The structure and geochemistry of gold mineralisation in the Mt Todd goldfield, Pine Creek Inlier, Northern Territory

by

Kim A.A. Hein B.Sc. (Hons)

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"Strange as it may seem, I cried the day I left...
We had been living at Mt Todd for close on twelve months, and I had made it home".

(From "No place for a woman: the autobiography of outback publican, Mayse Young", by Young and Dalton, 1991.)
STATEMENT

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ABSTRACT

The Mt Todd goldfield is located approximately 40 km northwest of the township of Katherine, in the Northern Territory, Australia. The goldfield is host to several discrete ore bodies that strike north-northeast within a broad northeasterly trending corridor of gold mineralisation. The largest ore body occurs at the Batman mine and constitutes an identified mineral resource of 92.9 million tonnes at 1.4 g/t Au.

The goldfield lies in the southern region of the Central Domain of the Pine Creek Inlier and is hosted by a Palaeoproterozoic sequence of rocks termed the Burrell Creek Formation. The formation is dominated by greywackes, siltstones, sandstones and shales that exhibit sedimentary features akin to those of a river-dominant delta front to pro-delta environment. The formation is conformably overlain by volcanoclastic and volcanolithic sediments of the Tollis Formation (~1890 Ma).

Three deformation events are recognised, viz., D₁, D₂ and D₃. The earliest deformation, D₁, is characterised by close to tight, northeast to northerly to northwesterly trending asymmetric folds (F₁), a continuous axial planar cleavage (S₁), and numerous strike-slip faults. The deformation is associated with the development of buck quartz veins and was preceded by the emplacement of the Yenberrie Leucogranite of the Cullen Batholith (1835-1820 Ma) which contact metamorphosed the sediments of the Burrell Creek Formation to hornblende-hornfels facies (H₁), with the development of cordierite porphyroblasts (type C₁). D₁ was coincident with peak regional metamorphism to greenschist facies.

D₂ is associated with westerly trending open folds (F₂), a spaced disjunctive to fracture cleavage (S₂) in transection to the folds, and strike-slip and normal faults. The deformation is associated with the development of quartz-tourmaline and gold-bearing quartz-sulphide veins and lodes. It was preceded by the emplacement of the Tennysons Leucogranite which contact metamorphosed the Yenberrie Leucogranite of the Cullen Batholith and the sediments of the Burrell Creek Formation to hornblende-hornfels facies (H₂), with the development of cordierite porphyroblasts (type C₂).

D₃ is characterised by the reactivation of strike-slip faults (mostly sinistral), a steeply dipping Type S₃-C type foliation, and mesoscopic en échelon folds (F₃) that trend oblique to the faults in a left stepping (sinistral) array. The faults offset D₁ and D₂ structures. Calcite-base metal veins and epithermal style quartz veins may have formed during this deformation.
The age of emplacement of the leucogranites, $D_1$ and $D_2$ is constrained by the age of emplacement of the Cullen Batholith at 1835-1820 Ma. $D_1$ and $D_2$ are correlated with deformation during the Maud Creek Event (~1850 Ma), while $D_3$ is correlated with deformation during the Shoobridge Event (~1780 Ma).

Mineralisation in the goldfield is associated with quartz-sulphide veins and lodes that comprise pyrrhotite, chalcopyrite, arsenopyrite, pyrite, marcasite, löellingite, with minor galena, sphalerite, bismuth-sulphosalts, ISS CuFeS minerals, and gold. Gold is associated with bismuth and CH$_4$-rich hypersaline fluid inclusions (29-33 wt% NaCl eq) in trails which cross-cut early quartz and sulphide. The general vein/lode alteration assemblage comprises quartz, biotite, muscovite, chlorite, sericite, rutile, accessory tourmaline, chalcopyrite, pyrrhotite, pyrite or marcasite.

The quartz-sulphide veins and lodes are typified by extension concomitant to hydrothermal fracturing and/or normal faulting. Across the goldfield, the veins and lodes exhibit a remarkable similarity with respect to strike geometry, morphology, vein forming process and isotopic composition. They are located within the thermal aureole of the Tennysons Leucogranite and strike north-northeast within a corridor that trends northeasterly above a basement strike-slip fault. They are preferentially hosted in competent rock types such as greywackes and siltstones. Detailed structural and petrographic analyses has indicated that the veins and lodes were generated at the same time by hydrothermal activity during retrograde (contact) metamorphism associated with cooling of the Tennysons Leucogranite, and early in $D_2$, prior to the development of the regional $S_2$ fabric.

Integrated petrologic, isotopic (sulphur and oxygen) and fluid inclusion studies suggest that the hydrothermal fluids were chiefly magmatic-metamorphic in origin. The fluids were enriched in metal and sulphur derived either as magmatic metal or sulphur (from the Tennysons Leucogranite), or scavenged during fluid-rock reactions from sources in the sedimentary pile, the Yenberrie Leucogranite, or from sulphide-bearing greisens in the Yenberrie Leucogranite. Fluid evolution was controlled by retrograde reactions which accompanied rehydration of the country rock and Tennysons Leucogranite, at low fluid-rock ratios.

The hydrothermal fluid was reduced in nature and exhibited a tendency to lower $fO_2$ and $fS_2$ concomitant to an overall decrease in temperature and salinity for a given pH. $\delta^{34}S$ values of 4.5% to 13.3% indicate that the source for sulphur was mixed (magmatic-sedimentary).

Initial precipitation of crack-seal fibre quartz from a silica-saturated brine occurred during early retrograde metamorphism by progressive fracturing and sealing of the host rocks. The
subsequent precipitation of quartz and tourmaline coincided with a significant vein dilation event and accompanied decompressive effervescence and boiling of a surface-derived hydrothermal fluid at a minimum pressure of 180-225 bars or a minimum depth of 1.8 to 2.6 kilometres. Fluid salinity ranged from 1 to 20 wt% NaCl eq: this fluid interacted with a mixed magmatic-metamorphic fluid.

The precipitation of a silicate-sulphide-carbonate assemblage accompanied an influx of hot (~400°C), acidic (pH ~ 4.0 at 330°C), hypersaline brine (30-50% NaCl-CaCl₂). This metal and sulphur enriched brine was mixed magmatic-metamorphic in origin. The principal cause of metal deposition was fluid unmixing at a shallow crustal level (minimum depth of 1.8 to 2.6 km) during a decrease in temperature from approximately 370°C to 240°C, a decrease in salinity from 30-50% NaCl-CaCl₂ to 29-33 wt% NaCl eq, and vertical transport of the brine along joints, fractures and faults. Gold precipitation occurred late in the development of the silicate-sulphide-carbonate assemblage from a hot (~250°C), acidic (pH of 2.3-4.9 at 250°C), hypersaline brine (29-33 wt% NaCl eq), and accompanied the precipitation of bismuth and the entrapment of CH₄-rich fluid inclusions.

A genetic model for mineralisation in the goldfield invokes sinistral reactivation of a north-easterly trending basement strike-slip fault under the influence of effective tensile stresses caused by cooling of the Tennysons Leucogranite. The reactivation of the fault caused brittle failure in the upper crust and/or dilation of existing north-northeasterly trending faults, fractures and joints in competent rock types. The generation of dilatant structures, coupled with a sudden reduction in pressure facilitated channelization of fluid flow into the upper crust, probably away from convection cells juxtaposed about the cooling leucogranite. Rising fluids decompressed causing phase separation with mineral precipitation. Throttling of the conduit or fluid pathways resulted in over-pressuring of the fluid, this giving way to hydrothermal fracturing and an enhanced permeability.
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Figure 3.11: Stratigraphic column for the Central Domain of the Pine Creek Inlier, as adopted in this study. Modified after Jagodzinski (1992) and Stuart-Smith (1987).

Figure 3.12a: Lithological interpretation - 1:10,000 scale map with stratigraphic column of the Burrell Creek Formation.

Figure 3.12b: Lithological interpretation (supplementary) - 1:10,000 scale map with stratigraphic column of the Burrell Creek Formation and the Tollis Formation.

Figure 3.13: The Yenberrie Leucogranite.

Figure 3.14: Meta-sedimentary xenolith in the Yenberrie Leucogranite (Yenberrie Field MG 11900N 5450E).

Figure 3.15: Spaced, stylolitic to anastomosing cleavage in the Yenberrie Leucogranite. Pencil is 12 cm long (Yenberrie Field MG 11150N 5800E).

Figure 3.16: The Tennysons Leucogranite.

Figure 3.17: Meta-igneous xenoliths in the Tennysons Leucogranite. The sample was collected along the Stuart Highway, 6 kilometres north of the junction with the Edith Falls Road.

Figure 3.18: Porphyritic rhyolite dyke (Batman and Robin MG 9440N 8700E).

Chapter 4

Figure 4.1: 1:5000 scale fact map of Horseshoe Creek.

Figure 4.2: 1:5000 scale fact map of Horseshoe Creek Tin Field.

Figure 4.3: 1:5000 scale fact map of Yenberrie Field.

Figure 4.4: 1:5000 scale fact map of North Batman.

Figure 4.5: 1:5000 scale fact map of Quigleys.

Figure 4.6: 1:5000 scale fact map of West Batman.

Figure 4.7: 1:5000 scale fact map of Batman and Robin.

Figure 4.8: 1:5000 scale fact map of Mount Todd.

Figure 4.9: 1:5000 scale fact map of Edith River Crossing.

Figure 4.10: 1:5000 scale fact map of Stow Creek.
Figure 4.11: A structural interpretation of the Mt Todd goldfield - 1:15,000 scale map.

Figure 4.12a: Regional structural map of the Mt Todd goldfield - 1:50,000 scale map.

Figure 4.12b: Schematic section across the Mt Todd goldfield (from the Tennysons Leucogranite to the Dick Grayson prospect, Figure 1.3) along mine grid northing 11000N showing the gross relationship of the main tectonic elements. The key to stratigraphic units shown is equivalent to that given in Figure 3.12b, Map - Lithological Interpretation (supplementary).

Figure 4.13: (a) Equal area stereographic projection of total S1 data. (b), (c), (d) and (e) consider the S2 data in domains of equal strike. The plots demonstrate the helical trace of S1 from the North domain through to the Access Road domain in the south of the goldfield.

Figure 4.14a, b, c, d: Equal area stereographic projection of the plunge of F1, as determined from the intersection of bedding and S1. The circled points in (a) correspond with fold hinges that have been rotated toward the southeast adjacent to northwest strike-slip faults.

Figure 4.15: Equal area stereographic projection of the poles to S2 from the Access Road domain. S2 lies in transection to F2 such that the total set of data in (a) consists of northwesterly trending S2 on the north and south limb of the fold, plots (b) and (d), and westerly trending S2 on the hinge of the fold, plot (c). The dip of the cleavage at the hinge is steeper than on the limbs.

Figure 4.16: Equal area stereographic projection and contour plot of poles to S2 planes.

Figure 4.17: Sinistral shear faulting in the Yenberrie Shear Zone (Batman and Robin MG 9400N 7500E). The shear faults are weathered to a buff-red colour and are associated with a mesoscopic S-C fabric (S3). This fabric is characterised by the development of jagged, lenticular slabs of rock that are positioned oblique to, and along the strike of the fault zone. (The camera lens cap is 5 cm in diameter.)

Figure 4.18: Buck quartz vein - a large single blow (Quigleys MG 13150N 11550E).

Figure 4.19: A photograph of a buck quartz vein in a road cut along the Mine Access Road (Stow Creek MG 6850N 11350E). The vein is lenticular and undulating in form and typical of buck quartz vein morphologies throughout the Mt Todd goldfield. Sinistral slip is indicated from offset on bedding. This vein is hosted by greywacke and volcanoclastic beds at the contact of the Burrell Creek Formation with the Tollis Formation (the greywacke beds are greyish in colour in this photo while the volcanoclastic beds are distinctly chocolate brown).

Figure 4.20: A. Block diagram of conjugate buck quartz veins. B. Equal area stereographic projection of poles to veins of buck quartz in the North domain with mean vein orientations shown as great circles. C. Equal area stereographic projection of poles to veins of buck quartz in the Access Road domain with the mean vein orientations shown as great circles.

Figure 4.21a: Equal area stereographic projection of poles to quartz-tourmaline veins of the Mt Todd goldfield.

Figure 4.21b: Photomicrograph of green to brown coloured tourmaline fibres in a quartz-tourmaline vein from Mount Todd. The vein is 4 mm wide and is cross-cut by a vein of calcite (Vq). Quartz is syntaxial to the wall rock and tourmaline masses are scattered throughout the wall rock assemblage (right). The opaques at top right are pyrrhotite, chalcopyrite, and goethite after pyrrhotite and chalcopyrite. (Thin section MT013.)

Figure 4.22a: Equal area stereographic projection of poles to calcite-base metal veins in the Mt Todd goldfield.

Figure 4.22b: Photomicrograph of colloform pyrite and marcasite (py - marc) lining the walls of a 2 mm wide calcite-base metal vein. Galena (gn) and calcite (cal) occur as fill between the colloform bands while chalcopyrite (ccp) and sphalerite (sph) form a discontinuous base to the bands. DDH BD087(1), 297.78-297.94 m.
Figure 4.23: a. Reflected light photomicrograph of herringbone texture in galena (gn) which is surrounded by sphalerite (sph) and rhombs of calcite (cal). Scale bar is 2 mm long.

Figure 4.23: b. Transmitted light photomicrograph of Figure 4.23a. Galena is surrounded by Fe-bearing sphalerite (honey-brown coloured) and Fe-poor sphalerite (white). Thin section BD008a (1-3) collected from DDH BD008, 98-99 m. Scale bar is 2 mm long.

Figure 4.24: Photomicrograph of an epithermal style quartz vein (V5). Densely packed layers of reniform or chalcedonic quartz forms a substrate to fine euhedral crystals of quartz (0.1 to 0.5 mm). The texture is interpreted as replacement of epithermal calcite by quartz. Scale bar is 1.0 mm long. (Thin section DC001, Batman and Robin MG 10830N 9940E.)

Figure 4.25: Epithermal style quartz vein (V5). Loosely packed layers of reniform quartz are arranged into a rhombohedron and Fe-oxides occur as fill between layers. The texture is interpreted as replacement of epithermal calcite by quartz and Fe-oxide. (J001, Batman North MG 12150N 9275E.)

Figure 4.26: Equal area stereographic projection of quartz slickenfibres on the footwall to gold-bearing quartz-sulphide lodes at the Quigleys deposit, Mt Todd goldfield.

Figure 4.27: A. Foliation in the Yenberrie Leucogranite (Yenberrie Field MG 11600N 4725E). B. Equal area stereographic projection of S1 in the Yenberrie Leucogranite. C. Equal area stereographic projection demonstrating the relationship of the mean S1 and S2 to the foliation in the leucogranite: the foliation correlates well with the regional S1 but is oblique to the regional S2.

Figure 4.28: A photograph of a shear fault cross-cutting the body of the Tennysons Leucogranite, 4 km southwest of Yinberrie Hill. The shear faults are associated with a narrow chloritised or silicified selvedge: thin section analyses of orientated blocks of the selvedge material has indicated that the shear faults are narrow zones of intense shear dislocation and flattening. View is towards the southeast. (Thank you D. Bremner.)

Figure 4.29: A photograph of a sheared and strongly fractured sample of the Tennysons Leucogranite. (Sample OR001, collected 3 km southwest of Yinberrie Hill.)

Figure 4.30: Map of the contact aureole zones of the Mt Todd goldfield - 1:50,000 scale.

Figure 4.31: A tracing of a photomicrograph of retrogressed cordierite spots (C1 and C2) in a laminated shale. C1 spots are completely retrogressed to sericite and chlorite and these minerals have adopted a preferred orientation (S1). C1 is elongate within S1 and is cross-cut by C2. The texture is indicative of a contact metamorphic event pre- to syn-S1. (Thin section GN001, Batman North MG 12550N 9900E.)

Figure 4.32: A sketch of a photomicrograph of a mesoscopic F2 fold. The cordierite spots are completely retrogressed to sericite and chlorite and these minerals have adopted a preferred orientation (S1). The spots are elongate within S1 and both are folded about the fold hinge. The texture is indicative of a contact metamorphic event pre- to syn-S1. (Thin section A005, Horseshoe Creek MG 13675N 9825E.)

Figure 4.33: The approximate P-T conditions for regional and contact metamorphism in the Mt Todd goldfield. Data from Ferguson et al. (1980) and Pattison and Tracy (1991).

Figure 4.34: The relationship of structural and metamorphic events of the Mt Todd goldfield.

Chapter 5

Figure 5.1: Geology of the Quigleys pits - 1:1000 scale.

Figure 5.2: Fact geology over a cleared area on Batman Hill - 1:50 scale.
Figure 5.3: Fact geology recorded on a sketch of the south-western face of the Quigleys south pit.

Figure 5.4: Equal area stereographic projection data from the Batman deposit. A. Bedding. B. Quartz-sulphide veins. C. Quartz fibres (crack-seal). D. Quartz slickenfibres.

Figure 5.5: Mount Todd area rainbow psuedocoloured aeromagnetics with regional structural geology overlay.

Figure 5.6: Sheeted quartz-sulphide veins of the Batman deposit. View is toward the south. (Batman and Robin MG 10375N 8505E.)

Figure 5.7: Photomicrograph of quartz-fibre fill. The straight fibres are syntaxial to quartz clasts in the wall rock (bottom right). The central fibre (white) is 2.6 mm long. (Thin section BD008c collected from DDH BD008, 90.9 m.)

Figure 5.8: Geometries associated with the main quartz-sulphide vein set in the Batman deposit. An approximately 3 metre normal displacement (down-throw to the east) is indicated for an orebody width of 150 metres.

Figure 5.9a: Photograph of the alteration selvedge surrounding the veins in the Batman deposit. View is toward the south. (Batman and Robin MG 10375N 8505E.)

Figure 5.9b: Sketch of Figure 5.9a. The chlorite/sericite alteration selvedge is uneven in distribution around or along the vein, and it may be partially or completely absent.

Figure 5.10: Sketch of the vein sets in the Batman deposit.

Figure 5.11:
A. $S_2$ fracture cleavage cross-cutting a quartz-sulphide vein in the Batman deposit. The vein was painted red at some stage prior to photography. Camera lens cap is 5 cm in diameter. (Batman and Robin MG 10000N 8450E.)
B. The $S_2$ fracture cleavage has caused blocking of quartz sulphide veins into a mosaic of rhombic shapes (Batman and Robin MG 10375N 8500E).

Figure 5.12: Tracing of a photomicrograph demonstrating the continuity between the fracture cleavage and the wall rock cleavage in the Batman deposit (Thin section BD007c collected from DDH BD007, 119.0 m.)

Figure 5.13: Equal area stereographic projection of the joint geometries in the Batman deposit. The overlap of points for set 1 and 2 is the result of a slight rotation in joint strike across the mapped area.

Figure 5.14:
A. Equal area stereographic projection of the orientation of the Batman orebody. B. Diagrammatic representation of the geometry associated with the orebody and the main quartz-sulphide vein set of the Batman deposit.

Figure 5.15: Structural interpretation of the Quigleys deposit: 1:5000 scale map.

Figure 5.16: Parallel set of buck quartz veins in the footwall of Quigleys 10550 pit. The largest is 12 metres long. (Thank you D. Wegmann.)

Figure 5.17: Quartz-sulphide lode of Quigleys 10550 pit. The lode is composed of intensely micro-fractured and brecciated boulders and clasts of quartz, in association with vuggy to massive red-brown-blue Fe-oxide.

Figure 5.18a: Photomicrograph of lode material from DDH QD008, 120 m: cataclasite of quartz (black and pink) and arsenopyrite (white). Scale bar = 2 mm.

Figure 5.18b: Photomicrograph of lode material from DDH QD008, 120 m: cataclasite of quartz (black), pyrite (yellow) and arsenopyrite (white). Scale bar = 2 mm.
Figure 5.19a, b: Complex zone of stockwork veining in the footwall to the lode of the Quigleys 10500 pit comprising a coarse in-situ breccia in which fine veins are connected via larger veins to the main lode. S1 in the wall rock (centre left) is cross-cut by quartz-sulphide stockwork veins. Late calcite base metal veins (sub-horizontal) cross-cut quartz-sulphide stockwork veins (sub-vertical). View is toward the north.

Figure 5.20: Equal area stereographic projection of slickenfibres at the Quigleys deposit. A. Reverse displacement is west block up. B. Normal displacement is west block down.

Chapter 6

Figure 6.1: Photomicrograph of fibrous pyrrhotite (polars half crossed). The fine deformation lamellae at bottom right indicate that the pyrrhotite has been deformed subsequent to its precipitation (DDH BD079, 264 m). Scale bar is 500 \( \mu \)m long.

Figure 6.2: Photomicrograph of quartz 2 (qtz 2) against a crack-seal fibre termination (qtz 1). The crack-seal fibre is highly decorated with 1-3 \( \mu \)m fluid inclusions. Opaque is pyrrhotite. DDH BD080(3), 286.95 m. Scale bar is 50 \( \mu \)m long.

Figure 6.3: Photomicrograph of 100 \( \mu \)m wide selvedge of qtz 2 between crack-seal quartz (qtz 1) which contains numerous fluid inclusions, and pitted pyrrhotite 3 (po 3). Transmitted and reflected light are used simultaneously in this view. DDH BD086, 143.0 m. Scale bar is 200 \( \mu \)m long.

Figure 6.4: Photomicrograph of euhedral quartz (qtz 2) against a crack-seal fibre termination (qtz 1). The crack-seal fibre is highly decorated with 1-3 \( \mu \)m fluid inclusions. Opaque is pyrrhotite. DDH BD080(3), 286.95 m. Scale bar is 50 \( \mu \)m long.

Figure 6.5: Photomicrograph of euhedral quartz terminations (qtz 2). The terminations exhibit growth zones containing tourmaline and liquid-vapour fluid inclusions (arrows). The opaque is pyrrhotite. DDH BD080(2), 286.95 m. Scale bar is 500 \( \mu \)m long.

Figure 6.6: Photomicrograph of growth zone in quartz 2 (qtz 2). Tourmaline (tour) and liquid-vapour fluid inclusions (arrows) occur in the same zone. DDH BD086, 143.0 m. Scale bar is 40 \( \mu \)m long.

Figure 6.7: Photomicrograph of co-precipitate minerals cubanite (cub), pyrrhotite (po 1) and talnakhite 1 (tal 1) as an inclusion in arsenopyrite 1 (asp 1). DDH BD079, 264 m. Scale bar is 25 \( \mu \)m long.

Figure 6.8: Photomicrograph of a fractured arsenopyrite euhedra (asp 1) enclosed by pyrrhotite 3 (po 3) and chalcopyrite 1 (ccp 1). The arsenopyrite is host to numerous inclusions of cubanite, pyrrhotite 1 and chalcopyrite 1 (DDH BD079, 264.0 m). Scale bar is 500\( \mu \)m long.

Figure 6.9: Phase relations in the Cu-Fe-S system at 350°C, modified after Sugaki et al. (1975). Electron microprobe analyses as follows: \( \odot \) = Cu-Fe-S minerals in the association pyrrhotite 1 and cubanite, plus talnakhite 1; \( X \) = chalcopyrite in equilibrium with pyrrhotite 3; \( \bullet \) = talnakhite 2 in equilibrium with bismuthinite. Cubanite and talnakhite approach solid solution with moolhoekite and haycockite, respectively.

Figure 6.10: Photomicrograph of pyrrhotite 2 (po 2) in pitted pyrrhotite 3 (po 3). The texture suggests that po 3-qtz replaced po 2. DDH BD027, 172.2 m. Scale bar is 0.5 mm long.

Figure 6.11: Photomicrograph of pyrrhotite 3 (po 3)-qtz replacing pyrrhotite 2 (po 2). Fractured monoclinic tablets of quartz (centre right) probably pseudomorph arsenopyrite 1. DDH BD027, 172.2 m. Scale bar is 500 \( \mu \)m long.

Figure 6.12: Photomicrograph of pyrrhotite 3 (po 3) and chalcopyrite 1 (ccp 1). The smooth embayed margins are indicative of co-precipitation of the two sulphides. DDH BD076, 473.0 m). Scale bar is 500 \( \mu \)m long.
Figure 6.13: Photomicrograph of euhedral pyrrhotite 5 (po 5) in intimate association with gold (Au), bismuthinite (bis), talnakhite 2 (tal 2), and bismuth (Bi). DDH BD080(3), 286.95 m. Scale bar is 50 µm long.

Figure 6.14: Photomicrograph of pyrrhotite 5 (po 5) intimately associated with talnakhite 2 (tal 2), bismuth (Bi) and bismuthinite (bis). DDH BD079, 264 m. Scale bar is 50 µm long.

Figure 6.15: Photomicrograph of pyrite 1 core (py 1) in pitted pyrite 2 (py 2). The pits in PY 2 are quartz filled inclusions and the white mineral is arsenopyrite (asp). Py 2-qtz also occurs as fill to fractures which cross-cut the py 1 core. The texture suggests that py 2-qtz replaced py 1. DDH BD008, 90.9 m. Scale bar is 100 µm long.

Figure 6.16: Photomicrograph: delicate fringe of pyrite 2 (py 2) and marcasite 1 (mare 1) around pyrite 1 (py 1). The marcasite is difficult to identify but is slightly whitish in colour relative to yellowish coloured pyrite. DDH BD008, 90.9 m. Scale bar is 0.5 mm.

Figure 6.17: Photomicrograph of a 10 µm wide colloform band of pyrite 2 (py 2) and marcasite 1 (marc 1) as an overgrowth on subhedral grains of pyrite 1 (py 1). Euhedra of marcasite 2 lie to the left of the colloform band and are interspersed with pyrite 3 (py 3). A single crystal of arsenopyrite 4 (asp 4) occurs at bottom left. DDH BD008, 90.9 m. Scale bar is 20 µm.

Figure 6.18: Photomicrograph of marcasite 2 (mare 2) with pyrite 3 (py 3) as interstitial fill. DDH BD023, 41.9 m. Scale bar is 0.5 mm.

Figure 6.19: Photomicrograph of löellingite (loell) in pyrite 1 (py 1). The polars are partly crossed to accentuate the strong anisotropy of the löellingite. DDH BD021, 117.5 m. Scale bar is 200 µm.

Figure 6.20: Photomicrograph of inclusions of löellingite (loell) and pyrrhotite 3-chalcopyrite 1 in arsenopyrite 2 (asp 2). The arsenopyrite is fractured and enclosed by sphalerite (sph). The texture suggests that arsenopyrite precipitation succeeded that of löellingite and po 3-ccp 1, but preceded a fracturing event and the precipitation of sphalerite. DDH BD018, 88.1 m. Scale bar is 400 µm long.

Figure 6.21: Photomicrograph of löellingite (loell) in arsenopyrite 2 (asp 2) and enclosed by arsenopyrite 3 (asp 3). These are in turn enclosed by pyrrhotite 3 (po 3). This is cross-cut by a fracture containing pyrrhotite 4 (po 4). The pinkish speckled mineral in the fracture cross-cutting the löellingite and arsenopyrite is bismuth (Bi). The texture suggests that the precipitation of löellingite was succeeded by that of arsenopyrite 2, arsenopyrite 3 and pyrrhotite 3. A fracturing event preceded the precipitation of pyrrhotite 4 and bismuth, respectively. DDH BD006, 336.1 m. Scale bar is 100 µm long.

Figure 6.22: Photomicrograph of sphalerite (sph) enclosing galena (gn) enclosing spongy anhedral chalcopyrite 2 (ccp 2). Gangue mineral is quartz. DDH BD004, 86.8 m. Scale bar is 50 µm long.

Figure 6.23: Photomicrograph of tiny blebs of chalcopyrite 3 (yellow) after chalcopyrite disease in sphalerite (sph). The white-grey coloured mineral at top left is galena (gn). The lamellar masses at centre and bottom right are supergene marcasite that exhibit birdseye texture. DDH BD087, 297.78-279.94 m. Scale bar is 200 µm long.

Figure 6.24: Photomicrograph of talnakhite (tal 2) partly enclosed by bismuthinite (bis). A gold grain (Au) occurs at centre left. Gangue minerals are quartz (flat grey) and calcite (pitted grey). DDH BD079, 264 m. Scale bar is 50 µm long.

Figure 6.25: Photomicrograph of gold (Au) as fill to fractures in arsenopyrite (asp). DDH BD079, 264 m. Scale bar is 50 µm long.

Figure 6.26: Photomicrograph of gold (Au) in intimate association with bismuth (Bi), as fill to a fracture in quartz. DDH BD080(2), 286.95 m. Scale bar is 0.5 mm long.
**Figure 6.27**: Photomicrograph of gold (Au), bismuth (Bi) and bismuthinite (bis) as fill to a fracture in quartz (flat grey). Note the bismuthinite and euhedral quartz grain in calcite (cal). The texture suggests that gold, bismuth, bismuthinite, calcite and quartz precipitated at the same time. DDH BD080(3), 286.95 m. Scale bar is 0.5 mm long.

**Figure 6.28**: Photomicrographs.
A. Trails of bismuth-gold 2-20 µm inclusions (black) along a healed fractures in quartz.
B. 10-15 µm gold-bismuth and bismuth only inclusions in quartz (reflected light).
C. Same as B (transmitted light). DDH BD080, 286.95 m. Photo-micrographs reproduced with permission of R.N. England (Thank you!!).

**Figure 6.29**: Photomicrograph of bismuth (Bi), gold (Au), hedleyite (had) and bismuthinite (bis) within a fracture in pitted pyrrhotite 3 (po 3). The gold and bismuth enclose and partly replace hedleyite and bismuthinite. The texture suggests that the co-precipitation of hedleyite and bismuthinite preceded the co-precipitation of gold and bismuth. DDH BD080(3), 286.95 m. Scale bar is 50 µm long.

**Figure 6.30**: Photomicrograph of bismuthinite (bis) replacing bismuth (Bi) in a fracture containing calcite (dark grey). A talnakhite grain (tal 2) lies at centre right and a gold grain (Au) lies at top right. DDH BD080(3), 286.95 m. Scale bar is 50 µm long.

**Figure 6.31**: Photomicrograph of an inclusion of bismuth-rich galena in coarse bismuth (Bi). DDH BD080(3), 286.95 m. Scale bar is 20 µm long.

**Figure 6.32**: Deformation texture. Photomicrograph of fractured arsenopyrite 1 (asp 1) in association with pyrrhotite 3 (po 3). The pyrrhotite is remobilised into the fractures. DDH BD083, 287.0 m. Scale bar is 500 µm long.

**Figure 6.33**: Deformation texture. Photomicrograph of decussate texture of quartz along a fracture that has offset a quartz-sulphide vein. The re-equilibration texture is the result of high strain along the fracture with thermal annealing. DDH BD001, 52.10 m. Scale bar is 1 mm long.

**Figure 6.34**: Deformation texture. Photomicrograph of deformed crack-seal quartz fibres. Corrugation twins occur at centre, while serrated and stylolitic contacts occur along fractures at centre right. DDH PD001, 156.51-156.61 m. Scale bar is 2 mm long.

**Figure 6.35a**: Deformation texture and supergene alteration. Photomicrograph of deformed pyrrhotite (po) in reflected light. Bird's eye supergene marcasite cross-cuts a pyrite (py) grain at bottom left. Scale bar is 500 µm long.

**Figure 6.35b**: Photomicrograph Figure 6.35a in reflected light. The deformed pyrrhotite exhibits dense corrugation twins and serrated grain margins when the polars are crossed. DDH BD080(3), 286.15 m. Scale bar is 500 µm.

**Figure 6.36**: Deformation texture. Photomicrograph of dislocation slip bands in sphalerite. Tiny blebs of chalcopyrite after chalcopyrite disease which are offset by slip have invaded cleavage planes in the sphalerite. An irregularity in the sphalerite crystal lattice can be seen at bottom left. DDH QD002, 144.76-144.89 m. Scale bar is 50 µm.

**Figure 6.37**: Deformation texture. Photomicrograph of kink bands in galena. DDH BD087, 267.78-297.94 m. Scale bar is 100 µm.

**Figure 6.38**: Deformation texture. Photomicrograph of BD001, 138.6 m. Diffusive mass transfer is indicated by the development of a pervasive fabric in an assemblage of pyrite (py), chalcopyrite (ccp), arsenopyrite (asp) and loellingite (loell). The fabric sub-parallels the regional S2. DDH BD001, 138.8 m. Scale bar is 500 µm.

**Figure 6.39**: Supergene alteration. Photomicrograph of the replacement of chalcopyrite (ccp) by bornite (bor), covellite (cov) and digenite (dig). Transition zone of the Batman deposit. DDH BD027, 391.95 m. Scale bar is 100 µm.
Figure 6.40: Photomicrograph of supergene marcasite. Complete pseudomorphs of pyrite occur at bottom left (large squarish mass) and bottom right: the remainder comprises cellular or box-work pseudomorph after pyrite. DDH BD021, 117.5 m. Scale bar is 200 µm.

Figure 6.41: A schematic of the vein paragenesis for quartz-sulphide veins of the Batman deposit. The sketch illustrates the complex array of cross-cutting textures and micro-fractures in thin sections BD006, 336.1 m; BD008, 90.9 m; BD018, 88.1 m; BD027, 172.2 m; BD079, 264.0 m; BD080(1), 286.95 m; BD081, 267.5 m; BD081, 385.0 m and BD083, 287.0 m. (Alteration and carbonate mineralogies are not illustrated). It should be noted that a single vein may or may not record the full paragenetic sequence. Indeed, in hand specimen a vein may comprise a series of cross-cutting sulphide and/or silicate micro-veinlets. Alternatively, a vein may comprise a single sulphide or silicate phase. Furthermore, calcite-base metal vein textures, S$_2$ and supergene alteration have substantially modified the quartz-sulphide assemblage.

Figure 6.42: Paragenetic sequence for quartz-sulphide and quartz-tourmaline veins of the Batman deposit, Mt Todd district.

Chapter 7

Figure 7.1: Group 2 fluid inclusions. The inclusions constitute a single compositional type (type A) and occur in growth zones in euhedral quartz (qtz 2). DDH B086, 143.0 m. Scale bar is 100 µm long.

Figure 7.2: Type A fluid inclusions in a growth zone in qtz 2. The inclusions contain liquid (L) and vapour (V), and are accompanied in the same growth zone by greenish-coloured tourmaline (T). DDH BD086, 143.0 m. Scale bar is 30 µm long. (Same photograph as Figure 6.31 but upside down.)

Figure 7.3a: Photomicrograph of type B fluid inclusions in quartz. The inclusions contain liquid (L), vapour (V), carbonate daughter mineral (C), unidentified acicular daughter mineral (N), and halite (H). DDH BD006, 147.25 m. Scale bar is 10 µm long.

Figure 7.3b: Photomicrograph of type B fluid inclusions in quartz. The inclusions contain liquid (L), vapour (V), a carbonate daughter mineral (C), an unidentified acicular daughter mineral (N), and halite (H). Nahcolite is orientated NE-SW beneath the halite daughter mineral. DDH BD006, 147.25 m. Scale bar is 10 µm long.

Figure 7.4: Photomicrograph of a trail of type C fluid inclusions (arrowed) in a healed fracture with gold-bismuth blebs (black) in quartz. Tourmaline (T) occurs as accidentally trapped inclusions in the wall of some fluid inclusions. DDH BD080(2), 286.95 m. Scale bar is 30 µm long.

Figure 7.5: Composite photomicrograph of a trail of type C fluid inclusions (arrowed) in the same healed fracture with gold-bismuth blebs (black) in quartz. The fluid inclusions occasionally wet the gold-bismuth blebs suggesting they formed at the same time. DDH BD086, 143.0 m. Scale bar is 20 µm long.

Figure 7.6: Photomicrograph of type C fluid inclusions in quartz "wetting" opaque blebs of gold-bismuth. The inclusions contain liquid (L) and vapour (V), and occasionally connect gold-bismuth blebs though fluid-filled tubes (large open arrow). Tourmaline (T) occurs as accidentally trapped inclusions in the wall of some fluid inclusions. DDH BD086, 14.0 m. Scale bar is 30 µm long.

Figure 7.7: Photomicrograph of a trail of type D fluid inclusions in quartz. DDH BD001, 56.0 m. Scale bar is 40 µm long.

Figure 7.8: Photomicrograph of a cluster of type E fluid inclusions in quartz. H = halite, L = liquid, and V = vapour. DDH BD086, 345.0 m. Scale bar is 30 µm long.
Figure 7.9: Serial photomicrographs of type E fluid inclusion in quartz. The inclusion is 25 μm long and contains an unidentified acicular mineral (N), halite (H) and liquid (L). DDH BD086, 345.0 m.
A. At 20.1°C.
B. At -12.6°C strongly birefringent hydrohalite is clearly visible.
C. At 8.6°C hydrohalite is metastable. The hydrohalite melted at 16.2°C with the simultaneous growth of a halite crystal.

Figure 7.10: Flow diagram of the observed cross-cutting relationships of types A, B, C, D, and E fluid inclusions, and their relationship to qtz-2-tourmaline and gold-bismuth.

Figure 7.11: Type A fluid inclusions in quartz.
A. Histogram of data for the final melting temperature of CH₄ (TmCH₄), the final melting temperature of CO₂ solid (TmCO₂), the first melting temperature of NaCl-H₂O ice (Te), and the final melting temperature of NaCl-H₂O ice (Tm).
B. Histogram of data for the temperature of total homogenisation (Thtot).

Figure 7.12: Type B fluid inclusions in quartz.
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Mineral key: qtz = quartz, tour = tourmaline, po = pyrrhotite, ccp = chalcopyrite, cub = cubanite, tal = talnakhite, asp = arsenopyrite, py = pyrite, marc = marcasite, sph = sphalerite, gn = galena, Bi = bismuth, bis = bismuthinite, pav = pavonite, hed = hedleyite, Bign = bismuth-rich galena, Au = gold.

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1.1 PREAMBLE

In the Pine Creek Inlier, Northern Territory, Australia (Figure 1.1), gold deposits in Palaeoproterozoic rocks exhibit a spatial association with granite plutons that is suggestive of a genetic relationship (Ewers & Scott, 1977; Wall, 1989; Wall & Taylor, 1990; Stuart-Smith et al., 1993). The deposits are associated with the development of quartz veins and lodes in shear zones, fractures, joints and faults that are located adjacent to, or within, antiformal or domal structures (Koerber, 1989; Wall, 1989; House, 1990; Partington, 1990; Nicol, 1991; Donaldson, 1992; Lawrie, 1993; Matthäi & Henley, 1993). They are typically situated in competent rock units, such as greywackes and siltstones (Koerber, 1989; Donaldson, 1992) or located proximal to carbonaceous metasediments (Matthäi & Henley, 1993) and pyritic dolomitic siltstone (Stuart-Smith et al., quoting Nicholson & Eupene, 1984), thus suggesting the deposits are, in part, stratigraphically controlled. The deposits are believed to form during metamorphic devolatilisation reactions associated with granite emplacement (Wall, 1989; Matthäi & Henley, 1993; Sheppard & McNaughton, in press), mixing of magmatic and metamorphic fluids (Koerber, 1989; Donaldson, 1992), or metamorphic devolatilisation of the lower crust coincident with underplating and generation of I-type magmas (Lawrie, 1993).

Recently, exploration interest in these deposits has intensified. From studies by Wall (1989) and Wall and Taylor (1990), thermal aureole gold deposits have been recognised as a unique deposit style. However, the genetic association between mineralisation and granite plutons is not clear and, as demonstrated by Lawrie (1993) and Sheppard and McNaughton (in press), may fail in identifying the source(s) of gold and ore-related metals. Furthermore, genetic associations are poorly constrained to regional and local tectonics, and to other local gold deposit styles.

The gold deposits of the Mt Todd region in the southern portion of the Pine Creek Inlier provide an excellent laboratory to test the genetic relationships between spatially associated gold deposits, and between deposits and adjacent granite plutons. The region is host to several extensive gold-bearing vein and lode systems that show some affinity with the thermal aureole gold style. The largest of these, the Batman deposit, has an identified mineral resource of 92.9 million tonnes of ore grading 1.4 g/t gold (Poxon & Hein, 1994). The vein and lode systems appear to have a similar structural and metallogenic history, although are dissimilar in local structural setting, in metal concentration, and in metal
distribution. They are hosted within similar rock types (greywackes and siltstones) and record geochemically and petrographically similar events. The geology of these gold deposits, their metallogeny, inter-relationships and association with local granites of the Cullen Batholith, are the subject of this thesis.

1.2 AIMS

This project was initiated in order to document the geology of gold deposits in the region surrounding Mount Todd, and to define the geological features significant to their locations. Specific aims were:-

1. To define the rock type(s) that host mineralisation through detailed mapping of the region, and to characterise the composition and competency of those rock types. The rock types had previously not been described in detail, but were considered significant to gold exploration in locating gold-bearing veins and lodes.

2. To establish the regional and local structures, particularly the location of shear zones, and major and minor folds, and to develop a structural and metamorphic history of the area.

3. To examine the influence of granite emplacement on rock type, particularly in relation to changes in rock competency and deformation. Given the spatial association of gold deposits to the Cullen Batholith, it was considered important to gold exploration to establish what relationship, if any, existed between the gold deposits, granite emplacement, and regional and local structures.

4. To define the mineralogy, fluid history and isotopic character of the vein system of the Batman deposit, and compare that system with others in the goldfield. The deposits appeared to be rather similar in style and mineralogy (although dissimilar in local structural setting, metal concentration, and metal distribution) and it was possible they were co-genetic, thus suggesting the presence of an extensive palaeo-hydrothermal system and further gold potential.

5. To establish the strata-tectonic history for the region, and propose a genetic model for gold mineralisation in the Mt Todd area.

1.3 LOCATION AND PHYSIOGRAPHY

The Mt Todd goldfield lies in the southern part of the Pine Creek Inlier (hereafter termed PCI) and is located 45 km NNW of Katherine and 45 km SE of Pine Creek, in the Northern Territory of Australia, at latitude 14° 08', longitude 132° 08' (Figure 1.1). The goldfield is wholly incorporated in the Edith River Region 1:100 000 scale geology map and the Cullen Mineral Field 1:125 000 scale geology map, and includes all the deposits surrounding the Cullen Batholith.
The topography of the Mt Todd goldfield can be divided into a western area which is composed predominantly of granite, a central region where bedrock consists of hornfelsed Palaeoproterozoic metasediments, and an eastern region where bedrock consists of poorly hornfelsed to unhornfelsed Palaeoproterozoic metasediments. The predominantly granitic region typically consists of low undulating plains with an average elevation of 120 m. The plains are dotted with tors, and granite bedrock is overlain by a mantle of grus (fine granite gravel or sand) (Twidale, 1981).

The topography of the central western region consists of actively eroding hills (maximum elevation of approximately 250 m) that are covered with colluvium and scree. The region does not contain significant areas of flat land either in valley floors or on hill tops. Bedrock consists of hornfelsed Palaeoproterozoic metasediments.

The topography of the central western region gradually gives way to the eastern region where bedrock consists of poorly hornfelsed Palaeoproterozoic meta-sediments. The geomorphology of this region is largely controlled by regional and local rock competencies such that a combined ridge and flat valley geomorphology is developed. Ridges are elongate and are covered with colluvium and scree. They are generally incised by numerous small channels. Maximum ridge elevation approximates 220 m. The ridges give way to elongate valleys and plains which have an average elevation of 130 m. The valleys and plains contain low rounded hills which are dissected by channels and occasional wide intermittent streams.

1.4 CONVENTIONS AND ACRONYMS

Throughout this thesis several formats and terms have been adopted in order to avoid confusion. These are as follows:-

1. The full extraction Mount Todd refers specifically to the geographic feature known by that name, and Mount Todd prospect refers to the prospect situated on that geographic feature. Mt Todd region refers to that area in Figure 1.2 bounded by the Fergusson River and Driffield Creek, the Arnhem Land Escarpment and the Stuart Highway. Mt Todd goldfield refers to that sub-area within the Mt Todd region which surrounds Mount Todd (Figure 1.1). It does not include the Yenberrie Wolfram Field or the Horseshoe Tin Field which are referred to separately.

2. The terms Batman deposit, Robin, Jones Brothers Workings, Golf, Regatta, Regatta South, Hotel, Penguin, Dick Grayson, Morris Mine Dam, Joker, Horseshoe, Hunters Resources, New Tollis, Old Tollis, Mt Todd Battery, Alpha, Bravo, Charlie, Delta, Q3-Q4 and Quigleys West refer to locations within the Mt Todd goldfield as shown in Figure 1.3. Quigleys deposit is the collective term for the Quigleys North Extended, Quigleys North pit,
Quigleys Central pit, Quigleys 10500 pit, and Quigleys South pit.

3. Since "Tollis" has been historically applied to three separate locations in the goldfield, its use has been abandoned in favour of Old Tollis for those historic workings described and documented by Hossfeld and Nye (1941), and New Tollis for the region immediately west of the Jones Brothers Workings, including "Tollis EM" of Kenny et al. (1990).

4. A prospect is defined as an area assessed to be a geological or geophysical anomaly by the Joint Venture Partners of Billiton Australia Gold Pty. Ltd. and Zapopan N.L., to December 1991.

5. Sedimentary rocks are classified according to grain size using the Udden-Wentworth scale as adopted by Adams et al. (1987). Thus a greywacke is a sandstone containing greater than 15% fine grained matrix. Sandstone greywacke and siltstone greywacke are non-quantitative terms used in field work to define units which cannot be immediately classified as either sandstone or greywacke, or siltstone or greywacke.

6. $S_n$ and $F_n$ are used respectively, to denote foliations and folds, which are developed during some deformation $D_n$. The subscript "n" is used to denote the relative age of each deformation for which $D_1$ is recognised as the earliest phase.

7. All structural readings are quoted with respect to grid north. True North lies 0.5°C west of Grid North. Stereographic projections are Equal Area, unless otherwise stated and are plotted to grid north.

8. The Precambrian time scale nomenclature is adopted in accordance with the IUGS-approved chronometric subdivision of Precambrian time and the Proterozoic Eon, of Plumb (1991).

Acronyms used in this thesis include:-

- cm centimetre(s)
- km kilometre(s)
- Ma million years ago
- mm millimetre(s)
- PCI Pine Creek Inlier
- °C degrees Centigrade
- µm micrometer(s)

1.5 **THESIS ORGANISATION AND SUMMARY**

This thesis is divided into text (Volume I) and appendices (Volume II), the latter containing maps on which a large part of the text is based, and thin section descriptions: this volume is included in accordance with an agreement between the author and the Joint Venture Partners of Billiton Australian Gold Pty Ltd and Zapopan NL to submit all relevant data in a concise form.
The text is subdivided into eight chapters excluding the introduction:

Regional setting (in review):- Chapter 2 reviews the lithological, structural and metallogenic setting of the Central Domain of the Pine Creek Inlier, relevant to the Mt Todd goldfield.

Lithologies of the Mt Todd goldfield: Chapter 3 describes and characterises the nature of clastic, volcanoclastic and intrusive rocks in the goldfield, based on detailed field and petrographic studies. The clastic sediments comprise a monotonous sequence of greywackes, siltstones and shales and these are host to lode and vein style gold mineralisation.

Structure and metamorphism: - Chapter 4 presents the results of detailed structural and metamorphic mapping across the goldfield and selected out-lying areas, and is supported by petrographic studies of metamorphic assemblages. Special attention has been paid to characterising the timing of gold mineralisation with respect to regional deformations, local plutonism and the formation of other vein types. A structural and metamorphic history is established.

Local structures: - Chapter 5 presents the results of a detailed structural analysis of two gold deposits in the Mt Todd goldfield. These are the large-tonnage, low-grade Batman deposit which is characterised by sheeted and stockwork quartz-sulphide veins (currently in operation), and the low-tonnage moderate-grade Quigleys deposit which is characterised by a series of parallel quartz-sulphide breccia-type lodes. The structural features which are considered fundamental to the formation of the deposits are established and compared, and the timing of the gold mineralising event is refined.

Mineralogy and micro-structures: - Chapter 6 details the mineralogy and microstructure of gold-bearing quartz-sulphide veins of the Batman deposit and establishes a genetic sequence of formation of silicate, sulphide and carbonate mineral assemblages. The character of vein deformation and supergene alteration is examined, and the physico-chemical conditions operating during vein formation are estimated using established phase relationship data for selected sulphide assemblages.

Microthermometry and Laser Raman analyses: - Chapter 7 presents the results of a fluid inclusion study of five characteristic and sequential fluid inclusion types using microthermometry and laser Raman techniques. The inclusions are constrained to the mineral paragenesis established in Chapter 6, through a temporal and spatial association between fluid inclusion type and mineral phase. Thereby, the physico-chemistry of the mineralising hydrothermal fluid, and the nature of the fluid from which the gold precipitated is characterised.

The source and geochemistry of ore fluid: - Chapter 8 constrains the sources of ore forming fluids and of sulphur, by combing the geologic, petrographic and fluid chemistry data outlined in the preceding chapters. The sources are refined through oxygen and sulphur isotope analyses, respectively. Sources for metal and ligands, the method of gold and metal transport, and the physical and chemical processes which may have been involved in metal deposition are evaluated.
Conclusions: Chapter 9 summarises the strato-tectonic history of the goldfield with respect to the genesis of gold mineralisation, and concludes the nature of the hydrothermal fluid. The character of gold deposits in the Mt Todd goldfield is briefly compared with the thermal aureole gold style.

In order to avoid breaking the continuity of the text, figures and tables are presented at the end of the chapter concerned, excluding the case where reference is made to an earlier figure or table.
Figure 1.1: Location map of the Mt Todd goldfield.
Figure 1.2: Definition: the Mt Todd goldfield.
CHAPTER 2
REGIONAL SETTING (IN REVIEW)

2.1 REGIONAL GEOLOGY

2.1.1 Introduction
The Mt Todd goldfield lies in the southern region of the Pine Creek Inlier (PCI) in the Central Domain (Figure 2.1). The domain consists of Palaeoproterozoic geosynclinal and transitional sequences, numerous granite intrusives, occasional basic intrusives, and acid and intermediate volcanics (Plumb et al., 1981). Sediments of the domain are obscured to the east, where they dip beneath Palaeo- to Mesoproterozoic sediments of the Arnhem Land Plateau; in the south and south-west, where they dip beneath Palaeozoic Daly River Basin sediments which unconformably overlie Neoproterozoic Victoria Basin sediments, and in the north, where they dip beneath Mesozoic sediments of the Bathurst Terrace. A stratigraphic column for the Central Domain is presented in Figure 2.2.

2.1.2 Basement Rocks
In its western region, the Central Domain consists of two dome-shaped inliers of Archaean basement complex, respectively named, the Rum Jungle Complex and Waterhouse Complexes (Figure 2.1). The Rum Jungle Complex is older than 2300-2400 Ma, this being a Rb/Sr age for the youngest known intrusive leucogranite in the Complex (Richards et al., 1977). An Archaean age for formation of the complex is supported by a U-Pb zircon discordia of 2450 Ma (Plumb et al., 1981, pp. 220). The complexes generally consist of phyllitic adamellite, biotite-muscovite granite, migmatites, gneisses, schist and metadolerite (Plumb et al., 1981) and are unconformably overlain by sediments of the Palaeoproterozoic Geosynclinal Sequence (Needham et al., 1988). The complexes are inferred to be laterally continuous at a depth of less than 5000 m throughout the Central Domain (Tucker et al., 1980).

2.1.3 Geosynclinal Sequence
The sediments of the Palaeoproterozoic Geosynclinal Sequence were deposited in several half-graben style, intracratonic extensional basins at approximately 2000 Ma by the process of rifting and subsidence of the Archaean basement (Needham et al., 1988). The sequence can be divided into a lower rift package, and an upper laterally extensive thermal subsidence package (sag phase), which in turn is covered by a thick pre-orogenic turbidite package (Wyborn, 1988). The thickness of each package, and the type of sediment deposited, varies across each half-graben.
The two oldest groups of the Geosynclinal Sequence in the Central Domain, the Namoona Group and the Mount Partridge Group, characterise what has been termed the "rift" phase of sedimentation (Needham et al., 1988). The Namoona Group consists of a sequence of carbonaceous pelites, calcareous sandstones and evaporites known as the Masoon Formation, which occasionally interfinger with volcanic breccias (Needham et al., 1980) and lavas of ocean and tholeiitic affinity (Stuart-Smith et al., 1980). The volcanic breccias and lavas have been termed the Stag Creek Volcanics. The Namoona Group is separated from the Mt Partridge Group by an unconformity that is the result of gentle warping of the older Namoona Group sediments (Stuart-Smith, 1987).

The Mount Partridge Group consists of two conformable formations, the Mundogie Sandstone and the Wildman Siltstone. The former comprises a sequence of conglomerate, arkose, sandstone and siltstones that thin southward from an inferred northern upland (Needham et al., 1988). The latter is composed of siltstone, black carbonaceous laminated shale and minor sandstones, and has been interpreted as the distal and vertical subtidal gradation of the fluvial fans of the Mundogie Sandstone (Stuart-Smith et al., 1980). The Mount Partridge Group is unconformably overlain by "sag" phase sediments (Needham & Stuart-Smith, 1985a), the unconformity marking a period of warping.

The "sag" phase of sedimentation of the Geosynclinal Sequence is represented by the South Alligator Group (Needham et al., 1988) and has been dated at 1885±2 Ma (U-Pb zircon age; Page and Williams, 1988). The group consists of three formations: (1) shallow marine transgressive sequence of the Koolpin Formation (pyritic carbonaceous shale, chert, carbonaceous siltstone, and dolomite); (2) extensive subaerial felsic volcanics of the Gerowie Tuff; and (3) greywacke and volcanic sequences of the Mount Bonnie Formation. The latter is host to stratiform massive sulphide mineralisation containing Au, Ag, Cu, Pb, and Zn (Stuart-Smith, 1987). It is conformably overlain by a pre-orogenic package of sediments termed the Finniss River Group (Stuart-Smith, 1985).

The Finniss River Group is the youngest group of the Geosynclinal Sequence in the Central Domain and is exclusively represented by the Burrell Creek Formation. The formation consists predominantly of interbedded greywacke, siltstone, sandstone and shale and is interdispersed with volcanic and gritstone horizons. The formation commonly exhibits turbiditic sedimentary features including graded beds, scour and fill marks, flute casts, flame structures and occasional dewatering structures (Needham et al., 1980). Deposition of the Finniss River Group terminated with the onset of a major period of deformation during the Top End Orogeny (Stuart-Smith, 1987).
2.1.4 Top End Orogeny

Between 1870 and 1780 Ma the PCI underwent a major period of deformation known as the Top End Orogeny (Needham et al., 1988). The orogeny consists of three compressional events; the Nimbuwah Event which has been correlated with the Barramundi Orogeny recognised throughout most of northern Australia by Etheridge et al. (1987), the Maud Creek Event and the Shoobridge Event; each has been interrupted by a phase of extensional tectonism. Johnston (1984) concluded that five deformation events were associated with compressional tectonism, viz, \( D_1 \) - westerly verging thrusts, \( D_2 \) - recumbent folding, \( D_3 \) - upright north-south folding, \( D_4 \) - east-west folding, and \( D_5 \) - kinking related to basement block movements.

The principal phase of compression took place during the Nimbuwah Event which has been dated by Page and Williams (1988) at 1885-1870 Ma (U-Pb zircon age). Compressional tectonism was accompanied by the development of early bedding parallel thrusts, recumbent to upright folds (Stuart-Smith et al., 1993) and retrogressive metamorphism to lower to upper greenschist facies (Stephansson & Johnson, 1976; Ferguson et al., 1980). Compression was synchronous with topographic uplift and erosion, and was preceded by the intrusion of a suite of tholeiitic sills known as the Zamu Dolerite (Needham et al., 1988).

The period of extensional tectonism following the Nimbuwah Event was accompanied by the development of a NW-SE trending shallow rift graben near the eastern margin of the Central Domain, in the vicinity of the present day Alligator River Valley. This graben became the locus for the felsic lava flows, volcanoclastic and flysch sediments of the El Sherana Group. Deposition of the Coronation Sandstone (conglomerate, sandstone and shale) and the Pul Pul Rhyolite (extensive sheets of felsic lava) within the graben, preceded the deposition of the Tollis Formation (beds of volcanolithic greywacke, siltstone, crystal tuff and ignimbrite) and the Big Sunday Formation (rhyolite, basalt, tuff, siltstone and feldspathic greywacke) within the graben and across surrounding uplands (Needham & Stuart-Smith, 1985b). These sediments and volcanics were subsequently folded and faulted during the second phase of compressional tectonism, the Maud Creek Event.

The Maud Creek Event, dated at 1850 Ma (Needham et al., 1988), was associated with the development of north to northeast trending upright folds which were refolded by broadly spaced open, east-west trending folds (Stuart-Smith et al., 1993). Folding was accompanied by faulting, uplift and erosion.

The Maud Creek Event was immediately preceded by the emplacement of large granitoids of the Cullen Batholith at between 1835-1820 Ma (Stuart-Smith et al., 1993). The granites have intruded the sediments of the Geosynclinal Sequence and the El Sherana Group, extensively contact metamorphosing them to K-feldspar cordierite facies and hornblende-
hornfels facies mineral assemblages. The granites are predominantly I-type in character (Ewers & Scott, 1977; Etheridge et al., 1987; Wyborn, 1988) although Ferguson et al. (1980) recorded the existence of S-type granites near the margins of the batholith. The I-type granites are characterised by the presence of biotite, some hornblende, allanite and sphene, and are devoid of meta-sedimentary xenoliths (Wyborn, 1988). The batholith is associated with numerous mineralised sites throughout the Central Domain including those anomalous for Sn, W, Au, Ag, Pb, Zn, Cd, Bi, Cu, U and Mo. The Tennysons Leucogranite and the Yenberrie Leucogranite, are those plutons of the Cullen Batholith located close to the Mt Todd goldfield.

Following emplacement of the Cullen Batholith, graben subsidence was once again initiated through rifting and downwarping of the peneplanated landscape. The Mount Callanan Basin, the Birdie Creek Basin, and the Edith Falls Basin were formed at this time by normal movement along NW-SE trending faults (Needham & Stuart-Smith, 1985b). These became the locus for rhyolitic ignimbrites and lavas, dacites, basalts, rare andesites, and tuffaceous sediments of the Edith River Group (currently accepted conventional U-Pb zircon age of 1830 Ma; Stuart-Smith et al., 1993).

The Shoobridge Event, the final deformational event of the Top End Orogeny, was largely confined to the western margin of the Central Domain; it has been dated at 1770 to 1780 Ma and was accompanied by the development of new and/or reactivation of pre-existing linear structures (Needham et al., 1988) such as the Pine Creek Shear Zone. The deformation has been associated with widespread retrogressive metamorphism, particularly within the Cullen Batholith (Stuart-Smith et al., 1993).

The emplacement of anorogenic intrusives, such as the Lewin Springs Syenite and the shallow lopoliths of the Oenpelli Dolerite (dated at 1688 Ma by Page et al., 1990), mark the final known geological events in the evolution of the Central Domain of the Pine Creek Inlier. The emplacement predates a period of extensive erosion and intense chemical weathering (Needham & De Ross, 1990).

### 2.1.5 Platform Sequence

The Palaeo- to Mesoproterozoic platform sediments of the Arnhem Land Plateau bound the eastern margin of the Central Domain of the PCI. The sediments consist predominantly of sandstone with minor siltstone, conglomerate and volcanic interbeds (Needham & De Ross, 1990). The sediments were deposited on a broad shelf (the Arnhem Shelf) adjacent to a subsiding north-south trending graben (Plumb et al., 1990) that occupied a position close to the eastern coastline of the present day Arnhem Land peninsula. In the southeastern region of the Central Domain, rifting reinitiated subsidence of the Edith Falls Basin along a set of NW-SE trending faults (Needham & De Ross, 1990; Plumb et al., 1990). The basin
became the locus for sediments of the Katherine River Sequence: the sediments consist of coarse quartz-sandstone, quartz-pebble to boulder conglomerate, minor polymict boulder conglomerate, andesite and basalt lava (Bagas et al., 1982). Deformation of the sediments was confined to gentle tilting and warping (Mulder & Whitehead, 1988).

2.1.6 Victoria Basin and Daly River Basin

The Early Cambrian lavas of the Victoria Basin, and the Middle Cambrian calcareous sediments of the Daly Basin, onlap or are faulted against the southern and southwestern margin of the Central Domain of the PCI. Faulted contacts exhibit normal displacement and drag folding, and fault blocks are down thrown to the southwest (Stuart-Smith et al., 1993).

The lavas of the Victoria Basin unconformably overlie, or are faulted against the Burrell Creek Formation, Cullen Batholith and Katherine River Sequence. Essentially, the lavas comprise massive continental flood-type basaltic flows that are interbedded with sandstone and alternating limestone and sandstone units (Mulder & Whitehead, 1988).

The calcareous sediments of the Daly River Basin unconformably overlie, or are faulted against the Geosynclinal and Platform Sequences, Cullen Batholith, and lavas of the Victoria Basin. They generally consist of interbedded limestone, sandstone, dolomite and shale which have remained relatively flat-lying since their deposition in the Cambrian (Plumb, 1979).

2.2 REGIONAL STRUCTURES

2.2.1 Folds

In the Central Domain of the PCI, two significant fold morphologies are associated with regional deformation during the Top End Orogeny:-

1. Upright north-south trending folds. These folds have been classified as D₃ folds by Johnston (1984) and are associated with regional metamorphism to greenschist facies.
2. East-west trending folds which have been classified as D₄ folds by Johnston (1984).

North-south trending D₃ folds are generally open to tight and non-cylindrical throughout the domain, but increase in intensity toward its eastern boundary (Stuart-Smith et al., 1980), i.e., folds become tight and overturned and finally, isoclinal (Needham et al., 1980). In trend, the folds display a megascopic sinusoidal swing which Johnston (1984) concluded could be neither attributed to an original orocline, nor to late folding. Folds trend north-northeasterly in the north of the domain, northwesterly in the central region, and north-northeasterly in the southern region in the vicinity of Mt Todd (Figure 2.3). Needham et al. (1980) noted that the change in the trend direction from north-northeasterly to north-northwesterly occurred about the northeast trending Grove Hill Cross Flexure. Recent work by Stuart-Smith et al. (1993)
has suggested that the north-south trending folds may resolve two fold generations, i.e., northwest trending, upright to overturned tight to isoclinal folds formed during the Nimbuwah Event (~1870 Ma), and north to northeast trending open to tight folds formed during the Maud Creek Event (~1850 Ma).

The existence of D₄ east-west trending folds throughout the Central Domain has not been widely recognised, however the folds are persistent across the region. They increase in intensity eastward from an open fold style in the centre of the Domain, to a tight fold style at the eastern margin of the Domain (Johnston, 1984). Stuart-Smith et al. (1993) noted that the folds were associated with emplacement of the Cullen Batholith and may be allied with east-west trending kinks or crenulation cleavage. Koerber (1989) documented the existence of macroscopic open style east-west trending folds in the Mt Todd goldfield in association with an axial planar disjunctive cleavage. East-west trending macroscopic and mesoscopic folds have also been observed in the Union Reefs prospect in association with an axial planar crenulating to disjunctive cleavage (Donaldson, 1992).

2.2.2 Faults

Tectonic activity in the Central Domain of the PCI, has been dominated by the development and reactivation of NW to NNW and NE to NNE trending strike-slip and normal faults (Needham et al., 1980; Needham et al., 1988; Plumb et al., 1990) and thrust faults (Johnston, 1984). Strike-slip and normal faults appear to have been active over a considerable period of time, from early in the Palaeoproterozoic, to well in the Cenozoic (Plumb, 1979; Needham et al., 1988). Plumb (1979) speculated that their orientations reflected a characteristic northwest to northeast fracture pattern that was established throughout the northern Australian craton by at least the Palaeoproterozoic. Simpson et al. (1980) noted the coincidence of fault orientations with dolerite dykes, some gravity trends, and with NE and NW trending Landsat linears, and concurred that the features were expressions of deep basement fractures.

Normal faults in the Central Domain are related to rifting and subsidence of the following:-

1. The Archaean basement during the Palaeoproterozoic,
2. The Geosynclinal Sequence and the Platform Sequence at the end of the Top End Orogeny (during the Palaeo- to Mesoproterozoic),
3. All Precambrian sequences during the formation of the Phanerozoic-aged Daly River Basin.

They are believed to have been active over a substantial period of time since they displace sediments deposited in the Palaeoproterozoic, Mesoproterozoic, Lower Palaeozoic, Mesozoic and the Eocene (Needham et al., 1988). The faults are considered to have exerted a strong influence over sedimentation patterns during deposition of the Geosynclinal Sequence and the McArthur Basin Platform Sequence (Needham & Stuart-Smith, 1985b).
There is an association between strike-slip faults and the major NW-NNW trending linear shear zones, such as the Pine Creek Shear Zone and the South Alligator Fault Zone. The northwesterly trending strike-slip faults, commonly exhibit oblique sinistral displacement, although minor dextral reversals are locally apparent (Plumb, 1979). ENE trending faults commonly exhibit dextral displacement (eg. the ABC Fault, One Mile Fault and the Two Mile Fault), as well as some late normal displacement with downthrow on the southern side of the fault. Generally, the faults have displaced sediments which were deposited during the Palaeoproterozoic, Mesoproterozoic and the Lower Palaeozoic. Activity along these faults may have been related to displacement on antithetic and synthetic basement fractures (Koerber, 1989).

Thrust faults are believed to have developed in response to tectonic sliding and east-west crustal shortening early in the Top End Orogeny (Johnston, 1984). Reactivation of thrusts during intrusion of the Cullen and Burnside granites, concomitant with upward migration of gold mineralising fluids along these structures, has been postulated as a mechanism for generating lode and stockwork mineralisation at the Goodall mine (Wall, 1989), Tom's Gully mine (Sheppard & McNaughton, in press) and the Mt Todd goldfield (Koerber, 1989).

2.2.3 Pine Creek Shear Zone
The Pine Creek Shear Zone is the dominant structural feature in the Central Domain of the PCI in the vicinity of the Mt Todd goldfield. It consists of a northwesterly trending complex zone of faults which extends from Spring Hill to the Cullen Railway Siding, a distance of some 70 km (Figure 2.3). It lies within the Noonamah-Katherine Lineament Zone (Simpson et al., 1980) which has a coincident TM signature, magnetic and gravity anomaly. In the region between Pine Creek and Spring Hill, the Pine Creek Shear Zone follows an embayment of the Burrell Creek Formation and divides the Cullen Batholith into two major lobes (Stuart-Smith et al., 1987). A net sinistral displacement of 3-5 km has been suggested by Koerber (1989). Donaldson (1992) demonstrated that sinistral activity accompanied reverse slip in the vicinity of the Union Reefs prospect, with east block up and north. Experimental work by Krowkowski and Olissoff (1990) has indicated that sinistral transcurrent displacement was synchronous with compression, ie., sinistral transpression, and that sinistral transpression preceded a late phase of dextral strike-slip activity along the shear zone.

In the region between Pine Creek and the southern boundary of the PCI, the Pine Creek Shear Zone traverses several plutons of the Cullen Batholith (Figure 2.3). In this region, the Pine Creek Shear Zone splays into a series of parallel to sub-parallel or anastomosing north-west trending faults. The faults have a considerable strike length and are associated with sinistral displacements of the order of hundreds of metres. In particular, the Phillips Creek Fault, which extends from the Cullen Railway Siding 25 km south of Pine Creek, to Leight Creek 13 km north of Katherine (a total distance of 55 km) exhibits a displacement of 1-2 km.
The Pine Creek Shear Zone is believed to have been active over a period greater than 200 million years. Stuart-Smith et al. (1987) determined that activity along the zone was least within the areas of granite and greatest along the margins of the granites. They concluded that it was most active prior to the emplacement of the Cullen Batholith (dated at 1835-1820 Ma, Stuart-Smith et al., 1993). An inferred asymmetry to the morphology of the Burnside Granite is suggestive of a synkinematic emplacement of this granite body. Left-hand offsets along faults in volcanic beds in the Kombolgie Formation (dated at about 1645 Ma, Stuart-Smith et al., 1987) at the southeastern extent of the Pine Creek Shear Zone (i.e., along the Phillips Creek Fault) indicates that the shear zone was active after the emplacement of that Formation.

2.3 REGIONAL MAGNETICS

Prominent magnetic features in the Central Domain of the PCI are those that are associated with the elongate anomalies described by Tucker et al. (1980) and Simpson et al. (1980). These anomalies include linear, curvilinear and circular or "bullseye" types.

The linear anomalies trend either northwesterly or northeasterly, and are parallel to, or coincide with, faults, mafic dykes and Landsat linears (Tucker et al., 1980). The northwest-trending anomalies include the 20-25 km wide Noonamah-Katherine Lineament (Figure 2.3) which has a coincident gravity feature, and is associated with the Pine Creek Shear Zone, numerous felsic and mafic dykes, and a northwest alignment of minor folds. Simpson et al. (1980) suggested that the lineament reflected a major crustal structure and noted its broad coincidence with the western margin of the Edith Falls Basin.

NE trending linear anomalies are relatively rare in the Central Domain, however, they occur beneath the Bathurst Terrace and in the vicinity of the Mt Todd goldfield (Tucker et al., 1980). The lineaments parallel or subparallel the strike of an extensive suite of dolerite dykes and/or faults which display right-hand offsets, such as the ABC Fault, the One Mile Fault and the Two Mile Fault (Koerber, 1989). A northeast trending magnetic lineament, extending from the Batman deposit in the Mt Todd goldfield, to the Driffield goldfield northeast of Mount Todd (Harley, 1988), is believed to reflect a deep crustal structure which may be related to antithetic fracturing in the basement during displacement along the Pine Creek Shear Zone (Koerber, 1989).

Curvilinear anomalies are associated with, or located over, steeply dipping units of the Geosynclinal Sequence which are known to contain magnetic pyrrhotite, the strongly magnetic sills of the Oenpelli Dolerite, as well as other variably magnetic volcanic units within the Platform Sequence (Tucker et al., 1980). South of Mount Todd, these anomalies are associated with magnetite-bearing volcanoclastic units.
Circular magnetic anomalies in the Central Domain are typically observed in areas of low magnetic relief and are located near to, or within, the granite complexes themselves (Tucker et al., 1980). These authors associated the anomalies with deeply buried mafic intrusions; however, shallow "bullseye" type and arcuate anomalies are found in association with pyrrhotite in sheeted veins and lodes in the Mt Todd goldfield.

2.4 REGIONAL GOLD METALLOGENY

2.4.1 Introduction

The nature of gold mineralisation in the Central Domain of the PCI has been described and discussed by Needham and Roarty (1980), Stuart-Smith and Needham (1984), Stuart-Smith (1985), Wall (1989), Needham and De Ross (1990) and Nicholson and Eupene (1990). Two deposit styles are significant and these are restricted to the thermal aureole surrounding the Cullen Batholith and its outlier plutons (Wall, 1989), and the Mt Bundey pluton. They include:

1. Quartz vein/lode gold mineralisation.
2. Alluvial gold.

2.4.2 Quartz vein/lode mineralisation

The most common form of gold mineralisation in the Central Domain is that associated with quartz vein or lode mineralisation (Figure 2.4). The veins and lodes generally occupy fractures and shear zones that are close to, or within, the axes of antiformal or domal structures, such as occurs at the Enterprise Mine (House, 1990), Spring Hill prospect (Nicol, 1991), Union Reefs prospect (Donaldson, 1992), Howley district (Partington, 1990), Mt Todd goldfield (Koerber, 1989), and Moline/North Hercules mines (Nicholson and Eupene, 1990). Quartz lodes are also located on shallow to moderately dipping faults (thrust) at the Tom's Gully mine adjacent to the Mt Bundy pluton (Sheppard and McNaughton, in press) and Quigleys deposit in the Mt Todd goldfield (Koerber, 1989). The fine quartz stockwork at the Goodall mine is believed to occupy a fracture zone that is positioned above a thrust duplex (Wall, 1989).

Quartz veins and lodes may be confined to particular stratigraphic units; the ore zones of the Cosmo-Howley mine have a strong association with calcareous iron silicate horizons within the Koolpin Formation (Wall, 1989). The style of quartz veins or lodes is dependent on lithology, i.e., sheeted quartz veins are, in general, favoured by massive and competent sandstone or siltstone greywacke units, while quartz lodes are favoured by shale units (that are often carbonaceous). Sheeted veins vary from a few millimetres to a metre in thickness,
and may strike for tens of metres. They are often stockworked where vein density and shearing is high.

Gold occurs as free gold along fractures in quartz and arsenopyrite, as inclusions within arsenopyrite and pyrrhotite, or as disseminations within the alteration selvedge around the vein. It is often been found in intimate alliance with bismuth and bismuthinite. Associated sulphides include arsenopyrite, pyrrhotite, chalcopyrite, marcasite, pyrite, galena, sphalerite. Alteration assemblages include quartz, biotite, sericite, tourmaline, chlorite, pyrite, arsenopyrite, pyrrhotite and carbonate.

2.4.3 Alluvial gold

Alluvial deposits of gold have been mined in the Howley district (Partington, 1990), Fountainhead mine (Nicholson and Eupene, 1990), Union Reefs prospect (Donaldson, 1992), Watts Creek mine, Driffield goldfield and adjacent to the Enterprise mine at Pine Creek (Figure 2.4). The alluvials are commonly located proximal to known vein/lode type deposits and occur within flat lying gravel beds of Tertiary stream channels.
Figure 2.1: Location map of the Mt Todd goldfield in the Central Domain of the Pine Creek Inlier.
Figure 2.2: Stratigraphic column for the Central Domain of the Pine Creek Inlier, modified after Stuart-Smith et al. (1987).
Archaean

Intrusive

Figure 2.3: Fold trends in the Central Domain of the Pine Creek Inlier (modified after Needham et al., 1980).
• Sheeted quartz vein and quartz lode gold mineralisation

△ Alluvial gold deposits

Figure 2.4: Location map of major gold deposits in the Central Domain of the Pine Creek Inlier (modified after Nicholson and Eupene, 1990).
CHAPTER 3
LITHOLOGIES OF THE MT TODD GOLDFIELD

3.1 INTRODUCTION
The lithologies of the Mt Todd goldfield are dominated by a Palaeoproterozoic sequence of greywacke, siltstone, sandstone and shale termed the Burrell Creek Formation. The sequence exhibits sedimentary features similar to those of a river-dominant delta front to prodelta environment, as described by Miall (1984) and Elliott (1986), and is conformably overlain by volcanolithic and volcanoclastic sediments of the Tollis Formation. The sequence is folded about northeast to north-northwest trending fold axes, and about east-west trending fold axes (Koerber 1989). It has been subjected to lower greenschist facies metamorphism (Mulder & Whitehead, 1988) and, where it is intruded by granitoids of the Cullen Batholith, it has been contact metamorphosed to hornblende-hornfels facies.

3.2 THE BURRELL CREEK FORMATION

3.2.1 Introduction
The main rocks units in the goldfield, in order of frequency, are greywacke, siltstone, shale and sandstone, with single occurrences of grainstone and graphitic chert. In weathered outcrop the rocks range in colour from pale buff-yellow to ochre-red and dark green, while in diamond drill core they are generally pale grey to dark green-grey-black. The units are laterally extensive and uniform, and are monotonously interbedded over an estimated 2 km of stratigraphic section (Figure 3.12a & b in Appendix A).

3.2.2 Greywackes
Greywackes are those sandstone rock units which contain more than 15% fine matrix material (Adams et al., 1987). In the goldfield they occur as massive blocky beds which range in thickness from 20 cm to 1.5 m. They are commonly chloritised and thus appear dark grey-green in outcrop colour, although freshly broken samples often reveal a paler colour ranging from buff yellow to red, or grey to dark grey. Although largely structureless, individual beds may exhibit base sharpness and an upward gradation into finer clast and matrix material. All or part of the Bouma sequence (Walker, 1984) may be present (Figure 3.1), together with sandy parallel laminations, dewatering structures (Figure 3.2), climbing ripples, flames, and
rare scour and load structures (Figure 3.3). Fine scale bed grading (<1-2 cm), parallel and convoluted laminations, mud balls (Figure 3.4) and rare slumps are developed in some finer grained units. Mud balls are generally ovoid to spherical in shape (1-8 cm long diameter) and are composed of concentrically zoned layers of mud, silt or sand surrounding a nucleus of mud, rock fragments, silt or sand. Petrographically, the greywackes consist of coarse silt to medium sand sized (50-500 µm), angular to subangular clasts of quartz, quartzite, chert, zircon and accessory sericitised potash and sodic feldspars, in a fine to coarse silty matrix (10-50 µm grain size) of quartz, sericite, chlorite, graphite and zircon. The greywackes are poorly sorted, and clasts are grain supported. The matrix comprises approximately 15-45% of the total rock volume. Muscovite and brown biotite plates (10 µm), clusters of rutile needles, green tourmaline needles (10 µm long), and rare sericite, chlorite and Fe-oxide spots (probable psuedomorphs of cordierite) are developed in greywackes which have been hornfelsed (Section 4.3).  

3.2.3 Siltstones
The siltstone units of the goldfield occur as fine to massive blocky beds which range in thickness from a few millimetres to more than a metre. They are commonly iron stained and thus appear buff yellow to red in colour, although chloritised and/or graphilic siltstones are coloured green-grey or dark grey in outcrop. They often exhibit bed gradation in which rarely division C, and generally division D and E of the Bouma sequence are present together with base sharpness or scouring, continuous and discontinuous sandy to silty laminations, dewatering features, and mud balls (0.5-6.0 cm long diameter). Massive beds are frequently structureless. Petrographically, the siltstones are composed of a very fine to medium-sized silty matrix (10-20 µm) of quartz, sericite, chlorite, graphite and accessory zircon. They regularly contain medium to coarse angular clasts of quartz (to 50 µm) and occasionally, quartzite, chert, potash feldspar, sodic feldspars and detrital biotite, however, the total clast content is generally less than 10% of the total rock volume. Fe-oxide + sericite + chlorite + quartz spots (psuedomorphs of cordierite), rutile needles, green tourmaline needles (10 µm), muscovite and brown biotite plates (15µm) are developed in siltstones which have been hornfelsed.  

3.2.4 Shales
Shales are developed as interbeds with greywacke and/or siltstone, or as division E of the Bouma sequence in graded greywacke or siltstone. The shales are generally deep red or green in outcrop colour, but distinctly dark green-black in diamond drill core. They occur as thin (<3 cm), continuous, laminated beds which occasionally exhibit plastic mass flow deformation structures such as slumps and convoluted laminations (Figure 3.5).
Petrographically, the shales are composed of fine sericite, chlorite and Fe-oxides, and silt-sized (20-30 µm) angular clasts of quartz, quartzite, chert and zircons may be scattered throughout. They are commonly graphitic and occasionally exhibit micro-cross beds and graded beds.

### 3.2.5 Sandstones

Sandstone beds crop out infrequently in the goldfield. They are largely indistinct from beds of greywacke, although are generally paler in colour, i.e., buff yellow to orange in outcrop and pale grey in diamond drill core. They are commonly structureless but may exhibit basal sharpness, a weak gradation toward the bed top, weak silty laminations, rare dewatering structures and asymmetric ripples.

Petrographically, the sandstones contain less than 15% matrix material and are composed of moderately sorted, angular to subrounded clasts of quartz, quartzite and zircons to 0.2 mm, and sericite and chlorite.

### 3.2.6 Arkosic grainstone

A single thin (< 4 cm) continuous bed of arkosic grainstone (gritstone) crops out on a ridge north of the Horseshoe prospect (Figure 1.3) as an interbed within a sequence of massive greywackes. It consists of coarse angular clasts of sericitised and hematised potash and sodic feldspar, devitrified glass, quartz, quartzite, chert, and lithic fragments (to 3 mm) cemented in a sandy matrix of quartz, quartzite, detrital biotite, sericite and chlorite. The matrix generally forms less than 40% of the total rock volume.

### 3.2.7 Graphitic chert

A competent bed of graphitic chert crops out in the vicinity of New Tollis (Figure 1.3) as an interbed within a sequence of siltstone and greywacke. It strikes for approximately 600 m and is less than 1 metre in width. It is predominantly composed of fine quartz (< 2 µm) and graphite, and occasionally contains angular to subangular clasts of quartz and zircon to 0.4 mm in size. It is laminated and finely cross bedded (Figure 3.6).

### 3.3 TOLLIS FORMATION

The Tollis Formation crops out along the southern boundary of the goldfield at the junction of the Mine Access Road and the Edith Falls Road (Figure 3.12a & b). In this location, the Formation conformably overlies greywackes and siltstones of the Burrell Creek Formation, and consists of massive beds of tuffaceous greywacke or gritstone (4-5 m thick) and laminated to finely bedded crystal tuff (Figure 3.7). These units are interbedded with thin horizons (30 cm) of siltstone or finely laminated shale.
The tuffaceous greywacke beds are dark grey to black in colour. They are characterised by pink spots (6 cm in diameter) that are dominated by potash feldspar (Figure 3.8a), and these often have a locus on 4-6 cm long, flattened clasts of crystal tuft or graphitic chert (Figure 3.8b). The finely bedded units are banded pink-yellow-black in colour and may show bed gradation at the centimetre scale. The older coarser portion of these beds is generally felsic in nature and dominated by lithic fragments and clasts of potash feldspar and quartz (2 mm) relative to the younger portion composed of chert, chlorite and graphite.

A U-Pb age of 1890±15 Ma has been determined for the Tollis Formation by Page and Williams (1988) from an interbedded andesitic lava flow near Edith Falls.

3.4 THE BURRELL CREEK FORMATION-TOLLIS FORMATION BOUNDARY

3.4.1 Introduction
In 1987, the Joint Venture Partners of Billiton Australia Pty. Ltd. and Zapopan N.L. undertook a stream-sediment sample program in the goldfield and subsequently identified the Batman gold anomaly, 3 km WSW of Mount Todd. At that time, the host rock greywacke, sandstone, siltstone and shale were allied with the basal sequence of the Palaeoproterozoic Tollis Formation (El Sherana Group), a Formation more characterised by the presence of volcanics such as cherty tuft, crystal tuft, pitchstone, andesite, and altered mafic and intermediate volcanics (Needham & Stuart-Smith, 1985a). It was generally accepted that the Batman deposit lay on the western edge of a peninsula shaped inlier of Tollis Formation that unconformably overlay, and was surrounded by greywacke, sandstone, siltstone and shale units of the Palaeoproterozoic Burrell Creek Formation (Finniss River Group). The position of the unconformity was sited from aerial photographs of the region; east and north of Mount Todd and immediately west of the Batman deposit (Needham et al., 1989) (Figure 3.9).

During the course of this study, the Mt Todd goldfield was extensively mapped and a stratigraphy based on lithofacies was established (Section 3.5). Special attention was paid to locating the Burrell Creek Formation-Tollis Formation boundary and several traverses were conducted across, and along, the recorded position of the unconformity. The results of the investigations do not support an argument for an unconformity in the Mt Todd goldfield.

3.4.2 Position
In the Mt Todd goldfield, bedding and/or lithofacies are not truncated along strike, except where faulted, and a distinct palaeo-erosional surface has not been observed as might be expected if an angular unconformity were present. West of the Batman deposit, hornfelsed sandstone and siltstone beds parallel or transect the recorded position of the unconformity,
as do siltstone and shale beds east of Mount Todd. The recorded northern position of the unconformity is obscured by regolith, and volcanic or volcanoclastic beds have only been observed at the southern margin of the goldfield adjacent to the Edith Falls Road. In this location greywacke beds are conformably overlain by tuffaceous greywacke, and subsequently, by tuffaceous gritstone and crystal tuff (Section 3.3). The contact between clastic rocks and volcanoclastic rocks is gradational.

On the basis of these observations, it is concluded that the boundary between the Burrell Creek Formation and Tallis Formation lies at the southern margin of the goldfield and is conformable in nature (Figure 3.10). Based on a similarity with lithologies elsewhere in the Pine Creek Inlier it is furthermore concluded that the rocks which host gold mineralisation at the Batman deposit belong to the Burrell Creek Formation.

### 3.4.3 Implications

Several implications follow from the above conclusions. Firstly, if the Burrell Creek Formation and the Tallis Formation are conformable, then the Tallis Formation must constitute the youngest formation in the Palaeoproterozoic Geosynclinal Sequence, i.e., it must be part of the pre-orogenic sequence of the Finniss River Group and not part of the syn- to post-orogenic El Sherana Group (Section 2.1.3 & 2.1.4). This is consistent with geochronological studies by Jagodzinski (1992) in which it was concluded that a discrepancy in age between the Gimbat Ignimbrite Member (1831 ± 2 Ma, U-Pb age) of the Pul Pul Rhyolite and the Tallis Formation (1890 ± 15 Ma, U-Pb age; Page and Williams, 1988) was the result of an incorrect correlation between the Tallis Formation and the overlying formation to the rhyolite, the Big Sunday Formation. She noted that the age of the Tallis Formation was consistent with a best estimate age for the South Alligator Group (Section 2.1.3) at 1885 ± 2 Ma (U-Pb age; Page & Williams, 1988) and recommended that the Tallis Formation be placed at the top of the Burrell Creek Formation. Her revised stratigraphy for the Geosynclinal Sequence is adopted here (Figure 3.11).

Secondly, in corollary, the 1890 ± 15 Ma U-Pb age of Page and Williams (1988) for the Tallis Formation provides an age constrain to sediment deposition in the upper portion of the Geosyncline Sequence.

### 3.5 Lithofacies Stratigraphy

In order to facilitate structural mapping, a stratigraphy was established for the goldfield based on 16 conformable lithofacies (definition from Bates & Jackson, 1987). The distribution of the 16 lithofacies, and a lithofacies column representing a greater than 2000 m stratigraphic section, is presented in Figure 3.12a & b in Appendix A. There are no distinctive marker
beds and the boundary between lithofacies are gradational. There is a broad, and perhaps imprecise, correlation between topography and lithofacies such that competent lithofacies crop out on hills and ridges, in contrast to less competent lithofacies in valleys.

3.6 DEPOSITIONAL ENVIRONMENT

The sedimentary structures observed throughout the goldfield indicate that deltaic processes and turbidite activity were operating conjointly during deposition of the Burrell Creek Formation. Beds containing sedimentary structures which are deltaic in origin, such as mud balls (Conybeare & Crook, 1968), fine scale cross beds in shales and chert, and asymmetric current ripples in siltstones and some greywackes, are intercalated with beds deposited by turbidity currents, ie., greywacke and siltstone units which exhibit graded bedding and part or all of the Bouma sequence in association with dewatering structures, base sharpness, rare scour and load casts, and discontinuous and continuous sandy/silty laminations (Walker, 1984). It is therefore envisaged that deposition of the Burrell Creek Formation occurred at the margin of an ocean basin in a delta slope to prodeltaic environment of a river dominant-delta system (eg., the Mississippi delta).

In this environment the influence is ostensibly fluvial, and waves, long shore currents and tidal currents are weak (Miall, 1984). Sediments are evenly distributed across the delta front and prodelta through avulsion, and massive, laterally extensive beds of poorly sorted mudstone and siltstone are produced which are intercalated with coarser siltstone and sandstone beds (Miall, 1984; Elliott, 1986). The sedimentation rate is rapid and progradation of facies may result in overstepping of the delta front, slumping and turbidity flows (Elliot, 1984; Sangree et al., 1992; Mitchum et al., 1993). Sediments which are deposited on the delta front are often reworked and emplaced in the prodeltaic environment (Miall, 1984; Mitchum et al., 1993). The sedimentary structures produced are thus very similar to those exhibited by marine turbidite deposits, but include structures associated with deltaic processes, as manifest in the Mt Todd goldfield.

3.7 IGNEOUS ROCKS

3.7.1 The Yenberrie Leucogranite

The Yenberrie Leucogranite (1835-1820 Ma) is a medium to coarse grained, two mica leucogranite (Figure 3.13) that crops out as two discrete cupolas centred approximately 3 km NW of Batman Hill. The leucogranite is in non-conformable contact with the Burrell Creek Formation and is extensively greisenised and veined. It contains numerous meta-sedimentary xenoliths (Figure 3.14) and occasional meta-igneous xenoliths, is deformed and fractured. A spaced, stylolitic to anastomosing cleavage transects the pluton (Figure 3.15).
In thin section, the leucogranite consists of subhedral to anhedral quartz (0.5-1.5 mm), pale brown to green-brown biotite (1-3 mm), pink to yellow muscovite (1-5 mm), anhedral to subhedral zircons (30 µm), subhedral sericitised laths of orthoclase/microcline (to 2 mm), andesine/oligoclase (Ab$_{28-34}$-An; 300-500 µm) and accessory apatite. Quartz, orthoclase/microcline and andesine/oligoclase crystals typically exhibit concavo-convex to serrate margins, and many biotite plates are chloritised. Some quartz crystals have undulose extinction and extensive subgrain development, and may be micro-fractured. Potash feldspars exhibit weak tartan twining. The pervasive tectonic fabric is defined by numerous Fe-oxide stained stylolitic and anastomosing fractures and a weak alignment of platey minerals.

A significant proportion of the leucogranite is greisenised. The greisens consist of an interlocking assemblage of coarse grained polygonal quartz (1-5 mm) and pink, pale yellow and clear muscovite (0.5-3 mm), in association with adularia (300-500 µm), masses of purple fluorite (20-30 µm), and rare lepidolite and oligoclase (100 µm). Clots of quartz, fluorite, epidote, muscovite, lepidolite and rare topaz mark the sites of altered country rock xenoliths. Primary biotite is generally altered to an aggregate of muscovite, chlorite, epidote and/or fluorite, and commonly retains relict zircons. Euhedral to subhedral rhombs of adularia (0.25-3.0 cm in length) overprint and replace orthoclase. Wolframite and molybdenite are disseminated through parts of the greisen. Wolframite and molybdenite, in association with cassiterite, bismuth sulphosalts, arsenopyrite, pyrite, copper sulphides and muscovite have been found in northwest and north-northeasterly trending quartz veins (30 cm in width) that cross-cut the greisens (Rattigan & Clark, 1954/55; Walpole et al., 1968). The veins and greisens are collectively known as the Yenberrie Wolfram Field and were worked for tungsten, molybdenite and bismuth from 1911 to 1919, and 1950 to 1953 (Rattigan & Clark, 1954/55; Walpole et al., 1968).

3.7.2 The Tennysons Leucogranite
The Tennysons Leucogranite (1835-1820 Ma) is a massive, porphyritic, medium to coarse grained, biotite leucogranite (Figure 3.16). It crops out on the western margin of the study area in fault and non conformable contact with the Burrell Creek Formation. The leucogranite contains numerous meta-igneous xenoliths (Figure 3.17) suggesting it is I-type in character (Wyborn, 1988), and is not compositionally zoned on a regional scale. The pluton is dissected by numerous parallel to sub-parallel or anastomosing NW trending shear faults of the Pine Creek Shear Zone (Section 2.2.3) which have modified primary igneous textures.

In thin section, the leucogranite consists of euhedral to subhedral laths of microperthitic microcline phenocrysts (to 3-5 cm) in a groundmass of green-brown pleochroic biotite (400 µm-3 mm), subhedral laths of andesine (Ab$_{38}$-An; 1-5 mm), anhedral quartz (0.25-5 mm), microperthitic microcline (1.5 mm), accessory zircon (120 µm) and apatite. Phenocrysts may
comprise between 15-30% of the total rock volume. Microcline phenocrysts may cluster as glomerocrysts up to 10 cm in diameter. Large biotite plates may be sub-poikiloblastic with chadacrysts of quartz, microcline and andesine (100-200 µm). Quartz crystals exhibit undulose extinction with subgrain development, and may be rimmed by fine polygonal quartz (5 µm). A concavo-convex to serrate texture characterises the margins of quartz, microcline and andesine crystals. Some twins in microcline and andesine feldspar are kinked, and biotite is partly chloritised.

The leucogranite has undergone post-crystallisation hydrothermal modification, and greisens, quartz-chlorite veins and quartz-hematite lodes are developed. Where greisens have formed, the leucogranite has altered to a quartz-muscovite-sericite-chlorite assemblage that contains minor bundles and rays of green euhedral tourmaline (0.75 mm long). Where quartz-hematite lodes have formed, the granite has been replaced by an assemblage of hematite, muscovite and quartz. Randal (1963) noted the presence of autunite, meta-autunite and torbernite. The lodes are enriched in uranium, and were mined on a small scale at the Yenberrie Mine, YMCA shaft No 1 and YMCA shaft No 2 (Figure 1.2).

3.7.3 Porphyritic rhyolite dyke
A porphyritic rhyolite dyke (1835-1820 Ma) crops out intermittently over a strike distance of 2.5 km, between the Mt Todd Battery and Batman Hill. The dyke is 2-4 m in width and intrudes rocks of the Burrell Creek Formation. It is fractured and veined, and cross-cut by a spaced cleavage.

In hand specimen, the dyke consists of subhedral to euhedral laths of orthoclase (5-12 mm), subhedral laths of andesine (Ab_{35-40}-An; 1-3 mm), and partially reabsorbed equant polygons of quartz (1 mm) in a fine groundmass (Figure 3.18). Orthoclase phenocrysts are zoned, finely perthitic, and partially sericitised. They may cluster as glomerocrysts. In thin section, the groundmass is composed of quartz (10-15 µm), orthoclase (20 µm), andesine (20 µm), chlorite after biotite (40 µm), and accessory zircon.

The dyke is fractured and altered, and a network of fine chlorite-sulphide veins that define a fabric element are characteristic (<1 mm). Anhedral sulphides grains are distinctly embayed, and include pyrite, chalcopyrite, sphalerite, galena, arsenopyrite, and occasional pyrrhotite. Chlorite is vermiform and spherulitic (1 mm). The dyke has been sampled along strike and assayed for Au, Cu, Pb, Zn, Sn, Bi, and As. Analyses indicate that the dyke is not prospective for the elements assayed (Table 3.1).
Figure 3.1: Graded bedding in a hornfelsed greywacke from a road cut along the Mine Access Road. The lower portion of the rock comprises CDE of the Bouma sequence in which D and E are repeated. The coarser upper portion of the rock comprises part A of a second turbidite cycle (Stow Creek MG 6858N 11250E).

Figure 3.2: Dewatering structures in a hornfelsed greywacke (facing is up). Pencil is 12 cm long (Batman and Robin MG 11155N 8200E).

Figure 3.3: Convoluted bedding, ripples, scour and load structures in a massive hornfelsed greywacke exposed in a road cut along the Mine Access Road (Stow Creek MG 6858N 11250E).
Figure 3.4: Mud balls in a bed of massive siltstone exposed in the north face of the Golf pit. The mud balls comprise alternating layers of mud and silt which formed when clots of sticky mud were rolled by a current over a silty surface (Conybeare & Crook, 1968). As the clots rolled they became rounded and increased in size through accretion of detritus. They are usually found in fluviatil to marine littoral environments: their association with graded bedding, ripples, laminations, convoluted laminations and slumps, and load and scour structures are interpreted to mean that deposition occurred in a shallow marine and perhaps deltaic environment. The mud balls are flattened parallel to the regional S₁ and S₂. Pencil is 8 mm wide.

Figure 3.5: Convoluted laminations in a chloritised shale (Mt Todd MG 9020N 11000E).

Figure 3.6: Graphitic chert: laminated and finely cross-bedded (Sample T001, North Batman MG 11400 N 9240E).
Figure 3.7: Laminated and finely bedded crystal tuff of the Tollis Formation (Stow Creek MG 6850N 11350E).

Figure 3.8a: Tuffaceous greywacke of the Tollis Formation. The greywacke is characterised by pink spots that are dominated by potash feldspar. (Stow Creek MG 6855N 11325E.)

Figure 3.8b: Tuffaceous greywacke of the Tollis Formation. The pink spots in the greywacke are composed predominantly of potash feldspar which has a locus on 4-6 cm long, flattened clasts of crystal tuff or graphitic chert. (Stow Creek MG 6855N 11325E.)
Figure 3.9: The position of the unconformity (Burrell-Creek Formation-Tollis Formation boundary, BCF/TF boundary) as sited from aerial photographs of the Mt Todd goldfield, after Needham et al. (1989).

Figure 3.10: The position of the Burrell Creek Formation-Tollis Formation boundary as determined by this study. The boundary is conformable in nature.
**Figure 3.1.1:** Stratigraphic column for the Central Domain of the Pine Creek Inlier, as adopted in this study. Modified after Jagodzinski (1992) and Stuart-Smith (1987).
Figure 3.13: The Yenberrie Leucogranite.

Figure 3.14: Meta-sedimentary xenolith in the Yenberrie Leucogranite (Yenberrie Field MG 11900N 5450E).

Figure 3.15: Spaced, stylolitic to anastomosing cleavage in the Yenberrie Leucogranite. Pencil is 12 cm long (Yenberrie Field MG 11150N 5800E).
Figure 3.16: The Tennysons Leucogranite.

Figure 3.17: Meta-igneous xenoliths in the Tennysons Leucogranite. The sample was collected along the Stuart Highway, 6 kilometres north of the junction with the Edith Falls Road.

Figure 3.18: Porphyritic rhyolite dyke (Batman and Robin MG 9440N 8700E).
Table 3.1

Assay results for 10 samples of a porphyritic rhyolite dyke that contains a network of fine chlorite-sulphide veins. The samples were collected at regular intervals along the 2.5 km strike length of the dyke.

<table>
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<th>SAMPLE NOS.</th>
<th>Au ppm</th>
<th>Cu ppm</th>
<th>Pb ppm</th>
<th>Zn ppm</th>
<th>Sn ppm</th>
<th>Bi ppm</th>
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<td>18</td>
<td>18</td>
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<td>143</td>
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<td>210</td>
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<td>51</td>
<td>12</td>
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<td>80</td>
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<td>8</td>
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<td>38</td>
<td>16</td>
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</table>
CHAPTER 4

STRUCTURE AND METAMORPHISM

4.1 PREAMBLE

The structure and metamorphism of the Mt Todd goldfield has been considered in several company reports by Kenny et al. (1989), Taylor (1990a, 1990b), England (1990), Kenny et al. (1990), Hein (1991a, 1991b, 1991c, 1992), and an honours thesis by Koerber (1989). This study differs from those considerations in providing a fuller analysis and description of the structural and metamorphic features of the goldfield, and is focussed toward understanding the:

-- The styles of deformation.
-- The intra-relationship of the several vein types developed in the goldfield.
-- The relationship between folds, faults and veins.
-- The nature of contact metamorphism, and the role of granite emplacement.
-- The structural and metamorphic history of the goldfield with specific emphasis on the timing of the gold mineralising event.

To that end the Mt Todd goldfield has been extensively mapped at 1:5000 scale by the author in four field seasons conducted between 1990-1994. The data collected has been recorded on 1:5000 scale fact maps of Horseshoe Creek, Horseshoe Creek Tin Field, Yenberrie Field, North Batman, Quigleys, West Batman, Batman and Robin, Mount Todd, Edith River Crossing and Stow Creek (Figures 4.1 to 4.10). These maps constitute a database for the following discourse and are presented in Appendix B. The main structural elements of the maps are presented in Appendix C on a 1:15,000 scale map entitled "A structural interpretation of the Mt Todd goldfield" (Figure 4.11), and are summarised on a 1:50,000 scale map entitled the "Regional structural map of the Mt Todd goldfield" (Figure 4.12). A schematic section along mine grid northing 11000N is presented in Figure 4.12b.

4.2 STRUCTURE

4.2.1 Introduction

The rocks of the Mt Todd goldfield have undergone two phases of brittle-ductile deformation, D₁ and D₂, and at least one phase of brittle deformation, D₃. D₁ resulted in the development of north-south trending megascopic and mesoscopic folds (F₁) and an axial
planar cleavage ($S_1$) restricted to the hinge of folds, while $D_2$ resulted in the development of megascopic folds ($F_2$) and a spaced disjunctive cleavage in transection to the folds ($S_2$). $D_3$ involved displacement along several strike-slip faults with the concomitant development of a steeply dipping $S_3$-$C$ type foliation and mesoscopic en échelon folds ($F_3$) trending oblique to the strike of the faults.

4.2.2 Folds
Folds formed during $D_1$ are non-cylindrical, asymmetric and open to tight in profile. They generally plunge to the south although plunge reversals occur adjacent to northwest trending strike-slip faults and where $F_1$ hinges are refolded by $F_2$. The eastern limb of $F_1$ is typically steep or overturned, the asymmetry suggestive of an eastward direction to tectonic transport during $D_1$. They have an average half wavelength of 4 km (Koerber, 1989).

The folds are discontinuous along the fold axis trend and overstep to the left (Figure 4.11). Parasite folds may be developed on the long limb or hinges of $F_1$ folds. The concurrence of parasitic folds with overstepped folds may result in a complex fold hinge such as that which occurs south of Jones Brothers Workings where a left-handed fold-overstep is subtended by parasitic folds on the long limb of the western fold and between the hinges of the western and eastern fold (Figure 4.7 & 4.11, MG 11000N 10000E).

The axial trace of $F_1$ is strongly curved from northeast in the north of the goldfield, north to northwest in the centre of the goldfield, and northeast in the south of the goldfield (Figure 4.11). The change in trend of fold hinges between the centre and south of the goldfield is the result of refolding of $F_1$ about an $F_2$ syncline. The change in trend of the fold hinges between the north and centre of the goldfield occurs about the Gotham City Fault (Figure 4.12) and is the result of a variation in attitude of fold axes, i.e., south of the Gotham City Fault fold axial planes dip steeply west, while north of the Gotham City Fault they dip both east and west.

$F_1$ have an associated axial planar cleavage ($S_1$) which is well developed at the hinge of folds, but weakly developed or absent on the fold limbs. It has not been observed in rocks which have been hornfelsed to hornblende-hornfels facies, indicating that $S_1$ developed prior to hornfelsing of the sediments. In shale beds, $S_1$ occurs as a continuous, fine, parallel to anastomosing fabric which is predominantly defined by the alignment of platey minerals. In siltstone and greywacke beds, $S_1$ is typically non-penetrative and discontinuous, and generally coarse and anastomosing. It is defined by flattened quartz clasts and/or by aligned platey minerals.

A stereographic projection of poles to $S_1$ is presented in Figure 4.13a. $S_1$ strikes north-northeasterly and dips steeply west. The spread of $S_1$ poles around the primitive circle results from the change in strike of the axial plane of $F_1$ throughout the goldfield, from north to south.
Stereographic projections which consider the S1 data in terms of domains of equal strike, are presented in Figure 4.13b-e and four domains are defined. In the northern domain (north of the Gotham City Fault), S1 has a range of orientations that define a weak girdle, indicating that F1 axes are sub-vertical (Figure 4.13b). The spread of points about the primitive circle is the result of refolding of S1 about F2, and an anticlockwise rotation of fold hinges adjacent to northwest trending strike-slip faults. The sense of displacement on these faults is sinistral or dextral with offsets of approximately 250 metres. In the Golf domain, which incorporates the region around the Golf pit, S1 strikes slightly east of north (Figure 4.13c), while in the Robin domain (the region incorporating the Robin prospect) S1 strikes northwest (Figure 4.13d). In the Access Road domain, which is located south of the North Batman Fault (Figure 4.11), S1 strikes northeast and is overturned to the east with a moderate to steep dip to the west (Figure 4.13e).

The plunge of folds, as determined from the intersection of bedding and S1, has been determined for each domain and is presented in Figure 4.14a-d. In the northern domain, the fold plunge is predominantly toward the south-southwest although local plunge reversals occur where F2 has refolded F1 (Figure 4.14a). The attitude of the F1 varies considerably from shallowly north and south plunging, to steeply plunging. The spread of data about the primitive circle toward the south-southeast is the result of anticlockwise rotation of fold hinges adjacent to northwesterly trending strike-slip faults. In the Golf and Robin domains, the plunge of F1 is southerly trending about the Golf Pit and south-southeasterly trending at Robin, and is generally shallow (Figure 4.14b & 4.14c). Plunge reversals may be attributed to slight changes in the orientation of S1 with respect to bedding, because bedding and S1 are sub-parallel in strike throughout these two domains. In the Access Road domain, the plunge of F1 is predominantly toward the south-southwest and is shallow to moderately steep (Figure 4.14d).

The general north-south orientation of D1 structures indicate that the maximum compressive stress (σ1) operating during D1 was orientated east-west.

4.2.3 D2

Folds formed during D2 are asymmetric, rounded and open in profile, with an estimated half wavelength and amplitude of 6 km and 1.5 km, respectively. They strike roughly east-west and are generally less than 3 km in strike length. The superposition of F2 on F1 has resulted in a Type 2 interference pattern (Ramsay, 1967) which is characterised by symmetric to non-symmetric, crescent shaped domes and basins (Thiessen & Means, 1980).

F2 is associated with a pervasive, transecting, spaced or fracture cleavage (S2). The fabric is east-west striking at the hinge of folds but north-westerly striking on fold limbs (Figure 4.15). It dips steeply south or southwest: a north to northeasterly direction to tectonic transport is
indicated. The transecting nature of $S_2$ may indicate that it formed before $F_2$ (Gray, 1981; Johnson & Woodcock, 1991). The fabric refracts convergently across quartz veins and competent rock types.

$S_2$ is disjunctive because it cuts across $S_1$, and there is no sense of continuity or relationship between $S_1$ and $S_2$ (Powell, 1979). In shale beds, it occurs as a continuous, finely spaced (<0.01 mm), parallel to anastomosing fabric which is defined by an alignment of platey minerals and by a flattening of detrital grains. $S_2$ is a regular, parallel array of joints where it traverses massive beds of greywacke or siltstone, and in rocks which have been contact metamorphosed to hornblende-hornfels facies. In the latter, the joints divide the rocks into a series of blocks and tabular microlithons that may ascribe a tombstone-like topography. The occurrence of $S_2$ within the hornfelsed aureole that surrounds the local granites indicates that hornfelsing was pre- or syn-$S_2$.

Overall $S_2$ has a northwesterly strike that dips steeply southwest (Figure 4.16a, b). The spread of data about the primitive circle reflects the difference in strike of $S_2$ between the fold hinge and limbs, refraction of $S_2$ between differing rock types, and minor rotation of $S_2$ into northwest trending strike-slip faults.

The attitude of $F_2$ is eastward in the west of the goldfield within the zone of hornblende-hornfels rocks and adjacent to the local granites, and westward in the east of the goldfield. The plunge is shallow in the north and steep in the south. An overall basin geometry is developed. The eastward plunge in the west of the goldfield may reflect fold drape against granites or doming in association with granite emplacement.

The general east-west trend of $D_2$ structures indicates that the principal compressive stress ($\sigma_1$) operating during $D_2$ was orientated roughly north-south.

4.2.4 $D_3$

A period of brittle deformation, $D_3$, has been recognised from offset of $F_1$ and $F_2$ hinges, and inferred from horizontal rotation of bedding, $S_1$ and $S_2$ along northwest trending strike-slip shear faults (e.g., Figure 4.11, Yenberrie Shear Zone, Gotham City Fault and Horseshoe Fault). The shear faults exhibit sinistral or dextral offset, the magnitude of which is less than 250 meters. They are associated with mesoscopic folds ($F_3$) that trend oblique to the strike of the faults in an en échelon array. The array is generally left stepping indicating a predominant sinistral strike-slip movement on the faults (Sylvester, 1988).

The shear faults are wide zones of disruption (<25 meters) or brecciation that are bound by discrete fault margins. The shear faults are intensely weathered and occasionally decorated by a network of fine quartz-chlorite veins (<2 mm). In the region of hornfelsed rock, the
zones are associated with a profound buff-red weathering and a mesoscopic $S_3$-$C$ fabric in which the $C$-surfaces define the dominant foliation. The fabric is characterised by the development of jagged, lenticular slabs of rock that are positioned oblique to, and along, the strike of the fault zone (Figure 4.17). The long axis of the slabs is orientated sub-vertically. The fabric is similar in nature to that described by Lister and Snoke (1984) for Type II $S$-$C$ mylonites. However, it is low grade in the development of mylonite. The development of the fabric suggests that ductile deformation accompanied block dislocation.

4.2.5 Veins

Five vein types are recognised in the Mt Todd goldfield:

- $V_1$ - Buck quartz veins.
- $V_2$ - Quartz-tourmaline veins.
- $V_3$ - Quartz-sulphide veins and lodes.
- $V_4$ - Calcite-base metal veins.
- $V_5$ - Epithermal style quartz veins.

Type $V_1$ veins are early and are cross-cut by $V_2$ and $V_5$; $V_2$ and $V_3$ are syn-genetic and cross-cut by $V_4$; $V_5$ are genetically late. Type $V_3$ are associated with gold mineralisation in the goldfield and are the focus of current mining operations at the Batman deposit.

4.2.5.1 Buck quartz veins ($V_1$)

Buck quartz veins are preferentially distributed in siltstone and greywacke-dominant lithofacies, near the hinge of $D_1$ folds and adjacent to faults, and are abundant outside the zone of hornblende-hornfels facies rocks. They range in size from centimetre scale veinlets to large single blows (Figure 4.18), the largest measuring 350 m long by 25 m wide.

They are, however, more commonly developed as small lenticular bodies (3 m x 3 m x 0.5 m) that are undulating in form (Figure 4.19). They may form conjugate vein sets, random parallel vein sets, or en échelon vein arrays. They occur as breccia fill and ladders (up to 4 m long and 30 cm wide) and may occupy faults (third order). The veins are cross-cut by $S_1$, $S_2$ and quartz-sulphide lodes and veins at the Quigleys deposit indicating that buck quartz veins formed prior to the development of the $S_1$ and $S_2$ fabrics. The veins are not associated with economic mineralisation.

The veins are composed of massive, anhedral, white and grey (rare) quartz. In thin section, the quartz is heavily decorated with large CO$_2$-rich fluid inclusions (5-100 µm in size) that are cross-cut by numerous trails of tiny H$_2$O fluid inclusions (< 3 µm in size).

The geometry of the buck quartz veins in the Access Road domain (south of the Batman North Fault) and North domain (north of the Gotham City Fault) have been studied: it has been established that the veins occur in two orientations that form a conjugate set.
Type 1 strikes northeast and dips moderately southeast, while type 2 strikes northwest and dips steeply southwest (Figure 4.20a, b, c). The acute angle between the vein types is 59° in the North domain and 56° in the Access Road domain and their intersection plunges moderately to the south-southeast. The veins are associated with offset of the bedding in the plane of the vein. For type 1 veins this offset is dextral, while for type 2 veins the offset is sinistral.

The principal palaeo-stress directions ($\sigma_1$, $\sigma_2$ and $\sigma_3$) that operated at the time of formation of the buck veins can be deduced from the geometry of the vein types. The orientation of $\sigma_2$ is parallel to the intersection line of the vein types (Hobbs et al., 1976) and perpendicular to the $\sigma_1\sigma_3$ plane. $\sigma_1$ should lie parallel to the bisector of the acute angle between the vein types (Figure 4.20b, c). From this, the following palaeo-stress orientations have been established for the domains: $\sigma_1$ plunged gently toward the west-southwest; $\sigma_2$ plunging moderately south-southeast; $\sigma_3$ plunging moderately north. The $\sigma_1$ direction is sub-normal to the local S$_1$ in each domain.

The modelled stress directions for vein type 1 and 2 in the North and Access Road domains are approximate to the stress directions for $D_1$, as indicated from the attitude of $F_1$ (Section 4.2.2), and it is argued that the vein orientations in those domains were the result of deformation during $D_1$. It is also contended that the buck quartz veins in those domains were formed during $D_1$. Furthermore, because the veins are cut by S$_1$, it is considered that they formed early in that deformation.

Lastly, the orientations are coincident with third order conjugate faults. Since conjugate sets of strike-slip faults form in pure shear (Sylvester, 1988), and the faults are occupied by conjugate sets of buck quartz veins, it has been concluded that the faults and veins formed in a coaxial stress field.

4.2.5.2 Quartz-tourmaline veins ($V_2$)
Veins composed of quartz and tourmaline crop out in massive beds of greywacke on Mount Todd and atop several steep hills 750 m northwest of the Quigleys North pit (QNP). The veins have also been identified by Taylor (1990a) in diamond drill core from the Batman deposit. The veins cross-cut buck quartz veins ($V_1$) and occur as a parallel set of vertical sheets that are irregularly spaced apart. Individual veins are continuous in strike ($\pm 25$ m) and range in width from 1 mm to 1.0 cm. They strike northwest with a mean strike and dip of 316° 89° W (Figure 4.21a). The veins are not associated with economic mineralisation.

In thin section, the veins are largely composed of anhedral quartz fibres (1.0 mm), anhedral to euhedral green to brown coloured tourmaline fibres (500 µm long) and rare biotite plates (200µm). Tourmaline occurs a discrete masses, bundles, or as a network of fine needles.
A narrow bleached alteration selvedge (1-2 mm wide) is composed of quartz, tourmaline, rutile, biotite, goethite after sulphide, and rare aggregates of pyrrhotite and chalcopyrite.

The formation of quartz-tourmaline veins preceded sulphide precipitation quartz-sulphide veins ($V_3$), but both vein types are syn-genetic (Chapter 6). Quartz-tourmaline veins are cross-cut by calcite-base metal veins ($V_4$).

4.2.5.3 Quartz-sulphide veins and lodes ($V_3$)

Gold mineralisation in the Mt Todd goldfield is associated with the development of quartz and sulphide minerals in discrete vein sets and lodes: the term "lode" is used to describe a zone of proto-cataclasite or cataclasite as detailed by House and Gray (1982). The location of the vein sets and lodes is presented in Figure 1.3 and the style of veining at each location, host lithology, orebody geometry (where known), production figures, and estimated reserves are presented in summary in Table 4.1. The type of testing and best interval intersected for each location is indicated where a reserve has not been estimated or assessed.

The vein sets and lodes in the goldfield strike predominantly north-northeast. They are hosted in greywacke or siltstone-dominant lithofacies and are located outside, or at the margin of the zone of hornblende-hornfels facies rocks. Their distribution across the goldfield resolves a broad northeast trending corridor that is coincident with a northeast trending magnetic lineament (Kenny et al, 1990). This lineament is believed to reflect a deep crustal structure that may be related to fracturing in the basement (Koerber, 1989).

As indicated in Table 4.1, gold-bearing vein sets occur at the Batman deposit, Quigleys deposit, Golf pit, Penguin prospect, New Tollis prospect, Regatta East prospect, Horseshoe prospect, Q3-Q4 costeans, and Quigleys West prospect (Figure 1.3). At these locations, vein sets consist of sheeted or stockworked veins. They range in width from less than a millimetre to 10 cm and are associated with an alteration selvedge of biotite, muscovite, sericite, chlorite, rutile and accessory tourmaline that is 1-3 times the width of the vein. The veins cross-cut the regional $S_1$ fabric and are cross-cut and deformed by the regional $S_2$ fabric. Sheeted veins form a parallel to anastomosing array that is generally discordant to bedding: they are discontinuous in strike and terminate as joints. Stockwork veins form an anastomosing or orthogonal array which may be concordant or discordant to bedding. The stockwork vein array may define a zone of brecciation in the footwall to a lode, or interconnect adjacent quartz-sulphide lodes.

The veins are dominated by the development of fibre quartz in association with pyrrhotite, chalcopyrite, arsenopyrite, pyrite, marcasite, löllingite, ISS cubanite, galena, sphalerite, bismuth, bismuthinite, talnakhite ($Cu_9Fe_6S_{10}$), Bismuth-rich galena, pavonite ($Ag_2S.3Bi_2S_3$).
hedleyite (Bi₁₄Te₆), and gold. This assemblage is developed as a two-stage fill: early fibre quartz lines the vein walls as a selvedge, and euhedral quartz and sulphide occur as a late fill. The ratio of early quartz to late fill is highly variable within any given vein set, or between discrete vein sets, as is the ratio of euhedral quartz to sulphide in late fill. Furthermore, the type of sulphide and amount present is variable. The nature of these variations, and the petrology and mineralogy of the veins is presented further in Chapter 6, along with a comprehensive paragenetic sequence of mineralisation.

Gold-bearing lodes occur at the Quigleys deposit, the Golf pit, Regatta prospect, Hotel, Old Tollis, Jones Brothers Workings, Quigleys West prospect, Alpha, Bravo, Charlie and Delta pits, and at MG 7700N 8350E. The latter has not been sampled. In outcrop, the lodes comprise a massive to vuggy proto-cataclasite or cataclasite that consists of silt to boulder-sized, intensely micro-fractured clasts of quartz and quartz/Fe-oxide set in a quartz and red-brown-blue coloured Fe-oxide matrix. The lodes are lenticular in form and most occupy faults that are concordant to bedding. The largest, at the Quigleys deposit, exceeds 350 m in strike length and ranges in width from 30 cm to 1.5 m. The biotite-chlorite-sericite alteration selvedge associated with this lode extends a metre into the footwall, and up to 3 metres into the hanging wall.

The mineralogy for the lodes is similar to that for veins perhaps indicating that they formed about the same time. Lode mineral texture is strongly modified by shearing and brecciation and a complex inter-relationship of cross-cutting fractures is developed. The nature of these deformation textures is outlined in Chapter 6.

In further evaluation of the structural setting of quartz-sulphide veins and lodes, two deposits have been selected for detailed structural study, the Batman deposit and the Quigleys deposit. This study is presented in Chapter 5.

4.2.5.4 Calcite-base metal veining (V₄)
Calcite-base metal veins have been observed in diamond drill core and in outcrop at the Batman deposit, Golf pit and Quigleys deposit. At the Batman deposit the veins occur within fault-fractures that strike northwest and northeast with the predominant strike to the northwest (Figure 4.22a). The veins (and host fault-fractures) range in width from <1 mm to 3 cm, but are generally less than 3 mm wide. They are spaced approximately one metre apart. Host faults are dextral or sinistral strike-slip. They offset quartz-sulphide veins (V₃) with maximum throw of 20 cm, thus indicating that calcite-base metal veins postdate quartz-sulphide veins. In thin section, the veins are observed to also cross-cut the S₂ fabric, indicating they formed after the development of that fabric element.
Calcite-base metal veins are composed of cockscomb and colloform calcite with minor euhedral quartz that is syntaxial to the vein-wall interface. Euhedral to anhedral galena, sphalerite, pyrite, arsenopyrite, and chalcopyrite occur as vein fill and may exhibit herringbone or colloform texture (Figures 4.22b & 4.23a,b). Spherulites of calcite, sphalerite inclusions in coarse rhombs of calcite (<5 mm), and calcite-sphalerite-chalcopyrite-pyrrhotite cauliflower texture have also been observed. Chlorite and rutile occur as rare aggregates in this assemblage. The presence of adularia, albite, fluorite and apatite as accessory minerals has been noted by England (1989) and Taylor (1990a).

Due to their limited size and distribution, the veins are not of economic significance.

4.2.5.5 *Epithermal style quartz veins* (*V*<sub>5</sub>)

Several large epithermal quartz veins crop out along the northerly trending ridge line of the Robin prospect, in cross-cutting relationship with buck quartz veins 300 m southeast of the Golf pit, and at a site 800 m north-northwest of the Golf pit. The vertical veins strike northerly, and measure 5-10 m in length and 3 cm to one metre in width.

In these veins, loose or densely packed layers of reniform or chalcedonic quartz form a substrate to fine euhedral crystals of quartz (0.1-0.5 mm) (Figure 4.24). The layers may be arranged into a rhombohedron, and Fe-oxides may occur as fill between layers (Figure 4.25). The textures are similar to those associated with epithermal mineralisation, as described by Lindgren (1933), and are interpreted as replacement of epithermal calcite by quartz and Fe-oxide.

The presence of epithermal style quartz veins may signify a potential for epithermal mineralisation in the Mt Todd goldfield.

4.2.6 *Faults*

4.2.6.1 *Introduction*

Faults have been difficult to locate in the goldfield due to the lack of distinctive stratigraphic marker beds in the Burrell Creek Formation. Nevertheless, a number of faults were identified that varied in type, strike length and magnitude of offset (Figure 4.11). The criteria used to establish the location of faults was as follows: they were identified from offset of bedding (where beds could be reasonably correlated), lithofacies, fold axes and quartz-sulphide veins and lodes, and were implied where quartz slickenside fibres and S<sub>3</sub>-C type foliations were developed. Some were recognised in outcrop from fault gouge, breccia and buck quartz veins in the fault plane, or where buck quartz veins extended away from an observed fault in a linear array parallel to the strike of the fault. Others were inferred from topographic irregularities such as an abrupt termination of a prominent ridge, or a significant horizontal shift in the ridge line. The direction of fault offset was established from facing of accretion
steps and extension fractures in association with quartz slickenfibres, and where a reasonable correlation between rock types, lithofacies and fold axes could be established.

4.2.6.2 First-order faults

First-order faults are northwesterly trending strike-slip shear faults that comprise the Pine Creek Shear Zone: they are located west of the Mt Todd goldfield (Section 4.2.8). The faults tens of kilometres in strike length and are associated with sinistral displacements of the order of hundreds of metres, eg., the Phillips Creek Fault (Section 2.2.3).

4.2.6.3 Second-order faults

The largest faults within the Mt Todd goldfield are second-order northwesterly trending, dextral and sinistral strike-slip faults, ie., the Yenberrie Shear Zone, Batman North Fault, Gotham City Fault, Regatta Fault, Quigleys Fault and Horseshoe Fault (Figure 4.11). These faults exceed three kilometres in strike length and are associated with the development of an S3-C type foliation in the zone of hornblende-hornfels facies rocks, and elsewhere, with zones of dislocation (<25 m) in which bedding and fold axes are dragged into the strike of the fault. Bed-dip reversal and overturning may occur within these zones. The magnitude of horizontal displacement is less than 250 m, while the calculated vertical displacement is approximately 25 m with south block down.

Northwesterly trending dextral strike-slip faults are meridonal to northwesterly trending sinistral faults and terminate against northerly trending reverse and normal faults, or as horsetail splays.

Second-order sinistral strike-slip faults are inferred to be conjugate of northeasterly trending dextral strike-slip faults that are approximately one kilometre in strike length. These faults offset bedding by less than 25 m.

4.2.6.4 Third-order faults

Third-order faults are the most common structural discontinuities across the goldfield and three major types have been recognised: strike-slip, reverse and normal faults. The faults vary in strike direction, magnitude and displacement (Table 4.2).

In outcrop, strike-slip faults are recognised by the presence of white to buff or orange-coloured gouge, quartz-filled breccias zones and massive buck quartz veins infill along the fault plane. These faults exhibit a maximum dextral and sinistral offset of 25 m and either displace, or terminate at, reverse or normal faults, bedding and fold hinges. They may be developed as thin zones of intense brecciation (1-8 cm wide) that refract through competent rock units.
Third-order dip-slip faults (reverse and normal) trend northerly to north-northeasterly in the goldfield. They are characterised by the development of quartz breccias, gossanous quartz and gold-bearing quartz-sulphide lodes in the plane of the faults, and quartz fibres orientated normal to sub-normal to F1 fold axes. The faults are partly concordant to bedding. Their distribution across the goldfield is coincident with a broad northeasterly trending corridor of gold mineralisation and a northeasterly trending magnetic lineament which Koerber (1989) related to fracturing in the basement.

Two quartz slickenfibre sets are developed on some dip-slip fault planes. The first exhibits up-dip facing on accretion steps and are smeared-out by the second which exhibits down-dip facing on accretion steps. The smearing is interpreted to mean that reverse movement preceded normal movement, or that reverse faults were re-activated as normal faults. An example from the northerly trending, west dipping Lode Fault at the Quigleys deposit is presented in Figure 4.26. Here, along the footwall of the fault, quartz slickenfibres with down-dip facing accretion steps have smeared-out quartz slickenfibres with up-dip facing accretion steps and, because both fibres sets plot in the same sector within the primitive circle it is concluded that activity on the Lode Fault was reserve and then normal. A flip in the approximate orientation of the local stress field is implied with firstly, $\sigma 1$ trending westerly and secondly, $\sigma 3$ trending westerly. This may be consistent with a change in the principal stress directions incumbent on moving from $D_1$ to $D_2$ with the corollary that the Lode Fault, and other northerly to north-northeasterly trending faults in the goldfield that exhibit similar activity, were reverse in $D_1$ and normal in $D_2$.

A lineament analysis of third-order faults has been undertaken to establish if a relationship exists between the strike of the faults and deformation. During this analysis it was possible to assign all faults to either $D_1$ or $D_2$ by comparing them with shears generated in a non-coaxial stress field. The rationale for the comparison with a non-coaxial stress field was based on a lack of evidence of broad scale conjugation of third order faults, this being fundamental to a comparison with a coaxial stress field (Sylvester, 1988). The stress directions for $D_1$ and $D_2$, as indicated from the attitude of folds and the geometry of buck quartz veins ($V_1$), was adopted in order to constrain the maximum compressive stress direction of the stress field, and the faults were evaluated in three domains. The results are listed in Table 4.3.

The table indicates that all strike orientation and displacement directions can be accommodated with a corresponding shear. This may suggest that:

1. $D_1$ was associated with sinistral movement on northwest and north-northwesterly trending faults (R and P faults), dextral movement of northeast and north-northeasterly trending faults (X and R' faults), reverse movement on northerly trending faults, and normal movement on easterly trending faults.
2. $D_2$ was associated with sinistral movement on northeast and east-northeasterly trending faults (R and P faults), dextral movement on northwesterly trending faults (R' faults), and normal movement on northerly trending faults.

Furthermore, it is probable that faults which were formed during $D_1$ were reactivated during $D_2$ (and $D_3$).

4.2.7 **Structure in the Yenberrie Leucogranite**

A spaced, stylolitic to anastomosing tectonic foliation cuts the Yenberrie Leucogranite (Figure 4.27a), meta-sedimentary xenoliths, and wolframite-molybdenite-bismuth-cassiterite-copper sulphide-bearing quartz veins that cross-cut the leucogranite. The foliation refracts through the western and eastern cupolas of the leucogranite increasing in dip on refraction. At the contact margin with the hornfelsed country rock it dips shallowly to moderately west or east and strikes normal to sub-normal to the granite/country rock interface. Away from the margin, the foliation trends north-northeasterly and increases slightly in dip. However, on transition further into the leucogranite it refracts abruptly toward the northeast, and finally toward the east, becoming steeply dipping to the north or south. In stereographic projection (Figure 4.27b), this refraction is resolved as a spread of points about the primitive circle. The foliation is not continuous with an apparent fabric in the hornfelsed country rock but is continuous in trend with an $F_1$ syncline adjacent to the eastern cupola, and with a $F_1$ syncline at the western margin of the western cupola.

In thin section, the foliation is defined by numerous Fe-oxide stained, coarse stylolitic and anastomosing fractures, a weak alignment of platey minerals, and a subtle general flattening and elongation of quartz crystals.

In Figure 4.27c, the relationship of the foliation to the mean trend of the regional $S_1$ and $S_2$ has been considered: a good correlation is achieved between the mean trend of the regional $S_1$ and the foliation in the leucogranite. The foliation is oblique to the regional $S_2$. On the basis of this data, the foliation in the leucogranite is probably $S_1$. Based on the morphology of the foliation patterns outlined by Paterson et al. (1991) it is concluded that the fabric formed at the same time or after emplacement of the leucogranite.

Further to this, because the fabric cross-cuts wolframite-molybdenite-bismuth-cassiterite-copper sulphide-bearing quartz veins, and these transect the leucogranite, it is concluded that the veins also formed at the same time or after emplacement of the leucogranite, perhaps as extension fractures and/or fluid conduits that facilitated cooling and crystallisation of leucogranite (Balk, 1937; Pitcher, 1979; Ramsay, 1987).
4.2.8 Structure in the Tennysons Leucogranite

The Tennysons Leucogranite is dissected by numerous parallel to anastomosing northwesterly trending first-order shear faults of the Pine Creek Shear Zone (Section 2.2.3 & 4.2.6.2). The shear faults are spaced approximately 100-200 m apart and divide the leucogranite into ribbon-like blocks. Where the faults cross into, or define a contact with the Burrell Creek Formation, they are associated with a zone of mylonitisation in the leucogranite, and a narrow zone of brecciation and complex folding in the hornfelsed sediments. The folds are arranged oblique to the strike of the shear faults in a left step en échelon pattern that is indicative of sinistral strike-slip activity, as described by Sylvester (1988).

Where the faults occur within the body of the leucogranite, a narrow chloritised or silicified selvedge is developed which is commonly 20 cm to 3 m in width (Figure 4.28). A lineation orientated 40° towards 330°, and a steep vertical foliation at an angle to the shear plane, is defined throughout the selvedge material by a crude alignment and elongation of crystals. Microcline and andesine phenocrysts are flattened, strongly fractured, and cross-cut by thin (<1 mm) quartz-chlorite veins (Figure 4.29). Thin section analysis of orientated blocks of selvedge material has indicated that the shear faults are narrow zones of intense shear dislocation and flattening. Oblique sinistral displacement is indicated, i.e., west block up and toward the southeast.

It is suggested that the shear faults in the leucogranite are equivalent in age to second-order northwesterly trending shear faults in the nearby Yenberrie Shear Zone; that is, they were formed in D3.

4.3 METAMORPHISM

4.3.1 Introduction

In the Mt Todd goldfield four metamorphic events are recognised:

1. Contact metamorphism to hornblende-hornfels facies (H1) during emplacement of the Yenberrie Leucogranite.

2. Regional metamorphism to lower greenschist facies in D1 with retrogression of hornblende-hornfels facies (H1) assemblages and the development of S1 in cross-cutting relationship.

3. Contact metamorphism to hornblende-hornfels facies (H2) during emplacement of the Tennysons Leucogranite, and an overprint of H2 on H1. The development of H1 obliterated or obscured S1 within the inner contact aureole.

4. Regional metamorphism to lower greenschist facies in D2 with retrogression of H2 assemblages, and the development of S2. The fabric overprints H1, S1 and H2.
Three isotextural contact metamorphic zones are developed in the goldfield: an inner, middle and outer aureole (Figure 4.30). The inner aureole lies west of the Batman deposit and extends 3-4 km from the nearest granite outcrop. Rocks in this zone are characteristically massive, fine grained and black to grey in colour. Primary sedimentary features are resolved as a compositional variation, and cordierite porphyroblasts (seen as spots in hand specimen) are developed in hornfelsed siltstone and shale beds. Pre-existing tectonic fabrics are obscured or have been obliterated (Section 4.2.2 & 4.2.7).

The middle aureole extends eastward from the inner aureole for 0.5-1.5 km. Within this aureole primary sedimentary features are preserved and coarse grained rocks, such as greywackes and siltstones, retain their sedimentary integrity. Compositional variation and cordierite porphyroblasts may be developed in fine grained siltstones or shales. Pre-existing tectonic fabrics are well preserved.

The outer aureole surrounds the inner aureole and is believed to extend to 8.0 km east of Mt Todd (Koerber, 1989). In this aureole the sedimentary integrity of the rocks is retained except where they are substantially modified by cross-cutting tectonic fabrics (eg., within D₃ shear zones).

**4.3.2 Metamorphic assemblages**

The metamorphic assemblages developed in rocks of the Burrell Creek Formation are presented in Table 4.4 and are based on petrographic observations of 41 thin sections. The assemblages reflect several metamorphic and retrogressive events that were incident in the goldfield. In the inner aureole, they are biased toward assemblages developed during contact metamorphism while in the outer aureole, they are biased toward assemblages developed during regional metamorphism.

The highest grades attained during contact metamorphism (generally) may be met within pelitic rocks of the inner contact aureole with the development of the assemblage quartz-cordierite-muscovite-biotite-chlorite (Qtz-Crd-Ms-Bt-Chl) of the hornblende-hornfels facies. The occasional occurrence of potash feldspar and/or andalusite indicate that higher metamorphic grades were attained locally. The partial or complete alteration of cordierite porphyroblasts (Section 4.3.3) to sericite, chlorite and Fe-oxide, and the alteration of biotite to chlorite and rutile (Section 6.3) indicate the assemblage suffered retrograde metamorphism during rehydration of the contact aureole (Sylvester et al., 1978). Retrogression of this quartz-cordierite-muscovite-biotite-chlorite assemblage is also associated with a regional metamorphic overprint.

The highest grade attained during regional metamorphism has been recognised by Ferguson et al. (1980) within meta-pelitic rocks of the Central Domain of the Pine Creek Inlier with the development of plagioclase-potash feldspar-biotite-muscovite-chlorite-epidote (Pl-Kfs-Bt-Ms-Chl-Epi).
4.3.3 The relative timing of metamorphic events

The number and relative timing of metamorphic events has been determined from petrographic studies of cross-cutting textures, in particular, the cross-cutting relationships developed in the middle aureole between cordierite porphyroblasts, $S_1$ and $S_2$.

Two sets of cordierite porphyroblasts are developed in rocks of the middle aureole. The first set, $C_1$, range in size from 100-400 $\mu$m and are completely retrogressed to chlorite, sericite and Fe-oxide, or chlorite only, and thereby occur as spots (Figure 4.31). The chlorite and sericite define a fabric in the spots and this fabric is continuous with $S_1$ in the wall rock. The spots have a diffuse rim and/or a core of Fe-oxide and are elongate parallel to $S_1$. The texture indicates that $C_1$ formed pre- to syn-$S_1$ (Vernon et al., 1993), or that contact metamorphism and retrogression of cordierite to chlorite-sericite-Fe-oxide assemblage preceded, or was synchronous to, the development of a fabric in $D_1$.

The second set of cordierite porphyroblasts, $C_2$, range in size from 100-600 $\mu$m and cross-cut the $C_1$ and $S_1$ (Figure 4.31). The porphyroblasts have grown without evidence of $S_1$, indicating they formed after that foliation (Vernon et al., 1993). They are partially or completely retrogressed to aggregates of quartz, sericite and chlorite that are rimmed by Fe-oxide and thereby occur as spots: the sericite and chlorite weakly define a fabric that is continuous with $S_2$ in the wall rock. The spots are weakly elongate parallel to $S_2$. Adjacent to gold-bearing quartz-sulphide veins, retrogressed porphyroblasts are replaced by chalcopyrite, pyrrhotite, pyrite or marcasite, or by chlorite and sulphides, or by chlorite only. The texture indicates that:

1. $C_2$ formed prior to the development of $S_2$ or, retrogression of cordierite porphyroblasts to a chlorite-sericite-Fe-oxide assemblage preceded the development of a fabric in $D_2$.

2. A contact metamorphic event ($H_2$) occurred after $D_1$ ($C_2$ cross-cuts $C_1$ and $S_1$).

3. A contact metamorphic event ($H_2$) preceded the mineralisation event. This is so because type-$C_2$ porphyroblasts are replaced by sulphides adjacent to gold-bearing quartz-sulphide veins, it can be concluded that the mineralising event postdated peak temperatures associated with the second thermal metamorphic event. This agrees with the observations by Mason (1989) of chlorite and sulphide replacing cordierite porphyroblasts in Sample BD007, 45.2 m.

Two other metamorphic textures have relevance here. Firstly, in several orientated thin sections, plates of sericite and chlorite that define $S_2$, cross-cut plates of sericite that define $S_1$ in type-$C_1$ spots. The texture supports a pre- to syn-$S_1$ contact metamorphic event.

Secondly, in a thin section of the hinge region of a mesoscopic $F_2$ fold (thin section GN002), cordierite porphyroblasts that are retrogressed to sericite and chlorite are elongate within a

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wall rock fabric that is folded (Figure 4.32). Sericite and chlorite, both within and without the retrogressed porphyroblasts, define this fabric. The fabric is orientated sub-parallel to bedding and has been interpreted as S1. This texture suggests that (a) the development of cordierite porphyroblasts preceded the S1, and (b) the cordierite porphyroblasts and S1 were folded during D2. It further supports the interpretation of a pre- to syn-S1 contact metamorphic event.

In summary, an early set of cordierite porphyroblasts (C1) were formed during contact metamorphism of the Burrell Creek Formation to hornblende-hornfels facies (H1). The contact metamorphic assemblage suffered retrograde metamorphism during rehydration of the contact aureole, and/or retrogression during regional metamorphism to lower greenschist facies in D1. A fabric was developed (S1). The H1-D1 mineral assemblage was modified during a second contact metamorphic event to hornblende-hornfels facies (H2) and cordierite porphyroblasts (C2) were formed in cross-cutting relationship with respect to type-C1 spot and S1. This event preceded retrograde metamorphism during rehydration of the contact metamorphic assemblage, and gold mineralisation. The final phase of regional metamorphism to lower greenschist facies in D2 saw the development of a cross-cutting fabric (S2) and the continued retrogression of pre-existing metamorphic assemblages.

4.3.4 Relative timing of granite emplacement
A reasonable correlation has been made between the relative timing of metamorphic events, as outlined above, and the emplacement of the Yenberrie Leucogranite and the Tennysons Leucogranite. The correlation is based on the following:-

1. S1 cross-cuts type-C1 spots in the middle aureole, and cross-cuts the Yenberrie Leucogranite (Section 4.2.7). From this, type-C1 porphyroblasts are correlated with the emplacement of the Yenberrie Leucogranite.

2. A second thermal metamorphic event is implied by the presence of type-C2 porphyroblasts and, because the porphyroblasts cross-cut C1 spots and S1, the event occurred after the emplacement of the Yenberrie Leucogranite and the development of S1. Secondly, S2 cross-cuts type-C2 spots and therefore the second thermal event preceded the development of the S2. Thirdly, this study has noted that the S2 fabric transects the hornfelsed region (i.e., the inner aureole) on the margin of Tennysons Leucogranite (Figures 4.3-4.4, 4.6-4.7 & 4.12). It is reasonable to argue that S2 would have been obliterated during thermal metamorphism to hornblende-hornfels facies if the leucogranite had been emplaced after the development of S2. Since this is not the case, it is inferred that the Tennysons Leucogranite was emplaced prior to the development of S2 and is therefore correlated with type-C2 porphyroblasts.

Based on these correlations, it has been concluded that (a) the emplacement of the Yenberrie Leucogranite occurred pre- to syn-D1, and (b) the emplacement of the Tennysons Leucogranite occurred pre- to syn-D2, and pre- to syn- sulphide mineralisation.
Lastly, the timing of the mineralisation event may be constrained from the following:
1. Sulphide mineralisation overprints type-C₂ cordierite porphyroblasts (Section 4.3.3).
2. The S₂ fabric cross-cuts type-C₂ spots (retrogressed cordierite porphyroblasts - Section 4.3.3) and quartz-sulphide veins (Section 4.2.5.3).
3. The Tennysons Leucogranite is correlated with type-C₂ cordierite porphyroblasts.

From this it may be concluded that the gold mineralising event occurred after peak thermal metamorphism (as indicated by type-C₂ porphyroblasts) associated with the emplacement of the Tennysons Leucogranite, but preceded the development of the S₂ fabric.

4.3.5 P-T Conditions of metamorphism
The approximate P-T conditions for regional and contact metamorphism in the goldfield is indicated in Figure 4.33. The peak conditions associated with contact metamorphism to hornblende-hornfels facies may be described by the assemblage quartz-chlorite-muscovite-biotite-cordierite (Qtz-Chl-Ms-Bt-Crd). The general presence of cordierite porphyroblasts (retrogressed) with a sympathetic absence of andalusite, is believed to imply that metamorphism occurred at low pressure, perhaps between 1-1.5 kbars (Pattison & Tracy, 1991, pp 124 and Table 7). The occasional occurrence of potash feldspar (England, 1990) provides an upper temperature constraint of approximately 575°C for a pressure of 1.5 kbars. These P-T conditions are in good agreement with regional P-T estimates by Ferguson et al. (1980) of P<1 kbar and T~500-680°C for contact metamorphism in the Central Domain of the Pine Creek Inlier, and are indicative of low to medium grade contact metamorphism.

Assuming average crustal densities of 1.0 kbar ~ 3.3 km, it is estimated that the granites in the goldfield were emplaced at a minimum depth of approximately 3.5-5.0 km.

The peak conditions associated with regional metamorphism to lower greenschist facies have been estimated at T=400°-500°C, P<4 kbar from an absence of cordierite, staurolite or almandine in mineral assemblages (Ferguson et al., 1980).

4.4 STRUCTURAL AND METAMORPHIC HISTORY
In the Mt Todd goldfield, five tectonic events have been recognised. The age of these events are constrained by the age of emplacement of the Yenberrie and Tennysons Leucogranites at 1835-1820 Ma, this being the age of the Cullen Batholith as stated by Stuart-Smith et al. (1993). These events are presented in Figure 4.34 and summarised below:

**Event 1**: Contact metamorphism to hornblende-hornfels facies (H₁) during emplacement of the Yenberrie Leucogranite, with concurrent development of cordierite porphyroblasts (C₁). Retrogression of hornblende-hornfels facies assemblages accompanied rehydration during cooling and crystallisation of the pluton.
**Event 2:** Regional metamorphism to lower greenschist facies in D₁, with retrogression of hornblende-hornfels facies assemblages. D₁ is characterised by close to tight northeast to northwest trending, asymmetric folds (F₁) that plunge gently toward the south, and a continuous axial planar cleavage (S₁). The cleavage is cross-cutting with respect to the Yenberrie Leucogranite. The east vergence of folds suggests that tectonic transport was toward the east.

The formation of conjugate buck quartz veins (V₁) occurred early in D₁ at which time, a coaxial stress field may have operated in which a₁ plunged gently toward the west-southwest; a₂ plunged moderately south-southeast; a₃ plunged moderately north.

Folding in D₁ was associated with the development of numerous third-order faults with sinistral movement on northwest and north-northwesterly trending faults (R and P faults), dextral movement on northeast and north-northeasterly trending faults (X and R' faults), reverse movement on northerly trending faults, and normal movement on easterly trending faults. The fault orientations and a left step in fold axes indicates that a non-coaxial sinistral strain history may have operated at this time in D₁ with a₁ directed westerly, a₂ directed south and a₃ directed sub-vertically.

The discordance between the type of stress field operating early in D₁ (during the formation of buck quartz veins) and that operating during folding and faulting, may be the result of a progressive change from coaxial mechanics to non-coaxial sinistral strike-slip mechanics during deformation, with the corollary that early structures were rotated anticlockwise in the latter non-coaxial sinistral stress field.

**Event 3:** Contact metamorphism to hornblende-hornfels facies (H₂) during emplacement of the Tennysons Leucogranite, with concurrent development of cordierite porphyroblasts (C₂) in cross-cutting relationship with respect to C₁ and S₁. Retrogression of hornblende-hornfels facies assemblages accompanied rehydration during cooling and crystallisation of the pluton.

**Event 4:** Regional metamorphism to lower greenschist facies in D₂ with retrogression of H₂ hornblende-hornfels facies assemblages. The second deformation, D₂, is characterised by westerly trending, open folds (F₂) and a spaced disjunctive cleavage (S₂) that transects the folds. The cleavage is cross-cutting with respect to the H₁, S₁ and H₂. The folds are asymmetric in profile, and a northerly vergence suggests that tectonic transport was toward the north. The plunge of the folds ascribes a basinal geometry that may be indicative of fold drape across the Tennysons Leucogranite, or doming in the west during emplacement of the Tennysons Leucogranite. The latter would imply that plutonism and deformation were coeval: this agrees with the conclusions of Stuart-Smith et al. (1993) that east-west trending,
open style folds are associated with the emplacement of the Cullen Batholith (1835-1820 Ma).

During $D_2$ a non-coaxial stress field may have operated with sinistral movement on third order northeast and east-northeasterly trending faults ($R$ and $P$ faults), dextral movement on northwesterly trending faults ($R'$ faults), and normal movement on northerly trending faults. The maximum compressive stress, $\sigma_1$, was orientated approximately north-south.

Based on cross-cutting relationships, it is concluded that gold-bearing quartz-sulphide veins and lodes ($V_3$), and possibly quartz-tourmaline veins ($V_2$), were formed early in this deformation and after the peak thermal metamorphism associated with the emplacement of the Tennysons Leucogranite.

**Event 5:** Reactivation of west-northwesterly and northwesterly trending sinistral strike-slip faults in $D_3$. The faults are associated with a steeply dipping $S_3$-C type foliation and mesoscopic en échelon folds ($F_3$) that trend oblique to the strike of the faults. The faults offset $D_1$ and $D_2$ structures and may be associated with the development of calcite-base metal veins ($V_4$) and epithermal style quartz veins ($V_5$).

### 4.5 REGIONAL VERSUS GOLDFIELD TECTONIC EVENTS

The age of emplacement of the Yenberrie and Tennysons Leucogranites is constrained by the age of emplacement of the Cullen Batholith at 1835 to 1820 Ma, as stated by Stuart-Smith et al. (1993). Similarly, the age of $D_1$ and $D_2$ is constrained by the age of batholith emplacement because $D_1$ occurred after the emplacement of the Yenberrie Leucogranite but before emplacement of the Tennysons Leucogranite, while $D_2$ occurred synchronous to, or immediately after emplacement of the Tennysons Leucogranite.

With respect to fold style, $F_1$ in the Mt Todd goldfield is similar to the regional $F_2$ which were formed in geosynclinal strata and the Tollis Formation during deformation accompanying the Maud Creek Event (~1850 Ma). Unfortunately, an imprecise correlation is established because the Maud Creek Event is believed to precede emplacement of the Cullen Batholith (Stuart-Smith et al., 1993), and clearly, $D_1$ in the Mt Todd goldfield was synchronous with pluton emplacement. In contrast, $F_2$ in the Mt Todd goldfield is equivocal with the regional $F_3$ which formed in the geosynclinal strata during emplacement of the Cullen Batholith (Stuart-Smith et al., 1993). $D_3$ in the Mt Todd goldfield (ie., $F_3$ and reactivated strike-slip faults) is equivocal with deformation accompanying the Shoobridge Event (~1780 Ma), of Needham et al. (1988).
Figure 4.12b: Schematic section across the Mt Todd goldfield (from the Tennysons leucogranite to the Dick Grayson prospect, Figure 1.3) along mine grid northing 11000N showing the gross relationship of the main tectonic elements. The key to stratigraphic units shown is equivalent to that given in Figure 3.12b, Map - Lithological Interpretation (supplementary).
Figure 4.13: Equal area stereographic projection of total $S_1$ data. (b), (c), (d) and (e) consider the $S_1$ data in domains of equal strike. The plots demonstrate the helical trace of $S_1$ from the North domain through to the Access Road domain in the south of the goldfield.
Figure 4.14a, b, c, d: Equal area stereographic projections of the plunge of $F_1$, as calculated from the intersection of bedding and $S_1$. The circled points in (a) correspond with fold hinges that have been rotated toward the southeast adjacent to northwest trending strike-slip faults.
Figure 4.15: Equal area stereographic projection of poles to $S_2$ from the Access Road domain. $S_2$ lies in transection to $F_2$ such that the total set of data in (a) consists of northwesterly trending $S_2$ on the north and south limb of the fold, plots (b) and (d), and westerly trending $S_2$ on the hinge of the fold, plot (c). The dip of the cleavage at the hinge is steeper than on the limbs.
Figure 4.16: Equal area stereographic projection of poles to $S_2$ planes.  
B. Contour plot of poles to $S_2$ planes.
Figure 4.17: Sinistral shear faulting in the Yenberrie Shear Zone (Batman and Robin MG 9400N 7500E). The shear faults are weathered to a buff-red colour and are associated with a mesoscopic S-C fabric ($S_3$). This fabric is characterised by the development of jagged, lenticular slabs of rock that are positioned oblique to, and along the strike of the fault zone. (The camera lens cap is 5 cm in diameter.)

Figure 4.18: Buck quartz vein - a large single blow (Quigleys MG 13150N 11550E).
Figure 4.19: A photograph of a buck quartz vein in a road cut along the Mine Access Road (Stow Creek MG 6850N 11350E). The vein is lenticular and undulating in form and typical of buck quartz vein morphologies throughout the Mt Todd goldfield. Sinistral slip is indicated from offset on bedding. This vein is hosted by greywacke and volcanoclastic beds at the contact of the Burrell Creek Formation with the Tollis Formation (the greywacke beds are greyish in colour in this photo while the volcanoclastic beds are distinctly chocolate brown).
Figure 4.20: A. Block diagram of conjugate buck quartz veins. B. Equal area stereographic projection of poles to veins of buckle quartz in the North domain with mean vein orientations shown as great circles. C. Equal area stereographic projection of poles to veins of buckle quartz in the Access Road domain with the mean vein orientations shown as great circles.
## Table 4.1

Quartz sulphide veins and lodes of the Mt Todd goldfield

<table>
<thead>
<tr>
<th>Mine/prospect working</th>
<th>Style of veining</th>
<th>Lithofacies and host rock type</th>
<th>Geometry of veins set or lode</th>
<th>Production figures</th>
<th>Estimated reserves</th>
<th>Best intercept (g/t Au)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Batman deposit</td>
<td>Bed discordant sheeted vein set</td>
<td>Greywacke, siltstone and minor interbedded shale</td>
<td>dip: 70-75° E, length: &gt; 1 km, width: 300 m, depth: &gt; 500 m</td>
<td>In production</td>
<td>92.9 million tonnes @ 1.4 g/t Au</td>
<td>n/a</td>
</tr>
<tr>
<td>Quigleys series of pts</td>
<td>Fault hosted lode, footwall stockwork veins mostly discordant to bedding</td>
<td>Greywacke and siltstone</td>
<td>dip: NNE, overall strike: 25-50° W, length: ~ 1 km, total width: up to 3 m, depth: &gt; 100 m</td>
<td>8: 252 tonnes @ 3.84 g/t Au (1986-88, Pacific Goldmines)</td>
<td>646,000 tonnes @ 3.8 g/t Au (Ore stockpile and waste dumps: 317,000 tonnes @ 1.2 g/t Au)</td>
<td>n/a</td>
</tr>
<tr>
<td>Golf pit</td>
<td>Fault hosted lode, minor footwall stockwork Sheeted vein sets (X2)</td>
<td>Lode at shale-siltstone interface</td>
<td>dip: 72° E</td>
<td>3000 tonnes @ 10.0 g/t Au (1986-88, Pacific Goldmines)</td>
<td>Pit vein set: 14,000 t @ 6.0 g/t Au West Golf vein set: 0.5 million t @ 0.7 g/t Au</td>
<td>n/a</td>
</tr>
<tr>
<td>Jones Brothers Workings</td>
<td>Fault hosted lode (Extension of Golf lode)</td>
<td>Lode at shale-siltstone interface</td>
<td>dip: 72° E</td>
<td>1908-1919, 7 Jones Brothers 879 tonnes @ 27.9 g/t Au</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Penguin prospect</td>
<td>7 sheeted vein set (as indicated from diamond drill data)</td>
<td>Siltstone</td>
<td>1.5 km BCL, anomaly</td>
<td>n/a</td>
<td>n/a</td>
<td>n/a</td>
</tr>
<tr>
<td>New Tolls (Tolls of Kenny et al., 1990)</td>
<td>Sheeted veins and stockwork veins</td>
<td>Greywacke and siltstone</td>
<td>dip: 007° N, length: 300 m, width: 100 m</td>
<td>n/a</td>
<td>1.6 million t @ 1.1 g/t Au</td>
<td>RC TP013, 7m @ 1.37 g/t Au RC TP016, 4m @ 5.5 g/t Au</td>
</tr>
<tr>
<td>Regatta prospect</td>
<td>3 parallel lodes (Not fault hosted and bedding discordant)</td>
<td>Greywacke and siltstone</td>
<td>dip: -10° N, strike: 150-200 m, width: &gt; 1 m</td>
<td>n/a</td>
<td>63,468 t @ 4.73 g/t Au</td>
<td>-</td>
</tr>
<tr>
<td>Regatta East prospect</td>
<td>Stockwork veins (Bedding discordant)</td>
<td>Greywacke</td>
<td>no data available</td>
<td>n/a</td>
<td>320,000 t @ 1.6 g/t Au</td>
<td>-</td>
</tr>
<tr>
<td>Hotel prospect</td>
<td>Single lode (bed discordant)</td>
<td>Shale and siltstone</td>
<td>dip: 080° E, length: 350 m, width: 0.5-1.0 m</td>
<td>n/a</td>
<td>-</td>
<td>Rock chip sample, 2 m @ 6.9 g/t</td>
</tr>
<tr>
<td>Quigleys west</td>
<td>Sheeted vein set and fault hosted lode (bed discordant)</td>
<td>Greywacke and siltstone</td>
<td>strike: 010° N, dip: vertical</td>
<td>-</td>
<td>-</td>
<td>Rock chip sample of lode, 0.3 m @ 10.0 g/t</td>
</tr>
<tr>
<td>C3-C4</td>
<td>Stockwork veining (bed discordant)</td>
<td>Greywacke and siltstone</td>
<td>strike: 010° N</td>
<td>-</td>
<td>-</td>
<td>Rock chip sample, C3: 10 m @ 10.79 g/t C4: 10 m @ 2.77 g/t</td>
</tr>
<tr>
<td>Old Tolls</td>
<td>Fault hosted lode (bed discordant)</td>
<td>Siltstone</td>
<td>strike: 340° N, length: 150 m, width: 10-30 cm</td>
<td>1908-1919, 7 Jones Brothers Production figures not known</td>
<td>-</td>
<td>Grab sample, 20 cm @ 7.7 g/t</td>
</tr>
<tr>
<td>Alpha, Bravo, Charlie, Delta</td>
<td>2 sets of X2 parallel fault hosted lodes and stockwork veins (bed discordant)</td>
<td>Siltstone</td>
<td>strike: 010° N, dip: vertical, length: &gt; 50 m each</td>
<td>Collectively ~2000 tonnes @ 4.0 g/t Au (1986-88, Pacific Goldmines)</td>
<td>-</td>
<td>R&amp;B hole intercept, 24 m @ 3.1 g/t Water borehole BW005, 6.0 m @ 3.95 g/t</td>
</tr>
<tr>
<td>Horseshoe prospect</td>
<td>Sheeted vein set (bed discordant)</td>
<td>Greywacke</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>MG 7700N 8350E</td>
<td>Fault hosted lode (bed discordant)</td>
<td>Hornfels?</td>
<td>strike: 010° N, dip: vertical, length: &gt; 50 m, width: 0.5 m</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

Data from Cottle (1937), Hosfield and Nye (1941), Kenny et al. (1989), Wegmann (1990), Kenny et al. (1990), and Poxon and Hein (1994).
Figure 4.21a: Equal area stereographic projection of poles to quartz-tourmaline veins of the Mt Todd goldfield.

Figure 4.21b: Photomicrograph of green to brown coloured tourmaline fibres in a quartz-tourmaline vein from Mount Todd. The vein is 4 mm wide and is crosscut by a vein of calcite (V4). Quartz is syntaxial to the wall rock and tourmaline masses are scattered throughout the wall rock assemblage (right). The opaques at top right are pyrrhotite, chalcopyrite, and goethite after pyrrhotite and chalcopyrite. (Thin section MT013.)
Figure 4.22a: Equal area stereographic projection of poles to calcite-base metal veins in the Mt Todd goldfield.

Figure 4.22b: Photomicrograph of colloform pyrite and marcasite (py-marc) lining the walls of a 2 mm wide calcite-base metal vein. Galena (gn) and calcite (cal) occur as fill between the colloform bands while chalcopyrite (ccp) and sphalerite (sph) form a discontinuous base to the bands. DDH BD087(1), 297.78-297.94 m.
Figure 4.23: a. Reflected light photomicrograph of herringbone texture in galena (gn) which is surrounded by sphalerite (sph) and rhombs of calcite (cal). Scale bar is 2 mm long.

Figure 4.23: b. Transmitted light photomicrograph of Figure 4.23a. Galena is surrounded by Fe-bearing sphalerite (honey-brown coloured) and Fe-poor sphalerite (white). Thin section BD008a (1-3) collected from DDH BD008, 98-99 m. Scale bar is 2 mm long.

Figure 4.24: Photomicrograph of an epithermal style quartz vein (V$_5$). Densely packed layers of reniform or chalcedonic quartz forms a substrate to fine euhedral crystals of quartz (0.1 to 0.5 mm). The texture is interpreted as replacement of epithermal calcite by quartz. Scale bar is 1.0 mm long. (Thin section DC001, Batman and Robin MG 10830N 9940E.)

Figure 4.25: Epithermal style quartz vein (V$_5$). Loosely packed layers of reniform quartz are arranged into a rhombohedron and Fe-oxides occur as fill between layers. The texture is interpreted as replacement of epithermal calcite by quartz and Fe-oxide. (J001, Batman North MG 12150N 9275E.)
Table 4.2: The trend, type of movement, and average length of third order faults in the Mt Todd goldfield.

<table>
<thead>
<tr>
<th>Fault trend</th>
<th>Type of movement</th>
<th>Fault length</th>
</tr>
</thead>
<tbody>
<tr>
<td>NNE</td>
<td>oblique dextral</td>
<td>&lt;400 m</td>
</tr>
<tr>
<td>NE</td>
<td>dextral</td>
<td>150-500 m</td>
</tr>
<tr>
<td>ENE</td>
<td>dextral</td>
<td>150-500 m</td>
</tr>
<tr>
<td>E-W</td>
<td>dextral</td>
<td>&gt; 600 m</td>
</tr>
<tr>
<td>NW</td>
<td>normal (inferred)</td>
<td>n/a</td>
</tr>
<tr>
<td>NWW</td>
<td>dextral sinistral</td>
<td>300 m-1 km</td>
</tr>
<tr>
<td>N-S</td>
<td>reverse to normal to sinistral</td>
<td>1-2 km</td>
</tr>
</tbody>
</table>

Table 4.3: A comparison of third order faults and type of movement, with shear directions in a non-coaxial stress field.
A. Foliation in the Yenberrie leucogranite (Yenberrie Field MG 11600N 4725E).

B. Equal area stereographic projection of S₁ in the Yenberrie leucogranite. C. Equal area stereographic projection demonstrating the relationship of the mean S₁ and S₂ to the foliation in the leucogranite: the foliation correlates well with the regional S₁ but is oblique to the regional S₂.

Figure 4.27: A. Foliation in the Yenberrie leucogranite (Yenberrie Field MG 11600N 4725E). B. Equal area stereographic projection of S₁ in the Yenberrie leucogranite. C. Equal area stereographic projection demonstrating the relationship of the mean S₁ and S₂ to the foliation in the leucogranite: the foliation correlates well with the regional S₁ but is oblique to the regional S₂.
Figure 4.28: A photograph of a shear fault crosscutting the body of the Tennysons leucogranite, 4 km southwest of Yinberrie Hill. The shear faults are associated with a narrow chloritised or silicified selvedge: thin section analyses of orientated blocks of the selvedge material has indicated that the shear faults are narrow zones of intense shear dislocation and flattening. View is towards the southeast. (Thank you D. Bremner.)

Figure 4.29: A photograph of a sheared and strongly fractured sample of the Tennysons leucogranite. (Sample OR001, collected 3 km southwest of Yinberrie Hill.)
Figure 4.31: A tracing of a photomicrograph of retrogressed cordierite spots (C₁ and C₂) in a laminated shale. C₁ spots are completely retrogressed to sericite and chlorite and these minerals have adopted a preferred orientation (S₁). C₁ is elongate within S₁ and is crosscut by C₂. The texture is indicative of a contact metamorphic event pre- to syn-S₁. (Thin section GN001, Batman North MG 12550N 9900E.)

Figure 4.32: A sketch of a photomicrograph of a mesoscopic F₂ fold. The cordierite spots are completely retrogressed to sericite and chlorite and these minerals have adopted a preferred orientation (S₁). The spots are elongate within S₁ and both are folded about the fold hinge. The texture is indicative of a contact metamorphic event pre- to syn-S₁. (Thin section A005, Horseshoe Creek MG 13675N 9825E.)

Figure 4.33: The approximate P-T conditions for regional and contact metamorphism in the Mt Todd goldfield.
| Table 4.4 |
| Metamorphic assemblages of the Mt Todd goldfield |

<table>
<thead>
<tr>
<th>Middle/Outer Aureole</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Shale</strong></td>
</tr>
<tr>
<td>Qtz-Ms #</td>
</tr>
<tr>
<td>Ms-Chl #</td>
</tr>
<tr>
<td>Qtz-Ms-Fe oxide *</td>
</tr>
<tr>
<td>Qtz-Chl-Fe oxide *</td>
</tr>
<tr>
<td>Qtz-Chl-Ms #</td>
</tr>
<tr>
<td>Qtz-Chl-Ms-Bt-Crd *</td>
</tr>
<tr>
<td>Pl-Kfs-Bt-Ms-Chl-Epi (Ferguson et al., 1980)</td>
</tr>
</tbody>
</table>

| **Siltstone**        |
| Qtz-Chl-Ms *#        |
| Qtz-Chl-Ms-Bt *      |
| Qtz-Chl-Ms-Crd *     |
| Qtz-Chl-Ms-Bt-Crd *  |

| **Greywacke/Sandstone** |
| Qtz-Chl-Ms *#           |
| Qtz-Chl-Ms-Bt *         |

<table>
<thead>
<tr>
<th>Inner Aureole</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Hornfels</strong></td>
</tr>
<tr>
<td>Qtz-Ms-Bt</td>
</tr>
<tr>
<td>Qtz-Ms-Chl-Bt</td>
</tr>
<tr>
<td>Qtz-Ms-Bt-Kfs $*$</td>
</tr>
<tr>
<td>Qtz-Ms-Bt-Crd $*$</td>
</tr>
<tr>
<td>Qtz-Chl-Bt-Crd $*$</td>
</tr>
<tr>
<td>Qtz-Bt-Cord $*$</td>
</tr>
<tr>
<td>Qtz-Crd-Kfs-And $*$</td>
</tr>
<tr>
<td>$*$ = assemblage determined from one sample.</td>
</tr>
<tr>
<td># = outer aureole</td>
</tr>
<tr>
<td>* = middle aureole</td>
</tr>
</tbody>
</table>

Abbreviations after Kretz (1983).
faults - F3, S3

Lower greenschist facies
F2 folds & S2

quartz-sulphide veins
(GOLD MINERALISATION)
quartz-tourmaline veins
C2 cordierite spots
(hornblende-hornfels facies)

Lower greenschist facies
F1 folds & S1

faults & buck quartz veins
C1 cordierite spots
(hornblende-hornfels facies)

Tollis Formation
c. 1890 Ma

Burrell Creek Formation

Figure 4.34: The relationship of structural and metamorphic events in the Mt Todd goldfield.
CHAPTER 5
LOCAL STRUCTURES

5.1 PREAMBLE

From the tectonic events outlined in Chapter 4, a temporal and broad structural relationship has been established between the several quartz-sulphide veins and lode systems of the Mt Todd goldfield. The systems were formed after peak thermal metamorphism associated with the emplacement of the Tennysons Leucogranite, and early in D2, prior to the development of the regional S2 fabric. However, it has not been established that the sheeted vein systems and the lode/stockwork systems are co-genetic, or whether the structures that host these systems are co-genetic.

To this end, the structural setting of the sheeted vein system of the Batman deposit and the lode and stockwork vein systems of the Quigleys deposit are outlined and contrasted. From the comparison, the structural features that may have underpinned vein and lode formation are delineated.

The rationale for selecting the Batman and Quigleys deposits has been based primarily on their size and economic significance (they are the largest known in the goldfield), a high diversity in geological setting (Section 5.2.3 and 5.3.3) and an availability of diamond drill core. The deposits have been mapped in detail, and the data collected has been recorded on:

1. 1:1000 scale fact "Geology of the Quigleys pits" (Figure 5.1).
2. 1:50 scale fact geology over a cleared area on Batman Hill (Figure 5.2).
3. Fact geology recorded on a sketch of the south-western face of the Quigleys south pit (Figure 5.3).

These maps constitute a database for the following discussion and are presented in Appendix D.

5.2 BATMAN DEPOSIT

5.2.1 Introduction

The structure of the Batman deposit has been discussed and described by Koerber (1989), Kenny et al. (1989, 1990), Taylor (1990a), Hein (1991b, 1991c). The deposit is situated on
a prominent keel-shaped, north-northeasterly trending ridge (Batman Hill) which has a topographic elevation of 222 m. It occupies a near central position in the goldfield (Figure 1.3) and directly overlies the southern end of an arcuate shaped magnetic high that is believed to result from pyrrhotite in quartz-sulphide veins in the deposit (Kenny et al., 1989). A detailed structural map of a cleared section atop Batman Hill is presented in Figure 5.2 and the structural elements of this map are presented in a series of stereographic plots in Figure 5.4.

5.2.2 Mining history

Since the discovery of alluvial gold in 1871, the Mt Todd region has been known to be prospective for gold, tin, tungsten, molybdenum and uranium. The first recorded gold workings were established in 1908 by the Jones Brothers who mined a gold-bearing quartz ironstone reef 2 km west of Mount Todd (Cottle, 1937). The mine consisted of five shafts, three open cuts and a number of trial holes. Figures for the 1908-1911 period show that 879 tonnes of ore were treated with a recovery of 867.14 oz or 24.54 kg of gold (ie., 879 tonnes at 27.9 g/t) (Cottle, 1937).

The quartz-ironstone breccias of the Quigleys reef and the iron-rich quartz of the Tollis reef were discovered between 1908 and 1919. Operations at the Tollis reef consisted of one shaft and numerous surface pits and costeans. Hossfeld and Nye (1941) stated that an initial test grade of 7.7 g/t over 20 cm (5.1 dwt over 8 inches) was proved at the reef, but problems with extracting gold from the "sulphidic" ore hindered operations and finally caused the cessation of gold mining in the Mt Todd goldfield in 1911 (Hossfeld & Nye, 1941). By 1919, the goldfield was abandoned to sporadic prospecting (Rattigan & Clark, 1954/55).

In 1987, the Joint Venture Partners of Billiton Australia Gold Pty. Ltd. and Zapopan N.L. undertook a gold stream-sediment sampling program in the Mt Todd goldfield, and subsequently identified the Batman gold anomaly, 3 km WSW of Mount Todd. Leases across the goldfield were acquired from Pacific Goldmines in 1988 and by May of that year an ambitious drilling program had commenced. In March 1990, the total metres drilled were recorded as 38,235 m in the vicinity of Batman Hill (diamond, RC and percussion), with another 6975 m drilled in associated areas (Kenny et al., 1990). Geostatistical analysis based on data collected during that drill program, allowed the delineation of a resource of 39 million tonnes at 1.5 g/t Au at Batman Hill.

In April 1991, Billiton Australia Gold Pty. Ltd. exercised a deal with Zapopan N.L. which effectively dissolved the Joint Venture Partnership Agreement. By January 1992, Zapopan N.L. acquired full control of leases in the goldfield and proceeded with the development of the project. Phase I of the project, which involves heap leaching of the weathered and transition zones of the Batman orebody, commenced in December 1993 (Anon, 1993).
current identified mineral resource is 92.9 million tonnes of ore grading 1.4 g/t gold (Poxon & Hein, 1994).

5.2.3 Geological setting
The Batman deposit is hosted by rocks of the Burrell Creek Formation that comprise greywacke and siltstone interbedded with shale (ie., Lithofacies GR). Bedding is conformable and consistently southwest dipping with a modal strike of 320°-330° NW and a mean strike and dip of 321° 50° W (Figure 5.4a). The beds young to the southwest, as indicated from graded bedding.

The deposit is located on the northeast face of a basin formed by the intersection of a northwest to north-northwesterly trending, southerly plunging, D1 syncline with an easterly trending, D2 syncline (Figure 4.12). It lies immediately east of a major northerly trending lineament, the lineament roughly dividing hornblende-hornfels facies rocks of the inner contact metamorphic aureole from those of the middle aureole. The deposit also straddles a northeasterly trending magnetic lineament that may reflect a deep crustal structure (Figure 5.5). Koerber (1989) postulated that the structure related to antithetic fracturing in the basement during displacement along the Pine Creek Shear Zone.

5.2.4 Vein types
Two vein types have been observed in the Batman deposit: quartz-sulphide veins (V3) and calcite-base metal veins (V4). Quartz-sulphide veins are composed of crack-seal quartz which lines the vein wall, dog-tooth quartz, pyrrhotite, chalcopyrite, löellingite, ISS cubanite, pyrite, marcasite, arsenopyrite, gold, bismuth, bismuthinite, talnakhite (Cu9Fe8S10), pavonite (Ag9S3Bi2S3), hedleyite (Bi4Te3), Bi-rich galena, sphalerite, galena, calcite and chlorite. A mineral zonation is not apparent. Calcite-base metal veins are composed of dog-tooth quartz, calcite, galena, sphalerite, pyrite, arsenopyrite, and chalcopyrite (Hein, 1991a). The calcite-base metal veins cross-cut and offset quartz-sulphide veins both in outcrop and thin section, and are thus genetically late.

5.2.5 Quartz-sulphide veins (V3)
The main quartz-sulphide vein set dips east with a modal strike of 010°-020° and a mean strike and dip of 018° 60° E (Figure 5.4b). The veins occur as a parallel to anastomosing set of sheets which are spaced no more than one metre apart and commonly less than 10 cm apart (Figure 5.6). Individual veins are not continuous in strike. Kenny et al. (1989) recorded that the majority of veins averaged 10-20 mm in thickness, however, over the area of detailed mapping (Figure 5.2) vein widths rarely exceed 8 mm. The number of veins per meter varies from 5-15 with an average of 12, over an orebody width of 150 meters.

The veins commonly terminate as joints but may emerge as veins further along the same
joint. This gives the appearance of pinch and swell structures. They are truncated by a west-northwesterly trending, regularly spaced, fault-fracture set (spacing is less than one metre), or by individual bedding surfaces. Truncation at bedding tends to occur where there are changes in rock type or where beds have been displaced by bedding parallel, sinistral-reverse faults.

The veins have a two stage fill: early quartz-fibre fill with a late open space quartz-sulphide fill. Early quartz-fibre fill is associated with the nucleation and growth of straight fibrous crystals (Figure 5.7), perhaps by crack-seal mechanisms as described by Cox and Etheridge (1983). The quartz fibres are syntaxial to quartz clasts in the wall rock and thus line the vein walls. They have a mean long axis orientation that is roughly parallel to the crystallographic C-axes (England, 1990): the mean long axis orientation plunges 18° towards 288° (Figure 5.4c).

The orientation of the long axis of the quartz fibres may be used as a record of the displacement history of the vein walls (Cox & Etheridge, 1983) provided that undeformed fibres connect markers across the vein (Urai et al., 1991). Of the quartz-sulphide veins examined in this study, most showed some signs of deformation including undulose extinction, subgrain development, shearing and kinking. Nevertheless, the fibres record the displacement history of the vein walls because:-

1. They connect matching vein wall irregularities.
2. Where fibres are cross-cut by late open space quartz-sulphide fill, fibre segments may be paired from equivalent optical extinction and matching vein wall irregularities.

In corollary, because the fibres have tracked the opening trajectory, they are a manifestation of the extension direction (\(\lambda_1\)) (Hancock, 1985), i.e., \(\lambda_1\) was directed shallowly west-northwesterly during vein formation at the Batman deposit.

With regard to the displacement history of the vein walls, when viewed along the strike of the veins, the mean plunge of the long axis of the fibres is perpendicular to the mean dip of the veins, with a clockwise deviance of 12° (Figure 5.8). When viewed along vein dip, the mean azimuth of the long axis of the fibres is perpendicular to the mean strike of the veins. From this, it is concluded that vein dilation had a component of normal dip-slip displacement with east block down. The magnitude of normal dip-slip has been estimated for an orebody width of 150 m at approximately 3.0 m displacement. This is considered significant because:-

1. It indicates that zone of veining in the Batman deposit was coincident with a broad zone of extension and micro-faulting (normal). This zone is north-northeasterly striking.
2. There is a coincidence with other quartz-sulphide bodies in the goldfield, i.e., northerly to north-northeasterly trending normal faults (third order faults) are often characterised by the development of gossanous quartz and quartz-sulphide lodes in the plane of the faults, Section 4.2.6.3).
The late open space fill of quartz-sulphide veins is associated with several phases of micro-fracturing of the early fibre quartz and/or wall rock, with fresh sulphide being precipitated at each micro-fracture event. The successive micro-fracture/precipitation events parallel, cross-cut, meander through, or offset pre-existing vein fill producing a complex array of cross-cutting relationships. The nature of these cross-cutting relationships is discussed further in Chapter 6.

The alteration selvedge surrounding the veins in the Batman deposit is characterised by a pronounced red staining in outcrop, and as a thin zone of chloritisation and/or subtle bleaching in fresh diamond drill core. In thin section, the selvedge is composed of biotite, rare coarse muscovite, sericite, chlorite, and accessory tourmaline. The selvedge half-width is one to three times the width of the vein, and is uneven in distribution around or along the vein, ie., the selvedge may be partially or completely absent (Figure 4.9a, b). Indeed, the alteration meanders across, parallels or cross-cuts quartz-fibre fill with a high degree of synchronicity with open space fill. From this it is concluded that the alteration selvedge is associated with the sulphidic phase of vein formation, ie., open space fill.

A total of eight vein sets have been identified in the Batman deposit (Figure 5.10 & 5.6):-

1. Set A: north-northwest or north-northeast striking
   (i) steeply east dipping (the main set).
   (ii) steeply west dipping.
   (iii) vertical.
   (iv) dipping shallowly east.
   (v) dipping shallowly west.

2. Set B: east-northeasterly striking, southeast dip.

3. Set C: easterly striking, vertical.

4. Set D: bedding parallel.

Although it has been possible to identify vein sets which cross-cut, it has not been possible to sort the cross-cutting relationships into a sequence of vein formation. Individual veins cross-cut and are themselves cross-cut. It is just as likely that set A will cross-cut set B, as it is that set B will cross-cut set A. More so, that set A and B will parallel or anastomose. This is consistent with the nature of micro-fracturing of early fibre quartz and the generation of successive cross-cutting micro-fracture/precipitation events, as outlined above. It has therefore been concluded, that all quartz-sulphide vein sets are co-genetic.

5.2.6 Quartz slickenfibres

The presence of quartz slickenfibres on some bedding planes in the deposit area is indicative of slippage on those surfaces. A sinistral reverse (oblique) sense of slip has been established from east facing accretion steps and from east dipping extension fractures, ie., movement was west block up and towards the east. The transport vector for bed
displacement, as inferred from the mean azimuth of the fibres, is sub-perpendicular to the bedding planes and directed toward 098° (Figure 5.4d). The approximate trend of the maximum compressive stress (\(\sigma_1\)) was directed westerly. This concurs with the orientation of \(\sigma_1\) as determined from the geometry of buck quartz veins and the attitude of \(\sigma_1\) folds (Section 4.2.5.1 & 4.2.2) and it is concluded that reverse-sinistral displacement on bedding, with the concomitant development of quartz slickenfibres, occurred during \(D_1\).

5.2.7 Cleavage
A cleavage in the wall rock that is defined by an alignment of platey minerals, is continuous with a fracture cleavage in outcrop that cross-cuts the quartz-sulphide veins (Figure 5.11a) and causes "blocking" of the veins into a mosaic of cubic and rhombic shapes (Figure 5.11b). The wall rock cleavage has a mean strike and dip of 319° 52° SW and parallels the regional \(S_2\) cleavage. The fracture cleavage strikes slightly more westward with a mode of 300°-310° and dip southwest, and a mean strike and dip of 305° 54° SW (Figure 5.11a). The difference in orientation between wall rock and fracture cleavage is interpreted to reflect a westward refraction of \(S_2\) on intersection with the quartz-sulphide veins.

The relationship of the fracture cleavage to the wall rock cleavage has been studied in thin section and it is concluded that they are continuous (Figure 5.12). In the wall rock, the fabric is defined by the alignment of sericite and/or chlorite and this alignment has continuity with a series of closely spaced anastomosing trails of tiny fluid inclusions (<1µm) in the quartz-fibre fill. In turn, the trails cross-cut sulphide filled micro-fractures and are continuous with deformation lamellae in those sulphides (Chapter 6). The overall texture is interpreted to mean that the fabric formed after, and deformed the quartz-fibre fill and the quartz-sulphide open space fill. Since the fabric parallels the regional \(S_2\) cleavage, it is concluded that the quartz-sulphide veins formed prior to the development of the regional \(S_2\) cleavage.

5.2.8 Jointing styles
Several joint geometries have been measured in the area of detailed mapping and are presented in stereographic projection in Figure 5.13. Dominant joint sets are:-

- Set 1: East-northeasterly trending (055°-079°).
- Set 2: Easterly trending (080°-100°).
- Set 3: North-northeasterly trending (010°-020°)
- Set 4: West-northwesterly trending (290°-310°)

Joint set 1 and 2 dip primarily northward, shallowly to steeply: the overlap between these sets is the result of a slight rotation in the strike of joints across the area of detailed mapping. Joint set 3 lies parallel to the modal strike direction of quartz-sulphide veins while joint set 4 is oriented parallel to the modal strike direction of calcite-base metal veins.

The joint geometries are coincident with mesoscopic strike-slip faults, normal faults and breccias. Normal dip-slip faulting along joint set 3 was synchronous to quartz vein dilation (see Section 5.2.5). Late sinistral strike-slip faulting along joint set 1 produced drag folds in
some quartz-sulphide veins, while late sinistral strike-slip faulting along joint set 1, 2 and 4 offset those veins. Late normal faulting along joint set 3 (parallel to the modal strike direction of quartz-sulphide veins) has been noted by Koerber (1989).

5.2.9 The orebody: morphology and definition
The zone of quartz-sulphide veining on Batman Hill defines an orebody which strikes north-northeasterly. It terminates south of Batman Hill against strongly hornfelsed rocks of the Burrell Creek Formation and appears to "die out" northward before terminating against the Batman North Fault (Figure 4.11). The orebody dips 70°-75° to the east (Kenny et al., 1989) which is steeper than the dip of the quartz-sulphide veins (i.e., 60° E). Koerber (1989) noted that the dip of the quartz-sulphide veins remained relatively constant with depth and suggested that the orebody may have steepened as a result of an en échelon displacement of the veins during flexural slip on bedding. However, it is not altogether clear that a steepening has occurred.

The quartz crack-seal fibres represent the extension direction during vein formation at the Batman deposit, i.e., 8° toward 288°. The λ1 direction should lie normal to the plane of extension: the geometry of the plane of extension may describe the geometry of a hypothetical orebody and has been calculated at 018° 72° E (Figure 5.14a). The dip of the hypothetical orebody is in good agreement with the dip of the Batman orebody of 70° to 75°, as noted by Kenny et al. (1989), suggesting that orebody has not been steepened.

The nature and morphology of the orebody may thus be defined. It is composed of a stack of discontinuous sheet-like veins that are orientated 018° 60° E within an "ore envelope" that is orientated 018° 72° E (Figure 5.14b). The nature of the "ore envelope" is that of a broad zone of normal micro-faults in which each vein describes a single dilated fault plane. The manner of dilation across each fault plane has been tracked during the growth of quartz fibres.

5.2.10 The timing of quartz-sulphide veins
It has been possible to constrain the time of formation of the quartz-sulphide veins at the Batman deposit by combining structural data in Sections 4.2, 4.4 and 5.2.

1. The veins are cross-cut by calcite-base metal veins (Section 5.2.4) and a set of east-northeast and easterly trending sinistral strike-slip faults (Section 4.2.5). The veins therefore predate those structures.

2. The veins are traversed and deformed by the regional S₂ fabric (Section 4.2.5) and therefore predate the development of that fabric element.
3. Reverse-sinistral displacement on bedding, with the concomitant development of quartz slickenfibres, occurred during D₁ when σ₁ was westerly trending (Section 5.2.6). Quartz-sulphide veins were formed when σ₃ was west-northwesterly trending (Section 5.2.5). The veins can not have formed synchronous to the development of quartz slickenfibres since this would imply that σ₁ was parallel to σ₃. That is to say, the maximum and minimum stress axes would have been directed approximately westward at the same time, and this would not be logical. It follows that the veins can not have formed during D₁.

4. The following palaeo-stress orientations have been established for the formation of buck quartz veins in D₁: σ₁ plunged gently toward the west-southwest; σ₂ plunged moderately south-southeast; σ₃ plunged moderately north (Section 4.2.5.1). In addition, the general north-south orientation of D₁ folds indicate that the maximum compressive stress (σ₁) operating during D₁ was orientated east-west (Section 4.2.2). In contrast, quartz-sulphide veins were formed when σ₃ was west-northwesterly trending (Section 5.2.5). It follows that the veins can not have formed during the development of buck quartz veins and folds in D₁ since this would imply that σ₁ and σ₃ were directed approximately westward at the same time. Again, this is not logical: the veins can not have formed during D₁.

Based on this data and that from Chapter 4, it is concluded that the quartz-sulphide veins in the Batman deposit formed early in D₂ and prior to the development of the regional S₂ fabric. The formation of the veins occurred during east-west extension. Subsequent to their development, the quartz-sulphide veins (V₃) were cross-cut by calcite-base metal veins (V₄) and offset along east-northeast and easterly trending sinistral strike-slip faults.

5.3 QUIGLEYS DEPOSIT

5.3.1 Introduction

The structure of the Quigleys deposit has been described and discussed by Kenny et al. (1990) and Wegmann (1990). The deposit comprises a 1-1.2 km long system of lodes (ie., quartz and sulphide-rich zone of proto-cataclasite or cataclasite that are located approximately 4-5 km northeast of the Batman deposit in a cluster of elongate north-northeasterly trending hills of 202 m maximum elevation. The lodes crop out in four pits: Quigleys North pit (QNP), Quigleys Central pit (QCP), Quigleys 10500 pit (10500P) and Quigleys South pit (QSP), and they strike approximately 015°-030°, 355°, 030°-150° and 340°, respectively. The pits have been mapped at a 1:1000 scale (Figure 5.1, Appendix C) and for convenience are referred to by their acronyms as indicated above. The fact geology and structural features of the southwestern face of the QSP are represented in Figure 5.3 (Appendix C). The structural elements of the pits are summarised in Table 5.1.
5.3.2 Mining history

The "quartz ironstone breccia" of the Quigleys lode system was discovered between 1908 and 1919 (Rattigan & Clark, 1954/55). Subsequent to its discovery several companies sampled and tested the lode system, but despite a promising gold potential, the reef was not prospected further.

In late 1981, CRA Exploration undertook resource reconnaissance and commenced a diamond drill program of 14 holes totalling 676.5 m (Kenny et al., 1989). A resource of 160,000 tonnes at 5.6-6.7 g/t Au over a strike length of 800 m, was defined (Wegmann, 1990). Once again, despite establishing the potential of the reef, operations were withdrawn from the goldfield.

During late 1986, Pacific Goldmines N.L. tested the lode system with 28 percussion pre-collared HQ diamond holes and 39 costeans (Wegmann, 1990). Open cut mining in a series of pits along the strike of the deposit commenced in that year and continued to February 1988, with 81,252 tonnes at 3.84 g/t Au being recovered from the resource.

In 1992 the resource was transferred to Zapopan N.L. when that company acquired the leases across the Mt Todd goldfield. The current resource estimate is 646,000 tonnes at 3.8 g/t Au, with a further 317,000 tonnes at 1.2 g/t Au in ore stockpile and waste dumps.

5.3.3 Geological setting

The Quigleys deposit are situated approximately 2 km due east of the zone of hornblende-hornfels facies rocks, at the margin between the middle and outer contact metamorphic aureoles (Figure 4.30). The deposit is hosted in rocks of the Burrell Creek Formation that are composed of greywacke, siltstone and shale (Lithofacies SH2-SG4). It overlies a northeast trending aeromagnetic lineament that has been related to fracturing in basement rocks (Koerber, 1989).

The deposit lies in a structurally complex region of the goldfield. It comprises several interconnected gold-bearing lodes and quartz-sulphide stockwork vein sets that are arranged in a steep *en échelon* array within a north-northeasterly trending corridor. The lodes and veins are cross-cut by an F2 anticline, but cross-cut an F1 anticline, indicating that they formed before folding in D2, but after folding in D1 (Figure 5.15). The position and general structure of the pits is as follows.

*QNP* is located on the southeastern limb of a northeasterly trending F1 syncline. Bedding is north to northeasterly trending and west-dipping, and (stratigraphic) facing is toward the west. At least three discrete lodes occur in this pit and they occupy bedding concordant faults that lie at the interface of siltstone and greywacke. The lodes are interconnected at
their northern and southern strike extent by weakly developed stockwork vein systems that form a right stepping array. At the southern end of the QNP, beds are drag folded and rotated eastward against one of the lodes (Figure 5.1).

**QCP** is situated immediately south of the QNP and comprises two lodes that trend northerly. The lodes are situated on the western limb of an F1 anticline and occupy bedding concordant faults that lie at the interface between shale and greywacke units (Figure 5.1). Bedding is north-northwesterly striking and westerly dipping, and (stratigraphic) facing is towards the west. The lodes are interconnected at their northern and southern strike extent by a stockwork vein system that forms a right stepping array. This right-stepping is associated with bed scale faults (up to 0.5 m in length) that are northeasterly striking and offset bedding sinistrally.

At the southern end of the pit, the lode/stockwork vein system cross-cuts a northeasterly trending F1 anticline and a westerly trending dextral strike-slip fault (third-order). The dextral fault offsets the anticline by approximately 30-40 m. The structural relationships are interpreted to mean that folding preceded faulting, and that veining postdated faulting. This suggests that the lodes and veins in the QCP are post- D1.

**10500P** is located south of the QCP on the western limb of a north to northeasterly trending F1 anticline (Figure 5.1). It comprises a single lode that is continuous into the QSP around a westerly trending F2 anticline, and is continuous with the lodes of the QCP via a stockwork vein system that right-steps across bedding. In the footwall, the lode is concordant to beds of greywacke that are northerly trending and westerly dipping: the lode is subtended by a zone of complex stockwork veining that is well developed about the hinge of the F2 anticline. In the hanging wall, beds of greywacke and siltstone are west, north and northeasterly trending and discordant to the lode.

Toward the southern end of the pit, the lode/stockwork vein system cross-cuts a second-order sinistral strike-slip fault that is defined by a zone of brecciation in the hanging wall (the footwall is absent). The fault, denoted the Quigleys Fault, may be traced north-westward for greater than 3.0 km into the inner contact metamorphic aureole (Figure 4.11 & 5.1). It is observed to offset D1 folds at West Quigleys but does not offset the margin of the inner contact aureole. The structural relationships are interpreted to mean that folding preceded faulting, and that veining and a contact metamorphic event postdated faulting. This suggests that the lodes and veins in the 10500P are post- D1 (and perhaps synchronous to a contact metamorphic event).

**QSP** is located in the southern face of a domal structure formed from the intersection of a northeasterly trending F1 anticline and a westerly trending F2 anticline (Figure 5.15). It
comprises a single northwesterly trending lode that is continuous into the 10500P around a westerly trending F2 anticline. It may be continuous with stockwork veining in the Regatta East prospect, 500 m south of the QSP (Wegmann, 1990).

At its northern end, the lode is concordant to northwesterly trending beds of greywacke in the footwall, but discordant to westerly trending beds of greywacke, siltstone and shale in the hanging wall. The hanging wall beds are dissected by numerous northwesterly trending faults that have reverse displacement, and an assemblage of imbricate blocks is developed. Block displacement is northward towards the hinge of the F2 anticline. Beds dip steeply north and south, but (stratigraphic) facing is southward indicating that some beds are overturned.

At its southern end, the lode cross-cuts westerly trending beds of siltstone and shale in the hanging wall and footwall, and (stratigraphic) facing is toward the south.

5.3.4 Vein types
Three vein types have been observed in the Quigleys deposit: buck quartz veins (V1), quartz-sulphide veins and lodes (V3) and calcite-base metal veins (V4).

**Buck quartz (V1):**- Buck quartz veins are developed in the hanging wall and footwall of the QNP and 10500P as conjugate pairs or parallel sets (Figure 5:16). They are lenticular in form with a maximum length of 12 m, and are weakly undulated in plan view. The veins are cross-cut by the regional S1 cleavage, quartz-sulphide lodes, stockwork veins, and by a set of easterly dipping, northwesterly trending normal faults. The buck quartz veins are composed primarily of brecciated anhedral white quartz, but may contain sulphide and supergene minerals where they are cross-cut by quartz-sulphide stockwork veining. It is concluded that the veins formed early in D1 (Section 4.2.5.1).

**Quartz-sulphide lode and vein stockwork (V3):**- The northerly trending lodes are subtended in the footwall, or interconnected by, and intricate stockwork of orthogonal or anastomosing veins and veinlets. Single veins also occupy northwesterly trending sinistral strike-slip faults. The lodes and veins are composed of anhedral quartz fibres, euhedral quartz, arsenopyrite, pyrite, marcasite, chalcopyrite, löllingite, bismuth, bismuthinite, gold, electrum, minor pyrrhotite, galena and sphalerite. Supergene hematite, chalcocite, digenite and bornite occur in the weathered horizon. Chlorite is well developed throughout the host wall rock.

**Calcite-base metal veins (V4):**- The veins are composed of euhedral quartz, calcite, galena, sphalerite, chalcopyrite and rare rutile needles. In outcrop and thin section, the veins cross-cut quartz-sulphide stockwork veins.
5.3.5 Quartz-sulphide lodes

Bedding concordant lodes are the main gold-bearing structures in the Quigleys deposit. They comprise massive lenticular proto-cataclasite or cataclasite bodies up to 350 m long and 1.5 m wide that are restricted to the interface of beds of greywacke and siltstone, or greywacke and shale. In outcrop, they are composed of boulders and clasts of quartz, Fe-oxide and Fe-oxide/quartz (0.2 mm - 4 cm) set in a quartz and red-brown-blue coloured Fe-oxide matrix (Figure 5.17). The matrix is vuggy to massive and forms 50-60% of the total rock volume. Clasts and boulders may be sub-rounded, angular or elongated. The lodes are associated with a pervasive alteration selvedge composed of biotite, chlorite, sericite, rutile and Fe-oxide that may extend up to 3 m into the hanging wall and one metre into the footwall.

In thin section of samples from diamond drill core, the lodes are generally composed of quartz, arsenopyrite and pyrite clasts that are cemented in a quartz or quartz-pyrite matrix (Figure 5.18a,b). A late cross-cutting fracture set is associated with development of calcite-base metal veins (V4). The nature of these cross-cutting relationships is discussed further in Chapter 6.

5.3.6 Stockwork veining

A complex zone of stockwork veining is developed in the footwall to the lode in the 10500P and within the hinge of the F2 between the 10500P and QSP. The zone comprises a coarse in-situ breccia (Figure 5.19a, b): fine veins (<1 mm wide) are connected via larger veins (5 mm to 4 cm wide) to the main lode. Three dominant vein orientations have been identified: (a) a northerly trending vertical set that conjugates with a northwesterly trending, (b) steeply west dipping set, and (c) a shallowly east-dipping set. The conjugate vein set is associated with quartz-sulphide mineralisation of the type developed in the lodes, and a narrow alteration selvedge of biotite, chlorite, sericite, rutile and Fe-oxide equal to that developed about the lodes. The set is continuous with the lodes and cross-cuts the regional S1 fabric (Figure 5.19a, b). This is interpreted to mean that the veins and lodes formed after the development of the fabric in D1.

The shallow east dipping vein set cross-cuts and displaces the conjugate vein set and S1. The veins consist of quartz and calcite with minor sphalerite and galena, and are coincident to a set of easterly dipping, northwesterly trending normal faults that offset buck quartz veins in the QNP and 10500P. On the basis of mineral composition and cross-cutting relationships, the veins have been assigned to vein type V4.

5.3.7 Quartz slickenfibres

In the Quigleys deposit, two quartz slickenfibre sets have been observed. The first set may be found on west and east dipping bedding planes throughout the deposit. They are oriented parallel to sub-parallel to the dip direction, and a reverse sense of slip has been established from the striae and extension fractures. The transport vector for displacement,
as determined from the mean azimuth of the fibres, is directed toward 102° with the maximum compressive stress (\(o_1\)) directed approximately westerly (Figure 5.20a). This agrees with the orientation of \(o_1\) as determined from the geometry of slickenfibres on bedding in the Batman deposit, the geometry of buck quartz veins in the North and Access Road domains, and the attitude of \(D_1\) folds (Section 5.2.6, 4.2.5.1 & 4.2.2). It is therefore concluded that the reverse displacement on bedding, with the concomitant development of up-dip facing quartz slickenfibres, occurred during \(D_1\).

The second set of fibres are restricted to bedding surfaces that form the footwall to the quartz-sulphide lodes in the QNP, QCP and 10500P. The fibres exhibit down-dip facing on accretion steps (west facing) with displacement toward 292°, and a normal sense of slip has been interpreted (Figure 5.20b). The approximate direction of the minimum compressive stress (\(o_3\)) was directed toward the west, and this is in good agreement with a \(o_3\) orientated westerly during formation of gold-bearing quartz-sulphide veins in the Batman deposit.

Two other features have relevance here. Firstly, where the two fibre sets occur together on the footwall to the lodes down-dip facing slickenfibres smear accretion steps on up-dip facing slickenfibres. This suggests that reverse slip preceded normal slip and argues for a maximum compressive stress (\(o_1\)) directed toward the west before a minimum compressive stress (\(o_3\)) directed toward the west.

Secondly, the restricted distribution of down-dip facing accretion steps to the footwall to the lodes, and the interconnected nature of the lodes with stockwork veins and bed scale sinistral faults, may suggest that dilation occurred along a single continuous normal-sinistral structure. This structure is referred to as the Lode Fault. It has been classified as an early \(D_2\) structure because it cross-cuts a northeast trending \(F_1\) anticline at a position south of the QCP, but is folded by a westerly trending \(F_2\) anticline at the southern end of the 10500P.

5.3.8 **The hanging wall**

The hanging wall to the lodes contain several northwesterly trending faults that cross-cut and offset the Lode Fault and the quartz-sulphide lodes (offset is less than 1.0 m). The faults strike northwesterly, dip steeply to the south, and are host to calcite-base metal veins. In the QNP and 10550P, the faults divide the hanging wall into a series of fault blocks that are rotated and internally internal deformation (brittle). When viewed in plan, the sense of rotation is anticlockwise relative to the unfaulited footwall such that a sinistral sense of displacement along the Lode Fault is implied (Figure 5.1). Conversely, in QSP, the faults divide the hanging wall into a series of blocks that are imbricated relative to the unfaulited footwall, such that a dextral sense of displacement along the Lode Fault is implied (Figures 5.1 & 5.3). The culmination of the displacement directions occurs about an \(F_2\) anticline located between 10500P and QSP suggesting that faulting and block rotation may have accompanied flexural-slip folding in \(D_2\). Given that the faults cross-cut the lodes and that the lodes are folded by and \(F_2\) anticline, it is concluded that flexural-slip folding, displacement
along the Lode Fault, and block rotation and imbrication in the hanging wall occurred subsequent to the mineralising event (or that mineralisation occurred prior to folding D2).

5.3.9 Cleavage
The mean strike and dip of S1 and S2 in each pit in the Quigleys prospect is presented in Table 5.1. S1 is developed as a fine continuous parallel to anastomosing fabric that is defined predominantly by an alignment of platey minerals and a flattening of quartz clasts. The cleavage is sub-parallel to the strike of the lodes, striking northerly in the QNP to northeasterly in the 10500P, and north-northeasterly in the QSP. S1 is folded about the F2 anticline and is steeply west dipping. It cross-cuts buck quartz veins in the QNP and 10500P and is cross-cut by stockwork veining at the hinge of the F2 anticline. This indicates that buck quartz veins formed pre-S1 (as concluded in Section 4.2.5.1), while quartz-sulphide stockwork veining formed post-S1 (as indicated in Section 4.2.5.3).

S2 is developed as a spaced coarse disjunctive cleavage that is defined by an alignment of platey minerals. It is associated with a fabric parallel joint set in massive greywacke beds. The fabric transects the hinge of the F2 fold but trends northwesterly on the north limb of the fold as observed elsewhere in the Mt Todd goldfield (Section 4.2.3). It dips steeply south at the fold hinge but shallowly to steeply on the fold limbs, therein forming a divergent fan.

5.3.10 Timing of quartz-sulphide lodes and stockwork veins
Based on the cross-cutting relationships outlined above, it is concluded that the quartz-sulphide lodes and associated stockwork veins at the Quigleys deposit were formed early in D2 before the development of F2 and S2. The formation of lodes and veins was synchronous with normal displacement and dilation of the Lode Fault. Subsequent to mineralisation, the lodes and veins were folded about an F2 anticline. Faulting, block rotation and fault block imbrication in the hanging wall to the lodes accompanied flexural-slip folding in D2 and minor faulting of the lodes. Northwesterly trending faults were host to late cross-cutting calcite-base metal veins.

5.4 THE BATMAN AND QUIGLEYS DEPOSITS: A COMPARISON

A summary of the geological features of the Batman and Quigleys deposits is presented in Table 5.2. The deposits are remarkably similar with respect to strike geometry, vein/lode morphology, vein forming process, alteration assemblage and relative geological time. There is a broad agreement in the type of host lithofacies, an association with a northeasterly trending regional magnetic lineament, the direction of the principle stress axes, and position within the contact aureole. The range of sulphide mineralogies at the Quigleys deposit is a subset of sulphide mineralogies at the Batman deposit. From this data it is concluded that the deposits formed at the same time, in similar lithofacies and by a similar process.
The structural features which may be considered as fundamental to the formation of the deposits are:-

1. Normal faults (microscopic and mesoscopic). Regardless of whether the host structures originated as a joints (at the Batman deposit) or reverse faults (at the Quigleys deposit), they have ultimately behaved as normal faults (microscopic and mesoscopic) during the mineralising process in D2.

2. The position of the deposits within the contact aureole. The deposits formed after peak thermal metamorphism associated with the Tennysons Leucogranite, and probably during cooling of the pluton and retrogression of contact metamorphic assemblages. They are therein confined to the thermal aureole of the pluton. Since the deposits formed after peak thermal conditions and during cooling of the pluton (collapsing isotherms) they may have formed within either the inner, middle or outer contact aureole. This highlights the potential for gold in these regions.

3. Competency of lithofacies. Stockwork and sheeted vein sets tend to be prevalent in lithofacies in which competent rock types predominate (ie., massive greywacke and siltstone beds) and this appears to be related to the capacity for those rocks to fracture or joint. In contrast, lodes are prevalent in those sites where the lithofacies have a significant competency contrast between their component rock types, ie., between greywacke and shale beds. The difference in competency may cause slip between individual beds or dilation at the interface of contrasting rock types. Competent rock types therefore provide suitable targets for gold reconnaissance surveys.

4. The position with respect to a northeasterly trending corridor. The corridor is chiefly defined from the distribution of BCL (bulk cyanide leach) soil gold anomalies (Kenny et al., 1990), and quartz-sulphide vein sets and lodes. Within the corridor, the vein sets and lodes strike north-northeasterly and form a right step en échelon array that is indicative of sinistral strike-slip movement along the corridor. Since the corridor is coincident with a distinct northeasterly trending magnetic lineament (Koerber, 1989), a zone of strike-slip displacement overlying a basement strike-slip fault is concluded. Sinistral displacement along this fault early in D2 (synchronous to the emplacement of the Tennysons Leucogranite) resulted in dilation of north-northeasterly trending faults, bedding planes and joints, and these became mineralised. Clearly, a position proximal to this structure provides a favourable exploration target.

5. Strike geometry. The mineralisation event was concurrent with dilation of faults, joints and bedding planes that strike approximately north-northeast, with little regard to the dip geometry of those structures.

The confluence of strike geometry, rock competency, position with respect to basement strike-slip faults, position within the contact aureole, and potential for dilation during D2, provide a useful exploration criteria for gold mineralisation in the Mt Todd goldfield.
The structural features which seem inconsequential to the formation of the deposits in the Mt Todd goldfield include:-

1. **The nature of the host structure.** Joints at the Batman deposit versus reverse faults at the Quigleys deposit suggest that the nature of the host structure is inconsequential.

2. **The relationship to the host lithology.** The deposits may be concordant or discordant to bedding.

3. **The relationship to folds.** In general, the deposits do not have a broad spatial association with antiformal or domal structures as occurs at the Enterprise Mine (House, 1990), Spring Hill prospect (Nicol, 1991), Union Reefs prospect (Donaldson, 1992), Howley district (Partington, 1990), and Moline/North Hercules mines (Nicholson and Eupene, 1990). However, these should not be precluded in any exploration model.
Figure 5.4: Equal area stereographic projection data from the Batman deposit. A. Bedding. B. Quartz-sulphide veins. C. Quartz fibres (crack-seal). D. Quartz slickenfibres.
REGIONAL GEOLOGY OVERLAY

Emerald Creek Mine

Quigleys

Horseshoe Fault

Gotham City Fault

Yamberra Fault

Tennysons Leucogranite

Stuart Highway

River

Tooronga Creek Fault

Tollis Formation

REGIONAL GEOLOGY OVERLAY

Figure 5.5
Figure 5.8: Geometries associated with the main quartz-sulphide vein set in the Batman deposit. An approximately 3 metre normal displacement (down-throw to the east) is indicated for an orebody width of 150 metres.

Figure 5.9a: Photograph of the alteration selvedge surrounding the veins in the Batman deposit. View is toward the south. (Batman and Robin MG 10375N 8505E.)

Figure 5.9b: Sketch of Figure 5.9a. The chlorite/sericite alteration selvedge is uneven in distribution around or along the vein, and it may be partially or completely absent.
Assume 12 veins per metre
at average width 8 mm
\[= 96 \text{ mm per metre}\]
\[= 14.4 \text{ m per 150 m}\]
\[
\tan = \frac{\text{opp}}{\text{hyp}}
\]
\[
\tan 12^\circ = \frac{X}{14.4 \text{ m}}
\]
\[= 3.06 \text{ m}
\]
3 metres

5.8

5.9a

5.9b
Figure 5.10: Sketch of the vein sets in the Batman deposit.

Set A: NNW-NNE strike
1. East dip (steep)
2. West dip (steep)
3. Vertical
4. Shallow dip east
5. Shallow dip west
Set B: ENE strike
Set C: Easterly strike
Set D: Bedding parallel

Figure 5.11:
A. S$_2$ fracture cleavage crosscutting a quartz-sulphide vein in the Batman deposit. The vein was painted red at some stage prior to photography. Camera lens cap is 5 cm in diameter. (Batman and Robin MG 10000N 8450E.)
B. The S$_2$ fracture cleavage has caused blocking of quartz sulphide veins into a mosaic of rhombic shapes (Batman and Robin MG 10375N 8500E).
Quartz fibre which has grown synclastically
Sulphide
Sphalerite
Chlorite
Highly strained quartz
\( S_2 \) in the quartz-sulphide vein
\( S_2 \) or \( S_1 \) in the wallrock

Figure 5.12: Tracing of a photomicrograph demonstrating the continuity between the fracture cleavage and the wall rock cleavage in the Batman deposit (Thin section BD007c collected from DDH BD007, 119.0 m.)

Set 1: ENE trending
Set 2: Easterly trending
Set 3: Parallel to quartz-sulphide veins
Set 4: Parallel to calcite-base metal veins

Figure 5.13: Equal area stereographic projection of joint geometries in the Batman deposit. The overlap of points for set 1 and 2 is the result of a slight rotation in joint strike across the mapped area.
Figure 5.14: A. Equal area stereographic projection of the orientation of the Batman orebody. B. Diagrammatic representation of the geometry associated with the orebody and the main quartz-sulphide vein set of the Batman deposit.
Table 5.1

Summary of the structural elements of the Quigleys deposit.

<table>
<thead>
<tr>
<th>Pit</th>
<th>Bedding</th>
<th>Facing</th>
<th>$S_1$</th>
<th>$S_2$</th>
<th>Lode trend</th>
<th>Lode dip</th>
</tr>
</thead>
<tbody>
<tr>
<td>QNP</td>
<td>012° 49° W (footwall)</td>
<td>West</td>
<td>003° 55° W</td>
<td>321° 70° W</td>
<td>000°</td>
<td>50°</td>
</tr>
<tr>
<td></td>
<td>024° 44° W (hangwall)</td>
<td></td>
<td>018° 73° W</td>
<td></td>
<td>015°</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>130°</td>
<td>45°</td>
<td></td>
</tr>
<tr>
<td>QCP</td>
<td>356° 46° W (footwall)</td>
<td>West</td>
<td>no data</td>
<td>294° 79° S</td>
<td>355°</td>
<td>45°</td>
</tr>
<tr>
<td></td>
<td>345° 45° W (hangwall)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>10500P</td>
<td>033° 47° W (footwall)</td>
<td>West</td>
<td>001° 56° W</td>
<td>289° 48° S</td>
<td>020°</td>
<td>40°</td>
</tr>
<tr>
<td></td>
<td>029° 35° W (hangwall) (287° 56° S - rotated block)</td>
<td></td>
<td>057° 83° W</td>
<td></td>
<td>043°</td>
<td>30°</td>
</tr>
<tr>
<td>QSP</td>
<td>303° 36° S (footwall)</td>
<td>SW in the footwall.</td>
<td>029° 74° W</td>
<td>271° 88° S</td>
<td>336°</td>
<td>25°</td>
</tr>
<tr>
<td></td>
<td>077° 77° S (hangwall)</td>
<td>SE in the hangwall</td>
<td>no data</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Figure 5.15: Structural interpretation of the Quigleys deposit: 1:5000 scale map.
Figure 5.16: Parallel set of buck quartz veins in the footwall of Quigleys 10550 pit. The largest is 12 metres long. (Thank you D. Wegmann.)

Figure 5.17: Quartz-sulphide lode of Quigleys 10550 pit. The lode is composed of intensely micro-fractured and brecciated boulders and clasts of quartz, in association with vuggy to massive red-brown-blue Fe-oxide.

Figure 5.18a: Photomicrograph of lode material from DDH QD008, 120 m: cataclasite of quartz (black and pink) and arsenopyrite (white). Scale bar = 2 mm.

Figure 5.18b: Photomicrograph of lode material from DDH QD008, 120 m: cataclasite of quartz (black), pyrite (yellow) and arsenopyrite (white). Scale bar = 2 mm.
Figure 5.19: a & b. The complex zone of stockwork veining in the footwall to the lode of the Quigleys 10500 pit comprises a coarse in-situ breccia in which fine veins are connected via larger veins to the main lode. S1 in the wall rock (centre left) is crosscut by quartz-sulphide stockwork veins. Late calcite-base metal veins (sub-horizontal) crosscut quartz-sulphide stockwork veins (sub-vertical). View is towards the north.

A Mean bedding for the Quigleys deposit
004° 39° W

B Mean bedding (footwall only)
019° 45° W

Figure 5.20: Equal area stereographic projection of slickenfibres at the Quigleys deposit. A. Reverse displacement is west block up. B. Normal displacement is west block down.
Table 5.2
Summary of the geological features of the Batman and Quigleys deposits.

<table>
<thead>
<tr>
<th></th>
<th><strong>Batman deposit</strong></th>
<th><strong>Quigleys deposit</strong></th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Type</strong></td>
<td>Sheeted to stockwork vein set</td>
<td>Lodes with stockwork veins</td>
</tr>
<tr>
<td><strong>Host lithology</strong></td>
<td>Greywacke &amp; siltstone</td>
<td>Greywacke, siltstone &amp; shale</td>
</tr>
<tr>
<td><strong>Lithofacies:</strong></td>
<td>GR</td>
<td>SH2-SG4</td>
</tr>
<tr>
<td><strong>Relationship to host</strong></td>
<td>Bed discordant</td>
<td>Bed concordant to discordant</td>
</tr>
<tr>
<td><strong>Aureole</strong></td>
<td>Inner to middle contact aureole</td>
<td>Middle to outer contact aureole</td>
</tr>
<tr>
<td><strong>Position</strong></td>
<td>On limb of F1 syncline</td>
<td>- QNP: on limb of F1 syncline</td>
</tr>
<tr>
<td></td>
<td></td>
<td>- QCP: on limb &amp; at the crest of F1 anticline</td>
</tr>
<tr>
<td></td>
<td></td>
<td>- 1050P: on limb of F1 anticline</td>
</tr>
<tr>
<td></td>
<td></td>
<td>- QSP: SW face of an F1 dome</td>
</tr>
<tr>
<td><strong>Magnetics</strong></td>
<td>Overlies bullseye anomaly, overlying NE magnetic linear</td>
<td>Overlies NE magnetic linear</td>
</tr>
<tr>
<td><strong>Geometry</strong></td>
<td>018° 72° E (orebody) 018° 60° E (main vein set)</td>
<td>~ 010° overall, dips 25°-50° west</td>
</tr>
<tr>
<td><strong>Mineralogy</strong></td>
<td>Quartz, pyrrhotite, chalcopyrite, löllingite, ISS cubanite, pyrite, marcasite, arsenopyrite, gold, bismuth, bismuthinite, talnakhite, pavonite, hedleyite, Bi-rich galena, sphalerite, galena</td>
<td>Quartz, arsenopyrite, pyrite, marcasite, chalcopyrite, löllingite, bismuth, bismuthinite, gold, electrum, minor pyrrhotite, galena, sphalerite</td>
</tr>
<tr>
<td><strong>Vein/lode morphology</strong></td>
<td>Crack-seal veins succeeded by open space fill. Veins formed syn-brecciation &amp; micro-fracturing</td>
<td>Crack-seal veins succeeded by open space fill. Lodes &amp; veins formed syn-brecciation &amp; shearing</td>
</tr>
<tr>
<td><strong>Host structure</strong></td>
<td>N-S trending joints</td>
<td>N-S trending reverse faults</td>
</tr>
<tr>
<td><strong>Principle vein forming process</strong></td>
<td>Dilation of the joints associated with normal dip-slip displacement with east block down</td>
<td>Dilation of the faults associated with normal dip-slip displacement with west block down</td>
</tr>
<tr>
<td><strong>Alteration</strong></td>
<td>Quartz, biotite, chlorite-rutile after biotite, coarse muscovite, sercite, chlorite, accessory tourmaline, Fe-oxides</td>
<td>Quartz, chlorite, sercite, biotite, rutile, Fe-oxides</td>
</tr>
<tr>
<td><strong>α1 direction</strong></td>
<td>- westerly (up-dip facing slicks)</td>
<td>- westerly (up-dip facing slicks)</td>
</tr>
<tr>
<td></td>
<td>- northerly (inferred)</td>
<td>- northerly (inferred)</td>
</tr>
<tr>
<td><strong>α3 direction</strong></td>
<td>- sub-vertical (inferred)</td>
<td>- sub-vertical (inferred)</td>
</tr>
<tr>
<td></td>
<td>- westerly (crack-seal fibres)</td>
<td>- westerly (down-dip facing slicks)</td>
</tr>
<tr>
<td><strong>Timing</strong></td>
<td>Early D2 (post-S1 to pre-S2)</td>
<td>Early D2 (post-F1, S1 to pre-F2)</td>
</tr>
</tbody>
</table>

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6.1 PREAMBLE

Gold mineralisation in the Mt Todd goldfield is associated with several discrete quartz-sulphide veins sets and lodes that formed co-genetically during east-west extension in \( D_2 \), and during emplacement of the Tennysons Leucogranite. In the Batman deposit, gold mineralisation is contained within an intense zone of quartz-sulphide veining that constitutes an orebody striking \( 018^\circ \) and dipping steeply toward the east. 117 polished thin sections of selected quartz-sulphide vein mineralisation from diamond drill core of this deposit have been examined in order to establish the timing of gold with respect to sulphide phases present: diamond drill hole numbers, locations and detailed section descriptions of 32 representative sections are included in Appendix E and F. The data collected has been used to establish:

1. The mineralogy of the veins.
2. The mineral textures.
3. The paragenetic sequence to mineralisation (with specific reference to the timing of the gold event).
4. The sequence of cross-cutting micro-structures.

The interpretation of sulphide textural equilibrium has been used to estimate the physico-chemical conditions operating during mineralisation. Mineral precipitation from a cooling acidic fluid is indicated.

6.2 MORPHOLOGY OF THE MINERALISATION

The morphology of the quartz-sulphide veins in the Batman orebody has been described in Sections 4.2.5.3 and 5.2. In summary, the veins are composed predominantly of fibre quartz in association with pyrrhotite, chalcopyrite, arsenopyrite, pyrite, marcasite, and traces of cubanite, löllingite, galena, sphalerite, talnakhite, bismuth, bismuthinite, and gold, and rare hedleyite, Bi-rich galena and pavonite.

The quartz-sulphide veins comprise a two stage fill: early fibre quartz lines the vein walls, while euhedral quartz and sulphide occur as late fill. The late fill is associated with several
phases of micro-fracturing of early fibre quartz and/or wall rock, with fresh sulphide being precipitated (or remobilised?) at each micro-fracture event. A complex array of cross-cutting relationships is produced because quartz and sulphide infills fractures which cross-cut, brecciate or offset early quartz-fibre fill.

Some sulphides are fibrous, particularly pyrrhotite (Figure 6.1), arsenopyrite and pyrite suggesting that crack-seal mechanisms may have accompanied the second stage of vein fill. The veins are cross-cut by $S_2$ and thereby exhibit weak to strong re-equilibration of the pre-existing vein textures in sympathy with the magnitude of deformation. The nature of deformation in the veins and the resultant textures is discussed in Section 6.4.4.

The majority of the gold is hosted within the vein quartz or in cross-cutting relationship to associated sulphides, with a small amount hosted at the vein wall rock interface, and even less within the alteration selvedge encompassing the veins. England (1990) recorded that 82 wt% of the gold occurred in 2.5-30.0 µm grains, his calculations being based on the size distribution of 1200 gold grains that he observed in 51 polished thin sections of selected quartz-sulphide vein samples from diamond drill core of the Batman orebody. Gold within vein quartz is intimately associated with bismuth in fluid inclusions (England, 1990; Hein, 1991a); the nature of this association is described further in Chapter 7. Zonation of sulphide minerals with respect to the quartz veins has not been observed.

6.3 ALTERATION MINERALOGIES

The general alteration assemblage comprises biotite, muscovite, chlorite, sericite, rutile, accessory tourmaline, chalcopyrite, pyrrhotite, pyrite or marcasite. Biotite is generally retrogressed to chlorite ± rutile. Muscovite, chalcopyrite, pyrrhotite, pyrite or marcasite, chlorite±sulphides, or chlorite only, may replace cordierite (Section 4.3.3), although cordierite is generally retrogressed to aggregates of chlorite and sericite that are rimmed by Fe-oxide. Mason (1989) recorded the presence of disseminated tourmaline towards the margin of the alteration zone.

6.4 MINERAL TEXTURES

6.4.1 Introduction

Mineral textures in the quartz-sulphide veins of the Batman deposit are dominated by chaotic aggregations of sulphide, silicate and carbonate minerals that are cross-cutting, fractured and brecciated. The aggregates are weakly to strongly deformed and/or are slightly to completely altered by supergene processes. For this reason, the mineral textures of the deposit are
considered in four sections:-

1. Silicate and carbonate mineralogy (Section 6.4.2)
2. Sulphide textures (Section 6.4.3)
3. Deformation textures (Section 6.4.4)
4. Supergene alteration (Section 6.4.5)

Description of textures in Sections 6.4.2, 6.4.3 and 6.4.4 are restricted to polished thin sections of diamond drill core from the zone of unweathered rock, while those for Section 6.4.5 are concerned with observations at outcrop and in polished thin sections of diamond drill from the weathered and transitional zones of Kenny et al. (1990). Mineral acronyms are used to avoid breaking the continuity of the text.

6.4.2 **Silicate and carbonate mineralogy**

Silicate mineralogy is dominated by quartz which is brecciated, fractured and in part, strongly deformed. Its predominance during the early stage of vein fill (quartz-fibre fill), and its occurrence together with pyrrhotite, pyrite and arsenopyrite, and as fill to fractures which host gold and bismuth, suggest that it was persistent throughout vein formation. Three quartz types are recognised:

1. **Quartz 1 (qtz 1)** comprises straight crack-seal fibres which line the vein walls. The fibres are densely decorated with wispy trails of tiny fluid inclusions (<1-3 \(\mu\)m) that lie sub-perpendicular or parallel to the c-axes of fibres. The inclusions are interpreted to have formed during irregular fibre growth or during sequential micro-fracture events (e.g., Cox & Etheridge, 1983).

2. **Quartz 2 (qtz 2)** is developed as a thin selvedge (<250 \(\mu\)m) that blankets crack-seal fibre terminations (Figures 6.2 & 6.3), or clusters of euhedral crystals that lie within, or at the margins of some sulphides, particularly pyrrhotite (Figure 6.4). The selvedge is optically continuous with quartz crack-seal fibres and commonly forms distinct crystal terminations. The texture indicates that a significant dilation event preceded qtz 2 precipitation. The quartz terminations occasionally exhibit growth zones containing tourmaline (<1-50 \(\mu\)m) and liquid-vapour fluid inclusions (Figures 6.5 & 6.6). The fluid inclusions are described in Chapter 7.

3. **Quartz 3 (qtz 3)** comprises anhedral grains intergrown with sulphides, and interpreted to be part of the sulphide paragenesis.

With regard to tourmaline, the associated mineral assemblage is equivalent to that developed in quartz-tourmaline veins which crop out on Mount Todd and atop several hills 750 m north of the Quigleys North pit (Section 4.2.5.2), i.e., quartz in association with tourmaline, biotite, and rare masses of pyrrhotite and chalcopyrite. Quartz-tourmaline veins have not been observed in cross-cutting relationship to quartz-sulphide veins, but sulphide minerals are developed in quartz-tourmaline veins. Indeed, quartz-tourmaline vein mineralogy forms a subset to quartz-sulphide vein mineralogy. It is therefore concluded that the two vein types formed during the same mineralising event.
The presence of brown biotite within the immediate wall rock to the veins, in contact with euhedral qtz 2, or in aggregation with pyrrhotite in the veins (England, 1990) suggests that biotite was a stable phase during early vein formation. Biotite forms large anhedral plates where it contacts pyrrhotite and is typically retrogressed to chlorite and rutile aggregates. Biotite and pinkish muscovite may occur as infill to rare fractures which cross-cut crack-seal quartz.

Chlorite and sericite dominate silicate mineralogy in the alteration selvedge surrounding the veins, and postdate the development of qtz 1, qtz 2, tourmaline and asp 1 (see below). Within the veins, chlorite is associated with pyrrhotite, chalcopyrite, pyrite, talnakhite and bismuthinite indicating it was persistent during precipitation of those sulphides. In the wall rock, fine aggregates of chlorite and sericite, and occasionally, coarse plates of chlorite, replace cordierite. Chlorite replacing biotite is common throughout the vein wall rock assemblage.

Calcite is abundant in the veins and wall rock of the Batman deposit. However, this mineral is also predominant in calcite-base metal veins which cross-cut quartz-sulphide veins (Section 4.2.5.4) and a significant calcite overprint is recognised. It is therefore difficult to ascertain the proportion of calcite precipitated during quartz-sulphide mineralisation. Nevertheless, calcite in intimate association with sphalerite, galena, ccp 2, bismuth, bismuthinite and tal 2, is regarded as quartz-sulphide vein gangue.

6.4.3 Sulphide textures

Sulphide textures in the Batman veins are composed of anhedral and occasionally euhedral grains and masses of sulphide that are fractured or brecciated. Anhedral mono-sulphide masses exhibit polygonal or granoblastic texture, crack-seal fibre texture, colloform texture, and growth zonation, while poly-sulphide masses exhibit simple to complex replacement of successive sulphide species (excluding replacement during supergene alteration), or are developed as mosaics of one or more sulphide species. Euhedral grain masses exhibit loose or densely-packed crystal intergrowths that partially or wholly enclose anhedral aggregations.

At least five phases of pyrrhotite precipitation have been identified:-
1. Pyrrhotite 1 (po 1) in equilibrium with anhedral cubanite and talnakhite 1 (tal 1) in 50 µm wide inclusions (Figure 6.7) within fractured euhedra of arsenopyrite (Figure 6.8). The texture is indicative of the coeval precipitation of po 1, cubanite and tal 1 prior to that of arsenopyrite (asp 1). Electron microprobe analyses of cubanite (CuFe2S3) and talnakhite (Cu8Fe8S16) indicate that the minerals may approach solid solution with haycockite (Cu4Fe5S8) and mooihoekite (Cu9Fe9S16), respectively (Table 6.1 & Figure 6.9).
2. Non-pitted sub-polygonal pyrrhotite 2 (po 2): occurs as isolated, sub-rounded cores in
pitted pyrrhotite-qtz 3 (Figure 6.10), or occasionally as fill to fractures in arsenopyrite (asp 1). Po 2 is replaced by pyrrhotite 3 and qtz 3.

3. Pitted Pyrrhotite 3 (po 3): arguably the most common form in the Batman deposit, especially at depth. It is characterised by numerous fine to coarse quartz-filled inclusions (qtz 3) that appear as pits in the polish surface (Figure 6.10). The po 3-qtz 3 assemblage replaces po 2 massively, and along grain boundaries: fractured monoclinic-shaped tablets of quartz in the po 3-qtz 3 matrix may be pseudomorphs after arsenopyrite 1 (Figure 6.11). Po 3 forms smooth embayed margins with chalcopyrite 1 (ccp 1) indicating coeval precipitation of the two sulphides (Figure 6.12), i.e., po 3-ccp 1-qtz 3 replaces po 2. The textural inter-relationship of asp 1, po 2, po 3 and ccp 1 suggest that the precipitation of po 2 succeeded that of asp 1, but preceded that of po 3-ccp 1-qtz 3.

4. Pyrrhotite 4 (po 4): occurs as fill to fractures which cross-cut arsenopyrite 2 and 3 (see below) and thus postdates the development of those minerals. The pyrrhotite does not exhibit a diagnostic texture.

5. Pyrrhotite 5 (po 5): occurs as subhedral to anhedral grains (approximately 200 µm wide) in intimate association with talnakhite 2 (Figures 6.13 & 6.14). Intergrowths of po 5 with talnakhite 2 and/or bismuthinite are common (see below). These textures suggest that the precipitation of po 5 was coeval with that of bismuthinite and talnakhite 2.

Pyrite is the principal sulphide in the upper levels of the Batman orebody, although marcasite (euhedral) may be locally abundant. These sulphides typically replace early pyrrhotite. England (1990) noted that pyrite, and less commonly marcasite replacement of pyrrhotite was extensive.

Three successive pyrite types have been noted:
1. Subhedral to anhedral, non-pitted sub-polygonal pyrite 1 (py 1). It is sometimes developed as sub-rounded cores in pyrite 2 (Figure 6.15). Py 1 replaces pyrrhotite and therein contains numerous inclusions of that sulphide. Py 1 is replaced by the assemblage pyrite 2-qtz 3.

2. Pitted pyrite 2 (py 2): is characterised by its strong, and somewhat unusual association with marcasite 1 (marc 1) with which it forms fine speckled and colloform intergrowths (Figures 6.16 & 6.17). The nature of this association is described in Section 6.6. Py 2 is also denoted by the numerous inclusions of quartz (qtz 3) that occur as fine to coarse pits in the polish surface (Figure 6.15). The pitted py 2-marc 1-qtz 3 assemblage replaces py 1 and pyrrhotite, but also forms delicate fringes to py 1 (Figure 6.16). The py 2-marc 1-qtz 3 assemblage is succeeded by euhedral marcasite (marc 2).

3. Pyrite 3 (py 3): anhedral to subhedral pyrite occurs as fill between coarse crystals of marcasite 2 (marc 2). The association typically forms densely packed aggregates (Figure 6.18), although loosely packed interspersions of py 3 and marc 2 may be developed (Figure 6.17). In this case, py 3 is euhedral.
The textural inter-relationships of marcasite and pyrite phases are consistent with the precipitation/replacement succession py 1 => py 2 => marc 1 => marc 2 => py 3.

In concert with pyrrhotite and pyrite, at least four phases of arsenopyrite precipitation, and one of lőellingite have been identified:-

1. Euhedral arsenopyrite 1 (asp 1): The monoclinic symmetry of arsenopyrite is strongly exhibited by this phase. Asp 1 is characterised by rare inclusions of po 1-tal 1-cubanite (Figures 6.7 & 6.8) and is cross-cut by fractures containing po 2, or po 3-qtz 3. The textural inter-relationship suggest that the precipitation of asp 1 succeeded that of po 1-tal 1-cubanite, but preceded that of po 2 and po 3-qtz 3.

2. Lőellingite typically forms embayed cores within asp 2, and occasionally py 1 (Figure 6.19). In asp 2, it accompanies or cross-cuts inclusions of po 3-ccp 1-qtz 3 (Figure 6.20). The texture suggests that the precipitation of lőellingite succeeded that of po 3-ccp 1-qtz 3 but preceded replacement by asp 2.

3. Anhedral arsenopyrite 2 (asp 2) and arsenopyrite 3 (asp 3): asp 2 encloses and replaces grains and/or aggregates of lőellingite, and is itself enclosed and replaced by asp 3 (Figure 6.21). Asp 3 lies in optical discontinuity to asp 2: the two phases are readily differentiated by their curious unequal anisotropy under crossed polars. Po 4 encloses and fills fractures which cross-cut aggregates of lőellingite-asp 2-asp 3. The textural inter-relationships are consistent with the replacement succession lőellingite --> asp 2 --> asp 3. The sulphides are cross-cut and enclosed by po 4 indicating that po 4 succeeded lőellingite-asp 2-asp 3.

4. Euhedral arsenopyrite 4 (asp 4) occurs in loosely packed aggregates in association with py 4 and marc 2 (Figure 6.17), or in cross-cutting relationship to those sulphides. The textures suggest that asp 4 precipitated after py 4 and marc 2.

The presence of chalcopyrite 2 (ccp 2) in fractures in pyrite indicates that it precipitated after that mineral. Ccp 2 occurs as spongy anhedral grains and is wholly or partially enclosed by anhedral galena and/or sphalerite (Figure 6.22): sphalerite typically encloses galena. The texture is suggestive of the precipitation/replacement succession ccp 2 => galena => sphalerite. Sphalerite may also enclose pyrrhotite, pyrite, marcasite and arsenopyrite indicating that it formed after those sulphides (Figure 6.20).

A third chalcopyrite 3 (ccp 3) phase is associated with sphalerite in which it occurs as tiny blebs (3-5 µm in size), along crystal cleavage planes and cross-cutting fractures (Figure 6.23). The texture is similar to chalcopyrite disease (Barton & Bethke, 1987) which is indicative of replacement of Fe-bearing sphalerite by Fe-poor sphalerite and chalcopyrite (Eldridge et al., 1988).

Talanakhite 2 (tal 2) in intimate association with po 5, occurs as infill to fractures that cross-cut vein quartz (qtz 1 and 2), suggesting that a brecciation event occurred during, or preceded,
talnakhite-po 5 precipitation. The green-brass-tan coloured mineral (optical colour) has similar optical properties to that of chalcopyrite, however, contains slightly less sulphur and Fe than that mineral. Electron microprobe analyses suggests it approaches solid solution with mooihoekite (Cu$_2$Fe$_9$S$_{16}$) (Table 6.1 & Figure 6.9). The tal 2-po 5 assemblage is partly enclosed and/or cross-cut by the assemblage bismuth-gold-bismuthinite (Figures 6.13 & 6.24) suggesting that tal 2-po 5 precipitation preceded that of gold, bismuth and bismuthinite.

The presence of gold, bismuth, bismuthinite, hedleyite (Bi$_{14}$Te$_6$), pavonite (Ag$_2$S.3Bi$_2$S$_3$), and bismuth-rich galena as fill to fractures that cross-cut the vein quartz-sulphide assemblage (Figures 6.13, 6.24-6.27) suggest that a brecciation event occurred during, or preceded the precipitation of these minerals. Gold is intimately associated with bismuth in hypersaline-CO$_2$-CH$_4$ fluid inclusions (Figure 6.28a-c), or closely associated with bismuth and bismuthinite as fill within fractures (Figures 6.24, 6.26, 6.28-6.30). England (1991a) recorded the presence of bismuth as metallic-liquid fluid inclusions in quartz. Smooth curved contacts between hedleyite and bismuthinite suggest that these minerals precipitated at the same time: they are enclosed by gold-bismuth grains (Figure 6.30). Bismuth-rich galena and pavonite occur as 5-10 µm inclusions within coarse bismuth (Figure 6.31). The textures indicate that the coeval precipitation of gold, bismuth and bismuthinite succeeded the precipitation of hedleyite, bismuthinite, pavonite and Bi-rich galena. The occurrence of gold and bismuth in fluid inclusions is discussed in Chapter 7.

The late replacement of bismuth by bismuthinite is indicated by numerous tiny blades of bismuthinite sprouting from bismuth grains, and bismuthinite mantles enclosing sub-grains within the bismuth grain matrix (Figure 6.29).

6.4.4 Deformation textures

The quartz-sulphide veins of the Batman deposit are cross-cut by S$_2$, and sulphide, silicate and carbonate mineral textures are weakly to strongly modified by deformation. Using the subdivisions adopted by Marshall and Gilligan (1987) and Lianxing and McClay (1992), the textures developed may be divided into three categories: (a) cataclasis textures, (b) dislocation textures, and (c) diffusive mass transfer textures.

Cataclasis textures are exhibited predominantly by quartz, pyrite, arsenopyrite and pyrrhotite, and comprise fracturing, cracking and brecciation of sulphide and silicate minerals. Grains are shattered or crushed and an assemblage of subrounded knots and fracture-bound slithers is developed (Figures 6.10, 6.11, 6.15, 6.32). The sub-rounded knots are penetrated and/or enclosed by sulphide and silicate, and fractures are infilled by later minerals. A mosaic of fine polygonal grains is developed along some fractures, particularly in quartz (Figure 6.33). Lianxing and McClay (1992) suggested that such textures were the result of thermal annealing of quartz during fracturing.
Based on cross-cutting features, at least nine distinct and mutually exclusive fracturing events have been recognised:-

1. Continuous cracking/fracturing during the formation of crack-seal veins, in the manner described by Cox & Etheridge (1983).
2. Po 2 occurs as fill to fractures which cross-cuts asp 2 and this is interpreted to mean that a fracturing event preceded the precipitation of po 2.
3. Po 3-ccp 1-qtz 3 replaces and/or cross-cuts po 2: the texture suggests that replacement and precipitation accompanied a fracturing event (Figure 6.10 & 6.11).
4. Po 4 occurs as fill to fractures which cross-cut asp 2 and 3, and encloses those sulphides (Figure 6.21). The texture suggests that a fracturing event occurred prior to, or during, precipitation of po 4.
5. Py 2-qtz 3 replaces and cross-cuts py 1: the texture suggests that replacement and precipitation accompanied a fracturing event (Figure 6.15).
6. Lööllingite cross-cuts inclusions of po 3-ccp 1-qtz 3 and this assemblage is enclosed by asp 2. The texture suggests that a fracturing event that was restricted in duration, preceded or accompanied the precipitation of lööllingite.
7. Ccp 2 occurs as fill to fractures in pyrite, and sphalerite occurs as fill to fractures in pyrite and arsenopyrite. The textures indicate that a fracturing event succeeded the precipitation of pyrite, but preceded the precipitation of ccp 2 or sphalerite.
8. Tal 2-po 5 occur as fill to fractures in crack-seal quartz, po 3-ccp 1-qtz 3, arsenopyrite and pyrite. The textures are interpreted to mean that a fracturing event preceded, or accompanied, the precipitation of tal 2-po 5.
9. Gold and bismuth occur in fluid inclusion trails which cross-cut quartz-sulphide vein assemblages: the development of the trails accompanied a fracturing event.

**Dislocation textures** are developed in quartz and pyrrhotite as corrugation twins (Figures 6.34 & 6.35a, b), in sphalerite as dislocation slip bands (Figure 6.36) and in galena as kink bands (Figure 6.37). In sphalerite, dislocation slip has displaced crystal cleavage planes that are decorated with tiny blebs of chalcopyrite (ccp 3) after "chalcopyrite disease" (Barton & Bethke, 1987). The texture indicates that dislocation slip occurred after chalcopyrite replacement of sphalerite.

Textures associated with **diffusive mass transfer** include: (a) the development of serrated and stylolitic contacts along fractures or grain margins in quartz and pyrrhotite (Figures 6.34 & 6.35b) and (b) the development of a pervasive fabric (Figure 6.38). The fabric sub-parallels the regional $S_2$ cleavage.

**6.4.5 Supergene alteration**

Supergene alteration in the transition and lower oxidised zone of the Batman deposit is dominated by the replacement of pyrrhotite, pyrite and marcasite by supergene marcasite...
and goethite, Cu-Fe-S minerals by chalcocite, bornite, digenite and covellite (Figure 6.39) and arsenopyrite by scorodite and Fe-oxide. Alteration to supergene marcasite is evidenced as:

1. Feathery fringes to sulphides especially where pyrrhotite and pyrite lie in contact.
2. Complete pseudomorphs where the original mineralogy is totally replaced by marcasite (Figure 6.40).
3. Cellular pseudomorphs (Figure 6.40). Here the penetration of supergene marcasite along cleavage planes, fractures and grain boundaries, and the complete removal of the parent sulphide, may result in the development of a series of connected box-like cells (Roberts & Travis, 1986).
4. Bird's eye texture (Figure 6.18): aggregates of laminated pyrite and marcasite (Ramdohr, 1980).
5. Fine mottling or pitting of the sulphide grain surface.

In the upper oxidised zone of the deposit, supergene alteration is total and the remnants of oxidated sulphide minerals form veins of gossan and quartz.

6.5 PARAGENESIS

From the textural relationships, it can be established that:-
1. Co-precipitation of po 1, cubanite and tal 1 preceded the precipitation of asp 1.
2. Asp 1 precipitation was interrupted by a fracturing event and ceased prior to the precipitation of po 2.
3. Precipitation of po 2 was interrupted by a fracturing event. The fractures were filled by po 3 and qtz 3. Po 3 replaced po 2 extensively.
4. Co-precipitation of po 3 and qtz 3 was succeeded by the co-precipitation of po 3, ccp 1 and qtz 3.
5. Co-precipitation of po 3-ccp 1-qtz 3 may have been synchronous with, or succeeded by, löellingite.
6. Löellingite was succeeded and/or replaced by asp 2 precipitation. In turn, asp 2 was succeeded and/or replaced by asp 3.
7. Precipitation of asp 3 was interrupted by a fracturing event and had ceased prior to the precipitation of po 4.
8. Po 2, po 3 and po 4 were replaced by py 1, particularly in the higher levels of the orebody.
9. Py 1 precipitation was interrupted by a fracturing event. The fractures were filled by py 2 and qtz 3, and/or py 1 was massively replaced by py 2 and quartz.
10. Py 2 was succeeded by the precipitation of marc 1 and marc 2, which in turn was succeeded by the precipitation of py 3.
11. Precipitation of py 3 ceased with, or prior to, the onset of asp 4 precipitation.
12. A fracturing event followed asp 4 precipitation.
13. Precipitation of ccp 2 preceded that of sphalerite and galena.
14. Fe-bearing sphalerite was replaced by ccp 3 and Fe-poor sphalerite to produce a "chalcopyrite disease".
15. A fracturing event preceded the co-precipitation of po 5 and tal 2.
16. Po 5-tal 2 precipitation was interrupted by a fracturing event.
17. The co-precipitation of bismuthinite, hedleyite, pavonite and Bi-rich galena was synchronous to a fracturing event, and was succeeded by the co-precipitation of bismuthinite, bismuth and gold. Gold and bismuth, in association with a hypersaline-CO$_2$-CH$_4$ brine (see Chapter 7), were trapped in fluid inclusions.
18. Bismuth was replaced by bismuthinite.

The majority of these cross-cutting relationships are illustrated in Figure 6.41.

In terms of silicate/carbonate mineralogy, the precipitation of quartz accompanied each depositional event. Quartz crack-seal texture (qtz 1) was developed during sequential cracking events. A significant fracturing and dilation event preceded the precipitation of euhedral quartz (qtz 2), tourmaline and biotite, which in turn, preceded the co-precipitation of qtz 3 and the various sulphide assemblages, as detailed in Section 6.4.2. Chlorite and sericite formed during the precipitation of po 2, and continued until after the cessation of bismuthinite. The formation of biotite and muscovite ceased with, or prior to, the formation of chlorite. Calcite formed soon after asp 4 and continued until after the cessation of bismuthinite.

The quartz-sulphide veins were deformed during and after their formation, with the development of a cross-cutting fabric element, ie., $S_2$. The quartz-sulphide veins and fabric element were cross-cut by calcite-base metal veins ($V_5$). All sulphide textures were weakly to strongly modified by supergene alteration.

The paragenesis outlined above, indicates that micro-fracturing occurred throughout vein formation, with fresh sulphide being precipitated at each micro-fracture event. It is likely that fracturing accompanied dissolution of early sulphide assemblages. The textures concur with field observations that vein formation occurred pre-to syn deformational, ie., pre to syn-$S_2$.

This paragenetic sequence is presented in Figure 6.42 in table form.

6.6 PHYSICO-CHEMICAL ESTIMATES

The physico-chemical conditions that operated during the formation of quartz-sulphide veins have been estimated using the phase relationship data of sulphides that are in textural equilibrium, and summary data for silicates that are observed in active geothermal systems. These data are:-

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1. Hydrothermal biotite is developed within the alteration selvedge around quartz-sulphide veins, within quartz-tourmaline veins, or in aggregations with pyrrhotite in quartz-sulphide veins. The associations lead England (1990) to conclude that quartz-sulphide veins were emplaced under biotite stable conditions. If so, it may be interpreted that biotite formed at >330°C, based on the range of measured temperatures over which biotite has been observed in active geothermal systems (Henley et al., 1984), assuming hydrostatic pressure during formation of the veins.

2. Paragenetic studies have indicated that po 2 was succeeded by the co-precipitation of po 3 and ccp 1 in equilibrium. A temperature of approximately 330°C has been estimated for precipitation of the pyrrhotite-chalcopyrite assemblage based on experimental studies of the Cu-Fe-S system by Yund and Kullerud (1966) and Sugaki et al. (1975). These authors concluded that chalcopyrite and pyrrhotite in equilibrium were stable under dry conditions at 328°C ± 5°C, with an upper temperature of 334°C ± 17°C (Yund & Kullerud, 1966).

3. During the course of textural studies, it was determined that py 2 => marc 1 & 2 => py 3 and that FeS₂ replaces pyrrhotite extensively.
   (a) Based on experimental data by Allen et al. (1914) and Schoonen and Barnes (1991, quoting Goldhaber & Statton, 1987) the FeS₂ succession may be considered indicative of fluctuations in the pH of the hydrothermal fluid.
   (b) Schoonen and Barnes (1991) suggested that at 300°-350°C, FeS₂ forms via nucleation and growth from solution, while at 250°-300°C FeS₂ forms via an FeS precursor from an FeS supersaturated solution.
   The data suggests that pyrite and marcasite formed by sulphidation of pyrrhotite (FeS precursor) from an FeS supersaturated solution during fluctuations in fluid pH. The temperature of precipitation of FeS₂ is thereby estimated at between 250°-300°C.

4. The phase relations for the telluride and gold-bismuth-telluride systems have been studied by Afifi et al. (1988a, quoting Elliot, 1965) and Gather and Blachnik (1974), respectively. Hedleyite (Bi₁₄Te₆) is stable between 312°-266°C (1 atmosphere), while below 266°C, hedleyite reacts to form hedleyite + bismuth and finally maldonite (Au₂Bi) + bismuth + Bi₂Te (unnamed): these temperature data provide an estimate for the temperature over which hedleyite and bismuth may have co-precipitated, ie., T < 266°C.

5. The phase relations for the gold-bismuth system have been studied by Hultgren et al. (1983). A melt eutectic occurs at 241°C for ~81 mol% Bi, and constitutes a minimum temperature for the co-precipitation of gold and bismuth. Below 241°C a gold-bismuth melt will precipitate maldonite (Au₂Bi) + bismuth. Further to this, T_melt of bismuth occurs at ~271°C (Hultgren et al., 1973; Angino, 1979) and constitutes an upper temperature for precipitation of this mineral.
6. The presence of sericite as a major component in alteration assemblages has enabled the calculation of the pH of the hydrothermal fluid from thermodynamic data of Pisutha-Arnond and Ohmoto (1983) for the reactions,

\[ 3 \text{K-spar} + 2H^+ = \text{sericite} + 6 \text{quartz} + 2K^+ \]

and,

\[ 3 \text{kaolinite} + 2K^+ = 2 \text{sericite} + 3H_2O + 2H^+. \]

Assuming \( a_{K^+} = 0.1 \), pH lies within the range 2.5 to 5.3 at 330°C, this being the upper temperature for the precipitation of sericite in equilibrium with chalcopyrite and pyrrhotite (see above), and 2.3 to 4.9 at 250°C, this being the approximate temperature for the precipitation of gold and bismuth, i.e., the hydrothermal fluid was acidic. As indicated above, the FeS$_2$ succession pyrite $\Rightarrow$ marcasite 1 & 2 $\Rightarrow$ pyrite 3 suggests that pH may have fluctuated. However, pH fluctuations can not have caused a significant shift outside the sericite stability field because neither potash feldspar, nor kaolinite occur in the alteration assemblage (cf., Heinrich & Eadington, 1986; Romberger, 1988; Giggenbach, 1992).

When compared with the mineral paragenesis outlined in Section 6.5, the data above indicates that sulphide, silicate and carbonate minerals precipitated from a cooling acidic brine. Early vein formation occurred under biotite stable conditions (\( T > 330^\circ \text{C} \)). A decrease in the fluid temperature to minimum biotite stable conditions accompanied the precipitation of pyrrhotite 3 and chalcopyrite 1 (\( T \sim 330^\circ \text{C} \)). The precipitation of pyrite and marcasite during sulphidation of pyrrhotite suggest that the fluid continued to cool (i.e., \( T = 250^\circ \text{C} - 300^\circ \text{C} \)): cooling was concomitant with fluctuations in the fluid pH. Further cooling of the fluid is indicated from co-precipitation of hedleyite and bismuth (\( T < 266^\circ \text{C} \)), and subsequently, bismuth and gold (melt eutectic, \( T > 241^\circ \text{C} \)). These data provide minimum temperature constraints to the precipitation of sulphide and silicate/carbonate minerals from the hydrothermal fluid and are compared with fluid inclusion data in Chapter 8.
Figure 6.1: Photomicrograph of fibrous pyrrhotite (polars half crossed). The fine deformation lamellae at bottom right indicate that the pyrrhotite has been deformed subsequent to its precipitation (DDH BD079, 264 m). Scale bar is 500 µm long.

Figure 6.2: Photomicrograph of quartz 2 (qtz 2) against a crack-seal fibre termination (qtz 1). The crack-seal fibre is highly decorated with 1-3 µm fluid inclusions. Opaque is pyrrhotite. DDH BD080(3), 286.95 m. Scale bar is 50 µm long.

Figure 6.3: Photomicrograph of 100 µm wide selvedge of qtz 2 between crack-seal quartz (qtz 1) which contains numerous fluid inclusions, and pitted pyrrhotite 3 (po 3). Transmitted and reflected light are used simultaneously in this view. DDH BD086, 143.0 m. Scale bar is 200 µm long.

Figure 6.4: Photomicrograph of euhedral quartz (qtz 2) in pyrrhotite (black). DDH BD086, 143.0 m. Scale bar is 500 µm long.

Figure 6.5: Photomicrograph of euhedral quartz terminations (qtz 2). The terminations exhibit growth zones containing tourmaline and liquid-vapour fluid inclusions (arrows). The opaque is pyrrhotite. DDH BD080(2), 286.95 m. Scale bar is 500 µm long.

Figure 6.6: Photomicrograph of growth zone in quartz 2 (qtz 2). Tourmaline (tour) and liquid-vapour fluid inclusions (arrows) occur in the same zone. DDH BD086, 143.0 m. Scale bar is 40 µm long.
Figure 6.7: Photomicrograph of co-precipitate minerals cubanite (cub), pyrrhotite (po 1) and talnakhite 1 (tal 1) as an inclusion in arsenopyrite 1 (asp 1). DDH BD079, 264 m. Scale bar is 25µm long.

Figure 6.8: Photomicrograph of a fractured arsenopyrite euhedra (asp 1) enclosed by pyrrhotite 3 (po 3) and chalcopyrite 1 (ccp 1). The arsenopyrite is host to numerous inclusions of cubanite, pyrrhotite 1 and chalcopyrite 1 (DDH BD079, 264.0 m). Scale bar is 500µm long.

Figure 6.10: Photomicrograph of pyrrhotite 2 (po 2) in pitted pyrrhotite 3 (po 3). The texture suggests that po 3-qtz replaced po 2. DDH BD027, 172.2 m. Scale bar is 0.5 mm long.

Figure 6.11: Photomicrograph of pyrrhotite 3 (po 3)-qtz replacing pyrrhotite 2 (po 2). Fractured monoclinic tablets of quartz (centre right) probably pseudomorph arsenopyrite 1. DDH BD027, 172.2 m. Scale bar is 500 µm long.
Figure 6.9: Phase relations in the Cu-Fe-S system at 350°C, modified after Sugaki et al. (1975). Electron microprobe analyses as follows: $\Delta$ = Cu-Fe-S minerals in the association pyrrhotite 1 and cubanite, plus talnakhite 1; $X$ = chalcopyrite in equilibrium with pyrrhotite 3; $\bullet$ = talnakhite 2 in equilibrium with bismuthinite. Cubanite and talnakhite approach solid solution with moolhoekite and haycockite, respectively.
analyses of cubanite and talnakhite were conducted at the Central Science Laboratory, University of Tasmania, using a Cameca SX50 Microprobe. Analyses conditions used a 20kV accelerating voltage and a 20nA beam current. Phases include (refer to Figure 6.42) -

1. Pyrrhotite (po 1) + cubanite + talnakhite (tal 1)
2. Pyrrhotite (po 3) + chalcopyrite (ccp 1)
3. Talnakhite (tal 2) + pyrrhotite (po 5)

M.S = metalsulphur
moo = moochoke
tal-moo = phase lying between the end members talnakhite and moochoke

Table 6.1: Electron microprobe data for cubanite, talnakhite and chalcopyrite from quartz-sulphide veins in the Batman deposit.
Figure 6.12: Photomicrograph of pyrrhotite 3 (po 3) and chalcopyrite 1 (cpp 1). The smooth embayed margins are indicative of co-precipitation of the two sulphides. DDH BD076, 473.0 m. Scale bar is 500 µm long.

Figure 6.13: Photomicrograph of euhedral pyrrhotite 5 (po 5) in intimate association with gold (Au), bismuthinite (bis), talnakhite 2 (tal 2), and bismuth (Bi). DDH BD080(3), 286.95 m. Scale bar is 50 µm long.

Figure 6.14: Photomicrograph of pyrrhotite 5 (po 5) intimately associated with talnakhite 2 (tal 2), bismuth (Bi) and bismuthinite (bis). DDH BD079, 264 m. Scale bar is 50 µm long.

Figure 6.15: Photomicrograph of pyrite 1 core (py 1) in pitted pyrite 2 (py 2). The pits in py 2 are quartz filled inclusions and the white mineral is arsenopyrite (asp). Py 2-qtz also occurs as fill to fractures which cross-cut the py 1 core. The texture suggests that py 2-qtz replaced py 1. DDH BD008, 90.9 m. Scale bar is 100 µm long.

Figure 6.16: Photomicrograph: delicate fringe of pyrite 2 (py 2) and marcasite 1 (marc 1) around pyrite 1 (py 1). The marcasite is difficult to identify but is slightly whitish in colour relative to yellowish coloured pyrite. DDH BD008, 90.9 m. Scale bar is 0.5 mm.
Figure 6.17: Photomicrograph of a 10 µm wide colloform band of pyrite 2 (py 2) and marcasite 1 (marc 1) as an overgrowth on subhedral grains of pyrite 1 (py 1). Euhedra of marcasite 2 lie to the left of the colloform band and are interspersed with pyrite 3 (py 3). A single crystal of arsenopyrite 4 (asp 4) occurs at bottom left. DDH BD008, 90.9 m. Scale bar is 20 µm.

Figure 6.18: Photomicrograph of marcasite 2 (marc 2) with pyrite 3 (py 3) as interstitial fill. DDH BD023, 41.9 m. Scale bar is 0.5 mm.

Figure 6.19: Photomicrograph of löllingite (loell) in pyrite 1 (py 1). The polars are partly crossed to accentuate the strong anisotropy of the löllingite. DDH BD021, 117.5 m. Scale bar is 200 µm.

Figure 6.20: Photomicrograph of inclusions of löllingite (loell) and pyrrhotite 3-chalcopyrite 1 in arsenopyrite 2 (asp 2). The arsenopyrite is fractured and enclosed by sphalerite (sph). The texture suggests that arsenopyrite precipitation succeeded that of löllingite and po 3-ccp 1, but preceded a fracturing event and the precipitation of sphalerite. DDH BD018, 88.1 m. Scale bar is 400 µm long.
Figure 6.21: Photomicrograph of loellingite (loell) in arsenopyrite 2 (asp 2) and enclosed by arsenopyrite 3 (asp 3). These are in turn enclosed by pyrrhotite 3 (po 3). This is cross-cut by a fracture containing pyrrhotite 4 (po 4). The pinkish speckled mineral in the fracture cross-cutting the loellingite and arsenopyrite is bismuth (Bi). The texture suggests that the precipitation of loellingite was succeeded by that of arsenopyrite 2, arsenopyrite 3 and pyrrhotite 3. A fracturing event preceded the precipitation of pyrrhotite 4 and bismuth, respectively. DDH BD006, 336.1 m. Scale bar is 100 μm long.

Figure 6.22: Photomicrograph of sphalerite (sph) enclosing galena (gn) enclosing spongy anhedral chalcopyrite 2 (ccp 2). Gangue mineral is quartz. DDH BD004, 86.8 m. Scale bar is 50 μm long.

Figure 6.23: Photomicrograph of tiny blebs of chalcopyrite 3 (yellow) after chalcopyrite disease in sphalerite (sph). The white-grey coloured mineral at top left is galena (gn). The lamellar masses at centre and bottom right are supergene marcasite that exhibit birdseye texture. DDH BD087, 297.78-279.94 m. Scale bar is 200 μm long.

Figure 6.24: Photomicrograph of talnakhite (tal 2) partly enclosed by bismuthinite (bis). A gold grain (Au) occurs at centre left. Gangue minerals are quartz (flat grey) and calcite (pitted grey). DDH BD079, 264 m. Scale bar is 50 μm long.
Figure 6.25: Photomicrograph of gold (Au) as fill to fractures in arsenopyrite (asp). DDH BD079, 264 m. Scale bar is 50 µm long.

Figure 6.26: Photomicrograph of gold (Au) in intimate association with bismuth (Bi), as fill to a fracture in quartz. DDH BD080(2), 286.95 m. Scale bar is 0.5 mm long.

Figure 6.27: Photomicrograph of gold (Au), bismuth (Bi) and bismuthinite (bis) as fill to a fracture in quartz (flat grey). Note the bismuthinite and euhedral quartz grain in calcite (cal). The texture suggests that gold, bismuth, bismuthinite, calcite and quartz precipitated at the same time. DDH BD080(3), 286.95 m. Scale bar is 0.5 mm long.

Figure 6.28: Photomicrographs.
A. Trails of bismuth-gold 2-20 µm inclusions (black) along a healed fractures in quartz.
B. 10-15 µm gold-bismuth and bismuth only inclusions in quartz (reflected light).
C. Same as B (transmitted light). DDH BD080, 286.15 m. Photo-micrographs reproduced with permission of R.N. England (Thank you!!).

Figure 6.29: Photomicrograph of bismuth (Bi), gold (Au), hedleyite (hed) and bismuthinite (bis) within a fracture in pitted pyrrhotite 3 (po 3). The gold and bismuth enclose and partly replace hedleyite and bismuthinite. The texture suggests that the co-precipitation of hedleyite and bismuthinite preceded the co-precipitation of gold and bismuth. DDH BD080(3), 286.95 m. Scale bar is 50 µm long.
Figure 6.30: Photomicrograph of bismuthinite (bis) replacing bismuth (Bi) in a fracture containing calcite (dark grey). A talnakhite grain (tal 2) lies at centre right and a gold grain (Au) lies at top right. DDH BD080(3), 286.95 m. Scale bar is 50 µm long.

Figure 6.31: Photomicrograph of an inclusion of bismuth-rich galena in coarse bismuth (Bi). DDH BD080(3), 286.95 m. Scale bar is 20 µm long.

Figure 6.32: Deformation texture. Photomicrograph of fractured arsenopyrite 1 (asp 1) in association with pyrrhotite 3 (po 3). The pyrrhotite is remobilised into the fractures. DDH BD083, 287.0 m. Scale bar is 500 µm long.

Figure 6.33: Deformation texture. Photomicrograph of decussate texture of quartz along a fracture that has offset a quartz-sulphide vein. The re-equilibration texture is the result of high strain along the fracture with thermal annealing. DDH BD001, 52.10 m. Scale bar is 1 mm long.

Figure 6.34: Deformation texture. Photomicrograph of deformed crack-seal quartz fibres. Corrugation twins occur at centre, while serrated and stylolitic contacts occur along fractures at centre right. DDH PD001, 156.51-156.61 m. Scale bar is 2 mm long.
Figure 6.35a: Deformation texture and supergene alteration. Photomicrograph of deformed pyrrhotite (po) in reflected light. Bird's eye supergene marcasite cross-cuts a pyrite (py) grain at bottom left. Scale bar is 500 µm long.

Figure 6.35b: Photomicrograph Figure 6.35a in reflected light. The deformed pyrrhotite exhibits dense corrugation twins and serrated grain margins when the polars are crossed. DDH BD080(3), 286.15 m. Scale bar is 500 µm.

Figure 6.36: Deformation texture. Photomicrograph of dislocation slip bands in sphalerite. Tiny blebs of chalcopyrite after chalcopyrite disease which are offset by slip have invaded cleavage planes in the sphalerite. An irregularity in the sphalerite crystal lattice can be seen at bottom left. DDH QD002, 144.76-144.89 m. Scale bar is 50 µm.

Figure 6.37: Deformation texture. Photomicrograph of kink bands in galena. DDH BD087, 267.78-297.94 m. Scale bar is 100 µm.
Figure 6.38: Deformation texture. Photomicrograph of BD001, 138.6 m. Diffusive mass transfer is indicated by the development of a pervasive fabric in an assemblage of pyrite (py), chalcopyrite (ccp), arsenopyrite (asp) and löllingite (loell). The fabric sub-parallel to the regional S2. DDH BD001, 138.6 m. Scale bar = 500 \( \mu \)m.

Figure 6.39: Supergene alteration. Photomicrograph of the replacement of chalcopyrite (ccp) by bornite (bor), covellite (cov) and digenite (dig). Transition zone of the Batman deposit. DDH BD027, 391.95 m. Scale bar = 100 \( \mu \)m.

Figure 6.40: Photomicrograph of supergene marcasite. Complete pseudomorphs of pyrite occur at bottom left (large squarish mass) and bottom right; the remainder comprises cellular or box-work pseudomorph after pyrite. DDH BD021, 117.5 m. Scale bar is 200 \( \mu \)m.
Figure 6.41: A schematic of the vein paragenesis for quartz-sulphide veins of the Batman deposit. The sketch illustrates the complex array of cross-cutting textures and microfractures in thin sections BD006, 336.1 m; BD008, 90.9 m; BD018, 88.1 m; BD027, 172.2 m; BD079, 264.0 m; BD080(1), 286.95 m; BD081, 267.5 m; BD081, 385.0 m and BD083, 287.0 m. (Alteration and carbonate mineralogies are not illustrated). It should be noted that a single vein may or may not record the full paragenetic sequence. Indeed, in hand specimen a vein may comprise a series of cross-cutting sulphide and/or silicate microveinlets. Alternatively, a vein may comprise a single sulphide or silicate phase. Furthermore, calcite-base metal vein textures, $S_2$ and supergene alteration have substantially modified the quartz-sulphide assemblage.
Figure 6.42: Paragenetic sequence for quartz-sulphide and quartz-tourmaline veins of the Batman deposit, Mt Todd district.
7.1 PREAMBLE

Microthermometry of fluid inclusions can provide useful information about the composition, pressure, temperature, and evolution of a hydrothermal fluid, provided that populations of inclusions can be identified that are specific to some event in the fluid history. Fluid inclusions within growth zones can be termed "primary" because they formed at the time the host mineral precipitated (Roedder, 1984). The analyses of fluid within these inclusions can provide specific information about the physico-chemical conditions at the time the host mineral precipitated. In contrast, trails of fluid inclusions that cross-cut grain boundaries are generally "secondary" in nature because they were not trapped at the time the host mineral precipitated, but at some time later (Roedder, 1984). For these inclusions it is difficult, if not impossible, to relate the fluids to any particular mineral phase and thus analyses can lead to difficulties in interpreting microthermometric data (Mccuaig & Kerrich, 1994). However, if a paragenesis of secondary fluid inclusions can be linked directly to a mineral paragenesis, it is possible to gain some information about changes in the fluid conditions during the mineralising process. This has been the aim of these investigations.

In this study more than 20 fluid inclusion types were identified in quartz in Batman veins (Hein, 1991a). From this number, five characteristic populations were selected for microstructural, microthermometric and laser Raman analyses. Based on cross-cutting relationships, the inclusion types were organised into a sequence of formation, and this genetic sequence was constrained to the mineral paragenesis outlined in Section 6.5 through a temporal and spatial association between fluid inclusion type and mineral phase. It was therefore possible to document the physico-chemistry of the mineralising fluid, and in specific, to document the nature of the fluid from which gold precipitated. Furthermore, the data collected from microthermometric and laser Raman analyses has been used to establish a fluid history.

7.2 FLUID INCLUSION PETROGRAPHY

Fluid inclusions in quartz were observed in 50 polished thin sections of quartz-sulphide veins from the Batman deposit. These fluid inclusions comprise three genetic groups:-
Group 1 fluid inclusions were generated during crack-seal growth of quartz (qtz 1) and are secondary in nature because the cross-cut grain boundaries (Figures 5.7, 6.2, 6.3). Group 2 fluid inclusions occur in growth zones within euhedral quartz terminations (qtz 2) to quartz crack-seal fibres (qtz 1) and are primary in origin. They are elongate parallel to the direction of growth of the host quartz (Figure 6.5, 6.6, 7.1 & 7.2). Group 3 secondary fluid inclusions occur in healed fractures that cross-cut qtz 1 and qtz 2.

Fluid inclusions in group 1 have not been studied because they formed during crack-seal growth of fibre quartz and before the precipitation of qtz 2 and vein sulphides (Section 6.5). They thereby do not bear a direct relationship to the mineralising event. In contrast, groups 2 and 3 fluid inclusions were studied in detail because they formed during precipitation of qtz 2 and immediately before the precipitation of vein sulphide (group 2), and during and after precipitation of vein sulphide (group 3). They therefore bear a direct relationship to the mineralising event (Figure 6.42). This relationship is detailed in Section 7.3.

Group 2 fluid inclusions constitute a single compositional type and are herein termed type A. The inclusions contain a liquid and vapour phase (Figure 7.1) and are irregular in shape with a maximum size of 40 x 18 µm. Liquid to vapour ratios (L:V) are 10-70% vapour. The fluid inclusions are accompanied within growth zones by subhedral to euhedral greenish-coloured tourmaline that occasionally project into the fluid inclusion cavity. The tourmaline, quartz and type A fluid inclusions are co-genetic.

Group 3 comprises 20 fluid inclusion types. Four characteristic populations were recognised within healed fractures that cross-cut type A fluid inclusions and/or qtz 2-tourmaline:-

**type B** contain liquid + vapour + 3 daughter minerals (NaCl + carbonate + an unidentified acicular mineral). The fluid inclusions are irregular in shape (Figure 7.3a, b) with a maximum size of 18 x 15 µm. Tourmaline may occur as accidentally trapped solid inclusions in the fluid inclusion wall and are not co-genetic with the inclusion contents. L:V ratios are 5-10% vapour. Daughter minerals are ubiquitous and show constant volume proportions to other phases, suggesting they are true daughter minerals (Roedder, 1984). The presence of halite indicates fluid saturation with respect to NaCl (Roedder, 1984; Reynolds, 1990), while the presence of carbonate daughter minerals indicates fluid saturation with respect to carbonate (Henley et al., 1984).

**type C** contain liquid + vapour and are accompanied by gold, gold-bismuth and bismuth grains in the same healed fracture (Figures 7.4 & 7.5). L:V ratios are 70-95% vapour (vapour-rich inclusions) and rarely, 1-2% vapour (liquid-rich inclusions) (Figure 7.6). Tourmaline may occur as accidentally trapped solid inclusions in the fluid inclusion wall (Figure 7.4). Occasionally, the inclusions wet gold and/or bismuth grains and, in some instances, fluid-filled tubes connect these grains (Figure 7.6). The inclusions are irregular in shape with a maximum size of 15 x 4 µm, and are petrographically similar to CO₂-rich fluid inclusions that
accompany gold and telluride grains in quartz from the Sigma mine in Canada (cf. Robert & Kelly, 1987).

**Type D** contain vapour, and in several cases, a thin film of liquid. The inclusions are smooth and irregular to crystal-negative in shape (Figure 7.7) with a maximum size of 20 x 20 µm. L:V ratios are 75 to >95% vapour. Tourmaline may occur as accidentally trapped solid inclusions in the fluid inclusion wall.

**Type E** contain liquid ± vapour ± NaCl daughter mineral ± unidentified acicular daughter mineral (Figure 7.8 & 7.9 a-c). The inclusions are irregular in shape and several types occur along the same healed fracture, or within the same cluster. They measure up to 36 x 5 µm in size. Where a vapour phase is present, L:V ratios are 0-5% vapour. Daughter minerals do not show constant volume proportions to other phases, or are absent. This suggests that the vapour phase and daughter minerals formed before the inclusions coalesced, i.e., the inclusions have "necked down" (Roedder, 1984). The presence of halite in some inclusions indicates fluid saturation with respect to NaCl (Roedder, 1984; Reynolds, 1990).

### 7.3 Fluid Inclusion Type and Mineral Paragenesis

The nature of cross-cutting relationships of the four group 3 populations were studied in detail, and their relationship to type A fluid inclusions is illustrated in Figure 7.10.

- Type A fluid inclusions in growth zones are cross-cut by type C and D.
- Type B are cross-cut by type C and D.
- Type C are cross-cut by type D and E.

Types A and C constrain the fluid inclusion sequence to the mineral paragenesis outlined in Section 6.5 through their temporal and spatial association with qtz 2-tourmaline and gold-bismuth, respectively. Type B, D and E fluid inclusions do not tie the inclusion sequence to the mineral paragenesis, but are constrained via the cross-cutting relationships outlined above.

### 7.4 Microthermometry

#### 7.4.1 Methods and procedures

Of the 50 polished thin sections originally examined, 22 contained group 2 and 3 fluid inclusions: doubly polished wafers were cut for these 22 sections from residue thin section blocks. However, only 9 wafers proved suitable for microthermometry of types A, B, C, D and E fluid inclusions. The deformed nature of the other samples, the dense secondary fluid inclusion distribution, and/or the juxtaposition of fluid inclusions to opaque minerals resulted in poor transmission of light and excluded them from analyses. Suitable wafers were carefully fragmented into chips and most chips were run only once (see Section 7.4.5):
freezing preceded heating. Occasionally, two to five fluid inclusions in the same growth zone or healed fracture were "batch" analysed. Microthermometry was conducted at the Department of Geology at the University of Tasmania using a USGS gas-flow heating/freezing stage, calibrated against synthetic fluid inclusions. The accuracy of Trendicator thermocouple measurements was ± 0.3°C for cooling and ± 3.5°C for heating.

7.4.2 Notation and calculations

Phase changes observed during freezing are recorded as (1) final melting temperature of CH₄ solid = TmCH₄, (2) final melting temperature of CO₂ solid = TmCO₂, (3) eutectic (first) melting temperature of ice = Te, (4) final melting temperature of ice = Tm, (5) final melting temperature of hydrohalite = Tmhydro, and (6) final melting temperature of CO₂ clathrate = TmCO₂clath. TmCH₄ and TmCO₂ are defined as the transition from S+L+V to L+V by the disappearance of the solid phase. The first melting temperatures of CO₂ and CH₄ are recorded as TiCO₂ and TiCH₄, respectively, and are defined as the transition from a solid coexisting with liquid or vapour (ie., S+L or S+V) to S+L+V. They are not discussed in the text.

Phase changes observed during heating are recorded as (1) homogenisation of the CO₂-rich vapour phase = ThCO₂, (2) homogenisation of the H₂O-rich vapour phase = Thvap, (3) final temperature of dissolution of the halite daughter mineral = Thhalite, (4) final temperature of the dissolution of the unidentified daughter phase = Thdt unk, (5) temperature at which the last phase homogenised = Thtot, and (6) temperature at which the fluid inclusion decrepitated = Td. For the special case involving the melt and solidification of bismuth grains in contact with type C fluid inclusions, data are recorded as (1) first melting temperature of bismuth = TeBi, and (2) temperature of solidification of bismuth = TsolBi. Data have not been recorded for the dissolution of carbonate daughter minerals because dissolution did not occur (Section 7.4.4).

Weight percent NaCl equivalent values (hereafter wt% NaCl eq) were calculated from:-

1. The final melting temperature of H₂O (Tm) using a least squares regressions curve calculated by Potter et al. (1978),

\[ ws = 0.00 + 1.76958 \theta - 4.2384 \times 10^{-2} \theta^2 + 5.2778 \times 10^{-4} \theta^3 \pm 0.028, \]

where \( ws \) is weight percent NaCl in solution, and \( \theta \) is the freezing point depression in degrees Celsius ( \( \theta = 0 \) - freezing point).

2. The homogenisation temperature of NaCl (Thhalite) using a least squares regression curve quoted by Reynolds (1990),

\[ \text{wt}\% \text{ NaCl eq} = 26.218 + 0.0072T + 0.000106T^2 \pm 0.5 \text{ wt\%}, \]

where \( T \) is the dissolution temperature of halite in degrees Celsius.

3. The final melting temperature of CO₂ clathrate (TmCO₂) using a least squares regression curve calculated by Bozzo et al. (1975),

\[ \text{wt}\% \text{ NaCl} = 0.05288(10-T)(T + 29.361), \]

where \( T \) is the melting temperature of CO₂ clathrate in degrees Celsius.
7.4.3 Type A fluid inclusions

Data gained from freezing of type A fluid inclusions are summarised in Table 7.1 and Figure 7.11a. On freezing, the vapour phase collapsed and the inclusion contents solidified. In two samples, residual liquid indicated that CH\textsubscript{4} and/or N\textsubscript{2} might be present (van den Kerkhof, 1988). TmCH\textsubscript{4} were -82.9 and -80.0°C which agrees with an experimentally determined pure CH\textsubscript{4} critical temperature of -82.1°C (Burruss, 1981) rather than a N\textsubscript{2} critical temperature of -147.0°C (van den Kerkhof, 1988), confirming that CH\textsubscript{4} is present in these samples. TmCO\textsubscript{2} range from -61.5 to -56.2°C, with a mode of -57.4°C which is slightly lower than the CO\textsubscript{2} eutectic of -56.6°C (Burruss, 1981), but a depression in the CO\textsubscript{2} critical temperature is expected given the presence of CH\textsubscript{4} in some inclusions. The first melting temperatures of NaCl-H\textsubscript{2}O ice (Te) range from -32.4 to -25.9°C (mode at -29.0°C), compared with a H\textsubscript{2}O-NaCl eutectic of -21.1°C (Shepherd et al., 1985): the lower temperature range is indicative of salts in addition to NaCl. The salts are likely to be MgCl\textsubscript{2}, CaCl\textsubscript{2}, LiCl, AlCl\textsubscript{3} and/or FeCl\textsubscript{3} which cause a depression in the NaCl eutectic to -35.0°C, -52.0°C, -78°C, -55°C and -55°C, respectively (Crawford, 1981; Roedder, 1984). Final melting temperatures (Tm) range from -17.3 to -0.8°C. Salinity estimates derived from Tm by using a least squares regression curve of Potter et al. (1978) range from 1.39 to 20.6 wt% NaCl eq indicating that these inclusions are weakly to moderately saline.

Homogenisation temperatures (T\textsubscript{htot}) for type A fluid inclusions are summarised in Figure 7.11b and Table 7.1. They plot as two groups that range from 357.8 to 418.8°C and 454.8 to 458.2°C, with homogenisation into the vapour or liquid phase. For fluid inclusions in the same growth zone (Sample BD086a), homogenisation occurred into the liquid or vapour phase at similar temperatures. This provides strong evidence of boiling of the hydrothermal fluid at the time of entrapment (Roedder, 1981 & 1984; Bodnar et al., 1985).

Several irregularities are noted in the microthermometric data for type A fluid inclusions (Table 7.1):-
(a) The wide range of L:V ratios (10-70 % vapour),
(b) The wide range in Tm data,
(c) The occurrence of a T\textsubscript{htot} peak between 454.8 and 458.2°C,
(d) The presence of CO\textsubscript{2} or CO\textsubscript{2}-CH\textsubscript{4} in some aqueous fluid inclusions,
(e) The presence of CO\textsubscript{2} in some aqueous fluid inclusions which show evidence of boiling.

In the first instance, they may be explained by considering the nature of a boiling hydrothermal fluid. Roedder (1984) noted that boiling systems are heterogeneous and therefore, it can be expected that individual fluid inclusions will trap different ratios of the phases present, ie., different ratios of aqueous saline fluid and/or gas. It follows that the inclusions will exhibit a range in L:V ratios. Furthermore, Roedder (1981 & 1984) stated that inclusions trapped from a boiling fluid will give highly irregular temperature, density (dependent on Tm) and pressure data, and a wide range in Tm and T\textsubscript{htot} data may therefore be predicted. For example, a fluid inclusion trapping a small aqueous vapour bubble will
homogenise at a higher temperature than that which existed at entrapment and thus give a meaningless Thtot. It is therefore possible that the wide range of L:V ratios and Tm values, and a Thtot peak between 454.8 and 458.2°C, is an artefact of boiling of the hydrothermal fluid.

However, a further complication is added by the presence of varying proportions of CO₂ or CO₂-CH₄ in some aqueous fluid inclusions, and furthermore, by the presence of CO₂ in some fluid inclusions which show evidence of boiling (Sample BD086a). Certainly, the CO₂ or CO₂-CH₄ content of these inclusions is negligible because clathrate formation was not observed (Sisson et al., 1981). The apparent absence of CH₄ in some CO₂-bearing inclusions is probably the result of small amounts of CH₄ remaining undetected, particularly because CH₄ is completely miscibility in CO₂ (Burruss, 1981, quoting Unruh & Katz, 1949, and Robinson & Hutton, 1962). Nevertheless, the existence of the gases in some inclusions, and in inclusions which show evidence of boiling, indicates that two immiscible fluids were present at the time of entrapment, and during boiling of the aqueous phase, ie., CO₂-CH₄-rich bubbles (very small) were present in a boiling aqueous saline brine. In this situation, the entrapped fluid will not only give highly irregular temperature, density and pressure data, and a wide range in Tm and Thtot data (Roedder, 1984), but also exhibit a range in composition. Therefore, it is considered that the range of L:V ratios and Tm data, a Thtot peak between 454.8 and 458.2°C, and the presence of CO₂ or CO₂-CH₄ in several aqueous fluid inclusions, some which show evidence of boiling, is the result of entrapment of immiscible fluids. In particular, it is considered that type A fluid inclusions record the entrapment of three coeval fluid phases; a CO₂-CH₄ rich vapour, H₂O-rich vapour (steam), and a saline aqueous brine.

7.4.4 Type B fluid inclusions

Freezing data for type B fluid inclusions is presented in Table 7.2 and summarised in Figure 7.12a, b. Complete solidification was not achieved in two samples suggesting that CH₄ and/or N₂ might be present (van den Kerkhof, 1988). TmCH₄ were -83.6 and -82.1°C and indicate that the residual liquid was CH₄ because the TmCH₄ values approximate the CH₄ critical temperature of -82.1°C of Burruss (1981). TmCO₂ range from -64.9 to -57.7°C with a mode at -61.8°C. The depression in TmCO₂ values compared with a CO₂ critical temperature of -56.6°C (Burruss, 1981) is the result of a small amount of CH₄ in addition to CO₂ (Crawford et al., 1979b).

Te data range from -51.9 to -41.9°C (Figure 7.12a) which is close to the NaCl-CaCl₂-H₂O melt eutectic of -52.0°C (Crawford, 1981) indicating that CaCl₂ is present in addition to NaCl (Crawford et al., 1979b). The salts AlCl₃, FeCl₃ and/or LiCl may also be present since they depress the NaCl eutectic to -55°C, -55°C and -75°C, respectively (Crawford, 1981; Roedder, 1984). In the case of MgCl₂, the NaCl-CaCl₂-H₂O eutectic is depressed to -57°C.
(Crawford et al., 1979b quoting Luzhnaya & Vereschchetina, 1946): the existence of these salts cannot be confirmed (NaCl and Ca\textsuperscript{2+} are indicated by the presence of halite and carbonate daughter minerals). Nevertheless, at the measured eutectic melting temperatures (Te) the inclusions contained rounded ice rafts, strongly birefringent hydrohalite (NaCl\textsubscript{2}H\textsubscript{2}O), liquid and a small vapour bubble (CO\textsubscript{2} + CH\textsubscript{4}). The final melting temperature of ice (T\textsubscript{m}) range from -33.2 to -24.4°C (Figure 7.12a), while the final melting temperature of hydrohalite (T\textsubscript{mhydro}) range from 4.1 to 11.2°C (Figure 7.12b). The final disappearance of hydrohalite was accompanied by the growth of a halite daughter crystal(s). The persistence of hydrohalite to temperatures higher than its true melt point of 0.1°C is typical of the metastable or sluggish melt behaviour of the salt hydrate (Roedder, 1984; Vanko et al., 1988).

Heating behaviour for type B fluid inclusions is complex with data recorded for the dissolution of halite (Thhalite) and an unidentified acicular mineral (Thdt unk), and the homogenisation of vapour phase (Thvap) (Table 7.2 and Figure 7.12c): Thdt unk range from 108.0 to 210.8°C, Thvap range from 217.4 to 380.1°C, and Thhalite range from 195.2 to 380.1°C. Th\textsuperscript{tot} range from 324.8 to 380.1°C. Typically Thdt unk preceded Thvap, which preceded Thhalite. However, in several cases halite dissolution preceded homogenisation of the vapour phase. Carbonate daughter minerals did not dissolve. This unusual behaviour does not appear to result from differential phase behaviour caused through "necking" of the inclusion. In fact, the carbonate minerals are ubiquitous and show constant volume proportions to other phases suggesting they are not accidentally trapped inclusions, or the product of necking. It is therefore concluded that carbonate daughter mineral did not dissolve because they did not attain equilibrium with the liquid (Roedder, 1984).

The bulk composition of type B fluid inclusions has been estimated in the system NaCl-CaCl\textsubscript{2}-H\textsubscript{2}O from T\textsubscript{m} and Thhalite data of three samples using the method of Vanko et al. (1988). The results are presented in Table 7.2 and summarised in Figure 7.13. NaCl/CaCl\textsubscript{2}/H\textsubscript{2}O bulk compositions are 26.0/9.0/65.0, 37.0/13.0/50.0 and 30.0/8.0/62.0 with total salinities of 35, 50 and 38 wt% NaCl-CaCl\textsubscript{2}, respectively. Na:Ca ratios are 2.9, 2.8 and 3.7, respectively, and are well below the sea water ratio of 26.0 (Vanko et al., 1988).

In analysis, the fluid inclusion data presented above and in Section 7.2 indicates that a single fluid type was present at the time of entrapment because (a) only one inclusion type occurs along a given healed fracture, (b) L:V ratios are relatively constant at 5 to 10 % vapour, (c) daughter minerals are ubiquitous, and (d) daughter minerals show constant volume proportions to other phases. Therein the fluid is summarised as a CO\textsubscript{2}-CH\textsubscript{4} bearing, aqueous brine of high total salinity (35-50 wt% NaCl-CaCl\textsubscript{2}) that is saturated with respect to NaCl and carbonate. This fluid has been modelled using experimental data for the H\textsubscript{2}O-CO\textsubscript{2}-NaCl system (Takenouchi & Kennedy, 1965; Bowers & Helgeson, 1983a; Franck, 1985, Gehrig et al., 1986), and the H\textsubscript{2}O-CH\textsubscript{4}-NaCl and H\textsubscript{2}O-CH\textsubscript{4}-CaCl\textsubscript{2} systems (Krader & Franck, 1987). The data indicates that:-
1. For the pure CO₂-H₂O system, the addition of small amounts of NaCl broadens the CO₂-H₂O solvus, and shifts the solvus peak to higher temperatures (Takenouchi & Kennedy, 1965; Hollister & Burross, 1976; Bowers & Helgeson, 1983a; Franck, 1985; Gehrig et al., 1986). Thereby, for a given pressure (< 3 kbars), the solubility of CO₂ in a saline aqueous brine is limited (< 15 mol %) at temperatures less than 400°C (Sisson et al., 1981; Naden and Shepherd, 1989). For example, the maximum solubility of CO₂ in a NaCl-H₂O solution of 20 wt% is less than 3 mole % for a pressure of 200 bars and a Thtot of less than 400°C (Figure 7.14). Furthermore, the solubility decreases further with increasing salinity for a given temperature and pressure (Naumov et al., 1974).

2. For the pure CH₄-H₂O system, the addition of small amounts of NaCl broadens the CH₄-H₂O solvus and shifts the solvus peak to higher temperatures (Franck, 1985; Krader & Franck, 1987). For example, the isopleth for a solution containing 17 mol % CH₄ at a given pressure (< 2 kbars) will shift from approximately 400°C to 500°C by the addition of only 0.53 wt% NaCl (Krader & Franck, 1987), and even further with the addition of more salt (Figure 7.15). The effect caused by the addition of CaCl₂ is similar to, but slightly smaller than that of NaCl, for equal molar salt concentrations (Franck, 1985). The data suggests that for a given temperature and pressure, CH₄ solubility will decrease as salinity increases or, as stated by Naden and Shepherd (1989), only small amounts of methane (< 5 mol %) are soluble on aqueous brines over a wide range of temperatures and pressures (up to 3 kbar and 400°C). Therefore, CH₄ is expected to have a very limited solubility in a hypersaline NaCl or CaCl₂ brine.

3. The effect of adding CaCl₂ to a CO₂-H₂O solution may be predicted from data by Franck (1985) and Krader & Franck (1987), i.e., it is expected to be similar to, but smaller than the effect caused by the addition of NaCl to a CO₂-H₂O solution: it should broaden the CO₂-H₂O solvus and shift the solvus peak to higher temperatures. It follows that the solubility of CO₂ in a highly saline CaCl₂-rich aqueous brine will be limited at temperatures less than 400°C.

In summary, the experimental data suggests that the solubility of CO₂ and CH₄ in undersaturated NaCl-H₂O or CaCl₂-H₂O solutions is dependent on salinity, and decreases with increasing salinity for a given pressure and temperature. It is therefore expected that the solubility of CO₂ and CH₄ in a saturated NaCl-CaCl₂-H₂O solution will be very low (at a given pressure and temperature). With respect to type B fluid inclusions, for a given pressure and Thtot range of 324.8 to 380.1°C, the high calculated total salinities preclude the possibility of significant quantities of CO₂ and CH₄, and this together with the fact that clathrate formation was not observed supports the conclusion that the CO₂-CH₄ content of fluid inclusions is negligible.
7.4.5 Type C fluid inclusions

Thermometric data for type C fluid inclusions is summarised in Figure 7.16a-c and Table 7.3. As described previously (Section 7.2), type C inclusions constitute a special case because they accompany gold, gold-bismuth, or bismuth grains in the same healed fracture, and because the inclusions occasionally wet gold and/or bismuth grains. This in itself caused a few technical difficulties and a large number of inclusions (> 20) were excluded from analyses because their juxtaposition to accompanying opaques resulted in the poor transmission of light.

Concerning the data set in Table 7.3, temperatures were recorded only when results could be reproduced. The temperature of melting of CH\textsubscript{4} solid (TmCH\textsubscript{4}) range from -85.7 to -80.1°C (Figure 7.16a) which agrees with an experimentally determined CH\textsubscript{4} critical temperature of -82.1°C (Burruss, 1981): the presence of N\textsubscript{2} is precluded because the values do not approximate the N\textsubscript{2} critical temperature of -147.0°C of van den Kerkhof (1988). This is consistent with laser Raman analyses of type C inclusions which indicated that N\textsubscript{2} was not present (Section 7.5.2). TmCO\textsubscript{2} range from -67.9 to -57.6°C, with a mode of -60.1°C, which is lower than the CO\textsubscript{2} critical temperature of -56.6°C. The depression in TmCO\textsubscript{2} is indicative of CH\textsubscript{4} in addition to CO\textsubscript{2}. The first melting temperature of NaCl-H\textsubscript{2}O ice (Te) range from -28.1 to -21.6°C, compared to a NaCl-H\textsubscript{2}O melt eutectic of -21.1°C (Shepherd et al., 1985). The lower temperature in Te are indicative of minor salts in addition to NaCl, and these are likely to be KCl, MgCl\textsubscript{2}, CaCl\textsubscript{2}, AlCl\textsubscript{3}, FeCl\textsubscript{3} and/or LiCl which depress the NaCl eutectic to -22.6, -35.0, -52.0, -55, -55 and -75°C respectively (Crawford, 1981; Roedder, 1984). Temperatures of final melt of ice (Tm) in three inclusions were -14.4, -13.4 and -13.1°C.

The warming of type C fluid inclusions subsequent to freezing resulted in the development of pale white CO\textsubscript{2} clathrate. TmCO\textsubscript{2} clath range from -7.8 to -4.0°C (Figure 7.16b). Salinity estimates calculated from TmCO\textsubscript{2} clath using a least squares regression curve of Bozzo et al. (1975) range from 18.7 to 20.2 wt% NaCl eq. However, these salinities may not be useful indicators of the true salinity because the presence of CH\textsubscript{4} is known to significantly affect the melt temperature of CO\textsubscript{2} clathrate and therefore, salinity estimates based on depression of the CO\textsubscript{2} clathrate melting point by NaCl (Hollister & Burruss, 1976; Collins, 1979). Collins (1979) has demonstrated that the addition of CH\textsubscript{4} shifts the CO\textsubscript{2} invariant melting point to higher pressure and temperature. Thus the effect of CH\textsubscript{4} is to counteract the depression of the CO\textsubscript{2} clathrate melting point by NaCl. The present salinity estimates therefore underestimate the true salinity, but constitute a minimum salinity for these inclusions.

Homogenisation temperatures for type C fluid inclusions are summarised in Figure 7.16b,c and Table 7.3. Two heating runs gave ThCO\textsubscript{2} of 16.7°C and 17.2°C which correspond to CO\textsubscript{2} densities of 0.8 to 0.85 g/cc (Roedder, 1984, Figure 8-9), with homogenisation to liquid CO\textsubscript{2}. ThCO\textsubscript{2} are lower than the pure CO\textsubscript{2} critical point of 31.1°C (Roedder, 1984) which
implies that the CO₂ in the inclusions is less dense than pure CO₂ (van den Kerkhof, 1988), but the TmCO₂ values lower than the CO₂ critical temperature had already indicated that this was not pure CO₂. Thtot have not been recorded because homogenisation of the vapour phase did not occur. However, several phase changes associated with melting and solidification of bismuth, and expansion and contraction of the vapour phase, were observed for type C fluid inclusions which wet bismuth and gold-bismuth grains.

On heating, the vapour bubble slowly contracted, but at TeBi there was a rapid expansion of the vapour phase. With further heating, it contracted again. Decrepitation of the inclusion preceded homogenisation (maximum heating was to 500°C). Where samples were not run to decrepitation, cooling of the inclusion accompanied expansion of the vapour phase to TsolBi, at which time it rapidly contracted. Further cooling to ambient temperature accompanied the expansion of the vapour phase and the inclusion returned to its original state as near as could be determined. During these heating runs, the shape of the inclusion and that of the associated opaques did not change.

In interpretation of the phase changes described, TeBi and TsolBi are assumed to correspond with the melt and solidification of bismuth, respectively, because:-
1. Only bismuth-gold, bismuth or gold were observed in intimate association with type C inclusions in the same healed fracture.
2. Bismuth on solidifying, increases its volume by approximately 3% (Vasilenko, 1973; Angino, 1979), or conversely decreases its volume on melting. A vapour bubble in an inclusion which is wetting the surface of a bismuth or gold-bismuth grain may be expected to expand on heating at the melting point of bismuth, or contract on cooling on solidification of bismuth. This appears to be what is happening.
3. TeBi range from 241.9 to 259.9°C (Figure 7.16c) which agrees with a gold-bismuth eutectic at X₇₉₋=0.81 (Figure 7.17) of 241.0°C. TsolBi are 191.6 and 220.6°C, and probably reflect solidification during metastable cooling of the bismuth-rich liquid.

The implication arising from the TeBi values is that each grain was composed of gold and bismuth rather than pure metal because the melt eutectic lies at >1300 K (>1025°C) for gold (Hultgren et al., 1963) and at 544 K (271°C) for bismuth (Hultgren et al., 1963; Angino, 1979), respectively. The proportion of gold and bismuth in bismuth-gold grains can be estimated from the phase diagram for the bismuth-gold system. Using an upper TeBi of 259.9°C, the X₇₉₋ in the gold-bismuth grains range from ~ 0.79 to 0.96, ie., 79 to 96% bismuth.

7.4.6 Type D fluid inclusions
Freezing of type D fluid inclusions resulted in the double collapse of the vapour phase and the partial solidification of the inclusion contents, thus indicating the CH₄ and/or N₂ is present. TmCH₄ range from -86.5 to -80.3°C (Figure 7.18a) with a reasonably well defined
maximum at -81.7°C. This is close to the CH₄ critical temperature of -82.1°C of Burruss (1981) and confirms that the residual unfrozen liquid is CH₄. TmCO₂ range from -64.0 to -56.7°C with a strong peak at -58.6°C, this being lower than the CO₂ critical temperature of -56.6°C (Burruss, 1981) and the result of the presence of CH₄. For inclusions in which a liquid phase could easily be identified, the formation of ice rafts were recognised from indents on the surface of the vapour phase. However, the ice rafts were extremely small in size, and coupled with the opaque nature of the fluid inclusions, were impossible to quantify in terms of first or final melting temperatures (of ice).

In contrast, the formation of a CO₂ clathrate was easily observed in three inclusions containing a liquid phase. TmCO₂ clath were -15.2, -15.0, and -12.1°C (Figure 7.18b) from which salinity of 20.0, 19.0 and 20.1 wt% NaCl eq were calculated using a least squares regression curve of Bozzo et al. (1975). However, as indicated previously (Section 7.4.5), these salinities under-estimate the true salinity because the effect of CH₄ is ignored in these calculations. Thus, the salinities at best, constitute a minimum for type D inclusions.

Data gained from heating type D inclusions is limited to those containing a liquid phase. Four heating runs gave ThCO₂ of 11.4, 13.5, 15.6 and 16.9°C (Figure 7.18b), which correspond to CO₂ densities of 0.8 to 0.85 g/cc (Roedder, 1984; Figure 8-9), with homogenisation to liquid CO₂. The values are lower than the pure CO₂ critical point of 31.1°C (Burruss, 1981) which implies that the CO₂ in the inclusions is less dense than pure CO₂ (van den Kerkhof, 1988), but this was apparent from the low TmCO₂ data. Thtot data range from 343.4 to 381.1°C (Figure 7.18c). These temperatures are expected to be high because it was difficult to determine the exact point at which the last droplet of liquid homogenised. Homogenisation was typically into the vapour phase.

### 7.4.7 Type E fluid inclusions

Type E fluid inclusions containing liquid and/or vapour and/or daughter minerals are typical of fluid inclusions that have "necked down" after entrapment (Roedder, 1984). Therefore, thermometric data can at best, only broadly indicate the nature of the parent fluid. The data are expected to show a spread of values in response to "necking" of the inclusion after formation of the daughter and/or vapour phase(s). The presence of halite in some inclusions is indicative of saturation with respect to NaCl (Figure 7.8 & 7.9a), or salinities of approximately 26.3 wt%.

TmCO₂ range from -64.1 to -52.1°C (Figure 7.19a & Table 7.5) which approximates the CO₂ critical temperature of -52.2°C of Burruss (1981). A single TmCH₄ of -83.5°C indicates that CH₄ might be present. First melting temperatures of ice (Tm) range from -49.7 to -42.8°C (Figure 7.19a), compared with a NaCl-H₂O melt eutectic of -21.1°C (Shepherd et al., 1985). The lower temperature range is significant of salt in addition to NaCl and this is likely to be CaCl₂ which depresses the NaCl eutectic to -52.0°C (Crawford et al., 1979b; Crawford, 1981). MgCl₂, AlCl₃, FeCl₃ and /or LiCl₃ may also be present (Crawford, 1981; Roedder,
In some inclusions, the characteristic development of salt hydrate in the presence of ice, liquid and vapour was similar to that exhibited by type B fluid inclusions (Figure 7.9 a-c), while for others, the salt hydrate did not form. Consequently, $T_m$ data show a broad range of values from -29.1 to -11.1°C (Figure 7.19a), the lower temperature values belonging to those inclusions in which hydrohalite did not form. $T_{m\text{hydro}}$ range from 4.9 to 19.6°C (Figure 7.19b). Dissolution of hydrohalite was accompanied by the growth of a daughter halite crystal(s).

Data gained from heating of type E inclusions is summarised in Figure 7.19c and Table 7.5, and are typically complex because the homogenisation of vapour and the dissolution of daughter minerals was not progressive but occurred in a random fashion. $T_{\text{hav}}$ range from 110.8 to 267.9°C and $T_{\text{halite}}$ range from 160.1 to 270.3°C. Th for two inclusions containing an unidentified acicular mineral occurred at 233.4 and 257.6°C. Th of all phases occurred over a wide range of temperatures from 121.7 to 267.9°C. It may be argued that this range broadly constrains the temperature at which type E fluid inclusions formed, and that type E formed at lower temperature relative to type A, B, C, and D fluid inclusions.

Salinities where calculated from $T_m$ for type E inclusions that did not contain halite daughter minerals from a least squares regression curve of Potter et al. (1978), and from $T_{\text{halite}}$ for halite-bearing inclusions from a least squares regression curve quoted by Reynolds (1990). These range from 15.1 to 25.7 wt% NaCl eq and 29.6 to 35.9 wt% NaCl eq, respectively, but only provide an approximate salinity because the inclusions have "necked".

7.5 LASER RAMAN ANALYSES

7.5.1 Introduction
Laser Raman spectroscopy has been used in this study to identify the composition and relative concentration of the gas phase(s) in type B, C and D fluid inclusions, salinity in type C inclusions, and to determine the nature of the carbonate daughter mineral in type B fluid inclusions. The data collected are summarised in Table 7.6a-e. Laser Raman analyses were conducted at the Australian Geological Survey Organisation in Canberra, Australia, using a Dilor Microdil-28® laser Raman spectrometer. A description of the principles and practical aspects of laser Raman spectroscopy is given by Roedder (1984), Pasteris et al. (1988), Dubessy et al. (1989), Memagh and Wilde (1989), and Khin Zaw et al. (1994).

7.5.2 Analyses of gas phase
The composition and relative concentration (in mol%) of the gas phases in type B, C and D fluid inclusions (1, 4 and 4 analyses, respectively) were calculated in the method of Chou et al. (1990) and Pasteris et al. (1988). The gas phase in type B and D inclusions contained CO$_2$ in excess of CH$_4$ (Table 7.6 a, c). N$_2$ was not detected. The CO$_2$ concentration for a single type B inclusion was 94 mol% compared with 6 mol% CH$_4$, while the CO$_2$ concentra-
tion for type D inclusions ranged from 85 to 89 mol% relative to 11 to 15 mol% CH₄.

The gas phase in type C inclusions were scanned for CH₄, CO₂, N₂ and H₂S, however, only CH₄ was detected (Table 7.6b). This is surprising considering that CO₂ is indicated from microthermometric data, i.e., TmCO₂, TmCO₂ clath and ThCO₂ data. The discrepancy may lie in the fact that CH₄ is strongly partitioned into the vapour phase with respect to CO₂ (Burruss, 1981; Roedder, 1984), but it seems unusual that CO₂ was not detected. Clearly, the type C fluid inclusions require further investigation.

7.5.3 *Total salinity measurements*

The total salinity of type C fluid inclusions were measured using laser Raman analyses after the method of Mernagh and Wilde (1989). Total salinities range from 28.8 to 33.9 wt% NaCl eq (n = 7) with an average of 29.5 wt% NaCl eq (Table 7.6d), and imply saturation with respect to NaCl. The values do not agree with salinity estimates calculated from TmCO₂ clath data of 18.7 to 20.2 wt% NaCl eq, but the clathrate-based salinities are expected to represent minimum salinities because they are calculated on depression of the CO₂ clathrate melting point by NaCl and assume XCH₄ is zero. In contrast, the spectra-based salinities provide a good estimate of the total fluid salinity, with an average error of <2 wt% NaCl (Mernagh & Wilde, 1989). It is therefore concluded that total salinity values for type C fluid inclusions average 29.5 wt% NaCl eq ± 2 wt%.

7.5.4 *Type B fluid inclusions: the nature of carbonate daughter minerals*

Raman spectral peaks were obtained in the 1083.9 to 1086.4 spectral window (in relative wavelength number, cm⁻¹) from carbonate daughter minerals of four type B fluid inclusions (Table 7.6e), and correspond with spectral peaks for calcite at 1087 cm⁻¹, rhodochrosite at 1087 cm⁻¹ (CaMn(CO₃)₂), kutnahorite at 1086 cm⁻¹ (MnCO₃) and aragonite at 1085 cm⁻¹, as stated by Herman *et al.* (1987). The data indicate that CaCO₃, MnCO₃ or (CaMn(CO₃)₂) may be significant in these inclusions. The presence of a Mn⁺⁺-bearing daughter phase would imply saturation with respect to that cation and thus, estimates of salinity based on the CaCl₂-NaCl-H₂O system may prove inadequate.

7.6 *INTERPRETATION*

Based on the cross-cutting relationships between the fluid inclusion types outlined in Section 7.3, it is concluded that at least five kinds of fluids were circulating in the veins, or at least five kinds of fluid were introduced at different times. The five palaeo-fluids are:-

1. **Type A**: An aqueous fluid of low to moderate salinity (1-20 wt% NaCl eq) containing Na⁺, Cl⁻ and possibly Ca²⁺, Mg²⁺, Li⁺, Al³⁺ and/or Fe³⁺, with small amounts of CO₂ and CH₄. The CO₂-CH₄ rich fluid was immiscible in the aqueous brine at the time of entrapment, and the aqueous brine was boiling.
2. **Type B**: An aqueous fluid of high salinity (35-50% total NaCl-CaCl₂) containing Na⁺, Ca²⁺, Cl⁻, and possibly Mg²⁺, Li⁺, Al³⁺ and/or Fe³⁺. Saturation with respect to calcite, rhodochrosite, kutnahorite, or aragonite suggests the fluid may also contain Mn⁺, HCO₃⁻, CO₃²⁻ or H₂CO₃. Laser Raman analyses indicate the presence of CO₂ in excess of CH₄, however, given the low solubility of CO₂ and CH₄ in saline aqueous fluids (Section 7.4.4), the CO₂ and CH₄ content is predicted to be negligible.

3. **Type C**: Vapour or aqueous-rich fluid of moderate to high salinity (28-34 wt% NaCl eq) containing Na⁺ and Cl⁻, and possibly Ca²⁺, Mg²⁺, K⁺, Li⁺, Al³⁺ and/or Fe³⁺. The vapour phase is dominated by CH₄ (as indicated by laser Raman analyses), although microthermometry has also identified the presence of CO₂.

4. **Type D**: CO₂-rich fluid of moderate salinity (>20 wt% NaCl eq) containing CH₄ and salts (unidentified).

5. **Type E**: An aqueous fluid of high salinity (15-34 wt% NaCl eq) containing Na⁺, Cl⁻, and possibly Ca²⁺, Mg²⁺, Li⁺, Al³⁺ and/or Fe³⁺, with small amounts of CH₄ and CO₂. The presence of unidentified acicular daughter minerals suggest the fluid contain other species.

The five palaeo-fluids may be summarised in terms of varying proportions of H₂O, CO₂, CH₄, Na⁺, Cl⁻, and Ca²⁺, with minor amounts of Mg²⁺, Mn⁺, K⁺, Li⁺, Al³⁺, Fe³⁺, HCO₃⁻, CO₃²⁻ or H₂CO₃ and this consistency in bulk fluid composition is suggestive of a common parent fluid and a common fluid source. With regard to variance, the fluids exhibit an overall shift in character from dominantly aqueous to vapour-rich to dominantly aqueous once again. The variance may be interpreted in terms of unmixing and/or entrapment of immiscible fluids.

As described in Section 7.4.4, experimental data for the H₂O-CO₂-NaCl system (Takenouchi & Kennedy, 1965; Bowers & Helgeson, 1983a; Franck, 1985, Gehrig *et al.*, 1986), and the H₂O-CH₄-NaCl and H₂O-CH₄-CaCl₂ systems (Krader & Franck, 1987) indicate that the addition of NaCl and/or CaCl₂ shifts the CO₂-H₂O or CH₄-H₂O solvus peak to higher temperatures and increases the size of the two phase liquid-vapour field. This is demonstrated in Figure 7.14 for the H₂O-CO₂-NaCl system: the upper boundary of the two-phase liquid-vapour field lies at approximately 400°C. Figure 7.14 indicates that at temperatures less than 400°C a parent H₂O-CO₂-NaCl fluid may (a) unmix to produce a saline aqueous brine containing small amounts of CO₂, and a CO₂-rich fluid containing small amounts of brine, (b) trap a homogeneous saline liquid containing small amounts of dissolved CO₂, or (c) trap a homogeneous CO₂-rich vapour containing small amounts of saline liquid. Bowers and Helgeson (1983b) have suggested that a high salinity brine may be generated in this fashion because the salts will fractionate into the aqueous phase.

With regard to the H₂O-CO₂-CaCl₂ and H₂O-CH₄-CaCl₂ systems, limited experimental data do not allow a full discussion of the phase relationships at low temperatures (< 400°C) and pressures (< 500 bars). However, for the H₂O-CH₄-NaCl system, isobaric diagrams of Krader and Franck (1987) and T-X diagrams of Naden and Shepherd (1989) depict a CH₄-H₂O-T plane that is dominated by a two-phase methane-rich vapour and H₂O-rich liquid field (Figure
7.20). The diagrams indicate that only small amounts of methane (≤ 5 mol %) are soluble in aqueous brine over a wide range of pressures and temperatures (up to 3 kbar and 400°C) and, at temperatures less than 400°C, CH₄ solubility is decreased with increasing salinity. Thus in similarity with the H₂O-CO₂-NaCl system, at T<400°C (P< 3 kbar) a parent H₂O-CH₄-NaCl fluid may trap homogeneous end-member fluids or unmix to produce and trap immiscible fluids. Further to this, because the solubility of CH₄ is lower than CO₂ for a given pressure and temperature (Naden & Shepherd, 1989), CH₄ immiscibility will be attained before CO₂-CH₄ immiscibility, and therefore unmixing can result in methane as the dominant volatile species.

Using this data it is possible to explain the sequence of five fluid inclusion types detailed above:-

**Type A** fluid inclusions represent mixed entrapment in the two-phase field of a low to moderately saline aqueous brine containing minor CO₂ and CH₄, and small CO₂-CH₄ vapour inclusions (as detailed in Section 7.3.3). The inclusions occur in qtz 2 within growth zones.

**Type B** fluid inclusions represent entrapment in the one-phase liquid field of a homogeneous saline aqueous brine containing minor CO₂ and CH₄. Hence, only one kind of fluid inclusion occurs along a healed micro-fracture, L:V ratios are relatively constant, and daughter minerals show constant volume proportions to other phases.

**Type C** fluid inclusions represent entrapment in the two-phase field of an CH₄-CO₂ rich, moderate to high salinity fluid (CH₄ dominant vapour-rich inclusions), and a moderate to high salinity aqueous brine containing minor CO₂ and CH₄ (liquid-rich inclusions). The inclusions occur along the same healed micro-fractures together with bismuth-gold grains.

**Type D** fluid inclusions represent entrapment of an CO₂-CH₄ rich, moderately saline fluid (CO₂ dominant vapour-rich inclusions), probably in the two-phase field because L:V ratios are not constant and range from 75 to >95% vapour. However, liquid-rich inclusions have not been identified along the same healed micro-fractures and this may indicate an exotic origin for this fluid, or entrapment in the one-phase vapour field.

**Type E** fluid inclusions represent entrapment in the one-phase liquid field of a homogeneous saline aqueous brine containing minor CO₂ and CH₄. During, or subsequent to entrapment, the inclusions suffered necking.

In analysis, types C and D fluid inclusions exhibit a variation in liquid or vapour components which may be explained by the mixed entrapment of two immiscible fluids, ie., entrapment in the two-phase field (Figure 7.14 & 7.20). In contrast, type B fluid inclusions do not exhibit a variation in liquid or vapour components, but constitute entrapment of a homogeneous hypersaline aqueous brine (containing minor CO₂-CH₄), or entrapment in the one-phase liquid field. Therein, the type B fluid may approximate a parent brine that unmixed to yield types C and D fluid inclusions. It is probable that unmixing of the fluid occurred during a decrease in temperature because there is a significant decrease in the (minimum) fluid temperature and salinity for these inclusions (Figure 4.21), and this is coincident with a waning in sulphide precipitation. Indeed, type E fluid inclusions may be explained as a
low(er) temperature derivative of the parent type B brine.

In contrast, type A fluid inclusions cannot be explained by unmixing of a type B parent brine. Rather, the inclusions appear to characterise a distinct fluid type, based on calculated salinities (Figure 4.21). Of relevance here is the fact that some type A fluid inclusions show evidence of boiling (Section 7.4.5). As stated previously (Section 7.2), type A fluid inclusions occur in growth zones with tourmaline in euhedral quartz terminations to crack-seal quartz fibres. Indeed, the formation of this quartz (qtz 2) is the first event in the open-space fill stage of quartz-sulphide vein development (Section 4.2.5.3, 5.2.5 & 6.4.2), and thus wise demarks a significant vein dilation event. In this context, unmixing to produce a CO₂-CH₄ rich vapour and a saline aqueous brine, and the boiling of that saline aqueous brine, may be attributed to the sudden reduction in pressure (and ?temperature) of a locally derived fluid during vein dilation, i.e., a locally derived fluid suffered decompressive effervescence and decompressive boiling prior to an influx of a type B parent brine.

In summary, a consistency in bulk fluid composition suggests that type B, C, D, and E fluid inclusions had a common parent fluid and a common parent source. Fluid variance between the fluid inclusion populations may be interpreted in terms of decreasing temperature and salinity, and (a) entrapment of a type B parent fluid (type B fluid inclusions), (b) unmixing of a type B parent fluid (types C & D fluid inclusions), or (c) entrapment and necking of a low temperature derivative of a type B parent fluid (type E fluid inclusions). The formation of type A fluid inclusions, quartz and tourmaline accompanied decompressive effervescence and boiling of a locally derived fluid during a sudden reduction in fluid pressure and as a result of micro-fracturing of the host veins.

7.7 ESTIMATE OF MINIMUM DEPTH AND PRESSURE

Where a set of inclusions in the same growth zone exhibit similar or identical Thtot into the liquid and vapour phase respectively, the Thtot is the temperature of entrapment (Roedder, 1981) and may be used to calculate pressure (Roedder, 1984). In this study, the minimum pressure of fluid entrapment has been estimated from the homogenisation temperature of type A fluid inclusions in the temperature range 357.8 to 390.1°C (n=5), i.e., type A fluid inclusions in sample BD086a (Table 7.1). Critical pressure values have been calculated from equation (4) of Knight and Bodnar (1989) for the NaCl-H₂O system because this system draws a close approximation to the fluid composition of type A fluid inclusions in sample BD086a,

\[ P = 2094 - 20.56T + 0.06896T^2 - 8.903 \times 10^{-3}T^3 + 4.214 \times 10^{-5}T^4, \]

where \( P \) is the critical pressure in bars (±21 bars) and \( T \) is the critical temperature in degrees Celsius. The minimum depth of boiling has been calculated using the relationship,

\[ P = H \times 1.0 \times 0.0981, \]

of Shepherd et al. (1985), where \( P \) is the critical pressure in bars, \( H \) is the depth in metres,
1.0 represents the specific gravity of water, and 0.0981 is a constant. Critical pressures range from 178.49 to 253.01 bars, which corresponds with a minimum depth range of boiling of approximately 1800 to 2600 metres, or 1.8 to 2.6 km.

7.8 FLUID HISTORY

The data presented above point to a progressive change in the fluid conditions during mineralisation, given the cross-cutting nature of the five fluid inclusions types as established in Section 7.3.

Based on petrographic, microthermometric and laser Raman analyses, the composition of the fluid was as follows:-

1. (Type A) An aqueous fluid of low to moderate salinity (1-20 wt% NaCl eq) in which the main species were Na+, Cl-, and possibly Ca²⁺, Mg²⁺, Li⁺, Al³⁺ and/or Fe³⁺, with small amounts of CO₂ and CH₄. The fluid was possibly derived locally during a significant vein dilation event. Precipitation of quartz (qtz 2) and tourmaline from this fluid accompanied decompressive boiling and effervescence at minimum pressures ranging from approximately 180 to 255 bars, at a minimum depth of approximately 1.8 to 2.6 km.

2. (Type B) An aqueous fluid of high salinity (35-60% total NaCl-CaCl₂) containing Na⁺, Ca²⁺, Cl⁻, and possibly Mg²⁺, Li⁺, Al³⁺ and/or Fe³⁺. Saturation with respect to calcite, rhodochrosite, kutnahorite, or aragonite suggests the fluid may also contain minor Mn⁺, HCO⁻₃, CO₃²⁻ or H₂CO₃. CO₂:CH₄ ratio is approximately 15:1, however, the overall CO₂-CH₄ content is negligible. The fluid is inferred to approximate the parent bring of types C, D and E fluid inclusions.

3. (Type C) Vapour or aqueous-rich fluid of moderate to high salinity (28-34 wt% NaCl eq) containing Na⁺ and Cl⁻, and possibly Ca²⁺, Mg²⁺, K⁺, Li⁺, Al³⁺ and/or Fe³⁺. The vapour phase is dominated by CH₄ (as indicated by laser Raman analyses), although microthermometry has also identified the presence of CO₂. CO₂ densities range from 0.8 to 0.85 g/cc. The fluid was accompanied by the precipitation of gold and bismuth with X₈₁ ~ 0.79 to 0.96, or 79 to 96% bismuth.

4. (Type D) A CO₂-rich fluid of moderate salinity (>20 wt% NaCl eq) containing CH₄ and salts (unidentified). The main dissolved species are not known. CO₂ densities range from 0.8 to 0.85 g/cc. CO₂:CH₄ ratios range from 6:1 to 8:1.

5. (Type E) An aqueous fluid of high salinity (15-34 wt% NaCl eq) containing Na⁺, Cl⁻, and possibly Ca²⁺, Mg²⁺, Li⁺, Al³⁺ and/or Fe³⁺, with small amounts of CH₄ and CO₂. The inclusions show evidence of necking subsequent to entrapment.

The nature and composition of types B and E fluid inclusions are very similar, however, Thet data indicate that the fluids were trapped at high (~ 380-325°C) and low temperatures (~270-120°C), respectively. Type C and D fluid inclusions are also similar being vapour rich (CH₄- CO₂ rich, then CO₂-CH₄ rich) suggesting there was an overall continuity in fluid composition.
during and after the precipitation of gold and bismuth. The change in the overall fluid composition from type A to type B fluid inclusions describes a period of increased salinity that was concomitant with the influx of a hypersaline brine and the first precipitation of sulphide minerals, while the change in the overall fluid composition from type B to perhaps type E fluid inclusions, describes a period of fluid unmixing, decreasing salinity and a waning in the precipitation of sulphide minerals. Based on Thot data, there is a significant decrease in the overall (minimum) temperature of fluid entrapment, i.e., the fluid cooled through time.

In summary, microthermometric and laser Raman analyses of the five fluid inclusion populations suggest that mineral precipitation occurred during a decrease in temperature from approximately 380° to 240°C from a metal enriched CO₂-CH₄-bearing, moderately saline to hypersaline brine (type B parent brine). Initial precipitation of quartz (qtz 2) and tourmaline from a moderate to low salinity aqueous brine was accompanied by decompressive effervescence and boiling at a minimum pressure of 180-255 bars, or a minimum depth of 1.8 to 2.6 km. The data support physico-chemical estimates determined in Section 6.6 which indicate that mineral precipitation resulted during cooling of the hydrothermal fluid.
Figure 7.1: Group 2 fluid inclusions. The inclusions constitute a single compositional type (type A) and occur in growth zones in euhedral quartz (qtz 2). DDH B086, 143.0 m. Scale bar is 100 µm long.

Figure 7.2: Type A fluid inclusions in a growth zone in qtz 2. The inclusions contain liquid (L) and vapour (V), and are accompanied in the same growth zone by greenish-coloured tourmaline (T). DDH BD068, 143.0 m. Scale bar is 30 µm long. (Same photograph as Figure 6.3 but upside down.)

Figure 7.3a: Photomicrograph of type B fluid inclusions in quartz. The inclusions contain liquid (L), vapour (V), carbonate daughter mineral (C), unidentified acicular daughter mineral (N), and halite (H). DDH BD066, 147.25 m. Scale bar is 10 µm long.

Figure 7.3b: Photomicrograph of type B fluid inclusions in quartz. The inclusions contain liquid (L), vapour (V), a carbonate daughter mineral (C), an unidentified acicular daughter mineral (N), and halite (H). Nahcolite is orientated NE-SW beneath the halite daughter mineral. DDH BD006, 147.25 m. Scale bar is 10 µm long.
Figure 7.4: Photomicrograph of a trail of type C fluid inclusions (arrowed) in a healed fracture with gold-bismuth blebs (black) in quartz. Tourmaline (T) occurs as accidentally trapped inclusions in the wall of some fluid inclusions. DDH BD080(2), 286.95 m. Scale bar is 30 µm long.

Figure 7.5: Composite photomicrograph of a trail of type C fluid inclusions (arrowed) in the same healed fracture with gold-bismuth blebs (black) in quartz. The fluid inclusions occasionally wet the gold-bismuth blebs suggesting they formed at the same time. DDH BD086, 143.0 m. Scale bar is 20 µm long.
Figure 7.6: Photomicrograph of type C fluid inclusions in quartz "wetting" opaque blebs of gold-bismuth. The inclusions contain liquid (L) and vapour (V), and occasionally connect gold-bismuth blebs though fluid-filled tubes (large open arrow). Tourmaline (T) occurs as accidentally trapped inclusions in the wall of some fluid inclusions. DDH BD086, 14.0 m. Scale bar is 30 µm long.

Figure 7.7 Photomicrograph of a trail of type D fluid inclusions in quartz. DDH BD001, 56.0 m. Scale bar is 40 µm long.

Figure 7.8: Photomicrograph of a cluster of type E fluid inclusions in quartz. H = halite, L = liquid, and V = vapour. DDH BD086, 345.0 m. Scale bar is 30 µm long.

Figure 7.9: Serial photomicrographs of type E fluid inclusion in quartz. The inclusion is 25 µm long and contains an unidentified acicular mineral (N), halite (H) and liquid (L). DDH BD086, 345.0 m.
A. At 20.1°C.
B. At -12.6°C strongly birefringent hydrohalite is clearly visible.
C. At 8.6°C hydrohalite is metastable. The hydrohalite melted at 16.2°C with the simultaneous growth of a halite crystal.
Figure 7.10: Flow diagram of the observed cross-cutting relationships of types A, B, C, D, and E fluid inclusions, and their relationship to qtz 2-tourmaline and gold-bismuth.
<table>
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<th>No.</th>
<th>Tour P or S</th>
<th>No. of phases</th>
<th>Primary inclusion</th>
<th>Secondary inclusion</th>
<th>Tmin (°C)</th>
<th>Tm (°C)</th>
<th>Tm CH4 (°C)</th>
<th>Tm CO2 (°C)</th>
<th>Te (°C)</th>
<th>TmH2O (°C)</th>
<th>TmCH4 (°C)</th>
<th>TmCO2 (°C)</th>
<th>Td (°C)</th>
<th>COMMENT</th>
<th>Wt% NaCl equiv</th>
<th>C.P.</th>
<th>Depth (metres)</th>
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</table>

VAPK = estimated vapour percent in the inclusion.

P = primary inclusion.
S = secondary inclusion.

X = tourmaline as a secondary inclusion that projects into the fluid inclusion cavity
XX = tourmaline in the growth zone together with the fluid inclusion.

Tmin = minimum temperature to which sample was cooled.
TmCH4 = temperature of first melting of CH4 (L+4 -> L+V+4).
TmCH4 = temperature of first melting of CH4 (L+4+V -> L+V).
TmCO2 = temperature of first melting of CO2 (L+4+V -> L+V).
TmCO2 = temperature of first melting of CO2 (L+4+V+4 -> L+V+4).

Te = temperature at homogenisation into the vapour phase.
Te = temperature at homogenisation into the liquid phase.
Td = temperature at decrepitation.

NaCl = weight percent NaCl equivalent calculated from Tm using a least squares regression equation of Potuz et al. (1978). Error is +/- 0.028 w/ NaCl.

C.P. = critical pressure in bars calculated from Tm of coexisting fluid inclusions that exhibit critical behaviour, using equation (4) of Knight & Bodnar (1989). Experimental uncertainty is +/- 21 bars.

Depth in metres is calculated from C.P. using data of Shepherd et al. (1985), and assuming hydrostatic pressure.

dbh = inclusion decrepitated before homogenisation.

* = denotes inclusions in the same growth zone.
"=" implies no data collected.

Table 7.1: Microthermometric data for type A fluid inclusions from the Batman deposit.
Figure 7.11: Type A fluid inclusions in quartz.
A. Histogram of data for the final melting temperature of CH₄ (TmCH₄), the final melting temperature of CO₂ solid (TmCO₂), the first melting temperature of NaCl-H₂O ice (Te), and the final melting temperature of NaCl-H₂O ice (Tm).
B. Histogram of data for the temperature of total homogenisation (Thtot).
### Table 7.2: Microthermometric data for type B fluid inclusions of the Batman deposit

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<th>INCLUSION No.</th>
<th>VAP No. of phases</th>
<th>Tour Carb</th>
<th>P, S</th>
<th>SIZE (µm)</th>
<th>TMin cooling</th>
<th>TICH4</th>
<th>TMCH4</th>
<th>TnCO2</th>
<th>TMCO2</th>
<th>Tm</th>
<th>Th</th>
<th>Thap</th>
<th>Tn halite</th>
<th>Thtot</th>
<th>Td</th>
<th>COMMENT</th>
<th>NaCl/CaCl2/H2O approximate</th>
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</tr>
</tbody>
</table>

VAP% = estimated percent of vapour in the inclusion.
Tour = tourmaline as an accidentally trapped solid in the wall of the fluid inclusion. X implies the mineral is present. Carb = carbonate daughter mineral. X implies the mineral is present. P = primary inclusion. S = secondary inclusion. TMin cooling = lowest temperature to which the sample was cooled. TICH4 = temperature of first melting of CH4 (L+V => L+V+L). TMCH4 = temperature of final melting of CH4 (L+S+V => L+V). TnCO2 = temperature of final melting of CO2 (L+V => L+V). Tn halite = temperature of the dissolution of halite daughter mineral. Th tot = temperature of the homogenisation of the total. Td = temperature of decrepitation. NS = not seen due to poor visibility. dbh = inclusion decrepitated before total homogenisation. *n* implies no data collected.
Figure 7.12: Type B fluid inclusions in quartz.
A. Histogram of data for the final melting temperature of CH$_4$ (TmCH$_4$), the final melting temperature of CO$_2$ solid (TmCO$_2$), the first melting temperature of NaCl-CaCl$_2$-H$_2$O ice (Te), and the final melting temperature of NaCl-CaCl$_2$-H$_2$O ice (Tm).
B. Histogram of data for the temperature of final melting of hydrohalite (Tm hydro).
C. Histogram of data for the temperature of homogenisation of vapour (Thvap), and the dissolution of halite (Th halite) and an unidentified daughter mineral (Thdt unk).
Figure 7.13: Ternary diagram contoured with respect to temperature in degrees Celsius in the NaCl-CaCl$_2$-H$_2$O system (modified after Vanko et al., 1988). The shaded area describes the bulk composition for type B fluid inclusions.
Figure 7.14: T-X diagram for the binary system H₂O-CO₂ at 200 bars pressure; modified after Bowers and Helgeson (1983). With increasing salinity the solvus broadens and shifts to higher temperatures, and the solubility of CO₂ is decreased.
Figure 7.15: P-T diagram for the ternary system H$_2$O-CH$_4$-NaCl, modified after Krader and Franck (1987). With increasing salinity, the isopleth for 17% CH$_4$ is shifted to higher temperatures. Conversely, it may be argued that for a given temperature and pressure, the solubility of CH$_4$ in the fluid decreases as salinity increases.
<table>
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<th>INCLUSION</th>
<th>VAP</th>
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<th>Solid</th>
<th>Tour</th>
<th>P, S</th>
<th>SIZE (µm)</th>
<th>Tmin cooling</th>
<th>TmCH4</th>
<th>TmCH4</th>
<th>TmCO2</th>
<th>TmCO2</th>
<th>Te</th>
<th>TmCO2</th>
<th>TmCO2</th>
<th>Telb</th>
<th>TsoBi</th>
<th>Td</th>
<th>COMMENT</th>
<th>Wt% NaCl equiv</th>
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<td>-</td>
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<td>Bi</td>
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<td>-250</td>
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<td>NR</td>
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<td>2</td>
<td>Bi-Au</td>
<td>-</td>
<td>S</td>
<td>15x4</td>
<td>-172.4</td>
<td>NS</td>
<td>-91.8</td>
<td>-80.1</td>
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<td>-</td>
<td>19.9</td>
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</table>

VAP = estimated percent of vapour in the inclusion.
Solid = solid phase wet by the fluid inclusion.
Tour = tourmaline as an accidental inclusion in the wall of the fluid inclusion.
P = primary inclusion.
S = secondary inclusion.
Tmin cooling = lowest temperature to which the sample was cooled.
TmCH4 = temperature of first melting of CH4 (L+S ⇄ L+V+S).
TmCO2 = temperature of first melting of CO2 (L+S ⇄ L+V+S).
TmCO2 = temperature of final melting of CO2 (L+V ⇄ L+V).
Te = temperature of the first melting of ice.
Tm = temperature of the final melting of ice.
TmCO2 clath = temperature of the final melting of CO2 clathrate.
ThCO2 = temperature of the homogenisation of CO2.
Telb = temperature of melt of bismuth.
TsoBi = temperature of the solidification of the bismuth.
Td = temperature of decrepitation.
wt% NaCl equiv = weight percent NaCl calculated from TmCO2 clathrate using a least squares regression curve of Bozzo et al. (1975).
NS = not seen due to poor visibility.
NR = no data recorded.
* = implies no data collected.

Table 7.3: Microthermometric data for type C fluid inclusions of the Batman deposit.
Figure 7.16: Type C fluid inclusions in quartz.
A. Histogram of data for the final melting temperature of CH₄ (TmCH₄), the final melting temperature of CO₂ solid (TmCO₂), the first melting temperature of NaCl-H₂O ice (Te), and the final melting temperature of NaCl-H₂O ice (Tm).
B. Histogram of data for the melting temperature of CO₂ clathrate (TmCO₂ clath) and temperature of homogenisation of CO₂ (ThCO₂).
C. Histogram of data for the melting temperature of bismuth (TeBi) and solidification temperature of bismuth (TsolBi).
Figure 7.17: Phase diagram for the gold-bismuth system, modified after Hultgren et al. (1963). L = liquid; G = gold; M = maldonite.
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<tr>
<th>INCLUSION No.</th>
<th>VAP %</th>
<th>No. of phases</th>
<th>TOUR</th>
<th>P, S</th>
<th>SIZE (μm)</th>
<th>Tmax cooling</th>
<th>T vap collapse</th>
<th>TIC02</th>
<th>TmC02</th>
<th>TIC02 clath</th>
<th>Th tot</th>
<th>Composition</th>
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<td>-</td>
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<td>-</td>
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<td>NS</td>
<td>-</td>
<td>V(C02)</td>
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</table>

VAP % = estimated percent of vapour in the inclusion.
Tour = tourmaline as an accidental inclusion in the wall of the fluid inclusion. X implies the mineral is present.
P = primary inclusion.
S = secondary inclusion.
Tmax cooling = temperature to which the sample was cooled.
T vap collapse = temperature at the vapour bubble collapsed.
TIC02 = temperature of first melting of CH4 (L+V = L+V+S).
TmC02 = temperature of first melting of CH4 (L+S+V = L+V).
TIC02 = temperature of first melting of CO2 (L+V = L+V+S).
TmC02 = temperature of final melting of CO2 (L+V = L+V).
TmC02 clath = temperature of final melting of CO2 clathrate
Th tot = temperature of the homogenisation of the total into the liquid.

(V) = homogenisation into the vapour.

wt% NaCl equiv = weight percent NaCl equivalent calculated from TmC02 clath using a least squares regression curve of Bozzo et al (1975)
NR = data was not recorded.
NS = not seen due to poor visibility.
# = implies no data collected.

Table 7.4: Microthermometric data for type D fluid inclusions of the Batman deposit
Figure 7.18: Type D fluid inclusions in quartz.
A. Histogram of data for the final melting temperature of CH$_4$ (TmCH$_4$) and the final melting temperature of CO$_2$ solid (TmCO$_2$).
B. Histogram of data for the melting temperature of CO$_2$ clathrate (TmCO$_2$ clath) and temperature of homogenisation of CO$_2$ (ThCO$_2$).
C. Histogram of data for the temperature of total homogenisation (Thtot).
Table 7.5: Microthermometric data for type E fluid inclusions of the Batman deposit
Figure 7.19: Type E fluid inclusions in quartz.
A. Histogram of data for the final melting temperature of CH₄ (TmCH₄), the final melting temperature of CO₂ solid (TmCO₂), the first melting temperature of NaCl-CaCl₂-H₂O ice (Te), and the final melting temperature of NaCl-CaCl₂-H₂O ice (Tm).
B. Histogram of data for the temperature of final melting of hydrohalite (Tm hydro).
C. Histogram of data for the temperature of homogenisation of vapour (Thvap), and the dissolution of halite (Th halite) and an unidentified acicular daughter mineral (Th dt unk).
### A. Type B fluid inclusions - gas analyses

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<th>TYPE OF ANALYSIS</th>
<th>SECTION NO.</th>
<th>HOST MINERAL</th>
<th>Analysed Composition</th>
<th>Position of major peak</th>
<th>Phase not detected</th>
<th>CO2/CH4</th>
<th>COMMENT</th>
</tr>
</thead>
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<td>V(CO2 + CH4) + L(H2O) + carb</td>
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<td>N2</td>
<td>94:6</td>
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</tr>
<tr>
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<td>CH4</td>
<td>BD086b</td>
<td>quartz</td>
<td>V(CO2 + CH4) + L(H2O) + carb</td>
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<td>N2</td>
<td>94:6</td>
<td>Paired analysis with KAH.36</td>
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</table>

L = liquid  
V = vapour

Sop = inclusion is wetting a gold-bismuth grain

### B. Type C fluid inclusions - gas analyses

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<th>TYPE OF ANALYSIS</th>
<th>SECTION NO.</th>
<th>HOST MINERAL</th>
<th>Analysed Composition</th>
<th>Position of major peak</th>
<th>Phase not detected</th>
<th>CO2/CH4</th>
<th>Comment</th>
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<td>quartz</td>
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<td>quartz</td>
<td>V(CH4) + L(H2O) + Sop</td>
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<td>CH4</td>
<td>BD086b</td>
<td>quartz</td>
<td>V(CH4) + L(H2O) + Sop</td>
<td>2915.9</td>
<td>CO2</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

L = liquid  
V = vapour

Sop = inclusion is wetting a gold-bismuth grain

### C. Type D fluid inclusions - gas analyses

<table>
<thead>
<tr>
<th>ANALYSIS NOS.</th>
<th>TYPE OF ANALYSIS</th>
<th>SECTION NO.</th>
<th>HOST MINERAL</th>
<th>Analysed Composition</th>
<th>Position of major peak</th>
<th>Phase not detected</th>
<th>CO2/CH4</th>
<th>Comment</th>
</tr>
</thead>
<tbody>
<tr>
<td>01.KAH</td>
<td>CO2</td>
<td>BD080q</td>
<td>quartz</td>
<td>V(CO2 + CH4) + L(H2O)</td>
<td>1384.9</td>
<td>85:15</td>
<td>N2</td>
<td>Paired analysis with KAH.02</td>
</tr>
<tr>
<td>02.KAH</td>
<td>CH4</td>
<td>BD080q</td>
<td>quartz</td>
<td>V(CO2 + CH4) + L(H2O)</td>
<td>2913.4</td>
<td>85:15</td>
<td>N2</td>
<td>Paired analysis with KAH.01</td>
</tr>
<tr>
<td>09.KAH</td>
<td>CO2</td>
<td>BD080q</td>
<td>quartz</td>
<td>V(CO2 + CH4) + L(H2O)</td>
<td>1384.1</td>
<td>88:12</td>
<td>-</td>
<td>Paired analysis with KAH.09</td>
</tr>
<tr>
<td>10.KAH</td>
<td>CH4</td>
<td>BD080q</td>
<td>quartz</td>
<td>V(CO2 + CH4) + L(H2O)</td>
<td>2912.2</td>
<td>88:12</td>
<td>-</td>
<td>Paired analysis with KAH.01</td>
</tr>
<tr>
<td>30.KAH</td>
<td>CH4</td>
<td>BD084d</td>
<td>quartz</td>
<td>V(CO2 + CH4) + L(H2O)</td>
<td>2912.8</td>
<td>89:11</td>
<td>-</td>
<td>Paired analysis with KAH.03</td>
</tr>
<tr>
<td>31.KAH</td>
<td>CO2</td>
<td>BD084d</td>
<td>quartz</td>
<td>V(CO2 + CH4) + L(H2O)</td>
<td>1387.3</td>
<td>89:11</td>
<td>-</td>
<td>Paired analysis with KAH.03</td>
</tr>
<tr>
<td>50.KAH</td>
<td>CH4</td>
<td>RX280x</td>
<td>quartz</td>
<td>V(CO2 + CH4) + L(H2O)</td>
<td>2912.8</td>
<td>89:11</td>
<td>-</td>
<td>Paired analysis with KAH.03</td>
</tr>
<tr>
<td>51.KAH</td>
<td>CO2</td>
<td>BD080q</td>
<td>quartz</td>
<td>V(CO2 + CH4) + L(H2O)</td>
<td>1386.5</td>
<td>89:11</td>
<td>-</td>
<td>Paired analysis with KAH.03</td>
</tr>
</tbody>
</table>

L = liquid  
V = vapour

### D. Type C fluid inclusions - salinity

<table>
<thead>
<tr>
<th>ANALYSIS NOS.</th>
<th>TYPE OF ANALYSIS</th>
<th>SECTION NO.</th>
<th>HOST MINERAL</th>
<th>wt% NaCl eq</th>
</tr>
</thead>
<tbody>
<tr>
<td>39.KAH</td>
<td>salinity</td>
<td>BD080q</td>
<td>quartz</td>
<td>27.7</td>
</tr>
<tr>
<td>40.KAH</td>
<td>salinity</td>
<td>BD080q</td>
<td>quartz</td>
<td>29.5</td>
</tr>
<tr>
<td>41.KAH</td>
<td>salinity</td>
<td>BD080q</td>
<td>quartz</td>
<td>28.8</td>
</tr>
<tr>
<td>43.KAH</td>
<td>salinity</td>
<td>BD080q</td>
<td>quartz</td>
<td>33.9</td>
</tr>
<tr>
<td>44.KAH</td>
<td>salinity</td>
<td>BD080q</td>
<td>quartz</td>
<td>26.5</td>
</tr>
<tr>
<td>45.KAH</td>
<td>salinity</td>
<td>BD086b</td>
<td>quartz</td>
<td>30.4</td>
</tr>
<tr>
<td>46.KAH</td>
<td>salinity</td>
<td>BD086b</td>
<td>quartz</td>
<td>29.9</td>
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### E. Type B fluid inclusions - carbonate daughter mineral

<table>
<thead>
<tr>
<th>ANALYSIS NOS.</th>
<th>TYPE OF ANALYSIS</th>
<th>SECTION NO.</th>
<th>HOST MINERAL</th>
<th>Composition</th>
<th>Position of major peak</th>
<th>CMWMNP</th>
<th>Phase not detected</th>
</tr>
</thead>
<tbody>
<tr>
<td>19.KAH</td>
<td>carbonate daughter</td>
<td>BD080q</td>
<td>quartz</td>
<td>L + V + carbonate</td>
<td>1083.9</td>
<td>A, K</td>
<td>graphite</td>
</tr>
<tr>
<td>26.KAH</td>
<td>carbonate daughter</td>
<td>BD084d</td>
<td>quartz</td>
<td>L + V + carbonate</td>
<td>1085.5</td>
<td>A, K, C, R</td>
<td>CO2, CO3, graphite</td>
</tr>
<tr>
<td>28.KAH</td>
<td>carbonate daughter</td>
<td>BD084d</td>
<td>quartz</td>
<td>L + V + carbonate</td>
<td>1085.5</td>
<td>A, K, C, R</td>
<td>HCO3, CO2, N2, H2S, graphite</td>
</tr>
<tr>
<td>34.KAH</td>
<td>carbonate daughter</td>
<td>BD086b</td>
<td>quartz</td>
<td>L + V + carbonate</td>
<td>1086.4</td>
<td>K, C, R, A</td>
<td>graphite</td>
</tr>
</tbody>
</table>

CMWMNP = carbonate minerals which match the major peak.
- A = Aragonite, peak at 1085 (Herman et al., 1987)
- K = Kutnahorite, peak at 1086 (Herman et al., 1987)
- C = Calcite, peak at 1087 (Herman et al., 1987)
- R = Rhodochrosite, peak at 1087 (Herman et al., 1987)

Table 7.6: Laser Raman analyses.
Figure 7.20: A. Isobaric diagram of P-T-X$_1$-X$_2$ phase space at 500 bars in the ternary system H$_2$O-CH$_4$-NaCl, modified after Krader and Franck (1987). The prism depicts a CH$_4$-H$_2$O-T plane that is dominated by a two phase methane-rich vapour and H$_2$O-rich liquid field. A temperature section through the prism (B.) demonstrates that this field broadens toward the NaCl end-member which indicates that CH$_4$ solubility is reduced with increasing salinity. C. Pseudo-binary temperature-composition section for the system CH$_4$-H$_2$O-NaCl (8 wt%) at 0.5 kbars (solid line), 1.5 kbars (dashed line), and 3.0 kbars (dotted line), after Naden & Shepherd (1989). The figure indicates that only small amounts of CH$_4$ are soluble in aqueous brine.
A

500 bars

Temperature

CH₄

H₂O

NaCl

Section of plane

B

(NaCl + CH₄)-rich fluid (unknown)

CH₄

Vapour

H₂O-rich liquid + CH₄-rich vapour

NaCl + NaCl-rich liquid + CH₄-rich vapour

Liquid

H₂O

C

Temperature (°C)

550

450

350

250

150

50

0 10 20 30 40 50 60 70 80 90 100

Mol% gas
CHAPTER 8

THE SOURCE & GEOCHEMISTRY OF ORE FLUID

8.1 SOURCE OF FLUIDS

8.1.1 Geological & fluid inclusion constraints

Any discussion involving the likely source of fluid(s) to the quartz-sulphide vein/lode systems in the Mt Todd goldfield must be constrained by the geologic data outlined in the preceding chapters. In summary, the data indicate that the vein/lode systems formed after peak thermal metamorphism associated with the shallow emplacement (approximately 3.5 to 5.0 km) of the Tennysons Leucogranite, and early in $D_2$, prior to the development of the regional $S_2$ fabric. The veins and lodes exhibit a spatial relationship to the leucogranite and lie at the contact margin between the inner and middle aureole. They occur within a northeast trending corridor and form a right stepping en échelon array which is indicative of sinistral strike-slip movement along the corridor. The veins and lodes are preferentially located in faults, joints and bedding planes that strike north-northeast, indicating they are structurally controlled. This suggests that fluid focus was also structurally controlled.

Sulphide, silicate and carbonate minerals were precipitated during a decrease in fluid temperature from approximately 380°C to below 240°C, from a metal enriched moderately saline to hypersaline brine (up to 50% NaCl-CaCl$_2$) containing minor CO$_2$-CH$_4$ (Section 7.6), during several phases of micro-fracturing of the wall rock. Precipitation of quartz (qtz 2) and tourmaline from a low to moderately saline brine (1-20 wt% NaCl eq) accompanied decompressive effervescence and boiling (Section 7.4.3 & 7.6) at a minimum pressure of 180-255 bars, or a minimum depth of 1.8 to 2.6 kilometres (Section 7.7).

From this geological data, it may be concluded that the quartz-sulphide veins formed during retrograde metamorphism associated with cooling of the Tennysons Leucogranite, and coeval with early prograde metamorphism associated with regional deformation during $D_2$.

Under these opposing metamorphic conditions, the volatiles released from the crystallising leucogranite, such as H$_2$O, H$_2$S, CO$_2$, HCl and HF (Burnham, 1979), contrast with the small to large amounts of H$_2$O consumed during rehydration of the country rocks and cooling leucogranite (Crawford et al., 1979a; Burnham, 1979, Giggenbach, 1992; Miyashiro, 1994), and the appreciable amounts of H$_2$O and CO$_2$ that may be liberated during dehydration and decarbonation of the deforming sediment-granite pile (Miyashiro, 1994). Thus, the advective and convective exchange or mixing of considerable volumes of potentially
occurred (Knapp & Norton, 1981; Ferry, 1991) with the possibility that large volumes of externally derived fluid may have infiltrated a site of potential ore formation (Miyashiro, 1994). Possible fluid source(s) to the vein or lode systems are therefore constrained by the degree to which one or several contrasting fluid types may have mixed and exerted an influence. Given this, the geologic data indicate that the source of fluids may have been:

(a) Metamorphic, being derived from devolatilisation early in D2,

(b) Magmatic, being derived from the crystallisation of the Tennysons Leucogranite,

(c) Mixed magmatic-metamorphic.

Alternatively, connate fluids (water trapped during sedimentation) or fluids derived from deep crustal infiltration of meteoric or seawater may also constitute a source because at depths shallower than 6 to 10 km, infiltration of surface-derived fluid, or mobilisation of connate fluids may be significant particularly if the crust is structurally weakened through faulting and fracturing (Sibson, 1989; Bennett & Barker, 1992). Further to this, connate or surface-derived fluids may mix with magmatic, metamorphic or mixed magmatic-metamorphic fluids thus providing a complex fluid source history. For example, a rising fluid from a mixed magmatic-metamorphic source may impinge on a shallow meteoric system (Furlong et al., 1991).

However, in assessment of these sources of ore-forming fluid, it is unlikely that connate fluid will be a recognisable contributor to the overall fluid budget in the hydrothermal systems of the Mt Todd goldfield: it will have been substantially modified and recycled during prograde and retrograde metamorphic reactions associated with D1 and the emplacement of the Yenberrie and Tennysons Leucogranites (Chapter 4). That is, it will be metamorphic.

Similarly, the moderate to high salinity (up to 50% NaCl-CaCl2) of the aqueous hydrothermal fluid associated with the precipitation of sulphides and gold, as characterised by types B, C, D, and E fluid inclusions (Section 7.6 & 7.8), make it difficult to recognise a meteoric or seawater source for this ore forming fluid. In fact, a Na:Ca ratio of 2.9 to 3.7 (Section 7.4.4) that is well below the seawater ratio of 26.0 (Vanko et al., 1988), and a Na:Ca cation ratio of 3:1 (calculated from Table 7.2) where Na > Ca >> K, would indicate that these source fluids, if involved, were substantially modified by fluid-wall rock reactions during crustal infiltration (Crawford, 1981; Vanko et al., 1988; Khin Zaw et al., 1994). In comparison, the low to moderate salinity (1-20 wt% NaCl eq) of locally-derived aqueous hydrothermal fluid, as characterised by type A fluid inclusions (Section 7.6 & 7.8), is consistent with a meteoric water or seawater source, particularly because type A fluid inclusions were trapped at the relatively shallow depth of 1.8-2.6 km (minimum).

Given this, it may be concluded that the source for moderate to high salinity aqueous hydrothermal fluid was either magmatic, metamorphic, or mixed magmatic-metamorphic in origin, or a substantially modified meteoric water or seawater. As indicated above, connate fluids are
unlikely to have survived modification during regional and contact metamorphism and will not be a (recognisable) contributor to the overall fluid budget. Meteoric water or seawater are a probable source for the locally-derived, low salinity aqueous fluids that precipitated quartz (qtz 2) and tourmaline.

8.1.2 Oxygen isotope data
To model the possible origin(s) of the hydrothermal fluids responsible for the formation of the quartz-sulphide vein and lode systems in the Mt Todd goldfield, oxygen isotope ratios were measured on 7 samples of crack-seal quartz from quartz-sulphide veins in the Batman and Quigleys deposits. $\delta^{18}O$ values were measured at the oxygen isotope facility in the Department of Geology at the University of Tasmania, using the method of Clayton and Mayeda (1963), and the results are presented in Table 8.1 and in Figure 8.1. Two samples of quartz from quartz-tourmaline veins on Mount Todd were also measured because the veins are believed to constitute a subset in the quartz-sulphide vein system (Section 6.4.2). $\delta^{18}O$ values (relative to SMOW) for the 7 samples from quartz-sulphide veins range from 10.2 to 15.2‰, while those for the 2 samples from quartz-tourmaline veins are 10.0‰ and 15.1‰.

From the $\delta^{18}O$ quartz values, $\delta^{18}O_{\text{H}_2\text{O}}$ has been calculated using the relationship,

$$1000 \ln \alpha_{\text{QW}} = 3.34(10^6T^{-2}) - 3.31 \quad \text{(between 250 and 500°C)}$$

of Matsuhisa et al. (1979), where $1000 \ln \alpha_{\text{QW}} = \delta_{\text{quartz}} - \delta_{\text{H}_2\text{O}}$, and $T$ is temperature in degrees Kelvin. For a temperature of 357°C (lowest $T_{\text{tot}}$ for type A fluid inclusions in sample BD086a), the $\delta^{18}O_{\text{H}_2\text{O}}$ values range from 5.1‰ to 10.1‰ for quartz-sulphide veins, and 4.9‰ and 10.1‰ for quartz-tourmaline veins relative to SMOW. $\delta^{18}O_{\text{H}_2\text{O}}$ values for the Batman and Quigleys deposits range from 8.2 to 10.1‰, and 5.1 to 9.2, respectively. In comparison, for a temperature of 390°C (highest $T_{\text{tot}}$ for type A fluid inclusions in sample BD086a), the $\delta^{18}O_{\text{H}_2\text{O}}$ values range from 5.9 to 10.9‰ for quartz-sulphide veins, and 5.7‰ and 10.8‰ for quartz-tourmaline veins relative to SMOW. $\delta^{18}O_{\text{H}_2\text{O}}$ values for the Batman and Quigleys deposits range from 9.0 to 10.9‰, and 5.9 to 10.0‰, respectively. The $\delta^{18}O_{\text{H}_2\text{O}}$ values for quartz-sulphide veins from the Batman deposit overlap with those of the Quigleys deposit, while $\delta^{18}O_{\text{H}_2\text{O}}$ values of quartz-tourmaline veins from Mount Todd overlap with $\delta^{18}O_{\text{H}_2\text{O}}$ values for quartz-sulphide veins from the Batman and Quigleys deposits. The similarity in isotopic signature between quartz-sulphide veins from the two deposits, and between quartz-sulphide and quartz-tourmaline veins supports structural data that suggests (a) the deposits formed at the same time by a similar process (Section 5.4) and probably from a common parent fluid (Section 7.6), and (b) quartz-tourmaline veins form a subset in the quartz-sulphide vein system.

With regard to a fluid source, the calculated $\delta^{18}O_{\text{H}_2\text{O}}$ values are consistent with those for magmatic water or metamorphic water, for which the typical $\delta^{18}O_{\text{H}_2\text{O}}$ values range between 6 to 12‰ and 3 to 20‰, respectively (Ohmoto, 1986). They overlap with $\delta^{18}O_{\text{H}_2\text{O}}$ values of 5.5
to 10.3% reported by Wygralak and Ahmad (1990) for fluid inclusion water from several mesothermal gold deposits in the Pine Creek Inlier, for which a mixed magmatic and metamorphic source for fluids has been postulated (Figure 8.2). However, the $\delta^{18}O_{H2O}$ values are also consistent with:-

(a) Isotopic exchange between the sediments of the Burrell Creek Formation, for which $\delta^{18}O$ values within a range of 10 to 40% would be expected based on overall $\delta^{18}O$ values for sedimentary rocks (Hoefs, 1973), and seawater, which has fluctuated in its $O^{18}$ value in the last 3 billion years from -5 to 5% (Ohmoto, 1986, quoting Karhu & Epstein, 1986).

(b) Isotopic exchange between sediments of the Burrell Creek Formation and meteoric water, for which $\delta^{18}O$ varies in accordance with the equation,

$$\delta D = \delta^{18}O + 10$$

(Craig, 1961)

(c) Values for juvenile waters which closely approximate 6 ± 1% (Ohmoto, 1986). The range of calculated values are greater than would be expected from a fluid of mantle origin, but this may result from mixing with an isotopically heavier fluid.

The oxygen isotope data therefore suggest that the fluid source was either magmatic, metamorphic, seawater, meteoric water, juvenile or mixed in origin, and hence do not help to characterise a common parent fluid source.

In part this ambiguity may be the result of a difficulty in collecting a pure quartz sample. As stated previously, each sample was collected from crack-seal quartz and ultimately the calculated $\delta^{18}O_{H2O}$ values characterise the source fluid to this quartz. However, each sample also contained a lesser fraction of quartz 2 and/or quartz 3 (<5%) that could not be separated during sample preparation. Therein, each measured $\delta^{18}O$ value and each calculated $\delta^{18}O_{H2O}$ value must constitute a sum. In corollary, if quartz 2 was precipitated from a locally-derived meteoric water or seawater source (a lighter isotope source), as is suggested from fluid inclusion and mineragraphic data, then the calculated $\delta^{18}O_{H2O}$ values may be light. Similarly, if quartz 3 was precipitated from a high salinity fluid of magmatic, metamorphic, seawater, meteoric water or mixed water source, then in the case of a meteoric water or seawater source (a lighter isotope source), the calculated $\delta^{18}O_{H2O}$ values may be light, while in the case of a metamorphic source (a heavier isotope source), the calculated $\delta^{18}O_{H2O}$ values may be heavy. As a result, although the oxygen isotope data support a common parent fluid, it has not facilitated the identification of the nature of the fluid source. As a result, an estimate of the source of fluids that is based on the oxygen isotope data alone may prove inadequate.

### 8.1.3 Discussion

As stated previously, the quartz-sulphide veins and lodes of the Mt Todd goldfield comprise a two-stage fill: early crack-seal fibre quartz and late silicate-sulphide-carbonate fill (Chapter 6). Oxygen isotope data has indicated that the possible sources for fluid which precipitated the crack-seal quartz include magmatic, metamorphic, seawater, meteoric water, juvenile or...
mixed water. The data supports a common parent fluid for crack-seal quartz in quartz-sulphide veins at the Batman and Quigleys deposit, and in quartz-tourmaline veins at Mount Todd. In comparison, geological data has indicated that the possible sources for moderate to highly saline aqueous fluids that precipitated silicate, sulphide and carbonate fill include magmatic, metamorphic, mixed magmatic-metamorphic water, or highly modified meteoric water or seawater. The hypersaline nature of this brine (Chapter 7) and the strong structural control to the location of veins and lodes (Chapters 4 & 5), is consistent with rehydration of early higher grade contact metamorphic assemblages at low fluid to rock ratios (Crawford et al., 1979a; Wall, 1989; Bennett & Barker, 1992).

As stated previously, a discussion of the likely sources of fluid to the hydrothermal system should ultimately find constraint in the available geologic data, and this suggests that the gold-bearing veins and lodes formed during retrograde metamorphism associated with cooling of the Tennysons Leucogranite, and coeval with early prograde metamorphism in $D_2$.

The sources of fluid are therefore constrained by the degree of interaction or mixing of (a) magmatic fluids that were released from the cooling leucogranite, (b) metamorphic fluids released during prograde metamorphism early in $D_2$ and (c) surface-derived fluid, and to the degree to which the source fluids are consumed during rehydration reactions. Clearly, during the precipitation of quartz (qtz 2) and tourmaline, low to moderately saline (1-20 wt% NaCl eq), surface-derived fluids (meteoric or seawater) dominated the hydrothermal system, albeit, for a short period of time.

In discussion, convective circulation is a likely consequence of (a) the intrusion of magma into the crust, (b) the liberation of fluids in the contact aureole during dehydration and decarbonation in the country rock, and (c) thermal buoyancy of heated crustal fluids (Knapp & Norton, 1981; Furlong et al., 1991). Convective circulation is greatest after peak thermal conditions have been reached within the inner aureole and decrease in time as the pluton cools (Furlong et al., 1991). The magnitude of convection is directly related to the rock permeability (Knapp & Norton, 1981). In turn, rock permeability is significantly enhanced by fractures formed as a result of an increase in effective tensile stress ($\alpha$-Pr) during cooling of the magma (Burnham, 1979; Pitcher, 1979; Ramsay, 1981; Giggenbach, 1992; Ferry, 1991). Overall, the progressive fracturing results in a decrease in volume. A concomitant increase in porosity via solution-reactions enables the leaching of volatiles dissolved in the rock matrix, or stored in vesicles, and a highly saline brine and/or a gas-rich fluid may be formed (Giggenbach, 1992). However, porosity is also reduced sympathetically through crack-sealing and pressure solution, mineral deposition, and ductile strain (Furlong et al., 1991). The latter may initiate ductile deformation in the country rock and/or cooling granite carapace. Rehydration reactions will further decrease porosity and enhance channelised fluid flow. If the rehydration rate is rapid, and fluid-rock ratios are low, then fluid salinity will increase as $H_2O$ is consumed (Crawford et al., 1979a; Bennett & Barker, 1992). It is
therefore possible to generate a hypersaline fluid of significant metal carrying capacity.

In corollary, the source(s) of fluid to the gold-bearing veins and lodes in the Mt Todd goldfield may therefore be explained by a convective circulation model involving cooling of the Tennysons Leucogranite. In this model, the morphology of the veins and lodes are interpreted as follows:-

(a) Crack-seal fibre quartz formed by progressive fracturing, sealing and pressure solution, from a silica-saturated brine that was probably derived from a mixed magmatic-(contact) metamorphic fluid source during early retrograde metamorphism.

(b) Euhedral quartz (qtz 2) and tourmaline formed from a surface-derived fluid (meteoric or seawater) during a significant vein dilation event, perhaps marking the first stage of channelised fluid flow. Because tourmaline is a typical metasomatic mineral (Heinrich, 1965; Battey, 1981), it is possible that this fluid mixed with a magmatic-metamorphic fluid. A mixed fluid would also explain the variable CO₂:CH₄ content and salinity if type A fluid inclusion which formed coeval with precipitation of tourmaline.

(c) The precipitation of a silicate-sulphide-carbonate-(gold) assemblage is associated with the influx of a moderately saline to hypersaline brine that was derived from a magmatic-(contact) metamorphic fluid source, during retrograde metamorphism.

Further to this, deformation early in D₂ may have been initiated during cooling of the leucogranite as a result of high contractual strain (Furlong et al., 1991). However, it is not expected that fluids derived from devolatilisation would be significant during D₂, and are likely to have been quickly consumed in rehydration reactions within the thermal aureole of the leucogranite.

With regard to an alternative source(s) of fluid, as stated previously, a connate fluid source is not indicated because it is unlikely to have survived modification during regional metamorphism in D₁ and contact metamorphism during emplacement of the Yenberrie and Tennysons Leucogranites. Similarly, a (direct) juvenile fluid source is not indicated because mantle-derived mafic melts were not emplaced at the time the gold-bearing veins and lodes were formed, ie., the Zamu and Oenpelli Dolerites were emplaced before and after the Cullen Batholith, respectively, and the emplacement of the Batholith is broadly synchronous with mineralisation in the Mt Todd goldfield (Section 2.1.4 & 4.3.4). Alternatively, it may be argued that an (indirect) juvenile fluid source is implied by the emplacement of the Cullen Batholith because the Batholith is believed to have formed by remelting of fractionated mantle-derived material (Wyborn, 1988). In this case, a magmatic fluid source would also imply a juvenile fluid source.

It is therefore concluded that the sources of fluid to the gold-bearing veins and lodes in the Mt Todd goldfield are predominantly mixed magmatic-(contact) metamorphic in origin. The
influx or mixing with surface-derived fluid is associated with a significant fracturing and vein dilation event.

8.2 SOURCE OF SULPHUR

8.2.1 Geological considerations

Burnham and Ohmoto (1980) stated that sulphur is dissolved in hydrous granitic melts primarily as SH⁺, but exists in large quantities in the aqueous phase of I-type granites as both H₂S and SO₂. Thus, a significant source of sulphur was potentially available to the hydrothermal systems in the Mt Todd goldfield as magmatic sulphur that was trapped within the rock matrix of the Tennysons Leucogranite (I-type granite, Section 3.7.2), or degassed from the leucogranite during cooling.

However, it is also possible that sulphur was inherited from sulphide bearing vein-greisens hosted by the Yenberrie Leucogranite (Section 3.7.1), and from the Yenberrie Leucogranite itself. As stated previously (Section 4.2.7), the Yenberrie Leucogranite crops out 2.5 km west of the Batman deposit (Figure 4.30) and therein, lies within the thermal aureole of the Tennysons Leucogranite. The Yenberrie Leucogranite was emplaced before D₁ and is cross-cut by S₁. It is host to numerous vein-greisens bearing wolframite, molybdenite, cassiterite, bismuth sulphosalts, pyrite, arsenopyrite and copper sulphides (Walpole et al., 1968) and these are also cross-cut by the regional S₁ foliation (Section 4.2.7). Given this, it is probable that the vein-greisens formed during metasomatic alteration of the Yenberrie Leucogranite (e.g., Evans, 1987 quoting Best, 1982). Nevertheless, a source of sulphur in the form of vein-greisen sulphide was immediately available to the hydrothermal system, and/or may have been available in volatiles (H₂S and SO₂ bearing) trapped in the rock matrix of the Yenberrie Leucogranite.

A third source for sulphur to the hydrothermal system was that available in carbonaceous or sulphide-bearing sediments. In this regard, it cannot be argued that sulphur was locally derived because the sediments of the Burrell Creek Formation which host quartz-sulphide vein mineralisation are generally devoid of sulphide-bearing sediments (Chapter 3). However, carbonaceous and/or pyritic sediments do occur within older stratigraphic units (Needham et al., 1988; Stuart-Smith et al., 1993) and these may have provided a source for sulphur, in particular, pyritic carbonaceous shales and carbonaceous siltstones of the Koolpin Formation, black carbonaceous laminated shales of the Wildman Siltstone, and carbonaceous pelites, calcareous sandstones and evaporites of the Masson Formation (Section 2.1.3). Further to this, the Mt Bonnie Formation may also have proved a significant source for sulphur because the formation is host to stratiform massive sulphide mineralisation containing Au, Ag, Cu, Pb and Zn (Stuart-Smith, 1987).
Thus, several sources of sulphur were readily available to the hydrothermal system: (a) magmatic sulphur derived from the Tennysons Leucogranite, (b) magmatic sulphur trapped in the Yenberrie Leucogranite, (c) sulphur from sulphide-bearing vein-greisens that cross-cut the Yenberrie Leucogranite, and (d) carbonaceous and pyritic sediments in older stratigraphic units.

8.2.2 Sulphur isotope analyses
To further constrain the source for sulphur to the hydrothermal system, sulphur isotope ratios were measured on 38 samples of pyrite, 31 samples of pyrrhotite, 12 samples of arsenopyrite, 2 samples of chalcopyrite, one sample of marcasite (crystalline), using the $\delta^{34}S$/$\delta^{32}S$ method of Robinson and Kusakabe (1975). The 84 total samples were collected from quartz-sulphide veins in diamond drill core from the Batman deposit, Penguin prospect, Golf pit, Quigleys deposit, Regatta prospect and 50 m south of the Delta prospect (Figure 1.3), and submitted to the Central Science Laboratory at the University of Tasmania for analyses. The results are presented in Table 8.2 and Figures 8.3a-d and 8.4a-b. Sulphur isotope geothermometric calculations, as described by Ohmoto and Rye (1979), have not been investigated because of textural disequilibrium exhibited between sulphides (Chapter 6).

$\delta^{34}S$ values for sulphides from quartz-sulphide veins range from 4.5 to 13.3‰, with a mean of 10.2‰ (Figure 8.3a). The range for pyrrhotite-chalcopyrite is 8.4 to 11.9‰, for pyrite-marcasite (crystalline) is 5.4 to 13.1‰, and for arsenopyrite is 4.5 to 13.3‰, with mean $\delta^{34}S$ values at 10.2‰, 10.1‰ and 10.5‰, respectively (Table 8.2 & Figure 8.3b-d). These values overlap with $\delta^{34}S$ values of 4 to 10‰ obtained by Wygralak and Ahmad (1990) during regional studies of sulphides from mesothermal gold deposits throughout the Pine Creek Inlier (Figure 8.4). The $\delta^{34}S$ values for pyrrhotite (Figure 8.4a) from the Penguin prospect (9.6 and 9.8‰), near Delta prospect (9.6-10.7‰), and Quigleys deposit (8.7-11.8‰) form a subset to $\delta^{34}S$ values for pyrrhotite from the Batman deposit (8.4-11.9‰), while $\delta^{34}S$ values for pyrite (Figure 8.4b) from the Quigleys deposit (7.5-11.3‰), and Regatta, near Delta and Penguin prospects (11.1‰, 10.5-11.3‰, and 5.4-10.5‰, respectively) overlap with, or form a subset within $\delta^{34}S$ values for pyrite from the Batman deposit (7.4-13.1‰). The similarity in isotope signatures for pyrrhotite and pyrite, respectively, supports oxygen isotope and structural data that suggests the deposits and prospects formed at the same time by a similar process (Section 5.4 & 8.1.2), and probably from a common parent fluid.

With regard to $\delta^{34}S_{fluid}$, the absence of sulphates in the sulphide mineral assemblage (Chapter 6), and a calculated pH of 2.5 to 5.3 at 330°C, and 2.3 to 4.9 at 250°C (Section 6.4), suggest the dominant sulphur species was $H_2S$ with $\delta^{34}S_{H_2S}=\delta^{34}S_{fluid}$ during sulphide deposition. The relationship of pyrrhotite and pyrite (as the dominant sulphides analysed) to
the fluid has been calculated for particular isotope fractionation factors between sulphides and H$_2$S (Ohmoto & Rye 1979). The temperature range 300 to 360°C has been selected for pyrrhotite as an approximation to the temperature window over which this mineral precipitated based on integrated paragenetic and fluid inclusion studies (Figure 8.5), while the temperature range 250-300°C has been selected for pyrite based on physico-chemical estimates of Schoonen and Barnes (1991) which indicate that the replacement of pyrrhotite by pyrite may occur at between 250-300°C (Section 6.6). For pyrrhotite, the fractionation is 0.2 to 0.3‰, while for pyrite it is 1.2 to 1.4‰. Given that the mean $\delta^{34}$S for pyrrhotite and pyrite is approximately 10.2‰ and 10.1‰ respectively, it appears the $\delta^{34}$S$_{\text{fluid}}$ was about 10.0‰ during the pyrrhotite deposition and about 8.8‰ during the pyrite deposition, thus suggesting a similar and relatively consistent source of sulphur.

Concerning a source for sulphur, the range of $\delta^{34}$S values and the calculated $\delta^{34}$S$_{\text{fluid}}$ indicate that the source is unlikely to have been exclusively magmatic because sulphur isotope fractionation between fluid and mineral is small in the temperature range of magmatic processes (Nielsen, 1979; Ohmoto & Rye, 1979). However, the range of $\delta^{34}$S values form a subset to measured $\delta^{34}$S values of 0-30‰ for disseminated iron sulphides in carbonaceous fine grained sediments of the Pine Creek Inlier that lie outside the contact aureole of Cullen Batholith (Donnelly & Crick, 1999), and are heavier than $\delta^{34}$S values for near basal sediments of 0 ± 4‰ (Figure 8.6). Thus several alternative sources for sulphur may be suggested:

(a) Sulphur derived exclusively from sulphides in carbonaceous fine grained sediments and/or pyritic sediments.
(b) Sulphur derived by mixing of sulphides in near basal sediments with those in carbonaceous and/or pyritic sediments.
(c) Sulphur derived from mixing of sulphur in the sedimentary pile with magmatic sulphur.
(d) Sulphur derived from a varied source rock composition which has suffered substantial homogenisation.

In discussion, if sulphur were exclusively derived from sulphides in the sediments, a range of $\delta^{34}$S values might be expected that reflect the source sulphur signature and this is not the case. Instead, the restricted $\delta^{34}$S values of 4.5 to 13.3‰, and the calculated $\delta^{34}$S$_{\text{fluid}}$ given above, suggest the source of sulphur was relatively consistent. It is therefore concluded that sulphur was not derived exclusively from the sulphides in carbonaceous fine grained sediments. In corollary, it is likely that the source of sulphur was mixed and, given the geological constraints outlined in Section 8.1.1, and the availability of magmatic, sedimentary and vein sulphur (Section 8.2.1), it is probable that $\delta^{34}$S values are the result of mixing of sedimentary, vein and magmatic sources. Alternatively, $\delta^{34}$S values may reflect homogenisation of a varied source rock composition, or mixing of sulphur in near basal sediments with sulphur in carbonaceous and/or pyritic sediments. Hence these sources cannot be discounted.
8.3 SOURCE FOR METALS/LIGANDS

The data presented above has indicated a mixed magmatic-metamorphic source for fluid to the gold-bearing veins and lodes of the Mt Todd goldfield, and either a mixed magmatic-sedimentary-vein source or sedimentary source for sulphur. It is therefore reasonable to assume that metals and ligands were also derived from a mixed source, that is, the sedimentary pile and local granites.

A significant source of chloride complexes must have been the Tennysons Leucogranite. Burnham and Ohmoto (1980) stated that except for silica, chloride complexes constitute the major portion of total dissolved salts in normal granites. Those complexes include HCl, NaCl and KCl (Barker, 1983; Burnham & Ohmoto, 1980; Giggenbach, 1992), with NaCl > KCl in typical granites. Less commonly CaCl₂, FeCl₂ and FeCl₃ may constitute a portion of the total salts (Burnham & Ohmoto, 1980), although in the Mt Todd goldfield, the major source of CaCl₂ (and perhaps NaCl) was probably calcareous and evaporitic sediments in the older stratigraphic units (ie., older than the Burrell Creek Formation). It is interesting that a significant quantity of KCl was not detected during fluid inclusion analyses, however, the occurrence of sericite in wall rock assemblages suggests that KCl was consumed during rehydration reactions, eg.,

\[ \text{Cordierite} + \text{Biottite} + \text{H}_2\text{O} = \text{Muscovite} + \text{Chlorite} + \text{Quartz} \quad (\text{Pattison} \& \text{Tracy}, 1991) \]

The Tennysons Leucogranites was probably the major contributor of CO₂ (Barker, 1983; Bergantz, 1991): given the inferred shallow depth of granite emplacement at 3.5 to 5.0 km minimum (Section 4.3.5), CO₂ would have segregated due to its low solubility in silicate liquids at low pressure (Barker, 1983). Alternatively, CO₂ may have been derived from devolatilisation of calcareous sediments, or by oxidation of graphite in carbonaceous sediments (Miyashiro, 1994), ie.,

\[ \text{C} + 2\text{H}_2\text{O} = \text{CO}_2 + [\text{H}], \]

where [H] represents hydrogen bound in methane and/or phyllo-silicates (Naden & Shepherd, 1989). CH₄ may also have been derived directly from the Tennysons Leucogranite during degassing of the pluton (Burnham & Ohmoto, 1980; Wedham, 1987; Giggenbach, 1992), but it is more likely that CH₄ was generated by reaction of the mineralising fluid with graphite in the sediments during rehydration (or devolatilisation D₂), ie.,

\[ \text{C} + 2\text{H}_2\text{O} = \text{CH}_4 + [\text{O}], \]

where [O] represents oxygen bound in carbon dioxide and/or carbonate (Naden & Shepherd, 1989; Ettner et al., 1994).

The principal source of metals to the hydrothermal system was likely the sedimentary pile, in particular, pyritic, carbonaceous, ferruginous, volcanic and volcanoclastic units in older
stratigraphic units may have contributed to the overall metal-ligand budget by providing Fe, Cu, As, Pb, Zn, Bi, Au, Ag, Te, Mg, and Mn. Sulphide-bearing vein-greisens of the Yenberrie Leucogranite were available as a further source for Fe, Cu, As, Bi and S. The Tennysons Leucogranite may have contributed Fe, Bi, Au, Ag, and Te.

8.4 METAL TRANSPORT

8.4.1 Gold transport

Romberger (1988), Hannington and Scott (1989), and Shenberger and Barnes (1989), stated that AuCl₂ is favoured by low pH, high fO₂, high chloride concentrations (≥ 1 mNaCl), at elevated temperatures. Huston and Large (1989) and Hannington and Scott (1989) stated that Au(HS)₂ is favoured in low temperature solutions of high aH₂S or aHs- of neutral to alkaline pH, and in equilibrium with pyrite. Huston (1991) indicated that gold is insoluble under low temperature, oxidising conditions, and under reduced conditions, and Seward (1973) demonstrated that the solubility of gold as Au(HS)₂ is low in acid solutions. Huston and Large (1989) suggested that gold was unlikely to be transported as a thio-complex or as a chloro-complex in the pyrrhotite field at moderate to high temperature (>200°C) and low fO₂ (<10⁻⁴⁰). From this, it may be concluded that gold solubility as AuCl₂ is high under high temperature saline oxidising conditions, but low under high temperature saline reducing conditions, while gold solubility as Au(HS)₂ is high within the pyrite field at low temperatures and high ∑S concentrations.

This data indicates that gold solubility is dependent on the temperature-pH-salinity-fO₂-fS₂ conditions of the hydrothermal fluid. In order to gain an understanding of the nature of gold solubility in the hydrothermal systems in the Mt Todd goldfield, the temperature-pH-salinity-fO₂-fS₂ conditions (for the Batman deposit) are summarised:-

1. Temperature: Based on the range of homogenisation temperatures for type A, B and E fluid inclusions (357-418°C, 324-380°C, and 121-267°C, respectively), there is a general decrease in temperature which suggests the hydrothermal system cooled through the paragenesis (Figure 8.5, from event 1 to E). This is supported by physicochemical estimates from Section 6.6 that are based on (a) phase relationship data of Yund and Kullerud (1966) and Sugaki et al. (1975) for the Cu-Fe-S system, Allen et al. (1914) and Schoonen and Barnes (1991) for FeS₂ (pyrite and marcasite), Afifi et al. (1988a) and Gather and Blachnik (1974) for the telluride and gold-bismuth-telluride systems, and Hultgren et al. (1963) for the gold-bismuth system, and (b) summary data of Henley et al. (1984) for silicates that are observed in active geothermal systems. In brief, early vein formation occurred under biotite stable conditions (T>330°C, Event 1 in Figure 8.5) with precipitation of quartz and tourmaline and entrapment of type A fluid inclusions (Ttot = 357-418°C). A decrease in the fluid
temperature to minimum biotite stable conditions accompanied the precipitation of pyrrhotite 3 and chalcopyrite 1 (T=330°C, Event 4 in Figure 8.5). The precipitation of pyrite and marcasite during sulphidation of pyrrhotite suggest that the fluid continued to cool (i.e., T=250°C-300°C, Event 5 in Figure 8.5). Further cooling is indicated from the co-precipitation of hedleyite and bismuth (T<266°C, Event 8 in Figure 8.5), and subsequently, bismuth and gold (melt eutectic, T>241°C).

2. Salinity: During precipitation of sulphide, fluid salinity decreased from 35-50 wt% NaCl-CaCl₂ to 29-33 wt% NaCl eq, at which time gold precipitated in equilibrium with bismuth (Chapter 7).

3. pH:- Assuming $a_{K^+} = 0.1$, the calculated pH for the hydrothermal solution lies within the range 2.5 to 5.3 at 330°C, and 2.3 to 4.9 at 250°C (Section 6.6). As stated previously, pH may have fluctuated slightly, but stayed within the sericite stability field because neither potash feldspar nor kaolinite occur in the alteration assemblage to gold-bearing veins and lodes (cf., Heinrich & Eadington, 1986; Romberger, 1988; Giggenbach, 1992). There is a slight increase in pH concomitant to a decrease in temperature.

4. $fO_2$-$fS_2$: It has not been possible to determine the $fO_2$-$fS_2$ conditions of the hydrothermal system, however, a broad trend in $fO_2$-$fS_2$ has been established using integrated fluid inclusions and paragenetic data (Chapters 6 & 7):-

(a) Sulphide precipitation is characterised by a change from a pyrrhotite-dominant system to a pyrite-dominant system concomitant to a decrease in temperature. This describes a shift to a higher $fO_2$ for a pH of approximately 4.0, irrespective of whether there is an increase or decrease in $fS_2$.

(b) Subsequent to the precipitation of pyrite, there was a change to a pyrrhotite-talnakhite-bismuthinite-bismuth-gold bearing system concomitant to a sharp decrease in the amount of sulphide precipitated and a decrease in temperature. This describes a shift to a lower $fO_2$ for a pH of approximately 4.0, and a decrease in $fS_2$.

Overall, there is a trend to lower $fO_2$ and $fS_2$ with decreasing temperature and salinity for a given pH.

The data permit the transport of gold as AuCl₂ at higher temperatures and salinities, and as Au(HS)₂ at lower temperatures and a lower overall $fO_2$ and $fS_2$. That is, from a high temperature relatively reduced hypersaline position within the pyrrhotite stability field, the fluid evolved along a path of decreasing temperature and salinity, and increasing pH and $fO_2$, to eventually cross the pyrrhotite-pyrite boundary into the pyrite stability field (Figure 8.7). It is therefore likely that gold as AuCl₂ became less soluble and significant relative to Au(HS)₂. Indeed, the solubility of gold as Au(HS)₂ may have increased preventing the precipitation of gold. However, with a subsequent decline in the temperature, salinity, $fO_2$ and $fS_2$, gold
solubility as Au(HS)$_2^-$ would have decreased leading to a precipitation of gold.

It is therefore postulated that gold was transported as AuCl$_2^-$ at higher temperatures and salinities, and as Au(HS)$_2$ at lower temperatures and a lower overall fO$_2$-fS$_2$. A fuller thermodynamic evaluation of the hydrothermal fluid is required in order to gain a comprehensive understanding of gold complexing in the hydrothermal solution.

8.4.2 Other metals

Based on calculations by Helgeson (1969) that Cu, Pb, Zn and Fe chloride complexes are more soluble than bisulphide complexes in saline weakly acidic solutions, it is suggested that Cu, Pb, Zn, Fe and Ag were transported in the hydrothermal fluid as chloro-complexes. Bismuth was probably transported as BiCl$_3$ (Angino, 1979), but transport as BiOCl is also possible because BiCl$_3$ is readily hydrolysed by water to BiOCl + 2HCl (Cotton & Wilkinson, 1976), i.e.,

$$\text{BiCl}_3 + \text{H}_2\text{O} = \text{BiOCl} + 2\text{HCl}.$$  

Because the maximum solubility of silver is the same as that for gold (Romberger, 1988), silver may have been transported either as AgCl$^0$, AgCl$_2^-$ or AgCl$_3^{2-}$ (Seward, 1973; Heinrich et al., 1989), with AgCl$_3^{2-}$ important at high temperature, and decreasing in relative abundance with decreasing temperature (Wood et al., 1987). The transport of silver as Ag(HS)$^0$ or Ag(HS)$_2^-$ (Jaireth, 1994) may have been significant at low temperature. The presence of tellurium is indicative of an input of magmatic H$_2$Te into the hydrothermal system (Afifi et al., 1988b), but tellurium may also have seen transport as a gold-telluride complex such as Au(Te)$_2^{2-}$, Au(TeS)$_2^{3-}$, Au(Te$_2$)HS$^{2-}$ or AuTe$_2^{3-}$ (Seward, 1991), or HTe$^-$, H$_2$TeO$_3$(aq) or HTeO$_3^-$ (McPhail, 1994).

8.4.3 Gold and bismuth transport

An alternative mechanism for the transport of gold at Mt Todd has been proposed by England (1990, 1991b) in which gold may be scavenged by a bismuth-rich liquid in a manner similar to that of a collector in a fire assay because gold is preferentially partitioned into the metal phase relative to a vapour and/or aqueous phase. England (1991b) speculated that once bismuth reached saturation it formed dense metallic melt droplets that moved down through the fluid conduit where it redissolved and re-ascended. He envisaged a solution in dynamic equilibrium in which liquid-bismuth became enriched in elements compatible with the low temperature metallic melt, eg., Au and Te. The enriched liquid was later trapped and solidified during, say, a decrease in temperature.

This mechanism is somewhat interesting because it allows for the transport of significant quantities of gold relative to a gold-saturated brine. Furthermore, it offers an explanation for the large amount of bismuth and gold trapped with CH$_4$-rich fluid inclusions in fractures in quartz in the Batman deposit (Chapter 7). That is, once a bismuth melt separated from the
CH$_4$-rich fluid (containing BiCl$_3$ and/or BiOCl solutes), it scavenged gold from Au(HS)$_2$ in the fluid, and concentrated that metal within fractures. At some later stage, the bismuth and gold were entrapped with the vapour-rich fluid during healing of the host fracture and frozen within the bismuth when temperatures dropped sufficiently for bismuth to solidify.

8.5 DEPOSITION

8.5.1 Physical processes

The physical processes involved in the formation of the Batman orebody were ultimately controlled by the decrease in temperature associated with cooling of the Tennysons Leucogranite, the collapse of isotherms about the pluton, and the decay of convective circulation. The processes have been described by Burnham (1979), Norton & Cathles (1979), Knapp & Norton (1981), Barton et al. (1991), Furlong et al. (1991), Giggenbach (1992), and in Section 8.1.3, and are summarised: the overall decrease in temperature following peak thermal conditions, is likely to have resulted in convective circulation and mixing of magmatic and metamorphic fluids. Cooling may have resulted in an increase in effective tensile stress within the pluton carapace and contact aureole leading to failure and an enhancement of permeability (Ferry, 1991). In turn, an enhancement of permeability may have accelerated cooling, and so on. This dynamic process may have accompanied sympathetic dissolution and deposition with retrogression of prograde metamorphic assemblages. An increase in fluid salinity may have resulted from the consumption of H$_2$O at low fluid-rock ratios.

Of significance was the combined effect of thermal cooling and rehydration. While the former may have facilitated the development of discrete, large scale fractures, the latter is likely to have facilitated a decrease in wall rock porosity (Furlong et al., 1991) with fluid flow becoming channelised. Indeed, it is postulated that fluid flow in the Mt Todd goldfield was enhanced by cooling-induced fracturing and faulting of the Burrell Creek Formation and basement, in particular, by dilation of north-northeasterly trending faults, fractures and joints within a northeasterly trending corridor above a basement strike-slip fault (sinistral). It is suggested that fluid flow was focussed into fractures within competent rock types (eg., massive greywackes and sandstones) producing vein and lode style mineralisation. Furthermore, it is speculated that deformation early in D$_2$ was initiated during cooling of the leucogranite as a result of high contractual strain about the pluton.

The role of fracturing and cooling in the process of ore formation was therefore pivotal to gold mineralisation in the Mt Todd goldfield and ultimately lead to:-

(a) The early precipitation of crack-seal quartz from a silica-saturated brine (Section 8.1.3).
(b) A significant vein dilation event that perhaps marked the first stage of channelised fluid
flow (Section 7.6).

(c) An influx of surface-derived fluid subsequent to channelised fluid flow.

(d) Phase separation with precipitation of quartz and tourmaline during decompressive effervescence and boiling (Section 7.6) at the commencement of channelised fluid flow.

(e) The subsequent influx of hot, hypersaline, metal and sulphur rich brine, bearing minor CO\textsubscript{2} and CH\textsubscript{4} either in an immiscible form, or in solution.

(f) Fluid unmixing and the development of several hybrid fluids, as characterised by types B, C, D and E fluid inclusions (Chapter 7).

(g) Deposition of the assemblage pyrrhotite, chalcopyrite, arsenopyrite, pyrite, marcasite, and traces of cubanite, löellingite, galena, sphalerite, talnakhite, bismuthinite, rare hedleyite, Bi-rich galena, pavonite, bismuth and gold. Metal deposition was accelerated by continued fluid unmixing during a decrease in temperature.

**8.5.2 Chemical processes**

In the preceding chapters, it has been established that the parent fluid to the Batman hydrothermal system was hot (\(-400\,^\circ\text{C}\)), acidic (pH \(-4.0\) at 330\,°C), hypersaline (30-50\% NaCl-CaCl\textsubscript{2}), reduced in nature with H\textsubscript{2}S as the dominant sulphur species, and enriched in metals and sulphur derived from the Tennysons Leucogranite and/or scavenged from the sediment-granite pile. The fluid contained minor quantities of CH\textsubscript{4} and CO\textsubscript{2}.

The deposition of metal from this brine was characterised by a decrease in temperature as exemplified by (a) a decrease in T\textsubscript{Thot} for types A, B, C, D and E fluid inclusions (380-240\,°C), (b) a slight increase in pH, with a mean pH of 4.0, (c) an overall decrease in fO\textsubscript{2} and fS\textsubscript{2}, (d) a decrease in salinity from 30-50 \% NaCl-CaCl\textsubscript{2} to 27-33 wt\% NaCl eq, and (e) fluid unmixing.

The principal cause of metal deposition was fluid unmixing at a shallow crustal level (1.8 to 2.6 km) during vertical transport of the saline aqueous parent brine along joints, fractures and faults. However, it cannot be established whether fluid unmixing was initiated by decompression associated with sequential fracturing events, a decline in temperature, or as a result of fluid-rock interactions (see below).

Nevertheless, the separation of volatile and liquid phases was concurrent with mineral precipitation and a saturation of sulphur (eg., Reed & Spycher, 1985). Mineral precipitation was characterised by the sequential development of (a) pyrrhotite, cubanite, talnakhite and quartz, (b) arsenopyrite and quartz, (c) pyrrhotite with minor chalcopyrite, löellingite and quartz, (d) arsenopyrite and quartz, (e) quartz and pyrite or marcasite, (f) arsenopyrite and quartz, (g) chalcopyrite, galena, sphalerite, quartz and calcite, (h) pyrrhotite, talnakhite, quartz and calcite, and (i) bismuthinite, bismuth, gold, pavonite, hedleyite, quartz and calcite. The precipitation of biotite in equilibrium with pyrrhotite was succeeded by the precipitation of sericite and chlorite in equilibrium with pyrrhotite, and finally pyrite. The presence of sericite in the alteration assemblage suggests that pH was buffered by hydrolysis-type
reactions (Heinrich & Eadington, 1986), such as,
\[
\text{Cordierite + Biotite + H}_2\text{O} = \text{Muscovite + Chlorite + Qtz} \quad (\text{Pattison & Tracy, 1991}).
\]

With respect to gold deposition, gold and bismuth grains occur together with CH$_4$-rich and aqueous-rich saline fluid inclusions (type C) along the same micro-fractures (Chapter 7). These inclusions represent entrapment of immiscible fluids (Section 7.6). Naden & Shepherd (1989) showed that for a fluid containing no more than a few mole percent CH$_4$ (<3 mol% at 250°C), immiscibility may occur over a wide range of temperature and pressure conditions by the addition of CH$_4$ to the parent brine through fluid interaction with graphite in the wall rock. They stated that the partition of H$_2$S into the vapour phase during separation of a CH$_4$-rich vapour phase could lead to gold precipitation. Krader and Franck (1987) demonstrated that CH$_4$ solubility is significantly decreased with increasing salinity and this suggests that very little CH$_4$ is needed to initiate immiscibility in a saline brine. From this data it may be suggested that a minor addition of CH$_4$ to the Batman hydrothermal fluid, as a result of fluid-rock reactions with graphite in the wall rock, would have lead to fluid unmixing and separation of a CH$_4$-rich and aqueous-rich saline brine, the partition of H$_2$S into the vapour phase, and a decrease in the solubility of gold as Au(HS)$_2$: Alternatively, a reduction in CH$_4$ solubility may have been achieved through a decrease in temperature or pressure, without further addition of CH$_4$. In any case, fluid immiscibility concomitant to gold precipitation is indicated. Liquid-bismuth may have scavenged gold and increased its concentration in the final precipitate (England, 1990).

**8.6 SUMMARY**

1. The source of hydrothermal fluid to the gold-bearing veins and lodes in the Mt Todd goldfield was mixed magmatic-metamorphic in origin. In particular, fluids derived d from the Tennysons Leucogranite during cooling of the pluton where mixed during convective circulation with metamorphic fluid liberated from the sedimentary pile and Yenberrie Leucogranite. These fluids may have interacted with locally derived meteoric water or seawater.

2. The sources of sulphur include: (a) magmatic sulphur derived from the Tennysons Leucogranite, (b) magmatic sulphur trapped in the Yenberrie Leucogranite, (c) sulphur from sulphide-bearing vein-greisens that cross-cut the Yenberrie Leucogranite, (d) sulphur from carbonaceous and pyritic sediments in older stratigraphic units and, (e) sulphur from a varied source rock composition which suffered substantial homogenisation.

3. $\delta^{18}O$ values (relative to SMOW) for the 7 samples from quartz-sulphide veins in the Batman and Quigleys deposits range from 10.2 to 15.2‰, while $\delta^{18}O$ values for the 2 samples from
quartz-tourmaline veins in Mount Todd are 10.0% and 15.1%. A similarity in calculated 
\( \delta^{18}O_{H_2O} \) values between deposits and between vein styles supports structural data that suggests (a) the deposits formed at the same time by a similar process and probably from a common parent fluid, and (b) quartz-tourmaline veins form a subset in the quartz-sulphide vein system.

4. \( \delta^{34}S \) values for sulphides from quartz-sulphide veins from the Batman deposit, Penguin prospect, Golf pit, Quigleys deposit, Regatta prospect and 50 m south of the Delta prospect range from 4.5 to 13.3% with a mean of 10.2%. The hydrothermal fluid was characteristically reduced with \( H_2S \) as the dominant sulphur species. \( \delta^{34}S_{\text{fluid}} \) values of 10.0% to 8.8% between pyrrhotite and pyrite, respectively, indicate that the source for sulphur was relatively consistent.

5. Metals and ligands were derived from a mixed source, that is, the sedimentary pile and local granites.

6. In evolution, the fluid exhibited a tendency to lower \( fO_2 \) and \( fS_2 \) concomitant to a decrease in temperature and an overall cooling of the hydrothermal system from approximately 380°C to below 240°C. A decrease in salinity from a high of 35 to 50 wt% NaCl-CaCl₂ during initial precipitation of sulphide, to a low of approximately 27-33 wt% NaCl eq during precipitation of gold and bismuth, corresponded with a decrease in solute content. The fluid pH fluctuated but was typically acidic (pH \( \sim 4.0 \)). There was a slight decrease in pH as the hydrothermal system cooled. Gold was probably transported as \( AuCl^- \) at higher temperatures and salinities, and as \( Au(HS)^- \) at lower temperatures and a lower overall \( fO_2-fS_2 \).

7. The role of fracturing and cooling in the process of ore formation was pivotal to gold mineralisation and lead to (a) precipitation of crack-seal quartz from a silica-saturated brine, (b) channelised fluid flow, (c) an influx of surface-derived fluid subsequent to channelised fluid flow, (d) phase separation with precipitation of quartz and tourmaline during decompressive effervescence and boiling at the commencement of channelised fluid flow, (e) the subsequent influx of hot, hypersaline, metal and sulphur-rich brine, bearing minor \( CO_2 \) and \( CH_4 \) either in an immiscible form, or in solution, (f) fluid unmixing and, (g) deposition of sulphides and gold.

8. Gold deposition concomitant to fluid unmixing was the result of either (a) decreasing pressure associated with sequential fracturing events, (b) decreasing temperature, and/or (c) fluid-rock interactions with graphite in the wall rock. Liquid-bismuth may have scavenged gold and increased its concentration in the final precipitate.
Figure 8.1: Histogram of the $\delta^{18}$O quartz values (per mil) for quartz-sulphide and quartz-tourmaline veins in the Mt Todd goldfield.

Table 8.1: Oxygen isotope analyses (per mil values) of selected quartz samples from the Mt Todd goldfield. $V_2 =$ quartz-tourmaline veins, $V_3 =$ quartz-sulphide veins.

Figure 8.2: $\delta^{18}$O$_{H_2O}$ values for magmatic, metamorphic, and sedimentary rocks, and $\delta^{18}$O$_{H_2O}$ values for the Pine Creek Inlier compared to $\delta^{18}$O$_{H_2O}$ values for the Mt Todd goldfield.

A. $\delta^{18}$O$_{H_2O}$ values for magmatic water (6-12‰) after Ohmoto (1986).

B. $\delta^{18}$O$_{H_2O}$ values for metamorphic water (3-20‰) after Ohmoto (1986).

C. $\delta^{18}$O$_{H_2O}$ values for Pine Creek Inlier (5.5-10.3‰) after Wygralak and Ahmad (1990).

D. $\delta^{18}$O$_{H_2O}$ values for sedimentary rocks (-5 to 5‰) after Hoefs (1973).

E. $\delta^{18}$O$_{H_2O}$ values for quartz-sulphide veins in the Batman and Quigleys deposits.

F. $\delta^{18}$O$_{H_2O}$ values for quartz-tourmaline veins from Mount Todd.
Quartz-sulphide  
Quartz-tourmaline

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V2 = quartz-tourmaline vein  
V3 = quartz-sulphide vein  
1, 2, 3, 4 = Batman deposit  
5 & 6 = Mount Todd  
7, 8 & 9 = Quigleys deposit  

A. magmatic  
B. metamorphic  
C. Pine Creek Inlier  
D. sedimentary rocks  
E. quartz-sulphide veins  
F. quartz-tourmaline veins
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Table 8.2: Sulphur isotope analyses (per mil values) of selected samples of quartz-sulphide veins from the Mt Todd goldfield.
Figure 8.3: Histograms of sulphur isotope data for sulphides from quartz-sulphide veins of the Batman deposit, Mt Todd goldfield.

A. Total $\delta^{34}$S data.
B. $\delta^{34}$S for pyrrhotite and chalcopyrite.
C. $\delta^{34}$S for pyrite and marcasite.
D. $\delta^{34}$S for arsenopyrite.
Figure 8.4: Histograms of sulphur isotope data for sulphides from quartz-sulphide veins in the Mt Todd goldfield.
A. $\delta^{34}S$ data for pyrrhotite.
B. $\delta^{34}S$ data for pyrite.
Figure 8.5: Combined petrographic, physicochemical and fluid inclusion data for quartz-sulphide and quartz-tourmaline veins in the Mt Todd goldfield.


(2) Paragenetic range over which sericite and chlorite occur in the alteration assemblage.

(A) Range of homogenisation temperatures for type A fluid inclusions that exhibit critical behaviour, and salinity data.

(B) Range of homogenisation temperatures for type B fluid inclusions containing (3) carbonate daughter minerals, together with salinity data.

(4) Temperature range for equilibrium assemblage pyrrhotite + chalcopyrite, after Yund & Kullerud (1966) and Sugaki et al. (1975).


(6) Paragenetic range over which calcite occurs in the alteration assemblage.

(7) Maximum temperature for co-precipitation of hedleyite and bismuth (Gather & Blachnik, 1974; Afifi et al., 1988a).

(8) Temperature range for the equilibrium assemblage gold + bismuth (Hultgren et al., 1973; Angino, 1979).

(C) Melt temperature of bismuth (TeBi) in type C fluid inclusions, and salinity data.

(D) Range of homogenisation temperatures for type D fluid inclusions, and salinity data.

(E) Range of homogenisation temperatures for type E fluid inclusions.

Mineral key: qtz = quartz, tour = tourmaline, po = pyrrhotite, ccp = chalcopyrite, cub = cubanite, tal = talmakhite, asp = arsenopyrite, py = pyrite, marc = marcasite, sph = sphalerite, gn = galena, Bi = bismuth, bis = bismuthinite, pav = pavonite, hed = hedleyite, Bign = bismuth-rich galena, Au = gold.
Figure 8.6: $\delta^{34}S$ data for the Pine Creek Inlier compared to $\delta^{34}S$ data from the Mt Todd goldfield.
A. The range of $\delta^{34}S$ values for disseminated iron sulphides in carbonaceous fine grained sediments of the Pine Creek Inlier, after Donelly and Crick (1988).
B. The range of $\delta^{34}S$ values for near basal sediments of the Pine Creek Inlier, after Donelly and Crick (1988).
C. The range of $\delta^{34}S$ values obtained by Wygralak and Ahmad (1990) for gold deposits throughout the Pine Creek Inlier.
D. Total $\delta^{34}S$ values for sulphides in quartz-sulphide veins in the Mt Todd goldfield, as indicated in Figure 8.3a.
Figure 8.7: Log $fO_2$-T diagram at pH = 4.5 showing speciation for sulphur bearing fluids, and solubility of Au(HS)$_2$ and AuCl$_2$ for a 0.5 molar NaCl-H$_2$O solution in equilibrium with pyrrhotite and pyrite: modified after Hein et al. (1994). The diagram is not specific for the Batman hydrothermal system, but is used to illustrate the possible evolution path of the hydrothermal fluid given the geochemical and mineragraphic data available. From a high temperature relatively reduced position within the pyrrhotite stability field, the fluid evolved along a path of decreasing temperature to eventually cross the pyrrhotite-pyrite boundary into the pyrite stability field. Gold as AuCl$_2$ became less soluble and significant relative to Au(HS)$_2$. With a subsequent decline in the temperature and $fO_2$, gold solubility as Au(HS)$_2$ decreased leading to a precipitation of gold.
CHAPTER 9
CONCLUSIONS

9.1 STRATO-TECTONIC HISTORY

9.1.1 Sedimentary history
The gold deposits of the Mt Todd goldfield are hosted in Palaeoproterozoic geosynclinal sediments of the Burrell Creek Formation that are composed of interbedded greywackes, siltstones and shales, with minor beds of sandstone, arkosic grainstone and graphic chert. Sediments containing primary structures that are deltaic in origin (mud balls, fine-scale cross beds, asymmetric ripples) are intercalated with those deposited by turbidity current, and this suggests that sedimentation took place in a deltaic to prodeltaic environment in a river-dominant delta system (Miall, 1984; Elliot, 1986; Mitchum et al., 1993).

In the south of the goldfield, the Burrell Creek Formation is conformably overlain by sediments of the Tollis Formation that are composed of spotted tuffaceous greywackes and bedded crystal tuff. The change in the sediment character from clastic to volcanoclastic, in transition from the Burrell Creek Formation to the Tollis Formation, is consistent with a shift from a relatively mature provenance to a juvenile (rifted) nearby provenance, as described by Needham et al., (1988).

The conformable nature of the Tollis Formation on the Burrell Creek Formation in the goldfield implies that the Tollis Formation sediments are the youngest in the Palaeoproterozoic Geosynclinal Sequence and, in contrast to an unconformable relationship observed elsewhere in the Pine Creek Inlier (Needham et al., 1988, Stuart-Smith et al., 1993), argues for a revision of the stratigraphy of the Central Domain of the Pine Creek Inlier, as suggested by Jagodzinski (1992). Furthermore, the 1890 ± 15 Ma U-Pb age of Page and Williams (1988) for the Tollis Formation provides an age constraint to deposition in the upper portion of the Geosyncline Sequence.

9.1.2 Structural and metamorphic history

9.1.2.1 Tectonic events in the Mt Todd goldfield
The Palaeoproterozoic rocks of the Mt Todd goldfield have been deformed and regionally metamorphosed to greenschist facies, and contact metamorphosed by pre- to syn-orogenic plutons and a porphyry dyke. P-T conditions for contact metamorphism are approximated at
575°C for a pressure of 1-1.5 kbars and, assuming average crustal densities of 1.0 kbar ~ 3.3 km, it is estimated that the minimum depth of granite emplacement was 3.5-5.0 km. P-T conditions for regional metamorphism are approximately 400-500°C for a pressure of <4 kbars (Ferguson et al., 1980).

Early emplacement of the Yenberrie Leucogranite was concomitant with contact metamorphism of the Burrell Creek Formation to hornblende-hornfels facies (H₁), and the development of type-C₁ cordierite porphyroblasts. Emplacement preceded metasomatic alteration of the leucogranite, and the development of greisens bearing wolframite, molybdenite, cassiterite, bismuth sulphosalts, arsenopyrite and copper sulphide, in the carapace of the leucogranite.

Subsequent to emplacement of the Yenberrie Leucogranite, the rocks of the Burrell Creek Formation and Tollis Formation were subjected to regional metamorphism to lower greenschist facies (D₁). Deformation accompanied retrogression of hornblende-hornfels facies assemblages (H₁), and resulted in the development of close to tight northeast to north to northwesterly trending, asymmetric folds (F₁) that plunge gently toward the south, and a continuous axial planar cleavage (S₁). The cleavage cross-cuts and refracts through the Yenberrie Leucogranite increasing in dip on refraction, but also cross-cuts greisens hosted in the leucogranite carapace. The east facing of folds suggests that tectonic transport was toward the east.

The formation of conjugate buck quartz veins (V₁), breccias, veinlets and large single buck quartz blows, occurred early in D₁ at which time, a coaxial stress field may have operated in which σ₁ plunged gently toward the west-southwest; σ₂ plunged moderately south-southeast; σ₃ plunged moderately north. Throughout the goldfield the veins are cross-cut by S₁ indicating they formed early in D₁ prior to development of the foliation.

Folding in D₁ was associated with the development of numerous third-order faults with sinistral movement on northwest and north-northwesterly trending faults (R and P faults), dextral movement on northeast and north-northeasterly trending faults (X and R' faults), reverse movement on northerly trending faults, and normal movement on easterly trending faults. The fault orientations and a left step in fold axes indicate that a non-coaxial sinistral stress field may have operated at this time in D₁ with σ₁ directed westerly, σ₂ directed south and σ₃ directed sub-vertically, i.e., folding was associated with sinistral strike-slip displacements along northwest trending faults (eg., Quigleys Fault).

The discordance between the type of stress field operating early in D₁ (during the formation of buck quartz veins) and that operating during folding and faulting, may be the result of the progressive change from coaxial mechanics to non-coaxial sinistral strike-slip mechanics during deformation, with the corollary that early structures were rotated anticlockwise in the latter non-coaxial sinistral stress field (eg., buck quartz veins).
Folding in D₁ was also associated with the development of slickenfibres on bedding planes throughout the goldfield, in an orientation perpendicular to sub-perpendicular to the hinges of F₁. In the Batman and Quigleys prospects, the transport vector for bed displacement, as inferred from the mean azimuth of fibres, is directed toward 098° and 102°, respectively, and slip was sinistral-reverse. This indicates that σ1 was directed westerly, and suggests that fold and fibre formation were concurrent, i.e., folding involved flexural slip (Hobbs et al., 1976; Koerber, 1989).

Subsequent to D₁, the sediments of the Mt Todd goldfield were intruded by the Tennysons Leucogranite and contact metamorphosed to hornblende-hornfels facies (H₂), with the concurrent development of type-C₂ cordierite porphyroblasts in cross-cutting relationship to type-C₁ cordierite porphyroblasts and the regional S₁ foliation. The pluton is transgressive with respect to D₁ folds, and probably transgressive at depth with respect to the Yenberrie Leucogranite.

Following, or concurrent with emplacement of the Tennysons Leucogranite, the rocks of the goldfield were metamorphosed to lower greenschist facies (D₂). Deformation accompanied retrogression of H₂ hornblende-hornfels facies assemblages and resulted in the development of westerly trending, discontinuous open-style folds (F₂) and a spaced disjunctive to fracture cleavage (S₂) in transection to the folds. Refolding of F₁ by F₂ resulted in the development of symmetric to non-symmetric, crescent-shaped domes and basins. F₂ are asymmetric in profile, and a northerly facing suggests that tectonic transport was toward the north. The plunge of the folds ascribes a basinal geometry that may be indicative of fold drape across the Tennysons Leucogranite, or doming of the sediments in the west of the goldfield during pluton emplacement. The latter would imply that plutonism and deformation were coeval: this agrees with the conclusions of Stuart-Smith et al. (1993) that east-west trending, open style folds are associated with the emplacement of the Cullen Batholith (1835-1820 Ma). S₂ is cross-cutting with respect to the (1) Yenberrie Leucogranite and type-C₁ cordierite porphyroblasts formed during emplacement of the leucogranite, (2) D₁ structures (F₁, S₁, and buck quartz veins), and (3) hornfelsed sediments in the contact aureole of the Tennysons Leucogranite. S₂ also cross-cuts a porphyry rhyolite dyke suggesting that the dyke was emplaced pre- to syn-D₂. The injection of the dyke (along a fault?) was probably synchronous with emplacement of the Tennysons Leucogranite.

The cross-cutting nature of gold-bearing quartz-sulphide veins with respect to D₁ structures (buck quartz veins, F₁ and S₁), and the cross-cutting nature of D₂ structures (F₂, S₂ and D₂ faults) with respect to the gold-bearing quartz-sulphide veins suggests the veins formed pre- to syn-D₂ (see Section 9.1.2.3).

During D₂ a non-coaxial stress field may have operated with sinistral movement on third-order northeast and east-northeasterly trending faults (R and P faults), dextral movement on northwesterly trending faults (R’ faults), and normal movement on northerly trending faults.
The maximum compressive stress, \( \sigma_1 \), was orientated approximately north-south.

The final tectonic event recognised in the Mt Todd goldfield is associated with the reactivation of west-northwest and northwesterly trending sinistral strike-slip faults in \( D_3 \) (eg., reactivation of first-order faults in the Pine Creek Shear Zone and second-order faults in the Yenberrie Shear Zone). The faults are associated with a steeply dipping (sinistral) Type II \( S_3 \)-C foliation (cf., Lister & Snoke, 1984) and mesoscopic en échelon folds (\( F_3 \)) that trend oblique to the strike of the faults in a left stepping array (sinistral array). The faults offset \( D_1 \) and \( D_2 \) structures and cross-cut the Tennysons Leucogranite in which they resolve narrow zones of sinistral shear dislocation. The faults may be associated with the development of calcite-base metal veins (\( V_4 \)) and epithermal style quartz veins (\( V_5 \)).

**9.1.2.2 Regional tectonic events and the Mt Todd goldfield**

The tectonic events of the Pine Creek Inlier, as detailed in Chapter 2, are compared to the tectonic events in the Mt Todd goldfield in Table 10.1. The regional \( F_2 \) and \( F_3 \) of the Maud Creek Event (1850 Ma), and fault reactivation in the Shoobridge Event (1770-1780 Ma) correlate with \( F_1 \) and \( F_2 \), and \( F_3 \) fault reactivation in the Mt Todd goldfield, respectively. Folds associated with the Nimbuwah Event (regional \( F_1 \)) do not occur in the goldfield.

The emplacement of the Yenberrie Leucogranite pre- to syn-\( D_1 \) in the goldfield suggests that the duration of emplacement of the Cullen Batholith may span a greater time interval than indicated by Stuart-Smith *et al.* (1993), ie., 1850-1820 Ma. Alternatively, the age of the Maud Creek Event may be incorrect and should be constrained by the age of the Cullen Batholith at 1835-1820 Ma.

**9.1.2.3 Timing of quartz-sulphide veins and lodes**

The timing of gold-bearing veins and lodes may be summarised as follows:-

1. Veins in the Batman deposit are cross-cut and deformed by the regional \( S_2 \) fabric and are offset along east-northeasterly trending sinistral strike-slip faults. They cross-cut type-\( C_2 \) cordierite porphyroblasts which formed during emplacement of the Tennysons Leucogranite, and these are also cross-cut by the regional \( S_2 \) fabric.

2. Lodes and stockwork veins at the Quigleys deposit are folded about an \( F_2 \) anticline and offset along northeast trending \( D_2 \) sinistral strike-slip faults.

3. Lodes at the Quigleys deposit cross-cut buck quartz veins, \( D_1 \) faults and the regional \( S_1 \) fabric.

Based on these cross-cutting relationships, it is concluded that gold-bearing veins and lodes were formed early in \( D_2 \) and after the peak thermal metamorphism associated with the emplacement of the Tennysons Leucogranite (Figure 4.34). The orientation of quartz crack-seal fibres in quartz-sulphide veins in the Batman deposit, and the mean azimuth of striae on the footwall to lodes in the Quigleys deposit (the striae record a normal sense of slip) indicate that the extension direction (\( \lambda_1 \)) during vein and lode formation was directed westerly.
9.1.2.4 Micro-structure of vein/lode formation

The quartz-sulphide veins of the Batman deposit (main vein set strike and dip 018° 60° E) are composed of a two stage fill: syntaxial quartz crack-seal fibres which line the vein walls and open space type silicate-sulphide-carbonate fill. The quartz crack-seal fibres (qtz 1) are indicative of the extension direction ($\lambda_1$) during vein formation and this was orientated 8°->288°. The $\lambda_1$ direction should lie normal to the plane of extension, the geometry of which describes the geometry of the Batman orebody at 018° 72° E. The orebody is therein composed of a stack of discontinuous sheet-like veins that are orientated 018° 60° E within an 'ore envelope' that is orientated 018° 72° E.

The open space silicate-sulphide-carbonate fill of the quartz-sulphide veins precipitated in the sequence:-

- euhedral quartz (qtz 2) + tourmaline + biotite.
- pyrrhotite 1 + cubanite + talnakhite 1 + arsenopyrite 1 + biotite + muscovite.
- pyrrhotite 2 + quartz 3 + sericite + chlorite.
- (pyrrhotite 3 + chalcopyrite 1 => löellingite => arsenopyrite 2 & 3) + quartz 3 + sericite + chlorite.
- (pyrrhotite 4 => pyrite 1) + quartz 3 + sericite + chlorite.
- (pyrite 2 => marcasite 1 & 2 => pyrite 3 => arsenopyrite 4) + quartz 3 + sericite + chlorite.
- (chalcopyrite 2 => sphalerite => galena => chalcopyrite 3) + quartz 3 + sericite + chlorite + calcite.
- pyrrhotite 5 + talnakhite 2 + quartz 3 + sericite + chlorite + calcite.
- (bismuthinite + bismuth sulphosalts => bismuthinite + bismuth + gold => bismuthinite) + quartz 3 + sericite + chlorite + calcite.

Open space silicate-sulphide-carbonate fill is associated with (a) crack-seal mechanics during the precipitation of the earlier sulphide assemblages, and (b) with several phases of micro-fracturing such that silicate, sulphide and/or carbonate minerals cross-cut, meander through, and offset pre-existing vein material thus producing a complex array of cross-cutting relationships. In this study at least nine mutually exclusive fracturing events have been recognised in association with silicate-sulphide-carbonate fill in the Batman deposit.

In the case of lodes and footwall stockwork veins at the Quigleys deposit, the mineralogy of the veins and lodes includes crack-seal fibre quartz (qtz 1) 10, euhedral quartz (qtz 2), arsenopyrite, pyrite, marcasite, chalcopyrite, löellingite, bismuth, bismuthinite, gold, electrum, and minor pyrrhotite, galena and sphalerite. However, lode/vein fill is significantly modified by syn- to post-depositional fault activity that produced proto-cataclasite to cataclasite type textures. This suggests that vein and lode formation at the deposit was typified by extension or dilation concomitant to fracturing and/or faulting.
9.1.2.5 Batman deposit versus the Quigleys deposit

The quartz-sulphide veins of the Batman deposit, and the lodes and footwall stockwork veins of the Quigleys deposit are remarkably similar with respect to strike geometry, morphology and vein forming process despite the considerable syn- to post-depositional modification of the footwall stockwork veins at the Quigleys deposit. In terms of position, the lodes and veins lie within the middle contact aureole of the Tennysons Leucogranite and typically occupy faults, joints and bedding planes that strike north-northeasterly within a corridor that trends northeasterly across the aureole. The deposits are therefore structurally controlled. There is a broad agreement in the type of host lithofacies (i.e., siltstone or greywacke dominant), and in the dominance of chlorite and sericite in the alteration selvedge. The pyrite-arsenopyrite dominant sulphide assemblage of the Quigleys deposit forms a subset to the pyrrhotite-pyrite-arsenopyrite dominant sulphide assemblage of the Batman deposit. $\delta^{18}$O$_{H_{2}O}$ values of 8.2-10.1% at 357°C and 5.9-10.9% at 390°C for the Batman deposit, overlap with $\delta^{18}$O$_{H_{2}O}$ values of 5.1-9.2% at 357°C and 5.7-10.8% at 390°C for the Quigleys deposit. $\delta^{34}$S values of 8.7-11.8% for pyrrhotite and 7.5-11.3% for pyrite from the Quigleys deposit form a subset to those from the Batman deposit of 8.4-11.9% for pyrrhotite and 7.4-13.1% for pyrite. Collectively, the data suggests that the deposits formed at the same time and by a similar tectonic process. This being the case, it is likely that the majority of quartz-sulphide vein and lode occurrences in the Mt Todd goldfield where formed in a similar time-tectonic regime because by and large, they are similar in strike geometry, morphology, metallogeny and isotopic signature to that exhibited in the Batman and/or Quigleys deposit.

9.1.2.6 Quartz-sulphide veins and quartz-tourmaline veins

The relationship between quartz-sulphide and quartz-tourmaline veins in the Mt Todd goldfield is summarised in Table 9.2. The mineralogy of quartz-tourmaline veins is composed of crack-seal qtz (qtz 1), euhedral quartz (qtz 2), tourmaline, biotite, and rare masses of pyrrhotite and chalcopyrite, and thus the veins form a subset to quartz-sulphide vein mineralogy, as detailed above. Quartz-tourmaline veins cross-cut buck quartz veins and $S_1$, and are cross-cut by $S_2$. They do not cross-cut quartz-sulphide veins and are not crosscut by those veins. Furthermore, $\delta^{18}$O$_{H_{2}O}$ values for quartz-sulphide veins of 5.1-10.1% at 357°C, and 5.9-10.9% at 390°C, overlap with $\delta^{18}$O$_{H_{2}O}$ values for quartz-tourmaline veins of 4.9-10.1% at 357°C, and 5.7-10.8% at 390°C. This data supports the contention that quartz-tourmaline veins form a subset within the quartz-sulphide vein group and suggest that quartz-tourmaline and quartz-sulphide veins are co-genetic.
9.2 The Hydrothermal Fluid

The hydrothermal fluid responsible for gold mineralisation in the Mt Todd goldfield was derived chiefly from a mixed magmatic-metamorphic source. In particular, fluids derived from the Tennysons Leucogranite during cooling of the pluton were mixed (during convective circulation) with metamorphic fluids that were liberated from the sedimentary pile and Yenberrie Leucogranite subsequent to thermal metamorphism and perhaps early regional metamorphism in D2. The fluids were enriched in metal and sulphur, and these were derived either as magmatic sulphur and metal, or scavenged during fluid-rock reaction from metal-sulphur sources deep within the sedimentary pile, the Yenberrie Leucogranite, or within greisens which cross-cut the Yenberrie Leucogranite.

The hydrothermal fluid was reduced in nature reflecting a reduced parent fluid and a reduced fluid source. H2S was likely the dominant sulphur species. Measured δ34S values range from 4.5% to 13.3% and indicate that the source for sulphur was mixed, i.e., the δ34S values reflect mixing of magmatic sulphur derived from the Tennysons Leucogranite, vein sulphur derived from sulphide-bearing greisens in the Yenberrie Leucogranite, or sulphur derived from carbonaceous and/or pyritic sediments located deep within the sedimentary pile. Alternatively, δ34S values reflect a substantial homogenisation of a varied source rock composition. Calculated δ34Sfluid values of 10.0% to 8.8% between pyrrhotite and pyrite, respectively, indicate that the source for sulphur was relatively consistent throughout fluid evolution.

The hydrothermal fluid was relatively consistent in bulk fluid composition and contained varying proportions of H2O, CO2, CH4, Na+, Cl-, and Ca2+, with minor amounts of Mg2+, Mn2+, K+, Li+, Al3+, Fe3+, H2CO3, HCO3- or CO3-. This suggests that the hydrothermal fluid was relatively homogeneous in character, and probably derived from a common parent fluid and a common fluid source (i.e., a mixed magmatic-metamorphic source as stated above).

In evolution, the hydrothermal fluid exhibited a tendency to lower fO2 and fS2 concomitant to a decrease in temperature and salinity for a given pH. Initial precipitation of crack-seal fibre quartz (qtz 1) from a silica saturated brine occurred during early retrograde metamorphism (cooling of the Tennysons Leucogranite) by progressive fracturing sealing of the host rocks. Oxygen isotope data suggest this fluid was derived from a mixed magmatic-metamorphic source.

The precipitation of euhedral quartz (qtz 2) and tourmaline occurred at an early stage of open space silicate-sulphide-carbonate fill. It coincided with a significant vein dilation event and accompanied boiling of a low to moderately saline (1-20 wt% NaCl eq), CO2-CH4 bearing,
surface-derived fluid, at an approximate minimum temperature of 370°C and at a minimum depth of 1.8 to 2.6 km. This was possibly preceded by decompressive effervescence (ie., an immiscible fluid of low salinity subsequently suffered decompressive boiling). This fluid apparently interacted with a mixed magmatic-metamorphic fluid.

The precipitation of the silicate-sulphide-carbonate assemblage outlined in Section 9.1.2.4 and Chapter 6 accompanied an influx of hot (approximate minimum of 380°C), acidic (pH ~ 4.0 at 330°C), metal and sulphur enriched, hypersaline brine (30-50% NaCl-CaCl₂). The principal cause metal deposition was fluid unmixing at a shallow crustal level (minimum depth of 1.8 to 2.6 km) during a decrease in temperature from approximately 380°C to 240°C, a decrease in salinity from 30-50% NaCl-CaCl₂ to 29-33 wt% NaCl eq, and vertical transport of the brine along joints, fractures and faults.

Gold precipitation occurred late in the development of the silicate-sulphide-carbonate assemblage from a hot (~ 250°C), acidic (pH of 2.3 to 4.9 at 250°C), hypersaline brine (29-33 wt% NaCl eq), and accompanied the precipitation of bismuth and the entrapment of CH₄-rich fluid inclusions. Liquid bismuth may have scavenged gold and increased its concentration in the final precipitate (England, 1990).

9.3 GENETIC MODEL

The gold deposits of the Mt Todd goldfield formed by hydrothermal activity concomitant with retrograde (contact) metamorphism and associated deformation during cooling and crystallisation of the Tennysons Leucogranite, and early in D₂. It is speculated that pluton cooling resulted in the development of effective tensile stresses which dilated and/or reactivated structures generated during pluton emplacement and/or during D₁ (Furlong et al., 1991), or which fractured the country rock carapace as is typical during cooling of shallowly emplaced plutons (Balk, 1937; Burnham, 1979; Knapp & Norton, 1981). In particular, this model invokes sinistral reactivation of a north-easterly trending basement strike-slip fault, causing brittle failure in the upper crust and/or dilation of existing north-northeasterly trending faults, fractures and joints in competent rock types such as greywackes and siltstones. The generation of dilatant structures above the basement structure (ie., along a northeasterly trending corridor overlying the basement fault) coupled with a sudden reduction in pressure (Sibson, 1989), may have facilitated channelization of fluid flow into the upper crust (Furlong et al., 1991). Rising fluids decompressed causing separation of the vapour and fluid phases concurrent with mineral precipitation (ie., the fluid boiled or effervesced). Throttling of the conduit or fluid pathways probably resulted in overpressuring of the fluid (Roedder, 1984), this giving way to further fracturing, and so on.
Phase separation accompanied a decrease in temperature, although ultimately, the hydrothermal system cooled as isotherms collapsed about the cooling pluton (Knapp & Norton, 1981).

Further to this, the evolution of the metal-sulphur enriched magmatic-metamorphic hydrothermal fluid (as described in Section 9.2) in terms of temperature, $fO_2$, $fS_2$, pH and salinity is the result of retrograde reactions accompanying rehydration of the country rock and pluton, and the release of appreciable amounts of Na, Ca, Fe, and O (Crawford et al., 1979a; Bennett & Barker, 1992). The initial rate of rehydration was rapid (Knapp & Norton, 1981) (because rehydration reactions are exothermic) and fluid-rock reactions were low (Bennett & Barker, 1992). The removal of $H_2O$ by mineral hydration resulted in high initial salinities (Crawford, 1981). Convective circulation facilitated fluid mixing.

### 9.4 THERMAL AUREOLE GOLD DEPOSITS

Thermal aureole gold deposits describe a class of gold deposits localised in the thermal aureole of granite plutons and formed during dilatant deformation that is synchronous with granite emplacement and thermal metamorphism (Wall, 1989). The deposits commonly form during prograde to peak thermal conditions, but also during retrograde thermal conditions. The metallogeny of thermal aureole gold deposits defines a type: fine grained gold (rarely nuggetty) in en échelon, stockwork, and single vein systems, or as disseminations and skarns, is typically associated with pyrite, pyrrhotite, arsenopyrite and/or loellingite, with lesser occurrences of galena, sphalerite, chalcopyrite, telluride and bismuth sulphides (Wall, 1989). Base metal sulphide content is typically low. Gold transporting fluids of low to high salinity are reduced in nature.

These characteristics are typified in the gold deposits of the Mt Todd goldfield. All are located within the thermal aureole of the Tennysons Leucogranite, and as described previously, formed during retrograde thermal metamorphism. They largely comprise sheeted, stockworked, or single vein systems (lodes): the latter form an en échelon array at the Quigleys deposit. Sulphide mineralogy is dominated by pyrrhotite or pyrite with arsenopyrite or loellingite, and base metal sulphide content is low. Gold is intimately associated with bismuth (in fluid inclusions), bismuth sulphosalts, and rare telluride. Fluid salinities were high to moderate, and the hydrothermal fluid was strongly reduced.

Given the geological and geochemical constraints outlined throughout this text, and the similarity in deposit style and metallogeny to thermal aureole gold deposits, it is concluded that the Mt Todd gold deposits collectively comprise a field of thermal aureole gold.
mineralisation. The unusual location of the Batman deposit on the northeastern face of a basin produced by the intersection of and F₁ and F₂ syncline, sets it apart with respect to other gold deposits in the Pine Creek Inlier since they are dominated by quartz vein or lode style mineralisation in fractures and shear zones located close to, or within the axes of antiformal or domal structures, eg., Enterprise Mine (House, 1990), Spring Hill prospect (Nicol, 1991), Union Reefs prospect (Donaldson, 1992), Howley district (Partington, 1990), and Moline/North Hercules mines (Nicholson & Eupene, 1990). This, together with the fact that gold is intimately associated with bismuth in fluid inclusions, suggests the Batman deposit constitutes a new deposit style. Similarly, the proto-cataclasite to cataclasite-"chain of lodes" style mineralisation of the Quigleys deposit sets it apart from other deposits styles in the Pine Creek inlier as unique, and highlights the exploration potential of fault systems.
### Table 9.1
Comparison between the tectonic events of the Pine Creek Inlier and the Mt Todd goldfield.

<table>
<thead>
<tr>
<th>Pine Creek Inlier (regional)</th>
<th>Mt Todd goldfield</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Nimbuwah Event, (1885-1870 Ma)</strong></td>
<td>Bedding parallel thrusts, NW trending recumbent to upright folds ($F_1$)</td>
</tr>
<tr>
<td><strong>Maud Creek Event, (1850 Ma)</strong></td>
<td>N-NE trending upright folds ($F_2$)</td>
</tr>
<tr>
<td><strong>Cullen Batholith (1835-1820 Ma)</strong></td>
<td>East-west trending open folds ($F_3$).</td>
</tr>
<tr>
<td><strong>Shoobridge Event (1770-1780 Ma).</strong></td>
<td>Reactivation of faults (principally NW trending)</td>
</tr>
</tbody>
</table>
Quartz-sulphide veins

- qtz 1 (crack-seal)
- qtz 2 + tour + bt
- po 1 + cub + tal 1 + asp 1 + bt + musc
- (po 2 => po 3 + ccp 1 => loell => asp 2 & 3) + qtz 3 + ser + chl
- (po 4 => py 1) + qtz 3 + ser + chl
- (py 2 => marc 1 & 2 => py 3 => asp 4) + qtz 3 + ser + chl
- (ccp 2 => sph => gn => ccp 3) + qtz 3 + ser + chl + cal
- po 5 + tal 2 + qtz 3 + ser + chl + cal
- Bi + hed + pav + Bign + qtz 3 + ser + chl + cal
- Au + Bi + bis + qtz 3 + ser + chl + cal
- bis + qtz 3 + ser + chl + cal

Quartz-tourmaline veins

- qtz 1 (crack-seal)
- qtz 2 + tour + bt
- po
- ccp

Cross-cuts buck quartz veins (at Quigleys)
- Cross-cuts S1
- Cross-cut by S2
- Do not cross-cut quartz-tourmaline veins
- Cross-cuts buck quartz veins (at Mount Todd)
- Cross-cuts S1
- Cross-cut by S2
- Do not cross-cut quartz-sulphide veins

δ¹⁸O_H₂O = 5.1-10.1% at 357°C
δ¹⁸O_H₂O = 5.9-10.9% at 390°C
δ¹⁸O_H₂O = 4.9-10.1% at 357°C
δ¹⁸O_H₂O = 5.7-10.8% at 390°C

Table 9.2: Quartz-sulphide veins and quartz-tourmaline veins are compared in three boxes. The first compares mineralogy; the second structural data; the third oxygen isotope data. Mineral key: qtz = quartz, tour = tourmaline, po = pyrrhotite, ccp = chalcopyrite, cub = cubanite, tal = talnakhite, asp = arsenopyrite, py = pyrite, marc = marcasite, sph = sphalerite, gn = galena, Bi = bismuth, bis = bismuthinite, pav = pavonite, hed = hedleyite, Bign = bismuth-rich galena, Au = gold, ser = sericite, chl = chlorite, bt = biotite, cal = calcite.
BIBLIOGRAPHY


Anon, 1993, Big raising for big project, in Gold Gazette (Ed: R. Louthean). Volume 3, No. 47, April 5, 1993, pp 38 & 42. Published for Resource Information Unit Ltd, Western Australia.


Bergantz, G.W., 1991, Chemical and physical characterisation of plutons, in Contact Metamorphism (Ed: D.M. Kerrick). Reviews in Mineralogy, Volume 26, pp. 13-42. (Mineralogical Society of America.)

Bowers, T.S. & Helgeson, H.C., 1983a, Calculations of the thermodynamic and geochemical consequences of nonideal mixing in the system $\text{H}_2\text{O}-\text{CO}_2-\text{NaCl}$ on phase relations in geologic systems: Equations of state for $\text{H}_2\text{O}-\text{CO}_2-\text{NaCl}$ fluids at high pressure and temperature. GEOCHIMICA ET COSMOCHIMICA ACTA, Volume 47, pp 1247-1275.

Bowers, T.S. & Helgeson, H.C., 1983b, Calculations of the thermodynamic consequences of nonideal mixing in the system $\text{H}_2\text{O}-\text{CO}_2-\text{NaCl}$ on phase relations in geologic systems: Metamorphic equilibria at high pressures and temperatures. AMERICAN MINERALOGIST, Volume 68, pp 1059-1075.


Donnelly, T.H. & Crick, I.H., 1988, Biological and abiological sulphate reduction in two Proterozoic Australian basins. TERRA COGNITA, Volume 8, No 3, pp. 222.


Franck, U., 1985, Aqueous mixtures to supercritical temperatures and at high pressures. PURE AND APPLIED CHEMISTRY, Volume 57, Number 8, pp. 1065-1070.


Hollister, L.S. & Burruss, R.C., 1976, Phase equilibria in fluid inclusions from the Khtada Lake metamorphic complex. GEOCHIMICA ET COSMOCHIMICA ACTA, Volume 40, pp. 163-175.


Krader, T. & Franck, E.U., 1987, The ternary systems H$_2$O-CH$_4$NaCl and H$_2$O-CH$_4$CaCl$_2$ to 800K and 250MPa. BER BUNSENES PHYS CHEM, Volume 91, pp. 627-634.


McCuaig, T.C. & Kerrich, R., 1994, P-T-t-deformation-fluid characteristics of lode gold deposits: evidence from alteration systematics, in Alteration and alteration processes associated with ore-forming systems (Ed: D.R. Lentz), Short Course Notes, Volume 11.


Page, R.W. & Williams, I.S., 1988, Age of the Barramundi orogeny in Northern Australia by means of ion microprobe and conventional U-Pb zircon studies. PRECAMBRIAN RESEARCH, Volume 40/41, pp. 21-36.


Powell, C.McA., 1979, A morphological classification of rock cleavage. TECTONOPHYSICS, Volume 58, pp. 21-34.


Reynolds, T.J., 1990, Workshop on application of fluid inclusions to minerals exploration. Course Notes for the Key Centre for Ore Deposit and Exploration Studies, University of Tasmania, 35 p.


Robert, F. & Kelly, W.C., 1987, Ore-forming fluids in Archaean gold-bearing quartz veins at the Sigma Mine, Abitibi Belt, Quebec, Canada. ECONOMIC GEOLOGY, Volume 82, pp. 1464-1482.


Robinson, B.W. & Kusakabe, M., 1975, Qualitative preparation of sulphur dioxide, for $^{34}$S/$^{32}$S analyses, from sulphides by combustion with cuprous oxide. ANALYTICAL CHEMISTRY, Volume 47, No 7, June 1975, pp. 1179-1181.


Sangree, J.B., Mitchum, R.M. & Vail, P.R., 1992, Application of sequence stratigraphy to hydrocarbon exploration, in Sequence Stratigraphy of European Basins, Abstracts Volume. Published by Institut Français du Pétrole, France, and Paléontologie analytique et Géologie sédimentaire, Université de Bourgogne, France. No pagination.


