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PROTEROZOIC GEOLOGY OF THE WESTERN CAPRICORN OROGEN — A FIELD GUIDE

by D. McB. Martin, S. Sheppard, A. M. Thorne, T. R. Farrell, and P. B. Groenewald

Geological Survey of Western Australia

GEOLOGICAL SURVEY OF WESTERN AUSTRALIA

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D. McB. Martin. S. Sheppard, A. M. Thorne, T. R. Farrell, and P. B. Groenewald

Introduction

This field guide accompanies the September 2006 excursion by the Geological Survey of Western Australia (GSWA) to the western part of the Capricorn Orogen. Tectonic units examined include the Ashburton Basin, Gascoyne Complex, Edmund Basin, and Collier Basin (Fig. 1). The excursion is to be carried out in two parts, the first of which (Part A) will examine:

- sedimentation and deformation within the upper part of the Paleoproterozoic Ashburton Basin, and the nature of the contact between the Ashburton Basin and the overlying Paleoproterozoic to Mesoproterozoic Edmund Basin;
- recent revisions to the stratigraphy of the Edmund Basin and the overlying Mesoproterozoic Collier Basin, and the effect of the major structural elements on basin evolution;
- field characteristics, U–Pb zircon and baddelyite geochronology, and geochemistry of two suites of Mesoproterozoic dolerite sills intruded into the Edmund and Collier Groups;
- deformation of the Edmund and Collier Basins during the 1070–750 Ma Edmundian Orogeny.
- a brief overview of the major regolith units within the excursion area.

The second part of the excursion (Part B) will examine:

- the nature of metamorphism, deformation, and magmatism during the 1830–1780 Ma Capricorn Orogeny in the Gascoyne Complex;
- metasedimentary rocks of the c. 1680 Ma Pooranoo Metamorphics;
- the contact between metasedimentary rocks of the younger-than-1840 Ma Morrissey Metamorphics and Pooranoo Metamorphics;
- the nature of metamorphism, magmatism, and deformation during the 1680–1620 Ma Mangaroon Orogeny;

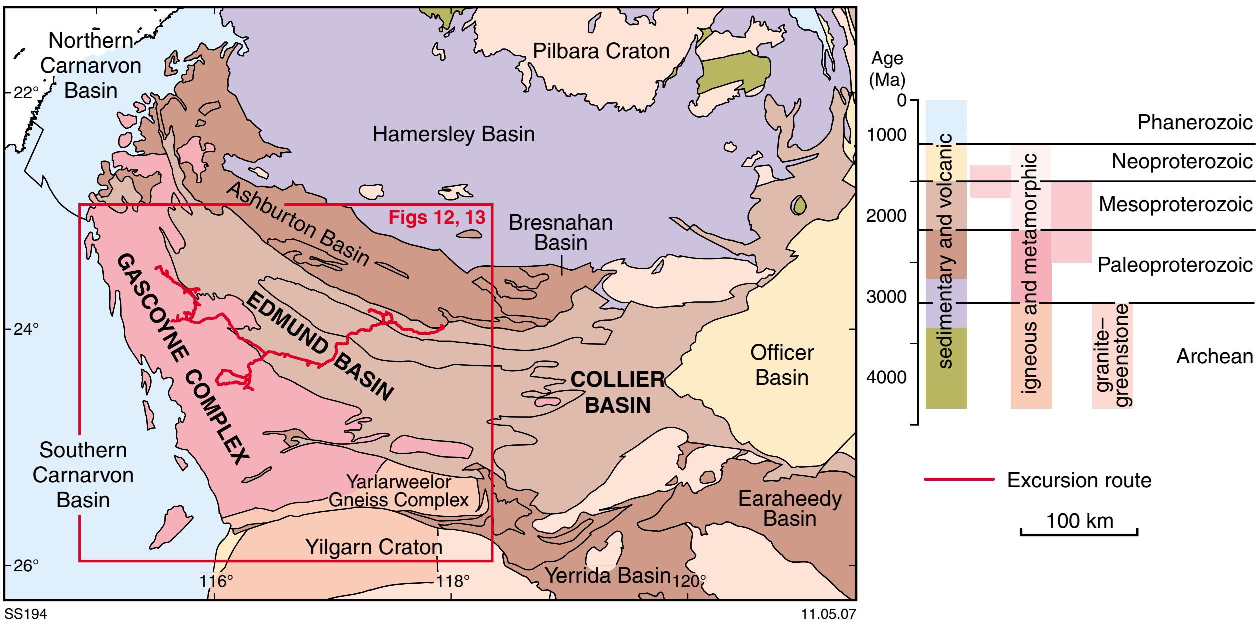

Figure 1. Tectonic units in the central part of Western Australia, with the excursion route shown

- newly recognized regional metamorphism and granitic magmatism dated at 1030–950 Ma in the central Gascoyne Complex;
- brittle–ductile strike-slip structures reflecting late Neoproterozoic (<755 Ma) reactivation in the Gascoyne Complex and Edmund Group.

The excursion illustrates the results of an ongoing GSWA program integrating regional mapping with geochronological, structural, stratigraphic, metamorphic, and geochemical studies. Parts of the guide are directly taken from an earlier field guide to the Capricorn Orogen (Occhipinti et al., 2004) and the explanatory notes describing the geology of the MAROONAH*, ULLAWARRA, CAPRICORN, MANGAROON, EDMUND, and ELLIOTT CREEK 1:100 000 map sheets (Martin et al., 2005).

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

Regional geology

The excursion area is in the western part of the Capricorn Orogen, which is a major zone of Proterozoic deformation, metamorphism, and magmatism between the Archean Yilgarn and Pilbara Cratons (Gee, 1979; Cawood and Tyler, 2004; Fig. 1). The orogen includes metamorphic and igneous rocks of the Gascoyne Complex, a number of sedimentary basins, as well as the deformed margins of the Yilgarn and Pilbara Cratons. Within the excursion route, Capricorn Orogen rocks comprise Paleoproterozoic rocks of the Ashburton Basin and Gascoyne Complex, Late Paleoproterozoic to Mesoproterozoic rocks of the Edmund Basin, and Mesoproterozoic rocks of the Collier Basin (Plate 1).

Part A: Ashburton, Edmund, and Collier Basins

by

D. McB. Martin and A. M. Thorne

Part A of the excursion will focus on Late Paleoproterozoic to Mesoproterozoic rocks of the Edmund Basin, and Mesoproterozoic rocks of the Collier Basin. In addition, the upper part of the underlying Ashburton Basin will be examined briefly.

Geological setting

Ashburton Basin

The Ashburton Basin (Thorne, 1990; Thorne and Seymour, 1991) corresponds to the present-day outcrop of the Wyloo Group, which is a 12 km-thick succession of low-grade, metasedimentary and metavolcanic rocks. The Wyloo Group is subdivided into a lower succession consisting of the Beasley River Quartzite and the overlying Cheela Springs Basalt, and an upper succession comprising the Mount McGrath Formation, Duck Creek Dolomite, June Hill Volcanics, and Ashburton Formation (Thorne and Seymour, 1991).

Tyler and Thorne (1990) and Thorne and Seymour (1991) interpreted the Ashburton Basin stratigraphy as a response to the transition from an active tectonic margin to a foreland basin, during the convergence of the Pilbara and Yilgarn Cratons. The strongly diachronous nature of sedimentation during the foreland basin stage was highlighted by Evans et al. (2003) in their comparison of geochronological data from the upper Wyloo Group and the overlying Capricorn Group. Krapež (1999) considered the Ashburton Basin fill in terms of two megasequences that record the opening and closure of an Atlantic-type ocean. Other workers have suggested that the lower part of the Ashburton Basin succession, along with the underlying Turee Creek Group of the Mount Bruce Supergroup, were deposited in the McGrath Trough (Blake and Barley, 1992; Horwitz, 1982; Martin et al., 2000; Powell and Horwitz, 1994), which is a foreland basin related to the development of the Ophthalmia Fold Belt (Cawood and Tyler, 2004).

The Ashburton Formation is the only part of the Wyloo Group that will be examined during this excursion. The age of these rocks is controversial. Sensitive high-resolution ion microprobe (SHRIMP) U–Pb data have been obtained from volcanic units within the formation, from volcanic rocks in the underlying June Hill Volcanics and the overlying Capricorn Group, and from the intrusive Boolaloo Granodiorite. Evans et al. (2003) obtained SHRIMP U–Pb zircon ages of 1799 ± 8 and 1786 ± 11 Ma from the June Hill Volcanics on WYLOO, whereas a felsic volcaniclastic sandstone within the Ashburton Formation has yielded a SHRIMP U–Pb zircon age of 1806 ± 9 Ma (Geological Survey of Western Australia, 2006; GSWA 148922). These dates are slightly younger than the SHRIMP U–Pb age of c. 1819 Ma reported from the youngest zircon population in seven Ashburton Formation sandstones (Sircombe, K. N., 2002, written comm.) and the SHRIMP U–Pb age of 1829 ± 5 Ma obtained from a felsic volcaniclastic unit within the Ashburton Formation (Sircombe, 2003). Hall et al. (2001) reported a SHRIMP U–Pb age of 1804 ± 7 Ma for a felsic volcanic unit in the unconformably overlying Capricorn Formation (now Capricorn Group). The younger age limit for the Ashburton Formation is also constrained by the SHRIMP U–Pb age of 1786 ± 5 Ma for the Boolaloo Granodiorite that intrudes these rocks on WYLOO (Krapež and McNaughton, 1999). The contradictory nature of some of these age data may reflect markedly diachronous foreland basin sedimentation during the Capricorn Orogeny (Tyler and Thorne, 1990; Thorne and Seymour, 1991; Evans et al., 2003).

The Ashburton Formation is composed of mudstone, siltstone, and immature sandstone, interbedded with minor amounts of conglomerate and volcanic rock. Paleocurrent and sandstone composition data from the Ashburton Formation in the excursion area suggest that sediment transport was toward the west-northwest (Thorne and Seymour, 1991). Thorne and Seymour (1991) interpreted the Ashburton Formation as a linear submarine fan system developed in a foreland basin setting.

Paleoproterozoic Capricorn Orogeny

The Capricorn Orogeny has been interpreted as reflecting oblique collision of the Archean Yilgarn and Pilbara Cratons (Tyler and Thorne, 1990), but is now thought to result from intracratonic reworking following an older Paleoproterozoic collision event (Sheppard, 2004). Deformation, low- to medium-grade metamorphism, and voluminous granite intrusion across the Gascoyne Complex and northern Capricorn Orogen occurred between 1830 and 1780 Ma (Sheppard and Occhipinti, 2000; Occhipinti and Sheppard, 2001; Occhipinti et al., 2001; Martin et al., 2004).

Ashburton Fold Belt

The Ashburton Fold Belt formed in response to the Capricorn Orogeny and includes the deformed Wyloo and Capricorn Groups. The structural history of the Ashburton Fold Belt has been presented by Tyler and Thorne (1990) and Thorne and Seymour (1991) who described two major deformation events — D_{1a} and D_{2a}. In addition, a third deformation event (D_{3a}) has been recognized by Martin et al. (2004). The earliest deformation, D_{1a}, occurred after deposition of the Ashburton Formation (upper Wyloo Group), but prior to deposition of the Capricorn Group. This deformation was accompanied, or slightly post-dated, by a regional metamorphic event (M_{1a}). The second and third events (D_{2a}/M_{2a} and D_{3a}) occurred after deposition of the Capricorn Group and following intrusion of the Boolaloo Granodiorite, but before deposition of the Edmund Group. On the basis of these relationships the timing of D_{1a}/M_{1a} was between c. 1806 and 1787 Ma, whereas D_{2a} and D_{3a} occurred between c. 1787 and c. 1620 Ma.

Much of the evidence for the D_{1a} deformation is based on the widespread S_{1a} foliation, the marked angular unconformity between the Ashburton Formation and the Capricorn Group, and rare F_{1a} folds. Throughout most of the Ashburton Basin the metamorphic grade of M_{1a} is low; however, there is a general increase in grade and schistosity towards the west and southwest. Thorne and Seymour (1991) noted that much of the Ashburton Basin is characterized by the mineral assemblage quartz–chlorite–muscovite(–sericite) in pelitic and psammitic rocks.

Most of the obvious folding and faulting in the Ashburton Fold Belt results from the second deformation event (D_{2a}). F_{2a} folds in the excursion area are small to large (1–6000 m wavelength), open to tight, noncylindrical, and trend west to northwest. Most plunge 10–40° (up to 80° locally) to the southeast or northwest. Axial planes dip steeply to the southwest or northeast. S_{2a} is a penetrative slaty cleavage in the more easterly outcrops, but is a crenulation cleavage further west. Numerous west-northwesterly to north-northwesterly trending dextral strike-slip faults are parallel to the F_{2a} fold axes or crosscut them at a shallow angle.

Late Paleoproterozoic to Mesoproterozoic Edmund and Collier Basins

The Edmund and Collier Basins are the youngest tectonic units within the Capricorn Orogen and correspond to the present-day outcrop of the Edmund and Collier Groups that together make up the Bangemall Supergroup. These rocks unconformably overlie the Paleoproterozoic Ashburton, Blair, and Bresnahan Basins to the northeast and rocks of the Paleoproterozoic Gascoyne Complex to the southwest (Plates 1 and 2). The Edmund Basin was initiated during intracratonic extensional reactivation of structures formed during earlier stages of the Capricorn Orogen. These structures were intermittently active throughout the evolution of the basin, but do not appear to have had a significant influence on the Collier Basin. Subsidence of the Edmund and Collier Basins was most likely driven by plate boundary forces generated in the Pinjarra, Albany–Fraser, and Paterson Orogens.

Edmund Group

The Edmund Group is about 4 km thick and consists, from bottom to top, of the Yilgatherra, Irregully, Gooragoora, Blue Billy, Cheyne Springs, Kiangi Creek, Muntharra, Discovery, Devil Creek, Ullawarra, and Coodardoo Formations. The Mount Augustus Sandstone, a thick succession of coarse-grained siliciclastic rocks that outcrops on northern MOUNT PHILLIPS, may correlate with the lower part of the Yilgatherra Formation.

The Edmund Group is younger than 1680–1620 Ma granites in the underlying Gascoyne Complex, and older than c. 1465 Ma dolerite sills that intrude it over large areas of EDMUND (Martin and Thorne, 2004).

Yilgatherra Formation

The Yilgatherra Formation (Martin and Thorne, 2002) is the oldest lithostratigraphic unit in the Edmund Group and has a maximum thickness of about 350 m. North of the Talga Fault the Yilgatherra Formation forms a discontinuous unit at the base of the Edmund Group, but is more continuous south of the fault.

The basal Yilgatherra Formation consists of an upward-fining succession, from localized pebble to cobble conglomerate (up to 20 m), through very coarse grained sandstone, to fine-grained sandstone and siltstone. North of the Talga Fault this succession fills topographic lows, and is locally interbedded with stromatolitic dolostone and planar-laminated siltstone. South of the Talga Fault, where conglomerates are generally thinner, the Yilgatherra Formation consists of an upward-fining succession from very coarse or coarse-grained sandstone to planar-laminated siltstone. Paleocurrent directions from trough cross-stratified sandstone are highly variable with paleoflow to the northwest, west, southwest, and southeast. Current directions may be unimodal, polymodal or bipolar.

Deposition of the Yilgatherra Formation occurred during a marine transgression in environments ranging from fluvial to shallow marine, and was controlled by active basement highs south of the Talga Fault.

Mount Augustus Sandstone

The Mount Augustus Sandstone (Muhling and Brakel, 1985) is a thick (>1150 m) succession of coarse-grained siliciclastic rocks that unconformably overlies metasedimentary and granitic rocks of the Mangaroon Zone (Gascoyne Complex) on MOUNT AUGUSTUS. There is some uncertainty regarding the true stratigraphic position of the Mount Augustus Sandstone. Muhling and Brakel (1985) assigned the formation to the basal part of the Edmund Group based on the apparently conformable upper contact with the Irregully Formation on the southwestern limb of the Cobra Syncline. As such the Mount Augustus Sandstone is therefore likely to be a thickened correlative of the lower Yilgatherra Formation. An alternative interpretation, based on the formation's thickness, lithology, and provenance, is that it may correlate with the Bresnahan Group that unconformably underlies the Edmund Group in the north-central Capricorn Orogen (Cooper et al., 1998).

The Mount Augustus Sandstone comprises an upward-fining succession that varies from localized pebble to boulder conglomerate at the base, through very coarse grained sandstone, to fine-grained sandstone. Most of the sandstones are feldspathic and tourmaline-bearing, and their internal structure is dominated by stacked sets of trough cross-strata. The latter record a consistent paleoflow from northwest to southeast.

The Mount Augustus Sandstone is interpreted as a localized fluvial deposit, laid down in a southeast-trending basin associated with the Lyons River Fault system.

Irregully Formation

The Irregully Formation conformably overlies the Yilgatherra Formation in most areas, and is conformably to unconformably overlain by either the Gooragoora Formation, Blue Billy Formation, or Kiangi Creek Formation. The principal lithologies are laminated and stromatolitic dolostone and dolorudite, minor sandstone, and dolomitic sandstone with rare conglomerate. A significant thickness of sandstone and siltstone is interbedded with the dolostone on MAROONAH and northeast MANGAROON (Martin and Thorne, 2000). The thickness of the Irregully Formation increases from about 350 m in the type area on Irregully Creek to more than 2000 m on MAROONAH and northeast MANGAROON. This thickness change is related to syndepositional movement along the Talga Fault. The thickness of the Irregully Formation decreases markedly towards the southeast corner of EDMUND due to erosion on a disconformity at the base of the Kiangi Creek Formation.

Stromatolite types recognized include *Conophyton garganicum australe* Walter 1972, *Paniscollenia* Koroljuk 1960, *Colonella* Komar 1964, and *Baicalia capricornia* Walter 1972. Paleocurrent directions in sandy dolostones on southeast MANGAROON were commonly to the south, with a wide spread between southwest and east-southeast (Daniels, 1966).

The depositional environment of the Irregully Formation ranges from a peritidal platform north of the Talga Fault, in an area known as the Pingandy Shelf, to mainly slope and deep-marine deposits with localized fluvial and shallow-marine facies south of the fault. Fluvial and shallow-marine facies south of the Talga Fault were deposited around the margins of paleotopographic highs on MAROONAH, MANGAROON, and EDMUND (1:100 000). The upper parts of the Irregully Formation on the Pingandy Shelf consists of slope lithofacies capped by a widespread stromatolitic bioherm that records the final shallowing of the carbonate platform to wave base. This shallowing is also recorded south of the Talga Fault. In the Mangaroon Syncline the Irregully Formation is dominated by slope-facies carbonates, suggesting localized steep paleoslopes, possibly due to movement along the major bounding faults.

Gooragoora Formation (not examined on this excursion)

The Gooragoora Formation sharply overlies the Irregully Formation and is in turn overlain by the Blue Billy Formation. A slight local angular discordance with the Irregully Formation, noted by Chuck (1984), suggests that the lower contact of the Gooragoora Formation may be unconformable. The upper contact with the Blue Billy Formation is marked by an abrupt change to pyritic carbonaceous siltstone. On southeastern EDMUND the Gooragoora Formation is truncated beneath a disconformity at the base of the Kiangi Creek Formation.

The Gooragoora Formation consists mainly of medium- to thick-bedded, trough cross-stratified, very coarse to fine-grained sandstone, with lesser amounts of conglomerate, planar-laminated siltstone, and dolostone. Paleocurrent directions in sandstone units on EDMUND are locally highly variable, but predominantly towards the southeast.

The Gooragoora Formation is interpreted to have been deposited by the progradation of coarse-grained deltaic sandstones during a period of high or rising sea level (Martin and Thorne, 2004). Deltaic sedimentation was restricted to the northwestern parts of EDMUND where uplift to the northwest exceeded the rate of sea level rise. Finer grained facies interbedded with planar-laminated siltstone were probably deposited in a proximal shelf environment.

Blue Billy Formation

The Blue Billy Formation is exposed throughout most of EDMUND and KENNETH RANGE, and is up to 500 m thick (Martin and Thorne, 2002), but generally less than 200 m thick. The Blue Billy Formation is truncated by a disconformity beneath the Kiangi Creek Formation on western EDMUND.

The formation consists primarily of a distinctive parallel-planar-laminated, black, pyritic, carbonaceous, fissile to flaggy siltstone that weathers to a blue-grey colour. Weathered surfaces are also commonly coated with a white efflorescence of gypsum, particularly in creek-bed exposures. Parallel-bedded gypsum layers, with crystals oriented normal to bedding, and carbonate are exposed in Irregully Gorge on ULLAWARRA. Pyrite is largely weathered to goethite and hematite, and is concentrated in thin beds or laminae, or disseminated throughout the rock. The characteristic blue-grey to black colour is due to finely disseminated opaque minerals, a high total organic carbon content (1.60–3.87%), and localized silicification. Goethite and hematite nodules and concretions are also present locally. On a regional scale, siltstones become more dolomitic towards the top of the Blue Billy Formation, and the contact with the overlying Cheyne Springs Formation. The Blue Billy Formation also contains tabular to lenticular bodies of thick-bedded, massive, medium- to coarse-grained quartz sandstone, dolomitic sandstone, and pebbly sandstone interbedded with planar-laminated siltstone and minor dolostone. Groove casts on the base of a sandstone bed on MAROONAH trend northwest–southeast, whereas ripple marks on ELLIOTT CREEK have a paleocurrent direction to the southwest.

The Blue Billy Formation is interpreted as a transgressive deposit of deep marine-shelf anoxic mudstone that records transgression and expansion of the Edmund Basin beyond the preserved limits of the Pingandy Shelf. Sandstones in the upper part of the Blue Billy Formation, which were deposited north and south of the Talga Fault on ULLAWARRA, MAROONAH, MANGAROON, and EDMUND (1:100 000), are interpreted as turbidite and mass-flow deposits (Martin and Thorne, 2004). The upward increase in grain size, sandstone bed thickness, and dolomite content suggests that the Blue Billy Formation is an upward-shallowing distal shelf succession.

Cheyne Springs Formation (not examined on this excursion)

The Cheyne Springs Formation comprises interbedded dololutite, dolomitic siltstone, and siltstone, together with minor fine- to medium-grained dolomitic sandstone and quartz sandstone. The formation ranges from 50 to 300 m thick on EDMUND, but seldom exceeds 200 m. It is thickest immediately adjacent to the Talga Fault, and thins rapidly to the south. The Cheyne Springs Formation is truncated beneath a disconformity at the base of the Kiangi Creek Formation. Paleocurrent directions determined mainly from ripple cross-lamination and current lineation on the soles of sandstone beds range from south to northwest.

The Cheyne Springs Formation has been interpreted as an association of shelf and pelagic dolostones (Chuck, 1984) that marks a slight shallowing of the Edmund Basin (Martin and Thorne, 2004). The presence of localized hummocky cross-stratification suggests shallowing to storm wave-base, and sandstones of possible turbidite origin indicate a broadly southwest-sloping shelf.

Kiangi Creek Formation

The Kiangi Creek Formation is defined as the succession of interbedded siltstone, sandstone, and local dolostone that disconformably overlies the Cheyne Springs Formation and lower units in the Edmund Group, and is overlain by either the Muntharra or Discovery Formations (Martin and Thorne, 2002). The formation is present throughout EDMUND, and typically ranges from 200 to 550 m thick, with the thickest sections southwest of the Talga Fault.

The basal contact of the Kiangi Creek Formation is a very low angle disconformity on EDMUND 1:100 000. In complete sections the basal contact overlies the uppermost carbonate in the Cheyne Springs Formation; however, it succeeds the Blue Billy Formation on parts of MAROONAH, MANGAROON, and EDMUND (1:100 000) and directly overlies the Irregully Formation near Pimbyana Hill on southeast EDMUND (1:100 000). The formation is overlain by the Muntharra Formation on ELLIOTT CREEK and ULLAWARRA, and by the Discovery Formation elsewhere on EDMUND.

The Kiangi Creek Formation consists mainly of varying proportions of interbedded planar-laminated siltstone and thin- to thick-bedded sandstone, with minor conglomerate and dolostone. The siltstones are commonly carbonaceous with goethite and hematite pseudomorphs (up to 5 mm) after pyrite, and are locally silicified. Sandstone beds are mainly massive to planar and ripple laminated, commonly with well-preserved sole marks. Dewatering dishes and pillars are also common. Conglomerate beds are very localized, generally less than 0.5 m thick, and consist mainly of tabular mudstone or fine-grained sandstone intraclasts commonly preserved as moulds.

With the exception of hummocky and trough cross-stratified units, the Kiangi Creek Formation on EDMUND was deposited mainly below storm wave-base. The planar-laminated carbonaceous and pyritic siltstones are interpreted to be deep-water anoxic deposits. The bulk of the sandstones display characteristics typical of turbidites and can be interpreted in terms of the classical Bouma sequence. The broadly lenticular geometry of sandstone packages, and the marked lateral and vertical variations in siltstone to sandstone ratio, were interpreted by Chuck (1984) as reflecting turbidite fan-lobes. The localized presence of hummocky stratification, trough cross-stratification, and dolostone and dolomitic sandstone indicates that parts of the Kiangi Creek Formation on KENNETH RANGE were deposited above storm wave-base. However, there is no evidence for intertidal or subaerial deposition.

Muntharra Formation (not examined on this excursion)

The Muntharra Formation conformably overlies the Kiangi Creek Formation, and is overlain by the Discovery Formation. It is confined mainly to the area southeast of Doolgarrie Creek on ULLAWARRA, and northwest of Thorpe Bore on ELLIOTT CREEK. The formation is commonly 30–50 m thick and appears to be restricted to an area immediately adjacent to the Talga Fault.

The Muntharra Formation consists mainly of thinly bedded, planar-laminated siltstone, dolomitic siltstone, and dolostone, with interbedded dolomitic sandstone and quartz sandstone. Lithofacies and sedimentary structures of the Muntharra Formation suggest background deposition predominantly below storm wave-base, with localized high-energy deposition of carbonate grainstone, dolomitic sandstone, and sandstone. The absence of stromatolites suggests deposition below the photic zone. This is supported by the presence of sulfide pseudomorphs that suggests deposition and diagenesis in an anoxic environment. The transition from the Kiangi Creek to the Muntharra Formation represents a shoaling of the basin.

Discovery Formation

The Discovery Formation is recognized throughout the Edmund Basin and has been regarded as a major regional marker horizon (Muhling and Brakel, 1985). Its thickness ranges from 50 to 70 m on northern EDMUND (Martin et al., 1999) to an estimated 280 m south on ELLIOTT CREEK (Muhling and Brakel, 1985). Isopachs presented by Daniels (1966) show thickening towards the southwest, with a maximum thickness for the siliceous component of 365 m (Daniels, 1975).

The basal boundary is sharp, and marked by the contact between predominantly carbonaceous siltstone and the top of the last dolostone or sandstone in the Muntharra Formation, or the last sandstone in the Kiangi Creek Formation where the former is absent. A number of features suggest that the basal contact is a regional disconformity. These include apparent very low angle truncation of underlying units, reflected in the localized distribution of the Muntharra Formation, and the presence of basal silicified conglomerate near Strama Bore on ELLIOTT CREEK (MGA 457900E 7397700N). The upper contact with laminated siltstone and silicified planar-laminated dolorudite and dololutite at the base of the Devil Creek Formation is sharp.

Throughout much of the region the Discovery Formation consists predominantly of planar-laminated, pyritic, silicified carbonaceous siltstone with cubic and rhombohedral mineral casts after pyrite or possible evaporite minerals, comprising a lower unit characterized by pinch-and-swell structures and an upper planar-laminated unit (Brakel and Muhling, 1976). The truncation of laminae in pinch-and-swell structures suggests that they may have behaved as concretionary structures with a planar basal surface and undulating upper surface. The pinch-and-swell structure characteristic of undeformed areas is superimposed by tight upright folding with a strongly undulose form surface in tightly deformed areas. Intrafolial folds and sheath folds with axial surfaces parallel to bedding are also common in these areas, particularly in the Mangaroon Syncline and north of the Lyons River Fault. Silicified beds of medium-grained sandstone, pebble conglomerate, and sedimentary breccia are exposed locally.

The origin and depositional environment of the Discovery Formation are controversial and previous authors have argued for both a primary and secondary origin for the chert. Suggested origins include a primary chemical precipitate (Daniels, 1966), silicification of algal mats and local primary deposition (Marshall, 1968), and deposition as a silica gel (Muhling and Brakel, 1985). The sedimentary and petrographic features of the Discovery Formation outlined above suggest that it is a silicified carbonaceous siltstone rather than a primary chert. Deposition appears to have occurred mainly in a starved anoxic environment below storm wave-base. Higher energy deposits are locally abundant, particularly at the base of the formation. Silicification was an early diagenetic feature. The Discovery Formation marks a major marine transgression that crossed the region from southwest to northeast. Silicified sandstone and conglomerate at the base of the formation along the southwestern margin of the Pingandy Shelf are possible remnants of a ravinement surface developed during the initial marine transgression.

Devil Creek Formation

The Devil Creek Formation is a prominent carbonate unit, with a distinctive striped airphoto pattern, that lies conformably above the underlying Discovery Formation and below the overlying Ullawarra Formation (Martin and Thorne, 2002). The succession is 85–200 m thick, and thickens from northeast to southwest across the Talga Fault. The thickest sections are on EDMUND (1:100 000) where up to 450 m of dolomitic siltstone and dolostone are preserved.

The Devil Creek Formation comprises thin- to thick-bedded dolostone, dolomitic siltstone, siltstone, dolostone breccia, dolomitic sandstone, stromatolitic dolostone, and minor chert. The distribution of these lithologies is strongly controlled by the Talga Fault. North of this structure the formation consists of interbedded planar to gently undulatory and hummocky laminated dololutite and massive to imbricated dolostone breccia that contain disaggregated columnar stromatolites locally. South of the Talga Fault the Devil Creek Formation consists mainly of thin- to medium-bedded dolomitic siltstone, with varying proportions of thin-bedded dololutite, fine- to medium-grained dolarenite, and siltstone. Paleocurrent directions recorded by Daniels (1966) are mainly towards the northwest and west.

Chuck (1984) identified four facies within the Devil Creek Formation: algal dolostones (subtidal), dolostone breccias (slope debris flows), turbidites, and parallel-laminated dolostones (deep-water suspension settling). Overall, the distribution of lithofacies in the Devil Creek Formation reflects platform carbonates north of the Talga Fault and basinal mudstones south of the fault. The fault itself is marked by slope-facies breccias, and may have been rimmed by a stromatolite reef. The alignment of columnar stromatolites at Strama Gap is interpreted as reflecting a paleoslope towards the southwest, perpendicular to the Talga Fault. Daniels (1966), however, documented a preferential orientation with tops toward 124°, which was interpreted as the product of a current from the south-southeast.

Ullawarra Formation

The Ullawarra Formation conformably overlies the Devil Creek Formation and its upper unit, the Curran Member, is transitional with the overlying Coodardoo Formation. The formation consists mainly of thin-bedded black and maroon siltstone and mudstone with lesser interbedded fine- to medium-grained sandstone, and minor chert and dolostone. A characteristic feature of the Ullawarra Formation is the large volume of dolerite sills intruded south of the Talga Fault. The sedimentary component varies in thickness from 100 m immediately north of the Talga Fault on ELLIOTT CREEK to 650 m on southern EDMUND.

Siltstones are predominantly thin bedded and parallel-planar to undulatory laminated, and commonly ferruginous to locally calcareous. Tabular to lenticular fine- to medium-grained sandstone beds that are interbedded with parallel-planar-laminated siltstone are commonly less than 0.5 m thick and mostly in the 10 to 20 cm range. These become the dominant lithology in the Curran Member. Sole marks consisting mainly of bounce, brush, prod, and flute marks are commonly well preserved at the base of these beds, and give paleocurrent directions towards the west and northwest, consistent with paleocurrent directions determined from ripple cross-lamination.

The general lack of shallow-water facies, characterized by sedimentary structures formed above wave-base, in the

Ullawarra Formation suggests that it was deposited in a deep-marine shelf environment. Background sedimentation is dominated by suspension settling to form parallel-planar-laminated mudstone, and was interrupted by the deposition of sandstone beds from dilute turbidity currents. Paleocurrent data further indicate that deposition occurred in response to uplift to the southeast of the Edmund Basin. This uplift represents a major change in paleogeography compared to underlying units that were primarily derived from the northeast and northwest. Deposition of the Curran Member marks a period of significantly increased siliciclastic-sediment supply, which culminated in the deposition of the Coodardoo Formation.

Coodardoo Formation

The Coodardoo Formation, the youngest stratigraphic unit in the Edmund Group, has a gradational contact with the underlying Curran Member of the Ullawarra Formation and is unconformably overlain by the Backdoor Formation (Collier Group). The maximum thickness of the formation ranges from 150 to 370 m (Daniels, 1975) and it thins to the southeast. It is dominated by very thick to thick-bedded, very coarse to fine-grained lithic quartz sandstone and pebbly sandstone, interbedded with siltstone.

Sandstone beds are commonly tabular, massive to normally graded, and either amalgamated or capped by very thin parallel-planar-laminated mudstone. Amalgamated beds are commonly massive, 0.3 – 2 m thick, with gently undulating contacts and localized scour-and-fill structures. Graded beds locally contain very coarse grained sandstone to granule–pebble conglomerate at the base, and diffuse horizontal lamination is highlighted by selective silicification. Well-preserved dewatering structures are a characteristic feature of the Coodardoo Formation, and include streaks, dishes, dykes, and disrupted lamination. Paleocurrent directions from sole markings are towards the southeast and northwest. However, it is not possible to determine which of these transport directions predominates, or whether they are axial or transverse to the basin margins.

The graded amalgamated sandstones and interbedded planar-laminated siltstones of the Coodardoo Formation are interpreted as turbidite fan deposits. However, the ambiguity surrounding the paleocurrent data makes it difficult to interpret the significance of these turbidites to basin evolution. Daniels (1966) had earlier suggested that during deposition of the Coodardoo Formation, uplift to the north had changed paleocurrent directions from easterly to southerly.

Collier Group

The Collier Subgroup was elevated to the Collier Group by Martin et al. (1999) during revision of the former Bangemall Group. It has an estimated thickness of about 2 – 2.5 km and is subdivided, in ascending order, into the Backdoor, Calyie, and Ilgarari Formations (Martin and Thorne, 2002).

Detrital zircons from the Backdoor Formation, near the base of the Collier Group, constrain a maximum age of about 1620 Ma for the group (Martin et al., in press). The minimum age of the Collier Group is constrained by a suite of c. 1070 Ma dolerite sills that were intruded into both the Collier Group and the underlying Edmund Group (Wingate, 2002). These sills were intruded into wet sediments (Martin, 2004a), indicating that the depositional age of the Collier Group is close to 1070 Ma and significantly younger than the Edmund Group. Current geochronological constraints imply a hiatus between the Edmund and Collier Groups that is possibly as much as 395 million years.

Backdoor Formation

The Backdoor Formation outcrops in the Wanna Syncline on central and southeastern EDMUND. The formation is about 1500 m thick on southern ELLIOTT CREEK and has a sharp regionally disconformable contact with the underlying Coodardoo Formation (Cooper et al., 1998). The gradational upper contact with the overlying Calyie Formation is characterized by a rapid increase in the sandstone to siltstone ratio. The formation is also intruded by locally discordant dolerite sills.

The Backdoor Formation consists of planar-laminated green, maroon, and cream-coloured siltstone and mudstone, interbedded with subordinate very coarse to fine-grained lithic sandstone, oolitic dolostone, and minor thin chert beds. Fossils of possible metaphytes (Grey and Williams, 1990; Grey et al., 2002) and impressions of microbial mats (Martin, 2004b) are also locally preserved on the soles of fine-grained sandstone beds. Paleocurrent directions determined from sole marks and cross-stratification are consistently towards the southwest in the lower Backdoor Formation sandstones. Paleocurrents in the upper Backdoor Formation are locally directed towards the east-southeast, particularly in the transition to the Calyie Formation.

The parallel-planar-laminated siltstones that form the bulk of the Backdoor Formation were most likely deposited below storm wave-base on a distal siliciclastic shelf. Interbedded fine- to medium-grained, thin-bedded sandstones in the lower Backdoor Formation are the product of storm-generated bottom currents that transported sediment offshore. Southwesterly directed paleocurrent data suggest a source area to the northeast of EDMUND and is consistent with the sediment slumping to the south noted by Daniels (1966). Very coarse to medium-grained, thick-bedded sandstones at the tops of upward-thickening packages in the lower Backdoor Formation, and throughout the upper Backdoor Formation, were probably deposited close to or above storm wave-base, as indicated by the hummocky cross-stratification. The Backdoor Formation therefore records a progradation from distal shelf to proximal shelf and distal delta-front depositional environments.

Calyie Formation

The Calyie Formation is a 190 m-thick, sandstone-dominated unit in the middle Collier Group. The basal contact with the Backdoor Formation is gradational, and characterized by an increasing proportion of sandstone in a succession of interbedded siltstone and sandstone. The upper contact with the siltstone-dominated Ilgarari

Formation is sharp and is placed at the top of the last major sandstone bed in the Calyie Formation.

The Calyie Formation consists mainly of very thick to medium-bedded, medium- to coarse-grained quartz sandstone that is typically massive to trough cross-stratified, and is interbedded with minor planar-laminated siltstone. Paleocurrent directions in the Calyie Formation are variable, with two major opposing components towards the southeast and northwest on ELLIOTT CREEK. Daniels (1966) recorded paleocurrents that were commonly to the north and south.

The rapid facies changes that characterize the Calyie Formation – Backdoor Formation transition are interpreted to reflect a change from a delta-front to delta-top depositional environment (Martin and Thorne, 2004). The bulk of the Calyie Formation was probably deposited in a shallow-marine to fluvial depositional environment, as suggested by the abundant evidence for intraformational reworking and subaerial exposure. Opposing paleocurrent directions recorded by trough cross-strata are possibly due to ebb and flood flow in tidal channels.

Ilgarari Formation (not examined on this excursion)

The Ilgarari Formation is the uppermost unit of the Collier Group. The formation is about 680 m thick on ELLIOTT CREEK, excluding the numerous dolerite sills (Martin et al., 1999). It consists almost entirely of greenish-grey to black, parallel-planar-laminated, pyritic and carbonaceous siltstone that is locally silicified. Thin interbeds of massive to planar-laminated fine-grained sandstone are present in the upper parts of the formation.

Muhling and Brakel (1985) interpreted the Ilgarari Formation as the lagoonal facies of a regressive marine shoreline. However, there is no evidence for lagoonal deposition on EDMUND, and the formation is now regarded as a transgressive, deep-marine shelf deposit (Martin and Thorne, 2004).

Depositional packages

In order to interpret the evolution of the Edmund and Collier Basins the 13 lithostratigraphic formations of the Bangemall Supergroup are arranged into six depositional packages (see numbers on right-hand side of Plate 2), which represent discrete stages in basin development (Martin and Thorne, 2004). The Edmund Group comprises packages 1–4, and the Collier Group comprises packages 5 and 6. Each depositional package consists of an assemblage of genetically related strata, with a basal contact that is either a regional unconformity or a major flooding surface.

Package 1 (Yilgatherra and Irregully Formations)

Package 1 was deposited unconformably on an eroded surface of Paleoproterozoic metamorphosed sedimentary and igneous rocks. The basal succession of fluvial to shallow-marine siliciclastic rocks (Yilgatherra Formation) is overlain by intertidal to shelf and slope carbonate rocks and interbedded fluvial to shallow-marine siliciclastic rocks (Irregully Formation). Paleocurrent data indicate the presence of local low-relief topographic highs in the southwest of the excursion area, although the consistency of Yilgatherra Formation thickness and facies suggests no uplift of these features at this time. Similarly, there were no significant facies or thickness changes across major faults, other than the Talga Fault.

Deepening of the Edmund Basin during deposition of the overlying Irregully Formation appears to have been controlled largely by subsidence across the major basement structures, particularly the Talga Fault. On the Pingandy Shelf, interbedded peritidal and supratidal facies at the base of the formation are overlain by peritidal facies that record an initial upward deepening and drowning. The presence of upward-shoaling cycles in the lower and middle parts of this succession suggests that this transgression was stepped, and possibly related to episodic fault movement. The upper part of the Irregully Formation in this area consist of slope lithofacies. These are capped by a widespread stromatolitic bioherm that records a final shallowing of the carbonate platform to wave base.

The relatively thin succession of peritidal platform facies that characterizes the Irregully Formation on the Pingandy Shelf contrasts markedly with the thick slope-marine-, and fluvial-facies deposits south of the Talga Fault. Immediately south of this structure the lower Irregully Formation is dominated by slope-facies breccias and planar-laminated dolomudstone and siltstone, with localized shallow-marine and fluvial sandstone. In contrast, the middle to upper Irregully Formation in this area is represented by shallow-marine sandstone and stromatolitic dolostone. Between the Lyons River Fault and the faulted southwestern margin of the Wanna Syncline the middle to upper parts of the Irregully Formation are characterized by a marked thickening, totalling several hundreds of metres. This increase was accompanied by the deposition of significant amounts of fluvial to marine siliciclastic facies. Differential movement along the basement faults appears to have stopped by the time the upper part of the Irregully Formation was deposited. This allowed stromatolitic shallow-marine-shelf carbonate facies to blanket much of the area both to the north and south of the Talga Fault.

Package 2 (Gooragoora, Blue Billy, and Cheyne Springs Formations)

Throughout the study area the contact between package 2 and the underlying package 1 is sharp, although it does not show evidence of significant erosion. Package 2 sedimentation occurred in response to a rise in relative sea level and drowning of the Pingandy Shelf, an event that resulted in deposition of planar-laminated siltstone at the base of the Gooragoora Formation throughout the study area. In contrast to package 1, thickness, facies, and paleocurrent trends in package 2 suggest that northwest-trending basement structures such as the Talga Fault had little influence on sedimentation at this time. Coarse-grained deltaic sedimentation in the Gooragoora Formation was restricted to the northwestern parts of the study area. Here, the progradation of deltaic sandstones during a period of high or rising sea level suggests that uplift to the northwest exceeded the rate of sea level rise.

Deposition of deep-marine anoxic mudstone of the overlying Blue Billy Formation records the maximum extent of the package 2 transgression and expansion of the Edmund Basin beyond the preserved limits of the Pingandy Shelf. Localized turbidite and mass-flow deposits, which are present in the upper part of the Blue Billy Formation in the northwest, were deposited both north and south of the Talga Fault. The gradational contact between the Blue Billy Formation and distal shelf facies of the overlying Cheyne Springs Formations marks a slight shallowing of the Edmund Basin. The fact that shallow-marine facies are most abundant in southeastern parts of the Pingandy Shelf suggests that there may have been a gentle east to west paleoslope during the latter stages of package 2 deposition.

Package 3 (Kiangi Creek and Muntharra Formations)

The interval between the deposition of packages 2 and 3 was marked by a period of uplift and erosion in southeastern parts of the Edmund Basin that was accompanied by basement uplift between the Lyons River Fault and the southwestern limb of the Wanna Syncline. On the Pingandy Shelf pre-package 3 erosion resulted in a gradual southeasterly thinning of the Cheyne Springs Formation, whereas the localized tilting and erosion associated with movement along the Lyons River Fault resulted in the removal of package 2 as well as the top part of package 1 in the area southeast of Edmund Homestead.

Pre-package 3 erosion was followed by relative uplift to the north and east, tilting to the southwest, and renewed southwest-block-down movement along the Talga Fault. The resulting rise in relative sea level over most of the area culminated in the deposition of a shallow to deeper marine-shelf deposit (Kiangi Creek Formation) and the overlying locally preserved carbonate unit (Muntharra Formation). On the southeastern part of the Pingandy Shelf the Kiangi Creek Formation comprises northeasterly derived deltaic facies with local stromatolitic carbonate deposits. When traced to the northwest and southwest, however, these facies are transitional into deeper marine-shelf deposits consisting largely of very thick to thin-bedded sandstone turbidites and mudstone. The thickness and proportion of sandstone increase significantly on the southwestern side of the Talga Fault; however, this trend was not observed across the other major structures. The upper levels of package 3 appear to have been deposited in response to a shallowing of the Edmund Basin along the Pingandy Shelf, which was accompanied by a marked decrease in siliciclastic supply to the basin. During this time high-energy carbonate grainstone of the Muntharra Formation was deposited along the line of the Talga Fault, whereas sulfidic mudstone accumulated in deeper water towards the southwest.

Package 4 (Discovery, Devil Creek, Ullawarra, and Coodardoo Formations)

The base of package 4 is marked by a major marine transgression that appears to have crossed the region from southwest to northeast. Despite the lenticular nature of the Muntharra Formation at the top of package 3, which may reflect original depositional controls, there is no compelling evidence for either fault movement or significant uplift and erosion in the interval immediately prior to deposition of package 4. Silicified sandstone and conglomerate that occur at the base of the Discovery Formation along the southwestern margin of the Pingandy Shelf are possibly remnants of a ravinement surface developed during the initial marine transgression. These basal coarse-grained units are overlain by silicified sulfidic mudstone and chert, which are interpreted to reflect deposition under the ensuing starved anoxic basinal conditions. Because of the widespread uniform nature of the Discovery Formation across the region it is likely that these starved basinal conditions persisted across much of the northwestern Capricorn Orogen. During this time there appears to have been no significant movement along the major basinal structures, other than a possible northeast-block-down movement of the Lyons River Fault to accommodate the increased thickness of the Discovery Formation towards this structure.

A more significant reactivation of the Talga Fault, which was accompanied by a slight shallowing of the basin, appears to have occurred during deposition of the overlying Devil Creek Formation. Platform-facies carbonates, including intraclast grainstone and stromatolitic dolostone, were deposited on the Pingandy Shelf, whereas the line of the Talga Fault was marked by slope-facies carbonates, including significant accumulations of intraformational breccia. Southwest of this line the slope facies pass into thick, fine-grained mixed siliciclastic–carbonate basinal deposits that are uniform across the southern part of the area, suggesting little influence from other basement structures in this part of the region.

Deposition of the remainder of package 4, comprising the Ullawarra and Coodardoo Formations, was characterized by further basin subsidence accompanied by a gradual increase in siliciclastic supply. Deep-marine-shelf deposits of the Ullawarra Formation were laid down in response to uplift to the southeast of the Edmund Basin, as well as a northwesterly tilting of the basin floor. As was the case with deposition of the Devil Creek Formation, the Talga Fault probably acted as a major growth structure at this time, resulting in a marked southwesterly increase in thickness of the Ullawarra Formation. Upper parts of the Ullawarra Formation were deposited during a period of significantly increased siliciclastic sediment supply, and localized shallowing to carbonate facies along the Talga Fault. During this time, however, basin subsidence continued to keep pace with the increased sedimentation rates. Thin- to thick-bedded turbidite facies in the upper part of the Ullawarra Formation show a southeasterly derivation and may reflect axial flow. However, limited paleocurrent data from the overlying deeper shelf sandstones of the Coodardoo Formation point to an increasing contribution from an upland area to the northwest of the Edmund Basin.

Packages 5 (Backdoor and Calyie Formations) and 6 (Ilgarari Formation)

Deposition of package 4 was followed by an interval during which the Edmund Group was tilted gently towards the west and southwest, uplifted and eroded in the east, and intruded by the first of two generations of dolerite sills at about 1465 Ma.

Package 5 comprises the Backdoor and Calyie Formations, and package 6 is represented by the Ilgarari Formation. Deposition of package 5 records an initial period of carbonate shelf sedimentation, followed by a major marine transgression (lower Backdoor Formation) and then by the progradation of a sand-dominated delta complex (upper Backdoor Formation and Calyie Formation) across the region. Facies distributions in the carbonates and paleocurrent data from sandstones in the Backdoor Formation indicate a southwesterly sloping sea floor that was probably a remnant of pre-package 5 tilting. Northwesterly progradation of deltaic facies in the Calyie Formation, at the top of package 5, records a further change in paleogeography of the region, and possibly reflects uplift to the southeast.

The deposition of package 5 was terminated abruptly by a further marked rise in relative sea level and transgression that resulted in the deltaic deposits of package 5 being blanketed by deep-marine-shelf mudstone of the Ilgarari Formation. Paleocurrent data from thin-bedded sandstone turbidite deposits indicate that the paleoslope was still to the northwest. Because of the limited present-day outcrop of the Collier Group, which is largely confined to the axis of the Wanna Syncline, the influence of the major structures during deposition of packages 5 and 6 is unclear. The presence of a southwesterly dipping paleoslope during deposition of the Backdoor Formation may in part reflect relative uplift of the Pingandy Shelf. Small northeast-trending normal faults, which cut the northeastern limb of the Wanna Syncline, appear to have had little if any influence on sedimentation at this time.

Dolerite sills

Two chemically and geochronologically distinct suites of dolerite sills have intruded the Bangemall Supergroup (Wingate, 2002; Geological Survey of Western Australia, 2006; Morris and Pirajno, 2005), but are indistinguishable in outcrop and thin section. The older suite of sills has been dated at c. 1465 Ma, and the younger suite at c. 1070 Ma (Wingate, 2002; Geological Survey of Western Australia, 2006) using SHRIMP U–Pb geochronology of zircon and baddeleyite extracted from granophyric zones within coarse-grained dolerite and gabbro. Furthermore, regional correlation of each suite is facilitated by their distinctive geochemistry (Morris and Pirajno, 2005) and paleomagnetic remanence (Wingate, 2002). On the basis of available data all 1465 Ma sills have intruded the Edmund Group, and the bulk of the 1070 Ma sills have intruded the Collier Group. This younger suite is the more widespread, and can be correlated with mafic intrusions of a similar age that have been identified well beyond the limits of the Bangemall Supergroup, and constitute the 1078–1070 Ma Warakurna large igneous province (Wingate et al., 2004).

Although most sills are locally concordant with the host sedimentary rocks, they are regionally discordant along strike and characterized by a step-and-stair geometry (Francis, 1982). They are generally moderately to poorly exposed, and tend to be highly weathered. The best exposed dolerites are characterized by poorly exposed chilled margins of fine-grained dolerite that are commonly covered with scree derived from the host rocks. However, good exposures of these contact zones indicate that in some cases they are marked by folding or brecciation of the host rocks, plastically deformed xenoliths, amygdales, vesicles, varioles, and rare peperitic textures. These features suggest that some of the sills were emplaced at relatively shallow depths into unconsolidated or partially lithified sediments.

Dolerite sills intruded into the Bangemall Supergroup generally consist of a granular intergrowth of approximately equal proportions of subhedral bladed to tabular pyroxene and plagioclase, with minor amounts of opaque oxide. Although they are similar in thin section the two suites of dolerite sills are compositionally distinct in terms of their major and trace element geochemistry (Morris and Pirajno, 2005). Both suites are broadly tholeiitic in composition, and show enrichment in low field strength elements (LFSE) relative to high field strength elements (HFSE) that is indicative of melting of a subduction-modified mantle source (Morris and Pirajno, 2005). Differences between the two suites are reflected in variations of absolute concentrations of Zr, Th, U, Cr, Ni, Mn, Cu, Ga, Sn, and Rb at a given magnesium number (Mg#). The available geochemical and geochronological data indicate that most of the sills intruding the Edmund Group have compositions consistent with the c. 1465 Ma suite, and those intruding the Collier Group with the c. 1070 Ma suite (Morris and Pirajno, 2005). The exceptions in the case of the Edmund Group are sills intruding the Pingandy Shelf, and a sill at the top of the Ullawarra Formation south of the Talga Fault.

Regional- and outcrop-scale observations suggest that dolerite sills were intruded into the Edmund and Collier Groups shortly after deposition of each group and prior to regional deformation during the Edmundian Orogeny. The anorogenic setting of the sills has led Wingate et al. (2004) and Morris and Pirajno (2005) to propose that the younger of the two suites was emplaced as part of the Warakurna large igneous province, which is interpreted as the product of a mantle plume. The regional distribution and hence origin of the older suite is less well known, but is clearly also related to large-scale anorogenic magmatism within the Mesoproterozoic West Australian Craton. Paleomagnetic evidence suggests that these sills were probably intruded during the early stages of assembly of the Rodinia supercontinent (Wingate, 2002; Morris and Pirajno, 2005).

Neoproterozoic Edmundian Orogeny

The Edmundian Orogeny (Halligan and Daniels, 1964) is an intracratonic event that resulted in the locally pervasive deformation of the Bangemall Supergroup, and the formation or reactivation of structures in the underlying basement. The orogeny largely resulted in north–south shortening that formed easterly to southeasterly trending, open to tight, upright folds and normal, reverse, and strike-slip faults. The area affected by this regional deformation is here referred to as the Edmund Fold Belt.

Bangemall Supergroup rocks in the Edmund Fold Belt are divided into three structural zones referred to informally as the northeastern, central, and southwestern zones. Within these zones two main periods of Edmundian deformation and low-grade metamorphism are recognized. The first of these (D_{1e} /M_{1e}) is responsible for much of the currently observed structural geometry and involved mostly north–south shortening, whereas the second (D_{2e}/M_{2e}) reflected initial compression, then extension along an east-southeast to west-northwest line, followed by dextral strike-slip movement.

The northeastern zone is northeast of the Talga Fault and corresponds broadly to the Pingandy Shelf of Muhling and Brakel (1985). It is characterized by a gentle regional tilt of 5° to 25° to the southwest. Rocks of the northeastern zone are cut locally by small, steeply dipping, southeasterly or east-southeasterly trending faults.

The central zone equates to the Wanna Syncline, which is an asymmetric southeasterly plunging synclinorium between the Talga Fault and the main Gascoyne Complex outcrop to the southwest. It appears to have developed over a reactivated system of late ?Paleoproterozoic to early Mesoproterozoic (Edmund Group) grabens and horst blocks (Muhling and Brakel, 1985). The Talga Fault is a major northwest-trending normal fault that has undergone significant reverse movement during the Edmundian Orogeny. Its throw decreases from the northwest, where it juxtaposes rocks of the upper Ashburton Formation (Wyloo Group) against the Kiangi Creek Formation (middle Edmund Group), to the southeast where it cuts the upper Edmund Group. The southwestern limb of the Wanna Syncline is more extensively deformed than the northeastern limb and comprises a series of large-scale anticlines and synclines. Most folds are upright open structures, although they may be tightened and inclined locally near exposed faults. Plunges are mostly gentle, toward either the southeast or northwest. Axial traces of the larger folds are commonly curvilinear, reflecting the influence of gentle refolding during D_{2e}.

The southwestern zone is represented by the more highly deformed rocks of the Mangaroon, Cobra, and Ti Tree Synclines. Bangemall Supergroup rocks in the southwestern zone experienced generally higher levels of D_{1e} strain than either the northeastern or central zones. Rocks within the major synclines show a marked contrast in competency, ranging from massive thick dolerite sills to thin-bedded siltstone and sandstone. The flanks and cores of the synclines are cut by steep, northwest-trending faults that have undergone initial normal movement followed by transpression. Away from the faults, bedding dips are 20–50°, whereas closer to these fractures dips may be vertical or locally overturned. Traces of major and minor fold axes are commonly subparallel to the major faults and the folds are characterized by a steeply dipping penetrative axial-planar cleavage. Bedding surfaces of tightly folded silicified siltstone may exhibit a strong movement striation (slickenside) lineation developed parallel to the direction of flexural slip. Locally this lineation may be folded about the hinges of steeply plunging small-scale intrafolial folds developed close to major faults.

On southwestern EDMUND (around MGA 413100E 7348200N) the northwestern extension of the Cobra Syncline is preserved between two major northwest-trending Edmundian faults. The most northerly of these, the Lyons River Fault (Lyons River Lineament of Pearson et al., 1996), preserves a south-block-down sense of movement and can be traced as an aeromagnetic lineament from southwestern EDMUND (around MGA 403100E 7358500N) to the eastern part of MOUNT AUGUSTUS. The unnamed bounding fault on the south side of the syncline extends from central-eastern MANGAROON (MGA 381600E 7379200N), where it juxtaposes Gascoyne Complex rocks to the south against lower Edmund Group on the north, to the southern limb of the Cobra Syncline on MOUNT AUGUSTUS where it is interpreted from aeromagnetic data. Both fault traces are aligned subparallel to fold-axial traces in the Cobra Syncline, suggesting a strong northeast–southwest transpressional stress regime during D_{1e}.

The metamorphic grade of M_{1e} was low throughout most of the northeastern zone, with generally good preservation of primary sedimentary and diagenetic textures. Buick et al. (1995) and Buick and Knoll (1999) recorded a southeast to northwest increase in kerogen colour index within the zone, corresponding to a slight increase in minimum metamorphic temperature from 100–125°C at Fords Creek (MGA 529000E 7370000N on KENNETH RANGE) to 150°C near Irregully Creek (around MGA 449800E 7419200N). The metamorphic grade of M_{1e} in the central zone, although generally low, appears to be higher than in the northeastern zone. The typical assemblage for siltstones within the Edmund Group consists of clay minerals, sericite, chlorite, quartz, feldspar (K-feldspar and albite–oligoclase), and epidote, together with variable amounts of iron sulfide, iron oxide, and highly mature kerogen. Rock-Eval pyrolysis of a sample of carbonaceous siltstone from the lower Discovery Formation on MAROONAH (MGA 382578E 7425833N) indicates a maximum temperature during burial of 200–250°C (Ghori, A., 2004, written comm.). The typical metamorphic mineral assemblage for siltstones within the southwestern zone is similar to those in the central zone and suggests a low metamorphic grade during M_{1e}.

Myers et al. (1996) suggested that the Edmundian Orogeny resulted from the collision of the North and West Australian Cratons between 1300 and 1100 Ma. However, dolerite sills dated at c. 1070 Ma (Wingate, 2002; Geological Survey of Western Australia, 2006) that intrude the Bangemall Supergroup are also deformed into easterly trending folds. The Bangemall Supergroup and the dolerite sills are refolded about a north to northeasterly trend to form dome-and-basin structures. Undeformed dolerite dykes that cut both sets of folds have been dated at c. 755 Ma (Wingate and Giddings, 2000). This date, and the age of the deformed dolerite sills, constrains the Edmundian Orogeny to between 1070 and 755 Ma.

The Edmundian Orogeny may be associated with the c. 900 Ma thermal event identified from the Errabiddy Shear Zone in the southern Capricorn Orogeny (Occhipinti, S. A., and Reddy, S. M., 2002, pers. comm., quoted in Cawood and Tyler, 2004). Cawood and Tyler (2004) interpreted the Edmundian Orogeny as a far-field reactivation of the Capricorn Orogen related to the break up of Rodinia and with events associated with Gondwana assembly (Fitzsimons, 2003).

Excursion localities — Edmund and Collier Groups

Day 1

The excursion will begin with a 2½ hour drive from the Mount Augustus Tourist Complex to 'Tchintaby Ridge' on the northern margin of the Edmund Basin (Plate 1), followed by the examination of three localities on the Pingandy Shelf. Depositional environments on the Pingandy Shelf are generally much shallower than those south of the shelf (Martin and Thorne, 2004), and one of the main objectives of this excursion will be to compare and contrast these different environments. The excursion will also examine subtle differences in the stratigraphy of the Edmund Group between these regions that are influenced by facies changes, and the presence of low-angle regional unconformities that bound the six depositional packages identified within the Bangemall Supergroup (Plate 2).

At Locality A1 on the northern slopes of 'Tchintaby Ridge' we will examine the unconformable contact between the Ashburton Formation and the Irregully Formation at the base of package 1. After a short drive south to Locality A2, we will examine siliciclastic and carbonate facies of the Kiangi Creek Formation, near the base of package 3, as well as a ferruginous duricrust that forms part of the extensive and complex regolith cover in the Bangemall Geomorphological Province. At Locality A3 we will examine the Blue Billy Formation at the base of package 2, and one of the numerous dolerite sills belonging to the 1070 Ma Warakurna large igneous province (Wingate et al., 2004) that intrudes the Bangemall Supergroup. The first night will be spent camping close to Locality A4.

Locality A1: Ashburton Formation – Irregully Formation unconformity at 'Tchintaby Ridge' (MGA 593915E 7347176N)

From the Mount Augustus Tourist Complex, drive about 138 km north along the Mount Augustus – Pingandy road to the intersection with the Ashburton Downs – Meekatharra road (at about MGA 594150E 7345520N). ***Drivers should take note that this road, as well as many others taken on this excursion, has a large number of dangerous turns, dips, and crests. Extra care should be taken at all times during the drive between excursion localities.***

Turn left (northwest) at the intersection, and drive a further 1.5 km to Locality A1, on the north side of 'Tchintaby Ridge'. Turn south off the road and park in the open area to the north of the ridge. From the vehicles, walk about 300 m west to a prominent gully draining north from the ridge. Locality A1a (MGA 593915E 7347176N) is in the Ashburton Formation in this gully, close to the foot of the slope.

Looking southward from the flats to the north of 'Tchintaby Ridge' provides an overview of the escarpment section (Fig. 2). 'Tchintaby Ridge' is a good illustration of the structural relationships between the Ashburton Formation and the overlying Bangemall Supergroup, with the latter providing the bulk of the vertical outcrop in the cliffs. At the top of the section the Bangemall Supergroup is represented at this location by the Irregully Formation, which is a stromatolitic dolarenite to dololutite with interbedded thin sheets of quartz sandstone that dips about 10–15° towards the southeast and south-southeast. The slopes below the Irregully Formation consist of highly deformed sandstone, siltstone, and minor conglomerate of the Ashburton Formation. On the horizon to the northwest (bearing 310°), a range of low hills marks Mount Boggola (681 m), where a series of mafic volcanic rocks in the Ashburton Formation is capped by the Capricorn Group. To the north the distinctly rounded range is Mount Bresnahan (683 m), which comprises Bresnahan Group conglomerates.

The Ashburton Formation at this locality consists of massive, medium-grained sandstone interbedded with granular sandstone, felsic volcaniclastic sandstone and breccia, and conglomerate. Beds dip 35–45° towards the south-southeast and local examples of normal grading and ripple cross-lamination indicate that they also young to the south. Locally, bedding defines a small-scale F_{2a} fold with S-vergence. Unlike the Ashburton Formation elsewhere in the region, there is only one prominent cleavage (S_{2a}) that dips steeply (76°) towards the south-southwest. This cleavage and associated quartz veins are truncated at the unconformity.

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Figure 2. View south of 'Tchintaby Ridge' showing the position of the unconformity between the Irregully Formation of the Edmund Basin (above) and the Ashburton Formation of the Ashburton Basin (below; dashed line). Note the light-coloured silcrete cap at the top of the ridge (MGA 593915E 7347176N)

From Locality A1a, walk about 300–400 m east along the foot of 'Tchintaby Ridge' where details of the unconformity and lithologies within the Irregully Formation can be seen in some large fallen blocks at Localities A1b (MGA 593886E 7347165N), A1c (MGA 593965E 7347154N), and A1d (MGA 593930E 7347124N), at the foot of the slope.

The fallen block at Locality A1b shows lithofacies typical of the Irregully Formation on the Pingandy Shelf. These include microbially laminated dololutite to dolorudite, low synoptic relief domical stromatolites (*Panniscollenia*), imbricate platy intraclast breccia, sandy dolostone, and thin very coarse to coarse-grained sandstone beds. These facies are interpreted to have been deposited on a shallow-marine shelf, above storm wave-base.

When viewed from the base of 'Tchintaby Ridge' the unconformity at the base of the Irregully Formation appears relatively planar, and there is little evidence of weathering. These features can be seen in the fallen block at Locality A1c (Fig. 3). However, the contact is locally irregular with pockets of conglomerate consisting of brecciated siltstone in a dolomite matrix extending up to 10 m below the main contact. The lowest continuous Irregully Formation unit is a 0.2 m-thick dolomitic quartz conglomerate overlain by wavy laminated dololutite, interbedded with flat-pebble dolorudite. These dolostone layers can contain up to 50% quartz sand and granules, as seen at Locality A1b.

Dolomitic units within the Edmund Group, particularly the Irregully Formation, are characterized by extensive development of silcrete and brecciated siliceous caprock. Scattered occurrences of ferruginous silcrete are also recorded within the Bangemall Geomorphological Province (Martin et al., 2005). Daniels (1967) considered these deposits to form part of the Bangemall Group (now Supergroup); however, they post-date Edmundian folding and intrusion of the Mundine Well dyke swarm at 755 Ma (Wingate and Giddings, 2000). An example of the silcrete cap that overlies the Irregully Formation, which forms the light-coloured unit at the top of the ridge (Fig. 2), can be seen in the fallen block at Locality A1d. This silcrete consists of silicified dolostone clasts cemented by quartz.

Figure 3. Fallen block showing the unconformity (dashed line) between the Irregully Formation (above) and Ashburton Formation (below) at 'Tchintaby Ridge'. Younging is towards the upper left (MGA 593886E 7347165N)

Locality A2: Sandstone and dolostone with *Conophyton* stromatolites in the Kiangi Creek Formation (MGA 587431E 7342615N)

From the 'Tchintaby Ridge' locality, drive 1.5 km south along the Ashburton Downs – Meekatharra road to the turn-off to Mount Augustus. About 8.5 km along the Pigandy – Mount Augustus road, turn off the road to the north (at about MGA 587180E 7341980N) and drive along the flats on the west side of a small creek for about 600 m to where a low ridge of dolostone is cut by the creek (at MGA 587428E 7342615N). From the west bank of the creek, walk about 300 m west-northwest to Locality A2a (MGA 587245E 7342732N) where Kiangi Creek Formation sandstone is exposed in a small creek bed north of a ferruginous duricrust.

On the Pingandy Shelf the Kiangi Creek Formation consists of interbedded sandstone, siltstone, and local stromatolitic dolostone. The base of the Kiangi Creek Formation is marked by a regional unconformity, which on the shelf cuts down through a large part of the Cheyne Springs Formation at the top of package 2, between Fords Creek (about 65 km to the west) and this locality.

Locality A2a consists of medium- to coarse-grained, planar, undulatory, and ripple-laminated sandstone at the top of the lowermost sandstone unit of the Kiangi Creek Formation. Ripple bedforms preserved on the bedding surface are indicative of combined flow to the southwest (Fig. 4), consistent with regional paleocurrent flow patterns in the Kiangi Creek Formation on the Pingandy Shelf (Martin and Thorne, 2004). In a measured section in Fords Creek this sandstone is about 10 m thick and trough cross-stratified. The sandstone is overlain by a unit of stromatolitic dolostone that will be examined at Locality A2c. The depositional environment is interpreted to be relatively shallow marine, above fair weather wave-base.

From Locality A2a, walk about 50 m southwest to A2b (MGA 587217E 7342684N) at the top of the low ridge of ferruginous duricrust.

Ferruginous duricrust is developed over a wide range of siliciclastic and carbonate rocks in the Edmund Group as well as the associated dolerite sills. The deposits range from thin cappings of bedded or vuggy ironstone with disseminated quartz veins to pisolitic laterite deposits up to 10 m thick. In many areas stratigraphic relationships indicate that ferruginous duricrust is the oldest recognizable component of the regolith, and is commonly overlain by relict alluvium and calcrete.

The sandstone outcrop at Locality A2a is overlain by a chaotic boulder scree of sandstone and siltstone clasts

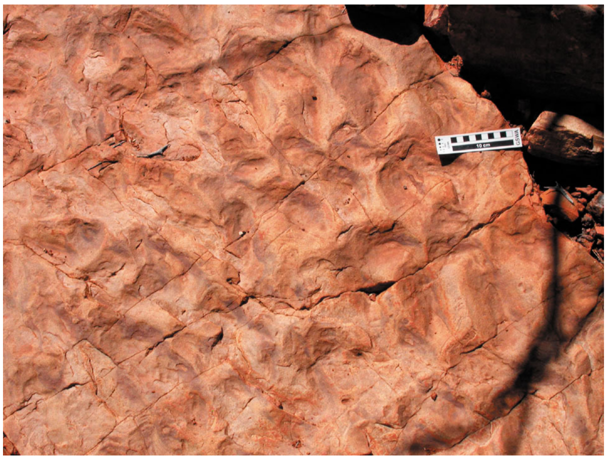

Figure 4. Combined flow ripples in Kiangi Creek Formation sandstone. Plan view of top surface (MGA 587245E 7342732N)

cemented by ferruginous duricrust. This scree is overlain at Locality A2b by a finer grained ferruginous duricrust consisting of iron-replaced bedrock and pisolitic fragments in a hematite-rich matrix. Although not present here, fossil wood fragments are common in the pisolitic ferruginous duricrusts of the region — a feature that allows them to be correlated with the Robe Pisolite (Channel Iron Deposits) of the Hamersley Basin. The views to the west and north show that this unit is up to about 25 m thick and forms a valley-fill that mantles much of the outcrop. The degree of incision of this unit by the modern drainage gives an indication of its relative age; this is considered one of the older relict regolith units in the region and is probably Cenozoic.

From Locality A2b walk about 200 m southeast to a prominent outcrop of stromatolitic dolostone at Locality A2c (MGA 587431E 7342615N).

At Locality A2c large conical bioherms of *Conophyton* stromatolites are within a unit of grey recrystallized dolomitic microspar that overlies the medium- to coarse-grained quartz sandstone seen at A2a. These two units correlate with similar units in the Kiangi Creek Formation mapped on KENNETH RANGE, about 56 km to the northwest, where the *Conophyton* bioherms are much smaller. At Locality A2c many individual cones are up to 2.5 m high and 2 m wide at the base, and clustered into bioherms. The sides of the cones dip at 58° and most cones have a narrow axial zone of highly steepened lamina crests that is commonly less than 10 mm wide and commonly slighter lighter in colour than the rest of the cone. Rare cones have small (finger-size) branches growing upward from the wall of the cone. The outcrop has been deeply weathered, and at first appearance many cones now resemble domes, but careful inspection will reveal the small axial zone in the core.

Each lamination is only a few millimetres thick and forms a uniform coating over the entire cone surface. The bacterial mats that formed the cones have modern counterparts in hot-spring environments like Yellowstone (U.S.A.) and Rotorua (New Zealand). The bacteria influence the local microenvironment and mediate precipitation of any minerals that are close to saturation in the water. In the case of modern hot springs this can be silica, but at this locality the mineral was probably carbonate. The uniform coating of the cone can only occur if the structure is fully below wave-base, therefore the height of the cones can be used to estimate minimum water depth (in the case of eroded cones the probable height can be calculated from the diameter and axial angle using simple trigonometry). The largest cones are more than 1 m wide at the base, and the original relief is estimated to have ranged from 3 to 9 m (Grey, 1985). Bioherms are up to 4 m in diameter. The difference in size of the cones at this stratigraphic level implies a general shallowing to the northwest, between 'Tchintaby Ridge' and Wandarry Creek (on KENNETH RANGE).

The stromatolites are composed of dense carbonate and must have lithified as they formed, but the intercolumn areas were probably much more porous and did not lithify until later. These areas between the bioherms could have provided a conduit for later mineralizing fluids.

Locality A3: Dolerite sill in the Blue Billy Formation (MGA 545457E 7361189N)

From Locality A2 return to the main road and turn right (west), then drive about 47 km to the turn-off to Scoophole Yard and Tea Tree Well (at about MGA 548350E 7351020N). Turn right (north) and drive for about 17 km along the station track to Locality A3, northwest of Tea Tree Well. Bear left at Tea Tree Well (about MGA 549690E 7358160N) and at the track junction (at about MGA 548020E 7359480N), and then right at the next track junction (at about MGA 546180E 7358980N. From the track walk about 70 m northeast to Locality A3a (MGA 545457E 7361189N).

The Blue Billy Formation is not well exposed in this area, due mainly to the fact that it is intruded by a thick dolerite sill, and is commonly covered with relict alluvium. However, Locality A3a provides a good example of the planar-laminated, fissile to flaggy siltstone that is typical Blue Billy Formation. In fresh exposures this siltstone is characteristically carbonaceous, but the bleached and partially silicified siltstone preserved here is more typical of Blue Billy Formation outcrops. The Blue Billy Formation is typically also pyritic, and has been considered highly prospective for base metal deposits. The Reiffel Zinc prospect is hosted within the Blue Billy Formation (formerly Jillawarra Formation) about 20 km southeast of this locality.

From Locality A3a, walk about 70 m northeast to Locality A3b (MGA 545506E 7361243N), which consists of some prominent boulders of dolerite.

The thick dolerite sill that intrudes the Blue Billy Formation in this area is well exposed at Locality A3b. Here the sill consists of even-textured, coarse-grained dolerite to gabbro, and contains small amounts of quartz. Chemical analysis of this sill indicates that it belongs to the 1070 Ma suite, as do all the sills that intrude the

Bangemall Supergroup north of the Talga Fault. Sills belonging to this suite are largely restricted to this area, and to the Collier Group. Sills that intrude the Edmund Group south of the Talga Fault belong almost exclusively to the 1465 Ma suite, with the exception of a sill locally preserved immediately below the Coodardoo Formation that has been dated at 1070 Ma (Wingate, 2002). Dolerite sills intruded into the Edmund and Collier Groups commonly display planar upper and lower contacts, with minimal contact metamorphism, indicative of intrusion into relatively dry lithified sediments. However, excellent evidence for intrusion into wet partially lithified sediment has been documented in the case of one 1070 Ma sill (Martin, 2004a) and one interpreted 1465 Ma sill.

From Locality A3 drive about 2.7 km south along the track, back towards Tea Tree Well, then turn right (west) onto the two-wheel track to Mothersday Yard (at about MGA 546180E 7358980N). After driving for about 5 km along this track, turn right (north) onto the graded track to Mothersday Yard (at about MGA 541900E 7358890N). Proceed along this track for about 5 km to an intersection where the track turns sharply north (about MGA 538030E 7360890N). Continue west along the track for about 6 km to where the track starts to cross a small ridge of chert (at about MGA 532970E 7362950N). Turn right off the track and drive about 800 m along the flats on the north side of the low chert ridge, to the first camp site.

Day 2

On day 2 the aim is to continue to examine the Edmund Group on the Pingandy Shelf, before moving on to the Collier Group south of the Talga Fault. At Locality A4 we will examine the Discovery Formation at the base of package 4. This unit has been interpreted as a primary chert and used as a regional marker horizon by previous workers (e.g. Muhling and Brakel, 1985), but is now considered to consist of a number of silicified clastic lithologies that overlie a subtle regional unconformity. At Locality A5 we will see the typical carbonate lithofacies of the Devil Creek Formation, and at Locality A6 we will examine silicification of these facies at the contact with the overlying Ullawarra Formation. The excursion will then move up to the contact between the Collier and Edmund Groups, and at Locality A7 will examine the unconformable contact between the Ullawarra and Backdoor Formations, in packages 4 and 5 respectively. At Locality A8 we will examine the enigmatic ‘string of beads’ fossils in the Backdoor Formation, and then proceed to Locality A9 where a silicified calcrete cap is preserved. The last outcrop of the day at Locality A10 is in sandstone of the Calyie Formation, near the middle of package 5. From Locality A10 we will return to the Mount Augustus Tourist Complex where we will spend the night.

Locality A4: Discovery Formation (MGA 532238E 7363253N)

From the camp site, walk about 100 m southwest to Locality A4.

The Discovery Formation on the Pingandy Shelf overlies a variety of lithological units within the Kiangi Creek Formation on a low-angle unconformity. Although the Discovery Formation is extensively silicified throughout the region, precursor conglomerate, sandstone, and carbonaceous siltstone can be identified on the Pingandy Shelf. In particular, this area records evidence for the formation of early diagenetic chert nodules in siltstone, as opposed to primary silica precipitation. In Fords Creek the Discovery Formation also contains large chert olistoliths that are several metres in diameter, as well as berthierine-bearing sandstone indicative of deposition in a deep-water (anoxic) environment. The Discovery Formation as a whole is interpreted as a transgressive deposit.

Locality A4 consists of planar to undulatory laminated, black to grey chert that is typical of the Discovery Formation. Silicified clastic texture is preserved locally, and there is also evidence for much younger silcrete brecciation. The primary lithologies are difficult to decipher at this locality, but were most probably siltstone and sandstone. At this locality the Discovery Formation overlies stromatolitic dolostone of the Kiangi Creek Formation, but further west in Fords Creek it overlies siltstone at a higher stratigraphic level in the Kiangi Creek Formation.

Locality A5: Devil Creek Formation in Fords Creek East (MGA 531321E 7362543N)

From the camp site, return to the main track and turn right (west), then drive about 2 km to where the track crosses Fords Creek East (about MGA 531250E 7362590N). Park in the creek and walk about 100 m east to Locality A5a (MGA 531321E 7362543N) on the north bank of Fords Creek East.

The Devil Creek Formation on the Pingandy Shelf consists mainly of parallel-planar to undulatory laminated dololutite interbedded with tabular dolostone intraclast breccia. Tabular microbial lamination is preserved in thin beds and stromatolites are generally absent, except along the shelf edge at Irregully Creek where they have been incorporated into slope breccias.

Locality A5a is a typical example of the dololutite facies, and consists of parallel-planar laminated dololutite with thin interbeds of microbially laminated dololutite. Local differential compaction is concentrated within the microbially laminated beds. This facies is interpreted to have been deposited below storm wave-base on a carbonate shelf.

From Locality A5a walk back to where the track turns onto the south bank of Fords Creek East, then walk a further 100 m south-southwest to Locality A5b (MGA 531224E 7362512N) on a prominent bench of dolostone exposed on the south bank of the creek.

Locality A5b is a typical example of interbedded dololutite and intraclast breccia (Fig. 5). The intraclast breccias are tabular to lenticular bedded and tabular intraclasts display foreset imbrication indicating southerly transport. Ripple cross-lamination is also preserved in

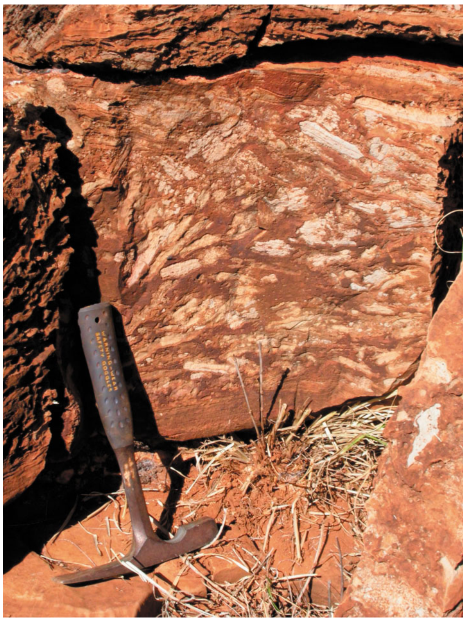

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Figure 5. Tabular intraclast dolomite breccia in the Devil Creek Formation. Note weak imbrication in the upper right corner (MGA 531224E 7362512N)

dololutite immediately overlying the intraclast breccia. The tabular intraclast breccias on the Pingandy Shelf are interpreted to have been deposited below storm wave-base by strong offshore storm surges (as indicated by the direction of clast imbrication).

Locality A6: Chert breccia at Devil Creek Formation – Ullawarra Formation contact (MGA 530744E 7361416N)

From Fords Creek East drive about 1.8 km south along the track to the flat area south of a prominent small hill of ferruginous duricrust. Locality A6 is about 150 m north of the track, on the south side of the duricrust outcrop.

On Kenneth Range and for some distance to the southeast the contact between the Ullawarra and Devil Creek Formations is marked by a prominent chert breccia that is restricted to this part of the Pingandy Shelf. This chert breccia locally resembles silcrete in outcrop, but is clearly a stratabound unit developed along this contact. The breccia consists of angular, black and white chert fragments set in a chert matrix that is locally pyritic. Silicification locally extends into the lower Ullawarra Formation to the west of Locality A6.

At this locality the breccia appears to be partly replacing a matrix-supported chert pebble conglomerate, and unaltered dolostone clasts are also locally preserved. However, there is also a component of secondary brecciation and silica replacement as indicated by open-space cavity fills. This breccia is interpreted as a secondary replacement of dolomite breccias in the Devil Creek Formation by hydrothermal fluids. Despite the evidence for hydrothermal fluid flow, geochemical analysis of a sample from Fords Creek failed to detect any anomalous metal content.

About 2 km southeast of Locality A6 the overlying Ullawarra Formation contains large (up to 20 cm) diagenetic barite nodules. Compaction of these dense nodules into the underlying siltstone has resulted in distinctive deformation of the laminae. The internal structure of the nodules consists of a medium-grained equigranular core surrounded by coarse radiating elongate crystals around the margin. Adjacent small concretions locally coalesce into larger concretions and there is a strong internal fabric parallel to bedding, indicative of growth in situ. In thin section the barite crystals are surrounded by a matrix of siltstone. These barite nodules are closely associated with ironstone nodules, up to 5 cm diameter, that appear to be pseudomorphs after pyrite or marcasite.

Locality A7: Collier Group – Edmund Group unconformity (MGA 542911E 7353274N)

From Locality A6 drive about 21 km back along the main station track, towards the Pingandy – Mount Augustus road. Park on the left (east) side of the track, about 300 m south of a windmill and yard (Poverty Bore), that are on the left-hand (east) side of the track. Walk about 50 m south to Locality A7, on a low rise of siltstone overlain by dolostone.

The contact between the Edmund and Collier Groups is marked by a regional low-angle unconformity that cuts down-section from west to east, and represents a hiatus of up to 395 m.y. Along the Pingandy Shelf this unconformity is marked by the contact between dolostones at the base of the Backdoor Formation and various units of the Edmund Group. The unconformity truncates the Coodardoo Formation (at about MGA 509300E 7370700N), and a dolostone in the upper Ullawarra Formation (at about MGA 510200E 7371200N) on Kenneth Range. South of the Talga Fault the unconformity is marked by the contact between basinal mudstone and siltstone at the base of the Backdoor Formation and turbiditic sandstone of the Coodardoo Formation.

At this locality planar-laminated siltstone of the middle to upper Ullawarra Formation is disconformably overlain by interbedded dolomite breccia (Fig. 6) and planar-laminated dolarenite of the basal Backdoor Formation. The Ullawarra Formation contains no sills here, in contrast to the region south of the Talga Fault where 1465 Ma dolerite sills predominate. There is also no sandstone within the Ullawarra Formation in this area, whereas sandstone beds are a common feature south of the Talga Fault. These siltstones are interpreted to have been deposited in a distal shelf to basinal environment, below storm wave-base. The dolostones at the base of the Backdoor Formation are interpreted as distal shelf to slope deposits.

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Figure 6. Intraclast dolomite breccia on the unconformity at the base of the Backdoor Formation (MGA 542911E 7353274N)

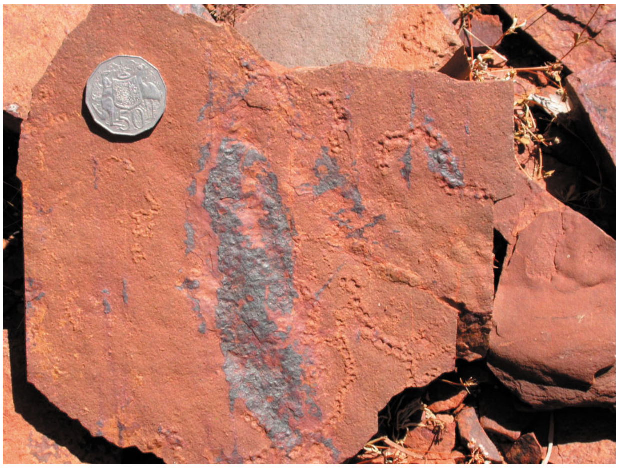

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Figure 7. 'Strings of beads' fossils on the sole of a thin sandstone bed in the Backdoor Formation (MGA 541350E 7350360N)

Locality A8: 'Strings of beads' fossils in the Backdoor Formation (MGA 541350E 7350360N)

From Locality A7 return to the Pingandy – Mount Augustus road, a distance of about 4 km, and turn right (west). Drive west for about 3 km, and stop on the sharp left-hand turn (at about MGA 541300E 7350300N). Here, the Backdoor Formation is exposed to the north of the road, on either side of a small north-flowing creek. Locality A7 is on the west bank of the creek.

The 'strings of beads' are megafossils that are abundant within the lower Backdoor Formation of the Collier Group, and in correlative horizons in the Stag Arrow Formation of the Manganese Group. The strings of beads have been found at more than 45 localities and are distributed over a strike length of more than 500 km within the lower 1500 m of the Backdoor Formation and the correlative Stag Arrow Formation (Martin, 2004b). Known localities of the fossils are commonly restricted to the Pingandy Shelf in sub-tidal below wave-base siltstone facies. They have not been found in any other units of the Bangemall Supergroup, and are considered to be multicellular fossils whose biological affinity remains uncertain. Factors supporting a biological interpretation include the regular size and spacing of the beads, their dissimilarity to known mechanically and chemically induced structures, their morphological complexity and presence across a range of depositional environments, and evidence for flexibility and possible tissue differentiation (Grey and Williams, 1990; Grey et al., 2002). This fossil is also found in the Belt Supergroup of North America, and probably in the Wade Creek Sandstone in the Osmand Range, east Kimberley.

At this locality planar-laminated siltstones and thinly interbedded fine-grained sandstones of the Backdoor Formation dip shallowly to the south. Sandstone beds range from about 1 to 5 cm thick, and are commonly massive with linguoid and lunate ripples on their upper surface. The soles of these sandstone beds are flat, with rare flute and groove casts, and abundant 'strings of beads' (Fig. 7). Each string of beads consists of sinuous serially aligned pits of constant diameter and spacing, forming strings of varying shape that do not appear to be aligned with measured paleocurrent directions. Individual beads are commonly surrounded by a small raised collar of sandstone. In exceptionally well preserved examples a thin strand connects each successive bead. The strings of beads are also commonly associated with enigmatic dimple marks, and fine wrinkle marks (referred to informally as 'elephant skin texture'). The dimple marks may be holdfasts to which the strings of beads were once attached, and the wrinkle marks are interpreted as moulds of microbial mats that colonized the underlying siltstone (Martin, 2004b). Bed thickness increases up-section, and the prominent scarp that can be seen south of the road marks the top of an upward-shallowing succession that is capped by thick-bedded sandstones that also contain the strings of beads. The depositional environment ranges from subtidal below storm wave-base at this locality, to between storm and fair weather wave-base at the top of the scarp south of the road.

Locality A9: Silicified calcrete overlying Backdoor Formation (MGA 534209E 7348310N)

From Locality A8 drive about 8 km west along the Pingandy – Mount Augustus road to where the road climbs up a low rise of silicified calcrete. Park on the right-hand side of the road, to the north of the calcrete. Locality A9 is on the east bank of a small northeast-trending creek incised into the calcrete, about 50 m west of the main road.

Calcrete is a common component of the regolith in this region, and ranges in age from relatively recent pedogenic calcrete developed along active drainage lines to highly incised relict units up to 30 m thick that are commonly perched high above the present drainage and locally overlie ferruginous duricrust. This locality is an example of the latter, which has also been largely replaced by opaline silica. Such replacement is a common feature of calcrete in the Bangemall Geomorphological Province, particularly in the upper parts of the calcrete profile. Most calcrete deposits are vuggy with irregular, lenticular, bedding-parallel cavities that give the rock a fenestral appearance. At this locality the calcrete has been almost entirely replaced by silica, although relict fragments of calcrete are preserved locally. The silicified calcrete overlies saprolite developed on an older alluvium of uncertain age, although it probably correlates with the extensive relict alluvium that overlies the Collier Group between here and Locality A10. Elsewhere in the region these calcretes have been considered prospective for uranium, and small occurrences are found in calcrete locally associated with the unconformity between the Yilgatherra Formation and the Gascoyne Complex.

On the drive to Locality A10 particular note should be taken of the thick relict alluvial deposits that overlie the Collier Group and are deeply incised by the modern drainage. These deposits commonly overlie a ferruginous duricrust, and are themselves affected by at least one period of lateritic weathering and saprolite formation. Clast compositions reflect localized derivation, and the matrix is commonly ferruginous.

Locality A10: Coarse-grained sandstone in the basal Calyie Formation (MGA 522392E 7343401N)

From Locality A9 drive about 17 km west along the Pingandy–Mount Augustus road, and park on the left-hand side of the road where it makes a sharp right-hand turn (about MGA 522400E 7343470N). Locality A10 is on a low ridge of sandstone about 70 m south of the road.

The Calyie Formation gradationally overlies the Backdoor Formation, although in this area the contact is commonly obscured by relict alluvium, and scree derived from the steep slopes of Calyie Formation above. This locality is on a structural culmination within the Wanna Syncline. The ridges of Calyie Formation visible on the skyline to the northwest mark the northern limb of the Wanna Syncline.

Locality A10 is in thick-bedded, trough cross-stratified, medium- to coarse-grained sandstone that is typical of the Calyie Formation. Paleocurrents within the Calyie Formation in this area were towards the west-northwest, in contrast to the predominantly southwesterly paleocurrents in the underlying Backdoor Formation. The transition from the Backdoor Formation to the Calyie Formation is interpreted as reflecting the progradation of delta front and delta top sandstones over prodelta to basinal mudstones and siltstones.

From Locality A10, drive about 56 km to the Mount Augustus Tourist Complex where we will spend the night.

Day 3

The first two stops on Day 3, Localities A11 and A12, will examine the Mount Augustus Sandstone and its unconformable relationship with the Gascoyne Complex around Mount Augustus. The correlation of this unit and its significance to the Bangemall Supergroup are enigmatic. The excursion will then begin to examine the differences in facies and stratigraphy of the Edmund Group south of the Talga Fault. The first of these differences will be examined at Locality A13 where a deep-water facies of the Kiangi Creek Formation is exposed. The trip will then move on to look at more distal, deeper water facies of the Devil Creek Formation at Locality A14, and the Discovery Formation at Locality A15. In addition, at Locality A15 we will see the effects of high-strain partitioning into the Discovery Formation. From Locality A15 there will be a drive of about three hours to our third campsite, close to Locality A16.

Locality A11: Mount Augustus Sandstone on northeastern Mount Augustus (MGA 487520E 7309190N)

From the Mount Augustus Tourist Complex drive southwest for 4.3 km, crossing the Cobra–Meekatharra road onto the access road marked 'Gum Grove/Warrarla', to Locality A11 at the foot of Mount Augustus. Follow the Kotka Gorge trail to where it crosses a small creek. Walk about 50 m up the creek to Locality A11, which is about 60 m southwest of the car park and on the east bank of the creek.

Mount Augustus is a northwest-trending, asymmetrical anticline, with steep and shallow dips on the northern and southern limbs respectively. It is composed of coarse-grained fluvial siliciclastic rocks of the late Paleoproterozoic to early Mesoproterozoic Mount Augustus Sandstone and the unconformably underlying granite and metasedimentary rocks of the Gascoyne Complex (Mangaroon Zone). The distribution of the Mount Augustus Sandstone is restricted to Mount Augustus, and the southern limb of the Cobra Syncline. Correlation of the Mount Augustus Sandstone is unclear. Facies and paleocurrent data suggest equivalence with the Bresnahan Group, but stratigraphic evidence and paleocurrent data from the southern limb of the Cobra Syncline indicate that it may correlate with the fluvial facies of the Yilgatherra Formation at the base of the Bangemall Supergroup. The Mount Augustus Sandstone has been explored for uranium at Centipede Range in the Cobra Syncline.

Locality A11 shows a typical outcrop of the Mount Augustus Sandstone. It is composed of interbedded pebble conglomerate and very coarse grained to granular pebbly sandstone. Bedding dips 30° towards 042°. A 1 m-thick sandstone unit overlies a parallel stratified conglomerate bed and shows distinct 0.1 – 0.3 m-thick, stacked sets of trough cross-strata. The laminae are commonly defined by dark, tourmaline-rich heavy mineral concentrations. Paleocurrent data from trough axes indicate that flow within this fluvial unit was towards the southeast.

Locality A12: Mount Augustus Sandstone unconformity at 'The Pound', on northwestern Mount Augustus (MGA 479178E 7312958N)

From Locality A11 return to the Cobra–Meekatharra road and turn left (west). Drive northwest for 9.5 km and take the turn-off to the southwest, signposted as 'Bowgada Drive'. Follow the road skirting the northwestern end of Mount Augustus for 5.3 km and take the turn off north-northeastwards (left) to the locality signposted as 'The Pound'. Drive to The Pound car park, and walk about 400 m northward along the marked trail to a large stone cairn. From the cairn, follow a small contour path that skirts the ridge to your right for about 200 m to Locality A12.

The unconformable contact between the Gascoyne Complex and the Mount Augustus Sandstone is exposed at several localities along the northwestern slopes of Mount Augustus. Gascoyne Complex rocks consist of foliated, medium- to coarse-grained biotite–muscovite monzogranite, and siliciclastic metasedimentary rocks. The granite is part of the Late Paleoproterozoic Durlacher Supersuite, whereas the metasedimentary rocks belong to the middle to late Paleoproterozoic Pooranoo Metamorphics.

The metasedimentary succession comprises quartz–mica phyllite, interbedded with metamorphosed sandstone and granule to pebble conglomerate. Bedding is tightly folded into steeply plunging small-scale folds with curvilinear axial surfaces. The folds are cut by a strong foliation dipping 80° towards 190°. Near the northern margin of the outcrop the unconformable contact with the Mount Augustus Sandstone is planar, but appears to be faulted. Here, a sliver of steeply dipping pebbly sandstone (Mount Augustus Sandstone) is bound to the north and south by metasedimentary rocks of the Pooranoo Metamorphics.

A lower strain contact between weathered conglomerate and pebbly sandstone and the underlying quartz–mica schist is present on the western slopes (around MGA 479180E 7312960N), although the actual contact is covered by fallen blocks. The strong anastomosing foliation in the schist dips 85° towards 010°. This is reflected by a spaced jointing of similar orientation in the conglomerate and pebbly sandstone. The conglomerate and pebbly sandstone is massive, matrix supported, and has clasts up to 35 cm across. The clasts consist mostly of vein quartz, quartz sandstone, and quartz–tourmaline rock, and angular fragments of quartz–mica schist.

The low hills that can be seen immediately north of the main road are composed of Kiangi Creek Formation sandstone, forming a southeast-plunging anticline, and are separated from Mount Augustus by the southeastern extension of the Lyons River Fault. Similarly, the hills of Discovery Formation visible in the distance are separated from the Kiangi Creek Formation by another major fault.

Locality A13: Kiangi Creek Formation at Gap Well in the Cobra Syncline (MGA 458193E 7319138N)

From Locality A12 return to the Cobra–Meekatharra road, and turn left (west). Drive about 21.8 km to the cattle grid at the Cobra – Mount Augustus round boundary fence. Turn left (south) off the main road about 200 m west of the boundary fence, onto a poorly marked station track and head for the creek crossing (at about MGA 460480E 7321470N). Continue 3.5 km southwest along this track to where a large creek cuts through the sandstone ridge at Gap Well; Locality A13 is on the west bank of this creek.

This locality contrasts with the Kiangi Creek Formation seen on the Pingandy Shelf at Locality A2. South of the Talga Fault the Kiangi Creek Formation consists mainly of planar-laminated siltstone interbedded with thick units of sandstone that were deposited in a deep-water environment. Sandstone units commonly consist of stacked $T_{a–d}$ turbidites or amalgamated T_a turbidite units. The units of stromatolitic dolomite that characterize the Kiangi Creek Formation on the Pingandy Shelf are not present south of the Talga Fault, although dolomitic siltstone units are locally present in this area.

Locality A13 is within a thick unit of massive, medium-grained, thick-bedded, amalgamated quartz–sandstone (T_a turbidites) that is typical of the Kiangi Creek Formation south of the Talga Fault. However, thick sandstone beds are not present everywhere, and their distribution appears to be fault controlled. The sandstone overlies poorly exposed interbedded siltstone and sandstone that probably correlates with the lower Kiangi Creek Formation. This sandstone is overlain by siltstone that is in turn overlain by the Discovery Formation.

Locality A14: Devil Creek Formation in the Cobra Syncline (MGA 448074E 7323600N)

From Locality A13 return to the main Mount Augustus – Cobra road and continue driving west for about 13 km. Turn to the south, off the main road, about 1.1 km east of Bangemall Creek (at about MGA 447900E 7323700N) where the Devil Creek Formation is exposed in the core of the Cobra Syncline, northeast of the Cobra airstrip. Locality A14 is on the low ridge of dolostone south of the road.

The Devil Creek Formation conformably overlies the Discovery Formation, which marks the base of the fourth depositional package in the Bangemall Supergroup. There is a marked change in facies within the Devil Creek Formation across the Talga Fault, from platform-facies stromatolitic dolostone and dolostone breccia, through slope facies adjacent to the fault, to predominantly subtidal below wave-base dololutite south of the fault.

At this locality thick- to thin-bedded planar-laminated dolarenite and interbedded parallel-planar-laminated and hummocky to swaley laminated dololutite of the

Figure 8. **Interbedded planar-laminated siltstone and dololutite, and blocky dolomite intraclast breccia in the Devil Creek Formation (MGA 447900E 7323700N)**

Devil Creek Formation are exposed on a low hill. Thick interbeds of fissile dololutite or dolomitic siltstone are also present (Fig. 8). The facies exposed in this section are typical of the Devil Creek Formation in this area, and are indicative of distal carbonate deposition at or below fair weather wave-base. The breccia beds exposed at this locality are interpreted as mass-flow deposits, as opposed to those seen on the Pingandy Shelf at locality A5. A dolerite sill, interpreted to belong to the 1465 Ma suite, intrudes the formation towards the top of this low ridge. The prominent ridges on the northern side of the road consist of the underlying Discovery and Kiangi Creek Formations.

Locality A15: Discovery Formation in Cobra Syncline (MGA 444067E 7323490N)

From Locality A14 continue driving west along the Cobra – Dairy Creek road for about 4.3 km. Stop (at MGA 444067E 7323490N) where the Discovery Formation of the Edmund Group is well exposed on both sides of the road, south of the Bangemall mining centre. Locality A15 is on the west side of the road.

The Discovery Formation is commonly used as a regional marker horizon in the middle Edmund Group due mainly to the presence of distinctive nodular chert. This chert was considered to be a primary precipitate by Muhling and Brakel (1985). However, evidence from the undeformed Pingandy Shelf suggests that the nodular appearance is partly due to diagenetic silicification of carbonaceous siltstone. This locality is in a high-strain zone in the core of the Cobra Syncline, on the southern limb of the upright overturned Bangemall Anticline. The core of the Bangemall Anticline consists of interpreted 1465 Ma dolerite, which contains northwest-trending quartz veins that are the primary source of alluvial gold mined at the Bangemall Mining Centre.

Tight, upright-folded, laminated siltstone of the Discovery Formation is well exposed on both sides of the Dairy Creek – Mount Augustus road, about 2.5 km west of Cobra Homestead. Stratigraphic evidence indicates that bedding is overturned here (S_0 = 86/046°) and youngs to the south. Cleavage is commonly subparallel to bedding, and also truncates small-scale, upright, doubly plunging intrafolial folds. A well-developed, steeply plunging slickenside lineation is also preserved here, and is cut by orthogonal tension gashes locally. The Discovery Formation at this locality is interpreted to be a silicified deep-water carbonaceous siltstone that has undergone excessive strain during the Edmundian Orogeny. Strain partitioning into the Discovery Formation, particularly in tightly folded areas, is common in the Bangemall Supergroup, but is less pronounced in fold closures.

From Locality A15 drive about 2.7 km southwest to the turn-off to Gifford Creek station, and turn right (northwest). Continue along the Cobra – Gascoyne Junction road for about 90 km to the intersection with a station track (at about MGA 376860E 7353290N). Turn right (north) onto the track and follow it for about 35 km to the turn-off to Deep Bore (at about MGA 374570E 7380560N). Drive eastwards along this track for about 8 km (to about MGA 381310E 7381830N), and turn right onto a track heading south. Drive south along this track for about 3 km (till about MGA 381760E 7379030N which will be our camp site for the night).

Day 4

The morning of Day 4 will be spent examining the lower part of package 1, and comparing it to what was seen on the Pingandy Shelf at Locality A1, and with what has been documented from the Jillawarra Sub-basin further east (Vogt, 1995). At Locality A1 we will examine the Yilgatherra Formation, which is only locally exposed on the Pingandy Shelf, and may be a partial correlative of the Mount Augustus Sandstone. Localities A16 and A17 provide excellent examples of the thick siliciclastic facies that are common in the Irregully Formation in this area, but are absent on the Pingandy Shelf. The part of the excursion dealing with the Edmund and Collier Basins will finish at Locality A19, where carbonate lithofacies typical of this area and the Pingandy Shelf will be examined. After finishing at Locality A19, participants will travel north to Maroonah Homestead where Part B of the excursion, dealing with the Gascoyne Complex, will begin.

Locality A16: Yilgatherra Formation unconformity on Gascoyne Complex (MGA 381832E 7378990N)

Locality A16 is about 100 m southeast of the camp site (at MGA 381760E 7379030N), on the south side of a small westward-flowing creek.

The Yilgatherra Formation in this area consists of a thin basal unit of fluvial coarse-grained sandstone, pebbly sandstone, and localized conglomerate, overlain by interbedded marine sandstone and siltstone that pass gradationally upward into dolostone of the Irregully Formation. The unconformity at the base of the Yilgatherra Formation overlies various units within the Mangaroon Zone of the Gascoyne Complex in this area, and is well exposed at this locality.

At this locality the Yilgatherra Formation unconformably overlies medium-grained, porphyritic biotite monzogranite of the Durlacher Supersuite, and consists of fluvial pebbly sandstone which fines rapidly upward into trough cross-stratified lithic and feldspathic quartz sandstone (Fig. 9). The granite is locally cut by veins of muscovite–tourmaline pegmatite and two-mica granite in this area, and also displays a locally well-developed flow fabric. The main paleocurrent direction in the Yilgatherra Formation at this stop is down-dip, to the southwest. Paleocurrents elsewhere in this area are directed radially off a basement paleotopographic high of Durlacher Supersuite granite. The outcrop break immediately southwest of this locality corresponds to the marine-facies siltstone and sandstone of the Yilgatherra Formation, and the gradational contact into Irregully Formation dolostone.

Locality A17: Interbedded siltstone and sandstone in the Irregully Formation (MGA 383674E 7382101N)

From Locality A16 drive 3 km back along the track to the north and turn right (east) at the junction (at MGA 381310E 7381830N). Drive a further 2.4 km east, then turn southeast off the track (at about MGA 383570E 7382300N) and drive about 200 m to locality A17.

This outcrop is near the top of a thick succession (hundreds of metres) of interbedded siltstone and sandstone that overlies lower Irregully Formation dolostone. These dolostones conformably overlie the Yilgatherra Formation

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Figure 9. Unconformable contact between trough cross-stratified sandstone of the Yilgatherra Formation (foreground) and porphyritic biotite monzogranite of the Durlacher Supersuite (background; MGA 381760E 7379030N)

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Figure 10. Large-scale trough cross-stratified sandstone within the Irregully Formation (MGA 384938E 7381339N)

and will be examined at Locality A19. This locality consists of interbedded planar-laminated siltstone, massive sandstone, and pebbly sandstone. The uppermost sandstone within this succession is not well exposed here, and will be examined at the next locality. This succession is interpreted to have been deposited mainly in a deep-marine shelf environment, below storm wave-base, although some sandstone units elsewhere in this area display features indicative of fluvial to shallow-marine deposition.

Locality A18: Trough cross-stratified sandstone in the Irregully Formation (MGA 384938E 7381339N)

From Locality A17 return to the main track and turn right (east) and drive a further 1.7 km to Locality A18, which is about 150 m south of the track on a prominent sandstone ridge.

This ridge is composed of thick-bedded, large-scale trough cross-stratified, coarse-grained feldspathic quartz sandstone that marks the top of the interbedded siltstone and sandstone unit seen at Locality A17 (Fig. 10). The sandstone locally contains cubic moulds after pyrite (up to 1 cm), and up to 30% sericitized feldspar. To the south this unit is underlain by highly weathered siltstone and fine-grained sandstone beneath a veneer of silcrete and relict colluvium. The paleocurrent direction at this locality is approximately down-dip, to the northeast off a basement high to the south. This sandstone is overlain by interbedded dolostone, siltstone, and sandstone — an assemblage contrasting with that of the Pingandy Shelf where siliciclastic facies are generally thin to absent in the Irregully Formation. The abundance of intercalated siliciclastic rocks within the Irregully Formation in this area led Martin and Thorne (2000) to draw comparisons with the Jillawarra Sub-basin further east, which hosts the Abra polymetallic deposit.

Locality A19: Irregully Formation dolostone (MGA 383958E 7381007N)

From Locality A18 drive back along the track to the west for about 650 m, and turn south onto a faint two-wheel track (at about MGA 384480E 7381850N). This track crosses the sandstone ridge, seen at Locality A18, south of Ginnabooka Bore. From the crest of the ridge head south-southwest across country for about 650 m (to about MGA 384110E 7381150N). Park here and follow a small creek to the south until it intersects a larger northwest-trending creek. Continue southwestwards across the northwest-trending creek for a further 100 m to a prominent low outcrop of dolostone at Locality A19, which is about 220 m southwest of where the cars are parked.

At this locality the Irregully Formation consists of interbedded quartz sand-bearing recrystallized dolostone, dolomitic breccia, and planar- to undulatory-laminated fine-grained dolostone (Fig. 11). Ripple cross-lamination is also locally preserved. These facies are also typical of parts of the Pingandy Shelf, with a notable difference being the thinner breccias and lower abundance of stromatolitic facies at this locality. The depositional environment for this locality is interpreted to be a proximal ramp or shelf-slope, close to fair weather wave-base. These facies are typical of the succession that overlies the fluvial to shallow-marine Yilgatherra Formation (Locality A16), and are overlain by deep-marine shelf, fine- to medium-grained siliciclastic facies of the Irregully Formation (Localities A17 and A18).

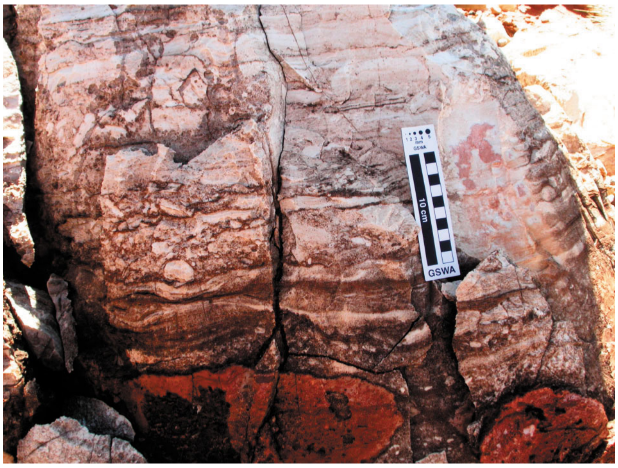

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Figure 11. Interbedded quartz sand-bearing recrystallized dolostone, dolomitic breccia, and planar- to undulatory-laminated fine-grained dolostone in the Irregully Formation (MGA 383939E 7380995N)

From Locality A19 return to the main station track at Ginnabucka Bore, and drive west for about 11 km to the Mangaroon–Maroonah access road. Turn right (north) and drive about 35 km to Maroonah Homestead (MGA 351660E 7402650N) where Part B of the excursion, dealing with the Gascoyne Complex, begins.

Part B: Gascoyne Complex

by

S. Sheppard, T. R. Farrell, and P. B. Groenewald

Geological setting

The Gascoyne Complex comprises Paleoproterozoic granitic rocks and medium- to high-grade metasedimentary rocks that form the high-grade core to the Capricorn Orogen (Williams, 1986). The Gascoyne Complex is juxtaposed against the northern margin of the Archean Yilgarn Craton by the Errabiddy Shear Zone. To the west the complex is overlain by Phanerozoic sedimentary rocks of the Carnarvon Basin, and to the east by Mesoproterozoic sedimentary rocks of the Edmund and Collier Basins (Fig. 1). At its northern end the Gascoyne Complex appears to be transitional to metasedimentary rocks of the upper Wyloo Group, in particular the Ashburton Formation.

The complex is divided into several east-southeasterly trending structural and metamorphic zones and one fault-bound terrane (Fig. 12). Recent U–Pb SHRIMP geochronological studies show that the complex was primarily shaped by four separate orogenic events:

- the 2005–1960 Ma Glenburgh Orogeny,
- the 1830–1780 Ma Capricorn Orogeny,
- the 1680–1620 Ma Mangaroon Orogeny,
- the 1030–950 Ma Edmundian Orogeny.

In addition, continental reactivation during the late Neoproterozoic is evident from an anastomosing network of dextral shear zones cutting the complex.

Glenburgh Orogeny

The 2005–1960 Ma Glenburgh Orogeny is known only from the Glenburgh Terrane at the southern end of the complex, and from the adjacent Errabiddy Shear Zone, because the remainder of the Gascoyne Complex comprises rocks younger than c. 1840 Ma. The Glenburgh Terrane (Fig. 12) is dominated by foliated and gneissic granites of the Dalgaringa Supersuite (Fig. 13), with crystallization ages of 2005–1970 Ma, that intruded Paleoproterozoic metasedimentary rocks of the Moogie Metamorphics and latest Archean granitic rocks (Sheppard et al., 2004) during the Glenburgh Orogeny. The supersuite comprises sheets, dykes, and veins of 2005–1985 Ma foliated and gneissic I-type tonalite, granodiorite, quartz diorite, and monzogranite, intruded by foliated c. 1975 Ma mesocratic and leucocratic I-type tonalite. In the northern part of the Glenburgh Terrane the supersuite was interleaved with Archean basement to form the Halfway Gneiss.

The Errabiddy Shear Zone (Fig. 12) comprises metasedimentary rocks of the Camel Hills Metamorphics, as well as tectonic slices of Archean granitic rocks of the Yilgarn Craton and possibly elements of the Glenburgh Terrane. These rocks were deformed and metamorphosed at medium to high grade during the Glenburgh Orogeny, with the formation of extensive migmatites in the pelitic rocks of the Camel Hills Metamorphics (Occhipinti et al., 2001). The end of the Glenburgh Orogeny was marked by intrusion of granites of the Bertibubba Supersuite at 1965–1945 Ma into the southern edge of the Glenburgh Terrane, the Errabiddy Shear Zone (Occhipinti et al., 2004; Sheppard et al., 2004), and the northern margin of the Yilgarn Craton.

Rocks of the Dalgaringa Supersuite do not intrude the adjacent Yilgarn Craton (Fig. 13), and the 2600–2450 Ma granitic basement to the supersuite has no equivalent in either the Yilgarn or Pilbara Cratons on either side of the Capricorn Orogen. The Glenburgh Terrane is exotic to the Yilgarn Craton to the south. The exposed northern extent of the Glenburgh Terrane is marked by the Chalba Shear Zone (Fig. 12), and the terrane may either be exotic or forms basement to the remainder of the Gascoyne Complex. The orogeny is interpreted to be the result of northwest-directed subduction beneath an Andean-type margin (Glenburgh Terrane), before accretion of the Glenburgh Terrane to the Yilgarn Craton at c. 1960 Ma (Occhipinti et al., 2004; Sheppard et al., 2004).

Capricorn Orogeny

Structures and metamorphic mineral assemblages related to the 1830–1780 Ma Capricorn Orogeny are recognized across the complex, and in adjacent tectonic units (Thorne and Seymour, 1991; Pirajno and Occhipinti, 2000; Sheppard and Occhipinti, 2000; Sheppard et al., 2003). Part of the northern margin of the Yilgarn Craton that was intensely deformed and metamorphosed during the Capricorn Orogeny is referred to as the Yarlarweelor Gneiss Complex (Fig. 12; Occhipinti et al., 1998).

Protoliths to medium-grade, mainly siliciclastic, metasedimentary rocks of the Morrissey Metamorphics were deposited across at least the northern two-thirds

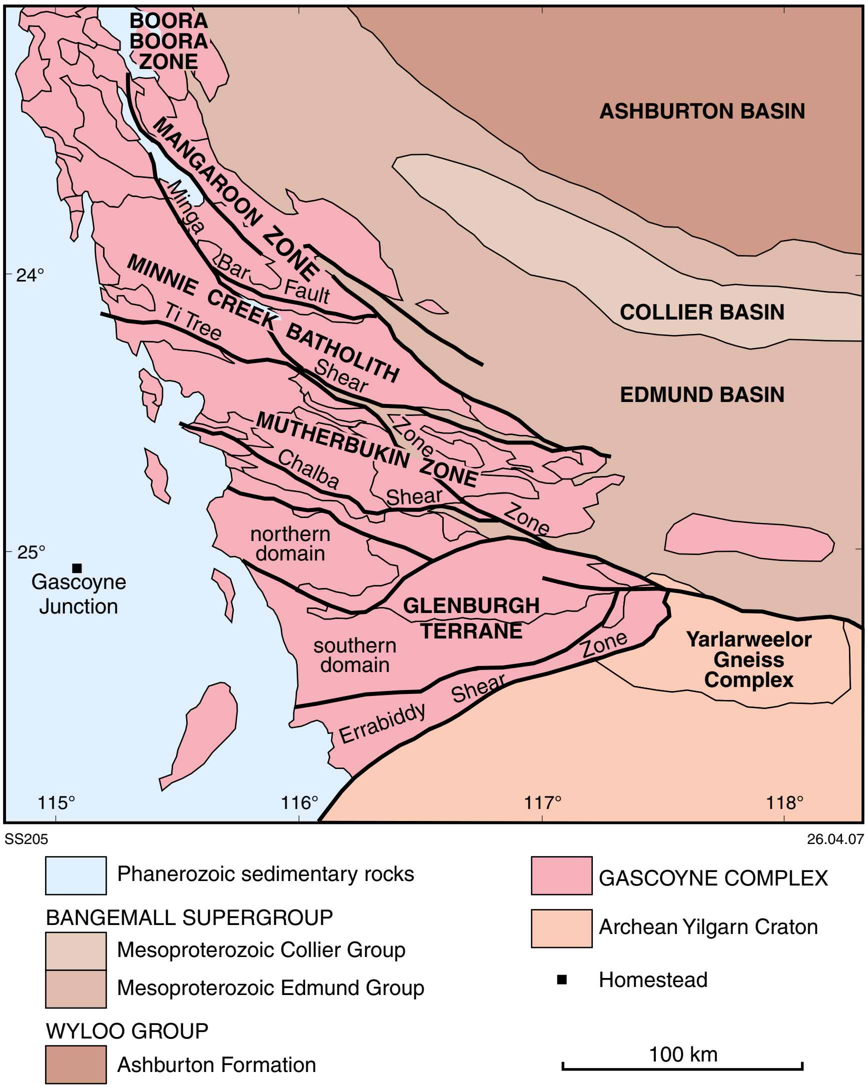

Figure 12. Structural zones and terranes in the Gascoyne Complex. Also shown are the Yarlarweelor Gneiss Complex and the main shear zones

of the Gascoyne Complex (Fig. 13). The maximum depositional age of the Morrissey Metamorphics is not well constrained, but SHRIMP U–Pb dating of detrital zircons in a sample from the central part of the complex have a maximum depositional age of c. 1840 Ma (Varvell, 2001). At the northern end of the complex, metasedimentary rocks in the Boora Boora zone probably grade northwards into lower grade metasedimentary rocks of the upper Wyloo Group, in particular the Ashburton Formation (Williams, 1986; Myers, 1990) which was deposited at about 1830–1800 Ma (see **Ashburton Basin**).

The Morrissey Metamorphics were first deformed and then intruded by granites of the Moorarie Supersuite, which includes the Minnie Creek batholith, during the Capricorn Orogeny. The Moorarie Supersuite was emplaced across the width of the complex and into the Ashburton Basin to the north, and the Errabiddy Shear Zone and Yarlarweelor Gneiss Complex to the south (Fig. 12; Occhipinti et al., 1998; Krapež and McNaughton, 1999; Occhipinti et al., 2001; Martin et al., 2005). The Moorarie Supersuite comprises monzogranite and granodiorite, with subordinate to minor syenogranite, tonalite, and quartz diorite. Most of the rocks contain biotite as the sole mafic mineral, but in the southern part of the complex, biotite(–muscovite–tourmaline)-bearing monzogranite and granodiorite are also common. Small gabbro intrusions and mafic inclusions in granites are present in the Minnie Creek batholith, but absent from elsewhere in the supersuite.

The nature and origin of the Capricorn Orogeny is not clear: it is widely interpreted as the result of oblique collision of the Yilgarn and Pilbara Cratons (Tyler and Thorne, 1990; Evans et al., 2003), but Sheppard (2004) suggested that there are a number of features of the collisional model that are at odds with observed rock associations and geochronological data. These include a lack of volcanic or volcaniclastic rocks in the Morrissey Metamorphics, no evidence of magmatism leading up to the collision, and the composition of the Moorarie Supersuite granites indicative of crustal recycling.

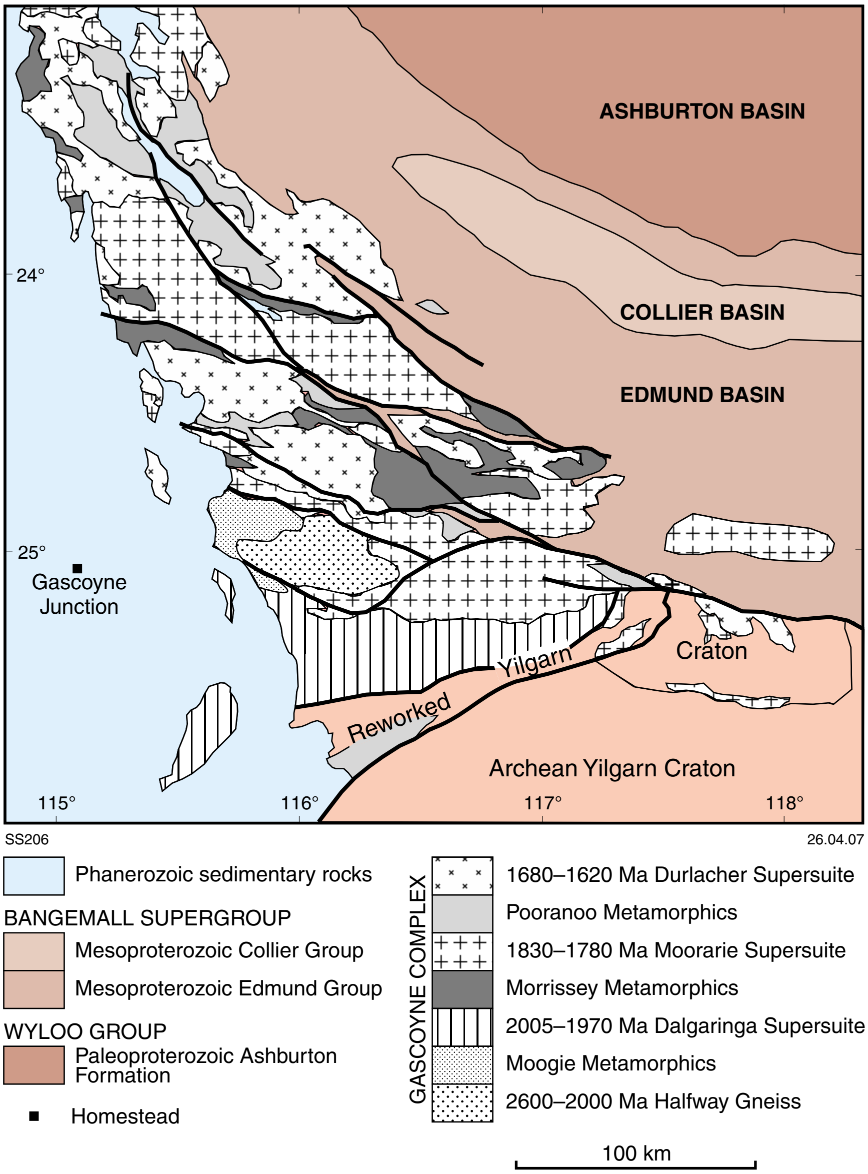

Figure 13. Simplified map showing the main metasedimentary successions and granite supersuites in the Gascoyne Complex

Granites of the Moorarie Supersuite at the northern end of the complex are associated with tungsten-skarn mineralization (Davies, 1998). In the Minnie Creek batholith, leucocratic monzogranite hosts molybdenum mineralization at the Minnie Springs prospect, which has been interpreted as porphyry style (Sullivan, 1996). Disseminated and veinlet-style molybdenite mineralization from the prospect has been dated at 1773 ± 6 Ma using Re–Os geochronology (GSWA unpublished data). This age is within error of younger phases of the Minnie Creek batholith dated at 1790–1780 Ma.

Mangaroon Orogeny

Structures and metamorphic assemblages related to the 1680–1620 Ma Mangaroon Orogeny are best developed in the northern Gascoyne Complex, in particular in the Mangaroon zone. The affects of this orogeny in the central part of the complex (that is, south of the Minnie Creek batholith) may have been largely obliterated by deformation and metamorphism related to the Edmundian Orogeny.

The psammitic, politic, and psephitic precursors to the Pooranoo Metamorphics were deposited across the complex at or before c. 1680 Ma. Low- to medium-grade metasedimentary rocks of the Pooranoo Metamorphics (Fig. 13) lie unconformably on older rocks, including the Morrissey Metamorphics and the Moorarie Supersuite. Lithologically similar low-grade metasedimentary rocks of the Mount James Formation (Hunter, 1990) have the same field relationships as the Pooranoo Metamorphics, and the two units are probably correlatives. During the 1680–1620 Ma Mangaroon Orogeny the Pooranoo Metamorphics, along with older rock units, were deformed and metamorphosed, and intruded by extensive granites of the Durlacher Supersuite (Fig. 13; Martin et al., 2005). In the Mangaroon zone the supersuite comprises biotite–muscovite-bearing monzogranite, granodiorite, and syenogranite, and some muscovite–tourmaline(–biotite) monzogranite. Metamorphism in the Mangaroon zone reached upper amphibolite facies, and appears to have been low-pressure and high-temperature in nature. South of the Minnie Creek batholith, the Moorarie Supersuite comprises biotite monzogranite and biotite–muscovite monzogranite and syenogranite.

The northern and central parts of the Gascoyne Complex appear to have similar early geological histories, suggesting that this crust is contiguous under the Mangaroon zone and that the zone formed roughly in its current position. The lack of any volcanic and plutonic activity immediately preceding the Mangaroon Orogeny, either within or flanking the Mangaroon zone, also precludes the orogeny being related to closure of an ocean. Instead, the Mangaroon Orogeny represents an episode of intracontinental reworking (Sheppard et al., 2005).

Edmundian Orogeny

A fourth event, the Neoproterozoic Edmundian Orogeny, reactivated faults and shear zones within the Gascoyne Complex, as well as folding the sedimentary rocks of the Bangemall Supergroup (Martin and Thorne, 2004). This orogeny is partly responsible for the anastomosing array of structures so prominent on small-scale maps (Fig. 12) and aeromagnetic images. Recent SHRIMP U–Pb geochronology of monazite and xenotime in pelitic schists from the central part of the complex suggests that the Edmundian Orogeny was also responsible for reworking of a southeast-striking corridor between the Chalba and Ti Tree Shear Zones. Monazite and xenotime that grew during greenschist- to amphibolite-facies metamorphism in schists of both the Morrissey Metamorphics and Pooranoo Metamorphics in this zone have been dated at between 1030 and 990 Ma (Sheppard et al., 2007). Previously the peak of regional metamorphism in this area had been interpreted as Paleoproterozoic.

Monazite from an undeformed rare-element pegmatite from the same belt gives a $^{207}Pb/^{206}Pb$ age of c. 950 Ma, suggesting that peak metamorphism was followed by pegmatite intrusion (Sheppard et al., 2007). The dated pegmatite is part of a large belt of pegmatites with the same field relationships and mineralogy that are associated principally with beryllium and tantalum–niobium mineralization. Regional metamorphic fabrics south of the Ti Tree Shear Zone are cut by three massive biotite–muscovite–tourmaline plutons, which are as yet undated. However, the field relationships imply that these plutons are Neoproterozoic.

Unnamed tectonothermal event

Rocks of the Gascoyne Complex and Bangemall Supergroup between the northeastern side of the Cobra Syncline and the Chalba Shear Zone are cut by a network of anastomosing shear zones and brittle–ductile faults. These commonly have well-developed dextral strike-slip shear-sense indicators, and they show dextral offset of dolerite dykes belonging to the c. 755 Ma Mundine Well dyke swarm. Between the shear zones, older structures and metamorphic assemblages appear to be unaffected. The deformation recorded by these shear zones represents an episode of late Neoproterozoic or younger reactivation of the Capricorn Orogen. A sample from the Chalba Shear Zone (Fig. 12), with abundant dextral strike-slip shear indicators and a schistosity in part defined by muscovite, was submitted for Ar–Ar geochronology. Of five small muscovite grains dated using ultraviolet and infrared laser analysis, four yielded an age of 570 ± 10 Ma, with one outlier at c. 610 Ma (GSWA unpublished data). Five larger muscovite grains yielded ages between c. 630 and 2790 Ma. The age of 570 ± 10 Ma is consistent with the field relationships of the shear zones, which cut c. 755 Ma dolerite dykes.

This late Neoproterozoic tectonothermal event may be responsible for some styles of mineralization in the Gascoyne Complex. At the Star of Mangaroon mine, silver–gold–lead–copper–vanadium mineralization is hosted by a quartz vein striking about 010° and dipping about 50° east (Sullivan, 2001). Mineralized veins cut the main east-southeasterly trending fabrics formed during the 1680–1620 Ma Mangaroon Orogeny. The 010° trending structures could be antithetic (R_2) Riedel faults related to east-southeasterly trending dextral strike-slip shear zones. In the Minnie Creek batholith there are a number of small copper prospects and shows hosted by quartz veins within narrow, east-southeasterly trending shear zones with dextral strike-slip kinematics. These shear zones offset dolerite dykes of the Mundine Well swarm.

Excursion localities — northern and central Gascoyne Complex

Part B of the excursion will examine the evidence for recent advances in our understanding of the geological evolution of the Gascoyne Complex. Some of this work has been published, but much is still in preparation or in press. The Gascoyne Complex was earlier thought to have been shaped primarily by one major orogenic event, the Capricorn Orogeny, which is loosely constrained by Rb–Sr dating to between 2000 and 1600 Ma (Williams, 1986). In the last 10 years, regional mapping by the GSWA, integrated with SHRIMP U–Pb zircon geochronology (and more recently SHRIMP U–Pb monazite and xenotime geochronology at the University of Western Australia), whole-rock chemistry, and Sm–Nd isotope studies, has found that the Gascoyne Complex was constructed and reworked episodically between about 2000 and 950 Ma. Fault and shear-zone reactivation continued until sometime after 755 Ma, which is the age of dolerite dykes belonging to the Mundine Well Dolerite Suite (Wingate and Giddings, 2000).

Excursion localities B1–B7 (Plate 1) will examine metasedimentary rocks of the c. 1680 Ma Pooranoo

Metamorphics and 1680–1620 Ma granites of the Durlacher Supersuite in the Mangaroon zone. It was in this zone that the scale and significance of intracontinental reworking during the 1680–1620 Ma Mangaroon Orogeny first became apparent (Sheppard et al., 2005). The Pooranoo Metamorphics is a newly defined unit that comprises pelitic gneiss and metamorphosed feldspathic sandstone, with minor amounts of metamorphosed conglomerate, quartz sandstone, amphibolite, and calc-silicate rock. The Pooranoo Metamorphics have a maximum depositional age of 1680 ± 13 Ma (Sheppard et al., 2005) and are, therefore, much younger than protoliths to the Morrissey Metamorphics (<1840–1800 Ma, Varvell et al., 2003) and the Moogie Metamorphics (>2005 Ma, Occhipinti and Sheppard, 2001) in the central and southern parts of the Gascoyne Complex respectively.

Localities B9–B10 (Plate 1) typify granites of the Minnie Creek batholith, which was constructed from numerous intrusions belonging to the 1830–1780 Ma Moorarie Supersuite. Granites of the Moorarie Supersuite were emplaced across the width of the Gascoyne Complex, and represent an important episode of crustal recycling. The batholith appears to have acted as a buttress during reworking of the Gascoyne Complex during the Mangaroon Orogeny and the 1030–950 Ma Edmundian Orogeny.

Localities B11–B16 (Plate 1) highlight the most recent changes to our understanding of the central Gascoyne Complex, much of which is currently unpublished. South of the Mesoproterozoic sedimentary rocks of the Edmund Group in the Ti Tree Syncline, low- to medium-grade metasedimentary rocks were all thought to belong to the Morrissey Metamorphics (deposited after 1840 Ma; Varvell, 2001) and to have been deformed during the 1830–1780 Ma Capricorn Orogeny (Varvell et al., 2003; Fitzsimons et al., 2004). However, field mapping in 2005 and subsequent SHRIMP U–Pb dating of detrital zircons indicate that the metasedimentary rocks belong to two separate packages — the Morrissey Metamorphics and the Pooranoo Metamorphics — that are separated by a low-angle unconformity, which is in part sheared. The main episode of regional metamorphism was then interpreted to be related to the Mangaroon Orogeny (Sheppard et al., 2006) using ages for granites that bracket the metamorphism. However, subsequent SHRIMP U–Pb dating of in situ metamorphic monazite and xenotime indicates that the main phase of regional metamorphism occurred at 1030–950 Ma (Sheppard et al., 2007). This apparent contradiction can be explained if the crosscutting leucocratic granite dated at 1652 ± 4 Ma (Culver, 2001), which provided the spurious younger age limit for regional metamorphism, contains only xenocrystic zircons.

Locality B1: Psammitic gneiss of the Pooranoo Metamorphics (MGA 347460E 7400450N)

From the meeting point at the gate (MGA 351423E 7402873N), just north of Maroonah Homestead, go south through the gate and drive 100 m before turning right through the gate to the west-southwest. Drive for 3.7 km (to MGA 348084E 7401633N) and turn left on an old station track for 1 km (to MGA 347933E 7400699N).

Psammitic gneiss and schist is a regionally extensive component of the Pooranoo Metamorphics. Here, small pavements and fresh boulders of psammitic gneiss and schist outcrop with minor pelitic horizons. Psammitic rocks contain scattered rounded quartz and K-feldspar grains with rare graded bedding. The protolith to the psammitic gneiss is feldspathic sandstone comprising coarse sand and granules of quartz in a fine sandy matrix dominated by quartz and plagioclase. Small-scale folds (F_{2m}) with S-vergence record upright folding of S_{1m} fabric. Throughout the Mangaroon zone F_{2m} folds are associated with a pervasive east-southeasterly striking, upright schistosity (S_{2m}), which is the dominant fabric in rocks of the Pooranoo Metamorphics. Granite veins subparallel to layering are also folded. A large dyke or sheet of medium-grained, sparsely porphyritic biotite–muscovite monzogranite intrudes the gneisses and post-dates folding. The monzogranite contains disseminated inclusions of biotite-rich pelitic gneiss.

A boulder of psammitic gneiss with preserved graded bedding was sampled here for SHRIMP U–Pb geochronology (Fig. 14). The sample, GSWA 178747, contains one zircon at 1756 ± 17 Ma (1σ) and 24 analyses of 23 zircons at 1803 ± 7 Ma, in addition to 10 zircons between 2328 and 1906 Ma (Geological Survey of Western Australia, 2006). Other samples from this unit have maximum depositional ages of c. 1800 Ma (Sheppard et al., 2005). However, given that the psammitic gneiss unit is interbedded with pelitic gneiss, a sample of which gives a maximum depositional age of 1680 ± 13 Ma (Sheppard et al., 2005), then the protolith to the psammitic gneiss must also have been deposited after c. 1680 Ma.

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Figure 14. Boulder of low-strain psammitic gneiss sampled for detrital zircon geochronology (GSWA 178747). The boulder shows weakly developed graded bedding with way-up towards the top of the photo (MGA 347460E 7400450N)

Locality B2: Psammitic gneiss and metaconglomerate of the Pooranoo Metamorphics (MGA 356600E 7399110N)

From Locality B1 return to the gate west of Maroonah Homestead. Turn right (south) and drive for 8.1 km to a fork in the road (MGA 354566E 7396047N). Take the track to Sheela Bore. Drive for 3.1 km until a gate (MGA 357599E 7395763N). Turn left along a poorly marked track along an old fenceline for about 2.5 km and then across open country toward Locality B2.

At this locality homogeneous psammitic schist and gneiss is extensively veined by medium-grained equigranular biotite granodiorite. The protolith to most of the outcrop is probably a fine- or medium-grained feldspathic sandstone, but there are thin (<0.5 m) intervals of metamorphosed granule sandstone. Bedding and an early gneissic fabric (S_{1m}) are folded about rare upright folds with a schistosity (S_{2m}) parallel to the axial surfaces of the folds. Granodiorite veins and dykes may have a weak magmatic foliation or schistosity parallel to S_{2m} in the metasedimentary rocks. A sample of granodiorite similar to these veins and dykes, about 4 km south-southwest of Locality B2, has an igneous crystallization age of 1678 ± 6 Ma (GSWA 178030; Martin et al., 2005).

Walk about 250 m towards folded layers of metamorphosed quartz-pebble conglomerate (MGA 356589E 7399334N). Lenses of metamorphosed conglomerate (Fig. 15), typically less than a few metres thick, are within the psammitic gneiss and schist. Pebbles define a steeply plunging lineation parallel to F_{2m} folds. Elsewhere in the northern part of the Mangaroon zone, metamorphosed pebble and cobble conglomerate and quartz sandstone form lenses up to 3 km long within pelitic and psammitic gneiss.

From Locality B2 return to main track beside gate. Turn right and drive for a few hundred metres to camp on left-hand side of track.

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Figure 15. Metaconglomerate of the Pooranoo Metamorphics showing stretched cobbles of vein quartz (MGA 356589E 7399334N)

Locality B3: Mingled gabbro and granite of the Durlacher Supersuite (MGA 372580E 7378920N)

From campsite drive west back to track junction (MGA 354566E 7396047N, and turn left (southeast). Drive for 24.6 km until old station track on right (MGA 372645E 7381872N). Take track and drive for 3.1 km. Park and walk about 50 m to exposures on west side of track.

In the Mangaroon zone the 1680–1620 Ma Durlacher Supersuite comprises biotite–muscovite monzogranite, granodiorite, and syenogranite, with subordinate leucocratic muscovite–tourmaline(–biotite) granite and biotite granodiorite. Gabbro is a very minor component of the Durlacher Supersuite, being restricted to several scattered exposures each with an area of less than 1 km^2. A gravity traverse was conducted across this locality to determine whether or not the gabbro here is part of a larger unexposed body. Preliminary results (GSWA unpublished data) suggest that the gabbro is only a very small intrusion.

At this locality boulders and tors consist of massive medium-grained xenocrystic metagabbro mingled with pale-grey monzogranite and pegmatite (Fig. 16). The metagabbro contains several percent biotite, in addition to partly rounded xenocrysts of K-feldspar from the comingled monzogranite and pegmatite. The main mafic mineral in the metagabbro is a pale-green actinolite, which contains cores of igneous pyroxene. Locally the pale-grey granite is veined by strongly porphyritic fine- to medium-grained monzogranite. Some pegmatite veins in gabbro contain acicular crystals of green actinolite up to 30 cm long.

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Figure 16. Net-vein structure developed between gabbro and monzogranite of the Durlacher Supersuite. The crenulate nature of the contacts and the widespread K-feldspar xenocrysts within the gabbro are indicative of mingling between mafic and felsic magma (MGA 372580E 7378920N)

Locality B4: Pelitic gneiss of the Pooranoo Metamorphics (MGA 372480E 7377960N)

From Locality B3 drive 0.9 km further south down the track and examine fresh boulders on left-hand side of track.

Pelitic gneiss and granofels is the other major component of the Pooranoo Metamorphics, besides the psammitic gneiss and schist. The hornfelsic texture of the pelitic gneiss at this locality is typical of much of the pelitic gneiss in this area north of the Mangaroon Syncline. The texture suggests that much of the pelitic gneiss here is underlain by granite, perhaps fine- to medium-grained porphyritic monzogranite of the Dingo Creek Granite, which extensively veins the gneiss in this area. A sample of pelitic gneiss from MAROONAH (GSWA 169094) has populations of zircons at 1680 ± 13 Ma (5 analyses of 4 grains: χ^2 = 0.70), 1741 ± 23 Ma (3 grains) and 1783 ± 6 Ma (18 grains: χ^2 = 1.46), in addition to older ungrouped zircons (Geological Survey of Western Australia, 2006). The population at 1680 ± 13 Ma is interpreted as detrital because, although these zircons are small euhedral grains or fragments without pitting or rounding typical of the older zircons, they do not have the low Th/U ratios common in many metamorphic zircons (Rubatto and Gebauer, 2000), and they contain complex growth zoning typical of granitic rocks (Corfu et al., 2003), and thus resemble those of the older detrital populations.

At this locality boulders of coarse-grained, unmelted, cordierite–quartz–muscovite–biotite–sillimanite–plagioclase gneiss are exposed (Fig. 17). Thin leucocratic melt veins cut the rock, and some enclosing layers of pelitic gneiss show migmatitic structures (metatexites). Cordierite and muscovite are poikilitic, and these minerals, along with biotite, are wrapped and form partly corroded remnants within seams and mats of fibrolite. Plagioclase and cordierite are partly recrystallized to fine-grained sericite and chlorite. There is no K-feldspar.

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Figure 17. Poikoloblastic cordierite–quartz–muscovite–biotite–sillimanite–plagiocase gneiss (MGA 372480E 7377960N)

Locality B5: Granites of Durlacher Supersuite (MGA 385750E 7367230N)

From Locality B4 drive back north along track to main station track and turn right. At 2.8 km take the main track, ignoring track to left. Drive for a further 14.5 km and turn left at the poorly used track at the junction east of Alma Outcamp (ignore well-worn tracks straight on and to right). Travel for 1.2 km to the gate, ignoring the track to left. Drive for a further 4.8 km.

The Durlacher Supersuite was emplaced into the Pooranoo Metamorphics and older rock units during the Mangaroon Orogeny, and it is a major component of the Mangaroon zone (Sheppard et al., 2005). In the Mangaroon zone, and in the central Gascoyne Complex, the bulk of the granites have crystallization ages of 1680–1620 Ma, but granites of the supersuite intruded the Gascoyne Complex and adjacent Yarlarweelor Gneiss Complex to the southeast until as late as 1620 Ma (Sheppard and Swager, 1999).

At this locality granites belonging to the two largest units of the Durlacher Supersuite are exposed (Fig. 18). The majority of the exposure consists of large pavements and whalebacks of coarse-grained strongly porphyritic biotite monzogranite belonging to the Pimbyana Granite (Pearson et al., 1996; Martin et al., 2005). This monzogranite contains 30–45% tabular K-feldspar phenocrysts 1.5 – 2 cm long commonly with a weak preferred orientation. This is by far the most abundant and widespread rock type within this unit in the Mangaroon zone. Other rock types, which include fine- to medium-grained tonalite and granodiorite, are only exposed east and southeast of the Star of Mangaroon mine and around the Frasers–Yangibana prospect on EDMUND.

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Figure 18. Coarse-grained megacrystic biotite(–muscovite) monzogranite of the Pimbyana Granite (lower two-thirds of photo) intruded by a dyke of medium-grained porphyritic biotite–muscovite monzogranite of the Dingo Creek Granite. Igneous flow banding in the Dingo Creek Granite, which is defined by oriented K-feldspar phenocrysts, is parallel to the dyke wall (subhorizontal in photo; MGA 385760E 7367230N)

The Pimbyana Granite is intruded by dykes and veins of leucocratic granite and pegmatite. All these phases are intruded by dykes and sheets of fine-grained strongly porphyritic biotite(–muscovite) monzogranite of the Dingo Creek Granite ('Dingo Granite' of Pearson, 1996). Here, as elsewhere in the region, the Dingo Creek Granite is characterized by 30% or more thin tabular K-feldspar phenocrysts up to 1.5 cm long with a trachytic texture. The phenocrysts are typically oriented parallel to dyke walls at the dyke margins, where they define a magmatic flow fabric. Towards the centre of the dykes the phenocrysts may have no preferred orientation. In the region, intrusions of the Dingo Creek Granite are composed of numerous southeast-striking sheets from a few metres to hundreds of metres wide. The sheets and the magmatic foliation typically strike parallel to the S_{2m} tectonic foliation in rocks of the surrounding Pooranoo Metamorphics (Martin et al., 2005).

Both the Pimbyana Granite and Dingo Creek Granite from this locality were sampled for SHRIMP U–Pb zircon geochronology. The sample of Pimbyana Granite (GSWA 178029) has an igneous crystallization age of 1675 ± 11 Ma, and the sample of Dingo Creek Granite (GSWA 178028) has an igneous crystallization age of 1674 ± 8 Ma (Geological Survey of Western Australia, 2006). The age for the Dingo Creek Granite here is identical to the age of 1674 ± 6 Ma determined by Pearson (1996) from a sample about 30 km to the southeast.

Locality B6: Gooche Gneiss basement (MGA 373300E 7359210N)

From Locality B5 return to the main station track near Alma Outcamp, then drive 6.3 km south until a fork in the track and take the right-hand track. Drive for a further 9.1 km (MGA 379603E 7356514N) and turn right at the track. Drive for 7.9 km until a track junction. Continue straight ahead and park on the left-hand side of the track about 100 m past the junction.

The Gooche Gneiss is exposed as a series of small inliers (<2 km^2) tectonically interleaved with the Pooranoo Metamorphics. The gneiss is also common as inclusions within schlieric, inclusion-rich granodiorite — the oldest phase of the Durlacher Supersuite. A sample of Gooche Gneiss (GSWA 169058) from the EDMUND (1:100 000) sheet area has been dated at 1776 ± 8 Ma (Geological Survey of Western Australia, 2006). The gneiss represents a deformed part of the 1830–1780 Ma Moorarie Supersuite.

At this locality a small whaleback and several pavements of strongly porphyritic foliated to gneissic granite (locally an augen gneiss) are exposed. Round K-feldspar phenocrysts are up to 3 cm in diameter. The east-southeasterly trending tectonic fabric here is a composite of an L-tectonite fabric and an S-foliation. In places an S–C fabric and asymmetric porphyroclasts indicate reverse sense of shear (that is, southwest-side up). The foliation is cut by a quartz-filled brittle fault striking southeast; quartz fibres in associated quartz veins are normal to vein walls and imply sinistral shear.

Locality B7: Migmatitic pelitic gneiss and psammitic gneiss of the Pooranoo Metamorphics, Star of Mangaroon mine

From Locality B6 continue along the track across a large creek then to the right and past an abandoned windmill. Continue along the track for 1.3 km and park near the base of a knoll with a large antenna. Walk to the foot of the knoll (Locality B7a).

High-grade pelitic gneiss and granofels of the Pooranoo Metamorphics outcrop over a wide area north of the Mangaroon Syncline, but have a more restricted distribution south of the syncline. North of the syncline, pelitic gneiss and granofels contain the assemblages biotite–muscovite–quartz–plagioclase–sillimanite, quartz–biotite–cordierite–plagioclase–muscovite(–sillimanite), and plagioclase–biotite–quartz–sillimanite–muscovite–cordierite. The appearance of sillimanite in nonmigmatitic pelitic gneiss is consistent with a metamorphic grade equivalent to amphibolite facies. In contrast, south of the Mangaroon Syncline, including the area around the Star of Mangaroon mine, migmatitic pelitic gneiss (metatexite migmatite and minor diatexite migmatite) marked by the appearance of K-feldspar is abundant. The co-existence of sillimanite and K-feldspar in these rocks is consistent with the onset of upper amphibolite-facies conditions (Bucher and Frey, 2002, p. 110).

At this locality we will examine stromatic pelitic migmatite (Locality B7a), unmelted pelitic gneiss (B7b), refolded folds in psammitic gneiss (B7c), and diatexite migmatite (B7d). The return walk from Localities B7a to B7d covers about 900 m. The amount of partial melting varies substantially within the sequence, probably depending on the bulk composition of the individual layers.

Locality B7a: Pelitic gneiss, including migmatite (MGA 372340E 7359950N): Stromatic pelitic migmatite is very well exposed at the foot of the south side of the knoll. The rocks contain leucosomes of microcline, quartz, and minor plagioclase alternating with melanosomes of coarse-grained biotite, cordierite, sillimanite, and accessory iron oxides and muscovite (Fig. 19a). The leucosomes are both parallel and locally discordant to S_{1m}. The major difference with nonmigmatitic pelitic gneisses north of the Mangaroon Syncline is the paucity of plagioclase and muscovite, and the appearance of microcline. The co-existence of sillimanite and K-feldspar, coupled with the scarcity of plagioclase and muscovite, suggests that melting occurred via either of the following reactions:

muscovite + plagioclase + quartz = sillimanite + K-feldspar + melt (1)

muscovite + plagioclase + quartz = sillimanite + K-feldspar + biotite + melt (2)

The temperature at which reaction (1) occurs is strongly dependent on pressure, but a minimum temperature of about 630°C at 200 MPa is required, or about 700°C at 500 MPa (Spear, 1993, p. 368). Reaction

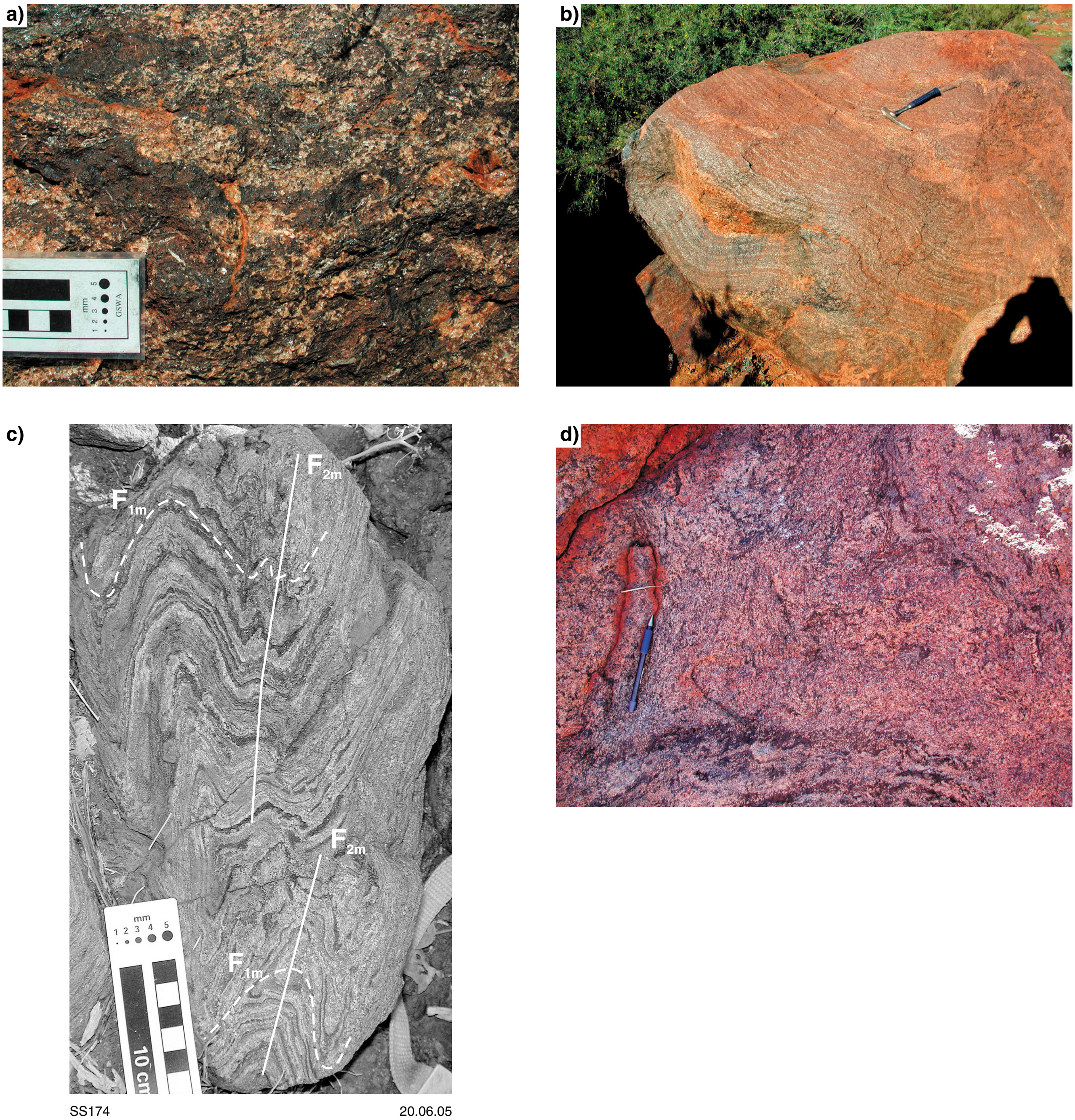

Figure 19. Migmatitic pelitic gneiss and psammitic gneiss of the Pooranoo Metamorphics: a) stromatic migmatite with leucosomes of microcline, quartz, and minor plagioclase alternating with melanosomes of coarse-grained biotite, cordierite, sillimanite, and accessory iron oxides and muscovite (MGA 372340E 7359950N); b) unmelted quartz–cordierite–biotite–microcline–sillimanite(–plagioclase) gneiss in hinge of small synform. This boulder was sampled for SHRIMP U–Pb zircon dating of detrital zircons (GSWA 178748; MGA 372332E 7359978N); c) small-scale fold-interference structures in pelitic and psammitic gneiss (MGA 372200E 7360090N); d) diatexite migmatite (granodiorite to tonalite), containing muscovite, biotite, and cordierite, with or without minor garnet and sillimanite. Schlieren of cordierite, biotite, sillimanite, and muscovite. This boulder was sampled for SHRIMP U–Pb zircon dating (GSWA 178749; MGA 372080E 7360170N)

(2) requires a temperature of about 720–750°C at 600 MPa (Patiño Douce and Harris, 1998).

Locality B7b: Pelitic gneiss (MGA 372340E 7359950N): Unmelted pelitic gneiss, which is interlayered with stromatic migmatite, is best exposed at the top of the knoll. Here, gneiss composed of cordierite–biotite–quartz–microcline–sillimanite(–muscovite) or quartz–cordierite–biotite–microcline–sillimanite(–plagioclase) is folded about an upright fold plunging to the west-southwest (Fig. 19b). A schistosity parallel to the fold-axial surface is defined by fibrolitic sillimanite. The grade of this schistosity is much higher than that which is axial planar to the regionally dominant southeast-striking F_{2m} folds, suggesting that this is an F_{1m} fold.

The assemblages developed in the Pooranoo Metamorphics are typical of high-Al pelitic rocks regionally metamorphosed at low-pressure amphibolite-facies conditions (Spear, 1993, p. 374–382). The absence of garnet (other than locally where it may have been stabilized by high whole-rock MnO), staurolite, and kyanite, combined with the abundance of cordierite, suggests that the rocks were not metamorphosed at intermediate to high pressures. The cordierite- and sillimanite-bearing assemblages noted above are stable at about 600–630°C at 200 MPa or 650–700°C at 500 MPa (Spear, 1993, p. 375–382).

Locality B7c: Psammitic gneiss (MGA 372166E 7360118N): Here a thin layer (<1 m thick) of psammitic gneiss contains rare, small, type 2 and 3 (Ramsay, 1967) fold-interference structures (Fig. 19c). Metamorphic assemblages are of cordierite–quartz–microcline–biotite–sillimanite–plagioclase – iron oxides. Cordierite commonly forms sieved porphyroblasts about 5 mm in diameter, with inclusions of quartz and biotite.

Locality B7d: Diatexite migmatite with inclusions of biotite–sillimanite-rich restite (MGA 372080E 7360170N): The diatexite migmatites are heterogeneous, medium- to coarse-grained granitic rocks with abundant biotite schlieren, as well as inclusions of metasedimentary rock (pelitic gneiss and metamorphosed feldspathic sandstone), with lesser amounts of vein quartz (with or without tourmaline), pegmatite, and augen gneiss (Gooche Gneiss). The diatexites range in composition from granodiorite to tonalite, and contain muscovite, biotite, and cordierite, with or without minor garnet and sillimanite. Schlieren consist of biotite–cordierite–sillimanite–muscovite–plagioclase or cordierite–plagioclase–quartz–sillimanite–muscovite. Diatexite migmatites commonly contain inclusions of gneiss with an S_{1m} fabric, indicating that melt transport and crystallization post-dated D_{1m}.

The intrusive nature of the diatexites suggests that they were not derived by partial melting of their immediate wallrocks, in contrast with the stromatic migmatites, such as seen at Locality B7a. The nature of the melting reaction that produced the diatexite migmatites is unclear. The muscovite dehydration reactions likely to be responsible for the stromatic migmatites do not produce more than about 5% partial melting (Clemens and Vielzeuf, 1987). The diatexite migmatites may have been generated below the current level of exposure at higher temperatures by biotite dehydration melting, which, on both theoretical and experimental grounds, can produce large amounts of melting (Clemens and Vielzeuf, 1987).

This locality consists of a weakly banded, cordierite-rich diatexitic migmatite that contains disrupted layers of pelitic gneiss, fragments of metamorphosed quartz sandstone, and biotite- and sillimanite-rich schlieren (Fig. 19d). A large boulder from here was sampled for SHRIMP U–Pb zircon geochronology in order to define the age of partial melting (and hence, peak metamorphism), and provide information on the provenance of the sedimentary protolith. Of the 57 zircons analyzed from this sample (GSWA 178749), 54 have $^{207}Pb/^{206}Pb$ ages of 1840–1724 Ma; there are two older zircon grains, as well as one grain with a $^{207}Pb/^{206}Pb$ date of 1657 ± 12 Ma (1σ; Geological Survey of Western Austrralia, 2006). The youngest grain, which has a very low Th/U value (0.02) typical of zircon crystallized during metamorphism (Rubatto and Gebauer, 2000), provides the best estimate of the age of peak metamorphism (Geological Survey of Western Austrralia, 2006). This sample contains xenoblastic crystals of garnet mantled by cordierite, implying at least partial re-equilibration of the primary assemblage, probably during decompression from peak metamorphism (Geological Survey of Western Australia, 2006).

Locality B8: Schlieric granodiorite of Durlacher Supersuite (MGA 366900E 7357422N)

Return via the track and turn right at the junction, then travel for 5.8 km down to the main road. Turn right and travel for 7.5 km until a faint station track on the right-hand side (MGA 367230E 7352304N). Drive north for 5.6 km along the old track and past a small pit across country to boulders and a small whaleback.

Schlieric, inclusion-rich granodiorite is the oldest unit of the Durlacher Supersuite. It forms irregular plutons composed of numerous steeply dipping to vertical sheet-like bodies that strike southeast with abundant screens of Pooranoo Metamorphics and Gooche Gneiss. The granodiorite typically contains a southeast-striking magmatic flow fabric or schistosity parallel to the dominant S_{2m} fabric in the enclosing Pooranoo Metamorphics. Veins and dykes of the granodiorite mostly intrude along the S_{2m} foliation. These features collectively imply that the granodiorite was intruded during D_{2m}. A sample of the granodiorite about 4 km west of this locality has an igneous crystallization age of 1677 ± 5 Ma (GSWA 178027; Geological Survey of Western Australia, 2006). This is probably the best estimate for the age of peak metamorphism in the Mangaroon zone.

At this locality, schlieric inclusion-rich granodiorite is exposed as a small whaleback and numerous boulders. The granodiorite intrudes, and has numerous inclusions of, psammitic and pelitic schist, as well as polycrystalline quartz. Inclusions may constitute up to 50% of the granodiorite in places. Locally the pelitic rocks contain sillimanite. A medium-grade S_{1m} fabric in country rock and inclusions of the schist is truncated by the granodiorite.

From Locality B8 drive a few hundred metres to the campsite in open ground.

Locality B9: Biotite granodiorite of the Minnie Creek batholith (Moorarie Supersuite; MGA 361697E 7351869N)

Return to the main road, turn right, and then travel for 6.2 km. Park on the left-hand side of the road and examine exposures on the right-hand side of the road.

The Minnie Creek batholith has an exposed length of more than 170 km and is about 30–50 km wide (Fig. 13). It is composed of granites of the 1830–1780 Ma Moorarie Supersuite; none of the intrusions in the batholith dated so far belongs to the younger Durlacher Supersuite. The majority of the batholith consists of monzogranite and granodiorite, with some tonalite and minor quartz diorite. Small gabbro intrusions are present throughout the batholith and many of the granites contain fine-grained mafic inclusions. The presence of mafic rocks distinguishes the batholith from other granites of the Moorarie Supersuite in the northern and southern parts of the Gascoyne Complex. The majority of granites in the batholith are massive, but rocks along the margin are commonly strongly foliated.

This locality is only a little over 1 km from the Minga Bar Fault, which is a large structure that separates the Minnie Creek batholith from the Mangaroon zone. On the north side of the fault, rock units were deformed and metamorphosed at grades up to upper amphibolite facies at 1680–1620 Ma, whereas granites of the Minnie Creek batholith show evidence only of static recrystallization at low to medium grade. At this locality, pale-weathering medium-grained sparsely porphyritic biotite granodiorite is exposed, with numerous small clots of fine-grained biotite. Plagioclase, which dominates the feldspars, is pale green because of the fine-grained replacement by sericite and epidote. The granodiorite is part of a unit of medium- to coarse-grained biotite tonalite, a sample of which has an igneous crystallization age of 1783 ± 5 Ma (GSWA 178024; Geological Survey of Western Australia, 2006). Also present at this locality are (?)dykes of a medium-grained, equigranular leucocratic biotite monzogranite. Both granites are cut by low-grade cleavages possibly related to the Minga Bar Fault.

Locality B10: Porphyritic monzogranite of the Minnie Creek batholith (Moorarie Supersuite; MGA 429932E 7313332N)

Drive back east towards the Gifford Creek Homestead for 45.5 km. Ignore the turnoff to the left towards Edmund Homestead, and continue past the Gifford Creek Homestead to the T-intersection with the Dairy Creek–Cobra road (104.7 km in total). Turn right towards Dairy Creek and drive for about 16 km. Granite boulders and tors are exposed on the right-hand side of the road.

Exposed at this locality is a foliated porphyritic biotite monzogranite with about 20–25% round and oval phenocrysts and porphyroclasts of K-feldspar. This is a very common rock type in the Minnie Creek batholith, and it gives igneous crystallization ages of between 1795 ± 8 and 1781 ± 6 Ma (Evans et al., 2003; Geological Survey of Western Australia, 2006). Here, the monzogranite contains scattered inclusions of fine-grained biotite-rich pelitic rock up to 20 cm long, but regionally it also has inclusions of mafic rock and porphyritic microgranite.

The monzogranite at this locality has a steeply plunging mineral lineation defined by the alignment of elongate aggregates of biotite (Fig. 20). The asymmetry of ∅-porphyroclasts and S–C relationships suggest reverse (north over south) movement. The age of the deformation at this locality is not well constrained because ductile deformation and biotite-grade metamorphism in shear zones is associated with the 1830–1780 Ma Capricorn Orogeny, the 1680–1620 Ma Mangaroon Orogeny, and the 1030–950 Ma Edmundian Orogeny. However, the absence of such deformation in rocks of the Edmund Group along strike from this shear zone suggests that the shearing visible at this locality is related to either the Capricorn Orogeny or Mangaroon Orogeny.

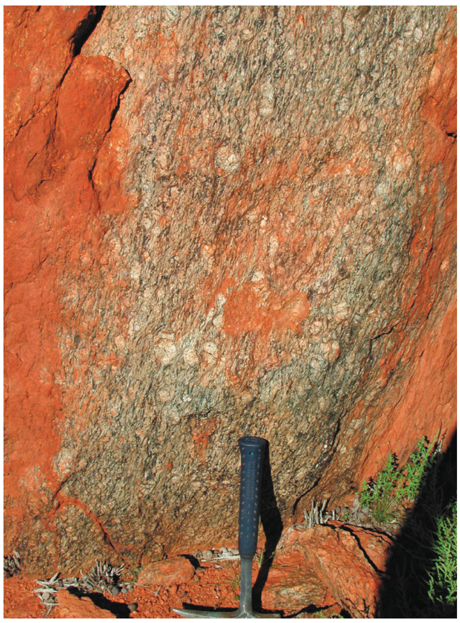

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Figure 20. Foliated porphyritic biotite monzogranite of the Minnie Creek batholith. The subvertical surface shows K-feldspar porphyroclasts with asymmetric tails suggesting dextral shear sense (MGA 429932E 7313332N)

Locality B11: A short traverse through medium-grade pelitic schist, psammitic schist, amphibolite, and quartzite of the Pooranoo Metamorphics to strongly sheared c. 1800 Ma granites (MGA 408973E 7293723N to MGA 408838E 7293140N)

Drive 16 km south along the Minnie Creek–Cobra road, then turn west off the road onto a station track. Take the left fork at 8.5 km, through an open gate at 15.2 km, and veer to the right of the microwave tower at 17.7 km. Continue on to Nardoo Well 21.7 km from the road. Stop on the west side of the tributary creek just west of the well.

Recent mapping, combined with SHRIMP U–Pb zircon and monazite dating, provides the basis for a major reinterpretation of the geological history in the Gascoyne Complex south of the Minnie Creek batholith. This work, much of which is in preparation or in press, suggests that two metasedimentary packages are present in the area, and that regional metamorphism and deformation is latest Mesoproterozoic to Neoproterozoic.

The presence of two metasedimentary packages is suggested by a sheared unconformity between metaconglomerate and the underlying pelitic schist at locality B15 that may be traced westwards in outcrop and aeromagnetic data (Fig. 21) through to this stop and beyond. The unconformity is inferred to represent the contact between the Pooranoo Metamorphics to the north and the Morrissey Metamorphics to the south. A quartzite from the interpreted base of the Pooranoo Metamorphics has a maximum age of 1806 ± 3 Ma, whereas protoliths to the Morrissey Metamorphics are intruded by granites of the 1830–1780 Ma Moorarie Supersuite. A sample of pelitic schist from the Morrissey Metamorphics south of the inferred unconformity has a maximum depositional age of c. 1840 Ma (Varvell, 2001). Other than an early deformation event in the Morrissey Metamorphics, the two packages seem to share a common deformation and metamorphic history, with prograde regional metamorphic assemblages forming between 1030 and 990 Ma.

The purpose of Locality B11 is to look at a transect through medium-grade rocks in the lowermost part of the Pooranoo Metamorphics, in the Nardoo belt (Williams, 1986), to the contact with strongly sheared c. 1800 Ma Moorarie Supersuite granites in the south. Regionally, metasedimentary rocks of the Pooranoo Metamorphics and older Morrissey Metamorphics dip steeply to the south (Fig. 21), although at this locality the rocks dip moderately. The Morrissey Metamorphics structurally overlie the younger Pooranoo Metamorphics south of the Ti Tree Syncline, suggesting that the sequence in

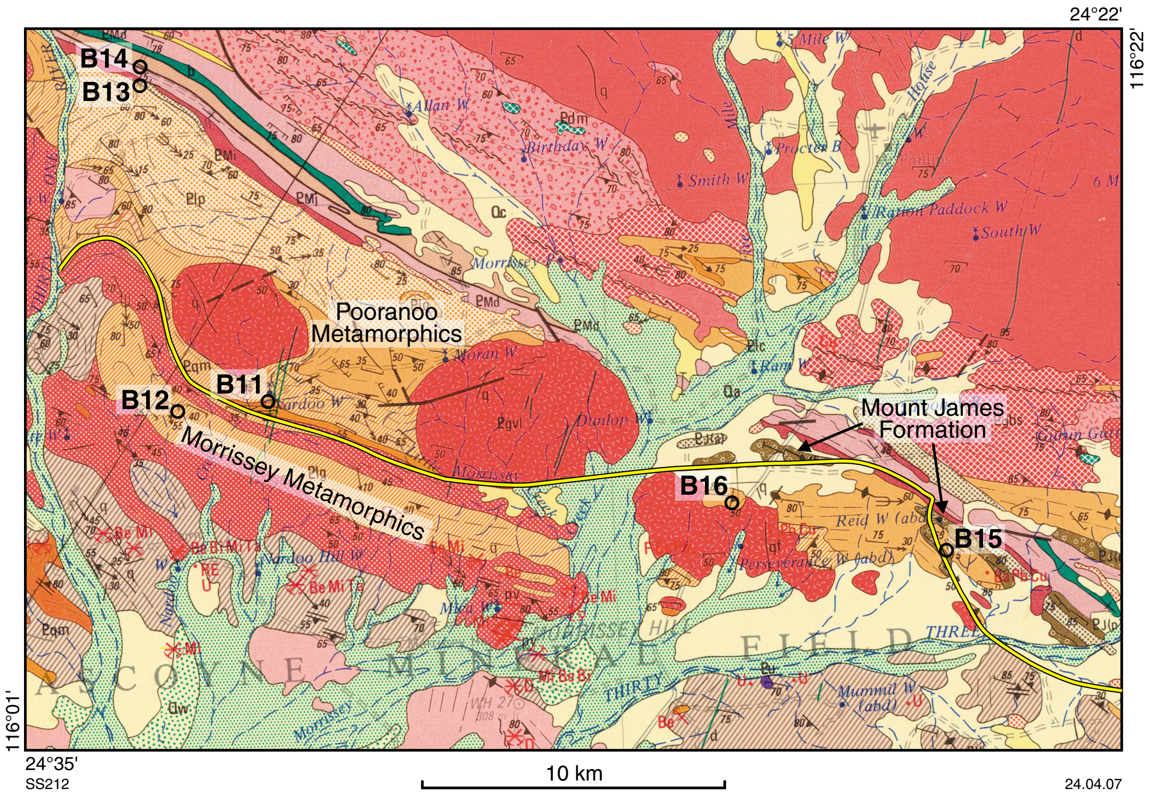

Figure 21. Map of the Nardoo Hills area showing the excursion localities. The figure consists of the first edition Mount Phillips 1:250 000 sheet (Williams et al., 1983) draped over the first vertical derivative of the total magnetic intensity. The dashed line shows the position of the interpreted unconformity between the Pooranoo Metamorphics to the north and the Morrissey Metamorphics. Siliciclastic rocks originally mapped as part of the Mount James Formation lie along strike from the Pooranoo Metamorphics

Figure 22. Crenulated schist of the Pooranoo Metamorphics with large garnet porphyroblasts (MGA 408980E 7293728N)

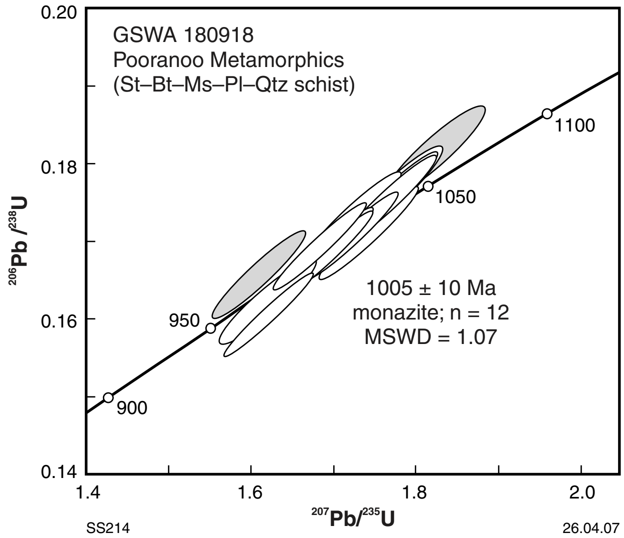

Figure 23. Concordia plot showing SHRIMP U–Pb data for metamorphic monazite from GSWA 180918 (Sheppard et al., 2007). Two shaded ellipses represent discordant data not used in age determination. Age uncertainty is 95% confidence limit and individual precision epllipses are 1σ

this area is overturned. Rare S–C fabrics in the quartzite at Locality B11c indicate thrusting to the north. These features are consistent with a large-scale recumbent fold verging to the north.

Locality B11a: Muscovite–staurolite–garnet schist (MGA 408973E 7293723N): At this locality the Pooranoo Metamorphics consist of medium-grade muscovite–biotite–staurolite–garnet(–plagioclase) pelitic schist with subordinate thin layers of cream-coloured garnet-bearing psammitic schist. The pelitic schist contains scattered dark red-brown garnet (Fig. 22) and black staurolite porphyroblasts, typically up to 20 mm in size. The porphyroblasts are enclosed by a strong composite fabric (S_{1-2}) and locally have remnants of S_1 in pressure shadows and inclusion tails. A prominent mineral lineation (L_2) is defined mainly by the alignment of 1–2 mm-sized staurolite crystals. The rocks have a localized crenulation with an associated weak crenulation cleavage (S_3).

In situ SHRIMP U–Pb dating of monazite in the foliation of a muscovite–biotite–staurolite–plagioclase schist, about 2 km west-northwest (GSWA 180918) along strike from this location, yielded a weighted mean age of 1005 ± 10 Ma (Fig. 23; Sheppard et al., 2007), which indicates that the medium-grade metamorphism in this area is related to the Edmundian Orogeny.

Locality B11b: Amphibolite (MGA 408931E 7293605N): Thin discontinuous layers of amphibolite are common in the Pooranoo Metamorphics along the southern margin of the Nardoo belt. They are locally associated with calc-silicate rocks and thin beds of metaconglomerate. Malachite staining has been observed in some of the amphibolite outcrops and scheelite mineralization is present in some locations. At this stop, however, the amphibolite is fine grained and compositionally homogeneous with a strong foliation and a weak mineral lineation. A narrow interval of strongly sheared porphyritic granite follows the southern boundary of the amphibolite.

Locality B11c: Quartzite of the Pooranoo Metamorphics (MGA 408923E 7293405N): The southern margin, and interpreted base, of the Pooranoo Metamorphics is marked in the Nardoo Well area by a distinctive, strongly outcropping foliated quartzite containing quartz (~90%) and accessory amounts of biotite, plagioclase, microcline, muscovite, epidote, and fine-grained opaque minerals. The accessory minerals are concentrated in foliation-parallel laminae, which are possibly relics of a primary bedding lamination. The quartzite has a strong planar foliation (S_{1-2}) dipping to the south-southwest at 30–40° (Fig. 24), a prominent lineation (L_2), and is locally interleaved

Figure 24. Strongly foliated quartzite at the interpreted base of the Pooranoo Metamorphics (looking east; MGA 408920E 7293430N)

with thin layers of deformed granite. The lineation is similar in orientation to the mineral lineation in the pelitic schists to the north, but the foliation dips more shallowly to the south approaching the southern margin of the Pooranoo Metamorphics, which is interpreted to be in sheared contact with leucocratic granites of the Moorarie Supersuite.

A quartzite sampled from Nardoo Creek (GSWA 191994), 500 m to the west, for SHRIMP U–Pb geochronology contained one zircon at 1782 ± 14 Ma (1σ) and 34 analyses of 34 zircons at 1806 ± 3 Ma, which is interpreted as the maximum depositional age of the rock. This sample has a similar maximum age and provenance to protoliths to psammitic gneiss samples from the Pooranoo Metamorphics in the Mangaroon zone north of the Minnie Creek batholith (Sheppard et al., 2005).

Locality B11d: Sheared leucogranite of the Moorarie Supersuite (MGA 408838E 7293140N): At this locality there are low exposures of a very strongly deformed, sheet-like body of leucocratic biotite monzogranite. The monzogranite contains sparse porphyroclasts of K-feldspar to 4 mm, and in some areas, boudinaged and isoclinally folded (F_2) pegmatite veins. The foliation in the monzogranite dips at about 50–60° to the south in this location and follows the curvilinear trend of the body, parallel to the S_{1-2} fabric in the Pooranoo Metamorphics to the north. In areas of better exposure the monzogranite displays a diffuse, ?magmatic layering defined by subtle variations in grain size, modal mineralogy, and abundance of coarser feldspars.

SHRIMP U–Pb zircon geochronology of a sample of the monzogranite (GSWA 191995) from 5 km to the northwest yielded a date of 1795 ± 7 Ma (Geological Survey of Western Australia, 2006).

A general trend in the sheet-like monzogranite is of increasing grain size and decreasing strain to the south and west, away from the contact with the Pooranoo Metamorphics. Combined with the southward increase in metamorphic grade in the Pooranoo Metamorphics, the dominance of D_2 structures in this zone, and the south to south-southwest dip of the foliation, this suggests that the contact between the monzogranite and the Pooranoo Metamorphics was a zone of uplift during D_2.

Locality B12: Medium-grade schists of the Morrissey Metamorphics (MGA 405782E 7293542N)

Drive west along the track, take the left fork at 2.1 km, and proceed a further 1.4 km before stopping at some low outcrops on the west side of a small gully.

The dominant rock type at this locality is coarse-grained pelitic schist (Fig. 25a) containing the assemblage quartz–muscovite–biotite–staurolite. Garnet is present in some parts of the unit. The staurolite porphyroblasts attain a maximum size of about 30 mm (Fig. 25b) and commonly have penetration twins. The rock has a strong mineral lineation and scattered coarse-grained quartz-rich aggregates, interpreted to be boudinaged quartz veins, enclosed by the foliation.

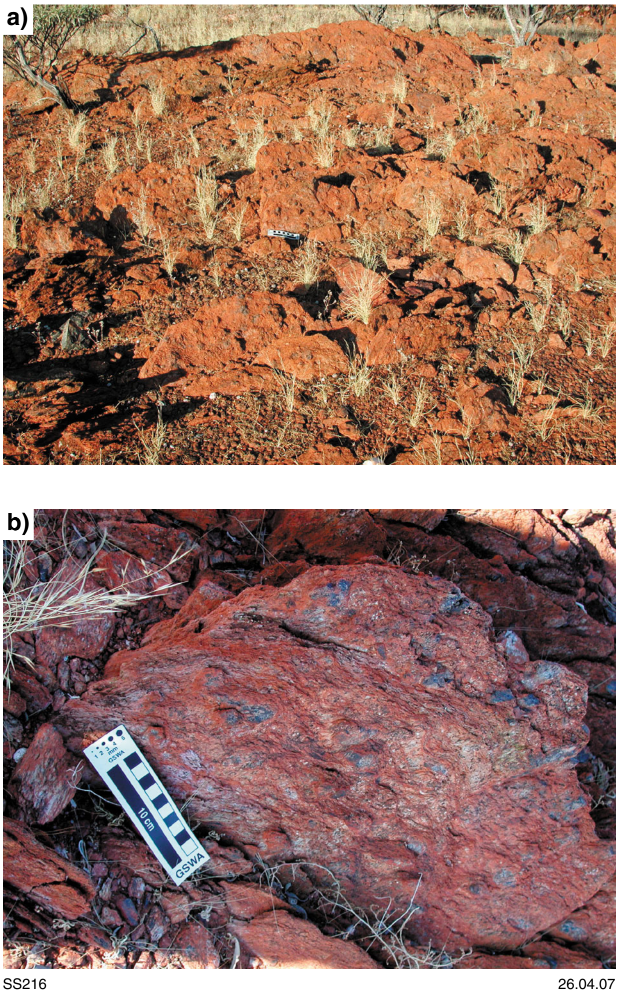

Figure 25. Quartz–muscovite–biotite–staurolite schist of the Morrissey Metamorphics: a) outcrops; b) close-up view of pelitic schist showing staurolite porphyroblasts (MGA 405775E 7293530N)

A sample of the pelitic schist from this site (GSWA 191977) yielded an in situ SHRIMP U–Pb monazite date of 1004 ± 8 Ma (Fig. 26; Sheppard et al., 2007), which is within error of the monazite date from the Pooranoo Metamorphics and indicates that medium-grade metamorphism of the two units was essentially coeval in this area.

From Locality B12 continue about 700 m west along track and camp in open country on the north side of the track.

Locality B13: Low-grade schists of the Pooranoo Metamorphics (MGA 405127E 7304279N)

From the campsite drive west along the track for about 3.3 km to White Well. Turn north just past the well, cross Thirty One River after 3.4 km and take the right fork after a further 0.5 km. Follow this track to the north for 5 km to Louden Well. Veer right at the well and follow the track for another 5 km. Stop on the east side of the track.

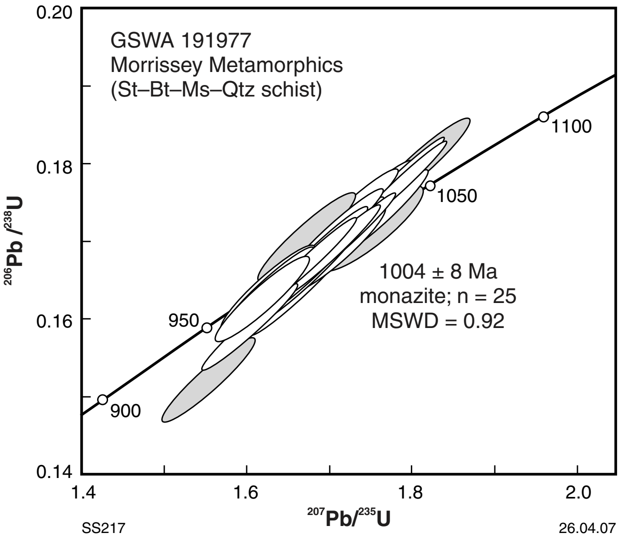

Figure 26. Concordia plot showing SHRIMP U–Pb data for metamorphic monazite from GSWA 191977 (Sheppard et al., 2007). Four shaded ellipses represent discordant data not used in age determination. Age uncertainty is 95% confidence limit and individual precision epllipses are 1σ

The metasedimentary rocks of the Morrissey Metamorphics and Pooranoo Metamorphics show a systematic decrease in metamorphic grade north towards the Ti Tree Syncline. In low-grade areas of the Nardoo belt, near the Ti Tree Syncline, the Pooranoo Metamorphics are typically fine-grained rocks comprising mainly slate and phyllite. The rocks have a pervasive cleavage (S_3), which is axial planar to kilometre-scale folds in the Pooranoo and Morrissey Metamorphics, and broadly parallel to structures in Edmund Group rocks within the Ti Tree Syncline.

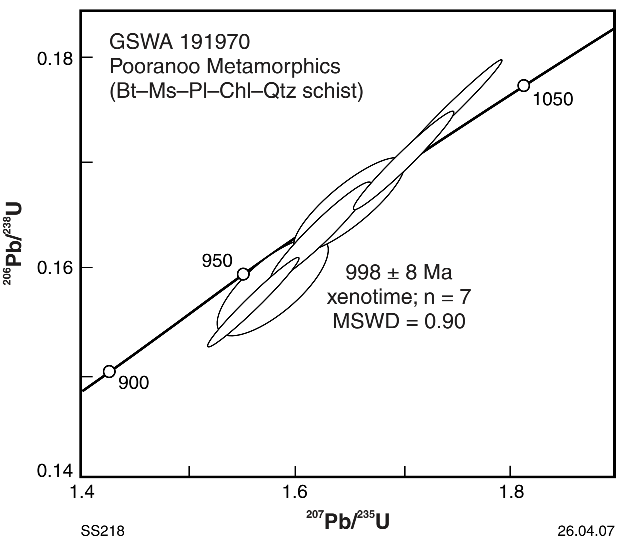

Figure 27. Concordia plot showing SHRIMP U–Pb data for metamorphic xenotime from GSWA 191970 (Sheppard et al., 2007). Age uncertainty is 95% confidence limit and individual precision epllipses are 1σ. Only the first analysis on each spot is plotted

This locality is about 120 m south of the faulted contact with Edmund Group rocks in the Ti Tree Syncline. The typical assemblage in the slaty rocks of the Pooranoo Metamorphics in this area is chlorite–muscovite–quartz (–albite–magnetite). The magnetite commonly forms small black porphyroblasts up to 2 mm. The rocks have a pervasive, planar continuous cleavage (S_3). Xenotime crystals aligned within this fabric in a sample (GSWA 191970) to the southeast of this locality yielded a SHRIMP U–Pb age of 998 ± 8 Ma (Fig. 27; Sheppard et al., 2007). At this locality the rocks are relatively uniform in composition, apart from a few more-siliceous zones. They commonly have dismembered, tight to isoclinally folded (F_3) quartz veins with axes subparallel to a weak mineral lineation (L_3). There is probably only a small contrast in metamorphic grade between the Pooranoo Metamorphics at this locality and Edmund Group rocks in the Ti Tree Syncline.

Locality B14: Low-grade dextral structures in Edmund Group rocks in the Ti Tree Syncline (MGA 405080E 7305039N)

Drive north along the track for 0.8 km. Stop on the west side of the track

The purpose of this stop is to examine evidence for late, post-D_3 dextral shearing in rocks of the Ti Tree Syncline. Narrow shear zones with dextral strike-slip kinematics are developed in rocks of the Bangemall Supergroup and Gascoyne Complex over a wide area. The shear zones strike east-southeast to southeast and cut north-northeasterly trending dolerite dykes of the c. 755 Ma Mundine Well swarm.

At this locality dark-grey siliceous, thin-bedded rocks of the Edmund Basin are foliated (S_3) and cut by numerous small-scale, anastomosing faults and striated surfaces that are concentrated in narrow zones where bedding is disrupted, folded, and rotated. Dextral indicators are best developed in fault-bound zones, typically less than 0.5 m wide, that show curvature of S_0 into the faults with dextral asymmetry, steep-plunging Z-shaped folds, and shallow-plunging fibres on slip surfaces.

Locality B15: Sheared ?unconformity between c. 1680 Ma Pooranoo Metamorphics and c. 1830 Ma Morrissey Metamorphics near Reid Well (MGA 430812E 7287905N)

Drive north along the station track for 1.4 km to Ti Tree Well. Turn east at the well and follow the track for 28.7 km to the Dairy Creek – Cobra road. Turn right and drive south along the road for 12.1 km. Turn east onto a station track just south of the cattle grid. Follow the track

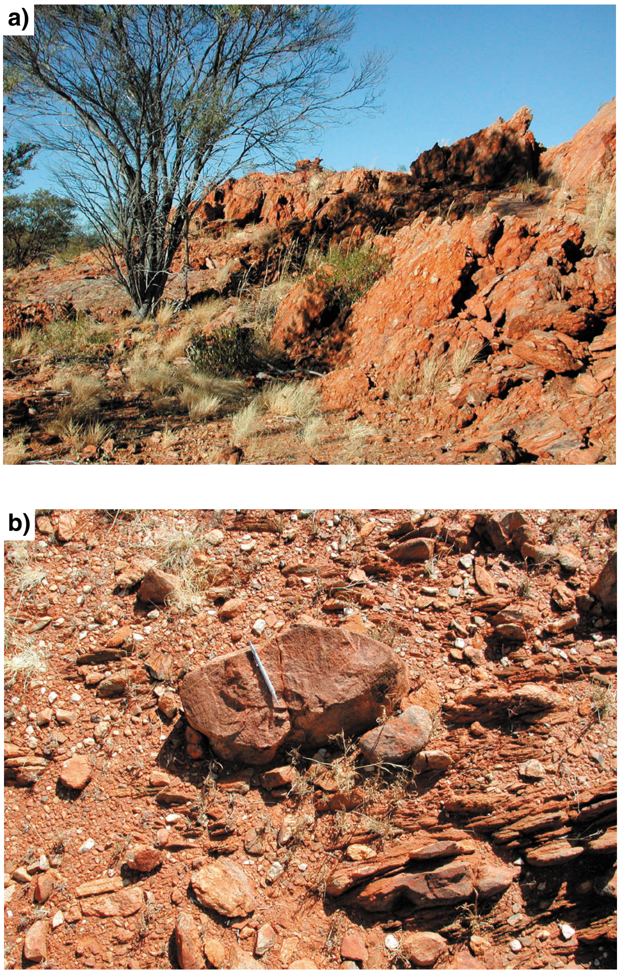

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Figure 28. Basal metaconglomerate of the Pooranoo Metamorphics near Reid Well. These rocks were formerly mapped as the Mount James Formation by Williams et al. (1983): a) outcrop (MGA 430760E 72879545); b) typical granite clasts near the basal unconformity of the Pooranoo Metamorphics

eastward to Reid Well at 6.7 km. Go past the well and turn south off the track 500 m beyond the well. Drive south-southwest cross country for about 500 m to the northeast side of a thickly wooded creek. Walk across the creek to the southwest for about 440 m, to the crest of the ridge.

At this locality there is a prominent outcrop of a strongly foliated metaconglomerate from the base of the Pooranoo Metamorphics, along the contact with the Morrissey Metamorphics. In contrast to the lithostratigraphic relationships at Nardoo well (Locality B11), the sheet-like Moorarie Supersuite monzogranite is absent and the two metamorphic units are in sheared contact. The metaconglomerate (Fig. 28a) contains quartz-vein clasts, and rare tourmalinized clasts of schist, indicating a metamorphic source area. Locally, the conglomerate contains cobbles of medium-grained granite (Fig. 28b). The foliation is oblique to the general trend of bedding, and there is a strong elongation or rodding of the clasts parallel to a mineral lineation.

On the south side of the conglomerate, semipelitic schists of the Morrissey Metamorphics contain muscovite and local biotite and garnet. A foliation and elongation lineation in the schists is similar in orientation to that in the metaconglomerate, indicating that the two units have undergone peak metamorphism in the same event. Two generations of late crenulations are present in the Morrisey Metamorphics.

Locality B16: Crenulated schist of the Morrissey Metamorphics and tourmaline granite (MGA 424033E 7289512N)

Return to the Dairy Creek – Cobra road and drive south along the road for 1.8 km. Continue driving for about 26.6 km towards Dairy Creek Station. Park in the open area on the right-hand side of the road. Walk across to exposures of crenulated schist.

Protoliths to metasedimentary rocks of the Morrissey Metamorphics were probably deposited after about c. 1840 Ma (Varvell, 2001), before being intruded by granites of the Moorarie Supersuite dated at 1810–1780 Ma in the central and northern Gascoyne Complex. Rafts of folded metasedimentary schist with two metamorphic fabrics are included within undeformed granites of the Minnie Creek batholith, indicating that the metasedimentary rocks were first deformed during the Capricorn Orogeny. Upright southeast-trending folds in the schists that are cut by granites of the Minnie Creek batholith are coplanar with F_3 folds south of the Ti Tree Shear Zone. However, monazite aligned within the S_2 fabric of the schists south of the shear zone yield in situ SHRIMP U–Pb ages of 1030–990 Ma (Sheppard et al., 2007), indicating that regional metamorphic assemblages and structures in the central Gascoyne Complex formed during the Edmundian Orogeny.

Structures formed during the Edmundian Orogeny in this area are cut by three plutons of massive biotite–muscovite–tourmaline granite, as well as a belt of pegmatites associated with rare-element mineralization (mainly Be and Ta–Nb). One of these granite plutons was interpreted to have an igneous crystallization age of 1652 ± 4 Ma (Culver, 2001), but this age is now thought to record a population of xenocrystic zircons in a low-Zr melt. One of the pegmatites, which contains beryllium–tantalum–niobium mineralization, has been dated at 954 ± 12 Ma using SHRIMP U–Pb in monazite (Sheppard et al., 2007). This age, and the field relationships of other dykes in the belt, which cut F_3 folds, suggests that all the rare earth element pegmatites are Neoproterozoic in age.

The crenulated muscovite–quartz–chlorite–magnetite schists exposed at this locality are typical of much of the Morrissey Metamorphics throughout the central Gascoyne Complex. Porphyroblasts of reddish-brown retrogressed ?andalusite up to 1 cm long are common. Some layers also contain clots up to 3 mm in diameter of retrograde chlorite. Here, pelitic schists are exposed in the hinge of a mesoscopic synform plunging moderately to the east-southeast. The crenulation cleavage (Fig. 29a) is parallel to the axial surfaces of the folds, which are interpreted to have formed during the Edmundian Orogeny. Magnetite porphyroblasts are enclosed by S_1 and have quartz pressure shadows, but the porphyroblasts also grow across S_1, suggesting growth during fabric development. Thin tourmaline–quartz veins cut the S_1 and S_2 fabrics, and are spatially associated with the Neoproterozoic tourmaline-bearing granite immediately to the south. The northern part of the granite and the contact between the schists and the granite are marked by intense tourmaline alteration (Fig. 29b).

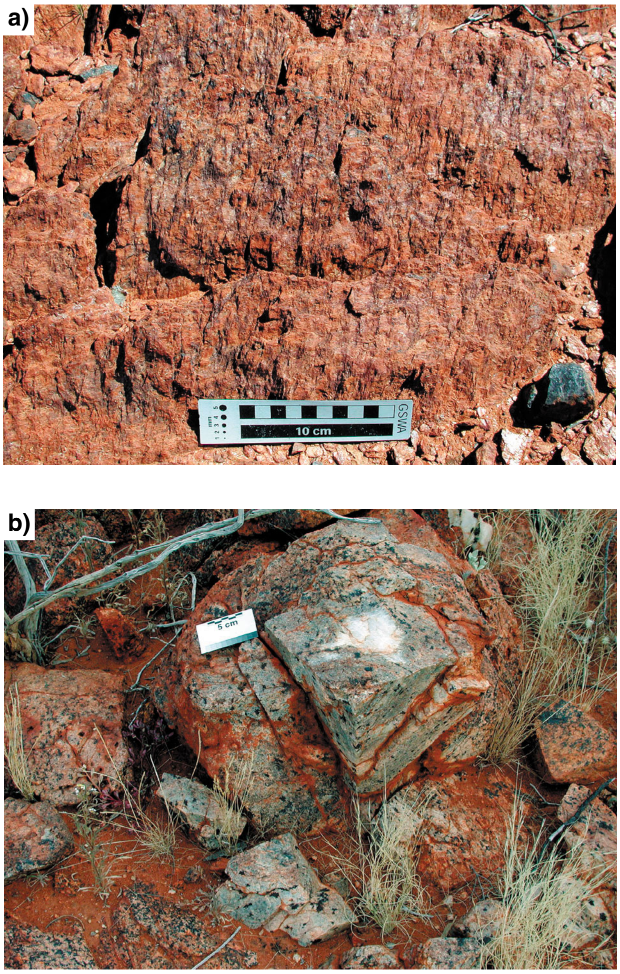

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Figure 29. Crenulated schist of the Morrissey Metamorphics and tourmaline granite: a) steeply dipping cleavage crenulating a schistosity oriented parallel to the scale bar. The crenulation cleavage is parallel to the axial surface of an F_3 fold (MGA 424033E 7289512N); b) tourmalinized and silicified Neoproterozoic granite adjacent to the contact with schists of the Morrissey Metamorphics (MGA 424069E 7289338N)

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— a field guide

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GEOLOGICAL SURVEY OF WESTERN AUSTRALIA
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Simplified geological map of the western Capricorn Orogen, showing the excursion route and localities

A9 Excursion locality
Mineral prospect or mine
Homestead
Bore or well
Topographic feature
Road
Creek
Fault
WYLOO 1:100 000 map sheet

Southern Carnarvon Basin
- Lyons Group
- Siliciclastic rocks, Devonian

- Dolerite dyke, sill, and plug
- Perseverance Supersuite
- Dolerite and gabbro sills intruded into Edmund Group and Collier Group

Collier Group
- Depositional package 6 (Ilgarari Formation)
- Depositional package 5 (Backdoor and Calyie Formations)

- Dolerite and gabbro sills intruded into Edmund Group

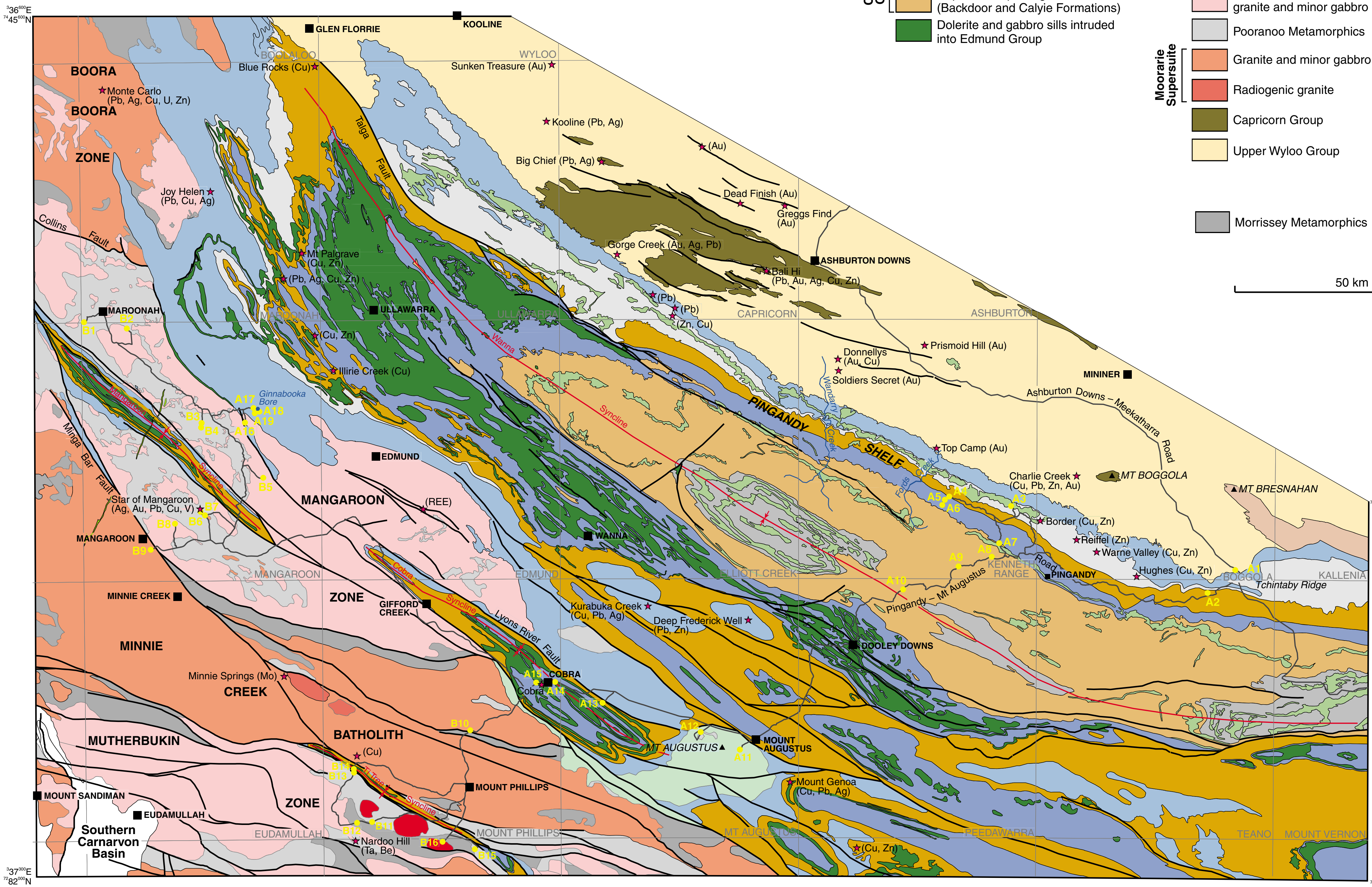
Edmund Group
- Depositional package 4 (Discovery, Devil Creek, Ullawarra, and Coodardoo Formations)
- Depositional package 3 (Kiangi Creek and Muntharra Formations)
- Depositional package 2 (Gooragoora, Blue Billy, and Cheyne Springs Formations)
- Depositional package 1 (Yilgatherra and Irregully Formations)

- Mount Augustus Sandstone
- Bresnahan Group
- Durlacher Supersuite, undivided: granite and minor gabbro
- Pooranoo Metamorphics

Moorarie Supersuite
- Granite and minor gabbro
- Radiogenic granite

- Capricorn Group
- Upper Wyloo Group
- Morrissey Metamorphics

50 km



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GEOLOGICAL SURVEY OF WESTERN AUSTRALIA
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Time–space distribution of bedrock geological units in the western Capricorn Orogen