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Subsea-floor replacement in volcanic-hosted massive sulfide deposits

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Abstract

Recent research on volcanic-hosted massive sulfide (VMS) deposits indicates that syngenetic subsea-floor replacement ores form an important component of many deposits. In the context of VMS deposits, subsea-floor replacement can be defined as the syn-volcanic formation of sulfide minerals within pre-existing volcanic or sedimentary deposits by infiltration and precipitation in open spaces (fractures, inter- and intra-granular porosity) as well as replacement of solid materials.

There are five criteria for distinguishing subsea-floor replacement in massive sulfide deposits: (1) mineralized intervals are enclosed within rapidly emplaced volcanic or sedimentary facies (lavas, intrusions, subaqueous mass-flow deposits, pyroclastic fallout); (2) relics of the host facies occur within the mineral deposit; (3) replacement fronts occur between the mineral deposit and the host lithofacies; (4) the mineral deposit is discordant to bedding; and (5) strong hydrothermal alteration continues into the hanging wall without an abrupt break in intensity. Criteria 1–3 are diagnostic of replacement, whereas criteria 4 and 5 may suggest replacement but are not alone diagnostic. Because clastic sulfide ores contain accessory rock fragments collected by the parent sediment gravity flow(s) during transport, criteria 2 can only be applied to massive, semi-massive, disseminated or vein style deposits, and not clastic ores.

The spectrum of VMS deposit types includes deposits that have accumulated largely subsea-floor, and others in which sedimentation and volcanism were synchronous with hydrothermal activity, and precipitation of sulfides occurred at and below the sea floor over the life of the hydrothermal system. Deposits that formed largely subsea-floor are mainly hosted by syn-eruptive or post-eruptive volcaniclastic facies (gravity flow deposits, water-settled fall, autoclastic breccia). However, some subsea-floor replacement VMS deposits are hosted by lavas and syn-volcanic intrusions (sills, domes, cryptodomes). Burial of sea-floor massive sulfide by lavas or sediment gravity flow deposits can interrupt sea-floor mineralization and promote subsea-floor replacement and zone-refining.

The distance below the sea floor at which infiltration and replacement took place is rarely well constrained, with published estimates ranging from less than 1 to more than 500 m, but mainly in the range 10–200 m. The upper few tens to hundreds of metres in the volcano-sedimentary pile are the favoured position for replacement, as clastic facies are wet, porous and poorly consolidated in this zone, and at greater depths become progressively more compacted, dewatered, altered, and less amenable to large scale infiltration and replacement by hydrothermal fluids. Furthermore, sustained mixing between the upwelling hydrothermal fluid and cold seawater is regarded as a major cause of sulfide precipitation in VMS systems, and this mixing process generally becomes less effective with increasing depth in the volcanic pile.

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The relative importance of subsea-floor replacement in VMS systems is related principally to four factors: the permeability and porosity patterns of host lithofacies, sedimentation rate, the relative ease of replacement of host lithofacies (especially glassy materials) and early formed alteration minerals during hydrothermal attack, and physiochemical characteristics of the hydrothermal fluid.

Keywords: Volcanic-hosted massive sulfide; Subsea-floor replacement; Sea-floor accumulation; Criteria; Subaqueous volcanism; Genetic models

1. Introduction

The literature on volcanic-hosted massive sulfide (VMS) deposits emphasizes formation on the sea floor by accumulation of sulfides precipitated from exhaling hydrothermal fluids, or from endogenic growth of a sulfide mound on the sea floor (e.g., Solomon and Walshe, 1979; Ohmoto and Skinner, 1983; Lydon, 1988; Large, 1992). Support for a sea floor exhalative origin initially came from studies of well-preserved ancient massive sulfide deposits (e.g., Ohashi, 1920; Solomon, 1976; Franklin et al., 1981; Lydon and Galley, 1986). The discovery of sulfide chimney and mound deposits on the modern sea floor gave credence to the hydrothermal exhalative model, but raised questions as to the viability of mineral precipitation from the exhaled fluids. Sulfide accumulation from buoyant hydrothermal plumes above black smokers is a highly inefficient process. It has been estimated that greater than 99% of the metal transported by venting hydrothermal fluids is dispersed in the water column by the plume and incorporated into distal sediments (Rona, 1984). Dense, saline hydrothermal fluids that form a bottom-hugging brine pool upon exhalation are required for a significant sulfide deposit to result from exhalation (Pottorf and Barnes, 1983; Solomon and Khin Zaw, 1997). Studies of the structure of modern sulfide chimneys and mounds (e.g., Goldfarb et al., 1983; Koski et al., 1984; Paradis et al., 1988), and the texture and metal zonation in ancient deposits (e.g., Large, 1977; Eldridge et al., 1983; Lydon, 1984a,b; Lydon and Galley, 1986), have highlighted the importance of sulfide accumulation by open space filling and replacement within the sulfide mound. There is increasing evidence to suggest that these processes may extend into the subsea-floor environment, and that some VMS deposits form largely by replacement of subsea-floor volcanic and sedimentary deposits. Subsea-floor replacement may be an efficient process for trapping a higher percentage of the total metal budget and hence may determine grade and tonnage.

Separate lenses or segments of a single deposit can form by different processes and both sea-floor massive sulfide accumulation and subsea-floor replacement may be involved in constructing a single deposit (Sangster, 1972; Large, 1977; Solomon and Walshe, 1979; Eldridge et al., 1983; Kuroda, 1983; Vivallo, 1985; Zierenberg et al., 1988; Galley et al., 1993; Humphris et al., 1995). Stringer zones beneath VMS deposits also provide clear evidence for precipitation of sulfide below the sea floor (Sangster, 1972; Franklin et al., 1981; Lydon, 1984a,b). However, in most cases this part of the deposit comprises veins and disseminations rather than massive replacements. The exceptions are sulfide lenses that form by infiltration and replacement of the more permeable zones, such as fine hydraulic-breccias, within the stockwork system (e.g., Galley and Koski, 1999; Tornos, 2000).

There are few detailed descriptions of large-scale subsea-floor replacement deposits in modern and ancient volcanic successions. Young subsea-floor sulfide accumulations have been mapped at Middle Valley (Goodfellow and Blaise, 1988; Goodfellow and Franklin, 1993) and on the Panarea platform, Aeolian volcanic arc, Italy (Marani et al., 1997; Gamberi et al., 1999). However, the best-studied examples of VMS deposits with subsea-floor replacement ores are confined to ancient volcanic successions. Published examples include deposits hosted by felsic to intermediate volcanic successions, mafic successions, and mixed felsic volcanic-limestone successions. However, only in a few cases are the replacement ores and ore–host rock relationships described in detail. Deposits hosted by felsic to intermediate successions include the Matsuki deposit and some lenses of the Hanaoka-Shakanai cluster, Japan (Kuroda, 1983; Ohtagaki et al., 1974), the Rosebery, Hercules, Mount Lyell, Benambra, Liontown, Highway-Reward, Mount Morgan, Sulfur Springs and Golden Grove deposits, Australia (Taube, 1986; Allen and Hunns, 1990; Allen, 1992, 1994a; Berry et al., 1992; Khin Zaw and Large, 1992; Doyle and McPhie, 1994; Bodon and Valenta, 1995; Morant, 1995, 1998; Messenger et al., 1997, 1998; Doyle and Huston, 1999; Sharpe and Gemmell, 2000, 2002; Corbett, 2001; Ulrich et al., 2002), the Mattabi (and other Sturgeon Lake deposits), Horne H lenses, Ansil, Kidd Creek, Coniagas, and some lenses of the Myra Falls deposits, Canada (Kerr and Mason, 1990; Morton et al., 1991; Kerr and Gibson, 1993; Doucet et al., 1994; Galley et al., 1995; Barrie et al., 1999; Hannington et al., 1999; Sinclair et al., 2000), the Renström, Kyrkvägen, Kankberg, Holmtjärn, Petiknäs North, Långdal, Långsele, Boliden and West, North and East Maurliden deposits, Sweden (Allen et al., 1996b; Bergman Weihed et al., 1996; Montelius et al., 2000), and the San Miguel, Salomon-Lago (Riotinto), San Platón, Concepción, Neves Corvo and Los Frailes-Aznalcóllar deposits of the Iberian Pyrite Belt, Portugal and Spain (Sáez et al., 1996; Almodóvar et al., 1998; Relvas et al., 2000; Tornos, 2000; Allen, 2001). Deposits hosted by mafic volcanic successions include the Turner-Albright deposits, U.S.A. (Zierenberg et al., 1988) and the Potter mine, Canada (Gibson and Gamble, 2000). The Garpenberg and Garpenberg Norra deposits, Sweden (Allen et al., 1996a), the Chisel Lake, North Chisel, Ghost Lake, Lost Lake, Errington and Vermilion deposits, Canada (Galley et al., 1993; Gray and Gibson, 1993; Stoness et al., 1993; Bailes and Galley, 1996; Galley and Ames, 1998), the Lynne deposit, North America (DeMatties, 1994), and the Henty-Mount Julia deposits, Australia (Halley and Roberts, 1997; Callaghan, 2001) are replacement-style VMS deposits in limestone and felsic volcanic rocks.

The extent of subsea-floor replacement in massive sulfide systems is only starting to be widely recognized, and no synthesis of the characteristics, different styles and criteria for recognition of subsea-floor replacement deposits has previously been presented. This paper focuses on distinguishing sea-floor and subsea-floor replacement-style VMS deposits, with the aim of constraining the diagnostic evidence for subsea-floor replacement and reviewing the role of subsea-floor replacement in the genesis of VMS deposits. Models for sulfide accumulation during the evolution of contrasting host successions are presented on the basis of published descriptions and our own studies.

1.1. Terminology

The term ''volcanic-hosted massive sulfide deposit'' is used for syngenetic accumulations of massive sulfide that are hosted by submarine volcanic successions (Solomon, 1976; Franklin et al., 1981; Lydon, 1984a; Large, 1992). In the context of VMS deposits, the term ''exhalation'' has been used to describe fluid emanations into the sea (or a brine pool) from the sea floor (Franklin et al., 1981; Solomon and Khin Zaw, 1997). Most primary textures in mound-style deposits are indicative of sulfide replacement and infilling of pore space, rather than precipitation from hydrothermal fluids that exhaled at the seawater/sea-floor interface (Barton, 1978; Eldridge et al., 1983; Lydon and Galley, 1986). For this reason, the term ''sea-floor massive sulfide'' is used here for the general case of all massive sulfides formed at the sea floor, including mound-style deposits, and the term ''exhalative'' is reserved for the specific case in which sulfide ores precipitated from hydrothermal fluid that exhaled into the sea (or a brine pool).

The position of the sea floor may change through the life of the hydrothermal system due to mound growth, sedimentation, volcanism, oxidation of sulfides and/or erosion. The term ''sea-floor massive sulfide accumulation'' refers to sulfide deposition on the sea floor, at one or successive sea-floor positions. These sea-floor positions can be identified as contacts between successive sedimentation units.

Replacement is defined as a ''change in composition of a mineral or mineral aggregate, presumably accomplished by diffusion of new material in and old material out, without breakdown of the solid state'' (Bates and Jackson, 1987). In the context of VMS deposits, subsea-floor replacement refers to the precipitation, from syn-volcanic hydrothermal solutions, of ore-forming minerals within pre-existing volcanic or sedimentary deposits. Replacement is synchronous with volcanism, sedimentation, and diagenesis in adjacent strata (herein referred to as syn-volcanic and syn-diagenetic replacement). The host deposits may be unconsolidated or lithified at the time of replacement. Subsea-floor replacement in most cases probably includes components of infiltration and precipitation in open spaces (fractures, inter- and intra-particle porosity) as well as replacement of solid materials. No restriction of depth beneath the sea floor is placed on the term. The term does not refer specifically to the process of zone-refining (Eldridge et al., 1983; Large et al., 1989) within the developing deposit, although this may be important during subsea-floor replacement.

2. Discriminating subsea-floor replacement from post-diagenetic replacement

Although obvious in some young VMS deposits, in ancient deposits it cannot be assumed that replacement occurred "subsea-floor" within the period of volcanism, sedimentation and diagenesis. Some massive sulfide deposits hosted in volcanic and sedimentary successions are attributed to later post-diagenetic, syn-tectonic replacement, although there is debate as to whether original syngenetic mineralization has been redistributed by later tectonic processes (Aerden, 1991, 1993; Perkins, 1984). For replacement ore deposits that show no clear syn-sedimentary features in their upper parts (e.g., clastic sulfides), the syn-volcanic or syn-diagenetic timing of replacement must be demonstrated via careful documentation of overprinting relationships among the textures, structures and mineral assemblages of hydrothermal alteration, diagenetic alteration, diagenetic compaction, tectonic deformation and metamorphism, and by consideration of the stratigraphic and structural setting of the ore deposit (Allen, 1994a; Doyle and Huston, 1999; Gifkins and Allen, 2001). Evidence suggestive of a syn-volcanic or syn-diagenetic timing for ore formation may include, but is not limited to: (1) an asymmetric alteration halo with strongest alteration in the stratigraphic footwall; (2) competent hydrothermal alteration assemblages (e.g., quartz- or feldspar-rich assemblages) contain relic uncompacted vitriclasts, whereas those in enclosing zones of less competent hydrothermal and diagenetic alteration assemblages (clays, phyllosilicates) have been compacted during diagenesis; (3) syn-volcanic intrusions that cut the ore deposit or intensely altered rocks are relatively unaltered; (4) the mineral deposit and alteration halo are overprinted by (i.e., pre-date) the first tectonic deformation observed in the enclosing host rocks; and (5) there is a strong stratigraphic and facies control on mineralization. The data and results presented in this paper are only relevant to ore deposits where a syn-volcanic or syn-diagenetic timing have been demonstrated.

3. Discriminating subsea-floor replacement from sea-floor accumulation

Our own studies and a review of the literature indicate that there are several criteria for distinguishing subsea-floor replacement in massive sulfide deposits:

(1) relics of the host rock (sedimentary, volcaniclastic, coherent volcanic facies) within the sulfide deposit,
(2) facies characteristics indicating very rapid emplacement of the host lithofacies,
(3) identification of replacement fronts between the sulfide deposit and host deposit,
(4) discordance with the enclosing host lithofacies,
(5) the presence of strong hanging wall alteration, similar in style and intensity to footwall alteration.

Criteria 1–3 are diagnostic of replacement, whereas criteria 4 and 5 may suggest replacement but are not alone diagnostic.

Criteria for distinguishing sea-floor VMS ores include:

(1) sedimentary clastic sulfide textures,
(2) sulfide chimney textures,
(3) exhalite at the ore horizon,
(4) fossil tube worms and bivalves,
(5) facies characteristics indicating slow accumulation rate of the host rocks, and/or occurrence of VMS ores between units of rapid emplacement rate,
(6) an asymmetric alteration pattern of strong footwall alteration and weaker hanging wall alteration.

Criteria 1–4 are diagnostic of sea-floor sulfide formation, whereas criteria 5 and 6 suggest sea-floor deposition but are not diagnostic.

These criteria for distinguishing sea-floor and subsea-floor massive sulfide deposition are discussed below.

3.1. Bedforms, sulfide clasts, rock fragments and host rock relics (pseudo-fragments)

The presence of sedimentary structures and clastic textures within some massive sulfide deposits has provided critical evidence in support of a sea-floor origin for VMS deposits. Graded bedding, cross bedding, soft-sediment deformation structures, and intercalations of laminated and fragmental ore have been identified in many Kuroko deposits (Kajiwara, 1970; Lee et al., 1974; Ishikawa and Yanagisawa, 1974; Ito et al., 1974; Kuroda, 1983), some Mesozoic–Palaeozoic deposits (e.g., Mt. Chalmers: Large and Both, 1980; Mathiati: Lydon, 1984b; Woodlawn: McKay and Hazeldene, 1987; Buttle Lake: Robinson et al., 1996; Filon Norte: Tornos et al., 1998; Eskay Creek: Roth et al., 1999; Sherlock et al., 1999; Hayl-as-Safil: Galley and Koski, 1999) and Archaean deposits (e.g., Vauze: Spence, 1975; Kidd Creek: Hannington et al., 1999). Mixtures of massive sulfide clasts and exotic rock fragments also occur. Rock fragments may be incorporated into clastic sulfide aggregates at source or be picked up during sedimentary transport via erosion of the sea floor by sulfide clast-rich gravity flows (e.g., Buchans: Binney, 1987; some Kuroko ores: Eldridge et al., 1983; Mathiati: Lydon, 1984b; Corbet: Gibson et al., 1993). At the other end of the spectrum, sulfide clasts can occur as accidental lithics in volcaniclastic mass flow deposits due to erosion of massive sulfide ores by mass flows transporting volcanic detritus. Mass flows can erode several metres down into the substrate. Consequently, both sea-floor and shallow subsea-floor sulfide deposits can be eroded.

These sedimentary structures and textures are good evidence for sedimentation of sulfide on the sea floor and that the sulfide clasts were sourced from sea-floor or shallow subsea-floor sulfide sources, *provided* that the sedimentary structures and clastic textures are indisputable primary sedimentary features (cf. Kajiwara, 1970; Eldridge et al., 1983; Lydon, 1984b; Lydon and Galley, 1986), rather than replacements of earlier formed sedimentary structures, or in situ post-depositional brecciation textures. Indisputable primary sedimentary features include sedimentary structures within the sulfide deposit that are different from structures in the host rocks outside the sulfide deposit, and cases where different sulfide clasts show a range of different textures, and sulfide textures are truncated (broken) at clast margins.

Apparent sedimentary bedforms, structures and textures also occur in replacement sulfide deposits. Apparent exotic rock fragments can result from: (1) incomplete replacement of pre-existing clastic facies that leaves relics of the host (clasts and aggregates of clasts) within the sulfide deposit (Bodon and Valenta, 1995; Galley et al., 1995; Hannington et al., 1999; Sharpe and Gemmell, 2002) or (2) incomplete replacement of coherent lava or intrusion facies, which leaves discrete relict patches (pseudoclasts) of the coherent facies within the deposit (Doyle and Huston, 1999). Replacement of pre-existing clastic host facies by sulfide can also preserve the original bedforms, structures and textures of the host rocks (Allen, 1994a; Galley et al., 1995; Bodon and Valenta, 1995). Furthermore, an apparent interbedding of sulfide and sediment can result if more permeable laminae are selectively replaced and fine-grained (impermeable) laminae remain unaffected (Bodon and Valenta, 1995; Galley et al., 1995). All of these structures and textures are diagnostic of replacement providing that: (1) beds can be traced from weakly mineralized host rocks into semi-massive or massive sulfide (Allen, 1994a; Bodon and Valenta, 1995; Galley et al., 1995) and/or (2) partially replaced beds, clasts, aggregates of clasts, or relics of coherent facies (e.g., phenocrysts), occur within the sulfide deposit, and these relics are similar in texture to the host rocks outside the sulfide deposit (Allen, 1994a; Bodon and Valenta, 1995; Galley et al., 1995; Doyle and Huston, 1999). Similar immobile element ratios in massive to semi-massive sulfide ore and the enclosing host rocks (Relvas et al., 2000) can suggest that the sulfide deposit has replaced a part of the enclosing host rocks, provided that the host rock is a rapidly emplaced facies and not, for example, a siltstone that may have accumulated synchronously with the sulfide deposit.

Breccia ore textures can develop in situ within both sea-floor deposits and subsea-floor replacement deposits. Brecciation can accompany the dissolution of anhydrite in the matrix to sulfide patches (TAG: Humphris et al., 1995), or during hydraulic brecciation and veining of massive sulfides by subsequent hydrothermal activity (e.g., Mathiati: Lydon and Galley, 1986; Hellyer: McArthur and Dronseika, 1990). Cross-cutting veins in massive sulfide can superficially resemble bedding. Strong tectonic foliation can also produce banding and sulfide mylonites that may easily be misinterpreted as bedding (Renström: Duckworth and Rickard, 1993; Benambra: Allen and Barr, 1990; Chisel Lake: Galley et al., 1993; Geco: Friesen et al., 1982; Zaleski and Peterson, 1995). At sheared contacts between ductile sulfides and more competent wall rocks, foliated wall-rock fragments may be tectonically liberated, rotated and enveloped within a remobilized sulfide matrix (Sulitjelma: Cook et al., 1993). These tectonoclastic (durchbewegung) textures could be misinterpreted as bedded clastic sulfides or as evidence for syn-volcanic subsea-floor replacement.

3.2. Exhalites

Some VMS deposits are associated with sulfide, oxide or carbonate exhalites that are more extensive than the massive sulfide deposit (e.g., Key Tuffite horizon deposits, Matagami district: Liaghart and MacLean, 1992; Brunswick horizon deposits, Bathurst: Saif, 1983; Peter and Goodfellow, 1996; some deposits of the Noranda district: Knucky et al., 1982; Foot-Mud horizon, Anderson Lake: Walford and Franklin, 1982; Bailes and Galley, 1996). These exhalites typically occur at, or above, the ore horizon and are attributed to discharge from the same hydrothermal system, but not necessarily the same hydrothermal vents, that formed the massive sulfide deposit. Where an exhalite occurs at the same stratigraphic horizon as the massive sulfide deposit, and surrounds the deposit, it provides diagnostic evidence that the sulfide deposit formed at the sea floor. However, where the exhalite does not occur at the ore horizon, it does not constrain whether or not the massive sulfide formed at the sea floor. Establishing contemporaneity between an exhalite and massive sulfide deposit can be difficult, especially for stacked systems, or where exhalites from low temperature, regional hydrothermal systems are overprinted by higher temperature, hydrothermal up-flow zones (e.g., Knucky et al., 1982; Davidson et al., 2001). For example, tuffaceous exhalites that are interpreted to mark palaeosea-floor positions for some Noranda orebodies (Amulet, Millenbach) are intersected and partially replaced by the stringer zones of massive sulfide orebodies emplaced at higher palaeosea-floor positions (Knucky et al., 1982). Furthermore, due to an original glassy siliceous composition, fine-grained felsic tuffaceous deposits are readily altered to a cherty appearance during diagenesis or by moderate silicification. These cherty silicified rocks and some strongly carbonate- or iron oxide–quartz-altered rocks and hematitic oxidized shallow-water volcaniclastic sedimentary rocks have been mistaken for silica, carbonate and iron oxide exhalites, respectively, at several VMS deposits (Allen, 1992; Bodon and Valenta, 1995; Allen et al., 1996b; Doyle and McPhie, 2001; Herrmann and Hill, 2001). Consequently, "exhalites" only provide evidence for sea-floor deposition if they are shown to be true hydrothermal sediments or mixtures of hydrothermal sediment and slowly deposited ambient sediments. True exhalites are fine-grained rocks that have the bedform of slowly deposited suspension sediments (sheet-like geometry, planar lamination and/or thin planar bedding) are chemically and texturally distinct from adjacent hydrothermally altered volcaniclastic rocks, and show textural evidence that the hydrothermal component is not purely replacing pre-existing deposits (Saif, 1983; Liaghart and MacLean, 1992; Peter and Goodfellow, 1996; Allen et al., 1996b).

3.3. Evidence for rapid emplacement of the host facies

This criterion involves interpretation of the transport and depositional mechanisms of the host facies, in those cases where it can be demonstrated that massive sulfide deposit occurs within, and not simply between, emplacement units of the host facies. Massive sulfide deposits are interpreted to take from tens of thousands of years to over a million years to accumulate (Rona et al., 1993; Hannington et al., 1999). Consequently, a sea-floor VMS deposit can only form within a sedimentary or volcanic facies that accumulates at a slower rate than the sulfide deposit (such as pelagic mud) or between facies that are rapidly emplaced (such as volcanic and sediment gravity flow deposits and lavas). Massive sulfides that are hosted within rapidly emplaced sedimentary or volcanic facies can only have formed by impregnation and replacement (Allen, 1994a; Allen et al., 1996b).

Rapidly emplaced host facies to massive sulfide deposits include syn-eruptive pumiceous gravity flow deposits (Allen and Hunns, 1990; Morton et al., 1991; Allen, 1994b), pumiceous water-settled pyroclastic fall deposits (Allen et al., 1996b), water-settled fall and syn-eruptive gravity flow deposits of andesitic–basaltic fire fountain breccia and hyaloclastite (Zierenberg et al., 1988; Allen, 1992; Bodon and Valenta, 1995; Gibson and Gamble, 2000), post-eruptive, crystal-rich volcanic sandstone gravity flow deposits (Allen, 1994a), post-eruptive, polymictic volcanic breccia and conglomerate gravity flow units (Hannington et al., 1999; Allen et al., 1996b), in situ and resedimented autoclastic breccias (Kerr and Mason, 1990; Kerr and Gibson, 1993; Hill, 1996; Galley et al., 1995) and the coherent facies of lavas (Galley et al., 1995; Allen, 1992) and syn-volcanic intrusions (Doyle and Huston, 1999; Montelius et al., 2000).

Parts of some massive sulfide deposits are hosted by successions of thin to thick beds of volcano-sedimentary facies rather than a single thick depositional unit (e.g., South Hercules: Khin Zaw and Large, 1992; Currawong: Bodon and Valenta, 1995; Gossan Hill: Sharpe et al., 1997). For these ores, the possibility remains that sedimentation was synchronous with hydrothermal activity and that precipitation of sulfides occurred at and below the sea floor over the life of the hydrothermal system.

3.4. Identification of replacement fronts

Replacement fronts are mineralization boundaries that overprint and transgress earlier formed structures, textures, or bedforms in the host rocks, thus indicating that the hydrothermal minerals replaced pre-existing deposits. The replacement fronts are commonly lobate or irregular in shape, and may be gradational or sharp. Sharp replacement fronts are especially common in carbonate rocks. Gradational replacement fronts are characterized by progressively increasing sulfide abundance and alteration intensity toward ore. This zone of increasing mineralization intensity may contain disseminated, spotty and blebby semi-massive sulfides that record the nucleation and growth of ore forming minerals within the host during replacement (Allen and Hunns, 1990; Khin Zaw and Large, 1992; Allen, 1994a; Miller, 1996).

3.5. Discordance with the enclosing lithofacies

Discordance between a mineral deposit and bedding in the enclosing rocks can provide evidence for replacement, provided that contacts are not faulted (e.g., Galley et al., 1993; Doyle and Huston, 1999). However, stratigraphic onlap, or interfingering, between sedimentary or volcanic units and a massive sulfide deposit can also produce discordant contacts (e.g., Middle Valley; Mottl et al., 1994). Accordingly, the criterion cannot be used alone and requires consideration of the nature of the contact and the facies characteristics of the host rocks.

3.6. Hanging wall alteration similar in style and intensity to footwall alteration

Sea-floor VMS deposits typically have an asymmetric alteration pattern comprising intense hydrothermal alteration in the footwall rocks and weaker alteration in the hanging wall, with or without local strong alteration directly above the massive sulfide (e.g., Kuroko deposits: Iijima, 1974; Mount Chalmers: Large and Both, 1980; Hellyer: Jack, 1989; Gemmell and Fulton, 2001). This pattern indicates that the main stage of ore-forming hydrothermal activity occurred after, or concurrent with, emplacement of the footwall rocks and before deposition of the hanging wall rocks. The hanging wall alteration records declining hydrothermal activity during accumulation of the hanging wall rocks (Iijima, 1974). The presence of strong alteration, stringer veining and disseminations in both the footwall and hanging wall rocks indicates either that ore-forming hydrothermal activity continued during deposition of the hanging wall succession or that massive sulfide deposition was entirely subsea-floor. Both scenarios are plausible and are discussed below.

3.7. Sulfide chimneys, biota, microtextures

Chimneys and chimney fragments provide strong evidence for sulfide accumulation at the sea floor (Lydon, 1988). These textures have been recognized in some ancient deposits (Oudin and Constantinou, 1984); however, there are few undisputed examples and some chimney-like structures (e.g., Sulfur Springs: Vearncombe et al., 1995) have been interpreted as mineralised conduits in subsea-floor replacement ores (Morant, 1995). Fossil tubeworms and bivalves in communities that colonize modern sea-floor sulfide mounds can be preserved by sulfide minerals and are also characteristic of sea-floor deposits (Haymon et al., 1984; Oudin and Constantinou, 1984; Jonasson and Perfit, 1999). Few other textures or structures in VMS deposits have genetic significance in distinguishing sea-floor deposits from subsea-floor deposits. Microaerophilic chemotrophic bacteria can colonize subsea-floor sediments and hydrothermal conduits (e.g., Jannasch and Mottl, 1985; Haymon et al., 1993; Parkes et al., 1994) so are not characteristic of a sea-floor position. Many massive sulfide textures form through open space filling (e.g., colloform banding, network textures). However, this can include infilling of intramound porosity, primary or secondary host-rock porosity, and/or space formed by fluid–rock–mineral interaction, such as dissolution of anhydrite (e.g., Lydon, 1984a,b; Embley et al., 1988; Humphris et al., 1995; Hannington et al., 1995; Galley and Koski, 1999).

4. Styles of subsea-floor replacement

The roles of subsea-floor replacement and sea-floor accumulation in the genesis of VMS deposits are now considered in light of the preceding discussion (Tables 1 and 2). Based on variation in ore deposit geometry and textures, distribution of alteration assemblages, and the facies associations of the host succession, 12 main styles (Fig. 1) and 5 specific volcanic settings of massive sulfide deposit can be recognized. More specific settings likely exist, but those selected here serve to highlight the main features and diversity of the deposits. There is also a continuum between the different styles and settings of deposit, even within a single complex ore deposit. The analysis highlights the influence of environment on the mineralization processes and consequently VMS deposit styles and ore types.

4.1. Sea-floor massive sulfide deposits

Three main styles of sea-floor VMS deposit are recognized (Fig. 1A–C): mound deposits (e.g., TAG: Rona et al., 1993; Humphris et al., 1995; Millenbach: Knucky et al., 1982; Gibson and Watkinson, 1990; Hellyer: McArthur, 1989; Gemmell and Large, 1992), clastic aprons and depression-fills that may or may not be located at the margins of mound deposits (e.g., Matsumine-Shakanai: Kajiwara, 1970; Ito et al., 1974; Buchans-MacLean lens: Binney, 1987; Hayl-as-Safil: Galley and Koski, 1999) and sheet deposits (e.g., Brunswick 12: van Staal and Williams, 1984; Tharsis: Tornos et al., 1998). Mound deposits are characterized by a mound morphology that rose tens of metres above the surrounding sea floor, endogenic growth and internal zone-refining, and their position directly above a "pipe" of footwall alteration and sulfide veins (Franklin et al., 1981; Eldridge et al., 1983; Rona et al., 1993; Humphris et al., 1995). They may contain clastic sulfides and sulfide chimneys. Clastic apron and depression-fill deposits are defined here as deposits with abundant clastic, sedimented sulfides that have bedforms indicating lateral transport, and the overall facies architecture of a talus apron, depression-fill or channel-fill. Parts of the deposits may be dispersed tens to hundreds of metres from the source stringer zone. Sheet deposits are characterised by their extensive tabular to sheet-like form and an extensive zone of footwall alteration and sulfide veining or impregnation. Some authors attribute some ancient VMS deposits, and especially the sheet style deposits, to precipitation of sulfides from exhalative sea-floor brine pools, analogous to Atlantis II Deep (Solomon and Walshe, 1979; Solomon and Khin Zaw, 1997; Tornos et al., 1998).

Many sea-floor massive sulfide deposits have a sharp top contact and strong alteration is restricted in extent above the deposits compared to the footwall. These characteristics indicate that the massive sulfide formed during a period of slow accumulation of volcanic and sedimentary deposits, and the hydrothermal system was waning by the time the massive sulfide became buried. Nonetheless, relatively rapid burial of sea-floor massive sulfide deposits is considered important in preserving them from oxidation and erosion (Herzig and Hannington, 1995). In contrast, the lower boundary of the mound and sheet style deposits is gradational into the footwall alteration-stringer zone, and shows evidence for extensive replacement of the host-rocks (replacement criteria 1–4 above). Evidence of replacement of host rocks is generally lacking in the uppermost parts of the massive sulfide. These characteristics indicate that the massive sulfide formed at the sea floor but grew downwards by replacement of host rocks in the upflow zone as well as upwards by precipitation at the sea floor (Lydon, 1988; Herzig and Hannington, 1995; Hannington et al., 1995). The clastic apron and depression-fill deposits may show similar evidence of replacement in the upper part of the footwall alteration pipe, but the clastic sulfide deposit may be displaced laterally from the footwall pipe and replacement sulfides. In all these sea-floor deposits, replacement massive sulfides formed at shallow depths (up to ?50 m below the sea floor). More complex deposits are discussed below.

4.2. Sea-floor massive sulfide deposits modified after burial

Some massive sulfide lenses display evidence for sea-floor accumulation, but also have strong hanging wall alteration. It can be inferred that volcanic or sedimentary strata rapidly buried these lenses while the hydrothermal system was still at its peak (Fig. 1D–E). Deposition of the hanging wall rocks interrupted ore formation. These deposits generally show evidence of strong modification compared to the simpler sea-floor deposits discussed above. Two main cases are considered: massive sulfides buried by lavas and those buried by clastic deposits.

4.2.1. Sea-floor massive sulfide deposits modified after burial by lavas

VMS deposits that were rapidly buried by lava flows have been documented in ancient (Fukazawa: Sato et al., 1979; Amulet and Millenbach deposits: Knucky et al., 1982) and modern settings (East Pacific Rise: Haymon et al., 1993). The well studied Noranda examples show discordant alteration zones and sulfide vein networks that extend up through the massive sulfide deposit and up to 300 m into the hanging wall lavas (Knucky et al., 1982; Gibson and Watkinson, 1990; Kerr and Gibson, 1993). The hanging wall lava flows formed a thermal insulator and impermeable cap that facilitated replacement in the buried sulfide deposit. The hanging wall sulfide veins indicate that the buried hydrothermal system hydraulically fractured and permeated its way into the overlying lavas. In some cases (Amulet A lenses; Millenbach Main lens and 19 lens) the system penetrated the hanging wall lavas and formed massive sulfide at the new sea floor position, resulting in a series of vertically stacked sulfide lenses. The more deeply buried lenses are more copper-rich, indicating that strong zone-refining or replacement occurred well below the sea floor (Fig. 2).

Consequently, this deposit style is characterised by: (1) evidence for sea-floor sulfide accumulation, (2) contact relationships and textural and facies evidence for emplacement of lavas onto sea-floor massive sulfides, such as peperitic contacts between lava and clastic sulfides, (3) strong hydrothermal alteration $\pm$ disseminated, vein, semi-massive or massive sulfides in the base of hanging wall lava flows, or similar zones that cut through the flows, (4) stacked deposits linked by stringer zones and strong hydrothermal alteration. The minimum required evidence to distinguish this style of VMS deposit is (1) and (3).

4.2.2. Sea-floor massive sulfide deposits modified after burial by clastic deposits

VMS deposits interpreted to have formed on the sea floor and to have been modified during or after rapid burial by clastic rocks include Que River (Large et al., 1988), the Gopher, South Trough and Battle Main lenses, Myra Falls (Robinson et al., 1996; Sinclair et al., 2000), Woodlawn (Petersen and Lambert, 1979) and Mount Chalmers (Large and Both, 1980). These VMS deposits generally coincide with intervals of mudstone or thinly bedded sedimentary rocks that accumulated slowly, or they occur between thick beds of rapidly deposited clastic facies, both of which are consistent with them having formed on the sea floor. Lithofacies characteristics (bedforms, structures, clast types) of high-concentration turbidity current deposits and other mass flow deposit types indicate that the clastic hanging wall rocks were mainly rapidly deposited. The rocks are generally syn-eruptive pumiceous deposits and syn-eruptive to post-eruptive lithic volcaniclastic deposits.

Similar to the massive sulfide deposits buried by lavas, these deposits have strong alteration zones (sericite–quartz–chlorite $\pm$ pyrite $\pm$ carbonate) including stringer, disseminated or semi-massive sulfides, that extend from the footwall up to 200 m into the hanging wall, and in some cases link stacked sulfide lenses (Fig. 3a; Battle, Que River). Alteration

Table 1
Characteristics of major Australian VMS deposits

Deposit	Style	Immediate host lithofacies	Precursor facies within mineral deposit	Rapidly emplaced host facies	Replacement fronts	Discordance >< ore and host rocks	Hanging wall alteration	Sulfide clast-bearing deposits	Interpreted environment[a]	References
Au type										
Henty-Mt. Julia	lenses	dacitic volcaniclastic rocks; massive carbonate; carbonate impregnated rocks	±	×	×	×	strong/ intense	×	subsea-floor (hybrid VMS/ epithermal)	(Halley and Roberts, 1997; Callaghan, 2001)
Cu–Au type										
Mt. Lyell	stockwork	pyrite–quartz–sericite–pyrophyllite schist; rhyolitic–andesitic volcanic rocks	✓	±	±	±	± strong/ intense	±	subsea-floor + sea floor (VMS/ epithermal)	(Corbett, 2001; Huston and Kamprad, 2001)
Mount Chalmers	lenses	tuffaceous sandstone, siltstone, mudstone	×	×	×	×	weak/strong	✓	sea floor ± subsea-floor	(Large and Both, 1980; Taube, 1990; Sainty, 1992; Hunns, 1994)
Mount Morgan	pipe	pumiceous–tuffaceous breccia–mudstone; jasper; rhyolite; carbonate	✓	±	✓	✓	weak–moderate	±	subsea-floor	(Taube, 1986; Messenger et al., 1997; Ulrich et al., 2002)
Highway-Reward	pipes	rhyolitic–dacitic syn-sedimentary intrusions	✓	✓	✓	✓	strong	×	subsea-floor	(Doyle and Huston, 1999; Doyle and McPhie, 1994, 2000; Doyle, 2001)
Balcooma	lenses	quartz ± chlorite ± muscovite ± staurolite ± garnet ± biotite schist	×	×	×	×	between lenses	×	sea floor ± subsea-floor	(Huston, 1990; Huston et al., 1992)
Gossan Hill	sheet	tuffaceous siltstone–sandstone; pumice breccia; chert	✓	±	✓	±	moderate/ strong	±	subsea-floor and sea floor	(Sharpe et al., 1997; Sharpe and Gemmell, 2000, 2001, 2002)
Whundo	lenses	chlorite–muscovite–quartz ± andalusite schist	×	×	×	×	not reported	×	indeterminate	(Reynolds et al., 1975; Barley, 1992)
Zn–Cu type										
Wilga	lens	siltstone, sandstone turbidites, dacite and hyaloclastite	✓	×	✓	×	strong ($\leq$ 10 m)	×	subsea-floor ± sea floor	(Allen and Barr, 1990; Allen, 1992)
Scuddles	lenses	polymictic breccia–sandstone, tuffaceous breccia–sandstone, chert	×	×	×	×	strong	✓	sea floor ± subsea-floor	(Ashley et al., 1988; Mill et al., 1990; Clifford, 1992)

Teutonic Bore	lens	pyritic chloritised schist, basalt	± (margin/ base)	±	±	×	moderate (~ 10 m)	×	sea floor ± subsea-floor	(Greig 1984; Hallberg and Thompson, 1985; Present study)
Zn–Pb–Cu type										
Woodlawn	lenses	black shale	×	×	×	×	intense	✓	sea floor ± subsea-floor	(Ayres, 1979; Petersen and Lambert, 1979; McKay and Hazeldene, 1987)
Captains Flat	lenses	shale, tuffaceous siltstone	?	×	×	×	not reported	×	sea floor?	(Davis, 1975; Bain et al., 1987; Davis 1990)
Currawong	sheet	dacite, basalt, hyaloclastite, turbidites and siltstone	✓	✓	✓	×	strong	×	subsea-floor ± sea floor	(Bodon and Valenta, 1995)
Rosebery	lenses	tuffaceous sandstone, pumice breccia	✓	✓	✓	×	moderate–strong	✓	subsea-floor ± sea floor	(Huston and Large, 1988; Lees et al., 1990; Aerden, 1991; Allen, 1994a)
Hercules	lenses	shale, tuffaceous sandstone, pumice breccia	✓	✓	✓	✓	strong	✓	subsea-floor ± sea floor	(Green et al., 1981; Lees et al., 1990; Aerden, 1993; Allen, 1994a)
South Hercules	lenses	tuffaceous sandstone–siltstone	±	±	✓	±	strong	×	subsea-floor ± sea floor	(Khin Zaw and Large, 1992)
Que River	lenses	andesite, coarse volcaniclastic units	×	×	×	×	between lenses	×	sea floor ± subsea-floor	(Large et al., 1988)
Hellyer	mound	polymictic–monomictic volcaniclastic breccia, sandstone	×	×	×	×	weak	✓	sea floor	(Drown, 1990; McArthur and Dronseika, 1990; Waters and Wallace, 1992)
Thalanga	lenses	rhyolite, dacite, rhyolitic breccia–sandstone beds, siltstone	±	±	±	×	between lenses	±	subsea-floor>sea floor	(Hill, 1996; Paulick and McPhie, 1999)
Liontown	lenses	pumice breccia, crystal-rich sandstone, siltstone	✓	±	±	×	between lenses	×	subsea-floor and sea floor	(Miller, 1996; Doyle, unpub. data)
Sulphur Springs	lenses	silicified siltstone–sandstone, perlitic dacite, pumice breccia	✓	±	✓	×	limited	×	subsea-floor ± sea floor	(Vearncombe, 1995; Vearncombe et al., 1995; Morant, 1995, 1998)

× : Feature not documented or absent.

[a] Principal mineralizing environment of system, excluding stringer zone.

Table 2
Characteristics of major subsea-floor replacement VMS deposits from Canada, North America, Japan, Portugal, Spain and Sweden

Deposit	Style	Immediate host lithofacies	Precursor facies within mineral deposit	Rapidly emplaced host facies	Replacement fronts	Discordance between ore and host rocks	Hanging wall alteration	Sulfide clast-bearing deposits	Interpreted environment[a]	References
Canada										
Ansil	lens	tuffaceous sandstone/siltstone, massive-pillowed andesite	✓	±	✓	±	strong (500 m)	×	subsea-floor > sea floor	(Galley et al., 1995)
Battle Mine	sheet/lens	tuffaceous sandstone/siltstone, pumice breccia, rhyolite sill	×	±	×	×	strong	×	sea floor and subsea-floor	(Robinson et al., 1996; Sinclair et al., 2000)
Kidd Creek	lenses	monomict–polymict volcaniclastic breccia–sandstone, rhyolite, argillite	✓	±	✓	✓	strong	✓	subsea-floor > sea floor	(Hannington et al., 1999; Barrie et al., 1999)
Horne	lenses	resedimented rhyolitic autoclastic breccia	✓	✓	✓	×	×	✓	subsea-floor > sea floor	(Kerr and Mason, 1990; Kerr and Gibson, 1993)
Mattabi	lenses	syn-eruptive, pumiceous mass-flow deposits	×	✓	×	×	strong	×	subsea-floor	(Gibson et al., 1999)
Coniagas	lenses	syn-eruptive felsic lithic–pumice breccia	✓	✓	✓	×	strong (~ 20 m)	×	subsea-floor	(Doucet et al., 1994, 1998)
Potter	lenses	basaltic fire-fountain deposits and sills, argillaceous mudstone, chert	✓	±	±	×	envelope	✓	subsea-floor > sea floor	(Gibson and Gamble, 2000)
Chisel Lake	lenses	limestone–dolomite skarn, dacitic tuffaceous units	✓	×	✓	±	limited (< 2 m)	×	subsea-floor	(Galley et al., 1993; Bailes and Galley, 1996)
Vermilion	lenses	carbonate exhalite, chert, turbidites	✓	×	✓	×	✓	×	subsea-floor	(Gray and Gibson, 1993; Stoness et al., 1993)
North America										
Turner–Albright	lenses	basaltic hyaloclastite, talus breccia	✓	✓	✓	±	×	✓	subsea-floor > sea floor	(Zierenberg et al., 1988)
Lynne	lenses	rhyolite breccia, carbonate rocks, skarn, chert	✓	×	✓	×	✓	×	subsea-floor	(DeMatties, 1994)

Japan										
Matsuki		tuffaceous sandstone, mudstone	✓	±	✓	✓	✓	×	subsea-floor	(Kuroda, 1983)
Portugal/Spain										
Los Frailes	lens	black shale	✓	×	✓ (lower part)	×	±	✓ (at top)	sea floor and subsea-floor	(Almodóvar et al., 1998; Tornos, 2000; Allen, 2001; Fernández Martinez, 2001)
Neves Corvo	lenses	rhyolitic dome–cryptodome–hyaloclastite complex	✓	×	✓	×	×	×	subsea-floor	(Relvas et al., 2000)
Sweden										
Kyrkvägen	lenses	rhyolitic pumice breccia	✓	✓	×	×	×	×	subsea-floor ± sea floor	(Allen et al., 1996b)
Renström	lenses	rhyolitic pumice breccia, peperite, siltstone	✓	✓	±	×	×	×	subsea-floor ± sea floor	(Allen et al., 1996b)
Renström East	lenses	basaltic andesite–andesitic scoria beds, sills	✓	✓	×	×	strong (50 m)	×	subsea-floor	(Allen et al., 1996b)
Långdal	lenses	rhyolitic pumice breccia	✓	✓	×	×	×	×	subsea-floor	(Allen et al., 1996b)
Långsele	lenses	rhyolitic pumice breccia	✓	✓	×	×	×	×	sea floor and subsea-floor	(Allen et al., 1996b)
Petiknäs North	lenses	lithic-rich bases of rhyolitic syn-eruptive mass-flow deposits	✓	✓	±	×	±	×	subsea-floor ± sea floor	(Allen et al., 1996b)
Boliden	lenses	dacite–basalt, pumice–lithic breccia–sandstone, mudstone	×	±	×	✓	strong (20–50 m)	×	subsea-floor + epithermal?	(Bergman Weihed et al., 1996; Allen et al., 1996b)
Holmtjärn	lenses	post-eruptive conglomerate, sandstone, mudstone	✓	✓	±	×	strong (150 m)	×	subsea-floor	(Allen et al., 1996b)
Maurliden	lenses	rhyolitic–dacitic lavas/intrusions, tuffaceous sandstone, mudstone	✓	✓	✓	×	✓	×	subsea-floor ± sea floor	(Allen et al., 1996b; Montelius et al., 2000)
Garpenberg	lenses	limestone–dolomite skarn, felsic volcaniclastic units	✓	±	✓	✓	locally strong	×	subsea-floor	(Allen et al., 1996a)

× : Feature not documented or absent.

[a] Principal mineralizing environment of system, excluding stringer zone.

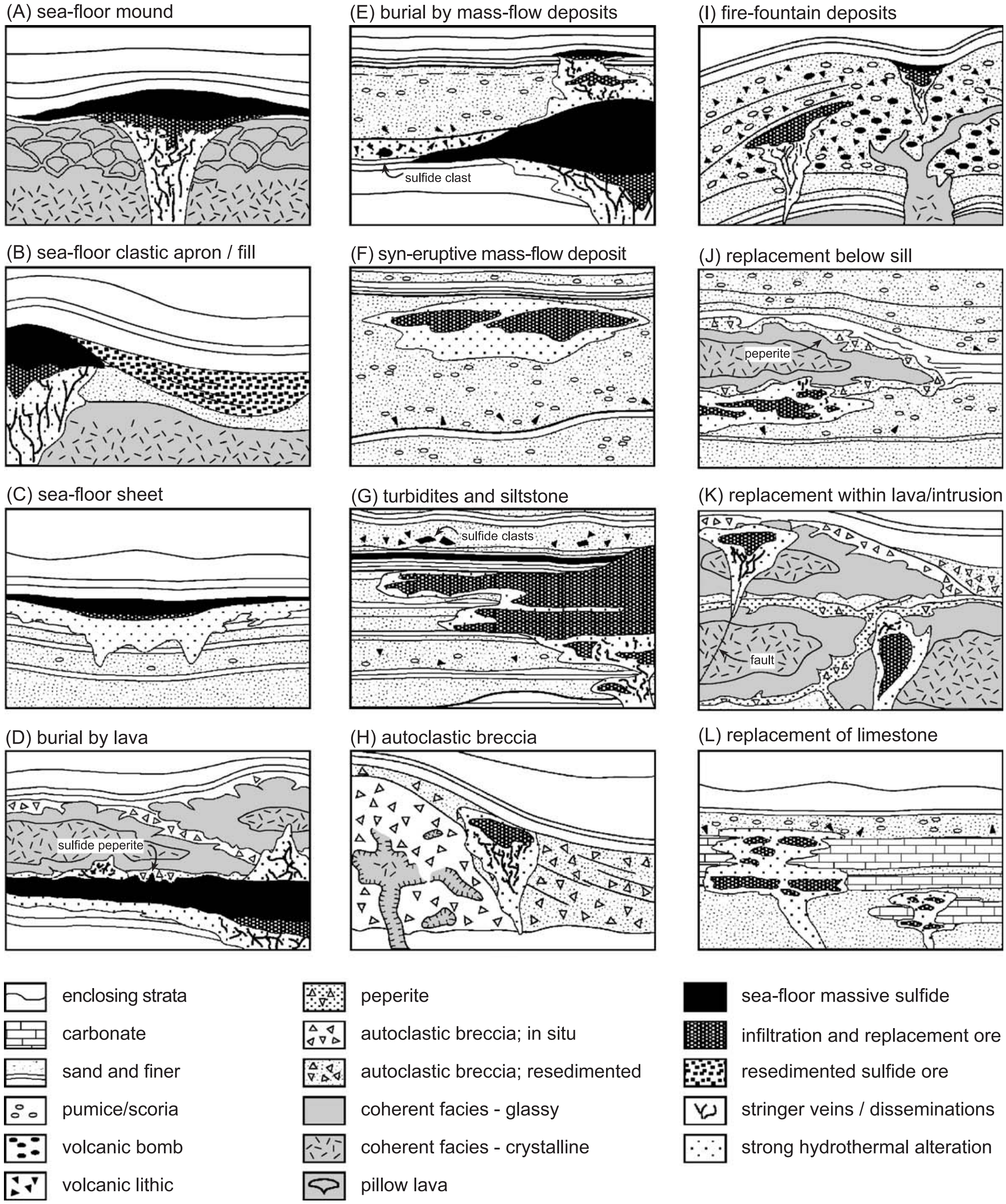


Fig. 1. Schematic representation of the 12 main styles of VMS deposits discussed in this paper. The relationships between lithofacies, hydrothermal alteration and massive sulfides are shown for sea-floor deposits (A–C), sea-floor deposits modified after burial by lavas or volcaniclastic deposits (D–E), and massive sulfide deposits dominated by infiltration and replacement (F–L).

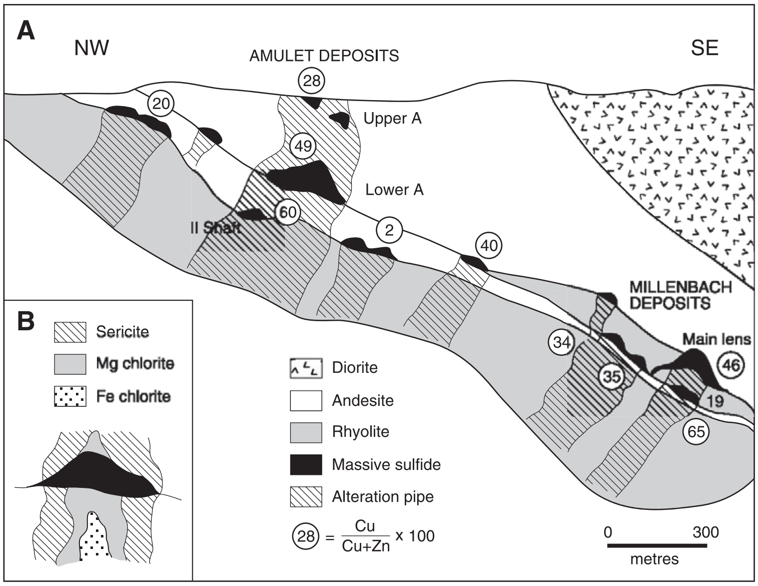


Fig. 2. Simplified geologic cross-section through the Amulet and Millenbach deposits showing the distribution of (A) lithofacies, strong hydrothermal alteration, and metal ratios for combined massive and stringer ore. The inset (B) illustrates the alteration zonation and assemblages for the stacked deposits. Modified after Knucky et al. (1982) and Kerr and Gibson (1993).

halos associated with most deposits are asymmetric, with the most intense and extensive hydrothermal alteration and stringer vein development in the footwall to the sulfide lenses. Strong hanging wall ore-associated hydrothermal alteration can form widespread zones between stacked lenses (Battle; Robinson et al., 1996), small discordant chimney-like silicified zones at the lateral margins of the sea-floor massive sulfides (Fig. 4A; Mount Chalmers: Large and Both, 1980), or form a concentrically zoned envelope around the deposit (Fig. 4B; Woodlawn: Petersen and Lambert, 1979). Compared to the ore deposits buried by lavas, these deposits generally have more extensive replacement sulfides in the hanging wall rocks.

It is conceivable that there are VMS deposits in which sea-floor sulfide accumulation was interrupted numerous times by the deposition of thin to thick mass flow beds. After each burial, sulfide deposition might migrate up to the new sea floor position by veining or infiltration and replacement of the porous clastic debris and then resume deposition of massive sulfide at the sea floor. Such a deposit might comprise numerous stratiform to stratabound sulfide lenses or sheets, alternating with ambient clastic rocks. Sulfide deposition would be both by sea-floor accumulation and replacement just below the sea floor throughout the life of the hydrothermal system. Although some authors have appealed to this type of scenario to explain VMS deposits that have sulfide lenses at several levels through a clastic host succession (Mattabi: Morton et al., 1991; Thalanga: Gregory et al., 1990; Rosebery: Lees et al., 1990; Currawong: Cox et al., 1990), there are few undisputed examples and these deposits have also been explained by subsea-floor replacement with relatively minor or no sea-floor sulfide accumulation (Mattabi: Gibson et al., 1999; Thalanga: Hill, 1996; Rosebery: Allen, 1994a; Currawong: Bodon and Valenta, 1995). It is possible that the diagnostic evidence for sea-floor accumulation of sulfides could be obscured by the upward migration of subsea-floor replacement fronts towards new sea floor positions. However, each sulfide lens that formed on the sea floor should still occur between rapidly deposited clastic beds and not within them (Allen, 1994a).

Sea-floor massive sulfide deposits modified after burial by clastic deposits are characterised by: (1) evidence for sea-floor sulfide accumulation, (2) textural and facies evidence for emplacement of clastic mass flow deposits onto sea-floor massive sulfides, such as rip-up clasts of massive sulfide within the

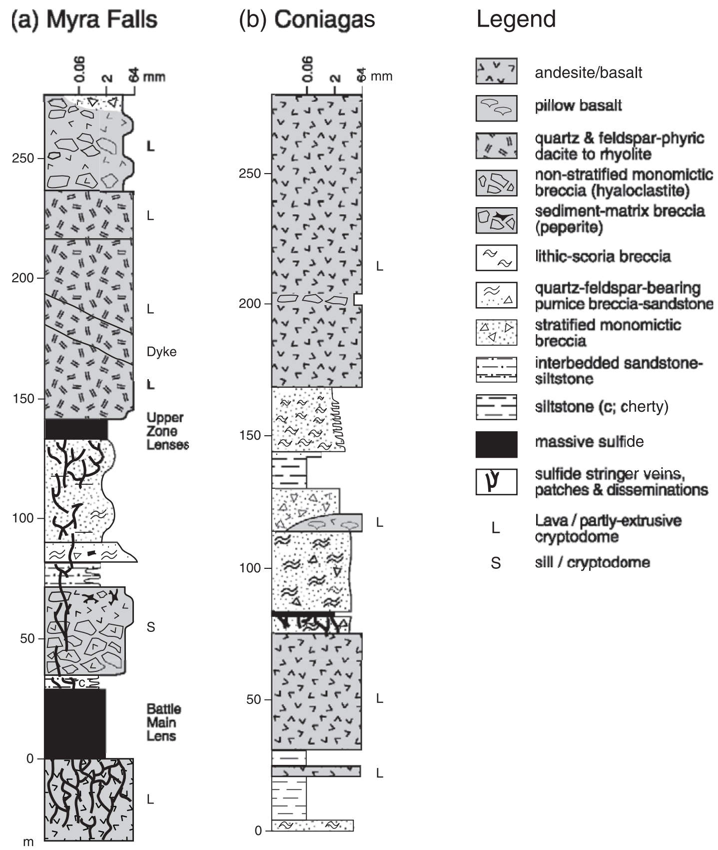


Fig. 3. Simplified stratigraphic columns for the Myra Falls and Coniagas VMS deposits in Canada. Modified after Robinson et al. (1996) and Doucet et al. (1998).

immediate hanging wall clastic unit, (3) strong hydrothermal alteration ± disseminated, vein, semi-massive or massive sulfides in the hanging wall rocks, (4) stacked deposits linked by replacement zones, stockworks, and strong hydrothermal alteration. The minimum required evidence to distinguish this style of VMS deposit is (1) and (3).

4.3. Massive sulfide deposits dominated by subsea-floor replacement

4.3.1. Subsea-floor replacement deposits in volcaniclastic rocks

Subsea-floor replacement VMS deposits have been interpreted to occur in a diverse range of host rocks, including massive to bedded clastic deposits, autoclastic facies associated with lavas and intrusions, and even within the coherent parts of lavas and intrusions. However, the majority of replacement-style VMS deposits are associated with volcaniclastic facies, and the characteristics of these deposits are summarized below.

4.3.1.1. In pumiceous deposits

Pumiceous gravity flow deposits. This style of VMS deposit contains sulfide lenses hosted within syn-eruptive, non-welded, generally felsic, pumice deposits, commonly near the top of a thick felsic eruption sequence that is overlain by post-eruptive, thinly bedded volcano-sedimentary rocks (Fig. 1F). The pumice deposits consist of one or more 5–500 m thick, laterally extensive (1–10's km) beds. Each bed has a lower massive part of breccia grain size (pumice breccia) and a normally graded or double-graded upper part of breccia to sand grainsize (pumice–breccia and ash–sandstone). The lithofacies charac-

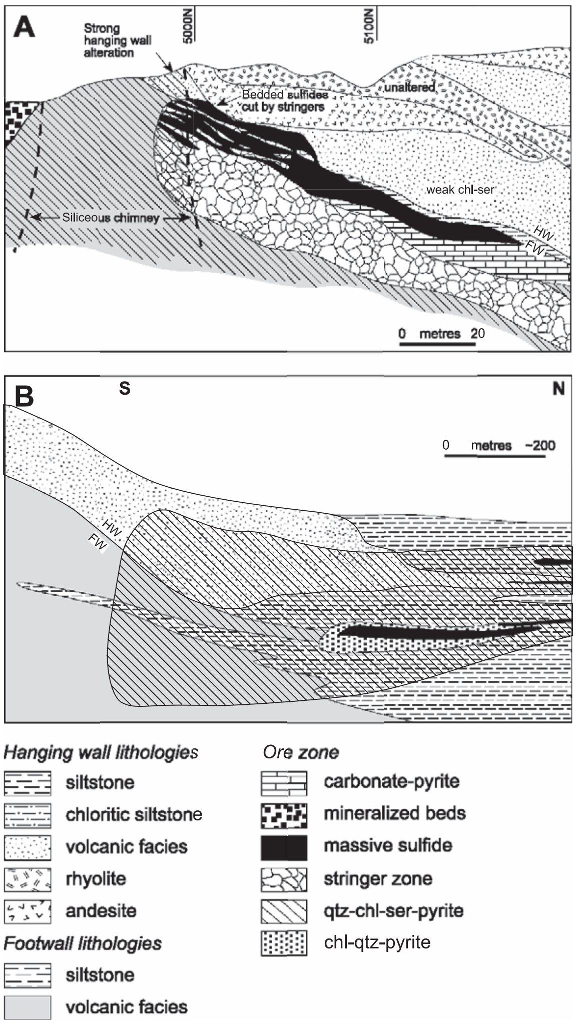


Fig. 4. Simplified geological cross-sections for (A) Mount Chalmers and (B) Woodlawn VMS deposits. The sections illustrate the character and arrangement of hydrothermal alteration zones, lithofacies and massive sulfides that developed in response to burial of sea-floor massive sulfides. Qtz = quartz, chl = chlorite, ser = sericite. Modified after Large and Both (1980) and Petersen and Lambert (1979), respectively.

teristics imply deposition from large-volume subaqueous gravity flows (mega-turbidity current flows) fed from pumiceous pyroclastic eruptions (Allen and Cas, 1990; Morton et al., 1991; Allen, 1994b). The instantaneous mass flow emplacement mechanism implies that the enclosed ores formed by infiltration and replacement, and precludes accumulation of the sulfide on the sea floor synchronous with volcaniclastic sedimentation (Allen, 1994a). The minimum depth below the sea floor that replacement occurred is the distance from the sulfide body to the top of the enclosing depositional unit.

Replacement ores of this type are the most widely documented type of VMS replacement deposit. Examples include, parts of the Rosebery, Hercules, Liontown, Gossan Hill, and Sulfur Springs deposits in Australia (Fig. 5a–c); Långdal, Långsele (Fig. 6a) and parts of the Renström (Fig. 6b) and Kyrkvägen deposits in Sweden; some lenses of the Hanaoka-Shakanai cluster in Japan; San Platón and Concepción in Portugal; and the Mattabi and other deposits in the Sturgeon Lake region of Canada (Tables 1 and 2). Each deposit comprises one or more massive sulfide lens and adjacent zones of low-grade disseminated sulfides. The ore lenses at a single deposit typically occur at more than one stratigraphic level within a stratigraphic interval of up to 100 m.

Separate ore lenses are linked by zones of strong hydrothermal alteration, disseminated sulfides and/or veining. However, alteration envelopes around most deposits are asymmetric, with the strongest and most extensive alteration in the footwall and laterally adjacent to the sulfide lenses. Carbonate alteration, especially dolomite, manganiferous carbonates and iron carbonates, are a common feature of this VMS deposit style (cf. Morton and Franklin, 1987). At Mattabi, there is a compositional zonation from Mg-rich to Fe-rich carbonate passing from the semi-conformable alteration into the footwall alteration pipe (e.g., Franklin et al., 1981; Morton and Franklin, 1987), whereas at Rosebery, the Mn content of semi-conformable stratabound spotty carbonate increases towards ore (Large et al., 2001). The massive to semi-massive sulfides and carbonate-rich alteration zones enclose relics (15–50% at Liontown) of, and grade laterally out into, the host rock. Spotty and blebby sulfide and carbonate textures are common and suggest that replacement commenced at scattered nucleation sites (commonly phenocrysts) within the pumiceous host (Fig. 7A–B).

Sulfide clasts in volcaniclastic mass-flow deposits overlying some of the ores (Hercules, Rosebery), or in the uppermost part of the massive deposit (some Hanaoka-Shakanai lenses) suggest that at least the uppermost part of the ore lens system formed close to the sea floor, or on the sea floor (Allen, 1994a). The distance from the base of the massive sulfide body to the base of the sulfide-clast bearing bed provides the maximum distance below the sea floor that replacement massive sulfide formed, and appears to be between ~ 100 and 150 m for Rosebery and Hercules. At Hanaoka-Shakanai, sedimentary structures in the upper part of some gypsum ores implies that replacement of the host pumice breccia–sandstone occurred less than ~ 50 m below the sea floor (Ito et al., 1974; Ohtagaki et al., 1974). At Liontown, replacement ores occur from ~ 5 m (Lower Lode) to 50 m (Carrington Lodes) below thin sea-floor massive sulfide lenses (Upper and Central Lodes) that are hosted in thinly bedded, post-eruptive volcano-sedimentary rocks (Miller, 1996; Fig. 5b).

Morton et al. (1991) suggested that the ore lenses of the Mattabi and nearby VMS deposits occur at bed boundaries in the thick pumiceous pyroclastic succession and that the ores formed in the short repose times between episodic eruptions; each eruption generating a major bed. However, there are no inter-eruptive interbeds in the succession and successive pumice beds are similar in composition. Consequently, the time interval between deposition of successive beds

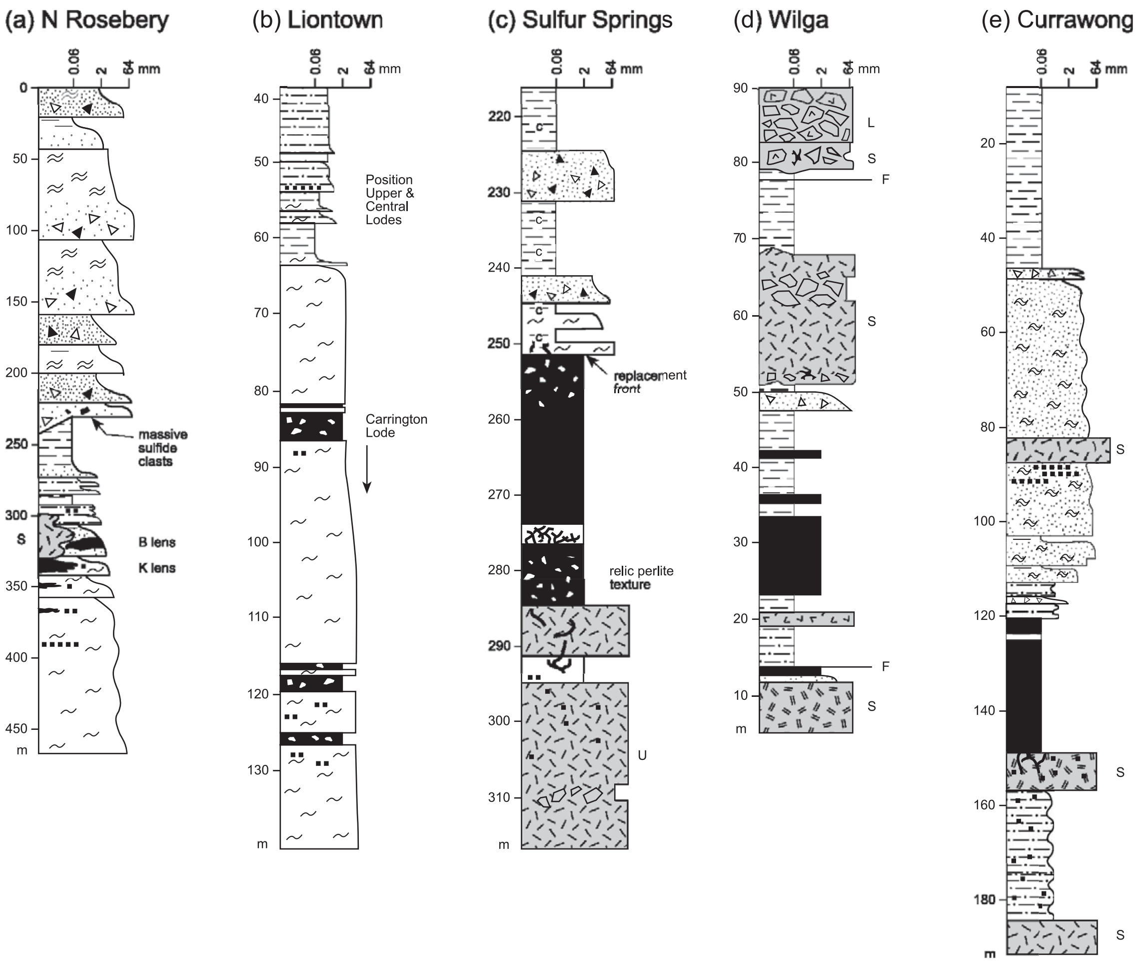


Fig. 5. Simplified stratigraphic columns for some Australian VMS deposits that are dominated by infiltration and replacement ores. The columns are presented as graphic lithological logs with grain-size profiles and are based on single or composite diamond drill hole logs (Rosebery North end: 120R, 120RD4, 49R, 74R, 78R; Liontown: LLD 101; Sulfur Springs: SSD14; Wilga: DDH 25; Currawong: BH45, BH59; Golden Grove; East Thalanga: E3204SD17, E3204SI52; Highway: REMM 560). Based on Allen (1992, 1994a), Hill (1996), Sharpe and Gemmell (2000) and the present study.

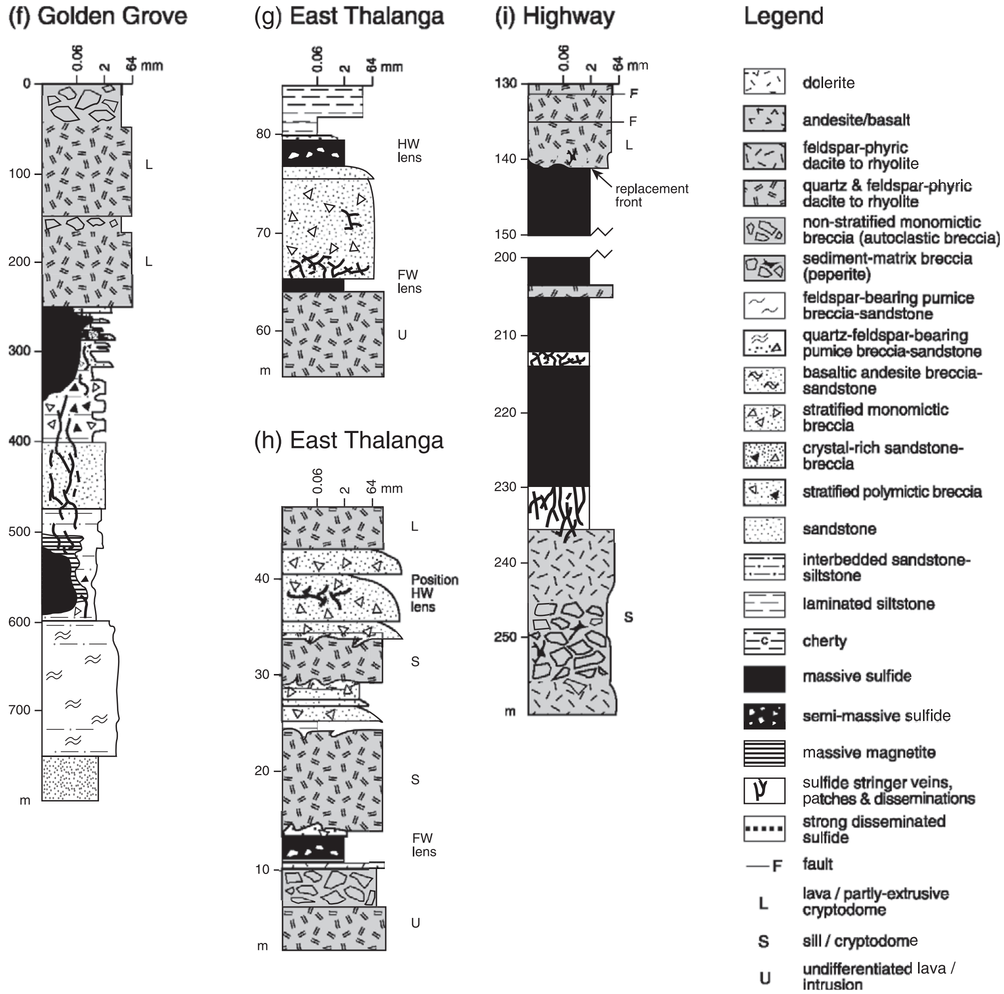


Fig. 5 (*continued*).

and for the whole package of related beds may have been very short. Gibson et al. (1999) interpret the Mattabi and other Sturgeon Lake Camp VMS deposits to have formed mainly by subsea-floor replacement within the pumiceous deposits. Evidence for subsea-floor replacement includes gradations between sulfide and host rock at the margins of the orebodies, and the presence of quartz phenocrysts and altered host rock relics within the ores (Allan Galley, written communication, 2003).

The small Renström East massive sulfide deposit (Allen et al., 1996b; Allen and Svenson, 1999) is a mafic-hosted analogue that occurs 150–350 m stratigraphically below the felsic-hosted Renström and Kyrkvägen deposits, and is part of the same hydrothermal upflow zone. Renström East comprises massive and semi-massive sulfide within a large zone of strong quartz–sericite–pyrite and chlorite–carbonate–pyrite alteration. The host rocks are a succession of thick bedded, andesitic scoria deposits, intruded by basaltic, andesitic and dacitic sills. Intense mineralization occurs preferentially in the volcaniclastic facies, suggesting a permeability control on the passage of mineralizing hydrothermal solutions and precipitation of sulfides.

Pumiceous water-settled fall deposits. Another pumiceous felsic facies is interbedded with the pumiceous gravity flow deposits described above at the

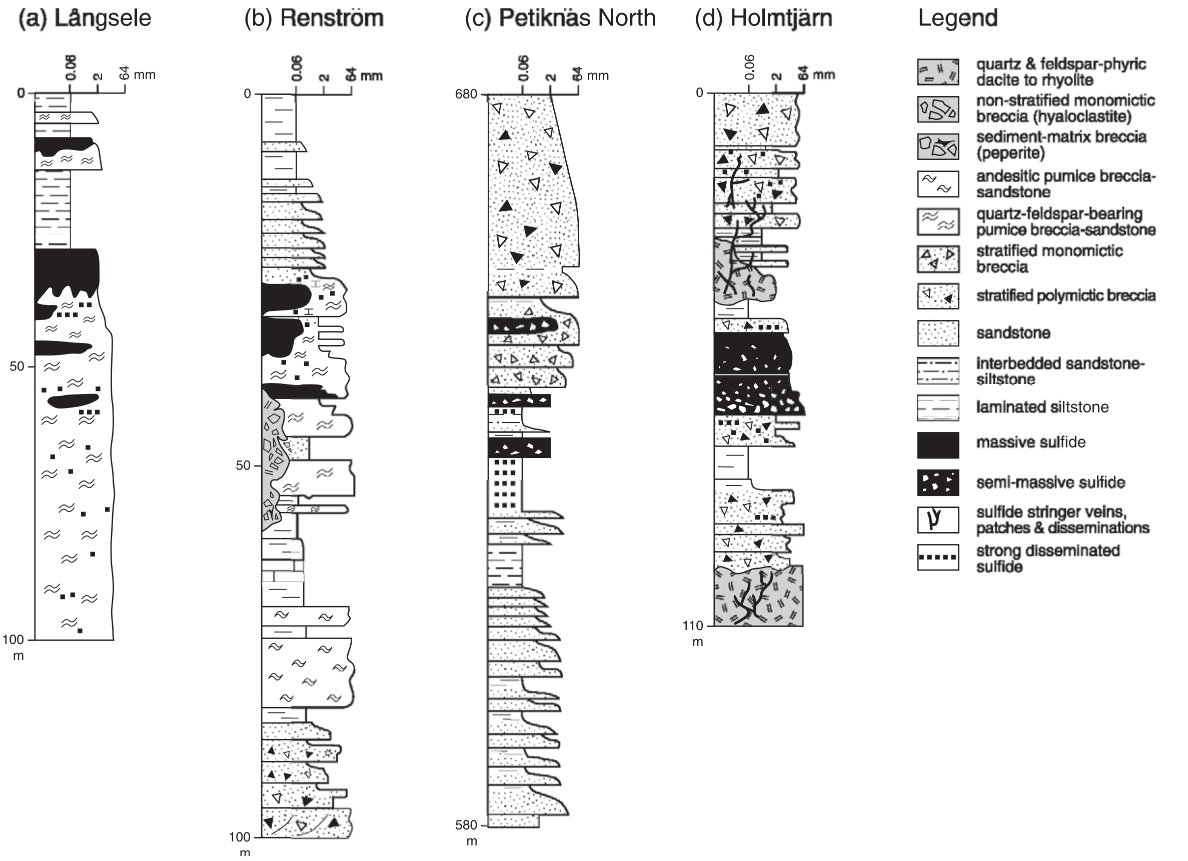
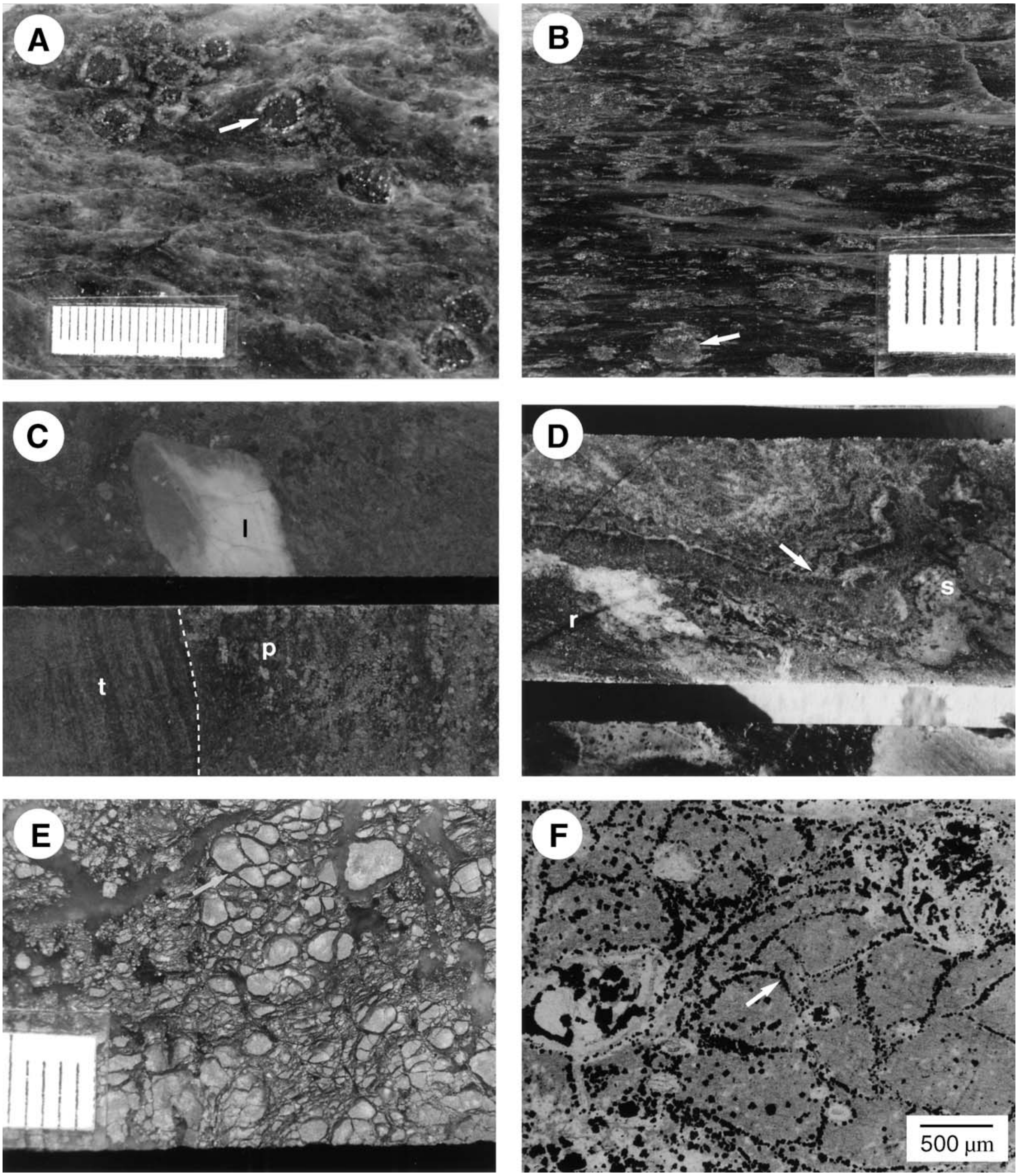


Fig. 6. Simplified stratigraphic columns showing the distribution of lithofacies and replacement ores for some VMS deposits in the Skellefte District, Sweden. The columns are single or composite diamond drill hole logs (Långsele: Lle220; Renström: 884G, 1377, 1243, 1036; Petiknäs North: 120R, 120RD4, 49R, 74R, 78R; Holmtjärn: 44, 336, 340). Modified after Allen et al. (1996b).

Renström and Kyrkvägen deposits (Allen et al., 1996b; Allen and Svenson, 1999) and also occurs in parts of the Rosebery and Hercules host successions. This facies comprises monomictic, thick (5–50 m), massive to graded beds of matrix-supported pumice blocks (up to 1 m diameter) in juvenile ash–sandstone/siltstone matrix. The grading is expressed by a decrease in the abundance of pumice blocks in the

Fig. 7. Examples of replacement and infiltration ores from VMS deposits. (A) Spotty sphalerite with pyrite and/or sericite rims (arrow) within altered rhyolitic pumice breccia-sandstone, from the spotty sulfide zone of the South Hercules deposit, Tasmania. Scale divisions are 1 mm. (B) Spotty sphalerite (arrow) within strongly foliated sericite–chlorite-altered pumice breccia from the margin of the Carrington lode, Liontown deposit, Queensland. LLD 114, 184 m (drill hole number; depth from collar). (C) A series of normally graded, sericite–chlorite-altered tuffaceous sandstone beds with coarse bases that have been selectively replaced by pyrite–chlorite (p). The semi-massive sulfides contain sphalerite microveinlets and relic lithic clasts (l). Replacement fronts (dashed line) cut lamination in the fine-grained sandstone top (t) of the bed. Core 35 mm wide. G96-184, 42.8–45 m, Golden Grove, Western Australia. (D) Finely banded pyrite–quartz–barite–sphalerite (arrow) replacing peperite at the margin of the Reward pipe. Rhyolite clasts (r) are pervasively chloritized and the siltstone (s) is silicified. Core is 5 cm wide. REMM 116, 182–190 m; Reward, Queensland. (E) The formerly glassy groundmass of this perlitic dacite has been near totally replaced by pyrite. Relic perlitic fractures (arrow) are delineated by quartz. SSD 14, 280.5 m; Sulfur Springs, Western Australia. (F) Lower contacts of the massive sulfide in (E) are marked by a zone of disseminated and spotty sulfides. Perlitic fractures in the dacite are outlined by pyrite (arrow), which has partially replaced the formerly glassy groundmass. Perlite kernels are now quartz. SSD 14, 288.2 m; Sulfur Springs, Western Australia. Plane polarized light.

upper part of each bed and normal grading of the matrix at the top of the bed. This facies is commonly more heterogeneous than the gravity flow deposits. Grain size of the matrix and blocks, abundance of blocks, and development of planar stratification in the matrix, vary irregularly up through the bed. The facies is attributed mainly to suspension settling (fallout) from heterogeneous, billowing, subaqueous clouds of



pumice and ash (subaqueous pyroclastic plume) (Allen et al., 1996b; Allen and Svenson, 1999). A few of the beds of pumice blocks in the Renström-Kyrkvägen succession have a reworked tuffaceous or grey mudstone matrix. These beds are attributed to pumiceous lava blocks that floated off the pumiceous carapace of a rhyolite dome and then became water-logged and sank (Allen and Svenson, 1999). Some of the Renström and Kyrkvägen ore lenses occur within the pumice block facies that is interpreted to be water settled pyroclastic fallout. These ores are similar to, and show the same relationships with their host facies, as the ores hosted by pumiceous gravity flow deposits described above. The rapid emplacement of the host facies indicates that these ores also formed by subsea-floor replacement.

A similar association of sediment-matrix pumice block facies, water-settled ash fall, jasper and pelagic siltstone, hosts the Mount Morgan Cu–Au pipe. The sediment-matrix pumice block facies is interpreted to record delayed settling of pumice blocks that entered suspension from pumiceous domes or from pumiceous mass flows that deposited the footwall pumice breccia–sandstone facies (Messenger et al., 1997). A quartz–sericite–pyrite alteration pipe and stringer zone extends more than 700 m below the massive ore within a rhyolitic cryptodome (Messenger et al., 1998). The ore deposit is interpreted to have formed by replacement along a syn-volcanic fault (Taube, 1986; Messenger et al., 1997, 1998; Ulrich et al., 2002). The sea-floor position during the early stage of mineralization is marked by laminated and bedded sulfide at the top of the pipe and overlies early-formed, barren massive sulfide mineralization that formed below the sea floor (Taube, 1986; Messenger et al., 1997, 1998; Ulrich et al., 2002). Later Au–Cu-rich quartz–chalcopyrite–pyrite stockwork veins are ascribed to a magmatic-fluid-dominated overprint associated with some sub-volcanic phases of the Mt. Morgan Tonalite (Ulrich et al., 2002).

4.3.1.2. Lithic volcaniclastic rocks. In this deposit style, the massive sulfide lenses are hosted by lithic-rich volcaniclastic facies, including mass flow breccia, mass flow conglomerate, sandstone turbidites and siltstone (Fig. 1G–I). Bedforms indicate deposition mainly from subaqueous gravity flows, including debris flows and high- and low-concentration turbidity currents. Examples of this deposit style include the Holmtjärn, Petiknäs North (Allen et al., 1996b), Matsuki (Kuroda, 1983), Coniagas (Doucet et al., 1998), Kidd Creek (Barrie et al., 1999; Hannington et al., 1999), Horne H lenses (Kerr and Mason, 1990; Kerr and Gibson, 1993), Ansil (Galley et al., 1995), Thalanga (Hill, 1996), Benambra deposits (Allen and Barr, 1990; Allen, 1992; Bodon and Valenta, 1995), and parts of Rosebery (Allen, 1994a), Hercules (Khin Zaw and Large, 1992) and Gossan Hill (Sharpe and Gemmell, 2001, 2002).

In common with the deposits formed by replacement of pumiceous host facies, these deposits have local strong hanging wall alteration and sulfide veins/disseminations, and show other evidence of replacement. The main difference is that the deposits commonly exhibit more obvious host rock relics within the ores, presumably because the lithic-rich rocks were more resistant to complete replacement than the glassy, chemically unstable and porous pumice deposits.

Syn-eruptive lithic breccia–conglomerate–sandstone. The Petiknäs North deposit comprises several thin sulfide lenses in the coarse-grained, lithic-rich bases of normal-graded mass flow breccia beds (Fig. 6c; Allen et al., 1996b). Various relic rock fragments (up to 20 cm diameter), enclosed by, or incompletely replaced by, sulfide occur within the ore lenses. The mineralized beds occur at the base of a thick package (300 m) of mass flow beds. The lower beds of the package are lithic-rich and polymictic with two or three dominant clast types, whereas the upper beds have lithic-rich bases and pumiceous tops. The similar composition of successive beds, the progressive changes up through the package of beds, the large volume of material, and the juvenile pumice component in the upper part of the package, suggest that the whole mass flow package was generated by pyroclastic eruptions (i.e., is syn-eruptive). Black, pyritic mudstone occurs directly below the lower sulfide lens and could represent barren pyritic sea-floor sulfide that was deposited early in the hydrothermal event. However, all of the massive to semi-massive sulfide occurs in the overlying rapidly emplaced clastic rocks and is therefore attributed to replacement of these beds.

The Coniagas deposit, Abitibi belt, Quebec comprises four sulfide lenses that are enclosed within a thick (37 m) massive felsic breccia facies (Fig. 3b). The breccia comprises dense and pumiceous juvenile clasts, and subordinate mafic fragments, that are interpreted as reworked pyroclastic products (Doucet et al., 1994). The poorly sorted and diffusely bedded character of the facies suggests rapid accumulation from weakly pulsing syn-eruptive gravity flows (principally debris flows), and indicates that the ores formed by subsea-floor replacement (Doucet et al., 1994, 1998). The footwall chlorite–sericite ± epidote ± spessartine alteration zone and hanging wall quartz–sericite ± epidote ± chlorite alteration zone are enclosed within the same depositional unit (Doucet et al., 1998). Doucet et al. (1998) interpret a spessartine-bearing zone at the base of the sulfide lens, as the metamorphic equivalent of Mn-rich hydrothermal alteration assemblages, rather than the palaeosea-floor where Mn-rich sediment and massive sulfide accumulated (cf. Bousquet and Val d'Or Camps). If this interpretation is correct, the minimum depth below the sea floor that replacement occurred is the distance from the sulfide body to the top of the enclosing depositional unit; approximately 30 m.

Syn- to post-eruptive lithic breccia–conglomerate–sandstone. The Ansil, Los Frailes-Aznalcóllar, Wilga-Currawong, Gossan Hill, Hercules, Rosebery and Matsuki VMS deposits comprise one or more stratabound sulfide lens, hosted by a variety of clastic facies including interbedded tuffaceous sandstone and siltstone (Fig. 5d–f). The sandstones are mainly normal graded, volcanic or non-volcanic sandstone facies (turbidites). The interbedded siltstones are the fine-grained tops of turbidites and hemipelagic sediment (Tables 1 and 2). The stratigraphic position of the sandstone–siltstone facies varies. For example, at Currawong, the lower part of the ore lens system occurs in the sandstone–siltstone facies, whereas the uppermost ore lies in resedimented basaltic hyaloclastite breccia (Allen, 1992; Bodon and Valenta, 1995). At Rosebery and Hercules, the lower ore lenses occur in a coarse syn-eruptive pumice deposit (see above) whereas the uppermost lenses occur in resedimented crystal–pumice–lithic sandstones (Allen, 1994a). The Gossan Hill deposit consists of two separate ore zones that are enclosed within syn- to post-eruptive crystal–pumice–lithic sandstone–siltstone facies, and connected by a stringer zone that cuts the intervening thick, syn-eruptive, pumice facies (Sharpe and Gemmell, 2002). At Matsuki, pyrite–barite-rich disseminated and vein ores (C ore deposits) are hosted in pumice breccia–sandstone beds, whereas massive and semi-massive pyrite–chalcopyrite–sphalerite–galena ± barite ± clay ores occur at the contact with, and enclosed within, the overlying tuffaceous mudstone (B and A ore deposits respectively) (Kuroda, 1983).

In all deposits, the sulfide lenses consist of massive, semi-massive, and banded sulfides, interleaved with strongly hydrothermally altered host rocks that contain disseminated, spotty and vein sulfides. Sulfide stringer veins extend below and locally above (Ansil, Currawong, Gossan Hill, Matsuki), the ore lenses. Parts of some ore lenses contain domains of strongly altered host rock (Ansil, Los Frailes-Aznalcóllar, Gossan Hill, Matsuki) or high matrix gangue contents (Rosebery, Currawong). Remarkably, the Matsuki A ore deposits and associated hydrothermal alteration zones even include sulfide pseudomorphs of benthonic foraminifera (Kuroda, 1983). Partly replaced beds are common (Fig. 7C) and bedforms (grading, flames; Galley et al., 1995; Bodon and Valenta, 1995) of the replaced host rocks are locally preserved. In the interbedded sandstone–siltstone facies, infiltration and replacement occurred preferentially in the coarser grained, rapidly deposited beds, rather than the intervening siltstones. The coarser beds were presumably more permeable.

Replacement fronts occur at the margins and tops of some of the ore lenses (Fig. 1G; Kuroda, 1983; Khin Zaw and Large, 1992; Allen, 1992; Bodon and Valenta, 1995; Sharpe and Gemmell, 2002) and strong hydrothermal alteration extends variable distances above the deposits (~ 8 m at Matsuki, Kuroda, 1983; 20 m at Currawong, Bodon and Valenta, 1995; >200 m at Gossan Hill, Sharpe and Gemmell, 2001; 3–500 m at Ansil, Galley et al., 1995). Hydrothermal alteration zones may include carbonate and siliceous nodules/lenses that formed by replacement (Matsuki, Gossan Hill, Hercules, Rosebery). The stacked Currawong lenses displays a repeated (cyclic) mineralogical zonation comprising a central pyrite zone passing upward and downward through pyrite–Zn ores into pyrite–Pb–Zn ores (Bodon and Valenta, 1995). Other deposits (Ansil) display textural evidence for progressive Cu-rich sulfide and magnetite replacement of earlier sphalerite-mineralized tuffaceous beds (Galley et al., 1995). In contrast, massive sulfide partially replaced early subsea-floor magnetite zones in the Gossan Hill deposit (Sharpe and Gemmell, 2002).

The Holmtjärn and Kidd Creek host facies are interpreted as post-eruptive, volcaniclastic sedimentary facies. Relics of the clastic host facies within, and especially at the margins of, the ore indicate that a major part of the ore deposits formed by replacement (Fig. 6d). At Kidd Creek, the occurrence of sulfide clasts of distinctive composition in some conglomerate beds, and of sulfide turbidite layers with load casts and flame structures, demonstrate that parts of a sulfide deposit were resedimented during formation of the Kidd Creek succession and imply that some sulfide accumulation occurred on the sea floor. Mass-wasting of sulfides and volcaniclastic detritus from one part of the deposit was followed by infiltration and replacement of the resedimented deposits by latter hydrothermal precipitates (Hannington et al., 1999). Sea-floor sulfide accumulation probably occurred during extended breaks in the mass flow sedimentation. These breaks are marked by argillite beds.

In situ and resedimented autoclastic rocks. The giant H massive sulfide orebodies at the Horne deposit, Noranda, and the Central and Eastern lenses at Thalanga, Australia occur in coarse-grained breccia beds of non-vesicular and pumiceous clasts that are interpreted to be mainly resedimented autoclastic debris shed from nearby rhyolite lava flows/domes (Fig. 1H; Kerr and Mason, 1990; Kerr and Gibson, 1993; Hill, 1996). The geometry of the orebodies, the occurrence of chloritized relics of rhyolite breccia within massive sulfide (Fig. 5g–h), and the truncation of stratigraphic units against the margins of the lenses, all suggest that the sulfide deposits grew by replacement of the rhyolitic volcaniclastic rocks and preexisting sulfides below the sea floor (Kerr and Mason, 1990; Hill, 1996). The H lenses are strongly zone-refined, which has been ascribed to the strongly insulated subsea-floor growth history of the deposit (Kerr and Mason, 1990; Kerr and Gibson, 1993). It is possible that the uppermost part of the orebodies lay at, or near, the sea floor, as the bodies grew downward. However, other sulfide lenses occur at higher stratigraphic positions and indicate that massive sulfides continued to form after burial of the H lenses. At Thalanga, the presence of massive sulfide clasts in breccia–sandstone beds overlying some ore lenses, and examples of barite ± sulfide and quartz–magnetite lenses between single rapidly emplaced units, also suggests that massive sulfides locally formed on the sea floor, and that replacement occurred at shallow depths below the subsea-floor (Hill, 1996). Replacement and zone-refining of the Thalanga and H lenses could have continued in the subsurface after complete burial of the deposits.

Subaqueous fire-fountain deposits. The Turner-Albright sulfide deposit, Oregon and the Potter and Corbet deposits, Canada are hosted by thick units of massive to bedded andesitic or basaltic breccia. These volcanic rocks are interpreted to be subaqueous fire-fountain (± Strombolian) deposits (Zierenberg et al., 1988; Gibson et al., 1993; Gibson and Gamble, 2000). Bedforms suggest rapid aggradation by fallout directly from a subaqueous eruption column, and/or from syn-eruptive gravity flows (Fig. 1I). The bulk of the massive to semi-massive sulfide lenses at Turner-Albright and Potter mines are interpreted to have grown by infiltration and replacement of the hyaloclastite–breccia matrix, and to a lesser degree, by replacement of clasts. The Corbet deposit comprises stacked sulfide lenses linked by stringer zones, strong hydrothermal alteration, and sulfide impregnated and replaced volcaniclastic facies (Gibson et al., 1993). Sulfide clasts occur in volcaniclastic mass-flow deposits overlying or flanking the massive sulfide ores (Potter, Corbet), or in the uppermost part of the massive sulfide deposit (Turner-Albright), and these stratigraphic intervals also contain facies of slow accumulation rate (mudstone, chert, sulfide beds). Consequently, it is likely that the uppermost parts of these VMS deposits formed on the sea floor. At Corbet, slumped and transported clastic sulfides formed a significant part of the ore (Gibson and Watkinson, 1990).

4.3.1.3. In volcaniclastic rocks directly below lavas or intrusions. There are several scenarios where massive sulfides can be directly in contact with lavas or syn-volcanic intrusions: (1) subsea-floor replacement below lavas, sills or cryptodomes (Allen, 1994a; Sinclair et al., 2000; Galley et al., 1995), (2) subsea-floor replacement within lavas or shallow intrusions (Doyle and Huston, 1999; Montelius et al., 2000), (3) burial of sea-floor sulfide deposits by lavas (Sato et al., 1979; Haymon et al., 1993; Knucky et al., 1982;

Gibson and Watkinson, 1990; Kerr and Gibson, 1993) and (4) massive sulfide that has been intruded by lava or a shallow intrusion (Lydon, 1984b; Kerr and Mason, 1990; Kerr and Gibson, 1993; Gibson and Gamble, 2000; Sharpe and Gemmell, 2002). Distinguishing the alternatives relies on carefully documenting the contact relationships and massive sulfide textures.

A massive sulfide lens can be attributed to replacement directly below a lava or intrusion if there is evidence that: (1) the sulfide lens grew by replacement of the host rocks, (2) strong hydrothermal alteration extends into the base of the overlying lava or intrusion, and (3) contact relationships indicate that the lava or intrusion was emplaced into the volcano-sedimentary succession before the sulfide lens formed (Fig. 1J). The K-lens at Rosebery (Fig. 5a), the Upper Zone lenses at Battle Mine (Fig. 3a) and some early-stage ores at Ansil Mine have been attributed to formation in this context (Allen, 1994a; Galley et al., 1995; Sinclair et al., 2000). In the case of the Rosebery K lens, the hanging wall rhyolite has a peperitic upper contact and a strongly hydrothermally altered base against the massive sulfide. Consequently, the rhyolite is an intrusion that was emplaced into unlithified clastic rocks before or during subsea-floor growth of the sulfide lens (Allen, 1994a). At Ansil, the basal autoclastic breccia facies of the hanging wall andesite is silicified and mineralized (Galley et al., 1995). In both cases, the lava or shallow intrusion is interpreted to have provided a cap-rock under which the sulfide body grew downwards by replacement of clastic host facies.

4.3.2. Subsea-floor replacement deposits within lavas and intrusions

Some massive sulfide deposits are hosted by coherent lava or intrusion facies and their associated autoclastic breccias (Fig. 1K). Examples include the Maurliden deposits, Sweden (Montelius et al., 2000); upper parts of the Ansil deposit, Canada (Riverin et al., 1990; Galley et al., 1995); and Highway-Reward (Large, 1992; Doyle and Huston, 1999), and parts of the Sulfur Springs deposit (Morant, 1995, 1998) in Australia. It is quite incredible that massive sulfide can form by replacement of originally coherent volcanic rock. However, careful application of the criteria listed above for distinguishing replacement sulfide bodies has demonstrated that in some special circumstances it does occur (Doyle and Huston, 1999; Montelius et al., 2000).

The facies architecture of the host volcanic centers is diverse, and includes intrusion-dominated volcanic centers (Highway-Reward, Maurliden), thick lava piles (upper part of Ansil), and mixed intrusion-lava-volcaniclastic successions (Sulfur Springs). The massive sulfides are stratabound and localised along contacts with overlying volcano-sedimentary facies (Fig. 1K; Fig. 5c, Sulfur Springs; West and East Maurliden), or are discordant to local bedding (Fig. 5i, Highway-Reward; North Maurliden) and occur within, or at the margins of, the lavas, sills, cryptodomes or partly extrusive cryptodomes (Doyle and Huston, 1999; Montelius et al., 2000; Doyle, 2001). The geometry of the Ansil middle-stage ores is more complex. Along the top of the semiconformable pyrrhotite–chalcopyrite–chlorite lens, a discordant sulfide spine cuts the hanging wall andesite and forms the keel to an overlying mineralized alteration zone. The margins of the hanging wall alteration and stockwork zone branch out along andesite hyaloclastite-rich interflow contacts and pass outwards from pyrrhotite–chalcopyrite–chlorite rich to sphalerite–pyrrhotite–sericite rich alteration assemblages (Galley et al., 1995). Probably in all cases, hydrothermal solutions were focused into the lava or intrusion by syn-volcanic faults, and spread out and replaced the lava or intrusion via permeable intense hydrothermal breccia zones, autoclastic breccia zones, or chemically unstable or reactive zones such as glassy margin facies.

The replacement ores contain relics (pseudoclasts) of the autoclastic or coherent host rocks (Fig. 7D). The Maurliden massive sulfides also contain relict quartz phenocrysts of the host intrusion (Montelius et al., 2000). Infilling of pore space, or replacement by sulfide minerals can locally preserve delicate volcanic textures (perlite, quench fractures), implying that the pre-existing volcanic-related fracture pattern influenced fluid flow during mineralization (Fig. 7E–F).

The massive and semi-massive sulfides are typically enclosed within a zoned hydrothermal alteration envelope that can include veins, disseminations, and locally stratabound sulfide/sulfate lenses. Hydrothermal alteration zones and vein or disseminated sulfides are most strongly developed in the footwall, but also extend a few tens of metres (Highway-Reward: Doyle and Huston, 1999; Doyle, 2001) or >500 m (Ansil: Galley et al., 1995) into the hanging wall.

4.3.3. Subsea-floor replacement deposits within limestone

VMS deposits hosted by limestone include: (1) hydrothermal replacements of sedimentary limestone facies (Garpenberg deposits, Sweden; parts of the Henty-Mt. Julia deposits, Tasmania) and (2) replacements of carbonate-rich exhalites (Chisel Lake, North Chisel, Ghost Lake and Lost Lake deposits in Snow Lake district; the Errington-Vermilion deposits, Sudbury; Lynne deposit, Wisconsin).

The Garpenberg deposits and other similar deposits in the Palaeoproterozoic Bergslagen region of central Sweden are hosted within and adjacent to regionally extensive sedimentary limestones that locally preserve stromatolite structures. The limestones are interpreted as marine biogenic carbonate facies unrelated to exhalative hydrothermal activity (Allen et al., 1996a). Carbonate exhalites (see above) that are spatially associated with other limestone-hosted VMS deposits (Chisel Lake deposits; Errington-Vermilion deposits; Lynne deposit) are interpreted as precipitates from the discharge of hydrothermal fluids at the seafloor. Progressions to Fe-rich carbonate–skarn–sulfide assemblages, reflect an overprinting of the seafloor precipitates by focused, higher temperature mineralizing fluids as the system evolved towards the thermal maximum (Galley et al., 1993; Gray and Gibson, 1993; Stoness et al., 1993; DeMatties, 1994; Bailes and Galley, 1996; Galley and Ames, 1998).

Both deposit styles are generally associated with footwall sequences dominated by originally permeable clastic strata (mainly tuffaceous breccia–sandstone facies). The Zn–Pb–Ag–Cu–Au deposits comprise massive to semi-massive lenses or pods, veins, and disseminations of pyrite–sphalerite–chalcopyrite–galena. The geometry of the orebodies, location of the largest orebodies in altered limestone (skarn, dolomite) and in volcaniclastic rocks directly beneath limestone, interleaving of sulfide–carbonate–skarn assemblages, gradations from Mg-rich skarn to dolomite to unaltered limestone (calcite marble) around the ore zones, and in some cases (Garpenberg district), localization of the best mineralized zones along contacts between the limestone and volcaniclastic rocks, all suggest that the sulfide deposits grew below the sea floor by replacement of reactive limestone strata and adjacent volcaniclastic rocks (Fig. 1L).

The Lynne and Chisel Lake deposits have an asymmetric alteration pattern, with strong semi-conformable to discordant alteration in the footwall rocks and weaker alteration in the hanging wall, with or without, strong hydrothermal alteration and disseminated or patchy sulfide at the top of the sulfide body (Galley et al., 1993; DeMatties, 1994; Bailes and Galley, 1996). This pattern suggests that the top of the replacement zone occurred at, or very close to, the seafloor. At Vermilion, footwall carbonate–silica–chlorite zones transgress the immediate host (exhalite) and overlying turbidites, suggesting replacement at deeper levels. Gray and Gibson (1993) suggest that the silicified turbidites acted as a cap rock, promoting thermal/chemical evolution of fluids into the mineralizing window, with consequent subsea-floor replacement of carbonates by base metal sulfides.

The Garpenberg succession, including the alteration zones, has been strongly deformed and metamorphosed to amphibolite facies. The Mg-rich alteration zones in the limestone are now dolomite and skarn, and the Mg- and K-rich alteration zones in the adjacent volcanic rocks comprise phlogopite–biotite–garnet and quartz–muscovite assemblages. Hydrothermal alteration is strongest and most extensive in the footwall volcanic rocks and base of the limestone, and diminishes upwards. However, the basal 50 m of the hanging wall pumiceous mass-flow unit is silicified over an extensive area and contains local zones of strong footwall-style alteration, implying that strong hydrothermal activity continued during emplacement of the hanging wall pumice breccias. Allen et al. (1996a) suggest that the ores and associated hydrothermal system may be genetically related to magmatic and volcanic processes that accompanied eruption and emplacement of the hanging wall pumice breccia. The majority of the footwall felsic volcanic rocks and the hanging wall pumice breccia are interpreted to have been deposited in a subaqueous (below wave base) environment. The host limestone is attributed to deposition both below and above storm wave base, but in the photic zone, and some associated volcaniclastic facies are attributed to above-storm-wave base depositional environments. The ore deposits thus formed in the subsea-floor, either under relatively shallow water conditions, or later after subsidence to deeper water conditions, or both (Allen et al., 1996a).

The Henty-Mt. Julia deposits comprise gold-rich sulfide veins, disseminations and semi-massive lenses within a stratabound quartz–sericite alteration zone that locally overprinted bedded carbonates and carbonate-impregnated host-rocks (Callaghan, 2001). The deposits have strong hanging-wall alteration up to 100 m above the mineralized horizon, widespread replacement textures, and do not have skarn-assemblages. These deposits could be a more gold-rich and lower metamorphic grade (lower greenschist facies) variant of the Garpenberg-type deposits.

Sulfide deposits not hosted by limestone, but with abundant calc-silicates include the Ansil deposit, Noranda, where early-formed replacement massive sulfides were replaced by late-stage magnetite. Host rock relics occur with the ore, and epidote and hedenbergite–andradite skarn assemblages occur below and along contacts with massive sulfide. Galley et al. (1995) and Galley and Ames (1998) attribute these relationships to the overprinting and telescoping of an early sulfide-rich hydrothermal system by a later high-temperature, magmatic-dominated hydrothermal phase, sourced from an underlying syn-volcanic intrusion. In some other deposits, carbonate-rich alteration zones have been metamorphosed to calc-silicate-bearing rocks (e.g., West Thalanga; Herrmann and Hill, 2001).

5. Replacement and the favourable horizon

The favorable horizon (Sangster, 1972), the horizon within a stratigraphic succession that massive sulfide deposits preferentially occur on, can be represented by sulfide clast-bearing volcaniclastic mass-flow deposits, pyritic mudstone, silica–pyrite or silica–iron oxide rocks, or carbonate-rich rocks (Saif, 1983; Peter and Goodfellow, 1996; Knucky et al., 1982; Duhig et al., 1992; Herrmann and Hill, 2001). Some of the iron oxide, silica-rich or carbonate-rich layers are exhalites and can provide evidence for a sea-floor exhalative origin for the massive sulfides. However, contact relationships and relict textures indicate that similar rocks can also from by subsea-floor replacement of pre-existing strata (see discussion above). In this case, they represent stratabound alteration zones that envelope or occur near the stratigraphic position of the ore deposits. They define a favorable stratigraphic interval rather than a specific horizon.

Silica-iron oxide replacements in the stratigraphic package hosting the Highway-Reward deposit and surrounding prospects occur as thin (centimetres to metres) stratabound or discordant lenses, pods and veins (metres to tens of meters long). The lenses can be hosted by a variety of volcanic and sedimentary facies, including water-settled fall beds, pumiceous gravity flow deposits, autoclastic breccia units, or coherent parts of lavas and syn-volcanic intrusions (Doyle and McPhie, 2001). Contacts between the quartz–hematite ± magnetite rocks and the host rocks are sharp, or else there are lateral and vertical replacement fronts passing from massive quartz–hematite ± magnetite, through quartz–hematite ± magnetite-altered coherent or volcaniclastic rocks, into phyllosilicate–hematite-altered rocks. The lenses are massive or semi-massive, with relic clasts or pseudoclasts of the host unit. Partly replaced beds are common, and lamination in the enclosing altered host lithofacies can be traced into the lenses (Fig. 8A–B). Major and trace element concentrations in the silica–iron oxide replacements reflect the composition of the precursor volcanic facies. The distance below the sea floor that replacement occurred is difficult to interpret. The best-constrained examples, where jasper clast-bearing mass-flow deposits overlie host rocks to the silica–iron oxide lenses, suggest that the palaeosea-floor was 10–25 m above the base of the lenses.

In cases where the favourable stratigraphic interval is marked by carbonate-altered rocks, these are hosted by pumice deposits and tuffaceous sandstones (e.g., Rosebery, Hercules, Snake Oil, Renström, Rakkejaur, Rävliden), or less commonly they occur at the top of thick lava piles (e.g., Thalanga). Some of the carbonate-altered zones are the products of early, regional-scale, hydrothermal systems that in some cases were overprinted by later, higher temperature systems (see above; Sturgeon Lake: Franklin et al., 1981; Morton and Franklin, 1987). Other examples are restricted to the margins of massive sulfide lenses (Rosebery, Hercules). In these cases, precipitation of sulfide

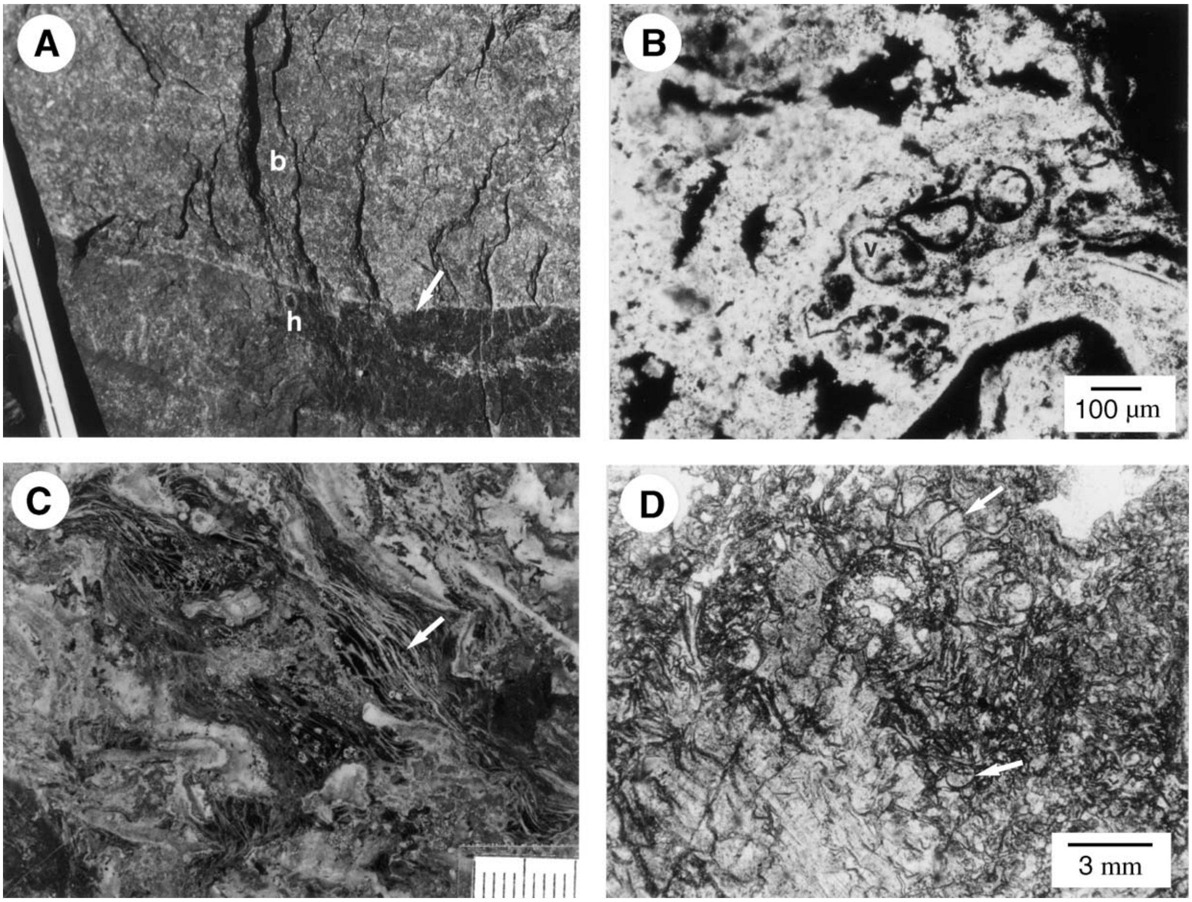


Fig. 8. (A) Upper margin (arrow) of a quartz-hematite lens (h), Trooper Creek prospect area, Mount Windsor Subprovince, Queensland. Spotty quartz-hematite has replaced diffusely stratified pumice breccia (b). Relic pumice fragments and shards are preserved in the lens. (B) Photomicrograph from a quartz-hematite lens replacing pumice breccia. A relic glass shard and contained vesicles (v) have been rimmed by hematite (dark) and replaced by quartz. Trooper Creek prospect area. Plane polarized light. (C) At the margins of a small sphalerite-galena deposit at Snake Oil prospect, massive and semi-massive spotty carbonate–hematite–sericite has replaced the host pumice breccia. Relict tube pumice textures (arrow) are preserved. Drill core SODD 002, 230.8 m; Snake Oil prospect, Mount Windsor Subprovince. (D) Moderately well preserved pumice from within the Rosebery massive sulfide deposit, Tasmania. Preservation of pumice is attributed to replacement of former glassy vesicle walls and infilling of vesicles by carbonate, quartz and feldspar, which were resistant to diagenetic compaction and tectonic deformation. Vesicles are outlined by phyllosilicate minerals that are probably the metamorphic equivalent of early clay minerals. Plane polarized light.

minerals within high-temperature upflow zones was probably accompanied by deposition of carbonate alteration facies in the cooler, lateral and upper margins of the system (Khin Zaw and Large, 1992). Accordingly, the spectrum of carbonate-altered host rocks associated with VMS deposits may encompass true subsea-floor replacements, examples of carbonate-impregnated rock below exhalites, as well as complex systems with stacked sea-floor and subsea-floor carbonate lenses. The subsea-floor replacements are characterized by relict textures or geochemical signatures of the host lithofacies (Fig. 8C–D) (e.g., Khin Zaw and Large, 1992; Orth and Hill, 1994; Allen et al., 1996b; Herrmann and Hill, 2001; Callaghan, 2001). In some cases, the carbonate-rich assemblage may have provided a permeability barrier or reactive host for subsequent subsea-floor sulfide precipitation.

6. Discussion

6.1. Importance of lithofacies analysis

Deciphering how a volcanic-hosted sulfide body formed depends partly on identification of the character and emplacement mechanism of the host facies, and on the recognition of any paleoseafloor positions at which mineralization was focused. Zone-refining, veining, deformation and metamorphism may obliterate primary sulfide textures and structures diagnostic of the original sulfide deposition. In other cases, the original sulfide textures may not be diagnostic of a specific style of sulfide deposit. Consequently, the origin of the immediate host lithofacies is a critical constraint as to whether the mineral deposit formed at the sea floor or subsea-floor. As a rule, massive sulfides hosted within rapidly emplaced sedimentary or volcanic facies can only have formed by impregnation and replacement (Allen, 1994a; Allen et al., 1996b). However, both the host successions and the sulfide deposits can be complex and varied, such that no one simple rule can be used to categorize all VMS deposits. It is necessary to determine both the eruptive and emplacement processes of the host lithofacies, and the nature and positions of contacts with mineralized intervals and alteration zones. The upper contacts and lateral margins of the mineral deposit provide most diagnostic information as to whether the deposit formed at the sea floor or in the subsurface. The lower contacts of both sea-floor and subsea-floor deposits can be similar and both generally show extensive evidence of replacement. It is especially necessary to identify and interpret relic volcanic textures in the mineralized zones, especially where drill core logs form the basis for reconstruction. Where drill core sections are widely spaced, critical spatial relationships (e.g., replacement fronts) may not be obvious, and interpretations are increasingly reliant on textural evidence.

Different mineral deposits vary widely in the extent to which their emplacement mechanisms can be deciphered. Most favorable are the deposits hosted by pumiceous gravity flow deposits, lavas, or syn-volcanic intrusions. Syn-eruptive gravity flow deposits sourced from explosive eruptions are typically widespread, rapidly emplaced, thick (McPhie and Allen, 1992), and traceable through zones of ore-associated hydrothermal alteration. They provide good stratigraphic markers and the contact relationships between single depositional units and ores can be determined. Host sequences dominated by lavas or intrusions also provide a good framework for reconstructing the sequence of mineralization. However, subaqueous lavas and intrusions typically involve small magma volumes and are spatially restricted, so that closely spaced observations and careful documentation of phenocryst assemblages (mineralogy, size, percentages) may be required to distinguish primary rock types and critical contact relationships (e.g., Doyle and Huston, 1999; Doyle and McPhie, 2000).

In host sequences dominated by thinly bedded deposits from suspension sedimentation and low-concentration turbidity flows, discriminating the relative importance of sea-floor sulfide accumulation and subsea-floor replacement can be very difficult. Furthermore, evidence for sea-floor sulfide accumulation may be masked by subsequent upward migration of subsea-floor replacement fronts. In deposits of this style, interpretations are dependent on detailed mineralogical and textural studies of the ores and host lithofacies, and identification of lateral and vertical replacement fronts.

6.2. Volcanic setting of subsea-floor deposits

The spectrum of VMS deposit styles described above indicates that deposits with a major component of subsea-floor replacement are quite diverse in their setting. The settings range from lava or syn-volcanic intrusion dominated, to those dominated by volcaniclastic or sedimentary rocks (Fig. 9). Nonetheless, particular host lithofacies associations, and to some degree particular volcano types, are more common than others. The most favourable rocks for hosting extensive sulfide replacement bodies are volcaniclastic rocks that had high initial porosities and especially those of originally glassy composition, such as pumice-ash deposits and the quenched margins of lavas and intrusions.

Among the volcaniclastic-dominant host sequences, associations of syn-eruptive felsic pumice breccia–sandstone units host a high proportion of known subsea-floor replacement deposits. The syn-eruptive facies are characterised by thick (10’s–100’s m)

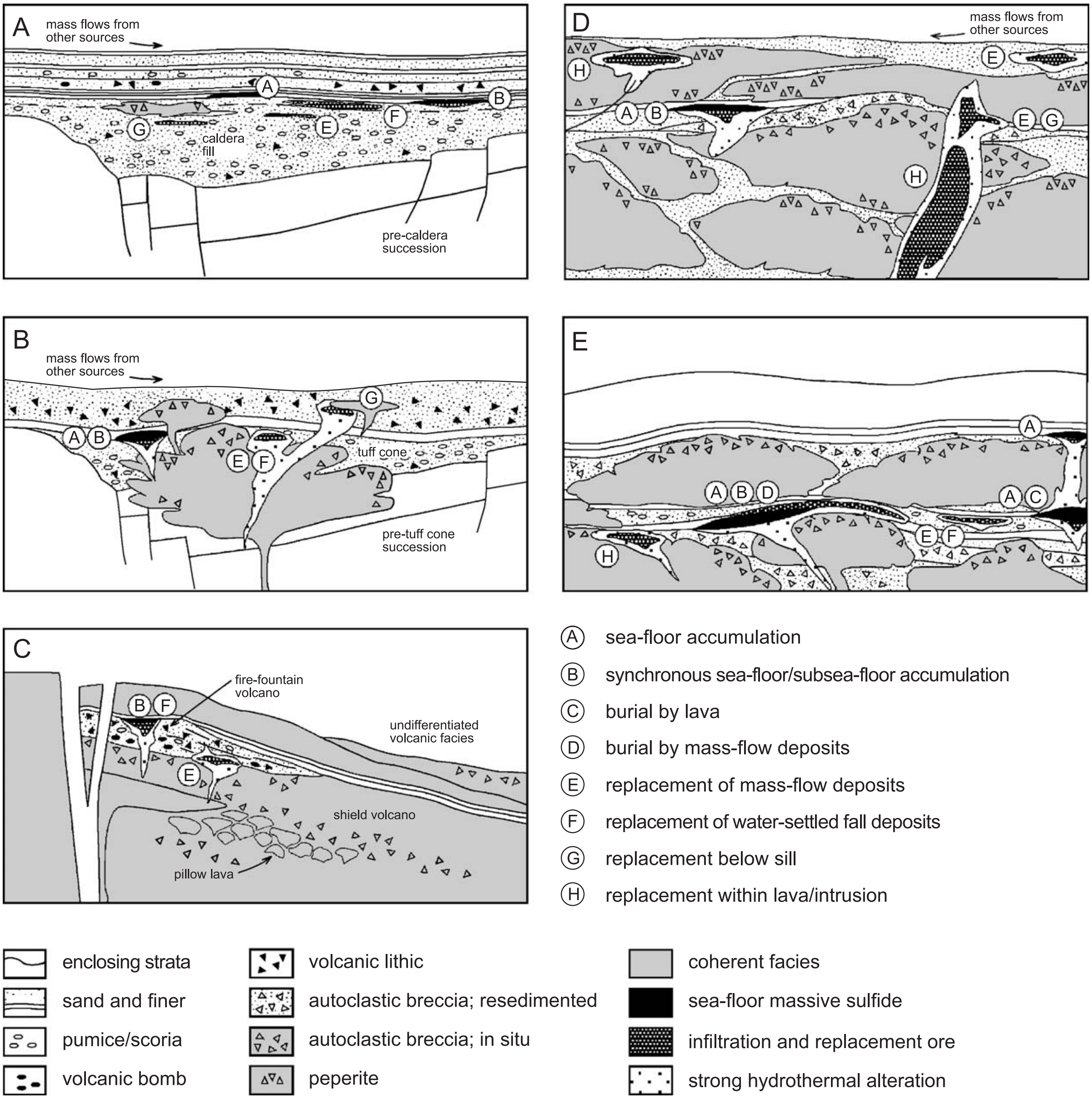


Fig. 9. Facies models for the volcanic environment of massive sulfide ores in (A) subaqueous felsic caldera volcanoes, (B) felsic cryptodome–dome–tuff cone volcanoes, (C) basalt–andesite fire-fountain volcanoes, (D) syn-volcanic intrusion-dominant volcanic centres, and (E) lava-dome complexes. Modified after Allen (1992), Allen et al. (1996b), Hill (1996), Gibson et al. (1999) and Doyle and Huston (1999).

deposits of unmodified pyroclasts sourced from explosive eruptions, and are often intercalated with lavas, sills, or cryptodomes. From analysis of the facies architecture, these successions are interpreted as the proximal–medial deposits of large submarine caldera volcanoes (e.g., Rosebery, Mattabi-Sturgeon Lake) or dome–cryptodome–tuff cone volcanoes (e.g., Renström), or medial to distal facies sourced from adjacent explosive volcanic centres (e.g., Liontown, Snake Oil). A range of massive sulfide styles may exist in any one of these volcanic settings. For example, observations at Rosebery suggest that replacement-style massive sulfides in submarine caldera settings can be hosted within thick pumice units (gravity flow deposits), near contacts with overlying thinly bedded tuffaceous siltstone–sandstone beds (tuffaceous turbidites, water-settled ash falls), or beneath syn-volcanic intrusions (sills, cryptodomes) within the pumice deposits (Fig. 9A). The massive sulfide deposits associated with subaqueous dome–cryptodome–tuff cone volcanoes typically occur within or immediately above the proximal (vent) facies association, which comprises lava domes and the thickest part of the pyroclastic deposits (Fig. 9B). The small Turner-Albright, Potter and Corbet deposits are mafic-hosted analogues that occur within the proximal to medial facies association of subaqueous volcanic cones characterised by fire-fountain and Strombolian eruptions (Fig. 9C).

The relationships of ore to lithofacies types are different in the lava- and intrusion-dominated settings. Subsea-floor massive sulfide deposits hosted by syn-volcanic intrusion-dominated volcanic centers comprise discordant pipes (Highway-Reward) or approximately bedding-parallel lenses (Maurliden) that are localized within the formerly permeable, brecciated and often glassy margins of the intrusions and lavas (Fig. 9D). Other deposits (e.g., Thalanga) are hosted in submarine volcanic centers dominated by lavas and domes, together with mass-flow-emplaced resedimented autoclastic breccia–sandstone facies derived from oversteepening of lava or dome margins. The replacement ore lenses are typically strata bound within mass-flow deposits that may fill sea-floor depressions on the top of the lava-dominated complexes (Fig. 9E).

The exact water depth at the time mineralization is very difficult to determine for VMS deposits hosted in the volcanic settings described above, but available geological and fluid inclusion data are generally consistent with sulfide accumulation in a relatively deep-water (below-storm-wave base) environment. In contrast, stratigraphic intervals with shallow water facies associations host the Garpenberg and Garpenberg Norra deposits, Sweden, and upper parts of the Henty, Mount Julia and Mount Lyell deposits, Australia (Allen et al., 1996a; Halley and Roberts, 1997; Callaghan, 2001; Corbett, 2001). Depositional environments near storm wave base are indicated by fossiliferous limestone beds. The Henty-Mount Julia deposits are hosted by a felsic lava succession intercalated with medial to distal volcanic facies and sedimentary facies (volcaniclastic sandstone, limestone) (Callaghan, 2001). The volcanic facies architecture of the Mt. Lyell deposits is poorly constrained. The Garpenberg Zn–Pb–Ag–Cu deposits are subsea-floor replacements within limestone beds that are intercalated with medial–distal facies from one large rhyolite–dacite volcano, and overlain by proximal pumice breccia facies from another felsic caldera volcano (Allen et al., 1996a).

Some subsea-floor replacement deposits display features that are transitional between VMS and high-sulfidation epithermal deposits (e.g., Mt. Lyell: Huston and Kamprad, 2001; Boliden: Bergman Weihed et al., 1996) or low-sulfidation epithermal deposits (e.g., Henty-Mt. Julia: Callaghan, 2001). These deposits are hosted within shallow-water facies associations, show evidence for extensive subsea-floor replacement mineralization, and have a metal association typical of epithermal mineralization (particularly Cu, Au, As, Sb, Te, Se) and local alumina-rich alteration. Some other deposits may be transitional between VMS and intrusion-related systems (e.g., Mt. Morgan: Ulrich et al., 2002).

6.3. Influence of porosity and permeability

Many subsea-floor replacement ores occur within tuffaceous rocks or resedimented flow breccias and hyaloclastites. Pumiceous mass-flow deposits appear to be a particularly good host for the development of economic deposits (Tables 1 and 2). The originally highly porous, permeable, water saturated and glassy nature of these lithofacies makes them favorable host rocks. Ascending hydrothermal fluids are probably focused at depth by syn-volcanic faults and/or volcanic vent structures, but loose focus and disperse within the overlying porous volcaniclastic deposits to produce widespread strata-bound alteration and lens- or sheet-style massive sulfide ores (cf. Gibson et al., 1999; e.g., Rosebery, Hercules, Liontown). Porosity near the sea floor is probably reduced early in the life of the hydrothermal system by cooling of the up-welling hydrothermal solution, and consequent precipitation of alteration minerals in pore space. This may ultimately seal the "roof" of the system and focus subsequent mineralization beneath the impermeable alteration zone, within more permeable strata (e.g., Gibson and Kerr, 1993; Galley et al., 1995; Doucet et al., 1998; Sharpe and Gemmell, 2001). Particular early-formed alteration styles such as sulfates and carbonates may also provide reactive subsea-floor hosts for the subsequent mineralizing stage (e.g., Rosebery, Hercules: Allen, 1994a). Deposit geometries and internal textural relationships suggest that the process involves the lateral expansion of alteration and mineralization fronts through receptive hosts, and the coalescence of spots, patches, and pods of sulfide, sulfate or carbonate. Mixing between the upwelling hydrothermal fluid and cold seawater is regarded as a major cause of sulfide precipitation in VMS systems, and this mixing process is no doubt strongly influenced by the permeability pattern developed by early hydrothermal alteration. In general, the mixing of hydrothermal fluids with cold seawater probably becomes less effective with increasing depth in the volcanic pile.

Within syn-volcanic intrusion and/or lava-dominated volcanic piles, fluids are focused along faults, autoclastic breccia zones, or within the fractured glassy margins of lavas and intrusions. The crystalline facies of lavas and syn-sedimentary intrusions, or intervening dewatered (indurated) sedimentary facies may act as barriers to ascending hydrothermal fluids, focusing hydrothermal fluid flow and mineralization (Kuroda, 1983; Allen, 1994a; Doyle and Huston, 1999; Doyle, 2001).

In host successions comprising complex associations of facies, permeability and porosity networks important for fluid migration may vary widely between units, so that an array of ore types and positions may be present. Argillites and carbonaceous argillites overlying volcanic-dominant sequences may act as thermal or hydrologic cap rock, or a chemically receptive host, trapping a greater percentage of the metal budget through promoting cementation and replacement (e.g., Matsuki: Kuroda, 1983).

6.4. Replacement and sedimentation rate

The rate of accumulation of strata influences the relative importance of sea-floor and subsea-floor processes in massive sulfide accumulation. If sedimentation rate exceeds the rate of accumulation of sulfide at the sea-floor, then the sulfide deposit may evolve from an initial sea floor deposit to a largely subsea-floor replacement deposit (Allen et al., 1996b). Furthermore, sulfide precipitation at sea-floor mounds may be intermittent and separated by long periods during which sediments accumulate (e.g., Rona et al., 1993). Volcanism, particularly explosive volcanism, has the potential to release large volumes of volcaniclastic detritus into submarine settings (Cas and Wright, 1991). Burial of sulfide mounds by volcaniclastic mass flows, water-settled ash fall, or lavas can occur during the life of the hydrothermal system, interrupting or terminating sea-floor sulfide deposition (Knucky et al., 1982; Gibson and Watkinson, 1990; Kerr and Gibson, 1993; Haymon et al., 1993). As the hydrothermal system attempts to advance upwards to the new sea-floor position, subsea-floor replacement of the intervening lithofacies by sulfides may become important (Large et al., 1988; Haymon et al., 1993; Humphris et al., 1995).

Although many replacement-style massive sulfide deposits may have initially started as sea-floor deposits and evolved into largely replacement-style subsea-floor deposits via downward growth from a sea-floor position or upward growth after burial, some deposits show evidence (see discussion above) that they never formed a significant sulfide accumulation at a sea-floor position and that they most likely commenced growth in the subsurface (e.g., Maurliden, Highway-Reward, Garpenberg Norra).

The literature on VMS deposits emphasizes the exploration significance of sedimentary facies that mark a long hiatus in volcanic activity for significant accumulation of sulfides (Sangster, 1972; Franklin et al., 1981; Lydon, 1984a). However, some giant VMS deposits (e.g., Kidd Creek, Horne) formed during building of the volcano-sedimentary sequence and are dominated by replacement and infiltration ores (Gibson and Kerr, 1993; Kerr and Gibson, 1993). Subsea-floor replacement may represent an effective mechanism for trapping a higher proportion of the metal budget and so generate large tonnage mineral deposits.

6.5. Depth of replacement below the sea floor

The distance below the sea floor at which infiltration and replacement take place is rarely well constrained. Best constrained are the examples where sulfide clast-bearing strata occur in the uppermost part of the sulfide deposit (many Kuroko deposits in Japan) or some tens of metres stratigraphically above the replacement ores (e.g., Rosebery-Hercules: Allen and Hunns, 1990; Allen, 1994a; Thalanga: Hill, 1996), suggesting that the top of the replacement zone probably occurred at or within a few metres of the sea floor. At Kidd Creek, clastic sulfide ores deposited at the sea floor were subsequently replaced and cemented by later hydrothermal precipitates, also implying shallow subsea-floor replacement (Hannington et al., 1999).

Conceptually it can be concluded that few if any replacement-style VMS deposits are likely to have formed entirely in the subsurface with no expression of mineralization at the sea floor, because the hydrothermal system involves the upward and outward movement of vast quantities of hydrothermal fluid that must vent at the sea floor somewhere. In general, it can be concluded that the minimum depth below the sea floor that replacement occurred is the distance from the sulfide body to the top of the enclosing volcanic or sedimentary depositional unit. In addition, the distance from the base of the massive sulfide body to the base of an overlying sulfide-clast bearing bed provides the maximum distance below the sea floor that replacement massive sulfide formed. Considering the available data for the sulfide deposits discussed in this paper, palaeosea-floor positions at the time of mineralization appear to have been within 10–500 m and mainly 10–200 m, above the base of the sulfide body.

This upper few tens to hundreds of metres in the volcano-sedimentary pile are probably the favored position for replacement because the strata are wet, porous and poorly consolidated, and at greater depths become progressively more compacted, dewatered, altered (e.g., Einsele, 1986) and less amenable to large scale replacement and infiltration by hydrothermal fluids. Ascending hydrothermal fluids meet and mix with cold seawater before reaching the sea floor, with the resultant decrease in temperature, and increase in pH and SO_4/H_2S promoting sulfide deposition. In general, the opportunities for sustained mixing between hydrothermal fluids and circulating cold seawater decrease with increasing depth in the volcanic pile, which further constrains the development of replacement-type sulfide deposits to the upper part of the volcanic pile. Facies characterized by low primary or secondary permeability and porosity, such as mudstone, crystalline zones of lavas or intrusions, zones sealed by hydrothermal alteration or by induration during magma-sediment interaction, can act as an aquiclude, enhancing mixing processes in the underlying strata, and trapping a greater percentage of the total metal budget (Kuroda, 1983; Gibson et al., 1999).

7. Conclusions

Sulfide ores which form by syn-volcanic subsea-floor infiltration and replacement of volcano-sedimentary facies can be distinguished by: (1) relics of the host rock preserved in the ore; (2) contact relationships implying the ores are enclosed within single lavas, intrusions or mass-emplaced clastic facies; (3) replacement fronts between the mineral deposit and the host rock; (4) discordance with the enclosing host lithofacies; (5) zones of strong hanging wall alteration, similar in style and intensity to footwall alteration. Criteria 1–3 are diagnostic of replacement, whereas criteria 4 and 5 may suggest replacement but are not alone diagnostic. These conclusions are only relevant to ore deposits where a syn-volcanic to syn-diagenetic timing has been demonstrated.

Based on variation in the lithofacies associations of the host succession, ore deposit geometry and textures, and arrangement of alteration assemblages, VMS deposits can be divided into 12 main deposit styles (Fig. 1). Evidence of replacement of the host rock can be found in most of these deposit styles. Some of these deposits commenced growth as sea-floor deposits and developed replacement-style subsea-floor mineralization via downward growth from a sea floor position (Horne H lens) or upward growth after burial (Amulet, Millenbach). However, at the other end of the spectrum, some deposits commenced growth in the subsurface and never formed a significant sulfide accumulation at a sea-floor position (Maurliden, Highway-Reward, Garpenberg Norra).

Subsea-floor replacement occurs mainly in volcaniclastic facies, and mainly within 200 m of the sea floor. Evidence of replacement of the host rock is least well developed in sea-floor deposits, and only weakly to moderately developed in sea-floor deposits that were modified after burial by lavas (Amulet, Millenbach, Fukazawa) or volcaniclastic deposits (Que River, Mount Chalmers, Woodlawn, and the Gopher, South Trough and Battle Main lenses, Myra Falls). Volcanic-hosted massive sulfide deposits that are dominated by subsea-floor replacement and infiltration can be grouped into seven main styles. These include deposits that formed largely subsea-floor with or without sea-floor lenses (Rosebery, Hercules, Highway-Reward, Liontown, Thalanga, Mount Lyell, Henty-Mount Julia, Mount Morgan, Sulfur Springs, Mattabi, Horne H lenses, Kidd Creek, Coniagas, Chisel Lake deposits, Errington, Vermilion, Renström, Kyrkvägen, Kankberg, Holmtjärn, Petiknäs North, Långdal, Långsele, Boliden, Maurliden, San Miguel and Salomon-Lago deposits, San Platón, Concepción, Neves Corvo, Lynne, Turner-Albright, parts of Myra Falls). Deposits hosted in stratigraphic intervals dominated by thinly bedded rocks may have accumulated at and below the sea floor (Matsuki, Los Frailes-Aznalcóllar, Ansil, Gossan Hill, Currawong, Wilga). The favorable horizon or stratigraphic interval to some of the replacement-style VMS deposits (Rosebery, Hercules, Thalanga, Highway-Reward) includes carbonate-rich lenses or quartz–hematite ± magnetite lenses which formed below the sea floor.

Subsea-floor replacement appears to be an effective mechanism for trapping a higher proportion of the total metal budget and thus contributes to the formation of large tonnage and/or high-grade VMS deposits.

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