

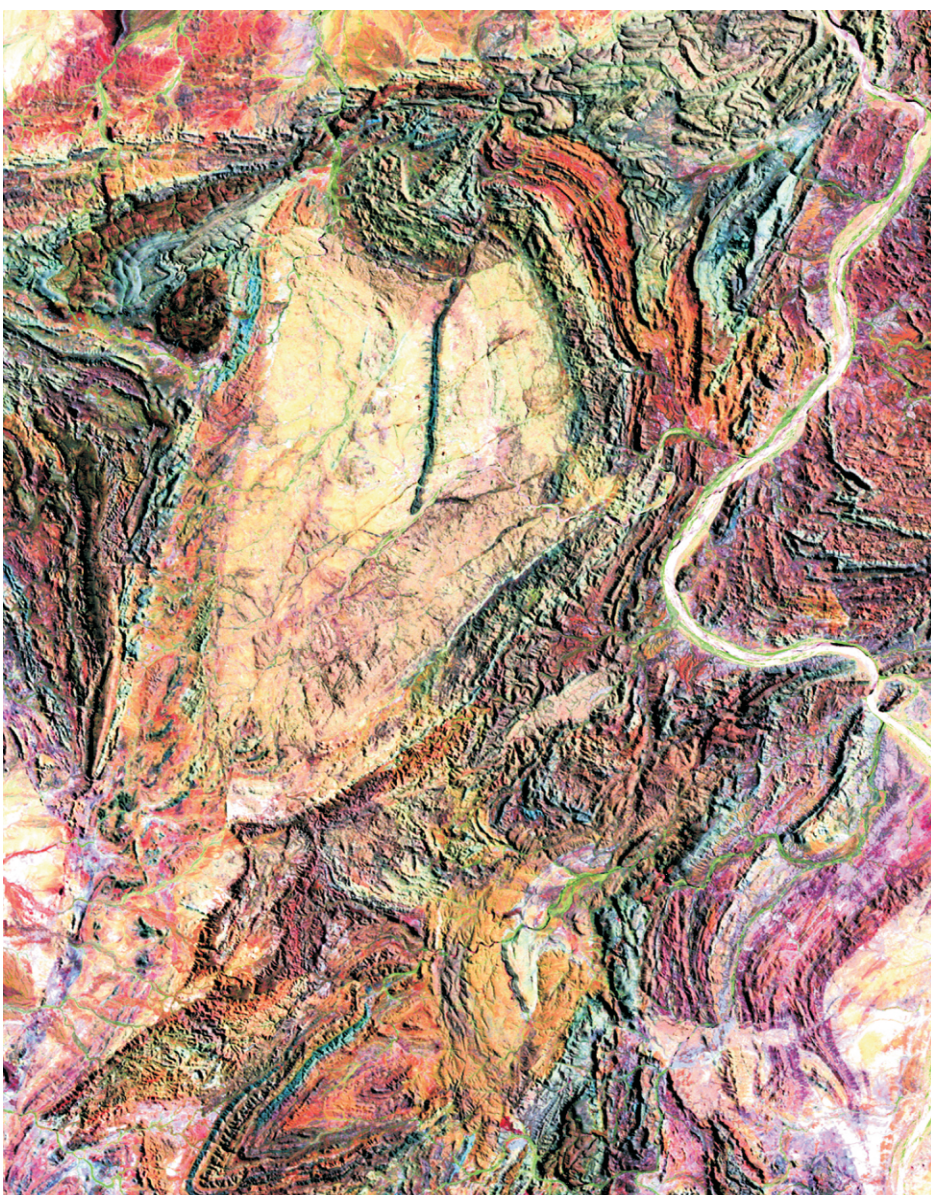
**EXPLANATORY  
NOTES**



# **GEOLOGY OF THE NORTH SHAW 1:100 000 SHEET**

**by M. J. Van Kranendonk**

**1:100 000 GEOLOGICAL SERIES**



**GEOLOGICAL SURVEY OF WESTERN AUSTRALIA**

**DEPARTMENT OF MINERALS AND ENERGY**



GEOLOGICAL SURVEY OF WESTERN AUSTRALIA

# **GEOLOGY OF THE NORTH SHAW 1:100 000 SHEET**

by  
M. J. Van Kranendonk

Perth 2000

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Cover photograph:

Landsat Thematic Mapper (TM) image of part (33 x 45 km) of the NORTH SHAW 1:100 000 sheet centred on the Strelley Granite. The curving white line to the right is the Shaw River.

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# Geology of the North Shaw 1:100 000 sheet

by

M. J. Van Kranendonk

## Abstract

The NORTH SHAW 1:100 000 sheet lies in the central-eastern part of the well-exposed North Pilbara Terrain of the Pilbara Craton. Apart from sparse Cainozoic cover, the sheet area is entirely underlain by Archaean rocks, including volcanic and sedimentary rocks of the Pilbara Supergroup, a variety of granitoid rocks that outcrop in parts of three domal complexes and three discrete plutons, minor intrusions, and outliers of the Mount Bruce Supergroup (Fortescue Group).

The Pilbara Supergroup is subdivided into five groups. The three lowermost groups are dominantly volcanic and include the c. 3515 Ma Coonterunah Group, the c. 3460 Ma Warrawoona Group, and the c. 3240 Ma Sulphur Springs Group. Disconformably overlying the Sulphur Springs Group are sedimentary and volcanic rocks of the undated Gorge Creek Group. These rocks are unconformably overlain by coarse clastic rocks of the c. 2940 Ma De Grey Group (Lalla Rookh Sandstone). Shelf-type supracrustal rocks of the unassigned Golden Cockatoo Formation are present as infolded remnants in the Yule Granitoid Complex, separated from the basal parts of the Sulphur Springs Group by the Numerous Scrapes Deformation Zone. Unconformably overlying rocks of the Fortescue Group ( $\leq$  c. 2772 Ma) are preserved in the southwestern part of the Marble Bar Outlier and in the Olympic Pool and Antarctic Outliers.

The Carlindi, Yule, and Shaw Granitoid Complexes are multicomponent structural domes with intrusive, sheared intrusive, or tectonic contacts with surrounding greenstones. The oldest phases are tonalite–trondhjemite–granodiorite–granite (TTG) in composition, whereas younger components are monzogranites. Dated components range in age from c. 3499 to 2936 Ma. Minor granitoid intrusions include the c. 3458 Ma North Pole Monzogranite — a synvolcanic laccolith to the Panorama Formation of the Warrawoona Group, the c. 3238 Ma Strelley Granite — a synvolcanic laccolith to the Kangaroo Caves Formation of the Sulphur Springs Group, and the c. 2936 Ma Keep It Dark Monzogranite, which cuts steeply dipping, amphibolite-facies greenstones of the North Shaw Belt and was emplaced during regional D<sub>3</sub> deformation.

Metamorphic grade in the greenstones decreases away from granitoid complexes, from a maximum of middle to lower amphibolite facies (hornblende–plagioclase, actinolite–garnet) adjacent to granitoid contacts, through widespread lower greenschist facies, to prehnite–pumpellyite facies. Relicts of low- to medium-pressure, granulite-facies assemblages (two pyroxenes, no garnet) are rarely preserved in amphibolite enclaves in the Shaw Granitoid Complex.

Structural data indicate five periods of deformation. D<sub>1</sub> occurred during accumulation of the Warrawoona Group and synvolcanic intrusion of TTG sills, between c. 3490 and 3400 Ma. The earliest components of D<sub>1</sub> deformation are manifest as synvolcanic growth faults, grabens, and ?syndepositional folds in the Dresser Formation. D<sub>1</sub> deformation also caused tilting, folding, and faulting of the Coonterunah Group prior to deposition of the unconformably overlying Strelley Pool Chert. During this time, c. 3460 Ma TTGs were locally transformed into isoclinally folded migmatitic gneisses.

D<sub>2</sub> structures (c. 3358–3235 Ma) include synvolcanic growth faults and synsedimentary slump folds in the Kangaroo Caves Formation, generated by emplacement and inflation of the Strelley Granite. In the Yule Granitoid Complex, a penetrative foliation developed during upright, non-cylindrical folding and synkinematic granitoid intrusion. Development of the Pilgangoora Basin to the north occurred at this time through horst and graben faulting. Elsewhere, largely cryptic D<sub>2</sub> structures include an increment of the regional foliation and tight folds generated under amphibolite-facies conditions both marginal and internal to the Shaw Granitoid Complex.

D<sub>3</sub> structures are the most prominent across NORTH SHAW, particularly within the Lalla Rookh – Western Shaw Structural Corridor, a 20 km-wide, fault-bounded zone of tightly folded and faulted rocks resulting from northwest–southeast oriented compression at c. 2936 Ma. This deformation led to the Strelley Granite and associated supracrustal rocks of the Soanesville Belt being pushed west,

rotated onto their side, and then translated to the north along the concave eastern margin of the Yule Granitoid Complex. Northerly translation was accompanied by deposition of the De Grey Group, and ended with collision against the Carlindi Granitoid Complex.

D<sub>4</sub> structures include faults and open to close folds in the Mount Roe Basalt of the Fortescue Group and are locally unconformably overlain by the Hardey and Kylenea Formations. D<sub>5</sub> structures include folds and faults that affect these stratigraphically younger formations. Faults are steep, curvilinear structures that are parallel to the contacts of granitoid domes, and have greenstone-side-down normal displacement.

The presence of regional unconformities between all groups requires an essentially autochthonous evolution for supracrustal rocks of NORTH SHAW. The geometry and kinematics of faults and folds associated with D<sub>1</sub>, D<sub>2</sub>, D<sub>4</sub>, and D<sub>5</sub> deformation events are consistent with the rise of buoyant granitoid rocks and concomitant sinking of greenstones during punctuated crustal overturn. D<sub>1</sub>–D<sub>2</sub> and D<sub>4</sub>–D<sub>5</sub> pulses of granitoid doming were separated by transpressional D<sub>3</sub> deformation at c. 2936 Ma. The effect of D<sub>3</sub> deformation decreases in intensity from west to east across the Shaw Granitoid Complex, and may have been caused by compressional tectonic events in the west Pilbara.

NORTH SHAW is extensively mineralized. Historically, the most prominent mining activity was for epigenetic gold at the Lalla Rookh, North Pole, and North Shaw Mining Centres, and until recently at the Normay mine. Current interest has sparked renewed activity in the North Shaw and Lalla Rookh Mining Centres. Palaeoplacer gold was extracted from the Keep It Dark (De Grey Group) and Virgin Creek (Fortescue Group) mines. Syngenetic barite was mined at the Dresser Mining Centre in the North Pole Dome, an area that still retains large reserves of barite. Chrysotile asbestos was removed from the Strelley prospect. Nickel-platinum group element mineralization formed in differentiated ultramafic sills of the Dalton Suite in the Soanesville Belt. The Breens felsic porphyry intrusion in the North Pole Dome hosts Cu, Ag, and Bi mineralization.

Numerous prospects of volcanogenic massive sulfides (Cu–Zn) are at the top of the Sulphur Springs Group near the Strelley Granite. The largest of these are the Sulphur Springs, Kangaroo Caves, and Bernts prospects, which contain a combined indicated and inferred resource estimated at about 5 Mt at 10% Zn and 0.6% Cu, and 2.5 Mt at 1.1% Zn and 4% Cu.

**KEYWORDS:** Pilbara Craton, greenstone, granitoid complexes, Pilbara Supergroup, Fortescue Group, revised stratigraphy, geochemistry, metamorphism, structural evolution, gold, massive sulfide deposits

## Introduction

The NORTH SHAW\* 1:100 000 sheet (SF 50-8, 2755) is bound by latitudes 21°00'S and 21°30'S and longitudes 119°00'E and 119°30'E. It lies within the central-northern part of the MARBLE BAR 1:250 000 sheet in the East Pilbara Mineral Field (Fig. 1). NORTH SHAW is underlain by folded and metamorphosed, early to late Archaean (c. 3.51 – 2.94 Ga) volcanic, sedimentary, and granitoid rocks of the East Pilbara Granite–Greenstone Terrane of the North Pilbara Terrain (Hickman et al., in prep.; formerly called the 'north Pilbara granite–greenstone terrane' by Griffin, 1990), and by outliers of little-deformed and weakly metamorphosed volcanic and sedimentary rocks of the unconformably overlying, late Archaean (c. 2.7 Ga) Fortescue Group (Mount Bruce Supergroup). The older supracrustal rocks have been subdivided into five unconformity-, disconformity-, or intrusion-bound groups collectively referred to as the Pilbara Supergroup, and the associated synvolcanic c. 3.45 Ga North Pole Monzogranite and c. 3.24 Ga Strelley Granite. Coeval granitoid rocks formed within parts of the domal, multicomponent (c. 3.49 – 2.94 Ga) Shaw, Yule, and Carlindi Granitoid Complexes, and the discordant, late-tectonic (c. 2.94 Ga) Keep It Dark Monzogranite.

These Explanatory Notes describe the regional geological setting, lithostratigraphic and lithostructural map elements, Archaean rock types, Cainozoic deposits, structures, metamorphism, geochronology, mineralization, and a brief geological evolution of the rocks on NORTH SHAW. A more detailed account of the geological evolution is presented in Van Kranendonk and Collins (1998) and Van Kranendonk (in prep.b), the latter of which also presents some detailed geochemistry.

## Access and land use

NORTH SHAW is relatively isolated and lies in the middle of a triangle of three main roads, including the unsealed BHP Iron Ore road to the west, the discontinuously sealed Marble Bar Road in the northeast, and the unsealed Woodstock – Hillside – Marble Bar track, which traverses the map area in the southwest. Much of the map area is only accessible by four-wheel drive vehicle on unmaintained tracks or by driving across country, while some middle parts of NORTH SHAW can only be visited by walking or helicopter. Much of NORTH SHAW has become accessible only recently because of tracks created during mineral exploration by Sipa Resources Pty Ltd around the Strelley Granite. Some of the tracks have now been abandoned and are quickly deteriorating due to erosion.

The main access points to NORTH SHAW are shown on Figure 1. These include a track that turns off from the

\* Capitalized names refer to standard 1:100 000 map sheets, unless otherwise indicated.

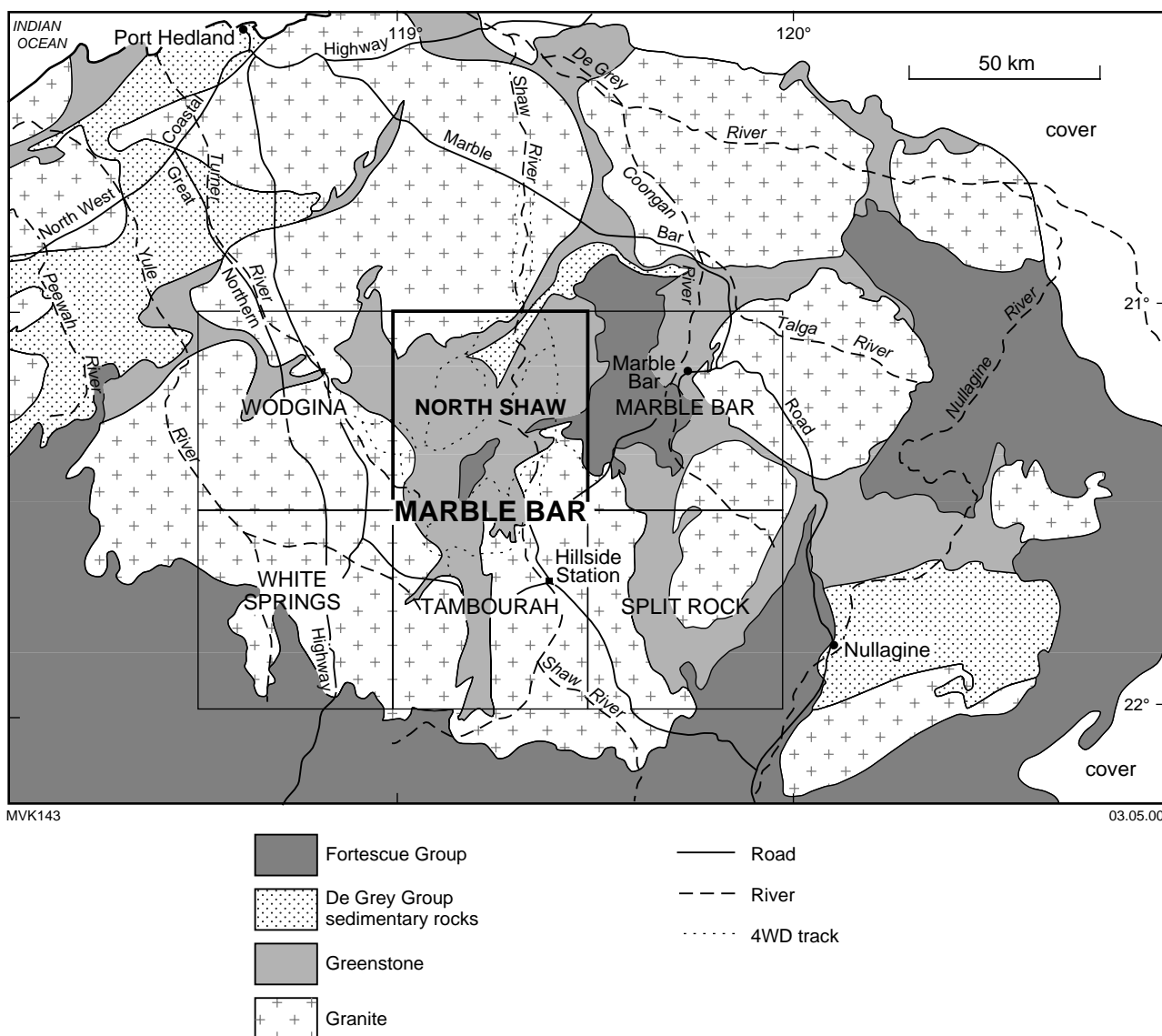


Figure 1. Location of NORTH SHAW in the east Pilbara, showing general geology, drainage, and access

Marble Bar Road about 50 km south of the North West Coastal Highway and heads southwest past the Lalla Rookh mine (abd) to the north-central map area. A second turn-off from the Marble Bar Road is located about 5 km southeast of the Shaw River crossing, following a good, unsealed road to the Normay mine (abd), Panorama Homestead (abd), and the Dresser Mining Centre (abd) in the middle of the North Pole Dome. This road links up with a track that continues south to the North Shaw Mining Centre (abd) and then southeast on good tracks maintained by graziers both north and south of Coolyia Creek, to the Hillside track about 5 km east of the map boundary. Another access point from the Hillside track is to the southeastern part of NORTH SHAW, from the turn-off to the Pilga Homestead (abd). A third access point from the Hillside track is at the turn-off to White Quartz Hill on TAMBOURAH, which leads to the south-central part of the map area. Another turn-off, 18 km west of Woodstock Homestead on TAMBOURAH, leads — via a more tortuous

path — to the southwestern part of NORTH SHAW. A track off the BHP Iron Ore road, 110 km south of the North West Coastal Highway, leads to the west-central part of the map area.

There are no extant settlements on NORTH SHAW, although it was the past site of the Pilga and Panorama homesteads, the North Pole, Dresser, and North Shaw Mining Centres, as well as several individual mines (see **Mineralization**). Ancient petroglyphs are a common site on granitoid tors throughout NORTH SHAW, indicating previous habitation by the nomadic aborigines of the northwest. Water is present in many permanent and temporary waterholes, and is clear and sweet to drink. Grazing is the only agricultural activity on NORTH SHAW, being run by Hillside Station on TAMBOURAH to the southeast and by the Lalla Rookh Homestead on CARLINDIE to the north. In the winter months, NORTH SHAW is host to many fossickers who pass through hoping for a show of gold.

## Physiography

NORTH SHAW has a bimodal topography that reflects the bedrock geology; greenstones outcrop as strike-controlled ridges with a maximum height of 462 m in the south-central part of the map area (Fig. 2a), whereas granitoid rocks are weathered flat with a subdued, undulating topography locally broken by kopjes and flat-capped laterite mesas (Fig. 2b). This is represented by the 'range' and 'low hills' divisions shown on Figure 3. Outcrop is extremely good, except in the northwest where the Carlindi Granitoid Complex is extensively covered by an alluvial–colluvial plain. The most rugged topography is in a north-northeasterly trending belt following the Lalla Rookh – Western Shaw Structural Corridor (LWSC) of deformed greenstones. This area represents a dissected Tertiary plateau, as indicated by the flat tops to many hills (Fig. 2c), and by their covering of ferruginous duricrust. Only the tops of the largest hills in the south have some relief, and even these are characteristically rounded and subdued. On these hills, the ferruginized horizon forms a lip partway down from the top of the hills (Fig. 2d), indicating an origin through groundwater processes. Elsewhere, as at Table Top West and 4 km west of Pyramid Hill, the ferruginous duricrust is a surface weathering feature on an eroded peneplain of the underlying rocks. The Agrippa and Panorama ridges form prominent features in the east-central map area, as does Sunset Ridge in the central-north (Fig. 3), which faces west and glows crimson as the sun sets. The area near the North Pole Mining Centre is flat, low, and completely encircled by steep-sided hills, including the Barite Range to the east.

NORTH SHAW is transected by innumerable small creeks, many larger creeks, and the Shaw River, which reaches a maximum width of 1 km in the southeastern part of NORTH SHAW (Fig. 3). Drainage flows dominantly northwards to the Indian Ocean, but the creeks are dry for most of the year, except during the wet summer months. The ground is evenly covered by several species of spinifex grass, the largest and most spiky of which is found in creekbeds or on their banks. Outside of creeks, the size and species of spinifex depend on the availability of near-surface water and when the area was last burned. Recently burned areas contain several species of flowering plants, and in sandy areas, the Sturt Desert Pea is common. Some valleys and plains are sparsely covered by native shrubs and bushes, including species of *Grevillea*, and wattle is common in sandy soil on granitoid rocks. Larger creeks and rivers are lined by ghost gums and paperbark trees in which corellas, galahs, and rarer coucal pheasants roost.

**Figure 2.** Physiography of NORTH SHAW: a) typical rugged terrain underlain by greenstones, which was formed by the dissection of a plateau, as indicated by equal heights of all hills; b) typical outcrop pattern of granitoid rocks with kopjes on a flat plain; c) flat, laterite-capped peneplain (left background) that has been dissected to form low-range topography (right foreground); d) laterite lip of rounded hill that has fresh outcrop at the top of the hill, indicating the local formation of laterite through groundwater processes



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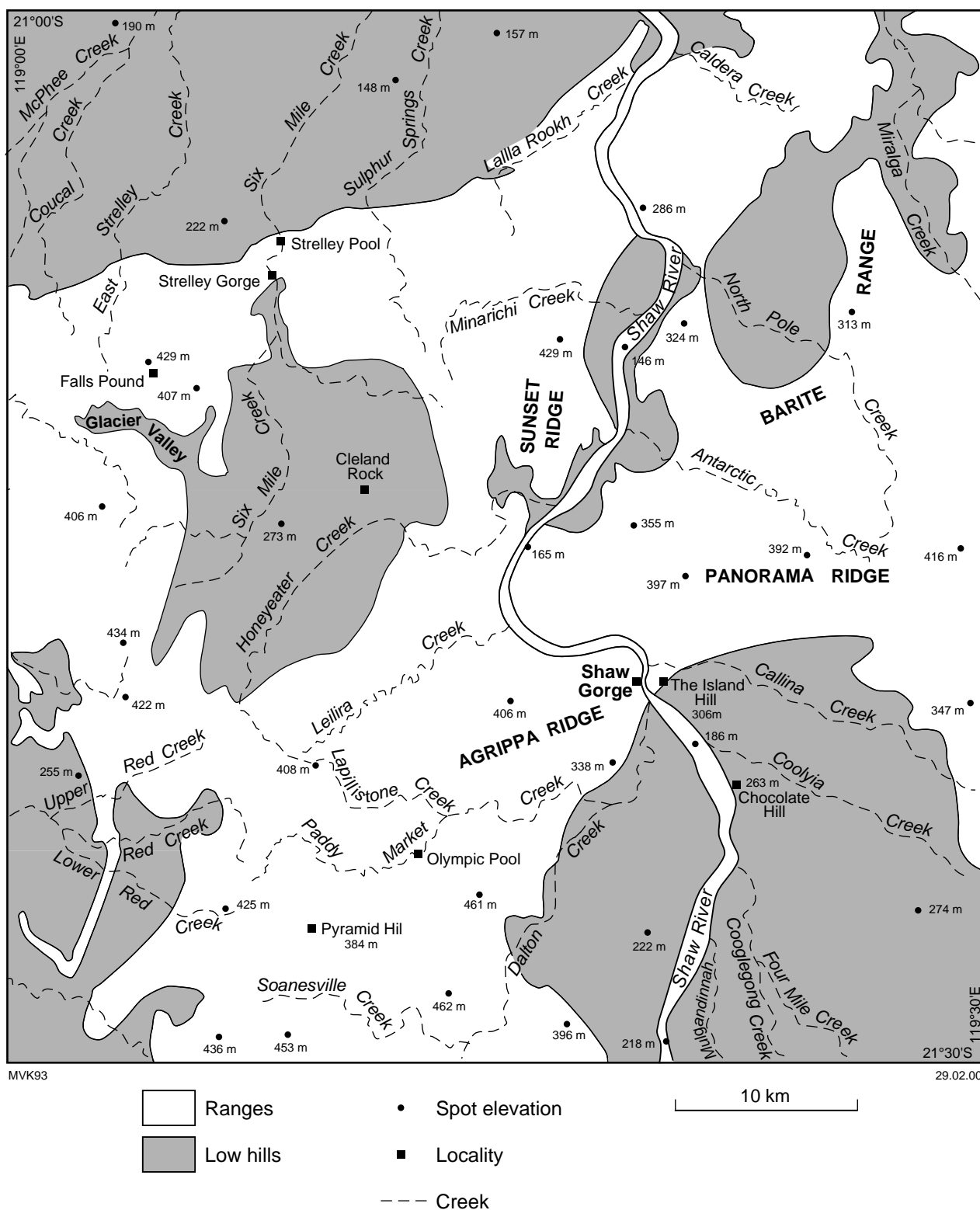


Figure 3. Physiography of NORTH SHAW showing the division into low hills and ranges, and the main drainage systems. Named physiographic features are indicated. Spot height elevations are in metres. Note how the topography is closely controlled by the geology with areas of low hills underlain by granitoid rocks, and ranges underlain by greenstones (compare with Fig. 5)

## Map reliability

NORTH SHAW is the result of mapping by several geologists from a variety of organizations (Fig. 4). Mapping by the author was conducted primarily at 1:25 000 scale and subordinatedly at 1:10 000 scale in complex areas. This was achieved as part of an Australian Research Council Post-Doctoral Fellowship at the University of Newcastle between 1994 and 1996 with Dr Bill Collins, who also contributed to mapping. During this period, 1:25 000-scale colour airphotos were generously provided by Sipa Resources Pty Ltd, who also provided field support in 1994. In subsequent years, field support was provided by CRA Exploration Pty Ltd (1995) and the Australian Geological Survey Organisation (AGSO; 1996). Sarah Van Kranendonk assisted mapping in 1994, and Lynn Pryer was a mapping assistant for two weeks in the 1996 season.

Mapping for the GSWA commenced in 1997 and continued through part of 1998, using 1:25 000-scale colour airphotos (1995, available from the Department of Land Administration; DOLA). Field mapping with AGSO and the Geological Survey of Western Australia (GSWA) was complemented by the use of processed Landsat Thematic Mapper (TM) imagery (Glikson, 1997) and 400 m line-spaced airborne magnetic and radiometric data (Mackey and Richardson, 1997a,b). Mapping was ably assisted by Mark Pawley, a PhD student at the University of Newcastle. Additional mapping assistance was provided in 1997 by Darcy Baker, also a PhD student at the University of Newcastle.

Sipa Resources Pty Ltd mapped the greenstones of NORTH SHAW at 1:25 000 scale between 1991 and 1999 as part of the 'Panorama Project' for the exploration of base metal sulfide deposits (Morant, 1998). Members of the project included (in alphabetical order) Nick Archibald, Roger Buick, John Maniw, Peter Morant, and Jim Thornett. Map and geochemical data from this project were used as a base for mapping and the compilation of local areas (see Fig. 4). Detailed mapping of the Strelley Granite and flanking greenstones of the Sulphur Springs Group was conducted by Peter Morant, chief exploration geologist of Sipa Resources Pty Ltd, and Carl Brauhart, a PhD student at the University of Western Australia, using 1:10 000-scale colour airphotos.

## Previous investigations

The early history of geological investigations in the east Pilbara region is detailed by Hickman (1983). The MARBLE BAR and NULLAGINE 1:250 000 sheets were remapped by the GSWA in 1972 and 1973, resulting in a number of publications, as cited by Hickman (1983).

## Greenstone succession

The Archaean layered succession of the east Pilbara was divided into the lower, dominantly volcanic Warrawoona Group and the overlying, dominantly sedimentary Gorge Creek Group, as formally defined by Lipple (1975). The type area of the Gorge Creek Group was thought to be on NORTH SHAW. Ingram (1977) divided the Warrawoona

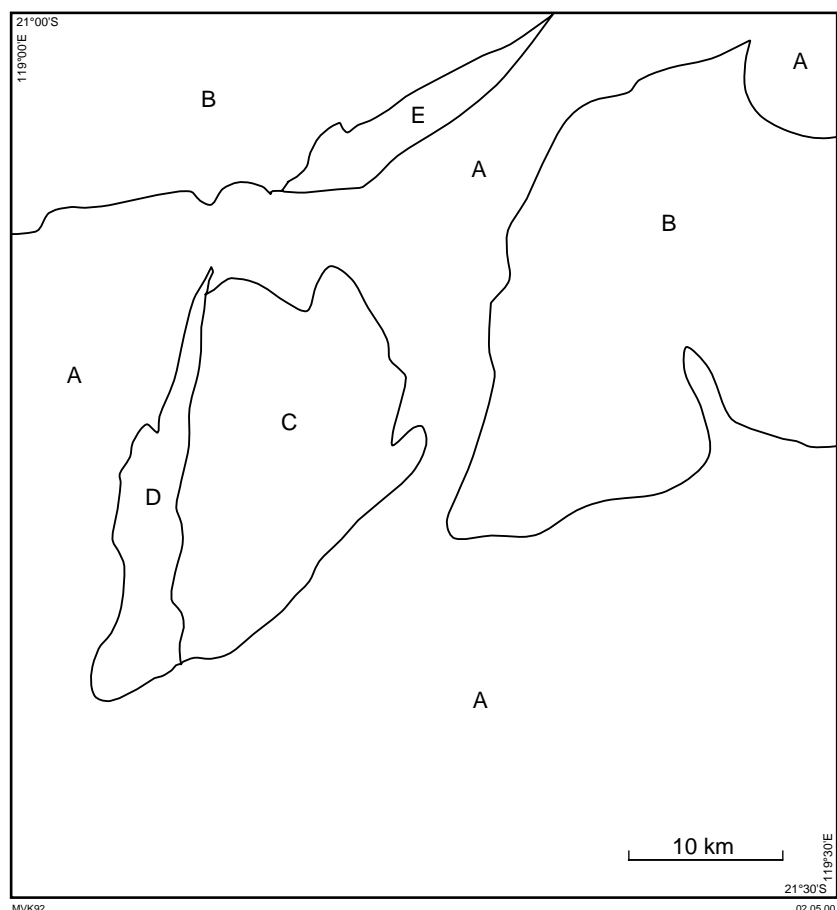
Group of the east Pilbara into four formations and the 'Mosquito Creek Succession', and recognized cyclicity in the volcanic formations.

The stratigraphy of the east Pilbara was modified by Lipple (1975) and Hickman (1977, 1983), who divided the Warrawoona Group into nine formations and the Gorge Creek Group into six formations, the latter including the Lalla Rookh Sandstone and Mosquito Creek Formation at the top. This stratigraphy was again modified when Hickman (1990) grouped together the sedimentary rocks of the Mosquito Creek Formation, Lalla Rookh Sandstone, and Mallina Formation into the De Grey Group.

Hickman (1983) identified a prominent unit of chert and subordinate, interbedded volcanic and sedimentary rocks within the Warrawoona Group, calling it the Towers Formation. Hickman (1990) used this unit in a craton-wide correlation of the volcano-sedimentary stratigraphy, equating the Marble Bar Chert in the Marble Bar Belt with the Strelley Pool Chert in the East Strelley Belt, the chert-barite association in the North Pole Dome, and with cherts in the west Pilbara. However, the nature of these correlations has been controversial. In particular, Lowe (1980, 1983) and Di Marco and Lowe (1989a) correlated the chert at Strelley Pool with chert overlying the Panorama Formation in the North Pole Dome and called it the Towers Formation, even though it is at a higher stratigraphic level than the type occurrence of the Towers Formation in the North Pole Dome (Hickman, 1990). Buick et al. (1995) recognized that the Strelley Pool Chert in the East Strelley Belt overlies a c. 3515 Ma sequence of dominantly tholeiitic greenstones, which they named the Coonterunah succession, across the oldest known unconformity on Earth. Van Kranendonk and Morant (1998) upgraded the older rocks to group status and the Strelley Pool Chert to formation status. Nijman et al. (1998) studied the facies architecture of barite in the Towers Formation and associated chert-barite dykes.

Detailed mapping of the Soanesville Belt by Sipa Resources Pty Ltd and graduate students at the University of Western Australia (Morant, 1995; Vearncombe et al., 1995) showed that a mafic to felsic volcanic sequence around the Strelley Granite was significantly younger (c. 3260 Ma) than the Warrawoona Group with which it had previously been correlated (Hickman, 1990). Informally named the Strelley succession, these rocks have since been assigned to the Pincunah and East Strelley Belts to the west (Van Kranendonk, 1997; Van Kranendonk and Collins, 1998). Van Kranendonk and Morant (1998) ascribed these rocks to four formations that comprise the Sulphur Springs Group. Van Kranendonk (1997) showed that the upper part of the Sulphur Springs Group was, at least locally, unconformable on the Warrawoona Group. Uranium-Pb zircon geochronology on both the volcanic rocks of the Sulphur Springs Group and the Strelley Granite indicate an identical age of c. 3240 Ma (Morant, 1998; Brauhart, 1999). Structural and geochemical studies have shown the Strelley Granite to be a synvolcanic laccolith to the Sulphur Springs Group (Vearncombe, 1996; Van Kranendonk, 1997; Brauhart, 1999).

Glikson and Hickman (1981a) and Hickman (1983) studied the geochemistry of mafic volcanic rocks in



- A: Detailed ground traverses by M. J. Van Kranendonk at 1–2 km spacing using 1:25 000-scale colour airphotos; additional map data from the Sipa–Outokumpu Panorama Project.
- B: Spaced ground traverses at 2–4 km intervals and airphoto interpretations by M. J. Van Kranendonk using 1:25 000-scale airphotos; additional map data from the Sipa–Outokumpu Panorama Project.
- C: Detailed ground traverses by C. Brauhart and P. Morant using 1:10 000-scale colour airphotos; reconnaissance ground traverses by M. J. Van Kranendonk.
- D: Detailed ground traverses by P. Morant using 1:10 000-scale colour airphotos; reconnaissance ground traverses and airphoto interpretations by M. J. Van Kranendonk.
- E: Detailed ground traverses by R. Buick using 1:25 000-scale colour airphotos; spaced ground traverses by M. J. Van Kranendonk at 2–4 km intervals.

Figure 4. Reliability diagram of NORTH SHAW

several transects across the Warrawoona Group. They identified progressive geochemical trends up-section and deduced from this that the greenstones had not been repeated by thrusting, a conclusion subsequently borne out by geochronology. Barley et al. (1979) and Barley (1993) studied the volcanic facies of the Warrawoona Group and proposed that it represents a volcanic arc and near-arc sequence of tholeiitic and calc-alkaline suites deposited in shallow to deep water, and capped by a felsic volcanic sequence developed, in part subaerially, from topographic domes.

DiMarco and Lowe (1989a) examined the stratigraphy and sedimentology of the main felsic volcanic horizons in the Warrawoona Group around the North Pole Dome and the Shaw and Corunna Downs Granitoid Complexes. These authors determined that the Duffer and Panorama Formations were successive facies of a single felsic volcanic event; the former deposited primarily on subaqueous volcanic aprons flanking felsic volcanic centres, the latter marked by explosive pyroclastic activity (DiMarco and Lowe, 1989b,c). Subsequent studies indicated that the Duffer and Panorama Formations are

statistically significantly different in age (3472–3463 Ma vs 3458–3454 Ma; Thorpe et al., 1992a; McNaughton et al., 1993; Nelson, 1997) and are locally separated by mafic volcanic rocks (Williams, 1998). Geochemical data indicate that these two felsic units were derived through different processes (Cullers et al., 1993). Hickman (1983) argued on geochemical grounds that the Duffer and Wyman Formations were derived from distinctly different sources (the former is predominantly dacitic whereas the latter is entirely composed of rhyolite), in much the same way as the pre- and post-tectonic granitoid rocks.

Hickman (1983) correlated sedimentary rocks in the core of the Pilgangoora Syncline (herein called the Pincunah Belt) with the Lalla Rookh Sandstone (De Grey Group), but more detailed studies indicated that they represent an upper part of the Gorge Creek Group (Wilhelmij and Dunlop, 1984). Eriksson (1981) interpreted these rocks to have been deposited in a platform (alluvial) to trough (submarine fan) environment and proposed a setting across a continental–marine transition with a narrow continental shelf. Sedimentary rocks of the Strelley Block (herein called the East Strelley Belt) were interpreted by Wilhelmij and Dunlop (1984) as parts of two depositional cycles, represented by the Gorge Creek Group and the Lalla Rookh Sandstone (De Grey Group) respectively. Wilhelmij (1986) showed that deposition of the Gorge Creek Group in both the Pincunah and East Strelley Belts was in a series of fault-bound basins during extension. Van Kranendonk and Morant (1998) identified two new formations in the Gorge Creek Group on NORTH SHAW, one at the base and one at the top of the group.

Krapez (1984, 1989, 1993) and Eriksson et al. (1994) presented a detailed stratigraphic analysis of the Lalla Rookh Basin and concluded that the coarse clastic sequence was deposited in a fault-bound, pull-apart basin during regional sinistral strike-slip faulting. Van Kranendonk and Collins (1998) confirmed that the Lalla Rookh Sandstone was deposited in a syntectonic fault-bound basin, but presented data that showed it formed as a small trough ahead of the north-moving and uplifting Strelley Granite. These authors also presented a detailed model of deformation on NORTH SHAW and TAMBOURAH, whereby volcanism and related granitoid doming at c. 3240 Ma (Van Kranendonk, 1997) were succeeded by sinistral transpression at c. 2950 Ma (regional D<sub>3</sub> deformation) within the Central Pilbara Structural Corridor (herein called the Lalla Rookh – Western Shaw Structural Corridor, as redefined by Van Kranendonk, 1998).

Hickman (1983, 1984) regarded the stratigraphy of the entire granitoid–greenstone terrane in the north Pilbara area as a single stratigraphic succession, despite the fact that a similar, earlier correlation of west and east Pilbara stratigraphy by Ryan (1965) was abandoned by Fitton et al. (1975). Horwitz (1990) suggested that the Warrawoona Group was restricted to an eastern, ancient nucleus to the craton, and identified two regional unconformities at the base of the Gorge Creek and Whim Creek Groups. In this model, he grouped the Mosquito Creek Formation with the Gorge Creek Group, and sedimentary rocks of the Lalla Rookh and Mallina Basins with the Whim Creek Group. Subsequent 1:100 000

mapping of the west Pilbara, combined with geochronology (Nelson, 1997, 1998), has supported the contention of Fitton et al. (1975) that the geology and stratigraphy of the west and east Pilbara are distinct and separate, having common elements only after the c. 3.0 Ga Whim Creek Group (Hickman, 1997; Smithies et al., 1999).

Krapez (1993) presented a sequence-stratigraphic model for the ‘north Pilbara granite–greenstone terrane’ based on observed changes in stratigraphy across a set of craton-wide, sinistral strike-slip faults first identified by Krapez and Barley (1987). Subsequently, it has been found that the proposed model relied too heavily on an *a priori* belief in uniformitarian principles (Trendall, 1995), incorrect correlations of undated stratigraphic sequences (Smithies et al., 1999), and an unfounded significance of the strike-slip faults as indicated by Van Kranendonk and Collins (1998), who showed that domain boundary 2–3 of Krapez (1993) is not a terrane boundary.

### Palaeontology

NORTH SHAW is famous as the location of Earth’s two oldest known stromatolitic horizons, as well as Earth’s oldest known microfossils. Nodular and columnar stromatolites were first discovered in cherts of the former Towers Formation of the North Pole Dome (herein the Dresser Formation) by J. Dunlop in 1977, who described textures resembling those found in stromatolitic carbonates (including stromatolitic, edgewise conglomerate with pre- or syn-depositional oncolitic overgrowth) in a paper devoted to the description of probable microfossils from this unit (Dunlop et al., 1978). The North Pole stromatolites were described by Walter et al. (1980), Buick et al. (1981), Groves et al. (1981), and Walter (1983). Further microfossils were discovered in cherts of the Apex Basalt immediately overlying the Dresser Formation in the North Pole Dome (Awramik et al., 1983), as well as in the Marble Bar Belt to the east (Schopf, 1992, 1993).

A second, slightly younger horizon of stromatolitic chert was discovered by Lowe (1980, 1983), who referred to this unit of silicified carbonates and evaporites as the Strelley Pool Chert for the type area around Strelley Pool in NORTH SHAW. Although the biogenicity of the distinctly conical structures in this unit has since been contested (Buick et al., 1981; Lowe, 1994) and debated (Buick et al., 1995; Lowe, 1995), recent analysis of these structures at a locality in the North Pole Dome has confirmed their biological origin (Hofmann et al., 1999). The discovery of probable microfossils from the Strelley Pool Chert in the type area (Schopf and Packer, 1987) further supports the contention for early life in these very old rocks.

### Granitoid rocks

Bettenay et al. (1981) outlined the component features of the Shaw Granitoid Complex and its protracted, complex tectonothermal history, identifying several phases of granitoid rocks and four phases of deformation. Hickman (1983) identified and named most of the component intrusive phases within the granitoid complexes, and identified the abnormally sodic composition of the early suites. He cited earlier work and presented new data,

indicating that the early sodic magmas could not have been produced through melting of sedimentary rocks. Bickle et al. (1983) described a little-deformed, c. 3500 Ma, calc-alkaline tonalite–trondhjemite–granodiorite–granite (TTG) suite in the northern part of the Shaw Granitoid Complex (North Shaw Suite), which they interpreted as a transitional upper crustal level equivalent to high-grade Archaean gneiss belts elsewhere. Bickle et al. (1993) studied these rocks in more detail and compared them favourably to modern calc-alkaline igneous suites developed in subduction settings, derived from melts in equilibrium with garnet in the lower crust. The age and geochemistry of this suite indicate that it represents the intrusive equivalents of the felsic volcanic rocks of the Duffer Formation.

Collins and Gray (1990) studied the Rb–Sr isotopic characteristics of the c. 3300 Ma suite of granitoid rocks that intrude an older suite of banded tonalitic gneisses in the Mount Edgar Granitoid Complex. They found that the younger suite yielded modelled source-rock ages of c. 3475 Ma and suggested that a substantial component was derived from Duffer Formation equivalents, a hypothesis subsequently confirmed by Collins (1993). Hickman (1983 and citations of earlier work therein) and Bickle et al. (1983) found that late- to post-tectonic granitoid rocks were derived by partial melting of pre-existing sialic crust.

## Deformation

Hickman (1975) identified five periods of deformation, the second of which was interpreted to be responsible for the dome-and-syncline pattern. Doming was thought to have formed by solid-state uplift of pre-existing granitoid bodies rather than by diapiric magmatic intrusion, as described in Hickman (1983, 1984) and supported by Collins' (1989) detailed study of the Mount Edgar Granitoid Complex.

In contrast, Bickle et al. (1980, 1985) and Bettenay et al. (1981) identified intercalations of greenstones within granitoid rocks of the Shaw Granitoid Complex and interpreted them to be the result of  $D_1$  thrusting, followed by  $D_2$  recumbent, isoclinal folding and the development of a bedding-parallel foliation. These two events, and associated medium-pressure, low-temperature metamorphism, were postulated to have arisen from Alpine-style thrusting (Morant, 1984; Bickle et al., 1985). Subsequent  $D_3$  shear belts and upright folds that developed syn- to post-intrusion of monzogranites were thought to represent the products of doming, which was later dated as c. 2966 Ma (Bickle et al., 1989). Boulter et al. (1987) also recognized the effects of horizontal tectonism and suggested that it occurred prior to deposition of the Gorge Creek Group. Hickman (1984), Collins (1989), Van Kranendonk and Collins (in prep.), and Van Kranendonk et al. (in press) pointed out that granitoid diapirism and interference between diapiric structures could account for the local development of horizontal structures, and that there was no large-scale duplication of stratigraphy. Collins and Van Kranendonk (1999) showed that the local development of high-pressure metamorphism at the margins of some granitoid complexes was consistent with

models of hot, rising diapirs as presented in Mareschal and West (1980). Van Haaften and White (1998) attempted to show that the type area of the Warrawoona Group in the Marble Bar Belt comprised an inverted stratigraphy, but this was discredited by Van Kranendonk et al. (in press). Collins et al. (1998) discussed the driving forces behind partial crustal overturn in the east Pilbara.

## Geochronology

A large volume of geochronological data now exists for the Pilbara Craton, such that only selected data relevant to NORTH SHAW are summarized. The Panorama Formation is 3458–3454 Ma in its type area of the North Pole Dome and Coppin Gap Belt respectively, the same age as the North Pole Monzogranite ( $3459 \pm 18$  Ma; Thorpe et al., 1992a; Nelson, 1997).

Vearncombe (1996) dated galena from the Sulphur Springs and Kangaroo Caves Zn–Cu prospects at the top of the Sulphur Springs Group, and obtained a Pb–Pb age of  $3257 \pm 8/-6$  Ma. She also identified post-c. 3.26 Ga mobility of the Pb, which she interpreted to have occurred at c. 2760 Ma, during eruption of basal Fortescue Group lavas. The isochron and error envelope of Pb from the genetically related Strelley Granite did not include the initial Pb composition of galena from the prospects, thus precluding exclusive derivation of the Pb for the mineralization from the granite. The higher  $^{207}\text{Pb}/^{204}\text{Pb}$  ratios from galena were thus interpreted to include a component derived from contamination by older sialic crust. Subsequent sensitive high-resolution ion microprobe (SHRIMP) U–Pb dating of zircons from the Strelley Granite and related felsic volcanic rocks of the Kangaroo Caves Formation has given ages of c. 3238–3235 Ma (Morant, 1998).

Pidgeon (1978) obtained a U–Pb zircon date of  $3417 \pm 40$  Ma on grey gneisses of the Shaw Granitoid Complex. Williams et al. (1983) obtained an ion microprobe U–Pb zircon age of  $3485 \pm 30$  Ma from grey gneisses in the Shaw Granitoid Complex, while Collerson and McCulloch (1983) obtained a Sm–Nd model age of 3240–3460 Ma on similar samples. Bickle et al. (1983) dated the North Shaw Suite at  $3499 \pm 22$  Ma (Pb–Pb isochron), an age apparently confirmed by a U–Pb zircon SHRIMP age of  $3493 \pm 4$  Ma from an unspecified locality (McNaughton et al., 1988). A  $3578 \pm 4$  Ma xenolith of gabbroic anorthosite (McNaughton et al., 1988) from the  $3431 \pm 4$  Ma South Daltons Pluton of the Shaw Granitoid Complex (McNaughton et al., 1993), and  $3524 \pm 6$  Ma and  $3509 \pm 15$  Ma xenocrystic zircons from the eastern margin of the Shaw Granitoid Complex (Zegers, 1996) lend support to the idea of older crust in the east Pilbara. Evidence of yet older crust was obtained from a c. 3724 Ma xenocrystic zircon in the Panorama Formation on NORTH SHAW (Thorpe et al., 1992a). Bickle et al. (1989) found that late tectonic (mainly adamellite) suites in the Shaw Batholith were c. 2960 Ma and that the post-tectonic Coonglegong Adamellite was  $2847 \pm 34$  Ma, with Sm–Nd model ages of 3200–3600 Ma.

Bickle et al. (1993) confirmed the extent of the c. 3500 Ma suite on NORTH SHAW, and also identified a

3338 ± 52 Ma plutonic phase based on a Pb–Pb isochron. Additional Pb-isotopic work was presented by Richards et al. (1981), Richards (1986), and Thorpe et al. (1992b). Whereas the 3490 Ma model age from bedded barite at North Pole reflects a depositional (or diagenetic) age for the Dresser Formation, the other data are from epigenetic veins and thus represent deformation–metamorphic events (e.g. dates from the Big Bertha and Lynas Find mines), or the effects of younger events contaminated with older crustal Pb (e.g. dates from the Soanesville and Lalla Rookh mines). It is noteworthy that a c. 3400 Ma model age on galena from the Normay mine is close to the age of c. 3380 Ma metamorphic zircons in the Mount Edgar Granitoid Complex (Williams and Collins, 1990), an Ar–Ar plateau age of 3398 ± 6 Ma on hornblende from the Coongan Belt (Davids et al., 1997), and c. 3410 Ma zircon overgrowths from the Shaw Granitoid Complex (Zegers, 1996; M. J. Van Kranendonk, unpublished U–Pb SHRIMP zircon data). These data suggest a widespread metamorphic event at this time. The North Pole barite sample has high  $\mu$  ( $^{238}\text{U}/^{204}\text{Pb}$ ) values of 9.45, a characteristic shared by the North Shaw Suite (Bickle et al., 1989). Tonalite and granodiorite from the North Shaw Suite have even higher  $\mu$  values of 9.66–9.9. Together, these results reflect recycling of significantly older, felsic, upper crustal rocks during eruption of the North Shaw Suite, and the possibility of basement to the Warrawoona Group that is ‘...appreciably older than the minimum 3724 Ma age for a xenocrystic zircon from the Panorama Formation’ (Thorpe et al., 1992b, p. 404).

Wijbrans and McDougall (1987) obtained two principal ages of metamorphism in greenstones of the Western Shaw Belt on TAMBORA from a study of Ar–Ar dating; namely c. 3240 Ma for amphibolite-facies metamorphism, and c. 2950 Ma for overprinting greenschist facies. Van Kranendonk and Collins (1998), on the basis of structural field mapping, ascribed these events to D<sub>2</sub> regional doming and D<sub>3</sub> sinistral, transpressional deformation respectively. Zegers (1996) confirmed a c. 2945 ± 5 Ma age for related D<sub>3</sub> sinistral shear deformation along the western margin of the Shaw Granitoid Complex (Mulgandinnah Shear Zone). She also obtained numerous Ar–Ar results on hornblende from in and around the Shaw Granitoid Complex. The oldest result is a partial plateau age of 3522 ± 13 Ma from a large amphibolite xenolith at the northern end of the complex on NORTH SHAW (AMG 483376\*), identical within error to more precise age determinations on the Coonterunah Group around the Carlindi Granitoid Complex (Buick et al., 1995). The same rock also yielded a fraction of hornblende with a plateau age of 3466 ± 13 Ma, which Zegers (1996) interpreted to be derived from coarse hornblende adjacent to an aplite dyke, and thus of contact-metamorphic origin related to intrusion of the Shaw Granitoid Complex. A maximum age for the reverse, east-side-up Split Rock Shear Zone along the eastern contact of the Shaw Granitoid Complex is c. 3200 Ma, and it was reactivated under greenschist-facies conditions at c. 2950 Ma (Zegers, 1996; Davids et al., 1997).

\* Localities are specified by the Australian Map Grid (AMG) standard six-figure reference system whereby the first group of three figures (eastings) and the second group (northings) together uniquely define position, on this sheet, to within 100 m.

## Mount Bruce Supergroup

The Mount Bruce Supergroup is a Late Archaean cover sequence that lies unconformably on the North Pilbara Terrain (Trendall, 1990). Noldart and Wyatt (1962) divided the rocks of the ‘Nullagine System’ — as it was then known — into several formations, as a precursor to MacLeod et al. (1963), who referred to the entire sedimentary and volcanic fill of the Hamersley Basin as the Mount Bruce Supergroup and defined the component Fortescue, Hamersley, and Wyloo Groups. Kriewaldt (1964) was the first to recognize the Mount Roe Basalt at the base of the Fortescue Group. Hickman (1983, table 5, p. 117) summarized the stratigraphy of the Fortescue Group.

The age of the Fortescue Group was constrained by a U–Pb zircon date of 2768 ± 16 Ma for the Spinaway Porphyry, which intrudes the base of the group (Pidgeon, 1984). Arndt et al. (1991) provided further geochronological age constraints on the Fortescue Group from U–Pb analysis of zircon, and showed that tuffs near the base of the group are c. 2765 Ma (maximum 2775 ± 10 Ma), the Pillingilli Tuff (which is correlated with the Tumbiana Formation) is 2715 ± 6 Ma, and the Jeerinah Formation at the top of the group is c. 2687 Ma. Nelson (1997) dated a dacite in the Maddina Basalt, which lies in between the Tumbiana and Jeerinah Formations, as 2717 ± 2 Ma. Wingate (1999) dated the Black Range Dyke at c. 2772 Ma.

Blake and Barley (1992) proposed a sequence-stratigraphic model for the Mount Bruce Supergroup, renaming it the Mount Bruce Megasequence Set and identifying component sequences, supersequences, and megasequences. Blake (1993) summarized the results of his extensive work in the Fortescue Group over the preceding ten years and, in particular, presented a detailed stratigraphic sequence analysis of the lower Nullagine and Mount Jope Supersequences, speculating that they represent a record of a protracted, two-phase, late Archaean continental break-up. In this model, the Nullagine Supersequence is interpreted to have formed during west-northwest–east-southeast directed crustal extension, and the subsequent Mount Jope Supersequence records south-southwest–north-northeast directed rifting of the southern margin of the Pilbara Craton.

## Regional geological setting

The Pilbara Craton is composed of the 3.6–2.8 Ga North Pilbara Terrain and the unconformably overlying volcano-sedimentary sequence (Mount Bruce Supergroup) of the 2.77–2.3 Ga Hamersley Basin (Arndt et al., 1991; Trendall, 1990). Hickman (1983) grouped the older volcano-sedimentary rocks into a single stratigraphic sequence — the Pilbara Supergroup. NORTH SHAW is located in the ancient, eastern nucleus of the North Pilbara Terrain, which is characterized by circular to elliptical, domal granitoid complexes with diameters of between 25 and 110 km (previously batholiths; Hickman, 1983; Trendall, 1990) and intervening, arcuate belts of commonly steeply dipping volcano-sedimentary rocks, referred to as greenstone belts (Fig. 1).

## Greenstone belts

Previous work has shown that greenstone belts of the North Pilbara Terrain are composed of four unconformity- or intrusion-bound supracrustal sequences deposited over c. 575 Ma (Hickman, 1983; Horwitz, 1990; Thorpe et al., 1992a; Krapez, 1993; McNaughton et al., 1993; Buick et al., 1995). The oldest rocks include the c. 3515 Ma, bimodal mafic and felsic volcanic Coonterunah Group, located around the southern margin of, and intruded by, the  $3468 \pm 4$  Ma Carlindi Batholith (Fig. 5; Buick et al., 1995; Van Kranendonk and Morant, 1998). Elsewhere, the oldest exposed rocks consist of undated mafic-ultramafic volcanic rocks at the base of the Warrawoona Group (Talga Talga Subgroup), and conformably overlying felsic volcanic rocks of the 3472–3463 Ma Duffer Formation (Hickman, 1983; Hickman, 1990; Thorpe et al., 1992a; McNaughton et al., 1993). These are, in turn, overlain by interbedded chert (Dresser Formation), the Apex Basalt, and c. 3458–3454 Ma felsic volcanoclastic rocks of the Panorama Formation, which all belong to the lower part of the Salgash Subgroup (Hickman, 1983; Thorpe et al., 1992a).

Conformably overlying the Panorama Formation in North Pole Dome and southeast of the Corunna Downs Granitoid Complex (southeast of NORTH SHAW), and unconformably overlying the Coonterunah Group south of the Carlindi Granitoid Complex is the Strelley Pool Chert, a distinctive unit of silicified epiclastic rocks and evaporites with common wavy- to domal-laminated structures and conical stromatolites (Lowe, 1983, 1994; Buick et al., 1995; Hofmann et al., 1999). Above this lies a thick section of mafic, ultramafic, and rare felsic volcanic rocks of the Euro Basalt in the upper part of the Salgash Subgroup. A felsic volcanic flow from within the upper part of the Salgash Subgroup, southeast of the Corunna Downs Granitoid Complex, is  $3324 \pm 4$  Ma (Kelly Formation; McNaughton et al., 1993), the same age as the felsic volcanic Wyman Formation ( $3325 \pm 4$  Ma; Thorpe et al., 1992a; McNaughton et al., 1993), which conformably, or locally unconformably, overlies the Salgash Subgroup east and southeast of NORTH SHAW (Hickman, 1983). The range in ages between dated components of the lower and upper parts of the Salgash Subgroup, and the angular unconformity at the base of the Strelley Pool Chert suggests that the Salgash Subgroup was deposited over about 130 m.y. (Van Kranendonk, 1998).

Northeast of the Yule Granitoid Complex are komatiites, pillow basalts, and felsic volcanic rocks of the c. 3240 Ma Sulphur Springs Group (Van Kranendonk and Morant, 1998; Brauhart, 1999; Buick, R., 1998, pers. comm.). An unconformable relationship of the Sulphur Springs Group on the upper part of the Salgash Subgroup was documented by Van Kranendonk (1997). The Sulphur Springs Group is disconformably overlain by the Gorge Creek Group in this area. A fourth sequence, unconformable on the Gorge Creek Group, is the coarse, clastic De Grey Group (Hickman, 1983), deposited at c. 2940 Ma in response to sinistral transpression across the craton (Krapez and Barley, 1987; Van Kranendonk and Collins, 1998).

A common characteristic of eastern Pilbara stratigraphy is that successive supracrustal sequences are deposited

unconformably above one another, and consistently dip and face away from the domal granitoid complexes (Hickman, 1983, 1984). Dips of bedding become gradually more shallow with decreasing age of the supracrustal succession and distance away from the granitoid complexes, suggesting that supracrustal sequences have wedge shapes that thicken away from the granitoid domes. These features were explained by Hickman (1975, 1984) as the result of progressive granitoid diapirism. The preservation of synclinal remnants of the 2765–2715 Ma Fortescue Group (Arndt et al., 1991; Nelson, 1998) over synclines of the older greenstones between domal granitoid complexes indicates that doming continued until after c. 2715 Ma (Hickman, 1984).

## Granitoid complexes

Granitoid complexes in the eastern part of the Pilbara Craton range from 25 to 110 km in diameter, with centres spaced an average of 60 km apart (Fig. 1). Older plutonic components fall into two distinct age groups, similar to those of felsic volcanic rocks in the adjacent synforms; c.  $3450 \pm 20$  Ma TTG plutons and gneisses (Bickle et al., 1983; Williams et al., 1983; McNaughton et al., 1988, 1993; Williams and Collins, 1990; Buick et al., 1995; Dawes et al., 1995), and a second suite of homogeneous, foliated granitoid plutons that vary in age from  $3310 \pm 10$  Ma in the Mount Edgar, Corunna Downs, Muccan, and Warrawagine Granitoid Complexes (Pidgeon, 1984; Williams and Collins, 1990; Bickle et al., 1993; Nelson, 1997; Barley, M. E., 1997, pers. comm.) to c. 3240 Ma in the Yule, Muccan, and Warrawagine Granitoid Complexes (M. J. Van Kranendonk, unpublished U–Pb SHRIMP zircon data; Nelson, 1999, in prep.). In the Mount Edgar Granitoid Complex, Collins (1993) provided geochemical evidence that the c. 3310 Ma suite of plutons was derived from partial melting of the c. 3450 Ma gneisses.

The Carlindi Granitoid Complex, North Pole Dome, and the northern part of the Shaw Granitoid Complex (Fig. 5) are largely composed of homogeneous, foliated TTG intrusions of the c.  $3450 \pm 20$  Ma suite. However, in other granitoid complexes, the c. 3450 Ma suite is preserved only as remnant gneissic screens around their outer margins, intruded by younger, c. 3300 Ma plutons that comprise the bulk of material in the cores of the domes. Late- (c. 3000 Ma) to post-tectonic (c. 2850 Ma) monzogranites are present in the cores of many granitoid complexes and represent the final stages of cratonization (Bickle et al., 1989).

## Lithotectonic and structural map elements

The Archaean rocks on NORTH SHAW may be divided into two broad classes of lithology, namely granitoids and greenstones. Van Kranendonk (1998) showed that these rocks are distributed in two distinct, overlapping classes of map elements, in addition to that of classical lithostratigraphy: lithotectonic and structural.

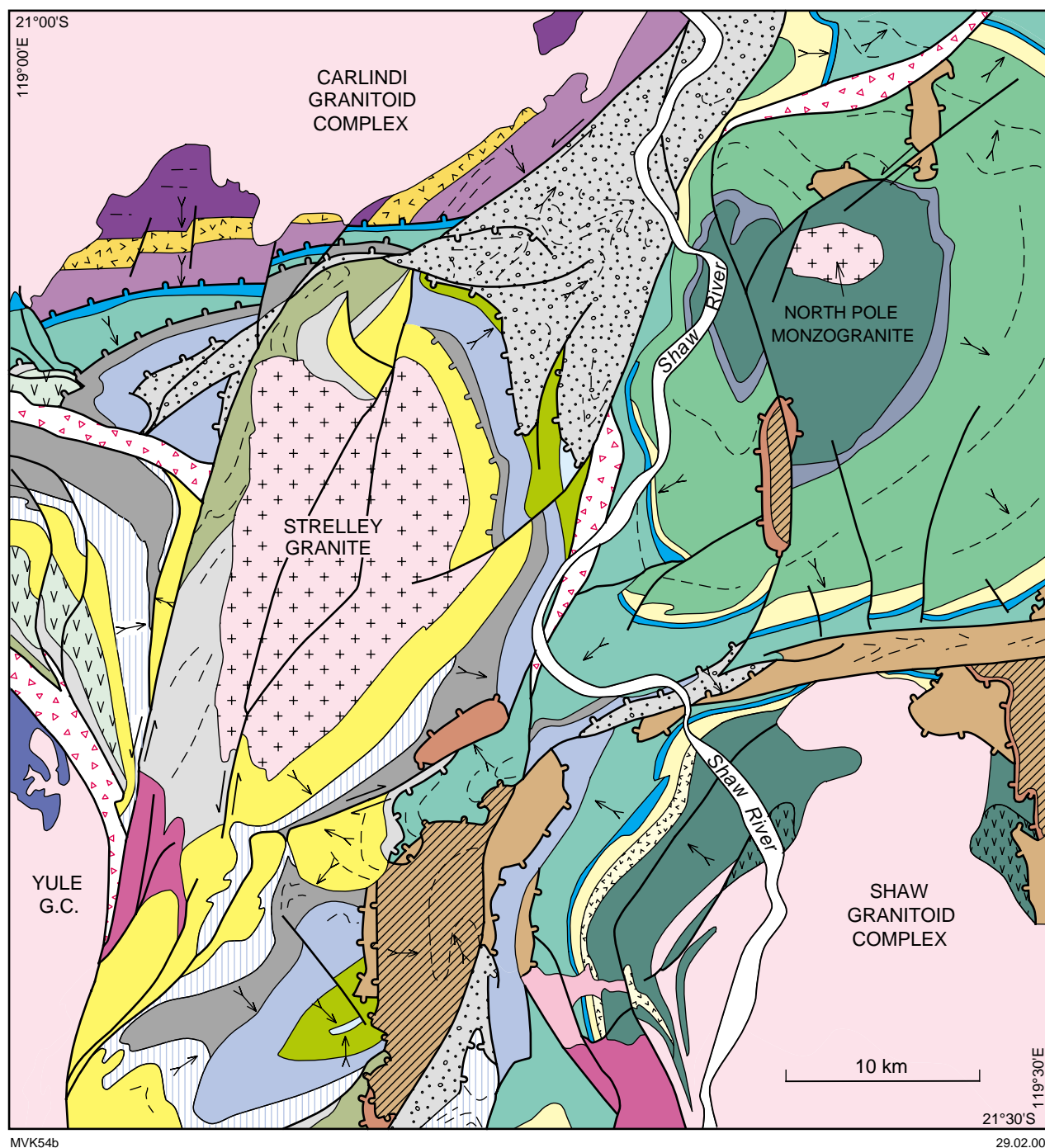


Figure 5. Lithostratigraphic geology of NORTH SHAW

Lithotectonic elements are defined as distinct structural elements for the two classes of lithology. They include, for greenstones, parts of eight greenstone belts, two greenstone complexes, and two synclinoria of late-tectonic clastic rocks. Granitoid rocks are subdivided into parts of three domal granitoid complexes and three discrete plutons.

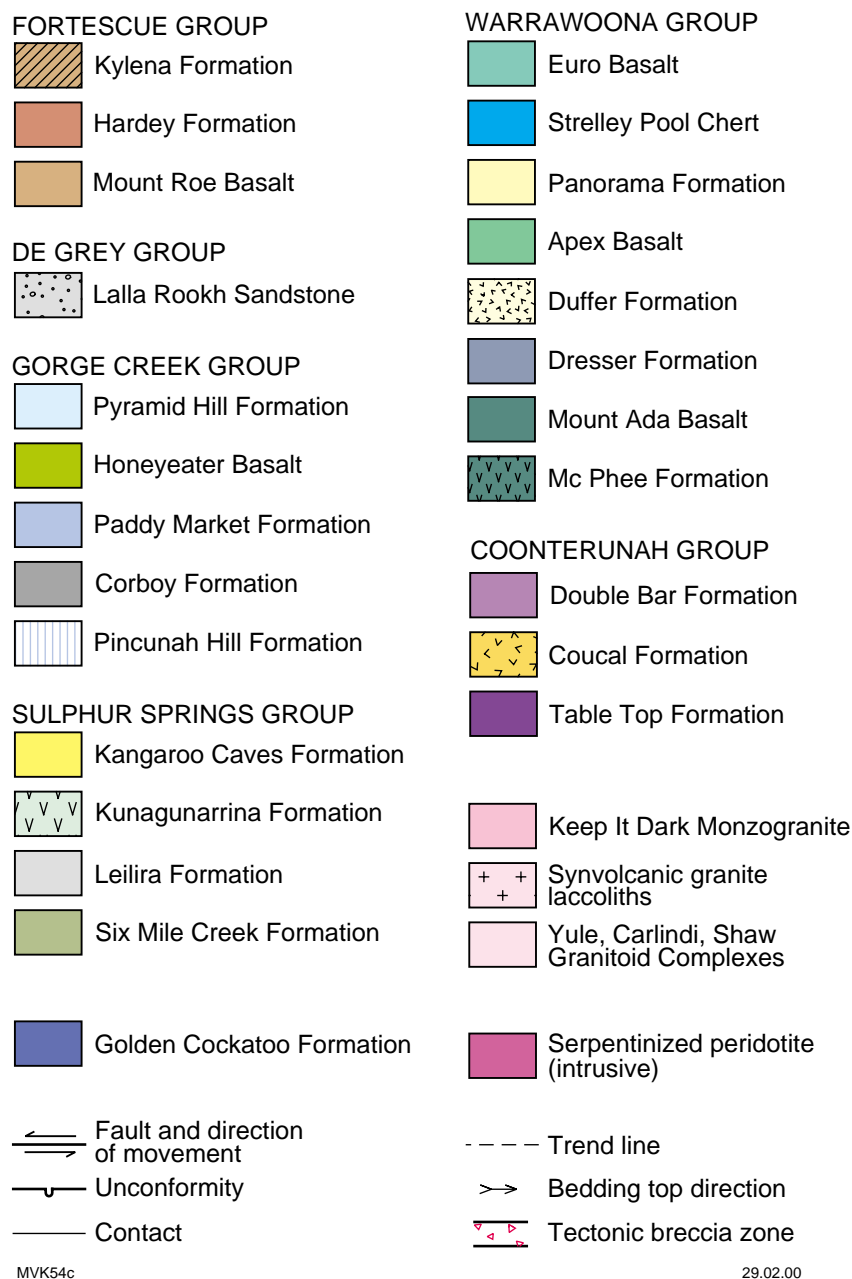
Purely structural map elements include parts of four domes and the northern half of the LWSC. These large-scale structures in turn contain, or are bound by, secondary structural elements such as folds, brittle faults, ductile

mylonite zones, and brittle-ductile deformation zones (see Van Kranendonk, 1998).

## Lithotectonic map elements

### Greenstone belts

Greenstone belts are defined as relatively well preserved tracts of commonly coherent greenstone stratigraphy bound by one or more of faults or intrusive or sheared intrusive contacts with granitoid complexes, or by unconformably overlying rocks. On NORTH SHAW, belts



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may contain one or more groups of stratified volcano-sedimentary rocks and synvolcanic granitoid intrusions of any age up to, but not including, the Fortescue Group.

Parts of eight greenstone belts fall on NORTH SHAW; the Soanesville, Western Shaw, North Shaw, Coongan, East Strelley, Pincunah, Abydos, and Panorama Belts (Fig. 6). The East Strelley and Pincunah Belts replace the northern and southern limbs respectively of Hickman's (1983) Pilgangoora Syncline, and are separated by the kilometre-wide Mount York Deformation Zone. The name Pilgangoora Syncline is reserved for a mappable synclinal fold in the northern part of the Pincunah Belt.

The Abydos Belt is the name given to a tract of amphibolite-facies supracrustal rocks infolded with granitoid rocks along the northeastern margin of the Yule Granitoid Complex, which are separated from the lower

grade rocks of the Pincunah Belt by the Numerous Scrapes Deformation Zone. The Panorama Belt is the name given to the greenstones that surround the North Pole Monzogranite within the large-scale structure known as the North Pole Dome.

### Greenstone complexes

Greenstone complexes are areas of high structural complexity in greenstones. Parts of two adjacent greenstone complexes are in the southern part of NORTH SHAW. The Emerald Mine Structural Complex, located west of the North Shaw Belt, is underlain by tightly folded mylonitic schists derived from Warrawoona and Gorge Creek Group protoliths and intrusive peridotites. This complex is separated from the adjacent Tambina Structural Complex to the west by the northerly striking South

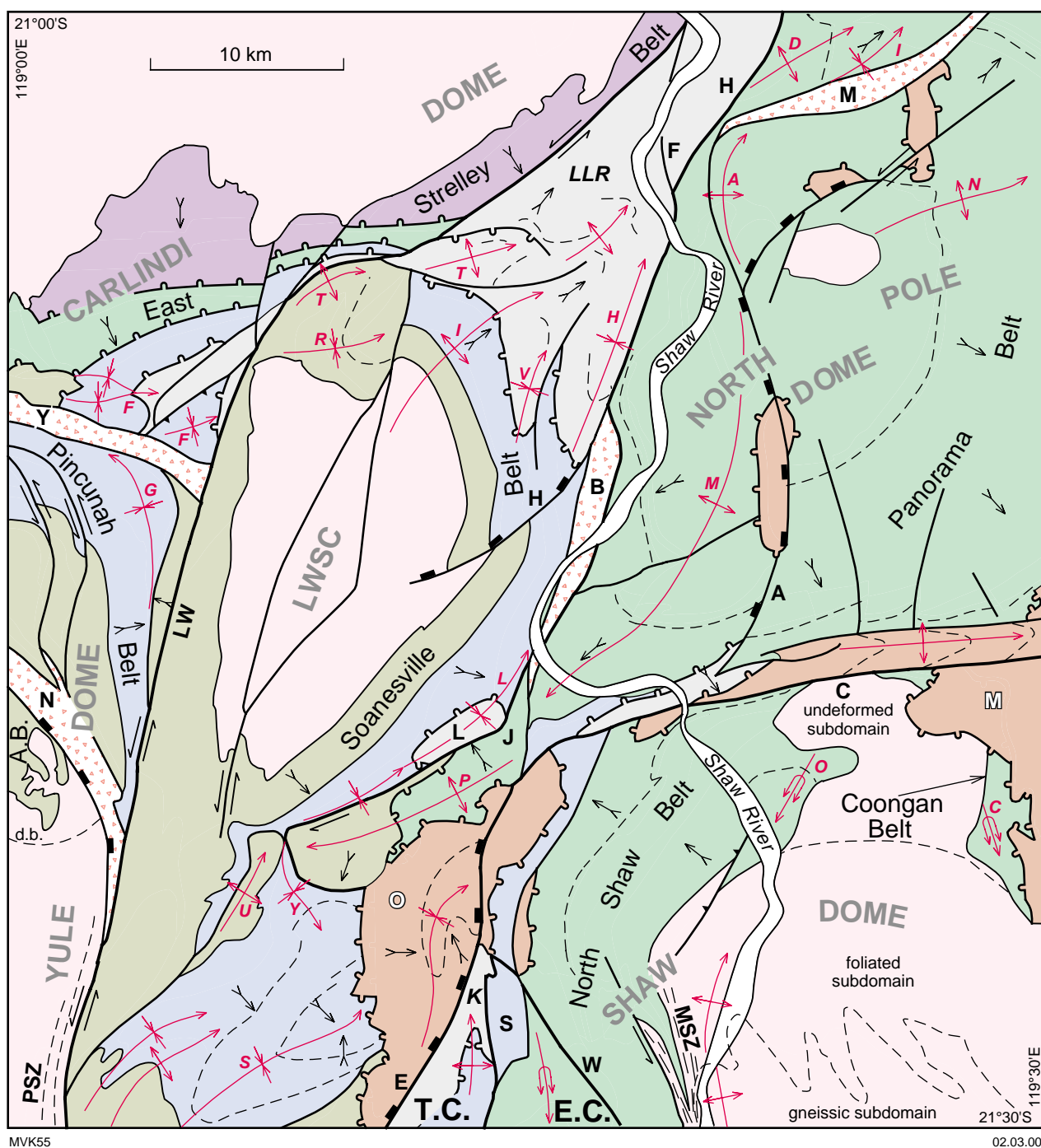





















Figure 6. Lithotectonic subdivisions of NORTH SHAW

Daltons Fault, which is composed of en echelon segments that juxtapose higher grade rocks in the east (upper greenschist to amphibolite facies) against lower grade rocks in the west. The Tambina Structural Complex contains folded and faulted, dismembered rocks of the Warrawoona and Gorge Creek Groups.

#### Synclinoria of the Lalla Rookh Sandstone

The largest occurrence of the Lalla Rookh Sandstone is located along strike northeast of the Soanesville Belt

(Fig. 6). Previously referred to as the Lalla Rookh Syncline (Hickman, 1983), or the Lalla Rookh Basin (Krapez, 1984, 1989), Van Kranendonk and Collins (1998) suggested that these rocks were originally part of a much larger basin and that the present structure is a fault-bounded trough in which only the upper parts were likely to have been deposited in situ. The complex folds and extensive faults that affect these rocks indicate that a more appropriate name is the Lalla Rookh Synclinorium. The Lalla Rookh Sandstone is also present in the Keep It Dark Synclinorium in the south-central part of NORTH SHAW and

	Fortescue Group		Trend line
	De Grey Group		Syncline
	Gorge Creek Group		Anticline
	Sulphur Springs Group		Synformal anticline
	Warrawoona Group		Fault; undefined displacement
	Coonterunah Group		Reverse fault
	Granitoid rocks		Normal fault
	Tectonic breccia zone		Strike-slip fault
			Structural domain boundary fault
			Unconformity
			Bedding top direction

**Folds (shown in red)**

A	Antarctic Anticline
C	Callina synformal anticline
D	Caldera Anticline
F	Falls Pound Syncline
G	Pilgangoora Syncline
H	Hogback Syncline
I	Sulphur Springs Anticline
L	Leilira Syncline
M	McLeods Arch
N	Normay Arch
O	Chocolate Hill synformal anticline
P	Potkoorok Anticline
R	Roadmaster Syncline
S	Soanesville Syncline
T	Strelley Anticline
U	Furies Anticline
V	Caves Syncline
U	Furies Anticline
Y	Pyramid Hill Syncline



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**Faults (marked in black)**

A	Antarctic Fault
B	Bernts Deformation Zone
C	Callina Fault
E	Escarpment Fault
F	Flying Fox Fault
H	Hogback Fault
J	Jamesons Fault
LW	Lalla Rookh – Western Shaw Fault
M	Miralga Deformation Zone
N	Numerous Scrapes Deformation Zone
S	South Daltons Fault
W	White Quartz Hill Fault
Y	Mount York Deformation Zone

**Shear zones**

MSZ	Mulgandinnah Shear Zone
PSZ	Pulcunnah Shear Zone

K	Keep it Dark Synclinorium
L	Leilira Synclinorium
LLR	Lalla Rookh Synclinorium
A.B.	Abydos Belt
E.C.	Emerald Mine Structural Complex
T.C.	Tambina Structural Complex
	Marble Bar Outlier
	Olympic Pool Outlier
d.b.	dome-and-basin subdomain
LWSC	Lalla Rookh – Western Shaw Structural Corridor

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in the tight syncline between the Panorama and North Shaw Belts (Van Kranendonk and Morant, 1998).

**Granitoid complexes**

NORTH SHAW is underlain by parts of the Shaw, Yule, and Carlindi Granitoid Complexes, all of which are composed of several distinct, mappable plutonic or gneissic units that together form a coherent whole. Granitoid complexes are in contact with greenstone belts or greenstone complexes across intrusive or sheared intrusive contacts. Structural and gravity data combine to suggest that the domal granitoid complexes link up at depth beneath intervening greenstones, and thus represent structural domes of a

continuous, mid-crustal granitoid layer beneath upper crustal greenstones (Hickman, 1984; Van Kranendonk and Collins, in prep.). This interpretation is supported by the fact that the different complexes are composed of similar-aged suites of plutonic rocks, as described below.

**Granitoid plutons**

In contrast to individual components of granitoid complexes, the North Pole Monzogranite, Strelley Granite, and Keep It Dark Monzogranite are discrete intrusions within greenstones. Despite a difference in the age (c. 3459 and c. 3240 Ma respectively) and orientation of preservation of the North Pole Monzogranite and Strelley

Granites (plan versus cross sectional respectively), both have convex upper contacts parallel to bedding in the overlying supracrustal rocks and are contemporaneous with felsic volcanic components of the overlying greenstones. Based on this combination of geometry and geochronology, these plutons are interpreted to represent synvolcanic laccoliths (Van Kranendonk, 1997) and are thus listed as component members of the lithostratigraphy (Table 1). The Keep It Dark Monzogranite, in contrast, is a discordant pluton that cuts across steeply dipping bedding and regional foliations in supracrustal rocks of the North Shaw Belt, is not directly associated with supracrustal rocks, and is thus not included in the lithostratigraphic table.

## Structural map elements

### Domes

NORTH SHAW is underlain by parts of the Carlindi, Yule, Shaw, and North Pole Domes (Fig. 6). Each dome comprises a granitoid intrusion or granitoid complex in their core, whose name is used for the dome (Hickman, 1983), and an outer rind of greenstones distributed in one or more belts; for example, the North Pole Dome is underlain by the North Pole Monzogranite and Panorama Belt. A fifth dome comprising the Strelley Granite and overlying volcano-sedimentary rocks of the Sulphur Springs and Gorge Creek Groups is within the Soanesville Belt, but has been flipped onto its side as a result of regional D<sub>3</sub> transpression (Van Kranendonk and Collins, 1998). Granitoid complexes coring the domes may be internally subdivided into structural subdomains; for example, the dome-and-basin subdomain of the Yule Granitoid Complex, and the undeformed, foliated and gneissic subdomains of the Shaw Granitoid Complex (Fig. 6).

### Lalla Rookh – Western Shaw Structural Corridor (LWSC)

A northerly striking zone of complexly deformed greenstones located between the Carlindi and Yule Domes in the west, and the Shaw and North Pole Domes in the east is referred to as the Lalla Rookh – Western Shaw Structural Corridor (LWSC; Van Kranendonk, 1998; Van Kranendonk and Collins, 1998; see Fig. 6). The LWSC is underlain by the Soanesville Belt, the Tambina Structural Complex, and the two synclinoria of the syntectonic Lalla Rookh Sandstone. The LWSC is characterized by northeasterly plunging folds, but may be subdivided into a southern area of tightly folded greenstones and a central area of little-deformed greenstones in the strain shadow around the Strelley Granite. The LWSC is bounded to the west by a curvilinear sinistral fault and the ductile Pulcunnah Shear Zone, and in the east by an echelon fault strands, some of which have a dextral component of displacement (Van Kranendonk and Collins, 1998).

### Folds and curvilinear high-strain zones

In the nomenclature scheme devised by Van Kranendonk (1988), the term syncline (or anticline) is used only to

describe a specific type of fold structure, rather than an area of greenstones as previously done by Hickman (1983). Synclines, as with other fold geometries, can form within any of the granitoid complexes, greenstone belts, greenstone complexes, synclinoria of late-tectonic clastic rocks, and in the Fortescue Group.

Three types of curvilinear, high-strain zones are recognized; narrow brittle faults (<5 m wide), wider brittle–ductile deformation zones (100–1500 m wide), and broad, ductile shear zones within granitoid rocks (>2 m wide; Fig. 6). Faults are commonly sharp discontinuities between different rock types, and may be silicified, or brecciated, or both. The most significant fault is the northerly striking, curvilinear sinistral fault that transects the map area just west of the Strelley Granite. This fault forms the northwestern boundary of the Lalla Rookh Synclinorium, curves around the northern end of the Strelley Granite where it has numerous splays, extends due south where it forms the boundary between the Soanesville and the East Strelley and Pincunah Belts, and then continues south as the ductile Pulcunnah Shear Zone along the margin of the Yule Granitoid Complex (Van Kranendonk and Collins, 1998). This fault was previously referred to as the Western Boundary Fault of the ‘Lalla Rookh strike-slip orogen’ (Krapez, 1989; Eriksson et al., 1994), the latter of which was an informal name subsequently shown to be inappropriate (Van Kranendonk and Collins, 1998) and superseded by the LWSC (Van Kranendonk, 1998).

Ductile shear zones are developed within granitoid rocks of the western margin of the Shaw Granitoid Complex and the eastern margin of the Yule Granitoid Complex, known as the Mulgandinnah and Pulcunnah Shear Zones respectively (MSZ and PSZ in Fig. 6). Deformation zones consist of large-scale tectonic breccia hosted by a matrix of ultramafic (typically talc–carbonate(–chlorite)) schist. Breccia fragments within these zones locally reach up to 2 km in length and consist of disaggregated and folded, schistose to undeformed packages of a variety of supracrustal rocks derived from adjacent greenstones. The ultramafic schist surrounding the breccia is derived both from high-Mg basalts within the stratigraphy, and locally from intrusions emplaced within actively deforming zones.

## Archaean rocks

The revised lithostratigraphy of NORTH SHAW is presented in Figure 5 and Table 1. Six groups and 25 formations of supracrustal rocks have been recognized. The component phases of the granitoid complexes are presented in Table 2.

### Coonterunah Group

Buick et al. (1995) identified the Coonterunah succession of dominantly tholeiitic mafic volcanic rocks around the southern flank of the Carlindi Granitoid Complex as distinct from the Warrawoona Group on the basis of a local, high-angle unconformity and an age of c. 3515 Ma on a felsic volcanic component. These rocks were

Table 1. Revised Archaean stratigraphy of NORTH SHAW

Deformation event	Group	Age (Ma)	Formation	Map code	Lithology
D <sub>4</sub> ~	<b>Fortescue Group</b>	c. 2747 <sup>(b)</sup>	Kylena Formation	<i>AFk</i>	Andesitic tuff, felsic tuff, tuffaceous sandstone, thick-bedded vesicular basalt
		c. 2756 <sup>(a)</sup>	Hardey Formation	<i>AFh</i>	Conglomerate, shale and sandstone, local tuff and agglomerate
		c. 2772 <sup>(a,b)</sup>	Mount Roe Basalt	<i>AFr</i>	Basalt, locally glomeroporphyritic; agglomerate
D <sub>3</sub>	<b>De Grey Group</b>	c. 2950 <sup>(b)</sup>	Lalla Rookh Sandstone	<i>ADl</i>	Conglomerate, sandstone, minor shale
D <sub>2</sub>	<b>Gorge Creek Group</b>	Soanesville Subgroup	‡ Pyramid Hill Formation	<i>AGy</i>	Banded iron-formation (BIF)
			‡ Honeyeater Basalt	<i>AGh</i>	Basalt
			‡ Paddy Market Formation	<i>AGp</i>	Fe-shale; locally silicified to grey and white chert
			‡ Corboy Formation	<i>AGc</i>	Sandstone, mudstone, minor conglomerate
			‡ Pincunah Hill Formation	<i>AGi</i>	Fe-shale, BIF, interbedded sandstone, felsic volcanic rocks
	‡ <b>Sulphur Springs Group</b>	3238–3235 <sup>(c)</sup>	‡ Kangaroo Caves Formation	<i>ASc</i>	Differentiated volcanic–volcaniclastic pile of mainly tholeiitic magmatic affinity varying from basalt to rhyolite, with comagmatic granite; includes chert, local polymictic megabreccia and iron formation, and calc-alkaline rhyodacite
			‡ Kunagunarrina Formation	<i>ASk</i>	Pillow basalt, komatiite, high-Mg basalt, chert
			‡ Leilira Formation	<i>ASl</i>	Wacke and intercalated rhyolite, sandstone, mudstone, chert
			‡ Six Mile Creek Formation	<i>ASx</i>	Mafic volcanic rocks, minor volcaniclastic rocks
			‡ Golden Cockatoo Formation	<i>Aj</i>	Cherty silicate-facies iron formation, rhyolite, quartzite and interbedded metapelite
D <sub>1</sub>	<b>Warrawoona Group</b>	3458 <sup>(e)</sup>	Euro Basalt	<i>Awe</i>	Pillow basalt and chert (‡Potkoorok Member)
					Pillow tholeiitic basalt and chert (*Miralga Creek Member)
			* Strelley Pool Chert	<i>AWs</i>	Komatiite and high-Mg basalt (‡Cloisters Member)
			Panorama Formation	<i>AWp</i>	Quartzite and chert, stromatolites
			Apex Basalt	<i>AWa</i>	Felsic lavas, tuffs and tuffaceous sandstone
	‡ <b>Coonterunah Group</b>	3471–3463 <sup>(d,e,g)</sup>	Duffer Formation	<i>AWd</i>	Carbonate-altered basalts, chert
			Dresser Formation	<i>AWr</i>	Dacitic tuff, agglomerate, and lava
			Mount Ada Basalt	<i>AWm</i>	Blue–black and white layered chert, barite, carbonate, mafic volcanic rocks
D <sub>1</sub>	‡ <b>Coonterunah Group</b>	3515 <sup>(f)</sup>	‡ Coucal Formation	<i>Aoc</i>	Basalts and cherts
			‡ Table Top Formation	<i>Aot</i>	Talc–chlorite schist, chert, BIF, pelite
			‡ Double Bar Formation	<i>Aod</i>	Basalts and volcanogenic sedimentary rocks

**NOTES:** \* Redesignation of existing name; ‡ New group/formation/member name and associated type area; † New type area designated

**SOURCES:** (a) Arndt et al. (1991); (b) Wingate (1999); (c) Unpublished SHRIMP U–Pb zircon data (Buick, R., University of Sydney, 1997, pers. comm.); (d) McNaughton et al. (1993); (e) Thorpe et al. (1992a); (f) Buick et al. (1995); (g) Nelson (in prep.)

upgraded to group status by Van Kranendonk and Morant (1998), who recognized three component formations including, from base to top, the Table Top, Coucal, and Double Bar Formations (Table 1). The group is preserved with a maximum intact stratigraphic thickness of 5.9 km (between 713000E and 717000E at 7665000N), but individual formations reach a maximum aggregate thickness of 6.7 km (Table Top Formation = 3.45 km; Coucal Formation = 1.38 km; Double Bar Formation = 1.87 km). The lower contact of the group is an intrusive

one with the Carlindi Granitoid Complex, whereas its upper contact is either an unconformity with the Strelley Pool Chert, or a faulted contact against the Lalla Rookh Synclinorium.

Glikson et al. (1986) presented major and trace element data from a north–south transect across the Coonterunah Group (then referred to as the Talga Talga Subgroup) in the East Strelley Belt, north of Strelley Gorge (AMG 170670–170620). The Table Top Formation

**Table 2. Component suites of granitoid complexes on NORTH SHAW**

Age (Ma)	Carlindi Granitoid Complex	Yule Granitoid Complex	Shaw Granitoid Complex
2851 ± 2		Abydos Monzogranite (on WODGINA)	Cooglegong and Spear Hill Monzogranites <sup>(a,c)*</sup> (on TAMBOURAH)
2930 ± 3		Woodstock Monzogranite <sup>(a)*</sup> (AgYwo)	Mulgandinnah Monzogranite <sup>(a)*</sup> (AgSmu)
c. 3240		Kavir Granodiorite <sup>(b)*</sup> (AgYka)	
3450–3400			Leucogranite <sup>(b,e)*</sup> (AgSl)
3465 ± 10	Leucogranite <sup>(e)*</sup> (AgLi) (on WODGINA)	Petroglyph Gneiss <sup>(b)*</sup> (AgYpe)	Coolyia Creek Tonalite <sup>(c,f)*</sup> (AgSco)
≥3480	Biotite granodiorite <sup>(e)*</sup> (AgLm) Wilson Well Gneiss (AgLwi)		North Shaw Tonalite <sup>(d)*</sup> (AgSns)

**NOTES:**

\* Dated components

**SOURCES:**

(a) Nelson (1998); (b) Van Kranendonk, unpublished SHRIMP U–Pb zircon data; (c) Bickle et al. (1989); (d) McNaughton et al. (1988); (e) Nelson (1999); (f) Nelson (in prep.)

(samples 69911–69934; Glikson et al., 1986) consists predominantly of tholeiitic basalt, with 4.3–7.8% MgO, Mg# = 40–51, and 11–15% Al<sub>2</sub>O<sub>3</sub>. One high-Mg basalt (10.4% MgO), one andesite, and three high-Al basalts (17–18% Al<sub>2</sub>O<sub>3</sub>) were also analysed. The Coucal Formation was found to consist of tholeiitic basalts, andesite, and dacite (samples 69935–69940, 69945, 69966, 69967; Glikson et al., 1986). Tholeiitic basalts are also in the Double Bar Formation, but they have higher MgO values of 7.65–8.15% than those of the Table Top Formation (samples 69941–69944; Glikson et al., 1986).

## Table Top Formation (Aot)

The Table Top Formation, named after Table Top Well (AMG 145699), is predominantly composed of fine-grained doleritic tholeiitic basalts and subvolcanic doleritic to gabbroic intrusions (Aot), with subordinate pillowed and variolitic flows, gabbro, and thin flows of high-Mg basalt near its base. The Carlindi Granitoid Complex intrudes out the lower parts of the formation, whereas the top of the formation is conformable with the overlying rocks of the Coucal Formation. Bedding is only rarely recognizable in this thick-bedded to massive unit. The massive nature of this rock is due, in part, to metamorphic recrystallization as a result of its proximity to the Carlindi Granitoid Complex, immediately adjacent to which basaltic rocks are affected by a hornfelsic metamorphic aureole up to 100 m wide. Metabasalts are fine- to medium-grained intergrowths of actinolite, plagioclase, and opaque minerals. Feldspar laths in most cases retain their igneous shape, but are sieved with sericite and have recrystallized margins that are intergrown with actinolite.

## Coucal Formation (Aocbi, Aocf, Aoci)

The base and top of the Coucal Formation are marked by one, two, or three 2–10 m-thick beds of centimetre-layered, black and white banded cherty iron-formation

(Aoci). The beds vary from evenly layered rocks with 1–2 cm-thick black and white layers, to millimetre-centimetre layered black, white, and less common red layers, to solid black rock with sparse, irregularly spaced, white chert layers. Variations in the amount of clear white chert between these units suggest that the formation was emplaced as sills into black, magnetite iron-formation.

Along the southern margin of the Carlindi Granitoid Complex, the Coucal Formation contains up to 1 km of massive, fine-grained doleritic andesite and basalt (Aocbi). In thin section, this unit has an interlocking texture of plagioclase laths within a fine-grained mat of actinolite and opaque minerals. Plagioclase is strongly sieved by actinolite, carbonate, and zoisite and has recrystallized boundaries. Scattered throughout the rock are large (1–2 mm), monocrystalline quartz crystals with subhedral outlines intergrown with plagioclase and the recrystallized mafic matrix.

Felsic volcanic rocks (Aocf) include massive, highly amygdaloidal dacite and subordinate dacite porphyry, and hyaloclastite and pumiceous rhyolite. The rocks, affected by metamorphic recrystallization and carbonate–sericite alteration, are commonly fine grained in thin section. Amygdales in dacite are typically filled by carbonate and epidote. Buick et al. (1995) dated a sample of brecciated hyaloclastic rhyolite from this formation (sample 70601, AMG 298684) at 3515 ± 3 Ma.

## Double Bar Formation (Aod)

The Double Bar Formation includes fine-grained tholeiitic basalt, pillowed tholeiitic basalt, and interbedded basaltic volcaniclastic rocks (Aod). Volcaniclastic units include fine-grained mafic tuff and in one locality, a 2 m-thick bed of crystal, lithic vitric tuff. In thin section, the rocks retain a basaltic texture of interlocking plagioclase laths, but all mafic minerals are recrystallized to a metamorphic mineral assemblage of chlorite or actinolite–chlorite–zoisite–

epidote – opaque minerals. Some flows contain up to 50% of stubby, rectangular crystals (probably originally clinopyroxene, but now carbonate, chlorite, and plagioclase) in a chlorite–plagioclase matrix with coarse clots of opaque minerals. The crystal, lithic vitric tuff contains elongate, but stubby plagioclase crystals and crystal fragments, fragments of doleritic basalt, pumiceous rhyolite and chert, and shards of altered volcanic glass (now chlorite and carbonate).

## Warrawoona Group

### McPhee Formation (*Awhba*, *Awhbc*, *Awhu*, *Awhc*)

The McPhee Formation only outcrops adjacent to the Shaw Granitoid Complex, and reaches a maximum of about 3 km thick in the southeastern Coongan Belt. The lower contact of the formation is an intrusive one with granitoid rocks of the Shaw Granitoid Complex, whereas its upper contact is conformable with the overlying Mount Ada Basalt. Rock types include fine- to medium-grained, amphibolite-facies metabasalts (*Awhba*), medium- to coarse-grained talc–chlorite–carbonate schists (*Awhu*), and carbonate-altered, chloritic metabasalts (*Awhbc*). Primary volcanic textures have been destroyed by penetrative alteration and metamorphic recrystallization. The top of the McPhee Formation is capped by a thin unit of finely bedded, red, black, and white layered banded iron-formation (*Awhc*).

### Mount Ada Basalt (*Awmb*, *Awmba*, *Awmbc*, *Awmc*)

The Mount Ada Basalt is a typically massive unit of tholeiitic basalt and dolerite intrusions (*Awmb*), between 3 and 4 km thick, which outcrops in the Panorama, North Shaw, and Coongan Belts. The lower contact is largely an intrusive one with granitoid rocks of the Shaw Granitoid Complex and North Pole Monzogranite, but it is locally conformable on rocks of the McPhee Formation. The upper contact is conformable with the overlying Dresser Formation (Panorama Belt) or Duffer Formation (North Shaw Belt). Volcanic textures are well preserved in outcrops located just west of the Dresser Mining Centre (AMG 496596), where interbedded, metre-thick flows of weakly vesicular, pillowed, and massive tholeiitic basalt are exposed. A spectacular outcrop of pillow breccia and basaltic agglomerate is visible on the trackside heading into the Dresser Mining Centre (AMG 518596). In this 300 m-thick unit, angular pillow fragments — locally with vesicular, chilled rims — vary from 10 cm up to 40 cm in diameter, and lie within a fine-grained matrix of basaltic glass shards. These rocks are extensively altered by dark-brown weathered carbonate, as are basaltic rocks (*Awmbc*) that typically underlie the Dresser Formation where they are transected by a boxwork of chert–barite dykes (see below) and depleted of SiO<sub>2</sub>.

A kilometre-thick rind of metabasalt that lies with intrusive contact against the western margin of the Shaw

Granitoid Complex was metamorphosed to medium-grained, foliated and lineated amphibolites (*Awmba*). Original textures in this rock have been obscured by complete recrystallization. Away from the granitoid complex, tholeiitic basalts of the Mount Ada Basalt are extensively pillowed and recrystallized to upper greenschist facies assemblages of chlorite–actinolite–epidote–carbonate and sericite after plagioclase. Local areas of basalt are tightly folded, schistose, and heavily carbonate altered (*Awmbc*).

Cherts are commonly absent from the Mount Ada Basalt, in contrast to other basaltic units of the Warrawoona Group, and in this way, it is similar to the Table Top Formation of the Coonterunah Group. One 10 m-thick chert (*Awmc*) northeast of the Shaw Granitoid Complex in the Coongan Belt is a grey-weathered, medium-grained quartzite with glassy fresh surfaces. The quartzite is thoroughly recrystallized to a polygonal granoblastic texture, but faint evidence of centimetre-scale and local graded bedding is preserved, which suggests that way up is to the north. Approximately 500 m north of this unit is a 50 cm-thick unit of coarse-grained, pink and light-green calcite marble with thin sandy interbeds (AMG 570386; not shown on map).

### Dresser Formation (*Awrb*, *Awrc*, *Awrct*)

The Dresser Formation formed only in the Panorama Belt of the North Pole Dome, where it consists of interbedded units of chert–barite (*Awrc*) and pillowed to massive mafic volcanic rocks (*Awrb*) sandwiched between the lower Mount Ada Basalt and overlying Apex Basalt. The Dresser Formation varies in thickness, reaching a maximum of about 1000 m at the Dresser Mining Centre, where there are three beds of chert–barite interlayered with mafic volcanic rocks. However, these beds taper laterally along strike to the north and south. In other parts of the belt, two or, in places, only one chert–barite unit is present. Individual chert–barite horizons reach a maximum thickness of about 40 m, and thin to a minimum of less than 5 m. Individual beds and lithofacies assemblages are discontinuous along strike. The chert–barite horizons are separated by highly altered, massive to pillowed mafic volcanic rocks (*Awrb*).

The chert–barite consists mainly of centimetre-layered, white and grey-blue or black chert, and laminated to coarsely crystalline barite. In detail, however, Buick and Dunlop (1990) recognized up to 22 sedimentary lithofacies, of which six are partly or wholly chemical in origin. Lithofacies have been divided into four facies assemblages, including a lower arenaceous assemblage of dominantly volcanogenic sand, one or more horizons of volcanogenic mud (lutaceous assemblage), a carbonate facies, and a sulfate facies. Original carbonate mineralogy is largely pseudomorphed by sulfates, including gypsum and barite. Both sedimentary and diagenetic barite have been recognized, the former as fine-grained detrital material, the latter forming coarse-bladed crystals pushing up sedimentary bedding (Fig. 7a). A summary of the main rock types and distribution of lithofacies assemblages originally described by Buick and Dunlop (1990) is presented in Barley (1993).



Figure 7. Features of the Dresser Formation: a) diagenetic barite crystals that have pushed up and folded sedimentary bedding; b) a small, domal stromatolite showing characteristic porous carapace and marginal sand drapes; c) wavy-laminated bacterial mats and domal stromatolites. Pen knife is 9 cm long

As noted previously, cherts of the Dresser Formation contain Earth's oldest stromatolites and possible microfossils (Dunlop et al., 1978; Walter et al., 1980; Buick et al., 1981; Groves et al., 1981; Walter, 1983). At one locality south of Panorama Homestead (Hickman, 1990), bacterial mats up to 4 cm thick overlie the tops of ripple-marked sandstones. The sequence dips at 40°E and is gently folded, but the rocks are otherwise undeformed. The bacterial mats are black with a dark-red weathering surface, finely laminated ( $\leq 1$  mm), and contain irregular crenulations and low-amplitude domes ( $< 1$  cm amplitude for 5 cm diameter domes; Fig. 7b). Higher up, similar finely laminated material outlines broad, low-amplitude domes 30 cm in diameter by 10 cm in height. Immediately overlying the finely laminated material are high-amplitude domes (10–15 cm wide by 10–15 cm high) that have conical to rounded crests. Several other types of stromatolite morphology are in this outcrop, including elliptical domes with moderate amplitudes (12 cm  $\times$  6 cm  $\times$  3–4 cm high) that are identical in shape to those in the Tumbiana Formation of the Fortescue Group, and other columnar stromatolites that have identical shapes and internal structures (e.g. gas bubbles) as extant forms in Shark Bay (Fig. 7c).

Barite was introduced through a network of chert–barite dykes (*AWrct*), which cut through underlying volcanic rocks of the Mount Ada Basalt. These dykes are distributed in a radial pattern out from the core of the North Pole Dome, and cut up along faults that disrupt the underlying stratigraphy (see geological map). These faults were subsequently reactivated and also cut up-section above the Dresser Formation. The dykes are most prominent east and southeast of the North Pole Monzogranite and reach a maximum of 25 m width. The dykes contain blue-black chert and coarse-bladed barite, with additional, local ferruginous gossan.

Nijman et al. (1998) showed that the dykes occupy growth faults and barite was deposited as mounds over the sites where the growth faults breached the surface. The presence of barite pebble conglomerates flanking the mounds, and of onlap relationships of arenaceous and lutaceous lithofacies against the barite mounds, indicate the formation and deposition of barite as an exhalative product emanating from a magma chamber in the core of the North Pole Dome.

### Duffer Formation (*AWd*, *AWda*, *AWds*)

Quartz–sericite schist (*AWd*) and foliated, but well-preserved felsic agglomerate (*AWda*) of the Duffer Formation outcrop prominently in the North Shaw Belt, immediately underlying the Panorama Formation. The more massive unit is commonly well foliated and rarely contains quartz phenocrysts. The agglomerate, or welded ignimbrite, is best exposed west of Dalton Creek (AMG 395357) where it contains angular, cusate, and rounded, tectonically flattened and lineated cobbles and pebbles of a variety of felsic lithology including weakly porphyritic dacite, massive purplish-coloured rock (chlorite-altered ?rhyodacite), and, in places, weakly vesicular, pumiceous felsic volcanic rock. Bedding is at a metre scale or more in this unit, and contains local beds

or lenses of sedimentary conglomerate, which consist of rounded quartzite cobbles. Both the massive schist and agglomeratic units are typically highly silicified.

The felsic volcanic unit (*AWda*) in the tight syncline adjacent to the southwestern margin of the Shaw Granitoid Complex (AMG 390265) is texturally different from the type agglomerate to the north and west. It is a heterogeneous, fine-grained, siliceous unit with a strong transposition foliation that overprints a prominent centimetre-scale layering, which varies in intensity and degree across strike. The layering is typically discontinuous and does not represent bedding, as it has ragged edges. In places, the layering is composed of highly elongate, lensoid patches of coarser textured material surrounding a slightly darker coloured core of chlorite after biotite. The origin of the lensoid patches in this unit is interpreted to represent lithophysae in a densely welded, rheomorphic ignimbrite (McPhee et al., 1993, plate 25-1) that partially recrystallized in an irregular pattern due to its own heat, and was then strongly deformed and recrystallized during intrusion-doming of the Shaw Granitoid Complex. An igneous zircon age of  $3466 \pm 3$  Ma from this unit confirms that it is part of the Duffer Formation (see **Geochronology**, p. 73).

Three thin slivers of conglomerate and cherty metasedimentary rock of inferred Duffer Formation age (*AWds*) formed in the southernmost extension of the Carlindi Granitoid Complex (AMG 080620 and 086623), a high-level quartz-feldspar porphyry dated at  $3467 \pm 4$  Ma (Buick et al., 1995). North-south subvertical bedding in the slivers contrasts sharply with the regional east-west strike and steep dip of bedding in the Coonerunah and Warrawoona Groups in this area (Fig. 8). The easternmost sliver is 1–2 m thick and contains sills of black chert intruding grey sandstone and conglomerate, in which graded bedding indicates top to the east. In contrast, the western slivers contain felsic agglomerate, green, fuchsite-bearing spherulitic to tuffaceous felsic volcanoclastic rocks, sandstone, and polymictic cobble conglomerate in which scours and graded bedding indicate top to the west.

### Apex Basalt (*Awab*, *Awabc*, *Awabm*, *Awaan*, *Awac*)

The Apex Basalt outcrops within the Panorama Belt between the Dresser and Panorama Formations. The unit

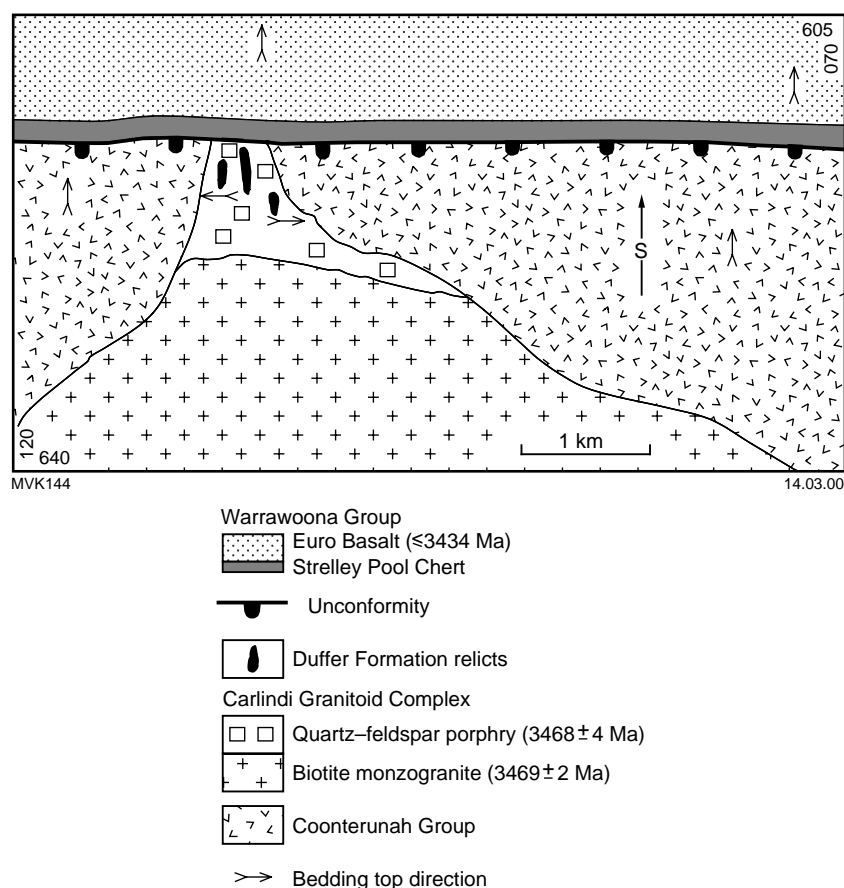


Figure 8. Simplified geological sketch showing the relationship between volcaniclastic slivers interpreted to be of the Duffer Formation within high-level quartz-feldspar porphyry in the southern promontory of the Carlindi Granitoid Complex. Note the opposite facing directions within the slivers and their high angle to adjacent rocks of the Coonerunah Group and unconformably overlying rocks of the Warrawoona Group

is dominantly underlain by extensively pillowed basalt, with both tholeiitic (*AWab*) and less voluminous high-Mg components (*AWabm*). Volcanic textures are well preserved throughout, although the rocks may be locally transected by a penetrative foliation. Exceptional exposures of pillows are common, showing vesicular rims, chilled margins, concentric cooling cracks, intrapillow hyaloclastite breccia, pillow tails, and even pillow breccia. Doleritic flows and intrusions are also common, as are rare doleritic dykes.

Extensively carbonate altered, high-Mg basalts (*AWabc*) lie immediately above the Dresser Formation in the southeastern part of the Panorama Belt (AMG 533575), and in the northwest part of the belt beneath the informally named Panorama volcano (AMG 457726 and 460690). In these places, carbonate alteration forms both at the grain scale and as intrusive dykes, but does not obscure original volcanic textures everywhere.

In the lower middle part of the formation is a 5–20 m-thick unit of chert, felsic volcanoclastic rocks, and carbonate, herein named the Antarctic Creek Member (*Awaan*). The type section of this member is 2.7 km southeast of the Dresser Mining Centre (AMG 548574) and consists of a number of distinct beds (Fig. 9). Here, the member consists of a lower 2 m of centimetre-layered, black, white, and jasperlitic chert. This is overlain by 2–3 m of felsic volcanoclastic rocks, including a lower, 30 cm-thick unit of porcellanite and then well-bedded (2–15 cm), commonly graded volcanoclastic sandstones and tuffs with beds of accretionary lapilli in a matrix of amorphous blue-grey silica. Overlying this is 20–30 cm of jasper and then a 15–20 cm-thick unit of finely bedded, grey shale. Above this lies 1–2 m of delicate, hummocky cross-stratified to thinly plane bedded carbonate–quartz–chlorite rock, which passes up into 3 m of massive carbonate–quartz–chlorite rock. The massive carbonate-bearing unit is capped by 1.5 m of 5–10 cm-bedded, carbonate-bearing siltstones, and this passes conformably up into weakly carbonate-bearing felsic tuffs. These felsic, carbonate-bearing rocks are conformably overlain by well-preserved pillowed basalts.

In thin section, the carbonate-bearing rocks have unusual textures. Carbonate and quartz are equally and evenly distributed throughout the rock, with about 5% of chlorite. The carbonate is in two forms; as large, clear, and locally twinned crystals mantled by an elliptical rind of inclusion-rich carbonate, and as smaller rhombs or groups of rhombs in quartz. Quartz has long and sutured grain boundaries and forms polycrystalline patches throughout the rock. Sparse chlorite forms irregular, small patches at the contacts between quartz and carbonate. The origin of carbonate-bearing rocks in this member is uncertain. There appears to be a mixture of textural relationships, which suggests that the carbonate was either introduced as a replacement assemblage, or deposited at the same time as quartz in the same rock. The stratabound nature of the carbonate horizon and presence of little-altered felsic volcanoclastic rocks above and below suggests that the carbonate may not be solely of replacement origin.

The Apex Basalt contains several thin (<10 m thick) horizons of chert, shown on the map as undivided chert

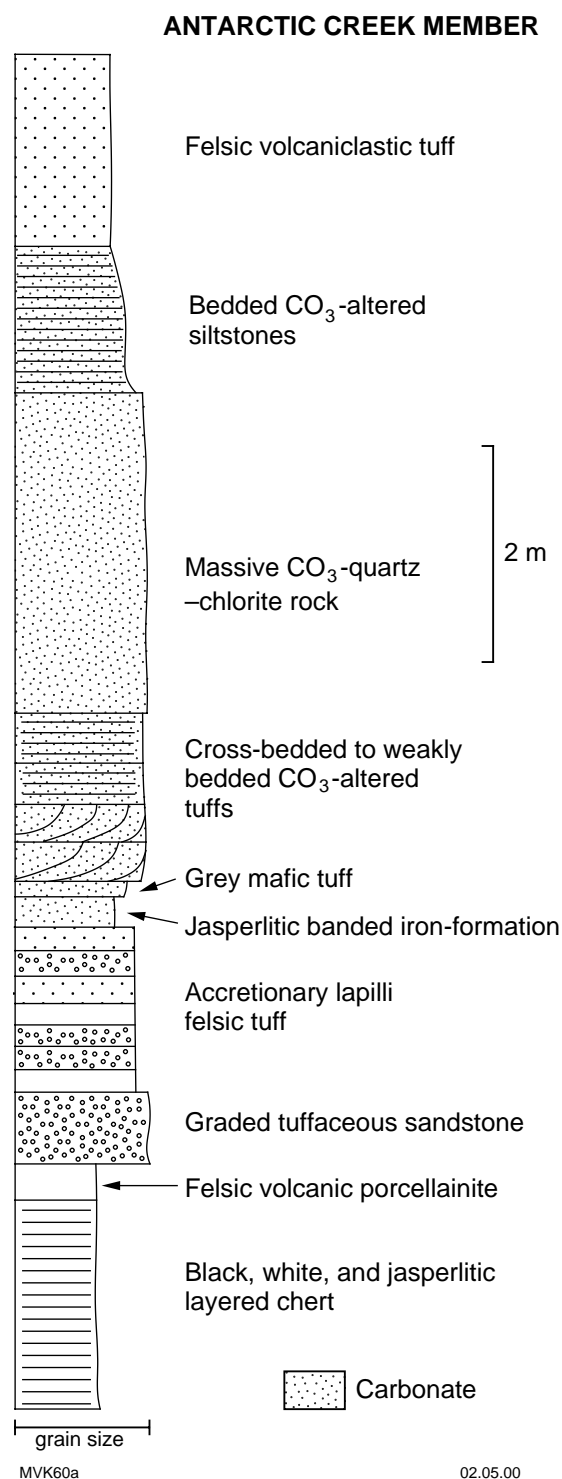


Figure 9. Stratigraphic section through the Antarctic Creek Member of the Apex Basalt in the Panorama Belt showing the interbedded nature of felsic volcanoclastic rocks and carbonate-bearing rocks

(*AWac*). The horizons include several compositional varieties, specifically grey chert with minor interbedded mafic volcanic rocks, ferruginous chert with minor interbedded banded iron-formation and clastic sedimentary rocks, silicified shale, sandstone and locally mafic tuff with local

jasper, and centimetre-layered, grey and white chert. At one locality (AMG 596635), Lowe and Byerly (1986) described bedded chert with impact-derived spherules.

### **Panorama Formation (*Awpcj*, *Awpx*, *Awpr*, *Awpp*, *Awpt*, *Awpca*, *Awpsc*, *Awps*, *Awpi*)**

The Panorama Formation is primarily composed of felsic volcanoclastic rocks with subordinate felsic lavas and chert (Hickman, 1983). DiMarco and Lowe (1989a,b) described the formation in several places throughout the east Pilbara. The type area is along Panorama Ridge (Hickman, 1983) in the southern part of the North Pole Dome, but the formation is also present in the northern part of the dome and to the east on MARBLE BAR. Thorpe et al. (1992a) dated a massive rhyolite from 20 m above the base of the formation at  $3458 \pm 2$  Ma. The Panorama Formation is identical in age to the North Pole Monzogranite exposed in the core of the North Pole Dome (Thorpe et al., 1992a), indicating that there is a synchronous relationship between volcanism and plutonism.

The Panorama Formation in the North Pole Dome comprises several mappable units. At the base of the formation, along Panorama Ridge, is a massive, orange-weathering rhyolite that locally contains small feldspar laths and sparse quartz phenocrysts (*Awpr*). In thin section, the rock contains abundant corroded and locally embayed phenocrysts of inverted  $\beta$ -quartz, up to a millimetre in diameter, set in a matrix of finely recrystallized quartz and feldspar that may contain sericite-altered plagioclase laths. There are also a few scattered muscovite phenocrysts, and some parts contain 1–2% leucoxene (after ilmenite).

In the northwest, adjacent to the Hogback Fault, a volcanic vent is preserved in plan view between splays of the Miralga Deformation Zone (Fig. 10). The vent is rimmed by up to 40 m of jaspilitic banded iron-formation, which is bedded on a centimetre scale (*Awpcj*). The banded iron-formation dips moderately inwards towards the vent and is cut by it. The vent facies (*Awpx*) is an unsorted breccia of weakly quartz and feldspar-phyric felsic rock that contains 5–10% of angular fragments of jasper and fine-grained felsic rock (Fig. 11a). Jasper fragments are up to 25 cm long and commonly elongate, whereas felsic fragments are commonly more equant. A 40 cm-diameter fumarole was observed in the east-central part of the vent (Fig. 11b).

Cutting the vent facies breccia in two places are small, circular pipes of carbonate breccia (see Fig. 10). The largest of these is 20 m in diameter and dark chocolate-brown, containing sparse and irregular fragments of fine-grained felsic rock up to 35 cm in size. The smaller pipe is 1 m in diameter and packed with angular jasper and less common felsic fragments in a chocolate-brown coloured, carbonate-bearing matrix (Fig. 11c). Thin-section petrography of the jasper breccia pipe shows the matrix to be composed of granophyric quartz–feldspar and intergrown carbonate, with subordinate, irregular-shaped, sharp-walled domains of pure quartz (Fig. 12a). The pure

quartz domains contain an outer margin of radial quartz fibre bundles that passes inwards to medium-grained cores with unstrained, polygonal textures (Fig. 12b). These textures indicate primary crystallization of quartz into a supersaturated magmatic fluid (Craig and Vaughan, 1981, p. 116). The radial quartz fibre bundles nucleated on the prismatic terminations of carbonate rhombs (Fig. 12b), thereby indicating that the carbonate was present as a magmatic component of the earlier matrix phase (Van Kranendonk, 1999).

Flanking the vent of the Panorama volcano to the north (AMG 475722) is a volcanoclastic ejecta apron that has been exposed in cross sectional profile around a northeasterly facing, vertical fold as a result of  $D_3$  deformation (see **Structure**). Overlying the jaspilitic iron-formation at the base of the ejecta apron is about 50 m of decimetre- to centimetre-scale, interbedded carbonate, jasper chert, and felsic tuff (Fig. 13a). Overlying this is the first of several fining-upward eruptive cycles. At the base is 50–100 m of densely welded rhyolitic ignimbrite with eutaxitic texture and massive felsic volcanic rocks. This is overlain by up to 150 m of medium- to thin-bedded felsic volcanoclastic tuffs, including cross-bedded siltstones, beds with accretionary lapilli, and volcanoclastic sandstones. At the top of the cycle is a 30–40 m-thick chert (*Awpca*) that is characterized by 5–50 cm-scale layering defined by alternating layers of red hematitic chert and white chert, with less common dark-blue chert layers and intervening volcanoclastic beds.

Next in the sequence, volcanoclastic rocks proximal to the vent show dark-brown weathering and have a matrix of abundant carbonate (ankerite or ferroan dolomite) and a microgranular, granophyric intergrowth of quartz and feldspar (*Awpsc*). Metre-wide dykes of dark-brown carbonate cut the rocks in this area. A section through this part of the sequence (Fig. 14) shows that it is composed of a basal section, 190 m thick, of two fining-upward cycles and up to 1 m of interbedded, carbonate-bearing and carbonate-free volcanoclastic rocks that include a unit of millimetre-bedded porcellanite (Fig. 13b). Thin-section petrography reveals that the amount and grain size of carbonate in the basal section varies in accordance with the size of the clastic grains across bedding. This basal sequence is overlain by another six fining-upward eruptive cycles of compositionally distinctive felsic volcanoclastic rocks. The lowest of these, cycle 4, contains poorly bedded, medium- to coarse-grained, conglomeratic to gritty calc-silicate rocks characterized by subrounded clasts of white, fine-grained felsic volcanic rock and chert up to 10 cm in diameter, in a medium-grained, equigranular matrix of carbonate and granophyric quartz–feldspar. This unit is similar in texture and composition to the vent facies breccia of the Panorama volcano (*Awpx*). Other beds are characterized by elongate, angular fragments of bedded volcanoclastic rocks in a gritty carbonate and quartz–feldspar matrix, and appear to have been affected by in situ brecciation (?syneruptive earthquakes; Fig. 13c). Cycle 5 contains a lower bed that is packed with angular jasper fragments in a dark-brown, carbonate-bearing matrix (Fig. 13d). The top of cycle 6 is capped by a unit of ultra fine grained grey chert (felsic ash) that is cut by black chert veins — the only unit in

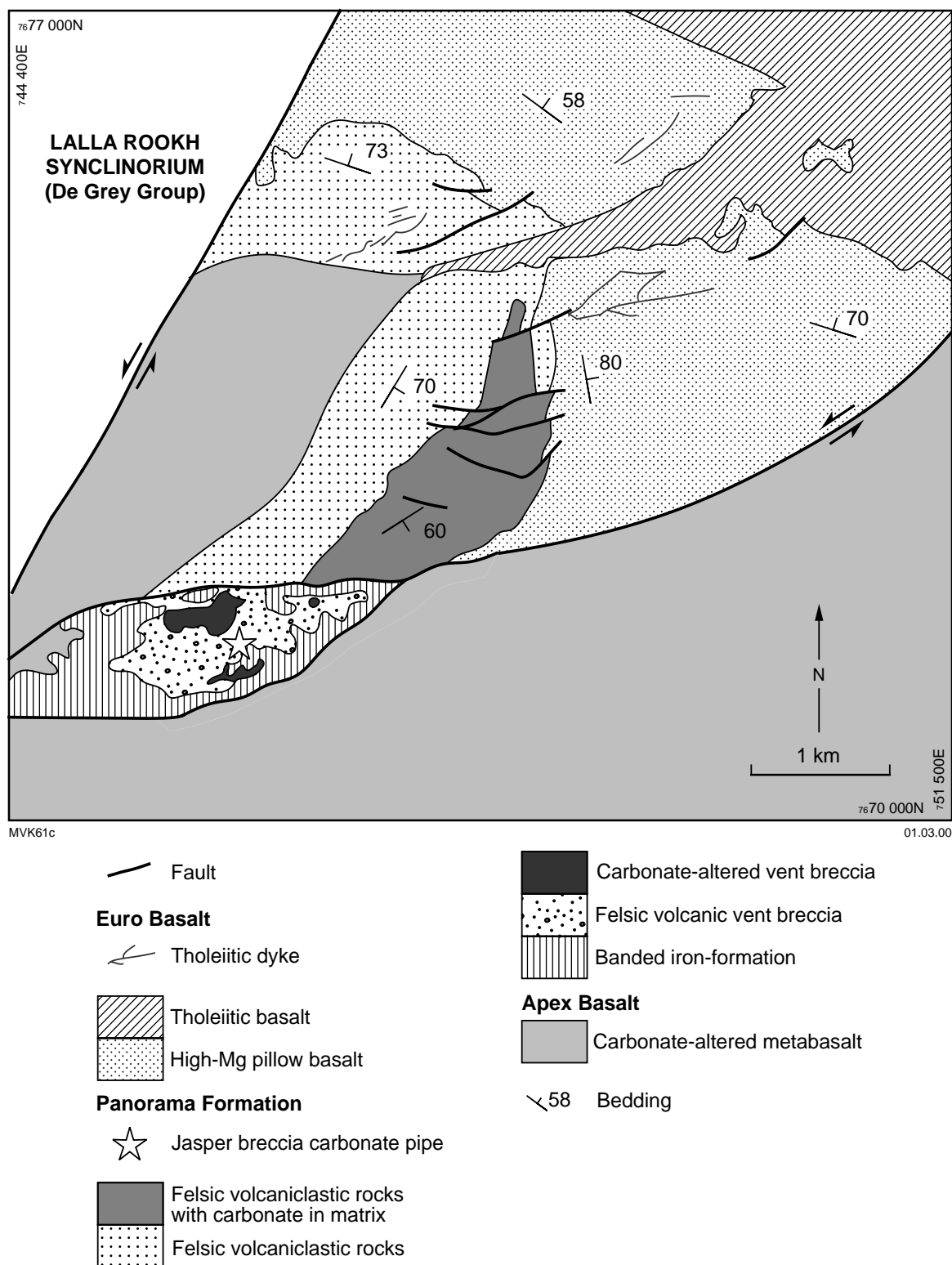
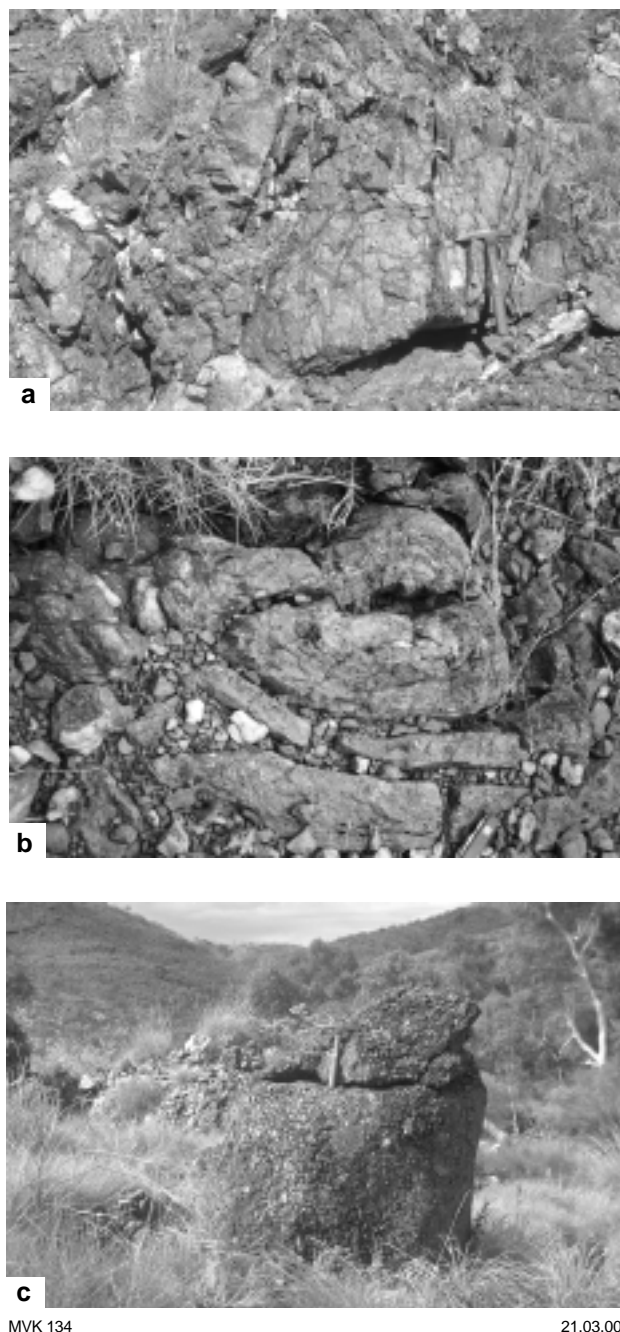


Figure 10. Simplified geological sketch of the Panorama volcano showing the location of the vent facies and its flanking rim of banded iron-formation, as well as the location of the carbonate-bearing breccia pipe shown in Figure 11c. Note the distribution of carbonate-bearing volcaniclastic rocks in the ejecta apron flanking the vent



**Figure 11.** Features of the Panorama volcano vent: a) vent facies breccia of jasper and fine-grained rhyolite fragments in massive, medium-grained, quartz-phyric rhyolite; b) plan view of a fumarole from the core of the vent (width of view 60 cm); c) jasper fragment breccia pipe from the core of the vent (see Fig. 10 for location)

the sequence to contain black chert. Volcaniclastic conglomerates and sandstones at the base of the overlying, carbonate-free cycle 7 contain black chert pebbles, indicating that black chert veining and silicification were at the end of cycle 6 and not some later time. Cycle 8 is the most coarse grained, with a basal debris flow containing packed, angular jasper and less common felsic volcanic fragments up to 15 cm in size in a dark-brown, carbonate-rich matrix. These rocks are crudely bedded on

a 1–4 m scale and display symmetrical, reverse, or normal grading. Cycle 9 is finer grained overall and capped by up to 250 m of millimetre-bedded, carbonate-bearing, tuffaceous siltstones (Fig. 13e). Carbonate-bearing volcaniclastic rocks are also south of the Panorama volcano and in the rocks underlying Panorama Ridge.

In the more distal part of the volcaniclastic ejecta apron along strike to the north and northwest, a similar sequence of rocks lacks carbonate in the matrix and has a white-weathering surface (*AWps*). Identical rocks along strike to the east, across the north end of the North Pole Dome, consist of sandy to pebbly volcaniclastic litharenites and grits, with a grain size up to 8 mm. Bedding varies from crude to well developed with cross-bedding present locally. Angular to subrounded lithic fragments include jasper, white chert, and quartz, which lie within a fine-grained quartz and lithic sand matrix.

Felsic tuff and agglomerate (*AWpt*) outcrop in the northwestern and southeastern parts of the Panorama Belt. In the northwest, there are fine-grained tuffaceous rocks including cross-bedded volcaniclastic sandstones and tuffs, beds of ultra fine grained porcellanite, and centimetre-bedded pumiceous deposits. Shards of clear crystalline quartz and devitrified felsic glass are common in these rocks. On Panorama Ridge, fuchsitic, clotty-textured agglomerate is characterized by rounded clasts of rhyolite, up to 2 cm in diameter, in a rhyolite matrix. These grade upward into fuchsite-altered, fine-grained cherty tuffs. Thin-layered, black and white cherts (*AWpca*) represent silicified, very fine grained, thinly bedded tuffaceous rocks.

The Panorama Formation is interpreted to also outcrop around the northern end of the Shaw Granitoid Complex (see also Di Marco and Lowe, 1989a). As elsewhere, these rocks are composed predominantly of felsic volcaniclastic and epiclastic rocks with subordinate, massive felsic rocks of indeterminate intrusive or extrusive origin. At The Island Hill (AMG 420406) and along strike to the southwest for 10 km, a thick unit ( $\leq 30$  m) of centimetre-to decimetre-layered, hematitic banded iron-formation (*AWpi*) caps the Panorama Formation.

A set of fine-grained, aphanitic to weakly quartz porphyritic dykes (*AWpp*) that cut through the underlying Apex Basalt in the North Pole Dome are interpreted to represent feeder dykes to the Panorama Formation. These represent higher level equivalents of granitic dykes that are proximal to the North Pole Monzogranite (*Agno*).

Cullers et al. (1993) studied the geochemistry of Panorama Formation felsic volcanic flows and volcaniclastic rocks in the Panorama Belt. Volcaniclastic rocks were strongly altered under maximum temperatures of 350°C and affected by intense silicification (83–92 wt% SiO<sub>2</sub>), sericitization, and baritization. The felsic volcanic rocks have highly fractionated rare earth element (REE) trends (high La/Lu<sub>N</sub> ratios), but no negative Eu anomalies. The REE data were interpreted to indicate formation through large-scale melting of eclogite — possibly deeply depressed portions of oldest Warrawoona Group basalts — without subsequent crystallization of plagioclase. This is in contrast to the origin of Duffer Formation lavas,

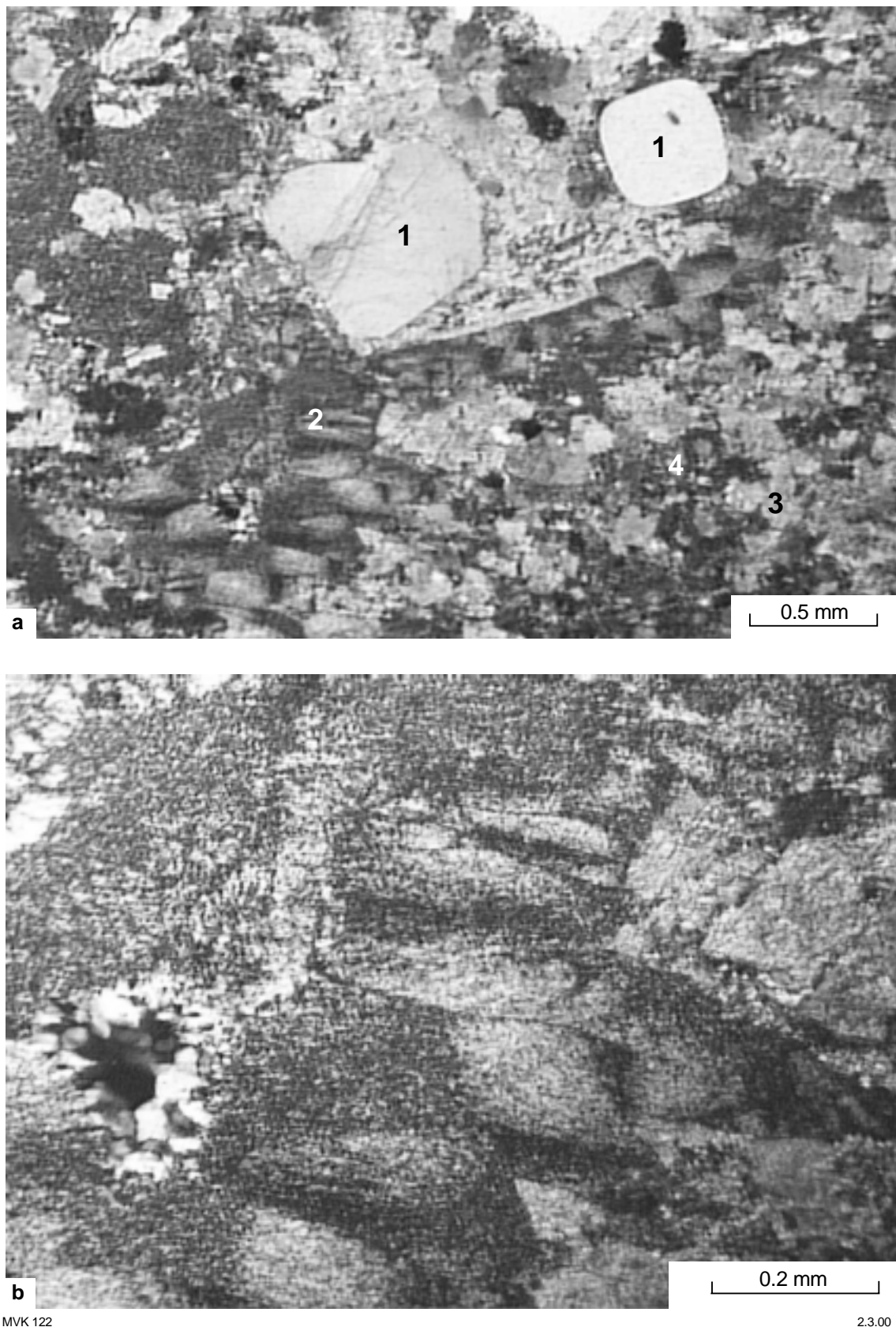
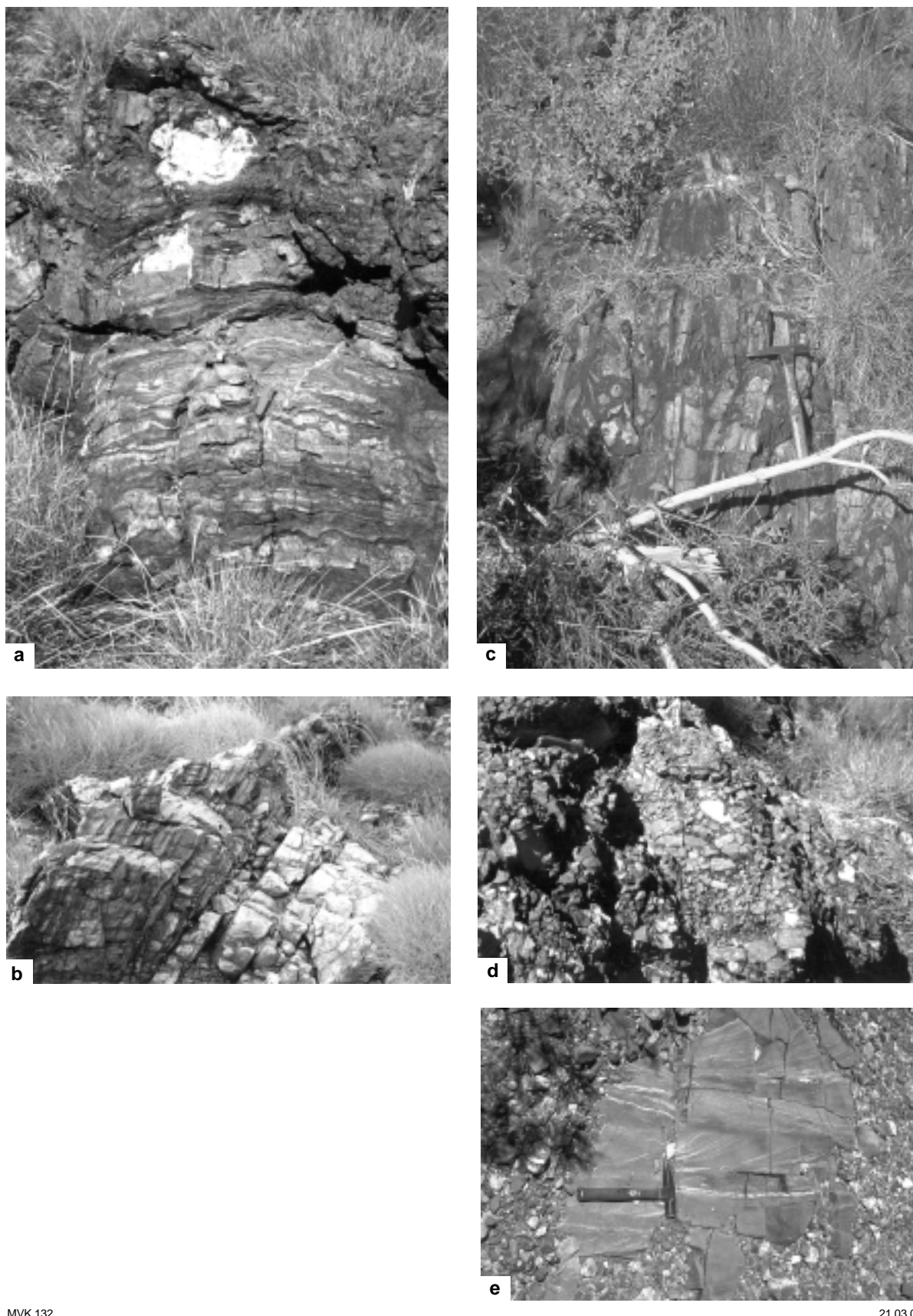


Figure 12. a) Thin section, cross-polarized view of the matrix of the jasper fragment breccia pipe (Fig. 11c). Note the subhedral quartz phenocrysts (1) and the sharp-edged domain of pure quartz (2) in a matrix otherwise containing scattered carbonate rhombs (3) in devitrified rhyolite glass (4); b) detail of the pure quartz domain, showing it is composed of radiating sheaths of fibre bunches nucleated on the prismatic terminations of carbonate rhombs and an inner core of coarser textured, granoblastic quartz. The fibre bunches are characteristic textures of precipitation from boiling fluids. The fact that the quartz fibres nucleate on carbonate rhombs indicates that carbonate was of magmatic origin



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Figure 13. Features of the ejecta apron proximal to the Panorama volcanic vent: a) basal unit of interbedded jasper, carbonate, and felsic volcaniclastic rocks; b) interbedded carbonate-bearing (left) and carbonate-free felsic volcaniclastic rocks; c) angular (?in situ) brecciation of bedding in carbonate-bearing felsic volcanic rocks; d) debris flow breccia of jasper fragments in a quartz-phyric, carbonate-bearing volcaniclastic matrix; e) cross-bedded, carbonate-bearing volcaniclastic siltstones. Penknife in a) and d) is 9 cm long

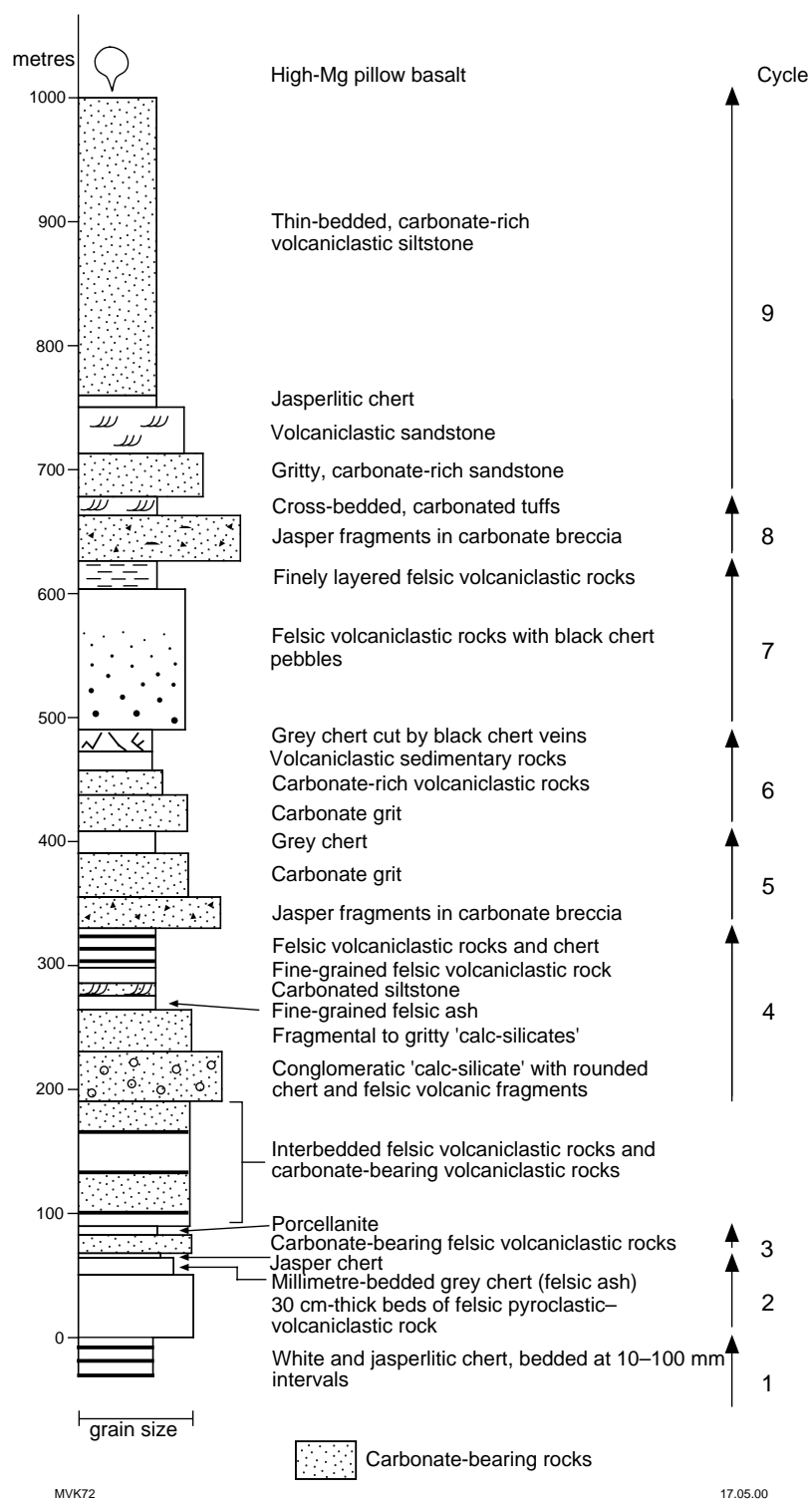


Figure 14. Stratigraphic section through the carbonate-bearing, volcanoclastic ejecta apron of the Panorama Formation, proximal to the vent of the Panorama volcano, showing nine fining-upward eruptive cycles

which may have been derived through fractional crystallization of basalt (Barley et al., 1984).

## North Pole Monzogranite (*Agno*)

The North Pole Monzogranite (*Agno*) is a synvolcanic laccolith of medium- to coarse-grained biotite monzogranite exposed over 6 km<sup>2</sup> in the core of the North Pole Dome. The rock is massive to weakly foliated and homogeneous. Principal minerals are oligoclase (zoned), microcline, quartz, and biotite. Weakly porphyritic texture and miarolitic cavities are locally present. Radiating out from the laccolith are metre-thick dykes of fine- to medium-grained biotite monzogranite, which grade upwards into fine-grained aphanitic dykes feeding the Panorama Formation (*Awpp*). Geochemical analyses of the main body are presented in Hickman (1983). The laccolith has been dated as  $3459 \pm 18$  Ma (Thorpe et al., 1992a), the same age as the Panorama Formation.

## Strelley Pool Chert (*AWs*)

The Strelley Pool Chert (*AWs*) was the name given by Lowe (1983) to a unit of silicified carbonate and siliciclastic rocks in the East Strelley Belt, where it lies disconformably to unconformably on the Coonterunah Group. Results of current mapping have shown that the Strelley Pool Chert also conformably to disconformably overlies the Panorama Formation in the Panorama and North Shaw Belts. Van Kranendonk and Morant (1998) upgraded the Strelley Pool Chert to formation status.

The Strelley Pool Chert is a characteristically white to white and grey layered chert that forms a steeply dipping ridge rising above the surrounding volcanic rocks. Silicification of the Strelley Pool Chert is locally a recent feature, as indicated by the fact that where topographically high ridges of pure chert are transected by a stream, the rocks pass along strike, down topography into pure brown carbonates, which display the identical textural features as their silicified counterparts higher up.

Lowe (1983) described five members in the type area (column 2 of Fig. 15). These are, from base to top: Member 1, a quartzite deposited in a high-energy, shallow-water environment; Members 2 and 3, a regressive succession of subaqueous, wavy-laminated, stromatolite and evaporite chert and intraformational detrital units deposited under intermittent to predominantly exposed conditions; and Members 4 and 5, a progradational sequence of volcanoclastic alluvial fringe. Barnes (1983) found the western part of the Strelley Pool Chert in the East Strelley Belt to contain a similar stratigraphy, except for the inclusion of a felsic volcanoclastic facies at the base (column 1 in Fig. 15). The main stromatolitic horizon in each locality was found to be about 8 m thick. Remapping of this area could not determine if the felsic volcanoclastic facies was part of the chert or belonged to the underlying Panorama Formation, to which it bears textural resemblance. In thin section, the volcanoclastic sandstone contains well-rounded grains of chert, fine-grained felsic volcanic rocks, and monocrystalline and polycrystalline grains of quartz and K-feldspar, which in the latter are

intergrown across long to sutured boundaries. A rind of quartz cement lines embayed, monocrystalline quartz shards. Where there are three grains of touching quartz detritus, they are in contact across 120° triple point boundaries, suggestive of annealing within a hot matrix. The matrix of the rocks is a fine-grained granophyric intergrowth of quartz and feldspar, with a small amount of fine-grained biotite concentrated along grain boundaries. Matrix silica locally displays a fine radiating structure resembling spherulitic texture, and, in places, invades clastic grains along irregular channels and veinlets. Both these textures, and the evidence of annealing of the clastic grains, suggest that the matrix was a hot felsic volcanic ash that has since devitrified.

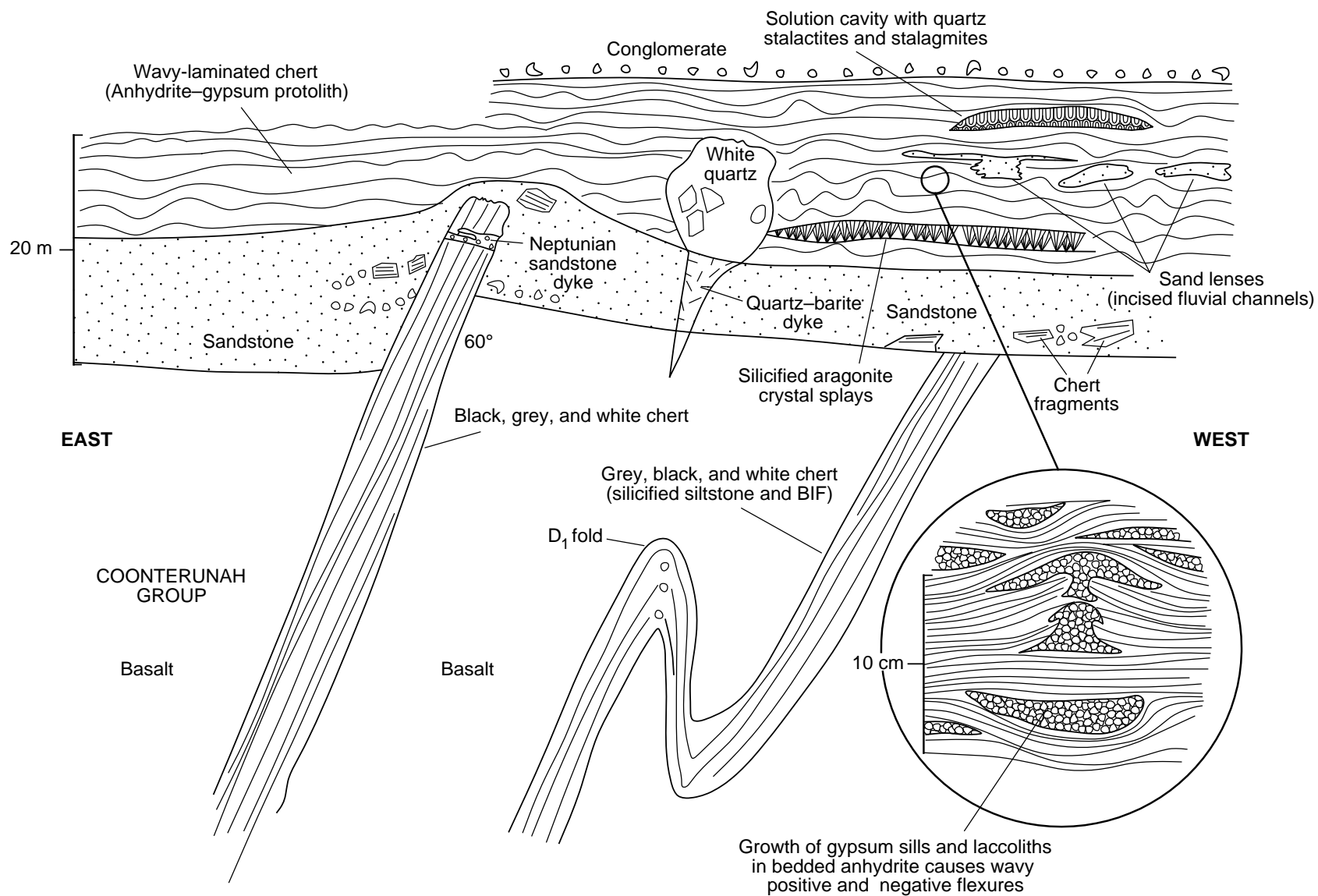
Volcanoclastic sandstones and conglomerates similarly comprise the basal section of the Strelley Pool Chert along the western side of the Panorama Belt (Column 4 in Fig. 15). Here, thick-bedded conglomerates with angular fragments of black chert, felsic volcanic rock, and jasper have a dark-brown matrix of carbonate and an identical appearance to carbonate-bearing felsic volcanoclastic rocks of the Panorama volcano (*Awpsc*).

In the East Strelley Belt, low total strain has resulted in excellent preservation of the unconformity at the base of the Strelley Pool Chert. Aside from the regional pattern outlined by the strike of different aged groups, the unconformity is marked by evidence of relict topography in the basement and erosional scouring. On the western bank of Sulphur Springs Creek (AMG 268652), a 2 m-thick unit of cherty banded iron-formation of the Coonterunah Group (*Aoci*) strikes towards the base of the chert at a 60° angle and extends up into it for up to 18 m (Fig. 16). This palaeohigh ridge is draped by conglomeratic lags and quartz sandstone beds that infill the adjacent small valleys, and the top of the chert bar is overlain by wavy-laminated cherts. A further 100 m west along strike, another cherty banded iron-formation of the Coucal Formation is sharply truncated by the basal sandstone and conglomerate of the Strelley Pool Chert, which contains angular boulders and cobbles of the iron formation 'downstream' to the west. On the eastern bank of Strelley Pool (AMG 223639), basal quartz sandstone of the Strelley Pool Chert fills potholes in silicified pillow basalt of the Coonterunah Group, in which suspended pebbles of the eroded basalt are still visible (Fig. 17). This feature indicates a high-energy, probably coastal marine environment for the basal sandstone of the Strelley Pool Chert.

In the Panorama Belt, excellent exposures of the Strelley Pool Chert are on the banks of the Shaw River. Stratigraphic sections from outcrops flanking both banks of the river show a similar stratigraphy to the Strelley Pool Chert in the East Strelley Belt, including an approximately 8 m-thick section of stromatolitic carbonates (columns 3 and 4 in Fig. 15). A significant difference between the sections is in the greater thickness of radial ?gypsum crystal arrays in the Shaw River localities, indicating possibly more extensive evaporative (exposed) conditions. On the northwestern bank of the Shaw River (column 3 of Fig. 15), a thicker section of well-preserved volcanoclastic rocks is exposed at the top of the formation.



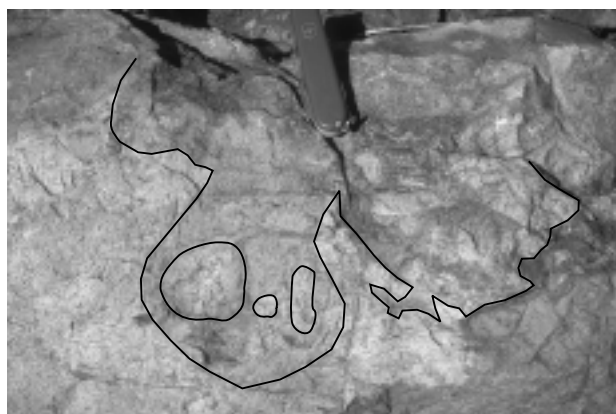
**Figure 15. Stratigraphic sections through the Strelley Pool Chert from the East Strelley Belt in the west (columns 1 and 2) and the Panorama Belt in the east, emphasizing the overall similarities in the position and thickness of the stromatolitic horizon**



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Figure 16. Diagrammatic sketch of the unconformity between the Strelley Pool Chert and Coonterunah Group in the East Strelley Belt



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**Figure 17.** Detail of the unconformity between the Strelley Pool Chert and Coonterunah Group in the East Strelley Belt, at Strelley Pool. This photograph shows a pothole developed in underlying silicified and altered metabasalt of the Coonterunah Group, filled by quartz sandstone of the base of the Strelley Pool Chert. The scouring pebbles of silicified metabasalt, which created the pothole, are seen suspended in the quartz sandstone (outlined)

Wavy-laminated chert is perhaps the most common and diagnostic textural feature of the Strelley Pool Chert, but it is also the feature most difficult to interpret. Lowe (1983) suggested that the chert may represent a bacterial mat community or even a distinct class of stromatolites, but then later recanted this view (Lowe, 1994). Close inspection of the wavy laminations in the chert shows that they are irregular, with small positive and negative flexures that have no consistent relationships to adjacent flexures. The wavy-laminated chert is composed of millimetre-bedded, evenly laminated, grey material that has been disrupted into a wavy pattern by the growth and discontinuous lateral and vertical intrusion of coarser grained, white material (Fig. 16, insert). The wedges of white material grow laterally into, and prise open, the finely bedded, grey chert forming local positive and negative bulges, which form the overall wavy-laminate texture. This is interpreted to result from the diagenetic transformation of plane-bedded anhydrite to amorphous gypsum, a process accompanied by a significant volume gain. The presence of identical wavy-laminate textures in pure carbonates is interpreted to suggest that the carbonate in these rocks may itself be a diagenetic replacement of the anhydrite–gypsum rocks.

Radiating crystal arrays form at scales from centimetre-size up to 20 cm long within crudely defined beds, but grow at right angles to bedding. The crystal splays are now almost entirely replaced by brown ferroan dolomite or ankerite, but may have originally been gypsum, aragonite, or barite (Buick and Dunlop, 1990). These textures again suggest secondary, or even tertiary, diagenetic carbonate replacement.

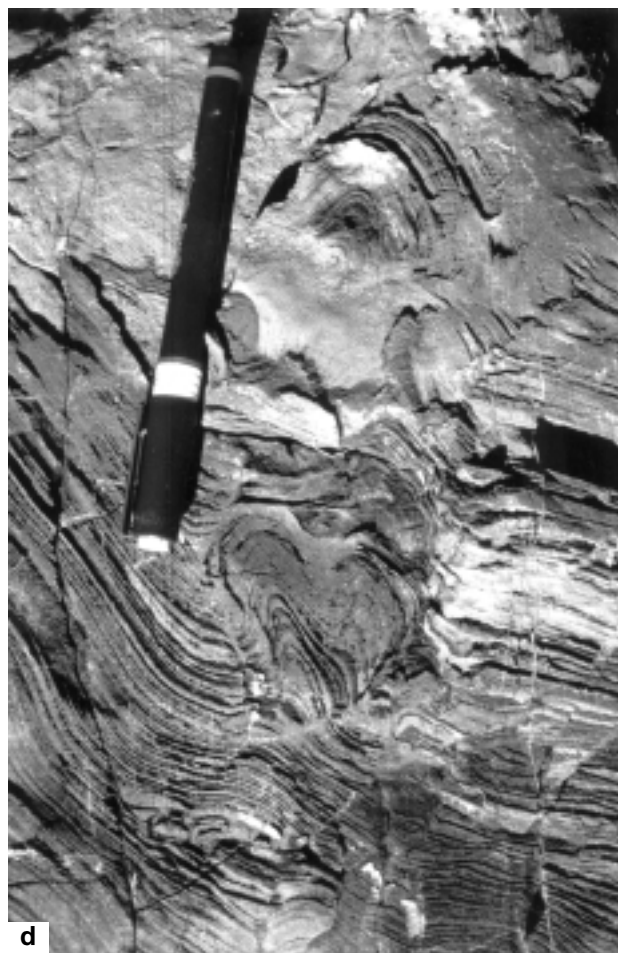
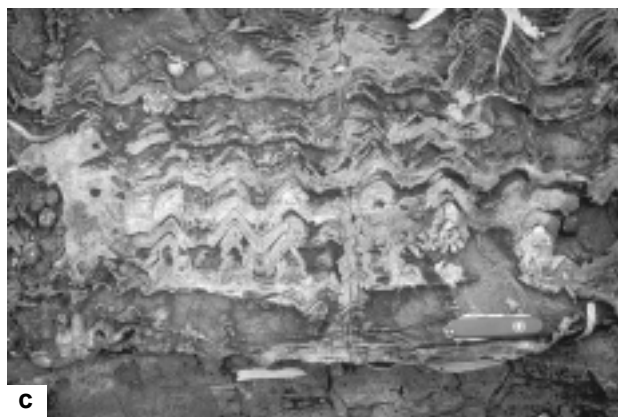
The Strelley Pool Chert contains distinctive ringed conical to domal structures that have generated much



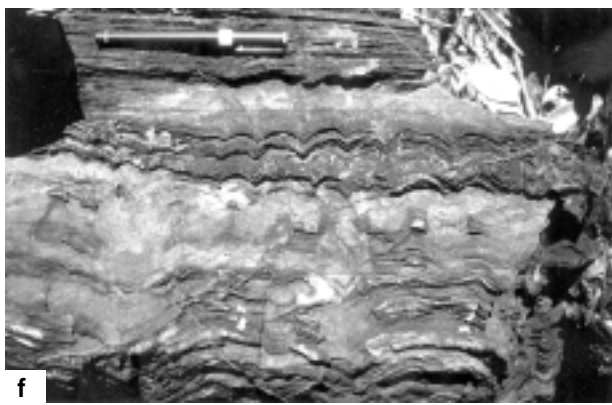
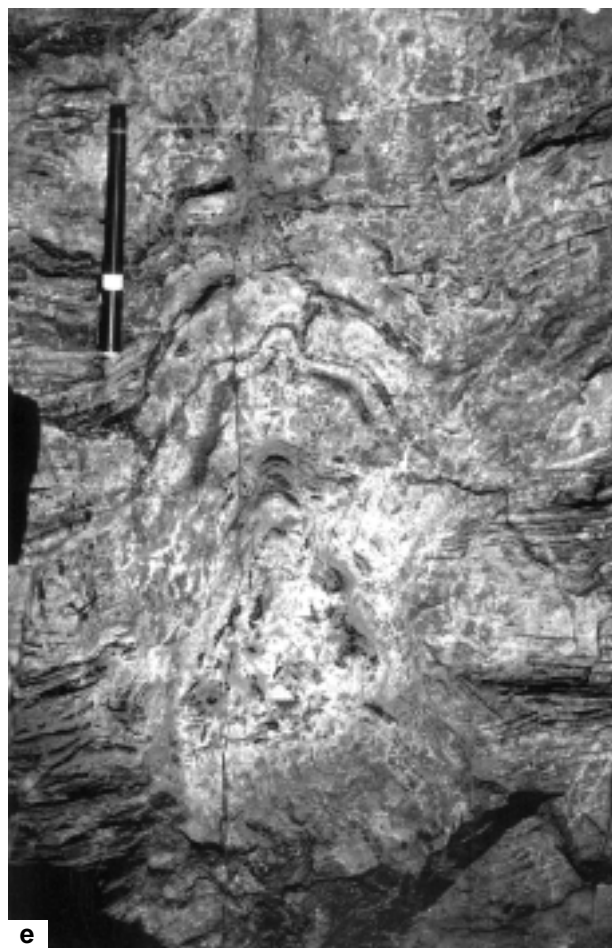
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**Figure 18.** Fossil stromatolites of the Strelley Pool Chert: a) plan view, on a bedding plane, of a section through conical stromatolites of various sizes in silicified carbonate. Note the random size and distribution of the forms, indicating that they are not interference folds; b) oblique view of conical stromatolites on a bedding surface; c) cross section through conical stromatolites in carbonate; d) a branching, bulbous stromatolite with smooth, rounded tops; e) cross sectional view of flat-crested, cusped stromatolites in carbonate; f) large stromatolite in bedded, silicified carbonate



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scientific debate (Lowe, 1980, 1994), with opinions varying between a biogenic origin as stromatolites, to forms developed during evaporate precipitation, to folds. However, well-preserved plan and cross sectional pavements in undeformed rocks display unequivocal features of a diverse biological origin (Fig. 18; Hofmann et al., 1999). In plan view, the ringed structures are circular to elliptical and may be of random size on a single bedding plane (Fig. 18a). In cross section, they are seen to be commonly conical structures (Figs 18b,c) that may branch up-section, as do modern stromatolites. Detailed inspection has uncovered a diverse assemblage of forms in the Strelley Pool Chert, including a branching, bulbous

form (Fig. 18d), others with broad, flat crests (Fig. 18e), and a larger, higher amplitude form (Fig. 18f), all of which suggest the presence of a diverse biological community of micro-organisms in these ancient rocks. A biological origin for these structures was further supported by the discovery of microfossils in the Strelley Pool Chert (Schopf and Packer, 1987).

The unit of chert that caps the Panorama Formation in the North Shaw Belt lacks the distinctive wavy-laminated and stromatolitic carbonates, and is at a higher metamorphic grade than correlatives in other belts. Instead, the Strelley Pool Chert in this area comprises thin outcrops

of quartz sandstone, grey chert, millimetre-layered chert-carbonate, and a local, thin capping unit of pink and white layered marble (AMG 353306). The sandstones are interpreted to represent Member 1 of Lowe (1983), whereas the fine-grained cherts may represent Member 5. At the northern apex of Shaw Granitoid Complex, the Strelley Pool Chert is layered on a millimetre scale and composed of alternations between recrystallized white quartz and carbonate, possibly representing a sinter-type, exhalative deposit.

### **Euro Basalt (*Aweb*, *Awebc*, *Awebm*, *Awebt*, *Awebk*, *Aweft*, *Awesh*, *Awec*, *Awed*)**

The Euro Basalt lies conformably above the Strelley Pool Chert in the East Strelley, Panorama, and North Shaw Belts. The unit is principally composed of pillowed basalt of interbedded tholeiitic (*Awebt*) and high-Mg (*Awebm*) composition, but contains a discontinuous basal unit of basaltic komatiite, and higher up contains thin beds of chert (*Awec*), fine-grained clastic sediment (*Awesh*), and felsic volcanoclastic rocks (*Aweft*). The maximum thickness of this formation is 9.4 km in the Panorama Belt. Except for the stratigraphically highest rocks of the Potkoorok Member, volcanic rocks of the Euro Basalt are remarkably fresh compared with the Apex Basalt underlying the Panorama Formation.

Two compositionally distinctive units within the Euro Basalt are designated member status. Discontinuous at the base of the formation is the Cloisters Member, a pyroxene spinifex-textured, massive and pillowed unit of high-Mg to komatiitic basaltic lavas, and up to 2250 m of overlying pillowed and locally ocellar and fragmented (pillow breccia agglomerate) high-Mg basalt. The lower basaltic komatiite (*Awebk*) is present in the southwestern part of the Panorama Belt and northern part of the North Shaw Belt, where it displays a spinifex texture with blades up to 4 cm long. In thin section, these rocks consist of scattered augite phenocrysts up to 1 mm long in a fine spinifex-textured matrix of curved, branching, feathery pyroxene. Augite phenocrysts have a pseudohexagonal cross section made up of long  $100^\circ$  cleavage faces, and meeting pairs of long and short  $010^\circ$  cleavage faces such that  $010^\circ$  faces are not developed. These phenocrysts are characteristically hollow with a core of some fine alteration mineral (?serpentine), suggesting the growth of augite phenocrysts on ?olivine or pigeonite nuclei. The rocks are characteristically altered to a fine-grained intergrowth of chlorite, serpentine, tremolite-actinolite, and carbonate. In the North Shaw Belt, good exposures of this unit (*Awebk*) outcrop 800 m west of Shaw Gorge (AMG 400400), and 6.3 km northeast of North Shaw Well (AMG 530429; *Awebm*), although the latter rocks are heavily carbonate altered despite preserving good pillow and spinifex textures. The thickest section of pillowed, high-Mg basalt is in the northern part of the Panorama Belt, stratigraphically above the Panorama volcano. In this area, the lowest rocks display spinifex texture of branching, curved augite crystals up to 1 cm long, whereas the bulk of the volcanic rocks higher up are pillowed high-Mg basalts.

In the East Strelley Belt, the Cloisters Member is up to 2.7 km thick and consists of pillowed, ocellar, and locally vesicular high-Mg basalts and related subvolcanic dolerite intrusions. The member overlies a basal section of interbedded volcanic rocks ( $\leq 1$  km thick) and up to five cherts at the far western part of the belt. Here, the stratigraphically lowest chert is a unit of silicified pale-grey to green tuffs. Above highly altered mafic rocks, the next chert is massive and pale grey, interbedded with pale-brown, centimetre-bedded tuffs and up to 2 m of pale-yellow weathering, cross-bedded volcanoclastic sandstone. In thin section, the cross-bedded sandstone is observed to be a matrix-supported volcanoclastic rock with 50% angular to ragged-edged, dagger-shaped clasts of quartz, chert, and dacite, and rounded clasts of sericite-altered feldspar. The matrix is predominantly microcrystalline quartz. Larger matrix domains display radial spherulite texture indicating that the rock matrix was a volcanic glass or very fine hot ash. The upper three cherts are pale grey and massive or faintly ?bedded at a centimetre scale, with thin blue-black chert sills that are interspersed with massive high-Mg doleritic basalts or dolerites. Further east, the basal part of the formation contains only two cherts. These are pale-grey-green, silicified, fine-grained clastic rocks, with the basal chert locally containing a lower breccia unit interpreted to be a silicified flow-top breccia. Overlying the cherts are up to 2 km of high-Mg pillow basalt and related dolerites. The volcanic rocks are well preserved and have microspinifex texture identical to their eastern counterparts, but are relatively fresh with abundant, unaltered augite. Local alteration includes carbonate-epidote-titanite, whereas chlorite forms in fractures.

In the Panorama and East Strelley Belts, rocks of the Cloisters Member are conformably to unconformably overlain by dark-brown weathering, pillowed to massive tholeiitic basalts (*Awebt*) and related synvolcanic dolerite intrusions (*Aweb*) of the Miralga Creek Member, whose type section is 8.4 km east-southeast of Lalla Rookh Well (AMG 500750). An eruptive vent of this member is preserved cutting through the core of the folded felsic volcanoclastic apron flanking the Panorama volcano (AMG 490750; Fig. 10). Here, a 600 m-wide, north-easterly striking dyke of schistose metadolerite (*Awed*) cuts through felsic volcanoclastic rocks of the Panorama Formation, and passes gradationally up-section into pillowed tholeiitic basalts (*Awebt*) that lie subconformably on underlying high-Mg basalts of the Cloisters Member. The transition zone between massive and pillowed varieties is within an asymmetrical, triangular-shaped area that has a subvertical western wall and a more shallowly inclined eastern slope. This is interpreted to be an asymmetrical eruptive vent, because as one heads east from the transition area, the size of pillows in the tholeiitic volcanic rocks rapidly decreases. Rafts of both a distinctive chert unit in the underlying Cloisters Member and of high-Mg basalt derived from the underlying unit are found within the asymmetrical transition zone, supporting interpretation as a vent. Equigranular, undeformed dolerite (*Awed*) forms dykes (?ring dykes) parallel to either side of the vent, and sills near the base of the formation (e.g. AMG 590741). These rocks are dark brown in colour, with a recrystallized doleritic texture

comprising plagioclase (30–40%), titanite (50–60%), quartz (<5%), and magnetite (1–2%) with extensive chlorite–sericite alteration.

In the North Shaw Belt where the Cloisters Member is absent, the Miralga Creek Member rests directly on the Strelley Pool Chert. Here, it comprises tectonically flattened, lower amphibolite to greenschist facies tholeiitic pillow basalts, chert (*AWec*), and a unit of silicified lithic sandstone, siltstone, and shale (*AWesh*). The sedimentary unit is up to 200 m thick and consists of lithic sandstone, variegated siltstones, quartz–muscovite schist, amphibolite, and 1–2 m of grey-black, irregularly laminated and brecciated chert (after shale).

Overlying the Miralga Creek Member is a thick section of tholeiitic to high-Mg pillow basalts. These lavas are interbedded on a kilometre scale, locally carbonate altered (*AWebc*), and interbedded with a few chert units and a unit of felsic volcaniclastic rock (*AWeft*). The felsic volcaniclastic unit (AMG 351439) is composed of feldspathic sandstone with cross-bedding, shale, finely rippled ash beds, felsic porcellanite, volcaniclastic sandstone and agglomerate, pebble conglomerate, and chert (silicified felsic ash?). Near the top of the unit are beds up to 1 m thick of spherulitic felsic volcanic rock and pumice. One of these contains 30 cm-long, radiating crystal splays perpendicular to bedding, similar to those in the Strelley Pool Chert and of probable evaporitic origin (i.e. gypsum, aragonite). South of the Shaw River (AMG 394422), the unit consists of thinly bedded, grey, tuffaceous chert (silicified felsic ash) with beds of

accretionary lapilli, volcaniclastic sandstone, and pebble conglomerate.

At the top of the Euro Basalt, in the southwestern extension of the Panorama Belt west of the Shaw River (AMG 285380), is an approximate 2800 m-thick sequence of interlayered pillow basalts ( $\text{MgO} < 5\%$ ) and cherts of varied character. This sequence, which unconformably underlies the Sulphur Springs Group, represents the uppermost part of the Warrawoona Group known in the east Pilbara Craton. Basalts are commonly well pillowed with weakly vesicular rinds, intrapillow hyaloclastite, and pillow shelves being quite common. A distinctive unit of chert (AMG 280365), up to 40 m thick, is layered on a 1–5 cm scale between alternating white (chert) and black (hematite) layers. The rock is recrystallized and has a saccharoidal texture. Other cherts in this part of the formation include massive blue-black chert, pale-brown lithic siltstones, and white and grey layered chert.

Glikson et al. (1986; samples 69946–69964) analysed a section through the Euro Basalt in the East Strelley Belt. Major and trace element data indicate vertical changes in magma composition within three units separated across two distinct boundaries (Fig. 19). The most striking of these changes is in the lower unit, in which  $\text{MgO}$ ,  $\text{Al}_2\text{O}_3$ , and  $\text{Ni}$  all decrease upwards and then become replenished again almost to initial values. All three zones show a saddle-shaped pattern of initial  $\text{MgO}$  depletion and then enrichment, and this is accompanied by changes in  $\text{FeO}_{\text{TOT}}$  (decreasing towards the top in the upper two cycles) and  $\text{Al}_2\text{O}_3$ . Such changes may reflect successive pulses of

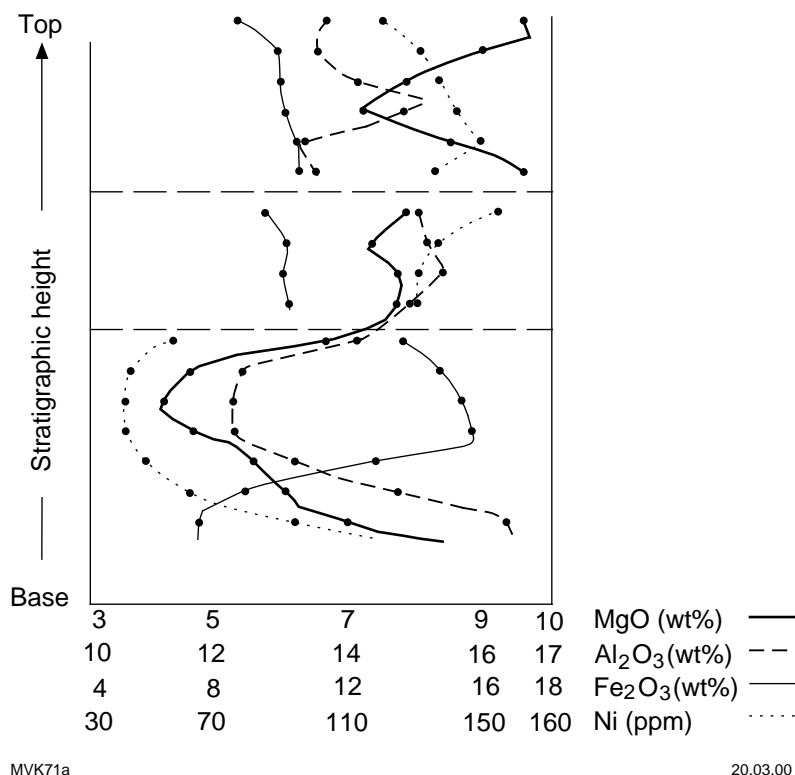


Figure 19. Geochemical transect through the Euro Basalt in the East Strelley Belt showing changes with stratigraphic height. Data from Glikson et al. (1986)

magma replenishment during eruption, and intervening fractional crystallization.

Pillowed volcanic rocks higher up in the Euro Basalt (*AWeb*), where it unconformably underlies wackes and felsic volcanic rocks of the Sulphur Springs Group in the Jamesons area (AMG 278382), have low MgO (1.88 – 4.95%) and flat to slightly light rare earth element (LREE)-enriched normalized REE profiles, with small negative Eu anomalies (M. J. Van Kranendonk, unpublished geochemical data). Trace element data show a similarity with mid-oceanic-ridge basalt (MORB) between Nb–Lu, but then a gradual rise in the high field strength elements (HFSE), punctuated by distinct K and Ba spikes.

## Golden Cockatoo Formation (*Aji, Ajp, Ajq, Ajx*)

The Golden Cockatoo Formation outcrops in the Abydos Belt and in a highly tectonized sliver along the margin of the Yule Granitoid Complex to the south. In the Abydos Belt, the lower parts of the formation are in intrusive or sheared contact with granitoid rocks. The upper parts of the formation are deformed within the Numerous Scrapes Deformation Zone. The formation, which is metamorphosed to lower amphibolite facies, consists of interbedded muscovite metapelite (*Ajp*), white- to beige-weathering metaquartzite (*Ajq*), brown-weathering, centimetre-bedded banded iron-formation (*Aji*), and interlayered mafic schist of indeterminate intrusive or extrusive origin. Tectonically interleaved rocks to the south include metapelite, metaquartzite, and mafic schists (*Ajx*).

Poorly defined bedding at a 10 cm scale is widely preserved in quartzites, subparallel to a penetrative metamorphic foliation defined by micas. Well-preserved cross-bedding is locally developed (e.g. AMG 083390) and indicates right way-up. Metapelites are strongly foliated to schistose and commonly contain 10 cm- to metre-thick interbeds of metaquartzite or metaquartz arenite. A 10 cm-wide layer on the western edge of NORTH SHAW contains 0.5 cm diameter garnets together with biotite and muscovite.

The banded iron-formation is distinct from those of other formations in having a regular, 1–5 cm-scale layering defined by alternating layers of blue-grey and white cherty rock where fresh and dark, rusty red-brown and grey cherty material where weathered. The formation is not commonly foliated, but is commonly tightly folded (see **Structure**).

## Sulphur Springs Group

The rock units of the Sulphur Springs Group are described below. Detailed descriptions of rocks of the Kangaroo Caves Formation and Strelley Granite may be found in Brauhart (1999), who also presented extensive, high-quality geochemical analyses of these rocks. Van Kranendonk (in prep.b) presented additional geochemical data on the lower formations of the group and a detailed discussion on the implications the data have on tectonic models of evolution of this group.

## Six Mile Creek Formation (*ASmb, ASmbm, ASmf, ASmc*)

The Six Mile Creek Formation includes basalts, cherts, and a unit of dacite that underlie wackes of the Leilira Formation in the Soanesville Belt west (AMG 185525) and north (AMG 242630) of the Strelley Granite, in the Soanesville Belt along the southern margin of NORTH SHAW (AMG 255221), and in the Pincunah Belt adjacent to the Yule Granitoid Complex (AMG 093428).

The main outcrop of the Six Mile Creek Formation west and north of the Strelley Granite consists of massive high-Mg basalt (*ASmbm*), pillowed basalt (*ASmb*), and chert (*ASmc*). Bright, light-green chert forms in the northwest, whereas cherts west of the Strelley Granite represent silicified, fine-grained siliciclastic rocks. Along the southern margin of NORTH SHAW, pillow basalts (*ASmb*) are interbedded with both grey, aphanitic cherts that represent silicified flow tops, and with less common, silicified, grey-green siltstones (*ASmc*). Volcanic rocks along the southern margin of the Pincunah Belt contain local ocelli.

Small, locally dismembered units of spherulitic to massive dacite (*ASmf*) outcrop within the southern margin of the Pincunah Belt and in the Numerous Scrapes Deformation Zone. These rocks are locally intrusive into ocellar basalt host-rocks, and are cut by chert dykes. However, the dacites are also locally subconcordant to layered, silicified sedimentary cherts and thus may be, in part, of extrusive origin as well. Dacites of the Six Mile Creek Formation are geochemically distinct from those higher up in the Sulphur Springs Group (Kangaroo Caves Formation), being less strongly calc-alkaline. However, like their stratigraphically higher equivalents, Six Mile Creek dacites contain local massive sulfide mineralization (Mad Hatters Zn, Pb, Brt prospect; AMG 103424). Sipa Resources Ltd of Perth conducted 8–10 Pb-isotopic analyses of galena from the Mad Hatters gossan, and the results indicate an age of c. 3.28 Ga, consistently older than that for gossans in the Kangaroo Caves Formation, which are c. 3.26 Ga (Morant, P., 1998, pers. comm.).

## Leilira Formation (*ASls, ASlf, ASlc*)

The Leilira Formation is predominantly composed of interbedded lithic arenite, wacke, and quartz sandstone (*ASls*). The rocks are typically sandstones, but locally vary from coarse to pebbly sand and local siltstone. Bedding is commonly not well developed, being restricted to crude partings at a metre scale and identified principally by trends on airphotos; no grading or cross-stratification was observed. Only near the top of the formation, north of the Strelley Granite (AMG 246605), do the rocks become well bedded and finer grained. These siliciclastic rocks contain abundant clasts of felsic volcanic material, mafic volcanic detritus, and milky white to clear quartz.

About 1.5 km east of the Jamesons prospect (AMG 292390), Leilira Formation wackes lie unconformably on pillow basalts of the Warrawoona Group (Fig. 20; Van Kranendonk, 1997). The base of the

formation in this area is composed of up to 1 m of thinly bedded, blue-grey felsic ash and discontinuous lenses of white quartzite. These rocks are overlain by 60 cm of felsic to intermediate lava with basal eutaxitic texture. Overlying this is another thin, felsic ash bed, and then about 40 m of massive, non-bedded wacke or welded volcanoclastic sandstone, and up to 1 km of massive felsic volcanic rock (*Aslf*). A further 200 m east, where the formation unconformably overlies cherts at a high angle, basal sedimentary rocks are only 1 m thick and composed of, from base to top, millimetre-bedded volcanoclastic sandstone with local fumaroles, 1 cm of ferruginous breccia, a eutaxitic-textured felsic volcanic rock, blood-red shale, and white quartzite (inset of Fig. 20b).

Following the basal unconformity of the Sulphur Springs Group along strike to the west, basal Leilira Formation sedimentary rocks are up to 2 m thick and composed of a distinctive light-brown unit of silicified, thin-bedded sandstone. This basal marker unit is locally intruded and disrupted by overlying felsic volcanic rocks (*Aslf*). In the southwest of this area (AMG 265348), felsic volcanic rocks are overlain by up to 1 km of bedded wacke and a thick blue-black and white layered chert, which represents silicified felsic volcanic ash and volcanoclastic siltstone (*Aslc*).

Along the southern margin of the Pincunah Belt, the Leilira Formation is represented by the same triad of massive felsic volcanic rocks (*Aslf*) interbedded with wacke and volcanoclastic sandstone (*Asls*), and capped by silicified, centimetre-bedded felsic ash and siltstone (*Aslc*).

### Kunagunarrina Formation (*Asks*, *Askb*, *Askuk*, *Askbm*, *Askbc*, *Askcw*, *Askcg*)

The Kunagunarrina Formation is predominantly composed of high-Mg to komatiitic basalt. The type section of this unit is in the well-exposed Pincunah Belt, where it is up to 2400 m thick. The lower basaltic unit (*Askbc*, *Askbm*) is about 500 m thick and composed of massive to locally schistose rocks. The lowermost part is heavily carbonate altered and featureless (*Askbc*), capped by a massive, grey to grey and black chert (*Askcw*). Overlying this are rocks that have a spinifex texture of fine, radiating crystal arrays visible in some outcrops, and a fine- to medium-grained, interlocking crystal texture elsewhere. In thin section, these rocks commonly have euhedral, but completely altered augite phenocrysts set in a matrix of microspinifex-textured augite, indicating that they are high-Mg basalts (7–16% MgO). Medium-grained doleritic basalts near the top of this unit (7–8% MgO) contain plagioclase laths and clinopyroxene phenocrysts in altered matrix glass, which is now composed of chlorite and small, euhedral crystals of epidote. This unit is capped by a 10 m-thick unit of chert that varies from green to grey massive chert and bright-green chert at the base, to laminated, black and bright-green chert at the top, the latter representing a silicified flow top (*Askcg*).

Immediately above the green chert is a distinctive unit (≤850 m thick) of dark-brown weathered basaltic komatiite with MgO between 8 and 19% (*Askuk*). The

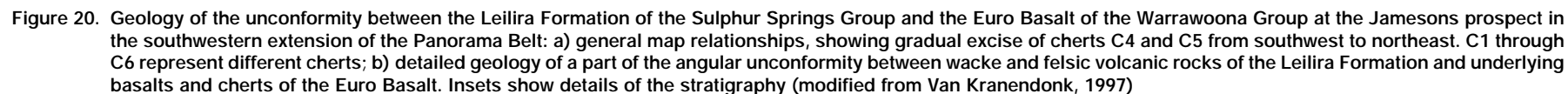
following textural variations occur from base to top: a basal, olivine cumulate layer; ocellar- and pillow-textured volcanic rocks; a unit of coarse fragmental tuff; pillowed and ocellar volcanic rocks; coarse spinifex-textured rocks; fragmented pillows; another layer of coarse spinifex-textured rocks; and massive rocks. In some areas, spinifex-textured augite crystals are up to 6–10 cm long, although no classical examples of sheath-bladed olivine were seen. The most common texture is that of skeletal, euhedral augite phenocrysts in a microspinifex-textured matrix (Fig. 21a), and forms in rocks with between 11 and 19% MgO. In coarse-textured spinifex rocks (12–16% MgO), large augite phenocrysts are typically composed of a thick mantle surrounding a core of altered material, possibly representing an altered nucleus of ?olivine or ?pigeonite (Fig. 21b).

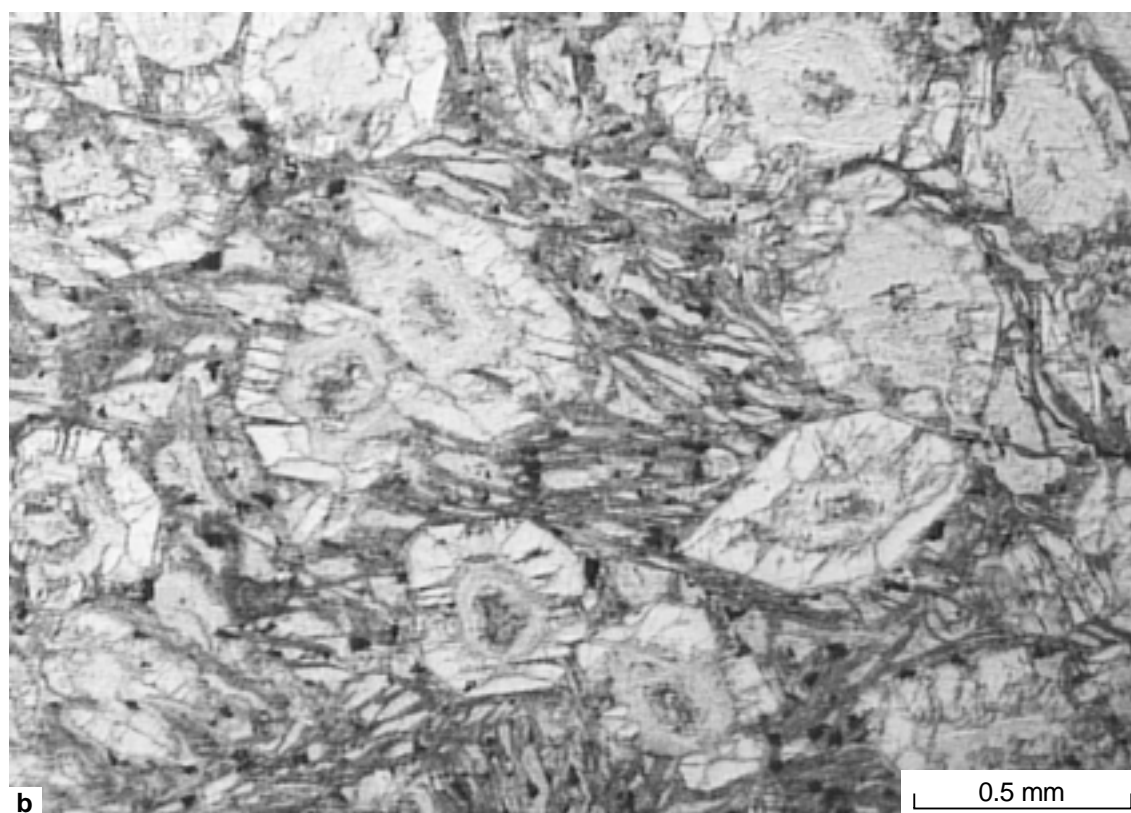
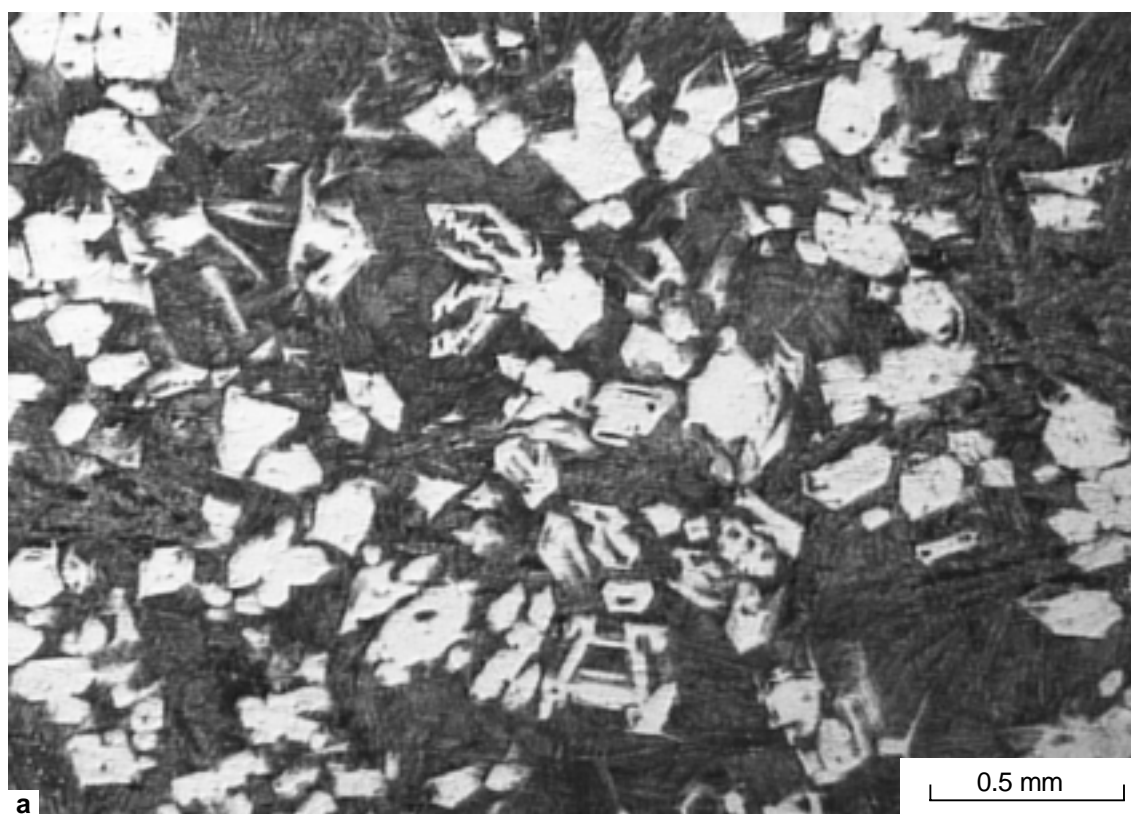
The Kunagunarrina Formation in the East Strelley Belt is composed of a basal unit of spinifex-textured basaltic komatiite (*Askuk*) and then up to 2 km of pillowed and massive basalt and derived mafic schist (*Askb*) with thin interbeds of chert. Just north of the Strelley Granite in the Soanesville Belt, immediately south of the Roadmaster prospect (AMG 235577), creamy-coloured, thin-bedded shales and siltstones (*Asks*) are near the top of the formation. The top of this unit is intruded by centimetre-thick sills of black and grey chert, and these increase upwards until the rock is a well-layered black and white chert (*Askcw*).

### Kangaroo Caves Formation (*Ascbi*, *Ascbt*, *Ascsh*, *Ascf*, *Ascfa*, *Ascfd*, *Ascfr*, *Ascc*, *Asci*, *Ascsx*)

The Kangaroo Caves Formation is up to 1.5 km thick and lies conformably on rocks of the Kunagunarrina Formation. From approximately bottom to top, the formation is composed of massive, pillowed, pillow-brecciated, and hyaloclastic andesitic volcanic rocks (andesite–basalt and andesite), and sills and lavas of dacite, rhyodacite, and rhyolite. These rocks are best developed in the Soanesville Belt, east of the Strelley Granite, where they are well exposed, effectively undeformed, and weakly metamorphosed at lower greenschist facies. Andesitic volcanic rocks are much more abundant in the northern part of this area, whereas felsic compositions are more common in the south. West of the Sulphur Springs prospect, spherulitic rhyodacite sills are common, whereas between Sulphur Springs and Kangaroo Caves, dacite sills (*Ascfd*) that are variably amygdaloidal, typically aphyric, and commonly perlitic, have been emplaced near the top of the formation. Individual sills reach a maximum of 250 m in thickness, and may be continuous for several kilometres along strike. South of Kangaroo Caves, rhyolite domes and pumiceous volcanoclastic rocks (*Ascfr*) increase in abundance (Morant, 1995, 1998).

The bulk of the volcanic succession around the Strelley Granite is underlain by andesite–basalt (*Ascbi*), a term used because detailed trace element geochemistry of these rocks indicates that they range across andesite and basalt (Brauhart, 1999). This unit varies from pillowed to massive and is variably amygdaloidal. Hyaloclastite is





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Figure 21. Plane-polarized thin section views of high-Mg basalt of the Kunagunarrina Formation: a) skeletal augite crystals in a microspinifex matrix; b) cross section through coarse spinifex-textured pyroxene needles, showing pseudo-hexagonal augite rims growing around altered nuclei in a matrix of chlorite-epidote

Figure 22. Outcrop features of the Kangaroo Caves Formation: a) pillowed basalt-andesite with flow banding and

commonly developed between pillows, at flow tops, and in some thicker sequences (Fig. 22a). The primary mineralogy of this unit is not preserved, but thin sections show primary volcanic textures of fine, interlocking plagioclase laths in a fine-grained, recrystallized groundmass of metamorphic chlorite and minor epidote, carbonate, and opaque minerals. Stubby, rectangular ferromagnesian phenocrysts are represented by rarely preserved shapes, but have been completely altered to chlorite.

In the southwestern extension of the Panorama Belt, the Kangaroo Caves Formation includes a lower, discontinuous unit of pillowed and tuffaceous andesite (*AScbi*, *AScbt*). At the base of the formation (AMG 258342), a 100 m-thick unit of coarse, andesitic lapilli tuff (*AScbt*) contains large (1–50 cm), angular to subrounded inclusions of felsic volcanic rocks (including flow-banded rhyolite) and basement country rocks enclosed by rinds of accreted volcanic glass (Fig. 22b). In thin section, the matrix consists of millimetre–centimetre size fragments of altered volcanic glass, quartz and feldspar crystals, chert, and other rock types in an ultra fine grained matrix. The volcanic glass and crystal fragments have curved, embayed margins. The volcanic glass fragments and matrix are altered to chlorite–carbonate. In this belt, the andesite–basalt unit is overlain by massive, columnar-jointed, and weakly bedded rhyolite (*AScfr*), which is quartz-phyric and displays local spherulitic textures.

In the Pincunah Belt, the Kangaroo Caves Formation comprises centimetre-bedded, white and blue-grey to black chert (*AScc*), thinly bedded shale (*AScsh*), well-bedded, pyritic andesite tuff (*AScbt*), and millimetre–centimetre bedded, fine-grained felsic volcanoclastic rocks (*AScf*). The chert comprises millimetre–centimetre bedded volcanoclastic rocks of felsic to intermediate composition in which some centimetre-thick beds contain 50–60% angular fragments of monocrystalline volcanic quartz. The shale varies in colour from grey to red, and locally contains sandy interbeds. The andesitic tuff is a dark-yellow weathering, pale-green, fine-grained rock, well indurated, and with bedding displayed at a 0.5–40 cm scale. Thicker beds are slightly coarser grained with distinct, clear crystal fragments of quartz. Pyrite grains are ubiquitous and form up to 2% of the rock. In thin section, the rock has a blotchy appearance defined by irregular-shaped, brownish, devitrified volcanic glass shards. The fine shards are overgrown by a reticulated set of fine, needle-shaped prisms of a colourless, birefringent mineral (possibly clinopyroxene or an amphibole). Larger shards are composed of chlorite. Sprinkled throughout the rock are angular to subangular fragments of monocrystalline volcanic quartz and cryptocrystalline chert. Some beds

intra-pillow hyaloclastite; b) armoured lapilli in andesitic tuff breccia; c) lobate sills of rhyodacite from the Kangaroo Caves dome; d) synsedimentary slump folds in banded iron-formation from the olistostrome breccia at the top of the formation, Sulphur Springs

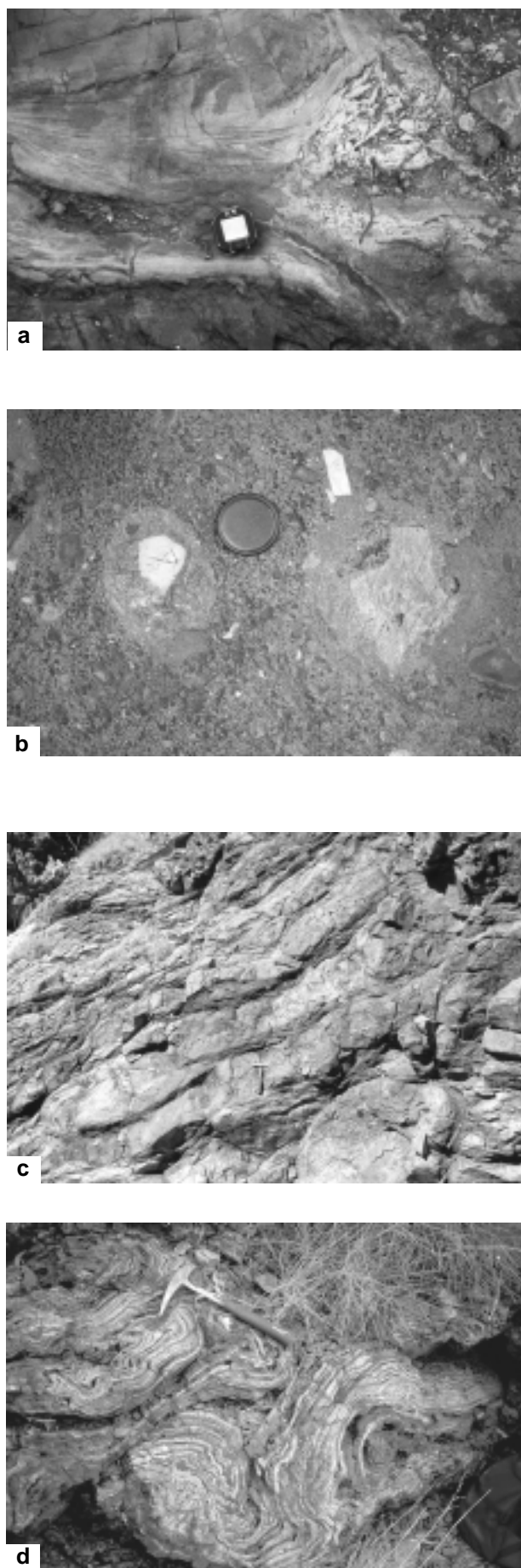


Figure 23. Magma mingling textures at the base of the inner phase of the Strelley Granite: a) mingling between

contain up to 10% of evenly distributed, but irregularly shaped, patches of carbonate.

Dacite near the top of the Kangaroo Caves Formation in the Soanesville Belt (*AScfd*) is weakly feldspar-phyric, commonly spherulitic, and locally amygdaloidal or spherulitic or both. The 250 m-thick dacite sill at the top of the volcanic pile between Sulphur Springs and Kangaroo Caves has a pepperitic upper contact, indicating intrusion into wet sediment. Below this contact, the dacite is commonly lobate consisting of metre-thick, coherent bodies in finer grained, sheared dacite (Fig. 22c). The base of this unit is massive and spherulitic with typically elongate amygdales up to 3 cm long. Rhyodacite sills northeast of the Roadmaster prospect (AMG 237585) are also spherulitic. Rhyolite (*AScfr*) is quartz- and feldspar-phyric with coarse leucoxene (after magnetite?) and accessory ilmenite and apatite. The rhyolite forms as both amygdaloidal, locally flow banded lavas and graded pumice breccia beds (*AScfa*) with volcanoclastic sandstones at their tops. In the southern part of the Soanesville Belt (around AMG 162295), felsic volcanic rocks are more strongly deformed and commonly transformed into nondescript quartz-sericite schists (*AScf*), whose original composition could not be determined.

The marker chert at the top of the Sulphur Springs Group (*AScc*) is up to 100 m thick. The chert is typically composed of centimetre-layered, grey-blue and white, silicified, fine-grained volcanoclastic (including andesite shard-rich sandstone) and epiclastic (including black mudstone, sandstone, and breccia) detritus intruded by discordant veins, and concordant sills of white chert and black, kerogenous chert. A thin (<1 m) basal unit of shard-rich sandstone is locally developed. Local quartz sandstone and banded iron-formation occur in part of this unit southeast of the Strelley Granite. Vearncombe (1996) presented a detailed account of the vein paragenesis involved in the transformation of this unit into chert.

At Sulphur Springs, the marker chert is overlain by a unit of polymict megabreccia (*AScsx*) across a sharp, erosional contact. The megabreccia is up to 600 m thick and commonly massive, although it locally displays crude bedding. The megabreccia is composed of blocks and fragmented rafts of felsic volcanic rocks, silicified fine-grained sediment (chert), banded iron-formation (*ASci*), and rare, massive sulfides within a matrix of fine-grained, grey mudstone and volcanic ash. Detailed maps of this unit were presented by Hill (1997) and Vearncombe et al. (1998). Clasts vary in size from centimetre scale to hundreds of metres in diameter, to kilometre-long panels of dismembered chert and banded iron-formation. The matrix is fine grained with up to 10% angular to subangular, scalloped shards of clear, monocrystalline volcanic quartz and variably altered fragments of felsic volcanic rock, including glassy lithic fragments. In some sections, no bedding could be observed in the matrix, but textural variations were observed within angular, blocky fragments identified by slightly different ratios of quartz detritus. The presence of abundant volcanic quartz and felsic volcanic detritus and the unbedded, blocky texture of the matrix suggest that much of the matrix is a volcanic ash. Parts of the matrix are cut by irregular to curving quartz-filled cracks, some of which change shape and

orientation across internal matrix boundaries. Such cracks are probably syneresis cracks, thought to be caused by earthquake-induced dewatering (Pratt, 1998).

Within the upper part of the megabreccia is a unit of red, black, and white layered, ferruginized siltstone and banded iron-formation (*ASci*). The top contact of the unit is eroded by the megabreccia and intruded by megabreccia in neptunian dykes. The ferruginous sediment layer is deformed at a broad scale into a tight, easterly facing synsedimentary fold in the eastern part of the megabreccia. Whereas bedding in the nose of the fold is regular and well preserved, bedding on the southern limb is strongly deformed into a contorted set of chaotic synsedimentary folds (Fig. 22d), and then ultimately into a tectonic conglomerate in which more-competent white layers have been boudinaged into a series of isolated, white silica lenses and blobs surrounded by a red silica matrix. To the south, the southern limb is cut out within massive megabreccia.

In the core of the synsedimentary slump fold are two rhyodacite sills that show internal quartz-phyric, spherulitic, and perlitic textures, and marginal rinds of pepperite. Both sills are capped by a thin unit of silicified siltstone (chert) derived from pepperite and volcanic ash. To the east, both the felsic sills and cherts become truncated by, and disaggregated within, the megabreccia.

At Kangaroo Caves, feldspar-phyric, calc-alkaline rhyodacite (*AScfr*) forms an elongate dome, 400 m high by 1500 m wide, above the marker chert and its associated mineralization. The dome is capped by another bed of silicified siltstone (chert), which is unmineralized. The dome is mostly massive to lobed, with amygdaloidal and perlitic textures, but contains volcanoclastic rocks at its top. Primary ferromagnesian minerals in these fine-grained rocks are altered to chlorite or sericite, and opaque minerals are pseudomorphed by leucoxene.

## Comagmatic microdiorite (*ASgd*)

At, or near, the base of the Kangaroo Caves Formation, along the upper contact of the Strelley Granite, is a sill of microdiorite (*ASgd*). This sill is a medium-grained, equigranular rock composed originally of pyroxene and plagioclase with accessory magnetite and apatite. Plagioclase is typically lath shaped, but locally plumose, defining an interlocking crystalline texture. Original clinopyroxene has been incipiently altered to actinolite and magnetite is commonly oxidized. Around the northern contact of the Strelley Granite, the microdiorite contains 2–3 cm-large epidote alteration spots. The microdiorite is cut by the outer phase of the Strelley Granite.

## Strelley Granite (*Agste*, *Agstey*, *Agsta*, *Agstp*)

The synvolcanic Strelley Granite is genetically related to the Kangaroo Caves Formation (Vearncombe, 1996; Van Kranendonk, 1997; Brauhart, 1999). Three compositionally and texturally distinct phases formed in a synvolcanic laccolith, which is exposed in cross section as a result of tilting associated with its northerly displacement during

D<sub>3</sub> deformation at c. 2940 Ma (see **Structure**). The three textural phases include: the main outer phase of medium-grained, equigranular to weakly porphyritic hornblende–biotite monzogranite (*Agste*); an outermost rind of microcrystalline granophyre (*Agstey*) at the top of the granite, differentiated from the outer phase; and a younger inner phase of commonly porphyritic hornblende–biotite monzogranite (*Agstp*). Both the outer and inner phases contain miarolitic cavities, and the original mineralogy of these rocks is extensively altered to chlorite–sericite (Brauhart, 1999). There is a physically separate and texturally distinct phase of the granite (*Agsta*) to the south of the laccolith in the southwestern part of NORTH SHAW.

The outer phase of the Strelley Granite laccolith is an equigranular intergrowth of coarse-grained quartz–K-feldspar–plagioclase with 3–5 mm aggregates of hornblende. K-feldspar laths are rimmed by albite (rapakivi texture), and plagioclase laths are invariably sieved with fine-grained sericite, zoisite, and minor carbonate. Hornblende is commonly rimmed by fine-grained, green–brown pleochroic biotite and rutile. The granophyre is a fine-grained intergrowth of feldspar, quartz, and mafic minerals that are thoroughly altered to chlorite–sericite. The outer rind of the laccolith, like the adjacent microdiorite sill (*ASgd*), commonly has up to 1.5 cm-large epidote alteration spots. Primary magnetite is commonly altered to rutile, titanite, or hematite. West of the laccolith, numerous thin sills of outer phase granite intrude the lower parts of the Sulphur Springs Group.

The inner phase of the Strelley Granite laccolith is typified by centimetre-size, rounded K-feldspar phenocrysts or xenocrysts rimmed by albite, although non-porphyritic areas are also present. In thin section, phenocrysts of K-feldspar, plagioclase, and quartz lie in a groundmass of fine-grained, intergrown K-feldspar, quartz, plagioclase, and biotite, with accessory magnetite, apatite, and zircon. Fine-grained biotite is the dominant mafic mineral, but biotite and hornblende were locally observed together. The rocks of this phase are commonly very fresh, although in thin section, plagioclase is invariably sieved with fine-grained zoisite, sericite, and minor carbonate.

The upper and northern parts of the outer phase of the Strelley Granite laccolith are characterized by numerous parallel to gently radiating joints oriented perpendicular to the top contact of the granite (see accompanying map). These joints are readily visible on airphotos, but less easily recognized in the field. However, in areas of continuous exposure, the joints can be seen to form along greisen veins, which contain massive sulfides in their core (Brauhart, 1999). Mineralization includes vein-hosted Cu–Zn–Pb–Ag–Sn (e.g. Wheal of Fortune prospect), gold, and gold with disseminated sulfides. In the uppermost part of the inner phase, greisen-hosted disseminated Mo has been located (AMG 275530).

West and south of the Strelley Granite (e.g. AMG 145310), sheets and sills of fine-grained felsic rock (*Agsta*) display a weak primary layering at a metre to decametre scale. Layering varies between aphanitic featureless rocks and those with a discernible igneous texture defined by flow aligned laths of plagioclase feldspar. In the eastern part of this unit, rocks have a bimodal texture of coarse

(1–5 mm), rounded, and strongly zoned plagioclase within a finer grained matrix of quartz, feldspar, green to straw-yellow pleochroic biotite, and opaque minerals. In the middle of the unit, the rock is dominated by plagioclase, which forms as fine- to medium-grained, stubby laths that have a weakly interlocking texture. Biotite in this rock is brown and rimmed by chlorite, and there is a moderate development of sericite and carbonate. Further west, the rock is composed primarily of quartz, with lesser amounts of microcline and some granophyre. Grain boundaries are strongly sutured and anastomosing strings of sericite are developed. Along the western contact, the rock consists of a fine-grained mosaic of sutured and recrystallized subgrains of quartz, plagioclase, and K-feldspar, with coarse bladed sheaths of intergrown epidote and sericite, as well as chlorite and titanite. These rocks all contain biotite and are interpreted, on the basis of textures observed in thin section, to represent a layered, subvolcanic granitoid intrusion. These are monocyclic rocks in which a single, penetrative schistosity of variable intensity has developed.

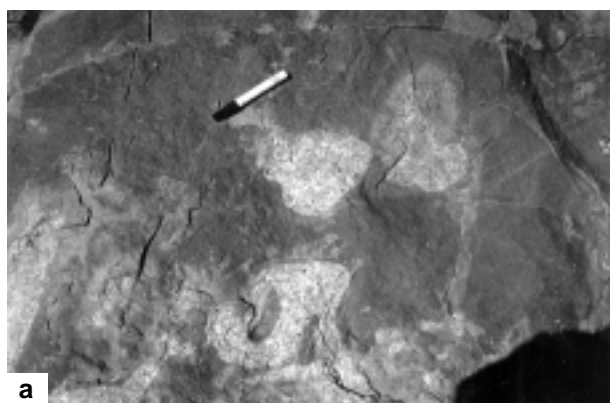
## Comagmatic dolerite (*ASd*)

The base of the inner phase of the Strelley Granite laccolith (*Agstp*) is mingled with a unit of comagmatic dolerite (*ASd*) across a hybrid zone up to 200 m wide. When approaching the dolerite from the granite, one first sees spherical inclusions, 10 cm to 1 m in size, of grey, hybrid granitoid rock that consists of aggregations of rapakivi-textured feldspar and quartz phenocrysts within a matrix of actinolite, plagioclase, and quartz. Further west, the hybrid rock occupies progressively larger areas, which start to host inclusions of fine-grained dolerite. The granite–dolerite contact is characterized by mutually crosscutting relationships (Fig. 23a) and by disequilibrium textures, such as reaction haloes in dolerite around granite inclusions and the presence of rapakivi-textured K-feldspar phenocrysts in contaminated dolerite. In dolerite marginal to the granite, disequilibrium textures of actinolite rims on quartz phenocrysts have developed (Fig. 23b). West of the Strelley Granite, intrusive sheets, sills, and dykes of uncontaminated, fine- to medium-grained dolerite intrude rocks belonging to the lower parts of the Sulphur Springs Group.

## Gorge Creek Group

### Pincunah Hill Formation (*AGih*, *AGihc*, *AGii*)

Rocks of the Pincunah Hill Formation lie conformably over the Kangaroo Caves Formation and conformably under the Corboy Formation in the Pincunah Belt and the southern part of the Soanesville Belt. At the base of the formation in the Pincunah Belt, there is up to 1 km of red and black, thinly bedded banded iron-formation (*AGii*), which includes subordinate, black and white layered chert, purplish-red to grey shale, and siltstone. Up to 5 m of soft, yellow-weathering felsic tuff and tuffaceous siltstone are exposed in the central part of the banded iron-formation, in the southern part of this unit.



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comagmatic dolerite and granite. Way up is to the top of the photo. Note the reaction rinds in dolerite around the granite inclusions, and the diapiric form of the lower granitoid lobe into the dolerite; b) disequilibrium textures in dolerite near the granite contact of actinolite rims on anhedral quartz phenocrysts

Figure 24. a) Cross-polarized thin section view of a harzburgite of the Dalton Suite showing poikilitic inclusion of

Overlying the banded iron-formation in the Pincunah Belt, but forming the base of the unit in the Soanesville Belt, is dark-red to purple ferruginous shale (*AGih*) with interbedded siltstone, local red and black banded iron-formation, and minor amounts of chert, sandstone, and conglomerate. Centimetre-bedded, variegated chert after shale (*AGihc*) formed along the northern limb of the Pilgangoora Syncline and in the Tambina Structural Complex along the southern margin of NORTH SHAW. These shales were silicified into high-standing ridges due to a combination of events including diagenesis and possibly Cainozoic weathering.

### Corboy Formation (*AGcb*, *AGcc*, *AGcg*, *AGcq*, *AGcs*, *AGct*)

The thickest succession of the Corboy Formation is in the Pincunah Belt, where several distinct lithofacies show significant variations in stratigraphic thickness across major faults, indicating fault activity during sedimentation (Wilhelmij and Dunlop, 1984; Wilhelmij, 1986). In the far

west of NORTH SHAW, the basal unit is represented by about 80 m of thin-bedded quartz sandstone that passes sharply up into 200 m of poorly sorted, thick-bedded pebble to cobble conglomerate and interbedded sandstone (*AGcc*), which shows graded beds and local fining-upward sequences. These rocks are overlain by about 20 m of red shale, and then about 400 m of coarse sandstone with local conglomeratic beds (*AGct*). Overlying this are about 200 m of sandstone, which is interbedded with conglomerate at the base and shale higher up. These pass up into massive sandstone and conglomerate.

In the central part of the Pincunah Belt (AMG 134507), banded iron-formation of the Pincunah Hill Formation (*AGii*) is overlain by a transitional basal unit, about 100 m thick, containing — from base to top — several metres-thick beds of grey quartzite interbedded with sandy siltstones, centimetre-bedded grey-brown siltstones, fine-grained quartzites, interbedded grey-brown wackes with 30 cm-thick conglomerate beds, and shaley siltstones (*AGct*). A lens of white quartz arenite and quartzite (*AGcq*) is present further south in the belt. These rocks are overlain by a homogeneous unit ( $\leq 960$  m thick) of brown-weathering lithic sandstone or greywacke (*AGcg*). Although this upper unit appears well bedded at a decametric scale on airphotos, bedding is difficult to see on the ground and can be determined only by the strike of ridges. These rocks commonly have sparsely distributed, angular pebbles of black chert supported in a medium sandy matrix, but these do not define bedding. At the top of this unit is less than 20 m of red shale overlain by a 5 m-thick bed of densely packed pebble to cobble conglomerate, in which pebbles are well sorted and well rounded. Pebbles consist mainly of black chert, finely layered brown-grey chert, and white quartz, but less commonly include white quartzite and subangular fragments of felsic volcanic rock. The conglomerate is directly overlain by a bed of felsic volcanoclastic sandstone and then by interbedded pebbly sandstone and wacke. These rocks, in turn, pass up into tightly folded, red-weathering, greenish-grey siltstone and shale with sandstone interbeds, the thickness of which is impossible to determine accurately. The shales are overlain by 100–200 m of quartz sandstone and thinly bedded sandstone, a thin bed of red shale, and then thick-bedded (5–10 m) quartz arenite with glassy quartz detritus.

In the East Strelley Belt, sandstones (*AGct*) of the Corboy Formation form the base of the Gorge Creek Group. The formation is about 500 m thick and grades up from medium-grained sandstones with rare, granule conglomerate beds, to thin-bedded sandstones and shales. Cross-bedding is developed in several places. In thin section, the sandstones are seen to be well sorted with angular to subangular quartz clasts and rare, detrital muscovite in a fine-grained sericite matrix containing secondary oxides. These rocks have been recrystallized such that mutually touching quartz grains have long contacts and common 120° triple points.

Thick-bedded turbiditic sandstones of the Corboy Formation (*AGct*) form the base of the Gorge Creek Group in the northern part of the Soanesville Belt, where they only the top marker chert of the Sulphur Springs Group around the northeastern end of the Strelley Granite. The

formation is up to 475 m thick in this area, but has a lensoid shape, thinning out along strike to the north and west over the top of the olistostrome breccia at the top of the Sulphur Springs Group (*AScsx*; see Van Kranendonk, 1997), and also thinning to the south over the top of the felsic volcanic dome at the Kangaroo Caves prospect. The Corboy Formation is also on the western side of the breccia as a less than 100 m-thick unit of coarse sandstone, indicating that the breccia was a topographic high during deposition of the Corboy Formation. Immediately above the thin eastern wedge of the olistostrome breccia, the Corboy Formation includes a basal matrix-supported conglomerate (*AGcc*), up to 30 m thick, which is interlayered with lenses and beds of coarse sandstone. The conglomerate contains subrounded cobbles and pebbles of layered black chert derived from the underlying marker chert bed. However, the main part of the formation in this area is composed of coarse sandstone and sparsely pebbly sandstone that is bedded on a 1–5 m scale and displays minor fining-upward textures within beds (Hill, 1997, fig. 3.2). The coarse, thick-bedded turbiditic sandstones gradually fine upwards to sandstones and siltstones at the top, and these have a gradational relationship into the basal silts and shales of the Paddy Market Formation. In this area, the Corboy Formation was interpreted by Wilhelmij (1986) and Hill (1997) to represent small, prograding fan deltas and related turbidite systems infilling a valley in a shallow-marine setting.

The Corboy Formation also drapes over the felsic volcanic dome at the Kangaroo Caves prospect. At this locality, a thin, basal unit of pebbly conglomerate along the flanks of the dome displays excellent cross-bedding. Neptunian sandstone dykes cut down into the marker chert at the top of the Sulphur Springs Group (*AScc*), indicating that the marker chert was already silicified and lithified prior to deposition of the Gorge Creek Group. This feature, and the onlap of the Corboy Formation, indicates that there was at least some time break in between the Sulphur Springs and Gorge Creek Groups. Further south within the Soanesville Belt, the Corboy Formation is predominantly composed of sandstones and lesser amounts of interbedded siltstone.

North and south of the Bernts prospect (AMG 364503), foliated sandstones and siltstones are tectonically interleaved with mafic and ultramafic schist (?Dalton Suite) within the Bernts Deformation Zone. The scale of interlayering is too fine to show at the map scale and thus the various rock types have been lumped together (*AGcb*).

In south-central NORTH SHAW (AMG 300210), a tectonic sliver of Gorge Creek Group rocks have been turned up on-end and folded. Initially described by Boulter et al. (1987), these rocks contain an unconformity between underlying centimetre-layered, blue-grey and white layered cherts (silicified shales) and overlying sandstone and cobble to pebble conglomerate (*AGcs*). The younger, coarse clastic sequence was named the Daltons Sandstone by Boulter et al. (1987) and equated with the Lalla Rookh Sandstone. However, Van Kranendonk (1998) argued on the basis of field and structural data, that these rocks are different from, and older than, the Lalla Rookh Sandstone, suggesting instead that they represent part of

the Corboy Formation. The unconformity at the base of this unit varies between having a high-angle and subconformable contact with the underlying cherts, and is locally overturned. The unconformity is marked by a basal angular breccia and 10 cm-thick zone of ferruginous weathering of sandstone, and overlain by pebble conglomerate. Conglomerates vary from well-rounded cobble conglomerate to subangular pebble conglomerate, but all are matrix supported. Quartzites form higher up in the sequence.

## Paddy Market Formation (*AGph*, *AGphc*, *AGpi*, *AGpf*)

Rocks of the Paddy Market Formation lie conformably above the Corboy Formation in the Soanesville and East Strelley Belts, and also in the Panorama and North Shaw Belts and Tambina Structural Complex. In the East Strelley Belt, the formation consists of thinly bedded, black, white, red, and grey cherty banded iron-formation and minor ferruginous shale (*AGpi*). This passes along strike to the east into ferruginous shale and mudstone with local siltstone and sandstone (*AGph*). Similar rocks are gradationally above the Corboy Formation in the Soanesville Belt, where they are overlain by the Honeyeater Basalt.

In the Tambina Structural Complex and North Shaw Belt, the Paddy Market Formation consists of tightly folded, centimetre-bedded, grey and white layered chert after shale (*AGphc*). There are similar rocks in the southwestern part of the Panorama Belt at Agrippa Ridge, and here it can be seen that the silicification is due to Cainozoic weathering, as cherts high up on the ridge pass along strike directly into red, ferruginous shales in a valley. The basal part of the Paddy Market Formation at this locality is heterogeneous, consisting of — from base to top — finely layered chert, weakly layered banded iron-formation, fissile red and green shale, centimetre-layered black, white, and brown chert (after shale), up to 10 m of sandstone, up to 100 m of dark-red and black banded iron-formation, and then more chert. The stratigraphically highest chert contains felsic volcanic breccias and silicified, well-bedded felsic volcanoclastic tuffs that underlie up to 200 m of massive to spherulitic dacite (*AGpf*) of uncertain intrusive or extrusive origin.

## Honeyeater Basalt (*AGh*)

The Honeyeater Basalt conformably overlies the Paddy Market Formation in the Soanesville Belt and in the Panorama Belt at Agrippa Ridge. In the core of the Soanesville Syncline, the Honeyeater Basalt (*AGh*) consists of high-Mg basalt and dolerite in the lower and central parts of the formation, whereas at higher levels dolerite is intercalated with tholeiitic basalt (Glikson and Hickman, 1981b, fig. 2.1d). Massive high-Mg basalts may contain centimetre-sized ocelli. The central and upper parts of the formation are pillowed, as is the formation east of the Strelley Granite, where pillows may contain coarse gas cavities. Metamorphism is of very low grade with fine-grained prehnite–pumpellyite visible in some

sections. Some basalt contains amphibole and serpentine pseudomorphs, probably after hypersthene phenocrysts, with the remaining minerals being augite and saussuritized plagioclase, with minor chlorite, clinozoisite, epidote, and secondary quartz.

Glikson and Hickman (1981a,b) analysed 40 samples of the Honeyeater Basalt from the Soanesville Syncline. They found that the unit contained both tholeiitic basalts (MgO approximately 6.3%) with high  $\text{Al}_2\text{O}_3$  (approximately 14.3%), and high-MgO basalts (approximately 11%) with low  $\text{Al}_2\text{O}_3$  (approximately 14.2%), and that both types had characteristically high Ba and Sr, and low  $\text{TiO}_2$  (Glikson et al., 1991). Glikson et al. (1986) presented further analyses of the Honeyeater Basalt from just west of the Bernts prospect (AMG 357515). They found that the unit contained  $7.45\% < \text{MgO} < 12.2\%$ , and had flat, normalized REE patterns at about 10 times primitive mantle values.

### Dalton Suite (*AaDo*, *AaDlx*, *AaDd*, *AaDp*, *AaDpd*, *AaDx*)

A number of layered ultramafic–mafic sills lie within the Gorge Creek and Sulphur Springs Groups and the Strelley Granite in the Soanesville Belt. Thin dykes (*AaDd*) shoot off the sill within the Strelley Granite, cutting up through the granite and overlying volcanic rocks of the Kangaroo Caves Formation, into the base of the Gorge Creek Group. Individual sills within, and to the east of, the Strelley Granite are up to 300–400 m thick, but reach a maximum thickness of 475 m in the Soanesville Syncline.

The sill within the Strelley Granite is differentiated into a lower (western) pyroxenite (*AaDx*; olivine websterite to harzburgite) and an upper (eastern) gabbro (norite (*AaDo*). Sills in the Gorge Creek Group east of Strelley Granite are differentiated upwards from olivine websterite at the base, through Mg-rich metadolerite, to norite at the top. In the Soanesville Syncline, sills are differentiated upwards from dunite ( $\leq 250$  m; *AaDpd*), through peridotite (*AaDp*), pyroxenite (*AaDx*), gabbro (norite (*AaDo*), to pyroxene leucogabbro (*AaDlx*) and anorthosite, and even locally, granophyric granite (not shown on map).

Olivine cumulate textures are preserved in the basal parts of the sills, with variable amounts and proportions of intercumulate clinopyroxene and orthopyroxene (Fig. 24a). Olivine is extensively altered to serpentine–magnetite in these rocks, whereas pyroxenes are typically fresh. In the upper, more mafic parts of the sills, alteration is much more extensive with saussuritization of plagioclase, bastite alteration of orthopyroxene, and growth of zoisite, fine-grained chlorite, and actinolitic amphibole. Vearncombe (1996) described prehnite from sills east of the Strelley Granite. At the top of one of these sills, an unusual granitic differentiate was observed, which contains cauliflower-textured quartz crystals in a fine, reticulated network of intergrown plagioclase and subordinate K-feldspar (Fig. 24b).

Flat, normalized REE patterns similar to Honeyeater Basalt were obtained from samples of the Dalton Suite ultramafic sill (*AaDx*) cutting the Strelley Granite (Brauhart, 1999, fig. 3.4r).

### Pyramid Hill Formation (*AGyh*, *AGyhc*, *AGyi*)

Shale and banded iron-formation of the newly identified Pyramid Hill Formation (Van Kranendonk and Morant, 1998) overlie the Honeyeater Basalt to the east of the Strelley Granite and in the core of the Soanesville Syncline. Shales of this formation (*AGyh*) are black to reddish in colour and commonly variegated. They are also commonly silicified to centimetre-layered, blue-black and white cherts (*AGyhc*). Two observations suggest that this silicification predated deposition of the unconformably overlying Lalla Rookh Sandstone. First, a ridge of silicified Pyramid Hill Formation shale forms a palaeohill at the base of the Lalla Rookh Sandstone 4 km northeast of the Sulphur Springs prospect (AMG 308620). On the northwestern side, this ridge is unconformably overlain by an unsorted, coarse, angular breccia (scree slope deposit) that contains fragments of the silicified shale, up to 50 cm in length, in a coarse, non-silicified sandy matrix. This, and the fact that there are similar clasts throughout the basal conglomerate of the Lalla Rookh Sandstone, suggests that silicification of Pyramid Hill shales occurred prior to deposition of the Lalla Rookh Sandstone.

Banded iron-formation (*AGyi*) forms the top of the Gorge Creek Group along the southern part of the Lalla Rookh Synclinorium and in the core of the Soanesville Syncline. These rocks are dark red-brown in colour and well layered at a 2–5 cm scale, with alternations of hematite-rich and cherty beds.

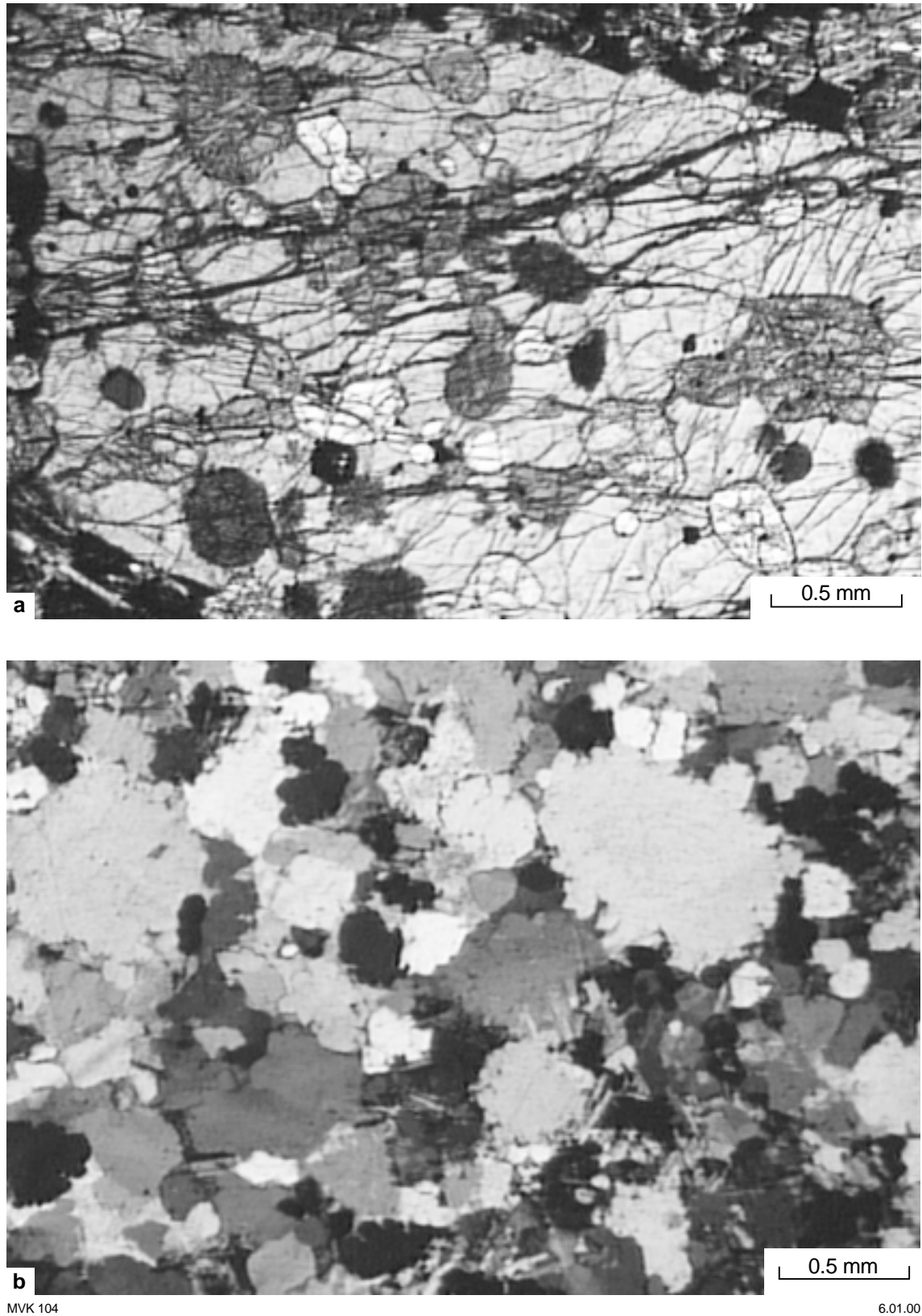
## De Grey Group

### Lalla Rookh Sandstone (*ADls*, *ADlc*)

The Lalla Rookh Sandstone consists of matrix-supported pebble to boulder conglomerate and diamictite (*ADlc*), and pebbly sandstone, quartzose arenites, feldspathic sandstone, and minor shale (*ADls*). Conglomerate clasts include primarily black chert, layered black and white, or black and grey chert, banded iron-formation, white quartz, various metasandstones, rare felsic metavolcanic rocks, and very rare pebbles of mafic–ultramafic schist.

Rocks of the Lalla Rookh Sandstone lie unconformably on rocks of the Gorge Creek Group in five localities, the largest being the Lalla Rookh Synclinorium in the north-central part of NORTH SHAW. This formation also outcrops in the East Strelley Belt, the Leilira Synclinorium in the Soanesville Belt, the Keep It Dark Synclinorium in the Tambina Structural Complex, and the fault-bounded syncline between the North Pole and Shaw Domes.

The unconformity at the base of the Lalla Rookh Synclinorium, east of the Strelley Granite, is subparallel to bedding in the underlying Gorge Creek Group, but cuts about 950 m down-section from northwest to southeast along strike as a result of erosion through the Pyramid Hill Formation and most of the Honeyeater Basalt. Islands of resistant chert 200 m high are present at the base of the conglomerates and are flanked by angular, boulder scree



fresh olivine in orthopyroxene megacryst. Width of view is 3.5 mm; b) granitic differentiate of the Dalton Suite, showing 'cauliflower' growth of quartz and blocky laths of plagioclase

Figure 25. Leucogranite diatexite from the Shaw Granitoid Complex: a) granitic pegmatite dyke formed from

deposits. At one locality (AMG 323604), the basal unconformity consists of a 20 cm-thick conglomerate bed lying unconformably on bleached, but undeformed Honeyeater Basalt, then about 3–4 m of fine-grained quartz–sericite–titanite felsic volcanic rocks with preserved perlitic cooling cracks, and then overlies cobble conglomerates of the Lalla Rookh Sandstone proper. Some 30 m along strike to the south, altered and ferruginized Honeyeater Basalt is overlain by a 2–5 m-thick unit containing blocks of chaotically folded and brecciated blue-black and grey chert (*AGyhc*) within a fine-grained, grey matrix that has a distinctly scoriaceous texture. In thin section, the matrix of this rock contains fragments of clear quartz crystals with embayed edges, in addition to angular fragments of the chert, and is vesicular, indicating a volcanic origin. The large blocks of chert in the grey volcanoclastic rock occupy topographic highs in the overlying conglomerates of the Lalla Rookh Sandstone and are draped by the basal beds of this formation. The age of these basal units is unknown.

The Lalla Rookh Sandstone reaches a maximum thickness of about 3000 m along the western margin of the Lalla Rookh Synclinorium (Krapez, 1984). The basal conglomerate contains packed (60–70%), well-rounded cobbles up to 20 cm in diameter within a coarse sand matrix. Bedding is at a metre scale or more and decreases gradually upwards, as the grain size decreases. Scours and channels are commonly visible. A significant intraformational unconformity (AMG 336646) suggests that folding initiated early during progressive sedimentation. The matrix of the conglomerates is coarse sand, which in alluvial fan deposits at the base of the sequence is feldspar free. Sandstones further up in the succession contain altered microcline; fresh microcline is present higher up in the sequence (Krapez, 1984). This upward change from lithic fanglomerates to quartzose sandstone, to feldspathic sandstones indicates greater exposure and erosion of granitoid rocks (probably the Strelley Granite) throughout basin evolution (Krapez, 1984). Krapez (1984) recognized alluvial-fan, braided-stream, floodplain, and fan-delta deposits, and interpreted the Lalla Rookh Synclinorium to represent a depositional basin formed in an en echelon dilational jog within a sinistral fault system. However, Van Kranendonk and Collins (1998) showed that the geometry of faults bounding the Lalla Rookh Synclinorium was incompatible with a dilational jog origin, and proposed that it represents a small trough developed in advance of the north-moving Strelley Granite during regional  $D_3$  deformation (see **Deformation**).

In the Keep It Dark Synclinorium, conglomerates drape off topographic highs of the silicified Pincunah Hill Formation. Sedimentological studies indicate that these outcrops of the Gorge Creek were already topographic highs and that they sourced proximal coarse detritus in alluvial fans during deposition of the Lalla Rookh Sandstone (Chan, 1998). The predominant fill of the basin is a sublitharenite and subordinate lithic feldspathic arenite in the south, derived from mixed granitic and volcanic sources, as well as recycled, sedimentary source rocks. Palaeocurrent analysis of this synclinorium suggests input of detritus from flanking highs and transport along the basin axis from the southeast (Chan, 1998).

## Keep It Dark Monzogranite (*Agk*)

Weakly foliated to undeformed monzogranite cuts amphibolite-facies greenstones of the North Shaw Belt (AMG 340275), both as a discrete pluton and flanking dykes. The pluton has intrusive northern and southern contacts, but is bound by north-northwesterly striking faults in the east and west. Another fault cuts right through the central part of the granite, but does not offset its intrusive northern contact, indicating that faulting and granite intrusion were synchronous at c. 2936 Ma (see **Geochronology**, p. 73).

The Keep It Dark Monzogranite varies from a medium-grained, equigranular rock in the east and south, to a slightly coarser grained and weakly porphyritic variety in the northwest, where it is least deformed and contains faint, igneous flow alignment of K-feldspar crystals. The eastern part of the granite is transected by a weak ( $D_3$ ) foliation defined by aligned chloritized micas, which is steeply dipping and subparallel to the bounding faults. In thin section, the rock has a medium-grained, interlocking texture of anhedral crystals of quartz, microcline, and plagioclase with highly sutured grain boundaries. Quartz shows ubiquitous undulatory extinction and minor subgrain development due to recrystallization. The composition of the plagioclase is  $An_{25}$  (oligoclase). The rock contains about 3% dark-green pleochroic chlorite after biotite, about 2% sericite, and trace epidote, titanite, carbonate, and zircon.

## Granitoid complexes

### Carlindi Granitoid Complex (*AgLwi*, *AgLm*, *AgLI*, *AgLp*, *AgLmh*, *AgLg*)

The Carlindi Granitoid Complex on NORTH SHAW is a weakly deformed body of granitoid rocks with well-preserved intrusive contacts into the surrounding Coonterunah Group. In the west, granitoid rocks intrude to high structural levels within a small cupola, and are unconformably overlain by the Strelley Pool Chert (Buick et al., 1995). Local, fine-grained hornfels is developed along some granite contacts, and in other places, granite intrusion caused the growth of feldspar clots in the host amphibolite. The fine- to coarse-grained, equigranular texture of amphibolites surrounding the granitoid complex reflects contact metamorphism associated with the emplacement of this body.

The eastern half of the Carlindi Granitoid Complex is underlain by homogeneous, leucocratic, equigranular biotite–hornblende granodiorite–granite (*AgLg*). The rock is composed of 5% of biotite–epidote–titanite–muscovite–actinolite clots in a coarse-grained matrix of quartz–plagioclase–K-feldspar (microcline and antiperthite). K-feldspars commonly show igneous growth zoning, whereas plagioclase does not. Both feldspars are heavily saussuritized, with local, coarse growth of epidote. Very coarse grained titanite is observed in some sections.

The dominant lithology in the western half of the Carlindi Granitoid Complex is a weakly foliated biotite–hornblende leucogranite (*AgLI*), which outcrops in flat

platforms. This rock is characterized by 5–8% of shiny, rectangular prisms of black hornblende up to 5 mm in length in a medium-grained, interlocking assemblage of 35–40% of normally zoned plagioclase feldspar (up to 1 cm in length), 30% K-feldspar, and 25–30% quartz. The hornblende is commonly altered to biotite–muscovite–epidote (2–3%). The cores of plagioclase crystals are heavily sericitized with local epidote, whereas the outer, typically untwinned rims are clear. The composition of one homogeneous plagioclase crystal is oligoclase. K-feldspar consists of microperthite and microcline, with local myrmekite, and is typically cloudy with epidote alteration, although many still show igneous growth zoning. Quartz has serrate grain boundaries with feldspars, undulose extinction, and 120° triple point boundaries with other quartz grains. The rock is unfoliated and only the undulose extinction and subgrain boundary formation in quartz attests to it having been affected by tectonic strain. A sample of this unit (GSWA 153190) has been dated at  $3469 \pm 2$  Ma (Nelson, 1999; see Table 3).

In between the western and eastern halves of the Carlindi Granitoid Complex are two other distinct varieties of granitoid rock, including hornblende–phyric monzogranite (*AgLmh*) in the south and schlieric to gneissic tonalites of the Wilson Well Gneiss in the north (*AgLwi*). The monzogranite is more mafic than the leucogranite further east, but is also a monocyclic rock, locally only weakly foliated. Compositions vary from hornblende monzodiorite to hornblende granite. The igneous hornblende is retrogressed to epidote–chlorite, and plagioclase is altered to sericite and epidote.

The Wilson Well Gneiss is a heterogeneous unit of weakly migmatitic tonalite gneiss containing sparse enclaves of homogeneous, medium-grained quartz diorite, and is cut by numerous dykes of various texture and composition. Typically, the bulk of the unit is well foliated and contains less than 10%, 0.5–2 cm-wide leucogranite veinlets that define a gneissic layering. The host rock is medium grained and thoroughly recrystallized to an equigranular, granoblastic texture. Relicts of feldspar–porphyritic texture are locally preserved. As many as six intrusive granitoid phases were recognized in outcrops of the Wilson Well Gneiss, in addition to enclaves of still older amphibolite. These include hornblende tonalite (phase 1) and component migmatite veins or local swarms of pegmatitic granite (phase 2). Cutting these early phases is a homogeneous, medium-grained hornblende granodiorite–granite (phase 3). These more voluminous components are cut by narrow dykes of strongly foliated, fine-grained, grey granite whose fabric is syntectonic with emplacement, since neither older nor younger components contain this foliation. The fine-grained dykes are cut by dykes of leucogranite (phase 4) and pegmatitic granite (phase 5), and these are cut by dykes of schlieric, wispy-layered leucogranite (phase 6). None of the younger intrusive phases (2–6) in these outcrops can be directly correlated with the main phases of the rest of the granitoid complex. An attempt at dating the older gneissic tonalite was unsuccessful.

Granitoid rocks that intrude to higher structural levels — for example, in the granite cupola in the west (AMG 083618), in the centre (AMG 160644), and in the

east (AMG 352703) — have a distinctly porphyritic texture (*AgLp*) defined by up to 6 mm large phenocrysts of clear, unstrained quartz, plagioclase, K-feldspar, and muscovite in a fine- to medium-grained quartzofeldspathic (+ local magnetite) matrix. Feldspar phenocrysts are commonly euhedral, with growth zoning around their edges, whereas quartz phenocrysts are typically round and locally show resorption embayments. In the western cupola, Buick et al. (1995) dated a sample of this rock at  $3468 \pm 4$  Ma (Table 3), within error of the deeper level leucogranite it was seen to cut.

Another distinctive phase of the complex is a blue-grey, fine-grained biotite monzogranite (*AgLm*), which outcrops in two localities in the northwestern corner of NORTH SHAW (AMG 082760 and 145725). This unit forms a distinctive area of outcrop characterized by rounded boulders with dark-brown weathering rinds, distributed in pyramidal tors of delicately balanced boulders. The rocks are commonly silicified and ring loudly when struck with a hammer. No foliation was observed in this unit. The monzogranite is identical in appearance to the Mulgandinnah Monzogranite of the Shaw Granitoid Complex, but has a very different age, having been dated at  $3484 \pm 4$  Ma (Nelson, 1999; GSWA 153188).

### Shaw Granitoid Complex (*AgSn*, *AgSns*, *AgSco*, *AgSl*, *AgSg*, *AgSp*, *AgSmh*, *AgSmu*, *AgSr*)

The Shaw Granitoid Complex occupies the core of the Shaw Dome in the southeastern corner of NORTH SHAW. The complex has intrusive contacts with flanking greenstones, except in the far southwest where it has a sheared intrusive relationship with adjacent mafic schists. An intrusive relationship is evident from the irregular northeastern and northwestern contacts of the complex, the presence of several thin granite sills in the North Shaw Belt, and the numerous amphibolite xenoliths (*Aba*) within the complex. The Shaw Granitoid Complex has been the focus of several previous investigations, including a mapping study by Bettenay et al. (1981) and geochemical and isotopic studies by Bickle et al. (1983, 1989, 1993). Zegers (1996) undertook structural studies along the margins of the complex and mapped the internal parts. A detailed mapping and geochemical study of the northern part of the Shaw Granitoid Complex is the subject of a current PhD thesis by Mark Pawley at the University of Newcastle.

The Shaw Granitoid Complex may be divided into several texturally and compositionally distinctive units that young inwards, down structural section, as deduced from crosscutting intrusive relationships and the results of U–Pb SHRIMP geochronology on zircon. The northern part of the complex is primarily composed of three subparallel units of meta-igneous rock, as originally identified by Bettenay et al. (1981) and collectively referred to as the North Shaw Suite (after Bickle et al., 1993). These rocks are deformed by a single foliation and shallowly plunging, mineral elongation lineation that strike and trend parallel to the greenstone contact. Igneous textures are widely preserved in the northern two units,

which contain amphibolite xenoliths and prised off roof pendants of adjacent greenstones, whereas the structurally lowest unit is diatexite.

The oldest component of the North Shaw Suite is the North Shaw Tonalite (*AgSns*; name from Bettenay et al., 1981), a mafic (colour index 15–30%) hornblende quartz diorite to trondhjemite. The best exposures of this unit (at AMG 560386, 546365, and north and south of AMG 530396) are intrusive dykes into amphibolite. The outcrop northeast of the latter point contains an unusual cumulate texture and resembles a gabbroic anorthosite. There is a large xenolith of North Shaw Tonalite within the younger Coolyia Tonalite (at AMG 553323). In thin section, hornblende is largely altered to an intergrown mixture of green-biotite–epidote–titanite–actinolite. The rock contains between 0 and 30% quartz, less than 5–10% microcline, and the remainder is highly altered plagioclase. Along the northwestern contact of the North Shaw Tonalite is a distinctive unit of porphyritic hornblende granite to granodiorite (*AgSp*) characterized by sparsely distributed plagioclase phenocrysts ( $\leq 1$  cm in length) in a fine-grained microgranite matrix. The distribution of this rock type suggests that it represents a fine-grained marginal phase to the North Shaw Tonalite.

The Coolyia Creek Tonalite (*AgSco*) is readily distinguished from adjacent units by its distinctive clotty hornblende texture defined by 15–20% of elongate and flattened (L–S) clots of mafic minerals, 1–3 cm in length. Bickle et al. (1993) described the local development of cumulate igneous layering within this body in its type area, in Coolyia Creek. The tonalite has intrusive contacts with the North Shaw Tonalite and marginal greenstones, and contains scattered inclusions of amphibolite derived from adjacent wallrocks along its western contact against greenstones where it cuts across bedding around Coolyia Creek (between AMG 465352 and 475368) and west of Shaw River (between AMG 404308 and 439340). The contact to the south of this area, and in the east, is subconcordant with bedding in the greenstones. In thin section, the mafic clots are composed of green-biotite–epidote–chlorite – opaque minerals (trace) and lie within a medium-grained, recrystallized groundmass of plagioclase and quartz, with less than 5% myrmekite and K-feldspar.

The origin of the clotty texture is enigmatic. The mafic clots have clearly been affected by the regional tectonic strain, which has deformed originally spherical clots into prolate spheroids. The origin of the spherical mafic mineral clots cannot be explained by assimilation of xenoliths alone, because their distribution and size are too even over most of the unit, and they have fuzzy or weakly gradational margins. Instead, this texture is possibly the result of mingling and mixing of comagmatic basaltic and granitic liquids (Pitcher, 1993, p. 133).

A sample of the Coolyia Creek Tonalite was collected for U–Pb SHRIMP dating of zircon from the same locality where Bickle et al. (1993) obtained a Pb–Pb isochron age of  $3338 \pm 52$  Ma. Results indicate a single population of zircons with an igneous age of  $3469 \pm 2$  Ma (Nelson, in prep.), indicating that the previous Pb–Pb age must reflect a period of metamorphic recrystallization.

In the northwestern part of the Coolyia Creek Tonalite and along the contact between it and the North Shaw Tonalite are two occurrences of a texturally distinct, medium- to coarse-grained hornblende monzogranite to granodiorite (*AgSmh*). Contact relationships with adjacent units are obscured by cover, and thus its relative age with respect to these units is unknown. This weakly strained granite has 30% quartz, scattered, centimetre-sized, euhedral plagioclase phenocrysts, and 10% hornblende.

Cutting the southern contact of the Coolyia Creek Tonalite is a sheeted complex of foliated, biotite leucogranite dykes. These dykes emanate from an associated body of schlieric, heterogeneous leucogranite diatexite further south (*AgSl*), and this rock type also forms the main migmatizing component of orthogneisses deeper within the granitoid complex (*AgSn*). Indeed, it is considered that this unit represents a ponded sill of leucogranite magma that was emplaced into the base of the overlying Coolyia Creek Tonalite during structural doming, and was derived from melting of the structurally lower orthogneisses. The texturally variable components of this unit share the characteristics of having biotite as the sole mafic phase (outside areas that are directly derived by melting of amphibolite), and weather to a distinctive chalky white colour.

Leucogranite diatexite contains three components, including relict enclaves of amphibolite, wispy-layered to schlieric leucogranite, and homogeneous leucogranite. These phases and their interrelationships are particularly well exposed in an outcrop on the east bank of the Shaw River (AMG 437237), which lies 50 m structurally above the contact with orthogneiss (*AgSn*). The bulk of the outcrop consists of schlieric to gneissic leucogranite diatexite, which varies from a gneiss with well-defined layering continuous over 2–5 m, to a texturally variable, medium- to coarse-grained rock that contains faint, wispy, discontinuous layering, abundant patches and veins of pegmatitic granite, and numerous large, round crystals of K-feldspar. Swarms of leucogranite melt veins intrude along a pre-existing foliation in amphibolite xenoliths, transforming these rocks into gneisses. Injection accompanied ductile folding of the enclaves, as seen by truncation surfaces along the enclave contacts against isotropic leucogranite. At one locality, a sharp-walled pegmatitic granite dyke appears to have formed from the collection of diffuse melt pressed out of a larger volume of rock along strike (Fig. 25a). Elsewhere, irregular patches of well-layered, remnant gneissosity lose definition both along and across strike, having been transformed in situ into isotropic, medium-grained schlieric leucogranite in which only faint, remnant traces of the original layering are preserved.

Cutting all of the above rocks are straight, continuous sheets of texturally and mineralogically differentiated leucogranite, approximately 40 cm wide and continuous across the 30 m width of the outcrop (Fig. 25b). Along the southern (structurally lower) margin of these sheets, about 10 cm of medium-grained isotropic monzogranite has a sharp, straight lower contact with diatexite, but a botryoidal upper contact against finer grained, grey leucogranite. The lower contact of the finer grained, grey leucogranite is marked by a 2–3 cm-wide zone of

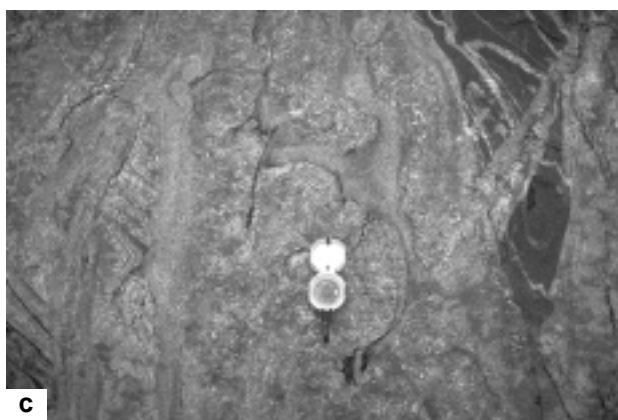
Table 3. Geochronology on NORTH SHAW

Sample number	AMG location	Lithology	Method	Age (Ma)	Reference
M95-327	374273	Keep It Dark Monzogranite ( <i>Agki</i> )	U–Pb SHRIMP zircon	2936 ± 5	M. J. Van Kranendonk, unpublished data
T94/87	483376	Amphibolite-facies Mount Ada metabasalt ( <i>AWmba</i> ) adjacent to Shaw Granitoid Complex	<sup>40</sup> Ar/ <sup>39</sup> Ar on hornblende	3522 ± 13 (P) <sup>(a)</sup> 3466 ± 13 (S) <sup>(b)</sup>	Zegers (1996)
49	501652	Galena in quartz vein ( <i>AWmb</i> ), Normay mine	Pb–Pb model age	c. 3400	Thorpe et al. (1992b)
100	473387	Galena in quartz vein ( <i>AWmb</i> ), Big Bertha mine	Pb–Pb model age	c. 2988	Thorpe et al. (1992b)
401	364708	Galena in quartz vein ( <i>AOcd</i> ), Lalla Rookh mine	Pb–Pb model age	c. 3188	Thorpe et al. (1992b)
415	526598	Galena in Dresser Formation barite ( <i>AWrc</i> )	Pb–Pb model age	c. 3490	Thorpe et al. (1992b)
70601	298684	Coucal Formation ( <i>AOcf</i> ), Coonterunah Group	U–Pb SHRIMP zircon	3515 ± 3	Buick et al. (1995)
81–631	448281	Coolyia Creek Tonalite ( <i>AgSco</i> ), Shaw Granitoid Complex	<sup>40</sup> Ar/ <sup>39</sup> Ar on hornblende	2907 ± 7 (S)	Wijbrans and McDougall (1987)
91084	471272	Leucogranite ( <i>AgSl</i> ), Shaw Granitoid Complex	U–Pb SHRIMP zircon	3493 ± 4	McNaughton et al. (1988)
91085	452280	Coolyia Creek Tonalite ( <i>AgSco</i> ), Shaw Granitoid Complex	U–Pb SHRIMP zircon	3467 ± 6	McNaughton et al. (1988)
94001	367422	Silicified felsic tuff ( <i>AWeft</i> ), Euro Basalt	U–Pb SHRIMP zircon	≤3463 ± 7 (D) <sup>(c)</sup> c. 3520–3510 (D) 3547 ± 6 (D)	Buick et al. (1995)
94058	080618	Quartz–feldspar porphyry ( <i>AgLp</i> ), Carlindi Granitoid Complex	U–Pb SHRIMP zircon	3468 ± 4	Buick et al. (1995)
100507	479466	Rhyolite ( <i>AWpr</i> ), Panorama Formation	U–Pb conventional zircon	3458 ± 2 c. 3724 (X) <sup>(d)</sup>	Thorpe et al. (1992a)
100510	467633	North Pole Monzogranite ( <i>Agno</i> )	U–Pb conventional zircon	3459 ± 18	Thorpe et al. (1992a)

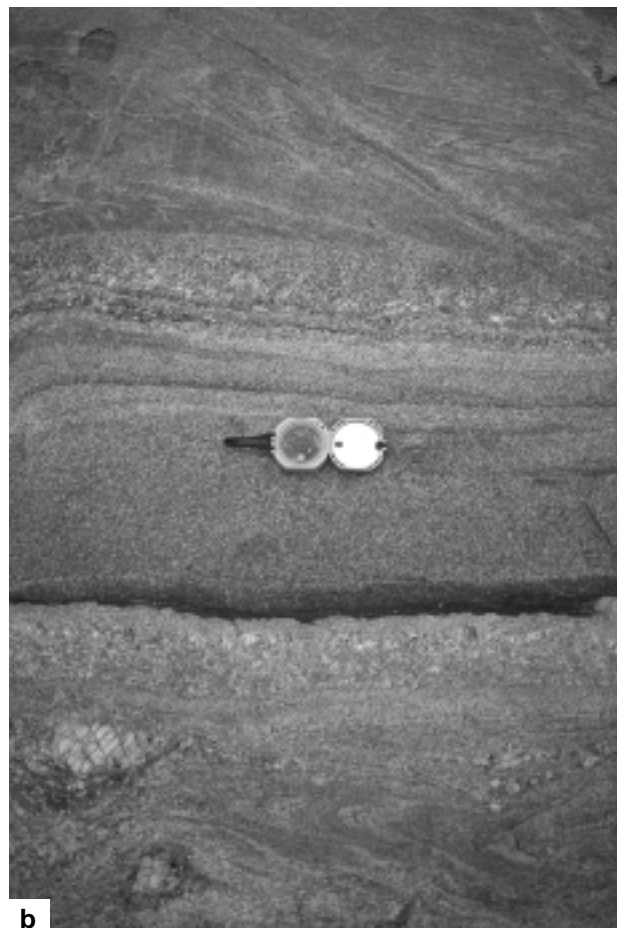
Table 3. (continued)

Sample number	AMG location	Lithology	Method	Age (Ma)	Reference
111868	149289	Dacite ( <i>AScf</i> ), Kangaroo Caves Formation	U–Pb conventional zircon	3240 ± 2 ≥3428 ± 2 (X)	Thorpe, R., 1996, written comm.
111868	149289	Galena from syngenetic Cardinal VMS gossan	Pb–Pb model age	c. 3260	Thorpe, R., 1996, written comm.
142881	430236	Leucocratic diatexite ( <i>AgSl</i> ), Shaw Granitoid Complex	U–Pb SHRIMP zircon	c. 3450 (P) c. 3400 (S)	Nelson, D., 1998, pers. comm.
142882	425223	Mulgandinnah Monzogranite ( <i>AgSmu</i> ), Shaw Granitoid Complex	U–Pb SHRIMP zircon	2928 ± 2	Nelson (1998)
142954	456715	Panorama Formation ( <i>AWpx</i> )	U–Pb SHRIMP zircon		Nelson (in prep.)
142962	442332	Coolyia Creek Tonalite ( <i>AgSco</i> ), Shaw Granitoid Complex	U–Pb SHRIMP zircon Pb–Pb whole rock	3469 ± 2 3338 ± 52	Nelson (in prep.) Bickle et al. (1993)
142964	388266	Rheomorphic ignimbrite ( <i>AWda</i> ), Duffer Formation	U–Pb SHRIMP zircon	3466 ± 3 (P) 3446 ± 8 (S)	Nelson (in prep.) Bickle et al. (1993)
153188	154716	Biotite granodiorite ( <i>AgLm</i> ), Carlindi Granitoid Complex	U–Pb SHRIMP zircon	3484 ± 4	Nelson (1999)
153190	121715	Leucogranite ( <i>AgLI</i> ), Carlindi Granitoid Complex	U–Pb SHRIMP zircon	3469 ± 2	Nelson (1999)
155910	165743	Wilson Well Gneiss ( <i>AgLwi</i> ), Carlindi Granitoid Complex	U–Pb SHRIMP zircon	Poor quality data	Nelson, D., 1998, pers. comm.
203366	183456	Strelley Granite, outer phase ( <i>Agste</i> )	U–Pb SHRIMP zircon	3239 ± 2	Brauhart (1999)
203368	276527	Strelley Granite, inner phase ( <i>Agstp</i> )	U–Pb SHRIMP zircon	3238 ± 2	Brauhart (1999)

NOTES: (a) P: primary igneous age  
 (b) S: secondary plateau, metamorphic age  
 (c) D: detrital zircon age  
 (d) X: xenocrystic zircon



MVK140



amalgamation of filter-pressed melt out of a large volume of rock; b) close-up view of zoning across a discordant granitoid dyke cutting migmatitic gneiss and leucogranite diatexite. Way up is to the top of the photograph, which is north; c) texturally and compositionally zoned leucogranite (at and around compass) cutting migmatitic gneissic protolith. As with b), the dykes display highly irregular zoning

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concentrated biotite, which grades up to more leucocratic schlieric granite in the core of the dyke. At the top of the sheet, the finer grained phase grades up into medium-grained granite, which resembles the lower phase of the sheet. In one sheet, up to 35 cm of topography is developed along the botryoidal contact between the finer grained and coarser grained phases (Fig. 25c). These sheets are themselves cut by later generations of

pegmatitic granite sheets derived by remelting of the diatexite. Therefore, the sheets are interpreted to form an integral part of the diatexite and not some unrelated, later phase.

The asymmetrical features of the texturally differentiated leucogranite sheets — specifically the convex-up nature of the botryoidal contact, the concentration of mafic

minerals at the base of the finer grained phase, and the gradation from fine- to coarse-grained granite at the top of these sheets — are used to infer that they were originally emplaced as sills, whose tops now face to the north. Their present dip of 70°N, combined with their age of emplacement (see below), indicates that these and adjacent rocks throughout the northern part of the Shaw Dome have undergone 70° of tilting to the north since c. 3410–3400 Ma.

A sample from a 2 m-wide, foliated biotite leucogranite dyke cutting migmatitic orthogneiss near the base of this unit just east of NORTH SHAW gave a concordant age of  $3445 \pm 3$  Ma (Nelson, 1998; GSWA 142878) for a single population of zircons. A sample of diatexite from the Shaw River exposure described above, contained zircons with clear, core-overgrowth relationships. Whereas cores gave c. 3450 Ma ages, rims gave concordant ages of between 3410 and 3400 Ma, and this latter age is interpreted as the time of melting, formation of the diatexite, and intrusion of the texturally differentiated sills. Combined, the two samples indicate that melting of tonalitic precursors in the core of the Shaw Granitoid Complex and generation of leucogranite magmas occurred between c. 3445 and 3400 Ma.

To the south of the leucogranite diatexite, across what varies from a sheeted intrusive contact to a 10 m-wide zone of porphyroclastic mylonite, is migmatitic tonalite orthogneiss (*AgSn*). The precursor to the gneiss is well preserved in many places. It consists of a sheeted sill complex of blue-grey tonalites that have slight variations in mafic mineral content and texture; some sheets are equigranular and leucocratic, others are weakly feldspar porphyritic, and still others are more mafic with less quartz (e.g. quartz diorites). The scale of sheeting varies from 30 cm to several metres, although large areas may be entirely homogeneous. The sheets are parallel to the margins of the granitoid complex, but they are penetratively foliated (and migmatized), such that the original orientation of their intrusive contacts may have varied.

The gneisses contain numerous xenoliths of layered amphibolite gneiss, and less commonly of anorthosite (e.g. at AMG 438235) and metaquartzite. Amphibolite rafts reach maximum dimensions of 3 km long by 100 m wide, and these help outline regional folds. Some amphibolites contain relict orthopyroxene and clinopyroxene, indicating that granulite-facies temperatures were reached prior to extensive amphibolite-facies retrogression, but pressure conditions during metamorphism were low as no evidence of relict garnet has been observed.

The gneisses include all types of migmatites through to diatexite, but these variations have not been differentiated at the scale of the map. Stromatic migmatites are the most common type with between 10 and 50% veins of leucogranite melt. Leucogranite melt veins were intruded from deeper levels (i.e. sharp-walled, injection veins) and also developed as in situ melt products (e.g. veins with biotitic palaeosome margins), with some of the two types occurring in the same outcrop. At deeper structural levels in the southeast are true diatexites with blocks of migmatitic tonalite and amphibolite ‘floating’

in ‘a sea’ of leucogranite, which may be seen to both derive from, and explode, amphibolite xenoliths.

A 500 m-long unit of strongly foliated, granitoid rock in the southwest part of the Shaw Granitoid Complex is underlain by equigranular, medium-grained leucocratic granodiorite (*AgSg*) and subordinate gneiss. This unit forms the northern extension of a more extensive body of this rock to the south on TAMBOURAH, which includes the c. 3430 Ma South Daltons Pluton (McNaughton et al., 1993). In this southern area, rocks of this type and age were observed to cut gneissic tonalites in several localities, indicating that gneissosity formed early in some places, consistent with the 3445 Ma age of some leucogranite dykes (*AgSl*) described above, which were interpreted to be derived from melting of tonalites.

Bickle et al. (1983, 1993) studied the geochemistry and Pb–Pb and Sm–Nd isotopic systematics of rocks from the northern part of the Shaw Granitoid Complex (*AgSns*, *AgSco*, *AgSl*, and *AgSn*). These calc-alkaline TTG rocks are characterized by highly fractionated REE patterns (LREE enriched, HREE depleted), high Ba and Sr, low Ti, and little or no negative Eu anomaly. Fractionation–cumulation processes were only locally recognized, but could not account for the variations in the North Shaw Suite. Instead, variations in composition were most likely the result of variable degrees of partial melting at depth.  $\epsilon_{\text{Nd}}(3450 \text{ Ma})$  values range from +1.9 to –0.3, indicating contamination by slightly older crust (Bickle et al., 1993). These authors concluded that rocks of the North Shaw Suite were derived from a depleted mantle source and contaminated by a component of slightly older crustal material.

A string of small plutons and related dykes of the  $2928 \pm 2$  Ma (Nelson, 1998; GSWA 142882) Mulgandinnah Monzogranite (*AgSmu*) intrude gneisses and leucogranite of the Shaw Granitoid Complex along the southern margin of NORTH SHAW. The rocks are blue-grey and weather to a dark chocolate-brown colour developed as a thin rind on the outer skin of rounded boulders, which form distinct kopjes that rise above the surrounding flat granitoid terrain. The rock is a fine- to medium-grained monzogranite with up to 8% biotite. Rocks vary from undeformed to penetratively foliated, the latter fabric defined primarily by elongate quartz in outcrop and the formation of subgrains of quartz and a fracture cleavage through feldspars in thin section. Plagioclase forms 40% of the rock, is oligoclase in composition, and typically has heavily saussuritized cores. Microcline forms 25% of the rock and there is also local myrmekite. Accessory minerals include titanite, apatite, and allanite, the latter mineral forming conspicuous dark spots in the rock mantled by a white, metamict halo. Zircon is abundant as very fine grains, and there is secondary growth of epidote, sericite, carbonate, and chlorite after biotite.

The youngest felsic unit in the Shaw Granitoid Complex is a swarm of ultra fine grained, undeformed rhyolitic dykes (*AgSr*) located in the northeast (AMG 571320). These dykes are oriented parallel to the contact of the complex and some are formed in the adjacent greenstones of the Coongan Belt. The pale-brown-weathering, blue-black dykes are highly siliceous and

form narrow ridges a few metres higher than the surrounding granitoid pavements. Well-developed flow banding was observed at the locality indicated above, consisting of centimetre-wide bands of darker and paler coloured rocks folded into a rootless isoclinal fold, 5 cm wide and facing to the northwest. The age of these felsic dykes is unknown, but must be less than c. 2772 Ma, as they are cut by the Black Range Dyke of the Fortescue Group.

### Yule Granitoid Complex (*Agype*, *Agyka*, *AgYwox*, *AgYwo*, *AgY*)

The Yule Granitoid Complex on NORTH SHAW is composed of three progressively younger rock types (*Agype*, *Agyka*, *AgYwo*). An area of undivided granitoid rocks (*AgY*) along the western margin of the map sheet was not visited. The oldest recognized unit is migmatitic tonalitic orthogneiss with sheeted, pegmatitic granite veins, herein named the Petroglyph Gneiss (*Agype*), after the fact that they were the favoured rock type for ancient Aboriginal art. The blue-grey rocks outcrop as low pavements with a thin, red-weathering rind and are characterized by a well-developed gneissic layering defined by sheeted veins of pegmatitic monzogranite. As with equivalent gneissic units in the Shaw and Carlindi Granitoid Complexes, the precursor to the gneiss was a sheeted complex of blue-grey tonalites with slight variations in texture (equigranular to weakly porphyritic) and composition (trondhjemite, tonalite, quartz diorite). However, at variance with the Shaw Granitoid Complex, the migmatite veins are largely of injection type (vs leucosome melts), pink (vs white), and monzogranites (vs trondhjemites). In thin section, the tonalites have a granoblastic texture of quartz and plagioclase, with aligned flakes of biotite that define a foliation parallel to the gneissic layering in the rocks.

Leucocratic biotite granodiorite to granite, called the Kavir Granodiorite (*Agyka*), intrudes gneisses that are locally strongly deformed and contain isoclinal folds of granitoid veins. This rock has an equigranular to weakly porphyritic texture and varies from massive to well-foliated rocks with a penetrative mineral elongation lineation. Van Kranendonk (1997) showed that the Kavir Granodiorite was syntectonic with D<sub>2</sub>, which formed a kilometre-scale dome-and-basin pattern on the north-eastern margin of the Yule Granitoid Complex. Two samples of this unit from just west of NORTH SHAW, on WODGINA, returned an age of c. 3240 Ma (M. J. Van Kranendonk, unpublished U–Pb SHRIMP zircon data). Xenocrystic zircons from one of the samples — a metre-wide dyke cutting the Petroglyph Gneiss — gave an indication of the age of the Petroglyph Gneiss at c. 3470 Ma (M. J. Van Kranendonk, unpublished U–Pb SHRIMP zircon data).

The Woodstock Monzogranite, a foliated to mylonitized, K-feldspar porphyritic to megacrystic biotite monzogranite (*AgYwo*), intrudes both of the older rock types described above. This dark-grey unit is well exposed and weathers to a deep orange-red colour. The rocks contain 0.5 – 2.5 cm K-feldspar phenocrysts with zoning outlined by poikilitic inclusions of biotite, plagioclase, and quartz. Phenocrysts comprise 10–20% of the rock and lie within an equigranular, medium-grained matrix (finer

grained where more strongly sheared) of quartz (30%), plagioclase (35%), and K-feldspar (including myrmekite). Quartz is recrystallized to fine-grained and strongly undulose subgrains, and the margins of K-feldspar phenocrysts are typically recrystallized, indicating ductile conditions during shearing. In the southwest, the Woodstock Monzogranite has a sheeted, irregular intrusive contact with the older rocks, forming a heterolithic igneous complex (*AgYwox*). In these areas, sheets of the Woodstock Monzogranite commonly have a faint schlieric texture inherited from the older gneisses, and this texture becomes much more pervasive over wide areas of the Woodstock Monzogranite further south on TAMBOURAH. This rock is affected by a penetrative shear foliation within the Pulcunnah Shear Zone along the eastern margin of the granitoid complex, characterized by sinistral asymmetry and shallow, northerly plunging elongation lineations defined by stretched quartz.

## Unassigned rocks

### Ultramafic rocks (*Auc*, *Aubs*, *Aup*, *Auph*, *Aur*, *Aux*)

A variety of ultramafic rocks throughout NORTH SHAW cannot be confidently correlated with either a particular intrusive suite or formation. Such unassigned ultramafic rocks are in the Numerous Scrapes and Mount York Deformation Zones, the Emerald Mine Structural Complex west of the Shaw Granitoid Complex, along the eastern contact of the Yule Granitoid Complex, and as dykes cutting the Carlindi Granitoid Complex.

The Numerous Scrapes Deformation Zone is composed of strongly sheared ultramafic rocks that envelop rafts, up to 2 km long, of disaggregated supracrustal rocks and less strongly deformed lumps of serpentinized peridotite (*Aup*). The ultramafic matrix to this megabreccia zone is a strongly foliated and commonly lineated talc–carbonate and talc–chlorite–carbonate schist (*Auc*), with abundant oxidized opaque minerals. The rocks are medium-grained metamorphic rocks with mineral elongation lineations defined by needles of talc and elongate blades of chlorite. Typically the rocks contain centimetre-scale lenses of relatively more competent, serpentinized and carbonate-altered peridotite, surrounded by an anastomosing network of talc–chlorite schist. Within the deformation zone are numerous dark-brown lenses, pods, dykes, and locally, whole hillsides composed of dark-brown weathering carbonate (ferroan dolomite or ankerite), indicating large-scale mobilization of carbonate during shear deformation. The protolith of these rocks is uncertain, but may include members of the Dalton Suite or high-Mg to komatiitic volcanic rocks of the Six Mile Creek or Kunagunarrina Formation of the Sulphur Springs Group. Much of the same type of rocks underlie the Mount York Deformation Zone, although not as heavily altered by carbonate. The Mount York Deformation Zone is a tectonic megabreccia of rafts of weakly deformed, serpentinized peridotite (*Aup*) and supracrustal rocks of the Kunagunarrina Formation in strongly deformed serpentine(–chlorite–talc) schists. Along the northeastern margin of the Bernts Deformation Zone is a smaller scale

ultramafic tectonic breccia, hosted by talc–carbonate–chlorite schist and containing decimetre- to metre-size blocks of mafic and ultramafic schist.

In the Emerald Mine Structural Complex, ultramafic rocks are exposed in two areas. In the eastern area, weakly deformed, fine-grained, massive peridotite (*Aup*) contains widespread relicts of cumulus olivine texture, although the olivine is altered to serpentine, chlorite, and opaque minerals. This dark-blue to greenish-black rock weathers to tan brown and forms low outcrops. A weak foliation is locally developed in this rock, primarily as a fracture cleavage, but where more strongly deformed in the north, it is altered to talc–carbonate–chlorite schist (*Auc*). Along the southern margin of the well-preserved rocks, ultramafic schists are interbedded with chloritic mafic schists at a scale of decametres (*Aubs*). In the western part of the Emerald Mine Structural Complex, interbedded mafic and ultramafic schists (*Aubs*) underlie a large area and probably represent the sheared remnants of interbedded tholeiitic and high-Mg basalts. Such intimately interbedded rocks also outcrop in the southeastern part of the Yule Granitoid Complex, where a supracrustal origin is indicated by thin interbeds of sheared metaquartzite or chert or both.

Tremolite–chlorite schists (*Aur*) outcrop in the Emerald Mine Structural Complex (AMG 325242) and as dykes within, and cutting across the northwestern margin of, the Shaw Granitoid Complex (AMG 560312 and 536242). These rocks are fine- to medium-grained, thoroughly recrystallized metamorphic rocks and typically strongly foliated to schistose. The age of dyke formation in the Shaw Granitoid Complex is unknown, bracketed only by the age of the host rocks (*AgSns*;  $\geq 3470$  Ma) and the unconformably overlying c. 2.76 Ga Fortescue Group.

Sheets and sills of metaperidotite and metapyroxenite (*Aup*, *Aux*) along the eastern contact of the Yule Granitoid Complex are interlayered with dolerites and gabbros (*Ao*) and host rocks of the Sulphur Springs Group, particularly the Leilira Formation. These weakly deformed, but tightly folded rocks contain some relict igneous textures, but are commonly thoroughly recrystallized to chlorite–serpentine(–talc–carbonate) schists. These rocks have not been assigned to a formation or an intrusive suite because of their state of deformation and recrystallization, but it is considered most likely that they belong to the Dalton Suite.

Serpentinized metaperidotites and serpentine–chlorite schists underlie much of the area affected by tight folds along the southern margin of NORTH SHAW, adjacent to the Yule Granitoid Complex. Although no volcanic textures are preserved in these schists, the fact that they are present at the level of the mafic–ultramafic Kunagunarrina Formation suggests they may have originally been flows of this formation.

Cutting the Carlindi Granitoid Complex is the Shilliman Dyke, a metamorphosed and faulted, medium- to coarse-grained harzburgite (*Auph*) with a curved intrusion pattern. The main part of the dyke contains serpentinized olivine and fresh to bastite-altered intercumulus orthopyroxene with a texture similar to that

in Figure 24a. The western contact of the north–south segment of the main dyke is differentiated into a coarse-grained gabbro with a well-preserved igneous texture of plagioclase laths and ?igneous hornblende phenocrysts (or metamorphic hornblende pseudomorphs after igneous clinopyroxene). Plagioclase laths are surrounded by an extensive rind of myrmekite and there is free quartz in the rock, suggesting partial contamination of the magma by the assimilation of granite. Two thin dykes of similar composition, texture, and altered mineralogy strike northeast–southwest just south of the main dyke.

The dykes are typically recrystallized to a metamorphic assemblage of chlorite–sericite and fine titanite. Fine zircon of probable metamorphic origin is present within chlorite, together with metamorphic titanite, and some larger areas of plucked out material with radiation-damage colour haloes in amphibole suggest the presence of igneous zircon, monazite, or baddeleyite. One probable baddeleyite grain was observed. The age of these dykes has not been established other than that it must be younger than the host leucogranite (*AgLl*; c. 3470 Ma). The differentiated nature and harzburgite composition suggests that the Shilliman Dyke may be part of the Dalton Suite.

### Mafic rocks (*Ao*, *Aolx*, *Ab*, *Aba*, *Abm*, *Abz*)

Strongly deformed and some structurally isolated mafic rocks, which cannot be placed in the stratigraphic context of the map with certainty, are present in several places across NORTH SHAW. The most common of these are greenschist-facies, massive to pillowed metabasalts (*Ab*), which are commonly deformed and metamorphosed into chloritic schists. Pillowed mafic volcanic rocks in the Emerald Mine Structural Complex are interbedded with horizons of blue-black massive and blue-black and white layered cherts, forming a succession similar to the Euro Basalt. High-Mg metabasalts are also present (*Abm*), discriminated from the former by a paler weathering colour, small-scale pyroxene spinifex texture, and ocelli. Amphibolite-facies equivalents (*Aba*), which are commonly massive or less commonly pillowed, formed slices and large rafts within the northern part of the Shaw Granitoid Complex (AMG 406295 and 515397 respectively). Whereas the former contains thin interbeds of metaquartzite, the latter is a massive, fine- to medium-grained, equigranular amphibolite with no textural features indicative of its origin. Several smaller xenoliths of amphibolite are present throughout the North Shaw Suite, with those near the granite–greenstone contact almost certainly representing inclusions of metabasalt.

Extensive alteration of basaltic rocks has occurred immediately beneath the Strelley Pool Chert in two places in the East Strelley Belt (AMG 124619 and 315666) and in the western part of the Panorama Belt (at, and along strike to the south of, AMG 388552). These rocks (*Abz*) weather to a golden-orange colour and are massive to schistose, but can be seen to grade into basaltic rocks across strike, down stratigraphic section, and locally into pillowed basalts. In thin section, relict doleritic to basaltic

textures are preserved, with coarse grains of ilmenite, square outlines of igneous pyroxene, and laths of plagioclase feldspar, but these original minerals have been altered to leucoxene, chlorite, and sericite, with sparse, coarse-grained quartz. In thin section, relict laths of twinned plagioclase were completely recrystallized to fine-grained myrmekite. This texture is interpreted to reflect the alteration of albite to anorthite, with a consequent liberation of excess silica that crystallized as 'worms' of free quartz.

Gabbroic and doleritic rocks (*Ad*) outcrop along the southeastern contact between the Carlindi Granitoid Complex and East Strelley Belt, and as a tightly folded dyke cutting the Coonterunah Group in this same area (AMG 360701). More strongly deformed rocks are completely recrystallized to an equigranular, fine-grained assemblage of plagioclase–chlorite–carbonate – opaque minerals. Dolerite of unknown age also outcrops in the Coonterunah Group in the western part of the East Strelley Belt and in the southernmost extension of the Strelley Granite (*Agsta*). In these areas, the rocks are fine- to medium-grained, recrystallized dolerites, but contain no distinctive textures to associate them with other intrusions.

A unit of coarse-grained leucogabbro (*Alx*) is exposed near the base of the Euro Basalt in the East Strelley Belt. This light-brown weathering rock has a conspicuous gabbroic texture defined by centimetre-scale, brown-weathering, prismatic crystals of pyroxene ( $\leq 25\%$  by volume) in a white matrix of plagioclase. The rock most closely resembles leucogabbro of the Dalton Suite south of the Soanesville Syncline (AMG 257214), but the age of this unit is unconstrained. In thin section, the rocks are coarse-grained intergrowths of plagioclase laths in subophitic augite, with 5% free quartz and 3–5% ilmenite. Variable alteration has resulted in the growth of leucoxene and chlorite–epidote–titanite(–serpentine–carbonate).

## Other rocks (*As*, *Ac*, *Af*, *Ad*, *Aq*, *go*)

A variety of rock types form disrupted breccia fragments within ultramafic schists of the Numerous Scrapes Deformation Zone. The topographically highest of these include kilometre-scale panels of thick-bedded, coarsely recrystallized metaquartzite and less common interbeds of metasiltstone (*As*). These rocks resemble quartzites from the Golden Cockatoo Formation. Metasedimentary rocks of uncertain lithostratigraphic affiliation are also in the Bernts Deformation Zone (AMG 368545), but these are distinctly more lithic in character and variably bedded, and composed of metres-thick beds of metasandstone in poorly bedded wacke and litharenite. These tightly folded rocks most closely resemble the Leilira Formation of the Sulphur Springs Group.

Unassigned cherts (*Ac*) of a variety of thickness and composition outcrop in the Emerald Mine and Tambina Structural Complexes, and in the wide deformation zones. The most common type is massive blue-black chert, but centimetre-layered, white and blue-black chert, aphanitic grey, and green cherts are also present. Blue-grey chert breccias (*Ac* with breccia overprint) form prominent ridges

along the southern margin of the Mount York Deformation Zone and in the Miralga Deformation Zone. These rocks consist of packed, angular fragments of silicified, fine-grained rock in an aphanitic, grey chert matrix. They lack any evidence of sorting or bedding, and because they outcrop only within deformation zones, are interpreted to represent silicified fault breccias.

Several thin, dismembered slices of massive felsic volcanic rock (*Af*) outcrop in the Numerous Scrapes Deformation Zone. Whereas these rocks almost certainly belong to the Sulphur Springs Group, a lack of supporting geochemical data means that they cannot be confidently ascribed to a specific formation. These rocks are fine grained, being silicified and composed of a finely intergrown, recrystallized, and locally polygonal mosaic of quartz and plagioclase feldspar.

A northeasterly striking swarm of fine- to medium-grained dolerite dykes (*Ad*) of uncertain age and affiliation cut the Carlindi Granitoid Complex in the northwestern part of NORTH SHAW. These massive, but metamorphically recrystallized, dykes lie within faults that offset the Shilliman Dyke, and are cut by massive, white quartz veins (*Aq*). In thin section, the dolerite dykes were observed to consist of chlorite-altered laths of clinopyroxene within interstitial plagioclase, which has been altered to sericite. Small grains of free quartz are in the dykes along with hematite-altered flakes or needles of original iron oxides. Little-deformed dykes cutting the Coonterunah Group (e.g. at AMG 337706) contain fresh clinopyroxene phenocrysts and plagioclase in a ground-mass of actinolite–epidote–rutile. Clinopyroxene is locally rimmed by relicts of brown–green pleochroic hornblende, and these by a secondary rim of actinolite, indicating a two-stage metamorphic overprint in this area.

A 1–2 m-wide dolerite dyke cuts west-southwest–east–northeast across the Pincunah Belt, the Lalla Rookh – Western Shaw Fault, and into the western margin of the Strelley Granite, indicating that it is post- $D_3$  in age (c. 2940 Ma). This rarely exposed dyke can be traced as a prominent lineament on airphotos. Where exposed, the dyke is a fine-grained dolerite with a smooth-weathering, grey-black appearance and is composed of a little-altered, igneous mineralogy of plagioclase laths in intercumulus clinopyroxene and ilmenite (2–3%). This post-tectonic dyke has a different appearance and orientation to those of the Black Range Doleritic Suite associated with the Fortescue Group, and is of unknown age.

Massive, white quartz veins (*Aq*) are present within faults cutting the Strelley Granite and as pods, veins, and dykes within the Numerous Scrapes Deformation Zone. Prominent white quartz veins are also present in faults in the Carlindi Granitoid Complex, the Pulcunah Shear Zone along the margin of the Yule Granitoid Complex, and the southwestern part of the Lalla Rookh Synclinorium. In these areas, quartz veining is intimately associated with  $D_3$  structures and thus pre-dates the Fortescue Group. One area where quartz veins may not be associated with  $D_3$  deformation is in the southeastern Strelley Granite, where they strike east-southeast, parallel to growth faults in the overlying volcanic stratigraphy.

Map-scale surface gossan (*go*) is exposed at the Bernts prospect (AMG 362506). The gossan consists of a dark red-brown weathering, massive rock with a deeply pitted, irregular surface. Broken samples are porous and white in colour, commonly consisting of a fine, reticulated network of silica veinlets, interstitial porosity, and dark-purple, centimetre-wide veins of hematite(–magnetite). The porosity is interpreted to result from the weathering out of sulfide minerals. Other gossans, the one at Sulphur Springs for example, are composed of similar rocks, but are too small to show on the map.

## Fortescue Group

Rocks of the Fortescue Group outcrop in three large synclinal outliers — the Olympic Pool, Marble Bar, and Antarctic Outliers — and as a discontinuous, thin veneer of basalt over the North Pole Dome. Basal rocks of the Fortescue Group lie on folded rocks of the Pilbara Supergroup across an angular unconformity. Intergroup unconformities are developed between the Mount Roe Basalt and Hardey Formation, and in one location between the Hardey Formation and Kylena Basalt. These unconformities indicate tectonic instability during deposition of the group.

## Mount Roe Basalt (*AFr*, *AFra*)

The Mount Roe Basalt (*AFr*) is commonly a thick-bedded, massive to weakly vesicular basaltic formation that lies unconformably over the full spectrum of older rocks (including the northeastern Shaw Granitoid Complex, the lower Warrawoona Group in the North Pole Dome, rocks of the Sulphur Springs and Gorge Creek Groups along the western margin of the Olympic Pool Outlier, and the De Grey Group of the Keep It Dark Basin). The best development of this unconformity is along the eastern side of the Olympic Pool Outlier (AMG 311290), where shallowly north dipping rocks of the Mount Roe Basalt lie unconformably on closely folded, steeply dipping, northerly striking sandstones and conglomerates of the De Grey Group. At the base of the formation in this area is a metre-thick unit of pebble to boulder conglomerate containing subrounded clasts of predominantly layered black and white chert derived from nearby outcrops of the Gorge Creek Group. The conglomerate is typically white and highly silicified, and locally cut by small, white quartz veins, in contrast to the fresh overlying basalt. In places, it passes along strike into a thin unit of white, silicified metaquartzite. Immediately above the conglomerate is about 1 m of red-weathering, ferruginized, rubbly basalt that probably represents a basal eutaxitic, brecciated contact of the lowest basalt flow. Silicification of the basal conglomerate was probably caused by heat from eruption of the overlying Mount Roe Basalt.

Basaltic rocks include massive, vesicular, feldspar-phyric, and coarsely amygdaloidal flows, the latter being particularly well developed in a small outlier in the North Pole Dome (AMG 555753) where pale-green basalt has coarse gas cavities up to 5 cm wide that are filled, or partly filled, by agate and clear quartz. Basalts in the Marble Bar Outlier locally have vesicular flow tops with clusters of

pipe vesicles from which it is possible to measure bedding. The Mount Roe Basalt commonly contains plagioclase phenocrysts, which in one section vary upwards from sparse, equant, euhedral phenocrysts, through 1–2 cm-large glomerophenocrysts, to larger aggregates of euhedral crystals. Locally, feldspars were observed as thin, feathery laths. The volcanic rocks are commonly altered to chlorite–epidote assemblages with coarse pyroxenes and titanite-ilmenites, and less common plagioclase and quartz phenocrysts.

Basaltic agglomerate (*AFra*) is widely developed at the base of the Mount Roe Basalt, possibly infilling small valleys. These rocks are characterized by irregular, commonly subrounded fragments of basalt in a fine-grained matrix with many numerous basaltic glass shards and quartz–chlorite filled vesicles. In thin section, the basaltic fragments are chlorite altered and contain abundant perlitic cooling cracks surrounded by quartz–chlorite–carbonate matrix with local to common augite phenocrysts and some plagioclase phenocrysts.

## Hardey Formation (*AFhh*, *AFhs*)

The Hardey Formation consists of clastic sedimentary rocks including, from base to top, boulder to pebble conglomerate and local interbedded sandstone, a distinctive unit of fissile green-brown shale (*AFhh*), and overlying quartz arenite (*AFhs*). The distribution of these rock types varies throughout the area. Conglomerate is developed only above the Mount Roe Basalt (AMG 573415), where it contains clasts of white quartz, laminated black and white chert, and sandstones of the Gorge Creek Group. Sandstone (*AFhs*) is the most common rock type in this formation, consisting of well-sorted detritus and displaying local cross-bedding. In a number of places, but most obviously northeast of the Virgin Creek prospect (AMG 592447), the Hardey Formation was observed to unconformably overlie folded rocks of the Mount Roe Basalt, but they also rest unconformably on the Pilbara Supergroup (AMG 283245 and 585355). Hardey Formation sandstones are conformably overlain by the Kylena Basalt, and in one outcrop (AMG 583419), were seen to intermingle with them, indicating local contemporaneity of deposition.

## Kylena Formation (*AFkb*, *AFkba*, *AFks*, *AFkbt*)

The Kylena Formation is dominated by a thick, lower unit of vesicular to massive basalt (*AFkb*), which is especially well developed in the Marble Bar and Olympic Pool Outliers. Local columnar jointing may be observed, but bedding is commonly difficult to measure. Less common are amygdaloidal basalt flows with flow-top breccias. In the Antarctic Outlier, the preserved part of the Kylena Formation is 580 m thick. From base to top the formation includes: 90 m of doleritic basalt; a thin sandstone with conglomeratic lag layers and cross-bedding; about 117 m of massive basalt, mafic tuff, and feldspar-phyric basalt; 30 m of sandstone, lapilli tuff, and 10 cm-scale bedded mafic tuff that has erosional bases, rare cross-beds, and

vesicular tops; and an upper unit of mafic tuff, basaltic agglomerate (*AFkba*), and vesicular and amygdaloidal flows.

In the Olympic Pool Outlier (see Fig. 6), about 790 m of the lower basaltic unit (*AFkb*) is conformably overlain by up to 27 m of cross-bedded sandstone to pebbly sandstone and conglomerate, and thinly bedded, felsic to intermediate, sandy to fine-grained tuff with local accretionary lapilli (*AFks*; Fig. 26). A basal unit of blue-grey pebble conglomerate, 0.5 m thick, contains centimetre-size oncolitic stromatolites with small branching arms (Fig. 27a), which have tentatively been identified as *Alcheringa narrina* Walter 1972, similar to those described from Mount Herbert in the northeast Pilbara by Walter (1972; Grey, K., GSWA, 1998, pers. comm.). Possible microfossils have been observed in thin section. This rock also contains common micritic carbonate pelloids and less common clasts of plagioclase-phyric basalt, and pyroxene spinifex-textured komatiitic basalt. Some of the exotic clasts are rimmed by a thin, locally multilayer rind of algal material.

Within the sandy tuff layers, cross-bedding and trough cross-bedding are common, indicating a high-energy environment. Palaeocurrent indicators (axis of trough cross-beds, asymmetric ripples) from the southern limb of the fold indicate water flow to the northeast. Both fining-upward textures within regular graded beds and local inverse grading were recognized, indicating the presence of floating pumice fragments. Supporting this interpretation is the presence, even in trough cross-bedded, sandy tuff beds, of coarse recrystallization textures in the matrix, indicative of it having been hot during sedimentation (Fig. 27b). Some fine, ash beds contain shards of clear felsic glass with embayed, scalloped edges. In accretionary lapilli beds, the cores of the lapilli are commonly fragments of clear volcanic quartz crystals and, less commonly, of feldspar and biotite (Fig. 27c).

Conformably overlying the epiclastic succession are up to 60 m of well-bedded, pale grey-green mafic tuffs (*AFkbt*). These smooth weathering, homogeneous rocks are very fine grained and display faint to well-developed bedding at a 10 cm scale. Beds commonly contain slightly paler patches of fluidized intraformational rip-up clasts that have been transected by a network of veins of the host matrix (Figs 28a, b). In thin section, there is no visible difference between the clasts and matrix in these mafic to intermediate tuffs, other than a subtle change in grain size. Some beds are more massive and contain crosscutting crystals or patchy, devitrification textures indicating that, although water-lain, some beds were still hot when deposited.

## Dolerite dykes (*AFd*)

Weakly (deuterically) altered doleritic to gabbroic rocks of the Fortescue Group (*AFd*) within straight-walled dykes primarily cut the Shaw Granitoid Complex and North Pole Monzogranite, but also cut greenstones of the Panorama Belt. These dykes strike north-northeast and are up to 25 m wide. A prominent dyke in the southeastern part of

NORTH SHAW is similar in composition, weathering pattern, texture, and strike to dykes of the older (c. 2772 Ma) Black Range Doleritic Suite, but unlike these older dykes, cuts both the Mount Roe Basalt (AMG 563394 and 580440) and the lower part of the Kylenea Basalt (AMG 576410) in the Marble Bar Outlier, thereby indicating that it must belong to a younger set.

## Cainozoic geology

Ferruginous duricrust (*Czrf*) includes massive, pisolitic, and nodular laterite, and locally, consolidated ferruginous alluvium. These rocks form a dissected laterite plateau atop many hills on NORTH SHAW, particularly on ferruginous shale and banded iron-formation of the Gorge Creek Group (Pincunah Hill and Pyramid Hill Formations) in the East Strelley, Pincunah, and southern part of the Soanesville Belts. In these areas, the laterite is rock destructive and forms pisolitic to nodular hematite, maghemite, and goethite–limonite crusts up to several metres thick. Local alteration below laterite may reach several tens of metres depth (e.g. at AMG 134632). On the highest hills in the south-central part of NORTH SHAW, laterite forms as a fringe around the margins of rolling hilltops, below the maximum topographic height. In these areas, the laterite would appear to have been formed by the actions of groundwater. Laterite is also present locally on flat hills in small valleys below other laterite-capped hills, suggesting the presence of a relict topography during laterite formation.

Ferruginous silt, sand, and gravel (*Czcf*), which are variably consolidated and dissected, formed along the flanks of laterite caps on the Yule Granitoid Complex in the southwestern part of NORTH SHAW. Locally, these rocks consist of cemented goethite–limonite pisolites and nodules weathered out of laterite (*Czrf*). On the Carlindi Granitoid Complex, a ferruginous silt and sand blanket covers large areas of flat ground.

Massive, nodular, and cavernous limestone (*Czrk*) cover large areas of granitoid rocks in the North Pole Monzogranite and the Shaw and Carlindi Granitoid Complexes. It is also developed over peridotites and a variety of lithology in the bends of streams. The limestone is a bedrock destructive feature, variably silicified, and developed in situ.

Dissected, consolidated colluvium (*Czc*) derived from adjacent rock outcrops is deposited in small areas over much of NORTH SHAW. Composed of clay, silt, and sand, it is most widely deposited on flat granitoid complexes and on low slopes or flat plains derived through erosion of topographically high points. A specific variety of variably consolidated and dissected colluvium (*Czcg*) is composed of quartz–feldspar clay, silt, sand, and gravel derived proximally from, and deposited on, granitoid rocks.

Unconsolidated, colluvial sand, silt, and gravel (*Qc*) formed on outwash fans and on scree and talus slopes in small pockets across the rugged greenstone terrain of

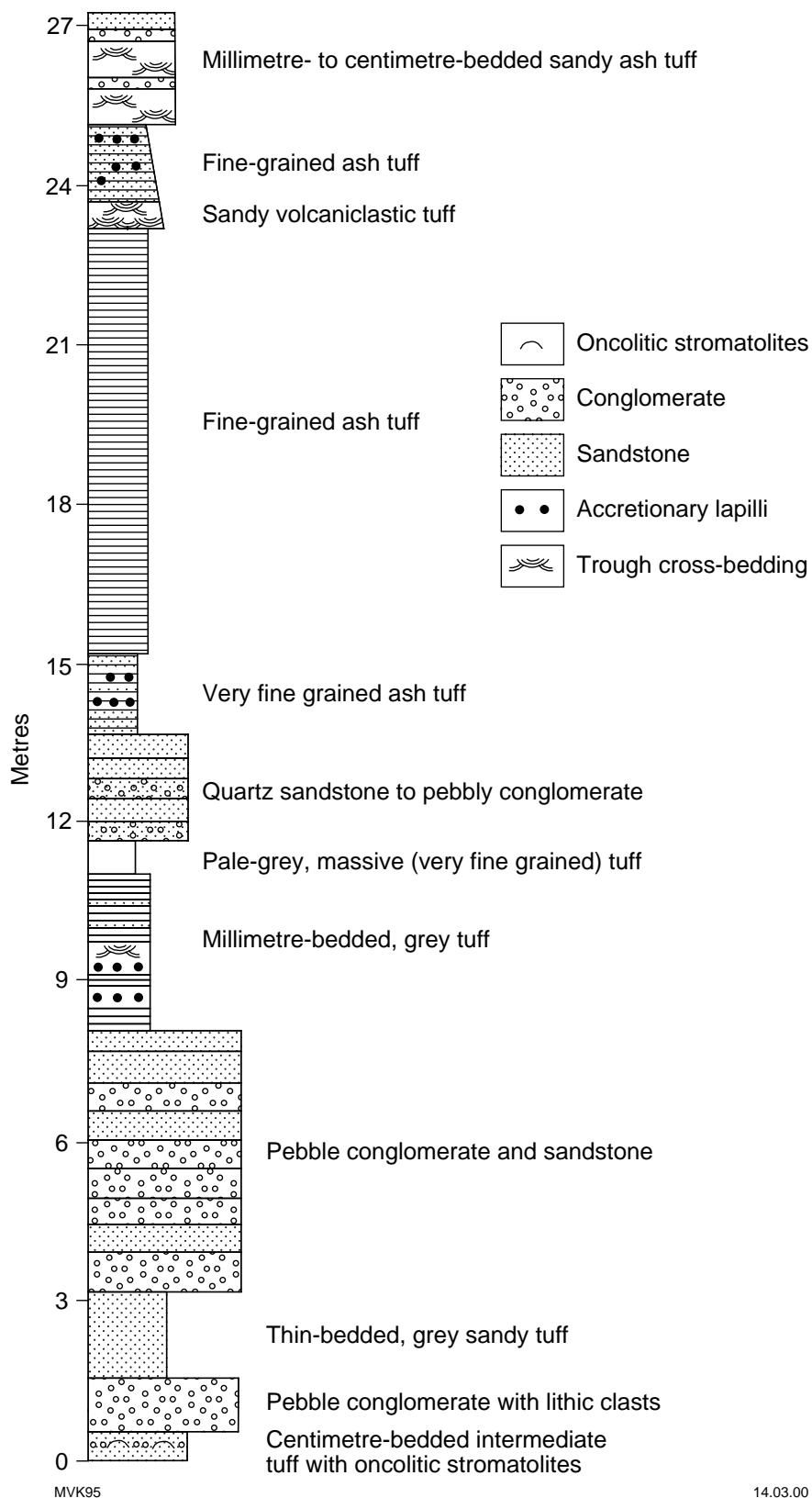


Figure 26. Stratigraphic section through volcaniclastic sandstones of the Kylena Formation in the Olympic Pool Outlier

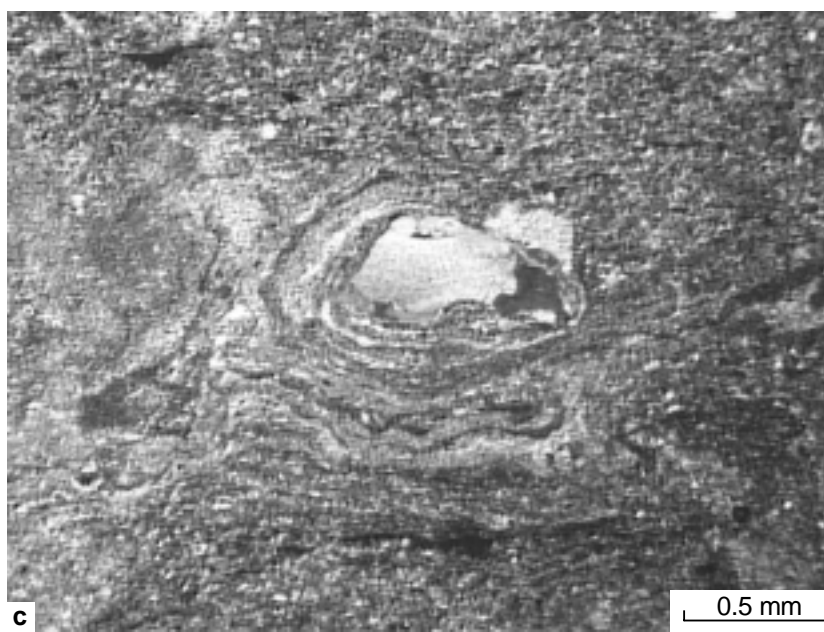
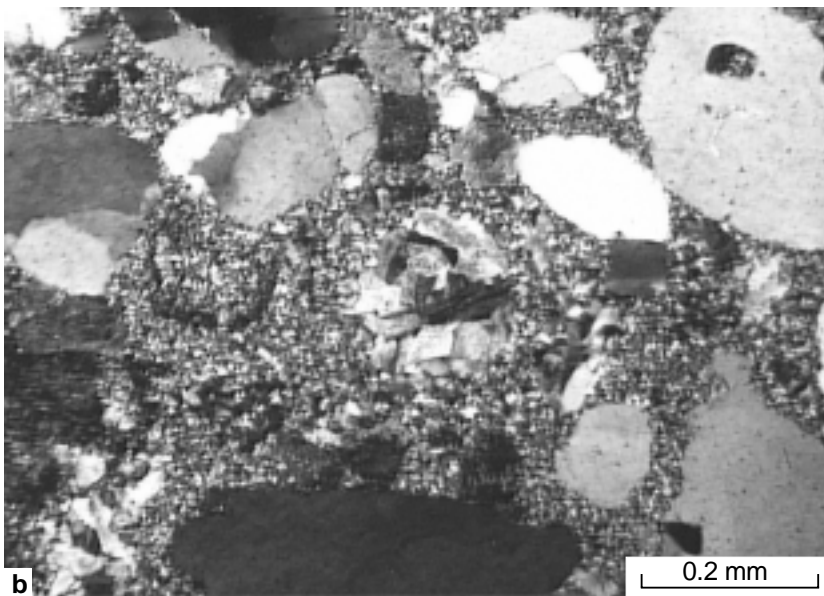
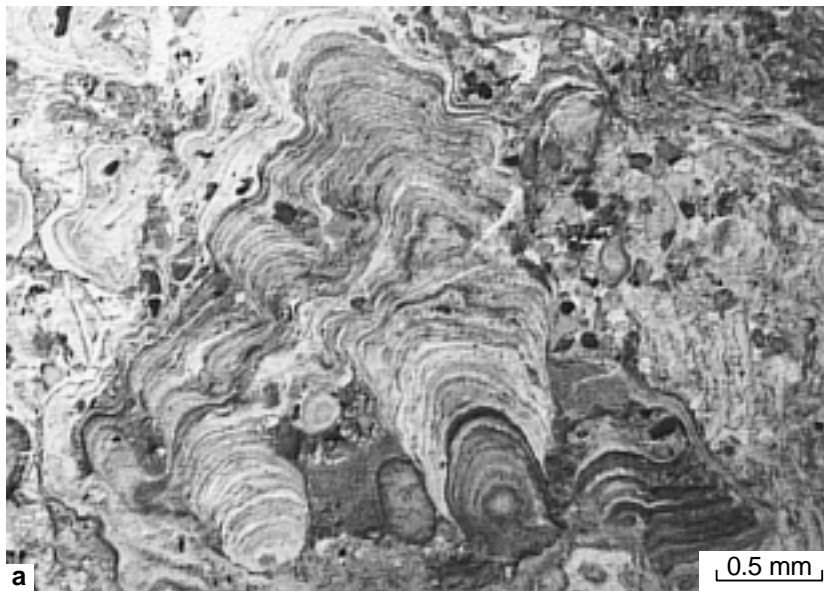
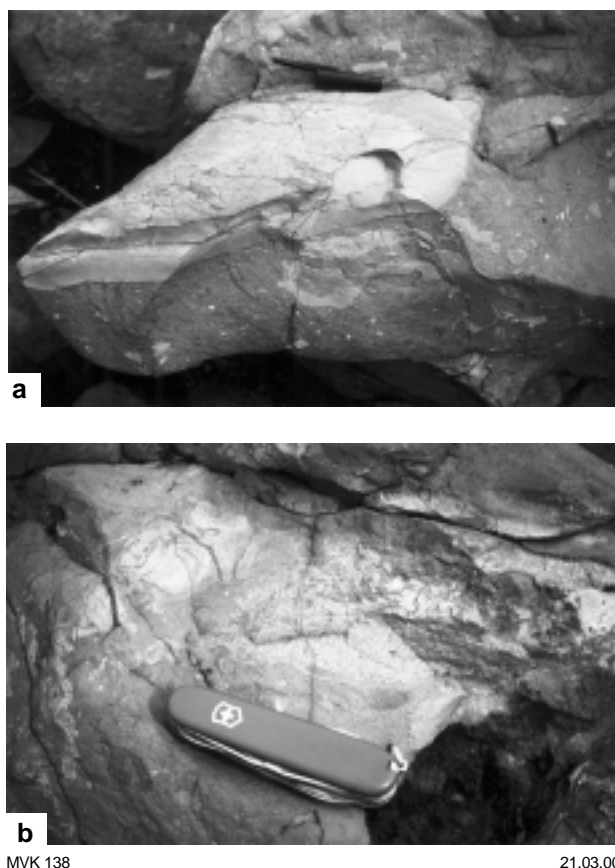


Figure 27. Thin section textures of the Kylene Formation in the Olympic Pool Outlier: a) plane-polarized light view of oncolitic stromatolite with entrained micritic carbonate pelloids in the Kylene Formation in the Olympic Pool Outlier. From the base of the section shown in Figure 26; b) cross-polarized light view of devitrification textures of radiating plagioclase laths developed in hot volcanoclastic sandstones; c) cross-polarized light view of accretionary lapilli tuff developed around a nucleus of volcanic quartz. Way up is towards the top of the photo, as indicated by erosion of the armoured lapilli rind

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**Figure 28.** Intermediate (?andesitic) tuff of the Kylena Formation in the Olympic Pool Outlier: a) coherent fragment of siltstone (light colour) in tuff matrix that contains irregular, partly liquified, rip-up clasts of siltstone that have zoned margins indicating reaction with a hot volcanic matrix (pen cap for scale); b) zone of liquified rip-up clasts (above 9 cm long penknife) and more coherent, but still irregular, clasts in tuff matrix

**NORTH SHAW.** These surficial deposits reflect the processes of proximal mass wasting due to uplift and erosion. Quartzofeldspathic eluvial sand (*Qrg*), with quartz and rock fragments, overlies, and was derived from, granitoid rocks in areas with low slopes. Sheetwash deposits (*Qw*) are composed of silt, sand, and pebbles, and were deposited on distal fans with no defined drainage, which are on low-gradient plains distant from the rugged greenstone hills. Sheetwash sand and quartz-pebble deposits derived from granitoid rocks (*Qwg*) directly overlie areas of granitoid rock. Orange to light-red coloured sand (*Qs*) forms in undulating sheets in small areas on the Carlindi Granitoid Complex, and is interpreted to be small outliers of eolian sand.

Rivers and creeks contain unconsolidated silt, sand, coarse sand, and gravel (*Qaa*). In the larger streams and the Shaw River, these deposits are commonly well sorted, although they may be variable across and along the main drainage channels, with broad sandy areas, gravel and pebble bars, and broad pebbly washes. The active stream channels are typically bound by lines of snappy gums,

beyond which are marginal, overbank deposits (*Qao*) partly stabilized by sparse undergrowth, which consist of alluvial sand, silt, and gravel. These deposits commonly form on flat floodplains adjacent to main drainage channels at bends in the streams, or on low islands within the streams.

## Structure

### Introduction

Five regional sets of structures are recognized on NORTH SHAW on the basis of crosscutting relationships between structures and magmatic phases in granitoid domes, by unconformities between volcano-sedimentary groups that contain different sets of structures, and by overprinting sets of structures in some outcrops. A brief overview of the structural history of NORTH SHAW and TAMBOURAH is presented in Van Kranendonk (1997), who showed that there are several different types of fold geometry and ages of folding. Nijman et al. (1998) described synvolcanic structures developed in the Panorama Belt during deposition of the Dresser Formation. Wilhelmij and Dunlop (1984) described syndimentary faults associated with deposition of the Gorge Creek Group in the Pincunah Belt, which Van Kranendonk (1997) ascribed to regional  $D_2$  deformation. Van Kranendonk and Collins (1998) described the regional  $D_3$  set of structures across NORTH SHAW and TAMBOURAH, and presented a detailed model linking a wide range of structures with deposition of the De Grey Group during sinistral transpression. These authors interpreted the age of this event as c. 2950 Ma and presented a revised tectonic setting for the Lalla Rookh Synclinorium and related outliers of the De Grey Group. Zegers et al. (1998) confirmed a c. 2950 Ma age for this event by dating a synkinematic pegmatite from the Mulgandinnah Shear Zone along the western margin of the Shaw Granitoid Complex. These authors also described an easterly striking shear zone in the central-northern part of the Shaw Granitoid Complex on NORTH SHAW — the Split Rock Shear Zone — which they interpreted to be a  $D_1$  shear zone. However, Van Kranendonk (1998) showed that this structure did not continue onto the NORTH SHAW map area, as suggested by Zegers et al. (1998).

### $D_1$ doming: c. 3470–3400 Ma

$D_1$  structures were observed in the Coonterunah Group in the East Strelley Belt, the Warrawoona Group in the Panorama Belt, and in the Shaw, Yule, and Carlindi Granitoid Complexes.  $D_1$  structures include tight folds of bedding in parts of the Coonterunah Group, growth faults in the lower part of the North Pole Dome formed during deposition of the Dresser Formation, and isoclinal folds of migmatitic leucosomes in parts of the granitoid complexes. The  $D_1$  structures are interpreted to have formed during deposition of the Warrawoona Group and widespread intrusion of early TTG's, now exposed in the domal granitoid complexes, and are, therefore, most likely to have developed between c. 3470 and 3400 Ma.

There is a set of mesoscale, tight, 'Z'-asymmetrical  $F_1$  folds in the southeasterly facing, partly overturned panel of the Coonterunah Group along the southeastern contact of the Carlindi Granitoid Complex. These folds have developed in banded iron-formation of the Coucal Formation (*Aoci*) on the banks of Sulphur Springs Creek (AMG 268653). Fold hinges are commonly brecciated and filled by white quartz, and fold limbs may be displaced across minor faults. The folds have shallow northeasterly and southwesterly trending plunges on steeply west-dipping axial planes (Van Kranendonk, 1997). Such structures do not form in the unconformably overlying rocks of the Salgash Subgroup and are thus constrained to be younger than 3515 Ma (the age of the Coonterunah Group; Buick et al., 1995), but older than, or equal to, the age of the c. 3458 Ma Salgash Subgroup (Thorpe et al., 1992a). Such folds are not present in the easterly striking panel of the Coonterunah Group along the southern flank of the Carlindi Granitoid Complex, where the older rocks lie conformably beneath the Strelley Pool Chert.

Van Kranendonk (1997) showed that when bedding in the unconformably overlying Salgash Subgroup is back-rotated to the palaeohorizontal, the orientation of bedding and  $F_1$  axial planes were originally moderately to steeply east dipping away from the Carlindi Granitoid Complex. In this reconstruction, the original sense of vergence of the folds was to the east, away from the Carlindi Granitoid Complex. The fold geometry and their vergence away from the Carlindi Granitoid Complex is consistent with their formation as gravity-driven cascade folds during the main phase of granitoid plutonism and uplift at c. 3467 Ma, as documented for folds adjacent to other Pilbara domes (Collins, 1989) and in centrifuge models (Dixon and Summers, 1983).

The stratiform chert–barite horizons of the Dresser Formation are fed by a conjugate boxwork set of chert–barite dykes that radiate out from the core of the North Pole Dome (Hickman and Lipple, 1978; Hickman, 1983; Nijman et al., 1998). In the Barite Range east of the North Pole Monzogranite, chert–barite dykes fill growth faults that bound structural grabens in which the Dresser Formation becomes notably thicker and consists of up to three chert–barite horizons, as opposed to only one horizon elsewhere. The bounding faults of the graben are listric structures that flatten and link together about 1 km beneath the graben. Nijman et al. (1998) showed that growth faulting and chert–barite dyke intrusion were contemporaneous with the deposition of chert–barite strata. Although faults filled by chert–barite veins beneath the Dresser Formation continue into the overlying strata, these structures at stratigraphically higher levels are not filled by chert–barite and are thus interpreted to represent later re-activation of the older faults. West of the  $D_5$  Antarctic Fault, which cuts off the western margin of the North Pole Monzogranite, a small circular area of the Mount Ada Basalt capped by the Dresser Formation is laced with chert–barite dykes. The circular distribution of the Dresser Formation and abundance of dykes in this area suggest that it was originally the cap over the North Pole Monzogranite, which was subsequently displaced down and to the west across the Antarctic Fault.

Age data from the overlying Panorama Formation show that the synvolcanic dyke intrusion and basin formation during deposition of the Dresser Formation occurred prior to c. 3458 Ma (Thorpe et al., 1992a). This is confirmed by the fact that palaeocurrent analysis of the 3458 Ma Panorama Formation indicated that the North Pole Dome was already, at least in part, a topographic dome during eruption (Di Marco and Lowe, 1989a). Evidence that the North Pole Dome was reactivated at c. 3410 Ma comes from a Pb–Pb date of this age on galena in a quartz vein associated with faulting (Thorpe et al., 1992b). Further amplification of the dome continued throughout  $D_2$ – $D_5$ .

$D_1$  structures in the granitoid complexes include isoclinal folds of migmatitic veins in c. 3460 Ma granitoid rocks (e.g. AMG 428252). Whereas such veins have not been directly dated, those in the Shaw Granitoid Complex are locally cut by c. 3430 Ma intrusions, as determined by preliminary SHRIMP U–Pb zircon data from rocks on TAMBOURAH. On NORTH SHAW, further evidence for  $D_1$  deformation is provided by c. 3410–3400 Ma zircon overgrowths on older zircon cores in leucogranite diatexite of the Shaw Granitoid Complex (*AgSl*; see **Geochronology**, p. 73). This interpretation is consistent with age data obtained from three samples of orthogneiss across the Shaw Granitoid Complex, which indicated migmatite generation at between  $3427 \pm 5$  Ma and  $3415 \pm 4$  Ma (Zegers, 1996). As described above (see **Shaw Granitoid Complex**), the porphyroclastic straight gneiss along the contact between the diatexite (*AgSl*) and orthogneiss (*AgSn*) units (AMG 430236), which displays north-side down kinematics, was interpreted to be synchronous with melting and diatexite formation and is thus a  $D_1$  structure. The penetrative foliation in rocks of the North Shaw Suite, which is parallel to the granite–greenstone contact, is interpreted as  $D_1$  in age because it is folded by  $D_2$  folds.

In the Yule Granitoid Complex, Van Kranendonk (1997) showed that migmatitic gneisses are cut by c. 3240 Ma synkinematic granitoid sheets and were thus transformed into migmatites prior to, or during, this time. In the Carlindi Granitoid Complex, homogeneous granitoid phases (*AgLm* and *AgLl*), dated at c. 3484 and 3467 Ma respectively (Buick et al., 1995; Nelson, 1998), cut weakly migmatitic tonalite of the Wilson Well Gneiss (*AgLwi*), thereby indicating an older age for magmatism and migmatization.

It is uncertain what the presence of migmatite veins ( $\geq 3400$  Ma) in orthogneiss and contact-parallel foliations in granitoid rocks mean in terms of tectonics. On the one hand, these planar foliations may record the effects of an early, craton-wide tectonic event that was not widely recorded in the greenstones, or they may have formed due to the heat of ongoing magmatism and compression during associated structural doming in the granitoid complexes (Hickman, 1984; Van Kranendonk, in prep.b). A magmatic doming origin is supported by the observation made by Zegers et al. (1996) that deposition of the c. 3467 Ma Duffer Formation on the adjacent MARBLE BAR sheet was accompanied by synvolcanic listric faulting during doming of the Shaw Granitoid Complex.

## D<sub>2</sub> continued doming and synvolcanic structures: between c. 3360 and 3240 Ma

Several different types of structures are grouped together under D<sub>2</sub>. The oldest of these include tight, overturned folds of gneissosity and foliated rocks of the North Shaw Suite in the Shaw Granitoid Complex, and synformal anticlines in the North Shaw and Coongan Belts. Whereas the ages of these structures have not been directly dated, circumstantial evidence suggests that they formed at c. 3359–3310 Ma, although it is likely that they formed progressively throughout D<sub>1</sub>–D<sub>5</sub> time by increments of coaxial deformation (Van Kranendonk, in prep.b). Three other types of structures related to D<sub>2</sub> deformation are constrained as less than or equal to 3240 Ma, including upright dome-and-basin folds, foliations and lineations in the Yule Granitoid Complex and Abydos Belt, folds and growth faults related to the emplacement of the Strelley Granite, and horst and grabens in the Pincunah Belt.

A set of close to tight, north-northwesterly trending folds in the southern part of the Shaw Granitoid Complex are interpreted as D<sub>2</sub> in age because they fold  $\geq 3400$  Ma rocks and structures, and are refolded by D<sub>3</sub> folds and transected at a high angle by S<sub>3</sub> foliations. These D<sub>2</sub> folds are overturned to the west, and have easterly dipping limbs and moderate to steep, northeasterly plunging fold axes (calculated from stereoplots of poles to folded gneissosity and foliations). The fold axes are subparallel to mineral elongation lineations in the central part of the Shaw Granitoid Complex, such that the lineations are interpreted as D<sub>2</sub> structures. The D<sub>2</sub> folds and lineations have been moderately realigned by compression during D<sub>3</sub> folding, such that the trend of the lineations varies from east to west across the granitoid complex.

In the northwestern part of the Shaw Granitoid Complex and across into the North Shaw Belt, amphibolite-facies mineral elongation lineations trend north-northeast with shallow plunges. These linear fabrics are developed in rocks also affected by a penetrative flattening fabric and associated foliation, such that pillows are deformed into flattened cigar shapes. Around the western convex contact of the Shaw Granitoid Complex, L<sub>2</sub> lineations plunge shallowly to the north and northeast, varying about the horizontal, but become steeply southerly plunging further north where bedding is overturned (AMG 470387).

Rocks of the North Shaw Belt are deformed into several tight D<sub>2</sub> folds (see Fig. 6). A tight, northwesterly plunging D<sub>2</sub> syncline affects a sliver of Duffer Formation 42 km west of Shaw River (AMG 390265), and is cut by a southerly plunging D<sub>3</sub> syncline and related fault at its northwestern end. West of this, a downward-facing, southerly plunging synformal anticline is present within the Duffer Formation (AMG 365265), as reflected in the map cross section between marker points K and L. Further north, two shallow, northeasterly plunging anticlines are present in the northeasterly striking part of the belt (AMG 390330 and 408317). The plunge of the fold adjacent to the granitoid complex becomes increasingly

steeper to the northeast, and in the related en echelon fold to the northeast, bedding around the closure of D<sub>2</sub> folds becomes overturned and the fold plunges to the southwest, indicating that it is a synformal anticline. A similar fold, symmetrically opposite across the Shaw Granitoid Complex in the Coongan Belt, plunges to the south-southeast. This structure passes along trend from a steeply north plunging anticline in the north, through a vertical plunge, to a synformal anticline in the south. The origin of these folds is discussed in **Geological evolution**.

The age of D<sub>2</sub> folds in the North Shaw and Coongan Belts, and in the Shaw Granitoid Complex is constrained between the age of rocks and structures they deform (c. 3400 Ma) and those structures that deform them (D<sub>3</sub>, c. 2940 Ma; see below). This leaves a wide age range of possible formation, although there are indications from the results of previous geochronology that they formed at c. 3340  $\pm$  20 Ma. The pertinent geochronological data includes a Pb–Pb isochron age of 3338  $\pm$  52 Ma from a suite of weakly foliated samples of the Coolyia Tonalite (*AgSco*) collected from the same place where Nelson (in prep.) recently obtained a U–Pb SHRIMP zircon age of 3469  $\pm$  2 Ma. This indicates that the c. 3340 Ma age must record the time of isotopic resetting associated with fluid mobilization and metamorphism. This age is widespread to the east of the Shaw Granitoid Complex, where it is related to the main phase of granitoid doming (see **Previous investigations**; e.g. Williams and Collins, 1990; Collins et al., 1998). Additional evidence for a metamorphic event at this time is presented by Zegers (1996), who showed that several samples from the Shaw Granitoid Complex experienced zircon recrystallization at between 3359 and 3251 Ma.

Several types of synvolcanic structures developed during deposition of the Sulphur Springs Group and emplacement of the Strelley Granite. These include growth faults, synsedimentary slump folds, and folds of cherts proximal to the Strelley Granite that were softened by the effects of contact metamorphism. Details of growth faulting around the top of the Strelley Granite are presented in Hill (1997), Vearncombe et al. (1998), and Brauhart (1999). Specific examples include those underlying the Sulphur Springs and Kangaroo Caves prospects, which provided passageways for mineral-rich fluids and resultant volcanogenic massive sulfide (VMS) mineralization. Another set of prominent growth faults is present in, and flanking, a graben dominated by basalt–andesite along the southeastern flank of the Strelley Granite (AMG 255415). Growth faults also formed around the Jamesons prospect, where they strike north–south (e.g. AMG 275380).

The northern apex of the asymmetrical Strelley Granite laccolith (sphenolith; AMG 269582) is flanked by rocks that display different types of synmagmatic structures. On the eastern side, just west of the Sulphur Springs prospect, the top of the Sulphur Springs Group is underlain by up to 1 km of olistostrome breccia in which synsedimentary slump folds are common. A kilometre-scale panel of iron formation within the breccia is deformed into a tight, easterly facing fold, the southern limb of which is transformed into a slump-induced tectonic conglomerate

and truncated by massive, unsorted breccia. Numerous, highly non-cylindrical, contorted folds (e.g. Fig. 22d) and small-scale thrusts in the iron formation all lack a tectonic cleavage and thereby attest to the fact that deformation occurred in the sedimentary environment, and not as a result of later tectonism. A rhyodacite sill (*AScfr*) occupies the core of the synsedimentary olistostrome breccia. The presence of pepperite along the contacts of the sill, irregular intrusion shapes, and the presence of floating blocks of felsic volcanic rock in the breccia (Hill, 1997) are used to interpret a syndeformational emplacement age for the felsic sill in the core of the breccia (Fig. 29).

West of the apex of the Strelley Granite, tight folds in chert of the Kunagunarrina Formation (*ASkc*) at the Roadmaster prospect have highly irregular forms and variably plunging axes, but verge primarily to the west. In this same area, numerous disrupted, isolated panels of chert are within felsic (columnar jointed in places) volcanic rocks and are locally juxtaposed against more continuous horizons of chert across knife-sharp contacts. These folded and disrupted rocks are only within 100 m of the contact of the Strelley Granite and contain the effects of contact metamorphism, notably coarse epidote spots. The extremely ductile and irregular nature of folds in the chert, their vergence away from the apex of the Strelley Granite, and the local disruption of chert by the intrusion of subvolcanic felsic sills are used to infer that these structures were developed during emplacement of felsic magmas and inflation of the Strelley Granite laccolith (Fig. 29). In this model, folding and disruption of the cherts is suggested to have developed due to slumping of overlying greenstones off the top of the rising sphenolith.

The opposite vergence of folds in chert (westerly verging) and in the olistostrome breccia (easterly verging) suggest that the northern apex of Strelley Granite was a topographic high during granite intrusion. This interpretation is supported by the observation that felsic volcanic rocks above the inferred apex at Sulphur Springs are relatively thin, but thicken to the east, 'downslope'. This model predicts that the area above the sphenolith apex would have been an eruptive centre of genetically related felsic volcanic rocks, a premise supported by the large volume of felsic volcanic sills in this area, by the fact that the largest tonnage of VMS ore is in the Sulphur Springs prospect, and the presence of an apron of banded iron-formation (*ASci*). Late-stage bulging of the middle part of the Strelley Granite has been inferred from the metallogeny of greisen veins in the outer rind of the Strelley Granite (Brauhart, 1999), and this may relate to the formation of the rhyodacite dome at Kangaroo Caves (Fig. 29).

D<sub>2</sub> deformation in the Yule Granitoid Complex and Abydos Belt resulted in the formation of a small-scale, dome-and-basin map pattern. Early structures related to this event include a penetrative schistosity in amphibolite-facies supracrustal rocks of the Golden Cockatoo Formation, and rare, recumbent, tight folds of bedding (Fig. 30a). These structures are refolded by upright, doubly plunging, open to tight folds on steeply dipping, north-northeasterly striking axial planes (Fig. 30b), which

developed in conjunction with a penetrative foliation and mineral elongation lineation. Van Kranendonk (in prep.a) documented a progressive development of recumbent to upright structures associated with the synkinematic emplacement and doming of granitoid rocks. A dyke of foliated granodiorite (*AgYka*) that intruded into the axial plane of an upright D<sub>2</sub> fold of orthogneiss, about 1 km west of the map area on WODGINA, was dated at c. 3240 Ma (M. J. Van Kranendonk, preliminary U–Pb SHRIMP zircon data) and interpreted to represent the time of folding and doming of these rocks (Van Kranendonk, 1997, in prep.a). The formation of foliations and downdip mineral elongation lineations of talc and chlorite in schistose ultramafic rocks of the Numerous Scrapes Deformation Zone is also interpreted as D<sub>2</sub> in age (Fig. 30c; Van Kranendonk, 1997, in prep.a).

Wilhelmij and Dunlop (1984) and Wilhelmij (1986) described syndepositional horst and graben faulting during deposition of the Gorge Creek Group in the Pincunah and East Strelley Belts. Although the minimum age of the Gorge Creek Group in this area is unknown, the field relationships suggest that it is not much younger than the Sulphur Springs Group and probably related to the same depositional cycle.

## D<sub>3</sub> regional transpression: c. 2940 Ma

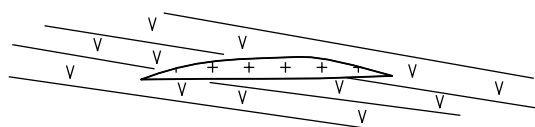
Structures formed during regional D<sub>3</sub> deformation at c. 2940 Ma are present across NORTH SHAW, but are most strongly developed within the 5–20 km-wide, north-northeasterly trending LWSC, which transects the central part of NORTH SHAW (Van Kranendonk, 1998).

## D<sub>3</sub> structures in the LWSC

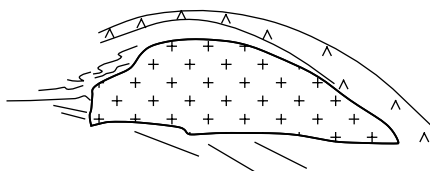
The low- to medium-grade rocks of the LWSC are deformed by an anastomosing network of greenschist-facies strike-slip faults, high-angle reverse faults, and commonly northeasterly trending folds related to regional D<sub>3</sub> transpressional deformation, as shown in Figure 31 (Van Kranendonk and Collins, 1998; Krapez, 1989). The northeasterly trending folds in the LWSC are wholly contained within the western and eastern boundary fault systems and do not affect the bounding faults, thus indicating that folding occurred during faulting (Van Kranendonk and Collins, 1998).

The western boundary of the LWSC is formed by the curvilinear, sinistral Lalla Rookh – Western Shaw Fault that extends north-northeast across NORTH SHAW. In the north, this fault forms the northwestern boundary of the Lalla Rookh Synclinorium, where it is a 5–10 m-wide zone of silicified chlorite schist with shallow, south-westerly plunging, striping lineations. From there, the fault continues west across the top of the Strelley Granite where it is up to 20 m wide and developed within talc–tremolite–carbonate schists, in which sinistral S–C–C' textures (Berthé et al., 1979) were locally observed. The fault then curves south-southwest, just west of the Strelley Granite, where it forms the boundary between the Soanesville and

## Synvolcanic emplacement of granite sill into mafic volcanic rocks

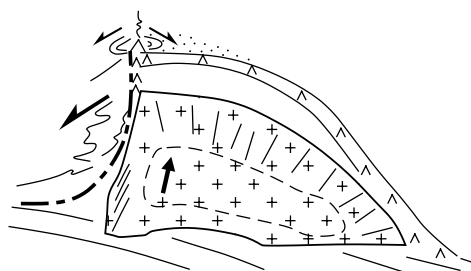


## Continued felsic magma injection into laccolith and surface extrusion



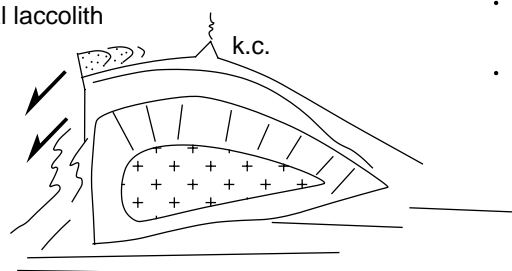
- Hydrothermal circulation
- Cu–Zn ore deposition

## Ballooning of laccolith through injection of inner phase



- Major volcanic centre developed over highest point of laccolith
- Slump folds generated in older rocks away from the topographic high
- Development of radial greisen veins in outer phase
- Zn–Cu deposition

## Bulging of central laccolith



- Late development of Kangaroo Caves Dome (k.c.)
- Slump folds



Olistostrome breccia with fold



Strelley Granite



Felsic volcanic rocks



Mafic volcanic rocks



Trend of bedding

MVK81

02.03.00

Figure 29. Diagrammatic sketch of the evolution of the Strelley Granite. Asymmetrical ballooning of the synvolcanic laccolith caused the escape of felsic magmas out of the top left of the sill, underneath Sulphur Springs. Ballooning caused the formation of slump folds in older rocks to the west, off the apex of the dome, and coeval deposition of the olistostrome breccia downslope to the east, which contains syndepositional slump folds



MVK 141

21.03.00

Figure 30. Structures from the Abydos Belt: a) recumbent, tight fold of banded iron-formation; b) upright  $D_2$  fold of cherty banded iron-formation; c) high-strain tectonic schist in the Numerous Scrapes Deformation Zone developed in talc-chlorite schist. Foliations dip away from the Yule Granitoid Complex to the northeast (left of photograph), and a contained mineral elongation lineation is downdip

East Strelley Belts, and then the boundary between the Soanesville and Pincunah Belts. Southwards, the fault is developed within the eastern margin of the Yule Granitoid Complex as a 2 km wide zone up to sinistral porphyroclastic mylonite developed in the synkinematic Woodstock Monzogranite (Fig. 32). At least 15 km of translation across this fault system is suggested by the offset of the unconformity at the base of the De Grey Group between the East Strelley Belt and Lalla Rookh Synclinorium.

Three en echelon fault segments form the eastern boundary of the LWSC. The most easterly of these is the northern extent of the northerly striking Mulgandinnah Shear Zone along the narrow, western sliver of the Shaw Granitoid Complex. This zone is composed of amphibolite- to greenschist-facies, sinistral porphyroclastic mylonite and straight gneiss. The mylonitized granitoid rocks are separated from adjacent greenstones by a narrow chloritic fault that locally contains tectonic fragments of the mylonitized granite. The central splay is marked by the South Daltons Fault along the eastern margin of the Keep It Dark Synclinorium of the De Grey Group and the Olympic Pool Outlier of the Fortescue Group (see Fig. 6). The western splay is made up of the Jamesons Fault in the southwest, the Bernts Deformation Zone, and the Hogback and Flying Fox Faults along the southwestern and western boundary of the North Pole Dome. Segments of the eastern boundary fault are interspersed with predominantly northeasterly trending folds, as described below (Van Kranendonk, 1998).

The northeastern part of the LWSC is characterized by a set of tight, northeasterly plunging, symmetrical folds that are in between the bounding Lalla Rookh – Western Shaw and Hogback Faults (Fig. 6). These include, from north to south, the Strelley Anticline, Roadmaster Syncline, Sulphur Springs Anticline, Caves Syncline, Sunset Ridge Anticline, and the Hogback Syncline. The steeply east plunging Strelley Anticline located immediately north of the Strelley Granite is interpreted as a drag fold to the shearing, consistent with sinistral asymmetric foliations in the fault zone (Van Kranendonk and Collins, 1998, fig. 4g). Folds in the Lalla Rookh Synclinorium, although of significant amplitude, do not extend westward into underlying rocks, as these have been protected from penetrative deformation within a strain shadow created by the more-competent Strelley Granite.

In the Lalla Rookh Synclinorium, northwest–southeast compression becomes more intense to the northeast, away from the shielding effects of the competent Strelley Granite, and this excess shortening is accommodated by two prominent sets of branching, flower-like faults. Tight folds are also developed immediately adjacent to, but outside of, the bounding faults of the LWSC in this area. These include folds in the Gorge Creek Group just south of a bend in Sulphur Springs Creek (AMG 268640) and those in the North Pole Dome, where felsic volcanoclastic rocks of the Panorama Formation are deformed into the steeply plunging, easterly facing Caldera Anticline – Miralga Syncline pair (AMG 472745).

Rocks north of the Strelley Granite are tightly folded and faulted by a set of sinistral and dextral conjugate

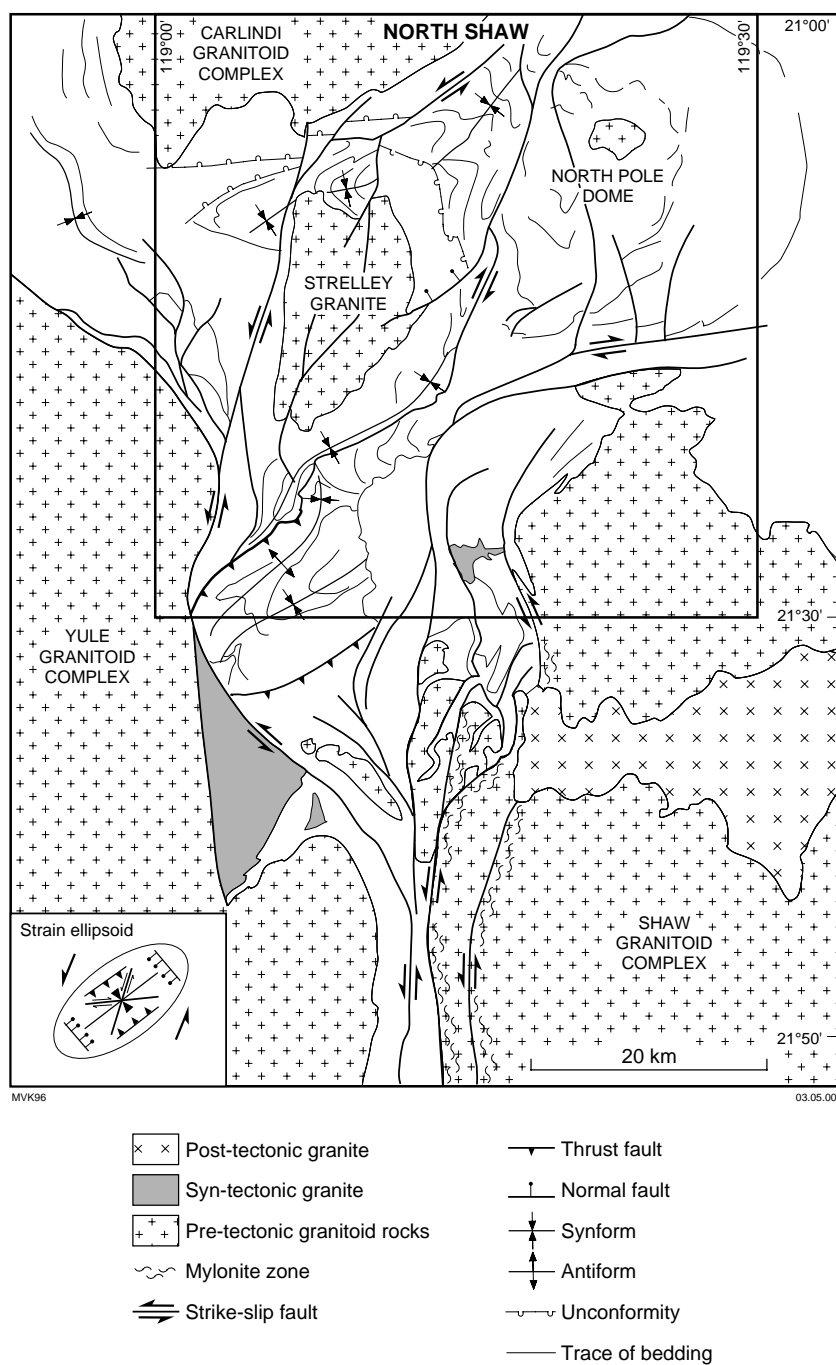
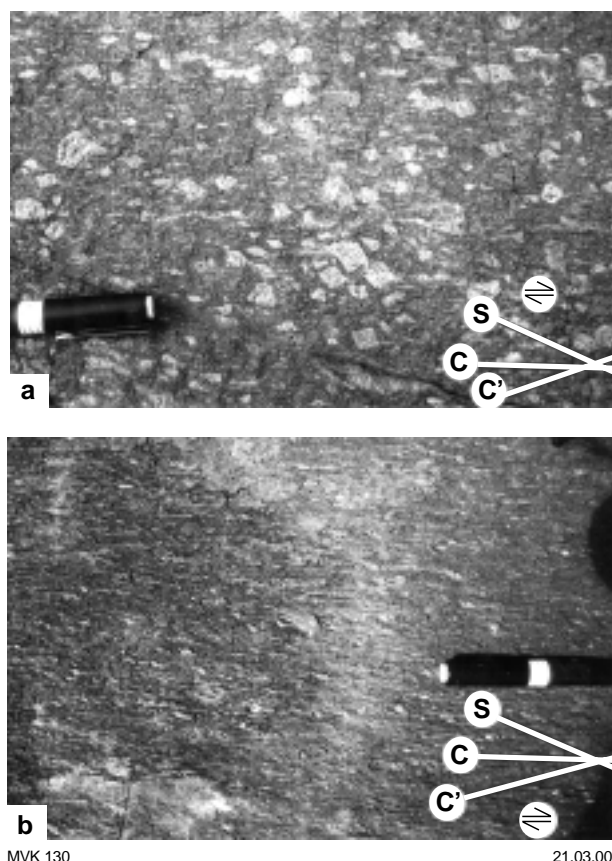


Figure 31. Principal D<sub>3</sub> structures in the LWSC formed during regional northwest-southeast compression. Note that northeasterly trending folds are bound by, and do not affect, the western and eastern boundary faults of the system, indicating coeval formation during sinistral wrenching. From Van Kranendonk and Collins (1998)

faults. These structures helped to accommodate significant shortening in the concave bend of the Lalla Rookh – Western Shaw Fault around the top of the Strelley Granite.

Kinematic indicators were not observed in the 38 km-long Hogback Fault bounding the southeastern margin of the Lalla Rookh Synclinorium. However, this subvertical

fault terminates, without splaying, in the Strelley Granite, a feature that suggests it is, at least in part, a scissor fault. Whereas the offset of the Strelley Granite contact and overlying units across the fault is apparently sinistral in plan view, the moderate northerly plunge of the related drag fold axis suggests a dominantly vertical sense of displacement. This, combined with a steep easterly dip of



**Figure 32.** Sheared Woodstock Monzogranite from the sinistral Pulcunnah Shear Zone along the western boundary fault of the LWSC: a) coarsely porphyroclastic protomylonitic granite with S-C-C' planes; b) porphyroclastic granite mylonite with a stronger development of C' extensional shear bands

the granite contact and stratigraphy, indicates a north-side-down sense of displacement. Further along strike to the northeast, a right-stepping array of fault segments along the southern margin of the Lalla Rookh Synclinorium suggests a component of dextral displacement (Davis, 1984, p. 300).

The Hogback Fault truncates the northern tip of the approximately 2 km-wide Bernts Deformation Zone of tectonic megabreccia and is therefore a relatively late  $D_3$  structure. The slightly older Bernts Deformation Zone forms the western limit of the Warrawoona Group in the North Pole Dome and juxtaposes it against steeply dipping, easterly facing younger rocks of the Gorge Creek Group in the Soanesville Belt (see Fig. 6). Within the deformation zone are kilometre-scale panels of dismembered, folded, and locally schistose volcanic and sedimentary rocks and chert, which, at least in part, belong to the Sulphur Springs and Gorge Creek Groups, and possibly includes rocks derived from the upper parts of the Warrawoona Group. A distinctive unit of steeply west dipping ultramafic tectonic schist and breccia along the eastern margin of the zone contains rare kinematic indicators of oblique dextral, reverse displacement. The western boundary of this zone is steeply east dipping, with

down-dip mineral elongation lineations and rare kinematic indicators of reverse, east-side-up displacement. These observations imply that the Bernts Deformation Zone represents a 'pop-up' structure in the core of the syncline between the Soanesville and Panorama Belts.

The Bernts Deformation Zone passes along strike southward into the Jamesons Fault, which forms the transposed southeastern limb of the Leilira Syncline in the Soanesville Belt and the northern limb of the southwesterly facing Potkoorok Anticline in the Panorama Belt. In this area, the fault has been reactivated with a component of normal, north-side-down movement after deposition of the Hardey Formation in the Leilira Outlier on folded rocks of the Gorge Creek Group. Westward, the Jamesons Fault curves south and southeast along the contact of the Sulphur Springs and Gorge Creek Groups at the western limit of the Potkoorok Anticline, where it dips west and has a reverse sense of displacement. West of this, Sulphur Springs and Gorge Creek Group rocks are deformed into tight, highly non-cylindrical folds. In this area, the axial traces of tight  $D_3$  folds fan out from a narrow cusp (AMG 210357), to the west and west-northwest, southwest, south (Pyramid Hill Syncline), southeast, east, and east-northeast (Leilira Syncline) (Fig. 33). The cusp is a steeply south plunging syncline, whereas folds of the Sulphur Springs Group to the west are doubly plunging structures, commonly with overturned limbs and locally with overturned hinges (synformal anticline). Other than a refolded syncline in the southwest, there are no overprinting relationships between the folds with different orientations, such that the folds in this area are considered to be coeval. This pattern is interpreted to reflect the effects of soft crust that has been pinched between two competent structural buttresses, specifically the southwesterly facing extension of the North Pole Dome in the east and the southern tip of the Strelley Granite in the north, which must extend further south in the subsurface than it does at surface (e.g. see also Van Kranendonk and Wardle, 1996, for another example of crust pinched between competent blocks).

In the southwestern corner of NORTH SHAW (AMG 112230), greenstones along the contact with the Yule Granitoid Complex are strongly deformed into a series of tight to isoclinal 'S'-asymmetric folds, the curving axial traces of which extend north and northeast from the Lalla Rookh – Western Shaw Fault and dissipate into more open structures higher up in the Gorge Creek Group. Several of these folds have highly attenuated, faulted western limbs in which a strong schistosity is developed with kinematic indicators of sinistral shear. The 'S'-asymmetry of the folds is consistent with the sinistral sense of shear within the Pulcunnah Shear Zone, and their development adjacent to the Yule Granitoid Complex is used to infer that they represent drag folds formed during  $D_3$  sinistral shearing along the Lalla Rookh – Western Shaw Fault (Van Kranendonk and Collins, 1998).

The Soanesville Syncline is a broad, upright, north-easterly plunging fold (AMG 167200 to 243280) outlined by up to four differentiated ultramafic–felsic sills of the Daltons Suite. Minor accommodation faults cut through the core of the syncline and a reverse fault is inferred for

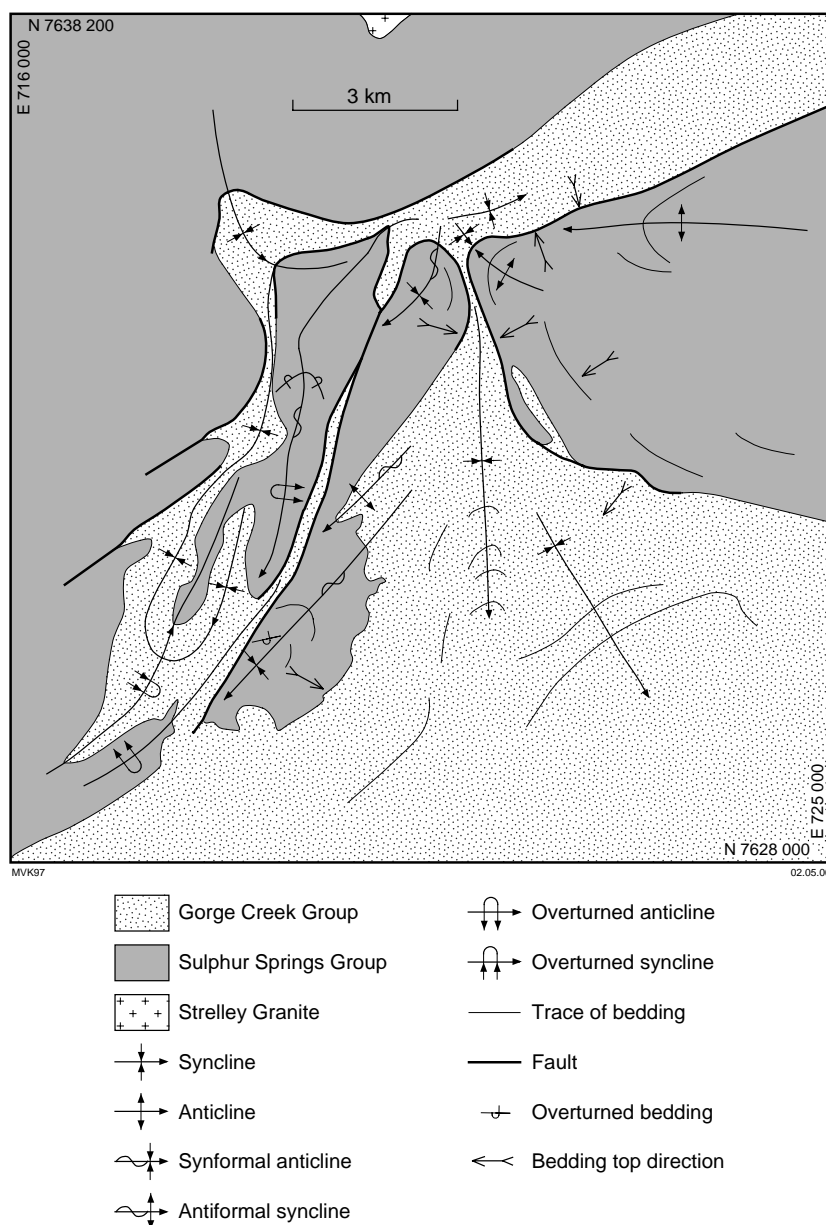


Figure 33. Detailed sketch showing the radiating nature of  $D_3$  folds in the Furies area, south of the Strelley Granite

the northwestern limb on the basis that bedding trends in the adjacent Corboy Formation sandstones (*AGct*) are tightly folded and butt up against the outermost sill of the Daltons Suite. Similarly, about 1 km to the west of the Soanesville Syncline (AMG 160225), a reverse fault is inferred along the boundary between Pincunah Hill Formation shale (*AGih*) and Corboy Formation sandstone (*AGct*) to accommodate the extreme attenuation of bedding thickness of the latter.

East of the Soanesville Syncline is the Escarpment Fault, a polyphase structure initiated during  $D_3$  and reactivated through  $D_4$  and  $D_5$  events, which separates the LWSC from the Tambina Structural Complex (see Fig. 6;

Van Kranendonk, 1998). The rocks east of this fault are tightly folded by  $D_3$  structures with north-northeasterly trending axial planes and northerly plunging axes. The Keep It Dark Synclinorium represents the faulted and folded remnant of a synorogenic clastic basin that is locally unconformable on the Gorge Creek Group and is itself, unconformably overlain by the Mount Roe Basalt. South of the Keep It Dark Synclinorium, along the very southern margin of NORTH SHAW (AMG 300210), is a small, fault-bounded sliver of the Gorge Creek Group that is deformed into a steeply dipping, northeasterly striking, and northwesterly facing panel (Boulter et al., 1987; Van Kranendonk and Morant, 1998). This small tectonic remnant is in thrust contact with mafic volcanic rocks and

cherts of the Warrawoona Group both to the north and south, and is folded into a synformal anticline plunging 60° towards 174°.

The eastern margin of the Keep It Dark Synclinorium is bounded by the South Daltons Fault, a prominent structure that marks the boundary between the Tambina and Emerald Mine Structural Complexes (see Fig. 6; Van Kranendonk, 1998). Rocks of the Emerald Mine Structural Complex are tightly folded, sheared, and commonly dismembered by strong D<sub>3</sub> deformation. The 2936 ± 5 Ma Keep It Dark Monzogranite, which lies across the northern tip of the Emerald Mine Structural Complex and cuts across upturned, amphibolite-facies greenstones of the North Shaw Belt, contains a variably developed S<sub>3</sub> foliation defined by greenschist-facies metamorphic minerals. The western edge of the Keep It Dark Monzogranite is bound by faults, but these structures do not displace the northern contact of the granite, indicating that it was emplaced during D<sub>3</sub> faulting. Thus the age of the Keep It Dark Monzogranite is interpreted to date the age of D<sub>3</sub> deformation in this area, and confirms a similar age of 2936 ± 2 Ma from the Mulgandinnah Shear Zone along the western margin of the Shaw Granitoid Complex on TAMBOURAH (Zegers, 1996).

## D<sub>3</sub> structures in the Shaw Granitoid Complex and North Pole Dome

In the southern part of the Shaw Granitoid Complex, numerous small intrusions of the 2934 ± 2 Ma Mulgandinnah Monzogranite contain a weak northerly to north-northeasterly striking, D<sub>3</sub> quartz foliation. This quartz foliation is also locally visible in the older host rocks across the southern part of the Shaw Granitoid Complex, and is axial planar to a tight, D<sub>3</sub> fold of orthogneiss (AgSn) and leucogranite diatexite (AgSl) just east of the Shaw River in the southern part of the map (AMG 428225). This fold refolds a set of earlier D<sub>2</sub> folds of gneissosity (AMG 430240).

A 500 m-wide zone of amphibolite-facies porphyroclastic mylonite and ultramylonite is developed in the thin western splay of the Shaw Granitoid Complex. This zone, characterized by subvertical mylonitic foliations and subhorizontal quartz-ribbon lineations, represents a northern splay of the Mulgandinnah Shear Zone and passes along strike into a narrow fault, which curves to the north and northeast, parallel to bedding.

Cutting the northwestern margin of the Shaw Granitoid Complex is a slightly curved, steeply east dipping reverse fault (AMG 410295). This structure cuts foliated and lineated amphibolite-facies rocks and is thus post-D<sub>2</sub> in age. The western end of this cusped fault terminates within greenstones of the North Shaw Belt, very near to where the mylonite zone along the western margin of the Shaw Granitoid Complex ends in a brittle fault. The geometry of these structures suggests that they may be linked. In this scenario, the cusped, northeasterly striking reverse fault accommodated northwesterly directed sinistral translation of rocks across the northwesterly striking strike-slip fault.

Mafic volcanic rocks of the Warrawoona Group on the southern flank of the North Pole Dome are affected by a penetrative (though patchy), steeply north dipping foliation, associated northwesterly plunging mineral elongation lineations, and rare minor folds. Although undated, these structures are interpreted to have formed during D<sub>3</sub> northerly translation of the Shaw Granitoid Complex across the sinistral Mulgandinnah Shear Zone (Van Kranendonk and Collins, 1998; see below). The northerly dip of the foliations is interpreted to reflect the northerly dip of the northern margin of the Shaw Granitoid Complex.

Just north of the North Pole Mining Centre (AMG 445657), Dresser Formation cherts are deformed into an easterly plunging anticline–syncline pair of tight folds with northerly dipping axial planes that strike parallel to bedding and dip away from the North Pole Monzogranite. These folds are ductile structures with rounded, non-cylindrical hinges that lack an axial planar foliation. The geometry of these folds indicates a reverse sense of vergence, or greenstone-side-up displacement relative to the North Pole Monzogranite, during compression related to D<sub>3</sub> deformation.

In the northwestern part of the North Pole Dome, rocks of the Panorama Formation outline a tight, easterly facing anticline with a subvertical fold axis. Limbs of this structure outline an overall 'Z' asymmetry, consistent with dextral D<sub>3</sub> shear on the bounding Hogback Fault.

## D<sub>3</sub> structures in the Pincunah and East Strelley Belts

Structures post-dating the D<sub>2</sub> formation of depositional sub-basins in the Pincunah and East Strelley Belts include a set of doubly plunging folds in sedimentary rocks of the Gorge Creek Group in the Pilgangoora Basin, on trends that parallel the contact of the Yule Granitoid Complex. These folds have curved axial traces and form isolated structures bound by D<sub>3</sub> sinistral fault splays. This peculiar geometry, combined with evidence of stratigraphic variations between fault-bounded blocks, suggests that the syndepositional (D<sub>2</sub>) horst–graben block architecture influenced the geometry of later deformation.

Other D<sub>3</sub> structures include easterly to southwesterly trending folds in Gorge Creek and De Grey Group sedimentary rocks in the East Strelley Belt. Fold axial traces follow the general outline of the Carlindi Batholith, from east–west in the west, to southwest–northeast in the east (see Fig. 6). Folds are open, easterly plunging structures in the west, but change to steeply westerly plunging structures across a northeasterly striking fault in the De Grey Group. Along the trend of these folds, axes revert to northeasterly plunges.

Blossoming from the southeastern tip of the Pincunah Belt is a positive flower structure of curvilinear northerly to northwesterly striking sinistral faults (Fig. 34). The point from which these faults splay is also the tight hinge of the northerly plunging Pilgangoora Syncline, and it is thus probable that the development of these structures was

contemporaneous with the faults accommodating some of the shortening across the fold.

The eastern limb of the Pilgangoora Syncline is underlain by a westerly facing panel of the Kangaroo Formation of the Sulphur Springs Group, which is cut off by the Lalla Rookh – Western Shaw Fault to the east. Restoration of sinistral displacement across the Lalla Rookh – Western Shaw Fault indicates that this panel links up with the truncated northern limb of the Soanesville Belt

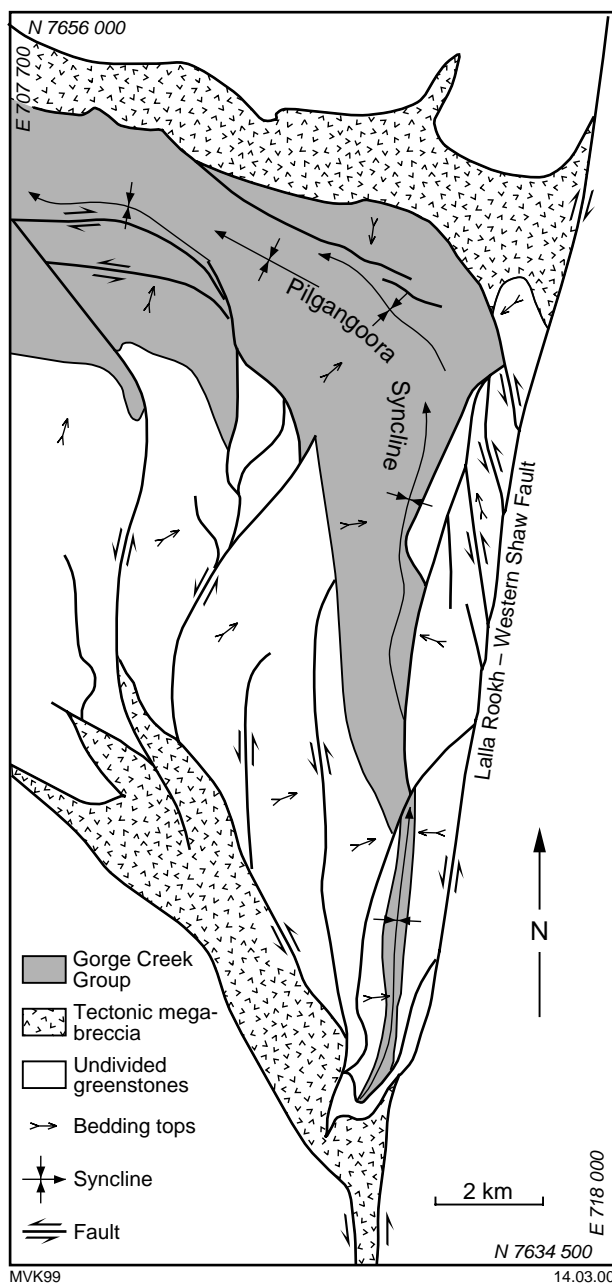


Figure 34. Sketch of  $D_3$  structures in the Pincunah Belt showing the flower-like geometry of faults and fold axes from a point in the south-central part of the area, bound in the east by the Lalla Rookh – Western Shaw Fault

immediately north of the Strelley Granite. In so doing, it is apparent that this panel represents the detached western limb of an  $\Omega$ -shaped regional fold cored by the Strelley Granite, and that the amount of displacement across the fault is about 18.5 km (Van Kranendonk and Collins, 1998).

## $D_4$ folding of the Mount Roe Basalt: <2765 Ma, >2715 Ma

The narrow septum of Mount Roe Basalt between the North Pole Dome and Shaw Granitoid Complex is folded into a tight  $F_4$  anticline and unconformably overlain in the east by shallow, easterly dipping conglomerate, sandstone, and shale of the Hardey Formation and conformably overlying basalts of the Kylenea Formation. A similar relationship is observed just to the south of the North Pole Dome (AMG 580423), where gently northeasterly dipping Mount Roe Basalt on the southern limb of a gently east-northeasterly plunging  $F_4$  fold is cut off by unconformably overlying, southeasterly dipping rocks of the Hardey and Kylenea Formations, themselves gently folded by an  $F_5$  structure. Although not as easily demonstrable, it is certain that faulting accompanied this folding, such as the normal faults bounding either side of the Mount Roe septum.

## $D_5$ faulting: <2715 Ma to Holocene

Several of the faults on NORTH SHAW experienced a component of movement after the deposition of the Kylenea Formation, most notably the Antarctic Fault, which cuts through the North Pole Dome, and the Escarpment Fault, which cuts through the Olympic Pool Outlier of the Fortescue Group. These undated fault movements document a normal sense of displacement of Fortescue Group rocks down, away from granitoid-cored domes (Fig. 35). Associated with the Antarctic Fault are splays that strike to the northeast and have sinistral movement. Two gentle folds are related to this event; a north-northwesterly plunging anticline that refolds an  $F_4$  syncline in the Olympic Pool Outlier, and a gentle syncline that affects the southern part of the Marble Bar Outlier.

The Leilira Outlier of the Hardey Formation is folded into a gentle, doubly plunging  $F_5$  syncline with a steeply dipping, drag-faulted southern limb. The  $F_5$  syncline axis is developed along the axis of the earlier  $D_3$  Leilira Syncline, but the plunge of these two fold generations varies greatly.

A recent fault scarp was observed in the Shaw Granitoid Complex (AMG 435207), marked by three parallel linear fractures within a low, smooth, red-weathering outcrop of granitoid gneiss ( $AgSn$ ). The faults strike  $027^\circ$ , dip  $80^\circ E$ , and have a reverse (granitoid or west-side-up) sense of displacement, elevating the surface of the outcrop a distance of up to 30 cm across individual faults. Metre-sized blocks with fresh blue-grey surfaces have fallen away from the faults up to a distance of 50 cm.

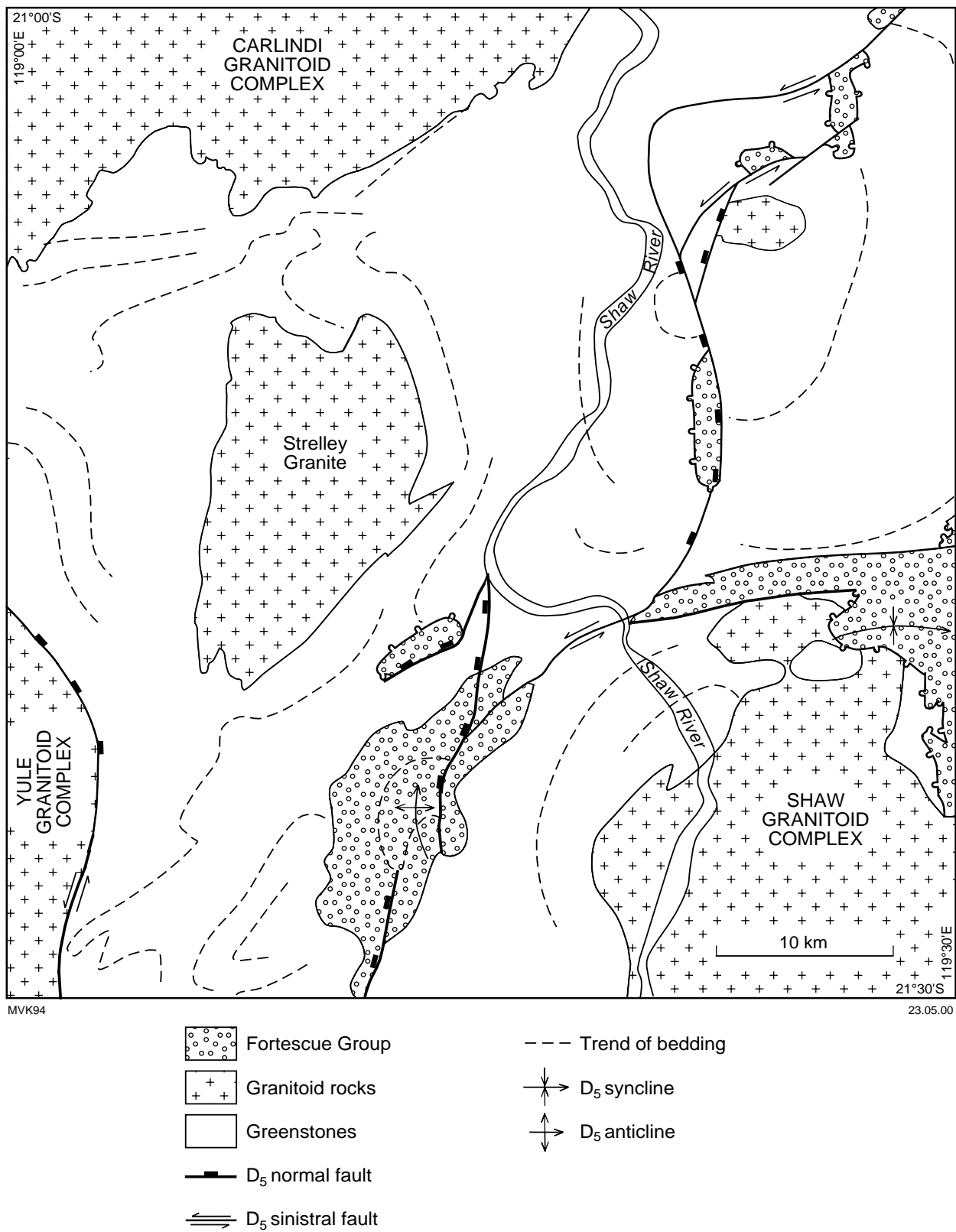


Figure 35. D<sub>5</sub> structures on NORTH SHAW showing west-side-down normal displacement during the latest phase of granitoid doming

## Metamorphism

Metamorphic grade in Pilbara Supergroup rocks varies throughout NORTH SHAW, from low temperature metamorphism in the Fortescue and De Grey Groups, through areas of prehnite–pumpellyite facies in the Gorge Creek Group, to greenschist facies throughout a large area of the map (Fig. 36). Prehnite–pumpellyite to lower greenschist facies mafic volcanic rocks outcrop in the East Strelley, Pincunah, Soanesville, and Panorama Belts. Higher grade rocks, including upper greenschist facies to lower amphibolite facies, outcrop around the margins of the North Pole Monzogranite (Mount Ada Basalt) and the Shaw and Carlindi Granitoid Complexes. Rocks of the Golden Cockatoo Formation within the Yule Granitoid Complex are also at amphibolite facies (local biotite–garnet–muscovite in metapelitic rocks), as are granitoid rocks of the Carlindi, Shaw, and Yule Granitoid Complexes. Relicts of clinopyroxene and orthopyroxene were locally observed in mafic enclaves in the Shaw Granitoid Complex (Fig. 36), indicating temperatures in excess of about 775°C (Spear, 1993).

A lack of garnet with high-temperature mineral assemblages indicates low to moderate pressures during metamorphism. The highest temperature amphibolites proximal to the Shaw Granitoid Complex contain hornblende, titanite, ilmenite, plagioclase, and quartz in a polygonal-textured mosaic. One metabasalt from the amphibolite to greenschist facies transition in the North Shaw Belt (point A on Fig. 36) contains the assemblage garnet–actinolite–tremolite–hornblende–muscovite–epidote–quartz–ilmenite. Outcrops containing relict pyroxenes within the Shaw Granitoid Complex lack textural evidence for the existence of garnet, indicating pressures of less than 8 kbars at 800°C (Spear, 1993).

The annular distribution of metamorphic facies around the granitoid complexes, and the decrease in grade with stratigraphic height, is interpreted to reflect the effects of contact metamorphism and progressive structural doming (see Van Kranendonk, in prep.b).

## Geochronology

Geochronological data for NORTH SHAW is presented in Table 3 and summarized in Figure 37. The data show that NORTH SHAW experienced a punctuated evolution, with major magmatic events at 3550–3500 Ma (Coonterunah and ?lower Warrawoona Groups), 3490–3400 Ma (upper Warrawoona Group), 3260–3240 Ma (Sulphur Springs Group), c. 2950–2930 Ma (De Grey Group), c. 2760 Ma (Mount Roe Basalt of the Fortescue Group), and c. 2715 Ma (Hardey and Kylena Formations of the Fortescue Group).

## Supracrustal succession

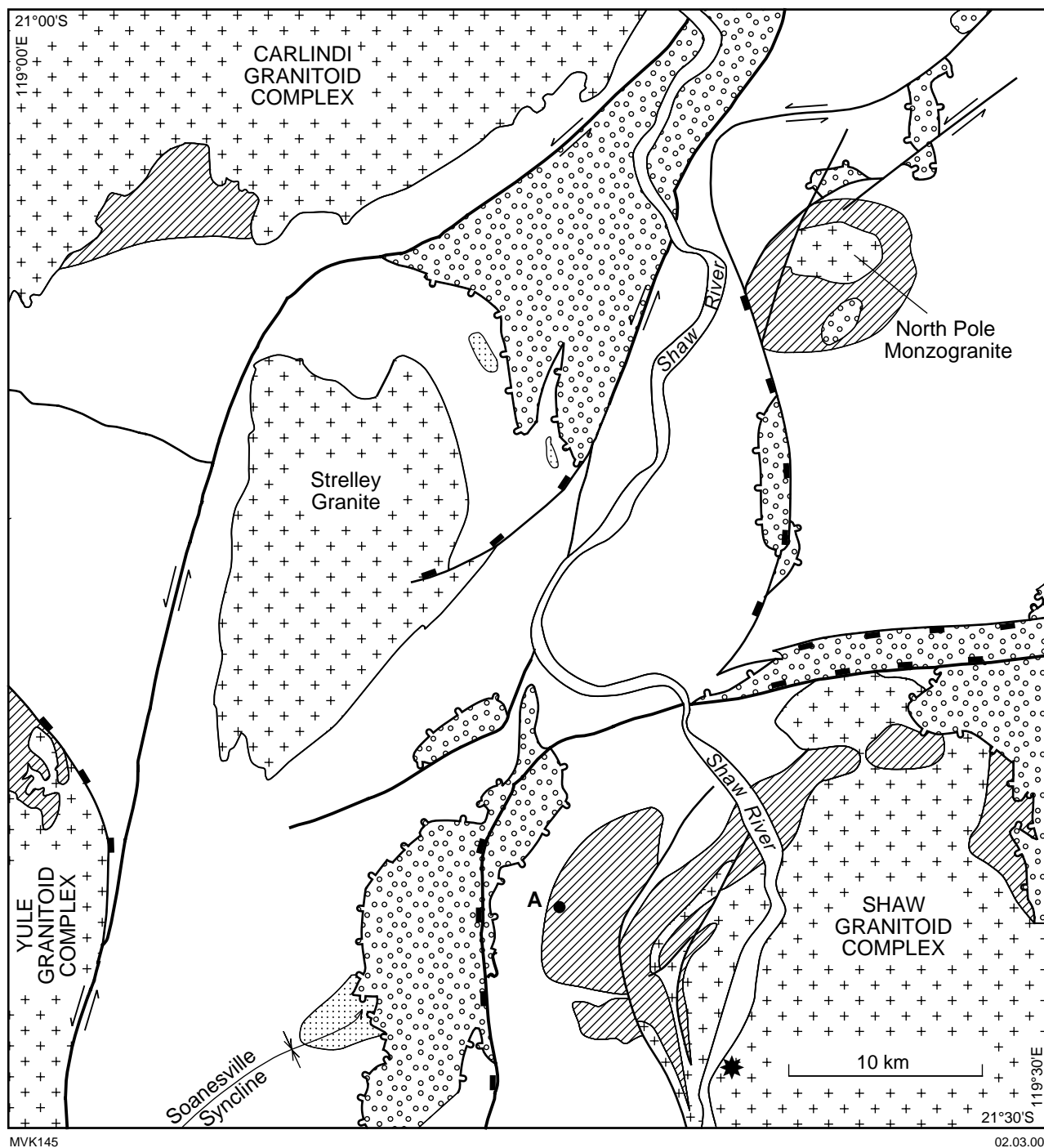
The oldest directly dated rock on NORTH SHAW is a brecciated rhyolite from the Coucal Formation of the Coonterunah Group (*AOcf*), which is  $3515 \pm 3$  Ma (sample

70601 in Table 3; Buick et al., 1995). Detrital zircons of this age were also extracted from a chert in the Euro Basalt (*AWeft*) in the Panorama Belt (sample 94001 in Table 3; Buick et al., 1995).

A similar age of  $3522 \pm 13$  Ma was obtained on hornblende from an amphibolite-facies metabasalt of the Mount Ada Basalt (*AWmba*), adjacent to the contact of the Shaw Granitoid Complex (sample T94/87 in Table 3; Zegers, 1996). This sample also yielded a semi-plateau age of  $3466 \pm 13$  Ma, which was interpreted as the hornblende cooling age of contact metamorphism during intrusion of the Coolyia Creek Tonalite (*AgSco*; Zegers, 1996), an interpretation confirmed by two U–Pb SHRIMP zircon dates on this rock of  $3469 \pm 2$  Ma (sample 142962) and  $3467 \pm 6$  Ma (sample 91085). Therefore, the  $3522 \pm 13$  Ma Ar–Ar age probably represents a true age of the basaltic rocks and suggests that the Mount Ada Basalt may belong to the Coonterunah Group. A potentially old age for basal volcanic rocks of the Warrawoona Group in the Panorama Belt is supported by a Pb–Pb model age of c. 3490 Ma (sample 415) on galena from depositional–diagenetic barite in the Dresser Formation (*AWrc*; Thorpe et al., 1992b). A minimum age for this formation is provided by the c. 3459 Ma age of the crosscutting North Pole Monzogranite.

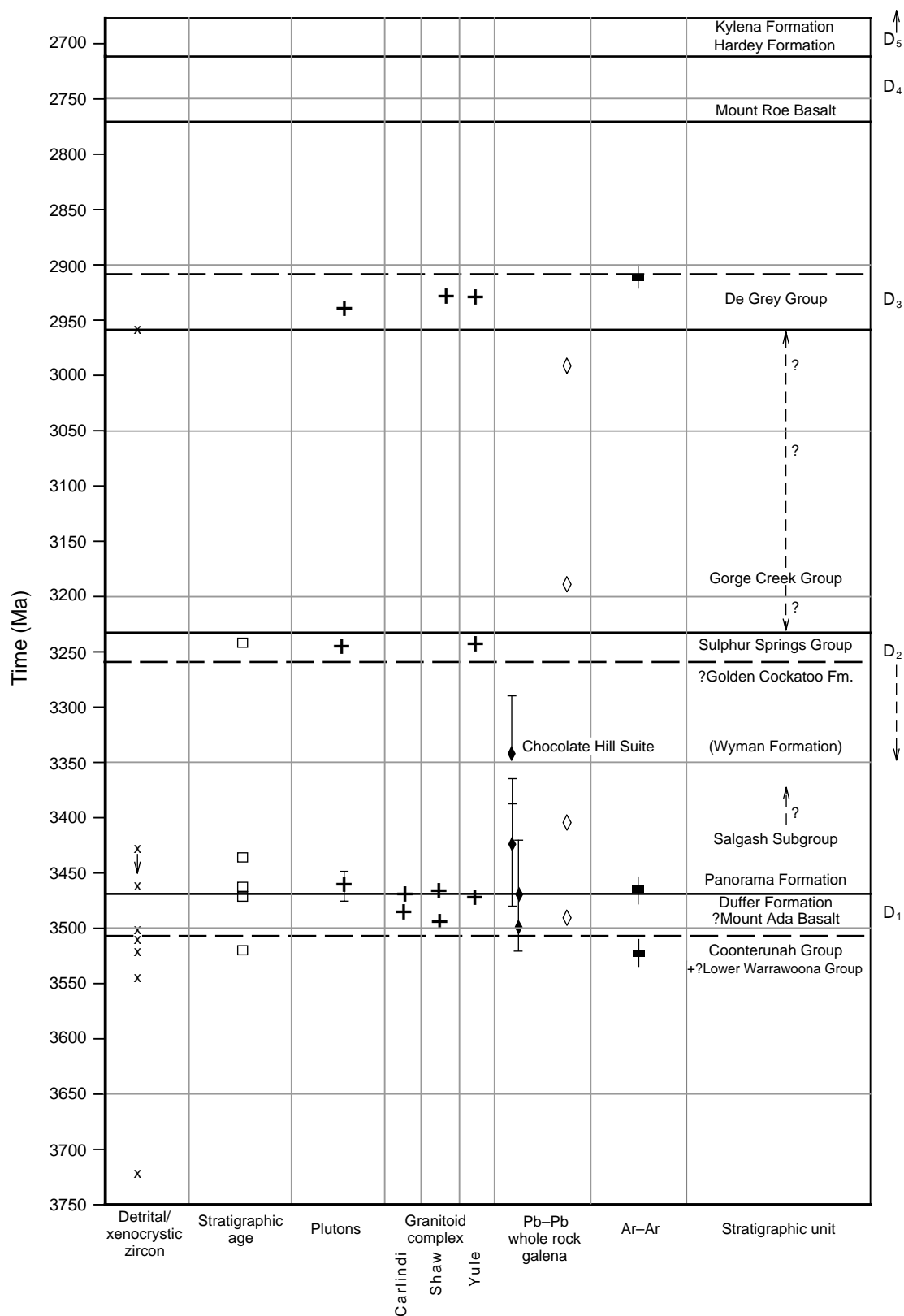
A precise age of  $3458 \pm 2$  Ma was obtained for a rhyolite from the Panorama Formation in the Panorama Belt (*AWpr*), on the southern flank of the North Pole Dome (Thorpe et al., 1992a). This sample contained a xenocrystic zircon dated at c. 3724 Ma, which is the oldest dated zircon in the Pilbara Craton. The Panorama Formation age is identical, within error, to the age of the North Pole Monzogranite ( $3459 \pm 18$  Ma; Thorpe et al., 1992a; sample 100510), indicating that the latter is a synvolcanic intrusion. The age of the overlying Euro Basalt is unconstrained, other than it must be younger than the c. 3458 Ma Panorama Formation.

The age of the Golden Cockatoo Formation is unknown, but must be older than c. 3240 Ma granitoid rocks of the Yule Granitoid Complex (*AgYka*), which intrude it (M. J. Van Kranendonk, unpublished U–Pb zircon data from WODGINA). The Sulphur Springs Group has been dated at between c.  $\leq 3255$ –3235 Ma by U–Pb SHRIMP zircon geochronology (Brauhart, 1999). The 3238–3235 Ma ages of volcanic rocks of the Kangaroo Caves Formation correspond with the  $3238 \pm 2$  Ma ages of the outer and inner phases of the synvolcanic Strelley Granite laccolith (samples 203366 and 203368 in Table 3; Brauhart, 1999). An identical zircon age has been confirmed for a dacite along strike, further south at the Cardinal gossan (c.  $3240 \pm 2$  Ma; sample 111868 in Table 3). Galena from the Cardinal gossan returned a Pb–Pb model age of c. 3260 Ma (Thorpe, R., Geological Survey of Canada, 1996, pers. comm.). Similar Pb–Pb model ages of c. 3260 Ma on galena have been obtained from the Sulphur Springs and Kangaroo Caves prospects (Vearncombe et al., 1995) and the Bernts prospect (Thorpe, R., Geological Survey of Canada, 1994, pers. comm.), whereas the Mad Hatters prospect returned a slightly older Pb–Pb model age of c. 3280 Ma (Morant, P., Sipa Resources, 1998, pers. comm.).



- |  |   |
|--|---|
| Low temperature metamorphism (anchizone) | Orthopyroxene-clinopyroxene relics in orthogneiss |
| Low greenschist to greenschist facies    | Granitoid rocks                                   |
| Amphibolite-facies greenstones           | Strike-slip fault                                 |
| Reported prehnite-pumpellyite facies     | Normal fault                                      |

Figure 36. Simplified sketch of the distribution of metamorphic facies on NORTH SHAW



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Figure 37. Geochronological data for NORTH SHAW showing clusters of ages at c. 3460 Ma, c. 3240 Ma, and c. 2950 Ma, which correspond with episodes of both magmatism and deformation

The age of the Gorge Creek Group is unknown, constrained only as being younger than the underlying Kangaroo Caves Formation (c. 3235 Ma) and older than the De Grey Group, which Van Kranendonk and Collins (1998) argued on the basis of structural data and correlations with existing geochronology, was c. 2950 Ma. The fact that the Gorge Creek Group fills in topographic depressions in the Sulphur Springs Group in the Soanesville Belt is used as evidence to suggest that the Gorge Creek Group is not much younger than the Sulphur Springs Group.

## Granitoid complexes and plutons

Several age dates have been obtained from the Shaw Granitoid Complex. The oldest age is a  $3499 \pm 22$  Ma Pb–Pb whole-rock isochron obtained for a suite of nine samples from the North Shaw Suite (Bickle et al., 1983). This age accords well with a U–Pb SHRIMP zircon age of  $3493 \pm 4$  Ma on an unspecified component of the North Shaw Suite (McNaughton et al., 1988; sample 91084). Individual components of the North Shaw Suite, dated by the Pb–Pb whole-rock isochron method, have given ages of  $3468 \pm 53$  Ma and  $3422 \pm 57$  Ma (Bickle et al., 1993). A Pb–Pb whole rock isochron of  $3338 \pm 52$  Ma from the Coolyia Creek Tonalite (*AgSco*) is interpreted to reflect a time metamorphic resetting rather than a true igneous age as previously suggested (Bickle et al., 1993), based on U–Pb zircon ages of  $3467 \pm 6$  Ma (McNaughton et al., 1988; sample 91085) and  $3469 \pm 2$  Ma (Nelson, in prep.; sample 142962) from this unit. A sample of leucocratic diatexite (*AgSl*; sample 142881) contains older zircon cores at c. 3450 Ma, overgrowths at c. 3410 Ma, and a third population of zircons at c. 3279 Ma (Nelson, D., GSWA, 1988, pers. comm.).

Samples of the Mulgandinnah Monzogranite (*AgSmu*) from Mulgandinnah Hill (AMG 456229) and just north of Pilga (AMG 498223) gave a combined Rb–Sr isochron age of  $2954 \pm 29$  Ma with an  $^{87}\text{Sr}/^{86}\text{Sr}_i = 0.70108 \pm 0.00073$  (Bickle et al., 1983). This is only slightly older than a U–Pb zircon age of  $2928 \pm 2$  Ma from a sample collected just west of Mulgandinnah Hill (sample 142882; Nelson, 1998).

Three samples of the Carlindi Granitoid Complex have been dated. The oldest is a fine- to medium-grained, undeformed monzogranite (*AgLm*) identical to the Mulgandinnah Monzogranite in the Shaw Granitoid Complex, but significantly older at  $3484 \pm 4$  Ma (sample 153188). A sample of biotite–hornblende leucogranite (*AgLl*) is  $3469 \pm 2$  Ma (sample 153190), and a quartz–feldspar porphyry (*AgLp*) interpreted to be a higher level equivalent of the leucogranite, is identical within error at  $3468 \pm 4$  Ma (Buick et al., 1995; sample 94058).

No samples of the Yule Granitoid Complex have been dated on NORTH SHAW, but immediately adjacent to the west on WODGINA, ages of c. 3470 Ma and 3240 Ma have been obtained on tonalitic orthogneiss (*AgYpe*) and homogeneous tonalite–granodiorite (*AgYka*) respectively (M. J. Van Kranendonk, unpublished U–Pb SHRIMP zircon data). Further south, on TAMBOURAH, Nelson (1998)

obtained an age of  $2927 \pm 3$  Ma on a sample of the Woodstock Monzogranite (*AgYwo*).

The small, discordant Keep It Dark Monzogranite (*Agki*) is exposed in the North Shaw Belt, outside of the granitoid complexes. A sample of weakly foliated granite (sample M95-327; Table 3) was found to contain abundant zircons comprising a single population of igneous-zoned, doubly terminated clear grains. Nine of the most concordant grains indicate a crystallization age of  $2936 \pm 5$  Ma, whereas a single grain is slightly older at  $2957 \pm 6$  Ma (M. J. Van Kranendonk, unpublished U–Pb SHRIMP zircon data).

## Metamorphic and mineralization events

The age of the D<sub>1</sub> folds in the Coonterunah Group must be older than the Strelley Pool Chert, and as this unit conformably overlies the Panorama Formation in the North Pole Dome, its age of 3458 Ma is interpreted as a minimum age for the deformation in the older rocks. An unpublished conventional U–Pb zircon age of  $3431.5 \pm 7.3$  Ma was obtained on the Breens Porphyry in the Panorama Belt (Thorpe, R., Geological Survey of Canada, 1992, pers. comm.), and this is interpreted to be the most likely time of associated mineralization in this area.

A metamorphic and magmatic event is interpreted to have occurred at c. 3400 Ma, based on the Pb–Pb galena date from the Normay mine and zircon overgrowths in diatexite of the Shaw Granitoid Complex (sample 142881). It is unknown if this event is related to regional thermotectonism or late Warrawoona Group magmatism, although the lack of evidence for regional deformation at this time supports the latter interpretation.

The Pb–Pb whole-rock isochron age of  $3338 \pm 52$  Ma is within error of known magmatic events further east in the Pilbara Craton, dated at between 3325 and 3304 Ma (Williams and Collins, 1990; Thorpe et al., 1992a; McNaughton et al., 1993). The presence of this age in the Shaw Granitoid Complex, together with evidence of zircon recrystallization at this time, is used to suggest that this was a time when doming occurred on NORTH SHAW (i.e. regional D<sub>2</sub> deformation).

The age of regional D<sub>3</sub> deformation was determined by Zegers et al. (1998) and Van Kranendonk and Collins (1998) as c. 2950 Ma, and this is reflected on NORTH SHAW by the c. 2930 Ma ages of the Keep It Dark, Woodstock, and Mulgandinnah Monzogranites. Final cooling of hornblende after this event occurred at c.  $2907 \pm 7$  Ma (sample 81-631; Table 3).

## Stratigraphy

NORTH SHAW is unique in containing six regional unconformities (see simplified geology insert on the NORTH SHAW map; and Van Kranendonk and Morant, 1998, fig. 1). The oldest, most high-angle unconformity

is between the Coonterunah Group and Strelley Pool Chert in the East Strelley Belt (AMG 268652 to 222639). This unconformity formed due to tilting of the Coonterunah Group by doming of the Carlindi Granitoid Complex at c. 3467 Ma (see **Structure**).

The known extent of the Coonterunah Group is restricted to within the East Strelley Belt, around the southeastern margin of the Carlindi Granitoid Complex, where it has been directly dated by Buick et al. (1995) at  $3515 \pm 3$  Ma. As described below, indirect geochronological data suggests that the Coonterunah Group may be more widespread, comprising the Mount Ada Basalt underlying the lower parts of the Panorama Belt in the North Pole Dome and North Shaw Belt. However, as direct igneous ages from these lower basaltic units have not yet been obtained, these rocks have been included as part of the Warrawoona Group until further information warrants change.

Within the lower Warrawoona Group, rocks previously interpreted to represent the Towers Formation have been renamed the Dresser Formation and placed lower down in the stratigraphy, as they lack the distinctive jasperlitic cherts of the Towers Formation in the type area of the Marble Bar Belt. This reclassification is supported by a 3490 Ma Pb–Pb age on galena from the Dresser Formation, which is significantly older than the Towers Formation in the Marble Bar Belt and is at most 3465 Ma. If the Dresser Formation lies within the lower part of the Warrawoona Group, then it may be possible that the felsic volcanoclastic Antarctic Creek Member of the Euro Basalt in the Panorama Belt may be a correlative of the Duffer Formation. U–Pb geochronology is being attempted in order to verify this supposition.

The second regional unconformity on NORTH SHAW developed at, or just before, c. 3240 Ma between the Leilira Formation of the Sulphur Springs Group and the underlying Euro Basalt of the Warrawoona Group (Van Kranendonk, 1997). This unconformity is preserved in the western part of the East Strelley Belt (AMG 080586), and in the southwestern extension of the Panorama Belt (AMG 288383).

A disconformity between the Gorge Creek and Sulphur Springs Groups locally developed into a low-angle erosional unconformity in the East Strelley Belt, where the Gorge Creek Group cuts out the Sulphur Springs Group (between AMG 105669 and 228634). Neptunian dykes of Corboy Formation sandstone locally intrude down into lithified marker chert at the top of the Sulphur Springs Group, indicating at least a minor time gap. However, the presence of relict volcanic topography in the underlying Sulphur Springs Group between Sulphur Springs and Kangaroo Caves, and the absence of evidence for any tectonic deformation in the Sulphur Springs Group prior to Gorge Creek Group deformation, suggests that there is no significant time gap between the deposition of these groups.

The fourth unconformity is between the Gorge Creek and De Grey Groups. This unconformity is particularly well developed at the base of the Lalla Rookh Synclinorium (between AMG 300623 and 330510), and in the

East Strelley Belt (between AMG 150554 and 183597). Local boulder-scrree deposits are preserved on the flanks of palaeotopographic highs (AMG 308622 and at AMG 303250). There are internal unconformities and disconformities within the Lalla Rookh Synclinorium, consistent with it having been deposited during D<sub>3</sub> deformation.

After the D<sub>3</sub> regional episode of transpressional deformation, the Mount Roe Basalt was deposited on an unconformity across the full spectrum of older rocks, indicating that the older rocks of the Pilbara Supergroup were already partly exposed by dome-and-syncline folding. However, parts of the Mount Roe Basalt are tightly folded within structures that had previously formed in the older rocks (e.g. the narrow, east–west septum in the eastern-central part of NORTH SHAW), indicating local, significant reactivation of pre-existing structures. After tight folding of the Mount Roe Basalt, rocks of the Hardey and Kylenea Formations were deposited across yet another unconformity that locally cuts down into the early Archaean basement.

As discussed in Hickman (1984) and Van Kranendonk and Collins (in prep.), the sequential deposition of supracrustal successions across progressively more shallowly dipping unconformities in between granitoid domes supports the influence of granitoid doming in the formation of the North Pilbara Terrain. The highest angle unconformities indicate prior periods of most significant doming, and these formed at c. 3460 Ma, >3240 Ma (probably at c. 3300 Ma), c. 2950 Ma, and c. 2765 Ma, as described in **Geological evolution**.

## Mineralization

### Volcanogenic massive sulfides (VMS)

The first discovery of VMS mineralization in the vicinity of the Strelley Granite was that of the Cardinal gossan in the early 1970s, from which the best intersection was 10 m of 0.9 wt% Cu, 5.9 wt% Zn, 0.4 wt% Pb, and 36 ppm Ag. H. Wilhelmij discovered magnesium-sulfate precipitates in the Sulphur Springs Creek in 1984, and this led to the subsequent discovery of the Sulphur Springs gossan by prospector D. O'Meara. Since then, intensive mineral exploration has been undertaken by Sipa Resources Ltd of Perth, in joint venture first with Ashling Resources NL (1989–1992), and subsequently with the Outokumpu Group. Exploration has primarily involved detailed mapping of the Soanesville Belt and adjacent areas, during the course of which several more Zn–Cu gossans were found. Succinct overviews of the geology of the prospects and exploration results are presented in Morant (1995, 1998). A more thorough treatment is presented by Ferguson (1999). Vearncombe et al. (1995, 1998) described the structural and tectonic settings of mineralization. Geochemical studies of the Kangaroo Caves Formation and Strelley Granite are presented in Vearncombe (1996), together with detailed petrography of ore textures. Hill (1997) studied the hangingwall alteration

above the Sulphur Springs gossan. Brauhart et al. (1998) and Brauhart (1999) described the regional alteration system associated with mineralization in, and above, the Strelley Granite, and provided additional geochemical data on the Kangaroo Caves Formation and Strelley Granite.

Five of the main Zn–Cu VMS prospects are near the top of the Sulphur Springs Group, spaced at regular 5–7 km intervals around the Strelley Granite. A sixth prospect, the Bernts prospect, forms at the same stratigraphic level but within the fault-bounded Bernts Deformation Zone of folded, sheared, and brecciated rocks located between the Soanesville and Panorama Belts. The Roadmaster, Cardinal, and Jamesons prospects are about 1 km stratigraphically below the other prospects.

Since 1989, 50 000 m of diamond and reverse circulation drilling by Sipa Resources Ltd has identified the following indicated and inferred resource estimates (from Morant, 1998): Sulphur Springs — 2.8 Mt at 10.7% Zn and 0.6% Cu, and 2.5 Mt at 1.1% Zn and 4.0% Cu; Kangaroo Caves — 1.7 Mt at 9.8% Zn and 0.6% Cu; and Bernts — 0.6 Mt at 7.8% Zn and 0.3% Cu.

The Zn–Cu mineralization at Sulphur Springs is hosted within a 200 m-thick dacite sill immediately beneath the marker chert, within the marker chert itself, and rarely in the hangingwall rhyodacite (Morant, 1998). The mineralization is zoned from Zn-rich within the marker chert, to Cu-rich beneath the marker chert. Mineralization at Kangaroo Caves is hosted within the marker chert, projecting down-plunge to the northeast for at least 1.5 km.

The sulfide assemblage at Sulphur Springs and Kangaroo Caves comprises pyrite, low-Fe sphalerite, chalcopyrite, and galena, with minor arsenopyrite and tennantite–arsenopyrite (Vearncombe, 1996; Morant, 1998). Well-preserved ore textures described by Vearncombe et al. (1995, 1998) and Vearncombe (1996) include dendritic, colloform, and botryoidal types that may have formed by open space precipitation of sulfides. Vearncombe et al. (1995) considered that the development of some delicate sulfide textures, such as spherical pellets and stromatolitic types, were analogous to black-smoker sulfide chimneys at present submarine hydrothermal vents.

The Zn–Cu mineralization at the five main prospects around the Strelley Granite is spatially related to growth faults emanating from the granite (Vearncombe et al., 1998). Mineralized prospects are surrounded by a zone of related feldspar-destructive chlorite–quartz alteration (Brauhart et al., 1998). Further away from the mineralized prospects, sericite–quartz(–carbonate) and albite–carbonate(–pyrite) alteration facies have been recognized. Sulfide mineralization forms in veins with greisen-altered margins in the outer phase of the Strelley Granite (*Agste*), and porphyry molybdenum in greisen has been observed locally near the top of the inner phase (*Agstp*) near the Greisen prospect (AMG 275530) (Folkert, S., University of Western Australia, 1998, pers. comm.). These data indicate that intrusion of the Strelley Granite caused extensive hydrothermal circulation and deposition of the associated mineralization (Vearncombe et al., 1998). The source of the mineralization has yet to be determined and

is the topic of a current PhD thesis by S. Folkert at the University of Western Australia. Geochronological and geochemical similarities between the Strelley Granite and felsic volcanic rocks of the Kangaroo Caves Formation (Vearncombe, 1996; Brauhart, 1999) indicate that the Strelley Granite was a synvolcanic laccolith during eruption of the Kangaroo Caves Formation. Such an interrelationship between volcanism, granite laccolith intrusion, and massive sulfide mineralization is characteristic of Archaean to Holocene VMS deposits (Franklin et al., 1981), although the Strelley Granite has slightly unusual geochemistry, as discussed below.

A model for the magmatic and structural development of the Kangaroo Caves Formation, Strelley Granite, and VMS mineralization is presented in Figure 38. In this model, eruption of the basal volcanic sequence was followed by intrusion of felsic subvolcanic sills near the top of the volcanic pile beneath a carapace of volcanoclastic and epiclastic sediment. Intrusion of the outer phase of the Strelley Granite as a sill only 1.5 km below the surface–water interface initiated extensive hydrothermal circulation, which resulted in leaching of metals from the volcanic sequence (Brauhart, 1999) and silicification of the sedimentary carapace to form an impermeable cap. This was followed by intrusion of the inner phase of the Strelley Granite (*Agstp*) and comagmatic dolerite (*Asd*), causing asymmetrical inflation of the sill into a sphenolith. Inflation was accommodated by the development of growth faults in the overlying volcanic sequence (Vearncombe et al., 1998), and of mineralized extension veins in the outer rind of the granite. As the extension veins and growth faults are both mineralized, it is thought that the bulk of mineralization in the main prospects beneath the marker chert was deposited at this time. Ore textures indicate that sulfide minerals were deposited as open-space precipitation, mainly within the marker chert, but some hydrothermal fluids may have escaped onto the surface, forming analogues of present-day sulfide chimneys (Vearncombe et al., 1995, 1998).

## Barite

Several million tonnes of barite were deposited in the Barite Range on the eastern flank of the North Pole Dome (Hickman, 1973, 1983). Between 1970 and 1990, 129 505 t of barite was mined at the Dresser Mining Centre (AMG 527590), which covered leases with total published reserves of 500 000 t (Australian Mining, 1983; Sargeant and Sampson, 1980). A review of the deposits and their origin is presented in Abeysinghe and Fetherston (1997).

The main deposits formed as beds and mounds for 8 km along strike within the Dresser Formation (*AWr*), and within coarse crystalline veins together with black chert, which intrude immediately underlying host rocks of the Mount Ada Basalt (*AWmb*). Barite is present as pebbles in basal diamictite layers, finely laminated rock containing stromatoloid laminae (Buick et al., 1981), coarsely crystalline mounds adjacent to syndimentary growth faults (Nijman et al., 1998), and coarsely bladed veins together with black chert (*AWrct*). The main deposits

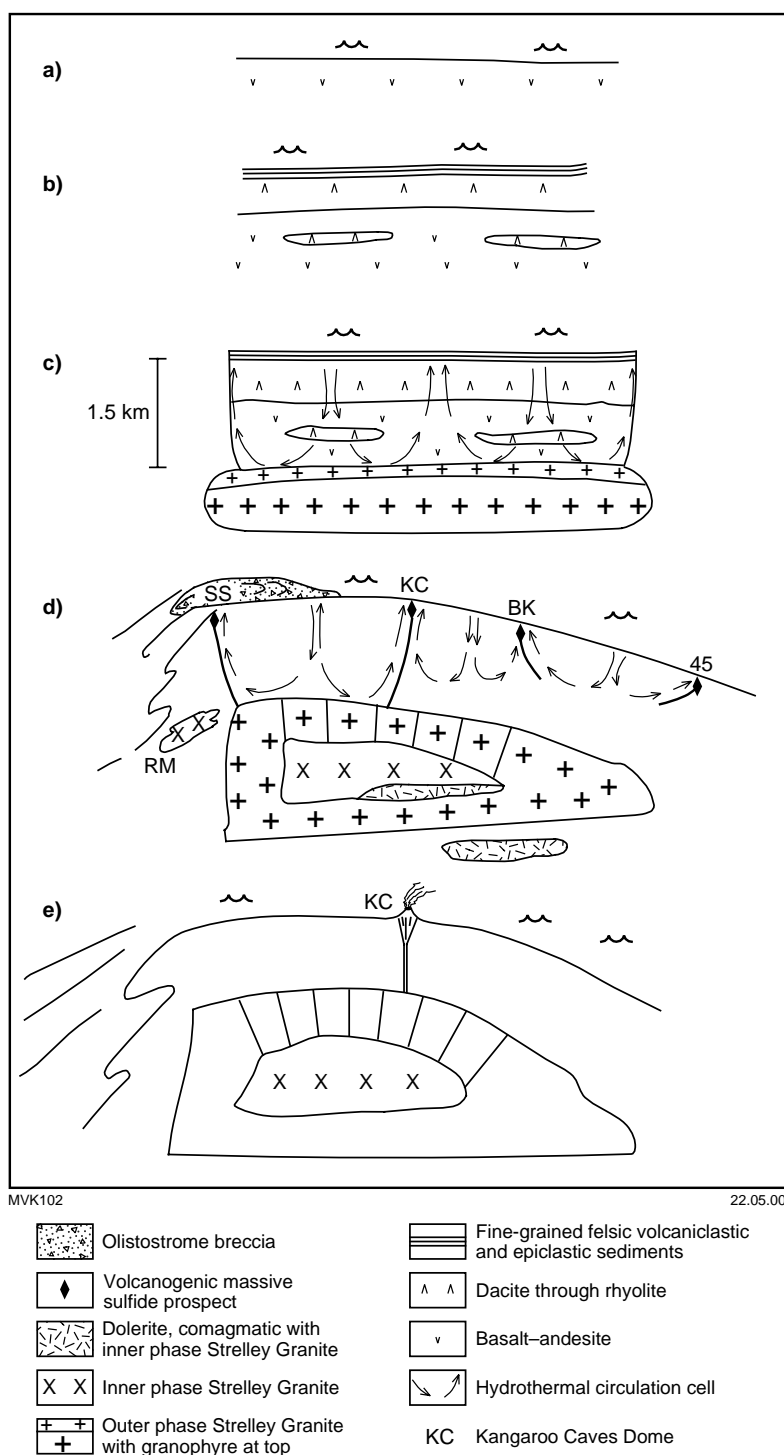


Figure 38. Schematic geological evolution of the Strelley mineralizing system (in part after Vearncombe, 1996; Morant, 1998; Vearncombe et al., 1998; Brauhart, 1999). Abbreviations: BK = Breakers prospect; KC = Kangaroo Caves prospect; RM = Roadmaster prospect; SS = Sulphur Springs prospect; 45 = Anomaly 45 prospect: a) deposition of the basal basalt-andesite volcanic suite; b) deposition of the upper volcanic suite, including further basalt-andesite volcanism, dacitic to rhyolitic volcanic rocks, and deposition of a fine-grained felsic volcaniclastic cover; c) intrusion of the outer phase of the Strelley Granite with differentiation into an outer rind of granophyre. Intrusion drives broadly distributed hydrothermal circulation and consequent leaching of metals in the overlying volcanic rocks. Hydrothermal circulation causes silicification of the epiclastic cover at the sediment-water interface; d) intrusion of the inner phase of the Strelley Granite, including granite and comagmatic dolerite. Intrusion causes asymmetrical ballooning of the laccolith and the intrusion of greisen veins in joints in the outer phase of the granite. Felsic magmas escape from the northern apex of the asymmetrical granitoid laccolith and intrude as sills beneath, and to the west of, Sulphur Springs, as well as in the Roadmaster area. Continued hydrothermal circulation is concentrated into cells whose output zones in growth faults become mineralized. Late uplift of the magmatic apex caused collapse of the Sulphur Springs area, with deposition of the synvolcanic olistostrome breccia and formation of ductile-brittle cascade folds in the Roadmaster area; e) late bulging of the central part of the laccolith caused eruption of the Kangaroo Caves dome

formed where the Dresser Formation is thickest and consists of three stratigraphic horizons of chert–barite interlayered with basalt. The three beds are bound by an array of listric normal faults, the main set of which passes just south of, and through, the Dresser Mining Centre. The geometry of the faults and their close relation to barite mineralization (Nijman et al., 1998) indicates that the barite was deposited during normal faulting and development of a small graben. Galena and sphalerite are finely disseminated through the system, suggesting the possibility of massive sulfides at depth. Sulfide mineralization was explored for in the underlying stratigraphy, but to no avail (Sargeant, 1979).

Hickman (1973) ascribed a sedimentary origin for the North Pole barite deposits and presented several points against a hydrothermal origin for the barite. The presence of barite in chert veins was interpreted to be the result of tectonic re-activation of sedimentary chert–barite during structural doming. Dunlop et al. (1978) and Sargeant (1979) suggested that the barite deposits formed by diagenetic replacement of an original gypsum–carbonate sequence; gypsum pseudomorphs in barite were reported by Lambert et al. (1978) and Buick and Dunlop (1990). Barite crystals can be clearly seen to have grown in semilithified rocks during diagenesis by the fact that they forcefully displace bedding (Fig. 7a).

Sargeant (1979) inferred that the deposits formed by exhalative chemical precipitation. In this model, the chert–barite veins were interpreted to represent the footwall vein stockwork system. A syngenetic, volcanogenic exhalative model was further supported by the detailed mapping studies of Nijman et al. (1998), who documented variations in both lithofacies type and thicknesses of barite and chert adjacent to growth faults, indicating a contemporaneous development. These authors estimated a water depth of less than 50 m, which supports the contention of Barley and Groves (1984) that volcanogenic massive sulfides would not form in this area because water depth was too shallow for boiling. The environment of deposition was interpreted as a shallow and probably restricted, inshore marine setting, with deposition of barite and diagenetic replacement of sulfate–carbonate rocks by barite during low-temperature, hydrothermal volcanic emissions (Pirajno, 1992, fig. 11.21).

In 1997, Haoma Mining NL discussed mining the barite from the Dresser mine with potential partners, but no further activity has developed.

## Epigenetic gold

Epigenetic gold was mined near the turn of the century in the North Shaw, North Pole, and Lalla Rookh Mining Centres. The North Shaw Mining Centre is host to six mineralized reefs in carbonate-altered metabasalts and subordinate metagabbro of the North Shaw Belt (*AWmb*). The mineralization is hosted by sulfide-rich quartz veins striking north and east within shear zones. From 1897 to 1937, 400.52 oz of gold was mined from 548 t of ore treated, with the majority coming from the North Pole Democrat mine (229 oz) in 1899–1900. Prior to

1973, the total mined was 4.23 kg from alluvials, 18.04 kg doliied, and 36.63 kg mined. In 1989, a further 4.5 kg was obtained from alluvials and 130.78 kg was mined, for a total historical production of 214.12 kg.

In recent times, interest was sparked by encouraging results from regional soil sampling and initial RAB drilling in 1995 and 1996. In 1997, Gold Partners NL farmed out the property to Geographe Resources Ltd, who conducted further soil sampling and RAB drilling. In May 1998, Haoma Mining NL bought the tenements and stockpiled ore from Gold Partners NL, but no further activity has been noted since.

The North Pole Mining Centre consists of five historical prospects from which 400.52 oz of gold were mined from 1899 to 1931. The prospects are in some of the most rugged terrain on NORTH SHAW, in sheared and altered metabasalts on either side of a chert bar. The Breens Reward mine is adjacent to a  $3432 \pm 7$  Ma felsic porphyry (see **Geochronology**, p. 73), which is interpreted to have been the source for the Au, Cu, Ag, and Bi mineralization.

The Normay mine, located in sheared metabasalt near the northern contact of the North Pole Monzogranite in the North Pole Dome, was operational until the early 1990s and placed under care and maintenance in 1994. In 1996, it was abandoned. Gold and silver mineralization are in quartz veins together with minor pyrite and copper–lead–zinc sulfides at depth within a steeply dipping, narrow shear that extends for up to 1 km along strike to the east, until it is cut out by a gabbro dyke associated with the Fortescue Group (*Afd*). The mine was instigated in 1940–41, when 70.1 t of ore were treated to retrieve 0.966 kg of gold at an average grade of 13.79 g/t Au. In 1949–1951, a further 3781.55 t of ore were treated for 46.62 kg Au and 8.089 kg Ag, at an average grade of 12.33 g/t Au. From 1952 to 1957, 44.664 kg Au and 54.59 kg Ag were extracted at the high grade of 54.59 g/t Au. The mine was re-opened in 1971–72, when 4.54 kg of gold were extracted from 679.1 t of ore, but by far the largest amount was mined between 1988 and 1995 when 809 kg of gold were produced.

Mineralization at the Lalla Rookh Mining Centre is hosted by chlorite–carbonate schist within a panel of relatively undeformed volcanic rocks (*Aot*). The ore lies within quartz veins emplaced into the hinges ( $\leq 15$  m thick, saddle reef) and limbs of tight, easterly trending, upright folds. Whereas the quartz veins in fold hinges are up to 15 m thick, those in the fold limbs are up to 5 m thick, suggesting that mineralization was controlled by saddle-reef structures formed during folding. Between 1901 and 1925, a total of 4175.59 oz of gold was produced, the majority of which was mined between 1904 and 1906 and 1919 and 1925. By 1967, a total of 12 603 oz or 392 kg of gold had been mined. Since then, only a further 7.057 kg have been mined. The leases are currently owned by Haoma Mining NL and there has been recent activity in 1997–98. Total measured mineral resources at the mine were re-evaluated in 1996–97, comprising 22 000 t at 2.5 g/t Au at North Reef, 199 000 t at 1.9 g/t Au at South Reef, and 47 000 t at 2.81 g/t Au at South Reef Shoot, for

a total of 268 000 t of gold at an average grade of 2.11 g/t Au for 18 170 oz. Haoma Mining NL said that potential remained to expand the South Reef resource at depth, as it was found to plunge 14°W.

Lynas Gold NL identified a shallow inferred mineral resource of 60 000 t of gold at 3 g/t Au at the Mercury Hill prospect within highly sheared and altered fuchsite schists along the contact between ultramafic and felsic volcanic rocks in the Numerous Scrapes Deformation Zone. Visible gold was recorded in numerous inter-sections, which had grades of up to 7.6 g/t Au, and on surface exposures of narrow, dark-grey quartz veins. Exploitation of this resource in 1997 was uneconomic given a haulage distance of 25 km to the mill at the Iron Stirrup mine. Given that the mill has since been removed to Paraburdoo, it is unlikely that this resource will be mined in the near future.

## Tin

Mining of alluvial tin in the vicinity of the Pilga Homestead (AMG 500215) followed its discovery in 1890 (Simpson, 1948). Mining continued through the 1970s, but has since been abandoned. The bulk of tin mineralization is in the form of cassiterite, and was extracted from the beds of creeks in and around the Cooglegong Mining Centre to a depth of 1.5 – 2.5 m. The origin of the mineralization is from albite pegmatites, which intrude host granitoid gneisses around the margin of the post-tectonic Cooglegong Monzogranite, located immediately south of the map area on TAMBOURAH, and dated as  $2851 \pm 2$  Ma (Nelson, 1998). Reserves and grades of tin mineralization in the Shaw River tin fields are discussed by Blockley (1980).

## Geological evolution

The geological evolution of NORTH SHAW is long and complex. It may be divided into five main geotectonic periods, based on geochronological data and the structural geology outlined above. The five sets of structures identified across NORTH SHAW are interpreted to have formed through two different tectonic styles. One dominant style is thermally driven gravitational inversion — or partial convective overturn (Collins et al., 1998; Collins and Van Kranendonk, 1999; Van Kranendonk and

Collins, in prep.) — of an upper crustal layer of relatively dense supracrustal rocks and an underlying, middle crustal layer of more buoyant granitoid rocks. Deformation events driven by this tectonic style are interpreted to include D<sub>1</sub>, D<sub>2</sub>, D<sub>4</sub>, and D<sub>5</sub>. In contrast, the geometry of structures formed during D<sub>3</sub> indicates formation due to northwest–southeast horizontal compression (Van Kranendonk and Collins, 1998). A detailed review of the geological evolution of NORTH SHAW is outside the scope of these notes and is presented in Van Kranendonk (in prep.b).

Evidence in support of granitoid doming includes the fact that all six groups of supracrustal rocks have conformable to disconformable relationships with older groups, or intrusive lower contacts with granitoid rocks, and are thus entirely (par-) autochthonous. This negates, for this area at least, an origin through accretionary processes as proposed by Krapez (1993) and Barley (1997). Successively younger groups of supracrustal rocks are deposited in progressively tighter synclines developed between domal granitoid complexes and laccolith-cored domes. The groups are preserved as wedge-shaped packages with decreasing dips of their respective basal unconformities up stratigraphic section, away from granitoid-cored domes. In opposite, but parallel, corollary, granitoid rocks in the Shaw Granitoid Complex (and probably the Carlindi and Yule Granitoid Complexes at a scale larger than that covered by NORTH SHAW) become progressively younger down structural section into the core of the dome, indicating progressive intrusion of subsequent generations of magmas into a progressively amplifying dome.

Parallel ages of granitoid and volcanic magmatism, and the coeval development of progressively deepening synclines and uplifting granitoid domes combine to indicate a linked, parallel evolution as originally proposed by Hickman (1975, 1984). Collins et al. (1998) discussed the driving force behind the generation of such linked dome-and-basin structures across the ancient nucleus of the eastern Pilbara Craton, and concluded that it was due to thermally driven, partial convective overturn of a density-inverted, greenstone-over-granitoid crust, and that it evolved through a punctuated history related to specific episodic mantle-driven thermotectonic events.

## References

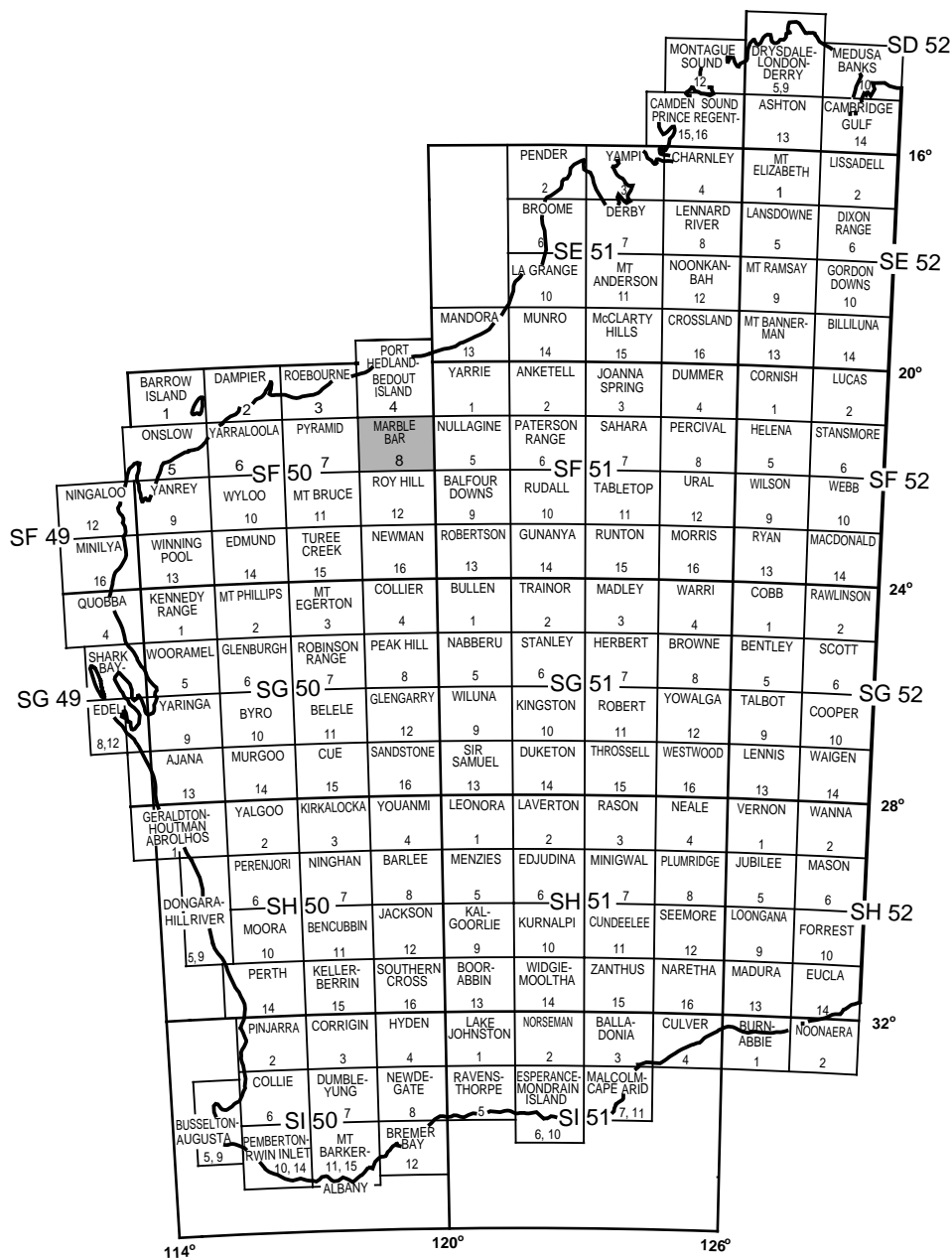
- ABEYSINGHE, P. B., and FETHERSTON, J. M., 1997, Barite and fluorite in Western Australia: Western Australia Geological Survey, Mineral Resources Bulletin 17, 97p.
- ARNDT, N. T., NELSON, D. R., COMPSTON, W., TRENDALL, A. F., and THORNE, A. M., 1991, The age of the Fortescue Group, Hamersley Basin, Western Australia, from ion microprobe zircon U–Pb results: *Australian Journal of Earth Sciences*, v. 38, p. 261–281.
- AUSTRALIAN MINING, 1983, The mineral, baryte: Australian Mining, p. 62.
- AWRAMIK, S. M., SCHOPF, J. W., and WALTER, M. R., 1983, Filamentous fossil bacteria from the Archean of Western Australia: *Precambrian Research*, v. 20, p. 357–374.
- BARLEY, M. E., 1993, Volcanic, sedimentary and tectonostratigraphic environments of the ~3.46 Ga Warrawoona Megasequence — a review: *Precambrian Research*, v. 60, p. 47–67.
- BARLEY, M. E., 1997, The Pilbara Craton, in *Greenstone belts edited by M. J. De WITT and L. ASHWAL*: Oxford University Monographs, Geology and Geophysics, no. 35, p. 657–664.
- BARLEY, M. E., and GROVES, D. I., 1984, Constraints on mineralization of the Warrawoona Group, in *Archaean and Proterozoic basins of the Pilbara, Western Australia — evolution and mineralization potential edited by J. MUHLING, D. I. GROVES, and T. BLAKE*: University of Western Australia, Geology Department and University Extension, Publication no. 9, p. 54–67.
- BARLEY, M. E., DUNLOP, J. S. E., GLOVER, J. E., and GROVES, D. I., 1979, Sedimentary evidence for an Archaean shallow-water volcanic-sedimentary facies, eastern Pilbara Block, Western Australia: *Earth and Planetary Science Letters*, v. 43, p. 74–84.
- BARLEY, M. E., SYLVESTER, G. C., GROVES, D. I., BORLEY, G. D., and ROGERS, N., 1984, Archaean calc-alkaline volcanism in the Pilbara Block, Western Australia: *Precambrian Research*, v. 24, p. 285–319.
- BARNES, K. R., 1983, The geology of part of the Strelley Pool Chert, Pilgangoora Syncline, Pilbara Block, Western Australia: University of Western Australia, BSc Honours thesis (unpublished).
- BERTHÉ, D., CHOUKROUNE, P., and GAPAIS, D., 1979, Orientations préférentielles du quartz et orthogneissification progressive en régime cisailant: l'exemple du cisaillement sud-américain: *Bulletin de Minéralogie*, v. 102, p. 265–272.
- BETTENAY, L. F., BICKLE, M. J., BOULTER, C. A., GROVES, D. I., MORANT, P., BLAKE, T. S., and JAMES, B. A., 1981, Evolution of the Shaw Batholith — an Archaean granitoid–gneiss dome in the eastern Pilbara, Western Australia, in *Archaean Geology edited by J. E. GLOVER and D. I. GROVES*: Geological Society of Australia, 2nd International Archaean Symposium, Perth, W.A., 1980, Proceedings, Special Publication, no. 7, p. 361–372.
- 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*, v. 8, p. 525–529.
- BICKLE, M. J., BETTENAY, L. F., BARLEY, M. E., CHAPMAN, H. J., GROVES, D. I., CAMPBELL, I. H., and de LAETER, J. R., 1983, A 3500 Ma plutonic and volcanic calc-alkaline province in the Archaean east Pilbara Block: *Contributions to Mineralogy and Petrology*, v. 84, p. 25–35.
- 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*, v. 101, p. 361–376.
- 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 geochemical constraints from the Shaw Batholith: *Precambrian Research*, v. 60, p. 117–149.
- BICKLE, M. J., MORANT, P., BETTENAY, L. F., 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 sequences edited by L. D. AYERS, P. C. THURSTON, K. D. CARD, and W. WEBER*: Geological Association of Canada, Special Paper 28, p. 325–341.
- BLAKE, T. S., 1993, Late Archaean crustal extension, sedimentary basin formation, flood basalt volcanism and continental rifting: the Nullagine and Mount Jope Supersequences, Western Australia: *Precambrian Research*, v. 60, p. 185–241.
- BLAKE, T. S., and BARLEY, M. E., 1992, Tectonic evolution of the late Archaean to Early Proterozoic Mount Bruce Megasequence Set, Western Australia: *Tectonics*, v. 11, p. 1415–1425.
- BLOCKLEY, J. G., 1980, Tin deposits of Western Australia with specific reference to the associated granites: *Western Australia Geological Survey, Mineral Resources Bulletin* 12, 184p.
- BOULTER, C. A., BICKLE, M. J., GIBSON, B., and WRIGHT, R. K., 1987, Horizontal tectonics pre-dating upper Gorge Creek Group sedimentation, Pilbara Block, Western Australia: *Precambrian Research*, v. 36, p. 241–258.
- BRAUHART, C., 1999, Regional alteration systems associated with Archaean volcanogenic massive sulphide deposits at Panorama, Pilbara, Western Australia: University of Western Australia, PhD thesis (unpublished).
- BRAUHART, C., GROVES, D. I., and MORANT, P., 1998, Regional alteration systems associated with volcanogenic massive sulfide mineralization at Panorama, Pilbara, Western Australia: *Economic Geology*, v. 93, p. 292–302.
- BUICK, R., and DUNLOP, J. S. R., 1990, Evaporitic sediments of early Archaean age from the Warrawoona Group, North Pole, Western Australia: *Sedimentology*, v. 37, p. 247–277.
- BUICK, R., DUNLOP, J. S. R., and GROVES, D. I., 1981, Stromatolite recognition in ancient rocks: an appraisal of irregular laminated structures in an early Archaean chert–barite unit from North Pole, Western Australia: *Alcheringa*, v. 5, p. 161–181.
- BUICK, R., THORNETT, J. R., McNAUGHTON, N. J., SMITH, J. B., BARLEY, M. E., and SAVAGE, M., 1995, Record of emergent

- continental crust ~3.5 billion years ago in the Pilbara Craton of Australia: *Nature*, v. 375, p. 574–577.
- CHAN, M., 1998, Sedimentology and structure of three small conglomeratic basins in the Lalla Rookh – Western Shaw structural corridor, Pilbara Craton, Western Australia: University of Western Australia, BSc Honours thesis (unpublished).
- COLLERSON, K. D., and McCULLOCH, M. T., 1983, Field and isotopic evidence on Archaean crust and mantle evolution in the east Pilbara Block, Western Australia: Geological Society of Australia, Abstract Series 9, p. 167–168.
- COLLINS, W. J., 1989, Polydiapirism of the Archaean Mount Edgar Batholith, Pilbara Block, Western Australia: *Precambrian Research*, v. 43, p. 41–62.
- COLLINS, W. J., 1993, Melting of Archaean sialic crust under high  $a_{H_2O}$  conditions: genesis of 3300 Ma Na-rich granitoids in the Mount Edgar Batholith, Pilbara Block, Western Australia: *Precambrian Research*, v. 60, p. 151–174.
- COLLINS, W. J., and GRAY, C. M., 1990, Rb–Sr isotopic systematics of an Archaean granite–gneiss terrain: The Mount Edgar batholith, Pilbara Block, Western Australia: *Australian Journal of Earth Sciences*, v. 37, p. 9–22.
- COLLINS, W. J., and VAN KRAENDONK, M. J., 1999, Model for the development of kyanite during partial convective overturn of Archaean granite–greenstone terranes: the Pilbara Craton, Australia: *Journal of Metamorphic Geology*, v. 17, no. 2, p. 145–156.
- COLLINS, W. J., VAN KRAENDONK, M. J., and TEYSSIER, C., 1998, Partial convective overturn of Archaean crust in the east Pilbara Craton, Western Australia: driving mechanisms and tectonic implications: *Journal of Structural Geology*, v. 20, p. 1405–1424.
- CRAIG, J. R., and VAUGHAN, D. J., 1981, Ore microscopy and ore petrography: New York, John Wiley and Sons, 406p.
- CULLERS, R. L., DiMARCO, M. J., LOWE, D. R., and STONE, J., 1993, Geochemistry of a silicified, felsic volcanoclastic suite from the early Archaean Panorama Formation, Pilbara Block, Western Australia — an evaluation of depositional and post-depositional processes with special emphasis on the rare-earth elements: *Precambrian Research*, v. 60, p. 99–116.
- DAVIDS, C., WIJBRANS, J. R., and WHITE, S. H., 1997,  $^{40}\text{Ar}/^{39}\text{Ar}$  laserprobe ages of metamorphic hornblendes from the Coongan Belt, Pilbara, Western Australia: *Precambrian Research*, v. 83, p. 221–242.
- DAVIS, G. H., 1984, Structural geology of rocks and regions: New York, John Wiley and Sons, 492p.
- DAWES, P. R., SMITHIES, R. H., CENTOFANI, J., and PODMORE, D. C., 1995, Sunrise Hill unconformity: a newly discovered regional hiatus between Archaean granites and greenstones in the northeastern Pilbara Craton: *Australian Journal of Earth Sciences*, v. 42, p. 635–639.
- DiMARCO, M. J., and LOWE, D. R., 1989a, Stratigraphy and sedimentology of an Early Archaean felsic volcanic sequence, eastern Pilbara Block, Western Australia, with special reference to the Duffer Formation and implications for crustal evolution: *Precambrian Research*, v. 44, p. 147–169.
- DiMARCO, M. J., and LOWE, D. R., 1989b, Shallow water volcanoclastic deposition in the Early Archaean Panorama Formation, eastern Pilbara Block, Western Australia: *Sedimentary Geology*, v. 64, p. 43–63.
- DiMARCO, M. J., and LOWE, D. R., 1989c, Petrography and provenance of silicified Early Archaean volcanoclastic sandstones, eastern Pilbara Block, Western Australia: *Sedimentology*, v. 36, p. 821–836.
- DIXON, J. M., and SUMMERS, J. M., 1983, Patterns of total and incremental strain in subsiding troughs: experimental centrifuged models of inter-diapir synclines: *Canadian Journal of Earth Sciences*, v. 20, p. 1843–1861.
- DUNLOP, J. S. R., MUIR, M. D., MILNE, V. A., and GROVES, D. I., 1978, A new microfossil assemblage from the Archaean of Western Australia: *Nature*, v. 274, p. 676–678.
- ERIKSSON, K. A., 1981, Archaean platform-to-trough sedimentation in the east Pilbara Block, Australia, in *Archaean Geology edited by J. E. GLOVER and D. I. GROVES*: Geological Society of Australia; 2nd International Archaean Symposium, Perth, W.A., 1980, Proceedings, Special Publication, no. 7, p. 235–244.
- ERIKSSON, K. A., KRAPEZ, B., and FRALIK, P., 1994, Archaean sedimentation: *Earth Science Reviews*, v. 37, p. 1–88.
- FERGUSON, K., 1999, Lead, zinc and silver deposits of Western Australia: Western Australia Geological Survey, Mineral Resources Bulletin 15, 314p.
- FITTON, M. J., HORWITZ, R. C., and SYLVESTER, G., 1975, Stratigraphy of the Early Precambrian in the West Pilbara, Western Australia: Australia CSIRO, Division of Mineralogy, Minerals Research Laboratories, Report FP 11.
- FRANKLIN, J. M., LYDON, J. W., and SANGSTER, D. F., 1981, Volcanic-associated massive sulfide deposits, in *Economic Geology — seventy fifth anniversary volume: 1905–1980 edited by B. J. SKINNER*: Texas, Economic Geology Publishing Company, p. 485–628.
- GLIKSON, A. Y., 1997, North Shaw — Landsat-5-TM Image, (1:100 000 scale map): Australian Geological Survey Organisation.
- GLIKSON, A. Y., DAVY, R., and HICKMAN, A. H., 1986, Geochemical data files of Archaean volcanic rocks, Pilbara Block, Western Australia: Australia Bureau of Mineral Resources, Record 86/14.
- GLIKSON, A. Y., DAVY, R., and HICKMAN, A. H., 1991, Trace metal distribution patterns in basalts, Pilbara Craton, Western Australia, with stratigraphic–geochemical implications: Australia Bureau of Mineral Resources, Record 1991/46, 37 p.
- GLIKSON, A. Y., and HICKMAN, A. H., 1981a, Chemical stratigraphy and petrogenesis of Archaean basic–ultrabasic volcanic units, eastern Pilbara Block, Western Australia in *Archaean Geology edited by J. E. GLOVER and D. I. GROVES*: Geological Society of Australia; 2nd International Archaean Symposium, Perth, W.A., 1980, Proceedings, Special Publication, no. 7, p. 287–300.
- GLIKSON, A. Y., and HICKMAN, A. H., 1981b, Geochemistry of Archaean volcanic successions, eastern Pilbara Block, Western Australia: Australia Bureau of Mineral Resources, Record 1981/36, 83p.
- GRIFFIN, T. J., 1990, North Pilbara granite–greenstone terrane, in *Geology and mineral resources of Western Australia: Western Australia Geological Survey, Memoir 3*, p. 128–158.
- GROVES, D. I., DUNLOP, J. S. R., and BUICK, R., 1981, An early habitat of life: *Scientific American*, v. 245, p. 64–73.
- HICKMAN, A. H., 1973, The North Pole barite deposits, Pilbara Goldfield: Western Australia Geological Survey, Annual Report 1972, p. 57–60.
- HICKMAN, A. H., 1975, Precambrian structural geology of part of the Pilbara region: Western Australia Geological Survey, Annual Report 1974, p. 68–73.
- HICKMAN, A. H., 1977, New and revised definitions of rock units in the Warrawoona Group, Pilbara Block: Western Australia Geological Survey, Annual Report 1977, p. 53.
- HICKMAN, A. H., 1983, Geology of the Pilbara Block and its environs: Western Australia Geological Survey, Bulletin 127, 268p.

- HICKMAN, A. H., 1984, Archaean diapirism in the Pilbara Block, Western Australia, in *Precambrian tectonics illustrated* edited by A. KRÖNER and R. GREILING: Stuttgart, E. Schweizerbart'sche Verlagsbuchhandlung, p. 113–127.
- HICKMAN, A. H., 1990, Geology of the Pilbara Craton, in *Third International Archaean Symposium, Excursion Guidebook No. 5: Pilbara and Hamersley Basin* edited by S. E. HO, J. E. GLOVER, J. S. MYERS, and J. R. MUHLING: University of Western Australia, Geology Department and University Extension, Publication no. 21, p. 2–13.
- HICKMAN, A. H., 1997, A revision of the stratigraphy of Archaean greenstone successions in the Roebourne–Whundo area, west Pilbara: Western Australia Geological Survey, Annual Review 1996–97, p. 76–82.
- HICKMAN, A. H., and LIPPLE, S. H., 1978, Marble Bar, W.A. Sheet SF 50-8 (2nd edition): Western Australia Geological Survey, 1:250 000 Geological Series.
- HICKMAN, A. H., VAN KRANENDONK, M. J., and SMITHIES, R. H., in prep., New tectonostratigraphic subdivisions of the Archaean Pilbara Craton, Western Australia: Australian Journal of Earth Sciences.
- HILL, R. M., 1997, Stratigraphy, structure and alteration of hanging wall sedimentary rocks at the Sulphur Springs volcanogenic massive sulfide (VMS) prospect, east Pilbara Craton, Western Australia: University of Western Australia, BSc Honours thesis (unpublished).
- HOFMANN, H. J., GREY, K., HICKMAN, A., and THORPE, R., 1999, Origin of 3.45 Ga coniform stromatolites in the Warrawoona Group, Western Australia: Geological Society of America, Bulletin, v. 3, p. 1256–1262.
- HORWITZ, R. C., 1990, The Archaean unconformity at Shay Gap, northeastern Pilbara Craton: Australia CSIRO, Institute for Minerals, Energy and Construction, Exploration Geoscience Restricted Report 123R.
- INGRAM, P. A. J., 1977, A summary of the geology of a portion of the Pilbara Goldfield, Western Australia, in *The Archaean, search for the beginning* edited by G. J. H. McCALL: Stroudsburg, Pennsylvania, Dowden, Hutchinson and Ross, p. 208–216.
- KRAPEZ, B., 1984, Sedimentation in a small, fault-bounded basin — the Lalla Rookh Sandstone, east Pilbara Block, in *Archaean and Proterozoic basins of the Pilbara, Western Australia — evolution and mineralization potential* edited by J. R. MUHLING, D. I. GROVES, and T. S. BLAKE: University of Western Australia, Geology Department and University Extension, Publication no. 9, p. 89–110.
- KRAPEZ, B., 1989, Depositional styles and geotectonic settings of Archaean metasedimentary sequences: Evidence from the Lalla Rookh Basin, Pilbara Block, Western Australia: University of Western Australia, PhD thesis (unpublished).
- KRAPEZ, B., 1993, Sequence stratigraphy of the Archaean supracrustal belts of the Pilbara Block, Western Australia: Precambrian Research, v. 60, p. 1–45.
- KRAPEZ, B., and BARLEY, M. E., 1987, Archaean strike-slip faulting and related ensialic basins — evidence from the Pilbara Block, Australia: Geological Magazine, v. 124, p. 555–567.
- KRIEVALDT, M. J. B., 1964, The Fortescue Group of the Roebourne region, North-West Division: Western Australia Geological Survey, Annual Report 1963, p. 30–34.
- LAMBERT, I. B., DONNELLY, T. H., DUNLOP, J. S. R., and GROVES, D. I., 1978, Stable isotopic compositions of early Archaean sulfate deposits of probable evaporitic and volcanogenic origins: Nature, v. 276, p. 808–810.
- LIPPLE, S. L., 1975, Definitions of new and revised stratigraphic units of the eastern Pilbara Region: Western Australia Geological Survey, Annual Report 1974, p. 58–63.
- LOWE, D. R., 1980, Stromatolites 3,400-Myr old from the Archaean of Western Australia: Nature, v. 284, p. 441–443.
- LOWE, D. R., 1983, Restricted shallow-water sedimentation of Early Archean stromatolitic and evaporitic strata of the Strelley Pool Chert, Pilbara Block, Western Australia: Precambrian Research, v. 19, p. 239–283.
- LOWE, D. R., 1994, Abiological origin of described stromatolites older than 3.2 Ga: Geology, v. 22, p. 387–390.
- LOWE, D. R., 1995, Abiological origin of described stromatolites older than 3.2 Ga: Reply: Geology, v. 23, p. 191–192.
- LOWE, D. R., and BYERLY, G. R., 1986, Early Archean silicate spherules of probable impact origin, South Africa and Western Australia: Geology, v. 14, p. 83–86.
- MacLEOD, W. N., de la HUNTY, L. E., JONES, W. R., and HALLIGAN, R., 1963, A preliminary report on the Hamersley Iron province, North-West Division: Western Australia Geological Survey, Annual Report 1962, p. 44–60.
- 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, v. 60, p. 69–98.
- McNAUGHTON, N. J., GREEN, M. D., COMPSTON, W., and WILLIAMS, I. S., 1988, Are anorthositic rocks basement to the Pilbara Craton?: Geological Society of Australia, Abstracts v. 21, p. 272–273.
- McPHEE, J., DOYLE, M., and ALLEN, R., 1993, Volcanic textures: A guide to the interpretation of textures in volcanic rocks: University of Tasmania, Centre for Ore Deposit and Exploration Studies, 198p.
- MACKEY, T. E., and RICHARDSON, L. M., 1997a, Total magnetic intensity (reduced to the pole) with westerly illumination colour pixel-image map of Marble Bar, W.A., (1:250 000 scale): Australian Geological Survey Organisation.
- MACKEY, T. E., and RICHARDSON, L. M., 1997b, Airborne gamma-ray spectrometry colour composite pixel-image map of Marble Bar, W.A., (1:250 000 scale): Australian Geological Survey Organisation.
- MARESCHAL, J.-C., and WEST, G. F., 1980, A model for Archaean tectonism. Part 2. Numerical models of vertical tectonism in greenstone belts: Canadian Journal of Earth Sciences, v. 17, p. 60–71.
- MORANT, P., 1984, Metamorphism of an Archaean granitoid–greenstone terrain, east Pilbara Block: the western Shaw Batholith and adjacent greenstones: University of Western Australia, PhD thesis (unpublished).
- MORANT, P., 1995, The Panorama Zn–Cu VMS deposits, Western Australia: Australian Institute of Geoscientists, Bulletin, v. 16, p. 75–84.
- MORANT, P., 1998, Panorama zinc–copper deposits, in *Geology of Australian and Papua New Guinean mineral deposits* edited by D. A. BERKMAN and D. H. MacKENZIE: Australasian Institute of Mining and Metallurgy, Melbourne, Monograph 22, p. 287–292.
- NELSON, D. R., 1997, Compilation of SHRIMP U–Pb zircon geochronology data, 1996: Western Australia Geological Survey, Record 1997/2, 189p.
- NELSON, D. R., 1998, Compilation of SHRIMP U–Pb zircon geochronology data, 1997: Western Australia Geological Survey, Record 1998/2, 242p.
- NELSON, D. R., 1999, Compilation of geochronology data, 1998: Western Australia Geological Survey, Record 1999/2, 222p.

- NELSON, D. R., in prep., Compilation of geochronology data, 1999: Western Australia Geological Survey, Record 2000/2.
- NIJMAN, W., de BRUIJNE, K. C. H., and VALKERING, M. E., 1998, Growth fault control of Early Archaean cherts, barite mounds and chert-barite veins, North Pole Dome, eastern Pilbara, Western Australia: *Precambrian Research*, v. 88, p. 25–52.
- NOLDART, A. J., and WYATT, J. D., 1962, The geology of a portion of the Pilbara Goldfield, covering the Marble Bar and Nullagine 4-mile map sheets: Western Australia Geological Survey, Bulletin 115, 199p.
- PIDGEON, R. T., 1978, 3450 m.y. old volcanics in the Archaean layered greenstone succession of the Pilbara Block, Western Australia: *Earth and Planetary Science Letters*, v. 37, p. 421–428.
- PIDGEON, R. T., 1984, Geochronological constraints on early volcanic evolution of the Pilbara Block, Western Australia: *Australian Journal of Earth Sciences*, v. 31, p. 237–242.
- PIRAJNO, F., 1992, Hydrothermal mineral deposits: principles and fundamental concepts for the exploration geologist: Berlin, Springer-Verlag, 709p.
- PITCHER, W. S., 1993, The nature and origin of granite: London, Blackie Academic and Professional, 321p.
- PRATT, B. R., 1998, Syneresis cracks — Subaqueous shrinkage in argillaceous sediments caused by earthquake-induced dewatering: *Sedimentary Geology*, v. 117, p. 1–10.
- RICHARDS, J. R., 1986, Lead isotopic signatures: further examination of comparisons between South Africa and Western Australia: *Transactions of the Geological Society of South Africa*, v. 89, p. 285–304.
- RICHARDS, J. R., FLETCHER, I. R., and BLOCKLEY, J. G., 1981, Pilbara galenas: precise isotopic assay of the oldest Australian leads; model ages and growth-curve implications: *Mineralium Deposita*, v. 16, p. 7–30.
- RYAN, G. R., 1965, The geology of the Pilbara Block: Australasian Institute of Mining and Metallurgy, Proceedings 214, p. 61–94.
- SARGEANT, D. W., 1979, Annual report on mineral claim numbers 45/8674 to 45/8684 inclusive North Pole area, Pilbara W.A.: Report by Newmont Pty Ltd (unpublished).
- SARGEANT, D. W., and SAMPSON, P. G., 1980, Final report on Temporary reserve 6712H North Pole area, Pilbara Goldfield, Report by Newmont Proprietary Ltd: Western Australia Geological Survey, M-Series, Item 2760 (unpublished).
- SCHOPF, J. W., 1992, Newly discovered microfossils from the Early Archaean Apex Basalt (Warrawoona Group), Western Australia, in *The Proterozoic biosphere: a multi-disciplinary study edited by J. W. SCHOPF and C. KLEIN*: Cambridge University Press, p. 25–39.
- SCHOPF, J. W., 1993, Microfossils of the Early Archaean Apex Chert: new evidence of the antiquity of life: *Science*, v. 260, p. 640–646.
- SCHOPF, J. W., and PACKER, B. M., 1987, Early Archaean (3.3 billion to 3.5 billion-year-old) microfossils from Warrawoona Group, Australia: *Science*, v. 237, p. 70–73.
- SIMPSON, E. S., 1948, Minerals of Western Australia, Volume 1: Perth, Western Australia, Government Printer.
- SMITHIES, R. H., HICKMAN, A. H., and NELSON, D. R., 1999, New constraints on the evolution of the Mallina Basin, and their bearing on relationships between the contrasting eastern and western granite-greenstone terrains of the Archaean Pilbara Craton, Western Australia: *Precambrian Research*, v. 94, p. 11–28.
- SPEAR, F. S., 1993, Metamorphic phase equilibria and pressure-temperature-time-paths: Mineralogical Society of America, Monograph, 799p.
- THORPE, R. I., HICKMAN, A. H., DAVIS, D. W., MORTENSEN, J. K., and TRENDALL, A. F., 1992a, U–Pb zircon geochronology of Archaean felsic units in the Marble Bar region, Pilbara Craton, Western Australia: *Precambrian Research*, v. 56, p. 169–189.
- THORPE, R. I., HICKMAN, A. H., DAVIS, D. W., MORTENSEN, J. K., and TRENDALL, A. F., 1992b, Constraints to models for Archaean lead evolution from precise zircon U–Pb geochronology for the Marble Bar region, Pilbara Craton, Western Australia in *The Archaean: Terrains, processes and metallogeny edited by J. E. GLOVER and S. E. HO*: University of Western Australia, Geology Department and University Extension, Publication no. 22, p. 395–407.
- TRENDALL, A. F., 1990, Hamersley Basin, in *Geology and mineral resources of Western Australia*: Western Australia Geological Survey, Memoir 3, p. 163–190.
- TRENDALL, A. F., 1995, Paradigms for the Pilbara, in *Early Precambrian processes edited by M. P. COWARD and A. C. REIS*: Geological Society of London, Special Publication, no. 95, p. 127–142.
- VAN HAAFTEN, W. M., and WHITE, S. H., 1998, Evidence for multiphase deformation in the Archaean based Warrawoona Group in the Marble Bar area, east Pilbara, Western Australia: *Precambrian Research*, v. 88, p. 53–66.
- VAN KRANENDONK, M. J., 1997, Results of field mapping, 1994–1996, in the North Shaw and Tambourah 1:100 000 sheet areas, eastern Pilbara Craton, northwestern Australia: Australian Geological Survey Organisation, Record 1997/23, 44p.
- VAN KRANENDONK, M. J., 1998, Litho-tectonic and structural map components of the North Shaw 1:100 000 sheet, Archaean Pilbara Craton: Western Australia Geological Survey, Annual Review 1997–98, p. 63–70.
- VAN KRANENDONK, M. J., 1999, Two-stage degassing of the Archaean mantle: evidence from the 3.46 Ga Panorama volcano, Pilbara Craton, Western Australia, in *GSWA 99 extended abstracts: New geological data for WA explorers*: Western Australia Geological Survey, Record 1999/6, p. 1–3.
- VAN KRANENDONK, M. J., in prep.a, Structural evidence for diapiric re-activation of the Yule Batholith during extension and basin formation in the central-eastern part of the Archaean Pilbara Craton, Western Australia: *Journal of Structural Geology*.
- VAN KRANENDONK, M. J., in prep.b, Geochemistry of rocks from the North Shaw 1:100 000 sheet, Pilbara Craton, Western Australia, and tectonic implications: Western Australia Geological Survey, Record.
- 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*, v. 88, p. 207–232.
- VAN KRANENDONK, M. J., and COLLINS, W. J., in prep., Partial convective overturn of the eastern part of the Archaean Pilbara Craton, Western Australia: Field evidence for the diapiric emplacement of granitoid domes: *Precambrian Research*.
- VAN KRANENDONK, M. J., and MORANT, P., 1998, Revised Archaean stratigraphy of the North Shaw 1:100 000 sheet, Pilbara Craton: Western Australia Geological Survey, Annual Review 1997–98, p. 55–62.
- VAN KRANENDONK, M. J., and WARDLE, R. J., 1996, Burwell domain of the Paleoproterozoic Torngat Orogen, northeastern Canada: Tilted cross section of a magmatic arc caught between a rock and a hard place, in *Precambrian evolution of the North Atlantic Region edited by T. S. BREWER*: Geological Society of London, Special Publication, no. 112, p. 91–115.

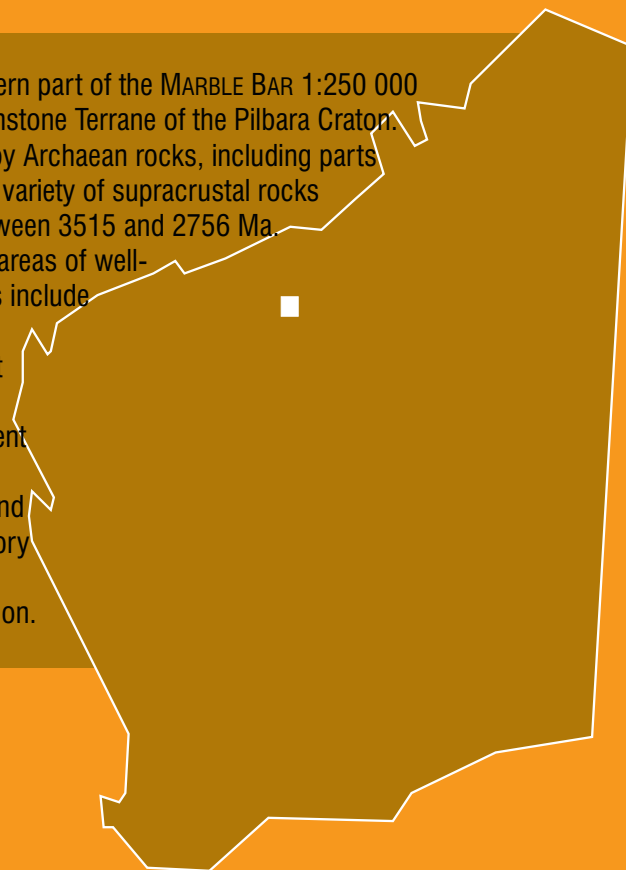
- VAN KRANENDONK, M. J., HICKMAN, A. H., and COLLINS, W. J., in press, Evidence for multiphase deformation in the Archaean basal Warrawoona Group in the Marble Bar area, east Pilbara, Western Australia: Comment: Precambrian Research.
- VEARNCOMBE, S., 1996, Volcanogenic massive sulfide-sulphate mineralisation at Strelley, Pilbara Craton, Western Australia: University of Western Australia, PhD thesis (unpublished).
- VEARNCOMBE, S., BARLEY, M. E., GROVES, D. I., McNAUGHTON, N. J., MIKUCKI, E. J. and VEARNCOMBE, J. R., 1995, 3.26 Ga black smoker-type mineralization in the Strelley Belt, Pilbara Craton, Western Australia: *Journal of the Geological Society of London*, v. 152, p. 587–590.
- VEARNCOMBE, S., VEARNCOMBE, J. R., and BARLEY, M. E., 1998, Fault and stratigraphic controls on volcanogenic massive sulfide deposits in the Strelley Belt, Pilbara Craton, Western Australia: *Precambrian Research*, v. 88, p. 67–82.
- WALTER, M. R., 1972, Stromatolites and the biostratigraphy of the Australian Precambrian and Cambrian: *Palaeontological Association (London), Special Papers in Palaeontology*, no. 11, 190p.
- WALTER, M. R., 1983, Archean stromatolites: evidence of the Earth's oldest benthos, in *Earth's earliest biosphere — its origins and evolution* edited by J. W. SCHOPF: New Jersey, Princeton University Press, p. 187–213.
- WALTER, M. R., BUICK, R., and DUNLOP, J. S. R., 1980, Stromatolites, 3 400–3 500 Myr old from the North Pole area, Western Australia: *Nature*, v. 284, p. 443–445.
- WIJBRANS, J. R., and McDOUGALL, I., 1987, On the metamorphic history of an Archaean granitoid-greenstone terrane, east Pilbara, Western Australia, using the <sup>40</sup>Ar/<sup>39</sup>Ar age spectrum technique: *Earth and Planetary Science Letters*, v. 84, p. 226–242.
- WILHELMIJ, H. R., 1986, Depositional history of the middle Archaean sedimentary sequences in the Pilbara Block, Western Australia — a genetic stratigraphic analysis of the terrigenous rocks of the Pilgangoora syncline: University of Western Australia, PhD thesis (unpublished).
- WILHELMIJ, H. R., and DUNLOP, J. S. R., 1984, A genetic stratigraphic investigation of the Gorge Creek Group in the Pilgangoora Syncline, in *Archaean and Proterozoic basins of the Pilbara, Western Australia — evolution and mineralization potential* edited by J. R. MUHLING, D. I. GROVES, and T. S. BLAKE: University of Western Australia, Geology Department and University Extension, Publication no. 9, p. 68–88.
- WILLIAMS, I. R., 1998, Muccan, W.A. Sheet 2956: Western Australia Geological Survey, 1:100 000 Geological Series.
- 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*, v. 97, p. 41–53.
- WILLIAMS, I. S., PAGE, R. W., FROUDE, D., FOSTER, J. J., and COMPSTON, W., 1983, Early crustal components in the Western Australian Archaean: Zircon U–Pb ages by ion microprobe analysis from the Shaw Batholith and Narryer Metamorphic Belt: *Geological Society of Australia, Abstracts Series* v. 9, p. 169.
- WINGATE, M. T. D., 1999, Ion microprobe baddeleyite and zircon ages for Late Archaean mafic dykes of the Pilbara Craton, Western Australia: *Australian Journal of Earth Sciences*, v. 46, p. 493–500.
- ZEGERS, T. E., 1996, Structural, kinematic and metallogenic evolution of selected domains of the Pilbara granitoid-greenstone terrain: implications for mid-Archaean tectonic regimes: The Netherlands, University of Utrecht, PhD thesis (unpublished).
- ZEGERS, T. E., WHITE, S. H., de KEIJZER, M., and DIRKS, P., 1996, Extensional structures during deposition of the 3460 Ma Warrawoona Group in the eastern Pilbara Craton, Western Australia: *Precambrian Research*, v. 80, p. 89–105.
- ZEGERS, T. E., de KEIJZER, M., PASSCHIER, C. W., and WHITE, S. H., 1998, The Mulgandinnah shear zone and Archean crustal-scale strike-slip zone, eastern Pilbara, Western Australia: *Precambrian Research*, v. 88, p. 233–248.



WODGINA 2655	NORTH SHAW 2755	MARBLE BAR 2855
MARBLE BAR SF 50-8		
WHITE SPRINGS 2654	TAMBOURAH 2754	SPLIT ROCK 2854

The NORTH SHAW 1:100 000 sheet covers the central-northern part of the MARBLE BAR 1:250 000 sheet, in the central part of the East Pilbara Granitoid–Greenstone Terrane of the Pilbara Craton. The area is well exposed and almost exclusively underlain by Archaean rocks, including parts of three domal, multicomponent granitoid complexes and a variety of supracrustal rocks belonging to six unconformity-bound groups deposited between 3515 and 2756 Ma.

Although structurally complex, NORTH SHAW contains large areas of well-preserved rocks, including Earth's oldest fossils. Structures include those related to granitoid doming at c. 3460 Ma, 3240 Ma, and 2760 Ma, and those related to sinistral transpression at c. 2930 Ma. Base metal prospects occur around the synvolcanic Strelley Granite laccolith, platinum-group element prospects occur in differentiated ultramafic sills, epigenetic gold has been mined from a variety of structural settings, and alluvial tin was mined from granite detritus. These Explanatory Notes describe the geology in detail, including detailed lithostratigraphy, structure, metamorphism, and mineralization.



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