Pergamon

Journal of Structural Geology, Vol. 20, No. 9/10, pp. 1405 to 1424, 1998

PII: S0191-8141(98)00073-X

0191-8141/98/$ - see front matter

Partial convective overturn of Archaean crust in the east Pilbara Craton, Western Australia: driving mechanisms and tectonic implications

W. J. COLLINS and M. J. VAN KRANENDONK*

Department of Geology, University of Newcastle, NSW 2308, Australia
(bcollins@geology.newcastle.edu.au)

and

C. TEYSSIER

Department of Geology and Geophysics, University of Minnesota, Minnesota, 55455, U.S.A.

(*Received* 25 *August* 1997; *accepted in revised form* 20 *April* 1998)

Abstract—Strain varies systematically from weakly-developed, outward-dipping, *S*-tectonites in the ~3320–3310 ± 10 Ma Mount Edgar Batholith to intensely deformed, subvertical, *L*-tectonites in greenstones of the Warrawoona syncline. A consistent 'greenstone-down/batholith-up' sense of shear is recorded in batholithic domal margins and adjacent high-grade supracrustal rims: lineations converge to a central zone of subvertical extension (zone of sinking) along the synclinal axis. At domal margins, early kinematic granitoid sheets and 'intrusive diatexites' are subconcordant to a well-developed, dome-parallel schistosity, but late- to post-kinematic intrusives are discordant, high-level plutons. All granitoids are the same age, within analytical error.

These syn-doming features conform with structural tests for diapirs, and differ from those expected during metamorphic core complex formation and cross-folding. Diapirism is part of a larger process involving partial convective overturn of the crust. Based on strain patterns and kinematic criteria, we argue that deformation was initiated by sinking of greenstones, which rapidly subsided; the domes then rose passively as accommodation structures. Ongoing doming (D_4) partitioned strain to the southern rim of the batholith and uplifted a wedge of the Warrawoona Syncline keel. We consider that crustal overturn occurred in response to mantle plume activity: Eruption of a 5–10 km thick, ~3325 Ma, mafic–ultramafic greenstone pile onto an older granite–greenstone terrain created a negatively buoyant crust, but convective overturn occurred some 5–20 Ma later, triggered by widespread anatexis (thermal softening). Convective overturn may have been a common Archaean process, occurring in response to a hotter mantle, and represents an end-member deformation mechanism that includes thrust-accretion of lithotectonic assemblages in other Archaean terranes, such as the Superior Province of Canada.

INTRODUCTION

Formation of dome-and-basin or 'dome-and-keel' structures in the Archaean granite–greenstone terrain (GGT) of the Pilbara Craton, Western Australia (Fig. 1) has been considered to result from diapiric uprise of batholithic domes, either covered by internal buoyancy forces associated with magma emplacement (Hickman, 1984; Collins, 1989; Williams and Collins, 1990), or by uplift following crustal overthickening (Bickle *et al.*, 1980, 1985). These high-amplitude, dome-and-keel structures and their high-strain marginal, steeply-dipping, concentric deformation zones, differ fundamentally from the overthrust belts that developed in some Late Archaean belts such as the Superior Province of Canada. There, sutures defining the sublinear distribution of alternate metasedimentary and metavolcanic subprovinces tend to correspond with deep-crustal seismic reflections, which are subconcordant with widespread penetrative, thrust-related foliations. These features are consistent with juxtaposition of microcontinent-scale blocks within a craton-scale, fold-and-thrust belt (e.g. Sawyer and Benn, 1993; Benn *et al.*, 1994), but they are absent in the east Pilbara. The Late Archaean strike-slip deformation zone, which anastomoses between the domes through the central Pilbara, is the only clear representative of regional-scale horizontal tectonics in the area (Van Kranendonk and Collins, 1998).

The absence of sublinear tectonic grain and juxtaposed metasedimentary/metavolcanic belts, along with the unusual subvertical, concentric deformation zones of the Pilbara, suggests that tectonic processes differed from those of the Superior Province. Other characteristic features of the east Pilbara, such as: (1) continuity of low-grade stratigraphy within and between domes over large (> 100 km) distances; (2) younging directions that typically trend away from the batholithic domes; and (3) the general weakly to very weakly deformed nature of the low-grade greenstones, where delicate Early Archaean stromatolites, non-deformed

*Now at: Geological Survey of Western Australia, 100 Plain St, E. Perth, W.A. 6004, Australia.

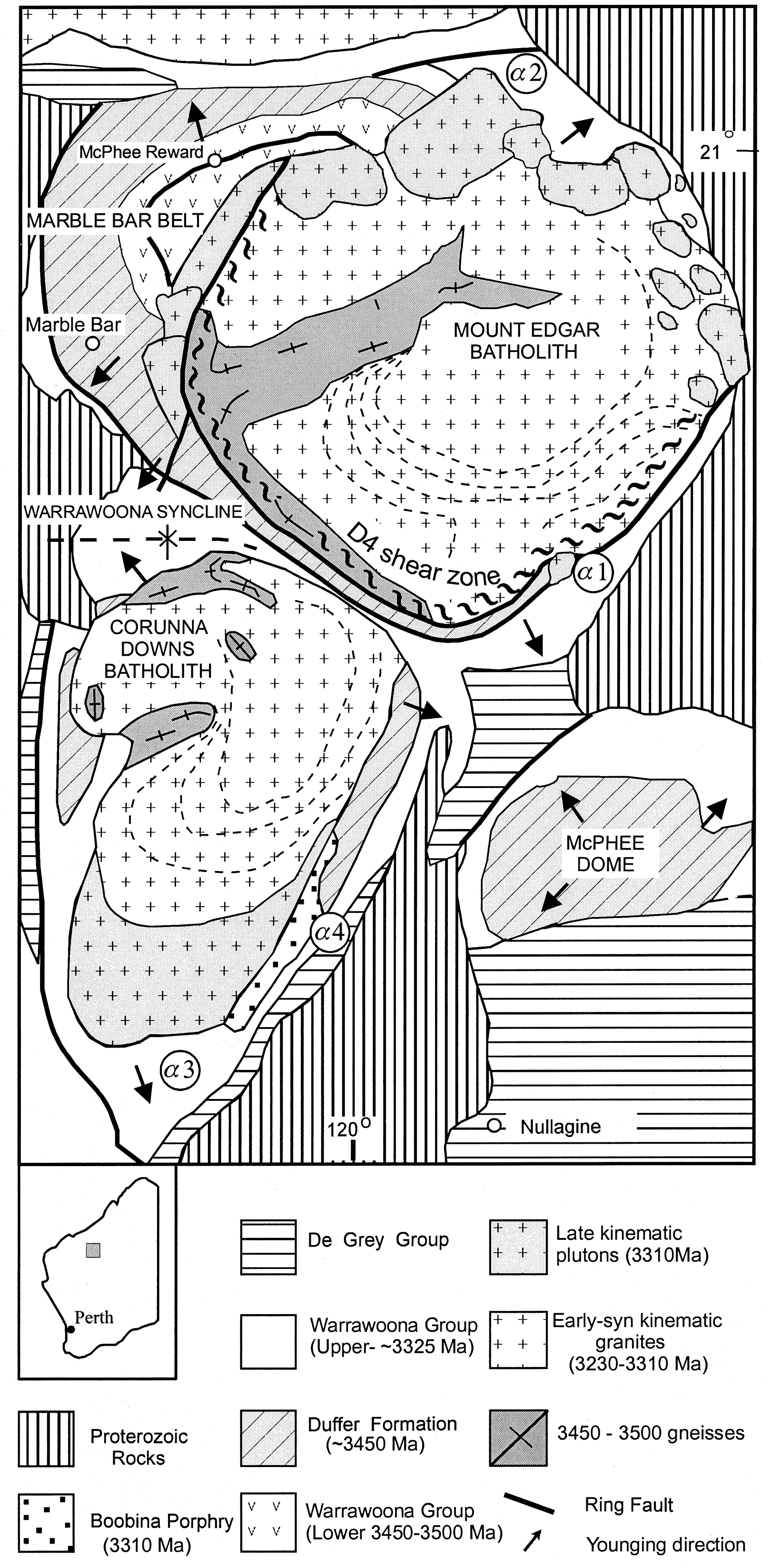

Fig. 1. Geological map of the east Pilbara, showing the Mount Edgar and Corunna Downs batholiths and Warrawoona Syncline. The post-kinematic Wilina pluton (α1) has a SHRIMP U–Pb zircon age of 3324 ± 6 Ma (Collins, unpublished data). Other sampling localities of inferred Salgash subgroup rocks (α2–4) are also shown (see text for discussion). Inferred flow lines (dashed) in batholiths reflect compositional variation (K:U:Th) in granitoids (from Mackey and Richardson, 1997).

vesicles in pillowed basalts and recrystallised glass shards in ignimbrites are all preserved, only add to the uniqueness of the craton.

In this paper, the specific cause for development of the Archaean dome-and-keel structures in the Pilbara and the driving mechanism for their evolution is discussed by using the magmatic, metamorphic and structural history of the Mount Edgar Batholith (MEB) and adjacent Warrawoona Syncline dome as an example. We consider that the dome results from convective overturn of the crust in response to density inversion of a thick (> 10 km) greenstone cover erupted over a previously stable configuration of ~3450 Ma greenstones and coeval tonalite–trondhjemite–granodiorite (TTG) intrusives.

GEOLOGICAL CONSTRAINTS

The MEB (Fig. 1) is dominated by syn- and late-kinematic ~3320–3310 Ma granites (Williams and Collins, 1990; Collins, unpublished data), with subordinate older (~3450 Ma) orthogneisses (< 20%) that rim the southern margin and form a septa extending through the centre of the batholith. The granitoids are tonalitic and trondhjemitic, with synkinematic types dominating the central and southern sectors and late-kinematic plutons dominating the northern and eastern sectors. Structurally, the batholith is a dome ~50 km in diameter, outlined by sub-concentric flow patterns, reflected by cm- to km-scale compositional differences which can be identified by remote-sensing techniques (Mackey and Richardson, 1997), and by solid-state foliations that are most strongly developed around the southern margin (Collins, 1989). The dome was tilted to the northeast during the last major deformation (D_4), which is reflected by the concentration of deep-level, early-kinematic granitoids and older gneisses in the southwest sector of the batholith (Collins, 1989). The late-kinematic granites have subvolcanic textures such as miarolitic cavities and β-quartz pseudomorphs, and are discordantly intrusive into low-grade, weakly-strained greenstones. They have intruded through the domal structure outlined by the greenstones and represent high-level plutons in the uppermost part of the batholith. In contrast, the southern contact is strongly tectonised and the granites are sheet-like, with the exception of the post-kinematic Wilina pluton along the southeast boundary (Fig. 1). Structural and magmatic relations between foliated 3320–3310 Ma granitoids, older orthogneisses and greenstones in this southwest zone are the focus of this paper.

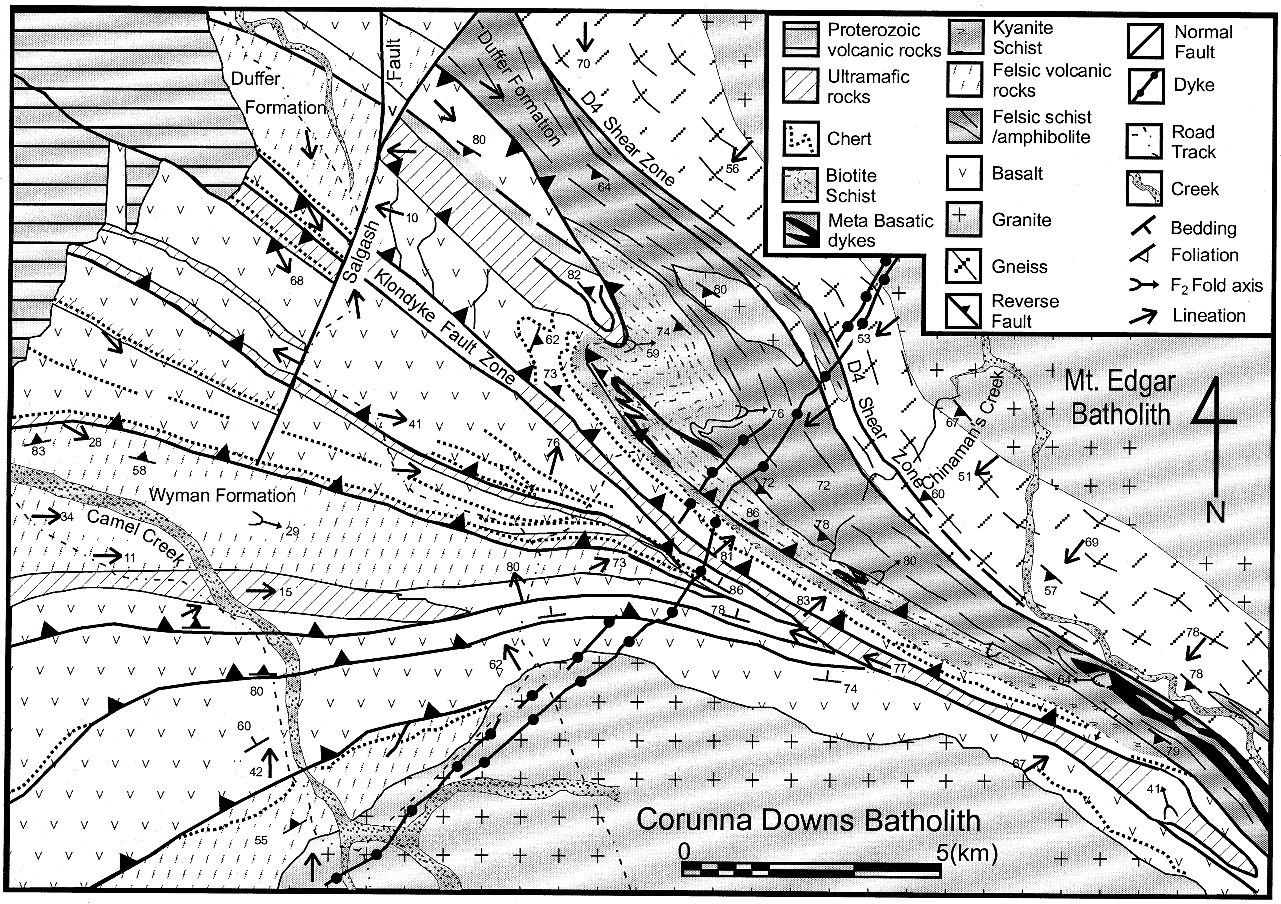

Fig. 2. Detailed map of Warrawoona Syncline, showing stratigraphically younger rocks toward the synclinal axis. Note the wedge of discordant, kyanite-bearing mylonitic schists located between the D_4 shear, Salgash and Klondyke faults.

The Corunna Downs Batholith (CDB) consists of TTG rocks that are similar to those found in the MEB. Orthogneiss remnants and amphibolites comprise < 10% of the batholith. The internal structure of the CDB is also broadly concentric (Fig. 1), with a large-scale, arcuate, outer intrusive phase (Hickman and Lipple, 1978). The contact of the CDB with low-grade greenstones of the Warrawoona Syncline is intrusive, broadly concordant but locally discordant, and only slightly deformed, indicated by a weak contact-parallel, outward-dipping foliation in the granite.

Between the MEB and CDB, the Warrawoona Syncline comprises greenstones (including komatiites) of the Salgash Subgroup, representing the upper part of the Warrawoona Group, and felsic volcanic rocks of the Wyman Formation, located in the synclinal axis (Fig. 2). Attitude of layering is steep to subvertical and up to 10 km of stratigraphy is lost eastward in the syncline, along layer-parallel normal faults. These faults and the synclinal axis are truncated by the Klondyke Fault, which separates high-grade, mylonitic schists from the more typical greenstone succession. The schists form a tectonic wedge also bounded by the Salgash Fault in the west and the D_4 shear zone of the MEB to the north (Fig. 2). The wedge comprises schistose and mylonitic greenstones metamorphosed to amphibolite facies. Felsic rocks from this zone, representing altered volcanics from the Duffer Formation of the lower Warrawoona Group, are kyanite-bearing (Collins and van Kranendonk, 1998).

STRAIN PATTERN DEVELOPMENT DURING EMPLACEMENT OF SHEETED GRANITOIDS

The relation between strain pattern within Archaean GGTs, development of the granite domes, and emplacement of the granitoids, is a crucial aspect of dome-and-keel evolution. To this end, we investigated the strain pattern across the high-strain, southern, granite–greenstone contact of the MEB, through the axis of the Warrawoona Syncline to the CDB in the south, reflecting the transition from domal wall to greenstone keel. It is characterised by the systematic change from weak, inclined, oblate (*S*-fabrics) to intense, subvertical prolate (*L*-fabrics) as the synclinal axis is approached from either batholith (Fig. 3a). The most lineated rocks approximately follow the axial trace and lineations define a central zone of the syncline. The following is an example of the changing character from *S*- to *L*-tectonite from within the MEB to the Warrawoona Syncline. It illustrates that doming was synchronous with granite emplacement, rather than occurring at a significantly later stage; an aspect which is crucial for the crustal overturn model presented later, and confirmed by recent SHRIMP dating of the Wilina pluton.

Granitoids of the MEB are typically homogeneous and variably foliated. The foliation is usually a tectonic fabric, characterised by recrystallised aggregates of quartz and biotite that anastomoses around plagioclase grains. Locally, biotite schlieren are well developed adjacent to xenoliths and rafts of amphibolite or orthogneiss, giving the granites a banded appearance. Given its structural continuity with xenoliths, extending as diffuse wisps from the gneissic rocks, the schlieren appear to result from flow sorting of restitic phases during intrusion. The schlieren and tectonic foliation are usually non-aligned within the interior of the batholith (Fig. 4a, locality 1), but are subconcordant nearer the margin (Fig. 4b & c).

Within 2.5 km of the southern margin of the MEB (Fig. 4b, locality 2), sheeted granites become more common, and are typically aligned concordantly with the batholith margin. They are characterised by mm- and cm-scale biotite/hornblende and/or plagioclase schlieren, which are traceable for up to tens of metres within otherwise homogeneous granites. Plagioclase is rarely aligned. The schlieren are considered to represent flow sorting, as they generally define the boundaries of individual granitic sheets. They are subconcordant to a weakly developed tectonic foliation, which is better observed at the metre-scale, rather than centimetre-scale (Fig. 4c, detail of Fig. 4b). The tectonic foliation is in turn concordant with the domal boundary, becoming more intense in that direction (see below), suggesting that sheeted intrusion and magmatic flow developed during doming. A lineation was not observed.

Relatively homogeneous, sheeted granites from locality 2 grade within a kilometre southward into a zone best described as ‘intrusive diatexite’. At this locality (3, Fig. 4d), granitic material is heterogeneous, consisting of decimetre- to metre-scale leucocratic and mesocratic sheets that contain persistent mm-scale schlieren and lenses of mafic minerals, some of which are isoclinally folded. That the sheets are intrusive is indicated by the low-angle discordance between them, with the latter sheets being equally heterogeneous. Subcordant, late pegmatite sheets with associated smaller discordant veins that grade into leucosome represent the last intrusive phase. Again, lineations were not observed.

Southward, closer to the batholith margin, the heterogeneous schlieric granites pass gradationally into a migmatitic phase, where coarse igneous-textured leucosome, rimmed by biotite melanosome, forms layers within gneissic palaeosome (Fig. 4e, locality 4). A weak lineation is evident. The migmatite is truncated by an even-grained, but more heterogeneous, ‘intrusive’ diatexite. The vertical view shown in Fig. 4(e), parallel with the lineation, with the batholith margin to the left of the photograph, shows a component of viscous drag of the migmatitic phase indicating ‘batholith-up’ sense of movement.

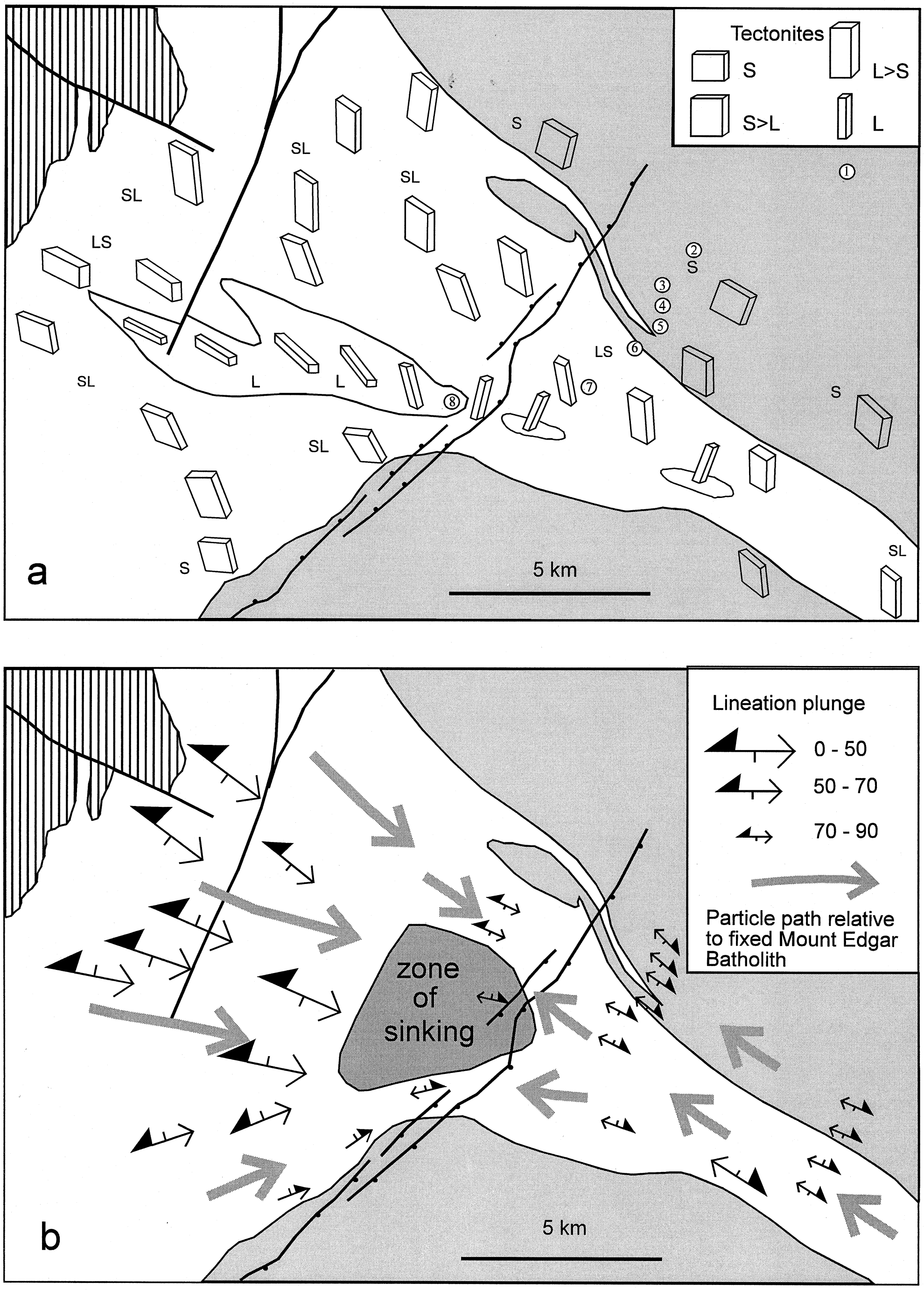

Fig. 3. (a) Strain pattern variation across the Mt Edgar batholith and Warrawoona Syncline (from Teyssier and Collins, 1990). Location of photographs in Fig. 4 also shown. (b) Lineation pattern showing convergence toward a central point along the synclinal axis. Combined with kinematic data, this indicates that the greenstones sunk through a zone of vertical constrictional strain in the axis of the Warrawoona Syncline (modified from Teyssier and Collins, 1990).

This is the last deformation affecting the area and the 'shear zones' are concordant with the domal boundary, located several kilometres away. This suggests that, during intrusion of the 'diatexite' sheets, the core of the batholith was moving upward relative to its domal margin.

A strongly deformed dioritic phase, dominated by plagioclase and hornblende, exists as a $\leq$400 m-wide intrusive sheet at the outermost margin of the MEB (Fig. 4f, locality 5). The sheet contains remnants of the greenstones, now amphibolite lenses, and veins of a fine-grained trondhjemitic phase which is tightly to isoclinally folded and commonly located along amphibolite contacts. Relative to locality 4, the foliation is much stronger and the down-dip lineation is much better developed. Thus, as the batholith margin is approached from localities 1 through 5, strain intensity increases, with granitoids progressively changing from weakly deformed *S*-tectonites to strongly developed *S–L*-tectonites.

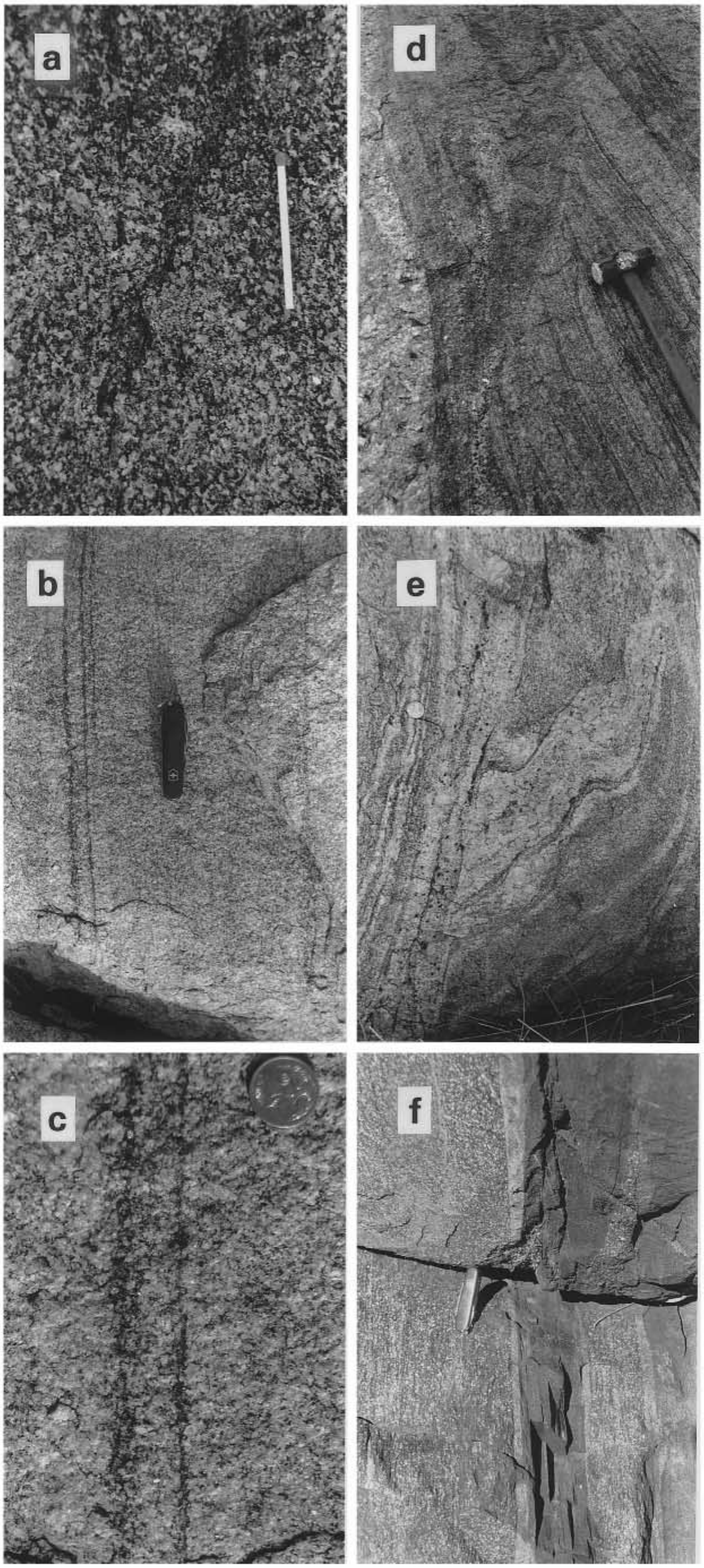

Fig. 4(a–f)

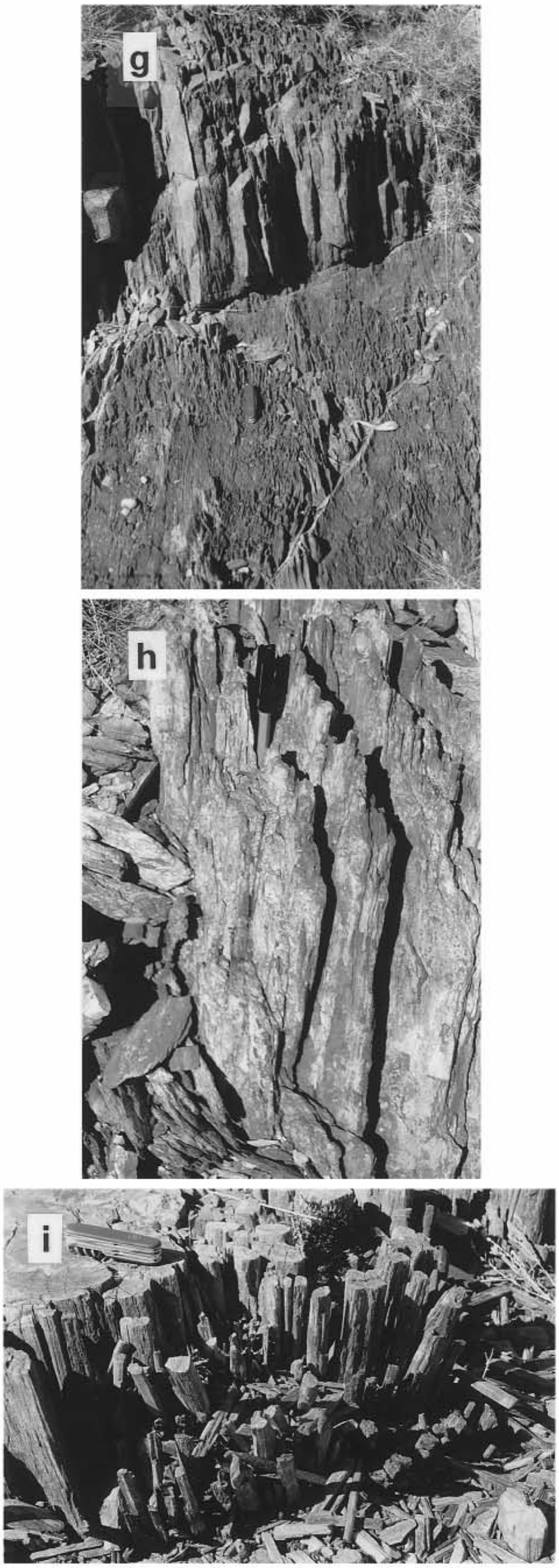

Fig. 4. Series of field photographs recording the progressive change in strain across the southern margin of the Mount Edgar Batholith into the axis of the Warrawoona Syncline. Variation is from weak oblate strain in the batholith interior (a–c), systematically to intense prolate strain in the axis. Alignment of magmatic flow and solid-state foliations (c) with sheeted granites and 'intrusive diatexite' (d & e), all subconcordant with the batholith margin, suggests doming occurred during granite emplacement and anatexis. (f–i) illustrate the progressive change from *S–L* tectonite at the batholith contact (f), though *L–S* fabrics in greenstones (g & h) to *L*-tectonite (i) in the Warrawoona Syncline axis. See text for details. Locations (1–8) are shown on Fig. 3.

Beyond the batholith margin for approximately 1 km, fine-grained amphibolites predominate (Fig. 4g, locality 6). They are strongly flattened, but a well developed, down-dip, mineral elongation lineation defined by cm-scale amphibole aggregates, is evident. The rock is an intensely developed $S > L$-tectonite and this pattern is dominant almost to the axis of the Warrawoona Syncline. The down-dip lineation generally intensifies from a well-defined mineral elongation to a rodding lineation, and the foliation tends to wrap around the lineation as incipient mullion structures defined by concentrations of quartz aggregates in schists (Fig. 4h, locality 7).

In the axis of the Warrawoona Syncline, across the Klondyke Fault, the rocks are extremely rodded L-tectonite metabasalts and cherts (Fig. 4i, locality, 8), although the metamorphic grade is much lower. The vertical rodding is defined by subcircular quartz ribbons, with axial ratios in excess of 1:1:50. At this locality, the chert is isoclinally folded, but the axial surface is strongly curviplanar, more akin to disharmonic folds developed in constrictional mylonite zones. This location corresponds with the 'zone of sinking' (Fig. 3b) identified along the axis of the syncline (Teyssier and Collins, 1990).

On the southern side of the synclinal axis, the intensity of deformation rapidly decreases, and the greenstones are weakly strained S–L-tectonites. They are intruded semiconcordantly by granites, which have a weak contact-parallel foliation that dips outward, away from the batholith core. This contact and the concentric outcrop pattern of the greenstone layers outline the domal shape of the CDB (Fig. 2).

AGE CONSTRAINTS

The field evidence described earlier suggests that sinking of the greenstones and rising of the MEB occurred while some granite sheets were being emplaced and the sialic crust was undergoing partial anatexis. The general alignment of sheets, subconcordant with the domal margin of the batholith, is consistent with doming (e.g. Fig. 4d), and the concordance of sheets with even a weak domal fabric (Fig. 4b), suggests a strong structural control during emplacement. Therefore, age determination of the granitic sheets in the MEB should constrain the timing of dome-and-keel formation in the east Pilbara.

Early- to syn-kinematic granites of the MEB have the same U–Pb zircon SHRIMP age (3304 ± 10 Ma) as the late-kinematic granites (3314 ± 10 Ma), within measured error limits (Williams and Collins, 1990). A recent U–Pb zircon SHRIMP age determination from the post-tectonic Wilina pluton ($\alpha1$, Fig. 1) is 3324 ± 6 Ma (Collins, unpublished data). The pluton is non-foliated and cuts the D_4 shear zone, which is contact metamorphosed. These structural and geochronological data confirm that dome formation and marginal shearing occurred during and shortly after intrusion of the vast majority of granitoids.

A 3307 ± 19 Ma age for the Boobina porphyry, a subvolcanic granitoid sill on the upper eastern margin of the CDB (Pidgeon, 1984) and 3310 ± 10 Ma ages from the CDB itself (Barley, personal communication), suggest that the CDB was coeval with the MEB. Coeval formation of the CDB and MEB is consistent with the symmetrical arrangement of greenstones in the intervening Warrawoona Syncline.

Although the age of the Lower Warrawoona Group is well established at 3470–3440 Ma (Thorpe *et al.*, 1992; McNaughton *et al.*, 1993), the age of the Upper Warrawoona Group, comprising greenstones of the Warrawoona Syncline, is conjectural. Thorpe *et al.* (1992) suggested that they are ~3450 Ma old, based on U–Pb age determinations of inferred Upper Warrawoona correlatives from NE of the MEB ($\alpha2$, Fig. 1), in the Coppin Gap greenstone belt. However, Nijman *et al.* (1998) indicated that the stratigraphic position of these units cannot be accurately established, because of tectonic repetition, and suggested that the analysed unit belongs to the lower Warrawoona Group.

A better approximation of the age of volcanic rocks in the upper Warrawoona Group, we suspect, is provided by greenstones of the Kelly Belt, which occur on the southern side of the CDB ($\alpha3$, Fig. 1). These rocks can be traced continuously around the CDB from the type section of the Salgash subgroup and they overlie the typical older association of the Duffer Formation and distinctive Marble Bar chert. Ages of dacitic volcanics in this belt, interlayered with basalts and subordinate komatiites, are 3325 ± 4 Ma and 3324 ± 4 Ma (McNaughton *et al.*, 1993). The data suggest that the upper part of the Warrawoona Group is ~150 Ma younger than the lower part.

The age of the Wyman Formation in the Warrawoona Syncline (overlying the upper Warrawoona Group) has been determined recently at *ca* 3325 Ma (Barley, personal communication), similar to correlatives on the southeastern side of the CDB ($\alpha4$, Fig. 1), which yield an age of 3325 ± 4 Ma (McNaughton *et al.*, 1993). As the Wyman Formation is part of the Warrawoona Syncline, structural development of the syncline must have occurred later than 3325 Ma. The stratigraphically younger Gorge Creek group is also strongly infolded along the southeastern extension of the Warrawoona Syncline, indicating that it also pre-dates doming. As the domal foliations of the MEB are concordant with those of the syncline, extend around the entire batholith in a concentric arrangement, and are tightly bracketed as ~3320–3310 Ma old, development of the Warrawoona Syncline is also restricted to ~3320–3310 Ma, and Gorge Creek group deposition is constrained between 3325–3310 Ma. Therefore, dome-and-keel development occurred at ~3320–3310 Ma, possibly up to 20 Ma after deposition of mafic and ultramafic (komatiitic) greenstones of the upper Warrawoona and Gorge Creek groups.

THERMO-MECHANICAL MODELS FOR DEFORMATION AND METAMORPHISM

Buckling mechanisms

The dome-and-keel structures of the Pilbara superficially resemble a series of concentric 'basement-cover' folds, which are characterised by tight, pinched synclines and broader, rounded anticlines that develop at the unconformity between sediments and underlying high-viscosity (crystalline) basement culminations (see Ramsay, 1967, p. 384). This style of deformation is not applicable for the Pilbara because the 'basement' was migmatitic during deformation (Fig. 4 d & e), and thus of much lower viscosity than a crystalline basement. Also, deformation was synchronous with voluminous intrusion at ~3320–3310 Ma within the MEB, indicated by the overlapping U–Pb zircon ages of early-, late-, and post-kinematic granites. Therefore, much of the granitoid substrate was younger and of lower viscosity than the 'cover' sequence during folding, so the dome-and-keel structures of the Pilbara cannot be disharmonic 'basement-cover' folds.

Cross-folding has been preferred to gravity mechanisms to explain the dome-and-keel structure of some GGTs, (Dimroth *et al.*, 1983; Myers and Watkins 1985; Hudleston *et al.*, 1988). Schwerdtner (1990) outlined several structural tests to distinguish between diapiric and cross-folding models and applied them to the granitoid domes of the Superior Province, Ontario. Here, we compare the Ontario domal structures with those of the Pilbara.

An obvious difference is the contrasting size of the typical domes of the Wabigoon subprovince in Ontario (~10 km diameter) compared with those of the Pilbara (~50–60 km diameter). More importantly, the early major fault systems in the Pilbara are extensional and lack associated upright folds (Collins, 1989; Zegers *et al.*, 1996), whereas those in Wabigoon are contractional and associated with abundant, isoclinal, upright folds. A well developed (axial planar) gneissosity and fold axis-parallel (intersection?) lineation exists throughout the Wabigoon structures, the orthogneisses having been subjected to 'severe ductile deformation' during horizontal shortening *before* doming (Schwerdtner, 1990). However, *S*- and *L*-fabrics are concentrated only in the rims of the Pilbara domes and are weak or not observable elsewhere. The lack of early upright, isoclinal folds in the Pilbara domes, particularly in the MEB and CDB, indicates that the cross-folding mechanism is inapplicable because the necessary earlier fold generation did not exist.

Lineation orientation is the other crucial difference between the Wabigoon and Pilbara domes. The Wabigoon lineations are ubiquitous, parallel to the hinge-line of small-scale buckle folds, typically shallow, and are traceable between domes. In contrast, lineations are not observed in batholith interiors of the east Pilbara, and are well-developed only at and beyond the dome margins. They do not represent intersection lineations as they are confined within a simple foliation which maintains a steep attitude concordant with the batholith margin, and they are defined by elongate mineral aggregates which form rods and mullions (Fig. 4 h & i). In addition, they are progressively better developed toward the greenstone synclinal axis (Fig. 4 g–i). A cross-folding model cannot easily explain the virtual absence of fabrics in the domal centre and the progressive development of strain toward the adjacent basin.

In the Wabigoon domes, the lineations are not subradial and trend lines pass through triple junctions, indicative of cross-folding. In contrast, individual Pilbara domes have independent lineation patterns: in the D_4 shear zone of the MEB it is a distinctive subradial pattern trending dominantly to the south-southwest (Collins, 1989, fig. 10), but in the Warrawoona Syncline the strongly developed lineations change systematically eastward along the axial trace from shallow easterly, ultimately to steep westerly (Fig. 5). This change is mimicked by a systematic variation in lineation orientation throughout the Warrawoona Syncline, as shown by the Camel Creek, Chinamans Creek and Horrigan traverses (Fig. 5). All lineations converge toward a central point within the syncline, but there is no change in foliation attitudes. Again, this is difficult to explain by a cross-folding mechanism.

At craton-scale, the general lineation trajectory for individual domes in the east Pilbara is toward the triple junction between the North Pole Dome, MEB, CDB and Shaw Batholith (Van Kranendonk and Collins, unpublished data). Therefore, the structural tests of diapir hypotheses outlined by Schwerdtner (1990) suggest that the Wabigoon domes are not diapiric structures, but that the Pilbara domes could be.

Metamorphic core complexes?

Zegers *et al.* (1996) suggested that the Shaw Batholith, of similar dimension to and located directly west of the CBD, initially formed as a metamorphic core complex. This model needs to be tested for MEB and CDB evolution, so the features of metamorphic core complexes (MCCs) are here compared and contrasted with the Pilbara domes.

Characteristic features of Cordilleran-style MCCs include upright, doubly-plunging foliation arches, with a high-grade gneiss/granite core up to 10 km in diameter, overlain by a gently-dipping, mylonitic décolle-

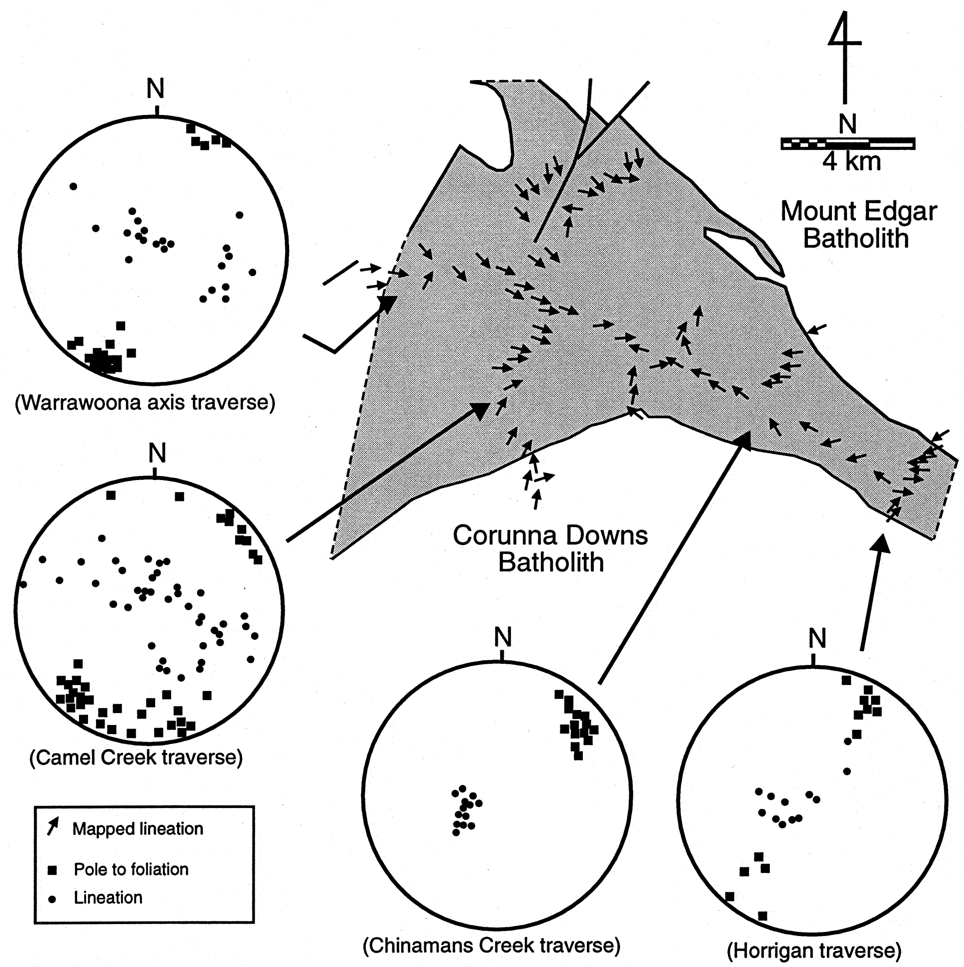

Fig. 5. Lineation orientation in the Warrawoona Syncline (from Teyssier and Collins, 1990). Note the convergence toward a central point along the synclinal axis, and the generally shallow plunges in the western, lower-strain section (Camel Creek) compared with the consistently steeper attitudes (subvertical extension) along the Horrigan and Chinamans Creek traverses, where significant losses of stratigraphy occur (Fig. 2) and strain is much greater.

ment horizon and lower-grade cover of supracrustal rocks (Coney, 1980). Strain is concentrated along the décollement horizon, which typically contains a unidirectional mineral lineation coaxial with the axes of arching, and which has a consistent sense of shear across the entire core complex: on one flank the shear-sense is normal, but on the other it is reverse, which is the opposite to that expected in diapirs, where buoyant uplift of the core occurs (Scott and Lister, 1995, fig. 1).

The Pilbara domes superficially resemble MCCs. They share common features such as: (1) synkinematic granitoids, pegmatite and migmatite generally restricted to domal cores; (2) marginal sheet-like granitoids; (3) a discontinuous high-strain migmatitic carapace; (4) décollement horizon(s); (5) normal arrangement of stratigraphy in the cover, with 'younger-on-older' bedding-parallel extensional faults (McPhee Reward: Collins, 1989; van Haaften and White, 1998); and (6) a normal metamorphic zonation.

Nonetheless, major differences exist between the Pilbara domes and MCCs. The most obvious difference is the size and amplitude of the Pilbara domes, up to an order of magnitude greater in diameter and with 'décollement horizons' generally tilted to steep or overturned attitudes. For example, the outer ring fault of the MEB, extending from west of Marble Bar around the southern margin of the MEB, is typically steep ($>70°$) and locally overturned (Collins, 1989, fig. 5). In addition, several 'décollement horizons' exist within the cover sequence, none of which are confined to the basement–cover interface (Fig. 1) and many have been intruded by late- and post-kinematic plutons. Thus, in contrast to typical MCCs, the east Pilbara décollement horizons have been involved in progressive reorientation and repeated late-intrusion.

Other important differences are the lineation trajectories, shear sense at the basement–cover interface, and the strain pattern. In contrast to the unidirectional pattern of MCCs, the lineation pattern around the MEB is subradial (Collins, 1989). Also, the sense of shear around the MEB is consistently normal (batholith-up, greenstone-down), whereas MCCs do not show this antithetic relation: one margin must show reverse movement. In addition, strain is typically localised to the décollement horizon at the basement–cover interface in MCCs, but the strain is most intense in the synclinal axes of the greenstones in the Pilbara, where it is generally constrictional and subvertical. To our knowledge, these features do not resemble any described MCC.

The ductility and metamorphic contrast across the 'basement–cover' interfaces in the east Pilbara are not extreme. Whereas the décollement horizon of MCCs separates rocks of highly contrasting structural and metamorphic character, with deformation in the cover rocks being typically brittle, ductile behaviour characterises many of the faults within the Pilbara greenstone succession: some developed at amphibolite facies, particularly in the Warrawoona Syncline. These ductile faults are typically layer-parallel structures that were responsible for removal of large sections of greenstone stratigraphy (Fig. 2), rather than forming high-angle detachment faults that define half-graben structures above MCCs. Around the MEB, the 'basement–cover' interface is only one detachment surface among several that telescope the overlying stratigraphy and metamorphic zonation pattern. Finally, the abundant syntectonic granitoids of the Pilbara are not confined to near the décollement, but extend into the deep interior of the domes and comprise over 90% of the MEB (Collins and Gray, 1990; Williams and Collins, 1990). Thus, the thermal anomaly that characterises only the 'mylonite front', the actual 'décollement surface' of MCCs (Reynolds and Lister, 1990), commonly extends to the core of the Pilbara domes. These domes bear some resemblance to core complexes, but they now represent much larger, higher amplitude, more evolved, magma-dominated structures where ductile deformation has been ultimately concentrated in the cover sequence along normal faults (Figs 3 & 4), rather than merely along the 'décollement horizon'.

The core complex model for the Shaw Batholith (Zegers *et al.*, 1996) requires that the dome developed during eruption of the Duffer Formation and intrusion of granitoids at ~3450 Ma. However, several features militate against this model: (1) despite our detailed mapping, we cannot trace the inferred décollement surface, the Split Rock Shear Zone (SRSZ), around the northern part of the dome (Zegers *et al.*, 1996, fig. 4); (2) the SRSZ should be characterised by sheeted intrusions and magmatic foliations, at least locally, but all the deformation appears to be solid-state, within homogeneous plutons; (3) available Ar–Ar hornblende ages from the SRSZ yield ages of 3220–3020 Ma (Davids *et al.*, 1997), not the predicted ages of 3450 Ma; (4) a marked unconformity should exist above the Duffer formation, outlined by detritus infilling the inferred overlying, half-graben structures, but we cannot find this feature; (5) *S–L* fabrics interpreted to be extensional and syndepositional with the Duffer Formation are identical to those in overlying formations, indicating that doming post-dated the Duffer eruption. Based on these contradictions, we consider that this core complex model is untenable in the Shaw Batholith.

Convective crustal overturn

The geometric and kinematic evolution of the MEB has been described by a diapiric model. Collins (1989) postulated that D_1 and D_2 structures in the MEB and adjacent greenstones formed during an early, major diapiric event that produced the 50 km diameter, domal structure of the MEB, followed by more restricted domal uplift (D_4) that tilted the dome, uplifting the SW margin. The domal geometry of the MEB is asymmetric, hinged in the northeast where low-grade rocks are preserved, and uplifted in the southwest where high-grade, high-strain supracrustal rocks marginal to the dome dip steeply away from, or underneath, the dome. Internally, the dome is inferred to be a doubly-lobed structure with a central spine striking to the northeast from the southwestern margin. It is considered to represent a shallow-plunging roof pendant synform (D_3 zone of Collins, 1989), as the D_3 folds do not extend to the SW into the D_4 shear zone. However, this diapiric model does not take into account the active role of greenstones in the deformation process.

Teyssier and Collins (1990) showed that the Warrawoona syncline, adjacent to the MEB, was the complementary keel structure to the batholith, based on strain patterns, lineation orientation and kinematic criteria. The strain pattern is highlighted by a progressive increase from a weakly inclined flattening to intense vertical constrictional strain as the Warrawoona Syncline is approached (Figs 3a & 4), and kinematic criteria from the *S–L* tectonites consistently indicated a 'MEB-up' sense-of-shear (Teyssier and Collins, 1990). In addition, the lineations converge toward a central point along the synclinal axis (Fig. 5), which corresponds to one of the two sites of preferential subsidence generated by sinking of a dense surface layer through a less dense substrate (Dixon and Summers, 1983, fig. 12). If the MEB is taken as a fixed reference point, then the inferred particle trajectories (approximated by lineation attitude) converge into the zone of intense vertical constrictional strain. Given that the sense-of-shear in the Warrawoona Syncline is consistently batholith-up, greenstone-down, Teyssier and Collins (1990) described this area as a 'zone of sinking' (Fig. 3b).

The 'zone of sinking' in the Warrawoona Syncline is also analogous to those modelled numerically by Mareschal and West (1980). Overall, strain is greater in the eastern half of the syncline, where significant sections of the greenstones have been removed along layer-parallel normal faults, or 'tectonic slides' (Fig. 2). This strain increase and progressive loss of stratigraphy corresponds to a progressive eastward steepening of lineation orientation (compare Camel Creek with

Chinamans and Horrigan traverses; Fig. 5). The normal faults and associated 'zone of sinking', characterised by intense subvertical constrictional strain, reflect vertical stretching and downward material flow of greenstones into the subsiding trough. Dixon and Summers (1983) predicted that the highest magnitude of vertical extension should straddle the axial plane of the trough, as developed in the axis of the Warrawoona Syncline.

A km-scale, subvertical, F_2 fold east of the Salgash Fault, outlined by the Towers Formation (Fig. 2), is considered to represent a 'cascading fold' that developed on the flanks of the MEB dome, similar to those developed at McPhee Reward (Collins, 1989, fig. 2). However, the fold is now strongly flattened, with its axis rotated into parallelism with the extension lineation (Fig. 2). This fold appression and rotation appears to be associated with horizontal shortening and down-drag that developed perpendicular to the synclinal axis as the greenstones were progressively drawn towards and into the subsiding trough (zone of sinking). In summary, the strain patterns, fold geometry, and inferred particle trajectories in the MEB and Warrawoona Syncline (summarised in Fig. 3) are those simulated by Dixon and Summers (1983) and suggest that deformation was triggered by vertically sinking greenstones, followed by lateral flow off the batholiths toward the developing syncline, eventually into the rapidly subsiding trough, or greenstone 'keel'.

Structures in the Warrawoona Syncline are considered to be high-strain equivalents of those developed in adjacent greenstone belts. For example, in the Marble Bar belt, localised, intermediate-strain rocks in the McPhee Reward area, northwest of the MEB dome (Fig. 1), contain open to tight, shallow-plunging, non-cylindrical folds. These are considered to be cascading folds that resulted from the greenstone overburden being shed off the rising dome (fig. 2 of Collins, 1989). van Haaften and White (1998) also demonstrated that the early *major* movement at McPhee Reward was extensional, with west-northwest normal movement, consistent with shedding of greenstones off the MEB, although they believed the faults formed earlier, at ~3450 Ma, during eruption of directly overlying felsic volcanics of the Duffer Formation. However, this interpretation is based simply on variations in thicknesses of the volcanic rocks, which is just as likely to reflect original variation on the flanks of a large felsic volcanic centre (Barley, 1981). Also, the normal sense of movement is defined by a 'rotated external foliation' at McPhee Reward (van Haaften and White, 1998, fig. 5), the existence of which would be surprising if the felsic lavas had erupted only 2 km above this detachment. Rather, this 'external foliation' is more indicative of ductile deformation in a deeper crustal section, and the *radial* lineations surrounding the Talga Anticline (van Haaften and White, 1998, Fig. 2) are more consistent with movement of a thick greenstone pile off the MEB during doming.

The structural evolution of the Warrawoona Syncline and adjacent batholith domes, therefore, is considered to be an intrinsic part of GGT evolution in the east Pilbara. Whereas Dixon and Summers (1983) only modelled deformation in the subsiding greenstone synclines, the overall dome-and-keel structures, their scale and geometry, and the strain pattern are more fully represented by the thermal–mechanical models of Mareschal and West (1980). The models require a gravitational instability generated by placing at least 10 km thickness of greenstone on a felsic (granitic) substrate, a condition that is satisfied in the Pilbara, where the average estimate of Warrawoona Group greenstones overlying old (>3400 Ma) sialic crust at ~3320–3310 Ma is >15 km (Hickman, 1990).

The thickness estimate of the Warrawoona Group comprises sections mostly from the Marble Bar Belt (Fig. 1), an area of good exposure and generally low strain, where original eruptive features, such as vesicles in pillow basalts and shards in ignimbrites, are typically not deformed. The few significant layer-parallel faults in the type section of the lower Warrawoona Group are extensional structures (see above) and the stratigraphy is not inverted, contrary to the speculation of van Haaften and White (1998). Given the general low-strain nature of this belt, the 15 km thickness estimate is unlikely to be exaggerated.

According to the models of Mareschal and West (1980), the greenstones sink up to 25 km, forming narrow (12–15 km wide) synclines that show intense vertical constrictional strain, whereas the granitoid domes are much broader in diameter (50–60 km) and rise up to 3 km within the crest (Fig. 6). Material flow into the syncline is compensated by lateral and vertical migration of the less dense sialic substrate into the domal crest, generating a pattern of convective overturn in the crust (Fig. 6a–d). The modelled particle trajectory for the syncline is vertically downward and for the domal axis it is vertically upward, with the crust 'circulating' about a central point located halfway between the two axes, which migrates systematically downwards toward the core of the syncline as the structure evolves. Greenstone downflow is almost compensated by batholith uprise, but shallow (~1 km-deep) interdome depressions might develop, which should fill with clastic sediments. Sandstones and conglomerates within isolated basins above greenstone synclines, representing outliers of the De Grey Group in the east Pilbara, are likely to represent these interdome sediments (Fig. 1).

Application of the convective overturn model to the MEB indicates that the supracrustal rocks are rapidly buried as the greenstone belt sinks into the underlying basement (Fig. 7). Initially, layer-parallel detachment faults are developed in the greenstone belts and foliations are restricted to these zones (Fig. 7a).

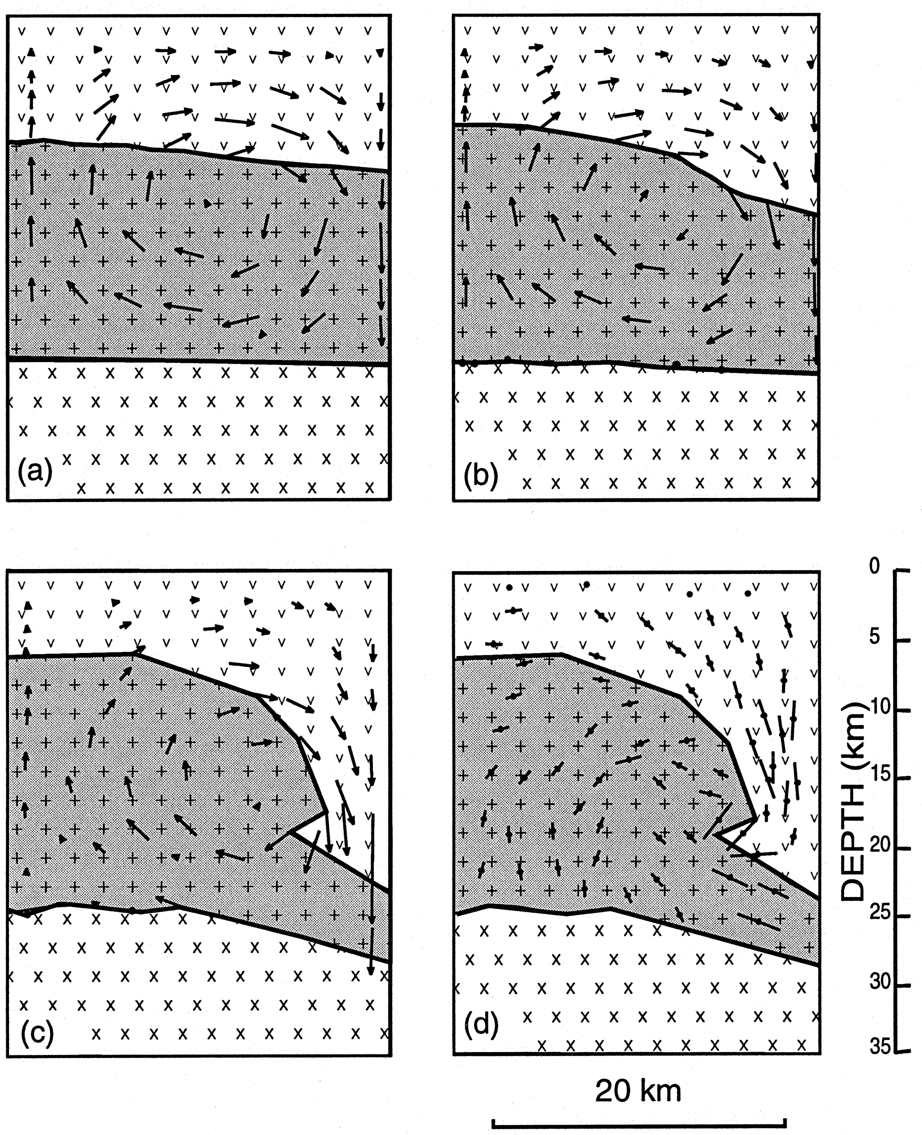

Fig. 6. Crustal overturn model, from Mareschal and West (1980), showing formation of greenstone syncline characterised by intense vertical constrictional strain, and passive convective flow of underlying sialic crust. Arrows of (a), (b) & (c) show particle trajectory and bars of (d) show instantaneous finite strain. The three major subdivisions are upper crustal greenstones, middle crustal sialic 'basement' and mafic lower crust.

Asymmetric, reclined to recumbent drag or 'cascading' folds are developed between or directly above the faults (McPhee Reward), but generally lack a penetrative axial planar foliation. As the greenstone flow into the synclines, strain intensity increases, foliation becomes more pervasive and lineations develop. Early folds are flattened and steepened, with axes rotated toward the principal flow direction, approximated by the stretching lineation (Fig. 7b). Progressive deformation in the synclines may produce coaxial interference folds, as seen in the Warrawoona Syncline, where F_1 and F_2 folds are subparallel to the well-developed stretching lineation. Within the core of the syncline, strain becomes intensely constrictional and lineations align subvertically, as material flow is toward and into specific zones of vertical sinking (Fig. 3b). Kyanite is the predicted early-, syn- and late-kinematic aluminosilicate polymorph in the syncline (Collins and Van Kranendonk, in press).

Specific structural features also develop within the dome. Early kinematic granites are emplaced as sheet-like bodies near the basement–cover interface (Fig. 7a), a zone of viscosity contrast. These granites are produced by partial melting of the older sialic crust, and the widespread 'thermal softening' provides the trigger for greenstone sinking (see below). As the greenstones sink, the early granites are deformed and rotated into semiconcordance with the steepening domal boundary. Other granites are intruded syn-kinematically, so that low-angle, sheeted intrusions are commonly developed (Fig. 4d). Some sheeted granitoids, which did not move far from the source, remain heterogeneous, and are best described as 'intrusive diatexites' (Fig. 4e). Later intrusives tend to be larger, more discordant,

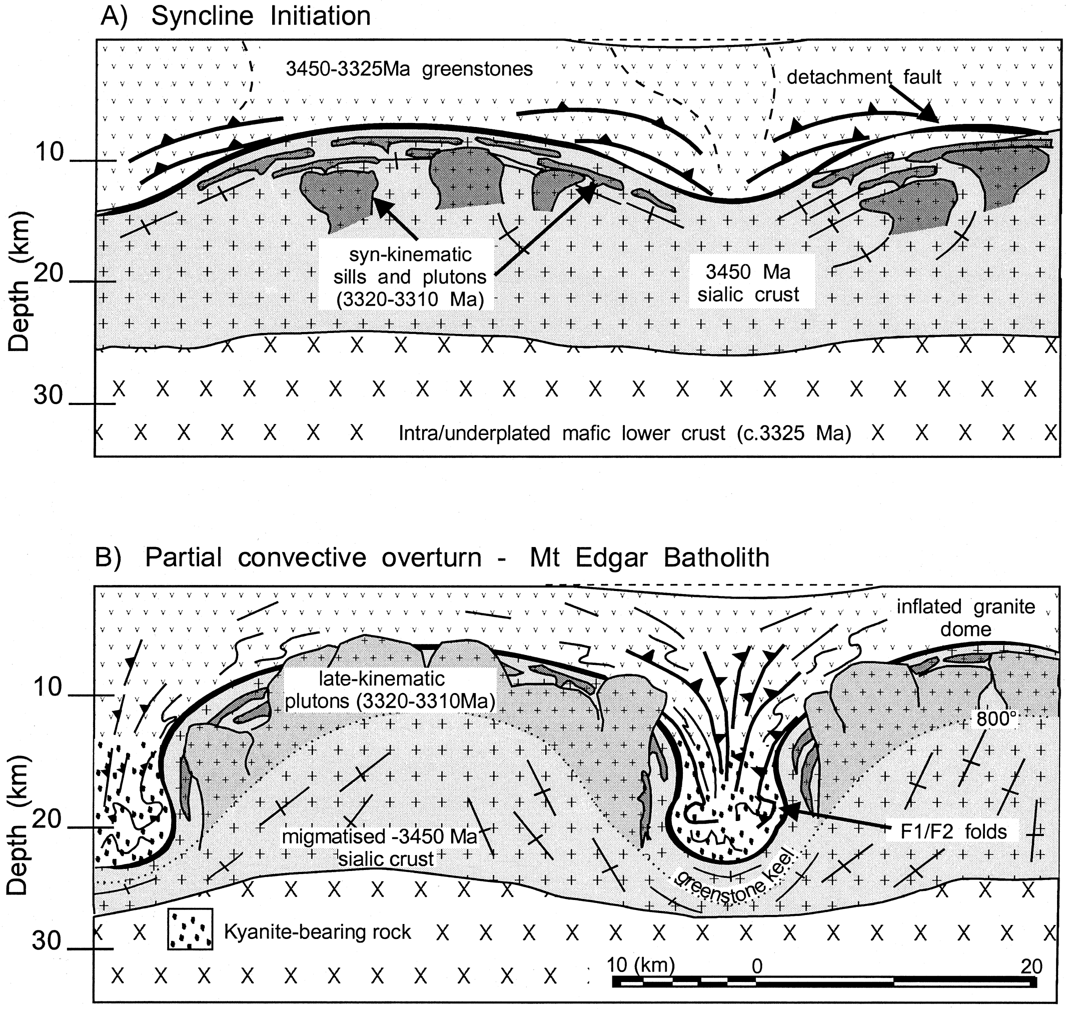

Fig. 7. Crustal overturn model applied to the east Pilbara. The geometry, kinematics and strain patterns are directly comparable with the Mareschal and West (1980) models. (a) Initiation of syncline associated with granite sheeting and pluton emplacement near basement–cover interface. (b) Continued granite emplacement after convection overturn. U–Pb SHRIMP ages for early-, syn- and post-kinematic granites of the Mount edgar Batholith are all within error. Note the formation of kyanite in the synclinal structure, in the zone of subvertical constriction, a feature that is difficult to simulate by other models of dome formation (see text). The 800°C isotherm is arbitrarily taken as the boundary above which granitic magma coalesces as plutons.

less strongly foliated bodies that are most obvious near the domal boundary (Fig. 1), where they cut the highly-deformed, basement–cover interface (Fig. 7b). The 800°C isotherm is arbitrarily taken as the boundary from which ~3320–3310 Ma granitoids segregate from older sialic crust.

Complete crustal overturn would result in the granitic layer extruding over the greenstone succession, but this is unlikely as crustal viscosities are generally high. Although convection is initiated during partial melting, when viscosity of the sialic substrate is relatively low, viscosities in the dome will increase with time because convective overturn is a cooling phenomena, particularly so if the dome reaches the Earth's surface where the upper thermal boundary layer is the atmosphere, an inadequate thermal buffer. Hence, the convective 'cell' will cool rapidly. Thus, Pilbara GGT evolution is considered to illustrate time-dependant convection, where the rate of overturn is controlled by the rate of cooling of the domal core. Typically, the process is arrested and structures become 'frozen-in' as the classic dome-and-keel geometry of Archaean GGTs. Hence, it is best described as *partial convective overturn*.

Thermally, partial convective overturn of the crust produces a 'cold-finger' effect in the synclines, with the geotherms depressed transiently during overturn, so that kyanite should be the stable aluminosilicate polymorph within the deeper parts of the syncline and synclinal wall (Fig. 7b; see also Collins and Van Kranendonk, in press). In addition, *P–T–t* paths for the deep synclinal wall and syncline follow near-isothermal burial in the kyanite stability field, suggesting that kyanite should be an early-, syn-, and late-kinematic mineral phase (Collins and Van Kranendonk, in press). Indeed, kyanite is the stable aluminosilicate polymorph in the high-strain rim of the Warrawoona Syncline, at *P–T* conditions of ~6 kbar and 500°C (Delor *et al.*, 1991), and it is an early-, syn-, and late- kinematic mineral (Collins and Van Kranendonk, in press, fig. 3). The metamorphic constraints confirm the suitability of the Mareschal and West models for the east Pilbara granite–greenstone terrain and the MEB dome in particular.

If the greenstone keel remains at depth, thermal recovery of the crust will raise the geotherm, causing kyanite to be overprinted by the higher-temperature aluminosilicate polymorphs, sillimanite and/or andalusite. However, if the syncline rapidly rebounds, kyanite might be preserved. Collins and Van Kranendonk (in press) argue that rapid rebound did occur in the high-strain zone of the Warrawoona Syncline as a result of outward migration of the main D_4 uplift faults. As the MEB tilted to the northeast, major uplift was transferred from the walls of the dome (D_4 shear zone; Figs 1 & 2) to the Klondyke and Salgash faults (Fig. 2), which now separate the low-grade greenstones from the deeper level, mylonitic kyanite schists. Thus, the gneisses and granites of the southwestern MEB rose during D_4 uplift, along with the kyanite aureole, but the rest of the Warrawoona Syncline and the entire Marble Bar Belt remained in their pre-D_4 domal configuration. In this way, the deep-level schists were juxtaposed against regional, greenschist facies, low-strain greenstones and kyanite was preserved.

The presence of kyanite in almost all Archaean granite–greenstone terrains (Percival, 1979) suggests a general applicability for the crustal overturn model. Although a rare mineral in Archaean GGTs, it usually occurs in the same structural setting, *viz*., only within the highly strained wallrocks near a batholith margin, which is its predicted location based on the crustal overturn model. An exception is the Pontiac Subprovince of Canada, where kyanite occurs within a large-scale duplex structure that represents the deeper part of an overthrust belt (Benn *et al.*, 1994). This setting for kyanite is entirely different to that of GGTs characterised by dome-and-keel structures like the Pilbara and reflects exhumation of mid-crustal sections during lateral accretion of lithotectonic assemblages.

As pointed out by Collins and Van Kranendonk (in press) restriction of kyanite to the domal margins suggests that it is unlikely to have formed by other doming mechanisms, such as core complex, gneiss dome or ballooning pluton development, for it should occur either within the dome and rim, or within neither. For example, if kyanite formed during crustal thickening and the domes represent diapiric uprise of thickened sialic portions (Bickle *et al.*, 1980, 1985; see also Kusky, 1993), then kyanite should be widespread within the Pilbara domes, but this has not been observed. Rather, andalusite and sillimanite are the dominant aluminosilicates within the batholiths (Bickle *et al.*, 1985). Therefore, the regional structural and metamorphic features of those Archaean GGTs characterised by dome-and-keel geometry are consistent with partial convective overturn of Archaean crust.

A THREE-DIMENSIONAL VIEW OF THE PILBARA

The added significance of the MEB kyanite-bearing rim is that it provides a three-dimensional glimpse of the deep crustal levels within Archaean GGTs. Most important is the concordance of the higher-level, subvertical greenschist facies rocks in the Warrawoona Syncline with the subvertical, deep-level kyanite-schists, which have been buried to at least 6 kbar (Delor *et al.*, 1991). This indicates that the entire greenstone keel had a similar subvertical orientation to ~20 km. The steep-sided wall of the keel accords with the subvertical foliations in amphibolitic 'xenolith trails' (Glikson, 1984), which commonly reflect deeper levels of exposure within the domes. Preservation of kyanite in the keel of the Warrawoona Syncline confirms that the greenstones sunk to the middle–lower crust, confirming that convective overturn in the Pilbara was a crustal-scale process, similar in magnitude to that recognised in the Dharwar craton, where the dome-and-keel structures can be observed at up to 8–9 kbar (Choukroune *et al.*, 1995).

These fragmentary pieces of evidence are the vital clues to the original three-dimensional structure of the domes. Seismic evidence suggesting that many greenstone belts do not extend to depth is inconclusive, because the synclines should be obliterated by partial melting during thermal relaxation, as discussed earlier. In addition, those seismic profiles that show shallow subhorizontal reflectors beneath greenstone belts (Calvert *et al.*, 1995) might reflect a subsequent subhorizontal tectonic event that is not directly related to greenstone belt formation (e.g. extensional collapse; Moser *et al.*, 1996). Until the age of such horizontal reflectors can be determined, their bearing on greenstone belt evolution remains in the realms of speculation.

PARTIAL CONVECTIVE CRUSTAL OVERTURN: A UNIQUE ARCHAEAN PROCESS?

The dome-and-keel pattern of GGTs is characteristic of the Archaean and has provided the focus for much geological debate. In this section, the possible conditions and processes which are considered to have given rise to crustal overturn in the east Pilbara are examined. As the presence of komatiites is taken to infer the presence of mantle plumes (Campbell *et al.*, 1989; Nisbet *et al.*, 1993; McDonough and Ireland, 1993; Abbott, 1996), the close association between komatiite and voluminous basalt eruption at ~3325 Ma, and granite generation and deformation at

~3320–3310 Ma, suggests that the dome-and-keel structures of the Pilbara formed in response to a hot plume head, emplaced near the base of the pre-existing (3450 Ma) crust. Although neither Mareschal and West (1980), nor Dixon and Summers (1983) considered external heat sources for their diapiric models, the close correspondence in age between komatiites and batholithic domes in the east Pilbara suggests a causal link, as discussed below.

Several features of Pilbara geology support a plume model for GGT evolution at ~3300 Ma, involving conductive heating of the crust. The ~5–20 Ma time-lag between greenstone eruption and granite generation is typical of the time required for heat to conduct through the crust above a plume head (Campbell and Hill, 1988). If the plume entered the crust (advective heating), melting would occur rapidly and efficiently, producing abundant granites of similar age to the greenstones, but this is not observed. Also, such crustal melts should mix efficiently with basaltic magma from the plume head to produce hybrid magmas of intermediate composition, and they should contain abundant mafic enclaves similar to those formed by crust–mantle mixing in present-day magmatic arcs. However, this also is not observed: 3300–3000 Ma magmas from the Pilbara show minimal Nd isotopic or chemical evidence of a substantial mantle contamination (Bickle *et al*., 1989, 1993; Collins, 1993) and mafic enclaves are rare to non-existent. The lack of mantle input in the granites, and their generation and emplacement into a slightly older, komatiite-bearing, greenstone succession, is evidence that suggests a plume model involving conductive rather than advective heating of the crust.

Plume head sizes greatly exceed tail sizes and as a consequence, the head rapidly delivers most of the heat, estimated to occur over a few million years (Griffiths *et al*., 1989). Mature plume heads typically vary in diameter between 1000–2000 km, and are capable of generating a thermal anomaly just beneath the lithosphere in the order of 10^6–10^7 km^2, or more. Given the general plume head size, the rapid rate of heat transfer, and the subsequent crustal overturn, little chance exists for the plume to develop a 'hot-spot track' in a region such as the Pilbara, where the areal extent of the east Pilbara GGT is less than 1×10^5 km^2, possibly as low as 3×10^4 km^2. It also suggests that the plume tail effects are unlikely to be seen within the same greenstone succession. Small komatiite-bearing greenstone belts, such as the 3240 Ma Strelley succession in the central Pilbara (Van Kranendonk and Collins, 1998), might represent the effects of the plume tail (Campbell *et al*., 1989).

Campbell *et al.* (1989) and Nisbet *et al.* (1993) suggested that Archaean mantle potential temperatures were probably 200°C hotter than present-day mantle. This inference was supported by Abbott *et al.* (1994), who synthesised chemical data from primitive mafic and ultramafic rocks erupted over the last 3.5 Ga and showed that the maximum temperature of eruption has declined exponentially over the Earth's history. They calculated ambient Archaean mantle temperatures (at > 3.0 Ga) to be at least 100°C greater than present.

A hotter Archaean ambient mantle probably produced a higher crust:lithosphere thickness ratio than the present-day mantle. Hotter mantle produces thicker oceanic crust (Bickle, 1986; Langmuir *et al*., 1992): present-day thicknesses of oceanic crust, produced by decompression melting at mid-ocean ridges, are 7–8 km, whereas Archaean thicknesses were probably ~24–30 km (McKenzie and Bickle, 1988; Abbott 1996). However, the lithosphere was probably much thinner than present-day, as it is largely a mechanical boundary layer produced by conductive cooling. Its thickness reflects the ~1200°C isotherm, which marks the rheological change from 'rigid' to 'convecting' mantle (Wilson, 1987). An estimate of Archaean continental lithosphere thickness can be gleaned from the 3.2–3.3 Ga inclusions in diamonds from Cretaceous kimberlite pipes in the Barberton GGT (Richardson *et al.,* 1984), which suggests that the lithosphere was at least 150 km thick *after* dome-and-keel development. Given that the inclusions post-date crustal stabilisation associated with dome-and-keel formation, a pre-doming thickness estimate of < 100 km seems appropriate, which is similar to that suggested by Campbell and Hill (1988). With the inferred exponential cooling of the Earth (Abbott *et al*., 1994), the lithosphere thickness should have generally increased with time, whereas the crustal thickness should have decreased, thereby increasing the crust:lithosphere thickness ratio.

Archaean greenstone belts are characterised by a relative abundance of komatiite eruptives, compared with present-day continental flood basalt (CFB) provinces, suggesting that hotter plumes probably also existed at this time. Campbell and Griffiths (1992) used primitive mafic magma geochemistry to calculate the maximum change in plume temperatures from the Archaean to present-day is 250 ± 100°C. Nisbet *et al.* (1993) suggested temperatures > 1600°C, compared with 1450–1500°C in present-day plumes (White and McKenzie, 1995). As a result, the average thickness of erupted Archaean 'flood basalts' provinces (Archaean GGTs) was probably greater than at present, which is supported by their generally greater thickness compared with younger CFB provinces. For example, the ~3325–3310 Ma 'flood basalts' of the upper Warrawoona and Gorge Creek groups in the Pilbara have a combined thickness of ~10 km (Hickman, 1990). Most well-documented post-Archaean CFB provinces have *maximum* thicknesses of 3.5 km or less (Wilson, 1987), with only the Keweenawan (12 km) and Karoo (9 km) successions possibly being of comparable thickness. Nonetheless, the estimate of Keweenawan thickness by White and McKenzie (1995) is only 8–9 km. The compilation indicates that post-Archaean CFB provinces rarely achieve the thickness of the Pilbara greenstone successions, and komatiites are rare to non-existent. However, the Keweenawan and Karoo crust did not overturn, even though dense mafic rocks overlie sialic crust, so deformation-inducing factors other than negative buoyancy must also be involved.

The conductive heating effects of a rising mantle plume on overlying sialic crust should have been much greater in the Archaean. The inferred high crust:lithosphere thickness ratio in the Archaean suggests that plumes were emplaced much closer to the crust–mantle boundary than at present. Impingement of a $>1600°C$ mantle plume head on the Archaean lithosphere would have resulted in considerable thermal erosion of the lithospheric plate, particularly if it was actively extending (White and McKenzie, 1995). Campbell and Hill (1988) calculated that plumes approached to within 20–40 km of the base of the crust, assuming a conductive model for heat transfer (see earlier). Such high emplacement levels implies considerable crustal extension and heat input during greenstone deposition.

Heat conducted upward to the base of the Pilbara crust by the mantle plume, perhaps combined with a thermal blanketing effect produced by the overlying >10 km thick insulating layer of greenstone (Mareschal and West, 1980), is likely to induce extensive thermal softening and widespread anatexis of sialic middle/lower crustal material. We demonstrated earlier in the discussion of Fig. 4 that doming occurred during sheeted intrusion along the southern periphery of the MEB and was accompanied by anatexis of the underlying sialic crust. Large-scale anatexis in the MEB, at $\sim$3320–3310 Ma, is indicated by the wholesale generation of the synkinematic granites and 'intrusive diatexites'. U–Pb isotopic analysis indicates that the diatexites are remobilised $\sim$3450 Ma orthogneisses that underwent ancient Pb loss no earlier than 3380 ± 8 Ma, based on the age of zircon overgrowths (Williams and Collins, 1990, p. 49). Syn- to late-kinematic granitoids of the MEB, representing $>80\%$ of the batholith, formed at $\sim$3320–3310 Ma, throughout the structural development of the dome.

Collins (1993) argued on petrological and geochemical grounds that the $\sim$3320–3310 Ma granitoids of the MEB are largely recycled sialic crust from the 3450 Ma TTG suite. A wholesale crustal recycling model is strongly supported by the Nd isotopic data of Bickle *et al.* (1989, 1993), which shows that the 3300–3000 Ma granites of the Shaw Batholith in the East Pilbara lie near the same Nd isotopic growth curve as the 3450 Ma granitoids, suggesting derivation from those rocks. The Nd isotopic growth curve for Archaean mantle diverges markedly from the granitoid growth curve, indicating minimal mantle contribution to granite magma composition. Therefore, combined field, structural, isotopic and chemical evidence indicates widespread softening and melting of the sialic basement, but not crust–mantle mixing, during doming. The evidence supports a conductive, rather than advective heating model of the crust, generated by plume emplacement at $\sim$3325 Ma, which induced widespread recycling of pre-existing sialic crust.

In the east Pilbara, the effect of massive crustal recycling was two-fold: It decreased the density of the substrate by generating abundant granitoid magma through crustal recycling and it thermally softened the substrate, reducing it to a migmatitic 'mush'. This two-fold effect markedly reduced the viscosity of the substrate, allowing the dense greenstone overburden to 'sink' through the sialic basement and initiate convective overturn.

If plumes can cause massive crustal recycling and crustal overturn in Archaean GGTs, why are the most recent examples of plume-related volcanism so ineffective? The Karoo and Deccan flood basalts, for example, have few if any coeval felsic lavas, and evidence for convective overturn is lacking. Massive CFB eruptions occur if plumes are emplaced during active lithospheric thinning, as large volumes of melt are generated by decompression of anomalously hot mantle in the plume head (White and McKenzie, 1995). A good example is the Keweenawan lavas, which were emplaced into an active intracontinental rift, although those lavas also appear to have been exceptionally hot. However, as it is generally considered that plates were smaller, mid-ocean ridges longer, and plate velocities more rapid in the Archaean (Bickle, 1986), the chance of these optimal conditions producing massive CFBs (= greenstone belts) was much greater at that time. Alternatively, if the lithosphere is near-stationary and thick, decompression melting is limited to convective flow in the plume centre, a more likely situation in post-Archaean times because the lithosphere has become progressively thicker as the Earth progressively cools. Accordingly, the chances of producing massive CFB provinces is significantly less in the post-Archaean, and the possibility of plume emplacement close to the base of continental crust, inducing pervasive anatexis and softening, has become extremely small. Therefore, even though thicknesses in some post-Archaean CFB provinces are comparable to the Archaean greenstone succession of the Pilbara (e.g. Keweenawan), the crust was not sufficiently thermally softened, remaining too viscous to yield to Rayleigh–Taylor instability. As a result, the possibility of convective crustal overturn has decreased with the progression of time.

Other models

The proposed sequence of events at *ca* 3300 Ma in the east Pilbara is comparable to the crust-formation model proposed for the Kalgoorlie–Norseman area of the Archaean Yilgarn Craton by Campbell and Hill

(1988). In both areas, granite intrusion and doming occurred up to 20 Ma after mantle plume activity. However, Campbell and Hill (1988) considered that the granitic magmas rose diapirically as plutons, and were the driving force for greenstone belt deformation. Rather, it is considered here that the deformation was initiated by the sinking of the overlying thick, negatively-buoyant greenstones, rather than rise of the granites.

A partial convective overturn model is also similar to the sagduction model proposed for the Late Archaean dome-and-keel structure of the Dharwar craton, India (Bouhallier *et al.*, 1995; Chardon *et al.*, 1996), although that terrain has been modified by synchronous and/or later strike-slip movement. The Dharwar craton is a tilted crustal section, and dome-and-keel structures can be observed from greenschist to granulite facies, showing the same type of geometry and strain pattern throughout. This evidence indicates that 'sagduction' was a crustal-scale process. However, we prefer to emphasise the convective response of the whole crust to Rayleigh–Taylor instability, facilitated by thermal softening associated with large-scale crustal recycling. We view the sinking, or sagduction of greenstones, as only a part of the whole process.

Our work from the east Pilbara generally supports the three-stage crustal evolution model of Choukroune *et al.* (1995), who emphasise that transient 'intense and total reheating' of pre-existing sialic crust caused thermal softening and a decrease in density, thereby promoting diapirism. To this model, we add evidence that reheating was caused by plume activity, that eruption of a ~10 km thick, successor greenstone belt upon the older (~3450 Ma), greenstone sequence magnified the crustal instability, that thermal softening of the sialic crust was caused by widespread anatexis, that the voluminous granites were generated from this crust, and that they were emplaced during crustal overturn.

We also consider that the east Pilbara is an end-member example of Archaean tectonic processes, where the typical dome-and-keel structure of GGTs is produced by Raleigh–Taylor instabilities within the crust, and does not require external body forces to initiate deformation. If White and McKenzie (1995) are correct, thick greenstone belts might require coeval crustal extension, perhaps induced by thermal upwelling, but coeval compression is highly unlikely. The other end-member tectonic setting is the alternating, sublinear, thrust-bound greenstone and supracrustal belts of the Superior Province, Canada, which suggests an assembly of lithological packages at the margin of a proto-craton during rigid plate convergence. Other Archaean GGTs appear to contain features of both processes, such as the Dharwar craton of India, the Yilgarn of Western Australia, and the Hebei Province of China (Choukroune *et al.*, 1995).

It is possible that the different GGTs reflect proximity to craton margins, with the Pilbara as the interior and the Superior as the periphery (Choukroune *et al.*, 1997), which is supported by the ~140 Ma paraconformity between the 3450 Ma and 3225 Ma greenstone belts of the Pilbara. Alternatively, the exclusive development of dome-and-keel structures might reflect proximity to a plume head centre, where body forces associated with thermal upwelling and gravitational instabilities transiently overcame horizontal forces associated with plate convergence. Another alternative is that GGTs such as the Pilbara may have simply formed during a period when strong external (horizontal) body forces were not operative, possibly on a plate bounded by mid-ocean ridges where much of the weak boundary force was 'ridge-push', much like the African continent is today.

Acknowledgements—Supported by an ARC post-doctoral fellowship to MVK and ARC-funded small research grants. Damian Kinder and Stephen Dick are thanked for computer drafting. The paper benefitted substantially from reviews by W. M. Schwerdtner, P. Choukroune and an anonymous reviewer, and from the editorial comments of E. W. Sawyer.

REFERENCES

Abbott, D. H. (1996) Plumes and hotspots as sources of greenstone belts. *Lithos* **37,** 113–127.

Abbott, D. H., Burgess, L., Longhi, J. and Smith, W. H. F. (1994) An empirical thermal history of the Earth's upper mantle. *Journal of Geophysical Research* **99,** 13 835–13 850.

Barley, M. E. (1981) Relations between volcanic rocks in the Warrawoona Group: continuous or cyclic evolution? *Geological Society of Australia, Special Publication* **7,** 361–372.

Benn, K., Miles, W., Ghassemi, M. R. and Gillett, J. (1994) Crustal structure and kinematic framework of the northwestern Pontiac Subprovince, Quebec: an integrated structural and geophysical study. *Canadian Journal of Earth Sciences* **31,** 271–281.

Bickle, M. J. (1986) Implications of melting for stabilisation of the lithosphere and heat loss in the Archaean. *Earth and Planetary Science Letters* **80,** 314–324.

Bickle, M. J., Bettenay, L. F., Boulter, C. A., Groves, D. I. and Morant, P. (1980) Horizontal tectonic interaction of an Archaean gneiss belt and greenstones, Pilbara Block, Western Australia. *Geology* **8,** 525–529.

Bickle, M. J., Bettenay, L. F., Chapman, H. J., Groves, D. I., McNaughton, N. J., Campbell, I. H. and de Laeter, J. R. (1989) The age and origin of younger granitic plutons of the Shaw Batholith in the Archaean Pilbara Block, Western Australia. *Contributions to Mineralogy and Petrology* **101,** 239–283.

Bickle, M. J., Bettenay, L. F., Chapman, H. J., Groves, D. I., McNaughton, N. J., Campbell, I. H. and de Laeter, J. R. (1993) Origin of the 3500–3300 Ma calc-alkaline rocks in the Pilbara Archaean: isotopic and chemical constraints from the Shaw Batholith. *Precambrian Research* **60,** 117–150.

Bickle, M. J., Morant, P., Bettenay, L., Boulter, C. A., Blake, T. S. and Groves, D. I. (1985) Archaean tectonics of the Shaw Batholith, Pilbara Block, Western Australia: Structural and metamorphic tests of the batholith concept. In *Evolution of Archaean Supracrustal Assemblages*, eds L. D. Ayers, P. C. Thurston, K. D. Card and W. Weber, pp. 325–341. Geological Association of Canada, Special Paper, **28**.

Bouhallier, H., Chardon, D. and Choukroune, P. (1995) Strain patterns in Archaean dome-and-basin patterns: The Dharwar craton (Karnataka, India). *Earth and Planetary Science Letters* **135,** 57–75.

Calvert, A. J., Sawyer, E. W., Davis, W. J. and Ludden, J. N. (1995) Archaean subduction inferred from seismic images of a mantle suture in the Superior Province. *Nature* **375,** 670–674.

Campbell, I. H. and Griffiths, R. W. (1992) The changing nature of mantle hotspots through time: implications for the chemical evolution of the mantle. *Journal of Geology* **92,** 497–523.

Campbell, I. H. and Hill, R. I. (1988) A two-stage model for the formation of the granite–greenstone terrains of the Kalgoorlie–Norseman area, Western Australia. *Earth and Planetary Science Letters* **90,** 11–25.

Campbell, I. H., Griffiths, R. W. and Hill, R. I. (1989) Melting in an Archaean mantle plume: heads it's basalts, tails it's komatiites. *Nature* **339,** 697–699.

Chardon, D., Choukroune, P. and Jayananda, M. (1996) Strain, patterns, décollement and incipient sagducted greenstone terrains in the Archaean Dharwar craton (south India). *Journal of Structural Geology* **18,** 991–1004.

Choukroune, P., Bouhallier, H. and Arndt, N. T. (1995) Soft lithosphere during periods of Archaean crustal growth or crustal reworking. In *Early Precambrian Processes*, eds M. P. Coward and A. C. Ries, pp. 67–86. Geological Society Special Publications, **95**.

Choukroune, P., Ludden, J. N., Chardon, D., Calvert, A. J. and Bouhallier, H. (1997) Archaean crustal growth and tectonic processes: a comparison of the Superior Province, Canada and the Dharwar Craton, India. In *Orogeny Through Time*, eds J.-P. Burg and M. Ford, pp. 63–98. Geological Society Special Publication, **121**.

Collins, W. J. (1989) Polydiapirism of the Archaean Mount Edgar Batholith, Pilbara Block, Western Australia. *Precambrian Research* **43,** 41–62.

Collins, W. J. (1993) Melting of sialic crust under high a_{H_2O} conditions: genesis of 3300 Ma old Na-rich granitoids in the Mount Edgar Batholith, Pilbara Block, Western Australia. *Precambrian Research* **60,** 151–174.

Collins, W. J. and Gray, C. M. (1990) Rb–Sr isotopic systematics of an early Archaean granite–gneiss terrain: the Mount Edgar Batholith, Pilbara Block, Western Australia. *Australian Journal of Earth Sciences* **37,** 9–22.

Collins, W. J. and Van Kranendonk, M. J. (1998) Model for the development of kyanite during partial convective overturn of Archaean granite–greenstone terrains: the Pilbara Craton, Australia. *Journal of Metamorphic Geology* (in press).

Coney, P. J. (1980) Cordilleran metamorphic core complexes: An overview. In *Cordilleran Metamorphic Core Complexes*, pp. 7–13. Geological Society of America Memoirs, **153**.

Davids, C., Wijbrans, J. R. and White, S. H. (1997) $^{40}Ar/^{39}Ar$ laserprobe ages of metamorphic hornblendes from the Coongan Belt, Pilbara, Western Australia. *Precambrian Research* **83,** 221–242.

Delor, C., Burg, J.-P. and Clarke, G. L. (1991) Relations diapirisme–métamorphisme dans la province du Pilbara (Australie occidentale): implications pour les régimes thermiques et tectoniques à l'Archéen. *Compte-Rendu de l'Acadamie des Sciences de Paris, Série II* **312,** 257–263.

Dimroth, E., Imreh, L., Goulet, N. and Rocheleau, M. (1983) Evolution of the south-central segment of the Archaean Abitibi Belt, Quebec. Part II: Tectonic evolution and geomechanical model. *Canadian Journal of Earth Sciences* **20,** 1355–1373.

Dixon, J. M. and Summers, J. M. (1983) Patterns of total and incremental strain in subsiding troughs: Experimental centrifuge models of inter-diapir synclines. *Canadian Journal of Earth Sciences* **20,** 1843–1861.

Glikson, A. Y. (1984) Significance of Early Archaean mafic–ultramafic xenolith patterns. In *Archaean Geochemistry*, ed. A. Kroner, pp. 262–281. Springer-Verlag, Berlin.

Griffiths, R. W., Gurnis, M. and Eitelberg, G. (1989) Holographic measurements of surface topography in laboratory models of mantle hotspots. *Geophysical Journal of the Royal Astronomical Society* **96,** 477–495.

Hickman, A. H. (1984) Archaean diapirism in the Pilbara Block, Western Australia, in *Precambrian Tectonics Illustrated*, eds A. Kroner and R. Greiling, pp. 113–127, E. Schweizerbartische Verlagbuchhandlung, Stuttgart.

Hickman, A. H. (1990) Geology of the Pilbara craton. In *Third International Archaean Symposium, Excursion Guide 5*, eds S. E. Ho, J. E. Glover, J. S. Myers and J. R. Muhling, Vol. 21, pp. 1–13. Geology Department and University Extension, The University of Western Australia.

Hickman, A. H. and Lipple, S. L. (1978) *Explanatory Notes on the Marble Bar Geological Sheet (SF50-8)*. Geological Survey of Western Australia.

Hudleston, P. J., Schultz-Ela, D. and Southwick, D. L. (1988) Transpression in an Archaean greenstone belt, northern Minnesota. *Canadian Journal of Earth Sciences* **25,** 1060–1068.

Kusky, T. M. (1993) Collapse of Archaean orogens and the generation of late- to post-kinematic granitoids. *Geology* **21,** 925–928.

Langmuir, C. H., Klein, E. M. and Plank, T. (1992) Petrological constraints of mid-ocean ridge basalts: Constraints on melt generation beneath ocean ridges. In *Mantle Flow and Melt Generation*, eds J. P. Morgan, D. K. Blackman and J. M. Sinton, pp. 183–280. Transactions of the American Geophysical Union, **71**.

Mackey, T. E. and Richardson, L. M. (1997) *Airborne gamma-ray spectrometry colour composite pixel-image map of Marble Bar 1:250 000, Western Australia.* Australian Geological Survey Organisation.

Mareschal, J.-C. and West, G. F. (1980) A model for Archaean tectonism: Numerical models of vertical tectonism in greenstone belts. *Canadian Journal of Earth Sciences* **17,** 60–71.

McDonough, W. F. and Ireland, T. R. (1993) Intraplate origin of komatiites inferred from trace elements in glass inclusions. *Nature* **365,** 432–434.

McKenzie, D. P. and Bickle, M. J. (1988) The volume and composition of melt generated by extension of the lithosphere. *Journal of Petrology* **29,** 625–679.

McNaughton, N. J., Compston, W. and Barley, M. E. (1993) Constraints on the age of the Warrawoona Group, eastern Pilbara Block, Western Australia. *Precambrian Research* **60,** 69–98.

Moser, D. E., Heaman, L. M., Krogh, T. E. and Hanes, J. A. (1996) Intracrustal extension of an Archaean orogen revealed using single-grain U–Pb zircon geochronology. *Tectonics* **15,** 1093–1109.

Myers, J. S. and Watkins, K. P. (1985) Origin of granite–greenstone patterns, Yilgarn Block, Western Australia. *Geology* **13,** 778–780.

Nijman, W., Willigers, B. J. A. and Krikke, A. (1998) Tensile and compressive growth structures: relationships between sedimentation, deformation and granite intrusion in the Archaean Coppin Gap greenstone belt, eastern Pilbara, Western Australia. *Precambrian Research* **88,** 83–108.

Nisbet, E. G., Cheadle, M. J., Arndt, N. T. and Bickle, M. J. (1993) Constraining the potential temperature of the Archaean mantle: a review of the evidence from komatiites. *Lithos* **30,** 291–307.

Percival, J. A. (1979) Kyanite-bearing rocks from the Hackett River area, N.W.T.: Implications for Archaean geothermal gradients. *Contributions to Mineralogy and Petrology* **69,** 177–184.

Pidgeon, R. T. (1984) Geochronological constraints on early volcanic evolution of the Pilbara Block, Western Australia. *Australian Journal of Earth Sciences* **31,** 237–242.

Ramsay, J. G. (1967) *Folding and Fracturing of Rocks*. McGraw-Hill, London.

Reynolds, S. J. and Lister, G. S. (1990) Folding of mylonitic zones in Cordilleran metamorphic core complexes: evidence from near the mylonitic front. *Geology* **18,** 216–219.

Richardson, S. H., Gurney, J. J., Erlank, A. J. and Harris, J. W. (1984) The origin of diamonds in old enriched mantle. *Nature* **310,** 198–202.

Sawyer, E. W. and Benn, K. (1993) Structure of the high-grade Opatica Belt and adjacent low-grade Abitibi subprovince, an Archaean mountain front. *Journal of Structural Geology* **15,** 1443–1458.

Schwerdtner, W. M. (1990) Structural tests of diapir hypotheses in Archaean crust of Ontario. *Canadian Journal of Earth Sciences* **27,** 387–402.

Scott, R. J. and Lister, G. S. (1995) Analogue modeling of detachment systems and core complexes: Comment. *Geology* **23,** 287–288.

Teyssier, C. and Collins, W. J. (1990) Strain and kinematics during the emplacement of the Mount Edgar Batholith and Warrawoona Syncline, Pilbara Block, Western Australia. In *Third International Archaean Symposium, Extended Abstracts*, pp. 481–483. Geoconferences (W.A.) Inc., Perth, Western Australia.

Thorpe, R. A., Hickman, A. H., Davis, D. W., Mortensen, J. K. and Trendall, A. F. (1992) U–Pb zircon geochronology of Archaean felsic units in the Marble Bar region, Pilbara Craton, Western Australia. *Precambrian Research* **56,** 169–189.

van Haaften, W. M. and White, S. H. (1998) Evidence for multiphase deformation in the Archaean basal Warrawoona Group in the Marble Bar area, east Pilbara, Western Australia. *Precambrian Research* **88,** 53–66.

Van Kranendonk, M. J. and Collins, W. J. (1998) Timing and tectonic significance of Late Archaean sinistral strike-slip deformation in the Central Pilbara structural corridor, Pilbara craton, Western Australia. *Precambrian Research* **88,** 207–232.

White, R. S. and McKenzie, D. (1995) Mantle plumes and flood basalts. *Journal of Geophysical Research* **100,** 17543–17585.

Williams, I. S. and Collins, W. J. (1990) Granite–greenstone terranes in the Pilbara Block, Australia, as coeval volcano–plutonic complexes; evidence from U–Pb zircon dating of the Mount Edgar Batholith. *Earth and Planetary Science Letters* **97,** 41–53.

Wilson, M. (1987) *Igneous Petrogenesis: A Global Tectonic Approach.* Unwin and Hyman, London.

Zegers, T. E., White, S. H., de Keijzer, M. and Dirks, P. G. H. M. (1996) Extensional structures during deposition of the 3460 Ma Warrawoona Group in the eastern Pilbara craton, Western Australia. *Precambrian Research* **80,** 89–105.