Geology and alteration of the Merrie greenstone belt

by T. R. Farrell[1] and N. G. Adamides

Abstract

The Archaean Merrie greenstone belt is located at the northern end of the Eastern Goldfields Province in the Yilgarn Craton. The belt contains a large syncline with a lower sequence of mafic and ultramafic rocks, and an upper sequence of felsic and sedimentary rocks. The belt has been deformed and metamorphosed to lower amphibolite facies adjacent to the greenstone margin, and greenschist facies in the centre. Alteration of the greenstones post-dates regional metamorphism and deformation. Mafic rocks are typically overprinted by patchy epidote alteration and quartz–epidote veins, or by diffuse biotite alteration adjacent to networks of pyrite-filled fractures (or both). Alteration took place at 410–430°C, under low-fluid CO_2 concentrations (X_{CO_2} <0.18), and resulted in the net addition of potassium. Alteration was either not intense enough, or occurred at fluid CO_2 levels that were too low, for the development of carbonate-bearing alteration.

KEYWORDS: Archaean, greenstone, metamorphism, hydrothermal alteration

The Merrie (or Cunyu) greenstone belt lies about 90 km north of Wiluna, on the northern edge of the Archaean Yilgarn Craton. The belt is poorly exposed and, in the past, the geology and structure were necessarily inferred from limited outcrop and low-resolution aeromagnetic data. In this paper, we present a revised interpretation of the structure, metamorphism, and alteration of the Merrie belt, with emphasis on the mineralogy and $T–X_{CO_2}$ conditions of alteration. This interpretation is based on recent 1:100 000-scale geological mapping, new aeromagnetic data, and petrographic work on drillhole samples.

Geological setting

The Eastern Goldfields Province (Griffin, 1990) is characterized by elongate, north to northwesterly trending greenstone belts and extensive areas of monzogranite and gneiss. The Merrie greenstone belt (Fig. 1) contains a sequence of deformed supracrustal rocks dominated by basalt (with thin units of ultramafic rock and minor gabbro), with less abundant overlying felsic and sedimentary rocks. The belt is flanked by poorly exposed granitoid rocks (mainly monzogranite), and unconformably overlain in the north by sedimentary rocks of the Proterozoic Yerrida and Earaheedy Basins. A major tectonic lineament, the Merrie Range Fault (Fig. 1; Myers and Hocking, 1998), lies to the east, and a possible splay off this fault runs along the eastern margin of the belt.

Stratigraphy and structure

The Merrie belt is poorly exposed and the sparse surface outcrops are deeply weathered. For this reason, most of the information on the geology of the belt has been derived from sampling of exploration drillholes. An aeromagnetic image of the area (Fig. 2) shows that the belt contains a large fold, outlined by thin units of high magnetic intensity that are interpreted to be ultramafic bodies (no banded iron-formation was identified in surface outcrops). Liu (1997), and Myers and Hocking (1998) interpreted this fold as an anticline. However, on the basis of way-up directions from poorly preserved pillow structures in metabasalt, it is interpreted here to be a syncline. This is consistent with the absence of granite in the core of the structure (common in antiformal greenstone belts, Myers, J. S., 1999, pers. comm.), and is further supported by geophysical modelling (Adamides, in prep.). Mafic rocks (dominantly metabasalt with subordinate, thin ultramafic units, and minor metagabbro, amphibolite and chlorite schist) dominate the outer, stratigraphically lower, parts of the syncline. Metamorphosed felsic and sedimentary rocks are interpreted from surface geology and drillhole information to occupy the upper part of the sequence in the centre of the syncline (Fig. 1). Weakly deformed, metamorphosed hornblende diorite was identified in one drillhole in the core of the syncline.

The Merrie belt is heterogeneously deformed, and zones of strongly deformed rock typically alternate with zones of little or no deformation. A zone of intense deformation is present along the faulted eastern margin of the belt. Most rocks in the belt have a weak to moderate, subvertical, north-northwesterly

[1] t.farrell@dme.wa.gov.au

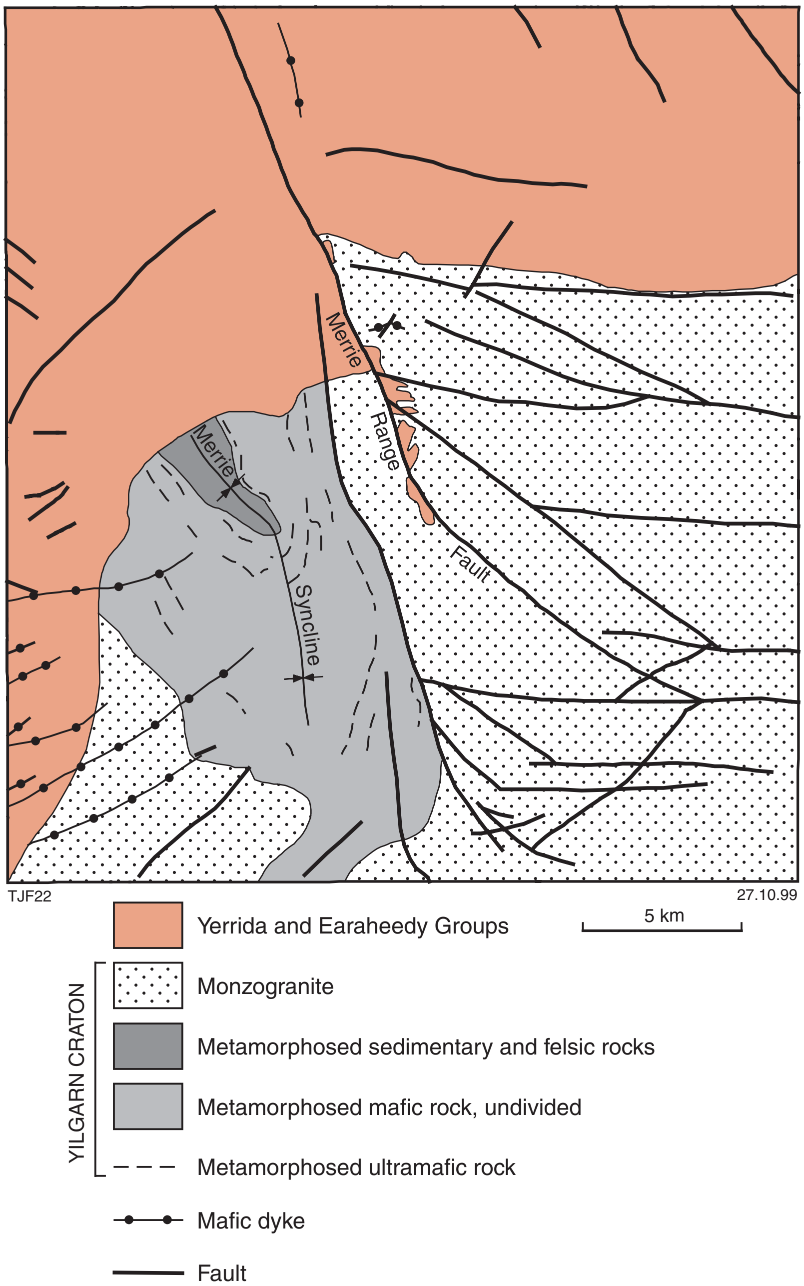

Figure 1. Interpreted geology of M*ERRIE* *showing the location of the Merrie greenstone belt*

trending foliation, defined mainly in mafic rocks by the preferred orientation of chlorite. A sub-horizontal mineral lineation is present in more highly strained zones, and a fine crenulation is also present in chlorite-rich mafic rocks.

Metamorphism

In common with other greenstone belts in the Eastern Goldfields Province (Myers and Swager, 1997), the Merrie belt has a narrow higher grade zone (amphibolite facies) along the margins, and a low-grade zone (greenschist facies) covering most of the central part of the belt.

In the amphibolite-facies zone, mafic rocks contain mineral assemblages that include hornblende and Ca-plagioclase. In contrast, in the low-grade zone, medium- to coarse-grained mafic rocks (metagabbro and metadolerite) contain lower greenschist-facies assemblages of actinolite–albite(–epidote). Metabasalt commonly contains vesicles filled by quartz(–epidote) and chlorite (Fig. 3a). The ground-mass in these rocks is typically composed of a fine-grained mixture of quartz, actinolite, albite, and chlorite. Intermediate rock types (?meta-andesite) were also identified, in some cases with relict plagioclase phenocrysts (Fig. 3b). Metamorphosed ultramafic rocks consist of various proportions of intimately intergrown chlorite, tremolite, anthophyllite, and talc.

Metamorphosed felsic porphyritic or volcaniclastic rocks (Fig. 3c), encountered in drillholes in the centre of the belt, typically have a recrystallized matrix of fine-grained, interlocking quartz and albite enclosing fragments of subhedral or embayed quartz, sericitized plagioclase, and, in some cases, felsic lithic fragments and relict white mica.

Metasedimentary rocks contain the assemblage quartz–actinolite–epidote–leucoxene (probably from the breakdown of detrital iron–titanium minerals). They are interpreted to be sedimentary in origin on the basis of a layering defined by the alternation of quartz- and epidote-rich domains, and the absence of any relict igneous textures.

The stability of Ca-plagioclase in mafic rocks close to the margin of the belt indicates a temperature greater than about 450–500°C in the amphibolite facies zone. In low-grade mafic rocks in the centre of the belt, the presence of chlorite–albite–actinolite(–epidote) indicates lower greenschist facies — probably at a temperature of 350–400°C (cf. Liou et al., 1985).

Hydrothermal alteration

The metamorphic mineral assemblages are overprinted by later veins and zones of hydrothermal alteration, particularly adjacent to crosscutting fractures. The most abundant alteration minerals are biotite, epidote, and quartz. Hydrous Fe-oxides (after ?pyrite) are present in some veins, and carbonates are usually scarce. Strongly epidotized rocks typically contain abundant

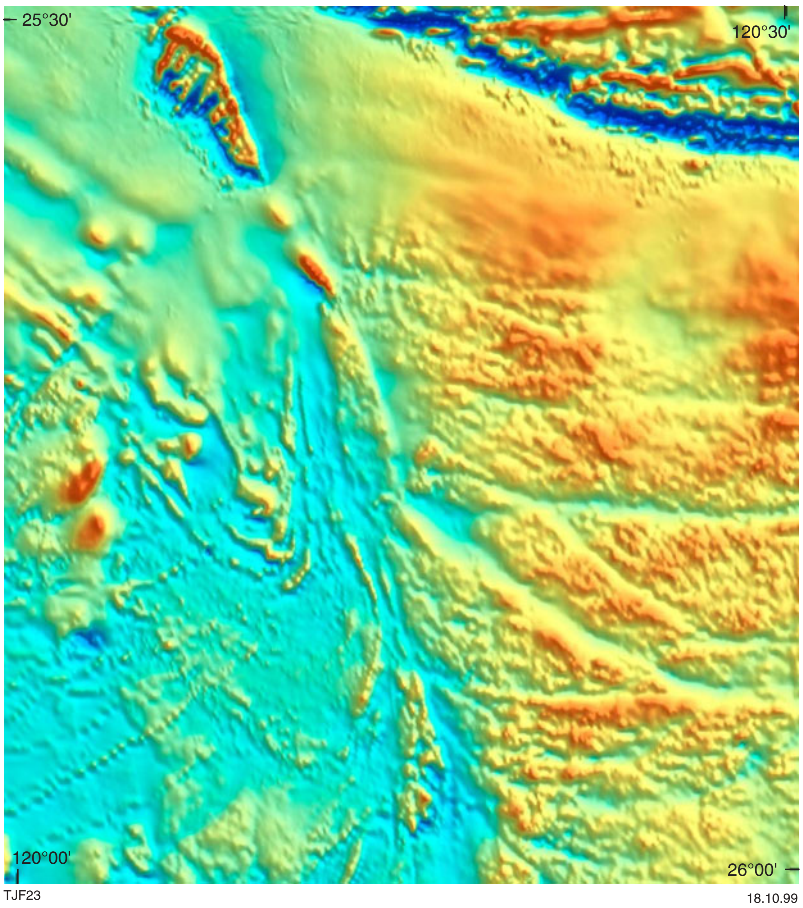

Figure 2. Total magnetic intensity aeromagnetic image of MERRIE*. Note the fold structure in the Merrie greenstone belt defined by thin highly magnetic units*

fine-grained epidote, and are cut by veins containing quartz and epidote (Fig. 3d).

Biotite is a conspicuous alteration mineral in many of the mafic rocks, and its presence is interpreted to be evidence of K-metasomatism. Biotite typically occurs in broad alteration haloes surrounding veins and fractures, commonly in association with fine-grained epidote. In strongly altered rocks, the biotite alteration is pervasive. Biotite is also present as an alteration product around the margins of actinolite crystals, as well as being retrogressively altered to chlorite.

The altered mafic rocks are commonly cut by an irregular network of fractures and thin veins that contain hydrous Fe-oxides (?limonite–goethite). The Fe-oxides in these veins are probably the product of weathering and oxidation of former pyrite. A biotite alteration halo typically envelopes the veins. In some samples, these (former) pyrite veins cut across quartz–epidote veins, and are therefore interpreted to be late in the paragenetic sequence. In all cases, the pyrite veins cut across the foliation and the vein minerals are undeformed.

The opaque minerals are usually broken down to leucoxene and, with progressive alteration, a transition is observed through to microcrystalline titanite, and rarely, to well-crystallized rutile. The presence of finely dispersed microcrystalline titanite throughout many altered rocks suggests that there was localized mobility of titanium during alteration.

Discussion

Alteration in the Merrie belt is interpreted to have taken place after the main regional deformation events (D_1–D_3, Farrell, 1997). This is based on the observation that the pyrite veins with associated biotite haloes are undeformed and cut across the main foliation, coupled with the random orientation of individual biotite crystals. Biotite is not a typical component of unaltered metamorphosed mafic rocks in the Eastern Goldfields Province, and its abundance in the Merrie belt is probably due to K-metasomatism, possibly related to fluids released during the crystallization of late granitoid intrusions (cf. McNaughton et al., 1993). In addition, biotite has not been reported from alteration zones at Jundee–Nimary (Phillips et al., 1998a, 1998b; Byass and Maclean, 1998) and Wiluna (Chanter et al., 1998) to the south. The presence of biotite in the Merrie belt suggests that alteration occurred at a higher temperature than in either of these two deposits.

Figure 4 shows some of the relevant mineral equilibria and the interpreted conditions of alteration in the Merrie belt on a T–X_{CO_2} diagram. The coexistence of epidote–chlorite–biotite–albite–quartz, along with the absence of Ca-plagioclase, limits the temperature to approximately 410–430°C for greenschist facies rocks (neglecting the effect of Fe^{2+} substitution for Mg) — slightly higher than indicated for regional metamorphism. The stable coexistence of epidote–biotite also indicates a low X_{CO_2} (0 – 0.15), which is consistent with the low carbonate content of the altered mafic rocks. This is also confirmed by the scarcity of rutile, which indicates that the CO_2 content of the fluid was only locally high enough (X_{CO_2} 0.15 – 0.18) to cause the breakdown of titanite.

These values of X_{CO_2} are broadly similar to typical values for lode-gold deposits in the Yilgarn Craton (0.05 – 0.25; Witt, 1991; Mikucki and Ridley, 1993). However, alteration in the Merrie belt is distinctly different in style to the alteration reported from gold deposits in the northern part of the Yandal and Agnew–Wiluna belts. The sequence at Jundee–Nimary (Byass and Maclean, 1998; Phillips et al., 1998a,b) is broadly similar to that in the Merrie belt (dominated by mafic rocks with thin units of ultramafic rock and subordinate

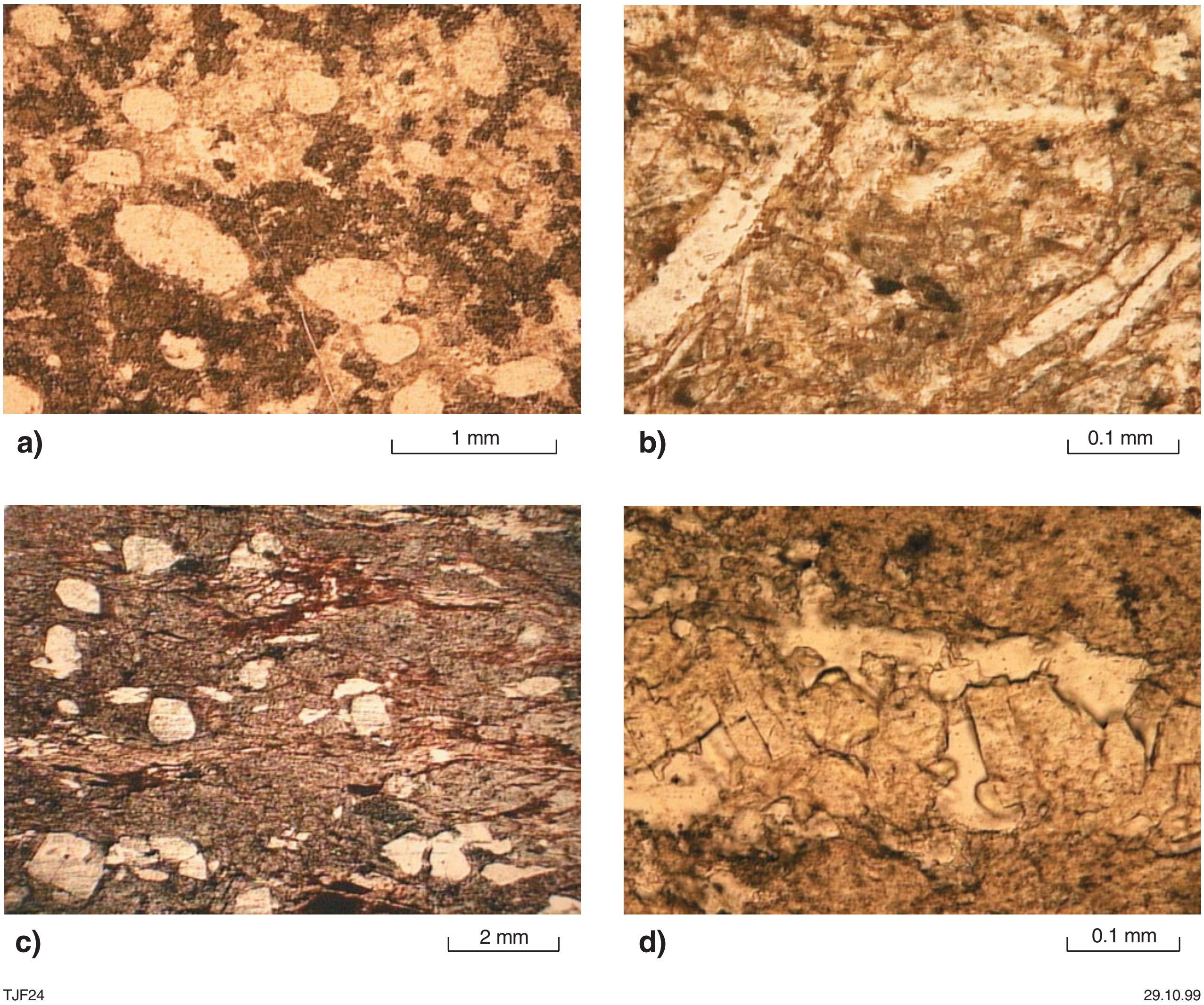

Figure 3. Textures of rock types of the Merrie greenstone belt
a) Relict vesicular texture in metamorphosed mafic lava. Vesicles are filled by quartz. The matrix of the rock is predominantly composed of actinolite, with patchy epidote alteration (dark areas). Plane polarized light. GSWA 152869
b) Plagioclase-phyric, epidotized intermediate rock, probably meta-andesite. The matrix is composed of epidote, biotite, and albite. Plane polarized light. GSWA 152875a
c) Typical felsic volcaniclastic rock, with relict subhedral and embayed quartz grains in a foliated fine-grained matrix. Plane polarized light. GSWA 152894
d) Epidote–quartz vein in fine-grained mafic rock. Plane polarized light. GSWA 152877

amounts of felsic porphyry), but the mineralization is associated with quartz–carbonate–sulfides(–chlorite –muscovite) alteration. Mineralization in the Bulletin deposit at Wiluna (Chanter et al., 1998) is associated with sericite–carbonate (–chlorite) alteration. The important difference to the Merrie belt is that carbonates are abundant in proximal alteration zones in both deposits, and in many other low-T (<450°C) lode-gold deposits in the Yilgarn Craton (Witt, 1991; Mueller and Groves, 1991; Mikucki and Ridley, 1993). The general absence of carbonates in the Merrie belt suggests that, either the sampled rocks are from areas of distal (weak) alteration at reduced X_{CO_2} (cf. Witt, 1991), or that the fluid CO_2 content was too low to stabilize carbonates.

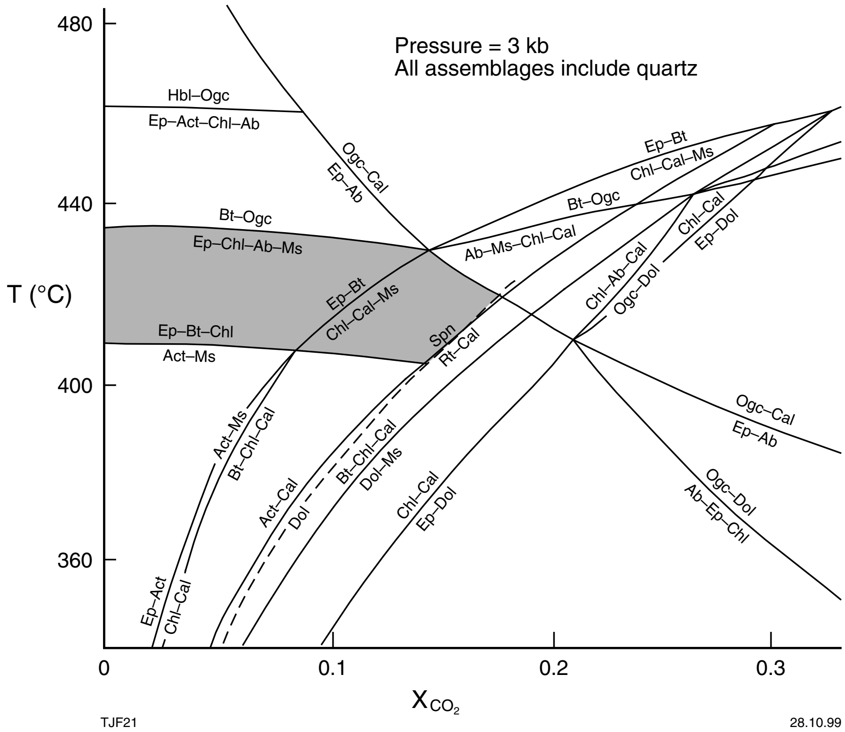

Figure 4. *Isobaric T–X_{CO_2} diagram showing relevant mineral equilibria in the system K_2O–Na_2O–CaO–MgO–Al_2O_3–SiO_2–H_2O–CO_2 (modified from Clark et al., 1986). Temperature–X_{CO_2} conditions for alteration in the Merrie belt are indicated by the shaded area. Minerals: Ab = albite, Act = actinolite, Bt = biotite, Chl = chlorite, Cal = calcite, Dol = dolomite, Ep = epidote, Hbl = hornblende, Ms = muscovite, Ogc = oligoclase, Rt = rutile, and Spn = sphene*

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