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The northern margin of the Capricorn Orogen, Western Australia—an example of an Early Proterozoic collision zone

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Abstract—The Early Proterozoic (2200–1600 Ma) Capricorn Orogen is a major zone of deformation, metamorphism and granite emplacement between the Archaean Yilgarn and Pilbara Cratons of northwest Australia. Capricorn Orogen structures are recognized along the southern margin of the Pilbara Craton deforming the Archaean granite–greenstone terrain of the Sylvania Inlier together with rocks of the Hamersley, Ashburton and Blair Basins.

In the Sylvania Inlier deformation is characterized by large-scale shear zones interpreted as NNE-directed thrusts. In the Hamersley Basin, orogenic deformation forms the Ophthalmia Fold Belt and can be divided into two groups of structures: in the southwest large-scale dome-and-basin folds (pre-Ashburton Basin) are present; in the southeast folds and faults are part of a younger (syn-Ashburton Basin) N-directed foreland fold and thrust belt directly linked to deformation in the Sylvania Inlier.

The deformed Ashburton Basin and adjacent parts of the Hamersley Basin form the Ashburton Fold Belt. Two periods of deformation are recognized: early recumbent folding (D_{1a}, post-Ashburton Basin–pre-Blair Basin), and later dextral wrench faulting and associated folding (D_{2a}, syn- or post-Blair Basin).

Capricorn Orogen structures are interpreted as the result of an oblique continent–continent collision between the Pilbara and Yilgarn Cratons. Along the southern Pilbara margin, collision first occurred in the southeast and migrated northwest, with the Ashburton Basin evolving as a foreland basin. Late-stage dextral wrench faulting was the result of westward extrusion of material caught between the opposing craton margins.

INTRODUCTION

The Capricorn Orogen (Fig. 1) was defined by Gee (1979, p. 352) as "a major orogenic zone involving geosynclinal sedimentation, metamorphism, basement reworking and granitoid emplacement in the region between the Yilgarn Block and the Pilbara Craton. It takes its name from the Tropic of Capricorn, at which latitude it is approximately located". It is synonomous with the "Median Belt" of Horwitz & Smith (1978). The Yilgarn Block has since been redefined as the Yilgarn Craton (Geological Survey of Western Australia 1990). The orogen is made up of several tectonic units namely the Ashburton Basin, the Gascoyne Complex and the Nabberu Basin, together with the basement rocks beneath the Bangemall Basin. In this paper it is taken to also include areas on the margins of the two cratons where foreland deformation has occurred, and involves rocks in the Sylvania Inlier, the Hamersley Basin, the Marymia Inlier and the Narryer Gneiss Complex (Fig. 1). Available geochronological data, reviewed and summarized by Libby *et al.* (1986), suggest that the orogen developed between 2200 and 1600 Ma ago.

This paper is concerned with the tectonic development of the northern margin of the orogen where it impinges on the Pilbara Craton (see Fig. 2). It deals primarily with foreland deformation in the Pilbara Craton margin and with deformation of the Ashburton and Blair Basins.

STRATIGRAPHY

Pilbara Craton

The Pilbara Craton (Fig. 1) consists of an Archaean (>2750 Ma) granite–greenstone terrain unconformably overlain by supracrustal rocks of the Late Archaean to Early Proterozoic (2750–2300 Ma) Hamersley Basin (Trendall 1983). The granite–greenstone terrain outcrops over most of the northern half of the craton and is also exposed in inliers in the southern part of the Hamersley Basin. The largest of these is the Sylvania Inlier (Tyler 1990a, in press).

The Hamersley Basin succession comprises three conformable rock units (Table 1) that together make up the Mount Bruce Supergroup (Trendall 1979). The lowest is the basalt-dominated Fortescue Group. This is overlain by the Hamersley Group, which consists of banded iron-formation, carbonate, fine-grained siliciclastic rock and acid volcanic rock. The uppermost unit is the Turee Creek Group comprising fine- to coarse-grained siliciclastic rocks with locally developed chemical deposits. This stratigraphy records the development of the basin from an initial rift (Blake & Groves 1987), to a stable shelf or platform open to the deep ocean and remote from terrigenous sediment supply (Horwitz & Smith 1978, Ewers & Morris 1981, Morris & Horwitz 1983, McConchie 1984). The final stages of basin development produced shallower water conditions combined with increasing tectonic activity (Trendall 1983).

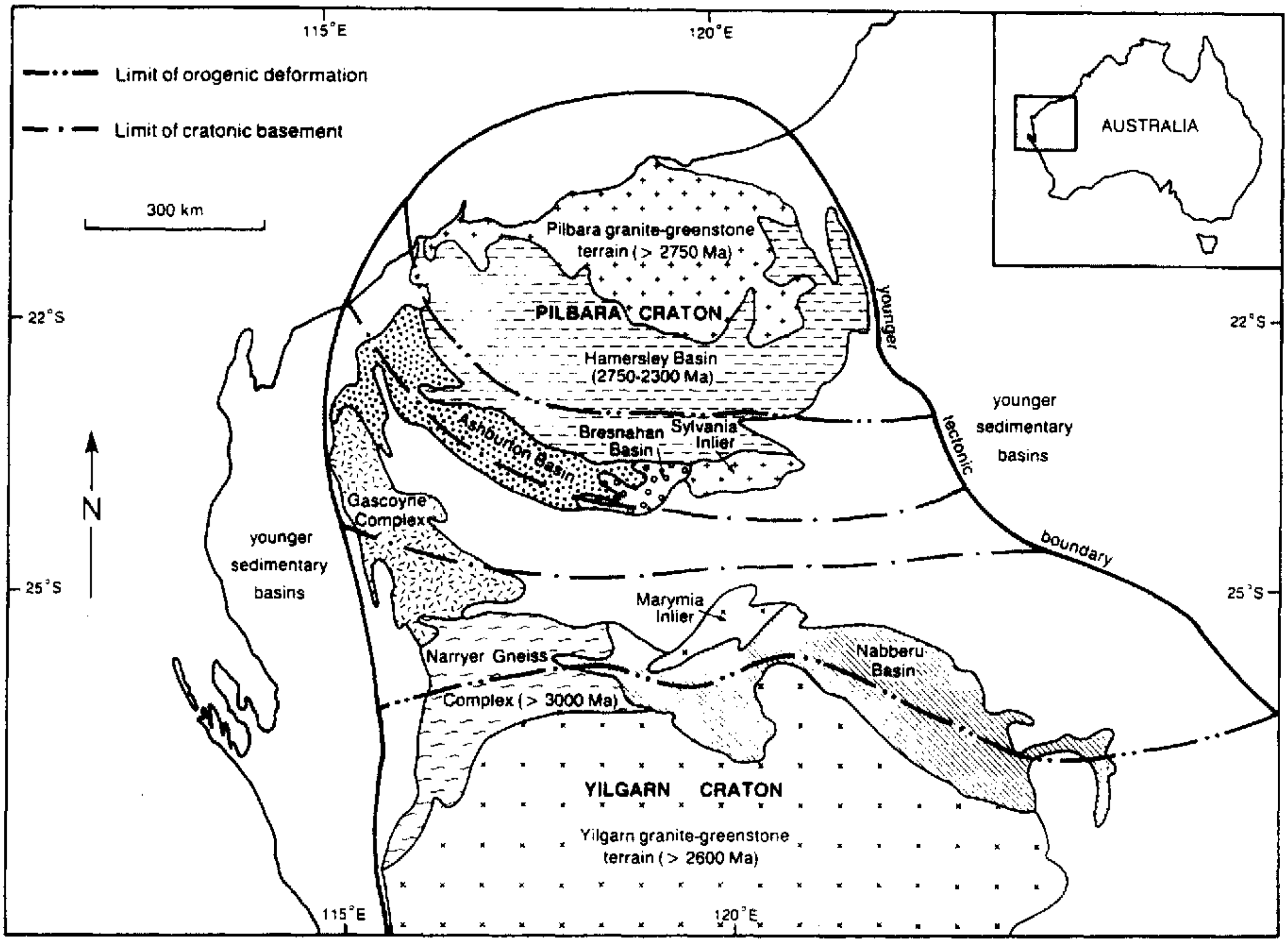

Fig. 1. Map showing the main tectonic units of the Pilbara Craton, the northern Yilgarn Craton and the Capricorn Orogen. The limits of cratonic basement and orogenic deformation beneath younger sedimentary basins are based on data from Wellman (1976, fig. 3) and Drummond (1981, fig. 10).

Ashburton and Blair Basins

The Ashburton and Blair Basins have been the subject of a major study by Thorne & Seymour (in press) (see also Thorne 1985, Thorne & Seymour 1986). Rocks deposited within the Ashburton Basin comprise the *ca* 2000 Ma old Wyloo Group (Table 1) that overlies rocks of the Hamersley Basin with both angular unconformity in the Hardy Syncline area (Trendall 1979, Seymour *et al.* 1988), and disconformity in the Turee Creek Syncline and Mount Maguire areas (Thorne *et al.* in press) (see Figs. 2 and 3). The Wyloo Group records a change from terrestrial and shallow-marine to 'deep-water' sedimentation. During this evolution the distribution of sedimentary facies was strongly influenced by local and regional tectonism concomitant with the transition from an active continental margin to a foreland basin (Thorne & Seymour in press).

The Blair Basin is occupied by the Capricorn Formation, which unconformably overlies folded Wyloo Group rocks. It represents fluvial, alluvial fan and subaqueous sedimentation in a relatively small fault bounded basin (Thorne & Seymour in press). The Mount Minnie Group is a sequence of conglomerate, sandstone, siltstone and mudstone slightly younger than the Capricorn Formation (Seymour *et al.* 1988).

Rocks of the Pilbara Craton and the Ashburton and Blair Basins are unconformably overlain by post-orogenic sedimentary rocks of the Middle Proterozoic Bresnahan Basin and the Middle to Late Proterozoic Bangemall Basin, as well as Phanerozoic rocks of the Carnarvon and Canning Basins.

CAPRICORN OROGEN STRUCTURES IN THE PILBARA CRATON MARGIN

Deformation in the Pilbara Craton margin was assigned originally to two fold periods, the Ophthalmian and the Rocklean, with both events thought to have taken place after deposition of the Wyloo Group (Halligan & Daniels 1964, MacLeod 1966). Subsequent recognition of a major unconformity between the Hamersley Basin succession and the Wyloo group (Trendall 1979) led Gee (1979) to recognize two separate fold belts. The oldest was the Ophthalmia Fold Belt, which affected Hamersley Basin rocks south of the Fortescue Valley. This was followed by the Ashburton Fold Belt coincident with the geographical extent of the Ashburton

Table 1. Stratigraphy of the northern margin of the Capricorn Orogen

Basin	Group	Formation
Blair Basin		Capricorn Formation
----------	unconformity	----------
Ashburton Basin (*ca* 2000 Ma)	Wyloo Group	Ashburton Formation
		June Hill Volcanics
		Duck Creek Dolomite
		Mount McGrath Formation
	--unconformity–disconformity--	
		Cheela Springs Basalt
		Beasley River Quartzite
----------	unconformity–disconformity	----------
Hamersley Basin (2750–2300 Ma)	Turee Creek Group	
	Hamersley Group	Boolgeeda Iron Formation
		Woongarra Volcanics
		Weeli Wolli Formation
		Brockman Iron Formation
		Mount McRae Shale
		Mount Sylvia Formation
		Wittenoom Dolomite
		Marra Mamba Iron Formation
	Fortescue Group	Jeerinah Formation
		Mount Jope Volcanics
		Hardey Sandstone
		Mount Roe Basalt
----------	unconformity	----------
	granite–greenstone terrain (>2750 Ma)	

Basin (Fig. 2). Tyler (in press) has identified numerous shear zones in the granite–greenstone terrain rocks of the Sylvania Inlier that can be directly related to folding in the Ophthalmia Fold Belt.

Capricorn Orogen structures in the Sylvania Inlier

The Sylvania Inlier has been described in detail by Tyler (in press). It can be divided into two parts (Fig. 4); a relatively undeformed western third, and an extensively deformed eastern two-thirds.

In the eastern two-thirds, numerous shear zones are present and granitoid is typically foliated. Rocks within shear zones often show a pronounced stretching lineation on foliation surfaces. Towards the southern margin of the inlier, the most intensely deformed rocks show complete recrystallization producing a schistose quartz–biotite–muscovite–chlorite ± kyanite rock.

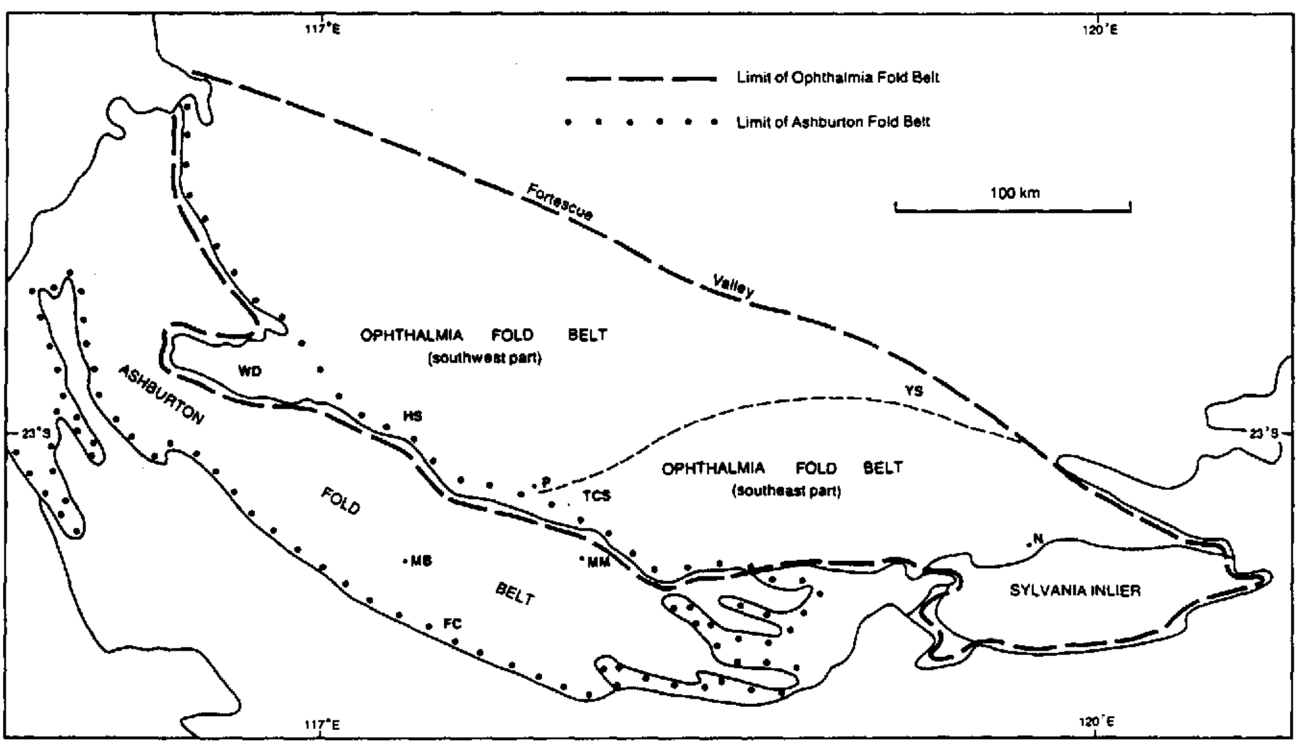

Fig. 2. Map showing the distribution of the Ophthalmia Fold Belt and Ashburton Fold Belt. Tectonic units as in Fig. 1. FC—Fords Creek, HS—Hardey Syncline, MB—Mount Blair, MM—Mount Maguire, N—Newman, P—Paraburdoo, TCS—Turee Creek Syncline, WD—Wyloo Dome, YS—Yandicoogina Syncline.

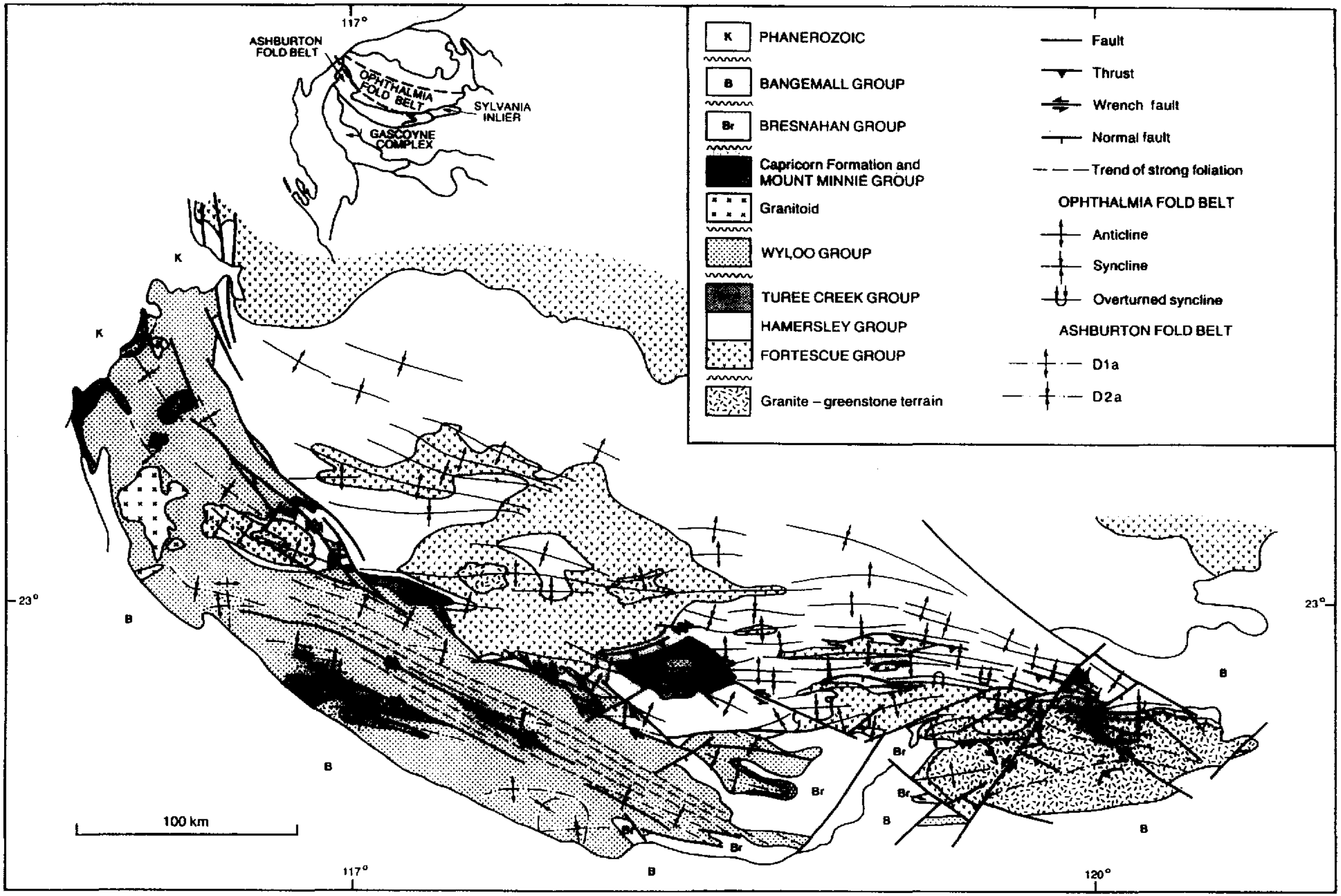

Fig. 3. Simplified geological map of the southern Pilbara showing the main structural features (folds, faults and cleavage–foliation trends).

Shear zones with mylonitic fabrics pass directly from granitoid rocks into greenstone belts and one such structure has been interpreted as a N-directed thrust (the Mindoona Bore Thrust; see Tyler & Thorne 1990, Tyler in press). The contact between the inlier and Fortescue Group (Figs. 3 and 4), has also been interpreted as a fault (Williams & Tyler 1989, Tyler in press). The fault is parallel to the shear zones and foliation in the inlier. Both the Mindoona Bore Thrust and the Painkiller Bore Fault die out to the east and at the eastern end of the inlier Hamersley Basin rocks lie unconformably on granitoid.

In the western part of the inlier only a few shear zones occur. A major E-trending fault, the Western Creek fault, offsets Fortescue Group rocks indicating an apparent sinistral movement of 10 km.

Ophthalmia Fold Belt

Two groups of folds can be identified within the Ophthalmia Fold Belt (Fig. 2). In the southwest the folds form broad-scale, open dome-and-basin structures having a mainly NW trend (Fig. 3; the central structural zone of MacLeod *et al.* 1963). In the southeast, folds have an easterly trend, are close to tight and have short wavelengths (the southern structural zone of MacLeod *et al.* 1963).

The principal domical structures in the southwest part of the fold belt contain outcrops of the underlying granite–greenstone terrain while basinal structures contain stratigraphic units up to, and including the Turee Creek Group. The limbs of the major structures dip at about 30° (MacLeod *et al.* 1963). Previous workers (e.g. Halligan & Daniels 1964, de la Hunty 1965, MacLeod 1966) suggested that the dome-and-basin structures resulted from the interference of 'Ophthalmian' folds trending NW and 'Rocklean' folds trending NE. However, small-scale folds of the 'Rocklean' fold period are not seen, and Gee (1979) regarded all folding as part of a single fold set with curvilinear axes. Tyler & Thorne (1990) have pointed out that regional-scale 'folds' shown on plate 2 of MacLeod (1966) as having NE trends, cross the axes of NW-trending smaller-scale structures. They interpreted these apparent 'folds' as plunge culminations and depressions in a single set of NW-trending, en échelon, non-cylindrical folds.

Beasley River Quartzite at the base of the Wyloo Group unconformably overlies folded Hamersley and Turee Creek Group rocks in the Hardey Syncline and Wyloo Dome areas (Trendall 1979, Seymour *et al.*

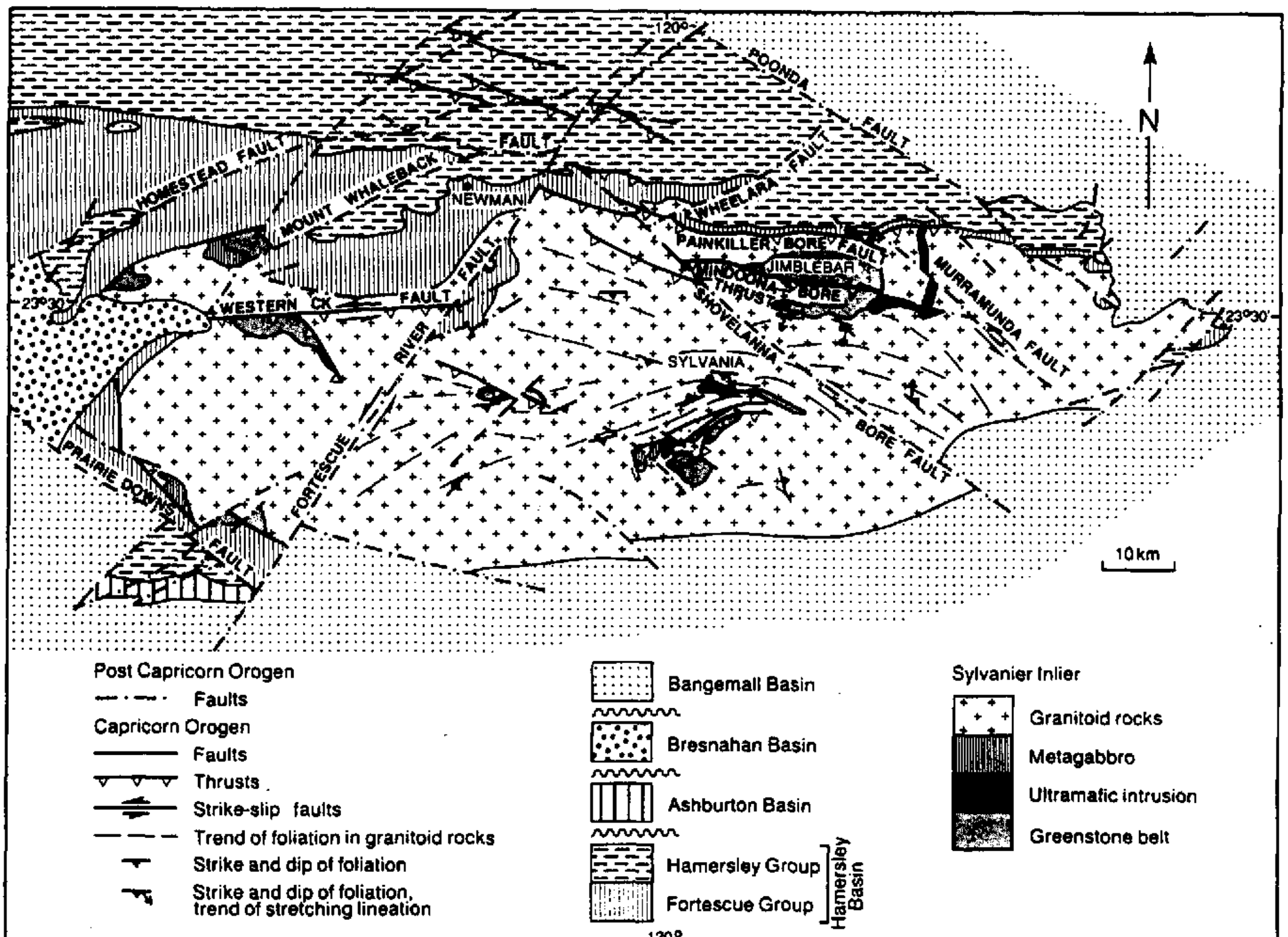

Fig. 4. Simplified geological map showing the main structural features within the Sylvania Inlier.

1988). Folding must therefore have taken place after deposition of the Turee Creek Group and before deposition of the Wyloo Group.

Two phases of deformation have been recognized in the southeast part of the Ophthalmia Fold Belt (Tyler 1986) and have been described in detail by Tyler (in press). An early event (D_{1c}), comprising small-scale layer-parallel folds and associated mylonites, and a later regional-scale fold event (D_{2c}).

D_{1c} structures are seen at only a few localities. Mylonitic rocks and small-scale layer-parallel folds (Fig. 5a) are restricted to particular stratigraphic horizons. They occur in the basal unit of the Fortescue Group, in the upper part of the Jeerinah Formation and basal part of the Marra Mamba Iron Formation, in the central BIF of the Woongarra Volcanics and at the base of the Boolgeeda Iron Formation. Although not abundant they have a wide geographical distribution and are recognized as far apart as the Newman area and the Turee Creek Syncline.

D_{2c} is the main deformation event in the southeastern part of the Ophthalmia Fold Belt and can be recognized from the eastern end of the Sylvania Inlier to the Turee Creek Syncline. The northern limit of folding is gradational, lying to the south of the Yandicoogina Syncline. The currently exposed southern limit is marked by contacts with the Bresnahan and Bangemall Basins.

Deformation is pervasive and is seen on all scales. It takes the form of gently inclined to upright (Figs. 5b & c), generally N-facing, open to isoclinal folds that often have a conjugate box-type form (Fig. 5c). Folds are typically of buckle-type with rounded hinges and are non-cylindrical and impersistent, dying out both laterally and vertically along their axial surfaces at all scales. Mesoscopic and macroscopic fold profiles vary from parallel to flattened parallel (Figs. 6a & b). Plunges are usually gentle to moderate either to the east or west.

Folding in the southeast Ophthalmia Fold Belt can be separated into two main areas on the basis of the intensity of the D_{2c} deformation: an area of high strain immediately to the north of the Sylvania Inlier; and an area of lower strain lying farther to the northwest and west (Fig. 3). In the area to the north of the Sylvania Inlier, folding passes northwards through asymmetrical overturned N-facing folds into reverse faulting and steeply inclined to upright folding (Fig. 7). An axial plane cleavage (S_{2c}) is well developed in rocks north of the Sylvania Inlier (e.g. Fig. 6b). To the northwest and west the cleavage is less prominant and is not present in the Paraburdoo area.

The occurrence of the lowermost Wyloo Group (Beasley River Quartzite and Cheela Springs Basalt) in the core of the Turee Creek Syncline provides a lower limit to the age of D_{2c} folding. Along the margin of the Hamersley Basin folds are cut by WNW-trending mafic dykes (Tyler 1990b). At Paraburdoo these dykes predate the formation of hematite ore in the Brockman Iron Formation (Morris 1980). Pebbles of hematite ore occur in the Mount McGrath Formation in the Paraburdoo area limiting the age of ore formation as pre-Mount McGrath Formation (Morris 1980). Folding must therefore have taken place after eruption of the Cheela Springs Basalt but before dyke intrusion and ore formation and before the deposition of the Mount McGrath Formation. The folding is therefore younger than that in the southwest part of the Ophthalmia Fold Belt.

Ashburton Fold Belt

The Ashburton Fold Belt (Fig. 2) formed as the result of post-Wyloo Group, pre-Bresnahan Group deformation (Thorne & Seymour in press). The Wyloo Group, Capricorn Formation, Mount Minnie Group and Mount Bruce Supergroup, were subject to open to isoclinal folding and normal, reverse and wrench faulting. Two generations of structure are recognized (D_{1a} and D_{2a}). D_{1a} structures post-date the Wyloo Group and pre-date the Capricorn Formation and Mount Minnie Group, as well as the granitoid intrusion (the Boolaloo Batholith, intruded into the Wyloo Group, see Fig. 3). D_{2a} structures post-date intrusion of the Boolaloo Batholith and deposition of the Capricorn Formation.

Deformation within the Ashburton Fold Belt is dominated by the D_{2a} event. Thorne & Seymour (in press) recognized three structural zones according to the varying intensity of deformation present. The first of these zones occurs along the boundary between the Hamersley and Ashburton Basins. It is dominated by large-scale open to tight folds and dextral wrench faults. It includes Hamersley Basin rocks and separation of Ashburton and Ophthalmia structures can be difficult with the later deformation tightening and re-orientating pre-existing folds and re-activating earlier-formed faults. Within the Paraburdoo orebody these earlier-formed faults are preserved as post-Ophthalmia Fold Belt, pre-ore low-angle normal faults (Morris 1985).

Faults trend WNW to NW and dextral offsets on fault systems of up to 11 km are recognized (Seymour *et al.* 1988). Southwest of Paraburdoo a major wrench fault system cross-cuts a set of E-trending en échelon D_{2a} folds. The fault system continues northwestward for 300 km, and for much of this distance forms the boundary between the Hamersley and Ashburton Basins. Subsidiary NW-trending synthetic, and N- to NE-trending antithetic faults (cf. Wilcox *et al.* 1973) are well developed.

The second zone occupies the central part of the fold belt and is characterized by relatively high strains with the development of tight to isoclinal, non-cylindrical folds with wavelengths ranging from 5 to 200 m. Folds are associated with a strong axial plane cleavage (S_{2a}) dipping steeply SW. Extensive dextral wrench faulting is present. Outliers of Capricorn Formation may be partially or completely fault-bounded producing distinctive rhombohedral outcrop geometry (Fig. 3).

The third zone has been subjected to lower strains and is distinguished from the central zone by the presence of large-scale D_{2a} folds having wavelengths of up to 15 km. Axial planes dip steeply NE or SW and folds are

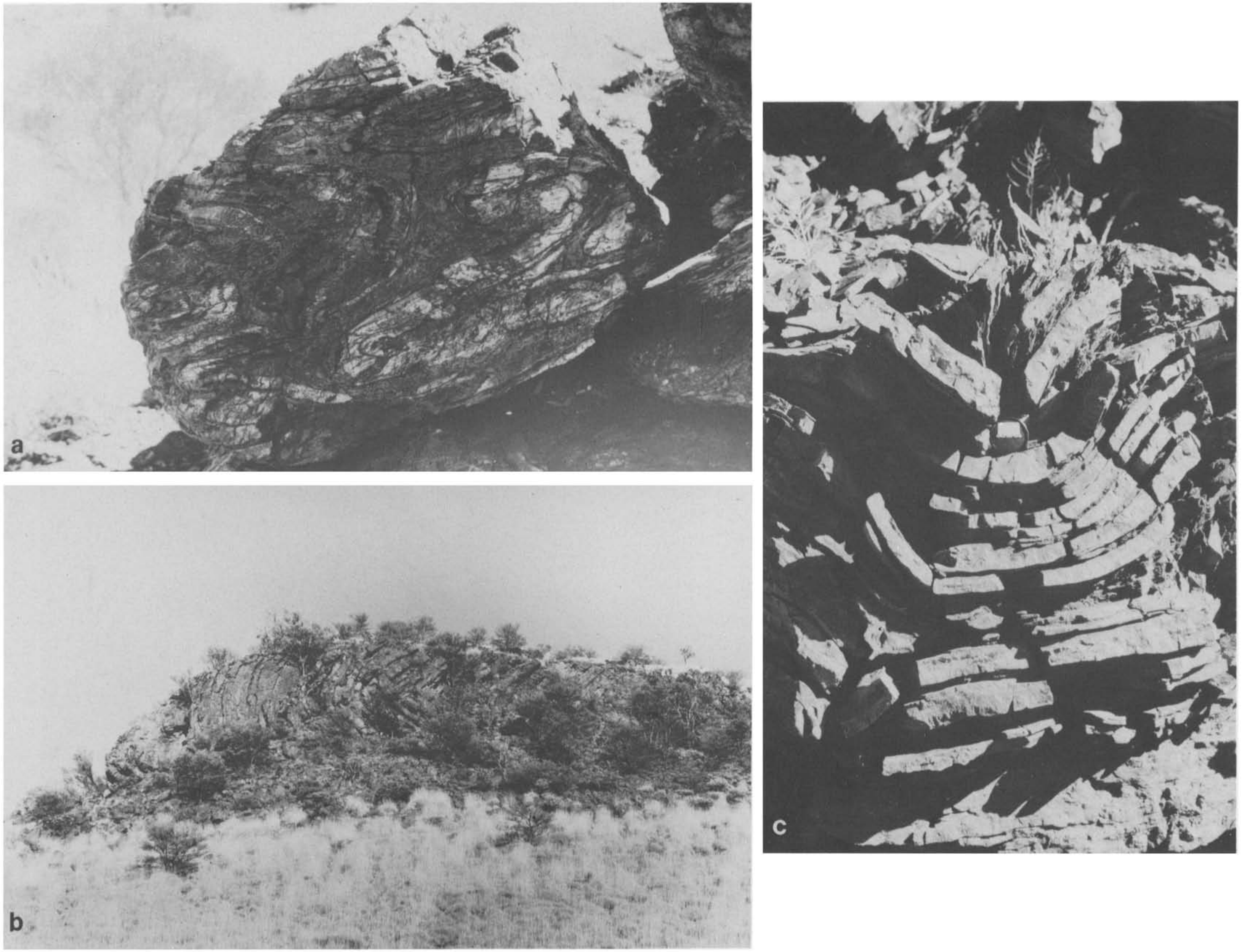

Fig. 5. (a) Tight to isoclinal D_{1c} folds in the upper Jeerinah Formation. The boulder is 1.5 m across. (b) Recumbent D_{2c} fold in Marra Mamba Iron Formation south of Newman. The cliff is 20 m high. (c) Upright conjugate D_{3c} fold in Weeli Wolli Formation BIF, north of the Sylvania Inlier. The lens cap is 5 cm in diameter.

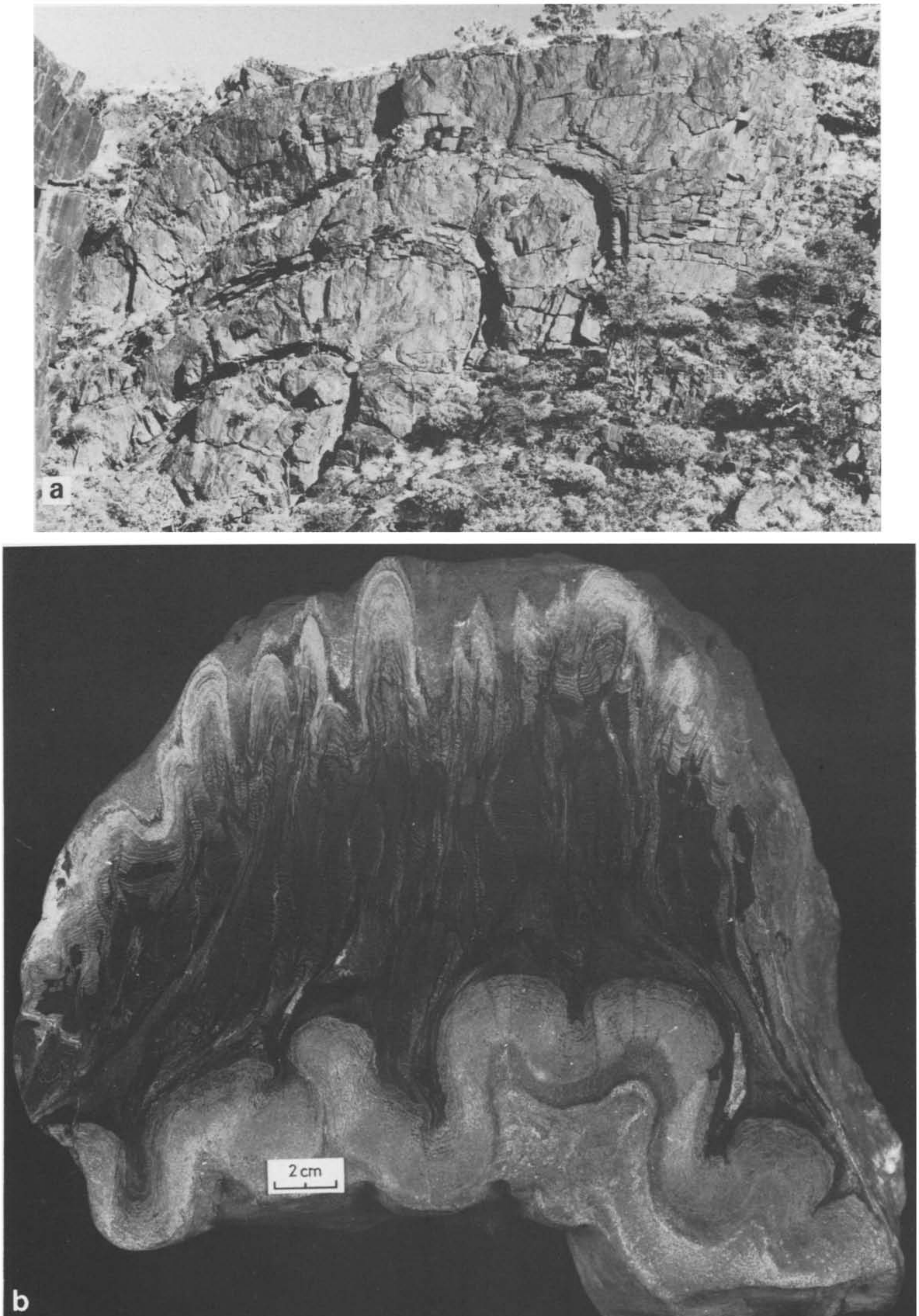

Fig. 6. (a) Large-scale overturned D_{2c} fold in Brockman Iron Formation, west of Newman. Note flattened parallel profile. The cliff is 30 m high. (b) Polished slab of folded Brockman Iron Formation, Mount Whaleback iron ore mine, Newman. Note parallel conjugate D_{2c} folding in lower part and compare with flattened parallel D_{2c} folds in upper part. A wavy anastomosing spaced cleavage (S_{2c}), picked out by seams of hematite, is well developed. GSWA sample 42229.

generally more open becoming tighter as the central high strain zone is approached. Downward-facing structures result from refolding of D_{1a} folds. D_{2a} structures are generally coaxial and may be coplanar with D_{1a} structures, producing 'hook' style refolding (Fig. 3). Dome-and-basin fold interference patterns are also present.

Evidence concerning the pre-D_{2a} orientation of D_{1a} structures is limited. At Mount Blair, unfolding of the Capricorn Formation unconformity indicates that S_{1a} dipped steeply SW while the D_{1a} folds faced NE. Beneath the Bangemall Group unconformity at Fords Creek there is evidence that large-scale, tight D_{1a} folds had moderate NE-dipping axial surfaces and faced SW (Thorne & Seymour in press).

METAMORPHISM

An axial plane cleavage is associated with D_{2c} folds in the southeast part of the Ophthalmia Fold Belt. Formation of this cleavage has taken place under low- and very low-grade metamorphic conditions. Smith *et al.* (1982) have interpreted low- and very low-grade metamorphic mineral assemblages throughout the Hamersley Basin as the product of burial metamorphism (M_h) that reached a peak at the end of Turee Creek Group times. The formation of the S_{2c} cleavage however, implies a later, Capricorn age metamorphic event (M_c) superimposed onto the earlier burial metamorphic event. Grades in the M_c event were similar to those present during M_h.

Mineral assemblages in mafic dykes and granitoid rocks in the Sylvania Inlier are indicative of conditions transitional between low- and medium-grade metamorphism (Tyler in press). In the western inlier a static low- to medium-grade recrystallization has been equated with the M_h event (Tyler in press).

In the eastern Sylvania Inlier metamorphism that can be related to the Capricorn Orogeny is recognized within shear zones. Chlorite–kyanite is stable in quartzofeldspathic schist derived from granitoid in shear zones in the southern part of the Sylvanier Inlier. Tyler (in press) has estimated a temperature of 550°C with pressure in excess of 500 MPa for this assemblage, equivalent to at least 18 km of overburden. Assemblages indicative of lower grades occur in shear zones towards the northern margin of the inlier (Tyler in press).

Capricorn age metamorphism has also affected Wyloo Group rocks in the Ashburton Basin (Thorne & Seymour in press). Grade ranges from the prehnite–pumpellyite facies adjacent to the Hamersley Basin to upper greenschist facies adjacent to the Bangemall Basin unconformity.

Very low-grade assemblages in Hamersley Basin rocks in the Paraburdoo and Hardey Syncline areas have been interpreted by Smith *et al.* (1982) as the result of Hamersley Basin burial metamorphism. An alternative explanation is that these assemblages represent the effects of a later burial under the Ashburton Basin.

STRUCTURAL SYNTHESIS

The Sylvania Inlier and southeastern Ophthalmia Fold Belt

Tyler (in press) has concluded that the Sylvania Inlier is not a simple domical structure analogous to other granite–greenstone terrain outcrops in the Hamersley Basin (e.g. Macleod 1966). Rather it is a composite structure.

The occurrence of extensive shear zones in the eastern inlier is indicative of large inhomogeneities of strain (Ramsay & Graham 1970) with simple shear as the principal deformation mechanism (cf. Rathbone *et al.* 1983). Strain variations are not consistent with an inhomogeneous pure shear (e.g. Coward 1984, fig. 10). Granitoid between the shear zones typically develops a penetrative foliation that is parallel to the shears. This is consistent with simple shear being superimposed on a homogeneous strain, the latter being responsible for the initial flattening.

Tyler (in press) interpreted deformation in both the Sylvania Inlier and the Ophthalmia Fold Belt as part of the same regional-scale N-directed thrust system. At least one shear zone in the inlier, the Mindoona Bore Thrust, can be interpreted as a N-directed thrust based on stratigraphic relationships where it cuts the Jimblebar greenstone belt (Tyler & Thorne 1990, Tyler in press). This is consistent with the N-directed fold and thrust-style deformation in the Ophthalmia Fold Belt to the north.

At the western end of the inlier, granite–greenstone terrain crops out within anticlinal fold cores (Fig. 3). Such a relationship rules out gravity slumping of the Hamersley Basin succession off a rising domical structure as proposed by Gee (1979), as this mechanism would require a decollement between cover and basement together with a zone of extension (the "tectonic gap" of Price 1971) that should balance the shortening at the front of a gravity-driven system. North-facing structures in the Hamersley Group to the south of the inlier (Tyler in press) are further evidence against a gravity slumping, as this model requires structures on the south side of the uplift to face south.

A feature of low-angle thrust systems is that they usually propagate towards their foreland so that the youngest detachment is the lowest (Boyer & Elliott 1982). This last detachment, usually referred to as the sole thrust, carries all the earlier thrust slices with it 'piggy-back' fashion. Where the sole thrust passes over a ramp in its footwall large-scale folding must take place in the hanging wall, deforming the earlier nappes and thrusts. This type of deformation can produce folding of the opposite vergence to the primary transport direction, particularly in early thrusts now high in the sequence (e.g. Price 1981, Boyer & Elliott 1982).

The western third of the inlier comprises little reworked, essentially autochthonous to parautochthonous granite–greenstone terrain. The eastern two-thirds of the inlier is characterized by ductile thrusting and is

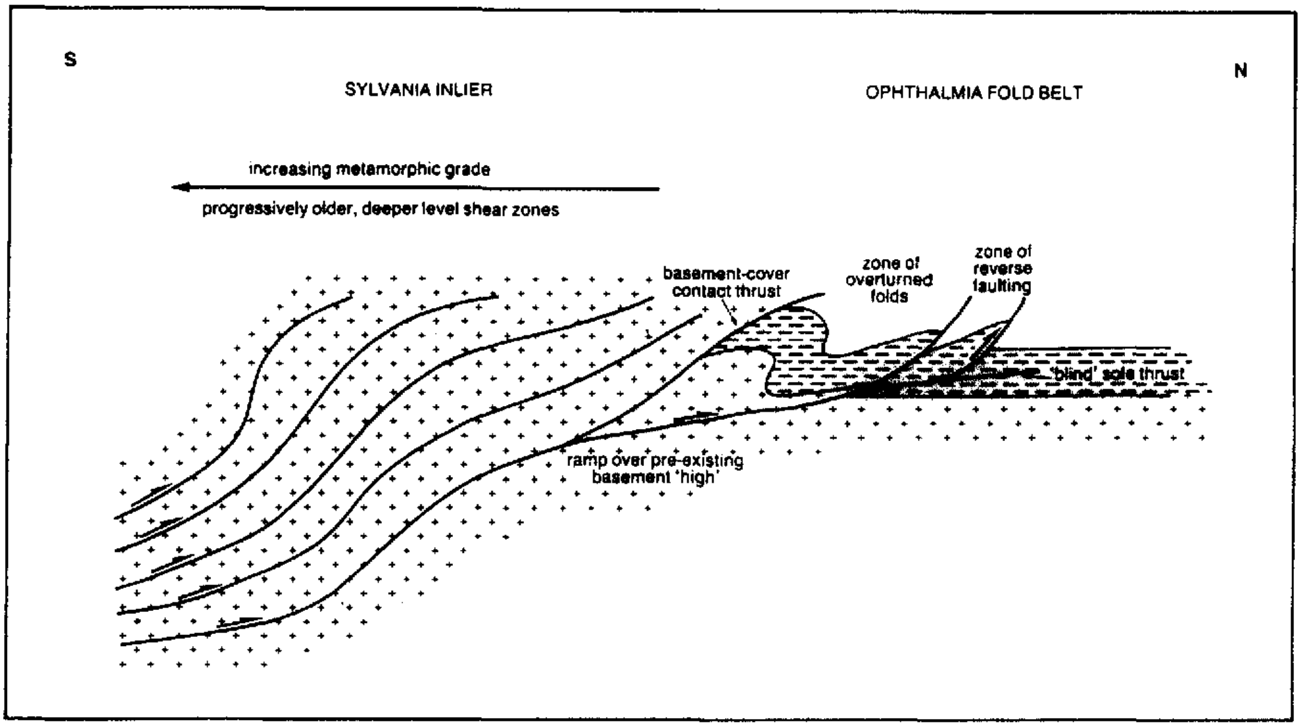

Fig. 7. Diagrammatic cross-section through the eastern part of the Sylvania Inlier and the southeast Ophthalmia Fold Belt.

interpreted as a stack of parautochthonous to allochthonous thrust sheets. Large-scale folding has produced N-dipping shear zones at the southern margin of the inlier and these are interpreted as the result of the passage of the thrust stack over a ramp in the footwall (Fig. 7). Ramping can be caused by irregularities in the basement–cover contact (Wiltschko & Eastman 1983) and such a feature is provided by the topographic high identified as being present in this area during Fortescue Group times (Horwitz & Smith 1978, Blake 1984, Tyler in press).

This interpretation is consistent with the apparent contrast in metamorphic grade between the inlier and the overlying Hamersley Basin succession (Tyler in press), and with the observed increase in metamorphic grade from north to south across the eastern part of the inlier. Such a pattern, together with the ductile nature of the shearing, is also consistent with the exposure of progressively deeper crustal levels (e.g. Rathbone *et al.* 1983).

Most of the faults exposed in the Ophthalmia Fold Belt north of the inlier are steep, dipping south at about 60°. In order to balance folding (cf. Dahlstrom 1969) a flat-lying sole thrust is required to occur beneath the fold belt (Fig. 7) with the steep faults forming part of an imbricate fan (cf. Boyer & Elliott 1982). Folding is present to the north of the fan, and its conjugate nature is consistent with a blind extension to the sole thrust (e.g. Thompson 1981). The sole thrust is considered to lie within the relatively incompetent Fortescue Group.

From the disposition of faults, shears and stretching lineations in the Sylvania Inlier and from the pattern of folding in the cover (Fig. 3) it is possible to reach some conclusions concerning the principal direction of thrust transport and the area in which thrust movements were concentrated. The most intense deformation in the inlier occurs towards the southern margin of its eastern part. To the east, both shears and foliation swing to the southeast. To the west, the trend is SW. Stretching lineations trend NNE throughout the inlier. In the western part of the inlier, the Western Creek Fault has a sinistral offset.

The most intense deformation in the Ophthalmia Fold Belt occurs north of the Sylvania Inlier. The pattern is consistent with the main locus of thrust movement being to the NNE and concentrated into the central part of the inlier where it was responsible for the greatest amounts of uplift (and the deepest exposure levels). To the east and west, the frontal thrusts are interpreted as passing into oblique ramps (cf. Hossack 1983) which may give relative dextral (eastern ramp) and sinistral (western ramp—the Western Creek Fault) offsets.

Origin of D_{1c} *structures*

The origin of D_{1c} structures is problematical. They have been recognized to the north of the frontal imbricate fan of the main thrust system. This makes them difficult to interpret as earlier formed thrusts deformed by later thrust movements. Although geographically widespread, any one occurrence appears to be limited in its extent. Deformation is restricted to bedding planes, and ramping, in which the deformation cuts up or down the stratigraphy, is not seen. Movement direction is unknown and they could represent either an extensional or a compressional feature. Their development as part of an early extensional fault system (cf. Gibbs 1984) associated with the formation of the Ashburton Basin is unlikely, however, as a direct link with faults on the basin margin is not seen.

At present, deformation is regarded as the result of gravity spreading (cf. Ramberg 1981) during the early stages of uplift associated with D_{2c} thrusting. A similar style of early, restricted, bedding-plane-controlled deformation has been described from the Southern Uplands of Scotland (Knipe & Needham 1986).

Influence of pre-existing structures at the craton margin

Rifting along the southern margin of the Pilbara Craton developed during Fortescue Group times (Blight 1985, Blake & Groves 1987). An earlier NNE oriented rifting event has also been identified by Blake (1984) (see also Blake & Groves 1987). Structures established during these rifting events have had a major influence on later sedimentation and the subsequent deformation of the craton margin.

The rhombohedral shape of many of the dome-and-basin structures in the southwest part of the Ophthalmia Fold Belt (Fig. 3) reflects control by a basement segmented along NW- and NNE-trending shears and faults. This control is the origin of the supposed 'Rocklean' cross-fold direction discussed earlier.

Prior to folding in the southeast Hamersley Basin, extensive uplift took place along the southern margin of the Pilbara Craton (up to 9 km according to Smith *et al.* 1982). Normal faulting is seen in the vicinity of the Wyloo Dome and may be responsible for localized fan-delta style sedimentation in the Beasley River Quartzite (Thorne & Seymour 1986). Faulting has a general WNW to NW trend, reflecting the influence of the earlier Fortescue Group rift.

Within the southeastern part of the Ophthalmia Fold Belt, regional plunge culminations can be seen exposing Fortescue Group rocks (Fig 3). A major plunge depression occurs in the Turee Creek Syncline area exposing lower Wyloo Group. This is interpreted as a reflection of the influence of basement block movement prior to D_{2c} thrusting. The blocks can be identified in the seismic refraction data of Drummond *et al.* (1981, section FDB). Any associated fold would have been very broad, and likely to be tightened up, rather than refolded, by the later thrust event.

Origin of Ashburton Fold Belt D_{1a} *structures*

The origin of the Ashburton Fold Belt D_{1a} folds is uncertain. At the southwest margin of the fold belt the D_{1a} folds are recumbent and as such may form part of a fold and thrust system. Although these folds face SW D_{1a} folds farther to the northeast face NE. Folding can be interpreted in terms of a NE-directed thrust system that ramped on the margin of the Pilbara Craton. Back-thrusting and SW-facing folds could then have developed above the ramp. The increase in metamorphic grade from northeast to southwest is consistent with crustal thickening due to thrusting and recumbent folding.

Dextral wrench faulting in the Ashburton Fold Belt

The D_{2a} deformation in the Ashburton Fold Belt can be interpreted as the result of major, regional-scale dextral wrench faulting parallel to the southern margin of the Pilbara craton (Tyler & Thorne 1990, Thorne & Seymour in press). Harris (1987) has suggested that thrusting in the southeast Ophthalmia Fold Belt has developed as part of this wrenching event. However, sedimentation patterns within the Ashburton Basin are inconsistent with the view that it represents a strike-slip basin. All direct evidence suggests the D_{2a} event is either synchronous with, or post-dates, deposition in the overlying Blair Basin. There is a clear time separation therefore between thrusting in the southeast Ophthalmia Fold Belt (D_{2c}), which occurred between the deposition of the Cheela Springs Basalt and the Mount McGrath Formation, and dextral wrenching.

Dextral wrench faulting appears to have been controlled by basement structures with the highest strain developing in the central part of the Ashburton Fold Belt, immediately southwest of the craton margin (cf. Drummond 1981). At the margin between the Hamersley and Ashburton Basins strike-slip movement is presumed to have taken place on pre-existing normal faults.

A re-orientation of Ophthalmia Fold Belt D_{2c} folds can be recognized across the Turee Creek Syncline (Fig. 3). This, together with the widespread occurrence of dextral wrench faulting, and with local NW-directed thrusting, is consistent with the Turee Creek Syncline having been subjected to clockwise, i.e. dextral, rotation during the D_{2a} event.

TECTONIC EVOLUTION

There has been much debate as to the mechanisms producing Proterozoic Orogeny. With the recognition of the importance of thrust tectonics and horizontal motions in the Proterozoic crust (e.g Coward 1984), attention has focused on the applicability of Phanerozoic-style plate tectonic models (for a review see Shackleton 1986). Some authors have applied the classical plate tectonic model (referred to as B-subduction or the Wilson Cycle) in its entirety (e.g. Burke *et al.* 1976, Hoffman 1980, Windley 1981, 1983). Others, noting the apparent absence of criteria which would imply B-subduction such as ophiolites, paired metamorphic belts and island arcs, have proposed a modified plate tectonic model of intracratonic orogeny (referred to as A-subduction e.g. Kroner 1981, 1983, Etheridge *et al.* 1987) driven by crust–mantle delamination.

Previous models for the evolution of the Capricorn Orogen (MacLeod 1966, Daniels 1975, Horwitz & Smith 1978, Gee 1979, Williams 1986) have generally envisaged the development and tectonic evolution of a geo-

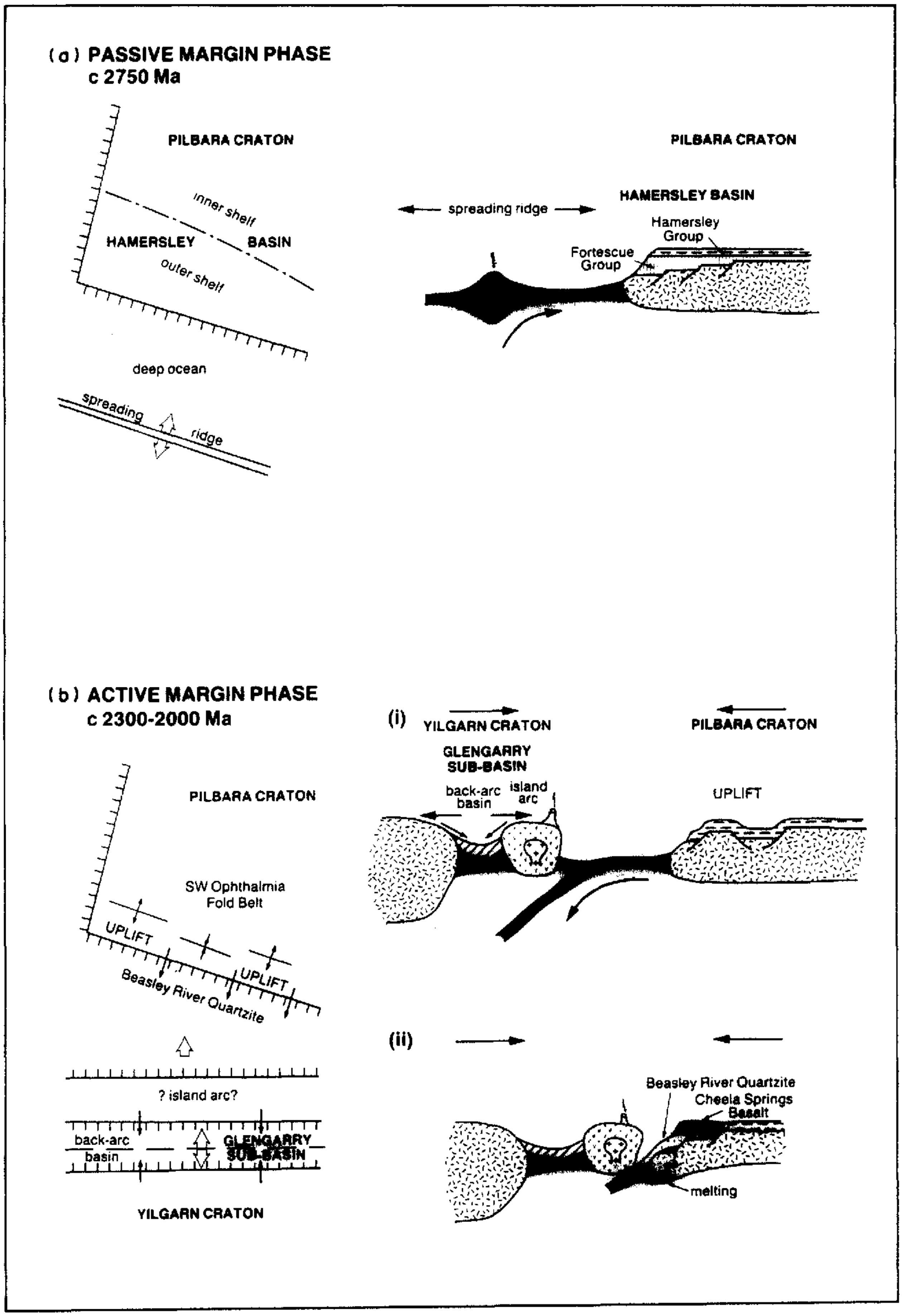
(a) PASSIVE MARGIN PHASE
c 2750 Ma
PILBARA CRATON
inner shelf
HAMERSLEY
BASIN
outer shelf
deep ocean
spreading
ridge
PILBARA CRATON
HAMERSLEY BASIN
spreading ridge
Hamersley
Group
Fortescue
Group
(b) ACTIVE MARGIN PHASE
c 2300-2000 Ma
PILBARA CRATON
SW Ophthalmia
Fold Belt
UPLIFT
UPLIFT
Beasley River Quartzite
? island arc?
back-arc
basin
GLENGARRY
SUB-BASIN
YILGARN CRATON
(i)
YILGARN CRATON
PILBARA CRATON
GLENGARRY
SUB-BASIN
back-arc
basin
island
arc
UPLIFT
(ii)
Beasley River Quartzite
Cheela Springs
Basalt
melting

Fig. 8. (a) and (b).

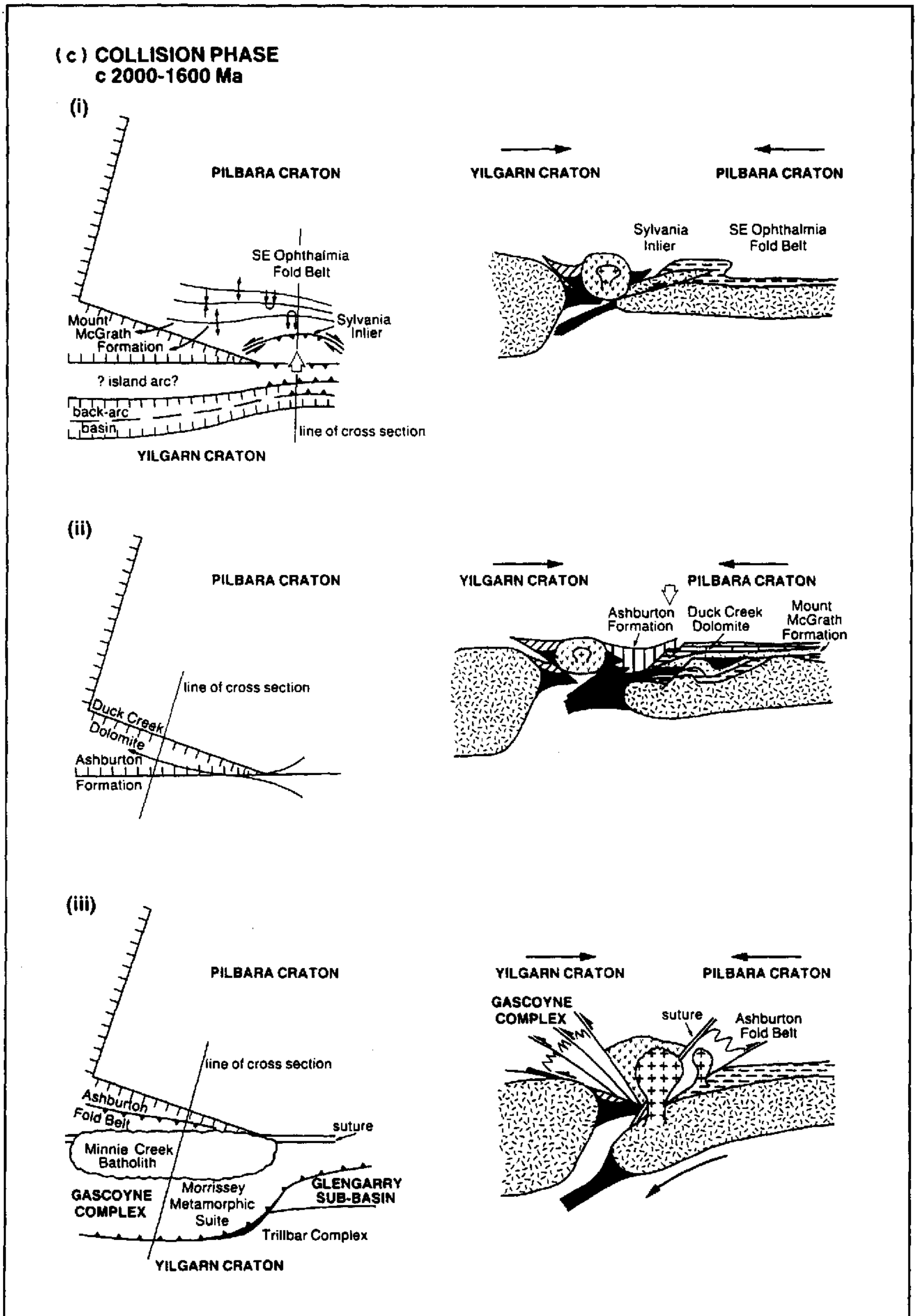

Fig. 8. Series of cartoons illustrating the development of the Capricorn Orogen as part of a B-subduction-style collision (after Tyler in press). (a) Passive margin phase. (b) Active margin phase: (i) Post-Turee Creek Group; (ii) Beasley River Quartzite and Cheela Springs Basalt. (c) Collision phase: (i) development of the southeast Ophthalmia Fold Belt; (ii) Duck Creek Dolomite–Ashburton Formation; (iii) Suturing–Ashburton Fold Belt D_{1a} and granitoid intrusion.

syncline within an intracratonic setting. In each of the models it is assumed the Pilbara and Yilgarn Cratons initially formed part of a single continuous craton.

Thorne & Seymour (in press) and Tyler (in press) have summarized evidence that illustrates the different geological histories of the northern margin of the Yilgarn Craton and the southern margin of the Pilbara Craton prior to the initiation of the Capricorn Orogeny. There are well documented differences in the evolution of the granite–greenstone terrains (e.g. Gee 1975, 1979; compare Gee *et al.* 1981 and Watkins 1990, with Hickman 1984; compare also geochronological data in Trendall 1983 and Blake & McNaughton 1984, with Fletcher *et al.* 1984 and McNaughton & Dahl 1987), crustal structure (Drummond 1981, Drummond *et al.* 1981), age of cratonization (Blake & Groves 1987), and in the nature of the Lower Proterozoic sedimentary sequences on each margin (compare Gee 1987, 1990, with Thorne 1990, Thorne & Seymour in press). Palaeomagnetic evidence tying the two cratons together before the orogeny is, at best, equivocal (McWilliams 1981, Idnurm & Giddings 1988).

Horwitz & Smith (1978) cited the occurrence of E-trending and N-trending dyke swarms in both cratons as evidence for a common stress system prior to initiation of the Hamersley Basin. Tyler (1990b) has pointed out that considerable age differences exist between the swarms. Those in the Pilbara either pre-date, or developed synchronously with, the Fortescue Group (i.e. *ca* 2750 Ma ago) while the Widgiemooltha dyke swarm in the Yilgarn, which is the major E-trending suite, has been dated at *ca* 2400 Ma (Fletcher *et al.* 1987). Horwitz & Smith (1987) also suggested that the Coobina ultramafic intrusion in the Sylvania Inlier shows both trends. However, the intrusion does not form part of any dyke swarm being intruded into, and deformed with, a greenstone belt, and intruded by Archaean granitoid (Williams & Tyler 1989, Tyler in press).

Sm–Nd and Sr isotopic data summarized by Libby *et al.* (1986), indicate that most of the igneous rocks involved in the orogeny were derived directly or indirectly from mantle-like reservoirs between 2400 and 1600 Ma ago. In the southern part of the orogen, older (>3000 Ma) Rb–Sr whole-rock and Sm–Nd model ages suggest that the Gascoyne Complex developed on Yilgarn basement. To the northwest, a decrease in the Sm–Nd model ages indicates either gradual migration of crust forming processes or progressive northward loss of contaminating Archaean crustal material. This suggests that Archaean basement may not have been continuous throughout the orogenic belt.

No unequivocal evidence is currently available that *requires* the Pilbara and Yilgarn Cratons to have formed part of a single, continuous craton prior to the development of the Capricorn Orogen. There is also, therefore, no *requirement* for intracratonic, i.e. A-subduction-style, orogeny. Any model for the tectonic evolution of the orogen *must* account for the differences that have been recognized between the Pilbara and Yilgarn Cratons. In an attempt to overcome this difficulty in an intracratonic model, Harris (1987) has suggested that suturing between the cratons occurred prior to the rifting phase and that evidence for this suture was masked by the later development of the orogen. The two cratons would have been juxtaposed by transcurrent movements along the line of the suture. No evidence has been presented to support such a model.

Myers (1990), in a re-assessment of the Gascoyne Complex, has attributed extensive shear zones and mylonite formation at the southern margin of the orogen to the development of a foreland thrust belt. This has formed mainly in rocks of the Early Archaean Narryer Gneiss Complex. They have been overridden by a SW-transported thrust sheet of deformed and metamorphosed gabbro and ultramafic rocks forming the Trillbar Complex. This was interpreted as possibly representing the lower part of an ophiolite sequence obducted onto the foreland (Myers 1989).

A feature of many Proterozoic orogenic belts is the occurrence of metamorphic mineral assemblages consistent with high-temperature and low- to medium-pressure metamorphism. *P–T*–time paths are interpreted in terms of slow isobaric cooling (e.g. Hobbs *et al.* 1984), in contrast with the rapid isothermal uplifts often associated with B-subduction (England & Thompson 1984). Baker *et al.* (1987), working on rocks from the southern foreland of the Capricorn Orogen, have identified corona textures between kyanite, garnet and gedrite in medium- to high-grade supracrustal rocks that are consistent with substantial uplift (10 km or more) accompanied by relatively little cooling. This was taken as evidence consistent with the development of the Capricorn Orogen in a B-subduction-style collision zone (Muhling 1988).

The Narracoota Volcanics of the Glengarry Group were deposited in the Glengarry Sub-basin on the northern margin of the Yilgarn Craton. They consist mainly of tholeiitic basalts with MORB affinities. However, mafic volcanic rocks whose chemistry resembles some boninites found associated with island arcs of the western Pacific, are present (Hynes & Gee 1986).

The above evidence suggests that at least some of the criteria are present necessary to establish the operation of B-subduction-style collision tectonics (cf. Shackleton 1986). The structural evolution of the northern margin of the orogen will therefore be discussed in terms of a continent–continent collision model (see also Thorne & Seymour in press, Tyler in press).

The earliest phase in the development of the orogen begins *ca* 2750 Ma ago with the initiation of the Hamersley Basin on the Pilbara Craton in an intracratonic rift (Blake & Groves 1987). The northern margin of the rift evolved into a passive continental margin adjacent to a deep ocean basin as crustal separation took place (Fig. 8a). The fate of the southern margin of the rift is unknown. There is no sequence of rocks preserved on the northern margin of the Yilgarn Craton or within the Gascoyne Complex comparable, either in age or stratigraphic sequence, with those deposited in the Hamersley Basin. A direct correlation of the Hamersley Basin

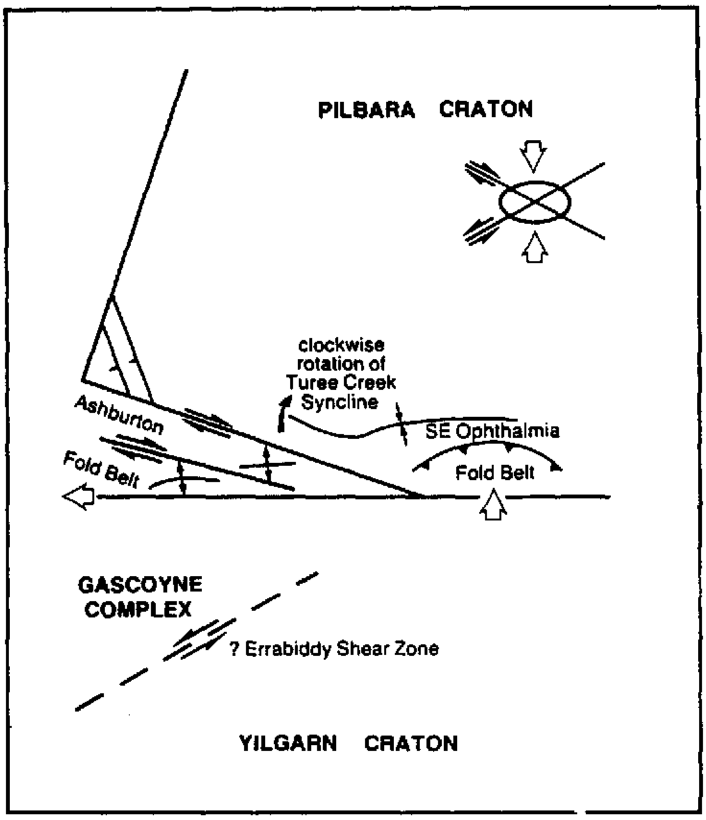

Fig. 9. Relationship of Ashburton Fold Belt D_{2a} wrench faulting to the overall structure of the Capricorn Orogen developed as part of an oblique collision.

succession with similar Late Archaean to Early Proterozoic sequences elsewhere in the world (e.g. the Ventersdorp Supergroup and Ghaap Group on the Kaapvaal Craton of southern Africa; Hickman & Harrison 1986, Grobler *et al.* 1989) does not provide a ready answer to the problem as younger rift events could be invoked as being responsible for the separation of these fragments of early crust. The crustal fragment carrying the southern side of the rift remains unrecognized.

The early stages of ocean-basin closure were marked by an active continental margin phase with the re-establishment of the supply of terrigenous sediment to the Hamersley Basin in Turee Creek Group times (*ca* 2300 Ma ago). Considerable uplift occurred along the craton margin, with large-scale folding in the southwest Hamersley Basin (the southwest part of the Ophthalmia Fold Belt) controlled by the movement of basement blocks (Fig. 8bi).

The stratigraphy of the Ashburton Basin has been interpreted by Thorne & Seymour (in press) as the evolution of the southern margin of the Pilbara Craton from an active margin to a foreland basin. The lower Wyloo Group was deposited during the final stages of ocean basin closure with extension along the southwestern part of the margin the result of arching of the crust (a flexural bulge) as the adjacent ocean floor was loaded by the approaching Yilgarn continent. Generation of mafic magma erupted as the Cheela Springs Basalt may be related to this flexuring (e.g. Baer 1981) (Fig. 8bii).

Collision between the two cratons occurred following deposition of the Cheela Springs Basalt. It was oblique, occurring first in the southeast with the establishment of a foreland fold and thrust belt (southeast Ophthalmia Fold Belt, Fig. 8ci). As a part of this belt, the Sylvania Inlier represents parautochthonous to allochthonous Archaean Pilbara basement collected from the edge of the craton and thrust back into the Hamersley Basin. Thrust transport was normal to the craton margin suggesting that movement of the Yilgarn Craton relative to the Pilbara was to the NNE.

Collision migrated to the northwest and migration of the associated flexural bulge produced further post-folding uplift and extension (Thorne & Seymour in press). A suite of WNW-trending mafic dykes was intruded at this time and was followed by subaerial exposure and formation of iron ore at Paraburdoo. Uplift supplied sediment to the Mount McGrath Formation.

The end of Mount McGrath Formation deposition was marked by 'deep-water' conditions. Continued loading produced downwarping and drowning of the craton margin cutting off the supply of terrigenous sediment. The Duck Creek Dolomite was deposited initially in relatively deep water that shallowed during subsequent progradation of the carbonate shelf (Thorne & Seymour in press). The end of Duck Creek Dolomite deposition was marked by oversteepening and/or normal faulting of the shelf slope and collapse of the carbonate shelf (Thorne & Seymour in press).

With collapse of the carbonate shelf, a deep-water basin was established parallel to the craton margin (Fig. 8cii). Material eroded from the uplifted granite–greenstone terrain of the Sylvania Inlier was shed into the basin's eastern end and was transported westwards by means of an elongate submarine fan that extended as far as the Wyloo Dome (Thorne & Seymour in press).

This submarine fan system is preserved as the Ashburton Formation.

Continent–continent collision now took place along the southern margin of the Ashburton Basin, and granitic detritus from the northern Yilgarn was incorporated into a second submarine fan system that prograded north-northeastwards into the basin, over the easterly-derived fan (Thorne & Seymour in press). A third fan complex is also recognized in the the northwest Ashburton Basin. Here sediment was derived from local uplift in the western Hamersley Basin.

Further cratonic convergence produced recumbent folding in the southern Ashburton Basin (D_{1a}, Fig. 8cii). Tectonic thickening of the sequence by folding and thrusting is indicated by the increase of metamorphic grade from northeast to southwest across the basin.

High-grade metamorphism in the Gascoyne Complex was followed by the intrusion of granitoid batholiths throughout the orogen (Williams 1986, Muhling 1988). The Boolaloo Batholith, which gives a Rb–Sr date of *ca* 1680 Ma (cf. Libby *et al.* 1986), forms part of this suite of granitoids.

The last stage of collision involved large-scale dextral wrench faulting parallel to the margin of the Pilbara Craton. This was responsible for the D_{2a} deformation in the Ashburton Fold Belt and for the clockwise rotation of the Turee Creek Syncline. The Capricorn Formation was probably deposited in a strike-slip basin formed during this event. The dominance of wrench faulting, together with normal faulting at the southwest corner of the Pilbara Craton, can be explained in terms of the progressive development of an oblique collision (Fig. 9). The initial collision in the southeast Pilbara was NNE-directed, normal to the craton margin. The northern margin of the Yilgarn trended more easterly (see Wellman 1976, fig. 3, Drummond 1981, fig. 10). Progressive closure of the oblique margins produced relative dextral movement on the Pilbara margin. Corresponding sinistral movements on faults and shear zones in the northern Yilgarn have been reported by Williams (1986). A model for this style of lateral extrusion of material caught between two opposing plates is provided by the collision between India and Asia (Tapponnier *et al.* 1986).

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