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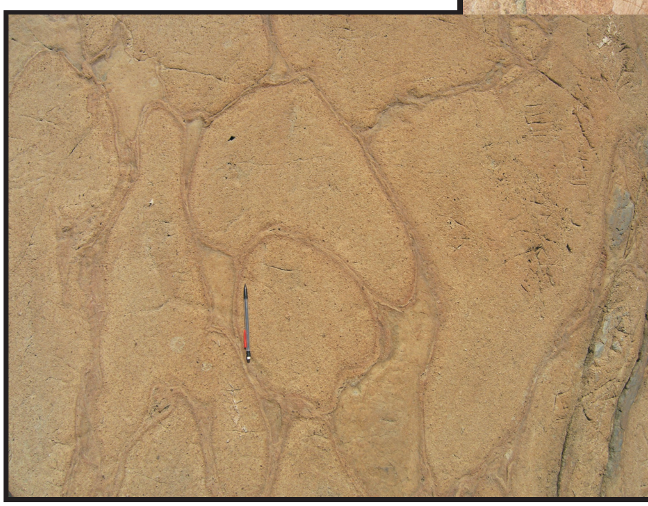
GEOSCIENTIFIC REPORT

Geology of the Flin Flon area, Manitoba and Saskatchewan (parts of NTS 63K12, 13)



By
R-L. Simard, K. MacLachlan, H.L. Gibson,
Y.M. DeWolfe, C.A. Devine, P.D. Kremer,
B. Lafrance, D. Ames, E.C. Syme, A.H. Bailes,
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Manitoba

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Author affiliations:

R-L. Simard, P.D. Kremer and E.C. Syme
Manitoba Innovation, Energy and Mines
Manitoba Geological Survey

K. MacLachlan
Saskatchewan Ministry of Energy and Resources
Saskatchewan Geological Survey

H.L. Gibson, Y.M. DeWolfe, B. Lafrance, K. Bailey,
E.M. Lewis, D. Lewis
Laurentian University
Department of Earth Sciences

Y.M. DeWolfe
Mount Royal University
Department of Earth Sciences

C.A. Devine
Hudson Bay Exploration and Development Co. Ltd.

D. Ames, S. Pehrsson and A.G. Galley
Geological Survey of Canada

A.H. Bailes
(formerly of the Manitoba Geological Survey)
Bailes Geoscience
Winnipeg, Manitoba

D. Price
Consultant
Flin Flon, Manitoba

Published by:

Manitoba Innovation, Energy and Mines
Manitoba Geological Survey
360–1395 Ellice Avenue
Winnipeg, Manitoba
R3G 3P2 Canada

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Cover illustration: Representative rock types and textures of the Hidden formation: plagioclase-phyric pillowed basalt of the Reservoir member (left photograph) and pillow breccia of the Carlisle member (right photograph).

Abstract

A five-year collaboration between the Manitoba Geological Survey, the Saskatchewan Geological Survey, the Geological Survey of Canada, Laurentian University and Hudson Bay Exploration and Development Company Limited, as part of the 2005–2010 Government of Canada Flin Flon Targeted Geoscience Initiative (TGI-3), has produced a new 1:10 000 scale bedrock geological map (Simard et al., 2010, Geoscientific Map MAP2010-1, back pocket) and a geological report for the Flin Flon mining camp.

The Flin Flon map area contains five of the Paleoproterozoic Flin Flon Belt's 27 producing and past-producing base-metal mines, making this area one of the most productive base-metal camps in Canada. The TGI-3 collaborative research project has led to the development of a coherent lithostratigraphic and structural framework that spans the border between Manitoba and Saskatchewan.

The structural architecture of the Flin Flon area and the geometry of the deposits have been shaped by seven deformation events (from D1 to D7) related to the ca. 1.87–1.80 Ga Trans Hudson Orogen, including several phases of early folding and thrusting and later phases of discrete faulting. This complex structural history has resulted in a map pattern of distinct stratigraphic and structural panels. Detailed studies have nevertheless defined coherent stratigraphic linkages across some of these structures.

The lithofacies, stratigraphy and geochemistry of volcanic strata hosting the main volcanogenic massive sulphide (VMS) deposits of the Flin Flon mining camp (Flin Flon–Callinan–777), west of the Channing–Mandy Road faults, record extension, subsidence and infilling of a complex subsidence structure or cauldron and the development of an intense hydrothermal system (Flin Flon formation) that terminated with a hiatus in active volcanism and the formation of the VMS deposits. Following VMS deposition there was resurgence in volcanism and continued subsidence marked by the development of one or more mafic shield volcanoes that buried this cauldron (Hidden and Louis formations). The mine-hosting succession and its hangingwall have been dated to ca. 1889 Ma, demonstrating rapid evolution in a dynamic volcanic environment.

The volcanic rocks between the Channing–Mandy Road faults and the Cliff Lake Fault to the east host the Schist Lake and Mandy VMS deposits. Although these rocks are economically significant, they have not been easily correlated with any single stratigraphic unit in the Flin Flon mining camp to date. The Western Hook Lake succession farther east of the Cliff Lake Fault is plausibly correlative with the rocks of the Flin Flon mining camp based on its lithostratigraphy and age. The Eastern Hook Lake succession is 10 m.y. younger than the rocks of the Flin Flon mining camp, but is similar to and overlaps in age with rocks that host the Trout Lake VMS deposit 10 km to the north.

Reconstruction of the volcanic architecture of the Flin Flon mining camp through detailed mapping, as well as geochemical and geochronological study, provides unequivocal evidence for extension of an arc-like volcanic edifice and concomitant cauldron subsidence during VMS formation. An improved understanding of the importance of subsidence structures and associated high-temperature hydrothermal alteration in the generation of the VMS deposits has led to the recognition of favourable host stratigraphy and exploration potential in areas well outside the immediate mine surroundings.

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Introduction

The Flin Flon map area, located in the eastern part of the Paleoproterozoic Flin Flon Belt (Figure 1), contains five of the belt's 27 producing and past-producing base-metal mines (Syme et al., 1996). These five mines are the currently producing 777 mine and the past-producing Flin Flon, Callinan, Schist Lake and Mandy mines, which makes this area one of the most productive base-metal camps in Canada.

The Flin Flon area has been mapped at various scales by geologists from several organizations over the past 70 years (see details below). Despite the fact that each of these mapping efforts was successful in increasing our understanding of this prolific base-metal area, the need for a comprehensive, cross-border, state-of-the-art, detailed lithostratigraphic bedrock map of the area, to support exploration for VMS deposits, still existed in the early 2000s.

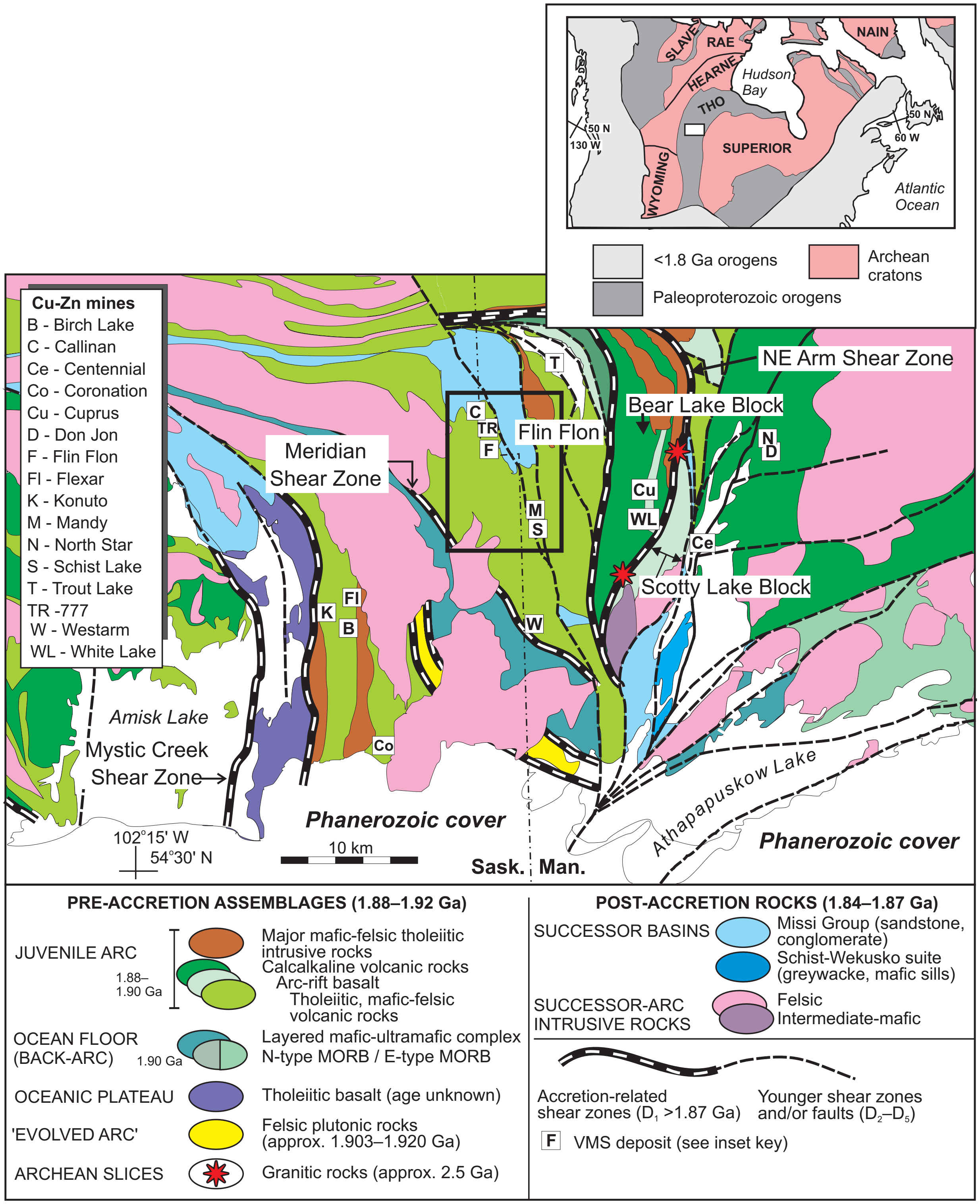


Figure 1: *Geology of the western portion of the Flin Flon Belt, showing locations of known volcanogenic massive sulphide (VMS) deposits (modified from Syme et al., 1999a); box indicates the area covered by Figure 3; inset map shows the location of the Flin Flon Belt within the Trans-Hudson Orogen (THO).*

With the intent of stimulating private-sector resource exploration in areas of high base-metal potential in established mining communities, the Government of Canada launched a five-year Targeted Geoscience Initiative (TGI-3) in 2005. As part of this initiative, the Manitoba Geological Survey, in collaboration with the Saskatchewan Geological Survey, the Geological Survey of Canada and researchers from Laurentian University, took on the production of a new 1:10 000 scale geological map of the Flin Flon area.

The Simard et al. (2010) map, found in the back pocket of this report (Geoscientific Map MAP2010-1), is the culmination of a team effort by 16 geoscientists from government, academia and industry over the five-year duration of the Flin Flon TGI-3 project. The new geological map of Flin Flon results from a combination of 1) detailed (1:500–1:5000 scale) new/recent mapping, mainly of the volcanic stratigraphy in the Flin Flon area, which took place as part of either the Flin Flon TGI-3 project described above or the Flin Flon TGI-1 project, a Natural Sciences and Engineering Research Council of Canada (NSERC)–HudBay Minerals–Laurentian University Collaborative Research and Development (CRD) Grant (supported by the Manitoba and Saskatchewan geological surveys) in the early 2000s; and 2) compilation of previous mapping efforts, mainly on the younger sedimentary and intrusive rocks and outlying areas, by various geological surveys and/or exploration companies over the years (see 'Previous work' section below for details). Given the underlying base-metal theme of the TGI-3 program, new mapping in the Flin Flon area was focused on the volcanic rocks that host and/or were the most prospective for volcanogenic massive sulphide (VMS) deposits, which is why most of the 'compiled' geology is from areas dominated by younger sedimentary and/or plutonic rocks, or in volcanic outliers far from the known VMS-hosting stratigraphic successions. Both Figure 2 and the map in the back pocket (Geoscientific Map MAP2010-1) include a 'Mapping Index' and a complete list of references detailing the extent of new mapping and compiled areas.

As with the map (Simard et al., 2010, Geoscientific Map MAP2010-1, back pocket), this report is a result of a collaborative effort by numerous co-authors over the last four years. Although R-L. Simard and K. MacLachlan are the principal authors of this report, the specific contributions of co-authors are indicated for each section (e.g., Flin Flon formation, 'H.L. Gibson, C.A. Devine and K. MacLachlan'). In areas where the geology was not entirely remapped as part of this study, the unit descriptions are based on previously published work, with the appropriate reference(s) provided after the title or subtitle of that section (e.g., 'from Thomas, 1989', 'R-L. Simard and P.D. Kremer; and from Bailes and Syme, 1989').

This report focuses on providing lithological and stratigraphic descriptions of the various rock packages found in the map area and their relationships/implications with regards to the VMS mineralization in the area. The rock packages in the map area are described from older to younger and from west to east in between major faults. The geochemistry and geochronology of the various rock packages are not discussed in detail in this report; the key geochemical and geochronological results are stated within the descriptions, including all original references to which the reader can refer for full details.

Previous work

The Flin Flon area is particularly suitable for detailed bedrock mapping for a number of reasons: bedrock exposure is commonly from 40 to 80%; the outcrops are virtually lichen-free; and the regional metamorphic grade is lower greenschist facies (epidote-chlorite, Bailes and Syme, 1989), resulting in good preservation of primary textures and structures. These exceptional characteristics have resulted in repeated mapping efforts in the area over the years.

The Flin Flon area was mapped first at 1:63 360 scale by Tanton (1941a, b), Buckham (1944) and Bateman and Harrison (1945). More detailed mapping at 1:12 000 scale was carried out by Stockwell (1960) in the vicinity of Flin Flon. The development of plate tectonic theory, inexpensive and precise geochemical analytical methods, and a better understanding of physical volcanology generated a second wave of bedrock mapping of the area during the late 1970s and early 1980s. A modern 1:20 000 scale bedrock map of the Flin Flon–White Lake area was produced by Bailes and Syme (1989) of the Manitoba Geological Survey, whereas a comparable 1:10 000 scale bedrock map of the Douglas–Phantom Lake area was produced by Thomas (1990, 1992) of the Saskatchewan Geological Survey. Stockwell's map is the only transborder map of the area, and is still in common use.

In addition to mapping, contributions to the understanding of the volcanic rocks in the Flin Flon arc assemblage have been made by Byers and Dahlstrom (1954), Byers et al. (1965), Stauffer et al.(1975), Ayres (1978, 1982), Van Wagoner and Van Wagoner (1982), Bailes and Syme (1989) and Thomas (1989, 1990, 1991, 1992, 1993, 1994). Recent work by Gibson et al. (2001, 2005, 2007, 2009, in press), Ames et al. (2002), MacLachlan et al. (2002), Devine (2003), DeWolfe and Gibson (2004, 2005, 2006), Bailey (2006), Gibson (2006), MacLachlan (2006a, b, c), Simard (2006a, b), Cole et al. (2007, 2008), Kremer and Simard (2007), MacLachlan and Devine (2007), Simard and Creaser (2007), Simard et al. (2007), DeWolfe (2008, 2009, 2010, 2011) and DeWolfe et al. (2009a, b) has shaped the newly proposed stratigraphy for the area (Table 1, back pocket).

Sedimentary rocks of the Missi Group have been described by Bruce (1918), Ambrose (1936a, b), Buckham (1944), Byers (1953), Byers and Dahlstrom (1954), Byers et al. (1965), Stauffer and Mukherjee (1971), Stauffer (1974) and Bailes and Syme (1989).

The structure of the area was described by Ambrose (1936b), Stockwell (1960), Stauffer and Mukherjee (1971), Bailes and Syme (1989), Thomas (1989, 1990, 1991, 1992, 1993, 1994), Fedorowich et al. (1995), and Gale et al. (1996, 1999). Recent work by Lewis et al. (2006, 2007), Cole et al. (2007), Lafrance et al. (2007, 2009, in press), Lewis (2011) and S. Pehrsson (unpublished data, 2006–2008), in conjunction with 3-D modelling of the area by HudBay Minerals Inc. and the Geological Survey of Canada (Schetselaar et al., 2010), utilizing recently acquired 2-D and 3-D seismic data has

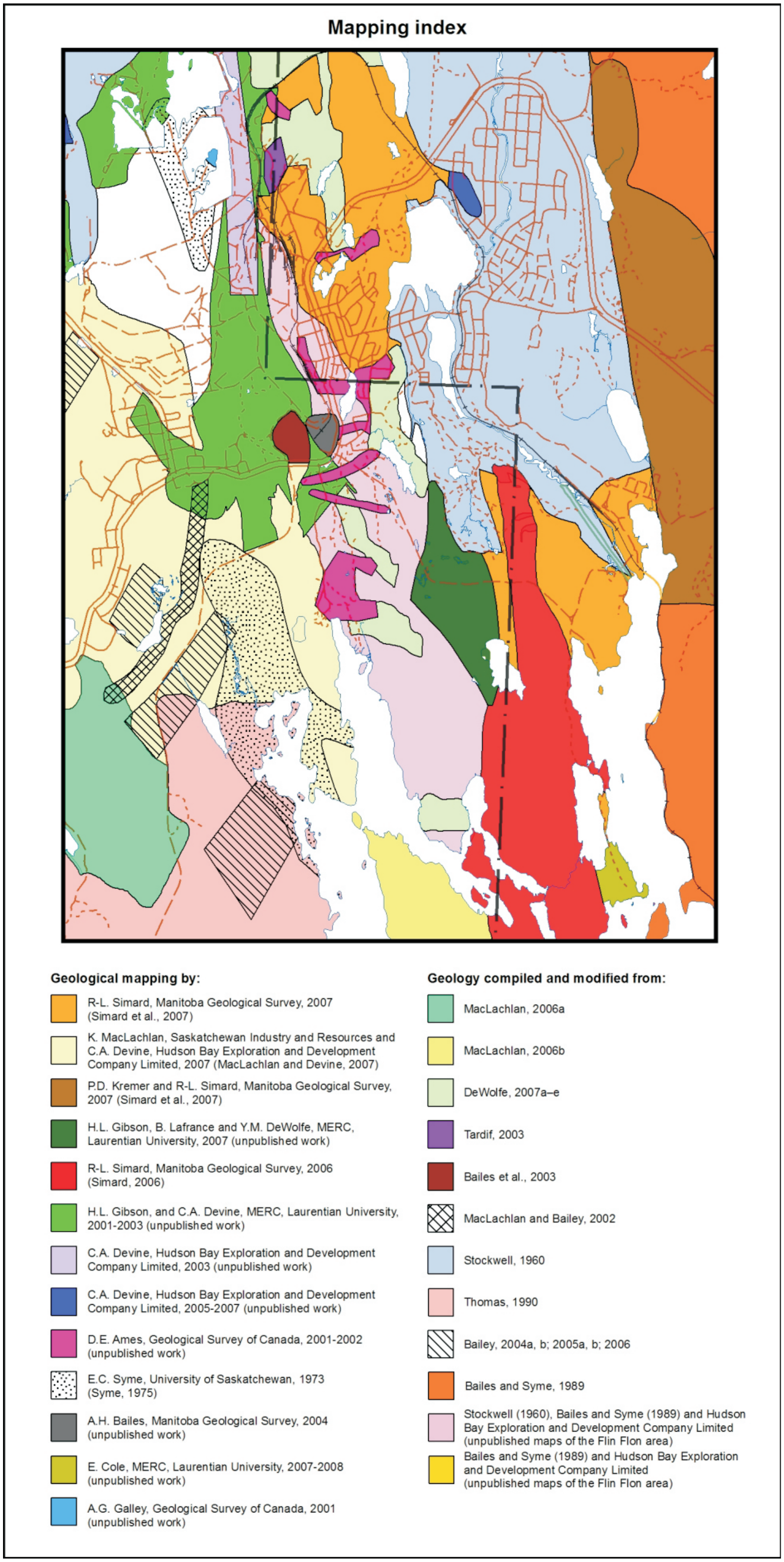


***Figure 2**: 'Mapping Index' detailing the extent of new mapping and compiled areas for the Simard et al. (2010) 1:10 000 scale bedrock map of the Flin Flon area, Manitoba and Saskatchewan. Refer to the lower portion of the map in the back pocket (Geoscientific Map MAP2010-1) for the complete list of references.*

advanced and refined our understanding of the structural history of the area (see 'Structural history' below).

Base-metal deposits, occurrences and related alteration have been described by several authors over the years. Bruce (1918), Byers et al. (1965), Sangster (1972), Koo and Mossman (1975), Price (1977) and Gale et al. (1980) provided the first descriptions of the Schist Lake, Mandy, Flin Flon, Callinan and 777 deposits. Detailed work by Bailes and Syme (1989) provided the first interpretations of the stratigraphic setting of individual sulphide deposits in the area (Syme and Bailes, 1993). Recent alteration studies of the Flin Flon–Callinan–777 massive sulphide deposits have been done by Ames et al. (2002, 2003) and Tardif (2003), whereas Cole et al. (2007, 2008), DeWolfe (2009, 2010, 2011) and Lewis (2012), investigated the alteration and stratigraphy of the Schist Lake–Mandy deposits.

Gold deposits (Newcor, Bootleg/Rio and Henning-Maloney) in the Douglas Lake area were described by Byers et al. (1965), and Ansdell and Kyser (1992). Information on specific occurrences and showings in the Flin Flon area is summarized in the Saskatchewan Mineral Deposit Index (Saskatchewan Energy and Resources, 2006) and in the Manitoba Mineral Deposit Series reports (Gale and Eccles, 1988a, b; see 'Economic geology' for more details).

Regional setting

The Paleoproterozoic Flin Flon Belt is part of the juvenile Reindeer Zone of the Trans-Hudson Orogen (Figure 1), a collision zone formed during the 2.0–1.8 Ga amalgamation of several Archean cratons into Laurentia (Hoffman, 1988). The Flin Flon Belt consists of a series of tectonostratigraphic assemblages (juvenile arc, juvenile ocean-floor back arc, ocean plateau, oceanic-island basalt and evolved plutonic arc) that range in age from 1.87 to 1.92 Ga (Syme et al., 1999a). The belt was assembled first through a period of intraoceanic accretion that amalgamated part of these arcs and ocean basins at around 1.88–1.87 Ga to form the Amisk Collage (Lucas et al., 1996). This accretionary complex was subsequently 'stitched' by calcalkaline plutons related to the development of a 1.87 to 1.84 Ga successor arc, with coeval subaerial volcanism (ca. 1.87–1.85 Ga, Lucas et al., 1996). During the waning stage of successor-arc magmatism, erosional denudation during uplift of this collage led to the development of a paleosol and deposition of the continental (alluvial-fluvial) sedimentary rocks of the Missi Group (ca. 1.85–1.84 Ga, Lucas et al., 1996). All of the VMS deposits mined to date in the Flin Flon area are hosted within the 1.9 Ga juvenile Flin Flon arc assemblage (Syme et al., 1999a).

Volcanic rocks of the Flin Flon arc assemblage consist of tholeiitic, subaqueous pillowed basalt and basaltic andesite, with lesser amounts of heterolithic mafic and lesser felsic volcaniclastic rocks, and minor dacite to rhyolite flows (Bailes and Syme, 1989). Within the dominantly mafic volcanic complex at Flin Flon, the VMS deposits are spatially associated with felsic volcanic units that occur in synvolcanic collapse structures (Bailes and Syme, 1989; Syme and Bailes, 1993; Figure 3).

Structural history

(B. Lafrance, P.D. Kremer and S. Pehrsson, modified from Lafrance et al., in press)

Seven deformation events from D_1 to D_7 are recorded in the rocks of the Flin Flon mining district (Lafrance et al., in press). They are summarized in Table 2, and the structures that formed during these events are labelled and shown in Figure 4.

The first two deformation events, D_1 and D_2, occurred during intraoceanic accretion of the Flin Flon arc to other volcanic terranes to form the Amisk Collage. These two events resulted in the formation of the D_1/F_1 Burley Lake syncline and the north-northwest-striking D_2/F_2 Hidden Lake fold system. The Burley Lake syncline is faulted along its hinge by the Burley Lake Fault. The younger Hidden Lake fold system includes the Beaver Road anticline, the Hidden Lake syncline, the Mandy Road anticline, the Carlisle Lake anticline and the Milliken Lake syncline(?). These folds, which represent the most prominent regional structures in the volcanic rocks, formed before the emplacement of ca. 1872 Ma intermediate dikes (Rayner, 2010) and prior to the deposition of fluvial sandstone and conglomerate of the Missi Group. A weak axial planar cleavage (S_2) is associated with the Hidden Lake fold system.

During D_3 deformation, the Flin Flon arc volcanic rocks were thrust and intercalated with the younger sedimentary rocks of the Missi Group, possibly in response to the final accretion of the Flin Flon Terrane to the Glennie Terrane. This produced an east-dipping thrust stack of basement and cover rock panels bounded by north-northwest-striking, west-directed, thrust faults represented by the 1920, Flin Flon Lake, Cliff Lake, Hook Lake(?) and Manistikwan Lake(?) faults (Figure 5a). West-verging, map-scale folds (Pipeline, Mud Lake and Grant Lake synclines) developed within the Missi Group cover rock panels during thrusting. The development of this D_3 thrust-and-fold belt occurred during or shortly after the deposition of the Missi Group cover rocks. Both events are bracketed in age between ca. 1.847 Ma, the age of the youngest detrital zircon in the Missi Group (Ansdell et al., 1992), and ca. 1842 Ma, the age of crosscutting Boundary intrusions (see 'Post–Missi Group intrusive rocks'). Thus, the basement volcanic rocks were uplifted during D_3, eroded, and covered by Missi basins, which were later deformed during westward migration and propagation of the thrust-and-fold belt.

A second, north-directed, D_4 thrust system formed during collision of the amalgamated Flin Flon–Glennie complex with the Sask craton as the oceanic crust that separated these two entities disappeared through subduction beneath the Flin Flon–Glennie complex. North-directed thrust faults (Club Lake, Railway and Catherine faults), east-trending folds (Flin Flon Creek syncline) and a strong, regional, southeast-plunging (30–50°) stretching lineation, formed during this thrusting event (Figure 5b). The development of this thrust system, which truncates the early west-directed D_3 thrust faults, was broadly coeval with, but outlasted the emplacement of, ca. 1838 Ma Phantom Lake dikes (see details below). The formation of the stretching lineation had a strong effect on the orientation of pre-existing regional folds whose axes were rotated into

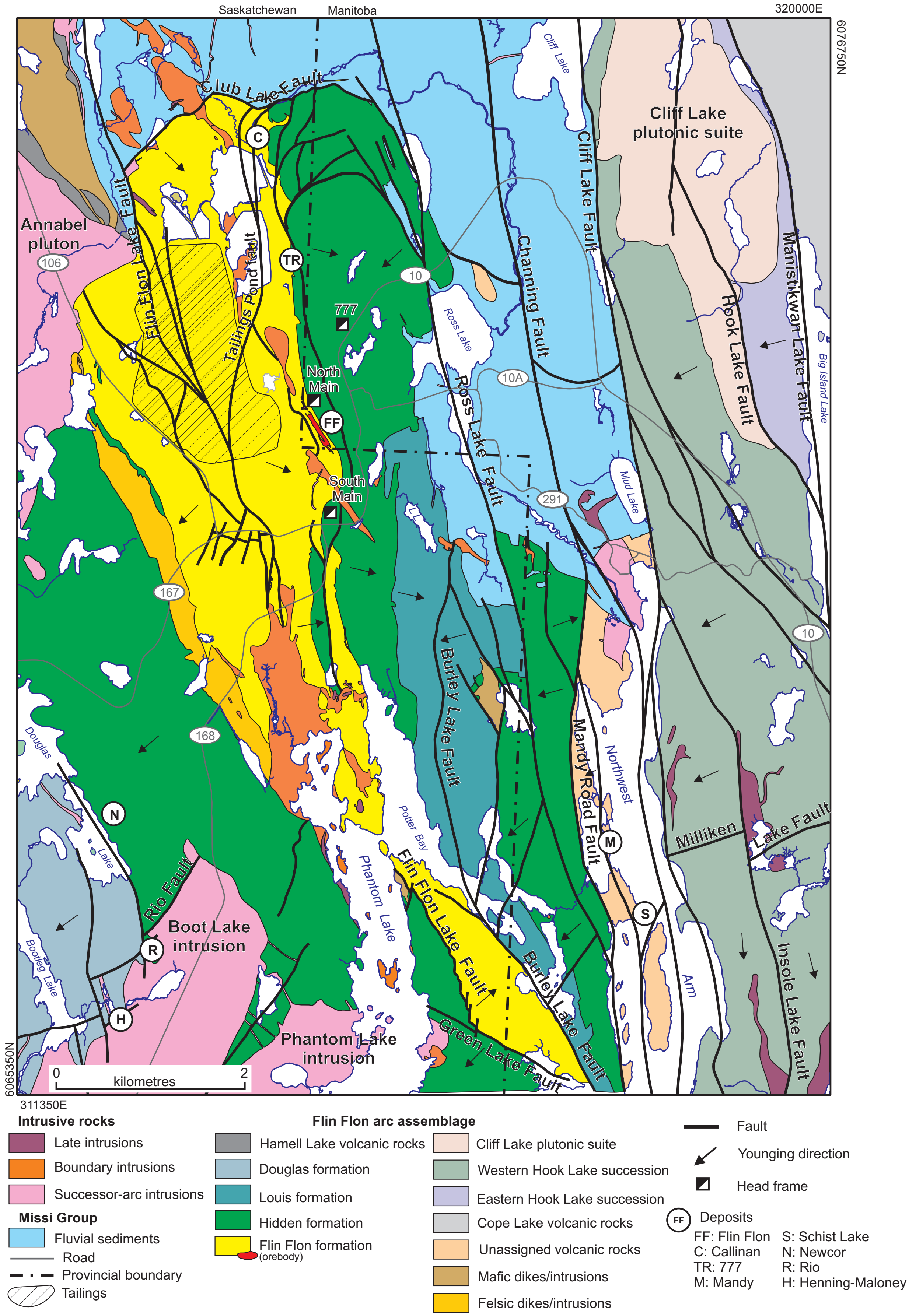


Figure 3: *Simplified geology of the Flin Flon area, showing the main stratigraphy and major structures in the area. Abbreviation: LL, Louis Lake.*

Table 2: Summary table of deformation events and their associated structures for the Flin Flon area and their relative timing and tectonic correlations (modified from Lafrance et al., in press).

Deformation event/Associated structures		Timing/Tectonic event
D_1		Pre-1872 Ma intraoceanic accretion of the 1888 Ma Flin Flon arc to other volcanic terranes
F_1	Fold	
	Burley Lake syncline Burley Lake Fault (?)	
D_2		
F_2	Folds	
	Hidden Lake syncline Beaver Road anticline Mandy Road anticline Carlisle Lake anticline Milliken Lake syncline (?)	
S_2	Axial plane cleavage	
D_3		
D_3	Thrust faults, NNW- to NNE-striking	Collision of Flin Flon and Glennie domains
	1920 fault and other unnamed faults; thrust imbrications and stacking of VMS ore horizons	Produced west-propagating thrust-fold belt and basin between 1847 and 1842 Ma in which fluvial sedimentary rocks (Missi Group) were deposited
F_3	Folds, map scale syncline	
	Pipeline syncline Grant Lake syncline Mud Lake syncline	
D_4		
D_4	Thrust faults, E-striking	Subsequent collision of the Flin Flon-Glennie Complex with the Archean Sask craton ~1840 Ma
	Club Lake Fault Railway Fault Catherine Fault	
F_4	Folds, E-trending synclines	
	Flin Flon Creek syncline Ross Creek syncline Unnamed syncline at the Missi outliers	Second truncating fold-thrust system
D_4	SE-plunging regional stretching lineation	
D_5		
D_5	Dextral shear zones	
S_5	Regional NNW-striking foliation	
D_5	Dextral reactivation of NNW-striking D_3 thrust faults and lithological contacts	
D_6		Terminal collision of the Sask craton and the Flin Flon-Glennie Complex with the Superior craton from 1.83 to 1.79 Ga
S_6	Regional NE-striking foliation	
D_6	Sinistral reactivation of following D_3 thrust faults	
	Cliff Lake Fault	
	Flin Flon Lake Fault	
D_6	Dextral reactivation of following D_4 thrust faults	
	Railway Fault	
	Club Lake Fault	
D_7		Late continued collision of the Sask craton, Flin Flon-Glennie Complex and the Superior craton characterized by strike-slip faulting from 1.79 to 1.75 Ga
D_7	Late brittle faults	
	Ross Lake Fault Channing Fault (?) Insole Lake Fault (?)	

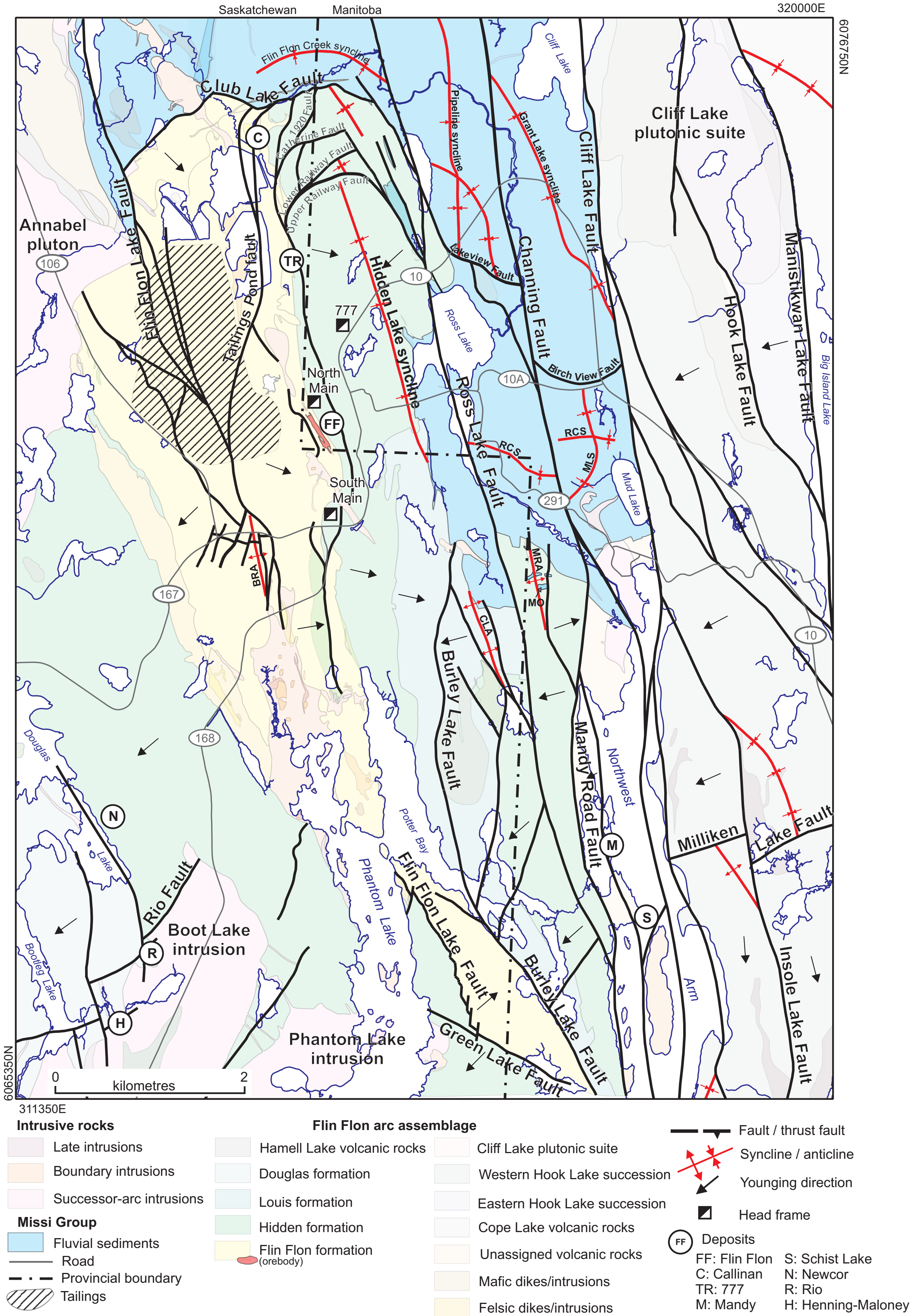


Figure 4: *Simplified geology of the Flin Flon area highlighting major faults and folds in the area. Abbreviations: BRA, Beaver Road anticline; CLA, Carlisle Lake anticline; MLS, Mud Lake syncline; MO, Missi Group outliers; MRA, Mandy Road anticline; RCS, Ross Creek syncline.*

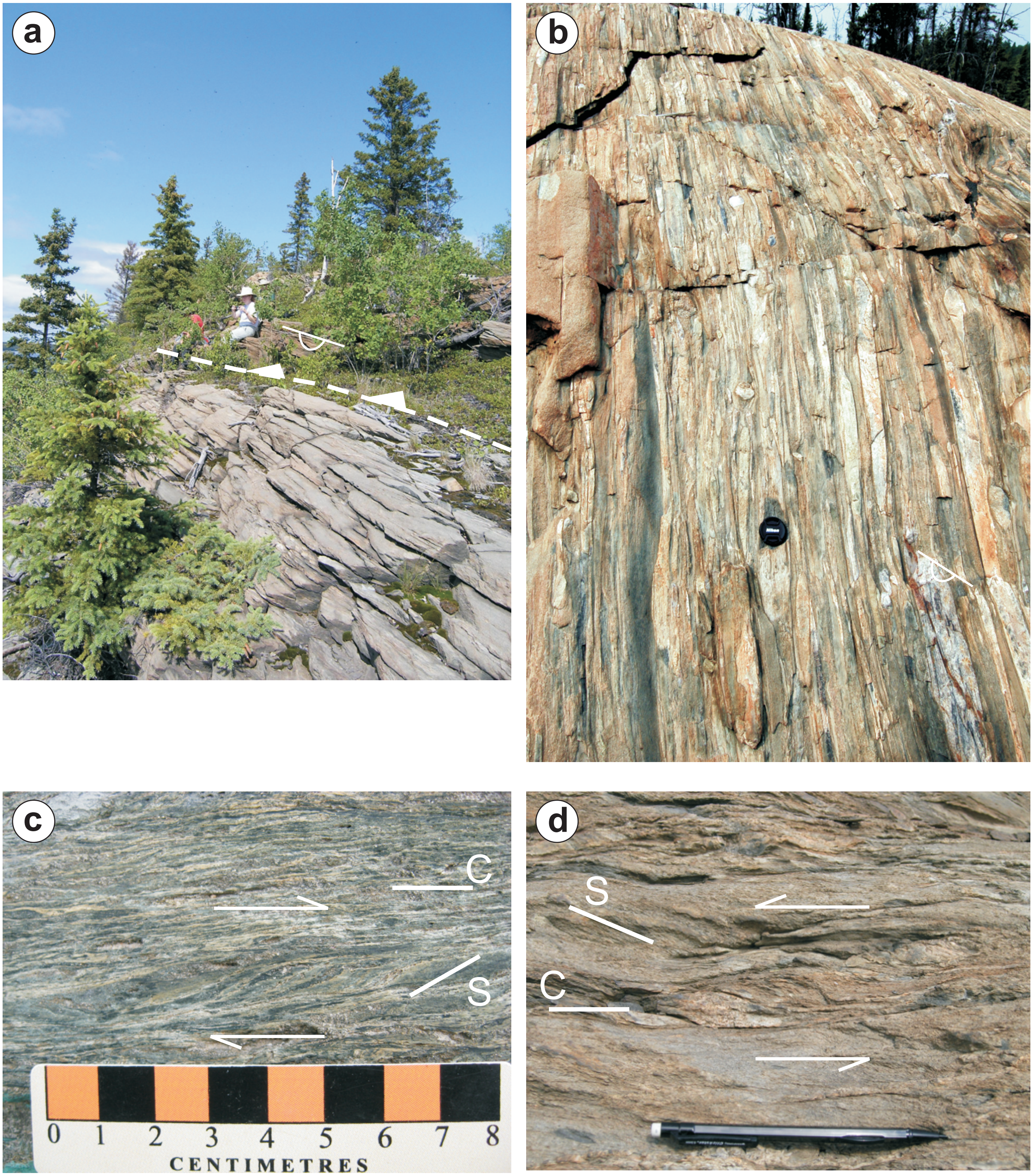


Figure 5: *Representative structures in the Flin Flon area:* ***a)*** *shallow-dipping, west-verging Cliff Lake Fault thrusting overturned mafic volcanic rocks of the Western Hook Lake succession (top) onto overturned Missi Group sedimentary rocks (bottom);* ***b****) stretching lineation defined by elongate pebbles in the hinge of the Flin Flon Creek syncline;* ***c)*** *late dextral C-S fabric in Cliff Lake Fault by Highway 10; and* ***d****) late sinistral C-S fabric in Cliff Lake Fault by Cliff Lake.*

near-parallelism with the lineation, deforming the F_3 Pipeline syncline into a kilometre-scale sheath fold.

The first penetrative regional cleavage (S_5) across the Flin Flon mining camp formed during D_5. The S_5 cleavage is expressed by the flattening of pebbles and cobbles in Missi Group conglomerate, as a disjunctive to continuous, chloritic and sericitic cleavage in sandstone, and as a continuous chloritic foliation that wraps around flattened pillows and volcanic fragments in the underlying volcanic rocks. The cleavage (S_5) strikes northwest to north-northwest (320–350°) and dips 50–80° to the northeast. It overprints the regional F_3 and F_4 folds in the Missi Group cover rocks and is associated with dextral shear reactivation of D_3 thrust faults and lithological contacts along the northwest-striking limbs and near the hinge of the Hidden Lake syncline (Figure 5c).

A second regional cleavage (S_6) formed during D_6 during terminal collision of the Sask craton and Flin Flon–Glennie complex with the Superior craton at ca. 1.83–1.79 Ga. This north-northeast-striking cleavage formed as a result of compression from the east-southeast and west-northwest, which caused dextral reactivation of east-striking D_4 thrust faults, and sinistral reactivation of northwest-striking lithological contacts and D_3 thrust faults, such as the Cliff Lake and Flin Flon Lake faults (Figure 5d).

The last deformation event D_7 involved the formation of subvertical faults, such as the Ross Lake Fault and maybe the Channing and Insole Lake faults, with oblique-slip displacements.

Terminology

The stratigraphic nomenclature used on the map and throughout this report is a combination of historical and new nomenclature. Where historical names existed, and were deemed appropriate following the most recent work done in the area and according to the North American Stratigraphic Code, these names were preserved (e.g., Flin Flon formation, Hidden formation; see Table 1, back pocket, for details). New/modified nomenclature was introduced to reflect the recent work done in the area and to ensure stratigraphic coherency throughout the map (e.g., merging of multiple 'formations' into one, new members added to various formations). In areas where recent work did not provide enough information to use 'formal' stratigraphic nomenclature, and there were no historical names in current usage, the term 'succession' was used to refer to the recognized lithostratigraphic unit (e.g., Western Hook Lake succession; see Table 1, back pocket, for details). In areas where no new work was performed in the course of this project, the latest published informal names were retained as is (e.g., Hamell Lake volcanic rocks).

The legend of the map presents rock packages from youngest at the top to oldest at the bottom, based on the known or inferred ages. Labels are composed of one or two upper case letter(s), generally derived from the overarching nomenclature in use for these rock packages (e.g., F, Flin Flon formation; HL, Hamell Lake volcanic rocks; CL, Cliff Lake plutonic suite) further refined, if needed, by a lower case letter(s) derived from further subdivisions of these packages (e.g., Fm, 'Millrock member' of the Flin Flon formation). Polygons on the map reflect units and subunits of these rock packages. A 'unit' is composed of individual or multiple lithofacies or rock types and is identified by adding a single digit number to the alphanumeric label (e.g., unit Fm4, 'Rhyolite flows, massive, locally flow banded' of the Millrock member of the Flin Flon formation). A 'subunit' further characterizes a unit, if needed, and is identified by adding a lower case letter at the end of the alphanumeric label (e.g., subunit Fm4b, 'quartz- and plagioclase-phyric' rhyolite flows of the Millrock member of the Flin Flon formation). Units and subunits are systematically presented from coherent to clastic lithofacies/rock types, and from mafic to felsic lithofacies/rock types. Hence, this presentation of the units and subunits does not necessarily reflect the internal stratigraphic order of any given rock package. In this report, units and subunits of each rock package are described either in stratigraphic order or from the most to the least abundant.

The level of detail in the linework on the map reflects the scale of mapping for each area, and is therefore not even throughout the map. Solid lines are used for contacts/facies/faults where the location of the feature is defined (observed or tightly constrained). Dashed lines are used for contacts/facies/faults where the location of the feature is approximate (loosely constrained). Dotted lines are used for contacts/facies/faults where the location of the feature is assumed. Facies (or lithofacies) contacts are only represented in areas where the mapping scale allowed and/or unit/subunit complexity warranted it.

In the following lithological descriptions, the nongenetic terminology of Fisher (1966) is used for volcaniclastic rocks strictly to distinguish clast size and the percentage of clasts versus ash-sized matrix, and do not imply a mechanism of formation through pyroclastic processes. Refer to Figure 6 or the lower portion of the map in the back pocket for Fisher (1966) volcaniclastic rocks grain-size classification diagram.

The term 'peperite' is used with a genetic meaning according to the definition of White et al. (2000): "peperite (n), a genetic term applied to a rock formed essentially in situ by disintegration of magma intruding and mingling with unconsolidated or poorly consolidated, typically wet sediments".

The term 'cryptoflow' is used to describe flows that were erupted just below the seafloor. It describes a massive or pillowed flow with peperite along the upper and lower contacts indicating emplacement into wet, unconsolidated volcaniclastic deposits (DeWolfe et al., 2009a).

The term 'included tuff' is used to describe inclusions of tuff observed within massive coherent flows, or between pillows (but not within individual pillows).

The term 'epidote-quartz alteration' refers to round or amoeboid patches consisting of a granular mosaic of epidote and quartz with minor actinolite in mafic volcanic rocks. Epidote-quartz alteration is attributed to relatively high-temperature (>300°C, <400°C) evolved seawater-rock interaction, and defines semiconformable alteration zones in some VMS districts (Galley, 1993; Gibson and Kerr, 1993).

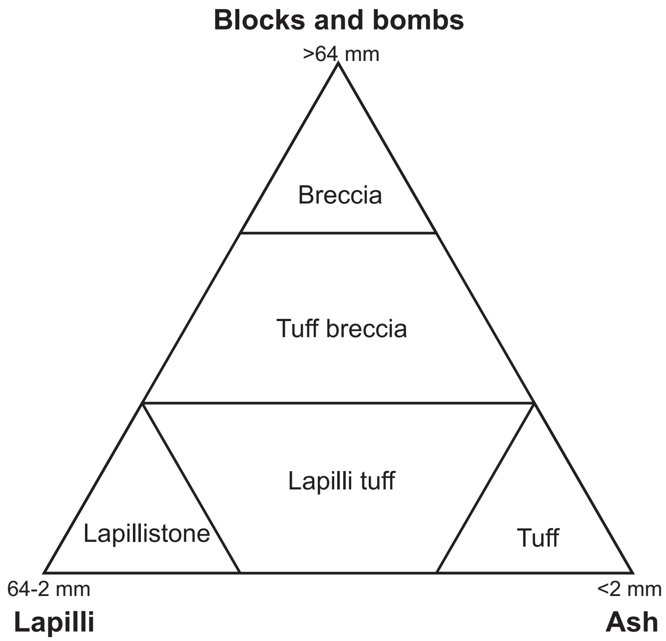


Figure 6: *Volcaniclastic rocks grain-size classification diagram by Fisher (1966).*

Geology of the map area

The map area is underlain by two main components: 1) >1.88 Ga juvenile oceanic-arc basalt, basaltic andesite, rhyolite, and synvolcanic intrusions of the Flin Flon arc assemblage (formerly Amisk Group, Syme and Bailes, 1993; see details below); and 2) <1.88 Ga calcalkaline 'early successor-arc' plutons (see 'Pre–Missi Group intrusive rocks' below), fluvial-alluvial sedimentary rocks of the Missi Group (see details below), and 'late successor-arc' intrusions (Bruce, 1918; Bailes and Syme, 1989; Stauffer, 1990; Stern et al., 1999; see 'Post–Missi Group intrusive rocks' below).

The structural architecture of the Flin Flon area and the geometry of the deposits have been shaped by multiple generations of folds, thrusts and strike-slip faults (see 'Structural history' above for details). This resulted in the map area being broken into several distinct stratigraphic and structural panels along major north-northwest-striking early thrust faults and late strike-slip faults (Figure 3).

Flin Flon arc assemblage (1.88–1.92 Ga)

The Flin Flon arc assemblage, which includes all the volcanic rocks in the area, extends in all directions beyond the boundaries of the map, to the Meridian Shear Zone to the west, the Mystic Creek Shear Zone to the east, and to the Phanerozoic cover rocks to the south (Figure 1).

The Flin Flon arc assemblage has been dismembered by structural complexities (folding and faulting). However, coherent portions of stratigraphy have been recognized across some of these structures (e.g., Flin Flon Lake and Ross Lake faults, Figure 7), whereas in other cases, the lack of data or major differences highlighted through mapping and more recent geochronological and geochemical work preclude stratigraphic

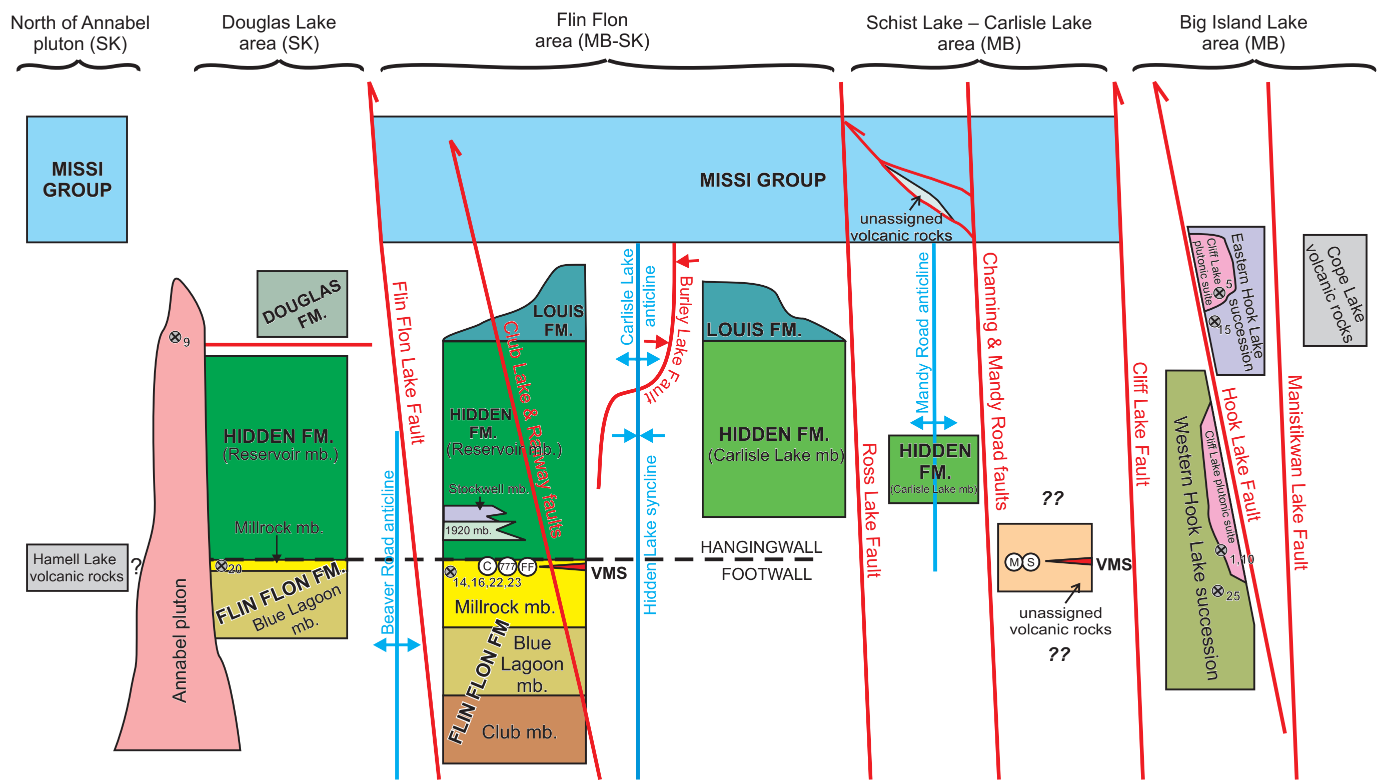


Figure 7: *Schematic representation of the stratigraphy and structure of the Flin Flon area (Devine, 2003; Simard, 2006a; MacLachlan and Devine, 2007; Simard and Creaser, 2007; DeWolfe, 2008; DeWolfe et al., 2009a; Gibson et al., 2009; Simard et al., 2009, 2010). Red lines represent faults, blue lines represent folds, arrows represent facing directions on each side of a given structure. Symbols/abbreviations: MB, Manitoba; SK, Saskatchewan; VMS, volcanogenic massive sulphide; VMS deposits (C, Callinan; FF, Flin Flon; M, Mandy; S, Schist Lake); X, geochronology samples listed in the table on the lower portion of the map (Geoscientific Map MAP2010-1, back pocket, Simard et al., 2010).*

correlation across faults (e.g., Channing–Mandy Road fault, Hook Lake Fault, Figure 7; see details below).

The geochemical signatures of the different volcanic rock packages of the Flin Flon arc assemblage in the Flin Flon area are all very similar (Figures 8, 9). Mafic flows in the area display low Nb/Y and Zr/TiO_2 ratios (Figure 8) as well as small to moderate light rare earth element (LREE) enrichment with relatively flat heavy rare earth element (HREE) patterns on the rare earth element (REE) and mantle-normalized trace-element diagrams (Figure 9). Pronounced negative Nb and Ti anomalies on mantle-normalized trace-element diagrams (Figure 9), low Ti/V ratios (not shown) and distribution on the Th-(Zr/117)-Nb/16 diagram (Wood, 1980, not shown) indicate that those mafic flows are the products of subduction-related processes in an oceanic-arc environment. Their relatively low $(La/Yb)_N$ ratios (Figure 9) and high FeO_{total}/MgO ratios (not shown) suggest a tholeiitic affinity (Syme et al., 1999b; Devine, 2003; Bailey, 2006; Simard and Creaser, 2007; DeWolfe et al., 2009b). Felsic volcanic rocks in the Flin Flon area display low Nb/Y values that suggest subalkaline affinity (Figure 8a). The content of high-field-strength elements (Nb, Ta, Ga, Zr, Hf, Y), relative to other incompatible elements within these rocks, are moderate to low and characteristic of rocks generated in an oceanic arc environment (Syme, 1998; Syme et al., 1999b; Simard and Creaser, 2007). The REE patterns of the felsic flows show slight LREE enrichment with a relatively flat HREE pattern (Figure 9). Similar to the mafic flows from the area, felsic flows of the Flin Flon area display pronounced negative Nb and Ti anomalies on the mantle-normalized trace-element diagrams (Figure 9). This, plus their Ti/V ratio (not shown) and their distribution on the Th-(Zr/117)-Nb/16 diagram (Wood, 1980, not shown) also suggest subduction-related genesis in an oceanic-arc environment (Syme, 1998; Syme et al., 1999b; Simard and Creaser, 2007).

Overall, all volcanic rocks in the area are the result of tholeiitic arc volcanism (Syme et al., 1999b; Devine, 2003; Bailey, 2006; Simard and Creaser, 2007; DeWolfe et al., 2009b) with relatively juvenile isotopic signatures (i.e., positive Nd values, ranging from +2 to +5; Stern et al., 1995a, b; Simard and Creaser, 2007). Although small geochemical differences do exist between formations/successions, the considerable overlap in geochemical signatures usually prevents differentiating most of them based solely on geochemistry (e.g., Hidden versus Louis formation, Figures 8, 9). One exception to this rule are the felsic rocks from the Eastern Hook Lake succession, which display a significantly distinct geochemical signature (higher Zr and LREE enrichment) from the other felsic rocks in the Flin Flon area (Figures 8, 9). The scarcity of analyses for 'unassigned volcanic rocks' (U) hosting the Schist Lake and Mandy VMS deposits prevents any conclusive comparison to the volcanic rocks hosting the Flin Flon–Callinan–777 VMS deposits to the west or the Western and Eastern Hook Lake successions to the east at this stage (Figures 8, 9).

West of the Channing–Mandy Road faults

Although rocks of the Flin Flon arc assemblage are regionally extensive, they are best exposed around the town of Flin Flon, west of the Channing and Mandy Road faults, where they include one of the most prolific volcanogenic massive sulphide (VMS)-hosting sequences in the Flin Flon Belt.

West of the Channing and Mandy Road faults, the Flin Flon arc assemblage is divisible into four formations, Flin Flon (F), Hidden (H), Louis (L) and Douglas (G) formations, along with the Hamell Lake volcanic rocks (HL) north of the Annabel pluton and abundant synvolcanic intrusive rocks (D; see Table 1, back pocket, for previously used nomenclature; Figure 3).

The lithofacies, stratigraphy and geochemistry of volcanic strata hosting the Flin Flon–Callinan–777 VMS deposits record extension, subsidence and infilling of a complex subsidence structure or cauldron and the development of an intense hydrothermal system (Flin Flon formation; Bailes and Syme, 1989; Syme et al., 1999a; Devine, 2003), which terminated with a hiatus in active volcanism and the formation of the VMS deposits. Following VMS deposition, there was resurgence in volcanism and subsidence marked by the development of one or more mafic shield volcanoes that buried this subsidence structure (Hidden and Louis formations; Syme et al., 1999a; DeWolfe and Gibson, 2004, 2005, 2006; DeWolfe, 2008; DeWolfe et al., 2009a).

This most recent definition of the VMS-hosting stratigraphy is based on the geology of the east-younging and east-dipping succession east of the Flin Flon Lake Fault (Ames et al., 2002, 2003; Devine et al., 2002a, b; Devine, 2003; Gibson et al., 2003b; DeWolfe, 2008; DeWolfe et al., 2009a). However, recent detailed mapping west of the Flin Flon Lake Fault suggests that, despite some differences in the abundance and distribution of rock types, the west-facing and west-dipping succession west of the Flin Flon Lake Fault and the Beaver Road anticline is a time stratigraphic equivalent of the mine succession (MacLachlan and Devine, 2007). Thus, the stratigraphic terminology for the Flin Flon–Callinan–777 mine succession has been extended to strata west of the Flin Flon Lake Fault.

Flin Flon formation (F)

(H.L. Gibson, C.A. Devine and K. MacLachlan)

The Flin Flon formation (F) is composed of heterolithic and monolithic volcaniclastic rocks, rhyolite flows, domes and associated volcaniclastic rocks, and massive to pillowed basalt flows and flow breccias (Figure 10). It records successive periods of extension, subsidence and infilling of the volcanic subsidence structure or cauldron that contains the Flin Flon, Callinan and 777 VMS deposits (e.g., Bailes and Syme, 1989; Syme and Bailes, 1993; Devine, 2003; Figure 11).

The Flin Flon formation is exposed on both sides of the Flin Flon Lake Fault (Figure 12). It can be subdivided in three mappable members from the oldest to the youngest: the Club member (Fc), the Blue Lagoon member (Fb) and the Millrock member (Fm; Devine et al., 2002a).

Club member (Fc)

The Club member (Fc) is only exposed north of the tailing ponds on the eastern side of the Flin Flon Lake Fault where is occurs as a shallowly dipping, south-southeast-facing

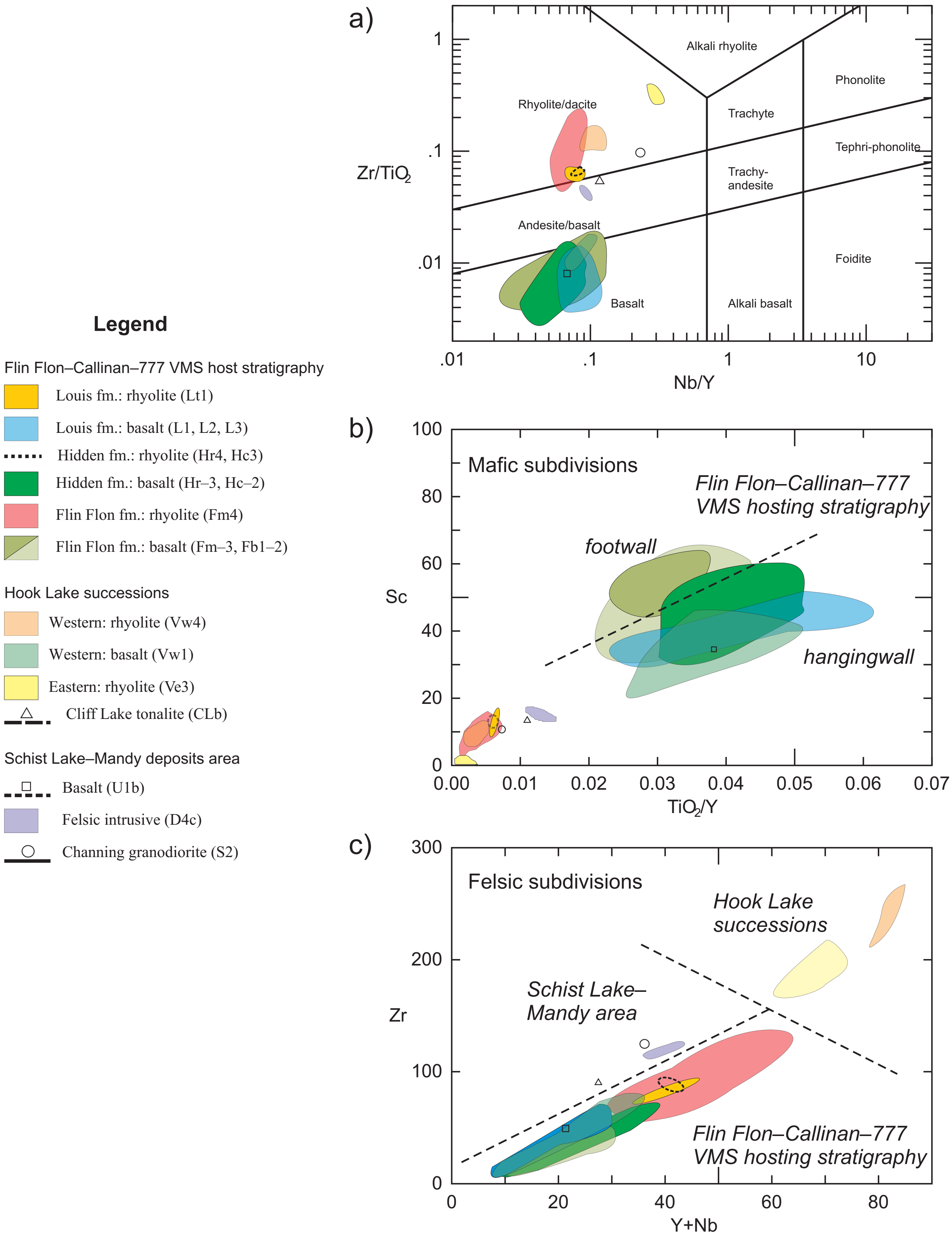


Figure 8: *Geochemical characteristics for volcanic and plutonic rocks of the Flin Flon area:* ***a)*** *Zr/TiO_2 vs. Nb/Y classification diagram (modified from Winchester and Floyd, 1977);* ***b)*** *Sc vs. TiO_2/Y diagram showing overlaps and differences between the various rock packages in the Flin Flon area; mafic rocks of the footwall rocks of the Flin Flon–Callinan–777 deposits generally show higher Sc values than those of the hangingwall rocks; and* ***c)*** *Zr vs. Y+Nb diagram showing overlaps and differences between the various rock packages in the Flin Flon area; felsic rocks of the Hook Lake successions are significantly more enriched in Zr, Y and Nb than other felsic volcanic rocks in the Flin Flon area. Selected data: Flin Flon formation, H.L. Gibson and K. MacLachlan (unpublished data 2001–2009); Hidden and Louis formations, K. MacLachlan (unpublished data 2001-2007), Simard and Creaser (2007), DeWolfe (2008), DeWolfe et al., (2009b); Hook Lake successions, R-L. Simard and P.D. Kremer (unpublished data 2007).*

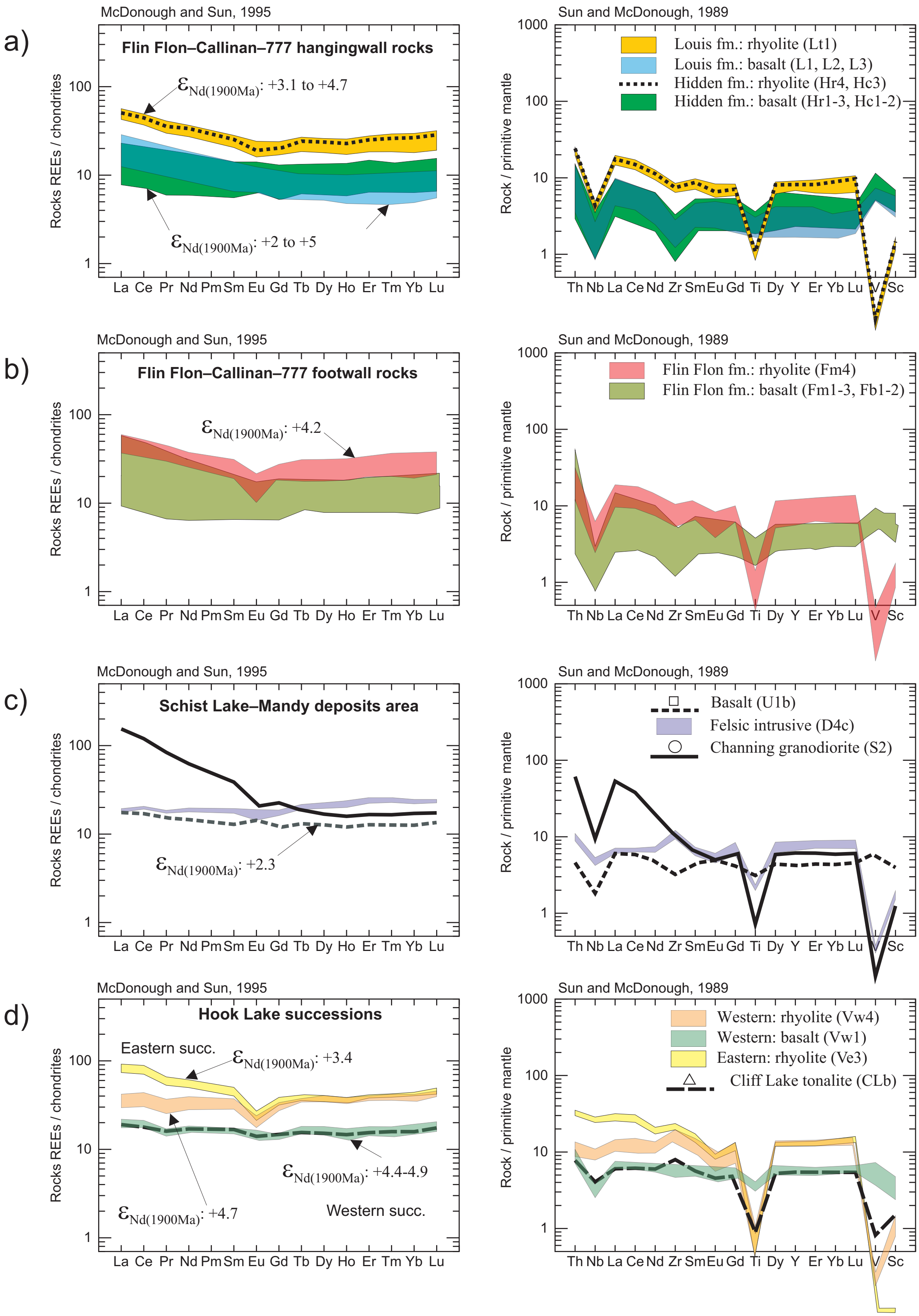


Figure 9: *Chondrite-normalized rare earth element patterns (left, normalization values from McDonough and Sun, 1995) and mantle-normalized incompatible trace-element patterns (right, normalization values from Sun and McDonough, 1989) for volcanic and plutonic rocks of the Flin Flon area:* ***a)*** *hangingwall rocks of the Flin Flon–Callinan–777 VMS deposits;* ***b)*** *footwall rocks of the Flin Flon–Callinan–777 VMS deposits;* ***c)*** *hostrocks of the Schist Lake and Mandy VMS deposits; and* ***d)*** *rocks of the Western and Eastern Hook Lake successions. Selected data: Flin Flon formation, H.L. Gibson and K. MacLachlan (unpublished data 2001–2009); Hidden and Louis formations, K. MacLachlan (unpublished data, 2001–2007), Simard and Creaser (2007), DeWolfe (2008), DeWolfe et al., (2009b); Hook Lake successions, R-L. Simard and P.D. Kremer (unpublished data 2007).*

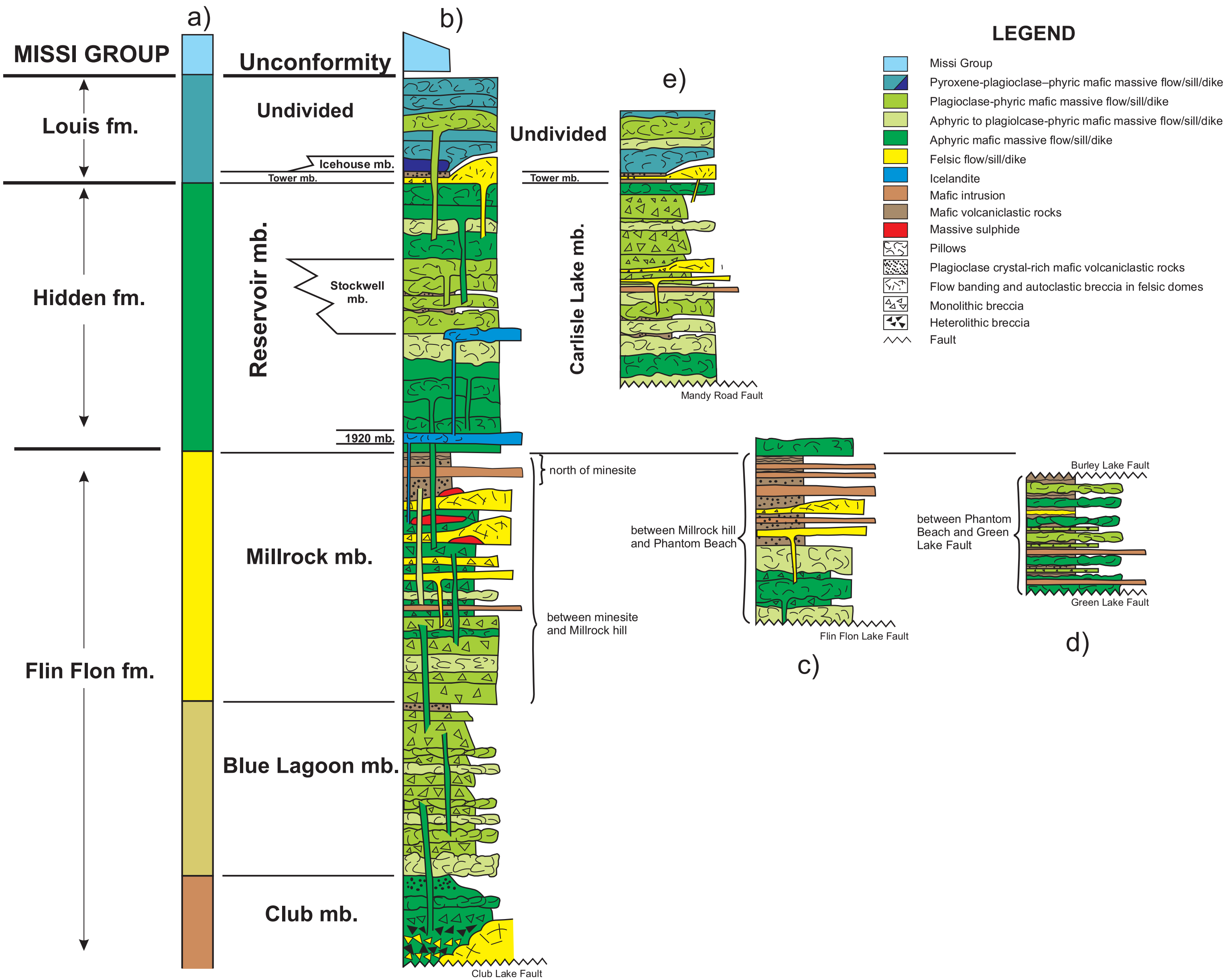


Figure 10: *Idealized stratigraphic column showing generalized lithofacies through the footwall and hangingwall to the Flin Flon–Callinan–777 volcanogenic sulphide deposits east of the Flin Flon Lake Fault (modified from Ames et al., 2002; Devine et al., 2002a; Devine, 2003; Gibson et al., 2003b; DeWolfe and Gibson, 2004, 2005; Simard, 2006a; DeWolfe et al., 2009a; DeWolfe, 2010, 2011):* ***a)*** *stratigraphic column of the formations in the area;* ***b)*** *Flin Flon, Hidden and Louis formations east of the Flin Flon Lake Fault and on the western limb of the Hidden Lake syncline;* ***c)*** *Millrock member between Millrock hill and Phantom Beach;* ***d)*** *Millrock member between Phantom Beach and the Green Lake Fault; and* ***e)*** *Hidden and Louis formations east of the Burley Lake Fault.*

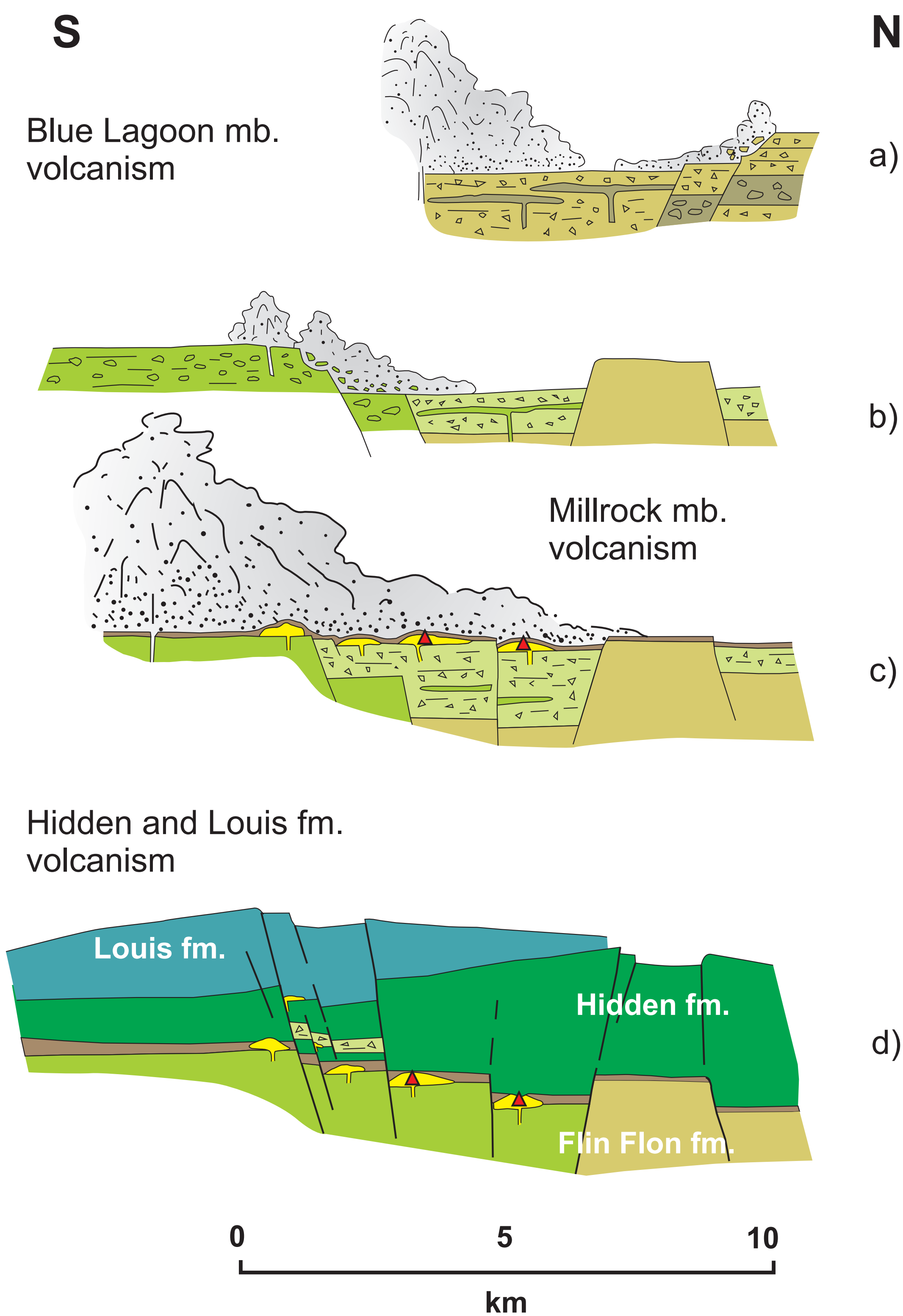


Figure 11: *Sequence of interpreted volcanic, subsidence and VMS ore-forming events (modified from Gibson et al., 2009):* ***a)*** *localized subsidence accompanied Blue Lagoon effusive and pyroclastic eruptions; piece-meal collapse into nested basins, redeposition of volcaniclastic lithofacies (megabreccias);* ***b)*** *initial effusive eruptions of the Millrock member constructed small basaltic lava shield on the margin of the cauldron; piece-meal collapse along margin of the cauldron triggered collapse of the lava shield to produce thick megabreccia deposits; collapse was accompanied by pyroclastic and effusive eruption, within the Flin Flon cauldron;* ***c)*** *eruption rhyolite localized to nested faults basins within the cauldron led to the construction of rhyolite flow-dome complexes; high-temperature hydrothermal discharge was localized in time and space to rhyolite dome construction to form the Flin Flon, Callinan and 777 deposits; ore formation was followed by voluminous basaltic pyroclastic eruptions that were centred along the southern margin, but tephra blanketed the entire cauldron and the margin; and* ***d)*** *dominantly effusive eruption of basalt (lesser andesite) constructed the Hidden and Louis formation lava shields that buried the Flin Flon cauldron and VMS deposits; lava shield construction was accompanied by localized and minor subsidence.*

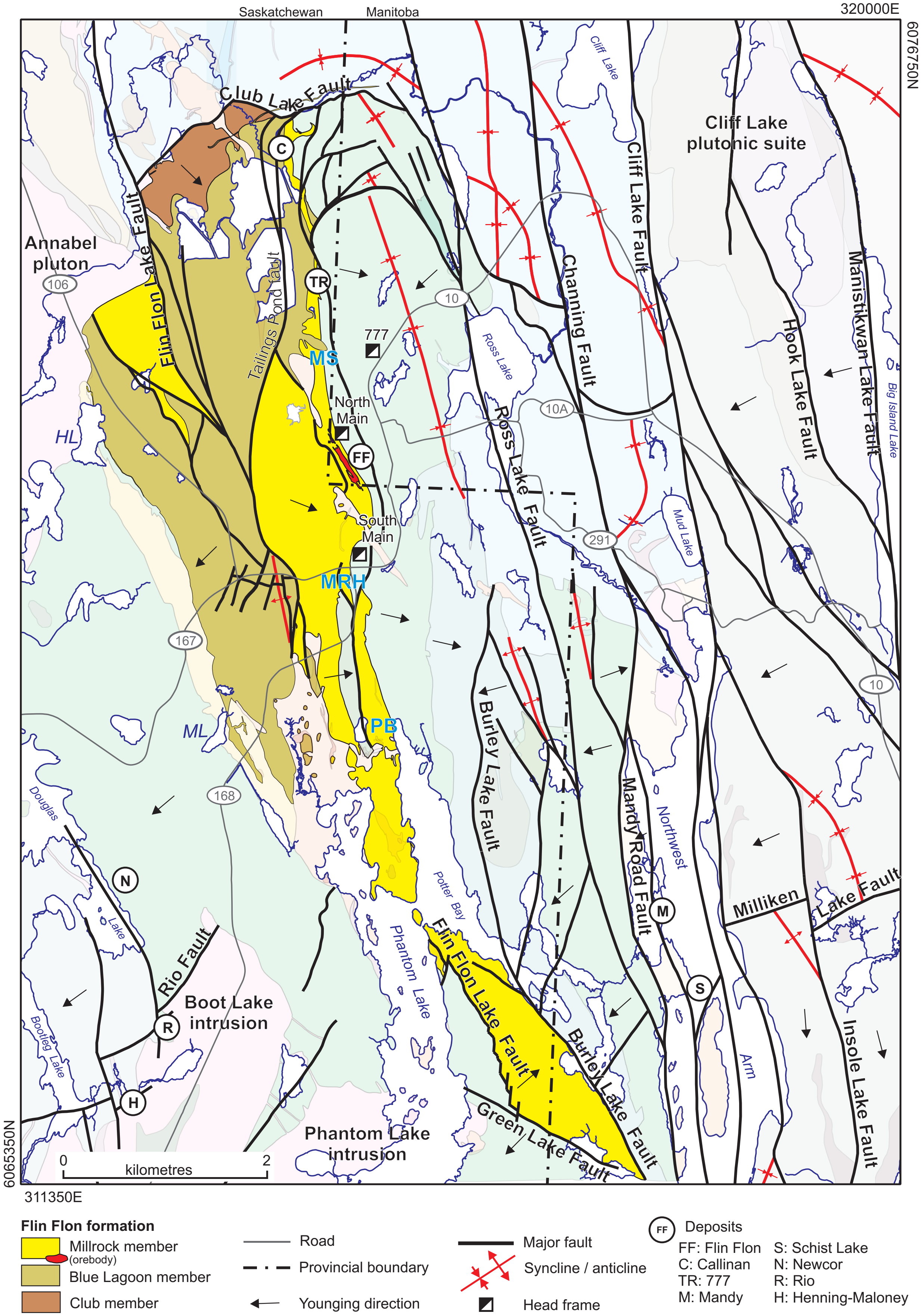


Figure 12: *Simplified geology of the Flin Flon area, highlighting only the three mappable members of the Flin Flon formation and major structures in the area (Abbreviations: MS, minesite; MRH, Millrock hill; PB, Phantom Beach; HL, Hilary Lake; ML, Myo Lake). For complete legend refer to Figure 3.*

succession ~500 m thick (apparent thickness; Figure 12). The Club member is truncated at its base by the Club Lake Fault to the north and it is conformably overlain by the Blue Lagoon member to the south (Figures 10, 12). In the following section, the units and subunits of the Club member will be described in relative stratigraphic order.

The Club member consists of massive, aphyric coherent rhyolite (unit Fc2), monolithic felsic breccia, and rhyolite-clast–bearing mafic volcaniclastic rocks and bedded tuff (unit Fc3; Figure 13a) intercalated with lesser aphyric mafic flows (unit Fc1). The coherent rhyolites (unit Fc2) are often associated with rhyolite-clast–bearing volcaniclastic rocks, which are interpreted as part of rhyolite flow or dome complexes with their in situ and flank breccias (Devine, 2003). The Club member felsic volcanic rocks were emplaced, along with minor mafic effusive volcanic rocks, within localized volcanic basins. The associated heterolithic mafic volcaniclastic rocks (subunit Fc3a) are interpreted to have been derived from the proximal resedimentation of pre-existing mafic and, to a lesser extent, felsic volcanic rocks and their localized accumulation within basin(s) that developed during subsidence. The bedded mafic tuff may be a product of suspension sedimentation from concomitant mafic pyroclastic eruptions that accompanied this effusive mafic and felsic volcanism.

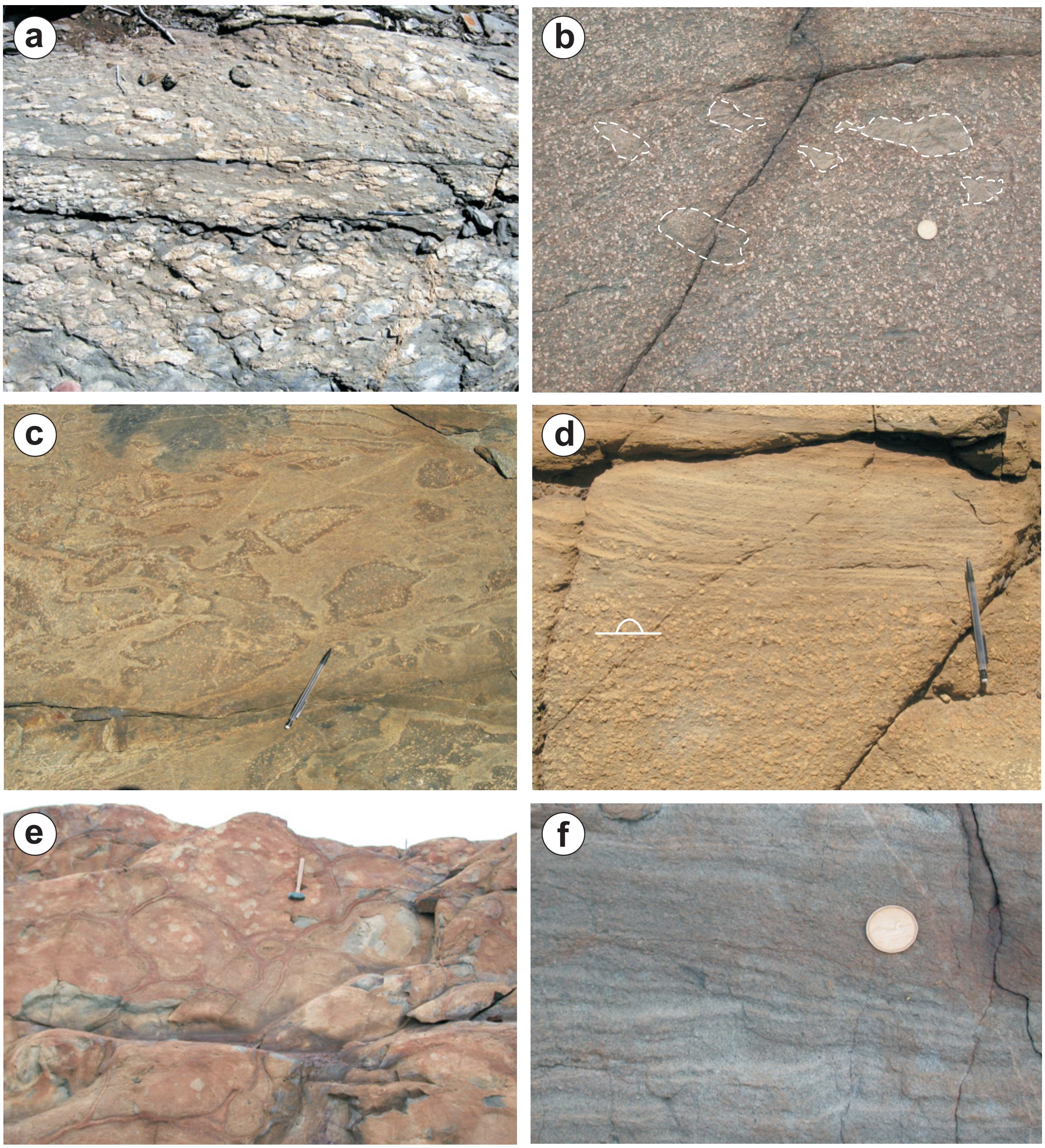


Figure 13: *Representative rock types, facies and textures of the Flin Flon formation:* ***a)*** *heterolithic felsic breccia of the Club member (subunit Fc3a);* ***b)*** *plagioclase-crystal–rich lapillistone of the Blue Lagoon member (subunit Fb3b);* ***c)*** *strongly plagioclase-phyric amoeboid pillow breccia of the Blue Lagoon member (subunit Fb2b);* ***d)*** *plagioclase-crystal–rich tuff at the top of the Blue Lagoon member (subunit Fb3c);* ***e)*** *plagioclase-phyric pillowed basalt flow of the Blue Lagoon member (subunit Fb2b); and* ***f)*** *plane-bedded and crossbedded tuff of the Millrock member (subunit Fm5e).*

Blue Lagoon member (Fb)

The Blue Lagoon member (Fb) is exposed on both sides of the Flin Flon Lake Fault (Figure 12). East of the Flin Flon Lake Fault it is best exposed around the tailings pond where it constitutes a shallowly dipping, south-southeast-facing succession from 50 to >500 m thick (apparent thickness; Figure 12). West of the Flin Flon Lake Fault, where it extends towards Hilary and Myo lakes, it constitutes a steeply dipping, south-southwest-facing succession >500 m thick (apparent thickness; Figure 12; see details below). In both areas, thrust fault repetitions have thickened parts of the Blue Lagoon member stratigraphy (Figure 12).

The Blue Lagoon member forms a distinctive succession of mafic, plagioclase-crystal–rich volcaniclastic units with up to 40% plagioclase crystals (unit Fb3; Figure 13b) and aphyric to plagioclase-phyric flows with phenocrysts up to 1 cm in size (units Fb1, Fb2; Figure 10). In the volcaniclastic units, the plagioclase crystals occur within both the clasts and the matrix. Although most of the volcaniclastic units and flows have a plagioclase-rich matrix or groundmass, some intercalated beds, clasts or flows do not contain plagioclase crystals or phenocrysts. Characteristic volcaniclastic units, such as the heterolithic breccia (subunit Fb3a), and the lapilli tuff and tuff breccia (subunit Fb3b) are strongly heterolithic, and include clasts of plagioclase-phyric basalt, aphyric basalt, scoriaceous basalt, and minor (<1%) aphyric rhyolite blocks.

East of the Flin Flon Lake Fault and north of the tailings pond, the Blue Lagoon member lies conformably atop the Club member and is conformably to unconformably overlain by rocks of the Millrock member (Figure 12). The base of the Blue Lagoon member is characterized by the intercalation of plagioclase-phyric flows (unit Fb2; Figure 13c) with the plagioclase-crystal–rich volcaniclastic units (subunits Fb3b, c), which suggests the latter may be transported autoclastic deposits of the former (Devine, 2003). The upper portion of the Blue Lagoon member is characterized by thick, heterolithic plagioclase-crystal–rich volcaniclastic rocks (subunits F3b, c; Figure 13d) and localized flows (units Fb1, Fb2). The volcaniclastic rocks are interpreted as primary pyroclastic deposits, their syneruptive redeposited equivalents and/or subsidence induced megabreccias, deposited by mass flows into rapidly subsiding, fault-controlled basins (Devine, 2003; Gibson et al., 2005; Figure 11a). The Blue Lagoon member is characterized by abrupt changes in the thickness and limited lateral continuity of units indicating the presence of several distinct synvolcanic fault blocks. The dominance of coarse volcaniclastic rocks versus flows within the Blue Lagoon member east of the Flin Flon Lake Fault is interpreted to be a product of extensive and long-lived, subsidence-triggered, mass-flow sedimentation (Devine, 2003).

West of the Flin Flon Lake Fault, the Blue Lagoon member is characterized by localized thick accumulation of coarse, plagioclase-crystal–rich heterolithic mafic volcaniclastic rocks (subunits Fb3a, b), megabreccia (subunit Fb3e), and tuffaceous beds (subunit Fb3c) interbedded with aphyric (unit Fb1) and plagioclase-phyric (unit Fb2; Figure 13e) mafic flows (Figure 10). These volcaniclastic-dominated successions occur within localized 'sub-basins' within successions dominated by mafic flows (units Fb1, Fb2). Along the margin of one of these inferred fault-bounded sub-basins is an asymmetric mound of silicified amoeboid breccia (subunit Fb3d) that is interpreted to represent spatter breccia from a fissure eruption localized along the bounding synvolcanic fault (MacLachlan and Devine, 2007). The volcaniclastic dominated successions are interpreted to define small, isolated fault basins within a larger subsidence structure (MacLachlan and Devine, 2007; Figure 11a). The proportion of volcaniclastic material to aphyric and plagioclase-phyric flows is much lower than on the western side of the Flin Flon Lake Fault suggesting less subsidence-induced mass flow sedimentation (Gibson, 2006).

Millrock member (Fm)

The Millrock member is exposed on both sides of the Flin Flon Lake Fault, although it is best developed on the eastern side, and so far, is only known to contain VMS deposits on the eastern side (Figure 12).

East of the Flin Flon Lake Fault, the Millrock member can be traced along strike for >5 km where it ranges in thickness from <10 m in the north to >500 m in the south (Figure 12). Its lithofacies are discontinuous along strike and, as such, define synvolcanic basins that range from tens to several hundreds of metres in width (Gibson, 2006; Gibson et al., 2009; Figures 10, 11b, c).

Because of this lateral variability in lithofacies thickness and extent, the Millrock member stratigraphy will be described in four main sections: 1) north of the minesite (MS); 2) between the minesite and Millrock hill (MRH); 3) south of Millrock hill to Phantom Beach (PB); and 4) between Phantom Beach and the Green Lake Fault (Figures 10, 12). Units and subunits are described in relative stratigraphic order.

North of the minesite, the Millrock member constitutes a steeply dipping, northeast-facing succession that is 10–150 m thick (apparent thickness; Figures 10, 12). It is composed of a 5–40 m thick, laterally continuous, well-bedded mafic tuff (subunit Fm5e; Figure 13f) that is either underlain by massive to crudely bedded heterolithic mafic breccia (subunit Fm5b) that ranges in thickness up to 150 m, or by Blue Lagoon volcaniclastic rocks. The base of the heterolithic mafic breccia (subunit Fm5b) occupies scours within the underlying crystal-rich beds of the Blue Lagoon member and locally contains fragments and plagioclase crystals derived from the underlying crystal-rich beds. The limited lateral extent of this heterolithic mafic volcaniclastic unit is consistent with emplacement into a distinct fault-bounded basin (Devine, 2003; Figure 11b, c). The lack of internal structure within the coarser volcaniclastic rocks suggests emplacement by mass flows into a rapidly subsiding structurally controlled basin (Devine, 2003). The well-bedded mafic tuff unit (subunit Fm5e) is dominantly composed of ash-sized or minor lapilli-sized clasts, plagioclase crystals, and some beds contain accretionary lapilli. The tuff is typically well bedded and often displays soft sedimentary deformation structures (Figure 14a; Devine, 2003). The lateral continuity of bedded tuff (along the entire strike length of the Millrock member), dominance of plane-parallel beds, and occurrence of accretionary lapilli (Figure 14b) support deposition by suspension sedimentation following mass flow events and/or

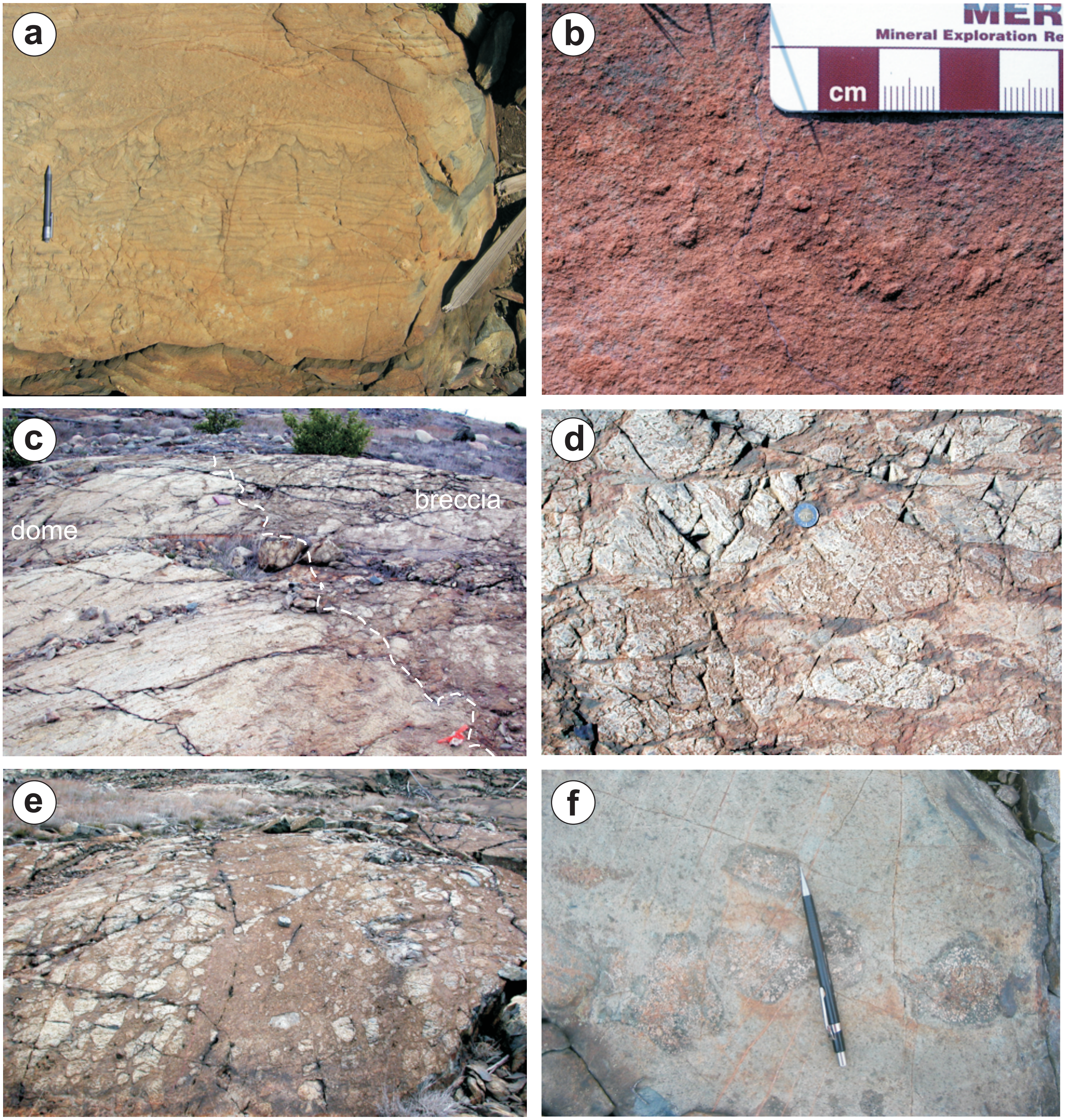


***Figure 14:** Representative rock types, facies and textures of the Millrock member of the Flin Flon formation: **a)** flames and load structures in mafic tuff at the top of the Millrock member (subunit Fm5e); **b)** accretionary lapilli (round clasts)in plane-bedded tuff in the upper portion of the Millrock member; **c)** rhyolite dome and its autoclastic breccia carapace on Millrock hill, Millrock member (subunit Fm4b); **d)** rhyolitic tuff breccia on Millrock hill, Millrock member (subunit Fm7b); **e)** heterolithic tuff breccia and lapillistone on Millrock hill (subunit Fm6a); and **f)** heterolithic mafic tuff breccia with abundant altered scoriaceous clasts of the Millrock member (subunit Fm6c).*

from subaqueous eruption columns that at times broke through to surface (Fisher and Schmincke, 1984; Cas and Wright, 1987; McPhie et al., 1993; Gibson et al., 1999; Figure 11b, c). Local crossbedding and scour structures are interpreted to result from bottom current reworking (Gibson, 2006; Gibson et al., 2009). The bedded tuff unit represents a significant hiatus in effusive volcanism that was punctuated by the emplacement of rhyolite domes and the formation of the VMS deposits farther to the south (see details below; Devine, 2003; Figure 11c).

Between the minesite and Millrock hill, the Millrock member is a moderately dipping east- to northeast-facing succession with a total apparent thickness, including thrust repetitions, of >1000 m (Figure 12). The base and bulk of the Millrock member is characterized by a thick succession of megabreccia and other coarse mafic volcaniclastic units (subunits Fm5a, f, g) locally intercalated with aphyric to sparsely plagioclase-phyric and lesser plagioclase-phyric (15–20% plagioclase) mafic flows (units Fm1, Fm3). The mafic volcaniclastic rocks, mainly tuff breccia and breccia (subunits Fm5a, f), are characteristically monolithic and composed of aphyric, scoriaceous, amoeboid- to fluidal-shaped basalt fragments, and more rounded, equant basalt fragments. The megabreccia (subunit Fm5g) consists mainly of angular to subangular blocks, up to tens of metres in size, of aphanitic, amygdaloidal, massive and pillowed basalt, with minor amoeboid and scoriaceous clasts in a finer-grained, tuff to lapilli-tuff matrix. Volcaniclastic units composed entirely of, or dominated by, fluidal, amoeboid and scoriaceous clasts have been interpreted as products of fire-fountain and/or Strombolian eruptions (Gibson, 2006; Gibson et al., 2009). Thick, poorly sorted and bedded deposits of megabreccia, previously referred to as the Creighton member of the Flin Flon formation (see Table 1, back pocket; Devine, 2003), are interpreted to represent subsidence-triggered, debris-avalanche deposits derived from collapse of fault scarp walls composed of pillowed flows and mafic volcaniclastic deposits (Gibson, 2006; Gibson et al., 2009). Lateral variations in the thickness of megabreccia units is consistent with their emplacement by high-concentration mass flows into localized fault-controlled depositional basins (Devine, 2003; Gibson et al., 2009).

A succession of monolithic mafic breccia (subunit Fm5a, represents a finer, more distal facies of the megabreccia), aphyric to porphyritic coherent rhyolite (unit Fm4) and associated felsic volcaniclastic rocks (unit Fm7), lenses of massive sulphide (unit Fm8), and minor heterolithic mafic and felsic breccia (subunit Fm6a) locally overlies the megabreccia units (subunits Fm5a, f, g). Coherent and volcaniclastic felsic units define 'felsic volcanic centres' that are locally intercalated with aphyric to sparsely plagioclase-phyric mafic flows (unit Fm1). The close association of coherent rhyolite flows/domes (unit Fm4; Figure 14c) with bedded monolithic felsic breccia and tuff breccia to lapillistone (subunit Fm7b; Figure 14d), and well-bedded, heterolithic mafic and felsic breccia and tuff breccia to lapillistone (subunit Fm6a; Figure 14e), suggests that the felsic and heterolithic volcaniclastic units represent carapace and flank breccias to the coherent rhyolite flows/domes (Devine, 2003; Gibson, 2006; Gibson et al., 2009). The bedded, nongraded to normally and reversely graded heterolithic and felsic volcaniclastic units were likely deposited by high concentration mass flows generated during mass wasting and/or collapse of the nearby rhyolite flows/domes (Devine, 2003; Gibson et al., 2009). Underlying and overlying these felsic centres are localized monolithic and heterolithic mafic volcaniclastic units (subunits Fm5a, b), locally intercalated with mafic flows and sills (±peperite; units Fm1–Fm3). These mafic volcaniclastic rocks are interpreted to have been emplaced by a mix of proximal mass flow and syneruptive resedimentation of locally sourced mafic detritus within restricted basins (Devine, 2003; Gibson et al., 2003a, 2009).

The massive sulphide orebodies (unit Fm8) formed principally at or near the top, but also within, below and lateral to rhyolite flow/dome complexes indicating a long-lived hydrothermal event where felsic volcanism was accompanied by ore deposition. Mineralization consists of massive, disseminated and stringer sulphides; clastic, transported sulphide ore lenses are recognized at the Callinan deposit (see details in 'Economic geology'). Two of the rhyolite bodies in close proximity to the mineralization have been dated. The massive base of a quartz-phyric rhyolite dome located just north of the minesite and interpreted to represent immediate footwall of the Flin Flon VMS deposits was dated at 1903 +7/–5 Ma (sample FF92-1, Stern et al., 1999). However, a massive to brecciated rhyolite dome near the South Main mine shaft also situated in the immediate footwall of the Flin Flon Mine yielded an age of 1889 ±2 Ma (sample PQB-1707-08, Rayner, 2010). More work is underway to clarify this ~10 m.y. age discrepancy (Gibson et al., in press).

The Flin Flon, Callinan and 777 deposits are interpreted to have formed within a large, subaqueous subsidence structure or cauldron in which they were spatially and temporally associated with the construction of rhyolite flow and dome complexes that mark the end of the Flin Flon formation volcanism (Devine, 2003; Figure 11c). Felsic volcanism and ore-formation were followed by a period of quiescence dominated by suspension sedimentation of predominately mafic ash to form the uppermost succession of mafic bedded tuff described previously (subunit Fm5e).

South of Millrock hill to Phantom Beach, the Millrock member constitutes a moderately steep to shallowly dipping, east-facing succession that has a total apparent thickness, including thrust repetitions, of >400 m (Figure 12). Immediately east of the Flin Flon Lake Fault, the Millrock member is characterized by a succession of predominantly aphyric to sparsely plagioclase-phyric mafic flows (unit Fm1) and lesser units of amoeboid-clast volcaniclastic units with local areas of intense quartz-epidote alteration. These rocks were formerly referred to as the Creighton member of the Flin Flon formation (Table 1, back pocket; MacLachlan and Devine, 2007). Overlying and locally interbedded with these aphyric flows is a package of thin-bedded to laminated mafic tuff (subunits Fm1f, Fm5e) and massive to flow-banded to in situ–brecciated quartz±plagioclase–phyric rhyolite (subunits Fm4, Fm6a, Fm7b; Figure 10). Locally between the flows and tuffs is a package of massive to bedded scoriaceous mafic breccias, in part interbedded with tuff and lapilli tuff (subunit Fm5c). The lack of exposed carapace/resedimented breccia along the contacts of some of these rhyolite bodies suggests that they are not all extrusive rhyolite flows, but more likely represent felsic synvolcanic sills or 'cryptoflows' (MacLachlan and Devine, 2007). However, some of the coherent rhyolite units have in situ–brecciated and clast-rotated breccias along their upper margins (with no peperite) and are associated with bedded felsic volcaniclastic deposits, which suggests they represent local flow-dome complexes. East of Millrock hill, the top of the Millrock member is marked by a thick section of bedded tuff (subunit Fm5e) that is intruded by amygdaloidal basalt sills and overlain by the aphyric massive mafic flows with amoeboid flow tops and large quartz-filled gas cavities of the Hidden formation (unit Hr1).

Unlike areas to the north that are described above, the Millrock member south of Millrock hill is dominated by numerous aphanitic, fine- to coarser-grained synvolcanic sills (some with peperitic margins; subunit D2b) that constitute sill complexes within the bedded tuff and felsic units. The sill complexes are variably epidote-quartz altered and often contain ‘screens’ of bedded tuff, coherent rhyolite and felsic volcaniclastic rocks Figure 10.

Between Phantom Beach and the Green Lake Fault, the Millrock member is composed of a complexly intercalated succession of mafic tuff, lapilli tuff and breccia (subunits Fm5d, e) intercalated with aphyric and plagioclase±pyroxene–phyric mafic flows (subunits Fm1d, Fm2a, Fm3c). This succession appears to thicken towards the south, although the base of the Millrock member in this area is removed by either the Boundary intrusions or the Flin Flon Lake Fault. Thin-bedded to laminated mafic tuff occurs throughout. In the lower part of the succession, mafic tuff is intercalated with abundant aphyric and porphyritic massive flows and/or sills (subunits Fm5d, h). The central part of the succession is characterized by massive to bedded scoriaceous breccias interbedded with mafic tuff and crystal-rich lapilli tuff (subunit Fm5c; Figures 10, 14f). The uppermost exposed part of the succession is characterized by thin-bedded to laminated mafic tuff, and thin-bedded mafic and felsic tuff and lapilli tuff (subunit Fm6c; Figure 15a), intercalated with aphyric and porphyritic massive and pillowed flows, which are locally peperitic (subunits Fm1f; Fm2b, c; Fm3a). In this area, the top of the Millrock member is truncated by the Burley Lake Fault, which juxtaposes the east-facing rocks of the Millrock member against the west-facing rocks of the Louis formation, or is unexposed beneath Potter Bay.

The abrupt lateral change from mafic coherent flows at Millrock hill to thick discontinuous mafic volcaniclastic units and megabreccias between Millrock hill and Flin Flon minesite is interpreted to define a structural margin within the large cauldron subsidence structure that hosts the Flin Flon–Callinan–777 VMS deposits (Syme and Bailes, 1993; Gibson et al., 2005; MacLachlan and Devine, 2007; Gibson et al., 2009, in press; Figure 11c). The package of coherent mafic flows and the overlying intercalated volcaniclastic rocks and mafic flows observed south of Millrock hill are interpreted to occur within the structural margin of the main subsidence structure (MacLachlan and Devine, 2007; Gibson et al., 2009, in press). The package of massive to bedded scoriaceous mafic breccias, in part interbedded with tuff and lapilli tuff, are interpreted to have infilled smaller structural basins within the margin of the main cauldron.

West of the Flin Flon Lake Fault, the Millrock member occurs in three different thrust panels. Immediately west of the Flin Flon Lake Fault by the Creighton landfill site, northeast of Hilary Lake, rocks of the Millrock member form a >400 m thick (apparent thickness; Figure 12) fault-bounded mafic dike and sill complex (subunit D2f) containing screens of monolithic and heterolithic mafic breccia (subunits Fm5a, b), felsic tuff and lapilli tuff (subunit Fm7a), and massive to brecciated quartz-phyric and quartz-plagioclase–phyric coherent rhyolite (subunits Fm4b, c). Due to the abundance of mafic dikes and sills, no clear primary layering is preserved in the area, which prevents identification of a younging direction for this panel. A sample of felsic, quartz-phyric lapilli to block breccia (part of a composite quartz-phyric massive to brecciated rhyolite body that may either represent a subvolcanic dome or an extrusive flow; subunits Fm4b, D4g) from this thrust panel yielded an age of 1889 ±9 Ma (sample PQB07-KM157-01-01, Rayner, 2010). This age is interpreted to represent the maximum age of the breccia and is consistent with the other dated Millrock member felsic rocks.

A similar fault-bounded thrust panel of Millrock and Blue Lagoon members was identified just south of the latter panel under what is now the southeast corner of the tailings pond (Figures 3, 12). Previous mapping by Stockwell (1960) and Hudson Exploration and Development Co. Ltd. (Figure 2) recognized about 100 m of monolithic felsic breccia, most likely of Millrock member affinity, sitting on plagioclase-phyric mafic flow typical of the Blue Lagoon member.

The Millrock member also occurs at Hilary Lake, where it forms a very thin, steeply dipping southwest-facing laterally continuous succession with an apparent thickness of 1.5–50 m (Figure 12; Geoscientific Map MAP2010-1, back pocket). Here, the Millrock member conformably overlies plagioclase-phyric basalt flows of the Blue Lagoon member (subunit Fb1c) and is conformably overlain by aphyric massive basalt flows of the Hidden formation (subunit Hr1g). The Millrock member is composed of a thin, up to 1.5 m thick, unit of finely bedded to massive felsic lapilli tuff (subunit Fm7a; Figure 15b) that can be traced along strike to the southeast for >500 m. Correlative with this thin felsic tuff, located at the north end of Hilary Lake, are thinly bedded to laminated mafic tuff with lesser massive, mafic lapillistone and breccia (subunits Fm5b, d, e). These tuffaceous units are intercalated with aphyric massive basalt flows with abundant amoeboid flow tops (subunit Fm1e) and bedded, sparsely quartz±plagioclase–phyric rhyolite breccia (subunit Fm7b). A sample of bedded lapilli tuff southeast of Hilary Lake yielded an age of 1886 ±4 Ma (sample PQB07-KM156-01-01, Rayner, 2010), which is consistent with the other dated felsic rocks from the Millrock member. A sample of a quartz-plagioclase–phyric felsic sill collected approximately 2 km to the southeast, near Myo Lake (subunit D4g) where it cuts across the southern extension of this felsic tuff, yielded an age of 1888.9 ±1.7 Ma (sample 05MYO-01, Bailey, 2006).

This succession of thin-bedded to laminated mafic and felsic tuff and felsic volcaniclastic rocks is interpreted to represent the same or a similar hiatus in volcanic activity to the mafic well-bedded tuff unit (subunit Fm5e) on the eastern side of the Flin Flon Lake Fault (Figure 11c).

Hidden formation (H)

(Y.M. DeWolfe, R-L. Simard and K. MacLachlan)

The Hidden formation (H) is composed of mafic flows, sills and volcaniclastic rocks, with subordinate basaltic andesite flows, rhyolite flows and felsic volcaniclastic rocks. It defines the onset of hangingwall volcanism atop the Flin Flon formation volcanic centre and Flin Flon–Callinan–777 VMS

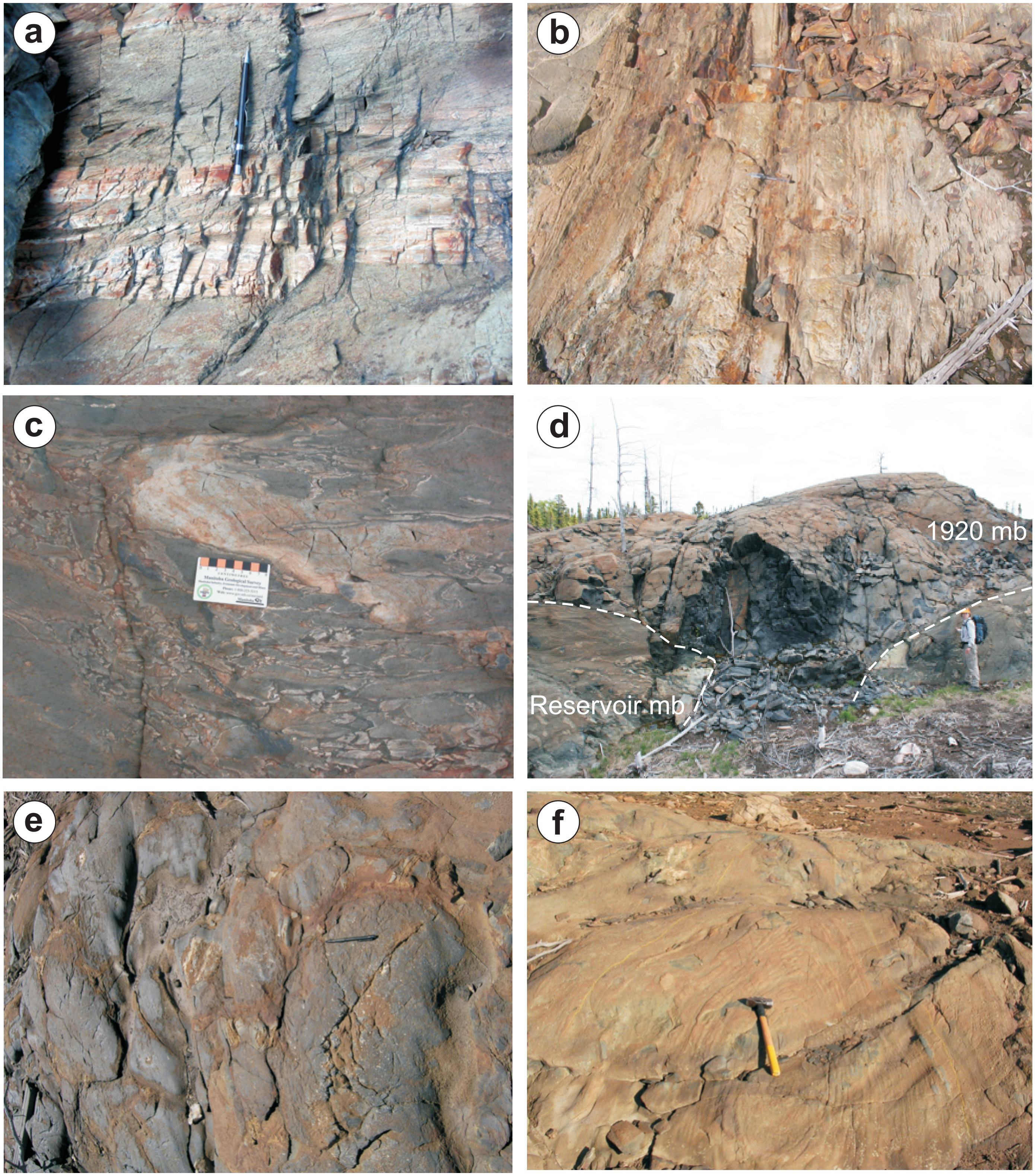


Figure 15: *Representative rock types, facies and textures of the Hidden formation and of the Millrock member of the Flin Flon formation:* ***a)*** *bedded mafic and felsic tuff of the Millrock member, Flin Flon formation (subunit Fm6c);* ***b****) well-bedded felsic tuff of the Millrock member, Flin Flon formation (subunit Fm7a);* ***c)*** *peperite at the upper contact of the 1920 member cryptoflow, Hidden formation (subunit Hi1c);* ***d)*** *basaltic andesite (icelandite) feeder dike of the 1920 member showing columnar jointing, Hidden formation (subunit Hi1a);* ***e)*** *pillowed aphyric basalt of the Reservoir member, Hidden formation (subunit Hr1b); and* ***f)*** *concentric banding in large lava tube of aphyric basalt of the Reservoir member, Hidden formation (subunit Hr1a).*

deposits (Stockwell, 1960; Bailes and Syme, 1989; DeWolfe, 2008; DeWolfe et al., 2009a; Figure 11d).

The Hidden formation is exposed on both sides of the Flin Flon Lake Fault as well as on the eastern side of the Burley Lake Fault (Figure 16). It can be subdivided in four mappable members: the 1920 member (Hi), the Reservoir member (Hr), the Stockwell member (Hs) and the Carlisle Lake member (Hc). Only the Reservoir member is present on both sides of the Flin Flon Lake Fault, the others are restricted to the eastern side of the Flin Flon Lake Fault.

The base of the Hidden formation is placed at the last occurrence of tuff and/or rhyolite of the underlying Millrock member of the Flin Flon formation, and is marked by 1) aphyric to sparsely plagioclase-phyric basalt flows and/or sills (unit Hr1, Reservoir member); or 2) ±basaltic andesite flows or sills containing amphibole porphyroblasts (units Hi1, Hi2, 1920 member; DeWolfe, 2008).

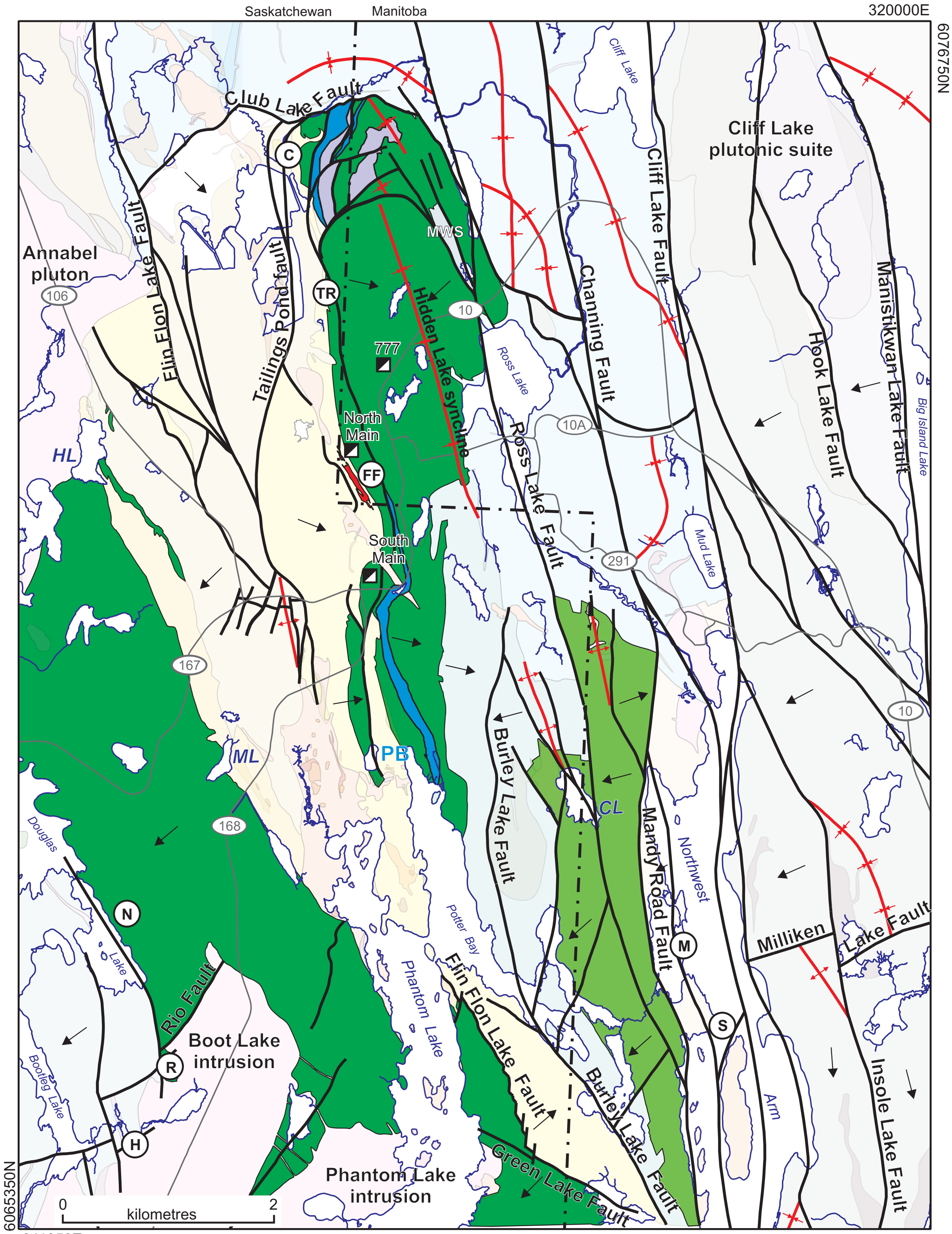
1920 member (Hi)

The 1920 member is restricted to an area east of the Flin Flon Lake Fault on the western limb of the Hidden Lake syncline (Figure 16). It has been recognized in a few different fault panels, mainly north of the Lower Railway Fault (subunits of Hi1), and one east of the Flin Flon open pit and south towards Phantom Beach (unit Hi2; Figures 4, 16). The 1920 member (Hi) is composed of aphyric, Fe-Ti-P–enriched, basaltic andesite cryptoflows (unit Hi1), and aphyric to sparsely plagioclase-phyric, Fe-Ti-P–enriched basalt flows and sills (unit Hi2).

The rocks of the 1920 member have a very distinctive geochemistry, which allows them to be differentiated from other mafic volcanic rocks in the area. The mafic flows/sills of the 1920 member, as well as a few more basaltic flows towards the south, plot in the basaltic andesite field of Pearce (1996). All of them are enriched in TiO_2, Fe_2O_3 and P_2O_5 and depleted in V and Sc relative to other mafic rocks in the Hidden formation (Ti/V 168–240 vs. 15–50, DeWolfe et al., 2009a). Rocks of the 1920 member, although more correctly identified as Fe-Ti basalt or Fe-Ti basaltic andesite, have been referred to as 'icelandite' in the immediate Flin Flon area because of their more andesitic composition relative to surrounding basalts, and their Fe-Ti–enrichment, which resembles true mid-ocean-ridge–derived icelandite as first described by Carmichael (1967). The Fe-Ti basalt or Fe-Ti basaltic andesite have been recognized elsewhere in arc environments and are interpreted to be derived from shallow-level (<2 km) magma chambers in localized rifts (Embly et al., 1988; Barrie and Pattison, 1999; Perfit et al., 1999; DeWolfe et al., 2009b). Such shallow magmatic chambers would have provided a high-temperature environment that could have generated and sustained the high-temperature convective hydrothermal systems necessary for the formation of massive sulphide mineralization (DeWolfe et al., 2009b), which make the recognition of 'icelandite-like' units a potential VMS exploration tool.

North of the Lower Railway Fault, rocks of the 1920 member appear in two separate fault blocks separated by two northeast-striking thrust faults (Figures 4, 16; Geoscientific Map MAP2010-1, back pocket). These thrusts structurally repeat the 1920 member and the bounding basalt sills or flows of the Reservoir and Stockwell members on the northwest limb of the Hidden Lake syncline around the Catherine Fault. North of the Lower Railway Fault, the 1920 member is easily recognisable by its abundant randomly oriented acicular amphibole porphyroblasts (25–50%) and by its light metallic blue colour on fresh surfaces. It forms a semi-intrusive, <100 m thick succession of sills (>95% volume) and a thick, massive to pillowed, aphyric basaltic andesite cryptoflow (subunits of Hi1; DeWolfe, 2008) with an approximate strike length of 1.1 km (Figure 16). In this area, the 1920 member cryptoflow has well-developed peperite along its upper contact (subunit Hi1c; Figure 15c), includes tuff within the upper 10–30 m, and has lenses of included tuff along its basal contact (subunit Hi1c). The base is characterized by massive 'feeders' (subunit Hi1a; Figure 15d) that have been emplaced through massive to pillowed flows/cryptoflows of the Reservoir member (subunits Hr1a, b) at the base of both thrust panels (Figure 16). The upper contact of the 1920 member is bounded by felsic or mafic volcaniclastic rocks of the Reservoir member (subunits Hr4e, Hr5) or by basalt sills or flows of the Reservoir and Stockwell members (units Hr1 or Hs1). In places, this cryptoflow shows abrupt changes in thickness coincident with facies changes, from thick massive cryptoflow (subunit Hi1a) to thin pillowed cryptoflow. The thick massive cryptoflow is interpreted as a ponded flow within a graben (DeWolfe, 2008; DeWolfe et al., 2009a). The thin pillowed cryptoflow represents the breaching of the margin of this graben by the magma to form a thin pillowed cryptoflow as a result of slower flow rates due to ponding of the main flow mass within the graben (DeWolfe, 2008; DeWolfe et al., 2009a).

East of the Flin Flon open pit and south towards Phantom Beach, the flows and sills of the 1920 member form a thin east-facing succession with a total apparent thickness of 50–200 m (Figure 16). It can be traced along strike for >2.5 km from the Flin Flon open pit south to Phantom Beach (Figure 16). The flows and sills lack the amphibole porphyroblasts so characteristic in the north; the regional 'amphibole-in' metamorphic isograd runs east-west around Hidden Lake, between the two 1920 member occurrences. The 1920 member east of the Flin Flon pit forms a thin, <50 m thick succession of Fe-Ti–enriched basalt flows with scoriaceous flow-top breccia (subunits Hi2a, b) sandwiched between mafic tuff and lapilli tuff of the Reservoir member (subunit Hr5a). The flows are massive, amygdaloidal at the base, with increasing amounts of amygdules and large gas cavities (up to 20 cm long) towards the top. They commonly have well-developed amoeboid flow-top breccia, and are locally peperitic. Farther south towards Phantom Beach, the flows forms a succession up to 200 m thick and are massive to pillowed (subunit Hi2a); locally mega-tubes or lobe structures are exposed. Similar to some of the flows in the north, these flows commonly have well-developed flow-top breccia (subunit Hi2b). This succession is either conformably overlying the uppermost tuff of the Flin Flon formation (subunit Fm5e, Millrock member) or within aphyric massive to pillowed flows and sills of the Hidden formation (unit Hr1, Reservoir member).



Hidden formation

- Carlisle Lake member
- Stockwell member
- Reservoir member
- 1920 member

- Road
- Provincial boundary
- Younging direction
- Major fault
- Syncline / anticline
- Head frame

Deposits

FF: Flin Flon	S: Schist Lake
C: Callinan	N: Newcor
TR: 777	R: Rio
M: Mandy	H: Henning-Maloney

Figure 16: *Simplified geology of the Flin Flon area, showing four mappable members of the Hidden formation and major structures in the area. Abbreviations: CL, Carlisle Lake; PB, Phantom Beach; HL, Hilary Lake; ML, Myo Lake; MSW, Missi wedge structure. For complete legend refer to Figure 3.*

Reservoir member (Hr)

The Reservoir member (Hr) is found on both sides of the Flin Flon Lake Fault (Figure 16). East of the fault, the Reservoir member forms a steeply to moderately dipping succession, with an apparent thickness of 400–800 m, which is folded around the Hidden Lake syncline and the Burley Lake Fault (Figures 4, 16). It can be traced along strike for >6 km from the Club Lake Fault in the north all the way to Phantom Lake in the south (Figure 16). North of the Upper Railway Fault, portions of the Reservoir member, along with rocks of the Stockwell member and 1920 member, are thrust-repeated by two northeast-trending thrust faults and the Upper and Lower Railway, and Catherine thrusts faults (Figures 4, 16). East of the Flin Flon open pit, portions of the Reservoir member are repeated along with the top of the Millrock member of the Flin Flon formation by a west-verging north-trending thrust fault (Figure 16). West of the Flin Flon Lake Fault, the Reservoir member forms a homoclinal, steeply dipping, southwest-facing succession with an apparent thickness of >1000 m (Figure 16).

The Reservoir member (Hr) is composed of aphyric (unit Hr1), plagioclase-phyric (unit Hr2) and pyroxene-plagioclase–phyric (unit Hr3) basalt flows, with minor volcaniclastic rocks (units Hr4, Hr5, Hr6). Although not always specifically mentioned, some of the rock units described below as flows may be more appropriately termed cryptoflows, especially towards the base of the Reservoir member.

Aphyric flows of the Reservoir member (unit Hr1) are aphanitic with <5% plagioclase phenocrysts (1–2 mm in diameter), and are black on weathered surfaces and dark green on fresh surfaces. Aphyric flows are mainly massive (>70% in volume; subunit Hr1a) and pillowed (>30% in volume; subunit Hr1b; Figure 15e), with some large 'lava tubes' (<5% in volume; subunit Hr1a), and commonly have peperitic margins (subunits Hr1d, e). Aphyric massive flows (subunit Hr1a) are usually 5–15 m thick, nonamygdaloidal near the base and contain up to 30% quartz-filled amygdules (up to 10 cm in size) near the top. Some of these massive flows are capped by well-developed amoeboid flow-top breccia (subunit Hr1g). Columnar jointing is observed locally. The majority of the flows are variably peperitic along their margins (subunit Hr1e) or contain included tuff (subunit Hr1d) at their top. These massive aphyric flows locally grade upward to peperitic pillowed flows (subunits Hr1b, e) and pillowed flows with included tuff (subunit Hr1d) or pillowed flows (subunit Hr1b). Aphyric pillowed flows of the Reservoir member (subunit Hr1b) contain pillows that are commonly closely packed, elongated in shape, amygdaloidal (5–15%, 2–5 mm in diameter) and 50–100 cm in diameter (up to 250 cm), with 1–3 cm thick recessively weathered chilled margins. In some areas, large ovoid lava tubes (subunit Hr1a; Figure 15f) displaying concentric cooling laminations, with amygdules concentrated towards their margins, are surrounded, but not truncated, by pillows (DeWolfe, 2008; DeWolfe et al., 2009a). Both massive and pillowed flows commonly show light green quartz-epidote alteration patches, usually elongated and parallel to flow contact in the massive flows, and ovoid to amoeboid in the pillowed flows.

Plagioclase-phyric flows of the Reservoir member (unit Hr2) have between 5 and 15% plagioclase phenocrysts (2–4 mm in diameter) with less than 3% pyroxene phenocrysts (1–3 mm in diameter), and are buff brown to brown on weathered surfaces and dark green on fresh surfaces. They are mainly pillowed (>80% in volume; subunit Hr2b; Figure 17a) with some lava tubes (<5% in volume; subunit Hr2a) and localized amoeboid pillow breccia (<5% volume; subunit Hr5e) and in situ breccia (<5% volume; subunit Hr2d; Figure 17b). Pillows within the plagioclase-phyric pillowed flows are usually closely packed, ovoid in shape, amygdaloidal (5–15%, 1–3 mm in diameter) and 20–50 cm in diameter (up to 80 cm) with 1–2 cm thick darker brown to reddish chilled margins. Pillow breccia and in situ breccia (subunit Hr2d) show plagioclase-phyric fragments, up to 30 cm in diameter, with chilled margin on one or more sides, sitting in plagioclase-crystal–rich matrix containing cuspate fragments with chilled margins. These breccia units are commonly massive to crudely bedded.

Plagioclase-pyroxene–phyric flows of the Reservoir member (unit Hr3) make up <5% of the Reservoir member. These flows have between 5 and 15% pyroxene phenocrysts (1–4 mm in diameter) and between 5 and 15% plagioclase phenocrysts (2–3 mm in diameter). The flows are buff brown on weathered surfaces and medium green on fresh surfaces. They are massive (>70% in volume; subunit Hr3b) and pillowed (>30% in volume; subunit Hr3b).

Volcaniclastic rocks (units Hr4, Hr5, Hr6) are present in the Reservoir member in thin fairly discontinuous packages either found as thin layers between flows, or included in flows or between pillows.

East of the Flin Flon Lake Fault on the western limb of the Hidden Lake syncline, the Reservoir member lies conformably atop the uppermost well-bedded mafic tuff of the Millrock member of the Flin Flon formation (subunit Fm5e). The base of the Reservoir member is characterized by the abundance of sills and aphyric massive and pillowed cryptoflows, peperites and flows (unit Hr1) intercalated with thin discontinuous layers of mafic and felsic volcaniclastic rocks (units Hr4, Hr5, Hr6). Sheet flows, thin extremely continuous flows along strike, are also observed in this area (DeWolfe, 2008). The local presence of peperite at the top of massive and pillowed basalt flows, near the base of the Reservoir member, suggests that they were in part emplaced as massive to pillowed high-level sills or cryptoflows into wet, unconsolidated volcaniclastic material (DeWolfe, 2008). Some of this volcaniclastic material was likely part of the uppermost well-bedded mafic tuff unit of Millrock member (subunit Fm5e), which was incorporated into the cryptoflows of the Reservoir member (MacLachlan and Devine, 2007; DeWolfe, 2008; Gibson et al., 2009). North of the Upper Railway Fault, early sills and cryptoflows of the Reservoir member are intruded by the basaltic andesite cryptoflow of the 1920 member, which is in turn overlain in places by more aphyric massive and pillowed flows of the Reservoir member. This intercalation of Reservoir member and 1920 member cryptoflows suggest that the two were emplaced at approximately the same time. Since the cryptoflows were emplaced just below the seafloor (as suggested by the presence of peperite) and are not laterally extensive (<200 m), their abundance/presence implies a vent proximal environment for the base of the Reservoir member (DeWolfe, 2008; DeWolfe

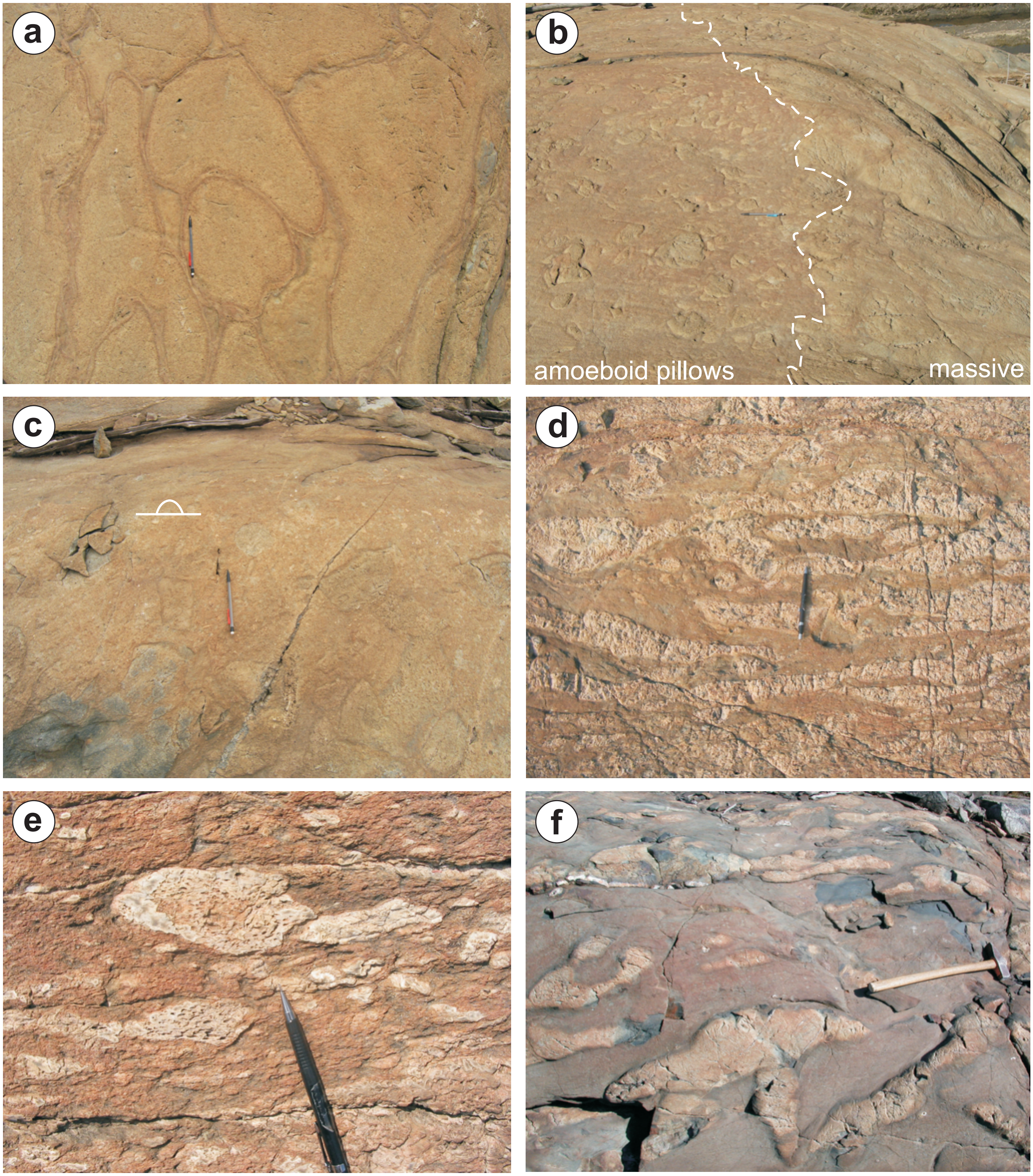


***Figure 17:** Representative rock types, facies and textures of the Reservoir member of the Hidden formation: **a)** plagioclase-phyric pillowed basalt (subunit Hr2b); **b)** massive to amoeboid pillowed plagioclase-phyric basalt (subunit Hr2c, d); **c)** normally graded mafic lapilli-tuff to tuff (subunit Hr5a); **d)** silicified amoeboid flow-top breccia of an aphyric flow (subunit Hr1g); **e)** silicified mafic scoriaceous lapilli tuff (subunit Hr5b); and **f)** epidote-quartz alteration patches in a massive aphyric basalt flow (subunit Hr1a).*

et al., 2009a). In this area, the exposed volcanic stratigraphy has been thickened by abundant synvolcanic sills and dikes (>15% in volume; unit D2), which in combination with the high concentration of large lava tubes is also indicative of a vent-proximal environment (DeWolfe, 2008; DeWolfe et al., 2009a). The dikes often have brecciated or irregular margins with the 1920 member and Reservoir member, suggesting that they are synvolcanic. The presence in the area of a large synvolcanic pyroxene-phyric mafic intrusion (subunit D2d) with peperitic contacts, at the base of the Reservoir member, further supports this interpretation (DeWolfe, 2008; DeWolfe et al., 2009a).

Higher in the stratigraphy, the Reservoir member is characterized by the intercalation of aphyric massive and pillowed flows (unit Hr1) and plagioclase-phyric pillowed flows (subunit Hr2b). South of the Lower Railway Fault, the aphyric flows (unit Hr1) and the plagioclase-phyric flows (unit Hr2) form a >800 m thick succession. The ratio of pillowed to massive flows increases significantly towards the top of the Reservoir member with a decrease in the occurrence of the large lava tubes. The large lava tubes or megapillows are interpreted to represent large lava tubes that grew through slow steady, sustained eruptions that fed advancing flows (DeWolfe, 2008; DeWolfe et al., 2009a). The decrease in the abundance of large lava tubes and the increase in the presence of pillowed over massive flows suggest a decrease in eruption rate or a less proximal setting for the higher part of the Reservoir member stratigraphy.

Localized plagioclase-pyroxene–phyric flows (unit Hr3) are also present in this succession. Mafic volcaniclastic rocks (unit Hr5) are found discontinuously at various levels throughout the upper stratigraphy of the Reservoir member. Mafic lapilli tuff and tuff (unit Hr5; Figure 17c), 4–20 m thick, occurs intercalated with the pillowed flows near the core of the Hidden Lake syncline. The lapillistone units are clast supported, massive to thinly bedded and composed of 70% amygdaloidal (5–85% quartz-filled amygdules), rounded to subangular basalt clasts (0.2–20 cm, with an average clast size of 2 cm); and 10% aphanitic, subrounded to subangular nonamygdaloidal basalt clasts (0.2–5 cm) in a light brown tuff matrix (DeWolfe, 2008). The lapillistone beds are interpreted as mass flow units deposited in localized (<100 m) basins (DeWolfe, 2008; DeWolfe et al., 2009a). Given the scoriaceous nature of the dominant clast type, these units may be the result of redeposition of primary pyroclastic deposits into localized basins (DeWolfe, 2008).

East and south of the Flin Flon open pit, towards Phantom Beach, the flows and sills of the Reservoir member (subunit Hr1) are intercalated with flows and sills of the 1920 member. Peperite is only locally observed in the area, which suggests that most of the flows were actually extruded to the seafloor. In this area, the volcanic stratigraphy of the Reservoir and 1920 members is much thinner than in the north, and contains fewer dikes. This suggests a decrease in eruption rate or a less proximal setting for this part of the Reservoir and 1920 members.

On the eastern limb of the Hidden Lake syncline, the Reservoir member forms a >400 m thick west-facing succession of plagioclase-phyric basalt pillowed flows (subunit Hr2b) with minor amounts of associated in situ breccia (unit Hr2d). The base of the Reservoir member has been faulted off along the north-trending Ross Lake Fault or the Missi wedge structure (MWS; Figure 16). Structural attenuation of the base of the Reservoir member, along the major north-trending strike-slip faults on the eastern limb of the Hidden Lake syncline, namely the Ross Lake Fault and the Missi wedge structure, most likely accounts for the difference in thickness and stratigraphy between the limbs of the syncline. However, the abrupt transition just east of the Hidden Lake syncline fold axis, from massive and pillowed flows with large lava tubes on the western limb, to mostly pillowed flows on the eastern limb, could also represent a synvolcanic subsidence structure (DeWolfe, 2008; DeWolfe et al., 2009a). In this scenario, the thicker, more proximal volcanic facies of the Reservoir member observed on the western limb of the Hidden Lake syncline may have been emplaced in a rapidly subsiding synvolcanic graben, whereas the pillowed flows of the eastern limb may have been erupted outside that graben; therefore, the transition from thicker more massive flows to thinner, pillowed flows may represent the synvolcanic fault marking the margin of the graben.

Immediately west of the Flin Flon Lake Fault, the Reservoir member is characterized by aphyric massive flows with abundant amoeboid flow-top breccia (subunit Hr1g). These massive flows have abundant large quartz-filled amygdules and gas cavities as well as pervasive silicification, particularly in the flow-top breccia facies (Figure 17d). Intercalated with these flows are thin discontinuous units of thin-bedded to laminated mafic to felsic tuff and felsic breccia (subunits Hr5b, d; Hr6a, b; Hr7b). Above the massive flows is a thick succession of aphyric, amygdaloidal massive to pillowed flows, locally with abundant interflow and intraflow peperite (subunit Hr1c, e) and discontinuous units of thin-bedded to laminated mafic to felsic ash tuff (subunit Hr6b). The abundance of peperite suggests that these flows are actually cryptoflows that were emplaced into unconsolidated tuff, most likely from the underlying tuffaceous units of Millrock member (MacLachlan and Devine, 2007; DeWolfe, 2008; Gibson et al., 2009). Most of the flows have minor to moderate quartz-epidote alteration and are pervasively weakly to moderately silicified. These aphyric flows are overlain by up to 100 m of coarser volcaniclastic rocks (subunit Hr5b) consisting of poorly bedded mafic volcaniclastic rocks with abundant silicified amygdaloidal to scoriaceous fragments (Figure 17e). The composition and texture of this unit varies considerably along strike, from strongly silicified basalt lapillistone in the south, to locally well-bedded and graded, clast-supported tuff breccia, to typically bedded, matrix-supported, heterolithic lapilli tuff in the north. This silicified volcaniclastic unit is overlain by massive to predominantly pillowed aphyric amygdaloidal flows (subunits Hr1b, c) that are intercalated with rare discontinuous thin- to thick-bedded mafic tuff to lapilli tuff (subunit Hr5a; MacLachlan, 2006c). Just north of the Boot Lake intrusion, a package of monolithic felsic breccia (subunit Hr7b) is also found intercalated with these mafic flows, which suggests some extrusive felsic volcanism within the Reservoir member.

Overlying these volcaniclastic rocks, the aphyric flows just east of Douglas Lake are typically massive to pillowed, moderately to strongly amygdaloidal, with abundant quartz-epidote alteration in the cores of the pillows and as irregular elongate patches or regular bands approximately parallel to bedding in the massive flows (Figure 17f). Semiconformable silicification is very common in this succession (MacLachlan, 2006c; MacLachlan and Devine, 2007), and is similar to that in the footwall of major VMS deposits in both Snow Lake, Manitoba and Noranda, Quebec. Several of the gold/copper-gold showings and past producers in the Douglas Lake area are associated with massive sulphides, bedding-parallel disseminated sulphides and copper- and zinc-bearing sulphides, suggesting potential for VMS mineralization at the top of the Reservoir member along Douglas Lake (see 'Economic geology' below).

The Reservoir member west of the Flin Flon Lake Fault, does not have abundant dikes like it has east of the fault, but rather is characterized by abundant sills and sill-like intrusions (units D2, D4; MacLachlan et al., 2002). Between Bomber, Key and Reddy lakes, and in other areas, mafic sills and sill-like fine- to medium-grained aphyric and plagioclase-phyric gabbro intrusions (unit D2) make up over 70% of the sequence. In places, the mafic sills are in such abundance that they form a sill complex, locally with screens of the hostrocks remaining (subunit D2f). Numerous thin and discontinuous, massive to flow-banded, aphyric to quartz- and plagioclase-phyric felsic units (unit D4), approximately parallel to bedding, are also common in the area. These felsic rocks are typically massive, coherent bodies lacking flow features, which suggests that they are intrusive and emplaced by single pulses of magma to form sheet-like sills (Bailey, 2006). Although some of these units do display flow banding and even breccia, their upper contacts are usually sharp and chilled with no evidence of the development of hyaloclastite or carapace breccia. In some localities, the felsic rocks truncate bedding on a centimetre-scale, which suggests intrusive emplacement (Bailey, 2006). A large, >300 m thick, sparsely quartz- and plagioclase-phyric felsic sill cuts the base of the Reservoir member and the top of the Blue Lagoon and Millrock members between Myo and Hilary lakes ('Myo intrusions' or 'Myo rhyolites'; Table 1) at a low angle. A sample of this felsic intrusive unit collected near Myo Lake yields an age of 1888.9 ±1.7 Ma (Bailey, 2006).

Just north of Myo Lake, within the predominantly massive Myo intrusion, is a quartz-plagioclase–phyric felsic extrusive unit (subunit Fm7b) recognized by Bailey (2006). At this location, monolithic felsic breccia is found in close spatial association with the flow-banded massive flow and displays gradational contacts where contorted flow bands break up to form a flow breccia suggesting extrusive emplacement (Bailey, 2006). This extrusive felsic unit is interpreted to be part of the Millrock member of the Flin Flon formation that was incorporated in the Myo intrusion during emplacement (MacLachlan and Devine, 2007).

Similar to the area east of the Flin Flon Lake Fault, the base of the Reservoir member west of the Flin Flon Lake Fault is interpreted to represent rapidly emplaced cryptoflows erupted into the unconsolidated tuffs of the Millrock member of the Flin Flon formation in a volcanic subsidence structure proximal to the magma source (MacLachlan and Devine, 2007). The interbedded coarser volcaniclastic rocks and nonpeperitic flows observed higher in the stratigraphy would suggest a decrease in eruption rate or a less proximal setting for the higher part of the Reservoir member in this area. The Reservoir member is in faulted contact with the Douglas formation to the west along the eastern shore of Douglas Lake. It is cut in the south by the Phantom Lake intrusive suite (unit S3).

Stockwell member (Hs)

The Stockwell member (Hs) is only found on the eastern side of the Flin Flon Lake Fault (Figure 16). It occurs north of the Lower Railway Fault where it forms a generally south-facing moderately dipping sequence with an apparent thickness of up to 300 m. The Stockwell member can be traced along strike for over 500 m around the nose of Hidden Lake syncline. On the western limb of the Hidden Lake syncline, the Stockwell member is thrust-repeated with portions of the Reservoir member and 1920 member by a set of northeast-trending thrust faults, which in turn are offset by the Catherine Fault (Figures 4, 16; Geoscientific Map MAP2010-1, back pocket). In the lower thrust panel, the Stockwell member occurs immediately above the 1920 member (Figure 16). In the upper thrust panel and on the eastern limb of the Hidden Lake syncline, the Stockwell member is located within aphyric flows and sills of the Reservoir member (Figure 16). In places, the upper and lower contacts of the Stockwell member with the Reservoir member flows are marked by the presence of finely laminated mafic tuff (subunit Hr5d).

The Stockwell member (Hs) includes massive, pillowed and breccia facies of strongly plagioclase (15–50%) and pyroxene (<5%) porphyritic basalt flows (unit Hs1) and is locally overlain by mafic volcaniclastic rocks (unit Hs2; DeWolfe et al., 2009a). Massive domains (subunit Hs1a) of fine- to medium-grained, nonamygdaloidal basalt containing 15–40% plagioclase phenocrysts (1–4 mm) and <5% pyroxene phenocrysts (1–2 mm) occur within the pillowed flows (subunit Hs1b). These massive domains are commonly crudely ovoid in shape, from 1 m to approximately 60 m wide, columnar-jointed, and grade over 10–20 cm into the pillowed facies. Their margins are locally flow-banded, commonly become finer grained closer to the contact, and do not truncate, but conform to, the shape of surrounding pillows, which suggests they are large (≤60 m) lava tubes or megapillows (DeWolfe, 2008; DeWolfe et al., 2009a).

Surrounding pillows (subunit Hs1b; Figure 18a) range in size from 0.2–1.0 m, are brown in colour and amygdaloidal; amygdules are equally dispersed throughout the pillows (DeWolfe, 2008). The pillows have dark reddish-brown selvages (1–2 cm wide), commonly have light brown epidote-quartz alteration patches in their cores and, locally, have light brown, finely laminated tuff inclusions between pillows. In places, 10 by 5 m domains of fractured or in situ–brecciated pillows can be observed showing angular fragments of pillows (1–10 cm in size) separated by light brown hairline fractures and displaying a 'jigsaw-puzzle fit'. These domains of fractured/in situ–brecciated pillows are surrounded by intact pillows and may be a product of local in situ fracturing during loading and/or cooling of the flow (pillow breccia; DeWolfe, 2008). Locally, pillowed flows of the Stockwell member grade upward from pillow facies into fluidal breccia units. The breccia contains brown, quartz-amygdaloidal (10–5%, 1–2 mm), plagioclase-phyric (15%), amoeboid-shaped, basalt fragments (2–60 cm

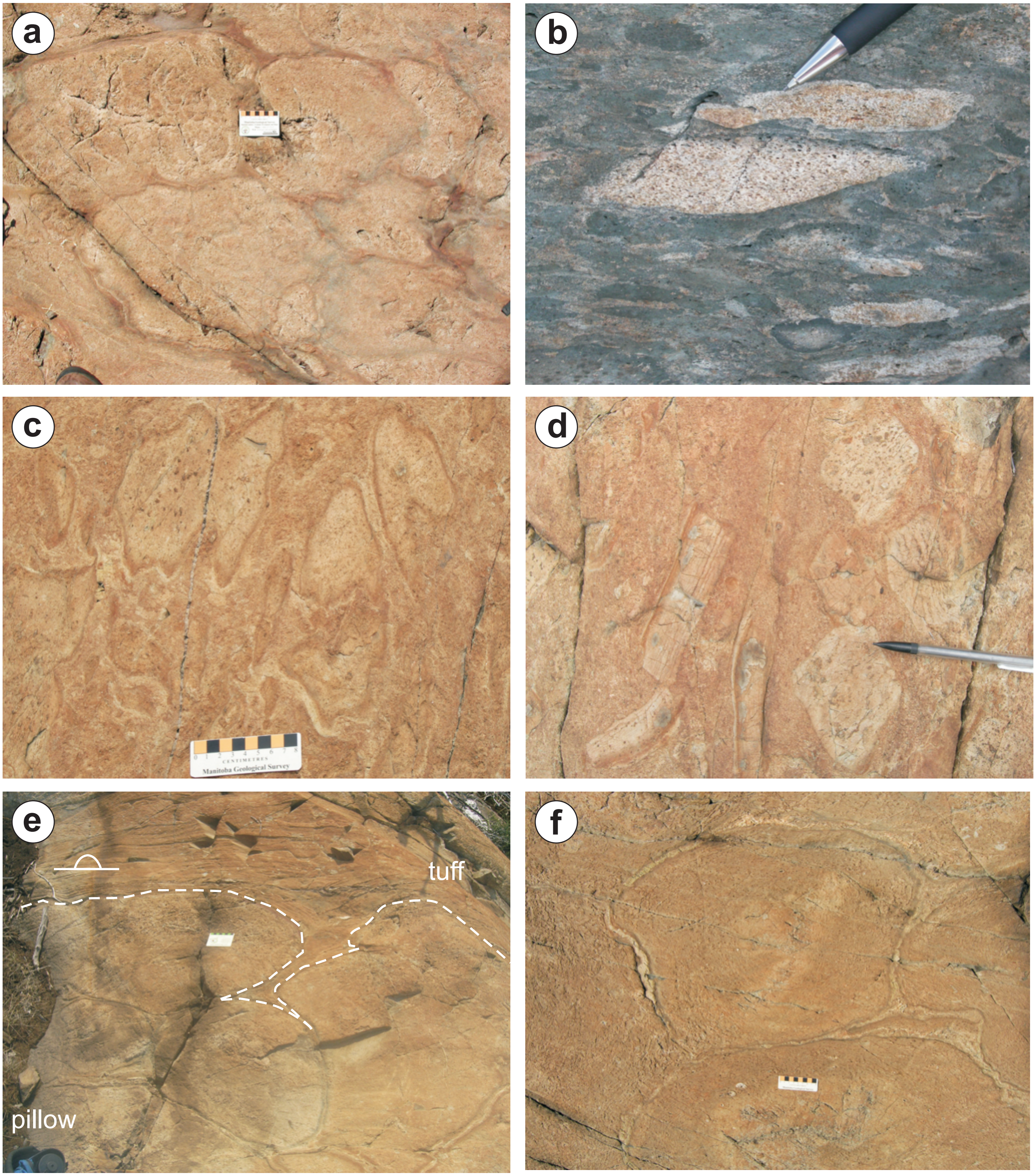


Figure 18: *Representative rock types, facies and textures of the Hidden formation:* ***a)*** *highly plagioclase- and pyroxene-phyric pillowed basalt of the Stockwell member (subunit Hs1b);* ***b)*** *heterolithic lapillistone bed between flows of the Stockwell member showing pumiceous and scoriaceous clasts (subunit Hs2a);* ***c)*** *amoeboid pillows of an aphyric to sparsely plagioclase-phyric basalt flow of the Carlisle member (subunit Hc1d);* ***d)*** *pillow breccia of the Carlisle member (subunits Hc4a, d);* ***e)*** *well-bedded mafic tuff draping an aphyric to sparsely plagioclase-phyric pillowed basalt flow of the Carlisle member (subunits Hc1b, Hc4c); and* ***f)*** *aphyric to sparsely plagioclase-phyric pillowed basalt flow of the Carlisle member showing concentric thermal contraction cracks (subunit Hc1b).*

long) in a fine-grained, dark grey matrix and is interpreted to be an autoclastic flow-top breccia (DeWolfe, 2008).

Mafic lapillistone beds (subunit Hs2a; Figure 18b), 4–20 m thick, occur intercalated with the pillowed flows. These lapillistone beds are clast supported, massive to thinly bedded and composed of rounded to subangular basalt clasts (0.2–20 cm with an average clast size of 2 cm) in a light brown tuffaceous matrix. The lapillistone beds are interpreted as mass flow units deposited in localized (<100 m wide) basins (DeWolfe, 2008; DeWolfe et al., 2009a).

The intercalation of the Stockwell member with cryptoflows and sills of the Reservoir member and 1920 member suggests that they were coevally emplaced (DeWolfe, 2008; DeWolfe et al., 2009a). The high concentration of large lava tubes, similar to those found at the base of the Reservoir and 1920 members, is indicative of a vent-proximal environment for the emplacement of the Stockwell member flows (DeWolfe, 2008; DeWolfe et al., 2009a).

Carlisle Lake member (Hc)

The Carlisle Lake member (Hc) is only found on the eastern side of the Flin Flon Lake Fault, between the Burley Lake Fault and the Channing–Mandy Road fault (Figure 16). South of Carlisle Lake, the Carlisle Lake member forms a homoclinal steeply dipping, west-facing succession with an apparent thickness >800 m including potential structural thickening by the Ross Lake Fault (Figure 16). North of Carlisle Lake, on the eastern side of the Ross Lake Fault, the rocks of the Carlisle Lake member are folded/faulted around the Mandy Road anticline and form an east-facing sequence >400 m thick that is truncated to the east by the Channing–Mandy Road fault (Figure 16). The Carlisle member stratigraphy can be traced along strike for >5 km from the Channing airstrip all the way south to Green Lake. The base of the Carlisle Lake member is truncated to the east by the West Mandy Road Fault and the top is exposed on the peninsula between Burley Lake and Potter Bay of Phantom Lake (Figure 16; Geoscientific Map MAP2010-1, back pocket).

The Carlisle Lake member (Hc) is composed of aphyric to sparsely plagioclase-phyric (unit Hc1), plagioclase-phyric (unit Hc2) basalt flows, mafic volcaniclastic rocks (unit Hc4) and minor amounts of felsic volcanic rocks (unit Hc3).

Aphyric to sparsely plagioclase-phyric basalt flows (unit Hc1), which account for >50% of the Carlisle Lake member rocks, show <7% plagioclase phenocrysts (<2 mm), are usually buff brown to brown on weathered surfaces, except in areas of higher strain where they are dark green, and are dark green on fresh surfaces. They usually form 3–10 m thick flows (subunit Hc1a), which usually display a massive base (subunit Hc1c; 1–5 m thick), grading upward into megapillows or pillows (subunit Hc1b; 0.3–5 m thick), in turn grading into amoeboid flow-top breccia (subunit Hc1d; Figure 18c), although they can be missing the massive or pillowed section (Simard, 2006a). They are locally intercalated with massive to well-bedded pillow breccia (subunit Hc4a; Figure 18d), heterolithic mafic breccia (subunits Hc4a, b) and mafic tuff (subunit Hc4c; Figure 18e). The massive parts of the aphyric flows may be slightly more plagioclase-phyric (up to 10% plagioclase phenocrysts, 1–2 mm in size), and show an increase in the amount and size of quartz amygdules from <2% (2 mm in size) near the base to >5% (up to 3 cm in size) close to the top. Gas cavities up to 7 cm long and 1 cm thick, partly filled with quartz, are locally observed towards the top of massive sections. Elongated to rounded, epidote-quartz alteration patches are also commonly observed in massive flows, and increase in size and abundance towards the top. Pillows are commonly irregularly shaped (amoeboid), from 0.2 to 3 m in size, and loosely packed with abundant interpillow material, including amoeboid pillow breccia. They are commonly highly amygdaloidal (up to 30% of 1–5 mm quartz-filled amygdules) and show thin (<1 cm), dark reddish-brown to dark green chilled margins and localized epidote-quartz alteration patches. Pillowed portions of the flows typically grade both vertically and laterally into amoeboid pillow breccia. Amoeboid pillow breccia is composed of poorly sorted, subangular, highly vesicular to scoriaceous fragments of amoeboid pillows. The matrix is composed of smaller broken pieces of the same material, as well as abundant angular fragments of the chilled pillow margins, plagioclase crystals and fine mafic material.

Thick aphyric to sparsely plagioclase-phyric pillowed basalt flows (subunit Hc1b) are also common in the Carlisle Lake member. They form up to 50 m thick pillowed sections of closely packed, well-shaped pillows, from 30 to 90 cm in size, with dark reddish-brown to dark green, from 1 to 2 cm thick chilled margins; locally, megapillows up to 3 m in size were observed. Pillows are commonly amygdaloidal (5–15%, 1–2 mm quartz-filled amygdules) with concentric thermal contraction cracks (Figure 18f). The larger pillows locally show radial pipe amygdules at their margins. Interpillow material typically forms <10% of the outcrops and consists of brownish fine-grained material with angular reddish-brown fragments of chilled pillow margin (hyaloclastite), or massive to well-bedded greenish mafic tuff.

Localized plagioclase-phyric pillowed flows (10% plagioclase, 1–1.5 mm) with amoeboid pillowed flow top (subunit Hc2a) are found intercalated with the aphyric to sparsely plagioclase-phyric basalt flows just east of Carlisle Lake. Pillows are closely packed, from 0.2 to 1.5 m in size, with dark reddish-brown, 0.5–1 cm thick chilled margins. They decrease in size upsection towards the west. The interpillow material is plagioclase-crystal–rich and mixed with angular reddish-brown fragments of chilled pillow margin (hyaloclastite).

Mafic volcaniclastic rocks of the Carlisle Lake member (unit Hc4), which account for ~40% of the Carlisle Lake member rocks, consist mainly of massive to crudely bedded, heterolithic, mafic volcanic breccia (subunits Hc4a, b) with lesser amounts of well-bedded, fine-grained mafic tuff (subunit Hc4c). Heterolithic mafic volcanic breccia (subunits Hc4a, b; Figure 19a) is mainly massive, clast supported and very poorly sorted near the base of the Carlisle Lake member, but becomes crudely bedded to bedded, matrix supported and displays a greater proportion of lapilli breccia and lapilli tuff higher up in the stratigraphy. The matrix of the heterolithic breccia is characteristically plagioclase-crystal rich (10–30% plagioclase, 1–5 mm in size; subunit Hc4b). Fragments are subangular to angular and range in size from <0.5 to 60 cm. All fragments are

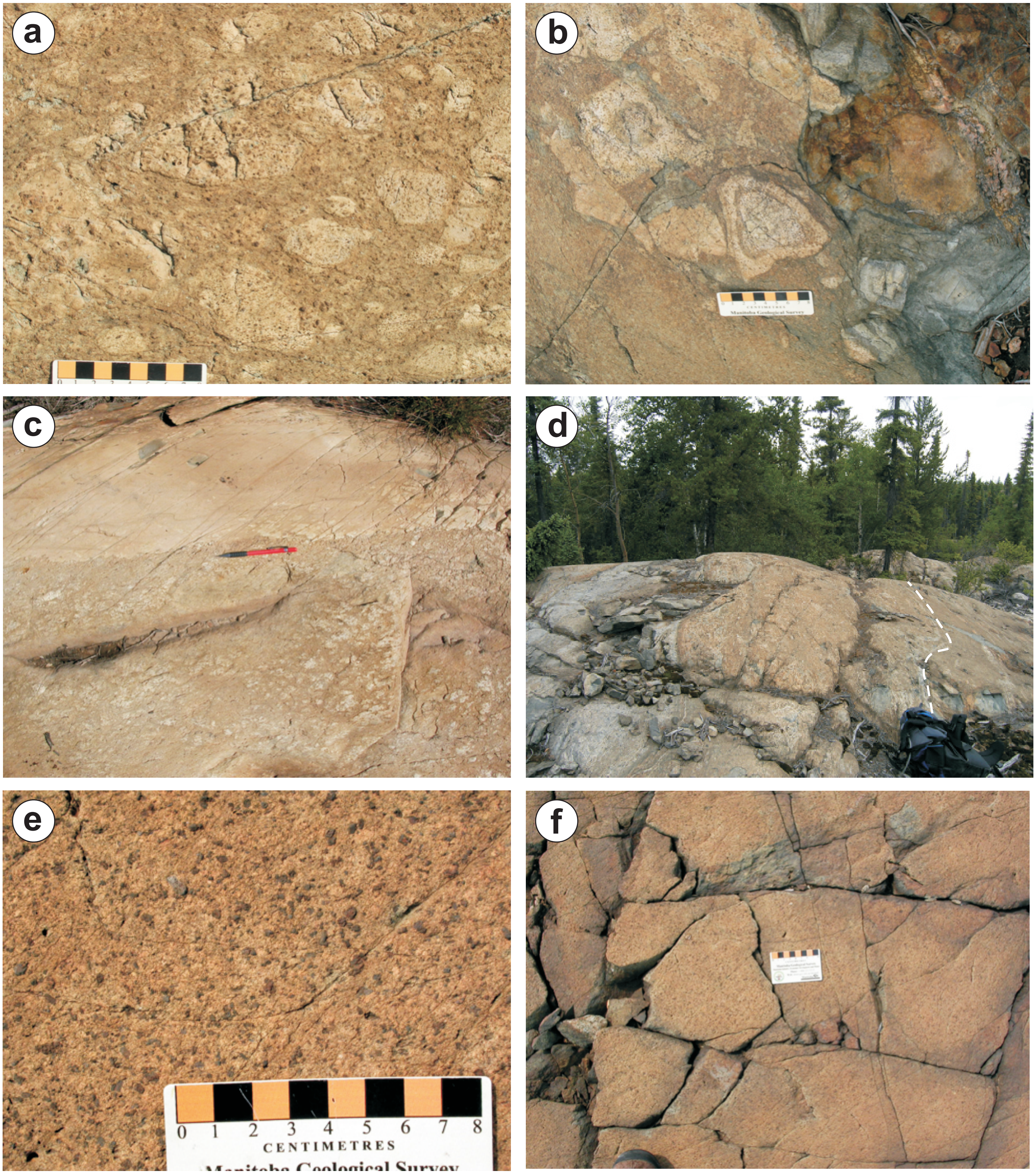


Figure 19: *Representative rock types, facies and textures of the Hidden and Louis formations:* ***a)*** *plagioclase-rich heterolithic mafic breccia of the Carlisle Lake member, Hidden formation (subunit Hc4b);* ***b)*** *heterolithic mafic breccia of the Carlisle Lake member, Hidden formation, showing felsic-cored clasts with mafic scoriaceous envelopes (subunits Hc4a, b);* ***c)*** *rhyolite-rich lapillistone bed overlain by mafic tuff of the Tower member, Louis formation (subunit Lt2a);* ***d)*** *flow-banded quartz-plagioclase–phyric rhyolite of the Tower member, Louis formation (subunit Lt1b);* ***e)*** *close-up of a massive strongly plagioclase- and pyroxene-phyric basalt flow of the Icehouse member, Louis formation (subunit Lc1a); and* ***f)*** *columnar jointing in massive strongly plagioclase- and pyroxene-phyric basalt flow of the Icehouse member, Louis formation (subunit Lc1a).*

mafic in composition but vary in texture: buff brown to beige-green aphyric basalt; buff brown to green amygdaloidal (5–35% of 0.5–4 mm quartz-filled amygdules) aphyric basalt; pillowed basalt fragments still showing part of their chilled margin; dark reddish-brown, chilled pillow-margin fragments; scoriaceous basalt fragments; rare plagioclase-phyric amygdaloidal to nonamygdaloidal basalt; and a few felsic-cored clasts with mafic scoriaceous envelopes (Figure 19b). Some localized pockets of the heterolithic breccia northeast of Carlisle Lake are not plagioclase-crystal rich (subunit Hc4a). Thin (<3 m), well-bedded, fine-grained mafic tuff intervals (subunit Hc4c) are found intercalated with the mafic flows of the Carlisle Lake member.

The rocks of the Carlisle Lake member are cut by several generations of mafic and felsic dikes throughout (units D2, D4). Southeast of Carlisle Lake, northwest-trending, thin felsic dikes run parallel to/along faults and are associated with multiple mafic dikes trending in the same direction. Feeding these mafic dikes to the east are dike complexes composed of >90% mafic dikes with <10% screens (1–5 m^2) of the hostrocks remaining (subunit D2f). These dikes locally inject interpillow spaces upsection, suggesting that their emplacement occurred prior to induration of the interpillow material. The northwest-trending felsic dikes stop just below a small quartz-plagioclase–phyric rhyolite lobe (subunit Hc3a) 250 m south of Carlisle Lake. The rhyolite lobe, which has felsic breccia associated with it, is interpreted to be part of a small extrusive dome (Simard, 2006a). A gossan zone is associated with this rhyolite lobe.

East of the Ross Lake Fault, the Carlisle Lake member varies along strike from mainly flows in the north to mainly breccia in the south. The transition is abrupt and spatially coincides with two large mafic dike complexes to the east and a small rhyolite lobe along a northwest-trending set of faults just southeast of Carlisle Lake (Simard, 2006a). North of these faults the Carlisle Lake member, which totals >500 m in thickness, is mainly composed of basalt flows with localized (<10%) lenses of heterolithic breccia and thin mafic tuffaceous intervals. South of these faults, heterolithic mafic breccia (subunits Hc4a, b) is the dominant rock type, where it constitutes >60% of the stratigraphy, and is up to 250 m thick. The abrupt transition from mainly flows in the north to mainly breccia in the south, along northwest-trending faults, suggests that these faults may be synvolcanic in origin, and form the margin of a synvolcanic subsidence structure (Simard, 2006a; DeWolfe, 2009). The mafic dike complexes and felsic magmatism are interpreted to have been emplaced along the margins of this synvolcanic structure (Simard, 2006a). The high angularity of the fragments and local preservation of fragments with delicate structures, such as scoriaceous and cored clasts, suggest that the heterolithic breccia unit underwent minimal transport (Simard, 2006a). It is interpreted to be the product of debris flows deposited in a fault-bounded subsidence structure, from local sources. The cored clasts, with scoriaceous envelopes, are typical of spatter deposits. Their proximity to the transition area suggests that the fault structures may have localized some ‘fountaining’ volcanism (Simard, 2006a; DeWolfe, 2009).

The presence of this graben-like structure just southeast of Carlisle Lake, with associated mafic and felsic magmatism, suggests potential for VMS mineralization in the Carlisle Lake member (Simard, 2006a). The presence of fine-grained volcaniclastic rocks within this sequence also provides evidence for a hiatus in effusive volcanic activity, which is interpreted to play a key role in the formation of a VMS deposit in the area.

West of the Ross Lake Fault, the Carlisle Lake member is >500 m thick and is composed of abundant heterolithic mafic breccia intercalated with aphyric to sparsely plagioclase-phyric basalt flows, and localized areas of mafic dike complex. This succession is offset along strike by several late northeast-trending faults that appear to be truncated by the Ross Lake Fault to the east (Figure 16). Overall, the heterolithic breccia fines upward and becomes more stratified to the south and west, which suggests a depositional environment similar to, but less proximal than, the one east of the Ross Lake Fault.

The Carlisle Lake member is interpreted to represent a lateral equivalent to the Reservoir member (Simard, 2006a; DeWolfe et al., 2009a; DeWolfe, 2010, 2011). The Reservoir member is interpreted to record the onset of mafic magmatism to form the Hidden shield volcano atop the Flin Flon subsidence structure, whereas the Carlisle Lake member most likely represents a subsidence structure on the flank of the main volcanic edifice (Simard, 2006a; DeWolfe et al., 2009a; DeWolfe, 2010, 2011; Figure 11d).

Louis formation (L)

(Y.M. DeWolfe and R-L. Simard)

The Louis formation (L) is composed of basalt flows and mafic volcaniclastic rocks, with subordinate rhyolite flows and felsic volcaniclastic rocks. It was deposited during a second episode of mafic hangingwall volcanism, atop the Hidden formation (DeWolfe, 2008; DeWolfe et al., 2009a; Figure 11d).

The Louis formation is exposed only on the eastern side of the Flin Flon Lake Fault (Figure 20). It is subdivided in two mappable members, the Tower member (Lt) and Icehouse member (Lc), as well as a suite of undivided basalt flows (L; DeWolfe et al., 2009a). The rocks of the Louis formation are present on both sides of the Burley Lake Fault, which defines the core of a faulted synform synclinal structure extending from Louis Lake southwards along the peninsula east of Potter Bay (Figure 3; Geoscientific Map MAP2010-1, back pocket).

The base of the Louis formation is placed at the first occurrence of rhyolite and associated volcaniclastic rocks of the Tower member (Lt; see details below). Where the Tower member is not present, the base of the Louis formation is defined by the first occurrence of plagioclase-pyroxene–phyric basalt flows. The top of the Louis formation is defined by the basal unconformity of the Missi Group to the north, and by the present-day erosion surface in the south.

Tower member (Lt)

The Tower member (Lt) occurs at the base of the Louis formation on both side of the Burley Lake Fault where it forms oppositely facing, steeply dipping successions <75 m thick (apparent thickness) that can be traced along strike for >4 km (Figure 20). It consists of massive to in situ–brecciated, aphyric

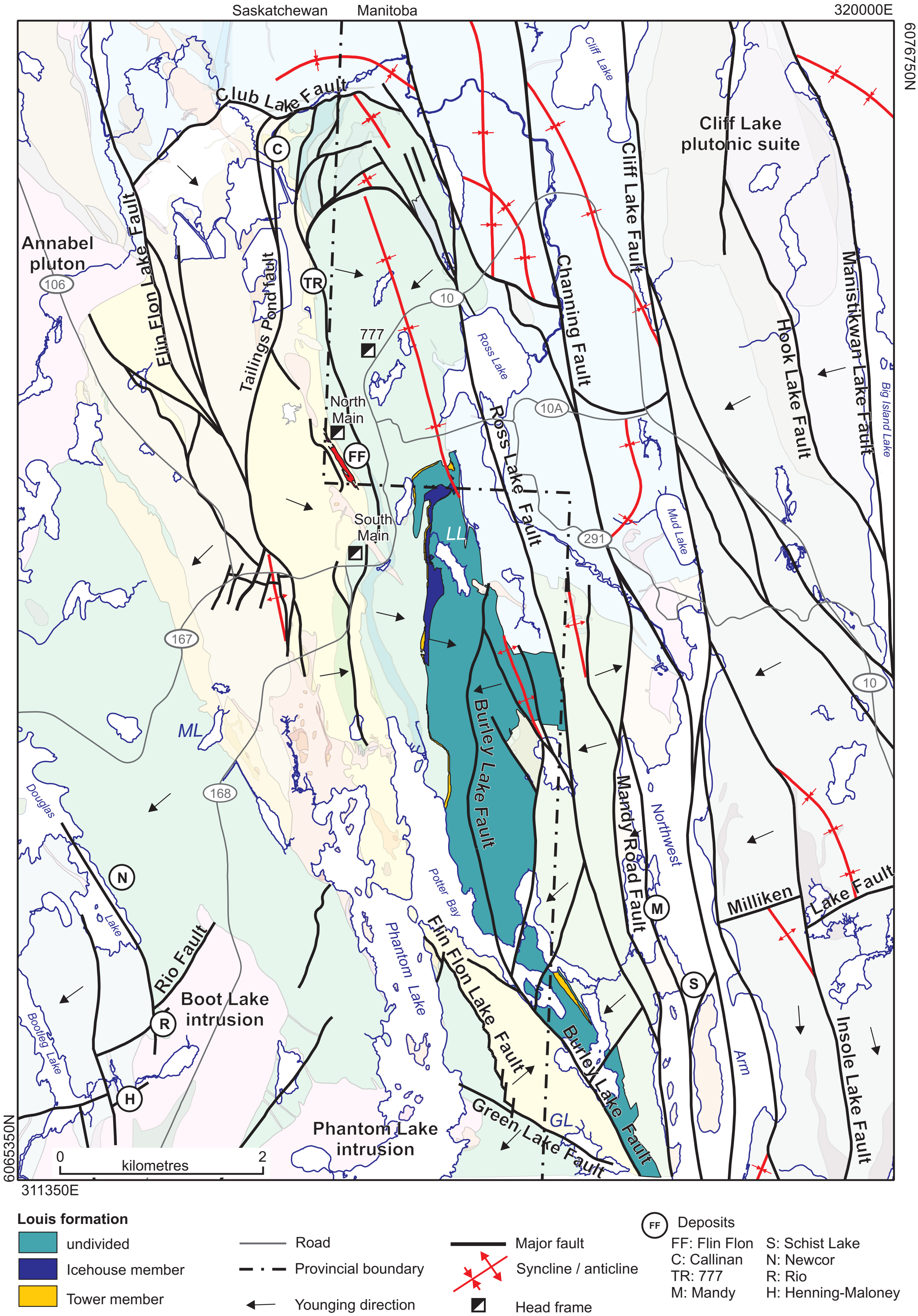


Figure 20: *Simplified geology of the Flin Flon area, showing the two mappable members of the Louis formation and major structures in the area. For complete legend refer to Figure 3. Abbreviations: GL, Green Lake; LL, Louis Lake; ML, Myo Lake.*

or sparsely plagioclase- and quartz-phyric rhyolite (unit Lt1) overlain by, and locally underlain by, mafic tuff (unit Lt2; Figure 19c; DeWolfe, 2008; DeWolfe et al., 2009a). Where the coherent rhyolite portion of the Tower member is not present, the member is defined by a laterally extensive mafic tuff, which includes lapilli-tuff beds containing rhyolite fragments derived from the coherent rhyolite (unit Lt2; see details below; DeWolfe, 2008; DeWolfe et al., 2009a).

The coherent portion of the Tower member rhyolite (unit Lt1) is areally restricted, occurring only in the areas north of Louis Lake and north of Burley Lake. The massive to brecciated rhyolite (subunits Lt1a, b) is commonly light brown to beige on weathered surfaces and grey on fresh surfaces. North of Louis Lake, it is aphyric to sparsely plagioclase- and quartz-phyric (subunit Lt1a), locally flow banded, strongly silicified and locally sericitized, and ranges from <1 to 20 m thick (apparent thickness), with in situ breccia at both its upper and lower contacts (DeWolfe, 2008; DeWolfe et al., 2009a). North of Burley Lake, the rhyolite unit (subunit Lt1b) is plagioclase- (<10%, 1–1.5 mm) and quartz-phyric (<5%, 1–1.5 mm), and is up to 75 m thick with at least two strongly silicified flows, 20–30 m thick, each displaying internal in situ breccia and flow banding (Figure 19d; Simard, 2006a). The in situ breccias are interpreted as autoclastic breccia (Simard, 2006a; DeWolfe, 2008). The rhyolite is interpreted to have been emplaced as a flow or dome, with associated autoclastic breccia, at the onset of a new magmatic episode in the hangingwall of the Flin Flon–Callinan–777 VMS deposits (DeWolfe, 2008). North of Louis Lake, the synvolcanic structures associated with the emplacement of the Tower member rhyolite flow or dome (Figure 11d), and the overlying Icehouse member flows (see 'Icehouse member' below) define a structural corridor similar to the one hosting the main Flin Flon massive sulphide orebody (DeWolfe, 2008).

Mafic tuff found under the rhyolite north of Burley Lake (subunit L4b) is composed of finely laminated mafic mudstone and siltstone, <10 m thick, and laterally continuous over more than 500 m (Simard, 2006a). Mafic tuff overlying the rhyolite north of Louis Lake is finely laminated to massive and contains lapilli-tuff beds with rounded lapilli of rhyodacite identical to the Tower rhyolite (<1–10 cm, <80%; DeWolfe, 2008). Along strike, the mafic tuff contains between 5 and 90% of these lapilli-tuff beds intercalated with mafic tuff. Plane-parallel bedding within the mafic tuff (subunit L4b) is interpreted to be a product of suspension sedimentation, whereas the massive lapilli-tuff beds with rhyolite clasts (subunit Lt2a) are interpreted as mass flow units shed from the adjacent massive coherent facies of the Tower rhyolite (DeWolfe, 2008).

Icehouse member (Lc)

The Icehouse member (Lc) is only found on the western side of the Burley Lake Fault where it forms a steeply dipping east-facing sequence from 25 to 100 m thick (apparent thickness) that can be traced along strike for >1 km west of Louis Lake (Figure 20). The Icehouse member forms a distinctive succession of strongly plagioclase-pyroxene–phyric basalt flows (unit Lc1) and mafic volcaniclastic rocks (unit Lc2) near the base of the Louis formation.

The basalt flows of the Icehouse member (unit Lc1; Figure 19e) are commonly brown on weathered surfaces and dark green on fresh surfaces, and are characterized by their high phenocryst content (~20–25% plagioclase phenocrysts, 1–5 mm in size; 15–20% pyroxene phenocrysts, 1–8 mm in size). From north to south, the Icehouse member changes from a thick (~100 m) single facies massive flow to a thin (~25 m) multifacies flow with a massive bottom and a pillowed top (DeWolfe, 2008; DeWolfe et al., 2009a). The massive facies (subunit Lc1a) is commonly columnar jointed (Figure 19f), does not have a flow-top breccia, and has a flow top marked by an increase in quartz-amygdule content from 5 to 30% (<1–5 cm) near the top. It is conformably overlain by a finely laminated, strongly epidote-quartz–altered mafic tuff to lapilli tuff (subunit Lc2b). There is a gradational contact along strike where the massive facies grades into pillowed facies (subunit Lc1b) over a distance of 1–2 m. Pillows are 0.2–1 m in size, have thin (≤2 cm) selvages and contain 5–10% quartz amygdules (<1 cm). Finely laminated epidote-quartz–altered mafic tuff commonly occurs between the pillows. The upper margin of the pillowed facies of the Icehouse member is irregular and broken, and grades over a distance of 1 m from intact pillows, to pillow breccia, to an overlying volcaniclastic facies.

North of Louis Lake, overlying the massive facies, is a plane-bedded mafic tuff or mafic lapillistone (subunit Lc2b). The mafic tuff is ≤2 m thick and is strongly silicified. The mafic lapillistone is ≤4 m thick, massive and well sorted, containing 70% well-rounded, plagioclase- and pyroxene-phyric basalt clasts in a mafic tuff matrix. West of Louis Lake, overlying the pillowed facies, is a 20 m thick, normally graded, crudely bedded, heterolithic volcaniclastic unit (subunit Lc2a). Overall, the volcaniclastic unit contains 50% lapilli-sized, rounded, aphyric to plagioclase- and pyroxene-phyric basalt clasts. The tuff-sized matrix is reddish-brown in colour and contains abundant plagioclase (20–25%) and pyroxene (15%) crystals. The lower 5 m is a clast-supported tuff breccia bed, containing 20% large blocks (6.4–50 cm) of plagioclase- and pyroxene-phyric, quartz-amygdaloidal basalt, and 5–10% angular to subrounded blocks of aphyric rhyolite. The overlying interval marks a 3–5 m thick transition from a clast-supported tuff breccia bed to a matrix-supported tuff breccia bed. In this interval, there are only 5% large plagioclase-pyroxene–phyric pillow fragments (>10 cm); 3–5% aphyric rhyolite clasts (rounded and ≤15 cm); and 10–20% lapilli-sized, aphyric to plagioclase-pyroxene–phyric basalt fragments. The upper 10 m of the volcaniclastic unit is a matrix-supported lapillistone bed. The lapillistone bed contains 30% rounded, lapilli-sized basalt clasts and 5–10% subrounded, aphyric rhyolite clasts in a tuff matrix. The plane-bedded mafic tuff north of Louis Lake is interpreted to be the product of suspension sedimentation from the overlying water column (DeWolfe, 2008). The heterolithic tuff breccia to lapillistone unit west of Louis Lake is interpreted to be mass flow(s) deposited in a localized basin (DeWolfe, 2008; DeWolfe et al., 2009a).

Overall, the massive Icehouse member flow is interpreted to have formed from the ponding of lava within a fault-bounded graben parallel to the feeding fissure (DeWolfe, 2008; DeWolfe et al., 2009a). The Icehouse member thins to the south where it grades laterally and vertically from massive facies into pillowed facies as the flow becomes thinner. This thinning of the flow and change to pillowed facies along strike could be the result of the flow overriding the wall of the graben that ponded the Icehouse member flows to the north (DeWolfe, 2008; DeWolfe et al., 2009a). Where the areally restricted, heterolithic, tuff breccia to lapillistone mass flow unit overlies the thick Icehouse member flow, the former is interpreted to have been deposited within the graben associated with the subsidence and volcanism during emplacement of the massive Icehouse member flows. Subsidence that exposed the underlying rhyolite (Tower member) may have triggered mass flows, resulting in a localized tuff breccia–lapillistone deposit, composed of basalt and Tower member-derived rhyolite detritus deposited in a fault-controlled basin (DeWolfe, 2008; DeWolfe et al., 2009a).

Undivided Louis formation rocks (L)

The undivided Louis formation rocks (L) account for >90% of the Louis formation, and are present on both sides of the Burley Lake Fault (Figure 20). They form oppositely facing steeply dipping sequences >500 m thick (apparent thickness) that can be traced along strike for >12 km in the core of this faulted synclinal structure from Louis Lake to Green Lake (Figure 20; Geoscientific Map MAP2010-1, in back pocket). The undivided Louis formation rocks consist of aphyric to sparsely plagioclase-phyric (unit L1), plagioclase-phyric (unit L2), and plagioclase-pyroxene–phyric (unit L3) basalt flows intercalated with subordinate amounts of mafic volcaniclastic rocks (unit L4). West of the Burley Lake Fault, the undivided Louis formation rocks sit conformably over either the mafic volcaniclastic unit of the Icehouse member, mafic tuff of the Tower member, or aphyric flows of the Reservoir member of the Hidden formation (Figures 3, 16, 20). On the eastern side of the Burley Lake Fault, they rest directly on aphyric flows of the Carlisle Lake member of the Hidden formation (Figures 3, 16, 20). The top of the undivided Louis formation rocks is represented by the basal unconformity of the Missi Group to the north or the present-day erosion surface to the south.

The basalt flows of the undivided Louis formation rocks are mainly plagioclase-pyroxene–phyric (unit L3), with lesser amount of aphyric to sparsely plagioclase-phyric (unit L1) and plagioclase-phyric (unit L2) flows. They are commonly brown on weathered surfaces and dark green on fresh surfaces, with individual flows commonly ranging from 10 to 50 m thick in the north (DeWolfe, 2008; DeWolfe et al., 2009a), and up to 100 m thick in the south. The plagioclase (>15%, 1–3 mm)-pyroxene (>5%, 1–4 mm)–phyric basalt flows (unit L3) consist of massive and pillowed facies. The massive flows (subunit L3a) account for ~25% of the flows, most of which are pillowed. Locally, flows grade upward from massive facies into flow-top breccias (subunits L3a, d); more commonly there is no flow-top breccia and the top of the massive flows are defined by the presence of volcaniclastic rocks (see below; DeWolfe, 2008; DeWolfe et al., 2009a). The pillowed flows (subunit L3b; Figure 21a) are typically amygdaloidal (0.1–2 cm quartz-filled amygdules), with pillows 0.5–2.5 m in diameter, commonly with epidote-quartz–altered mafic tuff between. Pillows commonly have thin (<1 cm), dark brown to red chilled margins. On the peninsula between Burley Lake and Potter Bay of Phantom Lake, up to 100 m thick coarse-grained plagioclase-pyroxene–phyric massive flows, with thin pillowed and/or amoeboid flow-top breccia (subunit L3d), sits conformably atop the Tower member rhyolite (Simard, 2006a).

Mafic volcaniclastic units (unit L4; Figure 21b) typically separate pillowed and massive basalt flows. They range from <1 to 5 m in thickness, and consist of clast-supported, massive lapilli tuff beds (subunit L4a) to finely laminated, plane-bedded mafic tuff beds (subunit L4b). Lapilli tuff beds contain up to 70% plagioclase-pyroxene–phyric, amygdaloidal basalt clasts that are rounded to subangular in shape and range in size from 0.2 to 6 cm, with an average clast size of 2 cm; they also contain 10% aphanitic basalt clasts (0.2–3 cm, subrounded to subangular), and both clast types occur in a grey to brown, tuff-sized matrix (DeWolfe, 2008; DeWolfe et al., 2009a).

Between Louis Lake and the northern shore of Phantom Lake, the undivided Louis formation rocks are composed of >90% plagioclase-pyroxene basalt flows intercalated with thin mafic volcaniclastic units. These basalt flows become thinner to the south, which suggests a source to the north, with a more distal environment in the south (DeWolfe, 2008; DeWolfe et al., 2009a).

Farther south and east of the Burley Lake Fault, on the peninsula between Burley Lake and Potter Bay of Phantom Lake, the undivided Louis formation rocks are composed of>70% thick plagioclase-pyroxene basalt flows intercalated with thin mafic volcaniclastic units, along with ~25% thick, aphyric to sparsely plagioclase-phyric and plagioclase-phyric basalt flows. The average flow thickness on this peninsula is >50 m, up to 100 m in places, which suggests a very proximal setting to the vent. Similar to the area north of Louis Lake, where the emplacement of the Tower member rhyolite flow or dome corresponds with the thickest part of the overlying Icehouse member flows, the very thick basalt flows of the undivided Louis formation are emplaced directly atop the >50 m thick pile of rhyolite flows of the Tower member on the eastern shore of the peninsula. Furthermore, the synvolcanic structures inferred for the emplacement of the Tower member rhyolite flows and the overlying thick basalt flows on this peninsula are located directly above the subsidence structure within the rocks of the Carlisle Lake member to the east.

Douglas formation (G)

(from Thomas, 1989)

The Douglas formation (G) is composed predominantly of volcaniclastic rocks intercalated with minor amounts of mafic flows (previously referred to as the Douglas Lake assemblage, see Table 1, back pocket, for details; Thomas, 1989). It has been interpreted as a possible time correlative unit to the Louis formation on the eastern side of the Flin Flon Lake Fault, but could also be equivalent to the Carlisle Lake member of the Hidden formation (Figure 7).

The Douglas formation is only exposed on the western side of the Flin Flon Lake Fault, south and west of Douglas Lake. It forms a steeply dipping homoclinal southwest-facing succession >2.5 km thick (apparent thickness, including potential structural

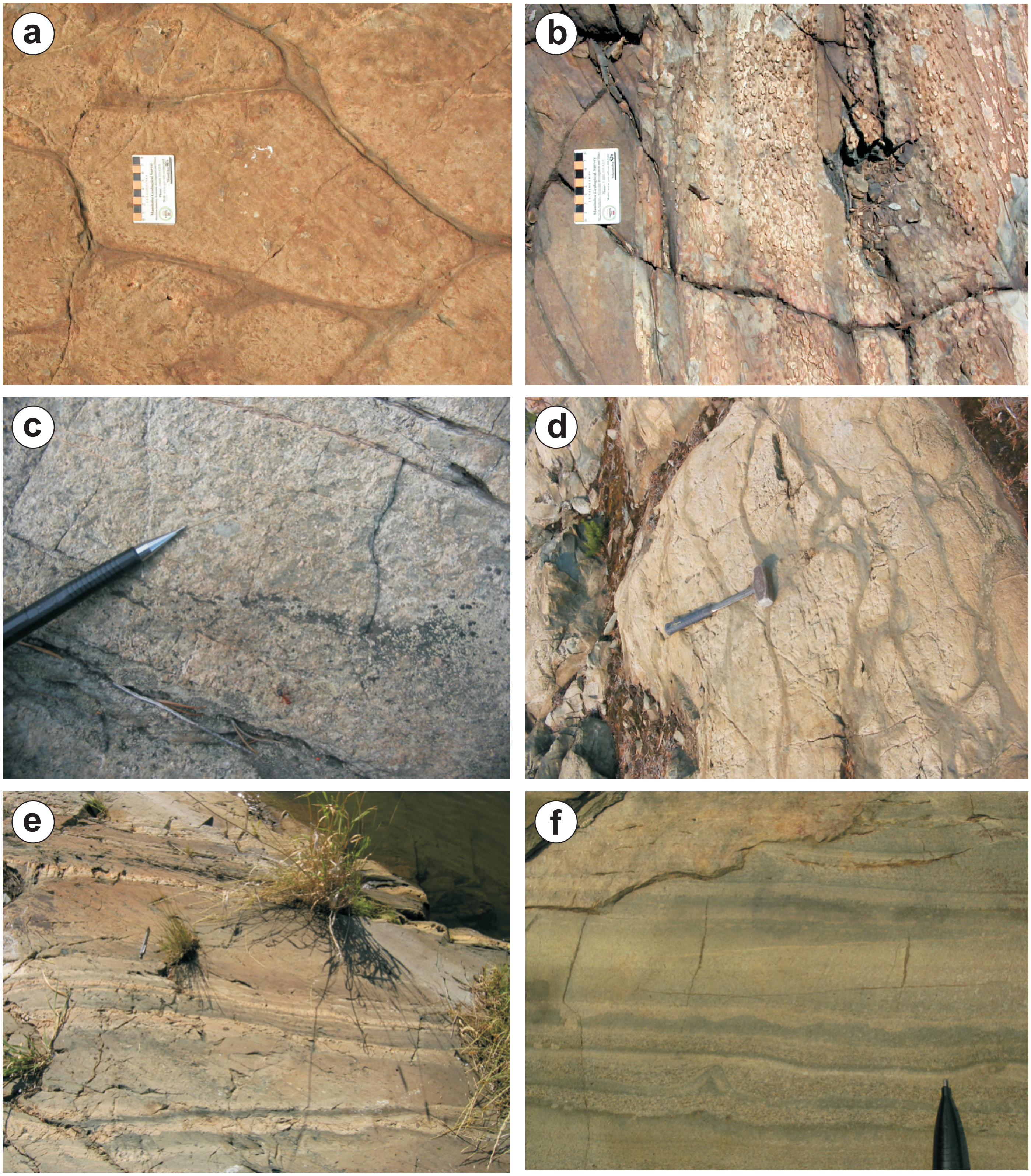


Figure 21: *Representative rock types, facies and textures of the Louis and Douglas formations, and of unassigned volcanic rocks between the Channing–Mandy Road and Cliff Lake faults:* ***a)*** *plagioclase- and pyroxene-phyric pillowed basalt of the undivided Louis formation rocks (subunit L3b);* ***b)*** *mafic tuff with accretionary lapilli of the undivided Louis formation rocks (unit L4);* ***c)*** *heterolithic plagioclase-crystal–rich mafic lapilli tuff of the Douglas formation (subunit G4a);* ***d)*** *altered plagioclase-phyric pillowed basalt flows of the Douglas formation (subunit G2a);* ***e)*** *well-bedded mafic tuff of the unassigned volcanic rocks between the Channing–Mandy Road and Cliff Lake faults (subunit U3b); and* ***f)*** *well-bedded mafic tuff of the unassigned volcanic rocks between the Channing–Mandy Road and Cliff Lake faults showing normally graded beds and load structures (subunit U3b).*

thickening), which is cut to the southeast by the Boot Lake–Phantom Lake intrusion (unit S3). The Douglas formation is interpreted to stratigraphically overlay the Reservoir member of the Hidden formation along the eastern shore of Douglas Lake, however, the contact is highly tectonized and altered by a late north-northwest-trending shear zone that completely obliterated the contact in this area (Figure 3; Thomas, 1989).

The Douglas Lake formation (G) consists of mafic (unit G3) and heterolithic (unit G4) volcaniclastic rocks, intercalated with minor aphyric (unit G1) and plagioclase-phyric (unit G2) basalt flows.

The base of the Douglas formation is characterized by a >400 m thick package of heterolithic breccia locally interbedded with minor tuff breccia and lapilli tuff (subunit G4a, b; Figure 21c). These volcaniclastic rocks are clast- to matrix-supported, and composed predominantly of variably amygdaloidal aphyric and plagioclase (±pyroxene)-phyric basalt fragments, with rare aphyric rhyolite and silicified amygdaloidal basalt fragments. The matrix in places is characteristically plagioclase-crystal–rich (15–20%, 1–4 mm; subunit G4a). Intercalated with these crystal-rich volcaniclastic rocks are variably plagioclase-phyric basalt flows (unit G2; Figure 21d). They are commonly light green to buff on weathered surfaces and dark green on fresh surfaces, and are sparsely to strongly plagioclase-phyric (1–10%, 1–3 mm). These plagioclase-phyric basalt flows are <100 m thick and usually pillowed (subunit G2a) with amoeboid flow tops, which are both heterogeneously silicified and commonly display quartz-epidote alteration patches.

On the western side of the Douglas Lake Fault, the Douglas Lake formation is composed of a >400 m thick package of mafic tuff, either finely laminated to thickly bedded tuff and lapilli tuff (subunit G3b) or massive to crudely bedded tuff that is pyroxene-plagioclase–crystal rich (subunit G3a), intercalated with heterolithic breccia with abundant felsic volcanic clasts in a plagioclase-crystal–rich matrix (subunit G4c; Thomas, 1989). Overlying these volcaniclastic rocks are about 100 m of extensively silicified plagioclase-phyric amygdaloidal pillowed basalt flows (Thomas, 1989). These flows are in turn overlain by a second package of mafic volcaniclastic rocks (subunits G3a, b) and heterolithic breccia with abundant felsic clasts (outside map area to the northwest). West of Bootleg Lake, aphyric, pillowed basalt flows (unit G1) are intercalated within the mafic volcaniclastic rocks (unit G3).

The volume of breccia and crudely bedded lapilli tuff and tuff within the Douglas formation suggests that these rocks are the result of debris flows (Thomas, 1989). Potentially, if the contact between the Douglas and Hidden formations is conformable, these volcaniclastic rocks could be derived in part from the underlying Reservoir member rocks within a volcanic subsidence structure, a setting similar to the Carlisle Lake member of the Hidden formation. The intense silicification observed in the plagioclase-phyric basalt flows of the Douglas formation (subunit G2a), intercalated within volcaniclastic beds between Douglas and Bootleg lakes as well as the volcaniclastic beds observed in some of the uppermost flows of the Hidden formation, just east of Douglas Lake, suggests that the rocks of the Douglas formation were likely deposited within a long-lived subsidence structure in proximity to an active hydrothermal system, which would suggest potential for VMS mineralization in this area (MacLachlan, 2006c).

Hamell Lake volcanic rocks (HL)

(from Thomas, 1993)

The Hamell Lake volcanic rocks (HL) consist predominantly of felsic and mafic flows intercalated with volcaniclastic rocks, all of which are intruded to the southwest by the Annabel pluton. The Hamell Lake volcanic rocks are exposed north of the Annabel pluton and west of the Flin Flon Lake Fault and Club Lake Fault (Figure 3), where they extend for >10 km to the northwest beyond the boundary of the map.

According to Thomas (1993), the lowest stratigraphic unit in the Hamell Lake volcanic rocks is the Hamell Lake rhyolite (unit HL3), which is composed mainly of coherent, locally flow-banded, aphanitic to quartz-plagioclase–phyric rhyolite. This felsic unit is stratigraphically overlain by the Hamell Lake volcaniclastic unit (unit HL4), which consists of poorly bedded to massive, heterolithic, matrix-supported tuff breccia to lapilli tuff, composed mainly of mafic clasts, with <5% of felsic volcanic clasts. Felsic fragments are particularly abundant in the lower part of the unit, so that locally, a narrow discontinuous, 1–10 m thick, clast- to matrix-supported felsic tuff breccia overlies the Hamell Lake rhyolite. The upper part of the Hamell Lake volcaniclastic unit is characterized by an extremely low percentage of felsic volcanic clasts and an increasing amount of amygdaloidal mafic volcanic clasts. A mixed sequence of plagioclase-phyric (unit HL2) and aphyric (unit HL1) basalt flows with minor plagioclase-crystal–rich mafic tuff, the Hamell Lake basalt unit, overlies the Hamell Lake volcaniclastic unit. These flows are mainly pillowed, locally massive, commonly silicified and commonly grade into amoeboid pillows and pillow breccia.

The overall composition and internal facies variation characterizing the Hamell Lake basalt unit, and its association with felsic volcanic rocks, are similar to the footwall of the Flin Flon–Callinan–777 VMS deposits east of the Flin Flon Lake Fault, suggesting that these rocks may be stratigraphically equivalent to the Flin Flon formation (Thomas, 1993).

Between Channing–Mandy Road faults and the Cliff Lake Fault

(R-L. Simard, Y.M. DeWolfe and H. L. Gibson)

Rocks between the Channing–Mandy Road faults and the Cliff Lake Fault host the Schist Lake and Mandy VMS deposits. Although these rocks are of economic significance, they have not been easily correlated with any single stratigraphic unit in the Flin Flon camp at this stage (Simard, 2006a). More field and analytical work is underway to clarify this issue (Cole et al., 2007, 2008; DeWolfe, 2009, 2010, 2011). They were not assigned a formal name, therefore are referred to as 'unassigned volcanic rocks' (U).

Unassigned volcanic rocks (U)

These unassigned volcanic rocks (U) are dominated by heterolithic mafic breccia, heterolithic mafic and felsic breccia, with minor felsic breccia, and subordinate amounts of basalt flows and late felsic intrusive rocks (Simard, 2006a; Cole et al., 2007, 2008; Simard and Creaser, 2007; DeWolfe, 2009, 2010, 2011).

The internal stratigraphy of these rocks has been disrupted by two or more north-trending faults and their associated splays. These faults are not well exposed because they are either beneath a local access gravel road or beneath the Northwest Arm of Schist Lake (Figure 3). Between the Cliff Lake Fault to the east and the West Mandy Road Fault to the west, at least three north-trending vertical to subvertical successions of volcanic rocks have been recognized (see Geoscientific Map MAP2010-1, in back pocket). Between fault splays they form elongated lozenges of coherent stratigraphy that vary in thickness from <150 to >500 m thick (apparent thickness) and that can be traced along strike from <1.5 to >5 km (Figure 3). The lack of continuous rock exposure south of the Schist Lake deposit, prevents direct correlation with the VMS host stratigraphy to the north.

The westernmost of these successions, between the West and East Mandy Road faults, is unique in the area as it is dominated by well-bedded mafic tuff (unit U3b). It forms a steeply east-dipping, west-facing succession <200 m thick (apparent thickness, including potential structural thickening). This mafic tuff (unit U3b; Figure 21e) succession consists of interbedded, finely laminated mafic tuff beds (10–60 cm), massive mafic tuff beds (30–100 cm) and plagioclase-crystal–rich tuff beds (10–100 cm; Figure 21f; DeWolfe, 2009). Bedded units within these mafic tuffs are commonly parallel laminated and seldom crosslaminated (DeWolfe, 2009). The tuff succession grades laterally northward into heterolithic, mafic, clast-supported lapilli breccia to breccia (unit U3a), which is massive, poorly sorted and commonly composed of angular to subangular mafic clasts that range in size from 1 to 50 cm (Simard, 2006a; DeWolfe, 2010). The clasts are composed of multiple types of basalt (aphyric, plagioclase-phyric, variously amygdaloidal), and seldom show evidence of chilled margins (pillow fragments). The matrix, which is typically plagioclase-crystal rich, is poorly sorted and composed of lapilli-sized fragments of similar composition to that of the contained clasts (Simard, 2006a; DeWolfe, 2010). This tuffaceous succession is cut by numerous mafic sills and dikes (unit D2), especially to the north; some of the mafic dikes within the tuff display peperitic margins, suggesting they were emplaced when the tuff was wet and unconsolidated (DeWolfe, 2009). Felsic dikes are also present in the sequence, but they are less common and do not display peperitic margins. Some of the felsic dikes are folded suggesting they predate the deformation, whereas others cut across some of the late structures, suggesting a later episode of felsic magmatism in the area. Although there is no direct exposure of either fault, the succession is interpreted as fault bounded (Figure 3).

The rocks between the East Mandy Road Fault and the Cliff Lake Fault form two steeply dipping successions dominated by heterolithic breccias (subunits U3a, U4a). The two successions have been interpreted to be west facing, although no younging indicators have been documented in the rocks of the easternmost succession. They are separated by an unnamed north-northwest-trending shear zone, which runs under the southern part of the access road to the former Mandy and Schist Lake minesites, and host most of the mineralization of the Schist Lake deposit.

West of the unnamed fault, primary features in the rocks are generally well preserved and show very little alteration, except for small localized patches of iron staining, sericitization and hematization (Simard, 2006a; Cole et al., 2007; DeWolfe, 2011). They consist mainly of massive to well-bedded, poorly sorted clast- to matrix-supported lapilli breccia to breccia with lapilli to tuffaceous bed intervals (unit U3a; Figure 22a; Simard, 2006a; DeWolfe, 2011). Clasts in this breccia unit are mafic in composition (aphyric, porphyritic, variably amygdaloidal, with chilled margins in places), angular to subrounded and range in size from 0.5 to 60 cm in clast-supported beds and from 3 to 15 cm in matrix-supported beds. The matrix is poorly sorted and composed of lapilli and finer fragments of composition similar to that of the larger clasts. A package (150 m thick) of aphyric to sparsely plagioclase-phyric, massive, pillowed and amoeboid basalt flows (subunit U1b; Figure 22b) is exposed north of the Mandy deposit (Simard, 2006a). The pillows are amygdaloidal and show very irregular shapes, with up to 30% interpillow mafic volcaniclastic rocks with well-developed peperitic texture in places (Simard, 2006a). Some well-bedded mafic tuff (subunit U3b) is intercalated with the breccia west of the Schist Lake deposit (Cole et al., 2008; DeWolfe, 2011). Two small felsic bodies intrude the breccia just west of the unnamed fault and the Schist Lake deposit.

East of the unnamed fault, the rocks are pervasively chloritized, sericitized (Cole et al., 2007, 2008; DeWolfe, 2011) and strongly foliated, with two to three well-developed foliations that have rendered most of the primary textures of the rocks indistinguishable (Figure 22c). The rocks consist mainly of heterolithic mafic breccia (subunit U3a; Figure 22d), with localized heterolithic mafic and felsic breccia (subunit U4a; Figure 22e); but given the degree of alteration and deformation, only locally can the rock types of the clasts within these breccias be determined with more confidence. Where exposed, the heterolithic mafic and felsic breccia unit (subunit U4a) consists of massive, poorly sorted, clast-supported breccia with subangular to subrounded clasts of basalt (aphyric, porphyritic, variably amygdaloidal, with chilled margins in places) and angular clasts of rhyolite (aphyric, plagioclase-phyric; Simard, 2006a). The felsic clasts usually represent <15% of the clast content. The matrix to the breccia is commonly chloritic and rich in plagioclase crystals (up to 30%, 1–4 mm in size). Near the Mandy and Schist Lake deposits, rocks are heavily foliated (schistose) and bleached by the sericite alteration (white-weathered surfaces), limiting documentation of the primary composition of the clast content of the breccia (Figure 22c). Although these rocks were previously mapped mainly as heterolithic felsic breccia, recent alteration work by E.M. Lewis (M.Sc. thesis, Laurentian University) suggests that most of the hostrocks are more mafic in composition (Cole et al., 2008; Lewis 2012). Along the shoreline of Schist Lake, ~300 m north of the Schist Lake mine shaft, a stockwork of chlorite veining is particularly well developed. It forms 5–25 cm wide veins that extend several metres in length and criss-cross each other to form a stockwork covering ~20 m^2 (Figure 22f; Bailes and Syme, 1989; Cole et al., 2007; DeWolfe, 2011). Strongly

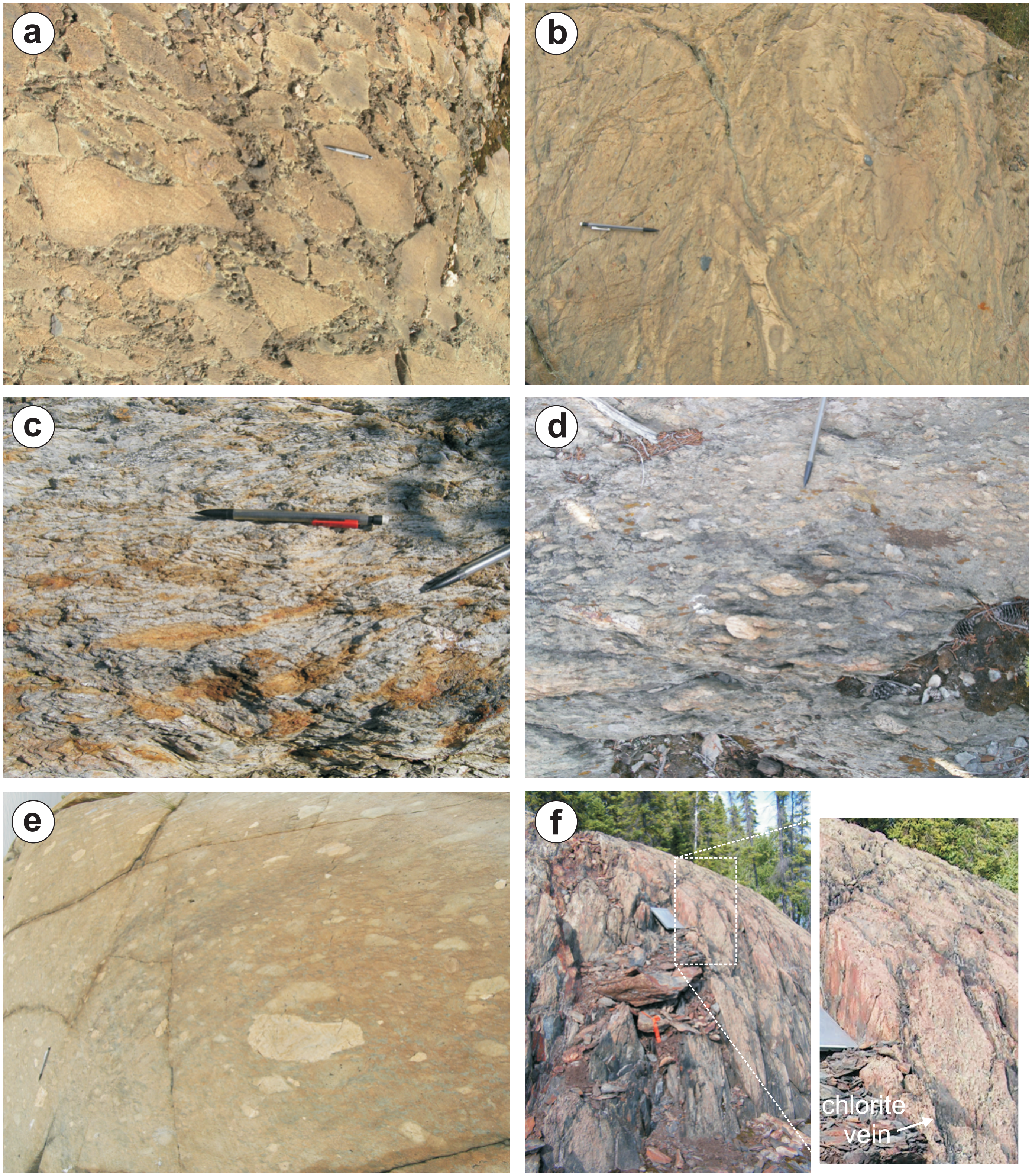


*Figure 22: Representative rock types, facies and textures of the unassigned volcanic rocks between the Channing–Mandy and Cliff Lake faults: **a)** heterolithic mafic breccia (subunit U3a); **b)** aphyric to sparsely plagioclase-phyric pillowed basalt flow (subunit U1b); **c)** well-developed foliations in a sericitized and silicified heterolithic mafic breccia (subunit U3a); **d)** sericitized, chloritized and foliated heterolithic mafic breccia (subunit U3a); **e)** heterolithic mafic and felsic breccia (subunit U4a); and **f)** stockwork chlorite veins in foliated heterolithic mafic volcanic breccia (subunit U3a).*

weathered massive sulphide stringers 0.1 to 3.5 m in width are exposed at surface halfway between Mandy and Schist Lake mines (DeWolfe, 2011). Each sulphide stringer has an associated alteration zone of intense chloritization and quartz-stockwork veining in the surrounding hostrocks (DeWolfe, 2011). These features are typical of those found in the footwall of many VMS deposits (Grifkins et al., 2005), which suggest that this easternmost sequence, east of the unnamed fault, most likely contains the footwall strata to the Schist Lake and Mandy deposits (C.B. Ko and J. Percel, pers. comm., 1987; Bailes and Syme, 1989; Cole et al., 2007; DeWolfe, 2011). For more details on the Schist Lake and Mandy deposits, see 'Economic Geology' below.

Most heterolithic breccia deposits, between the Channing–Mandy Road faults and the Cliff Lake Fault, display crude layering and angular to subrounded clasts, which suggests a moderate amount of transport from their source (Fisher and Schmincke, 1984). The observed thickness of these breccia successions is commonly ≥200 m suggesting they are the product of debris flows derived from local sources and deposited in a subsidence structure (Bailes and Syme, 1989; Simard, 2006a; Cole et al., 2007; DeWolfe, 2011). It remains to be determined whether all of these volcaniclastic successions were deposited in one bigger subsidence structure hosting the Schist Lake and Mandy VMS deposits that was later dismembered by faults, which would explain the variation in alteration and clasts content between the various successions, or whether they were deposited in unrelated smaller subsidence structures, one of which hosting the VMS deposits, that were later juxtaposed by faulting.

The precise stratigraphic relationship of these various volcanic rock successions with the adjacent rocks to the west and east also remains unclear. Similar stratigraphy and depositional volcanic environments have been documented both to the west, in the Flin Flon and Hidden formations, and to the east in the Western Hook Lake succession (Devine, 2003; Kremer and Simard, 2007; Simard and Creaser, 2007; DeWolfe, 2009, 2010, 2011; DeWolfe et al., 2009a; Gibson et al., 2009).

Stitt Island

(from Bailes and Syme, 1989)

Stitt Island, just south of the Schist Lake deposit, is underlain by a large body of variably carbonatized dacite. The dacite weathers white to buff and is light olive to pinkish on fresh surfaces. It is slightly quartz-phyric with 0–1% mauve quartz phenocrysts 0.5–1 mm in size. The dacite is commonly massive or weakly brecciated; polygonal columnar joints are exposed along 240 m of the central eastern shore of Stitt Island. The dacite is interpreted to be fault bounded, on the basis of its oblique trend with respect to adjacent units to the northwest, and the presence of extremely schistose zones on the southwestern, southeastern and northeastern shorelines. Its precise stratigraphic relationship to the heterolithic breccia that hosts the Schist Lake and Mandy mines is therefore unknown. Virtually all the dacite on Stitt Island has undergone some degree of carbonatization, ranging from relatively weakly developed carbonate blastesis and open-space filling by quartz-carbonate-hematite, to intense brecciation and veining by a high proportion of purple- to orange-weathering Fe-carbonate.

East of the Cliff Lake Fault

Volcanic rocks east of the Cliff Lake Fault also form part of the Flin Flon arc assemblage, and can be subdivided into three major fault-bounded successions: the Western Hook Lake succession (Vw) between the Cliff Lake and Hook Lake faults, the Eastern Hook Lake succession (Ve) between the Hook Lake and Manistikwan Lake faults, and the Cope Lake volcanic rocks (C) east of the Manistikwan Lake Fault. The first two successions were first described by Bailes and Syme (1989) as part of the Hook Lake Block, and the latter as part of the Cope Lake Block. The bounding faults are interpreted to have originated as west-verging thrust faults that have subsequently been reactivated as strike-slip faults.

Western Hook Lake succession (Vw)

(R-L. Simard and P.D. Kremer; and from Bailes and Syme, 1989)

The Western Hook Lake succession (Vw) is exposed between the Cliff Lake Fault and the Hook Lake Fault. Its internal stratigraphy was disrupted by the early east-trending Milliken Lake Fault and further dismembered into a series of north-northwest-trending panels by the Big Charlie Lake Fault, Insole Lake Fault and their associated splays (Geoscientific Map MAP2010-1, back pocket, see details below). North of the Milliken Lake Fault, the Western Hook Lake succession is steeply dipping, slightly overturned, west-facing and greater than 1200 m thick (apparent thickness). It can be traced along strike for more than 7 km (Figure 23). South of the Milliken Lake Fault, the Western Hook Lake succession occurs as a narrow, fault-bounded, moderately to steeply dipping, south-facing succession over 5 km thick (apparent thickness).

The Western Hook Lake succession consists largely (>70%) of aphyric to sparsely plagioclase-phyric (unit Vw1), plagioclase- and pyroxene-phyric (unit Vw2), and plagioclase-phyric (unit Vw3) basalt flows. It includes lesser amounts (~25%) of mafic (unit Vw5) and heterolithic (unit Vw6) volcaniclastic rocks, and <5% of variably porphyritic felsic volcanic rocks and their associated breccias (unit Vw4).

Aphyric to sparsely plagioclase-phyric basalt flows (unit Vw1) are aphanitic with <5% plagioclase phenocrysts (1 mm in diameter), and are brown to black on weathered surfaces and dark green on fresh surfaces. Individual flows include pillowed (subunit Vw1b) and/or massive (subunits Vw1a, c) sections with amoeboid flow-top breccia (subunit Vw1d). Measured flow thickness is variable, ranging from 10 to 75 m. The massive flows (subunit Vw1a) are generally <10 m thick and characterized by a lack of internal structure, oblate to irregular, elongate zones of quartz-epidote alteration, and an increase in quartz-feldspar–filled amygdules (10%) and quartz-filled gas cavities (up to 5 cm long) towards their top. Thick upper selvages and thermal contraction cracks occur locally at the tops of some massive flows. Pillowed flows (subunit Vw1b; Figure 24a) are 30–60 m thick, and form the basal unit of individual flows where the massive facies is

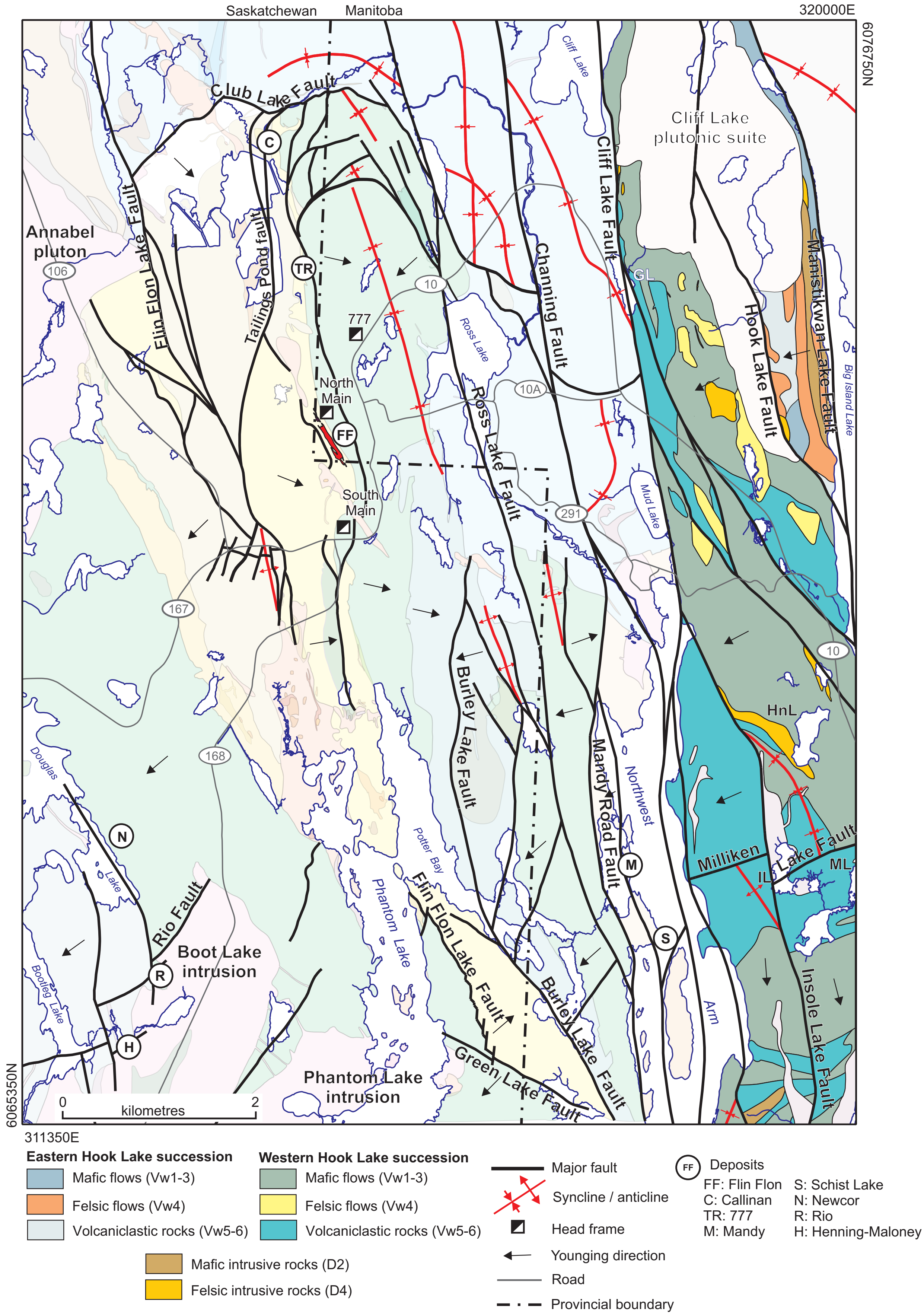


***Figure 23:** Simplified geology of the Flin Flon area, showing the main rock types of the Western and Eastern Hook Lake successions and major structures in the area. Abbreviations: GL, Grant Lake; HnL, Hartnett Lake; IL, Insole Lake; ML, Milliken Lake. For complete legend refer to Figure 3.*

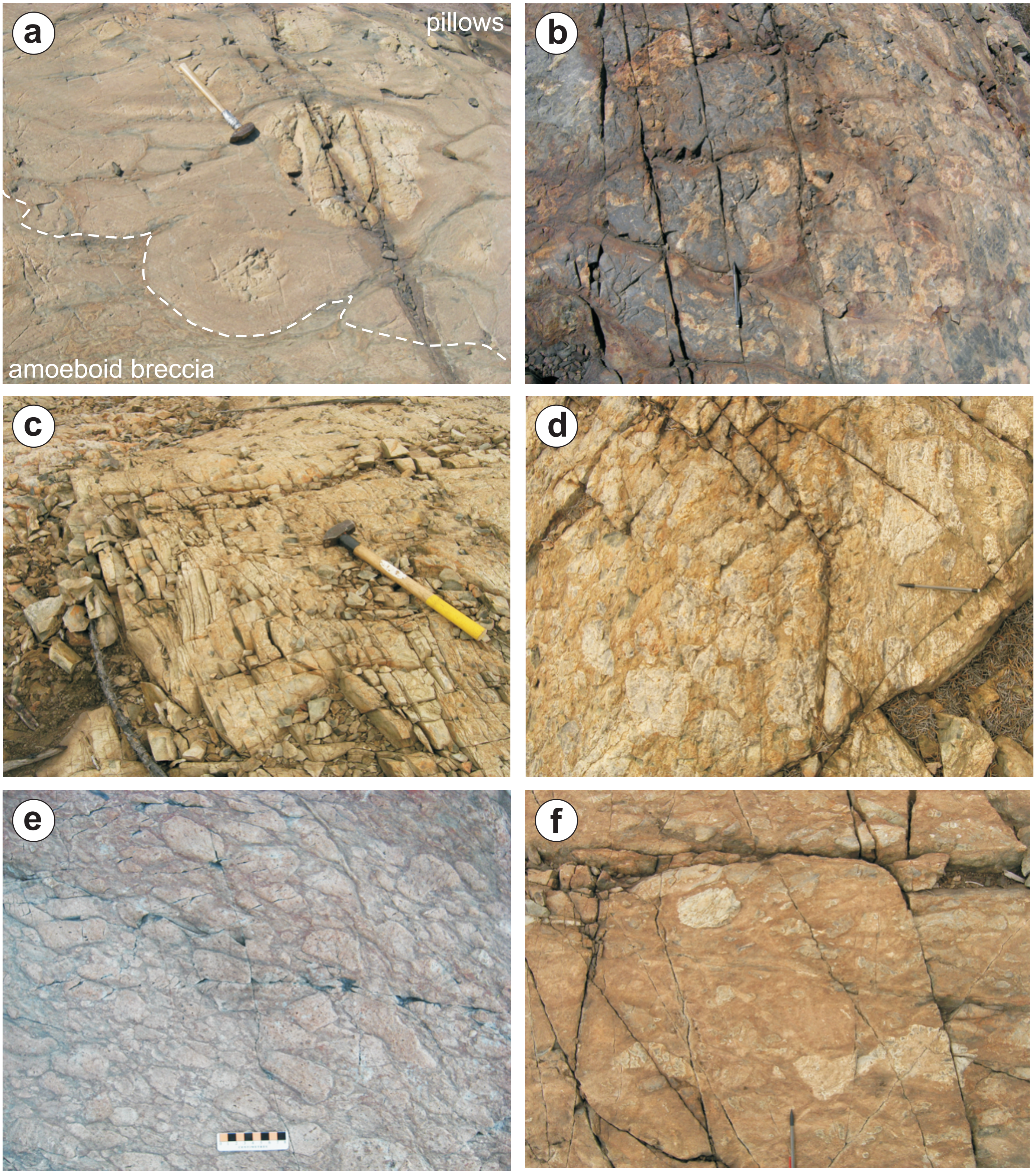


***Figure 24:** Representative rock types, facies and textures of the Western Hook Lake succession: **a)** aphyric to sparsely plagioclase-phyric pillowed basalt flows with amoeboid flow-top breccia (subunit Vw1b); **b)** silicified and hornfelsed aphyric pillowed basalt flow showing abundant epidote-quartz alteration patches (subunit Vw1b); **c)** quartz-phyric massive rhyolite (subunit Vw4c); **d)** quartz-phyric monolithic felsic breccia (subunit Vw4d); **e)** heterolithic mafic breccia (subunit Vw5a); and **f)** heterolithic mafic and felsic breccia (subunit Vw6a).*

absent. Hydrothermal quartz-epidote alteration patches occur throughout pillowed flows and are most pronounced in pillow cores (Figure 24b). Up to 30% quartz-, feldspar-, or chlorite-filled amygdules (1–10 mm) are common throughout pillowed flows, with larger amygdules concentrated at the rims and smaller ones concentrated in the cores of individual pillows. Amoeboid flow-top breccia (subunit Vw1d) occurs at the top of most aphyric flows and contains 10–40 cm irregularly shaped clasts within a highly epidotized matrix.

Plagioclase- and pyroxene-phyric basalt flows (subunit Vw2) have 7–20% plagioclase phenocrysts (2–3 mm in diameter), 10–15% pyroxene phenocrysts (2–6 mm in diameter), and are buff-brown on weathered surfaces and dark green to grey on fresh surfaces. These basalt flows are mainly pillowed (subunit Vw2b) with some massive flows (<5% in volume; subunit Vw2a) and are intercalated with significant amounts of in situ breccia (>50% volume; subunit Vw2c). Massive flows are highly amygdaloidal (up to 25%) towards the top of individual flows and locally contain gas cavities up to 15 by 7 cm in size. Pillows are typically closely packed, amygdaloidal and contain 1–3 cm thick brown to green chilled margins. They range in size from 0.1 to 2 m in diameter, with average size varying across stratigraphy, but generally decreasing towards the top of flows. Many pillowed flows contain various amounts of epidotized and/or laminated interpillow material. Quartz-filled amygdules and gas cavities (5–20%), locally reaching 6 cm in size, occur in the majority of pillows and are concentrated around the rims of individual pillows. Amygdules generally show an increase in size and abundance towards the top of flows. Some pillowed flows grade upward into an amoeboid pillow breccia. In situ breccia contains mainly plagioclase-pyroxene–phyric fragments up to 10 cm in diameter, with chilled margins preserved on one or more sides of some of the fragments, within a pyroxene- and plagioclase-crystal–rich matrix. These breccia units are commonly massive to crudely bedded, and are interpreted as pillow-fragment breccia (Bailes and Syme, 1989; Kremer and Simard, 2007).

Plagioclase-phyric basalt flows (unit Vw3) are coarsely plagioclase-phyric (20%, 0.4–4 mm in diameter), and are buff-brown to dark green on weathered surfaces and grey on fresh surfaces. These flows are predominantly massive and homogeneous, with a lesser amount of pillowed flows. Thick units of in situ breccia (subunit Vw3a) composed of monolithic, matrix-supported, poorly sorted, subangular to rounded pillow fragments are common.

Felsic volcanic rocks (unit Vw4) are relatively sparse in the Western Hook Lake succession. They are buff white on weathered surfaces and pale blue-grey on fresh surfaces. They are either aphanitic (subunit Vw4a), plagioclase-phyric (subunit Vw4b; <10%, 1–2 mm in diameter), or quartz-phyric (subunit Vw4c; <5%, <1 mm in diameter; Figure 24c). Aphyric and plagioclase-phyric felsic rocks are mainly massive and generally lacking features indicative of an extrusive origin (such as lobes and flow-top breccias), which suggests that they could be intrusive. However, the massive quartz-phyric felsic volcanic rocks (subunit Vw4c) usually have associated lenses of weakly to well-developed monolithic felsic breccia (Figure 24d) occurring at their top or on their lateral margins, suggesting that these rhyolite breccias are carapace and flank breccias of the coherent rhyolite flows (Kremer and Simard, 2007). Numerous gossan zones and sulphide occurrences were observed in close association with rhyolite flows and breccia units throughout the Hook Lake Block (Kremer and Simard, 2007).

Mafic volcaniclastic rocks (unit Vw5) consist mainly (>90%) of heterolithic mafic breccia (subunit Vw5a) with lesser amounts of well-bedded tuff (subunit Vw5b) and scoriaceous lapilli tuff (subunit Vw5c). The heterolithic volcaniclastic rocks (unit Vw6), in particular heterolithic mafic-felsic breccia (subunit Vw6a), are intercalated with heterolithic mafic breccia (subunit Vw5a).

Heterolithic mafic volcanic breccia (subunit Vw5a; Figure 24e) is massive, clast supported and very poorly sorted, but becomes crudely bedded to bedded, and matrix supported higher up in the stratigraphy, where it also includes a greater proportion of lapilli breccia and lapilli tuff. Bedding is defined by variations in fragment size and abundance, and crude layering is defined by 'trains' of blocks of a specific composition. Where the heterolithic mafic breccia directly overlies, or is intercalated with mafic flows, its clast content commonly reflects the composition of the nearby flows, displaying a greater proportion of clasts similar in phenocryst content to the underlying flows. The clasts range in size from 2 to 30 cm, are subangular to rounded, and are commonly scoriaceous or contain varying amounts of quartz- and feldspar-filled amygdules. Gradational changes in clast composition occur within these breccia units, with intercalated beds of more heterolithic to more monolithic material. The matrix of the heterolithic mafic breccia is commonly composed of lapilli-sized clasts of mainly aphyric basalt, plagioclase (5–10%, 1–2 mm in diameter) and pyroxene (<5%, <2 mm in diameter) crystals, in mafic tuffaceous material. Bedded tuff and scoriaceous lapilli tuff are locally present throughout the Western Hook Lake succession.

Heterolithic mafic-felsic breccia (subunit Vw6a; Figure 24f) is massive, matrix to clast supported and poorly sorted. It is composed of 5–20% angular to subangular, 2–40 cm clasts of aphyric to weakly porphyritic rhyolite mixed with a variety of subrounded, 0.5–100 cm mafic clasts, including highly amygdaloidal aphyric basalt, plagioclase-phyric basalt and plagioclase-pyroxene–phyric basalt. The matrix of these heterolithic breccias is composed of fine mafic material (chlorite±actinolite), with varying amounts of plagioclase (±pyroxene) crystals, crystal fragments, and lapilli-sized mafic fragments. The average clast size decreases upsection to less than 10 cm, the proportion of felsic clasts increases to approximately 20%, and the amount of matrix increases to a maximum of 15–20%. These heterolithic mafic-felsic breccia units are locally intercalated with heterolithic mafic breccia, and sharp bedding contacts are locally developed.

North of the Milliken Lake Fault (Grant Lake section of the Hook Lake Block by Bailes and Syme, 1989), the Western Hook Lake succession is <1.5 km thick, slightly overturned, southwest-facing, and can be traced laterally over 7 km from the Milliken Lake Fault to several kilometres north of the Cliff Lake plutonic suite (Figure 3). The base of the succession is characterized by over 700 m of aphyric to sparsely plagioclase-phyric basalt flows (unit Vw1) with minor plagioclase- and pyroxene-phyric basalt flows (unit Vw2). North of Mud Lake, the base of the succession is intruded to the east by an early dioritic phase of the Cliff Lake plutonic suite that has been dated at 1888 ±1 Ma (sample PQB-1705-08, Rayner, 2010), indicating that this volcanic package is older than 1888 Ma (Rayner, 2010). The mafic flows are best exposed between Highway 10 and the Northwest Arm of Schist Lake, just north of Hartnett Lake, and immediately west of the Cliff Lake plutonic suite (Figure 23). They form thick 30–60 m successions of pillowed flows intercalated with minor massive basalt flows and amoeboid flow-top breccia. All flows display variable amounts of quartz-epidote alteration. These basalt flows are crosscut by numerous felsic rocks. Some of these felsic rocks are massive, coherent bodies lacking flow features (unit D4), which suggests that they are intrusive and emplaced by a single pulse of magma. Others, like most of the quartz-phyric felsic rocks bodies (unit Vw4c), have monolithic felsic breccia found in close spatial association with massive rhyolite domes suggesting possible extrusive emplacement. Some of these felsic flows/domes, like the ones 500 m east-southeast of Grant Lake, are closely associated with a mafic dike swarm (subunit D2f), more intense hydrothermal alteration in the surrounding rocks, and locally have mineralized zones (gossan zones; Kremer and Simard, 2007). A quartz- and plagioclase-phyric massive rhyolite collected east of Mud Lake was dated at 1891 ±17 Ma (sample 107-07-1326, Rayner, 2010). These felsic rocks and dike swarms commonly underlie an abrupt lateral transition from mainly mafic flows to mainly heterolithic breccia, suggesting that they were emplaced at the margin of primary synvolcanic subsidence structure(s), which highlights the VMS potential of these rocks (Kremer and Simard, 2007). Infilling these subsidence structures, and overlying the basal basalt flows, is a >300 m thick sequence of mixed heterolithic breccia, both felsic and mafic, which is interpreted to represent proximal resedimentation of pre-existing mafic and felsic volcanic rocks into a subsidence structure/basin within the Western Hook Lake succession (Kremer and Simard, 2007). The southern portion of this succession has been dismembered by late north-northwest-trending thrust(?)–strike-slip faults, which preserved and repeated only parts of it (Figure 23).

South of the Milliken Lake Fault (Schist Lake, Milliken Lake, and Rennie Lake sections of the Hook Lake Block, by Bailes and Syme, 1989), the Western Hook Lake succession is roughly 5 km thick (apparent thickness), moderately to steeply dipping, south-facing and exposed in narrow (<700 m wide) fault-bounded panels between the Cliff Lake Fault and the Manistikwan Lake Fault (Insole Lake Fault, Figure 23). The internal stratigraphy and the location of the prehnite-pumpellyite isograd in each panel is similar, but has been slightly displaced by late strike-slip movement recorded along the various north-trending faults. The lowermost unit of Western Hook Lake succession, south of the Milliken Lake Fault, is a roughly 600 m thick package of heterolithic mafic breccias with subordinate thin (1–4 m) mafic flows, overlain by a roughly 1400 m thick sequence of aphyric and porphyritic mafic flows. These flows have been correlated with similar flows at the base of the Western Hook Lake succession north of the Milliken Lake Fault (Bailes and Syme, 1989). These flows are overlain by a 3 km thick succession of alternating mafic heterolithic breccias, scoria-rich tuff and tuff breccia (subunits Vw5a, c; ~200–500 m thick), and aphyric and porphyritic mafic flows (units Vw1, Vw2, Vw3; ~200–800 m thick). The succession does not display any large-scale systematic stratigraphic variations in flow thickness, vesicularity, or flow organization; virtually all of the flows are thick, pillowed and strongly amygdaloidal, some with large (1 cm) amygdules, gas cavities and radial pipe vesicles (Bailes and Syme, 1989). The predominance of pillowed flows suggests that this succession is not vent proximal. Also, the variation in phenocryst content could suggest one or more sources of magmatism, or replenishment of the magma chamber between magmatic events. The clasts in the breccia are dominated by various textures and colours of basalt fragments with portions of chilled margin preserved, suggesting that most of them were derived from fragmentation of pillowed flows transported by subaqueous density currents from local sources (Bailes and Syme, 1989). The presence of scoria-rich tuff and tuff breccia suggest that these rocks were deposited in a shallow water environment. This sequence is interpreted to represent an accumulation of coarse volcaniclastic material on the flank of a subsiding volcanic edifice (Bailes and Syme, 1989), similar to the interpretation north of the Milliken Lake Fault.

Eastern Hook Lake succession (Ve)

The Eastern Hook Lake succession (Ve) is exposed between the Hook Lake Fault and the Manistikwan Lake Fault north of Highway 10. It forms a steeply dipping, slightly overturned, west-facing succession, >700 m thick (apparent thickness), which can be traced along strike for over 5 km north of Highway 10 (Figures 3, 23), and extends for several kilometres north of the map area.

The Eastern Hook Lake succession (Ve) consists mainly of rhyolite flows and flow breccia (unit Ve3; ~30%) intercalated with heterolithic mafic and felsic breccia (subunit Ve5a; ~20%) and heterolithic mafic breccia (subunit Ve4b; ~15%), with a subordinate amount of mafic flows (units Ve1, Ve2; <15%) and mafic siltstone and mudstone (unit Se1; <10%; Kremer and Simard, 2007). This succession is intruded in the east by laterally continuous gabbroic sills (subunit D2b).

Rhyolite (unit Ve3) similar in appearance to rhyolite observed within the Western Hook Lake succession occurs in abundance in the Eastern Hook Lake succession. It outcrops as both laterally continuous layers and irregular discontinuous lobes. Basal massive rhyolitic flows, which weather white and have blue-grey fresh surfaces, are typically aphyric or contain 5–10%, 2–5 mm quartz and feldspar phenocrysts (Figure 25a). Massive rhyolite is typically capped by monolithic felsic breccia and locally overlain by well-bedded, fine to very fine grained felsic tuff. Felsic breccia is clast supported, with a general increase in the amount of matrix and decrease in the size of clasts upsection. Matrix material is primarily chlorite±sericite and is sometimes highly silicified. A sample of moderately to strongly silicified rhyolite from the Eastern Hook Lake succession was dated at 1882 ±1 Ma (sample 107-07-1381, Rayner, 2010) suggesting that the Eastern Hook Lake succession is significantly younger than the Western Hook

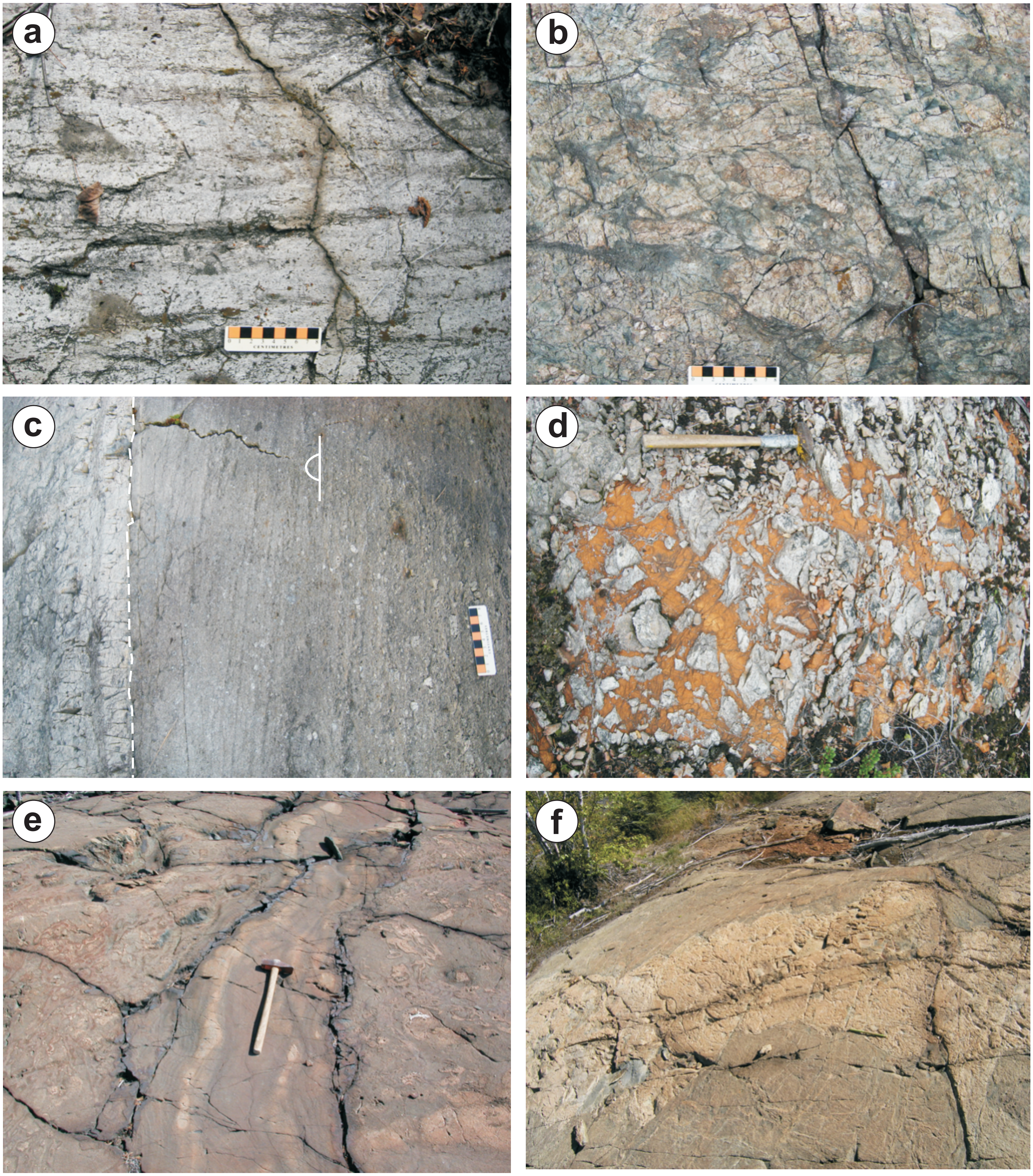


Figure 25: *Representative rock types, facies and textures of the Eastern Hook Lake succession volcanic rocks and some synvolcanic intrusive rocks of the Flin Flon arc assemblage:* ***a)*** *massive foliated plagioclase-phyric rhyolite of the Eastern Hook Lake succession (subunit Ve3c);* ***b)*** *heterolithic mafic breccia of the Eastern Hook Lake succession (subunit Ve4b);* ***c)*** *contact between pebbly greywacke and siltstone, showing normal grading (right, subunit Se1), and massive plagioclase-phyric rhyolite (left, subunit Ve3c) of the Eastern Hook Lake succession;* ***d)*** *intense iron-carbonate alteration of a quartz-phyric rhyolite (in situ pseudobreccia; subunit Ve3b);* ***e)*** *synvolcanic aphanitic mafic dike cutting across amoeboid flow breccia of the Hidden formation (subunit D2a); and* ***f)*** *aphyric felsic dike cutting across aphyric basalt flows of the Western Hook Lake succession.*

Lake succession. Note that this rhyolite unit was previously mapped as a shallow felsic sill by Bailes and Syme (1989), but remapped as an extrusive rhyolite by Kremer and Simard (2007). At one location southeast of the Cliff Lake plutonic suite, in an otherwise massive quartz-phyric rhyolite, a narrow zone of intense iron-carbonate alteration has formed an in situ pseudobreccia (Kremer and Simard, 2007). The 1882 Ma rhyolite is laterally continuous and extends north where it is clearly crosscut by quartz-phyric leucotonalite that makes up the main phase of the Cliff Lake plutonic suite. The plutonic suite east of the Hook Lake Fault has never been precisely dated (1859 ±22 Ma, Ansdell and Kyser, 1990), however, the Cliff Lake plutonic suite west of the Hook Lake Fault has yielded a reliable age at 1886 ±1 Ma (Stern et al., 1999). This is older than the age of the felsic volcanic rock in the Eastern Hook Lake succession and highlights a question about whether the Cliff Lake plutonic suite is really the same intrusive body on either side of the Hook Lake Fault. More work is underway to clarify this issue.

Heterolithic breccia (subunits Ve5a, Ve4b) in the Eastern Hook Lake succession consists of monolithic aphyric basalt breccia (subunit Ve4a), heterolithic mafic lapilli tuff to breccia (subunit Ve4b; Figure 25b), heterolithic mafic-felsic breccia (subunit Ve5a) and minor laminated mafic and felsic tuff (Kremer and Simard, 2007). The breccias are commonly clast to matrix supported and show variation in clast size and composition across section. Bedding is readily observed throughout the Eastern Hook Lake succession and can locally be used for reliable top determinations.

Two small outcrops of interbedded pebble conglomerate, greywacke and laminated siltstone (subunit Se1; Figure 25c) were observed on the western shore of Big Island Lake (known unofficially as Manistikwan lake), but they could not be traced due to large gabbroic intrusions in the vicinity. Large, fine-grained, equigranular, oikocrystic and plagioclase glomeroporphyritic gabbroic sills (subunit D2b) intrude the Eastern Hook Lake succession near the western shore of Big Island Lake. These sills are commonly up to 100 m thick, and can be followed laterally for more than 3 km along strike to the north.

The presence of abundant breccia in addition to the relative scarcity of flows suggest that these rocks, similar to the coarse volcaniclastic sequence in the upper portion of the Western Hook Lake succession, represent proximal resedimentation of pre-existing felsic and mafic volcanic rocks into a subsidence structure and/or basin (Kremer and Simard, 2007). However, the presence of massive rhyolite and its associated breccia indicate localized magmatism, possibly adjacent to the margin of the subsidence structure. These rocks have been intensively explored for copper and zinc since the 1930s. At one location southeast of the Cliff Lake plutonic suite in an otherwise massive quartz-phyric rhyolite, a narrow zone of intense iron-carbonate alteration has formed an in situ pseudobreccia (Figure 25d; Kremer and Simard, 2007). Disseminated to semimassive sulphides occur locally, either at the contact between massive and fragmental rhyolite and/or at the contact of rhyolite with the overlying volcanic-volcaniclastic rocks (Kremer and Simard, 2007). Assays of up to 12% zinc have been reported at the upper contact of the 1882 Ma rhyolite a few hundred metres west of Big Island Lake (Gale and Eccles, 1988b).

A similar package of rocks 5 km north of the map area hosts the Trout Lake VMS deposit. From bottom to top, the host stratigraphy for the Trout Lake VMS deposit includes mafic flows and mafic volcaniclastic rocks, aphyric to sparsely quartz-phyric rhyolite, graphitic argillite and rhyolitic tuff, rhyolitic volcaniclastic rocks and quartz-phyric coherent rhyolite, which host the VMS mineralization, and finally graphitic argillite and greywacke (Ordóñez-Calderón et al., 2009). Mafic dikes and sills (gabbros) commonly intrude the base of this volcanic stratigraphy (Ordóñez-Calderón et al., 2009). A synvolcanic rhyolite sill was dated at 1878.5 ±1.1 Ma (J.C. Ordóñez-Calderón, pers. comm., 2010), which, if representative of the age of the host volcanic succession, would make the Trout Lake volcanic succession, not only very similar in terms of rock types and potential depositional volcanic environment, but also coeval with the Eastern Hook Lake succession.

Cope Lake volcanic rocks (C)

(from Bailes and Syme, 1989)

The Cope Lake volcanic rocks (C) are composed of a folded succession of aphyric to strongly porphyritic mafic flows, subordinate synvolcanic dikes, and minor pillow-fragment breccia (historically referred to as the Cope Lake Block, Bailes and Syme, 1989). They occur at the northeastern corner of the map area between the Manistikwan Lake Fault (Figure 3) and Big Island Lake Fault (outside of the map area).

The base of the Cope Lake succession consists of aphyric to sparsely porphyritic pillowed flows, alternating with strongly porphyritic flows, with individual flow packages between 150 and 300 m thick. Aphyric to sparsely porphyritic basalt flows contain 0–10% plagioclase and pyroxene phenocrysts, from 0.2 to 2 mm in size; the phenocryst content of porphyritic flows varies with plagioclase 5–10% (1–2 mm) and pyroxene 0–5% (1–2 mm). The upper half of the sequence is dominated by a thick (>700 m) succession of strongly porphyritic pillowed and massive flows containing contain 15–30% plagioclase phenocrysts and glomerocrysts (0.3–5 mm) and 0–10% pyroxene phenocrysts (0.5–10 mm). Vesicularity varies considerably between flows, with the local presence of large (1 cm) amygdules, gas cavities and radial pipe vesicles.

The predominance of flows and the lack of volcaniclastic material in this succession suggests a relatively vent-proximal depositional environment for the Cope Lake volcanic rocks; the increase in proportion of massive flows versus pillowed flows upsection may reflect an increase in magmatic activity in the area. The Cope Lake volcanic rocks most likely formed on the flank of a mafic submarine volcano, relatively close to a vent.

Intrusive rocks of the Flin Flon arc assemblage

The volcanic rocks of the Flin Flon arc assemblage are intruded by a variety mafic to felsic dikes, sills, and other intrusive bodies. Some are clearly synvolcanic (peperitic margins, etc.), others are clearly postvolcanic (crosscut stratigraphy), but older than the Successor-arc rocks (<1.88 Ga; see 'Successor-arc rocks' below; Figure 26).

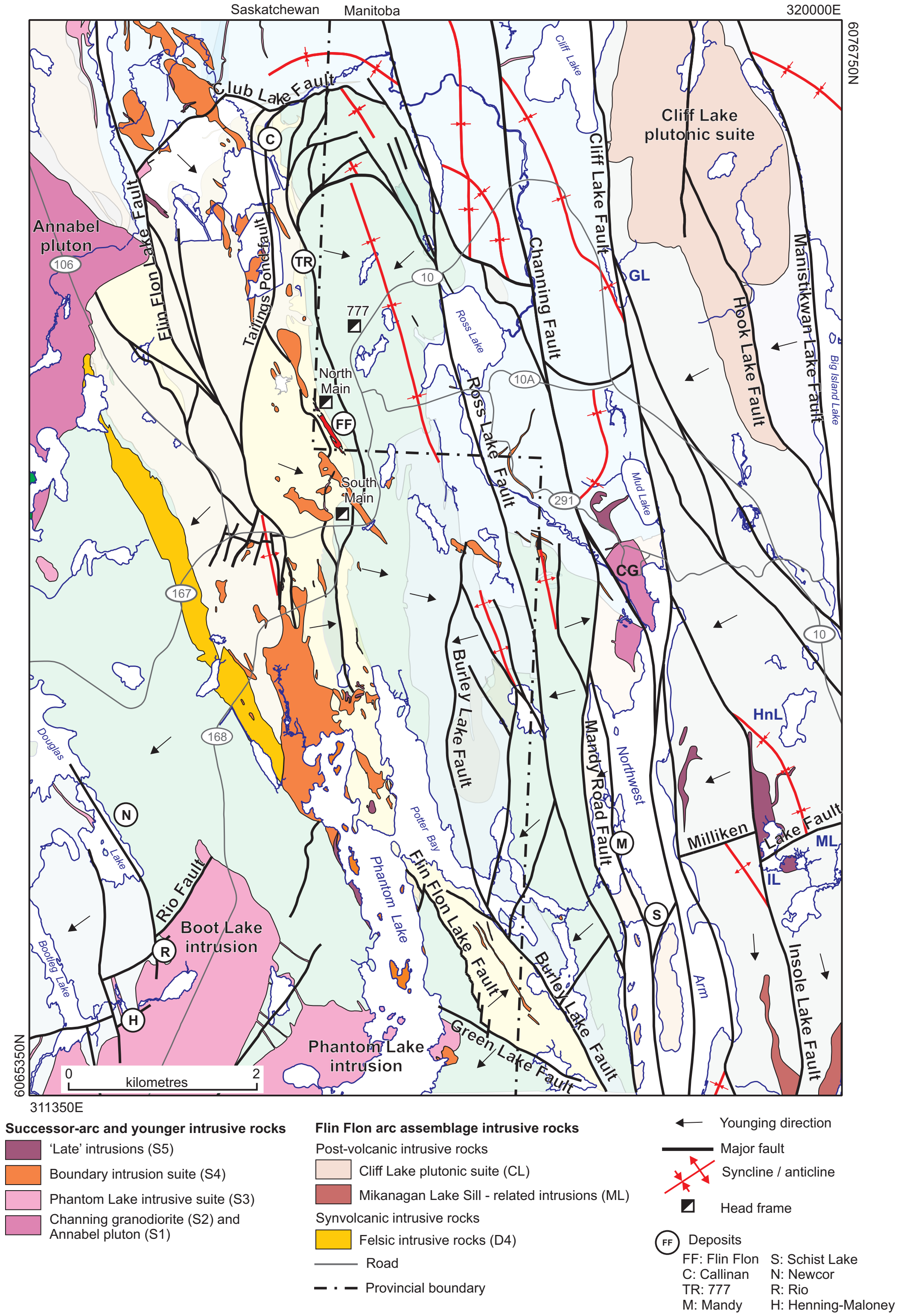


Figure 26: *Simplified geology of the Flin Flon area, showing the main intrusions and major structures in the area. Abbreviation: CG, Channing granodiorite; GL, Grant Lake; HnL, Hartnett Lake; IL, Insole Lake; ML, Milliken Lake. For complete legend refer to Figure 3.*

Synvolcanic intrusions (D)

Mafic dikes and sills are present in all the formations/successions of the Flin Flon arc assemblage in the area. Mafic sills are so common in parts of the stratigraphy, like the lower portion of the Hidden formation, that thicknesses of volcanic successions are commonly exaggerated because of dilation by mafic sills (Figure 26). Synvolcanic dikes commonly display irregular contacts with the flows (Figure 25e) and are characterized by complex crosscutting relationships (DeWolfe et al., 2009a). They follow pillow margins in places, filling the space between them. They vary in textures and colours (aphyric, porphyritic, gabbroic) and in size (generally <2 m wide). Many of these dikes and sills display distinctive centimetre-scale rhythmic banding, expressed by alternating grain-size variations or phenocryst content variations parallel to their walls (multiple injections/cooling events), with a massive core. Where dikes have been traced into volcaniclastic units, their margins are often peperitic, indicating that they were emplaced into wet unconsolidated volcaniclastic material (Gibson et al., 2003b). Dikes often indicate an extensional regime during emplacement, whereas sills are more characteristic of compression. In places, mafic dikes are so abundant that the hostrock appears only in localized screens, which indicates emplacement in a vent-proximal environment; sills are features that are more consistent with their emplacement farther from the volcanic vent (Gibson, 1990).

Felsic dikes are less common than mafic dikes in the Flin Flon arc assemblages (1:20). Multiples generations of felsic dikes can be recognized based on crosscutting relationships and textures. They commonly occur as isolated dikes, generally <2 m wide and vary widely in texture (aphyric, porphyritic). Their edges are usually straight, and they crosscut the stratigraphy without inflexion at contacts (Figure 25f). Peperitic margins in volcaniclastic units were not observed. West of the Channing–Mandy Road faults felsic dikes are common below the Louis formation, but do not occur within or above the Louis formation, which suggests that they may have been feeders to the Tower member rhyolite at the base of the Louis formation.

Felsic sills are common within the Hidden formation on the western side of the Flin Flon Lake Fault. They form 50–300 m bodies that roughly parallel stratigraphy for several kilometres and are often associated with abundant mafic sills with or without dikes. A massive quartz-plagioclase–phyric felsic sill, which is up to 400 m thick and extends over 5 km in length from Hilary Lake and past Myo Lake (subunit D4g; historically referred to as the Myo rhyolite/intrusion; Table 1, back pocket), cuts the top of the Flin Flon formation and the base of the Hidden formation at a very low angle west of the Flin Flon Lake Fault (Figures 3, 26). This sill yielded an age of 1888.9 ±1.7 Ma (sample 05MYO-01, K. Bailey, pers. comm., 2009), which is coeval with the surrounding volcanism in the area.

A massive V-shaped gabbro/diorite intrusion occurs just west of Ross Lake around the 777 mine headframe (subunit D2b; historically referred to as the Town diorite, Bailes and Syme, 1989; Figure 26). This intrusion is brown on the weathered surfaces and medium to dark green on fresh surfaces, medium grained and equigranular, with 0–10% quartz (Bailes and Syme, 1989). Its V-shape may be original or the result of folding, as the 'V' is coaxial with the hinge of the Hidden Lake syncline.

Post-volcanic intrusions

Felsic dikes and sills (unit D4)

Felsic dikes and sills very similar in appearance to the synvolcanic felsic dikes and sills of the Flin Flon arc assemblage described above, were emplaced within structural corridors that are known to be postvolcanic but pre–Successor-arc in age. They follow arcuate shear zones that have been interpreted as minor thrust faults that repeat portions of the volcanic pile.

Cliff Lake plutonic suite (CL)

(from Bailes and Syme, 1989)

The Cliff Lake plutonic suite is a multiphase, quartz-megacrystic tonalite that forms a composite body east of the Cliff Lake Fault and intrudes both the Western and Eastern Hook Lake successions (Figures 3, 26). The composite pluton is >6 km long and a maximum of 1.8 km wide. It is crosscut lengthwise by the Hook Lake Fault, which thrust the eastern side up onto the western side, and was later reactivated as a strike-slip fault. West of the Hook Lake Fault, the Cliff Lake plutonic suite is a multiphase laccolith <200 m thick (C.A. Devine, pers. comm., 2009); east of the Hook Lake Fault it forms a potentially deeper-rooted stock.

The Cliff Lake plutonic suite is composed of six discrete mappable units (subunits CLa–f) and minor aplite dikes. Other than an early quartz-diorite unit (subunit CLa), the remaining tonalite units are usually distinguished by the amount and degree of assimilation of contained recrystallized volcanic xenoliths.

Quartz diorite (subunit CLa) predominantly occurs west of the Hook Lake Fault, and is the oldest phase of the Cliff Lake plutonic suite. It composes ~15% of the plutonic suite, and occurs mainly as large xenoliths within quartz-phyric tonalite (subunit CLb) and locally along the western marginal of the intrusion. Quartz diorite weathers buff-brown to brown, is dark grey-green on fresh surfaces, and is equigranular with a grain size of 1–4 mm.

Quartz-phyric tonalite with abundant large xenoliths (subunit CLb; Figure 27a) occurs predominantly west of the Hook Lake Fault and composes ~25% of the plutonic suite. This distinctive phase occurs at the southwestern margin of the pluton and in a fault-bounded slice between the Big Charlie Lake and Hook Lake faults. It is characterized by abundant, slab-like, relatively unassimilated xenoliths of recrystallized mafic volcanic rocks and quartz diorite (subunit CLa). The tonalite has a sharp irregular contact with the host volcanic rocks of the Western Hook Lake succession to the west. Dikes of the tonalite locally intrude the volcanic pile. The tonalite weathers white, is pale grey on fresh surfaces, and is quartz-megacrystic (1–8 mm phenocrysts, 30–40%) with a medium-grained, quartz-plagioclase groundmass. Mafic xenoliths are the most conspicuous component of this phase. They are generally angular or slab-like to nebulous and lensoid, and are generally about 10–200 cm long. They vary widely in abundance, are variably recrystallized, and are commonly concentrated in linear, inclusion-rich zones in which xenoliths are aligned (Figure 27b). An intrusive contact between this phase of the tonalite and the subunit CLc is superbly exposed along virtually its entire length. This contact is sharp, nearly vertical, irregular/serrated and shows subunit CLc tonalite chilled against subunit CLb.

Quartz-phyric tonalite with abundant small, highly digested xenoliths (subunit CLc) is characterized by abundant (commonly 25–50%) small (2–5 cm), highly assimilated, rounded recrystallized mafic xenoliths. This phase occurs predominantly west of the Hook Lake Fault, and composes ~30% of the plutonic suite. This phase is relatively homogeneous, weathers white to creamy buff, is pale grey on fresh surfaces, and texturally and mineralogical is very similar to subunit CLb.

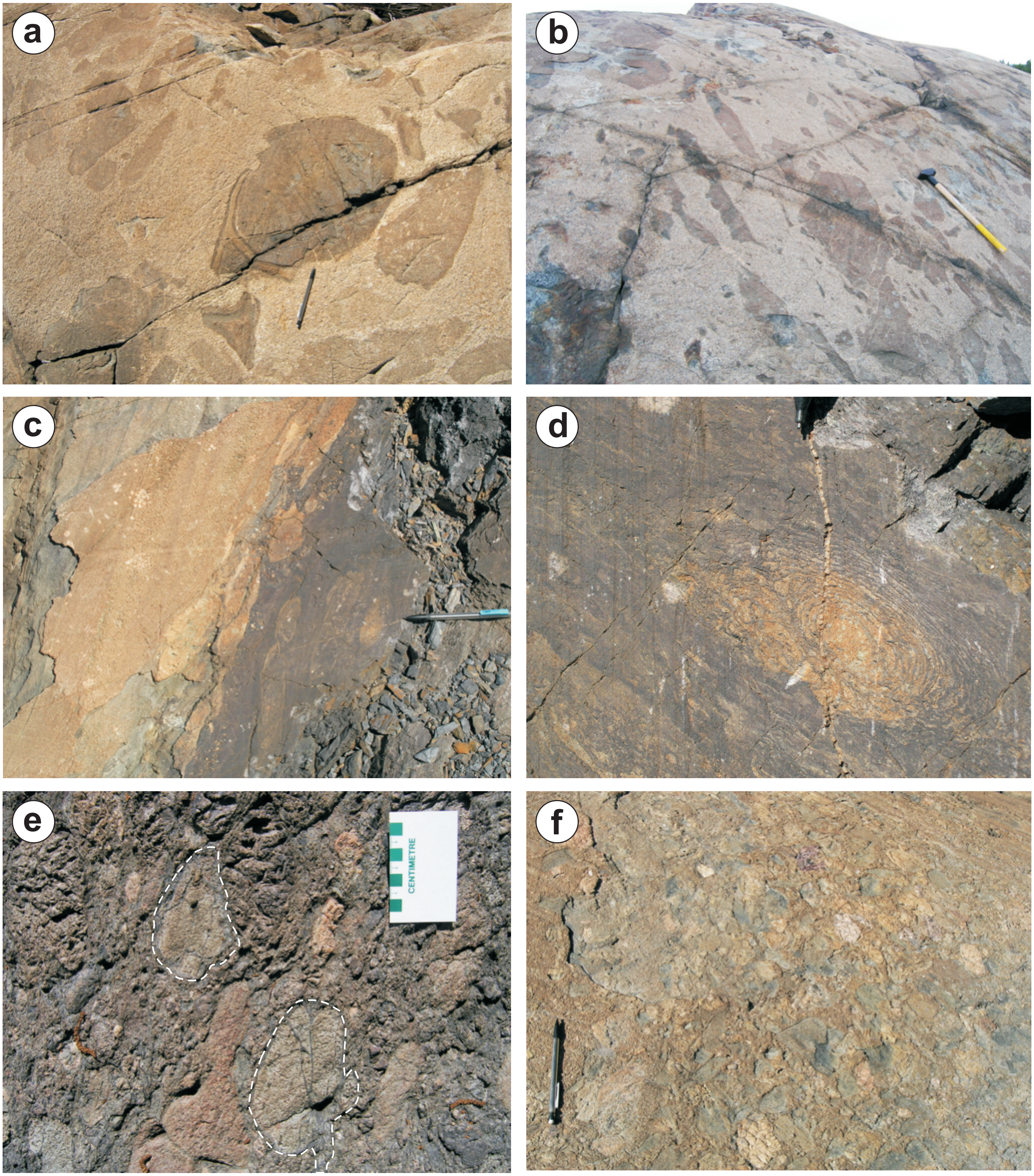


Figure 27: *Representative rock types, facies and textures of rocks of the Cliff Lake plutonic suite, regolith and Missi Group:* ***a)*** *quartz-phyric tonalite with abundant large variably digested mafic volcanic xenoliths (subunit CLb);* ***b)*** *quartz-phyric tonalite with abundant variably digested aligned mafic volcanic xenoliths (subunit CLb);* ***c)*** *unconformable contact between the Missi Group sandstone (left; unit M3) and very well developed regolith in pillowed mafic volcanic rocks of the Hidden formation (right; subunit Hc1b) showing strong hematization of the volcanic rocks (purple colour);* ***d)*** *well-developed Liesegang banding (onion peel–like) in dark purple regolith in mafic volcanic rocks of the Hidden formation (subunit Hc1b);* ***e)*** *dark purple Missi Group conglomerate (unit M1) showing abundant clasts of regolith; and* ***f)*** *greenish cobble to pebble conglomerate of the Missi Group (unit M1).*

Quartz-phyric leucotonalite and micrographic leucotonalite (subunit CLd) differs from subunit CLc in that it is generally more leucocratic, and contains very few, small, highly assimilated mafic xenoliths. This phase occurs predominantly east of the Hook Lake Fault where it forms a plug-like body. It also occurs farther to the east, outside the map area near Cope Lake, where it intrudes the Cope Lake volcanic rocks.

The leucotonalite marginal phase (subunit CLe) occurs as an irregular body of leucotonalite west of the Big Charlie Lake/ Hook Lake faults in the north. It contains rare small tabular xenoliths, and is not necessarily quartz-megacrystic.

The fined-grained, quartz-phyric micrographic tonalite (outside the map area to the north) forms two small, poorly defined plug-like bodies that occur east and west of the Big Charlie Lake/Hook Lake faults in subunits CLc and CLd. This tonalite has sharp contacts with the enclosing megacrystic tonalite. The presence of quartz-phyric micrographic tonalite as small plugs in the two main phases of the Cliff Lake plutonic suite is a definite indication of the multiphase and possibly long-lived nature of this intrusive system, which could explain the range in U-Pb crystallization ages (see details below).

Light pink micrographic aplite dikes occur in all phases of the Cliff Lake plutonic suite but are rarely observed outside of the intrusions; they are the youngest phase of this system. The dikes are generally only a few centimetres wide (up to 50 cm), straight-walled and continuous, and have no preferred orientation. Thicker dikes are weakly porphyritic.

The Cliff Lake plutonic suite is cut and crosscuts a wide variety of texturally and compositionally distinct mafic dikes and small mafic plugs.

A number of features in the Cliff Lake plutonic suite, including magmatic stoping of wall rocks, mineralized breccia pipes, and abundant volcanic xenoliths, suggest that it is a high-level intrusion. A cognate quartz-diorite xenolith from the western portion of the Cliff Lake plutonic suite was dated at 1886 ±1 Ma (Stern et al., 1999) along with an early quartz-diorite marginal phase of the intrusion dated at 1888 ±1 Ma (sample PQB-1705-08, Rayner, 2010). These ages suggest that the Cliff Lake plutonic suite is roughly coeval with the Flin Flon formation west of the Cliff Lake Fault.

Mikanagan Lake Sill and related intrusions (ML)

(from Bailes and Syme, 1989)

Several differentiated gabbroic intrusions have been recognized in the rocks of the Flin Flon area and farther eastward. These intrusions are all medium to coarse grained and have prominent compositional zoning. Individually, the intrusions differ in form (sill- to plug-like bodies), zoning, and parental magma composition. These intrusions are the largest mafic bodies in the area, with the largest, the Mikanagan Lake Sill, which was dated at 1881 +3/–2 Ma, well east of the map area in the Bear Lake Block (Stern et al., 1999).

One of these intrusions occurs south of the Milliken Lake Fault in the Western Hook Lake succession. It forms an elongated Y-shaped plug that extends over 3 km southward outside the map area (Figure 26). It consists mainly of gabbronorite (subunit MLa) with subordinate amounts of micrographic leucotonalite (subunit MLb). Gabbronorite (subunit MLa) weathers shades of buff and buff-brown and is mottled with grey-green fresh surfaces; typically grain size is 1–2 mm. Gabbronorite usually forms the base of most of these intrusions. The micrographic leucotonalite (subunit MLb) weathers white to pink and is light to dark grey on fresh surfaces. This phase is usually found towards the top of the differentiated gabbroic intrusions, and in this case, occurs on the western side of the gabbronorite in the map area. This compositional zoning would indicate that the top of this intrusion is to the west.

Successor-arc rocks (1.84–1.87 Ga)

Rocks of the Flin Flon arc assemblage were intruded by felsic and mafic intrusive rocks and overlain by sedimentary and volcanic rocks following the accretion of the assemblage to other tectonostratigraphic elements of the Flin Flon Belt (ocean floors, island arcs, oceanic plateaus, etc.) at around 1.88–1.87 Ga (Stern et al., 1999). These intrusive, sedimentary and volcanic rocks represent the postaccretion arcs and basins that formed between 1.87–1.84 Ga atop the more juvenile, recently accreted portion of the Flin Flon Belt (Stern et al., 1999). They are referred to as ‘successor’ to describe the temporal context, in that they are younger than and thus succeed the 1.88–1.92 Ga tectonostratigraphic assemblages, such as the Flin Flon arc assemblage (Stern et al., 1999).

In the Flin Flon area, only intrusive and sedimentary rocks of the Successor-arc rocks are present. They can be subdivided in three: the pre–Missi Group intrusive rocks (units S1, S2), the Missi Group sedimentary rocks (units M1, M3), and the post–Missi Group intrusive rocks (units S3–S5; Figure 26). None of these rocks were remapped in detail for this project. The section below summarizes previous work.

Pre–Missi Group intrusive rocks (units S1, S2)

It is difficult to definitively establish the ‘pre–Missi Group’ nature of the intrusive rocks observed in the Flin Flon area. The Missi Group sedimentary rocks occur as a single, slightly structurally dismembered sedimentary basin, which does not come into contact with all the documented intrusive rocks. Exposed nonconformable contacts and/or clear crosscutting relationships between the various intrusive and Missi Group rocks are rarely available making it difficult to determine their exact age relationship, and precise U/Pb ages are not always available.

Channing granodiorite (unit S2)

(from Bailes and Syme, 1989)

A small body of strongly deformed granodiorite, exposed at and south of Channing (Figures 3, 26; Geoscientific Map MAP2010-1, back pocket) intrudes the undivided volcanic rocks that host the Schist Lake and Mandy VMS deposits between the Channing–Mandy Road faults and the Cliff Lake Fault. It is nonconformably overlain by Missi Group sedimentary rocks to the north, which makes it one of the few unequivocally pre–Missi Group felsic plutons in the Flin Flon area. The granodiorite weathers rusty buff to buff and is rusty buff, grey-green or grey on fresh surfaces. It has a hypidiomorphic equigranular texture and 1–3 mm grain size, but the rock is typically strongly foliated so that primary textures are not well preserved. The Channing granodiorite is crosscut by a southern splay of the Channing Fault. Where exposed in the fault zone, it is cataclastic. The intrusion does not display a strong metamorphic aureole.

Annabel pluton (unit S1)

(from Thomas, 1993)

The Annabel pluton (unit S1) forms a northwesterly tapering pear-shaped body approximately 26 km long and as much as 5.5 km wide (Byers and Dahlstrom, 1954; Byers et al., 1965; Figure 26). The Annabel pluton is a multiphase intrusion, composed of two main rock types: biotite-hornblende granodiorite and biotite granodiorite, together with minor amounts of tonalite, granite and diorite.

Most of the intrusion within the Flin Flon area consists of biotite (±hornblende) granodiorite that is pinkish-grey to grayish-orange, coarse grained and inequigranular. It is commonly deformed, showing two cleavages and a pronounced rodding fabric. The biotite-hornblende granodiorite occurs mainly along the margin of the intrusion in the Flin Flon area, and commonly hosts abundant metavolcanic xenoliths ranging from small, subrounded inclusions to large, metre-sized, angular blocks. It is typically very pale orange to yellowish-grey, medium to coarse grained, inequigranular, and moderately to very well foliated. This pluton cuts the Flin Flon arc assemblage rocks to the east and is cut by the ca. 1800 Ma Annabel Lake shear zone to the west (Stern and Lucas, 1994). It has been dated at 1866 ±3 Ma (sample SLB-92-82, Stern and Lucas, 1994).

Missi Group (<1.845 Ga; M)

(from Bailes and Syme, 1989; Stauffer, 1990)

The Missi Group (M) is composed of sandstone, pebbly sandstone and conglomerate that lies with profound angular unconformity on the volcanic rocks of the Flin Flon arc assemblage, and nonconformably overlies a wide variety of intrusions that cut these volcanic rocks (Bailes and Syme, 1989; Figures 3, 28). The Missi Group sedimentary rocks are in turn cut by a limited number of ultramafic to felsic intrusions. Contacts and structural symbols that represent the Missi Group on the geological map (Geoscientific Map MAP2010-1, in back pocket) are from Stockwell (1960) with minor revisions along the volcanic-sedimentary contacts. Locations of fold axis are mainly from Stauffer and Mukherjee (1971) with some modifications based on recent work from B. Lafrance (Lafrance et al., in press; Figure 28).

All volcanic rocks at exposed, unfaulted contacts with the Missi Group display a hematiferous weathered zone (regolith) up to 10 m thick (Figure 27c). Regolith developed on volcanic rocks shows distinctive spheroidal weathering defined by concentric bands of reddish-purple to black hematite (Liesegang rings), and darkening of the rock between spheroids or, in their absence, of the rock as a whole (Figure 27d). Such darkened rock represents a paleoregolith that has been metamorphosed so that it now appears more highly altered (mainly chlorite) than surrounding rocks (Stauffer, 1990). The degree of hematization and spheroidal weathering decreases away from the unconformity surface. Abundant weathered volcanic regolith detritus is incorporated into the lowermost Missi Group sedimentary rocks; many pebbles display spheroidal weathering identical to that developed in underlying volcanic rocks (Figure 27e). Basal Missi Group conglomerates and sandstones that are particularly rich in hematiferous regolith detritus have a distinct mauve weathering colour relative to the normally grey weathering Missi Group. The evidence of pre–Missi Group in situ weathering is important because it shows uplift of the submarine volcanic rocks of the Flin Flon arc assemblage into a subaerial environment where weathering began some significant period of time prior to the development of the system that deposited the Missi Group (Stauffer, 1990). It is clear that considerable uplift, weathering, erosion and some folding occurred prior to the deposition of the Missi Group sedimentary rocks (see 'Structural history' below).

The Missi Group sedimentary rocks consist of three main rock units: pebble to cobble conglomerate with minor interbedded sandstone and pebbly sandstone (unit M1), pebbly sandstone (unit M2) and sandstone (unit M3; Figure 28).

The pebble to cobble conglomerate, with minor interbedded sandstone and pebbly sandstone (unit M1; Figure 27f) occurs as massive to normally graded beds composed of well-rounded pebbles and cobbles (Bailes and Syme, 1989). With the exception of jasper, chert and iron formation, which are rare in the presently exposed Flin Flon arc assemblage rocks, the great majority of clasts present in the conglomerate are similar to rock types that occur in the underlying rocks (e.g., mafic volcanic rock, felsic volcanic rocks, quartz vein, fine-grained mafic intrusive rocks). Granitoid clasts, although present, are not a common component of the coarse fraction. Intraformational sandstone clasts occur only in the upper portion of the Missi Group. Notably, pebbles of 'metamorphic rocks' (higher than the regional greenschist facies) and gneissic rocks are absent. All clasts are moderately to well rounded. Sand-sized grains in these rocks include quartz (mono- and polycrystalline), plagioclase and rock fragments. Sand-sized rock fragments are generally poorly preserved and include common felsic volcanic rocks, mafic volcanic rocks, chert and minor granitoid types. Accessory heavy minerals in the sandstone include specular hematite, zircon, apatite, allanite and tourmaline. The specular hematite is particularly abundant, and occurs as subrounded sand-sized grains in laminae, placers and crossbed foresets (Bailes and Syme, 1989). Most of the conglomerate and sandstone beds appear to be lensoid scours with erosive bases.

The pebble sandstone (unit M2) occurs as trough crossbedded coarse to very coarse grained sandstone and pebbly sandstone (Figure 29a). Pebbles occur in discrete beds, crossbeds and scours, and are dominated by quartz, chert, oxide-facies iron formation and felsic volcanic rocks. This unit is generally thinner and less common than the other units of the Missi Group (Figure 28).

The sandstone (unit M3; Figure 29b) is the most abundant unit of the Missi Group. It occurs at various levels within the Missi Group stratigraphy (Figure 28). Various sandstone units are defined by colour, grain size, composition and contained sedimentary structures. Most sandstone units are dominated

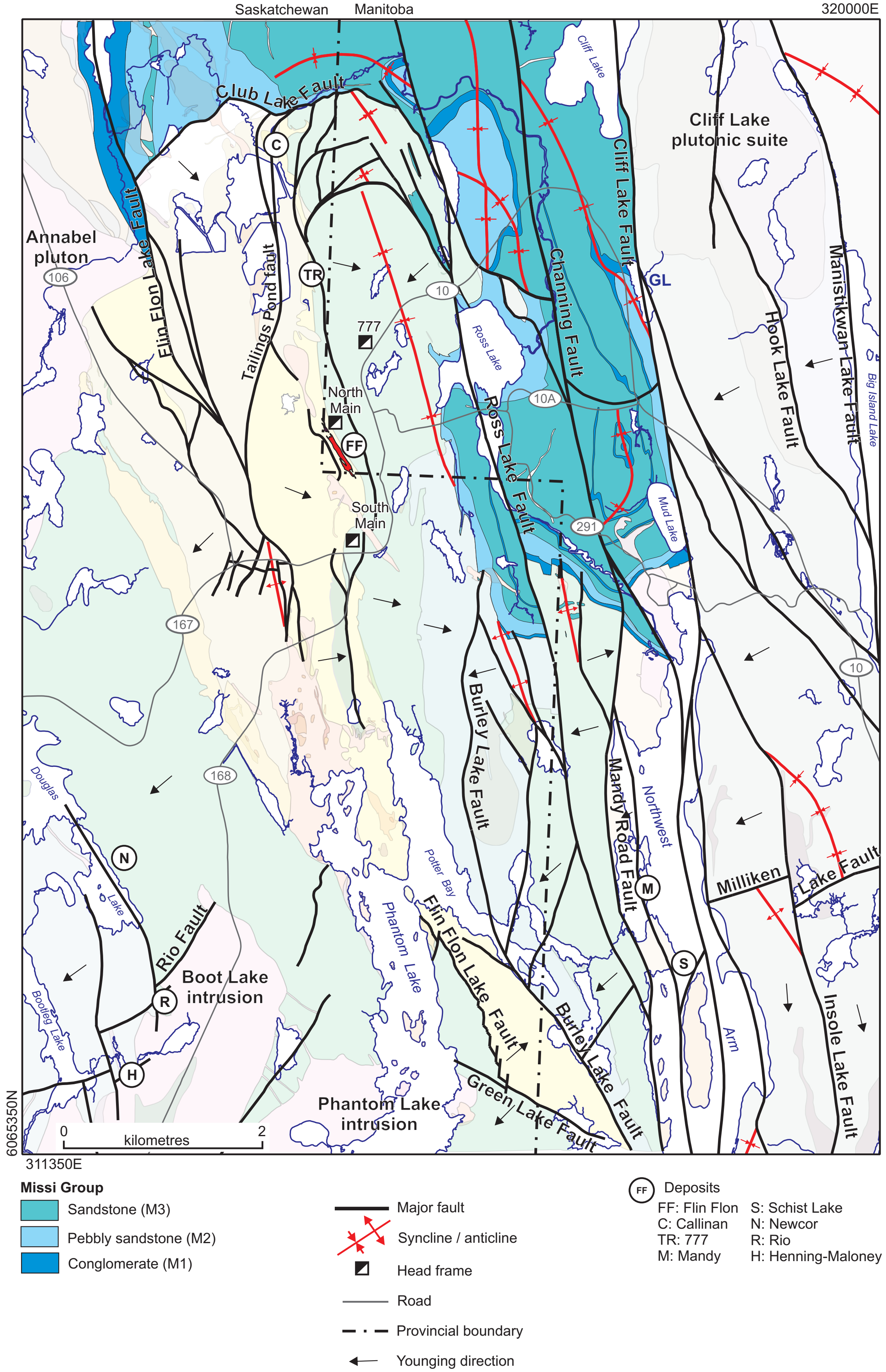


***Figure 28:** Simplified geology of the Flin Flon area, showing the main rock types of the Missi Group and major structures in the area. Abbreviation: GL, Grant Lake. For complete legend refer to Figure 3.*

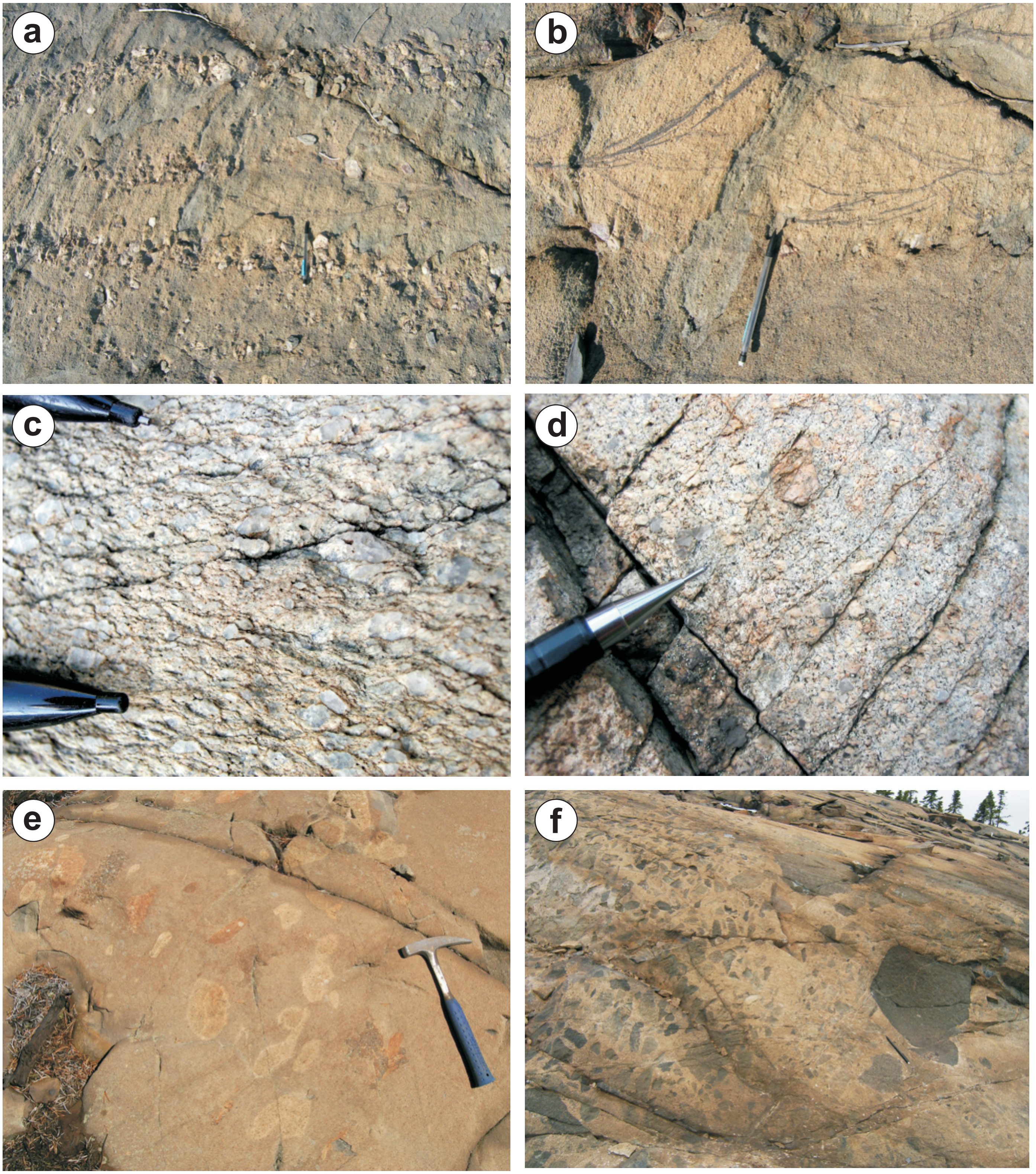


Figure 29: *Representative rock types, facies and textures of the Missi Group and post–Missi Group intrusive rocks:* ***a)*** *pebbly arkose of the Missi Group (unit M2) showing crossbedded structures and pebble lags;* ***b)*** *sandstone of the Missi Group (unit M3) showing crossbedded structures highlighted by the hematite lamina on crossbed foresets;* ***c)*** *close-up of very coarse sandstone of the Missi Group (unit M3) showing well-developed foliations;* ***d)*** *microcline-quartz–phyric granite dike of the Phantom Lake intrusive suite (Phantom Lake intrusion, subunit S3a);* ***e)*** *melagabbroic phase of the Boundary intrusions (subunit S4h) showing several rounded mafic to ultramafic xenoliths; and* ***f)*** *intrusion breccia of the Boundary intrusions (subunit S4i).*

by coarse to very coarse detritus (Figure 29c), with only the uppermost units being somewhat finer grained. Crossbedded sandstones predominate, and pebbles occur as rare isolated clasts generally composing less than 3% of the rock.

The stratigraphy of the Missi Group in the Flin Flon area has been described in detail by Stauffer (1990). The Missi Group contains two thick, fining-upward and maturing-upward sequences of fluvial conglomerates and sandstone, each with a polymictic conglomerate unit at their base (Stauffer and Mukherjee, 1971). The first basal conglomerate unit is 0–200 m thick and unconformably overlies the Flin Flon arc volcanic and intrusive rocks; this is clearly reflected in its clast content, many of which were derived from the underlying regolith. This conglomerate is overlain by an ~100 m thick pebbly sandstone unit that is in turn overlain by thick sandstone units with thin conglomeratic interbeds, with a total thickness of ~900 m (Bailes and Syme, 1989). The thin conglomerate units differ from the basal conglomerate in that they are characterized by an abundance of resistant rock types: quartz, chert, iron formation and felsic volcanic rocks. The sandstone units become somewhat finer grained towards the top of this succession. The second significant conglomerate occurs approximately halfway through the Missi Group stratigraphy (around the Pipeline and Grant Lake synclinal basins), and contains similar clast types to the first basalt conglomerate, except that it also contains 11–14% intraformational sandstone clasts (Bailes and Syme, 1989). Similar to the lower succession, the basal conglomerate of the second succession is overlain by a pebbly sandstone unit in turn overlain by thick sandstone units.

The Missi Group is interpreted as being deposited by fluvial processes in an alluvial system, adjacent to a highland source (Byers, 1953; Stauffer, 1974), which is consistent with the sedimentary structures observed in the Missi Group rocks. The presence of intraformational sandstone clasts at the base of the second succession indicates that lithified underlying Missi Group sedimentary rocks were uplifted, eroded and incorporated into the later portion of the group. This suggests abrupt uplift between deposition of the two successions (Stauffer, 1974; Bailes and Syme, 1989).

The rocks of the Missi Group are crosscut by intrusive rocks of the Phantom Lake intrusive suite, which provide a minimum age of sedimentation of 1842 ±3 Ma (sample 9112-0286, Heaman et al., 1992). Detrital zircons from the Missi Group sedimentary rocks were analyzed, and gave ages between 1950 and 1854 Ma (Ansdell et al., 1992). The Missi Group sedimentary rocks were thus deposited between 1854 and 1842 Ma (Ansdell et al., 1992).

Post–Missi Group intrusive rocks

Post–Missi Group intrusions are defined as those that are observed to cut the Missi Group, are emplaced within or across post–Missi faults, or have been precisely dated younger than 1842 Ma (Figure 26).

Boundary intrusion suite (unit S4)

(from Syme and Forester, 1977; Bailes and Syme, 1989)

Plug- and dike-like bodies of the Boundary intrusions only occur west of the Channing–Mandy Road faults, in a north-northwest corridor from Beaverdam Lake in the north to Phantom Lake in the south (Figures 3, 26; Geoscientific Map MAP2010-1, back pocket). Field relationships, petrography and chemical characteristics of the Boundary intrusions indicate that they are composed of three compositionally distinct, sequentially emplaced groups. From the oldest to the youngest these are 1) a mafic, augite- and biotite-bearing meladioritic group (subunits S4c–h); 2) a felsic group ranging from leucodiorite to granodiorite (subunits S5b–d; see comments below); and 3) an olivine-bearing group containing wehrlite and olivine gabbro (subunits S4a, b). The felsic group is chemically similar to post–Missi Group granodioritic plutons and is not genetically related to the mafic and ultramafic members of the Boundary intrusion suite, it was therefore labelled unit S5. Emplacement of the mafic intrusions was accompanied by significant dike emplacement and brecciation (intrusion breccia; subunit S4i) of both country rocks and earlier crystallized phases of the Boundary intrusion suite.

The mafic group is the most prevalent variety of Boundary intrusion in the area and is composed mainly of porphyritic biotite-augite meladiorite (subunit S4g) and porphyritic hornblende meladiorite (subunit S4d), and minor hornblende melamonzodiorite, melatonalite, and mela–quartz diorite (subunit S4c), hornblendite (subunit S4e), porphyritic gabbro and melagabbro (subunit S4h; Figure 29e), and highly strained (schistose) meta-pyroxenite (subunit S4f). These rocks are characterized by their abundance of intrusion breccia envelopes and radiating dike swarms peripheral to the intrusions (Figure 29f). The meladiorite dikes commonly contain up to 25% subrounded volcanic, plutonic and older Boundary intrusion xenoliths that range from a few centimetres to a metre in diameter.

The felsic group is characterized by leucodioritic to granodioritic rocks containing less than 15% total mafic minerals (i.e., hornblende, biotite and rare augite). This group represents less than 15% of the Boundary intrusions and generally occurs as small dikes and irregular-shaped intrusion breccias cutting the mafic Boundary intrusions.

The younger olivine-bearing group (subunits S4a, b) is composed of wehrlite (subunit S4b) and olivine gabbro (subunit S4a). The wehrlite (peridotite) forms a long, narrow body, for the most part completely enclosed by the mafic intrusive rocks forming the greater portion of a large Boundary intrusion by Phantom Lake. Small bodies of relatively unaltered peridotite also occur within the largest mafic Boundary intrusion. Small xenoliths of mafic Boundary rocks occur in the border zones of the wehrlite mass. Olivine gabbro forms a discontinuous rim from 30 to 70 m wide around the northern margins of the large wehrlite body at the north end of Phantom Lake. Olivine-bearing gabbroic rocks also occur, albeit rarely, as dikes cutting both wehrlite and other members of the mafic group. No felsic rocks were observed to cut any members of the olivine-bearing group.

A medium- to coarse-grained melanocratic gabbro of the Boundary intrusions was dated at 1842 ±3 Ma (sample 9112-0286, Heaman et al., 1992).

Phantom Lake intrusive suite (unit S3)

(from Galley and Franklin, 1987; Thomas, 1989)

The Phantom Lake intrusive suite, otherwise known as the Boot Lake–Phantom Lake intrusion (Byers et al., 1965; Pearson, 1984), is a multiphase intrusive body underlying the area between Phantom and Bootleg lakes (Figure 26) and extending about 3.5 km southwards to Boot Lake (outside the map area). This intrusive suite is composed of a sequence of four main intrusive phases from mafic to progressively more felsic rocks. Phase 1 is the most heterogeneous, consisting of gabbroic and dioritic rocks (subunit S3c; Boot Lake intrusion) mainly in the western portion of the intrusion, and a small mass farther south towards Boot Lake. Phase 2 occurs outside of the present map area, south of Boot Lake. It consists of fine-grained hornblende-biotite–phyric quartz monzodiorite and minor biotite-phyric syenite. Phase 3 rocks occur in two stocks of plagioclase-phyric granodiorite (subunit S3b; Boot Lake intrusion) in the western portion of the intrusion and in a small mass around Boot Lake. The rocks vary from fine to medium grained, are light grey to medium grey on fresh surfaces and weather pale yellowish-orange to yellowish-grey. Biotite and hornblende are present in about equal amounts and generally form less than 10% of the rock. Phase 4 is represented by a large body of microcline-quartz–phyric granite (subunit S3a; Kaminis or Phantom Lake intrusion; Figure 29d) underlying the northeastern part of the intrusion. The rocks, which are typically orangey pink on weathered surfaces and reddish-orange on fresh surfaces, vary from fine to medium grained and contain from 5 to 15 % euhedral microcline phenocrysts 1–4 cm (average 1.5 cm) long. Quartz phenocrysts, from 1 to 2 mm in diameter, compose up to 15% of the rock, whereas biotite generally forms less than 5%.

Contacts between the different phases are not always visible, but most appear gradational except for the porphyritic granite rocks (subunit S3a), which have sharp intrusive contacts with the earlier phases. All the Boot Lake–Phantom Lake phases are characterized by concentrations of xenoliths along contacts between the different phases. In places, the xenoliths form intrusion breccia zones several metres wide and over 10 m long. Xenoliths are rounded to angular and include an earlier phase of the intrusion itself, volcanic xenoliths, as well as xenoliths from the Boundary intrusions. At least four periods of dike emplacement have been observed: an early mafic phase mainly in the western margin of the complex; small aplite dikes concentrated along the margins between the porphyritic granite (subunit S3a) and intruded hostrocks; a northwest-trending dike swarm of porphyritic granite similar to subunit S3a that crosscuts all other phases of the intrusion and the surrounding hostrocks; and finally north-trending hornblende-plagioclase–phyric rocks that crosscut all other rock types in the zone delineated by the mafic dike swarm.

A fine- to medium-grained hornblende monzodiorite (subunit S3c; Boot Lake intrusion) that cuts a Boundary intrusion as well as a coarse-grained microcline-phyric granodiorite (subunit S3a; Phantom Lake intrusion) were dated. Both the monzodiorite and granodiorite gave an 1838 ±2 Ma age for the Phantom Lake intrusive suite (samples 9012-0370 and 9012-0371, Heaman et al., 1992), confirming that the Boot Lake and Phantom Lake intrusions are part of the multiphase intrusion that is the Phantom Lake intrusive suite (Galley et al., 1989).

'Late' intrusions (unit S5)

Several suites of intrusions that crosscut the younger intrusive rocks in the area, such as the Boundary intrusions or the Phantom Lake intrusive suite, have been grouped together and labelled as 'Late' intrusions (unit S5; Figure 26). None of them have been precisely dated.

The composition of the 'Late' intrusions varies greatly from gabbro (subunit S5e) and diorite (subunit S5d), to monzodiorite, monzonite and syenite (subunit S5c), to tonalite (subunit S5b) and granodiorite (subunit S5a). Because they are not genetically related to the mafic and ultramafic members of the Boundary intrusions, rocks of the felsic group of the Boundary intrusions were also labelled S5 (see 'Boundary intrusions' above for details). The 'Late' intrusions mostly occur as dikes or small irregularly shaped intrusions. One of the biggest of the 'Late' intrusions in the map area is the 'Insole Lake tonalite' (defined by Bailes and Syme, 1989) by Insole Lake in the southeastern corner of the map. This tonalite body (subunit S5b) is truncated to the west by the Insole Lake Fault and intruded by magnetite-bearing quartz diorite (subunit S5d). No age has been determined for this intrusion.

Economic geology

(D.E. Ames, K. MacLachlan and H.L. Gibson)

The Flin Flon area includes seven past-producing mines and one producing mine, including both volcanogenic massive sulphide and gold mines. In addition, there are over 150 known mineral occurrences of various types (Figure 30) in the area. The following section provides a general description of the volcanogenic massive sulphide and gold deposits, a brief discussion on the prospectivity for such deposits, as well as highlights of other mineral occurrences/showings in the area.

Volcanogenic massive sulphide (VMS) deposits

Since the first discovery of base metals in the Flin Flon area in 1915 a world-class mining district has developed, centred around the auriferous 62.5 Mt Flin Flon deposit (Galley et al., 2007; Mercier-Langevin et al., 2011). The Flin Flon mining district hosts six volcanogenic massive sulphide (VMS) deposits, with the 777 and Trout Lake mines as current producers and the Flin Flon, Callinan, Schist Lake and Mandy mines as past producers (Figure 30). The combined production for these six mines over the last 90 years has been 112.7 Mt with an average grade of 3.4% Cu, 6.02% Zn, 0.14% Pb, 2.13 g/t Au and 34.34 g/t Ag (Table 3).

Flin Flon, 777 and Callinan deposits and associated alteration

The Flin Flon, 777 and Callinan VMS deposits are associated with felsic volcanic rocks of the predominantly basaltic Flin Flon formation. The VMS deposits formed within an intra-arc rift basin or cauldron described previously (Syme et al., 1993; Devine et al., 2003; Gibson et al., 2009), in which

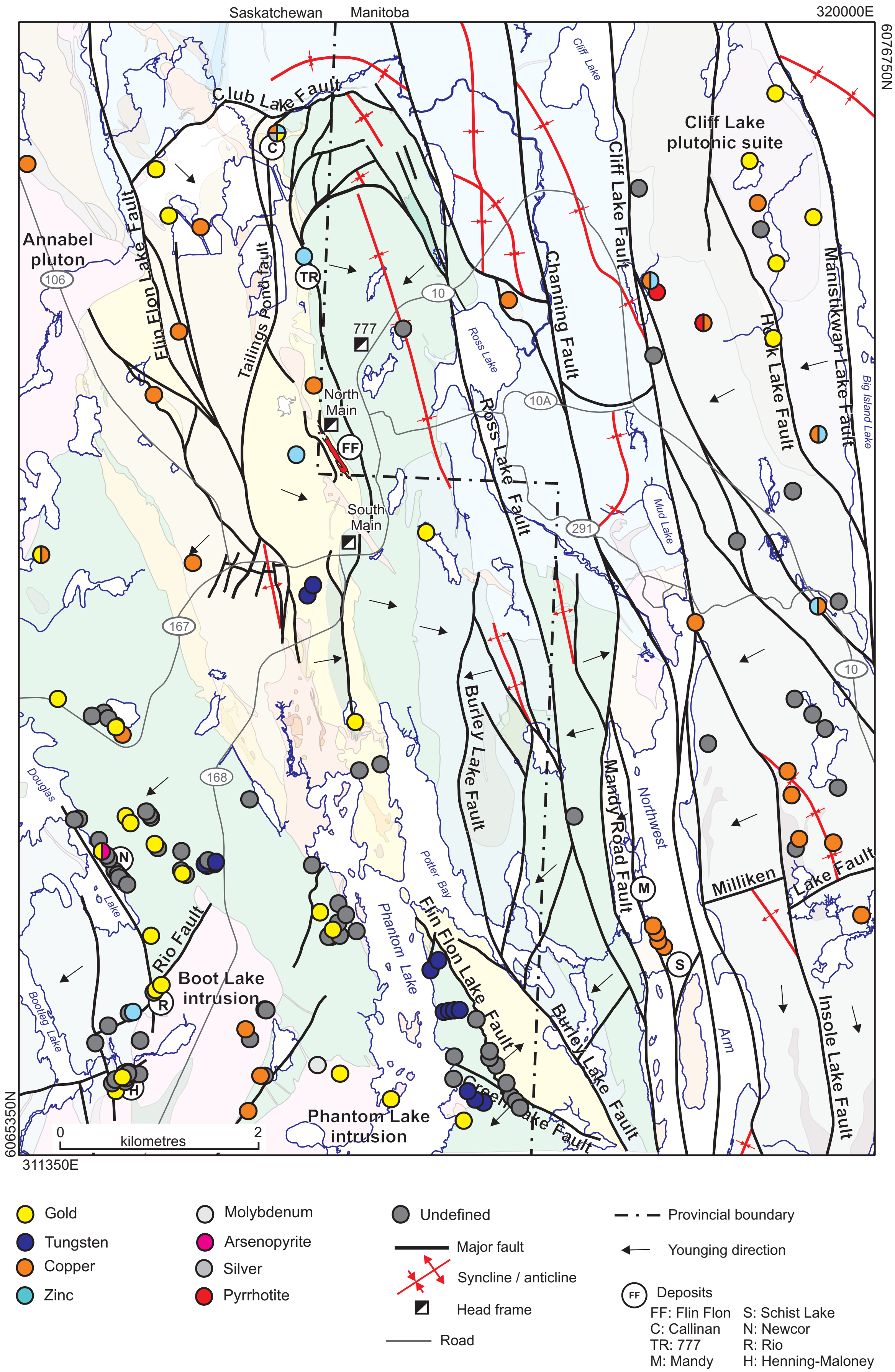


***Figure 30:** Generalized map of the Flin Flon area showing the eight past-producing and producing mines, as well as the majority of the 157 known mineral occurrences (from Simard et al., 2010). For complete legend refer to Figure 3.*

Table 3: Grade and tonnage data for the six VMS deposits in the Flin Flon area[1].

Deposit	Status	Host	Cu (%)	Zn (%)	Pb (%)	Cu+Zn (%)	Cu+Zn+Pb (%)	Cu/(Cu+Zn) (%)	Cu/Zn	Au (g/t)	Ag (g/t)	Au/Ag	Tonnes
Flin Flon[2]	CL	A-TH	2.2	4.1	0.01	6.30	6.31	34.92	0.54	2.7	41.28	0.07	62 485 362
Triple 7[2]	OP	A-TH	2.59	4.39	0.01	6.98	6.99	37.11	0.59	2.1	26.94	0.08	21 903 539
Callinan[2]	CL	A-TH	1.36	4	0.05	5.36	5.41	25.37	0.34	2.1	24.63	0.08	7 773 725
Mandy[3,4]	CL	A-TH	8.22	11.40	0.02	19.62	19.64	41.90	0.72	3.0	60.15	0.05	125 116
Schist Lake[2]	CL	A-TH	4.3	7.27	0.7	11.57	12.27	37.17	0.59	1.3	37.03	0.04	1 846 656
Trout[2]	OP	A-TH	1.74	4.97	0.06	6.71	6.77	25.93	0.35	1.6	16.02	0.10	21 612 296
Average Flin Flon area (10 km^2)			**3.40**	**6.02**	**0.14**	**9.42**	**9.56**	**36.10**	**0.52**	**2.13**	**34.34**	**0.06**	**115 746 694**

Abbreviations: A-TH, arc-tholeiite; CL, closed; OP, open
[1] Grade and tonnage data are not National Instrument 43-101 compliant.
[2] S. Pehrsson and K. Gilmore, pers. comm. 2012
[3] Bamburak, 1990
[4] Mandy was mined in 1917–1944 for Cu only, Zn remains.

individual deposits occur within local subsidence structures (DeWolfe et al., 2009a; Gibson et al., 2009). Known orebodies in the main Flin Flon camp extend over a distance of 7 km from north to south and occur in at least two thrust panels within the polyphase, imbricated sequence (Schetselaar et al., 2010). The zinc-copper-(gold) orebodies generally dip east and have a moderate plunge towards the southeast, parallel to the regional stretching lineation, with a maximum plunge length that ranges from 50 to more than 1000 m and a width, normal to plunge that ranges from 50 to 300 m (E. Schetselaar, pers. comm., 2009). Mineralization consists of massive, disseminated and stringer sulphides (Figure 31a–d); clastic, transported sulphide ore lenses are recognized at the Callinan deposit (Figure 31d). Major ore minerals are sphalerite, chalcopyrite, magnetite and pyrite, with minor cubanite ($CuFe_2S_3$), arsenopyrite (Fe[Co]AsS), tetrahedrite ($[Cu,Fe]_{12}Sb_4S_{13}$) and galena, and trace gold, electrum, antimonide Pb-telluride and selenide minerals (Au, Ag, Sb, Te, Se, respectively; Brownell and Kindel, 1935; Healy and Petruk, 1988, 1990; Ames et al., 2009). Tetrahedrite is a minor component of the ore that was recognized early at the Flin Flon deposit (Brownell and Kindel, 1935), more recently at Callinan (Petruk et al., 1990), but is found in all five deposits along with eight other antimonide minerals and probably accounts for most of the silver and antimony in the deposits (Ames et al., 2009). Gold occurs mainly as alloys in the five VMS deposits and as rare antimonide and telluride minerals. Electrum and gold contain significant mercury and antimony in the VMS deposits of the Flin Flon mining camp, however, electrum has few impurities in the mylonitized Schist-Mandy deposits (Ames et al., 2009).

Regional synvolcanic hydrothermal alteration in the rocks west of the Channing–Mandy Road fault consists of chlorite-biotite, epidote-quartz, actinolite-epidote, quartz-albite and quartz mineral assemblages (Ames et al., 2002; Tardif, 2003). Both pervasive, weak- to moderate-silicification zones and low-temperature silica deposition occurred. Regional hydrothermal alteration in the Flin Flon camp is typical of bimodal-mafic VMS deposits and is dominated by extensive zones of epidote-quartz alteration (Bailes and Syme, 1989; Franklin et al., 2005). Epidote is abundant (15–20%) throughout the camp with local anomalous zones that contain >30% epidote and localized zones of high-temperature clinozoisite-bearing alteration assemblages (Ames et al., 2011). Clinozoisite is associated with the 777-Callinan dike swarm in the Hidden formation and a discordant zone in the Hidden formation below a minor subsidence structure at the base of the Louis formation (Ames et al., 2011). It is less abundant away from the main subsidence structure (Ames et al., 2011). Stratabound, pervasive zones of silicification within the uppermost flows of the Hidden formation are recognized north-northwest of Douglas Lake, south of Carlisle Lake to Phantom Lake and as a localized zone southwest of Louis Lake. The silicified basalts are texturally, mineralogically and compositionally similar to silicified andesitic flows in the Noranda mining district where they underlie an area with significant VMS mineralization (Gibson, 1990).

Documentation of deposit-scale alteration assemblages in the main camp of the Flin Flon VMS deposit is incomplete because, unlike the hangingwall strata, the footwall strata are not well exposed. Koo and Mossman (1975) identified a discordant chlorite zone with quartz and carbonate in the subsurface of the Flin Flon deposit. Recent examination of mining company logs in a 1.8 km^3 block enveloping the 777-Callinan deposits identified proximal, synvolcanic, footwall hydrothermal alteration assemblages characterized by chlorite, sericite, talc and silicification that has been overprinted by carbonate-bearing assemblages (E. Schetselaar, pers. comm., 2009). The chlorite-biotite isograd, which marks the transition from greenschist to lower amphibolite facies metamorphism, overprints the alteration assemblages resulting in chlorite-dominant assemblages at the 777 and Flin Flon deposits, and biotite-dominant alteration at the Callinan deposit to the north. Recent detailed mapping has documented hydrothermal alteration in the footwall that consists of silicification, chloritization and

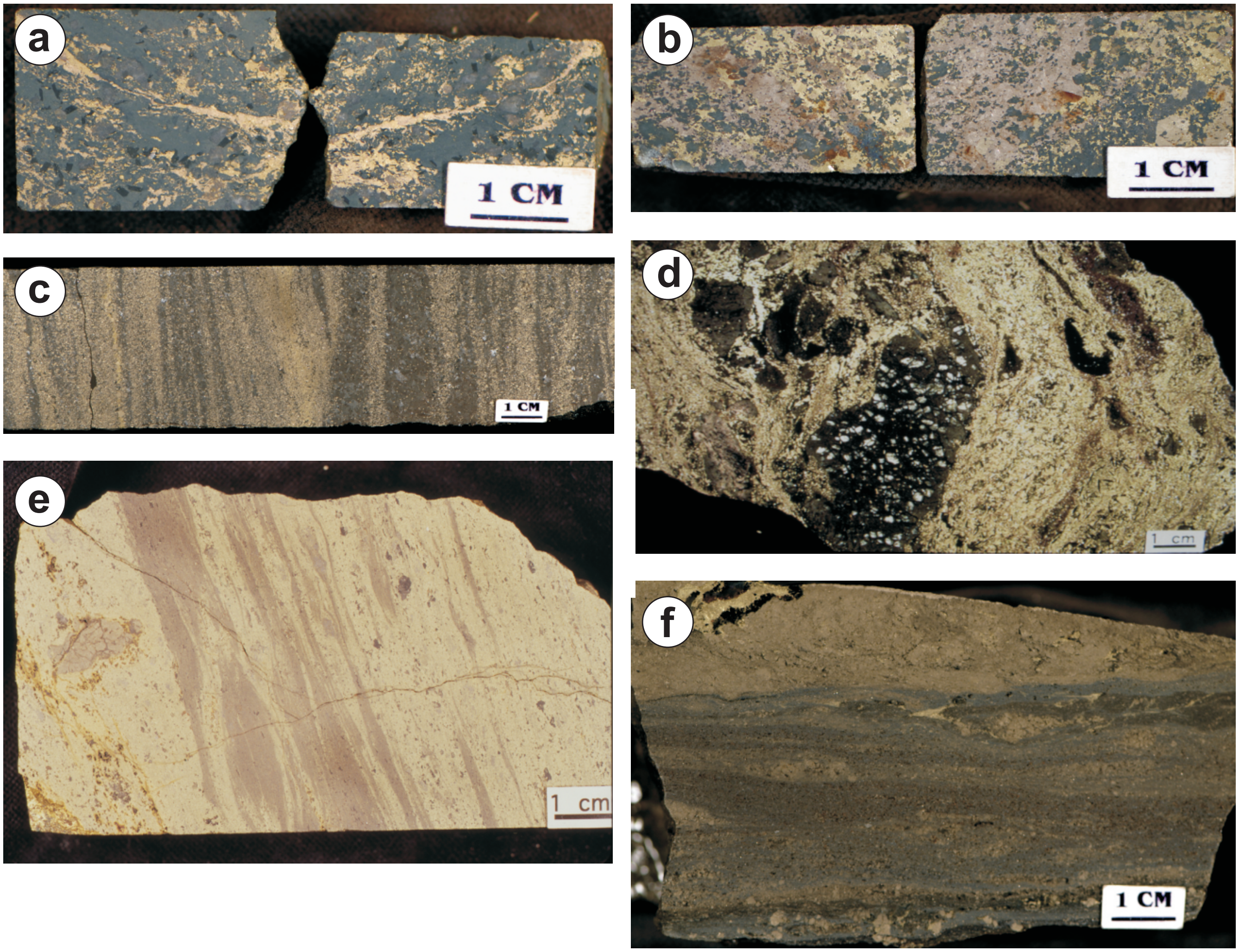


***Figure 31:** Samples of representative metamorphosed ore types from four VMS deposits in the Flin Flon area: **a)** 777 deposit, chalcopyrite-rich stringer ore containing chlorite-chalcopyrite-pyrite, minor biotite porphyroblasts, K-feldspar and trace pyrrhotite-magnetite-cubanite-sphalerite and berthierine; **b)** 777 deposit, massive chalcopyrite-pyrrhotite ore consisting of quartz-pyrrhotite-chalcopyrite, minor magnetite, pyrite and trace cubanite-sphalerite-sericite-carbonate and Fe-rich minerals berthierine and stilpnomelane; **c)** 777 deposit, massive pyrite-sphalerite banded ore with minor chalcopyrite and pyrrhotite intergrown with sphalerite interstitially to pyrite; ore contains trace arsenopyrite-magnetite-carbonate-stilpnomelane and chlorite; **d)** Callinan Cu-Zn-(Pb,Au) deposit, fragmental ore, altered quartz-phyric rhyolite fragments with semimassive chalcopyrite/sphalerite in matrix; **e)** Mandy Cu-Zn-(Au) deposit, example of mylonitized massive sulphide from the open pit, massive banded chalcopyrite-pyrrhotite ore with porphyroclasts and discontinuous streaks of sphalerite defining the tectonic fabric; and **f)** Schist Lake Cu-Zn-(Pb) deposit, tectonically banded sphalerite-pyrite ore in sharp contact with massive pyrite ore.*

epidotization all overprinted by magnetite veining. A broad zone of diffuse silicification is present within the footwall volcaniclastic lithofacies to the Flin Flon mine and a more localized, but poorly exposed and constrained, zone of intense chlorite alteration immediately below the deposit. Although outcrop exposure is limited, the chlorite alteration appears to be semiconformable, in part reflecting the volcaniclastic nature of the footwall lithofacies (Gibson and Kerr, 1993), but also reflecting transposition by a pronounced near-bedding parallel F_2 foliation and the location of the deposit along the limb of the Hidden Lake syncline.

Schist Lake and Mandy deposits and associated alteration

The sheared Schist-Mandy VMS deposits extend to surface on the western side of, and below, Schist Lake (Bruce, 1918; Cairns et al., 1957) and lie within a splay fault between the north-northwest-trending East Mandy Road and Cliff Lake faults (Figure 3, Geoscientific Map MAP2010-1, back pocket). Massive, banded sphalerite-chalcopyrite-pyrrhotite-pyrite ore with discontinuous streaks or bands of sphalerite define a pronounced tectonic fabric (Figure 31e, f). The Schist and Mandy deposits are hosted in sericite-carbonate and/or chlorite schistose rocks, and the lithostratigraphic position of the deposits with respect to the established stratigraphy for the Flin Flon deposits is uncertain (Simard, 2006a; Kremer and Simard, 2007; Simard et al., 2010; DeWolfe, 2011).

The Schist deposit was in production from 1950 to 1976. It has tabular to lensoid, northwest-striking ore lenses that dip steeply east and plunge 60° southeast (Cairns et al., 1957; Galley et al., 2007). Footwall hydrothermal alteration consists of pervasively chloritized and sericitized rocks and a chlorite stringer zone east of the deposit (Bailes and Syme, 1989; Cole et al., 2007, 2008; DeWolfe, 2011). All massive and disseminated sphalerite and chalcopyrite ore types are pyritic and display cataclastic textures and tectonic banding (Jonasson et al., 2009). The mylonitized sulphide ore minerals include major sphalerite, chalcopyrite, pyrite, minor tetrahedrite, arsenopyrite, galena and trace electrum, Ag and Co tellurides, Ni, Co, Fe and Ag antimonide minerals, pyrrhotite and cobaltite (Ames et al., 2009). Magnesian chlorite-dolomite-quartz alteration and minor sericite are prominent within the gangue and alteration zones.

Mineralization in the Mandy open pit consisted of a lensoid core of chalcopyrite-pyrite surrounded by banded pyrite-sphalerite enclosed by chlorite schist, within unassigned mafic volcaniclastic rocks (Bruce, 1918; Simard et al., 2010). The small (70 by 12 m wide) ore lens dips steeply (75–80°E) with a moderate (45–50°N) northerly plunge and is interpreted as a dragfold within a north-northwest-trending fault zone (Bruce, 1918) with late sinistral movement (Cole et al., 2007, 2008; refer to 'Structural history' above). Chlorite is the dominant alteration mineral, with local zones of sericite and late quartz, calcite, dolomite and selenite (Alcock, 1930). Major and minor sulphide mineral assemblages are similar to those in the Flin Flon camp but with differences in mineral chemistry, however, this deposit has fewer trace minerals and is characterized by the presence of Co, Ni and Fe antimonide minerals and lack of Ag, Au and Pb antimonide minerals (Ames et al., 2009). Gold occurs in the mineral electrum and trace sylvanite $(Au,Ag)_2Te_4$.

Volcanogenic massive sulphide (VMS) prospectivity

The exploration potential for VMS mineralization in the area encompassed by the map is very high and includes the immediate area of the Flin Flon–Callinan–777 deposits and areas outside of the main camp including the Schist-Mandy area, the Western and Eastern Hook Lake successions and the area west of the Flin Flon Lake Fault.

Flin Flon–Callinan–777 area

The Millrock member of the Flin Flon formation is the most prospective stratigraphic interval in the Flin Flon district as it contains the Flin Flon, Callinan and 777 VMS deposits. Reconstruction of the volcanic architecture, alteration and deformation history, based on mapping, geochemistry and geochronology, provides unequivocal evidence for extension of an arc-like volcanic edifice and concomitant cauldron subsidence during VMS formation as defined by 1) a thick but areally restricted package of predominately mafic megabreccia deposits containing blocks up to 50 m in size; 2) basaltic and rhyolitic dike swarms and vents localized within the presumed core of the cauldron; 3) numerous synvolcanic faults; and 4) high-temperature epidote-quartz alteration. The VMS deposits formed within proximal rhyolitic centres, defined by coherent rhyolite and distal rhyolitic volcaniclastic facies that occur within distinct fault blocks in the larger cauldron. This stratigraphic interval has been repeated by two sets of thrust faults (Club Lake and Railway faults) and, although the panels between thrust faults are intact, they have been complicated by folding making exploration challenging.

Schist Lake–Mandy area

The Schist Lake and Mandy deposits occur within an unassigned stratigraphic succession that cannot confidently be correlated with the stratigraphy established for the main mining camp (Bailes and Syme, 1989; Simard, 2006a; Cole et al., 2007, 2008; Kremer and Simard, 2007; Simard et al., 2010; DeWolfe, 2011). The deposits are hosted within a primarily west-facing heterolithic mafic volcaniclastic succession that lies on the eastern side of the Mandy Road Fault. Depending on the movement attributed to this fault, the Schist-Mandy succession may represent either a separate VMS-hosting stratigraphic interval distinct from that of the main camp or, if movement along the fault is limited, the Schist-Mandy mafic volcaniclastic succession would lie below a succession of basaltic flows that is correlated with the Hidden formation (Bailes and Syme, 1989; Simard, 2006a; Cole et al., 2007, 2008; Kremer and Simard, 2007; DeWolfe, 2009, 2010). In the latter case, the Schist-Mandy succession could be the stratigraphic equivalent of the Millrock member in the main camp. The heterolithic volcaniclastic successions east and west of the Mandy Road Fault are both similar in their clast population, depositional units and chemistry.

Western and Eastern Hook Lake successions

Reconstruction of the subvolcanic architecture based on recent mapping, results from 3-D seismic surveys, and new radiometric dating of the Cliff Lake plutonic suite and volcanic strata of the Western Hook Lake succession, into which the intrusion was emplaced, suggest that 1) the Cliff Lake plutonic suite is a multiphase, sill-like tonalite/trondhjemite-quartz gabbro synvolcanic, subvolcanic intrusion; 2) the western or upper margin of the intrusion is, in part, enveloped within a high temperature(?) reaction zone characterized by semiconformable silicification, epidote-quartz and chloritic alteration; 3) volcanic strata of the Western Hook Lake succession can be tentatively correlated with strata of the Flin Flon mining camp; 4) the Cliff Lake Fault is a reactivated thrust fault that placed somewhat older (deeper) strata of the Hook Lake successions, containing the Cliff Lake subvolcanic pluton, upon somewhat younger VMS-hosting strata of the Flin Flon mining camp; and 5) the Cliff Lake plutonic suite, like synvolcanic subvolcanic plutons in many other VMS districts (e.g., Snow Lake, Noranda) represents the intrusive equivalent of a magma chamber that was emplaced into its own volcanic pile during a period of magmatic resurgence that followed cauldron subsidence and VMS deposit formation (Gibson et al., 2009). Analogous to other VMS districts, strata of the Western Hook Lake succession that are within 2000 m of, and overlie, the Cliff Lake Plutonic suite, are most prospective for VMS deposits (Franklin et al., 2005).

Rocks of the Eastern Hook Lake succession have been heavily explored for copper and zinc since the 1930s. Primary sulphide occurrences associated with rhyolite flows are present as discontinuous lenses and lobes of disseminated (<10%), stringer to locally semimassive (20–30%) pyrite±pyrrhotite, sphalerite and chalcopyrite; mineralization occurs either at the contact between massive and fragmental rhyolite flows, within fragmental rhyolite or between fragmental rhyolite and overlying volcanic-volcaniclastic rocks. An example of the latter occurs approximately 300 m west of Big Island Lake (known unofficially as Manistikwan lake), just south of the Cliff Lake plutonic suite. This sulphide occurrence has been previously documented (Gale and Eccles, 1988a, occurence no. 61) and been the focus of drilling by Hudson Bay Exploration and Development Co. Ltd. in the early 1980s. Parts of the zone have prominent sphalerite mineralization, with assays up to 12% Zn reported by Gale and Eccles (1988a). The association of mineralization with the fragmental rhyolite flows suggests potential for economic volcanogenic massive sulphide–type mineralization in the area. Moreover, recent work on the Eastern Hook Lake succession demonstrated that these rocks are younger than the rocks of the Flin Flon mining camp, but lithostratigraphically and age correlative to those that host the Trout Lake VMS deposits 10 km to the north (Kremer and Simard, 2007; Simard, 2009; Simard and MacLachlan, 2009; Rayner, 2010).

West of the Flin Flon Lake Fault

Recent detailed mapping west of the Flin Flon Lake Fault (MacLachlan, 2006c; MacLachlan and Devine, 2007), and U-Pb geochronology (Rayner, 2010) have demonstrated that this succession is lithostratigraphically equivalent to that which hosts the 777, Callinan and Flin Flon deposits. Furthermore, this succession is also characterized by extension, synvolcanic faulting, localized synvolcanic dike swarms and a zone of semiconformable silicification, which is similar to that observed in the Noranda mining camp and in the footwall to other major VMS systems (MacLachlan, 2006c). Furthermore, the silicification occurs in the footwall of the Newcor deposit, which is characterized by massive arsenopyrite. Although this deposit is structurally controlled, MacLachlan (2006c) has suggested that it may have remobilized or overprinted VMS mineralization, however, there is not enough evidence to demonstrate this, as the deposit is not exposed and there is very little documentation of the deposit from the time it was explored and mined. Major zones of silicification have also been documented in other parts of this succession, most notably within spatter breccias adjacent to a synvolcanic fault in the Blue Lagoon member just east of the southern end of Hilary Lake.

Gold deposits

There have been three gold deposits mined in the Flin Flon area over the last 50 years, the Newcor, Rio and Henning-Maloney mines, all located in Saskatchewan in the vicinity of Douglas Lake (Figure 3). The Rio and Newcor mines briefly went into production in 1984 and 1947, respectively, and a small amount of gold was produced from all three mines during test production. The total gold production from the area is thus small and not well documented.

Newcor deposit

The Newcor deposit is a quartz-arsenopyrite vein-type deposit (Byers et al., 1965). It occurs in a north-northwest-striking, steeply southwest-dipping shear zone within the Hidden formation on the eastern shore of Douglas Lake (Figure 3). The following description of the deposit is from Byers et al. (1965):

> *Mineralization along the main fault consists of a lenticular vein up to 30 inches wide, composed of quartz interlayered with massive arsenopyrite and slices of altered wall rock. The ore has a banded structure consisting of alternating layers of quartz and arsenopyrite, locally with thin seams of pyrite and sphalerite. The mafic volcanic wall rocks are altered to chlorite schist over widths up to 6 feet. The schist contains stringers, lenses and veins composed of quartz and arsenopyrite, or massive arsenopyrite with very little quartz.*

There is pervasive synvolcanic silicification within the Hidden formation east of Douglas Lake (MacLachlan, 2006c; MacLachlan and Devine, 2007), which is similar to that in the footwall of major VMS deposits in the Noranda mining camp (MacLachlan, 2006c). The gold-copper-zinc-arsenics metal association of the Newcor deposit is similar to VMS deposits in the Flin Flon mining camp and suggests the possibility that this structurally controlled deposit could have remobilized synvolcanic VMS mineralization.

Rio deposit

The Rio deposit (Bootleg mine) occurs along the Rio Fault, which strikes north-northeast just south of Douglas Lake (Figure 3). The fault cuts across volcanic and volcaniclastic rocks of the Hidden member of the Flin Flon formation and the Douglas formation (Simard et al., 2010), which strike north-northwest and dip steeply west. South of the Rio Fault, the volcanic rocks are intruded by the Phantom Lake intrusive suite (Galley et al., 1989). Along much of its length, the Rio Fault forms the contact between volcanic rocks to the northwest and the Phantom lake intrusive suite to the southeast, however, the Rio deposit occurs in an embayment along the intrusive contact of the pluton, such that volcanic rocks occur on both sides of the fault.

The Rio Fault comprises two discrete north-northeast-trending structures, one to the northwest and one southeast of the deposit (Stockwell, 1960), which are characterized by zones of intense shearing about 5 m wide, with very little alteration (Pearson, 1984). In the area of the deposit, the two structures overlap and form a broad zone of intense shearing, fracturing and alteration up to 90 m wide (Pearson, 1984). Away from the structures, there are numerous minor fractures with several preferred orientations that are presumed to be related to movement on the Rio Fault (Pearson, 1984). According to Pearson (1984), the orebodies are associated with a wide zone of silica-ankerite alteration. The ankerite occurs as disseminations and veinlets within the silica alteration. The style of alteration in the Rio and Henning-Maloney deposits corresponds to the third phase of mineralization and alteration recognized regionally by Galley and Franklin (1987), which they interpreted to be controlled by a conjugate system of dextral north-northeast-trending and sinistral east-trending faults.

Henning-Maloney deposit

The Henning-Maloney deposit occurs in an unnamed, north-northeast-trending fault, less than 1 km south of, and parallel to, the Rio Fault just south of Douglas Lake (Figure 3). The fault and deposit are entirely hosted within diorite of the Phantom Lake intrusive suite (Byers et al., 1965). The deposit is characterized by shear zones in two orientations, one set strikes north-northeast and dips 65–80° southeast and the other strikes northeast and dips 58–70° southeast (Byers et al., 1965). Alteration is typified by quartz-chlorite-ankerite-pyrite (Ansdell and Kyser, 1992) with associated quartz-carbonate veins up to several inches wide (Byers et al., 1965). According to Galley and Franklin (1987), gold zones at Henning-Maloney plunge steeply southwest.

Mineral occurrences and showings

A number of gold, tungsten, copper, zinc and arsenic mineral occurrences/showings have been documented in the Flin Flon area (Figure 30). The following section highlights some of them.

The Cor, Unity and Bomber gold or copper-gold showings, located east of Douglas Lake (Figure 30), have similar mineralogy, structural setting and alteration to the Newcor deposit. According to Pearson et al. (1986), these showings and the Newcor deposit predate intrusion of the ca. 1840 Ma Phantom Lake intrusive suite (Galley and Franklin, 1987; Galley et al., 1989).

There are a number of tungsten and copper±gold showings south of the town of Flin Flon in the Phantom Lake area (Figure 30; Geoscientific Map MAP2010-1, in back pocket) including 1) the Dion Lake copper showings; 2) several gold showings along the northeastern extent of the Dion Lake Fault; 3) several tungsten showings along the western side of the Green Lake Peninsula; and 4) the Douglas east scheelite showings about a kilometre east of the Newcor Mine. The tungsten and copper±gold showings in the area are spatially associated with the Phantom Lake intrusive suite, and Galley and Franklin (1987) suggested that they formed as part of a high-level porphyry system during the final stages of intrusion and cooling. In some cases the porphyry related mineralization has been overprinted by later gold mineralization related to faulting along north-northeast-trending and east-trending faults such as the Rio and Dion Lake structures, resulting in deposits such as the Rio and Henning-Maloney, both past-producing mines (see previous section), and possibly the unnamed gold showings along the northern extent of the Dion Lake Fault.

Numerous copper occurrences were documented in the southern extension of the Western Hook Lake succession north of the Milliken Lake Fault (Figures 23, 30). These occurrences are found in mafic flows and heterolithic breccias, along with some felsic intrusive rocks, a stratigraphic succession commonly associated with synvolcanic subsidence structure(s) in this area, which highlights the VMS potential of these rocks (Kremer and Simard, 2007).

Some gold occurrences have been documented within the eastern lobe of the Cliff Lake plutonic suite (Figures 26, 30), where the quartz-phyric leucotonalite is cut by irregular veinlets of coarse, white quartz, and vein-like bodies of fine-grained chlorite. Trenches have exposed 15–25 cm white quartz veins with pyrite- and tourmaline-filled joints. Locally, the porphyry contains several chloritic and pyritic shear zones. The hostrock shows trace amounts of pyrite throughout in close proximity to the quartz veins.

Conclusion

The Flin Flon area provides a unique opportunity to study the hostrocks of volcanogenic massive sulphide (VMS) deposits. The metamorphic grade is low, the primary structures are well preserved, and there is a very high percentage of clean exposure. All of these factors were essential and facilitated detailed stratigraphic and structural mapping of the area, allowing for major breakthroughs in the understanding of local VMS-hosting stratigraphy, leading to a better understanding of the environment of and controls on VMS formation and the recognition of other prospective areas.

The stratigraphy and geochemistry of the volcanic rocks associated with the Flin Flon–Callinan–777 VMS deposits record the infilling of a subsidence basin with abundant volcaniclastic material, localized felsic magmatism and the development of an intense hydrothermal alteration system (Flin Flon formation; Bailes and Syme, 1989; Syme et al., 1999b; Devine, 2003), which terminated with a hiatus in volcanism and formation of the VMS deposits. Following VMS deposition there was resurgence in volcanism and subsidence marked by the development of one or more mafic shield volcanoes atop this subsidence structure (Hidden and Louis formations; Syme et al., 1999b; DeWolfe and Gibson, 2005, 2006; DeWolfe, 2008; DeWolfe et al., 2009a, b).

Reconstruction of the volcanic architecture, based on mapping, geochemistry and geochronology, provides unequivocal evidence for extension of an arc-like volcanic edifice and concomitant cauldron subsidence during VMS formation in Flin Flon as defined by 1) a thick but areally restricted package of predominately mafic megabreccia deposits containing blocks up to 50 m in size; 2) internal angular unconformities; 3) basaltic and rhyolitic dike swarms and vents localized within the presumed core of the cauldron; 4) numerous synvolcanic faults; and 5) high-temperature epidote-quartz alteration (Gibson et al., 2009). The VMS deposits formed within proximal rhyolitic centres, defined by coherent rhyolite and more distal rhyolitic volcaniclastic facies that occurred within distinct fault blocks in the larger cauldron.

Detailed facies mapping allowed for recognition of synvolcanic structures and associated synvolcanic subsidence/depositional basins, and proved to be critical in identifying new areas with increased prospectivity. Such features are

1. synvolcanic faults, which are required for cross-strata permeability and to act as magma and hydrothermal fluid pathways; in the Flin Flon area, these are marked by abrupt changes in volcanic/volcaniclastic lithofacies along strike;
2. synvolcanic grabens, which provided evidence of extension and rifting, or the thinning of the crust that is needed for high heat flow; the thick succession of

volcaniclastic rocks that overlies VMS deposits and, the fact that they change abruptly along strike, indicate different depositional basins;

3. hydrothermal alteration, which is evidence of an active, synvolcanic hydrothermal system; alteration in the Flin Flon camp occurs as zones of semiconformable, strong to pervasive chlorite alteration, and conformable quartz-epidote alteration; and
4. a hiatus in effusive volcanism, which allowed for the concentration of mineralization along a specific stratigraphic horizon; evidence of this is provided by the large volume of tuffaceous volcaniclastic rocks overlying the deposits (DeWolfe, 2011).

The increased understanding of the VMS-hosting stratigraphy achieved by the recent remapping has in turn allowed for a greater understanding of the complex polyphase structural history of the area. Although previously recognized, this study reinforces the importance of numerous thrust panels within the polyphase, imbricated sequence of the Flin Flon area at various scales, of which at least two host VMS mineralization.

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