

Geological and structural controls on gold mineralization in the Tanami District, Northern Territory

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Abstract Gold mineralization in the Tanami district is hosted within moderately northwest dipping turbiditic sedimentary and basaltic volcanic rocks of the Paleoproterozoic Mt. Charles Formation. The gold occurs within a complex sinistral wrench-fault array and associated veins and alteration haloes. The main mineralized faults have a northerly trend and dip steeply east. Subsidiary structures trend at 030° and 070° and dip towards the southeast. Paleostress calculations based on fault striation populations and geometry (strike and dip) of faults indicate that at the time of the mineralizing event, σ_1 was sub-horizontal and SE–NW directed with σ_2 subvertical. Structural studies indicate that the mineralization occurred after the regional folding event and synchronous with the emplacement of felsic dykes into the mine sequence. Gold veins in the Tanami district are interpreted to be part of an outer thermal aureole gold system that formed during the emplacement of granitoids in the nearby ~1,815 to ~1,799 Ma Frankenia and/or Coomarie domes. Economic gold mineralization occurred late in the paragenetic history of the district. Gold is hosted by quartz-carbonate veins within shear zones, and also in the surrounding sericite-quartz-pyrite $\pm$ carbonate-altered wallrocks. Gold-mineralized veins precipitated at depths of 3 to 6 km from high temperature (~300°C), low

salinity (~5 wt% NaCl equivalent) fluids with low CO₂ contents. Barren quartz, dolomite and calcite veins that occur in pre- and post-mineralization thrust faults formed from high salinity (~20 wt% NaCl equivalent), low temperature (~120–150°C) basinal brines. Pyrite in the gold mineralized veins and alteration halos has lower $\delta^{34}\text{S}$ values (6.8 to 12.5‰) than local diagenetic pyrite (17.8 to 19.2‰) or pyrite in pre-mineralization thrust faults (31.7 to 37.1‰). The mineralizing fluids are inferred to have contained a well-homogenized mixture of magmatic and sedimentary-derived sulfur.

Keywords Proterozoic · Tanami · Granite · Gold mineralization · Thermal aureole

Introduction

The Tanami goldfield, located 650 km northwest of Alice Springs, is one of several significant gold districts in the Paleoproterozoic Tanami region of the Northern Territory. This area has become one of the major gold provinces of Northern Australia. Historically, investigations into the geology of the region have been neglected because of deep weathering, lack of exposure, and remoteness. The Tanami region grades, without clear boundary, into the Arunta inlier to the southeast and is separated from the Halls Creek inlier and the King Leopold Mobile zone to the west by the overlying sedimentary units of the Middle Proterozoic Birrindudu Basin (Plumb 1990; Fig. 1).

Gold was originally discovered at Tanami in 1904, and the area has been worked intermittently since then. Large-scale, open-cut mining commenced in 1987 within mining lease 153 (MLS153) and was undertaken first by the Tanami Joint Venture and then by Zapopan N.L. until

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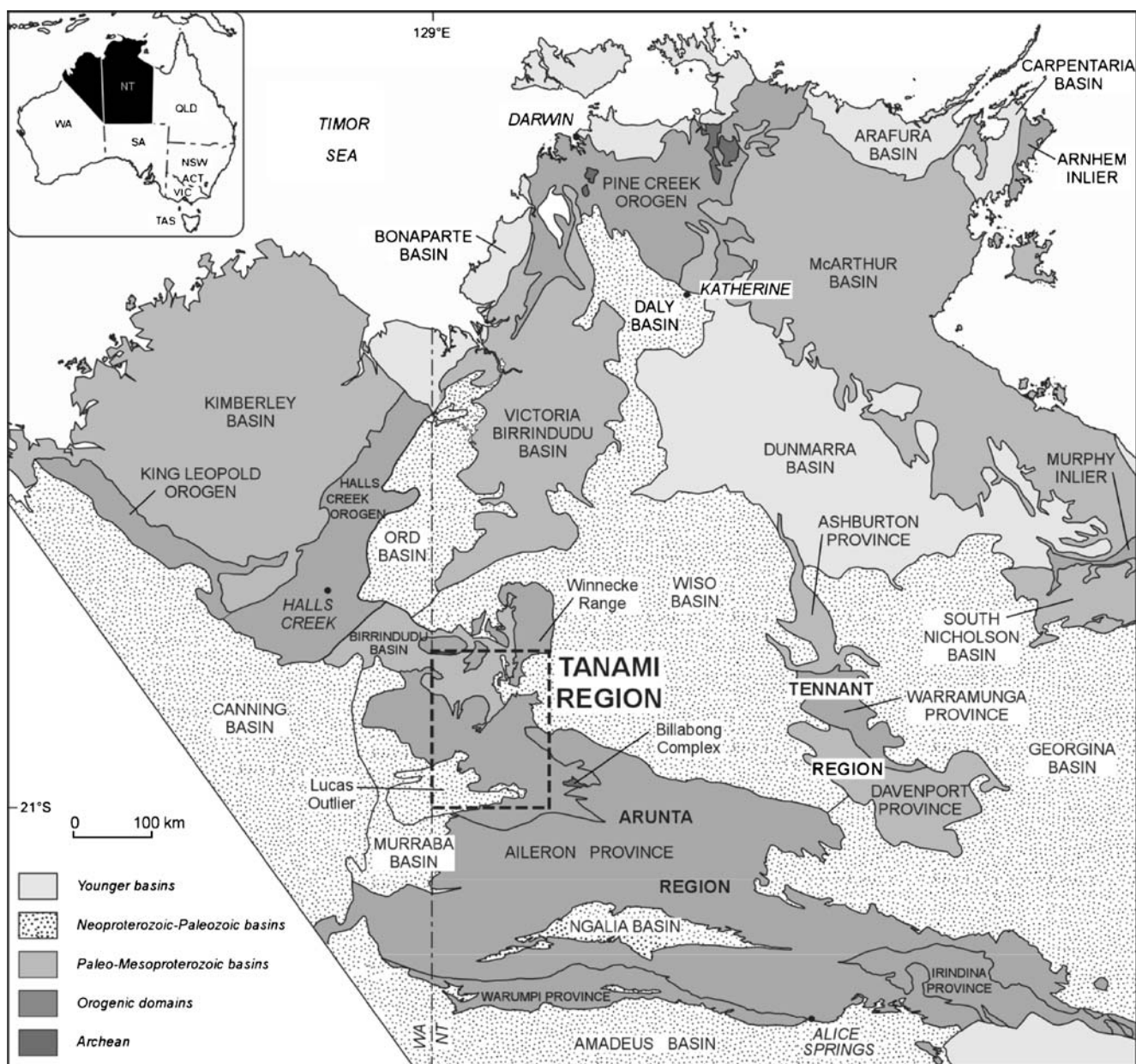


Fig. 1 Distribution of Paleoproterozoic inliers throughout Northern Australia, highlighting the location of the Tanami region (after Crispe et al. 2006)

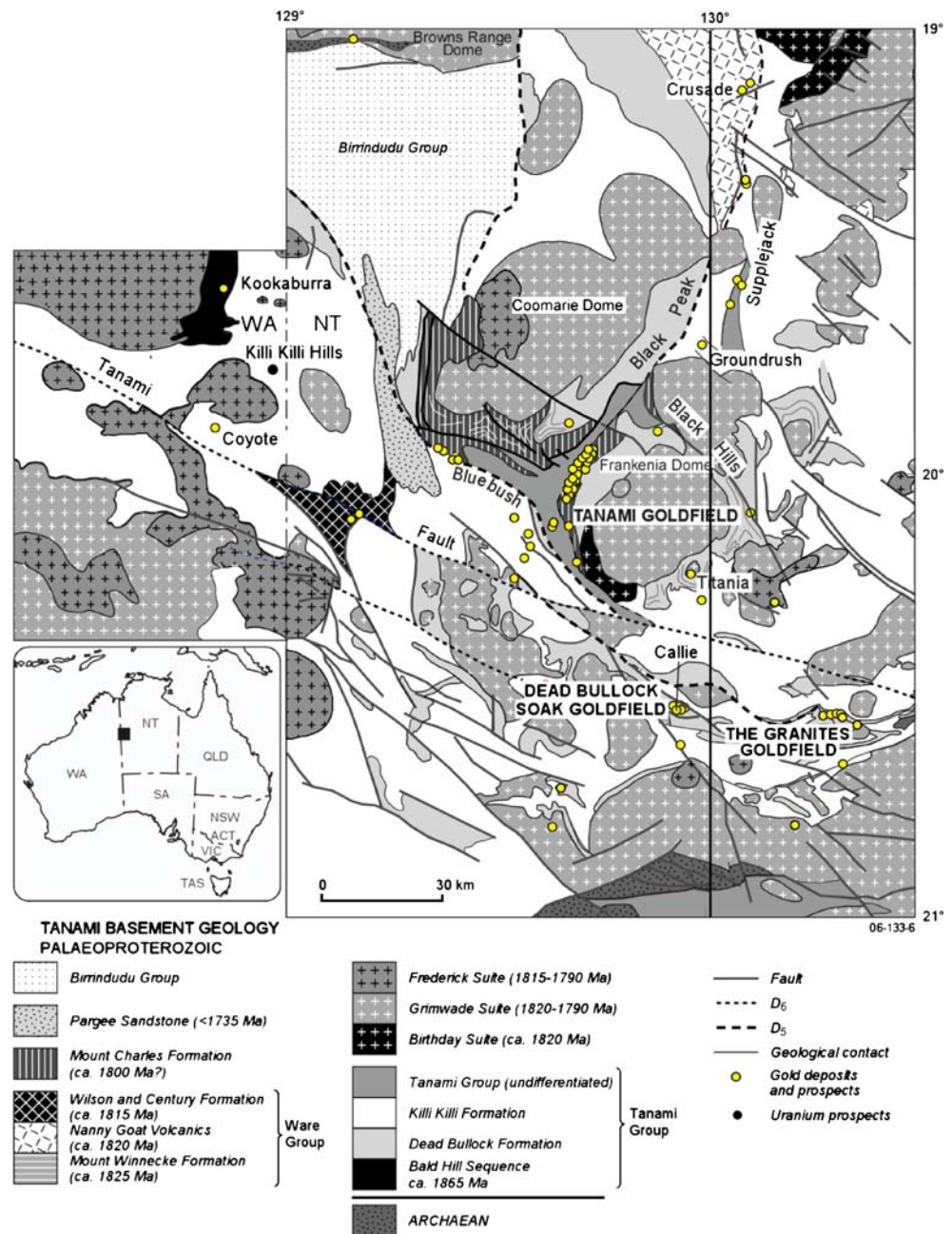
1994 (Tunks and Marsh 1998). More recent exploration and mining was undertaken by Otter Gold Mines and Acacia Resources/AngloGold as the Central Desert Joint Venture. Since 1987, approximately one million ounces was produced and four million ounces of Au discovered in a series of small tonnage, isolated ore shoots developed along the “Tanami Mine Corridor” (Fig. 2) at an average grade of 2.1 g/t Au. All known gold mineralization in the district is hosted by the Mt. Charles Beds, a subdivision of the Palaeoproterozoic Tanami Complex (Blake et al. 1979). In 2001, Newmont Australia began mining at Groundrush, 43 km NE of MLS153. They currently use the MLS153 mill to process ore from Groundrush (Huston et al. 2006).

The excellent exposure provided both within the open-pit mines and in exploration drill holes throughout MLS153 provided an ideal opportunity to study the structure and geological setting of gold mineralization in a way not possible previously. This paper describes work completed as part of a Ph.D. project and documents the mine-scale geology, alteration, and deformation associated with the gold mineralization.

Geological setting

Paleoproterozoic sedimentary and volcanic rocks of the Tanami region are underlain by Neoarchean rocks of the

Fig. 2 Distribution of rock units and major mineral deposits in the Tanami region (after Huston et al. 2006). Our study is focused on the northern part of the Tanami Goldfield, which is marked by the ENE-trending cluster of gold deposits that define the “Tanami Mine Corridor” between the Frankenia and Coomarie granite domes



Billabong Complex, which are exposed to the east of the Granites Gold Mine (Huston et al. 2006; Fig. 2). The Paleoproterozoic Tanami Complex, as defined by Blake et al. (1979), was originally divided into five separate units on the basis of differing proportions of lithologies and geographical location. The Killi Killi, Mt. Charles, Nanny Goat Creek, Nongra, and Helena Creek beds were all thought to be possible lateral correlatives. Recent work by Hendrickx et al. (2000) has suggested significant changes to the stratigraphy of the Tanami region and reclassified the host rocks to the mineralization as part of the newly defined Mt. Charles Formation. However, as we discuss

only the local geology of the Tanami goldfield, a simplified stratigraphy will be used in this paper.

Throughout the Tanami region, there has been a history of episodic granitic intrusion from the Archean to the Palaeoproterozoic (Page et al. 1995) that has produced large domal structures (e.g., Browns Range). Based on geochemical criteria, three suites of granites were emplaced in the region between 1,825 and 1,790 Ma, namely, the Birthday, Frederick, and Grimwade suites (Green, in Crispe et al. 2006). The Tanami gold deposits occur between two granitic domes, which are cored by the Coomarie and Frankenia Granites (Fig. 3). Parts of the Frankenia granite

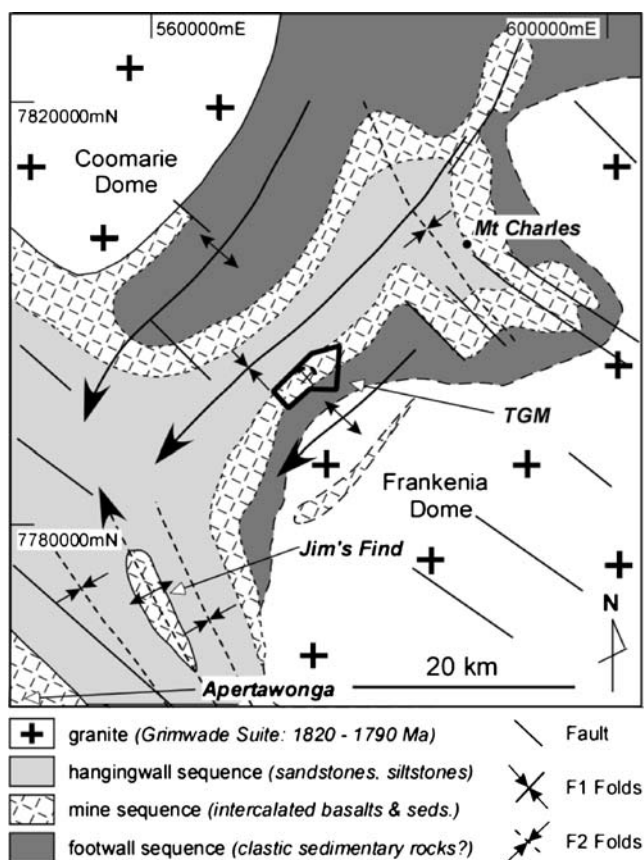


Fig. 3 Palaeoproterozoic basement geology of the Tanami district and surrounding environments, interpreted from regional mapping and aeromagnetic data

complex have been dated at $1,815 \pm 4$ Ma and $1,805 \pm 6$ Ma (Smith 2001; Cross et al. 2003; Wygralak et al. 2004). Both granites are part of the Grimwade Suite (Fig. 2), a series of weakly to non-magnetic peraluminous calcic monzogranites and syenogranites (Green, in Crispe et al. 2006).

The Paleoproterozoic basement rocks that make up the Tanami region can be separated into two distinct structural domains. The southern part preserves mainly NW-trending structures possibly related to the Arunta inlier (Plumb 1990). This trend can be traced under the sedimentary rocks of the Birrindudu basin and into the King Leopold Mobile Zone forming part of a major continental lineament (O'Driscoll 1990). At least four major orogenic deformational events have been recognized in the Granites mine area (Adams et al. 1994). Granite intrusion has occurred throughout the structural history of the Granites area, but is volumetrically dominated by the unfoliated Granites Granite suite (Ding and Giles 1993).

In the northern domain, the intensity of deformation is lower, with only two ductile events recorded in the Tanami Mine area. The first-order F_1 fold structures plunge shallowly southwest and have wavelengths of approximately 10 km. Shorter wavelength (<100 m) second-order folds

are abundant within the sedimentary packages away from the more competent mafic rocks exposed in drill core and the open pits. Although there is little evidence for a ductile D_2 event in the open pits, to both the north and south, F_1 folds are refolded about NW-trending upright open F_2 folds that are accompanied by a locally intense, millimeter spaced, crenulation cleavage. Fold interference between the F_1 and F_2 folds has created basin and dome style outcrop patterns (Fig. 3).

Mine scale geology

The mine sequence comprises up to four massive to pillowed basaltic packages, each approximately 80–150 m thick. Each basaltic package is made up of several, commonly pillow rich, flow units that vary from 2 to 30 m thick. The flow units are intercalated with thin sedimentary horizons that vary from less than 1 to 15 m thick. The major basalt dominated packages are separated by sediment-dominated packages, where basalt units are either rare or absent, that are between 80 and 200 m thick. The total thickness of the mine sequence exposed in the district is approximately 1,000 m.

In the open pits, the mine sequence dips moderately (35 – 55°) towards the NW and faces consistently upwards (Figs. 4 and 5a). There is no penetrative cleavage, and mesoscopic folds only occur adjacent to the major faults. The mine sequence is deeply weathered and the transition to fresh rock is approximately 60 m below the present surface. Beneath the weathering profile and away from the intense alteration halos associated with the Au mineralization, textural preservation of all rock types is good.

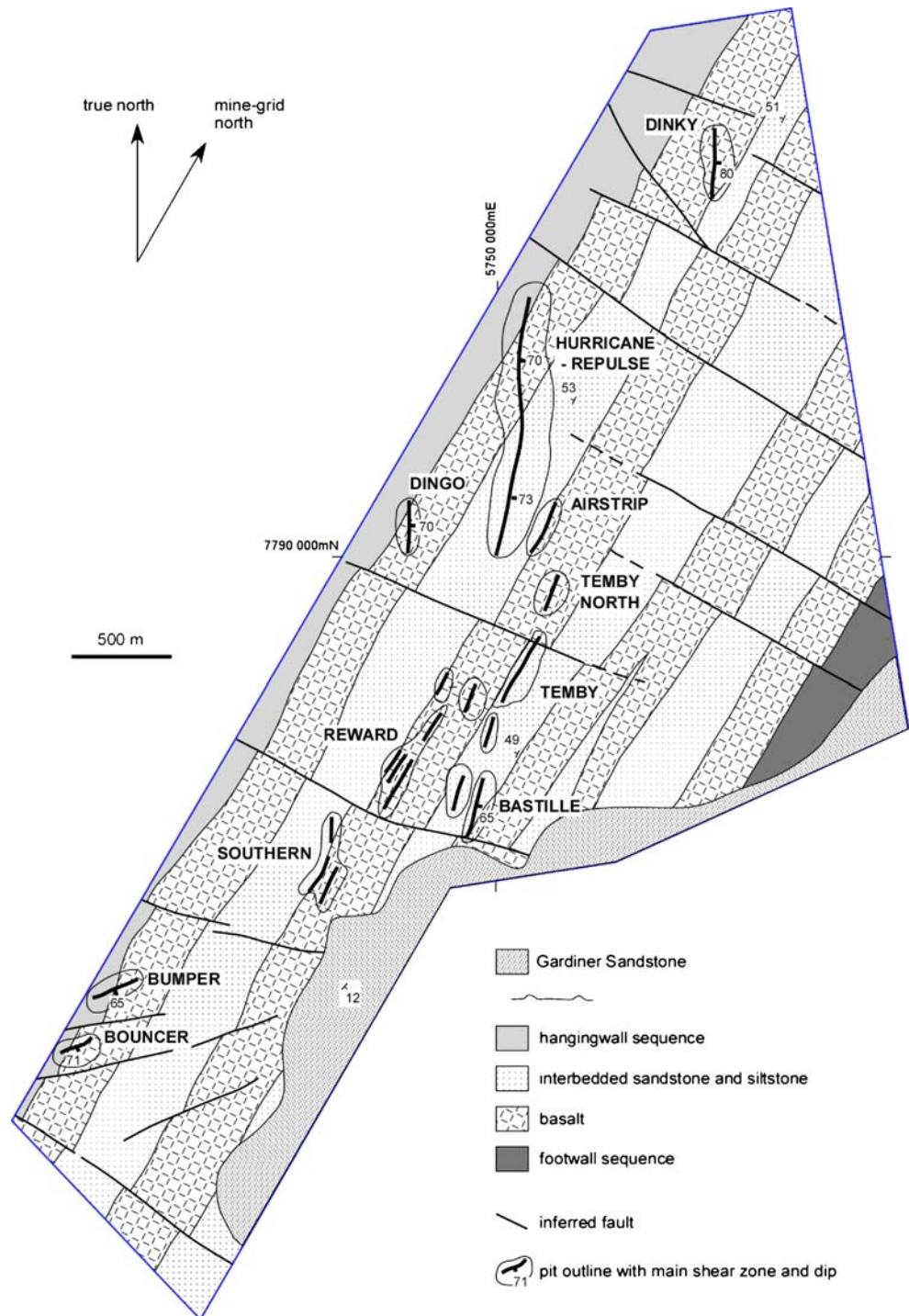
Basaltic rocks

Three distinct basaltic lithofacies can be recognized in the district. They are: massive basalts, brecciated basalts, and pillowed basalt.

Massive basalts are the most common basaltic facies present. Thick (20 m), internally structureless flow units are composed predominantly of fine-grained (<0.2 mm) albitized plagioclase laths in a groundmass of volcanic glass that has been altered to chlorite and epidote (Fig. 5b). Fine-grained magnetite is abundant and is responsible for the high magnetic response of the basalt.

Basaltic breccias mark the top of several massive units. In one case, a single breccia unit approximately 10 m thick can be traced for several kilometers, forming a useful marker horizon. The breccias are typically monomict with very angular to angular clasts, and grade from poorly sorted matrix-supported varieties to in situ, jigsaw-fit breccias that

Fig. 4 Geology of the northern part of the Tanami Goldfield showing the position of major open cut mines and the trends of major ore structures. The geology has been mapped from open pit exposures and extensive drilling across the mining lease



are clast-supported. These breccias are interpreted to be the products of autobrecciation during emplacement (McPhie et al. 1993).

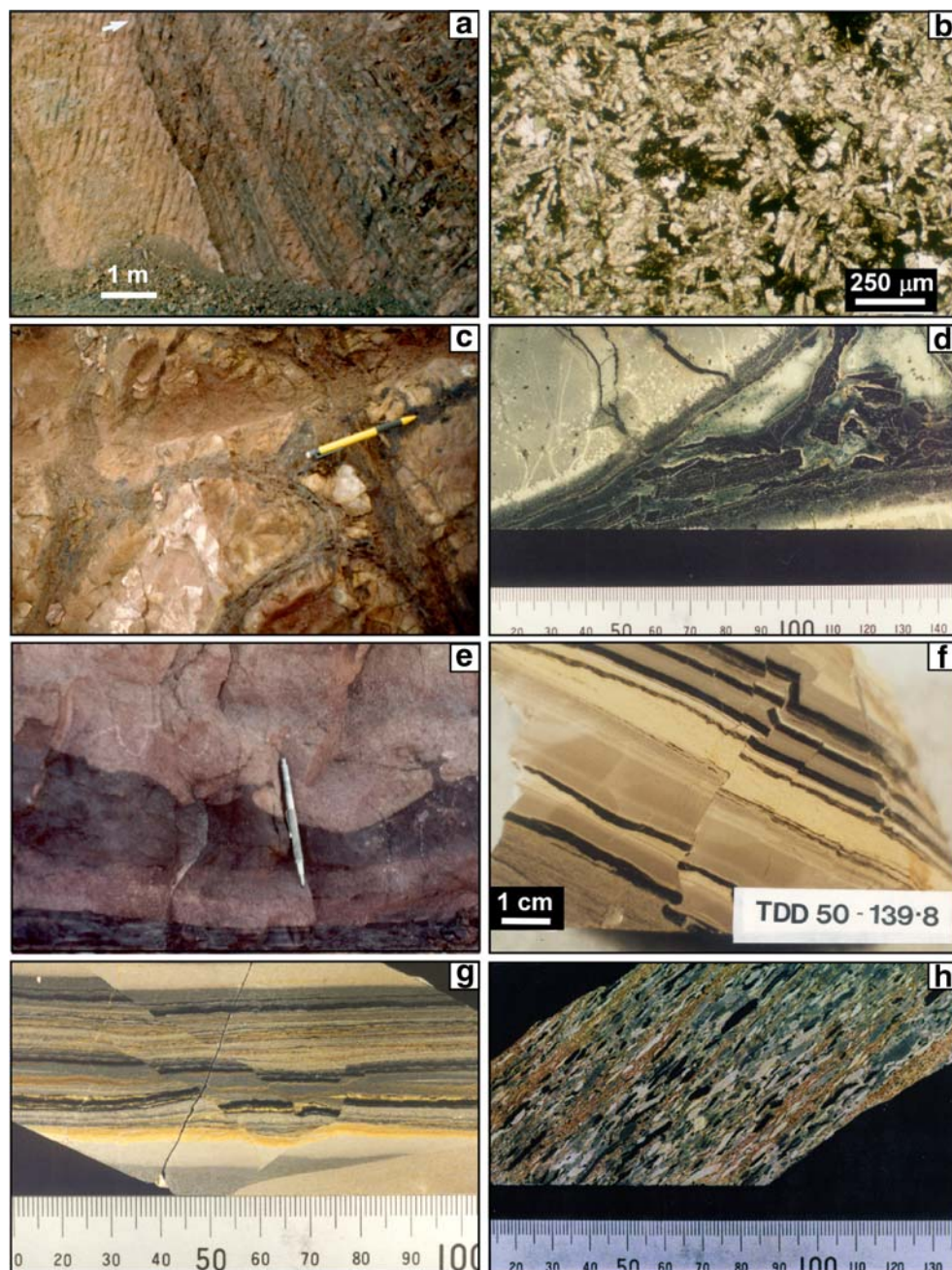
There are several essentially strata-bound zones of *pillow basalt* within the massive basalts. The zones are up to 8 m thick and individual pillows can be as large as 1 m in diameter. Triple point contacts between pillows are filled with quartz and carbonate (Fig. 5c,d). The petrology of the

pillowed basalts is similar to that described for the massive basalts.

Sedimentary rocks

Sedimentary packages comprise approximately half of the mine sequence. Four main lithofacies are present; sandstones,

Fig. 5 **a** Sharp irregular contact between a massive basalt flow BV and an overlying interflow sedimentary package within the Bouncer Open Pit. **b** Photomicrograph taken in PPL illustrating typical texture of the basalt. **c** Pillow basalt from the western wall of the Southern open pit. Note the massive quartz infill at the pillow triple point. **d** Close-up view of pillow triple point in drill core. Note the narrow alteration halo around chlorite veins and pillow edges. **e** Load cast at the base of a thick sandstone bed overlying a massive siltstone bed within the Hurricane-Repulse Pit. **f** Typical laminated carbonate/carbonaceous siltstone of the carbonate mudstone lithofacies. **g** Extensional microfaults within the finely laminated carbonate mudstone facies. Note the thickening of units within the central graben structure. **h** Intraclastic conglomerate from the uppermost carbonate mudstone horizon. Note the flattened clasts and clast imbrication



mudstones, carbonaceous mudstones, and intraclast conglomerates, all of which can be divided into separate facies.

The sandstone lithofacies comprises thin bedded (1 to 15 cm) sandstones with internal normal grading, cross and parallel laminations, and medium to very thick bedded (>10 cm to 2.5 m) sandstones and pebbly sandstones (Fig. 5e). Detrital mineralogy in the sandstones includes quartz, basaltic lithic fragments, bedding parallel aligned muscovite, rare biotite, and tabular feldspars that have been completely replaced by ankerite. Clays and fine-grained rutile are abundant in the matrix, as are fine iron hydroxides

(limonite) and minor amounts of hematite, which gives the rocks a characteristic red-brown color.

The mudstone lithofacies includes purple mudstone interlaminated with siltstone or fine-grained sandstone. It varies from dominantly siltstone to dominantly sandstone facies. The carbonate mudstone lithofacies occurs as a limited facies that can be traced laterally for up to 10 km, forming a distinctive marker horizon known locally as the “pale sediments”. Up to five separate packages of this group, varying in thickness from 1 to 5 m, occur interbedded with the *mudstone lithofacies*. The most

common sedimentary structure is that of interlaminated to interbedded light gray siltstones and black carbonaceous layers (Fig. 5f,g). Gray laminae are dominantly carbonate intraclasts within a carbonate mud and sericite matrix with rare detrital quartz grains. Framboidal pyrite is abundant within the carbonaceous layers.

Units of the *intraclast conglomerates* occur as distinct beds of variable thickness throughout the sequence (Fig. 5h). Ranging from thin horizons (<10 cm) to massive or graded beds from 20 cm to 2 m, they are dominantly matrix supported. Intraclasts vary in size up to 20 cm, but are generally less than 10 cm, and are typically round to sub-round shaped. The majority of intraclasts are composed of medium-to fine-grained siltstones, which typically preserve sedimentary structures such as grading or lamination. Basaltic clasts are rarely present. The presence of 10 to 40 m thick, thinning and fining upward cycles within the sedimentary packages are consistent with this model and could indicate gradual lateral lobe shifting from a lobe center to a lobe fringe environment within a submarine fan (Walker 1984).

Intrusive rocks

Several narrow (<3 m thick) weathered felsic dykes have intruded throughout the mine sequence and are exposed in pit walls. The dykes are now composed of sericite, clay, and spherulitic quartz and are interpreted to be aplitic dykes sourced from the nearby granitic plutons.

Most of the felsic dykes have been cut and displaced by the mineralized faults. This is evident at several localities, including: (1) Hurricane-Repulse, in the northern end of the pit, where a narrow dyke intruded subparallel to bedding and has been displaced by a mineralized fault (Fig. 6a,b); and (2) Hurricane-Repulse south, where a 070°-trending mineralized shear zone has crosscut and altered a narrow N-trending dyke (Fig. 6c,d). Other examples suggest synchronous or pre-mineralization timing for some felsic dykes. In the Bouncer open pit, a thin felsic dyke has intruded parallel to a mineralized shear zone, approximately 5–6 m in the hanging wall of the fault (Fig. 6e,f). A 2- to 3-m wide felsic dyke oriented at 094°/55°S has truncated the main N-trending mineralized zone in the central part of the Hurricane-Repulse Pit (Fig. 6g). The contact between this dyke and the mineralization is sharp. Thin (2–3 cm) quartz-carbonate veins and associated alteration occur on either side of the dyke but were not observed within the dyke (Fig. 6g). These conflicting contact relationships demonstrate that the timing of dyke intrusion is closely related to, and approximately synchronous with, the gold mineralizing event in the Tanami district.

Mine scale structure

Pre-mineralization faulting

Extensional faulting has resulted in abundant micro-faults with centimeter displacements in sedimentary units (Fig. 5g). Later, undisrupted sedimentary layering has draped these blind structures, indicating a syn-sedimentary origin. Although only small-scale (up to 20 cm long) micro-faults have been observed during the current study, larger scale syn-sedimentary faults are also believed to be present. This is because Carthew et al. (unpublished company report 1982) described larger growth faults with several meters of displacement that had been draped by later sedimentary layers in the (now back-filled) Reward open pit.

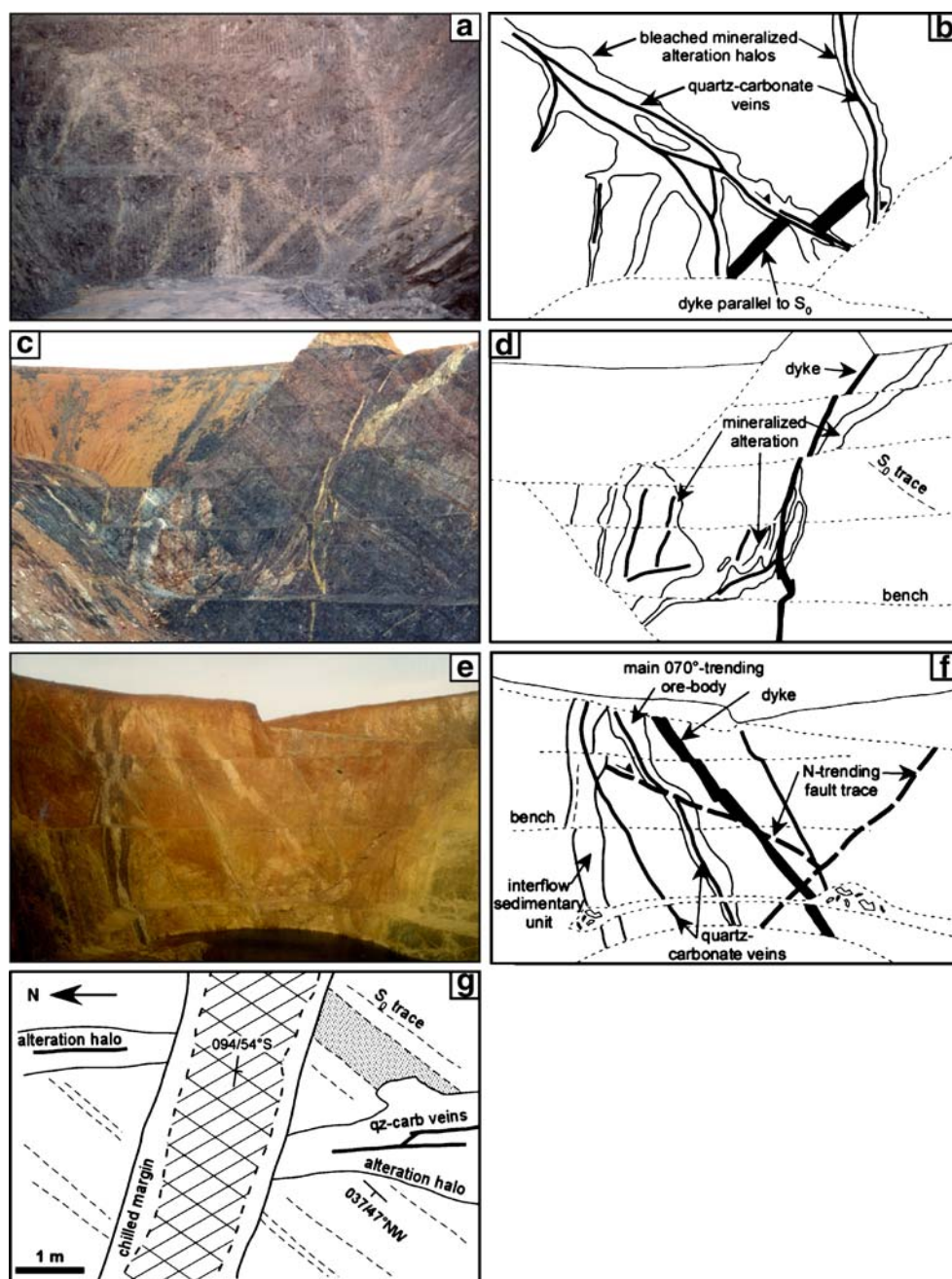
The major pre-mineralization faulting event comprises a series of low-angle thrusts that typically form ramp and flat geometry with respect to bedding (Fig. 7). These thrusts have sharp boundaries and show only minor, if any, evidence of cataclasis or foliation. The maximum observed displacement is approximately 25 m. Quartz fibers and fault plane striations are common, and these kinematic indicators, in combination with the dips of the faults, record a dominantly dip-slip reverse movement (Fig. 8a). The bedding-plane movement zones, including the thrust faults and duplex structures that contain fault striations sub-perpendicular to the regional F_1 fold axis (Fig. 3), are interpreted to have formed during flexural slip folding. However, this hypothesis is difficult to prove because the Tanami goldfield is situated on one limb of a regional scale fold.

Faults active during mineralization

Mineralized faults in the open pits form a complex array with considerable variation in style and orientation. The major mineralized structures trend 350–000° and 070° and dip steeply to the east and southeast, respectively (e.g., Fig. 6). Other steeply dipping subsidiary faults are oriented at 030–040° and are splays from the main 000° trend (Figs. 8b, 9 and 10). Distinctive bleached alteration halos (up to 3 m wide) characteristic of the mineralized faults are preferentially developed in the hanging wall. A weak continuous cleavage defined by sericite is locally developed within the alteration halo adjacent to the main mineralized faults.

Displacement across these faults is typically <5 m and is accommodated across sharply defined planar faults accompanied by only minor zones (<5 cm) of cataclasite. Thin (generally <5 m) quartz-carbonate fault veins typically have a similar orientation to the major structures. Mineralized veins exhibit crack-seal textures (Ramsay 1980), single-stage growth textures, and zones of intense hydraulic brecciation with wallrock clasts up to 2 cm in diameter sealed in quartz-carbonate cement. Within the crack-seal

Fig. 6 Photographs and interpretive sketches of relationships observed between mineralized shear zones and felsic dykes in open pits of the Tanami district. **a** and **b** Narrow felsic dyke emplaced subparallel to bedding has been displaced by a mineralized fault, northern end of the Hurricane-Repulse pit. **c** and **d** Narrow N-trending felsic dyke has been crosscut and altered by a 70°-trending mineralized shear zone, southern end of the Hurricane-Repulse pit, looking west. **e** and **f** Narrow felsic dyke has been intruded parallel to an 70°-trending mineralized fault zone, and both are cut by a late N-trending sinistral strike slip fault, Bouncer open pit, looking east. **g** Interpretive sketch of a 2–3 m wide felsic dyke oriented 094°/55°S that has cut the main N-trending mineralised zone in the central part of the Hurricane-Repulse pit. Thin (2–3 cm) quartz-carbonate veins and their related alteration halos have been cut by the dyke



style of veins, individual opening increments, are defined by bands of wall rock inclusions much less than 1 mm thick. In many cases, the primary vein textures have been overprinted by later deformation textures, including fine bands of breccia and cataclasite.

Petit's (1987) criteria were used for establishing movement vectors on brittle structures, but extensive weathering of the mineralized faults within the open pits has made the recognition of these kinematic indicators difficult. The most common criteria used were fiber veins and frictional striations, such as deepening grooves. In several cases, striated fault planes were observed, but no sense of shear could be deduced from the striations. In these instances,

offsets of sedimentary marker units were used to determine the displacement. No overprinting of two or more sets of striations with different orientations on the same fault plane was observed for the mineralized structures. The striations that were observed have shallow plunges consistent with dominantly strike slip movement.

The limited evidence recognized indicates that the N- and 030°- to 040°-trending structures underwent sinistral-normal movement, and the 070° trend underwent a dextral-reverse displacement. In both cases, strike-slip movement is dominant. As with the pre-mineralization structures, displacement across individual faults is generally <5 m. The highest grade and most continuous mineraliza-

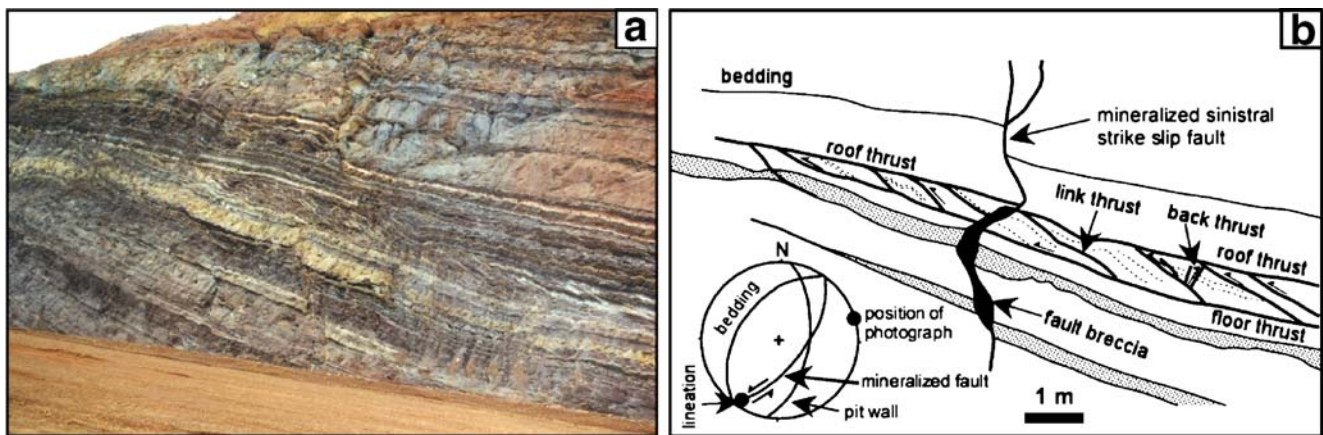


Fig. 7 **a** Photograph and **b** interpretive sketch of a thrust duplex located in the western wall of the Hurricane-Repulse pit. Note the flat-roofed nature of duplex and the crosscutting nature of the weakly mineralized

fault. The anastomosing nature of the mineralized fault is exaggerated by a combination of the similar orientations of the pit wall and the fault, and the uneven nature of the pit wall

tion was found in the Hurricane-Repulse Pit (Fig. 9) and is spatially related to both the N-trending faults and the intersection of the N- and 070°-trending faults, which has resulted in steeply southeast-plunging high grade (>0.8 g/t Au) ore shoots. Rare mineralized faults also occur along bedding planes within the sedimentary lithologies. An excellent example of this style of mineralization occurs on the north-eastern pit wall at the Bouncer Pit, where essentially strata-bound mineralization is confined to a thin (5 m) interflow sedimentary horizon (Fig. 10).

The array of brittle mineralized structures in the Tanami district shows some similarity to the Riedel model, even though the Riedel geometry experiments used a homogeneous medium, which differs from the layered stratigraphy of the Tanami goldfield. The 350–000° trend can be considered as the major D shears, whereas the 030–040° trend would be the Riedel shears and the 070° trend are the antithetic R' shears (Fig. 8b). The sense of displacements observed on the mineralized faults is consistent with a sinistral wrench model, because the R and D orientations have dominantly sinistral strike-slip movement, and the R' shears have dextral displacement. The association of crack-seal veins and a range of undeformed to strongly deformed vein fabrics in fault fill veins indicate that fault dilation, slip and internal deformation has been a cyclic process (Cox 1991). Processes of this nature have been attributed to cyclic variations in both the fluid pressure and the shear stress associated with fault valve behavior (Sibson et al. 1988).

Calculated stress conditions

The stress inversion methods of Etchecopar et al. (1981) have been used to calculate the stress field responsible for strain during the mineralizing event. This method requires that the orientations of (1) the fault plane, (2) the fault

striation, and (3) the sense of shear be known. A total of 36 fault striation measurements were collected from mineralized faults and veins (Fig. 11a). Striations on mineralized structures have shallow plunges consistent with dominantly strike slip movement. Within the three major orientation subdivisions, the 345–010° and the 030–040° striking faults mostly preserve evidence for sinistral displacement, whereas the 060–070° striking faults are dominantly dextral, with a minor component of reverse displacement (Fig. 11a). Stress inversion analysis of the mineralized faults show that σ_1 was plunging subhorizontal to the SE, with σ_2 subvertical and σ_3 subhorizontal to the SW (Fig. 11b).

Post-mineralization faulting

Post-mineralization deformation resulted in reactivation of early bedding-parallel reverse faults. Other post-mineralization reverse faults with E–W trends are also present. The maximum observed displacement on these structures is less than 5 m. The E–W faults are accompanied by extensive zones of supergene kaolinitization that overprint all other structures.

Mineralization and alteration

Three hydrothermal alteration types have been identified: 1) carbonate–chlorite–quartz alteration associated with pillow basalts (e.g., Fig. 5c,d); 2) a regional chlorite–carbonate alteration assemblage; and 3) sericite–quartz–carbonate–pyrite alteration associated with gold mineralization (e.g., Fig. 6). Supergene weathering has also affected the district (e.g., Fig. 6).

Five vein stages were identified and are shown schematically in Fig. 12. The complete paragenetic sequence has been inferred using crosscutting relationships from

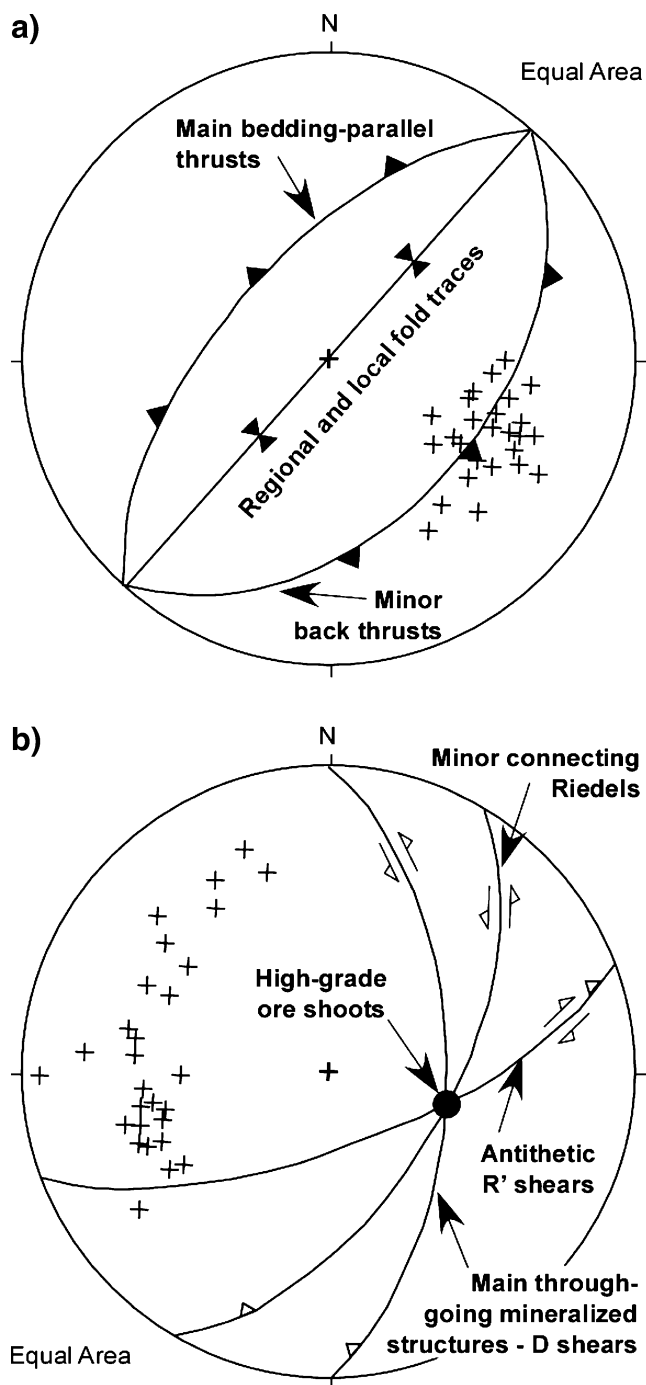


Fig. 8 Equal area stereonets. **a** Illustrates the orientation of the major pre-mineralization structures (teeth on hanging wall block, crosses represent poles to bedding). **b** The major structures active during the Au mineralizing event (crosses represent the poles to measured fault planes)

several areas across the mine lease. No single locality contained all five vein stages. The complete paragenesis is shown in Fig. 13. Stage III veins are associated with a regionally developed carbonate–chlorite alteration assemblage and stage IV veins are associated with Au mineralization. Stages I, II, and V veins do not possess significant

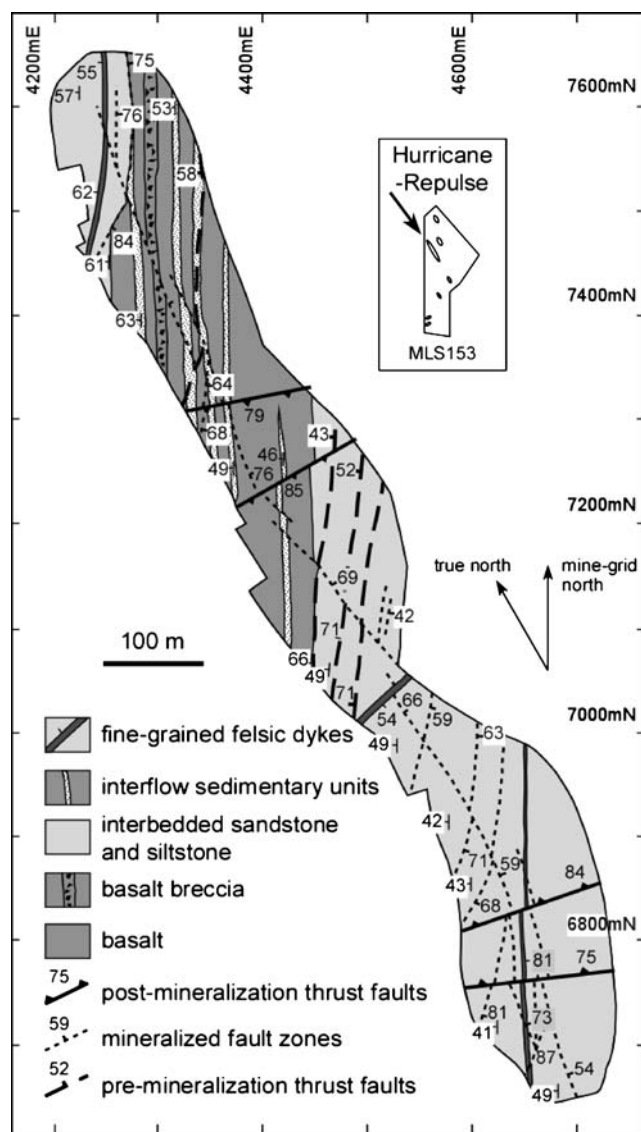


Fig. 9 Simplified geology of the Hurricane-Repulse Open Pit. Geology projected to a common RL

alteration halos and are not spatially or temporally associated with Au mineralization.

Stage I

Stage I veins are associated with extensional microfaults. These veins are only developed along fault planes within the carbonate–mudstone facies, and have no associated alteration halos. The veins consist of dolomite + hematite + sericite ± quartz, and are typically narrow (<1 mm wide) and up to 4–5 cm in length.

Stage II

Stage II veins are associated with bedding-parallel reverse faults and occur mostly within the sedimentary units. There

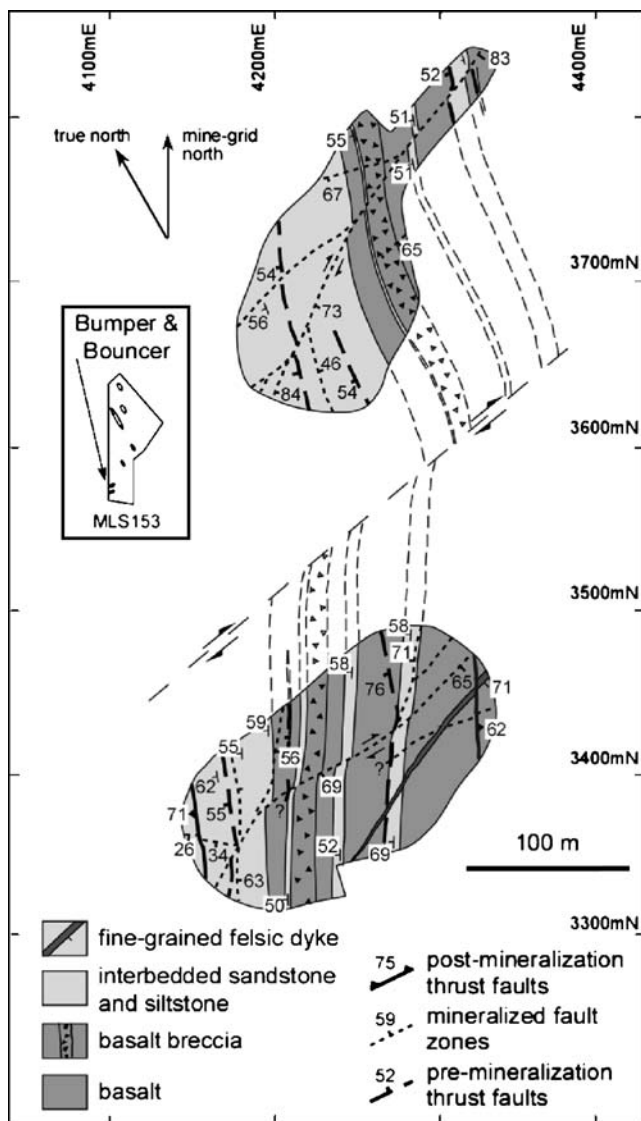


Fig. 10 Simplified geology of the Bumper and Bouncer open pits. Geology projected to a common RL

are no discernible alteration halos around stage II veins. Within quartz-rich sedimentary units, stage II veins contain quartz ± pyrite ± chalcopyrite ± covellite. The veins contain dolomite + chlorite ± quartz ± sericite ± hematite in the carbonate–mudstone facies.

Stage III

Stage III chlorite ± calcite ± ankerite ± sericite ± quartz veins are associated with chlorite–carbonate alteration within the basalts and carbonate alteration in the sedimentary rocks. Veins vary from 0.1 to 20 mm in width. A pervasive chlorite–carbonate alteration assemblage associated with stage III veins has affected the bulk of the mine sequence.

Basalts that have undergone low-intensity alteration have an assemblage of albite–relict pyroxene–magnetite–chlo-

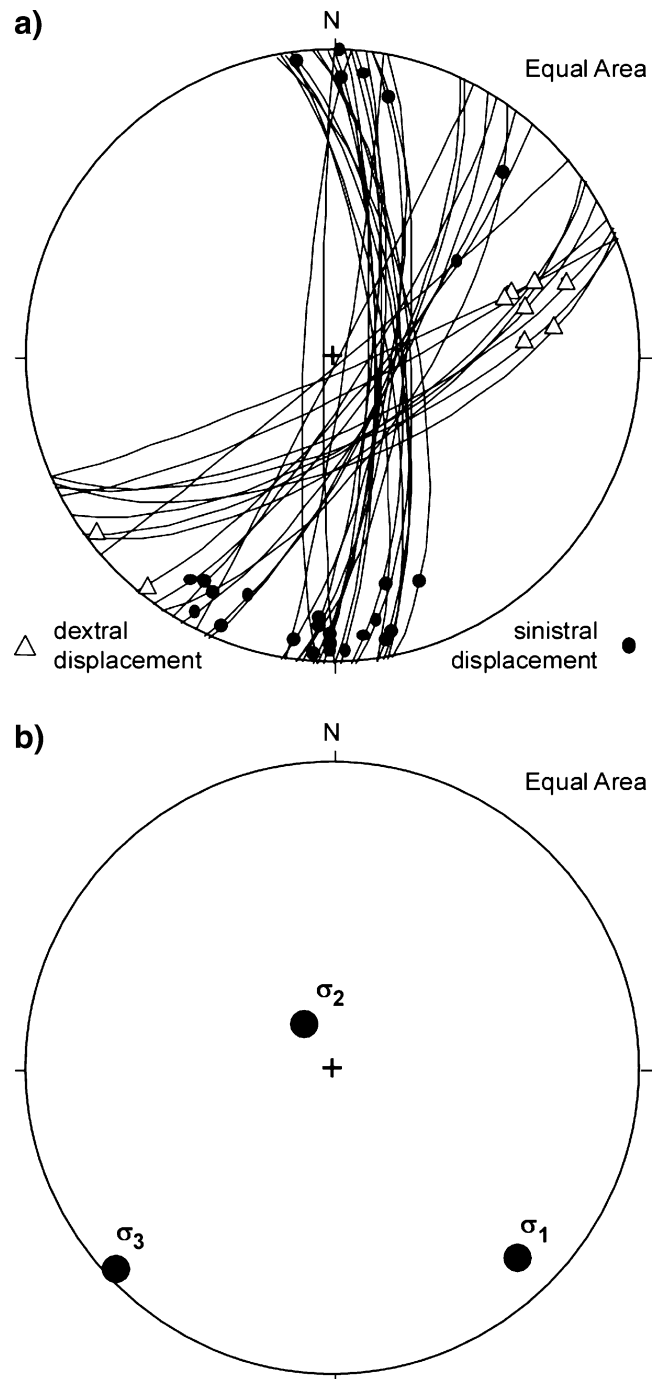


Fig. 11 Equal area stereonets. **a** All data used in stress inversion calculations for the mineralized shear zones. **b** Calculated principal stress orientations for the mineralizing event. '+' denotes pole to mineralized faults

rite. The groundmass is extensively altered to chlorite, with minor, ragged clots of calcite and dolomite. With increasing alteration intensity, pyroxene has been replaced by a fine-grained aggregate of chlorite, sericite, and minor carbonate. Plagioclase has been more intensely sericitized and large (up to 1 mm) clots of calcite are present throughout the matrix. Abundant fine-grained opaque phases (magnetite or

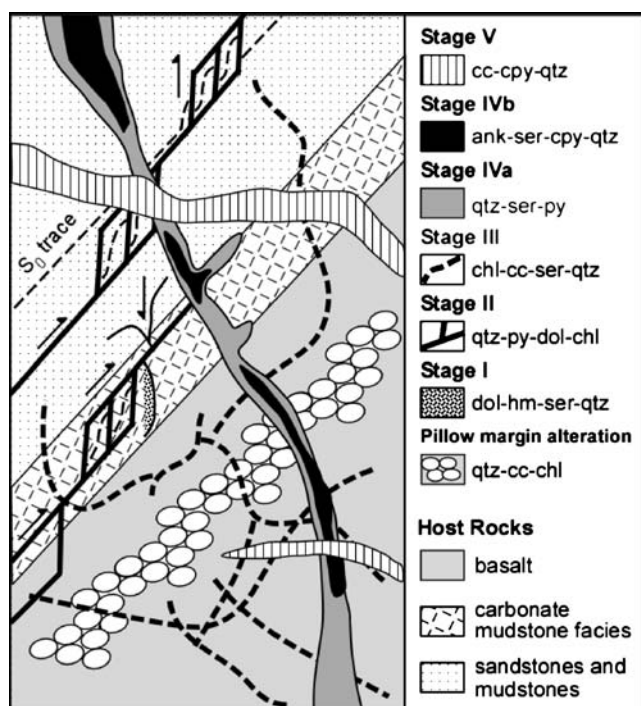


Fig. 12 Schematic illustration of paragenetic relationships in the Tanami district. The observed crosscutting relationships are shown as is the host rock control on the distribution of vein stages. The assemblages developed within each stage are summarized in the legend and discussed in the text. Abbreviations: *qtz* quartz, *cc* calcite, *ank* ankerite, *dol* dolomite, *ser* sericite, *hm* hematite, *py* pyrite, *cpy* chalcopyrite, *chl* chlorite. Gold mineralization is confined to stages IVa and IVb

titanomagnetite?) occur in the altered groundmass. Based on the presence and abundance of stage III veins, approximately 70–80% of the basalts have been altered to the chlorite–carbonate assemblage. Relict kernels of unaltered basalt are preserved at several localities.

Stage III alteration within the sedimentary rock types resulted in the complete replacement of detrital feldspars by ankerite $\pm$ minor sericite and chlorite. Ankerite is also present, as ragged clots scattered throughout the matrix. Chlorite alteration of the matrix varies from rare patches intergrown with sericite to pervasive chlorite alteration.

It has not been possible to link stage III veins to a specific deformational event. Stage III veins are typically extension veins, which in places have formed minor stockworks. The district-scale distribution of the chlorite–carbonate alteration is poorly constrained due to the lack of information available outside the immediate mine environment. Mapping within the Tanami district suggests that most of the basaltic and all the sedimentary rocks of the mine sequence have been affected by chlorite–carbonate alteration.

Stage IV (gold mineralization)

Stage IV veins and breccias and associated alteration halos are the hosts to economic Au mineralization. Stage IV vein and breccia assemblages are complex and appear to have

	Pillow margin alteration	Stage I	Stage II	Stage III	Stage IVa	Stage IVb	Stage V	
quartz			-----	-----	-----	-----		Abundance
chalcedony		-----	-----	-----	-----	-----		Trace (< 1%)
sericite		-----	-----	-----	-----	-----		Minor
illite			-----	-----	-----	-----		Major (> 5%)
chlorite			-----	-----	-----	-----		
calcite				-----				Distribution
ankerite			-----	-----				Noted in all lithologies
dolomite		-----	-----	-----				
arsenopyrite					?			 Confined to basalts
sphalerite					?			
pyrite			-----			?		
chalcopyrite			-----					 Confined to sediments
galena	 ?							
hematite		-----						 Confined to carbonate mudstone facies
rutile								
gold					?	?		

Fig. 13 Vein paragenesis for gold deposits of the Tanami district

evolved temporally. For this reason, stage IV has been subdivided into stage IVa and stage IVb.

Stage IVa veins—gray quartz

Stage IVa is characterized by the assemblage gray quartz $\pm$ sericite $\pm$ pyrite $\pm$ chlorite $\pm$ sphalerite $\pm$ arsenopyrite $\pm$ gold. Although all of these phases are present in stage IVa veins, quartz, sericite and pyrite are predominant and characteristically constitute 95% of the vein fill. Macroscopic vein textures include open-space fill, massive vein fill, and hydrothermal breccia zones with entrained wall-rock clasts up to 10 cm in diameter.

Quartz and sericite are abundant in stage IVa veins. Quartz occurs as subhedral to anhedral crystals up to 2 mm in diameter intergrown with fine-grained (<0.5 mm) patches of sericite. Minor amounts of fibrous chlorite (<5 vol.% of vein fill) are intergrown with quartz and sericite, particularly at vein margins. Euhedral to subhedral grains of pyrite range up to 0.5 cm, but are more commonly less than 1 mm in diameter. Individual grains have coalesced to form clots of pyrite up to 2–3 cm in diameter. These clots typically occur at vein margins, where pyrite is intergrown with sericite, quartz, and chlorite. Pyrite commonly contains small (<200 μm) inclusions of anhedral sphalerite and arsenopyrite. Gold occurs as minute inclusions up to 15 μm in diameter contained within pyrite grains.

Stage IVb veins—banded ankerite/quartz

Stage IVb veins and breccias contain ankerite—quartz $\pm$ chalcopyrite $\pm$ chlorite $\pm$ gold $\pm$ sericite $\pm$ pyrite $\pm$ calcite. Stage IVb occurs as discrete veins that crosscut, and therefore postdate, stage IVa veins. Stage IVb vein fill also occupies the central portions of some stage IVa veins.

Ankerite is the most abundant mineral phase in all stage IVb veins, constituting between 40 and 95% of the vein fill. Ankerite occurs as euhedral crystals up to 4 mm long, and as coarse (2–3 mm) anhedral grains. The presence of microscopic inclusions of wall rocks, and fluid inclusion trails within ankerite that are developed parallel to the vein margins are interpreted to indicate crack-seal processes operating during the formation of the veins (Ramsay 1980; Cox 1987, 1991). Quartz occurs as anhedral grains interstitial to the ankerite, as subhedral grains intergrown with ankerite, and as subhedral to euhedral crystals up to 10 mm long that have sealed the final stages of stage IVb veins. Some of the euhedral quartz crystals contain growth bands defined by fluid inclusions.

Chalcopyrite is the predominant sulfide phase in stage IVb veins, occurring as euhedral to subhedral grains up to 0.2 mm in diameter that have rimmed stage IVa pyrite. Chalcopyrite has also sealed microfractures in stage IVa

pyrite and quartz. Subhedral free gold up to 15 μm in diameter typically occurs as inclusions in chalcopyrite that have sealed fractures in pyrite, interstitial to stage IVa pyrite, and stage IVb chalcopyrite, and associated with carbonate inclusions that have infilled cavities in stage IVa pyrite.

Euhedral to subhedral pyrite and ragged patches of sericite are minor constituents of stage IVb, and generally occur as intergrowths with ankerite and quartz. Sericite together with ankerite and chlorite have also sealed microfractures in stage IVa pyrite.

Stage IV—sericitic alteration

Hydrothermal alteration associated with stage IV veins has resulted in the bleaching of the basaltic and sedimentary lithologies and has produced a sericite–quartz $\pm$ pyrite $\pm$ carbonate assemblage. Due to the common close spatial and temporal association of stage IVa and IVb veins, it is not possible to distinguish alteration assemblages associated with each vein substage. The alteration assemblages surrounding stage IVa and IVb veins have, therefore, been grouped together as a single-alteration type.

Sericitic alteration halos occur as broad (up to 10 m) pervasive zones surrounding mineralized shear zones (Fig. 6), and as narrow halos (<10 cm wide) surrounding individual veins. Mineral zonation is evident within many of the alteration halos ($\approx 70\%$ of observed cases) with an inner sericite–quartz–pyrite zone (SQP alteration) and an outer sericite–quartz zone (SQ alteration). At other localities, pyrite occurs throughout the sericitic halo.

The minerals that characterize the sericitic alteration assemblage are sericite–quartz $\pm$ pyrite $\pm$ ankerite $\pm$ siderite $\pm$ chalcopyrite $\pm$ leucoxene $\pm$ anatase $\pm$ rutile. Sericite is the most abundant alteration mineral (up to 50 vol.%) and has completely replaced plagioclase in the basalts. Fine-grained quartz (<0.1 mm) is intergrown with sericite throughout the bleached zone. Euhedral to subhedral pyrite is common (up to 80 vol.%) within the inner SQP halos, with individual grains up to 2 mm in diameter. Aggregates of pyrite up to 4 cm in diameter occur at several localities. Within sedimentary host rocks, there is a positive correlation between primary grain size and the grain size of secondary pyrite, which has replaced hematite, biotite, muscovite, and ankerite. Within the basalts, pyrite has replaced magnetite and chlorite. Magnetite destruction has resulted in a decrease in the magnetic susceptibility of the basalts to values less than 1.0×10^{-3} S.I. within the SQP alteration zones ($15\text{--}60 \times 10^{-3}$ S.I. is the range for non-altered basalts in the Tanami district).

Ankerite is the most abundant carbonate mineral in the sericitic alteration halos, with lesser amounts of dolomite and siderite also identified by electron microprobe. Secondary

carbonates occur as fine-grained anhedral intergrowths with quartz and sericite after primary plagioclase. The amount of carbonate within the sericitic alteration halos varies from 10 to 40 vol.% depending on the presence or absence of stage IVb veins. The carbonates have replaced plagioclase laths in the basalts and occur in the matrix of the sedimentary lithologies.

One of the major difficulties in determining the abundance of stage IV carbonates within the sericitic alteration halos is the presence of earlier generations of carbonate. Testing within basaltic lithologies with acid (10% HCl) has shown that the transition from calcite to ankerite as the predominant carbonate phase corresponds to the approximate onset of visual bleaching. However, the acid test does not work in the sedimentary lithologies, where both the stage III and stage IV carbonates are ankeritic.

Minor chalcopyrite is restricted to the inner ≈ 2 cm selvage around stage IVb veins, where it has rimmed and filled fractures within pyrite. Gold has not been observed petrographically in the alteration halos, but its presence has been confirmed by geochemical assay. Rutile, leucosene, and anatase are all present within stage IV alteration halos. Titanium-oxides occur as fine-grained solid inclusions ($<20 \mu\text{m}$) within quartz and/or intergrown with quartz and sericite. Fine-grained Ti-oxides have coalesced to form rounded aggregates to 1 mm in diameter at several localities.

Stage V

Thin (0.2 to 30 mm wide) stage V calcite $\pm$ chalcopyrite $\pm$ quartz veins have consistently crosscut and displaced stage IV veins. Approximately 80% of stage V veins have a massive texture, although both banded and crack-seal textures have also been observed. Assay results have confirmed that stage V veins do not contain gold (Tunks 1996).

Fluid chemistry

Fluid inclusions

Fluid inclusion data were obtained from stage II, IV, and V veins. Stage III veins did not contain workable fluid inclusions, and stage I was not examined. Four types of fluid inclusions have been recognized during this study. Type 1 are two-phase [liquid–vapor (L–V)], liquid-rich ($V < 20\%$) fluid inclusions that homogenize by vapor disappearance. Type 1A fluid inclusions are two-phase (L–V) with variable liquid–vapor ratios ($V \leq 80\%$) that homogenize by vapor disappearance. Type 2 fluid inclusions are two-phase (L–V) and vapor-rich ($V > 50\%$); these homogenize by liquid disappearance. Type 3 fluid inclusions are single-phase,

liquid-filled. None of the fluid inclusions examined during this study contain visible CO_2 or daughter minerals. Fluid inclusions were classified as primary or secondary based on the criteria of Roedder (1984).

A total of 270 fluid inclusions were analyzed microthermometrically from 19 thin sections at the University of Tasmania using a modified USGS gas flow heating–freezing stage supplied by Fluid Inc., Denver, CO, USA. (Stern and Bodnar 1984; Bodnar and Stern 1987). Calibration of the equipment was performed using synthetic fluid inclusions manufactured by Syn-FliC, and the precision of measured temperatures are $\pm 1^\circ\text{C}$ for heating and $\pm 0.3^\circ\text{C}$ for freezing. The accuracy of the ‘Trendicator’ thermocouple measurements was calibrated against the CO_2 triple point (-56.6°C), the melting point of water (0.0°C), and the critical point of water (374.1°C) in synthetic inclusions. Salinities were calculated from freezing point depression measurements by assuming a $\text{NaCl-H}_2\text{O}$ system, and using the algorithm of Bodnar et al. (1989). This method provides only an estimate of the true salinity, due to the effects of other dissolved salts in the fluids that are not taken into consideration. Freezing point determinations were obtained for approximately half of the 270 fluid inclusions that were analyzed, because the small size of many fluid inclusions prevented measurement of the temperature of final ice melting. Only stage IV fluid inclusion data are described in detail below. Fluid inclusion results for stages II and V are summarized with stage IV data in Fig. 14, and are reported in detail in Tunks (1996).

Stage II quartz

One sample of a stage II vein collected from the base of the Hurricane-Repulse open pit contained euhedral quartz crystals (up to 2 mm in diameter) with abundant, primary, Type 1 fluid inclusions in growth zones. Total homogenization temperatures ($T_{\text{HOM(L)}}$) range from 66 to 254°C , with most between 100 and 140°C (Fig. 14a). Salinity estimates based on final ice-melting temperatures vary from 15 to 21 wt% NaCl equivalent (Fig. 14a). Only five eutectic melting temperatures were recorded; all were between -52 and -48°C . Mernagh and Wygralak 2006 reported homogenization temperatures of 110 to 378°C and salinities of 0.1 to 19 wt% NaCl equivalent from two samples of pre-ore veins from the Southern open pit. It is not clear whether these veins correlate to stage II or stage III mineralization.

Stage IVa quartz

Fourteen doubly polished sections were prepared from this paragenetic stage. Stage IVa quartz is characterized by abundant secondary fluid inclusions ($<15 \mu\text{m}$ in diameter),

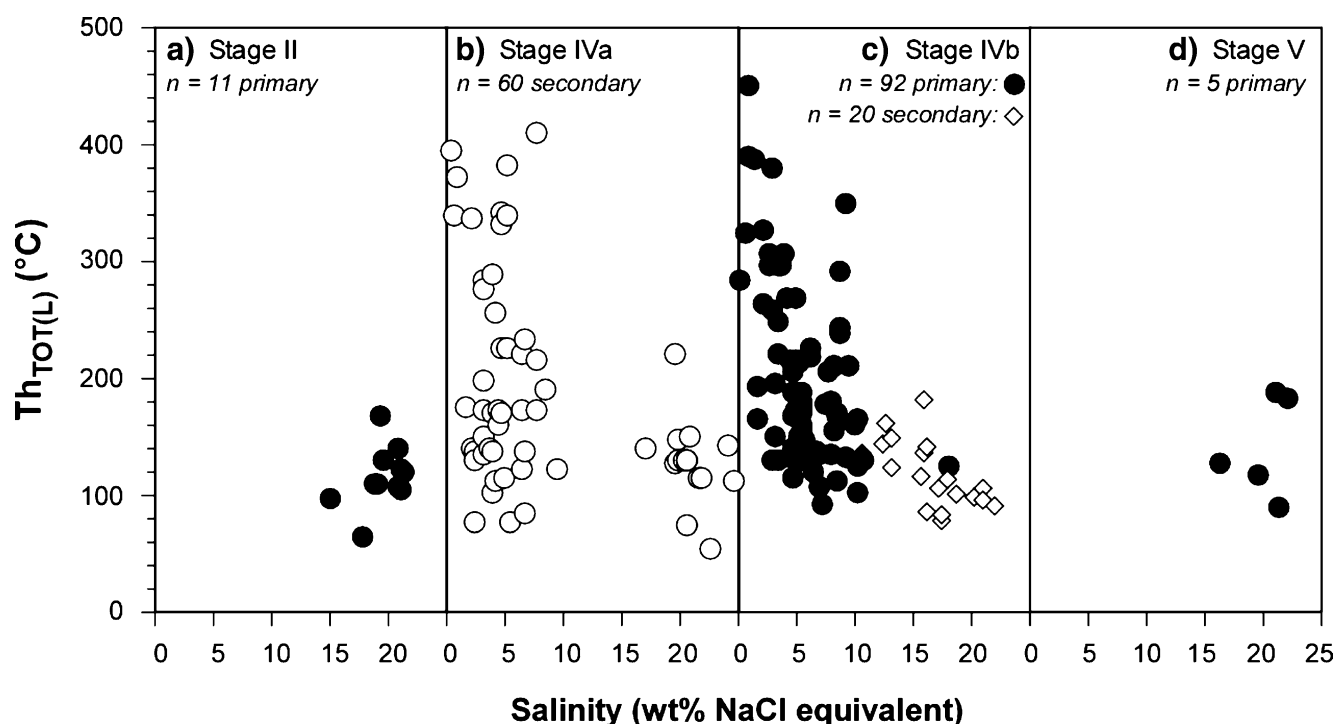


Fig. 14 $Th_{TOT(L)}$ (temperature of homogenization) vs salinity (wt% NaCl equivalent) for fluid inclusions from stage II, IV, and V veins. Data tabulated in Tunks (1996)

which occur as clusters and trails. Microthermometric data were obtained from secondary Type 1 (abundant) and Type 1A (rare) fluid inclusions. No primary fluid inclusions were analyzed. Rare growth bands defined by primary fluid inclusions were observed in stage IVa quartz, but the inclusions were too small for accurate measurement of microthermometric data.

$Th_{TOT(L)}$ ranges from 56 to 471°C (Fig. 14b; Tunks 1996). This large range probably relates in part to post-entrapment modification of some stage IVa fluid inclusions, although textural evidence for such modifications was not observed. Calculated salinities are bimodal, with a low- to moderate-salinity population (0 to 10 wt% equivalent NaCl) and a high-salinity population (17 to 25 wt% equivalent NaCl; Fig. 14b). Eutectic melting temperatures ranged from -52.8 to -21.0°C , with data peaks at -45 to -55°C (consistent with Ca-bearing brines), and at -20 to -30°C (indicative of Na–K–Cl brines with trace amounts of other salts). The high-temperature, low-salinity fluids are rich in Na–K, and are inferred in this paper to be related to mineralization due to their unique occurrence in stage IV veins, and their similarity to primary stage IVb fluid inclusions. In contrast, the high-salinity, low-temperature secondary fluid inclusions in stage IVa veins correlate with primary stage II and stage V fluid inclusion data (Fig. 14) and appear to relate to Ca-rich, low-temperature, saline basal brines that were unrelated to gold mineralization (Tunks 1996).

Stage IVb quartz

Primary fluid inclusions occur in growth zones within stage IVb quartz veins. The inclusions vary in size from 2 up to 40 μm . In some cases, fluid inclusions within growth bands are too densely clustered for microthermometric data to be obtained. Coexisting type 1 and type 1A fluid inclusions occur in some growth bands, providing local evidence of phase separation. Fluid inclusion shapes are variable, and include negative crystal, elongate, and irregular shapes. Data were also collected from secondary fluid inclusions that have formed along trails and healed fractures that cut primary growth zones.

Homogenization temperatures for primary fluid inclusions ranged from 90 to 451°C (Fig. 14c), although eight fluid inclusions were observed that did not homogenize below 500°C. A broad peak in the data occurs between 120 and 220°C (Fig. 14c), and these are considered representative of mineralization. The higher and lower temperature data may relate to post-entrapment modification of individual fluid inclusions, although no textural evidence was observed that would support this hypothesis. Salinity estimates, based on 89 final ice melting ($T_{m_{ice}}$) measurements, are mostly between 0 and 11 wt% equivalent NaCl with most around 5 wt% (Fig. 14c) and an outlier at 18 wt% NaCl. Eutectic melting temperatures range from -36.2 to -18.2°C , with a peak between -30 and -25°C . Mernagh and Wygralak 2006 reported a similar range of $Th_{TOT(L)}$

values of 101 to 375°C and salinities of 0 to 14.9 wt% equivalent NaCl for ore-stage aqueous fluid inclusions from MLS153 (Southern and Bastille pits), although they did not specify if these data were obtained from stage IVa or IVb veins.

Secondary fluid inclusions in stage IVb veins have compositions that are distinct from the primary stage IVb fluid inclusion data, but are similar to the primary stage II and stage V fluid inclusion data (Fig. 14). Homogenization temperatures range between 79° and 182°C, with a mean value of 120°C (Fig. 14c). Estimated salinities ranged from 9 to 22 wt% equivalent NaCl, and eutectic melting temperatures between −37° and −34°C.

Mernagh and Wygralak 2006 analyzed 15 CO₂-bearing fluid inclusions from gold-bearing veins in the Southern open pit. These inclusions had Th_{TOT(L)} values from 288 to 436°C, salinities of 4.0 to 11.22 wt% equivalent NaCl, CO₂ vapor mole fractions from 0.9 to 1.0 and CH₄ vapor mole fractions of 0.0 to 0.1. These are the only CO₂-bearing fluid inclusions that have been detected from MLS153. It is not clear whether they were hosted by stage IVa or IVb veins.

Stage V calcite

Two narrow (<2 mm) stage V calcite veins that crosscut stage IVa veins contained workable fluid inclusions. Both stage V veins contained rare, large (15–30 μm), isolated fluid inclusions that had negative crystal shapes; these textural features were used to classify them as primary (Roedder 1984). Eight workable fluid inclusions were found and analyzed by microthermometric techniques. Total homogenization temperatures range from 92 to 186°C, with a mean value of 149°C (Fig. 14d). Five salinities were calculated from final ice-melting temperatures. They range between 16 and 22 eq. wt% NaCl (Fig. 14d). Four eutectic temperatures were measured; values were between −25 and −30°C.

Depth of formation and temperature of mineralization

Pressure corrections are required to convert T_h values to trapping temperatures (i.e., the temperature of the fluid at the time of fluid inclusion formation). Metamorphism within the mine sequence has reached only diagenetic grades, based on the lack of metamorphic recrystallization and low illite crystallinity values (Tunks 1996), implying background temperatures of less than 250°C (Weaver 1984). A depth of burial can, therefore, be calculated based on an estimate of the geothermal gradient. Because Au mineralization is interpreted to be synchronous with the intrusion of felsic dykes, a high geothermal gradient (40°C) is assumed. This is considered to be reasonable, as regional metamorphism in the Australian Proterozoic is characterized

by high temperatures and low pressures (Loosveld and Etheridge 1990; Sandiford and Powell 1991; Scrimgeour and Sandiford 1993).

Based on the assumed geothermal gradient, a maximum depth of 6 km is estimated for the mine sequence during gold mineralization. Furthermore, the presence of crack-seal and extensional veins throughout the paragenesis is consistent with lithostatic fluid pressures (e.g., Cox et al. 1986, 1990). The transition from hydrostatic to lithostatic fluid pressures, varies widely depending on the presence or absence of impermeable seals, but typically occurs at approximately 3 km (Price 1966; Spear 1993; Cox et al. 1990). Au mineralization in the Tanami district is therefore inferred to have formed at depths between 3 and 6 km. This estimate is similar to that of Mernagh and Wygralak 2006, who used their fluid inclusion data to estimate depths of formation of 1.5 to 5.6 km for the Tanami goldfield. They preferred the shallower end of this depth range, due to the local occurrence of open-space fill vein textures.

Formation temperatures for stage IV veins can be estimated using the measured primary stage IV fluid inclusion homogenization temperatures and the estimated depth of formation. Lithostatic load can be calculated using the relationship $P = H\rho g$, where H = depth, ρ = density, and g = acceleration due to gravity. Using the range in depth estimates (3–6 km) and assuming rock densities of 2.5 g/cm³, lithostatic pressures between 0.8 and 1.3 kb are estimated. Based on the pressure–temperature curves of Potter (1977), the low- to moderate-salinity stage IVb veins require a temperature correction of between 60 and 130°C to the homogenization temperature data (Th_{TOT(L)} average = 210°C), giving a trapping temperature range of 270 to 340°C. A trapping temperature of ≈300°C for stage IV veins seems reasonable based on: 1) the observed alteration assemblage of sericite, ankerite, quartz, and pyrite; and 2) similar temperatures reported from other shear-zone-hosted gold deposits that have similar alteration assemblages (e.g., the Archean deposits of W.A., Mt. Charlotte, Golden Mile-Lakeview and North Kalgurli, and Ora Banda; Ho et al. 1990).

Sulfur isotopes

A total of 62 $\delta^{34}\text{S}$ analyses were obtained from pyrite using analytical facilities at the Central Science Laboratory, University of Tasmania. Six analyses were obtained using conventional (hand-drilling) techniques (Robinson and Kusakabe 1975). The rest were obtained by laser ablation using the method described by Huston et al. (1995). Three generations of pyrite were selected for $\delta^{34}\text{S}$ analyses: (1) disseminated diagenetic pyrite from outside the mineralized zones (one sample; seven ablation and one conventional analyses); (2) euhedral pyrite from a stage II quartz vein in

a pre-mineralization thrust fault (one sample; seven ablation and one conventional analyses); and (3) stage IV disseminated and vein-hosted pyrite from mineralized alteration zones in basalts (three samples, 11 ablation analyses) and sedimentary lithologies (seven samples; 24 ablation and four conventional analyses). Results are plotted on Fig. 15, as $\delta^{34}\text{S}\%$ relative to the Canyon Diablo Troilite (CDT). Details of the analytical procedures and all analytical data are provided in Tunks (1996).

$\delta^{34}\text{S}$ values of diagenetic pyrite from the host sequence have a narrow range (17.8 to 19.2‰; Fig. 15). The source of sulfur is inferred to have been seawater sulfate that was reduced during early diagenesis by sulfur-reducing bacteria (Tunks 1996). Stage II pyrite has $\delta^{34}\text{S}$ values between 31.7 and 37.1‰ (Fig. 15), which are interpreted to be a product of complete in situ reduction of an isotopically heavy sulfur source (evaporitic?). This interpretation is based on the fact that the stage II pyrite is completely enclosed by vuggy vein quartz, thus discounting the possibility of sulfate-reducing bacteria.

Stage IV pyrite hosted by quartz veins and related alteration halos have a limited range of $\delta^{34}\text{S}$ values (6.8 to 12.5‰; average=9.3‰), which are lower than the diagenetic and stage II pyrites (Fig. 15). The narrow range in $\delta^{34}\text{S}$ values is consistent either with the oxidation state of the stage IV mineralizing fluids being below the $\text{SO}_4^{2-}/\text{H}_2\text{S}$ buffer or with the $\text{SO}_4^{2-}/\text{H}_2\text{S}$ ratio remaining constant during the mineralizing processes (Ohmoto and Rye 1979; Sheppard 1986). The lower $\delta^{34}\text{S}$ values from stage IV pyrite hosted by sedimentary lithologies compared to basalts (Fig. 15) could be related to partial oxidation of sulfur in the mineralizing fluid via interaction with the locally hematite-bearing sedimentary rocks (Tunks 1996).

The lack of regional sulfur isotope data limits the conclusions that can be drawn concerning the ultimate sources of sulfur for stage IV mineralization. The restricted range of data is consistent with stage IV sulfur being derived primarily from a homogenized external fluid rather

than local sedimentary sources. Furthermore, the total reduced sulfur concentrations in the fluid is inferred to have been much greater than that derived by leaching of the local sedimentary host rocks, given the lower $\delta^{34}\text{S}$ values of stage IV pyrite compared to diagenetic pyrite (Fig. 15). It is possible that mixing, at depth, of a magmatically sourced fluid with a $\delta^{34}\text{S}$ of close to 0‰ with a sedimentary sulfur source could have generated a hybrid mineralizing fluid with a $\delta^{34}\text{S}$ value of approximately 12‰.

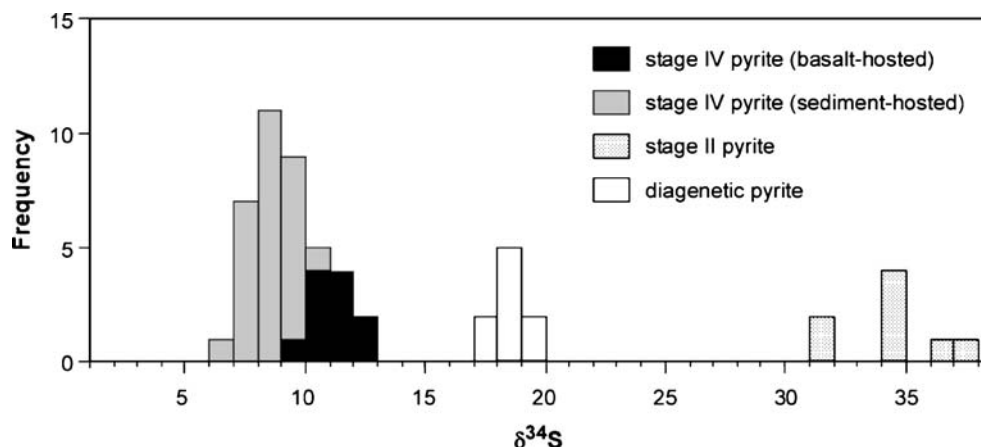
Discussion and conclusions

The Paleoproterozoic host rocks to gold mineralization in the Tanami district were deposited in a below-wave-base subaqueous setting. The combination of bimodal mafic/felsic volcanism and hematite-bearing clastic sedimentary rocks are consistent with an intra-continental rift setting at the time of deposition.

Interference between twofold generations has produced kilometer-scale basin and dome structures to the north and south of the mine lease. Thrust faults that are interpreted to have formed during F_1 flexural slip folding have a ramp and flat geometry with respect to bedding.

A pervasive chlorite–carbonate alteration assemblage is associated with stage III chlorite–carbonate veins. Because no workable fluid inclusions were found in stage III veins, the composition of the stage III fluid is unknown. Stage III veins do not appear to be spatially related to the major mineralized fault zones, and a systematic compositional and geochemical zoning from the SQP alteration out through SQ alteration to stage III chlorite–carbonate alteration could not be established. Stage III chlorite–carbonate alteration is mineralogically similar to weak regional carbonate metasomatism within the Golden Mile Dolerite, which has been interpreted to be an outermost halo of alteration associated with the Au mineralizing event at the Golden Mile (Phillips 1986).

Fig. 15 Results of sulfur isotope analyses of pyrite (expressed as $\delta^{34}\text{S}$ CDT) from the Tanami district. Data tabulated in Tunks (1996)



Stage IV faults and veins are well constrained within the structural history of the Tanami district. They crosscut the early extensional microfaults and the bedding parallel thrust faults. The mineralization has been interpreted to occur after the formation of F_1 folds because: 1) the mineralized structures crosscut the bedding parallel thrusts that formed during F_1 , and 2) alteration halos are preferentially developed in the present-day hanging wall of the faults. The positioning of the alteration halo in the hanging wall of a fault plane is to be expected as the density of the mineralizing fluids would have been less than half the density of the wall rocks, and this would cause the fluid to migrate upwards (Phillips 1972).

This would not be the case if the mineralized faults had been rotated passively into their current position by post-mineralization folding. The stage IV mineralized fault zones have mutually crosscutting relationships with felsic dykes intruded into the mine sequence, which are interpreted to indicate that the dykes and stage IV mineralization formed approximately synchronous. Dyke emplacement and mineralization is inferred to have taken place between 1,815 and 1,799 Ma, the time of emplacement of granitoids in the nearby Frankenia and Coomarie domes (Smith 2001; Cross et al. 2003; Wygralak et al. 2004), most likely towards the end of that cycle of intrusive activity (Huston et al. 2006).

Stress inversion techniques have been used to calculate the paleo-stress field for the sinistral wrench system that hosts Au mineralization. σ_1 was subhorizontal and directed northwest–southeast, with σ_3 oriented subvertical. This is in contrast with the stress field interpreted for the formation of F_1 folds and thrusts, which developed under a similarly directed σ_1 , but with σ_2 rather than σ_3 subvertical. Importantly, a change in stress conditions from σ_2 to σ_3 subvertical allows for the development of subvertical faults, which may substantially alter fluid migration patterns within the sequence (e.g., Cox 1991).

The internal textures observed within stage IV veins and breccias are indicative of multiple reactivation events. Such continual reactivation of fault zones allows for the regeneration of permeability within the faults and veins, and the potential for fluid–wall rock interaction as well as fluid mixing at various times within the failure cycle.

Alteration halos around stage IV structures are zoned from an inner sericite–quartz–pyrite $\pm$ ankerite (SQP) assemblage to an outer sericite–quartz–ankerite (SQ) assemblage. All high-grade, hypogene gold mineralization in the Tanami district occurs within the stage IV veins and inner alteration halos. Petrographic studies have indicated that gold occurs as inclusions in pyrite and chalcopyrite, and as coarse free-gold in quartz–carbonate veins. The outer SQ alteration is mineralized, with sub-economic gold grades typically up to one order of magnitude higher than unaltered wall rocks, which have Au grades of 1 to 20 ppb

(Tunks 1996). Gold grades in the veins and SQP alteration halos are up to four orders of magnitude higher than the unaltered wall rocks.

High-salinity (~20 wt% NaCl equivalent), low-temperature (~120–150°C) Ca-rich basinal brines were associated with the formation of stage II and stage V veins (Fig. 14). The elevated $\delta^{34}\text{S}$ values of stage II pyrite (31.7 and 37.1‰) are indicative of a sedimentary (possible evaporitic) sulfur source. Stages II and V brines contrasted markedly with the stage IV mineralizing solutions (Fig. 14), which had moderate temperatures (~300°C), low salinities, low CO_2 contents, were rich in Na and K, and which precipitated pyrite with much lower $\delta^{34}\text{S}$ values than stage II veins (Fig. 15). Mineralization during stage IV is inferred to have occurred at depths of 3–6 km under lithostatic load from mineralizing fluids that contained a homogenized mixture of magmatic and sedimentary sulfur.

Huston et al. (2006) advocate a continuum-style lode gold model for the Tanami region gold deposits. The Granites goldfield and the Goldrush deposit are inferred to have formed at depths of up to 11 km from CO_2 - and CH_4 -rich reducing fluids, based on the fluid inclusion data of Mernagh and Wygralak 2006. The gold-bearing veins and shear zones of the Tanami goldfield veins are inferred to represent the shallowest end of the lode gold continuum, due to their low CO_2 contents and estimated depths of formation (Huston et al. 2006). Although a lode (orogenic) gold model is advocated, the deposits are also noted to be related to the waning phases of granitic magmatism in the region, with deposit orientations varying systematically with respect to the Coomarie and Frankenia granite complexes (Huston et al. 2006). SHRIMP age determinations of hydrothermal xenotime from Callie and Coyote (~1,803 $\pm$ 19 and 1,791 $\pm$ 8 Ma, respectively; Cross et al. 2005; Bagas et al. 2006) overlap with the age of magmatism in the Frankenia Dome (~1,815 to ~1,799 Ma), supporting a magmatic relationship. Pb isotopic systematics of several deposits from the Tanami region (e.g., Groundrush, Coyote and Pebbles) are consistent with at least some Pb being derived from granitic sources (Huston et al. 2006).

The gold deposits of MLS153 have some characteristics that are compatible with an orogenic (lode) gold model, including an intimate relationship between mineralization and deformation, and vein and alteration mineral assemblages that are similar to Archean and Paleozoic lode gold systems (e.g., Groves et al. 1998; Hagemann and Cassidy 2000; Bierlein and Crowe 2000). However, we believe that other aspects of the geology of Tanami district (low CO_2 content of the fluid inclusions, shallow depth of formation, intimate cross-cutting relationships between gold-bearing shear zones and felsic dykes, elevated $\delta^{34}\text{S}$ values of ore-stage pyrite, distinctive low-temperature basinal brines in

pre- and post-mineralization veins hosted by thrust faults) illustrate that a metamorphic fluid and metal source was unimportant to mineralization and that magmatism was the critical element in triggering gold formation in the district. We, therefore, interpret the gold veins of MLS153 to be examples of “thermal aureole gold” (TAG) mineralization, based on the genetic model of Wall (1989, 2005). The TAG model describes a class of deposits, which form in the inner and outer thermal aureole of granite plutons, and can be considered a subset of the family of gold deposits classified as “intrusion-related gold” (e.g., Thompson and Newberry 2000; Lang and Baker 2001).

TAG deposits are inferred to have formed during dilatant deformation synchronous with granite emplacement and characteristically occur as en echelon, stockwork vein systems in the roof zones above plutons. Base metal contents of thermal aureole deposits are generally low, and gold is commonly associated with pyrite, pyrrhotite, or arsenopyrite. Australian gold deposits cited by Wall (1989, 2005) as possible examples of thermal aureole deposits include the Paleozoic Golden Mountain, Wonga and Nagambie deposits in Victoria and Proterozoic deposits of the Cullen Mineral Field in the Pine Creek Inlier, such as Cosmo Howley and Goodall. The Mt. Todd Gold Mine in the Pine Creek Inlier has also been classified as a thermal aureole gold deposit (Hein 1994, 2003). The Tanami district gold veins are best classified as outer thermal aureole systems because of the lack of obvious contact metamorphism evident within the mine sequence.

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