Petrology and U-Pb Geochronology of Footwall Porphyritic Rhyolites from the Wolverine Volcanogenic Massive Sulfide Deposit, Yukon, Canada: Implications for the Genesis of Massive Sulfide Deposits in Continental Margin Environments

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Abstract

Porphyritic rhyolite sills form an important component of the footwall of the Wolverine volcanogenic massive sulfide (VMS) deposit, Yukon, Canada, and occur proximal to mineralization in the immediate deposit area (Wolverine/Lynx zone) and at similar stratigraphic levels along strike (Fisher, Puck, and Sable zones). Porphyritic rhyolites are of two types: an older quartz-feldspar porphyritic (QFP) rhyolite suite; and a younger feldspar porphyritic (FP) suite. Both the QFP and FP suites of intrusions are semiconcordant, suggesting a sill-like morphology, and are altered and crosscut by veinlet mineralization, suggesting that they are pre- to syn-mineralization. The margins of QFP suite of intrusions contain minor xenoliths of surrounding shales and poorly developed chilled margins suggesting emplacement into partially consolidated sedimentary rocks, whereas the FP suite of intrusions shows well-developed chilled margins indicative of emplacement into fully solidified sedimentary rock. These features suggest that the QFP suite of intrusions represents an older phase of rhyolitic magmatism, whereas the FP suite represents a younger event. This is supported by U-Pb zircon ages, which indicate a 352.4 ± 1.5 Ma emplacement age for the FP suite and a ~347 to 346 Ma emplacement for the FP suite (two ages at 347.8 ± 1.3 and 346.0 ± 2.2 Ma). Both suites of porphyries have inherited Proterozoic zircon and have ratios of La/Sm (UCN ~1 and Nb/Th (UCN ~1 (UCN – upper continental crust normalized), indicating derivation from and/or extensive interaction with ancient upper continental crustal materials. The FP suite, however, has elevated high field strength element (HFSE) and rare earth element (REE) contents, high zircon saturation temperatures, and higher Nb/Ta ratios and lower Ti/Sc ratios than the QFP suite. These features are interpreted to reflect that the FP suite of magmas was hotter (>900°C) melts with a greater mantle component in their genesis. Both suites, however, are interpreted to have formed due to basaltic upwelling, crustal melting, and crust-mantle mixing during ensialic back-arc basin activity. The presence of mantle heat within the Wolverine basin from ~352 to ~347 to 346 Ma, a minimum of 5 m.y., suggests that sustained mantle heat flow was critical to the genesis of the Wolverine porphyries. It is also suggested that this sustained mantle heat was responsible for the Wolverine hydrothermal system and that upwelling mantle is essential in providing the heat to drive hydrothermal systems even in continental margin-type VMS environments (e.g., Bathurst, Iberian pyrite belt).

The Wolverine porphyries are among the most HFSE- and REE-enriched felsic rocks associated with VMS mineralization globally. The high HFSE and REE concentrations in these rocks are interpreted to be due to
high-temperature melting of continental crust and the efficient dissolution of HFSE- and REE-bearing accessory phases (e.g., zircon, monazite) during crustal melting. The presence of HFSE- and REE-enriched felsic rocks indicates thermally anomalous geodynamic settings, which may have had sufficient heat flow to form high-temperature crustal melts and, by association, the potential to generate robust and long-lived hydrothermal systems. Thus, in ancient continental margin settings identification of HFSE-REE–enriched rhyolites may be useful in outlining potentially prospective areas for VMS mineralization.

Formation of the Wolverine deposit, and other Devonian-Mississippian VMS deposits in the Finlayson Lake district, coincided with widespread extension in the Finlayson Lake region and elsewhere in the northern Cordillera. This extensional geodynamic activity was related to arc-rifting and ensialic back-arc basin activity in the Yukon-Tanana terrane and the opening of the Slide Mountain back-arc basin (ocean). The age of the Wolverine deposit is also broadly coincident with a global pulse of syngenetic VMS and SEDEX mineralization (e.g., Iberian pyrite belt, Sebelyn basin) and an episode of regional, and possibly global, ocean anoxia. The combination of (1) high-temperature magmatism capable of driving vigorous hydrothermal circulation, (2) widespread extension to provide structural conduits to focus hydrothermal discharge, and (3) anoxic bottom waters and abundant associated black shales to act as a trap to prevent hydrothermal fluid dissipation into the water column, were important factors in the formation of the Wolverine VMS deposit. These conditions existed elsewhere in the world during the Devonian-Mississippian and may explain the abundance of syngenetic sulfide deposits that formed at this time.

Introduction

The precious metal-rich Wolverine volcanogenic massive sulfide (VMS) deposit (Tucker et al., 1997; Bradshaw et al., 2001, in press) is one of the most recent base metal deposit discoveries in northern North America (6.2 Mt of 12.96% Zn, 1.53% Pb, 1.41% Cu, 359.1 g/t Ag, 1.81 g/t Au; Tucker et al., 1997). This deposit is one of five base metal deposits that have been discovered since 1995 within Devonian-Mississippian rocks of the Yukon-Tanana terrane in the Finlayson Lake district in southeastern Yukon, Canada (Fig. 1). In recent years the Wolverine VMS deposit has been the focus of numerous studies, including a study of the stratigraphy, geotectonic setting, and genesis of the deposit (Bradshaw et al., 2001, in press); the genesis and exploration significance of chemical sedimentary (exhalative) rocks (Peter, 2003); the distribution and controls on trace metals within the deposit (Layton-Matthews, 2005); and the origin and petrogenesis of hanging-wall basaltic rocks (Piercey et al., 2002b). In addition, this area was the focus of regional studies (Murphy et al., 2006; Piercey et al., 2006) as part of the Ancient Pacific Margin National Mapping (NATMAP) project (Colpron and Nelson, 2006, and papers therein).

The football feldspar (s’quartz) porphyritic rhyolites form very distinctive rocks within the stratigraphic footwall of the Wolverine VMS deposit (Piercey et al., 2001c). These porphyry bodies occur ~10 to 20 m below massive sulfide lenses (horizons?) of the Wolverine deposit and at similar stratigraphic horizons along strike from massive sulfide mineralization (e.g., Fisher, Sable, and Puck zones). Notably the porphyritic rhyolites are clustered proximal to the Wolverine deposit and occur only locally elsewhere in the Finlayson Lake district. Their origin is therefore essential for understanding the origin and localization of mineralization at the Wolverine deposit and has implications for exploration for similar deposits elsewhere in the Finlayson Lake district, Yukon-Tanana terrane, and other continental back-arc geodynamic environments.

In this paper we present an integrated study of the footwall porphyritic rhyolites from the Wolverine VMS deposit, including field and petrographic observations, lithochemical data, and U-Pb geochronological ages for these units. The objectives are to (1) provide a descriptive field-based geologic database for the rhyolite bodies, for comparison with other rhyolites in similar geologic environments, (2) elucidate the petrogenetic history of the porphyritic rhyolites and their role in the generation of the Wolverine VMS deposit, (3) provide temporal constraints on the porphyritic rhyolites and associated mineralization in the Finlayson Lake district, and (4) understand the regional-scale controls that led to the formation of the Wolverine VMS deposit and that might be applied to exploration for other VMS districts in continental margin environments.

Regional Geologic Setting

The Finlayson Lake district forms the central part of an isolated outlier of the Yukon-Tanana and Slide Mountain terranes and lies northeast of the Tintina fault in southeast Yukon (Fig. 1). The Yukon-Tanana and Slide Mountain terranes represent a continental arc and back-arc basin sequence that developed along the ancient Pacific margin of North America in the middle to late Paleozoic (late Devonian through Permian) similar to the present-day tectonic configuration of Japan, Japan Sea, and the Sino-Korean craton (Nelson et al., 2006; Piercey et al., 2006). The district lies near the inner, eastern margin of the outlier, where pericratonic rocks of the Yukon-Tanana terrane and oceanic rocks of the Slide Mountain terrane are juxtaposed against rocks of the North American continental margin along the post-Late Triassic Inconnu thrust (Murphy et al., 2006; Figs. 2, 3). The Yukon-Tanana and Slide Mountain terranes in the Finlayson Lake massive sulfide district comprise variably deformed and metamorphosed, lower greenschist to amphibolite facies metasedimentary and metavolcanic rocks and affiliated metaplutonic suites. The rocks of the terrane are foliated and variably folded; however, regionally extensive stratigraphic units have been defined by mapping and drill core data, and primary features and geochemical characteristics are locally well preserved (e.g., Murphy et al., 2002, 2006; Piercey et al., 2004, 2006).

The Yukon-Tanana and Slide Mountain terranes in the Finlayson Lake district have been subdivided into several informal fault- and unconformity-bounded groups and formations (Figs. 2, 3; Murphy, 2001; Murphy et al., 2002, 2006). The structurally deepest units of the Yukon-Tanana terrane are...
those in the footwall of the Money Creek thrust, an Early Permian east-northeast-vergent thrust fault with more than 35 km of displacement that juxtaposes broadly coeval but lithologically and geochemically distinct lithologic successions (Figs. 2, 3; Murphy and Piercey, 2000). Rocks in the footwall of the thrust include the mafic and felsic metavolcanic and metasedimentary rocks of Upper Devonian and older Grass Lakes Group, Late Devonian to early Mississippian granitic metaplutonic rocks of the Grass Lakes plutonic suite and metasedimentary and mafic and felsic metavolcanic rocks of the unconformably overlying lower Mississippian Wolverine Lake Group (Figs. 2, 3). The Grass Lakes Group is host to the Fyre Lake, Kudz Ze Kayah, and GP4F deposits, whereas the Wolverine Lake Group hosts the Wolverine deposit. The Grass Lakes and Wolverine Lake Groups have been interpreted to represent a continental back-arc rift to back-arc basin assemblage (Piercey et al., 2001a, b, 2002a, b, 2003; Murphy et al., 2006; Nelson et al., 2006).

The hanging wall of the Money Creek thrust is comprised of the Upper Devonian to lower Mississippian metasedimentary and felsic to intermediate metavolcanic rocks and granitoid rocks (Mortensen, 1992b, and references therein), Lower Permian limestone, and, locally, Lower Permian dark-gray basinal clastic rocks (Figs. 2, 3). These latter rocks are overlain by rocks of an upper thrust sheet which comprises undeformed, predominantly mafic Late Devonian volcanic
FIG. 2. Geologic setting of the Yukon-Tanana terrane and associated rocks in the Finlayson Lake region, southeastern Yukon. Modified from Murphy et al. (2006).
FIG. 3. Stratigraphic section illustrating the relationships of different stratigraphic units in the Yukon-Tanana terrane in the Finlayson Lake region. A. All the stratigraphic units within the Finlayson Lake region. B. Detailed stratigraphy of the Grass Lakes and Wolverine Lake Groups. ALK = alkaline basalt, Arc = basalts with arc affinity, AT = A-type felsic rocks, "BAB" = back-arc basin basalt, BON = boninite, EMORB = enriched mid-ocean ridge basalt, N-MORB = normal mid-ocean ridge basalt, OIB = ocean island basalt. Bars near plutons reflect the range of U-Pb ages in plutonic suites. Modified from Murphy et al. (2006).
rocks of the Cleaver Lake Formation, spatially associated and probably comagmatic felsic, mafic, and ultramafic metaplu- 
tonic rocks, and a crosscutting early Mississippian pluton of 
the Simpson Range plutonic suite (Figs. 2, 3). None of these 
rock units host significant accumulations of VMS mineraliza-
tion. The rocks of the hanging wall of the Money Creek thrust 
(Cleaver Lake Formation and Simpson Range plutonic suite) 
have been interpreted to represent magmatism within a con-
tinental arc sequence (Grant, 1997; Piercey et al., 2001a, b, 
2003, 2006; Murphy et al., 2006).

To the north and east, the imbricated rocks of the Yukon-
Tanana terrane are juxtaposed against rocks of the Slide 
Mountain terrane along the Jules Creek fault (Figs. 2, 3). In 
this area, the Slide Mountain terrane comprises the Missis-
sippian to Lower Permian Fortin Creek Group metasedi-
mentary and metavolcanic rocks. The Slide Mountain terrane 
also includes pristine to weakly foliated Lower Permian 
basalt, mafic, and ultramafic plutonic rocks, and minor sedi-
mentary rocks of the Campbell Range Formation. The Slide Mountain terrane also includes pristine to weakly foliated Lower Permian basalt, mafic, and ultramafic plutonic rocks, and minor sedimentary rocks of the Campbell Range Formation. The Slide Mountain terrane are host to the Ice VMS deposit (Figs. 2, 3) and are interpreted to have formed in a Permian back-arc basin environment (Plint and Gordon, 1997; Piercey et al., 2006).

Porphyritic Rhyolites

The Wolverine deposit consists of two lenses of Zn-Pb-
Ag-rich massive sulfide associated with abundant Cu-rich 
stringer mineralization and typical pipelike, chlorite alter-
ation (Bradshaw et al., 2001, in press). The deposit has a 
hanging wall that consists of aphyric rhyolite, carbonaceous 
sedimentary rocks, iron formation, and basalt (Bradshaw et 
al., 2001). The footwall contains felsic volcanioclastic rocks, 
abundant carbonaceous sedimentary rocks, and rhyolite por-
phyritic sills. Within the Wolverine Lake Group, the bulk of 
the porphyritic rhyolites are located proximal to the Wolver-
ine deposit. The deeper portions of the Wolverine Lake Group consist predominantly of felsic volcanioclastic and carbonaceous sedimentary rocks. Within the Wolverine deposit area the porphyritic rhyolites are present in four zones: (1) Wolverine/Lynx zone, (2) Fisher zone, (3) Sable zone, and (4) Puck zone (Fig. 4). The porphyries are concordant to semi-
concordant (e.g., Fig. 5), suggesting a sill-like morphology, and are of two types: quartz-feldspar porphyry (QFP) and 
feldspar porphyry (FP). The two types of porphyry occur in 
most zones, but the relative abundances of the types differ 
between each zone. The Wolverine/Lynx and Fisher zones 
contain predominantly FP with subordinate QFP, the Puck 
zone contains predominantly QFP with lesser FP, and the 
Sable zone contains only QFP (Figs. 5–7; additional graphic logs are available as a digital supplement at <http://www.geo-
sienceworld.org/> or, for members and subscribers, on the 
SEG website, <http://www.segweb.org/>). At present, signific-
ant mineralization is associated with porphyries only in the 
Wolverine/Lynx zones, although the Fisher, Sable, and Puck

![Diagram of tectonic relationships and geological succession](image)

**Fig. 3.** (Cont.)
Fig. 4. Geology of the immediate Wolverine deposit area, showing the locations of different zones of the Wolverine deposit and the locations of U-Pb geochronological samples. Geology from Murphy et al. (2006).
zones also contain iron formation, carbonate-pyrite exhalites, and minor massive sulfide.

Quartz-feldspar porphyry intrusions are typically massive and coherent, range in thickness from ~1 to 2 m up to 25 m (Fig. 6). Quartz and feldspar phenocrysts occur within a homogeneous “vitreous” groundmass. Quartz crystals are subhedral, clear to blue in color, comprise 7 to 12 vol percent of the rock, and are 2 to 8 mm in size. Feldspar grains are typically subhedral to euhedral, comprise <1 to 7 vol percent of the rock and are <1 mm to 1.3 cm in size (Figs. 6, 7). The QFP intrusions are massive, homogeneous, and are devoid of amygdules and flow banding. They have chilled contacts along their margins, and there is a progressive decrease in the size and abundance of quartz and feldspar crystals from the margins to the interiors of the sills (Fig. 6). Chilled margins are not always present along both the upper and lower contacts of the intrusions due to shearing and deformation along their contacts. The QFP intrusions intrude and are interlayered with felsic volcanic siltstone (felsic tuff), variably carbonateargillite, and mixed felsic volcanic siltstone and argillite (Fig. 6). Locally along the chilled contacts there are centimeter-scale argillite fragments and xenoliths, suggesting emplacement into sediment that was not completely lithified.

All QFP intrusions are variably altered, with most feldspars partially to completely replaced by sericite with a groundmass consisting primarily of recrystallized quartz and sericite. Millimeter-scale (1–4 mm wide) veinlets of quartz, sericite, pyrite, and sphalerite in varying proportions are present in the QFP; in some samples sulfide-rich veinlets constitute 5 to 10 vol percent of the rock.

Feldspar porphyries are similar in many respects to the QFP. The FP intrusions are massive and coherent with 5 to 15 vol percent subhedral to euhedral feldspar crystals that range in size from 2 mm to 1.3 cm within an aphanitic (vitreous) groundmass (Fig. 7). The intrusions are homogeneous and devoid of amygdules and flow banding (Figs. 6, 7). Chilled contacts with wall rocks are typically aphyric and are characterized by an increase in feldspar phenocryst size and percentage toward the interior of the intrusions (Fig. 6). Contacts are parallel to foliation or bedding in the adjacent host rocks consistent with a sill-like morphology (Figs. 6). The FP intrusions are variably altered, most commonly with patchy replacement of primary K-feldspar by gray to black secondary K-feldspar. Sericite commonly replaces both matrix and feldspar phenocrysts. Pyrite-quartz and pyrite-quartz-sericite veinlets crosscut many of the porphyries; in a
FIG. 6. Graphic logs of type sections of quartz-feldspar porphyries (A) and feldspar-porphyries (B) from the Wolverine VMS deposit. Additional graphic logs of feldspar and quartz-feldspar porphyries are available as a digital supplement to this paper at <http://www.geoscienceworld.org/>, or, for members and subscribers, on the SEG website, <http://www.segweb.org>. 
few localities chlorite-pyrite-quartz veinlets are also present. In some feldspar porphyries, Fe carbonate patches and veinlets are associated with the pyrite-quartz and pyrite-quartz-sericite of alteration.

Geochemistry

The geochemistry of the porphyritic rhyolites is summarized in Tables 1 and 2. The entire dataset is available as a digital supplement to this paper at <http://www.geoscienceworld.org/> (or, for members and subscribers, on the SEG website, <http://www.segweb.org>). Details of the sampling protocol, analytical methods, and precision and accuracy are presented in the Appendix. To characterize the primary geochemistry of the porphyritic rhyolites we have relied primarily on the immobile major elements Al₂O₃ and TiO₂, the high field strength elements (HFSE) Zr, Hf, Nb, Ta, Y, and Th, and rare earth elements (REE) La to Lu. We have not used most of the major elements or low field strength elements (LFSE), which substitute for Na, K, and Ca, due to their mobility during hydrothermal alteration and metamorphism. This is supported by their scatter on an alteration box plot in which the porphyries lie on arrays extending from the least altered rhyolite to the chlorite-pyrite and sericite-altered end members (Fig. 8). Furthermore, the large scatter of most major elements in plots against the immobile element Zr (Fig. 9) is consistent with their mobility during hydrothermal alteration.
Table 1. Mean and 2$\sigma$ Error Values for Geochemical Data for Porphyritic Rhyolites from the Wolverine VMS Deposit

| Sample Name | Average Wolverine/Lynx FP (n = 24) | 2$\sigma$ | Average P00-WV-2 Wolverine/Lynx QFP (n = 9) | 2$\sigma$ | Average Fisher FP (n = 2) | 2$\sigma$ | Average Fisher QFP (n = 2) | 2$\sigma$
|-------------|-----------------------------------|----------|-----------------------------------------|----------|-------------------------|----------|-------------------------|----------
| SiO$_2$ (wt%) | 65.35 | 2.45 | 71.40 | 66.61 | 4.86 | 61.50 | 21.20 |
| TiO$_2$ | 0.45 | 0.04 | 0.30 | 0.44 | 0.10 | 0.31 | 0.03 |
| Al$_2$O$_3$ | 15.53 | 1.15 | 14.00 | 16.00 | 3.71 | 1.27 | 12.45 | 14.91 |
| Fe$_2$O$_3$ | 5.53 | 1.29 | 8.87 | 5.67 | 3.71 | 1.27 | 12.45 | 14.91 |
| FeO* | 0.02 | <0.01 | 0.04 | 0.02 | 0.01 | <0.01 |
| MnO | 0.40 | 0.08 | 0.19 | 0.61 | 0.43 | 0.09 | 0.03 |
| MgO | 0.87 | 0.25 | 2.45 | 1.00 | 0.65 | 0.59 | 0.64 |
| Na$_2$O | 1.57 | 0.26 | 0.14 | 1.16 | 0.53 | 0.60 | 0.98 |
| K$_2$O | 6.62 | 0.64 | 5.05 | 5.70 | 1.16 | 1.16 | 1.16 |
| P$_2$O$_5$ | 0.14 | 0.01 | 0.15 | 0.14 | 0.03 | 0.09 | 0.04 |
| H$_2$O | 1.50 | - | - | 1.70 | 0.57 | NA | NA |
| CO$_2$ | 0.65 | 0.22 | 1.90 | 0.77 | 0.56 | 0.90 | 0.90 |
| Sr | 3.37 | 1.06 | 1.26 | 2.30 | 1.07 | 9.81 | 12.59 |
| LOI | 3.81 | 0.72 | 3.60 | 3.84 | 1.05 | 7.70 | 7.00 |
| Total | 100.31 | 0.47 | 99.00 | 99.28 | 0.73 | 102.85 | 5.30 |
| Cr (ppm) | 12 | 1 | <10 | 16 | 3 | <10 | <10 |
| Ni | 8 | 1 | 5 | 8 | 2 | <1 | - |
| Co | 9.8 | 1.3 | 4.0 | 9.7 | 3.2 | 4.0 | 0.7 |
| V | 26 | 5 | 17 | 34 | 31 | 25 | 10 |
| Cu | 12 | 1 | <10 | 14 | 2 | <1 | - |
| Pb | 31 | 7 | 18 | 40 | 23 | 60 | 28 |
| Zn | 29 | 4 | 228 | 29 | 8 | 24 | 5 |
| Bi | 0.81 | 0.57 | <0.2 | 0.38 | 0.12 | 0.30 | 0.00 |
| In | 0.10 | 0.01 | <0.05 | 0.09 | 0.02 | 0.06 | - |
| Sn | 5.31 | 0.64 | 13.00 | 6.39 | 3.39 | 5.55 | 0.70 |
| Mo | 1.90 | 0.30 | 0.80 | 2.34 | 0.94 | 3.25 | 2.90 |
| Sb | 4.10 | 1.28 | 3.00 | 2.78 | 1.54 | 7.05 | 5.90 |
| Ag | 0.52 | 0.30 | 0.20 | 0.08 | 0.45 | 0.30 |
| Rb | 144 | 15 | 110 | 141 | 28 | 69 | 16 |
| Cs | 2.33 | 0.24 | 2.00 | 4.29 | 2.36 | 1.50 | 0.20 |
| Ba | 4609 | 124 | 4300 | 2430 | 1624 | 1830 | 420 |
| Sr | 96 | 15 | 229 | 108 | 66 | 58 | 24 |
| Ti | 2.9 | 0.5 | 1.5 | 1.9 | 0.6 | 2.5 | 0.6 |
| Ga | 23.8 | 2.8 | 23.0 | 25.2 | 6.7 | 14.5 | 1.0 |
| Ta | 2.4 | 0.2 | 1.7 | 2.3 | 0.5 | 1.8 | 0.6 |
| Nb | 39.3 | 3.4 | 21.0 | 36.2 | 6.7 | 23.0 | 4.0 |
| Hf | 13.1 | 1.1 | 5.8 | 12.5 | 2.6 | 7.5 | 0.6 |
| Zr | 373.6 | 50.3 | 235.0 | 517.6 | 96.6 | 314.0 | 16.0 |
| Y | 58.3 | 4.1 | 41.0 | 54.7 | 18.1 | 34.5 | 1.0 |
| Tb | 35.7 | 4.0 | 18.0 | 37.2 | 5.6 | 21.0 | 8.0 |
| U | 5.4 | 0.4 | 5.4 | 6.4 | 3.7 | 5.2 | 0.8 |
| La | 98.08 | 8.86 | 48.00 | 101.44 | 20.38 | 43.50 | 11.00 |
| Ce | 214.04 | 20.48 | 100.00 | 212.78 | 44.07 | 92.00 | 16.00 |
| Pr | 24.63 | 2.37 | 12.00 | 24.36 | 5.06 | 10.95 | 2.10 |
| Nd | 69.56 | 6.87 | 42.00 | 66.22 | 17.14 | 37.30 | 3.00 |
| Sm | 16.69 | 1.58 | 8.60 | 15.94 | 3.23 | 7.70 | 0.20 |
| Eu | 1.78 | 0.20 | 2.80 | 1.68 | 0.46 | 0.93 | 0.10 |
| Gd | 13.49 | 1.22 | 7.20 | 12.91 | 2.65 | 6.85 | 0.90 |
| Tb | 1.95 | 0.16 | 1.20 | 1.89 | 0.38 | 1.05 | 0.10 |
| Dy | 10.48 | 0.83 | 7.10 | 10.31 | 2.10 | 6.10 | 0.40 |
| Ho | 2.06 | 0.16 | 1.40 | 1.99 | 0.39 | 1.20 | 0.00 |
| Er | 5.31 | 0.39 | 3.60 | 5.07 | 1.68 | 3.15 | 0.30 |
| Tm | 0.79 | 0.06 | 0.53 | 0.74 | 0.18 | 0.46 | 0.08 |
| Yb | 5.10 | 0.36 | 3.20 | 4.77 | 1.19 | 3.05 | 0.70 |
| Lu | 0.78 | 0.05 | 0.46 | 0.73 | 0.18 | 0.45 | 0.11 |
| T-ZrSat$^{1}$ | 930 | 15 | 852 | 941 | 32 | 851 | 68 |

Notes: $^{1}$Zircon saturation temperature (Watson and Harrison, 1983)
There is no difference in the compositions of the FP or QFP in the different zones of the Wolverine deposit. Therefore, we consider samples from different zones to belong to a single suite of FP or QFP intrusions.

The FP suite has Nb/Y ratios that straddle the subalkaline and/or alkaline boundary with Nb/Y systematics, indicative of a within-plate affinity (Fig. 11, Table 2). They have transitional to calc-alkalic Zr/Y and Zr/Nb ratios that border on peralkaline (Fig. 10D, F, Table 2). On primitive mantle-normalized plots the FP suite has downward-sloping profiles with LREE enrichment and distinctive negative Nb, Eu, and Ti anomalies and depletions in the compatible elements V and Sc (Fig. 12). They have flat upper continental crust-normalized patterns with variable Eu anomalies (Fig. 12). The QFP suite has flat upper continental crust-normalized patterns with weak negative Eu anomalies (Fig. 12) and La/Sm\textsubscript{UCN} ~1 (Fig. 13), similar to rocks derived from, or that have interacted with, upper continental crustal material. This is also supported by the Nb/Th\textsubscript{UCN} values close to 1 (Fig. 13; McLennan, 2001). The LREE enrichment in this suite overlaps the fields for Archean FI to FI rhyolites (Fig. 14A) and the fields for the Phanerozoic rhyolites from the Mount Windsor, Kuroko, and Que River districts (Fig. 14B).

The QFP suite has Nb/Y ratios that straddle the alkaline boundary and have within-plate affinities (Fig. 11). Unlike the FP suite, the QFP suite has Zr/Y ratios that are transitional between tholeiitic and calc-alkalic (Fig. 10F). The primitive mantle-normalized patterns are similar to the FP suite with LREE enrichment and negative Nb, Eu, and Ti anomalies, but the overall abundance of trace elements is lower (Fig. 12). The QFP suite has flat upper continental crust-normalized patterns with variable Eu anomalies (Fig. 12), likely reflecting variable mobility due to alteration (e.g., Wood and Williams-Jones, 1994). La/Sm\textsubscript{UCN} and Nb/Th\textsubscript{UCN} ratios are close to 1, as in the FP suite, but they are displaced toward lower absolute values of Nb, Th, La, and Sm (Fig. 13). The REE concentrations of the QFP suite are similar to F1 rhyolites from Archean volcanic sequences (Fig. 14A) and overlap the fields for Phanerozoic rhyolites from the Que River district (Fig. 14B).

U-Pb Geochronology

Four samples for geochronology were taken from both surface and drill core from the Wolverine deposit. The sampling protocol and analytical methods for U-Pb geochronology are outlined in the Appendix, and the results are outlined below and in Figure 15 and Table 3.

Wolverine/Lynx zone

This drill core sample (WW00-01, 439836E, 6811491N, 445-m depth) is a coherent FP rhyolite intrusion that yielded abundant clear, pale yellow, euhedral prismatic zircon grains. Analyses of six fractions of abraded zircons are shown in Figure 15A. Two fractions (A and G) yielded overlapping concordant analyses with a total range of 206Pb/238U ages of 347.8 ± 1.3 Ma, which is interpreted to be the crystallization age (Fig. 15, Table 3). Fractions D and F yielded slightly older 207Pb/206Pb ages, and fraction E yielded a much older 207Pb/206Pb age (802 Ma: Fig. 15, Table 3), indicating the
Fig. 9. Variation diagrams of major elements vs. the immobile, incompatible element Zr, including (A) SiO$_2$, (B) Fe$_2$O$_3$tot, (C) MgO, (D) Na$_2$O, (E) K$_2$O, (F) CaO, (G) CO$_2$tot, and (H) S$_{tot}$. The scatter of these elements vs. the immobile element Zr suggests variable mobility of these elements due to varying amounts of sericite, K-feldspar, carbonate, and pyrite alteration.
PETROLOGY AND U-Pb GEOCHRONOLOGY OF FOOTWALL RHYOLITES, WOLVERINE VMS DEPOSIT, YUKON

FIG. 10. Variation diagram of immobile elements vs. immobile and incompatible element Zr, including (A) Ti, (B) Al, (C) Sc, (D) Nb (diagram after Leat et al., 1986), (E) Ga, (F) Y (affinities based on work of Barrett and MacLean, 1999), (G) Th, (H) La, (I) Sm, and (J) Yb. The strong correlation with Zr concentrations indicates that these elements were immobile during hydrothermal alteration.
presence of a minor to major component of inherited zircon in these fractions. Fraction C gave much younger \( {^{206}\text{Pb}}/{^{238}\text{U}} \) ages than the two concordant fractions; however, the \( {^{207}\text{Pb}}/{^{206}\text{Pb}} \) age for this fraction overlaps those of the concordant fractions (Table 3). This fraction is interpreted to have been free of inheritance but likely contained zircon that had experienced Pb loss that was not completely removed by the abrasion.

**Sable zone**

This surface sample (Sable, 440338E, 6810613N) is of a QFP intrusion that yielded zircons similar in appearance to those from WW00-01. Four fractions of grains with clear tube-shaped inclusions passing through their centers were analyzed after strong abrasion. From previous experience in Yukon-Tanana terrane rocks zircons with tube-shaped inclusions in their cores typically do not contain inherited zircon and yield magmatic crystallization ages rather than inheritance ages. Fractions A and D yielded overlapping concordant data (Fig. 15B) with a total range of \( {^{206}\text{Pb}}/{^{238}\text{U}} \) ages of 352.4 ± 1.5 Ma (Table 3) interpreted to be the crystallization age of the sample. Fraction C yielded an older and slightly discordant \( {^{207}\text{Pb}}/{^{206}\text{Pb}} \) age, and fraction B yielded a much older \( {^{207}\text{Pb}}/{^{206}\text{Pb}} \) age (Table 3). Both of these fractions are interpreted to contain inherited zircon. Two fractions (E and F) of fine-grained, very elongate prisms were also abraded and analyzed. They gave the same \( {^{207}\text{Pb}}/{^{206}\text{Pb}} \) ages as the two concordant fractions but considerably younger \( {^{206}\text{Pb}}/{^{238}\text{U}} \) ages (Table 3), indicating that they have both experienced postcrystallization Pb loss.

**Fisher zone**

This surface sample (P98-69A, 433931E, 6816809N) is a coherent FP rhyolite that yielded two main populations of zircons: elongate square prismatic grains, and stubby prisms to equant grains. Six fractions of strongly abraded elongate prismatic grains were analyzed together with three fractions of stubby prismatic grains. The elongate grains generally had little or no inherited component, whereas the stubby to equant grains yielded slightly to much older \( {^{207}\text{Pb}}/{^{206}\text{Pb}} \) ages, reflecting the presence of a substantial inherited zircon (Fig. 15C). Two fractions of the elongate grains (A and I) yielded overlapping, concordant \( {^{206}\text{Pb}}/{^{238}\text{U}} \) ages of 346.0 ± 2.2 Ma (Table 3). This is considered to be the crystallization age of the sample. All other fractions yielded somewhat older \( {^{207}\text{Pb}}/{^{206}\text{Pb}} \) ages and a considerable range of \( {^{206}\text{Pb}}/{^{238}\text{U}} \) ages (Table 3), and these are thought to reflect the effects of minor amounts of inherited zircon in a small proportion of the grains analyzed, coupled with minor postcrystallization Pb loss.

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**Fig. 10.** (Cont.)

**Fig. 11.** A. Modified Zr/TiO₂-Nb/Y diagram of Winchester and Floyd (1977) (from Pearce, 1996). B. Nb-Y discrimination diagram of Pearce et al. (1984). These diagrams illustrate that the Wolverine porphyries straddle the subalkaline-alkaline boundary and have within-plate (A-type) affinities.
Puck zone

This sample of QFP from drill core (PUCK, 44384E, 6805995N, drill hole PK96-04, 225-m depth) yielded four concordant fractions (A, B, D, and E) but with a considerable range in 206Pb/238U ages (Table 3). We tentatively assign a crystallization age of 356.9 ± 0.5 Ma to this sample, based on the 206Pb/238U age for fraction D, which is the oldest concordant analysis. Fractions F and C give much older 207Pb/206Pb ages, reflecting the presence of a substantial inherited component. The interpreted age of this sample is considerably older than that of the three other samples and is based on only one concordant data point. As such, we cannot preclude the possibility that this fraction also contained a minor component of inherited zircon from slightly older rock units through which it was intruded. Based on the similarities geologically and geochemically with the Sable zone QFP intrusion, which has a more precise age of 352.4 ± 1.5 Ma, the assigned crystallization age for this intrusion is possibly overestimated.

Geologic, Temporal, and Petrotectonic Evolution of Porphyries in the Wolverine VMS Deposit

Based on field relationships, the QFP and FP intrusions represent two distinct magmatic events within the Wolverine basin, and U-Pb geochronology supports this interpretation. The QFP intrusions are interpreted to be the earliest phase of rhyolitic magmatism, as they have weakly developed chilled margins and xenoliths along their margins, indicative of intrusion into partially, but not completely, lithified sedimentary material (Fig. 7). In contrast, the feldspar porphyries show well-developed chilled margins (Fig. 7), suggesting that they were emplaced after the QFP intrusion into sediments that were completely lithified. The FP suite from the Wolverine/Lynx and Fisher zones have U-Pb ages that are ~5 m.y. younger than the QFP intrusions (Fig. 15, Table 3) and indicate that porphyry magmatism was episodic during the evolution of the Wolverine footwall sequence.

The episodic magmatism can be explained within the context of an evolving continental back-arc rift. In this model the porphyries formed in response to crustal melting and crust-mantle mixing caused by basaltic underplating during back-arc extension (Piercey et al., 2001b, 2003, 2006). In both porphyry suites, crustal melting and/or crustal contamination was quite important as indicated by the flat upper continental crust-normalized REE patterns (Fig. 12) and La/Sm$_{UCN}$ and Nb/Th$_{UCN}$ ratios near unity (Fig. 13, Table 2; McLennan, 2001). Furthermore, the inherited zircon (Fig. 16, Table 3) and existing Nd isotope data (Piercey et al., 2003) all point to
Fig. 13. Trace element plots for the Wolverine porphyries. Notably the FP suite has lower Ti/Sc (A) and higher Nb/Ta (B) ratios than the QFP suite, consistent with a greater mantle component in their genesis. Both suites, however, have Nb/Th$_{UCN}$ (C) and La/Sm$_{UCN}$ (D) values ~1, suggesting derivation from or extensive interaction with upper continental crust. UCN = upper crust normalized (values from McLennan, 2001). Values for enriched mantle and depleted mantle are those of ocean island basalt (OIB) and normal mid-ocean ridge basalt (N-MORB) from Sun and McDonough (1989).

Fig. 14. La/Yb-Yb plots of prospective vs. less prospective rhyolites, after Hart et al. (2004) diagram (A), modified from Lesher et al. (1986), and a modified diagram for Phanerozoic rocks (B) from Lentz (1998).
the role for significant ancient crust in the genesis of these porphyries.

There are subtle but important differences between the FP and the QFP. The FP suite has higher absolute abundances of HFSE and REE (Figs. 10–12, 16; Table 1). Furthermore, the FP has higher Zr/Ti, Zr/Y, Zr/Ga, Ti/Sc, and Nb/Ta ratios and lower Al/Zr ratios relative to the QFP (Figs. 10, 13; Table 2), suggesting variations in their petrogenic histories. Although subtle, these variations are significant, and variation in HFSE and REE abundances can be explained largely by variations in the temperature of melting between the two suites; the higher the temperature of melting of a source region the more HFSE and REE enriched the felsic magma will be (Clemens et al., 1986; Whalen et al., 1987; Creaser and White, 1991; Bea, 1996a, b; Watson, 1996; King et al., 1997; Hanchar and Watson, 2003).

Zircon saturation temperatures can be affected by alkali element mobility, a distinct possibility at Wolverine; however, in Figure 17 we have provided two plots of zircon saturation temperatures against Nb (a proxy for HFSE enrichment); one with all the data (except the chlorite-carbonate-pyrite–altered samples) and one screened for the least altered samples. From this diagram it is clear that both the altered and least altered samples comprise two populations: the FP with higher zircon saturation temperatures and higher Nb contents and the QFP with lower zircon saturation temperatures and Nb contents (Fig. 18). This diagram illustrates that there is a correlation between HFSE and zircon saturation temperatures implying that the higher temperature porphyries have higher Nb contents (and by association the other HFSE), consistent with existing experimental and empirical work (Clemens et al., 1986; Whalen et al., 1987; Creaser and White, 1991; Bea, 1996a, b; Watson, 1996; King et al., 1997; Hanchar and Watson, 2003).

\[
\ln \left( \frac{D_{Zr, \text{zircon}}}{D_{Zr, \text{melt}}} \right) = \left[ -3.8 - \left[ 0.85(M - 1) \right] \right] + 12.900/T, \tag{1}
\]

where \(D_{Zr, \text{zircon}}/D_{Zr, \text{melt}}\) is the ratio of Zr concentration (ppm) in zircon to that in a saturated melt (i.e., the mineral-melt partition coefficient for Zr from Watson and Harrison, 1983), M is the alumina saturation factor for the rock \([M = (Na + K + 2 \cdot Ca)/\text{Al} \cdot Si]\), and T is temperature in Kelvin. In rocks that have inherited zircon, like those at the Wolverine, zircon saturation temperatures provide maximum estimates of the temperature of the felsic rock at the source (Miller et al., 2003) and by association provide an estimate of the temperature of formation of the felsic rocks (e.g., Barrie, 1995). Zircon saturation temperatures can be affected by alkali element mobility, a distinct possibility at Wolverine; however, in Figure 17 we have provided two plots of zircon saturation temperatures against Nb (a proxy for HFSE enrichment); one with all the data (except the chlorite-carbonate-pyrite–altered samples) and one screened for the least altered samples. From this diagram it is clear that both the altered and least altered samples comprise two populations: the FP with higher zircon saturation temperatures and higher Nb contents and the QFP with lower zircon saturation temperatures and Nb contents (Fig. 18). This diagram illustrates that there is a correlation between HFSE and zircon saturation temperatures implying that the higher temperature porphyries have higher Nb contents (and by association the other HFSE), consistent with existing experimental and empirical work (Clemens et al., 1986; Whalen et al., 1987; Creaser and White, 1991; Bea, 1996a, b; Watson, 1996; King et al., 1997; Hanchar and Watson, 2003).
The high Zr/Ti, Zr/Y, Zr/Ga, and lower Al/Zr ratios also are consistent with higher temperatures of melting, as the Zr concentrations would increase to a greater extent than Ti, Y, Ga, and Al (e.g., Bea, 1996a, b). However, the temperature of melting cannot explain the differences in Nb/Ta and Ti/Sc ratios between the FP and QFP. The difference in Nb/Ta and Ti/Sc can be attributed to an increase in the mantle component in the QFP suite relative to the FP suite.

The variations in temperature and mantle contributions to the FP and QFP magmas can be explained by the evolving back-arc basin model. The early magmatism during initial back-arc rifting at ~352 Ma was caused by crustal melting due to basaltic underplating (e.g., Huppert and Sparks, 1988), resulting in the formation of the QFP suite (Fig. 18A). This stage was broadly contemporaneous with the formation of intrabasinal tuffs and the bulk of the footwall succession of the Wolverine VMS deposit; notably, the tuffaceous rocks from the footwall of the deposit have trace element signatures similar to the QFP suite (Piercey et al., 2001b).

Further support for extension and upwelling of basalt comes from the stratigraphy of the Wolverine deposit. The entire Wolverine Lake Group is capped by mid-ocean ridge basalt, and the volcanic rocks of the lower Wolverine Lake Group are interpreted to represent back-arc basin volcanic activity. The variations in temperature and mantle contributions to the FP and QFP magmas can be explained by the evolving back-arc basin model. The early magmatism during initial back-arc rifting at ~352 Ma was caused by crustal melting due to basaltic underplating (e.g., Huppert and Sparks, 1988), resulting in the formation of the QFP suite (Fig. 18A). This stage was broadly contemporaneous with the formation of intrabasinal tuffs and the bulk of the footwall succession of the Wolverine VMS deposit; notably, the tuffaceous rocks from the footwall of the deposit have trace element signatures similar to the QFP suite (Piercey et al., 2001b).

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Fig. 16. Frequency histograms for the HFSE, Zr, Nb, Th, and the REE La for the FP and QFP suites. Note the higher HFSE and REE in the FP suite relative to the QFP suite.
FIG. 17. Zircon saturation temperatures (T-ZrSat; Watson and Harrison, 1983) in relationship to Nb contents (Nb acting as a proxy for HFSE enrichment). In (A) the entire suite of FP and QFP are presented and unscreened for alteration (except for the chlorite-carbonate-pyrite altered porphyries which are not included). In (B) the least altered samples are presented. In both plots there is a separation into two populations: the higher temperature FP suite with higher concentrations of HFSE and the lower temperature QFP suite with lower concentrations of HFSE. This diagram strongly suggests that the elevated HFSE in the Wolverine porphyries are directly related to their temperature of formation.

FIG. 18. Block model outlining the petrogenetic and metallogenic evolution of the Wolverine back-arc basin in relationship to the two main phases of porphyry intrusion. A. The initial QFP phase (~352 Ma) involves predominantly crustal melting due to basaltic underplating and subsequent emplacement of the QFP sills within a back-arc rift. B. With continued extension within the rift basaltic magmas rise within the crust resulting in increased heat flow, increased crust-mantle interaction (thus explaining the greater mantle component in the FP suite) and genesis of the FP suite of intrusions. The increase in heat flow from the mantle was critical in elevating the geothermal gradient within the Wolverine back-arc basin and in generating the Wolverine VMS deposit.
basalt (MORB; see Piercey et al., 2002b; Figs. 4, 5), suggesting that the initial phase of rifting (Fig. 16A) and initial mantle upwelling (Fig. 16B), eventually gave rise to basaltic eruptions and sea-floor spreading.

Although mineralization occurred after the FP suite was emplaced (mineralization cross cuts the FP suite), mantle heat was likely key in forming both the QFP and FP intrusions (Fig. 15). Sustained mantle heat flow within the Wolverine back-arc rift from ~352 to ~347 to 346 Ma, therefore, likely contributed to the formation of the Wolverine hydrothermal system and VMS deposit (Fig. 15).

Implications of ages for the tectonic setting and timing of Wolverine VMS deposit genesis

The new U-Pb ages for the Wolverine porphyries complement existing data for the Finlayson Lake district (Mortensen, 1992a, b; Murphy et al., 2006) and provide a critical chronostratigraphic framework for the large-scale tectonic-magmatic and metallogenic history for a large part of the northern Cordillera.

The age of the lowermost part of the Wolverine Lake Group is constrained by the youngest detrital zircons in the lowermost conglomerate unit (~357 Ma) and a U-Pb zircon age for rhyolite just above this conglomerate (356 ± 1 Ma; Murphy et al., 2006). In contrast, the porphyries that comprise the immediate footwall to the Wolverine deposit range from ~352 to 347 Ma (Fig. 15). Both porphyry suites are cut by alteration and mineralization, which suggests that the Wolverine deposit is no older than 347.9 to 346.0 Ma, the age of the FP suite of intrusions (Fig. 15). The close spatial association of the porphyries to mineralization in the Wolverine/Lynx zone suggests that the age of the porphyries is also the likely age of deposit formation.

The ages obtained for the Wolverine deposit are different than other VMS deposits in the Finlayson Lake district (Fig. 19). For example, the mafic- and sediment-hosted Fyre Lake

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**Fig. 19.** Age distribution for VMS deposits in the Yukon-Tanana terrane, the Wolverine VMS deposit porphyritic rhyolites, VMS and sedimentary exhalative (SEDEX), and Mississippi Valley-type (MVT) deposits from the Kootenay terrane and the North American miogeocline, and globally significant (>50 Mt) VMS and SEDEX deposits/camps that formed during the Devonian-Mississippian. Age data for the VMS deposits in Yukon-Tanana terrane, Kootenay terrane, and North American miogeocline are from Nelson et al. (2002). Age data for Red Dog are from Morelli et al. (2004). Data for Rammelsberg and Meggen deposits are from Large and Walcher (1999) and references therein. Age data for the Iberian pyrite belt are from Nesbitt et al. (1999) and Barrie et al. (2002). Figure modified from Nelson et al. (2002). Timescale from Harlan et al. (1990).
deposit occurs within the Fyre Lake Formation of the underlying Grass Lakes Group (Figs. 2, 3) and has an age of ~365 Ma (Mortensen, 1992a, b; Murphy et al., 2006). The felsic-hosted Kudz Ze Kayah and GP4F deposits are hosted in the Kudz Ze Kayah Formation of the underlying Grass Lakes Group (Figs. 2, 3, 19) and have ages that are ~360 Ma (Mortensen, 1992b; Murphy et al., 2006). In contrast, the Ice deposit is hosted in the overlying Campbell Range Formation (Figs. 2, 3) in rocks that are ~275 Ma (Mortensen, 1992a, b; Murphy et al., 2006). The felsic volcanic petrology, radiogenic isotope systematics, and zircon inheritance patterns of the rhyolites at the Wolverine deposit are similar to those at Kudz Ze Kayah and GP4F (Mortensen, 1992b; Piercey et al., 2001b, 2003), yet there is at least 13- to 14-m.y. difference between the ages of the deposits (Fig. 19). These results clearly illustrate that VMS mineralization within the Finlayson Lake district was episodic.

The episodic nature of VMS mineralization in the Finlayson Lake district is similar to that found in the Bathurst mining camp in New Brunswick and the Iberian pyrite belt. In the Bathurst camp, the Brunswick deposits formed at ~471 to 469 Ma, whereas the Caribou deposits are younger than ~466 to 465 Ma (Sullivan and van Staal, 1990; van Staal et al., 1992). In the Iberian pyrite belt, recent U-Pb zircon dating has shown that the deposits formed at 356, 354, 352, and 346 Ma, spanning at least 10 m.y. (Nesbitt et al., 1999; Barrie et al., 2002). In contrast, deposit formation in some VMS districts occurs rapidly, often during a single event. For example, in the Noranda VMS camp most magmatism within the Blake River Group occurs between 2703 and 2697 Ma, and more likely all sulfide deposition occurred between 2701 and 2697 Ma (Mortensen, 1993; Galley and van Breenen, 2002). Similarly, in the Sturgeon Lake camp, mineralization formed over a very restricted interval between 2735 and 2734 Ma (Davis and Trowell, 1982; Davis et al., 1985; Galley et al., 2000).

The difference in ages between the older phase of Fyre Lake-, Kudz Ze Kayah- and GP4F-related hydrothermal activity and the younger Wolverine phase of hydrothermal activity (Fig. 19) has been attributed to tectonic disruptions during the evolution of an ensialic back-arc basin environment (e.g., Murphy and Piercey, 1999; Piercey et al., 2001b). Arc magmatism dominates the earliest phases of magmatism in the Finlayson Lake district at ~365 to 360 Ma (Mortensen, 1992b; Piercey et al., 2001a, b, 2003, 2004, 2006; Murphy et al., 2006) and was succeeded by ensialic back-arc basin magmatism and VMS deposit formation in the Kudz Ze Kayah Formation. Following mineralization, the entire basin was uplifted, deformed, and an unconformity was developed between the Grass Lakes and the Wolverine Lake Groups at ~357 Ma (Murphy, 1998; Murphy and Piercey, 1999; Murphy et al., 2006). The rocks of the lower parts of the Wolverine Lake Group and the footwall to the Wolverine deposit formed after the development of this unconformity. These rocks are petrologically identical to the overlying Grass Lakes Group but are less deformed (Murphy and Piercey, 1999; Piercey et al. 2001b, 2003). This implies that there was a continuation of ensialic back-arc rifting, which eventually resulted in seafloor spreading and the generation of the MORB-type basalt stratigraphically above the Wolverine deposit (Piercey et al., 2002b).

Regional implications

The ensialic back-arc basin activity within the Finlayson Lake region was broadly coincident with arc and continent rifting along the western North America margin in the Devonian-Mississippian, which led to the development of the Slide Mountain back-arc basin (Nelson, 1993; Nelson et al., 2002, 2006; Piercey et al., 2003, 2004, 2006), and continental rifting and magmatism within the North America cratonic margin (Fig. 19; Gوردley et al., 1987; Goodfellow et al., 1995; Paradis et al., 1998). This arc and continent rifting started at ~365 to 360 Ma (e.g., Nelson, 1993; Dusel-Bacon et al., 2004; Piercey et al., 2004) and continued after the Wolverine deposit formed as the Slide Mountain back-arc basin widened (e.g., Nelson, 1993; Nelson et al., 2006; Piercey et al., 2006). This extensional geodynamic regime along the margin of North America was likely a critical factor in regional metallogeny (e.g., Paradis et al., 1998; Nelson et al., 2002).

The formation of the Wolverine deposit is coincident with a significant number of exhalative, syngenetic sulfide deposits in the northern Cordillera that span the Devonian-Mississippian boundary, including both VMS and sedimentary exhalative (SEDEX) mineralization in both the Yukon-Tanana terrane and along the North American margin (Fig. 19; Paradis et al., 1998, 2006; Bailey et al., 2001; Nelson et al., 2002; Dusel-Bacon et al., 2004). In the southern Canadian Cordillera, similar styles of syngenetic sulfide mineralization formed in this time interval (Paradis et al., 2006). It was also a period when numerous VMS- and SEDEX-type deposits formed globally, including many giant and super giant deposits (Fig. 19; Large and Walcher, 1999; Nesbitt et al., 1999; Barrie et al., 2002; Morelli et al., 2004). It has been suggested that this global pulse of massive sulfide mineralization may have been related to global anoxia (Goodfellow, 1987). It is possible that black shales deposited during this anoxia, and which are prevalent in the Wolverine deposit, may have served as excellent traps for upwelling hydrothermal fluids (i.e., highly efficient cap rocks that could have prevented or significantly reduced the dissipation of metalliferous hydrothermal fluids into the overlying water column and promoted subsea-floor replacement-style deposition).

Significance of HFSE- and REE-Enriched Rhyolites to VMS Deposit Genesis

The Wolverine deposit porphyritic rhyolites are some of the most HFSE- and REE-enriched felsic rocks in any VMS camp in the world (see Lentz, 1998 and references therein). This geochemical relationship is not necessarily unique and has been recognized in VMS-related rhyolites from the Archean to the present (Lesher et al., 1986; Lentz, 1998; Piercey et al., 2001b; Dusel-Bacon et al., 2004). This common occurrence of HFSE- and REE-enriched rhyolites with VMS deposits suggests a link between the petrogenetic (and tectonic?) history of these rhyolites and the generation of massive sulfide mineralization.

It is not surprising that many ancient VMS systems, especially those associated with continental margin settings like the Wolverine deposit, are associated with HFSE- and REE-enriched felsic magmas (Mortensen and Godwin, 1982; Lentz, 1998; Piercey et al., 2001b; Dusel-Bacon et al., 2004).
These rocks formed in environments with elevated geothermal gradients and would have had the elevated heat flow necessary to drive long-lived and robust hydrothermal systems key to forming VMS mineralization (Cathles, 1981; Barrie et al., 1999).

Although the Wolverine HFSE- and REE-enriched porphyries are clearly associated with VMS mineralization, they have F1 to FII designations (Fig. 14), the least prospective types of VMS-associated rhyolites in Archean successions (Lesher et al., 1986; Hart et al., 2004). When initially conceived the F1 to FIII classification was intended for use in Archean rhyolite successions (Lesher et al., 1986). In recent years, however, some workers have suggested that it is also suitable for younger felsic sequences (Hart et al., 2004). The results provided herein imply that this suggestion is not completely valid. The designation of these rhyolites within the less prospective F1 to FII fields, however, has more to do with their elevated La (and Zr) than depletion in Yb (and Y). In the models of Lesher et al. (1986) and Hart et al. (2004) the F1 and FII designations, and depletions in HREE and Y, are properly interpreted to reflect the presence of garnet (F1) and amphibole (FII) in the melt source regions, which deplete Y and Yb, leading to higher Zr/Y and La/Yb ratios, reflecting deeper sources of melt generation. This assumption is valid in Archean crustal environments where the substrate is largely basaltic in nature (e.g., Hart et al., 2004), however, it is not completely valid for rocks derived from melting of evolved continental crust. In particular, the trace element budget of continental crustal melts is largely controlled by the efficiency of melting of HFSE- and REE-enriched accessory phases (e.g., Watson and Harrison, 1983; Bea, 1996a, b; Watson, 1996). Thus, for rhyolites formed from melting of continental crust, La/Yb and Zr/Y ratios are mainly controlled by the dissolution kinetics of HFSE- and REE-enriched phases during crustal melting and to a lesser extent the depth at which the melts were generated. Thus, in continental margin arc and back-arc rift environments it is more important to identify HFSE- and REE-enriched felsic melts, rather than F1 to FII signatures, when searching for new prospective VMS environments.

Summary and Conclusions

Porphyritic rhyolites form an important component of the footwall of the Wolverine volcanicogenic massive sulfide (VMS) deposit and include two types: an early, ~352 Ma suite of quartz-feldspar porphyritic (QFP) rhyolite suite, and a younger, ~347 to 346 Ma feldspar porphyritic (FP) suite. These porphyry sills are pre- to synmineralization, respectively, and suggest mineralization in the Wolverine VMS deposit formed at ~347 to 346 Ma. Both suites of porphyries have indications of derivation from and/or extensive interaction with ancient upper continental crustal materials. The FP suite, however, has elevated concentrations of high field strength elements (HFSE) and rare earth elements (REE), higher zircon saturation temperatures, and higher Nb/Ta ratios, and lower Ti/Sc ratios relative to the QFP suite. These features are interpreted to reflect that the FP suite of magmas were hotter (>900°C) melts with a larger mantle component in their genesis relative to the QFP suite. Both suites, however, are interpreted to have formed due to basaltic upwelling, crustal melting, and crust-mantle mixing during ensialic back-arc basin activity. The presence of mantle heat within the Wolverine basin from ~352 to ~347 to 346 Ma, a minimum of 5 m.y., suggests that sustained mantle heat flow was critical to the genesis of the Wolverine porphyries and Wolverine VMS deposit. Upwelling mantle, therefore, may be critical in providing the heat to drive hydrothermal systems even in continental margin-type VMS environments (e.g., Bathurst, Iberian pyrite belt), where the role of the mantle typically is not recognized.

When compared to Archean rhyolitic sequences the Wolverine rhyolites have F1 and FII signatures and would be considered less prospective hosts for VMS deposits (Lesher et al., 1986; Hart et al., 2004). This is due to the fact that in the original models of Lesher et al. (1986) the F1 and FII designations are controlled primarily by depth of melt generation and garnet and amphibole stability in the crustal residues from which the rhyolites were generated. Although this assumption is valid in many instances, it is not valid in areas dominated by mature continental crust where trace element budgets are largely controlled by HFSE- and REE-enriched accessory phases (e.g., zircon, monazite) and the extent to which these phases are dissolved during crustal melting. Thus, in younger belts underlain by mature continental crust, it is most important to identify HFSE- and REE-enriched felsic rocks.

The ages of formation of the Wolverine deposit, and other Devonian-Mississippian VMS deposits in the Finlayson Lake district, are coincident with widespread extensional geodynamic episodes in the Finlayson Lake region and elsewhere in the northern Cordillera. This extensional geodynamic activity accompanied arc-rifting and ensialic back-arc basin activity in the Yukon-Tanana terrane and was coincident with the opening of the Slide Mountain back-arc basin (ocean). The ages of the Wolverine deposit are also broadly contemporaneous with a widespread global pulse of syngeneric VMS and SEDEX mineralization (e.g., Red Dog, Iberian pyrite belt, Selwyn basin, Rammselberg-Meggen and Delta-Bonnefield districts), and one of the largest oceanic anoxic events in Earth's history (Goodfellow, 1987).

We suggest that the formation of the Wolverine deposit was the product of a number of factors including: (1) an extensional geodynamic regime (e.g., ensialic back-arc basin) that produced the conduits required for upwelling hydrothermal fluids and a focusing mechanism for these fluids (i.e., synvolcanic faults); (2) anomalously high-temperature felsic magmatism (>900°C) that provided the heat flow required to drive hydrothermal circulation; and (3) anoxic bottom waters and deposition of black shales, which served as physical (and thermal) traps and promoted the deposition and preservation of massive sulfides. In the Late Devonian to Early Mississippian these conditions may have been important for regional metallogenesis throughout the northern Cordillera (Fig. 19).

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Lithogeochemistry

Lithogeochemical samples were ~1 kg or larger in size and consisted of drill core from the Wolverine deposit. Samples were pulverized in a steel jaw crushe and powdered in a ceramic mill. Rock powders were analyzed using a combination of X-ray fluorescence (XRF), inductively coupled plasma emission spectroscopy (ICP-ES), inductively coupled plasma mass spectrometry (ICP-MS), and wet chemical methods at the Geological Survey of Canada, Ottawa, ON, Canada. Major elements were determined using fused bead X-ray fluorescence (XRF). Water (H2O_total) and CO2_total were analyzed by infrared spectroscopy, and FeO was analyzed by modified Wilson titration. Sample powders for trace element analysis were dissolved in a combination of nitric, perchloric, and hydrofluoric acid, and fluxed with lithium metaborate if any residual material existed after the first acid dissolution treatment. Trace elements were analyzed by ICP-ES (Ba, La, Pb, Sc, Sr, V, Y, Yb) and ICP-MS (remaining REE, Cs, Rb, Th, U, Ga, Hf, Ta, and Zr). Analytical precision calculated from repeat analyses of internal reference materials is given as percent relative standard deviation (%RSD = 100*standard deviation/mean), and was 0.43 to 6.52 percent for the major elements, 0.72 to 8.80 percent for the transition elements (V, Ni, Cr, Co), 2.21 to 5.92 percent for the HFSE (Nb, Zr, Hf, Y, Sc, Ga), 2.35 to 6.96 percent for the low field strength elements (LFSE) Cs, Rb, Th and U, but slightly higher for Ba and Sr (1.49–15.75%), and 2.15 to 6.47 percent for the REE (La-Lu; see also Piercey et al., 2001b, 2004). Accuracy for standards analyzed at the Geological Survey of Canada has been previously reported in Piercey et al. (2001b, 2004). Further details on the methodology can be obtained from the Geological Survey of Canada at: http://gsc.nrcan.gc.ca/labs/chem_e.php.

U-Pb geochronology

Samples for U-Pb geochronology consisted of 15- to 20-kg samples of porphyritic rhyolite from the Wolverine/Lynx zone, Fisher, and Puck zones, and a felsic tuffaceous sample from the Sable zone. Zircons were separated from 15- to 20-kg samples, using conventional crushing, grinding, Wilfley table, heavy liquids, and Frantz magnetic separation techniques. U-Pb analyses were undertaken at the Pacific Centre for Isotopic and Geochemical Research at the University of British Columbia, Vancouver, BC, Canada, on an Isomass VG 54R thermal ionization mass spectrometer (TIMS). The methodology for zircon grain selection, abrasion, dissolution, geochemical preparation, and mass spectrometry is described by Mortensen et al. (1995). Most zircon fractions were air abraded (Krogh, 1982) prior to dissolution to minimize the effects of postcrystallization Pb loss. Procedural blanks for Pb and U were 2 and 1 pg, respectively. Analytical data are listed in Table 3 and are shown in conventional U-Pb concordia plots in Figure 13. Errors associated with individual analyses were calculated using the numerical error propagation method of Roddick (1987). Decay constants used are those recommended by Steiger and Jäger (1977), and compositions for initial common Pb were taken from the model of Stacey and Kramer (1975). All errors are given at the 2σ level.

Most previously dated felsic metavolcanic rocks from the Yukon-Tanana terrane typically display complex U-Pb systematics characterized by inheritance (i.e., older zircon cores) as well as variable degrees of postcrystallization Pb loss, despite employing strong abrasion to remove the outer portions of the grains prior to dissolution. Most cores in the samples are “cryptic” (i.e., they cannot be distinguished visually under a binocular microscope), but their presence is clearly indicated by the U-Pb systematics of the sample. An effective approach has been to selectively analyze grains that contain rod- or tube-shaped inclusions, generally of apatite, that pass entirely through the grain centers, thus indication that no inherited cores are present. Normally zircons containing inclusions are avoided in U-Pb dating studies because inclusions typically contain a significant amount of common Pb, which results in a less precise analysis. However, in this study there was a trade-off between obtaining a less precise analysis and being able to eliminate complications related to the presence of inherited cores.