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ARTICLE

The setting, style, and role of magmatism in the formation of volcanogenic massive sulfide deposits

Stephen J. Piercey

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Abstract Throughout Earth's history, all volcanogenic massive sulfide (VMS)-hosting environments are associated with specific assemblages of mafic and felsic rocks with distinct petrochemistry (petrochemical assemblages) indicative of formation at anomalously high temperatures within extensional geodynamic environments. In mafic-dominated (juvenile/ophiolitic) VMS environments, there is a preferential association with mafic rocks with boninite and low-Ti tholeiite, mid-ocean ridge basalt (MORB), and/or back-arc basin basalt affinities representing forearc rifting or back-arc initiation, mid-ocean ridges or back-arc basin spreading, or back-arc basins, respectively. Felsic rocks in juvenile oceanic arc environments in Archean terrains are high field strength element (HFSE) and rare earth element (REE) enriched. In post-Archean juvenile oceanic arc terrains, felsic rocks are commonly HFSE and REE depleted and have boninite like to tholeiitic signatures. In VMS environments that are associated with continental crust (i.e., continental arc and back-arc) and dominated by felsic volcanic and/or sedimentary rocks (evolved environments), felsic rocks are the dominant hosts to mineralization and are generally HFSE and REE enriched with calc-alkalic, A-type, and/or peralkalic affinities, representing continental arc rifts, continental back-arcs, and continental back-arcs to continental rifts, respectively. Coeval mafic rocks in evolved environments have alkalic (within-plate/ocean island basalt like) and MORB signatures that represent arc to back-arc rift versus back-arc spreading, respectively. The high-temperature magmatic activity in VMS environments is directly related to the upwelling of mafic magma beneath rifts in extensional geodynamic environments (e.g., mid-ocean ridges, back-arc basins, and intra-arc rifts). Underplated basaltic magma provides the heat required to drive hydrothermal circulation. Extensional geodynamic activity also provides accommodation space at the base of the lithosphere that allows for the underplated basalt to drive hydrothermal circulation and induce crustal melting, the latter leading to the formation of VMS-associated rhyolites in felsic-dominated and bimodal VMS environments. Rifts also provide extensional faults and the permeability and porosity required for recharge and discharge of VMS-related hydrothermal fluids. Rifts are also critical in creating environments conducive to preservation of VMS mineralization, either through shielding massive sulfides from seafloor weathering and mass wasting or by creating environments conducive to the precipitation of subseafloor replacement-style mineralization in sedimented rifts. Subvolcanic intrusions are also products of the elevated heat flow regime common to VMS-forming environments. Shallow-level intrusive complexes (i.e., within 1–3 km of the seafloor) may not be the main drivers of VMS-related hydrothermal circulation, but are likely the manifestation of deeper-seated mantle-derived heat (i.e., ~3–10 km depth) that drives hydrothermal circulation. These shallower intrusive complexes are commonly long-lived (i.e., millions of years), and reflect a sustained thermally anomalous geodynamic environment. Such a thermally anomalous environment has the potential to drive significant hydrothermal circulation, and, therefore multi-phase, long-lived subvolcanic intrusive complexes are excellent indicators of a potentially fertile VMS environment. The absence of intrusive complexes, however, does not indicate an area of low potential, as they may have been moved or removed due to post-VMS tectonic activity. In some cases, shallow-level intrusive systems contribute metals to the VMS-hydrothermal system.

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S. J. Piercey (✉)
Department of Earth Sciences,
Memorial University of Newfoundland,
St. John's, NL A1B 3X5, Canada
e-mail: spiercey@mun.ca

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Introduction

Magmatic heat has long been advocated as the driving force for hydrothermal circulation to form volcanogenic massive sulfide (VMS) mineralization (Spooner and Fyfe 1973; Campbell et al. 1981; Cathles 1981, 1983; Franklin et al. 1981; Galley 1993, 1996, 2003; Cathles et al. 1997). Early workers showed that the ambient, elevated geothermal gradient in ridge environments would result in convective hydrothermal circulation through the crust with associated alteration and mineralization (Spooner and Fyfe 1973). Other workers have also stressed the importance of subvolcanic intrusive complexes that underlie the massive sulfide systems, suggesting that they are the thermal engines required to drive hydrothermal circulation responsible for forming VMS deposits (Campbell et al. 1981; Franklin et al. 1981; Galley 1993, 1996, 2003; Large et al. 1996; Whalen et al. 1998; Brauhart et al. 2001; Piercey et al. 2003; Whalen et al. 2004). Heat and fluid flow modeling of these intrusive complexes, and associated oxygen isotopic work on rocks proximal to the intrusions, have also documented the role of intrusive complexes may play in driving seafloor hydrothermal systems (Cathles 1981, 1983, 1993; Paradis et al. 1993; Cathles et al. 1997; Barrie et al. 1999a, b).

Similarly, numerous lithogeochemical and petrological studies of volcanic sequences hosting VMS deposits illustrate that hydrothermal activity is associated with eruptive cycles and assemblages of volcanic and intrusive rocks that have distinctive lithogeochemical and petrogenetic histories. In a landmark paper, Lesher et al. (1986) discriminated between ore-bearing and ore-barren felsic rocks in the Superior Province of Canada using high field strength elements (HFSE) and rare earth elements (REE). They found that most VMS-bearing rhyolites were associated with high-temperature melting at shallow levels within the lithosphere; this work has been tested and confirmed by numerous subsequent workers (e.g., Barrie et al. 1993; Barrie 1995; Lentz 1998; Syme 1998; Piercey et al. 2001b; Galley 2003; Hart et al. 2004; Piercey 2007, 2010). Early studies by Swinden and colleagues showed that mafic rocks associated with VMS deposits in the Central Mobile Belt of Newfoundland, Canada, have lithogeochemical signatures indicative of high-temperature origins, rifting, and generation at high levels within the lithosphere (Swinden et al. 1989; Swinden 1991, 1996). Studies of mafic rock signatures in other VMS belts have also yielded similar conclusions regarding the importance of high-temperature magmatism and extensional geodynamic setting of formation for VMS mineralization, for example, in the Flin Flon and Snow Lake belts (Bailes and Galley 1999; Syme et al. 1999), Kidd Creek (Wyman 1999; Wyman et al. 1999), and the Finlayson Lake district (Piercey et al. 2001a, 2004). More recently, the integrated analysis of mafic and felsic geochemistry has shown that most VMS districts and subclasses of VMS deposits are associated with specific assemblages of mafic and felsic rocks called petrochemical assemblages that are indicative of high-temperature magmatism within rift environments (Piercey 2007, 2010).

This contribution reviews the specific lithogeochemical signatures and petrological suites of mafic and felsic rocks associated with VMS deposit subclasses and stratigraphic settings, building on previous compilation work of Piercey (2007, 2010). Lithogeochemical signatures of volcanic and intrusive rocks provide insight into the nature of magmatism associated with, and the tectonic settings that VMS deposits form in. Furthermore, igneous lithogeochemical signatures provide insight into the potential heat engine that may have driven hydrothermal activity and the large-scale plumbing system that may have focused hydrothermal circulation. The key roles of mantle heat and extensional geodynamic activity (i.e., rifting) in the formation of VMS systems on a regional scale, the role of subvolcanic intrusive complexes in VMS deposit genesis are also discussed.

VMS deposits, lithogeochemistry, and petrochemical assemblages

VMS deposits form as a result of the syngenetic exhalation of metalliferous hydrothermal fluids upon or near the sea floor. These deposits are classified in numerous manners (e.g., metal content and type locality), but the most robust and widely accepted classification involves the utilization of host lithostratigraphy and geodynamic setting (e.g., Barrie and Hannington 1999; Franklin et al. 2005; Galley et al. 2007). Under the lithostratigraphic classification deposits are classified into five groups, including (Fig. 1): (1) mafic (i.e., Cyprus type); (2) mafic-siliciclastic (or pelitic-mafic, i.e., Besshi type); (3) bimodal-mafic (i.e., Noranda type); (4) bimodal-felsic (i.e., Kuroko-type); and (5) felsic-siliciclastic (or bimodal-siliciclastic, i.e., Bathurst-type). The first three groups are hosted by sequences dominated by mafic footwall rocks with varying amounts of siliciclastic and chemical sedimentary rocks, minor felsic rocks in the case of bimodal-mafic environments, and mafic to ultramafic intrusive rocks; these sequences are commonly juvenile and have very little continental crustal influence. In contrast, felsic and sedimentary rocks dominate the last two groups and these environments are commonly associated with evolved continental crust. The variation in substrate (i.e., juvenile versus evolved

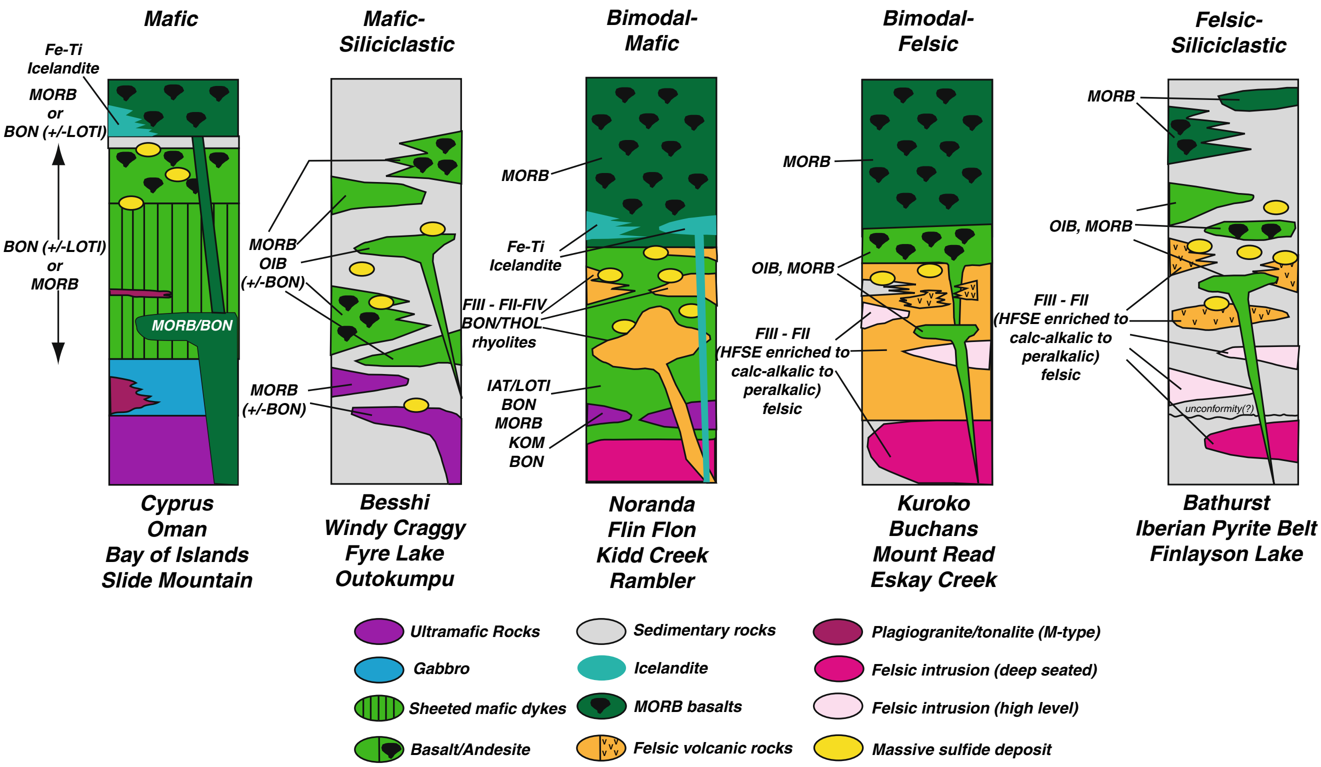


Fig. 1 Stratigraphic relationships and potential chemostratigraphic associations (i.e., petrochemical assemblages) for different VMS deposit groups. Modified from Piercey (2010)

crust) exerts a strong influence on the petrology and lithogeochemistry of felsic and mafic magmatism associated with VMS deposits and their environments (e.g., Piercey 2007, 2010).

The specific lithogeochemical attributes of volcanic rocks associated with the spectrum of VMS deposit environments and associated petrochemical assemblages are outlined below. Concentration is primarily on felsic and mafic rocks because most VMS districts are either bimodal in nature, or are associated with predominantly mafic or felsic rocks. The data presentation has also concentrated on the immobile HFSE and REE attributes of these rocks as these elements are generally immobile during alteration and metamorphism (e.g., MacLean 1990; MacLean and Barrett 1993; Barrett and MacLean 1999) and provide insight into primary igneous processes (e.g., mantle and crustal sources and igneous fractionation) and the tectonic settings of VMS deposit formation.

Mafic rocks associated with VMS deposits

Most VMS-bearing volcanic belts have specific petrological suites of rocks. Juvenile, mafic-dominated VMS environments (i.e., mafic, bimodal-mafic, mafic-siliciclastic deposit groups) contain boninite (BON) and low-Ti tholeiite (LOTI), island arc tholeiites (IAT), or mid-ocean ridge basalts (MORB) of both the normal and enriched (E-MORB) varieties (Figs. 2, 3, and 4; Table 1). There are specific associations of these signatures with different geodynamic environments and deposit associations, however. Boninite and associated rocks (i.e., LOTI and IAT) are associated with many ophiolite-hosted (mafic) VMS deposits, particularly those in forearc ophiolites, and are associated with some bimodal-mafic systems (e.g., Cycle 1 in Snow Lake), and more rarely in mafic-siliciclastic settings (Table 1). MORB are associated with VMS deposits in back-arc ophiolites, and modern mid-ocean ridges and back-arc basins (Figs. 2, 3, and 4). MORB-like rocks with weak negative Nb anomalies on primitive mantle normalized plots (Fig. 1), called back-arc basin basalts (BABB), are also present in mafic-type VMS environments including modern back-arc basins and back-arc ophiolites. In some mafic and bimodal-mafic environments (e.g., forearc or back-arc settings) MORB-type rocks either underlie boninite or overlie and/or cross-cutting boninite (Fig. 1; Table 1). In the case of forearc settings, MORB commonly underlies the boninite, whereas back-arc environments may have BON-LOTI-IAT overlain by MORB and BABB recording the transition from arc magmatism to back-arc rifting and spreading (Fig. 1). MORB-type rocks are also associated with mafic-siliciclastic environments in both the ancient record and modern sedimented ridges (Figs. 1, 4, and 5; Table 1). MORB, BABB, and BON-LOTI-IAT are all interpreted to have formed as a result of melting incompatible element-depleted (MORB, BABB, and IAT)

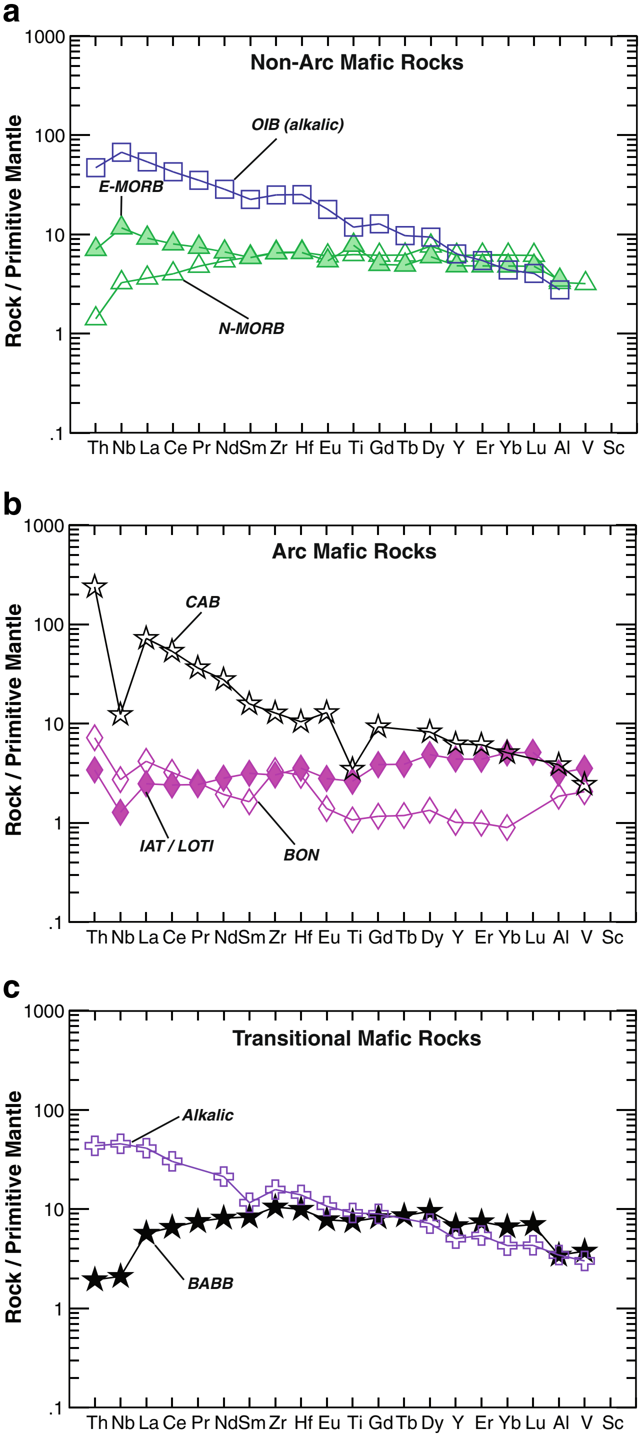


Fig. 2 Primitive mantle normalized trace element plots for mafic rocks: (**a**) non-arc rocks, (**b**) arc rocks, and (**c**) transitional rocks. *N-MORB* normal mid-ocean ridge basalts, *E-MORB* enriched mid-ocean ridge basalts, *OIB* ocean island basalts, *BON* boninite, *IAT/LOTI* island arc tholeiite/low-Ti island arc tholeiite, *CAB* calc-alkaline basalts, and *BABB* back-arc basin basalt. Data from Sun and McDonough (1989), Stoltz et al. (1990), Piercey et al. (2004), Kepezhinskas et al. (1997), and Ewart et al. (1994). Normalization values from Sun and McDonough (1989)

to ultra-depleted mantle (BON-LOTI) at elevated temperatures (i.e., T>1,200°C) (e.g., Crawford et al. 1981; McKenzie and Bickle 1988; Crawford et al. 1989; McKenzie and O'Nions 1991; Langmuir et al. 1992; Pearce et al. 1992; Stern and Bloomer 1992). In addition, these types of magmatic products form within extensional geodynamic settings, including MORB (e.g., Sinton and Detrick 1992), which are an extremely rare setting for VMS in the ancient record (Galley and Koski 1999), back-arc basins (BABB and MORB) (Hawkins 1995), forearcs (low Ca-BON to LOTI to IAT) (Pearce et al. 1992; Stern and Bloomer 1992), and nascent back-arc basins (high Ca-BON to LOTI to IAT to BABB to MORB) (Crawford et al. 1981, 1989; Piercey et al. 2001a).

In evolved continental arc to back-arc environments (i.e., bimodal-felsic and felsic-siliciclastic), VMS deposits are associated with minor amounts of mafic rocks that have MORB and alkalic (or within-plate/ocean island basalt (OIB)) signatures, which commonly overlie or cross-cut felsic volcanic and intrusive rocks (Figs. 1, 2, 4, and 5), and represent arc rifting to back-arc spreading. The MORB-like rocks in evolved environments are commonly E-MORB and less commonly there is a complete range of mafic rocks from incompatible element-depleted MORB, to weakly incompatible element-enriched E-MORB, to incompatible element-enriched OIB signatures (Fig. 5). The MORB- and OIB-like rocks commonly occur as sills and dykes that cross-cut mineralization, or as flows that overlie felsic rocks and the associated mineralization (i.e., they commonly post-date the main mineralization event). Furthermore, there is commonly a stratigraphic progression upwards from alkalic basalts to MORB (van Staal et al. 1991b; Shinjo et al. 1999; Piercey et al. 2002a, b). Alkalic and MORB-type basalts are associated with many bimodal-felsic and felsic-siliciclastic settings from both the modern (e.g., Bransfield Strait and Okinawa Trough) and ancient (e.g., Bathurst, Iberian Pyrite Belt, Finlayson Lake, and Eskay Creek) geological record (Figs. 4 and 5; Table 1), and are interpreted to represent the transition from melts derived from enriched, lithospheric mantle sources (alkalic) to depleted, asthenospheric mantle sources (MORB). The associated stratigraphic progression from alkalic basalt to MORB is commonly interpreted to reflect a shift from arc rifting to true back-arc seafloor spreading (e.g., van Staal et al. 1991b; Goodfellow et al. 1995; Barrett and Sherlock 1996; Almodóvar et al. 1997; Shinjo et al. 1999; Colpron et al. 2002; Piercey et al. 2002a, b).

Felsic rocks associated with VMS deposits

Felsic rocks associated with VMS deposits have been the topic of considerable research (e.g., Lesher et al. 1986;

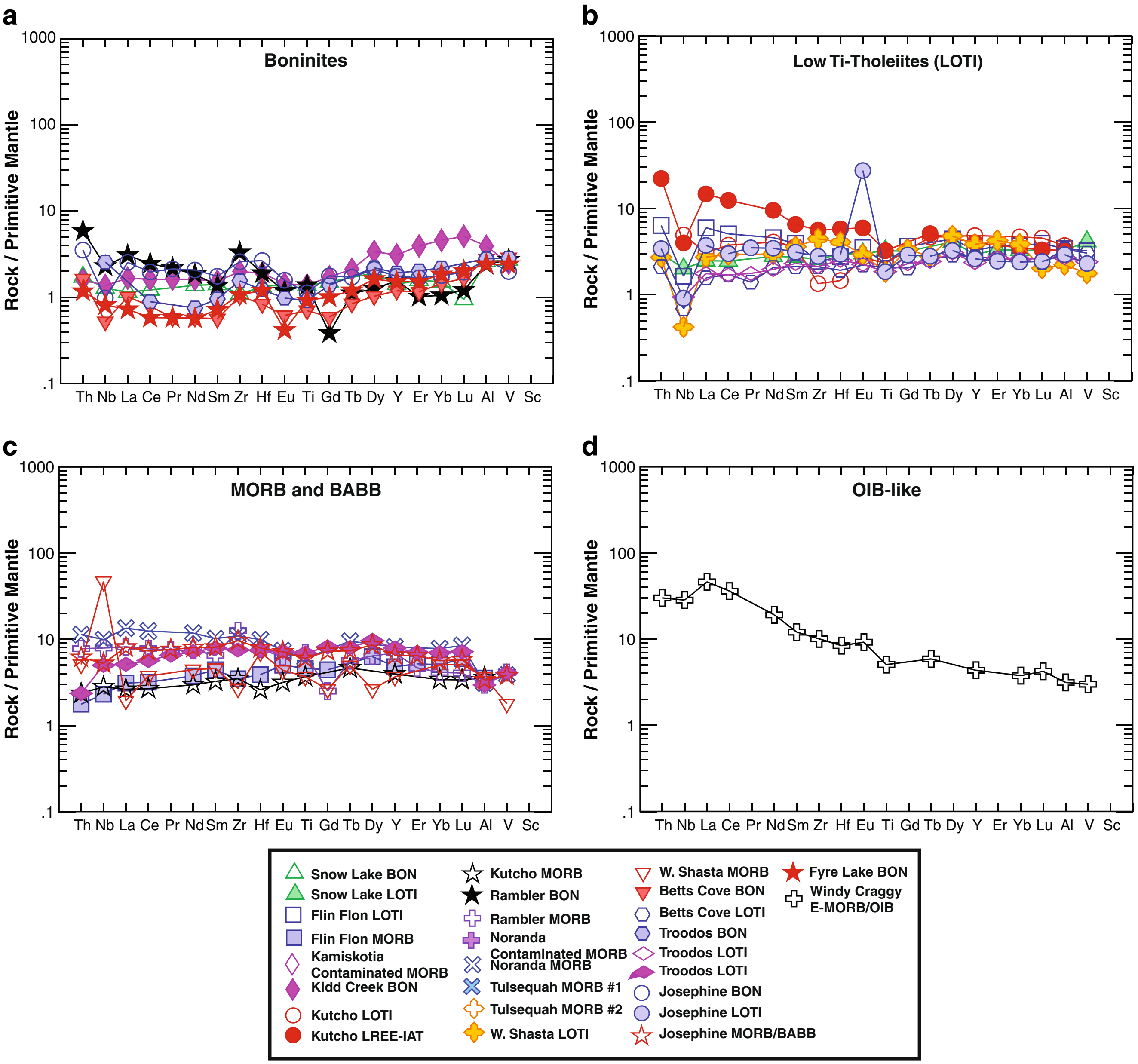


Fig. 3 Primitive mantle normalized trace element plots for mafic rocks from mafic-dominated (juvenile) VMS environments. Data sources in the Appendix. Normalization values from Sun and McDonough (1989)

Barrie et al. 1993; Lentz 1998; Hart et al. 2004). The petrology of felsic rocks associated with VMS environments is strongly dependent on the type of crust they are underlain by (i.e., juvenile or evolved) and this also varies as a function of age (i.e., Archean or post-Archean).

Much of our knowledge of Archean VMS-associated rocks comes from the Superior Province, where numerous workers have classified ore-bearing and barren assemblages based on the trace element geochemistry of felsic rocks (Lesher et al. 1986; Barrie et al. 1993; Kerrich and Wyman 1996; Kerrich and Wyman 1997; Hart et al. 2004). Lesher et al. (1986) outlined a threefold subdivision of VMS-fertile versus barren felsic rocks for the Superior Province—the FI to FIII suites of felsic volcanic rocks (Fig. 6a) that reflect increasingly more primitive terrains and hotter magmatic products formed at shallower levels in the lithosphere; Barrie (1995) further quantified the thermal histories of these rocks using zircon saturation temperatures. This classification was modified and revised by Barrie et al. (1993), and Hart et al. (2004) subsequently added a fourth suite to the classification, the FIV suite, and expanded it to post-Archean terrains (Fig. 6). In Archean environments VMS deposits are preferentially associated with FIII and FII suite rocks, with very few VMS occurrences associated

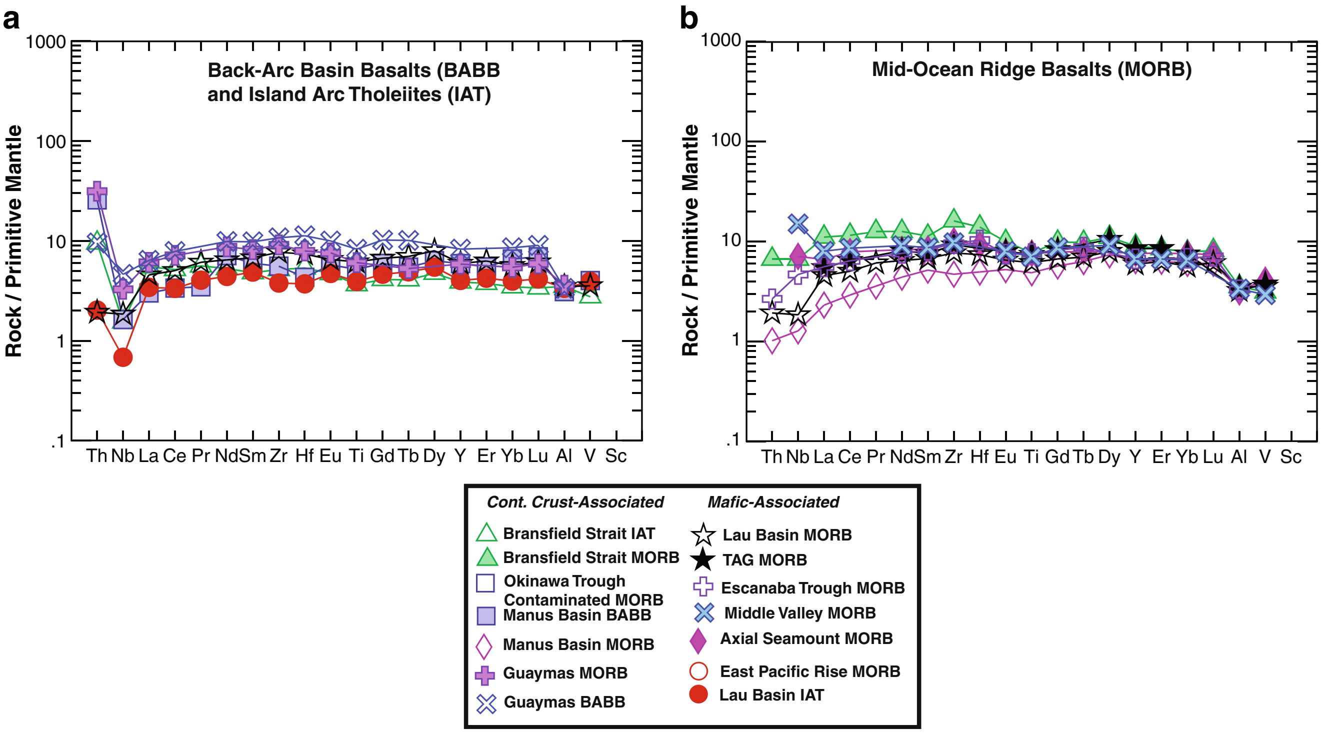


Fig. 4 Primitive mantle normalized plots for mafic rocks from modern VMS environments. Data sources in the Appendix. Normalization values from Sun and McDonough (1989)

with the FI suite (Fig. 5). The FIII felsic rocks have low La/Yb and Zr/Y, elevated HFSE and REE contents (e.g., Zr> 200 ppm), including high Y and Yb, and flat chondrite-normalized REE profiles (Figs. 6 and 7). The FI suite of felsic rocks has high La/Yb and Zr/Y, lesser HFSE and REE contents, low Y and Yb, and steep chondrite-normalized REE profiles, whereas the FII suite has characteristics transitional between the FI and FIII suites (Figs. 6 and 7). The majority of Archean VMS deposits is hosted by FIII and FII felsic rocks (Fig. 7), which are interpreted to have formed within Archean rift sequences from high-temperature melts (T>900°C) derived from melting of hydrated basaltic crust at shallow to mid-crustal depths during extension (Lesher et al. 1986; Barrie et al. 1993; Barrie 1995; Prior et al. 1999; Hart et al. 2004).

The composition of post-Archean felsic rocks associated with VMS deposits depends on whether they have formed in juvenile or evolved environments. Felsic rocks in evolved, post-Archean terrains are characterized by HFSE (e.g., Zr, Nb) and REE enrichment and within-plate (A-type) to peralkalic signatures (Figs. 8 and 9) (McConnell 1991; Lentz 1998, 1999; Whalen et al. 1998; Piercey et al. 2001b; Dusel-Bacon et al. 2004; Hart et al. 2004). These felsic rocks also have FIII to FII signatures with a greater abundance of FII signatures in post-Archean felsic rocks from evolved environments (Fig. 8) (Lentz 1998; Whalen et al. 1998; Lentz 1999; Piercey et al. 2001b; Hart et al. 2004). Geochemical and isotopic data for felsic rocks from these evolved settings are consistent with their derivation via partial melting of evolved continental crust, and sedimentary rocks derived thereof, at high temperatures (T>900°C) within rift environments (Lentz 1998, 1999; Whalen et al. 1998; Piercey et al. 2001b, 2003, 2008).

Post-Archean felsic rocks from juvenile terrains (e.g., Flin Flon, parts of Snow Lake, and Rambler) contrast with their evolved counterparts in that they are depleted in trace elements, commonly with very low HFSE and REE contents (e.g., Zr<50–100 ppm), FIV signatures, tholeiitic and M-type (i.e., derived from a mafic substrate) Zr-Y and Nb-Y systematics, and arc tholeiite- to boninite-like chondrite-normalized REE patterns (Figs. 10 and 11). These rhyolites are the result of partial melting of mafic (to andesitic) substrates during forearc rifting, intra-arc rifting, or rifting during the initiation of back-arc basin activity (e.g., Shukuno et al. 2006). In some cases, these M-type rocks are found associated with isotropic gabbro and tondjhemite-tonalite in ophiolite settings, but here they are not directly associated with VMS deposits (Galley and Koski 1999). The low overall trace element contents of the rocks above are likely a function of the low trace element abundances of their mafic source rocks (e.g., boninite and arc tholeiite) that were partially melted.

Table 1 Petrochemical assemblages of mafic and felsic rocks associated with different VMS deposit classes

VMS deposit group	Mafic	Felsic	Examples	Other relationships
Mafic (Cu–Zn)	Boninite, low-Ti tholeiite,IAT, MORB, BABB	–	Boninite/LOTI—Troodos, Semail, Turner-Albright, Betts Cove, Kidd Creek, Snow Lake, Rambler-Ming; MORB—East Pacific Rise, TAG; BABB—Lau Basin, Manus Basin, Semail	Forearc ophiolite: MORB underlies BON-LOTI-IAT. Back-arc ophiolite: BON-LOTI-IAT overlain by MORB and BABB
Mafic-siliciclastic (Cu–Zn–Co)	MORB, alkalic, boninite (rare)	–	MORB—Middle Valley, Guaymas, Escanaba Trough; boninite—Fyre Lake; alkalic/OIB—Windy Craggy	MORB, alkalic, boninite—often sill-sediment complexes. Geochemical diversity is very rare
Bimodal-mafic (Zn–Cu)	MORB, boninite, low-Ti tholeiite (calc-alkalic and island arc tholeiites present but rarer)	Archean—FIII rhyolites. Proterozoic-Phanerozoic—tholeiitic rhyolites, boninitic rhyolites	MORB, FIII rhyolite—Noranda; boninite/LOTI, FIII rhyolite—Kidd Creek; boninite/LOTI, arc tholeiite, tholeiitic rhylolites—Snow Lake (Cycle 1); boninite/LOTI and boninitic rhyolites—Rambler-Ming	Mafic rocks form bulk of stratigraphy, but deposits proximal to rhyolites
Bimodal-felsic (Zn–Pb–Cu)	MORB, alkalic	HFSE-enriched rhyolites (A-type), peralkaline and calc-alkalic rhyolites (rarer)	HFSE-enriched rhyolites, MORB/alkalic basalts—Eskay Creek; calc-alkalic rhyolites, MORB/alkalic basalts—Iberian Pyrite Belt, Mount Read	Rhyolitic rocks form bulk of stratigraphy, cross-cut and overlain by alkalic and/or MORB basalts
Felsic-silicilastic (Zn–Pb–Cu)	MORB, alkalic	HFSE-enriched rhyolites, peralkaline, and calc-alkalic rhyolites (rarer)	HFSE-enriched rhyolites, MORB/alkalic basalts—Bathurst, Finlayson Lake; Peralkaline rhyolites, MORB/alkalic basalts—Avoca, Delta-Bonnifield	Felsic rocks with abundant sedimentary rocks, commonly in greater abundance. Felsic rocks are commonly volcaniclastic in nature. Mafic rocks cross-cut and overlie felsic and/or sedimentary substrate

Petrochemical assemblages

Petrochemical assemblages are specific associations of petrological suites of mafic and felsic rocks present in specific VMS deposit environments (Table 1; Fig. 11). Not only are there petrological associations, there are specific chemostratigraphic relationships between these petrological suites of rocks (Fig. 1). In all VMS deposit environments, there is the common association of high-temperature magmatism, be it felsic or mafic, generated at shallow levels within the crust and mantle, and evidence for emplacement within rift environments (e.g., Lesher et al. 1986; Swinden et al. 1989; Swinden 1991; Barrie et al. 1993; Syme and Bailes 1993; Barrie 1995; Lentz 1998; Syme 1998; Bailes and Galley 1999; Bedard 1999; Prior et al. 1999; Syme et al. 1999; Piercey et al. 2001a, b; Hart et al. 2004; Piercey 2007).

In most mafic environments, there is the common association with boninite-low Ti tholeiites and/or MORB/BABB suite rocks, commonly with a progression from one suite to the other (e.g., Flin Flon-Snow Lake, Rambler-Ming, Turner-Albright, Fyre Lake, and Kidd Creek), indicative of forearc rifting or back-arc basin initiation (e.g., Cameron 1985; Piercey et al. 1997; Bailes and Galley 1999; Syme et al. 1999; Wyman et al. 1999; Bailey 2002; Harper 2003). In bimodal-mafic environments, the felsic rocks, be they FIII–FII in Archean environments or FIV in post-Archean environments, commonly occur at the transition between the arc and non-arc suite mafic rocks (e.g., Piercey et al. 1997; Bailes and Galley 1999; Prior et al. 1999; Syme et al. 1999; Wyman et al. 1999; Bailey 2002; Harper 2003). Volcanogenic massive sulfide deposits hosted by modern ridges, back-arc basins, and back-arc ophiolites have mafic rocks with MORB and BABB signatures (e.g., Pearce et al. 1981; Saunders et al. 1982; Swinden 1991; Hawkins and Allan 1994; Stakes and Franklin 1994; Petersen et al. 2004; Fretzdorff et al. 2006).

In most felsic-dominated environments there is an association of VMS deposits with HFSE- and REE-enriched (A-type) to calc-alkalic rocks (for the most part FII rocks with subordinate FI rocks), which are typically overlain and cross-cut by rocks of MORB and OIB affinities (Table 1; Fig. 1) (McConnell 1991; McConnell et al. 1991; van Staal et al. 1991a; Stolz 1995; Barrett and

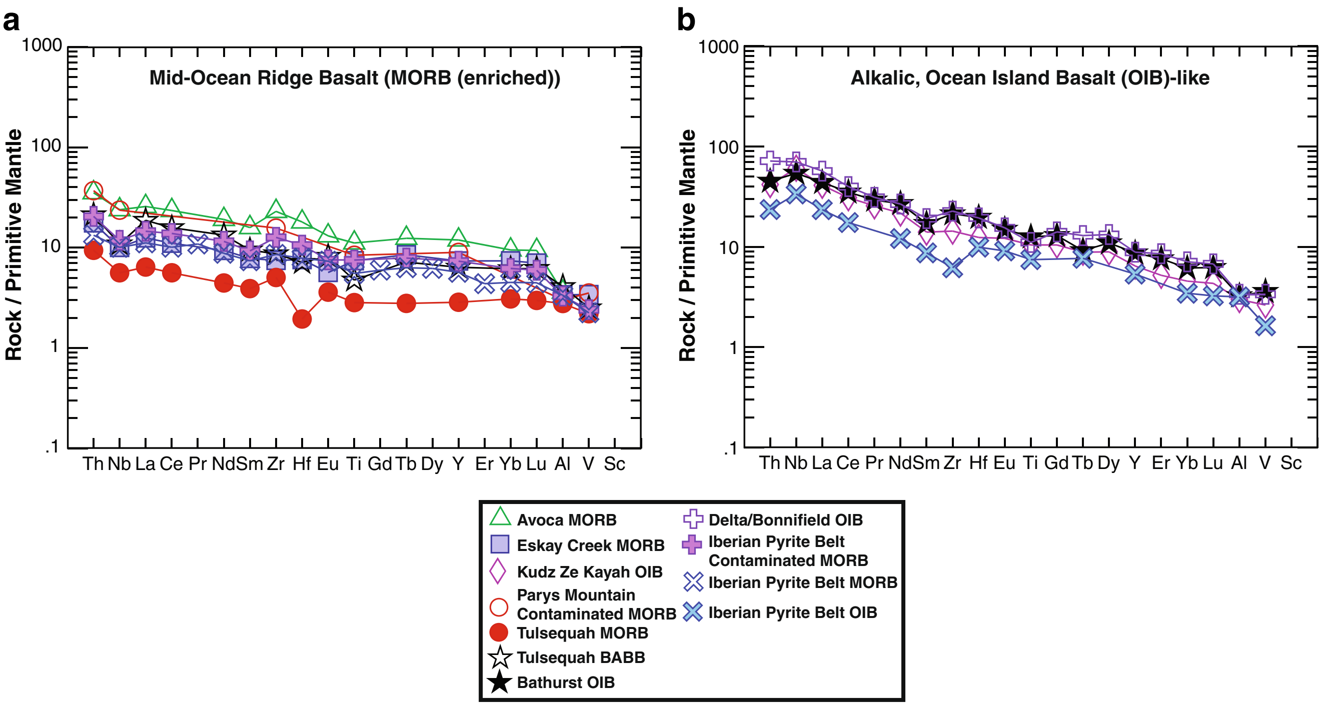


Fig. 5 Primitive mantle normalized trace element plots for mafic rocks from felsic-dominated (evolved) VMS environments. Data sources in the Appendix. Normalization values from Sun and McDonough (1989)

Sherlock 1996; Mitjavila et al. 1997; Thiéblemont et al. 1997; Lentz 1999; McConnell 2000; Piercey et al. 2001b; Piercey et al. 2002a, b; Rogers et al. 2003; Dusel-Bacon et al. 2004). Mineralization is commonly located at the transition from felsic-dominated to mafic-dominated magmatism; however, it can occur at any stratigraphic levels beneath the mafic assemblages, including, in some cases, within the sedimentary rocks intercalated with the volcanic rocks (e.g., Bathurst, Finlayson Lake, and Iberian Pyrite Belt) (Fig. 1).

Discussion

Mantle heat: a key in the generation of VMS systems

In all of the VMS districts mentioned, above there is a common association with mafic magmatism, with environments dominated by mafic magmatism forming the most of the footwall and in some cases hosting mineralization (e.g., mafic, mafic-siliciclastic, and bimodal-mafic), or with significant mafic magmatism post-dating mineralization in felsic-dominated environments (e.g., bimodal-felsic and felsic-siliciclastic). In mafic-dominated VMS environments, VMS deposits are associated with boninite-LOTI suite magmatic rocks and/or MORB/BABB (Table 1; Figs. 1 and 12a). These suites formed at high liquidus temperatures, in contrast to normal arc magmatic rocks, with most having liquidus temperatures >1200°C (e.g., MORB and BABB; McKenzie and Bickle 1988; McKenzie and O'Nions 1991; Langmuir et al. 1992; Hawkins and Allan 1994; Gribble et al. 1996; Falloon et al. 1999) to in excess of >1,400°C (e.g., boninites and LOTI; Crawford et al. 1989; Falloon and Crawford 1991; Falloon and Danyushevsky 2000).

The role of mafic magmatism in felsic- and sediment-dominated environments is less obvious; nevertheless, in these environments MORB/BABB and OIB suites are dominant and typically cross-cut and overlie felsic and sedimentary rocks (Table 1; Fig. 12b). The OIB-like rocks are commonly the first mafic magmatic products after felsic magmatism, but are generally low in volume, and they are commonly overlain by more voluminous MORB/BABB (van Staal et al. 1991b; Shinjo et al. 1999; Piercey et al. 2002a, b). Althought the mafic rocks typically post-date the VMS-hosting felsic rocks, commonly forming (and erupting) as the arc rift evolves to mature back-arc spreading, their source basaltic magmas are immediately responsible for causing partial melting of crustal material leading to the generation of the high-temperature felsic rocks ($T>900$°C) that host mineralization (Stoltz 1995; Lentz 1998; Whalen et al. 1998; Lentz 1999; Piercey et al. 2001b). A similar model has been proposed for felsic rocks in bimodal-mafic environments (e.g., Meijer 1983; Barrie 1995; Syme 1998; Prior et al. 1999; Schmitt and Vazquez 2006; Shukuno et al. 2006). In most cases, VMS deposits throughout geological time are associated with thermally anomalous geodynamic

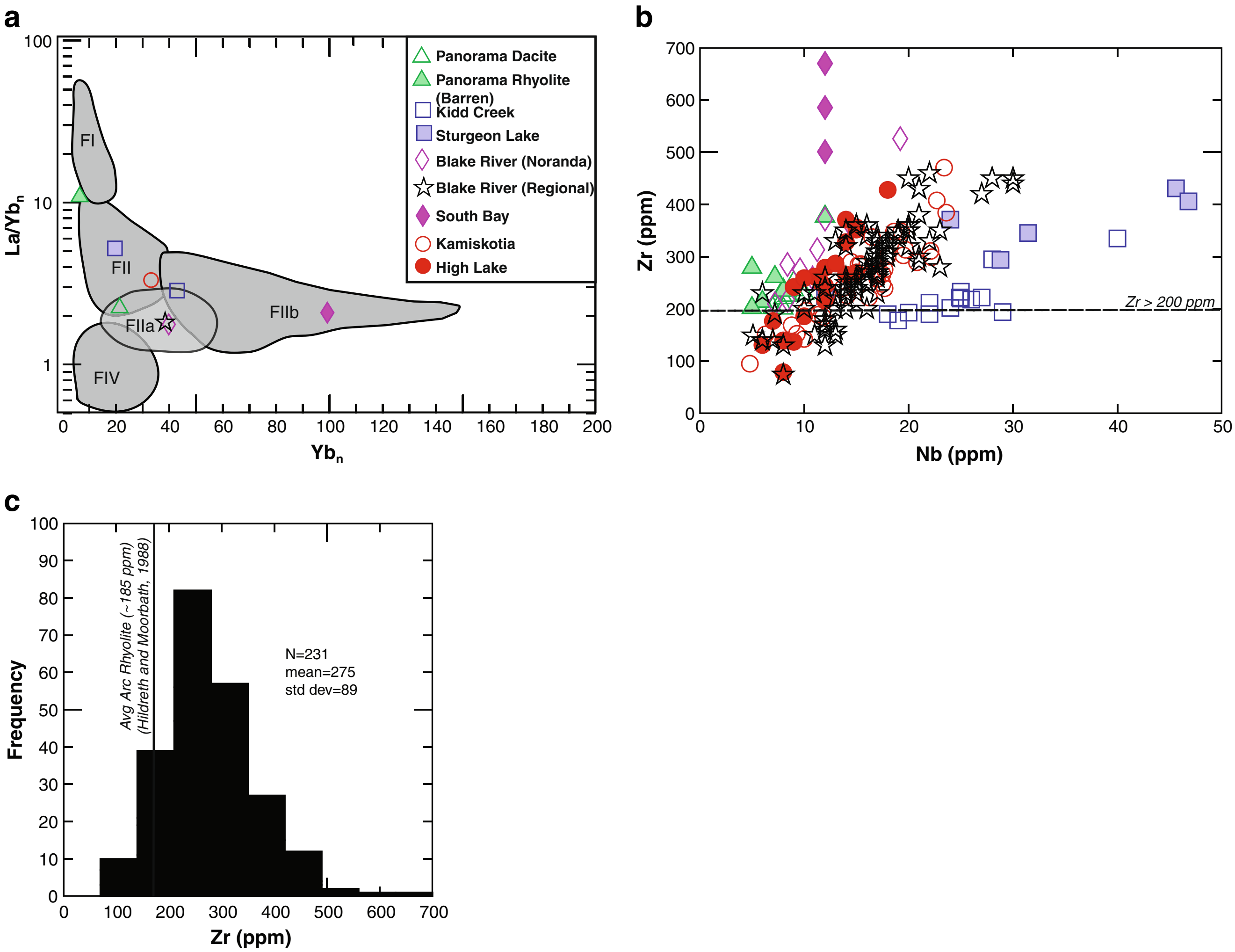


Fig. 6 Trace element diagrams for Archean VMS-associated rhyolites. **a** La/Yb$_n$–Yb$_n$ (n = chondrite normalized) plot outlining the fields for FI–FIV rhyolites (diagram from Lesher et al. 1986; Hart et al. 2004). **b** Zr–Nb plot from Leat et al. (1986). **c** Histogram of Zr contents of VMS-associated rhyolites. Notably, all Archean VMS-associated rhyolites have high HFSE and REE contents. Data sources in the Appendix. Data for Andean rhyolites from Hildreth and Moorbath (1988)

environments and high-temperature thermal corridors (e.g., Galley 2003), and are commonly associated with the upwelling of hot, mantle-derived magmas.

In addition to their high heat flow, the MORB/BABB and boninite-LOTI are indicative of generation at shallow levels within the mantle. Shown on Fig. 12 are plots of Nb/Yb and TiO$_2$/Yb ratios and the average values for the mafic rocks associated with VMS deposits in various environments. Niobium and Yb are strongly incompatible during mantle partial melting and ratios of these elements are relatively unaffected by partial melting, except at extremely low degrees of melting (i.e., <2–3%); hence, ratios of these elements can provide insight into the source region for the mafic rocks (Pearce 1983, 2008; Pearce and Parkinson 1993; Pearce and Peate 1995). TiO$_2$ and Yb, however, can be affected by the presence of garnet in the melt residue, leading to elevated TiO$_2$/Yb ratios in mafic rocks derived from deeper mantle in the garnet stability field (Pearce and Peate 1995; Pearce 2008). On the TiO$_2$/Yb-Nb/Yb diagram (Fig. 12) samples that lie within the MORB array are indicative of generation at shallow levels where spinel or plagioclase are stable in the residues (<50–100 km depth in the mantle; Wyllie 1981; McKenzie and Bickle 1988; Pearce and Peate 1995; Pearce 2008). In contrast, those that lie above the MORB array are generated where garnet is present in the residue leading to high TiO$_2$/Yb, indicating generation at greater depths (>100 km depth in the mantle; Wyllie 1981; McKenzie and Bickle 1988; Pearce and Peate 1995; Pearce 2008). In the mafic-dominated sequences and those from modern environments, most samples lie within the MORB array, indicative of shallow-level melting within the mantle (Fig. 12). Similarly, the majority of samples from the felsic- and sediment-dominated VMS environments also lie within the MORB array, with the exception

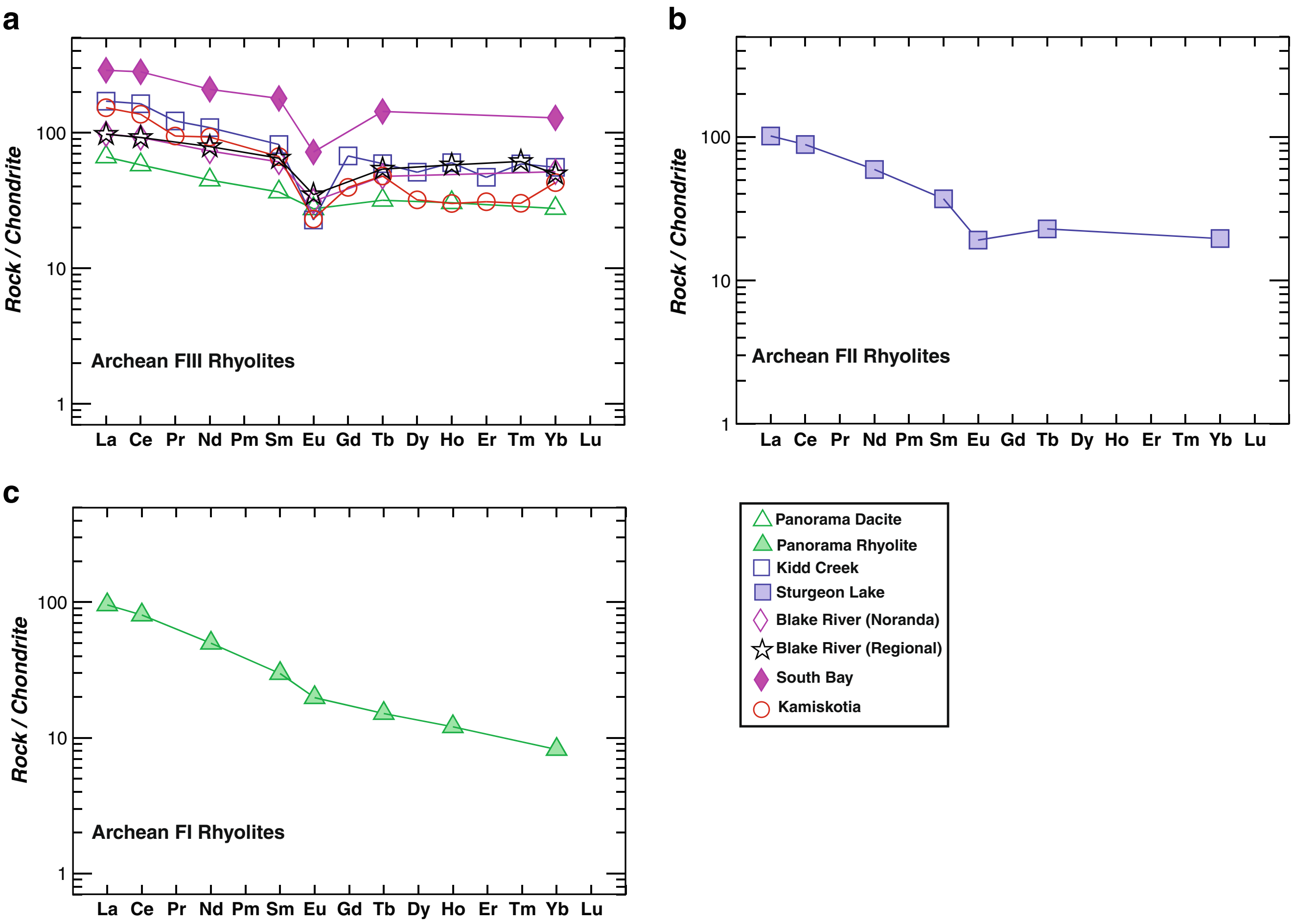


Fig. 7 Chondrite-normalized REE profiles for Archean rhyolites, including: **a** FIII rhyolites, **b** FII rhyolite, and **c** FI rhyolite. Data sources in the Appendix. Normalization values from Sun and McDonough (1989)

of the alkalic/OIB-like samples, that are also indicative of shallow-level, low degree partial melting of hydrated lithospheric mantle during continental arc rifting (Fig. 12).

There are two important consequences of the presence of high-temperature mafic magmas derived from shallow melting (i.e., <50–100 km depth) within the mantle. Firstly, shallow melting is likely to increase the probability of rapid transfer of mafic magma from the source of melting to the subvolcanic environment without significant loss of heat, in contrast to melts derived at depth; the latter would have greater probability of losing heat upon transit to the subvolcanic environment (all conditions being equal aside from the depth of melting). Secondly, although a number of factors can control the fraction of melt generated during partial melting (e.g., H_2O content and bulk rock composition), melting at shallow levels results in a greater volume of melt than melting at deeper levels in the mantle (Fig. 13) (McKenzie and O'Nions 1991; Ellam 1992; Williamson et al. 1995). The rise of significant volumes of hot, mafic magma to the base of the crust (and higher) will result in an elevated geothermal gradient (Fig. 14a) (e.g., Hyndman et al. 2005; Currie and Hyndman 2006), which will allow crustal partial melting and vigorous hydrothermal circulation, assuming suitable permeability (e.g., Cathles 1981, 1983; Cathles et al. 1997; Barrie et al. 1999a, b); this will then induce crustal melting in bimodal environments to produce the felsic volcanic and subvolcanic rocks associated with VMS deposits (e.g., Meijer 1983; Stoltz 1995; Lentz 1998, 1999; Syme 1998; Whalen et al. 1998; Prior et al. 1999; Piercey et al. 2001b; Schmitt and Vazquez 2006; Shukuno et al. 2006).

The presence of shallow-level, mantle-derived basaltic magma within thermal corridors defined by rifts or calderas is a major factor in generating VMS-hydrothermal systems on a geodynamic scale. Predictably, such magma suites do not only have specific geochemical compositions (e.g., boninite and MORB), but also predictable geological relationships and petrological associations, including: (1) mafic and felsic magmatism over extended periods of geological time (i.e., several millions of years) (e.g., Bleeker and Parrish 1996; Piercey et al. 2008); (2) evidence for polyphase subvolcanic intrusive complexes emplaced

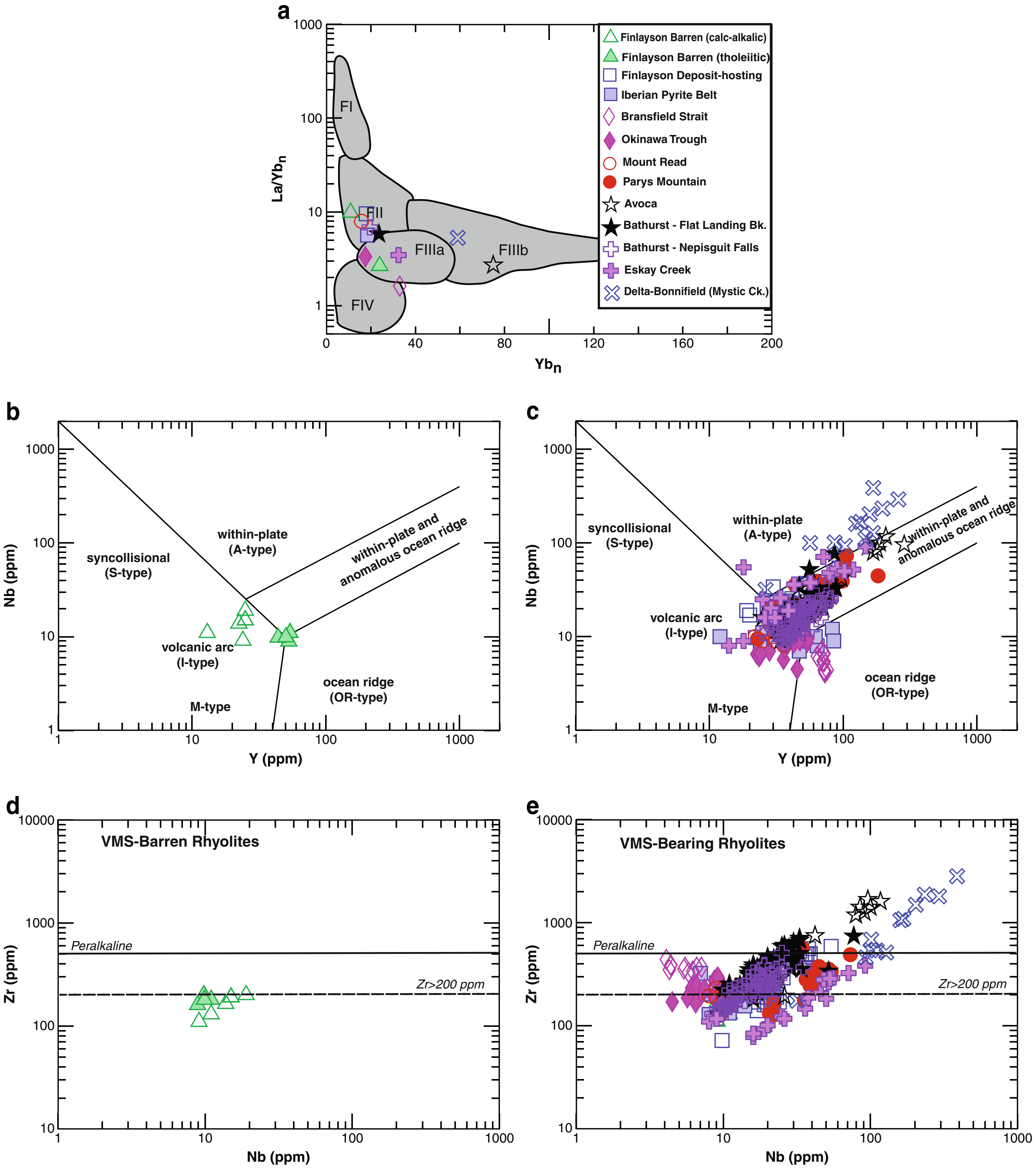


Fig. 8 Trace element diagrams for rhyolites from post-Archean felsic-dominated (evolved) VMS environments: **a** La/Yb$_n$–Yb$_n$ (n = chondrite normalized) plot outlining the fields FI–FIV rhyolites (diagram from Lesher et al. 1986; Hart et al. 2004). **b**, **c** Nb-Y diagram from Pearce et al. (1984) for VMS-barren and VMS-bearing rhyolites, respectively. **d**, **e** Zr–Nb diagram for VMS-barren and VMS-bearing rhyolites, respectively. Data sources in the Appendix

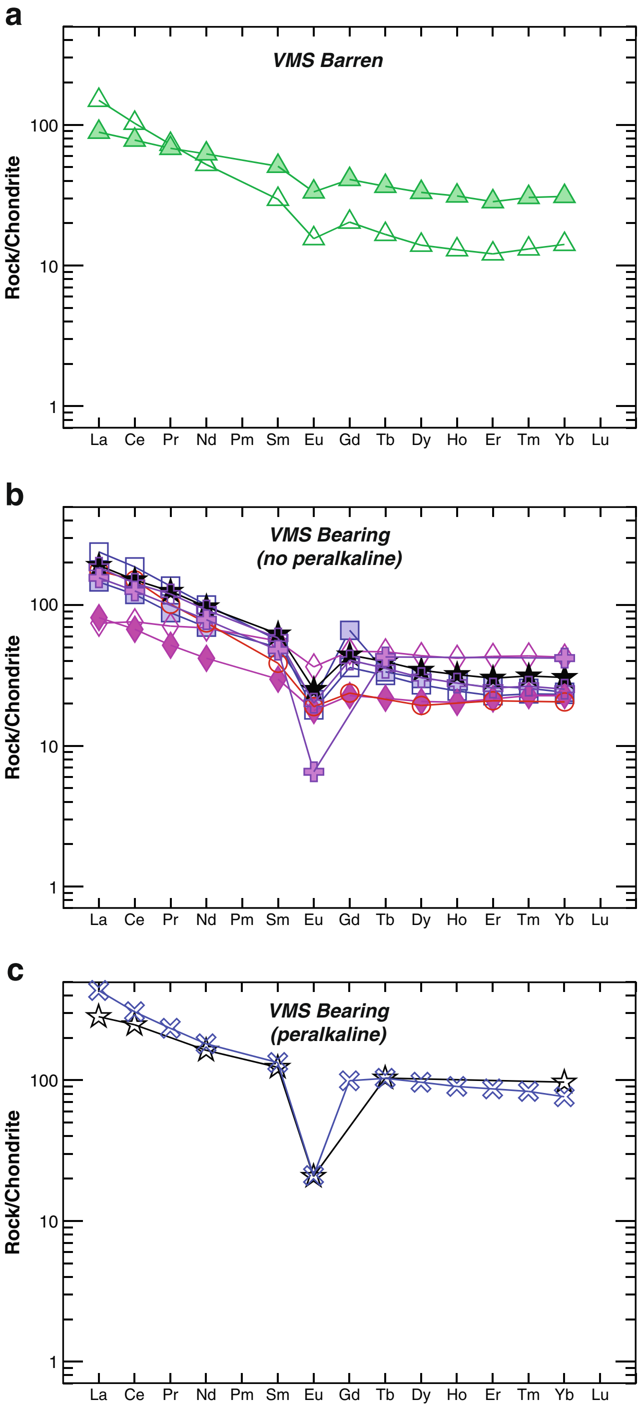


Fig. 9 Chondrite-normalized REE plots for rhyolites from post-Archean felsic-dominated (evolved) VMS environments: **a** VMS-barren rhyolites, **b** VMS-bearing rhyolites that are not peralkaline, and **c** peralkaline VMS-bearing rhyolites. Data sources in the Appendix. Symbols as in Fig. 8. Normalization values from Sun and McDonough (1989)

over extended periods of time (e.g., Galley et al. 2000; Galley and van Breemen 2002; Galley 2003); and (3) specific petrochemical assemblages indicative of high-temperature mafic and felsic magmatism within an exten-

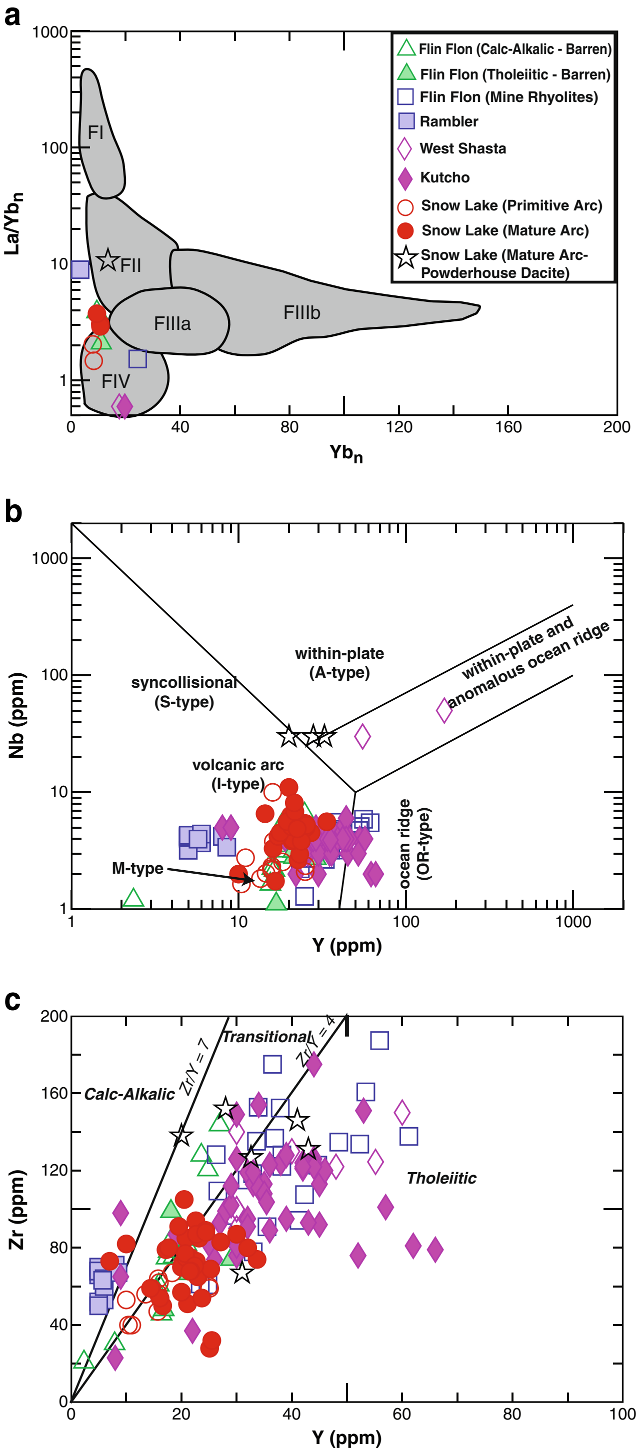


Fig. 10 Trace element diagrams for rhyolites from post-Archean mafic-dominated (primitive) VMS environments: **a** La/Yb$_n$–Yb$_n$ (n = chondrite normalized) plot outlining the fields for FI–FIV rhyolites (diagram from Lesher et al. 1986; Hart et al. 2004). **b**, **c** Nb-Y diagram from Pearce et al. (1984) for VMS-barren and VMS-bearing rhyolites, respectively. Data sources in the Appendix

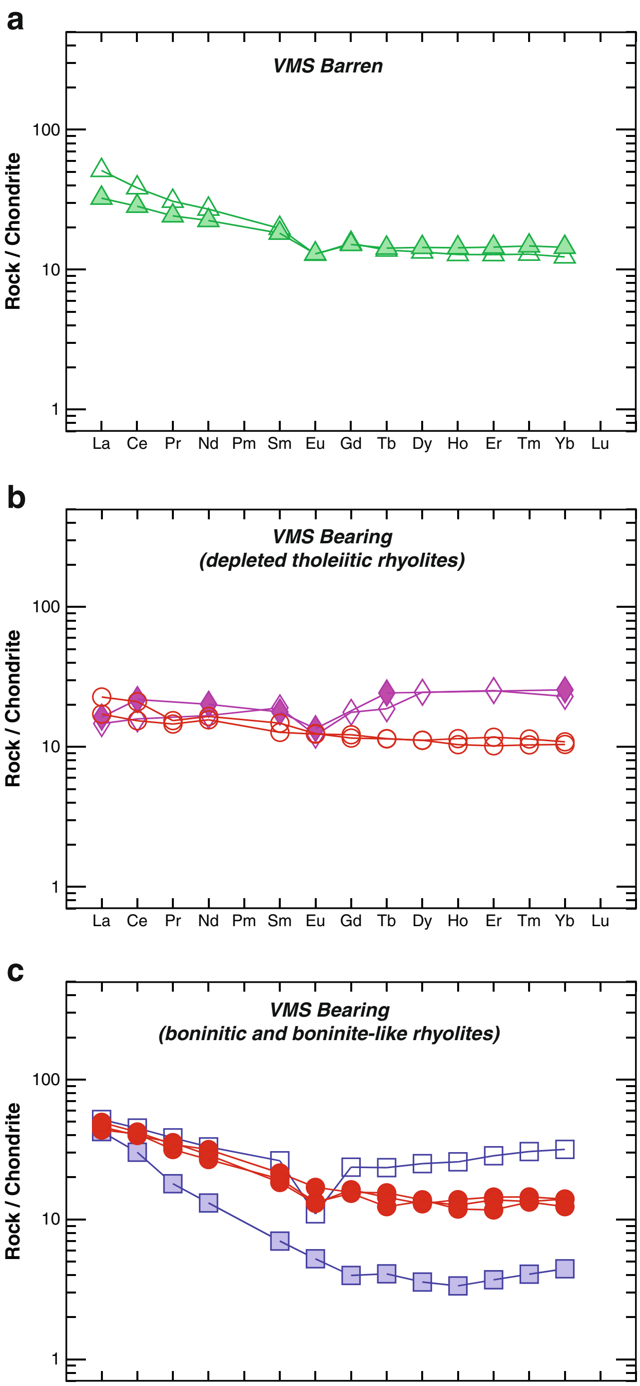


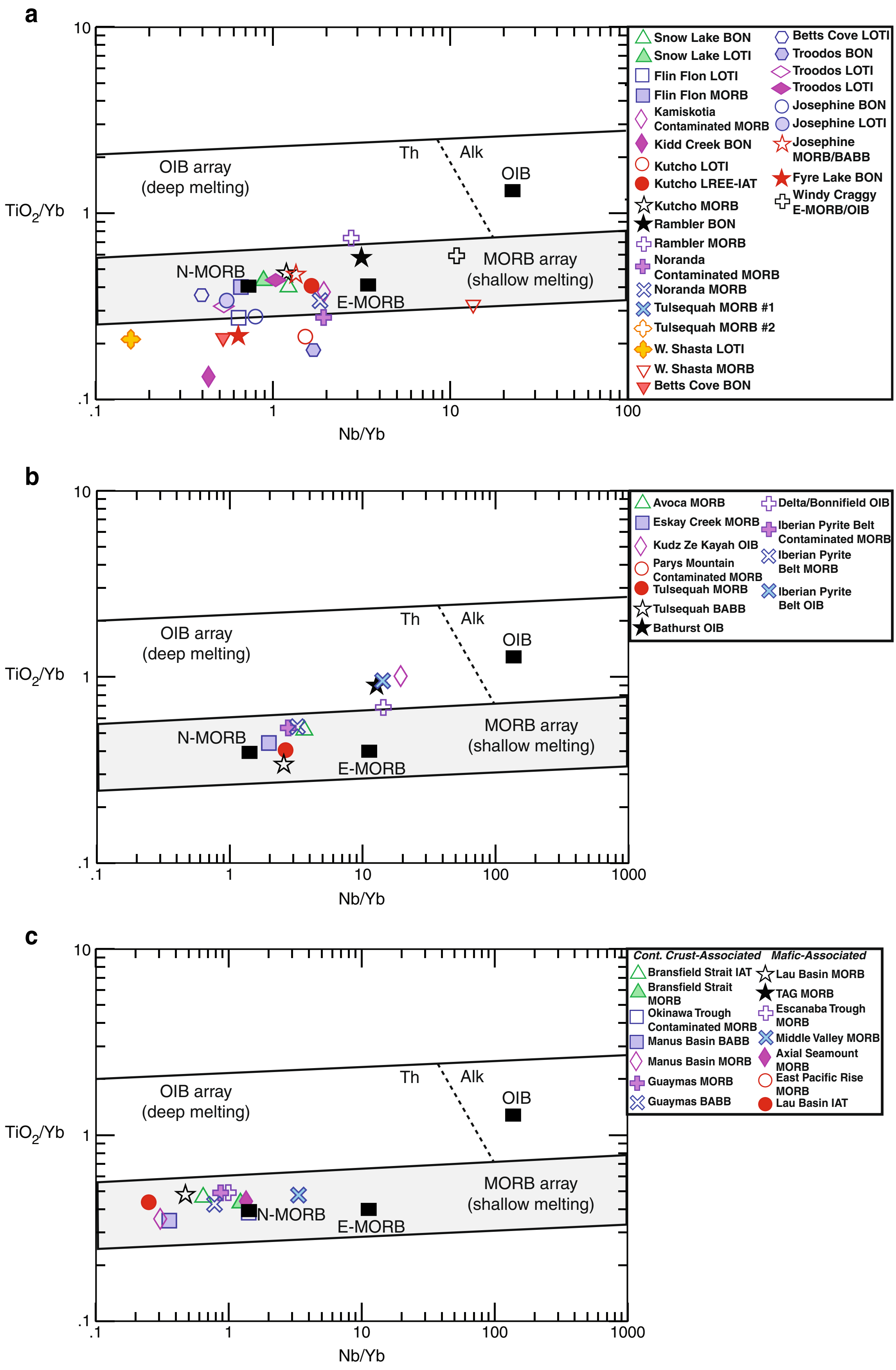
Fig. 11 Chondrite-normalized REE plots for rhyolites from post-Archean mafic-dominated (primitive) VMS environments: **a** VMS-barren rhyolites, **b** VMS-bearing rhyolites that are not peralkaline; and **c** peralkaline VMS-bearing rhyolites. Data sources in the Appendix. Symbols as in Fig. 10. Normalization values from Sun and McDonough (1989)

sional but restricted geodynamic environment (e.g., Lentz 1998; Piercey 2007). Identification of such features can outline environments with geodynamic and petrological attributes that may have had sufficient heat flow and crustal permeability to initiate, drive, and sustain hydrothermal circulation robust enough to generate metal-rich fluids and form VMS mineralization.

Extension and extension-related magmatism—the key to generating VMS-related hydrothermal activity

Throughout geological time VMS deposits are associated with high-temperature magmatism within well-defined thermal corridors (see above). As well, VMS deposits are preferentially associated with extensional geodynamic settings, such as mid-ocean ridges, back-arc basins, and intra-arc rifts (continental and/or intra-oceanic arcs) and magmatic products indicative of extensional activity (e.g., Sillitoe 1982; Cathles et al. 1983; Swinden 1991; Barrie et al. 1993; Lentz 1998; Prior et al. 1999; Syme et al. 1999; Piercey 2007). For example, in the modern oceans actively forming VMS deposits are associated with mid-ocean ridges, back-arc basins, and intra-arc rifts (e.g., Hannington et al. 2005, and references therein). Similar settings are observed for ancient VMS deposits, with the exception of those associated with boninite-LOTI assemblages. Boninite-LOIT assemblages are interpreted to represent extension associated with subduction initiation within a forearc, back-arc initiation, or plume-arc interaction (Stern and Bloomer 1992; Wyman et al. 1999; Piercey et al. 2001a); the boninite-LOTI association with VMS deposits has not been found in modern geodynamic environments; this may be because forearc environments in the modern record are covered by forearc sediments or have been overprinted by younger magmatic arcs (Galley, personal communication).

Rifting results in numerous features that are critical for hydrothermal system generation: (1) it increases permeability for fluid flow, providing the fluid conduits for hydrothermal flow; (2) allows for accommodation space in the crust and creates basins that provides physical and chemical traps for massive sulfide deposits; and (3) accommodation space in the crust also provides a location for upwelling mantle melts to occupy, thereby allowing the rift to have an elevated geothermal gradient. Extensional faulting of the crust associated with rifting results in abundant normal faults (e.g., synvolcanic and synsedimentary faults and it is these faults that provide the conduits that focus fluid flow. For example, numerous workers have shown via heat and fluid flow modeling that extensional faults within basins are the main controllers of upwelling hydrothermal fluids (e.g., Cathles 1981; Barrie et al. 1999a, b; Schardt et al. 2005, 2006; Yang 2006). Rifting also results in the formation of grabens, and in some cases calderas, whose bounding faults commonly control the localization of hydrothermal fluid flow and massive sulfide mineralization (e.g., Gibson and Watkinson 1990; Allen 1992; Kerr and Gibson 1993; Galley et al. 1995; Allen et al.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TiO2/Yb
Nb/Yb
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.1
1
10
100
OIB array
(deep melting)
Th
Alk
OIB
N-MORB
E-MORB
MORB array
(shallow melting)
Snow Lake BON
Snow Lake LOTI
Flin Flon LOTI
Flin Flon MORB
Kamiskotia Contaminated MORB
Kidd Creek BON
Kutcho LOTI
Kutcho LREE-IAT
Kutcho MORB
Rambler BON
Rambler MORB
Noranda Contaminated MORB
Noranda MORB
Tulsequah MORB #1
Tulsequah MORB #2
W. Shasta LOTI
W. Shasta MORB
Betts Cove BON
Betts Cove LOTI
Troodos BON
Troodos LOTI
Troodos LOTI
Josephine BON
Josephine LOTI
Josephine MORB/BABB
Fyre Lake BON
Windy Craggy E-MORB/OIB
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(shallow melting)
Avoca MORB
Eskay Creek MORB
Kudz Ze Kayah OIB
Parys Mountain Contaminated MORB
Tulsequah MORB
Tulsequah BABB
Bathurst OIB
Delta/Bonnifield OIB
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(shallow melting)
Cont. Crust-Associated
Mafic-Associated
Bransfield Strait IAT
Bransfield Strait MORB
Okinawa Trough Contaminated MORB
Manus Basin BABB
Manus Basin MORB
Guaymas MORB
Guaymas BABB
Lau Basin MORB
TAG MORB
Escanaba Trough MORB
Middle Valley MORB
Axial Seamount MORB
East Pacific Rise MORB
Lau Basin IAT

◀**Fig. 12** TiO$_2$/Yb-Nb/Yb plot (from Pearce 2008) for mafic rocks from: **a** mafic-dominated (primitive) VMS environments; **b** felsic-dominated (evolved) VMS environments; and **c** modern VMS environments. Notably, most mafic rocks from VMS environments are associated with shallow-level melting. Further details are given in the text. Data sources in the Appendix

1996; Gibson et al. 1999; Stix et al. 2003; Gibson 2005). Furthermore, the bathymetry and the presence of sedimentary/volcanic infill within seafloor volcanic depressions or grabens also enhances the preservation of deposits by preventing massive sulfide mass wasting and erosion, and enhances the potential for subseafloor replacement in permeable sediment- and volcaniclastic-filled grabens and calderas, a key factor in forming many large VMS deposits (e.g., Gibson and Watkinson 1990; Doyle and Allen 2003; Winter et al. 2004).

The accommodation space created during rifting and extension helps promote heat transfer from depths within the Earth to the subvolcanic environment. Shown on Fig. 14 are various isothermal profiles and heat flow distributions for different geological environments. For non-rift settings, like subduction zones (Fig. 14a) and normal oceanic crust (Fig. 14b), there are depressed geotherms below the arc in a subduction zone due to cooling of the mantle wedge by the subducted slab (e.g., Tatsumi and Eggins 1995), and relatively flat geotherms increasing with depth in normal ocean crust distal from the ridge axis (off rift location in Fig. 14b). In both cases, the ambient geothermal gradient is low or depressed (i.e., low heat flow).

In contrast, in the ridge axis location in mid-ocean ridges (Fig. 14b is also a proxy for forearcs, rifted arcs, and back-arc basins) the upwelling of hot, asthenospheric mantle beneath

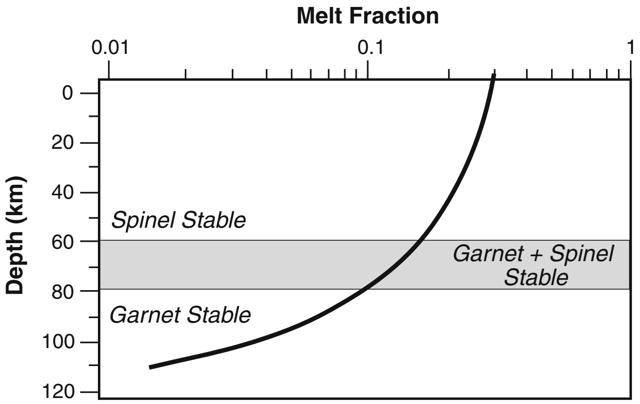

Fig. 13 Plot of percentage of melt fraction as a function of depth of melt generation within the mantle. Notably, at shallow levels of melting within the mantle there is a greater abundance of melt generated. With greater volumes of melting, there is greater ambient heat flow in the geodynamic environment and this explains why most VMS belts are associated with mafic rocks indicative of shallow-level mantle melting. Diagram modified from McKenzie and O'Nions (1991) and Ellam (1992)

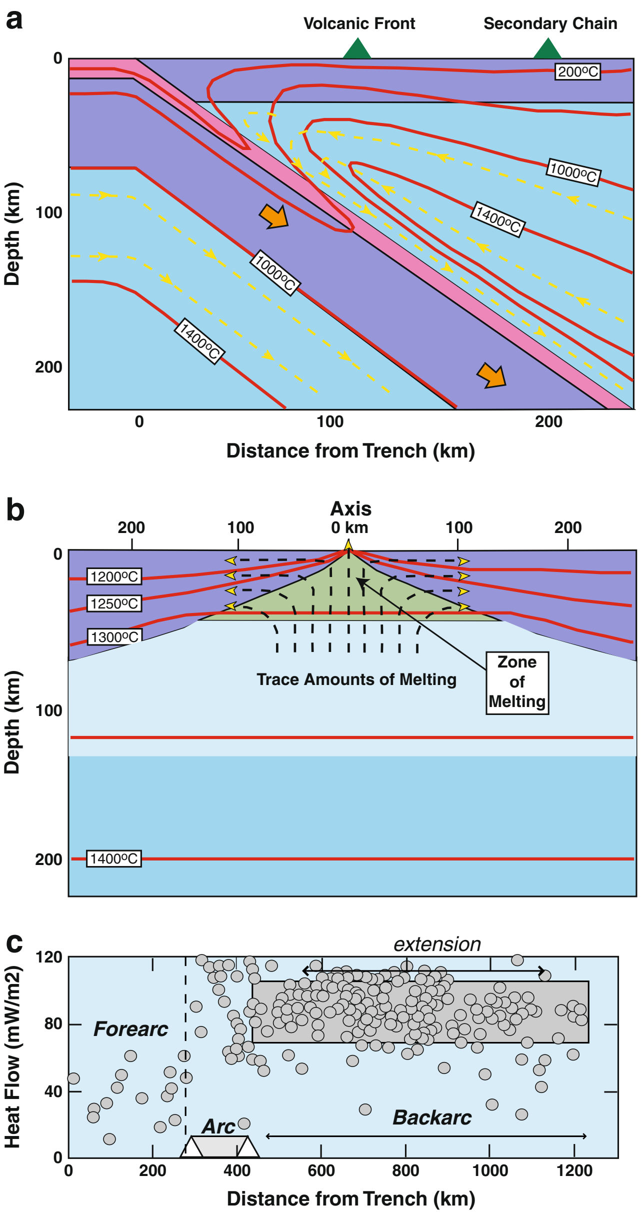

Fig. 14 **a** Cross section of a typical arc environment showing the pronounced downward warping of the geotherms. These environments are generally cooler than rift environments and this is why many VMS deposits do not occur in "arcs" sensu stricto, but rather form during later episodes of rifting within the arc. Diagram from Winter (2001). **c** Cross section of a typical mid-ocean ridge (this profile is also similar to a back-arc spreading center). As in rift environments, these spreading centers have upwelling mantle beneath the ridge and an upward warping of the geothermal gradient. This elevated heat flow due to mantle upwelling is responsible for driving hydrothermal circulation in these environments and explains the association of VMS deposits with ridges and mature (spreading) back-arc basins. Diagram from McKenzie and Bickle (1988). **c** Heat flow profile across the northeast Japan arc into the Japan sea back-arc region. Notably, the back-arc region is characterized by extension and elevated heat flow relative to the arc. These types of environments provide the key ingredients for VMS deposit formation: extension to form the permeability required for fluid flow and elevated heat flow to drive hydrothermal circulation. Diagram modified from Currie and Hyndman (2006)

the rift results in thinning of the crust and an upward warping of the geotherms immediately beneath rift (Fig. 14b). Furthermore, heat flow profiles for rift settings, such as back-arc basins (e.g., Hyndman et al. 2005; Currie and Hyndman 2006), show elevated heat flow compared with arcs and background crust (e.g., Fig. 14c). Thus, extensional activity allows deeper, mantle heat to be rapidly transferred to the subvolcanic environment and creates environments with greater potential to drive vigorous and long-lived hydrothermal circulation, and increases the potential to form VMS deposits.

Shallow subvolcanic intrusions: a passive product or an active participant in the VMS-hydrothermal system?

Considerable research has implicated shallow (i.e., 1–3 km below surface) subvolcanic intrusive complexes as drivers of hydrothermal circulation in VMS environments. Voluminous intrusions underlie some VMS districts, in some cases cross-cutting early formed deposits (e.g., Flin Flon and Noranda), and these typically have petrogenetic histories and lithogeochemical characteristics identical to the volcanic rocks hosting mineralization above the complexes (Campbell et al. 1981; Galley 1996, 2003). Many workers have suggested that these intrusive complexes are the drivers of hydrothermal circulation and some of the alteration within the VMS environment (Campbell et al. 1981; Cathles 1983; Galley 1996, 2003; Large et al. 1996; Brauhart et al. 1998; Barrie et al. 1999a). Many VMS environments, however, do not have associated subvolcanic intrusive complexes, including many large to giant VMS deposits and districts (e.g., Kidd Creek, Tambogrande, and Windy Craggy), or have intrusions at stratigraphic levels far below and not in the immediate vicinity of the deposits (e.g., Bathurst) (Whalen et al. 1998). This may be a function of insufficient exposure of footwall stratigraphy, or due to structural complexity and loss of stratigraphy in younger accretionary orogens. An alternative explanation is that the presence of high-level (i.e., within 1–3 km of the surface) subvolcanic intrusions is not critical to the formation of all VMS deposits.

Evaluating the importance of subvolcanic intrusive complexes is also important because in many VMS systems associated with subvolcanic intrusive complexes, the most voluminous phases of magmatism post-date VMS mineralization and associated alteration (e.g., Goldie 1978; Galley et al. 2000; Galley 2003). Similarly, recent work on the emplacement of intrusive complexes, based on detailed mapping and U–Pb geochronology, has suggested that some intrusive systems were likely emplaced in a series of small increments from small magma chambers, rather than as large plutons or batholiths generated from a single, large magma chamber (Coleman et al. 2004; Glazner et al. 2004; Whalen et al. 2004). This raises some doubt of the validity of models that assume a single, shallow-level intrusion as the driver of VMS-hydrothermal convection.

This does not, however, negate models of deeper magma emplacement (i.e., ~10 km) in the mid-crust as potential drivers of regional-scale hydrothermal circulation. For example, Barrie et al. (1999a) demonstrated that a deep, hot, ultramafic sill could generate sufficient hydrothermal fluid flow, with sufficient permeability and focusing of fluids to account for the large-tonnage Kidd Creek massive sulfide deposit. Similar models have been advocated for other VMS systems (e.g., Barrie et al. 1999b, and references therein). The deep-seated sill model for hydrothermal convection is also consistent with the upwelling of mantle-derived magmas beneath rifts as the main driver of hydrothermal circulation in the VMS environment, as suggested above. Furthermore, deeper mafic magma upwelling might also explain why in many VMS camps (e.g., Tambogrande, Kidd Creek, and Windy Craggy) there are no subvolcanic intrusive complexes, yet they have large massive sulfide deposits (i.e., the magmas ponded at deeper levels in the crust and never reached the shallow subvolcanic environment).

Shallow intrusions may not necessarily be a requisite heat source for driving hydrothermal circulation, but they are important proxies for the heat flow history of an area (e.g., thermal corridors of Galley 2003). Most of these high-level intrusive complexes have protracted magmatic histories, typically extending from the pre- to post-VMS formation stages (e.g., Galley 2003; Whalen et al. 2004). For example, in the Bieldeman Bay subvolcanic intrusive complex in the Sturgeon Lake VMS camp, magmatism extends over 14 Ma occurring syn- and post-VMS deposit formation (Galley et al. 2000). In the Flavrian complex in the Noranda VMS camp, syn-VMS intrusive phases are ~2,700 Ma, whereas post-VMS phases that host Cu–Mo occurrences are ~2,697 Ma, implying a minimum of 3 Ma of magmatic activity (Galley and van Breemen 2002). Similarly, in the Wolverine VMS deposit, early intrusive phases in the footwall to the deposit formed at ~352 Ma, whereas the syn-VMS intrusions were emplaced at ~347 Ma, with a minimum of 5 Ma of magmatic activity before VMS mineralization (Piercey et al. 2008). The extended magmatic activity in all of these intrusive complexes indicates that there was magmatic heat (i.e., ponded basalt) resident at deeper crustal levels for protracted periods of time during the formation of these VMS environments. It is this ponded basalt that is responsible for the long-term elevated geothermal gradient of a VMS environment and is likely the major cause for sustained, long-term hydrothermal circulation.

In addition to providing heat, subvolcanic intrusive complexes have been suggested to be potential contributors of metals to VMS-hydrothermal systems. Melt inclusion work in VMS-associated rocks, largely from modern systems, has lead many workers to that a shallow, degassing magma chamber (i.e., subvolcanic intrusive complex) likely plays a role in adding some metals to the VMS-hydrothermal system (e.g., Yang and Scott 1996; Kamenetsky et al. 2001, 2002; Yang and Scott 2002, 2005; Beaudoin et al. 2007). Elevated Sn and In contents in massive sulfides from Bathurst, Neves Corvo, and Kidd Creek have been suggested by some workers to be magmatic in origin and transported as volatiles from magma degassing (Boyle 1997; Hannington et al. 1999; Goodfellow and McCutcheon 2003; Relvas et al. 2006a, b). A contribution of metals transported by magmatic volatiles has been suggested in the genesis of many Au-rich VMS systems, as evidenced by the presence of aluminous alteration assemblages, precious metal (Au–Ag) enrichment, saline and magmatic halogen-bearing fluid inclusions, and epithermal suite of elements (e.g., Hg, Bi, Sb, and Ba) in the mineralization and alteration zones (Sillitoe et al. 1996; Hannington et al. 1999; Roth et al. 1999; Sherlock et al. 1999; Dubé et al. 2007; Mercier-Langevin et al. 2007). These data indicate that shallow-level (i.e., <1–3 km below the surface) subvolcanic intrusions actively contribute metals, fluids, and volatiles to some VMS systems.

This ultimately returns to the question of an active or passive role for subvolcanic intrusions in VMS deposit genesis. Many shallow-level intrusions (i.e., less than 3–4 km depth) are likely not major contributors to the heat budget of VMS systems, but are the manifestation of deeper heat (i.e., ~10 km depth) that drives hydrothermal circulation on a larger, geodynamic scale. Shallow-level intrusions, however, may play a role in contributing metals, fluids, and volatiles to VMS-hydrothermal systems, particularly in precious metal-rich VMS systems.

Conclusions

VMS deposits are spatially associated with specific suites of magmatic rocks that provide key information about the tectonic setting and magmatic history of the host rocks to VMS systems and provide critical information on the role that tectonics and petrology play in VMS deposit genesis. The specific assemblages of magmatic rocks associated with VMS deposits vary as a function of tectonic setting and geological environment of VMS formation, with different VMS groups having specific "petrochemical assemblages" indicative of a given geodynamic environment of formation. In all petrochemical assemblages and VMS environments there is the common association high-temperature magmatic activity and extensional geodynamic activity. The high-temperature magmatic activity in VMS environments is related to mantle upwelling beneath rifts in extensional geodynamic environments (e.g., mid-ocean ridges, back-arc basins, and intra-arc rifts). The underplated magmas provide the heat to drive hydrothermal circulation and extension results in the formation of extensional faults (e.g., synvolcanic faults) that create the permeability and porosity required for recharge and discharge of VMS-hydrothermal fluids. Extensional geodynamic settings are also critical in providing environments conductive to the preservation of VMS deposits via shielding of massive sulfides from seafloor weathering and mass wasting and/or by creating conditions that favor the precipitation of subsea-floor replacement-style mineralization in sedimented rifts.

Subvolcanic intrusions are also products of the elevated heat flow regime common to VMS environments occurring at various depths within the crust. Long-lived shallow (<1–3 km from surface) subvolcanic intrusive complexes are important elements in many VMS camps. While shallow intrusions may not be the main drivers of hydrothermal circulation, they are the record of the deeper-seated heat (i.e., basaltic underplating at ~3–10 km depth) that is the likely cause of hydrothermal circulation. The occurrence of shallow intrusive complexes is an excellent indicator of a potentially prospective VMS environment. Numerous VMS camps lack intrusive complexes, however, and the absence of a subvolcanic intrusive complex does not necessarily indicate an environment of poor VMS potential. Shallow intrusive complexes may also be important contributors of metals, fluids, and volatiles to the VMS-hydrothermal system.

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Appendix: sources of lithogeochemical data

Mafic rocks

Mafic-dominated VMS settings

Snow Lake and Flin Flon, Stern et al. (1995); Kamiskotia, Hocker et al. (2005); Kidd Creek, Kerrich et al. (1998) and Wyman et al. (1999); Kutcho, Barrett et al. (1996); Rambler/Ming, Piercey et al. (1997) and Bailey (2002);

Blake River Group (Noranda), Lafleche et al. (1992a, b); West Shasta, Brouxel et al. (1988), Bence and Taylor (1985), and Lapierre et al. (1985); Betts Cove: Bedard (1999); Troodos, Cameron (1985) and Rogers et al. (1989); Ice Deposit, Piercey (unpublished data); Josephine (Turner Albright), Harper (2003); Fyre Lake, Piercey et al. (2001a, 2004); and Windy Craggy, Peter and Scott (1999).

Modern VMS environments

Bransfield Strait, Keller et al. (2002); Okinawa Trough, Shinjo et al. (1999); Manus Basin, Sinton et al. (2003); Juan de Fuca (Axial Seamount), Rhodes et al. (1990); East Pacific Rise, Allan et al. (1987); Middle Valley, Stakes and Franklin (1994); Lau Basin, Ewart et al. (1994); TAG hydrothermal field (mid-Atlantic), Smith and Humphris (1998); Escanaba Trough: Saunders et al. (1982); Guaymas, Davis and Clague (1987).

Continental crust-associated VMS settings

Avoca, Leat et al. (1986) and McConnell et al. (1991); Eskay Creek, Barrett and Sherlock (1996); Kudz Ze Kayah (Finlayson Lake), Piercey et al. (2002a); Parys Mountain, Barrett et al. (2001); Tulsequah: Sebert and Barrett (1996); Bathurst, Rogers and van Staal (2003); Delta-Bonnifield, Dusel-Bacon et al. (2004); Iberian Pyrite Belt, Almodóvar et al. (1997) and Mitjavila et al. (1997).

Felsic rocks

Archean felsic rocks

Pilbara, Vearncombe and Kerrich (1999); Kidd Creek, Prior et al. (1999); Sturgeon Lake, Lesher et al. (1986); Noranda, Lesher et al. (1986) and Péloquin (1999)(regional); South Bay, Lesher et al. (1986); Kamiskotia, Hart (1984), Barrie and Pattison (1999); and High Lake, Petch (2004).

Post-Archean felsic rocks from mafic-dominated settings

Flin Flon, Syme (1998); Rambler (Ming), Bailey (2002) and Piercey et al. (1997); West Shasta, Bence and Taylor (1985) and Lapierre et al. (1985); Kutcho, Barrett et al. (1996); and Snow Lake, Bailes and Galley (1999, 2001).

Post-Archean felsic rocks from continental crust-dominated settings

Eskay Creek, Barrett and Sherlock (1996); Delta-Bonnifield, Dusel-Bacon et al. (2004); Finlayson Lake, Piercey et al. (2001b); Iberian Pyrite Belt, Almodóvar et al. (1997); Bransfield Strait, Petersen et al. (2004); Okinawa Trough, Shinjo and Kato (2000); Mount Read, Crawford et al. (1992); Parys Mountain, Barrett et al. (2001); Avoca, Leat et al. (1986) and McConnell et al. (1991) and Bathurst, Rogers et al. (2003).

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