A Structural Examination of the Telfer Gold-Copper Deposit and Surrounding Region, northwest Western Australia: The Role of Polyphase Orogenic Deformation in Ore-deposit Development and Implications for Exploration.

VOLUME 1

Thesis submitted by
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in October, 1996

for the degree of Doctor of Philosophy in the
Department of Earth Sciences at
James Cook University of North Queensland.

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Simon A.J. Hewson
3rd October, 1996.

I dedicate this thesis to my parents, John and Monica Hewson, without whose love, support and financial assistance over my life this would never have happened. I will remain forever indebted for their encouragement throughout my academic endeavours.

"The heights by great men reached and kept
Were not attained by sudden flight
But they, while their companions slept
Were toiling upward in the night".

HENRY WORDSWORTH LONGFELLOW

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PREAMBLE

The Telfer Au (+Cu) deposit is one of Australia's premier gold-producers, currently accounting for approximately 400,000oz. per year. It's discovery in 1972 heralded the recognition of the Proterozoic Paterson Province, in which it is located, as a polymetallic terrane with the potential for significant gold and base-metal mineralisation. The Paterson Province represents a small exposure ($\approx$ 36 000km^2) of the large NW-SE trending Paterson Orogen, a continental-scale orogenic belt that passes across northern Western Australia and into central Australia. Rocks in the Paterson Province have recorded a protracted Proterozoic history for this orogenic belt, which includes continent-continent collision at $\approx$ 1250Ma (Watrara Orogeny) and late-Proterozoic collisional tectonism (the Paterson Orogeny) between 700 and 600Ma. The latter generated considerable mineralisation in middle- to late-Proterozoic metasedimentary rocks of the Paterson Province.

Late-Proterozoic mineralisation in the Paterson Province is well developed in the NE region where numerous syn- and post-tectonic granitoids, gabbros and dolerites intruded the metasedimentary sequence. However, mineral exploration in this region is commonly hindered by extensive Phanerozoic and Tertiary cover that precludes observation of much of the Proterozoic sequence. Additionally, the province lies within the Great Sandy Desert, and is thus covered by extensive aeolian sand deposits. The large amount of younger cover has resulted in relatively small scattered "windows" that expose the mineralised Proterozoic rock sequence. The NE region of the Paterson Province provides one of the best exposures of this sequence and hosts the Telfer deposit.

Sampling of gossanous and stratabound quartz veining in the Telfer Dome (a regional antiformal fold) in 1972 by geologists from both Day Dawn Minerals and Newmont Pty Ltd identified gold enriched horizons within the pelitic sedimentary sequence (Telfer Formation) exposed in the dome. Subsequent drilling of these horizons confirmed an initial resource of 1 Million ounces of gold and mining activity commenced in 1975. During the early 1980s mining concentrated on one particular reef that outcropped in Main Dome, a sub-dome of the Telfer Dome. This reef, the Middle Vale Reef (MVR) exhibited strong secondary enrichment of gold and has historically comprised a significant resource in the Telfer deposit (Dimo, 1990). During the middle to late 1980s a shift to high-volume low-grade mining was facilitated by the ongoing success of dump leach extraction of gold from rocks previously too low-grade to be milled. This increased throughput caused production to expand to the second sub-dome (West Dome) as the E-Reefs were mined, and helped to make Telfer one of the top four gold producers in Australia during the late 1980s.

More recently, deep diamond drilling in Main Dome (commenced in 1992) has led to the discovery of approximately ten to twelve new reefs at depth in the dome. Underground mining, which had commenced in 1990, has been extended through an exploration decline to the upper of these newly discovered reefs (M10 and M30). This decline is currently being extended to reach the deepest reef, the I30, at approximately 1100m below the present ground surface. This should occur by the end of 1997. Another consequence of the deep drilling has been the further confirmation of an epigenetic genesis for the Telfer deposit, and particularly the identification of mineralisation in units other than the Telfer Formation.

Initial research on the stratabound reefs in the Telfer deposit suggested that they had formed through syngenetic exhalative processes (Tyrwhitt, 1979; Turner, 1982). A variation on this, whereby the MVR was considered to have formed as an evaporite horizon that was subsequently replaced by quartz-sulphide assemblages, was proposed by Royle (1985). However, the expansion of mining and deep drilling within the deposit provided increasing evidence for an epigenetic origin. This came both through structural observations and geochemical/fluid-inclusion studies that indicated the reefs were locally discordant, had associated stockwork veining in both the foot- and hanging-walls, contained magmatic elements in the ore-assemblage and that the ore-fluids were of variable salinity and temperature (Goellnicht, 1987). These observations led to an epigenetic model whereby magmatic fluids from the regional granitoids had mixed with cooler connate/formational waters, and had precipitated in structurally controlled, and compositionally favourable, sites within the Telfer Dome (Goellnicht, 1987; Goellnicht et al., 1989). Subsequent research downgraded the role of the granites, suggesting that they acted more as heat sources to convectively circulate connate/contact-metamorphic fluids that scavenge elements from the sedimentary sequence (Hall & Berry, 1989; Rowins, 1994).

The changing ideas on the genesis of the Telfer mineralisation are reflected in the changing focus of mineral exploration in the Telfer region and the Paterson Province. Initial exploration, utilising the syngenetic exhalative model, concentrated on locating further outcropping Telfer Formation. The inferred variable thickness of exhalative lenses was considered to have assisted the formation of the regional domal antiforms (Turner, 1982), and consequently those domes that exposed the Telfer Formation were targeted. However, recognition of an epigenetic genesis has focussed exploration activity towards targeting favourable host structures, such as regional folds and other ore-fluid traps. The change in exploration strategy means that a greater reliance is now placed on the structural geological setting of mineralisation in the Paterson Province. This study represents the first formal examination whereby the structural geological setting of mineralisation in the Telfer Mine is integrated with the regional- and orogenic-scale tectonic development of the Paterson Province.

TELFER (Au/+Cu) DEPOSIT

DEPOSIT STYLE: Epigenetic mesothermal lode-gold mineralisation hosted as stratabound quartz-carbonate reefs within a doubly-plunging domal antiform. (Elements: Au-Cu-As-Bi-Co±Ni±Pb±Mo)

AGE: Late-Proterozoic (≈650-600Ma)

LOCATION: 21°44'S 122°12'E

RESOURCE: (in situ) - 129Mt @ 1.24g/t
(Source: Sexton, 1994) - 5.1Moz (159t) Au
- Cu reserve undefined

DISCOVERED BY: Day Dawn Minerals / Newmont Pty. Ltd. (now owned and operated by Newcrest Mining Pty. Ltd.)

GEOLOGIC LOCATION : Paterson Province, Paterson Orogen northwest Western Australia

GEOGRAPHIC LOCATION : SW fringe of the Great Sandy Desert, W.A.

CURRENT PRODUCTION: ≈ 400,000oz. per annum

DEPOSIT HISTORY: - Outcropping stratiform gossans within the Telfer Dome were discovered and sampled in 1972 (named the Middle Vale Reef (MVR) and E-Reefs).
- During 1972 - 1977 reserve drilling confirmed a shallow ore-body of ≈ 1 Moz.
- Mine construction commenced in 1975.
- In the late 1980s Telfer becomes one of Australia's largest gold producers.
- Deep drilling commenced in Main Dome in 1992 and further stratabound reef mineralisation was discovered in the underlying Malu Quartzite and Isdell Formations

GENETIC MODELS: - 1979, 1982 - Syngenetic exhalative genesis; accumulation of dense metalliferous brine in small pools (Tyrwhitt, 1979; Turner, 1982)
- 1985 - MVR was an evaporite horizon that was replaced by quartz-sulphide (Royle, 1985)
- 1987 - Epigenetic origin; structurally controlled mineralisation; geochemical/fluid inclusion studies suggest that the ore-fluid was derived from the mixing of a magmatic and connate/formational fluid; most elemnts derived from magmas (Goellnicht, 1987)
- 1989,1994 - Convective model; Regional granites convectively circulate the ore-fluid through the host rocks; major elements derived from the sedimentary sequence (Hall & Berry, 1989;Rowins, 1994)

THESIS OVERVIEW

This thesis presents the results of a study of the structural geology of the Telfer Au (+Cu) deposit and surrounding region. The major aim of the project was to determine the timing and structural controls on mineralisation in the Telfer deposit in order to produce conceptual models that could be used for further mineral exploration in the Telfer region and Paterson Province. This aim was approached through an integrated study of four key structural problems in the Telfer region/Paterson Province and these have been written in this study as individual sections whose content is summarised below;

SECTION A - Details a structural analysis of middle- to late-Proterozoic metasedimentary rocks in both the Telfer Gold Mine and surrounding region, the NE Paterson Province. A deformation history involving multiple successions of sub-vertical and sub-horizontal foliation development has been observed, and is similar to modern examples of gravity-collapse orogenesis. This has established a regional tectonic history against which specific structural features in the Telfer mine and region can be related.

SECTION B - Presents the results of a further structural analysis of similar-aged rocks (to those studied in Section A) that lie approximately 40-50kms south of Telfer but are separated by extensive Permian fluvioglacial sedimentary cover. An identical deformation sequence has been observed in this southern sequence of rocks. However, they contain much better evidence for the first two deformation events because of the presence of schistose units. The first of these in particular could only be inferred in the Telfer region. Additionally, changes in the vergence directions for sub-horizontal deformations between both areas have been observed. These changes suggest that the orogen core was initially located well to the northeast but migrated during the middle and later stages of orogenesis southwestwards. They suggest that the Paterson Province may have lain in an ensialic back-arc tectonic setting.

SECTION C - Presents the results of paragenetic and structural timing of mineralisation (both micro- and meso-/macroscale) in the Telfer Au-Cu deposit. The discovery and development of new mineralised bodies in this ore-system has allowed access to unweathered mineralisation that was unavailable for previous studies. Structural timing of mineralisation in the Telfer deposit occurred synchronous with two weak deformation events, D_3 and D_4, which post-date the major deformation, D_2 that produced the large-scale host-structure. The hydrothermal veining/alteration paragenesis indicates that the ore-fluid homogenised and equilibrated with the regionally metamorphosing rock sequence prior to entering the Telfer deposit, suggesting that regional metamorphic fluids may have played a greater role, than was previously considered, in mineralisation genesis.

SECTION D - This section presents the results of a geometric analysis of the Telfer dome and the control that the two deformations associated with mineralisation, D_3 and D_4, identified in Section C have had on the emplacement of mineralisation into this host-structure. These two events produced distinct ore-body geometries within the Telfer Mine. Additionally, pre-D_2 folding controlled the subsequent geometry of the D_2 Telfer dome making it more conducive during later mineralising deformations to gaping in discrete locations. Much of the data for this section was gained from mapping of the Telfer Dome (Map 1).

SECTION E - This section applies models derived from Sections C & D to the NE Paterson Province in an attempt to provide a guide for future exploration in this area. The results of regional mapping (Maps 2 & 2B), which incorporates newly acquired geophysical data, assist the identification of late mineralising deformation effects across the region. The models integrate the key structural parameters identified in the Telfer region and conceptually extend and apply these to the geometrical relationships present within the larger-scale region.

ACKNOWLEDGMENTS

Firstly, I would like to gratefully acknowledge ***Newcrest Mining Pty. Ltd.*** for their financial and logistical support of the project, and in particular for allowing me access to data generated from the Main Dome Deep Drilling Project (which commenced in 1992) and high-resolution aeromagnetic data collected across the region in 1993. I would like to thank ***Don Hall*** (Chief Geologist) and ***Ray McCleod*** (Senior Exploration Geologist) who initiated the project in 1992, and to ***Andrew Richards*** (Chief Geologist), ***Peter Russell*** (Senior Exploration Geologist) and ***Colin McMillan*** (Senior Mine Geologist) who continued to support it from 1993 to completion.

Grateful acknowledgment must also go to the numerous staff at Telfer who assisted me whilst on site, including: ***Cameron Switzer, Myles Johnson, Jan DeVisser, John Foley, Sean Helm, Colin Moorhead, Julia Gleeson, Anna Timmins, Neil Simmons, Steve Reeves, Ingvar Kirchner, Campbell Mackey, Alex Eaves, Karen Johnson, Suzy Webb, Mike Sexton, Anne Woodgate,*** and ***Rob Cameron***. I would like to especially thank ***Hardy Cierlitza*** and ***Boris Tavcar*** for their boundless help and youthful enthusiasm with all things Telfer. Thanks are also owing to members of the Geological Survey of Western Australia for their advice and guidance on geological localities in the Rudall River National Park; including ***Ian Williams***, ***Leon Bagas*** and ***Arthur Hickman***.

Secondly I would like to extend grateful thanks to the "other half"; the staff and fellow students in the Department of Earth Sciences at James Cook University. In particular I gratefully thank my supervisor, ***Dr. Tim H. Bell*** (*alias* Reactivation), for his enthusiasm, guidance and help with the project. Additionally, I express grateful appreciation and thanks to my two "pseudo" supervisors (and close friends), ***Dr. Brett Davis*** (*alias* "Trumpet") and ***Dr. Ken Hickey*** (*alias* "MacTech", "Groundhog", "Kenny M-Hoggy",), and their respective partners ***Pamela*** and ***Frankie***, for their extensive support, guidance and party's. Thanks for the memories!! Brett in particular is thanked for all his valuable advice on diagram preparation, stylistics and proof reading. Ken is thanked for his incisive and encyclopaedic assistance with all matters (including, at times, structural geology).

Heartfelt appreciation is extended to ***Ms. Kaye Lake*** who supported me in the dying months of the thesis. Thankyou so very much Kaye for all your love and support. Thanks also to my sister, ***Alison***, who supported me whilst I was in Perth.

Lastly, I extend my sincerest thanks and appreciation to my parents, ***John*** and ***Monica***, for all their love, support and financial assistance throughout my life, and particularly this PhD. Thankyou for all your encouragement and guidance at times when I needed it; without your support this thesis would never have been completed.

SECTION A

Multiple Orthogonal Overprinting Deformation Events in the Telfer Region, W.A.: Preservation of a Complex Tectonic History in Weakly Deformed Rocks.

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ABSTRACT

Structural analysis of mesoscale folds and axial plane foliations indicates that multiple phases of orthogonally overprinting (ie. vertical and horizontal cleavage forming) deformation are preserved in low-grade meta-sedimentary rocks in the Telfer region. Horizontal compression, recorded as three generations of vertical deformation structures (D_{EF}, D_2 & D_4), was successively punctuated by horizontal fold/cleavage forming events (D_1 & D_3). A final event, D_5, encompasses two differently oriented phases of crustal compression, from that occurring previously.

Horizontal deformation events (D_1 & D_3) produced recumbent fold structures, the development of which involved a vertical shortening component. The cyclic change from horizontal compression to vertical shortening can be explained by a gravity collapse tectonic model with gravitational instability produced by uplift/thickening of a central plateau. Periodic collapse of this plateau would cause lateral spreading of rock material towards the margins of the orogen. The Telfer region appears to have lain on the margin of such an orogen.

Mesoscale fold and foliation development was extremely heterogeneous for deformation events that post-date the major folding phase, D_2. This partly resulted from folds in thick quartzite units producing vertical competent "struts" in the sequence that tended to prevent refolding. The heterogeneity was also due to partitioning of later tectonic stress into earlier structures causing them to be reactivated and re-used.

INTRODUCTION

Tectonic sequences involving a succession of vertical and horizontal deformations have been described for a number of orogenic belts around the world (*e.g.* Choukrane & Séguret, 1973; Meneilly, 1983; Mitra et al., 1984; Platt et al., 1983; Johnson, 1990a, 1992; Frotzheim, 1992; Jones, 1994). These sequences suggest that horizontal crustal shortening (vertical deformation) during compressional orogenesis is regularly punctuated by periods of vertical shortening or horizontal shearing. This appears to be related to the specific orogenic process of gravitational collapse (Bucher, 1956; Elliot, 1976; Ramberg, 1981; Bell & Johnson, 1989).

Multiple phases of orthogonal deformation have previously been recognised in higher metamorphic grade terranes, particularly where porphyroblasts have grown and preserve successive generations of foliations from destruction in the rock matrix during subsequent deformation (Meneilly, 1983; Platt et al., 1983; Johnson, 1990b; Bell & Hayward, 1991). Similar histories have also been recorded in strongly deformed lower grade rocks where foliations develop with greatest intensity against vein margins, and are subsequently preserved in strain shadows around the same veins during younger orthogonal overprinting events (*e.g.* Aerden, 1991,1993). In thin-skinned thrust belts, upright folds rotated and overprinted by thrusting deformation (Mitra et al., 1984; Mitra & Yonkee, 1985) also preserve evidence for episodes of orthogonal deformation. However, such histories are less commonly described for fold belts where axial plane cleavages are only weakly developed.

The Telfer region, in northern Western Australia, contains middle Proterozoic sedimentary rocks that were affected by convergent orogenesis during the late Proterozoic (Myers, 1990a,b). Previous structural analyses of these rocks in the Telfer region, host to the giant Telfer Au-Cu deposit, identified three potential deformation events (Chin et al., 1982; van Dijk, 1986; Goellnicht et al., 1989). However, these studies were limited to specific areas and did not integrate local, regional and orogen-scale observations. Superficially the Telfer region appears to have been affected by only one or two deformation events, and structural analysis is complicated by the poor exposure and highly weathered character of the Proterozoic sequence. Nonetheless, an analysis of mesoscale fold and foliation overprinting relationships encountered during systematic mapping of the Telfer mine and region has revealed a much more protracted deformation history and this is the subject of this section.

REGIONAL GEOLOGY AND PREVIOUS INVESTIGATIONS

The northern exposed section of the Paterson Orogen (Williams & Myers, 1990), the Paterson Province (Blockley & de la Hunty, 1975), comprises an early to middle Proterozoic gneiss complex, the Rudall Metamorphic Complex, unconformably overlain by middle to late Proterozoic sedimentary rocks of the Yeneena Basin (Fig. 1). Upper units of the Yeneena Basin, encompassing the north-eastern zone of the province (Fig. 1) are exposed in the Telfer region. The Yeneena Group was deformed and weakly metamorphosed during the late Proterozoic (Pan African) Paterson Orogeny (≈ 640-600Ma - see Section B for discussion of chronology) resulting in strong NW-SE striking folding, which in the Telfer region created a dome and basin structural setting north of the Karakutikati Range (Fig. 1). Late syn- and post-tectonic granitoid suites, as well as mafic intrusions, intruded the Yeneena Group in the Telfer region (Goellnicht et al., 1991).

Initial mapping of the Telfer region by the Geological Survey of Western Australia (GSWA) identified two folding events (Chin et al., 1982; Table 1), the first (D_1) comprising regional NW-SE trending folds, which were locally cross-folded by a second high angle deformation (D_2). Subsequent microstructural studies of rocks from the Telfer Mine identified two overprinting cleavages (van Dijk, 1986; Goellnicht, 1987; Goellnicht et al., 1989). However, whereas van Dijk (1986) interpreted these as two separate deformation events (Table 1), Goellnicht et al. (1989) considered both cleavages to be produced by one deformation (regional folding). Both authors interpreted S_2 as axial plane to the domes. Goellnicht et al. (1989) also identified a weak third fabric, comprising coarse muscovite growth, which postdated S_2 and that potentially represented the second deformation observed by Chin et al. (1982). Van

Dijk (1986) also inferred post-regional folding deformation to explain breccia veining and hydrofracturing in the Telfer ore-bodies.

Deformation	Chin et al. (1982)	van Dijk (1986)	Goellnicht et al., (1989)
D_1	NW-SE trending meso- to macroscale folding (domal) with an associated axial plane cleavage	A weakly spaced crenulation cleavage (S_1) lying at low angle to bedding.	Domal folding with two associated cleavages: S_1 (low angle to S_0) and S_2 (axial plane to folding).
D_2	Weak, high angle, cross-folding of regional folds with an associated coarsely spaced crenulation cleavage.	S_2 cleavage at a mod. to high angle to bedding, and axial plane to domal (regional) folding.	Mica overgrowth of the S_2 axial plane cleavage defining a weak overprinting cleavage
POST - D2		Breccia veining and fracturing.	

Table 1 : Summary of previous structural analyses and deformation histories proposed for the Telfer Region. Deformation notation is that of the respective authors.

DEFORMATION SEQUENCE

In the following section the term "deformation event" is used to signify a group of structures that are indicative of specific crustal kinematic conditions, and which can be compared with other different groups, into a relative chronological sequence on the basis of overprinting (*e.g.* Turner & Weiss, 1963; Hobbs et al., 1976). This is in contrast to an "orogeny" which describes a protracted period of crustal deformation that may comprise a number of individual deformation events. Structural terminology used to describe the following structures is after Bell & Duncan (1978), and summary stereographic data for each deformation event are presented in Appendix 1.

Deformation Event D_1

D_1 produced recumbent monoclinal folds ranging from centimetre (Fig. 2) to metre-scale (*e.g.* fig. 4 in Section D) with shallow SW dipping axial planes. The monoclines have short NE limbs, which are rarely overturned, open hinges and attenuated long (SW) limbs, the latter demonstrating a movement sense consistent with that observed on the axial plane cleavage (see further). A shallow SW dipping coarsely spaced cleavage (S_1), at low angle to bedding, lies axial plane to the monoclines (Fig. 2). Although S_1 is rarely sub-parallel to bedding, it is more commonly oblique, and generally ranges between 3° and 15° to S_0. S_1 varies from a weakly differentiated crenulation to a discontinuous/continuous pervasive slaty cleavage defined by recrystallised sericite and muscovite grains (Fig. 3). It is commonly coarsely spaced in those units where it overprints/crenulates a pre-existing bedding-parallel fissility. This fissility, marked by coarse muscovite and sericite grains, is a mimetic enhancement (Spry, 1969; Durney & Kisch, 1994) of bedding surfaces formed prior to D_1, rather than a tectonic feature (*e.g.*Holst, 1985), and is spectacularly developed locally along relict sedimentary cross-beds.

D_1 folds and S_1 cleavage verge (A. Bell, 1981) east in most locations throughout the Telfer region NW of the Karrakutikati Range. However, south-west of this range sporadic D_1 folds in the Isdell Formation verge west/south-west suggesting either a regional scale F_1^0 hinge in the vicinity of the Karrakutikati Range, or alternatively, coaxial deformation during D_1. Fold axes (F_1^0) and intersection lineations (L_1^0) lineations are re-folded during subsequent regional folding. However, D_1 folds preserved in sub-areas of horizontal bedding exhibit shallow-plunging or sub-horizontal F_1^0 axes that trend WNW to NW (310°,315°). A rarely observed weak mineral stretching lineation (L_1^1), defined by mineral streaking on S_1, is approximately perpendicular to F_1^0 axes. Differentiated crenulation asymmetry (*e.g.* Bell & Johnson, 1992) suggests that S_1 had a top-to-the-SW shear sense along the cleavage plane (Figs 4 & 5), which produced an extensional geometry (*e.g.* Platt & Vissers, 1980) consistent with the attenuation of the long (SW) limbs of D_1 monoclines (Fig. 4). Additionally, mesoscale normal faults (see fig. 10 - Section D) and tension veins, parallel to S_1 and pre-dating D_2, also suggest SW directed movement during D_1. Thus D_1 in the Telfer region appears to have involved SW directed sub-horizontal thrusting (shear) coupled with vertical shortening deformation.

Intrafolial Isoclinal Folding

Rare examples of isoclinal folding (Figs 6, 7A & B) developed in finely laminated units. These folds are generally symmetric and show parasitic folds along either limb (Figs 7B & 8) suggesting that they initially formed through coaxial shortening rather than through non-coaxial layer-parallel shearing (*e.g.* Sanderson, 1982). In some examples, the symmetric isoclinal folds terminate against enveloping laminae (Fig. 8) indicating post-formational disruption and potential rotation of the folds during a subsequent phase of layer-parallel shearing.

One mesoscale example of these isoclinal folds (Fig. 7A), located on the NE limb of Trotmans Dome (Fig. 9A; see fig. 1 & 5 - Section E, or Maps 2 & 2B for location of Trotmans Dome) has a steep axial plane that strikes perpendicular to the D_2 axial trend (Fig. 9A). This isoclinal fold was refolded by D_2, producing steeply plunging F_2 folds (Fig. 9A). It is not clear whether the steep plunge of D_2 axes is due to a pre-existing vertical axial plane orientation of the isoclinal fold, or is the result of post-D_2 refolding. The presence of high-angle D_5 cross-folding in the outcrop suggests the latter, and thus it is inferred that the isocline was recumbent prior to D_2. Consequently, it may be a D_1 fold, although it has a different style to the more common D_1 monoclinal geometry. An alternative explanation for the isocline is that it formed prior to D_1 as an upright fold that was subsequently rotated into a recumbent orientation by D_1, prior to refolding by D_2 (Fig. 9B).

Deformation Event D_2

D_2 constituted the major period of deformation that formed during the Paterson Orogeny and produced upright, open to tight, NW-SE trending micro- to macroscale folds, and macroscale domes, with an associated regionally penetrative axial plane cleavage (S_2). S_2 varies from a coarsely spaced weakly differentiated crenulation cleavage to a pervasive slaty cleavage in shale-mudstone units. In quartzite and crystalline carbonate units prominent fractures that formed along weathered traces of S_2 mark the D_2 axial plane. The development of S_2 was broadly synchronous with metamorphic matrix recrystallisation, and locally, a strong quartz grain elongation developed. In the Telfer deposit, S_2 is commonly indicated by small elongate quartz-albite (± pyrite) metasomatic aggregates that have grown along the cleavage.

Refolding of S_1 cleavage gives the relationships shown in Fig. 10, with both S_1 and S_2 having the same vergence on the south-western limbs of D_2 folds. A rare mineral stretching lineation (L_2^2), defined by a weak streaking of fine micas on the S_2 cleavage plane, has a sub-vertical pitch suggesting mainly vertical movement during D_2. Late reactivation (Bell, 1986; Aerdan, 1993) of bedding locally rotated S_2 towards parallelism with S_0, as well as crenulating it (Fig. 10). D_2 axial planes exhibit strike variations across the region (orogenic curvature), in addition to local anastomosing of S_2 which has arisen due to inhomogeneous shortening as well as refolding by subsequent deformation events.

Deformation Event D_3

The third deformation, D_3, was a weak, heterogeneously developed mesoscale folding event that formed asymmetric recumbent folds and crenulations of S_0 and S_2 (Figs 11,12 & 13). D_3 folds have sub-horizontal axial planes (S_3) with a consistent shallow NW to N dip and fold axes (F_3^0 , F_3^2) that plunge shallowly N to NNE. A penetrative axial plane cleavage is rarely developed, forming only in association with tight D_3 folds in finely fissile units (Fig. 13) and as discrete crenulations (Fig. 12; see also fig. 22 - Section C). Where present, S_3 is a coarsely spaced, differentiated crenulation cleavage or, less commonly, parallel fracture sets with associated en-echelon veining (*e.g.* Big Tree Costeans - see Maps 2 & 2B for location).

No macroscale D_3 folds were observed. However, refolding of D_2 folds in the Telfer Dome, resulting in local tightening and rotation of hinges, is interpreted to be the result of horizontal overthrusting deformation during D_3 (Section D). All D_3 folds verged NE with the exception of those in the Big Tree Costeans that verge to the SW. Additionally, differentiated crenulation asymmetry indicates a top-to-the-NE shear sense on S_3, suggesting that D_3 had a consistent NE transport direction in the Telfer region. The local vergence change may potentially arise from local coaxiality or local shear reversals due to shifting patterns in deformation partitioning (*e.g.* Bell & Johnson, 1989,1992) during D_3.

Deformation Event D_4

D_4 produced mesoscale cross-folding of D_2 folds, particularly in finer-bedded units of the Isdell Formation south of the Karakutikati Range (Fig. 1; see also Map 2B). These folds are upright open structures (Fig. 14), of small (2-4m) amplitude and limited axial plane extent (in both plan and section), and have a consistent orientation throughout the Karakutikati Range area. D_4 axial planes are sub-vertical or inclined towards the NW and strike 345-355°, which is 20-30° clockwise from the D_2 axial trend. In one example a broadly spaced differentiated crenulation cleavage (S_4) lies axial plane to D_4 folds (Fig. 15). No mineral stretching lineation (L_4^4) was observed, but the axial plane orientation suggests that D_4 comprised a weaker period of horizontal shortening that was oriented approximately WSW-ENE (255,260° - 075,080°). The limited occurrence of D_4 folds throughout the Telfer region probably reflects the similarity in the primary orientations of S_2 and S_4 preventing refolding of D_2 folds (*e.g.* Ghosh & Ramberg, 1968; Treagus, 1983; Ghosh et al., 1992). Overprinting relationships between D_3 and D_4 were assisted by the observation of structures produced by correlateable events further south of Telfer (Section B).

Deformation Event D_5

The fifth deformation, D_5, is characterised by small scale asymmetric monoclinal kink folds (Fig. 16A) and crenulation planes (Fig. 16B), which are also well developed in finely laminated units of the Isdell Formation. These kinks, which range from centimetre- to metre-scale, are discrete structures that commonly terminate at either end of the axial plane in small rounded monoclinal flexures (Fig. 16A). Kink axial planes and crenulations dip steeply (80°,85°) toward the WNW, NW (290°) although there is a marked variation in trend in some areas (see further). F_5^0 and F_5^2

axes are generally steeply plunging in the prevalent bedding plane. D_5 axial planes are marked by coarse dissolution seams and exhibit an apparent dextral shear sense. No conjugate kink sets were observed.

North of the Karakutikati Range, in coarser-bedded units, similar crenulations of both S_0 and S_2 occur, albeit rarely, in addition to sporadic examples of NNE-SSW trending vertical discontinuous cleavage. High angle cross-folding of D_2 folds occurs in the Malu Hills/17 Mile Hill localities (see Map 2 for location; *cf.* Chin et, al, 1982) with a similar axial plane orientation to other D_5 structures. The axial plane orientation of all these structures indicates D_5 involved weak compression oriented WNW-ESE (≈ 290-110°). The absence of conjugate kink sets suggests that D_5 compression was slightly oblique to the regional structural grain (Gay & Weiss, 1974; Reches & Johnson, 1976).

Late D_5 Deformation (D_{5b})

A group of distinctly oriented structures suggests that a separate NNW-SSE to N-S compressional event occurred. These include extension veins, normal faults and dolerite dikes, all of which trend N-S and crosscut D_2 folds. Small tensional fractures, which formed in the country rock along dike margins, lie parallel to the trend of the dikes. These suggest that the dikes formed through extensional opening perpendicular to σ_1, produced by an approximately N-S directed compression. Additionally, D_5-style kinks exhibit two distinct orientational groups (Fig. 17), one of which reflects a period of NNW-SSE to N-S compression. Kinks in both groups each have a similar crenulation/kink asymmetry and therefore are not members of a conjugate set.

Dissolution along the axial planes of both kink sets suggests that each formed through shortening perpendicular to the axial plane implying two separate compressive episodes (*ie.* NW-SE compression - D_5 described in previous section, and NNW-SSE to N-S compression - D_{5b} described in this section). As the overprinting relationships between these two episodes are unclear, the N-S compressional phase is designated D_{5b}, implying that it is stylistically similar to D_5, but represents different kinematic conditions. The development of similar structures (*ie.* kinks/crenulations) by these two compressive phases suggests that they may represent a progressive re-orientation of tectonic compression from 290-110° (WNW-ESE; D_5) through to NNW-SSE / N-S (D_{5b}) late in the deformation sequence.

INTERPRETATION

Structural analysis of mesoscale folds and foliations indicates that multiple fold-forming deformation events occurred across the Telfer Region (Fig. 18). These are preserved as distinct minor folds, with associated axial plane foliations, that record a complex kinematic history comprising horizontal shortening of the belt punctuated by periods of horizontal deformation suggestive of vertical crustal shortening.

Kinematic History of the Telfer Region

D_1 Monocline Formation; Implications for SW Directed Thrusting and Early Orogenic Shortening.

D_1 produced a penetrative sub-horizontal foliation (S_1) the geometry of which against bedding has ambiguous implications for F_1 monocline formation. Fold and cleavage vergence are consistent with formation during sub-horizontal thrusting (*e.g.* Elliot, 1976; Boyer & Elliot, 1982; Coward & Potts, 1983) towards the NE. However, S_1 cleavage has a SW directed shear sense, locally producing an extensional crenulation (Platt & Vissers, 1980) geometry along the attenuated long limbs of D_1 folds, whilst the short NE fold limbs represent low-strain zones in which bedding was not thinned. These fold-cleavage geometries are incompatible with formation during NE directed thrusting as the forelimbs (short NE limbs) should be overturned and thinned (Jamieson, 1992; Rowan & Kligfield, 1992; Bons, 1993), and cleavage (S_1) should exhibit a synthetic (*ie.* top-to-the-NE) movement sense in line with tectonic transport. However, the fold geometry is also difficult to reconcile with a SW directed thrusting regime that is suggested by the extensional shear sense on S_1.

One explanation for the conflicting genesis of D_1 folds is that they may comprise an earlier fold phase that was overprinted by sub-horizontal shearing/thrusting deformation (*e.g.* Sanderson, 1979; Rowan & Kligfield, 1992), with two possible scenarios capable of explaining the observed geometric relationships (Fig. 19). In the first, upright folds with an axial plane foliation are overprinted by a NE directed thrusting/shearing deformation (with or without vertical shortening) that rotates the folds and cleavage towards parallelism with bedding (Fig. 19A). Such rotation would also cause reactivation (Bell, 1986) of the axial plane cleavage (Fig. 19) producing a SW directed shear sense. The alternative path comprises the same early-formed folds being overprinted by SW directed thrusting/shearing with concomitant cleavage formation, the latter exhibiting a synthetic shear sense (*ie.* top-to-the-SW) consistent with the bulk tectonic movement (Fig. 19B).

Both models discussed above (Fig. 19) can explain the geometry of D_1 folds. However, they differ in the relative timing of cleavage formation and direction of thrusting/shearing. Additionally, each model implies the presence of a pre-D_1 fold phase. A similar development history is inferred for intrafolial isoclinal folds observed in this study. Such folds may represent the early fold phase on which D_1 monoclines could have formed during SW directed shearing. Structural analysis of Lower Yeneena Group rocks south of Telfer indicates that similar D_1-style monoclines, which have a SW shear sense on the axial plane cleavage (Section B), overprint small upright folds. These overprinting relationships were not unequivocal in the Telfer region. However, the presence of a pre-D_1 fold phase is interpreted to explain the geometry of both the isoclinal folds (*e.g.* Fig. 9) and D_1 folds. Consequently, it is interpreted that D_1 was a thrusting/shearing event, potentially driven by vertical shortening, with a SW directed tectonic transport consistent with the shear sense on S_1 (Fig. 19B).

The main implication of the interpretation discussed above is that upright and approximately symmetrical folds existed prior to D_1. Such folds could have been the product of layer-parallel shear during the initial stages of D_1 prior to being rotated under progressive deformation (*e.g.* Ghosh, 1966; Manz & Wickham, 1978;

Sanderson, 1979; Ramsay et al., 1983). Alternatively, they may have formed during pre-D_1 layer-parallel shortening, that was precursor to gravity-induced orogenic collapse (Bucher, 1956; Elliot, 1976; Ramberg, 1981) driving sub-horizontal D_1 deformation. The overprinting relationships further south of the Telfer region (Section B) support the latter, and hence in this study this *inferred* deformation is assigned the notation D_{EF} (early folding), signifying it's occurrence prior to D_1.

D_2 Deformation - Bulk Inhomogeneous Shortening and Dome Formation

Previous structural studies have inferred wrench- or transpressive tectonism for the Paterson Province based on the presence of regional domes (Harris, 1985; Bogacz, 1990; Baxter, 1991; see Sections D & E for discussion). However, the results of this study indicate that the domal axial plane cleavage (S_2) exhibits a sub-vertical stretching lineation (L_2^2) suggesting vertical movement during D_2. This is characteristic of inhomogeneous tectonic shortening (*e.g.* Ramsay, 1967; T. Bell, 1981), rather than transpressive or wrench environments where oblique movement would be expected (Woodcock & Schubert, 1994).

Late-D_2 Bedding Reactivation. - S_2 cleavage is rarely crenulated along bedding planes with an antithetic shear sense (Fig. 10), similar to that for flexural-slip deformation. Rather than signifying a separate deformation event, this crenulation is consistent with reactivation (Bell, 1986; Aerdan, 1993) of bedding during D_2. Reactivation occurs when bulk fold movement, initially accommodated by synthetic shearing on axial plane foliations, is locally partitioned into antithetic shear on the surface being folded (Bell, 1986). This will occur late in the folding event. However, it may also happen during subsequent and similarly oriented folding episodes that reuse the earlier fold (Davis & Forde, 1994).

D_3 Deformation - Implications for Vertical Crustal Shortening.

The crenulation/fold asymmetry of D_3 suggests that it involved NE directed tectonic transport related to a period of vertical crustal shortening. Evidence for this comes from the near-horizontal attitude of the axial surface, and vergence reversals suggestive of locally coaxial deformation, a feature characteristic of vertical inhomogeneous shortening (T. Bell, 1981; Bell & Johnson, 1989, 1992). Another indicator for vertical shortening is the development of sporadic crenulation planes, which exhibit a top-to-the-NE shear sense across a markedly inclined axial plane (Fig. 20). Potential development paths, which account for both the inclination and shear sense on the axial plane, are illustrated in Fig. 20. However, the kink-style of these folds, which occur in sub-vertically dipping and finely laminated bedding, suggests that shortening was slightly oblique, but near parallel, to the layers (*ie.* vertical; *e.g.* Gay & Weiss, 1974; Reches & Johnson, 1976; Ramsay & Huber, 1987, p. 428).

Structures of a similar style to D_3 folds in the Telfer region have been interpreted in other orogenic belts to result from ductile thrusting (Meneilly, 1983; Meneilly & Storey, 1986). However, no regionally penetrative thrust structures are observed in the Telfer region, and D_3 structures are only sporadically developed and thus not typical of extreme shearing deformation. In contrast, vertical shortening deformation could produce sub-horizontal cleavage and discrete structures, the latter occurring where pre-existing foliations were suitably oriented for shortening (*ie.* vertical). Such vertical shortening could also produce a shallowly oriented shearing component (T. Bell, 1981) producing horizontal differential movement that in the Telfer region was mainly NE directed.

Post - D_3 deformation - Progressive Re-orientation of Vertical Compression

Both D_4 and D_5 formed vertical structures indicative of horizontal crustal shortening. For D_4 this occurred at a low angle ($\approx$ 25-30°) to the D_2 shortening direction whilst D_5 and D_{5b} formed structures at higher angles (NW-SE and N-S respectively) indicating a progressive clockwise rotation of compressive forces. This suggests that the Telfer region would have become increasingly transpressive during the deformation history, characterised by left-lateral movements.

DISCUSSION

The Preservation of Complex Tectonic Histories in Multiply Deformed Rocks

Complex deformation histories resulting from multiple orthogonal overprinting of successive deformation events have been documented in many terranes, particularly modern convergent plate tectonic settings such as accretionary wedges/prisms (Platt, 1986; Sample & Moore, 1987) and collisional orogens/mountain belts (Dahlen & Suppe, 1988; Johnson, 1990a, 1992; Bell & Johnson, 1989, 1992; Aerden, 1994). Such sequences are recognisable in these settings because of their preservation within zones that have been protected from subsequent deformation, metamorphism and alteration. In higher-metamorphic grade terranes these include porphyroblasts, whose inclusion trails may preserve evidence for multiple foliation development (Meneilly & Storey, 1986; Johnson, 1990a,b; Bell & Hayward, 1991; Jones, 1994), and in some cases the original foliation orientation (Hayward, 1990; Bell & Johnson, 1992; Johnson, 1992). In lower grade rocks, quartz veining may preserve axial plane foliations (*e.g.* Forde, 1991), and record the strain history of the rocks through progressive quartz-fibre growth in strain shadows (Durney & Ramsay, 1973; Ramsay & Huber, 1983, p. 265; Kirkwood, 1995).

Similarly, boudinaged veins can preserve earlier formed matrix foliations in pressure shadows lying adjacent to boudin necks (Aerden , 1991; Davis & Forde, 1994). In the absence of rigid elements, deformation partitioning (*e.g.* T. Bell, 1981) within the rock matrix can also preserve evidence for the primary orientation of foliations in zones of shortening (coaxial deformation), whilst in adjacent non-coaxial shearing domains these earlier foliations get re-oriented (Bell et al., 1989). These examples all reflect the general tendency for relatively competent portions of rock, or structural elements, to preserve the foliation evidence for earlier deformations from subsequent destruction or re-orientation in the rock matrix. Previously, such features may have been rationalised by geologists as locally developed complexities against heterogeneities, rather than earlier developed and preserved structures indicative of a distinct phase of orogenic deformation.

Since the post-D_2 deformation intensity in the Telfer region was weak, fold and foliation evidence for successive orthogonal events has remained intact in the matrix. In very low-grade thrust belts, evidence for orthogonal deformation events can be preserved as initial upright symmetric folds (horizontal shortening) that have been subsequently rotated into a recumbent orientation and overprinted by a horizontal solution cleavage associated with thrusting deformation (Mitra et al., 1984; Mitra & Yonkee, 1985). In these terranes, thrusting appears to be the main deformation style, and has acted at a scale equal to or larger than folding. In contrast, the Telfer region is characterised by the development of regionally penetrative folding (vertical deformation) with minimal thrusting.

The absence of well developed thrust sheets, which could have accommodated much of the movement during sub-horizontal deformation, may be one reason why multiple successive orthogonal events have been preserved. In particular, vertical structures (*ie.* upright folds) would more likely have been preserved. Another reason may be that the limbs of upright D_2 folds in thick quartzite units acted as competent "struts" during subsequent deformations. This may have restricted the effects of further deformations that produced sub-horizontal axial planes. Younger deformations will tend to reactivate (Bell, 1986; Aerden, 1993) bedding between these "struts" rather than produce new folds or foliations, and consequently these events may have been very heterogeneously developed.

The Heterogeneity of Polyphase Deformation; Reactivation and Re-use of Regional Structures.

Heterogeneous structural development is common in multiply deformed terranes (*e.g.* Tobisch, 1967; Tobisch & Fiske, 1982; Davis, 1993; Davis & Forde, 1994). This is particularly so for fold phases that post-date the major period of deformation with one reason for this being the strong structural anisotropy produced by the main folding episode. In addition to preventing further strain (as discussed above), this anisotropy may cause subsequent tectonic strain to be accommodated dominantly by re-orientation and/or reactivation and re-use of pre-existing structures (*e.g.* Davis & Forde, 1994). This would be particularly true for deformation where there was a low angular difference between the primary compression direction for successive events. At Telfer, the low angular difference between D_2 and D_4 compression contributed to the lack of refolding of D_2 folds by this event (Ghosh & Ramberg, 1968). These folds would have been simply tightened and slightly re-oriented (Ghosh & Ramberg, 1968; Odonne & Vialon, 1987; Price & Cosgrove, 1990, p. 400; see Section D) thereby accommodating the D_4 strain. Since D_3 was weak, and did not grossly re-orient previous fabrics, the D_2 axial plane foliation (S_2) could be re-used during D_4 resulting in the development of a composite foliation (*e.g.* Tobisch & Fiske, 1982; Meneilly, 1983; Tobisch & Paterson, 1988).

Refolding, and subsequent development of new structures is most likely to occur where a high angular difference exists between two deformations. For example, it is interpreted that D_1 folds nucleated on a previous generation of upright folds (D_{EF}). These pre-D_1 folds locally produced a vertical bedding surface that was oriented at a high angle to the subsequent horizontal D_1 strain. Similarly, horizontal D_3 folding is best developed in steeply/vertically oriented bedding (Fig. 21), such as the NE limbs of D_2 folds. These limbs were in a forelimb position with respect to NE directed D_3 movement (*cf.* Ray, 1991; Alonso & Teixell, 1992). In contrast, no D_3 folds were observed on the SW limb of D_2 folds, reflecting the more favourable orientation of bedding in this limb for reactivation, rather than folding (Fig. 21). This preferential re-use of earlier structures according to their orientation relative to the prevailing deformation is another reason for the inherent heterogeneity of polyphase deformation.

Tectonic Implications - Gravity Collapse Orogenesis

Orthogonal overprinting deformation is suggestive of widely variable compressional directions during orogenesis of a rock sequence. Although horizontally directed shortening dominated in the Telfer region, consistent with convergent orogenesis, this was interspersed with repeated components of vertical crustal shortening. Such cyclic switching between horizontal compression and vertical crustal shortening in a convergent orogen is suggestive of gravity collapse tectonism (Bucher, 1956; Ramberg, 1981; Bell & Johnson, 1989, 1992; Aerden, 1994). The driving force for such tectonism is interpreted to be gravitational instability leading to collapse of the orogen core (Bucher, 1956; Price, 1973; Elliot, 1976; Ramberg, 1981). A similar process has been inferred for accretionary/orogenic wedges lying at the margins of an orogenic belt (Platt, 1986; Dahlen & Suppe, 1988).

In both mountain belts and accretionary wedges, collapse of an uplifted and thickened orogen core causes lateral spreading of material (horizontal deformation) towards the orogen margins (Bell & Johnson, 1989, 1992; Northrup, 1996). This is manifest as extensional deformation in the upper levels (Dewey, 1987; Aerden, 1994; Inger, 1994) and thrusting at lower levels (Bell & Johnson, 1992) and towards the margins (Price, 1973; Elliot, 1976; Northrup, 1996). The absence of thin-skinned thrusting in the Telfer region, generally indicative of processes in the extreme margins of an orogen (Platt, 1986; Northrup, 1996), coupled with locally penetrative cleavage development, suggests these rocks were at moderate depths during orogenesis (Elliot, 1976), and thus probably lay within an orogenic wedge adjacent to a thicker orogen core.

SIGNIFICANCE

Late deformations have been demonstrated in other terranes to be associated with, and control, epigenetic mineralisation (Kontak et al., 1990; Forde, 1991; Aerden, 1993, 1994; Wilkins, 1993; Forde & Bell, 1994). Their recognition at Telfer has enabled the timing and controls on gold-copper mineralisation within the Telfer deposit to be determined (Section C, Section D). The results of these models contrast with previous studies (*e.g.* Goellnicht 1987; Goellnicht et al., 1989) where the limited knowledge of the deformation sequence has meant that many aspects of the Telfer ore-system were considered only in terms of the most recognisable deformation, D_2. Consequently, it should be possible to develop more specific structural models to assist exploration for "blind" mineralisation occurrences. The results of this study suggest that structural analysis of other similar low-grade and highly weathered but mineralised terranes has the potential to identify orthogonal overprinting deformations, which in turn could prove beneficial in developing exploration strategies (*e.g.* Forde & Bell, 1994).

CONCLUSIONS

Structural analysis of mesoscale folds and axial plane foliations in the Telfer region has identified a tectonic history that consists of five deformation events that occurred during the late-Proterozoic Paterson Orogeny (≈ 640-600Ma). The deformation sequence comprises alternating vertical and horizontal fold/cleavage forming events. D_1 produced recumbent monoclinal folds, whose geometry is ambiguous and suggests that an earlier, but as yet unverified, fold phase was refolded by SW directed thrusting deformation. This inferred period of folding is termed D_{EF}. The strongest event , D_2, produced upright regionally penetrative NW-SE striking folds and axial plane cleavage that are the major structural elements of the Telfer region. D_3 was a period of horizontal deformation that produced folds with a horizontal axial plane, whilst D_4 comprised horizontal crustal shortening that was approximately 20-25° clockwise to that of D_2. The final event, D_5, comprised two periods of compression, each differently oriented and at a high angle to D_2 folding.

The presence of two sub-horizontal deformations, D_1 and D_3, in the sequence suggests that horizontal crustal shortening was interspersed by periods of vertical compression. This is evident from the development of folds and crenulations with sub-horizontal axial planes during D_1 and D_3. Vertical shortening in the tectonic sequence suggests that the Telfer region was affected by gravity collapse tectonism during the Paterson Orogeny. Such tectonism may have occurred when the orogen core collapsed as a result of gravitational instability causing lateral spreading of rock material to the margins of the orogen. The absence of thin-skinned thrusting in the Telfer region suggests that it lay at moderate crustal levels, probably marginal to a larger orogen. Additionally, late vertical deformations indicate an anti-clockwise re-orientation of horizontal compression over time, suggesting that the orogenic convergence changed from orthogonal to oblique, and that the Telfer region became transpressive.

One feature of the deformation sequence is the extreme heterogeneity in mesoscale fold and foliation development, particularly for post-D_2 events. This is interpreted to arise from the structural anisotropy produced by the major folding (D_2) phase. Additionally, the strain produced by subsequent deformations was accommodated through reactivation and re-use of earlier structures. Given that multiple phases of orogenic deformation may play a pivotal role in the development of some styles of epigenetic mineralisation, the recognition of such a teconic sequence in low-grade gold-bearing metamorphic terranes is an important first step in establishing a genetic model for deposit formation.

SECTION B

Structural Reconnaissance of the Lower Yeneena Group, Paterson Province, W.A.; Overprinting of an Ensialic Intracratonic Basin by Migrating late-Proterozoic Collisional Orogenesis and Resolution of Conflicting Tectonic Indicators.

SECTION B

Structural Reconnaissance of the Lower Yeneena Group, Paterson Province, W.A.; Overprinting of an Ensialic Intracratonic Basin by Migrating late-Proterozoic Collisional Orogenesis and Resolution of Conflicting Tectonic Indicators.

ABSTRACT

Structural reconnaissance of lower Yeneena Group sedimentary rocks, complements studies of the Telfer region (Section A) and confirms that a terrane-wide deformation sequence affected the Paterson Province during convergent orogenesis (Paterson Orogeny) in the late-Proterozoic (≈640-600Ma). This sequence comprised alternate horizontal and vertical-structure forming deformation events, indicating that horizontal compression was regularly punctuated by episodes of vertical crustal shortening. Vertical-structure forming episodes produced upright NW-SE trending folds at various scales, with the exception of later events whose primary compression rotated clockwise to become NW-SE trending. Horizontal-structure forming deformations produced distinctive recumbent folds and flat-lying cleavage. The regular alternation of shortening directions suggests that gravity collapse orogenesis occurred.

Vergence/shearing directions for the sub-horizontal deformations provide information on the gross horizontal movement of material during collapse phases, which in turn assists the identification of the position of the orogen core during orogenesis. These suggest that initially orogenesis commenced NE of the Paterson Province but migrated south-westward during the middle and latter stages of orogeny to encompass the province. This history provides an explanation for the conflicting tectonic indicators in the province by suggesting that the Paterson Province, and contained sedimentary rocks (Yeneena Group) initially formed in an ensialic back-arc/intracratonic setting that was subsequently overprinted by SW migrating collisional orogenesis. Migration of the orogen core craton-ward during late-Proterozoic orogenesis has implications for both the magmagenesis and metallogenesis of the Paterson Province.

INTRODUCTION

Within Australia, large Proterozoic mobile/linear belts (Davies & Windley, 1976), bordering stable Archean cratons (Fig. 1) exhibit long-lived episodic orogenic histories (*e.g.* Baer, 1983; Etheridge et al., 1987; Muhling, 1988; Wellman, 1988; Myers, 1990a,b; Tyler & Thorne, 1990; Clarke, 1991; Goscombe, 1992; Collins & Shaw, 1995) arising from repeated supercontinent aggregation and break-up (Bond et al., 1984; Gurnis, 1988; Hoffman, 1989). One such example, the Paterson Orogen (Williams & Myers, 1990; -previously the Paterson-Musgrave Structural Trend - Austin & Williams, 1978) spans the north-west and central regions of the Australian continent and records a long-lived Proterozoic orogenic history (Myers, 1990a, Clarke, 1991). Renewed orogenesis in the late-Proterozoic (≈700-600Ma - Paterson Orogeny) resulted in significant mineralisation in an exposed portion of this orogen, the Paterson Province in northwest Western Australia.

As much of the Paterson Orogen is now covered by Phanerozoic sedimentary basins, the Paterson Province represents only a very small marginal exposure of the large orogenic belt and thus the tectonic setting of this mineralised province is problematic. Although the middle-late Proterozoic sedimentary history is suggestive of an ensialic (intracratonic) setting (Turner, 1982; Goellnicht, 1992; Hickman et al., 1994), granitoid geochemistry, mineralisation and structural styles are all indicative of continent accretion or the development of an island/magmatic arc setting in the late-Proterozoic (Myers, 1990b; Goellnicht, et al., 1991). Additionally, structural analysis of the host sedimentary sequence in the NE areas of the province has indicated a sequence of orthogonally overprinting (*ie.* horizontal and vertical) deformation events suggestive of gravity collapse tectonism similar to that inferred in many modern collisional orogens (Section A).

Recent research suggests that successive foliation/foliation asymmetries produced by gravity collapse tectonism can be used to more accurately refine the position of a rock sequence within an orogenic belt (Bell & Johnson, 1989; 1992; Aerden, 1994; Northrup, 1996). Tectonic movement directions for horizontal (collapse stages) deformation events within such sequences provide information on the position of the orogen core, the locus of collapse (Bell & Johnson, 1989; Northrup, 1996). This study presents the results of a reconnaissance study of rocks in the SW area of the province which suggest that the Paterson Province lay marginal to a larger orogen, whose core migrated SW (inboard) during the middle and later stages of the Paterson Orogeny. Such a history may provide a resolution for conflicting tectonic indicators in the Paterson Province, and other deformed sedimentary basins marginal to large Proterozoic orogenic belts, by implying that collisional orogenesis may migrate inboard to deform and produce mineralisation in these basinal settings.

GEOLOGICAL SETTING AND PREVIOUS INVESTIGATIONS

The Paterson Province (Daniels & Horwitz, 1969; Blockley & de la Hunty, 1975), of northern Western Australia is an inlier of Proterozoic rocks lying on the SW margin of the NW-SE trending Paterson Orogen (Fig. 1). The province exposes two major lithotectonic units; the Rudall Metamorphic Complex which is unconformably overlain by the Yeneena Group (Fig. 2). The Rudall Metamorphic Complex comprises early-middle Proterozoic gneisses and meta-sediments that exhibit a complex tectonic history of crustal reworking culminating in orogenesis and continent-continent fusion at ≈1330Ma (Watrara Orogeny - Table 1; Clarke, 1991; Hickman et al., 1994).

The Yeneena Group comprises a mixed sequence of terrigenous and marine carbonate units (Table 2) that are divided into three geographic zones; the Western, Central and Northeastern zones (Fig. 2; Williams, 1990a). Deposition of the group occurred in fluvio-deltaic, shallow marine shelf and pelagic environments that progressively deepened north-east, in an interpreted continental margin (Hickman et al., 1994) or intracratonic (aulacogen/failed rift) setting (Turner, 1982). The Yeneena Group was multiply deformed during late-Proterozoic tectonism at ≈ 640-600Ma (Paterson Orogeny - Myers, 1990b; Table 1; see also Section A) that also resulted in some reactivation of basement (Rudall Metamorphic Complex) structures. This was accompanied by numerous syn- and post-tectonic granitoid and other igneous intrusions in the NE Paterson Province (Telfer district).

Lithotectonic Unit	Deformation	Characteristics
Rudall Complex	D_1	Penetrative layer parallel foliation; M1 ⇒ low pressure mid-amphibolite facies: crustal melting and granitoid intrusion
	D_2 (*W*)	SW and W directed thrusting/nappe folding with WNW trending axes; folds now overturned to SSW; M2 ⇒ med. pressure amphibolite facies (Barrovian)
	D_3* (*W*)	*upright NW trending folding*
Erosional unconformity; later tectonised		
Yeneena Group (and Rudall Complex)	D_3	W - NW directed recumbent folding with E-W/N-S trending fold axes; produced a layer-parallel foliation.
	D_4 (*P*)	Regional NW-SE trending folding throughout Yeneena Group with a penetrative axial-plane foliation; M4 ⇒ low greenschist facies
	D_5	Open recumbent crenulations suggesting NE directed stress release.
	D_6	NNE-SSW compression producing strain-slip cleavage axial plane to conjugate kink bands. Possibly related to granitoid intrusion.

Table 1: Summary of the deformation history proposed for the Paterson Province (including the basement Rudall Metamorphic Complex) from previous studies (after Hickman et al., 1994). Deformation notation is that of the cited authors. (W) is the Watrara Orogeny (= 1330Ma - Clarke, 1991; Hickman et al., 1994). (P) is the Paterson Orogeny (≈640-600Ma - see further). (*) denotes a deformation event identified only by Clarke (1991) in the basement rocks.

Unit	*Lithology*	*Max. Thickness (m)*
Kaliranu Beds	Silty dolomite, shale, sandstone	>350
Wilki Quartzite	Quartzite; minor shale, siltstone	1000
Puntapunta Fm.	Dolomite, calcarenite; minor clastics	2000
Telfer Formation	Sandstone, siltstone, shale	600-700
Malu Quartzite	Sandstone	0 - 1000
Isdell Formation	Dolomite, dolomitic shale; minor clastics	2000
Choorun Formation	Sandstone, shale, conglomerate; dolomite	0 - 2000
Yandanunyah Formation	Dolomite, siltstone, shale	80
Brownrigg Sandstone	Sandstone	350
Waroongunyah Formation	Dolomite, shale/siltstone, sandstone	250
Googhenama Formation	Sandstone, conglomerate	0 - 250
Broadhurst Formation	Graphitic shale, siltstone; dolomite	0 - 1000
Coolbro Sandstone	Sandstone, conglomerate, minor shale	4000

Table 2: Summary of the Yeneena Group stratigraphic sequence over the whole Yeneena Basin (from Williams, 1990a). See Fig. 2 for the geographic division/location of the various units.

LOWER YENEENA GROUP - PATERSON PROVINCE

Macrostructure

The rocks described herein comprise the basal units of the Central Zone of the Yeneena Group (Coolbro and Broadhurst Formations - Fig. 2; see also Table 2) that immediately overlie the Rudall Metamorphic Complex (Fig. 3). The contact between the Yeneena Group and Rudall Metamorphic Complex is erosional and marked by a boulder-cobble conglomerate that is locally strongly tectonised (Fig. 4). Regional-scale folding of the Yeneena Group is upright, NW-SE trending and formed during the major period of deformation in the Paterson Orogeny, D_3 (see further). Two areas, on either side of the Rudall Metamorphic Complex, were examined for mesoscale fold structures and associated axial plane foliations: the south-western margin and the Broadhurst Range (Fig. 3).

Folding along the south-western margin is open to tight, with long horizontal fold trains that are commonly overturned/inclined to the SW (Fig. 3). Numerous high-angle thrust faults disrupt the sequence and have interleaved basement and Yeneena Group rocks (Fig. 3). The relative intensity of faulting along this margin may reflect the surficial expression of larger, and deeper seated thrust planes, on which the basement and Yeneena Basin were transported south-westwards during the Paterson Orogeny (Chin et al., 1980; Clarke, 1991). In the Broadhurst Range, regional-scale anticlinal/synclinal folding over domed basement is upright or SW inclined, plunges SE, and is commonly faulted by high-angle thrusts along fold limbs (Fig. 3). Late N-S trending faults cross-cut the regional folds (Fig. 3).

Deformation Chronology

Reconnaissance structural analysis, conducted by vehicle traverses across the study areas, indicates six deformation events are preserved in the Lower Yeneena Group. Well developed datum structures are provided by regional (D_3) folds, enabling correlation between the two study areas. The structural terminology used is after Bell & Duncan (1978).

Pre - Regional Deformation (D_1, D_2)

The first deformation, D_1, is preserved as mesoscale intrafolial isoclinal folds (Fig. 5). These have axial planes parallel to the orogen trend and appear to have formed upright. However, all examples are now recumbent owing to a subsequent deformation, D_2. This deformation produced recumbent monoclines and rotated D_1 folds (Fig. 6), locally forming bedding parallel thrusts along D_1 fold limbs (Fig. 7). D_2, also produced a coarsely spaced sub-horizontal axial-plane cleavage (S_2) that overprints D_1 folds (Fig. 6). The shear sense on S_2 is top-to-the-SW, which produced an extensional geometry along the trailing limbs of D_1 folds consistent with the attenuation (Fig. 6) of bedding on the long-limbs of D_2 monoclines (Fig. 6). This suggests that D_2 involved horizontal SW directed movement. D_2 folds also have axes that lie sub-parallel to the regional fold/orogen trend (NW-SE), and early horizontal deformation in the Lower Yeneena Group rocks is consistent with observations made in other studies of bedding-parallel fabric development and strong deformation of basal conglomerates both prior to regional (D_3) folding (see Hickman et al., 1994).

The fore-limbs of D_1 folds were shortened by D_2 producing superposed forelimb folding (*e.g.* Ray, 1991; *cf.* Meneilly, 1983) whilst the trailing limbs were attenuated (Fig. 6), suggesting that D_2 involved a vertical crustal shortening component. This is illustrated by small rootless monoclines (Fig. 8), lying at a low-angle to bedding, that are strongly cleaved by S_2. These formed when rootless symmetric D_1 folds, probably produced by layer-parallel shortening (Fig. 9A), were rotated, flattened and overprinted by S_2 during D_2 (Fig. 9B). S_2 is especially well developed along the shorter (and more vertical) SW limb of the rotated D_1 folds (Fig. 9C) where vertical shortening of the limb would have occurred. In contrast, the trailing limb (NE limb of the D_1 fold) was attenuated with S_0 and S_2 sub-parallel (Fig. 9C). D_3 refolded these early D_1-D_2 folds as shown in Fig. 9D, but late antithetic reactivation of bedding during D_3 destroyed most of the long limbs preserving only short limb remnants (Fig. 9E).

Regional Deformation (D_3)

D_3 was the strongest event in the Yeneena Group and produced upright to weakly SW inclined open to tight NW-SE trending folds at all scales (*e.g.* Fig. 3) with a regionally penetrative axial plane cleavage, S_3. S_3 is variably a crenulation to pervasive slaty cleavage, depending locally on the presence or absence of earlier cleavage (*cf.* Williams, 1972). It was produced by sericite/muscovite recrystallisation (*e.g.* Bell, 1978; White & Knipe, 1978) and quartz elongation in the rock matrix during sub-greenschist facies metamorphism (quartz - albite - dolomite - chlorite - sericite).

A sub-vertical stretching lineation, L_3^3, defined by mineral streaking on S_3, and the formation of complementary high angle thrust and lag faults on D_3 fold limbs (Hickman et al., 1994) indicates mainly vertical movement during D_3. These high angle thrusts, which have a consistent SW vergence, developed late in D_3 following antithetic reactivation (Bell, 1986; Aerden, 1993) of bedding along fold limbs (Fig. 10). Across both study areas D_3 folds have variable plunges, but commonly they are either sub-horizontal or moderately SW plunging. Pre-existing nappe and thrust structures within basement gneisses were variably reactivated and steepened during D_3 deformation (Hickman & Clarke, 1993; Hickman et al., 1994).

Post-regional Deformation (D_4, D_5, D_6)

Deformation Event D_4 : D_4 produced mesoscale asymmetric folds of S_3 (Fig. 11) and bedding (Fig. 12), with an associated coarsely spaced crenulation cleavage (S_4: Fig. 13a, b). F_4 axial planes and S_4 cleavage are sub-horizontal or dip shallowly NW to N, and F_4^0 and F_4^3 axes are sub-horizontal and trend parallel to F_2 axes (NW-SE). D_4 effects are particularly evident in examples where upright S_3, preserved in fibrous quartz veins, is inclined and flattened in the matrix (Fig. 14). Locally, D_4 folds are symmetric (Fig. 13b), particularly where refolding steeply oriented foliations, suggesting layer-parallel compression arising from vertical shortening.

Differentiated crenulation (Fig. 13a) / fold asymmetry (Fig. 11) indicates top-to-the-SW shear along S_4 suggesting a bulk SW directed tectonic transport, although locally this may reverse (top-to-the-NE - *e.g.* Figs 12 & 14). Such reversals, which are not related to crossing macroscale F_4 fold axes, are characteristic of coaxiality in the deformation (*e.g.* Bell & Johnson, 1989, 1992) and suggest that D_4 comprised bulk vertical shortening that caused differential horizontal movement. D_4 folds and cleavage mainly occur on the SW limbs of D_3 antiforms (Fig. 15), where bedding was suitably oriented for folding. In contrast, the opposite NE limbs underwent reactivation (Bell, 1986) of bedding with subsequent crenulation of S_3 (Fig. 15).

Deformation Event D_5 : D_5 is characterised by finely- to coarsely-spaced crenulations (Fig. 16) and crenulation cleavage (S_5; Fig. 17) that overprint sub-horizontal S_0 and S_3, particularly in F_4 hinges (*e.g.* Figs 15 & 17). S_5 is sub-vertical and strikes NNW-SSE, approximately 20-30° clockwise from S_3 (Fig. 16), indicating a separate phase of horizontal shortening at a low angle to that which occurred in D_3. In thin section S_5 commonly resembles shear bands (*e.g.* Cobbold, 1977; Fig. 17). No meso- or macroscale D_5 refolding was identified, due to the low angle between D_3 fold axes and D_5 compression (*e.g.* Ghosh & Ramberg, 1968; Ghosh et al., 1992).

Deformation Event D_6 : D_6 produced late kink folds (Fig. 18A) and discrete crenulations/shear bands (Fig. 18B) at high angles to S_3. These have sub-vertical to steep NW to NNW dipping axial planes that are marked by fractures, micro-faults or dissolution seams. Asymmetric folding of S_0/S_3 across the axial plane exhibits a consistent sinistral shear sense and no conjugate sets were observed indicating that regional stresses during this deformation were oriented oblique to the S_0/S_3 foliation trend (Gay & Weiss, 1974; Reches & Johnson, 1976). This suggests that D_6 compression was approximately NNW-SSE.

STRUCTURAL ARCHITECTURE / GEOMETRY of the PATERSON PROVINCE

The Paterson Province lies at the margin of the Anketell Gravity Ridge (Wellman, 1976; Fig. 19). This ridge, which underlies much of the Paterson Orogen (Williams & Myers, 1990), terminates south-westwards against the Karara and Savory Basins (gravity lows - Fig. 19). Within it, semi-linear zones of intermediate density reflect basement anisotropies (*e.g.* doming, lithological variation, tips of thrust sheets) and parallel the gross trend of the orogen. The Rudall Metamorphic Complex, and areas examined in this study, lie marginal to this ridge suggesting that they represent the tip of a NE thickening wedge of Proterozoic crustal material (*e.g.* Chin et al., 1980; Clarke, 1991).

A particularly large ovoid shaped gravity anomaly (high) appears to be discordant against the basement grain (Fig. 19). This anomaly, located SSE of Telfer (Fig. 19), may represent upward doming of dense basement. However, the apparent discordancy suggests that it could be an igneous feature. Although the surface of the anomaly is concealed, the presence of sporadic gabbroic outcrops (Hickman & Clarke, 1993), coupled with evidence of strong Na alkali metasomatism in nearby mineral prospects (*e.g.* Grace Prospect - see Section E for location) may indicate that it represents a gabbroic complex.

The structural trend and style of the Yeneena Group, as indicated by magnetic linears that correspond to strike-lengths of individual units, varies across the Paterson Province (Fig. 19). The gross structural trend ranges from NW-SE in the Broadhurst Range area through to NNW-SSE (locally N-S) in the Throssel Range and Anketell regions (Fig. 19). Magnetic linears in the Throssel Range and Anketell regions are closely spaced, and truncated curvilinears (Fig. 19) suggest that regional folding is markedly truncated by faults, implying compressional thrust-folding (*e.g.* Harland, 1971; Jones, 1987). Other linears, at a low-angle to the structural trend, truncate the regional grain suggesting faulting of the sequence after regional folding (Fig. 19).

GEOCHRONOLOGY

Geochronological data for the Paterson Province are summarised in Table 3. Rocks of the Rudall Metamorphic Complex exhibit widespread dates (Table 3). However dating of orthogneiss mineral assemblages provides ages between 1333 ±44 Ma (Rb/Sr - Chin & deLaeter, 1981), 1250 Ma (U/Pb - Hickman & Clarke, 1993) and 1132 ± 21 Ma (Rb-Sr - Chin & deLaeter, 1981) for the Watrara Orogeny (see Table 1). Deposition of the Yeneena group therefore occurred post 1132 Ma but prior to 940 - 900 Ma epigenetic galena in the Broadhurst Formation, and intrusion of the late/post-tectonic Mt Crofton Granitoid Complex in the NE Paterson Province (Goellnicht et al., 1991) between 700 - 600 Ma (*e.g.* Trendall, 1974; McNaughton & Goellnicht, 1990; D.Nelson pers. comm., 1994; Table 3).

Geologic Feature	Age	Comments
Rudall Complex -		
Orthogneiss	2015 ± 26 -> 1765 ± 15Ma (U-Pb)[1]	Age of granitoid crystallisation
Orthogneiss	1333 ± 44Ma (Rb-Sr)[2]-	Watrara Orogeny (D_2)
Metadacite Dike	1250 Ma[3]	Pre - Watrara Orogeny
Orthogneiss	595 ± 27Ma (Rb-Sr)[2]	Alteration of orthogneiss (incl. sericitisation, saussuritization, chloritisation)
Yeneena Group -		
Depositional age -	1132 ± 21Ma (Rb-Sr)[2]	Pegmatite veins restricted to the Rudall Complex, maximum age of deposition
Epigenetic galena mineralisation	940 Ma (Pb-Pb)[4]; 900 (Pb-Pb)[5]	Minimum age of Yeneena Group deposition
Paterson Orogeny -		
Minyari Granite Complex	≈ 640 Ma (U-Pb)[6]	Crystallisation age of D_4 foliated Minyari "Gneiss" (see Goellnicht et al., 1991)
Mt Crofton Granite Complex; NE Paterson Province	690 ± 48 Ma (Pb-Pb)[7], ≈620 Ma (U-Pb)[1], 601±42 (Rb-Sr)[8]	Late to post D_4 deformation; broadly concurrent with mineralisation in the NE Paterson Province

***Table 3** - Summary of geochronological data for various geologic features in the Paterson Province (Note: "Comments" refer to the quoted sources' interpretation of the geological significance of the date). 1; D.Nelson quoted in Hickman et al., 1994. 2; Chin & deLaeter, 1981. 3; D. Nelson quoted in Hickman & Clarke (1993). 4; I. Fletcher quoted in Hickman et al., 1994. 5; I. Fletcher quoted in Blockley & Myers, (1990). 6; D. Nelson quoted by L. Bagas (pers. comm. 1995). 7; McNaughton & Goellnicht, (1990). 8; Trendall, (1974).*

The latter constrains the age of the Paterson Orogeny to older than 620-600Ma, although the dates are widespread (Table 3). U-Pb zircon dating of granites foliated by the regional S_3 fabric (*e.g.* Minyari Gneiss - NE Paterson Province; see Section E for location), gives an approximate crystallisation age of 640Ma (D. Nelson quoted by L. Bagas, pers. comm. 1995), thus providing an upper limit for the Paterson Orogeny. Additionally, geological and stratigraphic relationships in the Savory Basin (Figs 1 & 2) also constrain the age of deformation to between 640 and 600Ma (Williams, 1992).

METALLOGENESIS - YENEENA GROUP

Although poorly exposed the Paterson Orogen is currently being actively explored owing to the discovery and current mining of two major deposits, Telfer (Au - Cu) and Nifty (Cu), hosted within the Yeneena Group. Other significant prospects, including gold, gold-copper, porphyry-copper, tin-tungsten and potential Pb-Zn mineralisation occur in the NE Paterson Province in association with weakly calc-alkaline granites (Goellnicht, 1992; Rowins, 1994; see Section E for description). In the lower Yeneena Group, copper prospects are common in the Broadhurst Formation, as are numerous Pb-Zn gossan anomalies (Hickman & Clarke, 1993). The Nifty deposit is hosted within graphitic shales in the Western Geographic Zone (Fig. 2) near the Vines Fault. Only trace occurrences of gold have been reported in the Lower Yeneena Group (Hickman et al., 1994).

INTERPRETATION

Comparison between the tectonic sequences in the lower Yeneena group and the NE Paterson Province indicates they are the same, with each deformation directly correlative on the basis of style, orientation and relative overprinting relationships. This allows a province-wide polyphase deformation history to be proposed for the Yeneena Group (Table 4). Note however, that the earliest deformation in the NE Paterson Province has been inferred (Section A), as compared to this study where it is positively identified, and thus it and the rest of the sequence have a different notation to that presented in the left column of Table 4. Although stratigraphic uncertainty exists across the basin, the consistent tectonic history indicates that by the time of the Paterson Orogeny, the province was a uniform tectonic terrane (terminology as defined by Irwin, 1972; Gibbons, 1994).

Deformation Event	CHARACTERISTICS	Correlation NP	PS
D_1	Upright-inclined symmetrical mesoscale folding, now commonly recumbent, with NW-SE trending axes; ***NE-SW directed horizontal crustal compression.***	D_{EF}	
D_2	Recumbent monoclinal folding with a sub-horizontal axial plane cleavage (S_1) and NW-SE trending F_1 axes; ***vertical crustal shortening coupled with SW directed horizontal movement***	D_1	D_3
D_3	Upright NW-SE trending folding at all scales with a regionally penetrative axial plane cleavage (S_3); ***NE-SW horizontal crustal shortening.*** Doubly plunging F_3 axes in some areas (*e.g.* Telfer region).	D_2	D_4
D_4	Recumbent monoclinal and asymmetric mesoscale crenulation folding, NW-SE trending F_4 axes; ***vertical crustal shortening with horizontal tectonic transport*** (SW directed in the lower Yeneena Group, NE directed in the Telfer region.	D_3	D_5
D_5	Millimetre to centimetre scale upright crenulation folding and cleavage (S_5), with NNW-SSE trending F_5 axes; ***horizontal ENE-WSW crustal compression.***	D_4	
D_6	Sporadic kink and crenulation folding, with rare cross-folding of F_3 folds and axial plane cleavage; ***WNW-ESE directed crustal compression.*** A second period of ***N-S directed horizontal compression*** is suggested in the Telfer region.	D_5(a) D_5(b)	D_6

Table 4: *Summary of the proposed deformation sequence for the Paterson Province. The left column designates the province scale notation, whilst the right details the correlation between this sequence and those of the NE Paterson Province (Telfer district; Section A; NP) and of previous studies (Hickman et al., 1994; PS) . The kinematic interpretation for each deformation is italicised and bold.*

Multiple Phases of Horizontal and Vertical (orthogonal) Deformation; Implications for Compression and Gravity-driven Orogenesis.

The deformation sequence comprises alternate horizontal and vertical cleavage forming deformations. Vertical deformations, typified by upright folding/cleavage and steeply plunging stretching lineations, are interpreted to result from horizontal bulk inhomogeneous crustal shortening (*e.g.* Ramsay, 1967; T. Bell, 1981). In contrast horizontal deformations, marked by recumbent folds and shallow cleavage formation, may arise through either vertical shortening, shearing or a combination of the two (*e.g.* Coward & Kim, 1981; Meneilly, 1983; Platt et al., 1983; Dietrich & Casey, 1989; Ray, 1991; Frotzheim, 1992).

For both D_2 and D_4, evidence supporting bulk vertical shortening includes:

1. Sub-horizontal cleavage development, implying a vertical σ_1 (Turner & Weiss, 1963; Ramsay, 1967; Ghosh, 1995), that is not associated with pervasive high-strain zones (*e.g.* thrusts, shear/mylonite zones or nappe folding) in which a horizontal foliation could form through the intensification of the progressive shearing component (*e.g.* Ramsay & Graham, 1970; Coward & Kim, 1981; Sanderson, 1982).
2. The locally coaxial nature of D_2 and D_4, as marked by fold vergence/cleavage shear sense reversals, which were not related to changing positions on macroscale folds (*cf.* Platt et al., 1983), is typical of differential horizontal movement driven by vertical shortening (Bell & Johnson, 1992; Aerden, 1994; Northrup, 1996).
3. The character of D_2 and D_4 folds where steeply dipping elements are coaxially folded (buckled - *cf.* forelimb folding, Ray, 1991) and shallow limbs are attenuated (*e.g.* Fig. 9,15). This is analogous to bulk inhomogeneous vertical shortening of the steeply oriented structural elements (*ie.* they lie in shortening domains - T.Bell, 1981), and shearing (attenuation) of the shallow-dipping elements as a result of deformation partitioning (*e.g.* Bell, 1985; Bell et al., 1989)

A potential cause of the vertical shortening is gravitational collapse (*e.g.* Bucher, 1956; Price, 1973; Elliot, 1976; Ramberg, 1981) which occurs when an orogenic pile is sufficiently uplifted (or weakened through delamination) to induce gravitational instability and collapse. The regular alternation of horizontal and vertical cleavage forming deformations is a feature of gravity collapse tectonism (Platt et al., 1983; Bell & Johnson, 1989, 1992; Aerden, 1994) produced through periodic collapse of a compressional orogen core. Similar styles are also observed within accretionary and orogenic wedges (*e.g.* Platt, 1986; Dahlen & Suppe, 1988; Molnar & Lyon-Caen, 1988) where slope instability and underplating (Leggett et al., 1985; Casey-Moore et al., 1991) produce extension and thrusting in different parts of the wedge, which overprint each other through progressive orogenic movement.

Implications for a Migrating Orogenic Core

Ductile collapse is commonly viewed as a lateral symmetrical spreading of material towards the orogen margins (*e.g.* Bucher, 1956; Choukroune & Séguret, 1973; Ramberg, 1981; Northrup, 1996). However, recent microstructural studies have indicated that vertical shortening during collapse, through deformation partitioning, produces coaxial and non-coaxially deformed rocks at a larger range of scales (Bell & Johnson, 1989, 1992; Aerden, 1994). Consequently, the movement/shear sense directions of sub-horizontal deformations, as indicated by fold vergence and crenulation asymmetries, can provide important constraints on the position of a rock sequence with respect to an orogen core that is the locus of collapse.

The limited data for D_2, including those from the NE Paterson Province (D_1 - Table 3; Section A), indicate a consistent SW vergence (shear sense) across the Paterson Province (*ie.* non-coaxial on the province scale). This consistent vergence, and hence movement direction, implies that the core of the orogen lay external to the Paterson Province, most likely NE (further into the Paterson Orogen belt) of its present position. Consequently, initial D_1 shortening was weak, generating only minor folds, probably because of the distance from the orogen core (Fig. 20A). Collapse about the core during D_2 would subsequently produce SW directed thrusting onto the margins of the orogen (Fig. 20B) being everywhere consistent across the Paterson Province.

Renewed compressional shortening in the Paterson Province during D_3 was much stronger than for D_1 (Fig. 20C), and may have resulted from the migration of deformation to the SW, or by widening of the orogen (*e.g.* Molnar & Lyon-Caen, 1988). During D_4, the second collapse phase, there was a vergence (shear sense) reversal across the province. In the lower Yeneena Group, D_4 structures consistently verge SW, whilst in the northern Paterson Province they verge NE (D_3 in Section A). This suggests that sufficient instability/weakness in the orogenic pile occurred between the northeastern areas of the province and the lower Yeneena Group to cause a switching in the bulk movement during D_4. The opposing movement directions are similar to the thrusting geometries produced at deep crustal levels on either side of a collapsing orogen (Bell & Johnson, 1992), implying that the orogen core lay between Telfer and the Rudall Complex. However, this shear sense combination is unlikely to represent thrusting in the lower sections of an orogen given the relatively shallow crustal setting of the rocks in this study.

An alternative explanation is that the two areas lay on either side of the inflection line demarcating extension collapse in the upper half and thrusting in the lower (and marginal) sections of the orogen (Fig. 20D; Bell & Johnson, 1992). This implies that the inflection line shifted south-westwards during and after D_3 to lie between Telfer and the Rudall Metamorphic Complex (Fig. 20D). In the tapered wedge model of Platt (1986), this demarcation can be inferred as passing through the central portion of the wedge, rising to meet the surface between extensional thinning at the rear and thrusting deformation at the front (Fig. 20E). Such a wedge, encompassing the Paterson Province, could have existed in the foreland of an active collisional orogen and is consistent with the relatively low-grade metamorphic conditions that occurred in the Yeneena Group.

Therefore, the core of the Paterson Orogen may have shifted towards the north-eastern edge of the Paterson Province (Fig. 20D), possibly explaining the voluminous intrusion of granitoids in this locality (Telfer region), and is consistent with the oldest of these bodies (syn-D_3) occurring in the northernmost part of the Paterson Province (*e.g.* Minyari Gneiss - Goellnicht, 1992; see fig. 1 - Section E). Alternatively, or additionally, with the major centre of orogenic collapse further afield, a local gravitational instability may have occurred. Granite intrusion could have caused this through advection of heat into the mid crust, with subsequent weakening and ductility enhancement (Burchfiel & Davis, 1975; Molnar & Tapponier, 1981) inducing collapse (Fig. 20F) and thrust accretion (Armstrong, 1974; Burchfiel & Davis, 1975).

Late-Proterozoic Tectonic Development of the Paterson Province

Kinematic axes for vertical deformations D_1, D_3 and D_5 are broadly orthogonal to the structural grain of the Paterson Province indicating an overall NE-SW convergence during the Paterson Orogeny. Such convergence appears to have been widespread along the length of the Paterson Orogen at this time (Fig. 21) and could represent terrane accretion from the north/north-east (Myers, 1990a,b). Gravity data (see above) suggest that the present Paterson Province lay at the tip of a wedge of crustal material, which would have been thrust south-westwards on to the adjoining Pilbara Craton. This thrusting may also have produced the Karara Basin, and parts of the Savory Basin through thrust loading (*e.g.* Beaumont, 1981; Jordan, 1981) of the foreland continental margin (Fig. 21).

D_6 Deformation : D_6 deformation appears to post-date the Paterson Orogeny. Evidence for this comes from the Karara Basin, where the major structures (regional scale E/ENE plunging folds and associated fracture cleavage - Williams, 1990b) are consistent with D_6 compression. Given the likely foreland genesis of this basin during the Paterson Orogeny, D_6 structures could only have formed after a sufficient hiatus allowed lithification to occur. This implies that D_6 was not associated with the main period of NE-SW convergence and thus probably post-dates the Paterson Orogeny.

Macroscale Geometric Development of the Paterson Province

Major fold/structural styles vary throughout the Paterson Province (Fig. 22) providing information on the response of various localities to convergent deformation. These styles include fault-truncated folding (Anketell - Fig. 22A; south-west margin), open domal folding (Telfer region - Fig. 22B), and transpressive fault-folding (*e.g.* Harland, 1971; Sanderson & Marchini, 1984) along the Vines Fault (Fig. 22C). The differences reflect the internal orientation of rock units in each area within the province (*e.g.* transpression in the N-S oriented Vines Fault locality - Fig. 22) as well as the response of these to shortening deformation.

Later deformations (D_4, D_5 & D_6) appear to have locally modified the geometry of D_3 folds and faults, although these effects are largely controlled by the inherent structural anisotropy of D_3 folding. In the Telfer region, some D_3 domes are refolded by D_4 (Fig. 22B) producing a marked asymmetry. In the Broadhurst Range Hickman et al. (1994) have described prominent lag faulting along the limbs of SW inclined D_3 folds. Such a geometry is consistent with flattening, rotation and subsequent antithetic reactivation of the trailing limbs during D_4 (Fig. 22D).

D_5 shortening did not refold, or produce significant hinge migration (*e.g.* Odonne & Vialon, 1987), in D_3 folds owing to the similar axial orientation between the two events. However, D_3 folds are commonly truncated by late axial-plane parallel/sub-parallel faults that show evidence for repeated movement (Hickman & Clarke, 1993; Hickman et al., 1994; see also Section E), which is commonly sinistral (Fig. 22A,B & D). Coupled with large truncated magnetic linears (*e.g.* Figs 22 & 22A), which suggest post-D_3 "shuffling" of the Yeneena Group, these features are consistent with a shift towards transpressive deformation as the horizontal shortening component rotated clockwise from NE-SW (D_3) to ENE-WSW (D_5/D_6).

D_6 deformation may have also contributed to fault development including the development of prominent N-S trending faults in the Broadhurst Range area, and could have initiated dextral reactivation of pre-existing NW-SE trending faults (Fig. 22D). In the Throssel Range and Vines Fault areas, N-S trending faults would have been suitably oriented for sinistral reactivation resulting in potentially complex local fault geometries and movement histories (Fig. 22C).

DISCUSSION

Tectonic Setting of the Paterson Province

The nature and style of Proterozoic tectonism has been controversial, with ensialic orogenesis being proposed for many early-middle Proterozoic belts in Australia (Duff & Langworth, 1974; Rutland, 1976; Plumb, 1979; Kröner, 1981; Etheridge et al., 1987). Similarly, late-Proterozoic (Pan-African) belts were also considered to be ensialic (Kröner, 1977,1982; Martin & Porada, 1977; Engel et al., 1980). However, more recently Phanerozoic tectonic analogues have been interpreted in Pan-African belts leading to considerable consensus for modern-tectonic processes operating in these belts (*e.g.* Windley, 1983, and references therein; Key et al., 1989; Kröner et al., 1990; Kukla & Stanistreet, 1991; Murphy & Nance, 1991;).

A major problem in the Paterson Province has been the determination of a tectonic setting given the presence of seemingly conflicting tectonic discriminators (*e.g.* granitoid geochemistry, sedimentation, mineralisation styles). On the one hand the Yeneena Group contains no known volcanic material or ophiolitic rocks indicative of ocean closure. The depositional environment (shallow marine shelf - Williams, 1990a; Hickman et al., 1994) suggests either a continental margin/platform or intracratonic setting (Turner, 1982). Continent-continent collision and amalgamation during the Watrara Orogeny at 1330Ma, recorded in the basement Rudall Metamorphic Complex, occurred prior to deposition of the Yeneena Group (Clarke, 1991) and implies an ensialic intracratonic setting (Goellnicht, 1992).

Similarly, granitoid geochemistry, mineralisation and structural styles in the province are suggestive of a collisional setting during the Paterson Orogeny (Goellnicht et al., 1991; Goellnicht, 1992; Myers & Barley, 1992). In particular, the Telfer Au-Cu deposit bears many similarities to mesothermal slate-belt hosted lode-gold mineralisation (Section C) that develops in collisional/accretionary continental margin tectonic settings (*e.g.* Barley et al., 1989; Kerrich & Wyman, 1990). Similarly, large-scale thrust imbrication and deep syn-orogenic foreland basin development along the Paterson Orogen (Fig. 21) suggests collision/accretion of terranes to the north and north-east (Myers, 1990b, 1993). Additionally, the gravity-collapse nature of tectonism identified in this study appears to be a feature of modern collisional/accretionary orogenesis (Bell & Johnson, 1989).

The results of this study, which suggest that orogenesis migrated/expanded south-westwards, imply that the Paterson Province was only a small part of a much larger orogen that originated NE of the exposed terrane. With much of this orogen obscured by Phanerozoic basins, vital evidence for the tectonic setting is inevitably missing. Consequently, a collisional zone, as suggested by granitoid geochemistry (Goellnicht et al., 1991) and mineralisation styles, may be located further north than the presently exposed Paterson Province. Such a collisional margin, potentially involving subduction, is likely to be genetically unrelated to the initial development of the Yeneena Basin, which occurred around 1100-900Ma. This scenario would be consistent with the interpretation that orogenic activity migrated south-west into the Paterson Province. Initial orogenic activity would be only weakly transmitted through the relatively old, and therefore strong plate (Molnar & Tapponnier, 1981). However, as orogenesis fully developed, uplift of a central plateau to critical levels would result in gravitational collapse and widening of the zone of orogenic deformation (*e.g.* Burchfiel & Davis, 1975; Dahlen & Suppe, 1988; Molnar & Lyon-Caen, 1988), thus progressively intensifying deformation of the Paterson Province.

Consequently, the Yeneena Basin may represent an intracratonic (ensialic) back-arc or foreland basin located behind a collisional front (Fig. 23). Such a setting may initially have been well inboard of orogenic activity. Migration of the zone of orogenesis south/south-westwards would cause the Yeneena Basin to subsequently become part of an active collision/accretionary tectonic setting, thus explaining the apparent ensialic/intracratonic character of the Yeneena basin and the lack of volcanism and other rock types associated with subduction. Although the exact nature of orogenesis is speculative, paleogeographic reconstructions of the late-Proterozoic suggest the Paterson Orogen lay internally within a supercontinent (Hoffman, 1991; Moores, 1991). Thus orogeny could be of the interior-type described by Murphy & Nance (1991), characterised by continent-continent collision and, in many cases, pre-cursor oceanic crustal subduction.

Subduction or terrane accretion to the north of the Paterson Province would also provide a good explanation for the presence of calc-alkaline magmatism, porphyry-copper and mesothermal lode-gold mineralisation in the NE Paterson Province. Lithospheric delamination (*e.g.* Bird, 1978, 1979; Nelson, 1992) and resultant asthenopheric upwelling (Fig. 23A/B) is one mechanism that may have provided the necessary heat source for melting of the lower crust. Basic magma, transported along deep thrusts or décollements to underplate (Fig. 23C) the Paterson Province would have produced the mixed-source and highly fractionated felsic melts that subsequently intruded higher crustal levels in the NE Paterson Province.

Similarly, subduction may produce large volumes of fluid (Fyfe & Kerrich, 1985; Fig. 23D & E), which are a potential source for gold mineralisation (Goldfarb et al., 1986; Kerrich & Wyman, 1990). Coupled with metamorphic fluid generation in the deeper and hotter core of the orogen (Fig. 23F), such fluids could have migrated southwest along deep thrusts and décollements to subsequently rise through the crust generating mid-shallow level gold mineralisation in the Paterson Province. This may have occurred along fractures created or utilised by felsic magma, assisted by reactivated strike-slip faults during the switch from orthogonal to oblique convergence across the province (*e.g.* Kerrich & Wyman, 1990).

Periodic Extensional Opening of the Paterson Orogen

If subduction occurred north of the Paterson Province, then the Paterson Orogenic belt must have been sufficiently rifted apart to allow the formation of oceanic crust. Geological evidence from the Paterson Province, and adjacent sedimentary basins/orogens, suggests that repeated periods of compression parallel to the trend of the Paterson Orogen occurred during its history (*e.g.* Austin & Williams, 1978). In the Paterson Province this is suggested by D_6. However, similarly oriented compression also occurred prior to the Paterson Orogeny at ≈850-800Ma in the Savory Basin (Williams, 1992 - Blake Movement). South-east of the Paterson Province, the Amadeus Basin (Fig. 1) records a complex history of N-S contractional opening during the late-Proterozoic (Lindsay & Korsch, 1991; Shaw 1991).

The compressional episodes described above appear to broadly correspond with periods of dispersal and rifting of the late Proterozoic super-continent (Bond et al., 1984; Young, 1984; Lindsay et al., 1987). Consequently, they may provide a mechanism for weakening the Paterson Orogen leading to rifting (and ocean opening), sedimentary basin formation (Shaw et al., 1991) and initiation of orogenesis (*e.g.* Le Pichon et al., 1982). The reason for such compression, which suggests an approximate and periodic 90° change in plate motion, is speculative. It may be related to strain partitioning (*e.g.* Weijermars, 1993) within the "trans-Australia zone" during rotation of the northern part of Australia (Austin & Williams, 1978; Powell et al., 1994).

SIGNIFICANCE

The tectonic development of the Paterson Province discussed in this study has important implications for reconnaissance mineral exploration of deformed sedimentary basins within and marginal to Proterozoic orogenic belts. The migration of far-field orogenesis suggested by this study indicates a potential for such basins to be encompassed by the effects of crustal collision. As with the Paterson Province, the setting of these basins could be obscured by younger sedimentary cover leading to incorrect tectonic interpretations. Consequently, the potential for epigenetic mineralisation, associated with such settings, may be overlooked. During the late-Proterozoic many long-lived mobile belts underwent further tectonism (*e.g.* Windley, 1983; Bond et al., 1984; Lindsay et al., 1987), which coupled with an increase in gold mineralisation in this period (Hutchison & Vokes, 1987; Barley & Groves, 1992), suggests that reactivated late Proterozoic orogens, and their marginal sequences, may generally be important metallogenic provinces for further exploration.

The development of the Telfer Au-Cu deposit suggests that migration of a collisional orogen may produce mineralisation inboard of the orogenic front. This may be spatially associated with regional granitoids and thus be readily identifiable if the granites are exposed. However, mineralisation may actually have a metamorphic origin, or associated granites may not have reached the exposed level of the crust. In such cases, reconnaissance structural examination of collapse-related deformations may allow the position of the orogen core, and in which direction it potentially moved, to be determined (*e.g.* see Fig. 23/23G). This would facilitate the identification of areas with potential for mineralisation and the likely direction of metamorphic fluid migration.

CONCLUSIONS

1. Reconnaissance structural analysis in Lower Yeneena Group sedimentary rocks of the Paterson Province has indicated a polyphase sequence of orthogonally overprinting (vertical-horizontal) deformation events. This sequence correlates with that observed in the Telfer region and indicates that uniform tectonism occurred across the Paterson Province during the late Proterozoic Paterson Orogeny (≈ 640-600Ma).
2. Horizontal cleavage-forming deformation events in the sequence indicate that horizontal tectonic compression was interspersed by periods where vertical crustal shortening predominated. This tectonic style is suggestive of gravity collapse or accretionary wedge tectonism whereby gravitational instability causes the regular spreading of material to the margins of an orogen. Therefore the Paterson Province probably lay within an orogenic wedge. This wedge was strongly shortened during initially orthogonal convergence. However, over time this became oblique to the belt causing it to become weakly transpressive.

3. Compilation of the bulk tectonic transport directions for each of the horizontal deformations across the province suggests that orogenesis initiated considerably to the northeast of the presently exposed Paterson Province. Over time the core of this orogen migrated south-westwards, with subsequent widening of the orogenic welt, to encompass the Paterson Province generating regional deformation, granitoid intrusion and mineralisation.
4. Inboard migration of the orogen core may provide an answer to the problematic tectonic setting of the Paterson Province. The Yeneena Basin is interpreted to have developed in an intracratonic (ensialic) setting that, subsequent to orogenic migration, lay behind (back-arc) a collisional tectonic front. Consequently, the Yeneena Basin was incorporated into an accretionary wedge-type environment at the foreland margin of a migrating/widening orogen. Overprinting of this collisional orogenesis on the relatively inactive intracratonic basinal setting has produced conflicting tectonic indicators.
5. A collisional front, involving subduction and lithospheric delamination, to the NE of the Paterson Province provides an explanation for the source and generation of regional calc-alkaline granitoids and mineralisation styles in the Telfer region. Such a setting is also a potentially important source-region for auriferous fluids; these could have been involved in the genesis of the Telfer Au-Cu deposit.
6. Migration of orogenic activity, particularly collisional-styles, has important implications for other sedimentary basins lying marginal to or within trans-continental orogenic belts. These basins, which may appear to have an intracratonic setting, could have been affected by migrating far-field collisional/accretionary orogenesis. As a consequence, such basins may represent important target-districts for mineral exploration.

SECTION C

Late Structural Timing of Mineralisation in the Telfer Au-Cu deposit and the Role of Orogenic Deformation in Regional Fluid Flow and Mineralisation.

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ABSTRACT

A detailed paragenetic study of the late-Proterozoic Telfer Au-Cu deposit (northwest Western Australia) indicates a progressive sequence of hydrothermal vein/alteration mineral development. This comprises 1. Massive quartz veining; 2. Grey ferroan dolomite-epidote-chlorite-muscovite; 3. White-pink ferroan dolomite-chlorite-pyrite-galena; 4. Silica-chalcopyrite-calcite-chlorite-muscovite; 5. Late argillic and chalcedonic veining. Gold mineralisation is coeval with the later stages of ferroan dolomite precipitation and accompanied chalcopyrite and pyrite deposition. Structural timing of the vein phases indicates that, with the exception of Stage 1, they were precipitated during later, and weaker, orogenic deformation (D_3 and D_4) that post-dated the major terrane deformation (D_2).

Vein and alteration assemblages are similar to regional metamorphic assemblages, indicating that the ore-fluids equilibrated with the country rocks and homogenised prior to entering the deposit. This suggests that the ore-fluids may have had a significant metamorphic fluid component. Consequently, a granite-related genetic model, where ore-fluids were primarily derived from magmatic and connate/formational sources, as initially proposed, appears unlikely for the Telfer deposit. Instead, the deposit appears to have formed in a mid-crustal environment where there was a close association between metamorphic and magmatic fluid components. Similarly, the mineralogy of the vein/alteration assemblage is compatible with fluid immiscibility as a precipitational mechanism, suggesting that fluid-mixing may not have been as important as previously suggested.

The association between the release of ore-fluids from the sequence and late orogenic deformation suggests that the late timing of mineralisation in metamorphic-magmatic terranes may be controlled by changes in the predominant deformation mechanism. This could arise due to the inherent heterogeneity of late orogenic deformation that develops discrete zones of structural permeability in the form of gaping/dilation, faulting and veining. At a crustal-scale, ore-fluids would be focussed into these localised zones, increasing the fluid flux and thus potential to precipitate mineralisation.

INTRODUCTION

Epigenetic mesothermal lode-gold deposits are an important source of the worlds gold supply and developed predominantly in the Archean and Phanerozoic eras with only limited examples in the Proterozoic (Hutchison & Vokes, 1987; Barley & Groves, 1992; Kerrich & Cassidy, 1994). These deposits share many common characteristics including a convergent/accretionary tectonic setting (Cox et al., 1983; Kerrich & Wyman, 1990; Cox et al., 1991a), a regional metamorphic genesis (Phillips & Groves, 1983; Goldfarb et al., 1986; Groves & Phillips, 1987; Powell et al., 1991; Craw, 1992), strong structural control (Sandiford & Keays, 1986; Mueller et al., 1988; Vearncombe et al., 1989; Cox et al., 1991b) and a late mineralisation timing (Craw, 1989; Jemielita et al., 1990; Kontak et al., 1990; Robert, 1990; Forde, 1991). A strong relationship between regional metamorphism and terrane deformation in the genesis of lode-gold deposits is commonly suggested by the close similarity of ore-fluid and ore-mineral assemblages with those in the surrounding country rocks (*e.g.* Kerrich & Cassidy, 1994; Phillips et al., 1994)

The late-Proterozoic Telfer Au-Cu deposit displays some features suggestive of a magmatic-induced genesis, including a high-salinity ore-fluid component and a spatial/temporal association to both regional granitoids and their associated mineralisation (Goellnicht, 1987; Goellnicht et al., 1989). However, it also exhibits many characteristics of a typical mesothermal lode-gold deposit (as outlined in the previous paragraph), including a strong structural control and late mineralisation timing (Goellnicht, 1987; Goellnicht, et al., 1989). Previous studies of the timing of mineralisation indicated that it occurred late in or post- the formation of the Telfer deposit host-structure (van Dijk, 1986; Goellnicht et al., 1989). These studies were however, constrained by the then accepted regional tectonic history in which host-structure formation (D_2 - van Dijk, 1986) was the youngest defined deformation episode. However, subsequent regional structural analysis in the Telfer region has identified at least three deformation events post-dating D_2 (Section A), suggesting a potential for mineralisation to have formed after D_2

The recent (1992) discovery of further mineralisation at depth in the Telfer deposit has greatly expanded the scope and importance of this ore-system. Coupled with the recognition of a more protracted deformation history in the host terrane (Sections A & B), the role of orogenesis in the formation of the deposit requires further examination. This section presents the results of a paragenetic and structural timing of mineralisation study of the Telfer ore assemblage that has provided new constraints on both the genetic development of the deposit and on the timing of ore-fluid movement in shallow- to mid-crustal metamorphic/magmatic settings. The controls on mineralisation timing in such settings have been the source of considerable debate (see Kerrich & Cassidy, 1994 for a review), and are commonly interpreted as the result of tectono-thermal processes associated with regional metamorphism (Norris & Henley, 1976; Colvine, 1989; Hodgson et al., 1989; Craw, 1992). The late mineralisation timing at Telfer, where a degree of magmatic input is implied (Goellnicht, 1992), suggests that processes common to both metamorphic and metamorphic-magmatic environments at mid-crustal levels may operate to control the timing of ore deposition in a lode-gold deposit.

REGIONAL GEOLOGY AND PREVIOUS INVESTIGATIONS

The Telfer deposit is hosted in middle Proterozoic Yeneena Group metasedimentary rocks of the Paterson Orogen, in the Paterson Province, northwest Western Australia (see fig. 1 Section A). The Yeneena Group was deformed by convergent orogenesis late in the Proterozoic (the Paterson Orogeny ≈ 640-600Ma) involving a sequence of multiple orthogonal overprinting deformation events (Section A) accompanied by sub-greenschist facies regional metamorphism. The major deformation, D_2, formed prominent NW-SE trending folds throughout the Paterson Province, and in the Telfer region created a dome & basin setting. Numerous suites of syn- and post-tectonic highly fractionated monzogranites, syenogranites, discrete gabbro bodies and dolerite dikes intruded the Telfer region during the Paterson Orogeny (Goellnicht, 1992; see Section E for description; see fig. 1 - Section E and

Map 2 for distribution). Associated with the granitoids were local contact aureoles that commonly overprinted the regional metamorphic assemblage (Goellnicht, 1992).

One of the anticlinal domes, the Telfer Dome, is host to the Telfer deposit (≈ 8 Million ounces - Sexton, 1994) with mineralisation occurring as a series of stratabound quartz-carbonate-sulphide reefs hosted along the limbs of two lower order sub-domes, Main and West Dome (Fig. 1A; see Section D for a full description of the geology of the Telfer deposit). Initial study of the Telfer system proposed a syngenetic origin (Turner, 1982). However, further research and deposit development has indicated an epigenetic genesis that is typified by strong structural controls on veining and proposed genetic links with regional granitoids (Goellnicht, 1987; Goellnicht et al., 1989; Rowins, 1994). Goellnicht et al. (1989) favoured a direct magmatic influence, suggested by the Au-Cu-As-Bi-Co (± Ni, Pb, Mo) elemental association and Pb isotope composition of some ore sulphides, whilst Rowins (1994) considered the deposit to have formed distal (5-10kms) to cooling granitoids whose primary role was to circulate ore-fluids through the sedimentary sequence.

Previous studies of mineralisation timing interpreted it to be synchronous with dome formation (Hill, 1989; Windh, 1991; Rugless, 1993) as suggested by the formation of reefs along the domal limbs. Microstructural investigation indicated that mineralisation assemblages overprinted the domal axial plane cleavage (van Dijk, 1986 - Table 1; Goellnicht et al., 1989) and thus were either late- in or post-dated dome formation. Goellnicht et al. (1989) noted that muscovite alteration, coeval with sulphide-gold mineralisation, had grown aligned parallel to an undefined cleavage that overprinted the domal axial plane foliation, suggesting a possible post-D_2 mineralisation timing.

EVENT	HOSTROCK	VEINING
Syn D_2	Albite porphyroblasts	Minor albite veins
Late to post-D_2	Recrystallisation of albite porphyroblasts	
(Late to) post-D_2	Quartz/sericite/epidote replacement of albite. Carbonate replacement of albite and host rock.	
↓		Carbonate (breccia) veining. Quartz veining (reopening). Quartz replacement of carbonate.
↓	Pyrite replacement of albite, carbonate and hostrock.	
↓		Pyrite replacement of carbonate/quartz.
↓	Mica replacement (eg quartz-albite near pyrite. Mica recrystallises. Mica growth with [001] parallel to pyrite crystal faces.	

Table 1 : Summary of hydrothermal veining/alteration paragenesis identified by van Dijk (1986) in the Telfer system, based on a microstructural study of diamond drill-core from Middle Vale Reef mineralisation in Main Dome.

METAMORPHIC MINERALOGY/TEXTURES

Regional metamorphism is sub-greenschist facies (chlorite zone - Ferry, 1983; Miyashiro, 1994) with a quartz, albite, sericite, muscovite, chlorite, epidote and ferroan dolomite assemblage. Trace amounts of biotite, also noted by Turner (1982) and Rowins (1994), suggest that metamorphic conditions may locally have approached the biotite isograd (≈400°C - Ferry, 1984, 1986). The preservation of graphite inclusions in some of the reefs, coupled with the absence of Fe-oxides, indicates that metamorphic conditions were both low fO_2 and low sulphidation (a_s). The effects of the overprinting contact metamorphism, observed in many regional granite aureoles (Dimo, 1990; Rowins, 1994), have not been recorded in the Telfer Mine sequence.

Metamorphic recrystallisation accompanied deformation of the rocks and was coincident with the formation of S_1 and S_2 cleavages, although in low strain zones preservation of sedimentary structures is common. Some ferroan dolomite, epidote and chlorite formed in equilibrium with metamorphic recrystallisation of the matrix. However, the majority of these phases accompanied subsequent hydrothermal veining that overprinted regional metamorphic effects/textures. Fine-grained syn-metamorphic granular and fibrous quartz/quartz-albite veinlets, with fibres parallel to both S_1 and S_2 foliations, pre-date hydrothermal veining and alteration.

MINERALISATION VEINING/ALTERATION PARAGENESIS

Mineralisation related veining/alteration comprised one major hydrothermal event that was congruent in reef systems across both Main and West Domes (Rowins, 1994) and which shows a progressive mineral paragenesis. Three key suites that are representative of different reef/stockwork mineralisation styles in Main Dome were examined (Fig. 1B).

Suite 1 - Middle Vale Reef / Middle Vale Siltstone

The first veining phase, which produced the main body of the Middle Vale Reef (MVR), comprises infilling and replacive massive quartz that accompanied pervasive silicification of the MVR horizon. The quartz, which is strongly deformed, contains numerous, fine-grained, silty inclusions (Fig. 2) as well as small grains of ferroan dolomite, tourmaline and muscovite, and sporadic wallrock clasts. The abundance of silty inclusions, which preserve bedding (*e.g.* fig. 19 - Section D), suggest that much of this phase was replacive. Subsequent deformation, resulting in brecciation and dynamic recrystallisation of the massive quartz was accompanied by the first mineralisation phase. This comprised coarse-grained pyrite aggregates (±chalcopyrite - now altered to digenite) containing fine-grained inclusions of muscovite/tourmaline, that infilled the fractured/recrystallised quartz (Fig. 2), and which were accompanied by dolomite alteration (Fig. 2).

In the surrounding wallrocks, coarse-grained ferroan dolomite-quartz-albite-sulphide veining (the main-stage veins of Goellnicht, 1987), with sericite-muscovite-dolomite selvedges and limited carbonate-sericite-disseminated sulphide alteration halos (Fig. 3), overprint matrix fabrics and were synchronous with reef deformation. Associated with these is a pervasive matrix alteration comprising rhombic ferroan dolomite porphyroblasts (Fig. 4), tourmaline and rare biotite. Coarse-grained pyrite infilled/replaced the veins along dolomite-quartz grain boundaries (Fig. 3). This occurred late in the dolomite-quartz veining episode and was accompanied by continued quartz deposition in fractures and re-opened veins. Radial muscovite growth on sulphide crystal faces, and large pyrite porphyroblasts, overprint matrix fabrics.

Further mineralisation in the MVR occurred along the upper margins associated with continuing layer-parallel deformation. This resulted in strong quartz recrystallisation, evident as polygonal mosaics, as well as fracturing of sulphide grains and numerous fibrous quartz overgrowths (Fig. 5). A second phase of finer grained euhedral pyrite and granular quartz precipitated/recrystallised during this deformation, and was associated with minor infilling quartz-chlorite-calcite veining, and trace amounts of scheelite. Wallrock margins and clasts within the reef, already altered to carbonate-chlorite-muscovite, were further silicified and pyritised as the system became silica dominant. Late cryptocrystalline chalcedonic veinlets and vuggy infill, overprint earlier phases and are spatially/texturally associated with argillic veining in the wallrock.

Metasomatic Veinlets

Small ovoid quartz (± albite-dolomite) aggregates, containing helicitic cleavage textures (Fig. 6) and a progressive infill paragenesis similar to that in the reefs, occur throughout the matrix, and are important timing criteria (see below). These grow along planar anisotropies including bedding, S_1 and S_2 cleavage; the latter being the most common. Less commonly they comprise large individual quartz grains or granoblastic quartz-albite aggregates with diffuse boundaries suggestive of metasomatic replacement. The primary quartz-albite-dolomite assemblage, which is similar to the country-rock veining phase (main-stage - see above), was progressively infilled/replaced by dolomite/calcite followed by replacive sulphide (Fig. 6).

The aggregates were previously interpreted as syn-metamorphic albite porphyroblasts, that were subsequently recrystallised to quartz (van Dijk, 1986; see Table 1). However, there is no textural evidence indicating relict albite grains. Instead, the presence of fine-grained mica inclusion trails parallel to matrix foliations, coupled with the replacive nature and growth along differently oriented matrix anisotropies and progressive mineral paragenesis suggests they are a metasomatic feature resulting from pervasive fluid infiltration through the rock matrix (*e.g.* McCaig, 1987; Forde & Davis, 1994).

Suite 2 - M-Reefs

The M-Reefs, hosted within the Malu Quartzite Formation (Fig. 1), are newly discovered reefs that exhibit a range of compositions. The M10 and a subsidiary stringer (M12) contain a massive quartz veining phase, similar to the MVR, which is inclusion-rich, contains small clasts of wallrock and fine-grained tourmaline/biotite inclusions. This quartz phase was dynamically recrystallised during subsequent deformation accompanying dolomite incursion (see below). In contrast, the M8 (a 2-5cm thick concordant vein) and M30 reefs lack a massive quartz component, instead comprising coarse-grained grey ferroan dolomite/ pink dolomite/carbonate infill veining, although rarely small remnants of massive quartz vein are preserved within the dolomitic reef matrix.

The first carbonate phase comprises a grey, highly ferroan, coarse-grained rhombic dolomite that accompanied deformation of reef quartz in the M10/M12 reefs. This was introduced during and after dynamic recrystallisation along the reef margins (Fig. 7) and was coeval with pervasive dolomite matrix alteration, and coarse-grained dolomite-albite-quartz wallrock veining. In the M8/M30 reefs this phase exhibits coarse-grained euhedral aggregates (Fig. 8) that precipitated during multiple veining phases that accompanied strong host-rock brecciation, suggestive of regular fluid pressure fluctuations in the system (*e.g.* Robert & Brown, 1986; Sibson, 1987, 1992; Boullier & Robert, 1992).

A second carbonate phase, comprising a white-pink coloured ferroan dolomite infilled interstices in the earlier grey dolomite aggregates (Fig. 9) and fractures in the massive quartz reef host. This coloured dolomite phase also intergrew with, and replaced (Fig. 9), the grey dolomite commonly pseudomorphing large rhombs. The second dolomite phase was accompanied by chlorite-calcite-quartz-muscovite infill and alteration of dolomitised wallrock clasts (Fig. 10) with renewed wallrock silicification (Fig. 8). Accessory muscovite, tourmaline and scheelite are also coeval with this phase, commonly as fine-grained selvedges along reef margins that overprint incipient grey dolomite alteration.

Sulphide mineralisation, including pyrite and lesser chalcopyrite, replaced and infilled the first dolomite phase (grey) along grain boundaries and fractures and precipitated broadly coeval with the second dolomite (opaque) veining event. Chalcopyrite precipitated slightly later, coeval with calcite-chlorite-muscovite veining and silicification. This was also accompanied by gold mineralisation and coarse-grained muscovite growth, particularly along sulphide grain margins. Continued deformation along the upper M10 margin deformed and recrystallised pre-existing vein minerals (quartz and both dolomite phases) and was accompanied by further siliceous veining that commonly formed fine-grained fibrous overgrowths on sulphide grains. A second phase of finer grained idiomorphic pyrite, with associated gold mineralisation and accessory scheelite, chlorite and calcite, was coeval with this late silica veining. Scheelite, chlorite and calcite also accompanied multiple episodes of brecciation along the reef margins (see also Taylor & Myers, 1995).

Suite 3 - Telfer "Deeps"

The "deeps" suite was sampled from fold-hinge mineralisation that is commonly hosted in massive zones of brecciation in the vicinity of the I-Reefs (Fig. 1B). This brecciation/veining forms the major hydrothermal event identified in the "deeps". The initial veining phase comprises grey coarse-grained ferroan dolomite that infilled breccia interstices immediately following accessory epidote precipitation (Fig. 10). This was accompanied by pervasive dolomite-sericite-brown tourmaline-epidote wallrock alteration (Fig. 11), with lesser muscovite and quartz infill/alteration. Repeated dolomite-quartz veining occurred in re-opened veins and overprinted earlier cycles. A second carbonate phase comprising opaque white-pink ferroan dolomite replaced/infilled grey dolomite aggregates (Fig. 11). The opacity of this later dolomite is due to numerous fine-grained silty inclusions within the grains.

The opaque dolomite was accompanied by calcite (± chlorite) veining and alteration, and further siliceous veining/alteration that overprinted earlier wallrock alteration (Fig. 12). Small crack-seal (Ramsay, 1980; Cox & Etheridge, 1983) openings within the larger carbonate aggregates exhibit calcite-muscovite cores (broadly coeval with the second opaque carbonate) overgrown by fibrous quartz, indicating that silica veining was continual throughout the "deeps" paragenesis. In stockwork zones away from the reefs a similar paragenesis is observed although the major dolomite-sericite alteration phase is locally overprinted by late hematite alteration associated with chalcopyrite mineralisation.

Disseminated fine-grained and coarse-grained pyrite aggregate growth occurred late in the grey ferroan dolomite phase along grain margins and interstices and was broadly coeval with the second opaque dolomite-calcite-chlorite event. Chalcopyrite formed slightly later commonly replacing grey ferroan dolomite (Fig. 13), and disseminated sulphide mineralisation accompanied late- siliceous veining (Fig. 12). Sulphide mineralisation throughout the "deeps" was accompanied by random muscovite growth, particularly as a pervasive wallrock alteration along sulphide reaction fronts (Fig. 14).

Sulphide and Gold Mineralisation

The primary sulphide mineral assemblage consists of pyrite, chalcopyrite and galena with trace amounts of pentlandite, pyrrhotite, cassiterite (Rugless, 1993) and sphalerite (Rowins, 1994). Sulphide deposition occurred late in the grey ferroan dolomite vein phase (see above), broadly coeval with the second opaque dolomite-calcite and silica veining phases.

Sulphide Paragenesis

Pyrite, the earliest sulphide, occurs as an infill/replacement phase of ferroan dolomite and quartz reefs (*e.g.* Figs 2 & 13). Larger euhedral pyrite grains have small inclusions of chalcopyrite, galena, pentlandite, pyrrhotite and gold (see further) indicating that trace quantities of these minerals were present at the beginning of pyrite deposition. In the MVR, large euhedral and rounded pyrite grains and aggregates precipitated early, followed by a second phase of finer-grained pyrite, with chalcopyrite, that accompanied the late siliceous veining along the reef margins.

Galena occurs as an accessory sulphide and grew coeval with, and slightly later than, pyrite as subhedral overgrowing aggregates (Fig. 15). These were subsequently replaced by chalcopyrite (Fig. 16). However, galena occasionally overgrew chalcopyrite, and in rare cases the two intergrew, indicating that galena deposition spanned the sulphide paragenesis. Chalcopyrite, apart from fine-grained inclusions in large early-formed pyrites and galena aggregates, precipitated later than pyrite and infilled interstices and grain fractures. Anhedral chalcopyrite overgrew the first phase of coarse-grained euhedral pyrite (Fig. 17), replacing both dolomite phases (*e.g.* Fig. 13). Small euhedral pyrite grains, contained within larger chalcopyrite aggregates, indicate that pyrite deposition continued throughout the paragenesis.

Gold Paragenesis

Gold occurs within or in association with the following phases;

1. As fine-grained inclusions (*e.g.* Turner, 1982; Goellnicht et al., 1989) and along hairline fractures within large early-formed pyrites particularly those in the MVR. Gold is also contained within quartz-chlorite-calcite veining in fractured pyrite, and in recrystallised primary (massive inclusion-rich) quartz of the MVR. In the latter, gold has precipitated within healed fractures indicating that it was coeval with reef deformation.
2. Small gold grains are also associated with the second mineralisation phase, which comprised sulphide aggregates and fibrous quartz overgrowths, that formed in

response to renewed deformation along the upper margins of the MVR.

3. Gold in the M-Reefs is associated with late calcite-chlorite-quartz-chalcopyrite veins that infilled fractures in the quartz-dolomite host (Fig. 18) and were associated with the second opaque dolomite phase. Similarly, as for the MVR, gold is also associated with renewed deformation and further silica-sulphide ingress along the upper M10 margins.
4. In the "deeps" visible gold occurs in massive calcite-carbonate-chlorite infill zones that correspond with the first major sulphide veining phase, and a strong spatial association to chalcopyrite mineralisation has been noted.

INTERPRETATION

The three Main Dome suites exhibit a similar hydrothermal veining/alteration paragenesis. This is progressive (Fig. 19) and comprises five distinct phases (Table 2), three of which are associated with sulphide-gold-copper mineralisation in the reefs. Within each phase, veining commonly exhibits two to four individual stages, a feature common in mesothermal deposits and suggestive of cyclical fluid pressure fluctuations (Sibson et al., 1988; Boullier & Robert, 1992), and many of the paragenetic stages are associated with deformation.

Phase	*Veining*	*Alteration*
1	Massive inclusion-rich reef quartz veining, moderately to strongly deformed	Wallrock silicification
2	Ferroan dolomite, quartz, albite, ± muscovite, epidote, sericite - late sulphides	Sericite, Fe-dolomite, muscovite, epidote. Late sulphides
3	Pink/white dolomite (ferroan), calcite (Mg-rich), chlorite, muscovite, quartz, sulphide (pyrite, chalcopyrite) ± scheelite, tourmaline	Chlorite, calcite, sulphide, muscovite, silica
4	Repeated silica veining (MVR + M-Reefs), ± disseminated sulphide, ±scheelite, tourmaline, calcite, chlorite	Silicification
5	Chalcedonic, argillic, (calcite-hematite in the "deeps")	Argillic/ hematitic, minor sulphides, calcite

Table 2: *Summary of the successive vein-alteration stages and their component mineralogy across the three suites examined from Main Dome. The shaded areas represent the major mineralisation-related phases.*

Structural Timing of Phase 1 Quartz Veining.

Phase 1 quartz veining with associated pervasive silicification forms the main body of many of the Telfer reefs (*e.g.* MVR, M10, M12), and was strongly deformed prior to, and during, the commencement of mineralisation (Phases 2, 3 & 4). Replacive quartz along reef margins has rarely preserved S_1 cleavage as fine-grained inclusions (Fig. 20A), and the inclusion-rich nature of the quartz is similar in style to that observed in laminar bedding-concordant quartz veins in the mine sequence that formed syn-D_2. Both these features suggest a syn-D_2 genesis for Phase 1 veining.

The preservation of strain markers within the MVR (elongate carbonate spots and rotated S_2 cleavage) indicates that shear strain was partitioned into the reef horizon (Fig. 20B; see also Vearncombe & Hill, 1993 - fig. 3), and this appears to have occurred prior to silicification/quartz veining. This is because such silicification should preserve the carbonate spots from being markedly strained, as well preventing strong rotation of the cleavage (Fig. 20C) in a manner analogous to a porphyroblast preserving a foliation that has been deformed/destroyed in the matrix of a rock (*e.g.* Spry, 1969; Yardley et al., 1990). In the case of the MVR, silicification would have produced a competent horizon against which subsequent ductile shearing strain preferentially partitioned on the margins (Fig. 20C). However, the internal elements of the horizon are strongly deformed, suggesting that the majority of silicification/quartz veining occurred after the partitioning of penetrative shearing strain into the reef horizon (Fig. 20D).

Penetrative shearing within the limbs of the fold would have approximated flexural flow deformation and in many folds such strain is commonly partitioned into mechanically weaker horizons (Bayly, 1992). Therefore, Phase 1 quartz veining and silicification occurred after flexural flow deformation. The emplacement of the quartz may have coincided with the more brittle process of flexural slip deformation that generally occurs late in the development of a fold when the limbs reach orientations steep enough to accommodate discrete movement along individual bedding surfaces and bedding-parallel faults (Tanner, 1989, 1992; Cosgrove, 1993). Such deformation is commonly accompanied by elevated fluid pressures, due to stick-slip movement along individual bedding planes, and this would have assisted the dilation of reef horizons (*cf.* Goellnicht et al., 1989).

Structural Timing of Hydrothermal Veining (Phase 2-Phase 4)

The main stage mineralisation related veining (Phases 2, 3 & 4) commenced late pre- post-D_2. This is indicated by alteration selvedges, sulphide and dolomite porphyroblasts (*e.g.* Fig. 4) and muscovite growth, associated with sulphide mineralisation that overprint S_2 (Fig. 21A). Rare breccia clasts of cleaved wallrock (S_2) within veins are rotated (Fig. 21B; van Dijk, 1986; C. Switzer, pers. comm, 1995) also indicating a late- to post-D_2 vein timing. Further microstructural constraint is possible from the small metasomatic aggregates that occur in the Middle Vale Siltstone (Suite 1), which preserve a rudimentary vein paragenesis.

Structural Timing of Metasomatic Veinlets/Aggregates

Larger quartz grains along the margins of the aggregates preserve S_2 as fine-grained sericite inclusion trails (*e.g.* Fig. 6) indicating growth late in or post-D_2. However, these trails rarely have a weak curvature within either end of the aggregate (Fig. 21C) that corresponds to a weak seamy cleavage in the matrix, oriented at a high angle to S_2 (Fig. 21C), similar to that observed by Goellnicht (1987; 1989). Although the only samples of this come from unoriented drill core, the high angular relationship suggests that this overprinting fabric is S_3, and subsequently the weak asymmetry is due to a coarse D_3 crenulation. The preservation of this by the aggregates therefore suggests aggregate growth occurred early in D_3.

Similarly, the aggregates also rarely grow aligned parallel to a weak S_3 crenulation cleavage (Fig. 22, 21D), suggesting a syn-D_3 growth. In these examples, aggregate growth post-dates active S_2 formation, as an earlier formation is likely to have been prevented by shearing and dissolution along the S_2 cleavage (*e.g.* Bell & Cuff, 1989; Bell & Hayward, 1991). Although this does not preclude a very late D_2

growth, the S_3 crenulation cleavage geometry around the aggregates is strongly suggestive of an early D_3 timing for the initial quartz-albite assemblage. Therefore subsequent dolomite-calcite infill/replacement, leading to sulphide alteration (Fig. 21D), coeval with Phases 2 & 3 veining, must occur syn- to post-D_3 deformation.

Once formed, the aggregates underwent further deformation that caused a weak wrapping/anastomosing of pre-existing S_2 around their margins (Fig. 21E), tensional cracking/fracturing (Fig. 21E) and fibrous quartz overgrowths on either end (Fig. 21F). These features are consistent with renewed shortening across the sub-vertical S_2 cleavage (*e.g.* Bell & Rubenach, 1983), which, given a D_3 growth, is likely to have occurred during D_4. This is because D_4 is similarly oriented to D_2, and is thus likely to reactivate/re-use (Bell, 1986; Davis & Forde, 1994; Davis, 1995) the S_2 foliation. Calcite (+ dolomite) infill and quartz addition accompanied this deformation of the aggregates, and sulphides replaced this and the aggregates (Fig. 21G) indicating that veining/mineralisation continued through D_4.

Macrostructural Constraints on the Timing of Phases 2, 3 & 4 Veining.

Phase 2 veining in the "deeps" accompanied massive brecciation that was concentrated in the dome hinge. Brecciation appears to have developed synchronously with refolding and overturning of the D_2 fold hinge, which is interpreted to have occurred during D_3 deformation (Fig. 23; see also Section D). A similar pattern of increased brecciation and better developed mineralisation on steeper dipping NE dome limbs is common in many of the Telfer reefs, and is consistent with the favourable orientation of these limbs for gaping/dilation during D_3 (Fig. 23; see also Section D). This is particularly well illustrated by the M-Reef series, which were best developed in the hinge and NE limb of Main Dome, but were poorly developed, or non-existent, on the SW limb.

Structural Timing of Sulphide/Gold Mineralisation.

Sulphide mineralisation was introduced from late-Phase 2 through to Phase 4 (Table 2) veining. Given the timing constraints on veining/alteration to D_3/D_4, the timing of sulphide mineralisation in the Telfer deposit is interpreted to have commenced during D_3, as indicated by sulphide replacement of the syn-D_3 metasomatic aggregates. Sulphide mineralisation continued through D_4 as indicated by the replacement of D_4-related calcite in the metasomatic aggregates. The first sulphide phase (syn-late D_3) was deformed during deformation of the upper reef margins that accompanied Phase 4 veining. This deformation comprised layer-parallel shearing, which is interpreted to be due to slip/shear on reactivated bedding as a result of further tightening and re-use of the D_2 Telfer Dome during D_4 deformation. The second pyrite-silica (Phase 4) veining, with associated gold mineralisation, precipitated during the renewed deformation and was thus syn-D_4.

Structural Timing of Gold-Copper Mineralisation.

Gold mineralisation accompanied the early sulphides (Phase 3) and later reef deformation (Phase 4,5) stages, and is thus structurally timed as syn-D_3 through to, and including, D_4. This is consistent with macrostructural relationships that indicate a strong association between mineralisation and both D_3 and D_4 structures (Fig. 23). These include the asymmetry of reef development (see above) and the development of sulphide-gold-copper mineralisation within large right-stepping D_4 shear zones (West Dome Deeps - Fig. 23; see Section D). Middle Vale Reef mineralisation is truncated by the Graben Fault (Fig. 23), interpreted to be a D_5 structure, indicating that the development of mineralised reefs ceased prior to D_5.

D5-related Mineralisation - Implications for an Overprinting Epithermal Mineralisation Event

There is evidence that mineralisation also occurred during D_5. In Main Dome, dolomite-sulphide infill is present in fractures associated with the late-D_5 Graben Fault

(Fig. 23). There is also an association between mineral prospects and late-D_5 (D_{5b}), N-S trending lineaments/dolerite dikes. Structural analysis suggests that D_5 post-dated the Paterson Orogeny (Section B) thus implying that a second and separate mineralisation event may have occurred. This is consistent with a very late/overprinting hematitic alteration observed in the "deeps" ore-system that suggests a marked change in geochemical conditions, an observation also made by Taylor & Myers (1995). There is evidence that this event may be epithermal, and this includes the presence of large, apparently discordant, calcite bodies in drill core from the "deeps". These bodies contain spectacular visible gold and comprise multi-coloured calcite that exhibits crustiform/colloform vein textures, typical of open space infill and suggestive of higher crustal level epithermal conditions (Dowling & Morrison, 1989; Vearncombe 1993).

Given the variety of igneous intrusion across the Telfer region (*e.g.* Goellnicht, 1992), much of which is concealed, it is possible that some intrusions were high-level and post-dated the Paterson Orogeny. These could have produced an epithermal mineralising event that overprinted stratabound reef mineralisation. Whilst the evidence for a second mineralisation event is largely anecdotal, it has an important implication for further paragenetic/geochemical studies of the deposit, as failure to consider it, and sample accordingly, may lead to ambiguous and wide-spread data that produce conflicting indications of deposit genesis.

DISCUSSION

Constraints on Genetic Models for the Telfer Deposit.

The Telfer deposit represents an ore-system that seemingly comprises both magmatic and metamorphic/connate brine fluid components (Goellnicht et al., 1989; Rowins, 1994). On the one hand, the elemental Au-As-Cu-W-Bi-Mo-Co association and Pb isotope composition of ore sulphides coupled with a high salinity ore-fluid component (Goellnicht, et al., 1989) and a spatial relationship to Cu (± Pb, Zn, Au) porphyry and Au/Cu-W-Mo-Sn-Pb-Zn skarn mineralisation styles and calc-alkaline granites (Goellnicht et al., 1991) suggests a strong magmatic control. However, the deposit also exhibits many characteristics typical of metamorphic dominated greenstone/slate-belt hosted lode-gold deposits, such as a late mineralisation timing, similar metamorphic and veining/alteration mineral facies, and a strong inter-relationship between deformation and veining, the latter exhibiting textures characteristic of a mesothermal crustal setting (*cf.* Vearncombe, 1993).

More recent studies have down-played the role of regional metamorphism in the formation of Telfer, commonly because of a perceived lack of it (*e.g.* Goellnicht, 1987, 1992). Instead, these studies have inferred, on the basis of fluid inclusion and isotope data, that Telfer formed through the mixing of cool connate brines with hotter magmatic fluids (Goellnicht, 1987; Goellnicht et al., 1989). However, subsequent studies, including this one, indicate that the Telfer ore-fluid was in long lived contact and equilibrated with the regional sedimentary sequence (Rowins, 1994; Taylor & Myers, 1995) prior to entering the deposit. This is evident from:

1. Oxygen isotope data (Fig. 24, Table 3) that are consistent with pervasive infiltration of hydrous magmatic fluids into a regional calc-pelitic sequence (*e.g.* Ferry, 1983, 1986; Symmes & Ferry, 1991) with a characteristic shift in $\delta^{18}O$ values (Fig. 24). The data are also consistent with a long-lived evolution of connate brines in contact with marine carbonate units (Sheppard, 1986), and the tightly clustered field indicates that the ore-fluid was homogeneous.
2. The predominance of ferroan dolomite in the vein paragenesis, which forms under lowered Ca/Mg ratios, suggests that any connate brine component had a long-lived evolution, and indicates that the ore-fluids had a high xCO_2 content (Powell et al., 1984; Will et al., 1990) probably resulting from increased fluid participation in metamorphic decarbonation reactions (*e.g.* Kerrich & Fyfe, 1981).
3. Isotope data suggest that the majority of carbon, oxygen, sulphur and lead elements were derived from extensive fluid circulation and scavenging through the host sedimentary sequence (Rowins, 1994).

4. The similar mineral facies for both hydrothermal mineralisation-related veining and country rock metamorphic assemblages, identified in this study, is typical of fluids that have been in long-lived contact and reached equilibrium with the metamorphosing host-rocks (Norris & Henley, 1976; Yardley, 1983; Rumble, 1994).

ISOTOPE	*VALUES*
$\delta^{34}S$ **sulphide**	0 to 9.3 ‰ (Main Dome) -5 to 8 ‰ (West Dome)
$\delta^{13}C$ **carbonate**	-2.2 to 3.8 ‰ (Main Dome) -2.4 to 2.7 ‰ (Telfer Syncline) 2.5 ± 0.3 ‰ (primary carbonate - Puntapunta Fm.) * 0.1 ± 1.0 ‰ (primary carbonate - Isdell Fm.) *
$\delta^{18}O$ **carbonate**	16 ‰ (Main Dome)
$\delta^{11}B$ **tourmaline**	-15.4 to -12.6 ‰ (Hydrothermal tourmaline)

*Table 3: Summary of isotope values for mineralised veins/reefs in selected areas of the Telfer Deposit (after Rowins et al., 1992; Rowins, 1994; * = data from Goellnicht & McNaughton, 1989)*

Although previously unconsidered, a regional metamorphic (Yardley, 1983; Rumble, 1994) fluid component is compatible with many features of the ore-system. These include oxygen isotope data that lie in the metamorphic fluid field (Sheppard, 1986), the long-lived contact of the ore-fluid with the rock sequence and silica metasomatism, coupled with syn-metamorphic quartz veining, indicative of metamorphic fluid migration (*e.g.* Norris & Henley, 1976; Sawyer & Robin, 1986; Yardley & Bottrell, 1992) prior to mineralisation. Additionally, the sub-greenschist facies (chlorite) metamorphic grade is comparable with shallower mesothermal lode-gold mineralisation commonly in Archean terranes (*e.g.* Groves & Phillips, 1987; Colvine, 1989; Mueller & Groves, 1991; Mickuci & Ridley, 1993).

Significant metamorphic fluid generation in a rock sequence can occur both regionally (Gray et al., 1991; Cartwright et al., 1994) and from deeper crustal levels (Walther & Orville, 1982; Etheridge et al., 1983; Wood & Walther, 1986). Such fluids commonly infiltrate shallower crustal levels after peak metamorphism (*e.g.* Scrigemour & Sandiford, 1993). For the Telfer region, a metamorphic fluid component could have been sourced from devolatisation reactions in the regional rock sequence, coupled with potential influx derived from collisional orogenesis that may have occurred further NE of the region (Section B). Once at the mid- to shallow crustal levels presently exposed, extensive circulation and mixing of this fluid with magmatic and connate brine fluids is likely to have been enhanced by convective cooling (*e.g.* Norton & Knight, 1977; Fehn, 1985; Neiva et al., 1995) around the granitoids that intruded the Telfer region. As a consequence the composition of an original metamorphic fluid could have been modified through the addition of magmatic volatiles and elements, in particular Pb, Zn, and W, the latter producing scheelite in the Telfer ore assemblage.

Following the mixing and homogenisation of separate fluid phases in the Telfer region, pervasive movement of the resultant phase through the rock sequence would cause it to attain near-equilibrium with the country rocks. This would have occurred prior to its late release into trap-sites in the Telfer deposit. Consequently, the ore-fluid, regardless of its composition, would approximate in behaviour a true metamorphic fluid and therefore its release from the rock sequence late in the deformation history can be considered in terms of structural processes that are applicable to metamorphic terranes (see further). Additionally, a metamorphic fluid implicates potential mantle and deep crustal sources for gold (*e.g.* Kerrich & Fyfe, 1981; Meyer & Saager, 1985; Fyon et al., 1989; Goldfarb et al., 1989; Groves, 1993). Such a scenario is consistent with recent indications that collisional/accretionary orogenesis occurred within the Paterson Orogen (Goellnicht et al., 1991; Myers & Barley, 1992; Section B).

Genetic models for Proterozoic Cu-Au, Cu (±Au) and Au (±Cu) deposits, including Telfer, point to the strong spatial association between regional granitoids and mineralisation (Goellnicht, et al., 1989; Oreskes & Einaudi, 1992; Huston et al., 1993; Davidson & Large, 1994; Phillips et al., 1994). In view of a potential metamorphic genesis, such an association in the Telfer region may reflect a tendency for deep crustal auriferous metamorphic fluids to have risen through conduits utilised by the magmas.

This would provide an ideal way for a metamorphic fluid to gain magma-related elements. Alternatively, given the relatively shallow crustal levels, auriferous metamorphic fluids may have been much more effectively concentrated into the Telfer region through convective circulation caused by the presence of cooling igneous bodies.

Both recently proposed epigenetic models for the Telfer deposit consider it to have formed at a shallow-crustal setting in which magmatic-meteoric and/or formational waters combined to provide an ore-fluid (Goellnicht et al., 1989; Hall & Berry, 1989; Rowins, 1995; Taylor & Myers, 1995). However, a more appropriate genesis may be that related to a magmatic-metamorphic environment (*e.g.* Linnen & William-Jones, 1995), which appears to be common at deeper crustal levels. These levels, characterised by mid-crustal igneous intrusions, exhibit tin, molybdenum and tungsten deposits whose formation involved both magmatic and metamorphic fluid components (William-Jones et al., 1989; Kamilli et al., 1993; Linnen & William-Jones, 1994). Ore-fluid studies in these settings reveal multi-variable and co-existing phases (Linnen & William-Jones, 1994) with similarities to those observed for Telfer (Goellnicht et al., 1989). Such environments, given the input of a regional metamorphic fluid component, would also be highly conducive to the formation of typical metamorphic-induced mesothermal lode-gold deposits.

Constraints on Depositional Mechanisms

The results of this study indicate that some depositional mechanisms previously proposed for the Telfer deposit, including surface adsorption or reduction both caused by reaction with syngenetic pyrite or activated carbonaceous complexes (Goellnicht et al. 1989), are not applicable. Large rounded pyrite aggregates in the Middle Vale Reef were previously interpreted to be syngenetic (Turner, 1982; Goellnicht et al., 1989). This was because they were deformed and thus appeared to pre-date domal deformation. However, the textural relationships identified in this study, coupled with an increased knowledge of post-domal deformation, now indicates that these are epigenetic and associated with the first sulphide deposition (Phase 2-3; *e.g.* Fig. 5). Limited wallrock alteration haloes along reef margins also suggest that gold depositional mechanisms involving the country rocks, such as wallrock sulphidation (Mickucki & Ridley, 1993), were unlikely.

The high salinities of, and presence of copper in, the ore-fluids suggest that gold was probably transported as chloride complexes (Henley, 1973; Goellnicht et al. 1989; Davidson & Large, 1994). However, the paucity of iron oxides (hematite, magnetite) coupled with only minor pyrrhotite in the ore-assemblage indicates that Eh-pH conditions were close to the chloride-reduced sulphur complex boundary (*cf.* Fig. A8 of Romberger, 1988) and consequently, gold may also have been transported by reduced sulphur complexes (*e.g.* Seward, 1973). Depositional mechanisms for gold transported by chloride complexes include temperature and fO_2 reduction (Davidson & Large, 1994) or an increase in fluid pH (Romberger, 1988; Cuff & Doherty, 1989; Hayashi & Ohmoto, 1991). Temperature reduction and/or ore-fluid pressure fluctuations is supported by the variable fluid inclusion data (Goellnicht, et al., 1989) and extensive early silica veining (*e.g.* Rimstidt & Barnes, 1980). However, the prominence of dolomite in the paragenesis, which has a negative solubility coefficient (*e.g.* Fyfe et al., 1978; Kerrich & Fyfe, 1981), implies specific precipitative conditions, such as wallrock carbonation (Kerrich & Fyfe, 1981) or fluid immiscibility (Craw, 1992; Craw et al., 1993; Cartwright et al., 1994). The absence of extensive wallrock alteration indicating that the reefs were internally buffered suggests that the latter (immiscibility) was more likely.

Fluid immiscibility, with resultant boiling during pressure fluctuations, would have been enhanced by methane (Mernagh & Witt, 1994) that was present in the Telfer ore-fluids (Goellnicht et al., 1989). Such immiscibility provides an explanation for the variable and apparent highly saline character of the ore-fluids identified by Goellnicht (1987). Instead of representing the mixing of two fluids, as proposed by Goellnicht et al. (1989), the data may instead represent unmixing (*e.g.* Pichavant et al., 1982; Bowers & Helgeson, 1983; Bodnar et al., 1985; Robert & Kelly, 1987) of the ore-fluid during immiscibility into two or more separate phases. The presence of late argillic veining in wallrocks, indicative of sulphate production, is also consistent with fluid boiling/immiscibility (Henley, 1973). Such immiscibility releases CO_2 from the ore-fluid, causing a pH increase in the remaining ore-fluid that promotes gold deposition from chloride complexes. Excess CO_2 released from the fluid would also have promoted carbonate-dolomite alteration.

The Role of Polyphase Orogenic Deformation in Crustal Fluid Flow and Implications for Late Mineralisation Timing.

Late mineralisation timing is a common and defining feature of mesothermal lode-gold deposits (Barley. et al., 1989; Forde, 1991; Groves, 1993; Kerrich & Cassidy, 1994). In metamorphic environments this may arise when a time-lag occurs for volatile supply, from deep crustal sources, to the upper crust (see Kerrich & Cassidy, 1994 for a review) or through post-peak metamorphic fluid/volatile release resulting from upper crustal processes. Such processes include tectonic uplift (Norris & Henley, 1976; Craw, 1986), changing tectonic convergence (*e.g.* Kerrich & Wyman, 1990; Elder & Cashman, 1992) or permeability capping (Etheridge et al., 1983); the latter producing supra-lithostatic fluid pressures that breach the rock sequence.

In the case of Telfer, where an association with actively cooling granitoids is implied, an alternative explanation for the late timing of mineralisation could be related to the late- to post-D_2 emplacement of some regional granitoids (see Section E for summary of granitoid timing). However, the long lived circulation of ore-fluids, coupled with Telfer's distal (5-10kms) position to regional granitoids (Rowins, 1994) as indicated by the absence of contact aureole metamorphism in the mine suggests that the magmatic influence was minor thus precluding a direct input from a nearby late-intruding granite. Similarly, processes such as tectonic uplift and permeability capping are difficult to demonstrate for the Telfer region/Paterson Province. Instead, the supply of gold/copper-bearing fluids into the Telfer deposit was closely related to the development of discrete brittle-ductile structures and deformation of the host structure, as evidenced by the ongoing deformation of veining stages in the mineral paragenesis. Given that deformation is an important driving force for fluid movement in the crust (Yardley, 1983; Oliver & Wall, 1987; McCaig & Knipe, 1991; Yardley & Bottrell, 1992; Rumble, 1994) the late mineralisation timing may reflect differences in the deformation style throughout the tectonic sequence.

During the major terrane deformation, synchronous with prograde metamorphism, ductile processes such as shearing associated with cleavage formation could cause fluid to move pervasively (Etheridge et al. 1983; Bell & Cuff, 1989; McCaig & Knipe, 1991) along grain boundaries, microcracks and cleavage seams (Fig. 25). Consequently, the fluids are likely to be rock buffered enhancing their potential for equilibration (*e.g.* Wood & Walther, 1986) with the regional sequence. In contrast, later weak deformations are characterised by a more discrete and heterogeneous structure development (Fig. 25), which includes veining, faulting and localised gaping of pre-existing structures. Consequently, porosity and permeability are increased by this structural development (Cox et al., 1991a,b; Ridley, 1993) and fluid infiltration into the rock sequence becomes non-pervasive (McCaig & Knipe, 1992). Additionally, partitioning of regional deformation, coupled with an overall inhomogeneous stress distribution, will create zones of low mean stress into which ore fluids are subsequently focussed (Gray et al., 1991; Ridley, 1993). The increased fluid flux into these discrete zones will assist mineral concentration, and thus precipitation (Ferry & Dipple, 1991), in the process promoting further hydrofracturing (Phillips, 1972) and implosive brecciation (*e.g.* Bell et al., 1988; Forde & Bell, 1994) that enhance structural opening.

The development of dilational zones, and thus ore-fluid release, within the crust is commonly considered to result from increased fluid pressures that cause hydrofracturing and the subsequent generation of open space (Kerrich & Allison, 1978; Cox et al., 1991b). An example of this is the build-up of metamorphic fluid pressure that occurs from an accumulating volatile supply (*e.g.* Etheridge et al., 1983).

However, dilational zones may also be created by deformation that produces sites of structural opening into which fluids that are already at an increased pressure, due to metamorphic/tectonic processes, infiltrate rapidly. Such sites would contain implosive breccias and stockwork veins, suggestive of a static fluid pressure increase. However, their formation would have been controlled instead by deformation, rather than a static fluid pressure increase. As late orogenic deformations usually have a heterogeneous effect on the rock sequence, they would only produce discrete zones of opening and thus concentrate ore-fluids into areas of the crust that had effectively partitioned the strain arising from such deformation.

In summary it is proposed that the late timing of mineralisation in mesothermal crustal settings is the result of a change in deformation mechanisms between the major tectonic event and later weak events. These mechanisms initially comprise grain-scale effects (*e.g.* recrystallisation, penetrative cleavage formation) that along with the fluid are pervasively distributed through a rock sequence during the major period of terrane deformation and metamorphism. Subsequent deformations however, have a more heterogeneous strain distribution and produce meso- and macroscale brittle-ductile structures such as veins and faults that effectively channel the fluid into discrete zones of the crust. This focussing of ore-bearing fluids into specific sites in the crust, as a consequence of late structural timing, leads to the economic concentration of mineralisation.

SIGNIFICANCE

Implications for Exploration in Other Areas of the Paterson Province

In assessing the exploration potential of the whole Paterson Province for further Telfer-style mineralisation the role of the regional granitoids, which appear to have only intruded the Telfer region, in ore genesis is essential. The results of this study suggest that a direct genetic link (*ie.* mixing between a magmatic and other fluid phase) is unlikely and that a proximal relationship to actively cooling regional granitoids is not required for deposit development. Instead, the ore fluids were extensively circulated through the rock sequence, were in part sourced from metamorphosing rocks, and deposited mineralisation outside the direct influence of the granitoids. If the regional association between Telfer and the granitoids simply reflects the utilisation of magma conduits by auriferous metamorphic-sourced ore-fluids to access the Telfer region, then metamorphic Au deposits potentially may have developed in other areas of the Paterson Province. Such deposits would be unlikely to contain Cu or other magma-related elements and would have required a suitable province-scale fluid channelway to develop.

Implications for the Structural Controls on Mineralisation in the Telfer Deposit

Although the Telfer Dome (the host-structure) formed during D_2, the results of this study demonstrate that mineralisation actually occurred during D_3 and D_4 and this places constraints on models produced to account for the ore-body geometry. An example of this is the utilisation of quartz veining to determine structural controls and timing of mineralisation. Although gold mineralisation is spatially associated with late-D_2 quartz veining along the dome limbs, textural examination in this study has indicated that the ore-assemblage post-dates this veining. In the absence of this textural information, it would have appeared that the timing and structural controls on gold mineralisation were related to the late-D_2 quartz veining (*e.g.* Goellnicht et al., 1989; Windh, 1991), when instead it is actually related to two subsequent deformations. These deformations (D_3 and D_4) created precipitative openings in the quartz through internal deformation and gaping along the wallrock-quartz contact in which mineralisation developed.

MVR and E-reef Carbonate Equivalents

Previous studies identified coarse-grained dolomite-carbonate horizons that were lateral stratigraphic equivalents to reef mineralisation (Turner, 1982; Goellnicht, 1987). These were interpreted to be precursor chemical traps for quartz-sulphide reef mineralisation (Goellnicht et al., 1989). However, the results of this study indicate that carbonation occurred after reef quartz veining, and therefore did not provide a chemical control on mineralisation emplacement. Instead, the lateral carbonation of the reef horizons suggests that hydrothermal zonation may have occurred (Fig. 26). This would have arisen through the inability of Phase 2 & 3 carbonate-bearing solutions to permeate higher levels of the reef horizon due to a permeability reduction produced by the earlier silicification (Fig. 26). Such zonation suggests that fluid flow within the deposit was strongly channelled along the individual reef horizons.

CONCLUSIONS

A study of the Telfer ore-system, as represented in three suites from different stratabound reef systems in the Telfer Dome, has revealed a progressive mineral paragenesis across the major hydrothermal mineralising event. The paragenetic sequence comprises; **1.** Massive anhedral stratabound quartz veining; **2.** Grey ferroan dolomite-epidote-chlorite veining and pervasive wallrock alteration; **3.** Opaque white-pink ferroan dolomite-chlorite-pyrite-galena-chalcopyrite veining/alteration; **4.** Quartz-chlorite-muscovite-calcite veining plus pyrite and chalcopyrite mineralisation; and **5.** Late argillic and chalcedonic veining. Gold mineralisation is coeval with the latter stages of Phase 3 and 4 veining/alteration, and possibly Phase 5 as well.

The close similarity between the regional metamorphic and hydrothermal mineralisation assemblages suggests that the ore fluids were homogeneous and had equilibrated with the country rocks prior to entering the deposit. Coupled with the late-timing of mineralisation, and the strong structural control on veining, this suggests that the Telfer system had a metamorphic fluid component. This is a departure from previous genetic models where a direct contribution from spatially associated granitoids was inferred. The model proposed in this study emphasises the formation of a metamorphic mesothermal lode-gold deposit at a moderate crustal setting, where spatially related granites, and their related mineralisation, have contributed to, but not dominated, the development of the Telfer ore system.

Microstructural timing of the various veining stages indicates that Phase 1 occurred late in D_2, the host-structure (dome) forming event, whilst the subsequent phases precipitated during two later weak orogenic deformation events (D_3 & D_4). Therefore, the structural timing of gold-copper mineralisation is syn-D_3 and D_4. It is suggested that this late timing of mineralisation may be potentially due to a change in deformation mechanisms, between D_2 and D_3/D_4, and the effect this had on fluid movement in the upper crust. During D_2, fluid moved pervasively through the rock sequence, whilst in D_3/D_4 it was highly focussed into discrete deformation-induced zones of porosity/permeability. The latter arose from the heterogeneous character of late orogenic deformation that was highly partitioned in the rock sequence. This development of discrete fluid channelways increased the concentration and flux of the ore-fluids thus promoting precipitation in suitable trap-sites.

SECTION D

Progressive Structural Development of the Telfer Dome and Controls on Gold-Copper Mineralisation.

SECTION D

Progressive Structural Development of the Telfer Dome and Controls on Gold-Copper Mineralisation.

ABSTRACT

Structural mapping and analysis of the Telfer Dome, a doubly plunging antiformal fold that hosts the Telfer Au-Cu deposit, has revealed a progressive sequence of domal development. NW-SE trending regional folding, and dome formation, occurred during D_2. However, earlier folding phases influenced the D_2 geometry of the Telfer Dome. In particular, the refolding of monoclinal D_1 folds gave the Telfer Dome an asymmetric shape and a SW inclined axial plane that was transected by S_2. As well, post-D_2 deformation slightly modified the dome. The effects of this modification are important as post-D_2 deformations were broadly synchronous with mineralisation development.

A number of mineralisation styles were developed in the Telfer Dome, the most important being laterally extensive stratabound quartz-carbonate-sulphide reefs. Other styles include stockworks, discrete bodies known as pods, and vein systems. The stratabound reefs developed in specific stratigraphic horizons around the dome. These horizons were those that comprised transitional sedimentary units (ie. interbedded sandstone-siltstone packages) lying between thicker bedded units in the stratigraphic sequence. The best developed mineralisation in these horizons occurred where they had been extensively silicified, and internal reef textures indicate that the reefs had a protracted development that was associated with repeated deformation. Reef mineralisation developed preferentially on the NE limbs of D_2 folds, particularly where this limb had been steepened by subsequent deformation.

A model is proposed for the formation of the reefs during progressive polyphase tectonic deformation. This commenced with quartz veining and silicification of the reef horizons during late-D_2 flexural slip deformation. D_3 preferentially gaped the NE limbs of the Telfer Dome, especially those that were sub-vertical (a feature inherited from the refolding of D_1 folds), producing carbonate-sulphide-gold mineralisation in the fractured quartz reef host. Mineralisation continued during D_4 as further gold, sulphide and silica were added to the reefs during layer-parallel shearing of the reef horizon that arose from fold limb reactivation caused by renewed shortening of the Telfer Dome.

INTRODUCTION

Gold-bearing saddle reef structures are a well known example of fold-hosted mineralisation for which a strong degree of structural control has been inferred (*e.g. Victoria* - Cox et al., 1983; 1991a; Sandiford & Keays, 1986; Ramsay & Willman, 1988; *Meguma* - Keppie, 1976; Haynes, 1986; Mawer, 1986; *Haile* - Hayward, 1992). These, and other examples of fold-hosted mineralisation, are commonly associated with structural processes that occurred during active fold formation. Examples of such processes include bedding slip and hydraulic fracturing, which produce concordant quartz veins (Mawer, 1985,1987; Jessell et al., 1994) and en échelon tension veins (Wilson, 1982; Ramsay & Huber, 1987), limb and hinge dilation produced as accommodation structures during multi-layer folding (*e.g.* Chace, 1949; Cox et al., 1991b), and late-folding faulting with associated mineralised veining (*e.g.* Hodgson, 1989; Cox et al., 1991a,b).

The structural processes described above that lead to the development of suitable ore-fluid trap sites in regional folds are commonly considered to occur late in the main fold-forming event. However, the timing of mineralisation in many lode-gold deposits is commonly late (Craw, 1989; Kontak et al., 1990; Jemielita et al., 1990; Robert, 1990), and some microstructural studies have demonstrated that gold mineralisation actually accompanied late orogenic deformation events that post-dated development of the regional host-structure/fold (*e.g.* Wilkins, 1993; Forde, 1991; Forde & Bell, 1994; Section C). The latter observations imply that mineralisation emplacement in regional fold structures may be controlled by deformation conditions different to those that formed the fold. This raises an important question; "To what extent can late orogenic deformation interact with a pre-formed fold structure and how would this control the emplacement of mineralisation?"

The Telfer gold-copper deposit comprises a series of auriferous quartz-sulphide stratabound "reefs" developed along the limbs and hinge of an anticlinal dome that formed during the major period of terrane deformation, D_2 (Section A). Previous research has stressed the role of late D_2 folding-related processes and faulting in controlling mineralisation development within the deposit (Goellnicht, 1987; Hill, 1989; Bogacz, 1990; Laing, 1993; Vearncombe & Hill, 1993). However, current research indicates that mineralisation post-dated D_2 and occurred during two subsequent orogenic deformations, D_3 and D_4 (Section C). Consequently, structural processes invoked for D_2 folding cannot be wholly used to model the development and distribution of mineralisation in the Telfer Dome. Instead, the development of the Telfer Dome, and mineralisation within it, is likely to have been the result of the progressive and cumulative interaction of multiple phases of orogenic deformation.

REGIONAL GEOLOGY AND PREVIOUS INVESTIGATIONS

The Telfer deposit lies within the Paterson Province on the NW margin of the Paterson Orogen in the northwest of Western Australia (Williams & Myers, 1990). The province exposes middle-Proterozoic meta-sedimentary rocks of the Yeneena Group that were folded, metamorphosed and mineralised during convergent orogenesis in the late-Proterozoic (Paterson Orogeny, 640-600Ma - see Section B). The Paterson Orogeny comprised multiple phases of deformation (Section A). Of these, D_2 was the strongest and produced NW-SE trending folds across the province. In the Telfer region, these folds formed as doubly plunging antiforms and synforms producing a dome and basin structural setting. One of the antiformal domes, the Telfer Dome, hosts the stratabound reefs of the Telfer deposit within three stratigraphic formations, the Telfer, Malu Quartzite and Isdell (Fig. 1).

Initial research of the Telfer ore-system suggested an syngenetic exhalative origin for the laterally extensive stratabound mineralised reefs (Tyrwhitt, 1979; Turner, 1982). However, subsequent studies have demonstrated convincing evidence for an epigenetic origin based on isotopic data, structural timing and structural controls on mineralisation (Goellnicht, 1987; Goellnicht et al., 1989; Rowins, 1994; Section C). Numerous recent consultancies have proposed a variety of models for the formation of the Telfer Dome in different tectonic environments (Table 1). However, only limited study has focussed on the specific controls on reef development, and most consensus was for late-D_2 folding flexural slip/shear deformation producing dilation and stockworking in reefal horizons (Goellnicht et al., 1989; Windh, 1991; Vearncombe & Hill, 1993; Laing, 1993). Goellnicht et al. (1989) also suggested that reef formation may have been assisted by the replacement of carbonate-bearing units within the dome.

Dome Formation Model	References
Strike-Slip/Wrench (dextral) faulting in basement and cover sequence	Harris (1985); Goellnicht (1987); Rowins (1994)
Compressive Thin Skinned Thrust-Folding (NE vergence)	Hill (1989); Windh (1991).
Vertical Basement Movements (high angle reverse faults - NE verging)	Bogacz (1990)
Vertical Basement Movements (positive flower structure)	Baxter (1991); Hill (1989)
Inhomogeneous Shortening	Laing, 1993; this study

Table 1: Summary of the various models proposed for the formation of the Telfer Dome.

STRUCTURAL GEOLOGY/GEOMETRY OF THE TELFER DOME

The Telfer Dome is approximately 7 km long by 3km wide and contains two lower order sub-domes, Main and West Domes, arranged en échelon (Fig. 2; Map 1). The northern closure is moderately tight, plunging 25-30° NW, whilst the southern closure has a characteristic "swallow-tail" shape produced by two separate anticlines bounding the Main Dome Syncline (Fig. 2). The larger of these, the southern Main Dome closure, plunges shallowly (5-15°) SE whilst the smaller Pit 9S- Wallaby Hill closure (Fig. 2) is tighter with a steeper plunge (35-40°). This "swallow-tail" shape in the SW suggests that the Telfer Dome has an overall gentle plunge to the NW. The axial traces for both the Main and West Domes die out towards the centre of the overall Telfer Dome (Fig. 2), each passing into the limb regions of adjacent anticlinal folds. West Dome has a more complex axial trace geometry with a major axis passing out to the NW through the north-western Telfer Dome closure, and a south-eastern axis passing through Wallaby Hill. Various faults along hinge and limb regions offset marker units (Fig. 2) and these commonly have an apparent right-lateral movement sense.

Main Dome

Folding

Main Dome is 3.5km long by 1.5km wide and lies in the SE corner of the Telfer Dome (Fig. 2). It is an open doubly plunging anticline that is weakly asymmetric with a steeper dipping NE limb. However, axial plane cleavage (S_2) has a consistent sub-vertical orientation across the dome on both limbs (Fig. 2; Map 1). Recent deep drilling (commenced in mid-1992) has indicated that the Main Dome hinge tightens considerably in deeper levels, with local overturning of the fold limb and axial plane (S_2) cleavage (Fig. 3) that is consistent with refolding during D_3. The deep drilling has also indicated that the geometric axial plane is inclined approximately 75°SW. A large monoclinal recumbent fold (Fig. 4), previously considered to be the NW closure (Goellnicht, 1987; Hill, 1989; Gallo, 1991; Laing, 1993), is a D_1 fold that lies adjacent (NE side) to the Main Dome hinge (Fig. 2; see also Map 1). The southern closure has an opposite asymmetry to that of the dome (Fig. 5), and cleavage relationships suggest this was the result of local strain intensification on the SW limb during D_2 (Fig. 5).

Folding involved a combination of brittle and ductile deformation characterised by bulk shortening and penetrative cleavage formation. This was accompanied by late flexural slip deformation that produced slickenfibre fill, bedding-concordant veining, intra-bedding faulting, local fault-propagation folding and tension gash formation (Fig. 6). These structures also occur at a high angle to the domal axis (Fig. 6) indicating that flexural slip occurred longitudinally as well as laterally across the dome. Prominent jointing developed in the core of Main Dome and some of are consistent with formation during domal folding (Fig. 7). Very local and small disharmonic cross-folding has developed, rarely with an associated axial plane cleavage, and these correspond to either the subsequent D_4 or D_5 fold styles.

Faulting

Faulting in Main Dome exhibits three major trends (Table 2). The majority of faults form along dome limbs (Fig. 8), and are commonly parallel or sub-parallel to bedding along part of their length (Fig. 9A). Others produced small graben with ductile drag folding along the margins (Fig. 9B). A lack of suitable fault displacement criteria precludes effective kinematic analysis. However, the majority of faults exhibit a combination of dip-slip and transcurrent movement (Fig. 8, 9A).

Fault Trend	Comments
NE-SW striking, moderate (20-25° NW) to steeply dipping	Common in the Footwall Stockwork in the core of Main Dome where many are mineralised. Both normal and reverse movement types, with conjugate sets forming mini graben and horst structures
NW-SE striking, moderate (bedding parallel) to steeply dipping	Parallel to the domal axial plane forming along dome limbs; rarely curvilinear and passing into parallelism with bedding; truncate D_2 folds (Fig. 8). Apparent bedding offset suggests reverse (+ sinistral) movement.
NNW-SSE / N-S striking sub-vertical faults	Late faults with a normal (+ sinistral) movement. Form small graben (Fig. 9B), and associated chert veining (Smith, 1989).

Table 2: Summary of the major faulting trends and characteristics in Main Dome.

Graben Fault : This is a large 10-20m wide N-S striking fault zone that cuts the middle and SW corner of Main Dome (Fig. 2 & 9B; Map 1). The zone dips steeply WNW (≈70-72°) and has an apparent normal (+ sinistral) movement sense, which diminishes at lower structural levels suggesting a rotational component (M. Johnston, pers. comm., 1994). In-situ fault plane breccias contain poorly sorted angular clasts supported by a clay matrix, and zones of intense fault-parallel fracturing have formed in the competent hangingwall rocks. Large clasts of Middle Vale Reef (MVR) mineralisation within the breccia zones (C. Moorhead, pers.comm, 1993) indicate that the Graben Fault post-dates reef formation.

Pit 8 - Timing of Reverse Faulting

Pit 8, located on the north-westerly limb of Main Dome (Fig. 2; Map 1), contains flexural slip duplex (*e.g.* Tanner, 1992; Fig. 10A). These, coupled with high-angle reverse faulting, initiated as bedding parallel and curvilinear faults in finely laminated siltstone immediately overlying E-reef (see below) horizons (Fig. 10A & B). Timing criteria, including in-situ fault breccias and discontinuous cleavage developed marginal to the faults (Fig. 10A), suggest a late or post-S_2 timing. However, disharmonic folds within the duplex (Fig. 10A) and in laminae adjacent to fault planes (Fig. 10B), have a D_4 orientation (*Axial Plane:* sub-vertical and striking 165 - 175°) suggests that some faulting may have occurred during D_4 shortening of the Telfer Dome.

Disharmonic folding of laminae adjacent to the fault planes also produced dilation/gaping and quartz-sulphide infill (Fig. 10B) indicating that high-angle faulting was associated with mineralisation development in the Telfer Dome. In other areas of the E-Reef package small normal faults developed parallel to S_1 (*ie.* at a low-angle to bedding). However, some of these now exhibit reverse movement suggesting they were antithetically re-used during subsequent compressional deformation (Fig. 10C). The faults are interpreted to have initially formed during layer-parallel extension in D_1 and, along with S_1, had the potential through reactivation (*e.g.* Sibson, 1985) to assist subsequent reef and concordant vein formation by providing ore-fluid channelways to the base of the reefs (see also below).

West Dome

Folding

West Dome exhibits greater structural complexity than Main Dome and is located on the NW corner of the Telfer Dome. Overall the dome is a broad open asymmetric anticline characterised by a steeply dipping NE limb (Fig. 11A & B) and a complex internal geometry of smaller antiformal and synformal closures (Fig. 11B). Whilst S_2 is generally vertical across the dome, axial planes are commonly inclined to the SW (≈75°; Fig. 11A & B). Deep diamond drilling in the Pit 11 area (see Fig. 2 for location) has indicated a tightening of the fold hinge and a marked anticlinal asymmetry (steeply dipping NE limb) in that area. The main West Dome axial trace (Fig. 11A; Map 1) passes northward through the northern closure of the Telfer Dome (Fig. 2) but dies out on SW limb of the Main Dome Syncline (Fig. 2). A second subsidiary anticline to the south-west has produced a southern extension of West Dome and a map-view domal asymmetry. The two anticlines are separated by a shallow asymmetric syncline that lies on the SW limb of the overall West Dome.

Locally, fold hinges in the main West Dome Anticline display an asymmetric shape that mimics the bulk dome asymmetry (Fig. 12). Axial plane cleavage (S_2) across the hinges remains vertical suggesting that the asymmetry is the result of strain intensification during D_2 (Fig. 12). However, other asymmetric fold shapes are recumbent D_1 folds, and these are commonly preserved within the NE limb of West Dome (Fig. 13). Similarly, cleavage relationships along the NE limb of the main West Dome Anticline in the eastern benches of Pit 10 indicate that local D_1 vergence reversals occur. These reversals are associated with locally overturned strata and are up-strike of the recumbent folds shown in Fig. 13 suggesting that a mesoscale F_1 fold is preserved within and along the steeply dipping NE limb of the main West Dome Anticline (Fig. 14).

Folding is typically affected by faulting with small fault-propagation folds nucleating on curviplanar bedding-parallel faults that cut upwards through the sequence (Fig. 15A, B & C). Parasitic folds in West Dome typify complex multilayer behaviour with concentric (Fig. 15C) and box (Fig. 15D) styles common, and many exhibit inflexions or small kinks in the crestal regions (Fig. 15E). Most folds also exhibit marked hinge curvature (Fig. 15F). Relict monoclinal flexures are preserved within moderately dipping strata (Fig. 15G), and these are commonly truncated by layer-parallel shear resulting from bedding reactivation during D_2. Other flexures have a NE vergence and appear to be relict D_1 folds (Fig. 15H & I). As for Main Dome, folding was a combination of bulk shortening with penetrative axial plane cleavage development, coupled with late flexural slip and faulting deformation.

Faulting

The majority of faults in West Dome parallel the axial plane (Fig. 2; Map 1; *ie.* NW-SE striking) and developed in high strain zones where folds were tightened and then faulted. These zones are typified by numerous discontinuous sub-vertical fault planes, with associated breccia/gouge and are commonly infilled with anhedral quartz. Bedding deflections and slickenside lineations (sub-vertical pitch) indicate mainly dip-slip reverse movement in the Pit 9S area. In Pit 10, thicker fault zones exhibit both oblique dextral-normal and sinistral transcurrent movement histories. A set of right-stepping shears, termed the West Dome Deeps, have been identified through deep drilling; these lie along individual fold axial planes in the dome (see also below). Some axial faults have metre-scale normal movements (A.Eaves, pers.comm. 1994), which may be due to reactivation during NW-SE compression during D_5 (Section A), when σ_1 would have been parallel to the fault trend.

Pit 9S Fault Zone : This is a major high-angle reverse fault zone that lies in the hinge and along the NE limb of the Pit 9S anticline (Fig. 16; see Fig. 2/Map 1 for location). It comprises a steeply SW dipping 15-20m wide zone of discontinuous fault surfaces with interstitial breccia/gouge. Ductile drag folding and slickenside lineations indicate mainly dip-slip movement. The strong asymmetry of folding suggests that faulting controlled folding (*e.g.* fault-propagation folding - Suppe & Medwedeff, 1984; Erslev, 1991). However, vertical axial-plane cleavage (S_2) orientations are preserved across the closure (Fig. 16). Additionally, bedding and cleavage (S_2) on the steeply dipping NE limb are locally overturned (Fig. 15), and breccia clasts that contain cleaved (S_2) siltstone are rotated. This indicates that faulting post-dated S_2 formation and occurred late or post-D_2 folding, most likely modifying the original fold geometry.

MINERALISATION STYLES/SETTINGS

Gold-copper mineralisation is hosted in four different styles/settings; **1.** stratabound bedding-concordant reefs, **2.** stockworks (tabular and block), **3.** discrete bodies, referred to as pods and **4.** mineralised faults/shears. Of these, the stratabound reefs are the most economically significant.

Bedding Concordant Reefs.

A typical Telfer reef comprises a laterally extensive, conformable (locally slightly discordant), stratabound layer of quartz-limonite-pyrite-chalcopyrite mineralisation enclosed by siltstones. Near surface reefs (Middle Vale Reef/E-Reefs) are invariably altered to limonite-goethite-hematite during secondary enrichment, which markedly increased gold grade and obliterated primary textures. Reefs may locally be stratiform and comprise both infill and replacement veining styles, and are located at specific positions within the stratigraphic sequence (Table 3; see Fig. 1 for distribution).

REEF	Host Stratigraphic Unit	Stratigraphic Separation (*)	COMMENTS
E-Reefs (E1, E1A, E2, E3)	Outer Siltstone Member (Telfer Formation)	+ 65-75m	Currently mined in West Dome
Rim Reef	Rim Sandstone Member (Telfer Formation)	+ 55m	Minor economic importance
Middle Vale Reef (MVR)	Middle Vale Siltstone (Telfer Formation)	0 m	Prior to the discovery of deeper reefs, the MVR accounted for ≈70% of the Telfer resource.
M-Reefs (M10-M70)	Malu Quartzite Formation	M10 (-135m) M70 (-620m)	Recently discovered series of reefs
I-Reefs (I10-I30)	Isdell Formation	I10 (-860m) I30 (-910m)	Recently discovered by deep diamond drilling

Table 3: *Summary of stratabound reefs currently identified within the Telfer deposit and their stratigraphic positions. Note: (*) Stratigraphic separation is relative to the MVR; + denotes metres above, - denotes metres below. (See Fig. 1 for the distribution of reefs in Main Dome)*

Middle Vale Reef (MVR).

The Middle Vale Reef (MVR) is a 0.3 to 1m thick horizon consisting of massive quartz veining (0.2 - 0.4m) that is infilled with medium-coarse grained pyrite-chalcopyrite aggregates (Fig. 17), particularly along the upper contact. Additional disseminated pyrite, carbonate and argillic veining occur throughout the reefal horizon. However, the primary sulphide assemblage is extensively supergene altered to chalcocite and digenite, and heavily oxidised in near surface exposures, enriching the gold grade (Dimo, 1990). The MVR varies laterally ranging from areas with two to three well defined textural zones (see below) to thin (≈0.1m) chalcocite horizons devoid of the massive quartz protolith. This protolith may also split into two or more thin stringer veins along strike, and commonly exhibits a "pinch & swell" style, a feature common in stratabound veins (*e.g.* Koistinen, 1981; Wilkins, 1993).

The MVR is best developed along the eastern and south-eastern limbs of Main Dome. In these areas it formed along thick quartz veining and commonly exhibits textural zonation (see below). Elsewhere in Main and West Domes it is a heavily oxidised horizon comprising disseminated pyrite within strongly silicified-carbonated-sericitised siltstones with abundant fine-grained quartz and argillic veining. Massive quartz veining associated with MVR development is only generally weakly developed in the hinge regions of domal folds, and in the absence of this veining the MVR commonly formed as a thin horizon of disseminated sulphide replacement. The MVR is hosted within a finely laminated siltstone horizon (Middle Vale Siltstone; Fig. 18) that lies towards the base of the Telfer Formation (Fig. 1). This horizon is characterised by alternating sandstone and siltstone laminae, and lies between more massive units (Fig. 18).

The upper contact of the MVR is sharply planar and the hanging-wall is locally stockworked (argillic-carbonate-silica veins), silicified and carbonated. Enveloping laminae in the hanging-wall rarely exhibit D_4-oriented disharmonic folding in which the fold hinges are gaped and infilled with quartz-sulphide. In contrast, the lower margin is commonly irregular and replacive (Fig. 19) with massive quartz veining truncating weakly contorted footwall siltstone laminae. Wallrock alteration is restricted to the immediate reef margins. Small-scale faulting, coupled with pockets of silicified breccia that transgress reef margins along both contacts, was coeval with sulphide mineralisation (Fig. 20) indicating elevated fluid pressures and concomitant brittle deformation during progressive reef mineralisation. These faults, which are both high and low-angle to reef margins, exhibit dip-slip and oblique (reverse-sinistral) movement.

Textural Zoning in the MVR : In thicker, well developed, sections of the MVR two or three distinct textural zones occur (summarised in Table 4 (overpage) & Fig. 21). This zoning, which has only been observed on the central-eastern limb of Main Dome, comprises upward younging zones of reef deformation that resulted in dynamic recrystallisation of the massive quartz component, accompanied by fine-grained euhedral pyrite growth. Quartz-fibre pressure shadows on pyrite porphyroblasts (Fig. 21) are commonly elongate parallel to bedding indicating that syn-tectonic layer-parallel shearing (Durney & Ramsay, 1973; Ramsay & Huber, 1983) nucleated and intensified across the upper reef margin, and this was accompanied by cyclical fluid pressure increase and further quartz-sulphide precipitation.

E-Reef Mineralisation

The E-Reefs comprise four separate horizons hosted within the Outer Siltstone Member close to the Rim Sandstone contact (Fig. 22A). They are best developed in West Dome (Fig. 23) where strong supergene enrichment has increased the gold-grade to economic levels. Individual E-Reefs are hosted within finely laminated horizons (0.8 - 1.3m thick) that comprise alternate sandstone and siltstone laminae whose thickness ratio is approximately 1:1. These horizons lie between more massive bedded sandy siltstone units that lack the marked lithologic variation of the E-reef sequence.

ZONE	DESCRIPTION
1 (Upper)	Strongly fractured inclusion-rich quartz clasts (see Zone 3) are surrounded by a fine-grained recrystallised quartz mosaic, which contains fine-grained sulphides that exhibit a weak undulose layering. The massive quartz was dynamically recrystallised during strain intensification along the upper reef. The lower boundary of this zone against Zone 2 is undulose.
2 (Middle)	Scattered large (1-2mm) rounded pyrite grains are surrounded by a weakly laminated semi-massive pyrite aggregate, within a siliceous matrix comprised of fine-grained recrystallised quartz containing layers of fine-grained euhdral pyrite (0.08-0.12mm) parallel to the MVR margins. Large rounded pyrites have prominent quartz-fibre pressure shadows whose maximum elongation is parallel or oblique to the dip-slip vector. The degree of layering varies from well layered (differentiated qtz-pyrite layers) to wispier pyrites scattered through a recrystallised quartz matrix. A fine-grained chalcedonic phase overprints everything.
3 (Lower)	Comprising up to 70% of the MVR, the lower zone consists of massive coarse grained clear-grey anhedral quartz infilled by large subhedral-euhedral pyrite/chalcopyrite/chalcocite aggregates (5-25mm); a weak undulose layering of sulphide minerals is sometimes developed. Reefal quartz has been highly strained, brecciated and veined during subsequent deformation induced recrystallisation.

Table 4: Textural zonation observed in the Middle Vale reef (also summarised in Fig. 21); these represent the full spectrum. However, commonly only Zones 1 & 3, or Zones 2 & 3, are developed. The lower zone (3) is the oldest, whilst the upper zone (1) is the youngest.

E-reef mineralisation is strongly developed in tightened asymmetric hinge regions and on steeper NE dipping limbs (Fig. 22A; *e.g.* Pit 9S, Pit 10). Here, mineralisation is hosted as massive breccia/stockwork zones that are locally discordant, and high-grade quartz-goethite (after sulphide) lenses (pods - Fig. 22B), some of which are associated with sub-vertical faults (*e.g.* Pit 9S fault - see above). Breccias consist of angular siltstone fragments, supported in a quartz-sulphide-goethite matrix, indicating strong hydro-/implosive brecciation. Away from fold hinges the E-reefs generally comprise 1-1.5m thick horizons that contain irregular gossanous concordant (locally discordant) quartz-limonite-goethite veining, in-situ brecciation, strong silicification, and disseminated sulphide-silica-sercite alteration.

E-reef Veining Styles: Reefal veining involved both infill and replacement, and massive quartz veining in the MVR was only very locally developed (*e.g.* Pit 14). Veining styles in the E-reefs change across antiformal folds (Fig. 22B), ranging from infill/breccia in the hinge to replacement dominated on the limbs. Many veins nucleated at the upper contact of upward-fining sedimentary sequences suggesting a mechanical control on vein formation. The upper margins are planar, with a locally enhanced bedding fissility suggesting shear intensification, whilst the lower margins are commonly irregular as a result of disharmonic folding of individual laminae at the base and middle portions of the reef.

Rim Reef

The Rim Reef is hosted within a massive quartzite unit (Rim Sandstone), and comprises discrete stratabound lenses with associated supergene mineralisation in discordant fracture sets (Dimo, 1990). Primary reef mineralisation consists of coarse grained pyrite aggregates that replaced the sandstone-quartzite matrix. These aggregates were subsequently fractured and replaced by finer-grained quartz - sulphide veinlets with minor brecciation and incipient sericite alteration on the margins. Recently, pockets of breccia-hosted mineralisation have been identified in the Rim Sandstone (C. Switzer, pers.comm, 1995) and these are best developed on the NE limb of Main Dome.

Deep Reefs (M & I Reef Series)

Recent deep drilling in Main Dome has identified further reef mineralisation in the underlying Malu Quartzite and Isdell Formations (see Table 3; Fig. 1). Two of these reefs (M10, M30) were examined in newly developed underground mining operations, and form part of the M-Reef series which appears to have preferentially developed on the NE limb of Main Dome (Fig. 1).

M10 : This reef ranges from 0.4 to 0.9m in width and comprises a massive quartz protolith that is variably infilled with coarse-grained dolomite/carbonate and sulphide aggregates. Reef morphology varies laterally, commonly by thinning of the massive quartz veining plus an increase in coarse-grained sulphide-dolomite aggregates. The reef thickness also varies with the thickest mineralised reefs occurring where the quartz protolith was well developed. In other areas, the quartz is absent and the M10 comprises thick dolomite-sulphide veining. Reef veining involved infill with lesser replacement, the latter more common for the massive quartz component. The M10 is hosted in a finely laminated siltstone unit, which lies within massive sandstone-quartzite units of the Malu Formation (Fig. 24). Subsidiary reefs, including the M8 (a thin dolomite-sulphide stringer) and M12 (a thinner version of the M10) formed above and below the M10 respectively.

A crude zonation, similar to the MVR, occurs along the upper margins of the M10 reef and comprises zones of cataclastic/milled sulphides within a fine-grained recrystallised silica-dolomite (±scheelite) matrix. Additionally, local and late cyclic brecciation, coupled with intense fracturing occurred in hanging-wall siltstones. These textures suggest repeated layer-parallel shear deformation was localised into the upper reef. Extensional fibre growth in asperities indicates mainly dip-slip and oblique (WNW-ESE trending) movement. Reef margins are sharply planar (Fig. 24), particularly where thick quartz veining developed, whilst in sulphide-rich zones they are undulose. Wallrock silica-dolomite alteration and argillic stockwork veining are restricted to the immediate reef margins.

M30 : This reef generally lacks a massive quartz vein protolith. Instead it comprises a 1 -1.5m wide zone of tectonically disrupted finely laminated siltstone, which were variably folded and brecciated/replaced by dolomite-carbonate and sulphide aggregates (Fig. 25). Internal textures are predominantly infill (coarse interlocking euhedral crystals) and veining exhibits many repeat cycles indicative of cyclic fluid pressure fluctuations. Present drilling and mine development suggests that the gold grades in the M30 are not as great as in the MVR or M10 reefs.

I-Reefs : The three I-Reefs (I10, I20 and I30) comprise quartz-carbonate (± dolomite)-pyrite-chalcopyrite horizons. Less commonly, they are 0.05 - 0.1m thick zones of massive sulphide with associated angular breccias and stockworking. Mineralisation styles include both replacement and infill, and massive breccia zones appear to have formed in tightened domal hinge regions (Fig. 26; see also Fig. 1). Present drilling has targeted the hinge region and consequently little is known about the lateral extent and style of these reefs. The following points (also summarised in Fig. 26) are pertinent to the I-reef system:

1. The I-reefs occur at contacts between coarse grained quartz sandstone (base) and black calcareous (+/- carbonaceous material) siltstone within siltstone dominated sections of the stratigraphic sequence.
2. The Main Dome hinge is continuous and locally overturned in central regions of the dome. However, southwards the hinge becomes increasingly faulted accompanied by increased brecciation and stockworking. Correspondingly, reef veining changes from replacement dominated in the central region to infill dominated in the south (Jan de Visser, pers.comm.,1994).
3. Replacment styles predominate on limb regions, where reef thickness generally increases, whereas hinge regions are characterised by infill (breccia/veining).
4. A dolomitic unit, known as the Lower Limy Unit (LLU) appears to be a lateral equivalent to I-reef mineralisation on the eastern limb. It comprises finely laminated coarse grained dolomite that is overprinted by large aggregates of white-cream coloured epigenetic dolomite.

Grade Distribution Within the Stratabound Reefs

Internal grade distribution in the MVR is characterised by two orthogonal linear high-grade trends (Fig. 27A); the more pronounced of these (N-S trending) corresponds with a major vein trend (*e.g.* Goellnicht, 1987 - see below) indicating that ore-fluids utilised vein-systems to access the reefs. Also evident is the thick and well developed MVR in the SE corner of Main Dome. This corresponds with an overall asymmetry in gold grade within the Telfer deposit, with higher than average grades in both the SE (Main Dome-MVR) and NW (West Dome, Pit 9 - E-reefs) regions of the Telfer Dome. Limited data for the M10 and M30 indicates a "poddy" internal grade distribution characterised by linear zones parallel to both the strike and dip direction (Fig. 27B), of which the former is more prominent.

Stockworks/Faults-Shears/Pods and Vein Systems

Stockwork Veins/Systems

Tabular stockwork zones developed spatially adjacent to stratabound reefs in underlying competent sandstone units (*e.g.* Footwall Stockwork; I-30 stockwork - Fig. 26). The largest of these, the Footwall Stockwork, is hosted within the Footwall Sandstone Member in the core of Main Dome (see Fig. 2). The major vein trends, which are commonly controlled by prominent joint sets, include (after Smith, 1989);

1. Moderate to steeply south dipping quartz-sulphide veins (5 - 20mm thick) that contain 1 - 10 ppm Au.
2. Massive or laminated quartz-oxide veins (5-30cm thick) that dip preferentially west ($\approx 50°$), and less commonly south (50°). Wallrock alteration envelopes consist of Fe-oxides, silica and sericite, and spot gold values from 0.5 to 160ppm Au are recorded.

Veining was mainly extensional or shear-extensional, the latter containing marginal brecciation and deformed internal fibres suggesting that many veins either formed in a shear environment or were subjected to renewed deformation/reactivation after their formation. The I30 stockwork, which developed approximately 60m below the I30 reef in a massive sandstone unit (Fig. 26), comprises quartz-carbonate-sulphide veins with sericite-carbonate-chlorite alteration halos. Both stockwork zones also appear to have provided ore-fluid sinks (*e.g.* Ridley, 1993) immediately below their respective reefs.

Faults/Shears

Large faults/shear systems in West Dome host low-moderate grade gold mineralisation. The largest of these is the West Dome Deeps, a set of three right stepping axial plane-parallel fault zones (Fig. 23) that contain breccia hosted quartz-carbonate-sulphide mineralisation. Low grade mineralisation is also hosted along high angle reverse faults (*e.g.* Pit 9S Fault Zone - see above) in breccia zones and discrete fault-bounded bodies (also known as Pods - see below), and at the intersection of these with reef horizons.

Pods

Pods are discordant bodies of fault and fold hosted mineralisation that only developed in West Dome. They comprise bedding concordant lenses (Fig. 22B & Fig. 28A), linear and tabular bodies (Fig. 28A), some of which lie along fold hinges or limbs (Fig. 28B), and discordant quartz-breccia structures (Fig. 28C). Pod mineralisation developed at discrete stratigraphic-structural locations in West Dome (Fig. 23); in Pit 10 pods overlie E-reef mineralisation, whilst the reverse occurs in Pit 9 (Fig. 22B) and many are associated with faults along the West Dome anticline hinges.

Vein Morphology/Trends

Goellnicht et al. (1989) summarised the major vein types in Main Dome (Fig. 29) and similar styles also occur in West Dome. Large sub-vertical sheeted vein arrays also developed in West Dome (Leeder Hills/Daves Vein arrays - Fig. 23) perpendicular to the domal axial plane. These consist of extensional quartz-limonite-goethite veins in sheeted sets that commonly intersect E-Reef mineralisation. Where this occurs, gold grade is locally enriched in both the E-reefs and vein arrays.

Bedding Concordant Quartz Veins : Siltstone dominated units (*e.g.* the Outer Siltstone Member - Fig. 2) contain numerous laterally extensive 2-8cm thick bedding-concordant laminar quartz veins. These are rarely mineralised except where they intersect reefs or vein systems. The veins are both massive and laminar and formed at the contact between finely cleaved siltstone and micaceous sandstone in upward fining beds (Fig. 30A). They were folded by D_2 and pre-date cleavage rotation during late-D_2 bedding reactivation (Fig. 30B). Consequently, they formed during the early stages of domal (D_2) folding. Like the reefs, sequences with a high veining density are finely laminated (Fig. 30C) and comprise alternate sandstone-siltstone laminae with a thickness ratio of approximately 1:1.

Individual vein laminations range from 3-15mm thick and exhibit slickenfibre growth along their surface. These laminations may occur throughout the vein or be restricted to the margins, which are sharply planar. In thin section the veins are composed of massive subhedral interlocking inclusion-rich quartz crystals that grew perpendicular to vein margins and are optically continuous (Fig. 30D). Vein quartz is highly strained and dynamically recrystallised along laterally continuous fractures that formed, with increasing density towards the vein margins imparting a laminar nature to the veins (Fig. 30D).

The large vein quartz crystals are offset along the fractures/laminae. The presence of oblique fine-grained mica growth along the laminae (Fig. 30D) suggests that shearing, rather than pressure solution, occurred across the fractures. This is consistent with small asymmetric tensional pull-aparts (*e.g.* Peacock & Sanderson, 1995; Fig. 30E) along fracture surfaces that contain micas whose 001 face is aligned parallel to the pull-apart walls (≈90° to those on the fracture planes) indicating open space growth. Mica alignment on the fracture plane and the opening sense of the pull-aparts is consistent with flexural slip, suggesting that the laminations were the result of superimposed deformation (shearing) on pre-formed massive quartz veins.

INTERPRETATION AND DISCUSSION

Progressive Development of the Telfer Dome

Previous models of dome formation (see Table 1) fall into two broad categories; those supporting basement-fault controlled folding (*e.g.* Harris, 1985, Bogacz, 1990; Baxter, 1991) and those proposing compressive thrust-fold belt folding (*e.g.* Hill, 1989; Windh, 1991). Basement-fault controlled folding is generally incompatible with the development of regionally penetrative folding and axial plane cleavage accompanying metamorphic recrystallisation (*e.g.* Rowan & Kligfield, 1992; Hedlund et al., 1994; Narr & Suppe, 1994). Although Main and West Domes are arranged in a left-stepping en échelon manner, regional domes are not arranged en échelon (Section E), and fold axes in West Dome actually have right-stepping en échelon pattern (Fig. 2 & 11A; *cf.* Harris, 1985). Additionally, a sub-vertical mineral elongation lineation (L_2^2), coupled with the absence of fault-extension along domal axes, is inconsistent with fold and structure styles expected in a wrenching environment (*e.g.* Harding, 1974; Sylvester, 1988; Jamieson, 1991; Woodcock & Schubert, 1994).

Similarly, a number of factors oppose thrusting/thrust-fold belt models for dome formation, including:

1. The absence of strata thinning on fold limbs, particularly the NE ones, that would signify forelimb deformation arising from thrusting (*e.g.* Alonso & Teixell, 1992; Jamieson, 1992).
2. The Telfer Dome does not exhibit a kinked hinge or duplex imbrication, typical of the "snakes-head" geometry observed in fault-bend folds (Suppe, 1983; Suppe & Medwedeff, 1984; Zoetemeijer & Sassi, 1992).
3. Faulting post-dates shortening and cleavage development (*e.g.* Pit 9S fault zone - Fig. 16). Although this can occur in areas of detachment folding (Jamieson, 1992), such a model is not suggested by regional and province-scale structural geometry (see Sections B & E).
4. Regional structural analysis indicates that thin-skinned deformation did not occur in the Telfer region, and that there are no large thrust faults or ramps associated with regional domes (Section E) that would signify thrust–fold models.

Therefore, it is interpreted that the Telfer Dome formed through bulk tectonic shortening (*e.g.* Ramsay, 1967; T.Bell, 1981). In particular, the development of a penetrative axial plane cleavage indicates that the amount of shortening was considerably greater than previously suggested (*e.g.* 9% - Vearncombe & Hill, 1993) for the Telfer Dome . The doubly plunging nature of folding is interpreted to result from heterogeneous vertical strain (*e.g.* Sanderson, 1973) during D_2. However, the present geometry of the Telfer Dome provides evidence that pre-D_2 structures controlled folding, and that post-D_2 deformation modified the Telfer Dome.

Pre-D_2 Folding and its Control on Dome Asymmetry

The asymmetric nature of folding in the Telfer Dome (*e.g.* Figs 3, 11 & 12) was previously attributed to NE directed thrusting during dome formation (Hill, 1989; Windh, 1991). However, axial plane cleavage (S_2) remains vertical across the dome, except where it is locally refolded by D_3, rather than being inclined on the NE limb as would be expected for thrust-folding (*e.g.* Boyer, 1986; Alonso & Teixell, 1992). Additionally, the geometric domal axial planes are inclined SW and are thus transected by the vertical S_2. D_1 produced monoclinal folds (Fig. 31A & B; see also Section A) and refolding of these by D_2 is one explanation for the dome asymmetry and transection of the geometric axial plane by S_2 (Fig. 31C), by producing a relict F_1 fold shape in the D_2 dome. This is also consistent with the preservation of D_1 folds in near hinge positions throughout the dome (*e.g.* Figs 13 & 14).

The development of earlier fold phases (*ie.* D_{EP}/ D_1; Fig. 31A & B) could have potentially controlled the nucleation of D_2 folding throughout the region and of sub-dome axes within the Telfer Dome. In addition to controlling the geometry of D_2 domes, earlier folds may also have affected the partitioning of deformation during D_2. For example, centimetre scale D_1 folds commonly exhibit strong S_2 parallel to the steeper NE dipping limb (inset in Fig. 31). This intensification of the deformation can occur during upright refolding of pre-existing steeply oriented fabrics (Bell, 1986), such as bedding on the steeply NE dipping limb of a pre-existing D_1 fold. On a domal scale this would have caused higher D_2 strain along the NE dome limbs (*ie.* NE limbs of F_1 folds), thus accentuating the D_2 dome asymmetry.

Post-D_2 Domal Modification

Post-D_2 deformation subtly modified the Telfer Dome (Table 5); of these D_3 and D_4 are the most significant as these were coeval with the emplacement of gold-copper mineralisation (Section C).

Deformation	Structures developed/modified
D_3	Potential weak rotation of parts of the dome limbs causing bedding reactivation, particularly on the SW limb (see text). Local refolding and overturning of D_2 hinges (*e.g.* Fig. 3).
D_4	Continued amplification of D_2 folding, but with a slight oblique (non-coaxial) motion, sinistral strike-slip faulting, and renewed bedding slip/fault movement.
D_5	Weak cross folding (highly localised), with potential amplification of the double-plunging attitude. Steep reverse faulting at high angle to axial trend.
	(D_{5b}) N-S striking normal/extensional faulting (*e.g.* Graben Fault) at high angle to dome axis, with local graben development. Potential further shortening of the dome but with an oblique (dextral) motion.

Table 5: Summary of the observed/interpreted modifications produced by post-D_2 deformations on the Telfer Dome. The shaded rows indicate reef mineralisation - synchronous deformations.

D_3 Deformation : This caused tightening and refolding of D_2 hinges, producing overturned NE limbs and S_2 (Main Dome - Fig. 3; Pit 9S anticline - Fig. 16), although these effects were local as bulk dome rotation/refolding does not appear to be significant given the consistent sub-vertical S_2 orientation on either limb. However, minor rotation of the SW limb (Fig. 32A) may have occurred, as evident from intra-bedding tension veins whose asymmetry indicates they formed during D_3 rather than D_2 (Fig. 32B & C). This is because rotation and reactivation of this limb during D_3 (Fig. 32B) has the right sinistral shear sense to gape the tension gashes, whereas that during D_2 does not (Fig. 32C). In contrast, the NE limbs and fold hinges were deformed (gaped) by D_3 owing to their high-angle with respect to the prevailing NE directed sub-horizontal tectonic movement (Fig. 32A & D; see also fig. 21 in Section A) .

D_4 Deformation : Shortening during D_4 was oriented similar to that during D_2. Consequently, D_2 folds/foliation were further developed and re-used during D_4 to form composite structures (*e.g.* Meneilly, 1983; Tobisch & Paterson, 1988; Davis & Forde, 1994; Fig. 33A) that cannot be unambiguously separated from those of D_2. Local structural features indicate renewed shortening and flexural slip in the Telfer Dome, including:

1. S_2 intensification coupled with unfolding of D_3 folds (Fig. 33B), in the SW corner of Main Dome.
2. Reverse faulting arising from tightening of the southern Main Dome hinge (*cf.* Price & Cosgrove, 1990 - fig. 15.25, p. 400) at depth which is synchronous with mineralisation (*ie.* syn- to post-D_3; see Section C).
3. Large axial plane shears (West Dome Deeps) have a right stepping arrangement suggesting a sinistral movement component (Fig. 33A) that is consistent with oblique D_4 regional stresses (Fig. 33A). Additionally, many other faults throughout the dome exhibit an apparent sinistral offset, and associated disharmonic drag folds have a D_4 orientation (*e.g.* Pit 8).
4. Renewed flexural slip deformation occurred late in the mineralisation event, post-dating D_3 (Section C; see also below). Additionally, small asymmetric drag folds and low-angle reverse faults in reefal horizons exhibit both a dip-slip (reverse) and strike-slip (sinistral) movement component consistent with oblique flexural slip during D_4 (Fig. 33A). Small disharmonic folding of reefal horizons has a D_4 orientation indicating renewed shortening.

Structural Controls on Gold-Copper Mineralisation

Gold-copper mineralisation developed primarily in a series of stacked stratabound reefs within antiformal folds of the Telfer Dome, although individual reefs occupy distinct stratigraphic horizons in the sedimentary sequence. Within these horizons the degree and style of reef mineralisation is variable and appears to reflect its position within the dome. Consequently, two specific questions regarding reef formation arise;

1. What processes controlled the localisation of mineralisation into distinct stratigraphic horizons? and

2. What controlled the degree of reef development in these horizons?

Within the Telfer Dome there is a strong association between mineralisation and zones of deformation (see below), and paragenetic studies indicate that ore-veining was not controlled by pre-existing and compositionally favourable sites (Section C). Consequently, the above controls are likely to have represented the interaction between regional/orogenic deformation and the Telfer Dome resulting in specific internal-fold processes, such as gaping and layer-parallel shearing, that controlled mineralisation emplacement.

1. Site-specific Controls on Reef Development - Formation of Massive Quartz Veining.

Telfer reefs invariably formed in finely laminated and strongly interbedded sandstone-siltstone horizons that lay at the contacts of massive bedded units, and which would have represented mechanically incompetent layers within the mine sequence. Flexural slip generally follows flexural flow deformation, as fold limbs steepen (Tanner, 1989), and shearing strain associated with both would have been partitioned into mechanically incompetent layers (Johnson & Page, 1976; Tanner, 1989; Cosgrove, 1993). This occurred late in D_2 in the Telfer Dome prior to reefal silicification (Fig. 34A; Section C). Such partitioning would have increased the permeability of the reef horizons through internal deformation (see below) leading to increased fluid infiltration and quartz veining/silicification (Fig. 34A & B).

Potential mechanisms that created open space within the reefal horizons included hydraulic separation of bedding (Fig. 34B; *e.g.* Henderson et al., 1990; Cosgrove, 1993), asperity opening (*e.g.* Johnston, 1940, quoted in Hodgson, 1989;

Guha et al., 1983) along undulose bedding (Fig. 34B) as well as gaping induced by disharmonic folding (*cf.* Ramsay & Huber, 1987). Silica replacement would also have been enhanced by extensional micro-cracking (*e.g.* Cox & Etheridge, 1989; De Roo, 1989; Onasch, 1990) of pre-formed quartz veining. Sealing of the reefal horizons following silicification would have increased their competency causing further deformation to gape the contacts and promote internal fracturing allowing repeated fluid infiltration and further reef development (Fig. 34C).

2. Development of Mineralisation in the Reefs During Progressive Dome Modification - the Role of Changing Regional Deformation.

Following late D_2 silicification, hydrothermal dolomite-sulphide veining and alteration developed within the reefal horizons. The geometrical distribution of this veining and alteration suggests that its development was controlled by the interaction of D_3 and D_4 on the host structure.

D_3 Deformation : Mineralised reefs, and other bodies, in the Telfer deposit are generally thicker and best developed on NE fold limbs reflecting the favourable orientation of these limbs for gaping during D_3 (Fig. 35A). This is particularly evident in steeply dipping/overturned limb regions that exhibit strong brecciation (*e.g.* E-reefs & E-Reef Pods - Fig. 22; I-Reefs - Fig. 26) suggesting a rapid influx of ore-fluids and implosive breccia (*e.g.* Sibson, 1986; Forde & Bell, 1994) development concomitant with gaping (Fig. 35A & B). In contrast the SW limbs would have been under compression during D_3 thereby preventing gaping. However, reactivation of these would have caused gaping of sub-vertical fractures and faults (Fig. 35A), and tension vein formation (see Fig. 32). Additionally, the development of opposing shear couples (*e.g.* Forde & Bell, 1994) may have gaped the wallrock-reef margins extending reef silicification and mineralisation along the SW limb (Fig. 35C). These couples would have occurred where synthetic fold movement in homogeneous quartz veining interacted with antithetic shearing on reactivated bedding.

Gross reef development in Main Dome appears to be strongly controlled by D_3 (Fig. 35D). In near surface reefs, where Main Dome is open, the best development occurred on the NE limb. This is illustrated particularly well by the M-reefs which appear, from current drilling, to have only formed in a near-hinge and NE limb position (Fig. 35D). Reef horizons on the NE limb (*e.g.* M8, M30) were strongly dilational (Fig. 35E), and breccia veining along reef margins would have occurred where the quartz-wallrock contact gaped. In other reefs on the NE limb, thick quartz veining accommodated D_3 shortening/shearing strain resulting in internal fracturing of these competent horizons and subsequent infilling and replacement mineralisation in the quartz host (*e.g.* MVR, M10). At depth, the I-Reefs formed around a tightened and refolded (D_3 - see previous section) hinge, with spectacular breccia development in overturned (and gaped) regions on the NE limb (Fig. 35D). For all reefs, back-reef replacement along the SW limb would have been assisted by reactivation-induced shearing causing tensional cracking of pre-silicified horizons allowing ore-fluid infiltration (Fig. 35F).

D_4 Deformation : The second deformation that would have controlled mineralisation development was D_4 (Fig. 36A). Reefs on the NE limb of Main Dome exhibit a late phase of layer-parallel shear and associated mineralisation (*ie.* textural zonation in the MVR and M10 - see above). Such deformation, which post-dates the initial syn-D_3 mineralisation phases, is consistent with renewed shortening across the Telfer Dome during D_4 resulting in reactivation and shearing along the reef horizons (Fig. 36A). This shearing is heterogeneously developed along reef horizons on the NE limb suggesting that D_4 reactivation was strongly partitioned, and that its occurrence may also have been controlled by the undulose geometry of reef margins (*cf.* Hodgson, 1989). D_4 shearing strain is likely to have been relatively more intense along the steeper dipping NE limbs as these were more favourably oriented for accommodating shearing strain ($\approx 45°$ - Fig. 36B; Chapple & Sprang, 1974; Bayly, 1992), in contrast

to the SW limb (Fig. 36C), thereby producing a greater mineralisation effect on the NE limb.

Renewed D_4 shortening of the Telfer Dome would also have reactivated D_2 faults, fractures and veins/stockworks (Fig. 36A) and re-fractured pre-silicified pods (Fig. 36D) increasing fluid infiltration. Reactivation of large axial plane faults, particularly under elevated fluid pressures, would have promoted seismic pumping (Sibson et al., 1988) of ore-fluids though the deposit. The obliquity of D_4 regional stress against the dome is likely to have produced high-strain areas both in the NW and SE corners (Fig. 36E). Consequently, mineralisation enhancement through reactivation and shearing of reefal horizons would have been greater in these areas (Fig. 36F & G), and this is consistent with higher than average gold grades in both these areas (see above). Similarly, the large West Dome Deeps shears would have been favourably positioned for sinistral movement during D_4, in the process hosting mineralised breccia along their length (Fig. 36H).

Deposit Development During Polyphase Tectonic Deformation

The Telfer Deposit exhibits a progressive development of mineralisation that spanned both the major and subsequent deformational events in the host terrane (Fig. 37). This illustrates the potentially complex evolution of epigenetic fold-hosted deposits in other terranes that have undergone a protracted deformation history. At Telfer, deposit formation involved two major components:

1. *Development of the D_2 host-structure (Telfer Dome) as a "composite-fold" whose geometry was controlled by pre-D_2 fold phases.*
2. *The creation of "traps" for rising/circulating ore-fluids from structural opening (gaping) that arose from the interaction of post-D_2 deformation with the D_2 host-structure.*

The term "composite-fold" refers to a fold structure that contains elements from more than one deformational phase; similar to the principle of a composite foliation

(Tobisch, 1967; Meneilly, 1983; Tobisch & Paterson, 1988). Such folds develop when an initial fold phase acts as a nucleus for subsequent deformation that refolds and modifies the pre-existing fold. This process may happen repeatedly in a multiply deformed terrane so that the final structural expression is the product of a number of deformation events. This style of folding appears to be well developed in multiply deformed terranes (Meneilly, 1983; Davis & Forde, 1994; Davis, 1995; Bell & Hickey, 1997), particularly those that exhibit orthogonally overprinting (*ie*. vertical cleavage-forming followed by horizontal cleavage-forming) deformation sequences. At Telfer, the deposit host-structure (the Telfer Dome) was largely formed by D_2. However, both pre-D_2 and post-D_2 deformations controlled and modified the domal geometry such that it represents a composite fold.

Pre-D_2 folding produced specific geometric features in the Telfer Dome that appear to have enhanced the potential of the dome to host mineralisation. The most obvious of these was the D_2 domal asymmetry that arose from the refolding of a recumbent D_1 fold. This produced a steeper dipping NE F_2 limb that was more favourably oriented for gaping later during D_3. It is also likely that D_1 fold axes exerted some control on the axial development/localisation of D_2 folds. In West Dome this appears to have resulted in numerous smaller parasitic and asymmetric folds/flexures that subsequently became the nucleus for E-reef and Pod development (*e.g.* Figs 22 & 35). A similar effect, but at a larger scale, is interpreted to have produced the asymmetry in Main Dome that was so conducive to stratabound reef formation on the NE limb. Therefore, the initial stages of composite fold development in the Telfer Dome produced a resultant geometry in this host-structure that enhanced the subsequent emplacement of epigenetic mineralisation.

The development of mineralisation within the Telfer deposit was controlled by the effects of post-D_2 deformations as these were synchronous with ore-fluid release from the rock sequence (Section C). Mineralisation precipitation from the ore-fluids occurred in distinct areas of the Telfer Dome, which correspond to structural elements expected to gape during the mineralising deformation (*e.g.* the NE limbs of D_2 folds during D_3). Such gaping is likely to occur where a high-angular difference exists between the pre-existing structural element and the subsequent strain of the mineralising deformation. The creation of even small amounts of opening by this method would cause regionally pervasive ore-fluids, at elevated pressures, to enter such sites thus increasing the fluid flux and precipitative potential (*e.g.* Ferry & Dipple, 1991). High-pressure fluid influx into discretely gaped zones may also produce implosive brecciation (*e.g.* Sibson, 1986; Bell et al., 1988; Forde & Bell, 1994) that would locally increase structural permeability promoting further ore-fluid infiltration.

Cyclical fluid-pressure fluctuation appears to be common in many lode-gold deposits (Robert & Brown, 1986; Sibson et al., 1988; Boullier & Robert, 1992; Sibson, 1992) and is viewed as a major contributory factor in mineralisation development. This is because the periodic sealing of fluid channelways generates supra-lithostatic fluid pressures that promote hydrofracturing (*e.g.* Phillips, 1972) and the creation of ore-fluid "trap-sites" (Kerrich & Allison, 1978; Cox et al., 1991b). However, the creation of ore-fluid trap-sites may also be controlled by deformation as well as, or instead of, fluid pressure. In rocks where the ore-fluid is at a high pressure, such as those being regionally metamorphosed/deformed, deformation-induced gaping and/or micro-cracking creates fluid sinks (*e.g.* Ridley, 1993) into which the fluids rapidly move. This creates implosive brecciation (Bell et al., 1989; Forde & Bell, 1994) that further increases the structural permeability. Although high-pressure fluids assist implosive breccia formation, the major control on such opening appears to be structural. Consequently, the predictive geometry of crustal deformation means that ore-body geometries can be assessed more accurately than for those inferred from static fluid pressure increase alone.

Strain associated with late orogenic deformation is commonly heterogeneously partitioned in the Earth's crust, and in many cases a nucleus is required to sufficiently partition and concentrate the strain from such deformation into localised zones that gape/dilate. One potential nucleus is thick quartz veining and silicification, which is a common component of lode-gold deposits. Such veining provides a rigid competent nucleus in which fracturing and micro-cracking, produced by subsequent mineralising deformations, will concentrate thus focussing ore-fluid infiltration (Hayward, 1992; Vearncombe, 1993). As well, the competency contrast between quartz veining and the wallrocks may promote gaping at the contact between the two (Forde & Bell, 1994). In both cases, thick quartz veining acts as a mechanical nucleus for the effects of subsequent deformation events. This is borne out in paragenetic and mineralisation timing studies in lode-gold deposits that indicate gold mineralisation developed in deformed quartz veining (Forde, 1991; Forde & Bell, 1994; Section C). Therefore the development of thick quartz veining in a regional-scale host-structure can provide an important control on mineralisation development.

To summarise, the Telfer deposit exhibits a complex history of development that represents the interaction of specific crustal processes. One process involved the host-structure (the Telfer Dome) forming as a composite fold during multiple episodes of host-terrane deformation. These deformations progressively developed and then modified the dome producing, late in the tectonic sequence, local zones of gaping and dilation at geometrically favourable sites. This coincided with another crustal process; the late release from the regional rock sequence, and focussing, of a pervasive ore-fluid phase into the gaped sites (Section C). Such focussing, of both the ore-fluid and the effects of late-orogenic deformation, was facilitated by the development of extensive quartz veining, which appears to have been an integral part of host-structure formation. The result was the emplacement and concentration of precious metal mineralisation in the Telfer Dome.

SIGNIFICANCE

Implications for Deposit/Regional Exploration Drilling

The subtle complexities of ore-body geometry in the Telfer Deposit have important implications for exploration strategies designed to locate further fold-hosted mineralisation. Of most significance is the potential for an asymmetric mineralisation distribution within a regional fold structure, which occurs both in cross-section and in plan. The preferential gaping of the NE dome limbs by D_3 suggests that drilling programs in other domes should target this limb. In Main Dome, this limb has an approximate 45° dip indicating that D_3 gaping would occur in moderately open folds. However, gaping is likely to have been enhanced in those folds that had a steeper NE limb, such as those that refolded a D_1 fold. Consequently, folds that are markedly asymmetric with steep NE limbs would make ideal targets.

Exploration strategies also need to consider the map-view asymmetry of reef mineralisation arising from D_4. This preferential enrichment of mineralisation in opposite corners of the dome arose from the intensification of D_4 strain into these areas of the fold. Field indicators for such higher strain include an increased fault density, as well as strong cleavage (S_2) development, in both the SW and NE corners of regional domes. In these cases, drilling programs should target the NW and SE corners of the dome, especially the latter where both D_3 and D_4 effects would be enhanced.

CONCLUSIONS

Structural analysis of the Telfer Dome, a large domal antiform hosting the Telfer Au-Cu deposit, indicates that it developed as a composite fold structure during protracted regional deformation. Although the main antiformal structure was produced by bulk inhomogeneous crustal shortening during D_2, its nucleation appears to have been controlled by recumbent monoclinal D_1 folds. Refolding of the D_1 folds by D_2 produced a dome that was markedly asymmetric (steep NE limb), and which has a SW inclined geometric axial plane that is transected by vertical S_2. Post-D_2 deformation locally modified the Telfer Dome by refolding and overturning domal hinges (D_3), and generating renewed flexural slip deformation as the fold was further shortened during D_4.

Gold-copper mineralisation is primarily hosted in the Telfer Dome as stratabound quartz-carbonate-sulphide reefs, with subsidiary stockwork, sheeted vein and pod mineralisation. The reefs form a vertical stacked series along the domal axial plane, and individual reefs developed in specific stratigraphic horizons that were mechanically anisotropic and thus favourable for the intensification of flexural slip/shear deformation. Initial quartz veining in the reefal horizons occurred late in D_2 as a consequence of flexural-slip deformation. This silicification assisted the nucleation of subsequent mineralising deformations (D_3 & D_4), resulting in gaping and fracturing of the quartz host that provided suitable precipitative sites for sulphide-gold mineralisation. D_3 preferentially gaped the NE dome limbs causing thick quartz-carbonate-sulphide reef development, whilst renewed layer-parallel shearing in the reefs during D_4 shortening of the Telfer Dome enhanced this mineralisation.

The effects of D_3 and D_4 on the Telfer Dome have produced ore-body geometries that differ from those expected to be developed by the dome-forming deformation, D_2. This illustrates the importance of late orogenic deformation in controlling mineralisation development in regional host-structures. Such deformation, can, through gaping of internal-fold structures, produce discrete zones of ore-fluid infiltration and subsequent mineralisation. As a result, ore-body geometries may be significantly different from those predicted by major terrane folding processes alone and this must be considered when developing exploration strategies.

SECTION E

Exploration Models for Fold-hosted and Other Epigenetic Mineralisation Styles in the NE Paterson Province.

SECTION E

Exploration Models for Fold-hosted and Other Epigenetic Mineralisation Styles in the NE Paterson Province.

ABSTRACT

The Paterson Province consists of a typical compressional fold-belt that formed during polyphase deformation arising from gravity collapse orogenesis in the late-Proterozoic. The major terrane deformation, D_2, produced regional-scale folds that are important hosts for epigenetic mineralisation (e.g. Nifty - Cu & Telfer - Au). Earlier folding controlled the nucleation and subsequent geometry of D_2 folds resulting in inclined axial planes and asymmetric fold shapes. Post-D_2 deformation events (particularly D_3 and D_4) subtly modified the D_2 domes by refolding suitably oriented internal structures and reactivating bedding. D_4 involved horizontal compression, which was similarly oriented to D_2, and produced a weakly transpressive environment in which regional fold structures were re-used and faulted. This transpressive deformation is evident from a greater than average density of faults that lie at low-angles to the regional structural grain.

Post-D_2 fold modification was broadly coeval with mineralisation in the region. Consequently, structural models utilising this have been developed that can allow one to identify and predict further mineralisation occurrences throughout the province. Similar models can be generated for epigenetic mineralisation in shear zone systems and at granite-country rock contacts. All of these emphasise the relationship between heterogeneous (and discrete) deformation phases, and their role in causing selective gaping/dilation of pre-existing structural elements within regional fold structures. They have a wide range of applicability at different scales, from orogen reconnaissance to detailed prospect evaluation, allowing a greater precision in targeting sub-surface exploration programs. Discrete concentrations of metal mineralisation (ore-deposits) may be localised in the earth's crust by the heterogeneous effects of late orogenic deformation on pre-folded terranes.

INTRODUCTION

Sub-greenschist/greenschist grade fold belts that formed during convergent orogenesis are ideal host terranes for epigenetic mesothermal lode-gold deposits (*e.g.* Cox et al., 1983; Mawer, 1986; Goldfarb et al., 1986; Mueller et al., 1988; Colvine, 1989; Vearncombe et al., 1989; Kerrich & Cassidy, 1994). Within these terranes lode-gold deposits are invariably hosted in regional-scale structures, such as folds or shear zone systems (Boulter et al., 1987; Eisenlohr et al., 1989), which formed during the major period of terrane deformation. Consequently, models accounting for the development and exploration of lode-gold deposits often focus on the kinematic conditions inferred for host-structure development. A typical example is a mineralised saddle-reef (*e.g.* Chace, 1949) where quartz veining and gold mineralisation are commonly interpreted to form synchronously with folding (Chace, 1949; Cox et al., 1983; Ramsay & Huber, 1987; Hodgson, 1989; Cox et al., 1991b).

Recent studies have however, demonstrated that mineralisation development may post-date the major period of terrane deformation/accretion (Kontak et al., 1990; Jemielita et al., 1990; Robert, 1990), and in some instances is associated with late orogenic deformation phases (Forde, 1991; Section C). As a consequence, the development of mineralisation in regional-scale host-structures is likely to be controlled by kinematic conditions different to those that produced host-structure. In some terranes this is manifest by increased fluid flows during changing tectonic conditions (Kerrich & Wyman, 1990; Elder & Cashman, 1992) and mineralisation development in reactivated/refolded regional host-structures (*e.g.* Vearncombe et al., 1989; Forde & Bell, 1994; Section D). Therefore, exploration models based simply on the kinematics of host-structure formation may neglect the potentially different crustal-strain conditions under which mineralisation actually occurred.

The Paterson Province is one such terrane in which epigenetic Au-Cu mineralisation was associated with late orogenic deformation. In the northeastern area of this province fold-hosted mineralisation developed in doubly-plunging antiformal domes, and one of these hosts the Telfer Au-Cu deposit. Although this dome formed during D_2 (Section A, D), mineralisation development was controlled by two subsequent deformations, D_3 and D_4 (Section D). Structural mapping of the area encompassing these domes suggests that they were progressively developed during protracted orogenic deformation. This history enables conceptual structure-based models to be proposed for a number of epigenetic mineralisation styles in the Paterson Province to assist the identification of exploration targets and the planning of drilling programs. Such models are particularly relevant for deep drilling programs that target sub-surface mineralisation that lacks a surficial geochemical anomaly.

REGIONAL GEOLOGY AND PREVIOUS INVESTIGATIONS

The NE region of the Paterson Province (see fig. 1 - Section A), in the northwest of Western Australia, contains upper units of the middle- to late-Proterozoic Yeneena Group. These rocks, which consist of alternate calcareous and pelitic/psammitic units (Table 1), were deposited in a marine environment that prograded to the NE (Turner, 1982). They were subsequently deformed and metamorphosed (sub-greenschist facies) during the late Proterozoic Paterson Orogeny (≈640 - 600Ma). This was accompanied by numerous syn- and post-tectonic granitoid, gabbro and dolerite dike intrusions (*e.g.* Hall & Berry, 1989; Goellnicht et al., 1991; Goellnicht, 1992), and associated epigenetic mineralisation, across the region. Regional deformation comprised six orthogonally overprinting events suggestive of compressional/gravity collapse tectonism (see Section A).

Initial mapping of this region identified a dome and basin setting comprising upright NW-SE trending anticlinal and synclinal folds (Chin & Hickman, 1977; Chin et al., 1982) that were produced during the major terrane deformation, D_2 (Section A). Subsequent smaller scale mapping of exploration tenements by mining company personnel and consultants has lead to a variety of tectonic models being proposed for the region, and in particular formation of the domes (*e.g.* table 1 - Section D), with most consensus on compressive fold-thrust (Hill, 1989; Windh, 1991) or basement wrench tectonics (Harris, 1985; Goellnicht et al., 1989). In both models dome formation was considered the major and only significant deformational event

Unit	*Lithology*	*Thickness (m)*
Kaliranu Beds	Silty dolomite, shale, minor sandstone	> 350
Wilki Quartzite	Massive quartzite, minor shale/sandy interbeds	1000
Puntapunta Fm.	Medium to thickly bedded dolomite, limestone, calc-arenite, sand-silt interbeds in transitional areas, numerous small scale concave and sigmoidal trough cosets (a).	2000
Telfer Fm.	Sandstone, siltstone, shale	600-700 (b)
Malu Quartzite	Sandstone, quartzite	700 - 1000
Isdell Fm.	Dolomitic shale, dolomite, high energy calc-arenite	2000

Table 1 - Summary of the stratigraphic sequence in the north-eastern region of the Paterson Province (after Turner, 1982; Williams, 1990a; personal observation). (a) terminology after Jacob, 1973. (b) See Section D for detailed summary of Telfer Formation.

Exploration strategies in the NE Paterson Province were initially based on the early syngenetic exhalative model for mineralisation genesis in the Telfer Au deposit proposed by Turner (1982); hence regional antiformal domes that exposed these rocks were targeted. However, compelling evidence for an epigenetic genesis (Goellnicht, 1987; Goellnicht et al., 1989; Hall & Berry, 1989; Rowins, 1994; Section C), coupled with the discovery of mineralisation in other sedimentary formations throughout the region, has meant that exploration strategies now focus on locating favourable structural traps, such as regional domes and faults/shears that are spatially associated with igneous intrusions.

REGIONAL STRUCTURAL GEOLOGY/GEOMETRY

The regional structural geometry is characterised by laterally extensive and uniformly dipping NW-SE trending sequences of both Wilki and Malu Quartzite Formations (Fig. 1; *e.g.* Karakutikati Range). These sequences commonly form the limbs of upright sub-horizontally plunging synclinoria (*e.g.* O'Callaghans Syncline, Colonel Syncline - Fig. 1) and anticlinoria that have 10-15km wavelengths. Between these large folds the sequence is folded into doubly plunging anticlinal domes and synclinal culminations that range from 5-8kms in axial length (regional domes) to 1-2km long parasitic/ pod (*e.g.* Price & Cosgrove, 1990, p. 380) folds on dome limbs. Seismic sections indicate that regional folding is upright to steeply SW inclined and open (Fig. 2).

Antiformal and synformal folds retain their form along the strike of their axial planes resulting in laterally continuous fold axial traces across the region. However, some contain lower-order folding that is arranged in an en échelon manner resulting in complex internal dome geometries (*e.g.* Camp/Pajero Domes - see below; Telfer Dome - Section D). Fold axis plunges at either end of the domes are moderate, ranging from 10-35°, although locally some steepening occurs, and axial trends exhibit a weak curvature across the district ranging from NNW-SSE trending in the north-east to WNW-SSE in the south-west (Fig. 1). Fold styles commonly reflect the predominant lithology with smaller and tighter structures in finely bedded units, such as the Isdell Formation south of the Karakutikati Range (Fig. 1).

Wilki "ring" Structure

A large ring shaped feature, comprising a circular shaped exposure of Wilki Quartzite north of Trotmans Dome (Fig. 1), forms a significant regional structure that appears to be a disrupted synformal D_2 fold. Bedding around the margins dips steeply inwards and a relict synclinal closure is preserved in the lower SW corner (Fig. 3). D_2 folds and cleavage anastomose around the structure (Fig. 1 & 3) suggesting these were further shortened around a large rigid body. Sporadic outcrops of monzogranite (Wilki Granite - see also below), coupled with local contact aureoles (garnet/cordierite bearing schist) and strong internal magnetic signatures (Fig. 3) suggest that this is a large granitoid. The regional (S_2) cleavage, axial plane to domal folding, overprints the central SE and NW dipping limbs (Fig. 3), which in addition to the relict D_2-F_2 closures suggests that granite intrusion occurred early in D_2.

Faulting

Most faulting throughout the region is parallel, or at a low angle, to the bedding/fold trend, but is commonly difficult to observe in outcrop and requires aerial images (*e.g.* photographs, radar, magnetics) to identify subtle bedding truncations. Three main trends are observed:

1. NW-SE trending (fold/bedding parallel); these are the most common and occur along fold limbs and through axial planes. They are particularly frequent in the O'Callaghans Corridor (see below). Most are sub-vertical or steeply SW dipping and exhibit components of reverse and sinistral movement (Fig. 1).
2. NNW striking; a sub-set of the above, these truncate regional structures (*e.g.* Mathews Dome) at a low angle, commonly passing into parallelism with bedding (Fig. 1), and exhibit similar movement components.
3. NNE - SSW & N-S trending; these truncate regional folding at a high angle. Many of these are normal faults (*e.g.* Graben Fault - Section D) and are probably related to late N-S compressional deformation (D_{5b} - Section A). Others however, appear to have been produced by extensional stresses arising from low angle transpressive faulting of the sequence (*e.g.* extensional faulting along the Karakutikati Range - see O'Callaghans Corridor below).

Granitoid/Igneous Emplacement

Surface mapping, coupled with geophysical data interpretation, suggest that approximately 20% of the region is underlain by granitoids and other igneous bodies (Goellnicht, 1992). These consist mainly of highly fractionated metaluminous I-type granitoids (Fig. 4) that were derived from partial melting of lower crustal sources (Goellnicht et al., 1991; Goellnicht, 1992). The granitoids form large outcropping/sub-cropping complexes, whose positions can be inferred from magnetic geophysical data (Map 2; see also Fig. 1 and below). One complex, the Mt. Crofton Trend (Fig. 1; Map 2), is a dumbbell shaped batholith that trends N-S across the region truncating regional (D_2) folding. Other individual plutons and smaller stocks/cupola's intrude the cores of regional folds (*e.g.* 17 Mile Hill, O'Callaghans, Hasties) that are commonly synclines (Fig. 1; Map 2).

Contact aureoles are of limited width, suggesting a relatively shallow emplacement, and overprint regional metamorphic assemblages (Goellnicht, 1992; Fig. 4); this indicates a late- to post-D_2 emplacement timing. However, some granitoids are locally strongly foliated (*e.g.* Minyari Gneiss, Wilki Granite - see Figs 1 & 4 for location) suggesting an early D_2 emplacement. The sub-surface form of the granitoids is conjectural, although gravity data suggest that many are laccolithic (*e.g.* Mt Crofton, Wilki Granite - Goellnicht, 1992; Fig. 4), based on the absence of an underlying gravity depression. Large gravity depressions, with NW-SE trending long axes, occur under the O'Callaghans Granite and the northeastern end of the Mt Crofton Trend and may represent root zones for the batholith.

Sub surface dolerite and gabbros are commonly only observed in drillholes; these comprise both syn-tectonic bodies that have been affected by regional metamorphism (Goellnicht, 1992), and others that post-date metamorphism. Examples of the latter include prominent N-S trending dolerite dikes that outcrop infrequently but are readily apparent as magnetic lineaments on aeromagnetic images (*e.g.* Hardy's Dike, Minyari Dolerites - Fig. 1). These truncate regional folds and appear to be related to N-S compressional deformation (D_{5b}) that post-dates the Paterson Orogeny.

Lineaments

The major lineament orientations in the region include regional scale N, NNW to N, NNE and ENE to ESE trending sets, with district scale NE and NW striking sets (Harris, 1985; Langsford, 1989). Leeson (1992) identified three major Western Australian lineament corridors (*e.g.* O' Driscoll, 1981) that intersect within the Paterson Province. He also observed two regional linears, parallel to the gross structural trend, that defined the boundaries of a cohesive structural corridor that broadly corresponds with the O'Callaghans Corridor (see below). A lesser set of

NNW trending discontinuous linears that overprinted the major tectonic corridor were also observed (Leeson, 1992).

O'Callaghans Corridor

The O'Callaghans Corridor is a NW-SE trending linear zone that encompasses a large synclinorium (O'Callaghans Syncline - Fig. 5) and associated regional domes (Fig. 5; Map 2 & 2B). The SE end of this synclinorium lies between two anticlinal domes (Trotmans & Connaughtons - Fig. 5) and its SW limb is the Karakutikati Range. Folding within it is open with a steeply SW dipping axial plane (Fig. 2 - Seismic Line 2) and is doubly plunging, with numerous parasitic domes and folds developed immediately adjacent to the Karakutikati Range (*e.g.* Fallows Field - Fig. 5 & Map 2B - see further). The O'Callaghans Granite intruded the synclinorium closure (Fig. 5) late in the regional folding episode (ie. late- to post-D_2) and seismic interpretation indicates that the axial trace was disrupted resulting in a false closure (Fig. 2 - Seismic Line 4). South of the Karakutikati Range, regional folds have a horizontal plunge and exhibit numerous parasitic and disharmonic folds, some of which are attenuated along the NE limb.

The most notable features within this corridor are the numerous low-angle discordancies and faulted contacts (Fig. 5; Map 2 & 2B), most of which can only be observed on high-resolution aero-magnetic images. These discordancies occur along both fold axial planes (*e.g.* O'Callaghans syncline - Fig. 5 & Map 2B) and fold limbs (*e.g.* SW limb of Connaughtons Dome), and commonly pass into parallelism with bedding. The smaller Mathews Dome, an axial extension of Connaughtons Dome, is truncated by low- and high-angle faults and appears to unfold into the Karakutikati Range. Domal folds in the Puntapunta Formation, immediately NE of the Karakutikati Range, are also commonly truncated by axial-plane parallel faults (*e.g.* Fallows Field - Fig. 5 & Map 2B; see also below). Additionally, high resolution aero-magnetic images suggest that bedding is locally slightly discordant against the NE side of the Karakutikati Range.

As most truncations of bedding and regional folds are not clearly exposed, determining the sense of displacement on the faults is difficult. However, where it can be determined it involves combined sinistral and reverse movement. In plan view the movement sense suggested is invariably sinistral. The larger of these truncations/discordancies lie parallel to the Karakutikati Range suggesting that this region provided a competent locus for much of this faulting. The competent quartzite/sandstone formations that comprise the Karakutikati Range are cut by high-angle (ENE-WSW trending) oblique normal faults (Fig. 5) suggesting this was subjected to sinistral transpression along its length. This is consistent with the overall sinistral-reverse displacement suggested by faulting in this region.

Karakutikati Range

The Karakutikati (Parallel) Range is a laterally extensive (30-40km) steeply NE dipping sequence of quartzite/sandstone formations, lying on the SW limb of the O'Callaghans Syncline, which is not repeated on the limb of an anticline to the south (Fig. 5). Stratigraphic analysis indicates that the Malu Formation is conformable with the underlying Isdell Formation and thins S/SW across the district (Turner, 1982) suggesting that the Karakutikati Range sequence may wedge out to the SW (Chin et al., 1982). However, the lateral strike continuity of the sequence suggests that it may instead be a post-lithification structural discontinuity. Cleavage (S_2) - bedding relationships and younging data indicate a consistent SW vergence across the range.

Fallows Field

The Fallows Field (Au - Cu ± Sn) deposit, the only other active mine in the region, is hosted in a parasitic anticline within the Puntapunta Formation NE of the Karakutikati Range (Fig. 5 & 6A). This anticline is asymmetric (Fig. 6B) with a sub-vertical to NE inclined axial plane and locally overturned short SW limb. Fold axes are sub-horizontal through the mine but plunge steeply at either end. The surrounding units exhibit numerous low-angle discordancies across which bedding and folds are offset (Fig. 6A), a feature common throughout the Puntapunta Formation in this area (Turner, 1982). These discordancies, which are interpreted to be faults, suggest that "shuffling" of the sequence occurred post- regional folding, and stratigraphic offsets suggest this had a sinistral component of movement.

Gold mineralisation (silica-dolomite-sulphide - Johnson, 1991) is hosted within a massive calc-arenite unit that is tightly folded in the core of the anticline (Fig. 6B). Finely laminated calcareous/dolomitic siltstones that enclose this unit are generally not mineralised. Within the massive unit, mineralisation occurs in two major linear stratiform pods that are controlled by the anticline hinge (C. Moorhead pers. comm.1993), and which strike slightly oblique to the axial trend (Fig. 6C). Bedding within the ore/hinge zones is also oblique (≈ 25° anti-clockwise) to that on the NE flank of the anticline. Low grade gold mineralisation is also hosted in a wide breccia zone along the fold axial plane (Johnson, 1991).

17 Mile Hill Corridor

The NW-SE trending 17 Mile Hill Corridor lies between the O' Callaghans and Colonel Synclines (Fig. 1; Map 2). The corridor contains large regional synclinal and anticlinal folds and is truncated at either end by large granitoid intrusions, the Mt. Crofton Trend in the NW and the Wilki Granite to the SE (Fig. 7; Map 2). The emplacement of granitoids into the corridor, particularly at either end, caused fold axial trend to be distorted resulting in a pronounced curvilinear structural trend (Fig. 7). In the NW end of the corridor fold axes and bedding are sharply deflected towards the NE against the Mt. Crofton Trend granitoids, whilst in the SE regional fold axial traces curve into the Wilki "ring" structure (Fig. 7).

The sedimentary sequence in the 17 Mile Hill corridor is regularly folded with the large 17 Mile Hill - Thompsons East domal complex dominating the northeastern part of the corridor, whilst to the south the Telfer and Tims Domes lie along the anticlinal axial trend (Fig. 7). The synclinal closure between these contains small parasitic domes (Fig. 9), and one of these, which lies in the central hinge area, hosts the Thompsons deposit (see below). The sequence in this corridor is less truncated by low-angle faulting in contrast to the O'Callaghans Corridor. Larger domes are occasionally faulted along their axial planes (*e.g.* Camp Dome) or limbs, but most folds and bedding are continuous. The intensity of faulting that lies at a low-angle to the regional trend increases in the S/SE margin marking the beginning of the O'Callaghans Corridor.

Camp-Pajero Domal Complex

The large Camp-Pajero dome exhibits a complex geometry comprising a number of parasitic domal folds (Fig. 7). Surface folding patterns are complex with anticlinal folds merging into the limbs of other anticlines, and synclinal folds are commonly faulted/sheared out. Regional folds are cross folded in the north-western end of Camp Dome resulting in Type 1 (Ramsay, 1967; Thiessen & Means, 1980) interference patterns, and an associated weak crenulation cleavage (see also Chin et al., 1982). This cross-folding has an axial trend that is consistent with formation during NW-SE directed D_5 compression, and is also similar to the trend of the nearby Mt. Crofton Granite Trend.

Limited diamond drilling in the Camp Dome indicates that it is asymmetric and has a SW inclined axial plane (Fig. 8A; *cf.* Section D). Cleavage, marked in drillcore by strongly aligned biotite and pyrite, exhibits a variable SW dip and has been interpreted as overprinting the domal fold (Anderson, 1989; Rowins, 1994). However, its angular relationship to bedding is variable across the dome (Fig. 8A), and within each limb, suggesting that two separate cleavages may be represented in the drill-core data. The geometry of these (low- and high-angle to bedding) is similar to the S_1-S_2 relationship across regional D_2 folds (*e.g.* fig. 10 - Section A), and coupled with the inclined axial plane suggests that the Camp Dome may be a refolded (D_2) D_1 monoclinal fold (Fig. 8C).

Thompsons Prospect

The Thompsons prospect is a small resource (≈ 32,000oz.- Rowins, 1994) that is exposed as a series of gossanous stratabound reefs hosted in silty dolomite and sandstone units of the Puntapunta Formation. Whilst the prospect is located along the axis of a regional syncline (Fig. 7; Map 2), examination of bedding-cleavage relationships indicate that it encompasses a small anticlinal closure. This is borne out with interpretation of high-resolution aero-magnetic images that indicate the anticline is a parasitic closure lying within the syncline hinge.

MINERALISATION STYLES

A number of other deposits/prospects occur throughout the region (Fig. 1, 9) and can be divided into three broad types:

1. Epigenetic quartz-carbonate-sulphide vein hosted prospects/deposits (*e.g.* Big Tree, Black Hills, Thompsons and Triangle) that comprise stratabound and discordant veining with associated alteration (Fig. 9).
2. A large porphyry Cu (+ Au) deposit (17 Mile Hill - Rowins, 1994) hosted within the Camp Dome. Mineralisation is hosted within fracture sets that occur in a well-zoned alteration halo overlying an inferred granite intrusion (Anderson, 1989; Rowins, 1994).
3. Granite-related skarn mineralisation (*e.g.* Minyari, O'Callaghans). This is locally hosted within fracture and vein sets with mineralisation accompanying pervasive metasomatism.

DEVELOPMENT AND MODIFICATION OF REGIONAL FOLDING

The gross regional structural geometry is controlled by pervasive upright D_2 compressional folding. However, pre-D_2 structures appear to exert a control on the nucleation of D_2 folds, whilst post-D_2 deformation modifies the gross regional fold geometry.

Pre-D_2 Deformation (D_{EF}, D_1)

Mapping across the region suggests that some stratigraphic units have a variable thickness across the region. In some areas this may represent primary sedimentary variations (*e.g.* Chin et al., 1982; Turner, 1982), however, it may also reflect faulting. In particular, faulting during either D_{EF} (see Section A) or D_1 could include discrete low-angle thrusts (Fig. 10A), tectonic slides (Hutton, 1979; 1981) or shallow extensional faults (Dewey, 1988; Fig. 10B), which nucleated along stratigraphic/lithologic contacts. This would result in juxtaposition and possible local thinning of sedimentary units (Fig. 10A & B), which when folded by D_2 would produce variable unit thicknesses across regional folds (Fig. 10A & B). As thin skinned thrusting is not a feature of the region (Fig. 2), such faulting would probably have been only metres to hundreds of metres in scale causing only local stratigraphic variation.

Formation of the Karakutikati Range

The lateral continuity of the Karrakutikati Range suggests that it may be a large structural feature/discontinuity that was subsequently folded by D_2. Two possible structures, which would also account for the absence of the sequence on the SW limb of an antiform to the southwest, are either a primary growth fault (Fig. 10C) or alternatively a thrust fault cutting up sequence towards the SW (Fig. 10D). However, an alternative explanation is that it may have been part of a regional-scale D_1 monoclinal fold/flexure that was folded by D_2. This could have occurred following inclination of the regional sedimentary sequence prior to D_1 due to shortening D_{EF} (Section A) that produced broad anticlinal warps across the Paterson Province, with the NE region lying along the NE limb of one of these (Fig. 10E). Inclined bedding on this limb would be favourably oriented for monocline formation by SW directed thrusting during D_1 (Fig. 10F; see also Section A), and F_1 folds could subsequently have acted as nuclei for D_2 folding (Fig. 10G).

Folding of a large D_1 monocline, as part of the O'Callaghans Syncline, could produce the Karakutikati Range as a relict short NE F_1 limb preserved within the D_2 fold (Fig. 10G), as well as imparting an inclined geometric axial plane to the D_2 fold (Fig. 10G; see also Fig. 8C). Such a process may also have contributed to the formation of other laterally continuous quartzite units in the region (*e.g.* SW limb of the Colonel Syncline), and is consistent with the asymmetric shape of some domes through the preservation of a steep NE F_1 limb. Inclination of the sequence prior to D_1, as inferred above, is also consistent with the general angular relationship of S_1 to bedding (Section A), which suggests that bedding was inclined before the formation of horizontal S_1. Another geometric consequence of prior D_{EP}/D_1 folding would be a stepping downwards towards the NE of the Yeneena Sequence, amplifying that which already existed from prograding sedimentary deposition. Erosion would thus have removed the Karakutikati Range formations to the SW explaining their absence south of the region, although this may also be due to primary depositional thinning of the units (*e.g.* Chin et al., 1982).

Deformation Event D_2 and Models for Regional Dome Formation

Previous models of dome formation including basement wrench-controlled and, more recently, compressive fold-thrust tectonic styles (see table 1 - Section D) are generally incompatible with the regional structural geometry. In particular, antiformal folds retain their form along axial trends as opposed to an en échelon arrangement that would typify a wrench tectonic environment (*e.g.* Sylvester, 1988). Similarly, compressive thrust-fold (Harding & Lowell, 1979; Jones, 1987) or fault propagation folding (*e.g.* Faill, 1973; Williams & Chapman, 1983) is not applicable because regional deformation is not thin-skinned. Seismic interpretations (Fig. 2) indicate that regional folds are not truncated by faults nor do they exhibit forelimb deformation (*e.g.* Alonso & Teixell, 1992), which, coupled with the absence of laterally extensive tip-lines (Fig. 1), would mark emergent thrust faults (Boyer & Elliot, 1982).

In summary, regional domal folding is interpreted to have formed as a result of horizontal compressive strain during inhomogeneous crustal shortening of the Paterson Orogen. The doubly-plunging nature of folding, spectacularly displayed in the Camp-Pajero Dome complex, is interpreted as resulting from strain heterogeneity (Sanderson, 1973) rather than as a separate cross-folding event. Many smaller-scale doubly plunging folds formed in relatively incompetent interbedded strata that lay within larger synclinoria and anticlinoria, whose limbs are defined by thick competent quartzite units (Fig. 1), typifying the control exerted by lithological competency on the scale and character of regional folding.

Post-D_2 Deformation (Regional Dome Modification)

Modification of D_2 folds and their regional distribution occurred during the D_3 and D_4 deformations. D_3 has refolded dome hinges and caused gaping of the NE limbs in some domes (Section D). However, these effects can only be identified in areas of sufficient three dimensional exposure. For most regional folds the limited exposure has made identification of the overprinting effects of D_3 refolding extremely difficult. Limited deep drilling of the Big Tree deposit, within Connaughtons Dome, indicates that fold axial planes are vertical (H. Cierlitza, pers. comm., 1994) suggesting that D_3 had little effect on this particular dome.

D_4 is interpreted as having initiated and reactivated faults throughout the region. The majority of regional faults lie at a low angle to the structural grain and are controlled by the D_2 fold geometry (*ie.* they lie along axial-planes or fold limbs). This suggests that these faults may have formed in late-D_2 or D_4, although distinguishing which event produced them is difficult owing to the lack of overprinting criteria. Compressive stresses during D_2 would have produced faults that exhibited reverse movements with minimal transcurrent component (Fig. 11A). However, many of the faults exhibit a consistent plan-view sinistral displacement (Fig. 11A) in addition to reverse movement, suggesting that they formed during D_4 when compressive stresses were slightly oblique to the regional structural grain (Fig. 11B).

Renewed compression during D_4 would have been weakly transpressive (Fig. 11B), resulting in further shortening and reactivation/re-use of regional folds. Coupled with sinistral faulting, this appears to have produced a "shuffling" of the folded sequence that was accommodated through movement on low-angle faults. Although some faults may have initiated during D_2, these would have also been reactivated during D_4, particularly as the regional stresses would have been oblique to the pre-existing fault plane. The regional variation in fault density, with some linear corridors exhibiting a relatively high number of faults (*e.g.* O'Callaghans Corridor), suggests that the weakly transpressive D_4 strain was partitioned by the pre-existing structural architecture.

Multiple Granite Emplacement in the NE Paterson Province?

D_5 cross-folding in the Camp Dome has an NE-SW trending axial orientation parallel to the Mt. Crofton Granite Trend suggesting a potential association between D_5 and regional granitoid emplacement. Inferred root zones for the regional granitoids (see above) have long axes that would have lain perpendicular to the regional σ_1 direction during D_2 (Fig. 12A), suggestive of emplacement into ductile crust (Vigneresse, 1995). However, the Mt Crofton Granite Trend lies parallel to the regional D_2 σ_1, suggesting shallower and brittle (Andersonian) crustal conditions during emplacement, and thus conflicting models for regional granitoid emplacement are indicated (Fig. 12B & C).

One way to reconcile the different emplacement indicators is to evoke two separate stages of granite intrusion. Initially, plutons associated with the root zones and those within the cores of regional folds (*e.g.* 17 Mile Hill, O'Callaghans and Wilki granitoids) would have been emplaced into ductile crust during D_2-D_4 (the Paterson Orogeny). Emplacement of the Mt. Crofton Granite Trend, particularly the central regions, may then have occurred under cooler brittle Andersonian crustal conditions during NE-SW shortening (Fig. 12B) late in the Paterson Orogeny, or alternatively, by shortening (*e.g.* Castro, 1986; Davis, 1993) during D_5 at a high angle to the regional grain (Fig. 12C). The latter is consistent with an increase in D_5 ductility adjacent to the Mt. Crofton Granite Trend (Fig. 12D), although emplacement may have been episodic and encompassed both time periods.

Thus two potential episodes of granite emplacement may have occurred given that D_5 appears to post-date the Paterson Orogeny (Section B). This is compatible with the relative timing and geochronological ages of the various granitoids (Fig. 4), where recent U-Pb dating of the Mt. Crofton Granites suggests they are ≈ 20 m.y. younger than other syn-tectonic intrusions (Fig. 4). Similarly, geophysical data suggest that the whole Mt. Crofton Granite Trend comprises numerous separate plutonic bodies (R. Bateman, pers. comm. 1994), each of which could have had different crustal emplacement times.

CONCEPTUAL STRUCTURAL EXPLORATION MODELS

Epigenetic mineralisation in the Telfer deposit developed during, and was controlled by, the interaction of two late-orogenic deformation events (D_3 & D_4) acting on a pre-formed host structure (Section C; Section D). This mechanism allows one to develop a conceptual approach that can be applied to exploration of the other domes and structures in the region.

Fold-hosted Deposit Exploration Models

The Effects of D_3 Deformation

D_3 preferentially gaped the NE limb of the Telfer Dome during bulk vertical shortening combined with NE directed horizontal movement. Consequently, drilling should preferentially target the NE limbs of regional domes (Fig. 13A), particularly those that can be identified as being asymmetric with a steeper dipping NE limb. This relationship would switch to the SW limb if D_3 reversed vergence, and consequently the rocks should be examined for this feature prior to drilling. If D_3 deformation was overall coaxial in the area being explored, characterised by alternating vergence directions (*e.g.* Section B), then only limited gaping of the near-hinge regions would occur (Fig. 13B). An example of this latter feature may be the Big Tree deposit hosted in Connaughtons Dome. Here, sporadic SW-verging D_3 folds were observed which in an overall NE verging regional D_3 deformation suggest that local coaxiality occurred. This is consistent with deposit drilling confirming that no preferential gaping of either limb occurred in Connaughtons Dome; instead mineralisation developed as small pods within tight fold hinges (H. Cierlitza, pers. comm., 1995).

Another factor that may have enhanced D_3 gaping in some regional domes is the presence of refolded D_1 folds. As discussed above, D_1 folds may have controlled D_2 domal geometry, producing a steepened NE limb (Fig. 13C) that was favourably oriented for D_3 gaping. Alternatively, smaller-scale D_1 monoclinal flexures could have controlled the placement of linear ore-bodies along dome limbs (*e.g.* figs 3 & 15 - Haynes, 1986).

The Effects of D_4 Deformation

D_4 is interpreted to have reactivated D_2 folds causing renewed bedding slip and stratabound reef formation in the Telfer deposit (Section D). The clockwise obliquity of regional D_4 stress to D_2 folding implies that the NW and SE corners of domal folds would have experienced higher D_4 strain and, consequently, enhanced fold-limb reactivation (Fig. 13D; see also Section D). Therefore, drilling programs should preferentially target these corners for stratabound reef mineralisation. In particular the SE corner should be drilled first because both D_3 and D_4 effects could be well developed here. Additionally, a variety of veining styles along fold limbs may have been produced by re-fracturing of earlier silicified ore zones (Fig. 13E) as a result of renewed flexural slip deformation in the domes.

At the Fallows Field deposit, similar mineralised stratabound pods, oblique to the fold hinge (Fig. 13F; see also Fig. 6) were in the correct orientation and location for shear-extensional dilation (*e.g.* Hodgson, 1989) during D_4 sinistral transpression (Fig. 13F). However, in this deposit an additional lithological control on mineralisation occurred through the development of veining and brecciation within a thick competent unit in the core of the anticline. In other regional folds, particularly smaller-scale parasitic domes, the presence of internal competent units, or massive alteration (*e.g.* silicification), may have localised epigenetic mineralisation (*e.g.* Hayward, 1992; Vearncombe, 1993).

Competent units have been favourable hosts for vein/breccia-hosted and replacement mineralisation given their tendency to deform brittlely enhancing fracture permeability. Such units can become ideal conduits for fluid-focussing (*e.g.* Oliver & Wall, 1987; Dubè et al., 1989; Belkabir et al., 1993), and may also modify stress-strain paths (*e.g.* Treagus, 1983, 1988; Dubè et al., 1989) resulting in a variety of internal vein geometries, the orientation of which can be predicted by use of the D_4 instantaneous strain ellipse (Fig. 13G). These deposits are likely to be moderate to large, depending on the host structure size, but of low-grade if mineralisation is disseminated, and hence economic only at shallow levels.

Shear/Fault Zone Exploration Models

Zones of high D_4 strain will have an increased propensity for shear zone formation (Dubè et al., 1989; Poulsen & Robert, 1989), through the initiation or reactivation of faulting. This promotes seismic pumping (Sibson, 1988) that focuses ore-fluids into favourable trap-sites within the shear/fault system (*e.g.* Hagemann et al., 1992). Such shears appear to have formed in the NE Paterson Province around regional domal folds (Fig. 11) and along some lithological/structural contacts (such as Karakutikati Range). Penetrative ductile shear zones could host disseminated mineralisation along the main shear, within cataclastic-mylonitic rocks (Fig. 14A), or else in subsidiary structures such as dilational jogs (Fig. 14B, second-order shears and en échelon vein arrays (Fig. 14A & B).

Favourable NW-SE striking shear/fault systems in the NE Paterson Province will be those with left-stepping geometries as oversteps, bends and jogs will be releasing (Gamond, 1987; Woodcock & Schubert, 1994) and thus dilational regional sinistral transpression in D_4 (Fig. 14B & C). Left-stepping fracture/shear arrays may also produce tensional bridges (Gamond, 1987; Fig. 14C). Furthermore, the orientation of tensional features within a shear/fault zone will depend on the bulk movement vector (Fig. 14D), which is likely to be oblique for sinistral D_4 transpression. Consequently, dilational lenses should have moderate NW to NNW plunges (Fig. 14D) and will be effectively targeted by inclined drilling towards the SE.

Granitoid/Intrusive Exploration Models

Mineralisation spatially associated with granitoids in the region, such as skarn deposits, may exhibit local structural-control on both ore-placement and fluid flow. In particular, the effects of the weaker mineralising deformations (D_3 and D_4) are likely to have been enhanced in contact aureoles given the ductility increase. Examples of potential dilation geometries include the gaping of inclined strata during D_3 (Fig. 15A), and the development of opposing shear-sense couples (Forde & Bell, 1994) across primary lithological contacts during D_4 (Fig. 15A, B). The latter will occur when synthetic fold movement (synthetic shear) within a homogenous medium, such as a granite, opposes an antithetic shear developed on a reactivating (Bell, 1986) foliation. Where these two opposite shears meet, such as at a lithological contact, gaping/dilation and subsequent ore deposition can occur (Fig. 15A, B)

Other structures within contact aureoles include vein sets and stockworks whose orientations can be predicted by the instantaneous strain ellipse (*e.g.* Fig. 13G); these include WSW-ENE trending vertical sets during D_4. Additionally, local strike-slip faulting, with associated jogs (*e.g.* Fig. 14B) may preferentially nucleate in ductile aureoles (Fig. 15C). These structures should be preferentially targeted in the NE and SW corners as these would have been low-strain/pressure shadow zones during regional sinistral D_4 transpression, and thus ore-fluids may have preferentially focussed in these areas (*e.g.* Ridley, 1993; Fig. 15C).

Larger igneous complexes, such as Grace (Fig. 1, 5), contain dolerite dikes on which mineralisation may have nucleated. This may have occurred either in auriferous shears that nucleated on dike margins (Fig. 15D; Belkabir et al., 1993), or as internal vein sets with fracture-controlled replacement (Dubè et al, 1989; Oliver & Wall, 1987; Oliver et al., 1990). The trend of such dikes may be identifiable from ground-based geophysical surveys, and favourable orientations for shear development can be predicted from the instantaneous strain ellipse (Fig. 13G); these will lie within the shear-extensional veining field.

D_5 (D_{5b}) - related Mineralisation

Mineralisation also appears to be related to D_5 structures particularly those that trend N-S (*e.g.* Graben Fault - Section D), and evidence for a second later mineralising event has been observed (Section C). Additionally, some smaller prospects appear to lie along N-S trending magnetic lineaments (*e.g.* Occarina - Fig. 5) that correspond with late dolerite dike intrusion. Such mineralisation is likely to have formed only discrete bodies where earlier structures were reactivated or intersected (Table 2).

Deformation	Mineralisation Geometries/Occurrence
D_5	High angle cross-folding of domal folds. Mineralisation as pods within cross-folded D_2 dome hinges. Local tensional opening of dome limbs with subsequent mineral addition. NW-SE striking extensional vein sets. Tensional opening of strike-parallel (D_2, D_4) faulting.
D_{5b}	Mineralisation associated with N-S trending dolerite dikes, fractures (magnetic lineaments ?) and other structures. Poddy mineralisation at the intersection of the above structures with folded layers. Potential weak reactivation of regional folds (very locally)

Table 2: Summary of the potential mineralisation geometries associated with D_5 and D_{5b} compression.

Exploration Strategy: the Identification of Favourable Structural Targets

The conceptual models discussed above can be drawn together into a flow-chart that details the successive steps required to identify favourable structural targets (Fig. 16). This strategy, commencing either from reconnaissance or regional exploration, requires input through the specific identification of key structural attributes including the major D_3 vergence direction and zones of increased D_3/D_4 strain, as gaping effects within the latter will be enhanced.

Identification of D_3 Characteristics

Identifying the horizontally overprinting effects of D_3 requires prior three-dimensional knowledge of the area in question, coupled with detailed ground-based examination for minor D_3 folds to determine the vergence direction. However this is hindered by the general lack of outcrop and topographic relief NE Paterson Province. As for mesoscale examples, favourable positions for D_3 overprinting, given a NE vergence, will be at the hinge and north-eastern limb positions of regional anticlinoria (see fig. 21 - Section A). Both the Telfer and Fallows Field anticlines are located in such positions (Fig. 1). Additionally, careful analysis of cleavage-bedding relationships and minor-fold vergences in either outcrop or drill core (Fig. 17) may also assist the identification of sub-surface fold geometries.

Zones of Increased D_4 Transpressive Strain

Strain-field diagrams for bulk inhomogeneous shortening (T.Bell, 1981) predict that transpressive D_4 strain may partition, and thus locally intensify, into linear zones/corridors. The increased potential for fold re-use and shear formation, coupled with reactivation of deeper faults/lineaments to tap and channel ore-fluids makes such corridors ideal exploration targets, both regionally and at orogen-scale. Additionally, as these zones are vertical they are more readily identified during reconnaissance orogen exploration using air-borne plan-view geophysical data.

The partitioning of transpressive strain would have been controlled by pre-existing structural anisotropies (Jones & Tanner, 1995), such as those resulting from regional D_2 folds. Additionally, such strain would also have preferentially partitioned and intensified into rocks of lower viscosity (Treagus, 1983). Consequently, linear zones or corridors of relatively incompetent rock, sandwiched between thicker competent units are likely to have undergone relatively greater D_4 strain and thus are favourable areas for D_4 controlled mineralisation. These zones can be identified from the regional D_2 fold geometry of rock units of varying competency (Fig. 18). An example is the O'Callaghans Corridor (described above), which contains a high density of faults (relative to other areas in the region) and laterally continuous fold axes (*e.g.* Fig. 1 & 5). Its curvilinear form is suggestive of increased shortening and anastomosing of the corridor around the large Wilki "ring" structure (Fig. 18) during D_2 and D_4 (NE-SW to ENE-WSW) orogenic compression.

In contrast, the 17 Mile Hill - Thompsons corridor is cut into a smaller zone by granites. This may have caused transpressive strain to be diminished, allowing only limited and local transpressive effects within the corridor. Such disruption may have also occurred in the vertical dimension, due to fold closure of competent units or igneous intrusion at depth. Consequently, the scale of a corridor will be an important factor in determining its exploration potential. Although zones of transpression could form at any scale, larger corridors that completely enclose the targeted host-structure will be more favourable, as D_4 effects will be maximised across the structure. As an example, both currently mined deposits are located fully within transpressive corridors, although in each the host antiform is of different extent and scale (Fig. 18)

DISCUSSION

Structure-based exploration models are increasingly becoming standard components of modern mineral exploration. This has come about largely because surficial methods such as soil geochemistry may not work in some areas, such as those where thick cover sequences obscure geochemical signatures. Alternatively, such methods may have already been extensively used with little result. However, even with the identification of a geochemical or geophysical anomaly the task of designing an effective sub-surface exploration program, one which increases the probability of intersecting mineralisation, still remains. Finally, mineralisation might be too deep to be identified by surface methods, but still economic to mine if it can be targeted in deep drilling programs. In all these cases accurate structural modelling of ore-body geometry has considerable applicability.

Detailed and precise structural modelling enables the explorationist to more effectively understand the key structural geological elements of the host terrane. At the district/regional-scale favourable tracts or corridors may be delineated using regional-scale mapping/geophysics, allowing more effective focussing of the limited exploration resources. These zones/corridors will invariably be those that adequately partitioned the mineralising deformation phases amongst pre-existing structural anisotropies. Within these "exploration corridors", favourable targets may subsequently be identified through a combination of surficial and sub-surface surveys. Structural models will assist in predicting those structures likely to have gaped or opened to produce either a trap-site or channelway for ore-fluids during the mineralising deformations. Once a favourable host structure has been identified, and exploration reaches the drilling stage, the accurate prediction of potential ore-body geometries, and zones of gaping, allows a greater precision with targeting of drilling programs. This becomes particularly vital when deep drilling programs, targeting blind mineralisation or deep seated anomalies, are being undertaken given the increased expense of such programs.

Previous models of fold-hosted mineralisation in the NE Paterson Province considered the major structural control on ore emplacement to be bedding-slip induced dilation along dome limbs late in D_2 (Goellnicht et al., 1989; Bogacz, 1990; Laing, 1993). However, these models predict only a relatively simple mineralisation distribution within regional domes. Detailed study of two key parameters, the structural timing and structural controls on mineralisation in the Telfer deposit (Sections C & D), has demonstrated subtle, yet distinct, differences in the ore distribution, and thus a refined exploration model is required. Such models are presented in this study, which integrate the results of studies of known and well exposed ore-system (Sections C & D) with those gained from structural mapping of the district, so as to better predict zones of economic mineralisation concentration in regional folds and other structures.

The structural exploration models presented in this study will become important for future exploration in the Paterson Province, as many areas have currently been only surficially examined and drilling activities to the present have only targeted shallow anomalies. The lack of surficial indicators for a significant mineral concentration suggests either that it does not exist, or alternatively that it does exist but at a greater depth and lacks a geophysical or geochemical anomaly. Exploration for such a deposit can only be effectively targeted, in the absence of anomalies, through structural methods. Furthermore, numerous sub-economic deposits/resources have previously been identified (*e.g.* Fig. 9), and these may be upgraded economically utilising more accurately targeted drilling.

In other mineralised terranes similar structural models could be applied to the targeting of concealed/deep mineralisation concentrations (*cf.* Forde & Bell, 1994). Such models require the integration of structural geological parameters across both mine- and regional-scales, including;

1. An accurate and detailed knowledge of the deformation history of the host terrane supplemented by wider ranging reconnaissance studies of the host-orogen.
2. Detailed paragenetic and structural timing of mineralisation studies in known deposits where the specific ore-veining stages can be related to deformational episodes identified in the regional structural analysis.
3. An examination of the structural controls on mineralisation, and particularly the interaction of mineralising deformations on the regional host structure.

Structural Targeting for Gold-Copper Mineralisation in the NE Paterson Province - Some Considerations

The use of structural criteria in predicting favourable trap-sites, and ultimately ore-body geometry, relies on the important assumption that ore-fluids and deformation are intimately associated. For typical mesothermal lode-gold deposits, where fluid generation and movement are controlled by deformation/metamorphism (*e.g.* Kerrich & Allison, 1978; Robert & Brown, 1986; Forde, 1991) such an approach should yield valid results. However, a similar principle may also be applied to other epigenetic mineralisation styles, such as those directly associated with igneous intrusives (*e.g.* skarns), if these are contemporaneous with regional tectonism, and trap-site and permeability creation are structurally controlled.

In applying the structural models discussed above for further fold-hosted mineralisation the role played by regional granitoids in ore genesis is important as this will determine if other parts of the Paterson Province, which generally lack late-Proterozoic intrusives, are viable for exploration (see discussion in Section C). However, where granitoids are present Telfer-style mineralisation is generally distal (5-10kms) from actively cooling intrusives (Rowins, 1994; Section C) and its precipitation thus may have been controlled by temperature gradients away from the intrusions. In some terranes where mid- to upper- crustal granites are present, gold precipitation occurs at well defined physico-chemical conditions such as the chlorite-biotite isograd or the silica precipitation boundary (*e.g.* Meguma terrane - Mawer, 1986). Such areas could be delineated within the NE Paterson Province by petrological studies of surface outcrops/rock chips. In particular, the presence of extensive quartz veining/silicification may signify the silica precipitation boundary (≈ 300°C - Rimstidt & Barnes, 1980) and mark the development of extensive fluid channelways in the rock.

Another factor also likely to affect deposit development is the association between mineralisation and antiformal folding. Although, in theory, the above structural models apply equally to synformal folds, and base-metal mineralisation is hosted in a synform at Nifty (Dare, 1994), gold mineralisation in the NE Paterson Province is invariably sited within antiformal folds. This is a feature common to many fold-hosted deposit terranes (*e.g.* Mawer, 1986; Hodgson, 1989; Cox et al., 1991b; Hayward, 1992; Forde & Bell, 1994) and suggests that ore-fluids probably rise along highly permeable units (*e.g.* Oliver et al., 1990; Gray et al., 1991; Cartwright et al.,

1994) and are trapped beneath impermeable units in antiformal crests (Cox et al., 1991b). Alternatively, such antiforms may be relatively larger than their associated synforms and therefore experience more strain enhancing the effects of brittle deformation and hence mineralisation (Hayward, 1992).

In determining which regional antiforms have the potential to host a significant deposit, the position of such antiforms with respect to larger scale folds is also likely to be important (Fig. 19). Domal folds in the NE Paterson Province represent local culminations in which fluid may be trapped. However, these domes are generally parasitic to larger anticlinoria and synclinoria which have sub-horizontal axial traces. The Telfer Dome for example, lies at the crest of an anticlinorium (Fig. 19), whilst the Trotmans and Connaughtons domes both lie at a shallow part of the O'Callaghans synclinorium (Figs 1 & 5). If similar ore-fluid movements apply at the scale of the anticlinoria/synclinoria folds, and there is no evidence to suggest otherwise, then ore-fluids should migrate to the crests of anticlinorium and domal folds located in this position will be more favourable targets (Fig. 19).

An important indicator to potential mineralisation in the NE Paterson Province may be crustal-scale lineaments. Analysis of these in the region identified prominent structural corridors (Leeson, 1992), some of which correspond to zones of high-D_4 strain inferred in this study. Consequently, lineament analysis may be beneficial in determining linear zones that potentially partitioned transpressive deformation during mineralisation. Such zones may be characterised by major bounding linears parallel to the regional structural grain, with a relatively high internal concentration of shorter linear features sub-parallel to the boundaries. Additionally, individual lineaments, or groups of lineaments, may signify deeper crustal faulting, or major terrane structures, both of which could have been important channelways for rising ore-fluids (*e.g.* Kerrich, 1986; O'Driscoll, 1986; Eisenlohr et al., 1989; Elder & Cashman, 1992).

Other factors that may be important for internal deposit development include the rheological nature of the host-rocks. Stratabound reef formation in some fold-hosted deposits occurred within units that were mechanically favourable for flexural slip deformation (Mawer, 1986; Section D). These are typically finely laminated sedimentary rocks that exhibit a high internal lithological variation, which are sandwiched between thickly bedded units. Such units, which may mark a regional stratigraphic transition (*e.g.* Meguma terrane - Mawer, 1986), provide well developed competency contrasts that facilitate bedding-concordant vein formation (Jessell et al., 1994; Section D) and silicification on which subsequent mineralisation can develop.

The Role of Deformation in the Localisation of Precious Metal Mineralisation

The conceptual structural models presented herein illustrate important concepts regarding the economic concentration of metal mineralisation in the earth's crust. These models emphasise the relationship between deformation and the resultant creation of "traps" in which physico-chemical processes act to precipitate mineralisation. However, late orogenic deformation is commonly heterogeneous (Forde, 1991; Section A), and strongly partitioned, and therefore gaping of a pre-formed host-structure is likely to be very discrete. Whether gaping occurs will depend on the complex interaction between the mineralising deformation and the geometry of the host-structure (*e.g.* Bell et al., 1988; Forde & Bell, 1994).

The phenomenon of apparently very discrete mineralisation is well illustrated in the NE Paterson Province. Most antiformal domes invariably contain anomalous (with respect to average crustal abundances) gold mineralisation. However, no others have been identified that compare to the Telfer deposit. Why is this so? If this deposit had formed simply as the result of late-brittle processes during host-structure formation then all domes that have characteristics complying, within reason, to the considerations discussed above should exhibit spectacular reef formation. That this does not occur suggests that other factors have controlled mineralisation development. One factor could be the heterogeneity of later mineralising deformations, whose overprinting effects were limited to only very specific folds in the region resulting in an extremely localised mineralisation distribution. In the case of precious metals, the discrete development of structural traps fluid channelways by late heterogeneous orogenic deformation may provide the means to focus the considerable volumes of ore-fluid required to produce an economic concentration of these elements.

CONCLUSIONS

Structural development in the NE Paterson Province occurred during protracted compressive orogenic deformation. Although the gross structure of the region resulted from D_2 folding, pre-D_2 folds are interpreted to have, in part, controlled the nucleation and subsequent geometry of D_2 domes in the Paterson Province. This has resulted in asymmetric D_2 fold geometries in some domes, and may have produced the Karakutikati Range and other laterally continuous quartzite ridges throughout the region. Subsequent deformations slightly modified the D_2 domes, and these effects include gaping and rotation of the NE limbs and hinges during D_3. D_4 was a period of weakly oblique transpressive shortening that re-used/reactivated regional domes (F_2) in a sinistral transcurrent manner.

The importance of D_3 and D_4 lies in their association (timing) with ore-bearing fluids in some deposits (Sections C & D). Consideration of the effects of these deformations on pre-formed regional domal folds has led to conceptual models of structurally hosted epigenetic mineralisation that predict ore-body geometry and occurrence. These models emphasise the highly discrete nature of D_3 and D_4 with respect to the pre-existing D_2 architecture, and the specific gaping geometries that are expected to be encountered. Such gaping/dilation is considered, given the control on ore-fluid timing by D_3 and D_4 (Section C), to be of prime importance in creating favourable traps for regionally pervasive ore-fluids.

Many factors may ultimately control the localisation of economic grades of mineralisation in the Earth's crust. However, a particularly strong control appears to be that caused by orogenic deformation and in particular its influence on the creation of trap-sites and fluid channelways for epigenetic mineralisation. The recognition of polyphase deformation histories in mineralised terranes, coupled with the late timing of epigenetic Au and other mineralisation, suggests that the development of suitable host-structures may be more complex than previously thought. Exploration strategies need to account for such complexity, and in particular the potential for discrete host-structure development during heterogeneous late orogenic deformation.

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