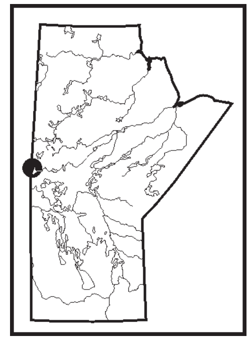
GS-2 Physical description of the 1920 member, Hidden formation, Flin Flon, Manitoba (NTS 63K16SW)

by Y.M. DeWolfe[1] and H.L. Gibson[1]



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Summary

The 1920 member occurs at the base of the Hidden formation, which defines the onset of hangingwall volcanism and occurs stratigraphically above the volcaniclastic rocks of the Flin Flon formation. The 1920 member, or icelandite unit, occurs as massive, columnar jointed, pillowed and peperite facies; is overlain locally by felsic or undifferentiated volcaniclastic rocks; and occupies a small, localized, synvolcanic subsidence structure.

Economic implications with respect to the 1920 member include the following:

- Structures that controlled the graben location may have been present during footwall volcanism and therefore may have controlled mineralization.
- Thrust faults that have structurally repeated the stratigraphy of the Hidden formation may also have repeated the horizon that hosts the deposits of the Flin Flon camp.
- The volcaniclastic rocks that overlie the 1920 member may be prospective along strike at depth, since they represent a hiatus in volcanism and occur within a topographic basin, near a volcanic vent and in an area of hydrothermal activity.

Introduction

The Hidden and Louis formations, which together constitute a voluminous basaltic hangingwall succession to the Flin Flon, Callinan and Triple 7 volcanogenic massive sulphide (VMS) deposits (Figures GS-2-1, -2), are an integral part of the volcanic and subsidence history that characterizes the ore-forming environment at Flin Flon. Reconstructing the volcanic, magmatic, subsidence and hydrothermal history of the basaltic hangingwall, particularly the identification of synvolcanic structures that may mark hydrothermal corridors localizing buried VMS deposits, will not only aid exploration in the Flin Flon camp but add to the understanding of VMS deposit formation and the factors that control the localization of VMS deposits during the evolution of submarine volcanic complexes.

The objectives of this project are to

- reconstruct the volcanic, subsidence and hydrothermal history of the Hidden and Louis formations, which constitute a large shield-like volcanic edifice, and determine their magmatic evolution;
- establish a litho- and chemostratigraphy that will aid in stratigraphic correlation within the formations and to more distant areas within the volcanic edifice (i.e., Myo Lake and Schist Lake–Mandy areas);
- define synvolcanic structures that may control the location of more deeply buried VMS deposits; and
- provide a preliminary assessment of the VMS potential of the Hidden and Louis formations.

This project builds on the previous work of Bailes and Syme (1989), and that of the Flin Flon Targeted Geoscience Initiative, in particular the work of Ames et al. (2002) and Tardif (2003). The Flin Flon Targeted Geoscience Initiative was a three-year (2000–2003) federal initiative designed to develop new criteria to assist exploration and mining companies in the expansion of current ore reserves and in the exploration for new deposits, in an effort to stabilize the economy of Flin Flon and surrounding northern communities. The purpose of this paper is to describe the field characteristics of the 1920 member of the Hidden formation.

Methodology

In order to reconstruct the volcanic and subsidence history, it is essential that individual flow units within the Hidden and Louis formations be defined and mapped. This is particularly true for the 1920 member, which is a lithologically distinct unit located in the lowermost Hidden formation. Its distribution and emplacement mechanisms will provide

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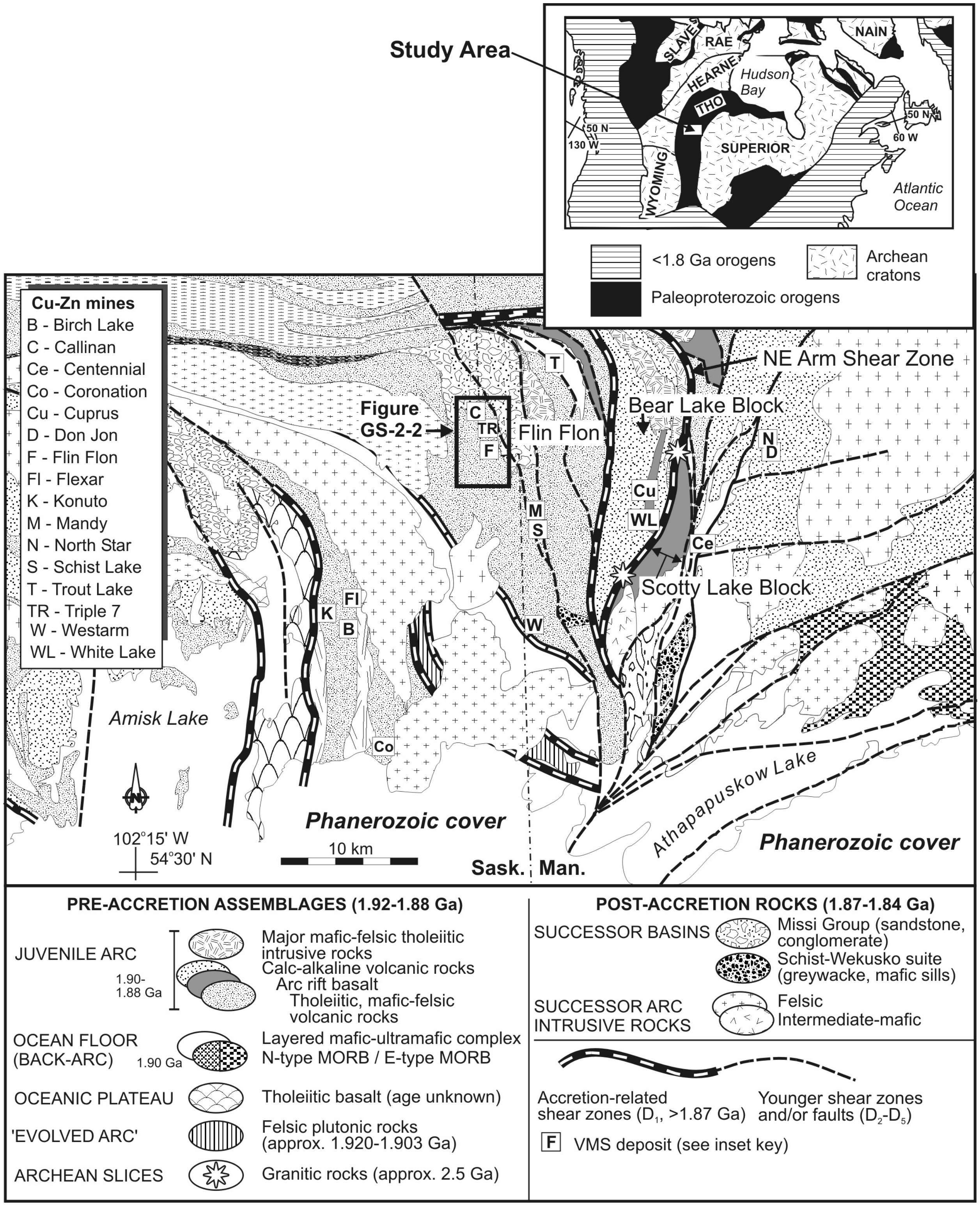

Figure GS-2-1: *Geology of the Flin Flon Belt, showing the locations of known volcanogenic massive sulphide (VMS) deposits (*modified from *Syme et al., 1996). Box indicates area covered by Figure GS-2-2.*

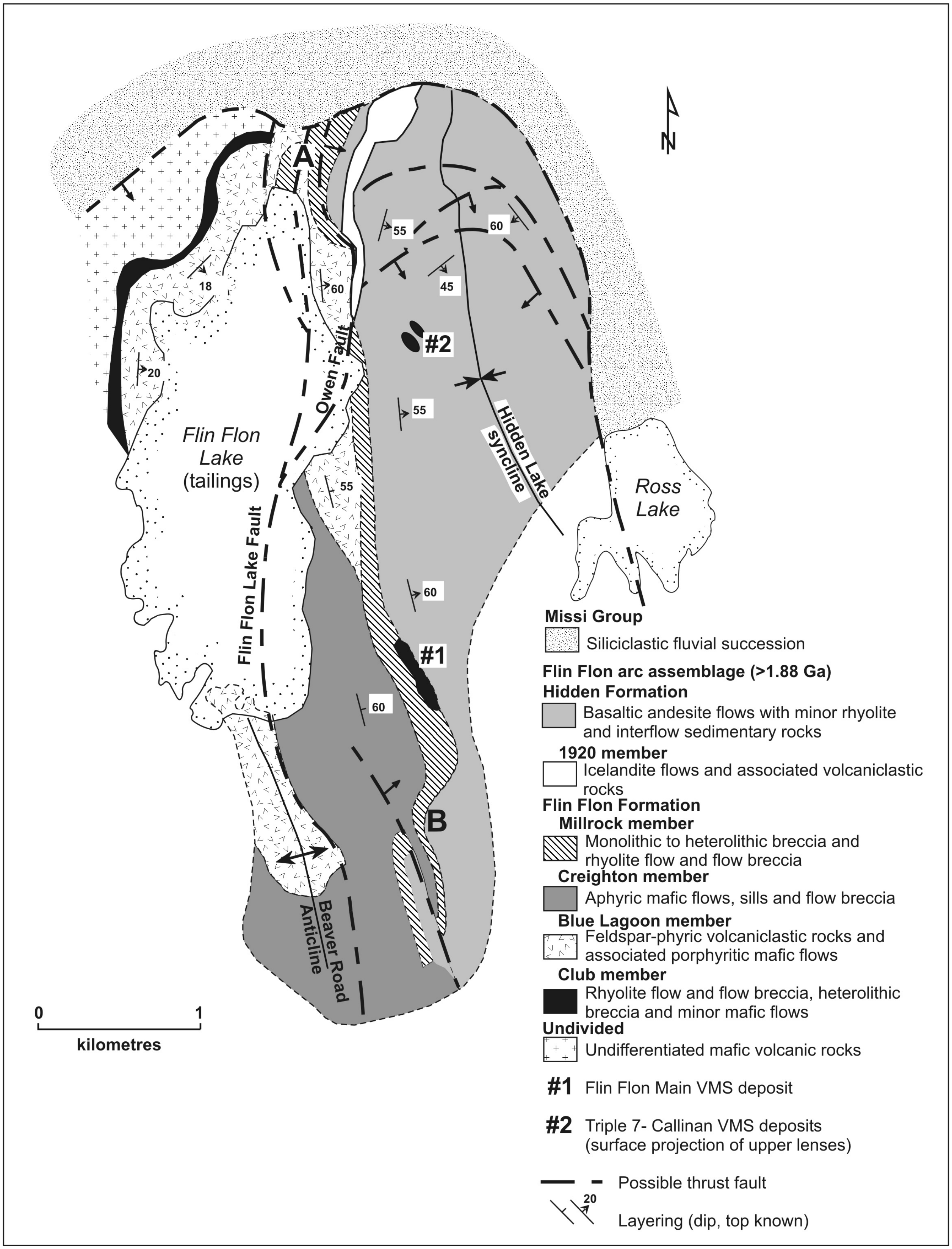

Figure GS-2-2: *Preliminary geology of the Flin Flon area, showing the surface exposure of the 1920 member or icelandite unit (*modified from *Bailes and Syme, 1989). See Figure GS-2-1 for location.*

valuable information regarding the earliest stages of Hidden formation volcanism and subsidence.

The 1920 member, or icelandite unit, has a strike length of approximately 1 km and is 0.7 km wide; all surface exposures of the icelandite unit were mapped at a scale of 1:1000. In addition, a smaller area (100 m by 50 m) of massive to in situ–brecciated icelandite and volcaniclastic rocks was mapped in detail (1:250 scale) to better understand the mechanisms of emplacement of the 1920 member and peperite formation. Representative samples (82) from all flows, sills and dikes within the mapped area of the Hidden formation were collected for geochemical and petrographic analyses.

In the description of rock types that follows, the terms tuff, lapilli tuff, lapillistone and tuff-breccia are used in a nongenetic sense, following the granulometric classification of volcaniclastic rocks proposed by Fisher (1966).

Regional geology

The Paleoproterozoic Flin Flon greenstone belt contains 27 known VMS deposits and is part of the southeastern Reindeer Zone of the Trans-Hudson Orogen (Figure GS-2-1). The greenstone belt consists of a series of terranes that range in age from 1.9 to 1.84 Ga (Stern et al., 1995) and include arc, back-arc, ocean-floor and successor-arc successions. The Flin Flon and Snow Lake ocean-arc assemblages contain the majority of the 27 known VMS deposits.

The study area is located in the Flin Flon arc assemblage, which is composed of juvenile metavolcanic rocks (1.91 to 1.88 Ga) unconformably overlain by fluvial sedimentary rocks of the Missi Group (ca. 1.84 Ga). The rocks of the Flin Flon arc assemblage are interpreted to have been erupted and emplaced in an island-arc–back-arc setting (Syme and Bailes, 1993), and consist of basalt, basaltic andesite flows and breccia, with lesser rhyolite flows. The Flin Flon, Callinan and Triple 7 VMS deposits, which total more than 85.5 million t grading 2.2% Cu, 4.3% Zn, 2.49 g/t Au and 38.16 g/t Ag, are interpreted to have formed during a period of localized rhyolitic volcanism in a synvolcanic subsidence structure, or caldera, within a much larger, dominantly basaltic, central volcanic complex (Bailes and Syme, 1989; Devine et al., 2002). The Hidden and Louis formations appear to have been erupted during a period of resurgent basalt volcanism and subsidence that immediately followed a hiatus in volcanism marked by VMS ore deposition.

Stratigraphy of the hangingwall

The hangingwall to the Flin Flon orebodies consists of the Hidden and Louis formations. From oldest to youngest, the Hidden formation consists of the 1920, Bomber, Newcor and Tower members (Figure GS-2-3). The Hidden formation is dominated by basaltic and andesitic flows, with minor interflow tuff and rhyolite (Ames et al., 2002). The exposed thickness of the Hidden formation ranges from 400 to 880 m and is complicated by D_2 folding and D_1 thrust faulting. The lower mafic flows of the Hidden formation are in contact with thick deposits of plane-bedded, water-laid tuff that constitutes the uppermost unit of the Millrock member of the Flin Flon formation (Devine et al., 2002). The upper contact of the Hidden formation is defined by the first appearance of the pyroxene-plagioclase–phyric basalt of the Louis formation (Ames et al., 2002).

The Louis formation consists of pyroxene-plagioclase–phyric basaltic flows and sills with lesser aphyric pillowed basalt, scoriaceous amoeboid breccia and scoriaceous lapilli tuff. Globular peperite is common in the lower Louis formation, and interpillow tuff extends up to 120 m above the contact between the Hidden and Louis formations (Ames et al., 2002).

The Hidden and Louis formations contain epidote-quartz alteration, silicification, chloritization, feldspar-quartz alteration and sulphide mineralization. A pervasive middle greenschist facies metamorphic mineral assemblage of actinolite-epidote-albite-quartz±biotite is characteristic of the regional metamorphism (Bailes and Syme, 1989).

1920 member

The 1920 member is critical in understanding the Hidden formation, as it is thought to define the onset of hangingwall volcanism and is readily recognizable in the field as an amphibole-porphyritic or amphibole 'needle'–bearing rock. The amphibole may be phenocrysts or porphyroblasts; future petrographic work will answer this question. The 1920 member is relatively enriched in TiO_2, Fe_2O_3 and P_2O_5 compared with the mafic volcanic rocks of the Hidden formation. It has an average SiO_2 content of 55.50 wt. % and can therefore be described as icelandite (basaltic andesite; Wyman, 1993).

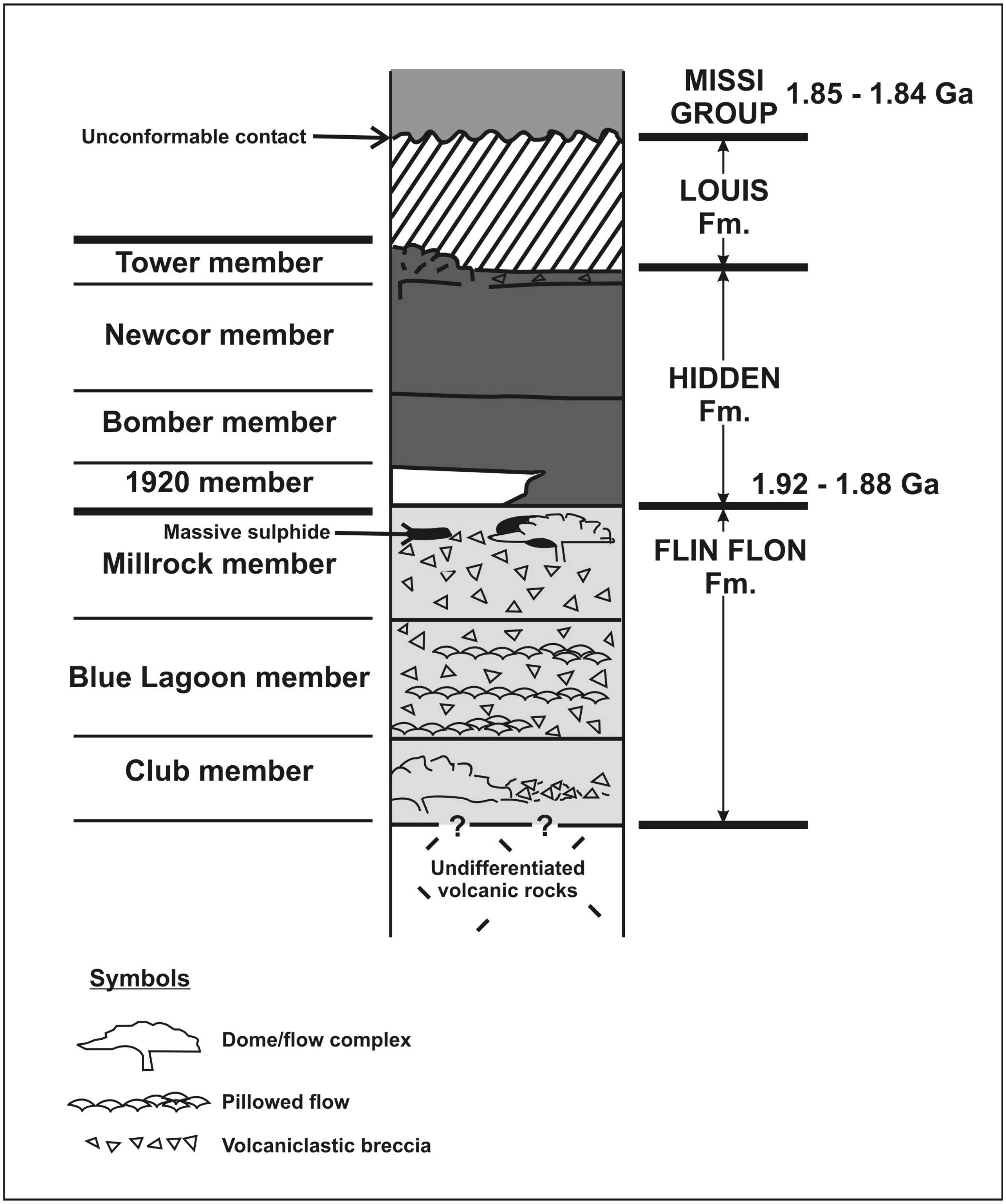

Figure GS-2-3: *Schematic stratigraphic section of the Flin Flon mining camp (*modified from *Devine, 2003).*

Distribution

The 1920 member is an aerially restricted unit located on the west limb of the Hidden Lake syncline and north of the Railway Fault (Figure GS-2-2; Price, 1997). It has a northeast strike, a maximum thickness of 90 m (not true thickness) and an approximate strike length of 1.1 km (Figure GS-2-2).

Stratigraphically, the 1920 member is located at the base of the Hidden formation, where it occurs stratigraphically above volcaniclastic rocks of the Millrock member of the Flin Flon formation (Figure GS-2-3). The top of the 1920 member is marked by a felsic volcaniclastic unit and by mafic sills of the Bomber member or mafic volcaniclastic rocks overlain by plagioclase-porphyritic flows of the Newcor member. The 1920 member and these 'conformably' overlying units are recognized in two separate fault blocks separated by two northeast-striking thrust faults that structurally repeat the 1920 member and the bounding basalt sills or flows.

Lithology

The 1920 member ranges from fine- to medium-grained icelandite and occurs as massive, columnar jointed, pillowed and peperite facies.

Contact relationships

The 1920 member within the western fault block has an upper contact that grades into an intermediate to mafic volcaniclastic facies that is, in turn, conformably overlain by the massive and pillowed, plagioclase-phyric basalt flows of the Newcor member. The lower contact of the icelandite is in fault contact with aphyric basalt in the northernmost exposure of the map area. Elsewhere along this contact, aphyric basalt occurs in sharp contact with the icelandite. Locally, there are large (0.5–1.0 m) lenses of finely laminated mafic tuff along this contact. Within 5 m either side of the contact, both the aphyric basalt and the icelandite contain irregular pods, lenses and 'stringers' (fractures) of mafic tuff. On the maps, these units are referred to as having 'included tuff'.

In the eastern fault block (across a northeast-striking and east-dipping thrust fault), the 1920 member has an upper contact that grades from massive icelandite to an icelandite peperitic facies over the upper 10 m. This is conformably overlain by a pumice-bearing, felsic volcaniclastic unit that is approximately 2 m thick (Figure GS-2-4a). Blocks, pods and irregular lenses of the felsic volcaniclastic unit occur within the upper 10 m of the icelandite. The contact with the underlying aphyric basalt is sharp; however, the upper 5 m of the basalt contains inclusions of mafic tuff and is peperitic.

Massive facies

The massive facies of the icelandite is fine grained and contains 20–40% black, needle-shaped amphibole crystals, ranging from less than 1 to 5 mm long, that are randomly oriented within the groundmass (Figure GS-2-4b). The weathered surface of the icelandite is brown and the fresh surface has a very distinctive metallic blue cast. The massive facies ranges from nonamygdaloidal near the base of the unit to a content of 30% amygdules near the top (Figure GS-2-4c). Amygdules are quartz filled and oval to round in form, and range from 0.2 to 8 cm in size. Amygdules are commonly stretched and aligned almost perpendicular to the strike of the unit, and occur concentrated in metre-wide bands that define lobe-like forms. The orientation of the stretched amygdules and bands of amygdules is approximately 135°, parallel to the strike of S_2 foliation.

Within the massive icelandite there are also localized areas (10 m by 10 m) of columnar jointing (Figure GS-2-4d). These areas are not amygdaloidal. Locally within the massive icelandite are blocks of mafic tuff that are finely laminated, light brown in colour and less than 50 cm in size (Figure GS-2-4e).

Both massive and columnar-jointed icelandite contain weak to strong, patchy epidote-quartz alteration, with the abundance of patches increasing upward within the unit. The epidote-quartz patches are light brown to cream coloured on the weathered surface, 5–80 cm in size and round in form (Figure GS-2-4f). The alteration patches are commonly elongate parallel to the dominant S_2 foliation in the area (Figure GS-2-4a).

Pillowed facies

At one location, the icelandite unit contains pillows that are 30–50 cm wide, grey (weathered) and metallic blue (fresh) in colour, with 1–2 cm wide, dark grey chloritized selvages (Figure GS-2-5a). The pillows are aphanitic to fine grained and only locally exhibit the distinct, black, needle-shaped amphibole crystals observed in the massive facies.

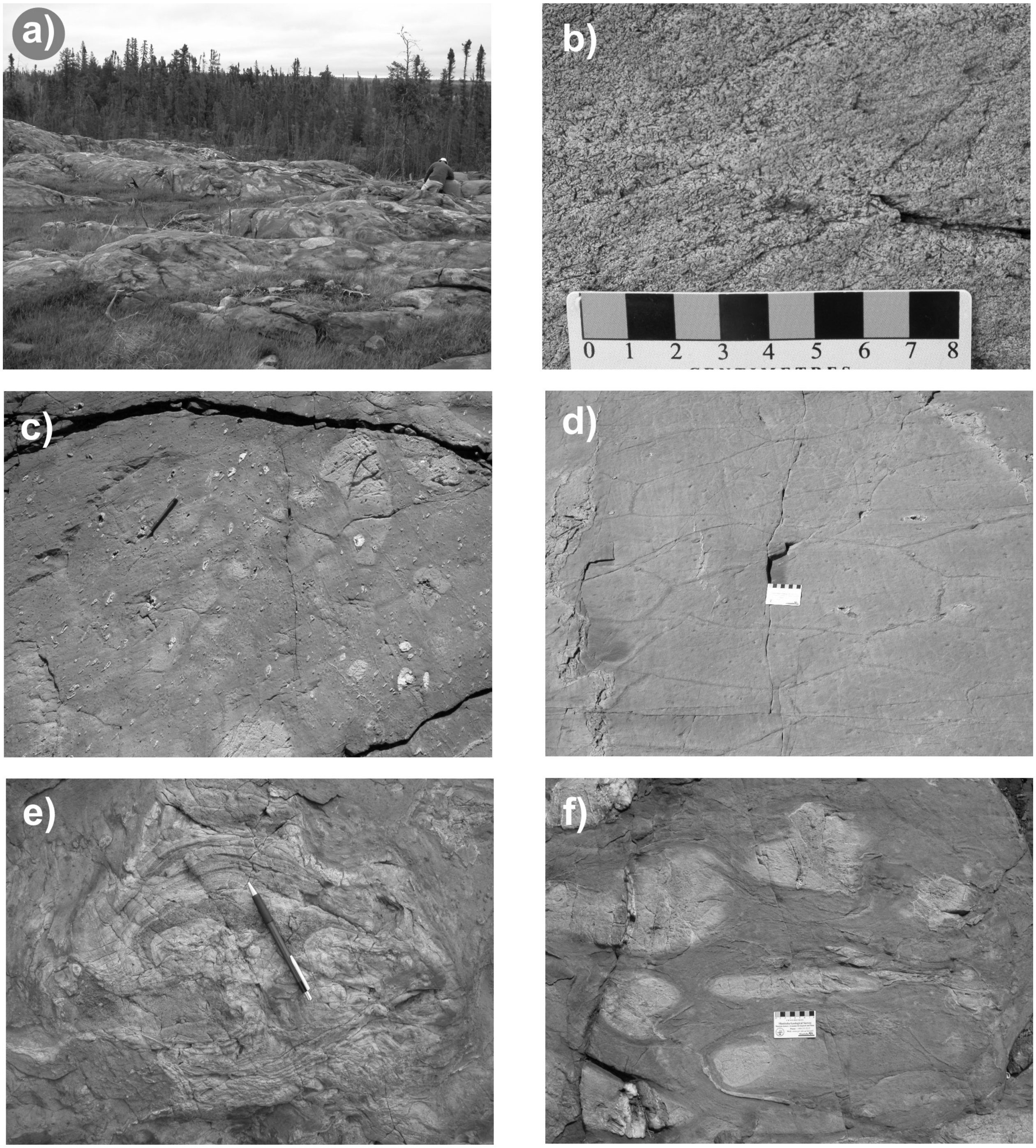

Figure GS-2-4: *Photographs of the 1920 member, Hidden formation, Flin Flon:* ***a)*** *strong, patchy epidote-quartz–altered icelandite; person marks the upper contact of the icelandite with the overlying felsic volcaniclastic unit;* ***b)*** *randomly oriented amphibole crystals in massive icelandite;* ***c)*** *massive, quartz-amygdaloidal icelandite with patchy epidote-quartz alteration; pencil is 12 cm long;* ***d)*** *columnar-jointed and weakly quartz-amygdaloidal icelandite;* ***e)*** *inclusion of finely laminated mafic tuff in massive icelandite; and* ***f)*** *patchy epidote-quartz alteration in icelandite.*

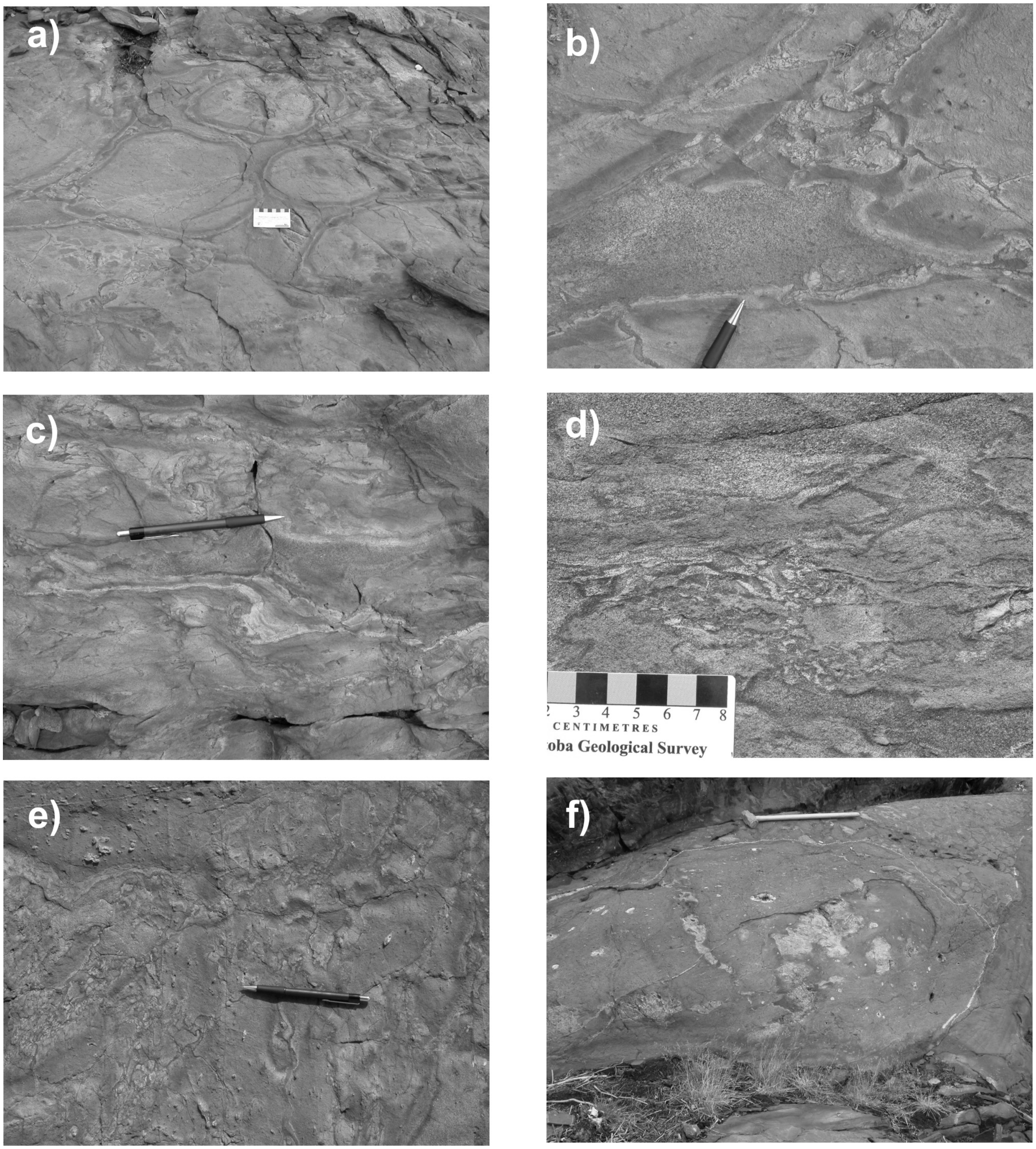

Figure GS-2-5: *Photographs of the 1920 member, Hidden formation, Flin Flon:* ***a)*** *pillows of icelandite with interpillow hyaloclastite and interpillow breccia;* ***b)*** *interpillow hyaloclastite;* ***c)*** *included mafic tuff in pillowed icelandite flow;* ***d)*** *in situ–brecciated icelandite with hyaloclastite between the angular fragments;* ***e)*** *angular, jigsaw-puzzle–fitting fragments of in situ–brecciated icelandite with hyaloclastite between fragments;* ***f)*** *massive lobe of quartz-amygdaloidal icelandite (outlined in chalk) surrounded by in situ–brecciated icelandite (mafic dike on right side of photo); note that the amygdules are concentrated around the outside and parallel to the margin of the lobe or lava tube; hammer is 1 m long.*

Amygdules are concentrated near the margins of the pillows, and concentric cooling laminations filled with quartz are also common at the pillow margins. Between the pillows there is a dark grey, fine-grained material that is possibly altered hyaloclastite (Figure GS-2-5a, b). Locally, blocks of light brown, finely laminated tuff are included in the pillowed flow (Figure GS-2-5c).

In situ–brecciated facies

In situ–brecciated icelandite commonly occurs near the top (in the upper 20 m) of the unit. Icelandite fragments are angular or blocky in form, 1–25 cm in size, aphanitic and grey (weathered) or metallic blue (fresh) in colour, and have a jigsaw-puzzle fit (Figure GS-2-5d, e). Between the fragments is a fine-grained, dark grey, granular-textured matrix which was probably hyaloclastite. The hyaloclastite often displays angular or cuspate shards that are less than 2 cm in size and surrounded by fine-grained matrix (Figure GS-2-5d). The fine-grained matrix may be a mix of tuff and hyaloclastite.

These areas of in situ–brecciated icelandite are also moderately to strongly epidote altered, the alteration manifesting itself as 5–50 cm, light green to light brown, rounded patches. Within the area of in situ–brecciated icelandite are massive, 1–3 m wide lobes that have quartz-filled amygdules concentrated towards their margins; concentric cooling laminations filled with quartz also occur at the lobe margins (Figure GS-2-5f). The lobes extend downward into the massive facies of the icelandite unit.

Peperite facies

Peperite is 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 or volcaniclastic deposits (Skilling et al., 2002).

As previously mentioned, the upper contact of the icelandite is commonly peperitic. The peperite is a mix of aphanitic to fine-grained, dark grey, blocky or angular, commonly quartz-amygdaloidal (1–20% amygdules less than 2 cm in size) fragments of icelandite (1–30 cm) and finely laminated (millimetre scale), tan coloured, strongly epidote altered, blocky to fluidal fragments of tuff that are 1–40 cm in size (Figure GS-2-6a–d). The matrix to the icelandite and tuff fragments is hyaloclastite, which consists of cuspate, angular grey fragments, less than 1 cm in size, in a dark grey to black, fine-grained tuff matrix (Figure GS-2-4a, b).

Volcaniclastic rocks dominated by icelandite fragments

The peperite locally grades upward into an icelanditic tuff-breccia (Figure GS-2-6e), and mafic lapilli tuff and tuff. The tuff-breccia contains fine-grained, quartz-amygdaloidal (1–3% amygdules less than 1 cm in size), light grey or tan blocks or fragments of icelandite that are 1–30 cm in size and have 1–4 mm wide, dark grey chilled margins (Figure GS-2-6e). The matrix is a light brown or tan, lapilli-tuff–size volcaniclastic material that is strongly epidote altered and silicified (Figure GS-2-6e). The tuff-breccia is massive and its lower contact with the peperite is sharp but irregular where it appears to define a depression, with a maximum thickness of approximately 3 m, within the upper surface of the peperite.

The tuff-breccia fines upward over its upper 50 cm into a finely bedded, mafic lapilli tuff and tuff deposit that has a total thickness of approximately 2 m (Figure GS-2-6f). Individual lapilli tuff and tuff beds range from 0.1 to 1 m in thickness. The beds are grey and the tuff beds are finely laminated (millimetre scale; Figure GS-2-6f). The thicker lapilli tuff beds are massive and contain 5% angular to subrounded, fine-grained, dark grey clasts of aphyric basalt that are greater than 2 cm in size.

Interpretation

In the northern part of the map area, the icelandite unit has a thickness of 90 m, whereas, to the south, the maximum thickness of the unit is only 10 m (Figure GS-2-2). The significant thickening of the unit to the north is not due to late faulting and folding, and may therefore indicate its location within a primary paleotopographic feature, such as a subsidence structure or graben to the north.

The local presence of peperite along the upper contact of the massive and in situ–brecciated icelandite, and the inclusions of mafic tuff within the base of the icelandite, which itself is in contact with basaltic units containing inclusions of tuff and locally peperite, both indicate that the icelandite was emplaced as a high-level sill or sills into wet, unconsolidated volcaniclastic rocks on the seafloor. The volcaniclastic deposits (either intermediate to mafic or felsic) that

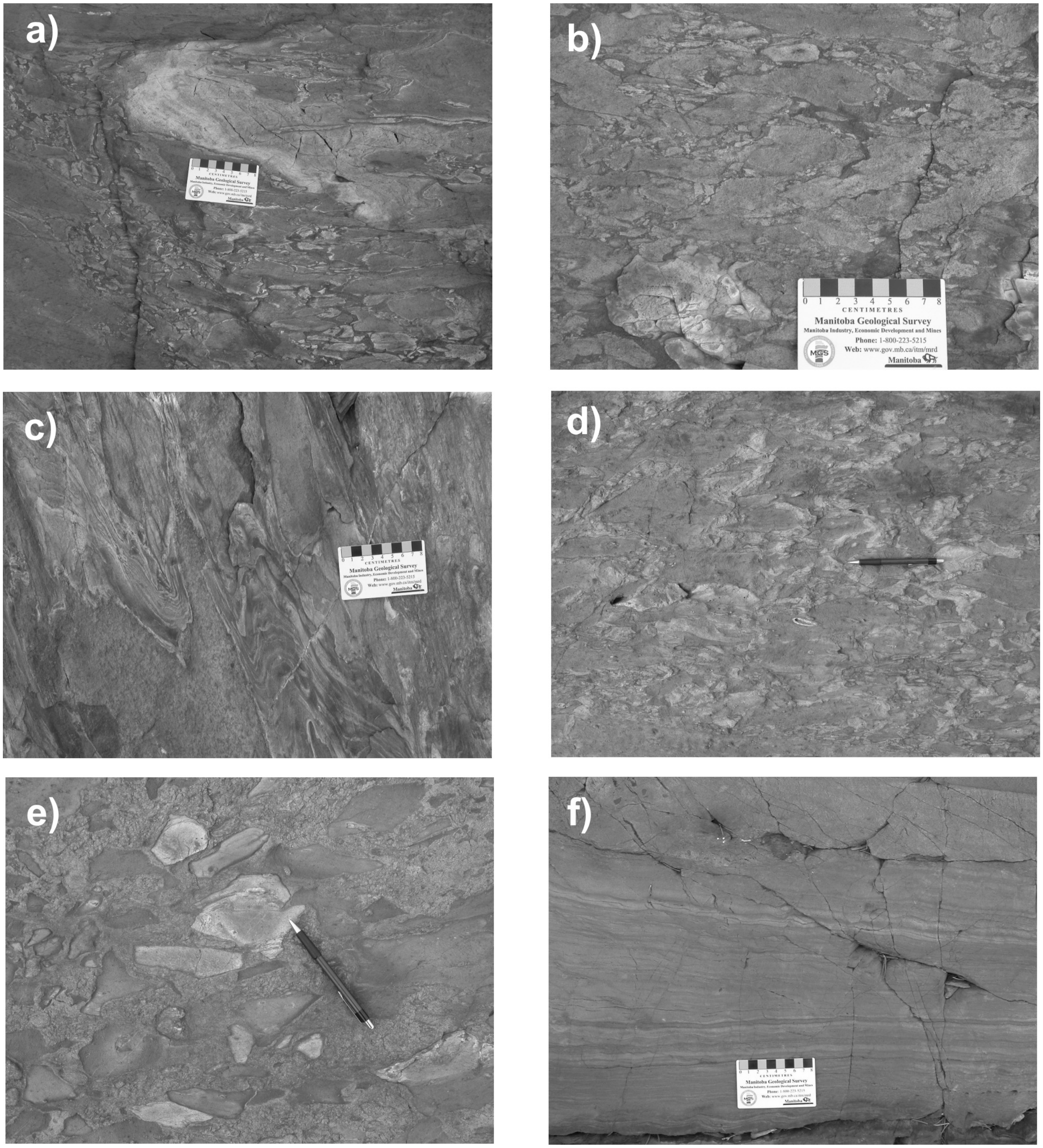

Figure GS-2-6: *Photographs of the 1920 member, Hidden formation, Flin Flon:* ***a)*** *peperite, comprising angular fragments of icelandite in a dark grey, fine-grained tuff and hyaloclastite matrix; lighter grey area is a block of epidote-quartz–altered tuff;* ***b)*** *peperite, comprising angular fragments of icelandite in a dark grey matrix of tuff and hyaloclastite;* ***c)*** *peperite, comprising fluidal-shaped blocks of finely laminated mafic tuff in in situ–brecciated icelandite;* ***d)*** *peperite, comprising a mix of angular fragments of icelandite (grey), epidote-quartz–altered tuff (light grey) and hyaloclastite (dark grey);* ***e)*** *angular fragments of icelandite in a tuff matrix;* ***f)*** *thin-bedded to finely laminated mafic tuff.*

now overlie the 1920 member within the two fault blocks were also deposited within, and define the topographic basins in which, the icelandite sill or sills were emplaced. In fact, using the volcaniclastic unit as a datum, the abrupt along-strike change in thickness from approximately 90 m in the north to only 10 m in the south suggests emplacement of the icelandite into a synvolcanic structural basin that deepened to the north. Emplacement of the icelandite resulted in the upward displacement of the volcaniclastic units during inflation to partially infill the structure. The occurrence of pillow facies to the south suggests that the icelandite unit locally broke out onto the surface, perhaps where it overran the walls of a confining basin or subsidence structure to the north.

Deposition of the 1920 member close to a basaltic volcanic vent is indicated, not only by its emplacement in an area of subsidence, but also by the abundance of basaltic sills and dikes within the member. These dikes often have very irregular and brecciated margins or contacts with the icelandite unit, suggesting that they are synvolcanic. An increase in the number of dikes to the north where the icelandite is thickest suggests a vent location coincident with the inferred graben to the north.

The peperite found in the 1920 member is comparable to other peperite, described in the literature, that involved the mixing of relatively fine grained sediment and rhyolitic to basaltic lavas, such as those at Kidd Creek, northeastern Ontario (DeWolfe, 2003); Onedin, northwestern Australia (Orth and McPhie, 2003); Passamaquoddy Bay area, southwestern New Brunswick (Dadd and Van Wagoner, 2003); Puna Highland, northeastern Argentina (Coira and Pérez, 2003); the Mount Read Volcanics, Tasmania; and the Green Tuff Belt, Japan (Gifkins et al., 2003).

In summary, field observations suggest that the 1920 member, or icelandite unit, was emplaced within, and largely confined to, a small synvolcanic graben that developed during subsidence at the onset of Hidden formation volcanism.

Economic considerations

The 1920 member, or icelandite unit, clearly defines a small, localized, synvolcanic subsidence structure that formed during the onset of Hidden formation volcanism. Structures that control the location of the subsidence structure may have also been operative during emplacement of the underlying Flin Flon formation, and may therefore have controlled the location of massive sulphide mineralization. Thrust faults that repeat hangingwall stratigraphy, particularly the 1920 member, may also structurally repeat the main mineralized horizon of the Flin Flon camp in the immediate mine area and elsewhere. Patchy yet strong epidote and quartz alteration within the 1920 member indicates that this area has also been an area of intense hydrothermal activity.

In addition, the volcaniclastic rocks that overlie the icelandite unit represent a hiatus in volcanism. Their occurrence within a topographic basin, near a volcanic vent, and in an area of hydrothermal activity indicates they may be prospective along strike at depth.

Acknowledgments

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