heterogeneous arc-volcanic rocks mapped by Syme et al. (1995), as well as thick volcanic units west of Morton Lake (Bailes, 1980a) and an extensive unit of uniform pillow basalt (Moody Lake basalt, unit 2b) to the northwest. The sheared structural base of the Fourmile Island assemblage is exposed at Dow Lake, where the tectonites include the remnants of the Reed Lake mafic–ultramafic complex. Mafic flows with highly flattened pillows are recognized on the sheared western part of the assemblage near North Star Lake (Norquay, 1997). Assigning the northern units to the Fourmile Island assemblage is based on several factors consistent with the structure section established on the southwestern shore of Reed Lake (Syme et al., 1995): 1) the units are arc related, including andesite, rhyolite and basaltic tholeiite; 2) they are intruded everywhere by the same widespread sills of Josland Lake gabbro; 3) their base is the West Reed–North Star Shear Zone, one of the major structural discontinuities in the Amisk Collage (WRNSZ); 4) in the northwest, they are separated from the Elbow-Athapapuskow ocean-floor assemblage by the extension of WRNSZ and the Star Lake syncline; 5) the Fourmile Island assemblage is everywhere structurally juxtaposed against the Burntwood Group to the northeast along the Loonhead Lake Fault and the Morton Lake Fault (Syme et al., 1995; Connors, 1996; Zwanzig, 1999).

The Fourmile Island assemblage is 1800 m thick west of Morton Lake and includes mafic to felsic volcanic rocks: Preaston andesite (unit 2c) at the base, stratigraphically overlain by Dickstone rhyolite (unit 2d, Figure 9c) and, in turn, the Storozuk volcanics (unit 2e), which are dominated by basalt. A volcaniclastic unit, the Yakymiw Formation (Figure 9d) forms the top of this succession (Bailes, 1980a; column 9A in Figure 8c).

Poorly preserved units of amphibolite that are possible members of the Fourmile Island assemblage surround the Dow Lake domal structure in the present hangingwall of the northern extension of the WRNSZ. These units include weakly layered amphibolite (Martell Lake basalt, unit 3c) and garnetiferous amphibolite (Dow Lake andesite, unit 3d). They are overlain in the east by minor rhyolite and the Moody Lake basalt (unit 2b). The adjoining Storozuk volcanics (column 8B in Figure 8c) have an unknown field relationship. Moody Lake aphyric basalt is interlayered in the west with SE Dow porphyritic basalt (unit 2f). Martell Lake basalt may lie stratigraphically beneath the Moody Lake basalt, but its facing is uncertain because the Moody Lake–Dow Lake area has a complex structural history. West of Moody Lake, the Fourmile Island assemblage consists of only Moody Lake basalt. This unit is moderately well exposed as pillow basalt (Figure 9e) and gabbro south of the lake. It is intruded by Josland Lake sills and unconformably overlain to the south by the Missi Group in a distinctive lithological package that extends west to Kisseynew and Kississing lakes, stratigraphically above the Loonhead Lake Fault (Figure 8d). Moody Lake basalt is recognizable in this package even where pillows are extremely flattened (Photo 66, Appendix 4, Table 1, DVD).

3.1.2.3 Flin Flon arc assemblage

Metavolcanic rocks lying at a deeper structural level are generally poorly preserved and occur in thin roof zones at the margins of various plutons. Their basaltic to andesitic composition and the presence of volcanic breccia and rhyolitic rocks suggest that they are part of a volcanic-arc assemblage, herein tentatively correlated with the Flin Flon arc assemblage (Stern et al., 1995b). This part of the assemblage includes Puffy Mine andesite (unit 3a) and Koscielny andesite flows and breccia (unit 3b), which are preserved in metamorphosed supracrustal remnants occurring, respectively, northwest of Puffy Lake and southeast of Ponton Lake (columns 2A and 2C in Figure 8b, back pocket). These enclaves of volcanic rock are a few hundred metres wide and about 3 km long. Koscielny andesite occurs at the margin of the large area of early successor plutons on the south side of the Kississing–File lakes area and is overlain by Ponton Lake basalt (Schledewitz, 1993c). Puffy Mine andesite occurs between deformed granitoid plutons at the southern margin of the Sherridon–Hutchinson Lake Complex. It is unconformably overlain by the Missi Group.

Other areas of supracrustal rocks along the eastern and northern margins of the Gants Lake Batholith and extending west as far as Kisseynew Lake contain mafic to felsic volcanic rocks that probably belong to the Flin Flon arc assemblage, but the geochemistry of these has not been determined. The rocks are assigned to units 4, 4a and 4b, representing mafic, mixed and felsic compositions, respectively (below).

3.1.3 Amphibolite to felsic rocks of uncertain affinity

The major granitoid intrusive bodies and gneissic domes commonly have supracrustal screens of mafic to felsic composition or partial mantles of such units. The structural relations of these rocks are like those of the early volcanic assemblages, forming narrow side walls on the early plutons, but the rocks may include small deformed successor-arc intrusive and coeval volcanic rocks. Layered and uniform amphibolite of basaltic composition is most common, but garnetiferous varieties include compositional andesite or volcaniclastic rocks. These undivided rocks (unit 4) may be part of the Amisk Collage or the Sherridon–Meat Lake assemblage. Narrow belts of amphibolite with local pillow structure occur along the northern and eastern margins of the Amisk Collage. These are designated as unit 4 because no geochemical data are available. Massive and pillowed aphyric basalt with epidosite alteration domains occurs 5 km north of North Star Lake (Norquay, 1997). Generally massive, plagioclase-phyric amphibolite occurs with minor dacite to rhyolite (unit 4b) at Spider Lake. Another belt of locally pillowed basalt (unit 4) and the underlying felsic rock (unit 4b) extend from Syme Lake west to Weldon Bay on Kisseynew Lake.

A more heterogeneous unit (4a) forms a belt, up to 2 km wide, extending from Fairwind Lake into the main body of the Walton Lake nappe. This belt may also extend southeast of Fairwind Lake around the File Lake synform north of File Lake to Folster Bay, where it contains fine- to medium-grained amphibolite with local pillow structure (Bailes, 1980a). Unit 4a generally includes layered amphibolite, hornblende-plagioclase gneiss, minor felsic gneiss, and calcsilicate gneiss with diopside (0–50%), epidote and minor calcite. These rocks were probably derived from carbonatized and epidotized basalt, as well as from volcaniclastic or felsic layers (Zwanzig and Schledewitz, 1992). Rocks with porphyroblasts of garnet±cordierite±gedrite were produced by hydrothermal alteration. The same rock types (unit 4a), bounded by Batty Lake trondhjemite gneiss (unit 9f), occur in the 100–500 m wide belt extending from the centre of Batty Lake for 12 km to the southwest. Unit 4a also occurs on the northern margin of the Sherridon structure where it is interpreted as part of the Sherridon–Meat Lake assemblage.

A predominantly mafic volcanic unit with lesser intermediate rock also includes siliciclastic members and iron formation (Norquay 1997). This belt of unit 4a, along with its flanking rhyolitic units, is interpreted as part of the Flin Flon arc assemblage. It follows a complex fold pattern extending from 1 to 10 km north of North Star Lake. The iron formation, which is generally not found elsewhere in the Flin Flon Domain, includes garnet-rich (silicate-facies) and magnetite-rich (oxide-facies) members up to 10 m thick. These are part of a sedimentary succession up to 40 m thick (Norquay, 1997).

The rhyolitic rocks (unit 4b) flanking unit 4a in the North Star Lake area occur in two belts, 600 and 100 m wide. The rocks in the wider belt vary from laminated to massive, commonly with areas of rusty weathering. The massive to layered rocks in the narrow belt contain small biotite-rich lenses (Norquay, 1997). The felsic unit (4b) between Syme Lake and Kisseynew Lake is gneissic and contains garnet. Areas with sillimanite and large porphyroblasts of garnet suggest local hydrothermal alteration. Similar rocks on the east shore of Yakushavich Island in Kisseynew Lake are possibly part of the Sherridon gneiss (unit 5), which hosts the 'Yak' Cu-Zn occurrence.

Better preserved aphyric dacite on Morton Lake (unit 4 of Bailes, 1980a) is herein mapped as unit 4a of unknown age and affinity because of its lack of contact relationships and because the geochemical data of Bailes (1980a) indicate that it is unrelated to any known unit in the Kississing–File lakes area.

3.1.4 Sherridon–Meat Lake assemblage

A bimodal succession of felsic gneiss and amphibolite (Sherridon Group of Bateman and Harrison, 1946; Robertson, 1953; Froese and Goetz, 1981), which extends from File Lake 60 km northwest to Sherridon (Figure 8a, back pocket), is interpreted to have been derived by high-grade metamorphism of predominantly volcanic rocks (Ashton and Froese, 1988; Zwanzig and Schledewitz, 1992). The gneissic units share characteristics with the rocks exposed northeast and northwest of Snow Lake, including 1) predominance of felsic rock types (Sherridon gneiss, units 5–5d) that were derived largely from rocks with a rhyolitic composition; 2) local preservation (in the felsic rocks) of breccia structure and layering, interpreted as contacts between flows or tuff units and transposed flow layering; 3) subordinate amphibolite (units 4c–e), locally showing pillow structure or breccia (Figure 9f); 4) abundant regional Fe-Mg and calcsilicate alteration (units 5b–e) that is locally focused and mineralized; 5) fault contacts with the Burntwood Group, suggesting that the rocks are allochthonous; and 6) high metamorphic grade like the felsic rocks in the mantle of the domal structures north of Snow Lake. Apparently, the Sherridon–Meat Lake assemblage had a similar type of tectonic history to the rocks in the Snow Lake subdomain. However, the younger preliminary isotopic age determined for the Sherridon felsic gneiss suggests that it is not related to any other volcanic assemblage in the region.

A present thickness of 1600 m of felsic gneiss and amphibolite is exposed in the Walton Lake–Meat Lake area, of which at least 500 m contains abundant garnet and may be tuff altered during hydrothermal activity and possibly sedimentary reworking (column 5 in Figure 8e). The predominantly felsic gneiss is structurally juxtaposed against the Burntwood Group and Batty Lake Complex along the Batty Lake Shear Zone in the northeast, and the Burntwood Group along the Fairwind Lake Fault in the south (Zwanzig, 1999). The gneiss has been interpreted to occupy a small nappe (Walton Lake nappe of Zwanzig, 1999). The window in the domal Sherridon–Hutchinson Lake Complex contains a very similar succession of gneiss (Froese and Goetz, 1981). This unit, which was little sampled for geochemistry for this report, is correlated with the gneiss in the Walton Lake nappe and included in the Sherridon–Meat Lake assemblage based on field appearance and abundant geochemical data from Halo Resources Ltd. (L. Bloom, pers. comm., 2007; S. Timpa, pers. comm., 2008).

Some units of amphibolite and intermediate gneiss extend for 10 km along strike and follow the folded outline of their felsic host (Maps GR2010-1-1 and -2, back pocket/DVD). These are slightly more abundant in the Sherridon structure, but more detailed sections, such as the one shown in column 5, have not been traced along strike for this report. Facing direction and the possible presence of structural repetitions are presently unknown in the Walton Lake nappe. The presence of cordierite±gedrite–bearing gneiss (hydrothermal alteration) above and below VMS deposits suggests internal isoclinal folding within the larger bodies of this unit. The long southeast extension of the Walton Lake nappe, which is dominated by intermediate to mafic gneiss, shares the highly altered felsic gneiss that characterizes the main part of this structure. Pillow and fragmental shapes are locally preserved, most prominently in the narrow body at Folster Bay on File Lake (column 9E in Figure 8e). Photographs of the volcanic-derived felsic rocks and amphibolite are available in Appendix 4, Table 1 (Photos 52–63, DVD).

3.1.4.1 Walton Lake and Meat Lake amphibolite

Amphibolite units intercalated with Sherridon (felsic) gneiss include two varieties: fine-grained amphibolite with local pillow or breccia structure (Meat Lake amphibolite, unit 4c); and coarse- to fine-grained amphibolite to ultramafic rock (Walton Lake amphibolite, unit 4d), derived in part from pillow basalt and lapillistone with variable phenocryst and amygdule distribution but also from local gabbro. Where intrusive contacts are exposed, metagabbro is designated as unit 4e. Overall, there are probably equal proportions of extrusive and intrusive components. These mafic units occur as layers up to 40 m thick in the Walton Lake nappe (column 5 in Figure 8e, back pocket). No internal stratigraphy is mappable in the nappe because of poor exposure, but Meat Lake amphibolite is more abundant and Walton Lake amphibolite has been confirmed only in a single discontinuous layer.

The persistent presence of the amphibolite in the Sherridon–Meat Lake assemblage makes the stratigraphy in the Sherridon structure and the Walton Lake nappe bimodal, thus supporting the correlation of rocks in these two structures.

3.1.4.2 Mafic calcsilicate gneiss

Relatively mafic calcsilicate rock and associated amphibolite in the core of the Sherridon structure were previously mapped as impure marble and interlayered calcsilicate (Froese and Goetz, 1981). This unit was interpreted as highly carbonatized mafic volcanic rocks (Ashton and Froese, 1988) or layered intrusive rocks, supported herein by geochemical data. It is therefore designated as unit 17c to indicate its geochemical and structural affinity with unit 17, but its age is unknown and may be as old as unit 1. Lithologically and geochemically similar rocks with remnants of volcanic structure are mapped elsewhere on the margin of the Kisseynew Domain as highly altered metabasalt and ultramafic tuff breccia, interpreted to be about 1.9 Ga (Zwanzig, 2008). The favoured affinity is with late intrusive rocks (unit 17) because it has retained a geochemical signature similar to an ultramafic–intermediate intrusive suite (unit 17b, c).

3.1.4.3 Sherridon Gneiss

The felsic gneiss is dominated by quartz-rich rocks (Sherridon gneiss, units 5–5d in this report), which is commonly variably garnetiferous but has a predominantly rhyolitic composition. The felsic gneiss makes up the largest parts of the Walton Lake nappe and the Sherridon structure. These units were previously interpreted entirely as paragneiss (Robertson, 1953; Froese and Goetz, 1981) based on their locally layered nature, high quartz content and the common presence of scattered garnet porphyroblasts and local sillimanite. In this report, much of the layering is interpreted as metamorphic or structurally transposed, attenuated units of tuff that are commonly altered.

The best preserved felsic gneiss (unit 5a) is weakly layered, locally with variations in quartz phenocryst populations or remnants of fragmental structure indicative of a volcanic origin. Uniform, coarsely porphyritic varieties may represent endogenic domes (Photos 54, 55, Appendix 4, Table 1, DVD). One of the early clues that the Sherridon gneiss comprises igneous rock is its previously unexplained uniform chemical composition, published by Froese and Goetz (1981). This characteristic is supported by our work. The high silica content is similar to that of silicified rhyolite in the main part of the Flin Flon Domain (Syme, 1998).

Altered rocks containing calcsilicate minerals (unit 5b), occurring in the structurally lower part of the Sherridon–Meat Lake assemblage, are interpreted to have been derived from rhyolite and possibly mafic rock. Most of the weak regional alteration (unit 5c or part of unit 5, undivided), however, is marked by garnet porphyroblasts, some up to 2–3 cm in diameter. The garnet may have been derived from hydrothermal chloritic alteration or from biotite by high-grade metamorphism (transitional to granulite) and incipient anatexis. Patches where garnet and/or amphibole and local sulphides are concentrated (unit 5d) support an origin of metamorphosed domains of regional hydrothermal alteration in rhyolite or reworked tuff (Figure 10a; Photos 57, 58, Appendix 4, Table 1).

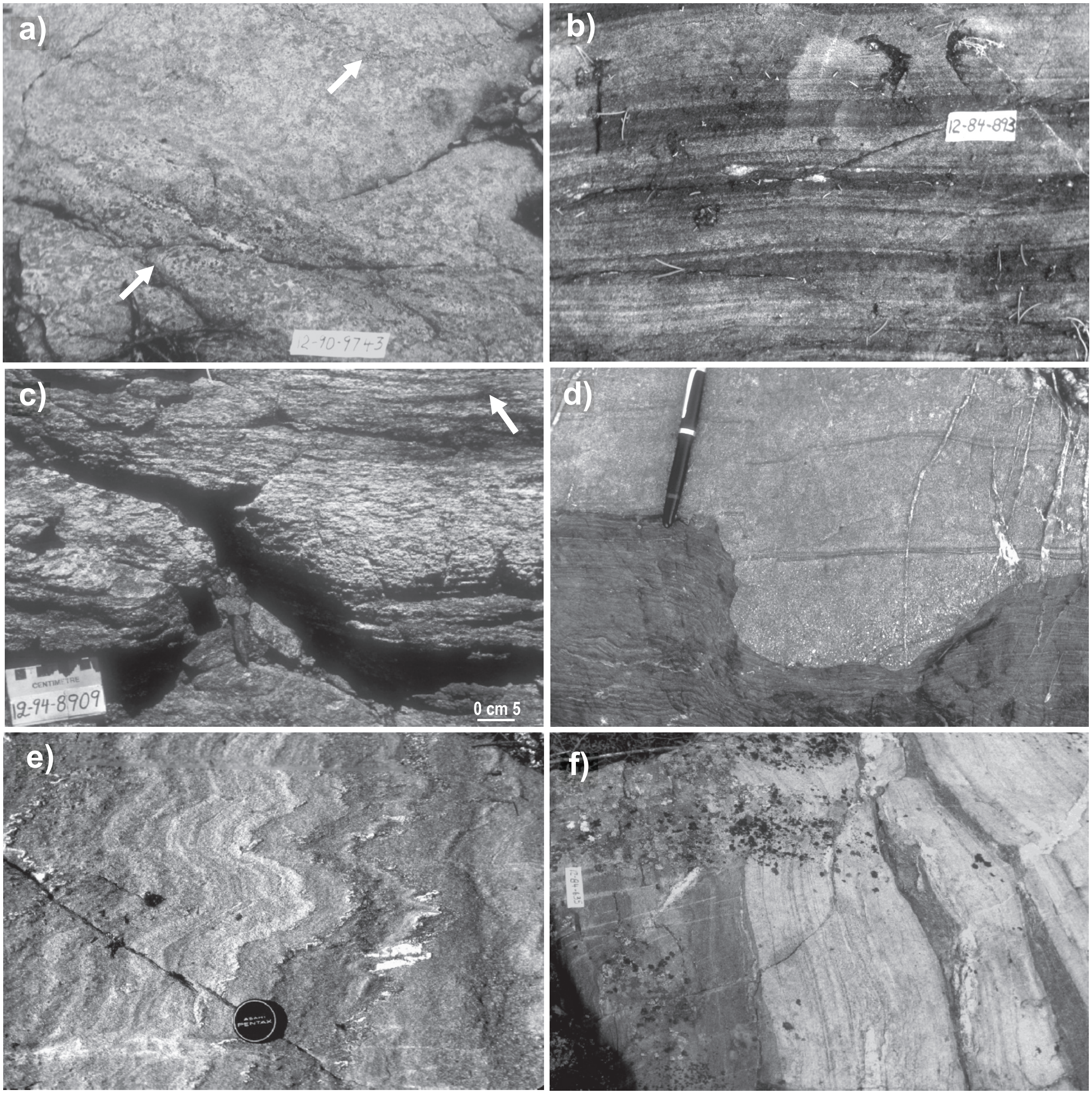


***Figure 10:** Early sills and successor-arc rocks (see Appendix 4 for these and additional photographs in colour): **a)** Sherridon gneiss with patches of hornblende-rich alteration (arrows), probably derived from rhyolite; tape is 10 cm long. **b)** Modal layering in Josland Lake mela- to leucogabbro, Nokomis Lake; tape is 10 cm long. **c)** Felsic mottled gneiss derived from trondhjemite with local deformed alteration along early fractures (arrow), underlain by garnet amphibolite derived from ferrogabbro (at scale card), Nokomis Lake. **d)** Scour in Burntwood Group greywacke-mudstone turbidite from upper-slope fan environment, File Lake. **e)** Folded Burntwood Group garnet-biotite gneiss derived from thin sandy (light) to mainly shaly (dark) beds in a lower-slope environment, north of Kisseynew Lake; lens cap is 5 cm. **f)** Interlayered pelite (dark) and quartzite from upper beds in the Burntwood Group, probable regressive shoreline environment below the nonmarine Missi Group, southwest of Cleunion Lake; tape is 10 cm long.*

The most garnet-rich varieties of Sherridon gneiss with cordierite-sillimanite±gedrite±hercynite (unit 5d) have been interpreted as products of more focused hydrothermal alteration (Ashton and Froese, 1988; Zwanzig and Schledewitz, 1992). Relatively biotite rich garnetiferous varieties of Sherridon gneiss may represent alteration or, more likely, epiclastic rocks as originally mapped. These rocks occur locally in the northeastern and northwestern areas. Cordierite-gedrite–bearing rocks (unit 5d) also occur along the margins of the Batty Lake Complex, suggesting hydrothermal alteration in the roof of the trondhjemite protolith and its volcanic cover, which may be nearly coeval.

3.1.5 Volcaniclastic and epiclastic rocks

Fine- to coarse-grained clastic rocks occur in the vicinity of Woosey, Morton, Evans and Hutchinson lakes. They are intruded by either the Josland Lake sills or the earliest successor-arc plutons, and are therefore considered to be synvolcanic and part of the early assemblages. The coarser, lower grade rocks in the southeast are mapped as unit 6a, and greywacke-derived biotite±hornblende gneiss in the northwest as unit 6b. The lower grade rocks are described in detail by Bailes (1980a) and briefly summarized with the high-grade rocks below.

A wedge of paraconglomerate, up to 600 m thick, outcrops on Woosey Lake (Parisian Formation of Bailes, 1980a). It is herein considered part of the Snow Lake assemblage and overlies altered, fragmental, mafic volcanic rocks of that assemblage. It is intruded at the top by a Josland Lake sill. The predominantly felsic volcanic and lesser subvolcanic and vein-quartz pebbles and cobbles are supported in a fine-grained matrix. The unit was probably deposited by debris flows from a mature volcanic highland.

The Yakymiw Formation (Bailes, 1980a), which outcrops west of Morton Lake, comprises laminated mudstone, siltstone and fine-grained sandstone with interbeds of pebbly greywacke. It overlies, with apparent conformity, the subaqueous volcanic rocks of the Fourmile Island assemblage, both the Storozuk volcanics and the Dickstone rhyolite. It forms the core of an early syncline whose originally recumbent axial surface was intruded by a thick, layered Josland Lake sill (Sections 3.2.1 and 4.1.1). Although much of the bedding is planar or lenticular, beds grading from pebbly to laminated, scour structures and recognizable Bouma divisions indicate deposition by high- to low-density turbidite flows (Figure 9d). Preservation of delicate pumice clasts suggests direct input of neovolcanic detritus from a nearby source. Although overlying different volcanic sequences, the Yakymiw and Parisian formations both represent local volcanic-basin fill.

Metasedimentary rocks converted to biotite±garnet±hornblende gneiss (unit 6b) occur south of Evans Lake in the roof of the Gants Lake Batholith and at the margin of the Ragged Lake pluton in the Sherridon–Hutchinson Lake Complex. They form domains up to 3 km long and 500 m wide at the margins of the plutons. Near Evans Lake, they include metagreywacke, biotite schist (pelite) with common fine-grained felsic beds, and local gritstone. These are commonly interlayered with amphibolite and intruded by tonalite dikes, but appear to have an origin as turbidite, like the Yakymiw Formation. The rocks northwest of the Ragged Lake pluton range from relatively uniform felsic gneiss with biotite+garnet in the north, through garnet-biotite paragneiss locally interlayered with amphibolite, to recognizable metagreywacke in the south.

Biotite±garnet±hornblende gneiss of uncertain age (unit 10a), located on the west side of Yakushavich Island and on Collins Point in northern Kississing Lake, is interpreted as metasedimentary. It occurs in the same structure as felsic gneiss interpreted to include tuff and hosts showings of Fe-sulphides and Zn-Cu±Pb occurrences. A small lens of similar gneiss occurs at the top of the assemblage (unit 7d) of Josland Lake sills and Moody Lake basalt on the east shore of Nokomis Lake.

In general, the coarser grained volcaniclastic rock units in the southeast appear to form a local sedimentary top on well-defined volcanic assemblages. They may be interpreted as part of the clastic fill of local synvolcanic basins. The Yakymiw Formation is the cover of the hangingwall volcanic rocks to the Dickstone VMS deposit. The very coarse, wedge-shaped Parisian Formation may have formed along a synvolcanic fault scarp. The gneissic units on Yakushavich Island may represent a lateral and stratigraphic transition between felsic tuff and sedimentary rocks. The underlying volcanogenic rocks at all of these sites are prospective for VMS deposits (Section 7.1.3.10).

3.2 Early synvolcanic intrusions

The oldest phases of arc intrusions from the northeastern margin the Gants Lake Batholith have been dated as about 1890 Ma old (Whalen, pers. comm., 2007), similar to the ages of synvolcanic tonalite south of Snow Lake (Bailes et al., 1991) and the granitoid gneiss domes north of Snow Lake (Gordon et al., 1990; David et al., 1996). A more clearly defined early age (1894 ±5 Ma, Percival et al., 2006) was obtained from gneissic tonalite in the core of the Reed Lake pluton, which has younger phases that intrude the Northeast Reed assemblage directly south of the File Lake area. Whereas none of the sampled granitoid bodies within the Kississing–File lakes area are known to fall within the early arc and back-arc magmatism, the mafic Josland Lake sills were intruded within its ca. 1886 Ma waning stage.

3.2.1 Josland Lake sills

A suite of mafic layered sills (Josland Lake sills of Bailes, 1980a; unit 7) intrudes the Fourmile Island, Northeast Reed and Snow Lake assemblages (Figure 8c, d, back pocket). Although the sills are best developed and best preserved south and southwest of File Lake, they have been shown to extend northwest to Kisseynew and Kississing lakes for a total strike length of 100 km (Zwanzig, 1994). The sills in the north are more highly metamorphosed but have preserved the same modal layering (Figure 10b; unit 7a) and major compositional change to garnet amphibolite and mottled felsic gneiss derived from highly fractionated ferrogabbro (unit 7b) and Na-rich quartz diorite to trondhjemite (unit 7c; Figure 10c; Photos 45–50, Appendix 4, Table 1, DVD). The sills are locally in fault contact with the Burntwood Group but do not intrude it. Those in the northwest are intruded exclusively into Moody Lake basalt. Sills west of Moody Lake have a thin roof of the basalt, with which they are mapped together as unit 7d.

The sills may be closely related to the Ponton Lake basalt in the Amisk Collage (discussed in Section 5.76). Ferroan quartz diorite of a large differentiated sill south of Josland Lake has yielded a U-Pb zircon age of 1885.8 ±3 Ma (Zwanzig et al., 2001). This is important because the Josland Lake sills provide a reliable maximum age for the numerous units that they intrude. The composition and intrusive habit are similar for the various Josland Lake sills and for the Mikanagan Lake sills, which occur northeast of Flin Flon (Bailes and Syme, 1989). A Mikanagan Lake sill has yielded U-Pb titanite ages of 1881 +3/–2 Ma (Stern et al., 1999), suggesting that the two suites are products of the same regional magmatic event.

Tectonic relations suggest that the Josland Lake sills may have been involved in the opening of a basin between the Fourmile Island and Snow Lake assemblages. The suite of sills extends along a series of structural breaks from Reed Lake to Kississing Lake, straddling the eastern margin of the Amisk Collage, the Northeast Reed (ocean-floor) assemblage, and the western and northern margins of the Snow Lake assemblage. The various sills show complex pre- and postkinematic relationships to the earliest deformation of their volcanic hostrocks (Bailes, 1980a). The dated sill postdates the early deformation and provides a new minimum age for deformation. The distribution and geochemistry of the sills lead to a revised tectonic model for the Kississing–File lakes area. The structural model involves an overturned thrust sheet traced out, in part, by the distribution of the Josland Lake sills (Section 6.1.4).

3.3 *Early successor-arc intrusions (1876–1845 Ma)*

About half of the Kississing–File lakes area is underlain by felsic to intermediate plutonic rocks and lesser mafic intrusions, all equivalent to those assigned to 'successor-arc' plutons in the main part of the Flin Flon Domain (Lucas et al., 1996; Whalen et al., 1999). There are few U-Pb zircon ages for the metaplutonic rock on the South flank of the Kisseynew Domain and northern margin of the Flin Flon Domain, but field relations provide a clear subdivision into two suites of plutons: those that predate deposition of the Burntwood and Missi groups and those that are younger. In this report, these two groups of plutons are designated as early and late successor-arc intrusions, using 1845 Ma as the division. The latter suite is the same as that of Lucas et al. (1996) and Whalen et al. (1999); in this report, the early suite (>1845 Ma) includes the early and middle successor-arc intrusions of Lucas et al. (1996) and Whalen et al. (1999).

The early successor-arc intrusions range from microdiorite, gabbro and quartz diorite dikes (unit 8a) to small bodies with ultramafic phases (units 8b), but are dominated by plutons ranging from foliated granite to gneissic tonalite–quartz diorite (units 9a–h). These intrusions were emplaced almost exclusively into the Amisk Collage southwest of the Loonhead Lake Fault and the West Reed–North Star Shear Zone (Figure 8a, back pocket). The plutons are unconformably overlain by rocks of the Missi Group (Figure 11a). Many reach batholithic dimensions but may be highly flattened; some were folded during penetrative polyphase deformation at mid-crustal levels. Their structural style varies from foliated stocks and batholiths in the south, where primary igneous textures are preserved, to tongue-shaped recumbent fold complexes of orthogneiss in the north, where granoblastic to annealed mylonitic textures and leucosome veins are common (Zwanzig and Schledewitz, 1992). Short petrographic comments with average modal mineral content are summarized from descriptions given in Zwanzig and Schledewitz (1992). Uranium-lead zircon crystallization ages were obtained from five of the early successor-arc plutons in the Kississing–File lakes area, but similar intrusive ages are found also directly to the south.

The western tonalite–granodiorite phase (unit 9d) of the Gants Lake Batholith has yielded 1876 +7/–6 Ma igneous zircons (Whalen and Hunt, 1994). The unit is medium grained and subporphyritic with plagioclase grains (<10 mm) among about 60% total plagioclase in the north and 12% combined amphibole and biotite (modal averages). Another main phase (unit 9a, granodiorite–granite) becomes progressively more strongly foliated in the east. It contains variable amounts of K-feldspar (12% in the north) and biotite as the only mafic mineral. Foliated to gneissic tonalite–granodiorite (unit 9d) forms a domal area at the north end of the batholith. It is medium grained with nearly 60% plagioclase and 9% biotite. Highly foliated and sheared quartz diorite to tonalite (unit 9h) occurs in the northeast end of the Gants Lake Batholith.

Tonalite–granodiorite gneiss (also mapped as unit 9d) in the Moody Lake dome, north of the Gants Lake Batholith, is probably unconformably overlain by the Missi Group, but the contact is intruded by granite. The rocks in the dome are strongly foliated and locally garnetiferous. Potassium-feldspar (~6%) defines a weak gneissic layering. Quartz (29%) and plagioclase (55%) form elongate (<7 mm) aggregates. Similar rock in the Defender Lake dome (Gall, 1981) on Kisseynew Lake is also considered to be an early successor-arc intrusion. A granodiorite phase (unit 9b) in the Moody Lake dome contains up to 30% K-feldspar, 8% biotite and local traces of garnet and sillimanite.

The Sherridon–Hutchinson Lake Complex, which forms the northwestern extension of the Gants Lake Batholith, includes 1873 ±4 Ma, strongly foliated, recrystallized granite (unit 9a, Ragged Lake pluton; Hunt and Zwanzig, 1990) and 1874 ±2 Ma tonalite gneiss (unit 9e, Star Lake pluton; Jungle Lake pluton in Machado et al., 1999). These various units represent highly flattened and folded nested plutons. The southern and eastern margins of the Star Lake pluton comprise strongly foliated to gneissic hornblende granodiorite and quartz diorite (unit 9g). The various plutons and intrusive phases are unconformably overlain by the basal Missi conglomerate (unit 12). The very large Star Lake pluton generally contains an average of 57% plagioclase, 3% K-feldspar and 10% biotite, plus hornblende, but has local variations in mineral content and grain size. The Ragged Lake granite typically has a weak augen texture of K-feldspar (24%) with 7% biotite. Small, elongate patches of massive leucogranite cut the foliation, often in the axial plane of small open folds. The western part of the Sherridon–Hutchinson Lake Complex includes the 1860 ±2 Ma, foliated Archie Lake tonalite–granodiorite (unit 9d) and undated gneissic leucogranite (Jay Lake granite, unit 9b; Hunt and Zwanzig, 1990, 1993; Machado et al., 1999). These bodies show intrusive contacts with only the early assemblages. The Archie Lake pluton contains about 6% each of hornblende and biotite, whereas the Jay Lake granite contains about 5% biotite.

Tonalite and granite plutons southwest of the Gants Lake Batholith have yielded ages of 1864 ±3 Ma and 1845 ±3 Ma (Whalen and Hunt, 1994). The belt that contains these intrusions extends north into the Kississing–File lakes area, where tonalite to granodiorite and quartz diorite (units 9d, g) are undated but intrude only early supracrustal rocks. Structures to the north suggest that this intrusive belt continues with the Archie Lake and Jay Lake plutons. Similarly, the gneissic granodiorite (unit 9c) in the Big Island dome intrudes only its amphibolite

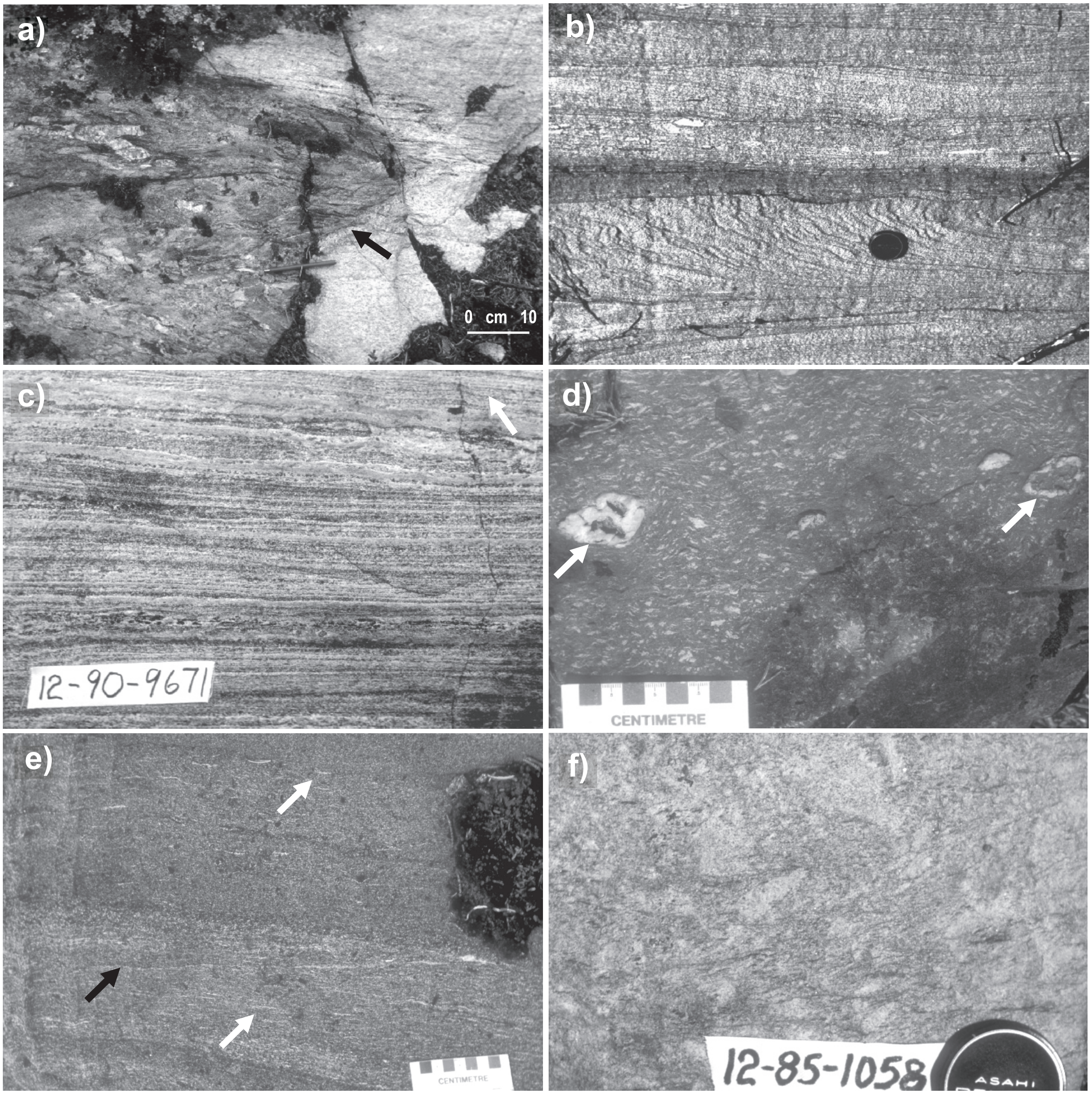


***Figure 11:** Late successor-arc rocks of the Missi Group (see Appendix 4 for these and additional photographs in colour): **a)** Basal conglomerate unconformably overlying the early successor-arc Ragged Lake granite, Puffy Lake; unconformity (arrow) is folded. **b)** Crossbedded meta-arenite and pebbly arenite (flatted pebbles, white) of lower fluvial-alluvial sequence north of Kisseynew Lake; foreset lamination at lens cap (5 cm) is oversteepened and folded by oblique compression. **c)** Thin-bedded to laminated, intermediate meta-arenite/siltstone with metamorphic amphibole in upper part of lower sequence, Evans Lake area; tape is 10 cm long. **d)** Plagioclase-phyric massive flow with some exceptionally large amygdules (arrows), indicating subaerial extrusion. **e)** Layering and flattened felsic fragments (white arrows) in Jungle Lake basaltic andesite, suggesting an origin as bedded tuff or reworked tuff, Puffy Lake; note that flattening is oblique to bedding and parallel to foliation and incipient veins (black arrow). **f)** Angular fragments in felsic breccia or monomictic conglomerate derived from Cleunion rhyolite, Puffy Lake; tape is 10 cm long.*

mantle (unit 4), which is overlain by basal Missi conglomerate (Zwanzig and Schledewitz, 1992).

The Batty Lake Complex comprises quartz-rich trondhjemite gneiss (unit 9f) that grades into a tectonite mantle and contains large supracrustal screens within. This orthogneiss has yielded 1859 +14/–7 Ma stubby igneous zircons and 1810 ±8 Ma elongate zircons (Hunt and Zwanzig 1993) that probably dated, respectively, the main intrusive phase and the diffuse leucosome (Photo 44, Appendix 4, Table 1, DVD). The main phase has an average mode of ~40% quartz and 50% sodic plagioclase, with the rest being intergrown amphibole and biotite, minor garnet and magnetite, and interstitial K-feldspar. The mafic minerals occur as elongate trains that help to define the foliation. Quartz and plagioclase form thin polygonized grains that are partly annealed to a fine granoblastic texture. Large garnet or magnetite porphyroblasts occur locally in patches of leucosome. The texture is the result of metamorphism, deformation and partial melting, but the uniformity of the rock indicates its igneous protolith, and the size of the mafic, quartz and plagioclase aggregates (interpreted as polygonized grains) points to an intrusive origin.

Unit 9f in the Defender Lake dome has a similar composition but more variable texture and structure, such as layering. It probably includes supracrustal rocks. It is noteworthy that, on the northern margin of the Flin Flon Domain, these intrusive complexes form gneiss domes or tongue-shaped recumbent structures with mantles of early assemblages or nearly coeval cover associated with the 1850–1855 Ma Sherridon–Meat Lake assemblage. The margin of the Defender Lake dome has a number of sulphide showings, and the margin of the Batty Lake Complex has local mineral assemblages that are the result of hydrothermal alteration, as in the Sherridon–Meat Lake assemblage.

3.4 Rocks of uncertain age and affinity

3.4.1 Bob gneiss

The core zone of the Sherridon structure is partly occupied by garnet-biotite±hornblende gneiss (unit 10a, Bob gneiss) that is described by Tinkham and Karlapalem (2008) as

…medium to coarse grained, mesocratic to melanocratic, black to medium grey and characterized by subhedral, 2–5 mm, homogeneously distributed red garnets. Compositional layering is moderate to weak, defined by varying biotite and hornblende content. Graphite is locally abundant (up to 3%)…near the contact with the graphitic biotite quartzofeldspathic gneiss unit. The contact with this unit is gradational….

[The graphitic gneiss is] leucocratic to mesocratic, white to buff to grey-green, commonly oxidized and showing a red to orange weathering colour, and dominantly medium grained (locally coarse to very coarse). Amphibole is locally present, and local sulphide concentrations up to 3% [are observed] near contacts with [mafic to calcsilicate rocks that adjoin felsic Sherridon gneiss].

Although the preferred origin of the Bob gneiss of Tinkham (pers. comm., 2008) is igneous because of its uniformity, it has a greywacke composition (Froese and Goetz, 1981). Its weathering colours are also typical of parts of the Burntwood Group. Also, detrital zircon ages are similar to those of the Burntwood Group (N. Rayner and D. Tinkham, pers. comm., 2009).

More prominently layered, migmatized, garnet-hornblende-biotite gneiss on Yakushavich Island and Collins Point on Kississing Lake, also mapped as unit 10a, may be as young as or only slightly older than the Burntwood Group, which locally has a similar composition. This occurrence of the unit also has an affinity to Sherridon gneiss (see also Section 3.1.4). The relationship between these rocks may be resolved with more detailed mapping and more detrital zircon ages. One model can be suggested: unit 10a originally lay stratigraphically above the Sherridon gneiss and stratigraphically below the Burntwood Group. The entire succession may have formed at the margin of the Sherridon-age successor-arc terrane and extended into the early Kisseynew basin as a resedimented volcaniclastic deposit. This interpretation is consistent with the close similarity between units 10a and 11c, the latter clearly part of the Burntwood Group. The gradation between the Sherridon gneiss and the Burntwood Group on the north side of the Walton Lake nappe may represent a preserved stratigraphic relationship that is obscured by thrusting, early in the history of the Fairwind Lake Fault, of the felsic volcanic assemblage over the Burntwood Group in the early Kisseynew basin.

3.4.2 Tectonite

A prominent unit of felsic and mafic tectonite (unit 10b), forming the West Reed–North Star Shear Zone (WRNSZ; Syme et al., 1995), extends for a strike length of 50 km along the west side of the Fourmile Island assemblage. The belt is 1 km wide at North Star Lake, widens to 5 km south of the Kississing–File lakes area and tapers toward the north where it merges with the younger Loonhead Lake Fault east of Dow Lake. A splay of the shear zone re-emerges west of Dow Lake. The tectonite includes felsic rocks locally recognizable as rhyolite (Norquay, 1997), but thin felsic layers have been interpreted as sheared synkinematic dikes by Syme et al. (1995). The mafic rocks consist of schist in the southeast, but flows and breccia are recognized on the west side of the WRNSZ, along with layered amphibolite and mafic schist (Norquay, 1997). Mylonitic fabric is annealed and the tectonite is folded at all scales in the Dow Lake area, where metamorphic grade is higher. Only the strongly layered characteristic of fine-grained straight gneiss is preserved in the north. A lens of biotite±garnet schist, similar to the Burntwood Group or older metagreywacke, occurs south of Dow Lake. Ultramafic lenses (described above) occur on the west side of the tectonite in the north. Strands of the WRNSZ surround the Reed Lake mafic–ultramafic complex west of Reed Lake from, which the ultramafic lenses were derived.

A long and narrow belt of felsic to intermediate gneiss (unit 10b) forms part of a shear zone at the northeastern contact of the Batty Lake Complex. This zone contains the local garnet-cordierite-gedrite gneiss that was probably derived from hydrothermal alteration.

3.4.3 Interlayered felsic and mafic rocks

Local areas of felsic and mafic gneiss, without strong fabric but of uncertain origin (unit 10c), occur on the west side of the WRNSZ. Fine-grained layered amphibolite and intercalated fine-grained rock of rhyolitic or granitic composition occur south of North Star Lake. Four to six kilometres north of North Star Lake, felsic to intermediate fragments that may be clasts or tectonic pseudoclasts occur in a mafic matrix.

3.4.4 Felsic rocks

Generally uniform, fine- to medium-grained felsic rock with biotite±garnet and rare magnetite (unit 10d) occurs in the core of the Dow Lake dome. It has a tonalitic to granodioritic composition but is strongly foliated, and igneous texture has generally not been preserved, except for local K-feldspar porphyroclasts that suggest an intrusive origin for unit 10d.

Another intrusive rock of uncertain age and affinity is quartz-plagioclase–phyric tonalite of unit 10e. This rock forms small high-level intrusions in the Fussey Lake basalt south of File Lake (Bailes, 1980a). An early age is suggested by the presence of crosscutting mafic dikes that may belong to the Josland Lake gabbro.

3.5 Late successor-arc and -basin assemblages (1.85–1.83 Ga)

The younger assemblages have three important components: 1) the Burntwood Group metagreywacke, 2) the Missi Group fluvial-alluvial to lacustrine arenite with intercalated subaerial volcanic rocks, and 3) the late successor-arc intrusions. The three components are partly coeval, the Burntwood Group likely being the distal marine facies of the Missi Group (Syme et al., 1995; Connors, 1996; Zwanzig, 1999) and some late successor-arc plutons likely being coeval with Missi Group volcanic rocks (Ansdell, 1993). These late successor-arc assemblages occur in the northwestern, northern and southeastern parts of the Kississing–File lakes area, where they lie either near the northern and eastern limit of the Amisk Collage or structurally below the Northeast Reed (ocean-floor) assemblage and Sherridon–Meat Lake assemblage (Table 4). The southern exposures of Missi Group rocks in the Kississing–File lakes area form the core of a long, tight syncline extending from Martell Lake to Weldon Bay (Zwanzig, 1999).

Where depositional contact relationships are known, the Burntwood Group turbidite forms the lower part of the late basinal deposits. The Missi Group is interpreted as a syntectonic overlap assemblage shed over the margin of the early volcanic to successor-arc massif and prograded over the Burntwood Group. Detrital zircon data indicate that the upper distal sequences of the Missi Group shallow-water and terrestrial deposits are only slightly younger than the upper part of the Burntwood Group, but the latter is partly coeval with considerably older, lower and more proximal sequences in the Missi Group. Onset of turbidite sedimentation may have been ca. 1855 Ma (David et al., 1996) or even earlier, as suggested by the 1855 Ma Marie Lake pluton that appears to intrude the Burntwood Group in Saskatchewan (NATMAP Shield Margin Project Working Group, 1998). The youngest age from zircon collected in the proximal Missi Group at Flin Flon is 1847 ±2 Ma (Ansdell, 1993) and, higher in the section at Puffy Lake, it is 1837 ±4 Ma (Machado et al., 1999). The youngest Missi-age intrusive to extrusive rock is 1832 ±2 Ma from the Nelson Bay dome (David et al., 1996). The youngest grains in the Burntwood Group are nearly euhedral, devoid of percussion marks and 1842 ±2 Ma old (Machado et al., 1999), consistent with the conclusion that the base of

Table 4: Field characteristics of late successor-arc and -basin assemblages in the Kississing–File lakes area.

Thickness (m)	Unit	Name	↑Upper contact ↓Lower contact	Lithology
MISSI GROUP (ca. 1850–1830 Ma)				
	14	Missi Group volcanic rocks		Metabasalt to rhyolite
<300	14a	Kississing River basalt	↑Conformable	Massive mafic flows and sills
<60	14b	Evans Lake basalt	↑Conformable	Massive mafic flows
2.5-50	14c	Jungle Lake basaltic andesite	↑Conformable	Intermediate gneiss (mafic and plagioclase phenocryst pseudomorphs)
	14d	Puffy Lake dacite	↑Conformable	Felsic ribbon gneiss (breccia and tuff)
<400	14e	Cleunion rhyolite	↑Conformable	Felsic gneiss (ash flows or reworked tuff)
<200	14f	Lobstick rhyolite	↑Conformable	Felsic gneiss (ash flows or reworked tuff)
1200	13	Missi Group sedimentary rocks	↑Top unexposed,	Arenite/paragneiss +magnetite
	13a	Arkose, derived paragneiss	possible internal faults and unconformities	Unnamed metasedimentary facies of 13:
	13b	Lithic arenite, derived gneiss		a) locally crossbedded meta-arkose
	13c	Intermediate composition arenite to pelite	↓ Base conformable with Burntwood Group, unconformable with Amisk Collage, structural in Snow Lake area	b) biotite-bearing meta-arenite c) hornblende-bearing meta-arenite
	13d	Dacititic to polymictic conglomerate		d) felsic to intermediate fragmental
	13e	Felsic tuffaceous metasediment		e) fine-grained leucrocratic, uniform
	13f	Siltstone to pelite		f) thin layered, common epidote
				(13 a–f locally in stratigraphic order)
5–20	12	Basal polymictic conglomerate		Intermediate–felsic upward
	12a	Conglomerate and sandstone	(close to base)	Mainly felsic clasts
	12b	Protoquartzite	(local basal unit)	Weakly feldspathic
BURNTWOOD GROUP (ca. 1855–1840 Ma sedimentation, ca. 1840 Ma amalgamation)				
0-10	11e	Burntwood gradation to Missi Group	Gradational–disconformable	Interlayered quartzite and pelite
>50	11g	Corley Lake (member of File Lake Formation of Bailes, 1980)	Conformable with greywacke above and below	Metamudstone
>700-1000	11a–f	File Lake Formation of Bailes (1980),	↑(Disconformable ↓Unexposed base	Metagreywacke-mudstone (turbidite), garnet-biotite gneiss, migmatite

the Missi Group is diachronous: an unconformity in the south but a prograding sedimentary facies boundary at the margin of the Kisseynew Domain. The Burntwood Group is therefore interpreted as the deep-water marine facies of the prograding Missi Group (Bailes, 1971; Syme et al., 1995; Connors, 1996; Zwanzig, 1999). These relations suggest that the base of the Missi Group and the occurrence of Burntwood Group can be used to reconstruct the depositional settings and paleogeography of the Kisseynew basin margin. This reconstruction also allows an assessment of the original geometry and vergence of the earliest structures that predate the Burntwood Group. These tectonostratigraphic relations are illustrated in Figure 7b and discussed more fully in Section 6.1.61.

The Missi Group is characterized by a prominent basal conglomerate that unconformably overlies the adjacent Amisk Collage and early successor-arc plutons (Figure 8b, back pocket). Missi Group overlying the Fourmile Island assemblage at the northeastern margin of the Amisk Collage contains thin basal conglomerate units generally interlayered with arenite and gritstone, and protoquartzite beds. The latter may represent a more distal facies, possibly deposited at a basin margin (Figure 8d; Zwanzig and Schledewitz, 1992). Missi Group meta-arenite exposed in higher structural levels comprises the most distal, finest grained facies and lies either disconformably on the Burntwood Group or grades conformably downward into it (Figure 8e; Zwanzig, 1999). Stratigraphy within the Missi Group in the Kississing–File lakes area is poorly understood because of the high metamorphic grade, complex structure, and facies and thickness changes among sub-basins that were probably fault bounded. There appear to be up to three fining-upward sequences, generally suggested by an increase in mafic mineral content related to increased clay fraction. The upward fining is confirmed locally by the presence of observed facing directions in coarse, crossbedded or conglomeratic bases, and thin-bedded to laminated sequence tops. Where the rocks are coarsely recrystallized and gneissic, however, such details are cryptic.

The presence in the Missi Group of mafic to intermediate flows and felsic tuff, as documented throughout the Kississing–File lakes area, indicates that sedimentation occurred during the last stage of early successor-arc magmatism and peaked during the late successor-arc magmatism and the onset of continental collision. This relationship is supported by the detrital zircon data and the similar trace-element contents of the Burntwood and Missi groups when compared to each other and the various felsic volcanic rocks and intrusions. All Missi Group volcanic rocks are subaerial, as indicated by massive mafic flows with contacts that contain large amygdules, intermediate tuff that is interlayered with fluvial beds, and felsic rocks that are locally interpreted as ignimbrite. The subaerial origin is in distinct contrast to the dominantly subaqueous deposition of the early volcanic assemblages. A suite of Missi-age to slightly younger plutons intrudes the basinal sedimentary rocks at the margin of the Flin Flon Domain, locally cutting structures interpreted as syncollisional (Section 4.1.3).

The successor-arc subaerial volcanism started with the onset of Missi sedimentation: a bed interpreted as tuff or reworked tuff within the basal conglomerate near Puffy Lake yielded zircons with a discordia age of 1848 ±4 Ma from four single-grain ID-TIMS analyses of a population of clear, sharply euhedral and subequant zircons (Machado et al., 1999). If this represents the age of crystallization of the tuff, as suggested, it provides an age for the start of sedimentation (proximal facies) where the Missi Group lies directly on the Amisk Collage. The peak of late successor-arc magmatism occurred during the advanced stage of sedimentation, as indicated by an 1836 ±2 Ma rhyolite extrusion at Wekusko Lake (Ansdell et al., 1999) and an 1832 ±2 Ma subvolcanic intrusion (Nelson Bay dome in this report) near File Lake (David et al., 1996). The volcanic rocks are restricted to the Missi Group sedimentary rocks in the Flin Flon–Kisseynew transition. Only one or two possible beds of reworked tuff have been observed in the Burntwood Group. However, the Burntwood Group contains abundant, little-abraded neovolcanic detritus (Bailes, 1980b), including the 1842 Ma zircon grains (Machado et al., 1999). This indicates that the Burntwood Group is a marginal basin deposit. Its origin in the Kisseynew basin in a possible fore-arc, back-arc or arc-rift environment is addressed in Section 6.1.6.1).

3.5.1 Basinal sedimentary and volcanic rocks

3.5.1.1 Burntwood Group

The Burntwood Group forms a monotonous succession of graphitic metagreywacke-mudstone (unit 11a) and derived garnet-biotite gneiss (unit 11b) and migmatite (unit 11c). It is the most widespread deposit in the Kisseynew Domain and extends from the Flin Flon Domain north to the Lynn Lake Belt, east to the TNB and northwest to the La Ronge Belt in Saskatchewan (Figure 1).

Much of the Burntwood Group in the Kississing–File lakes area has been interpreted to be in structural contact with other units, which include the Fourmile Island, Northeast Reed, Snow Lake and Sherridon–Meat Lake assemblages, the Josland Lake gabbro and the Missi Group (Figure 8, back pocket; NATMAP Shield Margin Project Working Group, 1998). However, there is local evidence that the Burntwood Group may have been deposited directly on parts of the early volcanic assemblages after a long hiatus. It may also lie conformably on parts of the early successor-arc volcanic rocks (Sherridon–Meat Lake assemblage). It is overlain by the Missi Group with a low-angle unconformity, a disconformity and/or a gradation. It is intruded by the late successor-arc plutons and bodies of synkinematic leucogranite to tonalite and quartz diorite. A structurally restored section where the base is not exposed suggests that the (structurally flattened) sequence of unfolded turbidite is presently ≥700 m thick at File Lake, although Bailes (1980a) estimated a pre-fold thickness of 1050 m.

A thin unit of typical Burntwood Group greywacke-mudstone overlies the Ponton Lake basalt in the Amisk Collage northwest of Puffy Lake (e.g., on strike with column 2B in Figure 8b). It grades upward into the regional basal Missi conglomerate (unit 12) through interbedded greywacke and conglomerate. The greywacke is traceable for nearly 6 km to the east, but quartzite occurs at the same stratigraphic position 5 km to the southwest. Although the contact between the greywacke and the volcanic rock is not exposed and may be a fault, it can also be interpreted as an unconformity, thus making the basalt the possible local basement to the Burntwood Group (Zwanzig, 1999).

A thin unit of amphibole-bearing metagreywacke occurs between the Moody Lake basalt and the basal Missi conglomerate, tuff and gritstone sequence on Nokomis Lake (column 3 in Figure 8b). This section is also sheared, but well-preserved conglomerate and tuff 1.5 km to the southwest lie directly on Moody Lake basalt and Josland Lake gabbro, so the shearing must be local. This gabbro-basalt-conglomerate-tuff sequence extends at least 10 km southwest to Puffy Lake and must contain the major unconformity at the base of the Missi Group. The several lenses of metagreywacke at this autochthonous Missi unconformity suggest that a unit of the Burntwood Group may have covered parts of the early arc massif at the margin of the Kisseynew basin. A similar lens, interpreted as allochthonous, between ocean-floor basalt and the Missi Group occurs along the Roberts Lake Fault, northeast of Wekusko Lake (NATMAP Shield Margin Project Working Group, 1998).

Better evidence for a depositional base of the Burntwood-type (1860–1856 Ma) metaturbidite occurs north of Flin Flon in the Trout Lake succession (Rayner, pers. comm., 2009). Graded bedding provides sedimentary tops for the Burntwood Group lying next to the early volcanic assemblages in the well-preserved rocks at File Lake, Morton Lake and Woosey Lake. At all of these sites, facing is away from the older rocks, consistent with a depositional contact, but this contact has been mapped as a fault because it generally represents a prominent shear zone. Consequently, the contact has been widely mapped as a major fault (Zwanzig and Schledewitz, 1992; Connors, 1996; Connors et al., 1999; Kraus and Williams, 1999; Zwanzig, 1999). Bailes (1980a), without knowledge of isotopic ages, indicated a conformable depositional contact. Structural considerations (Section 4.1.3) suggest a complex tectonic history for the contact.

The contact between the hornblende-bearing Burntwood Group gneiss (unit 11d) and the Sherridon–Meat Lake assemblage is also mapped as a fault (Fairwind Lake Fault, column 5 in Figure 8a, e) and seems to cut off units in the Walton Lake nappe. There are two areas, however, where a gradation may exist between felsic gneiss and the Burntwood Group. One site is the northeastern margin of the Walton Lake nappe, where the felsic 'volcanic' gneiss cannot be distinguished from the garnet-biotite gneiss of the Burntwood Group. The other area is on Yakushavich Island and Collins Point in Kississing Lake, where the apparent gradation is between felsic gneiss (units 5, 5d), garnet-biotite±hornblende±graphite gneiss with Fe-sulphides (unit 10a), and Burntwood Group migmatite (unit 11c). It is noteworthy that unit 11c, also comprising garnet-biotite±hornblende gneiss with graphite and Fe-sulphides, occurs at the east side of the Walton Lake nappe, locally adjoining felsic gneiss with a margin of rock containing biotite, cordierite and garnet, as on Yakushavich Island. The single zircon age of about 1850-1855 Ma from the Sherridon gneiss is consistent with a successor-arc age for all of these rocks.

South of File Lake, where greenschist-facies metamorphism prevails, the Burntwood Group (File Lake Formation of Bailes, 1980b) has well-preserved primary structures, including graded bedding, parallel and ripple laminations, convolute bedding and sole markings, that indicate its origin as a turbidite deposit. These are organized into turbidite A–E divisions (Bouma, 1962) that suggest a transition from an upper to a lower fan slope (Figure 10d, e), commonly with a prominent graded A division of high-density flows (Bailes, 1980b). With increasing metamorphic grade north across File Lake, turbidite was transformed to paragneiss that contains 10–15% garnet porphyroblasts up to 5 mm in diameter and some sillimanite-enriched knots (faserkiesel, containing microscopic fibrolite). Locally preserved rhythmic layering, alternating between quartz-plagioclase–rich gneiss and more garnet-biotite–rich gneiss, is interpreted to be derived from the graded greywacke-mudstone beds seen in the south. Locally occurring calcsilicate lenses, up to 5 cm thick, were probably derived from calcareous concretions. To the north, in the main part of the Kisseynew Domain, the Burntwood Group consists of migmatite with 10–75% granitoid leucosome in the form of sheets, veins and lenses of various relative ages (Zwanzig, 1999). The oldest high-grade leucosome (LS1) typically contains biotite, garnet and sillimanite±cordierite, which are also concentrated in selvages (melanosome). In the Northeast Kisseynew subdomain, there is leucosome with local hypersthene (Growdon et al., 2006). The relicts of the compositionally graded bedding are locally preserved in the northern part of the Kississing–File lakes area. Photographs of sedimentary facies of the Burntwood Group and its gneissic to migmatitic derivatives are available in Appendix 4, Table 1 (Photos 35–41, DVD).

A unit of Fe-rich metamudstone (Corley Lake mudstone in this report; Corley Lake member of Bailes, 1980a) occurs near the top of the section on both limbs of an early major syncline on File Lake (columns 9D and 9E in Figure 8e). The unit is up to 50 m thick and remarkably uniform, characterized by prominent euhedral garnet and staurolite porphyroblasts. Rock similar to the Corley Lake mudstone is reported from 20 and 60 km east of File Lake (Bailes, 1975, 1980a). The rock may be an overbank deposit or, more likely because of its wide extent, a distal basinal or hemipelagic deposit, not necessarily a turbidite but possibly deposited in a filled trench (Section 6.1.6.1). Its consistent position near the top of the group suggests that the overlying Missi Group has, at best, a low-angle unconformity and is not in fault contact everywhere, as postulated by the NATMAP Shield Margin Project Working Group (1998). The Walton Lake amphibolite (unit 4d) south of Walton Lake is structurally underlain by amphibole-garnet–bearing metagreywacke (unit 11d) with abundant graphite and scattered Fe-sulphides. An increase in graphite and cordierite, and the local presence of amphibole in the greywacke-gneiss near the Loonhead Lake Fault suggest that this thrust fault (and possibly the Fairwind Lake Fault) may be flats in the upper part of the Burntwood Group. Amphibole-bearing paragneiss occurs near the known top of the Burntwood Group at Puffy Lake, but no definitive stratigraphic position has been mapped for the amphibole-bearing rock. Nevertheless, the common upward increase in graphite in the gneiss suggests a fining upward (increase in graphitic pelite) in the part of the succession that directly underlies the Missi Group.

There is an important facies change in the Burntwood Group from coarser, locally thick (2 m) pebbly beds, commonly high-energy A–E beds, on Reed Lake (Syme et al., 1995) through beds showing most Bouma divisions on Morton Lake and into File Lake (Bailes, 1980a, b). At Morton Lake, there is a decrease in the greywacke:mudstone ratio and bed thickness from south to north and top to bottom of the exposed section. On the southwestern shore of File Lake, however, the greywacke: mudstone ratio and bed thickness decrease upward (Bailes, 1980a). This represents downslope changes and migration of upper to lower fan facies. The monotonous layering in the north may represent the basin plane (Figure 10e), but most turbidite structures there have been destroyed by deformation and high-grade metamorphism. The widespread Burntwood Group in the Kisseynew Domain was interpreted to have been deposited in coalescing submarine fans (Bailes, 1980b). The fault slice at Morton Lake may recur at the same structural level on the north shore of Martell Lake, where there is a gradation from uniform garnet-biotite gneiss (mudstone-greywacke) to Missi Group arenite. These relations suggest that a turbidite fan may have extended from Reed Lake into an open basin or trench to the north, and that the Missi Group continental deposits have prograded over the turbidite at the margin of the Kisseynew basin. This upper fan facies of the Burntwood Group contains volcanic pebbles and is best interpreted as para-autochthonous with the adjoining volcanic units.

Northeast of the Loonhead Lake Fault, the Burntwood Group locally grades upward into a unit of interlayered quartzite and pelite that is overlain by the Missi Group (unit 11e; Figure 10f) This unit is interpreted as a prograding shoreline deposit (Zwanzig, 1999). A thin bed of felsic tuff lies in the upper metre of Burntwood Group pelite within this transition at Puffy Lake (column 2E in Figure 8e). These beds are followed by thin chert, interbedded tuff and arenite (Appendix 4, Table 1, Photo 40, DVD), and, 25 m up the section, by a thin quartz-pebble conglomerate. Chert and tuff may represent a depositional hiatus or disconformity followed by the coarsening-upward succession, which is interpreted to represent the basinward prograding of the nonmarine facies (Missi Group). Although this type of sequence has been traditionally included entirely with the Missi Group (e.g., column 4c in Figure 8e; Bailes, 1980a; Zwanzig, 1999), it has been interpreted as the coarsening-upward top of the Burntwood Group in a well-exposed fan-delta sequence on the North flank of Kisseynew Domain (Zwanzig, 1990). Consequently, the interlayered quartzite and pelite (unit 11e) are herein considered as the shallow-water marine facies that forms the top of the Burntwood Group.

3.5.1.2 Missi Group

The Kississing–File lakes area contains only medium- to high-grade metamorphic rocks of the Missi Group, typically comprising magnetite-bearing quartzofeldspathic gneiss in which pebbles and crossbedding with magnetite-rich placers are locally preserved (*unit 13a*[4]). The protoliths were probably hematite-bearing sandstone (redbeds), siltstone and conglomerate, interpreted to have been deposited in fluvial-alluvial environments, as determined at Flin Flon (Bailes and Syme, 1989; Stauffer, 1990). A 5–20 m thick, deformed conglomerate (unit 12) commonly occurs at the unconformable base of the Missi Group, overlying the Amisk Collage and early successor-arc plutons (Figure 8b, back pocket; Figure 11a; Appendix 4, Table 1, Photos 29–32, DVD). This unit has been traced from Kississing Lake to File Lake, over an estimated area of >600 km^2. A thinner basal conglomerate or ribbon gneiss (highly deformed conglomerate) with interbedded sandstone-derived gneiss (unit 12a) is in contact with the Fourmile Island assemblage throughout the Loonhead Lake thrust sheet, over an estimated area of 1000 km^2 (Zwanzig, 1999). Protoquartzite and thin conglomerate (unit 12b) occur locally at the base of the Missi Group. Elsewhere, Missi Group meta-arenite overlies quartzite and pelite (unit 11e) of the Burntwood Group (column 4D in Figure 8e). Mafic and felsic volcanic rocks and subvolcanic intrusions are widely intercalated with the arenite, generally in thin units recognized in low-strain areas as tuff or massive flows with coarsely amygdaloidal margins. Because of the high metamorphic grade, the protoliths in many places were deduced from composition, transposed bedding or flow forms, and locally preserved sedimentary or volcanic structures.

The thickest, most proximal part of the Missi Group occurs in folded fault blocks north of the Amisk Collage on the north shore of Kisseynew Lake. A tentatively restored stratigraphic section suggests that more than 600 m of crossbedded arkose, pebbly arkose or lithic arenite and matrix-supported conglomerate (*unit 13a*; Figure 11b) are overlain by a fining-upward, metamorphic hornblende-bearing sandstone–siltstone (*unit 13c*) succession (Figure 11c) with more than 300 m of basalt (unit 14a, Kississing River basalt) and related mafic sills (columns 1A and 1A' in Figure 8b). Another 600 m thick fault-bounded section on Weldon Bay is interpreted to represent a stratigraphically higher sequence with local felsic conglomerate (*unit 13d*), and felsic tuffaceous arenite to siltstone with local felsic pebbles (*unit 13e*) and with a thick unit of felsic tuff (Lobstick rhyolite, unit 14f; column 1B in Figure 8b). The unconformable base of this section has the basal conglomerate exposed on Ponton Lake; the conglomerate extends east to Martell Lake (columns 2B–8A, Figure 8). Part of the section is interpreted as alluvial, probably deposited during synsedimentary normal faulting, such as in the hangingwall of what is presently the Defender Lake dome (Figure 8a; Zwanzig and Schledewitz, 1992). This part of the Missi Group is also similar to that found in fault-bounded blocks northeast of Wekusko Lake (Gordon and Lemkow, 1986) and at Flin Flon (Bailes and Syme, 1989). Amalgamation of several sequences appears to occur across Weldon Bay, similar to that observed in more modern terranes adjacent to synsedimentary faults. This and other regional variations and prominent local facies changes are attributed to sedimentation in a tectonically and volcanically active area. The presence of various blocks with different thicknesses of sedimentary rocks and abrupt changes in the thickness, composition and proportion of volcanic rocks indicates block faulting in an extensional environment.

[4] Units shown in *italics* do not appear on Map GR2010-1-1 in back pocket. They are interpreted to represent sedimentary facies but are not mappable as stratigraphic units. They are combined on the map, together with thin local volcanic horizons, as unit 13.

Possible sequence stratigraphy

A section was restored from the complex folds between Lobstick Narrows and Cleunion Lake. This section is interpreted to occupy the core of the Star Lake syncline (Zwanzig, 1999) and represents the highest stratigraphic level preserved in the Loonhead Lake thrust sheet (Figure 7b). The section is estimated to be 1600 m thick, but tectonically flattened pebbles suggest it may have originally been more than 5000 m thick. West of Cleunion Lake, where dips are more moderate, the total true present thickness is 1400 m, but units are structurally more attenuated. The aggregate thickness estimate for the sequence may be excessive because it is derived from more than one fault block. Nevertheless, the section is tentatively divided into three different fining-upward sequences (columns 1C–1C″ in Figure 8d, back pocket). The lower two of these correlate with column 1A–1B and the upper one may grade upward into calcareous (now mainly epidote-bearing) alluvial-plain or playa-lake deposits (*unit 13f*). Photographs of the various sedimentary facies and interlayered volcanic rocks are available in Appendix 4, Table 1 (Photos 8–34, DVD).

About 60 m of interlayered basal conglomerate and protoquartzite (unit 12) overlie amphibolite derived from basalt and Josland Lake gabbro at the base of the first sequence (column 1C in Figure 8d). About 300 m of overlying arkose with crossbedded to pebbly sections (*unit 13b*) locally contain a 7 m member of pink rhyolite grit interpreted as reworked tuff. The arkose is capped by an 80 m thick succession of amphibolite derived from aphyric to plagioclase-phyric basalt (unit 14a, Kississing River basalt) to porphyritic andesite (unit 14c, Jungle Lake basaltic andesite). Flows are massive (nowhere pillowed) and have locally preserved contacts with calcsilicate alteration and amygdules up to 30 mm (Figure 11d). These features indicate that the flows are subaerial. Basaltic andesite has only locally preserved primary features that include a variety of phenocryst populations and large amygdules. Layering suggests the presence of bedded or reworked tuff (Figure 11e). Basaltic andesite is interlayered at the top with dacite (unit 14d, Puffy Lake dacite) at one locality north of Weldon Bay. About 2 m of pink gneiss, interpreted as rhyolitic ignimbrite (unit 14e, Cleunion rhyolite), joins the volcanic unit farther northeast. The top of the lower sequence is 50–100 m of thin-bedded calcareous siltstone/sandstone *(unit 13f)* with amphibole and epidote as characteristic metamorphic minerals. This unit also contains a thin rhyolite member.

The second (probably overlying) sequence has a crossbedded arkose at the base, overlain by 150 m of felsic to lithic tuff of unit 14e (Cleunion-type rhyolite; column 1C’ in Figure 8d). The sequence has a local lens of fine-grained granite- to rhyolite-cobble conglomerate and 200 m of fine-grained felsic sandstone–siltstone (*unit 13e*) overlain by pink to brick red felsic gneiss (unit 14f, Lobstick rhyolite). The fine-grained felsic sandstone in sequence 2 and in the lower part of sequence 3 is very uniform and rhyolitic in composition. It probably contains a large proportion of rhyolite tuff reworked from the adjacent felsic complexes. The main felsic bodies are also locally uniform and are likely subvolcanic intrusions in part, but are interpreted to contain aprons of breccia, conglomerate and mass-flow deposits, as suggested from structures preserved in the upper greenschist– to amphibolite-facies metamorphic rocks southwest of Cleunion Lake. In felsic fragmental rocks that are highly flattened or coarsely recrystallized, breccia cannot generally be distinguished from monomictic conglomerate except on a few exposures (Figure 11f).

The uppermost sequence of the Missi Group at Cleunion Lake comprises 80 m of crossbedded sandstone (*unit 13b*) and 300 m of thin-layered, calcareous to argillaceous metasedimentary gneiss, typically containing epidote and, very locally, layers with corundum interpreted as Al-rich argillite (column 1C’ in Figure 8d).

The two lower sequences appear to be basin wide and probably represent tectonic control during syndepositional deformation. The overall upward and northerly fining of this and other sections provides a regional direction of sediment transport and dispersal direction of reworked tuff toward the Kisseynew basin, probably from an intermittently uplifted volcanic terrain.

Central restored tectonostratigraphic sections

Tectonostratigraphic sections within the Loonhead Lake thrust sheet, overlying the Sherridon–Hutchinson Lake dome, contain much thinner sequences of only about 80–150 m, but show similarities to sequences 1 and 2 (compare column 2D and columns 1C–1C″ in Figure 8d, back pocket). The sections were reconstructed from rocks farther removed from the large felsic centres that must have existed near Cleunion Lake. Their aggregate volcanic thickness is generally less than 50 m, and various thinner volcanic units (Puffy Lake dacite, Jungle Lake andesite and Evans Lake basalt) occur in different parts of these sections. The similar abbreviated section directly overlying the early volcanic and intrusive rocks of the Sherridon–Hutchinson Lake Complex at Jungle Lake (column 4A in Figure 8b) suggests that the sequences in the Loonhead Lake thrust sheet were deposited in the same basin as those resting on the main part of the Amisk Collage. It should be noted that the high flattening strain indicated in the northern gneissic units suggests that their present thicknesses should be multiplied by four or five to obtain approximate primary thicknesses.

A section of the Missi Group north of File Lake that is more than 800 m thick (estimated as 1300 m thick by Bailes, 1980a) disconformably overlies the Burntwood Group in the large early syncline that extends from File Lake for 60 km northwest to Jungle Lake (columns 4D and 9D in Figure 8e). The section has a base of thick-bedded metamorphosed subgreywacke or lithic arenite containing scattered pebbles and local porphyroblasts of garnet±staurolite or aggregates of fibrolitic sillimanite (faserkiesel). Ripple current laminations are preserved locally. The main part of the group is thick-bedded meta-arenite rich in plagioclase. The section includes 200 m of amphibolite in the middle, interpreted as metagabbro (Bailes, 1980a), and a thin felsic volcanic unit in the upper part. This is a distal part of the Missi Group (Zwanzig, 1999). About 40–50 m of amphibolite occurs at Jungle Lake in a structurally abbreviated section (column 4C) that contains units similar to those at File Lake. The amphibolite at Jungle Lake is clearly exposed as a series of thin mafic flows with amygdaloidal contacts and many showing a distinctive population and size of phenocrysts, which are typical characteristics of flows and tuff. Where such units are highly appressed within a sedimentary sequence, they can be easily mistaken for dikes or sills; this may have been the case for the amphibolite bodies interpreted as metagabbro at File and Kississing lakes.

A conformable base of the same succession that occurs north of File Lake is exposed at Puffy Lake (column 2E in Figure 8e), where the sandstone and siltstone overlying the basal unit are relatively rich in metamorphic hornblende and epidote (*unit 13f*), suggesting original calcareous cement that has reacted with biotite to form K-feldspar and calcsilicate minerals. This is interpreted as the most distal section, consistent with a previous structural restoration (Figure 7b; Zwanzig, 1999).

Similar K-feldspar– and amphibole-bearing rocks also occur near the top of sections at File Lake and Cleunion Lake. These were probably deposited in a distal basin plain. The K-rich composition of these rocks is interpreted as being due to a component of rhyolite tuff, whereas the mafic component probably came from the intermediate–mafic volcanic rocks in the underlying parts of the Missi Group.

3.5.1.3 Late successor-arc intrusions

The youngest suite of plutons resulting with certainty from arc magmatism (units 15a–h) intrudes the Fourmile Island and Northeast Reed assemblages, as well as the Burntwood and Missi groups and older gneissic granitoid rock. The composition of these plutons ranges from gabbro to granite, with local smaller mafic–ultramafic intrusions. Both intermediate and felsic phases generally contain a higher proportion of biotite as a mafic phase than the older intrusions. Intrusions are subdivided into units that occur in one or several plutons on Maps GR2010-1-1 and -2 (back pocket/DVD), based on composition, texture and state of deformation.

The late successor-arc intrusions occur in the southeastern and western parts of the Kississing–File lakes area, where they lie east of the West Reed–North Star Shear Zone or northeast of the Loonhead Lake Fault (Figure 8a, back pocket). The plutons have very different intrusive habits and structure in the two areas. Those in the southeast have round to oval outlines and are upright stocks or batholiths rooted in the volcanic rocks or the structurally underlying Burntwood Group. They are crosscutting and some have a metamorphic aureole (Bailes, 1980a). The core of the simple plutons is massive or only moderately foliated. The late intrusions in the northwest are foliated, flattened, sheet-like bodies structurally emplaced with their hosting paragneiss. They have been highly metamorphosed with their hostrocks. Late successor-arc intrusions north of File Lake and in the Dow Lake area just north of the sillimanite-biotite-garnet isograd are transitional between the two habits. They show that the structural style in the Kisseynew Domain was imposed by penetrative deformation after intrusion and that the late deformation was restricted to shear zones and pluton margins in the south.

The Reed Lake 'pluton' is a complex body that has a core of older gneissic tonalite (1894 ±5 Ma; Percival et al., 2005). Strongly foliated plagioclase-phyric granodiorite east of Reed Lake with abundant xenoliths has yielded 1839 +5/–4 Ma zircons (Kraus and Menard, 1997). This phase and diorite from the shore of Woosey Lake appear to represent late 'crescentic' plutons forming the margin of this complex. The margin phases are separated by a shear zone from well-preserved Burntwood Group to the west, and intrude the Burntwood Group to the north with evidence of stoping. The gneissic core also contains migmatitic greywacke, probably of the Burntwood Group. These relationships suggest that the Reed Lake 'pluton' may be an extensional core complex. The age of the intrusive margin indicates that this dome formed during successor-arc magmatism. The various phases, however, have only been partially mapped (Syme et al., 1995), and only the chemistry of the quartz diorite phase is presented in this report.

Many large granitoid plutons in the File Lake area, originally mapped as syn- to late kinematic (Bailes, 1980a), are included with the late successor-arc intrusions in this report because subsequent dating of them has yielded late Missi Group ages. The few dated bodies include 1832±2 Ma gneissic granite (Nelson Bay dome; David et al., 1996) and the 1830 +27/–19 Ma, massive to gneissic granodiorite to leucotonalite and tonalite (Ham Lake pluton; Gordon et al., 1990). The 1826 +11/–5 Ma (Hunt and Schledewitz, 1992) Thunderhill Lake quartz porphyry, which closely resembles the Missi Group rhyolite, may also belong to this suite. The large errors in the ages of the latter intrusions permit their classification as 1830 Ma or older successor-arc intrusions.

The 1832 Ma Nelson Bay dome (unit 15a) is clearly related to Missi Group rhyolite and dacite by its continuity into the felsic volcanic rocks and its geochemistry (discussed in Section 5.9.5.1). The northern extension of the dome may consist of the younger leucogranite to tonalite (unit 19) but has not been remapped. Bailes (1980a) had interpreted the main body as metavolcanic orthogneiss, but Harrison (1949) had mapped it as foliated granite and the NATMAP Shield Margin Project Working Group (1998) showed it as gneissic granodiorite to tonalite. Bailes (1980a) described it as granoblastic oligoclase-quartz-biotite±microcline gneiss that is medium grained compared the adjacent fine-grained rocks currently mapped as Missi Group metarhyolite and dacite. The domal gneiss is generally homogeneous but has a lensoid gneissic foliation with faint local layering that is herein interpreted as secondary. The close relationship with Missi Group metarhyolite on its margin suggests that the coarser phase may be a resurgent dome intruding its volcanosedimentary cover.

The ca. 1830 Ma Ham Lake pluton (unit 15d) comprises massive to gneissic granodiorite and leucotonalite–tonalite that are medium to coarse grained (Bailes, 1980a). Evidence of strain and recrystallization increases at the margin and north across the body. Minerals include andesine, quartz, microcline, biotite and hornblende, with minor muscovite, chlorite and epidote, and garnet±sillimanite in the north.

The southeastern part of the Batty Lake Complex is a quartz-rich leucotonalite similar to the Ham Lake pluton, but it may be part of the older Batty Lake trondhjemite (unit 9h). Screens and inclusions of garnet-biotite gneiss interpreted as Burntwood Group have suggested the younger age, but the high error in the U-Pb zircon age of the trondhjemite in the north and the uncertain age of the onset of Burntwood Group sedimentation do not allow a definite conclusion.

The ca. 1826 +11/–5 Ma Thunderhill Lake porphyry (unit 15e) lies northeast of Weldon Bay and is related to compositionally similar Lobstick rhyolite in the Missi Group (Hunt and Schledewitz, 1992). The relatively young age is consistent with the position of Lobstick-type rhyolite in sequence 2 of the Missi Group near Cleunion Lake (column 1C″ in Figure 8d). The rock contains 1–1.5 mm quartz phenocrysts in a fine-grained matrix of microcline, plagioclase, quartz, biotite (6%), magnetite and local garnet. Microcline is segregated into layers, at least on the margin of the body.

The ca. 1839 Ma outer phase of the Reed Lake 'pluton' (unit 15g) includes leucocratic quartz diorite that has been sampled at the north margin (Bailes, 1980a). It is rich in plagioclase and features mafic grains with a core of hypersthene and metamorphic coronas of cummingtonite and hornblende.

The undated Norris Lake pluton (unit 15b) comprises plagioclase-rich hornblende-biotite leucotonalite that is generally massive except at the contacts with the surrounding Amisk Collage. It is considered a late stock (Bailes, 1980a). The intrusion contains small phenocrysts of andesine and, in the matrix, primary brown hornblende and secondary green hornblende and biotite.

Most of the undated bodies in the late successor-arc suite intrude the Burntwood Group or Missi Group. Granite bodies include the porphyritic Barron Lake pluton (unit 15c), which intrudes the Burntwood Group at Woosey Lake. It also intrudes the southern margin of the Ham Lake pluton. The stock contains granite and leucogranite with 10–20 mm phenocrysts of microcline (Bailes, 1980a). A finer grained margin phase and metamorphic aureole indicate a high level of intrusion.

Mafic to intermediate successor-arc intrusions include the East Big Island gabbro (unit 15h), which locally intrudes the Missi Group on Kississing Lake. Quartz diorite to tonalite and hornblende-biotite gneiss (unit 15g) intrude the Burntwood and Missi groups where these surround the Big Island dome and Yakushavich Island on northern Kississing Lake. The rocks are medium to coarse grained and contain 10–15% quartz and 20–30% hornblende, with <1% titanite and <2% magnetite. Schistosity, layering and the stratabound field relationship suggest that parts of the unit may be volcanic rocks belonging to the Missi Group.

3.6 Synkinematic intrusions (1.83–1.77 Ga)

Intermediate to ultramafic intrusions (*unit 16a*)[5] that are commonly hypersthene bearing in the Central Kisseynew subdomain have been tentatively correlated with similar bodies containing mainly amphibole as the mafic mineral (unit 16b) in the Kississing–File lakes area (Zwanzig, 1999). In the central subdomain, these intrusions are represented by the ca. 1830 Ma Touchbourne intrusive suite (TIS) of enderbite and related rocks (Manitoba Energy and Mines, 1986), considered to be early synkinematic (Baldwin et al., 1979). They have been included in the latest volcanic-arc plutons by Hollings and Ansdell (2002). In light of a proposed older age for collision tectonics in the Flin Flon and Kisseynew domains (Zwanzig, 1999) and the presence of an arc signature in some intrusive rocks with no contemporaneous subduction (Morris and Creaser, 2008), this conclusion is reassessed in this report by a closer examination of the geochemistry of the TIS, its correlatives in the Kississing–File lakes area and younger granitoid rocks.

The probable syn- to postcollisional origin of the TIS is important to understand because these intrusions have been interpreted to predate all other intrusions in the Kisseynew Domain (Baldwin et al., 1979), thus making the other bodies postcollisional. The 1830 Ma age with its high error (+11/–5 Ma) in Manitoba (Gordon et al, 1990) has been confirmed by the precise 1830 ±2 Ma age of enderbite in Saskatchewan (Ashton et al., 1999). Sheets of garnetiferous amphibolite or quartz ferrodiorite (unit 16c, Dow Lake quartz diorite) up to 100 m thick were intruded into an imbricate of the Loonhead Lake Fault or a late northern extension of the West Reed–North Star Lake Shear Zone on the west shore of Dow Lake. This undated unit and possibly an adjacent garnet-bearing felsite (unit 10d) are also interpreted as early synkinematic. The geochemistry and Nd-isotope composition of the synkinematic mafic to intermediate rocks (discussed in Section 5.10.1) are of particular interest in the tectonic evolution of the Kississing–File lakes area because of their lower-crustal and mantle origin.

A set of mafic sheets (unit 17a, Weldon Bay gabbro) with a minor volume but wide distribution has a field relationship that indicates syntectonic intrusion. They comprise fine-grained amphibolite, intruded into the Loonhead Lake Fault at Weldon Bay; into the Missi Group conglomerate, adjacent to the Loonhead Lake Fault at Evans Lake; and into the Burntwood Group at Burntwood Lake in the Central Kisseynew subdomain. The mafic sheets have experienced the 1815–1805 Ma peak metamorphism and deformation (Zwanzig, 1999) but postdate the ca. 1830 Ma (late Missi) thrusting. In the Sherridon structure, unit 17a and b mafic and ultramafic bodies (Cree Lake intrusions) have a similar geochemistry to Weldon Bay gabbro and may be correlated with it. Unit 17 (undivided) high-Ti amphibolite (Halo Resources Ltd., pers. comm., 2008) is also tentatively correlated with the Weldon Bay gabbro because of its close geochemical affinity to the intrusive rocks and its gabbroic appearance.

A tonalite sheet (Burntwood Lake tonalite), which has a trace-element composition similar to that of the garnetiferous tonalite (*unit 18a*, Highrock Lake Batholith) and Flatrock Lake granite (*unit 18b*) that intrude the TIS in the central part of the Kisseynew Domain, has yielded a zircon crystallization age of 1823 ±4 Ma (Machado et al., 1999). The sheet is interpreted as part of the same period of felsic magmatism as these large intrusions. The age postdates the collisional underthrusting by the Sask craton beneath the Flin Flon–Glennie Complex, which is estimated to have been prior to the 1826 Ma intrusion of a quartz diorite dike into the décollement in the roof of the craton (Ashton et al., 1999; Ashton et al, 2005). Consequently, the large central plutons can be considered as also being postcollisional. The U-Pb zircon age of 1814 +17/–11 Ma (Gordon et al., 1990) on variably porphyritic granite at Burntwood Lake represents a common late phase of the central plutons and

[5] Units in italics are not shown on Map GR2010-1-1 in back pocket.

provides their minimum age. Granitoid sheets of similar composition (unit 19c) that intrude the Burntwood Group in the Kississing–File lakes area have synkinematic field relationships and are probably also coeval with the large plutons in the north. All of these rocks are briefly assessed or reassessed in this report because of this uncertain origin: they have been considered to be part of the latest arc magmatism (Hollings and Ansdell, 2002) and, alternatively, as (postcollisional) crustal melts (White, 2005).

The youngest sheets and lenses are leucogranite, pegmatite and biotite granodiorite (units 19a, b) associated with the advanced stages of high-grade metamorphism and migmatization (Zwanzig, 1999). They were examined only locally. Studies of the relatively young pegmatitic phases, which provided U-Pb monazite- and zircon-discordia ages as well as concordant single-grain monazite ages, yielded dates ranging from 1818 to 1789 Ma (Parent et al., 1999). Some of these sheets have sharp contacts with the Burntwood Group or margins intruded by pegmatite. Other margins are raft complexes of the Burntwood Group surrounded by sill complexes (Zwanzig and Schledewitz, 1992). Mineral contents are K-feldspar (generally greater than plagioclase), biotite (1–5%) and muscovite (<4%). Garnet is locally present where rafts of metagreywacke are abundant. Very similar sheets (*unit 19c*) contain little K-feldspar and have a composition of granodiorite or leucotonalite. Some of these intrude the mafic rocks but have not been distinguished on Maps GR2010-1-1 and -2 (back pocket/DVD). The tonalitic and later granitic leucosomes are also of interest because they provide clues to the origin of the late granitoid rocks and the thermal history of the area.

3.7 Discussion of tectonostratigraphy

The highly deformed and metamorphosed tectonostratigraphic units in the transition from the Flin Flon Domain to the Kisseynew Domain are extensions of units in the better preserved part of the Flin Flon Domain and had the same early path of evolution. New data from the Kississing–File lakes area therefore have a direct application also to the low-grade rocks. The oldest units divide into 1) arc assemblages, which contain mafic to felsic volcanic rocks, including breccia and sedimentary rocks; and 2) ocean-floor, arc-rift or back-arc–basin assemblages, which consist entirely of thick mafic volcanic units and mafic intrusions. These assemblages were structurally juxtaposed in alternating belts during early tectonic intraoceanic-arc accretion (Lucas et al., 1996). The data also support an evolution of the Amisk Collage in the southwest separate from that of the assemblages to the north and east (Syme at al., 1995). However, the new data suggest an earlier, belt-wide accretion event. Differences in tectonic evolution appeared only during an intermediate stage of greenstone-belt evolution, not during the earliest and latest stage. This complex history is addressed in this section and in Section 6.1.4.

The oldest volcanic rocks, subvolcanic intrusions and minor sedimentary rocks in the Kississing–File lakes area belong to the Amisk Collage (arc and ocean-floor assemblages), the Northeast Reed Lake (ocean-floor) assemblage and the Snow Lake (arc) assemblage. Josland Lake sills have intruded all of these assemblages, which must therefore have a minimum age of 1886 Ma. The Amisk Collage is also intruded by a suite of early successor-arc intrusions, and the intruded terrane is unconformably overlain by the late successor-arc deposits belonging to the Missi Group. The Fourmile Island assemblage forms the northern and eastern parts of the Amisk Collage and extends from its type area at Reed Lake northwest for 100 km, at least partly across Kississing Lake. It is everywhere intruded by the 1886 Ma Josland Lake sills and in fault contact with the younger (ca. 1.85 Ga) Burntwood Group. It is interesting that the Claw Lake gabbro in the Reed Lake mafic–ultramafic complex, structurally above the Fourmile Island assemblage, has an age of 1901 ±6 Ma (Stern et al., 1995a). Thus, the earliest episodes of ocean-floor and arc magmatism appear to have the same range of ages as in the vicinity of Flin Flon.

The Northeast Reed and Snow Lake assemblages are in fault contact with the Burntwood Group. This relationship supports the conclusion (Syme et al., 1995) that the eastern part of the Flin Flon Domain was structurally assembled much later than the Amisk Collage. The intrusion of the Josland Lake sills into the volcanic assemblages, however, suggests that earliest tectonic assembly may have occurred before ca. 1886 Ma and involved the entire Flin Flon Domain. Large-scale recumbent folding at ca. 1886 Ma was probably the result of the early accretion. The Sherridon–Meat Lake assemblage, which may also be in fault contact with the Burntwood Group, experienced late accretion and may be related to the eastern part of the Snow Lake subdomain. The various assemblages are probably para-autochthonous, developing in adjoining terranes that were telescoped or affected by transpression. The eastern part of the Flin Flon Domain may have rifted subsequently and become separated from the Amisk Collage by a back-arc basin. It was reassembled into the present composite terrane after deposition of the Burntwood Group, during Missi Group sedimentation and the latest arc magmatism. This speculative history of evolution of the southeastern part of the Reindeer Zone is tested for consistency using the geochemical data in Section 5.7.7.

Two ages of successor-arc intrusions can be distinguished by crosscutting relations and U-Pb zircon (ID-TIMS) dating in the Kississing–File lakes area. Many plutons in the early group are unconformably overlain by the Missi Group, whereas the late plutons cut the Burntwood and Missi groups. The early plutonism ranged from 1876 ±2 Ma (Machado et al., 1999) to 1845±3 Ma (Whalen and Hunt, 1994) on a regional scale. The area underlain by the Amisk Collage was host to these intrusions and thus must have remained a contiguous composite terrane during this prolonged early successor-arc magmatism. The late intrusions occur mainly in the eastern part of the Flin Flon Domain, where final assembly took place after deposition of the Burntwood Group. Their regional range of crystallization ages (1842 Ma to ca. 1830 Ma) indicates that this period of plutonism was largely coeval with deposition of the Missi Group sedimentary and volcanic rocks, which may have started ca. 1848 Ma (Machado et al., 1999) but with the main pulse of volcanism at about 1836 Ma (Ansdell et al., 1999).

The Burntwood Group comprises coalescing turbidite fans (Bailes, 1980b) and shows northward fining, thinning of beds and presumably northward sediment transport from Reed Lake into the Kisseynew basin. Although it is in fault contact with the early volcanic assemblages in the east, it shows an upward gradation through a prograding shoreline deposit into the non-marine sedimentary rocks of the Missi Group in the Kississing–File lakes area. A structural restoration (Zwanzig, 1999) is consistent with the conclusion that the Burntwood Group is the distal facies of the Missi Group.

Although the Missi Group in the Kississing–File lakes area is highly metamorphosed and strongly deformed, it appears to preserve a stratigraphy controlled by local fault-bounded basins or sub-basins. A basal conglomerate generally overlies older volcanic and intrusive assemblages in the southern basins, but a downward gradation to metapelite and greywacke forms the contact with the Burntwood Group in the Kisseynew basin to the north. The sequences along the northern and eastern margins of the Flin Flon Domain differ from the older and coarser, more widely derived clastic deposits without intercalated volcanic rocks of the Missi Group in the Flin Flon area. Nevertheless, these deposits formed in the same orogenic continental- or emerged-arc environment that existed for more than 15 m.y. Like the late plutonism, the Missi-age volcanism is restricted to the Flin Flon Domain–Kisseynew Domain transition and to the eastern part of the Flin Flon Domain that experienced the post–Burntwood Group accretion. The thickest part of the succession (~1800 m, deformed) is interpreted to comprise three fining-upward fluvial-alluvial sequences of meta-arenite to calcareous sandstone or pelite that include about 450 m of intercalated volcanic rocks. Locally, cobble conglomerate occurs near cryptic normal faults that may have formed along sub-basin margins. The distribution of thick volcanosedimentary units and equivalent condensed sequences in block-like areas suggests a rugged depositional topography of upthrown and dropped fault blocks, typical of extensional terrains. The lateral gradation of felsic tuff and subvolcanic domes into breccia and tuffaceous sedimentary rocks indicates that the fault basins were filled during active volcanism. The range of mafic to felsic compositions and the abundance of tuff suggest that deposition occurred in an arc environment, apparently an extensional subaerial arc. The latest arc magmatism is synkinematic and apparently occurred after the onset of continental collision.

Minor felsic to ultramafic intrusions in the Kississing–File lakes area that correlate with larger plutons in the Central Kisseynew subdomain may represent even younger arc magmatism. However, field relationships and the tectonic history of the area with early collision suggest that this was postcollisional magmatism. The question is tested using geochemistry in Section 5.10 and addressed in the regional tectonic synthesis in Section 6.1.

4 Structure and metamorphism

The area between Kississing Lake and Reed Lake forms the transition from northerly trending upright folds, faults and shear zones, typical of the main part of the Flin Flon Domain (Bailes, 1980a; Ryan and Williams, 1999), to northwest-trending refolded recumbent structures in the gneissic units of the Kisseynew Domain (Zwanzig and Schledewitz, 1992; Zwanzig, 1999). These structures record the history of synvolcanic deformation, arc-terrane accretion, protracted plutonism, continental collision and intracontinental transpression (Lucas et al., 1996) as seen at a range of crustal levels. The change in structural style from south to north represents the transition from rocks that remained in the upper crust (suprastructure) to those that have experienced midcrustal thermotectonism (infrastructure). Such structural changes are known world-wide, including Precambrian terrains, and demonstrated by thermal-mechanical models (Culshaw et al., 2006). In the Kississing–File Lake area, this change is a rapid gradation that lacks regional discontinuities but retains a continuity of units with only modest ductile offsets. High-strain zones, although concentrated along the transition, have a limited strike length and do not generally break the continuity of units. They represent a broad zone of ductile shear and step folding. The volcanic rocks occupy early upright folds and steeply dipping elongate fault blocks at low metamorphic grade in the suprastructure to the south (Bailes and Syme, 1989). These northerly trending structures curve abruptly to the northwest and west at the northern margin of the Flin Flon Domain, where the dip shallows to nearly recumbent structures in orthogneiss. Such highly attenuated recumbent structures and ductile shear zones in the absence of faults have been interpreted to represent a structural style common to many orogenic belts (Williams and Jiang, 2005). They are a fundamental feature that allows units and their mineral potential to be traced across the Kississing–File lakes area.

Two- and three-dimensional seismic data in the vicinity of Flin Flon show a similar change in structural style in vertical crustal sections and blocks. Upright and inclined folds and faults are seen as imbricate structures that become shallower dipping with increasing depth. Structural panels extend through the change in orientation without a break. The structural transition across the Kississing–File lakes area is therefore consistent with a change in structural level that provides a view of the full crustal architecture.

The protracted deformation at all crustal levels has produced the northeast-dipping stack of lithotectonic packages that controls the present distribution of units in the southeastern part of the THO in Manitoba. Archean gneiss of the Sask craton occurs in its lowest level, and this is structurally overlain by the early arc and back-arc massif (Flin Flon–Glennie Complex) that is unconformably overlain by the Missi Group. The structurally uppermost, northeast-dipping, refolded sediment-derived gneiss of the Kisseynew Domain is locally interleaved with the orthogneiss developed from the Flin Flon–Glennie Complex.

In the Kississing–File lakes area, the upright structures are preserved south and west of File Lake (Butler Lake and Dickstone blocks), whereas younger, refolded recumbent structures dominate the high-grade metamorphic rocks in the north. The geometry and relative age of major folds and faults that affect these areas were discussed in detail by Bailes (1980a) and Zwanzig (1999) and are summarized in Figure 12. The complex large-scale geometry of the recumbent structures is illustrated in east-west and north-south structure sections constructed by down-plunge projection of contacts based on structural analysis (Zwanzig and Schledewitz, 1992; Zwanzig, 1999) and reproduced in Figure 13. Simplified structure sections and stratigraphic restorations across File Lake were published by Bailes (1980a). They are updated and reassessed in this section.

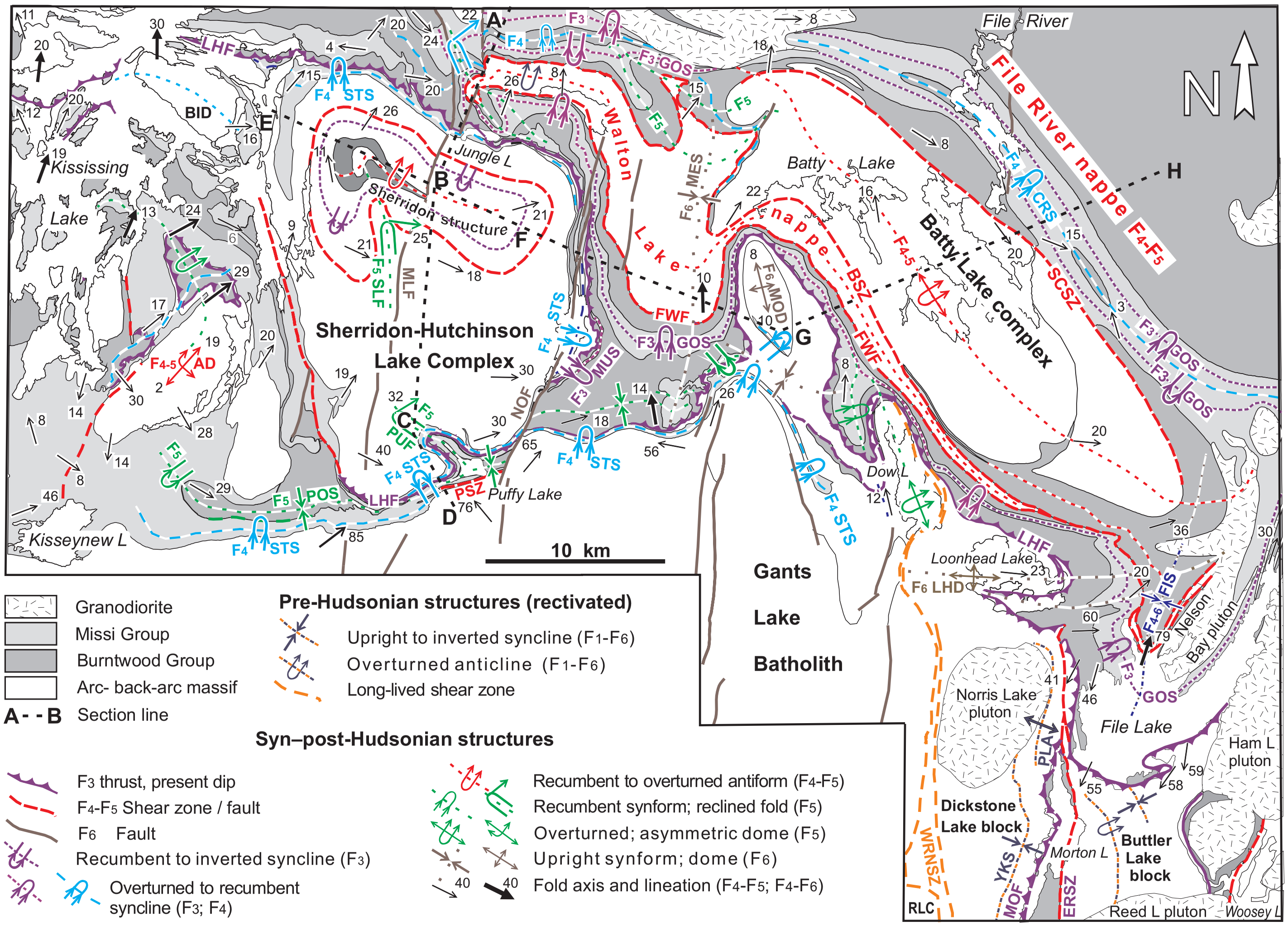


Figure 12: *Structural interpretation of the Kississing–File lakes area after Zwanzig (1999). The map shows the major folds, faults and shear zones inferred from reversals and breaks in the stratigraphic succession and from structural measurements plotted on maps and stereograms (not shown). Approximate axial traces were drawn for synclines and orthogneiss culminations. The plunge of the major structures is taken from β-poles to girdles, minor fold axes and linear structures in local subareas (not shown, but see Zwanzig and Schledewitz, 1992). Lineations and fold axes at File Lake are after Connors (1996), major fold traces after Bailes (1980a). F_1 pre-Missi fold: YKS, Yakymiw syncline. Long-lived shear zone: WRNSZ, West Reed–North Star Shear Zone. D_3 thrust faults (possibly reversed during D_4): LHF, Loonhead Lake Fault; MOF, Morton Lake Fault. F_3 synclines: GOS, Gohl Lake syncline; MUS, Mud Lake syncline. F_4 synforms: CRS, Craik Lake synform; STS, Star Lake syncline. Long-lived (D_{3-5}) shear zones and faults: ESRZ, East Reed Shear Zone; FWF, Fairwind Lake Fault. F_4–F_5 complexes: AD, Adamson Lake dome; Batty Lake Complex; Sherridon–Hutchinson Lake Complex. D_4–D_5 shear zones: BSZ, Batty Lake Shear Zone; PSZ, Puffy Lake Shear Zone; SCSZ, Shineton Creek Shear Zone. F_4–F_6 structure: FIS, File Lake synform. F_5 folds: POS, Ponton Lake synform; PUF, Puffy Lake fold; SHF, Sheila Lake fold. F_6 structures: LHD, Loonhead Lake dome; MES, Meat Lake synform; MOD, Moody Lake dome. F_6 faults: MLF, Molly Lake Fault; NOF, Nokomis Lake Fault; A–D, E–H, structure section lines on Figure 13.*

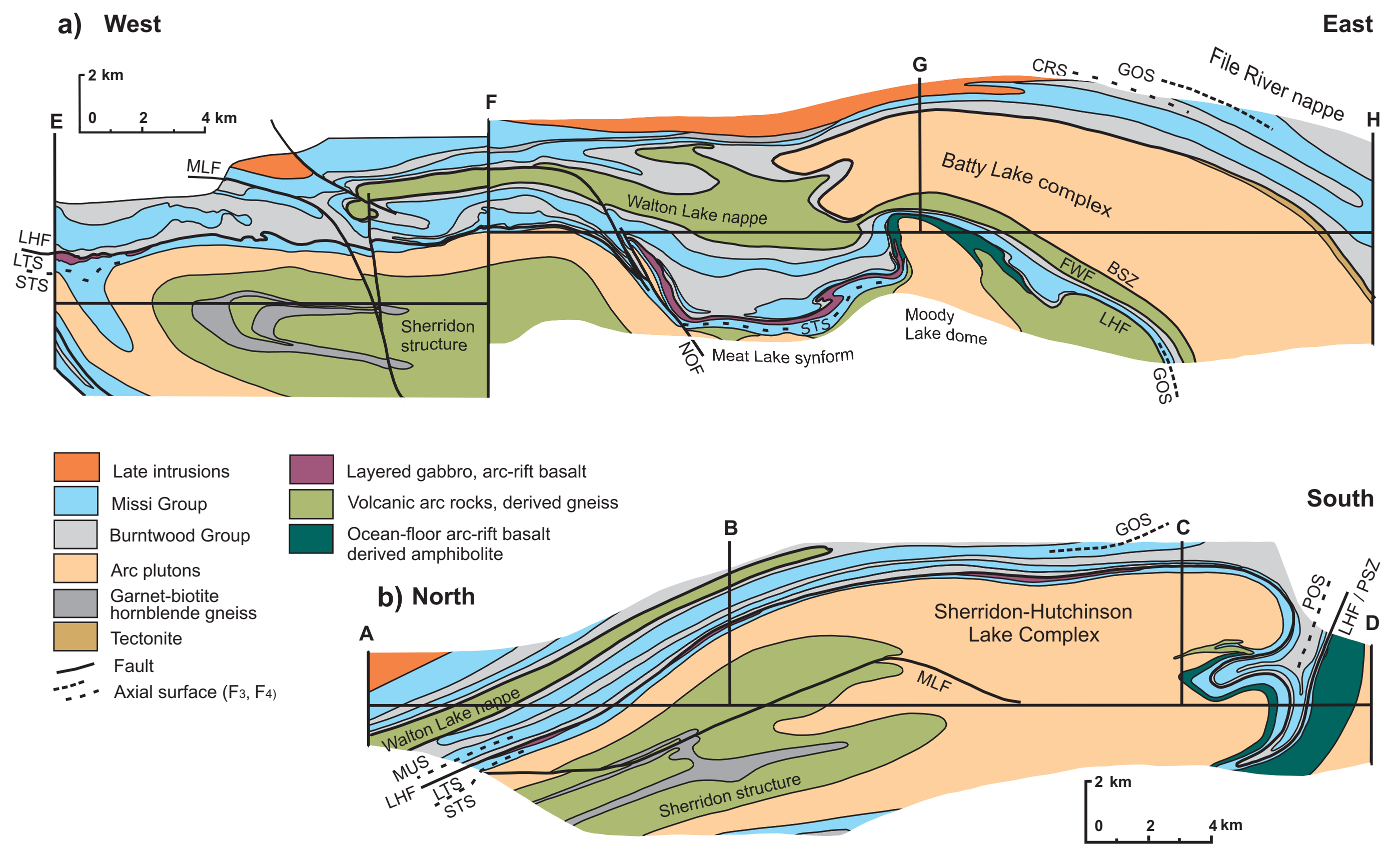


***Figure 13:** Simplified vertical structure sections drawn as down-plunge projections in segments with relatively uniform plunge: **a)** east–west section across the Sherridon-Hutchison Complex, Walton Lake nappe and Batty Lake Complex; **b)** north–south section across the Sherridon–Hutchinson Lake Complex and Star Lake synform. The sections follow structural culminations and are projected out of the adjacent synforms. This provides only a schematic depiction below the line of section (horizontal line). The Molly Lake Fault (MLF) and the parasitic folds below the Walton Lake nappe lie in the plane of the section, and are schematic. The amplitude of the Meat Lake synform (MES) is exaggerated. Areas close to bends in the section (vertical lines), where the plunge changes, are distorted. Projection planes and plunge were determined from a set of stereoplots of structural data (Zwanzig and Schledewitz, 1992). The lines of cross-sections A to H and acronyms of the major structures are defined in Figure 12. Additional abbreviation: LTS, Loonhead Lake thrust sheet.*

A summary of present knowledge of the structural evolution of the area is given below and in Table 5.

4.1 Structural evolution

The structural evolution of the Kississing–File lakes area occurred during the various episodes of arc and back-arc magmatism, accretion, sedimentary basin development and continental collisional to postcollisional convergence. Deformation took place over a period of 110 m.y. or more (Figure 14). The main stages were accompanied by regional metamorphism throughout the Flin Flon and Kisseynew domains at all exposed structural levels. Early structures that are preserved in the lower grade metamorphic rocks in the south are generally transposed in the north by younger structures formed at deeper levels. Relative ages in such a large volume of crust over this long period cannot simply be determined from overprinting relationships of structural elements, although these provide critical data in more restricted areas (Stauffer and Mukherjee, 1971; Bailes, 1980a; Zwanzig and Schledewitz, 1992; Norman et al., 1995; Syme et al., 1995; Connors, 1996; Norquay, 1997; Connors et al., 1999; Ansdell et al., 1999; Kraus and Williams, 1999; Ryan and Williams, 1999; Zwanzig, 1999). Consequently, the timing of various stages of the structural evolution is defined, in large part, by U-Pb dating of zircon and monazite that crystallized during consecutive episodes of magmatism and metamorphism.

Midcrustal deformation during the thermal peak at approximately 700°C and 5.5 kb in the southern Kisseynew Domain (Gordon et al., 1993; Froese, 1997) is interpreted to have tightened, overturned and/or rendered recumbent most earlier structures, while generating new folds parallel to the direction of stretching, normal and parallel to changing directions of tectonic transport. Large-scale sheath folds developed locally by flow and attenuation of domes that had formed by fold interference and possible early extension. Late flexural upright folds, developed during cooling and decompression, may have formed on synchronous, nearly orthogonal axes (Zwanzig, 1999) guided by regional shear zones. This complex and protracted synmetamorphic deformation is reanalyzed in this section using existing and new data from the Kississing–File lakes area and surrounding areas, concepts generated by new U-Pb ages and tectonostratigraphic analysis, and new mapping in the vicinity of Flin Flon. The work at Flin Flon (R-L. Simard et al., work in progress, 2010) has yielded a similar history of deformation that serves to confirm the model presented here within the present limits of understanding.

Table 5: Structural history of the transition between the Flin Flon and Kisseynew domains (1885–1770 Ma). Deformation phases in italics are for the south-central Flin Flon Domain (after Ryan, 1998). Shaded areas indicate the Trans-Hudson orogeny.

Age (Ga)	Phase		Structural style	Main structures (e.g.)	Tectonic stage
~1.885	D_1	*D_1*	High-strain corridors, fault zones (shallow dip?)	West Reed-North Star Shear Zone (WRNSZ)	Amalgamation of early volcanic arc and back-arc assemblages
			Recumbent folding	Yakymiw syncline (YKS)	
1.87 –1.85	D_2	*D_2*	Northeast- to northwest-trending steep foliation	Regional foliation, shear zones	Early successor-arc magmatism (probably due to shallow subduction of young back-arc lithosphere)
1.845 –1.825	D_3		Fold-and-thrust system of pro-wedge directed toward the Kisseynew Domain, then retro-wedge thrusting toward Sask craton	Morton Lake Fault Zone (MOF), West Reed Shear Zone (WRSZ), Loonhead Lake Fault (LHF), Gohl Lake syncline (GOS), Mud Lake syncline (MLS)	Closing of Kisseynew basin during Flin Flon-Glennie Complex collision with Sask craton. Renewed arc-related magmatism and basinal sedimentation; probable delamination
			Early dome development	Reed Lake pluton, Northern domal complexes	Possible back-arc core complexes similar to Namew Gneiss Complex south of Flin Flon
			Midcrustal flow (directed over Sask craton)	Pelican thrust zone (exposed in Saskatchewan)	Collision of Flin Flon–Glennie Complex with Superior craton in the southeast
~1.82	D_4		North: Fold nappes, large folds, now recumbent, originally west verging	File River nappe (early stage), Sherridon-Hutchinson Lake Complex, Walton Lake nappe, Star Lake syncline (STS), Batty Lake Complex	Northeast convergence of Superior craton and its overriding of the east side of the Reindeer Zone
			File Lake–Snow Lake: Thrust faults with westerly cut-offs and early transport, and start on north-northeast-trending folds	File Lake synform (FIS), Fairwind Lake Shear Zone (FWSZ), McLeod Road Fault (MRF), Snow Lake Fault (?; SLF), start of Berry Creek Fault (BCF)	
1.815 –1.79	D_{5a}	*D_3*	Synmetamorphic folding South: upright folding with northeast-trending foliation	Main regional foliation, steep dip in the south but shallow dip at deeper level in the north	Overriding by the northern part of the Reindeer Zone (attached to Hearne craton); southwest mid-crustal (channel?) flow in the Kisseynew Domain driving the Flin Flon–Glennie Complex farther over the Sask craton; further unroofing of domal complexes; north-east convergence of Superior craton
	D_{5b}		North: south-southwest-verging recumbent folding with northeast stretching; late, east-west upright folding and sinistral shear zones in structural transition	File River nappe (late stage), domal culminations (flattened), Ponton Lake synform (POS), Puffy Lake Shear Zone (PUSZ), Kississing River Shear Zone (KSZ)	
~1.80 –1.77	D_{6a}	*D_4*	Upright folds, steep shear zones, retrograde foliation	Meat Lake synform (MES), Moody Lake dome (MOD),	Oblique intracontinental convergence of the Reindeer Zone and the Superior craton (probably distal orogeny)
	D_{6b}	*D_5*	North-northeast- and east- to northeast-trending, east-side-up sinistral faults Late easterly dextral faults	Loonhead Lake dome (LHD), Nokomis Lake Fault (NKF), Molly Lake Fault (MLF), Puffy Lake Shear Zone	

Given that fold phases may be diachronous and that upright high-level structures were formed while recumbent folding took place at depth, U-Pb monazite and zircon ages are vital to establishing a coherent structural history. To date, however, different groups of structural geologists have rarely come up with exactly the same scheme of deformation. For example, thrusting of early volcanic assemblages over the basinal deposits in the direction of the Kisseynew Domain, although well established in the Flin Flon area (Gale et al., 1999; R-L. Simard et al., work in progress, 2010), is not generally recognized in the northeastern part of the Flin Flon Domain. Our contribution, keyed to Figures 12 to 14 and summarized in Table 5, represents another snapshot of our current understanding. An important part of this is the ages of the various stages of deformation, even though ages are under revision. The six stages and their respective tectonic origins proposed herein have affected both the Flin Flon Domain and the Kisseynew South flank. They indicate several changes in vergence. Further complications or simplifications will come with future data. Nevertheless, this should not detract from the main purpose of this structural analysis, which is to help trace major lithological packages through the complex terrain and thereby assign a mineral potential to them (Section 7.1.3).

4.1.1 D_1 deformation: early assemblage amalgamation and rifting

The earliest recognized structures (F_1; Bailes, 1980a) are the north-trending Yakymiw syncline (YKS; southwest of File Lake in Figure 12) and Podruski Lake anticline (PLA), whose axial surfaces are intruded by the 1886 Ma Josland Lake sills (Zwanzig et al., 2001). An earlier but petrographically identical sill is folded in the PLA, indicating an overlap in the period of D_1 folding and the magmatism. At the type section of the Fourmile Island assemblage on southwestern Reed Lake, a major anticline occurs southeast of the dated Josland Lake sill (Syme et al., 1995). This fold is also interpreted as a D_1 structure. The

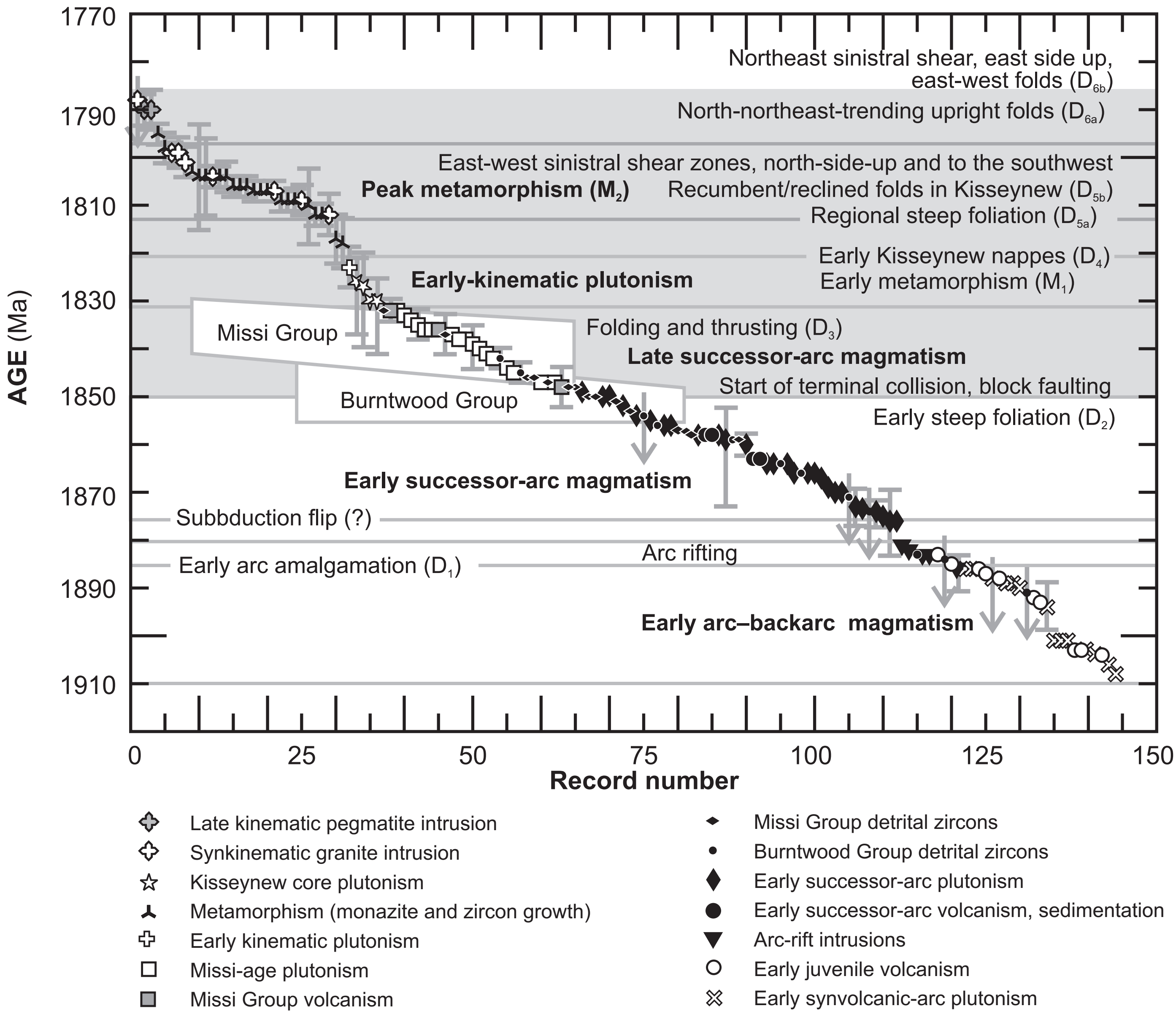


Figure 14: *U-Pb isotope dilution–thermal ionization mass spectrometry (ID-TIMS) ages of igneous and detrital zircons, and metamorphic monazites and zircons from the Flin Flon Domain and southern Kisseynew Domain in Manitoba. Bracketed is the timing of progressive stages of magmatic evolution from early arc and back-arc volcanism to late successor-arc plutonism, basin development (Burntwood and Missi groups) and synkinematic (collisional and postcollisional) intrusion. Stages of structural evolution are compiled from the literature cited in the text. The Trans-Hudson orogeny (grey shading) is herein defined as the terminal collisional and postcollisional continental convergence stages between the Archean cratons (and crustal fragments) surrounding the orogen and caught up within it. The 143 U-Pb ages are from the Manitoba Geological Survey geochronology database (Manitoba Geological Survey, 2006) and include those with errors (2σ) less than ±15 Ma, generally ±2 to ±5 Ma. Error bars are shown for the 52 ages from the Kississing–File lakes area (many of these being metamorphic). Arrows indicate moderately discordant (minimum) detrital ages. There are Archean detrital ages but they are not shown.*

West Reed–North Star Shear Zone (WRNSZ), which bounds the Dickstone block on the west, and a north-trending anticline-syncline pair in the Butler Lake block probably started to form at a similar early age (Syme et al, 1995). Because both the Amisk Collage and the Snow Lake subdomain are intruded by the Josland Lake sills, these structures probably formed during early amalgamation (D_1) or subsequent rifting of the oldest oceanic-arc and back-arc assemblages in the Central Flin Flon and Snow Lake subdomains. Some of the older sills, deformed in the F_1 folds, are interpreted to belong to the Josland Lake suite because of their proximity to less deformed sills and geochemical characteristics that are identical to the others (Section 5.7.6). The relationship between the folded and planar sills suggests that the deformation occurred ca. 1886 Ma, overlapping the time of their intrusion. Coeval extension and compression in adjacent areas could result from microplate rotation, which is common in the oceanic-arc environment (Wallace et al., 2005). Alternatively, sill intrusion may have occurred in two phases, one before and another after D_1 deformation. All the sills were originally emplaced as horizontal sheets of magma that became strongly differentiated and layered. Sills that occupy axial planes show that the host folds were recumbent at the time of their intrusion (Bailes, 1980a). Such early subhorizontal structures were predicted by Ryan and Williams (1999) from the shallow plunge of lineations in the central part of the Flin Flon Domain. Similar shallow-plunging lineations were recorded by Norquay (1997). A less likely possibility is that the recumbent structures and coeval sills are related to the uplift and extensional collapse of part of the Flin Flon–Glennie Complex during the development of arc rifting (Zwanzig et al., 2001). The early basin-fill units, the Yakymiw Formation in the west and the Parisian Formation in the east (Bailes, 1980a), were also intruded by Josland Lake sills before and after folding. The Parisian Formation and the Northeast Reed (ocean-floor) assemblage contain stratigraphically upward facing (early) Josland Lake sills, but the Yakymiw Formation and underlying volcanic units were folded before the intrusion of the latest (1886 Ma) Josland Lake sills in the west. Where the regional metamorphic grade is low, albite-epidote hornfels occurs at the contact with the sills (Bailes, 1980a), whereas the early recumbent folding in the File Lake area was followed by early greenschist-facies metamorphism (Bailes, 1980a).

The 1.90 Ga Reed Lake mafic–ultramafic complex (RLC in Figure 5), associated with the Elbow-Athapapuskow ocean-floor assemblage (Syme et al., 1995), occurs within the West Reed–North Star Shear Zone (WRNSZ). It faces west and structurally overlies the Fourmile Island assemblage if tops are restored in the complex and in the largest Josland Lake sill. This relationship suggests that the Elbow-Athapapuskow assemblage was thrust over the Fourmile Island assemblage during the D_1 amalgamation and recumbent folding. Structural fragments of the RLC also occur in the northern extension of the tectonite of the WRNSZ at Dow Lake. They suggested that the tectonite formed a regional detachment zone during dismemberment of an ophiolite complex and its obduction onto the Fourmile Island arc assemblage. The recumbent Yakymiw syncline and the anticlines east of the Josland Lake sill were therefore in the footwall of the shear zone and suggest that the assembly was west over east (present co-ordinates).

4.1.2 D_2 deformation: synplutonic compression or transpression

Steeply dipping, northeast- to northwest-trending structures (D_2) were formed in the early volcanic assemblages during the emplacement of several generations of successor-arc plutons (Ryan and Williams, 1999). These structures are preserved in the south-central part of the Flin Flon Domain, probably as far east as the WRNSZ. Well-preserved to strongly foliated volcanic rocks occur in northerly trending subvertical homoclines and large folds separated by corridors of high strain, marked by shear zones and brittle faults (Bailes and Syme, 1989; Ryan and Williams, 1999). The WRNSZ was probably active at that time. The 1876–1845 Ma old successor-arc plutons, which dominate the Central Flin Flon subdomain, show progressively stronger foliation with age and suggest that D_2 was a major episode of compression and shear (Ryan, 1998). In the vicinity of Flin Flon, F_2 folds also form the main regional structures that predate the Missi Group (R-L. Simard et al., work in progress, 2010). Shallowing subduction of relatively young back-arc–basin floor during this period (Whalen et al., 1999) may have caused the deformation, although a transpressional environment may be more favoured for the intrusion of abundant volcanic-arc plutons and is consistent with later deformation (Ryan and Williams, 1999). Such transpression may have produced large-scale repetition of structural packages. A general absence of plutons in this age range at Snow Lake and the retention of shallow dipping upward-facing structural panels suggest that the Snow Lake subdomain may have been little affected by D_2 deformation. Southeast of Snow Lake, however, steeply east-southeast-dipping foliation (parallel to the earliest foliation in the southern part of the Central Flin Flon subdomain) may have started to form during D_2.

The end of the D_2 transpressional deformation is marked by block faulting and local crustal extension during the development of basins and sub-basins filled with the Missi Group sedimentary and volcanic rocks. Sedimentary facies and thickness changes across such faults (proposed in Section 3.5.1.2) are further discussed in Section 6.1.6.2.

4.1.3 D_3 deformation: early collision

During the onset of D_3 in the File Lake area (ca. 1.845–1.832 Ma), the early arc and ocean-floor assemblages are interpreted to have occurred in elongate structural blocks between northerly trending faults, shear zones and plutons, as still preserved in the least deformed parts of the Flin Flon Domain (Figure 5). The blocks were accreted to the Amisk Collage along with narrow belts of Burntwood Group successor-basin deposits. The D_3 deformation produced the earliest folds in the Burntwood Group (generally called F_1 in earlier works). Major D_3 structures are northerly trending folds and faults, including the Morton Lake syncline, the Morton Lake Fault (MOF) and the East Reed Shear Zone (ERSZ) that involve and bound the Burntwood Group on Morton Lake (Figure 12; Bailes, 1980a; Syme et al., 1995). In the main part of the Kississing–File lakes area, the D_3 structures were reactivated, overturned and refolded (Zwanzig, 1999). The D_3 'event' may represent the closing of an ocean basin generated during arc rifting and spreading that occurred 40 m.y. earlier. The deformation probably occurred during the onset of collision of the Flin Flon–Glennie Complex with the Sask craton and underthrusting of the craton beneath the Flin Flon Domain (Ansdell et al., 1995). This appears to have led to continental collision with the Superior craton margin (White et al., 2002).

Syme et al. (1995) interpreted the D_3 Morton Lake Fault (MOF) as a thrust, along which the Burntwood Group was carried south over the Fourmile Island assemblage. If the dated Josland Lake sill is used to restore the F_1 recumbent folds, however, the interpretation of Syme et al. (1995) requires a second phase of recumbent folding to invert the sill before D_3 thrusting (Figure 15). A sum of absolute rotations required in the inverted limb of the F_1 Yakymiw syncline would be 360° and an additional 90° to bring it to the present orientation. Thus, the displacement on the Morton Lake Fault is more simply interpreted as being toward the Kisseynew Domain with the Burntwood Group in the footwall, consistent with the displacement history determined by Zwanzig (1999) for the Loonhead Lake Fault. It is likely that the Morton Lake syncline (MOS in Figure 15)

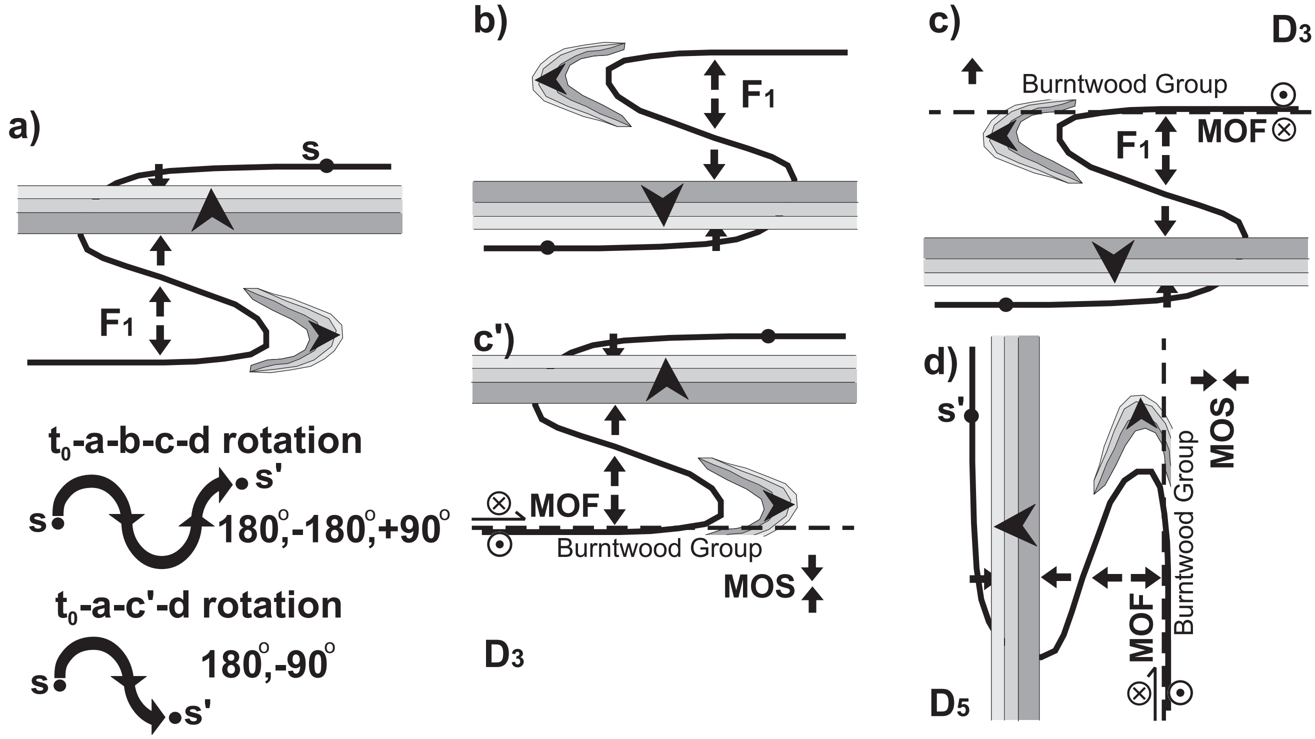


Figure 15: *Section views, with observer looking north, of alternative progressive-development models of the Morton Lake Fault (MOF) involving* ***a)*** *D_1 recumbent folding (Yakymiw syncline and Podruski Lake anticline) and intrusion of an early and a late Josland Lake sill;* ***b)*** *complete inversion; required by* ***c)*** *overthrusting of the Burntwood Group toward the south (after Syme et al. 1995); or* ***c')*** *Amisk Collage directly overthrusting the Burntwood Group with a northeast component of displacement, as in this report; and* ***d)*** *late development of the Morton Lake syncline (MOS) required by Syme et al. (1995); or early (D_3) development of the MOS as originally determined by Bailes (1980a) with later (D_5) steepening of all structures during development of peak metamorphic foliation. Total external rotation at s during folding is less in the preferred model with c'. The figure also shows that the D_1 vergence was easterly.*

formed in the footwall of the MOF. This took place after deposition of the Burntwood Group.

Timing and contact relationships between the late successor-arc intrusions and the D_3 folding and thrusting are apparent in the File Lake–Woosey Lake–eastern Reed Lake area as elsewhere in the Flin Flon Domain. The intrusive contact of the Reed Lake pluton has a thermal aureole that affected the Northeast Reed assemblage and the Burntwood Group (Bailes, 1980a). This indicates that the eastern fault contact with them formed before the 1839 +5/–4 Ma intrusion of the margin phase but after the ca. 1850 Ma deposition of the Burntwood Group. A series of south-side-up, south-dipping faults in the area north of Flin Flon (Club Lake, Catharine and Railway faults in Gale et al., 1999; Simard et al., 2010) also indicates thrusting of the Amisk Collage over the late successor-arc sedimentary rocks. The faults that carried the volcanic rocks over the Missi Group also include a syncline in the footwall sedimentary rocks, as proposed for the MOS (Flin Flon Creek syncline). The D_3 faults at Flin Flon are cut by the 1842 ±3 Ma Boundary Intrusion dikes (Heaman et al., 1992) and involve the part of the Missi Group that contains the youngest detrital zircon grain, dated at 1847 ±2 Ma (Ansdell, 1993), thus establishing a regional age and kinematics for D_3. A similar relationship exists with the Wekusko granite, southwest of Wekusko Lake, but possibly indicating later thrusting (before the 1834 +8/–6 Ma age of intrusion; Gordon et al., 1990). Some of the earliest folds in the Burntwood Group at Snow Lake and Wekusko Lake have also been interpreted as F_3 structures (F_1 in Kraus, 1998, who did not recognize pre-Missi structures).

Syncollisional D_3 structures in the Kississing–File lakes area include the northwest-trending Loonhead Lake Fault (LHF in Figure 12; Connors, 1996). This fault was traced to Kississing Lake by Zwanzig (1999) and reinterpreted to have had an early history of thrusting directed toward the Kisseynew Domain with later reorientation and generally southwesterly overturning. In this scenario, major F_3 folds, the Gohl Lake syncline and Mud Lake syncline (GOS surrounding the Batty Lake Complex, and MUS east of the Sherridon–Hutchinson Lake Complex in Figure 12) formed in the original footwall of the early Loonhead Lake thrust. It is also possible that the early thrust faults involved the inversion of syndepositional normal faults. This helps to explain the consistent synclinal structure of the Burntwood Group. The synclines open toward the northeast (present orientation), There the GOS was refolded by the F_4 Craik Lake synform (CRS), overrun by the Burntwood Group in the F_4–F_5 File River nappe, and made recumbent as a result.

The LHF truncates the northerly trending structures at File Lake. A probable interpretation is that the LHF and related folds, which involved a full thickness of the Missi Group, including the 1832 Ma Nelson Bay dome, are younger (D_4 or late D_3) than the faults to the south. The northwestern part of the fault experienced the southwest overturning, but the eastern (younger?) part of the LHF may have a simpler southwest vergence that was the cause of the overturning and thrust reversal in the northwest. This history is supported by the decapitation of the north-trending faults, including the WRNSZ east of Dow Lake, and by the consistent facing of the Burntwood Group away from the fault in the east. In general, the more westerly trending faults in the north are younger (late D_3–D_4) than the north-trending faults to the south and carried younger and older successions out of sequence toward the southwest. The earlier D_3 faults carried only older over younger, as expected for the first faults to involve the Burntwood Group. It also suggests that the East Reed Shear Zone (Syme et al., 1995; ERSZ in Figure 12), which placed the Northeast Reed ocean floor against the younger Burntwood Group, may have been a west-verging D_4 thrust similar to structures at Wekusko Lake to the east (Ansdell et al., 1999). The apparent younging of the basins and then the thrusts to the north and east are consistent with the sediment transport, facies prograding and the advance of a mountain front. The reversal in vergence from easterly to westerly would have been the result of successive collision with Archean cratons (Section 6.1.7).

North of Snow Lake (to the east of the Kississing–File lakes area), 'recumbent' D_3 folding is implied to be southeast vergent on the structure sections of Bailes (1975) but called F_1 in his Figure 13. Vergence was directed toward the area underlain by the Burntwood Group at Wekusko Lake, and toward the southeastern part of the Kisseynew Domain (Figure 5). A large early fold in the east, a tight syncline in the Missi Group, is illustrated to have been overturned to the west by D_4 folding and asymmetric, southwest-verging D_4–D_5–D_6 doming (current terminology). Thus the early tectonic transport after basinal sedimentation was apparently toward the Kisseynew Domain throughout the Flin Flon Domain. Vergence was later reversed during transposition and westerly thrusting, which was also on a regional scale.

4.1.4 D_4–D_5 deformation: terminal collision

The development of large-scale folds, small nappes and thrusts, as detailed by Zwanzig (1999), has produced the crustal structural stack in the Kississing–File lakes area. This architecture is interpreted to have formed during protracted deformation by a combination of west-northwest shortening by thrusting and sinistral shear at high crustal level, and subhorizontal, variably directed flow at depth. It was probably the result of collision and continued convergence of the Reindeer Zone with the Superior craton in the east (Zwanzig, 1999). The advanced phases (D_4–D_5) started in the Kississing–File lakes area before the ca. 1815–1805 Ma thermal peak but probably continued to the end of high-grade metamorphism ca. 1790 Ma (Figure 14). The D_4 deformation probably occurred during the prominent hiatus in magmatism between ca. 1823 and 1814 Ma, and marked a period of east-west compression and crustal thickening with westerly thrusting and fold vergence. The onset age is evident from the northerly increase in metamorphic grade overprinting the allochthonous uppermost structural sheet in the Kississing Lake area (i.e., File River nappe) as well as the File Lake synform. Large-scale asymmetric folds, mainly S-shaped pairs, below the File River nappe continued to carry the nappe piggyback after its initial emplacement. These folds show a change to more southerly vergence during a phase designated as D_5. Structures similar in style and relative age to D_4 and early D_5 have been widely recognized to have formed before the regional high-grade metamorphism also beyond the Kississing–File lakes area, such as to the northeast (Bailes, 1975). The D_5 structures, however, are not recognized in the lower grade rocks to the south.

Ubiquitous intrafolial minor folds in the well-layered units of gneiss in the study area form arrays with consistent asymmetry (Zwanzig, 1999). They indicate distributed shear over structural packages whose thicknesses were on the kilometre scale and that are typical of mid- to lower crustal flow capable of transposing early upright structures to recumbence and rotating fold axes toward the direction of flow (Williams and Jiang, 2005). The importance of recumbent folding and ductile shear in the infrastructure of orogens, rather than thrusting (with development of horses), has been stressed by Williams and Jiang (2005). LITHOPROBE seismic sections and matching geological structure sections in the Sherridon area (Zwanzig, 1993) show that the more generally accepted interpretation of reflectors with apparent cut-offs as an indication of thrusting is unlikely, but that the main structures are isoclinal recumbent folds where observed at surface. Reflectors traced to the surface have been identified as sheared lithological contacts. Where faults are present, these are older structures transposed with the layering and involved in the folding. Shear zones are commonly reactivated segments of such boundaries.

The entire F_4–F_5 recumbent fold stack is about 7 km thick and includes the prominent tongue-shaped or sheath-like structural culminations (Sherridon–Hutchinson Lake Complex and Batty Lake Complex in Figures 12, 13). These culminations of orthogneiss are separated by highly attenuated recumbent folds developed in the younger paragneiss. The attenuated folds generally have a core of Missi Group and limbs containing Burntwood Group that are in sheared contact with the orthogneiss units. Folds may show an inherited (F_3) northeasterly vergence, and an F_4 westerly and later (F_5) southerly vergence, illustrated in Figure 13a and b. The orthogneiss that developed from the Flin Flon–Glennie Complex and the younger high-grade sedimentary rocks therefore became structurally interleaved below the File River nappe and structurally inverted along with the D_3 Loonhead Lake Fault and part of the F_3 Gohl Lake syncline. The recumbent structures are interpreted to be related to the D_4 westerly emplacement of the File River nappe, but the deformation continued during further transport (D_5) to the southwest and south. Stratigraphic inversion took place above the recumbent axial surface of the F_4 Star Lake syncline (STS in Figure 13), below the orthogneiss culminations. A second, higher level of large-scale inversion is interpreted to lie directly below the File River nappe in the F_4 Craik Lake synform (CRS), probably resulting from the emplacement of the nappe (Zwanzig, 1999). The F_4 folds have amplitude to wavelength ratios of 10:1 to 40:1. They have caused attenuation and recumbence of the F_3 Gohl Lake syncline, which is preserved as an upright fold at File Lake but forms a thin, gently north-dipping refolded sheet to the north.

The reversal in tectonic transport direction, from the overriding by the margin of the Central Flin Flon subdomain onto the Kisseynew Domain during D_3 to the overriding by the metasedimentary rocks of the Kisseynew Domain onto the Flin Flon Domain during D_4–D_5, is a main feature of the kinematics of the Kississing–File lakes area. Although poorly constrained in the Kississing–File lakes area, the tectonic transport appears to have been directed west during D_4–D_5 but southwest at deeper structural levels (Figure 16a, b) and southerly during late D_5. Shortening may have remained predominantly northwest-southeast during D_4–D_5 but with intermittent constriction due to north-south compression or stain partitioning north of the sillimanite-biotite-garnet isograd.

Structures formed during D_5 include recumbent, reclined[6] and upright folds with shorter amplitudes than the earlier structures. Their relative age (F_5) is identified by their overprinting of one or both limbs of the F_4 folds (Figure 12). Shear zones that acted as detachment occur between changes in the orientation and style of these folds. In numerous areas, fold axes of

[6] A reclined fold is one in which the axis plunges down the dip of the axial surface.

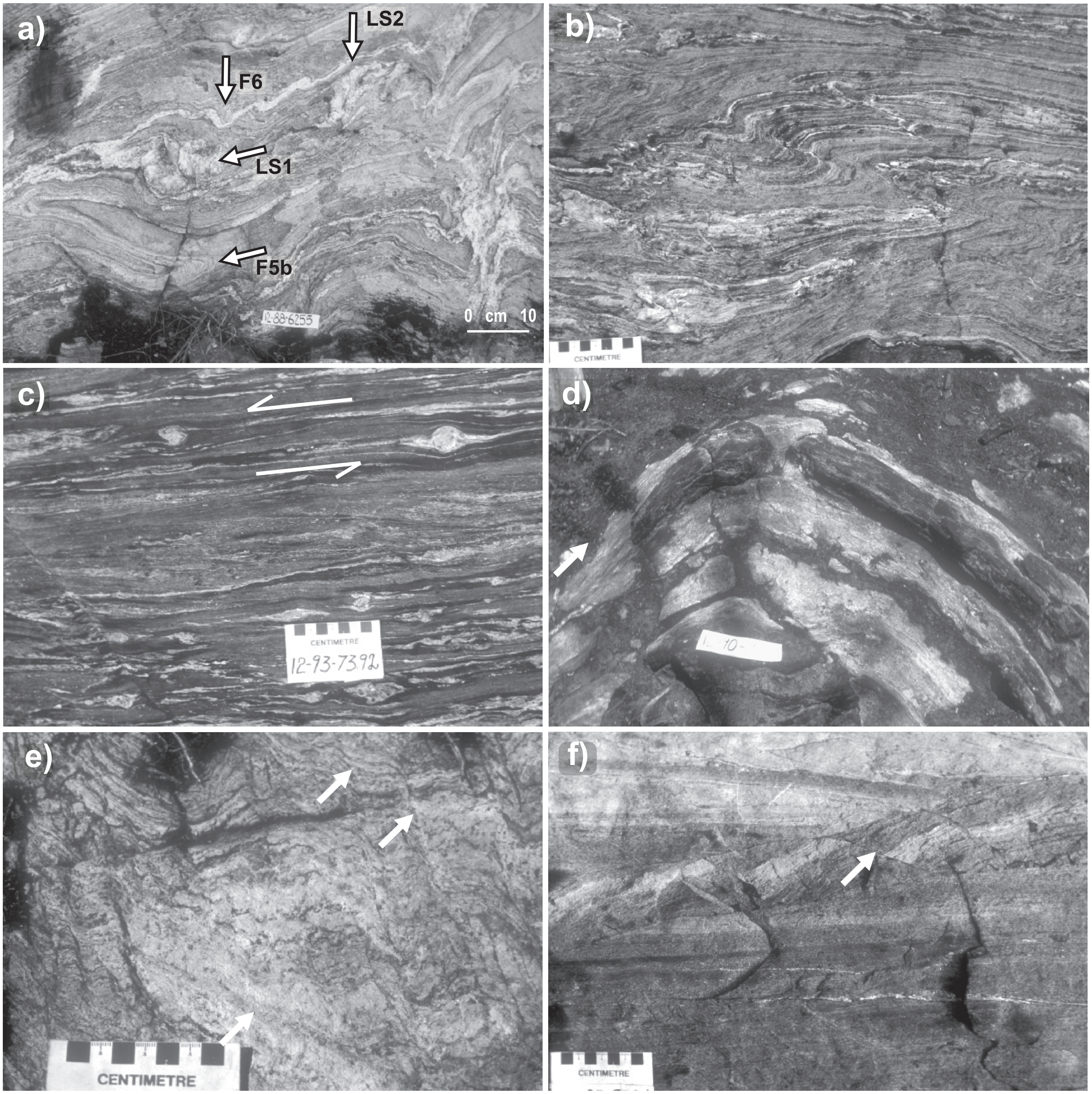


***Figure 16:** Minor structures and leucosomes (see Appendix 4, Table 6, DVD) for these and additional photographs in colour): **a)** Peak thermal F_{5b} folds involving LS1 leucosome and partly cut by LS2 leucosome refolded by open F_6 structures, Jungle Lake; tape is 10 cm long. **b)** S-shaped fold pairs and sinistral shears involving and forming with LS2 leucosome, Puffy Lake area. **c)** Sheared and rolled LS2 leucosome in late D_5 Puffy Lake Shear Zone with sinistral reverse movement. **d)** F_6 post–peak metamorphic, upright flexural fold with little or no hinge thickening but slip striations on limbs, Evans Lake; tape is 10 cm long. **e)** S_6 spaced cleavage (arrows) axial planar to open F_6 folds, Meat Lake synform. **f)** Late D_6 dextral faults (arrow) in mylonitic rocks along the Puffy Lake Shear Zone.*

several ages and lineations lie at the intersection of the D_4 and D_5 planes of deformation and represent the long axis of dome-and-basin interference structures, parallel to the direction of stretching due to superposed compression. The regional structural data (e.g., Bailes, 1975; Zwanzig and Schledewitz, 1992; Kraus, 1998; Zwanzig, 1999) have shown that the most common regional mineral and stretching lineation plunges toward the north-northeast, parallel to the intersection lineation but possibly redirected by sinistral shear. Peak thermal stretching is indicated by prominently elongate garnets with quartz-filled fractures and younger overgrowth at their long ends (Zwanzig, 1999). These lineations are consistent with north-northeast stretching of higher level crustal layers in the Central Kisseynew subdomain, as indicated also by upright west-northwest-trending granite dikes. Fold axes can be highly dispersed in the plane of flattening in some areas, but folds show a symmetry distribution that indicates regional west-northwest compression occurring at high structural levels, and are designated herein as D_{5a} deformation (e.g., Bailes, 1980a). West- and southwest-directed flow at deeper structural levels (Zwanzig, 1999) is herein designated as D_{5b}. These structures occur in the gneissic terrane north of Loonhead, Puffy and Kisseynew lakes, where they contributed to vertical flattening and subhorizontal flow, and include most of the minor structures (Figure 16a, b). West- to northwest-trending upright folds formed in the late stages of migmatite generation in the Central Kisseynew subdomain (Zwanzig, 1990), specifically in the area north of the Herblet Lake dome (Bailes, 1975) and along the transition from upright to recumbent folding north of the sillimanite-biotite-garnet isograd. These folds are herein designated as late or immature F_{5b} structures. They typically become shallow downward into east-northeast-dipping inclined or recumbent isoclines. Long-range, down-plunge projections of structures across the Kisseynew Domain (Zwanzig, 1990) indicate that the upright structures (D_{5a}) consistently overlie the recumbent folds but do not generally interfere with them. Thus, the isoclinal recumbent folds may have formed an active regional detachment to the upright structures above them and been elevated to the present level of the southern structures only during the last stage of D_5.

The D_5 deformation may have lasted for the full 40 m.y. of metamorphic monazite and zircon crystallization (Figure 14). A composite foliation ($S_{4–5}$) may have started to form early but developed continuously during the 1.82–1.79 Ga regional metamorphism (P_2 of Stauffer and Mukherjee, 1971; S_2 in Kraus, 1998; S_3 in Ryan and Williams, 1999; and S_3 Zwanzig, 1999). Regional, northerly trending, steeply dipping foliation and upright folds were developed during D_{5a} or D_4 in the greenschist- and lower amphibolite–facies rocks. They are preserved at high level in the south. They formed before the regional metamorphic peak (Bailes, 1980a, 1985; Kraus, 1998), whereas the D_{5b} structures transposed, enhanced and folded the high-grade foliation and migmatite layering formed at depth (Zwanzig, 1999). The onset of midcrustal high-grade metamorphism preceded the thermal peak in the upper crust: heat transfer is generally slow compared to the propagation of strain. Consequently, the relative age of D_{5a} and D_{5b} structures is uncertain and partly coeval. The D_{5a} style of folding and development of strong foliation were followed by a regional counter-clockwise bending of structures toward the northwest and west near the sillimanite-biotite-garnet isograd during the late stages of D_{5b}, probably as the deep structures in the north were elevated to the present higher level that existed from early on in the south.

The D_4 and D_{5a} structures are similar in geometry, age and kinematics to north-northwest-trending structures at Flin Flon that include faults, F_4 folds (e.g., Pipeline syncline; Simard et al., 2010; R-L. Simard et al, work in progress, 2010), strong foliation and lineation. They overprint the 1838 ±3 Ma Phantom Lake dikes near Flin Flon (Heaman et al., 1992) and are synmetamorphic, probably coeval with D_5 at File Lake. The faults at Flin Flon (e.g., Cliff Lake Fault) are imbricate thrusts that become shallow dipping toward the east at depth. These structures merge with the regional midcrustal seismic fabric imaged on LITHOPROBE sections (Hajnal et al., 1994). They are consistent with east-over-west transport in the infrastructure and west-northwest shortening in the suprastructure as suggested here for the Kississing–File lakes area. The thrusts at Flin Flon were later reactivated as sinistral reverse faults in a fashion similar to D_6 structures in the Kississing–File lakes area.

Kraus and Williams (1999) suggested large-scale sheath folding on the Kisseynew Domain margin. This is herein interpreted to have formed locally by flow and attenuation of domes generated by fold interference. Small sheath folds and oval fold interference at various scales are restricted to special environments, such as the hangingwalls of the long-lived shear zones; Kraus, 1998), the soles of nappes and some margins of domal structures (Bailes, 1975; Zwanzig, 1999). Sections drawn by W.A. Barclay (pers. comm., 2008) and the work of J. Krebs (pers. comm., 2008) indicate that the Sherridon structure is a well-developed sheath fold, more so than shown by Zwanzig (1999) and reproduced in Figure 13. There is likely extreme thickening in the hinges and attenuation in the limbs of the nappe-like folds. The high penetrative strain explains the shape and size of the large body of calcite-bearing calcsilicate (unit 17c) in the fold hinge 2.5 km north-northeast of Sherridon. The sheath is most likely rooted in the east-northeast, suggesting early or deep-level westerly flow, as indicated by the asymmetry and vergence in Figure 13a, as well as the divergence of fold hinges and their curving around the structure (Zwanzig and Schledewitz, 1992). South-southwest transport is indicated, on the other hand, by late D_5 kinematic indicators in the Meat Lake synform and late D_{5b} folds (Zwanzig, 1999).

4.1.5 Major D_3–D_5 structures

The uppermost structural level in the Kississing–File lakes area (File River nappe) has been interpreted as a fold nappe because it is preserved mainly as a large inverted limb in which the Burntwood Group occurs structurally above the recumbently folded Missi Group. The nappe apparently carried the Burntwood Group and its intrusions of the ca. 1830 Ma Touchbourne and Duval suites west and southwest over the Kisseynew South flank and onto the north side of the Flin Flon Domain. The intrusions contain hypersthene in the migmatitic Central Kisseynew subdomain but only amphibole as mafic mineral on the South flank. The thermal gradient must have been established after their structural emplacement along with their Burntwood Group host during D_4 because the underlying gneiss shows the same northerly increase in metamorphic grade (Zwanzig, 1999). The south margin of the File River nappe at Duval Lake in the west has retained an intermediate grade of metamorphism (Jungwirth, 1995). The underlying structural culminations of volcanic and intrusive orthogneiss (Sherridon–Hutchinson Lake Complex, Batty Lake Complex and Walton Lake nappe in Figures 12 and 13) are interpreted to have formed progressively during D_4 and D_5.

The Fairwind Lake Fault (FWF in Figures 12 and 13) occurs at the sole of the Walton Lake nappe, where it overrides the Burntwood Group. Like other 'faults' in the Kississing–File lakes area, it is a tectonostratigraphic discontinuity with a protracted history and an unknown amount of displacement. It extends southeast to File Lake, where it has been called the Beltz Lake Fault (Connors, 1996) and places highly metamorphosed pillow basalt on the northeast over the Burntwood Group. It is interpreted to be coeval and congruent with the McLeod Road thrust fault (MRF in Figure 5; Kraus, 1998), located farther to the east at Snow Lake, which also carried a panel of early metavolcanic rocks over the Burntwood Group. The earliest movement on the fault, however, may have been west over east as the early Walton Lake nappe was thrust into the Kisseynew basin and formed a klippe like the Northeast Reed assemblage. This interpretation of long-lived motion on the Fairwind Lake Fault is consistent with the presence of the volcanic rocks within the distal part of Burntwood Group. Alternatively, the early relationship may have involved a marine unconformity at the base of the Burntwood Group that was sheared to become the FWF.

Reinterpretation of (D_4) displacement along the McLeod Road Fault (MRF) is consistent with regional westerly or southwesterly thrusting and nappe emplacement. The fault dips north but cuts up-section toward the west-northwest in the hangingwall, from early volcanic rocks through the Burntwood Group and into the Missi Group. If its northerly dip at Snow Lake is regarded as the result of crossfolding, the cut-offs indicate west-directed thrusting (see, however, Kraus, 1998). Crossfolding, as suggested here, is consistent with the bend in the trace of the MRF north of Snow Lake and its southerly dip north of the Herblet Lake structural basin (HLB in Figure 5). The early fold in the footwall of the McLeod Road Fault, above the Snow Lake Fault (SLF in Figure 5; Kraus, 1998), appears to be a syncline that closes to the east. This closure is east of Snow Lake and reappears on the north side of the HLB. It indicates that the hinge trends north, parallel to the hinge of the F_4 Star Lake syncline in the Kississing–File lakes area (Zwanzig, 1999) and parallel to the strike of the fault cut-off near Snow Lake. This is normal to the direction of thrusting proposed herein, as expected for a frontal thrust ramp. The reinterpretation is consistent with the north-northeast trend of the major faults southeast of Snow Lake (e.g., the Berry Creek Fault [BCF] in Figure 5) as imbricates of thrusts; see, however, Connors et al. (1999), Ansdell et al. (1999) and Kraus and Williams (1999). Southwesterly thrusting, however, is suggested by cut-offs in the overturned northeast-dipping hangingwall of a major thrust separating the upper and lower Chisel sequences above the Lalor Lake deposit, farther to the south (Bailes et al., 2009).

Fabrics and minor structures along the Fairwind Lake Fault, McLeod Road Fault and adjacent structures are early to late metamorphic but appear to postdate the westerly emplacement of nappes. These fabrics include consistent down-dip stretching lineations and local sheath folds that indicate D_5 south-southwesterly tectonic transport (Kraus, 1998; Zwanzig, 1999). Ductile displacement along deeper structures may have been mainly during D_5 (e.g., Batty Lake Shear Zone [BSZ] and Shineton Creek Shear Zone [SCSZ] in Figure 12). Late-metamorphic (D_{5b}) amplification of the structural culminations in the Kississing–File lakes area is also indicated by the stronger recrystallization in these domes and possibly higher metamorphic grade than in the surrounding synforms. Diapirism has been suggested for the domes north of Snow Lake by Bailes (1975) and is supported by abundant boudinage and extensional shears filled with leucosome in the domes, whereas folding is concentrated in the orthogneiss around them. This suggests ballooning during amplification of the domes. Late movement on the bounding shear zones is indicated by dated synkinematic pegmatite sheets (Machado et al., 1999).

The northerly trending File Lake synform (FIS in Figure 12) was formed during D_4–D_{5a} and caused the refolding of the Loonhead Lake Fault and the Gohl Lake syncline, but is cut by the east-trending high-grade isograds (Bailes, 1980a). The FIS changes in style from an upright, south-plunging antiform on the south shore of File Lake to a steeply north-plunging synform at the north end of the lake, and a gently northeast-plunging, reclined fold (F_{5b}) in the gneissic rocks to the north. The change takes place across the isograds with the development of assemblages that include muscovite, chlorite, biotite and garnet in the south, to successive assemblages northward with 1) staurolite+biotite, 2) sillimanite+biotite, 3) cordierite+garnet, 4) sillimanite+garnet+biotite, 5) cordierite+garnet+sillimanite, and 6) sillimanite+melt. The assemblages developed at intermediate to low pressure (Froese and Gasparrini, 1975; Bailes, 1980a; Gordon et al., 1994; Menard and Gordon, 1997; Kraus and Menard, 1997). They formed mainly after the upright folding in the south, but high temperature existed already at deeper levels during the recumbent folding in the north. Longer lasting, higher grade metamorphism is a result of the deeper burial and slow unroofing in the north (M. Growdon, pers. comm., 2008). The shallowing in plunge and in dip of the axial surface of the FIS toward the north supports the conclusion that the F_3 Gohl Lake syncline (GOS) had an original northeasterly vergence and was overturned to the southwest during D_4 (Figure 17). Moderate D_6 tightening of the FIS is seen from the bending of the F_4–F_5 'axial surface' of the Batty Lake Complex farther north and from foliation curved over porphyroblasts at File Lake (Bailes, 1980a). The general change from upright to nearly recumbent folds occurs all along the northern margin of the Flin Flon Domain about 3 km north of the sillimanite-biotite-garnet isograd (Figure 5). Fold axes and lineations are transposed from steeply plunging in the south to moderately plunging in the north.

Commonly (e.g., south of Puffy Lake), the upright structures are detached from the reclined and recumbent folds along east-west D_5 mylonite zones with sinistral reverse displacement (Zwanzig and Schledewitz, 1992). These shear zones occur directly north of the zone of the curving from northerly trending

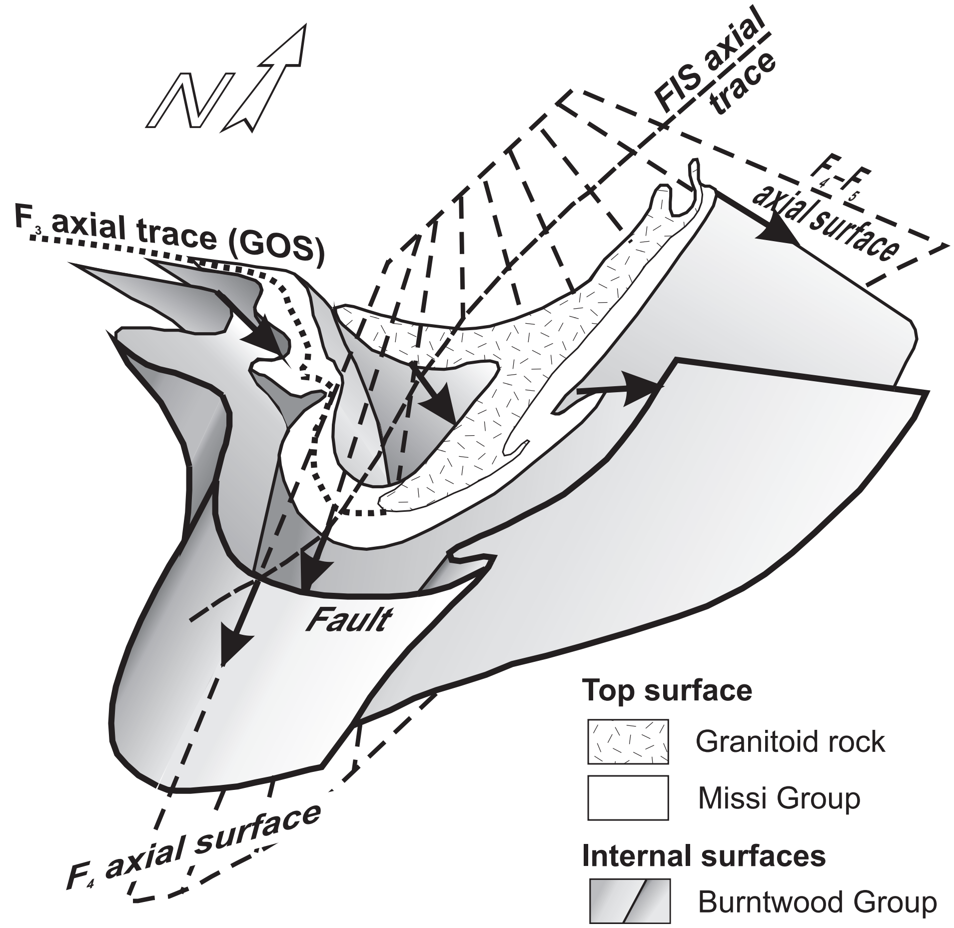


Figure 17: *Three-dimensional schematic diagram of the simplified File Lake (FIS) synform, showing the refolding of the F_3 Gohl Lake syncline (GOS) and the twisting of the F_4 axial surface (dashed lines) from upright in the south to moderately inclined in the higher grade rocks in the north. The F_4 hinges plunge south in the south and gently northeast in the north (arrows). They represent the intersection of the F_3 fold limbs and the F_4–F_5 axial surface, partly preserving the north-northeast F_3 vergence. East-trending F_5–F_6 folds occur on the west side of the structure and die out against the F_4–F_5 axial surface.*

structures to west-trending structures, close to the sillimanite-biotite-garnet isograd. This suggests that the structural change is a result of changing rheology, with rocks formed at higher temperature and deeper in the crust being exhumed on the north side of the mylonite zones during D_5.

The Puffy Lake Shear Zone (PSZ in Figure 12) is interpreted (Zwanzig, 1999) as a ductile imbricate structure that carried the deeper structural levels obliquely up and to the southwest (Figure 16c). Similar structures occur along strike (e.g., Kississing River Shear Zone [KRSZ]; Norman et al., 1995). The sense of shear is consistent with the sense of rotation and increasing P-T conditions across the structural bend in the Flin Flon Domain. A major, late F_{5b} synform (Ponton Lake synform [POS] in Figures 12 and 13) formed in the hangingwall of the Puffy Lake Shear Zone. The incompetent paragneiss in the core of the synform accommodated much of the change in structural style and detachment of the steep- and shallow-dipping structures. However, tectonostratigraphy and geochemistry leave little doubt that the volcanic units in the limbs of the synform correlate and are connected at depth (Figure 13b). South of the PSZ, directly north of the structural bend, the westerly trending volcanic rocks and their pillow structure are relatively well preserved, showing moderate compression normal to the shear zone. This is interpreted to be the result of strain partitioning with intense, non-coaxial strain (high component of simple shear) in the oblique transpression zone but mainly flexure and moderate attenuation of fold limbs to the south, where west-trending foliation developed in the plane of flattening. Areas with the same trend and good preservation also occur elsewhere within the bend in structural grain along the northern margin of the Flin Flon Domain (e.g., a small area south of Dow Lake and an area south of Kisseynew Lake in Gilbert (2004). These structures are also interpreted to be due to strain partitioning along adjacent shear zones. East-trending upright folds (late F_{5b}) occur also east of the Kississing–File lakes area in the Guay–Wimapedi lakes area (Bailes, 1975).

Upright, east-trending structures probably started to form in the late stages of D_5 all along the medium- to high-grade metamorphic transition on the north side of the Flin Flon Domain. This includes prominent structures such as the Loonhead Lake dome (LHD in Figure 12) and adjacent synform, and the Herblet Lake basin (HLB in Figure 5) north of Snow Lake, but these were amplified and became prominent during D_6 (Section 4.1.6). Upright, east-trending F_{5b} folds occur also in the transition from paragneiss to orthogneiss north of the Herblet Lake dome (HLD in Figure 5). The development of these structures during D_5 and high-grade metamorphism are indicated by late axial-planar and early folded leucosomes (Bailes, 1975). The originally recumbent structures in the dome to the south were lifted against the higher level upright structures in the north during D_6 dome amplification.

4.1.6 D_6 deformation: intracontinental transpression

Late deformation with retrogression and reorientation of peak-thermal porphyroblasts in the northerly trending foliation and refolding of the F_5 recumbent structures indicate a continuation or reactivation of southeast-northwest compression (D_{6a} in Figure 14). The File Lake structure was tightened at this time and the Walton Lake nappe was refolded by the open Meat Lake synform and Moody Lake dome (MES and MOD in Figure 12; sections in Figure 13). Whereas the recumbent and reclined D_5 structures represent midcrustal flow, the F_6 folds are flexural, generally open and locally have an axial-planar spaced cleavage (Figure 16d, e). Down-dip striations on the fold limbs and associated late foliation and shear zones in the gneissic rocks show retrogression to muscovite- or chlorite-bearing assemblages. The northerly trending cleavage associated with these structures retains microlithons, with high-temperature assemblages in the gneiss separated by lamellae containing muscovite and/or chlorite. In the zone of curving from northerly to northwest-trending folds, such as at Loonhead Lake, late, upright, west-trending folds (F_{6b}) have developed or have tightened and steepened late F_5 structures.

Ductile to brittle D_6 structures include the north-northeast-trending Nokomis Lake and Molly Lake fault zones (NOF and MLF in Figures 12 and 13). These have east-side-up sinistral displacement and have been linked to late deformation along the TNB (Zwanzig, 1990; Kraus and Williams, 1999). Late shear zones with similar kinematics occur throughout the Flin Flon Domain (Syme et al., 1995). This late-stage deformation is congruent with the folding and sinistral reverse shear in the TNB (Ryan, 1998; Zwanzig, 1999) that has been interpreted to represent a period of intracontinental transpression (Lucas et al., 1996).

The latest evidence of the D_6 deformation is faults that indicate continued northwest compression with regional northeast extension, the importance of which has not been well assessed in the Flin Flon and Kisseynew domains. Dextral fault motion with a normal component occurred during late stages on the west-southwest-trending Puffy Lake Shear Zone (PSZ in Figure 12) and the parallel western extent of the Berry Creek Fault (Syme et al., 1995; Ryan, 1998) on the south side of the Flin Flon Domain. The PSZ contains local, late, fine striations with a gentle northeast plunge in mylonite. Late minor structures indicate dextral north-side-down slip (Figure 16f). The western part of the Berry Creek Fault separates the low-grade Flin Flon Domain from the Namew Gneiss Complex to the south (Figure 5). Downthrow on its north side probably helped to preserve the highest structural levels, such as the prehnite-pumpellyite metamorphic area near Flin Flon and some of the weakly deformed rocks in the Reed Lake–southern Snow Lake area. The rapid increase in metamorphic grade to the south and the gradual increase to the north suggest that the main Flin Flon Domain is a brittle-ductile half graben. The structural bend on the north side of the Flin Flon Domain may have acted as a hinge zone. The north-northeast-trending D_6 faults, such as the NOF, do not show the reversal in slip like the PSZ but would have acted as conjugates to the dextral zones in providing northeast extension and northwest shortening. From an economic standpoint, D_6 faults are important in the preservation of the high structural levels with abundant mineralized volcanic rocks in the Flin Flon Domain.

4.2 Regional structural packages

Zwanzig and Schledewitz (1992) recognized panels comprising distinctive rock packages in the gently northeast-dipping structural stack of the southeastern THO in Manitoba. The structural packages are exposed at successively deeper structural levels in late domes between File Lake and Sherridon. Several structural panels were traced southeast into fault blocks with subvertical volcanic packages in the Morton Lake area (Zwanzig, 1996b). Many of these regional structural packages are separated by major faults or high-strain zones, as in the main parts of the Flin Flon Domain. Other panels represent the homoclinal limb of early major folds (Table 7; Zwanzig, 1999). Structural analysis has helped considerably in the recognition of individual tectonic assemblages because most are confined to one or two panels in this crustal-scale stack. Generally, assemblages do not extend from one package to another. These individual structural packages are the fundamental tectonic elements in the Kississing–File lakes area. Each package has not only a unique tectonostratigraphic succession but generally a distinctive set of intrusions that helps in tracing the complexly refolded panels. Different packages generally also contain different facies of the late successor-arc sedimentary and volcanic rocks, particularly the units at the base of the Missi Group. An analysis of the map-scale fold pattern of these major units, coupled with analysis of structural field measurements (Zwanzig and Schledewitz, 1992; Zwanzig, 1999), has provided a model of the structural geometry consistent with the polyphase deformation history of the area, as described above and shown in Figure 12. An understanding of the complex structure of the area provides an explanation for how rocks from disparate tectonic environments and with different geochemical patterns were juxtaposed. The structure also provides the rationale for tracing units (in part using geochemistry) that are locally only a few tens of metres thick over distances up to 80 km. The exposed structural stack is generally 6–8 km thick but only about 1.5 km thick between Dow Lake and Loonhead Lake. Structural attenuation and deletions in the section along faults and ductile high-strain zones have reduced the original thickness (Zwanzig, 1996b, 1999). A very similar structural style exists farther west along the northern margin of the Flin Flon Domain, at Tartan Lake, where thinning of units is also prominent (Gilbert, 1989). Delineation of the complex regional tectonic layering, even at the 10 m scale, may provide a control for VMS and gold deposits.

4.2.1 Infrastructure

The deepest structural level in the Kississing–File lakes area is a window in the core of the Sherridon–Hutchinson Lake Complex where the Sherridon gneiss is exposed below the orthogneiss derived from the early successor-arc plutons

Table 7: Major structural panels, faults and folds in the Kississing Lake–Snow Lake region.

Northern and central structural panels and folds (top to base)		Southeastern structural blocks to Snow Lake subdomain (NE to SW)	Assemblages (contact relationship)
File River nappe			Burntwood Group
Gohl Lake syncline (sequence as below but refolded)			
Walton Lake nappe			Sherridon-Meat Lake
		Herblet Lake basin	Snow Lake?, arc-rift and Missi
Fairwind Lake Fault		*McLoed Road Fault*	
Gohl Lake syncline			Missi Group
Mud Lake syncline	Structurally inverted		*(conformable / faulted)*
		Snow Lake syncline	Burntwood Group
Loonhead Lake Fault		*Snow Lake Fault*	
			Missi Group
Loonhead Lake thrust sheet			*(low-angle unconformity)*
			Four Mile Island[1] (arc, arc-rift)
		Butler Lake block	Northeast Reed (ocean floor)
		East Reed Shear Zone	
		Morton Lake block	Burntwood Group
		Morton Lake Fault	
		Dickstone block	Four Mile Island[1] (arc)
		(Podruski Lake anticline, Yakymiw syncline)	
West Reed-North Star Shear Zone			
Infrastructure			Missi Group
Central Flin Flon subdomain			*(angular unconformity)*
(Star Lake syncline)			Flin Flon[1] (arc)
			Elbow-Athapapuskow[1] (back-arc)
Sherridon–Hutchinson Lake Complex and Gants Lake Batholith			Successor-arc intrusions
Major structural break?			Structural lens of Burntwood Group
Sherridon structure			Sherridon-Meat Lake
Structural layer of Bob gneiss			Probable Burntwood Group
Shading indicates structural breaks, generally with Burntwood Group			

[1] included in Amisk Collage

(Figure 12; Zwanzig and Schledewitz, 1992). This is the (type) Sherridon gneiss, in large part identical to the gneiss in the Walton Lake nappe and interpreted as the Sherridon–Meat Lake assemblage. A narrow lens of garnet-biotite gneiss, compositionally identical and probably equivalent to the Burntwood Group paragneiss, occurs on the northeast margin of the Sherridon window. This suggests that a major structural break exists within the infrastructure (i.e., the roof of the Sherridon window is probably a D_3 shear zone). The presence of the Sherridon–Meat Lake assemblage below and above the rocks of the Amisk Collage (which form roof pendants in the Sherridon–Hutchinson Lake Complex) is consistent with the D_3 history of thrusting toward the Kisseynew Domain followed by overturning and tectonic transport toward the Sask craton. The garnet-biotite±hornblende gneiss (Bob gneiss of Tinkham, 2008) is tentatively correlated with the Burntwood Group (Section 5.9.1) This interpretation further supports an early period of thrusting of the volcanic rocks of the Flin Flon Domain over the Burntwood Group turbidite in the Kisseynew Domain.

Elsewhere, the infrastructure consists of the Amisk Collage, stitched by early successor-arc intrusions and unconformably overlain by the Missi Group. This structural package is generally upright, with intrusive rocks at the base, overlain by remnants of the Amisk Collage and capped by the Missi Group. A relatively thick, north-facing succession of Amisk Collage and Missi Group, south of Ponton Lake, is correlated with a structurally thinned, equivalent, south-facing succession that is truncated at the base along the Archie Lake pluton north of Ponton Lake (columns 2A–2A' and 2B in Figure 8b, back pocket). These sections represent the limbs of the major F_5 Ponton Lake synform (POS in Figure 12). In many areas, however, the Missi Group lies directly on the early gneissic intrusions (e.g., column 4A in Figure 8b).

Structural culminations (Sherridon–Hutchinson Lake Complex, and Adamson Lake and Big Island domes) comprise units of gneiss derived in the infrastructure from the successor-arc plutons with enclaves and sidewalls of amphibolite and intermediate gneiss that were probably derived from rocks of the Amisk Collage. The Batty Lake Complex and possibly the

Defender Lake dome may have Sherridon-age supracrustal mantles. Clearly recognizable early volcanic rocks occur only on the southern margin of the Sherridon–Hutchinson Lake Complex. The most southerly part of the Amisk Collage that was inverted during D_4 occurs above the recumbent axial surface of the Star Lake syncline (STS in Figure 13). Structural panels that were inverted during much earlier deformation (D_1) but later steepened occur east of the West Reed–North Star Shear Zone (WRNSZ in Figure 12).

4.2.2 Southeastern structural blocks

Structures east of the West Reed–North Star Shear Zone comprise north-trending, steeply dipping early assemblages separated by narrow belts of the younger Burntwood Group successor-basin deposits. A thick (<1800 m) volcanic succession (Bailes, 1980a) makes up the northern part of the Fourmile Island arc assemblage in the Dickstone block (Figure 12), where it forms the east-facing limb of the large F_1 Yakymiw syncline (YKS in Figure 12; column 9A in Figure 8c, back pocket). The west-facing limb adjacent to Morton Lake is thinner, heavily intruded by Josland Lake sills and cut off by the Burntwood Group along the Morton Lake Fault (MOF in Figure 12; column 9B in Figure 8d). The folded volcanic succession extends north to Dow Lake, where the plunge changes from east to northerly across a saddle structure between domes at Dow Lake and Loonhead Lake (Figure 12). Part of the Fourmile Island assemblage is exposed again in the northwest and continues to Evans Lake and beyond, in the Loonhead Lake thrust sheet.

The Northeast Reed (ocean-floor) assemblage is juxtaposed along the prominent East Reed Shear Zone against the Burntwood Group in the fault slice at Morton Lake (ERSZ in Figure 12). The Fussey Lake basalt unit in that assemblage is more than 3000 m thick in the Butler Lake block. The basalt is overlain by a 600 m thick sill of Josland Lake gabbro (column 9C–9C' in Figure 8c). The units occupy large northerly trending folds that affected a Josland Lake sill and may be late F_1 structures. The measured section is believed to lie on a single west-facing fold limb, but there are few top indicators in the lower part of the section, so repetitions cannot be ruled out.

4.2.3 Loonhead Lake thrust sheet

A northwest-tapering thrust sheet (Loonhead Lake thrust sheet) comprises early volcanic rocks belonging to the Fourmile Island assemblage, Josland Lake sills and overlying Missi Group. This rock package extends from Dow Lake to Kisseynew Lake and north to Jungle Lake (Zwanzig, 1999). The thin basal Missi conglomerate in the thrust sheet appears to mark only a low-angle unconformity. This relationship is suggested by the apparent stratigraphic concordance between 1) a distinctive basalt unit (Moody Lake basalt) in that part of the Fourmile Island assemblage, 2) the Josland Lake sills that intrude it, and 3) the stratigraphically overlying Missi Group. The consistent topping of the sills toward the Missi Group supports this suggestion. Alternatively, the volcano-intrusive panel may represent the upright limb of a large F_1 recumbent fold whose lower limb was cut off by the Loonhead Lake Fault. In either interpretation, the package was little affected by D_2 steepening.

The fault brings the original base of the thrust sheet into contact with the Burntwood Group. Much of the thrust sheet is presently inverted in the lower limbs of the large southwest-verging recumbent F_4–F_5 folds. However, the thrust sheet is rooted in the thick part of the Fourmile Island assemblage in the Dickstone block, where there are also prominent Josland Lake sills. The thrust cuts up-section toward Kississing and Kisseynew lakes, where the volcanic rocks in the sheet taper to a total thickness of less than of 10 m northwest of Lobstick Narrows and at Jungle Lake (compare columns 1C and 9B in Figure 8d, back pocket).

4.2.4 Gohl Lake and Mud Lake synclines

The third structural level is the large F_3 Gohl Lake syncline (GOS in Figure 12), occupied by only the distal facies of the late successor-arc units (Figure 8e, back pocket). Approximately 700 m of the Burntwood Group greywacke-mudstone turbidite and 800 m of overlying Missi Group sandstone and sills occur in the core of the syncline north of File Lake. The lower contact of the Missi Group is interpreted as gradational with Burntwood Group pelite. Basal Missi Group quartzite or lithic arenite extend northeast to Jungle Lake, where they are exposed along a logging road east of the lake. A similar, distal stratigraphic section is exposed also in the southern and eastern mantle of the Sherridon–Hutchinson Lake Complex, in the Ponton Lake synform and the Mud Lake syncline, respectively (POS and MUS in Figure 12).

4.2.5 Walton Lake and File River nappes

A thick succession of Sherridon gneiss and local amphibolite is in sharp structural contact with the distal facies of the Burntwood Group along the Fairwind Lake Fault, in the central part of the Kississing–File lakes area. This small F_4 fold nappe, which evolved into an F_4–F_5 thrust nappe (Walton Lake nappe), extends from File Lake (Folster Bay) as a thin fault sheet northwest to Batty Lake and, from there, thickens to an estimated 1600 m in the core of the nappe (columns 5 and 9E in Figure 8e, back pocket). Nearly identical felsic gneiss occurs in the core of the Sherridon window (Froese and Goetz, 1981; Zwanzig and Schledewitz, 1992) at the deepest structural level (Table 7). The uppermost structural panel (File River nappe) contains a thick succession of Burntwood Group migmatite and late intrusions, typical of the Central Kisseynew subdomain, where the nappe is rooted.

4.3 Metamorphism

The northerly increase in metamorphic grade from medium to high and to migmatite is a complicating feature that accompanies the change from predominantly volcanic to sedimentary supracrustal rocks from the Flin Flon Domain to the Kisseynew Domain. The higher grade of regional metamorphism has an affect not only on the structural style (see above), but also on the fundamental characteristics of all units, whether igneous or sedimentary. Clearly recognizable primary rock types in the south have changed to coarsely recrystallized ortho- and paragneiss at the northern margin of the Kississing–File lakes area. Neither the high metamorphic grade nor the complex folding, however, has destroyed mineral deposits or hydrothermal alteration domains. In fact, the latter are typically made more obvious by coarse recrystallization (Froese, 1989; Bailes and Galley, 1999). Because geochemistry is such an important tool for tracing units and determining tectonic and metallogenic environments, a better understanding of the high-grade metamorphism leads to more effective mineral exploration.

Changes in bulk chemical composition are evident in certain rocks, which have developed lit-par-lit structure. Migmatization occurred largely in open systems (Zwanzig and Shwetz, 1994b; Jungwirth, 1995; Jungwirth et al., 2000) and, as a consequence, interpretation of the rock types and their geochemistry must take into account the metamorphic grade and presence of leucosome. The sporadic to abundant veining with quartz or granitoid leucosome, which occurs in some of the high-grade rocks, required extreme care in geochemical sampling if the results were to be meaningful. Units that have granitoid veining, such as the Burntwood Group, Missi Group, K-rich andesite and rhyolite, are invariably altered in the north. Units of amphibolite derived from mafic volcanic and intrusive rocks may have thin veins of quartz-plagioclase leucosome, but these can generally be avoided during sampling. In general, the best data were obtained from the southern third of the area, south of the migmatite front (Figure 18), although good preservation of units is locally present in the north. Chemical plots in Section 5 concentrate on the least fluid-mobile elements and the most refractory rock types in order to address the problem of metamorphic alteration. Scatter diagrams and regression were used to test for self-consistency. Outlying analyses are either shown with special symbols or eliminated from the plots (but included in tables). Our results indicate that the high-grade metamorphic rocks have retained many chemical patterns and ratios that are consistent with those of the adjacent low-grade rocks or with published analyses of fresh rocks from modern environments. One of the important findings regarding migmatization is that melting occurred north of the first occurrence of leucosome. Sharp contacts of the granitoid sheets at the migmatite front suggest limited interaction with the hostrocks there.

Some felsic rocks (arkose, rhyolite, rhyodacite and granite) are expected to have changes in bulk composition during high-grade metamorphism. For example, a thin stratabound sheet of bright pink granite in the Missi Group north of the cordierite-garnet-melt isograd (13 in Figure 18) may be intrusive or *in situ* remelted rhyolite. Slabs of foliated metarhyolite, taken from areas north of the migmatite front, that were etched with HF and stained for K-feldspar show strong layering not evident in

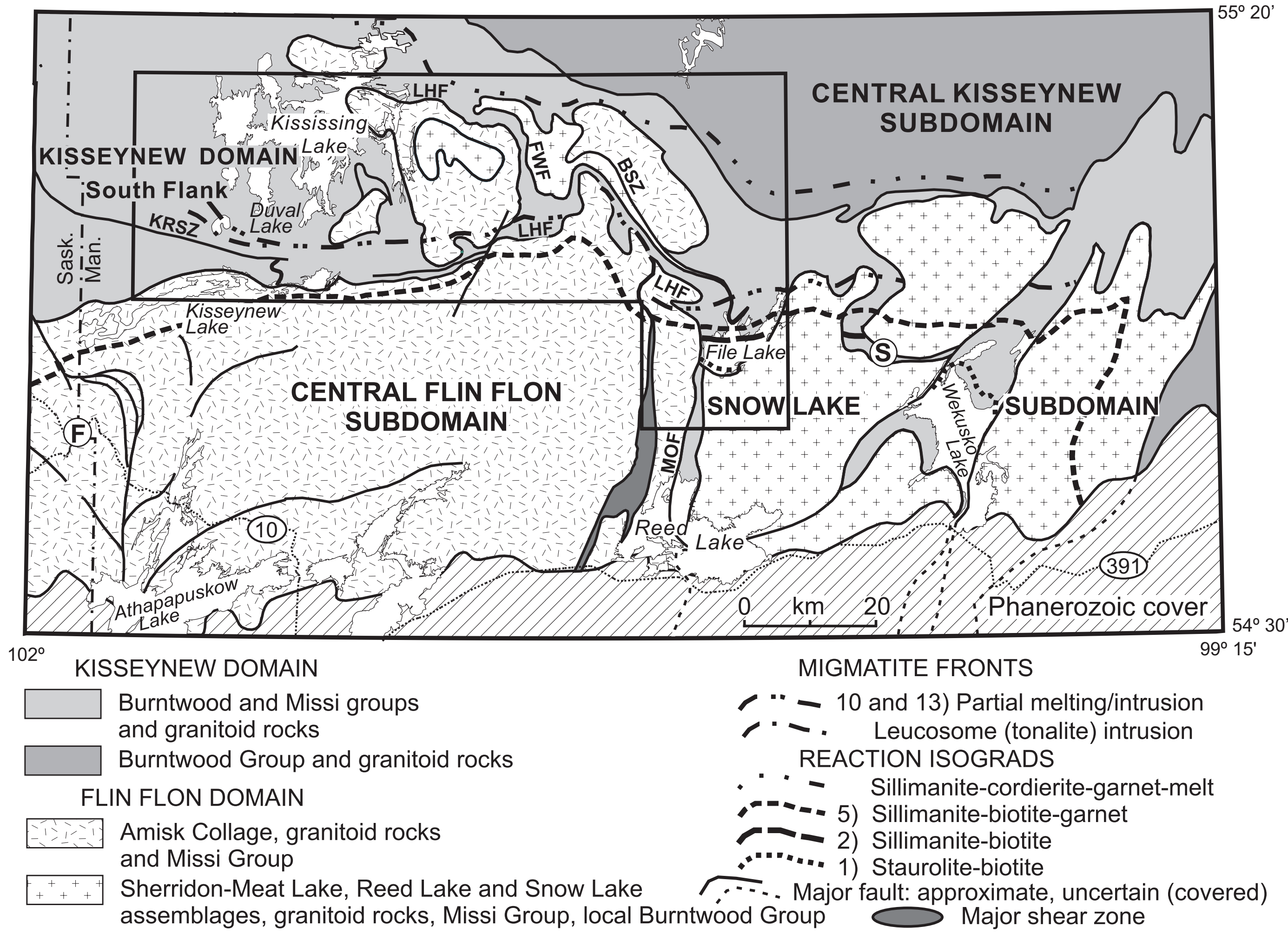


Figure 18: *Reaction isograds from intermediate to high grade and the line of first melting in the Kississing–File lakes area. Numbers in the legend are cited in the text. Abbreviations: BSZ, Batty Lake Shear Zone; F, Flin Flon; FWF, Fairwind Lake Fault; KRSZ, Kississing River Shear Zone; LHF, Loonhead Lake Fault; MOF, Morton Lake Fault; S, Snow Lake.*

the field. Relatively coarse layers comprise mainly plagioclase and quartz, and these are interpreted as leucosome. Directly at the migmatite front, pegmatite lenses with thicknesses between 0.1 and 10 m are locally very prominent, some replacing part of a rhyolite unit. Geochemical data suggest that the pegmatite is a leucosome representing a fraction that may have been removed from the host rhyolite or adjacent sediment (Section 5.10.4.1). Clearly visible leucosome occurs in some areas of K-rich andesite of the Missi Group. These observations indicate that extreme caution must be exercised in making interpretations based on data from such rocks.

4.3.1 Contact metamorphism

Contact metamorphism was documented (Bailes, 1980a) for several granitoid plutons and sills in the File Lake area where the regional metamorphic grade is sufficiently low. Contact metamorphic aureoles have been described for the Reed Lake pluton, Norris Lake pluton and Josland Lake sills. The volcanic rocks are little affected by the Josland Lake gabbro or any effects are obscured by later metamorphism. Metasomatic alteration and local digestion of the Yakymiw Formation occur in the roof zone of the largest sills. The foliated and gneissic intrusions farther north are highly recrystallized, most showing granoblastic fabrics that have overprinted and destroyed any contact effects.

The northern (intrusive) contact of the Reed Lake pluton at Woosey Lake has a partially melted to agmatitic contact zone with abundant rafts of mafic volcanic rock (Bailes, 1980a). Burntwood Group metagreywacke contains andalusite-biotite–bearing assemblages up to 1 km from the contact, and sillimanite adjacent to it. Bailes (1980a) concluded that this represents the roof zone of the intrusion. The assemblages suggest that P-T conditions were >500°C and near 4 kb. The Norris Lake pluton has an observable contact aureole directly south of the Kississing–File lakes area (Bailes, 1980a). These plutons are part of the late successor-arc intrusions and probably started to contribute to the regional high heat flow—another explanation for the wide metamorphic aureole.

4.3.2 Regional metamorphism

Regional metamorphism along the margins of the Flin Flon and Kisseynew domains was determined mainly from the metamorphic assemblages and migmatitic structure in the Burntwood Group, which has a history of thorough studies (Currie, 1947; Pollock, 1964; Moore and Froese, 1972, Froese and Gasparrini, 1975; Bailes and McRitchie, 1978; Bailes, 1975, 1980b, 1985; Froese and Moore, 1980; Gordon, 1989; Zwanzig and Schledewitz, 1992; Gordon et al., 1993, 1994; Jungwirth, 1995; Froese, 1997; Menard and Gordon, 1997; Kraus and Menard, 1997; Kraus, 1998; Kraus and Williams, 1999; Zwanzig, 1999; Jungwirth et al., 2000). Data from the most pertinent of these studies are briefly reviewed in this section and the evolution of some of the migmatitic and veined rocks is discussed, with emphasis on the element mobility that has affected the composition of the units.

An early period of greenschist-facies regional metamorphism in the Burntwood Group (M_1 in Bailes, 1980a) preceded the D_{5a} folding. This may have been simply the low-grade onset of the main protracted period of metamorphism. Alternatively, the limited data cited by Leclair et al. (1997), namely titanite (cooling) ages of 1828 ±3 to 1819 ±4 Ma for the high-grade Namew Gneiss Complex directly south of the Central Flin Flon subdomain (Figure 5), may indicate early regional metamorphism. New tectonic considerations in this report (Section 6.1.7.2) suggest that this moderately higher pressure (<8.5 kb), high-temperature gneiss complex surrounded by andalusite-bearing gneiss may have started to develop as an early (>1.83 Ga) thermal dome, back-arc core complex or half-horst, possibly coeval with the onset of doming in the Herblet Lake dome and Sherridon–Hutchinson Lake Complex (HLD and SHC in Figure 5). Higher level block faulting, as suggested by the abrupt lateral facies changes in the Missi Group (Section 3.5.1.2), and thermal effects from late plutons (e.g., Reed Lake pluton [RLP]) may have been associated with late Missi-age (M_1) metamorphism.

The present reaction isograds formed during the thermal climax (M_2 in Bailes, 1980a), which is presently considered to have been part of a protracted period of high heat-flow from ca. 1820 Ma to ca. 1800 Ma (e.g., Lucas et al., 1996). This period of high heat flow and recrystallization represents the main part of the Trans-Hudson orogeny with abundant metamorphic and synkinematic (D_5) igneous monazite and zircon formed between 1812 and 1803 Ma (Figure 14). Bailes (1980a) divided the Burntwood Group at File Lake into five M_2 metamorphic assemblage zones, shown by isograds based on discontinuous reactions (1, 2, 5, 10 and 13 in Table 8 and Figure 18).

The relative age of M_2 is understood within the limits of our current knowledge of the Trans-Hudson deformation history (Table 5) and the progressive evolution of foliation defined, in the south, mainly by biotite and muscovite alignment and concentration. Veins, sheets and lenses of leucosome (LS2), commonly encased in melanosome, are aligned with the biotite foliation in mesosome in the north (Zwanzig, 1999). In the south, the M_2 isograds cut the north-trending F_4 major fold (e.g., File Lake synform [FIS] in Figure 12). The M_2 porphyroblasts are interpreted to have been deformed during D_4–D_{5a}, as designated in this report. Porphyroblasts include a straight to moderately curved internal foliation, interpreted herein to have formed during D_3–D_4. In the south, the northerly trending foliation cuts bedding in the Burntwood Group in F_3 fold closures (Bailes, 1980b) but curves slightly around the porphyroblasts. This late strain is interpreted to have occurred during D_{5a} and continued during D_6. Farther north, sillimanite aggregates (faserkiesel) are flattened in the F_5–F_6 axial planes. In most areas, the isograds are parallel to the F_{5b} structures, which formed during the metamorphic peak. In the northern half of the Kississing–File lakes area, veins and sheets of somewhat younger leucosome (LS3) were openly to tightly folded with the main foliation during D_{5b} and D_6. Biotite-bearing rocks in the cores of tight F_5 folds have developed an axial-planar foliation during the late stage of D_5. In shear zones, a distinctive greenschist-facies overprint occurred during D_6 block uplift and cooling.

4.3.2.1 Reactions

The M_2 reactions that define metamorphic zones at File Lake are shown in Table 8 (reactions 1–10). Reactions 1–5 take place between about 500 and 600°C and between 3.5

Table 8: Compilation of isograds from the Kississing–File lakes and Snow lake areas. References are given in the text.

	Reaction	Lithology
1)	Chlorite+muscovite+garnet → staurolite+biotite+H2O	Greywacke-mudstone
2)	Chlorite+muscovite+staurolite → sillimanite+biotite+H2O	Greywacke-mudstone
3)	Chlorite+sillimanite+quartz → staurolite+cordierite+H2O	Greywacke-mudstone
4)	Chlorite+staurolite+quartz → cordierite+garnet+H2O	Greywacke-mudstone
5)	Muscovite+staurolite+quartz → sillimanite+garnet+biotite+H2O	Greywacke-mudstone
6)	Staurolite+quartz → cordierite+garnet+sillimanite+H2O	Greywacke-mudstone
7)	Muscovite+plagioclase+quartz → K-Na–feldspar+sillimanite+H2O	Mainly arkosic sandstone
8)	Muscovite+plagioclase+quartz+H2O → sillimanite+melt	Mainly arkosic sandstone
9)	Quartz+plagioclase+H2O → tonalitic melt	Felsic rocks
10)	Na-K–feldspar+plagioclase+quartz+H2O → granite melt	Felsic rocks
11)	K-feldspar+garnet+cordierite+H2O ← biotite+sillimanite	Greywacke-mudstone
12)	White mica+albite+quartz ← sillimanite+melt	Felsic rocks
13)	Biotite+sillimanite±plagioclase+quartz → melt+cordierite+garnet±K-feldspar	Greywacke-mudstone

and 5 kb, according to the reaction grid compiled by Froese (1997) and pressure-temperature (P-T) work by Gordon et al. (1993, 1994). At the eastern end of the Kisseynew Domain, northeast of Wekusko Lake, reaction 2 produced andalusite, indicating that the pressure was about 3 kb (Bailes, 1985). Reaction 4 occurs below 5 kb. In muscovite-free assemblages, staurolite is consumed by reaction 4 but, in some compositions of muscovite-free rocks, staurolite may persist beyond reaction 5 up to reaction 6, forming fibrous sillimanite knots.

Recrystallization of arkosic sandstone in the Missi Group involves mainly coarsening of the grain size, nearly until the generation of large quartz-sillimanite knots (faserkiesel) by the muscovite-consuming reaction 7 at <3.5 kb, and undersaturated melting (reaction 8) at higher pressure. The last reactions occurred at about 700°C and 6 kb, and may buffer the temperature in the Kisseynew Domain migmatite complex. The reactions are also expected to take place in Burntwood Group pelite, but early leucosome in the greywacke-migmatite is tonalitic and sillimanite decreases (Bailes, 1980a) rather than increases with increasing melt fraction (Section 4.3.2.2).

4.3.2.2 Migmatite front

The beginning of a migmatite zone was described by Jungwirth (1995) from the Duval Lake area, north of Kisseynew Lake, and was mapped by Zwanzig and Shwetz (1994a, b) east of Nokomis Lake. Certain observations in these areas (Figure 18), which appear to apply along the entire migmatite front in the Burntwood Group, are consistent with observations and experimental considerations from migmatite terrains in other parts of the world (e.g., Fornelli et al., 2002). Moreover, the Duval Lake area has been interpreted (Zwanzig, 1999) to contain the edge of the File River nappe, emplaced at a relatively high crustal level and reactivated along a basal shear zone (KRSZ in Figure 7) during D_{5b}. The Nokomis Lake area lies at a deeper structural level, below the Walton Lake nappe, originally below the Loonhead Lake Fault and presently adjacent to the long-lived (D_5–D_6) Nokomis Lake Fault (NOF in Figure 7). The distribution, history and origin of leucosome and granitoid sheets and lenses are of particular interest because they demonstrate the existence of a major intrusive component in the migmatite with a range of ages and implications for heat flow during the M_2 metamorphism. The way that the migmatite developed impacts our understanding of the origin of larger synkinematic plutons farther north. Moreover, changes in bulk chemical compositions due to the high-grade metamorphism and metasomatism can be studied in these areas.

A melt isograd is shown by Jungwirth (1995) at Duval Lake, but a series of mineralogical changes there have been tentatively attributed to metasomatism. A northerly increase in metamorphic grade is indicated only by coarsening grain size and the appearance of thinly stromatic (layered) migmatite. Staurolite, generally with sillimanite in the absence of muscovite, persists in a distinctive unit and no cordierite is reported from the area. This suggests that the Duval Lake area is above reaction 5 and muscovite has been consumed in some compositions of the Burntwood Group before staurolite, but reaction 6 has not been passed in some compositions. Melting may have occurred by reactions 9 and 10. Jungwirth (1995) has favoured metasomatic alteration and folding to form the local distribution of units. However, the mappable units at Duval Lake have the characteristics and repetitive distribution that are common throughout the Kisseynew Domain. Thus, units at Duval Lake can also be interpreted as primary changes in the turbidite lithology involving the original ratio of clay and quartzofeldspathic sand. At Granville Lake on the North flank of the Kisseynew Domain, where such units have preserved primary structures (Zwanzig, 1990), the psammitic units are clearly coarse grained and thick bedded, whereas the more pelitic units comprise alternating Al-rich pelite and greywacke-siltstone. One reason for near-isochemical reactions may be that much of the tonalite melt was intruded and crystallized quickly, reacting little with the wallrocks because melting of the entrained minerals requires considerable heat and leads to a rapid increase in crystallinity and viscosity (Glazner, 2007).

A unit that is garnet and staurolite free, however, contains veins and sheets of granodiorite to leucotonalite that have sharp intrusive contacts. There are also thick layers of metagreywacke with little or no granitoid rock but commonly containing sillimanite. The adjacent staurolite-bearing units are interpreted herein as metapelite. These units structurally overlie other well-preserved, relatively low grade rocks of the Missi Group, which occur south of the Kississing River Shear Zone (sole of the File River nappe). The well-preserved Burntwood Group units lie adjacent to and beneath stromatic migmatite (metatexite), much as at Granville Lake. The preservation of sharp contacts and the similar composition of tonalite sheets and early leucosome (LS1) in stromatic migmatite indicate that the tonalite veins and sheets are largely intrusive and have not mixed with most of the sedimentary rocks. The early tonalite generally forms <10% of outcrops on the Kisseynew flanks. Detailed studies (e.g., Fornelli et al., 2002; Cruciani et al., 2007) have indicated that such a component of tonalitic leucosome has formed from greywacke and pelite by the fluid-present melting reaction 9 at 700° and 5–6 kb, similar to conditions near the migmatite front in the Kississing–File lakes area. This reaction was also adopted for the Burntwood Group in the Duval Lake area (Jungwirth, 1995) and farther east by Froese (1997). Sharp boundaries on granite sheets in vein-free rocks indicate intrusion directly above the migmatite front. Melt was apparently generated at a deeper crustal level and injected up the northerly dipping foliation and layering. Where the migmatite front has been studied in detail at Nokomis Lake (e.g., Zwanzig and Shwetz, 1994b), tonalite to granodiorite generally forms sheets and thin leucosome (LS2), but thicker, less deformed sheets of granite and granodiorite are coeval with younger (LS3) leucosome.

The mafic minerals in the tonalite–granodiorite leucosome are variable but include the same phases as the hosting mesosome and melanosome. They comprise biotite, garnet±cordierite and accessory graphite, tourmaline, titanite, ilmenite, monazite and zircon. They are interpreted as partly entrained from the mesosome or melanosome and, where subjected to microprobe analysis, have the same chemical composition as in those components (Fornelli et al., 2002). Thin veins of tonalite can remain largely *in situ* and are more common in pelite than in psammitic greywacke. Similar veining in the arkosic arenite of the Missi Group or in rhyolite consists of granite. This leucosome probably formed by a fluid-induced melt reaction involving K-feldspar (reaction 10).

Discrete leucogranite to granodiorite intrusive sheets can occur in any proportion to the country rock. Previous studies have assumed all these to be derived from *in situ* melts because they are intimately associated with rafts of the Burntwood Group and generally appear above the first occurrences of sillimanite (e.g., Bailes, 1975, 1980a). However, even where early melting occurred, such leucosome may be partly allochthonous, formed by dehydration melting with biotite breakdown at orthopyroxene stability (~800°C and 7–8 kb) during the generation granulite restite (Fornelli et al., 2002). This is below the depth of migmatite occurrence in the Kississing–File lakes area. Experiments in fluid-present metapelite melting (Patiño Douce and Harris, 1998) produced sodic melts at 6 kb and potassic melts at 10 kb. Orthopyroxene, which indicates biotite breakdown, is a rare but widely distributed phase in the Northeast Kisseynew subdomain (Growdon et al., 2006). Scattered orthopyroxene throughout the heavily intruded Central Kisseynew subdomain suggests that this mineral was widespread there too before reacting with fluids left after granite crystallization. Orthopyroxene occurs with biotite in the Burntwood Group at all known localities in the Kisseynew Domain. A granulite massif, however, is suspected at depth. The granite sheets also contain the same mafic phases as the country rock, and these grains are interpreted as partly entrained from the country rock. Whereas granite hosted in the Missi Group contains magnetite, tonalite–granodiorite sheets in the Burntwood Group may contain graphite. Strongly poikiloblastic textures suggest late magmatic to subsolidus growth of garnet and cordierite. Much of the granitoid rock was probably never entirely fluid. The melanosome is apparently enriched in garnet and cordierite like some of the distant orthopyroxene-bearing gneiss, and this argues for a partly *in situ* and partly allochthonous origin for migmatite in the north.

The widespread decrease in the abundance of sillimanite in greywacke mesosome and melanosome across the migmatite front, and the lack of increased sillimanite with increased leucosome, requires an explanation. Bailes (1985) suggested the sillimanite-consuming retrograde reaction 12 and the hydration melting reaction 13, which also consumes biotite. Reaction 12 occurred in the upper part of the migmatite complex, with the consumption of sillimanite and crystallization of 'retrograde' muscovite (*sensu stricto* white mica), during either injection or crystallization. This would explain the abundant late muscovite in parts of the Kississing–File lakes area. White mica replaced much of the fibrous sillimanite in faserkiesel throughout the Kisseynew Domain. Plagioclase is partly replaced by sericite in most parts of the complex. Similarly, the replacement of garnet by biotite was part of back-reaction. Back-reaction probably occurred repeatedly as successive melts crystallized.

There are up to four or five phases of leucosome distinguished in many areas of the Kisseynew Domain. These are marked by crosscutting relationships in which the older sheets show progressively tighter folds and stronger foliation. Mechanisms that drive the periodic crystallization may be 1) intermittent pressure release as an interconnected stockwork of leucosome breaks into higher levels due to hydraulic fracturing; and/or 2) periodic cooling in D_5 nearly recumbent antiforms that have carried migmatite into higher crustal levels; and, most importantly, 3) the high heat requirement of reaction with wall-rock and entrained grains (Glazner, 2007), which buffered the temperature of the migmatite complex. The field relations of the leucosome suggest that tonalitic leucosome may have dominated during the early stages of M_2 whereas granite rose higher in the crust during hydraulic fracturing in the later stages. The sharp-walled sheets of tonalite to the south of Duval Lake may have experienced both conditions as the File River nappe climbed obliquely up the ramp marked by the Kississing River Fault. Migmatite in the Nokomis Lake area was transported first (D_5) up and to the southwest, and later (D_6) up and to the north. In general, leucosome formation was diachronous with the slow advance of the migmatite front buffered by the high heat budget required for melting. Thus, the age relationships among the various leucosomes in the Kississing–File lakes area are purely local.

A significant decrease in the abundance of sillimanite or its complete consumption, however, occurs in the absence of secondary muscovite in the northern part of the Kississing–File lakes area. This takes place by reactions 11 and 13 over a wide zone near the migmatite front (Figure 18). Conspicuous diatexite makes an appearance in the Burntwood Group in this zone. This is a type of relatively uniform migmatitic mesosome (or neosome) containing large porphyroblasts of plagioclase, garnet and cordierite±K-feldspar, and, in the adjacent granite sheets, large crystals of microperthite K-feldspar and sieve-textured garnet and cordierite. A decrease in biotite content compared to the rocks to the south and the closeness to orthopyroxene-bearing intermediate compositions (e.g., enderbite) indicate that the prograde biotite-consuming reaction 13 is important here. This is the most southerly region of notable granite melt generation by dehydration of pelitic rock in the Burntwood Group and its migration in veins of leucosome. Granite melt produced in reaction 10 in originally K-feldspar–bearing rocks of the Sickle Group and Sherridon gneiss occurs up to 15 km farther south and at lower temperature. In the absence of significant amounts of H_2O, however, reaction 13 is interpreted to have started in the Sherridon gneiss but with less visible segregation and probably much less migration. Nevertheless, major- and trace-element abundance may have been affected, particularly in the deep-level Sherridon structure. The reaction may have been restricted to areas with Mg increase due to early hydrothermal alteration.

Where late metasomatic alteration is documented (Zwanzig and Shwetz, 1994b), there is a very close positive correlation between the amount of granitoid material and prominent dark green biotite, which replaces garnet but often leaves small cores. This is considered to reflect the amount of fluid generated during granite crystallization. Areas with alteration are commonly associated with shear zones and late pegmatitic granite that provided the alkali-bearing fluid during crystallization. This has occurred, for example, in the hangingwall of the Nokomis Lake Fault. The wider presence of late coarse-grained muscovite at Duval Lake and elsewhere indicates metasomatism at the migmatite front. Such altered rocks were avoided during geochemical sampling. The injection migmatite clearly has lost its original bulk composition. The lack of mixing of mesosome and leucosome, however, suggests that some layers may have been little affected because fluid flow was channelled in the veins. Protracted flow in the veins is suggested by the transition from quartz filling in low-grade areas to quartz with tonalite rims and, farther north, tonalite with boudins of vein quartz. In the main migmatite complex, tonalite veins that are separated into widely spaced boudins form the oldest leucosome (Zwanzig, 1999).

The origin of muscovite near the migmatite front is problematic (compare Bailes, 1980a; Zwanzig and Schledewitz, 1992; Jungwirth, 1995). Currently, we prefer the interpretation of Jungwirth (1995) that muscovite is secondary, and we suggest that it crystallized from fluids produced by solidifying granitoid magma and leucosome. This suggests that SiO_2, K_2O, Na_2O and associated minor elements were transported to the migmatite front from the granulite at depth and that the present composition of rocks near the migmatite front is suspect. Avoiding schistose samples and those with visibly secondary muscovite or biotite is thus an important precaution in geochemical sampling. Conclusions made from fluid-mobile elements should be avoided or treated with great caution.

4.3.2.3 Pressure-temperature conditions

The various discontinuous M_2 reactions seen in the rocks of the Kississing–File lakes area provide an approximate metamorphic field gradient (P-T conditions across the isograds at the present level of erosion), as plotted on a metamorphic reaction grid (Bailes, 1980a; Froese, 1997). For the Snow Lake and File Lake areas, this gradient appears to pass near the triple point of the aluminum silicates shown at 525°C and 4.2 kb by Froese (1997). The peak conditions are about 700°C and 6 kb in the range of medium- to low-pressure metamorphism. The illustrations of Froese (1997) from the Weldon Bay area show that, at lower temperature conditions, a higher pressure is suggested than at Snow Lake, although this may be arbitrary. There is little control on pressure.

Geothermobarometry values obtained near Snow Lake using the TWQ 1.02 program with thermodynamic data from Berman (1988, 1991) by Kraus (1998) provide medians of 550°C and 4.3 kb (from five sets of values) below the sillimanite-biotite-garnet isograd, 600°C and 5.6 kb (three sets) above this isograd and 680°C and 6.3 kb (three sets) above the sillimanite-garnet-biotite isograd.[7] At 600°C, the data fall on the field gradient drawn by Froese (1997) and are consistent with the reactions of Bailes (1980a). The higher pressure is from areas approaching the structural domes north and northwest of Snow Lake. The data are consistent with about 14 km of unroofing at Snow Lake and 20 km of unroofing in the mantle of the domes. This is also consistent with the post–peak-metamorphic origin of the domes suggested by Bailes (1985) and Kraus (1998). It should be noted, however, that the top of the Batty Lake Complex is about 10 km above its root (Figure 13), showing that nearly half the present structural relief was produced before D_6. Doming had started during D_4 and is also consistent with data from the Namew Gneiss Complex south of Flin Flon (Section 4.3.2).

Northeast of Wekusko Lake in the eastern part of the Snow Lake subdomain, the presence of andalusite+biotite below the first appearance of sillimanite (Bailes, 1985) indicates a pressure of about 3.5 kb at 520°C. These conditions suggest that the Wekusko Lake area was at a relatively shallow crustal level during the thermal peak. The abrupt south turn of isograds east of there (Figure 18) and higher pressure to the west indicate D_6 block faulting or folding (after the thermal peak).

Results obtained or cited by Jungwirth (1995) in the west, between Weldon Bay and Duval Lake, north of the sillimanite-garnet-biotite isograd using the method of Gordon et al. (1994) have median values of 590°C and 4.6 kb (set of six).

Reaction 13 occurs between 770 and 800° at 5 to 6 kb in experimental work cited by Pattison et al. (2003), as is the case in the northern part of the Kississing–File lakes area. This high temperature heralds the onset of granulite-facies metamorphism with development of orthopyroxene in intermediate and mafic compositions at about 830° in the main part of the

[7] Only medians are cited here because of the high error of such calculations as determined by Gordon et al. (1994). Medians place little weight on outlying values, which are most prone to errors if the variance is from multiple sources.

Central Kisseynew subdomain. The rocks with biotite depletion show abundant evidence of melt loss or gain and are generally not suitable for geochemical analysis. Lenses of melanosome with large, partly retrogressed garnet and cordierite crystals and sillimanite suggest that reaction 13 may have occurred as far south as Nokomis Lake but later reversed with the addition of melt from the north. Thus, the southern limit of granite-melt generation is not properly constrained.

5 Geochemistry and chemostratigraphy

The use of major- and trace-element geochemistry and large-scale structure is crucial in defining and mapping the fundamental units (Syme et al., 1999; Zwanzig, 1999). Geochemistry is particularly important where the metamorphic grade is highest and the rocks have lost most of their primary structures but, as will be shown, have retained many of their original immobile trace-element and rare-earth element (REE) geochemical characteristics. This makes the geochemistry a necessary component of the stratigraphic, tectonic and metallogenic analysis. It allows the well-preserved rocks in the south to be traced for up to 80 km across metamorphic isograds. The combined study of structure and geochemistry provides a new tectonic framework and a 1:100 000 scale geological map (Maps GR2010-1-1 and -2, back pocket/DVD).

Geochemistry is also used extensively to determine the changing tectonic environments and crustal evolution of rock sequences exposed in the area that transects the Flin Flon and Kisseynew domains. Tracking gradual geochemical changes of closely related rocks locally allows the identification of autochthonous stratigraphic sequences. An abrupt major change in the inferred tectonic environments suggests a structural break. The evolution of arc and back-arc assemblages is tracked for a period of 70 m.y., the last 20 m.y. of which overlap the development of mainly sedimentary successor basins. This period is followed by another 30 m.y. of deformation, metamorphism and synkinematic magmatism.

The rocks of the Kississing–File lakes area are divided into four major groups of assemblages based on age, and 50 individual units based on geochemistry and field appearance. These units include mainly high-grade metavolcanic and meta-intrusive rocks that have lost many of their primary features. The metasedimentary rocks of the Burntwood and Missi groups have retained primary features in modified form and are also recognized by their metamorphic mineral contents, which reflect primary compositions. Limited geochemical data for the sedimentary rocks are used to compare them to the igneous units.

Raw geochemical data are included in Appendix 4, Table 9 (DVD); normalized volatile-free values with various ratios and averages are in Appendix 4, Table 10 (DVD); Nd isotope data and a summary of unit averages are given in Table 11. Where data from the Kississing–File lakes area were insufficient, additional analyses are included from Hollings and Ansdell (2002) and archived pulps for the Central and Northeast Kisseynew subdomains.

Geochemical diagrams are shown on 23 figures: Figures 19–23 in Appendix 1 for major-element variations; Figures 24–31 in Appendix 2 for minor- and trace-element ratios; and Figures 32–42 in Appendix 3 for chondrite-normalized rare earth element (REE) diagrams (Sun and McDonough, 1989) and multi-element diagrams, normalized with normal mid-ocean-ridge basalt (N-MORB) or primitive mantle (Sun and McDonough, 1989). These figures have been placed in the appendices for easy access and comparison. Unit symbols are consistent for all diagrams where units are not combined. The symbols and unit numbers and names are shown in Appendix 1, Table 12. Individual units are shown on combined and separate diagrams, frequently superimposed for comparison purposes, on tectonic fields from the literature or specific fields constructed from data of known modern environments taken from recent literature. This supports the validity of using the diagrams in the high-grade metamorphic rocks of the Kississing–File lakes area and can provide analogous tectonic environments for the local units. Fields with distinctive patterns from other parts of the Flin Flon domain are also used for comparisons and to suggest regional correlations. Diagrams based on several methods show that most units can be clearly identified in different ways using geochemical data. The duplication helps to overcome the mobility of various elements in the predominantly high-grade metamorphic rocks. The use of fluid-mobile elements is avoided as much as possible or, where they have been included, scatter has been eliminated by using only average values. For example, in MORB-normalized diagrams, elements are shown only up to Th on the left, thus excluding fluid-mobile elements that plot to the left of Th. The sharply defined multi-element signature of most units indicates that the data treatment is, in fact, rigorous.

An effort has been made in this report to provide brief qualitative insights into the petrogenetic processes that formed some of the major igneous units. An important purpose of detailed igneous geochemical analysis is to model these processes and estimate the composition of possible parent magmas and sources of melting. Such work is a powerful tool for defining tectonic and metallogenic environments. However, this requires little-altered rocks, high-precision analyses and knowledge of the composition of primary minerals, none of which are available here. Instead of modelling, compositional data and trends of the major units in the Kississing–File lakes area are compared to those of modern environments, thereby deriving their probable processes of formation by analogies with those cited in the literature. These processes are generally discussed in a separate, concluding paragraph for each unit and some of this is necessarily speculative.

5.1 Sampling method

Extensive geochemical sampling was carried out on the meta-igneous units from Kississing Lake to Dow Lake during detailed mapping. Where possible, samples were taken in areas with primary features; rocks showing signs of alteration or veining were generally avoided. This has restricted the main sampling area to the corridor of least mobilized rocks at middle- to upper-amphibolite facies, south of the migmatite front and extending from Kisseynew Lake to Dow Lake, and at upper-greenschist facies south of File Lake. Representative samples were collected, however, throughout the entire area (Map GR2010-1-2, DVD). Groups of several samples were collected over a type section of the more prominent units,

Table 11: Average geochemical composition of units, Kississing–File lakes area.

	Fussey Lake basalt	Ponton Lake basalt	Moody Lake basalt	Preaston andesite	Dickstone rhyolite	Storozuk basalt	Puffy Mine andesite	Koscielny andesite	Martell Lake basalt	Dow Lake andesite	SE Dow Lake porphyritic basalt	Meat Lake amphibolite	Walton Lake ampholite	Sherridon rhyolite gneiss (Meat Lake)	Felsic gneiss (Sherridon structure)	Josland Lake sills (gabbro-norite)
Unit	**1**	**2a(1)**	**2b(1)**	**2c**	**2d**	**2e**	**3a**	**3b**	**3c**	**3d**	**3e**	**4c**	**4d**	**5a**	**5(2)**	**7a**
SiO_2	49.5	49.5	51.3	55.2	76.2	53.6	59.6	55.4	49.8	57.6	52.1	51.2	46.1	76.9	75.2	49.6
Al_2O_3	14.37	14.81	14.99	14.57	10.94	14.88	14.48	15.64	14.76	15.31	17.68	17.69	15.72	12.19	11.95	15.78
FeO	12.16	10.84	12.73	12.99	5.22	13.02	10.25	10.53	12.82	10.80	11.92	12.12	11.02	2.18	3.33	8.09
CaO	13.05	11.65	10.48	8.27	1.12	8.79	8.01	8.37	10.99	8.20	10.40	10.02	10.20	1.76	3.22	12.52
MgO	7.83	9.58	6.48	4.08	1.98	5.63	4.05	5.19	8.23	3.50	4.50	5.16	12.24	0.84	1.41	11.27
Na_2O	1.70	2.04	2.50	3.20	3.07	2.65	2.13	3.49	2.11	2.95	2.21	2.48	2.29	2.93	2.05	1.66
K_2O	0.14	0.38	0.24	0.25	0.93	0.30	0.72	0.56	0.35	0.28	0.50	0.55	0.48	2.80	2.13	0.37
TiO_2	1.00	0.90	0.93	1.03	0.28	0.78	0.45	0.58	0.73	0.82	0.44	0.49	0.46	0.19	0.22	0.41
P_2O_5	0.08	0.10	0.08	0.15	0.05	0.10	0.11	0.08	0.08	0.32	0.05	0.10	0.06	0.04	0.07	0.06
MnO	0.20		0.22	0.21	0.26	0.22	0.17	0.18	0.19	0.19	0.18	0.22	0.18	0.05	0.08	0.19
Ni	91	156	48	47	8	29	19	38	59	14	16	11	168		10	121
Cr	317	500	111	68	2	17	50	127	181	15	34	16	524		19	713
V	336	298	363	333	2	395	316	308	419	184	330	345	331	12	20	202
Rb	1.3	3.7	2.8	4.0	10.6	3.1	12.8	7.2	3.6	1.9	4.7	10.8	5.4	36.7	26.0	6.7
Sr	154	198	92	121	44	134	294	265	261	145	227	237	98	90	87	150
Y	19.3	19.7	22.0	31.7	69.1	21.4	10.3	14.1	22.1	22.0	7.7	10.7	10.1	19.2	23.8	8.5
Zr	47.8	55.9	47.7	68.5	165.2	44.0	28.9	39.5	30.0	42.3	28.5	17.5	11.8	84.7	81.2	21.8
Nb	1.48	3.29	1.02	1.15	3.34	1.18	2.84	2.99	1.03	1.97	1.74	1.26	0.79	3.37	2.69	1.49
Cs	0.08	0.25	0.14		0.40	0.29	0.10	0.19	0.05	0.08	0.03	0.65	0.09	0.44	0.57	0.17
Ba	22	89	54	66	127	31	144	252	114	38	105	150	140	438	375	83
La	1.90	4.08	1.39	2.82	7.85	2.07	5.82	3.03	2.64	5.53	1.62	3.57	1.75	11.57	7.76	1.14
Ce	5.85	11.15	4.69	8.01	21.92	5.90	12.79	8.77	6.59	12.66	4.58	7.60	4.05	24.88	18.08	2.80
Pr	1.04	1.67	0.88	1.30	3.45	1.00	1.65	1.18	0.99	1.82	0.72	1.11	0.61	3.14	2.29	0.42
Nd	5.69	7.46	4.91	7.31	17.70	5.39	6.97	5.46	4.98	8.96	3.47	4.76	3.05	12.16	10.20	2.26
Sm	2.10	2.35	1.94	2.62	5.95	1.98	1.71	1.65	1.65	2.54	1.08	1.32	0.95	2.85	2.72	0.77
Eu	0.84	0.88	0.76	1.09	1.93	0.74	0.50	0.66	0.72	0.96	0.38	0.53	0.35	0.68	0.76	0.41
Gd	3.07	3.13	2.98	3.62	8.61	2.99	1.77	2.01	2.59	3.35	1.18	1.66	1.35	3.12	3.26	1.10
Tb	0.54	0.52	0.55	0.75	1.57	0.54	1.02	0.35	0.49	0.56	0.21	0.27	0.23	0.50	0.64	0.21
Dy	3.65	3.65	3.91	5.20	11.47	3.79	1.89	2.43	3.67	3.60	1.33	1.89	1.80	3.53	4.11	1.43
Ho	0.79	0.78	0.86	1.14	2.65	0.85	0.39	0.52	0.88	0.80	0.29	0.41	0.39	0.76	0.86	0.31
Er	2.26	2.37	2.62	3.50	8.16	2.52	1.21	1.53	2.70	2.39	0.86	1.29	1.25	2.40	2.77	0.95
Tm	0.33	0.36	0.39	0.55	1.28	0.39	0.19	0.23	0.42	0.37	0.12	0.20	0.19	0.39	0.44	0.14
Yb	2.11	2.28	2.64	3.54	8.36	2.51	1.28	1.49	2.82	2.37	0.85	1.38	1.25	2.67	2.94	0.90
Lu	0.30	0.34	0.40	0.52	1.32	0.41	0.20	0.24	0.47	0.38	0.13	0.20	0.19	0.43	0.47	0.13
Hf	1.41	1.72	1.41	2.08	4.97	1.36	0.97	1.28	0.91	1.25	0.94	0.54	0.46	2.76	2.36	0.59
Ta	0.44	1.20	0.51	0.04	4.97	1.36	1.53	0.69	0.04	0.10	0.15	0.62	0.56	1.79	0.25	0.10
Tl		0.06	0.05		0.08	0.03	0.07	0.06	0.04	0.04	0.04	0.10	0.06	0.08	0.09	0.06
Pb	0.6	2.4	0.8		3.2	1.6	4.6	4.5	0.6	1.3	1.6	2.3	1.3	6.7	5.2	8.8
Bi	0.07	0.51	0.30		0.08	0.16	0.41	0.12	0.09	2.31	0.17	0.47	0.43	0.19	0.22	
Th	0.16	0.88	0.21	0.31	0.81	0.20	1.40	0.97	0.29	0.75	0.29	0.66	0.35	3.35	1.20	0.16
U	0.08	0.30	0.10	0.21	0.56	0.25	0.54	0.36	0.16	0.32	0.13	0.33	0.17	1.22	0.75	0.04

(1) Fe-rich basaltic andesite samples omitted
(2) Fifty surface samples from Halo Resources Ltd. (2007)

Table 11: Average geochemical composition of units, Kississing–File lakes area. (continued)

	Josland Lake sills (ferro gabbro)	Josland Lake sills (quartz ferrodiorite–trondhjemite)	Martell Lake gabbro	Ragged Lake granite	Jay Lake granite	Big Island granodiorite	Archie Lake granodiorite	Star Lake to-nalite gneiss	Batty Lake gneiss	Burntwood Group meta-greywacke	Kississing River basalt	Evans Lake basalt	Jungle Lake basaltic andesite	Puffy Lake dacite	Cleunion rhyolite
Unit	**7b**	**7c**	**8a**	**9a**	**9b**	**9c**	**9d**	**9e**	**9f**	**11a**	**14a**	**14b**	**14c**	**14d**	**14e**
SiO_2	48.8	65.3	49.5	76.9	76.8	75.4	69.9	69.4	73.3	64.6	52.4	52.0	55.2	66.9	73.4
Al_2O_3	13.57	10.77	14.63	12.36	12.56	14.14	15.32	16.28	13.48	16.27	17.78	15.64	16.98	16.44	13.46
FeO	16.56	10.42	14.18	1.48	1.81	0.82	3.27	2.82	4.00	7.39	9.19	8.68	8.45	4.51	2.74
CaO	9.91	4.79	10.30	0.99	1.12	1.23	3.41	3.70	3.13	3.27	10.03	10.51	7.58	3.28	1.32
MgO	6.54	0.77	7.49	0.34	0.19	0.09	1.34	0.97	0.78	3.38	6.24	9.08	5.35	1.11	0.53
Na_2O	2.16	3.66	1.79	2.72	3.98	4.36	3.84	4.73	3.94	2.71	2.77	2.35	3.31	3.89	3.38
K_2O	0.32	0.31	0.38	4.98	3.31	3.55	2.36	1.63	1.04	1.48	0.47	0.76	1.73	3.09	4.75
TiO_2	1.79	0.89	1.27	0.14	0.14	0.05	0.39	0.34	0.22	0.63	0.83	0.63	0.93	0.49	0.29
P_2O_5	0.09	0.19	0.20	0.03	0.03	0.02	0.13	0.13	0.05	0.20	0.12	0.21	0.34	0.17	0.07
MnO	0.26	0.15	0.26	0.02	0.04	0.04	0.05	0.05	0.09	0.08	0.16	0.15	0.13	0.09	0.04
Ni	59		62	1	2	<50	7			38	103	191	115		4
Cr	94	26	232	6	2	<0.5	19			151	240	488	242	20	8
V	605	22	310	6	6		53	44	35	204	204	186	181	5	21
Rb	6.7	3.6	8.8	142.0	68.0	61.0	38.5	38.0	13.9	29.1	5.2	10.7	33.3	53.3	110.6
Sr	159	243	125	50	75	148	461	742	123	409	262	479	689	300	131
Y	21.1	72.5	30.3	35.5	96.0	18.0	9.9	6.0	20.9	13.7	23.5	19.7	22.8	31.1	38.4
Zr	54.0	196.0	76.7	128.5	162.9	75.0	155.3	127.5	91.3	102.3	70.1	77.9	168.1	329.1	217.2
Nb	4.35	10.93	6.24	10.62	10.70	5.00	10.63	3.50	3.56	7.22	4.48	4.88	8.32	11.28	11.67
Cs	1.06		0.44	1.37	0.50	<0.2	1.28	1.10		0.89	0.22	0.46	1.52	0.89	1.88
Ba	93	69	155	637	849	660	681	628	394	335	217	405	704	1212	1109
La	3.41	9.99	9.69	32.42	21.19	10.00	27.04	18.35	5.51	13.47	7.61	17.32	42.71	55.91	36.78
Ce	10.77	26.76	23.31	61.94	44.75	19.00	49.22	38.10	12.91	29.44	18.39	41.87	95.30	118.73	79.20
Pr	1.42	4.10	3.28	7.40	7.21		5.17	4.36	1.59	3.55	2.50	5.51	12.08	14.40	11.56
Nd	7.48	21.94	15.39	27.26	23.00	8.00	18.08	13.80	7.78	13.79	11.32	21.79	45.77	54.79	36.90
Sm	2.43	6.98	4.00	5.73	6.21	1.50	2.81	2.85	2.09	2.71	3.13	4.31	8.42	9.37	7.37
Eu	0.95	2.21	1.17	0.50	0.94	0.32	0.75	0.80	0.65	0.86	1.04	1.21	2.06	2.36	1.06
Gd	3.26	9.36	4.97	5.99	11.11		2.11	1.85	2.51	2.54	3.78	4.25	6.85	7.83	7.78
Tb	0.59	1.80	0.82	0.98	1.87	0.30	0.24	0.25	0.50	0.38	0.64	0.56	0.79	0.97	1.07
Dy	3.82	11.87	5.60	6.73	12.97	1.30	1.32	1.25	3.41	2.43	4.42	3.65	4.56	6.17	6.48
Ho	0.81	2.58	1.25	1.46	3.12		0.24	0.20	0.81	0.51	0.96	0.74	0.86	1.14	1.40
Er	2.43	8.26	3.61	4.46	9.57		0.67	0.60	2.51	1.47	2.85	2.24	2.39	3.43	4.14
Tm	0.36	1.26	0.56	0.68	1.50		0.10	0.09	0.39	0.23	0.44	0.33	0.35	0.52	0.61
Yb	2.28	8.20	3.54	4.31	8.28	1.46	0.62	0.50	2.49	1.52	2.81	2.19	2.14	3.73	3.62
Lu	0.34	1.23	0.55	0.66	1.28	0.24	0.09	0.07	0.39	0.24	0.42	0.32	0.32	0.56	0.55
Hf	1.67	5.97	2.72	4.38	5.36	2.30	3.79	3.20	2.78	2.59	2.22	2.18	4.30	8.62	6.36
Ta	0.63	1.95	0.37	1.72	1.73	2.20	0.78	0.35	2.29	2.59	0.88	0.89	0.95	1.28	2.42
Tl	0.08	2.77	0.08	1.08	0.47		0.37	0.35			0.06	0.08	0.33	0.47	0.70
Pb	1.2	11.0	4.5	17.4	8.2		8.8			6.3	2.9	5.0	11.7	15.5	12.0
Bi	0.24	1.45	0.08	0.05	0.06		0.18			0.27	0.41	0.38	0.37	0.21	0.31
Th	0.42	1.54	1.67	17.88	4.87	3.50	5.50	2.95	0.69	2.83	1.13	2.32	6.45	8.21	10.42
U	0.17	0.59	0.36	3.33	1.08	1.50	1.74	0.80	0.34	1.48	0.63	1.20	3.52	3.07	3.68

(1) Fe-rich basaltic andesite samples omitted
(2) Fifty surface samples from Halo Resources Ltd. (2007)

Table 11: Average geochemical composition of units, Kississing–File lakes area. (continued)

	Lobstick rhyolite	Barron Lake pluton	Ham Lake pluton	Touchbourne suite (felsic)	Toucbourne suite (mafic–ultramafic)	Duval Lake sill	Dow Lake quartz ferrodiorite	Highrock Lake tonalite–granodiorite	Flatrock Lake granite–granodiorite	Late dabase–gabbro	Synkinematic granodiorite–tonalite	Synkinematic leucogranite
Unit	**14f**	**15c**	**15d**	**16a**	**16a**	**16b**	**16c**	**17a**	**17b**	**18a**	**19c**	**19b**
SiO_2	73.7	75.3	67.4	60.5	53.3	56.7	58.4	67.5	72.9	48.6	71.7	73.7
Al_2O_3	13.97	13.23	16.20	16.25	8.90	16.00	13.11	16.03	14.61	13.82	14.83	14.80
FeO	2.29	2.03	4.28	6.63	8.20	8.40	13.61	4.73	2.29	14.38	2.85	1.21
CaO	1.40	1.17	3.44	6.04	10.74	6.41	6.48	3.12	1.43	11.67	2.40	1.30
MgO	0.29	0.37	1.48	3.79	16.18	4.78	2.79	1.91	0.60	7.43	1.20	0.33
Na_2O	3.73	3.17	3.69	2.82	0.36	3.12	1.67	3.69	3.20	1.61	3.82	3.36
K_2O	4.36	4.56	2.77	2.67	1.42	2.41	1.47	2.12	4.49	0.57	2.65	5.12
TiO_2	0.17	0.12	0.54	0.94	0.56	1.40	1.73	0.65	0.34	1.61	0.40	0.12
P_2O_5	0.03	0.03	0.16	0.27	0.22	0.65	0.47	0.19	0.10	0.15	0.09	0.09
MnO	0.02	0.04	0.06	0.09	0.14	0.10	0.26	0.04	0.03	0.20	0.04	0.02
Ni				46	391	24	13	22	11	109	10	13
Cr	9			67	1633	54	29	63	29	236	28	14
V	24	9	56	124	207	228	287	60	32	396	52	12
Rb	73.0	107.4	51.2	79.6	9.4	58.9	45.3	120.1	117.5	10.5	54.3	100.7
Sr	276	91	367	437	257	732	196	184	163	149	358	138
Y	11.2	31.5	15.0	15.5	15.2	23.4	47.1	15.5	17.6	23.6	9.8	10.3
Zr	187.2	132.1	183.5	185.1	64.1	193.8	276.2	189.0	182.9	89.4	133.2	71.6
Nb	10.93	6.41	5.34	11.38	4.17	14.23	19.90	14.00	10.66	6.80	5.32	5.54
Cs	1.17	8.82	1.18	2.01	0.20	3.49	1.23	3.08	1.18	0.72	1.57	1.67
Ba	1523	858	1155	648	80	709	498	502	670	98	987	697
La	33.35	20.55	22.55	25.41	13.69	51.28	47.06	28.56	39.26	7.31	25.14	19.59
Ce	66.42	49.23	47.40	52.62	37.09	104.04	101.43	58.94	79.57	19.16	48.53	38.12
Pr	5.62	5.72	5.26	6.83	5.44	14.01	12.16	7.13	9.08	2.64	5.48	4.25
Nd	32.56	22.90	20.68	28.02	24.34	58.16	49.28	27.04	32.86	13.16	20.20	16.40
Sm	5.76	4.61	3.67	5.44	5.17	9.81	9.89	5.49	5.84	3.79	3.48	3.60
Eu	1.23	0.63	0.98	1.57	1.38	2.70	2.15	0.97	0.90	1.34	1.01	0.73
Gd	3.10	4.91	3.25	4.45	4.54	7.77	9.46	4.69	4.63	4.41	2.80	2.95
Tb	0.48	0.72	0.46	0.60	0.57	0.85	1.35	0.62	0.62	0.74	0.35	0.42
Dy	1.64	4.38	2.57	3.13	3.01	4.68	8.61	3.33	3.48	4.37	1.90	2.14
Ho	0.34	0.99	0.51	0.59	0.55	0.83	1.77	0.60	0.65	0.88	0.36	0.36
Er	0.87	3.35	1.47	1.64	1.45	2.13	5.03	1.52	1.83	2.40	0.93	0.87
Tm	0.12	0.54	0.22	0.23	0.20	0.29	0.76	0.20	0.26	0.35	0.14	0.11
Yb	0.45	3.36	1.32	1.43	1.22	1.74	4.61	1.21	1.64	2.01	0.87	0.66
Lu	0.05	0.56	0.19	0.21	0.17	0.28	0.75	0.17	0.23	0.31	0.13	0.09
Hf	5.16	3.95	4.18	5.00	1.74	5.64	8.01	5.32	5.46	2.70	3.75	2.54
Ta	2.28	3.52	1.68	1.33	0.81	0.85	1.12	0.62	0.95	0.35	1.19	1.44
Tl	0.45	1.04	0.35	0.29		0.64	0.32			0.13	0.39	0.80
Pb	12.5	13.0	5.7	7.5	2.5	12.3	12.9	17.1	26.6		19.1	31.3
Bi	0.10					0.18	0.18			1.70	0.13	
Th	6.54	7.46	3.57	2.22	0.55	7.41	10.09	8.27	15.22	0.60	9.69	5.98
U	2.45	4.19	1.66	0.84	0.38	2.92	1.55	2.16	4.06	0.21	1.34	3.81

(1) Fe-rich basaltic andesite samples omitted
(2) Fifty surface samples from Halo Resources Ltd. (2007)

generally in the thickest and best preserved part of the unit and where field relationships were known. Much of this sampling was carried out on rocks that were exposed by forest fire and became clean in following years. This allowed the selection of samples devoid of early alteration, veins and pillow selvages, from clearly recognizable flows with a known tectonostratigraphic position. The results of scattered sampling from lower quality outcrops along strike or across the complex fold structures of the more intensely sampled sections were used to help delineate the full extent of units defined on the detailed maps. Samples were taken by sledgehammer from joint ledges or large *in situ* joint blocks where possible. They were broken up on the same outcrop and all weathered and most internal joint surfaces were trimmed off on the same bedrock. Veined and unusual looking pieces were discarded. About 1 kg of uniform pieces was generally collected, with more (typically 1.5 kg) collected for medium- to coarse-grained rocks. This sampling procedure has provided very consistent results where a series of samples was taken from a single unit over a limited area. The procedure is deemed critical in analyzing high-grade metamorphic rocks, all of which have been altered to various degrees.

5.2 Analytical methods

Samples were analyzed either by wet chemical method in the Manitoba Geochemical Laboratory or using x-ray fluorescence (XRF) for major elements, and a subset was analyzed for trace elements using inductively coupled plasma–mass spectrometry (ICP-MS) to provide a complete geochemical fingerprint.[8] The subset was selected to reduce the number of samples containing anomalous values; however, since alkalis are particularly mobile, their scatter was not always reduced. The analytical work was carried out at the University of Saskatchewan for the early samples and at Activation Laboratories Ltd. for most of the remainder (included in Appendix 4, Table 10, DVD). The trace elements of samples with low Zr contents from the University of Saskatchewan were reanalyzed in the same laboratory using the sinter method. Pulps were produced at the Manitoba Geological Survey rock lab in a tungsten-carbide mill, so Ta was not plotted because it is significantly contaminated, as indicated by lower values for samples from the same unit crushed by agate mill.

A small subset of samples was analyzed for Sm-Nd isotopes at the University of Alberta in Edmonton. All sample locations are shown on a simplified regional compilation map (Map GR2010-1-2, DVD). Details of the analytical techniques, detection limits and laboratories employed are given in Appendix 5 (DVD).

5.3 Major-element evolution

The geochemistry of the igneous rocks in the Kississing–File lakes area is traced through the evolution of the four major age divisions: 1) 1.89–1.88 Ga (early assemblages), 2) 1.88–1.85 Ga (early successor-arc magmatism), 3) 1.85–1.83 Ga (late successor-arc and basinal units of Burntwood/Missi age), and 4) 1.83–1.79 Ga (synkinematic intrusive units). The volcanic and intrusive rocks are tracked separately. Subdivision 4 consists of only intrusive rocks. The various assemblages in the age divisions generally show distinctive individual geochemical trends, but these common features allow only a tentative age-assignment of some of the units. Silica contents of our limited sample set are bimodal at various ages (Appendix 1, Figure 19).[9] Early volcanic assemblages are mainly mafic, whereas early successor-arc magmatism is distinctly bimodal but more commonly felsic; the younger intrusive suites include a significant proportion of intermediate compositions, with felsic rocks dominant in the synkinematic intrusions. Sampling of the volcanic rocks was dominated by mafic rocks, which form a larger number of small units in the south, and felsic rocks, which form the Sherridon gneiss in the north. When subvolcanic intrusions and reworked tuff are included, the Missi-age magmatic rocks are predominantly felsic in the Kississing–File lakes area. This evolution from predominantly mafic to bimodal volcanism to mainly felsic plutonism is also typical in the main part of the Flin Flon Domain (Lucas et al., 1996).

Total alkali contents are highly variable, reflecting original compositions and alkali mobility during seafloor alteration and metamorphism. There is a trend of increased average K_2O+NaO_2 with time at all values of SiO_2 from the early assemblages to the late successor-arc volcanic and intrusive rocks (Appendix 1, Figure 20a–d).[10] The early units and early successor-arc units are subalkaline, and the younger rocks range from subalkaline to (possibly) mildly alkaline (Section 5.5). This trend is moderated for the youngest (synkinematic) intrusion. Contents of K_2O increase from predominantly low-K in the early assemblages to predominantly medium-K in the early successor-arc intrusions, and to medium- to high-K for the youngest intrusions (Appendix 1, Figure 20e–h). The Missi Group volcanic rocks have the widest range of K_2O. Trends of increasing K_2O and other large-ion lithophile elements (LILE) with time are well established in the Flin Flon Domain (Lucas et al., 1996; Whalen et al., 1999). The geochemical trends may be related to any of 1) crustal thickening, 2) cooling of the subduction zone (Jolly et al., 2001), 3) melting deeper down the subduction zone, with 4) a decrease in the degree of partial melting (e.g., Woodhead and Johnson, 1993), 5) mixing of melts from the mantle wedge and slab, and/or 6) greater sediment supply to subduction (Jolly et al., 2001), or 7) crustal melting (e.g., Ferrari et al., 2002; Richards, 2003).

There is also a prominent evolutionary trend from early tholeiitic mafic volcanic rocks to later calcalkaline felsic intrusions with minor Fe-rich intermediate to mafic rocks. On an AFM diagram, the tholeiitic Fe fractionation in the early mafic volcanic rocks is followed by a calcalkaline trend in the felsic volcanic and intrusive rocks (Appendix 1, Figure 21a). The Josland Lake sills are distinguished by exceptionally strong Fe enrichment, with high FeO^t/MgO ratios of mafic to felsic upper members and Mg-enriched lower members (Appendix 1, Figure 21b). A sample from a chilled margin and volume

[8] Geochemical laboratories used are listed in Appendix 4, Table 10 (DVD); methods are summarized in Appendix 5 (DVD).
[9] All major element values shown in the text and figures are given as weight percent, normalized to volatile free.
[10] For clarity, some symbols that overlapped other symbols in the diagrams may have been moved a short distance.

calculations of the different fractionated layers (Bailes, 1980a) suggest that the source magma was relatively Fe rich and had undergone earlier low-pressure crystal fractionation. This was followed by the extreme fractionation in the sills. The successor-arc assemblages and synkinematic intrusions, in contrast, are dominated by calcalkaline rocks and generally include tholeiite with only moderate or weak Fe enrichment (Appendix 1, Figure 21c–f). Minor synkinematic mafic dikes are also tholeiitic. Quartz diorite bodies among the younger intrusions are highly Fe enriched (Dow Lake quartz diorite, Appendix 1, Figure 21f), but their affinity may be better assessed by trace-element ratios. Some of the youngest (synkinematic) dikes and sills are mafic–ultramafic tholeiitic intrusions that form a minor but tectonically important suite.

Simple variation diagrams help to define tectonic-assemblage types of individual units mapped in the Kississing–File lakes area. Diagrams with fields from various modern tectonic environments using the least mobile element ratios show that rocks of various ages in the Kississing–File lakes area have volcanic-arc (or crustal)– to MORB-like signatures with moderate and low TiO_2 contents, respectively, and moderate to high Al_2O_3 contents, respectively (Appendix 1, Figure 22). On the TiO_2 versus MgO plot, the early mafic to intermediate rocks divide into two groups: those with MORB-like values or with a weak arc signature and those with a clear volcanic-arc signature (Appendix 1, Figure 22a, b). A significant proportion of the low-K tholeiite in the early volcanic rocks and some in the Missi Group have higher TiO_2 contents than common subduction-related lavas. Josland Lake sills belong to the higher TiO_2 assemblages, as is evident from samples of a chilled margin and a fine-grained phase. The highly fractionated ferrogabbro samples are highly enriched in Ti, whereas the more mafic units are slightly depleted in Ti compared to MORB (Appendix 1, Figure 22c). The plagioclase-rich SE Dow porphyritic basalt is similarly depleted in Ti but belongs to the assemblages with weak arc signature, as seen on other plots (Section 5.4).

The thickest and laterally most continuous units of the early rocks with moderate to relatively high contents of TiO_2 1) fall into the field of modern MORB (Fussey Lake basalt, Ponton Lake basalt and Josland Lake chilled gabbro), 2) straddle the boundary of the MORB and arc fields (Moody Lake basalt), and 3) plot near the upper limit of modern arc-volcanic rocks (Preaston andesite and Storozuk volcanics; Appendix 1, Figure 22a, c). The TiO_2 content of most of these units plots in the field of Phanerozoic back-arc–basin basalt to andesite (BABB on Appendix 1, Figure 22a). They also show a weak arc signature on various tectonic diagrams based on trace-element ratios (Section 5.4), consistent with BABB (e.g., Stern et al., 1990; Sinton at al., 2003). The majority of mafic to intermediate units have less than 1% TiO_2 and fall into the field of modern arc-volcanic rocks (Appendix 1, Figure 22b).

Most early mafic to intermediate units have normal to low values of Al_2O_3 (Appendix 1, Figure 22e), but late successor-arc units with slightly elevated TiO_2 contents (e.g., Kississing River basalt and Jungle Lake basaltic andesite in the Missi Group) contain flows with high Al_2O_3 contents as evidence of their origin in a volcanic arc or marginal basin (Saunders and Tarney, 1978). The high H_2O content in mafic melts, as expected in a subduction environment, delayed crystallization of plagioclase and thus raised the Al_2O_3 content (Fretzdorff et al., 2002). Both early and late felsic successor-arc units extend into the high Al_2O_3 range (>15% at 70% SiO_2; Appendix 1, Figure 22f–g). The origin of some of these rocks is also consistent with melting of hydrated ocean floor or lower crustal amphibolite (Whalen et al., 1999, 2004). Alumina saturation is measured by the Shand index (modified after Maniar and Piccoli, 1989): A/CNK = molar Al/(Ca+Na+K). Values scatter about 1.0 for most granitoid rocks except the synkinematic (youngest) intrusions, which are slightly peraluminous and scatter about 1.1 (Appendix 1, Figure 23j, k).

The major-element geochemical evolution of the volcanic and intrusive rocks in the Kississing–File lakes area is consistent with various early oceanic environments and later emerged arcs, as described by Lucas et al. (1996) and Whalen et al. (1999) for the main part of the Flin Flon Domain. Although K_2O plotted on variation diagrams for the individual high-grade volcanic units produces a wide scatter, the main data cluster increases from the MORB-like Fussey Lake basalt (0.1–0.2% K_2O), through units with a weak arc signature (approx. 0.1–0.56% K_2O) to arc tholeiite (0.2–1.3% K_2O) and rhyolite (averaging 1–2% K_2O); Missi Group basaltic andesite averages 1.7% K_2O and associated rhyolite ranges to >5.9% K_2O (Appendix 4, Table 10, DVD). The data spread is consistent with the preservation of petrogenetic trends due to crystal fractionation, fractional melting or magma mixing from various depths or sources. Trends of slightly increasing or decreasing K_2O with TiO_2, most likely due to fractionation involving pyroxene, are preserved in basalt–andesite suites. On the other hand, different K_2O contents, at nearly the same TiO_2 content, most likely involved fractionation of high-temperature feldspar. The latter trend is typified in the high-K_2O basalt–andesite and rhyolite suites. Kississing River basalt shows a spread for both K_2O and TiO_2, presumably due to fractionation of a Ti-bearing phase as well as feldspar. This is consistent with the highly variable content of mafic and plagioclase phenocrysts in these units (Section 5.9.4.1 and Appendix 1, Figure 23g).

The fluid-soluble, large-ion lithophile elements (LILE), such as the major element K and the minor elements Rb, Sr and Ba, are important indicators of subduction-related processes. Thus, the recognizable increase in K_2O with the progressive tectonic evolution suggests that these elements are also useful in distinguishing volcanic-arc units from BABB, even for the highly metamorphosed flows. The Ba content of basalt and andesite is very useful in this distinction, but the high variation within units from the Kississing–File lakes area is the result of alteration. Nevertheless, a plot of early basalt to andesite with low K_2O contents and a weak arc signature shows that most of these units fall into the field of BABB (Appendix 1, Figure 23i), as determined from the Manus back-arc basin (Sinton et al., 2003). From that group, only the Ponton Lake basalt has a range of Ba content that extends into the field of arc lavas such as the Puffy Mine andesite. Neodymium isotope data (Section 5.6) may provide an explanation for the high Ba values in the Ponton Lake basalt, but they are most likely due to alteration.

5.4 Trace element patterns

Relatively fluid mobile major elements such as SiO_2 and K_2O (and other LILE) are subject to alteration in the high-grade metamorphic rocks of the Kississing–File lakes area. They generally display considerable scatter, and diagrams involving them have provided only the general characteristics of the broad groups defined by age and tectonic affinity. Where gain or loss of mobile elements is significant, the weight percentages of all major elements are affected by alteration. Plots of many trace elements, including high-field-strength elements (HFSE; e.g., Ti, Hf, Zr, Nb), rare earth elements (REE) and Th, in contrast, show considerably less scatter. The least mobile elements (HFSE and heavy REE [HREE]) provide a subset of the geochemical data for the Kississing–File lakes area that is little altered; they reliably characterize the geochemical fingerprint of individual units and define tectonic affinity. Consequently, chondrite-normalized plots of REE and extended-element plots normalized to normal mid-ocean-ridge basalt (N-MORB) that omit elements more mobile than Th are used extensively in this report, as was done previously for the geochemistry of Flin Flon Domain rocks (Stern et al., 1995a, 1995b). Ratios of the 'immobile' elements also provide a powerful tool to correlate samples, compare units and avoid variations related to fractional crystallization and fractional melting.

The cause of variations in the trace-element patterns in modern and older Phanerozoic arcs and back-arc basins is related to the ionic potential of the elements and hence their mobility in various media. Whereas the 'conservative' elements (HFSE and HREE) provide a direct measure of their abundance in the mantle or mafic crustal source rock, light to medium REE (LREE, MREE) are transported in silicate melts to the final source rock, and LILE are carried to the site of melting in predominantly aqueous fluids (Pearce, 1995; Jolly et al., 2001). The LREE, Th and LILE provide data related to the subducting mafic slab (where present), its sediment cover and possible granitoid crustal contaminants. The abundance of LREE can increase by input from the subducting slab, mantle composition and degree of partial melting. Fluid flux is considered to be strongest at the base of the thin frontal part of the mantle wedge above the subduction zone, whereas silicate melts have a stronger influence in deeper and/or hotter slab environments. Decompression melting of nearly unmodified mantle occurs in back-arc basins. Transitional geochemical characteristics can occur at the basin margins. Thus, plots using all of these elements can 1) distinguish assemblages from different tectonic environments (Winchester and Floyd, 1977), 2) track along- and across-arc to back-arc variations (Woodhead and Johnson, 1993; Ewart et al., 1998; Fretzdorff et al., 2002; Sinton et al., 2003) and 3) distinguish different igneous units that had sources of different composition or differences in the processes leading to magma production. As such, they provide conclusions critical to this report.

5.4.1 Incompatible element ratios

Of the 43 individual igneous units that have had more than one sample analyzed for trace elements, most form compact fields on several commonly used variation diagrams. These fields show little overlap and therefore help to define the units, particularly where primary volcanic features have been obliterated. For instance, the contents of REE and the ratios of LREE to HREE (e.g., Yb versus $[La/Yb]_N$) commonly provide separate fields for the various units (Appendix 2, Figure 24). In general, units form fields with 1) steep trends toward high Yb, or 2) shallow trends toward high $[La/Yb]_N$ (chondrite-normalized after Sun and McDonough, 1989). Felsic Archean rocks have been divided into these categories on similar plots (representing barren and mineralized groups; Lesher et al., 1986). The high–$[La/Yb]_N$ trend has been attributed to the presence of garnet±amphibole in the residue of differentiating magma (Goodwin et al., 2006), but the origins may be more complex. Regardless of cause, the plots are highly effective in defining assemblage types and individual comagmatic igneous units in the Paleoproterozoic metamorphic samples. The early arc-volcanic rocks generally form a 'high-Yb' trend at low $[La/Yb]_{N.}$ The successor-arc plutons define the 'high–$[La/Yb]_N$' trend, whereas the early successor-arc volcanic rocks and Batty Lake gneiss fall in between. This change is part of the chemical evolution in the Kississing–File lakes area from compositions similar to MORB and primitive arc-volcanic rocks toward a modern crustal composition.

The most meaningful units for mapping are formed from one or two unique sources of melt, which allows them to be distinguished from other units. Samples of these vary only due to fractional crystallization. Because the REE are largely incompatible with the major phases during fractional crystallization or fractional melting, they are concentrated at nearly the same rate, thus giving constant ratios. Some ratios will change, but only when phase assemblages change, and these show a bend in slope on variation diagrams with limited scatter. The absolute contents of the incompatible immobile elements in mafic to intermediate rocks, however, increase with the extent of fractional crystallization, and decrease with the extent of fractional melting (arrows in Appendix 2, Figure 24). Simple fractionating comagmatic suites, therefore, produce linear plots of increasing Yb at nearly fixed $[La/Yb]_N$. Minor fractionation of REE during crystal fractionation is indicated by the steep positive slope of Yb versus $[La/Yb]_N$. The intermediate to felsic rocks and some late mafic to intermediate suites, on the other hand, show high–$[La/Yb]_N$ trends that are nearly horizontal in Appendix 2, Figure 24. The origin of these magmas may involve vapour-absent partial melting of eclogite or amphibolite at somewhat shallower depth than in modern subduction (Goodwin et al., 2006), but this melt would be variably admixed with melt from the mantle wedge above the ancient subduction zone. Some late (synkinematic) granitoid rocks, as well as the Lobstick rhyolite in the Missi Group, can show trends toward low Yb at high $[La/Yb]_N$ that indicate garnet in the melt residue.

The absolute content of Yb can be a measure of the 'fertility' of the source mantle and therefore an important guide to the tectonic origin of basalt and andesite, and their role in the mineral potential of a unit. The variation in Yb within a unit can be constrained to provide a better comparison between unrelated highly and weakly fractionated units by estimating the REE content at a fixed value of MgO (e.g., Stern et al., 1990; Pearce, 1995; Ewart et al., 1998; Jolly et al., 2001; Fretzdorff et al., 2002). Between 6 and 9% MgO in the liquid composition, abundant olivine has completed crystallization in favour of pyroxene, plagioclase and, in arc lavas, Ti-magnetite. This changes the slope in many variation diagrams (e.g., Appendix 1, Figure 22a–d). Because slopes are more uniform at higher and lower MgO contents, variation trends of most mafic–intermediate rocks can be projected toward 7% MgO. Plots of Yb versus MgO and regression lines or visually fit curves drawn through the points for early basalt to andesite are summarized as Yb7 values in Table 13 and as crossbars in Appendix 2, Figure 24. These values form a group with Yb7 ≈ 2.0–3.2 and generally $[La/Yb]7_N$ <0.8 (near MORB), and another group with Yb7 ≈ 1.1 and $[La/Yb]7_N$ between 1.0 and 2 for more typical arc-like units.[11] The first group includes Fussey Lake MORB-like basalt and several adjacent units with a weak arc signature. The MREE- to HREE-depleted group includes Puffy Mine andesite, Koscielny andesite and SE Dow Lake porphyritic basalt from the Amisk Collage, as well as Meat Lake and Walton Lake amphibolites. Missi Group volcanic units are generally LREE enriched and range to much higher La/Yb ($[La/Yb]7_N$ >2 and <15) than the early units but with similar contents of HREE expressed as Yb7 (Appendix 2, Figure 24c). The plots show which units are closely related and how they differ from other related groups.

Possible petrogenetic interpretations, which contribute to understanding the tectonic evolution of the igneous rocks in this study, may be derived from the incompatible data. Because La/Yb and Yb7 are little affected by fractional crystallization or melting, their values are characteristic of the magma source and environment. The chondrite-normalized ratio $[La/Yb]_N$ versus Yb7 provides a powerful tool to distinguish various cogenetic meta-igneous units. Flows from sources that were similar to MORB formed where decompression was more important in melting than fluid fluxing. This would have occurred in an extensional tectonic environment: a rifting arc or back-arc basin. The REE-depleted lava with strong arc signature (highly depleted HFSE) would have formed from previously depleted mantle but with newly added fluid. Such magmas are generated in an arc that has a prominent back-arc basin as subduction draws the depleted mantle into the wedge where it is reprocessed (Gamble et al., 1993; Woodhead et al., 1993). Higher degrees of melting (due to fluid fluxing) also reduce the melts in HREE (e.g., Yb) and HFSE because these elements are highly incompatible with the crystallizing phases.

Further tests of the behaviour of trace-element ratios are plots against Mg number (Mg# = molar Mg / [Mg + $0.85Fe^t$]), a measure of fractionation or melting, as shown on Appendix 2, Figure 25a–c. Ratios of La/Yb form mainly linear plots that show only a slight increase with fractional crystallization (lower Mg# in Appendix 2, Figure 25a). Ponton Lake basalt and Evans Lake basalt may involve processes that changed the La/Yb ratio somewhat during progressive melting, magma- or source-mixing, or contamination (e.g., Saunders and Tarney, 1978; Huijsmans et al., 1988; Ewart et al., 1998; Sinton et al., 2003). The La/Zr ratio is little affected by fractionation of the early volcanic rocks (Appendix 2, Figure 25b) and was therefore used to test for trace-element preservation or alteration within some of the better sampled units (Appendix 2, Figure 26).

Incompatible element ratios are commonly used in metamorphic rocks as a substitute for major-element variation diagrams, such as the AFM diagram. A clear distinction between

[11] $La/Yb7_N$ is the chondrite-normalized ratio La/Yb estimated at 7% MgO.

Table 13: Ytterbium contents at 7% MgO (Yb7), Kississing–File lakes area.

Assemblage	Yb7 (Yb at 7% MgO; ppm) (1)	Data quality(2) for Yb7
Fussey Lake basalt	2.3	Moderate
Ponton Lake basalt	2.3	Moderate
Moody Lake basalt	2.2	Moderate
Storozuk basalt	~2	Low
Preaston andesite	~1.5	Low
Martell Lake basalt	~3.2	Low
SE Dow Lake porphyritic basalt	~0.4	Low
Meat Lake amphib	0.5	Moderate
Walton Lake amphibolite	1.4	Moderate
Josland Lake sills	~1.7	Low
Kississing River basalt	2.6	Moderate
Evans Lake basalt	2.5	Moderate
Jungle Lake basaltic andesite	1.8	Moderate
Dow Lake quartz diorite	~3	Low
Touchbourne suite (mafiic)	~2	Low
Duval Lake sills	1.7	Moderate
Weldon Bay gabbro	~2.0	Low

(1) Estimated maximum errors of Yb7 range from 0.2 to 0.8 ppmt.

(2) Quality of data for Yb7 was ranked on the scatter of points used to visually fit curves through the points.

tholeiitic and calcalkaline (and transitional) meta-igneous units is achieved using the ratios of Zr/Y versus Th/Yb (Ross and Bédard, 2009). This classification is applied to the rocks in the Kississing–File lakes area (Appendix 2, Figure 27) and is particularly useful for those with highly mobilized alkali elements. A comparison with Appendix 1, Figure 21, nevertheless, shows that most units have retained their major-element composition sufficiently to plot in the same fields on an AFM diagram as on the incompatible-element diagram. Where garnet is involved in melting, moreover, the latter diagram is not as useful.

Without the use of LILE in the metamorphic rocks, the ratio of Th/Nb is an important indicator of tectonic environment in mafic–intermediate volcanic-arc and non-arc melts. The ratio increases with the influence of subduction processes because Th is made available from the sinking slab. In the Kississing–File lakes area, Th/Nb_N (normalized to N-MORB) changes from about 2 for MORB-like basalt, to 2–4 in units with weak arc signatures, to generally >6 for arc-tholeiite and calcalkaline rocks. The ratio is highest for some Missi Group volcanic rocks. When anomalous values are ignored, most early units show a systematic decrease in Th/Nb_N with increased fractionation or decreased melting (decreasing Mg# in Appendix 2, Figure 25c). This is mainly because of increased Nb. Some Missi Group volcanic rocks show an increase in Th/Nb_N with fractionation. The behaviour of Th/Nb is probably related to greater partitioning of Nb into the liquid in the early tholeiitic rocks but into oxides in the late calcalkaline rocks. Other units and some individual samples show a wide scatter that may be a result of alteration. The common presence of coherent trends, however, indicates that Th/Nb ratios are still useful for assigning tectonic origins to the high-grade metamorphic rocks.

Plots of REE ratios for most units also show a clear evolution trend over time, from ratios relatively closer to MORB to higher La/Yb ratios (Appendix 2, Figure 24). Most early assemblages have $[La/Yb]_N$ <3, and early successor-arc volcanic rocks (Sherridon–Meat Lake assemblage) overlap this trend but extend to $[La/Yb]_N$ of 7. The early successor-arc plutons have $[La/Yb]_N$ up to 35, and the ratios for late successor-arc assemblages range from 1.5 to 100. These trends are even more obvious when Yb7 values are used to remove the effects of fractionation or melting (see above). These evolution trends have been well established in the main part of the Flin Flon Domain (Lucas et al., 1996; Stern at al., 1999; Whalen et al., 1999) and are common in the geological record of Phanerozoic volcanic arcs (e.g., Jolly et al., 2001). The late-tectonic (syn- to late-kinematic) intrusions lie mainly within the field of high La/Yb, low Yb assemblages. The increase in La/Yb with time accompanies the increase in K_2O (Appendix 1, Figure 23) and other LILE such as Rb, Ba (Appendix 2, Figure 28), as well as Th/Yb and Zr/Y (Appendix 2, Figure 29). Although conclusions from LILE data must be treated with great caution, a trend is apparent from MORB-like and oceanic-arc assemblages to continental-arc assemblages. The increasing contents of LREE and LILE are the same as in the main part the Flin Flon Domain, as determined from a larger data set and ascribed to gradual development of granitic crust in the juvenile zone of the THO (Lucas et al., 1996; Whalen et al., 1999).

5.5 Alteration and element mobility

Alteration of the volcanic units, which is widespread and locally very intense, is interpreted to result from synvolcanic hydrothermal activity beneath the seafloor and metamorphism during orogeny. The purpose of identifying rocks that have suffered alteration is twofold: 1) it can serve as a guide to mineral exploration, and 2) as is important in this report, altered samples can be excluded from element-variation plots or ignored when drawing variation trends and calculating averages.

The most highly altered volcanic rocks, relevant to exploration, contain porphyroblasts such as garnet, cordierite, anthophyllite/gedrite, cummingtonite, sillimanite, hercynite or magnetite. These minerals suggest a loss of alkalis and relative gain of Mg, Fe and Al during alteration. Such rocks are particularly prominent in belts of regional stratiform alteration in the Sherridon gneiss and in some of the associated amphibolite. They are related to VMS deposits (Froese and Goetz, 1981; Zwanzig and Schledewitz, 1992; Bailes and Galley, 1999). Some basalt units have abundant epidote- and/or diopside-rich domains interpreted as the metamorphic product of seafloor alteration (Bailes and Syme, 1989). Alteration to calcsilicate also occurs along massive, subaerial flow contacts (locally identified by strongly vesicular layers). Highly metamorphosed felsic volcanic rocks containing calcsilicate minerals over large areas are also considered to be altered. Proximity to carbonate veins or calcsilicate domains may also be linked to changes in the contents of HFSE. Some of these altered samples are included in Appendix 4, Table 10 (DVD), but they were generally avoided during the sampling. However, less obvious early alteration could not be avoided in the main dataset.

Alteration during metamorphism is also expected, and has been shown to occur in the metasedimentary rocks, and in felsic and some intermediate volcanic rocks (Section 4.3). Sampling of rocks with visible secondary muscovite or green retrograde biotite, as well as rocks with small veins and thin oxidation surfaces, was generally avoided. Sharp-walled, narrow veins were trimmed off during sampling when no rocks without veins could be found. The Sherridon gneiss and Missi Group metarhyolite commonly contains veins or patches of quartz-plagioclase–rich leucosome and local garnet-rich schlieren in mesosome. Although samples were not collected from these areas, a nugget effect due to small, irregularly scattered garnet porphyroblasts in the Sherridon gneiss may have affected the HREE contents of the sampled units. An attempt has been made to isolate such altered samples, and to avoid using them in defining typical trace-element patterns and determining the tectonic origin of a unit. The allochthonous, sharp-walled sheets of tonalitic leucosome were assumed to have minimal effect on the immobile elements in the adjacent rocks. The effect of metasomatic back-reaction of leucosome-derived fluid (Section 4.3.2.2) is expected to be confined mainly to major elements and LILE. However, to test this hypothesis, element ratios were plotted and a regression analysis carried out where sufficient data were available. Analyses that fall off the main trend were not plotted in some petrogenetic diagrams, or were tracked and not considered in conclusions regarding the origin of the affected rocks.

Igneous units that have been identified by similar field characteristics, petrography and geochemistry are considered to have been derived from the same type of parent magma or magmas (in the case of mixing). Samples from such units can show linear or smoothly curved plots on variation diagrams due to processes of crystal fractionation, fractional melting and possibly assimilation and mixing. During fractional crystallization, elements that are compatible with crystallizing phases are removed from the melt, whereas, during melting, the most compatible elements tend to remain in the restite until advanced stages of melting. Incompatible elements, on the other hand, are accumulated in the fractionating melt during crystallization but are depleted from the melt during progressive melting. When two compatible elements are plotted, there can be a wide dispersion (e.g., Kississing River basalt in Appendix 1, Figure 23g). Pairs of elements, one of which is relatively fluid immobile, generally do not show good correlation if the variation is strongly affected by alteration due to hydrothermal or metamorphic processes.

In Appendix 1, Figure 23, a fluid-mobile LILE (K_2O) plotted against an immobile HFSE (TiO_2) shows a wide scatter for some units but a linear trend for others. Outlying values (some of which are identified as large symbols in Appendix 1, Figure 23) are interpreted to be affected by significant alkali metasomatism. Nevertheless, the data show a systematic overall increase in K_2O from early MORB to late successor-arc andesite and basalt (Table 14). The widest scatter of points is in the K_2O-rich rhyolite units and Jungle Lake basaltic andesite. This suggests that alkali flux was highest in these units, whereas some basaltic units suffered only weak alkali metasomatism. The data are consistent with the preservation of a fine grain size in many of the mafic rocks and incipient development of leucosome in the rhyolite and K_2O-rich andesite. For example, even after omitting two outlying values from the determined K_2O contents of Jungle Lake basaltic andesite, the average value with scatter is 1.54 +2.8/–0.8%, whereas unaltered medium-K andesite from a single volcanic field in east Java (Gertisser and Keller, 2003), at the same SiO_2 value, has 1.65 ±0.15 wt.% K_2O. Thus, the value indicating ultra-high K and the two values indicating high K for Jungle Lake basaltic andesite in Appendix 1, Figure 20g are not meaningful; the unit is better classed as a medium-K basaltic andesite. Much of the basaltic andesite and rhyolite in the Missi Group appears to be tuff, and some outcrops show metamorphic differentiation or veining. Consequently, these rocks are among the most altered in the database of the Kississing–File lakes area, and other units provide more reliable alkali-element data.

Loss (but not gain) of alkali elements in the felsic rocks has been monitored using alkali/alumina ratios. Samples to the left of the alumina saturation line ([2Ca + Na + K] / Al <1 in Appendix 2, Figure 25d) have undergone loss of alkali elements if they were originally metaluminous like the tested units. Prominently altered samples, including Sherridon rhyolite gneiss and Cleunion metarhyolite, are shown as large symbols. The two most altered samples of Sherridon gneiss have lost all primary features in outcrop; the most alkali-depleted rock contains minor sillimanite and is from the margin of a large domain of regional hydrothermal alteration characterized in its core by garnet-, cordierite- and anthophyllite-rich mineral assemblages. Two samples of Dickstone rhyolite from the vicinity of the Dickstone VMS deposit are even more altered than the sillimanite-bearing gneiss at Sherridon. The data show that high-grade metamorphism contributed less to rock alteration than did synvolcanic alteration or hydrothermal alteration.

Table 14: Range in K_2O content of early units, Kississing–File lakes area.

Assemblage type	K_2O (%)	Figure
MORB	~0.1–0.2	23a
Weak arc affinity	~0.1–0.56	23a–c
Arc basalt to andesite	~0.2–0.88 [(1)]	23d
Late successor-arc basalt to andesite	~0.2–5.0	23g–h

[(1)] disregarding one outlier

To test for alteration among the trace elements in the mafic–intermediate units, ratios of various incompatible elements were plotted against Mg# in Appendix 2, Figure 25a–c, where Mg# is mainly a function of fractionation or melting processes. Random, independent changes in the ratio involving FeO and MgO are assumed to be due to alteration rather than changes produced by magma genesis and fractional crystallization. Where sufficient samples were analyzed, data scatter is moderate and generally less than variations between units. Ratios that do not fall on the main trends are depicted by enlarged symbols and have been excluded from the regression (shown as dotted lines in Appendix 2, Figure 25a–c). The La/Zr ratios, for example, produce little scatter and generally low slopes for the early low-K units. This indicates that these elements are fully incompatible and their values little affected by alteration. The Th/Nb ratios produce more scatter, particularly for Ponton Lake basalt that contains abundant epidote-diopside–rich domains produced during seafloor alteration. At least 12 of the 56 samples tested from various units have anomalous Th or Nb, about half showing an increase in Th over the main trend, as indicated by Appendix 4, Table 10. The other altered samples show a loss in Th or gain in Nb. Because Th/Nb ratios are important tectonic indicators, caution is required in any interpretation based on diagrams using these elements. Nevertheless, the trend of increasing Th/Nb from the MORB-like Fussey Lake basalt ($[Th/Nb]_N$ = 2 × MORB) to Jungle Lake medium- to high-K andesite (~13 × MORB) is clearly apparent. The slope of the regression lines indicates that these elements were weakly compatible with crystallizing phases and that Th accumulated more slowly than Nb in the fractionating, early low-K units, whereas the reverse occurred in the younger, K-rich arc-volcanic rocks. The diagrams indicate that alteration does not fully mask original petrogenetic trends, even of the relatively fluid mobile elements.

Because La and Zr are immobile and incompatible in mafic rocks, these elements have linear plots with high correlation coefficients (R^2). This is the case for the Ponton and Moody Lake basalts, and the Preaston-Storozuk volcanics, where samples with little scatter indicate a common history of melting and fractionation with little alteration (Appendix 2, Figure 26b–d). Most other mafic–felsic units show linear plots with distinctive rates of La versus Zr fractionation but slightly more scatter (Appendix 2, Figure 26e–h). With the alkali-depleted samples (large symbols) omitted from the regression, Sherridon volcanic gneiss also forms a linear plot.

The ratio Sr/Y is an important petrogenetic measure in felsic–intermediate melts from the crust, mantle wedge and slab, but Sr is generally considered to be subject to alteration. Plots of these elements were tested for scatter in Sherridon rhyolite-derived gneiss and Missi-age metarhyolite with related basalt (Appendix 2, Figure 26l). The data form linear plots with generally moderate scatter of Sr in four out of five samples. Two samples of Sherridon gneiss with low Sr are also anomalous on the La/Zr plot and have the lowest alkali versus alumina ratio in Appendix 2, Figure 25d. These were clearly depleted of alkalis during alteration and contain small amounts of garnet or sillimanite. Alteration is interpreted as a relatively early (synvolcanic) hydrothermal event. High Sr contents in two samples of Missi Group rhyolite occur in gneissic rocks at File Lake and northeast of Walton Lake, north of the migmatite front. The samples are depleted in CaO and Na_2O, also marking them for unusual alkali loss and exchange. The plots indicate that the majority of Sr/Y ratios are reasonably consistent with a primary igneous origin and are meaningful for petrogenetic analysis.

A conclusion of this study is that these high-grade metamorphic rocks and even some tectonites, provided they are devoid of obvious veins, layering and alteration, retain many of their primary geochemical characteristics. We attribute this to the careful screening process in choosing the sample sites and to only limited flux of fluids in the fine-grained (volcanic) rocks during high-grade metamorphism, which occurred at a high rock/fluid-volume ratio. For clarity, several of the more strongly altered samples are not shown on diagrams that follow Appendix 2, Figure 26 and are so noted. These samples are shown on other diagrams, but alteration suggested by anomalous trace-element patterns is taken into consideration in any conclusions.

5.6 Neodymium-isotope composition

Most early volcanic and intrusive units in the main part of the Flin Flon Domain have yielded Nd-isotope ratios that indicate juvenile origins; that is, they formed from depleted mantle in an oceanic environment or on recently evolved Paleoproterozoic crust with little or no influence of older (Archean) continental crust (Stern et al., 1995a, b; Syme et al., 1998; Whalen et al., 1999). Some early plutons, however, are strongly contaminated with Archean crust (or crust with an extended time-integrated history of REE enrichment), and some volcanic assemblages are slightly contaminated. Late, postcollisional intrusive rocks are variably contaminated.

The rare earth elements Sm (a parent isotope) and Nd are preferentially enriched in the same phases, have small bulk partition coefficients for most petrogenetic processes and are therefore relatively immobile, even at high metamorphic grade. Thus, Nd isotopes yield additional data that relate to the primary characteristics of the metamorphosed igneous units, and provide important information about their source(s) of melting. In the Kississing–File lakes area, they can be used to confirm that geochemically similar units are comagmatic or at least formed by melting of the same source(s) under the same conditions.

The Nd isotopic evolution is expressed as epsilon units, where

$\varepsilon_{Nd} = ((^{143}Nd/^{144}Nd_{initial}/^{143}Nd/^{144}Nd_{CHUR,\ t}) - 1) \times 10^4)$. Relative to CHUR (chondritic uniform reservoir = bulk Earth composition), the depleted mantle shows increasing $^{143}Nd/^{144}Nd$ (increasingly positive ε values) with time, whereas the continental crust extracted from the mantle shows increasingly negative ε values with time. The time (t) of the calculated ε_{Nd} is the crystallization age of the igneous unit.

Neodymium-isotope ratios and ε_{Nd} were determined for representative samples of the major volcanic assemblages and intrusive units in the Kississing–File lakes area. Time of crystallization of the early volcanic units was taken as 1890 Ma, similar to the average age of early volcanic assemblages in the Flin Flon Domain. The ages of the two early successor-arc intrusions are taken as 1870 and 1860 Ma, approximations of their U-Pb age. The age of Missi Group volcanic units is taken as 1840 Ma, late successor-arc intrusions as 1830 Ma and synkinematic granite as 1820 Ma. The uncertainty in ε_{Nd} is about 0.5 and little affected by limited changes in age.

Epsilon-neodymium has a wide overall range of values for the tested units in the Kississing–File lakes area but generally a very limited range for each individual assemblage from which more than one sample was analyzed. Variations are systematic and related to the tectonic environment in which the assemblages were formed. Most ε_{Nd} values are >0 and indicate little contamination by Archean continental crust, consistent with the crustal history of the main parts of the Central Flin Flon, Snow Lake and Batty Lake subdomains (Table 15; Appendix 2, Figure 29f). The MORB-like Fussey Lake basalt and most of the assemblages with weak arc signatures (unaltered) show near-mantle values (ε_{Nd} = +4.6 to +5.3). This is consistent with their proposed back-arc and oceanic arc-rift origins (Stern et al., 1995a). Assemblages with strong arc signatures (including the early arc-volcanic rocks, the Sherridon–Meat Lake assemblage, the Missi Group volcanic rocks and the late successor-arc intrusions) show little or slight contamination (ε_{Nd} = +1.4 to +3.5). Ponton Lake basalt and the chilled base of a Josland Lake sill (altered?) are slightly more contaminated with older crust than most assemblages. The geochemical signatures of these units are consistent with arc rifting, but their ε_{Nd} values of +1.2 to +1.5 suggest the presence of local Archean basement rather than contamination by Archean detritus. The values are consistent with the presence of thin Archean crustal slices in major shear zones produced during early accretion in the Amisk Collage (Stern et al., 1995b). An Archean melt component is also suggested by the low ε_{Nd} values (–2.1 to +1.1) of two early successor-arc plutons. One sample of Puffy Mine andesite may have been contaminated with leucosome that was partly derived from the adjacent contaminated Ragged Lake granite or from the same Archean source that contributed to the granite. Synkinematic intrusions are slightly contaminated (ε_{Nd} = –0.6 to +1.6) and probably had a sedimentary melt component from a mixed source such as the Burntwood Group. Late quartz diorite has the lowest ε_{Nd} value in the area (–5.6) and may postdate collision with the Archean Sask craton. These largely mantle-derived intrusions are discussed in more detail in Section 5.10.1.3.

Table 15: Neodymium-isotope compositions and ε_{Nd} values of rocks from the Kississing–File lakes area.

Sample	Unit	Sm (ppm)	Nd (ppm)	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	Uncertainty (±σ)	TDM(1) (Ga)	~t(2) (Ma)	ε_{Nd}(3)	$^{145}Nd/^{144}Nd$
07-95-1219-1-1	Fussey Lake basalt	2.01	5.42	0.223836	0.513219	0.000005	N/A	1890	4.77	0.348405
12-93-7495-5	Ponton Lake basalt	2.76	9.21	0.181260	0.512510	0.000006	N/A	1890	1.26	0.348430
12-93-7394-B	Ponton Lake basalt	3.22	10.17	0.191215	0.512633	0.000006	N/A	1890	1.25	0.348426
12-92-2729-2	Moody Lake basalt	1.40	3.34	0.254201	0.513613	0.000010	N/A	1890	5.09	0.348421
12-92-2729-1	Moody Lake basalt	3.15	8.36	0.228001	0.513297	0.000006	N/A	1890	5.28	0.348420
07-00-1725-1-1	Preaston andesite	4.47	12.25	0.220586	0.513170	0.000005	N/A	1890	4.61	0.348429
F07-1619-1	Dickstone rhyolite	5.52	16.41	0.203163	0.513061	0.000006	N/A	1890	6.72	0.348431
F07-7006-1	Josland Lake sill, tonalite	6.58	20.25	0.196424	0.512842	0.000008	N/A	1890	4.07	0.348436
F07-551-1	Chilled base	1.62	4.50	0.217412	0.512970	0.000006	N/A	1890	1.45	0.348426
12-93-7713-1	Puffy Mine andesite	1.35	5.30	0.153811	0.512229	0.000004	N/A	1890	2.44	0.348400
12-93-7715	Puffy Mine andesite	2.10	9.56	0.132945	0.511771	0.000005	2.61	1890	-1.45	0.348429
12-96-4835	Ragged Lake granite	4.99	20.82	0.144945	0.511895	0.000004	N/A	1870	-2.07	0.348428
12-96-4838	Archie Lake tonalite	2.99	19.40	0.093027	0.511426	0.000004	2.18	1860	1.12	0.348428
12-92-3448-1	Meat Lake amphibolite	1.02	3.68	0.167344	0.512396	0.000009	N/A	1855	2.27	0.348422
12-92-3426	Sherridon rhyolite gneiss	2.91	12.07	0.145646	0.512072	0.000009	N/A	1855	1.12	0.348430
12-SH-08-1	Sherridon felsic gneiss	3.11	10.88	0.172925	0.512517	0.000007	N/A	1855	3.31	0.348411
12-SH-08-2	Sherridon felsic gneiss	2.62	8.89	0.178238	0.512588	0.000007	N/A	1855	3.44	0.348412
12-SH-08-4	Sherridon felsic gneiss	2.32	8.39	0.167088	0.512448	0.000009	N/A	1855	3.37	0.348423
12-SH-08-5	Sherridon felsic gneiss	2.35	9.58	0.148169	0.512245	0.000009	N/A	1855	3.91	0.348426
12-SH-08-6	Bob gneiss	3.48	17.81	0.118286	0.511807	0.000005	2.16	1855	2.48	0.348417
12-93-3786-5	Kississing River basalt	2.71	10.35	0.158350	0.512349	0.000006	N/A	1840	3.44	0.348441
12-88-6997	Evans Lake basalt	3.68	17.40	0.127822	0.511942	0.000005	2.16	1840	2.70	0.348423
12-93-3667	Jungle Lake basaltic andesite	8.57	45.74	0.113317	0.511787	0.000007	2.08	1840	3.11	0.348424
12-83-394	Lobstick rhyolite	7.10	30.60	0.140324	0.512127	0.000006	N/A	1840	3.37	0.348426
7-87-1770-1	Ham Lake pluton	3.76	21.98	0.103329	0.511680	0.000005	2.04	1830	3.25	0.348429
F07-953-1	Nelson Bay pluton	7.63	40.17	0.114778	0.511784	0.000003	2.12	1830	2.60	0.348425
12-96-9409-2	Dow Lake quartz diorite	11.8	58.83	0.120793	0.511438	0.000005	2.81	1830	-5.60	0.348426
12-93-3613	Burntwood Lake tonalite	2.31	12.84	0.108981	0.511665	0.000005	2.17	1830	1.63	0.348427
12-84-1199	Syntectonic granite	4.70	18.42	0.154332	0.512158	0.000005	N/A	1820	0.54	0.348424

(1) TDM = integrated time of separation from depleted mantle

(2) ~t = estimated or approximate time of crystallization

(3) ε_{Nd} = epsilon unit at time t

5.7 Early volcanic to successor-arc assemblages

In the Kississing–File lakes area, eleven metavolcanic units belonging to the early assemblages and six units in the younger Sherridon–Meat Lake assemblage were identified based on field and geochemical criteria. Their subdivision is based on dominant geochemical characteristics, inferred tectonic environment of deposition, and intensity of hydrothermal alteration. These categories include a unit with geochemical characteristics similar to modern MORB and others with weak to prominent characteristics of oceanic-arc volcanic rocks. The units are most easily distinguished using chondrite-normalized REE and N-MORB–normalized extended-element plots, discussed in this section and summarized in Section 6.1.3. Element plots of the various assemblages are shown on adjacent diagrams to facilitate comparison. Arc tholeiite of mafic–intermediate composition and felsic calcalkaline rocks are most abundant (Appendix 1, Figure 21a, b, d). Distinguishing characteristics are the TiO_2/MgO ratio (Appendix 1, Figure 22a–b), and the Th/Nb/Zr ternary ratio (Appendix 2, Figure 30a–c). Because of local Th mobility but retention of the systematic behaviour of La (Appendix 2, Figure 26), a tectonic discriminant diagram using La/Y/Nb was used to provide further evidence that a number of units are geochemically transitional between volcanic-arc tholeiite (VAT) and MORB Appendix 2, Figure 31a). Among the early volcanic assemblages, the MORB-like unit and those with weak arc signatures have high ε_{Nd} similar to proposed Flin Flon MORB mantle (Stern et al., 1995a). Units with strong arc signatures show an estimated 0–4% contamination by Archean material (Appendix 2, Figure 29f).

Some units of amphibolite, particularly those near the base of the highly attenuated succession at Dow Lake, have lost their primary structure and are separated into different units of arc tholeiite, based mainly on geochemical data. The conclusion that these represent primary volcanic units, even where they have a limited strike length and only two or three samples were analyzed for trace elements, is strongly supported by their individually uniform multi-element patterns. Nevertheless, because no regression analysis could be carried out to identify outlying values for these lightly sampled minor units, generalizations of their geochemical characteristics and origin are necessarily tentative. All units between Puffy Lake and Dow Lake display alkali-element alteration, as indicated by the large variation of Na_2O contents (Appendix 4, Table 10, DVD), and this alteration may have also affected some LREE. The mobility of Na_2O and SiO_2 is predicted from the development of minor tonalitic leucosome directly north of the sampling corridor.

The different geochemical categories are interpreted to have formed in correspondingly different tectonic environments above Paleoproterozoic oceanic subduction zones. These interpretations are discussed with the data presented for the various geochemical subdivisions and for individual major units in this section and Sections 5.8–5.9. The MORB-like composition is interpreted to represent back-arc-basin basalt (BABB), whereas compositions with a weak arc signature, transitional between MORB/BABB and VAT, are interpreted to have formed at the margin of such a basin or in a rifting arc. Volcanic rocks with a strong arc signature are interpreted as products of the main arcs. These conclusions are supported by the position of data on the plots of numerous multi-element diagrams that show generalized fields derived from modern tectonic environments. Individual units in the Kississing–File lakes area are also compared to specific Recent to Mesozoic volcanic formations that serve as tectonic analogues. Processes of fluid-fluxed melting and decompression melting, as well as speculative mantle and crustal sources of melts, are suggested to be similar to those in the analogous modern volcanic rocks. Because such conclusions are poorly constrained by our limited data, they are necessarily speculative and several options are therefore generally provided. Nevertheless, some generalizations can be made wherein specific processes with possible metallogenic implications lead to the specific chemical fingerprints found in this study.

5.7.1 MORB-like assemblages

The volcanic assemblage northeast of Reed Lake (including the Fussey Lake basalt) comprises uniform pillow basalt, generally devoid of interlayered sedimentary rocks or felsic volcanic rocks. These basalt units (Northeast Reed assemblage) have an affinity with modern MORB (Syme and Bailes, 1996).

5.7.1.1 Fussey Lake basalt (unit 1a)

A monotonous, possibly 2.5 km thick succession of massive and pillowed mafic volcanic flows (Fussey Lake basalt) occupies large northerly trending folds within a fault-bounded block south of File Lake (column 9C–9C' in Figure 8c, back pocket). It is surrounded by the (younger) Burntwood Group metagreywacke on File Lake and the west arm of Woosey Lake. Very sparse data from pillow tops raise the possibility that the structurally lower part of the 'succession' faces southeast and therefore that not much more than a 1 km thickness may be exposed. No samples were taken for geochemical analysis in this southern area.

Fussey Lake basalt shows little variation in MgO, TiO_2 or Al_2O_3 contents (Appendix 1, Figure 22a, e); the rocks fall into the field of MORB or BABB. The LILE do not the show significant enrichment that would be typical of arc assemblages (Appendix 1, Figure 23a; Appendix 2, Figure 28a, e). Although it is somewhat Nb depleted, the unit falls mainly into the field of N-MORB on plots of Th/Yb versus Nb/Yb and Th/Nb/Zr (Appendix 2, Figure 29a, 30a); the samples shown to be altered on Appendix 2, Figure 25a–c are omitted from these tectonic diagrams. Contents of REE (Appendix 2, Figure 24a), as well as Th/Yb and Nd/Yb ratios (Appendix 2, Figure 29a), are similar to N-MORB. Most samples of Fussey Lake basalt plot at the apex of the BABB field on the La/Y/Nb diagram (Appendix 2, Figure 31a); all samples are shown, including one enriched in Nd, probably through carbonate or calcsilicate veining. A chondrite-normalized plot of REE shows a tight distribution, with slightly depleted LREE compared to MREE (Appendix 3, Figure 32a), a distinctive pattern that is not seen in typical arc-basalt units (e.g. Appendix 3, Figures 32g–33d). An extended-element plot shows a negative Nb anomaly (for all but the altered sample) and slightly depleted Zr and Hf (Appendix 3, Figure 34a). These are typical features of rocks interpreted as back-arc–basin basalt (BABB) elsewhere in the Flin Flon Domain (Stern et al., 1995a) and in the more modern environment (Sinton et al., 2003). The mantle-like ε_{Nd} value of 4.8 is consistent with the origin of the basalt in an oceanic back-arc environment.

5.7.2 Volcanic rocks with a weak arc signature (of arc-rift or back-arc basins)

Laterally extensive units of aphyric basalt and fine-grained gabbro (Ponton Lake basalt and Moody Lake basalt) on the northern and eastern margins of the Amisk Collage show field characteristics similar to those of the ocean-floor units in the Central Flin Flon subdomain MORB (Stern et al., 1995a; Syme et al., 1999) but have chemical characteristic that are intermediate between MORB and arc tholeiite. They contain no known interlayered sedimentary or felsic volcanic rocks, as would be expected in volcanic-arc assemblages; rather, they comprise pillow basalt and abundant high-level sills or thick flows more typical of extensional tectonic regimes. Another thick and laterally extensive volcanic succession with very similar geochemical characteristics extends from an area west of Morton Lake (column 9A in Figure 8c, back pocket) to Loonhead Lake. This succession forms part of the Fourmile Island assemblage, which includes mafic to felsic units (Storozuk volcanics, Preaston andesite and Dickstone rhyolite). As a group, the volcanic rocks with a weak arc signature have many similarities with the MORB-like Fussey Lake basalt and are possibly related to this ocean-floor unit in their origin. Unlike most ocean floor, however, they show a wider range of fractionation and evidence, although limited, of fluid-fluxed melting related to subduction.

With the exception of one sample of Fussey Lake basalt (as discussed) and four samples of Ponton Lake basalt (taken near calcsilicate alteration domain), the MORB-like to weakly arc-like samples plot in tight clusters or bands of data on trace-element variation diagrams. The plots suggest little alteration of trace and minor elements, and can be considered to provide reliable geochemical evidence for their original tectonic origin.

On a TiO_2 versus MgO diagram (Appendix 1, Figure 22a), Storozuk and Preaston basalt and andesite flows have strong Fe fractionation and a wide range of MgO and TiO_2, and plot within the field of modern BABB (after data from Sinton et al., 2003). They overlap arc basalt that has somewhat elevated TiO_2. Moody Lake basalt shows even wider Fe fractionation and more elevated TiO_2. Scatterplots (Appendix 1, Figure 23a–c) show that the rocks are low-K tholeiite, with up to 30% of samples falling slightly above the regression lines (i.e., indicating minor K_2O contamination). The units have a wide range of Mg#, with fractionation causing slightly increased La/Yb (Appendix 2, Figure 25). Ratios of La/Zr show a linear plot against Mg#, indicative of little-altered comagmatic suites, but Th/Nb ratios show scatter due to alteration. The units are characterized by variable Th/Nb/Zr ratios that plot across the boundary between the fields of modern arc basalt and MORB and into the field of high Zr/Th arc-volcanic rocks (Appendix 2, Figure 30a). The contents of Ni, Cr and V are similarly intermediate between arc tholeiite and MORB.

There is a general change in geochemical affinity of the Fourmile Island assemblage from arc like at Reed Lake, typical of arc-volcanic rocks in the main parts of the Flin Flon Domain (Syme and Bailes, 1996), to more MORB like to the northwest at Moody Lake. The transition in geochemical affinity is illustrated in the La/Y/Nb diagram (Appendix 2, Figure 31a): Preaston andesite lies within the VAT field; the Storozuk volcanics plot with Fussey Lake N-MORB–like basalt at the apex of the BABB field; Moody Lake basalt is slightly more depleted in La at the border of the N-MORB field; and Ponton Lake basalt lies near the Nb-rich corner of the BABB field. Data presented in this section suggest that this regional change represents the transition from a central-arc to an arc-rift and back-arc environment, a conclusion that has economic implications, as is discussed in Section 6.1.3.

5.7.2.1 Ponton Lake basalt (unit 2a)

Ponton Lake basalt forms an approximately 1.5 km thick, northwest-facing, sediment-free succession of basalt flows and comagmatic sills south of Ponton and Puffy lakes. It may form a northern extension of the Elbow-Athapapuskow assemblage, structurally underlain by andesite and unconformably overlain by the Missi Group (column 2A–2A' in Figure 8b, back pocket). It is considered to be part of the Amisk Collage because it is intruded by satellite plutons of the 1890–1876 Ma Gants Lake Batholith.

The type unit comprises mafic–ultramafic flows and shallow intrusions south of Puffy Lake. The mafic flows weather dark grey-green and those with the highest Mg content weather pale brown. They are dominantly aphyric, pillowed and amygdaloidal, with lighter grey-green diopside-epidote alteration. Pillow rinds have a plagioclase-rich core with amphibole-rich margins (Figure 9b). Hypabyssal intrusions, ranging from differentiated mafic sills to microdiorite dikes, are considered to be, respectively, syn- and postvolcanic.

More coarsely recrystallized and deformed pillow basalt, which is otherwise identical to the north-facing flows, forms the roof to the Archie Lake pluton in the south-facing section north of Ponton Lake (column 2B in Figure 8b). Ten kilometres to the east, near Martell Lake, two screens of foliated fine-grained amphibolite, 200 and 300 m thick, structurally overlie mafic tectonite on the northeastern margin of the Gants Lake Batholith (column 8A in Figure 8b). Sheared, hornblende-rich selvages are preserved in parts of the unit at Martell Lake that are recognized as pillowed. The geochemistry of the amphibolite in the eastern, central and northern outcrop areas indicates that they are all part of the same unit, even though the batholith separates the eastern and central parts, and the late successor-arc units in the core of the Ponton Lake synform separate the central and northern parts (Zwanzig, 1999; Maps GR2010-1-1 and -2, back pocket/DVD). Because of the strong recrystallization and deformation in the northern and eastern outcrop areas, no correlation would have been possible, however, without first obtaining the major- and trace-element geochemistry that characterizes this unit.

Ponton Lake basalt, which has primitive to strongly fractionated members, shows the widest range of MgO contents (5.6–14.8%) and among the highest TiO_2 contents (<1.83%) of the units in the Kississing–File lakes area (Appendix 1, Figure 22a). Geochemical analyses of Ponton Lake basalt straddle the E-MORB and arc fields (Appendix 2, Figure 30a), or the BABB and E-MORB fields (Appendix 2, Figure 31a) on tectonic element-ratio diagrams. The three most Th-enriched samples, which plot well off the regression line in Appendix 2, Figure 25c and have been interpreted as altered by gain of Th, are not shown on Appendix 2, Figure 30a. The anomalous samples do not occur in one particular area or stratigraphic interval, and are thus not considered to be a separate unit. Lanthanum, which is plotted in Appendix 2, Figure 31a, is considered to be less prone to alteration under high-grade metamorphic conditions than Th. All data are shown but only the central cluster is considered to be meaningful. The REE profile shown on a chondrite-normalized diagram is smooth and negatively sloped to flat, showing increased total REE and slightly fractionated LREE with decreasing MgO (Appendix 3, Figure 32b, all samples shown). The most primitive sample is from a pillowed flow at the top of the unit in the northwest. The most fractionated flows (Fe-rich basaltic andesite) occur in the east. Increased LREE in the fractionated samples is shown by the ratio $[La/Yb]_N$ ranging from 0.73 to 2.5, with the low numbers in the high-Mg basalt (MgO = 12–15%) and the highest in the basaltic andesite. Weak negative HFSE anomalies and a prominent Th spike (only partly due to the anomalous samples) are evident on MORB-normalized extended-element plots (Appendix 3, Figure 34b). The Th/Nb_N ratios increase (2.4–4.0, excluding altered samples) concomitantly with $[La/Yb]_N$ ratios. These characteristics are highly distinctive in the map area and are taken as evidence that the analyzed samples belong to a single comagmatic unit.

Slight enrichment LREE, as in Ponton Lake basalt, is a feature observed in several modern back-arc basins (Stern et al., 1990; Fretzdorff et al., 2002). The elevated Th and the HFSE troughs, although shallow, suggest that this basalt formed close to the main arc (e.g., Gribble et al., 1998). The geochemistry is similar to that of the Scotty Lake (contaminated) basalt in the southwestern part of the Central Flin Flon subdomain (Syme et al., 1999). The occurrences of the Ponton Lake basalt directly north of Elbow Lake, as well as on the opposite side of the Gants Lake Batholith, suggest that these units may have formed as part of the Elbow-Athapapuskow E-MORB assemblage in the same back-arc basin (Maps GR2010-1-1 and -2). The geochemistry of Ponton Lake basalt is particularly similar to McDougalls Point basalt (Stern et al., 1995a). The high 'stratigraphic' position and a similarity with the 1886 Ma Josland Lake gabbro, however, suggest that the Ponton Lake basalt may be younger than the Elbow-Athapapuskow assemblage but had a similar mantle source.

The trace-element variations suggest that another process, in addition to fractionation, has affected the magma. Data in Appendix 2, Figure 24a indicate that Ponton Lake basalt has two different geochemical trends. The high-Mg basalt flows have a weak $[La/Yb]_N$–enrichment trend at constant Yb, and this is followed by an absolute Yb-enrichment trend due to fractionation at constant $[La/Yb]_N$. Melting of variably contaminated mantle, or crustal assimilation followed by high-level fractionation may explain these trends as well as the elevated Th and LREE with slight depletion of HFSE. These processes may also explain the slightly higher SiO_2 and Ba contents of Ponton Lake basalt compared to other BABB (Appendix 1, Figure 23i), as well as the high ratio of Th/La (average of 0.19, compared to 0.08 and 0.10 for Fussey Lake and Moody Lake basalts). High Nb/Yb ratios suggest an enriched mantle source (Appendix 2, Figure 29a) and/or crustal contamination. The occurrence of the highest Nb/Yb (2.1 and 2.9) in the most fractionated rocks (Mg# of 43 and 45) but lower Nb/Yb ratios (1.5 and 2.0) in primitive basalt (Mg# of 71 and 74) is consistent with weak contamination of a high-Mg parent during fractionation. A trace-element pattern similar to primitive mantle (which has Nb/Yb of 1.49) indicates an undepleted mantle source even for the least fractionated flows (Appendix 3, Figure 34b). A high degree of partial melting is indicated by the high Mg# in the most primitive flows (<74 in flows and <81 in sills).

Ponton Lake basalt has only moderately positive ε_{Nd} values (+1.2 to +1.3), which indicate the presence of an older crustal component in the contaminant. About 4% contamination by Neoarchean rock similar to that found at the margin of the Scotty Lake block (David and Syme, 1994) is consistent with the Nd isotope composition (Appendix 2, Figure 29f). Isotopically evolved Paleoproterozoic rock is an alternative contaminant. The high-Mg basalt also shows anomalous values of Pb and a trend from low contents of Th to the high contents found in the fractionated Ponton Lake basalt flows (Appendix 4, Table 10, DVD). The mantle source(s) may have been variably contaminated with older crustal material, and different batches of melt acquired different levels of contamination.

5.7.2.2 Moody Lake basalt (unit 2b)

The Moody Lake basalt comprises pillowed and lesser massive flows and high-level gabbro that are best exposed southwest of Moody Lake in the saddle structure between the Moody Lake granodiorite dome (Zwanzig and Schledewitz, 1992; column 7B in Figure 8d, back pocket) and the Gants Lake Batholith southeast of Evans Lake. This basalt can be traced northwest into fine-grained amphibolite with layered structure. The basalt forms the main units in the northwestern part of the Fourmile Island assemblage, which separates the Amisk Collage from the Burntwood Group to the east and north. Near Moody Lake, the basalt has a minimum thickness of 300 m, but its stratigraphic base is not exposed and the unit is truncated on the north and south sides of the saddle along the sheared, overturned unconformity with the Missi Group (Zwanzig, 1999; column 7B in Figure 8d). The Moody Lake basalt is interpreted to be stratigraphically underlain by Martell Lake basalt (an arc tholeiite) southeast of Moody Lake. Southeast of Dow Lake, the Moody Lake 'basalt' includes Fe-rich basaltic andesite. In an area of poor exposure close to Loonhead Lake, unexplored in this study, the contact between Moody Lake basalt and the Storozuk volcanics is unexposed, so the thickness there of 400 m is also a minimum. A small outcrop area on Morton Lake, west of the Morton Lake Fault, has furnished one slightly altered sample correlated with the main outcrop areas and overlain by west-facing Storozuk volcanics (column 9B in Figure 8d). This relationship is also suspect because it lies close to the Morton Lake Fault. Moody Lake basalt west of Moody Lake is only 5–70 m thick in the northern areas. In those areas, the basalt and the unconformably overlying Missi Group occupy the Loonhead Lake thrust sheet, which was emplaced over the Burntwood Group along the Loonhead Lake Fault and subsequently overturned to the southwest (Zwanzig, 1999; Section 4.2.3). The Moody Lake basalt is locally cut out at Dow Lake, along the D_5-reactivated part of the Loonhead Lake Fault (Figure 8a).

The Moody Lake basalt generally occurs as dark green amphibolite with hornblende (0.2–3.0 mm) slightly in excess of plagioclase (Zwanzig and Schledewitz, 1992). Fine-grained amphibolite was derived from aphyric and finely plagioclase-phyric basalt. Medium-grained massive amphibolite is locally identified as sills, dikes or thick flows. The basalt retains local remnants of pillow rinds or is weakly layered to uniform. Pillows are highly flattened where recognized and have hornblende-rich rinds; the facing direction is not known except on Morton Lake. Thin layers and lenses rich in epidote or diopside, which represent transposed alteration domains in the basalt, are common south of Moody Lake.

The Moody Lake basalt is intruded by large sills of Josland Lake gabbro throughout its 80 km long eastern outcrop belt. These sills make up the major part of the Amisk Collage in the Loonhead Lake thrust sheet, so the basalt forms only a thin roof zone to the sills, stratigraphically below the Missi Group. The Dismal Lake suite contains local members at its margins that have a geochemical signature similar to that of chilled Josland Lake gabbro, thus strengthening the correlation of the entire package.

Despite the generally poor preservation of primary structures, the unit is recognized as a variably fractionated tholeiite with a geochemical signature transitional to MORB but different from Ponton Lake basalt (Appendix 1, Figure 22a). Its distinctive chemistry allows it to be traced for 50 km from Morton Lake north and west to Puffy Lake, and structural mapping indicates that it probably continues for another 30 km to the centre of Kississing Lake. South of Kisseynew Lake, a chemically nearly identical amphibolite unit (Dismal Lake suite of Gilbert, 2004) extends an additional 20 km west of Lobstick Narrows, making the total strike length of Moody Lake–type basalt a minimum of 100 km. The belt south of Kisseynew Lake, which forms the Dismal Lake suite, is up to 4 km wide and probably contains a thickness of more than 1000 m of amphibolite correlative with the Moody Lake basalt.

Samples of Moody Lake basalt plot across the boundary between fields of modern arc tholeiite and N-MORB. For example, they fall in the range of BABB on tectonic compatible and incompatible element-variation diagrams (Appendix 2, Figure 30a, three anomalous samples omitted). The ratios among Ni, Cr, and V (not shown) are also transitional between MORB and Arc. The basalt is typified by smooth, positively sloped, convex-upward REE plots (Appendix 3, Figure 32c). The plots also suggest that samples near Preaston and Storozuk lakes should be included in the Moody Lake basalt. Chondrite-normalized $[La/Yb]_N$ ratios are very low (0.29–0.48), with average Yb of 2.68 ppm (similar to N-MORB) and moderately low Yb7 (2.1 ppm, estimated at MgO of 7%; Appendix 2, Figure 24a). The unit has small negative HFSE anomalies and a Th spike on MORB-normalized plots (Appendix 3, Figure 34c). These distinctive features for basalt in the Flin Flon Domain make it an outstanding marker unit for structural mapping.

The geochemistry of Moody Lake basalt is atypical of volcanic-arc basalt but nearly identical to low-K basalt extruded at the Manus back-arc–basin spreading centre (type M1, classified as MORB in Sinton et al., 2003) and similar to arc-rift basalt that was interpreted to have formed in a transtensional fore-arc in central California (Snow, 2007). The presence of arc-rift basalt elsewhere in the Flin Flon Domain was suggested by Syme et al. (1999). The geochemical similarity with the adjacent Storozuk volcanics and the interlayering with SE Dow porphyritic basalt (Sections 5.7.2.5 and 5.7.2.6) indicate that the Moody Lake basalt was part of the same volcanic chain as the enclosing units. Its presence may represent the onset of arc rifting in the volcanic pile.

Moody Lake basalt is interpreted to have formed from a depleted mantle source. Although it might have formed by a high degree of melting to achieve its low contents of incompatible elements, it was most likely derived from a highly depleted mantle source that yielded lower REE parent magma than modern N-MORB. This source must have undergone previous melting, either at an early ridge or within the arc or growing back-arc basin prior to rifting at Moody Lake.

High ε_{Nd} values (+5.1 to +5.3), similar to the most depleted Flin Flon Domain mantle (Appendix 2, Figure 29f), support the MORB-like origin of the Moody Lake basalt. The high ε_{Nd} and low Nb/Yb indicate that it had a different mantle source (depleted) compared to the enriched, somewhat contaminated Ponton Lake basalt (Table 15). Thick layers of tectonite separate these units northwest of Martell Lake, suggesting a major break between them.

5.7.2.3 Preaston andesite (unit 2c)

The Preaston andesite forms the lower part of the Fourmile Island assemblage on the west limb of the Yakymiw syncline, where Bailes (1980a) estimated a thickness of 2 km of mafic to intermediate volcanic rocks recrystallized to upper greenschist facies and called the Preaston formation. These rocks are predominantly massive amphibolite but locally contain pillows, phenocrysts of plagioclase and thin layers of breccia where they were examined in the upper part of the succession (Bailes, 1980a). Southwest of Storozuk Lake, a transition from basalt to andesite is interlayered with thin felsic units near the top. This is overlain by Dickstone rhyolite and Storozuk volcanics (column 9A in Figure 8c, back pocket). The upper part of the Preaston andesite has been resampled and analyzed. The lower part has not been mapped and may contain unknown units east of the West Reed–North Star Lake Shear Zone.

All sampled rocks in the upper section are basaltic andesite (54–57% SiO_2) and are chemically similar to the most fractionated members of the Moody Lake basalt (Appendix 4, Table 10, DVD in back pocket). In fact, a sample collected as Preaston andesite and another as Storozuk volcanics have been regrouped with the Moody Lake basalt and are taken as evidence that these units are interlayered and possibly consanguineous.

Although scattered by element mobility, Ba contents are low, similar to BABB (Appendix 1, Figure 23i)., Preaston andesite is distinguished from Moody Lake basalt, however, by its lower content of TiO_2 at the same MgO content (Appendix 1, Figure 22a) and marginally higher ratios of La/Yb with slightly lower Yb7 (Appendix 2, Figure 24a). In plots of La/Yb, La/Zr and Th/Nb versus Mg#, only values that do not fall close to regression lines (and are considered to be altered) fail to show higher ratios for Preaston andesite (Appendix 2, Figure 25b–c). Although alteration plays a crucial role in scattering the data, when Zr is taken as the independent variable representing fractionation, relatively high values of R^2 in a regression of La indicate that Preaston andesite has about 65% higher La content than Moody Lake basalt and is therefore tentatively interpreted as derived from different parent magma. The higher La, Ce, Th/Yb and Th/Nb represent a greater subduction-zone component.

Preaston andesite data plot in the arc tholeiite field on a Th/Zr/Nb diagram (Appendix 2, Figure 30a) and a La/Y/Nb diagram (Appendix 2, Figure 31a). Although it shows the same range of fractionation of REE as Moody Lake basalt, it has slightly more pronounced negative HFSE anomalies, consistent with a stronger arc affinity (Appendix 3, Figure 32d, 34d). Similar flows, but with less fractionated LREE and weaker Ti anomaly, have been dredged from the Manus back-arc basin in the south Pacific (Sinton et al., 2003). The comparison suggests that the Preaston andesite was more influenced by subduction zone fluids or melts.

A sample of Preaston andesite has an ε_{Nd} value that is high (+4.6) but slightly lower than that of Moody Lake basalt and probably more influenced by subduction-zone contamination. Nevertheless, the juvenile Nd-isotope composition and weak arc-like character suggest that, like Moody Lake basalt, decompression in a rifting arc, as well as fluid fluxing (typical of arcs), contributed to the melting of the mantle source of this unit.

5.7.2.4 Dickstone rhyolite (unit 2d)

A mixture of dacite and rhyolite flows and fragmental rocks reaches a maximum thickness of 1500 m and a strike length of 15 km (Dickstone Formation of Bailes, 1980a). It is part of the steeply dipping, east-facing volcanic succession in the Fourmile Island assemblage, which forms the western limb of the Yakymiw syncline west of Morton Lake (column 9A in Figure 8c).

The three analyzed samples of rhyolite are relatively sodic (K_2O/Na_2O = 0.20–0.45), but at least two of them are depleted in alkali elements (Appendix 2, Figure 25d). The unit has elevated REE and lies on the fractionation trend of Preaston andesite (Appendix 2, Figure 24a), with positively sloped chondrite-normalized profiles similar to the underlying Preaston andesite and overlying Storozuk volcanics but showing a negative Eu anomaly related to feldspar fractionation (Appendix 3, Figure 32d). An extended-element plot shows a high negative Ti anomaly (Appendix 3, Figure 34d), typical for a fractionated felsic melt. These characteristics are good evidence that the Dickstone rhyolite formed by fractionation of the same parent magma as the hosting mafic and intermediate flows. Low Th/Yb makes the unit similar to the tholeiitic association in the southwestern part of the Central Flin Flon subdomain (Appendix 2, Figure 29c). High Mg# (35–52) and TiO_2 content at moderate Zr content (Appendix 2, Figure 29d) are typical of volcanic-arc rhyolite; however, total Y and Nb place it in the field of ocean-ridge granite (Appendix 2, Figure 25e). A sample of dacite fractionated from MORB on the Manus spreading centre (Sinton et al., 2003) has a MORB-normalized extended-element profile similar to that of the Dickstone rhyolite.

A very high ε_{Nd} value (+6.7) also indicates that the rhyolite was fractionated from a mantle-derived melt without influence from older crust. The exceptionally high value may be related to the fractionation of Nd and Sm in this very highly differentiated rock.

5.7.2.5 Storozuk volcanics (unit 2e)

An east-facing unit of mafic arc tholeiite, over 400 m thick, overlies Preaston andesite and Dickstone rhyolite at Storozuk Lake. The same unit faces west where it overlies probable Moody Lake basalt west of Morton Lake. Storozuk volcanics thus occur on both limbs of the Yakymiw syncline, which is cored in the volcaniclastic basin fill of the Yakymiw Formation (columns 9A and 9B in Figure 8c, back pocket; Bailes, 1980a). The Storozuk volcanics are heavily intruded by Josland gabbro sills, and the package of volcanic rocks and sills extends from Reed Lake 30 km north to Loonhead Lake, where exposure is poor and little mapping was done.

The Storozuk volcanics comprise basalt and andesite flows, intermediate breccia, and felsic flows and tuff (Bailes, 1980a). The type section is the steeply dipping, east-facing, 1200 m thick panel of predominantly pillowed flows southeast of Storozuk Lake. Metamorphic grade increases from upper-greenschist facies in the south to upper-amphibolite facies at Loonhead Lake. Early epidote and carbonate alteration is common in the pillows; fragmental units are locally gossanous.

Storozuk volcanics have the same geochemical characteristics as Preaston andesite. They are distinguished only by their stratigraphic position above the Preaston andesite and probably represent continued construction of the same oceanic volcanic rise. The three analyzed samples extend into the less fractionated (basaltic) fields on variation diagrams, but otherwise they plot on the same element-ratio trends as Preaston andesite (Appendix 1, Figure 23c; Appendix 2, Figures 24a, 25a, b, 26d). Storozuk volcanics also plot in the same arc-tholeiite field, with relatively low Th/Zr and low Nb, as Preaston andesite (Appendix 2, Figure 30a). Storozuk volcanics also have the same chondrite-normalized REE pattern with a weakly positive convex-upward slope, showing mild depletion of LREE, as Preaston andesite (Appendix 3, Figure 32e). A sample taken at Loonhead Lake allows the unit to be extended north of its type area. A MORB-normalized plot of Storozuk volcanics is flat, like N-MORB, but has the distinctive HFSE trough and Th spike of arc tholeiite (Appendix 3, Figure 34e). The Loonhead Lake sample has a pattern identical to that of Manus spreading-ridge basalt, classified as BABB by Sinton et al. (2003).

The presence of Moody Lake–type basaltic andesite, with its geochemical characteristics transitional to N-MORB, in the volcanic-arc succession that also contains Preaston andesite, Dickstone rhyolite and Storozuk volcanics suggests that at least two volcanic edifices were coeval and interlayered at their flanks.

5.7.2.6 SE Dow porphyritic basalt (unit 2f)

An approximately 130 m thick unit of basalt has a distinctive plagioclase-phyric texture that allows it to be traced for 5 km in the area southeast of Dow Lake. It includes massive and pillowed flows, as well as breccia with 5–20% plagioclase phenocrysts. Because the western contact is the West Read–Northstar Shear Zone (WRNSZ) and its eastern contact with the Dow Lake andesite is not exposed, the field relationship is uncertain. However, the unit is interlayered with aphyric basalt that is interpreted as part of the Moody Lake basalt. Nevertheless, there is a distinctive geochemical break between these interlayered units. The stratigraphic affiliation of the SE Dow porphyritic basalt is therefore uncertain, but its interlayering suggests that it is part of the Fourmile Island assemblage.

The basalt is depleted in TiO_2 (Appendix 1, Figure 22a), and has slightly elevated Al_2O_3 (Appendix 1, Figure 22e) and a low content of MREE to HREE (e.g., Yb in Appendix 2, Figure 24a), similar to rocks classified as arc basalt. Plots in NB/Yb, Th/Zr/Nb and La/Y/Nb space, however, are close to E-MORB (Appendix 2, Figure 29a, 30a, 31a), Moreover, the only two analyzed samples of the unit lack HFSE anomalies on MORB-normalized extended-element plots, and have no negative Zr anomaly, like other depleted arc tholeiite (Appendix 3, Figure 34b). Basalt with a similar extended-element profile has been classified as ‘MORB-like’ near the Kisseynew Domain margin north of Flin Flon (Gilbert, 2004, Figure GS-1-11e). The MORB-normalized profiles show depleted HREE, typical of arc-volcanic rocks. The low contents of HREE may be related, instead, to a breakdown of oxides along the WRNSZ in non-arc basalt. Plagioclase-rich gabbro in the Josland Lake sills, interpreted as an arc-rift unit, also has a very similar extended-element pattern (Section 5.7.6).

5.7.3 Oceanic-arc tholeiite (Amisk Collage)

About half of the early volcanic units in the Kississing–File lakes area have low TiO_2 contents, similar to modern arc-volcanic rocks (Appendix 1, Figure 22b). The SE Dow porphyritic basalt also has a low TiO_2 content, but is an exception (discussed above). The rest of the early units with low Ti are arc tholeiite, which is also characterized by low Zr and low Zr/Y, Th/Yb, Nb/Yb and La/Nb ratios. The early arc assemblages are distinguished in the field by the common presence of small garnet porphyroblasts caused by higher Al_2O_3/alkali element ratios than the MORB-related basalt units (compare A/CNK in Appendix 4, Table 10, DVD). Interlayered epiclastic rocks and breccia are scarce, and pillow structure is relatively abundant. These rocks occur in the Amisk Collage as well as in the (younger) Sherridon–Meat Lake assemblage. Four of the early mafic–intermediate volcanic units and two younger units are arc tholeiite. Their La/Yb ratios are generally higher than N-MORB and their Yb contents are usually lower (Appendix 2, Figure 24a). The slopes of chondrite-normalized REE plots are therefore generally negative to flat (Appendix 3, Figures 32g–33e).

5.7.3.1 *Puffy Mine andesite (unit 3a)*

A unit of Puffy Mine andesite about 300 m thick is preserved at the abandoned Puffy minesite in a large screen of Amisk Collage in the Sherridon–Hutchinson Lake Complex. It occurs between the Fire Lake granodiorite in the south and the Ragged Lake pluton in the north (Maps GR2010-1-1 and -2, back pocket/DVD). The andesite is unconformably overlain by the Missi Group in the core of a tight syncline (column 2C in Figure 8b, back pocket). Relatively well preserved pillows of the andesite were sampled, but much of the unit is strongly foliated and contains abundant calcsilicate alteration. The unit is intruded by the adjacent granitoid plutons, by the Puffy Lake dacite dikes and by mafic–ultramafic dikes.

Puffy Mine andesite contains 0–20% garnet, 0–10% biotite and commonly small amounts of leucosome veins. These characteristics suggest major-element and general LILE alteration, which is confirmed by the K_2O content of at least one sample (Appendix 1, Figure 23d). The anomalous sample is from a black-weathering alteration domain with high MgO. All samples are probably slightly altered or contaminated, producing trends of increasing La (e.g., Appendix 2, Figure 31b). Nevertheless, the La/Zr ratio is similar to that of other units with a strong arc signature (Appendix 2, Figure 26e), suggesting little alteration of the immobile elements. Near the surface exposure of the Puffy Lake gold deposit (abandoned), however, the rocks are visibly altered and contain abundant biotite and locally abundant garnet or calcsilicate minerals. These rocks were not analyzed.

Puffy Mine andesite has a range in Mg# (56–40) but shows no increase of TiO_2 with decreasing MgO (Appendix 1, Figure 22b). There is variation in La/Yb rather than a clear fractionation trend of REE (Appendix 2, Figure 24a). Negative HFSE anomalies and elevated Th are clearly defined on a MORB-normalized plot (Appendix 3, Figure 34f), similar to arc tholeiite units in the main Flin Flon Domain (Stern et al., 1995b).

Two samples of Puffy Mine andesite yielded different ε_{Nd} values (+1.4 and –2.4, Table 15). The latter has anomalously high contents of LREE, Th, Pb and U that may have been added from an Archean source, possibly from the adjacent Ragged Lake pluton, which is highly contaminated with older crust (as described in Section 5.8.2.1). Alternatively, the parent magma may have become variably contaminated by melt from older crust at depth. The higher ε_{Nd} value of the more 'normal' andesite sample is typical of the units that have a strong arc-like geochemical signature with only a very minor Archean component.

5.7.3.2 *Koscielny andesite (unit 3b)*

A variably pillowed, locally brecciated andesite–basalt unit (Koscielny andesite) was mapped south of Puffy Lake by Schledewitz (1993c). This unit lies beneath Ponton Lake basalt on the south side of the Ponton Lake synform and is possibly correlative with Puffy Mine andesite. However, unlike Puffy Mine andesite, the Koscielny andesite is plagioclase and hornblende (after pyroxene?) phyric, and it is separated from the Ponton Lake basalt by a thin unit of plagioclase-phyric basalt and probably a fault (Schledewitz, 1993c). Koscielny andesite also has a slightly different LREE content than Puffy Mine andesite (Appendix 2, Figure 24a). Both andesite units, however, have among the highest Th/Yb and Nb/Yb ratios in the area (Appendix 2, Figure 29a) and a similar high Th/Zr ratio (Appendix 2, Figure 30b). The high scatter in Appendix 2, Figure 31b, however, indicates that alteration or contamination has affected the trace elements. Nevertheless, three of the samples for the two units have a similar flat REE pattern (compare Appendix 3, Figure 32g and 32h). The two samples of Puffy Mine andesite show similar Th spikes and generally similar negative HFSE anomalies in MORB-normalized plots (Appendix 3, Figure 34f). These are characteristics of island-arc volcanic rocks. The lack of a negative Nb anomaly, slightly elevated Hf and relatively low La in one sample are thus interpreted as the result of alteration and illustrate the problem of dealing with high-grade metamorphic rocks. Only the combined use of numerous elements in the various diagrams provides a satisfactory level of confidence that these andesite units have their origin in a volcanic arc and are possibly related.

The similarity of the Puffy Mine and Koscielny andesite units and the correlation of the Ponton Lake basalt on both limbs of the Ponton Lake synform (POS in Figure 12) are consistent with the conclusion (Zwanzig and Schledewitz, 1992; Zwanzig, 1999) that the volcanic rocks of the southern Central Flin Flon subdomain (Amisk Collage) are contiguous with the volcanic-derived gneiss and amphibolite in the Sherridon–Hutchinson Lake Complex. This makes the regional trend of these units north to northwest despite the superposed westerly structural trend.

5.7.3.3 *Martell Lake basalt (unit 3c)*

Very fine grained amphibolite, interpreted as metabasalt (Martell Lake basalt), overlies mafic and mixed tectonites and Burntwood Group in the extension of the West Reed–North Star Lake Shear Zone (column 8B in Figure 8c, back pocket). Thus, the unit forms the structural base of the Fourmile Island assemblage between Moody Lake and Dow Lake. About 150 m are exposed above its tectonite base at Martell Lake. The unit is structurally thinned and truncated along the basal shear zone north of Dow Lake.

Martell Lake basalt is defined mainly on the strength of its trace-element pattern. The unit is an example where geochemistry is a critical mapping tool in the strongly deformed, high-grade metamorphic rocks. Monitoring possible alteration using incompatible-element ratios is therefore particularly important to avoid spurious results. It is noteworthy that ratio plots of incompatible elements in Martell Lake basalt, despite the low number of analyses (three), have slopes and dispersions similar to the better sampled, less tectonized units (Appendix 2, Figure 25a–c).

Martell Lake basalt shows Fe enrichment (Mg# = 59–32) with somewhat elevated TiO_2, compared to modern arc tholeiite (Appendix 1, Figure 22b). It contains one sample of Fe-rich basaltic andesite that is also REE enriched (Appendix 2, Figure 24a) and slightly Nb enriched (Appendix 2, Figure 31), but all samples have the same concave-upward profile with a positive slope of MREE to HREE (Appendix 3, Figure 33a). The Fe-rich sample was collected close to the basal tectonite, but its chondrite-normalized pattern, which is identical to that of the other samples (but elevated), suggests a uniform increase due to fractionation but little alteration of REE. The slightly increased Nb is typical of fractionated (Fe-rich) tholeiite. The data also show HFSE, Zr and Hf depletion in an N-MORB–normalized plot that is typical of arc tholeiite in the Central Flin Flon subdomain (Appendix 3, Figure 34g).

5.7.3.4 Dow Lake andesite (unit 3d)

Dow Lake andesite is another fine-grained amphibolite defined largely on its geochemistry; however, the presence of small garnet porphyroblasts in this unit serves as a field-mapping tool. About 150 m of the andesite structurally overlie SE Dow porphyritic basalt southeast of Dow Lake (column 8B in Figure 8c, back pocket). The unit may also overlie Moody Lake basalt at the margin of the Loonhead Lake dome and thus occupy a saddle structure. All of these units are truncated along the overturned, northeast-dipping Loonhead Lake Fault in the structurally inverted section. Consequently, the Dow Lake andesite is probably older than the surrounding units.

Dow Lake andesite has a trace-element signature of volcanic-arc tholeiite (e.g., Appendix 2, Figure 31b). It has a flat slope of MREE to HREE and a weakly negative slope of LREE on a chondrite-normalized plot (Appendix 3, Figure 33b). This, plus negative HFSE anomalies on an N-MORB–normalized plot (Appendix 3, Figure 34b), are similar to Koscielny andesite, a pattern that is also typical of better preserved arc tholeiite in the southern parts of the Flin Flon Domain.

5.7.4 Amphibolite (unit 4), mafic to mixed gneiss (unit 4a) and felsic rocks (unit 4b)

Widespread narrow screens of amphibolite (unit 4) bordering various early plutons have generally not been analyzed but are assumed to be part of the early volcanic assemblages. A sample from the Sherridon structure (Appendix 2, Figure 24a) and unpublished surface data from Halo Resources Ltd. (L. Bloom, pers. comm., 2007) of the most common variety of low-Ti amphibolite have trace-element characteristics similar to the Meat Lake and Walton Lake amphibolites but with less fractionated LREE (Appendix 3, Figure 35a).

The belt of mafic–intermediate rock (unit 4a) extending southeast from Batty Lake represents part of the Walton Lake nappe. Some of these rocks are highly altered and were not analyzed; they are assumed to be equivalent to Walton or Meat Lake amphibolite and Sherridon gneiss because of their structural continuity from Fairwind Lake to Meat Lake. An apparently weakly altered sample, obtained from Folster Bay, resembles Walton Lake amphibolite (Appendix 2, Figure 24a) except for Nb (Appendix 3, Figure 35b). This sample was therefore tentatively included with unit 4d.

Felsic rocks (unit 4b) on the southern margin of the Kississing–File lakes area were not sampled, but their spatial association with amphibolite, local intermediate rocks and what may be volcaniclastic rocks indicates that they are part on an arc assemblage, their location in the south suggesting the Flin Flon assemblage.

5.7.5 Arc-derived amphibolite to felsic gneiss: Sherridon–Meat Lake assemblage (units 4c–5d)

Units of highly recrystallized amphibolite (units 4c–e) were sampled in the Sherridon–Meat Lake area, mainly in the Walton Lake nappe. These amphibolite units are interlayered with the regionally predominant felsic gneiss (Sherridon gneiss), local hornblende-plagioclase gneiss and calcsilicate rock—all part of the Sherridon–Meat lake assemblage. Amphibolite constitutes <10% of the section (e.g., column 5, Figure 8e, back pocket) in a strongly bimodal assemblage (Appendix 1, Figure 19a). The mafic units are 6–10 m thick and contain variable amounts of plagioclase and mafic phenocrysts. Flattened pillows and, locally, breccia structure and gabbroic texture are preserved in the three largest belts. These occur near the western, southern and northeastern margins of the main part of the Walton Lake nappe.

These volcanic-derived mafic to intermediate rocks, located on the most northerly inlier of the Flin Flon Domain, have undergone alkali mobility, as illustrated by the scatter on the K_2O versus TiO_2 plot (Appendix 1, Figure 23d). Nevertheless, plots of La versus Zr (Appendix 2, Figure 26g) indicate that even the highly recrystallized amphibolite in the Sherridon–Meat Lake assemblage can retain a coherent immobile-element pattern. It is characterized by particularly low TiO_2 (<0.5%, excluding one anomalous sample; Appendix 1, Figure 22b), and generally low REE (Appendix 2, Figure 24a). Immobile trace-element ratios clearly show the amphibolite units to be tholeiitic and the felsic units to be calcalkaline–transitional (Appendix 2, Figure 27b). Zirconium contents and Zr/Y ratios of the mafic rocks are very low, generally lower than but overlapping those of the Amisk Collage in the study area (Appendix 2, Figure 29e); they fall clearly in the oceanic-arc field. Ratios of Th/Zr are high but La/Y ratios are relatively low (Appendix 2, Figure 30c, 31c), the latter supporting a tholeiitic to transitional origin. The deep troughs of Nb and Zr-Hf, low contents of MREE to HREE, and fractionated LREE yield a distinctive MORB-normalized trace-element profile (Appendix 3, Figure 35a–b). Similar arc tholeiite in a modern environment is considered to be the product of melting of a source with a history of previous melt extraction (Gamble et al., 1993; Woodhead et al., 1993; Jolly et al., 2001) but involving the addition of Th-bearing fluid or melt from the slab.

This general trace-element pattern is shared with several units in the main part of the Flin Flon Belt, including the Threehouse basalt in the upper part of the arc assemblage at Snow Lake (Appendix 3, Figure 35b). However, the young U-Pb age of the interlayered felsic volcanic gneiss does not support a correlation. In addition, high-Ti amphibolite in the Sherridon structure, probably forming sills, has fractionated REE and is unrelated to the Josland Lake sills (Appendix 3, Figure 35c), but is probably the same as the syntectonic gabbro that intruded the Burntwood Group (Weldon Bay gabbro, Section 5.10.2.1), thus providing no minimum age for the Sherridon–Meat Lake assemblage. The presence of such amphibolite in the Kississing–File lakes area in an assemblage that has yielded a young zircon age suggests that the primitive oceanic-arc environment was maintained locally during early successor-arc magmatism. Local pillows, widespread hydrothermal alteration and VMS deposits indicate that these rocks have economic potential. Thus, the chemical characteristics of these rocks appear to provide a better guide to their mineral potential than their age.

5.7.5.1 Meat Lake amphibolite (unit 4c)

Where geochemical data are available, two mafic varieties with similar trace-element patterns are distinguished: Meat Lake amphibolite, which has a moderate content of MgO and is of predominantly volcanic origin; and Walton Lake amphibolite, which is generally Mg rich and contains both volcanic and intrusive components. Both units have a highly variable state of preservation from local exposures.

Meat Lake amphibolite locally retains pillow rinds or breccia structure (Figure 9f), but in other places resembles a layered mafic tectonite with long thin lenses of plagioclase-hornblende amphibolite separated by garnetiferous amphibole-rich ribbons derived from pillow rinds. Many amphibolite lenses contain areas rich in plagioclase and diopside that are interpreted to have been derived from epidosite seafloor-alteration domains.

Despite alteration of alkali major elements and minor LILE, the Meat Lake amphibolite is considered to have retained important aspects of its geochemistry to allow its recognition as arc tholeiite. It is chemically evolved, with low contents of MgO (4.4–6.7%) and Ni (<30 ppm), and relatively low Mg# (43–56). This suggests that the incompatible elements should be elevated, but this is not the case, with REE ≈ 10× chondrite and less than N-MORB (Appendix 3, Figures 33c, 35a). Incompatible element contents are low except for Th and some LREE. The amphibolite is therefore considered to have been derived by highly fluid-fluxed melting of depleted mantle in a mature-arc environment that had previous melt extraction.

Neodymium isotope ratios, which are resistant to the high metamorphic grade as in the sampling area of Meat Lake amphibolite, yield an ε_{Nd} value of 2.4, similar to the least contaminated Puffy Mine andesite and other early arc-volcanic rocks (Appendix 2, Figure 29f).

5.7.5.2 Walton Lake amphibolite (unit 4d)

Walton Lake amphibolite (unit 4d) occurs in narrow belts within Sherridon gneiss south and southeast of Walton Lake, and a sliver at Folster Bay on File Lake is tentatively included. The unit preserves rare remnants of pillow structure from lavas but is mainly medium-grained amphibolite without flow contacts, possibly derived from intrusive rock. Walton Lake amphibolite may also include cumulate gabbro phases, but no local differentiation is apparent in it or the coarser grained bodies of recognizable gabbro.

The Walton Lake amphibolite is the most highly recrystallized mafic unit sampled in the Kississing–File lakes area. The amphibolite occurs where greywacke–pelite of the Burntwood Group has entered the zone of migmatization and changed bulk composition. The internal consistency and similarity of the trace-element patterns of Walton and Meat Lake basalt units to those of lower grade units like Threehouse basalt at Snow Lake (Bailes and Galley, 1999, 2001, 2007), however, attest to the remarkable preservation of the immobile trace elements in mafic rocks, even at high metamorphic grade. The unit has a larger range of MgO contents (5.9–16.9%) than the Meat Lake amphibolite, indicating fractionation and possible presence of cumulates. Walton Lake amphibolite is further distinguished from the Meat Lake by its slightly lower ratio of La/Yb and higher Yb7 (Appendix 2, Figure 24a). The trace-element data suggest that Walton Lake amphibolite is an arc-volcanic rock, with one sample (Folster Bay) having anomalously high Nb, probably due to alteration (Appendix 2, Figure 31c). The normalized patterns of REE and extended elements are similar to those produced by the Meat Lake amphibolite and gabbro in the Walton Lake nappe (compare plots in Appendix 3, Figures 33c–e and 35a, b). Walton Lake amphibolite has a geochemical pattern that most closely matches that of the Threehouse basalt and gabbro, part of the Snow Lake assemblage that is considered to have formed in a mature-arc environment (Appendix 3, Figure 35b; Bailes and Galley, 2001).

The trace-element similarity of the Walton and Meat Lake amphibolite units with low-Ti amphibolite from the Sherridon structure supports the correlation of the gneissic volcanic succession in the Walton Lake nappe with the Sherridon structure. The amphibolite in the vicinity of Sherridon is clearly tholeiitic, and fractionation of LREE is even less pronounced (Appendix 1, Figure 21a; Appendix 3, Figure 35a, b). The rocks are less evolved than the Meat Lake amphibolite, more like the lower part of the volcanic succession at Snow Lake, and are considered to represent a separate volcanic centre from rocks in the Walton Lake nappe.

5.7.5.3 Walton Lake gabbro (unit 4e)

Where local gabbroic texture is preserved or contacts are crosscutting or margins are chilled, several coarse-grained intrusive samples (Walton Lake gabbro) provided a geochemical signature similar to the Walton Lake and Meat Lake amphibolites. Trace elements of two fully analyzed samples suggest a subvolcanic origin related to Walton Lake amphibolite flows (e.g., Appendix 2, Figure 31c; Appendix 3, Figure 35b).

5.7.5.4 Sherridon rhyolite, derived gneiss and alteration (unit 5, unit 5a)

The thick succession of felsic gneiss in the Walton Lake nappe and Sherridon structure (Sherridon gneiss) is interpreted to have a generally volcanic origin and to form a bimodal suite with the Meat Lake, Walton Lake and equivalent low-Ti amphibolite units (column 5 in Figure 8e, back pocket). The best preserved samples in the Walton Lake nappe (unit 5a) have high SiO_2 (average 77%, range 72–81%) and low Al_2O_3 (average 12%). The more gneissic rocks (unit 5), excluding the most altered samples (units 5b–d), in the Walton Lake nappe average 75% SiO_2 (range 73–76%). Both types fall into the field of rhyolite on alkali-silica plots (Appendix 1, Figure 20a). On an AFM diagram, they form a calcalkaline trend (Appendix 1, Figure 21a) or, using immobile elements, a calcalkaline to transitional trend (Appendix 2, Figure 27b). The recently more widely sampled Sherridon gneiss in the Sherridon structure (unit 5) averages 75% SiO_2 (range 65–84%) for a selection of 49 surface samples from Halo Resources Ltd. (L. Bloom, pers. comm., 2007) that are least likely to be strongly altered. This suite apparently includes rhyolite and dacite.

Some of the rocks have undergone SiO_2 metasomatism, whereas others show evidence of alkali mobility and some hydrothermal alteration, resulting in elevated Fe or Mg. Altered samples are indicated by outliers with exceptionally high SiO_2 and variable contents of K_2O and other fluid-mobile elements (Appendix 1, Figure 23f), or elevated mafic elements with loss of alkalis. Large parts of the Sherridon gneiss have been recrystallized to weakly banded felsic gneiss with complete loss of primary structure and development of garnet±amphibole or epidote±calcite. These rocks have low K_2O contents and commonly an alumina index (A/CNK) of >1.2 or <0.8. This suggests alkali loss or gain during metamorphism and incipient partial melting. Alteration is also evident from the molar alkali/Al plot, in which more than half the strongly recrystallized samples show a small degree of alteration (Appendix 2, Figure 25d). The K_2O content of the older, lower grade rhyolite in the Snow Lake area is even more alkali depleted than that of the Sherridon gneiss (e.g., K_2O content of Photo rhyolite averages 1% and can be as low as 0.2%; Bailes and Galley, 2001). This suggests that alkali elements were mobile during early alteration as well as during metamorphism. The plot of fluid-immobile elements (La versus Zr) shows a wide scatter that suggests fractionation of accessory minerals or mobility during incipient partial melting (Appendix 2, Figure 26h); however, the alteration did not entirely conceal the fingerprint of subduction-related fluid-fluxed melting as opposed to slab melting (Appendix 2, Figure 28a, e).

Trace-element ratios are better preserved than the alkali elements and SiO_2 in the Sherridon gneiss. Ratios of REE from the Walton Lake nappe fall into the low-Yb, intermediate $[La/Yb]_N$ range (Appendix 2, Figure 24a), similar to type II rhyolite of Lesher et al. (1986) and all of the 'mature-arc' rhyolite in the Snow Lake area. Ratios of Th/Yb plot near the Central Flin Flon belt arc rhyolite units in the subdomain but with slightly higher Th and Nb (Appendix 2, Figure 29c). On the Zr versus TiO_2 plot, they fall into the field of arc rhyolite (Appendix 2, Figure 29d). A chondrite-normalized plot of REE from the Walton Lake nappe is nearly identical to that of the most common rhyolite near the top of the Snow Lake assemblage (Appendix 3, Figure 33f; 17 of 20 analyzed samples of Photo rhyolite from Bailes and Galley, 2001). Because this rhyolite unit northwest of Snow Lake is associated with the Threehouse basalt and gabbro, which have chemistry similar to that of the Walton Lake amphibolite (which, in turn, is interlayered with Sherridon gneiss), the two bimodal suites are interpreted to have formed from a similar mantle source in a similar tectonic environment and with a similar high VMS potential, although apparently at different times.

Typical felsic gneiss from the Sherridon structure (unit 5) also follows the pattern of the samples from the Walton Lake nappe but shows less fractionated LREE (Appendix 2, Figure 24a; Appendix 3, Figure 33f). Only about half of the selected group of surface samples from the dataset provided by Halo Resources Ltd. (L. Bloom, pers. comm., 2007) with high SiO_2, like the samples from the Walton Lake nappe, have $[La/Yb]_N$ >2. Apparently much of the rhyolite gneiss in the Sherridon structure is chemically and isotopically less evolved, with lower La/Yb and higher Yb, Ti and ε_{Nd} (+3.4 to +4.2), than that in the Walton Lake nappe. A sample of rhyolite gneiss taken southeast of Meat Lake has an ε_{Nd} value of +1.4 (Table 15; Appendix 2, Figure 29f). This value is considered to be typical of the Sherridon gneiss from the Walton Lake nappe because it has Th and U contents similar to the average, and these elements are correlated with ε_{Nd}.

The geochemical differences can have various explanations. Rocks in the Walton Lake nappe were carefully selected on clean surface outcrops to avoid visible alteration and thus may have higher contents of felsic minerals with associated accessory minerals (containing LREE). The rocks also may have gained LREE-enriched leucosome. There is similar fractionation of LREE for the associated mafic rocks (units 4–4d; Appendix 3, Figure 35a), which are not veined or enriched in leucosome. The difference in ε_{Nd} in the felsic rocks, however, suggests a primary difference in volcanic composition. The larger part of the volcanic products from the centre near Sherridon was probably less evolved than the rocks in the Walton Lake nappe. The overlap in trace-element geochemistry but with different ranges in the two areas suggests a similar origin but from different volcanoes. It is of interest that the rocks with the higher ratio of La/Yb but associated with the more depleted rhyolite gneiss appear to form the most altered and mineralized packages.

5.7.5.5 Calcsilicate rock (unit 5b)

Calcsilicate rock was analyzed from unit 5b (interlayered felsic gneiss and calcsilicate gneiss) on the southern margin of the Sherridon structure. The sample has trace-element characteristics of the Puffy Mine andesite (Appendix 2, Figure 24a) but is too altered and insufficiently sampled for a meaningful correlation. Two samples of mafic calcsilicate gneiss (unit 17c) from the large body near the core of the structure have a geochemical affinity with the intrusive rocks of unit 17 (Section 5.10.2) but may be an alteration product of much older rocks.

5.7.6 Josland Lake sills (unit 7)

The 1886 Ma age of the Josland Lake sills and their widespread intrusion into the Fourmile Island, Northeast Reed and Snow Lake assemblages make these intrusions one of the most important units in the area because they provide a minimum age for their volcanic hosts. The proposed correlation of the various sills is based, in part, on their similar chemical composition and trace-element ratios. An origin from a single parent magma has been suggested (Zwanzig et al., 2001). If the various Josland-type sills are, in fact, closely related and nearly coeval, a fundamental rethinking of the evolution of the Flin Flon Domain and Kisseynew Domain may be required. In order to test the wide correlation, the geochemistry of distinct zones occurring in three major sills was compared in detail over the 100 km extent of the intrusions.

The best preserved Josland Lake sills extend from an area of middle-greenschist facies west of Reed Lake to middle-amphibolite facies at File Lake. These layered differentiated bodies average 500–700 m thick (Bailes, 1980a). They have a thick lower zone of mela- to leucogabbro (unit 7a), a central zone of ferrogabbro (unit 7b) and an upper zone of granophyric quartz diorite to trondhjemite (unit 7c). In the large, extensively sampled sill dated at 1886 Ma near Josland Lake, these zones are, respectively, 350, 250 and 150 m thick (column 9B in Figure 8d, back pocket). The gabbro zones have sharp lower contacts and chilled margins. Weak rhythmic modal layering occurs, particularly near the top of the unit, and cumulate textures are locally preserved (Bailes, 1980a). Original minerals were augite, labradorite and ilmenite, but were preserved only as pseudomorphs replaced by metamorphic minerals.

The upper-amphibolite–facies equivalents of the sills in the northern part of the Flin Flon Domain have the same zoning as sills in the southeast but range in thickness from 150 m to only a few metres (Zwanzig et al., 2001). Sills in the high-grade terrain locally show rhythmic layering (Figure 10b) and their lowermost division is melagabbro that contains only minor amounts of plagioclase. The gabbroic main phase is recrystallized to medium-grained amphibolite with platy and fibrous amphibole, diopside and lesser plagioclase. The ferrogabbro zones weather nearly black at all metamorphic grades and contain mainly secondary amphibole. Their base is gradational over a few metres and shows strong rhythmic layering in the low-grade areas. The ferrogabbro is mainly uniform at high grade and contains scattered garnet porphyroblasts with local centimetre-scale garnet-magnetite layering at the top. The top horizon was previously mapped as iron formation (Zwanzig and Schledewitz, 1992) before fire removed the lichen cover and allowed closer inspection.

The upper zones grade over a short distance from ferrogabbro to quartz diorite (<40% mafic minerals), capped with trondhjemite or felsic gneiss rich in albite (<5% mafic minerals). Contacts are also gradational; some upper contacts show signs of roof-zone contamination, but thinner sills have chilled upper margins. Granophyric intergrowth between sodic plagioclase and quartz is preserved only in the lower grade rocks. Radiating clusters of metamorphic fibrous amphibole in the lower grade rocks (Bailes, 1980a) occur as a mottled gneissic texture in the north (Figure 10c). Pale pink–weathering albite is abundant in the mottled felsic gneiss.

The Josland Lake sills show extreme Fe enrichment in the ferrogabbro, quartz diorite and Fe-rich trondhjemite zones, similar to some back-arc–basin flows and gabbro (Appendix 1, Figure 21b). The TiO_2 versus MgO of the lower chill zone at Josland Lake and of a fine-grained amphibolite at the top of the gabbro at Evans Lake plot in the MORB field (Appendix 1, Figure 22c). The TiO_2 is fractionated into the ferrogabbro and depleted in the underlying gabbro units. Fractionation from a single parent magma is suggested by the limited range of $[La/Yb]_N$ ratios (0.7–1.4) over a wide range of REE contents (e.g., 0.6–10 ppm Yb in Appendix 2, Figure 24b). The chilled gabbro falls into the field of E-MORB on Th/Yb versus Nb/Yb and Th/Nb/Zr diagrams (Appendix 2, Figures 29a, 30d) but closer to N-MORB on a La/Y/Nb diagram (Appendix 2, Figure 31d). The trondhjemite and quartz diorite at the top of the sills plot in the field of ocean-ridge granite (ORG), based on the elevated Y and Nb contents (Appendix 2, Figure 25e).

The REE contents of the Josland Lake sills show a series of smooth parallel profiles, with the chill zone being 10× chondrite, the same as a flow at the top of the Ponton Lake basalt west of Ponton Lake (Appendix 3, Figure 35f). The wide range of total REE is the result of the strong fractionation. The positive Eu anomaly is caused by plagioclase accumulation in the lower part of the sills, where leucogabbro is present. The matching negative Eu anomaly of the ferrogabbro and intermediate–felsic differentiates represents plagioclase loss (sinking); the chilled rock has no Eu anomaly. The modal layering, common below the ferrogabbro, shows the accumulation of plagioclase, whereas garnet and magnetite banding in highly metamorphosed ferrogabbro shows Ti-magnetite±ilmenite accumulation higher in the sills. Elevated Cr occurs in melagabbro (~14–16% MgO) at the base of sills west of Morton Lake and north of Puffy Lake. The highest contents of Cr are 1370–1710 ppm (compared to 325 ppm for the basal chill zone and 105 ppm for a fine-grained top).

Mid-ocean-ridge basalt (MORB)–normalized extended-element plots of all units of the Josland Lake sills have a negative slope, typical of E-MORB, and elevated Th, typical of contaminated MORB (Appendix 3, Figure 35d–f). This is consistent with the composition of the chill zone and the indication of melting in the roof zone. There are no negative Nb, Zr and Hf anomalies that would indicate an origin as normal arc tholeiite. Chilled Josland Lake gabbro and the Snow Creek and Tramping Lake arc-rift basalt units, and possibly SE Dow porphyritic basalt, have a nearly identical multi-element pattern (Appendix 3, Figure 35d; Bailes and Galley, 2001). The plots suggest that the parent magma of the gabbro was similar to back-arc–basin and arc-rift basalt at the top of the Amisk Collage, and arc-rift basalt at the top of the Snow Lake assemblage. All of these were derived from an enriched source very similar to primitive mantle but with fractionation affecting the total content of the least compatible elements and disturbing the pattern between Zr and Ti (Appendix 3, Figure 34d–f).

Titanium was clearly involved in the differentiation process, probably in ilmenite and/or Ti-magnetite crystallization. This produced Ti depletion in the trondhjemite and accumulation in the ferrogabbro, thus showing large complementary Ti excursions in Appendix 3, Figure 35d–e. These element patterns are very similar for series of samples across sills from three widely separated areas. They support the contention that there was a large area in a similar environment producing Josland Lake–type magma. However, the calculated mean composition (Bailes, 1980a) of the chilled margin phase of the largest sill (i.e., west of Morton Lake) has a low Mg# and high FeO content that indicate previous fractionation in a deeper magma-plumbing system. The various sills consequently have different proportions of the three common zones, depending on the degree of fractionation of the local batch of magma. For example, the lower sill at Nokomis Lake comprises mainly ferrogabbro and has an upper zone of trondhjemite, whereas the upper sill comprises mainly normal gabbro and has a thin lower zone of melagabbro.

A sample of chilled gabbro and one of trondhjemite have ε_{Nd} values of 1.5 and 4.1, respectively. Taken at face value, this suggests that the trondhjemite was fractionated from juvenile parent magma but that the contact phase was weakly contaminated by country rock (possibly the Yakymiw sediments) that had an Archean clastic component. This is supported by a slightly higher K_2O content of the chill zone, but the samples are from different sills and more data are required to prove such an explanation. The thick sill probably formed from multiple pulses of magma and only the early pulse(s) were contaminated, and not necessarily at the site of intrusion, which may have been juvenile.

5.7.7 Early- to successor-arc volcanism: discussion

The tectonostratigraphy of the early volcanic rocks, based partly on geochemistry, supports the conclusion that the Flin Flon Domain divides into two major composite terranes: 1) the Central Flin Flon subdomain (underlain by the Amisk Collage), and 2) the Snow Lake subdomain. Isotopic dating shows that both of these tectonic complexes were formed by the amalgamation of ca. 1.90–1.88 Ga arc, back-arc and arc-rift volcanic assemblages. The intrusion in both subdomains of the ca. 1886–1881 Ma Josland Lake and identical Mikanagan sills, with arc-rift geochemistry, suggests that there may have been early arc–back-arc amalgamation followed by rifting.

Early successor-arc magmatism also occurred in both subdomains, but intrusions are generally restricted to the Central Flin Flon subdomain; volcanic rocks occur mainly in the eastern part of the Snow Lake subdomain and in the north, where they define the separate Batty Lake subdomain.

The Kississing–File lakes area includes the northeastern margin of the Amisk Collage, the southwestern part of the Snow Lake subdomain and the intervening Morton Lake–Loonhead Lake structural break (MOF and LHF in Figure 5). This break is marked by fault slices of the 1.85–1.84 Ga collision-related sedimentary rocks of the Burntwood Group. Assemblages are traced for up to 100 km along this break, showing it to be a crustal-scale feature. Units on either side of the break are recognized by their distinctive geochemical patterns despite the high to moderate grade of metamorphism and widespread alteration of LILE. Identifying and understanding these units is critical in reconstructing the early volcanic and metallogenic evolution of the Flin Flon Domain.

A major new contribution of this report is in establishing the likelihood that the entire Flin Flon Domain was accreted as a single arc–back-arc complex (Flin Flon–Glennie Complex) early in its history, but was subject to widespread arc rifting associated with the early development of the Kisseynew basin and the 1886–1881 Ma intrusion of the Josland Lake and Mikanagan sills and extrusion of their volcanic equivalents. The Morton Lake–Loonhead Lake structural break, with its early rift margin, BABB and younger (syncollisional) sedimentary rocks, is one of the more prominent post-Burntwood sutures along which the Flin Flon–Glennie Complex was reassembled during collision tectonics.

At a low structural level, the northeastern margin of the Amisk Collage is underlain by juvenile tholeiite with a strong geochemical arc signature. This includes the Puffy Mine and Koscielny andesite units, whose origin in a volcanic arc is well documented (e.g., Appendix 3, Figure 34f) despite signs of chemical alteration and local isotopic mixing by veining. These units are remnants of the relatively old volcanic edifices that developed on both sides of the same arc terrane and presently occur on opposite limbs of the Ponton Lake synform. Other small remnants of arc tholeiite (Martell Lake basalt and Dow Lake andesite) occur farther east, at Dow Lake. The early arc units are structurally or stratigraphically overlain by the Ponton Lake basalt and SE Dow basalt, units with weak tholeiite arc signatures and fractionated REE patterns. The Ponton Lake basalt is an extensive unit with some very MgO-rich flows and abundant sills, derived from a slightly enriched mantle and herein interpreted as having formed at the margin of a back-arc basin, possibly during recurring rifting. Geochemical plots of this unit on tectonic element-ratio diagrams straddle the fields of E-MORB and arc (e.g., Appendix 2, Figure 30a) or BABB, contaminated with crust (i.e., in the continental field; Appendix 2, Figure 31a).

Moderately enriched basalt similar to the Ponton Lake unit is a common feature in modern back-arc basins (e.g., Fretzdorff et al., 2002). The Ponton Lake basalt probably formed part of the margin of such a basin. It may be part of the 1.90 Ga Elbow-Athapapuskow ocean-floor assemblage, which has a similar E-MORB affinity (Stern et al., 1995a) and extends southwest for 150 km across the Flin Flon Domain into the Namew Gneiss Complex. Alternatively, Ponton Lake basalt may be genetically related to the 1886 Ma Josland Lake sills, which are geochemically very similar to a flow in the upper part of the unit. Another shared feature of these units is the moderately low ε_{Nd} of Ponton Lake basalt and chilled Josland Lake gabbro, which indicates slight contamination by Archean crust. If the back-arc basin was long lived (≥10 m.y.), these alternatives are not mutually exclusive, however, because the Elbow-Athapapuskow ocean-floor assemblage, Ponton Lake basalt, Josland Lake sills and geochemically similar basalt units all formed from similar, weakly enriched or 'primitive' mantle that produced normal, Fe-rich and high-Mg basalts. This includes the SE Dow porphyritic basalt, which is separated from the eastern exposures of the Ponton Lake basalt only by intrusions and the WRNSZ, and probably had a similar mantle source.

The outermost eastern margin of the Amisk Collage consists of the Fourmile Island assemblage (Syme et al., 1995). This assemblage has a geochemical arc signature at its type locality on Reed Lake (Syme and Bailes, 1996) but a progressively weaker arc signature, similar to arc-rift basalt, in its northern to northwestern extension along the full length of the Kississing–File lakes area. The transition zone between rocks with arc and arc-rift geochemistry is west of File Lake, where the Fourmile Island assemblage comprises Preaston andesite, Dickstone rhyolite and Storozuk volcanics. These genetically related units are characterized by positively sloped to flat MORB-normalized extended-element plots with variable HFSE troughs (Appendix 3, Figure 34c–e). The upward transition to the more mafic and more MORB-like Storozuk basalt suggests a change from fluid-fluxed melting to decompression melting during the inception of a back-arc basin. The ocean-ridge granite (ORG)–like Dickstone rhyolite was fractionated from a Preaston-like mafic parent, consistent with extension-related volcanism, for example, at the tip of an arc-rift basin. The Dickstone VMS deposit occurs at this favourable site, similar to other VMS deposits in the Flin Flon Domain (Syme et al., 1999). The Yakymiw volcaniclastic unit of Bailes (1980a) is interpreted as the rift-basin fill.

The Moody Lake basalt is interpreted as a 70 km long western extension of the Fourmile Island assemblage. The unit has low Nb but depleted LREE, commonly less than MORB but similar to units in the Fourmile Island assemblage to which it is related (Appendix 3, Figure 34c–e). Very similar basalt to that at Moody Lake is the Dismal Lake suite of Gilbert (2004), which occurs south of Kisseynew Lake at the northern margin of the Amisk Collage. Like the Moody Lake basalt, this suite is in direct contact with arc tholeiite and locally intercalated with higher La/Yb basalt similar to the SE Dow porphyritic basalt (which is interlayered with the Moody Lake basalt). These nearly identical tectonic and chemostratigraphic sequences are clearly part of a single package. The Moody Lake basalt is in direct (structural) contact with the Burntwood Group. These units are therefore interpreted to trace out the rifted edge of a basin on the margin of the Kisseynew Domain. Similar modern basalt occurs in the Manus back-arc basin in the south Pacific, where it is considered to be MORB (Sinton et al., 2003).

Early volcanic assemblages that occur north and east of the Morton Lake–Loonhead Lake structural break show some geochemical characteristics that are subtly different from those of the adjacent Amisk Collage. The Fussey Lake basalt (part of the Northeast Reed assemblage of Syme et al., 1995) is very similar to N-MORB and is interpreted in this report as BABB. The MORB-normalized plots are flat but have small negative HFSE anomalies (Appendix 3, Figure 34a), like Manus back-arc–basin basalt (BABB) in Sinton et al. (2003). It should be noted that basalt with a similar composition in the Phanerozoic fore arc of central California has also been interpreted to be associated with fore-arc rifting, but that unit passes stratigraphically upward into arc basalt (Snow, 2007).

The Josland Lake and Mikanagan sills form distinctive, highly fractionated bodies that are chemically similar to some modern back-arc basin flows and gabbro (Appendix 1, Figure 21b). The various sills are interpreted to have formed from slightly enriched high-Mg parent magmas, all from the same mantle source. The chilled gabbro is similar to arc-rift basalt at the top of the Amisk Collage (uppermost Ponton Lake basalt), and at the top of the Snow Lake assemblage (Snow Creek and Tramping Lake basalts). The Fe-rich chilled Josland Lake gabbro, with a high Mg# (65), must have had very similar parent magma to the upper flow in the Ponton Lake basalt or the main part of the Elbow-Athapapuskow assemblage (Stern et al., 1995a). Mantle upwelling and pressure release during arc rifting and back-arc–basin extension lead to high degrees of melting (e.g., Carn and Pyle, 2001) with high Mg, Ni and Cr, as in such parent magma.

The Josland Lake sills intrude the units in the Amisk Collage, Northeast Reed assemblage and Snow Lake assemblage. They are particularly abundant in the Fourmile Island assemblage, parallel to the Morton Lake–Loonhead Lake break, at the margin of the Amisk Collage where earlier rifting had already occurred. If the sills resulted from a single protracted event, as suggested by their similar geochemistry and broadly overlapping ages (1886 ±3 and 1881 +3/–2 Ma), this supports the possibility that the Central Flin Flon and Snow Lake subdomains formed as a single accretionary complex that was fragmented during progressive arc rifting. The intrusion of the E-MORB–like Josland Lake sills into the older N-MORB– or BABB-like Fussey Lake basalt requires repeated extensional oceanic tectonics. The attendant high heat flow and development of structural breaks may have led to the exceptionally high mineral potential of the Flin Flon Domain.

Despite its preliminary age, which places it as part of the early successor arc, the Sherridon–Meat Lake assemblage has geochemical characteristics similar to those of the older oceanic volcanic rocks, particularly those in the Snow Lake assemblage. The rocks at Sherridon and Meat Lake are strongly bimodal, predominantly felsic. The amphibolite units are depleted in HFSE and MREE to HREE, whereas the felsic volcanic–derived gneiss has Lesher type II geochemistry. The geochemical characteristics, shared with the upper part of the Snow Lake arc assemblage, also include fractionated LREE and elevated Th (Appendix 3, Figure 35a–b). In modern environments, mafic units with a similar strong arc signature, depleted in mantle components but enriched in fluid components as are the Walton and Meat Lake amphibolites, occur in arcs that have a prominent back-arc basin. Neodymium-isotope data suggest a juvenile origin for the rocks at Sherridon or, in the case of the Walton Lake nappe, mild contamination by Archean granitic crust or sedimentary components.

The rocks in the Sherridon structure are more primitive than those at Meat Lake, with higher ε_{Nd} and less fractionation of LREE. These rocks are locally associated with the more evolved rhyolitic compositions, particularly where hydrothermal alteration is prominent. In the origin of modern arc basalts with depleted HFSE and enriched Th, the upper mantle has been depleted before it is drawn into the mantle wedge below the arc, where it is reprocessed (Gamble et al., 1993; Woodhead et al., 1993). Considering the young preliminary age of the Sherridon–Meat Lake assemblage and the apparent absence of young BABB, however, depletion of the mantle source may have occurred during the earlier magmatism. Regardless of age, the most important geochemical characteristic of the amphibolite units is their affinity to an oceanic arc—one prerequisite for the high VMS potential of the associated felsic volcanic rocks.

5.8 Early successor-arc intrusions

Early successor-arc intrusions in the Kississing–File lakes area constitute the greatest proportion of the igneous rocks on the northern margin of the Flin Flon Domain. Their chemical characteristics are examined here only to provide a cursory comparison with the less recrystallized plutons in the south (Whalen et al., 1999). Compared to the earliest juvenile-arc intrusions, the early successor-arc plutons are much more felsic (Appendix 1, Figure 19) and have higher Al_2O_3 contents (65–75% SiO_2; compare Appendix 1, Figure 22e and f). They have higher La/Yb ratios (compare Appendix 2, Figure 24a and b) and higher Rb contents (Appendix 2, Figure 28b). However, like the early volcanic rocks, they include tholeiitic and calcalkaline units (Appendix 1, Figure 21c). Individual plutons and plutonic phases were distinguished by their petrographic characteristics (Zwanzig and Schledewitz, 1992) but also show distinct fields on plots of immobile trace elements and trace-element ratios (e.g., Appendix 2, Figure 24b).

5.8.1 Mafic intrusions

Gabbro and diorite represent the relatively small volume of mafic magmatism preserved with the early successor-arc plutons (Appendix 1, Figure 19c). They have low K_2O contents, consistent with an oceanic- or island-arc environment (Appendix 1, Figure 20f).

5.8.1.1 Martell Lake gabbro, microdiorite (unit 8a)

Several large sheets of gabbro south of Martell Lake have been mapped as early mafic intrusions because one of these bodies is truncated at the unconformity at the base of the Missi Group. They intrude the lower part of the Amisk Collage, which is generally more strongly foliated than the gabbro and had apparently been deformed during tectonic accretion before emplacement of the gabbro intrusion. Two analyzed samples of Martell Lake gabbro have a similar composition, with 7–8% MgO and moderate Fe enrichment, suggesting that the sills are tholeiitic (Appendix 1, Figure 21c). Contents of LILE are low (e.g., Appendix 2, Figure 28b, f) and high Th/Zr ratios (Appendix 2, Figure 30b). Although slightly elevated in Nb/La, the sills are clearly distinguished from the (older) Josland Lake gabbro (Appendix 2, Figure 31d). Fractioned REE and negative HFSE anomalies in the normalized plots (Appendix 3, Figure 36a, b) indicate that this was volcanic-arc magma.

Fine-grained intermediate dikes (microdiorite) that cut the volcanic rocks of the Amisk Collage are also relatively Fe rich (Appendix 1, Figure 21c), but some have calcalkaline trace-element ratios (Appendix 2, Figure 27d). Their age relative to the late successor-arc magmatism is uncertain. They have a lower La/Yb ratio than intermediate volcanic rocks of most of the late intrusions and the Missi Group (Appendix 2, Figure 24). They have a MORB-normalized extended-element pattern similar to that of Martell Lake gabbro (Appendix 3, Figure 36b), suggesting that the microdiorite belongs to the early intrusive suite.

5.8.2 Granitoid plutons

Seven granitoid plutons or intrusive phases from which geochemical data were collected have U-Pb ages or field relations indicative of early successor-arc magmatism. These plutons occur on the northern margin of the Flin Flon Domain and have developed moderate to strong foliation, some with gneissic layering and incipient remelting. They provide a comparison to the less strongly metamorphosed intrusions that have been interpreted to form belts extending across the Flin Flon Domain (Zwanzig, 1999). Although metaplutonic rocks of granitoid composition are commonly considered to have suffered less alteration than felsic volcanic rocks because of their coarser grain size (e.g., Whalen et al., 1999), little is known of chemical mobility in high-grade meta-igneous rocks, and our data suggest that the gneissic plutonic rocks have experienced significant alkali-element alteration. This is an impediment to assigning petrogenetic and tectonic origins to these rocks. As with the felsic volcanic gneiss, however, some of the general characteristics can still be discerned, particularly the preserved coherent REE patterns, interpreted to represent the igneous composition. The data indicate that most plutons follow a calcalkaline trend and are volcanic-arc intrusions similar to those belonging to the southern part of the Flin Flon Domain.

The early granitoid rocks are calcalkaline, predominantly medium K with a wide range of SiO_2 (Appendix 1, Figure 20f, 21c). They have a range of Al_2O_3 contents and an Al-saturation index scattered around 1.0 (Appendix 1, Figure 22f, 23j), typical of volcanic-arc plutons. They generally show strong negative anomalies of Nb and Ti, typical of such plutons (Appendix 3, Figure 37). They belong mainly to a low-La group with $[La/Yb]_N$ <12, except the Archie Lake and Star Lake plutons, which have $[La/Yb]_N$ ratios of 30–33 and 25–28, respectively, at low Yb (0.57–0.67 and 0.50 ppm, Appendix 2, Figure 24b). These two plutons also have elevated Sr/Y (Appendix 2, Figure 28b, f) suggesting a slab-melt component like other plutons in the Flin Flon Domain (Whalen et al., 1999). Values for ε_{Nd} of +1.1 and –2.1 indicate that at least some of the plutons have an old crustal-melt component.

5.8.2.1 Ragged Lake (granite) pluton (unit 9a) and Big Island (granodiorite) dome (unit 9c)

These plutons, which are unconformably overlain by the Missi Group, show the calcalkaline trend of decreasing mafic content with increasing alkali content (Appendix 1, Figure 21c). They have variable alumina saturation (Appendix 1, Figure 23j) and locally contain garnet, but the few scattered weakly peraluminous samples are considered to be altered by alkali loss during metamorphic reactions. Some display patterns that are also typical of felsic calcalkaline plutons in the main Central Flin Flon subdomain, with slightly declined to flat slopes of HREE, fractionated LREE and negative Eu±Sr anomalies, (Appendix 3, Figure 36c, 37a, c).

The 1874 Ma Ragged Lake pluton has the highest contents of K_2O (4.9–5.1%), other LILE and Th of the early successor-arc intrusions sampled in the Kississing–File lakes area (Appendix 4, Table 10, DVD). This body also has the negative ε_{Nd} (–2.1, Table 15) indicative of contamination by Archean crust during early successor-arc magmatism. The negative Eu and Sr anomalies are consistent with crustal melting retaining plagioclase in the residue (Defant et al., 1991).

5.8.2.2 Jay Lake (leucogranite) pluton (unit 9b)

The leucocratic monzogranite that occurs at Jay Lake and nearby in the southwestern part of the Sherridon–Hutchinson Lake Complex (Zwanzig and Schledewitz, 1992) has not yielded zircon for isotopic dating but is inferred to be an early intrusion because of its occurrence with other strongly foliated, early plutons in the complex. The granite has high REE contents and low La/Yb ratios (Appendix 2, Figure 24b), as well as negative Eu anomalies (Appendix 3, Figure 36d), all characteristics of highly fractionated granite or crustal melt that leaves plagioclase in the residuum (Fornelli et al., 2002). Their position in the tholeiitic field in Th/Yb versus Zr/Y space is probably due to fractionation of accessory minerals. Whereas most of the early successor-arc granitoid rocks fall into the 'volcanic-arc granite' (VAG) field on a plot of Rb versus Y+Nb, the Jay Lake granite falls into the 'within-plate granite' (WPG) field (Appendix 2, Figure 25f) and may have formed in a mildly extensional environment after the amalgamation of the Amisk Collage hostrock.

5.8.2.3 Fire Lake gneiss (unit 9c)

A minor unit of strongly foliated, fine-grained granodiorite (Fire Lake gneiss, sample locations 217 and 221, Map GR2010-1-2, DVD) occurs adjacent to the larger, early successor-arc plutons in the Puffy Lake area and south of Martell Lake at the margin of the Gants Lake Batholith. Only major-element analyses are available for this unit and these follow the relatively Fe-rich trend for felsic rocks (Appendix 1, Figure 21c) of the Batty Lake Complex (Section 5.8.2.6).

5.8.2.4 Fire Lake granodiorite and Archie Lake (granodiorite–tonalite) pluton (unit 9d)

The 1860 ±2 Ma Archie Lake tonalite pluton (Machado et al., 1999) belongs to the high Al_2O_3, high La/Yb, low Yb suite (Appendix 2, Figure 24b). The positive Sr anomaly in Archie Lake tonalite on a primitive-mantle–normalized plot is weak compared to modern adakite (Appendix 3, Figure 37d). The slightly elevated K_2O/Na_2O ratio (0.62) and moderate Sr/Y ratio (59) suggest that Archie Lake tonalite may be derived from a calcalkaline melt mixed with adakite-like melt, a common feature in some volcanic arcs (e.g., Samaniego et al., 2002). These ratios and most element compositions are nearly identical to the median values for Proterozoic tonalite-trondhjemite-granodiorite (TTG) plutons, as selected by Condie (2008); see Table 16. The concave-upward REE pattern, showing fractionation of MREE (Appendix 3, Figure 36e), was probably caused by hornblende (Defant et al., 1991), being in the lower crust or slab. Such origins are, in general, chemically indistinguishable (Xu et al., 2002; Richards and Kerrich, 2007). Moreover, unlike typical adakite but like some TTG, Archie Lake tonalite has an ε_{Nd} of 1.1, lower than MORB but weakly contaminated by older crust. Generation of TTG may involve melting of a thickened slab (Condie, 2008) or, more likely, amphibolite-dominated lower crust in an arc environment (Whalen et al., 2002). The older crustal contamination supports the possibility that melting occurred in predominantly mafic crust stacked with local, older felsic crust, similar to relations that occur in shear zones in the Central Flin Flon subdomain.

Table 16: Average contents and ratios of selected elements for various granitoid units in the Kississing–File lakes area.

Unit	SiO_2	Al_2O_3	TiO_2	MgO	K_2O/Na_2O	A/CNK	Ni	Cr	La/Yb_n	Sr/Y	Nb/Ta
Star Lake tonalite gneiss[1]	69.4	16.3	0.34	1.0	0.34	1.00			26	124	12
Archie Lake pluton[1]	69.9	15.3	0.39	1.3	0.62	1.02		19	31	59	14
Touchbourne suite (two most felsic)[1]	62.8	14.7	0.90	3.2	0.97	0.76	50	63	10	30	9
Highrock Lake Batholith[1]	67.5	16.0	0.65	1.9	0.58	1.14	22	29	17	12	25
Flatrock Lake granite[1]	72.9	14.6	0.34	0.6	1.40	1.15	11	29	17	9	11
SW Duval Lake synkinematic leucogranite[1]	73.8	15.0	0.12	0.3	1.25	1.09			20	12	8
Burntwood Lake tonalite dike[1]	68.2	16.7	0.43	1.7	0.39	1.08	20	38	17	73	5
Hobbit Lake road synkinematic tonalite[1]	71.2	15.3	0.30	0.9	0.45	1.04	3	6	23	128	2
Proterozoic calcalkaline granitoid (CA)[2]	66	15.3	0.60	1.3	1.10	1.01	18	40	12	9	10
Proterozoic tonalite-trondhjemite-granodiorite (TTG)[2]	69	15.5	0.19	1.4	0.66	1.00	19	29	23	49	13
Cenozoic adakite, California[3]	69.3	15.5	0.31	1.5	0.16	0.98	(10)	(25)	23	164	
Tonalitic leucosome, Sardinia[4]	71.9	16.1	0.24	0.6	0.48	1.30	11	20	16	33	
Granodiorite leucosome, southern Italy[5]	74.1	13.6	0.30	0.7	1.10	1.17	9	44	17	11	
Tonalitic leucosome, southern Italy[5]	72.0	16.1	0.20	0.7	0.18	1.09			12	29	

[1] Averages as in Table 11

[2] Medians (Condie, 2008)

[3] Most felsic adakite, Baja, California (Aguillon-Robbles et al., 2001); in brackets: Austral Andies (Richards and Kerrich, 2007)

[4] Selected analysis (Cruciani et al., 2007)

[5] Selected analysis (Fornelli et al., 2003)

Fire Lake gneissic granodiorite–tonalite near the Puffy minesite (sample locations 209 and 210, Map GR2010-1-2, DVD) forms a marginal phase in the Archie Lake pluton and occurs near Martell Lake at the margin of the Gants Lake Batholith (sample location 807, Map GR2010-1-2). Its more moderate negative slope on chondrite and primitive-mantle diagrams, with a concave-upward shape and minor Eu anomaly, suggest an origin as a calcalkaline melt (Appendix 3, Figure 36e, 37e).

5.8.2.5 Star Lake (tonalite) pluton (unit 9e)

The 1874 ±2 Ma Star Lake gneissic tonalite (Machado et al., 1999) is the largest pluton in the Sherridon–Hutchinson Lake Complex, where it forms the structural roof of the Sherridon gneiss. The tonalite belongs to the high Al_2O_3 (>15% at 70% SiO_2), high La/Yb, low Yb suite (Appendix 2, Figure 24b) that has been interpreted to have been derived possibly from slab melt (Whalen et. al., 1999) but most likely from mafic to tonalitic lower crust (Whalen et al., 2002). The high Sr/Y ratios (111–136) and steeply declined REE pattern lacking a Eu anomaly (Appendix 3, Figure 36e) are found in modern adakite (Defant et al., 1991) and in modern to ancient tonalite-trondhjemite-granodiorite (TTG), as defined by (Condie, 2008). The Star Lake tonalite and Archie Lake granodiorite show less fanning of HREE than more modern adakite. The plutons were likely triggered by subduction of the relatively hot, ca. 1.89 Ga back-arc basin basalt that is found throughout the THO internal zone. The voluminous generation of such granitoid batholiths, however, required thickened (arc) crust that contributed significantly to the melts.

5.8.2.6 Batty Lake (trondhjemite gneiss) Complex (unit 9f)

The Batty Lake trondhjemite gneiss forms the main, originally medium-grained, intrusive phase in the Batty Lake domal complex (Zwanzig and Schledewitz, 1992). The rock follows the tholeiite boundary of Irvine and Baragar (1971), being slightly more Fe enriched or alkali depleted than typical calcalkaline plutons (Appendix 1, Figure 21c); it also trends into the tholeiitic (or high-Yb) field in Th/Yb versus Zr/Y space (Appendix 2, Figure 27d). It shares this trend with other orthogneiss that contains local garnet, including slightly alkali-depleted Sherridon gneiss, which suggests alteration related to melt loss or alkali removal during shearing or, possibly, early hydrothermal alteration. Batty Lake gneiss also shows considerable scatter across and along the divide between the metaluminous and peraluminous fields in the Al-saturation diagram (Appendix 1, Figure 23j), a scatter probably amplified by alkali element mobility during regional metamorphism, with development of garnet and leucosome.

Despite alteration, the major- and trace-element geochemistry of the Batty Lake gneiss is similar to low-Al (13–14%) trondhjemite from modern arc terranes (Stork, 2004). It has high SiO_2 and Na_2O, and low K_2O/Na_2O and MgO (Appendix 4, Table 10, DVD). It has low La/Yb and a gentle REE slope with a weak Eu anomaly (Appendix 2, Figure 24b; Appendix 3, Figure 36f). The gneiss has only a weak negative Nb anomaly (Appendix 3, Figure 37f), which, along with its high modal quartz content (up to 50%) and variable Rb and Ba, gives it an unusual composition for a volcanic-arc pluton. Two of three samples, however, retain a high Ba/Nb ratio. This is best explained by variable loss of partial melt from an arc-related trondhjemite. An island-arc origin in a bimodal suite like the apparently coeval Sherridon–Meat Lake assemblage has been suggested for similar trondhjemite formed in the Phanerozoic (e.g., Leo, 1985).

The Batty Lake gneiss has trace- and major-element contents (except MgO and K_2O) similar to those of the Sherridon gneiss from the Sherridon structure (Appendix 4, Table 10; Appendix 3, Figure 37f), with which it also has an overlapping U-Pb zircon age within the determined high error. Moreover, the northeastern and northwestern margins of the complex and the narrow internal belt of unit 4a supracrustal gneiss all contain local gedrite-bearing alteration rocks like those at Sherridon. At the northeastern margin of the complex, these have become part of the outer schistose zone, mapped as unit 10b tectonite and probably consisting of supracrustal rock. On the northwestern margin of the complex, well-developed cordierite-gedrite–bearing gneiss was considered as early alteration stoped into the trondhjemite magma (Zwanzig and Schledewitz, 1992). However, the trace-element geochemistry and age, similar to the Sherridon rhyolite gneiss, suggest that the trondhjemite had a similar melt source in the mantle wedge and crust. It is therefore more likely that hydrothermal alteration occurred in the roof zone and volcanic cover of the pluton, and that the pluton provided the heat source. The presence of unit 9f in the Defender Lake dome and the uncertainty of an extrusive origin for some of the coarse-grained Sherridon gneiss of similar composition suggest that related trondhjemite–granodiorite intrusions also occurred outside the Batty Lake Complex.

The deep crustal environment suggested by the sheath-like structural culminations would have been conducive to partial melt removal and crystallization of the younger (1810 Ma) igneous-looking zircon in the Batty Lake trondhjemite (Hunt and Zwanzig, 1993). If the higher level Walton Lake nappe formed the cover of such a complex but the Sherridon structure remained at depth, the differences in geochemistry of these rocks may have been partly the result of upward migration of melt and fluid. Synmagmatic hydrothermal alteration, evidenced by the gedrite-bearing rocks, may have affected the geochemistry earlier on.

5.8.3 Early successor-arc magmatism: discussion

Although voluminous felsic volcanism and plutonism occurred during the early arc evolution of the Snow Lake subdomain, the main period of granitoid crustal growth in the Central Flin Flon and Batty Lake subdomains was during the early successor-arc magmatism. The felsic composition and large size of the plutons, such as the Gants Lake Batholith and its northwestern extension into the Ragged Lake and Star Lake bodies, as well the younger Batty Lake Complex and Sherridon felsic volcanic gneiss, indicate that the mafic–intermediate island-arc crust had attained a considerable thickness during amalgamation of the early assemblages. The highly varied geochemical affinities of the plutons suggest an origin from fluid-fluxed melt components derived from the mantle wedge and from the complexly stacked mafic–felsic arc crust.

The ca. 1874–1860 Ma (early successor-arc) plutons in the Kississing–File lakes area have the geochemical and Nd-isotope characteristics of predominantly medium-K calcalkaline to adakite- or TTG-like rocks. They originated in a volcanic arc that was built on the Amisk Collage. The latter, which acted as basement, must have included not only the early juvenile assemblages but also local Archean crustal fragments or structural slices that contaminated some of the successor-arc magma. Parts of such crust may have descended with the subducting slab or been stacked with a thick collage in which melting or assimilation took place at the base. The Ragged Lake granite, which shows significant contamination by Archean crust, indicates that the tectonic assembly occurred before 1873 Ma. One of the sampled plutons (the Star Lake tonalite) is adakitic; another (the Archie Lake pluton) has TTG geochemistry. The Star Lake tonalite is interpreted to have formed by melting of mafic crust or a hot, young (shallow?) slab, probably formed in an adjacent back-arc basin, as suggested for similar plutons in the main part of the Central Flin Flon subdomain (Whalen et al., 1999). The Archie Lake pluton, which intruded the Ponton Lake basalt (arc-rift) and the Puffy Mine andesite (arc), may have a mixed origin. Other plutons in the Sherridon–Hutchinson Lake Complex have the geochemical fingerprint of arc magma derived by fluid-fluxed melting of the mantle wedge, although none of the sampled plutons are high-Ba granitoid rocks (Appendix 2, Figure 28f). The Jay Lake granite may have involved some melting of felsic crust to produce non-arc chemistry during extension tectonics. The Batty Lake trondhjemite probably formed from depleted arc crust and mantle during successor-arc magmatism. It may have been altered by incipient melting during metamorphism.

If the preliminary U-Pb age of the Sherridon Group is valid, successor-arc volcanic rocks were formed in an oceanic tectonic setting similar to that of the early juvenile-arc assemblages. They have a geochemistry similar to that of older rocks in the Snow Lake subdomain, as well as to that of coeval trondhjemite at Batty Lake. Successor-arc plutonism in the Central Flin Flon subdomain was probably in a subaerial arc and did not promote widespread VMS deposition, whereas the Batty Lake subdomain remained as an oceanic environment. Its volcanism was also coeval with plutonism and subject to extensive hydrothermal alteration and VMS deposition.

5.9 Basinal and late successor-arc rocks

The early successor-arc plutonism and attendant compressional tectonics turned the mafic–intermediate crust formed by the Amisk Collage and Snow Lake assemblage into relatively thick crust of continental composition. The subsequent evolution of the Kississing–File lakes area entailed a change to widespread erosion of elevated terrains and sedimentation in marine and terrestrial basins (Burntwood and Missi groups). The subduction of a remnant ocean basin under the thickened arc crust resulted in the intermediate- to high-K, late successor-arc magmatism of the Missi Group and led to closing of the Kisseynew basin and collision of the arc terranes with the Superior craton. The metasedimentary rocks have geochemical patterns similar to some units of felsic tuff, consistent with the close stratigraphic relationship of these rocks discussed in Section 3.5.1.2.

Highly metamorphosed flows, dikes and units of tuff, which are identified using stratigraphic and geochemical criteria as belonging to the Missi Group, occur throughout the Kississing–File lakes area. They are generally intercalated with the fluvial-alluvial sedimentary rocks that dominate the Missi Group and there appear to be lateral transitions from felsic tuff and breccia to tuffaceous and lithic arenite. Missi-age plutons are concentrated in the Snow Lake subdomain and were analyzed mainly from the vicinity of File Lake. They share many of their geochemical characteristics with the volcanic rocks. Dikes with similar geochemistry also intrude the margin of the Amisk Collage but are difficult to distinguish from older dikes. The geochemistry suggests that the late successor-arc magmas may have acquired a melt component from subducted early Burntwood Group sedimentary rocks or early volcanoplutonic rocks carried down from the base of the arc massif with the subducting slab. The juvenile Nd signature of the late arc magmas indicates that mainly earlier juvenile crust was recycled with little input of Archean material.

5.9.1 Bob gneiss, rocks of uncertain origin (unit 10a)

The garnet-biotite±hornblende±graphite gneiss (Bob gneiss), which occurs as a unit (10a) within Sherridon felsic gneiss (unit 5), has a greywacke composition (Froese and Goetz, 1981) but is preferentially interpreted by Tinkham (pers. comm., 2008) as igneous because of its uniformity. A selected element pattern is used to distinguish different groups of paragneiss as normalized to Paleoproterozoic continental pelite and semipelite (average P2 member of the Pipe Formation in the Ospwagan Group; after Zwanzig et al., 2007). The pattern for the Bob gneiss closely resembles that of the Burntwood and Missi groups (Appendix 3, Figure 38h). Moreover the ε_{Nd} value (Table 15) and detrital zircon population (N. Rayner, pers. comm., 2008) of Bob gneiss are consistent with those of the Burntwood Group. The unit may be correlated with sedimentary rocks of Burntwood age that structurally underlie the Sherridon gneiss, and mafic to locally carbonate-bearing Walton Lake amphibolite south of Walton Lake. The contact may be a fault, but a disconformity is also possible because the preliminary zircon age of the Sherridon gneiss is similar to that of the Bob gneiss.

A unit of garnet-biotite paragneiss containing variable amounts of amphibole and scattered Fe-sulphide is designated as unit 10a on Yakushavich Island and Collins Point in the northern part of Kississing Lake. This unit may be related to the Sherridon gneiss because it contains local cordierite-gedrite rock (unit 5d, alteration), but elsewhere it resembles unit 11d of the Burntwood Group (below). Several base-metal occurrences are hosted in this unit, thus supporting of volcanogenic affinity.

5.9.2 Burntwood Group: basinal rocks (unit 11)

The wide extent of the Burntwood Group in the Kisseynew Domain and across the eastern part of the Flin Flon Domain makes this an exceptionally important unit in understanding the late tectonic and metallogenic evolution of the Kississing–File lakes area (Zwanzig, 1990, 1999). Well-preserved metagreywacke and mudstone with clastic texture and turbidite structure occur south of File Lake, where these rocks have been studied in detail (File Lake Formation of Bailes, 1980a, b). Archived samples collected during the original mapping have been reanalyzed and provide a small geochemical database for these rocks. These data are useful for stratigraphic correlation and assessing the possible source of crustal melt components of the younger igneous rocks.

The gneissic–migmatitic part of the Burntwood Group, which forms its vast majority in the THO north of the Flin Flon Domain, has also been sampled beyond the Kississing–File lakes area, where moderately well preserved beds still exist. Aspects of their geochemistry are presented for comparison with the low-grade rocks. This small database includes samples from the southern and eastern flanks of the Kisseynew Domain. The data are moderately affected by the high-grade metamorphism and incipient migmatization (Appendix 3, Figure 38g): LREE are nearly identical and little affected by metamorphism and the HREE pattern is more scattered in the lower grade rocks, probably by the variable distribution of heavy minerals in the turbidite, as indicated by the feldspar-rich bed. The primary variation appears to be greater than that caused by any incipient migmatization.

Neodymium-isotope analyses from a subset of these samples have yielded Paleoproterozoic model ages (generally 2.1–2.5 Ga and mainly with positive ε_{Nd} values; Böhm et al., 2007): they are consistent with the main source of detritus coming from the surrounding coeval and older igneous rocks, mixed with a small Archean detrital component.

The origin of the Kisseynew Domain and the important Morton Lake–Loonhead Lake structural break, which divides the Central Flin Flon subdomain from the Snow Lake subdomain and the Kisseynew Domain, depends partly on the interpretation of the tectonic setting of the Burntwood Group (e.g., Zwanzig, 1999).

5.9.2.1 Greywacke-mudstone (unit 11a)

A suite of samples of the Burntwood Group from the File Lake area and one sample from Snow Lake were reanalyzed to include trace-element data. Four of these samples are greywacke, four are mudstone and two are Corley Lake mudstone from the upper part of the Burntwood Group, all originally assigned to the File Lake Formation and collected by Bailes (1980a). The Burntwood Group geochemical data occupy a central field among the Missi-age volcanic rocks in REE space and on a Yb versus $[La/Yb]_N$ diagram, and overlap some of the early felsic volcanic and intrusive rocks as well as the felsic Sherridon gneiss (compare Appendix 2, Figure 24b and 24c). A significant number of samples, however, have La/Yb ratios that are higher than those of the Amisk Collage and the Sherridon felsic volcanic gneiss. These were probably eroded from the early successor-arc plutons that lie below the Missi unconformity. This is also consistent with the high plagioclase and low K_2O contents that the Burntwood Group shares with many early plutons. One sampled plagioclase-rich bed is particularly depleted in HREE. Ratios of La/Yb also lie between the most depleted Missi-age basalt (Kississing River basalt) and the highly fractionated Lobstick rhyolite. This is consistent with coeval ash as the source of some of the detritus. Most samples form a simple chondrite-normalized pattern with flat MREE and HREE and moderately fractionated LREE, similar to Puffy Lake dacite and Cleunion Lake rhyolite (Appendix 3, Figure 38), as well as Sherridon gneiss and older rhyolite (Appendix 3, Figure 33f).

The age of detrital zircon grains also indicates that a significant volume of detritus from early successor-arc igneous rocks is present (David et al., 1996; Machado et al., 1999). The presence of poorly rounded crystal fragments, including pyramidal quartz grains, has been interpreted to indicate a neovolcanic or synvolcanic component (Bailes, 1980b). This is confirmed by the youngest detrital zircons (1842 and 1845 Ma), which show few signs of sedimentary abrasion (Machado et al., 1999). It is also consistent with a component of felsic ash present in the Burntwood Group. Thin beds interpreted as felsic tuff are recognized only at the top of the Burntwood Group near Puffy Lake and possibly within the group farther northeast. Such beds are impossible to recognize in the migmatitic part of the group and therefore may have been more widespread.

5.9.2.2 Greywacke-derived gneiss (unit 11b) and migmatite (unit 11c)

The high-grade part of the Burntwood Group is characterized by well-layered, graphitic garnet-biotite gneiss, in which graded bedding is only very locally preserved on the South flank of the Kisseynew Domain. It is geochemically very similar, however, to its turbiditic greywacke-mudstone protolith (Appendix 2, Figure 24c; Appendix 3, Figure 38g; Zwanzig, 1990). Much of the gneiss contains 3–10% quartz-plagioclase–rich pods and veins of leucosome oriented parallel to original bedding and yielding enhanced layering. North of the migmatite front, granitoid veins, lenses and sheets of various thicknesses increase in volume toward the Central Kisseynew subdomain and toward high-strain zones that apparently formed during melt-enhanced deformation. The habit of the leucosomes (discussed in Section 4.3.2) and their geochemistry (discussed in Section 5.10.4) may provide clues to their origin by *in situ* melting, injection or flow-through of melts. The leucosome generally has thin, biotite-rich selvages, and the resulting neosome has therefore been avoided in geochemical sampling. Limited geochemical data suggest that the paleosome was little affected by limited generation or intrusion of the veins. Alteration of garnet to biotite where small granitoid bodies are abundant suggests local alkali metasomatism during crystallization of leucosomes and back-reaction of remaining fluid. The most altered mesosome contains only biotite as the main mafic mineral and has late muscovite that cuts earlier fabrics.

5.9.3 Missi Group conglomerate- and sandstone-derived gneiss (units 12–13)

The paragneiss units derived from the shallow-water sedimentary rocks of the Missi Group are generally more felsic and have higher SiO_2 and K_2O contents than the Burntwood Group (Appendix 4, Table 10, DVD), consistent with their more arkosic appearance and higher quartz and K-feldspar contents. The Missi Group sedimentary rocks were likely eroded from largely granitoid crust exposed in the surrounding greenstone belts and batholithic hinterland. However, trace-element contents, which are dominated by fine-grained components, are nearly identical in the Burntwood, Missi and correlative groups, suggesting a similar source of silt and clay in a largely juvenile Paleoproterozoic arc(s) (Zwanzig et al., 2007). A comparison of their geochemistry with the intermediate–felsic volcanic rocks in the Missi Group is also consistent with a significant component of coeval tuff (Appendix 3, Figure 38).

5.9.4 Missi Group volcanic rocks: emerged arc (unit 14, units 14a–f)

Volcanic rocks in the Missi Group are divided into five units, generally 1–50 m thick in their deformed (flattened) state but locally 400 m thick. They are most prominent in the large area of Missi Group gneiss north of Kisseynew Lake in the core of the Star Lake syncline, and in a fault block along the Kississing River. Composite units at the Kississing River are 100–300 m thick and constitute 20–35% of the regional section. The best preserved samples for geochemical analysis originate from thin volcanic units interlayered with the arenite units at Puffy Lake (Maps GR2010-1-1 and -2, back pocket/DVD). Some relatively well preserved tuff and flow units occur as far north as Jungle Lake, thus allowing a wide-ranging correlation of individual units.

Fine-grained felsic units up to 300 m thick, interpreted as metarhyolite, occur in the Missi Group 2–5 km north of Kisseynew Lake. Thicker units, which are farther north and east, may be partly intrusive domes, whereas thinner (0.2–3 m) units, which extend for several tens of kilometres along strike, were locally recognized as rhyolitic ignimbrite (Zwanzig and Schledewitz, 1992). The metarhyolite is very fine grained, locally with 1 mm phenocrysts of quartz or feldspar. Dacitic units, which are interlayered with the rhyolite and Missi Group sedimentary rocks, form felsic ribbon gneiss interpreted as possible tuff and breccia. Less abundant dacite units form sills in the Amisk Collage. Other felsic bodies, which are fine to coarse grained, are interpreted as subvolcanic intrusions broadly coeval with the rhyolite. These include a porphyritic body at Thunderhill Lake and the Nelson Bay gneissic dome northeast of File Lake. The latter may be a resurgent dome of the same composition as units that extend to its west and are herein interpreted as ignimbrite.

The composition of the Missi Group volcanic rocks is bimodal, including basalt to andesite and rhyolite but with few rocks in the 60–70% SiO_2 range (Appendix 1, Figure 19d). They are distinguished as a group not only by their stratigraphic position and subaerial extrusive habit but by their elevated total alkali content, which is variable but, on average, much higher than the early volcanic assemblages and also significantly higher than the early successor-arc intrusions. The K_2O content ranges from low in one mafic unit, to dominantly intermediate in basaltic andesite, to high in the rhyolite (Appendix 1, Figure 20g). The wide scatter of the K_2O data within units is attributed to alkali mobility (Section 5.5). Nevertheless, K_2O values vary systematically among the different units (Appendix 1, Figure 23g, h), which is interpreted to reflect primary compositional variation, albeit not accurately.

Most mafic units do not show strong Fe fractionation, but the Kississing River basalt has a range of Mg# (47–67). This widespread unit plots mainly in the tholeiite field on an AFM diagram (Appendix 1, Figure 21d), whereas other units typically show a calcalkaline trend. The tholeiitic to calcalkaline trend is also clearly seen on the immobile-element ratio diagram (Appendix 2, Figure 27e). Several units, including Kississing River basalt, have members with high Al contents, particularly the plagioclase-phyric rocks (Appendix 1, Figure 22g). Titanium contents are relatively high for arc-volcanic rocks; more than half the analyses fall into the BABB field on Appendix 1, Figure 22d, but this is probably the result of an enriched mantle source or low degree of melting (Sections 5.9.4.1–3).

Compared to the earlier juvenile-arc rhyolite, the Missi Group rhyolite gneiss units are less mafic (average 0.41% MgO) and have higher K_2O (average 4.6%) and lower average Mg# (24), the early rhyolite averaging 0.84% MgO, 2.8% K_2O and Mg# 46 (Appendix 4, Table 10, DVD; Appendix 1, Figure 23). Based mainly on trace-element data, rhyolite gneiss in the Missi Group divides into two types: Cleunion rhyolite and Lobstick rhyolite. Field observations indicate that the latter is generally a deeper brick red rather than the pink hue of the former. Average MgO content (0.29%; Mg# 0.19) indicates that Lobstick rhyolite is more leucocratic than Cleunion rhyolite (0.53% MgO; Mg# 0.29), but element mobility has affected these numbers even when samples with anomalous values were excluded. Lobstick rhyolite is best distinguished by its low Yb content, whereas Cleunion rhyolite has the same REE content as Puffy Lake dacite (Appendix 2, Figure 24c).

The five Missi Group volcanic units are distinguished by REE contents and REE ratios, and form separate fields on Yb versus $[La/Yb]_N$ plots (Appendix 2, Figure 24c). These units were identified most clearly by the geochemical data. Once they were recognized, the units were also evident from field descriptions based on mineral content, colour and phenocryst populations. A sixth unit (dacite) is distinguished from a rhyolite unit by its greyer appearance in the field and its lower hardness and SiO_2 content. the units are, in order of increasing La/Yb ratio, Kississing River basalt (the most abundant mafic unit), Evans Lake basalt (the most Mg rich), Cleunion rhyolite, Puffy Lake dacite (partly in dikes), Jungle Lake basaltic andesite and Lobstick rhyolite (with elevated Sr and very high La/Yb). One or two of the units have the steep Yb trend of high-level fractionation, whereas the rest show increasing La/Yb related to more complex processes. Most of these units plot on linear trends in Appendix 2, Figure 26i–l, where the incompatible element La shows a consistent increase along with the more weakly incompatible element Zr (which tracks crustal input or fractionation except in its most advanced stage). Mobility of La in these units during high-grade metamorphism and incipient melting is interpreted as the cause of the greater scatter in plots involving this element than for the early volcanic rocks that are less prone to melting. However, this does not detract from the usefulness of Appendix 2, Figure 24c in separating the units.

The LILE (Rb, Ba) plotted against Sr/Y show extended overlapping fields but with different mean concentrations for the various units. They suggest that these ratios, which are altered from the originals, still provide limited but useful data

(Appendix 2, Figure 28c, g). The intermediate to felsic units show weakly to moderately elevated Ba; the felsic units also show elevated Rb, consistent with a magma origin by fluid-fluxed melting of the mantle wedge.

The highly varied geochemical signatures of the Missi Group volcanic rocks are interpreted to indicate that multiple sources, including crust and mantle, contributed to their origin. The high Nb/Yb and Zr/Y ratios and elevated Zr contents of the units are consistent with the field evidence that the Missi Group volcanic rocks erupted in a continental-type or very mature island-arc environment, where various sources can be tapped (Appendix 2, Figure 29b, e). High Th/Yb ratios of the felsic rocks are also consistent with a calcalkaline affinity (Appendix 2, Figure 29c), as are high La/Y ratios of the mafic rocks (Appendix 2, Figure 31e). In general, the high LILE, Th and LREE contents of most of the Missi Group volcanic rocks suggest dramatic crustal thickening *preceding* the volcanism. This is consistent with their nonmarine environment of deposition. The bimodal magmatism, however, suggests crustal extension and local mantle upwelling *during* Missi Group magmatism. The mantle source must have been highly contaminated with subduction components, possibly accumulated over an extended period or provided by a sudden influx of sediments into the subduction zone. The elevated LILE and fractionated REE are also consistent with the addition of fluid and silicate melt from a subducted sediment source to the mantle wedge (e.g., Gertisser and Keller, 2003). Generally LILE- and LREE-enriched arc magmatism occurs above a deeply subducted portion of the slab, according to Woodhead and Johnson (1993) and Churikova at al. (2001). These authors favour decreased degrees of melting with depth and strong partitioning of LILE and LREE into the early partial melt. This tectonic history is consistent with the volcanic evolution of other remnant ocean basins during terminal continental collision.

Missi Group basalt and basaltic andesite have relatively high ε_{Nd} (+2.7 to +3.4; Appendix 2, Figure 29f), indicating little or no contamination by Archean crust or subducted sediments with a high Archean component. The similarity of these values to those of the Burntwood Group is consistent with a significant input of these sediments into the subduction environment but equally consistent with a deep crustal source directly from the arc massif that was elsewhere exhumed as the main erosion source of detritus for the Burntwood Group.

A sample of Missi Group rhyolite gneiss has a relatively high ε_{Nd} of 3.4, similar to coeval felsic plutons and mafic volcanic rocks (Table 15; Appendix 2, Figure 29f). This rules out a significant contribution by Archean crust, even though the rhyolite generally has high SiO_2 content (average of 74% with a spread to 77%, excluding two outliers; Appendix 4, Table 10). Such rocks are more commonly found in an isotopically evolved continental arc during periods of high magma flux (Ducea and Barton, 2007). Juvenile bimodal basalt to rhyolite, albeit with slightly different chemical composition to Missi Group volcanic rocks, were deposited during alternating compression and extension in the Andean foothills of central Chile (Nyström et al., 2003). The Chilean environment is, nevertheless, consistent with that of the Missi Group in 1) the bimodal character of the Missi Group volcanic rocks, 2) prominent caldera development, 3) interlayering of volcanic rocks (mainly tuff) with fluvial sedimentary rocks, 4) basin development during block faulting, and 5) syndepositional deformation and uplift of the hinterland.

The high ε_{Nd} constrains the origin of Missi Group rhyolite to partial melting of mantle or juvenile crust. Low-pressure fractionation is in accord with the parallel REE patterns and increased negative Eu anomaly in Cleunion rhyolite (Appendix 3, Figure 38e). The difference between the Cleunion and Lobstick rhyolites can be explained by the presence of garnet in the source of melting in the latter. Heavy REE are partitioned preferentially into garnet, which may have been absent or all of which may have been consumed in Cleunion rhyolite melt but not in the less mafic Lobstick rhyolite melt, leaving the latter deficient in HREE (Appendix 3, Figure 38f). A lower fraction of batch melting and possibly a higher felsic crustal component for Lobstick rhyolite are supported by the higher Sr content (e.g., Mahlburg Kay et al., 2005), as well as by a higher average Na_2O/K_2O ratio (0.85 compared to 0.71). Many individual samples do not show a clear correlation, but this is to be expected given the high alkali element mobility. Significant alteration due to development of migmatitic leucosome and garnet is ruled out because Missi Group metarhyolite is garnet free and most of the samples were taken south of the migmatite front.

5.9.4.1 Kississing River basalt (unit 14a)

The widespread unit of Kississing River basalt is the most primitive member of the Missi Group volcanic rocks. It generally occurs throughout the sequence 50–1000 m above the stratigraphic base (Figure 8b–e, back pocket), where it is commonly associated with Cleunion rhyolite. A thick pile (>300 m, structurally flattened) of porphyritic basalt and related gabbro lies south of the east-flowing reach of the Kississing River below the exit from Kisseynew Lake. The aggregate thickness of the unit may be about 400 m of mainly basalt at Jungle Lake. In contrast, members of Kississing River basalt at Puffy Lake range from <5 to 10 m thick. Individual measured flows range from 0.15 to 8 m. Flow contacts are commonly amygdaloidal or marked by calcsilicate alteration. Bodies of mafic porphyry similar to but coarser than the basalt are interpreted as intrusive or have uncertain field relationships. The flows include aphyric, hornblende- or originally pyroxene-phyric, and plagioclase-phyric varieties. A distinctive variety, possibly intrusive, contains plagioclase phenocrysts, up to 50 mm long, that are locally skeletal or glomeroporphyritic.

A significant number of sampled flows have low K_2O, moderate Fe enrichment and weakly elevated TiO_2 contents (Appendix 1, Figure 20g, 21d, 22d). The content of La is three times that of N-MORB (Appendix 2, Figure 26i). The least fractionated samples are similar to slightly enriched MORB (from weakly depleted mantle) with higher Nb and relatively lower Th contents than most arc-volcanic rocks (Appendix 2, Figure 29b, 30e). The data suggest that Kississing River basalt is transitional between volcanic-arc and continental basalt (Appendix 2, Figure 31e). Fractionation of LREE is small (Appendix 3, Figure 38a), but negative HFSE anomalies are prominent and give an extended-element pattern similar to that of the older Ponton Lake arc-rift basalt (compare Appendix 3, Figures 39a and 34b). These characteristics are also identical to the Ferro Mine basalt, a prominent arc tholeiite in the main outcrop area of Missi Group volcanic rocks at Wekusko Lake, on the west end of the exposed Flin Flon Domain (Gordon and Lemkow, 1987). Original mapping at Wekusko Lake (Stockwell, 1937) and sampling by Gordon and Lemkow (1987) place this unit at the top of the volcanic succession (see, however, Ansdell et al., 1999). The original stratigraphic interpretation is preferred in this report, and, coupled with the geochemistry, suggests an influx of primitive magma during the block faulting and sedimentation in the Missi basins.

Similar tholeiitic basalts occur in volcanic arcs in various settings, such as on Java, in the Mediterranean and in the Andean foothills of Chile (Huijsmans et al., 1988; Mitropoulos and Tarney, 1991; Carn and Pyle, 2001; Nyström et al., 2003). These settings are extensional, either in retro-arc environments (Churikova et al., 2001), in the front part of arcs during rapid subduction (Jodan et al., 2001) or as terminal arc magmatism during progressive continental collision (Mitropoulos and Tarney, 1992). These environments have been related to extensional basin development (Jordan et al., 2001) and asthenosphere upwelling (Carn and Pyle, 2001). Terminal arc magmatism during extension, related to the onset of collision in the THO, can explain the geochemistry and the juvenile Nd-isotope signature of Kississing River basalt.

5.9.4.2 Evans Lake basalt (unit 14b)

Evans Lake basalt occurs as generally aphyric massive flows with vesicular margins. These extend from Evans Lake to the northwest, mostly in the lower part of Missi Group volcanic sections. Units are up to 60 m thick but locally only 1–2 m thick. They have the highest contents of MgO (7.8–10.6%), Ni (138–276 ppm) and Cr (249–758 ppm) in the Missi Group. They show no Fe enrichment, indicating that they may be calcalkaline, but plot in a transitional field in Th/Yb versus Zr/Y space (Appendix 2, Figure 27e). Ratios of La/Yb are distinctly higher than those in Kississing River basalt and increase with only a slight increase in fractionation (i.e., Yb in Appendix 2, Figure 24c) but with an increase of SiO_2, LILE, Nb and Zr (i.e., crustal elements in Appendix 4, Table 10, DVD; Appendix 2, Figure 26j). A Th/Zr/Nb plot indicates a volcanic-arc affinity (Appendix 2, Figure 30e), similar to modern continental arcs (Appendix 2, Figure 29b). A La/Y/Nb plot confirms the calcalkaline trend (Appendix 2, Figure 31e). Light REE are fractionated (Appendix 3, Figure 38b) and negative HFSE anomalies as well as a Th spike are very prominent, similar to most of the Missi Group volcanic rocks and their coeval intrusions (Appendix 3, Figure 39). This pattern is distinctly different from Kississing River basalt, to which Evans Lake basalt cannot be related by crystal fractionation.

A possible origin of Evans Lake basalt is by mixing of melts from mantle and crustal sources enriched in LILE and LREE, either from early-subducted sediments or from the Flin Flon–Glennie arc massif. This explains the observed fanning of the REE pattern and SiO_2 that is relatively high at elevated MgO and increases with La/Yb. The mantle melt may have been tholeiitic, whereas the crustal source was calcalkaline to arrive at a transitional composition. The ε_{Nd} that is slightly lower than Kississing River basalt (+2.7 compared to +3.4) also suggests contamination by a crustal component. A similar geochemical pattern could result from melting of highly metasomatized heterogeneous sub-arc mantle, also explaining the high Ni and Cr contents. The geochemical and Nd-isotope data indicate that, regardless of origin, the light- to intermediate-REE components are close to juvenile and came partly from an originally crustal source, whereas the HREE and Ti reflect a depleted mantle wedge.

5.9.4.3 Jungle Lake basaltic andesite (unit 14c)

Porphyritic basaltic andesite occurs in a 2.5–50 m thick unit, or in several units, most commonly in the upper part of the mafic-volcanic sections. Some layers are recognized as massive flows with amygdaloidal contacts, whereas others are interpreted as tuff because they grade into overlying arenite, with contacts suggesting sedimentary reworking. Some units in the north have developed weak gneissic layering. The geochemistry of the basaltic andesite is characterized by elevated K_2O+Na_2O, with nearly half the samples containing between 5.0% and 6.9% (Appendix 1, Figure 20c). The same sample suite has elevated P_2O_5 (averaging 0.34%) compared to the older Preaston andesite at the same SiO_2 content (average of 55% SiO_2), but with the older andesite having averages of 3.5% K_2O+Na_2O and 0.15% P_2O_5. Jungle Lake basaltic andesite is calcalkaline (Appendix 2, Figure 27e) and may be transitional to alkaline if its composition was not heavily affected by alteration. The K_2O contents range from about 0.8% to 2.8% (excluding an outlier), placing the unit in the medium-K range of volcanic rocks. About half the samples of Jungle Lake basaltic andesite also have relatively high TiO_2 and Al_2O_3 contents (Appendix 1, Figure 22d, g). Contents of Ni and Cr are high, very similar to Evans Lake basalt at similar MgO contents, and distinctly higher than Kississing River basalt (Appendix 4, Table 10, DVD).

In addition to a steeply negative slope of REE with fractionated LREE (Appendix 3, Figure 38c), Jungle Lake basaltic andesite has high Nb/Yb (averaging ~3.9) and elevated Na_2O (averaging ~3.3%) and Sr (Appendix 2, Figure 28c), suggesting a possible low degree of partial melting (Fretzdorff et al., 2002). Some of the indicative elements are highly mobile, however, and plots show a wide scatter. Moreover, arc magmas generally have high contents of H_2O that promote melting (Sinton et al., 2003).

A more common explanation of the origin of high LILE and highly fractionated LREE in arc-volcanic rocks is the effect of sediments carried down the subducting slab and contaminating the mantle wedge that subsequently yields LREE-enriched melt (Jolly et al., 2001; Clift et al., 2005). Mixing of magma from a mantle source with an LILE- and LREE-rich crustal melt has the same effect. Contamination by sediments, however, is supported by trends of increasing Th/La from 0.08 in Evans Lake basalt to an average of 0.15 in Jungle Lake basaltic andesite and 0.35 in Puffy Lake dacite dikes (Section 5.9.4.4). The Th/La ratio in island arc suites is generally controlled by terrigenous sediments (Jolly et al., 2001; Clift et al., 2005). The relatively low contents of Nb and Ti in Jungle Lake basaltic andesite, close to the values in Evans Lake basalt, and the similar contents of HREE represent the mantle-wedge composition, as suggested for similar modern rocks (Woodhead et al., 1993; Jolly et al., 2001). This provides further support for processes other than fractionation in the formation of the highly enriched Jungle Lake basaltic andesite and the closely related Evans Lake basalt (Appendix 3, Figure 39b, c). At Evans Lake, the basaltic andesite lies stratigraphically atop the high-Mg Evans Lake basalt, forming a progressively evolved volcanic suite (column 7A in Figure 8d, back pocket).

Thus, by analogy with modern environments, we conclude that Jungle Lake basaltic andesite had multiple sources of melting or a highly subduction-enriched mantle wedge above a deeply subducted slab with a high melt column. The crustal component must have been relatively juvenile. In the genesis of similar modern rocks, some of these components are believed to contaminate the mantle wedge with phases, such as amphibole and phlogopite, that melt upon further heating as mantle is dragged deeper down the subduction zone (e.g., Woodhead and Johnson, 1993; Gertisser and Keller, 2003). The high Nb/Yb ratio of Missi basalt to dacite (Appendix 2, Figure 29b) suggests that the mantle wedge also had an enriched (E-MORB) component.

5.9.4.4 Puffy Lake dacite (unit 14d)

Puffy Lake dacite is relatively K_2O rich (3–4%; i.e., rhyodacite) and has REE contents between those of Jungle Lake basaltic andesite and Cleunion rhyolite, with which it is associated in the field and in chemical composition (e.g., column 9D in Figure 8e, back pocket; Appendix 2, Figure 24c). The Rb content and Sr/Y ratio are also intermediate (Appendix 2, Figure 28c), suggesting a fractionation or mixing trend between these units. The chondrite-normalized REE pattern is similar to those of Jungle Lake basaltic andesite and Cleunion rhyolite, with a negative Nb anomaly intermediate between them (Appendix 3, Figure 38c–e) and indicating a relationship by feldspar fractionation or melting. These REE patterns are also very similar to those of Missi Group metasandstone and equivalent metasedimentary rocks from other margins of the Kisseynew Domain (Appendix 3, Figure 38d, h). Puffy Lake dacite may have acquired its composition by magma mixing or sediment influx to the mantle wedge, whereas the sediments could have acquired a similar composition but with lower REE content by mixing of volcanic detritus and more quartzose sand.

5.9.4.5 Cleunion rhyolite (unit 14e)

Cleunion rhyolite occurs in widespread thin units and larger bodies within the clastic rocks of the Missi Group throughout the Kississing–File lakes area. It forms the main (felsic) component of the Missi-age bimodal volcanism and is distinguished by its relatively high content of unfractionated HREE and moderate La/Yb ratio (Appendix 2, Figure 24c). The rhyolite straddles the boundary between volcanic-arc granitoid rocks and within-plate granite (Appendix 2, Figure 25g), with elevated Rb content giving it a strong crustal signature (Appendix 2, Figure 28c). The high Th/Yb ratio is similar to those in extension-related rhyolite in the main Central Flin Flon subdomain (Appendix 2, Figure 29c). The elevated Zr content and generally high Zr/TiO_2 ratio also place the majority of samples into a tectonic field indicative of crustal extension (Appendix 2, Figure 29d). This lends strong support to the local crustal extension suggested for the calcalkaline to tholeiitic evolution of the mafic rocks in the late successor-arc (Section 5.9.6).

Cleunion rhyolite shares its chondrite-normalized REE pattern and negative Eu anomaly and deep trough of Nb, Sr and Ti on mantle-normalized plots with felsic intrusions of similar age in the area, some of which are interpreted to have formed in the same felsic centres as the rhyolite (Section 5.9.5). The rhyolite also shares this REE pattern with the metasedimentary rocks of the Missi and Burntwood groups, which are interpreted to contain a significant component of felsic ash and reworked felsic-volcanic detritus, providing a much greater volume of felsic material than do the volcanic rocks alone (Appendix 3, Figure 38g, h). By virtue of their wide distribution in the Kisseynew Domain, these sediments may include the products of large felsic eruptions and the erosion of a more distant granitoid terrane.

5.9.4.6 Lobstick rhyolite (unit 14f)

Lobstick-type rhyolite generally has a lower content of MgO and TiO_2 than Cleunion rhyolite, indicating that it is more fractionated, has a lower melt fraction and/or contains a higher crustal melt component. It plots in the field of high–La/Yb, low-Yb granitoid rocks, geochemically distinct from the other volcanic units in the Missi Group (Appendix 2, Figure 24c). This rhyolite, which has moderate Rb and Y+Nb, plots in the field of volcanic-arc granite/rhyolite (Appendix 2, Figure 25g). Elevated Zr and Ba contents in some samples give Lobstick rhyolite a strong to variable crustal signature (Appendix 2, Figure 26l, 28c). One way of producing this would be by remelting of the early successor-arc plutons, and magma mixing with mantle-derived melt (e.g., Evans Lake basalt) to acquire the present range of compositions during fractional crystallization. The high Sr content (averaging 276 ppm) and weak or missing Eu anomaly (Appendix 3, Figure 38f) are consistent with a high feldspar content, as these elements are partitioned into feldspar. The strongly depleted and fractionated HREE indicate that garnet remained in the restite from the crustal-melt component.

Without analyses and geochemistry of unaltered (and unmetamorphosed) whole rock and phenocrysts, the complex origin of this unit cannot be clearly determined. The favoured model from the present data is fractionation of mixed mafic melt from the mantle and feldspar-rich crustal melt that left garnet in their residuum of lower crustal rocks or subducted sediments. The chemistry of these rocks is very similar to porphyry intrusions in Chile determined to be fractionated crustal melts mixed with mafic magma underplated at the base of the crust from the mantle wedge (Appendix 3, Figure 41b; Richards, 2003). As is the case for the other volcanic rocks in the Missi Group, the ε_{Nd} value (+3.4 for Lobstick rhyolite) suggests that the crustal melt component came from the juvenile Paleoproterozoic arc basement or its detritus, although recycling of a limited amount of Burntwood Group sedimentary rocks is a possibility.

5.9.5 Late successor-arc intrusions of Missi age (unit 15)

Dated intrusions that range from ca. 1.84 to 1.83 Ga occur in the Burntwood and Missi groups and cut only the earliest post-Burntwood structures. Most of these intrusions are interpreted as late successor-arc plutons, which include the 1839 Ma phase of the Reed Lake pluton (Kraus and Menard, 1997). The youngest includes the 1832 Ma Nelson Bay gneissic dome. The plutons, dikes and Missi Group volcanic rocks of comparable major-element composition have nearly identical REE and multi-element patterns that strongly suggest comagmatic origins for some of these (Appendix 3, Figures 38, 40, 41a–c). The Ham Lake pluton and, in the Central Kisseynew subdomain, the Touchbourne intrusive suite, both yielding ages of about 1830 Ma, may represent the end of arc magmatism during terminal collision.

The late successor-arc intrusions from the Flin Flon Domain have felsic to mafic compositions with only weakly bimodal silica contents (Appendix 1, Figure 19e). The mafic members are more highly sampled than the felsic members, yet their age is less well known. Large mafic–intermediate intrusions assigned to this grouping include gabbro east of Big Island on Kississing Lake, and the Reed Lake diorite southeast of File Lake. Late successor-arc granitoid plutons that have been analyzed include the Norris Lake pluton, Ham Lake pluton, Nelson Bay dome, Thunderhill Lake pluton and Barron Lake pluton (listed with increasing SiO_2 from 64% to 76%). They occur in the File Lake area, except for the Thunderhill Lake body, which is east of Kisseynew Lake. The plutons form a major part of the felsic component of the Missi-age bimodal magmatic suite, with a range in composition similar to those of medium-K dacite to high-K rhyolite in the Missi Group (Appendix 1, Figure 20c, g). Like the coeval volcanic rocks, they include tholeiitic and calcalkaline members with slightly higher FeO^t than the early arc magmatic rocks (Appendix 1, Figure 21e; Appendix 2, Figure 27f).

The late successor-arc felsic intrusions range from metaluminous to weakly peraluminous but show scatter on Appendix 1, Figure 23j due to alkali-element mobility at high metamorphic grade or during a subsolvus fluid circulation. All except the Thunderhill Lake pluton have moderately fractionated REE (Appendix 2, Figure 24c) with moderate Y+Nb and Rb contents (Appendix 2, Figure 25g). They plot in the volcanic-arc granite field, with low Sr/Y showing little evidence for the presence of slab melt (Appendix 2, Figure 28c). Most commonly, chondrite-normalized plots of REE have flat HREE, a negative Eu anomaly and fractionated LREE, typical of calcalkaline-arc plutons (Appendix 3, Figure 40b–d). The Ham Lake pluton and the Thunderhill Lake porphyry have weakly and strongly fractionated HREE, respectively. The two sets of patterns of the late successor-arc intrusions (calcalkaline and high La/Yb) are similar to those of the two types of Missi Group rhyolite varieties and probably developed from the same mantle source and crustal sources above a subduction zone in the same tectonic environment as the volcanic rocks (Appendix 3, Figure 41a–d).

5.9.5.1 Nelson Bay dome (unit 15a), Norris Lake (unit 15b) and Barron Lake (unit 15c) plutons

The Nelson Bay dome is directly connected to a thin but extensive unit of felsic gneiss on File Lake that is interpreted as ignimbrite (Bailes, 1980a). Two phases were sampled in both the pluton and the felsic volcanic gneiss on File Lake. These phases are chemically like Cleunion rhyolite and Puffy Lake dacite in major elements (Appendix 4, Table 10, DVD) and, where analyzed, also in trace elements (Appendix 2, Figure 24c, d; Appendix 3, Figure 38e, 41a, c). The chemical similarity and field relations indicate that the Nelson Bay dome is probably a resurgent dome and the surrounding thin felsic volcanic units are extra-caldera ignimbrites. The pluton, dated at 1832 ±2 Ma (David et al., 1996), is slightly younger than 1836 ±2 Ma Missi Group rhyolite on Wekusko Lake (Ansdell et al., 1999) and significantly younger than reworked (?) 1848 ±4 Ma dacite tuff at the base of the Missi Group near Puffy Lake (Machado et al., 1999). The chamber containing the Nelson Bay dome may have erupted felsic tuff and caused pluton emplacement during protracted Missi-type magmatism. The Nelson Bay dome and Ham Lake pluton (Section 5.9.5.2) have the same juvenile ε_{Nd} values as the volcanic rocks of the Missi Group (Appendix 2, Figure 29f).

The undated Norris Lake and Barron Lake plutons have a geochemical fingerprint very similar to the Nelson Bay dome and are also interpreted as Missi age (Appendix 3, Figure 40b, c, 41c). The Barron Lake pluton intrudes both the Ham Lake pluton and the Burntwood Group, providing a relationship consistent with this age.

5.9.5.2 Ham Lake pluton (unit 15d)

The large, elongate Ham Lake pluton is interpreted to cut the Loonhead Lake Fault (Connors, 1996) and thus, with its sheared margin, belongs to the early synkinematic intrusions that probably represent the end stage of arc magmatism in the THO. It comprises variably foliated hornblende-biotite granodiorite (Bailes, 1980a). Its most strongly metamorphosed and gneissic northern portion contains minor garnet and sillimanite, but other samples are metaluminous and therefore included with the latest volcanic-arc plutons (Appendix 1, Figure 23j).

However, as with its age (ca. 1830 Ma) and structural habit, the Ham Lake pluton has a composition that is transitional between the more typical Missi-age plutons and the younger synkinematic intrusions. Its content and fractionation of REE lie between these intrusive suites, resembling most closely the intermediate members of the ca. 1830 Ma Touchbourne intrusive suite (Appendix 2, Figure 24d), with a moderately steep slope of REE, including depletion of HREE (compare Appendix 3, Figure 40b, f). Unlike most of the granitoid intrusions, it lacks Sr and Eu anomalies (Appendix 3, Figure 41d); that is, it shows no evidence of feldspar fractionation.

5.9.5.3 Thunderhill Lake porphyry (unit 15e)

A body of rhyolitic quartz porphyry, more than 1000 m wide, occurs northeast of Weldon Bay. The porphyry has a trace-element chemistry similar to that of Lobstick rhyolite, including Sr (>250 ppm; Appendix 4, Table 10, DVD) and depleted HREE (compare Appendix 3, Figures 40c and 37f), although LREE are slightly less fractionated. Like most Lobstick rhyolite, it shows strong negative Nb and Ti anomalies with only small Sr and Eu anomalies (Appendix 3, Figure 41b). The porphyry is interpreted as a subvolcanic intrusion. It yielded a U-Pb (ID-TIMS) age of 1826 +11/–5 Ma (Hunt and Schledewitz, 1992), within error of the age of the Nelson Bay dome and Ham Lake pluton.

Other large bodies of fine-grained felsic rock, which occur northeast and northwest of Weldon Bay, may also be subvolcanic domes or caldera. The occurrence of these large bodies at Cleunion Lake, Thunderhill Lake and Nelson Bay indicate the presence of three felsic volcanic centres in the Kississing–File lakes area. Another large felsic centre probably includes the Chickadee rhyolite and Lost Frog Lake granite, farther east on Wekusko Lake (Gordon and Lemkow, 1987). The combined volume of felsic tuff, reworked tuff in the Missi Group metasandstone, and coeval high-level intrusions and deeper seated plutons must have been considerable.

5.9.5.4 Reed Lake quartz diorite (unit 15g)

The Reed Lake 'pluton' or complex is partly Missi in age and forms a large dome southeast of the Kississing–File lakes area. At the northern margin of the pluton, where a relatively young phase intrudes Fussey Lake basalt south of File Lake, a sample of ferrodiorite is an Fe-rich tholeiite (Appendix 1, Figure 21e) with a very low Mg# (18). The trace-element signature of the quartz diorite is like those of some Missi Group volcanic rocks and is most comparable to Evans Lake basalt (Appendix 2, Figure 24c, d, 30e, f; Appendix 3, Figure 39b, d), thus supporting a similar origin consistent with its Missi age of about 1.84 Ga.

5.9.5.5 East Big Island gabbro (unit 15h)

Elongate bodies of uniform gabbro recrystallized to medium-grained foliated amphibolite occur in the Missi Group east of Big Island on Kississing Lake and may form a comagmatic suite with smaller dikes or sills and flows elsewhere in the Missi Group. Some highly foliated bodies are indistinguishable from Missi Group basalt in the field.

One sample of relatively massive gabbro has a major-element composition very similar to some Fe-rich flows of the Kississing River basalt (compare Appendix 1, Figure 21d and e; Appendix 2, Figure 27e and f). However, East Big Island gabbro has more fractionated LREE like Evans Lake basalt (Appendix 2, Figure 24c; compare Appendix 3, Figures 38a and 40a). The chemistry of the analyzed sample is supporting evidence that the gabbro is consanguineous with the Missi Group flows and tuff.

5.9.6 Missi-age magmatism: discussion

The complex geochemical signature of the Missi group volcanic rocks indicates that multiple sources contributed to their origin in a rapidly changing tectonic environment. In general, Missi-age magmatism is distinguished from the early arc–back-arc volcanism and early successor-arc magmatism by its higher content of LILE, Th and LREE, geochemical characteristics that suggest a deeper source of melting and a lower degree of melting (Woodhead and Johnson, 1993; Churikova at al., 2001; Nyström et al., 2003) for all mafic–intermediate units except the Kississing River basalt and related gabbro. A greater degree of sediment participation in subduction has a similar effect (Jolly et al., 2001; Gertisser and Keller, 2003; Clift et al., 2005), and the high MgO content at relatively high SiO_2 of Evans Lake basalt favours this origin. The different processes, however, are complementary, not mutually exclusive.

The change to the more complex magma-generation environment is interpreted to be the result of crustal thickening preceding the volcanism and leading to the widespread subaerial or continental environment that prevailed in the THO at that time. The crustal thickening is interpreted to be related to the extensive early successor-arc plutonism (Whalen et al., 1999) and possibly early collision with the Sask craton (Zwanzig, 1999). A high regional volume of siliciclastic sedimentary rocks (e.g., Burntwood Group) became available for subduction during the onset of collision tectonics shortly before the Missi Group magmatism. The high contents of LILE and LREE relative to HFSE indicate that magmatism still occurred by fluid-fluxed melting of the mantle wedge above an active subduction zone or rising asthenosphere into the previously contaminated mantle wedge. Continental-type arc magmatism is consistent with the higher Nb/Yb and Zr/Y ratios and elevated Zr contents of the Missi-age magmas. An increase in subducted sediments is consistent with the raised LILE, Th and LREE, although the sediment must have been nearly juvenile, like the Burntwood Group, to retain the high ε_{Nd}^{t} values in the Missi-age volcanic rocks and plutons.

The Burntwood Group is not as enriched in LREE as some of the volcanic-arc rocks. Accordingly, other processes may have taken place, such as earlier phlogopite and amphibole crystallization in the mantle wedge and their subsequent melting during continued heating or asthenosphere injection. The progressive increase in MREE to LREE, Hf, Zr and Th at constant Yb for Jungle Lake basaltic andesite from the high-Mg Evans Lake basalt (Appendix 3, Figure 39c) suggests mixing of melts from the mantle and crust. The dacite and rhyolite are probably related to Evans Lake basalt by magma mixing and fractionation of plagioclase to create the Sr and Eu anomalies (Appendix 3, Figure 41a). Regardless of the petrogenetic details, the subaerial volcanism was not conducive to VMS deposition, but played an important role in the heating and fluid flux that led to subsequent orogenic gold deposition.

Although stratigraphic successions are complicated by block faulting, thrusting and folding, there still appears to be a recognizable evolution trend from the mafic and intermediate calcalkaline magmatism to mafic and felsic bimodal magmatism. Tholeiitic basalt (Kississing River basalt and the similar Ferro Mine basalt) occur in thick successions with other volcanic rocks of the Missi Group in isolated fault blocks and in thin units that are intercalated with the other volcanic rocks. The tholeiite occurs partly up the succession and near the top. The evolution from medium- and high-K calcalkaline lava to more mafic, lower K tholeiite of the Missi Group is analogous to that on Santorini, in the Aegean arc (Huijsmans et al., 1988). Much of the Aegean is a region of extension, with crust and lithosphere thinning most in the central part of the arc beneath Santorini, where rising asthenosphere has contributed to the magmatism (Mitropoulos and Tarney, 1992). The TiO_2 content, relatively elevated compared to most mafic and intermediate arc-volcanic rocks (Appendix 1, Figure 22d), is also consistent with moderate crustal extension in a continental arc (e.g., Nyström et al., 2003).

The elevated Nb/Yb and slightly elevated Ti in Kississing River basalt are consistent with a weakly enriched mantle source, although MREE to HREE have a flat MORB-like pattern (Appendix 3, Figure 39a). The slab was probably at considerable depth beneath the Missi-age volcanic arc, as in the Aegean island arc and the back of the New Britain island arc (Mitropoulos and Tarney, 1992; Woodhead and Johnson, 1993). A variety of processes can take place within the high column of melting, producing the range of compositions from Kississing River low-K tholeiitic basalt to Evans Lake medium-K calcalkaline basalt and Jungle Lake basaltic andesite to felsic tuff. These processes would include the dewatering and melting of subducted sediments that became available in abundance during the onset of continental collision, a feature of other collisional orogens (e.g., Brown and Spadea, 1999). The influence of subducted sediments or melted crust on the composition of the Missi-age calcalkaline magmas is consistent with the wide range of Th/La and La/Yb ratios among Evans Lake high-Mg basalt, Jungle Lake basaltic andesite and Puffy Lake dacite. It is noteworthy that abundant juvenile sediments became available at a subduction zone facing the Flin Flon Domain only during the Missi-age collisional stage.

The Missi-age magmatism is bimodal, unlike most modern evolved or continental-arc volcanism but consistent with the proposed episodic extensional arc environment following crustal thickening. When all volcanic, intrusive and related sedimentary products are considered, the magmatism was predominantly felsic. At least three major felsic volcanic centres with domes (Cleunion Lake–Fay Lake) and possible resurgent caldera (Nelson Bay) or associated plutons (Wekusko Lake) occurred at 40–60 km intervals. The chemically identical thin units between these centres would have extended over the caldera rims. The inferred volume of felsic magma was large, possibly similar to the 'ignimbrite flare-up' that occurred on much larger scale in western North America and the Andes during Tertiary and Cretaceous intervals of fast seafloor spreading and increased arc magmatism, in some cases associated with or followed by slab detachment (Mpodozis and Allmendinger, 1993; Ferrari et al., 2002, Nyström et al., 2003; Ducea and Barton, 2007). Slab rollback after arc-continent collision in the THO (Ansdell et al., 1995; Zwanzig, 1999) is a likely cause of the Missi-age extensional magmatism that is dated from the 1836 Ma Chickadee rhyolite at Wekusko Lake and the 1832 Ma dome at Nelson Bay.

The wide distribution of rhyolite and the associated mafic–intermediate rocks recognized in this report indicate that the processes of asthenosphere upwelling associated with slab rollback occurred along the entire margin of the Flin Flon–Glennie Complex. The strong crustal signature and HREE depletion, yet high ε_{Nd}, of the Lobstick-type rhyolite and the Thunderhill Lake porphyry suggest melting at the base of the previously thickened juvenile-arc crust or a substantial input of juvenile sediment. The high heat flow required in this process was probably provided by tholeiitic magma, either ponded at the base of the crust or in crustal slices dragged down the subduction zone. The final compositions of the Kississing River and Ferro Mine basalt units were produced by fractionation. Similar mafic melts may have mixed with felsic melts derived from subducted sediments and/or early juvenile-arc crust, producing the observed range in lithophile crustal-element contents at relatively high LILE contents (Appendix 3, Figure 41a).

5.10 Synkinematic intrusions

The time of change from arc magmatism to postcollisional magmatism is poorly constrained by the field relations and geochemistry of the plutons in the Kississing–File lakes area. In this report, we tentatively consider the 1830 Ma Touchbourne intrusive suite (Manitoba Energy and Mines, 1986; Gordon et al., 1990), Ham Lake pluton and related intrusions to mark this change in tectonic environment. Members of the Touchbourne intrusive suite (TIS), which occur mainly in the Central Kisseynew subdomain, comprise 'enderbite' (quartz diorite with associated granodiorite containing igneous or metamorphic hypersthene) and pyroxene-rich mafic–ultramafic rocks interpreted as cumulates. Intrusions of similar composition but containing metamorphic amphibole rather than pyroxene occur west of Kississing Lake (Duval Lake salient of the File River nappe of Zwanzig, 1999). They include the Duval Lake sills. The 1842 Ma age of the Boundary intrusions (Heaman et al., 1992), which are compositionally similar to members of the TIS, suggests that the 1.84–1.83 Ga period may have involved overlapping subduction-related and postcollisional magmatism or non-arc magmatism, consistent with the geochemistry of these rocks (Section 5.10.1.1). Most likely, the period represents a slow retreat of magmatism from the Flin Flon Domain into the Kisseynew Domain. A north-dipping slab may have flattened (Hollings and Ansdell, 2002; Ansdell, 2005). However, it is more probable that such an event, if it happened, would have taken place later, during the 1823 to1814 Ma magmatic gap (Figure 14).

The 1830 Ma Ham Lake pluton intruded the early Loonhead Lake Fault east of File Lake (Connors et al., 1999), probably during a period of compressional tectonics. Its REE characteristics are transitional between the latest arc plutons on the margin of the Flin Flon Domain and the synkinematic intrusions at Highrock and Flatrock lakes, farther north in the Kisseynew Domain (Appendix 2, Figure 24d). In the Kississing–File lakes area, the Dow Lake quartz diorite, which is interpreted to have a similar or younger age, marks the beginning of extensive magma contamination by Archean crust that was probably underthrust during collision. Because these intrusions postdate the early post-Burntwood faulting, the onset of the Trans-Hudson orogeny and collision tectonics are interpreted to have occurred earlier (Zwanzig, 1999). A later resumption of arc magmatism is not strongly supported by the geochemistry or tectonic history.

Nearly all intrusions that are interpreted as younger than 1830 Ma are felsic and weakly peraluminous, and form the prominent high SiO_2 peak in Appendix 1, Figure 19f. These rocks are probably related to collision tectonics and possible slab foundering. The Highrock Lake Batholith and Flatrock Lake granodiorite (units 10 and 11 in Baldwin et al., 1979) intrude the TIS (unit 9 in Baldwin et al., 1979). They are herein interpreted to represent the younger, postcollisional magmatism.

These intrusive suites have no known mafic members and thus provide no direct evidence of arc magmatism that involved melting of an underlying mantle wedge. Their postcollisional origin is consistent with the U-Pb zircon age of ca. 1816–1814 Ma for granite and granodiorite in the Burntwood Lake area (Gordon et al., 1990) and 1823 Ma for a tonalite dike southwest of Burntwood Lake (Machado, et al., 1999). Their relatively young age is consistent with the synkinematic characteristic of foliated granite sheets concentrated in major shear zones (Zwanzig, 1990). Hollings and Ansdell (2002) considered the batholith and related granite in the Kisseynew Domain to be volcanic-arc plutons. Trace-element geochemistry, however, fails to distinguish volcanic-arc granites from postcollisional granitoid rocks (Pearce et al., 1984). It has been pointed out (e.g., Barnes et al, 2002) that granitoid rocks often reflect the geochemistry of their crustal protoliths rather than their tectonic environment. Thus, the arc-like compositions and relatively low Rb content of the late felsic intrusive rocks (Appendix 2, Figure 25e–g) are interpreted to represent the early Trans-Hudson crust.

Syn- to late-kinematic minor intrusions of pegmatite and leucocratic granite–tonalite have a wide range of trace-element patterns, including some with flat but much depleted MREE to HREE patterns and those with steep slopes and HREE depletion (Appendix 3, Figure 42). These reflect fractionation as well as differences in crustal and mantle sources, and more than one type of partial melting. A set of late-kinematic mafic tholeiitic dikes that were still affected by high-grade metamorphism indicates at least a minor involvement of postcollisional (non-arc) melts from the mantle. If these ubiquitously Fe-rich dikes (Weldon Bay gabbro) represent fractionation of widespread mafic underplating, they may be a key to the high heat flow and granite production that characterized the late evolution of the Kisseynew Domain.

5.10.1 Intermediate–mafic intrusions (unit 16)

Intermediate–ultramafic intrusions that share some geochemical characteristics with the Missi-age intrusive rocks on the northern margin of the Flin Flon Domain are widely distributed in the Central Kisseynew subdomain and occur locally on the South flank of the Kisseynew Domain. The elongate bodies in the Central Kisseynew subdomain range from hypersthene-rich ultramafic rock to enderbite that typically includes igneous breccia phases. They are known from the Burntwood Lake area, directly northeast of the Kississing–File lakes map area, as the TIS, which also includes abundant garnet-bearing retrograde equivalents (Baldwin et al., 1979; Manitoba Energy and Mines, 1986). Geochemical data for these rocks collected in the Burntwood Lake area were added to our dataset from Hollings and Ansdell (2002) and their geochemistry compared to similar looking rocks (but without hypersthene) in the Kississing–File lakes area. These intrusions and related felsic rocks are ca. 1830 Ma (Gordon et al., 1990). Slightly more potassic rocks than enderbite (opdalite) from the deepest level of the Kisseynew Domain in Saskatchewan have yielded the more precise age of 1830 ±2 Ma (Ashton et al., 1999). These ultramafic to felsic intrusions are chemically similar to the (older) Boundary intrusions, which have been interpreted to mark the end of arc magmatism in the vicinity of Flin Flon (O'Hanley and Kyser, 1994).

Sheet-like intrusions on the South flank of the Kisseynew Domain (Duval Lake sills) at the west end of the project area have an overlapping composition with the TIS. These sheets occur in the uppermost, farthest overriding part of the File River nappe on the South flank of the Kisseynew Domain. The Duval Lake sills have the same intrusive habit and locally abundant cognate xenoliths and igneous breccia with ultramafic–felsic phases as the Touchbourne intrusive suite (TIS), features that suggest that the sills share the same origin. A large body northwest of Duval Lake resembles the TIS but contains amphibole instead of pyroxene or retrograde garnet where it extends south into the map area. It may be less highly metamorphosed than the enderbite to the north (Zwanzig, 1999) or intruded at a higher structural level if the hypersthene had an igneous origin. These bodies are medium- to high-K calcalkaline rocks and had no Fe enrichment during fractionation (Appendix 1, Figures 20h, 21f). They plot from the calcalkaline into the transitional field in Th/Yb versus Zr/Y space (Appendix 2, Figure 27g).

Dow Lake quartz diorite is highly Fe enriched (Appendix 1, Figure 21f) but plots with the calcalkaline rocks in Th/Yb versus Zr/Y space (Appendix 2, Figure 27g).

5.10.1.1 Touchbourne intrusive suite (unit 16a)

Hypersthene-bearing sills or intrusive sheets mapped in the Central Kisseynew subdomain by Baldwin et al. (1979) range from 10 m to 3 km wide and up to 40 km long. They appear to form a bimodal suite dominated by a quartz diorite–tonalite composition with a lesser mafic to ultramafic phase. Rare, particularly well preserved, thick sills have ultramafic rock near one contact, evidence of *in situ* fractionation and evidence of mafic replenishment at the base of the magma chamber. The most abundant phase is foliated, brown-weathering metatonalite–diorite containing hypersthene that is commonly retrogressed to biotite and garnet. Fresh hypersthene occurs as elongate anhedral clusters and skeletal aggregates that are in optical continuity. This is interpreted as a metamorphic texture, but recrystallization of igneous hypersthene cannot be ruled out. Sills intruded the Burntwood Group, which was later migmatized, and they contain partly assimilated inclusions of the greywacke host and other rocks. The more mafic phases are olivine normative; all mafic phases have clinopyroxene in the norm and the most felsic phases are quartz normative. All samples have hypersthene, plagioclase and orthoclase (2–13%) in the norm.

Hypersthene also occurs locally in the adjacent hostrock, possibly as the result of contact metamorphism. The high abundance of adjacent leucosome and diatexite (very coarse grained structureless migmatite) may indicate contact migmatization and hybridization of the sills (compare Barnes et al., 2002). Generally, the sills are cut by abundant dikes of pegmatite and granite; some sills are agmatite, dominated by younger leucogranitoid rock. Hypersthene in these bodies is altered to garnet +biotite±talc±amphibole.

The Touchbourne intrusive suite (TIS) shows a calcalkaline to transitional trend (Appendix 2, Figure 27g). Ratios of K_2O/Na_2O, $[La/Yb]_N$ and Nb/Ta for the most felsic compositions are similar to typical Proterozoic calcalkaline granitoid rocks, but Sr/Y is elevated like TTG (Table 16). The suite shows a range of LREE fractionation ($[La/Yb]_N$ of 6.1–16) at relatively constant Yb (~1.3 ppm; Appendix 2, Figure 24d). The more felsic rocks also show increased LILE (Appendix 2, Figure 28d) but at relatively elevated MgO, Ni and Cr, suggesting a melt component from a strongly contaminated mantle wedge (Table 16). The $[La/Yb]_N$ ratio shows a positive covariance with Zr, Nb and Th (Appendix 4, Table 10, DVD), which are abundant in the crust; this suggests mixing with a crustal melt component. The abundant rounded felsic–intermediate inclusions in the more mafic phases may indicate crustal assimilation, but the largest sills are generally free of visible inclusions.

The mafic–ultramafic rocks of the suite do not plot in the field of arc-volcanic rocks in Appendix 2, Figure 30f, but have a similar position in Th-Zr-Nb space to the Ponton Lake (E-MORB–type) BABB (compare Appendix 2, Figure 30a and f). The La-Y-Nb plot suggests a calcalkaline-arc affinity (Appendix 2, Figure 31f), which may be true only if Th was significantly depleted in some sample sites during metamorphism. If Th was mobile, however, it was more likely enriched in the enderbite because surrounding and crosscutting granitoid rocks have high Th contents. Crystal fractionation probably affected the compositions too much for these diagrams to be useful in suggesting a tectonic origin.

The distinctive shapes of REE plots are considered to better reflect liquid compositions. Two types of pattern are present: 1) moderate La/Yb (Appendix 3, Figure 40e) of mainly mafic–ultramafic members, and 2) high La/Yb of intermediate members (Appendix 3, Figure 40f). The two types show, respectively, 1) a sinusoidal pattern for the high-Mg rocks, and 2) fractionation of middle REE and LREE, giving a concave-upward pattern for HREE in the intermediate–felsic rocks. The sinusoidal pattern of REE in type 1 (moderate La/Yb) is similar to that of water-deficient pyroxene-bearing plutonic rocks in the northeastern Superior Province (Percival and Mortensen, 2002). The elevated middle REE content and relatively high SiO_2 at high MgO (calcalkaline trend) may also be due to amphibole accumulation in the sills. The higher La/Yb trend (type 2), due to increased fractionation of LREE (more steeply negative slopes) of the two groups of samples, is the result of processes other than low-pressure crystal fractionation. An increase in REE with mafic content (e.g., Yb versus MgO, Appendix 3, Figure 40g) among the intermediate intrusive members may indicate mixing of an enriched mafic component and a depleted intermediate or felsic component. One explanation of these element variations is that there was more than one parent magma, the first from a deeper enriched mantle source and the second from a component of melted crust depleted in HREE, such as an arc massif. Strong negative HFSE anomalies in the MORB-normalized plots for the entire suite (Appendix 3, Figure 39e, f) may indicate ongoing subduction at 1830 Ma, as suggested by Hollings and Ansdell (2002), but the interpretation preferred herein is a non-arc process of dehydration and melting of enriched mantle that had a long previous history of subduction and lower crustal heating by rising asthenosphere (Pearce et al., 1984). Mixing with crustal melt or contamination is suggested by the abundance of felsic inclusions and igneous breccia with ultramafic–felsic components. This is also consistent with the origin suggested for the Duval Lake sills (next section).

5.10.1.2 Duval Lake sills (unit 16b)

Intrusive sheets containing amphibole as principal mafic mineral and occurring in weakly migmatitic Burntwood Group greywacke (some containing staurolite±sillimanite) have intrusive habit and geochemistry similar to those of the TIS. They are bimodal, and magma mixing or contamination is suggested by a gradational hybrid contact zone in one of the sills. Assimilation of mafic inclusions and Burntwood Group greywacke xenoliths is evident in the quartz diorite phase. The other (northern) sill appears to be differentiated mafic diorite–quartz diorite. It has a high Mg-Ni-Cr diorite phase that may be a cumulate. However, most rock types occupy discrete bodies. The sills may be lower grade equivalents of the TIS, as suggested by their structural position in the File River nappe (Zwanzig, 1999) and by the same calcalkaline trend and immobile element ratios as the TIS (Appendix 1, Figure 21f; Appendix 2, Figure 27g). No ultramafic rocks occur in the sampled Duval Lake sills, but an ultramafic dike altered to amphibolite at Jungle Lake has a major-element composition nearly identical to that of the pyroxene-rich members of the TIS. The most felsic rock southeast of Duval Lake has a low Yb content and high $[La/Yb]_N$ and Sr/Y ratios, similar to a ca. 1823 Ma tonalite sheet (Burntwood Lake tonalite) in the Central Kisseynew subdomain (Appendix 2, Figures 24d, 28h). The most leucocratic intrusive phase (67% SiO_2) has relatively high contents of Ni and Cr (40 and 99 ppm), similar to the most felsic analyzed phase in the TIS. These rocks also have high total alkali contents, plotting close to the boundary with the field of alkaline rocks (Appendix 1, Figure 20d) and consistent with a low degree of melting of deep fertile mantle. The high Sr/Y ratio (71) is consistent with a melt component from mafic crust (Section 5.10.3.2).

Although the Duval Lake sills also plot in the arc-volcanic field on tectonic discriminant diagrams (Appendix 2, Figures 30f, 31f), they show the same two sets of patterns as the TIS on MORB-normalized multi-element plots (Appendix 3, Figure 39e, f) and probably have a similar complex origin. The two types of REE pattern are 1) sinusoidal with weakly depleted HREE, and 2) more fractionated LREE and moderately depleted HREE with a concave-upward curve (Appendix 3, Figure 40e, f), caused by hornblende stability at the site of melting (Defant et al., 1991) or hornblende removal during fractional crystallization (Richards and Kerrich, 2007). The origin of the two types may be explained by melting of enriched mantle and hornblende±garnet–bearing, deep lower crust to produce the intermediate rocks. A probable association with magmatic hornblende in the original mineralogy of the Touchbourne Lake 'enderbite' supports the conclusion that its orthopyroxene formed during high-grade metamorphism from an amphibole-bearing rock such as the Duval Lake sills.

5.10.1.3 Dow Lake quartz diorite (unit 16c)

Intermediate–mafic intrusive rock (Dow Lake quartz diorite) forms large folded sheets on the west flank of the flattened dome at Dow Lake and extends across the syncline south of Martell Lake (Maps GR2010-1-1 and -2, back pocket/DVD).

Major early shear zones occur at the inferred top of the quartz diorite complex where it contains long narrow rafts of the Burntwood Group metagreywacke (column 8B in Figure 8c, back pocket). The field relationships, including local differentiation to tonalite, indicate that the intrusions were probably emplaced in early subhorizontal fault zones (D_3 thrusts) after juxtaposition of the volcanic assemblages with the Burntwood Group. The intrusions are sheared, highly metamorphosed, folded and cut by late-kinematic pegmatite. They are synkinematic intrusions that are postcollisional.

The mineral content and geochemistry of the quartz diorite are distinctive. It generally contains garnet, which can be relatively abundant and occur with biotite and significant amounts of accessory minerals, including magnetite, titanite and apatite. The mineralogy is reflected in the high contents of FeO, TiO_2 and P_2O_5 (Table 11). The unit is ferroan quartz diorite–diorite, a highly fractionated tholeiite (Appendix 1, Figure 21f) with low Mg# (21–34). Strong fractionation in a deeper chamber or a low degree of melting is indicated by the high REE content of the entire suite. The trend of increasing La/Yb occurs at nearly constant Yb (Appendix 2, Figure 24d) but with decreasing MgO (Appendix 3, Figure 40g) and increasing SiO_2. Ytterbium is constant as MgO varies. This suggests that the increase in LREE is not related to fractionation but rather to mixing of a mafic component from the mantle with a high–La/Yb felsic component from melting mafic crust. High Th contents place Dow Lake quartz diorite into the calcalkaline-arc field with the Duval Lake sills on Appendix 2, Figures 27g and 30f. Its extended-element pattern is shaped like that of Evans Lake basalt but with three to four times the trace-element concentration (Appendix 3, Figure 39d). Contents of Zr and Hf, however, show no arc-like depletion but are three to five times MORB. Similar geochemical characteristics occur in the mafic–ultramafic rocks with crustal contamination and intruded in a non-arc environment in the TNB (Zwanzig, 2005a). High contents of Th and U (Appendix 4, Table 10, DVD) are features of contamination by Archean crust (cf. Ragged Lake pluton and see below). Deep negative Nb anomalies are a general crustal feature not necessarily related to arc magmatism (Appendix 3, Figure 41).

Dow Lake quartz diorite has the lowest ε_{Nd} value (–5.6) of the tested igneous units in the Kississing–File lakes area. This distinguishes it from the intrusions of known and assumed Missi age. The quartz diorite is interpreted to contain mantle melt because of its fractionated intermediate–mafic composition, but it is contaminated with Archean crust that apparently provided at least part of the felsic component during magma mixing. The old crustal source may be a structural remnant that was already present at the margin of the Amisk Collage during the early successor-arc magmatism (cf. Ragged Lake pluton). Alternatively, the Archean component may be from the Sask craton, which had not been present during all previous juvenile magmatic episodes in the Dow Lake area but had been underthrust during collision with the Flin Flon–Glennie Complex. The Nd-isotope data and field relations show that the quartz diorite is not comagmatic with any Missi-type units. The similar trace-element geochemical pattern is probably inherited from the mantle wedge, which was strongly metasomatized and injected before and during Missi-age magmatism. The crustal contamination also contributed to the apparent arc-like geochemical signature of Dow Lake quartz diorite and related intrusions. Thus, the unit and related rocks are interpreted to be postcollisional, as are the younger felsic intrusions (below).

5.10.2 Late mafic intrusions, calcsilicate gneiss (unit 17)

This suite of rocks, which has been studied and sampled only in reconnaissance, requires further work to assign a relative age and origin. A simple interpretation is prohibited by the highly strained and altered state of some of its members. Locally better preserved diabase, gabbro and pyroxenite suggest a late intrusive origin (see, however unit 5b, Section 5.7.5.4).

5.10.2.1 Weldon Bay gabbro (unit 17a)

Sheets of mafic intrusive rocks, 1–2 m thick, are interpreted as synkinematic. Some of these were intruded into the Loonhead Lake thrust fault between the Burntwood Group and the Josland Lake gabbro at Weldon Bay; others into the Missi Group conglomerate at Evans Lake, slightly higher in the thrust sheet; and one or more into the Burntwood Group at Burntwood Lake.[12] Their common fine grain size, sharp contacts in a fault zone and geochemistry distinguish them from older units. Despite their fresh appearance, they have undergone the 1812–1805 Ma thermal peak. The sheets have been sampled on Weldon Bay (Lobstick Narrows) and on Burntwood Lake (Central Kisseynew subdomain).

The sheets represent part of the small mafic component of the otherwise granitoid synkinematic intrusions (Appendix 1, Figure 19f). They are relatively Fe- and Ti-rich tholeiite (Appendix 1, Figures 21f, 22d) that plot into the field of E-MORB on a tectonic discrimination diagram (Appendix 2, Figure 30 f). They form a smooth, negatively sloping, N-MORB–normalized plot with few or no HFSE anomalies or Th spike, similar to modern E-MORB or continental tholeiite (Appendix 2, Figure 31f; Appendix 3, Figure 39d).

The sheets are interpreted to represent postcollisional mafic magma that has found its way into the upper crust along major fault zones at the boundary of the Kisseynew Domain and within that domain. The high Fe and Ti contents at 7–8% MgO indicate tholeiitic fractionation of high-Mg parent magma. Its similarity with older suites suggests the presence of a long-lived, enriched mantle reservoir beneath the evolving Kississing–File lakes area.

5.10.2.2 Cree Lake intrusions (units 17a, b) and high-Ti amphibolite, calcsilicate (unit 17 undivided, unit 17c)

In and around the Sherridon structure, tholeiitic to transitional gabbro (unit 17a) and presumed cumulate pyroxenite (unit 17b), along with the high-Ti amphibolite (unit 17 undivided) and a large body of calcsilicate rock (unit 17c), are

[12] For locations of unit 17a, see locations 129, 162 and 663 on Map GR2010-1-2 (accompanying DVD).

tentatively interpreted to be cogenetic with the synkinematic (postcollisional) Weldon Bay gabbro (previous section; Appendix 1, Figure 21f; Appendix 2, Figure 27i).[13] These rocks plot in the E-MORB field in Th-Zr-Nb space (Appendix 2, Figure 30f) and the continental-basalt field in La-Y-Nb space (Appendix 2, Figure 31f). They have a unique geochemical signature in the Kississing–File lakes area, with uniformly fractionated REE (e.g., Appendix 3, Figure 35c). Their depletion of HREE is shared with the ca. 1830 Ma TIS and the Duval Lake sills. However, the Cree Lake intrusions are poorly sampled and the largest body near the core of the Sherridon structure is a heterogeneous calcsilicate to mafic gneiss that locally has high calcite contents. This rock is interpreted as highly carbonatized but has retained the trace-element signature of the recognizably intrusive members. The entire suite has a similar N-MORB–normalized multi-element pattern (Appendix 3, Figure 39d) in which negative HFSE anomalies are minor or lacking. The origin of the calcsilicate body is presently unknown. Its occurrence at the boundary between Sherridon gneiss and Bob gneiss suggests an age and origin similar to those of the Weldon bay gabbro.

5.10.3 Late granitoid intrusions

Granitoid rocks emplaced in large volumes in the Central Kisseynew subdomain extend from Highrock Lake into the Kississing–File lakes area (Figure 3), but these bodies are not adequately dated. Mapping of Baldwin et al. (1979) and U-Pb ID-TIMS dating of Gordon et al. (1990) have shown that the ca. 1830 Ma TIS is the oldest unit and may provide a maximum age for synkinematic intrusions. Some phases of the TIS may grade into the Highrock Lake Batholith. A problem exists, however, with contact migmatite: this generally includes granitoid phases that cut and surround the older units, making field relations difficult to interpret. A tonalite sheet (Burntwood Lake tonalite) has yielded a zircon crystallization age of 1823 ±4 Ma (Machado et al., 1999), and must postdate the TIS in the central part of the Kisseynew Domain, as does the 1814 +17/–11 Ma Flatrock Lake–type granite at Burntwood Lake (Gordon et al., 1990). The granite provides a minimum age for the Highrock Lake Batholith, which occupies the most central part of the Kisseynew Domain, and is cut by the Flatrock Lake granite in the south.

Pegmatitic granitoid rocks emplaced during protracted deformation have yielded U-Pb ages from monazite and zircon of ca. 1812–1789 Ma (Parent et al., 1999). Only the youngest pegmatite was not folded during a subsequent deformation phase (D_6). Small and large sheets of granite and leucogranite, up to 100 m thick and generally containing biotite and garnet or muscovite, form part of the large-scale migmatite layering in the Burntwood Group. They are foliated and affected by various stages of deformation. The sheets are abundant and were sampled near the base of the File River nappe southeast of Duval Lake, where they are interpreted as synkinematic. The sheets have been considered to represent the leucosome phases of the migmatite units in the Kisseynew Domain (Bailes, 1975; Baldwin et al., pers. comm., 1979; Zwanzig and Shwetz, 1994b; Zwanzig, 1999). In this report, they are considered to have a more complex origin involving melting of crust and possibly mantle. Such deep-seated magmas do not require active subduction but may contain a component of melt that has been derived, in part, from mantle modified during earlier subduction.

Data from Hollings and Ansdell (2002) were selected from archived samples collected by Baldwin et al. (1979) from the Highrock Lake Batholith (Zwanzig, 1990) and the Flatrock Lake granite (Figure 3). Although both units comprise several intrusive phases recognized in the field by colour, texture and estimated mineral composition, one intrusive phase was chosen from each unit. A tonalite phase has a major-element composition similar to (though less sodic than) the dated (1823 Ma) Burntwood Lake tonalite dike. This intrusive phase was compared to (younger) variably porphyritic granite–granodiorite that forms another important phase in the Highrock Lake–Flatrock Lake–Burntwood Lake area. The granite phase is very similar to some syntectonic dikes and sills whose field relationship is known to be postcollisional and synkinematic in the Kississing–File lakes area. Thus, field relationships based on newer mapping may be applied to the larger plutons that were analyzed from archived samples.

Granitoid rocks make up the overwhelming proportion of the synkinematic intrusions in the Kississing–File lakes area and the Central Kisseynew subdomain (Appendix 1, Figure 19f). They have a wide range of K_2O contents, are weakly peraluminous (>14% Al_2O_3; Al-saturation index of 1.0–1.2) and feature low Yb and high La/Yb (Appendix 1, Figures 20h, 22f, i; Appendix 2, Figure 24d). Epsilon-neodymium has the widest range among the various igneous units in the Kississing–File lakes area, indicating variable, generally limited but widespread involvement of older crust in the young melts (Appendix 2, Figure 29f). These characteristics, particularly the consistent depletion of HREE (Appendix 3, Figure 41e–g) and Al oversaturation distinguish the younger granitoid rocks from the Missi-age and older intrusions.

The geochemical characteristics of these young granitoid rocks are consistent with the conclusions of Hollings and Ansdell (2002) that they formed during continued subduction. However, their postcollisional age, widespread unfocussed distribution and sheet-like intrusive habit, as well as some geochemical features such as higher alumina saturation (ANK >1), require explanation. The late granitoid rocks are interpreted to represent postsubduction magmatism that was either related to the advanced collision and crustal thickening in the THO (Clarke et al., 2005; White, 2005) or was triggered by melting induced by mafic underplating and lithospheric extension (Kraus and Menard, 1997). A simple geochemical distinction between the two origins is not possible (Appendix 2, Figure 25h; Pearce et al., 1984), but the geochemistry can provide a preference between the competing models.

5.10.3.1 Burntwood Lake tonalite (unit 18)

The felsic sheet adjacent to Burntwood Lake, near the File River, that yielded the age of 1823 Ma is an Na-rich tonalite containing scattered garnet and no visible hypersthene, which

[13] For locations of unit 17a see locations 160 and 161, Map GR2010-1-2 (accompanying DVD).

distinguishes it from nearby enderbite. It has a high Sr/Yb ratio, similar to one Duval Lake sill with similar SiO_2 content (Appendix 2, Figure 28h). Its Sr/Y ratio of 73 also matches the 71 of this Duval Lake sill, suggesting a similar origin. The tonalite sheet has REE and extended-element patterns similar to the intermediate phase of the TIS and the Duval Lake sills (Appendix 3, Figures 39f, 40f) but a lower K_2O/Na_2O ratio. Although the Burntwood Lake tonalite is 170 km east of Duval Lake, all these bodies occur in the frontal part of the File River nappe (Zwanzig, 1999). A correlation suggests that some Duval Lake sills may be younger than the TIS, but similar magma of mixed crustal and mantle origin (Sections 5.10.3.2, 5.10.3.3) was widely generated in the Kisseynew Domain over a considerable period after subduction had slowed or ceased.

The Burntwood Lake tonalite has the characteristics of modern adakite-like rock produced by fractionation of amphibole (Richards and Kerrich, 2007) or, as recently redefined, Phanerozoic–Archean trondhjemite-tonalite-granodiorite (TTG; Condie, 2008; Table 16). This suggests that Burntwood Lake tonalite was derived by melting of garnet amphibolite in intermediate to lower crust. However, the extended trace-element pattern of these rocks is also similar to hydrous lower crustal melts such as tonalitic leucosome from southern Italy (Fornelli et al., 2002; Appendix 3, Figure 42c), as well as intermediate–felsic porphyry intrusions found in the continental arc in Chile (Appendix 3, Figure 41b). The latter are interpreted as a mix of subduction-related mantle melts and lower crustal melts (Richards, 2003). With ε_{Nd} of 1.6 and high Nd/Yb, the Burntwood Lake tonalite plots barely below the Flin Flon MORB-OIB mantle array on Appendix 2, Figure 29f. The melt source of a crustal component, if present, must have been juvenile.

5.10.3.2 Highrock Lake Batholith (unit 18a)

The main phase of the Highrock Lake Batholith is a light grey tonalite–granodiorite but with a characteristically higher colour index than the youngest of the granitoid rocks. Typically, the tonalite is moderately foliated and contains biotite and scattered garnet porphyroblasts 2–3 mm in diameter. The microtexture is metamorphic, showing strained plagioclase, amoeboid interstitial quartz and rare microcline preserved in myrmekite. The tonalite contains rounded cognate inclusions and locally abundant rafts of the Burntwood Group hostrock. The batholith includes several adjacent intrusions that have been flattened into large sheets, refolded and intruded by leucogranitoid units. Their contact zone is generally slightly more mafic than the core (i.e., they show the normal zoning of I-type batholiths). They are mantled or separated by sheeted injection complexes with abundant rafts of Burntwood Group migmatite.

Inclusions of retrogressed TIS enderbite indicate that the Highrock Lake Batholith is younger than ca. 1830 Ma. The 1823 ±4 Ma (Machado et al., 1999) Burntwood Lake dike, which has an REE pattern somewhat similar to that of the tonalite phase in the Highrock Lake Batholith, provides a possible age for the batholith. The 1814 +17/–11 Ma porphyritic granite (Gordon et al., 1990) that intrudes it provides a minimum age. Direct U-Pb zircon dating of the tonalite has yielded an errorchron age of 1841 +35/–15 Ma (Gordon et al., 1990). This is indistinguishable from the other constraining ages but less certain.

The main intrusive phase is medium-K calcalkaline tonalite–granodiorite, generally with only small amounts of interstitial K-feldspar. Hornblende is absent and the unit is slightly peraluminous. Average contents and ratios of selected elements from four representative samples are shown in Table 16 and on Appendix 1, Figures 20h, 21f and 23k. Although metamorphosed, the uniformity of the rocks and absence of veining over large areas suggest that the composition is characteristic of the original igneous rock. It is a calcalkaline granitoid but with moderately high Sr/Y typical of TTG suites. The bodies belong to the high-$[La/Yb]_N$, low-Yb intrusions, falling within the field of the synkinematic granite on Appendix 2, Figure 24d. Average $[La/Yb]_N$ is exceeded only in Missi Group rhyolite, Duval Lake sills, Archie Lake and Star Lake tonalites, and some younger synkinematic intrusions. The Highrock Lake Batholith plots in the arc field in Rb-Y+Nb space, but the upper boundary of that field is not defined for postcollisional granite. A subfield in Appendix 2, Figure 25h includes all the examples defined as postcollisional in the original presentation of the diagram by Pearce et al. (1984). This subfield encompasses most of the latest volcanic-arc and synkinematic granites in the Kississing–File lakes area, as well as the Highrock Lake Batholith and nearby granite in the Kisseynew Domain to the north. The field boundary of syncollisional granite, moreover, may not apply to early Precambrian rocks (Pearce et al., 1984). A significant addition of Rb-rich fluid is required to shift the composition of a melt derived from greywacke (e.g., the Burntwood Group) into the syncollisional field. The Trans-Hudson collisional environment, which was dominated by arc-volcanic rocks rather than continental rocks, would not have provided such fluid.

The REE profile of the Highrock Lake Batholith is steeply declined (Appendix 3, Figure 42a). A prominent negative Eu anomaly indicates plagioclase fractionation such that the parent magma must have been more basic than the batholith. This, and the low K_2O/Na_2O ratio (0.46–0.75), may be related to the fractionation of a mafic parent-magma component, whereas the overwhelming felsic composition with moderately elevated LILE (e.g., Appendix 2, Figure 28d, h) suggests significant melt component from crustal rocks. The ε_{Nd} of –0.6 indicates an Archean or earlier Paleoproterozoic component to the crustal melt.

The Highrock Lake Batholith shows strong negative anomalies of Nb and Ti on a primitive-mantle–normalized plot, a crustal signature that is typical of volcanic-arc granitoid rocks but also of all the late granitoid rocks in the Kississing–File lakes area (Appendix 3, Figure 41e–g). Although a volcanic-arc origin cannot be ruled out, that interpretation would contradict the synkinematic origin of granodiorite and granite sheets (e.g., southeast of Duval Lake), as well as the fact that leucosome generated in midcrustal greywacke–migmatite in southern Italy has geochemical characteristics similar to the Highrock Lake Batholith (Appendix 3, Figure 41f, g). The synkinematic granitoid rocks in the Kississing–File lakes area are interpreted to postdate the deformation (D_3) associated with the early collisional event in the THO. Consequently, the 'arc' interpretation of Hollings and Ansdell (2002) can be questioned. Alternatively, the Highrock Lake tonalite had the same postcollisional origin as similar Devonian tonalite–granite of southern Nova Scotia that shows evidence of magma mixing from crustal and mantle sources (Currie et al., 1998). These rocks have the same weakly peraluminous composition and extended-element pattern as the Highrock Lake Batholith and Flatrock Lake granite (Appendix 3, Figure 41e).

5.10.3.3 Flatrock Lake granite (unit 18b)

The other widespread granitoid suite in the Highrock–Burntwood lakes area occurs mainly in large, flat-lying or folded intrusive sheets and lenses that comprise a variety of biotite leucogranite–biotite tonalite phases with textures ranging from equigranular to porphyritic (microcline-phyric) and variably containing scattered or clustered garnet. A porphyritic variety has yielded a U-Pb zircon age of 1814 +17/–11 Ma (Gordon et al., 1990).

One compositional phase of these sheets (Flatrock Lake granite; data selected from Hollings and Ansdell, 2002) is chemically similar to some of the intrusive sheets in the Burntwood Group in the Kississing–File lakes area, comprising high-K, high-SiO_2 leucogranite–biotite granodiorite (Appendix 1, Figure 20h) with a K_2O/Na_2O ratio of 1.0–1.9. The phase has low Sr/Y, similar to average calcalkaline Proterozoic granitoid rocks (Table 16). Alumina oversaturation and high Rb and REE contents are similar to the Highrock Lake Batholith (Appendix 1, Figure 23k; Appendix 2, Figures 28d, 24d; Appendix 3, Figure 42a). The unit's chemistry also features deep troughs of Nb, Ti, Sr and Eu on an extended-element plot, as does the Highrock Lake Batholith (Appendix 3, Figure 41e). The data suggest that the granite was formed by fractionation of similar parent magmas but probably with lower proportions of mantle or mafic lower-crustal melt than melt of biotite-bearing felsic–intermediate crust. The higher K_2O content may be the result of a stronger influence of dehydration melting of micas, possibly from a greywacke-mudstone source such as the Burntwood Group. This origin would be consistent with the abundant screens of Burntwood Group in the granite and separating the sheets, but less so with the modest alumina saturation index (A/CNK of 1.1–1.2). The somewhat isotopically evolved nature of the Flatrock granite (ε_{Nd} of 0.0; Appendix 2, Figure 29f) is consistent also with remelting of the early successor-arc plutons and Burntwood Group sedimentary rocks. The rock has geochemistry very similar to granodiorite leucosome that was interpreted to have formed by biotite dehydration of metagreywacke in the lower crust of the Hercynian in southern Italy (Appendix 3, Figure 42a) and belonging to the low–Sr/Y calcalkaline rocks (Table 16).

A sheet of synkinematic two-mica leucogranite intruding well-preserved Burntwood Group gneiss southwest of Duval Lake has a major-element and REE composition very similar to that of Flatrock Lake granite (Table 16; Appendix 3, Figure 42a), particularly compared to a leucocratic sample (cf. samples 12-84-1199 and 25-2-7113 in Appendix 4, Table 10, DVD). The sheet has sharp contacts and clearly intruded the metagreywacke south of its transition to migmatite, showing that the granite was not generated *in situ* but formed deeper in the crust. Larger batches of similar melt intruded the Central Kisseynew subdomain to form the Flatrock Lake granite sheets.

5.10.4 Synkinematic granitoid sheets, leucosomes and pegmatite (units 19, 19a–c)

The youngest granitoid rocks form small- to moderate-sized intrusions and the felsic components of migmatite (leucosome), comprising leucotonalite to granodiorite and leucogranite, with textures ranging from fine grained to finely porphyritic and medium grained to pegmatitic. Most intrusive units and leucosomes are foliated and have granoblastic textures that show only relict primary grain size as mineral aggregates (Zwanzig and Schledewitz, 1992). They are dominated by leucogranitoid and pegmatitic sheets and lenses with biotite as the main mafic phase, some with additional muscovite and/or sillimanite and others containing scattered garnet or, in the north, garnet and/or cordierite. They exhibit a range of relative ages; some were affected by folding during different stages of deformation and others are planar and cut folds (Zwanzig, 1999). The sheets provide a minimum relative age (postcollisional, synkinematic) for stocks and batholiths of similar composition in the Central Kisseynew subdomain. Field relations of veins of leucosome and coeval larger sheets and lenses generally allow at least three relative ages to be distinguished locally: LS1, which is highly boudinaged leucotonalite; LS2, which comprises more coherent, highly leucocratic granodiorite to tonalite veins; and LS3, which ranges from fine-grained to pegmatitic sheets and lenses of tonalite to younger granite (possible LS4 leucosome).

The presence of these rocks as part of the migmatite of the Burntwood and Missi groups has been interpreted previously to indicate that the granitoid sheets are leucosomes formed from anatectic melts derived from these sedimentary rocks (Bailes and McRitchie, 1978). Southeast and southwest of Duval Lake, however, similar sheets separate screens of nonmigmatitic, locally staurolite-bearing metagreywacke of the Burntwood Group. Many sheets and lenses have relatively sharp contacts with the greywacke and can be interpreted as intrusive, rather than developed *in situ* (Photos 1, 2, 4, Appendix 4, Table 1, DVD). If these were anatectic melts, they must have originated from adjacent rocks or a deeper crustal level. An increase in the proportion of granitoid rock over paleosome north toward the Central Kisseynew subdomain, where garnet- and cordierite-rich diatexite is abundant, suggests that the granite accumulated in the upper part of the migmatite complex below a roof of paragneiss with more discrete granitoid sills. Another area with abundant granitoid sheets and lenses extends from Nokomis Lake north and northwest to Jungle Lake in a (D_5–D_6) mobile zone. These sheets and relatively older tonalite leucosome (LS1) have been dated at Jungle Lake, yielding U-Pb monazite and zircon ages that range from 1812 to 1789 Ma (Parent et al., 1999). Their monazite and rare zircon ages are younger than the poorly constrained zircon ages of the larger intrusions in the Central Kisseynew domain, but the latter have similar monazite ages that are interpreted as uplift and cooling ages (ca. 1817–1804 Ma; Gordon et al., 1990). Thus, the central intrusions directly preceded or overlapped generation of the sheets in the south. The younger ages of similar rocks in the Kississing–File lakes area suggest a southerly advance of the migmatite complex, probably during a prolonged period of crustal heating and deformation.

5.10.4.1 Synkinematic tonalite to granodiorite leucosomes (LS1, LS2)

The oldest and most strongly boudinaged and folded leucosome (LS1) common in the Burntwood Group in the Kississing–File lakes area is tonalitic and contains biotite+sillimanite±garnet (Figure 16a). This deformed leucosome has yielded a U-Pb monazite age of 1809 ±2 Ma, interpreted as the time of formation at Jungle Lake (Parent et al., 1999). Its generation probably started earlier with the advanced development of the main high-temperature S_5 foliation during M_2 metamorphism and D_5 deformation (Figure 14). A diffuse network of granodioritic leucosome in early successor-arc plutons is suggested to have also formed at that time, as indicated by ca. 1810 Ma zircon growth (Hunt and Zwanzig, 1990; Photos 42–44, Appendix 4, Table 1, DVD) No geochemical study was made of the LS1 leucosome, which may have formed purely by local fluid-present melting of quartz, plagioclase and minor K-feldspar. A slightly more coherent vein of tonalitic leucosome (LS2) has an REE plot identical to that of its adjacent Burntwood Group host and may have been generated *in situ*, with full entrainment of mafic and accessory grains (Photos 4, 7, Appendix 4, Table 1; Appendix 3, Figure 42b). The LS1 and LS2 leucosomes make up a total of <10 vol. % of the migmatite, similar to tonalitic leucosome generated by fluid-present melting of quartz and plagioclase in metagreywacke and pelite in midcrustal (5–10 kb) Hercynian migmatite of southern Italy (Fornelli et al., 2002) and Sicily (Cruciani et al., 2007).

5.10.4.2 Synkinematic granodiorite to tonalite sheets, LS3 leucosome (unit 19c)

The thicker synkinematic granitoid sheets and lenses in the Kississing–File lakes area are medium- to high-K calcalkaline rocks (Appendix 1, Figures 20h, 21f; Appendix 2, Figure 27i). They are weakly peraluminous tonalite–granodiorite (where locally analyzed) with A/CNK ranging from 1.0 to 1.2 and the majority of samples ~1.1 (Appendix 1, Figure 23k), making them slightly less aluminous than the Hercynian leucosomes associated with metagreywacke (Table 16). Most of our samples fall into the field of high–La/Yb, low-Yb granitoid rocks, overlapping the field of the tested Highrock Lake Batholith phase and including the ca. 1823 Ma Burntwood Lake tonalite and samples of relatively young (LS3) leucosome at Nokomis Lake (Appendix 2, Figure 24d). The granitoid sheets have moderate Y+Nb and Rb contents that plot within the field of volcanic-arc granite and postcollisional granite (Appendix 2, Figure 25h). Leucogranodiorite sheets with moderate La/Yb and negative Eu anomalies (Appendix 3, Figure 42b) are highly enriched in SiO_2 and depleted in mafic components, Ca, Sr, Zr and REE (synkinematic granodiorite in Appendix 4, Table 10, DVD). Their chemistry is consistent with advanced fractionation.

Locally crosscutting synkinematic granitoid sheets and leucotonalite–granodiorite leucosome (LS3) near the migmatitic front have high $[La/Yb]_N$ ratios, strong negative Nb anomalies and moderately depleted HREE with a concave-upward profile similar to the 1823 Ma tonalite sheet at Burntwood Lake (Appendix 3, Figure 42c). Granodiorite shows a strong negative Sr anomaly and is similar to Highrock Lake granodiorite. Tonalite has the positive Sr anomaly and high Sr/Y ratio with only moderate HREE depletion of adakite-type suites formed by hornblende-present fractionation, melting of mafic crust or fluid-present melting of quartz and plagioclase in the middle crust (Table 16). The field relationships suggest that many of the veins and larger sheets formed by intrusion during protracted metamorphism and deformation, probably with a component of melt from amphibolite with only minor garnet involvement. Mantle involvement is suggested by the association and similar trace-element geochemistry of an Mg-Ni-Cr–rich part of a mafic diorite sill in the Duval Lake suite with high LILE and REE (Appendix 3, Figure 41f; sample 12-96-4832-2, Appendix 4, Table 10). The geochemical patterns are not highly diagnostic (compare fields of Phanerozoic intrusions from the different environments in Appendix 3, Figure 41b, e and f).

5.10.4.3 Syn- to late-kinematic pegmatite and leucogranite (units 19a, b)

The youngest (late-kinematic), relatively parallel-sided veins, sheets and lenses consist of weakly peraluminous biotite±garnet granite–granodiorite. Near the sillimanite-biotite isograd, some are two-mica granite. These young intrusions are deformed only by F_6 folds and cut by weakly deformed planar granite pegmatite, with monazite yielding an age of 1789 ±2 Ma near Jungle Lake (Parent et al., 1999). The youngest muscovite-bearing granite in the Nokomis Lake area contains monazite that yielded a concordant and nearly concordant U-Pb TIMS age of 1809 ±2 Ma. This is tentatively taken as the crystallization age. Zircon grains in this rock have a high U content and do not yield a precise age. Similar synkinematic, variably deformed sheets locally yielded a U-Pb zircon age of 1798 +3/–2 Ma, but they may have started to develop throughout the main parts of the Kisseynew Domain earlier, shortly after the main foliation (S_5).

The young sheets and lenses are weakly peraluminous leucogranite–granodiorite with minor muscovite or garnet±cordierite (with A/CNK ratio of ~1.1). These are prominent in the south and in a larger body farther north. They have steep negative REE slopes with strong negative Nb and Ti anomalies on primitive-mantle–normalized plots, consistent with an origin as leucocratic fractionated melts largely derived from the felsic crust (Appendix 3, Figure 41g). The granite shows negative Sr and variable Eu anomalies related to loss of feldspar relative to accessory minerals (that contain the REE), either during melting or as the result of fractionation during transport. Depletion of HREE indicates garnet stability in the paleosome at the site of melting. This is most apparent in weakly foliated late-kinematic granite bodies, which have the strongly fractionated REE of leucosome that is interpreted to result from dry melting of biotite in the lower crust (Appendix 3, Figure 42e; Fornelli et al., 2002). These geochemical characteristics are consistent with fractionation and a high component of crustal melt in these bodies. The biotite breakdown allows LREE-bearing accessory minerals, which are commonly hosted in biotite, to become available to the melt. Garnet, which contains the HREE, remains in the melanosome (Fornelli et al., 2002). Leucogranite with depleted HREE has a major- and trace-element composition similar to that of the most fractionated Flatrock Lake granite, and represents emplacement and possible later development of melt during similar conditions in the Kississing–File lakes area (Table 16; Appendix 3, Figure 42a, d). Low ε_{Nd} marks the presence of an old component, a characteristic also found in Flatrock Lake granite (Appendix 2, Figure 29f). The ε_{Nd} of 0.5 (in the range of the Burntwood Group) for a tested sheet southeast of Duval Lake indicates that only a small proportion of Archean material was involved in the melt, but a larger component of nearly juvenile Paleoproterozoic sedimentary rocks and/or igneous rocks may have been present.

Pegmatite in the Burntwood Group migmatite was sampled east of Nokomis Lake (Zwanzig and Shwetz, 1994b) in the same zone as the pegmatite that yielded the 1809–1789 Ma monazite age near Jungle Lake. The bodies at Nokomis Lake are variably depleted in REE because mafic and accessory minerals were separated during fractionation and melting (and may have been missed during sampling); however, Eu is retained in the abundant feldspar (Appendix 3, Figure 42f). This created an REE pattern typical of granite leucosome (Fornelli et al., 2002; Clarke et al., 2005). Their variety of inherited zircon and monazite grains (Parent et al., 1999) is consistent with melting for a long period and over an extended thickness of crust. Analyzed granite with geochemistry similar to the pegmatite is restricted to sample 04-2-2201 from the Central Kisseynew subdomain but is common in other parts of the Trans-Hudson Orogen in peraluminous sheets formed ca. 1814 Ma from crustal melt (e.g., Clarke et al., 2005).

An important feature of the migmatitic leucosome and larger sheets of various granitoid compositions and textures is their considerable variety in the geochemical signature of trace elements, only a few of which match the consistent pattern of the phases chosen from the large plutons in the core of the Kisseynew Domain. This suggests that the leucosomes formed from small batches of melt that were rapidly removed. Their geochemistry suggests 1) a variety of melt sources, 2) more than one process of melting, and 3) that they are largely allochthonous and, in some cases, formed by a process similar to that responsible for the large plutons in the Central Kisseynew subdomain.

5.10.5 Late intrusive rocks: discussion

The late intermediate–mafic intrusive rocks in the Kississing–File lakes area and in the Central Kisseynew subdomain have melt components that formed in the mantle and imparted a geochemical pattern of strongly fractionated REE and negative HFSE anomalies (on MORB- or primitive-mantle–normalized plots). At first glance, this suggests continued oceanic-lithosphere subduction and arc magmatism during the 1830 Ma to ca. 1814 Ma span of the dated large intrusions. However, the interpretation of evidence presented in this report has stressed that the previous Missi-age period involved bimodal magmatism with local crustal extension and rise of asthenosphere, as well as uplift marked by a change to permanently continental sedimentation and volcanism. This was the result of collision tectonics, quite possibly accompanied by lithospheric thinning and followed by postcollisional convergence of cratonic blocks and crustal thickening that also caused the long period (ca. 1823–1789 Ma) of nearly exclusively felsic plutonism with a regional metamorphic peak at about 1815 Ma.

Although the well-organized process of subduction and the more enigmatic (collision-triggered) slab rollback and possible mantle delamination cannot be distinguished with geochemical data alone (Pearce et al., 1984), some data point to a non-arc process. Mafic–intermediate and felsic intrusive suites in the Kississing–File lakes area and Central Kisseynew subdomain are best explained by mixing of mantle and crustal components. This is particularly evident from the low ε_{Nd} ratio in the Dow Lake quartz diorite, which must have experienced mixing of mafic magma with Archean crustal melt, possibly from the underthrust Sask craton. This synkinematic intrusive suite is tholeiitic with respect to its elevated contents of FeO and TiO_2 (Appendix 1, Figure 21f) and low Al_2O_3, but its high Th and La place it in the field of common calcalkaline arc-volcanic rocks (Appendix 2, Figures 27g, 30f, 31f). Dow Lake quartz diorite and probably other units in that age division are equally well explained by decompression melting involving 1) rising asthenosphere; 2) the remaining mantle wedge; and 3) the lower crust—in the case of the quartz diorite, Archean crust. An alternative process of generating the late, mildly peraluminous granitoid rock is radioactive heating of thickened crust (White, 2005). This may have taken place even without the unproven 20 km original thickness of the Burntwood Group used in the model of White (2005). The felsic–intermediate arc crust provides a similar heat flow. The total content of Th and REE in the early successor-arc intrusions, which make up the main part of the arc massif, is similar to that of the Burntwood Group. (A weighted mean of total REE contents related to outcrop area of the various early successor-arc plutons is 103 ppm, compared to 109 ppm for the average nonmigmatitic Burntwood Group.) The geochemistry of the various late intrusive rocks, however, suggests melt components from the mantle, lower crust and middle crust that most likely require an additional heat input from the asthenosphere or fluid from active subduction.

Regardless of the exact tectonic environment in which these melts formed, the intrusions with low Yb and a trend of increasing $[La/Yb]_N$ were formed from multiple sources and preceded peak metamorphism, which was a period of high heat flow in the crust and probably the mantle. The Fe-rich (tholeiitic) synkinematic mafic dikes or sills (Weldon Bay gabbro) are products of decompression melting of enriched, probably asthenospheric mantle (Appendix 2, Figures 30f, 31f). They reached fault zones in the upper crust along the Kisseynew Domain margin, perhaps from more widespread mafic underplating that triggered the extensive melting of the crust.

The partly coeval to subsequent period of widespread, unfocussed felsic magmatism may have little mantle component but can be explained by three types of melting at several levels in the crust: 1) midcrustal fluid-present; 2) deeper dehydration-melting with biotite breakdown; and 3) deepest melting of amphibolite±garnet, mixed with mantle-derived melt. These have been proposed to generate, respectively, leucotonalite, granite, and adakite- or TTG-type to mafic calcalkaline melts (e.g., Fornelli et al., 2002). The large granite–tonalite plutons in the Central Kisseynew subdomain are possibly the result of mixing of these types of melt and final melting in the mantle wedge. The purely granitoid rocks cannot be distinguished from volcanic-arc intrusions using only geochemical data. Two phases selected from the large plutons in the Kisseynew

Domain interior (Highrock Lake tonalite and Flatrock Lake granite) have uniformly distinctive multi-element plots (e.g., Appendix 3, Figure 41e) consistent with biotite breakdown, leaving garnet in the residuum to form their elevated LREE and depleted HREE. Their Sr and Eu depletion indicates loss of plagioclase during fractionation; their negative HFSE anomalies can be explained equally well by inheritance of earlier arc characteristics from the crust and mantle, or melting above an active subduction zone. The absence of mafic–intermediate arc intrusions and presence of high-LILE, high–Mg-Ni-Cr, low-HREE rocks, however, suggests a non-arc process. The apparent absence of sheets of typical Eu-enriched tonalitic leucosome suggests that most melts formed deep in the crust where little fluid was present. The relatively low ε_{Nd} indicates that a limited Archean component was present. Such melting takes place after continental collision and attendant crustal thickening, or during arc magmatism in thick continental crust. In the case of the Kisseynew Domain, the latter is suggested by the late synkinematic age of these plutons formed during terminal collision tectonics.

5.10.6 Discussion of the igneous geochemistry

A surprising conclusion of this study is that the high-grade metamorphic rocks and even some tectonites in the Kississing–File lakes area have retained many of their primary geochemical characteristics. Of about 40 different igneous units for which more than one sample has been analyzed for major and trace elements, most form compact fields that overlap little with those of other units on some of the various kinds of element diagrams used in this report. Weight percent of Yb plotted against the ratio of $[La/Yb]_N$ provides a powerful tool for distinguishing the various volcanic units, particularly when MgO contents are taken into consideration to minimize the effects of crystal fractionation and fractional melting (Appendix 2, Figure 24). Ratios of incompatible trace elements generally plot as linear trends that are unique for many volcanic units, whereas different units generally plot on different lines (Appendix 2, Figure 26). The geochemical data are therefore very useful in identifying comagmatic units, even where primary structures are no longer preserved. Thus, the geochemistry serves as an important mapping tool in tracing units and units, more than 70 of which have been identified in the Kississing–File lakes area. Once identified, units can be traced along strike or north through the complex structures in the overlap area between the Flin Flon and Kisseynew domains, thus pointing to new areas for mineral exploration where units occur that are known to contain deposits in the south. A summary of the geochemical stratigraphy and structural continuity of units partly defined by geochemistry is given in restored columns arranged in the currently overlapping structural sheets and early fold limbs (I–V in Figure 43, back pocket).

One of the most important geochemical correlations throughout the Kississing–File lakes area is of well-preserved Josland Lake sills southwest of File Lake with the widespread, more highly metamorphosed sills extending northwest to Kississing Lake (columns in structural level III, Figure 43). They provide the minimum age (1886 Ma) for the Fourmile Island arc assemblage. Not only is there a ubiquitous origin by decompression melting of similar, slightly enriched-mantle that experienced exceptionally strong fractionation to ferrogabbro and trondhjemite, but there is, everywhere, the uniquely low La/Yb ratio of the volcanic host units (Fussey, Moody Lake, Preaston, and Storozuk volcanics). This allows a correlation of these rocks over distances of 100 km along the local structural trend and makes this combination of rocks an outstanding marker for structural mapping. This package of rocks, with its unconformably overlying Missi Group, defines a D_3 thrust sheet that has been transported over the Burntwood Group turbidite along the Loonhead Lake Fault during its early phase of thrusting.

A second important correlation of units is of the pillow basalt and sills of the Ponton Lake basalt south of Ponton Lake with the geochemically identical, more highly deformed and metamorphosed amphibolite north and west of the lake (columns 2A and 2B, Figure 43). This correlation, along with the underlying andesite units, shows the northeast structural continuity beneath the higher nappes and helps to define the late (F_5) Ponton Lake synform. The metavolcanic rocks in the outer limb of this synform carry the high mineral potential of the Central Flin Flon subdomain into the northern part of the subdomain.

Although the alkali elements have been mobile, major evolution trends of increasing K_2O (and other alkali elements) with time have been shown to be similar to trends resulting from early juvenile oceanic-arc and back-arc magmatism to continental-arc magmatism in the better preserved parts of the Flin Flon Domain (Lucas et al., 1996; Whalen et al., 1999). There is also a trend in the Kississing–File lakes area from early tholeiitic mafic volcanism to late calcalkaline felsic plutonism, along with increasing La/Yb ratios and LILE contents. Such trends and other distinctive geochemical features help to identify and characterize many of the major age-groupings of the igneous rocks in the Kississing–File lakes area: 1.89–1.88 Ga, early assemblages; 1.88–1.85 Ga, early successor-arc intrusions; 1.85–1.83 Ga, late successor-arc and basinal units; and 1.83–1.79 Ga, synkinematic, postcollisional intrusions. The Sherridon–Meat Lake assemblage is an exception: it comprises volcanic rocks with relatively low trace-element contents and moderately fractionated REE, but has yielded U-Pb zircon ages similar to those of the early successor-arc intrusions. These later juvenile volcanic rocks, however, formed in an oceanic tectonic environment, much like the early arc-volcanic deposits.

The general coherence of the geochemical data shows that various plots of elements and ratios provide a meaningful assessment of the tectonic environments of the different volcanic and intrusive units. A simple plot of TiO_2 versus MgO separates MORB-like units and back-arc–basin basalt (BABB) from the main arc-derived units (Appendix 1, Figure 22). The plots using Th, Zr and Nb to identify the tectonic environment of basalt and andesite (Appendix 2, Figure 30) are supported by plots using La, Y and Nb to avoid the effects of Th contamination caused by incipient melting during high-grade metamorphism, or crustal contamination during magma ascent (Appendix 2, Figure 31). The data show that mafic–intermediate igneous rocks in the Kississing–File lakes area were tholeiitic during the early magmatism; calcalkaline suites make up most of the late successor-arc and synkinematic rocks. The various plots of relatively immobile major and trace elements indicate a division of early units into 1) MORB (or BABB formed dominantly by decompression mantle melting) and those with 2) weak and 3) strong arc-like signatures. This subdivision identifies assemblages with low and high mineral potential according to tectonic origin (i.e., non-arc and arc). Several units that fall between the fields of MORB and arc may have formed during arc extension or at the margin of a back-arc basin, also with significant mineral potential. The conclusions based on standard diagrams are supported by the shape of chondrite-normalized REE plots, as well as by the presence or absence of the typical negative HFSE (Nb, Hf, Zr and Ti) anomalies in MORB-normalized or primitive-mantle extended-element plots (Appendix 3, Figures 32–42). The strongly arc-related assemblages (formed by strongly fluid-fluxed melting of the mantle wedge) include LREE-enriched and MREE-depleted units with high VMS potential where they occur in the Sherridon–Meat Lake area.

Given a preliminary age of 1850–1855 Ma for the Sherridon–Meat Lake assemblage (N. Rayner and D. Tinkham, pers. comm., 2009), these rocks are remarkably similar to the ca. 1890 Ma Snow Lake assemblage. Their mantle and crustal melt sources, as well as their tectonic environment, may have been very similar. Both successions contain pillow basalt and are interpreted as marine. They are depleted in MREE and HFSE, extending from relatively primitive to chemically evolved compositions. Felsic volcanic rocks are abundant and regional, and focused hydrothermal alteration is widespread. Both prominently felsic packages occur on the margin of the Flin Flon Domain facing the Kisseynew Domain. Our data indicate that these characteristics are considerably more important than the ages of the rocks—a most significant finding of this report.

Structural and stratigraphic considerations in this report indicate that the late successor-arc magmatism was syncollisional, occurring during the collision of the Flin Flon–Glennie Complex, the Sask craton and probably the Superior craton (Zwanzig, 1999; Section 6.1.6.2). This is inconsistent, however, with the relatively low contents of Rb in these rocks (Appendix 2, Figure 25g) and other geochemical characteristics. Consequently, the period of collision also must have involved the subduction of a sizable remnant ocean basin in the THO in Manitoba. The first contact of thick blocks may have been in Saskatchewan, where the oldest unconformity has been suggested to occur at the base of the syncollisional sedimentary rocks that are possibly correlative with Burntwood Group. Such a collision probably involved microplate rotation but no subduction of a continental plate or its margin (Section 6.1.8).

A related aspect of this study is the conclusion that synkinematic plutons with apparently 'arc-like' geochemistry were more likely produced by syn- to postcollisional, non-arc magmatism than in a late volcanic arc. This includes large tonalite–granite plutons in the Central Kisseynew subdomain and migmatitic sheets throughout the Kisseynew Domain. These intrusions are distinguished by their mildly peraluminous composition and general absence of mafic to intermediate members. They range to high alkali and LILE contents. They are also distinguished from melts generated in the upper (garnet-free) mantle wedge by their HREE depletion and may consist largely of melted crust. Mantle-derived ultramafic to intermediate intrusive members are present only in the earliest synkinematic (Touchbourne) suite, for which a postcollisional origin is also preferred. Rare late mafic dikes are continental-type tholeiite and suggest an input of mafic within-plate magma and mantle heat. The predominantly felsic synkinematic intrusions were probably generated by 1) dehydration melting involving mica in the middle to lower crust, as well as possible melting of 2) mafic lower crust or 3) slab remnants, 4) previously hydrated and contaminated mantle, and 5) limited fluid-present melting of quartz+plagioclase in the middle crust. Granulite apparently underlies the migmatite complex and is exposed in the Northeast Kisseynew subdomain (Growdon and Percival, 2008). These rocks contain hypersthene, particularly evident in older orthogneiss and in the leucosome of the Burntwood Group. Partial melts from this level in the crust would have been relatively dry, mobile and potassic. This is consistent with melts formed at the newly proposed high temperature of metamorphism (Growdon and Percival, 2008) and intruded as sheets higher in the migmatite complex.

Our preferred model of a heat source during the synkinematic magmatism is from the mantle (Kraus and Menard, 1997) in addition to radioactively generated heat from thickened crust (White, 2005). Involvement of decompression melting of the mantle and possible mafic underplating is suggested by the set of widely scattered synkinematic tholeiitic sheets (Weldon Bay and Cree Lake sheets). The arc-like geochemistry of the synkinematic felsic–intermediate intrusive rocks is believed to be inherited from the melting of the underlying juvenile arc-massif (Flin Flon–Glennie Complex), its detritus (Burntwood Group) and local Archean crust (possibly Sask craton), as well as remaining metasomatized or asthenospheric mantle. The exact age and origin of these late magmas is, however, uncertain.

The preserved geochemical data in the medium- to high-grade metamorphic rocks prove to be sufficient to trace the evolutionary history of the Kississing–File lakes area from the economically important early stage of juvenile-arc and back-arc magmatism, through volumetrically important but still economic successor-arc magmatism. This was followed in the Kisseynew Domain by local mafic–intermediate magmatism and widespread late felsic magmatism with a complex crustal origin. Each of these stages had a different metallogenic association. Newly proposed ideas in this report include 1) early (>1886 Ma) assembly of the Flin Flon–Glennie Complex followed by rifting and later reassembly; and 2) an extended period of crustal uplift during collision tectonics and final arc magmatism, local extension and development of sedimentary sub-basins and possible back-arc core complexes. This probably started ca. 1850 Ma and was punctuated by a pulse of felsic-dominated bimodal magmatism (ca. 1836–1832 Ma Missi Group volcanic and intrusive rocks). Uplift may help to explain the onset of orogen-wide continental sedimentation and block faulting. Subsequent, predominantly felsic, weakly peraluminous magmatism was synkinematic during crustal thickening. This occurred in a continental environment, after remnant-ocean basins had been consumed but the surrounding Archean cratons continued to converge.

6 Tectonic evolution

This section is a discussion of a tectonic model based on conclusions made in previous sections. The new information on the geology, geochronology and geochemistry of the rocks in the Kississing–File lakes area supports a more complex tectonic model than previously proposed for the Flin Flon Domain (e.g., Syme et al., 1995; Lucas et al., 1996), and this has repercussions on the metallogeny of the area. The model of early paleogeography suggests that the VMS-rich rocks of the Flin Flon Domain may have extended as a single arc across the western Trans-Hudson Orogen. Thrusting of the Amisk Collage over the Burntwood Group greywacke-mudstone turbidite may have been important in providing a fluid-rich infrastructure that resulted in the orogenic gold deposition during subsequent metamorphism. The evidence of possible mantle delamination during collision and the emplacement of the Sask craton with its ancient subcontinental lithospheric mantle provide a new incentive for diamond exploration in Manitoba. The tectonic model is discussed below and the possible economic ramifications are given in Section 7.

6.1 New model

This study of the Kississing–File lakes area provides a new interpretation of the 130 m.y. tectonic evolution of the mainly juvenile Paleoproterozoic crust in the southern part of the Trans-Hudson Orogen (THO) in Manitoba. Of considerable importance is the high structural relief that provides access to rocks originally part of the near-surface to midcrustal levels in the internal zone of the THO. Thus, volcanism, sedimentation, plutonism and deformation can all be related to processes that involved the entire crust and generally the mantle. Although interrupted by faults and younger plutons, early ocean-floor rocks may be traced from an intrusive level (Reed Lake Complex) to thick pillow basalt (e.g., Ponton Lake basalt), unconformably overlain by the youngest supracrustal rocks (Missi Group). The map area is sufficiently large that remnants of several volcanic edifices are present in the longer belts of related rocks. Changes along strike provide important clues to early paleogeography, whereas successive structural events give an indication of how the present crustal architecture may have formed and may be retrodeformed to generate the tectonic model. Although a unique path of tectonic evolution cannot be established rigorously, many of its aspects are supported by several lines of evidence, including tectonostratigraphy, structure and, in particular, lithogeochemistry. A better understanding of the early regional evolution guides construction of the new tectonic model of the THO.

The tectonic evolution is divided into five stages: 1) early arc, arc-rift and back-arc volcanism with early assembly of these disparate terrane fragments; 2) arc rifting and mafic intrusion with non-arc geochemistry; 3) early successor-arc magmatism; 4) late successor-arc magmatism and basin evolution during collision tectonics; and 5) orogeny involving synkinematic magmatism, terminal collision and deformation with final convergence and block faulting. These stages are first summarized in the tectonic model, then their effects are assessed in different tectonic subdomains and described separately. Because each stage is associated with different types of mineral deposits, the tectonic history is important for understanding the controls on the deposits and major showings.

Lithogeochemistry is instrumental in identifying tectonic environments and showing how these evolved. Preservation of the geochemistry of some major elements and many trace elements has been demonstrated even in highly metamorphosed, strongly deformed rocks in the study area. Values of certain fluid-mobile elements are variable but apparently still meaningful. The least mobile trace-element ratios that do not vary with fractionation are particularly useful in comparing geochemical processes with those in modern tectonic environments. Analogies with modern orogens are made on the assumption that most tectonic processes in the THO have only limited differences from those that control modern plate formation, destruction and collision. This assumption is generally verified by the remarkable similarity of multi-element patterns in the metavolcanic rocks with those of the more modern arc and back-arc volcanic rocks that are reported as unaltered. Some nearly ultramafic flows (in Ponton Lake basalt) and the general abundance of early bimodal arc assemblages of tholeiitic basalt and rhyolite, however, suggest higher mantle temperatures and more extension tectonics than in modern arcs. This may have contributed to the abundance of VMS deposits in the Flin Flon Domain. The Sherridon–Meat Lake assemblage, with its oceanic-arc geochemistry and VMS deposits but a preliminary successor-arc age of 1850–1855 Ma, points to a relatively young episode of volcanism with a high economic base-metal potential. The presence of successor-arc plutons with chemistry similar to that of Archean TTG suites is consistent with slab and crustal melting (Whalen et al., 2002; Condie, 2008) and may be related to the apparent absence of economic porphyry-copper deposits (Richards and Kerrich, 2007) in the Flin Flon and Kisseynew domains. Similarly, the major- and trace-element geochemistry of the late successor-arc bimodal magmatism indicates extensive involvement of juvenile Proterozoic crust or sediments in the generation of melt. Contamination by Archean crust and mantle was sporadic in the early stages and nearly absent in the late successor-arc stage, but more common in postcollisional magmatism during advanced convergence of the cratons.

Prominent sedimentary deposits (Burntwood and Missi groups) appeared only during the final stages of arc magmatism, the attendant sub-basin development and subsequent collision tectonics with basin closure. The juxtaposition of the Burntwood Group and slices of older MORB or BABB between the advancing Superior craton and the late successor arc(s) in the present Flin Flon, Glennie, La Ronge and Lynn Lake–Leaf Rapids areas suggest that these metaturbidites may have covered a wide suture zone that did not develop blueschist or mélange but was rapidly involved in high-temperature tectonics during continental collision. Sedimentation may have started by filling a shallow trench as an accretionary wedge and continued to prograde over the closing Kisseynew basin. Hints of depositional contacts at the base of the Burntwood Group on 40 m.y. older back-arc and arc-volcanic rocks suggest extension and foundering of the margin of the early volcanic terrane. A possible base of the Burntwood Group and its shallow-water equivalents on the continental-margin sedimentary rocks overlying slivers of Superior-type basement in the Northeast Kisseynew subdomain suggest deposition in a peripheral foreland basin (Sinclair, 1997). However, the earlier proposed origin as a back-arc basin (Ansdell et al., 1995) cannot be ruled out because the subduction polarity is uncertain. This part of the tectonic model and the importance of early amalgamation, arc rifting, disruption and reassembly are newly proposed and require testing.

The youngest igneous rocks in western THO, mainly synkinematic granitoid rocks, generally have low contents of Rb, Nb and Ti, and elevated Th. They are mildly peraluminous, however, and herein proposed to be postcollisional. Their geochemistry suggests significant recycling of juvenile-arc crust rather than mature continental crust. This gives an arc-like geochemistry to these late granite bodies.

For ease of reading, only a few critical references are cited in this section of the report. The remaining references and new data are given in Sections 3–5.

6.1.1 Tectonic subdomains

Five tectonic subdomains with different histories are recognized in the Kississing–File lakes area. They are defined in this report as the Central Flin Flon, Snow Lake, Batty Lake and Central Kisseynew subdomains, and the Kisseynew South flank (Figure 2). Each subdomain has a broadly similar, early coeval suite of arc, arc-rift and back-arc volcanic rocks formed in stage 1. Successor-arc and collision-related rocks of the subsequent stages can vary considerably among the subdomains, suggesting a different tectonic evolution of the domains during stage 2. The Central Flin Flon subdomain is dominated by the early volcanic and early successor-arc intrusions (stages 1 and 2). These are unconformably overlain by the Missi Group. The eastern part of the Snow Lake subdomain includes the Schist-Wekusko assemblage (stage 2), which is unconformably overlain by the Missi Group. Intrusions of known age are younger there (stage 3). The Batty Lake subdomain includes stage 2 successor-arc volcanic rocks (Sherridon–Meat Lake assemblage), according to preliminary U-Pb zircon isotopic data, and nearly coeval granitoid rocks. The Kisseynew subdomains are dominated by stage 4 syncollisional sedimentary rocks, but these also occur locally throughout the Flin Flon Domain. The Burntwood Group is generally in fault contact with rocks of stages 1 and 3 in the Snow Lake and Batty Lake subdomains, but forms only minor slivers to the southwest (Figure 8a, back pocket). The Burntwood Group appears to grade upward into the Missi Group on the Kisseynew South flank. Only Burntwood Group migmatite and late plutons, including abundant synkinematic, weakly peraluminous granite, are exposed in the Central Kisseynew subdomain.

The results of protracted deformation are also variable across the subdomains. The early history of terrane amalgamation is preserved throughout the Flin Flon Domain. Recumbent D_1 folds and the West Reed–North Star Shear Zone occur on the east side of the Central Flin Flon subdomain, whereas early upright folds and faults occur in the west. Steep S_2 foliation due to east-west compression occurs throughout the subdomain. Evidence of early post-Missi (D_3) thrusting of the Amisk Collage and early plutons over the younger Burntwood Group is found at the eastern and northern margins of the Central Flin Flon subdomain (Morton Lake and Loonhead Lake faults). These early faults form a structural discontinuity that extends across the Kississing–File lakes area, bounding the Central Flin Flon subdomain at its eastern and northeastern margins (Figure 2). The main crustal architecture, however, was probably developed during D_4 east-over-west stacking due to collision of the Flin Flon and Kisseynew domains with the Thompson Nickel Belt (TNB) and Superior craton (Figure 6). Late- to postcollisional structures (D_5) strongly affect areas in the north and east that have experienced high-temperature midcrustal deformation. In the medium- to low-grade part of the Flin Flon Domain, this phase appears to be restricted to shear zones. Postcollisional (D_6) open folding, faulting and reactivation of shear zones is interpreted as the result of renewed northwest compression and final north-northeast extension that may have helped to preserve the high-level part of the Flin Flon Domain due to relative down-throw. Thus, the Kississing–File lakes area provides continuous magmatic, sedimentary and structural histories of the internal zone of the THO over a time span of 130 m.y. Understanding this history and the resulting crustal geometry should provide support for future mineral exploration.

6.1.2 Tectonostratigraphic restoration

The work leading to this report (including Zwanzig, 1999) has provided a better understanding of the large-scale structural geometry of rocks in the Kississing–File lakes area. This has allowed a partial restoration of tectonostratigraphic packages to the end of stage 2, after the early collisional fold-and-thrust stack had formed during D_3 (Figure 7b; Figure 43, back pocket). Figure 43 shows five structural levels (I–V) of internally correlated packages in their present order, but used to analyze and partly restore stage 2 in the tectonic evolution of the Kississing–File lakes area. The restoration is based on remnant stratigraphy and structure cross-strike sections, but even more on correlation of geochemical units. The partial retrodeformation is qualitative, consisting of a series of stratigraphic columns, but accurate unit thickness or original separation of columns could not be determined because of the high penetrative strain and noncylindrical folding. Columns were restored using the base of the Missi Group as datum and matching volcanic successions below the unconformity, as well as using the proximal to distal nature of the lower part of the Missi Group. All columns were placed upright as at the end of F_3 thrusting. For clarity, those presently in the west and north are shown on the left side of Figure 43, but they are interpreted to have originally lain farther north and east (present co-ordinates). Also for clarity and alternative interpretations of the structural history, the present structural level of each package is maintained. Major faults and fold hinges are named as in Figure 12. A more complete restoration from column 2A to column 4D is attempted in schematic fashion in Figure 7. The columns in Figure 43 are used to illustrate the correlation of units along strike and the present upward changes from the early volcanoplutonic rocks of the Central Flin Flon subdomain in the south (starting at column 2C, level II) through the Kisseynew basin fill (at column 4D, level IV) to the structurally overlying Sherridon–Meat Lake assemblage (column 5, level V).

The lowest structural level (I) occurs near Sherridon and is composed of early successor-arc plutons and the Sherridon–Meat Lake assemblage in the Sherridon-Hutchinson Complex.

Structural level II is formed by the Amisk Collage, early successor-arc plutons and unconformably overlying Missi Group in the southwest, and the Fourmile Island and Northeast Reed assemblages in the southeast. It is apparent that these rocks were mainly flat lying and upright during the intrusion of early Josland Lake sills in the southeast and remained so during deposition of the Missi Group farther northwest.

Structural level III is composed of the arc-rift units (northern part of the Fourmile Island assemblage and Josland Lake sills) and unconformably overlying Missi Group, which together form the Loonhead Lake thrust sheet. The Loonhead Lake thrust sheet represents the structurally inverted northern and eastern margins of the Central Flin Flon subdomain. It presently trends west to northwest as the Loonhead Lake Fault (LHF in Figure 7a). It is restored to an upright pre-D_4 position in Figure 43. Columns 1B and 1C are interpreted to have been located north of column 9B with the present sinistral bend at the northern margin of the Kisseynew Domain restored to a straight line.

Structural level IV comprises the conformable (distal) Burntwood and Missi groups, mainly sedimentary rocks belonging to the Kisseynew South flank. These rocks are interpreted to have been folded under and in front of the early Loonhead Lake Fault during D_3 and are therefore restored to the northeast.

The fifth package (level V) contains the Sherridon–Meat Lake assemblage in the Walton Lake nappe (columns 5 and 9E, Figure 43). Its correlation with the rocks that lie at the deepest level in the Sherridon structure confirms that a thick package of rocks must have been inverted to the southwest in the middle crust during the advanced stages of deformation (D_4–D_5). Our preferred correlation of the garnet-biotite±hornblende gneiss (unit 10a) in the core of the Sherridon structure with the Burntwood Group is consistent with D_3 final amalgamation of the Flin Flon–Glennie Complex. A sixth structural level (not shown in Figure 43 but continuous with column 9D) is composed of distal parts of the Burntwood and Missi groups, which lie below the File River nappe and overrode the other structural packages during D_4–D_5 (Figure 13).

The interpretation of Zwanzig (1999) used in the restoration in Figure 43 implies that there was an east- to northeast-directed D_3 fold-and-thrust belt extending from column 9B at Morton Lake to column 4C at Jungle Lake, and from column 9D at File Lake to column 2E at Puffy Lake. This structural stack was overturned during vergence reversal and flattened below the D_4–D_5 File River nappe, Batty Lake Complex and Walton Lake nappe (Figure 13a). The stack was also rotated counterclockwise to the southwest and west during the progressive deformation. The presently inverted panels are labelled in Figure 7b. The overturning to recumbence probably involved east-over-west and northeast-over-southwest distributed shear at midcrustal level (cf. Williams and Jiang, 2005), and was likely related to underthrusting by the Sask craton from the southwest. The base of the inverted section is the F_4 Star Lake syncline (STS in Figure 12 and 13), where evidence of high strain indicates strong sinistral shear. Sheath folding apparently took place within small nappes, and vertical flattening probably resulted from tectonic burial. Deep tectonic burial and subsequent prograde metamorphism with uplift are consistent with the pressure-temperature data and modelling by Growdon (pers. comm., 2008). The D_4–D_5 deformation can also be related to overthrusting by the Superior craton from the east, and by the Northeast Kisseynew subdomain from the north and northeast.

A widely accepted, simpler interpretation involving long-lived tectonic transport directed from northeast over southwest is based largely on the present northeast dip of exposed structures and deep seismic reflectors (Zwanzig, 1990; Zwanzig and Schledewitz, 1992; Ansdell et al., 1995, 1999; Syme et al., 1995; Connors, 1996; Lucas et al., 1996; Kraus and Williams, 1999; Ansdell, 2005). However, such thrusting would have placed younger rocks of the Kisseynew Domain directly over the older rocks of the Flin Flon Domain during an early basin inversion, in a direction opposite to sediment transport. The model would imply that many of the sections shown in Figure 43 had to be overturned before thrusting. This model is herein deemed less likely than vergence reversal.

The tectonic model that involves a reversal in vergence is more consistent with the following implications of restoring the Flin Flon–Kisseynew overlap area:

1) The abundance of clasts from the Flin Flon–Glennie Complex in the Burntwood and Missi groups, and an absence of Burntwood clasts in the Missi Group, do not support an early basin inversion.
2) The critical taper required for thrusting, as indicated by the paleoslope away from the uplifted Flin Flon–Glennie Complex and its mountain front, is toward the Kisseynew Domain, in the direction of sediment transport of both turbidite and nonmarine facies.
3) The overlap of the early volcanic rocks in the Loonhead Lake thrust sheet on the Burntwood Group is indicated from the facing of the unconformity at the base of the Missi Group within the sheet.
4) A history of overturning of the axial-planar Josland Lake sill west of Morton Lake favours thrusting from the southwest (Figure 15).
5) Well–known, early, northerly directed thrusting placed the Amisk Collage over the Missi Group also at Flin Flon (R-L. Simard et al., work in progress, 2010).
6) The only southwest-directed kinematic indicators occur in S_5 foliation formed during the ca. 1810–1800 Ma thermal peak (Zwanzig, 1999).
7) Southwest-directed faults and shear zones such as the Fairwind Lake Fault and, in the east, the McLeod Road Fault cut, respectively, the Loonhead Lake thrust and the Snow Lake Fault to the southwest.
8) The major structural bend on the north side of the Flin Flon Domain is the result of advanced deformation during the thermal peak.

These features are consistent with deposition of the Burntwood Group as orogenic flysch in an evolving fore-arc and trench environment on subducting ocean floor that was attached to a Superior plate to the east. The Missi Group molasse and volcanic rocks have prograded over the Burntwood Group during final arc magmatism and the onset of continental collision.

Southwest-directed tectonic transport was the result of advanced collision, during which earlier structures were inverted and transposed at a crustal scale.

6.1.3 Early arc, arc-rift and back-arc volcanism

The geochemical characteristics and tectonic origins of the early volcanic rocks (units 1a–3d), which formed before their widespread intrusion by the 1886 Ma Josland Lake sills in the Kississing–File lakes area, are summarized in Table 17. Similarities in field relationships, geochemistry and Nd-isotope ratios with rocks in the low-grade parts of the Flin Flon Domain, which have yielded high-precision ID-TIMS ages between 1904 and 1885 Ma (Manitoba Geological Survey, 2006), are consistent with units 1a–3d also forming as products of arc and back-arc magmatism. Thus, the inferred early history of these rocks likely involved not only arc and back-arc magmatism, but also the structural amalgamation, protracted or repeated arc rifting and final displacement of arc-terrane segments previously recognized in the lower grade portions of the Flin Flon Domain (Lucas et al., 1996). The arc rifting is interpreted here to have led to the development of ocean floor that became part of the basement of the Kisseynew Domain. Thus, although the early tectonic history is shared by rocks in the Central Flin Flon subdomain and the adjoining Snow Lake subdomain, their common history likely ended with separation of the two subdomains during the arc rifting recorded by units 1a–3d, with both subdomains following different paths of evolution during subsequent tectonic events.

Units of arc-derived, low-Ti amphibolite (Puffy Mine, Koscielny and Dow Lake andesites and Martell Lake basalt) occur at low structural and stratigraphic levels (columns 2A and 8B in level II, Figure 43, back pocket). This early juvenile assemblage is broadly correlated with the arc assemblage near Flin Flon. Like their modern counterparts, many early units have a limited extent and probably represent the remnants of single submarine volcanoes, each with a slightly different range of compositions and trace-element fingerprint but with strong characteristics of arc-volcanic rocks (Table 17; Appendix 3, Figure 34f–h). Their products include fragmental rocks, pillowed flows and local volcaniclastic sedimentary rocks. All of these formed in a primitive or weakly extensional oceanic-arc environment. Unlike most modern island-arc deposits, they are relatively mafic or bimodal, with local felsic rocks, and lack abundant calcalkaline andesite. A unit with a weaker arc signature (Preaston andesite), which occurs in the lower part of the Fourmile Island assemblage, probably formed with the overlying Dickstone rhyolite at the onset of arc rifting.

Early back-arc magmas and some units of arc-rift basalt erupted in basins that were distant from sources of detritus and felsic volcanic centres, thus developing monotonous successions of pillow basalt and consanguineous high-level intrusions. As with the ocean-floor assemblages in low-grade parts of the Flin Flon Domain, no sedimentary rocks and felsic volcanic rocks are intercalated with these flows and sills. The great strike length and abundance of these units suggest that the basins were large, constituting a major part of the preserved juvenile Paleoproterozoic oceanic environment. Back-arc volcanic rocks and transitional to arc basalts occur both in the Amisk Collage (Ponton Lake and Moody Lake basalts) and in the Northeast Reed assemblage (Fussey Lake basalt).

Ponton Lake basalt has fractionated REE with only moderate HFSE anomalies (Appendix 3, Figure 34b) and is assigned to the Elbow-Athapapuskow ocean-floor assemblage, which contains similar rocks. It is likewise transitional to E-MORB (Appendix 2, Figure 30a). The Ponton Lake basalt is also similar to the Scotty Lake contaminated arc-rift basalt in the central part of the Flin Flon Domain (Syme et al., 1999). Very similar geochemistry also developed in the south Atlantic on parts of the East Scotia ridge, in a purely oceanic back-arc–basin environment (Stern et al., 1990; Gribble et al., 1998; Fretzdorff et al., 2002). The Ponton Lake unit appears to lie stratigraphically above the early arc deposits but may have been overthrust from the west with the Elbow-Athapapuskow ocean-floor assemblage in the Amisk Collage before the intrusion of the Gants Lake Batholith. Although there are remnants of coeval mafic–ultramafic layered intrusions in the Flin Flon Domain (Reed Lake mafic–ultramafic complex and small bodies in the Dow Lake area), no coherent ophiolite assemblages are preserved. Nevertheless, the presence of the ultramafic tectonic blocks along a major structural break (West Read–North Star Shear Zone) suggests that the Elbow-Athapapuskow ocean floor was obducted over the Fourmile Island assemblage. This is consistent with the easterly vergence suggested for the D_1 deformation (see caption of Figure 15).

The Moody Lake and Fussey Lake units had a different mantle source than the Ponton Lake basalt: the Moody Lake basalt source was depleted. The two sets of assemblages are separated by a narrow belt of volcanic-arc units north of North Star Lake and probably had different ages. The Moody Lake basalt and the slightly more arc-like Storozuk volcanics in the northern Fourmile Island assemblage occur along the northern and eastern margins of the Central Flin Flon subdomain (Figure 2). They are LREE depleted (Appendix 3, Figure 32c–e) and most samples have only small HFSE anomalies, similar to dredge samples from the Manus back-arc basin in the south Pacific (Sinton et al., 2003). They also have high ε_{Nd} and are transitional to the Fussey Lake basalt, which is similar to back-arc N-MORB. Their geochemistry compares well with samples from south Pacific and other back-arc–basin ridges and basin margins that provide modern analogues to ocean basins developed behind island arcs (Appendix 3, Figure 34a, c–e). Evidence that the LREE-depleted northern part of the Fourmile Island assemblage formed during arc rifting of the underlying arc-volcanic rocks (Martell Lake basalt and Dow Lake andesite) is provided by local stratigraphic interlayering with SE Dow porphyritic basalt and the gradual chemical transition from arc tholeiite to BABB. Preservation of an extensive arc-rift margin is suggested by the 100 km long tract with a weak arc signature that extends from Preaston andesite (southwest of File Lake) through Moody Lake basalt (toward Kississing Lake) to the Dismal Lake suite of Gilbert (2004), south of Kisseynew Lake. The arc signature becomes progressively weaker along this tract, showing a progressive decrease in Th/Zr and slight decrease in Th/Nb (Appendix 2, Figure 30a).

The onset of arc rifting is marked by the Dickstone rhyolite, which plots in the field of ocean-ridge granite (ORG in

Table 17: Geochemical, tectonic and metallogenic characteristics of early juvenile-arc and back-arc assemblages in the Kississing–File lakes area.

Unit	Assemblage	Type location	Geochemical characteristics	Isotope evolution	Tectonic origin	Mineral deposits
MORB-like basalt (back-arc basin floor)						
1a	Fussey Lake basalt	South of File Lake	Flat N-MORB–normalized extended-element plots, minor positive Th anomaly	Juvenile	Back-arc basin	
Volcanic rocks with weak arc signature (arc rift)						
2a	Ponton Lake basalt	Ponton Lake	Elevated Th and LREE (contaminated?); weak or lacking HFSE anomalies; elevated Mg, Nb/Yb (plume related)	Juvenile to contaminated	Back-arc basin	
2b	Moody Lake basalt	Dow Lake to Kississing Lake	Depleted LREE; negative Nb anomaly; small negative Zr, Ti anomalies; low Nb/Yb (N-MORB like)	Juvenile	Depleted back-arc basin (margin?)	
2c	Preaston andesite	W of File Lake	Depleted LREE; negative Nb, Ti anomalies; low Nb/Yb (N-MORB like)	Juvenile	Arc, transition to arc-rift	Dickstone VMS deposit
2d	Dickstone rhyolite	W of File Lake	REE pattern as Storozuk basalt but elevated (related by fractionation); low Rb, Th/Yb; high Y+Nb	Juvenile	Arc, transition to arc-rift	Dickstone VMS deposit
2e	Storozuk basalt	W of File Lake	Flat N-MORB–normalized element plot; negative HFSE anomalies	Juvenile	Arc, transition to arc-rift	Dickstone VMS deposit
2f	SE Dow Lake porphyritic basalt	SE of Dow Lake	Depleted REE, negative REE slope; elevated Th, little/no HFSE anomalies			
Arc tholeiite (Amisk collage)						
3a	Puffy Mine andesite	N of Puffy Lake	Weakly depleted HREE; variably elevated LREE; negative HFSE anomalies; elevated Th, high Th/Nb	Juvenile to weakly contaminated	Primitive tholeiitic arc, high subduction component	Puffy Lake Au depososit
3b	Koscielny andesite	SE of Puffy Lake	Weakly depleted HREE; variably elevated LREE; negative HFSE anomalies; elevated Th, high Th/Nb	Juvenile to weakly contaminated	Primitive tholeiitic arc, high subduction component	Puffy Lake Au depososit
3d	Dow Lake andesite	SE of Dow Lake	Weakly depleted HREE; variably elevated LREE; negative HFSE anomalies; elevated Th, high Th/Nb	Juvenile to weakly contaminated	Primitive tholeiitic arc, high subduction component	Puffy Lake Au depososit
3c	Martell Lake basalt	NE Martell Lake	Concave-up N-MORB–normalized element plot	Juvenile to weakly contaminated	Primitive tholeiitic arc, high subduction component	Puffy Lake Au depososit
Arc tholeiite (Sherridon–Meat Lake assemblage)						
4c	Meat Lake amphib.	Meat–Walton lakes area	Depleted MREE; concave-up N-MORB on normalized profile; strong negative Nb, Zr, Hf anomalies; high Th spike	Weakly contaminated	Extensional, mildly depleted to evolved arc (prominent subduction component)	Sherridon VMS deposits
4d	Walton Lake amphib.	Meat–Walton lakes area	Depleted MREE; concave-up N-MORB on normalized profile; strong negative Nb, Zr, Hf anomalies; high Th spike	Weakly contaminated	Extensional, mildly depleted to evolved arc (prominent subduction component)	Sherridon VMS deposits
Rhyolite (Sherridon–Meat Lake assemblage)						
5	Sherridon gneiss	Sherridon–Meat Lake area	Weak LREE fractionation; weak negative Eu anomaly; low Zr/TiO_2	Juvenile	Extensional, mildly depleted to evolved arc (prominent subduction component)	Sherridon VMS deposits
5a	Sherridon gneissic rhyolite	Meat Lake area	Weak–moderate LREE fractionation; weak negative Eu anomaly; low Zr/TiO_2	Juvenile to weakly contaminated	Extensional, mildly depleted to evolved arc (prominent subduction component)	Sherridon VMS deposits
Early sills and dikes						
7	Josland Lake sills	Reed Lake to Kississing Lake	Weakly elevated Th and LREE (contaminated?); flat to E- MORB element profile with weak or lacking HFSE anomalies of chilled margin phase; elevated Nb/Yb	Juvenile to weakly contaminated	Rifting arc	Nokomis and Evans Lake minor Au deposits ±PGE

Appendix 2, Figure 25e) and is chemically similar to back-arc spreading-centre dacite (Appendix 3, Figure 34d; Sinton et al., 2003). The Dickstone VMS deposit may be related to faulting and attendant hydrothermal fluid flow during the early stages of arc rifting. The less evolved flows of Storozuk basalt in the upper part of the Fourmile Island assemblage suggest a late influx of more primitive magma, derived from the same source as the more arc-like Preaston andesite and Dickstone rhyolite. The Storozuk volcanics possibly provide a transition from the arc-volcanic rocks of the Fourmile Island assemblage in the south (Syme and Bailes, 1996) to the more MORB-like compositions of Moody Lake basalt and Fussey Lake basalt. The MORB-like composition of Moody Lake basalt in the north and its persistent structural contact with the Burntwood Group suggest that the arc-rift basin opened in the direction of the future Kisseynew Domain and may have provided part of an early oceanic volcanic basement for the Burntwood Group. In this scenario, the volcaniclastic Yakymiw Formation forms the local rift-basin fill, which conformably overlies the Storozuk volcanics (Bailes, 1980a).

The Fussey Lake basalt, which shows only weak HFSE and LREE depletion, formed farther within the rift or back-arc basin, less under the influence of fluid-fluxed melting of the mantle wedge. It has a flatter N-MORB–normalized geochemical pattern than the basalts to the west (Appendix 3, Figure 34a). It is structurally isolated in the Burntwood Group and probably originated in a basin that lay between the Central Flin Flon and Snow Lake subdomains.

The conclusion of Syme et al. (1995) that the arc-volcanic rocks of the Amisk Collage in the Central Flin Flon subdomain formed separately from the Snow Lake assemblage was based on their different ages of structural amalgamation (Table 17) and is supported by the presence of the rift-margin volcanic rocks and the allochthonous Burntwood Group and Fussey Lake MORB-like basalt between them. The presence of the Josland Lake sills in the Snow Lake and Fourmile Island assemblages, as well as in the Fussey Lake basalt, however, suggests a more complex tectonic history.

The F_1 recumbent folds in these rocks (e.g., Yakymiw syncline) formed before 1886 Ma intrusion of the Josland Lake sills, possibly in the footwall of the obducted ocean floor whose remnants include the Elbow-Athapapuskow assemblage and the Reed Lake mafic–ultramafic complex. Such recumbent structures and even mega–sheath folds also occur beneath obducted ocean floor in the modern accretionary environment (Searle and Alsop, 2007). The accreted complex apparently continued to rift with the intrusion of the sills. During the D_3 onset of arc–back-arc reassembly, it was thrust over the Burntwood Group in the footwalls of the Morton Lake and Loonhead Lake faults (a–c′–d in Figure 15; restoration in Figure 43, back pocket).

6.1.4 Arc rifting and mafic intrusion

The Mikanagan and Josland Lake sills in the Central Flin Flon subdomain are correlated with each other and with sills in the Snow Lake subdomain, as well as with arc-rift basalt in these areas. The correlation is based on their distinctive Fe-rich composition, similar trace-element patterns and, where determined, their consistent age, which indicates widespread rifting in the early Flin Flon Domain. Very similar, highly fractionated sills and flows occur also in modern back-arc environments (i.e., Manus basin; Appendix 1, Figure 21b). The widespread occurrence of the sills in the Flin Flon Domain and the associated D_1 deformation (discussed in Section 4.1.3) suggest an early amalgamation of the entire domain that caused folding and a change to arc-rift magmatism. Splitting of the Flin Flon Domain appears to have occurred during protracted or repeated rifting and marginal-basin opening. This major tectonic episode involved 1) the ca. 1881 Ma Fe-rich Mikanagan Lake sills near Flin Flon, 2) arc-rift ferrobasalt conformably below 1885 Ma shoshonite at the top of the Flin Flon arc assemblage (Stern et al., 1995b), 3) similar arc-rift basalt at the top of the Snow Lake assemblage (Bailes and Galley, 1999, 2007), and 4) similar rock high in the section of Ponton Lake basalt. All of these units have a similar multi-element pattern (Appendix 3, Figure 35c–f) and, where known, similar U-Pb ages that overlap within error.

The chilled sill margins and basalt compositions of the rift-related rocks show previous Fe enrichment from fractionation in deeper magma chambers. Chilled Josland Lake gabbro has 15% FeO^t with a high Mg# (65). This suggests widespread sub-arc intrusion of high-Mg parent magma that experienced abundant olivine-crystal fractionation at depth before the emplacement of the sills and flows. The parent magma was enriched similar to E-MORB (Appendix 2, Figure 30d), such as the older Mg-rich basalt in the main (southern) part of the Elbow-Athapapuskow assemblage (Stern et al., 1995a) and the Ponton Lake basalt. The flows and sills must have formed from the same type of enriched mantle source, most likely within a single rifting plate. The Snow Lake subdomain probably separated from the Central Flin Flon subdomain during subsequent arc-terrane disruption that may also have led to the development or widening of the ocean floor that became part of the Kisseynew basin basement. Fault slices of similar enriched basalt, also formed by repeated rifting, are preserved around much of the Kisseynew Domain (Zwanzig et al., 1999; Zwanzig, 2000b, 2005), but most of the basin floor would have been consumed during closing of the basin and the resulting successor-arc magmatism. The presence of >1886 Ma enriched MORB-type basalt at the margin of the TNB and the prominent 1883 Ma Molson mafic–ultramafic dike swarm and ultramafic sills (Hulbert et al., 2005) suggests an early link between the Superior plate and arc terrains in the THO (see also Percival et al., 2005). Mantle upwelling or plume activity probably originated in an ocean basin (i.e., Kisseynew forerunner) that linked these terranes.

Arc-terrane disruption probably affected the paleogeography of the entire Reindeer Zone of the THO to such an extent that little can presently be recognized of the original arc configuration or polarity. This may have important implications for correlations within the orogen. The nearly identical evolution of the Flin Flon–Glennie and La Ronge–Lynn Lake subzones in the Reindeer Zone for a period of 130 m.y. opens the possibility that they formed in the same arc–back-arc system with similar metallogeny and high economic potential.

6.1.5 Early successor-arc magmatism

The early arc magmatism had resumed in the Kississing–File lakes area after a magmatic hiatus of about 10 m.y. following the early arc rifting (Figure 14). Dated intrusions have crystallization ages ranging from 1874 to 1860 Ma in the Kississing–File lakes area (Machado et al., 1999) and from 1876 to 1845 Ma regionally (Whalen and Hunt, 1994). These occur almost exclusively in the Central Flin Flon subdomain. The eastern subdomains (Snow Lake and Batty Lake) contain early successor-arc volcanic rocks, parts of the 1876 Ma Schist-Wekusko assemblage in the east and the apparently 1850–1855 Ma Sherridon–Meat Lake assemblage in the north. The overall geochemical and Nd-isotope characteristics of the early successor-arc plutons show a similarity to arc-plutonic rocks formed in the Early to Middle Ordovician from mantle, slab and crustal components, as in the Notre Dame Arc of the Appalachian Orogen (Whalen et al., 1997). The copious granitoid intrusion into the Amisk Collage and any attendant mafic underplating must have caused crustal thickening, uplift and erosion. The Sherridon–Meat Lake and Schist-Wekusko assemblages are marine and suggest little previous uplift in the north and east, compared to the rest of the Flin Flon Domain. Thus, a higher VMS potential was apparently preserved.

6.1.5.1 Early successor-arc plutonism

The oldest of the successor-arc plutons form a north- to northwest-trending belt along the eastern margin of the Central Flin Flon subdomain. They comprise the Gants Lake Batholith, with 1876 Ma granodiorite, in the south and the Sherridon–Hutchinson Lake Complex, with the 1873 Ma Ragged Lake granite and the 1874 Ma Star Lake tonalite, in the north. These plutons intruded the Elbow-Athapapuskow ocean-floor assemblage, as well as the arc assemblages in the Amisk Collage, and therefore postdate the collision and amalgamation of these assemblages. The calcalkaline chemistry of these and related plutons is consistent with the arc origin ascribed to them by Whalen et al. (1999). They have elevated LILE and negative Nb anomalies (Appendix 3, Figure 37) typical of volcanic-arc plutons formed by fluid-fluxed melting of the mantle wedge above a subduction zone but also involving a crustal melt component and possibly slab melts (Table 18). The low ε_{Nd} of the Ragged Lake granite indicates the local presence and melting of Archean crust during granite formation or pre–1873 Ma amalgamation of the early juvenile-arc terranes and Archean crustal fragments.

The 1874 Ma Star Lake pluton has the characteristics of adakite or slab melt (Appendix 3, Figure 37d). Similar rocks have been ascribed to the subduction of the relatively young ocean lithosphere, probably formed in the early juvenile back-arc basins (Whalen et al., 1999). The 1860 Ma Archie Lake pluton has some adakite-like characteristics, most likely produced by lower crustal melting. Such magmatism may have occurred during compression above a shallow slab in the emerged island arc. The D_2 deformation (summarized in Section 4.1.1) occurred at that time. An extensive literature review of adakite-like rocks (Richards and Kerrich, 2007) has concluded, however, that slab melting is rare and that processes of crustal melting, fractionation and fractional mantle melting can account for such compositions. The Star Lake and Archie Lake plutons most closely resemble Proterozoic TTG suites, as defined by Condie (2008). He suggested a significant melt component from existing mafic and felsic crust to form such rocks, in the THO probably from the earliest arc massif. Other plutons, such as the Jay lake granite, may have formed by predominantly crustal melting.

The Batty Lake trondhjemite gneiss is best explained by variable loss of partial melt from an arc-related trondhjemite. Although the isotopic age is approximate, it overlaps the preliminary age of the Sherridon–Meat Lake volcanic assemblage. These rocks also have a similar geochemistry. The trondhjemite is tentatively interpreted as a subvolcanic pluton related to the successor-arc volcanic rocks and may have provided a heat source for their widespread hydrothermal alteration and VMS deposits.

Thickening of the crust and conversion to a continental composition had a profound influence on the tectonics and metallogeny of the succeeding evolution of the Central Flin Flon subdomain. Successor-arc volcanism, there, was probably mainly subaerial and did not promote widespread VMS deposition in the west, nor are the adakitic compositions favourable for porphyry Cu deposits (Richards and Kerrich, 2007). The Batty Lake and eastern Snow Lake subdomains, conversely, formed in an environment similar to that of the early arc-volcanic rocks. The Batty Lake gneiss may have formed under a mainly oceanic-arc cover, with VMS deposits being preserved in the Sherridon–Meat Lake assemblage. Their geochemistry is very similar to that in the early Snow Lake arc assemblage and its synvolcanic plutons.

Table 18: Geochemical characteristics and tectonic origin of early successor-arc intrusions in the Kississing–File lakes area.

Unit	Intrusion	Location	Geochemical characteristics	Isotope evolution	Tectonic origin
Mafic–intermediate intrusions			**Arc tholeiite**		
8a	Martell Lake gabbro	South of Martell Lake	Elevated Th, fractionated LREE, negative HFSE anomalies		Mature arc
	Microdiorite	Ponton Lake			
Granitoid intrusions			**Calcalkaline**		
9a	Ragged Lake granite	Northeast of Puffy Lake	Strongly fractionated LREE, negative HFSE, Sr and Eu anomalies	Evolved	Mature arc
9c	Big Island granodiorite	Kississing Lake			
9b	Jay Lake granite	Jay Lake	Elevated REE, low La/Yb, negative HFSE, Sr and Eu annomalies		Possible arc rifting
9d	Fire Lake granodiorite	Northwest of Puffy Lake	High La/Yb, negative Nb and Ti anomalies	Weakly contaminated	Mantle wedge and lower crustal or slab melt
9d	Archie Lake pluton	West of Puffy Lake			
9e	Star Lake tonalite gneiss	Nokomis–Kississing lakes	Adakite-like: high Al, Rb and La/Yb, no Eu anomaly		Lower crustal or slab melt
9f	Batty Lake gneiss	Batty Lake	Trondhjemitic: low La/Yb, elevated Ba, Nb and Ti trough		Mantle wedge and crustal or slab melt

6.1.5.2 Successor-arc or earlier volcanism and sedimentation

The Sherridon–Meat Lake arc assemblage occurs along the present northern margin of the Flin Flon Domain (Figure 8a, back pocket), both at the deepest level (Sherridon structure) and at a high level (Walton Lake nappe; Figure 43, back pocket). The separation of the Walton Lake nappe from the Sherridon structure by the Burntwood Group may have taken place early (D_3) if the nappe had a history of thrust transport into the Kisseynew basin as a klippe, as suggested in Section 4.1.5. This assemblage, with a preliminary age of 1850–1855 Ma, apparently formed in the final stage of early successor-arc magmatism (stage 2). This age, however, is not entirely resolved and the high proportion of felsic rocks, geochemistry and widespread alteration of the assemblage suggest that it may be part of the Snow Lake assemblage, which formed during stage 1. Non-arc (high-Ti) amphibolite with fractionated REE in the Sherridon area is interpreted as late sills that cannot be related to the older Josland Lake sills. They are similar, instead, to relatively young synkinematic dikes and do not provide a minimum age for the Sherridon–Meat Lake assemblage. The geochemistry of samples from this assemblage in Appendix 4, Table 10 (DVD) and Halo Resources Ltd. exploration data indicate a derivation by high-grade metamorphism of a strongly bimodal volcanic suite dominated by felsic rocks (Appendix 1, Figure 19a). The mafic units (Meat Lake and Walton Lake amphibolite) have a strong arc-tholeiite signature with strongly depleted middle REE, Zr and Hf. It is noteworthy that they are very similar to the Threehouse basalt, which overlies Photo rhyolite in the upper part of the Snow Lake assemblage (with calcalkaline geochemistry similar to that of the Sherridon felsic gneiss). These rocks at Snow Lake are ca. 1890 Ma in the lower part and there is a broad similarity throughout the sequences, consistent with the early volcanism (stage 1). Thus, the possible correlation of the Sherridon–Meat Lake assemblage with the Snow Lake assemblage contradicts the preliminary age of the former and needs to be resolved.

Regardless of age, the felsic-dominated bimodal succession at Sherridon and related clastic rocks probably formed on thickened arc crust but as submarine volcanoes in an oceanic-arc environment. This crust, which was probably a major contributor of felsic melt, was juvenile Paleoproterozoic: most ε_{Nd} values for the Sherridon–Meat lake assemblage are high, with minor Archean crustal contamination evident only in a single analysis from the Meat Lake area. The volcanic area was probably undergoing mild extension and local resedimentation of felsic tuff during synvolcanic block faulting with abundant flux of hydrothermal fluids. This led to widespread alteration of the felsic rocks and deposition of base metals. Such synvolcanic intrusions as the Batty Lake Complex and Defender Lake dome, if these are indeed coeval, may have acted as a source of heat (see above). Volcaniclastic detritus and felsic ash may have formed some of the garnet-bearing, slightly more biotite-rich felsic units in the Sherridon gneiss that are interpreted as metasedimentary. Such detritus may also have formed the Bob gneiss (unit 10a) and rocks in the Yakushavich Island–Collins Point area, as well as a gradational base of the Burntwood Group on the north side of the Walton Lake nappe. An older age for some of these rocks would require an unconformity or fault contact.

Bob gneiss (unit 10a), north of Sherridon, is a possible equivalent of the Burntwood Group, with its similar detrital zircons, geochemistry and ε_{Nd}. Petrography of this unit is similar to some of the Burntwood Group surrounding the Walton Lake nappe, although its thick indistinct layering is unique. This possibly formed very close to a single neovolcanic source (Sherridon–Meat Lake assemblage?) without resedimentation, or possibly as high-density turbidite with thick amalgamated beds.

6.1.6 Marginal basin evolution and late successor-arc magmatism

Detrital zircons in the widespread Burntwood Group turbidites show that a major influx of clastic sediments to the Kisseynew basin began about 1855 Ma from the adjacent arc(s) during the end stage of early successor-arc magmatism (Figure 14; David et al., 1996). The locally pebbly, proximal (upslope) facies of the Burntwood Group on Reed Lake, south of the File Lake area (Syme et al., 1995), probably represents a point source or upper fan. The northward gradation through Morton Lake and File Lake into Dow Lake, with decreasing bed size and increasing proportion of mudstone, gives the local direction of sediment transport, away from the Flin Flon Domain into the Kisseynew Domain. An unconformity at the base of metasedimentary rocks that are correlated with the Burntwood Group, in the La Ronge Domain in Saskatchewan, was also dated as 1855 Ma. It overlies metasedimentary rocks formed during the early volcanism (Maxeiner and Sibbald, 1995). The unconformable succession comprises interlayered crossbedded arenite, conglomerate and greywacke, and may represent a proximal facies from another point source approximately coeval with the earliest turbidite at the margin of the Flin Flon Domain. Such sources have been interpreted by Bailes (1980a) to form coalescing fans in the Kisseynew Domain that continued to collect neovolcanic and eroded detritus from the active arc and its basement. Deposition lasted at least until 1842 Ma (the age of the youngest detrital zircon dated by Machado et al., 1999).

The terrestrial sedimentation and coeval subaerial volcanism of the Missi Group occurred in a series of sub-basins during block faulting along the margin of the Flin Flon–Glennie Complex facing the Kisseynew Domain. The prograding nonmarine facies (Missi Group), which lies unconformably on the Amisk Collage and early plutons in the Central Flin Flon subdomain, extends through a shoreline facies of quartzite and pelite to crossbedded arenite, and stretches over the turbidite fans in the Kisseynew Domain. This facies change is also marked by the appearance of felsic tuff, including the apparently reworked dacitic tuff that yielded 1848 Ma zircons. This is interlayered with the basal metaconglomerate, thus dating the base of the Missi Group lying on the Amisk Collage. The sedimentation and magmatism of the Missi Group lasted beyond the crystallization age of rhyolite tuff higher in the succession (dated as 1836 Ma in Chickadee rhyolite east of Wekusko Lake) and probably to the age of the 1832 Ma Nelson Bay dome that has been interpreted in this report as caldera fill with possible intrusive resurgence. Burned-over portions of the Nelson Bay gneiss dome examined in the 1980s included some fragmental units that are consistent with this interpretation. The proximal part of the Missi Group in the basin at Flin Flon contains 1847 Ma zircons partway up in the section, and was deformed and then cut by 1842 Ma dikes (Section 4.1.3). It was clearly older than much of the distal part of the Missi in the Kississing–File lakes area. The Burntwood Group turbidite deposition, lasting beyond the 1842 Ma age of the youngest detrital zircon, was coeval with the deposition of the Missi Group at Flin Flon and partly coeval with the deposition of the distal Missi in the Kississing–File lakes area.

These deposits have a similar trace-element geochemistry and ε_{Nd}, which confirms their similar provenance from the adjacent arc massif (Zwanzig et al., 2007; Böhm et al., 2007). Major-element geochemistry has suggested that the Missi Group was derived from a mature-arc to active continental-margin environment (Zwanzig, 1990). Contents of SiO_2 and K_2O that are commonly higher than those of the Burntwood Group (Zwanzig, 1990) are attributed to the large volume of reworked felsic tuff proposed to occur in the Kississing–File lakes area, as well as the change from erosion of more mafic supracrustal rock to that of the underlying felsic plutons. A further change, marked by the decrease in biotite content, represents the separation and basinward transport of clay in the suspended load of sediments. The base load, stored preferentially in the Missi Group, would have contained more quartz, feldspar and heavy minerals.

6.1.6.1 The Kisseynew basin problem and the THO tectonic model

A proximal part of the Missi Group at Flin Flon has been interpreted as molasse (Stauffer, 1990). Accordingly, the coeval Burntwood Group would represent the associated orogenic flysch. However, whether that flysch occupied a back-arc, inter-arc or fore-arc position has been under discussion (Zwanzig, 1997). The Missi-age volcanic belt occurs at the basin margin between the shallow- and deep-water facies. The belt has been interpreted as a volcanic arc developed at the active margin of the older arc massif (Zwanzig, 1999). Sediment transport and early thrusting was from the Flin Flon Domain toward the Kisseynew Domain (Zwanzig, 1997, 1999, this report). The direction of subduction beneath the Flin Flon–Glennie Complex during this stage of arc evolution in the THO has been interpreted as southerly (Ansdell et al., 1995) or southwesterly (Zwanzig, 1999), but northerly by Hollings and Ansdell (2002). The sediment filling of the Kisseynew Domain has generally been ascribed to the opening of a late back-arc basin during northerly directed subduction of ocean floor leading the Sask craton (Ansdell et al., 1995; Ansdell, 2005; Corrigan et al., 2005). Continuing northerly directed subduction beneath the Flin Flon and Kisseynew domains, possibly above a shallowing slab, is inferred from the geochemistry and location of synkinematic granitoid intrusions in the Kisseynew Domain by Hollings and Ansdell (2002). Collision has been inferred to have occurred between the Flin Flon–Glennie Complex and the Archean Sask craton during the ca. 1840–1830 Ma Missi-age magmatism (Ansdell et al., 1995; Ansdell, 2005) or ca. 1850 Ma during the onset of Missi Group sedimentation and magmatism resulting from subduction directed beneath the Flin Flon Domain, away from the Kisseynew Domain (Zwanzig, 1999).

The late-stage back-arc model encounters problems, particularly in view of recent data from the southeastern part of the THO in Manitoba. A model must explain the following observations and inferences:

- The best estimate of the age of basaltic back-arc basement (BABB) in the Kisseynew Domain and Flin Flon margin is 1.89 Ga in the south (this report) and on the North flank (Zwanzig, 2000b). Ocean-floor–type basalt in contact with the Burntwood Group is probably older still along the Superior Boundary Zone (Zwanzig, 2005a). This indicates that early ocean floor developed in the Kisseynew Domain about 50 m.y. before sediment filling by the Burntwood and Missi groups. In other words, the sediments were shed onto the margin of an existing ocean basin trapped during continental collision.
- Arc rifting in the Flin Flon Domain continued at least to the 1886–1881 Ma intrusion of the Mikanagan and Josland Lake sills and basalt of similar composition, which postdated the earliest arc- and back-arc–fragment amalgamation (Amisk Collage) in the Central Flin Flon subdomain and probably caused continued widening of the precursor to the Kisseynew basin.
- The onset of renewed arc magmatism occurred ca. 1876 Ma, possibly involving a new subduction zone with a westerly dip, as indicated by arc magmatism north and west of the Kisseynew Domain. Subduction of the plate that carried the Sask craton from the southwest continued as well, and led to collision.
- The high sediment flux in the Kisseynew Domain, starting ca. 1855 Ma, represents uplift and early collision of the Flin Flon–Glennie Complex in the southwest with a full thickness of crust (Sask craton) in the subduction zone and deposition of molasse and flysch, possibly over the margin of the subducting plate attached to the Superior craton, similar to a Phanerozoic example documented by Brown and Spadea (1999).
- Uplift was orogen wide and continued during deposition of the Missi Group and equivalent meta-arenite on all flanks of the Kisseynew Domain, as well as on the eastern external zone of the THO, thus representing the onset of terminal continental collision that also involved the Superior craton.
- Narrow belts of Burntwood Group turbidite (and equivalents) occur between blocks with disparate arc and back-arc tectonic origins, and indicate that the final (re)assembly of the Flin Flon–Glennie Complex occurred during terminal collision and closing of the Kisseynew basin. Assembly included parts of the Amisk Collage, the coeval Northeast Reed and Snow Lake assemblages, and successor-arc magmatic rocks and the coeval sedimentary rocks (Burntwood and Missi groups).
- The Burntwood Group separates all of the arc–back-arc domains from the Superior Boundary Zone but contains only detritus from the arc-related terranes, without a known contribution from the Superior craton (Zwanzig et

al., 2007). That craton must have been distant or on a low-lying subducting plate.

- Block faulting and 1848–1832 Ma continental-arc to bimodal magmatism occurred along the previously uplifted margins of the volcanic domains and helped to feed the Burntwood and Missi sedimentation, variously in different sub-basins.
- Early (D_3) thrusting of the arc margin (Moody Lake basalt) was over Burntwood Group turbidite toward the Kisseynew Domain (Zwanzig, 1999 and this report), consistent with subduction from the Kisseynew side.
- The Sask craton had already underthrust the Flin Flon–Glennie Complex from the south by 1824 Ma (Ashton et al., 1999, 2005) during advanced stages of collision (D_5).
- Seismic reflectors dip east and northeast in the western THO in Manitoba, suggesting a final crustal stacking order downward from northeast to southwest.

These relationships suggest early (ca. 1886 Ma) intraoceanic arc–back-arc amalgamation and nearly coeval rifting. The onset of terminal collision is marked by the ca. 1855 Ma unconformity at the base of the Burntwood Group in Saskatchewan. This was followed by the 1850 Ma termination, or significant reduction, of arc magmatism in the La Ronge Domain, but not on the northern and eastern margins of the Flin Flon Domain nor on the margin of the Lynn Lake–Leaf Rapids Domain. The relationships further suggest a northwesterly direction of subduction (present co-ordinates) during late successor-arc magmatism from the Kisseynew Domain but toward the Flin Flon–Glennie Complex and the Lynn Lake–Leaf Rapids Domain. The ocean slab was apparently attached to the approaching Superior craton (Zwanzig, 1999; White et al., 2000) and its pericontinental margin, the Northeast Kisseynew subdomain (Zwanzig, 2008). The permissible paleogeographic evolution of the Reindeer Zone and surrounding Archean cratons is presented in six schematic diagrams that represent the 130 m.y. evolution of the THO (Figure 44).

The sedimentary-facies changes and the inferred direction of sediment transport of the Burntwood and Missi groups are consistent with westerly directed subduction and deposition in a fore-arc and trench environment along the (present) northeastern margin of the Flin Flon–Glennie Complex. The age of deposition and detrital zircons, as well as the geochemistry of the sedimentary rocks and recognizable clasts, indicate that they were derived from the active margin and earlier arc to successor-arc massif of the Flin Flon–Glennie Complex. The Corley Lake mudstone, near the top of the Burntwood Group (Section 3.5.1.1), may represent the trench-fill stage; and the coarser overlying turbidite, locally grading into pebbly shallow-water deposits, may represent the prograding fan-delta to nonmarine facies at the margin of the Flin Flon Domain. The sedimentology of the Burntwood Group and its relation to an adjacent arc terrane are very similar to those of the Neoarchean sedimentary rocks of the McKellar Harbour Formation and Quetico subprovince south of the Wabigoon greenstone belt in northwestern Ontario (Fralick et al., 2006). Those have also been interpreted as fore-arc basin and trench fill in an accretionary wedge fed from the arc terrane. A similar Cretaceous analogue is the Matanuska Formation of turbidite deposited on a trenchward-dipping ramp directly after accretion of Wrangel composite (arc) terrane to central Alaska (Trop, 2007). Important sources in Alaska were older and rapidly unroofed very young plutons, as is also indicated in the Burntwood Group. Coeval conglomerate occurs unconformably on the Alaskan arc, and fluviolacustrine strata (cf. Missi Group) occupy basins farther inboard. According to the geochemistry and Nd-isotope data of the late successor-arc igneous rocks, abundant juvenile sediments became available at a subduction zone facing the Flin Flon Domain only during the Missi-age collisional stage. Consequently, the first sediment-dominated accretionary wedge in the Kississing–File lakes area may have involved the Burntwood Group.

A late west or northwest subduction implies that the present shape of the Kisseynew Domain is a result of terminal collision and continued convergence, not back-arc–basin opening (Zwanzig, 1999). The domain boundaries and units north of the major structural bend in the Kississing–File lakes area apparently underwent large-scale sinistral rotation, overturning and overthrusting during syn- to postcollisional shortening (D_4–D_6, Figure 14; Figure 44). West-directed thrusts occur in front of the Superior craton and south-directed thrusts in front of the escaping Kisseynew Domain. In this scenario, terminal continental collision was between the Superior and Rae-Hearne cratons, trapping between them the arc–back-arc domains, as well as the Sask craton and other Archean crustal fragments. Convergence continued during closing of the Kisseynew basin and deformation of its rocks up to the end stage of high-grade metamorphism (ca. 1790 Ma) and subsequent (1780–1770 Ma) intracontinental transpression. The present crustal architecture of the Kisseynew and Flin Flon domains was formed by the late convergence of the three surrounding Archean cratons with the Sask craton at the deepest level, followed by the Superior craton and overrun by previously accreted domains on the Rae-Hearne craton (Figure 6).

6.1.6.2 Missi-age volcanism and plutonism

Latest successor-arc volcanism occurred in the Missi Group in the Kississing–File lakes area ca. 1848–1832 Ma, and arc plutonism in the Snow Lake subdomain ca. 1840–1830 Ma. Although some plutons remain undated, and some of the younger intrusions occur also in the Central Flin Flon subdomain, the vast majority of the Missi-age igneous rocks occur only in the northern and eastern parts of the Flin Flon Domain, facing the Kisseynew Domain. The line of arc magmatism fringing the Kisseynew basin suggests subduction directed toward the Flin Flon–Glennie Complex. Although ca. 1.84 Ga bimodal intrusions occur near Flin Flon, these are probably back-arc intrusions without evidence of arc volcanism. The collisional tectonics during the Missi-age magmatism and the subsequent magmatic hiatus, as well as the geochemical changes of later magmatism, indicate an end stage of oceanic lithosphere subduction.

Missi Group sedimentation and late arc volcanism occurred in a number of nonmarine sub-basins, the largest of which extended along the present northern and eastern margins of the Flin Flon Domain. Stratigraphic analysis using tentatively restored sections (Figure 8, back pocket) and maps (NATMAP

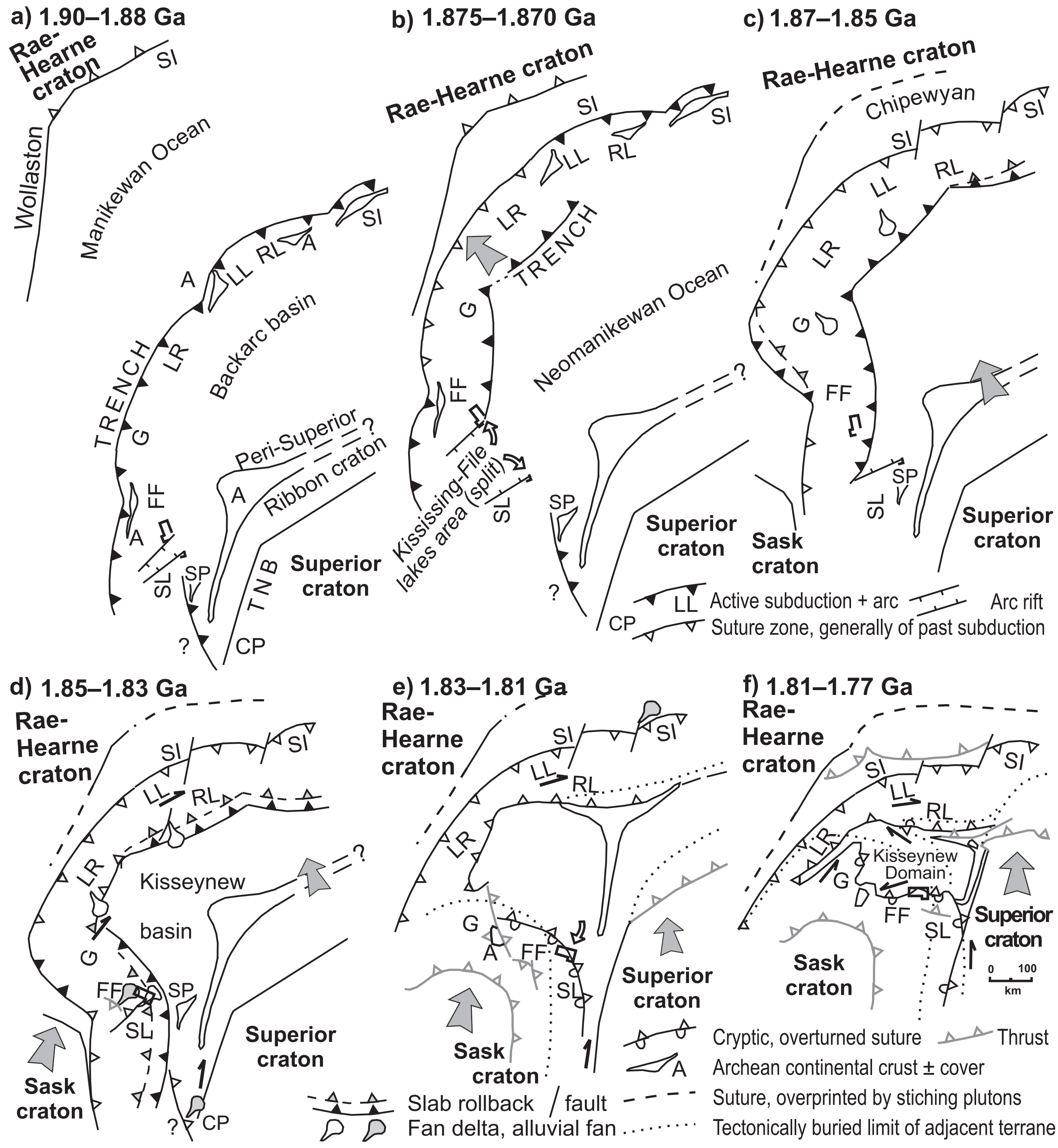


Figure 44: *Stages of permissible paleogeographic evolution of the Reindeer Zone (Trans-Hudson Orogen [THO] internides) and surrounding Archean cratons. The assumption that all greenstone belts were part of a single oceanic arc or a number of contiguous arcs is based on their identical products of the same age and similar tectonic events for a period of 70 m.y. (a–d), as well as the final near continuity (f). Subduction polarity in the La Ronge–Lynn Lake–Rusty Lake–Southern Indian arc–back-arc domains* ***(a)*** *is after Corrigan et al. (2005). The early polarity in the Snow Lake–Flin Flon subdomains and Glennie Domain are after Ansdell (2005) but rotated 90° to conform with the structural trends in the best preserved part of the Flin Flon Domain (Figure 5). The subduction flip* ***(b)*** *is inferred to be in response to collision with Rae-Hearne craton, causing the termination of early arc magmatism and the onset of early successor-arc magmatism. Northwest-directed subduction* ***(c)*** *is indicated by 1865–1850 Ma arc magmatism from La Ronge to Southern Indian and craton-stitching batholiths to the northwest. Late successor-arc magmatism* ***(d)*** *is by continued west-directed subduction, producing Missi-age volcanic rocks and abundant intrusions at Snow Lake and similarly aged plutons from La Ronge to Southern Indian. Overturning, inversion and rotation of major structures on a crustal scale* ***(e)*** *is in front of the colliding Sask craton. Final crustal architecture (nearly to scale) is the result of continued convergence of the cratons* ***(f)****. Archean crust–subcontinental mantle beneath much of the Reindeer Zone is indicated by Nd data. Small polygon indicates the Kississing–File lakes area. Domain abbreviations: A, Archean; CP, Clark pluton; FF, Central Flin Flon subdomain; G, Glennie Domain; LL, Lynn Lake Domain; LR, La Ronge Domain; RL, Rusty Lake Domain; SI, Southern Indian Domain; SL, Snow Lake subdomain; SP, Saw Lake pluton.*

Shield Margin Project Working Group, 1998) shows different successions in the various potential sub-basins. Some of these comprise thick successions of almost entirely siliciclastic rocks, whereas others contain widely differing thicknesses of volcanic rocks among mainly sedimentary rocks. Significant thicknesses of metarhyolite occur at Cleunion Lake and File Lake. Other potential sub-basins, such as those north of Kisseynew Lake and east of Wekusko Lake, are filled almost entirely with mafic–intermediate volcanic rocks. Boundaries between sub-basins have been interpreted as thrust faults (e.g., Ansdell et al., 1999) or have been obscured by polyphase deformation and high-grade metamorphism. The abrupt changes in stratigraphy, thickness and volcanic composition, however, indicate that the sub-basins are related to syndepositional block faulting. The bimodal and other geochemical characteristics of the Missi Group magmatic rocks indicate an extensional tectonic environment consistent with the proposed block faulting. The thrust-related displacement on the sub-basin–bounding faults is interpreted as the result of inversion during advancing collision tectonics. This history is consistent with the synclinal structure in most of the belts of basinal sedimentary rocks.

Characteristics of the Missi-age magmatism are summarized in Table 19. The complex geochemistry, with medium to high K_2O, variably fractionated REE and generally high LILE, of the Missi-age calcalkaline magmas stems from multiple sources and is best explained by a long melting column above a steep, deeply subducted slab. Such a melting column can also produce the high-Mg content of the Evans Lake basalt. Contamination by juvenile crustal sediments, quite possibly the Burntwood Group, may have contributed to the arc magmas. A melting, assimilation, storage and homogenization (MASH) process (Hildreth and Moorbath, 1988) at the base of the juvenile-arc crust may have contributed considerably. Slab rollback during the onset of collision probably caused the crustal extension that formed the various basins and sub-basins (Figure 44d). Repeated subsidence, which produced the fining-upward sedimentary cycles, was probably related to lithospheric thinning with rising asthenosphere that may have promoted the change to mafic tholeiitic magmatism in the two sub-basins (Kississing River and east of Wekusko Lake) filled with predominantly volcanic rocks. The best preserved high-K intermediate–tholeiitic mafic cycle is the eastern sub-basin,

Table 19: Geochemical characteristics and tectonic origin of late successor-arc and basinal rocks in the Kississing–File lakes area.

Unit	Name / lithology	Type location	Geochemical characteristics	Isotope evolution	Tectonic origin
Burntwood Group					
11a	Greywacke-mudstone	Morton Lake–File Lake			Marginal basin, or fore-arc and trench?
11b	Greywacke gneiss	Stephanuik Lake			
11c	Greywacke migmatite	Nokomis Lake			
Missi Group					
12	Conglomerate	North of Lobstick Narrows			Syncollisional extension
13	Arenite				
14a	Kississing River basalt, gabbro	Logging road north of Kisseynew Lake	High-alumina, low-K tholeiite; slighly elevated HFSE for an arc tholeiite, transitional to continental basalt	Juvenile	Extensional, emerged arc in island arc–continent collision
14b	Evans Lake basalt	Evans Lake	High-Mg calcalkaline; high La/Yb; strong negative Nb and Ti anomalies	Juvenile	
14c	Jungle Lake basaltic andesite	Pufffy Lake	Moderate- to high-K calcalkaline; high La/Yb; strong negative Nb and Ti anomalies	Juvenile	
14d	Puffy Lake dacite	West of Ragged Lake	High Rb, La/Yb; negative Sr anomalies		
14e	Cleunion rhyolite	South of road past Cleunion Lake			
14f	Lobstick rhyolite	Lobstick Narrows	High Sr, Ba, La/Yb; depleted HREE	Juvenile	
Late successor-arc felsic intrusions (Missi age)					
15a	Nelson Bay pluton	File Lake	Like Cleunion rhyolite and Puffy Lake dacite	Nearly juvenile	Extensional, emerged arc in island arc–continent collision
15b	Norris Lake pluton	West of File Lake			
15c	Barron Lake granite	Southeast of File Lake			
15d	Ham Lake pluton	File Lake	Similarities with Lobstick rhyolite; less depleted HREE		
15e	Thunderhill Lake porphyry	Thunderhill Lake	Like Lobstick rhyolite		
Late successor-arc mafic–intermediate intrusions (Missi age)					
15g	Reed Lake diorite	Woosey Lake	Trace elements like Evans Lake basalt, majors like Dow Lake quartz diorite		
15h	East Big Island gabbro	Kississing Lake	Similar to fractionated Evans Lake basalt		

possibly lying at the most outboard part of the volcanic arc, similar to modern Santorini in the Aegean Sea (Huijsmans et al., 1988). This paleogeography is consistent with the proposed subduction from the Kisseynew Domain beneath the Flin Flon–Glennie Complex, and also with the development of the Reed Lake 'pluton' as a back-arc core complex and a giant retro-arc batholith (Cormorant Batholith in Figure 5) farther south. Rapid erosion of subaerial felsic tuff led to the deposition of abundant sand–clay that is preserved in large parts of the Missi Group. The total volume of felsic magmatism was apparently considerable and may have represented a modest ignimbrite flare-up and granite bloom. This short (1836–1832 Ma) magmatic pulse may have been caused by final rapid subduction (Jordan et al., 2001) and/or slab fragmentation and break-off.

Epsilon-neodymium data (Appendix 2, Figure 29f; Whalen et al., 1999, Figure 16) are consistent with purely juvenile-melt sources for the Missi-age magmatism by subduction of ocean floor with mainly juvenile sediments from the side of the Kisseynew Domain rather than from the impinging Sask craton from the southwest. The early volcanic rocks and early arc and early successor-arc plutons, on the other hand, contain members with Archean components that were possibly derived from the subducting plate that carried the Sask craton and its probable foreland basin of mixed detritus. The change to the purely juvenile-arc magmatism can be explained by the subduction flip and eastward subduction step-back between 1860 and 1850 Ma. The Sask craton may have docked against the Flin Flon–Snow Lake arc segments at that time and the south branch of the Kisseynew basin may have closed (Figure 44c, d). This would have caused uplift and faulting of the Flin Flon–Glennie Complex in the southwest and subsidence along its margin with the Kisseynew Domain—a scenario that is supported by the nature and direction of the sedimentation and volcanism in the eastern sub-basins of the Missi Group.

The geochemistry of the late successor-arc magmatism may be explained by its generation on thicker crust above a deeper slab (e.g., Churikova et al., 2001) and involvement of crustal melting (e.g., Goodwin et al., 2006). Most likely, a multisource MASH process took place at the base of the crust (Hildreth and Moorbath, 1988; Richards, 2003). This type of process at an active margin is strongly supported by isotopic data from Phanerozoic continental arcs (e.g., Ducea and Barton, 2007) and would have yielded juvenile melts at the margin of the Flin Flon–Glennie Complex, which contains little Archean crust.

6.1.7 Orogeny: synkinematic magmatism, continental collision and deformation

Orogenesis progressed in the Central Flin Flon subdomain, caused by collision with the Sask craton in the south (Ansdell et al., 1995). This is consistent with the formation of local intermountain basins filled with Missi Group molasse that contains abundant conglomerate and pebbly sand, derived from the emerged arc terrane. Such basins occur in the backs of Phanerozoic continental-margin arcs (e.g., Trop, 2007). Subsequent northerly thrusting (D_3) of the Amisk Collage over the Missi Group has been documented near Flin Flon and is constrained to have taken place before 1842 Ma (Gale at al., 1999; R-L. Simard et al., work in progress, 2010). Late Missi-age thrusting (possibly after 1832 Ma magmatism) is indicated for the Kississing–File lakes area as northeasterly or easterly transport of Amisk Collage over the Burntwood Group at the Kisseynew basin margin. The intrusion of the Missi Group by plutons during ongoing sedimentation and volcanism indicates rapid tectonic burial. The presence of young magmatic zircons (e.g., 1837 Ma) indicates rapid uplift. Pressure-temperature modelling of the Kisseynew Domain and the Flin Flon Domain margin supports deep burial and subsequent heating of the crust (Growdon, pers. comm., 2008).

A change in fold vergence and thrusting toward the west (represented by D_4) is interpreted to have produced the crustal-scale easterly dip that is prominent in large-scale east-west structure sections (Figure 13a; Zwanzig, 1999) and in the regional seismic profiles (Hajnal et al., 1994). This is interpreted to have been the result of collision between the Superior craton and the THO internal zone (e.g. Clark pluton [CP], Figure 44e; White et al., 2002). Mildly alkaline to sanukitoid plutons that cut across the boundary between the Thompson Nickel Belt (TNB) and the Northeast Kisseynew subdomain have been dated at 1836–1835 Ma (Zwanzig et al., 2003), indicating that the Superior craton collided with the Kisseynew Domain before that time. This is interpreted to have led to the reversal in vergence and the transposition of the crustal structure. The lower crust and mantle beneath the southern TNB may have inherited their northeasterly dip from early continental-arc magmatism (Figure 44a; Percival et al., 2005).

The ca. 1830 Ma intermediate–ultramafic Touchbourne intrusive suite (TIS) in the Central Kisseynew subdomain and the similar Duval Lake sills on the South flank of the Kisseynew Domain have been interpreted to represent a change from syncollisional-arc to postcollisional magmatism because their chemistry suggests formation by dehydration and melting of enriched mantle that had a previous history of subduction, as well as by heating of the lower crust through rising asthenosphere. The low ε_{Nd} of Dow Lake quartz diorite indicates melting of Archean crust that may have been emplaced in the form of the Sask craton at this time. Thus, much of the juvenile Paleoproterozoic crust presently exposed in the Kississing–File lakes area may have become allochthonous at this time and come to lie on Archean crust and mantle. Further crustal melting in the THO involved sporadic participation of the Archean infrastructure.

6.1.7.1 Syn- to postcollisional magmatism and metamorphism

Another result of this investigation is the description of, and deciphering the probable origins of, the abundant late granite to tonalite sheets and rare mafic dikes that extend from the Kississing–File lakes area into the Central Kisseynew subdomain, where they are joined by stocks, laccoliths and batholiths (Table 20). These intrusions are characterized by the general absence of coeval mafic or intermediate igneous phases and by their weakly peraluminous and LILE-enriched, HREE-depleted composition. Their origin has previously been ascribed to the latest arc magmatism (Hollings and Ansdell, 1992), or to melting of tectonically thickened crust that was sufficiently radioactive to produce a high internal heat flow (White, 2005).

Table 20: Geochemical characteristics and tectonic origin of syn- to postcollisional intrusions in the Kississing–File lakes area.

Unit	Intrusion	Type location	Geochemical characteristics	Isotope evolution	Tectonic origin
Intermediate–ultramafic intrusions			Weakly peraluminous, arc-like inheritance		
16	Burntwood Lake tonalite	Track to Burntwood Lake	Medium-K calcalkaline, strong negative Nd, Ti anomalies; mafic–ultramafic trend toward high La/Yb, Hf and Zr; intermediate–felsic rocks have high La/Yb, moderately high Sr/Y and concave-up REE trend	Weakly Archean contaminated	Probably syncollisional mid- to lower crustal and mantle melts
16a	Touchbourne suite	Burntwood Lake			
16b	Duval Lake sills	Road to Duval Lake			
16c	Dow Lake quartz diorite	West of Dow Lake	High-Fe, meduim-K tholeiitic, trend toward high La/Yb, Ba, P, Ti, Hf, Zr, and REE	Evolved (high Archean REE component)	Probably late-collisional contamin-ated or Archean-mantle+crustal melt
17a	Weldon Bay diabase	Weldon Bay	High-Fe tholeiite (crystal fractionated or low melt fraction); weakly declined, straight MORB-normalized profile; elevated Nb/Yb (similar to E-MORB or continental basalt)		Postcollisional decompression mantle melts
	Cree Lake gabbro	Cree Lake			
17b	Pyroxenite	Cree Lake			
Tonalite–granite intrusions			Weakly peraluminous, arc-like inheritance		
18a	Highrock Lake Batholith (tonalite–granodiorite)	Highrock Lake	Medium- to high-K calcalkaline; weakly peraluminous; strong negative Nd, Sr, Ti anomalies; moderate negative Eu anomalies; high Rb, La/Yb.	Moderately Archean contaminated	Probably postcollisional with high crustal melt component
18b	Flatrock Lake granite	Highrock–Burntwood lakes			
Leucogranitoid sheets and pegmatite			Weakly peraluminous		Postcollisional
19, 19c	Synkinematic granodorite	West of Duval Lake	Moderate- to low REE and La/Yb, negative Eu anomaly; Al-saturation like Burntwood mesosome		Probable S-type anatectic melt
	Leucosome (LS2)	Nokomis Lake			
19c	Synkinematic granite to tonalite	SE of Duval Lake SW of Dow Lake	Like Burntwood Lake tonalite (adakite-like)		Probable lower crustal ± mid-crustal melts
	Leucosome (LS3)	Nokomis Lake			
19b	Synkinematic leucogranite	Kisseynew Domain South flank	High La/Yb, low Yb, variable Eu anomalies, low Sr	Moderately Archean contaminated	Crustal melts
	Late-kinematic biotite granite	Nokomis Lake	Very high La/Yb, depleted HREE		
19a	Pegmatite	Nokomis Lake	Low- to moderate REE and La/Yb, positive Eu anomaly		

A third model, which is preferred here, is that the heat was derived from anomalously hot mantle (Gordon, 1989) but not directly from magma. High heat flow was possibly a result of earlier lithospheric extension and subsequent or coeval (Kraus and Menard, 1997) crustal thickening. Rising heat flow occurred widely, nearly simultaneously throughout the THO, aided by rising granite magma (White, 2005) that produced the late (ca. 1814 Ma) granite bloom, migmatization and rare premetamorphic tholeiitic magmatism (e.g., Weldon Bay gabbro and Cree Lake intrusions).

Most late granitoid rocks have HREE depletion that indicates garnet in the residuum. The various units have different trace-element patterns similar to small-fraction melts of leucosomes from the granulite–amphibolite facies transition in other migmatite complexes (e.g., Fornelli et al., 2002). The more discrete, sharp-walled sheets in the Kississing–File lakes area indicate that the melts migrated into the somewhat lower grade roof of the migmatite complex. The larger plutons in the Central Kisseynew subdomain with similar geochemistry to the sheets in the south occur also at relatively high structural levels (Zwanzig, 1990). The larger batches of melt formed widespread homogeneous phases in the Highrock–Burntwood lakes area. Granodiorite and tonalite, including LS3 leucosome (Section 5.10.4.2), with flatter, concave-upward REE profiles (Appendix 3, Figure 42c) suggest relatively high-pressure, fluid-present melting of mainly quartz and plagioclase in a mid-crustal source and/or involvement of amphibole from melting of the arc massif deeper in the crust.

6.1.7.2 Final convergence and block faulting

Large parts of the volcanic domains were affected by only medium- to low-grade metamorphism. These areas show low-pressure metamorphic mineral assemblages (commonly 3.5 kb; Digel and Gordon, 1995) and may be interpreted as the upper crustal carapace that elsewhere was uplifted and eroded during the unroofing of some of the higher pressure (5–8 kb) domains, such as the Namew Gneiss Complex and the Northeast Kisseynew subdomain. Seismic-refraction and -reflection profiles show the low-grade south side of the Flin Flon Domain as a block that was obliquely downthrown in the south along the Berry Creek Fault Zone, relative to the adjacent Namew Gneiss Complex (Figure 5; Leclair et al., 1997). This late- to post-tectonic faulting with north-northeast extension and west-northwest compression formed the last stage of the D_6 deformation that occurred during intracontinental transpression, probably triggered by distal forces, possibly from the Yavapai Province, which was undergoing orogeny at that time south of the Superior craton (Whitmeyer and Karlstrom, 2007). This phase is characterized by north-northeast-trending, east-side-up sinistral faults and north-side-down dextral faults, represented by, respectively, the Nokomis Lake Fault and the latest displacement on the Puffy Lake Shear Zone (NOF and PSZ in Figure 12). The late fault pattern explains the late development of the easterly trend of the Flin Flon Domain and is independent of any early paleogeography or plate-tectonic configuration.

What may be a late half graben coincides approximately with a mantle-transition zone marked by northerly extension (Figure 1). The east-northeast-trending, highly oblique normal faults overlie and are parallel to the seismically anomalous mantle transition zone that has been interpreted to have formed ca. 1770 Ma—the end stage of the D_6 deformation. The Fort à La Corne kimberlite field and the Wekusko Lake carbonatite lie along the southern margin of this zone. Differences in Moho reflectivity and relief in the northwest and southeast (Németh et al., 2005) suggest that this is a regional feature and may prove important for diamond exploration. The mantle structure is subparallel to the northeastern termination of the Superior craton and the Owl River Shear Zone (OSZ in Figure 1), the latter probably forming the termination of activated Archean crust thrust beneath the (Archean-contaminated) Waskaiowaka Domain (Figure 4). This late stage of the tectonic evolution is also critical for preserving the high crustal level that contains the largest volume of volcanic rocks and their hosted mineral deposits in north-central Manitoba.

6.1.8 Possible general tectonic implications

The new data and a reinterpretation of earlier works presented in this report suggest that the proposed tectonic model for the development of the margins of the Flin Flon and Kisseynew domains has implications for the tectonic evolution of surrounding domains in Manitoba, the wider THO and other early Precambrian orogens. At the same time, the model gains support from correlations with the eastern THO, extending to northern Quebec, Labrador, Greenland and the Baltic Shield (St-Onge et al., 2006, 2009). The proposed tectonic model has implications for the origin of the entire Kisseynew Domain. The suggested early continuity of a single oceanic arc, including the Lynn Lake and Rusty Lake greenstone belts, implies that a greater dataset can be integrated to better understand the origin of, and mineral distribution in, the Reindeer Zone of the THO. For example, an easily recognizable accretionary wedge of sedimentary rocks, proper mélange, full ophiolite assemblages and blueschist are absent, as in similar early Precambrian collisional orogens. This can be partly explained in the western THO by evidence of higher Precambrian mantle temperatures, as are generally suggested (e.g., Herzberg et al., 2010), that produced high-Mg (up to 18% MgO in Ponton Lake basalt) arc and OIB-like volcanic rocks in the THO. Hot mantle may also have inhibited subduction of continental crust (e.g., Sask craton) and resulted in its high temperature and moderate- to low-pressure metamorphism.

The primitive-arc tectonism suggested by the geochemistry and structure of the early juvenile volcanic rocks in the Kississing–File lakes area has involved arc extension, rifting and foundering, as well as possible subduction erosion—processes that may have inhibited the formation of some modern tectonic features. The overlap in time of arc rifting and the collision of arc fragments suggested by the Josland Lake sills and the early recumbent folds may be related to the nearby location of a vertical plate-rotation pole (Euler pole), causing microplate rotation (e.g., Wallace et al., 2005). Thus, collision could have changed rapidly to extension at the passage of a triple junction. The subduction of small plume-related features, suggested by the abundance of E-MORB, may have caused intermittent crustal shortening in front of approaching oceanic-rise areas. Both processes would have contributed to high heat flow, bimodal magmatism, hydrothermal activity and repeated VMS deposition in the extending parts of the arc terrane. Parts of the arc crust would have remained relatively thin and hot.

The separation of early arc fragments and development of ocean floor in the early Kisseynew Domain, as preserved on its flanks, are large-scale effects implicit in the proposed model (Figure 44a). The presence of E-MORB adjacent to the Burntwood Group on the Kisseynew North flank, (Zwanzig, 2005b) suggests the existence of a sizable ocean basin (that can be called 'Neomanikewan') as part of an early Superior plate. The outliers of Archean basement and cover, with plutons derived from Archean mantle and crust (Zwanzig et al., 2003; Percival et al., 2005, 2006; Whalen et al., 2008), in the Northeast Kisseynew subdomain suggest that the Flin Flon Domain was separated from the Superior craton by a Japan-type back-arc sea and trapped ocean floor, as in the Songpan Ganzi turbidite basin in the Triassic Himalaya (Yuan et al., 2010). The Morton Lake–Loonhead Lake structural break marks a smaller rift that extended into the main ocean tract (Figure 44b). The large volume of successor-arc magmatism north and west of the Kisseynew Domain formed by subduction of an ocean large enough to have contained oceanic islands and to have sustained subduction for 40 m.y. (Figure 44b–d). This and the presence of the large Elbow-Athapapuskow (ocean-floor) assemblage in the Flin Flon Domain suggest subduction of several ocean tracts formed, in part, from back-arc basins; other early Precambrian orogens were probably similar.

The wider THO trends east-northeast across Hudson Bay (key map in Figure 1). Northeast of the bay, progressive growth was toward the south due to 1.88–1.865 Ga terrane accretion (St-Onge et al., 2006, 2009). This occurred at an originally lower plate margin in the north. Following collision, this margin became active with early (1.865–1.845 Ga) and late (1.84–1.83 Ga) arc magmatism above north-dipping subduction. The geodynamic evolution in the northeast is consistent with the ca. 1.87 Ga subduction flip proposed by Corrigan et al. (2005) for the western THO and shown on Figure 44b. The suture that formed at this time between the Rae-Hearne craton and the Reindeer Zone can be correlated with the coeval Baffin suture on central Baffin Island. The subducting slab(s) was trailed by the northern and northwestern margins of the Superior craton, where no arc magmatism is recorded during that time.

The voluminous successor-arc magmatism in the Central Flin Flon subdomain between 1.87 and 1.845 Ga was accompanied during its peak by major granodiorite intrusion in the northwestern Reindeer Zone (Chipewyan Batholith). Magmatism became greatly diminished at the re-amalgamation with the Snow Lake subdomain along the Morton Lake (northwestern)–Loonhead Lake faults. This structural break is probably the suture of the ocean tract that is minimally preserved in the Northeast Reed assemblage and the basalt units that are chemically similar to the Josland Lake sills. Similar breaks along some block-bounding faults may occur within the Central Flin Flon subdomain (Figure 2), one of them containing a slice of greywacke-mudstone similar in age to the Burntwood Group, as at the Trout Lake mine (Rayner, pers. comm., 2009). The break(s) can be correlated with the coeval (ca. 1.845 Ga) Soper River suture (St-Onge et al., 2009) on the south shore of Baffin Island. Collision tectonics must have been orogen wide during deposition of the Missi Group. The Snow Lake subdomain, with its lack of 1.87–1.845 Ga magmatism, probably occupied a lower plate position. Subduction apparently stepped back to the east after the re-amalgamation, and late successor-arc magmatism (Missi-age) was initiated ca. 1.845–1.840 Ga, as in the THO northeast of Hudson Bay. Collision with the Superior craton probably started at almost the same time in the south. This is suggested by the northerly fining and presumed sediment transport in the Grass River Group, which was coeval with deposition of the Missi Group and overlapped the Kisseynew-TNB boundary (Zwanzig et al., 2007). Collision-triggered rollback of the slab adjoining the Superior craton at the northwestern and southwestern margins of the Kisseynew basin started inward rotation and basin closure (Figure 44c, d). Granite batholiths in the upper (western) plate, such as the Cormorant Batholith and possibly a core complex with Missi-age intrusions at its margin (Reed Lake pluton), probably formed in an extensional continental back-arc environment. Thus, westward subduction of the last ocean floor attached to the Superior craton is strongly suggested, in addition to the ongoing northerly plate consumption that ended in underthrusting of the Flin Flon–Glennie Complex by the trailing Sask craton.

The abundant sediment flux from the arc terrane, which was in collision with the Rae-Hearne, Sask and Superior cratons, filled the closing Kisseynew basin with marine turbidite and shallow-water arenite during the terminal collision (Figure 44d). This eventually created a typical migmatitic turbidite–granite terrane, such as commonly interpreted in the THO and elsewhere as a back-arc basin. In our tectonic model, however, the Missi and Burntwood groups are interpreted as the fore-arc and trench deposits of sedimentary rocks that originated from and bypassed the late successor arc built on the margin of the older primitive arc massifs. A shallow trench, filled with juvenile detritus, may have formed parts of the Kisseynew basin margin and receded across the basin to the east during slab rollback caused by the collision. The youngest turbidite sedimentation (Burntwood Group) in the Northeast Kisseynew subdomain is estimated at ca. 1830 Ma (Growdon, pers. comm., 2010). The basement of the Burntwood Group near the trench is interpreted as the trapped ocean floor of the 1.9 Ga back-arc basin(s), but the base of the sedimentary rocks is generally not known to be exposed in the Kississing–File lakes area. If the active 1850 Ma successor-arc margin was foundering and sending volcanic klippen down the trench slope, this may explain the early origin of the Walton Lake nappe, Butler Lake ocean-floor block and other allochthons enclosed in the Burntwood Group. Subducted sediment probably melted in the hot, extensional Mediterranean-type environment. A high Precambrian heat flow, possibly augmented by rising asthenosphere during fragmentation of the slab that caused the Missi-age bimodal magmatism, would have promoted rapid recycling of juvenile sediments and crust into late, high-Na and high-K arc magma in the western THO. The juvenile isotopic character of the mature early arc magmatism (Sherridon–Meat Lake assemblage and Batty Lake Complex) and the Missi-age rocks, however, implies that the Superior craton did not contribute detritus or have its margin subducted during collision. That collision was probably highly oblique and 'hard', rather than along a marine basin. The past location of the craton and the importance of the pericontinental terrane that is partly preserved in the Northeast Kisseynew subdomain are unknown. Archean cratonic fragments are a common feature of the THO and other Proterozoic mobile belts involved in the collision of larger Archean cratons, and may contribute to their generally irregular shape compared to Neoarchean belts.

The pre- and early-collisional paleogeography of the western THO is therefore considered unrelated to the present orientation of the Kisseynew Domain and the present crustal seismic pattern. The north- and east-dipping crustal stack is probably related to the advanced stages of collision tectonics (D_4–D_6 in this report). However, subduction polarity during late successor-arc magmatism is uncertain (e.g., Ansdell et al., 1995; Zwanzig, 1997, 1999; Ansdell, 2005; Corrigan at al., 2005).

Regardless of subduction polarity, block faulting with subbasin development occurred on the active margin(s) surrounding the Kisseynew Domain, possibly because of convergence slowdown, and slab steepening and rollback due to collision. Ignimbrite and caldera development may have marked the change from rapid to slow convergence, as in similar, more recent situations (Mpodozis and Allmendinger, 1993; Ferrari et al., 2002). This probably led to rising asthenosphere producing the change from calcalkaline to bimodal tholeiitic and felsic magmatism in the Missi Group. Slab break-off and possibly lower-crustal and lithospheric-mantle delamination (along with radiogenic heat) probably triggered the high-grade metamorphism and the postcollisional granite intrusion throughout the THO that are also common in other Precambrian orogens.

The composition of syn- and postcollisional granite seems to have been similar to volcanic-arc plutons, except that the former are weakly peraluminous. Modern collisional environments involve old deep crust and large continents, whereas 'continental' crust in the Reindeer Zone formed mainly in an oceanic environment low in Rb and other elements that characterize modern collision. Consequently, tectonic fields derived from Phanerozoic data may not be appropriate in analyzing synkinematic Paleoproterozoic granitoid rocks. Moreover, Morris and Creaser (2008) have shown, in the Paleozoic of the northern Cordillera, that an arc signature of intrusive rocks is no guarantee of contemporaneous subduction but only of crustal contribution and/or of mantle that was altered by subduction at some unknown time. This has implications for the metallogeny of the late granite bodies in the Kisseynew Domain.

The final underthrusting of the internal zone by the Sask craton and presumably its Archean lithospheric mantle probably required removal of the juvenile-arc mantle and lower crust. Archean subcontinental lithosphere that buoyed up and stabilized the cratons would have a high content of magnesian olivine, as would be the case under most Archean cratons. A less depleted sub-arc mantle lithosphere would have been more Fe rich and therefore denser. A thick subcontinental lithosphere probably also prevented the partial subduction of the Sask and Superior cratons such that no blueschist could develop—a situation similar to other early Precambrian collisional orogens. The extent of the Sask craton in the lower crust is probably considerable (Figure 1), and Archean subcontinental lithospheric mantle is also interpreted to have existed, at least until 1885 Ma, in the Northeast Kisseynew subdomain (Figure 3; Whalen et al., 2008). These conclusions and the crustal structure strongly suggest that the juvenile Paleoproterozoic rocks in the Reindeer Zone are allochthonous and that most of this zone has an Archean subcontinental lithospheric mantle (Figure 6). This model, if correct, would have a profound influence on the kimberlite and diamond potential of Manitoba (Section 7.2).

In this model, the suture at the present northeastern margin of the Flin Flon–Glennie Complex was inverted, became cryptic and continued to rotate counter-clockwise after terminal continental collision (Figure 44e, f). This occurred due to mid-crustal flow after tectonic burial, during the high-grade metamorphism. The present east-west orientation of the Kisseynew Domain and the final direction of crustal stacking would therefore be the result of syn- to postcollisional convergence and tectonic wedging (Zwanzig, 1990) during D_4–D_5. This involved the accreted margins of the Rae-Hearne, Sask and Superior cratons (Figure 6). The fault between the para-autochthonous strata of the Superior Boundary Zone and the allochthonous rocks of the Reindeer Zone may be correlated with the Bergeron suture (St-Onge et al., 2009) in northern Quebec and Labrador. The north-south convergence (present co-ordinates) proposed for the eastern THO is consistent with the terminal stages of convergence in the western THO, but may require clockwise rotation of the Rae-Hearne craton or a late shift of its rotation pole.

The last stage of northwest compression ca. 1780–1770 Ma was probably caused by distal forces, possibly from the eastern THO or the Yavapai terrane in the northern midwest United States. An important feature for mineral exploration is the preservation, in the southern part of the Flin Flon Domain, of high crustal levels with low-grade metamorphism and exceptional exposure of volcanic rocks (including VMS and gold deposits). This is related to a postcollisional and postmetamorphic structure similar to an east-trending, south-dipping half graben. The late faulting may have occurred during the D_6 deformation as a conjugate displacement to the sinistral-reverse reactivation of the suture along the TNB.

7 Economic geology

Descriptions of mineral deposits and mineral occurrences in the Kississing–File lakes area are available in Zwanzig and Schledewitz (1992), Zwanzig (1994) and Manitoba Geological Survey (2010), but are summarized herein. New data and interpretations presented in this report provide a better understanding of the origin of some of these deposits. New relationships are established between metallogeny and the tectonic evolution of the area, each stage of evolution being associated with different types of mineral deposits. Deposit and occurrence locations are shown on Map GR-2010-1-2 (DVD).

7.1 Mineral deposits and occurrences

Mineral deposit data presented in this section are summarized mainly from the following sources:

- Mineral Deposit Series reports published by Manitoba Energy and Mines, namely Ostry and Trembath (1992) and Ostry et al. (1998)
- assessment files that have been released from confidentiality after publication of the original Mineral Deposit Series report in the Manitoba Mineral Deposits Database (Manitoba Geological Survey, 2010)

These data and the summary of the deposits presented here are abstracts from the online Manitoba Mineral Resources database (Manitoba Geological Survey, 2010). The Halo Resources Ltd. website (Halo Resources Ltd., 2009) was a major source of information on the Sherridon deposits. Other references are cited below. All UTM co-ordinates are given as Zone 14, NAD 83.

Although the deposit data have been published, the senior author (Zwanzig) has provided a new regional interpretation of the Dickstone volcanogenic massive sulphide (VMS) deposit and reconsidered the setting of the Nokomis gold deposit and related occurrences after mapping the hostrocks. The probably young age of the Sherridon deposits and occurrences is assessed, and some new thoughts are given on the sulphide occurrences on Kississing Lake. A major shear zone is associated with the Puffy Lake gold deposit.

Metallogenic interpretation indicates that six individual or groups of mineral deposits and smaller occurrences are associated with different tectonic settings in the Kississing–File lakes area:

- Sulphide showings and associated hydrothermal alteration that are most likely part of the early juvenile Flin Flon arc assemblage occur between Kisseynew Lake and Syme Lake.
- A VMS deposit (Dickstone deposit) is associated with early volcanism in the transition from arc to arc rift in the Fourmile Island assemblage southwest of File Lake.
- The Sherridon VMS deposits occur in the highly metamorphosed volcanic assemblage that is suggested to be part of a successor arc. Regardless of their age, geochemical data and rare primary features show that the hostrocks to these deposits have similar crustal and mantle melt-sources and formed in an oceanic environment, like the early arc assemblages.

- Iron-sulphide–graphite showings and drill intersections occur in the more distal, purely sedimentary, downslope parts of the Sherridon-age depositional system. Occurrences of volcanogenic sulphides and associated alteration appear to have formed at the margin of the successor-arc terrane leading into the early Kisseynew basin.
- Small, subeconomic gold deposits at Nokomis Lake and Evans Lake, and occurrences in the west toward Kisseynew Lake, are hosted in the exceptionally highly fractionated Josland Lake layered mafic sills, formed during early arc rifting. The gold may have become concentrated as a result of mineralizing fluids interacting with contained oxide phases.
- The Puffy Lake gold deposit may have veins related to the adjacent syn- to postcollisional Puffy Lake Shear Zone. Metamorphic fluids from the Burntwood Group at depth may have contributed to their formation.

7.1.1 Kisseynew–Syme lakes sulphide occurrences

Sulphide mineralization is traced intermittently for more than 8 km along the contact between felsic biotite±garnet gneiss (unit 4b) and amphibolite (unit 4) with rarely preserved deformed pillows. This contact extends from areas south of the Kisseynew Lake narrows and south of Weldon Bay to Syme Lake (Froese and Gall, 1981; Froese, 1997). Local, poorly preserved fragments in the felsic gneiss indicate a volcanic origin, and anthophyllite, some in layers with garnet, indicates that these rocks underwent premetamorphic hydrothermal alteration.

Mineralization consists mainly of disseminated pyrite, pyrrhotite, chalcopyrite and sphalerite (Froese and Gall, 1981). Drilling at the west end of the felsic gneiss shows “sheared micaceous quartzite” that may be silicification (Manitoba Geological Survey, 2010, deposit M63K/14-002 at UTM 345995E, 6093914N). The drillcores intersected only 1.2 m with traces of chalcopyrite in “sheared rhyolitic porphyry with moderate amounts of pyrrhotite” and elsewhere up to 45 cm of solid pyrrhotite. South of Weldon Bay, the host is finely layered siliceous rock with locally preserved quartz phenocrysts (Froese and Gall, 1981). The sulphide zone is approximately 2 m thick and has disseminated and vein-hosted pyrite, sphalerite, chalcopyrite, pyrrhotite and arsenopyrite (<30% in total) found in quartz-sericite schist in trenches. Early reports indicated that quartz-rich gneiss formed the structural footwall and quartz-bearing garnetiferous amphibolite the hangingwall to the deposit (Manitoba Geological Survey, 2010, deposit M63K/14-001 at UTM 348607E, 6093489N).

A trench southeast of Weldon Bay exposes a 5 m thick anthophyllite-bearing layer with 10–15% pyrite between rhyolitic tuff and garnet-biotite-quartz alteration (Gale, 1980; Manitoba Geological Survey, 2010, deposit M63K/14-005 at UTM 352732E, 6094139N). A nearby 20 m drill intersection returned 0.29–1.59% Cu over a total of 9 m with mineralized intervals. Core samples assayed generally <1% Cu (up to 5.7%) over 20 cm and 1–20% Zn over 25 cm, with <0.7 g/t Au.

These volcanogenic sulphide occurrences probably represent points that focused weak hydrothermal activity above felsic volcaniclastic rocks that acted as a stratabound fluid channel. This channel was probably capped by the mafic flow unit. Whole-rock geochemistry of rhyolitic compositions indicates a median K_2O/Na_2O ratio of 1.4, suggesting a rhyolite precursor in a volcanic-arc assemblage (Manitoba Geological Survey, 2010, deposit M63K/14-001 at UTM 348607E, 6093489N).

7.1.2 Dickstone VMS deposit

The Dickstone deposit, a past producer, lies 3 km west of Morton Lake and is hosted in the Dickstone rhyolite (Bailes, 1980a). The area was staked before 1935, after which Sherritt Gordon Mines Ltd. outlined about 1.8 million tonnes of ore to a depth of 100 m averaging 3.13% Cu. A second orebody was found by Hudson Bay Mining and Smelting Co. Ltd. in 1965. When production started in 1968, total reserves were 575,000 tons (522 000 tonnes) averaging 2.53% Cu, and still stood at 519,300 tons (471 000 tonnes) averaging 2.82% Cu, 2.8% Zn, 0.009 oz./ton (0.309 g/t) Au and 0.3 oz./ton (103 g/t) Ag in 1973, but mining ceased in 1975 after 775 210 tonnes of ore had been extracted.

The lens-shaped, Cu-Zn massive sulphide No. 1 orebody is hosted by felsic fragmental rocks near the stratigraphic base of the Dickstone rhyolite. Pyrrhotite, pyrite, chalcopyrite and sphalerite are coarsely recrystallized.

> *The body plunges steeply to the south, has a strike length of 235 m, averages 3 m in thickness (at surface), is 105 m deep and open at depth. The sulphide lens is stratigraphically underlain by a crosscutting Mg-Fe alteration pipe that includes a limited chalcopyrite stringer zone immediately below the orebody. Common minerals within the alteration zone include anthophyllite, chlorite, staurolite, almandine and cummingtonite.* [Manitoba Geological Survey, 2010, deposit M63K/16-058])
>
> *The No. 2 orebody is stratiform, dips close to vertical and plunges steeply to the south. It occurs at the stratigraphic top of the Dickstone formation in contact with overlying pillowed mafic metavolcanic rocks* [Storozuk volcanics]. *The orebody is 0.3–3.0 m thick, approximately 180 m in length and 350 m in depth.* [Bailes, 1980a; Manitoba Geological Survey, 2010, deposit M63K/16-058 at UTM 404009E, 6079546N]

7.1.2.1 Metallogeny

The Dickstone orebodies have the typical local setting of a VMS deposit and an interesting, wider regional setting. They are associated with felsic fragmental volcanic rocks and underlain by stringer ore and a metamorphosed hydrothermal alteration zone. The geochemistry of the intermediate and mafic rocks below and above has a moderate to weak volcanic-arc signature (Section 5.7.2.4). They lie in a region where the strong arc signature of the southern Fourmile Island assemblage (Syme and Bailes, 1996) changes along strike into the weak arc or arc-rift signature of the northern part of the assemblage. The Dickstone rhyolite and its deposit occur at the main transition, apparently marking the onset of arc rifting with attendant normal faulting. The epiclastic Yakymiw Formation (Bailes, 1980a), which overlies the upper mafic unit (Storozuk volcanics), is interpreted here as part of the rift-basin fill.

7.1.3 Sherridon VMS deposits

The Sherridon–Meat Lake area, as reviewed by Ferreira (2006) for Halo Resources Ltd., is a past producer still considered highly prospective for volcanogenic massive sulphide (VMS) Cu-Zn-Au-Ag mineralization. It encompasses 'the Sherridon structure', a window into a low level contiguous with the Flin Flon metavolcanic domain to the south, and similar to the Walton Lake nappe, an allochthon isolated within the migmatitic greywacke-mudstone unit of the Burntwood Group (Figure 12). Both of these structures were developed in the bimodal, felsic-dominated volcanogenic gneissic rocks of the Sherridon–Meat Lake assemblage. Mineralization is hosted in the felsic to lesser intermediate Sherridon gneiss (unit 5). The past producers were the Sherridon deposits; prospects include the Lost Lake, Bob Lake, Fidelity, Jungle Lake, Park Lake, Cold Lake and Ake zone deposits (Maps GR2010-1-1 and -2, back pocket/DVD). The Park Lake and Jungle Lake deposits were owned by Hudson Bay Exploration and Development Co. Ltd., and the Bob Lake deposit was owned by W. Bruce Dunlop (NPL) Ltd. (Halo Resources Ltd., 2009).

7.1.3.1 Sherridon East and West ore lenses

After the first discovery of base-metal mineralization in 1922, a mine operated by Sherritt Gordon Mines Ltd. from 1931 to 1951 produced 7.7 million tonnes of ore grading 2.46% Cu and 0.8% Zn, as well as 2800 kg Au and 9100 kg Ag (Ostry et al., 1998). Ore was recovered from two stratabound sulphide lenses (East and West ore lenses; Manitoba Geological Survey, 2010, deposit M63N/03-001), each 2400 m long, 5 m thick and reaching, respectively, 75 and 460 m vertical depth (Froese and Goetz, 1981). Both lenses occur at the same 'stratigraphic' position but are separated by a barren interval for approximately 1100 m along strike. Mineralization includes pyrite, pyrrhotite, chalcopyrite, sphalerite, magnetite and rare gahnite. This mineral assemblage occurs in all the Sherridon-type deposits in the region, although gahnite is not generally recognized in reports and may be missing. Both lenses at Sherridon are hosted in felsic gneiss (unit 5), with biotite-garnet-sillimanite–bearing gneiss occurring along a part of the East lens footwall contact (J. Krebs, pers. comm., 2010). This includes local cordierite-gedrite rocks of focused alteration (unit 5d) found also below the West ore lens. Overlying mafic rocks are volcanic-derived amphibolite (unit 4c). This rock association confirms a volcanogenic massive sulphide (VMS) origin. Crude metal zonation, identified by Goetz (1980) as an increase in Cu/Zn ratio downward, and generally sharper contacts at the top support upright facing of the mineralized section. Given the overall structure of the domal Sherridon–Hutchinson Lake Complex (Zwanzig, 1999; Healy and Krebs, 2008; this report), the Sherridon ore zones have been twice overturned.

The volcanic origin of the hosting gneissic units was recognized by Ashton and Froese (1988) and Zwanzig and Schledewitz (1992). Structural patterns that have placed strata into their present positions are key components in identifying new mineralization on the Sherridon property. Sheath folding has been identified by current workers as being responsible for the final geometry (Healy and Krebs, 2008; W.A. Barclay, pers. comm., 2008). The VMS deposits occur at or near the same two or three structural horizons. All but one of the deposits lie on horizons, probably a repeated single layer, on the inner and outer sides of the main succession of felsic gneiss of the Sherridon–Meat Lake assemblage. The Cold Lake deposit appears to lie at an intermediate structural level (J. Krebs, pers. comm., 2010).

7.1.3.2 Cold Lake deposit

The Cold Lake deposit lies 1 km northeast of the Sherridon West lens. The area was drilled by Halo Resources Ltd. in 2008. Intersections include 7.8 m containing 2.1% Cu, 6.2% Zn and 11 g/t Au. Resources total 1.2 million tonnes grading 0.7% Cu, 1.4% Zn, 0.3 g/t Au and 9.0 g/t Ag. From current data, the mineralization has about 400 m of strike length, dips northeast and can be traced 150 m down-dip in mainly felsic gneiss. It dips about 40°E near surface, steepening to 80° at depth, opposite to the adjacent Sherridon West ore lens, whose dip shallows at depth. Mineralization remains open along strike and also at depth (Halo Resources Ltd., 2009). Southwest of the deposit is a unit of Meat Lake–type amphibolite in the footwall.

Gale (1980) mapped a gahnite-bearing quartzite layer (herein interpreted as silicification) adjacent to an exposure of garnet-biotite-cordierite-anthophyllite rock in the vicinity. Cordierite-anthophyllite rocks were also documented in the area by Froese and Goetz (1981) and indicate Mg alteration. Krebs (Healy and Krebs, 2008; J. Krebs, pers. comm., 2010) has interpreted the Cold Lake deposit to be stratigraphically stacked on top of the horizon containing the Sherridon East and West lenses and that, structurally higher up, the succession becomes inverted. Thus, the Fidelity occurrence (Section 7.1.3.4) and Bob Lake deposit (Section 7.1.3.5) must face away from the core of the Sherridon structure toward the Cold Lake deposit.

7.1.3.3 Lost Lake deposit

The Lost Lake zone is a new discovery directly northeast of Sherridon between Lost Lake and Sherlett Lake, on strike with the Cold Lake deposit. Intersections along a strike length of >500 m encountered mineralization within 150 m of surface, including 6.3 m with 1.3% Cu and 7.8% Zn. Newly corrected dip is about 50°NW. The 'inferred resources' (NI43-101) are 2.2 million tonnes grading 0.9% Cu, 2.7% Zn, 0.6 g/t Au and 8.9 g/t Ag (Halo Resources Ltd., 2009). Mineralization occurs in felsic gneiss (unit 5) and at the hangingwall contact of interlayered intermediate gneiss (also shown as unit 5 on Maps GR2010-1-1 and -2, back pocket/DVD). Meat Lake–type amphibolite (unit 4c) occurs in the footwall at a variable distance below the ore zone. The deposit lies on the same horizon as the Cold Lake deposit and is therefore assumed to face upward.

7.1.3.4 Fidelity occurrence

The area was first staked by D.C. Barr in about 1930 as the Copper and Gold occurrences and, similar to the Sherridon ore lenses, had a history of follow-up work with drilling by Sherritt Gordon Mines Ltd. and Fidelity Mining Investments Ltd.

Disseminated and local massive mineralization occurs in quartzofeldspathic rocks that contain variable amounts of quartz, garnet, sillimanite and probably chloritized cordierite (reported as chlorite). Two southeast-dipping sulphide-rich layers, 5 and 4 m in core length, were intersected in one hole (Manitoba Geological Survey, 2010, deposit M63N/03-005 at UTM 368903E, 6114718N). These layers contain up to 80% pyrite and pyrrhotite with lesser amounts of chalcopyrite and sphalerite. The first layer is overlain by approximately 19 m of disseminated Fe-sulphide±chalcopyrite mineralization in rock that is reported to contain variable amounts of sillimanite, biotite, garnet and hornblende. Near-solid sulphide mineralization intersected in various holes included 1.36–1.94% Zn over 1.5–1.7 m and 2.14–1.56% Cu over 1.0–1.5 m.

Froese and Goetz (1981) showed a unit of "pelitic gneiss" in the structural footwall to the southeast, identified by Tinkham and Karlapalem (2008) as sillimanite-bearing garnet-biotite±cordierite±anthophyllite±hornblende gneiss with a very coarse core zone rich in cordierite-anthophyllite. This zone is best interpreted as premetamorphic VMS-related hydrothermal alteration, but if it is used to determine facing, its position is inconsistent with the fold model of the Sherridon structure. This may be explained by stratigraphic stacking of the deposits and alteration near the Cold Lake level that is younger than the Fidelity occurrence. The Fidelity occurrence can be interpreted to lie in the original footwall of a (D_3?) thrust that overrode the Bob gneiss. Restored to pre-F_4 folding, the Fidelity occurrence lay on the east limb of a large early syncline (F_3?) and faced the Sherridon East and West ore lenses and the Lost and Cold lakes deposits on the opposite limb. Thus, a single horizon containing most deposits formed two parallel 'tracks', one in each fold limb (terminology of Healy and Krebs, 2008). The Fidelity occurrence was probably overturned during F_4 (sheath) folding, and the Sherridon lenses were probably overturned in the Sherlett Lake reclined structure during F_5 refolding. Westward flow in the core of the sheath could have placed the early east-limb deposits (Fidelity and Bob) on the inner 'track'.

7.1.3.5 Bob Lake deposit

The Bob Lake deposit lies on the upper limb of the Sherlett Lake fold, east of and probably in the same horizon as the Fidelity occurrence (inner track). It consists of four lenses of solid and disseminated sulphides that occur intermittently over a strike length of approximately 490 m (Manitoba Geological Survey, 2010, deposit M63N/03-002 at UTM 369801E, 6114503N). They dip northeast and have the Bob gneiss (unit 10a) in the structural hangingwall. Part of the sulphide mineralization has been mobilized into adjacent granitic rocks during deformation and attendant metamorphism.

The Bob Lake deposit was discovered by Sherritt Gordon Mines Ltd. in 1941, and is currently held by W. Dunlop Ltd. and optioned to Halo Resources Ltd. It is hosted in felsic and intermediate Sherridon-type gneiss directly south of the north-northeast-dipping Bob Lake garnet-biotite±hornblende± graphite gneiss (metagreywacke?). The contact is marked by a unit of more highly graphitic gneiss. A thin spur of the mineralization extends into the Bob gneiss. Published reserve figures for three of the four lenses are 2.16 million tonnes of 1.33% Cu, 1.18% Zn, 0.31 g/t Au and 8.45 g/t Ag. The outcome of drilling by Halo Resources Ltd. up to June 2008 suggested that a continuous mineralized sulphide layer plunges 10–20°SE to a vertical depth of 260 m, pinching and swelling but open at depth. The 'indicated resources' (NI43-101) are 1.43 million tonnes grading 0.89% Cu, 0.88% Zn, 0.10 g/t Au and 2.58 g/t Ag (Giroux and Moore, 2007).

7.1.3.6 Jungle Lake deposit

The Jungle Lake deposit is also located in quartzofeldspathic±garnet±hornblende Sherridon-type gneiss (unit 5) and quartz-hornblende gneiss (unit 4a), but on the north-northeast-dipping margin of the Sherridon structure, south of Jungle Lake. It is structurally overlain by layered amphibolite and intermediate and felsic gneiss (unit 4a). This and a probable fault marked elsewhere by high graphite and Fe-sulphide concentrations separates it from the structurally overlying, older Star Lake trondhjemite gneiss farther to the north. The section is probably inverted, and the Jungle Lake deposit may lie on the same horizon as the Sherridon ore lenses (outer track). Weak alteration in the present footwall (original hangingwall) is probably related to stacked mineralization, as at Cold Lake. Because no stratigraphically higher mineralization is recorded anywhere northeast of Cold and Lost lakes, this horizon may have been relatively short.

The Jungle Lake deposit was discovered and then drilled by Hudson Bay Exploration and Development Co. Ltd. (HBED) from 1958 to 1967 (Ferreira, 2006). According to HBED, the deposit has an easterly strike length of 400 m and an average true width of 5.6 m, and extends to at least the 400 m level. The zone dips 40°N and plunges 35°NE. Chalcopyrite and sphalerite occur as stringers and blebs in near-solid pyrrhotite and pyrite (Manitoba Geological Survey, 2010, deposit M63N/02-003 at UTM 374408E, 6115079N). Historical reserves are 3.36 million tonnes grading 1.42% Cu and 1.1% Zn. After further drilling by Halo Resources Ltd., the Jungle Lake deposit (on option) was found to contain 'indicated resources' of 830 000 tonnes grading 0.99% Cu, 0.73% Zn, 0.39 g/t Au and 6.7 g/t Ag (Giroux and Moore, 2007).

7.1.3.7 Park Lake deposit

The Park Lake deposit, discovered at the northwest end of Park Lake and drilled from 1959 to 1964 by Hudson Bay Exploration and Development Co. Ltd., continues to be held by HBED. It is located on the PAR claims north and contiguous with Halo's Sherridon claims. The deposit lies 6.5 km west-northwest of the Jungle Lake deposit and is probably on the same mineralized horizon.

The Park Lake deposit comprises four lenses striking west-northwest, dipping 45°N and explored to a depth of 670 m. Historical resource estimates calculated by HBED are 6.14 million tonnes grading 0.42% Cu and 2.16% Zn (Ostry et al., 1998; Ferreira, 2006; Manitoba Geological Survey, 2010, deposit M63N/03-003 at UTM 368442E, 6117615N). Drilling by Halo Resources Ltd shows continuity of the thicker Zn-rich ore below 300 m from surface. Intersections of 17.9 and 12 m grade 2.5% and 3.7% Zn, respectively (Halo Resources Ltd., 2009). Mineralization is hosted in quartz-rich biotite±garnet gneiss (unit 5) interlayered with felsic hornblende-biotite gneiss and calc-silicate gneiss of the Sherridon–Meat Lake assemblage. These rocks occur at the northern, overturned margin of the Sherridon structure, below and south of the western extension of the older Star Lake trondhjemite gneiss.

7.1.3.8 Ake zone deposit

The Ake zone is located 4.5 km north-northeast of Meat Lake on the east side of the Walton Lake nappe. It is hosted in moderately northwest-dipping felsic Sherridon-type gneiss (unit 5) at the contact with garnet-biotite±hornblende gneiss (unit 5c). The latter is considered to include hydrothermal alteration of felsic and intermediate rocks (unit 5). Cordierite-gedrite rock (unit 5d) is interpreted as focused alteration and occurs southeast of the mineralization, suggesting the presence of a northwest-facing VMS deposit.

The Meat Lake area has had an exploration history with several claims since 1956. The Ake zone deposit, discovered by HBED in 1971, is held by W. Bruce Dunlop (NPL) Ltd. The deposit contains well-mineralized to near-solid pyrrhotite and pyrite layers that contain blebs and stringers of chalcopyrite and subordinate sphalerite. It has a strike length of about 180 m and width of up to 4.6 m, dipping 50°WNW but shallowing at depth across the F_6 Meat Lake synform (Zwanzig, 1999; Ferreira, 2006; Manitoba Geological Survey, 2010). The mineralized zone plunges for at least 915 m toward 345°.

7.1.3.9 Minor Sherridon-type Cu-Zn showings

Small surface sulphide occurrences extend for about 1 km along the strike of the local gneissic foliation near the southwest end of the Sherridon structure. One schistose zone, 3–6 m thick and 150 m long, in “fine-grained, bedded quartzite” contains small lenses of sphalerite and chalcopyrite. An airborne electromagnetic survey performed by Sherritt Gordon Mines Ltd. in 1972 did not delineate any anomalies (Manitoba Geological Survey, 2010, deposit M63N/02-011 at UTM 375762E, 6110109N).

A 21 m drill intersection of quartz-biotite gneiss in a hole about 1 km west of Star Lake contains disseminated pyrrhotite and pyrite, and local disseminated chalcopyrite (Manitoba Geological Survey, 2010, deposit M63N/02-027 at UTM 376665E, 6114167N).

Various claims have been held since 1926 on the present Halo Resources Ltd. property 1.8 km southwest of the Park Lake deposit and at the same structural level. Minor chalcopyrite was found in 1980 by Shell Canada Resources Ltd. in an exploration shaft in cordierite- and anthophyllite-bearing hornblende-plagioclase gneiss at the contact of felsic Sherridon gneiss and “pelitic gneiss” (Manitoba Geological Survey, 2010, deposit M63N/03-077 at UTM 365826E, 6117455N). If the pelitic gneiss represents Al-Mg alteration in the structural hangingwall, the showing supports the conclusion that the outer mineralized zone in the Sherridon structure is inverted and faces the other, sheared and generally graphite–Fe-sulphide–rich contact with the surrounding granitoid gneiss. Because the same contact contains local slivers of garnet-biotite paragneiss, the originally upright mineralized section is interpreted to be overthrust by the older granitoid gneiss with lenses of the Burntwood Group in the hangingwall. This must have occurred during D_3 and/or later deformation but before D_5 sheath folding.

An intersection (by HBED in 1958) of 3.5 m of moderate pyrite and disseminated sphalerite mineralization above 11.5 m of moderate to near-solid pyrrhotite with disseminated pyrite and sphalerite (Manitoba Geological Survey, 2010, deposit M63N/02-037 at UTM 375762E, 6110109N) occurs 1 km north of Jungle Lake in felsic gneiss (unit 5) at the west end of the Walton Lake nappe (UTM 374322E, 6119769N). This lies on strike southwest of a large alteration zone that includes a thin silicified layer above sillimanite-cordierite-garnet±anthophyllite/gedrite gneiss.

7.1.3.10 Metallogeny and deposit controls

The Sherridon deposits and occurrences are volcanogenic massive sulphides, although they were previously considered to be stratabound sediment hosted due to the abundance of quartz and presence of garnet±sillimanite in their felsic host. Rarely preserved primary structure, as seen in the Sherridon gneiss on clean burned-over outcrops, however, includes finely quartz-phyric texture. Locally, there are changes in size and abundance of the phenocrysts between different layers, as can occur only in volcanic rocks. Occurrences of slightly coarser porphyry are best interpreted as subvolcanic intrusive rocks. Poorly preserved breccia structure occurs northwest of the Ake zone. Interlayered mafic rocks have locally preserved pillow structure or fragmental texture. These are clearly volcanic rocks; although coarser grained amphibolite is intrusive or of uncertain igneous origin, much of the garnet-bearing felsic–intermediate rock has lost all its primary features.

The local clues to a volcanic origin for the Sherridon–Meat Lake assemblage, however, are strongly supported by geochemistry that is very similar to that of rhyolite, dacite and basalt in the lower grade parts of the Flin Flon Domain. The trace-element fingerprint, particularly of the mafic units (Meat Lake and Walton Lake amphibolite), has strong arc-volcanic characteristics. The mafic and felsic modes are particularly similar to units in the upper part of the Snow Lake arc assemblage (Section 5.7.7). The similarity of these bimodal assemblages, in fact, pointed to a correlation of these rocks (Zwanzig, 2008) until a preliminary age in the range of 1850–1855 Ma was obtained by U-Pb TIMS zircon geochronology (N. Rayner and D. Tinkham, pers. comm., 2009), which is about 40 m.y. younger than that of the Snow Lake assemblage (David et al., 1996).

The origin of the aluminous and mafic minerals in much of the felsic–intermediate Sherridon gneiss is interpreted as widespread, metamorphosed, regional hydrothermal alteration and locally focused alteration. Addition and removal of incipient leucosome has also caused alteration. Neither type of alteration, however, has had a significant effect on most incompatible elements and particularly their ratios (Section 5.5). The altered rocks occur either in mappable layers, highly elongate lenses or local decimetre- to metre-scale patches within the little-altered volcanic-derived gneiss. Patches generally have gradational margins and the mafic mineral content increases toward a core that may contain Fe-sulphide minerals. This structure and the presence of garnet, cordierite, anthophyllite/gedrite, sillimanite and/or various species of amphibole are typical of metamorphosed alteration in volcanic rocks. In many of these altered zones, a relatively high quartz content is retained and indicates a felsic volcanic protolith. The abundance of regional alteration indicates fluid circulation in a high-temperature volcanic regime. Areas with high contents of aluminous and mafic minerals are considered to be the result of intense hydrothermal alteration focused along synvolcanic faults or other fluid conduits. This is confirmed by the presence of such alteration preferentially adjacent to known base-metal deposits or sulphide showings. Both the Sherridon ore lenses and the Fidelity occurrence, as well as others, show such alteration nearby. Areas with abundant sillimanite (that may occur as fibrolite in quartz-rich faserkiesel) are interpreted as the metamorphic product of sericitized alteration domains, which commonly adjoin the strongly Mg-enriched and Na-depleted alteration.

Where recent mapping or new reconnaissance has been done, several areas also show a relatively thin layer of highly siliceous rock near a base-metal occurrence. These are interpreted to be the silicified carapace above hot circulating fluids and below cooler fluids from circulating seawater, as is suggested by common VMS models.

Deformation and generation of pegmatite have clearly imposed a dominant control over the geometry of alteration zones and deposits. The interpretation of the Sherridon structure as a large sheath fold implies very high non-coaxial regional strain. This has been interpreted to have occurred during peak metamorphism at a midcrustal level during continental collision. Deformation was probably enhanced by the generation of leucosome layers and pegmatite bodies during the start of unroofing of the Kisseynew migmatite terrane, which covered the Sherridon and Walton Lake structures at that stage. The main results of this deformation were intense attenuation and transposition of all existing structures. Stretching occurred in directions that curve smoothly around the Sherridon–Hutchinson Lake Complex and its Sherridon-structure core. This has resulted in all contacts being subparallel to primary lithological layering and stratabound alteration. Moreover, all features are much longer and thinner than original dimensions. Focused alteration zones generally appear to be subparallel to deposits and regional layering. Where investigated, the plunge of all orebodies conforms largely to the pattern of stretching lineations around these structures. This includes the shallow plunge of the Cold Lake deposit, the gentle southeast plunge of the Bob Lake deposit, the northeast plunge of the Jungle Lake deposit and the downward-shallowing north-northwest plunge of the Ake zone deposit. These trends follow the hairpin curve of the major sheath fold and the shape of the Sherridon–Hutchinson Lake Complex and Walton Lake nappe. The acute angle between the relatively late (D_5) stretching lineation and the plunge of the deposits is probably related to intense earlier deformation that may have resulted from folding during D_4. The plunge at Bob Lake, which is shallower than the adjacent F_5 Sheila Lake fold, suggests repeated folding with a shallow plunge. The curved Ake zone deposit indicates F_6 refolding in the Meat Lake synform.

The tectonic environment of the Sherridon deposits is clearly shown to be a juvenile oceanic arc by its geochemistry and Nd-isotope ratios. Bimodal volcanism was apparently followed by non-arc (extensional) mafic intrusions. Their geochemistry suggests that these are part of the syntectonic, post-Burntwood set of Weldon Bay dikes. The early events are inferred to be associated with unmapped synvolcanic faulting that caused the local focusing of hydrothermal alteration and generation of the VMS deposits. The abundant alteration and abundant deposits are probably the result not only of the pervious felsic stratigraphy, but also of the presence of subvolcanic intrusions and generally high heat flow associated with juvenile extensional arcs in Precambrian terranes.

The Batty Lake Complex, although possibly older than the Sherridon–Meat Lake assemblage, has a geochemical pattern very similar to that of the least potassic of the Sherridon gneiss. It is suggested (Section 5.8.2.6) that this large, partly remobilized batholith had the same sources of melting in the crust and mantle wedge as the Sherridon gneiss. Its trondhjemitic melt with unfractionated REE could have been generated in a purely oceanic mantle and crustal environment, although its size indicates a thickness of mature-arc crust. This is consistent with the stage of advanced successor-arc magmatism that is indicated by its recently determined preliminary age range of about 1850–1855 Ma for the volcanic rocks. The Mg-rich alteration in the trondhjemite yielded anthophyllite/gedrite and cordierite during peak metamorphism after sub-seafloor alteration similar to that in the Sherridon gneiss. Although not directly related to the Sherridon deposits, such plutons may have been abundant in the successor-arc crust, as is suggested by the similar composition of a phase in the Defender dome. These and other plutons were most likely parts of the heat engine driving the extensive alteration and sulphide deposition in the Sherridon–Meat Lake assemblage, in the cover of the Batty Lake Complex and, to a lesser extent, in the cover of the Defender dome. Thus, the entire Kisseynew–Kississing–Batty lakes area has a VMS potential. Further geochemical work, isotopic dating and VMS exploration is recommended for this region. The association of depleted rhyolite gneiss with felsic rocks having higher La/Yb ratios and with highly altered packages may warrant investigation. Perhaps the key to the high mineral potential of this part of the mature, early successor arc was its oceanic environment, as opposed to the more emergent main part of the Flin Flon Domain during the end of early successor-arc magmatism.

7.1.4 Possible sediment-hosted or distal VMS deposits

7.1.4.1 Yakushavich Island–Collins Point base-metal occurrences

These showings in north-central Kississing Lake were known and staked in 1925. Gale (1980) documented a near-solid sulphide layer composed of sphalerite, pyrite and chalcopyrite that is 25 cm or more thick on the east side of Yakushavich Island. The ‘Yak zone’ was mapped to adjoin ‘quartzite’ with minor biotite- and quartz-rich gneiss ±garnet, rock types with a mineralogy similar to silicified and Fe-enriched Sherridon felsic volcanic gneiss. From 1993 to 1995, Canmine Resources Corporation conducted geological mapping, geophysical surveys and diamond-drilling, and identified eight Zn-Cu–rich zones in the vicinity of Yakushavich Island and Collins Point.

In a drillhole at the Yak zone, siliceous biotite paragneiss is underlain by ~1 m of near-solid to solid pyrite, sphalerite and chalcopyrite. The sulphide-rich layer contained 5.29% Zn and 0.25% Cu. A 1.5 m core sample from the siliceous biotite gneiss contained 5.49 g/t Au (Manitoba Geological Survey, 2010, deposit M63N/03-044 at UTM 351727E, 6118034N). Southwest along strike, near-solid sulphide veins of pyrrhotite and pyrite±chalcopyrite, up to 2 cm thick, were documented over approximately 7 m, and 3% disseminated chalcopyrite was reported in granitoid rocks (Manitoba Geological Survey, 2010, deposit M63N/03-045 at UTM 351223E, 6117818N). An occurrence on the west side of Yakushavich Island (Manitoba Geological Survey, 2010, deposit M63N/03-050 at UTM 350502E, 6118419N) includes, for example, a 30 cm intersection that contains up to 50% pyrrhotite±chalcopyrite within amphibolite. Other short mineralized sections are in granitoid rocks.

Schledewitz (1995) mapped the area at 1:20 000 scale, concluding that the Yak zone lies in a refolded, early recumbent structure that extends to Collins Point. Local leucocratic quartz-feldspar gneiss with zones of porphyroblastic gedrite±sillimanite±cordierite gneiss is best interpreted as felsic metavolcanic rock with focused hydrothermal alteration (unit 5d) similar to that associated with metamorphosed VMS deposits elsewhere. Adjoining garnet-biotite gneiss (unit 10a) with local Fe-sulphide, which occurs widely within the recumbent fold, may represent weaker regional alteration or a variety of Sherridon-type gneiss with an uncertain volcanic or, more likely, sedimentary affinity. Surrounding amphibolite (unit 4) is interpreted as being derived from volcanic rocks. Garnet-biotite±hornblende ±graphite gneiss (unit 10a), surrounding the more felsic gneiss on Collins Point, hosts several occurrences of Fe-sulphide±galena±sphalerite±chalcopyrite±gahnite (e.g., Manitoba Geological Survey, 2010, deposits M63N/03-041 at UTM 347789E, 6119305N, M63N/03-042 at UTM 348929E, 6119287N, M63N/03-043 at UTM 347641E, 6120152N and M63N/03-054 at UTM 348453E, 6121338N). These may represent more distal mineralization deposited in volcanogenic sedimentary rocks that are generally hornblende bearing, also typical elsewhere of unit 10a. The base-metal occurrences in the Yakushavich Island–Collins Point area possibly occur in structural outliers or reworked volcanic rocks similar to those at Sherridon.

7.1.4.2 Ideal deposit

Geological mapping in eastern Kississing Lake by Cominco Ltd. in 1978 outlined a 600 m long, 50–60 cm thick, near-solid, stratabound pyrrhotite-pyrite-sphalerite-galena-chalcopyrite layer. Mineralized zones encountered in drilling contained pyrite-pyrrhotite-sphalerite and chalcopyrite, with the best reported intersections containing 1.65% Cu, 3.88% Zn and 0.65 g/t Au over 1.5 m, and 0.57% Cu with 2.04% Zn over 3 m (Manitoba Geological Survey, 2010, deposit M63N/03-006 at UTM 361851E, 6111611N). The hostrock is currently thought to be unit 10a garnet-biotite+hornblende+graphite gneiss gradational into amphibolite. Gale (1980) described an anthophyllite-rich rock. This and the occurrence of quartz-feldspar-biotite-garnet-sillimanite gneiss suggest a setting similar to that of the Yakushavich Island occurrence but classified as stratabound sediment associated, like the showings at Collins Point. The presence of chlorite, sericite and breccia in sheared rock indicates late faulting at the deposit.

7.1.4.3 Minor Cu-Zn occurrences

Near-solid sulphide contains small concentrations of Cu and Zn (i.e., up to 1690 ppm Zn and 662 ppm Cu in disseminated pyrrhotite, pyrite and possibly chalcopyrite). This is exposed in trenches cut into sheared amphibolite (unit 4) on the east side of the Adamson Lake dome (Manitoba Geological Survey, 2010, deposit M63N/03-059 at UTM 354886E, 6104126N).

7.1.4.4 Metallogeny and mineral controls

The small, subeconomic, commonly Pb-bearing and Zn-rich base-metal occurrences are closely associated with garnet-biotite gneiss that contains graphite-amphibole±Fe-sulphide (unit 10a). These may be Sherridon-age sediment-hosted volcanogenic sulphides. They occur on Kississing Lake at the outer margin of the mineralized area but are apparently coeval with or slightly older than the Burntwood Group. Amphibole-bearing greywacke gneiss and migmatite with disseminated sulphide occur also as unit 10a within the Burntwood Group near the Fairwind Lake and Loonhead Lake faults. This suggests that unit 10a paragneiss is an early basinal deposit that is stratigraphically overlain by the Burntwood Group. In Figure 44c, the Sherridon–Meat Lake assemblage is interpreted as part of the front of the early successor arc. The thin layers of Sherridon-type gneiss at the margin of the Kisseynew basin may be proximal aprons of resedimented material from this arc.

Structural control in large early folds is important in the distribution of these base-metal occurrences (see Yakushavich Island–Collins Point base-metal occurrences). The extensive, probably gradational contact with the Burntwood Group, however, indicates a predominantly stratigraphic control that forms a distinctive metallotect. This is consistent with an age for the distal deposits similar to that of the Sherridon deposits. Anywhere east of Kississing Lake, the amphibole-bearing metagreywacke units contain only Fe-sulphide and graphite but no base metals, and this appears to be the case also for the Bob gneiss in the core of the Sherridon structure. Because the more distal layers were thrust from the east and north during advanced deformation, the metallotect originally became barren downslope into the basin or, presently, structurally upward in the stack of folds.

7.1.5 Nokomis-type gold deposits and occurrences

7.1.5.1 Nokomis Lake

The Nokomis Lake deposit is located on the southeast side of Nokomis Lake in mottled felsic gneiss that adjoins garnetiferous amphibolite, which is locally layered and was originally interpreted to be iron formation in a sedimentary package. Its owner, Pioneer Metals Corporation, categorizes the deposit as a "shear-related intrusive-hosted (tonalite) lode gold system" (Manitoba Geological Survey, 2010, deposit M63N/02-002; however, see below in this section).

Sherritt-Gordon located arsenopyrite-bearing zones that locally contained visible gold. After several changes in interested parties, Rio Tinto Canadian Exploration Ltd. located several mineralized zones along a strike length of 1000 m in 1960. Drilling in 1961 indicated 90 700 tonnes grading 10.29 g/t Au. Dome Exploration (Canada) Ltd. conducted geophysical surveys and diamond-drilling between 1972 and 1987, and a new owner, Pioneer Metals Corporation, confirmed the presence of gold mineralization to depths of 60 and 185 m, with resources of 350 000 tonnes grading 6.1 g/t Au in 2002 (Manitoba Geological Survey, 2010, deposit M63N/02-002 at UTM 380008E, 6104066N). Traces of tungsten are common.

Zwanzig (1994) found the mineralization, which is exposed as rusty lenses in trenches, to occur in the trondhjemite gneiss at the stratigraphic top of an (overturned) layered Josland Lake–type sill. This felsic hostrock stratigraphically overlies garnet-bearing amphibolite derived from ferrogabbro. Stratigraphically lower diopside-bearing amphibolite was derived from gabbro–norite. The base of the mineralized sill is separated from Burntwood Group greywacke-migmatite to the east by the overturned Loonhead Lake Fault. The mineralized zone is stratigraphically overlain by a second sill that is not significantly differentiated at the top. This sill is intruded into fine-grained amphibolite with local pillow structure (Moody Lake basalt), which is disconformably overlain by the Missi Group. The entire section is overturned and dips moderately east. Mottled trondhjemitic gneiss at the top of the originally lower sill contains mineralized lenses with up to 15% mainly disseminated pyrite and pyrrhotite with arsenopyrite and carrying gold. Clean outcrops, burned over at the time of remapping, show that the amphibolite and the felsic unit are generally massive igneous rocks. Layering is the metamorphic reflection of original modal layering and sheared-out quartz veining. Tops were easily determined on the sills. Lithogeochemistry has clearly confirmed that all intrusive fractions at Nokomis Lake have compositions identical to the lower grade rocks of the type Josland Lake sills.

7.1.5.2 Evans Lake

Lenses of gold-arsenopyrite mineralization are stratabound within mottled trondhjemite gneiss at the top of a Josland Lake sill, about 500 m west of Evans Lake (UTM 386997E, 6101631N). Mineralization comprises up to 5% pyrite, pyrrhotite±arsenopyrite and rare chalcopyrite, disseminated and in thin veins. Up to 1% disseminated fine-grained pyrite and/or pyrrhotite with rare chalcopyrite occurs locally within the amphibolite (Manitoba Geological Survey, 2010, deposit M63N/02-004). Drillcore assays by International Mining ranged from 0.07 g/t Au over 2 m to 0.45 g/t Au over 3 m.

7.1.5.3 Lobstick Narrows to Adamson Lake

Small gold showings in the most strongly recrystallized Josland Lake sills extend to areas north of Kisseynew Lake. The protolith of the hostrock and metallogeny of none of these showings has been previously recognized, but a ‘stratigraphic’ control is mentioned. Anomalous Au assays in the parts per billion range are reported in layered amphibolite (unit 7d) 2.5 km north of Lobstick Narrows (Manitoba Geological Survey, 2010, deposit M63K/14-007 at UTM 342502E, 6097349N). Rusty layers contain up to 20% pyrite and up to 5% arsenopyrite. Higher quartz content towards their centres than at their margins may represent the sheared trondhjemite at the top of a Josland Lake sill. Along strike and 300 m northeast of Weldon Bay (Manitoba Geological Survey, 2010, deposit M63K/14-008 at UTM 343882E, 6095424N), disseminated pyrite occurs in rusty zones and minor quartz veins, mainly along the top (overturned) of the felsic gneiss (unit 7c trondhjemite) above the stratigraphic top of layered garnet amphibolite (unit 7b ferrogabbro).

About 2.4 km to the northwest along strike, up to 20% pyrite and minor arsenopyrite occur as disseminations and thin veins in layered amphibolite (herein reinterpreted as unit 7d). A trace of gold has been reported from this occurrence (Manitoba Geological Survey, 2010, deposit M63K/14-008 at UTM 342502E, 6097349N).

An approximately 5 m long mineralized drill intersection by Elken Exploration Limited in 1981 within siliceous biotite gneiss comprises layers, up to 2 cm thick, of pyrite±arsenopyrite and contains 0.69 g/t Au and 0.69 g/t Ag. Shoreline exposures on a small lake in the northeast mantle of the Adamson Lake dome (Manitoba Geological Survey, 2010, deposit M63N/03-060 at UTM 355796E, 6105208N) include amphibolite that is herein interpreted as unit 7d and tonalite (possibly including unit 7c). This occurrence with pyrite and arsenopyrite suggests a Nokomis-type metallogeny.

7.1.5.4 Metallogeny and comment

The small low-grade deposit and numerous occurrences are hosted mainly by the identical ‘mottled gneiss’ derived from ferroan quartz diorite to trondhjemite at the top of layered Josland Lake gabbro sills. ‘Siliceous’ host may be misidentified sheared trondhjemite or transposed quartz veins in such a rock. A network of mafic septa lying at acute angles to foliation at the Nokomis deposit resembles deformed facture filling. This suggests structural control in a competent layer. The highly Fe-enriched ferrogabbro that underlies the felsic rocks probably acted to oxidize the ore fluid and cause precipitation of gold. Local, above-background platinum group elements (PGE) with the gold, the lack of showings outside the sills, the >500 km^2 surface area of sills with persistent local gold occurrences, and the high-Mg back-arc parent magma source fractionated at depth and extremely fractionated again near surface (Section 5.7.6) together suggest that the ore fluid may have been magmatic. Regardless of the origin of the gold, the felsic unit in the sills is the best exploration target.

7.1.6 Puffy Lake gold deposit

The Puffy Lake deposit area, which is 8 km southeast of Sherridon, was first staked by A. Olson in 1960. After HBED drilling, Maverick Mountain Resources Ltd. with Gränges Exploration Aktiebolag drilled along a strike length of about 1 km. In 1986, Maverick Mountain merged with Pioneer Metals Corporation as major holder. In 1987, they reported reserves of 3.54 million tonnes grading 7.88 g/t Au, in four parallel zones. Advanced mine development and processing, including milling, were suspended after a drop in metal prices and damage due to forest fire. A 1993 study by Kilborn Engineering Ltd.

found proven and probable reserves of 1.3 million tonnes grading 8.57 g/t Au, and possible reserves of 833 700 tonnes grading 7.15 g/t Au (Manitoba Geological Survey, 2010, deposit M62N/02-001 at UTM 373296E, 6100193N).

Auriferous sulphide-bearing quartz veins and zones of quartz-free, gold-sulphide mineralization occur in predominantly biotite-rich mafic gneiss and schist derived by alteration of the Puffy Mine andesite (unit 3a). Gold-sulphide mineralization also occurs in quartzofeldspathic gneiss of the Missi Group (unit 13).

> *Fine- to coarse-grained blebs, disseminations, crystal clusters and mobilizate veins of arsenopyrite, pyrrhotite and pyrite±chalcopyrite, sphalerite and galena form up to 15% of white to smoky grey quartz veins. A fine-grained, mineralized, schistose, biotite-rich (<50%) quartzofeldspathic rock is an immediate host to the quartz vein(s) in three of the four zones. This rock commonly contains up to 20% arsenopyrite as thin wisps, laminae and lenticular clots (<5 by 20 mm). Visible sulphide mineralization within the host rocks is restricted to 1–2 m from vein margins. The fourth zone is hosted by medium- to coarse-grained felsic migmatitic quartzofeldspathic paragneiss (Missi Group) that contains erratic concentrations of disseminated sulphide.* (Manitoba Geological Survey, 2010, deposit M62N/02-001 at UTM 373296E, 6100193N)

The zones are parallel to the overturned limb of the F_5 Puffy Lake reclined fold but apparently predate it (Manitoba Geological Survey, 2010, deposit M62N/02-001). The D_4 generation of deformation is indicated by early flattening of conglomerate clasts followed by folding in the hinge zone of an F_5 fold. Burntwood Group and Missi Group sedimentary rocks in the hangingwall of the long-lived Puffy Lake Shear Zone (PSZ) underlie the deposit and may have provided the ore fluids with fluids from the shear zone during metamorphism. The presence of biotite and calcsilicate minerals in alteration suggests a relatively deep and hot environment for gold deposition. If this is an orogenic lode gold deposit, the moderately dipping mineralized zones were in the field of extension that caused dilation and vein formation with respect to the steep, sinistral, reverse D_5 displacement on the PSZ.

7.2 Regional economic application of new data and conclusions

The new data and conclusions in this report can assist future discoveries of volcanic massive sulphide (VMS) and gold deposits in the Kississing–File lakes area, as well as in a wider region in the Trans-Hudson Orogen (THO) of Manitoba. The new division into tectonic subdomains (Figure 2) helps to define the mineral potential and metallogeny of different areas. Tracing the tectonic packages and units may provide local controls for the deposits.

Each newly inferred tectonic environment is shown to have existed during a different stage of evolution in the THO, and is associated with a different suite of mineral deposits. Volcanogenic massive sulphide (VMS) deposits (e.g., Dickstone deposit) and smaller occurrences are hosted in early oceanic-arc to arc-rift volcanic rocks, and in volcanic-arc–derived gneiss (Sherridon–Meat Lake assemblage) with a preliminary age of successor-arc magmatism. Precious-metal deposits (e.g., Nokomis Lake) are hosted in layered mafic intrusions (Josland Lake sills) within MORB-like units that formed in a back-arc or arc-rift environment. There may have been a direct source of gold-bearing fluids and platinum-group elements (Peck et al., 2000; Bailes and Theyer, 2006). A gold deposit (Puffy Lake) is interpreted to be late synorogenic and associated with a major shear zone (Puffy Lake Shear Zone).

Newly proposed ideas pertinent to mineral exploration in the evolution of the western THO include the following:

- An early (>1886 Ma) assembly of the entire Flin Flon Domain was probably from a single oceanic arc with an exceptionally high VMS potential that may have spanned the entire western THO.
- This was followed by arc rifting and possible terrane dispersal, which may have left only parts of the arc (Batty Lake subdomain) in a younger fertile marine environment. Later (post–Burntwood Group) reassembly brought these crustal fragments back together, but they can still be identified by their tectonic history.
- The importance of geochemistry, even more than age, is a guide for mineral exploration, as suggested by the Sherridon deposits.
- An extended period of crustal uplift during collision tectonics and final arc magmatism started ca. 1850 Ma and included subaerial arc volcanism not conducive to VMS deposition. The barren Missi and Burntwood groups are distinguished by their age, geochemistry and stratigraphic position.
- Syndepositional block faulting indicated by facies changes in the Missi Group, however, was probably important in preserving early volcanic blocks with their VMS deposits.
- Subsequent, predominantly felsic, weakly peraluminous magmatism was synkinematic during crustal thickening. This occurred during collision-related metamorphism that liberated fluids for orogenic gold deposition.
- The exceptionally good preservation of the main part of the Flin Flon Domain with its mineral deposits is due to late normal faulting and tilting.

Because geochemistry is such an important tool for tracing units with mineral potential and for defining tectonic and metallogenic environments, it is important to have demonstrated, in this report, that many geochemical characteristics are relatively well preserved in the medium- to high-grade metamorphic rocks. Neither the high metamorphic grade nor the complex folding has destroyed mineral deposits or hydrothermal-alteration domains that may lead to future discoveries. The primitive-arc tectonism suggested by the geochemistry of the early juvenile volcanic rocks has involved arc extension, rifting and foundering. High-Mg flows in the Ponton Lake basalt, and mafic-tholeiite–rhyolite assemblages suggest high temperatures that would have contributed to the general abundance of VMS deposits in the arc-volcanic rocks of the Flin Flon Domain. The temporal and spatial transition from more arc-like Preaston andesite to more MORB-like Storozuk basalt, with intervening extension-related Dickstone rhyolite, suggests the onset of arc rifting, a favourable time and site for VMS deposition (Syme et al., 1999). The Dickstone VMS deposit may be related to faulting and attendant hydrothermal-fluid flow during the early stages of arc rifting.

Subaerial successor-arc volcanism, which occurred on thickened crust during the advanced stages of arc magmatism, would not have promoted VMS deposition. However, if the Sherridon–Meat Lake assemblage has a successor-arc age, the Batty Lake subdomain remained as a primitive oceanic environment. The newly obtained preliminary age of about 1850–1855 Ma for the Sherridon rocks suggests that other relatively young rocks may also have potential to host base-metal deposits. Their primitive to chemically evolved compositions and widespread hydrothermal alteration resemble characteristics of the VMS-hosting rocks of the Snow Lake assemblage. Thus, the chemical characteristics of these rocks appear to provide a better guide for their mineral potential than their age. If the trondhjemite of the Batty Lake Complex, with its age and local geochemistry similar to those of the Sherridon–Meat Lake assemblage, is a subvolcanic pluton, it may have provided a heat source for hydrothermal alteration. Its altered mantle and a similar gneiss on Yakushavich Island, farther west, may represent environments more distal to the main Sherridon VMS deposits.

The earlier structural architecture due to (D_3) thrusting of the Amisk Collage over the Burntwood Group greywacke-mudstone turbidite may have been important in providing metamorphic fluids for orogenic gold deposits. Both D_3 (late Missi-age) and D_6 (postcollisional) normal faulting may have been important in preserving abundant mineralized volcanic rocks. The possibility of late mantle delamination during collision and emplacement of the Sask craton is consistent with seismic data that suggest Archean crust exposed as domes in Saskatchewan is structurally overlain by juvenile, Proterozoic, mantle-derived igneous rocks. Thus, the present crustal level of exposure is considered allochthonous over the Archean crust. This old crust has probably retained its thick ancient subcontinental lithospheric mantle, and therefore provides an incentive for diamond exploration in Manitoba.

8 Acknowledgments

We owe a great deal to the many people who did mapping over a long period in, and adjacent to, the Kississing–File lakes area; who assisted us in the field and in the Manitoba Geological Survey office; or who worked together with us from outside agencies. Mapping and sampling in the central part of the Kisseynew Domain in the early 1970s by D. McRitchie, D. Baldwin and T. Frohlinger provided the emplacement order and character of the large intrusions (Baldwin et al., 1979). The geochemical work on these rocks by Hollings and Ansdell (2002) served as an important database, and we thank K. Ansdell for sharing this and shepherding much of our trace-element geochemical work through the laboratory at the University of Saskatchewan.

The Ph.D. dissertation of the junior author at the University of Manitoba, completed in 1979, first established the close link between the lower grade rocks south of File Lake and the gneisses to the north. Thanks to our excellent staff in the rock preparation laboratory, then led by D. Berk, the archived samples and powders were easy to retrieve and send out for analysis.

The senior author thanks E. Froese (now emeritus) of the Geological Survey Canada for patiently introducing him to the ‘chaos in the Kisseynew’ in 1981 after completing his mapping with P. Goetz at Sherridon (Froese and Goetz, 1981) and for continuing this work at Kisseynew and Fay lakes (Froese, 1997). His recognition—much at the prompting of K. Ashton of the Saskatchewan Geological Survey—that large areas of orthogneiss originated as volcanic rocks is another crux of this report (Ashton and Froese, 1988). We also thank him for commenting on Section 4.3 and for questioning the confusing concept of a ‘transition zone’ used in an earlier version of this report. His written comments and questions send periodically allowed us to believe that others were also interested in this complex terrane. The present interpretations reflect the views of the senior author and could be all wrong, but we never gave up.

Between 1985 and 1990, D. Schledewitz did the mapping of the Kississing–Fay lakes area, and P. Lenton mapped most of Batty Lake area, Limestone Point Lake area and adjacent areas to the northeast (Zwanzig and Schledewitz, 1992). Their contributions and discussions were invaluable to this report. Thanks, Dave, for maintaining that some sort of retrodeformation is needed to understand this area. We gave it a shot.

After the forest fire in 1989, when the area had just been mapped at 1:50 000 scale, D. McRitchie, then director, agreed to continue to funding and encouraged us to remap critical areas at 1:20 000 or 1:10 000 scale, now that we could actually see the rocks without their usual moss and lichen cover. We dedicate this report to his memory and his love for high-grade metamorphic rocks.

The senior author greatly enjoyed working with the late N. Machado and his students from the Université du Québec à Montréal (UQUAM) in teasing out some critical isotopic ages (Machado et al., 1999).

The recent VMS exploration and mapping by Halo Resources Ltd., their consultants (E. Buhlmann and B. Barclay) and D. Tinkham of Laurentian University provided another critical source of information on the important Sherridon area. L. Bloom (CEO of Halo Resources Ltd.), J. Krebs (head of Sherridon exploration), T. Jørgensen (principal mapping geologist) and S. Timpa (geochemist) were most generous in sharing their proprietary data and discussing the geology of their holdings.

We thank the many student field assistants during the years of mapping. When we were not crawling through the burned-over deadfall or slogging through swamps and mosquitoes, they cheerfully led our traverse lines, carried many of our rocks and cooked for us back in camp. Some of our common antics are fondly remembered. N. Brandson, our expeditor, always did his best.

We thank our GIS and drafting staff, so ably led by P. Lenton and L. Chackowsky. Thanks, B. Lenton, for helping the senior author when the program crashed once again in preparing the many figures.

C. Böhm (our section head) is thanked for his persistence in finally obtaining a U-Pb zircon age at the University of Alberta

for the critical unit, the Josland Lake gabbro, after attempts by others (Zwanzig et al., 2001). He weeded out some final problems in the latest version of the manuscript after two heroic sessions of editing by the junior author.

9 References

Annesley, I.R., Madore, C. and Portella, P. 2005: Geology and thermotectonic evolution of the western margin of the Trans-Hudson Orogen: evidence from the eastern sub-Athabasca basement, Saskatchewan; Canadian Journal of Earth Sciences, v. 42, p. 573–597.

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, Manitoba and Saskatchewan; *in* Radiogenic Age and Isotopic Studies: Report 7, Geological Survey of Canada, Paper 93-02, p. 49–57.

Ansdell, K.M. 2005: Tectonic evolution of the Manitoba–Saskatchewan segment of the Paleoproterozoic Trans-Hudson Orogen, Canada; Canadian Journal of Earth Sciences, v. 42, p. 741–759.

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

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; *in* NATMAP Shield Margin Project, Volume 1, S.B. Lucas (ed.), Canadian Journal of Earth Sciences, v. 36, no. 2, p. 293–312.

Ashton, K.E. 1999: A proposed lithotectonic domain reclassification of the southeastern Reindeer Zone in Saskatchewan; *in* Summary of Investigations 1999, Volume 1, Saskatchewan Geological Survey, Saskatchewan, Energy and Mines, Miscellaneous Report 99-4.1, p. 92–100.

Ashton, K.E. and Froese, E. 1988: Could the Sherridon Group at Sherridon be a high-grade equivalent of the Amisk Group?; *in* Report of Field Activities 1988, Manitoba Energy and Mines, Minerals Division, p. 183–184.

Ashton, K.E., Heaman, L.M., Lewry, J.F., Hartlaub, R.P. and Shi, R. 1999: Age and origin of the Jan Lake Complex: a glimpse at the buried Archean craton of the Trans-Hudson Orogen; Canadian Journal of Earth Sciences, v. 36, p. 185–108.

Ashton, K.E., Lewry, J.F., Heaman, L.M., Hartlaub, R.P., Stauffer, M.R. and Tran, H.T. 2005: The Pelican thrust zone: basal detachment between Sask craton and the Paleoproterozoic Flin Flon–Glennie Complex, western Trans-Hudson Orogen; Canadian Journal of Earth Sciences, v. 42, p. 685–706.

Bailes, A.H. 1971: Preliminary compilation of the geology of the Snow Lake–Flin Flon–Sherridon area; Manitoba Department of Mines and Natural Resources, Mines Branch, Geological Paper GP71-1, 27 p.

Bailes, A.H. 1975: Geology of the Guay–Wimapedi lakes area; Manitoba Department of 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, Geological Report GR78-1, 134 p.

Bailes, A.H. 1980b: Origin of Early Proterozoic volcaniclastic turbidites, south margin of the Kisseynew sedimentary gneiss belt, File Lake, Manitoba; Precambrian Research, v. 12, p. 197–225.

Bailes, A.H. 1985: Geology of the Saw Lake area; Manitoba Energy and Mines, Geological Report GR83-2, 47 p.

Bailes, A.H. and Galley, A.G. 1999: Evolution of the Paleoproterozoic Snow Lake arc assemblage and geodynamic setting for associated volcanic-hosted massive sulphide deposits, Flin Flon Belt, Manitoba, Canada; Canadian Journal of Earth Sciences, v. 36, p. 1789–1805.

Bailes, A.H. and Galley, A.G. 2001: Geochemistry and tectonic setting of volcanic and intrusive rocks in the VMS-hosting Snow Lake area assemblage, Flin Flon Belt, Manitoba: a preliminary release of the geochemical data set; Manitoba Industry, Trade and Mines, Manitoba Geological Survey, Open File OF2001-6, 1 CD-ROM.

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, Geological Services, Geological Report GR87-1-1, 1 map at 1:20 000 scale.

Bailes, A.H. and Theyer, P. 2006: Wonderland gabbro: U-Pb age and geological implications, Manitoba (NTS 63K12NE); *in* Report of Activities 2006, Manitoba Science, Technology, Energy and Mines, Manitoba Geological Survey, p. 35–41.

Bailes, A.H., Hunt, P.A. and Gordon, T.M. 1991: U-Pb zircon dating of possible synvolcanic plutons in the Flin Flon belt at Snow Lake, Manitoba; *in* Radiogenic Age and Isotopic Studies: Report 4, Geological Survey of Canada, Paper 90-02, p. 35–43.

Bailes, A.H., Gilmore, K., Levers, J. and Janser, B. 2009: The Lalor deposit—surprise at depth (abstract); Manitoba Innovation, Energy and Mines, Manitoba Geological Survey, Manitoba Mining and Minerals Convention, URL <http://www.gov.mb.ca/stem/mrd/minerals/convention/abstracts.html#lalor> [June 2010].

Baldwin, D.A., Frohlinger, T.G., Kendrick, G., McRitchie, W.D. and Zwanzig, H.V. 1979: Geology of the Nelson House–Pukatawagan region (Burntwood Project); Manitoba Department of Mines, Natural Resources and Environment, Mineral Resources Division, Geological Report GR78-3 (pt. III), Maps 78-3-1 to 78-3-22, scale 1:50 000.

Barnes, C.G., Yoshinobu, A.S., Prestvik, T., Nordgule, Ø., Karlsson, H.R. and Sundvoll, B. 2002: Mafic magma intraplating: anatexis and hybridization in arc crust, Bindoal Batholith, Norway; Journal of Petrology, v. 43, p. 2171–2190.

Bateman, J.D. and Harrison, J.M. 1946: Sherridon, Manitoba; Geological Survey of Canada, Map 862A, scale 1:63 360.

Beaumont-Smith, C.J. and Böhm, C.O. 2002: Structural analysis and geochronological studies in the Lynn Lake greenstone belt and its gold-bearing shear zones (NTS 64C10, 11, 12, 14, 15 and 16), Manitoba; *in* Report of Activities 2002, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 159–170.

Berman, R.G. 1988: Internally consistent thermodynamic data for minerals in the system Na_2O-K_2O-CaO-MgO-FeO-Fe_2O_3-Al_2O_3-SiO_2-TiO_2-H_2O-CO_2; Journal of Petrology, v. 29, p. 445–522.

Berman, R.G. 1991: Thermobarometry using multi-equilibrium calculations: a new technique, with petrological applications; Canadian Mineralogist, v. 76, p. 1223–1231.

Bleeker, W. 1990: Evolution of the Thompson Nickel Belt and its nickel deposits, Manitoba, Canada; Ph.D. thesis, University of New Brunswick, 444 p.

Böhm, C.O., Zwanzig, H.V. and Creaser, R.A. 2007: Sm-Nd isotope technique as an exploration tool: delineating the northern extension of the Thompson Nickel Belt, Manitoba, Canada; Economic Geology, v. 102, p. 1217–1231.

Bouma, A.H. 1962: Sedimentology of some Flysch Deposits: A Graphic Approach to Facies Interpretation; Elsevier Publishing Co., New York, 168 p.

Brown, D. and Spadea, P. 1999: Processes of forearc and accretionary complex formation during arc-continent collision in the southern Ural Mountains; Geology, v. 27, p. 649–552.

Burnham, M., Halden, N., Layton-Matthews, D., Lesher, M., Liwanag, J., Heaman, L., Hulbert, L., Machado, N., Michalak, D., Pacey, M., Peck, D., Potrel, A., Theyer, P., Toope, K. and Zwanzig, H. 2009: CAMIRO Project 97E-02, Thompson Nickel Belt: final report March 2002, revised and updated 2003; Manitoba Science, Technology, Energy and Mines, Manitoba Geological Survey, Open File OF2008-11, 1 CD-ROM.

Cabanis, B. and Lecolle, M. 1989: The La/10–Y/15–Nb/8 diagram: a tool for distinguishing volcanic series and discovering crustal mixing and/or contamination; Comptes Rendus de l'Academie des Sciences, Serie 2, Mecanique, Physique, Chimie, Sciences de l'Univers, Sciences de la Terre, v. 309, no. 20, p. 2023–2029.

Carn, S.A. and Pyle, D.M. 2001: Petrology and geochemistry of the Lamongan volcanic field, east Java, Indonesia: primitive Sunda Arc magmas in an extensional setting; Journal of Petrology, v. 42, p. 1643–1683.

Churikova, T., Sorendorf, F. and Wörner, G. 2001: Source of fluids in the mantle wedge below Kamchatka, evidence from across-arc geochemical variation; Journal of Petrology, v. 12, p. 1567–1593.

Clarke, D.B., Andrew S.H. and Hamilton, M.A. 2005: Composition, age, and origin of granitoid rocks in the Davin Lake area, Rottenstone Domain, Trans-Hudson Orogen, northern Saskatchewan; Canadian Journal of Earth Sciences, v. 42, p. 599–633.

Clift, P.D., Draut, A.E., Kelemen, P.B., Blusztajn, J. and Greene, A. 2005: Stratigraphic and geochemical evolution of an oceanic arc upper crustal section: the Jurassic Talkeetna volcanic formation, south-central Alaska; Geological Society of America Bulletin, v. 117, p. 902–925.

Condie, K.C. 2008: Did the character of subduction change at the end of the Archean? Constraints from convergent-margin granitoids; Geology, v. 36, no. 8, p. 611–614; Geological Society of America, Data Repository item 2008148.

Connors, K.A. 1996: Unravelling the boundary between the turbidites of the Kisseynew Belt and the volcano-plutonic rocks of the Flin Flon Belt, Trans-Hudson Orogen, Canada; Canadian Journal of Earth Sciences, v. 33, p. 811–829.

Connors, K.A., Ansdell, K.M. and Lucas, S.B. 1999: Coeval sedimentation, magmatism, and fold-thrust development in the Trans-Hudson Orogen: propagation of deformation into an active continental arc setting, Wekusko Lake area, Manitoba; *in* NATMAP Shield Margin Project, Volume 1, S.B. Lucas (ed.), Canadian Journal of Earth Sciences, v. 36, no. 2, p. 275–291.

Corrigan, D. and Rayner, N. 2002: Churchill River–Southern Indian Lake Targeted Geoscience Initiative (NTS 64B, 64C, 64G, 64H), Manitoba: update and new findings; *in* Report of Activities 2002, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 144–158.

Corrigan, D., Hajnal, Z., Németh, B. and Lucas, S.B. 2005: Tectonic framework of the Paleoproterozoic arc-continent collision zone, Trans-Hudson Orogen, from geological and seismic reflection studies; Canadian Journal of Earth Sciences, v. 42, p. 421–434.

Cruciani, G., Francesschelli, M., Elter, F.M., Puxeddu, M. and Utzeri, D. 2007: Petrogenesis of Al-silicate–bearing trondhjemitic migmatites from NE Sardinia, Italy; Lithos, v. 102, p. 554–574.

Culshaw, N.G., Beaumont, C. and Jamieson, R.A. 2006: The orogenic suprastructure-infrastructure concept: revisited, quantified, revived; Geological Society of America Bulletin, v. 34, p. 733–736.

Currie, J.B. 1947: Zones of metamorphism in the greenstones of the Morton Lake area, Manitoba; M.Sc. thesis, University of Toronto, 35 p.

Currie, K.L., Whalen, J.B., Davis, W.J., Longstaffe, F.J. and Cousens, B.L. 1998: Geochemical evolution of the peraluminous plutons in southern Nova Scotia, Canada—a pegmatite-poor suite; Lithos, v. 44, p. 117–140.

David, J. and Syme, E.C. 1994: U-Pb geochronology of late Neoarchean tonalites in the Flin Flon Belt, Trans-Hudson Orogen: surprise at surface; Canadian Journal of Earth Sciences, v. 31, no. 12, p. 1785–1790.

David, J., Machado, N., and Bailes, A.H. 1996: Evolution of the Snow Lake portion of the Proterozoic Flin Flon and Kisseynew belts, Trans-Hudson Orogen, Manitoba, Canada; Precambrian Research, v. 82, p. 107–124.

Defant, M.J., Richerson, P.M., De Boer, J.Z., Stewart, R.H., Maury, R.C., Bellon, H., Drummond, M.R., Feigenson, M.D. and Jackson, T.E. 1991: Dacite genesis via both slab melting and differentiation: petrogenesis of the Yeguada volcanic complex, Panama; Journal of Petrology, v. 32, p. 1101–1142.

Devaney, J.R. and Williams, H.R. 1989: Evolution of an Archean subprovince boundary: a sedimentological and structural study of part of the Wabigoon-Quetico boundary in northern Ontario; Canadian Journal of Earth Sciences, v. 26, p. 1013–1023.

Digel, S. and Gordon, T.M. 1995: Phase relations in metabasites and pressure-temperature conditions at the prehnite-pumpellyite to greenschist facies transition, Flin Flon, Manitoba, Canada; *in* Low-Grade Metamorphism of Mafic Rocks, P. Schiffman and H.W. Day (ed.), Geological Society of America, Special Paper 296, p. 67–80.

Drummond, M.S. and Defant, M.J. 1990: A model for trondhjemite-tonalite-dacite genesis and crustal growth via slab melting: Archean to modern comparisons; Journal of Geophysical Research, v. 95, p. 21503–21521.

Ducea, M.N. and Barton, M.D. 2007: Igniting flare-up events in Cordilleran arcs; Geology, v. 35, p. 1047–1050.

Ewart, A., Collerson, K.D., Regelous, M., Wendt, J.I. and Niu, Y. 1998: Geochemical evolution within the Tonga–Kermadec–Lau Arc backarc system: the role of varying mantle wedge composition in time and space; Journal of Petrology, v. 39, p. 331–368.

Ferrari, L., Lopez-Martinez, M. and Rosas-Elguera, J. 2002: Ignimbrite flare-up and deformation in southern Sierra Madre Occidentali, western Mexico; Tectonics, v. 21, no. 4, p. 17-1–17-24.

Ferreira, K.J. 2006: Technical report on the Sherridon VMS property, north-central Manitoba, Canada; prepared for Halo Resources Ltd., URL <http://www.halores.com/investors/pdf/sherridon-43-101.pdf> [June 2010].

Fornelli, A., Piccarreta, A., Del Moro, A. and Acquafredda, P. 2002: Multi-stage melting in the lower crust of the Serre (southern Italy); Journal of Petrology, v. 43, no. 12, p. 2191–2217.

Fralick, P, Purdon, R.H. and Davis, W.D. 2006: Neoarchean trans-subprovince sediment transport in the southwestern Superior Province: sedimentological, geochemical and geochronological evidence; Canadian Journal of Earth Sciences, v. 43, p. 1055–1070.

Fretzdorff, S., Livermore, R.A., Devey, C.W., Leat, P.T. and Stoffers, P. 2002: Petrogenesis of the backarc East Scotia ridge, South Atlantic Ocean; Journal of Petrology, v. 43, no. 8, p. 1435–1467.

Froese, E. 1989: Metamorphosed alteration zones; *in* Investigations by the Geological Survey of Canada in Manitoba and Saskatchewan during the 1984–1989 Mineral Development Agreements, A.G. Galley (ed.), Geological Survey of Canada, Open File 2133, p. 1–2.

Froese, E. 1997: Metamorphism in the Weldon Bay–Syme Lake area, Manitoba; *in* Current Research 1997-E, Geological Survey of Canada, p. 35–44.

Froese, E. and Gall, Q. 1981: Geology of the eastern vicinity of Kisseynew Lake, Manitoba; *in* Current Research, Part A, Geological Survey of Canada, Paper 81-1A, p. 311–313.

Froese, E. and Gasparrini, E. 1975: Metamorphic zones in Snow Lake area, Manitoba; Canadian Mineralogist, v. 13, p. 162–167.

Froese, E. and Goetz, P.A. 1981: Geology of the Sherridon Group in the vicinity of Sherridon, Manitoba; Geological Survey of Canada, Paper 80-21, 20 p.

Froese, E. and Moore, J.M. 1980: Metamorphism in the Snow Lake area, Manitoba; Geological Survey of Canada, Paper 78-27, 16 p.

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

Gale, G.H. 1980: Mineral deposit studies: Flin Flon–Kisseynew; *in* Report of Field Activities 1980, Manitoba Energy and Mines, Mineral Resources Division, p. 51–64.

Gale, D.F., Lucas, S.B. and Dixon, J.M. 1999: Structural relations between the polydeformed Flin Flon arc assemblage and Missi group sedimentary rocks, Flin Flon area, Manitoba and Saskatchewan; Canadian Journal of Earth Sciences, v. 36, no. 11, p. 1901–1915.

Gall, Q. 1981: Geology of the Defender Lake gneissic dome, Kisseynew metamorphic complex, Manitoba; Ph.D. thesis, Carlton University, Ottawa, 33 p.

Gamble, J.A., Smith, I.E.M., McCulloch, I.J. and Kokelaar, B.P. 1993: The geochemistry and petrogenesis of basalts from the Taupo volcanic zone and Kermadec Island arc, S.W. Pacific; Journal of Volcanology and Geothermal Research, v. 54, p. 265–290.

Gertisser, R. and Keller, J. 2003: Trace element and Sr, Nd, Pb and O isotope variations in medium-K and high-K volcanic rocks from Merapi Volcano, central Java, Indonesia: evidence for involvement of subducted sediments in the Sunda Arc magma genesis; Journal of Petrology, v. 44, p. 457–489.

Gilbert, H.P. 1989: Geological investigations in the Tartan Lake–Embury Lake area; *in* Report of Field Activities 1989, Manitoba Energy and Mines, Minerals Division, p. 19–30.

Gilbert, H.P. 2004: Geological investigations in the northern Flin Flon Belt, Manitoba (parts of NTS 63K13NE and 63K14NW); *in* Report of Activities 2004, Manitoba Industry, Economic Development and Mines, Manitoba Geological Survey, p. 9–23.

Giroux, G.H. and Moore, C.M. 2007: Technical report on Jungle Lake deposit, north central Manitoba, Canada; NI 43-101 report prepared for Halo Resources Ltd. by Scott Wilson Roscoe Postle Associates Inc., URL <http://www.halores.com/projects/sherridon/jungle-lake.html> [June 2010].

Glazner, A.F. 2007: Thermal limitations on incorporation of wall rock into magma; Geology, v. 34, no. 4, p. 319–322.

Goetz, P.A. 1980: Depositional environment of the Sherridon Group and related mineral deposits near Sherridon, Manitoba; Ph.D. thesis, Carleton University, 248 p.

Goodwin, A.M., Lambert, M.B. and Ujike, O. 2006: Geochemical and metallogenic relations in volcanic rocks of the southern Slave Province: implications for late Neoarchean tectonics; Canadian Journal of Earth Sciences, v. 43, no. 12, p. 1835–1857.

Gordon, T.M. 1989: Thermal evolution of the Kisseynew sedimentary gneiss belt, Manitoba: metamorphism at an early Proterozoic accretionary margin; *in* Evolution of Metamorphic Belts: Proceedings of the 1987 Joint Meeting of the Metamorphic Studies Group and IGCP Project 235, J.S. Daly, R.A. Cliff and B.W.D. Yardley (ed.), Geological Society, Dublin, Ireland, Special Publication 43, p. 233–243.

Gordon, T.M. and Lemkow, D.R. 1987: Geochemistry of the Missi Group volcanic rocks, Wekusko Lake, Manitoba; Geological Survey of Canada Open File 1442, 38 p.

Gordon, T.M., Aranovich, L.Y. and Fed'kin, V.V. 1993: Mineral assemblage and thermobarometric constraints on the temperature-depth history of the Kisseynew and Flin Flon belts, Manitoba; *in* LITHOPROBE: Trans-Hudson Orogen Transect: Report of Third Transect Meeting, Z. Hajnal and J.F. Lewry (ed.), LITHOPROBE Secretariat, University of British Columbia, LITHOPROBE Report 34, p. 131–138.

Gordon, T.M., Aranovich, L.Y. and Fed'kin, V.V. 1994: Exploratory data analysis in thermobarometry: an example from the Kisseynew sedimentary gneiss belt, Manitoba, Canada; American Mineralogist, v. 79, no. 9–10, p. 973–982.

Gordon, T.M., Hunt, P.A., Bailes, A.H. and Syme, E.C. 1990: U-Pb zircon ages from the Flin Flon and Kisseynew belts, Manitoba: geochronology 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.

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, p. 116, p. 281–322.

Gribble, R.F., Stern, R.J., Newman, S., Bloomer, S.H. and O'Hearn, T. 1998: Geochemical and isotopic composition of lavas from the northern Mariana Trough: implications for magma genesis in backarc basins; Journal of Petrology, v. 39, p. 125–154.

Growdon, M.L., Percival, J.A., Rayner, N. and Murphy, L. 2006: Regional granulite-facies metamorphism in the northeastern Kisseynew Domain, Manitoba (parts of NTS 63O); *in* Report of Activities 2006, Manitoba Science, Technology, Energy and Mines, Manitoba Geological Survey, p. 104–115.

Growdon, M., Percival, J., Rayner, N. and Wintsch, R. 2008: Rapid loading and heating to granulite facies in the eastern Trans-Hudson Orogen, Manitoba, Canada; Geological Society of America, 2008 Annual Meeting, October 5–9, 2008, Houston, TX, Abstracts with Programs, v. 40, no. 6, p. 204.

Hajnal, Z., Green, A., Lucas, S.B., White, D., Stauffer, M., Ashton, K., Weber, W. and Clowes, R. 1994: Structure of the Paleoproterozoic continent-continent collision zone: a LITHOPROBE seismic reflection profile across the Trans-Hudson Orogen, Canada; *in* Proceedings of the 5th International Conference on Seismic Reflection Probing of Continents and their Margins, R.M. Clowes and A.G. Green (ed.), Tectonophysics, v. 232, p. 143–160.

Hajnal, Z., Lewry, J., White, D., Ashton, K., Clowes, R., Stauffer, M., Gyorfi, I. and Takacs, E. 2005: The Sask Craton and Hearne Province margin: seismic reflection studies in the western Trans-Hudson Orogen; Candian Jounal of Earth Sciences, v. 42, no 4, p. 403–419.

Hajnal, Z., Lucas, S.B., White, D.J., Lewry, J., Bezdan, S., Stauffer, M.R. and Thomas, M.D. 1996: Seismic reflection images of strike-slip faults and linked detachments in the Trans-Hudson Orogen; Tectonophysics, v. 15, p. 427–439.

Halo Resources Ltd. 2009: Sherridon VMS property; Halo Resources Ltd., URL <http://www.halores.com/projects/sherridon/index.html> [April 2009].

Harrison, J.M. 1949: Geology and mineral deposits of File–Tramping lakes area, Manitoba; Geological Survey of Canada, Memoir 250, 92 p.

Healy, T. and Krebs J. 2008: From zero to 13 million tonnes in 18 months: the Sherridon success story [abstract]; Manitoba Mining and Minerals Convention 2008, Winnipeg, Manitoba, November 20–22, 2008, Program, p. 49.

Heaman, L.M., Machado, N., Krogh, T.E. and Weber, W. 1986: Precise U-Pb zircon ages for the Molson dyke swarm and the Fox River sill: constraints for early Proterozoic crustal evolution in northeastern Manitoba, Canada; Contributions to Mineralogy and Petrology, v. 94, no. 1, p. 82–89.

Heaman, L.M., Kamo, S.L., Ashton, K.E., Reilly, B.A., Slimmon, W.L. and Thomas, D.J. 1992: U-Pb geochronological investigations, 1992, Saskatchewan Energy and Mines, Miscellaneous Report 92-4, p. 120–123.

Herzberg, C., Condie, K. and Korenaga, J. 2010: Thermal history of the Earth and its petrological expression; Earth and Planetary Science Letters, v. 292, p. 79–88.

Hildreth, W. and Moorbath, S. 1988: Crustal contribution to arc magmatism in the Andes of central Chile; Contributions to Mineralogy and Petrology, v. 98. p. 455–489.

Hollings, P. and Ansdell, K. 2002: Paleoproterozoic arc magmatism imposed on an older backarc basin: implications for the tectonic evolution of the Trans-Hudson Orogen, Canada: Geological Society of America Bulletin, v. 114, p. 153–158.

Holm, D.K., Van Schmus, W.R., MacNeill, L.C., Boerboom, T.J., Schweitzer, D. and Schneider, D. 2005: U-Pb zircon geochronology of Paleoproterozoic plutons from the northern midcontinent, USA: evidence of subduction flip and continued convergence after geon 18 Penokean orogenesis; Geological Society of America Bulletin, v. 117, p. 259–275.

Huijsmans, J.P., Barton, M. and Salters, V.J.M. 1988: Geochemistry and evolution of the calcalkaline volcanic complex of Santorini, Aegean Sea, Greece; Journal of Volcanology and Geothermal Research, v. 34, p. 283–306.

Hulbert, L.J, Hamilton, M.A., Horan, M.F. and Scoates, R.F.J. 2005: U-Pb zircon and Re-Os isotope geochronology of mineralized ultramafic intrusions and associated nickel ores from the Thompson nickel belt, Manitoba, Canada; Economic Geology, v. 100, no. 1, p. 29–41.

Hunt, P.A. and Schledewitz, D.C.P. 1992: U-Pb zircon age for a quartz porphyry in the Thunderhill Lake area, Kisseynew gneiss belt, Manitoba; *in* Radiogenic Age and Isotopic Studies: Report 5, Geological Survey of Canada, Paper 91-02, p. 51–54.

Hunt, P.A. and Zwanzig, H.V. 1990: Pre-Missi granitoid domes in the Puffy Lake area, Kisseynew gneiss belt, Manitoba; *in* Radiogenic Age and Isotopic Studies: Report 3, Geological Survey of Canada, Paper 89-2, p. 71–76.

Hunt, P.A. and Zwanzig, H.V. 1993: U-Pb ages of intrusion, metamorphism and deformation in the Batty Lake area, Kisseynew gneiss belt, Manitoba; *in* Report of Activities 1993, Manitoba Energy and Mines, Geological Services, p. 34–37

Irvine, T.N. and Baragar, W.R.A. 1971: A guide to the chemical classification of common volcanic rocks; Canadian Journal of Earth Sciences, v. 8, p. 523–548.

Jenner, G.A., Longerich, H.P., Jackson, S.E. and Fryer, B.J. 1990: ICP-MS—a powerful tool for high precision trace-element analysis in Earth sciences: evidence from analysis of selected USGS reference samples; Chemical Geology, v. 83, p. 133–148.

Jolly, W.T., Lidiak, E.G., Dickin, A.P. and Wu, T-W. 2001: Secular geochemistry of central Puerto Rican island arc lavas: constraints on Mesozoic tectonism in the eastern Greater Antilles; Journal of Petrology, v. 42, p. 2197–2214.

Jordan, T.E., Burns, W.M., Veiga, R., Pángaro, F., Copeland, P., Kelly, S. and Mpodozis, C. 2001: Extension and basin formation in the southern Andes caused by increased convergence rate: a mid-Cenozoic trigger for the Andes; Tectonics, v. 20, p.308–324.

Jungwirth, T. 1995: Metamorphism of the Duval Lake area, Kisseynew gneiss belt, Manitoba; M.Sc. thesis, University of Calgary, 165 p.

Jungwirth, T., Gordon, T.M. and Froese, E. 2000: Metamorphism of the Burntwood Group in the Duval Lake area, Manitoba; *in* Tectonometamorphic Studies in the Canadian Shield (Part II), R.G. Berman, R.M. Easton and R.F. Martin (ed.), The Canadian Mineralogist, v. 38, pt. 2, p. 435–442.

Kreis, L.K., Ashton, K.E. and Maxeiner, R.O. 2000: Geology of the Precambrian basement and Phanerozoic strata in Saskatchewan, Lower Paleozoic Map Series, Sheet 1 of 8; Saskatchewan Energy and Mines, Open File Report 2000-2.

Kraus, J. 1998: Structural and metamorphic studies in the Snow Lake area, Trans-Hudson Orogen, Manitoba, central Canada; Ph.D. thesis, University of New Brunswick, Fredericton, 226 p.

Kraus, J. and Menard, T. 1997: A thermal gradient at constant pressure; implications for low- to medium-pressure metamorphism in a compressional tectonic setting, Flin Flon and Kisseynew domains, Trans-Hudson Orogen, central Canada; *in* Tectonometamorphic Studies in the Canadian Shield: Part 1, R.G. Berman and R.M. Easton (ed.), The Canadian Mineralogist, v. 35, no. 5, p. 1117–1136.

Kraus, J. and Williams, P.F. 1999: Structural development of the Snow Lake allochthon and its role in the evolution of the southeastern Trans-Hudson Orogen in Manitoba, central Canada; Canadian Journal of Earth Sciences, v. 36, no. 11, p. 1881.

Le Bas, M.J., Le Maitre, R.W., Streckeisen, A. and Zanetti, B. 1986: Chemical classification of volcanic rocks based on the total alkali–silica diagram; Journal of Petrology, v. 27, p. 745–750.

Leclair, A.D., Lucas, S.B., Broome, H.J., Viljoen, D. 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, no. 5, p. 618–634.

Leo, G.W. 1985: Trondhjemite and metamorphosed quartz keratophyre tuff of the Ammonoosuc Volcanics (Ordovician), western New Hampshire and adjacent Vermont and Massachusetts; Geological Society of America Bulletin, v. 96, no. 12, p. 1493–1507.

Lesher, C.M., Goodwin, A.M., Campbell, I.H. and Gorton, M.P. 1986: Trace element geochemistry of ore-associated and barren felsic metavolcanic rocks in the Superior Province, Canada; Canadian Journal of Earth Sciences, v. 23, p. 222–237.

Lewry, J.F., Collerson, K.D., Bickford, M.E. and Van Schmus, W.R. 1986: An evolutionary model of the western Churchill Province and western margin of the Superior Province in Canada and the north-central United States—discussion; Tectonophysics, v. 131, no. 1–2, p. 183–188.

Lewry, J.F., Hajnal, Z., Green, A.H., Lucas, S.B., White, D.J., Stauffer, M.R., Ashton, K.E., Weber, W. and Clowes, R.M. 1994: Structure of a Paleoproterozoic continent-continent collision zone: a LITHOPROBE seismic reflection profile across the Trans-Hudson Orogen, Canada; *in* Seismic Reflection Probing of the Continents and their Margins, R.M. Clowes and A.G. Green (ed.), Tectonophysics, v. 232, no. 1–4, p. 143–160.

Longerich, H.P., Jenner, G.A., Fryer, B.J. and Jackson, S.E. 1990: Inductively coupled plasma–mass spectrometric analysis of geological samples: a critical evaluation based on case studies; Chemical Geology, v. 83, p. 105–118.

Lucas, S.B., Syme, E.C. and Ashton, K.E. 1999: New perspectives on the Flin Flon Belt, Trans-Hudson Orogen, Manitoba and Saskatchewan: an introduction to the special issue on the NATMAP Shield Margin Project, Part I; *in* NATMAP Shield Margin Project, Volume 1, S.B. Lucas (ed.), Canadian Journal of Earth Sciences, v. 36, no. 2, p. 135–140.

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

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 Report OF2006-33, CD-ROM.

Machado, N.D., Zwanzig, H.V. and Parent, M. 1999: U-Pb ages of plutonism, sedimentation, and metamorphism of the Paleoproterozoic Kisseynew metasedimentary belt, Trans-Hudson Orogen (Manitoba, Canada); Canadian Journal of Earth Sciences, v. 36, no. 11, p. 1829.

Mahlburg Kay, S., Godoy, E. and Kurtz, A. 2005: Episodic arc migration, crustal thickening, subduction erosion, and magmatism in the south-central Andes; Geological Society of America Bulletin, v. 117, p. 67–88.

Maniar, P.D. and Piccoli, P.M. 1989: Tectonic discrimination of granitoids; Geological Society of America Bulletin, v. 101. p. 635–643.

Manitoba Energy and Mines 1986: Granville Lake, NTS 64C; Manitoba Energy and Mines, Geological Services Branch, Bedrock Geology Compilation Map Series, Map 64C, scale 1:250 000.

Manitoba Geological Survey 2006: Manitoba Geochronology Database; Manitoba Science, Technology, Energy and Mines, Manitoba Geological Survey, Open File Report OF2006-34, digital web release, version 1.51, URL <http://www.gov.mb.ca/iedm/mrd/info/libmin/OF2006-34.zip> [July 2010].

Manitoba Geological Survey 2010: Manitoba Mineral Deposits Database; Manitoba Innovation, Energy and Mines, Manitoba Geological Survey, GIS Map Gallery, deposit numbers [July, 2010]:
Deposit M63K/14-001, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63k_14_001.pdf>
Deposit M63K/14-002, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63k_14_002.pdf>
Deposit M63K/14-005, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63k_14_005.pdf>
Deposit M63K/14-007, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63k_14_007.pdf>
Deposit M63K/14-008, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63k_14_008.pdf>
Deposit M63K/16-058, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63k_16_058.pdf>
Deposit M63N/02-001, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_02_001.pdf>
Deposit M63N/02-002, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_02_002.pdf>
Deposit M63N/02-003, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_02_003.pdf>
Deposit M63N/02-004, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_02_004.pdf>
Deposit M63N/02-011, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_02_011.pdf>
Deposit M63N/02-027, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_02_027.pdf>
Deposit M63N/02-037, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_02_037.pdf>
Deposit M63N/03-001, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_03_001.pdf>
Deposit M63N/03-002, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_03_002.pdf>
Deposit M63N/03-003, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_03_003.pdf>
Deposit M63N/03-005, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_03_005.pdf>
Deposit M63N/03-006, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_03_006.pdf>
Deposit M63N/03-041, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_03_041.pdf>
Deposit M63N/03-042, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_03_042.pdf>
Deposit M63N/03-043, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_03_043.pdf>
Deposit M63N/03-044, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_03_044.pdf>
Deposit M63N/03-045, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_03_045.pdf>
Deposit M63N/03-050, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_03_050.pdf>
Deposit M63N/03-054, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_03_054.pdf>
Deposit M63N/03-059, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_03_059.pdf>
Deposit M63N/03-060, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_03_060.pdf>
Deposit M63N/03-077, URL <http://www.gov.mb.ca/stem/mrd/geo/gis/mds/m63n_03_077.pdf>

Maxeiner, R.O. and Sibbald, T.I.I. 1995: Controversial rocks in the Hebden–MacKay lakes area, southern La Ronge Domain; *in* Summary of Investigations 1995, Saskatchewan Geological Survey, Saskatchewan Energy and Mines, Miscellaneous Report 95-4, p. 79–85.

Menard, T. and Gordon, T.M. 1997: Metamorphic P-T paths from the eastern Flin Flon Belt and Kisseynew Domain, Snow Lake, Manitoba; *in* Tectonometamorphic Studies in the Canadian Shield: Part 1, R.G. Berman and R.M. Easton, R.M. (ed.), The Canadian Mineralogist, v. 35, no. 5, p. 1093–1115.

Mitropoulos, P. and Tarney, J. 1992: Significance of mineral composition variations in the Aegean Island Arc; Journal of Volcanology and Geothermal Research, v. 51, p. 283–303.

Moore, J.M. and Froese, E. 1972: Geological setting of the Snow Lake area; *in* Report of Activities, Part B, Geological Survey of Canada, Paper 72-1B, p. 78–81.

Morris, G.A. and Creaser, R.A. 2008: Correlation of mid-Cretaceous granites with source terranes in the northern Canadian Cordillera; Canadian Journal of Earth Sciences, v. 45, p. 389–403.

Mpodozis, C. and Allmendinger, R.W. 1993: Extensional tectonics, Cretaceous Andes, northern Chile (27°S); Geological Society of America Bulletin, v. 105, p. 1462–1477.

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

Németh, B., Clowes, R.M. and Hajnal, Z. 2005: Lithospheric structure of the Trans-Hudson Orogen from seismic refraction–wide-angle reflection studies; *in* The Trans-Hudson Orogen Transect of LITHOPROBE, Z. Hajnal, K.M. Ansdell and K.E. Ashton (ed.), Canadian Journal of Earth Sciences, v. 42, no. 4, p. 435–456.

Norman, A.R., Williams, P.F. and Ansdell, K.M. 1995: Early Proterozoic deformation along the southern margin of the Kisseynew gneiss belt, Trans-Hudson Orogen: 30 million year progressive deformation; Canadian Journal of Earth Sciences, v. 32, p. 875–894.

Norquay, L.I. 1997: Structural and metamorphic evolution of the North Star Lake area, Manitoba; M.Sc. thesis, University of Manitoba, Winnipeg, 244 p.

Nyström, J.O., Vergara, M., Morata, D. and Levi, B. 2003: Tertiary volcanism during extension in the Andean foothills of central Chile (33°15'–33°45'); Geological Society of America Bulletin, v. 115, p.1523–1537.

O'Hanley, D.S. and Kyser, T.K. 1994: Petrogenesis of the Alaskan-type mafic and ultramafic Boundary Intrusions, Flin Flon Domain, Trans-Hudson Orogen; *in* LITHOPROBE Trans-Hudson Orogen Transect: Report of Fifth Transect Meeting, Z. Hajnal and J.F. Lewry (ed.), April 3–4, 1995, LITHOPROBE Secretariat, University of British Columbia, LITHOPROBE Report 48, p. 80–89.

Ostry, G. and Trembath, G.C. 1992: Mineral deposits and occurrences in the Batty Lake area, NTS 63N/2; Manitoba Energy and Mines, Mineral Deposit Series Report No. 19, 264 p.

Ostry, G., Athayde, P. and Trembath, G.D. 1998: Mineral deposits and occurrences in the Sherridon area, NTS 63N/3; Manitoba Energy and Mines, Geological Services, Mineral Deposit Series Report No. 17, 157 p.

Parent, M., Machado, N. and Zwanzig, H.V. 1999: Comportement du systeme U-Pb dans la monazite et chronologie de la deformation et du metamorphisme des metasediments du domaine de Kisseynew, orogene transhudsonien (Manitoba, Canada); Canadian Journal of Earth Sciences, v. 36, no. 11, p. 1843–1857.

Patiño Douce, A.E. and Harris, N. 1998: Experimental constraints on Himalayan anatexis; Journal of Petrology, v. 36, p. 707–738.

Pattison, D.R.M., Chacko, T., Farquhar, J. and McFarlane, R.M. 2003: Temperature of granulite-facies metamorphism: constraints from experimental phase equilibria and thermobarometry corrected for retrograde exchange; Journal of Petrology, v. 44, p. 867–900.

Pearce, J.A. 1995: Tectonic implications of the composition of volcanic arc magmas; Annual Reviews of Earth and Planetary Science, v. 23, p. 251–285.

Pearce, J.A. and Peate, D.W. 1995: Tectonic implication of the composition of volcanic arc magmas; Annual Reviews of Earth and Planetary Science, v. 23, p. 251–285.

Pearce, J.A., Harris, N.B.W. and Tindle, A.G. 1984: Trace element discrimination diagrams for the tectonic interpretation of granitic rocks; Journal of Petrology, v. 25, p. 956–983.

Peccerillo, A. and Taylor, S.R. 1976: Geochemistry of the calcalkaline volcanic rocks from the Kastamonu area, northern Turkey; Contributions to Mineralogy and Petrology, v. 58, p. 63–61.

Peck, D.C., Theyer, P., Hulbert, L., Xiong, J., Fedikow, M.A.F. and Cameron, H.D.M. 2000: Preliminary exploration database for platinum-group elements in Manitoba; Manitoba Industry, Trade and Mines, Manitoba Geological Survey, Open File Report OF2000-5, URL <http://www.gov.mb.ca/stem/mrd/info/libmin/OF2000-5.zip> [June 2010].

Percival, J.A. and Mortensen, J.K. 2002: Water-deficient calcalkaline plutonic rocks of the northeastern Superior Province, Canada: significance of charnockitic magmatism; Journal of Petrology, v. 43, p. 1617–1650.

Percival, J.A., Whalen, J.B. and Rayner, N. 2005: Pikwitonei–Snow Lake Manitoba transect (parts of NTS 63J, 63O and 63P), Trans-Hudson Orogen–Superior Margin Metallotect Project: new results and tectonic interpretation; *in* Report of Activities 2005, Manitoba Industry, Economic Development and Mines, Manitoba Geological Survey, p. 69–91.

Percival, J.A., Zwanzig, H.V. and Rayner, N. 2006: New tectonostratigraphic framework for the northeastern Kisseynew Domain, Manitoba (parts of NTS 63O); *in* Report of Activities 2006, Manitoba Science, Technology, Energy and Mines, Manitoba Geological Survey, p. 74–84.

Pollock, G.D. 1964: Geology of the Duval Lake area; Manitoba Mines and Natural Resources, Mines Branch, Publication 61-6, 59 p.

Richards, J.P. 2003: Tectono-magmatic precursors for porphyry Cu-(Mo-Au) deposit formation; Economic Geology, v. 98, p. 1515–1533.

Richards, J.P. and Kerrich, R. 2007: Adakite-like rocks; their diverse origins and questionable role in metallogenesis; Economic Geology, v. 102, no. 4, p. 537–576.

Robertson, D.S. 1953: Batty Lake map-area, Manitoba; Geological Survey of Canada, Memoir 271, 55 p.

Ross, P-S. and Bédard, J.H. 2009: Magmatic affinity of modern and ancient subalkaline volcanic rocks determined from trace-element discriminant diagrams; Canadian Journal of Earth Sciences, v. 46, p. 823–839.

Ryan, J.J. 1998: Structural analysis of the Elbow–Cranberry–Iskwatum lakes area: a multiply reactivated deformation corridor in the Trans-Hudson Orogen of Manitoba; Ph.D. thesis, University of New Brunswick, Fredericton.

Ryan, J.J. and Williams, P.F. 1999: Structural evolution of the eastern Amisk Collage, Trans-Hudson Orogen, Manitoba; *in* NATMAP Shield Margin Project, Volume 1, S.B. Lucas (ed.), Canadian Journal of Earth Sciences, v. 36, no. 2, p. 251–273.

Samaniego, P., Hartin, H., Robin, C. and Monzier, M. 2002: Transition from calcalkaline to adakitic magmatism at Cayambe volcano, Ecuador: insight into slab melts and mantle wedge interactions; Geology, v. 30, p. 967–970.

Saunders, A.D. and Tarney, J. 1978: The geochemistry of basalt from a backarc spreading centre in the East Scotia Sea; Geochimica et Cosmochimica Acta, v. 41, p. 555–572.

Schledewitz, D.C.P. 1988: Kississing Lake (63N/3); Manitoba Energy and Mines, Geological Services, Preliminary Map 1988K-1, scale 1:20 000.

Schledewitz, D.C.P. 1993a: Web Lake–Fay Lake (west half); Manitoba Energy and Mines, Preliminary Map 1993K-3, scale 1:20 000.

Schledewitz, D.C.P. 1993b: Web Lake–Fay Lake (east half); Manitoba Energy and Mines, Preliminary Map 1993K-4, scale 1:20 000.

Schledewitz, D.C.P. 1993c: Geology of the Webb Lake–Fay Lake area (NTS 63K/14NE and 63K/15NW); *in* Report of Activities 1993, Manitoba Energy and Mines, Geological Services, p. 29–32.

Schledewitz, D.C.P. 1995: Collins Point and Yakushavich Island, Kississing Lake, NTS 63N/3NW; Manitoba Energy and Mines, Preliminary Map 1995K-3, scale 1:20 000.

Scoates, R.F.J. 1981: Volcanic rocks of the Fox River Belt, northwestern Manitoba; Manitoba Energy and Mines, Mineral Resources Division, Geological Report GR81-1, 109 p.

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

Searle, M.P. and Alsop, G.I. 2007: Eye-to-eye with a mega–sheath fold: a case study from the Wadi Mayh, northern Oman Mountains; Geology, v. 35, p. 1043–1046.

Shapiro, L. and Brannock, W.W. 1962: Rapid analysis of silicate, phosphate and carbonate rocks; United States Geological Survey, Bulletin 1144-A.

Simard, R-L., MacLachlan, K., Gibson, H.L., DeWolfe, Y.M., Devine, C., Kremer, P.D., Lafrance, B., Aimes, D.E., Syme, E.C., Bailes, A.H., Bailey, K., Price, D., Pehrsson, S., Cole, E., Lewis, D. and Galley, A.G. 2010: Geology of the Flin Flon area, Manitoba and Saskatchewan (part of NTS 63K12, 13); Manitoba Innovation, Energy and Mines, Manitoba Geological Survey, Geoscientific Map MAP2010-1 and Saskatchewan Ministry of Energy and Resources, Geoscience Map 2010-2, scale 1:10 000.

Sinclair, H.D. 1997: Flysch to molasse transition in peripheral foreland basins: the role of the passive margin versus slab breakoff; Geology, v. 25, p. 1123–1126.

Sinton, J.M., Ford, L.I., Chappell, B. and McCulloch, M.T. 2003: Magma genesis and mantle heterogeneity in the Manus backarc basin, Papua New Guinea; Journal of Petrology, v. 13, p. 159–195.

Snow, C.A. 2007: Petrotectonic evolution and melt modeling of the Peñon Blanco arc, central Sierra Nevada foothills, California; Geological Society of America Bulletin, v. 119, no. 7–8, p. 1014–1024.

Stauffer, M.R. 1990: The Missi Formation: an Aphebian molasse deposit in the Reindeer Lake Zone of the Trans-Hudson Orogen, Canada; *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. 121–141.

Stauffer, M.R., and Mukherjee, A. 1971: Superimposed deformations in the Missi metasedimentary rocks near Flin Flon, Manitoba; Canadian Journal of Earth Sciences, v. 8, p. 217–242.

Stern, R.A., Machado, N.D., Syme, E.C. Lucas, S.B. and David, J. 1999: Chronology of crustal growth and recycling in the Paleoproterozoic Amisk Collage (Flin Flon Belt), Trans-Hudson Orogen, Canada; Canadian Journal of Earth Sciences, v. 36, p. 1807–1827.

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

Stern, R.A., Syme, E.C. and Lucas, S.B. 1995b: 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.J., Ping-Nan L., Morris, J.D., Jackson, M.C., Fryer, P., Bloomer, S.H. and Ito, E. 1990: Enriched backarc basalts from the northern Mariana Trough: implications for the magmatic evolution of backarc basins; Earth and Planetary Science Letters, v. 100, p. 210–225.

Stockwell, C.H. 1937: Gold deposits of Herb Lake area, northern Manitoba; Geological Survey of Canada, Memoir 208, 46 p.

Stork, A. 2004: The Colo Plutonic Group and the pre–Middle Miocene upper crustal composition of Viti Levu, Fiji (abstract); American Geophysical Union, Eos, Transactions, Western Pacific Geophysical Meeting Supplement.

St-Onge, M.R., Searle, M.P. and Wodicka, N. 2006: Trans-Hudson Orogen of North America and Himalaya-Karakoram-Tibetan Orogen of Asia; structural and thermal characteristics of the lower and upper plates; Tectonics, v. 25, no. 4, 22 p., TC4006, doi:10.1029/2005TC001907.

St-Onge, M.R., Van Gool, J.A.M., Garde, A.A. and Scott, D.J. 2009: Correlation of Archaean and Paleoproterozoic units between northeastern Canada and western Greenland: constraining the pre-collisional upper plate accretionary history of the Trans-Hudson Orogen; Geological Society of London, v. 318, p. 193–235.

Sun, S.S. and McDonough, W.F. 1989: Chemical and isotopic systematics of oceanic basalts: implications for mantle composition and processes; *in* Magmatism in the Oceanic Basins, A.D. Saunders and M.J. Norry (ed.), Geological Society (London), Special Publication 42, p. 313–345.

Syme, E.C. 1988: Athapapuskow Lake project; *in* Report of Field Activities 1988, Manitoba Energy and Mines, Minerals Division, p. 20–34.

Syme, E.C. 1991: Elbow Lake project—Part A: supracrustal rocks and structural setting; *in* Report of Activities 1991, Manitoba Energy and Mines, Minerals Division, p. 14–27.

Syme, E.C. 1998: Ore-associated and barren rhyolites in the central Flin Flon Belt: case study of the Flin Flon mine sequence; Manitoba Energy and Mines, Geological Services, Open File Report OF98-9, 26 p.

Syme, E.C. and Bailes, A.H. 1993: Stratigraphic and tectonic setting of early Proterozoic volcanogenic massive sulfide deposits, Flin Flon, Manitoba; Economic Geology, v. 88, no. 3, p. 566–589.

Syme, E.C. and Bailes, A.H. 1996: Geochemistry of arc and ocean-floor metavolcanic rocks in the Reed Lake area, Flin Flon Belt; *in* Report of Activities 1996, Manitoba Energy and Mines, Geological Services, p. 52–65.

Syme, E.C., Bailes, A.H. and Lucas, S.B. 1995: Geology of the Reed Lake area (parts of 63K/9 and 10); *in* Report of Activities 1995, Manitoba Energy and Mines, Minerals Division, p. 42–60.

Syme, E.C., Lucas, S.B. and Stern, R.A. 1999: 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, no. 11, p. 1767–1788.

Syme, E.C., Lucas, S.B., Zwanzig, H.V., Bailes, A.H., Ashton, K.E. and Haidl, F.M. 1998: Geology, NATMAP Shield Margin Project area, Flin Flon Belt, Manitoba/Saskatchewan: accompanying notes; Geological Survey of Canada Map 1968A, Manitoba Energy and Mines Map A-98-2, Saskatchewan Geological Survey Map 258A, scale 1:100 000.

Tarney, J. and Jones, C.E. 1994: Trace element geochemistry of orogenic igneous rocks and crustal growth models; Journal of the Geological Society of London, v. 151, p. 855–868.

Tinkham, D.K. and Karlapalem, N. 2008: Hydrothermal alteration and metamorphism of the Sherridon structure, Sherridon area, Manitoba (part of NTS 63N3); *in* Report of Activities 2008, Manitoba Science, Technology, Energy and Mines, Manitoba Geological Survey, p. 79–86.

Tran, H.T., Ansdell, K., Bethune, K., Watters, B. and Ashton, K. 2003: Provenance and tectonic setting of Paleoproterozoic metasedimentary rocks along the eastern margin of Hearne craton: constraints from SHRIMP geochronology, Wollaston Group, Saskatchewan, Canada; Precambrian Research, v. 167, p. 171–185.

Trop, J.M. 2007: Late Cretaceous forearc basin development along an accretionary convergent margin: south-central Alaska; Geological Society of America Bulletin, v. 120, no. 1–2, p. 207–224.

Wallace, L.M., McCaffrey, R., Beavan, J. and Ellice S. 2005: Rapid microplate rotations and backarc rifting at the transition between collision and subduction; Geology, v. 33, p. 857–860.

Weber, W. 1990: The Churchill-Superior boundary zone, southeast margin of the Trans-Hudson Orogen: a review; *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. 41–55.

Whalen, J.B. and Hunt, P.A. 1994: Geochronological study of granitoid rocks in the Elbow Lake map area, Manitoba: a portion of the Flin Flon Domain of the Trans Hudson Orogen; *in* Radiogenic Age and Isotopic Studies: Report 8, Geological Survey of Canada, Current Research 1994-F, p. 87–96.

Whalen, J.B., Jenner, G.A., Longstaffe, F.J., Gariépy, C. and Fryer, B.J. 1997: Implications of granitoid geochemical and isotopic (Nd, O, Pb) data from the Cambrian–Ordovician Notre Dame arc for the evolution of the Central Mobile Belt, Newfoundland Appalachians; *in* The Nature of Magmatism in the Appalachian Orogen, A.K. Sinha, J.B. Whalen and P.J. Hogan (ed.), Geological Society of America, Memoir 191, p. 367–395.

Whalen, J.B., McNicoll, V.J. Galley, A.G. and Longstaffe, F.J. 2004: Tectonic and metallogenic importance of an Archean composite high- and low-Al tonalite suite, western Superior Province, Canada; Precambrian Research, v. 132, p. 275–301.

Whalen, J.B., Percival, J.A., McNicoll, V.J. and Longstaffe, F.J. 2002: A mainly crustal origin for tonalitic granitoid rocks, Superior Province, Canada: implications for Late Archean tectonomagmatic processes; Journal of Petrology, v. 43, no. 8, p. 1551–1570.

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; *in* NATMAP Shield Margin Project, Volume 1, S.S. Lucas (ed.), Canadian Journal of Earth Sciences, v. 36, no. 2, p. 227–250.

Whalen, J.B., Zwanzig, H.V., Percival, J.A. and Rayner, N. 2008: Geochemistry of an alkaline, ca. 1885 Ma K-feldspar–porphyritic, monzonitic to syenogranitic suite, northeastern Kisseynew Domain, Manitoba (parts of NTS 63O); *in* Report of Activities 2008, Manitoba Science, Technology, Energy and Mines, Manitoba Geological Survey, p. 66–78.

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, p. 325–343.

White, D.J. 2005: High-temperature low-pressure metamorphism in the Kisseynew Domain, Trans-Hudson Orogen: crustal anatexis due to crustal thickening?; Canadian Journal of Earth Sciences, v. 42, p. 741–759.

White, D.J., Lucas, S.B., Bleeker, W., Hajnal, Z. Lewry, J.F. and Zwanzig, H.V. 2002: Suture-zone geometry along an irregular Paleoproterozoic margin: the Superior Boundary Zone, Manitoba, Canada; Geology, v. 30, no. 8, p. 735–738.

White, D.J., Thomas, M.D., Jones, A.G., Hope, J., Németh, B. and Hajnal, Z. 2005: Geophysical transect across a Paleoproterozoic continent-continent collision zone: the Trans-Hudson Orogen; Canadian Journal of Earth Sciences, v. 42, p. 385–402.

White, D.J., Zwanzig, H.V. and Hajnal, Z. 2000: Crustal suture preserved in the Paleoproterozoic Trans-Hudson Orogen, Canada; Geology, v. 28, no. 6, p. 527–530.

Whitmeyer, S.J and Karlstrom, K.E. 2007: Tectonic model for the Proterozoic growth of North America; Geosphere, v. 3, no. 4, p. 220–259.

Williams, P.F. and Jiang, D. 2005: An investigation of lower crustal deformation: evidence for channel flow and its implications for tectonics and structural studies; Journal of Structural Geology, v. 27, p. 1486–1504

Woodhead, J.D. and Johnson, R.W. 1993: Isotopic and trace element profiles across the New Britain island arc, Papua New Guinea; Contributions to Mineralogy and Petrology, v. 113, p. 479–491.

Woodhead, J.D., Eggins, S. and Gamble, J. 1993: High field strength and transition element systematics in island arc and backarc basin basalts: evidence for multi-phase melt extraction and depleted mantle wedge; Earth and Planetary Science Letters, v. 144, p. 491–504.

Yuan, C., Zhou, M-F., Sun, M., Zhao, Y., Wilde, S., Long, X. and Yan, D. 2010: Triassic granitoids in the eastern Songpan Ganzi fold belt, SW China: magmatic response to geodynamics of the deep lithosphere; Earth and Planetary Science Letters, v. 290, p. 481–492.

Xu, J., Shinjo, R., Defant, M. and Rapp, R. 2002: Origin of Mesozoic adakitic intrusive rocks in the Ningzhen area of east China: partial melting of delaminated lower continental crust?; Geology, v. 30, no. 12, p. 1111–1114.

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. 1992a: Meat Lake; Manitoba Energy and Mines, Preliminary Map 1992K-1, scale 1:20 000.

Zwanzig, H.V. 1992b: Evans Lake; Manitoba Energy and Mines, Preliminary Map 1992K-2, scale 1:20 000.

Zwanzig, H.V. 1993: Matching LITHOPROBE sections and geological structure sections: a simple tool for analyzing upper crustal seismic data in Kisseynew and Flin Flon belts; *in* LITHOPROBE, Trans-Hudson Orogen Transect: Report of Third Transect Meeting, Z. Hajnal and J.F. Lewry (ed.), LITHOPROBE Secretariat, University of British Columbia, LITHOPROBE Report 34, p. 113–120.

Zwanzig, H.V. 1994: Geologic setting of the Nokomis Lake gold deposit (NTS 63N/3); *in* Report of Activities 1994, Manitoba Energy and Mines, Geological Services, p. 35–38.

Zwanzig, H.V. 1996a: Dow Lake; Manitoba Energy and Mines, Preliminary Map 1996K-1, scale 1:20 000.

Zwanzig, H.V. 1996b: Geology of the Dow Lake–Martell Lake area (parts of NTS 63K/15 and 63N/2); *in* Report of Activities 1996, Manitoba Energy and Mines, Geological Services, p. 21–28.

Zwanzig, H.V. 1997: Kisseynew metasedimentary gneiss belt, Trans-Hudson Orogen (Canada), backarc origin and collisional inversion: comment and reply; Geology, v. 25, no. 1, p. 90–92.

Zwanzig, H.V. 1999: Structure and stratigraphy of the south flank of the Kisseynew Domain in the Trans-Hudson Orogen, Manitoba: implications for 1.845–1.77 Ga collision tectonics; *in* NATMAP Shield Margin Project, Volume 2, Canadian Journal of Earth Sciences, v. 36, no. 11, p. 1859–1880.

Zwanzig, H.V. 2000a: Mystery Lake high-strain zones: sinistral and dextral, east-side-up displacements, Thompson Nickel Belt; *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 18–23.

Zwanzig, H.V. 2000b: Geochemistry and tectonic framework of the Kisseynew Domain–Lynn Lake belt boundary (part of NTS 63P/13); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 91–96.

Zwanzig, H.V. 2000c: Geochemical and tectonic evolution of the Flin Flon Belt–Kisseynew Domain Boundary Zone (abstract); Manitoba Industry, Trade and Mines, Manitoba Mining and Minerals Convention 2000, Winnipeg, Manitoba, November 16–18, 2000, Program, p. 39.

Zwanzig, H.V. 2005: Geochemistry, Sm-Nd isotope data and age constraints of the Bah Lake assemblage, Thompson Nickel Belt and Kisseynew Domain margin: relation to Thompson-type ultramafic bodies and a tectonic model (NTS 63J, O and P); *in* Report of Activities 2005, Manitoba Industry, Economic Development and Mines, Manitoba Geological Survey, p. 40–53.

Zwanzig, H.V. 2008: Correlation of lithological assemblages flanking the Kisseynew Domain, Manitoba (parts of NTS 63N, 63O, 64B, 64C): proposal for tectonic/metallogenic subdomains; *in* Report of Activities 2008, Manitoba Science, Technology, Energy and Mines, Manitoba Geological Survey, p. 38–52.

Zwanzig, H.V. and Parent, M. 1993: Jungle Lake; Manitoba Energy and Mines, Preliminary Map 1993K-1, scale 1:20 000.

Zwanzig, H.V. and Schledewitz, D.C.P. 1992: Geology of the Kississing–Batty lakes area: interim report; Manitoba Energy and Mines, Minerals Division, Open File Report OF92-2, 87 p.

Zwanzig, H.V. and Shwetz, J.V. 1994a: Nokomis Lake; Manitoba Energy and Mines, Preliminary Map 1994K-2, scale 1:10 000.

Zwanzig, H.V. and Shwetz, J.V. 1994b: Synkinematic migmatites in the vicinity of Nokomis Lake (NTS 63N/3); *in* Report of Activities 1994, Manitoba Energy and Mines, Geological Services, p. 39–41.

Zwanzig, H.V., Schledewitz, D.C.P and Parent, M. 1993: Puffy Lake; Manitoba Energy and Mines, Preliminary Map 1993K-2, scale 1:10 000.

Zwanzig, H.V., Syme, E.C. and Gilbert, H.P. 1999: Updated trace element geochemistry of ca. 1.9 Ga metavolcanic rocks in the Paleoproterozoic Lynn Lake belt; Manitoba Industry, Trade and Mines, Geological Services, Open File Report OF99-13, 46 p.

Zwanzig, H.V., Bailes, A.H. and Böhm, C.O. 2001: Josland Lake sills: U-Pb age and tectonostratigraphic implications (parts of NTS 63K and 63N); *in* Report of Activities 2001, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 28–32.

Zwanzig, H.V., Böhm, C.O., Protrel, A. and Machado, N. 2003: Field relations, U-Pb zircon ages and Nd model ages of granitoid intrusions along the Thompson Nickel Belt–Kisseynew Domain boundary, Setting Lake area (NTS 63O2, 63J15); *in* Report of Activities 2003, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 118–129.

Zwanzig, H.V., Murphy, L. Percival, J.A., Whalen, J.B. and Rayner, N. 2006: Thompson Nickel Belt–type units in the northeastern Kisseynew Domain, Manitoba (parts of NTS 63O); *in* Report of Activities 2006, Manitoba Science, Technology, Energy and Mines, Manitoba Geological Survey, p. 85–103.

Zwanzig, H.V., Macek, J. J. and McGregor C. R. 2007: Lithostratigraphy and geochemistry of the high-grade metasedimentary rocks in the Thompson Nickel Belt and adjacent Kisseynew Domain, Manitoba: implications for nickel exploration; Economic Geology, v. 102, p. 1197–1216.

Appendix 1

Map units and diagram symbols and major-element diagrams

Table 12: Map units and diagram symbols

Early volcanic to early successor-arc assemblages

MORB-like

Unit	
1a	Fussey Lake basalt

Weak arc signature (BABB/Rift)

Unit	
2a	Ponton Lake basalt
	high–Mg variety
2b	Moody Lake basalt
2c	Preaston andesite
2d	Dickstone rhyolite
2e	Storozuk basalt
2f	SE Dow porphyritic basalt
7	Josland Lake sills (all)
7a	Gabbro–melagabbro
	chilled margin
7b	Ferrogabbro
7c	Quartz diorite/trondhjemite

Arc tholeiite (Amisk collage)

Unit	
3a	Puffy Mine andesite
3b	Koscielny andesite
3c	Martell Lake basalt
3d	Dow Lake andesite

Sherridon–Meat Lake arc assemblage

Unit	
4	Amphibolite, undivided
4c	Meat Lake amphibolite
4d	Walton Lake amphibolite
	(Folster Bay)
4e	Gabbro

Sherridon felsic–intermediate gneiss

Unit	
5	(Sherridon structure)
5a	(Walton Lake nappe)
5b	Calcsilicate gneiss
5c	Moderate alteration
5d	Strong alteration

Early successor-arc intrusions

Unit	
8a	Martell Lake gabbro
	Microdiorite
9a	Ragged Lake granite
9b	Jay Lake granite
9c	Big Island granodiorite
9	Fire Lake granodiorite
9d	Archie Lake pluton
9e	Star Lake tonalite
9f	Batty Lake gneiss

Successor-arc–basinal rocks

Unit	
10a	Bob gneiss (probable Burntwood Group equivalent)
11a-c	Burntwood Group
	(greywacke and gneiss)

Late successor-arc units

Unit	
14a	Kississing River basalt
14b	Evans Lake basalt
14c	Jungle Lake basaltic andesite
14d	Puffy Lake dacite
14e	Cleunion rhyolite
14f	Lobstick rhyolite
15a	Nelson Bay pluton
15b	Norris Lake pluton
15c	Baron Lake pluton
15d	Ham Lake pluton
15e	Thunderhill Lake porphyry
15g	Reed Lake diorite
15h	SE Big Island gabbro

Synkinematic intrusions

Transitional and calcalkaline

Unit	
16a	Touchbourne suite (all)
	(mafic–ultramafic
	(intermediate–felsic)
16b	Duval Lake sills
16c	Dow Lake quartz ferrodiorite

Tholeiitic

Unit	
17a	Weldon Bay gabbro
17a/b	Cree Lake gabbro/pyroxenite
17c	Mafic calcsilicate gneiss

Weakly peraluminous

Unit	
18	Burntwood Lake tonalite
18a	Highrock Lake tonalite
18b	Flatrock Lake granite
19a	Late-kinematic pegmatite
19b	Leucogranite
	Late-kinematic granite
19c	Leucogranodiorite
	LS3 Leucosome
	Leucogranite–tonalite
	LS2 Leucosome

Symbols in this table are used in Figures 19 to 42 except where units are combined or subdivided and symbols shown locally.
Unit numbers, except those in grey, are used on Maps GR2010-1-1 and -2 (back pocket/DVD).

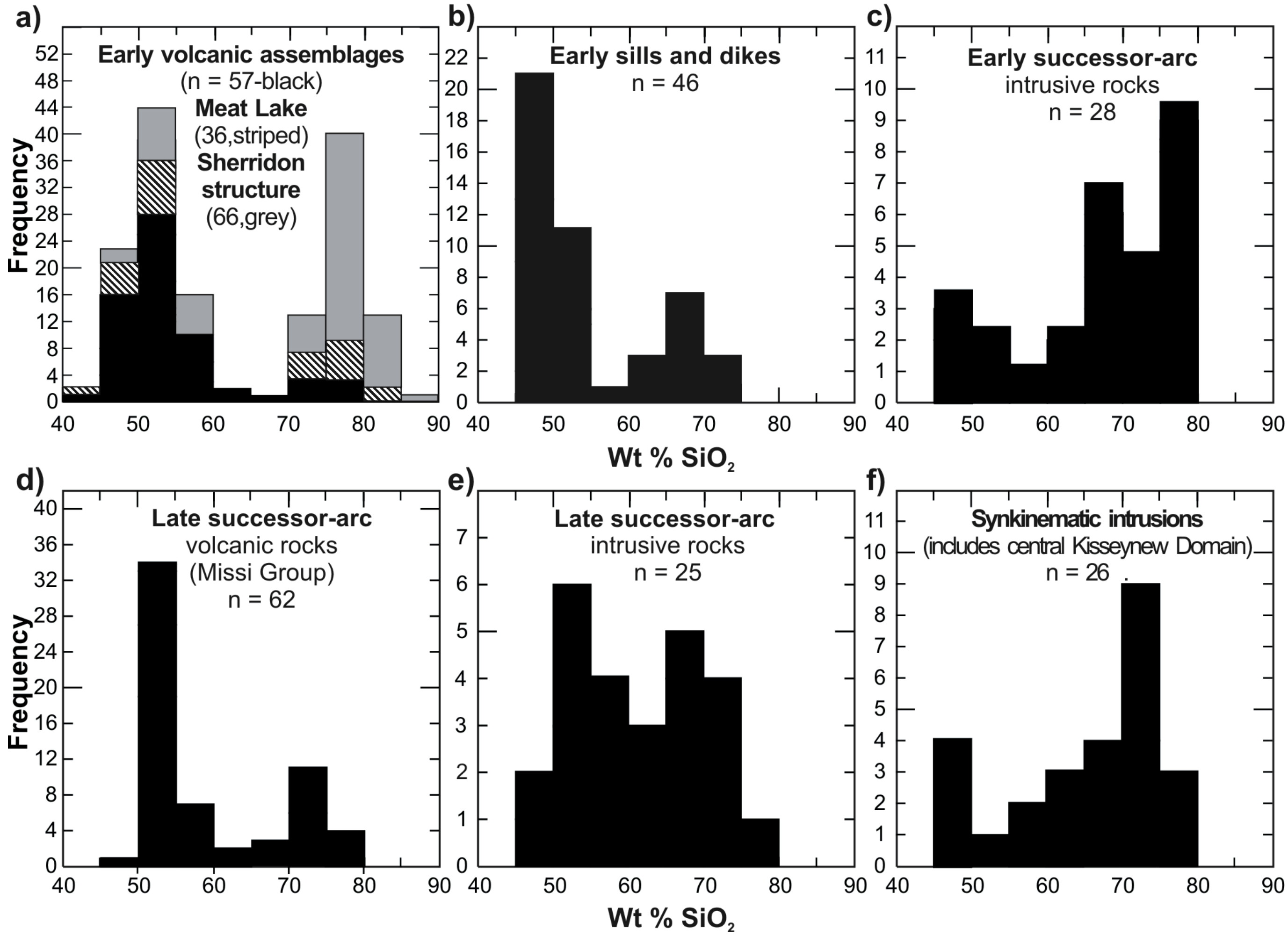


Figure 19: *Silica histograms showing a general bimodal distribution that is most pronounced for volcanic rocks and least pronounced for successor-arc intrusions:* ***a)*** *early volcanic assemblages are shown in black, and the Sherridon–Meat Lake assemblage in stripes (Meat Lake area) and grey (Sherridon structure); the low frequency of felsic data from Meat Lake is a sampling artifact;* ***b–f)*** *data are considered to show only a general trend to more intermediate and felsic compositions. Data from this study, augmented by data of Hollings and Ansdell (2002), are in black and stripes; 2007 data of Halo Resources Ltd. from the Sherridon structure are in grey.*

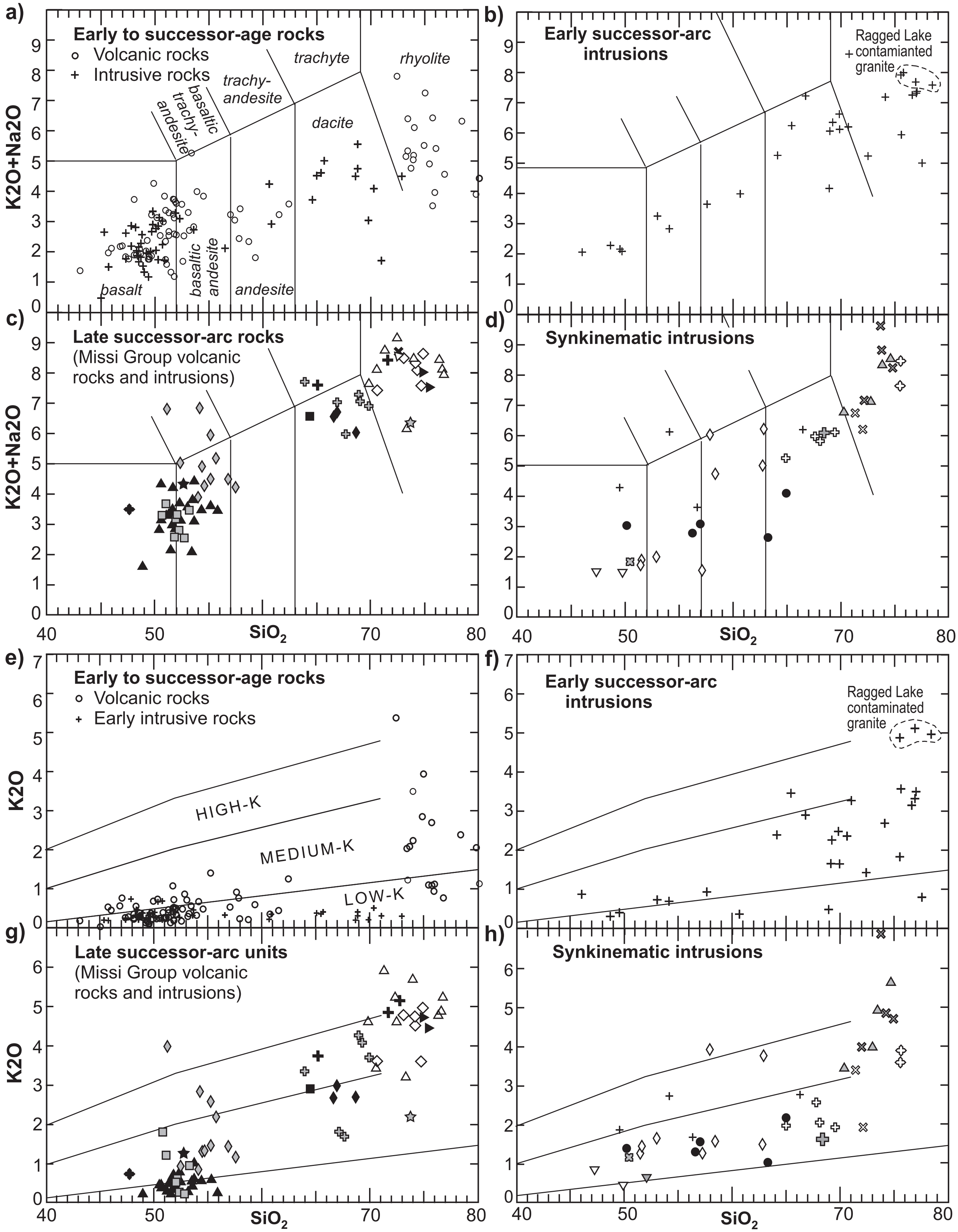


Figure 20: *Alkalis vs. SiO_2:* ***a–d)*** *K_2O+NaO_2 vs. SiO_2 (Le Bas et al., 1986), showing increasing total alkalis in arc-related rocks with time;* ***e–h)*** *K_2O vs. SiO_2 (Peccerillo and Taylor, 1976), showing a progressive change from predominantly low-K mafic and intermediate rocks with variable K in rhyolite (e) to medium-K and high-K arc-related rocks (f, g), followed by medium-K and minor high-K intrusions. Much of the scatter is due to alteration. Symbols legend for parts c, d, g and h in Appendix 1, Table 12.*

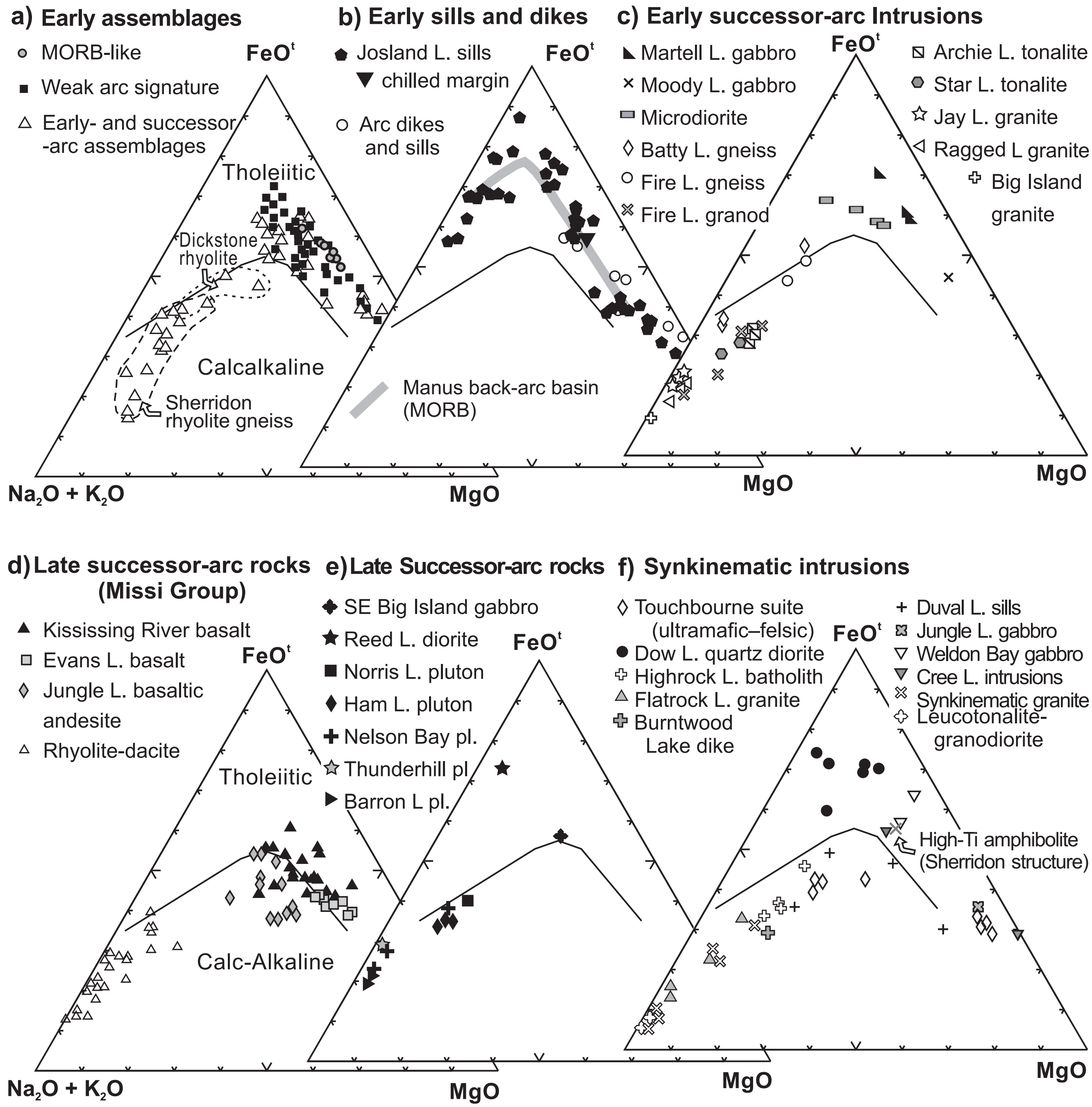


Figure 21: a–f) *AFM diagrams (Irvine and Baragar, 1971), showing progression from more tholeiitic units to more calcalkaline units with time. Field of Manus back-arc basin (MORB) from Sinton et al. (2003). Also shown is the high-Ti amphibolite from the Sherridon structure (average of five samples from Halo Resources Ltd., 2007). Symbols legend in Appendix 1, Table 12. Abbreviations: L., Lake; pl., pluton.*

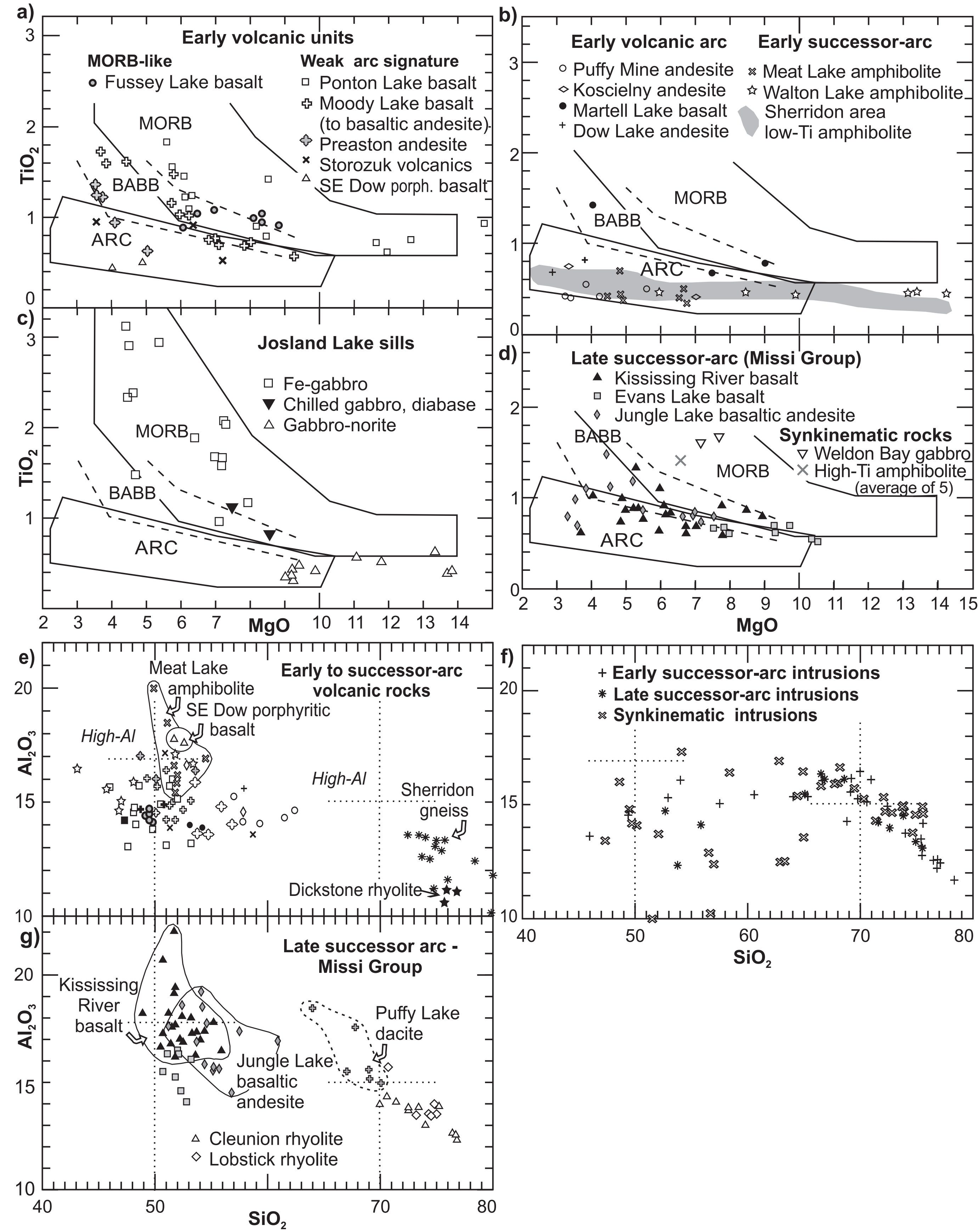


Figure 22: *Variation diagrams of selected major elements to distinguish MORB-like units from volcanic-arc units and late intrusions:* ***a–d)*** *TiO_2 vs. MgO;* ***e–g)*** *Al_2O_3 vs. SiO_2, with units that include high-Al volcanic rocks outlined as fields. Fields of MORB and arc after Stern et al. (1995a) and BABB from data of Sinton et al. (2003); High-Al division modified from Whalen et al. (1999). Sherridon area amphibolite from 2007 surface data of Halo Resources Ltd. Symbols legend in Appendix 1, Table 12.*

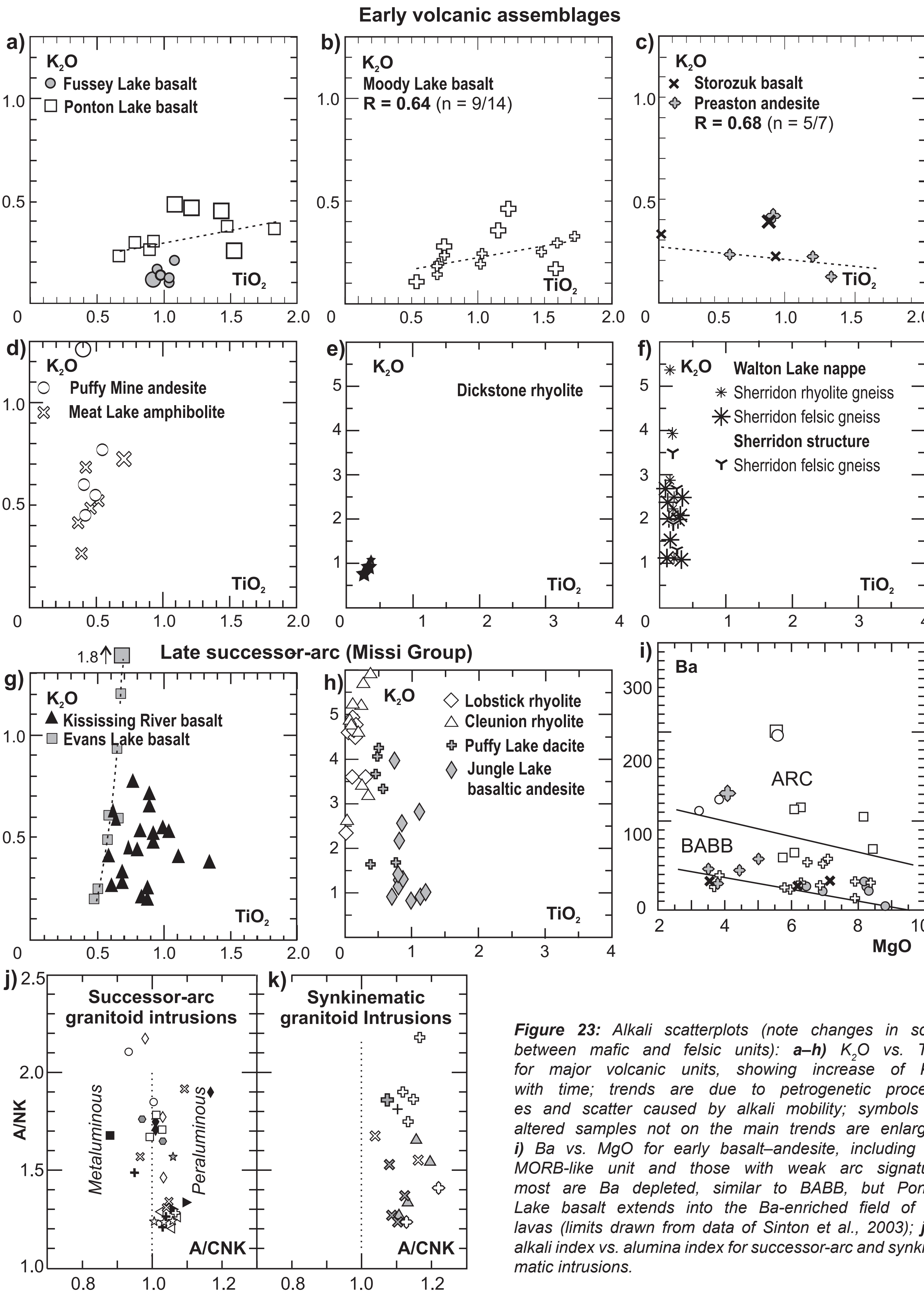


Figure 23: *Alkali scatterplots (note changes in scale between mafic and felsic units):* ***a–h)*** *K_2O vs. TiO_2 for major volcanic units, showing increase of K_2O with time; trends are due to petrogenetic processes and scatter caused by alkali mobility; symbols for altered samples not on the main trends are enlarged;* ***i)*** *Ba vs. MgO for early basalt–andesite, including the MORB-like unit and those with weak arc signature; most are Ba depleted, similar to BABB, but Ponton Lake basalt extends into the Ba-enriched field of arc lavas (limits drawn from data of Sinton et al., 2003);* ***j–k)*** *alkali index vs. alumina index for successor-arc and synkinematic intrusions.*

Appendix 2

Trace- and minor-element diagrams

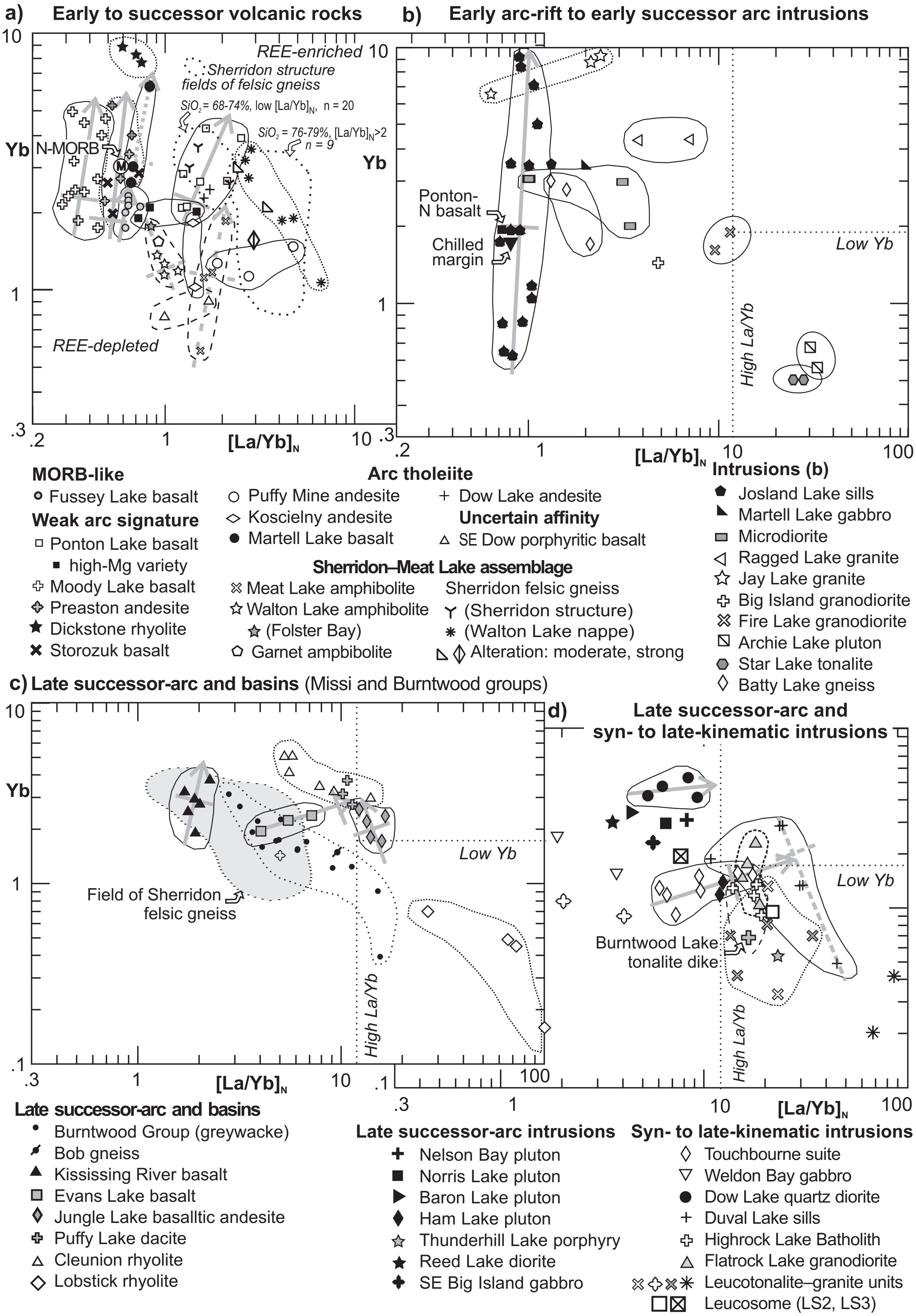


Figure 24: Scatterplots of Yb (ppm) vs. $[La/Yb]_N$ (chondrite-normalized), with distinguishing fields of units based on the contents and fractionation of REE. Grey lines are regressed trends due to fractional crystallization (up arrows), fractional melting (away from up arrows) or mixing with a high–La/Yb melt component (right arrows); large symbols indicate altered samples not used in regression; crossbars indicate Yb7 contents (at 7% MgO, fitted from Yb vs. MgO scatterplots, not shown); N-MORB (Sun and McDonough, 1989) shown for reference.

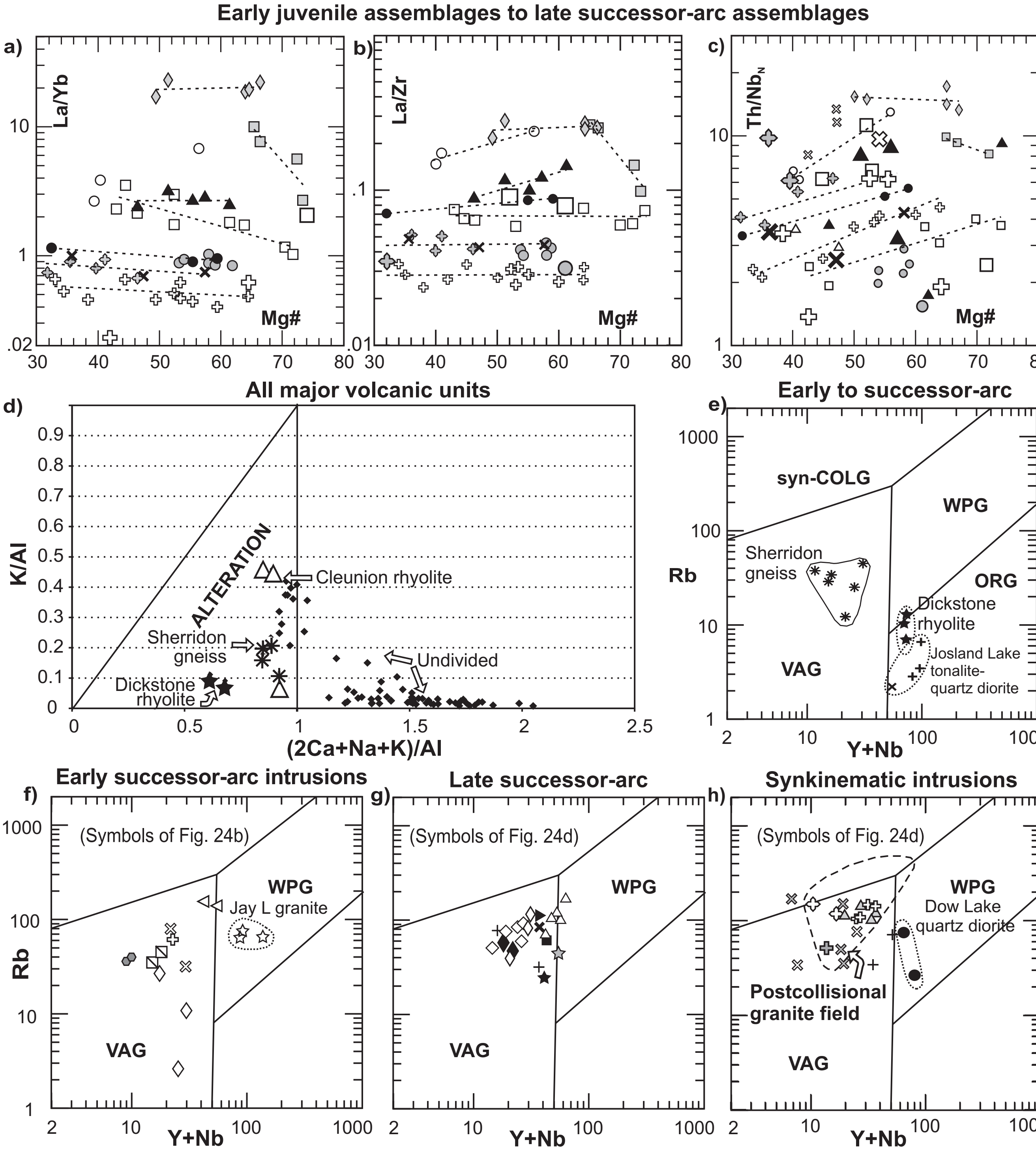


Figure 25: *Ratio scatterplots of selected trace and alkali elements:* ***a–c)*** *La/Yb, La/Zr and Th/Nb$_N$ (N-MORB–normalized) vs. Mg# (molar Mg / [Mg + 0.85Fe^t]), showing fractionation trends, with outlying values as enlarged symbols;* ***d)*** *Pearce element plot of molar K/Al, showing alkali-depleted samples (large symbols);* ***e–h)*** *tectonic granite diagram (Pearce et al., 1984) of felsic to intermediate rocks; COLG, WPG, VAG and ORG are collisional, within-plate, volcanic-arc and ocean-ridge granites. Symbols as in Appendix 1, Figure 24.*

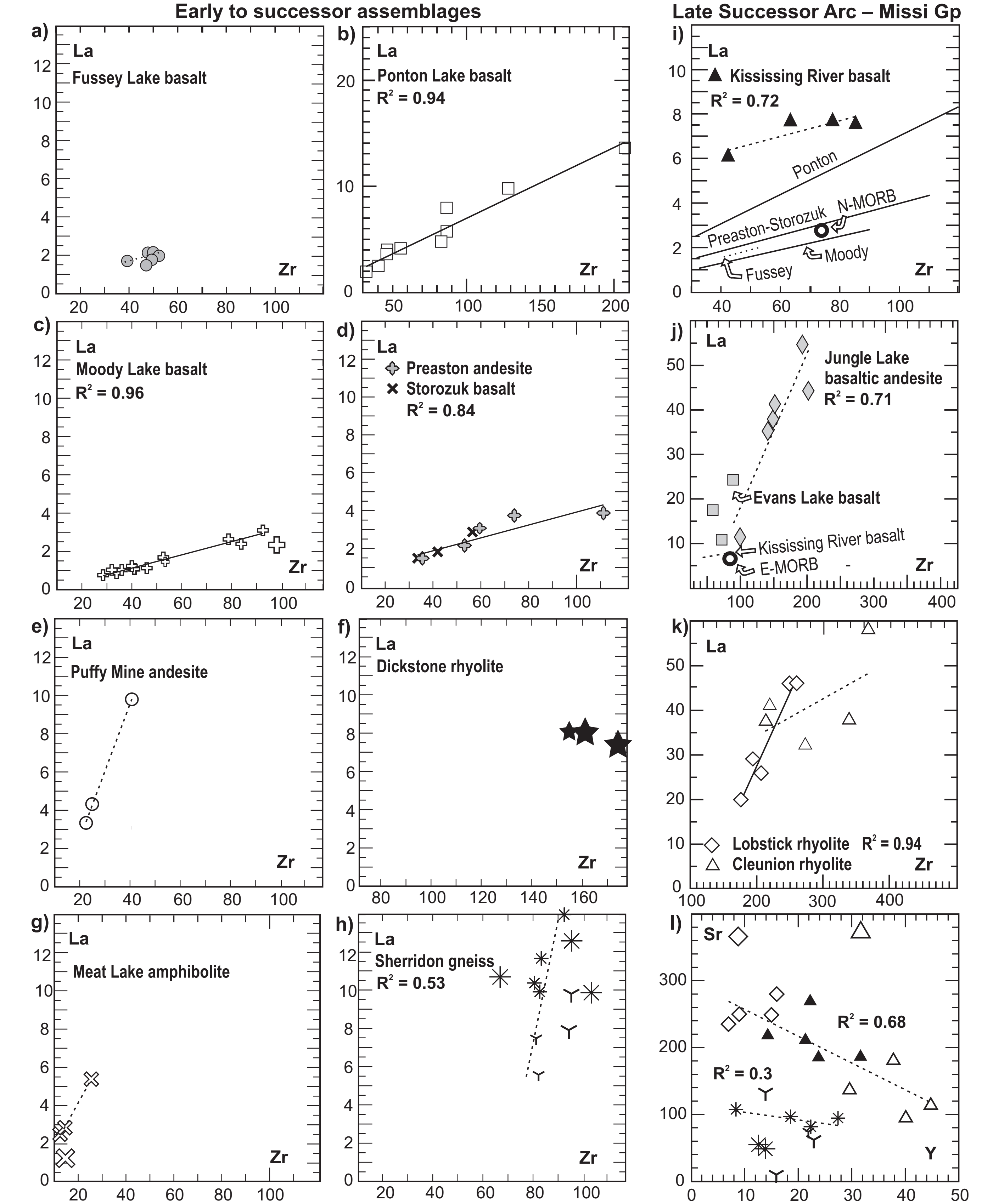


***Figure 26: a–k)** Scatterplots of La vs. Zr showing little scatter due to altered samples (large symbols) except for rhyolite units; linear trends are due to petrogenetic processes, mainly fractionation; regression lines are solid for high correlation (R^2) and dotted for small, poorly defined populations with lower R^2; successive plots are distinctive and indicate increased La (at comparable Zr) with time; N-MORB and E-MORB (Sun and McDonough, 1989) are shown for reference. **l)** Sr vs. Y scatterplot for related rhyolite and basalt, units showing linear pattern with mainly moderate scatter. Symbols as in Appendix 1, Table 12.*

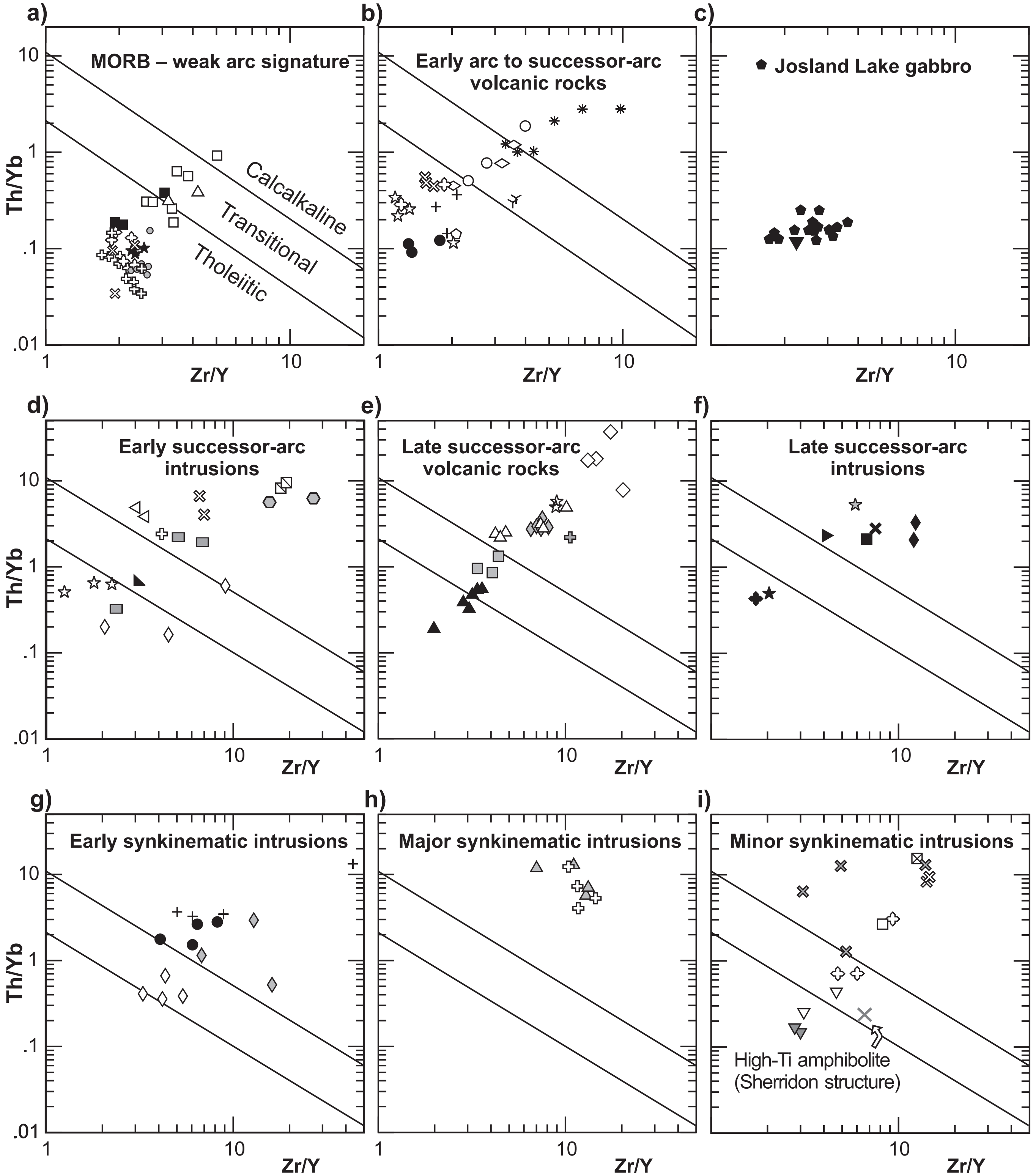


Figure 27: *Scatterplots of Th/Yb vs. Zr/Y discriminate meta-igneous rocks of tholeiitic and calcalkaline affinity (Ross and Bédard, 2009). There is an evolution from dominantly tholeiitic (mainly mafic) to dominantly calcalkaline (mainly felsic) compositions. All arc magmas include both types. Transitional units may have formed by magma mixing or fractionation, and crustal contamination (see text). Synkinematic intrusions are calcalkaline, but xenocrystic garnet in minor felsic melts shifts compositions into the transitional field (compare part i of this figure to Appendix 1, Figure 21f). Synkinematic tholeiitic dikes are present but not abundant. Symbols as in Appendix 1, Table 12.*

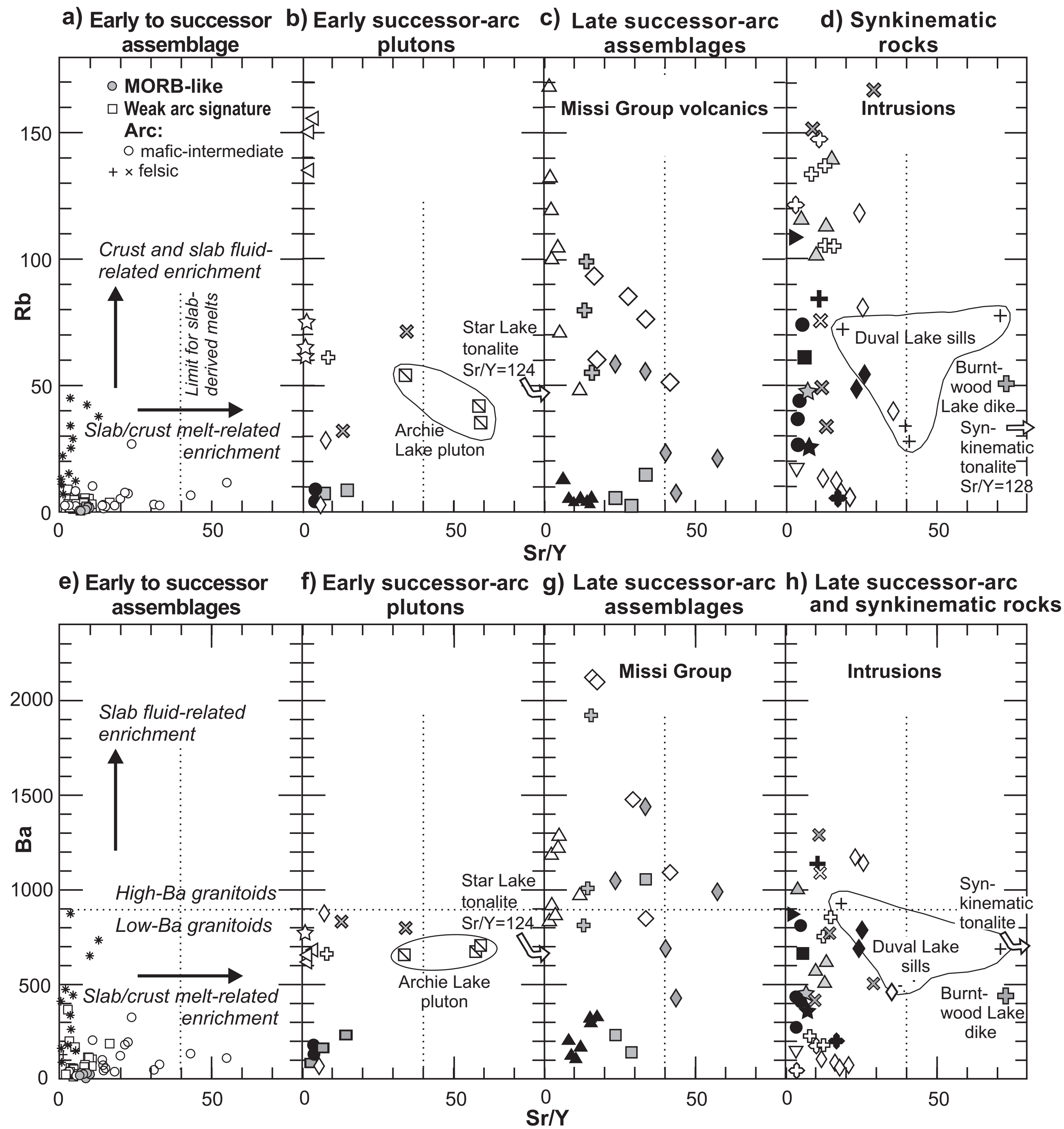


Figure 28: Scatterplots of alkali elements (Rb and Ba) vs. Sr/Y (after Whalen et al., 1999), showing increasing Rb content with time and highest Ba content in the Missi Group. Sr/Y ratios are generally low except for several successor-arc intrusive suites (outlined), suggesting only a local component of slab or lower crustal melt. Limit for slab melts from Drummond and Defant (1990); Ba division from Tarney and Jones (1994). Symbols as in Appendix 1, Table 12.

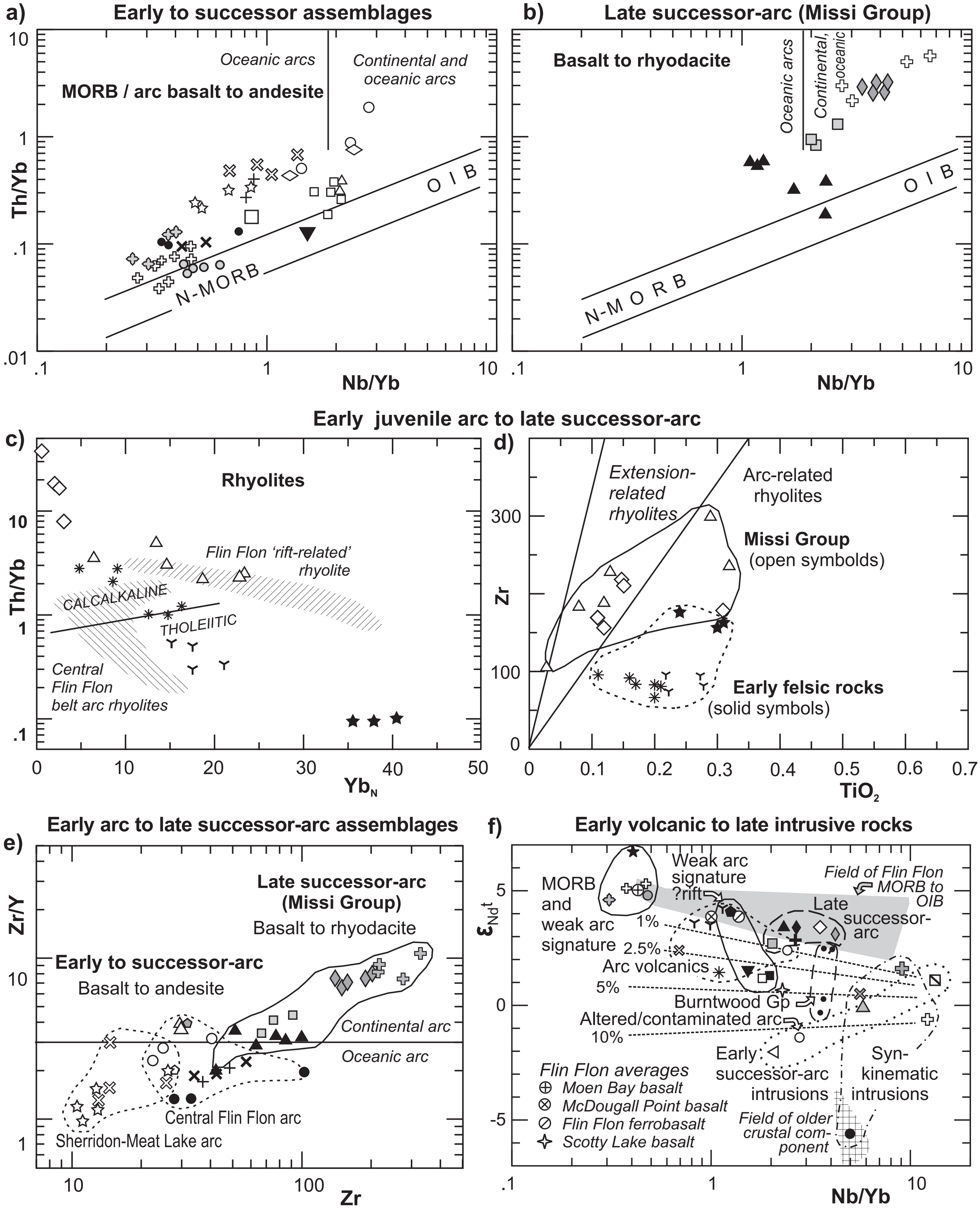


Figure 29: *Scatterplots of incompatible trace-element ratios and Nd-isotope data (symbols as in Appendix 1, Table 12):* ***a–b)*** *Th/Yb vs. Nb/Yb (Pearce and Peate, 1995) of mafic–intermediate units;* ***c)*** *Th/Yb vs. Yb_N (chondrite normalized) of rhyolite units, showing the fields of rhyolite types from the main Flin Flon Domain (Syme, 1998);* ***d)*** *Zr vs. TiO_2, showing the field of extension-related rhyolite (Syme and Bailes, 1993) containing most of the Missi Group felsic volcanic–derived gneiss;* ***e)*** *Zr/Y vs. Zr, showing the distinction between early oceanic-arc units and Missi Group rhyolite;* ***f)*** *ε_{Nd}^{T} (at time of crystallization) vs. Nb/Yb, showing a modelled amount of contamination by Archean crust and the Flin Flon mantle array. Symbols as in Appendix 1, Table 12.*

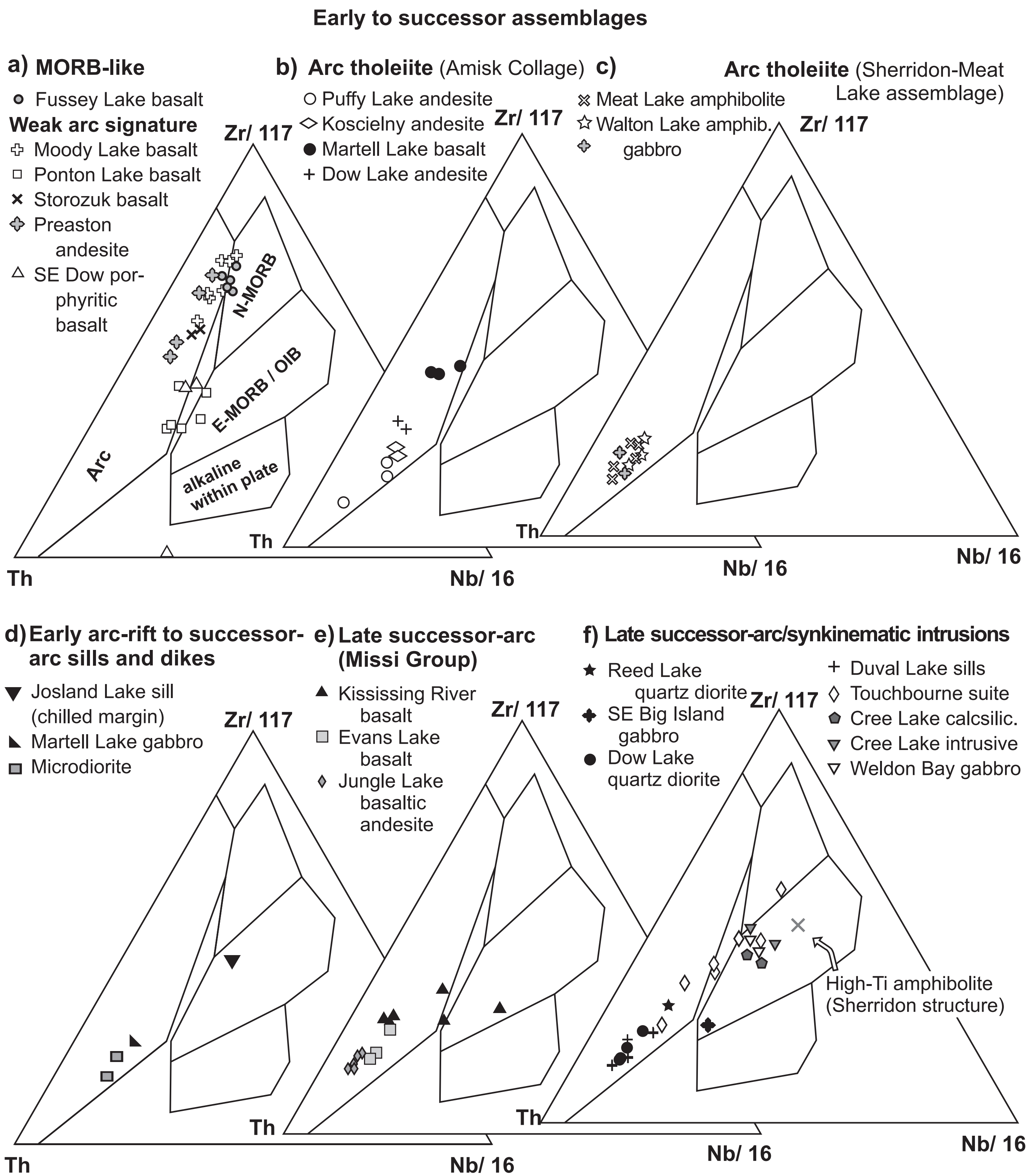


Figure 30: *Tectonic discriminant diagram (Th/Zr/Nb) for basalt and andesite (Pearce et al., 1984) grouped by age and tectonic affinity. The earliest units show trends from arc to N-MORB and E-MORB* ***(a)*** *or dominantly arc* ***(b–c)****; later units show trends from arc to E-MORB* ***(d–f)****, suggesting a change in the mantle source. A spread to higher Th/Zr ratios due to sea-floor and metamorphic alteration has been reduced by omitting some altered samples from (a).*

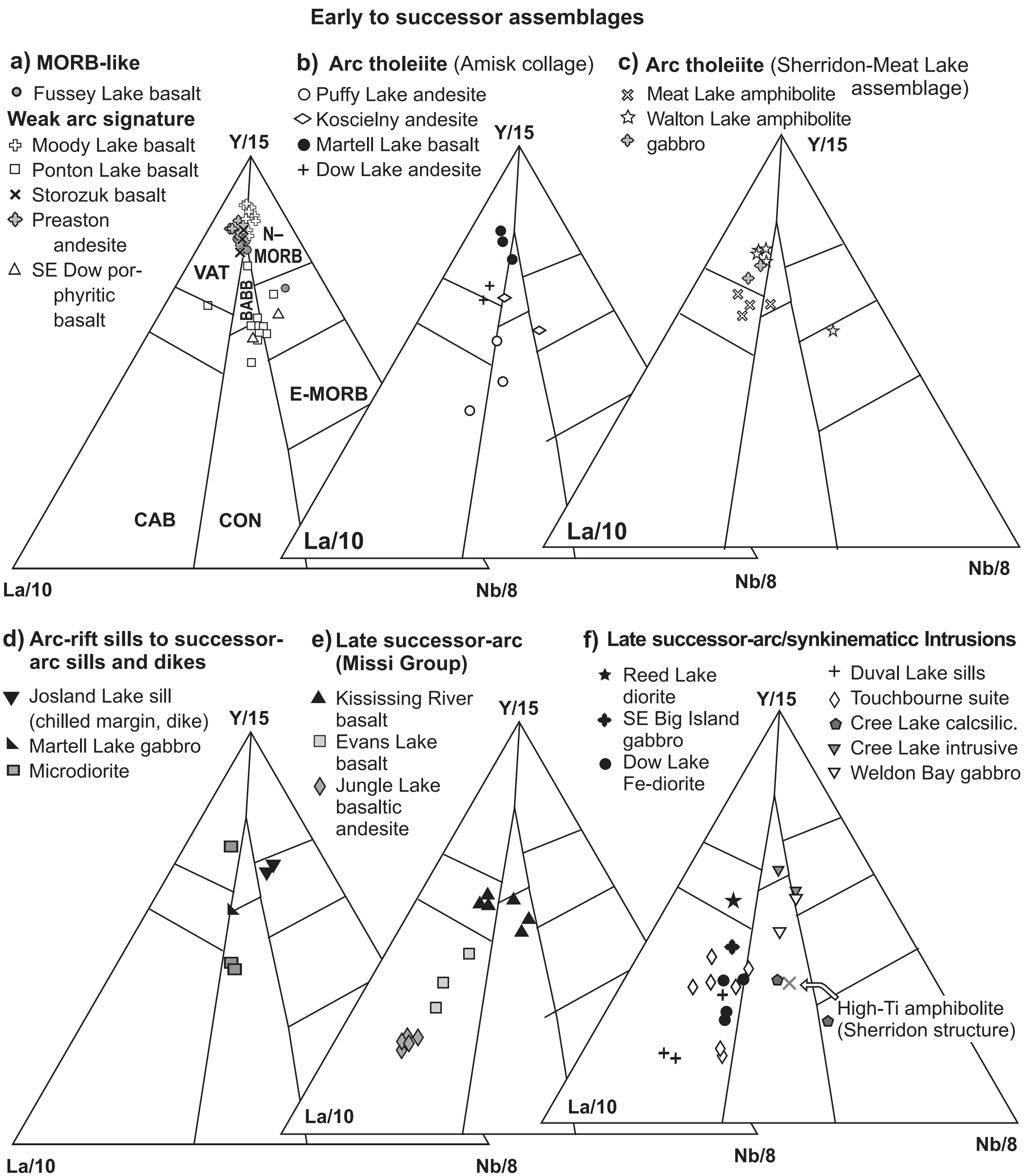


Figure 31: *Tectonic discriminant diagram (La/Y/Nb) for basalt and andesite (Cabanis and Lecolle, 1989) grouped by age and tectonic affinity. The earliest units show* ***a)*** *trends from the boundary of N-MORB and E-MORB to back-arc basin basalt (BABB) and volcanic-arc tholeiite (VAT),* ***b–c)*** *predominantly VAT, with scatter to higher Y and Nb probably due to alteration; or* ***e–f)*** *in later units, trends from calcalkaline arc basalt (CAB) toward continental basalt (CON), suggesting crustal evolution.*

Appendix 3

Chondrite-normalized rare earth element diagrams and N-MORB-normalized multi-element diagrams

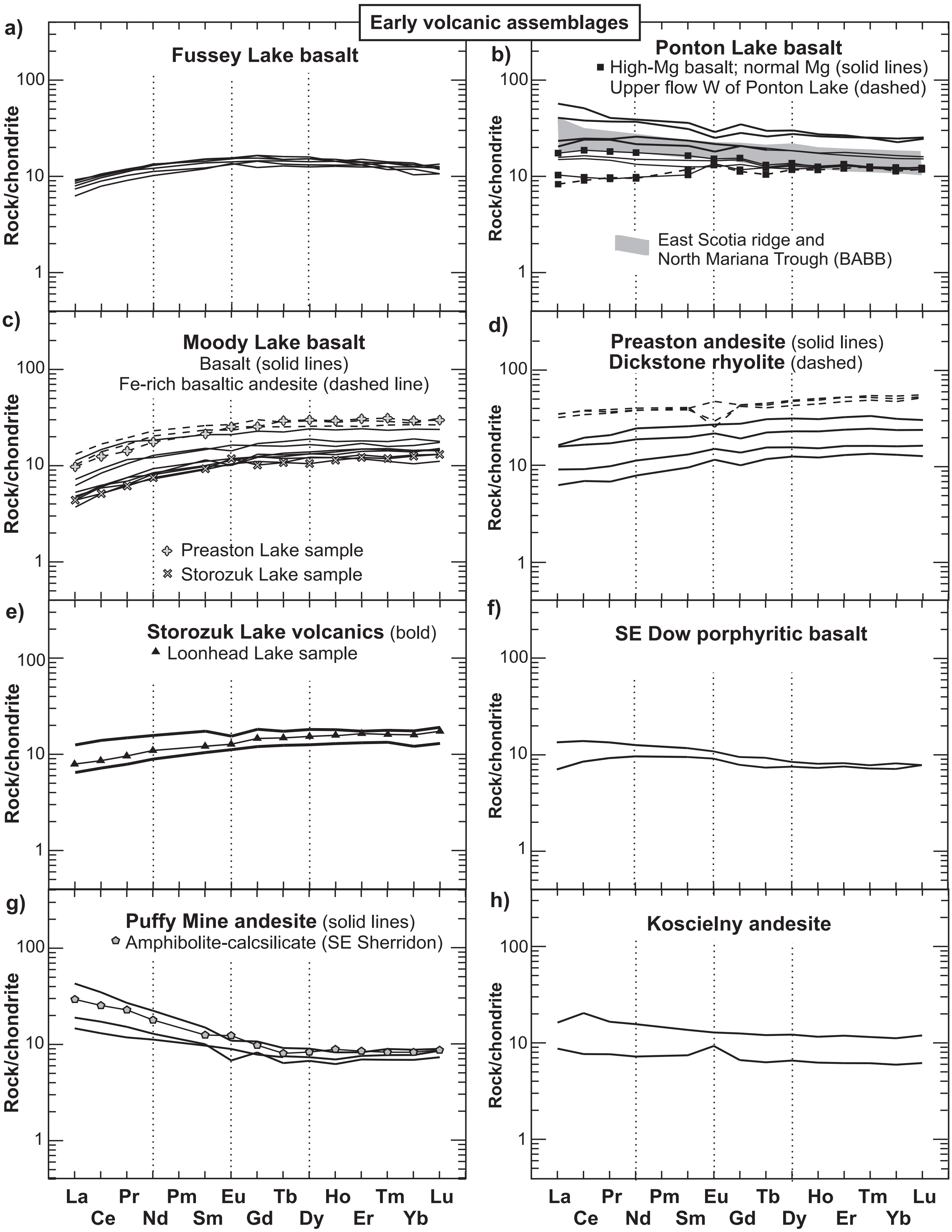


Figure 32: *Chondrite-normalized plots of REE from early assemblages. Increasing REE within each unit in b–h is probably due to fractionation. Fanning of LREE in b and g suggests crustal assimilation–fractional crystallization (AFC) and/or changes in fractional melting or source composition. MORB to weak arc signature has depleted or flat LREE (a–f); full arc signature has fractionated LREE (g, h). Comparison field (grey) of BABB from data of Fretzdorff et al. (2002) and Stern et al. (1990).*

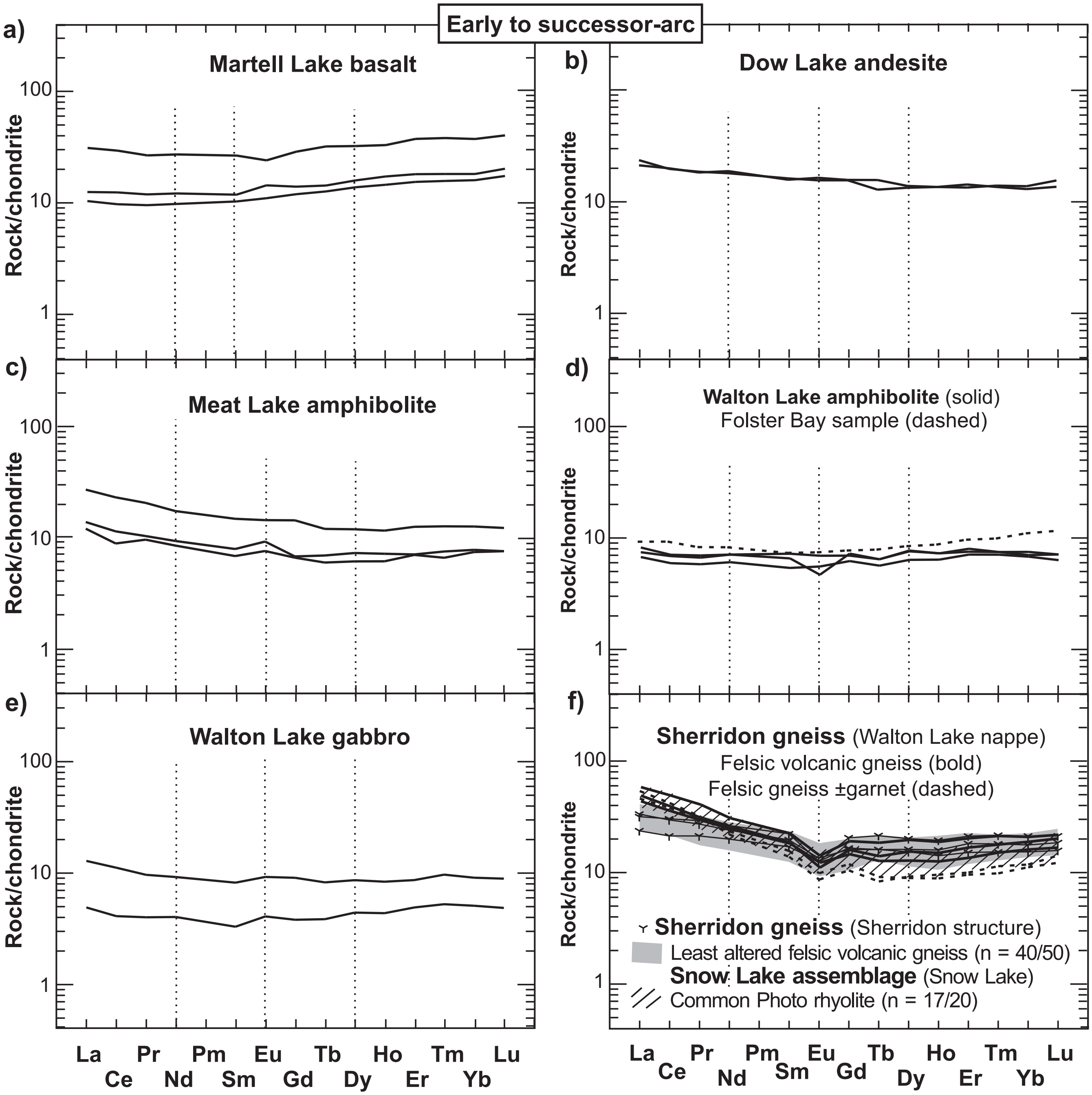


Figure 33: *Chondrite-normalized REE plots of early volcanic to successor-arc assemblages, showing flat MREE to HREE and fractionated LREE. Comparison field of the most common variety of rhyolite gneiss. Forty of the apparently least altered surface samples from the Sherridon structure are from data of Halo Resources Ltd. (L. Bloom, pers. comm., 2008), and 17 of the 20 samples of Photo rhyolite from the Snow Lake area are from Bailes and Galley (2001).*

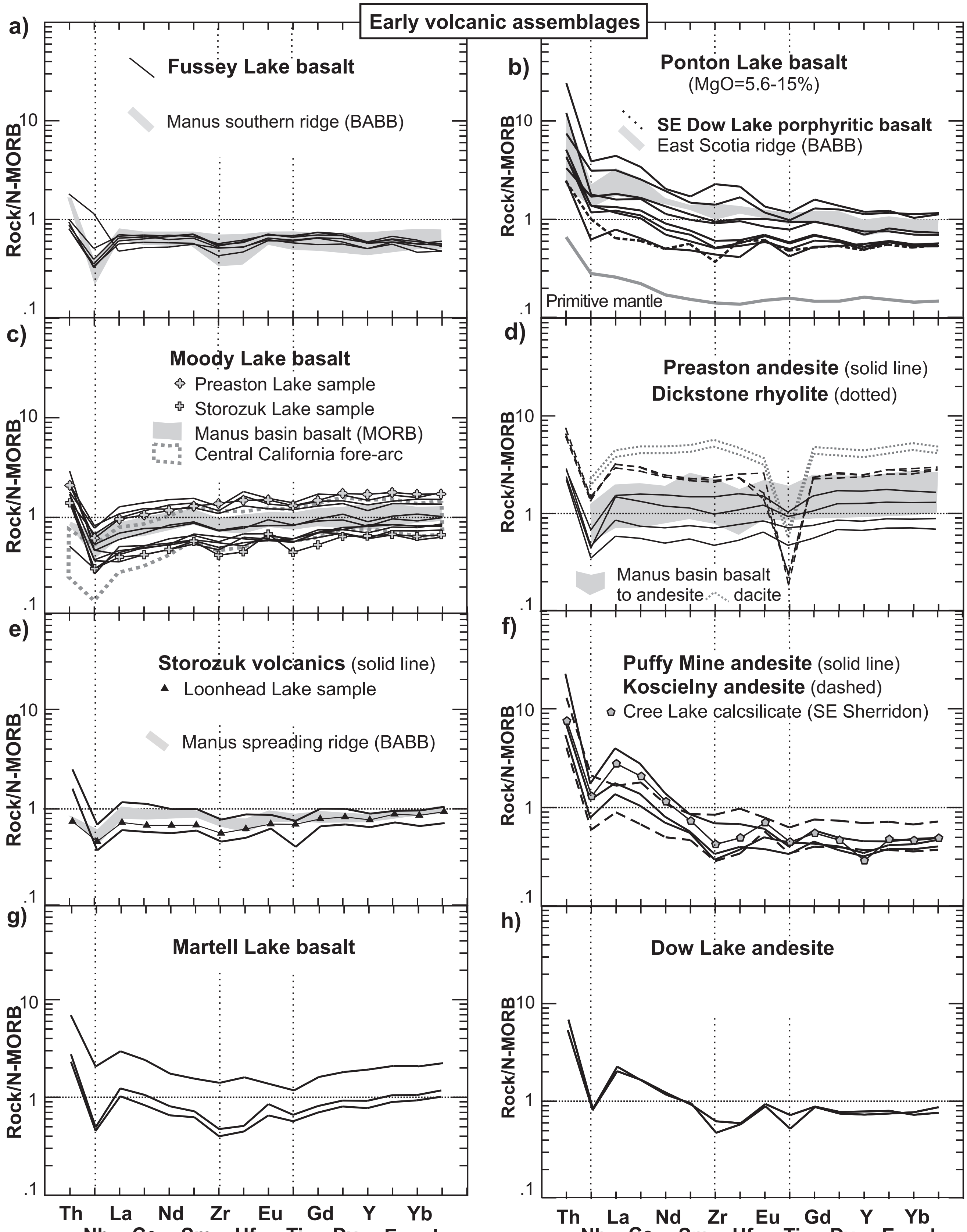


Figure 34: *N-MORB–normalized extended-element plots of early volcanic assemblages:* ***a–e)*** *low to moderate negative anomalies in HFSE and elevated Th are consistent with BABB origin, but each unit has a distinctive profile and different slope of REE (data for fields of comparison from Stern et al., 1990; Fretzdorff et al., 2002; Sinton et al., 2003;);* ***f–h)*** *the early arc assemblages have fractionated LREE and generally pronounced HFSE anomalies.*

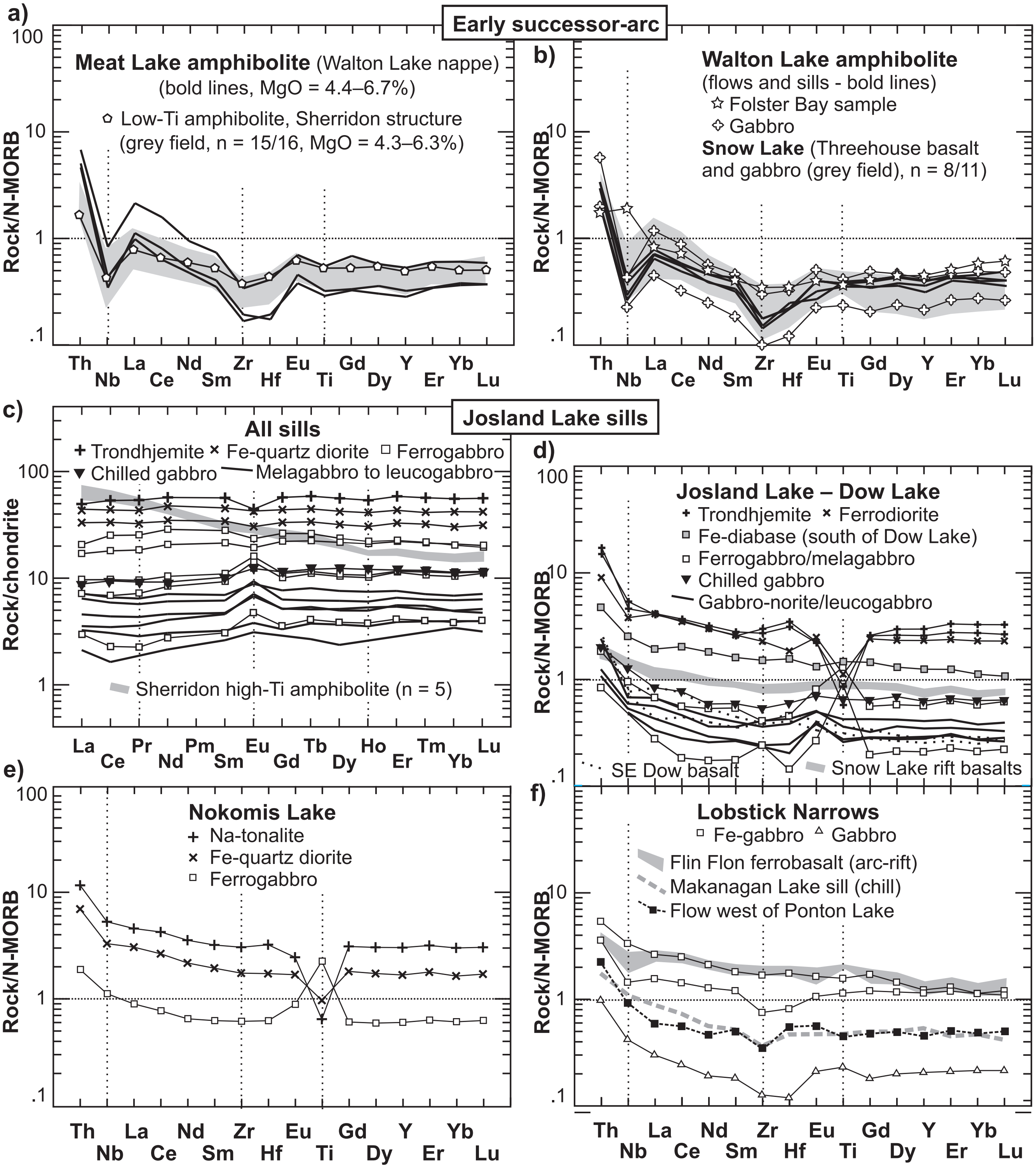


***Figure 35:* a, b)** *N-MORB–normalized extended-element plots of arc tholeiite, Sherridon–Meat Lake assemblage.* **c–f)** *Chondrite- and N-MORB–normalized extended-element plots of Josland Lake sills, showing strong crystal fractionation and similar patterns for widespread sills. Opposing anomalies above and below the chill phase show involvement of plagioclase for Eu (c), and Ti-bearing oxide for Ti (d, e). Comparative fields for Threehouse basalt and Snow Lake arc-rift basalt from Bailes and Galley (2001) suggest a similar tectonic origin for the amphibolite in the Walton Lake nappe and arc tholeiite in the Snow Lake area, whereas Sherridon high-Ti amphibolite is unrelated to the Josland Lake sills in the Amisk Collage but chemically similar to syntectonic mafic dikes (unit 18). Fields of amphibolite from the Sherridon structure (grey shading) are based on data from Halo Resources Ltd. (L Bloom, pers. comm., 2007).*

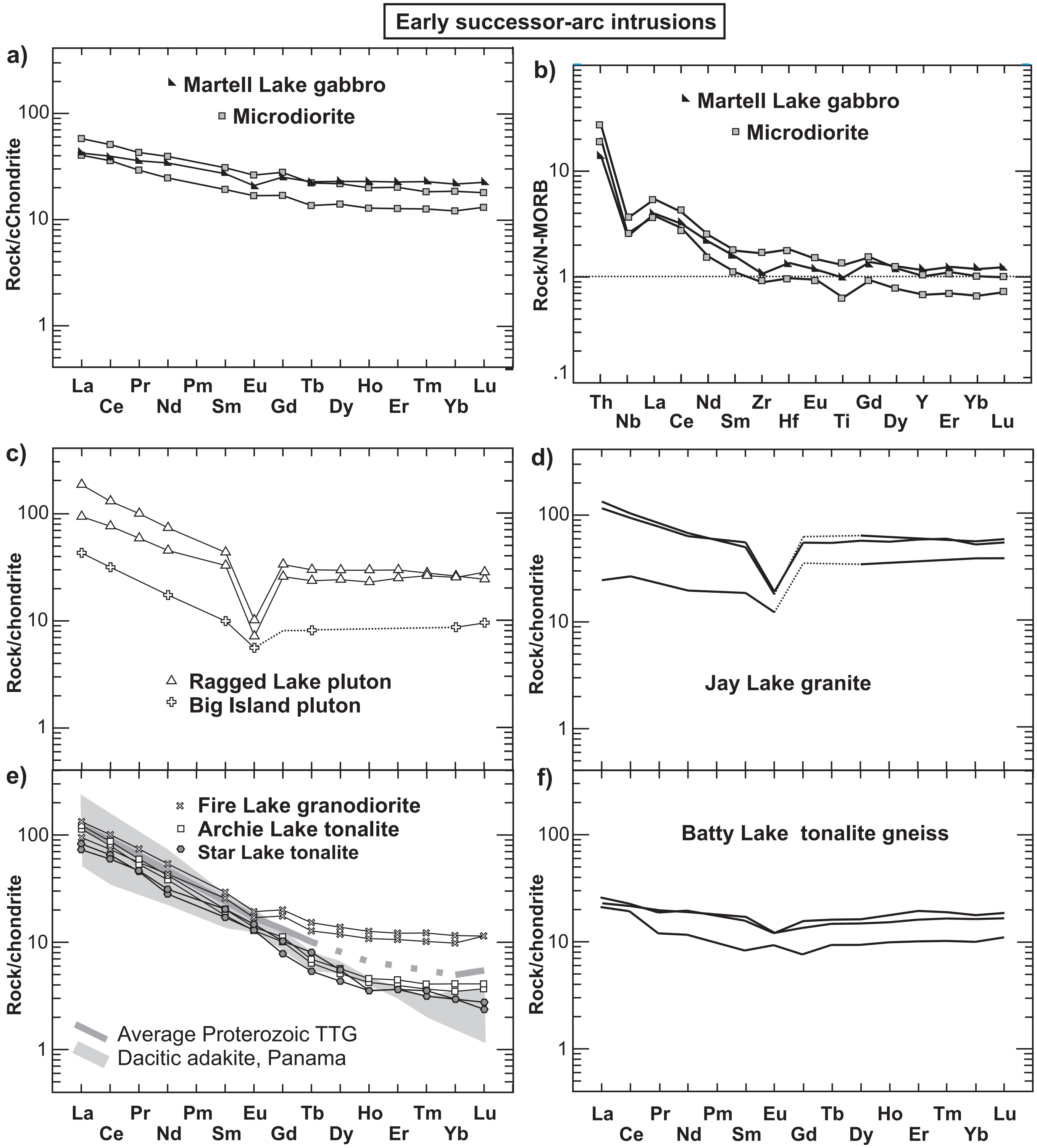


Figure 36: *Chondrite-normalized REE and N-MORB–normalized elements of early successor-arc intrusions with patterns of mafic–intermediate rocks similar to early arc-volcanic rocks (Appendix 3, Figure 33) but with higher REE contents* ***(a, b)*** *and HREE patterns of felsic plutons ranging from flat* ***(c–f)*** *to variably depleted* ***(e)****, probably by mixing with slab melts (adakite) or melts of early mafic arc crust. Field of Proterozoic tonalite-trondhjemite-granodiorite (TTG) from Condie (2008) and field of adakite from Defant et al. (1991).*

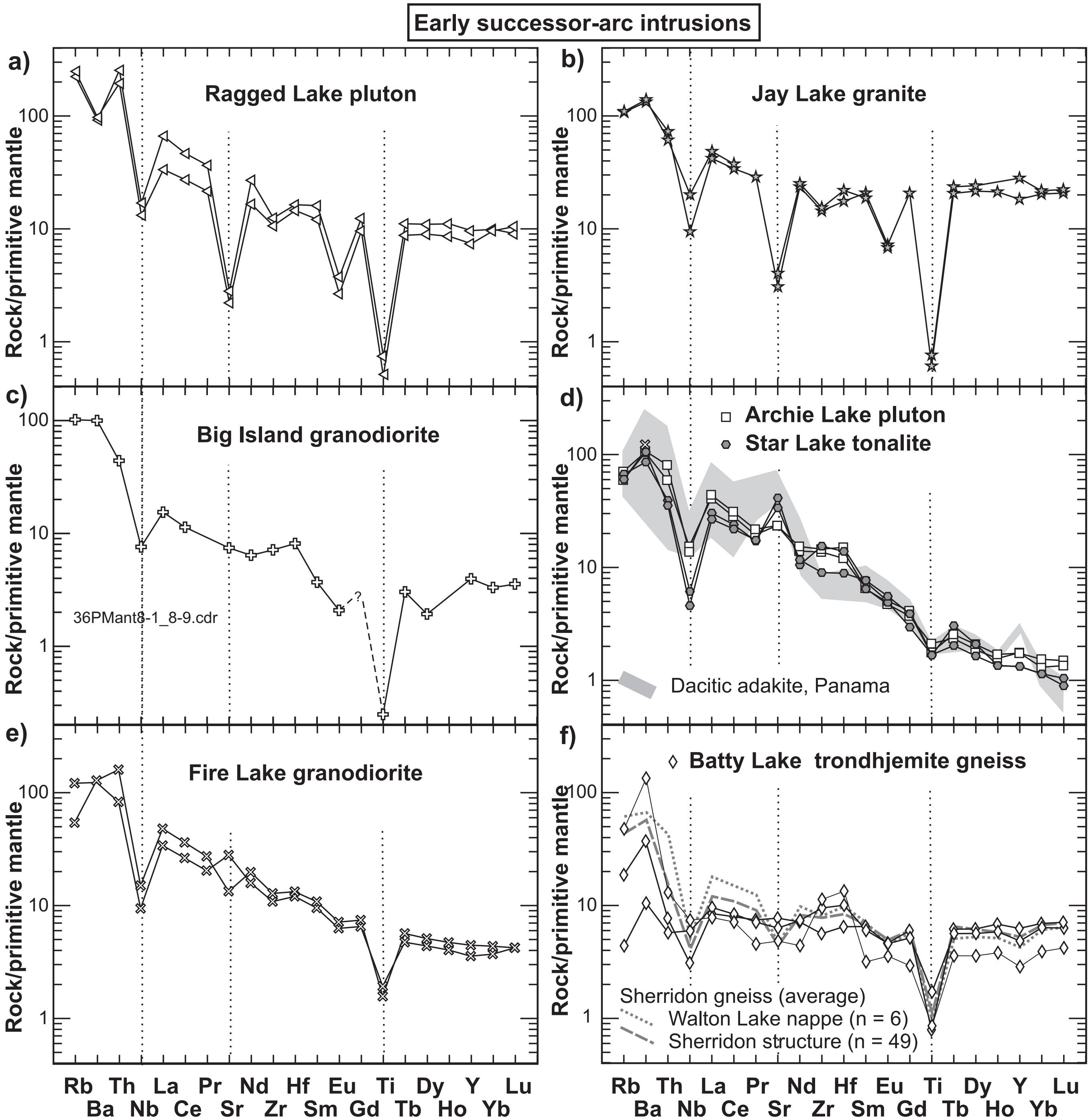
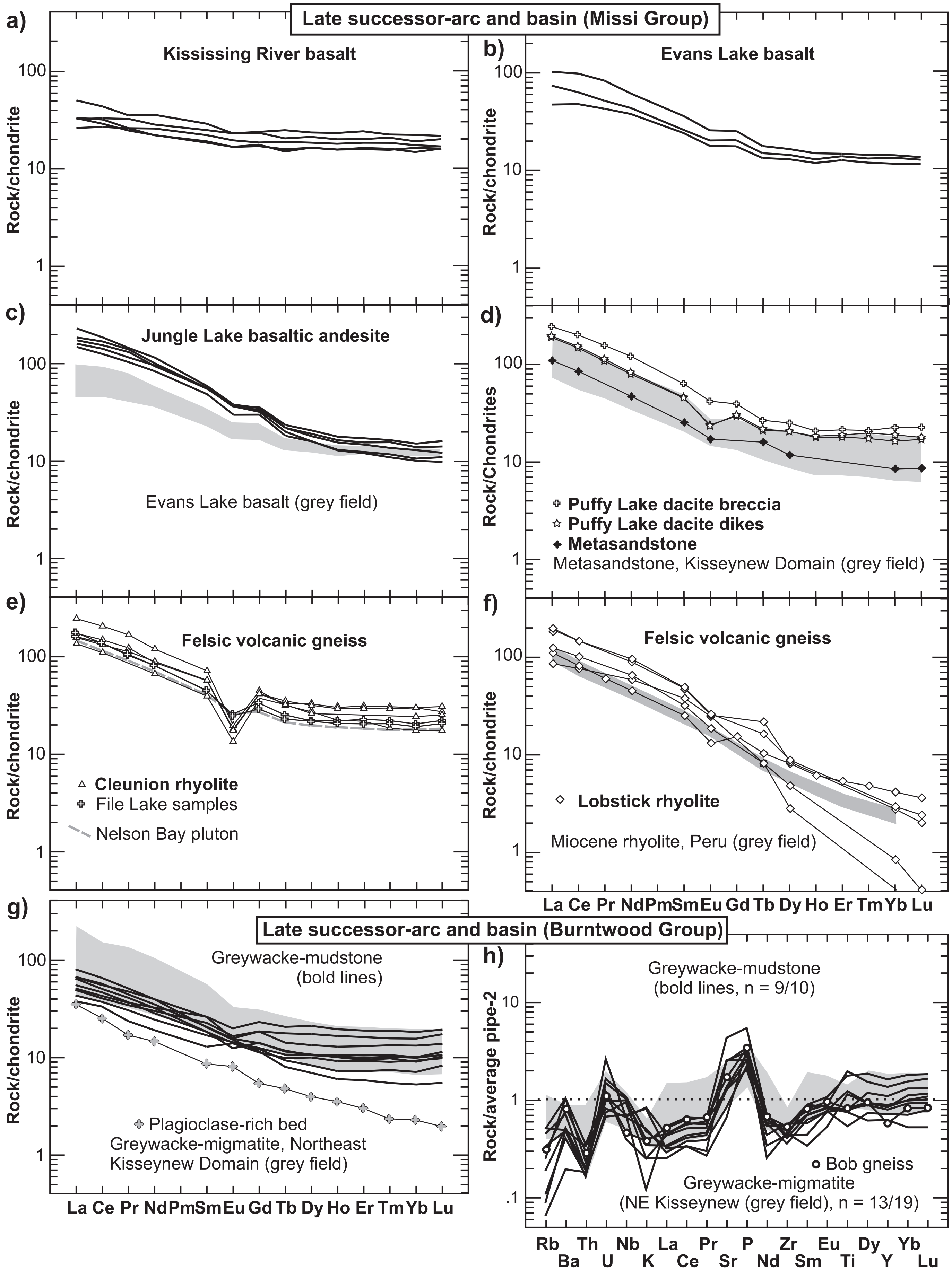


***Figure 37:** Primitive-mantle–normalized multi-element plots of early successor-arc plutons, some showing troughs of Sr and Eu that indicate plagioclase fractionation **(a, b)**, and most showing elevated Rb and Ba, and strong negative HFSE anomalies, typical of volcanic-arc granites. Batty Lake gneiss is similar to Sherridon felsic gneiss but variably depleted in mobile elements and LREE with respect to rhyolite gneiss in the Walton Lake nappe. Field of adakite (grey shading) from Defant et al. (1991). Dashed line is assumed; n is number of samples in average.*


Late successor-arc and basin (Missi Group)
a)
Kississing River basalt
b)
Evans Lake basalt
c)
Jungle Lake basaltic andesite
Evans Lake basalt (grey field)
d)
Puffy Lake dacite breccia
Puffy Lake dacite dikes
Metasandstone
Metasandstone, Kisseynew Domain (grey field)
e)
Felsic volcanic gneiss
Cleunion rhyolite
File Lake samples
Nelson Bay pluton
f)
Felsic volcanic gneiss
Lobstick rhyolite
Miocene rhyolite, Peru (grey field)
g)
Late successor-arc and basin (Burntwood Group)
Greywacke-mudstone
(bold lines)
Plagioclase-rich bed
Greywacke-migmatite, Northeast
Kisseynew Domain (grey field)
h)
Greywacke-mudstone
(bold lines, n = 9/10)
Bob gneiss
Greywacke-migmatite
(NE Kisseynew (grey field), n = 13/19)
Rock/chondrite
Rock/Chondrites
Rock/average pipe-2
La Ce Pr Nd Pm Sm Eu Gd Tb Dy Ho Er Tm Yb Lu
Rb Ba Th U Nb K La Ce Pr Sr P Nd Zr Sm Eu Ti Dy Y Yb Lu

Facing page

***Figure 38:** Chondrite-normalized REE plots of late successor-arc volcanic and sedimentary rocks of the Missi and Burntwood groups (a–g). The plots show basalt units with weakly and strongly fractionated REE **(a, b),** similar patterns of intermediate and felsic volcanic units **(c–e),** and a clear distinction between rhyolite units with undepleted and strongly depleted HREE **(e, f)**. Plots of Burntwood Group greywacke-mudstone in **(h)** are normalized using the average P2 member of the Pipe Formation in the Thompson Nickel Belt of pelite eroded from the Archean Superior craton (dotted line). The pelite average and fields from rocks in the northeastern Kisseynew Domain (excluding plots with outlying values) are from Zwanzig et al. (2007). The number of plotted samples (having no greatly anomalous values) over the total number of analyzed samples is designated as 'n' in (h).*

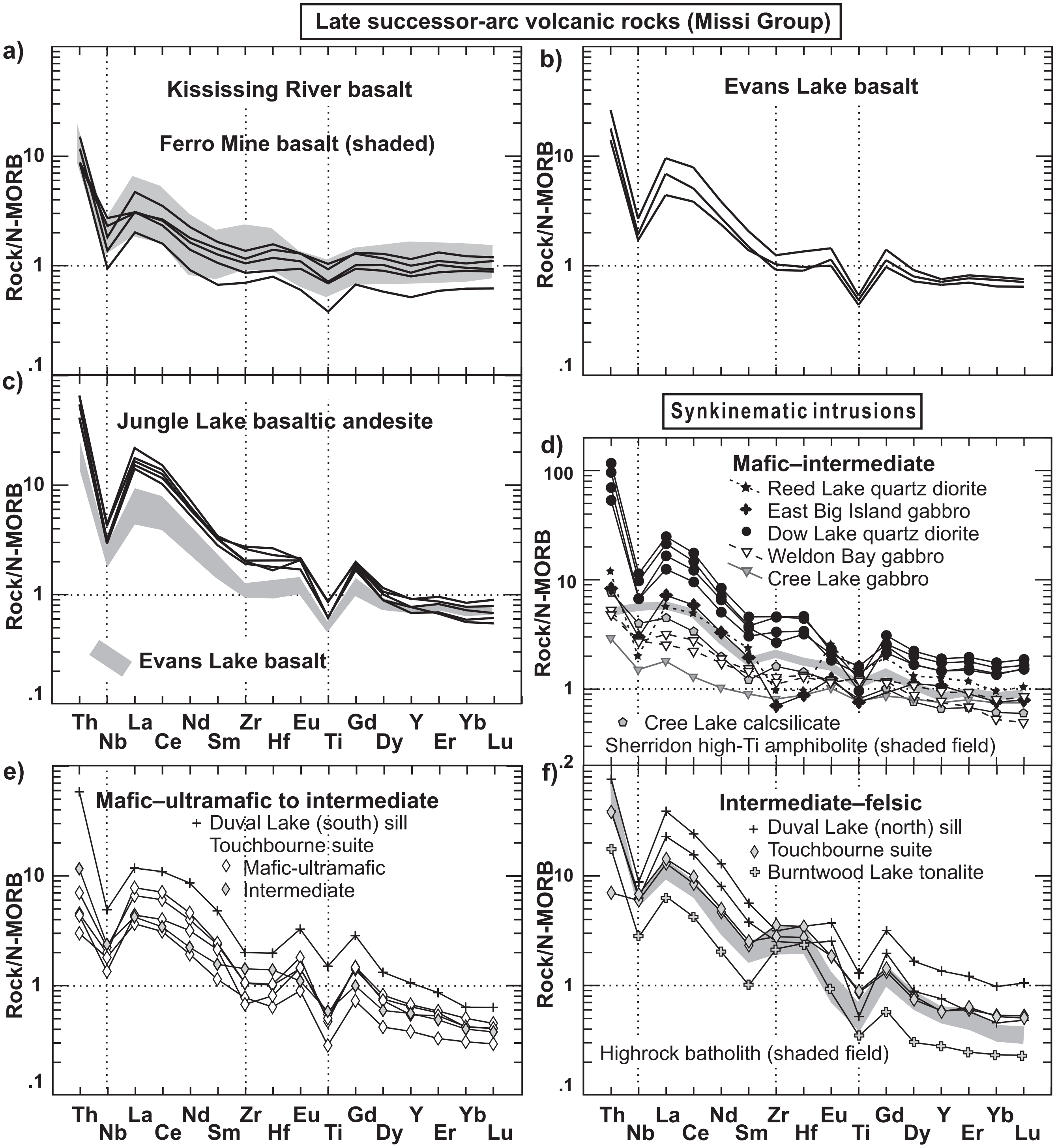
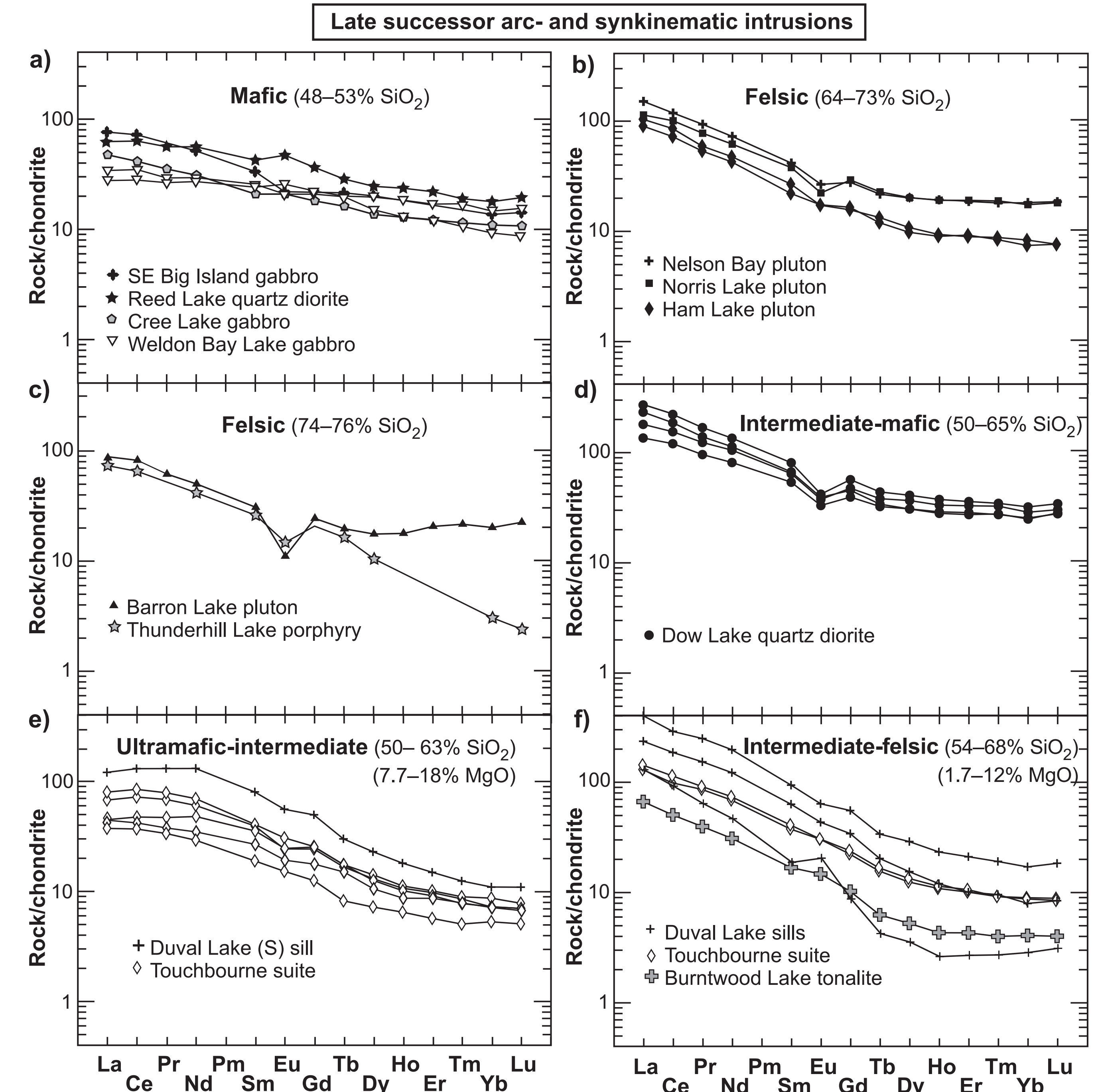


***Figure 39:** N-MORB–normalized extended-element plots of mafic late successor-arc igneous rocks, and mafic–intermediate synkinematic intrusions. Extrusive and intrusive units show similar strong negative HFSE anomalies in all but the youngest (postcollisional) gabbro (d); also distinct are two patterns for the intrusive rocks, possibly related to mixing of mantle and crustal melts and marking the end of arc magmatism. Data of Hollings and Ansdell (2002) are included (e, f).*



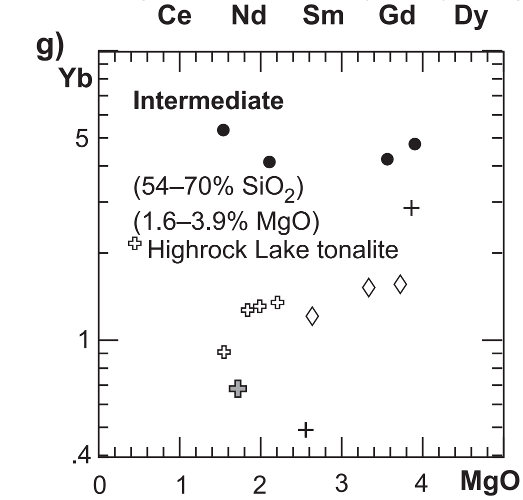


***Figure 40: a–f)** Chondrite-normalized REE plots of late successor-arc and synkinematic intrusions; Missi-age mafic and felsic intrusions **(a–c)** mimic those of the coeval volcanic rocks shown in Appendix 3, Figure 38; synkinematic gabbro units (**d**; Cree Lake, Weldon Bay) have slopes similar to, or shallower than, late arc gabbro; weak and strong fractionation patterns of LREE in late intrusions **(e–f)** occur equally on the Kisseynew South flank (Duval Lake sills) and in the Central Kisseynew subdomain (Touchbourne intrusive suite). **g)** Plot of Yb vs. MgO, showing REE increasing with mafic content in the intermediate compositions, suggesting mixing of felsic melt with high-REE mafic melt. Data from Hollings and Ansdell (2002) are included in parts e and g.*

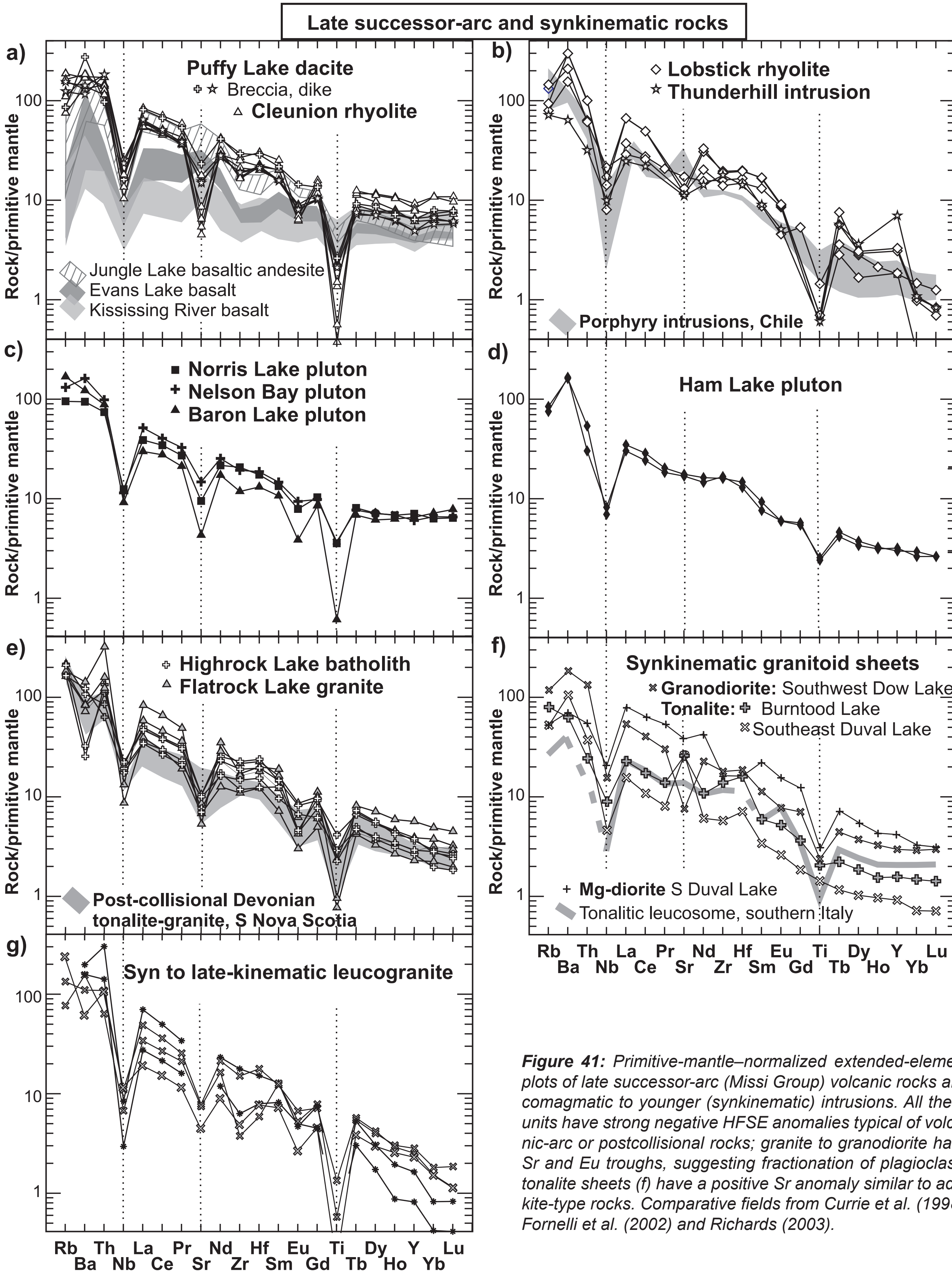


Figure 41: *Primitive-mantle–normalized extended-element plots of late successor-arc (Missi Group) volcanic rocks and comagmatic to younger (synkinematic) intrusions. All these units have strong negative HFSE anomalies typical of volcanic-arc or postcollisional rocks; granite to granodiorite have Sr and Eu troughs, suggesting fractionation of plagioclase; tonalite sheets (f) have a positive Sr anomaly similar to adakite-type rocks. Comparative fields from Currie et al. (1998), Fornelli et al. (2002) and Richards (2003).*

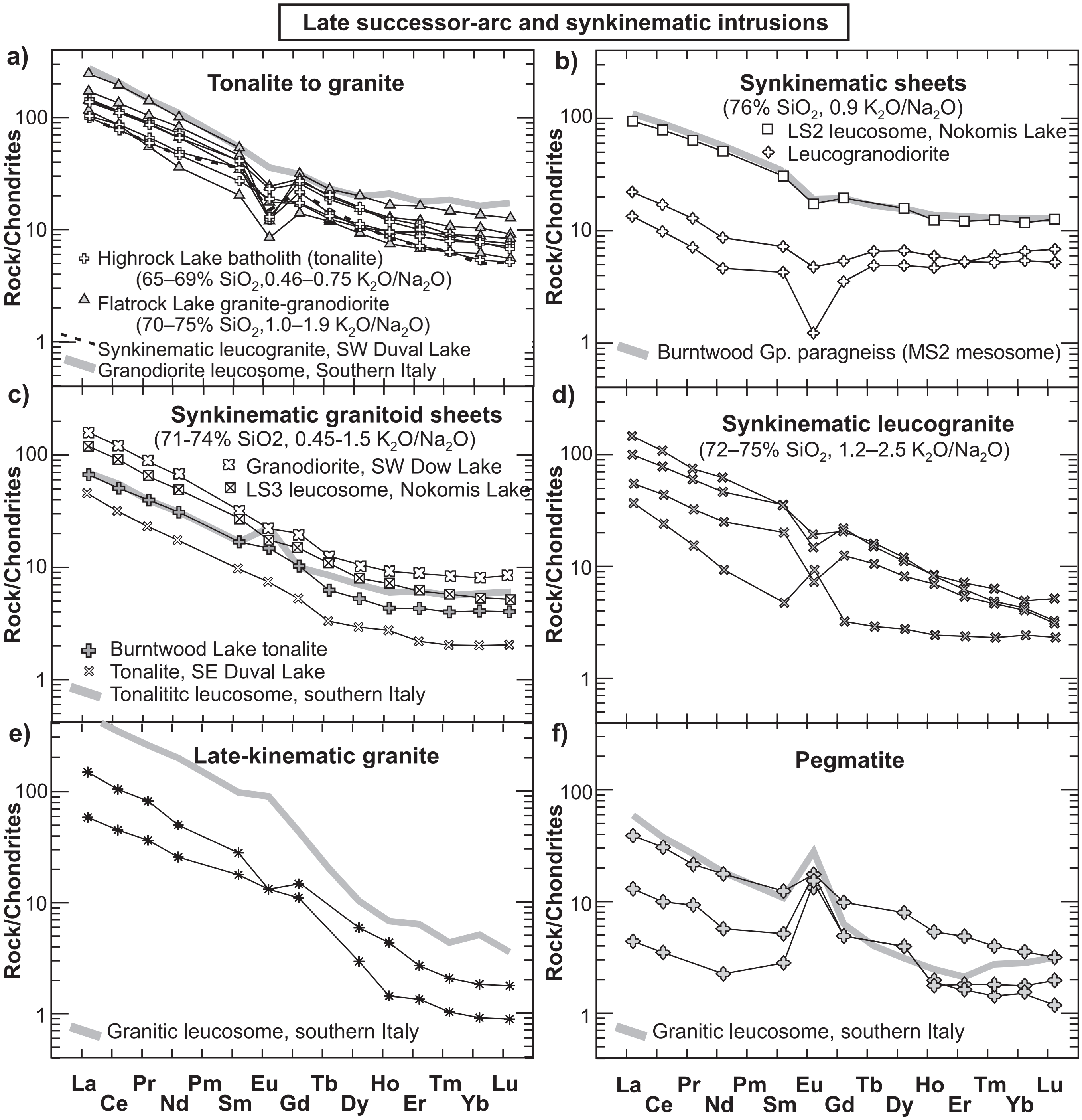


***Figure 42:** Chondrite-normalized REE plots of synkinematic granitoid intrusions: **a)** tonalite–granite plutons in the Central Kisseynew subdomain have similar patterns and origins; **b–f)** synkinematic granitoid sheets have a wide variety of patterns resulting from local migmatization injection of allochthonous lits (b); probably deeper crustal melting occurred with variable involvement of garnet in the residuum (low HREE in c–e) and loss of mafic phases or feldspar fractionation indicated by Eu anomalies (a, b, d, f). Comparative analyses of leucosome from Fornelli et al. (2002).*