The relation between iron-formation and low temperature hydrothermal alteration in an Archean volcanic environment


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Abstract

The uppermost section of the Hunter Mine group (HMG) (2728 Ma), a bimodal volcanic complex in the Abitibi greenstone belt, contains both oxide and carbonate facies banded iron-formation (BIF). This paper explores the relationship between volcanic activity and the development of the two types of iron-formation. The oxide facies, represented by chert-jasper-magnetite iron-formation is widespread and recurs at several stratigraphic levels. Field evidence suggests it deposited directly on the sea floor during periods of volcanic quiescence, and is most probably derived from fluids seeping from volcaniclastic sediments and the compaction of felsic shard-rich tuffs. Seeping occurred along synsedimentary and synvolcanic faults and then along bedding planes. Carbonate (siderite) facies iron-formation, in contrast, formed locally below a silica cap rock by in situ low-temperature hydrothermal replacement of chert-tuff beds and even of the oxide formation. Thus the two types of iron-formation, although spatially and stratigraphically juxtaposed, are not the result of the same process, and only the carbonate facies appears to be directly related to a low temperature hydrothermal volcanic process. The contrasting genesis of oxide and carbonate iron-formations indicates that current simplistic models relating banded iron-formation to volcanic massive sulphide deposits need to be re-evaluated. The notion that oxide iron-formation represents a distal manifestation of volcano-generated hydrothermal activity may be the exception and not the rule, because, as indicated by this study, BIF occurs in the central part of a volcanic edifice, rather than on the flanks. © 2000 Elsevier Science B.V. All rights reserved.

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1. Introduction

The origins of iron-formations remain controversial (Kimberley, 1979, 1989; Dimroth, 1986). Banded iron-formations (BIF), prominent during the Archean-Paleoproterozoic, have been used to indicate reducing atmospheric conditions during Earth’s evolution (Holland, 1984). The most recent classifications of iron-formations, or chert-bearing ironstones, (Kimberley, 1979, 1989; Fyon et al., 1992) distinguish those of shallow- and deep-water settings (Simonson, 1985), as well as directly deposited and replacement types of iron-formations. In Archean volcanic arc settings Fyon et al. (1992) recognize a shallow-water type which may deposit on eroded volcanic edifices, and the
more common deep-water type with a proximal sulphide-graphitic shale facies and the more common distal oxide-turbidite facies. Lithologic associations may determine the depositional setting of an iron-formation, but the possible replacement origin may be difficult to determine in metamorphosed sequences.

James (1954) first documented facies associations (oxide, silicate and carbonate) in shallow-water iron-formation, attributing the changes to water depth at the time of deposition. However, subsequent work has indicated that the mineralogy may be early diagenetic (Dimroth and Chauvel, 1973; Kimberley, 1979). The source of the diagenetic fluid is more contentious but a primary or secondary volcanic source is generally indicated (Dimroth, 1986). Deep-water banded iron-formation with lutite interbeds are less well understood, but their formation, by analogy with that of the shallow-water type, is generally attributed to basinal exhalation and subsequent chemical sedimentation. The Fe- and Si-rich fluids are either of direct volcanic origin (Gross, 1980; Lydon, 1984) or deep circulating waters enriched by water-rock reactions (Kimberley, 1994). Facies variation appears to be related to source proximity (Simonson, 1985; Gross, 1980). Gross (1980) suggested BIF’s were products of volcanic exhalations, and that facies variations may point to the centre of metal-bearing volcanic-induced hydrothermal cells. This has become a widely used exploration criterion (Lydon, 1984; Fyon et al., 1992), although little has been documented in the literature. The revival of gold exploration in the last three decades added a possible sulphide facies iron-formation to the sequence. James (1954) original facies series included black shale with diagenetic pyrite as a possible sulphide facies; this is in startling contrast to the interbedded chert and sulphide minerals of the observed sulphidic iron-formation. This sulphidic facies was supposedly gold bearing initially (Fripp, 1976) and the gold was later remobilized during deformation into viable deposits. Work by various authors (Phillips et al., 1984) suggests that the sulphur and gold were introduced synchronous to deformation, and that cherty sulphidic iron-formation is replacement rather than a depositional facies.

While studying volcanism and hydrothermal activity in the Archean Hunter Mine group (HMG),

Fig. 1. General geology of the Abitibi (after Chown et al., 1992).
the authors encountered oxide and carbonate facies BIF at the same stratigraphic level. Volcanic activity was waning and there was clear evidence of penecontemporaneous low temperature hydrothermal activity causing subsurface replacement. By incorporating these observation with the general literature, the following article strives to demonstrate the sometimes equivocal relation between hydrothermal activity (exhalation) on the sea floor and the development of iron-formations.

2. Abitibi geology

The 2730–2726 Ma HMG volcanic rocks (Mortensen, 1987; unpublished data) constitute an integral part of the northern volcanic zone (NVZ) in the Archean Abitibi greenstone belt (Fig. 1), the largest coherent Archean supracrustal sequence in the world. The study area represents the southern segment of the NVZ, adjacent to the E-trending Destor–Porcupine–Manneville fault, which separates the NVZ from the southern volcanic zone (Fig. 1). The 2730–2705 Ma NVZ, composed of two distinct volcanic cycles, probably accumulated in a diffuse arc setting (Chown et al., 1992).

Volcanic cycles 1 and 2 represent incipient and mature stages of arc evolution, respectively. Volcanic cycle 1, 1–5 km thick, is an extensive, subaqueous basalt plain with local mafic-felsic or felsic edifices that are interstratified with or overlain by intra-arc, volcaniclastic turbidites (Mueller et al., 1996). The volcanic edifices are submarine composites or stratovolcanoes generally of bimodal composition. These volcanic centres, dispersed...
throughout the Abitibi greenstone belt, range in age from 2730 to 2720 Ma (Mortensen, 1993a,b). The 1–3 km-thick, volcanic cycle 2 (2720–2705 Ma; Mortensen, 1993a,b), is preserved in the northern and southern extremities of the NVZ (Fig. 1) and displays, in the Chibougamau area, the evolved stages of arc development, with the emergence and unroofing of the arc (Mueller et al., 1989). Coarse clastic deposits (Mueller and Donaldson, 1992a) and contemporaneous shoshonitic volcanism (Picard and Piboule, 1986) are indicative of arc maturity. The HMG, focus of this study and traceable along a strike of 50 km, is part of volcanic cycle 1 in the southern segment of the NVZ (Fig. 2).

3. Hunter Mine group

The inferred 4–5 km thick HMG (HMG; Fig. 3) is a complex submarine volcanic edifice overlain conformably by komatiites and komatiitic basalts of the Stoughton–Roquemaure group (SRG; Eakins, 1972; Mueller et al., 1997). The basal to medial parts of the HMG are composed of 5–50 m thick, massive to brecciated rhyolitic lava flows and lobes and an extensive, 1.5 km wide, felsic-dominated, dyke swarm that can be traced for 2.5 km up-section (Mueller and Donaldson, 1992b). Collectively, these units have a calc-alkaline arc signature (Dostal and Mueller,
Fig. 4. Outcrop map showing a portion of the upper part of the Hunter Mine group. A large enclave of oxide iron-formation occurs as a rip-up within a volcanic breccia. The enclave is folded, but remained a coherent block. Note the chert jasper magnetite veins along synvolcanic faults (after Mueller et al., 1997).
1996). Tholeiitic, mafic dykes of SRG-affinity are a minor component of the felsic-dominated dyke swarm (Dostal and Mueller, 1996). Abundant pyroclastic rocks either deposited from primary flows or remobilized and deposited downslope via mass flow processes characterize the volcanic complex (Mueller and Donaldson, 1992a,b). In addition to pyroclastic debris, magnetite iron-formation and jasper-magnetite iron-formation constitute a minor but significant part of the early constructive phase of the edifice.

The upper 500–1000 m thick transition zone of the HMG, directly above the dyke system (Fig. 2), is a highly variable association of pyroclastic, shale and volcaniclastic sedimentary rocks, associated with iron-formation, mafic to intermediate dykes and sills and mafic and felsic lava flows (Eakins, 1972; Mueller et al., 1997; N’dah, 1998). The mafic flow component increases towards the SRG, displaying on-lapping (Eakins, 1972) and interstratification relationship between groups which is common to volcanic arc terranes (Mueller et al., 1989). Combined, iron-formation, chert beds and tuffs up to 40 m thick can be traced 5 km along strike. The summits part of the HMG is characterized by intense silicification and carbonate alteration, especially evident in the volcaniclastic rocks and shale. The presence of synvolcanic and synsedimentary faults parallel to the N-trending dyke systems which served as conduits for hydrothermal fluids, attest to significant volcanic activity. Numerous faults are injected by dykes with dyke margins acting as vents for hydrothermal activity (Fig. 4). Evidently, hydrother-

Fig. 5. Photographs of typical BIF. (A) Jasplite iron-formation preserved in rip-up (Fig. 4) note white areas of recrystallization. Pencil 15 cm. (B) Typical oxide facies iron-formation (see Fig. 6) white bands recrystallized chert, black, oxide-rich layers and grey graded tuffaceous silt beds. Coin 2 cm. (C) Continuous cherty layers in tuff-turbidites with the onset of carbonate flooding (dark grey). Pencil is 15 cm. (D) Typical carbonate facies iron-formation, disconnected layers of white recrystallized chert (note lamination) in a seemingly homogeneous matrix of siderite. Pencil is 15 cm.
mal activity recurred throughout the history of the Hunter Mine volcanic complex.

The contact of the SRG, although locally characterized by a zone of high strain, is depositional, and interstratification between tholeiitic mafic and calc-alkaline felsic flows is observed at some localities (Fig. 4). The contact between the HMG and SRG is defined by the first appearance of komatiitic basalts. Dynamic contacts of this nature are common to volcanic sequences of which the
Hunter Mine and Stoughton–Roquemaure groups are no exception. The iron-formation, discussed in this study, is prominent in the summital part of the sequence and is associated with a massive sulfide deposit (N’Dah, 1998) of Mattabi-type alteration (Morton and Franklin, 1987), felsic lava flow breccias, and volcaniclastic sedimentary rocks.

The Hunter Mine and Stoughton–Roquemaure groups represent a south-facing homoclinal sequence (Fig. 2) characterized by steeply dipping, W to WNW-striking strata, locally overturned to the north. The rocks were affected by sub-green-schist facies metamorphism, characterized by a prehnite-chlorite-epidote (zoisite) assemblage. The prefix meta, has been omitted to simplify rock description.

4. Sedimentary lithofacies and iron-formation deposits

The volcaniclastic sedimentary lithofacies represent an integral part of the summital sequence and must be described in order to place the iron-formation deposits in a proper facies context. The tuffs vary in thickness from 0.03 to 5 m and are intercalated with massive and brecciated felsic flows, forming continuous or lenticular units. They may be massive, or bedded and laminated. Beds of tuff 3 cm thick generally alternate with chert beds of the same thickness. Although the tuff is very fine-grained and recrystallized, the presence of ghosts of devitrified glass shards, quartz crystals and composite quartz and matrix grains argues in favour of a pyroclastic origin. The bedding and fine lamination indicate deposi-

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Fig. 7. (A) Photomicrograph of chert layer showing fine laminations (horizontal). Random porphyroblasts and veins are composed of carbonate. (B) Photomicrograph of silicified black shale ‘cap rock’. (C) Photomicrograph of stylolitic chert being replaced by ankerite porphyroblasts. (D) Photomicrograph of volcanic fragment completely replaced by siderite.
Fig. 8. Outcrop map showing section through the transition zone of the Hunter Mine group. Silica cap rock overlies the carbonate alteration zone in which volcanic breccia as well as intercalated iron-formation and chert-bearing tuff are all heavily carbonatized. Scale bar 15m. (after N’Dah, 1998)

silica may have been focused fluids passing through the soft sediment to deposit principally at the sediment–water interface.

5. Oxide facies iron-formation

Oxide iron-formation is widespread in the uppermost section, or transition zone of the HMG and also occurs irregularly in sedimentary interlayers throughout the volcanic section. It is best preserved in a large, folded rip-up clast within a volcanic breccia (Fig. 4) where it appears as typical red banded jaspilite iron-formation. Elsewhere in the succession even the low grade metamorphism has resulted in recrystallization of the jasper.

The jaspilite iron-formation consists of delicately banded, brown to red, jasper layers up to 1 cm thick, alternating with thinner (0.5 cm) magnetite layers and 1 cm thick graded silt layers (Fig. 5A). Repetition of layers is not regular. Incipient recrystallization along the edges of the rip-up clast results in conversion of the jasper to white microcrystalline quartz, but the magnetite and silt layers show no visible change.

Although this large clast displays the BIF in its pristine state, it represents less than 1/2 m of section at best, and the slightly recrystallized iron-formation interbedded in the volcano-sedimentary strata gives a better idea of the overall depositional environment. The preserved section of iron-formation seldom exceeds 2 m in thickness (Fig. 6). The more or less alternating black, dark grey and white layers (Fig. 5B) are composed of magnetite, graded tuffaceous silt and chert. Although outwardly appearing as magnetite-bearing iron-formation, magnetite layers comprise only a minor portion. In the depicted section (Fig. 5B; Fig. 6) only the lower 10 cm actually contains oxide interlayers, and the remainder is composed of alternating tuffaceous silt and chert.

In thin section, the graded beds are seen to be composed almost exclusively of volcanic material, shards, lithic and mineral fragments, seemingly the very fine-grained equivalents of the graded units higher in the section. The chert layers are composed of microcrystalline quartz whose contacts with the silt are somewhat blurred and irreg-
ular as a result of recrystallization. Fine lamina-
tions shown by trains of minute opaque inclusions
are still faintly visible in the chert.

In summary, the oxide formation is a regular
part of the sedimentary units intercalated in the
HMG. It appears to represent normal pelagic
sediment deposited in periods of volcanic quies-
cence. Once deposited, the sediment is ripped up
and preserved within subsequent volcanic brec-
cias, much as was described for turbiditic se-
quences by Meyn and Palonen (1980). The fine
parallel layering in the chert suggest primary de-
position, or diagenesis at the sediment–water in-
terface synchronous with sedimentation, possibly
favouring the silica-rich (glassy) fine layers. The
magnetite layers may be the result of diagenetic
alteration of an original iron precipitate, although
a syn-sedimentary diagenesis is possible for these
layers, and might explain the irregular iron-rich
layers in the primarily chert sequence.

6. Carbonate facies iron-formation

Carbonate facies iron-formation occurs over a
1 km strike length of the HMG transition zone
east of the best-preserved oxide iron-formation
localities. It is spatially associated with evidence
of widespread hydrothermal alteration and minor
sulphide mineralization of the entire transition
zone (N’dah, 1998).

Carbonate facies iron-formation is a buff and
white coloured banded rock composed of distinc-
tive cm-thick white chert layers, initially continu-
ous (Fig. 5C), but most commonly slightly
discontinuous, interbedded with medium-grained
siderite layers up to 4 cm thick (Fig. 5D). In
extreme cases the iron formation is a chert breccia
in a carbonate matrix. Some corroborative detail
may be gained from a microscopic examination of
the iron-formation. Layers of minute opaque in-
cclusions within the chert (Fig. 7A), identical to

![Fig. 9. Compositional analyses of carbonates from the alteration zone in the Hunter Mine group (N’dah, 1998).](image-url)
those of the oxide iron-formation witness the fine laminations largely lost in recrystallization. Siderite everywhere replaces chert, giving the layers a somewhat ragged appearance and resulting in the discontinuous chert layering. Some visible mineral grains (chiefly quartz) and ghosts of shards visible in the carbonate indicate that much of the carbonate replaces original tuffaceous silt. A few bedding-parallel lines of oxide grains suggest that original oxide layers may also be replaced.

A typical section through the transition zone of the HMG (Fig. 8) shows the extent of the hydrothermal alteration throughout the section described by N’dah (1998). The intercalated shale, BIF and volcanic breccia are almost completely hydrothermally altered. The uppermost part of the section is silicified (Fig. 7B) and with associated Fe-chlorite represents the first phase of hydrothermal alteration (N’dah, 1998), which created an impervious silica cap to the sequence. The second stage of alteration completely converted volcanic breccia and iron-formation-banded chert and silt below this cap, to carbonate (Fig. 9). Alteration fronts preserved in the volcanic breccias show the first appearance of random porphyroblasts of ankerite (Fig. 7C) followed by complete replacement by siderite, commonly partly preserving the original texture of the rock (Fig. 7D). The tuffaceous silt interbeds with the chert in the oxide iron-formation were easily replaced, producing the carbonate facies iron-formation. It is impossible to say if all rocks experienced an initial conversion to ankerite, or if it was formed only at the end stages of carbonatization. Sulphide mineralization (chiefly marcasite and pyrite) occurs principally in the altered volcanic breccia and little occurs in the iron-formation. The presence of spherulitic marcasite suggests a low temperature (≥100°C) mineralizing fluid (Ames et al., 1993).

7. Summary

The oxide and carbonate facies iron-formation in the HMG are not the result of the same process, nor are they necessarily the direct result of volcanic processes. The oxide iron-formation is both laterally widespread and common at various stratigraphic levels in the HMG and is clearly a common pelagic sediment at times of volcanic quiescence. The most direct relation to volcanism is clearly the quiescence. Although some massive chert is present, the very regular nature of most of the chert interbeds suggests primary deposition or replacement at the sedimentary–water interface. The presence of quartz–jasper-bearing veins parallel to synvolcanic faults (Fig. 4) is a strong argument for a fluid source of local volcanic derivation, although their precise timing with respect to the BIF is not clear. The regular interaction between hot seawater and glass-rich, chemically-active volcanic breccias and sediments can hardly be ruled out as the ultimate source for the Si and Fe. It is a fact, however, as is shown dramatically in metallogenic maps of sedimentary rock belts northeast of the study area, that there is a strong correlation between thick sedimentary sequences and the occurrence of BIF in the Abitibi belt, and the oxide BIF may well be the end result of the action of cold seeps filtering up through the fertile volcanogenic sediments to the sea floor (cf. submarine fans, Fyon et al., 1992). They may also be the result of a hydrothermal system not related to sulphide deposition as described by Leistel et al. (1998).

The carbonate facies iron-formation in the HMG is clearly the result of in situ low temperature hydrothermal replacement which formed a silica cap and then replaced the volcaniclastic rocks and the chert tuff formation below the cap with siderite as the hydrothermal system evolved. The hydrothermal activity is probably the result of late-stage volcanic activity and seawater circulation, but the BIF itself is a replacement deposit and not the result of direct seafloor precipitation induced by volcanism. A similar system is documented in the Iberian Sulphide belt (Leistel et al., 1998) where Fe–Mn carbonates replace chert interbeds, and the Helen Formation, long considered the type volcanic exhalative iron-formation formed in a similar manner (Morton and Nebel, 1984).
8. Conclusion

The two iron-formations in the HMG are spatially, but not temporally related. Oxide facies iron-formation formed first at the seawater interface, partly replacing the finer-grained portions of the turbiditic tuffs. Subsequently, a low-temperature hydrothermal system developed, which first formed a silica cap rock, and, as the hydrothermal solution evolved, replaced the underlying volcanioclastic sequence with siderite. The interbedded chert-tuff sequence became interbedded chert and siderite, an iron-formation. This example of oxide and carbonate facies iron-formation demonstrates the difficulty of inferring a simplistic one-to-one correlation between iron-formations and volcanic activity. The oxide iron-formation appears to be a normal pelagic deposit in a volcano-sedimentary environment. It may be the result of direct volcanic exhalation, but is more likely produced from exhalative fluids derived from rock-water interaction. As demonstrated in this study, it may occur within the central volcanic complex, rather than being a distal manifestation. The carbonate iron-formation, on the other hand, is of replacement origin and cannot be used as an indicator of sedimentary facies, although its obvious relation to a hydrothermal cell makes it alone a promising prospecting tool. These results show that the broad interpretation of banded iron-formation as a distal equivalent of volcanogenic massive sulphide deposits is unwarranted.

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