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***Cover:** Banded iron formation comprising 'tigereye' (fibrous yellowish brown: silicified crocidolite or blue asbestos), haematite–magnetite (grey), ferruginous chert and jasper (yellow and red). Sample from Western Australia; supplied by Dean Hoatson.*

The Western Tharsis deposit

A 'high sulphidation' Cu–Au deposit in the Mt Lyell field, of possible Ordovician age

DL Huston & J Kamprad

The Western Tharsis deposit in the Mt Lyell Cu–Au district of western Tasmania has been reinterpreted as an Ordovician 'high sulphidation' Cu–Au deposit. Mapping of alteration assemblages associated with chalcopyrite-rich and bornite-rich ore types suggests that these deposits formed in a single mineralising event, not in two disparate events as suggested previously. The presence of pyrophyllite, topaz, zunyite and woodhouseite within alteration zones associated with the deposit is diagnostic of 'high sulphidation' Cu–Au deposits. The use of PIMA was essential in mapping the alteration zonation as it identifies minerals such as pyrophyllite, topaz and zunyite effectively. High field strength and rare earth elements—generally considered immobile during alteration—were highly mobile during mineralisation at Western Tharsis and may have direct application as lithogeochemical indicator elements in 'high sulphidation' Cu–Au deposits.

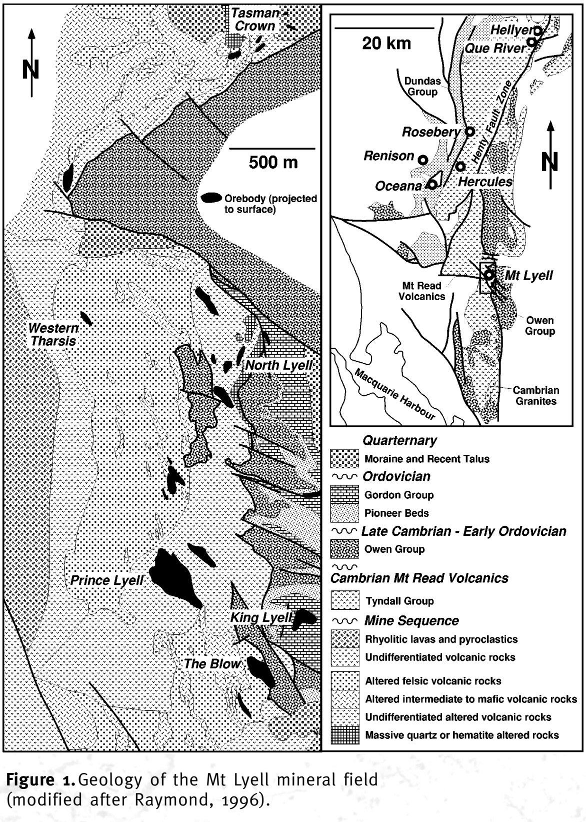

Figure 1. Geology of the Mt Lyell mineral field (modified after Raymond, 1996).

The Cambrian Mount Read Volcanic Belt of western Tasmania is divided into two distinct metallogenic provinces by the north-north-east trending Henty Fault Zone. One domain is dominated by Zn–Pb-rich volcanic-hosted massive sulphide (VHMS) deposits to the north-west, and the other is dominated by Cu–Au volcanic-hosted disseminated sulphide deposits to the south-east (figure 1). The Mount Lyell field, just to the east of Queenstown, contains 22 dominantly Cu–Au deposits and constitutes the major district of the south-eastern province.

Ore deposits in the Mount Lyell field can be subdivided into five groups:

1. disseminated pyrite–chalcopyrite orebodies such as Prince Lyell;
2. bornite–chalcopyrite orebodies such as North Lyell;
3. massive pyrite–chalcopyrite orebodies such as The Blow;
4. stratiform, massive pyrite–galena–sphalerite–galena–pyrite orebodies such as Tasman Crown; and
5. copper–clay orebodies such as King Lyell.[1]

The first two types are by far the most significant. Previously they were inferred to have formed by a Cambrian syn-volcanic mineralising event (disseminated pyrite–chalcopyrite ores) and by an overprinting event related to Devonian granitoid intrusion (bornite–chalcopyrite ores).[2]

As part of AMIRA project P439 (Studies of VHMS-related alteration: Geochemical and mineralogical vectors to mineralisation), geochemical patterns and the zonation of alteration assemblages were investigated around the Western Tharsis deposit. This deposit was chosen because it is the last undeveloped orebody in the district, and it is one of the few orebodies in which pyrite–chalcopyrite and bornite–chalcopyrite ores are juxtaposed.

Geological setting

The Mount Lyell mineral field (figure 1) occurs largely within the Central Volcanic Complex, the volcanic unit that forms the basal member of the Middle to Late Cambrian Mount Read Volcanic Belt. In the Queenstown area, the Central Volcanic Complex (~503 million years[3]) consists of felsic volcaniclastic rocks and lavas intercalated with intermediate volcaniclastic rocks, lavas and sills, together with minor siltstone and epiclastic sandstones. The Central Volcanic Complex is overlain unconformably by the volcaniclastic rocks of the Tyndall Group which, in turn, is overlain by siliciclastic conglomerate and sandstone of the Late Cambrian to Early Ordovician Owen Group. An internal unconformity (Haulage Unconformity) within the Owen Group separates older strata (~493 million years[4]) from a younger unit (~462 million years[5]) termed the Pioneer Beds. Carbonate rocks of the Ordovician Gordon Group overlie the Pioneer Beds (figure 1).

Three deformation events are recognised in the Mount Lyell area. The earliest, interpreted as an Ordovician event, is characterised by the development of inclined, asymmetric folds in the Owen Group and by erosion to form the Haulage unconformity that separates the Pioneer Beds from the rest of the Owen Group. Movement along the Great Lyell Fault, that forms the contact between the volcanic rocks and the Owen Group through much of the Mount Lyell field, began at this time.[2] Another two deformation events occurred during the Devonian Tabberabberran Orogeny.[6]

Most Mount Lyell ore deposits, including Western Tharsis, occur within the Mine Sequence. This unit, comprising mainly felsic volcaniclastic rocks and lavas with lesser intermediate volcanic rocks, has a maximum thickness of 800 metres. Much of the Mine Sequence is overprinted by a pyrite alteration zone that extends along strike for eight kilometres.[6] Nineteen of the 22 known deposits occur within this altered zone.

Figure 2A illustrates the lithological variation of the ore-bearing sequence at Western Tharsis. The host volcanic sequence dips steeply to the west and is overturned. The deposit, which has an indicated and inferred resource totaling

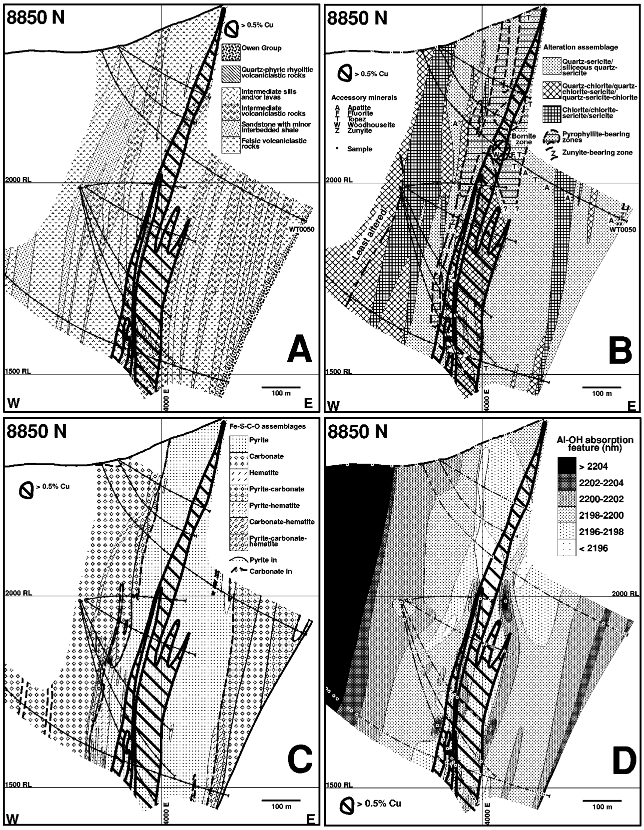

Figure 2. Cross section 8850N showing **A.** the geology of the Western Tharsis deposit **B.** the distribution of silicate alteration assemblages **C.** the distribution of Fe–S–C–O minerals and **D.** variations in the wavelength of the Al–OH absorption feature from PIMA analysis.

12.4 million tonnes at 1.3 per cent Cu, is a stratabound, mainly disseminated, pyrite-chalcopyrite body. The bornite-bearing assemblage that replaces the pyrite–chalcopyrite assemblage also contains minor to trace quantities of chalcocite, mawsonite, chalcopyrite, molybdenite, hematite, enargite, barite and woodhouseite.[7] The pyrite–chalcopyrite assemblage has a simpler mineralogy, with minor to trace barite, apatite and molybdenite, and with local zones of massive magnetite.

Zonation and mineral assemblages

Alteration at Western Tharsis was characterised by the development of zoned silicate (figure 2B) and Fe–S–O–C (figure 2C) mineral assemblages, each having significantly different zonation patterns. Three major silicate alteration assemblages are present at Western Tharsis. The lower part of the ore zone is centered on a pyritic quartz–chlorite±sericite zone that grades outward into a pyritic quartz–sericite±pyrophyllite zone. Within this zone, pyrophyllite, which was identified using PIMA (portable infra-red mineral analyser) analysis, forms a shell that wraps around the upper part of the ore-related quartz–chlorite± sericite zone. The pyrophyllite-bearing zone also contains topaz and zunyite, identified using PIMA. The quartz–sericite±pyrophyllite zone passes outward to a quartz–chlorite–sericite–carbonate zone that lacks pyrite (figures 2B & 2C). Hematite is commonly present along the contact between pyrite-bearing and carbonate-bearing zones. PIMA analysis also indicates that the composition of sericite (as measured by the wavelength of the Al–OH absorption feature) becomes more phengitic marginal to the orebody (figure 2D).

Geochemical dispersion

In concert with core logging, mineralogical and PIMA studies, a suite of more than 200 fresh drill core samples were analysed for major elements and trace elements. Based on contoured sections and inter-element correlations,[#] the elements that show systematic patterns with respect to the orebody can be subdivided into three major groups:

1. elements (As, Bi, Cu, Mo, Ni, S and Se) that are strongly enriched in the ore zone, and in the pyrite-bearing alteration zones, but depleted in the marginal carbonate-bearing alteration zone;
2. elements (K and Cs) that are characterised by uniform values except for extreme depletion within the pyrophyllite-bearing zone that forms a shell around the orebody; and
3. elements (C, Mn, Ca, Zn and Tl) that are enriched in the carbonate halo but depleted in the pyrite halo.

Figure 3A illustrates the distribution of Cu as established using routine company analytical data and data acquired during this study. Molybdenum, Ni and S correlate with Cu and form part of the ore metal assemblage. Although the correlation between Bi and Cu is not great (r = 0.107), Bi is enriched in the ore zone and, to a lesser extent, in the pyrite halo. Although Zn is locally enriched within the ore zone (figure 3B), it is mainly enriched in the carbonate halo (see below). This local Zn enrichment does not appear to be related to late flat veins as in the Prince Lyell orebody.[8]

The distribution of K is characterised by extreme depletion (0.06–0.9% K_2O; figure 3C) in narrow (20–30 m) zones that flank the orebodies both in the footwall and the hanging wall and envelop the orebody at surface. Elsewhere the abundance of K is stable and much higher (K_2O = 2–4%). Cesium correlates with K.

The K and Cs depletion anomalies correlate spatially with the pyrophyllite-bearing alteration zone that wraps around the orebody (figure 2B). The aluminous minerals in this zone developed at the expense of sericite, which accounts for K depletion.

The marginal carbonate-bearing alteration zone is characterised by low order Zn (100–1000 ppm; figure 3B) and Tl (0.5–1.2 ppm) enrichment. This contrasts with the inner pyritic zone, which generally contains less than 50 ppm Zn. In addition to Zn and Tl, the marginal zone is also enriched in C, Mn and Ca.

Mobility of 'immobile' elements

This study shows elements that are generally considered immobile, such as high field strength (HFSE) and rare earth elements (REE), were mobile in the Western Tharsis mineralising environment. Samples from drill hole WT0050, that passed through the bornite zone, were analysed for a suite of elements including HFSE and REE. The HFSE Ga and Y (and Be) were found to be depleted in the pyrophyllite-bearing zone (F and Sr were enriched; figure 4).

Figure 4 illustrates the spatial variation in chondrite normalised REE concentrations of rhyolitic rocks (identified using TiO_2/Zr) in WT0050. More mafic rocks were excluded as they have significantly different initial REE contents to rhyolitic rocks (cf. Crawford et al.[9]). Medium rare earth elements (MREE), represented by Ho and, to a lesser extent, heavy rare earth elements (HREE), represented by Yb, are strongly depleted in the pyrophyllite-bearing alteration zone surrounding the bornite-bearing ore zone. Marginal to this zone, MREE and HREE values rapidly approach those typical of Central Volcanic Complex rhyolites (cf. Crawford et al.[9]).

The distribution of light rare earth elements (LREE), as represented by Ce, is more complicated. From the stratigraphic footwall, LREE concentrations initially increase up to, and including, the ore zone, possibly due to the presence of LREE-bearing woodhouseite. In the hanging wall, LREE are moderately depleted, with zones of more extreme depletion corresponding to extreme F enrichment.

To more critically assess the dispersion of Ce and remove the effects of lithological variations, figure 3D was constructed to display variations in normalised Ce enrichment. Cerium concentrations were normalised to primary Ce concentrations (Ce/Ce_{PR}) of unaltered precursors from the Mt Read Volcanics (precursor data from Crawford et al.). Use of this technique indicates that positive Ce anomalies are developed within and adjacent to the orebody.

The slope of REE patterns, as measured by $[La/Yb]_{CN}$ (figure 4), largely follow the pattern observed for LREE, increasing towards ore except in F-rich samples. Figure 4 also shows variations in the Eu anomaly (Eu/Eu*). A narrow baseline with values of 0.50–0.69 is defined mainly by carbonate-bearing and some quartz–sericite–pyrite altered rocks. Values above 0.70, rising to 2.36 adjacent to the ore zone, define a major symmetrical anomaly about the ore position and a second zone about a smaller pyrophyllite- and zunyite-bearing zone near the base of the drill hole. The anomalously high Eu/Eu* values extend 100–150 metres

Raw analyses were contoured in preference to ratios or mass transfer as they require no assumptions about protoliths and as 'immobile' elements used to calculate mass transfer appear to be mobile at Western Tharsis.

from the ore zone into both the footwall and the hanging wall. This anomalous footprint is the widest of any REE-based anomalies, and is similar in extent to the anomalies associated with most other group one and two elements above.

Genetic implications

Paragenetic and spatial relationships indicate that the disseminated pyrite–chalcopyrite assemblage is associated mainly with a quartz–chlorite±sericite silicate alteration assemblage with accessory hydrothermal apatite. In contrast, the bornite-dominant ore assemblage is associated with a quartz–pyrophyllite±sericite alteration zone that wraps around the core quartz–chlorite±sericite zone. This spatial relationship suggests that the pyrophyllite- and chlorite-bearing assemblages, and the associated pyrite–chalcopyrite and bornite-dominant sulphide assemblages, formed during a single hydrothermal event. This is in contrast to the generally accepted two-event models such as that of Arnold and Carswell.[2] The association of woodhouseite, zunyite, topaz and fluorite with the pyrophyllite-bearing assemblage is diagnostic of hypogene advanced argillic alteration assemblages that occur in deep-level, higher temperature, 'high sulphidation' Cu–Au deposits.[10, 11]

Hart demonstrated that pyrophyllite-bearing alteration assemblages extend through the Owen Group, including the Pioneer beds and, possibly, the base of the Gordon Group.[12] He also reported the presence of bornite within the Owen Group. This suggests that the mineralising event in the Mt Lyell area may have post-dated deposition of the majority of the Owen Group, as suggested by Raymond.[13] As the alteration assemblages are all affected by Tabberabberan cleavages, this mineralising event must have predated this Devonian deformation event. Although an early Tabberabberan timing is possible, all known Tabberabberan granitoids in Tasmania are post-tectonic. This is not consistent with the close relationship with granites required for 'high sulphidation' deposits. These data suggest that much of the mineralisation in the Mt Lyell district may not be Cambrian or Devonian in age, but rather Ordovician in age. This hypothesis is currently being tested using Pb isotopes in apatite.

The geochemical dispersion around the deposits is also consistent with a 'high sulphidation' origin. For instance, White and Hedenquist indicate that K and Zn are depleted, and As, Cu, Bi and Mo are enriched in 'high sulphidation' deposits[10]—as at Western Tharsis. Arribas et al. documented extreme REE mobility and found that advanced argillic and residual silica altered samples from the Rodaqualar 'high sulphidation' deposit in Spain had $[La/Yb]_{CN}$ significantly higher than the ratios in less intensely altered samples.[14]

Conclusions

Zoned alteration assemblages in the Western Tharsis deposit indicate that chalcopyrite-rich and bornite-rich ores formed at essentially the same time, not at two separate times as previously suggested.

The overprinting of ore-related alteration assemblages on Ordovician rocks and the presence of Tabberabberan foliations in the alteration assemblages strongly suggest that Western Tharsis and, by inference, most of the Mt Lyell field formed during the Ordovician.

The presence of pyrophyllite, topaz, zunyite and woodhousite in the alteration assemblages suggests that the Western Tharsis is a deep-level 'high sulphidation' Cu–Au deposit.

Lithogeochemical studies of the deposit indicate proximal enrichment of As, Bi, Cu, Mo, Ni, S and Se, and distal enrichment of Zn, Tl, C, Mn and Ca. Significant K and Cs depletion is associated with pyrophyllite-bearing alteration assemblages.

Extreme alteration in the Western Tharsis deposit has resulted in the mobility of REE and some HFSE. Rare earth element mobility, in particular Ce and Eu may be useful as pathfinders in this and other 'high sulphidation' Cu–Au deposits.

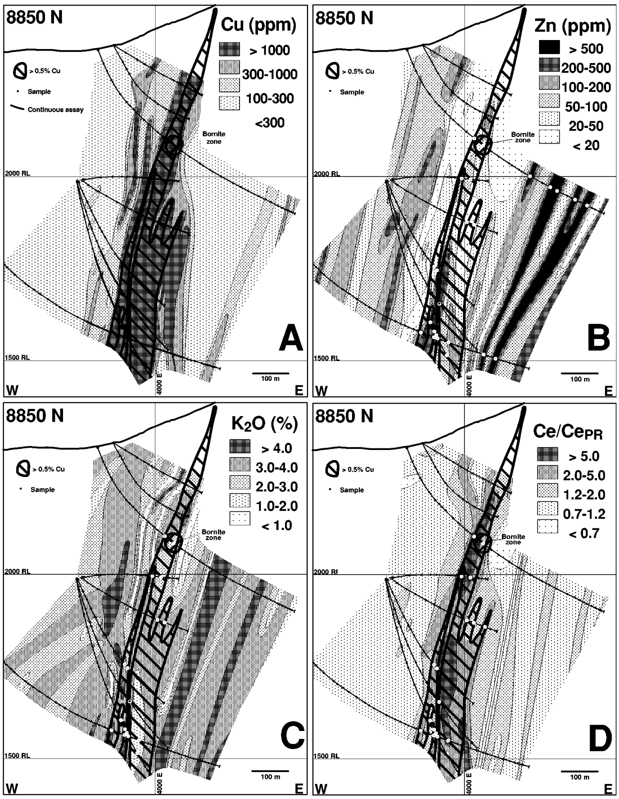

Figure 3. Cross section 8850N showing lithogeochemical dispersion of **A.** Cu **B.** Zn **C.** K and **D.** normalised Ce.

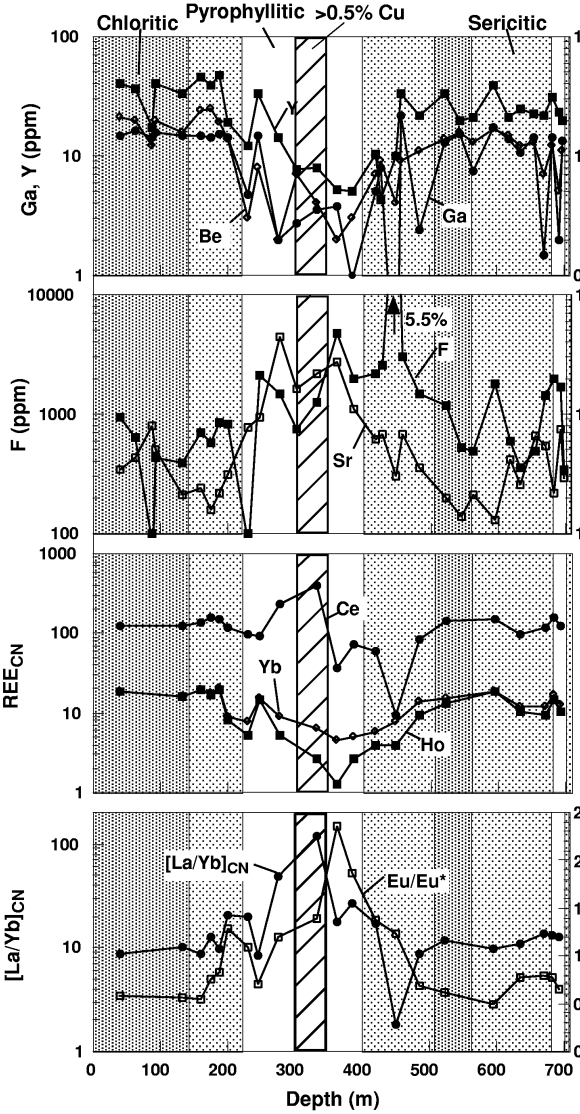

Figure 4. Variations in the concentrations of **A.** Be, Ga and Y **B.** Sr and F **C.** Ce, Ho and Yb and **D.** La/Yb and Eu/Eu* as a function of depth in WT0050. REE analyses are chondrite normalised and restricted to rhyolitic rocks.

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Exploration strategies for Precambrian layered mafic–ultramafic intrusions in the East Kimberley

DM Hoatson

The Palaeoproterozoic layered mafic–ultramafic intrusions in the Halls Creek Orogen (HCO) of the East Kimberley, Western Australia, represent one of the most extensively mineralised igneous associations of their type in Australia. The intrusions contain a range of magmatic and hydrothermal deposits of platinum-group elements (PGEs), chromium, nickel, copper, cobalt, titanium, vanadium, iron and gold. Despite intensive exploration during the past four decades and the discovery of many prospects, none of the deposits associated with the layered intrusions has proved to be economic.

AGSO and the Geological Survey of Western Australia, as part of the National Geoscience Mapping Accord Kimberley–Arunta project, have extended the distribution of prospective layered intrusions and highlighted the potential for different styles of mineralisation with prominent surface expressions and remote sensing–geophysical signatures.

Major findings relating to the intrusions are summarised in the following *AGSO Research Newsletter* articles: geology,[1, 2] mineralisation,[3–6] geochronology,[7] depths of emplacement,[8] and also the soon-to-be-released AGSO Bulletin 246. (See contents list on page 12.)

Geological setting and classification

The layered intrusions are restricted to the central and western zones of the HCO—a well-exposed north-north-east trending (~400 km long by 120 km wide) orogenic belt. The belt comprises variably deformed and metamorphosed sedimentary, volcanic, and intrusive Palaeoproterozoic rocks, and overlapping Proterozoic and Palaeozoic basinal sequences. The intrusions have been assigned to seven major groups (I–VII) on the basis of rock types, U–Pb geochronology, contact relations with country rocks, metamorphic–structural history, types of mineralisation, trace-element chemistry, and Sm–Nd isotopic composition. They form folded sheets, shallow-dipping basinal bodies, composite multi-chambered bodies, funnel-shaped bodies, steeply plunging plugs, fault-bounded blocks, narrow dyke-like bodies, and screens between granite plutons. Chilled and contaminated margins, contact aureoles, ribbon-like comagmatic satellite intrusions, net-vein complexes (resulting from mingling of mafic and felsic magmas), and feeder conduits indicate that the intrusions crystallised in situ, rather than being tectonically emplaced blocks that crystallised elsewhere.[1]

Most intrusions were derived from olivine tholeiite and quartz tholeiite parent magmas of basaltic affinity (with mg number = 100 x Mg/(Mg+Fe^{2+}) = 67 or less). Incompatible trace-element abundances are consistent with compositions ranging from mid-ocean ridge basalt to continental tholeiite.[9] High average S contents (520–1570 ppm) in cumulate rocks for most groups of intrusions indicate early S saturation for the parent magmas. The intrusions were emplaced into the crust at depths ranging from approximately eight to 23 kilometres (2.4–6.7 kb) and there were at least three major periods of emplacement (1855 Ma, 1845 Ma and 1830 Ma). Emplacement of the 1855 million-year intrusions (groups I–III) was contemporaneous with granite plutonism (part of the Bow River Batholith) and felsic volcanism (Whitewater Volcanics) during a major magmatic event that represented a large flux of heat into the crust. Most of the layered intrusions can be classified as late orogenic to postorogenic bodies emplaced during more quiescent phases between periods of active tectonism in a complex orogenic belt.

Mineralisation

Mineral deposits hosted by the layered intrusions can be broadly classified into three major orthomagmatic associations (chromite, sulphide, and Fe–Ti oxide) reflecting an apparent secular evolution of mineralising systems (i.e. Cr–PGEs→Ni–Cu–Co–PGEs→Fe–Ti–V). A fourth association relates to hydrothermal polymetallic deposits.

Chromite association (Cr–PGEs–Ni–Cu±Au)

Chromite is a ubiquitous cumulus mineral in the 1855 million-year, group I mafic–ultramafic intrusions (Panton, Big Ben, West Panton, Highway, Melon Patch, South Melon Patch, West McIntosh, Mini, West Robin Soak*) and in some coeval group II mafic intrusions (e.g. Springvale and Wilagee). In group I, chromite is associated with cumulus olivine in cyclic units of dunite, peridotite and harzburgite. In group II, it occurs with plagioclase in troctolite, anorthosite and leucogabbro. Of greatest economic significance are stratabound chromitite layers at similar stratigraphic levels in the upper parts of the ultramafic stratigraphy of group I intrusions (figure 1). In the Panton intrusion, these chromitites (Middle Group) extend for up to 12 kilometres along strike and are enriched in Pt (up to 2850 ppb), Pd (2470 ppb) and Au (153 ppb) relative to the chromitites from other stratigraphic levels in the intrusion. The chromitites range from single massive isomodal layers up to 2.4 metres thick, to very thin grain-size laminations, stacked cyclic packages of alternating chromitite and dunite layers, and thin discontinuous

* Locations of all intrusions are shown in figure 10 of Trudu & Hoatson 1996.

stringers and lenses. Average Pt+Pd abundances of chromitites for the East Kimberley intrusions decrease from: Panton (Middle Group), Big Ben, Panton (Lower Group), South Melon Patch, West McIntosh, Panton (Upper Group), West Panton, Mini, to Wilagee. This order generally correlates with decreasing thickness of the ultramafic component in the intrusion, implying that the PGE abundances of the chromitites are related to the volume of the ultramafic magma.

The PGE-enriched chromitites generally occur from two to 150 metres below the contact of the ultramafic and overlying mafic zones. There is a correlation between the thickness of these chromitite layers and the thickness of the ultramafic sequence. The thickest layers are in the Panton and Big Ben intrusions, which contain the thickest sequences of olivine bearing cumulates—650 and 1700 metres respectively. PGE-poor chromitite layers are also developed in the lower parts of these thicker bodies. There also appears to be a relationship between the stratigraphic level of the PGE-enriched chromitites and the relative thickness of the ultramafic sequence. For example, the distance of the most PGE-rich chromitite layer below the mafic–ultramafic contact generally increases progressively with increasing thickness of the ultramafic sequence.

Sulphide association (Ni–Cu–Co±PGEs)

Sub-economic concentrations of Ni–Cu–Co±PGE sulphides are hosted by the 1845 million-year group V mafic–ultramafic intrusions (Corkwood, Dave Hill, Fletcher Creek, Keller Creek, McKenzie Spring, Sally Malay, Spring Creek, and Wilson Creek—see figure 1). Massive sulphides generally occur in fractionated gabbroic and noritic rocks, which form basal units five to 40 metres thick, between the overlying peridotite and the paramigmatite footwall rocks. Matrix and disseminated sulphides persist into the overlying peridotite and gabbroic rocks (olivine gabbro, troctolite, leucogabbro, anorthosite). Indicated resources are typically one to four million tonnes @ 1–2% Ni+Cu, ~0.1% Co, ~0.5 ppm total PGEs. There is also significant sulphide mineralisation in mafic granulite bodies of the Tickalara Metamorphics (e.g. Norton and ?Bow River), but these deposits have lower Ni/Cu ratios (0.5:1 to 2:1) than those hosted by the group V intrusions (2:1 to 5:1).

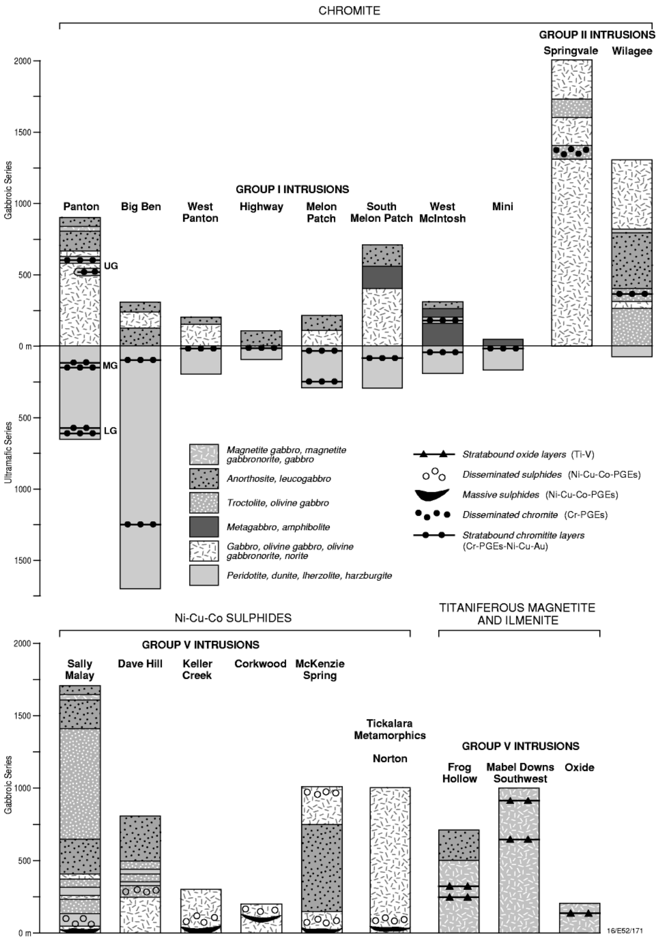

Figure 1. Stratigraphic distribution of chromite, Ni–Cu–Co sulphides, and titaniferous magnetite and ilmenite in the East Kimberley layered mafic–ultramafic intrusions. The vertical axis shows the thicknesses of the Ultramafic Series and Gabbroic Series for each intrusion. Chromitites in the Panton intrusion are shown as Upper Group (UG), Middle Group (MG) and Lower Group (LG).

Sulphide textures in the group V intrusions generally show little evidence of strain or recrystallisation. Those in the granulite bodies, however, consist of discontinuous massive recrystallised bands and thin lenticular segregations subparallel to the foliation of the host metagabbros.

Massive sulphides show a clear association with the lower parts of the stratigraphy—along, or very near, the basal contact of the intrusion or in the feeder conduit. The geometry of the basal contact and conduit is important because the massive sulphides need to be concentrated in restricted environments if economic grades are to be attained. Massive sulphides in the Sally Malay intrusion are concentrated in a keel-shaped depression of an interpreted feeder conduit.[6] In the McKenzie Spring intrusion they occur in a 1600-metre-long, fault-controlled structural embayment that is transgressive to the footwall country rocks. Accumulation of these sulphides is also greatest below the thickest part of the overlying cumulates. The widespread (and thinness of) mineralisation in the Norton and Bow River granulite intrusions indicates that there may not have been a suitable basal depression for concentrating dense sulphides by gravitational processes.

Significantly, the major footwall rocks for all mineralised group V and mafic granulite intrusions are high-grade metasediments (generally garnet–cordierite–sillimanite–spinel–biotite–feldspar–quartz paragneisses) of the Tickalara Metamorphics. The siliceous character and, presumably, ease of assimilation of these rocks make them favourable country rocks for changing the sulphide solubility of the mafic magmas—an important requirement for the precipitation of sulphides.

Fe–Ti oxide association (Fe–Ti–V)

Concentrations of titaniferous magnetite and ilmenite are hosted by gabbroic rocks in the middle to upper fractionated parts of the 1845 million-year group V (Frog Hollow, Mabel Downs Southwest, Ord Crossing, Oxide) and 1830 million-year group VI (McIntosh) mafic intrusions (figure 1). Their distribution highlights the evolved and tholeiitic character of the intrusions. Their association with the younger intrusions is consistent with the general progression from early, more primitive mineralising magmatic systems to later, more evolved systems in the HCO.

Most Fe–Ti–V deposits appear to be small and have little potential for a large-tonnage resource. The Fe–Ti oxides are disseminated or form stacked sequences of thin (0.5–5 cm) cyclic units, thin discontinuous irregular layers, and thicker massive lenses. The Frog Hollow mafic intrusion contains a number of massive oxide lenses (up to 39.4% TiO_2, 0.99% V, 60.6% total Fe as Fe_2O_3, and 0.93% Cr) hosted by ferrogabbro and leucogabbro.[5] The lenses form small, prominent hills, five to 15 metres high, incised by creeks containing abundant oxide alluvium. The coarse grain-size of the Fe–Ti oxides and high TiO_2 grades may reflect high-temperature subsolidus annealing and grain enlargement processes related to later metamorphism. Fe–Ti oxides in the large McIntosh intrusion form disseminations (up to 10 vol %) in troctolite, olivine gabbro and gabbronorite, and thin cyclic units (up to 50 vol %) in leucogabbroic rocks in the upper half of the intrusion.

Hydrothermal polymetallic association (Pd, Pt, Cu, Zn, Pb, Au)

Hydrothermal remobilised polymetallic deposits historically have not been a priority target for exploration in the HCO. However, this province contains many favourable ore-forming components (e.g. fluids, structural conduits, heat and metals) for such deposits. Indicators include widespread serpentinisation of ultramafic sequences, reactivated faults that cut rocks of different composition, multiple phases of metamorphism and deformation, high crustal heat flux from coeval gabbroic and granitic intrusions, and high regional background concentrations of PGEs.

The Emull Zn–Cu–Pb–Ag prospect is an example of hydrothermal remobilised mineralisation spatially associated with a mafic intrusion.[10] This polymetallic prospect (demonstrated resource of 4.7 Mt @ 4.5% Zn, 0.33% Cu, 0.2% Pb, and 19 ppm Ag) occurs near the contact between the differentiated Emull Gabbro and poorly exposed lateritised metasediments of the Koongie Park Formation. The mineralisation consists of a series of en echelon lenses and pods of disseminated and massive sulphides comprising sphalerite–galena–chalcopyrite–pyrite–pyrrhotite in 'serpentinite' and a diopside–carbonate–chlorite calc–silicate rock interpreted by Griffin et al. to be assimilated and metasomatised country rock in the gabbro.[10]

The coeval relationships of many mafic and felsic magmatic systems in the HCO highlights the potential for the secondary redistribution of base and precious metals by volatiles or volatile-enriched magmas (e.g. breccia-type and constitutional zone refining mineralisation described by Barrie[11]). Investigations for these deposits should focus on zones of alteration, hybridisation and brecciation near the contacts of coeval granite and gabbro bodies. Since serpentinisation alteration is an efficient mechanism for concentrating PGEs, thick sequences of serpentinised and faulted olivine-rich cumulates should also be targeted.

Exploration strategies

Some of the more important considerations when exploring for PGE-enriched chromitite layers and basal segregations of sulphides in the HCO are listed below.

Stratabound PGE-enriched chromitite layers

Group 1 differentiated intrusions (figure 2a) in the central zone that display open fractionation systems with repeated large pulses of primitive magma into the chamber (Panton, Big Ben, West McIntosh, Melon Patch group) are thought to be the most prospective. Intrusions formed from the most primitive magmas (mg >58, corresponding to olivine compositions with mg >82) and which contain thick sequences of olivine cumulates should be high priority.

The most PGE-enriched chromitites generally occur in the stratigraphic interval two to 150 metres below the contact of the Ultramafic and Gabbroic Series.

Detailed geological mapping and closely spaced sampling (initially 10–20 m spacing) are needed to determine magma-chamber processes that are important for the formation of chromitite layers, such as the mixing of primitive magma with more fractionated magma. Magma mixing can be indicated by:

- rapid changes in the composition of cumulus minerals and the volume of intercumulus melt in the cumulates (rock porosity) across a layered sequence;
- development of porphyritic, pegmatoidal and orthocumulus textures in associated rocks;
- occurrence of hybrid rocks and accidental blocks in slump deposits;
- lateral variation of rock types, such as interfingering, thinning or thickening; and
- unconformities representing magmatic erosion/disruption of footwall sequences.

Olivine and/or orthopyroxene cumulates are more favourable than clinopyroxene cumulates for hosting stratabound chromitite layers because of the relative high partitioning of Cr into clinopyroxene.

Proximity of chromitite layers is indicated by the presence of disseminated chromite in associated ultramafic cumulates, blocks of chromite in soil, detrital chromite grains in stream sediments, and magnesite alteration of host ultramafic cumulates. Major geochemical pathfinder elements are Cr, Pt, Pd, Cu, Au, Mg, mg number, and Ni/Cu ratios.

Narrow thicknesses (most <0.5 m) and low primary magnetite and sulphide contents of the chromitite layers means electrical methods (e.g. electromagnetics, induced polar-isation) have limited application (in contrast to basal Ni–Cu–Co sulphide deposits).

Aeromagnetics, gamma-ray spectrometry, and Landsat imagery (figure 2b) are useful for defining macroscopic features of intrusions, namely:

- aeromagnetics can help determine

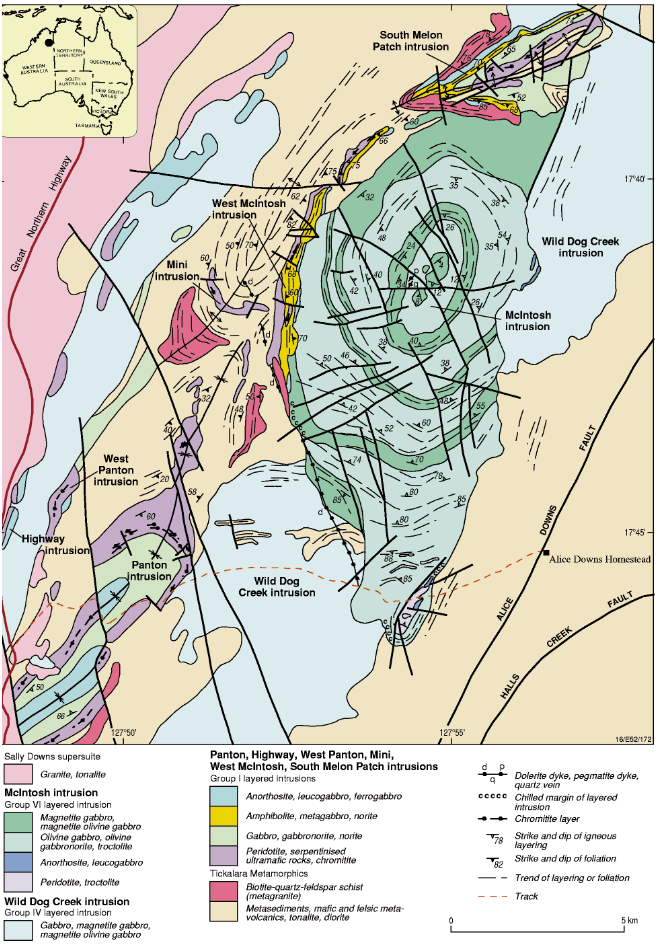

Figure 2a. Geological map of the group I (Panton, Highway, West Panton, Mini, West McIntosh, and South Melon Patch) mafic–ultramafic intrusions, group IV (Wild Dog Creek—part of) mafic intrusion, and group VI (McIntosh) mafic intrusion.

gross younging directions in gabbroic zones (coarse-grained primary Fe–Ti oxides) and the intensity of serpentinisation in ultramafic zones (fine-grained magnetite after olivine);

- composite gamma-ray spectrometry can define the regional distribution of mafic and ultramafic rocks, since these have low concentrations of K, Th and U; and
- Landsat 5 Thematic Mapper imagery, in particular, band-ratio (5/7, 5/3, 5: figure 2b) and directed principal component (2(4/3,5/7), 5/4, and 1+7) images are extremely useful for discriminating ultramafic, mafic and felsic rock types, and identifying favourable contacts between ultramafic and gabbroic sequences.

Basal segregations of Ni–Cu–Co±PGE sulphides

Differentiated layered mafic–ultramafic intrusions of group V (Sally Malay, Corkwood, Keller Creek, Wilson Creek, McKenzie Creek, etc.) and large massive mafic granulite bodies of the Tickalara Metamorphics (Norton) are considered the most prospective.

Establish whether the magma(s) that formed the intrusion were S saturated by determining:[12]

1. parent magma compositions (from chilled marginal rocks, comagmatic dykes and sills);
2. modal distribution of sulphides in cumulates and normalising S content of the intercumulus liquid (trapped melt) in unaltered cumulates to 100 per cent melt; or
3. S/Se ratios of unaltered cumulates. In general, most basaltic parent magmas with S content exceeding 1000 ppm are S saturated. If the intrusion is in part S undersaturated, the stratigraphic level of S saturation will need to be determined for stratabound PGE-enriched sulphide layers (cf. Munni Munni Complex-type in the west Pilbara Craton[12]).

Favourable country rocks are paramigmatites (Tickalara Metamorphics) and metasediments (Halls Creek Group), rather than metavolcanics or granites. Assimilation of metasediments containing S-bearing minerals (e.g. pyrrhotite, pyrite, gypsum, anhydrite) and graphite may induce the precipitation of sulphides in the contaminated magma.

Obtain evidence for crustal contamination of magma—for example:

- association of sulphides and alkali minerals;
- fractionated marginal rocks;
- abundance of accidental xenoliths and xenocrysts in the basal zones;
- mixed Sm–Nd and Re–Os isotopes;
- S-isotope compositions outside normal magmatic values;
- variable compatible versus incompatible elemental ratios in an intrusion; and
- different crystal fractionation trends in the same magmatic province.

Field investigations should carefully reconstruct the configuration of the basal contact, particularly below the thickest part of the mafic–ultramafic cumulates. Depressions and fault-controlled structural embayments are important for the accumulation of massive sulphides from gravitational and/or crustal contamination processes.

Feeder conduits to the intrusions are also favourable environments for the concentration of sulphides. The dynamics of the magma flow may be important for the deposition of massive sulphides; for example, fluid dynamic contrasts (fast, slow, turbulent, passive) associated with the change from narrow vertical conduits to broad sub-horizontal open magma chambers.[13]

Remobilised discordant and

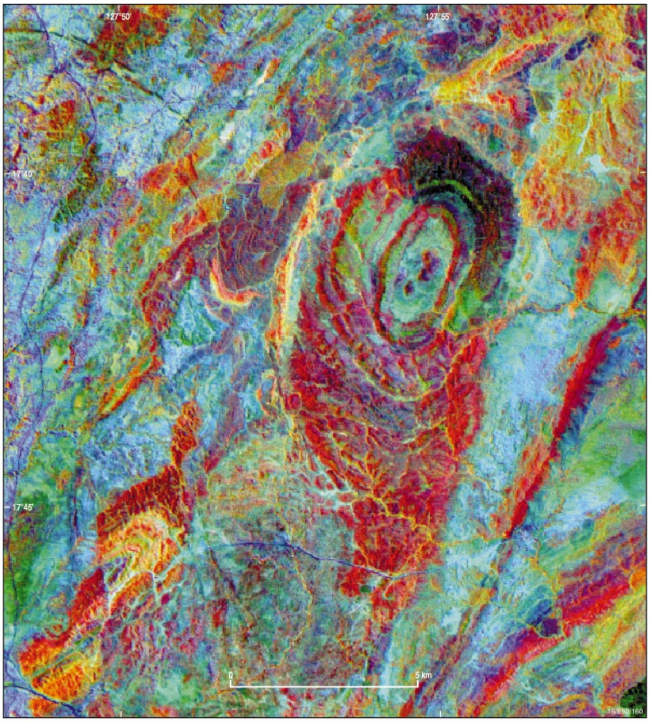

Figure 2b. Landsat-5 TM (path 107 row 72) colour composite 5/7 (to discriminate clays), 5/3 (iron oxides) and 5 (vegetation) image, displayed as red, green and blue, respectively, for the same area shown in Figure 2a. The image highlights the differentiated nature of the mineralised Panton and smaller group I intrusions further north, and the prospective contact region (for PGE-enriched chromitites) between the ultramafic cumulates and overlying gabbroic cumulates. The dark area in the north-east of the McIntosh intrusion is a fireburn. Image generated by LF Macias, AGSO.

fractionated ores enriched in Cu, Au, and/or Pd can occur far from the basal contact of the intrusion or in the country rocks.

Gossan search, stream sediment sampling, and rock geochemistry are recommended geochemical techniques. As well, electromagnetics, ground magnetics and induced polarisation are useful since most ore types contain pyrrhotite and magnetite. Aeromagnetics and gravity can delineate the regional extent, geometry and major structures of poorly exposed intrusions; gravity, subject to sufficient density contrast with the country rocks, can define feeder conduits.

Economic potential

The East Kimberley layered mafic–ultramafic intrusions are generally regarded as having greater potential for hosting an economic resource of Ni–Cu–Co sulphides rather than for PGEs associated with chromitite layers. Economic chromitite layers, such as those in the Great Dyke of Zimbabwe and the UG-2 in the Bushveld Complex of South Africa, typically have strike lengths of tens to hundreds of kilometres. They are hosted by large, layered bodies containing very thick sequences of olivine and orthopyroxene cumulates—none of which is typical of the East Kimberley intrusions. The economic viability of these deposits is usually dependent on the uniformity of PGE and Cr grades over long strike distances.

A major problem relating to the chromitite layers in the HCO is their variable morphology along strike. For example: the major chromitite layers in the Panton intrusion vary in thickness from a few centimetres to 240 centimetres over a few hundred metres along strike; are locally tightly folded; vary from a single massive layer to a package of several thin layers; and are displaced by high-angle cross-cutting faults. All these features potentially affect the lateral continuity of ore grades and hinder mining operations. In contrast, economic Ni–Cu–Co sulphide deposits are often hosted by small- to medium-sized mafic intrusions that may be layered or massive (common in the HCO). These types of deposits are also less dependent on the volume of magma and the dynamics of magma flow.

One of the major challenges in exploring for Ni–Cu–Co sulphide deposits in the HCO is to locate prospective intrusions that may be covered by shallow alluvium or underlie thick sequences of country rocks. The identification of two parallel north-east-trending metallogenic corridors containing the major mineralised intrusions (groups I, II and V) of the HCO will help focus exploration for these deposits.[8] Large prospective areas covered by shallow alluvium occur within these corridors and also in the southern parts of the HCO between Louisa Downs and Halls Creek. The identification of favourable basal contact and feeder conduit environments in these poorly exposed intrusions will most likely be achieved through advanced geophysical (airborne electromagnetic and gravity) and remote-sensing (e.g. ASTER—Advanced Spaceborne Thermal Emission and Reflection Radiometer) techniques in association with closely spaced grid drilling and, where possible, detailed mapping. Hoatson et al. have highlighted the potential of the East Kimberley and other Proterozoic provinces in Australia, such as the Musgrave Block in central Australia, for Ni–Cu–Co sulphide mineralisation associated with mafic–ultramafic intrusions.[6]

Other potential styles of mineralisation in the HCO that should be considered include:

- structurally controlled hydrothermal and constitutional zone refining mineralisation (such as Lac des Iles Intrusion, Ontario[11]);
- Ni–Cu–PGE sulphides in mafic sills related to flood basalts (such as Noril'sk-Talnakh, Russia[14]); and
- 'stratabound' Platinova Reef-type Au–PGE mineralisation (such as Skaergaard Intrusion, east Greenland[15]) related to ascending fluids in the upper parts of thick mafic intrusions.

Concluding statement

Exploration programs for Precambrian layered mafic–ultramafic intrusions vary considerably for different styles of precious and base-metal mineralisation. It is important not to be 'blinkered' by a particular model, but to maintain a flexible innovative approach and consider different styles of orthomagmatic and hydrothermal mineralisation at different stratigraphic levels in the intrusion.

It took more than 20 years of intensive exploration to define the J-M Reef of the Stillwater Complex, and it was not until late in the 1990s that a significant Au–PGE layer (Platinova Reef) was found in the Skaergaard Intrusion, east Greenland—an intrusion that had been investigated in great detail for more than 60 years.

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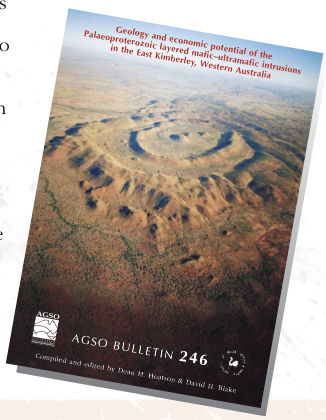

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Geology and economic potential of the Palaeoproterozoic layered mafic–ultramafic intrusions in the East Kimberley, Western Australia

A geological systems approach to understanding the processes involved in land and water salinisation

The Gilmore project area, central-west New South Wales

KC Lawrie, TJ Munday, DL Dent, DL Gibson, RC Brodie, J Wilford, NS Reilly, RA Chan & P Baker

Australia is experiencing widespread degradation of land and water as a consequence of salinisation, and the problem is growing. Urgent action is required to derive appropriate management strategies that will minimise future socioeconomic and environmental impacts of salinity. However, while the causes of dryland salinity have been established, processes operating at a catchment scale are not as well understood, and a new approach that integrates all the relevant datasets is required for mapping and predicting salinity.

A multi-disciplinary 'geological systems' approach has been tested to better map and quantify key physical properties/factors and processes that determine the susceptibility of a given catchment to salinisation. The approach benefits from the recent development of new geophysical technologies—notably airborne electromagnetics (AEM) systems (e.g. TEMPEST), and significant improvements in AEM data processing software (e.g. EMFlow) and data visualisation techniques. New insights into landscape evolution have been equally important. The new airborne geophysical datasets, calibrated by surface and targeted drillhole information, permit rapid delineation of the sub-surface geology and regolith architecture, and salt distribution. These datasets allow linkages between salt stores and groundwater distribution systems to be examined, and the salt hazard to be assessed.

The pilot study (Gilmore project) area straddles the Lachlan/Murrumbidgee watershed in the Murray–Darling Basin, central-west NSW. The area contains a complex regolith developed in floodplain, incised undulating hill and upland landscapes. Ground calibration of AEM datasets has enabled the spatial and vertical conductivity structure to be modelled with reasonable precision. Previously unidentified salt stores have been recognised, and saline–sulphate and 'fresh' groundwater delivery systems mapped in the sub-surface. Saline groundwater has been identified in transported materials and in saprolite. Differential weathering and erosion of bedrock has produced a complex buried palaeo-topography that controls valley-fill architecture. Lateral groundwater flow occurs through a network of sand and gravel-filled palaeo-channels and buried alluvial fans within Cenozoic cover sediments. Narrow, restricted connections between buried (drainage) basins confine lateral groundwater flows. A similar geological systems approach in upland areas, requiring a different mix of technologies, has delineated sub-catchments where saline groundwaters are sourced and stored.

Some three million hectares of Australian farmland suffer from dryland salinity, or saline seepage, caused by saline groundwater rising to the surface. The costs in terms of lost agricultural production and infrastructure are substantial.[1] The costs of contamination of shallow freshwater resources and the impact on freshwater ecology are also significant. In NSW alone, approximately 120 000 hectares of land are affected by salinity,[2] with a further five million hectares considered at risk.[3] The salt concentration in many streams and rivers, particularly in the southern half of the Murray–Darling Basin, is thought to be rising steadily.[3, 4] Urgent action is required to derive appropriate management strategies that will minimise the future socio-economic and environmental impacts of salinity.

The accumulation of salt in the regolith is a natural phenomenon that has occurred in Australia over many thousands (or millions) of years. The processes responsible for the development of saline land and water are complex, and relate to the chemical processes of weathering, deposition and redistribution of soluble salts in groundwater flow systems.[5] Most salt originated from the oceans. It was deposited in rainfall and redistributed in the landscape through surface and groundwater flow over time. Dryland salinity has been attributed to human disruption of the hydrologic cycle. Clearance of native vegetation and a system of agriculture dependent on shallow-rooted annual crops and pastures that use less water than the natural vegetation have resulted in an increase in groundwater recharge. The consequence is rising watertables and in places a mobilisation of salts stored in the regolith. Saline waters flow to lower parts of the landscape along preferential paths. Where watercourses (including major rivers) intercept these seepages, their salt loads are increased. Where saline groundwaters reach close to the soil surface, they restrict crop growth and damage roads and buildings.

While the causes of dryland salinity have been established, processes operating at a catchment scale are not as well understood. A new approach therefore is required for mapping and predicting salinity. A new methodology, termed a 'geological systems' approach, takes advantage of new geophysical technologies and increased understanding of landscape evolution. It provides a better understanding of processes responsible for salinisation in areas of complex regolith cover.[3, 6, 7] The methodology draws on previous studies that recommended an integrated approach to the problem, including the use of high-resolution airborne geophysics.[3, 6] Initial results from the application of the geological systems approach in central-west NSW suggest the methodology has particular relevance in areas of complex bedrock geology, regolith cover and landscapes considered prone to salinisation, such as that of the Murray–Darling Basin of eastern Australia.

Geological systems approach

Conventional methods of assessing salinity include soil surveys, airphoto interpretation, regional geological mapping, ground geophysics and drilling. In areas of complex regolith cover in particular,[7] these methods do not allow the building of an accurate three-dimensional sub-surface picture of the spatial distribution of saline groundwaters and/or the flow systems that deliver salts to discharge sites. The integration of conventional datasets with ground and borehole geophysics and studies of drillhole materials has similarly been hampered by the difficulty of interpolation between calibration points.

New methods are needed to provide a sound basis for management of salinisation. The National Airborne Geophysics project concluded that airborne magnetics has the ability to map geological structures not always apparent from outcrop or airphoto interpretation, and to map geological structure at a paddock scale.[8] Airborne electromagnetics (AEM) have the potential to map the sub-surface distribution of salt and variations in the nature of regolith materials.[6] However AEM data is difficult to interpret, with high conductivity measurements attributable to a complex interplay between saline and/or sulphate-rich waters and the host regolith materials. Expert teams are required to derive meaningful results and to avoid spurious conclusions.[9]

A holistic, multi-disciplinary geological systems approach, previously recommended by a national study,[6] is being tested in central-west NSW.[10] This builds on a methodology recommended in development of a national catchment classification scheme.[3] A key objective is to better map and quantify key physical properties/factors that determine the susceptibility of a given catchment to salinisation. These factors include:

- the hydrogeomorphology;
- bedrock architecture including composition and structure;
- regolith framework including palaeo-topography, sediment facies distribution, saprolith thickness, and saprolith and sediment composition, texture and fabric;
- the distribution and composition of soils;
- the spatial distribution, connectivity and hydraulic conductivity of groundwater flow systems;
- salt sources and stores and their connectivity to the groundwater distribution systems; and
- the identification of recharge and discharge areas.

Critically, the geological systems approach provides a better understanding of the groundwater aquifer systems, and their connectivity and spatial variability. In turn, this provides a framework within which the impact of other more variable factors such as land use, vegetation type/condition, climate, palaeo-climate, and groundwater recharge rates can be considered.[3] These datasets should help constrain groundwater distribution systems, and the water–rock interactions that lead to salinisation in a catchment, and assist with construction of predictive models.

The methodology integrates geophysical, bedrock geological, regolith, soil, hydrogeological and hydrogeochemical datasets. The approach benefits from recent development of significantly improved geophysical technologies, notably AEM systems (e.g. TEMPEST[11]), parallel improvements in processing software (e.g. EMFlow[12]), and data visualisation techniques[13]. High-resolution airborne magnetic data are also useful in delineating bedrock structure and (where there is a magnetic sediment fill) for identifying palaeo-channels that may be part of a sub-surface drainage network.[14, 15]

Airborne geophysical datasets assist in the rapid delineation of the sub-surface geology and regolith. Of particular importance is their value in helping determine the architecture of aquifer and groundwater flow systems in the regolith. However, experience has emphasised the importance of ground calibration to help constrain the modelling of geophysical data and the hydrogeology. This necessitates gathering available surface data, and a limited aircore and diamond drilling program to recover pore fluids and materials. Multi-parameter geophysical and geological logging of boreholes is deemed essential for calibration and modelling of AEM and airborne magnetic datasets.

New insights into landscape evolution have been equally important. It is generally recognised that the regolith (and bedrock) framework exerts critical controls on salt distribution and hydrogeological models of groundwater flow and salinisation in areas of complex regolith cover.[16] The integrated datasets allow linkages between salt sources and stores, groundwater distribution systems to be examined, and the salt hazard to be assessed. An expert decision-support system is required to reduce the complexity of the data and deliver an effective toolbox for land managers and communities. This approach has the potential to provide a basis for defining appropriate management options from paddock to catchment scales.

Gilmore project

The geological systems approach is being tested in a pilot study (the Gilmore project[17]) in an area on the eastern margin of the Murray–Darling Basin in central-west NSW. The project area was chosen because of its overlapping mineral exploration (Au–Cu) and salinity management issues, the availability of high-resolution geophysical datasets and drillhole materials, and datasets available from the minerals exploration industry.

The project, coordinated by the Australian Geological Survey Organisation, involves more than 50 scientists from 14 research organisations. Research partners include:

- Cooperative Research Centres for Advanced Mineral Exploration Technologies (CRC AMET) and Landscape Evolution and Mineral Exploration (CRC LEME), the CRC for Sensor Signal and Information Processing, and the Australian Geodynamics Cooperative Research Centre (AGCRC);
- Land and Water Sciences Division, Bureau of Rural Sciences (BRS);
- NSW Department of Land and Water Conservation and Department of Mineral Resources;
- various universities including the Australian National University, University of Canberra, Macquarie University, Monash University, University of Melbourne and Curtin University of Technology; and
- Australian National Seismic Imaging Resource (ANSIR).

The project has research agreements with the minerals exploration industry and is collaborating with rural land-management groups and the Grains Research and Development Corporation.

The study area (100 x 150 km), straddles the Gilmore Fault Zone, a major north-north-west-trending crustal structure that separates the Wagga–Omeo and the Junee–Narromine Volcanic Belts in the Lachlan Fold Belt. The project area includes tributaries of the Lachlan and the Murrumbidgee Rivers (figure 1) which are considered two of the systems most at risk from rising salinities. In parts of the

Murrumbidgee, salinities are increasing at 15 per cent per annum. This is having significant impact on downstream uses such as drinking water and irrigation.[3] The project area was chosen to compare and contrast salt stores and delivery systems in floodplain (in the Lachlan catchment) and incised undulating hill landscapes (Murrumbidgee catchment). The study area is characteristic of other undulating hill landscapes on the basin margins, areas within the main and tributary river valleys, and the footslopes and floodplains of the Murray–Darling Basin itself.

The bedrock geology in the study area has a complex architecture. The Gilmore Fault Zone consists of a series of subparallel, west-dipping thrust faults[18] that juxtapose, from west to east, Cambro-Ordovician meta-sediments and granites of the Wagga Metamorphics and, further to the east, a series of fault-bounded packages comprising volcanics and intrusions, and siliciclastic meta-sediments. Large-scale hydrothermal alteration and structural overprinting, particularly in the volcanics, has added to the complexity within the bedrock architecture.

Two AEM surveys were flown in smaller areas within the two catchments. In the Lachlan catchment, alluviation of the palaeo-Lachlan River system during the late Tertiary[19] largely buried the Bland Creek palaeo-valley. Up to 120 metres of sediment infill is recorded in the north-east of the AEM survey area. However, sediment thickness is markedly variable because of complex bedrock palaeo-topography. This has resulted from differential weathering and erosion of bedrock lithologies in the palaeo-landscape.

In the Murrumbidgee catchment, there are undulating hills incised by two main valleys largely infilled with up to 60 metres of transported sediments. There is an elevation difference of 100–200 metres from current valley floors to the crest of local divides. Siluro-Devonian volcanic and intrusive lithologies form higher hills in the east. The two main present-day creek systems within the western and eastern valleys, Houlaghans and Billabung Creeks respectively, flow south to the Murrumbidgee River.

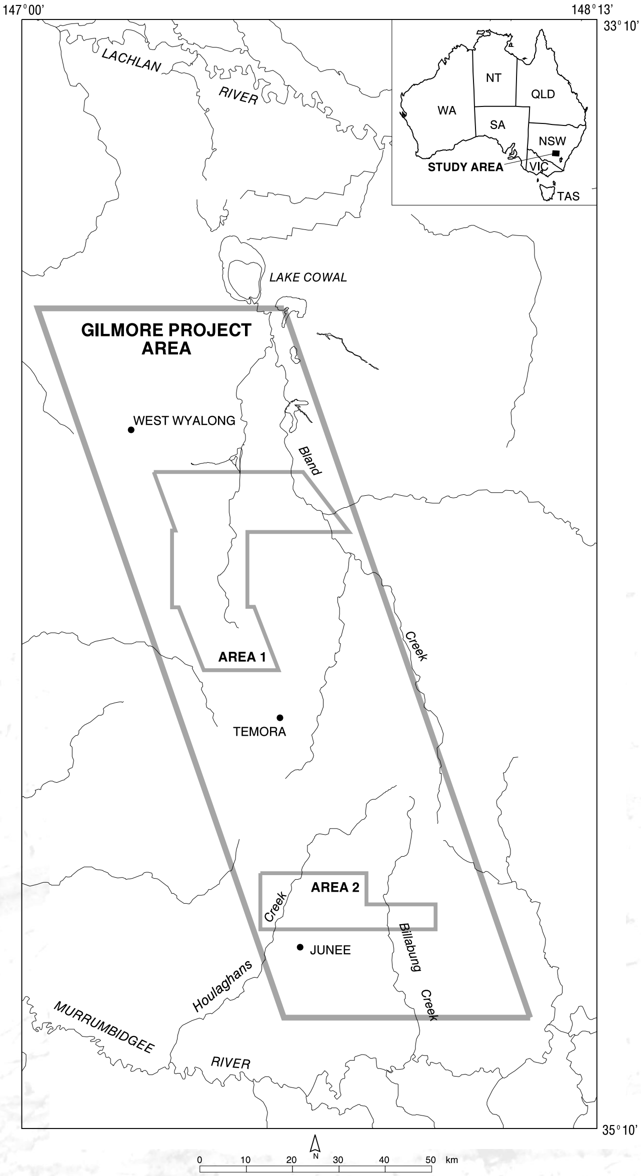

Figure 1. Diagram showing the location of the Gilmore project area relative to the Lachlan and Murrumbidgee Rivers. The two smaller AEM survey areas are also outlined.

Datasets

The Gilmore project has acquired new datasets including high-resolution airborne geophysical surveys, ground geophysics, hydrological datasets (surface stream flow and sub-surface borehole data), and regolith and bedrock geological mapping from extensive industry and government-funded drilling. Regional digital orthophoto and high-resolution digital elevation models provide an accurate reference framework for field studies and GIS referencing. Ground geophysical surveys, including shallow-penetrating seismic reflection,[20] seismic refraction,[21] ground penetrating radar, microgravity and electromagnetic surveys were carried out to attempt to calibrate palaeo-channel features evident on airborne survey datasets.

Hydrological datasets include surface stream flow and sub-surface borehole data measurements.[10] Multi-

➡ *Continued page 26*

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parameter geophysical borehole data acquisition comprised induction, gamma-ray and magnetic logs. Mapping of bedrock and regolith materials in 3D, analysis by Portable Mineral Infrared Analyser (PIMA), X-ray diffraction (XRD), X-ray fluorescence (XRF) and thin section studies of regolith and bedrock materials also were included.

At the start of the project, the minerals exploration industry gave access to core and chip materials from hundreds of drillholes within the study area. After examination of these materials, it was decided that a new drilling program was required to recover both regolith materials and pore fluids in order to calibrate the spatial and vertical conductivity structure within the AEM data. A drilling program was carried out between March and May 2000. Many of the holes are being completed as piezometer monitoring stations. Three drilling methods were used:

- a Rotary Air Blast (RAB) drilling program of 15 short holes was undertaken in upland areas and in the floodplains and incised valleys where drilling results from depths less than 25 metres were anticipated;
- an aircore drilling program of more than 30 holes drilled to saprock where possible (maximum depth 100 m); and
- a diamond core drilling program of five holes (maximum depth 80 m).

Drillhole materials were collected for laboratory analysis at one-metre intervals. Drilling with an air compressor rather than using water- or oil-based fluids was considered essential to help preserve materials and pore fluids.

To date, detailed descriptions, and PIMA, XRD and XRF analysis of regolith materials, limited grain-size analysis, moisture and electrical conductivity (EC1:5) measurements, and S:Cl ratios of pore fluid analysis have been carried out. Preliminary analysis of borehole fluid compositions has identified both fresh and saline (up to 2500 mS/m—half seawater salinity) groundwaters. EC1:5 and compositional analysis show that the pore fluids have saline–sulphate compositions, with S:Cl ratios varying from 1:6 to 1:1 (Astolfi E, 2000, pers comm). Regolith descriptions have been compared to downhole geophysical logs and the pore fluid and mineralogical data.

These data were acquired to compare and contrast salt stores and delivery systems in floodplain and incised undulating hill landscapes, and to develop a methodology for rapidly delineating areas in upland landscapes that generate and store saline groundwaters. Some preliminary results of the integration of Gilmore project datasets are given below.

TEMPEST AEM survey

From January to March 1999, 7721 line kilometres of electromagnetic (and aeromagnetic) data were acquired over two areas in the Gilmore project area using the TEMPEST system.[11, 22] The TEMPEST data set for Gilmore was transformed to earth conductivities using one-dimensional layered earth inversions (LEI)[23] and conductivity at depth, as well as spatial images of derived parameters such as average conductivities over nominated depth intervals (interval conductivities) and conductive unit parameters (e.g. thickness of conductive layer).

The inversion results were presented as both vertical cross-sections and spatial images. The vertical cross-sections are generated by splicing together results from individual inversions at each sample point (12-m intervals) along each flight line.

Layered earth inversions

A layered earth inversion (LEI) of the TEMPEST data was based on procedures previously described.[23] The conceptual model used was a three-layer earth consisting of a conductive layer sandwiched between a resistive upper layer and an infinitely thick resistive layer below. The inversion process calculates the conductivity of all three layers and the thickness of the upper two layers that best represents the observed data. In a geological sense, this model could represent a resistive alluvium–colluvium, overlying conductive transported material or saprolite, which sits over a resistive unweathered basement.

In the Gilmore data, the layered earth conductivity-depth sections show good line-to-line correlation. This suggests that at the scale of the AEM survey (hundreds of metres laterally and a few metres vertically), the electrical structure across the study area is essentially layered. The best results were obtained in the northern survey area, where much of the landscape is characterised by an extensive layer of conductive regolith material comprising both transported and in-situ regolith materials. In these landscapes the conceptual model is suitable and the results are effective. In areas of extreme conductance, the chosen starting model influences the inversion results by increasing the thickness of the resistive first layer and reducing the conductance of the second layer. This problem was encountered particularly in the northern survey area.

Conductivity depth imaging

EMFlow software, which is based on the approximate source image algorithm for layered earths, was used to produce a set of CDIs.[12, 24] CDIs allow subtle conductivity variations to be modelled more realistically by comparison with those produced from three-layered model inversions as described above. Interval conductivity slices, formed by averaging conductivity values over a discrete interval, formed the principal product for interpreting spatial variations in conductivity as a function of depth across the study areas. An 'EM response map' of the Gilmore area is essentially a map of the variation in the depth to, and thickness and/or conductivity of, the transported cover and saprolite overlying varying basement lithologies.

Preliminary results

Comparison of LEI and CDI models of the Gilmore AEM data shows good agreement in the spatial conductivity distribution. Figure 2 (a–d) shows conductivity depth slices generated from EMFlow-modelled CDIs for the northern AEM survey areas. The CDI models have been validated by borehole measurements. Borehole induction logs and moisture and EC1:5 measurements of regolith materials were compared with vertical conductivity profiles displayed on conductivity-depth sections derived from both CDI and LEI modelling of the AEM data (e.g. figures 3a, 3b). Validation using borehole data assumes that variations observed in the drill hole data are representative of variations measured by an AEM system. However, an AEM system with the geometry of the TEMPEST AEM system resolves conductivity variations of the order of 100 metres horizontally and tens of metres vertically. With the drillhole data, variations of the order of more than one metre are measured. Therefore, when comparing parameters such as conductivity or regolith thickness derived from the inversion of the AEM data with those obtained from drill-hole measurements, the scales at which these natural variations are likely to occur becomes important—particularly when interpreting the results.

By their very nature, LEIs do not give adequate definition of the subtle differences in vertical conductivity structure observed in borehole

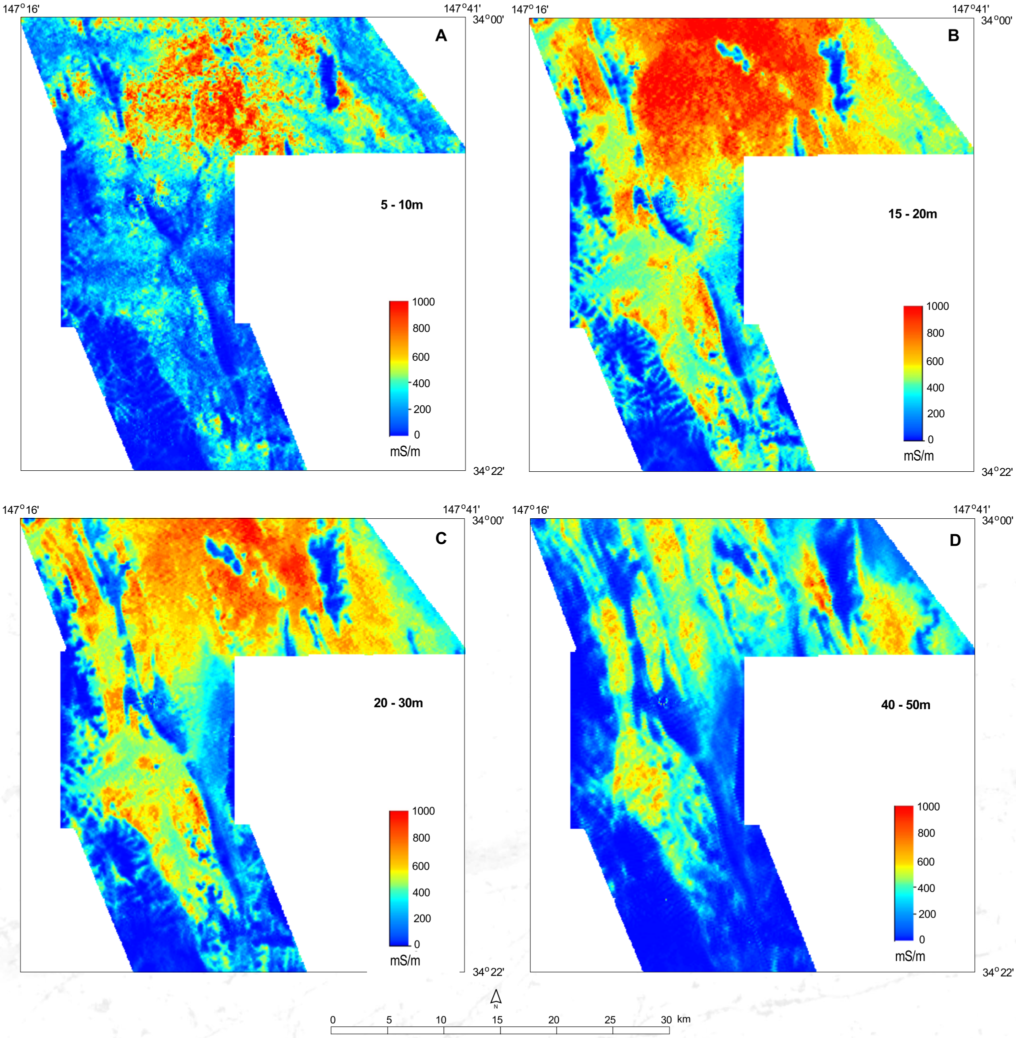


Figure 2. Conductivity depth slices for the intervals **a.** 5–10 metres **b.** 15–20 metres **c.** 20–30 metres and **d.** 40–50 metres, generated from EMFlow-modelled CDIs for the northern AEM survey area. Higher conductivities are indicated in red colours; resistive areas are blue.

measurements within the conductive layer. Hence no discrimination is possible between conductive transported sediments and saprolite. However, comparison with borehole data indicates that, overall, there is a good correlation between the modelled base of the conductor and the saprolite–saprock interface (e.g. figure 3b).

Initially, a poor vertical correlation was obtained from CDI models using standard survey parameters derived from experience in Western Australia, and EMFlow software. Re-examination of survey acquisition parameters, software modelling inputs, and datum corrections followed. A significant improvement in modelling vertical electrical structure was obtained. The re-calibration of survey parameters and EMFlow model inputs has resulted in a correlation coefficient of 0.79 between average borehole-conductivities and CDI model conductivities over 10-metre vertical intervals (figure 4).

The limitations of airborne systems in imaging the conductivity structure in the top 10 metres has been noted previously.[6] However in this study, significant improvement in the modelling of the vertical conductivity structure was produced by corrections to survey geometry (on the basis of comparison with borehole induction logs), height corrections to radar altimeter readings, and both EMFlow inputs and output use. There is now reasonable confidence that the spatial and vertical patterns observed in the CDI-modelled AEM data are a good representation of the conductivity structure within five to 10 metres of the surface (figure 2a), and that the zero- to five-metre depth slices are representative of bulk conductivities over this interval.

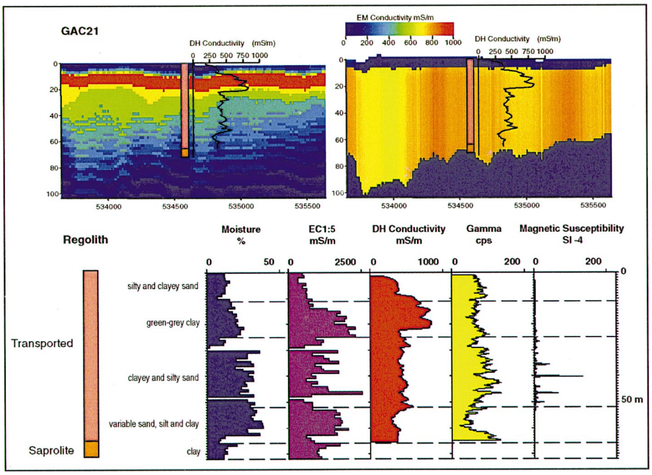
Problems still exist with modelling more complex vertical conductivity structures in the data, particularly in areas where several highly conductive layers are stacked above one another. This occurs in the northern survey area, where there are saline–sulphate groundwaters in near-surface aquifers and at depth in saprolite. In these cases the vertical resolution of the AEM system is simply not high enough.

Floodplain landscapes

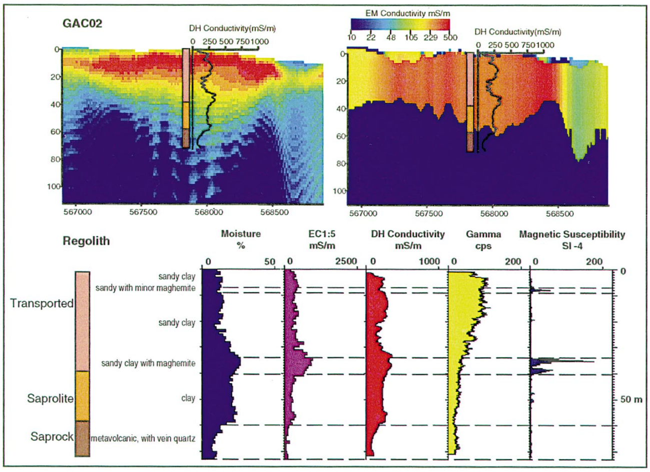
The northern area lies within a shallow inland basin, the Bland Creek palaeo-valley. It is a north-south-trending palaeo-valley system 60 kilometres across and 130 kilometres long (figure 1). This palaeo-valley controlled the discharge of Tertiary palaeo-rivers northwards into the main westward-flowing Lachlan River system. The northern AEM survey area lies within the western flank of the palaeo-valley, in an area of relatively flat alluviated plains with a few low hills. The latter consist of granites or silicified hydro-thermal alteration zones associated with Au and Au–Cu deposits. North-north-west-trending, discontinuous topographic ridges consist of siliciclastic meta-sediments and/or granites. The streams flowing from the hills mostly disappear into alluvial fans or into the alluvium of the flood plains. The main north-flowing stream, Bland Creek, varies in salinity, receiving low salt waters from its left bank but, occasionally, very high salinity waters from the right bank.[10]

Mineralogical, grain-size and textural (fabric) analysis of regolith materials are compared with EC1:5s and moisture content of materials, and multi-parameter geophysical borehole logs results (e.g. figures 3a, 3b). These data displays are used to help with interpretation and correlation of regolith materials and to calibrate models of geophysical data. Preliminary results suggest that the sediment infill in the Bland palaeo-valley involved deposition in low-angle aggrading fans and in palaeo-river systems. Present-day analogs occur on the perimeter of the present Bland alluvial plain. Significant vertical and lateral variations in clay mineralogy are mapped. Throughout the study area, a zone of weathered bedrock (saprolite) that varies from 10 to 100 metres thick, forms the base of the regolith. Preferential groundwater flow is interpreted as being through more transmissive sediment-fill in buried palaeo-channels, and by inter-layer flow in alluvial fan deposits, and at the transported sediment–saprolite interface, and through fracture networks (macropores) in saprolite.

3a



3b



Figures 3a & 3b. Borehole induction logs, and moisture and EC1:5 measurements of regolith materials were compared with vertical conductivity profiles displayed on conductivity-depth sections derived from both CDI and LEI modelling of the AEM data. Mineralogical, grain-size and textural (fabric) analysis of regolith materials are compared with EC1:5s and moisture content of materials, and multi-parameter geophysical borehole logs results. These data displays are used in interpreting and correlating regolith materials and calibrating models of geophysical data.

Interval conductivity slices of the AEM data reveal a complex pattern beneath the valley and floodplain surfaces (figure 2a–d). A geological systems approach to interpreting these patterns is essential. Ground-truthing of AEM data demonstrates that conductivity structure exhibits a strong lithodependence. A layered vertical structure is evident with, in general terms, a highly conductive sediment infill overlying a variably conductive saprolith, and resistive bedrock. This is evident in the conductivity depth slices which show increasing correlation with bedrock geological lithology distribution with depth (figure 2).

A very thin (<1 m) upper resistive soil layer is present over much of the floodplain. Within the transported sediments, AEM images displayed as interval conductivities suggest that extensive stores of saline (and sulphate-rich) groundwater are perched within clay-rich sediments within five to 30 metres of the surface. Spatial patterns in the AEM data that relate to variations within the transported regolith sediments are caused by a complex relationship between saline–sulphate groundwater content and variations in the nature of the saprolite and/or sedimentary facies.

A strong lithodependence is observed between bedrock lithologies and the later-time AEM data (figure 2d).[25] The Wagga Metamorphics, the Combaning Formation (Fm), and Late Devonian intrusions correlate with regolith landform units that comprise topographic highs and form relatively electrically resistive ridges. In contrast, volcanics, intrusions and meta-sediments of the Gidginbung and Belimebung Volcanics have a more conductive electrical structure. Drilling results show that the conductive areas correlate with an increase in conductive transported overburden and an increase in the total depth to base of saprolite.

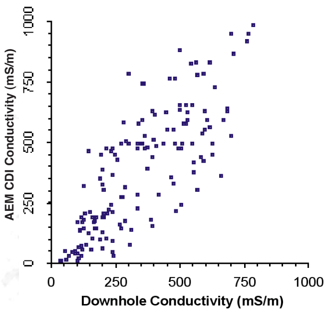


Figure 4. A correlation coefficient of 0.79 is observed between average borehole-conductivities and CDI model conductivities over 10-metre vertical intervals.

Although there is a significant lithodependence between the volcanic/intrusive belts and high conductance, a second-order patterning is also discernible in the data. Traverses along the Gidginbung Volcanics reveal a complex patterning at the same depths below surface. Areas of higher near-surface conductance appear to correspond with saprolite that is not siliceous. Other domains of moderate conductance coincide with thicker transported regolith sediments, and appear to indicate the presence of restricted, shallow, depositional basins (figure 5) that were confined to the west by Wagga Metamorphics, and to the east by Combaning Fm sediments. Similar 'basins' are identified at depth over the Belimebung Volcanics further to the east. Drainage from these restricted basins appears to have been through narrow gorges (figure 5) incised to relatively shallow depths in the Combaning Fm. Both saline–sulphate and 'fresh' groundwaters have been sampled at depth in different palaeo-channels (figure 5), and attest to the complexity of the sub-surface groundwater distribution systems in this area.

On the western palaeo-valley margin, numerous magnetically delineated east-north-east- to north-north-east-trending buried palaeo-channels cut through palaeo-ridges of Combaning Fm siliciclastic meta-sediments.[15] Some of these palaeo-channels are perched within alluvial sediments; some occur at the base of the transported sediment fill; and some channel-fill sediments were deposited in drainages incised into saprolith.[14, 15] Some palaeo-channels, targeted by drilling anomalies in the AEM data, have no magnetic signature (figures 5 & 6). The AEM signature of palaeo-channel deposits appears to vary depending on the character of the channel fill, its landscape position, and connection or disconnection with sub-surface hydrogeology. Palaeo-channels north of West Wyalong, delineated by maghemite channel-fill deposits, are broad, shallow channel-fill deposits with a low conductivity response. Higher conductivities are observed in narrow bands within saprolite bordering both banks. The channel-fill deposits appear to have a relatively high hydraulic conductance. Ephemeral run-off may flush any salts from these channel-fill deposits, with some salt accumulation in less-transmissive saprolite banks.

Upland landscapes

Several small upland basins occur within the Siluro-Devonian bedrock in the south-eastern margins of the Gilmore project area. These upland landscapes have concave lower slopes at elevations between 450 and 600 metres above sea level and carry a thin, silty topsoil over clay on deeply weathered colluvium (<8 m) and a variable saprolith. Surface stream flow salinity data, combined with soil and regolith mapping, demonstrate that in the project area salt is exported from these upland landscapes to the plains and major rivers. It comes mainly from sediment-filled upland basins with restricted outflows, and from areas that are overlain by thick clay soils that coat the footslopes and gently sloping upland basins.[10] Springs issuing from upper slopes with granite saprolite at the surface are clear and fresh. Saline stream flow from the hills mostly disappears into alluvial fans or into the alluvium of the Billabung Creek of the southern area, and flood plains of the Bland palaeo-valley in the north.

A different mix of technologies was used as part of a geological systems approach to mapping salinity within these upland landscapes.[26] This involves interpretation and modelling of high-resolution airborne gamma-ray spectrometry, primary and compound topographic indices derived from a high-resolution digital elevation model, and bedrock geological data. These data are used to map regolith materials and quantify geomorphic and hydrological processes in individual catchments. Field calibration of the derived regolith and soil data was required to validate interpretations. The study found relationships between regolith type, depth of weathering, geomorphic process and hydrologic gradients with salt mobility and storage in the landscape.[26] For remedial management purposes, catchments in these upland landscapes can be ranked according to their dryland salinity risk or potential risk.

Incised undulating hill landscapes

Houlaghans Creek

The modern morphology of Houlaghans Creek is a relatively linear, broad, gently concave valley floor bounded by steeper slopes developed on saprolith. The incised palaeo-valley appears to be localised by north-north-east-trending basement fault structures that transect Ordovician granites and Palaeozoic meta-volcanic and meta-sediments on a regional scale. Sediment infill of this palaeo-valley has alluvial fan and palaeo-channel deposits up to 60 metres thick. Some tributary palaeo-channel deposits, evident in the AEM data, occur at depth where there are no current surface creeks. The axial system lies beneath or offset by less than 200 metres from the present surface channel of Houlaghans Creek.

Upstream of Junee, Houlaghans Creek is a freshwater drainage. Like most of the creeks in the area, surface flows are ephemeral. However, drilling of the main and tributary palaeo-channels, targeted using interval conductivity slices derived from EMFlow processing of AEM datasets, confirmed the presence of saline–sulphate groundwaters within channel fill materials (figure 6). North-west-trending dykes, evident on aeromagnetic images, exhibit a smaller-scale control on conductivity structure. Some, but not all of these dykes appear to localise saline groundwater flow and localise ponding of groundwaters where they intersect the main palaeo-channel.

Billabung Creek

The present day Billabung Creek valley has a broad, flat floor characterised by alluvial deposition. Alluvial fan sediments are evident on the eastern valley footslopes. The valley is at the base of higher hills to the east, and appears to have etched out a course parallel to north-north-west-trending bedrock faults within more erodable Lower Palaeozoic lithologies. The Billabung Creek system, which is eight to 12 kilometres in width, is fed by both fresh and saline surface stream flow from tributaries. The palaeo-valley beneath is filled with at least 50 metres of sediment.

Saline surface stream flow waters are sourced from small, restricted, clay-filled upland basins in the higher landscapes in the hills to the east. Surface flows from the west are generally fresh, but recharge by saline groundwaters appears to occur through sub-surface drainage networks. Alluvial fan deposits in the footslopes of the hills on the eastern valley margins may be important aquifers for delivery of recharge groundwaters.

Undulating hill landscapes

The hills between the Houlaghans and Billabung Creeks are largely covered with thin sediments that increase in thickness downslope (<15 m). Higher crests generally have thin veneers of sediment over a saprolith that is between 10 and 80 metres thick. The sediment cover includes 'parna' deposits that are preserved on the leeside (eastern side) of drainage divides.[27] These re-worked aeolian deposits are linked to salt introduction and storage elsewhere in NSW.[28] In the Gilmore project area, multi-parameter borehole geophysical studies demonstrate significant differences between parna and underlying saprolite.[29] EC1:5 measurements and downhole induction logs suggest that conductivity is primarily controlled by textural variations rather than soluble salt and moisture contents of these materials. Ground EM (EM31) survey areas covered by parna exhibit marked lateral variations in average conductivity, attributable to differences in the thickness and proportion of parna in these materials.[29] Salts, once entrained in parna deposits, may have been mobilised into lower parts of the landscape and stored in accumulations of transported regolith in palaeo- and contemporary drainage.

Conductivity anomalies between the two main creeks were drilled, and four main types of anomaly were found. Conductive regolith materials drilled near Illabo contain up to 40 metres of fine-grained sediment-fill (figure 3b). Saline–sulphate groundwaters within this basin are

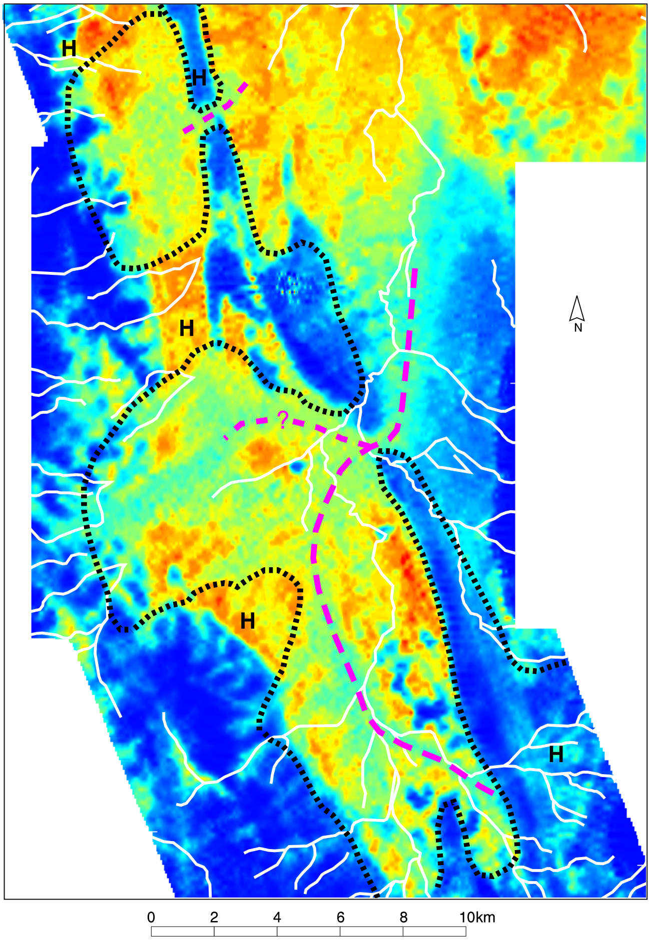


Figure 5. Interval conductivity slice (30–40 m) from the northern AEM survey area. Annotations point out a palaeo-drainage, the former course (pink dashed line) of which is picked out by moderate conductivities (green) at depth. Saline waters were intercepted at depth within this old river channel. Drainage basins containing thicker alluvial sediments are also evident from the spatial conductivity patterns. The blue areas are mostly resistive bedrock outcrop and buried ridges (area inside dotted line and 'H' symbol), except where there is flushing of regolith sediments by freshwater run-off from adjacent hills (eastern side of diagram).

largely perched within the transported sediment-infill, with some seepage of these waters into underlying saprolite. There is evidence of erosion of this landscape infill material.

Other conductive materials in the areas of relatively thin sediment veneer relate to the ponding and channelling of saline–sulphate groundwaters in thin regolith sediments, and ponding in saprolith adjacent to large quartz veins that form topographic highs and groundwater barriers. Significant saline groundwaters were found towards the saprolite–saprock boundary up-slope of quartz veins. Magnetic dykes also localise conductivity anomalies. Where these dykes transect drainage lines, they appear to confine the flow of saline groundwaters, in both hill and valley landscapes. Localised conductivity highs were also found to correlate with saline groundwaters at the base of deeply weathered outliers of siliciclastic Combaning Fm Palaeozoic meta-sediments. Conductivity lows relate to areas where there is no sediment veneer, with saprolite and/or saprock at the surface.

Conclusions

Initial results from applying a geological systems approach to problems of land and water salinisation in the Gilmore project area in NSW are proving to have particular value in delineating groundwater aquifer systems, their connectivity and spatial variability. One consequence is a better appreciation of the processes controlling dryland salinity in an area of complex regolith cover. New geophysical (particularly electrical) technologies have identified previously unknown salt stores and groundwater distribution systems (palaeo-channels) in the sub-surface. Studies are continuing with the intent of establishing the nature of the physical connections between salt stores and groundwater movement. Particular emphasis is being placed on establishing the hydraulic properties of materials in the area and recognising the aquifers.

Outputs of the geological systems approach provide an effective framework onto which more variable attributes such as land use, vegetation type/condition, climate, palaeo-climate and groundwater recharge rates can be added. When linked to a knowledge of the hydrological character (including water- and salt-balance relationships) of catchments, the methodology has value in defining appropriate and effective salinity management options.

An expert decision-support system is being developed to reduce the complexity of geological and geophysical data and deliver an effective toolbox for land managers and communities. The datasets generated by this approach will enable a more accurate prediction of salinity in these landscapes, and make targeted management of the hazards possible at catchment and paddock scales (figure 6).

The Gilmore study found that flexibility in the choice of specific technologies and research methodology is required for the landscapes studied. In footslope, valley floodplain or subdued landscapes with complex regolith cover, high-resolution AEM and airborne magnetic surveys are particularly useful in conjunction with new approaches to the mapping and modelling of bedrock and regolith materials and stratigraphy. In these landscapes, ground calibration of geophysical datasets by targeted drilling, along with near-surface sampling, is required to calibrate geophysical responses, bedrock geology architecture, regolith facies analysis, and groundwater distribution. These datasets provide a framework for understanding groundwater flow systems and hydrogeological models, and assist in the spatial delineation (in three dimensions) of salt stores and delivery systems in the sub-surface. Airborne gamma radiometric surveys are less important in these landscapes.

In contrast, airborne gamma-ray spectrometry is potentially a much more effective tool in upland landscapes, due to a generally thinner conductive regolith cover in upland areas. These data have been used to map regolith materials, and quantify geomorphic and hydrological processes in individual catchments. This requires integration of gamma-ray spectrometry with terrain attributes derived from a high-resolution digital elevation model. Field calibration of the derived regolith and soil data is required to validate interpretations. Examining the relationships between materials (soil and regolith) and landscape processes with salt loads in streams draining from these catchments provides new insights into salt movement and storage.

A similar geological systems approach is likely to provide similar benefits for hydrogeological and salinity modelling in other complex regolith terrains. The approach needs to be considered within the framework of a national catchment classification,[3] and tested in catchments in different climatic and geological regions. This multi-disciplinary approach can also be applied to the mapping of ore deposit mineral systems under cover. In particular, the integration of datasets has provided new insights into the modelling of hydromorphic dispersion of metals from concealed ore deposits.

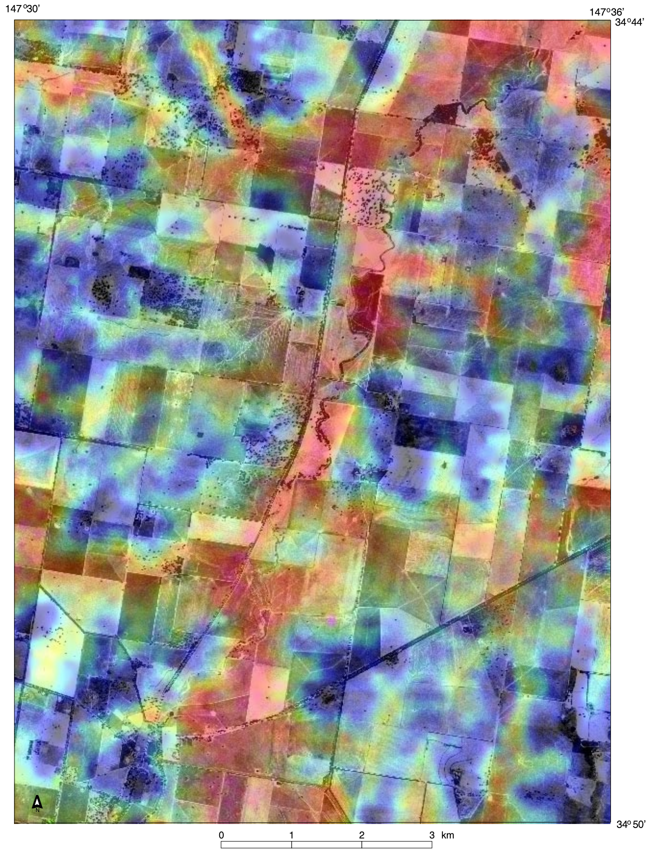


Figure 6. AEM (principal components) image draped over an orthophoto that shows creeks, roads and paddock boundaries. Red colours correlate with more conductive, saline groundwaters 20–40 metres underground in old river channels. In contrast, water is fresh in present-day creeks on the surface.

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Standard database entry of sequence stratigraphic units in AGSO

AT Brakel

Last issue comments were sought, particularly from industry, on AGSO's preliminary scheme for entering sequence stratigraphic units into its databases. Consequently, there were changes and the scheme AGSO follows is outlined below.

To address the urgent problem of entering sequence stratigraphic units into AGSO's databases, the following guidelines for standardised database entry of these units have been drawn up. These focus on those aspects that are the particular immediate concern of our databases, namely the definition and naming of these units. It does not deal with other aspects of sequence units such as Vail-Exxon v. Galloway types of units, or the basis on which units are ranked in hierarchies. The aim is to fulfil immediate needs, and to be able to enter those units already published or about to be published.

In the longer term, moves are underway to reach consensus on a national scheme for standards in the definition and naming of units.[1] If this produces a different scheme, the existing units in the databases can be linked to the new conventions.

Principles

The scheme is based on the following principles:

1. All sequence stratigraphic units are to be formally defined in a way similar to lithostratigraphic units, but with variations that are peculiar to sequences—for example: representative or type localities for sequence boundaries that can include outcrops, seismic line locations, and/or well intervals; biostratigraphic constraints; and methodology.
2. The basal sequence boundary is to be used as the defining parameter for sequence units (not type sections, lithology, etc.).
3. Names for sequence units must be unique and distinguishable from lithostratigraphic units.
4. The term sequence, supersequence, megasequence or subsequence must always be attached to the name of a sequence unit.
5. Names of sequences, supersequences and megasequences must be proper names, not alphanumeric codes or part alphanumeric.
6. Preferably invent appropriate names that do not contravene the other rules.
7. Using an abbreviation of the proper part of associated lithostratigraphic names is permitted. For example, a sequence that contains mainly Riversleigh Formation sediments can be called the River Sequence. However, a Mt Langsborough or Langsborough Creek Formation cannot be abbreviated to Langsborough Sequence. Langs Sequence or Borough Sequence would be acceptable in this case.
8. The use of geographic names, including well names, is discouraged because geographic names are required for lithostratigraphic units. Already there is an acute shortage of geographic names in some areas. If sequence units compete with lithostratigraphic units for the available geographic names, the situation will be exacerbated.
9. Avoid names that are age specific (e.g. *M. australis* Sequence, Jurassic 2 Sequence, Santonian Sequence).
10. Use of a digital hierarchy for subsequences and subordinate units is permitted (e.g. Dingo 1.2.3).
11. Use the sequence name or an abbreviation to identify the basal sequence boundary.
12. The parent unit of a sequence unit must be stored in the database, if a parent unit exists. (A parent unit is the next one up in the rank hierarchy to which the unit belongs; e.g. a supersequence is the parent unit of a sequence.)
13. A standard abbreviation, unique to the basin or province, must be recorded in the database for use when a name is too long for such purposes as maps and sections.

Hypothetical example

Name and rank:
White Gull Sequence.

Derivation (optional):
A bird common in the region.

Synonymy (if any): Upper part of Vindaloo Sequence of Jones (1985).

Distribution: Wildcat Basin, except for the north-east side where it has been eroded away.

Lower sequence boundary:

Type locality: Depth of 1532 metres in the Whitmore 1 well (lat 11° 30'S, long 140° 30'E), which corresponds to SP551, TWT 498 milliseconds in seismic section AZCO 1996–3.

Identifying features: 8° dip discordance on dipmeter log, strong spike on gamma-ray log, stratal termination surface in seismic section AZCO 1996–3.

Adjacent lithologies at the type locality: Limestone below the boundary, mudstone above.

Lithostratigraphic units at the type locality: Johns Limestone below, Bintang Formation above.

Underlying sequence unit at the type locality: Sea Hawk Sequence.

Age of rock below: Valanginian, E. torynum Zone.

Age of rock above: Valanginian, S. areolata Zone.

Regional aspects: Angular discordance decreases towards the centre of the basin.

Upper sequence boundaries: In the Whitmore 1 well, the unit extends upwards to the sequence boundary below the Pelican Sequence. This occurs at a depth of 1102 metres (still in the *S. areolata* Zone), which corresponds to SP551, TWT 365 milliseconds in seismic section AZCO 1996–3. However 25 kilometres to the east, the unit is overlapped by the Shearwater Sequence.

Sequence regional aspects: Subaerially deposited, sandstone-dominated succession along the eastern basin margin interfingers with paralic mudstone and sandstone westwards, and passes further west into deep offshore mudstone-dominated rocks. Sequence thickens towards the palaeo-shelf margin, but thins again in the deeper water facies.

Constituent units: Composed of three fourth-order units, the White Gull 1 Subsequence, White Gull 2 Subsequence, and White Gull 3 Subsequence.

Parent unit: Frigate Bird Supersequence.

➥ *Continued page 36*

Epithermal deposits of the Central Pilbara tectonic zone

Description and exploration significance

DL Huston, B Keillor, J Standing, R Blewett & TP Mernagh

Epthermal textures have recently been recognised in north-north-west-trending quartz veins hosted by Mesoarchaean rocks of the Central Pilbara tectonic zone. These veins occur near the inferred unconformity with the overlying Neoarchaean Fortescue Group. Geological relationships and Pb isotope model ages of possibly correlative mineral-isation suggest that this epithermal mineralising event was associated with the opening and early development of the Hamersley Basin at around 2750 million years. These results have significance not only to the potential for the Pilbara Craton and overlying Fortescue Group, but also to other Archaean terranes; preservation, not age, is the determining factor for Archaean epithermal mineralisation.

In 1997, Resolute Ltd geologists recognised the presence of epithermal veins in the Indee district within the 3.1–2.95 million year old Central Pilbara tectonic zone of the Pilbara Craton.[1] These veins are located just below the inferred position of the unconformity between the Central Pilbara tectonic zone and the overlying 2.78 million-year-old Fortescue group. Elsewhere in the Pilbara, epithermal veins may be present in the basal units of the Fortescue Group.[2]

The purposes of this communication are to:

1. describe the geology, vein textures and mineralogy of selected deposits;
2. document the trace element geochemistry of the veins;
3. provide preliminary constraints on the composition and temperature of the mineralising fluids; and
4. discuss the significance of these deposits to the Pilbara Craton and other Precambrian provinces.

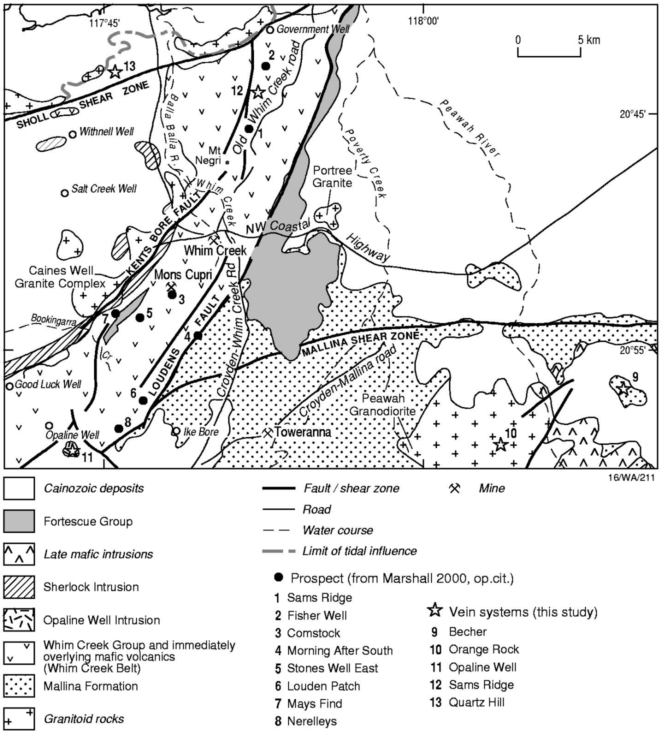

Figure 1. Location of epithermal deposits in the Central Pilbara tectonic zone (modified on Smithies, WA sheets 2456 & 2556—see references 5 & 13)

Figure 1 shows the regional geology along with locations of epithermal deposits visited by the authors and described by Marshall.[2]

Becher prospect (621500 mE, 7683700 mN)*

Epithermal textures are best developed at the Becher deposit, a zone of anastomosing quartz veins in sandstone of the Mallina Formation (figure 2). Although the vein system has an overall strike of 330°, it is comprised of two vein sets, with the more dominant set striking 310° and the subordinate set

* Zone 50 AMG locations using AGD66 datum

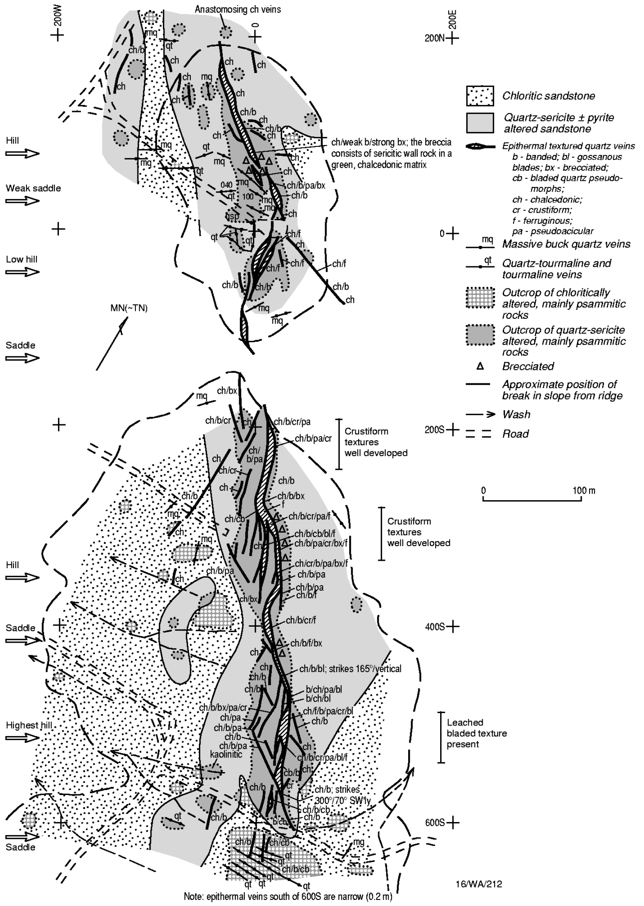

Figure 2. Surface geology of the Becher deposit

striking 350°. Wall rocks adjacent to the veins have been altered to a quartz–sericite–(ex)pyrite assemblage that can be traced at least 50 metres from the vein system, where it grades into a 'background', chloritic assemblage (figure 2). Kaolinitic zones locally occur within the quartz–sericite–(ex)pyrite assemblage.

The Becher vein system is dominated by massive chalcedony. Weakly banded chalcedony, also relatively abundant, is characterised by diffuse or vague colour banding, but no distinct crustiform banding is developed. Other habits of quartz include pseudoacicular quartz, quartz pseudomorphs replacing bladed minerals, and colloform–crustiform quartz. Brecciation of these other textures is also present locally.

Radial, needle-like quartz (pseudoacicular[3] or 'mold' textures[4]) is present both at the hand specimen (figure 3a) and the microscopic scales (figure 3b). Both the pseudoacicular textures and quartz pseudomorphs replacing a coarsely bladed mineral (possibly carbonate) are restricted in distribution (figure 2). Where present, bladed textures tend to occur along the hanging wall (e.g. western) side of the vein. Within these zones, the blades are typically five to 20 centimetres in length and one to five millimetres in thickness.

Crustiform quartz (figure 3c), which has a very restricted distribution (figure 2), is characterised by two- to 20-millimetre banding with a colloform habit. Locally this banding has been brecciated, resulting in complex overprinting relationships both at the hand specimen and microscopic scales.

Bladed gossan (figure 3d) is characterised by iron oxides that have replaced massive zones of an unoriented bladed mineral (possibly Fe-rich carbonate). Bladed gossan, present only at the highest hill in the Becher vein system, appears to be paragenetically late as it fills the centre of individual veins and locally cross-cuts other quartz types. Zones of coarsely bladed quartz appear to occur lateral to this zone of bladed gossan.

Brecciated textures occur to varying degrees throughout the Becher vein system. These textures vary from wall rock breccias with chalcedonic infill to breccias in which crustiform clasts are infilled by later chalcedonic quartz (figure 3e), and to breccias with complex clast and matrix types.

In thin section other textures are present, including cockade and plumose textures (figure 3f). The plumose quartz is associated with sericite that occurs both as seams and in a stubby mass, possibly pseudomorphous after a stubby mineral such as feldspar (possibly adularia).

The presence of abundant chacedonic quartz and bladed quartz pseudomorphs—combined with the presence of limited crustiform–colloform quartz and the absence of crystalline quartz—suggests that the exposed Becher vein system is in the chalcedonic superzone or the upper part of the crustiform–colloform superzone of Morrison et al.[4] These observations suggest that a boiling zone, if present, is at depth. The presence of multiple episodes of brecciation in some veins is also encouraging for mineral potential for the Becher vein system.

Orange Rock deposit (610250 mE, 7678900 mN)

The Orange Rock deposit is located 12 kilometres west-south-west of the Becher deposit within high-Mg granodiorite of the (~2950 Ma) Peawah Granodiorite.[5] The Orange Rock vein system has an overall trend of 345°. Like the Becher vein system, it is segmented with two short segments in the northern and central parts of the vein system having trends in the range from 350° to 015° (figure 4).

Most of the Orange Rock vein system is characterised by well-developed vein breccias. These breccias typically contain one- to 20-centimetre clasts of silicified granodiorite and chalcedonic quartz in a chalcedonic matrix. Other textural types are much more restricted in their distribution. The most abundant high-level textures in the Orange Rock vein system occur near the northern extremity. These vein textures occur in an approximately 50-metre-long, locally highly gossanous interval within a 345° trending segment just to the north of a jog from a 015° trending segment (figure 4).

Textures present in this gossanous interval include breccia, coarse bladed quartz psuedomorphs, and bladed gossan. The breccias contain silicified granodiorite clasts set within a gossanous matrix. The most unusual texture present in this zone is Fe-oxide pseudomorphs after a bladed mineral that has been folded (figure 3g). However, the lack of coherence within the folds and the lack of any other structural overprint suggest that the folds are not tectonic. Where the vein jogs to the south of this interval, the vein textures revert to the chalcedonic breccias characteristic of the vein system as a whole. This relationship implies that changes in vein trend have an important control on the development of epithermal textures in the Orange Rock vein system.

Bladed pseudomorphs of quartz are also present one kilometre to the south of the gossanous interval where the vein system bifurcates into a 350° trending main vein and a 330° trending branch. Chalcedonic quartz with disseminated pyrite is also locally present along the branch.

Veins in the Opaline Well Granite (575400 mE, 7679200 mN)

The 2765 million-year-old[6] Opaline Well Granite is cut by a series of 0.2–3 metre quartz veins that trend north-west to north-north-west (320–345°) and have strike lengths up to several hundred metres. Although evidence of sulphide minerals (e.g. gossanous patches) is lacking, these veins typically have well-developed epithermal textures dominated by bladed pseudomorphs of quartz (figure 3h), with lesser, weakly banded chalcedonic quartz. An unusual characteristic of these bladed textures is that locally the blades are composed of fluorite and not quartz. In outcrop the fluorite weathers recessively, leaving residual silica ridges (figure 5a). In thin section, the selvedges to some of the weakly banded chalcedonic veins are comprised of potassium feldspar (figure 5b). Marshall also reports the presence of epithermal textures in these veins, and notes the presence of colloform and crustiform textures.[2]

Sams Ridge deposit (590300 mE, 7707100 mN)

One of the more interesting prospects described by Marshall, the Sams Ridge deposit, is hosted by high-Mg basalt of the Louden Volcanics.[2] A brief visit was made to the Sams Ridge prospect at the location indicated by Smithies.[3] This location is 2.3 kilometres north of the Sams Ridge prospect as identified by Marshall.[2]

The site visited is characterised mainly by grey-white chalcedonic quartz with minor gossanous zones and minor brecciated zones. Locally malachite and disseminated pyrite are present. The veins are generally narrow (<1 m) and strike due north. Marshall's Sams Ridge prospect (590100 mN, 7704700 mN) consists of a north-trending 150-metre-wide zone of quartz veining and alteration along the same quartz vein system as the deposit visited by the authors. Marshall reports chalcedonic quartz, colloform banding, crack-seal textures and bladed pseudomorphs of quartz.[2] Gossanous zones, some of which contain oxide Cu minerals, occur along selvedges of some of the chalcedonic veins.

Quartz Hill vein (578700 mE, 7708000mN)

The authors also visited a north-trending chalcedonic vein near Quartz Hill Well just to the north of the Sholl Shear Zone. This occurrence is characterised by a brecciated quartz vein up to 30 metres in width and three kilometres long. Other than the extensive brecciation, no epithermal textures were observed.

From page 33 **Standard database entry of sequence stratigraphic units in AGSO**

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Re-naming sequences

The naming of sequences in AGSO and elsewhere has been based on a number of methods in the past—most of which are inappropriate to the standards set out here. Names typically are not unique and are commonly based on alphanumeric codes or age-specific names. Use of the geographic parts of formally defined lithostratigraphic names is also widespread, even though the units often are not identical. Ideally, only where there is an exact vertical correspondence between a sequence stratigraphic unit and a lithostratigraphic unit should the geographic name also be adopted.

An example of how current usage *might* be revised is set out below, using the AGSO guidelines. Alongside the published version is a possible revised scheme, based on carriage names.

Published scheme	Possible revised scheme
BB12	Chariot Sequence
BB12C	Chariot 3 Subsequence
BB12B	Chariot 2 Subsequence
BB12A	Chariot 1 Subsequence
BB11	Coach Sequence
BB10	Phaeton Sequence
BB9	Brougham Sequence
BB8	Wagon Sequence
BB7	Rickshaw Sequence

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Acknowledgments: This scheme was developed from the preliminary scheme[2] by a forum of AGSO staff, who are thanked for their input.

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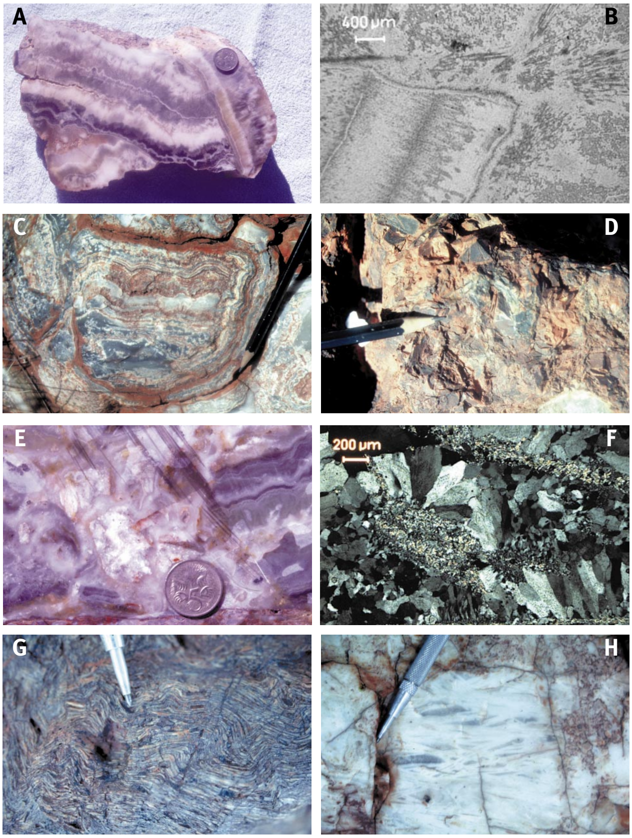

Figure 3. Photographs and photomicrographs showing textures in epithermal vein systems from the Central Pilbara tectonic zone.

A. Complex vein characterised by multiple generations of banded, pseudoacicular quartz
B. Banded clast with pseudoacicular quartz overgrowths (plane light)
C. Crustiform clast overgrown by crustiform quartz
D. Bladed gossan
E. Vein breccia, with crustiform quartz clasts in a chalcedonic quartz matrix
F. Plumose to subhedral quartz grains growing perpendicular to elongate seams of sericite. (Note the presence of a stubby a shape of sericite mat in the centre of the frame and the fluid inclusions that decorate growth zones just above this sericite mat; crossed polars.)
G. Fe oxides, pseudomorphous after a bladed and folded mineral
H. Bladed pseudomorphs of quartz

Geochemical studies

Twenty-four grab samples were analyzed at Analabs' Welshpool laboratory for a suite of elements (Au, Ag, Hg, As, Bi, Sb, Te, Tl, W, U, Cu, Pb and Zn) considered characteristic of the epithermal environment using a combination of AAS (atomic absorption spectrometry) and ICP-MS (induction coupled plasma-mass spectrometry) techniques. The samples were selected to be representative of the vein textures present at each deposit. Table 1 summarises the results.

The most anomalous samples were from the Becher prospect, where Au assays ranged upwards to 0.22 ppm. Arsenic, Bi, Sb, Te, W, Cu and Zn were also anomalous. Arsenic, Sb and Te were most elevated, with values to 6610 ppm, 178 ppm and 3.3 ppm, respectively. Analyses from the Orange Rock prospect, although lower than Becher, are still anomalous. The maximum assay for Au was 0.06 ppm, but As (to 208 ppm), Sb (to 14 ppm) and Te (to 0.6 ppm) were significantly elevated relative to average crustal abundances.

With the exception of weak Sb and Te anomalism, the results from the Opaline Well and Quartz Hill Well veins are uniformly low. Marshall, however, reported anomalous Au, Sb and Cu in streams draining the contact between the Opaline Well Granite and surrounding turbidites.[2] In contrast, results from Sams Ridge are enriched in Ag, Hg, As, Bi, Sb, Te and Cu, but Au assays are low (Table 1). Marshall reported highly anomalous assays from his Sams Ridge prospect, with values up to 0.44 ppm Au, 15 ppm Ag, 90 ppm As, 8100 ppm Sb and 9800 ppm Cu.[2] These latter results are the most significant reported for an epithermal prospect from the Central Pilbara tectonic zone.

Fluid inclusion studies

Preliminary fluid inclusion studies were undertaken on two samples from the Becher deposit. In the first sample, equant fluid inclusions that decorate growth banding in plumose quartz yielded eutectic temperatures of -50 to -46°C, indicating the presence of other salts (e.g. KCl, $CaCl_2$, $MgCl_2$, $FeCl_2$) besides NaCl. Salinities ranged between 9.2 and 17.8 wt % NaCl equivalent with a mode around 15.5 wt % NaCl equivalent. The inclusions homogenised over the range from 92 to 161°C with a mode at 146°C.

In the second sample, characterised by rhythmically banded chalcedonic quartz, primary inclusions decorate growth bands in euhedral quartz crystals that occur on the boundary of the chalcedonic quartz. These primary fluid inclusions are rounded to irregular in shape. Eutectic melting could not always be observed. When it was it ranged from -16.4 to -12.5°C. Salinities of inclusion fluids were all less than 3 wt % NaCl equivalent. Homogenisation occurred over a range of temperatures from 130 to 299°C with the majority of inclusions homogenising below 210°C.

No definitive evidence for boiling was found in either sample, but the vapour phase is commonly not trapped in epithermal systems.[7] The large scatter in homogenisation temperatures in the second sample may be the result of necking down or other post-entrapment modifications. Temperatures near the minimum homogenisation temperature (i.e.

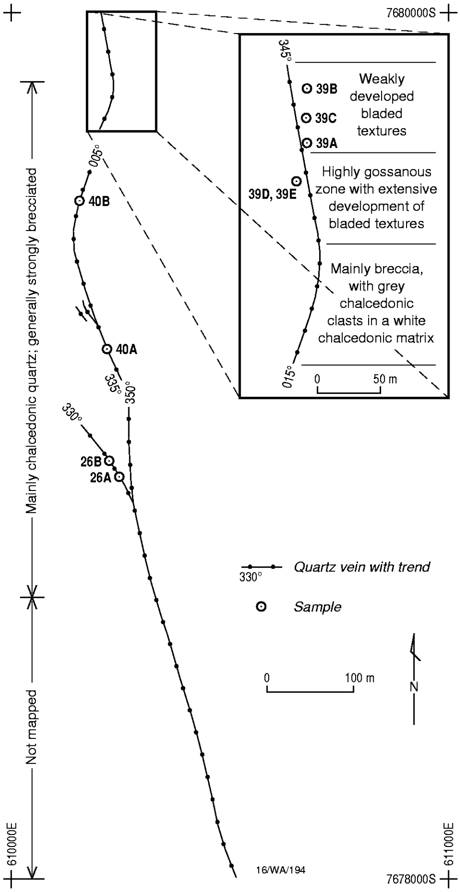

Figure 4. Geology of the Orange Rock vein system

130–150°C) most likely represent trapping temperatures as post-entrapment modifications commonly increase observed homogenisation temperatures. The salinity of the fluid in the first sample is clearly higher than that reported for most epithermal systems. Although early high-salinity fluids have been previously reported from the Creede and Summitville deposits (Colarado, USA),[8] the significance of these high-salinity fluids at Becher is not understood. The lower fluid salinities (reported from sample 98044088A) are more akin to those from typical epithermal systems.[7]

Discussion

Vein textures, trace element associations and the preliminary fluid inclusion data are all consistent with an epithermal origin for the Becher, Orange Rock, Opaline Well and Sams Ridge vein systems. The combination of chalcedonic, pseudoacicular, crustiform and bladed pseudomorphs is diagnostic of the epithermal environment,[9] and the trace element assemblage As–Sb–Te±Bi±W is also consistent with such an environment.[10] Moreover, the presence of potassium feldspar at the Opaline Well veins suggests a low sulphidation character to some of the deposits. The fluid inclusion characteristics of the Becher system are also consistent with an epithermal origin.[7]

The age of the mineral deposits has not been established directly. However, inferences can be made based on geological relationships, particularly from the Opaline Well occurrences. The veins at Opaline Well cut 2765 million-year granite set a maximum age of the mineralisation. Fluorite in these veins is found elsewhere in the Pilbara, in veins that have model Pb isotope ages of 2700–2750 million years.[11] Moreover, all Pilbara veins occupy north-west- to north-trending structures associated with the opening of the Fortescue basin (at ~2780 Ma).[12] These geological data are most consistent with, although not definitive of, an age of around 2700–2750 million years for the epithermal veins. If true, this suggests that Precambrian terranes, generally not considered prospective for these deposits, may contain epithermal deposits if a high crustal level has been preserved.

Finally, figure 1 shows that many of the veins, although north-north-west-trending, are located near north-east-trending structures such as the Kents Bore and Loudens Faults.

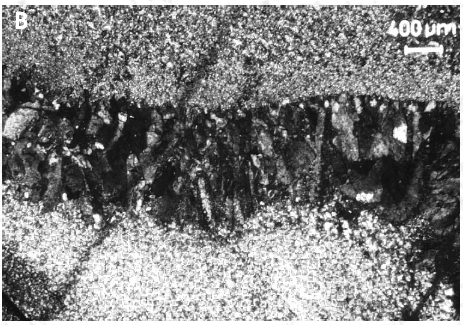

Figure 5. Photographs and photomicrographs showing textures in epithermal vein systems from the Central Pilbara tectonic zone. **A.** Fluorite pseudomorphs after a bladed mineral (recessively weathered) **B.** Low-temperature potassium feldspar forming selvedges to chalcedonic quartz vein

Conclusion

Quartz veins containing textures typical of epithermal vein deposits are present in a number of locations in the Central Pilbara tectonic zone. These veins cut a variety of host rocks including turbiditic sandstone, granite, high-Mg basalt, and high-Mg granodiorite. Elsewhere in the Pilbara, these veins cut basal units of the overlying Fortescue Group.

These veins appear to be related geologically to the opening of the Fortescue Basin, with a probable age of approximately 2700–2750 million years.

The presence of epithermal veins in the Pilbara indicates the possibility of epithermal deposits in other Precambrian terranes—provided that high-level environments have been preserved and not eroded.

Table 1. Range in trace element analyses from epithermal deposits of the Central Pilbara tectonic zone (all in ppm).

Deposit	No	Au	Ag	Hg	As	Bi	Sb	Te	Tl	W	U	Cu	Pb	Zn
Becher	12	<0.01–0.22	<0.1–1.8	<0.005–0.36	15–6610	0.2–3.3	2.6–178	0.2–3.3	<0.5–0.6	1.6–34.9	<0.05–11.9	4–740	<5–170	<4–409
Orange Rock	7	<0.01–0.06	<0.1–0.3	0.012–0.024	6–208	0.2–1.1	5.7–13.5	0.5–0.6	<0.5	1.1–2.4	0.08–2.9	5–13	<5–22	5–26
Opaline Well	2	<0.01	<0.1–0.2	0.012	8–27	0.4–0.5	6.2–8.4	0.5 0.8	<0.5	1.3–1.6	<0.05–0.07	4–5	16–23	4–5
Quartz Hill	1	<0.01	<0.1	0.035	13	0.5	14.3	0.7	<0.5	0.8	0.05	6	5	5
Sams Ridge	2	<0.01–0.02	2.3–4.5	1.64	73–868	0.3–7.9	69–131	0.6–0.8	<0.5	1.0–1.3	0.16–4.3	1135 2820	<5–187	39–40

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