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Chapter 5: Pine Creek Orogen

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Chapter 5: PINE CREEK OROGEN

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INTRODUCTION

The Pine Creek Orogen (PCO) is exposed over 47 500 km^2 and comprises a thick (>4 km) succession of Palaeoproterozoic clastic, carbonate and carbonaceous sedimentary and volcanic rocks, unconformably overlying Neoarchaean (ca 2670–2500 Ma) granitic and gneissic basement. These experienced regional metamorphism and deformation of varying grades and intensities in different parts of the orogen in the period ca 1867–1850 Ma and syn- to post-tectonic granite emplacement at ca 1830–1800 Ma. The Palaeoproterozoic strata of the PCO are extensively mineralised, hosting over 1000 known mineral occurrences. It is a multi-commodity province with the major commodities including gold, uranium, lead-zinc-silver, platinum-group elements, copper-cobalt-nickel, iron, tin-tantalum-tungsten and phosphate. Notably, the PCO contains ca 20% of the world's low-cost uranium resources (Ahmad 1998, 2007).

The PCO is broadly correlated with other Palaeoproterozoic domains of Northern Australia (eg Tanami and Tennant regions, Arnhem and Murphy provinces and the Halls Creek Orogen). Younger strata conceal its margins and the total extent is unknown (**Figure 5.1**). It is unconformably overlain by the Palaeo to Mesoproterozoic McArthur Basin to the east, the Palaeo to Neoproterozoic Victoria, Birrindudu and Fitzmaurice basins to the west, and Cambro–Ordovician and Mesozoic successions (Daly and Bonaparte basins) to the southwest and northwest. Mesozoic and Palaeozoic sediments of the Arafura and Money Shoal basins overlie parts of the PCO to the north and the Mesozoic onshore Carpentaria Basin overlies other parts to the south. Unconsolidated sand, silt and clay of Cenozoic age also cover much of the area. During the upper Cretaceous to mid Cenozoic, much of the land surface was subjected to intense chemical weathering, leading to laterite formation.

The PCO has been broadly subdivided into three domains (Needham *et al* 1988, Hollis *et al* 2009a) and this subdivision is used in this volume. These are the greenschist-facies Central Domain (including the Rum Jungle region and the South Alligator Valley), the amphibolite-facies Nimbuwah Domain to the east, and the amphibolite- to granulite-facies Litchfield Domain (formerly 'Litchfield Province') to the west (**Figure 5.2**). In addition to differences in metamorphic grade and structural styles, these regions are also distinct in the timing and nature of metamorphism and the timing and chemistry of the main phases of magmatism.

Needham *et al* (1980) subdivided the PCO into seven structural entities: Chilling Platform, Western Fault Zone, Batchelor Shelf, South Alligator Trough, South Alligator Hinge Zone, Nanambu High and Kakadu Shelf (**Figure 5.3**). These have little or no expression either in the regional-scale gravity or the airborne magnetic data.

Ahmad and McCready (2001) also subdivided the orogen into four depositional domains, on the basis of basement highs identified in the gravity data. From west to east these are the Batchelor Shelf, the Central Trough, the Nanambu Shelf, and the Eastern Trough (**Figure 5.4**). In the gravity data, the Central Trough appears as a northwest-trending depression between the South Alligator River Valley and Darwin, and the Eastern Trough as a narrow belt east of Jabiru. Depth to basement (granite and gneisses) in these centres is up to 4 km. Elsewhere, it is less than 2 km, with basement exposed at Rum Jungle, Nanambu (west of Jabiru), west Arnhem Land, and subcropping at Woolner (see **Archaean**). The interpretation of basement does not distinguish between Archaean rocks and intrusive granites.

The Palaeoproterozoic stratigraphic succession of the PCO (**Figures 5.2**, **5.5**) comprises the Woodcutters and Cosmo supergroups (Ahmad and McCready 2001). These correspond, respectively, to the P1–P2 and P3–P4 divisions of Ahmad (2000). No strata of equivalent age to the Woodcutters Supergroup are recognised outside of the PCO in the Northern Territory. The succession is well constrained in the Central Domain with more tentative correlations into the Litchfield Domain and Nimbuwah Domain. Needham *et al* (1988) attributed basin development and deposition of Woodcutters Supergroup-aged strata in the Central Domain to rifting of and deposition onto Archaean basement, the timing of which is constrained to ca 2020 Ma, based on U-Pb zircon dating of the Wildman Siltstone and Stag Creek Volcanics (Worden *et al* 2008b).

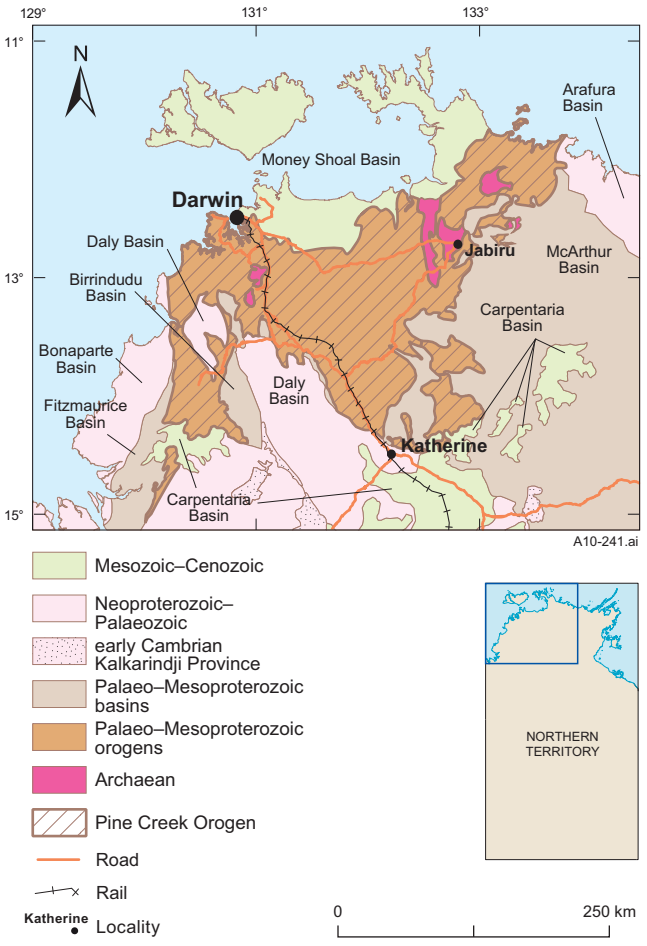


Figure 5.1. Geological setting of PCO. Inset shows location of PCO within Northern Territory.

The Woodcutters Supergroup is represented by the Manton, Mount Partridge, and Namoona groups in the Central Domain and by the Kakadu Group in the Nimbuwah Domain (Ahmad and McCready 2001, Needham *et al* 1980, 1988, Stuart-Smith *et al* 1980). Detrital zircon studies indicate a similar, dominantly Neoarchaean provenance for Woodcutters Supergroup sedimentary rocks (Cross *et al* 2005, Worden *et al* 2008a, Hollis *et al* 2009b). The Cahill Formation has also been proposed as a possible equivalent of the upper Woodcutters Supergroup in the Nimbuwah Domain (Needham *et al* 1980, Stuart-Smith *et al* 1980), although recent U-Pb detrital zircon work indicates a distinct provenance from similarly aged sedimentary rocks of the Central Domain (Hollis *et al* 2009b). Rocks of similar

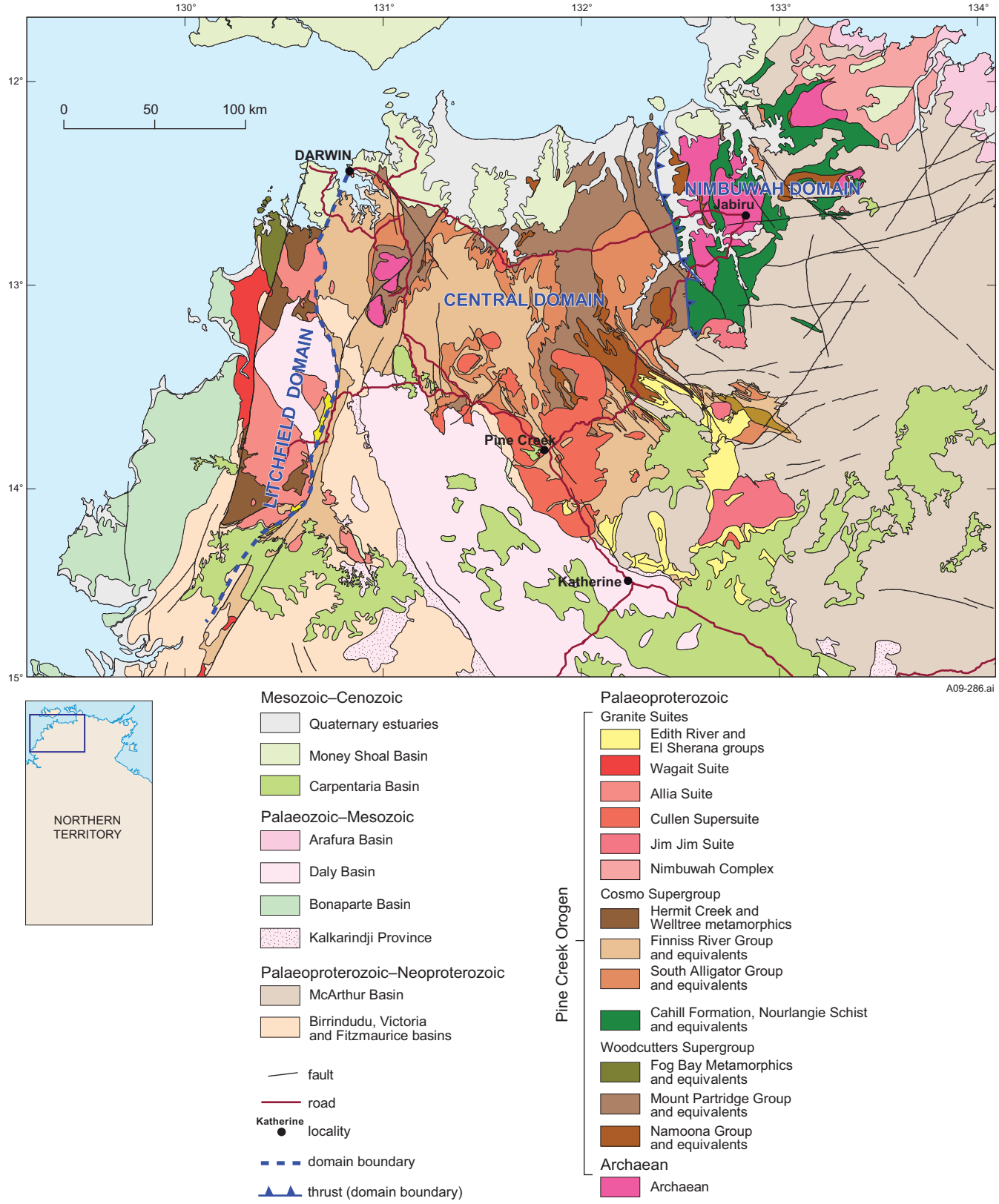


Figure 5.2. Generalised geology of PCO. Greenschist-facies Central Domain is separated from amphibolite-facies Nimbuwah Domain to east by Jim Jim Fault Zone. Boundary with amphibolite- to granulite-facies Litchfield Domain to west defined by westernmost extent of Finniss River Group.

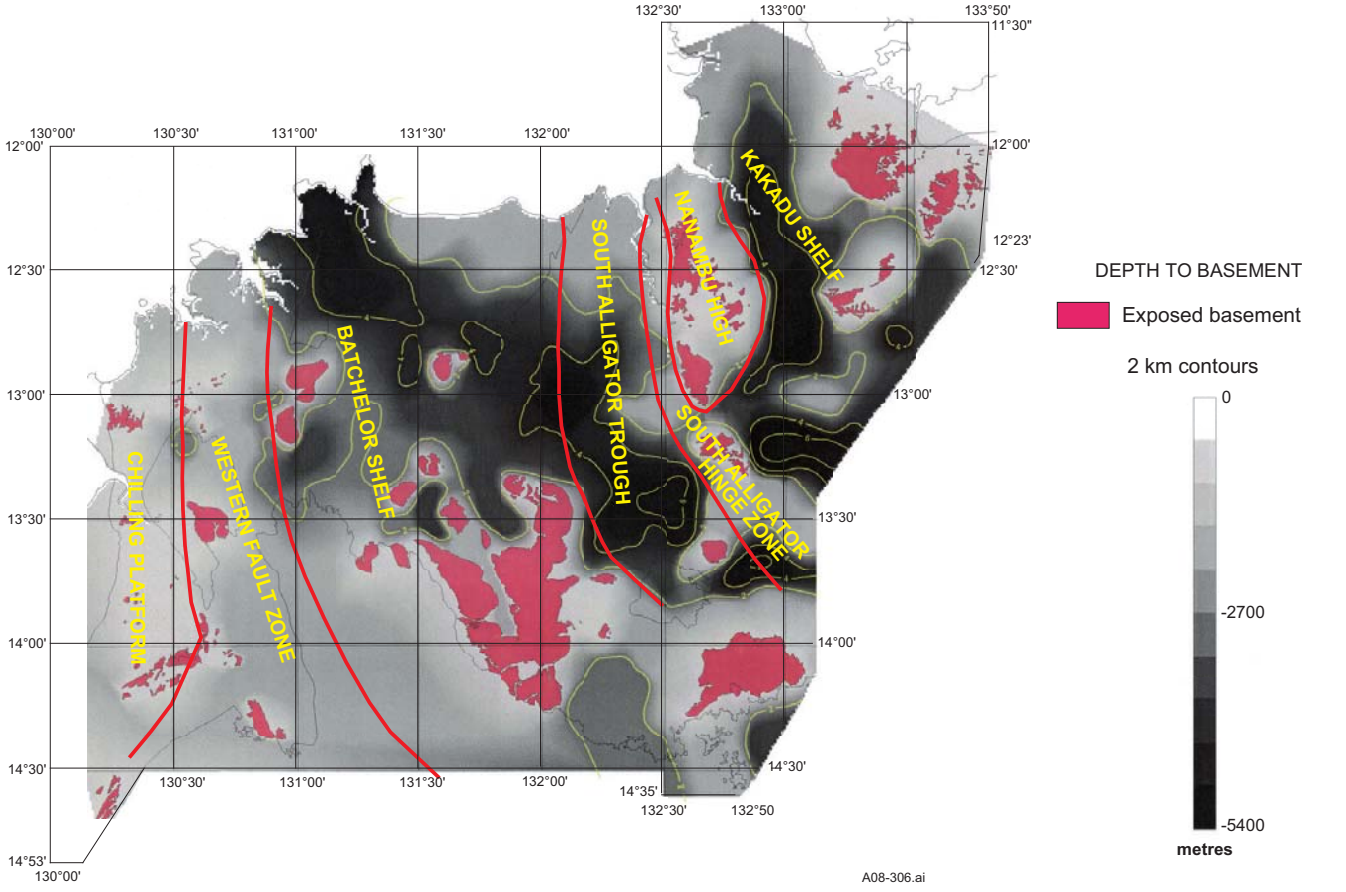


Figure 5.3. Tectonic subdivisions of PCO proposed by Needham *et al* (1980) superimposed on depth-to-basement interpretation derived from regional gravity data (Lewis *et al* 1995). After Ahmad and McCready (2001).

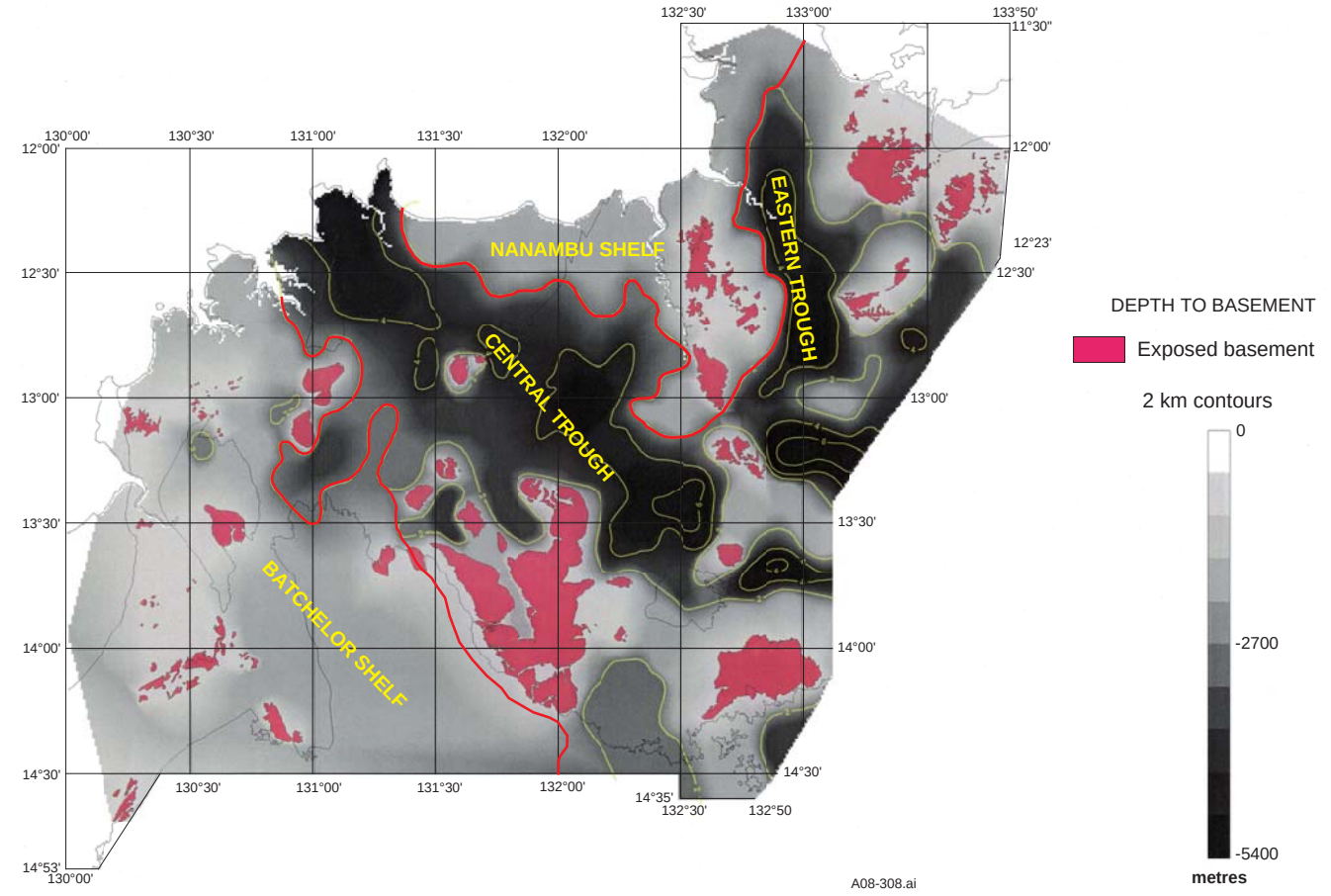


Figure 5.4. Palaeogeographic subdivisions of PCO of Ahmad and McCready (2001) superimposed on depth-to-basement interpretation derived from regional gravity data (Lewis *et al* 1995).

age to the Woodcutters Supergroup may be represented by the Fog Bay Metamorphics in the Litchfield Domain, based on detrital zircon spectra dominated by Neoarchaean and early Palaeoproterozoic ages with a maximum depositional age of ca 2028 Ma (Carson *et al* 2009). Otherwise, strata of this age are not currently recognised in the Litchfield Domain.

The Cosmo Supergroup is more widespread than the underlying Woodcutters Supergroup and unconformably overlies the latter (Walpole *et al* 1968, Needham *et al* 1980, 1988, Stuart-Smith *et al* 1980, Ahmad and McCready 2001). The unconformable contact represents a time gap of ca 160 Ma, as is shown by U-Pb zircon ages of ca 2020

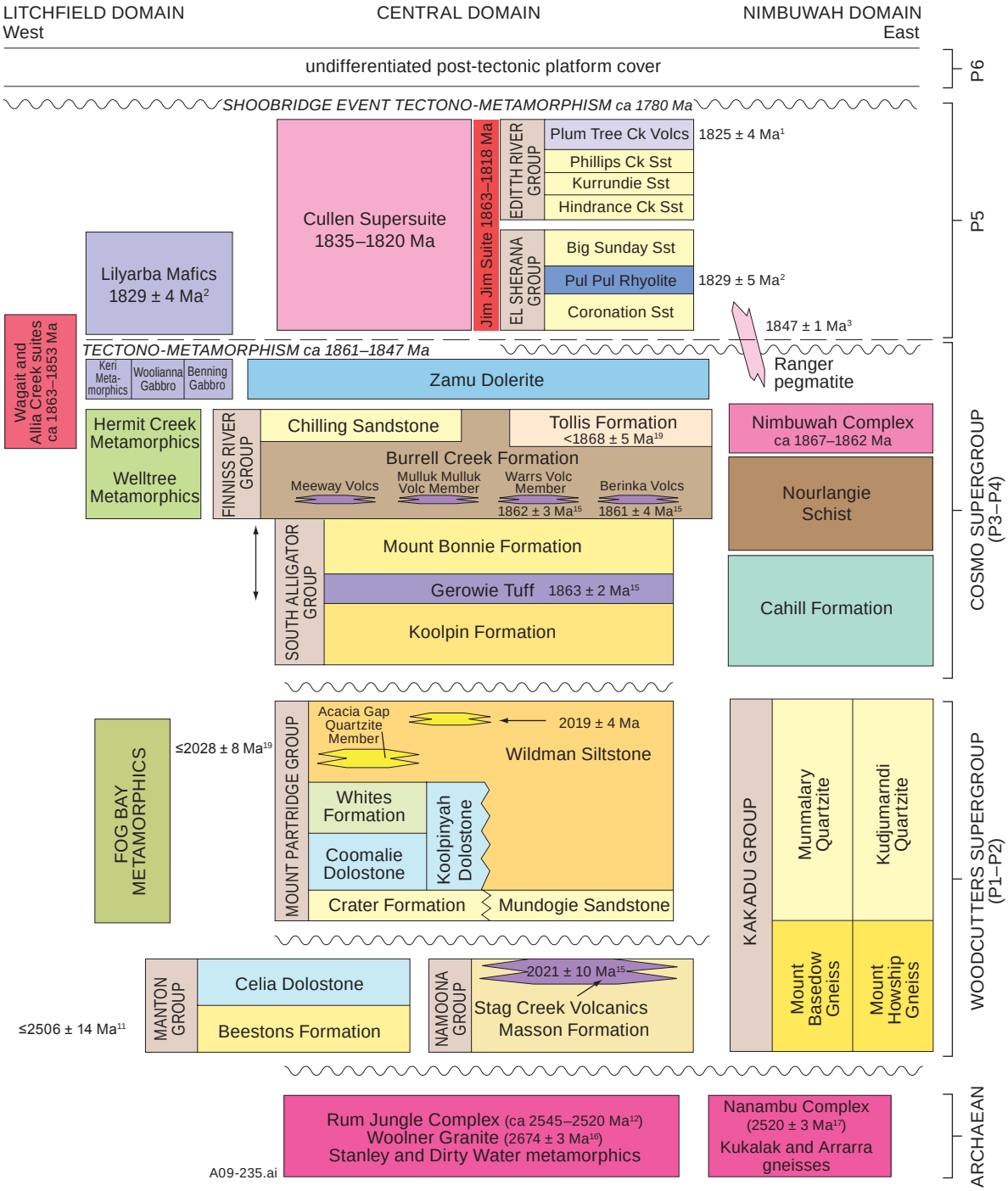


Figure 5.5 Simplified stratigraphic column of PCO (modified from Worden *et al* 2008b). Superscripts to geochronology refer to: (1) SHRIMP zircon extrusion age (Page 1996c); (2) SHRIMP zircon extrusion age (Jagodzinski 1998); (3) ID-TIMS monazite intrusion age (Annesley *et al* 2002); (4) SHRIMP zircon emplacement age (Stuart-Smith *et al* 1993, Page 1996e); (5) ID-TIMS zircon and xenotime zircon and xenotime emplacement age (Page *et al* 1985); (6) SHRIMP zircon emplacement age (Page 1966b); (7) SHRIMP zircon upper intercept deposition age (Page and Williams 1988); (8) various ID-TIMS zircon upper intercepts emplacement/metamorphic ages (Page *et al* 1980); (9) ID-TIMS zircon upper intercept deposition age (Needham *et al* 1988); (10–11) SHRIMP zircon maximum depositional ages (Cross *et al* 2005); (12) SHRIMP zircon emplacement age (Cross *et al* 2005); (13) SHRIMP zircon emplacement age (McAndrew *et al* 1985); (14) ID-TIMS zircon emplacement age (Page *et al* 1980); (15) SHRIMP zircon deposition age (Worden *et al* 2008b); (16) SHRIMP zircon emplacement age (Glass *et al* 2010); (17) SHRIMP zircon emplacement age (Hollis *et al* 2009c); (18) Carson *et al* 2009); (19) SHRIMP zircon maximum deposition age (Worden *et al* 2008b). Ck = Creek; Sst = Sandstone; Volc = Volcanic; Volcs = Volcanics. P1–6 terminology after Ahmad (2000).

and 1863 Ma for volcanic and tuff units within the two Supergroups (Worden *et al* 2008a, b). The Cosmo Supergroup succession comprises the South Alligator Group and the regionally extensive Finniss River Group in the Central Domain, with the Hermit Creek and Welltree metamorphics representing probable equivalents of the Finniss River Group in the Litchfield Domain (Worden *et al* 2008a). The Nourlangie Schist has been proposed as a possible equivalent of the Cosmo Supergroup in the Nimbuwah Domain (Needham *et al* 1980, Stuart-Smith *et al* 1980); however, recent detrital zircon data indicate distinct differences in provenance compared with Cosmo Supergroup strata in the Central Domain (Hollis *et al* 2009b).

In contrast to the historical grouping of tectonometamorphism in the NAC during the period ca 1880–1850 Ma into a single episode (Barramundi Orogeny), geochemical, petrographic, and U-Pb monazite and zircon studies in the PCO have isolated chronologically distinct thermal events in different regions, with different styles of metamorphism, and indications of significant crustal thickening and of subduction-related tectonism (Glass 2007, 2011, Worden *et al* 2008a, b, Carson *et al* 2008, Hollis *et al* 2009a, b). Therefore, the term 'Barramundi Orogeny' is abandoned here in favour of the ca 1865 Ma Nimbuwah Event and the ca 1855 Ma Litchfield Event (see also Hollis *et al* 2009a). The ca 1865 Ma Nimbuwah Event involved the emplacement of granodioritic Nimbuwah Complex plutons into, and amphibolite-facies, moderate-P metamorphism and deformation of granitic Archaean basement and overlying Palaeoproterozoic strata that were deposited at, or after ca 1900 Ma (Walpole *et al* 1968, Wyborn *et al* 1997, Worden *et al* 2008a, b, Hollis *et al* 2009a). This was coincident with felsic volcanism in the Central Domain and Litchfield Domain (Worden *et al* 2008a, b). In contrast, the ca 1855 Ma Litchfield Event involved emplacement of S-type granites of the Allia Creek Suite (Walpole *et al* 1968, Page *et al* 1980, Wyborn *et al* 1997, Worden *et al* 2008a), and amphibolite- to granulite-facies, low-P metamorphism of Palaeoproterozoic strata (Carson *et al* 2008), possibly associated with subduction-related mafic magmatism (Glass 2007, 2011).

In the Central Domain, detrital zircon data suggest that sedimentation of the Burrell Creek Formation continued beyond ca 1855 Ma and thus, that greenschist-facies metamorphism and tight, upright northwest-north-trending folding in this domain occurred after this time (Hollis *et al* 2009b), but prior to the 1829 Ma deposition of the El Sherana Group (Jagodzinski 1998, 1999). El Sherana Group sedimentary and volcanic rocks were deposited unconformably on the South Alligator Group in the South Alligator Valley (South Alligator Trough), a rift-related shallow northwest-trending graben structure (Needham and Stuart-Smith 1985a, Needham *et al* 1988, Jagodzinski 1998, 1999). Shortly after deposition, tight folding of the El Sherana Group ensued during the Maud Creek Event, prior to the deposition of disconformable, Edith River Group sedimentary and volcanic rocks at 1825 Ma (Needham and Stuart-Smith 1985a, Page 1996c). During the same period, the Cullen Event involved emplacement of I-type granites of the Cullen and Jim Jim suites in the Central Domain from ca 1835–1820 Ma (Walpole *et al* 1968, Stuart-Smith *et al* 1993, Bajwah 1994, Jagodzinski and Wyborn 1997, Wyborn *et al* 1997, Wyborn *et al* 2001). The El Sherana and Edith River groups both correspond to the P5 division of Ahmad (2000).

Subsequently, the Shoobridge Event resulted in widespread retrogression and the development of northwest-north-trending shear zones in the Central Domain at ca 1780 Ma, as shown by the resetting of Rb-Sr isotopic systems (Needham *et al* 1988). Significant uplift of the eastern part of the PCO, the Nimbuwah Domain, had occurred by ca 1723 Ma (Page 1996d), as is shown by the emplacement of lopoliths of the Oenpelli Dolerite, at 1–2 km depth, in rocks that had previously been buried to mid to deep crustal levels during the ca 1865 Ma Nimbuwah Event (Stuart-Smith and Ferguson 1978, Needham *et al* 1988, Hollis *et al* 2009a). Historically, the entirety of this geological evolution, from the ca 1865 Ma Nimbuwah Event through to either the emplacement of the Cullen and Jim Jim suites (Needham and DeRoss 1990) or through to the Shoobridge Event (Needham *et al* 1988), was referred to as the 'Top End Orogeny', but for the reasons already outlined, this terminology, along with the term 'Barramundi Orogeny', is abandoned here.

STRATIGRAPHIC SUCCESSION

Currently used stratigraphic subdivisions within the PCO are compared with earlier studies in **Table 5.1** and **Figure 5.5**, (Walpole *et al* 1968, Needham *et al* 1980, Needham and Stuart-Smith 1985b, Needham *et al* 1988, Pietsch and Edgoose 1988, Ahmad *et al* 2006, Worden *et al* 2008b). Detailed descriptions of stratigraphic units are provided in **Table 5.2**. Walpole *et al* (1968) and Needham *et al* (1988) considered that sedimentation within the PCO took place in a number of basins, characterised by different depositional settings. This subdivision was simplified by Ahmad and McCready (2001) into the Batchelor Shelf, Central Trough, Nanambu Shelf and Eastern Trough, the terminology employed here. Below, these strata are described in the context of the three main subdivisions of the PCO: the Central Domain, Litchfield Domain and Nimbuwah Domain.

CENTRAL DOMAIN

The Central Domain is the most well exposed and well characterised part of the PCO. The Palaeoproterozoic succession in the Central Domain unconformably overlies Neoarchaean basement, exposed as the Rum Jungle and Waterhouse domes (ca 2545–2520 Ma) and subcropping as the Woolner Granite (2674 Ma, see **Archaean**). The overlying succession varies from an inferred original thickness of ca 2 km to ca 18 km, deepening to the east (Ahmad *et al* 1993). It is subdivided into the Woodcutters Supergroup, deposited at or before ca 2020 Ma, and the Cosmo Supergroup, deposited at ca 1865–1855 Ma. These are separated by an unconformity representing at least 160 Ma (Ahmad *et al* 2006, Worden *et al* 2008a, b). Both supergroups are dominated by carbonate, carbonaceous, clastic, and volcaniclastic sedimentary rocks with subordinate mafic and felsic volcanic rocks.

Table 5.1 Generalised comparison of stratigraphic nomenclature of PCO followed in this study with those of previous publications. For details refer to **Table 5.2**.

Walpole *et al* (1968)			**Needham and Stuart-Smith (1984)**			**This study**		
Chilling Platform and Western Fault Zone	***Central Trough***	***Eastern Trough***	***Litchfield Block***	***Rum Jungle, Centre and South Alligator Valley***	***Alligator Rivers***	***Litchfield Domain***	***Central Domain***	***Nimbuwah Domain***
		Katherine River Group			**Katherine River Group**			**Katherine River Group**
	Edith River Volcanics	Edith River Volcanics	Chilling Sandstone Meeway Volcanics Berinka Volcanics Henschke Breccia	**Edith River Group** **El Sherana Group**			**Edith River Group** **El Sherana Group**	
GRANITE INTRUSIONS						**Allia Suite**	**Cullen Suite**	**Jim Jim Suite**
						LITCHFIELD EVENT 1855 Ma		
					BARRAMUNDI OROGENY 1880–1850 Ma			**NIMBUWAH EVENT 1865 Ma**
							COSMO SUPERGROUP	
Chilling Sandstone Berinka Volcanics			Welltree Schist Hermit Creek Metamorphics	**Finniss River Group** Burrell Creek Formation	**Nimbuwah Complex** Myra Falls Metamorphics Nourlangie Schist	Welltree Metamorphics Hermit Creek Metamorphics	**Finniss River Group** Chilling Sandstone Mulluk Mulluk Volcanics Warrs Volcanic Member Berinka Volcanics Burrell Creek Formation **South Alligator Group** Mount Bonnie Formation Gerowie Tuff Koolpin Formation	**Nimbuwah Complex** Nourlangie Schist upper Cahill Fm lower Cahill Fm
Finniss River Group Noltenius Formation	**Finniss River Group** Burrell Creek Formation	**South Alligator Group** Fisher Creek Siltstone Gerowie Chert Koolpin Formation		**South Alligator Group** Mount Bonnie Formation Gerowie Tuff Koolpin Formation				
							WOODCUTTERS SUPERGROUP	
	Goodparla Group Golden Dyke Formation	**Goodparla Group** Golden Dyke Formation		**Mount Partridge Group** Wildman Siltstone Mundogie Sandstone Whites Formation Coomalie Dolomite Crater Formation **Namoona Group** Stag Creek Volcanics Masson Formation Celia Dolomite Beestons Formation	Cahill Formation (upper) Cahill Formation (lower) **Kakadu Group**	Fog Bay Metamorphics	**Mount Partridge Group** Wildman Siltstone, Mundogie Sandstone Whites Formation, Coomalie Dolostone Crater Formation **Manton/Namoona groups** Stag Creek Volcanics Masson Formation Celia Dolostone Beestons Formation	**Kakadu Group** Munmarlary Quartzite Mount Howship Gneiss Kudjumarndi Quartzite Mount Basedow Gneiss
	Batchelor Group Coomalie Dolomite, Crater Formation Celia Dolomite, Beestons Formation	Masson Formation Mount Partridge Formation						
Hermit Creek Complex	**Rum Jungle Complex**	**Stag Creek Volcanics** **Nanambu Granite** **Myra Falls Metamorphics**		**Rum Jungle Complex** **Waterhouse Complex**	**Nanambu Complex**		**Rum Jungle Complex** **Stanley and Dirtywater metamorphics** **Woolner Granite**	**Nanambu Complex** **Kukalak Gneiss** **Arrarra Gneiss** **Njibinjibinj Gneiss**

Unit/thickness location	Lithology	Depositional environment
NIMBUWAH DOMAIN		
Nourlangie Schist <3000 m	Quartz-mica schist, minor quartzite.	Marine
Cahill Formation <400 m		
upper Cahill Formation	Quartzite, feldspar-quartz schist, garnet-mica schist, mica-quartz schist, minor conglomerate	
lower Cahill Formation	Quartzite, haematite-quartz-mica schist, pyritic carbonaceous schist, dolostone, magnesite, calc-silicate gneiss	
Kakadu Group		
Munmarlary Quartzite >200 m	Gneissic orthoquartzite, minor schist	Fluvial to shallow marine
Mount Basedow Gneiss >1500 m	Muscovite-biotite gneiss, granitoid gneiss, minor schist	Fluvial to shallow marine
Kudjumarndi Quartzite ca 150 m	Orthoquartzite, gneiss, minor schist	Fluvial to shallow marine
Mount Howship Gneiss >2000 m	Feldspathic leucogneiss, rare schist	
LITCHFIELD DOMAIN		
Welltree Metamorphics	Quartz-feldspar-muscovite schist and gneiss, minor quartzite, marble, graphitic schist, calc-silicate gneiss, marble	
Hermit Creek Metamorphics	Sillimanite gneiss, garnet-cordierite gneiss, biotite-quartz schist, metamorphosed mafic igneous rocks	
Fog Bay Metamorphics	Biotite gneiss, minor quartzite	
CENTRAL DOMAIN		
Geolsec Formation <400 m	Haematitic quartzite breccia, haematitic mudstone, siltstone and sandstone, minro phosphatic siltstone and breccia, fault scarp breccia	Shallow marine environment
Edith River Group		
Plum Tree Creek Volcanics ca 900 m South Alligator region	Massive, porphyritic rhyodacitic ignimbrite, glassy rhyolite, amygdaloidal andesite; minor lenses of pebbly quartz sandstone	Subaerial (volcanic rocks); minor braided fluviatile (sedimentary rocks)
Kurrundie Sandstone ca 350 m South Alligator region	Thickly bedded to massive, coarse to pebbly lithic quartz sandstone with interbeds of polymictic pebble to boulder conglomerate; minor siltstone	Fluviatile: juvenile braidplain
El Sherana Group		
Big Sunday Formation <480 m South Alligator region	Volcanilithic to arkosic sandstone, volcaniclastic siltstone, amygdaloidal basalt; rhyolite and rhyolitic ignimbrite in upper portion	Lacustrine (sedimentary rocks) and subaerial (volcanic rocks)
Pul Pul Rhyolite <1300 m South Alligator region	Felsic welded vitric and non-welded ignimbrite, polymictic volcaniclastic conglomerate, massive volcaniclastic sandstone, intrusive massive rhyolitic porphyry; minor siltstone lenses	Subaerial and intrusive (volcanic rocks); minor fluviatile (sedimentary rocks)
Coronation Sandstone <600 m South Alligator region	Cross-bedded, coarse to very coarse, quartz to sublithic sandstone; polymictic, pebble to cobble conglomerate; lithic sandstone; amygdaloidal basaltic lava; minor intrusive rhyolitic porphyry	Juvenile braided fluviatile (sedimentary rocks), subaerial volcanic rocks
Scinto Breccia ca 100 m South Alligator region	Pink-brown, quartz-haematite, polymictic pebble breccia, commonly phosphatic; minor haematitic sandstone	Subaerial talus
COSMO SUPERGROUP		
Dolerite bodies *Zamu Dolerite* <1500 m, *Goodparla Dolerite* <300 m	Grey, medium to coarse quartz dolerite; minor granophyre; rare biotite-hornblende-augite pegmatite	Intrusive
Finniss River Group		
Tollis Formation ca 1000 m South Alligator region	Poorly sorted, fine to coarse, volcaniclastic metagreywacke; laminated slate and mudstone; minor cherty shale and laminated black crystal tuff	Proximal mid-submarine fan
Chilling Sandstone ca 6500 m Western region	Quartz sandstone, minor conglomerate, siltstone and tuff	Shallow marine to fluvial
Henschke Breccia ca 1500 m Western region	Conglomerate, breccia, sandstone	Tallus breccia
Mulluk Mulluk Volcanics <2000 m Western region	Rhyolite and dacite	Aerial to subaqueous lava
Warrs Volcanic Member ca 500 m Western region	Acid volcanics, siltstone, volcaniclastic sandstone	Submarine lava

Table 5.2. Proterozoic stratigraphic succession of the Pine Creek Orogen (**continued next page**).

Unit/thickness location	Lithology	Depositional environment
Berinka Volcanics <75 m Western region	Rhyodacite, andesite, agglomerate, interlayered mudstone	Fluvial to subtidal
Dorothy Volcanics 200m Edith River region	Basalt, tuff	
Burrell Creek Formation ca 3000 m Central region	Siltstone, shale, greywacke	Proximal mid-submarine fan (turbidite)
South Alligator Group		
Mount Bonnie Formation 1200 m, Central region	Siltstone, shale, greywacke, crystal and vitric tuff, rare banded ironstone, dolomudstone	Shallow-marine inner shelf to deeper mid-shelf (turbidite)
Gerowie Tuff <750 m Central region	Siltstone, cherty shale, black cherty crystal tuff and lithic tuff; minor felsic ignimbrite, chloritic volcaniclastic shale, lapilli tuff; porphyritic dacite	Subaerial, shallow marine
Shovel Billabong Andesite <300 m South Alligator region	Green-grey, massive pitchstone; variolitic andesite, microdiorite	Submarine
Koolpin Formation <350 m Central and South Alligator region	Haematitic metasiltstone and phyllite with chert bands, lenses and nodules; graphitic phyllite and slate; silicified dolomudstone; minor quartz metagreywacke	Low-energy supratidal to subtidal shallow marine
WOODCUTTERS SUPERGROUP		
Mount Partridge Group		
Wildman Siltstone <2000 m Central and South Alligator regions	Carbonaceous shale, haematitic siltstone, quartz sandstone, quartzite, greywacke	Shallow marine: prodelta or mid-shelf
Mundogie Sandstone ca 1400 m Central and South Alligator regions	Feldspathic quartz sandstone, siltstone, conglomerate	Mainly braided fluviatile; shallow marine in upper portion
Yarrawonga Volcanic Member ca 40 m Central region	Dacite, ignimbrite, rhyolite	
Mount Dean Volcanic Member <150 m Rum Jungle region	Basic volcanic rocks	
Acacia Gap quartzite Member <300 m Rum Jungle region	Quartzite, sandstone, shale	Shallow marine
Whites Formation >1000 m Rum Jungle region	Calcareous and carbonaceous shale, dololutite, dolarenite, rare quartzite	Marine, euxenic
Koolpinyah Dolostone Central region	Dolomitic marble, dolomitic mica schist, dolomitic limestone, calcareous quartzite, conglomerate at base.	Shallow marine to fluvial
Coomalie Dolostone <600 m Rum Jungle region	Stromatolitic dolostone, magnesite and minor shale	Subtidal, evaporitic
Crater Formation <900 m Rum Jungle region	Conglomerate, arkose, sandstone, and minor siltstone and shale	Fluvial
Namoona Group		
Stag Creek Volcanics <1000 m South Alligator region	Basalt, andesite, agglomerate, volcaniclastic shale and sandstone	Submarine flows
Masson Formation <2800 m South Alligator and Central regions	Carbonaceous-pyritic and dolomitic shale, carbonaceous pyritic dolomudstone, quartz sandstone, feldspathic-quartz greywacke, dolostone , marble, minor mafic volcaniclastic shale and rhyolitic tuff	Mainly shallow marine (sub-wavebase)
Manton Group		
Celia Dolostone <300 m Rum Jungle region	Stromatolitic dolostone and magnesite	Subtidal, evaporitic
Beestons Formation <300 m Rum Jungle region	Quartz conglomerate, grit, arkose, quartz sandstone and minor siltstone	Fluvial
ARCHAEAN BASEMENT		
Archaean basement: **Central Domain** – Rum Jungle Complex (2545–2521Ma), Woolner Granite (2674 Ma), Dirtywater Metamorphics, Stanley Metamorphics; **Nimbuwah Domain** – Nanambu Complex (2520 Ma), Kukalak Gneiss (2527–2510 Ma), Arrarra Gneiss (2640 Ma), Njibinjibinj Gneiss (2671 Ma)		

Table 5.2. Proterozoic stratigraphic succession of the Pine Creek Orogen (**continued from previous page**).

Woodcutters Supergroup

In the Central Domain, the lower parts of the Palaeoproterozoic succession comprise the Manton Group (Batchelor Shelf), Namoona Group (Central Trough) and Mount Partridge Group (Batchelor Shelf, Central Trough, Nanambu Shelf). Collectively, these are termed the Woodcutters Supergroup and they unconformably overly Neoarchaean basement, exposed as the Rum Jungle and Waterhouse domes, and the sub-cropping Woolner Granite. The Woodcutters Supergroup typically has smooth REE patterns, consistent with provenance from calc-alkaline andesitic rocks (Ahmad and McCready 2001). It has detrital zircon spectra indicative of dominantly Neoarchaean sources, with no contribution from sources younger than ca 2000 Ma (Cross *et al* 2005, Hollis *et al* 2009a).

Manton and Namoona groups

The Manton Group comprises the Beestons Formation and Celia Dolostone, and is mainly exposed around the Rum Jungle and Waterhouse domes. The *Beestons Formation* comprises quartz conglomerate, coarse-grained sandstone arkose, quartz sandstone and minor siltstone. It unconformably overlies the Rum Jungle Complex. Exposures of the unconformable contact with the Rum Jungle Complex exist 2 km north and 3.25 km north-northwest of Batchelor (**Figure 5.6**). Zircon grains from a sample of arkosic sandstone near the unconformity have yielded SHRIMP U-Pb ages in the range 3615–2500 Ma and are dominated by a single mode at ca 2536 ± 5 Ma; this is reasonably consistent with the age of the underlying granite of 2525 ± 5 Ma (data in Cross *et al* 2005). The maximum age of deposition is constrained by the weighted mean average of the youngest six grains in the population of 2506 ± 14 Ma, though the true age of deposition is probably considerably younger, based on correlation with the Namoona Group. The *Celia Dolostone* conformably overlies the Beestons Formation and comprises mainly stromatolitic dolostone and magnesite. This formation is commonly silicified at the surface. Irregular to lensoidal bodies of magnesite are present and attain economic importance in some localities. Talc and tremolite commonly accompany magnesite.

The Namoona Group includes the Masson Formation and conformably overlying Stag Creek Volcanics. The *Masson Formation* comprises a succession of laminated, black pyritic and dolomitic slate, with interbeds of rhyolitic tuff, volcaniclastic slate, metagreywacke and dolarenite. Reduced pelitic and dolomitic sedimentary rocks dominate this formation and suggest deposition in a sub-wavebase marine shelf or broad, shallow basin environment. The *Stag Creek Volcanics* contain mainly basaltic to andesitic lavas and agglomerate that are discontinuously exposed in the South Alligator Valley region. Exposures are generally poor, due to deep weathering and extensive Cenozoic cover. The mafic volcanic succession comprises amygdaloidal to massive basaltic lava, mafic agglomerate and minor interbedded volcaniclastic arenite and shale. Rare pillow structures and the presence of interbedded tuffaceous shale indicate a subaqueous depositional environment (Stuart-Smith *et al* 1984). An agglomeratic sample from 1 km north of El Sherana homestead, just south of Gunlom Falls, has yielded a xenocrystic population with an age of 2021 ± 10 Ma, interpreted as a maximum extrusion age. This provides a maximum age of deposition for the Namoona Group (Worden *et al* 2008a, b). Zircons obtained from an outcropping mafic tuff have yielded a SHRIMP U-Pb age of 2048 ±13 Ma (Page 1996a, Ferenczi and Sweet 2005). This is a maximum extrusion age, based on the interpretation that these were xenocrystic zircons, and is consistent with the results of Worden *et al* (2008a, b).

Mount Partridge Group

The lower greenschist-facies Mount Partridge Group, in the Central Trough, Batchelor Shelf and Nanambu Shelf, unconformably overlies the Manton and Namoona groups and is subdivided into a number of stratigraphic units.

Around the Rum Jungle and Waterhouse domes, the succession commences with fluvial to shallow-marine basal conglomerate, arkose, sandstone and minor siltstone of the *Crater Formation*. Detrital zircon geochronology of Hollis *et al* (2010) indicates that this was dominantly derived from ca 3125 Ma, ca 3500 Ma and ca 3670 Ma sources, which are currently unrecognised in the Northern Territory. The Crater Formation is succeeded by the *Coomalie Dolostone* (locally magnesitised), which is a succession of stromatolitic dolostone and minor shale, deposited in subtidal, evaporitic environments (**Figures 5.7**, **5.8**). These units are overlain by carbonaceous shale, dololutite, dolarenite and minor quartzite assigned to the *Whites Formation*. In the Central Trough to the east, away from the basement highs, the succession comprises the Mundogie Sandstone conformably overlain by the Wildman Siltstone. The *Mundogie Sandstone* was laid down under shallow-marine conditions and comprises arkose, subarkose, feldspathic arenite, quartzite, conglomerate, phyllite and siltstone. Recent detrital zircon U-Pb geochronology has yielded a maximum depositional age of 2284 ± 18 Ma, with the bulk of analysed grains falling in the range 2577–2472 Ma, dominant age populations at



Figure 5.6. Unconformity surface (dashed line) between poorly sorted coarse to very coarse arkose and conglomerate of Palaeoproterozoic Beestons Formation and underlying Archaean Rum Jungle Complex (52L 718270mE 8559490mN). Inherited zircons from Beestons Formation near this outcrop have been dated by SHRIMP U-Pb method and are essentially identical to zircons of Rum Jungle Complex: age spectrum ranges from ca 3615 Ma to ca 2500 Ma and is principally dominated by a single mode at 2536 ± 5 Ma.

ca 2555 and ca 2520 Ma, and possible limited metamorphic zircon growth at ca 2470 Ma (Worden *et al* 2008a, Hollis *et al* 2009b). The overlying *Wildman Siltstone* is a succession of laminated, red-brown and cream, banded silty carbonaceous phyllite, sandy ferruginous siltstone, shale and fine crenulated quartz-sericite schist, deposited under shallow-marine conditions (**Figure 5.9**). Between the Mary and McKinlay rivers, the Wildman Siltstone has been subdivided (Stuart-Smith *et al* 1986, 1987) into a lower shale-dominated member and an upper shale-sandstone member. Thin (ca 40 cm) tuffaceous sediment beds were identified in fresh exposures of the Wildman Siltstone in a railway cutting, about 8 km east of Batchelor (**Figure 5.10**). SHRIMP U-Pb zircon dating of samples from two beds from this locality have yielded depositional ages of 2019 ± 4 Ma and 2030–2020 Ma (Lally and Worden 2004, Ahmad *et al* 2006, Worden *et al* 2008a, b). This gives a stratigraphic age for the top of the Mount Partridge Group that is synchronous with the maximum age of deposition of the Stag Creek Volcanics of the underlying Namoona Group (Worden *et al* 2008a, b). The Wildman Siltstone contains interbedded mafic (*Mount Dean Volcanic Member*) and felsic volcanic rocks (*Yarrawonga Volcanic Member*). In the Rum Jungle area, a 30–300 m-thick succession of pyritic sandstone near the base of the Wildman Siltstone has been named the *Acacia Gap Quartzite Member* (**Figure 5.11**). U-Pb SHRIMP dating of detrital zircons from this unit has yielded a continuum of zircon ages in the range 2710–2495 Ma (Cross *et al* 2005). The youngest zircon in this sample gave an apparent age of 2498 ± 24 Ma and this can be used as the maximum deposition age for the Acacia Gap Quartzite Member. Alternatively, a maximum deposition age can be calculated from an arbitrarily chosen six youngest analyses. These provide an age of 2512 ± 13 Ma. The dominant detrital age peaks are similar to those from the Beestons Formation of the underlying Manton Group and indicate a provenance in a Rum Jungle Complex-type Archaean terrane. The tuff bed



Figure 5.7. Coomalie Dolostone with chertified hummocky stromatolites comprising alternating bands of carbonate minerals and siliceous talc draped over one another in mound-like fashion (52L 722300mE 8556300mN).

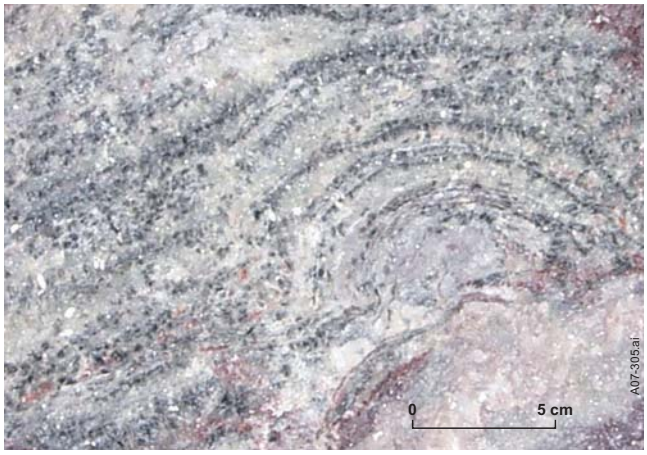


Figure 5.8. Coomalie Dolostone from Winchester magnesite deposit (52L 723532mE 8556909mN). Note domical stromatolite structure preserved by talc layering.



Figure 5.9. Thinly-bedded Wildman Siltstone in railway cutting about 8 km east of Batchelor.



Figure 5.10. Tuff bed (centre of image where note book is placed) in Wildman Siltstone in railway cutting about 8 km east of Batchelor. SHRIMP U-Pb zircon dating of this tuff yielded a depositional age of 2019 ± 4 Ma (Worden *et al* 2008a).



Figure 5.11. Tightly-folded Acacia Gap Quartzite Member, Mauries Quarry (52L 7307340mE 8586640mN).

dated from the Wildman Siltstone occurs stratigraphically above the Acacia Gap Quartzite Member and also contains some detrital grains of this age (ca 2600–2500 Ma; Worden *et al* 2008a, b), consistent with some inheritance from a similar source to that from which the Acacia Gap Quartzite Member and Mundogie Sandstone were derived.

From the area surrounding the Archaean Woolner Granite, Pietsch and Stuart-Smith (1987) described the *Koolpinyah Dolostone*, intersected mainly in drillholes. This unit is overlain by up to 80 m of Mesozoic and Cenozoic sediments and sedimentary rocks. It unconformably overlies the Archaean Woolner Granite and Dirtywater Metamorphics, and has been correlated with the Coomalie Dolostone in the Rum Jungle area. The Koolpinyah Dolostone comprises dolostone, marble, chlorite mica schist, quartz mica schist, dolomitic quartzite and sandy intraclastic limestone.

Dolostone units have also been intersected in water bores in the Humpty Doo, Palmerston, Gunn Point and Berry Springs areas. The stratigraphic position of these units is uncertain. Dolostone in the first three areas was considered to be equivalent to the Coomalie Dolostone by Verma and Qureshi (1980), Qureshi (1983) and Jolly (1985). That in the Berry Springs area was included in the South Alligator Group by Verma (1994).

The stratigraphic positions of the dolostone units in the Humpty Doo, Palmerston and Berry Springs areas are uncertain. To correlate the Humpty Doo dolostone with the Koolpinyah Dolostone is erroneous, as the latter has variable lithologies, unconformably overlies Archaean basement and is overlain by the Whites Formation. It is suggested here that the Humpty Doo, Palmerston and Berry Springs dolostones possibly represent an extension of carbonate deposition of younger basins.

Cosmo Supergroup

The Cosmo Supergroup includes the South Alligator and Finniss River groups, and is separated from the underlying Woodcutters Supergroup by a significant unconformity. Tuffaceous sedimentary rocks from the South Alligator Group have been dated at ca 1863 Ma and the underlying Wildman Siltstone (Woodcutters Supergroup) has been dated at ca 2020 Ma, indicating that the unconformity represents a time break of approximately 160 million years (Worden *et al* 2008a, b). A distinct provenance for sedimentary rocks of the Cosmo Supergroup (relative to the Woodcutters Supergroup) is indicated by both detrital zircon patterns, which are dominated by a ca 1865 Ma source (Worden *et al* 2008a, Hollis *et al* 2009b), and by REE patterns that typically show a negative Eu anomaly, consistent with a source region that has undergone plagioclase fractionation (Ahmad and McCready 2001).

South Alligator Group

The South Alligator Group is exposed from east of the Rum Jungle Complex to the South Alligator Valley. It is a succession of iron-rich sedimentary rocks, tuff, carbonate rocks, shale, greywacke and siltstone. It is divided into three formations, the Koolpin Formation, Gerowie Tuff and Mount Bonnie Formation. The thickness of the South Alligator Group increases from 1000 m in the west near Rum Jungle to as much as 5000 m in the South Alligator Valley.

The *Koolpin Formation* consists of ferruginous siltstone, banded ironstone, pyritic carbonaceous shale, ferruginous breccia and minor dolostone. An ironstone breccia up to 10 m in thickness is developed at the base and is termed the *Ella Creek Member* (Pietsch and Stuart-Smith 1987, Crick 1987). It comprises ferruginous breccia, quartzite breccia, minor ferruginous siltstone, and rare pebble and boulder conglomerate. Crick (1987) regarded it to be a Palaeoproterozoic regolith, but Pietsch and Stuart-Smith (1987) considered that at least some brecciation and ferruginisation may be due to recent surficial lateritisation and should not be taken as evidence for an ancient regolith. **Figure 5.12** shows stratigraphic sections of the Koolpin Formation from various localities in the Central Domain. Zamu Dolerite sills that intrude the Koolpin Formation in several of these sections have been excluded in order to provide a better idea of variations in the lithology and thickness of the unit across the region. The succession is informally divided into lower, middle and upper members (Nicholson 1980, Nicholson and Eupene 1984). The lower member comprises up to 100 m of carbonaceous mudstone, mudstone, siltstone and limestone, and extends from the base to the first appearance of banded ironstone. The middle member contains banded ironstone, mudstone, carbonaceous mudstone and siltstone, and may attain a thickness of up to 200 m. It also contains bands, lenses and nodules of saccharoidal quartz, ranging in size from a few to several tens of centimetres (**Figure 5.13**). The origin of these nodules has been variously explained. Walpole *et al* (1968) interpreted them to represent diagenetic replacement of carbonate-rich bands by chert, whereas Nicholson (1980) considered that the nodules represent syndepositional colloidal accretions or lumps of silica gel. The upper member comprises predominantly carbonaceous mudstone, but minor mudstone and siltstone are also present. It contains abundant pyrite and pyrrhotite, and appears as a distinct interval on aeromagnetic images, especially in the contact aureoles of granites. Minor lenses of tourmalinite are common in the lower and middle members in the Golden Dyke Dome and Cosmo Howley areas.

The *Gerowie Tuff* comprises green to grey mudstone, siliceous shale, siltstone, dark grey chert and tuff. Thin beds of iron-rich sedimentary rocks and minor quartz nodules (as described above under Koolpin Formation) are also present. Tuff constitutes about 25% of the formation and contains varying amounts of curved or angular crystal fragments of quartz, alkali feldspar and minor titanite, biotite and zircon in a matrix of devitrified glass shards and recrystallised alkali feldspar, sericite, chlorite, iron oxide and carbonates. Many of the tuffaceous intervals form 30–150 cm-thick graded beds, which have massive basal and medial parts and laminated tops with flaser bedding and soft-sediment slump structures. These graded beds are interpreted as ash fall accumulations, each from a single eruption (Stuart-Smith 1985). Many pulses of sub-aerial volcanism are indicated by the interbedding of tuff (with ripple drift in some fine-grained beds) and mudstone with chert-banded shale. It has been suggested that the source area for this sub-aerial volcanism was situated in the South Alligator

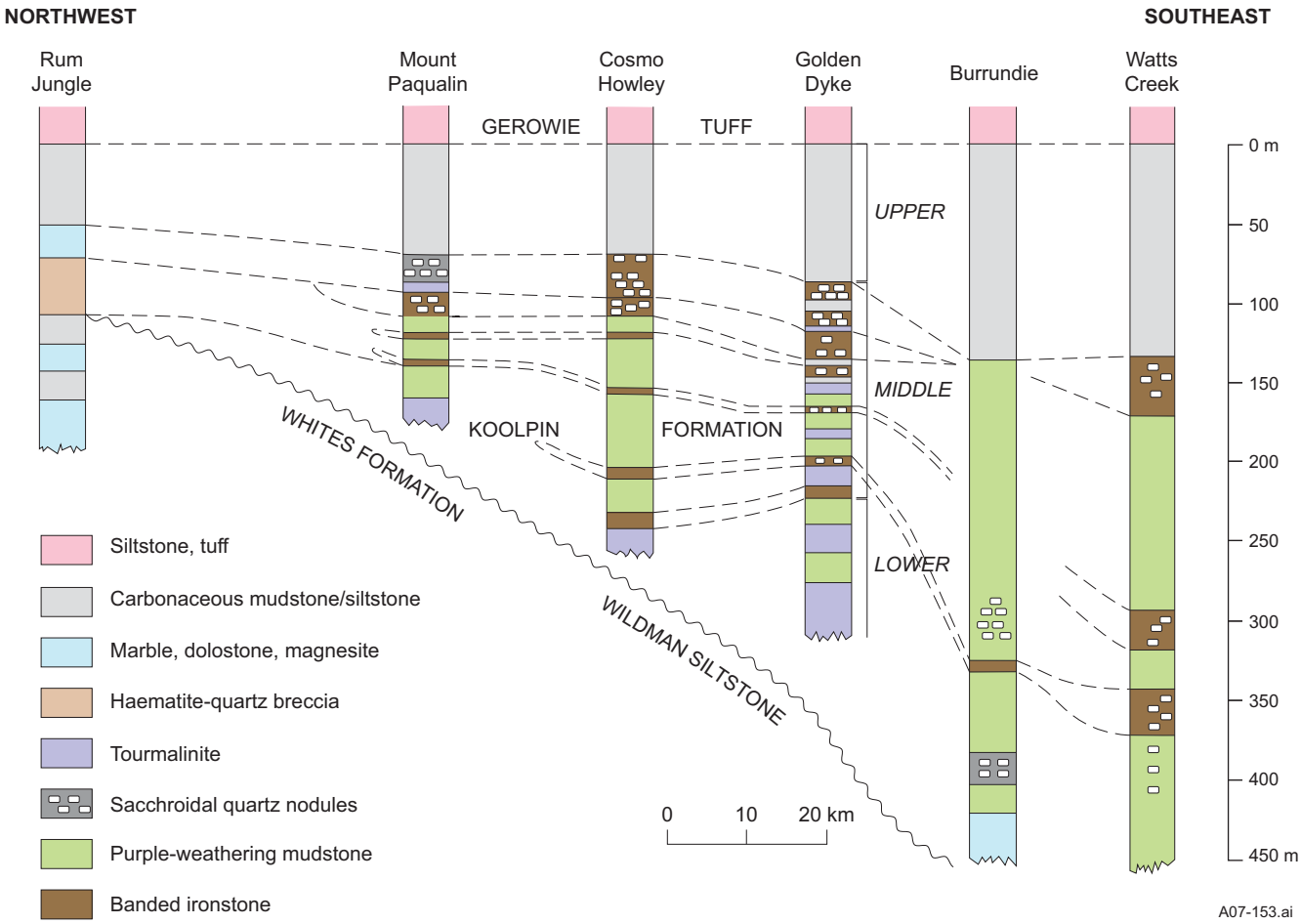


Figure 5.12. Regional section through Koolpin Formation from east of Rum Jungle to South Alligator Valley (after Ahmad *et al* 1993). The base of the Gerowie Tuff has been assumed to be at the same elevation across the region.

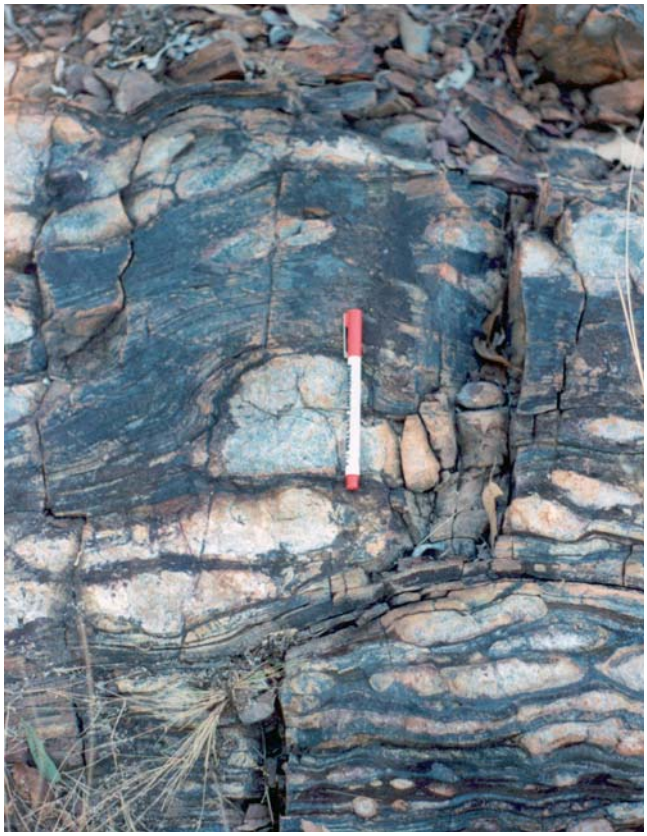


Figure 5.13. Chert nodules in thinly laminated pyritic shale of Koolpin Formation (near Golden Dyke mine, 52L 772629mE 8498361mN).

River Valley (Needham *et al* 1988). SHRIMP U-Pb zircon geochronology of two samples yielded a combined age of 1863 ± 2 Ma, interpreted as the depositional age of the Gerowie Tuff (Lally and Worden 2004, Worden *et al* 2004, 2008b). Worden *et al* (2008b) suggested that inheritance may have contributed to the older conventional U-Pb zircon age of 1885 ± 2 Ma obtained by Needham *et al* (1988) for selected zircon data from the Gerowie Tuff and Mount Bonnie Formation.

The *Mount Bonnie Formation* comprises a succession of interbedded slate, mudstone, phyllite, siltstone, feldspathic greywacke and minor tuffaceous chert, vitric crystal tuff, lithic crystal tuff, and rare banded ironstone and dolostone. It represents a transition from the low energy, shallow water environments of the Koolpin Formation and Gerowie Tuff to high energy, turbiditic environment of the overlying Finniss River Group.

Finniss River Group

The Finniss River Group represents a thick succession of sedimentary rocks of turbiditic origin. In the Central Trough, these sedimentary rocks are generally finer grained and are dominated by shale and sandstone/shale in contrast to the Batchelor Shelf, where conglomerate and sandstone are relatively more significant. The Finniss River Group in the Central Domain is subdivided into the Burrell Creek and Tollis formations. Several units of felsic volcanic rocks occur, interbedded with the Burrell Creek

Formation in the westernmost part of the Central Domain on the Batchelor Shelf.

The *Burrell Creek Formation* conformably overlies the Mount Bonnie Formation and is conformably overlain by the Tollis Formation. It consists of interbedded phyllite, slate, mudstone, feldspathic greywacke and minor pebble conglomerate. Laminated to thinly bedded, well cleaved grey or brown phyllite, slate, mudstone and quartz-mica schist are the dominant rock types. The Burrell Creek Formation is a turbidite succession that was deposited in deeper marine (shelf and shelf–slope) environments during a period of basin subsidence. Detrital zircon U-Pb geochronology on several samples from across the Central Domain and at various depths in the unit indicate that the Burrell Creek Formation is dominated by a ca 1868–1860 Ma source, which typically also defines the youngest detrital age population, thus giving a maximum depositional age (Worden *et al* 2008a, Hollis *et al* 2009b). A sample from the top of the unit, immediately below the Tollis Formation, has yielded a slightly younger maximum depositional age of 1851 ± 9 Ma, further constraining the maximum timing of deposition (Hollis *et al* 2009b).

The Mulluk Mulluk, Berinka, and Meeway volcanics, and the Warrs Volcanic Member are considered to be co-magmatic and are variably interbedded with the basal parts of the Burrell Creek Formation, where it outcrops west of the Giants Reef Fault, near the boundary of the Central Domain with the Litchfield Domain. Based on correlations of the Burrell Creek Formation with the Hermit Creek and Welltree metamorphics further west (Pietsch 1986b, Worden *et al* 2008a), the volcanics may also be coeval with these units.

The *Mulluk Mulluk Volcanics* comprise spherulitic rhyolite, rhyolitic tuff, rhyodacite and metadacite, and are inferred to conformably underlie the Burrell Creek Formation, immediately west of the Giants Reef Fault near Daly River (Dundas *et al* 1987). They typically show strong sericitic alteration. Worden *et al* (2008a) attempted to constrain the age of the Mulluk Mulluk Volcanics, but insufficient zircons were obtained from the zircon mineral separate for analysis.

The *Warrs Volcanic Member* outcrops in the Daly River region immediately west of the Giants Reef Fault. It comprises felsic volcanic flows (in some cases amygdaloidal), metadacite, tuffaceous sedimentary rocks, and volcaniclastic lithic sandstone (Dundas *et al* 1987). In the vicinity of the Giants Reef Fault, it is typically strongly altered to actinolite-, chlorite- and carbonate-bearing assemblages. The deposition or extrusion age of the Warrs Volcanic Member is constrained by a zircon U-Pb age of 1862 ± 3 Ma for a sample from an outcrop near the Anomaly 50 prospect near Daly River (Worden *et al* 2006a 2008b).

The *Berinka Volcanics* comprise rhyolite, tuff and agglomerate, spherulitic or porphyritic dacite, basic lavas, and interbedded sandstone that outcrop immediately west of the Giants Reef Fault in the western margins of the Central Domain (Dundas *et al* 1987, Edgoose *et al* 1989a, b). In the vicinity of the Giants Reef Fault, they are strongly sheared, altered and weathered. The volcanics are interbedded with conglomeratic rocks that were previously referred to as the 'Noltenius Formation' (a defunct term), a basal unit of the Burrell Creek Formation (Walpole *et al* 1968, Sweet *et al* 1974a, Sweet 1977, Edgoose *et al* 1989b). The age of the Berinka Volcanics is constrained by a zircon U-Pb age of 1861 ± 4 Ma for a sample from a locality approximately 50 km south-southwest of Daly River (Worden *et al* 2006b, 2008b).The *Meeway Volcanics* outcrop near the western margins of the Central Domain and comprise porphyritic felsic lava and tuff. Worden *et al* (2008a) attempted to constrain the age of this unit, but insufficient zircons were obtained from the zircon mineral separate for analysis.

Also mapped near the western margins of the Central Domain are the Henschke Breccia and Chilling Sandstone. Pietsch and Edgoose (1988) considered that the *Henschke Breccia*, a polymictic sedimentary breccia, represents the base of the Burrell Creek Formation. Needham and Stuart-Smith (1984) correlated the *Chilling Sandstone*, which consists of sandstone and conglomerate with minor pelitic interbeds, with the El Sherana Group in the South Alligator Valley, but it is also possible that it is a correlative of the Tollis Formation. Worden *et al* (2008a) dated detrital zircon from a sample south of Murra-Kamangee Hill and derived a conservative maximum depositional age of 1861 ± 4 Ma, consistent with the latter interpretation.

The *Tollis Formation* has a limited aerial extent. It is mapped in the Edith River and South Alligator areas, and comprises a succession of interbedded mudstone, slate, metagreywacke and minor felsic volcaniclastic shale that conformably overlies the Burrell Creek Formation. It was previously assigned to the El Sherana Group (Needham and Stuart-Smith 1985a, b, Stuart-Smith *et al* 1993), based on differences between deformation and metamorphism in the unit, as compared to the underlying Burrell Creek Formation. However, Poxon and Hein (1994) contended that the Tollis Formation is a part of the Finniss River Group, based on stratigraphic/lithologic similarities and conformity with the underlying Burrell Creek Formation. Detrital zircon U-Pb geochronology on a sample from the Katherine Gorge Road yielded a conservative maximum depositional age of 1868 ± 5 Ma (Worden *et al* 2008a). This is consistent with a re-interpretation by Kruse *et al* (1994) of data for a dacitic tuff in the Tollis Formation (from Page and Williams 1988), which excluded an inherited component to give a ca 1870 Ma maximum depositional age for the formation. Given the known age data for the Burrell Creek Formation (Worden *et al* 2008a, Hollis *et al* 2009b, see above), the true depositional age of the Tollis Formation is likely to be somewhat younger.

The *Dorothy Volcanics* comprise a ca 200 m-thick succession of basaltic lavas, pyroclastic rocks and tuffaceous sediments, ca 37 km east of Katherine. They overlie and are folded into the Burrell Creek Formation and are tentatively included within the Finniss River Group (Walpole *et al* 1968).

Mafic intrusive rocks

Hypabyssal mafic and felsic rocks, corresponding to the P4–P5 divisions of Ahmad (2000), are widespread in the Central Domain and include the Zamu Dolerite, Goodparla

Dolerite, Bludells Dolerite, Maud Dolerite, Shovel Billabong Andesite, lamprophyre and other dolerite dykes.

The *Zamu Dolerite* represents a suite of continental tholeiites with a composite thickness of about 1.5 km. It was intruded mainly as sills, which were folded and metamorphosed with the enclosing succession. These sills are most abundant in the South Alligator Group, but are also present in other groups. A sample of the Zamu Dolerite, obtained from the Fisher Landing Ground in the South Alligator Valley, has yielded a SHRIMP U-Pb zircon date of 1870 ± 6 Ma (Page 1996b). Since the age of the South Alligator Group, which the Zamu Dolerite intrudes, is now constrained to ca 1863 Ma (Worden *et al* 2008a), the SHRIMP age of the Zamu Dolerite is considered to have been complicated by inheritance. The main lithology is massive, medium- to coarse-grained, grey quartz dolerite, in which primary textures and some primary minerals are usually preserved. A weak metamorphic foliation is developed in some thin sills. The dolerite has an ophitic texture, consisting of plagioclase, clinopyroxene and interstitial intergrowths of quartz and K-feldspar, anhedral quartz, minor euhedral K-feldspar, magnetite, biotite and rare apatite. Fractionated phases of the dolerite include hornblende-bearing granophyre, granophyric quartz dolerite and metadiorite (Ferenczi and Sweet 2005).

The *Goodparla Dolerite* intrudes the Masson Formation in northwestern MOUNT EVELYN[1], in the Barramundi Dome area and between Old Goodparla ruins and Goodparla. Several magnetic amphibolite units within the Cahill Formation are included within this dolerite unit (Ferenczi and Sweet 2005). It comprises dark grey, medium-grained olivine metadolerite with an ophitic texture and is distinguished from the Zamu Dolerite by its darker colour, absence of quartz grains and magnetic properties.

The *Shovel Billabong Andesite* is a up to 300 m-thick concordant body within the Gerowie Tuff in the South Alligator Valley. It contains mainly sodic plagioclase and minor augite phenocrysts in a holocrystalline groundmass of the same composition. The intimate stratigraphic and spatial association of the Gerowie Tuff with the Shovel Billabong Andesite and a concordant dacite intrusive rock suggests a close proximity to a Palaeoproterozoic eruptive centre (Needham *et al* 1988).

The *Bludells Dolerite* (previously Bludells Monzonite) occurs as irregularly shaped bodies up to 5 km across within the McCarthys Granite, Frances Creek Leucogranite and Allamber Springs Granite. It also occurs as dykes in the latter (Stuart-Smith *et al* 1993). The dolerite comprises clinopyroxene, orthopyroxene, olivine, biotite, opaques, plagioclase and minor quartz and K-feldspar. Pyroxenes are typically altered to actinolite. It is distinguished from more mafic phases of the Cullen Supersuite by high Ni (165 ppm) and Cr (up to 2781 ppm), and much lower Ba and Sr. Late-stage Cullen Supersuite pegmatite and aplite veins, cutting both the dolerite and enclosing granites at contacts, indicate that the Bludells Dolerite is probably coeval with the Cullen Supersuite granites. As the Dolerite is usually enclosed by the Frances Creek Leucogranite, the latter may represent a localised melt formed by intrusion of the dolerite into a granitic intrusion (Stuart-Smith *et al* 1993).

[1] Names of 1:250 000 and 1:100 000 mapsheets are in large and small capital letters, respectively, eg MOUNT EVELYN, FOG BAY.

The *Maud Dolerite* is exposed in a small area about 15 km east of Katherine. It is a massive, medium- to coarse-grained to gabbroic, dark-grey quartz dolerite (Kruse *et al* 1994). The dolerite intrudes the Tollis Formation, largely as sills, in a number of places and is overlain unconformably by the Plum Tree Creek Volcanics. The Yeuralba Granite has contact metamorphosed the Tollis Formation and Maud Dolerite in this area (Kruse *et al* 1994). It is possible that the Maud Dolerite is of similar age to the Zamu Dolerite.

Lamprophyre dykes, typically minette, occur within faults and other pre-existing structures at a number of localities in the Central Domain. They are not visible on the surface, but are seen at mine workings within the Brocks Creek, Goodall, Mount Paqualin, Mount Todd, Rustlers Roost, Toms Gully, Woodcutters and Yam Creek deposits. These dykes postdate deformation and metamorphism in the Central Domain. At Brocks Creek, lamprophyres preferentially occupy an anticlinal hinge zone, but are not folded (Miller *et al* 1998). At Goodall, there is evidence that lamprophyres were emplaced after deformation, but prior to gold mineralisation (Quick 1994). The Toms Gully lamprophyres intrude both the Mount Bundey Granite (1831 ± 6 Ma) and the older Mount Goyder Syenite, but are co-genetically and temporally related to the granitoids (Sheppard 1992, Sheppard and Taylor 1992). At the Woodcutters deposit, they cross-cut the bedding and contain late-stage mineralisation (Smolongov 1988).

Geolsec Formation

The Geolsec Formation is a haematite-quartz breccia from the Rum Jungle region that unconformably overlies the Coomalie Dolostone and Whites Formation, and is unconformably overlain by the Depot Creek Sandstone of the Tolmer Group (Birrindudu Basin). It is considered to be younger than the Finniss River Group, based on the observation that its basal unconformity is not affected by folding preserved in the Finniss River Group (Ahmad *et al* 2006) and probably corresponds to either the P4 or P5 division of Ahmad (2000). The unit is 40–200 m thick and comprises haematitic quartzite breccia, milky quartz and chert breccia, haematitic mudstone, siltstone and sandstone, minor phosphatic siltstone and breccia. The suggested modes of origin include palaeokarst collapse (Crick 1987), reworked talus slope deposit (Paterson *et al* 1984, Beardsmore 1983) and fault scarp breccia deposited in a shallow-marine environment (Ahmad *et al* 2006).

El Sherana and Edith River groups

Widespread post- to late orogenic granites, coeval felsic volcanic rocks, sandstone and minor shale between the orogenic strata of the PCO and the McArthur Basin were originally incorporated in the "Transitional Domains" of Plumb *et al* (1980). This volcani-sedimentary succession is divided into the El Sherana and unconformably overlying Edith River groups.

El Sherana Group

The El Sherana Group outcrops in the South Alligator Valley and the Edith River area. It is a >2.5 km-thick

succession of mainly interbedded felsic and mafic volcanic rocks, sandstone, conglomerate and sedimentary breccia, unconformably overlying the South Alligator Group. Its depositional age is constrained by SHRIMP U-Pb zircon dates of 1829 ± 5 Ma for a rhyolite porphyry within the Pul Pul Rhyolite (Jagodzinski 1999) and 1825 ± 4 Ma for the overlying Plum Tree Creek Volcanics of the Edith River Group (Page 1996c).

The basal *Scinto Breccia* is a pink-brown siliceous and commonly phosphatic rock, containing poorly sorted, angular pebble-sized clasts of white, polycrystalline quartz and rare phyllite, metagreywacke and altered rhyolite, in a matrix and secondary cement of haematite, quartz and varying amounts of apatite. Stuart-Smith *et al* (1988) suggested that the breccia is an *in situ* regolith that developed over carbonate-bearing rocks (mainly Koolpin Formation) at the unconformity surface. This interpretation does not account for the presence of the breccia over non-carbonate rocks at several localities (eg metasiltstone at Saddle Ridge, rhyolitic ignimbrite southwest of Sleisbeck), the variable lithological composition of framework clasts (eg at Coronation Hill) and the unusual metal chemistry. A more probable origin is that the breccia denotes local talus deposits (Walpole *et al* 1968), situated within and adjacent to synsedimentary faults that have acted as conduits for various post-depositional hydrothermal fluids (Valenta 1991, Freidmann and Grotzinger 1991).

The *Coronation Sandstone* is a valley-fill succession of conglomerate, sandstone, volcaniclastic siltstone and shale, mafic volcanic rocks and minor felsic volcanic rocks. This unit unconformably overlies deformed and metamorphosed rocks of the Finniss River Group, or is interbedded with the Scinto Breccia. A SHRIMP U-Pb zircon age of 1827 ± 4 Ma was obtained for a rhyolitic quartz-feldspar porphyry that intrudes some mafic volcanics (Jagodzinski 1999).

The *Pul Pul Rhyolite* represents a succession of amalgamated felsic ignimbrites and volcaniclastic rocks with partially intrusive rhyolite porphyry and rhyolite lava. It unconformably overlies the Coronation Sandstone and various basement units. The top of the succession is interbedded with and disconformably overlain by polymictic volcaniclastic conglomerate, sandstone and minor siltstone lenses. SHRIMP U-Pb zircon geochronology has yielded ages of 1827 ± 4 Ma and 1828 ± 4 Ma for rhyolite porphyries, and 1829 ± 5 Ma for a lithic-poor ignimbrite (Jagodzinski 1999). Interpretation of SHRIMP U-Pb age data obtained from a lithic-rich ignimbrite (sample 89123138) by Jagodzinski (1999) suggests the presence of two zircon populations that have ages of 1862 ± 16 Ma (zircon xenocryts) and 1810 ± 27 Ma (syn-crystallisation zircon).

The *Big Sunday Formation* represents a valley fill-type succession of interbedded lithic to arkosic sandstone, volcaniclastic siltstone, amygdaloidal basalt, rhyolite and rhyolitic ignimbrite. A local disconformity is present at the contact with the underlying Pul Pul Rhyolite. An angular unconformity is apparent between the Big Sunday Formation and the overlying Kurrundie Sandstone.

Edith River Group

The Edith River Group overlies the El Sherana Group with only minor angular discordance and is represented by the Kurrundie Sandstone, Phillips Creek Sandstone, Hindrance Creek Sandstone and Plum Tree Creek Volcanics. The *Kurrundie Sandstone* forms the base of the Edith River Group and consists predominantly of thickly bedded to massive, purple to brown, coarse to pebbly lithic quartz sandstone interbedded with conglomerate and minor siltstone. The *Phillips Creek Sandstone* is unconformably overlain by the Plum Tree Creek Volcanics and rests with a marked angular unconformity on the Tollis Formation. It consists of interbeded, labile and tuffaceous, poorly sorted sandstone, arkose, conglomerate and minor siltstone. The *Hindrance Creek Sandstone* rests unconformably on the Tollis Formation and is either conformable beneath or is faulted against the Plum Tree Creek Volcanics. It comprises medium- to coarse-grained, poorly sorted, pebbly quartz sandstone and arkose. The *Plum Tree Creek Volcanics* conformably overlies the three sandstone units mentioned above and comprises massive rhyodacitic ignimbrite, massive glassy pink rhyolite, flow-banded ignimbrite and minor crystal tuff. Minor discontinuous clastic interbeds include fine to coarse quartz sandstone, volcaniclastic pebble conglomerate, micaceous sandy siltstone and tuffaceous siltstone. The succession includes interbedded mafic amygdaloidal basalt and andesite. A SHRIMP U-Pb zircon age of 1822 ± 6 Ma has been obtained from an ignimbrite within the Plum Tree Creek Volcanics (Kruse *et al* 1994).

Summary of basin development in the Central Domain

The initiation of sedimentation in the Central Domain is considered to be the result of rifting of Neoarchaean basement at ca 2020 Ma. The Central Trough was formed as a result of a northwest-trending structure and another depocentre was probably developed in the Eastern Trough (**Figure 5.4**). The form of these initial structures may be half grabens, as suggested by Needham *et al* (1988). These authors divided the overall succession into a rift phase (Manton, Namoona and Mount Partridge groups), sag phase (South Alligator Group) and pre-orogenic phase (Finniss River Group). Geochronological studies on tuffaceous and volcanic units in the Mount Partridge and South Alligator groups (Lally and Worden 2004, Worden *et al* 2008a, b) suggests a hiatus in sedimentation of ca 160 Ma between these two groups. Therefore a single rift, sag, pre-orogenic cycle is unlikely and two major depositional cycles are more probable. The Woodcutters and the Cosmo supergroups represent these older and younger cycles, respectively.

Figure 5.14 is modified from Needham and Stuart Smith (1984) and provides a palaeogeographical reconstruction of the Palaeoproterozoic successions of the Central Domain, prior to ca 1860–1840 Ma deformation and metamorphism. No detailed sedimentological studies on the Palaeoproterozoic strata have been undertaken and the following discussion provides a preliminary account of the sedimentary environments.

Along the basin shelf, Manton Group sedimentation was mainly fluvial to shallow marine, with the source of the detritus being Archaean basement highs, as is indicated by Neoarchaean inherited zircons in the Beestons Formation and Acacia Gap Quartzite Member. Towards the centre of the basin, the Masson Formation was deposited in a deeper-water low-energy environment that was dominated by carbonaceous shale with interbeds of calcareous sandstone and carbonate rock. Volcanic activity, manifested by the Stag Creek Volcanics, was apparently confined to the Central Trough. The Rum Jungle Complex formed palaeohighs surrounded by algal reef colonies that developed in a supratidal to evaporitic environment (Bone 1985). The Nanambu Complex and Woolner Granite/Dirtywater Metamorphics also probably formed palaeohighs that provided a similar palaeogeographic setting. It is possible that most of the Batchelor Shelf was emergent and that deposition never took place in the western PCO at this time.

The overlying Mount Partridge Group succession commenced with fluvial to shallow-marine sandstones that are widespread throughout the PCO and followed a period of source rejuvenation and non deposition (Needham and Stuart-Smith 1984). The sediments are considered to have been derived from the north. Coarse marine sandstone alternated with shale and volcanic interbeds. These were succeeded by a thick succession of shale deposited in a low-energy, possibly subtidal environment. In the Rum Jungle area, sedimentation extended westward and comprised a thick succession of fluvial sandstone overlain by subtidal to supratidal stromatolitic carbonate rock. This is followed by a thick succession of pyritic carbonaceous mud interbedded with dololutite that was deposited in a low-energy environment. A major depositional break of ca 160 Ma was followed by the deposition of the South Alligator Group.

Unlike the underlying deposits, there are no basal sandstones in the South Alligator Group succession, which begins with a basal breccia and is overlain by pyritic carbonaceous shale, interbedded carbonate rocks and banded ironstone, chert beds and tuffaceous intervals.

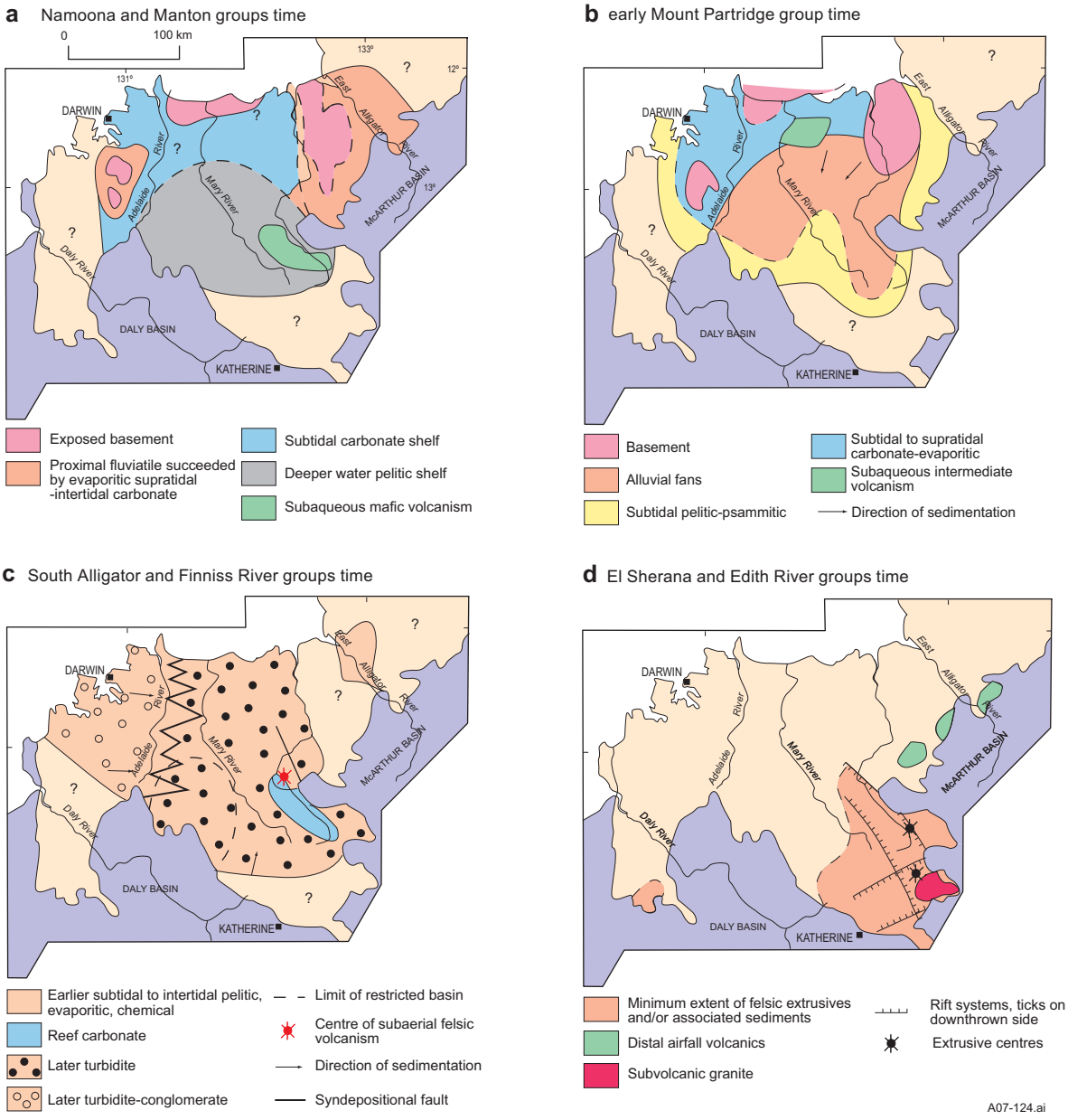


Figure 5.14. Palaeofacies reconstruction of four Palaeoproterozoic time slices in PCO (modified from Needham and Stuart-Smith 1984).

These sediments represent low-energy environments and slow sedimentation.

The upper part of the succession (Finniss River Group) is represented by deeper-water, high-energy mass-flow deposits. Clast compositions indicate a source area of high relief, little weathering and generally finer grain size (Needham *et al* 1988). This succession is dominated by monotonous turbidites and heralds the onset of orogenic activity (see below). The sedimentary rocks are coarser to the west, indicating a source in that direction.

The El Sherana and Edith River groups are envisaged to have been deposited in fault-bounded shallow grabens, within which a valley and ridge topography was developed in the South Alligator Valley area (Needham *et al* 1988, Needham and Stuart-Smith 1985). These grabens were filled with coarse fluviatile sediments derived from the adjoining highlands. Extrusion of ignimbritic rocks and rhyolitic flows (Pul Pul Rhyolite) subsequently filled the grabens and extended into adjoining areas beyond the bounding faults. The Pul Pul Rhyolite volcanic succession is interpreted to have accumulated in a small continental basin ('Jawoyn Sub-basin') under subaerial conditions (Needham and Stuart-Smith 1985a). The thick ignimbrite pile and abundant co-magmatic intrusions south of the Malone Creek Granite are thought to represent the preserved remnant of an ancient caldera structure (Jagodzinski 1999). Volcaniclastic turbiditic sediments and interbedded volcanics of the Big Sunday Formation subsequently spread over the surrounding uplands. The crystal-rich volcanic sandstone, feldspathic greywacke and volcaniclastic siltstone of this formation were probably deposited in a lacustrine environment (Freidmann and Grotzinger 1994).

Further uplift and erosion were followed by renewed subsidence and deposition of the Edith River Group. The initial sediments were mainly sandstones, possibly deposited as coalescing fans flanking marginal fault scarps (Needham *et al* 1988). The succession of facies represents a juvenile fluvial system with basal valley-fill deposits, followed by medial braidplain (pebble conglomerate and trough-bedded sandstone), through to distal braidplain (siltstone and well bedded fine sandstone) deposits. Palaeocurrent measurements from exhumed trough axes indicated a flow direction to the northwest (Freidmann and Grotzinger 1994). This was followed by the extrusion of the Plum Tree Creek Volcanics. These volcanics were subaerial lavas and ignimbrites that were intercalated with minor fluvial sediments. The position of the compositionally similar Grace Creek Granite probably coincides with the effusive centre of Plum Tree Creek rhyodacitic volcanism (Needham *et al* 1988).

Granitoids

On the basis of geochemical characteristics, Wyborn *et al* (1997) subdivided the PCO granites into the Nimbuwah Suite (Nimbuwah Complex, Nimbuwah Domain); Cullen Supersuite and Jim Jim Suite (Central Domain); and Allia Creek and Wagait suites (Litchfield Domain). Salient petrological and geochemical characteristics of these granitoids are given in **Table 5.3**. A summary of the characteristics of individual plutons is provided in **Table 5.4** and locations of various granite suites are given in **Figure 5.15**.

Cullen Supersuite

Cullen Supersuite granite plutons were emplaced into the Central Domain in the period ca 1835–1820 Ma (Wyborn *et al* 1997). This is believed to be synchronous with a major extensional event that led to the deposition of the El Sherana and Edith River groups and probably also the lower Katherine River Group. The granites form an I-type (granodiorite) suite that has undergone significant fractionation. They have large contact aureoles and are spatially associated with gold deposits. The granites are predominantly calc-alkaline and the presence of magnetite as an important accessory, together with magnetic susceptibility measurements (Bajwah 1994), indicates that they belong to the 'Magnetite Series' of Ishihara (1977). Petrological and geochemical data provide evidence that crystal fractionation played a dominant role during the crystallisation history.

On the basis of petrology, geochemistry and associated mineral deposits, Stuart-Smith *et al* (1993) and Wyborn *et al* (1997) subdivided the Cullen Supersuite into three groups, each containing a number of suites:

1. *Leucogranite-dominated plutons:*
   a. Saunders Suite: Foelsche Leucogranite, Mount Davis and Saunders granites
   b. Burnside Suite: Burnside, Wandie and Douglas granites, Frances Creek Leucogranite
   c. Tennysons Suite: Tennysons, Yenberrie, Umbrawarra leucogranites, Wolfram Hill and Fenton granites
2. *Granite-dominated plutons:*
   a. Fingerpost Suite (hornblende dominated): Fingerpost Granodiorite, Minglo and McCarthys granites
   b. McMinns Bluff Suite (biotite dominated): McMinns Bluff Granite
   c. Margaret Suite (hornblende-biotite granites): Margaret, Prices Springs, Mount Porter and McKinlay granites
   d. Mount Bundey Suite: Mount Bundey Granite and Mount Goyder Syenite
3. *Concentrically zoned transitional granites and leucogranite-dominated plutons:*
   a. Shoobridge Suite (zoned hornblende dominated): Shoobridge, Driffield, Bonrook and Tabletop granites
   b. Allamber Springs Suite (zoned biotite dominated): Allamber Springs Granite

Palaeoproterozoic country rocks have been contact-metamorphosed in the vicinity of the Cullen Supersuite granitoids. Albite-epidote hornfels is present in all contact aureoles, commonly with a narrow, inner continuous zone of hornblende-hornfels (Stuart-Smith 1985). The effects of contact metamorphism are seen up to 10 km away from the exposed granite contacts (Stuart-Smith 1985), but an emphasis on lateral distance may be misleading, as geophysical data (Tucker *et al* 1980) suggest that granitic rocks are present virtually everywhere in the PCO at a depth of 1–4 km. However, the geophysical data cannot

Suite and geochronology	Characteristics and component plutons
NIMBUWAH COMPLEX 1866 ± 8 Ma[1], 1864 ± 4 Ma[2], 1860 ± 4 Ma[2], 1866 ± 3 Ma[2]	Dominantly restite suite with little or no fractionation. **Geochemistry**: SiO_2: 58–76 wt%, no evidence of silicification. K_2O/Na_2O: some evidence of sodic alteration. Th/U**:** U loss during regional metamorphism. $Fe_2O_3/(FeO+Fe_2O_3)$: relatively reduced. Rb: weak increase with increasing SiO_2. Th, U, Y, Rb/Sr: no increase with increasing SiO_2. P_2O_5: decrease with increasing SiO_2. K/Rb: scattered. Rb-Ba-Sr: plot in anomalous granite field. Sr: moderately high and decreases with increasing SiO_2. Ba: scattered. F: weak decrease with increasing SiO_2. **Mineralogy**: Biotite, hornblende, plagioclase, K-feldspar, quartz; accessory apatite, zircon and allanite; rare fluorite, magnetite and titanite. In migmatised area in Beatrice Inlier, orthopyroxene, clinopyroxene and garnet have been noted.
ALLIA CREEK SUITE **Murra-Kamangee Granodiorite**, 1862 ± 2 Ma[2], 1854 ± 4 Ma[2], 1853 ± 3 Ma[3] **Jammine Granite** 1858 ± 5 Ma[2] **Mount Litchfield Granite**, 1768 ± 16 Ma[4] **Two Sisters Granite** 1862 ± 5 Ma[2] **Allia Creek Granite** 1806 ± 7 Ma[2] **Fish Billabong Adamellite** 1850 ± 3 Ma[2]	S-type suite containing peraluminous minerals, such as andalusite, cordierite and muscovite. **Geochemistry**: SiO_2: 57–76 wt%. Th/U: most ratios are within normal range. $Fe_2O_3/(FeO+Fe_2O_3)$: some more fractionated samples are more oxidised, reflecting haematite, which has been recorded as being common in late-stage phases. Rb: increases with increasing SiO_2 in Two Sisters, Jammine and Soldiers Creek granites. U: late increase in some plutons with increasing SiO_2. P_2O_5, Th, Y: no increase with increasing SiO_2. K/Rb: decreases with increasing SiO_2 in all plutons with exception of Murra-Kamangee Granodiorite and Allia Creek Granite. Rb-Ba-Sr: some samples plot in strongly differentiated field. Sr: high and variable. Rb/Sr: increases with SiO_2 in Two Sisters, Jammine and Soldiers Creek granites. Ba: decreases with increasing SiO_2. **Mineralogy**: Characteristically S-type, with andalusite and cordierite being recorded. Biotite and muscovite are intergrown. Some sulfides have been recorded, attesting to reduced nature of parts of this suite, whereas some late veins contain haematite, suggesting that the suite became more oxidised with fractionation. **COMPONENT PLUTONS: Murra-Kamangee Granodiorite**, **Mount Litchfield Granite, Two Sisters Granite, Allia Creek Granite, Jammine Granite, Fish Billabong Adamellite, Soldiers Creek Granite**
WAGAIT SUITE **Wagait Granite** 1863 ± 3 Ma[2] **Koolendong Granite** 1860 +2/-5 Ma[2]	Restite-rich, with some late fractionation of felsic phase. Mainly medium-grained granodiorite, monzogranite and granite. Suite is probably an extension of Paperbark Suite of Halls Creek Orogen. **Geochemistry:** Little geochemical data. **Mineralogy:** Dominant minerals are quartz, K-feldspar, plagioclase, biotite and hornblende, with accessory epidote, apatite and zircon. Peppimenarti Granite is more felsic and is dominated by monzogranite to granite. Some aplite and pegmatite phases suggest late-stage evolution of magmatic phase. Reynolds River Granite is a small stock of grey to pink, medium- to coarse-grained granite to monzogranite. No mineralisation is associated with this Suite, suggesting that its potential is low. **COMPONENT PLUTONS: Wagait Granite, Peppimenarti Granite, Reynolds River Granite, Koolendong Granite**
CULLEN SUPERSUITE **Margaret Granite** 1860 ± 25 Ma[1] **Fingerpost Granodiorite** 1843 ± 22 Ma[1], 1823 ± 3 Ma[2] **McMinns Bluff Granite** 1835 ± 6 Ma[2], 1818 ± 3 Ma[1] **Mount Bundey Granite** 1831 ± 6 Ma[1] **Allamber Springs Granite** 1826 ± 18 Ma[1], 1826 ± 32 Ma[1], 1822 ± 6 Ma[2], 1818 ± 5 Ma[2] **Umbrawarra Leucogranite** 1825 ± 7 Ma[2], 1805 ± 12 Ma[1] **Prices Springs Granite** 1804 ± 50 Ma[1] **Burnside Granite** 1800 ± 5 Ma[2], 1750 ± 56 Ma[1]	I-type (granodiorite) suite with significant fractionation. Clear evidence of late-stage release of magmatic fluids. Significant thermal metamorphism and large contact aureole. **Geochemistry**: SiO_2: 54–79 wt%. Some silicification. K_2O/Na_2O: remarkably coherent, considering degree of alteration observed in thin sections. Th/U: plot in normal range. $Fe_2O_3/(FeO+Fe_2O_3)$: relatively reduced. P_2O_5, U, Y, Rb, Rb/S, F: increase with increasing SiO_2. Th: weak increase with increasing SiO_2. K/Rb, Sr, Ba: decrease with increasing SiO_2. **Mineralogy**: Dominant minerals are quartz, plagioclase, K-feldspar, biotite and hornblende. Plagioclase grains are almost universally altered to sericite, and biotite is commonly altered to chlorite. **COMPONENT PLUTONS: Saunders Granite, Foelsche Leucogranite, Burnside Granite, Douglas Granite, Frances Creek Leucogranite, Wandie Granite, Tennysons Leucogranite, Yenberrie Leucogranite, Wolfram Hill Granite, Fenton Granite, Umbrawarra Leucogranite, Fingerpost Granodiorite, McCarthys Granite, Minglo Granite, McMinns Bluff Granite, Margaret Granite, Prices Springs Granite, Mount Porter Granite, McKinlay Granite, Shoobridge Granite, Tabletop Granite, Driffield Granite, Mount Davis Granite, Allamber Springs Granite, Bonrook Granite, Mount Bundey Granite**. **Mount Goyder Syenite**
JIM JIM SUITE **Grace Creek Granite** 1863 ± 5 Ma[1] **Plum Tree Creek Volcanics** 1858 ± 25 Ma[1], 1825 ± 4 Ma[2], 1833 ± 43 Ma[1] **Ironbark Member** 1857 ± 61 Ma[1] **Jim Jim Granite** 1838 ± 7 Ma[2] **Gimbat Formation** 1827 ± 5 Ma[2], 1827 ± 4 Ma[2] **Malone Creek Granite** 1823 ± 9 Ma[2], 1808 ± 20 Ma[2], 1736 ± 5 Ma[2] **Tin Camp Granite** 1846 ± 13 Ma[2] **Nabarlek Granite** 1818 ± 8 Ma[2]	Dominated by felsic end-members with significant fractionation. Compared to other PCO felsic igneous suites, Jim Jim suite is predominantly oxidised. **Geochemistry**: SiO_2: 50–79 wt%, most samples are in 63–79 wt% range. K_2O/Na_2O: dominant sericite alteration is reflected in samples with low to moderate K_2O and very low values of Na_2O. Some samples have been albitised. Th/U: ratios have been only weakly affected. $Fe_2O_3/(FeO+Fe_2O_3)$: some samples have been strongly oxidised. F, Rb/Sr, Th, Y, U, Rb: increase rapidly and exponentially to very high values with increasing SiO_2. Ba, K/Rb, P_2O_5: decrease with increasing SiO_2. Rb-Ba-Sr: most samples plot in strongly differentiated field. Sr: Sr values are very low. **Mineralogy**: Biotite is dominant ferromagnesian mineral, whereas hornblende is recorded only in more mafic plutons. Fluorite is commonly recorded. **COMPONENT PLUTONS: Nabarlek Granite, Tin Camp Granite, Jim Jim Granite, Malone Creek Granite, Grace Creek Granite, Eva Valley Granite and Yeuralba Granite.**

Table 5.3. Summary of PCO granite suites. Geochronology superscripts: 1 = conventional U-Pb zircon (TIMS), 2 = SHRIMP U-Pb zircon, 3 = U-Pb xenotime, 4 = Rb-Sr.

Name	Texture	Mineralogy	Petrology / geochenmistry
NIMBUWAH COMPLEX	Migmatitic, foliated, porphyritic coarse granodiorite.	Bt, Hbl, Pl, Kfs, Qtz, Ap, Zrn, Aln, Fl	Granodiorite, tonalite, quartz monzonite, A-type. Restite-dominated suite with little or no fractionation.
ALLIA CREEK SUITE			
Jammine Granite	Medium-grained tourmaline-muscovite leucogranite	Qtz, Mc, Ms, Tur	Leucogranite; high SiO_2, K_2O.
Two Sisters Granite	Medium-grained granite with coarse garnet-bearing pegmatitic variety.	Qtz, Kfs, Pl, Or, Bt, Ms, Grt, Tur, Ap, Ttn, Py, Po, Zrn.	Quartz monzonite to granite; pegmatitic variety is low in TiO_2, FeO, MnO, Sr, Zr, Ba and La, but is high in SiO_2, Na_2O, Al_2O_3, Rb, Bt, Li and Sn.
Mount Litchfield Granite	Fine to medium foliated to non-foliated granite.	Qtz, Kfs, Pl, Mc, Bt, Ms, Tur, Ap, Zrn,	Quartz monzonite to granodiorite. Large SiO_2 and Rb range, low K_2O and Sr, high CaO, MgO, Rb, Ba and Zr. S-type.
Murra-Kamangee Granodiorite	Medium-grained variable foliated granodiorite.	Qtz, Kfs, Pl, Mc, Bt, Grt, Ap, Ttn, Ilm, Mag, Ms, And, Sil, Crd, Zrn	Tonalite to granodiorite, quartz monzonite, large SiO_2 range, low K_2O and Na_2O. S-type.
Fish Billabong Adamellite	Medium to coarse, equigranular, partly foliated quartz monzonite.	Qtz, Kfs, Pl, Mc, Bt, Adl	Quartz monzonite.
Allia Creek Granite	Coarse porphyritic granite.	Qtz, Mc, Pl, Bt, Ms, Ap, Zrn	Granodiorite with large microcline phenocrysts.
Soldiers Creek Granite	Medium-grained porphyritic granite.	Qtz, Kfs, Mc, Pl, Bt, Ms	Quartz monzonite and granodiorite.
CULLEN SUPERSUITE ***Leucogranite-dominated plutons***			
Saunders Suite			
Saunders Granite	Fine- to medium-grained, equigranular leucogranite.	Qtz, Mc, Or, Pl, Bt, Hbl, Ap, Mag, Py, Ttn, Aln, Zrn	Leucogranite; large SiO_2 range, high FeO, TiO_2, MgO, CaO, Sr and Ba. Low Rb.
Foelsche Leucogranite	Fine- to medium-grained leucogranite, granophyric intergrowths of Qtz and Kfs.	Qtz, Kfs, Bt, Py	Leucogranite; low Li, Rb, U and Y.
Mount Davis Granite	Coarse-grained porphyritic leucogranite.	Qtz, Kfs, Pl, Bt, Hbl, Aln	Leucogranite; high TiO_2, MgO, Sr, and Ba, low Rb, U.
Burnside Suite			
Burnside Granite	Fine- to medium-grained granite, porphyritic to equigranular.	Qtz, Kfs, Pl, Ms, Bt, Aln, Mag, Ap, Zrn, Fl	Leucogranite; high SiO_2, Al_2O_3, Na_2O and Rb. Low MgO, Sr.
Douglas Granite	Fine- to medium-grained porphyritic granite.	Qtz, Kfs, Pl, Bt, Ap, Mag, Zrn, Mnz	Leucogranite; large SiO_2 range. Low Al_2O_3, Th, Zr, Nb, Ga, La and Ce.
Wandie Granite	Medium-grained equigranular granite.	Qtz, Kfs, Bt, Ms	Leucogranite; high SiO_2, Na_2O. Low MgO, Sr, Nb.
Frances Creek Leucogranite		Qtz, Pl, Mc, Bt, Ms, Aln, Ap	Leucogranite; high SiO_2, Al_2O_3. Low MgO, Nb, U.
Tennysons Suite			
Tennysons Leucogranite	Coarse-grained porphyritic leucogranite with interspersed phase of medium to fine leucogranite.	Qtz, Kfs, Pl, Bt, Ms, Ap	Leucogranite; high TiO_2, Fe, K_2O, P_2O_5, V, Ni, Rb, U, Y and Sn. Low CaO, Na_2O, Ba, Sr, Th, Zr, Nb, La and Ce.
Yenberrie Leucogranite	Fine-grained leucogranite and greisen.	Qtz., Kfs, Pl, Bt, Ms, Ap	Similar to fine-grained phases of Tennysons Leucogranite.
Wolfram Hill Granite	Coarse-grained porphyritic quartz monzonite.	Qtz, Pl, Kfs, Bt, Fl	Granite and quartz monzonite; very depleted in CaO (0.56%) and enriched in Al_2O_3 (17.26%).
Fenton Granite	Medium to coarse grained commonly porphyritic and partly foliated leucogranite and quartz monzonite.	Qtz, Kfs, Bt, Pl, Hbl, Mag, Aln, Ttn, Ap, Grt, Zrn	Leucogranite and quartz monzonite; low TiO_2, MgO, P_2O_5, Th and high Na_2O, Sr, Pb, Ba, Zr, La and Ce; more fractionated parts show a trend towards high Rb, U and Y.
Umbrawarra Leucogranite	Coarse grained porphyritic leucogranite.	Qtz, Pl, Kfs, Bt, Ms, Hbl, Mag, Ap, Ttn, Tur	Leucogranite , low CaO, Na_2O, Ba, Sr, Pb, Th, Zr, Nb, La, Ce, V, Cr and high K_2O, P_2O_5, As and Sn, Rb, U, Y and Li.

Table 5.4. Summary of some significant plutons within PCO granite suites. Abbreviations: Adl = Adularia, Aln = Allanite, And = Andalusite, Ap = Apatite, Bt = Biotite, Crd = Cordierite, Fl = Fluorite, Grt = Garnet, Hbl = Hornblende, Ilm = Ilmenite, Kfs = K-Feldspar, Mc = Microcline, Ms = Muscovite, Mag = Magnetite, Mnz = Monazite, Or = Orthoclase, Plg = Plagioclase, Po = Pyrrhotite, Py = Pyrite, Qtz = Quartz, Sil = Sillimanite, Tou = Tourmaline, Ttn = Titanite, Zrn = Zircon (**table continued on next page**).

Name	Texture	Mineralogy	Petrology / geochenmistry
Concentrically zoned transitional granites and leucogranite-dominated plutons ***Hornblende-dominated suites***			
Shoebridge Suite			
Shoobridge Granite	Medium-grained, zoned from granodiorite through quartz monzonite to leucocratic quartz monzonite.	Qtz, Kfs, Pl, Bt, Hbl, Mag, Ap, Zrn, Aln.	Leucogranite, granodiorite and quartz monzonite; high Al_2O_3, Na_2O, K_2O, P_2O_5, Ba, Sr, Pb, Th, U and Nb. Low in CaO and Fe.
Tabletop Granite	Coarse-grained porphyritic granite and leucogranite, fine equigranular granite.	Qtz, Kfs, Pl, Bt, Hbl, Aln, Ttn, Zrn, Mag, Fl	Granite and leucogranite; high CaO, MgO and Sr and low TiO_2, K_2O. Pb, Th, Zr, Nb, Y, La, Ce, V, Ga and F.
Mount Davis Granite	Equigranular and porphyritic coarse-grained leucogranite.	Qtz, Kfs, Pl, Bt, Hbl, Aln, Ap	Leucogranite; low TiO_2, Al_2O_3, K_2O and high Fe, CaO, Na_2O, V, Sn, Cu.
Driffield Granite	Marginal phase of fine- to medium-grained granite. Coarse-grained core with Kfs megacrysts.	Qtz, Kfs, Pl, Bt, Ms, Zrn, Ap, Py	Granite; high MgO, CaO and V; low Sr, Pb, Th, U, Zr, Nb, La and Ce. Rb, U, and Li increase with increasing SiO_2.
Biotite-dominated suites			
Allamber Springs Suite			
Allamber Springs Granite	Coarse-grained equigranular leucogranite with porphyritic granite at margins.	Qtz, Pl, Kfs, Bt, Hbl, Aln, Ttn, Mag, Zrn, Ap, Fl	Leucogranite and granite; low TiO_2, Al_2O_3, MgO, and CaO and high Na_2O, Sr, Th, Zr, La, Ce and V.
Bonrook Granite	Coarse-grained porphyritic granite and fine equigranular granite.	Qtz, Kfs, Pl, Bt, Ms, Aln, Zrn, Mag	Leucogranite and granite; low Al_2O_3, Ba, Th, Zr, La and Ce and high Rb, U and Th.
Granite-dominated plutons ***Hornblende-dominated plutons***			
Fingerpost Suite			
Fingerpost Granodiorite	Coarse-grained porphyritic granodiorite grading into porphyritic granite with up to 40% Kfs megacrysts.	Qtz, Kfs, Pl, Hbl, Bt, Ap, Mag, Zrn, Aln	Granodiorite and granite; high MgO, CaO, Sr, V, Cr, Ni, Cu and Sn and low TiO_2, Al_2O_3, K_2O, P_2O_5, Li, Rb, Zr, Nb, Y, La, Ce, Zn and Ga. One of the least fractionated plutons of Cullen Batholith.
McCarthys Granite	Coarse-grained porphyritic to equigranular granite gradational to leucogranite.	Qtz, Kfs, Pl, Hbl, Bt, Mag, Ap, Fl, Ttn, Grt, Ms	Granite and leucogranite; low TiO_2, K_2O, Li, Rb, U, Y and Nb. High MgO, CaO, Na_2O, Ba, Sr, Zr, V, Cr and Ni.
Minglo Granite	Coarse-grained porphyritic granite, monzonite and leucogranite.	Qtz, Kfs, Pl, Hbl, Bt, Mag, Ttn, Aln, Ap, Zrn	Granite, monzonite and leucogranite; mafic samples have high Al_2O_3, CaO, Na_2O, K_2O, P_2O_5, Sr, Pb, U, Th, Zr, Nb, La, Ce and V. Low TiO_2 and MgO.
Biotite-dominated plutons			
McMinns Suite			
McMinns Bluff Granite	Coarse- to fine-grained porphyritic granite.	Qtz, Kfs, Pl, Bt, Hbl, Mag, Ttn, Ap, Aln, Zrn	Granite; western lobe has high TiO_2, MgO, Na_2O, K_2O, P_2O_5, Rb, Sr, Ba, Nb, Zr, Th and U; eastern lobe has high TiO_2, Fe, K_2O, P_2O_5, Pb, Zr, Nb, Y, La, Ce, Zn, Ga, F and is low in MgO, CaO, Na_2O, Th, U, V and Ni.
Hornblende – biotite equal			
Margaret Suite			
Margaret Granite	Medium- to coarse-grained porphyritic to equigranular quartz monzonite.	Qtz, Kfs, Pl, Bt, Hbl, Aln, Mag, Zrn, Ap.	Quartz monzonite.
Prices Springs Granite	Medium- to coarse-grained porphyritic granite.	Qtz, Kfs, Pl, Bt, Hbl, Ap, Mag.	Granite; high TiO_2, Al_2O_3, K_2O, P_2O_5, Sr, Pb, Th, Zr, Nb and Bt; low CaO, MgO, Na_2O.
Mount Porter Granite	Coarse-grained porphyritic granite.	Qtz, Kfs, Pl, Ms, Bt, Hbl	Granite
McKinlay Granite	Medium- to coarse-grained mostly equigranular granodiorite.	Qtz, Pl, Kfs, Bt, Hbl, Mag, Ttn, Ap.	Granite and granodiorite; high TiO_2, Al_2O_3, K_2O, P_2O_5.
Mount Bundey Suite			
Mount Bundey Granite	Massive medium-grained granite.	Qtz, Pl, Hbl, Bt, Ap, Aln	I-type granite; high Th, U, Ni, Cr.
Mount Goyder Syenite	Cumulate-textured coarse-grained syenite.	Kfs, Hbl, Cpx, Qtz, Bt, Ttn, Ap, Mag, Zrn	Shoshonitic syenite; high alkalis, LILE, LREE, Th, and U, high but variable Cr and Ni.

Table 5.4. Summary of some significant plutons within granite suites of the Pine Creek Orogen (**continued from previous page** and **continued on next page**).

Name	Texture	Mineralogy	Petrology / geochenmistry
JIM JIM SUITE			
Jim Jim Granite	Coarse-grained equigranular hornblende-biotite granite and coarse-grained equigranular to porphyritic biotite granite.	Qtz, Kfs Pl, Bt, Hbl, Aln, Ap, Zrn	Granite, leucogranite and granophyre.
Malone Creek Granite	Porphyritic to equigranular medium-grained granite.	Qtz, Kfs, Pl, Bt, Aln, Ttn, Fl, Zrn	Alkali feldspar granite.
Grace Creek Granite	Fine- to coarse-grained equigranular to porphyritic alkali feldspar granite and greisens.	Qtz, Kfs, Pl, Bt, Hbl, Mag, Zrn	Microgranite and granophyre.
Eva Valley Granite	Porphyritic biotite leucogranite.	Qtz, Kfs, Pl, Bt, Aln, Ttn, Zrn, Fl	Leucogranite.
Yeuralba Granite	Coarse-grained equigranular leucocratic leucogranite.	Qtz, Kfs, Pl, Bt, Ms, Tur	Leucogranite.
Nabarlek Granite	Coarse-grained equigranular granite.	Qtz, Pl, Kfs, Bt, Aln, Zrn, Fl	Granite; very high in U content (up to 25 ppm).
Tin Camp Granite	Biotite granite, trondhjemite, porphyritic granite.	Pl, Kfs, Qtz, Bt, Zrn	Granite.

Table 5.4. Summary of some significant plutons within granite suites of the Pine Creek Orogen (**continued from previous page**).

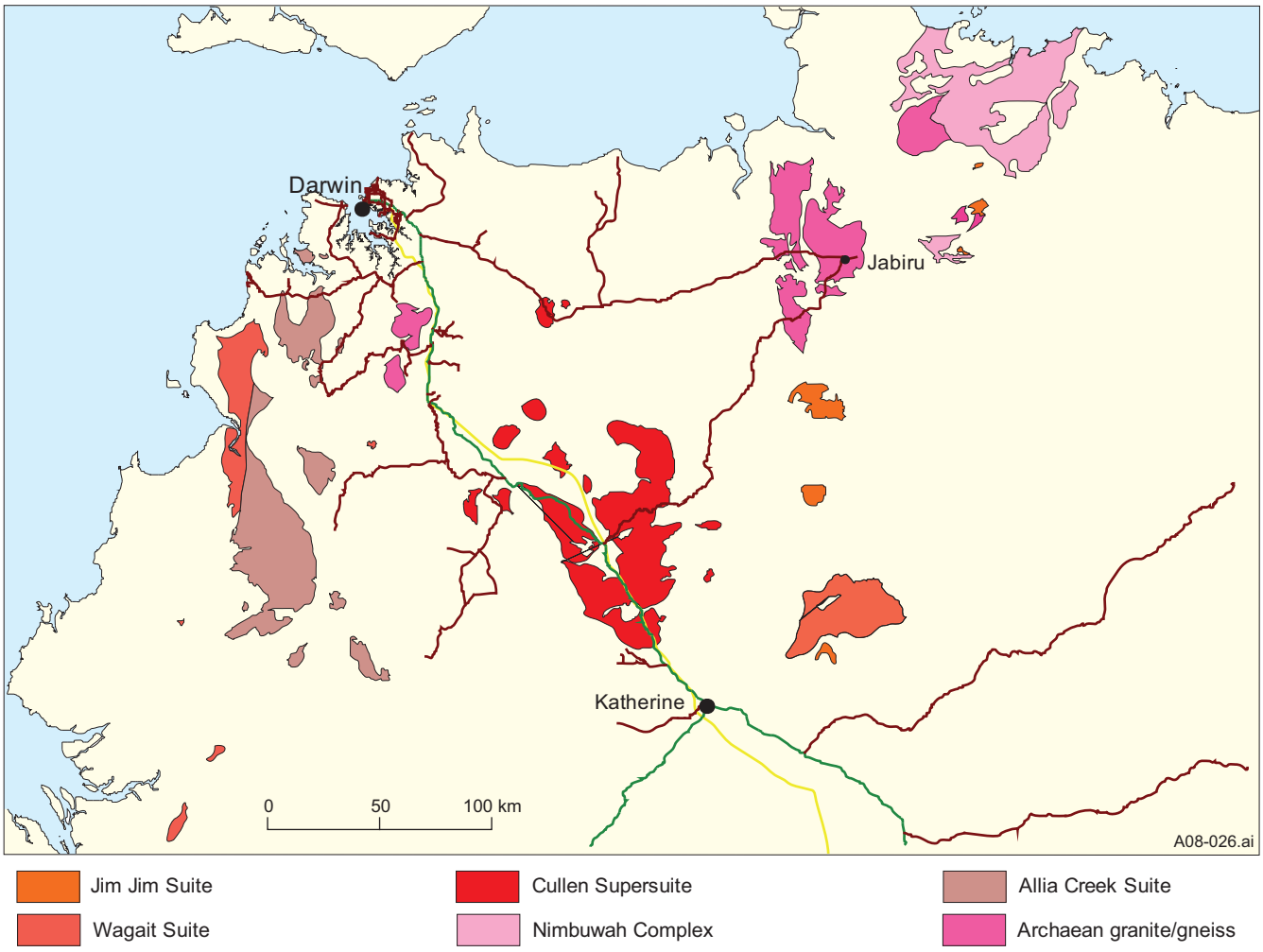


Figure 5.15. Spatial distribution of various PCO and Archaean granite suites. Units in the Litchfield Domain are largely undercover.

readily distinguish between Archaean basement and Palaeoproterozoic granitoids.

Deformation associated with the emplacement of the granitoids is not obvious in outcrops because of their large dimensions. The folds are open, have low amplitudes, trend east, may be associated with poorly developed crenulation cleavages, and are responsible for the formation of elongated basins and domes such as the Golden Dyke Dome, which is the result of F_4 interference with an F_3 fold (Stuart-Smith 1985, Stuart-Smith *et al* 1993). The *Maud Creek Event* (Needham *et al* 1988) could be regarded as a localised early phase of the Cullen Event. It is represented by folding in the El Sherana Group. These folds are typically open to tight and north- to northeast-trending.

The early crystallisation history of granitoids that are spatially related to gold and base metals mineralisation,

(Allamber Springs, Burnside and Tabletop granites) is dominated by the crystallisation of anhydrous minerals. Water saturation in the residual melt, causing sericitisation and chloritisation of primary mineral phases, is indicated by the late appearance of biotite and hornblende. The perthitic nature of K-feldspar in these granitoids was probably also caused by a reaction with late hydrothermal fluids (Finch *et al* 1990).

A number of attempts have been made to obtain radiometric ages of the different phases of the Cullen Supersuite. A detailed sampling program for U-Pb zircon, conventional and SHRIMP geochronological studies was carried out by RW Page in the early 1990s. The results were summarised in Stuart-Smith *et al* (1993) and Wyborn *et al* (1997), and details can be obtained from the Geoscience Australia OZCHRON database. U-Pb zircon SHRIMP ages range from 1835 Ma (McMinns Bluff Granite) to 1800 Ma (Burnside Granite) and are more precise than either conventional U-Pb zircon ages (1860 to 1750 Ma), or earlier Rb-Sr whole rock or K-Ar ages (ca 1780 Ma; Hurley *et al* 1961, Walpole *et al* 1968, Steiger and Jäger 1977, Riley 1980).

The brief petrological account of the various granite suites given below is largely derived from Ahmad *et al* (1993), Stuart-Smith *et al* (1993), Bajwah (1994) and Wyborn *et al* (1997), where more in-depth descriptions can be found.

Leucogranite-dominated plutons

These plutons occur mainly as circular to elliptical intrusions. They are generally pink to grey and fine to medium grained, and are usually porphyritic with marginal equigranular phases. Microcline, quartz, albite, plagioclase are the main minerals, with minor biotite, hornblende and muscovite. Megacrysts comprise perthitic microcline, albite, quartz and minor biotite. Accessory minerals are apatite, zircon and rare titanite, allanite, monazite, magnetite and fluorite. Mafic minerals are generally less than 5%. Some plutons have patches of greisens and veins of quartz, pegmatite and microgranite. A predominantly fine-grained, porphyritic phase is usually confined to the margins, and a medium- to coarse-grained predominantly equigranular phase occupies the central parts. Many of these plutons are spatially associated with Sn, W, Mo and Cu mineralisation.

Leucogranite-dominated plutons are characterised by >70 wt% SiO_2. The Saunders Suite shows little variation in K/Rb values, or Rb, Li, U or Y contents with increasing SiO_2. These granites have an ASI <1.1 and $Fe_2O_3/(FeO + Fe_2O_3)$ values >0.24. The Burnside Suite shows increasing Rb, Li, U, and Y, and decreasing K/Rb values with increasing SiO_2. The $Fe_2O_3/(FeO + Fe_2O_3)$ values range from 0.3 to 0.1 and the ASI is <1.1. The Tennysons Suite shows increasing Rb, U, Y and decreasing K/Rb values with increasing SiO_2. The ASI is >1.1 and $Fe_2O_3/(FeO + Fe_2O_3)$ values are <0.24.

Granite-dominated plutons

These plutons are generally irregular in shape, but some circular plutons are also present. A majority are pink and green in colour, coarse-grained and strongly porphyritic. Some phases have up to 50% K-feldspar megacrysts which are generally <6 cm in size. K-feldspar (microcline, orthoclase), albite/oligoclase, quartz, biotite, hornblende and rare muscovite are the major constituents. Accessory minerals include zircon, apatite, rare allanite, titanite and fluorite. Mafic minerals are in the range 5–15%. The more mafic plutons can be subdivided, depending on the relative proportions of biotite and hornblende, and the wt% at which hornblende disappears with increasing SiO_2. These plutons are spatially associated with some Cu and Au occurrences.

Hornblende-dominated plutons of the Fingerpost Suite are high in CaO, Sr and V, and low in TiO_2. K/Rb values are high and show little variation with SiO_2 contents, and ASI values are mostly <1.1. The *Fingerpost Granodiorite* is one of the least fractionated plutons and is high in MgO, CaO, Sr, Cr, V, Ni, Cu and Sn, and low in TiO_2, Al_2O_3, K_2O, P_2O_5, Li, Rb, Zr, Y, La, Ce, Zn and Ga. The *McCarthys Granite* shows little evidence of fractionation and has geochemical characteristics similar to the Fingerpost Granodiorite. The *Minglo Granite* has fractionated phases similar to the Burnside Granite.

Biotite-dominated plutons of the McMinns Bluff Suite are typified by the *McMinns Bluff Granite*. This pluton forms two chemically and petrographically distinct bodies, distinguished as eastern and western lobes, that are separated by younger sedimentary rocks. The western part contains hornblende in samples with up to 72 wt% SiO_2, whereas in the eastern part, hornblende is only present in samples with <64% SiO_2 (Stuart-Smith *et al* 1993). The eastern part has high TiO_2, total Fe, K_2O, P_2O_5, Pb, Zr, Nb, Y, La, Ce, Zn, Ga and F, and low MgO, CaO, Na_2O, Sr, Th, U, V and Ni. The western part of the McMinns Bluff Granite has high TiO_2, MgO, Na_2O, K_2O, P_2O_5, Rb, Sr, Ba, Nb, Zr, Th and U.

Mixed biotite-hornblende plutons of the Margaret Suite have characteristics intermediate between the hornblende- and biotite-dominated types. There are only a few analyses available (Wyborn *et al* 2001). They have high TiO_2, Al_2O_3, K_2O, P_2O_5 and low MgO and CaO.

The Mount Bundey igneous suite (informal name) comprises the *Mount Bundey Granite*, the *Mount Goyder Syenite* and a swarm of potassic *lamprophyre* dykes, together comprising the Mount Bundey pluton. The pluton covers an area of 80 km^2 and is dominated by the 1830 ± 6 Ma Mount Bundey Granite (Sheppard 1992, Page pers comm 2000 in Rasmussen *et al* 2001), which intrudes the syenite. Both granitoids are intruded by the lamprophyre dykes. This suite was not included in the classification scheme of Wyborn *et al* (1997), but is assigned here to the granite-dominated plutons of that scheme. The Mount Bundey Granite has a chemistry consistent with I-type calc-alkaline granites and has high abundances of Th, U, Ni and Cr. The Mount Goyder Syenite is shoshonitic, with high alkalis, LILE, LREE, Th, and U, and high but variable Cr and Ni. All of these elements are also abundant in the lamprophyre dykes, which are also shoshonitic. The Mount Goyder Syenite and the lamprophyre dykes are compositionally continuous, but there is a silica gap between the syenite and the granite. The Mount Goyder Syenite is thought to be a hybrid pluton, formed by interaction between the granite and lamprophyre magmas (Sheppard 1995).

Concentrically zoned transitional granites and leucogranite-dominated plutons

These plutons have a wide range of SiO_2 contents, with mafic end-members showing similarities to either biotite- or hornblende-dominated granite suites. The more felsic members show similarities to the leucogranite suites. Most varieties are coarse grained, porphyritic, and pink and green in colour. The concentric zones are arranged from oldest at the margins to youngest at the centre. In the *Tabletop Granite*, Stuart-Smith (1985) identified six major phases that were simplified by Bajwah (1994) into three: (i) a fine- to medium-grained porphyritic facies generally occupying the margins of the pluton and grading into (ii) medium-grained equigranular rocks with rare areas of porphyritic varieties, and (iii) a coarse-grained equigranular facies constituting the inner core of the pluton.

Most geochemical data is for the Tabletop Granite, which is high in CaO, MgO, and Sr, and low in TiO_2, K_2O, Pb, Th, Zr, Nb, Y, La, Ce, V, Ga and F. Among the biotite dominated suites, the *Allamber Springs Granite* has low TiO_2, Al_2O_3, MgO and CaO, and high Na_2O, Sr, Th, Zr, La, Ce and V contents, a low $Fe_2O_3/(FeO+Fe_2O_3)$ ratio and an ASI of 1.1. Some leucogranites of this suite resemble the Tennysons Leucogranite and show an increase in Rb, U, and Y, and a decrease in K/Rb values with increasing SiO_2.

Jim Jim Suite

The Jim Jim Suite includes plutons in the South Alligator Valley and Nabarlek area. It has also been referred to as the David Suite (Ferenczi and Sweet 2005) to avoid confusion with the Jim Jim Granite that forms part of the suite, though this revised name has not been formally recognised. The Nabarlek and Tin Camp granites of this suite are discussed in **Nimbuwah Domain**. Although apparently coeval with the Cullen Supersuite, the Jim Jim Suite has some distinctive characteristics. It contains granites associated with mainly subaerial volcanic and sedimentary rocks. These volcanic and sedimentary rocks generally infill rift valleys and have been included in the El Sherana and Edith River groups, discussed above. The granites were intruded at shallow levels. Some plutons grade into volcanic rocks and others have faulted contacts that are filled with quartz breccia (Wyborn *et al* 1997). The Jim Jim Suite is restite dominated; however, after restite-separation, the remaining fluid has fractionated significantly to produce greisenisation and late-stage alteration. The Jim Jim Suite has not been studied in as much detail as the Cullen Supersuite and there is only limited mineralogical and petrological data on the granites. The following description is mainly from Wyborn *et al* (1997).

The *Jim Jim Granite* comprises granite, leucogranite, porphyritic granite and granophyric granite. The main minerals are biotite, hornblende, plagioclase, K-feldspar, quartz, allanite, apatite and zircon. The *Grace Creek Granite* comprises porphyritic microgranite, medium-grained granite and porphyritic granophyre. The main constituent minerals are hornblende, biotite, K-feldspar, plagioclase, quartz, magnetite and zircon. The *Malone Creek Granite* consists of porphyritic alkali-feldspar granite, even-grained granite. The main constituent minerals include quartz, K-feldspar, plagioclase, biotite, allanite, titanite, fluorite and zircon. The *Eva Valley Granite* comprises biotite leucogranite and porphyritic biotite leucogranite with the main mineral constituents being quartz, K-feldspar, plagioclase, biotite, allanite, titanite, fluorite and zircon. The *Yeuralba Granite* is a coarse equigranular leucogranite. The main mineral constituents are quartz, K-feldspar, plagioclase, biotite, muscovite and tourmaline.

NIMBUWAH DOMAIN

The Nimbuwah Domain (Needham *et al* 1988) comprises the eastern part of the PCO, east of the Partridge Thrust/Jim Jim Fault (approximately coincident with, to slightly west of the South Alligator River). Its western boundary is defined by the westernmost extent of amphibolite-facies metamorphism of the Palaeoproterozoic succession, the Cahill Formation and Nourlangie Schist, which distinguishes this domain from the lower-grade greenschist-facies Central Domain. The easternmost extent of the Nimbuwah Domain is unknown, as it is concealed beneath Palaeo- to Mesoproterozoic cover of the McArthur Basin. Palaeoproterozoic strata in the Domain overlie 2670 Ma, 2640 Ma and 2530–2500 Ma Neoarchaean basement exposed as the Nanambu Complex west of Jabiru, and as the Njibinjibinj, Arrarra and Kukalak gneisses in western Arnhem Land (Hollis *et al* 2009a, b, see **Archaean**). Contacts are rarely exposed and are strongly tectonised, so that their original nature is unclear. The Palaeoproterozoic strata comprise the Kakadu Group, Cahill Formation and Nourlangie Schist. The formerly mapped Myra Falls Metamorphics (Needham 1988a) is now known to comprise components of Neoarchaean basement gneisses and Cahill Formation, and hence, this term has been abandoned (Hollis *et al* 2009a). Both the Palaeoproterozoic strata and the structurally underlying Neoarchaean basement are strongly ductily deformed and are metamorphosed typically to middle amphibolite facies. The heat source for the metamorphism is attributed to the emplacement of late deformational granitic, granodioritic and subordinate tonalitic gneiss and granite of the ca 1867–1862 Ma Nimbuwah Complex, which intruded as a series of east–west-elongated lobes to the east of approximately 133.20°E. Localised migmatisation of Palaeoproterozoic metasedimentary rocks has occurred immediately adjacent to plutons of the Nimbuwah Complex. The Tin Camp Granite, a late undeformed granite, was emplaced by 1846 Ma, giving a minimum constraint on the cessation of ductile deformation. Correlation between the Palaeoproterozoic succession of the Central Domain and that of the Nimbuwah Domain is hampered by the relatively high metamorphic grade and ductile deformation of the latter, combined with poor exposure, particularly in the region northeast of Oenpelli to the coast.

Kakadu Group

The Kakadu Group forms the oldest Palaeoproterozoic strata in the Nimbuwah Domain (Nanambu Shelf and Eastern Trough). Stratigraphic correlation of this succession across the PCO is tentative. Needham (1988a) correlated the Kakadu Group and lower Cahill Formation

with the Namoona Group. However recent detrital zircon geochronology data show that ca 1900 Ma erosive sources contributed to the formation of the Cahill Formation and Nourlangie Schist (Hollis *et al* 2009a). Therefore, the stratigraphic succession of the Nimbuwah Domain is distinct from that of the Central Domain. Because of higher degrees of deformation, scanty outcrops and metamorphic grade (amphibolite facies) the interrelationships between the individual units in the East Alligator River region is tentative.

The Kakadu Group comprises mainly fluvial arkose and sandstone, and includes the Mount Basedow Gneiss, Munmarlary Quartzite, Mount Howship Gneiss and Kudjumarndi Quartzite (**Table 5.2**). These units are exposed on the flanks of the Nanambu Complex (Nanambu Shelf) and in and around the Myra Falls Inlier in western Arnhem Land (Eastern Trough). The *Mount Basedow Gneiss* comprises pink-grey, quartzofeldspathic metamorphic rocks including meta-arkose, metaconglomerate, quartzite, gneiss and minor schist. The rocks are strongly foliated parallel to compositional layering (**Figure 5.16**). The meta-arkose and metaconglomerate are radioactive and locally have up to 0.03% U_3O_8 (Livingstone 1958, Needham 1988a). Cross-bedding is also preserved in places. The *Munmarlary Quartzite* is essentially similar to the quartzitic units within the Mount Basedow Gneiss and comprises muscovite and feldspathic quartzite. It is exposed as scattered outcrops on the western flank of the Nanambu Complex. The *Mount Howship Gneiss* has been reported from exposures of limited areal extent in the Myra Falls Inlier. It comprises granular quartzo-feldspathic gneiss, exposed in the core of an antiform, where it is overlain by the Kudjumarndi Quartzite. Based on recent work that has identified Neoarchaean basement structurally below the Kudjumarndi Quartzite (Hollis *et al* 2009a, b), it is possible that this is in fact part of the Neoarchaean basement. The *Kudjumarndi Quartzite* forms ridges in the Myra Falls Inlier and consists of orthoquartzite ranging from monomineralic to micaceous, hornblende-rich and feldspathic types. Recent U-Pb detrital zircon geochronology shows strong similarities in provenance between the Kakadu Group and the Manton and Mount Partridge groups, with dominant age peaks at ca 2673–2572 Ma and ca 2530–2483 Ma for the Kudjumarndi Quartzite and 2518 Ma for the Mount Basedow Gneiss (Worden *et al* 2008a).



Figure 5.16. Mount Basedow Gneiss: coarse quartzofeldspathic gneiss with dark, coarse to very coarse biotite-muscovite bands (near Mount Basedow, 53L 250430mE 8561760mN).

Cahill Formation and Nourlangie Schist

The lower of two 'members' of the *Cahill Formation* (referred to herein as *lower Cahill Formation*) represents a succession of carbonaceous schist, marble, calc-silicate rock, para-amphibolite, dolostone, magnesite, schist and quartzite. The lower Cahill Formation hosts uranium mineralisation at Ranger, Jabiluka, Koongarra and a number of other uranium deposits in the Alligator River region. It has been subdivided into a number of sub-units in the mine areas. Away from the mine areas, the lower Cahill Formation is distinguished generally by lower Fe/Mg ratios and higher values of Mg, F and Th.

The *upper Cahill Formation* has a conformable transitional contact with the lower Cahill Formation and comprises interlayered feldspathic quartz schist, feldspathic schist, feldspathic quartzite, conglomerate, and minor mica schist and quartzo-feldspathic gneiss. With an increase in the proportions of mica schist, it has a gradational contact with the overlying *Nourlangie Schist*, which is locally garnet- and/or staurolite-bearing. At its type locality at the northern scarp of Mount Brockman, a succession of crenulated quartz-muscovite schist is unconformably overlain by the Mamadawerre Sandstone of the Kombolgie Subgroup (**Figure 5.17**). Garnet-, staurolite- and kyanite-bearing assemblages in the Cahill Formation and Nourlangie Schist indicate amphibolite-facies metamorphic conditions and yield quantitative P-T estimates in the range ca 500–600°C and ca 8–10 kbar (Ferguson *et al* 1980, Hollis *et al* 2009b). The former Myra Falls Metamorphics (Needham 1988a), exposed in the Myra Falls Inlier, are now recognised as a composite unit of quartzofeldspathic gneiss, quartz-mica schist, mica schist, garnet-amphibolite schist and quartzite, which includes components of Neoarchaean felsic basement, Cahill Formation and Nourlangie Schist (Hollis *et al* 2009b).

The upper Cahill Formation and parts of the Nourlangie Schist were formerly correlated with the Mount Partridge Group in the lower greenschist facies terranes of the Central Trough and Batchelor Shelf (Needham 1988a, Ahmad *et al* 1993). However, U-Pb detrital zircon

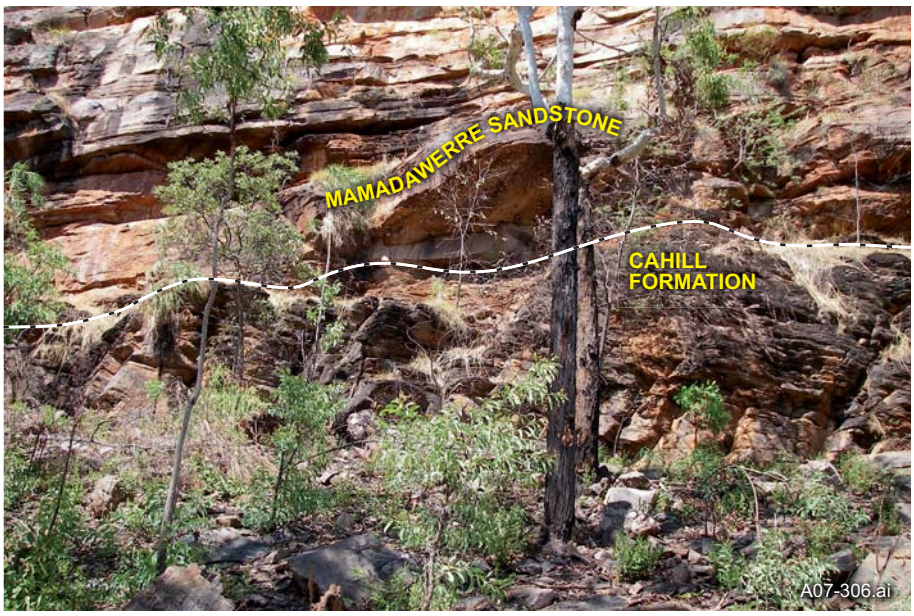


Figure 5.17. Unconformity surface (dashed line) between sandstone and conglomerate of Mamadawerre Sandstone and underlying schist of lower Cahill Formation (53L 265500mE, 8579700mN).

geochronology indicates a different provenance and age for the Cahill Formation and Nourlangie Schist, compared with the Mount Partridge Group, and these are therefore distinct successions. The Cahill Formation and Nourlangie Schist are dominated by detrital age peaks at ca 2500 Ma and an additional significant spread in detrital ages from ca 2500–1900 Ma (Hollis *et al* 2009b). This indicates deposition at ≥1870 Ma, based on the timing of migmatisation in pelitic schist, which is thought to be associated with intrusion of the Nimbuwah Complex (Carson *et al* 2010). This is prior to deposition of the Finniss River Group in the Central Trough at ≤1860 Ma, based on maximum depositional ages (Worden *et al* 2008a, Hollis *et al* 2009a).

Granitoids

Granitoids in the Nimbuwah Domain are represented by the syn-orogenic Nimbuwah Complex and late-orogenic Nabarlek and Tin Camp granites. The late-orogenic granites have been assigned to the Jim Jim Suite of the Central Domain, on the basis of geochemical and petrographic characteristics (Wyborn *et al* 1997).

The *Nimbuwah Complex* (**Table 5.3**, **Figure 5.15**) comprises syn-orogenic granodioritic plutons that form a series of east–west-elongated lobes, exposed in western Arnhem Land (Needham 1988a). Their sub-surface extent is recognisable in regional magnetic data. The Nimbuwah Complex (Nimbuwah Suite of Wyborn *et al* 1997) comprises granitic, gneissic and migmatitic components, all thought to have been emplaced during, or in the late stages of regional ductile deformation. This occurred at 1868–1860 Ma, based on SHRIMP U-Pb zircon geochronology (Page *et al* 1980, Wyborn *et al* 1997, Worden *et al* 2008a, Hollis *et al* 2009b, Carson *et al* 2010). These plutons were emplaced into Archaean felsic gneiss basement and the overlying Palaeoproterozoic succession. This is thought to have resulted in localised migmatisation immediately adjacent to plutons, constrained by a SHRIMP U-Pb zircon date of 1871 Ma (Hollis *et al* 2009a, Carson *et al* 2010). It may also have been responsible for and/or coeval with regional amphibolite-facies fabric development in the Nimbuwah Domain.

The Nimbuwah Complex is dominantly a restite suite, comprising hornblende-biotite granodioritic rocks that show little or no fractionation. Components include granodioritic migmatite and tonalite, with lesser monzogranite (Ferguson *et al* 1980, Needham 1988a, Wyborn *et al* 1997, Glass *et al* 2009). Leucocratic components comprise granitic to granodioritic migmatite and granitoids, containing quartz, feldspar, biotite, hornblende, titanite, zircon and apatite ± allanite. Migmatitic varieties commonly have hornblende-bearing leucosomes that cross-cut, but are also deformed within the regional gneissic fabric, indicative of migmatisation during ductile deformation. The essential minerals in the melanocratic variety are plagioclase (andesine/labradorite), quartz, hornblende, biotite, and potash feldspar ± hypersthene ± clinopyroxene ± almandine. Interpretation of geochemical data (Ferguson *et al* 1980, Wyborn *et al* 1997) suggests that the Nimbuwah Complex shows depletion in compatible elements, including U, and has low $^{87}Sr/^{86}Sr$ and $Fe_2O_3/(FeO + Fe_2O_3)$ ratios, indicating that these have not been affected by oxidised fluids.

The *Nabarlek Granite* is pink and coarse-grained with the main minerals being plagioclase, quartz, orthoclase, biotite, rare opaques, and accessory allanite, apatite, zircon and fluorite (Needham 1988a). It was emplaced after the main deformation phase into the Cahill Formation (formerly mapped as Myra Falls Metamorphics) and is cut by numerous quartz breccia-filled fault zones. The granite is extensively altered, with zones of overprinting haematite-chlorite and chlorite-sericite. Orthoclase is commonly sericitised, plagioclase saussuritised, and biotite altered to chlorite. It is significantly more radioactive than typical granitoids, averaging 50 ppm Th and 25 ppm U (Ferguson *et al* 1980). U-Pb SHRIMP zircon geochronology constrains its emplacement to 1818 ± 8 Ma (Worden *et al* 2006b). The *Tin Camp Granite* outcrops in the north of the Caramal Inlier and in the southeast of the Beatrice Inlier, and comprises biotite granite, trondhjemite and porphyritic granite. It was emplaced into Neoarchaean basement and the Palaeoproterozoic Nimbuwah Complex,

and is unconformably overlain by the Kombolgie Subgroup of the Katherine River Group. The granite is pervasively altered and enriched in Th and U (average 62 and 10 ppm, respectively; Ferguson *et al* 1980) and is cut by quartz breccia zones (Needham 1988a). The main mineral constituents are plagioclase, quartz, orthoclase, biotite and rare opaques. Recent U-Pb SHRIMP zircon geochronology constrains its emplacement to 1846 ± 13 Ma, though this is based on only 8 analyses (Worden *et al* 2008b).

Mafic intrusive rocks

Ductily deformed and regionally metamorphosed *undifferentiated amphibolites* have previously been grouped and mapped as Zamu Dolerite throughout the Nimbuwah Domain, based on correlations with lower-grade sills deformed at, or after ca 1860 Ma within the Palaeoproterozoic succession of the Central Domain. This correlation is now known to be erroneous, based on geochemical data for amphibolites from the Nimbuwah Domain, which contrast strongly with the geochemistry of the Zamu Dolerite as defined from the Central Domain (Stuart-Smith *et al* 1984, 1988, Warren and Kamprad 1990, Ferenczi and Sweet 2005, Hollis and Glass 2012). Based on this data, it appears to be unlikely that the Zamu Dolerite is present within the Nimbuwah Domain. Although further work is required to delineate the extent of geochemically distinct, deformed and regionally metamorphosed amphibolites in the Nimbuwah Domain, several units are now known to occur based on (a) their distinct geochemical signatures, (b) the distinct mineralogy of amphibolites within the Cahill Formation (commonly calc-silicate-bearing) versus those entrained within Archaean felsic gneiss, and (c) the preservation of older ductile deformation fabrics in amphibolites within Archaean felsic gneisses relative to those within the structurally overlying Palaeoproterozoic succession (Hollis and Glass 2012).

The Oenpelli Dolerite is a series of sills, intrusive into the Palaeoproterozoic succession, that outcrop discontinuously throughout Arnhem Land and the Alligator Rivers region (Ferenczi and Sweet 2005). Originally thought to be overlain by the Kombolgie Subgroup (Needham 1988a), the unit is now known to be intrusive (Carson *et al* 1999, Sweet *et al* 1999, Drever *et al* 2000), the misinterpretation arising from the common emplacement of the Oenpelli Dolerite at the base of the subgroup. The sills are typically flat-lying, except where emplaced along faults. They are continuous over tens of kilometres and have a maximum thickness of 500–1500 m (Smart *et al* 1975). The unit is best exposed at this level in the succession as ridges or low hills, whereas elsewhere, it is poorly exposed and typically recognised as subcropping, based on the presence of red-brown soils and thicker vegetation, or by highs in magnetic data. The sills comprise porphyritic to ophitic olivine dolerite, minor olivine-plagioclase cumulates that locally define rhythmic layering, and narrow discontinuous granophyric dolerite within zones of quartz dolerite (Stuart-Smith and Ferguson 1978). Intrusions show chilled margins with their host rocks and locally, have developed contact metamorphic aureoles. Slight to extensive alteration to mineral assemblages involving chlorite, prehnite, epidote, and calcite is common (Needham 1988a). Based on the subalkaline nature and differentiated character of the intrusions, Stuart-Smith and Ferguson (1978) suggested a source derived from partial melting of the upper mantle and emplacement at 1–2 km. Ferenczi and Sweet (2005) made reference to a ca 1723 ± 6 Ma U-Pb baddelyite age for the Oenpelli Dolerite from a sample from the Nabarlek uranium mine. This requires further verification as the original data are missing from the OZCHRON database referred to by those authors. However, it is consistent with a Rb-Sr whole rock and mineral isochron minimum age of 1688 ± 13 Ma obtained by Page *et al* (1980).

LITCHFIELD DOMAIN

The Litchfield Domain comprises the westernmost region of the PCO, separated from the Central Domain to the east partly by the southern extension of the Giants Reef Fault and bounded to the west by the onlapping Mesozoic Bonaparte Basin. The Province is elongate roughly north–south and comprises relatively sparsely outcropping amphibolite- to granulite-facies metasedimentary rocks, and syn- to post-tectonic granitoids and mafic rocks (Berkman 1980, Pietsch and Edgoose 1988). High-grade metamorphic rocks in the Litchfield Domain were originally thought to be Archaean (Walpole *et al* 1968), but are now known to be Palaeoproterozoic (Dundas *et al* 1987, Carson *et al* 2008, 2009, Worden *et al* 2008a). No Archaean basement to the Palaeoproterozoic succession in the Litchfield Domain has been identified.

The Palaeoproterozoic strata comprise the Hermit Creek, Welltree and Fog Bay metamorphics. On the basis of detrital zircon work, the Hermit Creek and Welltree metamorphics have been correlated with the Burrell Creek Formation of the Cosmo Supergroup in the Central Domain (Worden *et al* 2008b, Hollis *et al* 2009b), whereas the Fog Bay Metamorphics may correlate with the Woodcutters Supergroup of the Central Domain (Carson *et al* 2009). Only the Welltree Metamorphics preserve a well developed ductile fabric associated with metamorphism, whereas the Hermit Creek and Fog Bay metamorphics show no significant fabric development and commonly preserve primary bedding structures (Carson *et al* 2008). Mid- to high-grade metamorphism at low pressure is constrained to ca 1855 Ma, based on *in situ* monazite geochronology (Carson *et al* 2008). This corresponds to the timing of emplacement of mafic rocks (ca 1860 ± 6 Ma; Carson *et al* 2009) and granitoids of the Allia Suite (ca 1865–1850 Ma; Worden *et al* 2008a). Geochemical and isotopic work is consistent with the formation of some of these mafic rocks in an arc or back-arc environment (Glass 2007, 2011), strengthening correlations made with the central and western zones of the Halls Creek Orogen (Carson *et al* 2008).

Fog Bay Metamorphics

The Fog Bay Metamorphics are located in the eastern part of Fog Bay in the northwestern Litchfield Domain. The unit is entirely subsurface, concealed by Mesozoic and Cenozoic sediments, and has been intersected only in drill core (Pietsch and Edgoose 1988). They are inferred to be faulted against

the Welltree Metamorphics along the north-northeast-trending Tom Turners Fault. The Fog Bay Metamorphics comprise quartz-feldspar-biotite gneiss and schist commonly containing garnet and sillimanite, amphibolite, minor calc-silicate, marble, and quartzite (Hickey 1985). Carson *et al* (2008) described a pelitic sample from drill core with banded compositional layering, but no pervasive fabric development. The same authors obtained a peak P-T estimate of 596 ± 26ºC and 4.6 ± 1.1 kb (2σ). Detrital zircon age spectra for the Fog Bay Metamorphics show a broad distribution in the range ca 2500–2000 Ma and a maximum depositional age of 2028 ± 8 Ma (Carson *et al* 2009). The detrital age spectra contrast with those of the Hermit Creek and Welltree metamorphics and the Burrell Creek Formation, but are consistent with a similar provenance to the Woodcutters Supergroup of the Central Domain. Monazite and metamorphic zircon rims give consistent age constraints on amphibolite-facies metamorphism of 1854 ± 3 Ma and 1853 ± 4 Ma, respectively (Carson *et al* 2008, Worden *et al* 2008a).

Hermit Creek Metamorphics

The Hermit Creek Metamorphics are a composite package of pelitic to psammitic schist and gneiss, migmatite (**Figure 5.18**), amphibolite, and lesser calc-silicate and cordierite hornfels in the southern part of the Litchfield Domain (Walpole *et al* 1968, Pietsch and Edgoose 1988). Carson *et al* (2008) subdivided the metamorphics into three units, but the relationships between these remain unclear. These are (i) granulite-facies cordierite migmatite (± garnet ± sillimanite ± spinel), (ii) cordierite–andalusite hornfels and (iii) mid-amphibolite-facies ferruginous biotite–muscovite–quartz schist, associated amphibolite and rare calc-silicate. The amphibolites probably represent extrusive or hypabyssal mafic igneous rocks and are composed of titaniferous hornblende, tremolite, actinolite and plagioclase. Graded bedding in the cordierite migmatites is well preserved, on a 150–200 mm scale, with no evidence of significant tectonic deformation. Carson *et al* (2008) obtained peak P-T estimates for mineral assemblages from the cordierite migmatites of 727 ± 50ºC at 5.5 ± 0.6 kb (2σ). Metamorphism is constrained by $^{207}Pb/^{206}Pb$ monazite geochronology on two samples of cordierite migmatites to 1855 ± 2 Ma and 1853 ± 4 Ma, respectively (Carson *et al* 2008). Detrital zircon age spectra for the Hermit Creek Metamorphics are indistinguishable from those of the Burrell Creek Formation of the Central Domain and, on that basis, the two units are likely to be correlatives (Worden *et al* 2008a, Hollis *et al* 2009b).

Welltree Metamorphics

The Welltree Metamorphics are poorly exposed in the northeastern Litchfield Domain, east of Tom Turners Fault. They consist mainly of mid- to upper amphibolite-facies quartzofeldspathic to pelitic schist and gneiss, which commonly contain porphyroblasts of garnet and sillimanite or andalusite, and which have a pervasive tectonic fabric (Pietsch 1986a, Carson *et al* 2008). The lower part of the succession is composed of marble, calc-silicate rocks, para-amphibolite and quartzofeldspathic gneiss and is assigned to the *Sweets Member*. The succession also includes para-amphibolite composed of hornblende + feldspar + quartz ± biotite ± diopside that is interpreted to be metamorphosed carbonate rocks, but may represent an intrusive of pyroxenite to peridotite composition (Pietsch 1986b). Carson *et al* (2008) obtained a peak P-T estimate for a garnet-sillimanite-bearing pelitic sample of 714 ± 143ºC at 3.9 ± 2.2 kb (2σ). Metamorphism and associated fabric development is constrained to 1813 ± 3 Ma, on the basis of $^{207}Pb/^{206}Pb$ monazite geochronology (Carson *et al* 2008), although the regional significance of this event is unclear. There is an apparent metamorphic gradient from the Welltree Metamorphics east into greenschist-facies rocks of the Burrell Creek Formation of the Central Domain, consistent with the interpretation of these as stratigraphic equivalents (Pietsch 1986b). This is supported by detrital zircon spectra for the Welltree Metamorphics, which are indistinguishable from those of the Burrell Creek Formation (Carson *et al* 2005, Hollis *et al* 2009b). However, the timing of greenschist-facies metamorphism of the Burrell Creek Formation is constrained to >1829 Ma by the deposition of the El Sherana Group and therefore, this apparent metamorphic gradient cannot be attributed to a single thermal event.

Granitoids

The granites of the Litchfield Domain include the Allia Creek and the Wagait suites (Wyborn *et al* 1997). U-Pb SHRIMP zircon, monazite, and xenotime ages typically range from 1863–1850 Ma for these suites. The limited data available for the Wagait Suite place it in the oldest part of this range (1863–1860 Ma), whereas the Allia Creek Suite is dominated by slightly younger granites (1854–1850 Ma).

Allia Creek Suite

The Allia Creek Suite is predominantly S-type and contains peraluminous minerals such as andalusite, cordierite and muscovite (Wyborn *et al* 1997). The granitoids are high-level intrusions (1–1.5 kb), marked by the development of numerous pegmatites and greisens. Petrological data,

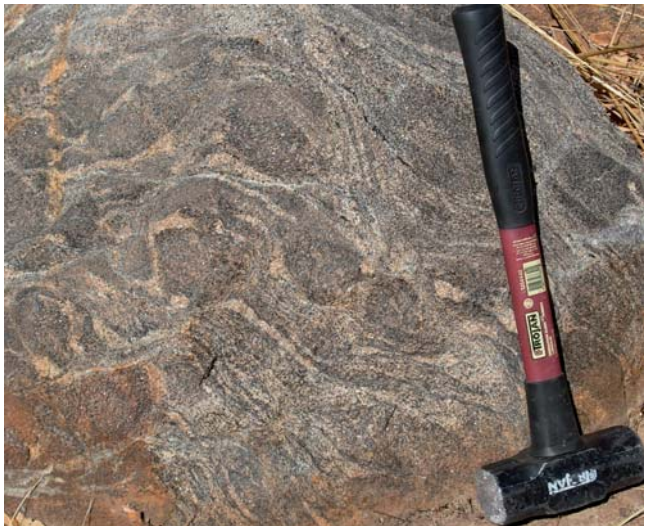


Figure 5.18. Migmatite of Hermit Creek Metamorphics. Foliation is defined by mafic and felsic minerals. Leucosomes wrap around mafic biotite/cordierite augens (52L 674446mE, 8448405mN).

combined with high concentrations of Rb and low values of Sr, indicate that crystal fractionation played a major role in the early crystallisation history of the granitoids. The following summary of the component plutons largely relies on Ahmad *et al* (1993) and Bajwah (1994) and Wyborn *et al* (1997).

The *Fish Billabong Adamellite* is predominantly fine to medium grained, partly heterogranular and is composed of quartz, K-feldspar, plagioclase and biotite. SiO_2 shows a narrow range (68.19–69.60 wt%) and CaO ranges from 1.58–1.82 wt%. Na_2O is high and ranges from 3.84–4.70 wt%. Rb is low, ranging from 91–112 ppm, and is comparatively low in comparison to the other Litchfield Domain granitoids. The Fish Billabong Adamellite has yielded a U-Pb SHRIMP zircon magmatic crystallisation age of 1850 ± 3 Ma (Worden *et al* 2008a).

The *Jammine Granite* comprises quartz-muscovite and feldspar-tourmaline greisens. It is medium grained and is predominantly composed of quartz and microcline, with subordinate coarse muscovite and minor tourmaline. Plagioclase and biotite show partial to complete alteration to sericite and chlorite. The Jammine Granite is high in SiO_2 and K_2O, and poor in FeO, MgO and TiO_2. Similarly, Rb concentrations are the highest amongst the Allia Suite, ranging from 366–571 ppm and averaging 494 ppm. On the other hand, Sr concentrations are low (range 75–330 ppm, average 165 ppm). All the data suggest that the Jammine Granite is highly fractionated. Sn and Li contents are also high and therefore, may be a potential source of Sn mineralisation in the area. The Jammine Granite has yielded a U-Pb SHRIMP zircon maximum intrusion age of 1858 ± 5 Ma (Worden *et al* 2008a).

The *Mount Litchfield Granite* is a fairly consistently foliated, fine- to medium-grained granite to quartz monzonite. Quartz occurs as anhedral strained grains, which may be elongated parallel to the foliation. Plagioclase ($An_{40–50}$) in the less foliated rocks is generally intergrown with quartz and is often zoned. K-feldspar is present mainly as large crystals, enclosing plagioclase and biotite. Biotite forms tabular crystals up to 2 mm long and usually defines a moderate foliation. Muscovite, a minor constituent, is generally associated with biotite. SiO_2 shows a large range, varying from 64.04–73.94%. Na_2O ranges from 2.60–4.86%. K_2O is low (1.64–5.16%), Rb ranges from 6228–271 ppm, averaging 152 ppm, Ba from 156–1780 ppm and Sr from 139–489 ppm.

The *Murra-Kamangee Granodiorite* is characterised by a moderate to strong foliation. It is composed of medium-grained garnet-bearing granodiorite and tonalite. Plagioclase ($An_{45–55}$) is the most common feldspar and forms up to 1 mm-long, tabular crystals with some zoning. Microcline is the predominant K-feldspar variety and forms euhedral megacrysts, usually less than 1 cm long. Biotite constitutes 10–15% of the rock and occurs as subhedral to anhedral laths. Garnet is almost always present as euhedra, up to 2 cm in diameter. Accessory minerals are apatite, titanite, ilmenite, magnetite, muscovite, andalusite, sillimanite, cordierite and zircon. SiO_2 ranges from 61.75–72.90%. K_2O and Na_2O are low compared with the other granitoids. Rb ranges from 120–269 ppm, Ba from 360–890 ppm and Sr from 75–157 ppm. The Murra-Kamangee Granodiorite has yielded a ^{207}Pb-^{208}Pb SHRIMP monazite magmatic crystallisation age of 1853 ± 3 Ma (Worden *et al* 2008a).

The *Two Sisters Granite* has two distinctive phases: a fine to medium variably foliated granite and coarse garnet-bearing pegmatitic granite. The less-foliated granite is equigranular and generally contains quartz, plagioclase, orthoclase and subordinate biotite. Quartz occurs as irregular, mainly interstitial grains up to 4 mm across and, in more foliated varieties, it is generally elongated parallel to the foliation. Plagioclase ($An_{40–50}$) forms subhedral to anhedral crystals. The larger phenocrysts (up to 3 mm) are usually zoned and more extensively altered. K-feldspar includes both orthoclase and microcline. Myrmekitic development of orthoclase is also common. Biotite is always strongly pleochroic and locally altered to chlorite. Muscovite is a minor constituent in the foliated granite and is generally interleaved with biotite, or present as randomly oriented small flakes within feldspar and biotite. However, towards the northeast, where the granite is cut by pegmatitic veins, up to 10% muscovite may be present. Garnet occurs as subhedral to anhedral, generally fractured crystals disseminated in the matrix, or is present in pegmatitic veins. The pegmatitic granite consists of an early-stage quartz-albite-muscovite sodic granodiorite and a quartz-feldspar graphic leucogranite. The latter is intruded by garnet- and tourmaline-bearing pegmatites. The Two Sisters Granite is characterised by high SiO_2 (69.9–74.9 wt%) and low Na_2O (2.2–4.9 wt%). Rocks that are relatively high in Na_2O are generally enriched in albite. K_2O ranges from 2.07–5.88 wt%. The pegmatitic variety is low in TiO_2, FeO, MgO, Sr, Zn, Ba and La, and relatively high in SiO_2, Na_2O, Al_2O_3, Rb, Bi, Li and Sn. The Two Sisters Granite has yielded a U-Pb SHRIMP zircon maximum intrusion age of 1862 ± 5 Ma (Worden *et al* 2008a).

The *Allia Creek Granite* is a biotite quartz monzonite and is often porphyritic, with large phenocrysts (up to 60 mm in length) of microcline. The phenocrysts often display a crude platy alignment, defining a flow structure that appears to be parallel to the margins of the granite (Edgoose *et al* 1989b). Biotite occurs in clusters and is associated with apatite and zircon. Muscovite is rare. The Allia Creek Granite has yielded a U-Pb SHRIMP zircon maximum intrusion age of 1806 ± 7 Ma (Worden *et al* 2008a).

The *Soldiers Creek Granite* represents a coarse porphyritic muscovite leucogranite associated with greisen veins. Numerous xenoliths (up to 30 cm in size) occur throughout the granite body. Quartz, orthoclase, microcline, plagioclase, biotite and muscovite are the main minerals. The age of the Soldiers Creek Granite is constrained to <1830 Ma, based on the age of a mafic sample entrained within the granite (Worden *et al* 2008a).

Wagait Suite

The Wagait Suite comprises the Wagait, Peppimenarti, Reynolds River and the Koolendong granites. Relatively little data is available for this suite. They are assumed to be restite-rich with some late fractionation of a felsic phase. Recent U-Pb zircon geochronology studies of the Koolendong and Wagait granites, yielding ages of 1860 +2/-5 Ma and 1863 ± 3 Ma respectively (Worden *et al* 2006b, 2008b), suggest that the Wagait Suite may be slightly older than the Allia Creek Suite.

The *Wagait Granite* consists of mainly medium-grained granodiorite, monzogranite and granite. The dominant

minerals are quartz, K-feldspar, plagioclase, biotite and hornblende, with accessory epidote, apatite and zircon. Saussuritisation and chloritisation of biotite is common. The Wagait Granite has yielded a U-Pb SHRIMP zircon emplacement age of 1863 ± 3 Ma (Worden *et al* 2008a).

The *Peppimenarti Granite* is more felsic than the Wagait Granite and is dominated by coarse-grained, massive and equigranular monzogranite to granite, with some aplite and pegmatite phases present, suggesting late-stage evolution of a magmatic phase. The dominant minerals are plagioclase, quartz, biotite and poikilitic orthoclase. Saussuritisation and chloritisation of biotite is common.

The *Reynolds River Granite* is a small stock of light grey to pink, medium- to coarse-grained biotite granite to monzogranite. It outcrops as low pavements and small bouldery hills. In places, the outcrop is distinctively layered.

The *Koolendong Granite* is a biotite-hornblende granodiorite and leucocratic porphyritic quartz monzonite. Quartz and pegmatite veins are abundant. It intrudes the Burrell Creek Formation (Morgan 1972, Sweet *et al* 1974a). The Koolendong Granite has yielded a U-Pb SHRIMP zircon emplacement age of 1860 +2/-5 Ma (Worden *et al* 2008a).

Mafic intrusive rocks

Previously, all mafic and ultramafic intrusive rocks in the Litchfield Domain were assigned to the *Wangi Basics* (Dundas *et al* 1987), a term that is now discontinued. The intrusive rocks comprise gabbro, leucogabbro, gabbronorite, dolerite, ultramafic rocks (the majority of which have been metamorphosed to amphibolite facies) and rare basalt (Edgoose *et al* 1989a, b, Glass 2007, 2011). Typical mineral assemblages include plagioclase, clinopyroxene, orthopyroxene, and hornblende ± quartz, although greenschist-facies retrogression to tremolite, actinolite, chlorite and epidote is common (Carson *et al* 2008, Glass 2011). These rocks occur as scattered outcrops and subcrops largely to the west of the Giants Reef Fault, with magnetic data indicating that they may be considerably more extensive under shallow cover. On the basis of significant geochemical and isotopic differences between suites of mafic rocks in the region, Glass (2011) subdivided these mafic rocks into four suites: a southern *Woolianna Gabbro* (high-alumina gabbro and dolerite, depleted low-Ti gabbro, low-Ti dolerite); the *Benning Gabbro* (high-Ti rift tholeiite); a northern *Keri Metamorphics* (metamorphosed noritic cumulates, metaferrogabbro) and, to the east of the Giants Reef Fault, the younger *Lilyarba Mafics* of probable shoshonitic/lamprophyric affinity (Glass 2011) dated at 1829 ± 4 Ma (Worden *et al* 2008a). With the exception of the northern Keri Metamorphics, the other suites show geochemical affinities with island-arc magmatism, with the low-Ti gabbro rocks of the Woolianna Gabbro probably being representative of back-arc basin oceanic crust (Glass 2007, 2011). The age of at least some of these mafic rocks is constrained to >1853 ± 3 Ma (Worden *et al* 2008a), on the basis of mafic enclaves within the Murra-Kamangee Granodiorite. Similarly, Carson *et al* (2009) obtained a magmatic crystallisation age of 1860 ± 6 Ma for a garnet-bearing amphibolite. However, given the recently identified geochemical subdivisions, it is possible that some of the sub-units are of different ages.

STRUCTURE

Deformation and structure in the PCO has been the subject of several studies. Johnston (1984) provided a comprehensive structural study. Hammond *et al* (1984) and Pietsch and Edgoose (1988) described the structural elements of the western PCO (Litchfield Domain). Stuart-Smith *et al* (1993), Needham *et al* (1988) and Ahmad *et al* (1993) provided regional-scale summaries, but essentially focused on the Central Domain. Lally (2002) discussed the structure of the Rum Jungle Mineral Field. The following description of the structure of the PCO relies on these earlier studies. The main structural elements are given in **Figure 5.19**.

Structural, tectonometamorphic and geochronological studies of the PCO indicate that there were distinct styles and possibly unrelated tectonism in different parts of the orogen (Johnston 1984, Needham 1988a, Pietsch and Edgoose 1988, Carson *et al* 2008, Hollis *et al* 2009b). Thus, any representation of the structural evolution as a single sequence of events applicable to the whole orogen is unlikely to be realistic. Even within a single structural domain, there is potential for complexity, owing to the possibility of diachroneity and variations in structural style between thrust sheets and within deformational phases (eg Johnston 1984). Nonetheless, in order to allow comparison of the relative timing of events and possible correlations across the orogen, the major stages of structural development are presented as a sequence of events from D_n–D_5 (Johnston 1984). Little mention is made of the structural evolution of the Litchfield Domain, which is relatively little-studied and is dominated by subvertical northeast–southwest-trending, composite S_1/S_2 foliations, as these are defined by high-grade low-P mineral assemblages (Pietsch and Edgoose 1988, Carson *et al* 2008) that do not have any clear equivalents in the Central or Nimbuwah Domains.

Archaean–Palaeoproterozoic deformation

Dn

The earliest fabrics are those preserved in weakly deformed Neoarchaean granitoids of the Rum Jungle Complex and in high-strain, multiply ductily deformed Neoarchaean gneisses in the Nimbuwah Domain, collectively referred to here as D_n fabrics. In the Rum Jungle region, the earliest structures are represented by tightly refolded folds within metasedimentary enclaves entrained within the Neoarchaean granitoids. The granitoids also contain a weak steep D_n foliation, defined by biotite and plagioclase (Johnston 1984). In the Nimbuwah Domain, Neoarchaean felsic gneisses preserve evidence for multiple phases of ductile deformation, in the form of isoclinally refolded folds and variably transposed gneissic fabrics, and heterogeneous development of high-strain fabrics; these are collectively assigned to D_n (Needham 1988a, Hollis *et al* 2008).

D_1

D_1 in the Central Domain is represented by northwest-trending stretching lineations, defined by clast elongation or flattening or both, recumbent folds, and a high-strain S_1 foliation parallel to bedding, which can also be found in the basement in the Rum Jungle region. It is associated

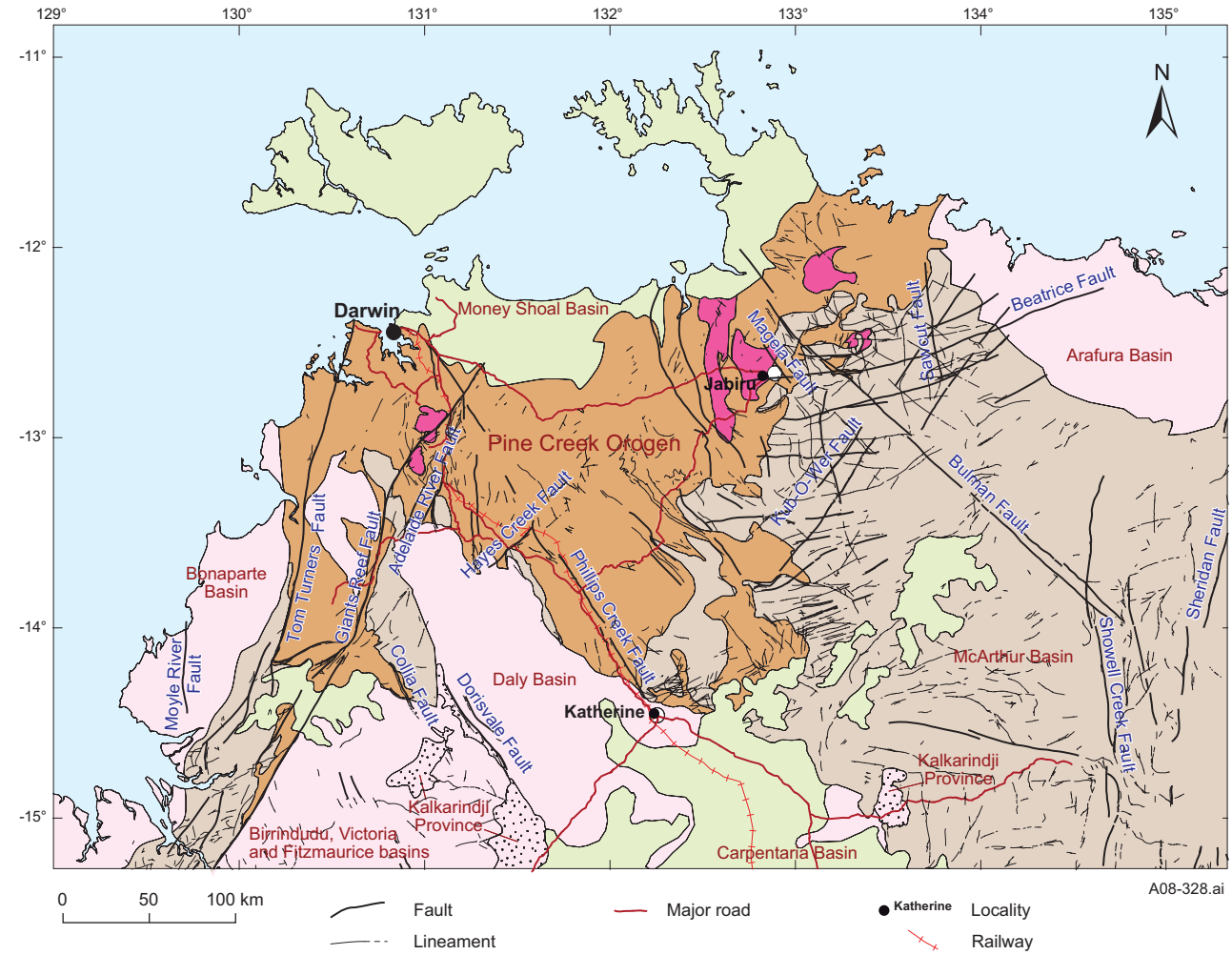


Figure 5.19. Major faults and lineaments of the PCO. NT geological regions from NTGS 1:2.5M geological regions GIS dataset.

with north-trending monoclinal warping and resulted in northwest-vergent tectonic sliding along the Neoarchaean–Palaeoproterozoic contact. D_1 produced a concordant slaty cleavage (F_1), elongation of cobbles, and mylonitic textures in some sandstones within the Crater Formation (Johnston 1984). A map-scale fold, defined by the Celia Dolostone and Crater Formation to the southeast of the Woodcutters mine, is also a product of D_1 deformation (Lally 2002). D_1 foliations in the South Alligator River region, Central Domain, are associated with northwest-trending monoclinal warps and are folded by subsequent fold generations (Johnston 1984). This warping predates deposition of the South Alligator Group and probably reflects pre-Nimbuwah Event deformation (Stuart-Smith 1993). In the East Alligator River region of the Nimbuwah Domain, D_1 is represented by amphibolite-facies S_1 foliations parallel to bedding (Johnston 1984).



Figure 5.20. F_3 folding in road cutting near Moline Mine (53L 191930mE, 8488577mN).

D_2

D_2 is more pronounced in the western and eastern parts of the PCO, where metamorphic grades are higher. D_2 structures are parallel to bedding for tens of kilometres, but are locally steepened so as to cut through the succession. In the low metamorphic grade regions of the Central Domain, D_2 thrusts are brittle structures several metres wide with associated 1–2 m-wide duplex structures. In the Nimbuwah Domain, amphibolite-facies D_2 foliations are at a low angle to bedding and are associated with northwest-verging, low-angle high-strain thrusts. D_2 is also expressed as recumbent F_2 folds, an S_2 schistosity and tectonic slides. These fabrics have associated east-trending stretching lineations, consistent with a top-to-the-west sense of movement (Johnston 1984, Hollis *et al* 2008, Hollis *et al* 2009b). In fault zones, D_2 is represented by mylonites, transposed fabrics and cataclastic zones.

D_3

D_3 is represented by north- to northwest-trending, macroscopic to mesoscopic, noncylindrical close to tight folds (**Figure 5.20**). These F_3 folds are usually symmetrical and are either upright or inclined to the southwest. A

penetrative near-vertical slaty cleavage, parallel to the axial surface of the fold, is usually well developed in shale. These folds control the outcrop pattern and typically have wavelengths ranging from 1–3 km, amplitudes of 1–1.5 km and plunges that are gently inclined to the north or south (Johnston 1984). In the Rum Jungle region, F_3 folds refold D_1 décollement structures (Johnston 1984). In the Nimbuwah Domain, F_3 folds are upright to reclined and open to isoclinal (Johnston 1984).

D_4

D_4 produced east–west-trending open folds in the Central Domain and similarly trending, open to tight F_4 folds in the Nimbuwah Domain. Domical structures of the Neoarchaean basement in the Rum Jungle region and in the Nimbuwah Domain are attributed to the interference of D_3 and D_4 fold structures (Johnston 1984).

D_5

D_5 resulted in kinking and box-folding in the Rum Jungle region, associated with basement block movement.

Faulting

Although no systematic or detailed studies have been conducted on the history and types of faulting in the PCO, several generations and types can be identified and these are tentatively correlated with the structural evolution discussed above.

Syn-sedimentation faults

Initial sedimentation in the PCO is likely to have been in response to the rifting and extension of Archaean crust. Stuart-Smith (1985) interpreted a Central Trough palaeoenvironment, flanked by shallow shelves. Several east-dipping half graben tilt blocks, separated by major west-dipping normal faults, were envisaged (Needham *et al* 1988). Subsequent sedimentation largely covered these faults. The South Alligator, Mount Shoobridge and Hayes Creek faults and possibly the Pine Creek Shear Zone may represent some of these early basin development faults, which have been subsequently reactivated.

Previous studies (Walpole *et al* 1968, Stuart-Smith *et al* 1980, Needham *et al* 1988, Pietsch and Edgoose 1988) have identified several major basement faults that also controlled sedimentation and the configuration of the PCO. The *Western Fault Zone* is a possible tilt-block boundary that separates the Litchfield Domain from the rest of the PCO. It appears to have played an important control on sedimentation and tectonics in the western half of the PCO. The *Giants Reef Fault* is possibly a post-Palaeoproterozoic expression of the Western Fault Zone that extends over 200 km and is part of the laterally extensive faults in the Halls Creek and Fitzmaurice mobile zones. Palaeoproterozoic movements on these faults were essentially dip slip, in the normal sense, with the downthrown side to the west. Later movements (post-middle Proterozoic) were mainly along strike slip faults.

The *Stapleton, Adelaide River and Mount Shoobridge* faults may also represent basement structures that were probably subjected to vertical movements during sedimentation, as is reflected by the presence of a coarse clastic facies (formerly the Noltenius Formation) of the Burrell Creek Formation, mainly to the west of the Mount Shoobridge Fault. Further east, the northwest-trending faults in the *South Alligator River Valley*, and the *Bulman Fault* in the Alligator Rivers region are also an expression of basement faults (Needham *et al* 1988).

D_1 Faults

In most areas, faults produced during the D_1 deformation cannot be distinguished from those caused by later deformation episodes. The radial faults along the margin of the Rum Jungle and Waterhouse domes probably belong to this category. These faults continue for short distances in both directions away from the granite–sedimentary rock contact, and in places, the sedimentary rocks are displaced laterally along the faults. Most of the faults only affect strata of the Namoona and Mount Partridge groups.

D_2 Faults

Faults formed during the D_2 deformation represent zones of cataclasis, mylonitisation and tectonic sliding. They are common in amphibolite-grade terranes in the Litchfield Domain and the eastern PCO. D_2 faults are zones of thrusting and, although extending over several tens of kilometres, they are not clearly discernable on geological maps because of bedding parallel geometry. They have a non-planar surface, controlled by the anisotropy of the enclosing rocks.

D_3 Faults

Faults associated with the D_3 deformation episode trend northwest, have steep dips or are vertical and cut Palaeoproterozoic sedimentary rocks throughout the Central Domain. In most cases, these faults terminate at the contacts between the sedimentary rocks and granitoids, thus suggesting a pre-granitoid age. However, in some instances (eg southern extension of the Pine Creek Shear Zone), they continue through the granitoids and suggest a post-granitoid age or the reactivation of pre-granitoid faults.

D_3 faults run parallel to the regional trend and axial plane cleavage (S_3) of the Palaeoproterozoic sedimentary rocks and possibly developed during, or after F_3 folding. These faults commonly show a history of reactivation since Palaeoproterozoic time (Stuart-Smith 1985).

The *Pine Creek Shear Zone* trends 140–150° and forms a part of a major northwest-trending strike slip fault system, extending from Darwin to Katherine, that Simpson *et al* (1980) termed the *Noonamah–Katherine Lineament*. The shear zone follows the embayment of two lobes of the Cullen Batholith. Metasedimentary rocks within the shear zone are chloritic and show vertical cleavage (Stuart-Smith 1985). Upon intersecting the granite, the lineament splays into many discrete fault breccias and mylonite zones. Granitoids within the Pine Creek Shear Zone are foliated, strained and extensively altered. Movements along the shear zone, as shown by displacements, have probably taken place before, during and after granitoid emplacement (Stuart-Smith *et al* 1987), and it is therefore possible that the shear zone originated during the D_3 deformation event and was reactivated during the D_4, D_5 and later episodes.

Simpson *et al* (1980) identified another major lineament to the east of and approximately parallel to the Pine Creek Shear Zone and termed it the *Noonamah–Mount David Lineament*. Several northwest-trending faults correspond to this lineament and clearly terminate at the granitoid–sedimentary rock contact.

D_4 Faults

D_4 faults trend west-northwest, northeast or north. Several quartz, greisen, tourmaline, aplite and microgranite veins within and outside the granitoids also represent these structures. These faults are probably associated with granitoid emplacement. Most have a strike length of less than 5 km, but some may extend to up to 10 km.

D_5 Faults

These faults trend northeast or northwest and some have caused displacement of the platform cover strata. They may be reactivated D_3 or D_4 faults.

The *Hayes Creek* and *Mount Douglas* faults are major northeast-trending reverse faults. Both faults are probably continuous under cover and form part of the 'Grove Hill Cross Flexure' (Walpole *et al* 1968), a regional northeast-trending structure, about which the direction of plunge changes and the F_3 fold axes are rotated from northwest to north.

Northwest-trending faults along the eastern margin of the Daly Basin are steep normal faults along which movement took place during the Mesoproterozoic, Palaeozoic and possibly Mesozoic, down-faulting sediments to the southwest (Stuart-Smith *et al* 1987).

TECTONOTHERMAL EVOLUTION

Regional metamorphism of the PCO, and of the North Australian Craton in general, in the period ca 1880–1850 Ma has historically been referred to as the *Barramundi Orogeny*, a broad period of tectonic activity relating to cratonisation of northern Australia, involving intra-plate continental rifting, basin formation, and subsequent compression (Etheridge *et al* 1987, Needham *et al* 1988, Page and Williams 1988). This interpretation was based on similarities in metamorphic and deformational styles across the PCO, indicating a broad regional extent for the orogeny, a lack of evidence for significant crustal thickening, the absence of paired metamorphic belts, and the absence of magmatic rocks with geochemical characteristics indicative of subduction-related processes (Etheridge *et al* 1987). However, subsequent geochemical, petrographic, and U-Pb monazite and zircon studies in the PCO have identified distinct thermal events in different domains, with different styles of metamorphism, and indications of significant crustal thickening and of subduction-related tectonism (Glass 2007, 2011, Worden *et al* 2008a, b, Carson *et al* 2008, Hollis *et al* 2009a, b). Previous studies have also recognised evidence for diachronous metamorphism and subduction-related tectonics in other parts of the North Australian Craton (Tyler *et al* 1995, Zhao and Bennett 1995, Myers *et al*, 1996, Page 1997, Sheppard *et al* 1999, Tyler *et al* 1999). Consequently, attributing magmatism, volcanism, metamorphism, deformation and sedimentation across the North Australian Craton in this period to a single orogenic event is misleading. With reference to the PCO, the term Barramundi Orogeny is therefore abandoned here and the tectonic evolution of the PCO during this period is redescribed in terms of two significant tectonic events: the ca 1865 Ma Nimbuwah Event and the ca 1855 Ma Litchfield Event (see also Hollis *et al* 2009a).The Central and Nimbuwah domains share Neoarchaean basement of similar ages (ca 2670 Ma and 2530–2500 Ma, Page *et al* 1980, Cross *et al* 2005, Hollis *et al* 2009a) and geochemical affinities (Glass *et al* 2009). No such Neoarchaean crust has yet been identified in the Litchfield Domain. On this basis, it seems likely that the Central and Nimbuwah domains shared a common Neoarchaean history. However, differences in the provenance of post-Woodcutters Supergroup sedimentary rocks (Hollis *et al* 2009b), the style of Palaeoproterozoic structural development, and the nature and grade of metamorphism between the domains support a model in which these domains were independent of each other after ca 2020 Ma, until emplacement of the Cullen and Jim Jim granite suites at ca 1830 Ma. The distinct post-Woodcutters Supergroup Palaeoproterozoic depocentres in the Central and Nimbuwah domains (Central and Eastern troughs, respectively; Ahmad and McCready 2001) may have developed as a result of ca 2020 Ma rifting of the Neoarchaean basement, forming a series of half grabens in an evolving rift-subsidence setting (Stuart-Smith *et al* 1980, Needham *et al*, 1980, 1988, Plumb *et al* 1981, Etheridge *et al* 1987, Wyborn 1988, Worden *et al* 2008b).

The ca 1865 Ma *Nimbuwah Event* affected only the Nimbuwah Domain. Regional, moderate-T, moderate-P amphibolite-facies metamorphism (Hollis *et al* 2009b) is thought to have been induced by the emplacement of I-type granodioritic plutons of the Nimbuwah Complex, at depth, into Neoarchaean basement and the overlying Palaeoproterozoic succession at 1868–1860 Ma (Needham 1988a, Wyborn *et al* 1997, Page *et al* 1980, Worden *et al* 2008a, Hollis *et al* 2009b). This was coincident with felsic volcanism in the Central Domain (ca 1863 Ma Gerowie tuff, ca 1863 Ma Berinka Volcanics and Warrs Volcanic Member; Worden *et al* 2008a, b). Moderate-P amphibolite-facies metamorphism in the Nimbuwah Domain (in contrast with the Central Domain and Litchfield Domain) is indicated by occurrences of garnet-bearing amphibolite and garnet-kyanite-bearing pelitic schist in Palaeoproterozoic strata in the Nimbuwah Domain (Ferguson 1980, Hollis *et al* 2009b). This may be related to the development of D_2 northwest-vergent folding and thrusts, which are prevalent throughout the Nimbuwah Domain (Johnston 1984, Needham 1988a, Hollis *et al* 2008, Hollis *et al* 2009b). Reaction textures, involving garnet-breakdown to plagioclase-biotite in amphibolites, are consistent with subsequent decompression (Hollis *et al* 2009b).

The *Litchfield Event* is well constrained by U-Pb and Pb-Pb dating of metamorphic zircon and monazite to ca 10 million years later at ca 1855 Ma (Worden *et al* 2008a, Carson *et al* 2008). This event is distinct from the Nimbuwah Event in that it involved high-T low-P amphibolite- to granulite-facies metamorphism of pelitic to psammitic and mafic units, locally with little or no associated fabric development (Dundas *et al* 1987, Pietsch and Edgoose 1988,

Carson *et al* 2008). The presence of biotite-, cordierite- and sillimanite-bearing assemblages and the absence of both staurolite and kyanite are consistent with high-T low-P metamorphism. This event was synchronous with the emplacement of S-type granites of the Allia Creek Suite at 1862–1850 Ma (Walpole *et al* 1968, Page *et al* 1980, Wyborn *et al* 1997, Worden *et al* 2008a). On the basis of geochemical and isotopic characterisation, Glass (2007, 2011) identified several suites of mafic rocks (formerly assigned to the Wangi Basics) in the Litchfield Domain that show subduction-related affinities. The magmatic crystallisation age of one of these has been constrained to 1860 ± 6 Ma (Carson *et al* 2009), giving rise to the possibility that metamorphism at ca 1855 Ma in the Litchfield Domain was induced by extensional mafic magmatism (Carson *et al* 2008).

In the Central Domain, the sedimentary rocks were regionally metamorphosed to lower greenschist facies. Regional metamorphism of shale produced fine-grained, weakly foliated sericitic and microcrystalline quartz-bearing rocks that show a phyllitic texture. Psammitic rocks usually have fractured and strained quartz grains that exhibit recrystallised, optically continuous quartz overgrowths. Minor metamorphic sericite, chlorite or muscovite is present in these rocks. Eutaxitic textures are commonly preserved in tuffaceous sedimentary rocks, which are devitrified and altered to chlorite, sericite and quartz (Stuart-Smith 1985). Calcareous units contain calcite + tremolite ± phlogopite ± talc and or dolomite + calcite + magnesite ± tremolite assemblages (Ferguson 1980, Pietsch 1986a, 1989). The absence of cordierite, staurolite, or both and the presence of sericite, muscovite, chlorite and tremolite suggest P-T regimes of less than 2 kbar and 400–500°C (Ferguson 1980). The timing of greenschist-facies metamorphism and associated northwest–north-trending upright folding in the Central Domain (**Figure 5.21**) is constrained only by the maximum depositional ages of the Finniss River Group and emplacement of the Cullen and Jim Jim granite suites. The majority of detrital zircon data for the Burrell Creek Formation gives maximum depositional ages coincident with the age of the main erosive sources at ca 1865–1860 Ma (Worden *et al* 2008a). However, a single sample from immediately underlying the Tollis Formation, at the top of the Burrell Creek Formation, has yielded a slightly younger maximum depositional age of ca 1851 Ma, although with a relatively large error of 9 Ma (Hollis *et al* 2009a). Nonetheless, this suggests that deposition in the Central Domain is likely to have continued until after the ca 1865 Ma Nimbuwah Event into at least the same period as the ca 1855 Ma Litchfield Event. A minimum constraint is given by emplacement ages of ca 1835–1820 Ma for the Cullen and Jim Jim granite suites (Jagodzinski and Wyborn 1997, Wyborn *et al* 1997, Wyborn *et al* 2001) and by the 1829 Ma deposition of the El Sherana Group (Jagodzinski 1998, 1999). Thus, the timing of low-grade metamorphism and upright folding in the Central Domain is likely to have occurred at, or after the ca 1855 Ma Litchfield Event.

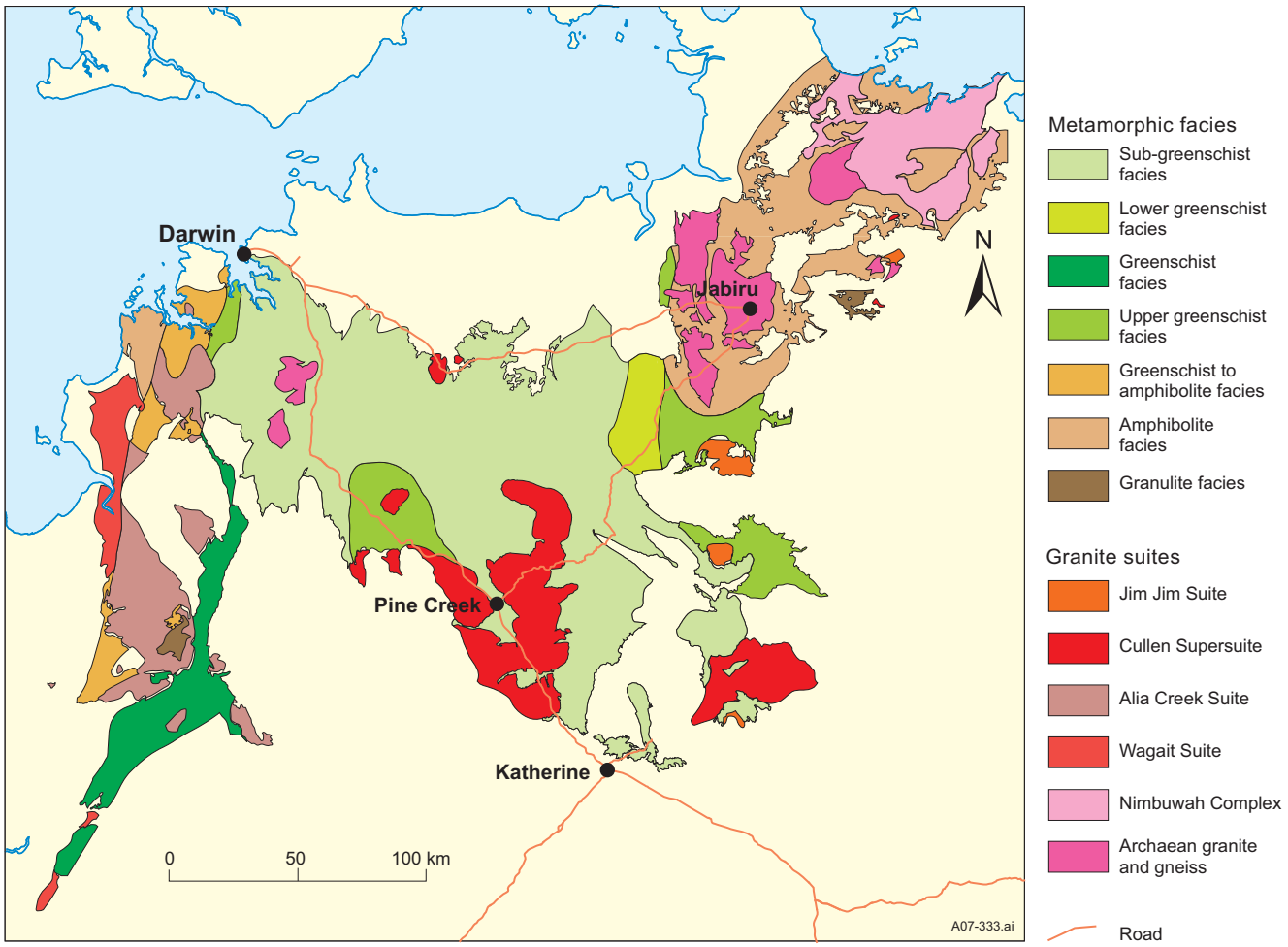


Figure 5.21. Spatial distribution of regional metamorphic facies in the PCO, modified from the Pine Creek Metallogenic GIS package 1:500 000 scale (Geoscience Australia 1995).

Contact metamorphism of metasedimentary rocks of the Central Domain is associated with emplacement of the Cullen and Jim Jim granite suites (**Figure 5.22**). Around the Burnside and Shoobridge granites, calcareous rocks of the Koolpin Formation contain tremolite, garnet, quartz and biotite. Iron-rich sedimentary rocks contain ferroactinolite, garnet, siderite and iron-rich chlorite. The boundary between the albite-epidote and the hornblende-hornfels facies is marked by the hornblende isograd, which occurs at the surface within 2 km of the granitoid boundary (Stuart-Smith 1985, Stuart-Smith *et al* 1993). The characteristic mineral assemblages comprise muscovite, quartz, diopside and rare K-feldspar and cordierite. The absence of sillimanite suggests P-T regimes of 2–4 kbar and 550–680°C. The albite-epidote facies could be divided into two parts by the biotite isograd (Stuart-Smith 1985). The inner, high-grade biotite-bearing zone is generally massive and hornfelsic, and contains readily visible muscovite, biotite and chiastolite. The pelitic rocks of the outer, lower-grade zone typically contain recrystallised foliated sericite, muscovite, chlorite and quartz. The presence of sericite, quartz, chiastolite, tremolite and calcite, and absence of diopside, K-feldspar and biotite suggests P-T regimes of 2–4 kbar and 500–600°C. Pelitic hornfels rafts and xenoliths within the granitoids and some pelitic rocks of the Burrell Creek Formation, close to the McMinns Bluff Granite, contain cordierite, andalusite, K-feldspar, biotite, muscovite and quartz. The presence of K-feldspar and andalusite, muscovite and biotite, and the absence of sillimanite suggest temperatures higher than the albite-epidote facies, probably in excess of 650°C.

The ca 1780 Ma *Shoobridge Event* (Needham *et al* 1988, Stuart-Smith *et al* 1993) is characterised by the development or reactivation of linear shear zones parallel to the regional north–south to northwest–southeast strike within both granites and metasedimentary rocks. Widespread retrograde metamorphism is also attributed to this event, as is reactivation and chloritisation within the Pine Creek Shear Zone. The age of this deformation event is reflected by Rb-Sr resetting at ca 1780 Ma in both the Cullen Supersuite granites and the metasedimentary rocks. Monazite dating (Sener *et al* 2003, Sener 2004, Carson *et al* 2005) also provides evidence for this younger retrograde event, which may have coincided with the initiation of the McArthur Basin. The retrograde and or hydrothermal effects of the Shoobridge Event are evident in alteration effects, such as the replacement of andalusite and cordierite porphyroblasts by muscovite and quartz, sericitisation of K-feldspar, and alteration of biotite and hornblende to chlorite and quartz.

MINERAL RESOURCES

The PCO hosts over a thousand mineral occurrences and is amongst the most prospective geological regions of Australia. It contains about 20% of the World's low-cost uranium resource, has a known resource of about 9 Moz of gold and has produced approximately 3.2 Moz to 2007 since 1870. Considerable resources of nickel-cobalt-lead-

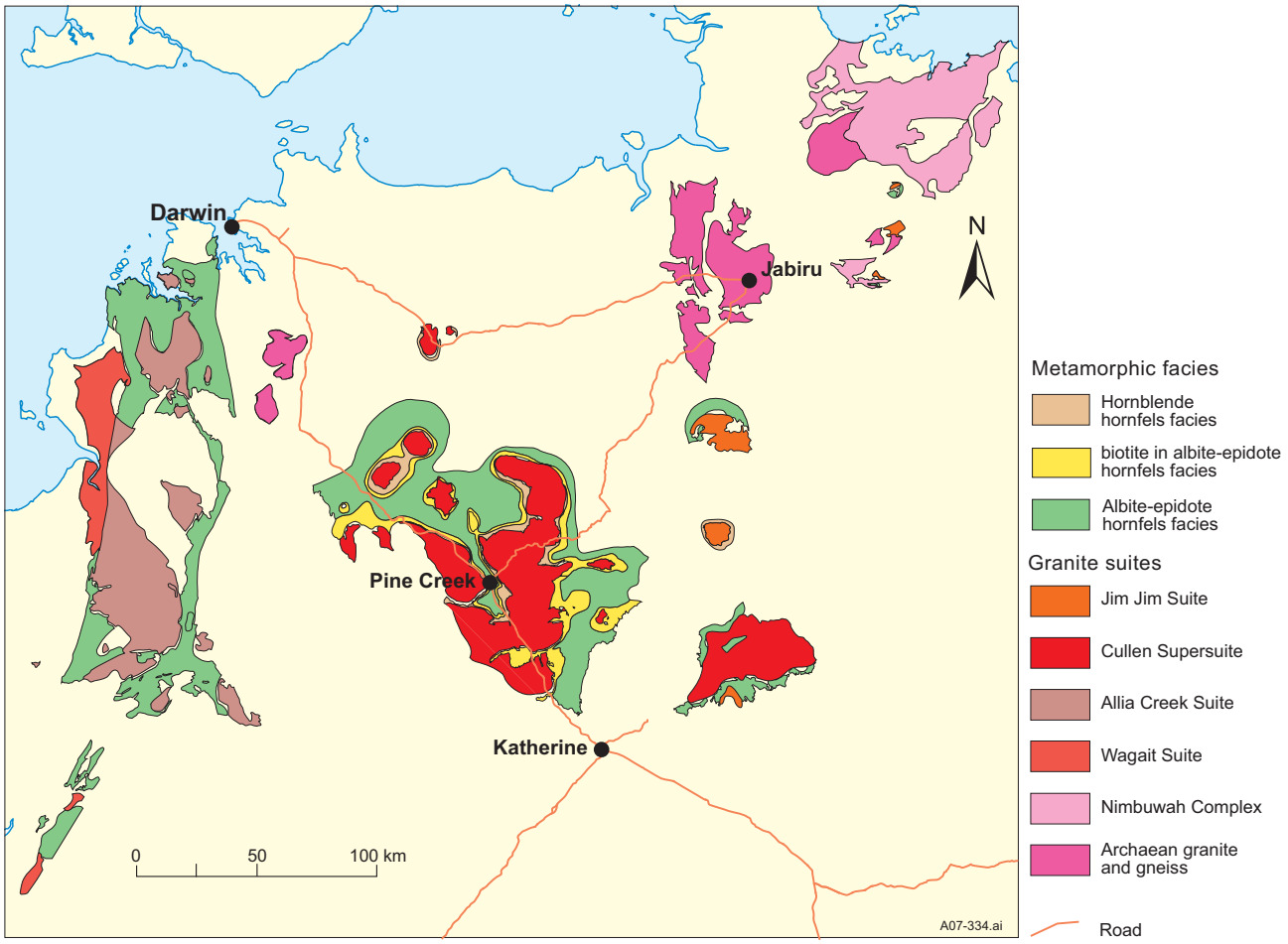


Figure 5.22. Spatial distribution of contact metamorphic facies in the PCO, modified from the Pine Creek Metallogenic GIS package 1:500 000 scale (Geoscience Australia 1995).

copper, lead-zinc-silver, platinum-palladium, tin-tantalum-tungsten, iron, magnesite, phosphate and various other commodities also exist. Economically, uranium and gold are the most dominant commodities. Details of these occurrences are contained within the Mineral Occurrence Database of the Northern Territory (MODAT).

Gold

FH Litchfield made the first discovery of gold in the Territory in 1865 in the Finniss River area of the PCO. Three years later, specks of gold were found at Tumbling Waters (Jones 1987). These early finds did not have any economic significance. It was the discovery, in 1870, of coarse gold near Yam Creek, during construction of the Overland Telegraph line, which resulted in a gold rush and several subsequent discoveries, including the deposits at Pine Creek, Union Reefs, Cosmopolitan (Cosmo) Howley, Mount Todd, Fountain Head, Zapopan, Spring Hill, Brocks Creek and Woolwonga. Production commenced from these deposits soon afterwards and continued until 1915, when most mines ceased production. This period also saw a significant influx of Chinese miners into the Territory (Jones 1987). Eluvial mining by Chinese miners continued until 1914 and they were also employed by English companies undertaking lode mining in the area (Jones 1987, Miller *et al* 1998). A total production of 16.86 t of gold was recorded for this period.

Minor production (3500 kg of gold) took place in the PCO from 1916–1968. A significant boost in exploration during the early 1980s, resulting from an increase in gold prices and improvements in gold mining and extraction techniques, led to the opening of the first modern mining venture, the Enterprise (Pine Creek) mine, in October 1985. Several other mining operations, including Brocks Creek, Cosmo Howley, Golden Dyke, Goodall, Moline, Mount Todd, Rustlers Roost, Toms Gully, Union Reefs, and Woolwonga, commenced production shortly afterward. Total gold production from the PCO to March 2010 is recorded as 101.3 t (Ahmad *et al* 2009, Mine Production Records).

The PCO gold deposits, along with deposits in the Tennant and Tanami regions, have been included in a distinctive class of deposits, termed 'orogenic gold deposits' by Goldfarb *et al* (2001). These deposits have common geological, geochemical, stable isotopic, mineralogical and thermochemical characteristics. Gold deposits of this class are recognised in most continents, in both Phanerozoic terranes and older cratonic regions.

On a local scale, Ahmad *et al* (1993) divided the gold deposits of the PCO into six types. These are, in order of abundance:

- Gold-quartz veins, lodes, sheeted veins, stockworks and saddle reefs, eg Union Reefs, Goodall, Mount Todd.
- Gold in iron-rich sediments, eg Golden Dyke, Cosmo Howley.
- Gold in association with platinum group elements, eg Coronation Hill, Gold Ridge, Sargents North.
- Gold in association with uranium, eg Jabiluka, Koongarra.
- Polymetallic gold deposits eg Iron Blow and Mount Bonnie.
- Placer deposits.

Gold-quartz veins, lodes, sheeted veins, stockworks and saddle reefs

These deposits are usually contained within folded, faulted and regionally metamorphosed (usually lower-greenschist facies) turbidite successions (Mount Bonnie and Burrell Creek formations) of the Central Domain. A few are in intermediate to basic intrusive rocks. Post- to late orogenic granites of the Cullen Supersuite have intruded these successions, resulting in superimposed contact metamorphism. Vein quartz is the principal gangue mineral and is rarely accompanied by white mica, chlorite and minor K-feldspar. The width of the gold-bearing quartz veins ranges from a few millimetres to a few metres. Pyrite and arsenopyrite are the main sulfides, together with minor pyrrhotite, chalcopyrite, sphalerite and galena. The base-metal sulfides formed paragenetically later than the gold veins. Late-stage carbonate veining is noted in many deposits. Wall rock alteration effects are negligible and are confined to a few tens of millimetres along vein edges. Within this alteration zone, silicification and sericitisation are the main alteration processes. Gold is fine grained to microscopic in size and is rarely visible. In most deposits, free gold is common in the oxidised ore, but in the primary ore, it is mostly contained as sub-microscopic grains in arsenopyrite and to some extent, in pyrite and pyrrhotite. The grade is generally low, usually less than 3 g/t gold, and individual deposits may have less than 0.5 tonnes to more than 50 tonnes of gold.

The auriferous quartz veins may be concordant or discordant and are usually located within, or close to anticlinal hinge zones**.** Fault-filling and en echelon veins are also present in shear zones. Some occurrences are within stockworks of millimetre-thin quartz veinlets. However, most are in quartz veins of 0.1–2 m thickness.

Most lodes trend northwest or north-northwest and have steep dips or are vertical. The spatial distribution suggests that most deposits are in the vicinity of the Pine Creek Shear Zone. Lodes in the northern extension of the Howley Anticline trend north and in the Zapopan-Fountain Head area, have a west-northwest trend, but in both of these areas, the lodes follow fold axes.

Almost all deposits are within the contact aureole of post-orogenic granites, some distance away from the granite–sedimentary rock contact. None of the deposits is hosted within the granite and the auriferous quartz veins do not continue into the granite. It appears that the northwest-trending shear zones, major lineaments and anticlinal hinges, as well as the contact metamorphic aureole, have played a significant role in the localisation of the gold-bearing quartz veins.

A summary of some significant deposits is provided in **Table 5.5**. The location of gold occurrences within the PCO is given in **Figure 5.23** and some important deposits are discussed in the following section, which relies heavily on Ahmad *et al* (1999, 2009). For smaller occurrences the reader is referred to MODAT.

Brocks Creek goldfield

The deposits of the Brocks Creek goldfield are located along the northwest-trending, east-dipping Brocks

Common name	Lat	Long	Size	Shape	Stratigraphic unit	Resource (Mt)	Resource (g/t Au)	Production (Mt)	Production (g/t Au)	Comments; resources reference
Brocks Ck field						5.1	1.83			
Zapopan (Brocks Creek)	-13.48	131.44	Small	Vein	Gerowie Tuff	0.126	10.2	0.041	20.4	Gillman *et al* (2009)
Alligator	-13.47	131.42	Small	Vein	Gerowie Tuff					Resource included in Zapopan; Gillman *et al* (2009)
Faded Lily	-13.47	131.42	Small	Vein	Gerowie Tuff					as above
North Point	-13.43	131.43	Small	Vein	Gerowie Tuff	0.79	1.5			Includes historical Yam Creek deposit; Gillman *et al* (2009)
Tally Ho	-13.49	131.46	Small	Vein	Gerowie Tuff	0.90	2.61			Gillman *et al* (2009)
Fountain Head	-13.47	131.50	Small	Vein	Burrell Ck Fm	0.652	1.75			Gillman *et al* (2009)
Princess Louise	-13.50	131.55	Small	Vein	Gerowie Tuff	0.679	1.2	0.002	52.3	Gillman *et al* (2009)
Rising Tide	-13.45	131.43	Small	Vein	Koolpin Fm	2.19	1.4			Gillman *et al* (2009)
Howley field										
Big Howley	-13.50	131.34	Small	Vein	Mt Bonnie Fm	0.647	1.6			Gillman *et al* (2009)
Chinese Howley 1, 2, 3	-13.51	131.35	Small	Vein	Gerowie Tuff	5.32	1.4			Gillman *et al* (2009)
Chinese South						4.2	1.5			Gillman *et al* (2009)
Cosmo Howley	-13.54	131.38	Medium	Vein	Koolpin Fm			6.98	2.1	McKibben *et al* (2008)
Cosmo Deep	-13.54	131.38	Medium	Vein	Koolpin Fm	8.743	4.38			Gillman *et al* (2009)
Golden Dyke field										
Davies No 2	-13.56	131.56	Small	Vein	Koolpin Fm			0.050	2.89	Ahmad *et al* (1993)
Golden Dyke	-13.57	131.52	Small	Vein	Koolpin Fm			0.11	7.66	Ahmad *et al* (1993)
Langleys	-13.57	131.55	Small	Vein	Koolpin Fm			0.050	3.9	Ahmad *et al* (1993)
Moline field										
Kendergarden	-13.67	132.15	Small	Vein	Mt Bonnie Fm			0.050	3	Ferenczi and Sweet (2005)
Moline Dam	-13.68	132.15	Small	Vein	Mt Bonnie Fm	0.338	2.4	0.595	2.63	Ferenczi and Sweet (2005)
Northern Hercules	-13.67	132.15	Small	Vein	Mt Bonnie Fm	0.164	4.1	0.507	4.1	Ferenczi and Sweet (2005)
Southern Hercules	-13.67	132.16	Small	Vein	Mt Bonnie Fm			0.227	3	Ferenczi and Sweet (2005)
Tumbling Dice	-13.69	132.15	Small	Vein	Mt Bonnie Fm			0.300	2.6	Ferenczi and Sweet (2005)
Mount Todd field										
Golf & Golf West	-14.13	132.11	Small	Vein	Tollis Fm	*0.121*	*1.22*	0.094	3.6	Poxon and Hein (1994)
Mount Todd	-14.13	132.10	Large	Vein	Tollis Fm	284	0.81	10.643	0.953	Vista Gold (2010)
Quigleys North	-14.11	132.13	Small	Vein	Tollis Fm	*5.5*	*1.73*	0.041	3.07	General Gold (1999)
Quigleys Extended	-14.11	132.13	Small	Vein	Tollis Fm	*Included above*				General Gold (1999
Quigleys South	-14.11	132.12	Small	Vein	Tollis Fm	*Included above*		0.019	6.4	General Gold (1999)
Pine Creek Field										
Coxs	-13.82	131.84	Small	Vein	Burrell Ck Fm	0.53	1.4			Gillman *et al* (2009)
Czarina	-13.83	131.83	Small	Vein	Burrell Ck Fm	1.844	1.69			Gillman *et al* (2009)
South Czarina	-13.83	131.83	Small	Vein	Burrell Ck Fm	0.17	1.5			Gillman *et al* (2009)
Enterprise	-13.83	131.83	Medium	Vein	Mt Bonnie Fm	1.241	2.6	19.4	2.7	McKibben *et al* (2008)
South Enterprise	-13.83	131.83	Small	Vein	Mt Bonnie Fm	1.24	2.6			Gillman *et al* (2009)
North Gandys Hill	-13.82	131.82	Small	Vein	Mt Bonnie Fm	0.163	2.8			Gillman *et al* (2009)
International	-13.82	131.82	Small	Vein	Burrell Ck Fm	1.141	1.7			Gillman *et al* (2009)
Kohinoor	-13.84	131.84	Small	Vein	Burrell Ck Fm	0.26	2.6	0.001	30.7	Gillman *et al* (2009)

Table 5.5. Summary data on some significant PCO gold deposits. Resources in italics are non-JORC-compliant. Abbreviations: Lat = latitude; Long = longitude; Fm = Formation; Dol = Dolostone; Sst = Sandstone; Sltst = Siltstone; Mt Bonnie = Mount Bonnie; Ck = Creek; Unconf = Unconformity (**continued on next page**).

Common name	Lat	Long	Size	Shape	Stratigraphic unit	Resource (Mt)	Resource (g/t Au)	Production (Mt)	Production (g/t Au)	Comments; resources reference
Ungrouped deposits										
Bons Rush	-13.33	131.32	Small	Vein	Zamu Dolerite	0.54	2.5			Gillman *et al* (2009)
Bridge Ck	-13.44	131.32	Small	Vein	Gerowie Tuff	1.038	1.6			Gillman *et al* (2009)
Coronation Hill	-13.59	132.61	Medium	Vein	Coronation Sst	*6.36*	*6.1 Au 0.90 Pd 0.27 Pt*			Mernagh *et al* (1994)
Esmeralda	-13.76	131.83	Small	Vein	Mt Bonnie Fm	1.55	1.8			Gillman *et al* (2009)
Glencoe	-13.44	131.50	Small	Vein	Mt Bonnie Fm	0.704	1.9	0.049	2.3	Australasia Gold (2010)
Goodall	-13.21	131.37	Medium	Vein	Burrell Ck Fm			4.095	1.99	Quick (1994)
Jabiluka 2	-12.50	132.90	Medium	Unconf-related	Cahill Fm	*1.1*	*10.7*			Hancock *et al* (1990)
Koongarra	-12.86	132.84	Small	Unconf-related	Cahill Fm	*1.04*	*3.1*			Snelling (1990)
Kazi	-13.35	131.33	Small	Vein	Gerowie Tuff	0.68	2.91			Gillman *et al* (2009)
Maud Creek 1	-14.44	132.44	Medium	Vein	Tollis Fm	10.36	3.02	0.18	3.5	Gillman *et al* (2009)
Mount Porter	-13.62	131.83	Small	Vein	Koolpin Fm	0.355	3.02			Arafura Resources (2005)
Quest 29	-12.95	131.57	Small	Vein	Koolpin Fm	1.24	2.17			Renison Consolidated (2006)
Rustlers Roost	-12.92	131.49	Medium	Vein	Mt Bonnie Fm	16.60	1.34	4.71	1.05	Valencia (2005)
Spring Hill	-13.61	131.72	Small	Vein	Mt Bonnie Fm	3.6	2.3			Tennant Creek Gold (2005)
Sundance	-13.05	131.05	Small	Irregular	Coomalie Dol			0.02	10.57	Simpson (1994)
Toms Gully	-12.83	131.56	Small	Vein	Wildman Sltst	0.68	11.1	0.419	8.4	Gillman *et al* (2009)
Union Reefs	-13.71	131.78	Medium	Vein	Burrell Ck Fm	0.25	2.7	20.225	1.47	Gillman *et al* (2009)
Western Arm	-13.41	131.30	Small	Vein	Mt Bonnie Fm	1.79	1.4			Gillman *et al* (2009
Woolgni 2	-14.08	131.97	Small	Vein	Burrell Ck Fm	*0.27*	*2.7*			Adelong Consolidated (1999)
Woolwonga	-13.41	131.55	Small	Vein	Mt Bonnie Fm			1.4	2.86	Northern Gold (2004)

Table 5.5. Summary data on some significant PCO gold deposits. Resources in italics are non-JORC-compliant. Abbreviations: Lat = latitude; Long = longitude; Fm = Formation; Dol = Dolostone; Sst = Sandstone; Sltst = Siltstone; Mt Bonnie = Mount Bonnie; Ck = Creek; Unconf = Unconformity (**continued from previous page**).

Creek Anticline, adjacent to the southern margin of the Burnside Granite. The anticline is cut at a low angle by the west-northwest-trending Brocks Creek shear zone. Several deposits are located along this shear zone within steeply dipping quartz-pyrite-arsenopyrite veins, hosted by the Gerowie Tuff and Mount Bonnie Formation. This goldfield includes the Brocks Creek (formerly Zapopan), Homeward Bound, Faded Lily, Alligator, Crocodile, John Bull, Rising Tide, Fountain Head, Princess Louise, Britannia and North Point deposits. The John Bull, Brocks Creek, Faded Lily and Zapopan line of workings extends discontinuously for about 5 km along the southern limb and hinge zone of an east-southeast-trending anticline. Mineralisation is present as concordant veins and stockworks, and rare stratabound replacement mineralisation, and as late vuggy veins and minor breccias (Sener 2004).

Post-1980 exploration included soil and auger geochemical sampling and limited diamond drilling (Dann 1982, Miller *et al* 1998). In 1988–1989, Zapopan NL delineated an indicated resource of 193 000 t at 3.9 g/t Au over a strike length of 200 m and to a depth of 75 m. An additional inferred resource of 200 000 t at 6–7 g/t Au was established along the down-plunge extension (Zapopan 1989). Subsequent exploration in the 1990s was conducted by Solomon Resources NL, Cyprus Gold Australia Corp (Miller *et al* 1998), Acacia Resources Ltd, Harmony Gold Mining Co Ltd, Anglo Gold Ltd, and Northern Gold NL. Mining at these deposits took place from 1996–2001. Total production from the Brocks Creek deposits amounted to 9800 kg of gold. Total resources are estimated at 5.1 Mt averaging 1.83 g/t Au (**Table 5.5**).

In March 2006, the leases were transferred to GBS Gold International Inc (GBS), a Canadian-based mining company. This company consolidated most of the gold deposits of the Central Domain and commenced an extensive drilling campaign and other exploration activities. GBS successfully confirmed the down-plunge continuity of mineralisation at the Zapopan mine and announced a total ore reserve (JORC-compliant) of 0.247 Mt at 13.05 g/t Au. This was later revised to 0.126 Mt at 10.2 g/t Au (Gillman *et al* 2009). Underground mining of the deposit commenced in 2007, but operations ceased in late 2008, when wholly-owned subsidiary GBS Gold Australia Pty Ltd (GBS Gold) went into administration. Total production from the GBS Gold operations to 31 December 2007 was 1.084 Mt of ore at an average head grade of 2.43 g/t Au. A total of 80 092 oz Au was produced since the commencement of commercial production on 1 April 2007 (McKibben *et al* 2008). The leases are currently held by Crocodile Gold Australia Ltd

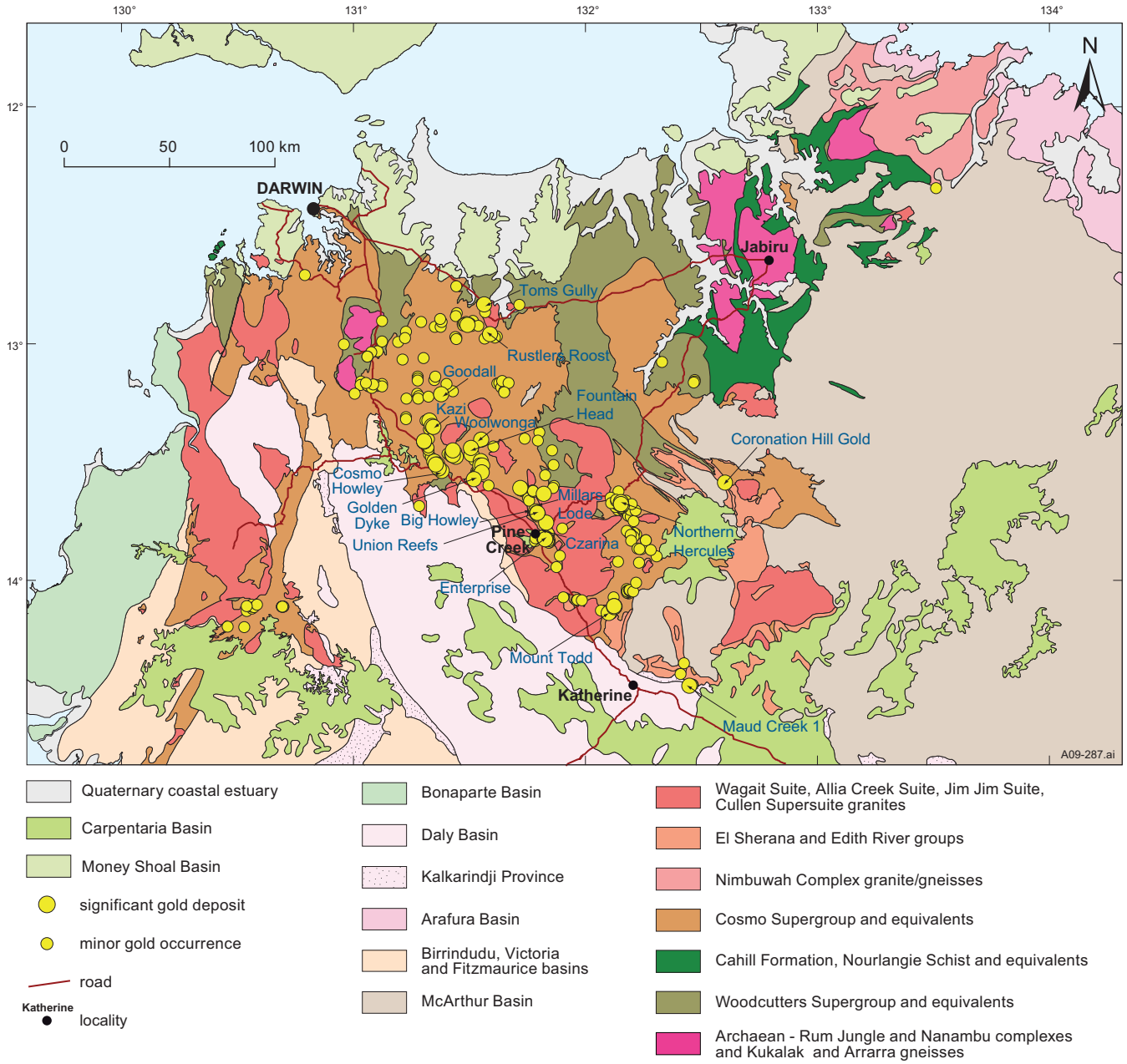


Figure 5.23. Distribution of gold occurrences within the PCO. Data is from MODAT. Significant deposits are labelled. Geology simplified from 1:2.5M geological map of NT (Ahmad and Scrimgeour 2006).



Figure 5.24. Zapopan open cut. Mineralisation is within quartz veins in anticlinal zone, roughly marked with white dashed line.

(Crocodile Gold), who commenced underground mining at Brocks Creek and has plans to produce 23 300 oz in 2010 (Crocodile Gold 2010).

At the *Brocks Creek (Zapopan)* mine, the auriferous quartz veins are in tuff and siltstone of the Gerowie Tuff. The lodes occupy the hinge zone of the tight, west-northwest-striking, 40° east-plunging Zapopan Anticline, which is offset towards the west by a northeast-oriented fault (**Figure 5.24**). The mineralisation is confined and controlled by the intersection of east-southeast-trending, steeply southwest-dipping faults ('slides') and the northern limb and axial hinge region of the anticline (Archibald and Bettenay 1990). The *Faded Lily, Alligator* and some of the *Zapopan* lodes are largely concordant quartz vein sets or stockwork, localised along the Brocks Creek Shear Zone and trending parallel to the Zapopan Anticline. Some stratiform mineralisation is noted at *Zapopan* within pyrite-chlorite-garnet-chert units. The lodes at

Rising Tide are within the Koolpin Formation and appear to be stratabound (Miller *et al* 1998). At *Faded Lily*, mineralisation is localised at the faulted contact between massive black shale and thinly bedded, grey, white and black siltstone and shale units. At *John Bull*, mineralisation occurs in a series of thin (up to 1.5 m-wide), steep, parallel quartz veins, 10–20 m apart. These veins occur off the main Brocks Creek Shear Zone and appear to be related to a conjugate set of quartz veins. At *Brocks Creek, Alligator and Crocodile*, a thin- to medium-bedded sandstone and siltstone succession abuts the siltstone-slate succession. Historically, mining in these localities was focused on the alluvials and on hard-rock mining of individual, rich auriferous veins. Mineralisation at Faded Lily and Alligator is present within quartz veins and along vein margins within graphitic shear zones, and has affinities with arsenopyrite and pyrite. Up to 10% pyrite and 5% arsenopyrite is present in the ore zones. Small visible grains of gold are commonly observed in higher-grade zones at *Faded Lily*, *Alligator* and *Zapopan*. Geochemical analysis of ore samples from the Alligator deposit on average contain about 4300 ppm As, 111 ppm Cu, 134 ppm Pb, 468 ppm Zn and less than 5 ppm Ag. In contrast, those from Faded Lilly have significantly lower values, averaging 1000 ppm As, 36 ppm Cu, 104 ppm Pb and 289 ppm Zn (Miller *et al* 1998). The *North Point* prospect is located approximately 10 km southeast of the Zapopan mine and was discovered in 2002. Mineralisation is within ironstone and fine-grained carbonaceous to tuffaceous rocks of the Mount Bonnie Formation, which have been folded about a north-northeast-trending syncline. The mineralisation occurs within a linear, northerly-oriented, multiple lode system that dips, conformably with bedding, at about 45°W. The system is reasonably continuous along strike for about 300 m and down-dip for 60 m. Drilling by GBS Gold identified two main lodes at *North Point*. The main lode comprises a steeply dipping, continuous zone of mineralisation with an average downhole thickness of 3.4 m. The second lode has similar geometry, with most of the 3.2 m-thick lode located in the northern half of the deposit. Resources (JORC-compliant) were estimated at 0.79 Mt averaging 1.5 g/t Au (Gillman *et al* 2009).

Goodall

The *Goodall* mine is situated about 30 km east of Adelaide River township at the northern extension of the Howley Anticline. It had a pre-mining resource of 4.25 Mt at 2.35 g/t Au. The deposit was mined between 1988 and 1993, producing 7.1 t of gold from 4.095 Mt of ore with an average head grade of 1.99 g/t Au (Quick 1994). The auriferous quartz veins are within the Burrell Creek Formation and are controlled by an upright anticline striking 320° and plunging 30–35° to the northwest. In the main open cut, the host rock comprises a greywacke-dominated succession. Small zones of shale-dominated units are common and a prominent 30–35 m-thick shale-dominated unit can be traced throughout the open cut. Hydrothermal alteration associated with the mineralised zone includes the development of sericite, carbonate, K-feldspar, tourmaline and apatite. Six extremely altered 060°-striking, southeast-dipping pre-mineralised lamprophyre dykes cut across the fold axis. The gold mineralisation is within a north-trending zone of quartz stockwork, consisting of sub-parallel thin quartz-sulfide veins (**Figure 5.25**). The mineralised zone is 750 m long, up to 50 m wide and at least 400 m deep, but grades diminish with depth. It is sub-parallel to the fold and is centred about 60 m east of the anticlinal axis. Quartz veins within the mineralised zone are thin (centimetre to millimetre scale). Both discordant and concordant veins are present. Gold occurs mainly in pyrite and arsenopyrite as 5–10 μm-size blebs. Minor amounts of gold also occur in quartz and chlorite. Other ore minerals include sphalerite, galena, chalcopyrite and pyrrhotite, as well as minor tetrahedrite, tennantite, bismuthinite, native bismuth and gersdorffite. Oxidised minerals include digenite, chalcocite, covellite, bornite, scorodite and wittichenite. Gangue minerals are quartz, chlorite, sericite, carbonates, K-feldspar, tourmaline and apatite.

SHRIMP ^{207}Pb-^{206}Pb geochronology of monazite associated with the gold veins has yielded a mean age of 1751 ± 15 Ma, which is considered to be the age of gold mineralisation (Sener *et al* 2003, Sener 2004). This age is younger than that of the overlying El Sherana/Edith River/Katherine River groups. It is also about 100 Ma younger than the intruding granites. The significance of this age is ambiguous and it is possible that it has been affected by later alteration processes.

Moline goldfield

The *Moline goldfield* contains both gold-quartz vein deposits and iron-rich sediment-hosted deposits (discussed below). It extends over a northwest-trending, elongate area, some 12 km long and 6 km wide, centred about 2.5 km southeast of Moline Dam. The area includes some 30 abandoned gold mines and prospects that together have produced about 4.5 t Au between 1882 and 1992. Most of the production has been from recent mining (3.1 t during 1988–1992) and came from the *Northern Hercules*, *Moline Dam*, *Southern Hercules*, *Kindergarden* and *Highway* pits (Pacific Goldmines 1987, 1988, Stuart-Smith *et al* 1988).

The *Northern Hercules* mine lies within interbedded greywacke, siltstone and carbonaceous phyllite of the Mount Bonnie Formation. Bedding is overturned and dips steeply (60°) to the south. Gold mineralisation is



Figure 5.25. Sheeted auriferous quartz sulfide veins (dark brown) at Goodall Mine. Veins are typically about 5 mm thick.

associated with discordant, steeply dipping, sub-parallel, sulfide-bearing (pyrite ± arsenopyrite, chalcopyrite, sphalerite, galena) vein quartz lodes. Pyrite constitutes 10–30% of the ore. Four Au-bearing quartz lodes with maximum widths of 23 m have been mined within a shear zone up to 80 m wide (Miller 1990). High-grade ore is commonly located in the bends of the shear, suggesting a dilatational jog control mechanism for the gold mineralisation (Moline Management 1991, 1992). Gold is present as 1–80 μm inclusions in arsenopyrite and occasionally, in pyrite and chalcopyrite. Visible gold in pyrite aggregates has been seen in drill cores and erratic repeat assay data indicate that coarse gold is significant (Miller 1990). Narrow hydrothermal alteration haloes adjacent to the auriferous veins, containing quartz, muscovite, K-feldspar, chlorite, sulfides and carbonate, have been identified by Wygralak (1996).

Mineralisation at *Moline Dam* is also hosted in the Mount Bonnie Formation and occurs in the core of an overturned south-plunging anticline. Gold is associated within banded and network-veined pyrite ± pyrrhotite, sphalerite, galena, arsenopyrite and bismuthinite, in laminated chert/banded ironstone and shale. Pyrite forms 10–15% of the ore. Sericite, K-feldspar, tourmaline, carbonate and chlorite are the main alteration minerals. Gold is present as 1–30 μm size inclusions within the arsenopyrite, galena and sphalerite (Miller 1990).

The *Southern Hercules* mineralisation represents an extension of that of Northern Hercules, but the mineralisation style there is similar to that of Moline Dam.

Maud Creek goldfield

The Maud Creek goldfield is located about 20 km east of Katherine. It includes the historic *Maud Creek* mine and the *Gold Creek* deposit, discovered in 1985 and located about 1 km west of the old mine workings. The Maud Creek mine was worked between 1890–1892 and 1932–1940, and produced 16.8 kg of gold (Cottle 1937a, Crohn 1961, Walpole *et al* 1968). The Gold Creek deposit is hosted by quartz-haematite breccia sub-cropping over approximately 1 km strike length in north–south direction. Unlike the lodes at the Maud Creek, which are hosted within a dolerite body, those at the *Gold Creek* deposit are within the Tollis Formation. This formation consists of lithic sandstone and siltstone to the west and mafic tuffs to the east. The latter probably belong to the Dorothy Volcanic Member (Morrison and Treacy 1998).

At the Gold Creek Deposit, mineralisation is associated with intense quartz veining, largely concentrated within a discrete lode at the southern end of a north-trending fault. The latter separates the mafic tuffs to the east from the sedimentary rocks to the west. Fine-grained mafic to intermediate dykes are present in the mine succession, specifically in the immediate footwall to the deposit. The orebody is characterised by stockworks and massive quartz veining, silica flooding, brecciation and intense graphitic and chloritic alteration. Sulfides constitute roughly 5% of the orebody and comprise pyrite and arsenopyrite with gersdorffite. These minerals are present as disseminations and there are also massive intervals containing up to 50% pyrite (Morrison and Treacy 1998). The orebody is 10–20 m thick, 150–200 m long, east dipping and steeply doubly plunging.

In 2000, AngloGold Ltd mined about 180 000 t of oxide ore averaging 3.5 g/t Au and processed it at the Union Reefs mill. The current JORC-compliant Resource is estimated at 10.36 Mt averaging 3.02 g/t Au (Gillman *et al* 2009).

Mount Todd goldfield

Gold was discovered at the Mount Todd goldfield in 1889 and was subsequently mined from three reefs, namely *Tollis, Jones Brothers* and *Quigleys*. These produced some 27 kg of gold from 893 t of ore averaging 30.2 g/t Au (Cottle 1937b). In 1937, the Mines Branch drilled five diamond drillholes at the *Jones Brothers* reef (Hossfeld and Nye 1941). In 1986–1988, Pacific Goldmines NL commenced open-cut mining at several locations (*Alpha, Bravo, Delta, Golf* and three pits in the Quigleys reef), producing 94 000 t ore at 3.6 g/t Au (Poxon and Hein 1994).

In 1988, Billiton Australia Gold Pty Ltd, in joint venture with Zapopan NL, discovered the *Mount Todd* deposit. In 1992, Zapopan acquired Billiton's equity and continued major exploration until 1996, when it was acquired by Pegasus Gold Australia Pty Ltd. Exploration involved geochemical surveys, geological mapping, geophysical surveys and an extensive drilling program, leading to the delineation of the *Batman* deposit (also known as *Yimuyn Manjerr*). This deposit was announced to contain a total resource of 193.55 Mt at 1.03 g/t Au (Pegasus Gold 1995). Vista Gold Corporation is currently carrying out exploration over the leases and has recalculated the resources at 284 Mt averaging 0.81 g/t Au (Vista Gold 2010). Total resource at the nearby Quigleys deposit was estimated at 5.5 Mt averaging 1.73 g/t Au (General Gold Resources 1999).

Mining took place at the Batman pit between 1993 and 2000, and at the Golf and Tollis pits in 1996. To the end of March 2000, when the production ceased, the deposit produced 408 859 oz gold.

Gold mineralisation in the Mount Todd area occurs along a 5 km-long, north-northeast-trending zone within greywacke and siltstone of the Burrell Creek Formation. At the Batman open cut (**Figure 5.26** and **5.27**), this succession consists of interbedded greywacke, shale and minor tuff. Four upward-fining cycles have been identified, each commencing with greywacke that grades upward to more shale-dominated units (Ormsby *et al* 1998). The rocks strike 325° and dip 55° SW. They were weakly to moderately hornfelsed by nearby intrusions of the Yenberrie and Tennysons leucogranites. The dominant metamorphic assemblage consists of cordierite, muscovite, biotite and chlorite. At places (eg Golf pit), the host rock shows a distinctive banding produced by thin laminae of iron oxides. The Batman deposit lies on a northeast-trending magnetic lineament, which possibly reflects a deeper crustal structure related to the Pine Creek Shear Zone (Koerber 1989, Poxon and Hein 1994).

Three sets of faults are recognised in the area: (i) first-order strike-slip northwest-trending shear faults;

(ii) second-order northwest-trending strike-slip faults; and (iii) prevalent third-order northwest- to northeast-trending strike-slip, normal and reverse faults. The last two types are closely associated with the mineralisation.

Two types of veining are present in the area: (i) early auriferous quartz-sulfide veins of crack-seal-type, and (ii) late calcite-base metals veins, composed of dogtooth quartz, calcite, galena, sphalerite, pyrite, arsenopyrite and chalcopyrite.

There are also two types of mineralisation. The first type comprises bedding-parallel stockwork zones of 1–30 mm-thick stringers and veinlets, composed predominantly of iron oxides with varying amounts of quartz. Such mineralisation occurs, for example, at the Batman, Alpha, Bravo and Delta deposits. At the Batman deposit, the zone of stockwork extends over an area 1500 m long, 300 m wide and at least 450 m deep. To the south, mineralisation terminates against strongly hornfelsed sedimentary rocks, and to the north, the mineralised zone appears to be cut off by a northwest-trending fault.

Most gold is within quartz veins and their margins, with only a minor amount extending into the wall rock. The primary mineral assemblage includes pyrrhotite, pyrite, chalcopyrite, arsenopyrite, marcasite, bismuth, bismuthinite, galena, sphalerite, cubanite, talnakhite $Cu_9(Fe,Ni)_8S_{10}$, hedleyite ($Bi_{14}Te_6$) and loellingite. In spite of this complex assemblage, most of the gold occurs as 2–60 µm-size inclusions in vein quartz. Minor amounts of gold are associated with chalcopyrite, bismuth and bismuthinite (Poxon and Hein 1994). Gangue minerals comprise quartz, tourmaline, biotite, muscovite and chlorite.

The second type of mineralisation comprises 0.5–1 m-thick, well defined lodes of massive brecciated iron oxides (eg Jones Brothers reef, Golf). In some places (eg Golf, Quigleys), both types occur together, ie massive lodes are surrounded by zones of ferruginous stockwork.

In the primary zone, massive arsenopyrite with minor pyrite and chalcopyrite are the predominant sulfides, with small amounts of marcasite, pyrrhotite, native bismuth and bismuthinite. Some gold also occurs as free metal.

The mineralised zones are usually located on moderately dipping (45–65°) limbs of folds. In some locations (eg Golf), bedding is almost vertical. At the Batman deposit, ferruginous concordant gold-bearing veins strike 10° and dip 70° east, and are offset by vertical faults striking 300°.

Detailed studies indicate that despite the differences in the two styles of mineralisation, the deposits are similar in vein/lode morphology, tectonic history, alteration assemblages and relative geological timing (Hein 1993a). The differences in mineralogy are attributed to gross metal and sulfide zonation around the Cullen Batholith (Hein 1993a).

Hein (1993b) recognised three deformation events in the Mount Todd area. D_1 is represented by conjugate buck quartz veins and close to tight northeast- to north- to northwest-trending asymmetric folds (F_1). This deformation is interpreted to have preceded the intrusion of the Yenberrie Leucogranite and coincided with peak deformation and metamorphism. D_2 is associated

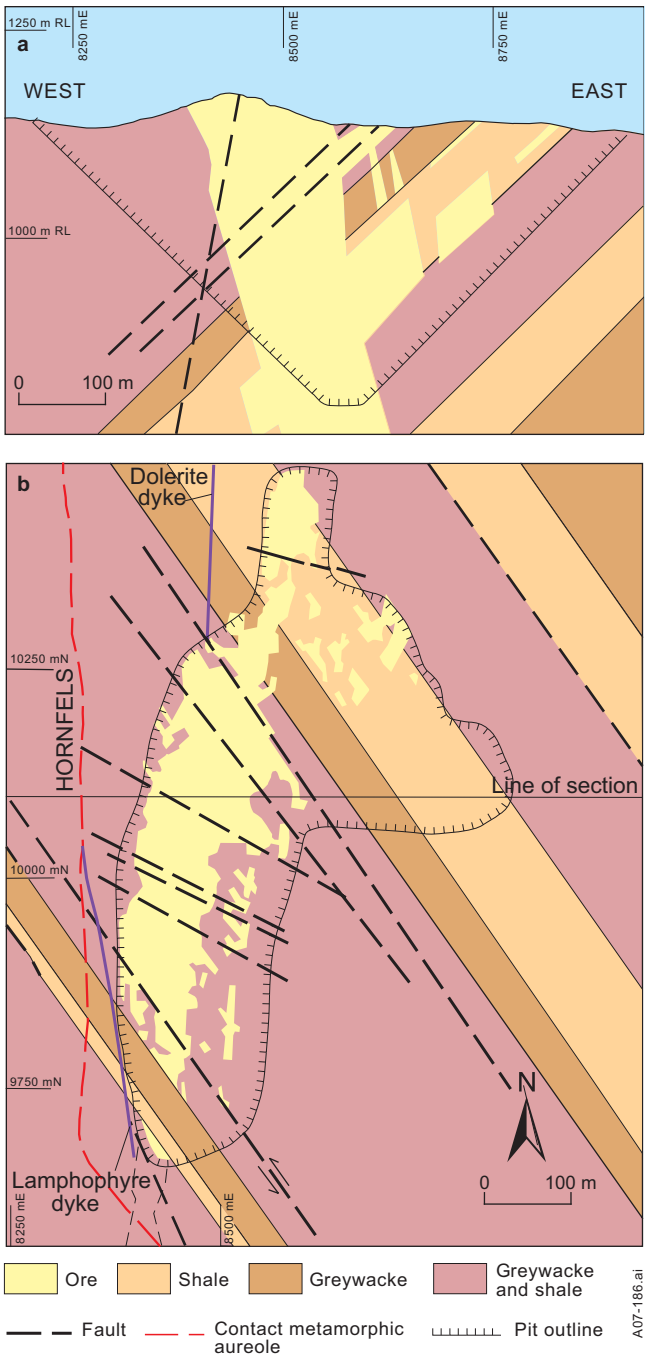


Figure 5.26. Geological section (**a**) and plan (**b**) of Mount Todd (Batman) mine (modified after Ormsby *et al* 1998).



Figure 5.27. General view of the Batman open cut in November 1996.

with westerly-trending open folds and preceded the emplacement of the Tennysons Leucogranite. It is associated with the development of auriferous quartz veins. D_3 is characterised by the reactivation of strike slip faults.

Sener (2004) conducted SHRIMP dating of monazite within the auriferous quartz veins. Approximately two-thirds of the analyses had unacceptably high common Pb and some data were affected by technical problems. After omitting the problematic data, 11 analyses on 4 monazite grains yielded a mean age of 1854 ± 16 Ma with low excess scatter (MSWD = 0.4). These ages are interpreted as representing an earlier generation of monazite that was incorporated into the veins from the wallrock by mechanical processes.

Sener (2004) also provided SHRIMP geochronological data on gold-associated hydrothermal xenotime for the Mount Todd deposit. Twenty-three analyses on seven xenotime grains from three polished thin-sections yielded a mean age of 1812 ± 11 Ma with high excess scatter (MSWD = 3.4). Rejecting eight distinct age outliers (<1800 Ma and >1835 Ma) from the main group left 15 analyses that yielded an age of 1819 ± 8 Ma with low excess scatter (MSWD = 0.4), which is considered a definitive age of gold mineralisation at Mount Todd. This age grouping coincides with the intrusion of granites of the Cullen Supersuite.

Pine Creek goldfield

The Pine Creek goldfield comprises a narrow, about 1 km wide and 6 km long, northwest-trending belt centred about 0.5 km west of Pine Creek township. It includes some 15 hard rock and numerous alluvial workings, which together constitute the most productive goldfield in the PCO (**Figure 5.28**).

The old workings (which are now mostly destroyed by recent mining) comprised numerous open cuts, shafts, adits and alluvial/eluvial diggings. In the larger mines, such as *Enterprise*, *Maid of Erin*, *Ellsinore*, *Kohinoor* and *Eleanor*, underground development reached a depth of about 80 m.

From the discovery of gold in the early 1870s to 1915, 124 960 t of ore were treated, yielding an average of 32 g/t Au from the batteries and 7.8 g/t Au from the cyanide works. Reported total production to 1915 is 2300 kg Au (Hossfeld 1936a). From 1980 untill 1984, Goldfield Exploration Pty Ltd, under a joint venture with the Enterprise Mines NL, carried out exploration including extensive drilling and resource estimation. Open cut operations at the Enterprise mine commenced in October 1985 by a joint venture that was named Pine Creek Goldfield Ltd. Mining of ore from this open cut lasted for a decade until 1995, producing 19.4 t Au from ore averaging 2.7 g/t Au (Ahmad *et al* 1990). During early 2008, some low-grade ore (0.7 g/t Au), treated as waste by the Pine

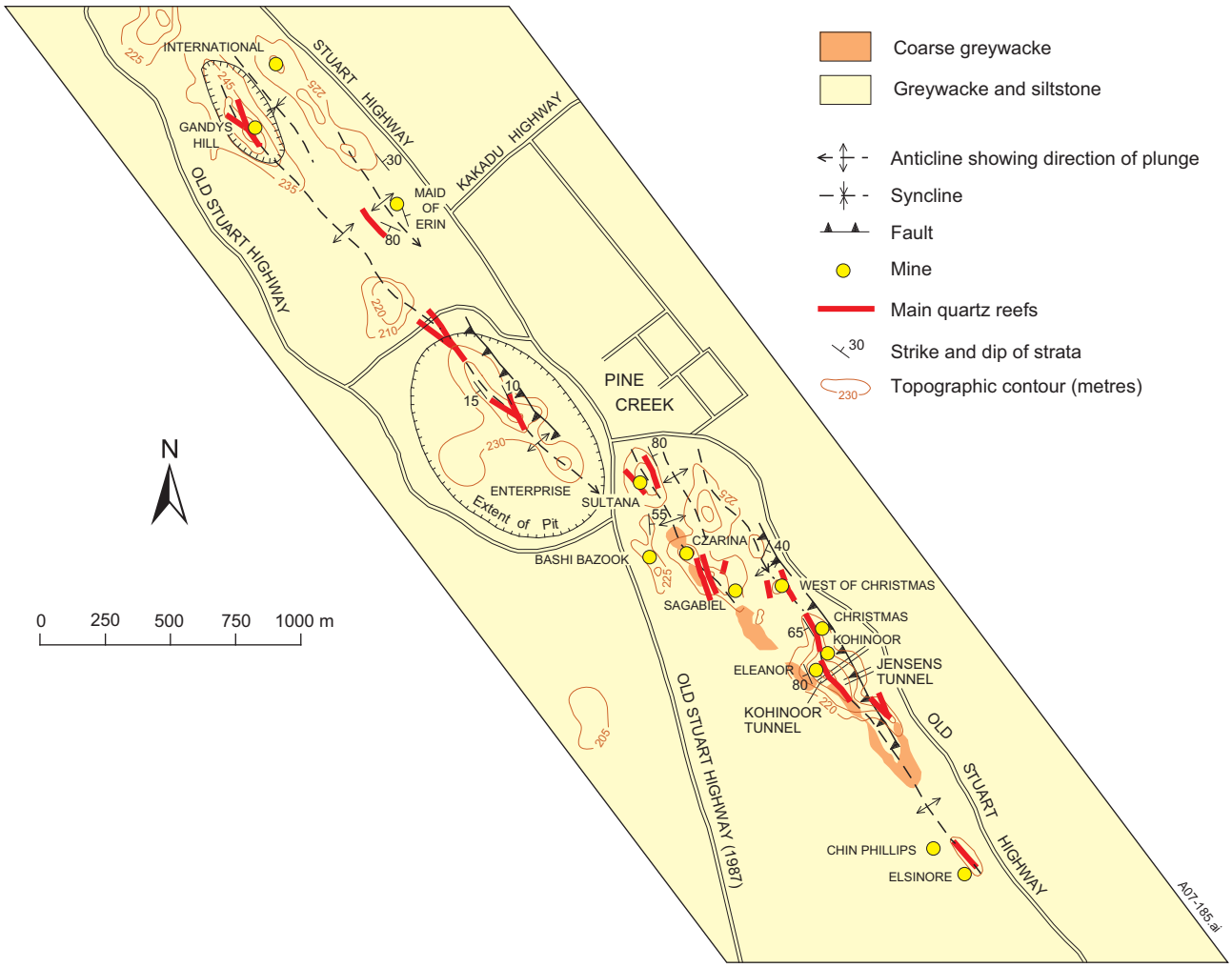


Figure 5.28. Gold occurrences and simplified geology of the Pine Creek goldfield (after Ahmad *et al* 1999).

Creek Goldfield Ltd, was used by GBS Gold to blend with higher-grade ore from the Brocks Creek (Zapopan) mine and was treated at the Union Reefs plant.

GBS Gold recalculated a remaining JORC-compliant resource of 6.59 Mt, with grades ranging from 1.4 to 3 g/t Au. This resource is contained in the *Czarina*, *Coxs*, *South Czarina, Enterprise, South Enterprise, North Gandys, International and Kohinoor* properties. The total resource at the Enterprise deposit is given at 1.24 Mt averaging 2.6 g/t Au (McKibben *et al* 2008). The leases are currently held by Crocodile Gold.

The Pine Creek goldfield lies adjacent to the western margin of the Pine Creek Shear Zone in sheared and contact-metamorphosed shale, siltstone and greywacke of the Mount Bonnie and Burrell Creek formations. The Mount Bonnie Formation, exposed in the northwestern part of the field, consists of pelitic sedimentary rocks interbedded with greywacke and tuffaceous units. A thin carbonaceous shale interval, containing chert nodules, has been used as a marker horizon in defining a moderately tight, southeast-plunging F_3 anticline (Enterprise anticline) at the Enterprise mine. Two other similar anticlines (Czarina and Kohinoor anticlines) are present to the east. The axes of all three structures trend northwest and plunge southeast (**Figure 5.29**).

Mineralisation is predominantly structurally controlled; payable lodes form saddle reefs and less commonly occur as discordant quartz veins, or in faults and shear zones. Minor amounts of gold are also disseminated in the wallrock adjacent to quartz veins. Hossfeld (1936a) listed 16 saddle reefs (eight in the Enterprise anticline, three in the Czarina anticline, five in the Kohinoor anticline) and these were mined extensively. The western limbs of these reefs are usually more persistent laterally and at depth. Old workings have usually followed the contact zone between the reef and wallrock, leaving the central part of the quartz veins untouched.

Mineralisation occupies the transitional zone between the Mount Bonnie and Burrell Creek formations. The wallrock to the gold-quartz lodes comprises greywacke, shale, mudstone and chert. Most of these rocks commonly contain up to 1% disseminated sulfides, mainly pyrite, but also arsenopyrite and pyrrhotite. Four informal lithological units (**Figure 5.29**) have been identified in the mine area (Dann and Delaney 1984, Cannard and Pease 1990). Although the sedimentary rocks have been subjected to deformation, as well as low-grade regional and contact metamorphism (as is shown by the common presence of cordierite porphyroblasts), primary sedimentary structures, including load casts, flame structures and sometimes cross-bedding, are still preserved.

Mineralisation at the Enterprise open cut (**Figure 5.30**) is confined to the Enterprise anticline, which strikes 315º and plunges 5–10º southeast (Cannard and Pease 1990). Several faults exist in the mine area; the largest is the Eastern Fault Zone, which can be traced for 600 m, dips 60–70° southwest and strikes northwest. This fault is displaced by younger north-trending faults, and in the southern part, it is displaced by a near-vertical 060°-trending fault. These younger faults do not carry auriferous quartz veins, but may host some late-stage galena-quartz veins.

The gold is free milling or is contained in arsenopyrite. Sulfide minerals include pyrite, arsenopyrite, pyrrhotite, marcasite, chalcopyrite, galena, sphalerite, bismuthinite, tetrahedrite and covellite. In addition, rare native copper and bismuth are present.

The *Gandys Hill* deposit (also known as *Carlton Hill*) is situated about 2 km to the northwest of Enterprise. A resource of 2.4 Mt averaging 2.5 g/t Au was established by diamond drilling and was subsequently mined by Pine Creek Goldfield Ltd. This deposit occupies a structural setting similar to that of the Enterprise mine and has similar ore mineralogy.

Rustlers Roost

This deposit is situated about 90 km southeast of Darwin and was discovered in 1948 or 1949 by Jim Escreet. The areas of high gold concentration were termed, from north to south, *Sweat Ridge, Dolly Pot, Beef Bucket* and *Backhoe*. It is estimated that 200–250 t of ore was subsequently mined by Jim Escreet from these deposits for the production of about 3.7 kg Au (Rabone 1995). Exploration by a number of other companies was conducted from 1978, and in 1993, Valdora Minerals NL (Valdora) announced a resource of 8.1 Mt of ore at 1.2 g/t Au. Additional exploration by Valdora increased this to a JORC-compliant resource of 34 Mt at 1.17 g/t Au (Rabone 1995). Between 1994 and 1998, Valdora produced about 113 000 oz Au (4.71 Mt at 1.05 g/t Au) from the deposit by heap leaching. The remaining mineral resource, at a cut-off grade of 0.70g/t, amounts to an indicated resource of 15.14 Mt, grading 1.34g/t Au, and an inferred resource of 1.46 Mt grading 1.23g/t Au (Valencia Ventures Inc, press release 14 April 2005).

Gold is present in a stockwork of thin quartz-sulfide stringers and stratiform iron-rich beds within the Mount Bonnie Formation. Dolerite dykes intrude the host succession. Mineralisation is mainly stratabound and occurs within three, spatially separate stratigraphic intervals of fine- to medium-grained, thinly bedded, dolomitic clastic sedimentary rocks. It extends over an area of 1.5 km by 0.5 km and is cut off to the south by a north-trending fault that dips 75° east. The rocks are only weakly metamorphosed to sub-greenschist facies and there are no visible contact metamorphic minerals. Structurally, the deposit is located at the crest of the south-plunging Dolly Pot Anticline. This anticline is asymmetrical with limbs dipping 35° east and 50–70° west. Mineralisation occurs in both limbs.

Several gold-bearing lenses with variable strikes are present within the area. Most gold occurs in dark dolomitic, carbonaceous and pyritic shale. Some gold is also present in flat-lying, discordant quartz veins. The ore assemblage includes gold (grains 1–50 μm in size), pyrite, arsenopyrite, chalcopyrite, marcasite, pyrrhotite and sphalerite. The oxidised ore extends to a depth of about 80 m.

Toms Gully

This deposit is located about 90 km southeast of Darwin and was discovered in 1986, as a result of a stream sediment geochemical survey by Carpentaria Exploration Company Pty Ltd. It was extensively drilled and evaluated in the

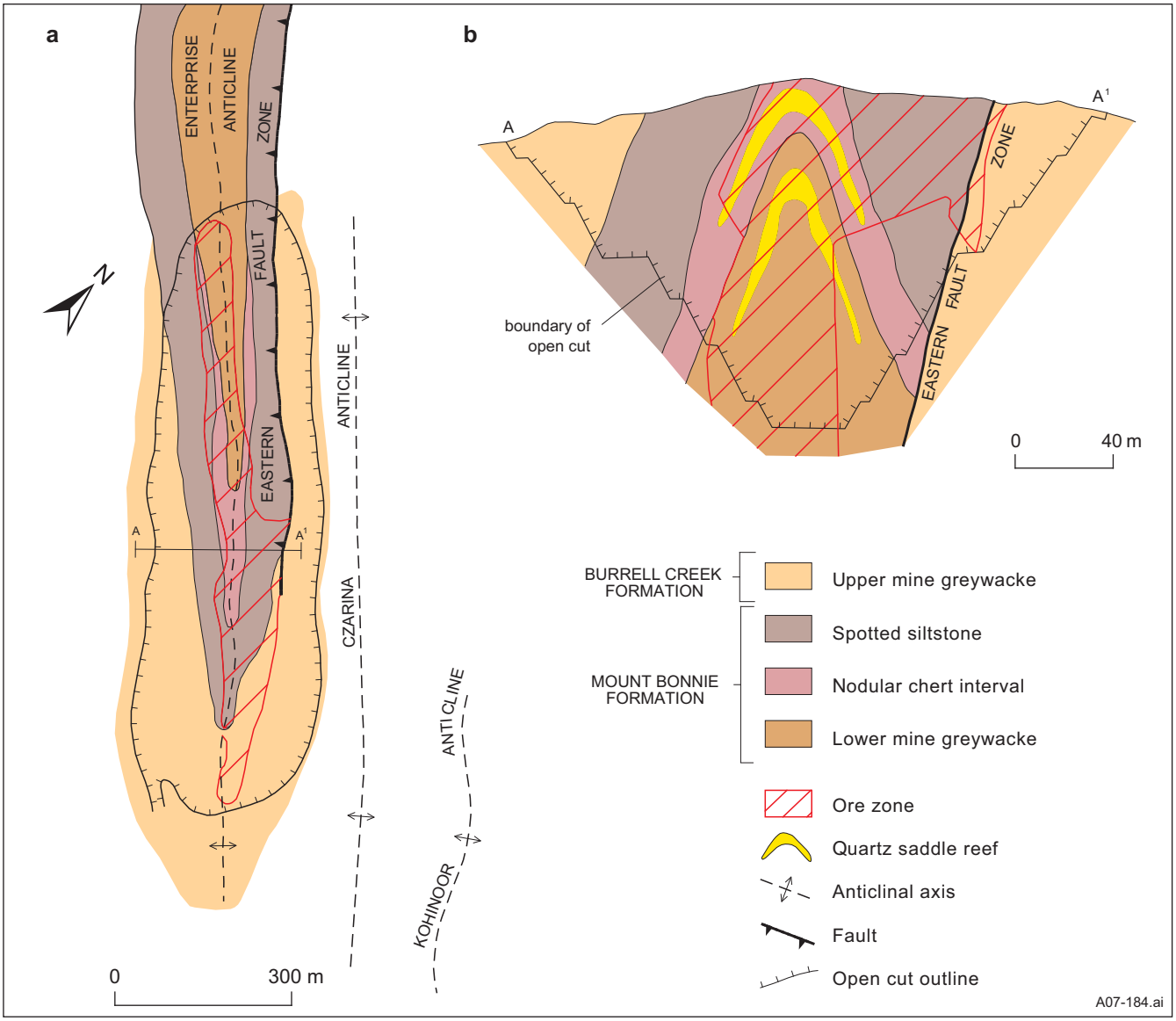


Figure 5.29. Generalised geological (**a**) map and (**b**) cross-section of the Enterprise (Pine Creek) mine (modified after Cannard and Pease 1990).



Figure 5.30. Open cut at Enterprise mine in 1991. Note anticline (dashed lines) through middle of pit.

following year, leading to the establishment of ore reserves of 0.28 Mt averaging 8.4 g/t Au (Simpson 1990, Sheppard 1992). Open-cut mining commenced in 1988 and continued until 1991, for a total production of 75 000 oz Au from 356 651 t of ore averaging 9.23 g/t Au [Intierra[1] records July 2005]. In 2005, Renison Consolidated Mines NL acquired the mine and produced a further 63 000 t of ore at 3.7 g/t Au for 8200 oz Au (MacKenzie 2007). On 8 May 2008, GBS Gold announced to the Toronto Stock Exchange a Canadian NI 43-101-compliant indicated resource of 1.28 Mt at 8.4 g/t, an inferred resource of 0.592 Mt at 7.4 g/t Au and a probable reserve of 0.74 Mt at 7.1 g/t Au. This has been further refined to a resource of 0.68 Mt at 11.1 g/t and a probable reserve of 0.37 Mt at 8.3 g/t by Gillman *et al* (2009).

This orebody comprises a quartz-sulfide vein striking 90° and dipping near the surface 30–40°S, which gradually flattens to sub-horizontal at depth (Tsuda *et al* 1994). The quartz-sulfide vein averages about 1 m in thickness, but the mineralisation continues into the adjacent wall rock giving an average thickness for the orebody of about 2 m. Only the eastern half of the exposed 800 m strike length is of ore grade, but this ore shoot extends southwest down plunge for over 1 km. The quartz vein is poorly banded, showing multiple generations of blue-grey quartz. Mineralisation is restricted to the central and eastern parts of the reef. The fault hosting the mineralised lode has been interpreted as a thrust fault (Sheppard 1992). Patchy sub-economic mineralisation is present in altered siltstone in the hangingwall. A steeply dipping, discontinuous Pb-Zn-Ag-Cu lode, up to a metre thick, overprints gold mineralisation at the eastern end of the lode. The north-northeast and northeast faulting has resulted in metre-scale or decametre-scale displacements of the mineralised vein.

The host rock comprises folded carbonaceous siltstone and mudstone of the Wildman Siltstone that has been regionally metamorphosed to lower greenschist facies. Intrusion of the nearby Mount Bundey Granite has resulted in the development of andalusite-bearing hornfels. Pyrite and arsenopyrite are the main sulfides and are present in the ratio 2:1. Minor loellingite, galena, sphalerite, chalcopyrite and pyrrhotite are also present.

Union Reefs goldfield

The Union Reefs goldfield is about 15 km north of the Pine Creek township and was discovered in 1873. It produced some 1.76 t Au from 0.58 Mt of ore during 1880–1910. The old workings consisted of some 1600 pits, open cuts and shafts concentrated in an area about 5 km long and 450 m wide, in two sub-parallel northwest-trending zones about 200 m apart. The western zone is known as the Union Reefs Line and the eastern one as the Lady Alice Line (Hossfield 1936b, Shields *et al* 1967, Newton *et al* 1998). The Lady Alice Line hosts the *Millers, Ping Ques, Lady Alice* and *Lady Alice North* workings. The Union Reefs Line hosts the *Union South, Crosscourse, Union Central, Prospecting Claim* and *Union North* workings (**Figure 5.31**).

In 1984, Enterprise Gold Mines NL undertook an appraisal of the deposit, including structural and geological interpretations and diamond drilling. This company identified a possible resource of 1.3 Mt averaging 1.72 g/t Au as oxide ore and 0.6 Mt averaging 1.62 g/t Au as sulfide ore in a smaller central part of the field, within a series of discrete and interconnected ore zones that extended over a distance of 250 m and reached a depth of 80 m (Enterprise Gold Mines 1986).

In 1991, the Shell Company of Australia Ltd purchased the Union Reefs tenements and carried out detailed exploration to delineate a resource. Open-cut mining by Acacia Resources Ltd (later AngloGold Ltd) commenced in January 1995 and continued to 2003, with the production of 935 975 oz Au from 20.225 Mt of ore averaging 1.47 g/t Au. The leases have passed through Northern Gold NL (50%) and Harmony Gold Mining Company Ltd (50%) to GBS Gold and are now held by Crocodile Gold. The remaining resource at Union Reefs is given at 0.25 Mt at 2.7 g/t Au (Gillman *et al* 2009). Although no mining has been carried out at Union Reefs goldfield since 2003, the treatment plant was used in 2007–2008 by GBS Gold to process ore from Brocks Creek gold deposits. After a short break, in December 2009, Crocodile Gold re-commenced production from the Union Reefs plant, also treating ore from Brocks Creek.

The lodes lie within the Pine Creek Shear Zone and are hosted by greywacke, slate and minor conglomerate of the Burrell Creek Formation. The rocks are strongly deformed into a series of isoclinal folds overturned to the northeast and in many places, are dislocated by bedding-parallel faults or shears (Turner 1990). The isoclinal folds form an anticlinorium with an average plunge of 25°S. In places, the sedimentary rocks are intruded by dolerite, pegmatite and amphibolite dykes.

Mineralisation is associated with quartz-sulfide veins (**Figure 5.32**), which are present as stockworks, sheeted veins and quartz reefs (Hellsten *et al* 1994). The thick quartz reefs are generally located in sheared, dominantly pelitic wall rocks. Stockwork-style veining is commonly restricted to greywacke. Sheeted veins are commonly present in the thinly interbedded intervals (**Figure 5.33**). The three vein styles may occur separately, overlap or merge. The lodes generally outcrop as en echelon, pinching and swelling, northwest-plunging lenses that are parallel to near-vertical shears trending 010°, 330° and 355°. Steeply plunging saddle reefs are up to 3 m thick. They comprise massive barren quartz in the centre and have mineralised margins (**Figure 5.34**). Most old workings are confined to the oxidised zone, which extends to a depth of about 50 m. The primary minerals consist of pyrite, arsenopyrite, galena, sphalerite, marcasite and pyrrhotite. Gold is present as rare small grains and as sub-microscopic inclusions in sulfides. Some coarse visible gold is also present; this is mainly confined to selvages of larger quartz veins, and is also present as single grains and as clusters of small grains up to 5 mm in diameter in small cracks within quartz-sulfide veins (**Figure 5.35**). A crude mineral zonation is present, whereby sphalerite and galena tends to be marginal to gold ore that is rich in pyrite and arsenopyrite. Gangue minerals comprise quartz, chlorite, calcite, dolomite and albite. In places, relatively small rich

[1] Intierra (full name Intierra Resource Intelligence) is a mining industry database that was formed in 2001 through the merger of three businesses – Intierra Ltd, Minmet Australia, and Mineral Information Maps (http://www.intierra.com/).

shoots with grades of up to 60 g/t Au and 25 g/t Ag over a true width of 2 m have been intersected at 100 m depth (Shields *et al* 1967, Turner 1990).

Sener (2004) carried out SHRIMP U-Pb geochronological studies on xenotime from the Union Reefs deposit. Approximately half of the analyses showed unacceptably high common Pb contents and were discarded. The remaining 10 analyses on five small xenotime grains from three polished thin-section fragments yielded a mean age of 1698 ± 18 Ma with moderate excess scatter (MSWD = 1.5). This significance of this date is uncertain. It may reflect a late alteration event.

Woolwonga

This deposit is situated about 50 km northwest of Pine Creek township and was first worked during 1871–1908 by underground methods. Total production for this period was 205 kg of gold from 7457 t of oxidised ore and a further 26 kg of gold was recovered from the cyanide leaching of 4600 t of tailings (Stuart-Smith 1985). The grade of the oxidised ore averaged 27.1 g/t Au and the primary ore carried about 3 g/t Au (Walpole *et al* 1968).

In 1985, exploration by Dominion Mining Ltd outlined recoverable mining reserves of 2.1 Mt averaging 2.78 g/t Au and a preliminary resource estimate was 5 Mt at 3 g/t Au (Kavanagh and Vooys 1990). Mining commenced in July 1991 and the ore was hauled some 28 km to the Cosmo Howley CIL treatment plant. The mining operation was completed in early 1995 and produced 130 000 oz of gold at an average grade of 2.8 g/t Au from 1.4 Mt of ore (Northern Gold 2004).

Mineralisation is structurally controlled (**Figure 5.36**) and occurs in quartz veins associated with faults, shears and zones of brecciation within a moderately tight anticline, striking 310° and plunging 35–40° to the southeast (**Figure 5.37**). A shear zone striking 330° cuts the anticline; this zone is probably related to the Pine Creek Shear Zone. Quartz veining comprises saddle reefs, sub-vertical veins, stockworks associated with shear zones and sub-vertical

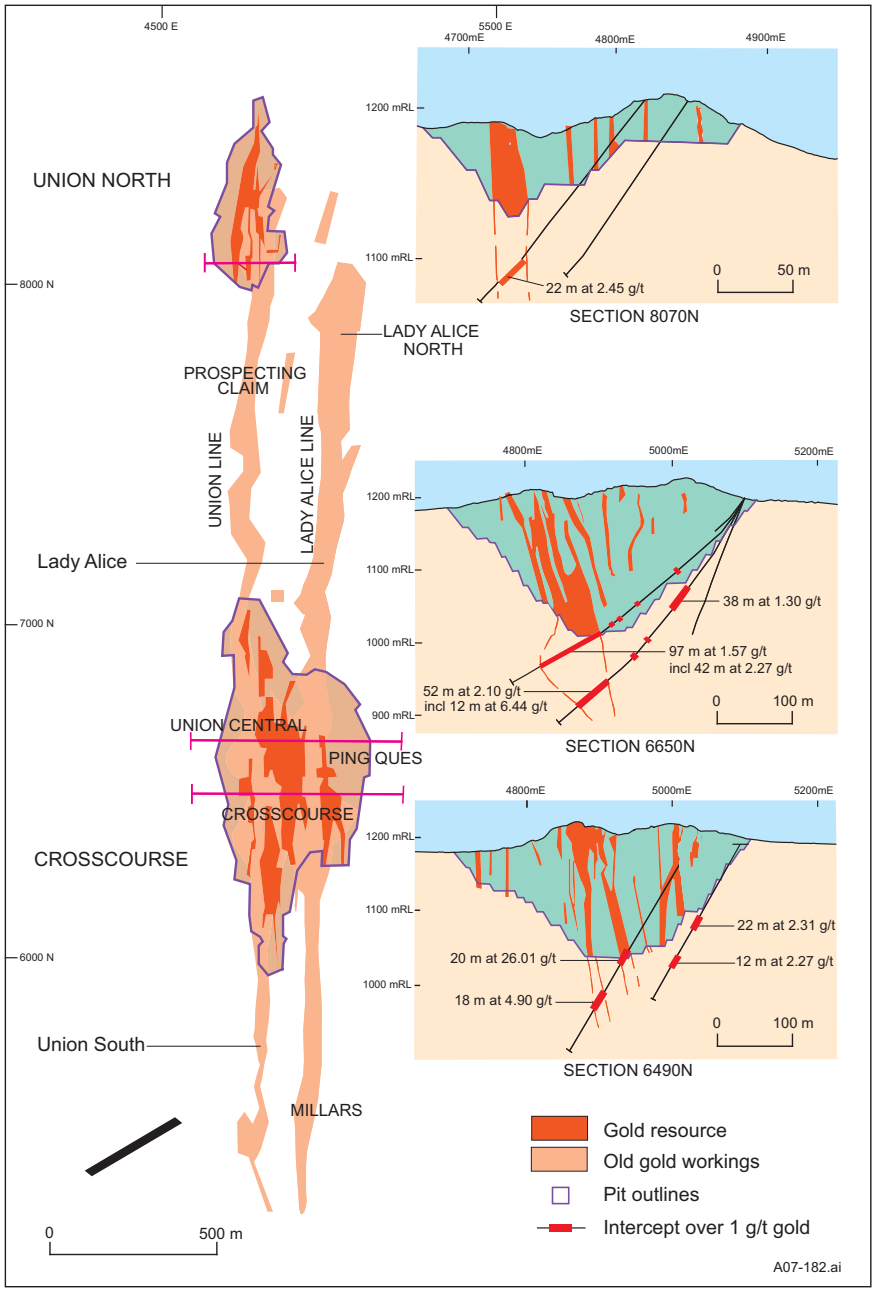


Figure 5.31. Generalised geology of Union Reefs mine (modified after Acacia Resources Annual Report to shareholders for 1996).

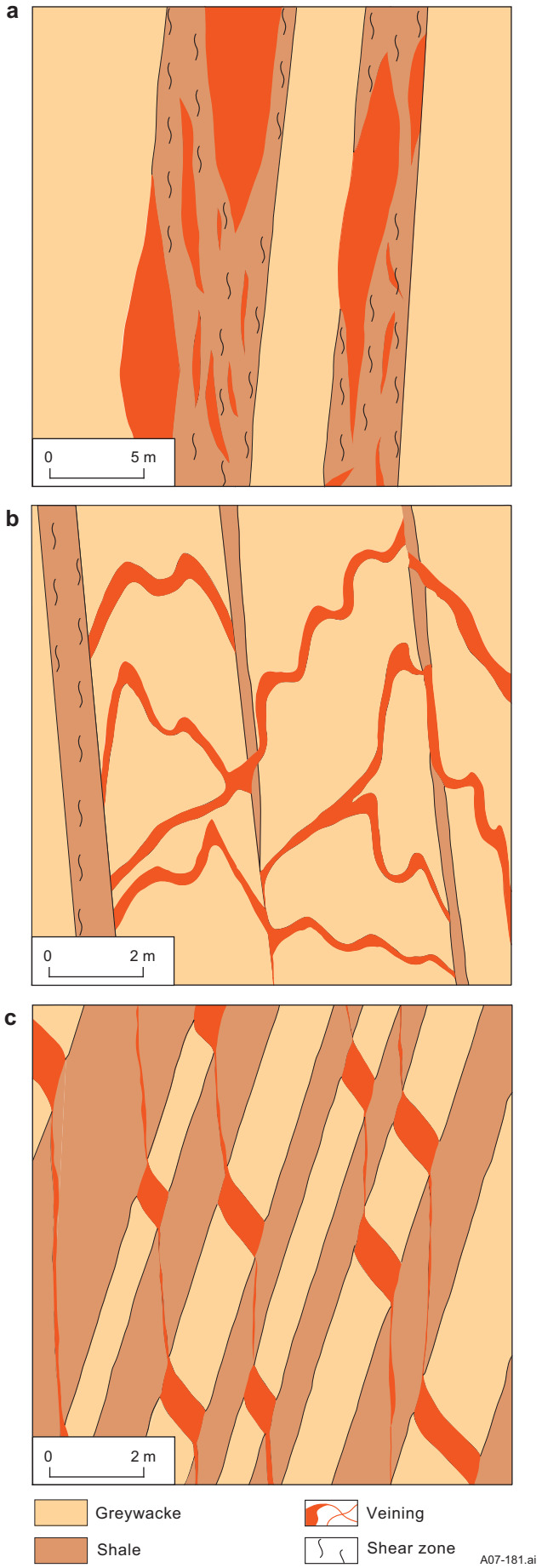


Figure 5.32. Styles of auriferous quartz veins at Union Reefs (modified after Hellsten *et al* 1994). (**a**) Load-style veins within sheared, dominantly shale wall rock. (**b**) Folded stockwork veins. (**c**) Sheared veins typical of Union North mine.



Figure 5.33. Crosscourse lode: near-vertical, thin, bedding-parallel, sheeted auriferous quartz veins in sheared southeast wall of Crosscourse Lode. Width of photograph is about 5 m.



Figure 5.34. Steeply plunging saddle reef at old workings north of Union Reefs mine. Note that old workings follow margins of reef, leaving barren central part untouched.

veins that are parallel to the axial plane of the anticline and to the dominant cleavage.

The richest gold mineralisation occurs at the intersection of the 330°-striking shear zone with the anticlinal axis and at the brecciated contacts of quartz saddle reefs. The predominant ore mineral association is arsenopyrite and pyrite, with subordinate amounts of marcasite, galena, native bismuth, pyrrhotite, chalcopyrite, sphalerite, covellite and chalcocite.

Gold occurs as small particles of free metal in quartz, or as minute blebs in arsenopyrite. Gangue minerals comprise mostly quartz, with minor siderite, K-feldspar and Mg-rich tourmaline (dravite).

The host rocks at this mine consist of tuffaceous greywacke, mudstone and carbonaceous mudstone of the Mount Bonnie Formation, frequently in upward-fining turbiditic successions. The rocks are black in colour, due to the high carbon content.

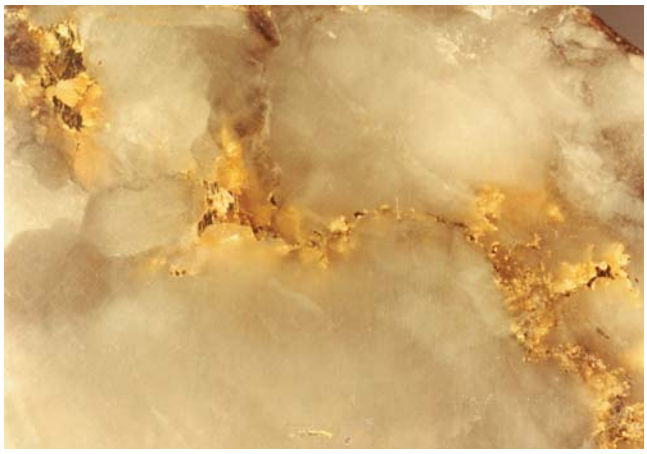


Figure 5.35. Coarse free gold in quartz from Union Reefs mine (length of photograph 15 mm).



Figure 5.37. Woolwonga open cut. Note white bedding-parallel quartz veins and anticlinal structure.

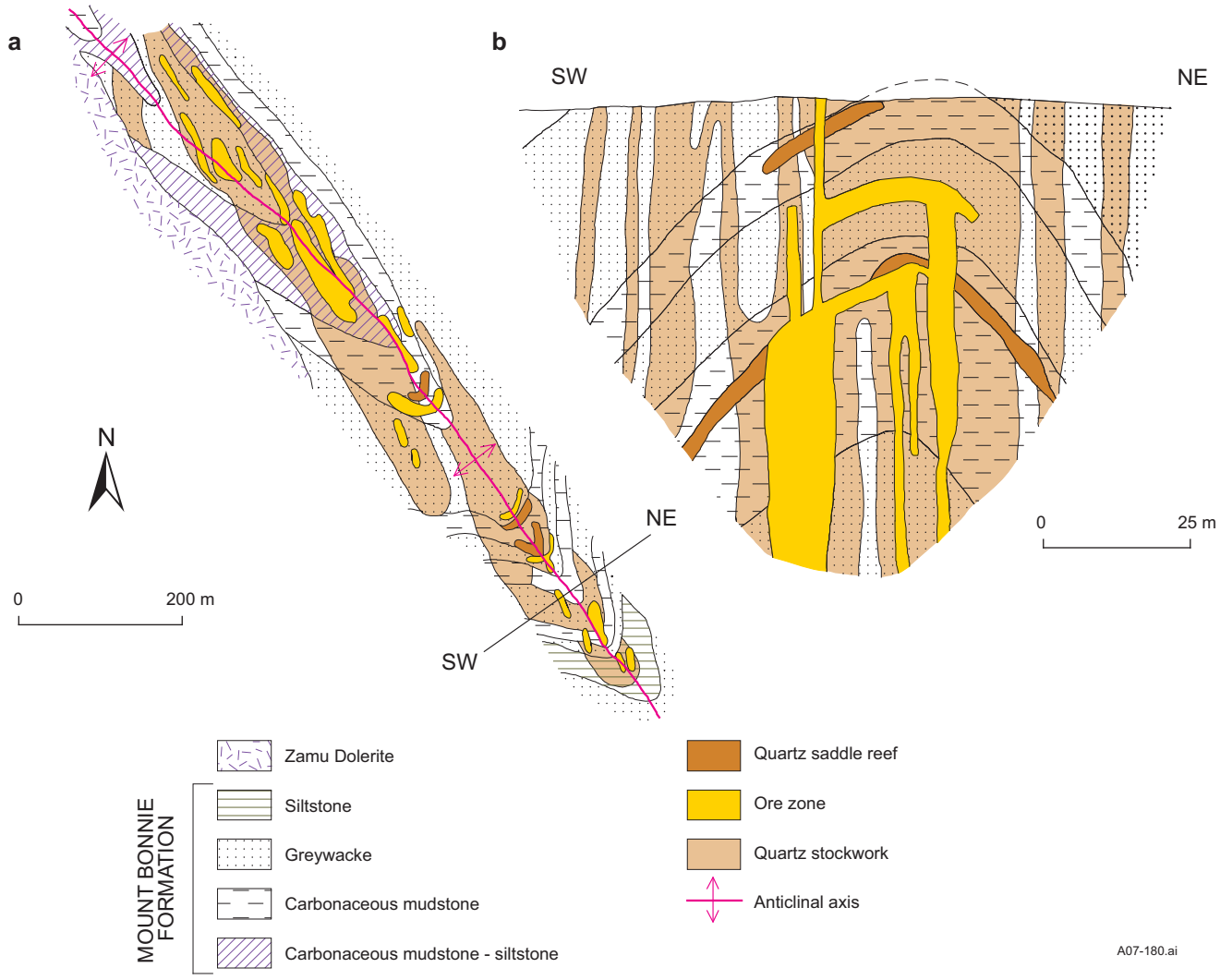


Figure 5.36. Generalised geological (**a**) plan and (**b**) cross-section of Woolwonga mine (modified after Kavenagh and Vooys 1990).

Gold in iron-rich sedimentary rocks

These deposits are associated with iron-rich sedimentary rocks [variously named as banded iron formations (BIF), ironstone, iron formations and silicate iron facies], which are generally capped by carbonaceous shale in the hangingwall. The ore interval is generally bedding concordant and comprises a variety of gangue minerals, including, quartz, siderite, ankerite, calcite, chlorite, cummingtonite-grunerite, ferroactinolite, almandine and tourmaline. Gold-bearing quartz veins are present in some deposits and were considered by Matthäi *et al* (1995a, b) to have provided most of the gold (eg at *Cosmo Howley*).

Sulfides include pyrite, arsenopyrite, marcasite and traces of galena and sphalerite. Gold is sub-microscopic in size and a large part of it generally occurs as native gold; however, a close association with arsenopyrite is also notable. In the oxidised zone, the bulk of the gold is free milling and has been extracted by conventional metallurgical methods.

These deposits were described briefly by Walpole *et al* (1968) and Stuart-Smith (1985), and some detailed studies are also available (Nicholson 1978, 1980, Wilkinson 1982, Alexander *et al* 1990, Nicholson and Eupene 1990, Ahmad *et al* 1993, 1999). A brief summary of significant deposits is given here.

Cosmopolitan Howley

The Cosmopolitan Howley (usually abbreviated to *Cosmo Howley*) mine has been one of the major gold producers in the PCO with an historic production of about 1.05 t of gold during 1879–1915, at an average grade of 22 g/t Au (Hossfeld 1942).

In 1984, Dominion Mining Ltd started a systematic appraisal of the deposit and commenced a new phase of production in 1987. The mining operation involved two open pits to a depth of 120 m. Mining was completed in early 1994, with the operation producing 6.98 Mt of ore averaging 2.1 g/t Au. GBS Gold subsequently conducted further exploration and defined a JORC-compliant resource of 8.7 Mt averaging 4.38 g/t below the present open cut, which is termed *Cosmo Deeps*. Reserves were given as 2.2 Mt at 5 g/t (McKibben *et al* 2008). The leases are currently held by Crocodile Gold, who are planning to commence underground mining, with planned production in mid-2011.

Mineralisation is hosted by the Koolpin Formation, which is divided into three informal members (Alexander *et al* 1990). The 'lower member' is up to 250 m thick and consists of carbonaceous mudstone and siltstone. The 'middle member' is largely composed of interbedded ironstone, mudstone and carbonaceous mudstone, reaching a thickness of up to 100 m. The 50–150 m-thick 'upper member' consists predominantly of carbonaceous mudstone. The Zamu Dolerite locally intruded the mine succession, mainly as sills, prior to folding.

The auriferous 'middle member' of the Koolpin Formation consists of five banded ironstone intervals (named I_1 to I_5) which are 5–20 m thick and separated by four intervals of mudstone (M_1 to M_4). The ironstones comprise greenschist-facies metamorphic assemblages, including chlorite and actinolite, with minor mica, quartz, garnet, graphite and fine-grained pyrite. They contain 10–15% iron (including iron present as sulfides). Common features of the ironstone horizons are recrystallised chert nodules and bands. The ironstone comprises laminae of iron-rich silicate minerals up to 5 mm thick, alternating with sulfide-rich bands. The intervening mudstone horizons comprise chlorite, quartz and muscovite, with minor pyrite and rare garnet. Lenses containing up to 60% Mg-rich tourmaline (dravite) are present towards the base of the 'middle member'. Nicholson (1980) interpreted these lenses as metamorphosed, syngenetically deposited boron-rich sedimentary rocks. Gold mineralisation at Cosmo Howley occurs in all ironstone horizons, but economic grades are confined mostly to the I_5 and M_4 intervals (**Figure 5.38**).

Structurally the mine is located on the major Howley Anticline, which in the mine area, strikes 315° and plunges steeply northwest (50–75°). The anticline has a complex internal structure and is asymmetric, with the east limb steeper than the west. Faulting is intense, especially in the axial plane and on the east limb; fault zones are 1–10 m wide and are sub parallel to the axial plane of the fold.

Most of the gold occurs as sub-microscopic inclusions in arsenopyrite and pyrite. Other sulfides include minor chalcopyrite and pyrrhotite. Free gold is rare. On a broad scale, the gold mineralisation is stratabound and is confined to the ironstone and mudstone intervals. However, it is also spatially related to the discordant and concordant quartz veins, as well as to quartz-filled fractures and faults (Matthäi *et al* 1995a). The quartz veins are usually up to several centimetres thick, but thin veinlets and stockworks are also present. High-grade mineralisation usually occurs in zones immediately adjacent to veins, rather than within the veins themselves. Gold grades range from 3 g/t on the flank of the anticline to 8 g/t in the fold noses to over 20 g/t in the oxidised portions of the veins and breccias (Stuart-Smith 1985).

Several other gold mines occur along the crest of the Howley Anticline. They include *Chinese Howley, Big Howley, Howley North, Bridge Creek* (described separately) and *Mount Paqualin*. Together with *Cosmo Howley* these mines form a 24 km-long line of lodes known as the Howley Line. The mines are located close to the hinge zone of the Howley Anticline in the Koolpin Formation (*Mount Paqualin*), Gerowie Tuff (*Chinese Howley*) or Mount Bonnie Formation (*Big Howley*). In these mines, mineralisation is essentially not associated with ironstone beds. It is largely structurally controlled and is within gold-quartz veins.

Golden Dyke Dome area

The mines in this area lie along the west limb of a doubly plunging, northwest-trending anticline known as the Golden Dyke Dome (Nicholson 1980). The area was first prospected in 1872, when some coarse alluvial gold was discovered. Mining commenced shortly afterwards and by 1936, some 43 kg of gold had been produced from underground mining (Hossfeld 1936c, Walpole *et al* 1968).

Exploration in the 1980s led to open-cut mining at *Golden Dyke, Langleys, Davis 2, Afghan Gully* and *Fisher*

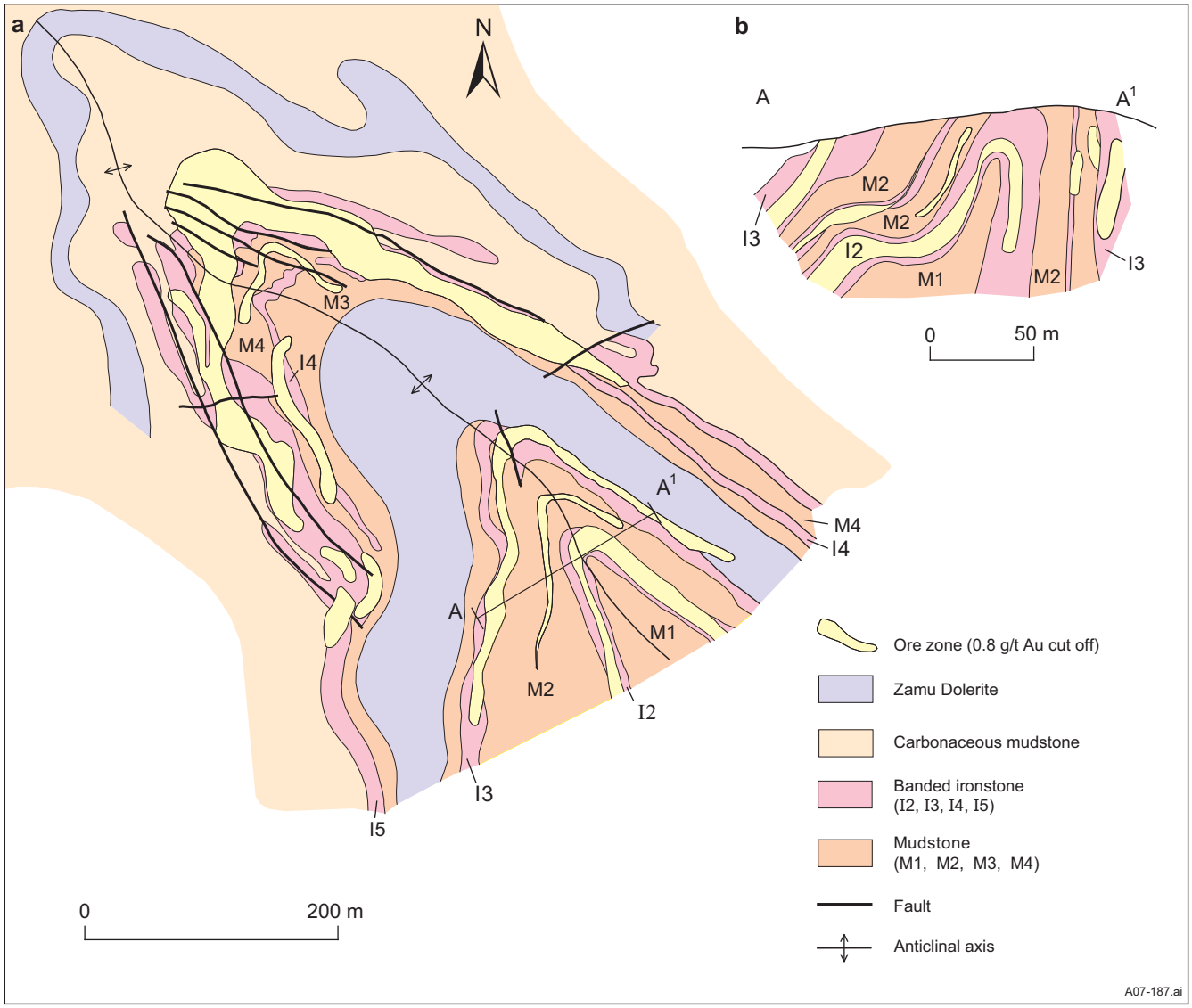


Figure 5.38. Generalised geological (**a**) plan and (**b**) cross-section at Cosmo Howley mine (adapted from Dominion Mining Ltd Annual Report to shareholders for 1998).

Lode. Other old mines in the area, namely *Corbetts* and *Good Shepherd*, were explored, but not mined.

Total post-1980 production from the Golden Dyke, Davis 2, Langleys, Afghan Gully and Fisher Lode open pits is 1.18 t Au. Golden Dyke was the main producer (0.11 Mt ore at an average grade of 7.66 g/t Au) and the open cut reached to a depth of about 150 m, well into the primary ore zone (**Figure 5.39**). The other open cuts remained in the oxidised zone and are 20–30 m deep.

As at the Cosmo Howley mine, mineralisation in the Golden Dyke Dome area is associated with ironstone intervals in the 'middle member' of the Koolpin Formation, but quartz veining is rarely present. Like the Cosmo Howley mine, five ironstone horizons (I_1–I_5) have been identified. The Fisher Lode is within I_5; the Langleys, Golden Dyke and Davis 2 mines are in I_4 and Corbetts prospect is in I_2. The Afghan Gully lode is in a tourmalinite lens, but most mineralisation is in quartz carbonate stringers. The I_3 horizon is known as the Buck Reef and can be traced almost continuously around the Golden Dyke Dome as a series of sugary quartz ridges in ferruginous material. It has been explored by costeans in many places, but gold values are generally low. Texturally, these quartz 'blows' are identical to the smaller chert pods, and Nicholson (1978) considered that the 'blows' were deposited as a relatively thick continuous bed.

At the Golden Dyke mine, the mineralised lode is situated on the western limb of an anticline. The host rocks strike 325° and dip 80° to the southwest; the lode is concordant and follows the ironstone interval. The host rocks are composed of alternating 2–7 mm-thick green ferroactinolite laminae and usually thinner green-brown bands, composed of chlorite, ferroactinolite and stilpnomelane.

The mineralogy of the ironstone units is essentially similar to that in the Cosmo Howley area and Nicholson (1978) has described it in detail. Large (up to 5 mm across) euhedral crystals of arsenopyrite cut across the laminae and are invariably present in the mineralised gold-rich part of the ironstone. Other minerals present in the mineralised zone are pyrite and minor chalcopyrite, pyrrhotite, galena and sphalerite.

Moline Dam

The Moline goldfield contains both gold-quartz vein deposits (discussed above) and iron-rich sediment-

hosted deposits. In this section two occurrences that are associated with iron-rich sediments are described. These deposits were discovered and developed by Cyprus Gold Australia Corporation (later Moline Management Pty Ltd) between 1984 and 1991.

Mineralisation is associated with pyritic chert and pyritic carbonaceous shale beds within the lower Mount Bonnie Formation, along the crests of F_3 folds. The chert beds are commonly laminated and often contain sugary chert nodules and tourmaline (Miller 1990). The gold is closely associated with Fe-As-Zn-Pb sulfides in mineralised zones up to 23 m wide.

The *Moline Dam* orebody is the largest documented example of this class. It was discovered in 1985 and subsequently mined by open-cut methods between 1989 and 1991, producing about 1.5 t of gold from 595 500 t of ore. The following description is largely based on Miller (1990).

Laminated pyritic chert and shale of the Mount Bonnie Formation host the mineralisation which is present in four sub-parallel lodes, individually up to 23 m wide, in a zone with a strike length of 1100 m and a total width of up to 65 m. These lodes comprise banded and network-veined pyrite ± pyrrhotite, sphalerite, galena, arsenopyrite and bismuthinite, within micro- and macro-shear zones produced by bedding plane slip during folding (Moline Management 1992). The wallrock alteration assemblage includes quartz, sericite, K-feldspar, albite, adularia, chlorite tourmaline and carbonate.

Gold occurs as 2–30 µm-sized particles within or between grains of arsenopyrite, sphalerite and galena. This base metals association is also common in the other stratabound Au-sulfide vein deposits (eg *Tumbling Dice, Dingo* and *Simple Dreams*). Moline Dam lodes average about 2700 ppm As, 10 ppm Ag, 1500 ppm Pb and 2400 ppm Zn.

The *Southern Hercules (School)* deposit is located about 200 m southeast of the Northern Hercules pit. Some 227 200 t of ore, averaging about 3 g/t Au, was mined between 1989 and 1990 (Ferenczi and Sweet 2005). The Mount Bonnie Formation at this site occupies the hinge area of an overturned anticline; bedding strikes northwest (305°) and dips steeply (60°) to the southwest. Gold mineralisation is associated with two parallel 330°-trending shear zones that dip steeply (40–65°) to the southwest (Moline Management 1992). Localised thickening of the ore zones occurs where the main shear intersects cross-shears and massive quartz veins to produce plunging elliptical chutes of higher grade (Moline Management 1992).

Gold in association with platinum group elements

Coronation Hill, in the South Alligator River Valley region, is the well known example of this type of deposit. Other Au-PGE prospects include *Gold Ridge, Sargents North* and a number of minor occurrences in the Rum Jungle area south of the Waterhouse Dome and in the western Arnhem Land area. At Coronation Hill, mineralisation is in microfractures, veinlets and disseminations in quartz-feldspar porphyry, volcaniclastic siltstone, debris flow conglomerate, sedimentary breccia and diorite. At Gold Ridge, it is associated with a shear zone within the Fenton Granite. At Sargents North, mineralisation is within haematite-quartz breccia.

Coronation Hill

Au-PGE mineralisation was discovered at Coronation Hill in 1984 by BHP Minerals, and lies beneath and to the east of the old uranium open-cut mine of the same name. Since its discovery, several other Au-Pt-Pd ± U prospects have been identified by both the Coronation Hill Joint Venture Partners and the Bureau of Mineral Resources (BMR) in the South Alligator Valley mineral field.

All of these prospects lie within, or near the Rockhole-Palette Fault System. They are located close to the unconformity between the Coronation Sandstone (El Sherana Group) and pre-Nimbuwah Event basement rocks, and are surrounded by alteration haloes (sericite-chlorite-haematite) that may extend over 1 km (Wyborn *et al* 1990a, b).

The Coronation Hill deposit (**Figures 5.40**, **5.41**, **5.42**) contains a total resource of 6.36 Mt grading 6.1 g/t Au, 0.27 g/t Pt and 0.90 g/t Pd (Mernagh *et al* 1994). It is hosted within quartz-feldspar porphyry, chloritic volcaniclastic rocks, quartz diorite and sedimentary breccia. Wyborn *et al* (1990b), Carville *et al* (1990) and Ahmad *et al* (1993) described the host lithologies as quartz-feldspar porphyry (assigned to the Gerowie Tuff), chloritic volcaniclastic rocks (Shovel Billabong Andesite), quartz diorite (Zamu

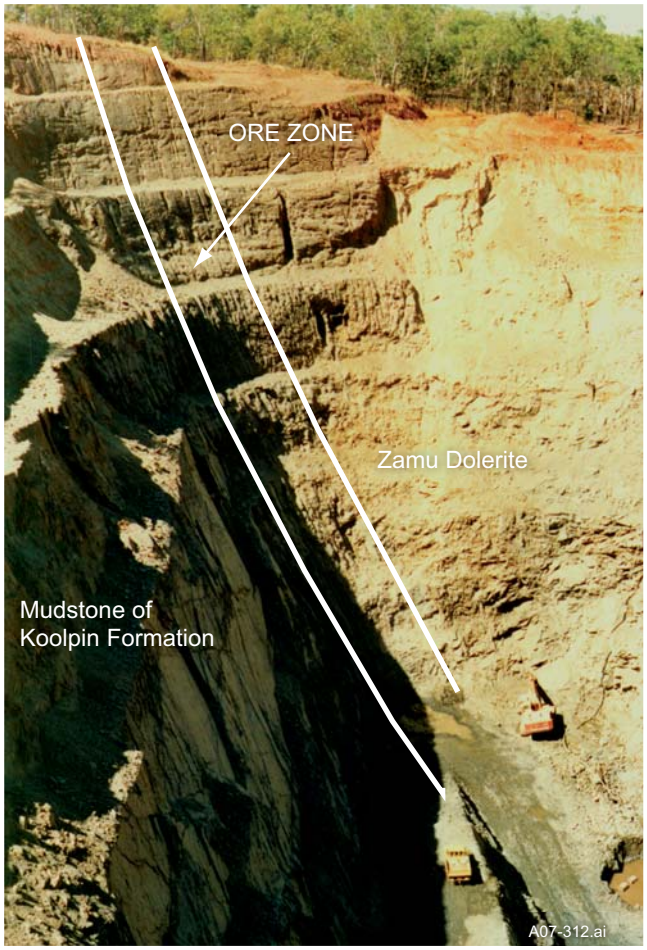


Figure 5.39. Open cut at Golden Dyke mine, looking south.

Dolerite) and sedimentary breccia (Coronation Sandstone). An alternative interpretation (Needham 1988b) is that all of these rock units are part of the Coronation Sandstone that has been intruded by quartz-diorite (Ferenczi and Sweet 2005). Eupene (2003) prepared a 3D model of the deposit based on geological logs. He considered that the lowermost unit intersected in drillholes is the Koolpin Formation; this is represented by a dolostone, which is conformably overlain by chloritised volcaniclastic sedimentary rocks that may possibly represent the Gerowie Tuff. This succession is overlain by a unit, called Type C breccia, which is characterised by clasts of amygdaloidal basic volcanic rocks, quartz feldspar porphyry, quartz diorite and chloritised volcaniclastic sedimentary rocks. The Type C breccia correlates with the breccia at the base of the Coronation Sandstone and is overlain by a sandstone unit (the capping sandstone) belonging to the Katherine River Group. Two types of debris flow breccias have been identified. These have faulted contacts with other rock types and probably postdate the capping sandstone. Type A breccia is a polymictic breccia and contains angular to sub-rounded clasts of porphyry, siltstone, quartz sandstone and volcaniclastic sedimentary rocks. Type B breccia is essentially monomictic and is dominated by clasts of porphyry, with minor black shale, green siltstone and quartz. This succession has been intruded by quartz feldspar porphyry and basic intrusive rocks, which must be older than the Type C breccia, because of the presence of porphyry clasts in the breccia.

The Au-Pt-Pd mineralisation is structurally controlled by the proximity of the Coronation Sandstone–South Alligator Group unconformity and two pre-mineralisation fault systems. The east-trending reverse faults have a vertical movement of 300–400 m and dip 60–70° to the south. West-northwest-trending vertical faults and related shears postdate the east–west reverse faults and have an

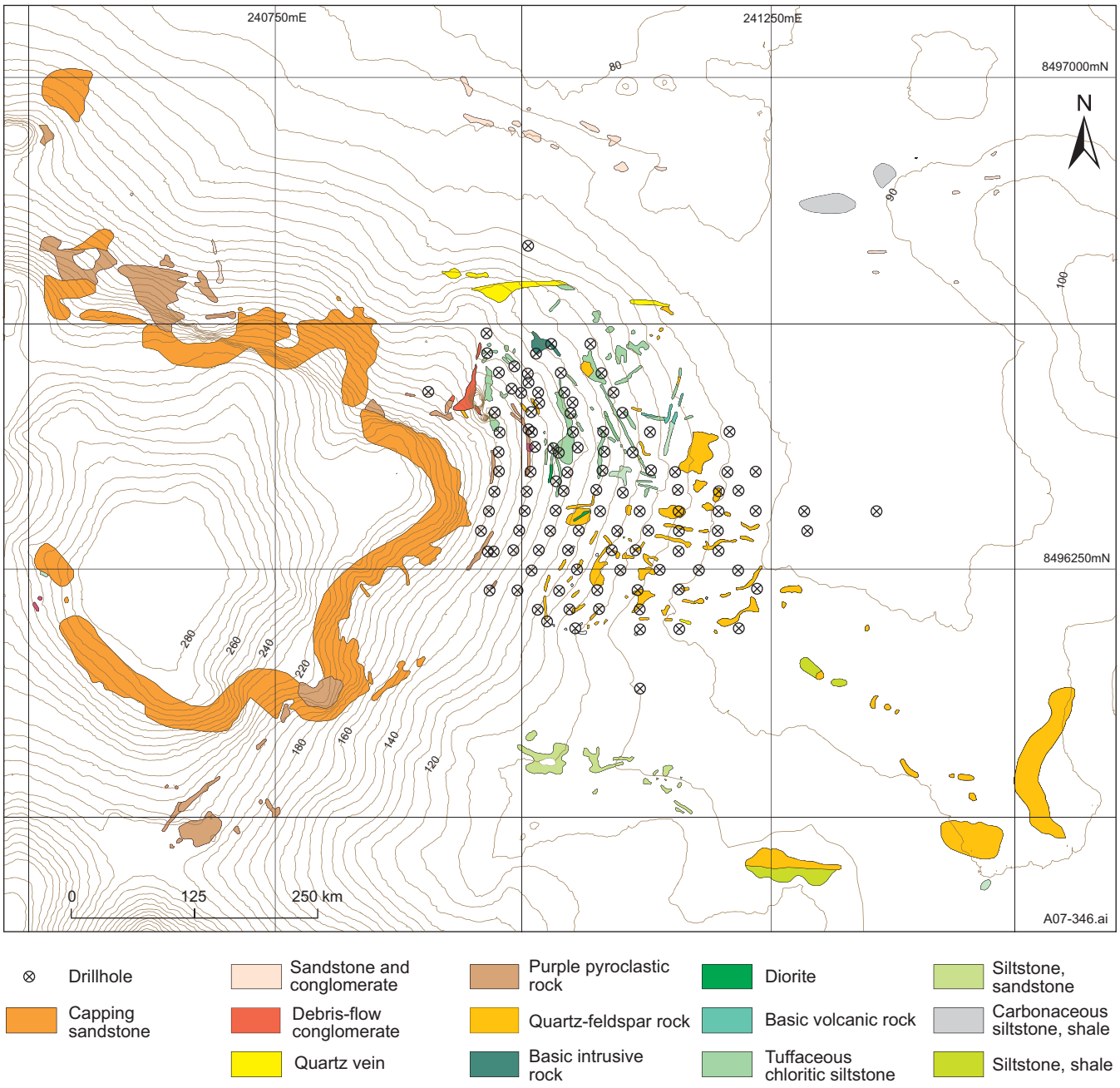


Figure 5.40. Outcrop geology at Coronation Hill (modified after Wyborn *et al* 1990a: plate 11).

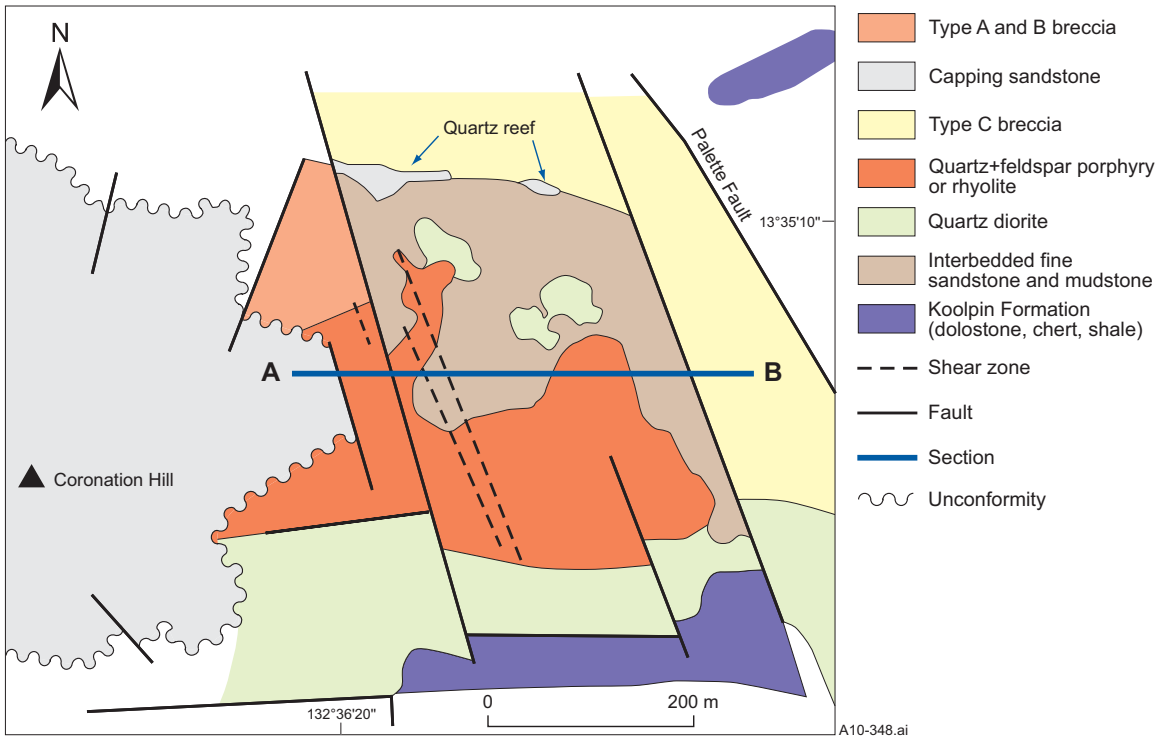


Figure 5.41. Interpreted solid geology of the Coronation Hill area, after Mernagh *et al* (1994) and Orth *et al* (2010). Line A–B denotes cross section shown in **Figure 5.42**.

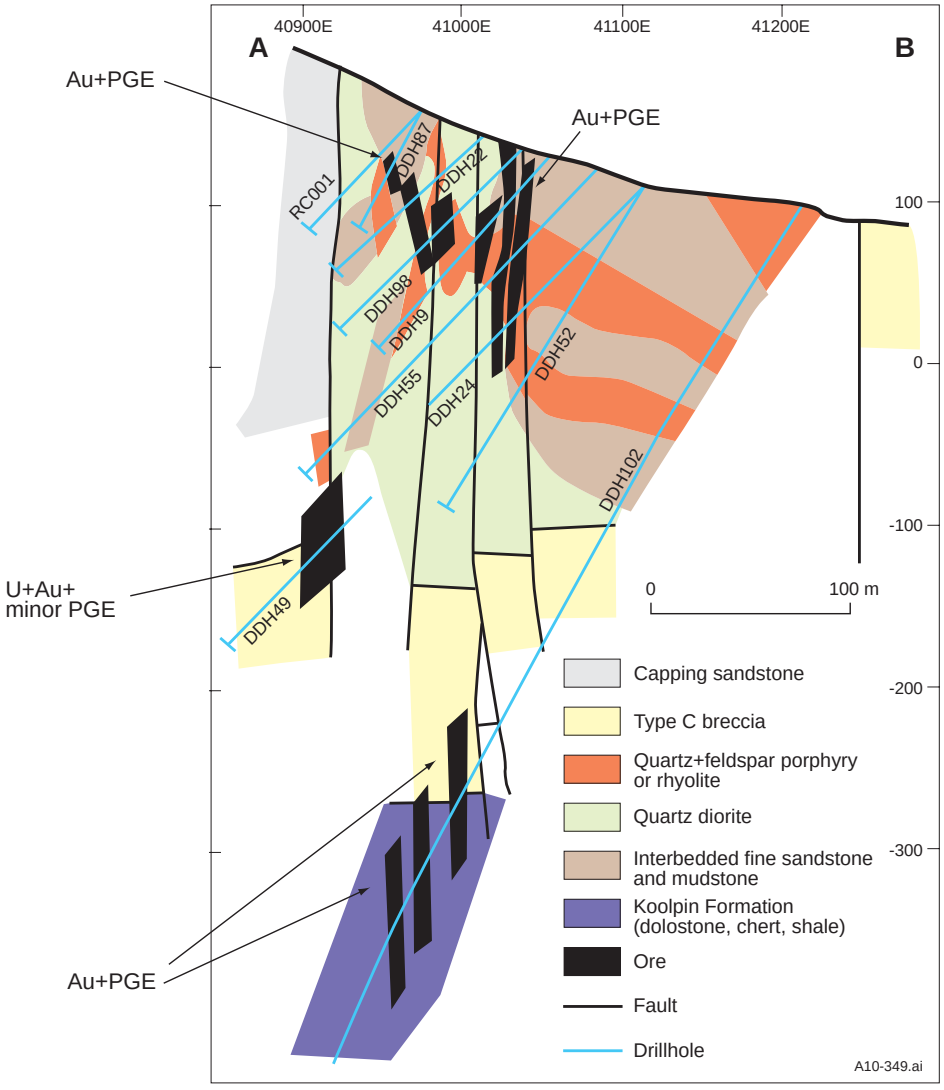


Figure 5.42. Geological cross-section at the Coronation Hill deposit along line A–B, after Mernagh *et al* (1994) and Orth *et al* (2010).

east-block-up sense of movement. Bedding dips steeply (50–60°) to the southwest.

The Coronation Hill deposit has a pipe-like geometry, with a strike length of about 250 m, and is 50–100 m wide. Drillholes have intersected mineralisation at depths in excess of 600 m. Detailed drilling has outlined a mineralised resource, some 240 m in length, up to 80 m wide and to a depth of 250 m. Mineralisation comprises several narrow (average width 10 m) tabular bodies that are sub parallel to north-northwest-trending faults. Mineralisation is present along microfractures, as quartz-carbonate-haematite microveinlets and as disseminations within the alteration matrix of the host rocks.

The gold is microscopic (3.5–332 μm) and is associated with selenium ± replacive pyrite. Sulfides are rare and include pyrite and trace amounts of marcasite, pyrrhotite, sphalerite, chalcopyrite and galena. Precious metals mineralisation is present in approximately the following order of decreasing abundance: gold, stibiopalladinite (Pd_5Sb_2), sudburyite (PdSb), native palladium, an unidentified palladium antimony selenide species, platinum palladium selenide [$(PtPd)Se_2$, possibly a new mineral phase], palladseite ($Pd_{17}Se_{15}$), porpezite (Au Pd phase), Bi Pd Sb phase, Bi Pd Sb selenide, Pt Pd (Fe) alloy phase, Bi Ag selenide, Pd Hg selenide, and very rare Pd As selenide. Other selenide minerals recorded include: clausthalite (PbSe), Ni Co (Fe) selenide, timeannite (HgSe) and very rare Hg Sb Ni selenide (Gilbert 1986, 1987).

Iron and base metals sulfides/arsenides include pyrite/nickeliferous pyrite, marcasite, galena, chalcopyrite, sphalerite, rare Ni Co Fe As sulfide (gersdorffite), rare Ni Cu sulfide, rare bornite, rare pyrrhotite and rare chalcocite. Uranium and rare earth minerals include uraninite, coffinite, rare U Th Al phosphate phase, rare U Th phosphate silicate phase and rare U Th RE oxide, florencite [$(CeLaCa)Al_3\ (PO_4)_2\ (OH)_6$], xenotime ($YPO_4$) and rare monazite [$(CeLa)PO_4$]. Intergrowths of gold/clausthalite/stibiopalladinite are common. Pt mineralisation comprises microveinlets and porous patches of Pt Pd selenide, generally closely associated with florencite crystals and clausthalite. Native Pt has only been recorded once (Gilbert 1986, 1987).

The Au and Pt group mineral phases show a distinct selenide association. Another association is also present where gold/clausthalite/stibiopalladinite are associated with replacive pyrite in some of the altered igneous rocks. These are later than pyrite in the mineral paragenesis.

Mernagh *et al* (1994) identified two styles of mineralisation at Coronation Hill:

- U-Au-PGE ore found at or below the basal El Sherana Group unconformity, in a sub-vertical zone containing sporadic lenses and disseminations. This is generally restricted to conglomerates containing carbonaceous clasts and to chloritic alteration zones in quartz feldspar porphyry. Sometimes, visible gold grains occur in association with disseminated and patchy pitchblende.
- Au-PGE ore, which occurs in irregular quartz-calcite-chlorite veins or breccias in the quartz feldspar porphyry, the quartz diorite and the volcaniclastic rocks. This style is confined to units below the basal El Sherana Group unconformity, but otherwise shows no control by lithology.

Needham and Stuart-Smith (1987) considered the deposit to be of epigenetic sandstone-type. Fluid inclusions, stable isotope studies and chemical modelling indicate that the ore constituents were transported in descending oxidised, low-pH, calcium-rich brines at a temperature of about 140°C. They were precipitated (except U) as a result of a decrease in fO_2 and an increase in pH, due to a reaction with feldspathic rocks. Uranium was precipitated due to the reaction of oxidised descending fluids with reduced methane-bearing fluids (Mernagh *et al* 1994, Mernagh and Wyborn 1994).

Gold Ridge

The *Gold Ridge* prospect is located about 23 km southwest of Hayes Creek Inn on Tipperary pastoral lease. Gold mineralisation was discovered and drilled by Mount Isa Mines Ltd between 1991–1992 (McGeough 1992). A small inferred resource of 32 000 t averaging 4.5 g/t Au, 0.3g/t Pt and 0.5 g/t Pd to a vertical depth of 36 m was estimated. Some mining has taken place via a shallow open cut, but there are no records of production. Some 30 percussion holes have been drilled and 12 costeans excavated in this vicinity.

The prospect lies in a roof pendant of Wildman Siltstone within the Fenton Granite (**Figure 5.43**). On the surface, a northwest-trending (335°), quartz breccia-filled fault zone that dips between 50–55° to the northeast can be discontinuously traced over 500 m. Au-Pt-Pd mineralisation is largely confined to a fine-grained, dark brown to black, graphitic (± pyrite) mylonite unit above the vein quartz breccia zone. Significant Au-Pt-Pd mineralisation is present over a strike length of 150 m and to a vertical depth of 55 m. Diamond drilling carried out by NTGS in 2004 to obtained core samples of the mineralised lode was unsuccessful, due to high hole deviation and jamming of drill rods (Doyle *et al* 2006).

Rum Jungle Area

A number of Au-Pt-Pd prospects were located in the late 1990s in the Coomalie Dolostone, in the area surrounding the Waterhouse Dome. Au-PGE mineralisation has been intersected in drillholes at the *Sargents North*, *Kylie* and *Hardtop* prospects. At *Sargents North* a potential shallow resource of 20 000–30 000 t at 3–4 g/t Au with associated PGE was anticipitated (Williams 1999a). This prospect is located in a 45° east-dipping succession of tremolite chlorite schist assigned to the Coomalie Dolostone. At the Hardtop prospect, the best reported drillhole intersections were 4 m at 2.2 ppm Au, 0.6 ppm Pt and 0.3 ppm Pd (Williams 1999b). At *Kylie*, the best intersection in drillhole KY78/03 was 2 m at 1.29 ppm Au, 0.58 ppm Pt and 0.78 ppm Pd. No primary Pt-Pd minerals have been described from these prospects. Mineralisation is close to the unconformity with the overlying Tolmer Group, and haematite quartz breccia of the Geolsec Formation is developed at the contact. Diamond drilling by NTGS in

2004 at Sargents North intersected the mineralised lode in a highly weathered gossanised zone, which assayed at 360 ppb Au and 36 ppb Pd (Doyle *et al* 2006).

Gold in association with uranium

Gold is associated with uranium deposits in the South Alligator Valley area and at Jabiluka and Koongarra in the Alligator Rivers region. Several other minor occurrences have been located by Cameco Australia Pty Ltd in the late 1990s in western Arnhem Land and these are described in **McArthur Basin**. Although considerable information is available regarding the mineralogy and genesis of the uranium deposits, similar information about the associated gold is lacking.

South Alligator Valley

Thirteen deposits were mined in this area by United Uranium NL between 1955 and 1964, producing some 875 t of U_3O_8 and 359 kg Au as a by product from 146 500 t of uranium ore averaging 0.12–2.5% U_3O_8 (Fisher 1968). Ore was extracted from a number of small stopes to a depth of about 90 m. The mineralisation is contained within a major northwest-trending, dextral strike-slip fault (Rockhole-Palette) system, at or close to the unconformity between the El Sherana and South Alligator groups. Almost all the mineralisation is located in open spaces created by faulting and deformation (Valenta 1991).

The uranium mineralisation consists of stringers, pods or lenses of massive pitchblende or sooty fracture coatings in shoots which vary from 25 mm to 2 m in thickness (Taylor 1968). Other ore minerals include clausthalite (PbSe), eskebornite ($CuFeSe_2$), pyrite, marcasite, chalcocite, native gold and secondary uranium minerals (Ayres and Eadington 1975). Threadgold (1960) examined gold-bearing uranium ores from the *El Sherana*, *Palette* and *Rockhole* mines. Gold is present as minute (average 10 µm) inclusions within pitchblende and partly along shrinkage cracks filled with secondary minerals and as minute inclusions (3–80 µm) within galena-clausthalite veinlets, cutting the pitchblende.

Geochemical and fluid inclusion studies (Ayres and Eadington 1975, Wyborn *et al* 1990a, b) suggest that the uranium and presumably gold was originally leached from the felsic volcanics by highly oxidised groundwaters. These brines descended along near-vertical faults and fractures into the carbonaceous basement rocks. Ore mineral precipitation was by reduction caused by the carbonaceous sediments.

Jabiluka

The *Jabiluka* uranium deposit, discussed in **Uranium**, also contains a resource of 1.1 Mt gold at 10.7 g/t Au Hancock *et al* 1990). Gold mineralisation is confined to the western half of the orebody and is associated with the graphitic intervals, particularly with areas of strong uranium mineralisation. Gold in excess of 500 g/t has been reported from some parts of the orebody. The bulk of the economic gold mineralisation occurs within the main mine succession. Significant gold mineralisation is also intersected within the hangingwall schist, but in this case, it is not associated with major uranium mineralisation. Gold occurs as inclusions (10 to 100 µm in size) or veins in uraninite and is also found associated with lead and nickel tellurides (Hills 1973).

The genesis of this deposit is discussed in **Uranium**, which deals mainly with the transport and deposition of uranium mineralisation. It is likely that gold species formed an integral part of the uranium generation system.

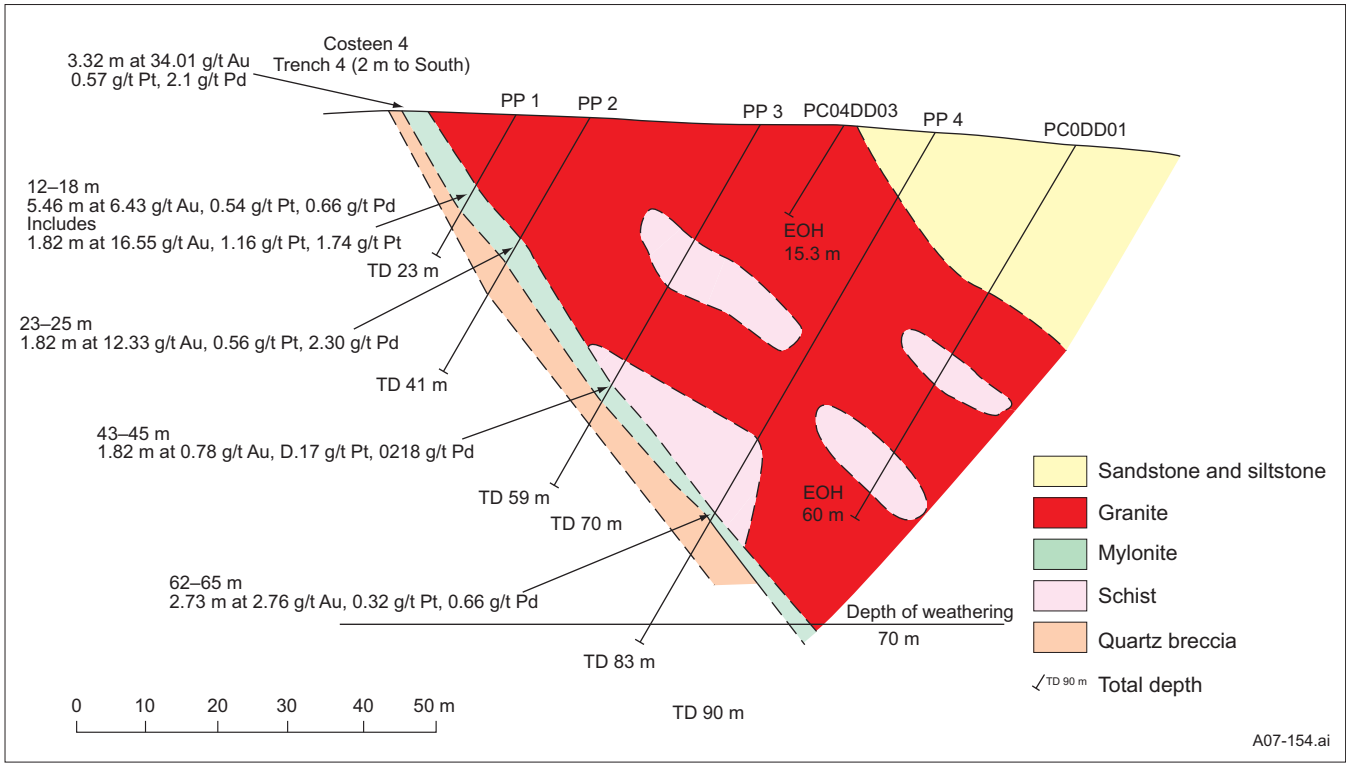


Figure 5.43. Geological cross-section through Gold Ridge deposit, showing significant drill intersections through orebody (modified from McGeough 1992).

Koongarra

The *Koongarra* uranium deposit is situated some 25 km south of Jabiru. It was discovered by Noranda Australia Ltd in 1970 and sold to Denison Mines Australia Ltd in 1980. The current total uranium resource is 0.624 Mt of ore averaging 10.55% U_3O_8 (AREVA 2008). The gold resource is estimated at 1.04 Mt of ore averaging 3 g/t Au (Snelling 1990). Native gold is present in clusters of rounded grains (less than 5 µm in diameter) in uraninite. There are no further details about the nature of the gold mineralisation.

Polymetallic gold deposits

These deposits are contained as concordant lenses within interbedded pyritic shale, dolomitic siltstone and tuff of the Mount Bonnie Formation and, to some extent, they resemble gold deposits hosted in iron-rich sedimentary rocks (see **Gold in iron-rich sedimentary rocks**). In addition to gold, they also contain lead, zinc, copper and silver. Quartz veining is almost non-existent. The common sulfide minerals in the primary ore are sphalerite, galena, arsenopyrite, pyrite, chalcopyrite, pyrrhotite and tetrahedrite. Zinc is the dominant element, followed by silver and gold. The *Iron Blow* and *Mount Bonnie* mines are the only representatives of this type of deposit in the Northern Territory.

Iron Blow

The *Iron Blow* mine is located on the western limb of the Margaret Syncline, which trends and plunges gently to the north. The host rocks are interbedded siltstone, shale (pyritic and carbonaceous), greywacke, hornfels, chert and minor conglomerate and carbonate rocks of the lower Mount Bonnie Formation. Carbonaceous slate is the most common rock type exposed in the open cut, with silicified siltstone, hornfels and greywacke units exposed to the west.

The gossan at the Iron Blow deposit was discovered in 1873 and first worked in 1886, when 100 t of ore was mined (McDonald 1901). During 1898–1906, Northern Territory Goldfields of Australia Ltd produced 15 000 t of oxide and sulfide ore by underground and surface mining methods (Eupene and Nicholson 1990).

In 1975, a Geopeko Ltd and BP Minerals Ltd joint venture carried out a detailed appraisal by additional drilling. They outlined a resource in the sulfide zone of about 0.98 Mt of ore averaging 6.8% Zn, 0.9% Pb, 117 g/t Ag and 2.1 g/t Au (Nicholson and Eupene 1990).

In 1984, the Henry and Walker Group Ltd mined 10 000 t of oxidised ore, averaging 9 g/t Au and 250 g/t Ag; and 25 000 t of sulfide ore, averaging 7 g/t Au and 360 g/t Ag. The estimated inferred resource remaining at Iron Blow is 3.17 Mt, grading 2.08 g/t Au, 100.9 g/t Ag, 0.76% Pb, 3.28% Zn, and 0.19% Cu (Crocodile Gold 2009).

The primary ore at Iron Blow is typically massive and medium to coarse grained, with a distinct, weak, bedding-parallel mineral foliation. Rare laminations of sphalerite in carbonaceous slate have been observed in drill core (DDH S/9: 139.45 m) and this was used as strong evidence for a syngenetic origin by Goulevitch (1980b).

Mount Bonnie

The *Mount Bonnie* mine lies on the eastern limb of the Margaret Syncline, at the same stratigraphic level as Iron Blow. The host rocks are interbedded shale, siltstone, greywacke, hornfels and carbonate rocks with minor pebble breccia of the lower Mount Bonnie Formation. The effects of both regional and superimposed contact metamorphism are evident in the mine succession.

This mine was initially worked in 1902 by Northern Territory Goldfields of Australia Ltd, using underground mining methods, but there are no records of production. Exploration in 1973 by Horizon Exploration Ltd and Jingellic Minerals Pty Ltd outlined a possible 0.48 Mt of ore, averaging 7.67% Zn, 0.4% Cu, 1.8% Pb, 186 g/t Ag and 1.5 g/t Au (Ivanac 1974). About 12 000 t of ore was mined from an open cut and transported to the Mount Wells Battery in 1979 (Rich *et al* 1984).

In 1975, Geopeko Ltd and BP Minerals Ltd carried out a detailed appraisal, including drilling. During 1979–1980, a period of high gold and silver prices, diamond drilling and sampling of old workings outlined reserves of 100 000 t at 8 g/t Au and 230 g/t Ag (Rich *et al* 1984). During 1983–1985, the Henry and Walker Group Ltd mined the oxidised ore yielding a total of 110 000 t, averaging 7 g/t Au and 230 g/t Ag (Nicholson and Eupene 1990). The sulfide zone contains approximately 480 000 t of ore ranging 1–5 g/t Au and 12–15% combined Pb and Zn (Stuart-Smith 1985). Currently available resources are given at 0.65 Mt of ore, averaging 1.7 g/t Au, 280 g/t Ag, 9% Zn, 2% Pb and 0.5% Cu (Northern Gold 2005).

The Mount Bonnie lode dips 40° west and is up to 15 m thick. The original gossan could be traced as discontinuous exposures for over 100 m along strike and consisted mainly of limonite, haematite and limonitic clay, and minor mimetite, duftite, cerussite, coinchalite, malachite, plumbojarosite and scorodite (Rich *et al* 1984). The upper 70 m of the gossan was relatively enriched in gold, silver, lead, bismuth, arsenic, antimony, mercury and tin, with a chalcocite-rich zone near the base (Rich *et al* 1984).

Exposure of the Mount Bonnie lode during open-cut mining revealed that the gossan was formed by the oxidation of two stratabound sulfide lenses. The lode mineralogy and textures are very similar to that of the Iron Blow sulfide lenses; dolomite, chlorite, talc, actinolite and quartz dominate the gangue mineralogy.

Other primary gold deposits

A unique gold deposit, the *Sundance* mine (**Figure 5.44**), is located 2.5 km east of Batchelor. It was discovered by Pancontinental Mining Ltd in 1978, while exploring for uranium (Pancontinental 1986, Simpson 1993, 1994). Analyses of ferruginous quartz breccia boulders ranged up to 55 g/t Au. In 1979–1980, Pancontinental conducted a program of costeaning, mapping, drilling, and ground magnetic and radiometric surveys, which revealed that mineralisation is spread over a large aerial extent, but does not continue below a depth of about 20 m. In 1986, a geological resource of 200 000 t at 4 to 5 g/t Au was estimated by Paterson (1986) for N Byrne and Associates

Pty Ltd. This was refined to a drilling-delineated 14 000 t of proven and indicated ore, averaging 8.16 g/t. Mining in 1986 and in 1993–1994 produced a total of 17 800 t of ore averaging 10.57 g/t. The ore was treated at the Mount Bonnie plant in 1986 and at the Dominion Mining plant at Cosmo Howley in 1993–1994.

Mineralisation is located at the contact between Coomalie Dolostone and overlying Whites Formation. The ore consists mostly of large lumps and clusters of boulders of dark brown, silicified haematite-quartz breccia up to 5 m in size. The ore comprises angular to irregular quartz clasts in a matrix of partly porous silicified iron oxide-rich breccia, including gossanous quartz vein fragments (**Figure 5.45**). Other ore types are sub-gossanous clayey haematite, siliceous breccia with clasts of chalcedonic and flinty silica and light, porous almost pumice-like gossanous rock. Minor pyrite is enclosed by vein quartz and within hard shells of silica and iron oxides. Larger and more deeply buried boulders contain massive pyrite. Free gold can be observed in some gossanous specimens.

A diamond hole drilled through the auriferous gossan averaged 45.2 g/t Au between 7.1 and 9.5 m, and then passed through a zone of massive auriferous pyrite with intervals of talc and magnesite. This pyritic zone assayed 12.1 g/t over a length of 11.5 m. Apparently, the high-grade ore resulted from oxidation and enrichment of an auriferous massive sulfide stock, which is surrounded by a talc alteration halo. Drill core analysis indicated up to 930 ppm Sn in the primary ore and 4850 ppm Sn in the secondary ore (Simpson 1994).

Another deposit named *Sundance East* is located about 5 km northeast of Sundance. Drill intersections of 18 m at 1.88 g/t and 7 m at 2.98 g/t Au have been reported, but follow-up drilling could not define any grade continuity. Mount Grace Resources NL reported a similar drill intersection of 7 m at 8.8 g/t Au within the Coomalie Dolostone at the southeastern corner of the Winchester magnesite deposit at 53L 723420mE 8556500mN (Bruce Uren, Mount Grace Resources, pers comm 2002).

Placer deposits

Many of the hard-rock quartz vein-type gold deposits described above contain associated eluvial/alluvial gold concentrations, which have been mined in most goldfields. The gold is usually fine-grained, generally less than 1 mm in size, but some coarse nuggetty gold has been reported from a few localities. There are few, if any, reported placer gold deposits associated with other deposit types. Major alluvial workings are at *Howley, Fountain Head, Yam Creek, Sandy Creek, Margaret Diggings, Wandie, Union Extended, Pine Creek (Silver Coin)* and in the *Union Reefs* area.

Genetic constraints

Considerable information is available on the genesis of vein quartz-hosted gold deposits within the PCO, including a wealth of data on fluid inclusions, stable isotopes, wall rock alteration and paragenesis (Sheppard 1992, Ahmad *et al* 1993, Wygralak 1996, Matthäi *et al* 1995a, b, Sener 2004). A summary of some important genetic parameters is provided below.

Wall rock alteration

Wall rock alteration effects in vein quartz-hosted gold deposits are minor and are confined to a few tens of millimetres wide fringe of altered rock along the edges of veins. The main wall rock alteration minerals are sericite,

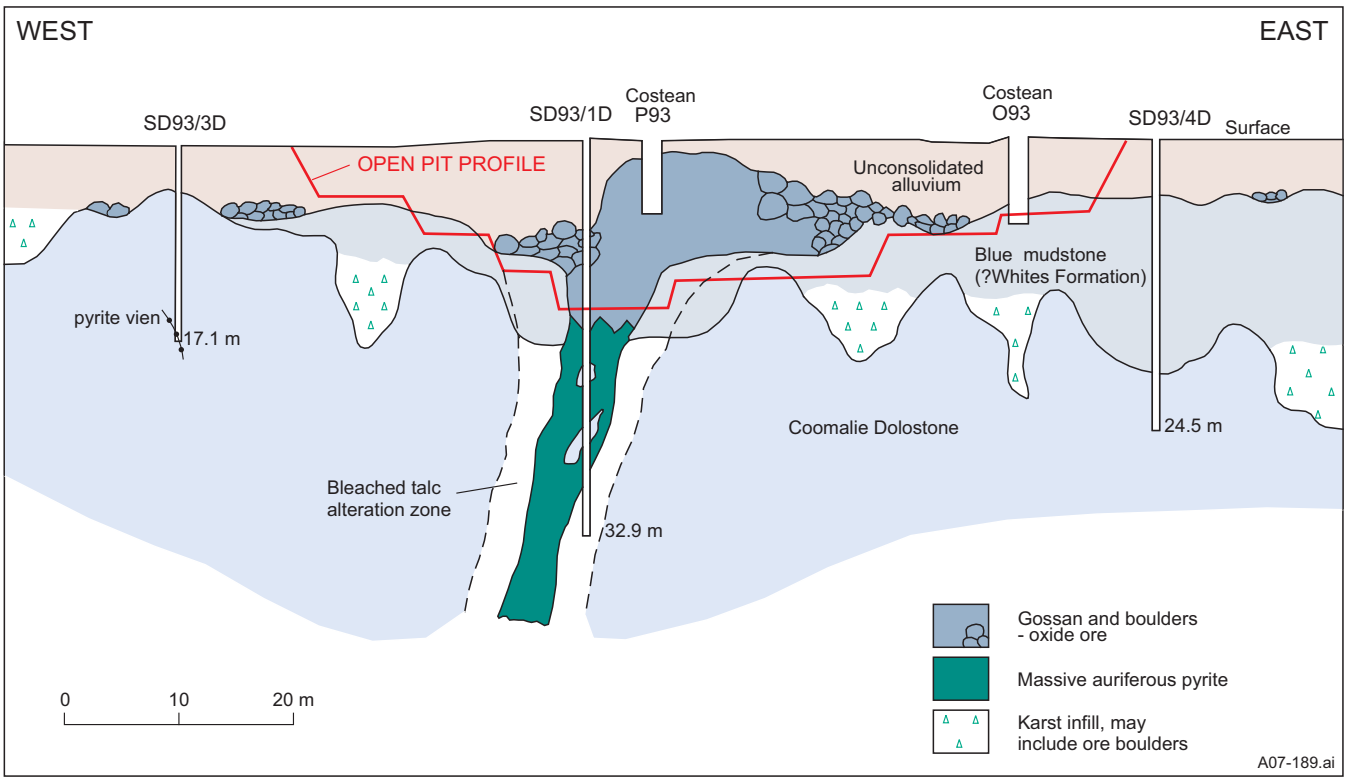


Figure 5.44. Generalised geological section through Sundance mine (modified from Simpson 1994a).

quartz, chlorite, minor carbonate and pyrite. A study of the veining sequence (Ahmad *et al* 1993) indicates that at least four stages of veins are present in most deposits (**Figure 5.46**). Early gold-bearing (Stage 1) veins are cut by later quartz veins (Stage 2), which are mostly barren, but often contain base metals sulfides. Stage 3 quartz veins are generally devoid of any mineralisation and usually form euhedral crystals in vuggy cavities. In some deposits, late carbonate veins (Stage 4) cut all stages of quartz veins.

In the banded ironstone-hosted deposits, chlorite is the dominant alteration mineral and much of it formed at the expense of amphibole. Early, possibly retrograde chlorite is light green, weakly pleochroic, occurs as fibrous aggregates and is often veined by a darker green chlorite, composed of radiating fibres. The late chlorite is probably of hydrothermal origin. Sericite is the next most common alteration mineral. It seems to have formed later than early chlorite (Ahmad *et al* 1993). The Mount Todd and Toms Gully deposits do not have large alteration haloes, suggesting that mineralisation in these deposits was introduced while the rocks were in thermal and chemical equilibrium with the hydrothermal fluids (Sener 2004). The only significant deposit that does not match the alteration types discussed above is Maud Creek. This deposit has sericite-carbonate-fuchsite-graphite alteration within the gold orebody (Morrison and Treacy 1998).

In stratabound polymetallic deposits, the predominant alteration processes are silicification, chloritisation and sericitisation. The most evident form of alteration is silicification of carbonate beds and this includes a quartz-carbonate-talc-tremolite-actinolite assemblage within and adjacent to the massive sulfide ore zones. Chloritisation of the wallrocks is often both pervasive and intense, producing a massive fine-grained chlorite rock and chlorite veins adjacent to the massive sulfide zones. Other effects include partial and selective alteration of primary minerals, such as the replacement of carbonates, biotite and pyroxenes by chlorite. Sericitic alteration is represented by the presence of randomly oriented flakes of phlogopite and muscovite, which often form spots and bands in altered wall rocks. Minor amounts of microcline are also often present.

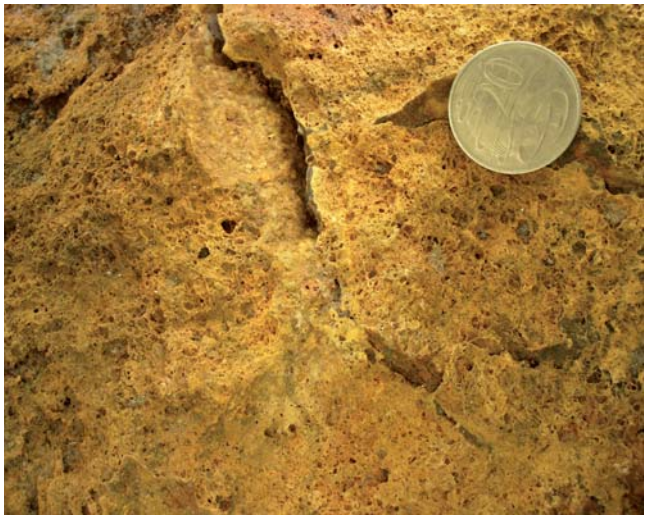


Figure 5.45. Gossanous ore at Sundance mine.

Fluid inclusions

Ahmad *et al* (1993) and Wygralak (1996) have carried out regional studies of a number gold deposits from the PCO. Sheppard (1992) studied fluid inclusions at the Toms Gully deposit, and Ho (1993, 1994) and Ho and Shepherd (1993) studied the characteristics of fluid inclusions at the West Arm deposit. The following summary is based on these reports.

Fluid inclusion studies on Au-quartz veins, Sn-quartz veins, Sn-sulfide veins, Sn-Ta-Pegmatites, Pb-Zn-Ag veins and Au-quartz veins in stratiform Au deposits show four types (**Figure 5.47**) of inclusions: $H_2O + CO_2 + CH_4$-bearing low-salinity Type A inclusions, H_2O + ca 20 volume% vapour-bearing Type B inclusions, H_2O + <15 volume% vapour-bearing Type C inclusions of variable salinities, and H_2O + NaCl + vapour-bearing Type D inclusions. Inclusion types C and D are paragenetically later than and are not related to the main ore-forming stages (Ahmad *et al* 1993). However, for the *West Arm* deposit, these high-salinity low-temperature low-to-nil CO_2-bearing inclusions are considered to be contemporaneous with the gold mineralisation event and are suggested to have evolved from the types A and B fluids at 240–320°C and ca 1 kbar (Ho 1993, 1994, Sener 2004).

Filling temperatures on these inclusions are regionally consistent and there is no significant inter- or intra-deposit variation. Type A inclusions were trapped on the H_2O-CO_2 solvus, mostly within the 250–350°C mode. Pressure-corrected (600–1600 bars) temperatures on Type B inclusions also fall within this narrow range. The salinity modes on Type A inclusions are mostly within a narrow range of 3–5%. Type B inclusions have higher salinities that vary between 1–27% and average about 10%. The P-T and salinity range of these deposits is comparable with the mesothermal deposits.

Type A and B inclusions occur together, but whereas the latter are devoid of CO_2 and CH_4, in the former, these species are abundant. The studied mineral occurrences are spatially closely associated with granite intrusive rocks and a genetic link is likely. Hydrothermal fluids of predominantly granitic parentage (eg porphyry copper deposits) are essentially high-salinity Na-K-Cl brines and, although minor CO_2 is

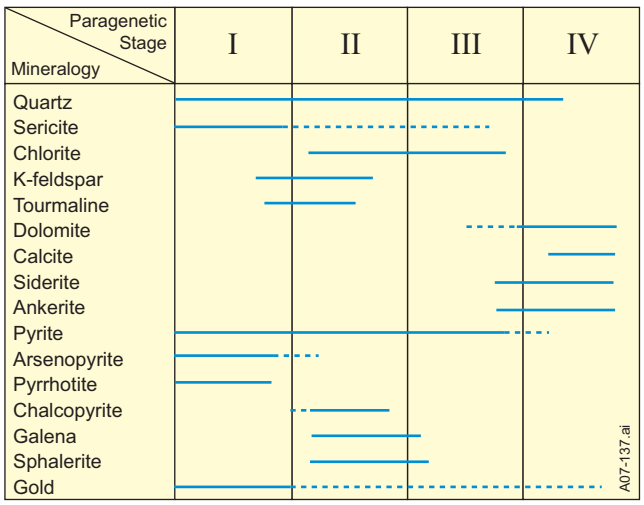


Figure 5.46. Generalised regional paragenetic sequence for PCO gold deposits (after Ahmad *et al* 1993).

present, CH_4 is insignificant or absent in some deposits. The tin-tungsten vein deposits, which are considered to be of magmatic origin, also lack CH_4, although CO_2 is present in some deposits. Fluid inclusions in most rare earth and Sn-Ta-bearing pegmatites are also poor in CO_2 and lack CH_4 (Roedder 1984).

Both CH_4 and CO_2 are abundant in fluid inclusions from low to medium-grade metamorphic environments (Swanenberg 1979, Roedder 1984). These metamorphic fluids also contain higher concentrations of calcium and magnesium (Crawford *et al* 1979). Almost all of the PCO occurrences are located within the contact aureole of granitic intrusions and a large proportion of CO_2, CH_4, $CaCl_2$ and $MgCl_2$ in the fluid inclusions could therefore be of contact-metamorphic origin.

Type B inclusions lack both CO_2 and CH_4, and are also more saline than Type A inclusions. Up to 2 mole% CO_2 is soluble in NaCl solutions at temperatures and pressures as low as 25°C and 5 bars (Collins 1979). Therefore, for a clearly visible CO_2 phase at room temperature, probably 5–10 mole% CO_2 is required to be present. Theoretically, it is possible to derive fluids trapped in Type B inclusions from those trapped in Type A inclusions by CO_2 loss due to a drop in pressure, or an increase in salinity, or both. For example, at a pressure of 1 kbar, CO_2 and H_2O are completely miscible above 275°C. If the pressure drops below 1 kbar, H_2O and CO_2 are no longer miscible and at a pressure of, say, 575 bars, a CO_2-rich fluid containing 60 mole% CO_2 and an H_2O-rich phase containing 11 mole% CO_2 can co-exist (Roedder 1984). At a pressure of about 250 bars the water-rich phase will have only about 5 mole% CO_2. For saline fluids, say about 6% NaCl, such low concentrations of CO_2 are achievable at pressures of about 500 bars (Takenouchi and Kennedy 1965).

Another possibility requires separate sources for Type A and Type B fluids. In this model, low-salinity Type A fluid is generated by contact metamorphism and may be present as a single H_2O-CO_2-CH_4 phase above the solvus. The higher salinity H_2O-CO_2-free Type B fluid of magmatic origin probably arrived later and mixed with the Type A fluid, causing CO_2 + CH_4 effervescence due to phase separation.

Sulfur isotopes

Ahmad *et al* (1993) summarised $\delta^{34}S$ measurements from a number of deposits from within the PCO. Pyrite and arsenopyrite from the Toms Gully mine gave a very restricted $\delta^{34}S$ range of 9.2 to 10.3‰ (average 9.7‰). Stage I (**Figure 5.46**) pyrite for other deposits (Enterprise, Kohinoor, Fountain Head, Maid of Erin, Union Reefs, Zapopan, Woolwonga and Faded Lily) gave a range of 3.9–8‰ (average 5.47‰) and there is no apparent inter-deposit variation in the $\delta^{34}S$ values. Two galena samples (Stage III) returned $\delta^{34}S$ values of 2 and 2.6‰ and four sphalerite samples gave values of 5.8, 4.7, 6.8 and 5.6‰. Sphalerite and galena (Stage III) co-exist in one sample, for which the $\delta^{34}S$ temperature has been calculated at 244°C, which is in general agreement with the fluid inclusion temperatures on Stage II veins.

$\delta^{34}S$ values have been determined on samples of pyrite, arsenopyrite and marcasite from the Golden Dyke and Cosmo Howley mines. Two marcasite samples gave values of -0.3 and +0.5‰. Pyrite 1 (fine disseminate) and Pyrite 2 (coarse, thin stringers) gave $\delta^{34}S$ values in the range +1.5 to +3‰ (average 2.2). Arsenopyrite values are distinctly heavier and range between +4.4 and +5.7‰. Twenty-eight samples of pyrite, pyrrhotite, chalcopyrite and sphalerite from the Mount Bonnie mine averaged at 2.0 ± 1.3‰.

The ranges of $\delta^{34}S$ values from various deposits in the PCO are plotted in **Figure 5.48**, which also includes $\delta^{34}S$ data on sulfides from various PCO rock units. $\delta^{34}S$ values

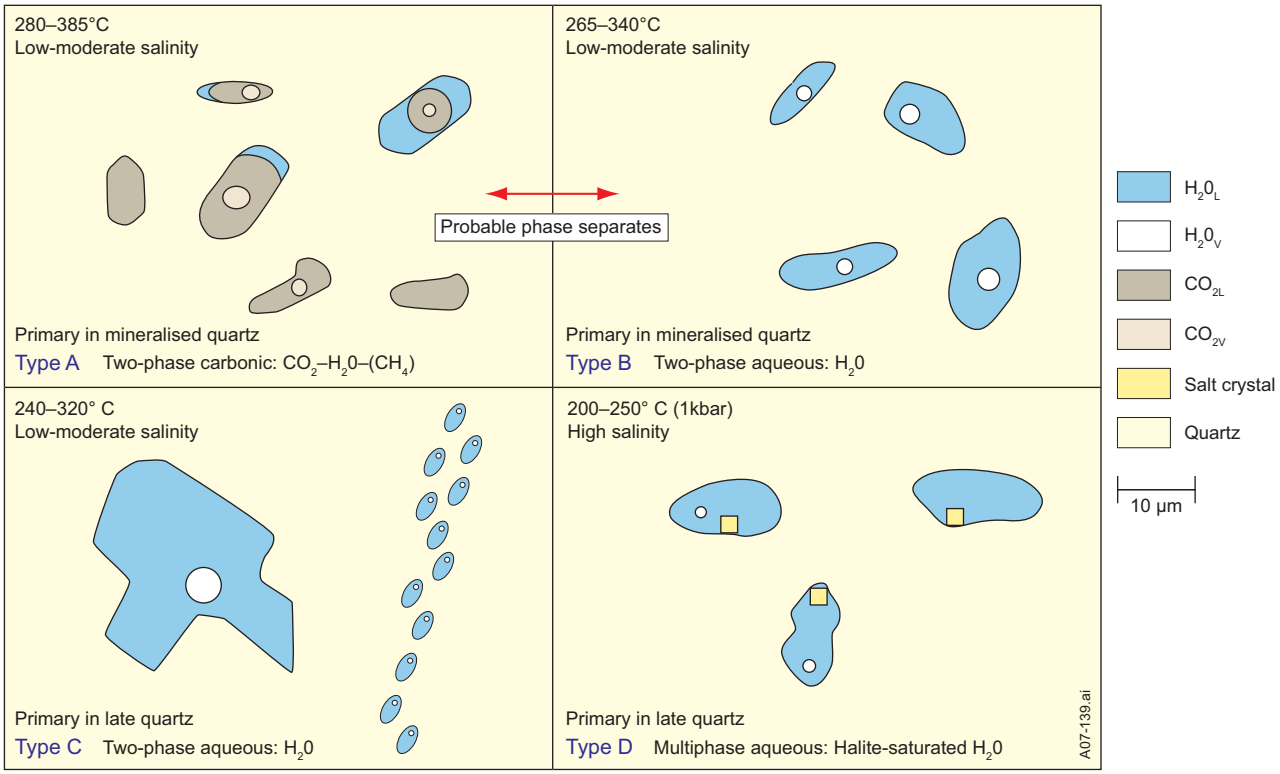


Figure 5.47. Types of fluid inclusions in PCO gold deposits (modified from Wygralak 1996).

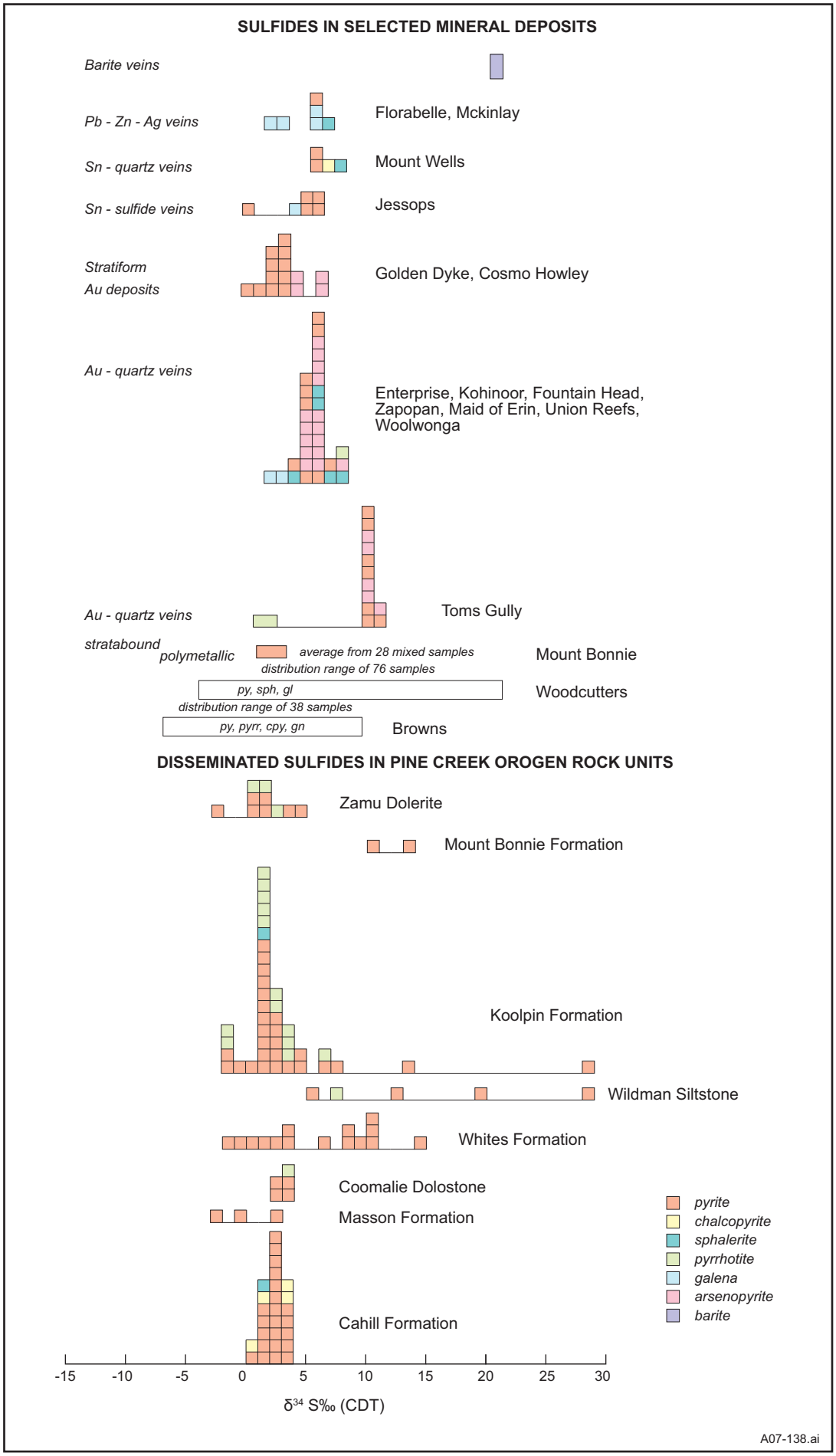


Figure 5.48. Sulfur isotope data from selected PCO mineral deposits (after Ahmad *et al* 1993). Vertical axes plots relative frequency of analyses. CDT = standard Canyon Diablo Troilite.

of vein-type deposits of gold, base metals and tin are all within the magmatic range. The Toms Gully and Northern Hercules deposits show relatively higher values, which may have resulted from the incorporation of sedimentary sulfides. Stratiform gold and polymetallic deposits also have a narrow range of $\delta^{34}S$ values that fall within the magmatic range. Previous studies on these deposits (Nicholson 1980, Goulevitch 1980b) have suggested a syn-sedimentary exhalative origin. Considering that sulfate was present in the seawater and the deposits are contained within a predominantly sedimentary succession, the $\delta^{34}S$ values support a magmatic sulfur source for these deposits.

Hydrogen and oxygen isotopes

Ahmad *et al* (1993) and Wygralak (1996) provided data on $\delta^{18}O$ determinations on the vein quartz and host rock silicate minerals and δD on fluid inclusion water from PCO gold deposit samples. $\delta^{18}O$ values of water in equilibrium with vein quartz were calculated using homogenisation temperatures and experimentally determined quartz-H_2O fractionation factors. The δD and $\delta^{18}O$ ranges of the deposits from the PCO are shown in **Figure 5.49**, along with ranges for the magmatic and metamorphic waters, taken from Taylor (1979).

The $\delta^{18}O$ H_2O values of the hydrothermal fluids show a narrow spread and are regionally similar throughout the PCO. Quartz from gold, tin, base metals, uranium and Sn-Ta pegmatites all fall within a narrow range of 3–12‰ which is compatible with either magmatic or metamorphic waters, or a mixture of these components.

The δD values for magmatic waters range from -40 to -80‰ (Taylor 1979). δD values for fluid inclusion waters from gold, tin, base metals and Sn-Ta pegmatite deposits from the PCO range from -19 to -55‰, and are heavier than the magmatic waters and within the metamorphic waters range. Such heavier values may be due to the formation of CH_4 by the reaction of hydrothermal solutions with carbonaceous material, or may be due to the mixing of isotopically lighter waters with heavier waters, eg magmatic and metamorphic waters.

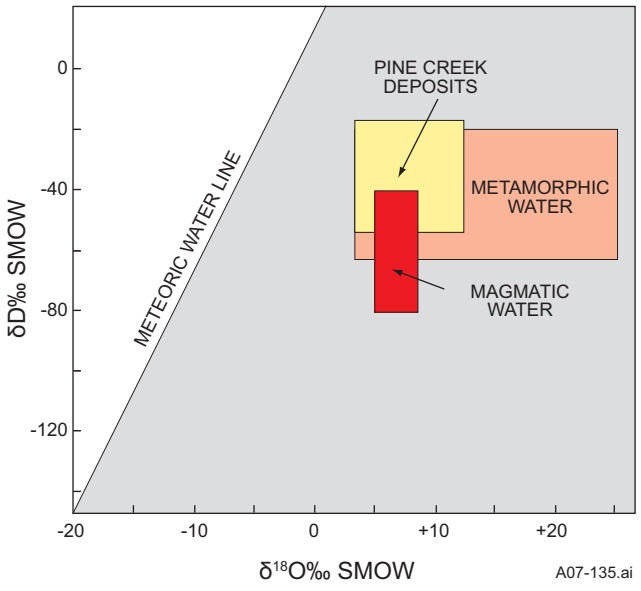


Figure 5.49. Oxygen and hydrogen isotope data of fluid inclusion water from PCO gold deposits. Also shown are magmatic and metamorphic regimes, and meteoric water line.

$\delta^{13}C$ of fluid inclusion CO_2

$\delta^{13}C$ measurements on fluid inclusion CO_2 were provided by Ahmad *et al* (1993). They range from +1 to -32‰ PDB and are plotted in **Figure 5.50**. The distinctly negative values of most samples suggest the incorporation of organically derived carbon.

Pb isotopes

Lead isotope measurements have been made on galena from several deposits in the PCO by CSIRO. A plot of the data (kindly supplied by Graham Carr, CSIRO, pers comm 2002) is reproduced in **Figure 5.51**. These data show a wide spread and can not be used for geochronology. Data on pyrite and arsenopyrite from the Toms Gully deposit are consistent with derivation from mixed granitic and country rock sources (Sheppard 1992). In contrast, Matthäi *et al* (1995a), studied $^{207}Pb/^{204}Pb$ ratios of sulfides from Cosmo Howley and showed that the lead was not derived from potential magmatic sources and that the Th/U ratios were consistent with the derivation of lead from the country rock. This was confirmed by Partington and McNaughton (1997), who evaluated the lead isotope signature of a variety of gold deposits and granitoids in the PCO. They concluded that the dominant source of lead in the deposits is probably either Palaeoproterozoic or Archaean basement rocks.

Age of mineralisation

Although there is strong spatial association of Cullen Event granites, thermal metamorphism and gold mineralisation, there is insignificant data on the actual geochronology of the gold veins. Matthäi *et al* (1995a) conducted a lead isotope study of arsenopyrite from the Cosmo Howley deposit and determined a $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ isochron age of 1820 ± 110 Ma. Although there is a large uncertainty, the mean age is indistinguishable from the SHRIMP U-Pb zircon age of the nearby McMinns Bluff Granite. An apparently better defined age of 1810 ± 10 Ma was obtained by SHRIMP U-Pb methods on a mineral separate of monazite and xenotime associated with gold

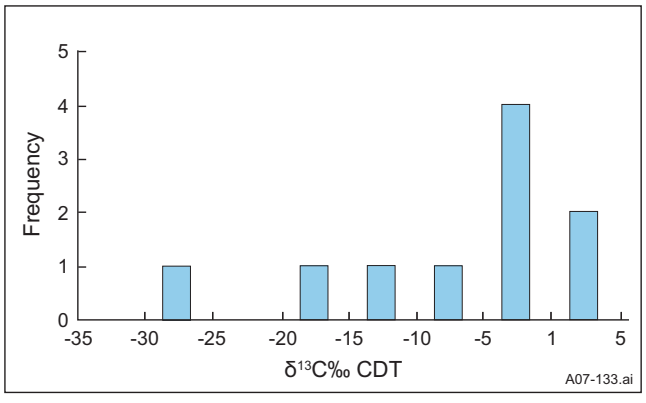


Figure 5.50. Carbon isotope data on CO_2 from fluid inclusions from PCO gold deposits (after Ahmad *et al* 1993). PDB = reference standard Pee Dee Belemnite.

mineralisation at the Goodall deposit (Compston and Matthäi 1994). However, there are doubts as to whether the monazite and xenotime are related to the ore-stage minerals or gold (Sener 2004).

Sener (2004) carried out SHRIMP U-Pb geochronological studies on monazite and xenotime from the Union Reef, Goodall and Mount Todd deposits. After the exclusion of several grains, because of unacceptably high common lead, xenotime from Union Reefs yielded a mean age of 1698 ± 18 Ma. The significance of this age is uncertain, as it does not coincide with any known event in the PCO. Monazite within the auriferous quartz veins from the Mount Todd deposit also contained a large population with unacceptably high common Pb contents. Selected analyses yielded a mean age of 1854 ± 16 Ma. This age was interpreted as representing an earlier generation of monazite that was incorporated into the veins from the wallrock by mechanical processes (Sener 2004). Xenotime from the Mount Todd deposit has yielded a mean age of 1819 ± 8 Ma, which is considered a definitive age of gold mineralisation at Mount Todd. This age coincides with the intrusion of the granites of the Cullen Supersuite. Monazite associated with the gold veins from the Goodall deposit has yielded a mean age of 1727 ± 11 Ma, after the exclusion of several analyses. Although Sener (2004) considered it as a reliable age for the gold vein formation at the Goodall deposit, this age is younger than that of the overlying El Sherana/Edith River/Katherine River groups. These units have no auriferous veins. It is also about 100 Ma younger than the intruding granites. The significance of this age is ambiguous and the analysed monazite may have been affected by later alteration processes.

Genesis of PCO gold deposits

The various genetic constraints discussed above are summarised in **Table 5.6**. A variety of genetic models ranging from exhalative syngenetic (for banded ironstone-hosted deposits) to magmatic to hydrothermal have been proposed for the formation of the PCO gold deposits. Data on fluid inclusions, stable isotopes and lead isotopes suggest a mixed metamorphic/magmatic source. Geochronological data on gold vein systems is diverse and mostly inconclusive.

Partington and McNaughton (1997) provided observations constraining the geological timing of PCO gold deposits. Gold veins cross-cut the Zamu Dolerite and are therefore younger than this intrusion. This dolerite is constrained to <1863 Ma by cross-cutting relationships with the well dated Gerowie Tuff (Worden *et al* 2008a). At Cosmo Howley, contact-metamorphic garnet, andalusite and cordierite grains are cut or fragmented in places by gold-bearing veins, providing evidence of a post-peak contact-metamorphic timing for the gold mineralisation (Matthäi *et al* 1995a). At the Enterprise deposit, cordierite porphyroblasts are commonly replaced by a sericite-biotite-quartz assemblage, which is the dominant hydrothermal alteration assemblage at this deposit (Cannard and Pease 1990). Fragmentation and the incorporation of contact metamorphic assemblages into the veins during incremental growth also provide strong evidence for a post-contact-metamorphic timing at this deposit (Matthäi *et al* 1995a).

Most geological models for gold deposits in the PCO take into consideration the spatial relationships to granitoid intrusions and proposed a mixed contact-metamorphic and magmatic source for the auriferous hydrothermal fluids, which were liberated during the

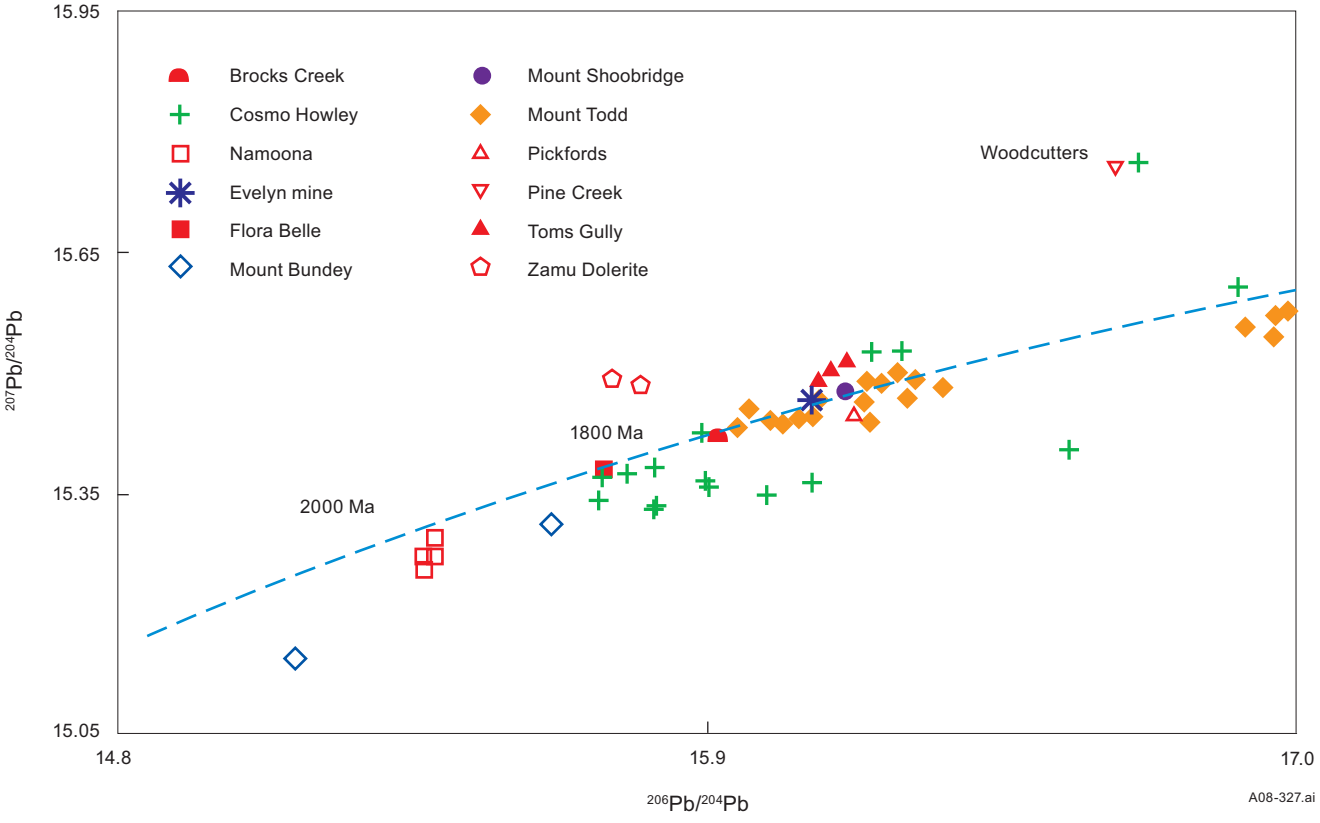


Figure 5.51. Lead isotope data from PCO mineral deposits, modified from figure supplied by Graham Carr (CSIRO, pers comm 2002).

emplacement of the granitoids (Stuart-Smith *et al* 1993, Ahmad *et al* 1993, Wygralak 1996, Wall 1989, Bajwah 1994, Matthäi *et al* 1995a, b, Partington and McNaughton 1997). On basis of monazite and xenotime geochronology, Sener (2004) argued that the PCO gold deposits formed after the youngest phase of granitoid intrusion and peak contact metamorphism, several tens of millions of years after granitoid magmatism. However as discussed above, the geochronological data are generally inconclusive. Partington and McNaughton (1997) considered that the presence of high-heat-producing granites is a prerequisite for the formation of PCO gold deposits. The heat from the granite intrusions and a prolonged cooling history coupled with pre-existing duplex thrust fold structure allowed regional-scale long-lived hydrothermal systems to channel fluids both from granites and the surrounding rocks.

Gold deposits hosted within the iron-rich sedimentary rocks are considered by some authors (Nicholson 1978, Wilkinson 1982, Nicholson and Eupene 1984) to have been derived by sedimentary exhalative processes. Matthäi *et al* (1995a, 1995b) suggested a model in which magmatic fluid interacted with CH_4+CO_2-bearing metamorphic fluid generated within the hanging wall carbonaceous slate, causing the precipitation of gold. They implied that mineralisation was caused by the mixing of magmatic and metamorphic fluids, and that the reaction of these fluids with carbonaceous matter at a temperature range of 550–620°C caused the ore mineral precipitation. Such temperatures are substantially higher than the 250–350°C range determined from fluid inclusion studies on other gold deposits in the PCO (Wygralak 1996) and on mineralised quartz veins in the Cosmo Howley and Golden Dyke mines (M Ahmad unpublished data).

The polymetallic gold deposits are considered to be of sedimentary syngenetic origin, with ore fluids derived from the devitrification of underlying tuffaceous sediments (Goulevitch 1980b, Nicholson and Eupene 1990). However, sulfur isotope data are similar to that from the other gold deposits of the PCO and suggest a magmatic sulfur source (Ahmad *et al* 1993).

Most gold deposits are located in rocks that are regionally metamorphosed to lower greenschist facies. The high-grade metamorphic rocks in the western and eastern parts of the PCO are mostly devoid of gold deposits and there are no known gold deposits hosted within the granite bodies. Another common factor is the association of gold deposits with I-type granites and their contact metamorphic aureoles. A third common factor is the presence of structural conduits for the transport of ore fluids (major lineaments, shear zones, anticlines, thrusts).

The above observations could be incorporated into a model (**Figure 5.52**) involving contact metamorphic and magmatic fluids. During the early stage (I), granitic intrusions of the Cullen Supersuite caused contact metamorphism followed by the release of oxidised, moderately saline gold-bearing fluid; at the same time, devolatilisation reactions in the surrounding rocks produced reduced, low-salinity, CO_2 ± CH_4-bearing contact-metamorphic fluids. Magmatic fluid was partly reduced while ascending through the graphitic sedimentary rocks of the Koolpin Formation and some gold was precipitated. In other deposits, mixing of fluids caused the reduction of oxidised auriferous fluids and the final precipitation of gold in favoured structures.

An alternative explanation is that the gold mineralisation events occurred several tens of million years after the intrusion of the granites (Partington and McNaughton 1997). In the high heat-producing granites, like those in the PCO, the decay of radioactive elements provides a permanent heat engine, lasting for hundred of millions of years after the crystallisation of the magma. This long-lasting thermal anomaly may have created paragenetically and chronologically complicated pulses of hydrothermal activity. This may also explain the subsequent resetting and spread of Rb-Sr, Ar-Ar data and the alteration/precipitation of monazite and xenotime, yielding anomalously younger U-Pb ages, as discussed above.

Parameter	Results	Remarks
Spatial association	Most deposits are in close proximity to ca 1825 Ma Cullen Event granites.	A genetic link with granites is possible.
Fluid inclusion composition	High concentration of CO_2 and CH_4. Halite saturation. High Ca, Mg.	Possible metamorphic influence. Possible igneous link or evaporate dissolution. Possible metamorphic source.
Hydrogen and oxygen isotopes	$\delta^{18}O$ vein quartz 10 to 17‰. $\delta^{18}O$ H_2O 3 to 12‰. δD Fluid inclusions -50 to -85‰.	Possibly magmatic or metamorphic. Possibly magmatic. Possibly mixed magmatic or metamorphic.
Sulfur isotopes	$\delta^{34}S$ Vein sulfides 0.5 to 15‰.	Possible magmatic source.
Carbon isotopes	$\delta^{13}C$ Fluid inclusion +1 to -32‰.	Interaction with carbonaceous rocks.
Lead isotopes		Possibly derived from metasedimentary rocks or Archaean basement.
Geochronology	Cosmo Howley 1820 ± 110 Ma lead isotope. Cosmo Howley1810 ± 10 Ma monazite. Mount Todd 1854 ± 16 Ma monazite; 1819 ± 8 Ma xenotime. Goodall 1727 ± 11 Ma monazite. Union Reefs 1698 ±18 Ma xenotime.	Can be linked to granite. Can be linked to granite. Can be linked to granite. Can be linked to granite. Possible age of alteration. Possible age of alteration.

Table 5.6. Summary data on genetic constraints of PCO gold deposits.

Uranium

The PCO hosts World-class uranium deposits in the East Alligator River region at the Ranger 3, Jabiluka and Koongarra deposits. Two other significant deposits, Nabarlek and Ranger 1, have been mined out. Important uranium deposits are also located in the Rum Jungle and South Alligator River Valley regions. Several other smaller deposits are known from elsewhere.

The first recorded discovery of uranium mineralisation was made in 1949 at Rum Jungle at the Whites deposit. Soon after, in 1953, BMR discovered uranium at Coronation Hill. Several deposits were later discovered and mined in the Rum Jungle and South Alligator River regions between 1954 and 1971. Notable amongst these are Dysons, Whites, Intermediate, Rum Jungle Creek South, Mount Fitch and Mount Burton in the Rum Jungle region and El Sherana, Rock Hole, Coronation Hill and Sleisbeck in the South Alligator region. The next phase of uranium exploration and discoveries was from 1969–1971, when the major uranium deposits at Nabarlek, Ranger, Jabiluka and Koongarra were discovered.

A resurgence in uranium exploration has coincided with a rise in uranium prices over the past few years. Extensive exploration is currently underway in the PCO, Murphy Inlier, and the Ngalia and Amadeus basins, as well as in several other parts of the Northern Territory.

Uranium deposits have been classified on the basis of host rocks, structural setting, mineralogy and geochemistry. Nash *et al* (1981) proposed a classification scheme based on the geological setting and genesis of mineralisation, which is similar to that developed by the International Atomic Energy Agency (IAEA) in 1988–1989 (OECD/NEA and IAEA 2000) and to that used by Geoscience Australia in their publication on uranium deposits in Australia (McKay and Miezitis 2001). The uranium deposits of the PCO have been grouped into four main types: unconformity-related, vein-type, intrusive-related and surficial deposits (Lally and Bajwah 2006). Resources at major deposits are given in **Table 5.7** and the spatial distribution of uranium occurrences is given in **Figure 5.53**. Major deposits cluster around the East Alligator, South Alligator and Rum Jungle regions.

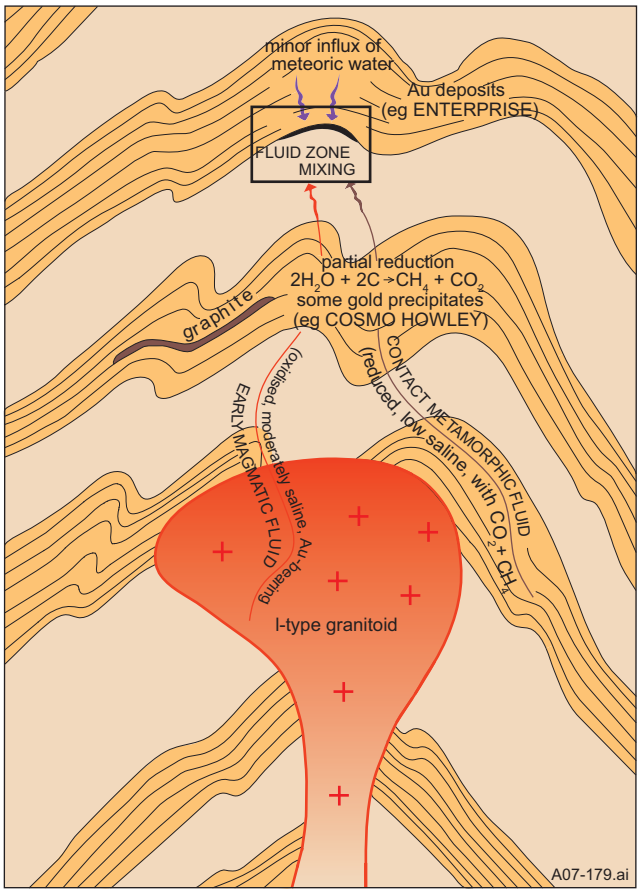


Figure 5.52. Model for genesis of gold deposits of PCO (after Ahmad *et al* 1999). See text for more detailed explanation.

Unconformity-related uranium deposits

Major deposits are within Palaeoproterozoic rocks close to the unconformity with overlying sandstones of the Katherine River Group (Alligator rivers areas) or Tolmer Group (Rum Jungle area). In both of these areas, Palaeoproterozoic rocks unconformably overlie Neoarchaean granite and gneiss (Nanambu and Rum Jungle complexes). Deposits range in size from a few tonnes to more than 100 000 t of contained U_3O_8. The largest deposits are in the East Alligator River area and include Ranger 1 and 3, Jabiluka, Koongarra and Nabarlek (**Table 5.7**).

The Rum Jungle uranium deposits (Whites, Dysons, Intermediate, Rum Jungle Creek South) are located in the vicinity of the Rum Jungle Complex and are hosted within carbonaceous and pyritic shale of the Whites Formation, adjacent to the contact with the underlying Coomalie Dolostone.

Minor smaller deposits occur in clusters in the South Alligator Valley (**Figure 5.53**). Gold, platinum and palladium are associated with some deposits and may be present in economic quantities. These deposits are generally hosted within the Koolpin Formation or Coronation Sandstone, near the unconformity with overlying sandstone of the Katherine River Group. Unlike the major deposits in the East Alligator River and Rum Jungle areas, there is no exposed Archaean basement in the South Alligator Valley area, but this may be present at depth.

East Alligator River area

Exploration interest in this region followed mapping by BMR and the publication in 1968 of a 1:500 000-scale map of the PCO, which revealed striking geological similarities between the East Alligator River area and the uraniferous Rum Jungle area. From 1969–1971, the major unconformity-related deposits at Nabarlek, Ranger 1 and 3, Jabiluka and Koongarra were discovered, as well as a number of other significant prospects (**Figure 5.54**).

Production started from Nabarlek in 1979 and from Ranger 1 in 1981. The Nabarlek mill was decommissioned in 1988 after producing about 11 084 t of U_3O_8. Mining at Ranger 1 was completed in 1994 and full-scale mining of the Ranger 3 orebody commenced in 1997.

Ranger 1 and 3

These two deposits are often known as *Ranger 1 No 1* and *Ranger 1 No 3* orebodies (originally anomalies 1 and 3). For the sake of simplicity, they will be referred as the *Ranger 1*

and *3* deposits. They are located 10 km east-southeast of the township of Jabiru. Both are in the lower Cahill Formation, close to the faulted and sheared contact with the underlying Nanambu Complex (**Figure 5.55**). The mine succession was informally divided by Eupene *et al* (1975) into footwall (FWS), lower mine (LMS), upper mine (UMS) and hangingwall (HWS) sequences (**Figure 5.56**). The FWS comprises schist, gneiss, amphibolite and migmatite of the Nanambu Complex. The LMS comprises recrystallised dolostone, magnesite and chlorite-sericite schist. The latter contains lenticles of chlorite, is often referred to as lenticular schist and is used as a marker bed. The UMS comprises quartz feldspar-mica schist and dolomitic marble, and is locally carbonaceous. The HWS comprises amphibolite and schist.

The mine succession dips 15° to 40° towards the east and there is a strong schistosity sub-parallel to bedding that

Common name	Lat	Long	Size	Stratigraphic unit	Shape	Model	Production U_3O_8 (t)	Resource (Mt)	Grade U_3O_8 (%)	Reference
ABC	-14.33	132.35	Small	Kombolgie Sgp	Tabular	Vein-type		*0.001*	*0.4*	Lally and Bajwah (2006)
Adelaide River	-13.27	131.11	Small	Burrell Ck Fm	Sheet	Vein-type	17.2	*0.006*	*0.32*	Plumb (1960)
Austatom	-12.62	132.67	Medium	Cahill Fm	Irregular	Unconf-related				Lally and Bajwah (2006)
Beatrice	-12.62	133.20	Occurrence	Cahill Fm?	Irregular	Unconf-related				Lally and Bajwah (2006)
Caramel	-12.50	133.36	Medium	Cahill Fm?	Tabular	Unconf-related		*0.758*	*0.33*	McKay and Miezits (2001)
Coronation Hill	-13.59	132.61	Medium	Koolpin Fm	Tabular	Unconf-related	75	*0.34*	*0.54*	Lally and Bajwah (2006)
Dam	-13.57	132.02	Medium	Masson Fm	tabular	Vein-type		0.58	0.028	JORC Compliant Atom Energy (2008)
Dysons	-12.99	131.02	Medium	Whites Fm	Tabular	Unconf-related	534			Lally and Bajwah (2006)
El Sherana	-13.51	132.52	Medium	Koolpin Fm	Tabular	Unconf-related	226			Lally and Bajwah (2006)
El Sherana West	-13.51	132.52	Medium	Koolpin Fm	Tabular	Unconf-related	185			Lally and Bajwah (2006)
Fleur de Lys	-13.53	131.37	Small	Gerowie Tuff	Sheet	Vein-type	0.21			Lally and Bajwah (2006)
George Creek	-13.34	131.14	Small	Burrell Ck Fm	Sheet	Vein-type	0.32			Lally and Bajwah (2006)
Hades Flats	-12.59	132.92	Small	Cahill Fm	Tabular	Unconf-related		*0.36*	*0.2*	Lally and Bajwah (2006)
Jabiluka 1	-12.50	132.94	Medium	Cahill Fm	Tabular	Unconf-related		1.3	0.25	Lally and Bajwah (2006)
Jabiluka 2	-12.50	132.90	Large	Cahill Fm	Tabular	Unconf-related		29.24	0.48	Energy Resources (2010)
Koongarra	-12.87	132.84	Large	Cahill Fm	Tabular	Unconf-related		0.812	1.102	AREVA (2008)
Mount Burton	-12.98	130.96	Small	Whites Fm	Tabular	Unconf-related	12.8			Lally and Bajwah (2006)
Mount Fitch	-12.95	130.95	Medium	Whites Fm	Tabular	Unconf-related		5.03	0.03	Compass Resources (2008)
Nabarlek	-12.31	133.32	Large	Nourlangie Schist?	Tabular	Unconf-related	11084			Lally and Bajwah (2006)
O'Dwyers	-13.47	132.47	Small	Koolpin Fm	Tabular	Unconf-related	152			Lally and Bajwah (2006)
Palette	-13.53	132.56	Small	Koolpin Fm	Pipe-like	Unconf-related	124			Lally and Bajwah (2006)
Ranger 3	-12.68	132.92	Large	Cahill Fm	Tabular	Unconf-related	49194	123.23	0.09	Energy Resources (2010)
Ranger 1	-12.69	132.92	Large	Cahill Fm	Tabular	Unconf-related	60961			Lally and Bajwah (2006)
Ranger 68	-12.51	132.87	Medium	Cahill Fm	Tabular	Unconf-related		1.5	0.357	Browne (1990)
Rockhole	-13.47	132.47	Small	Coronation Sst	Tabular	Unconf-related	152			Lally and Bajwah (2006)
Rum Jungle Creek South	-13.04	131.00	Medium	Whites Fm	Tabular	Unconf-related	2859			Lally and Bajwah (2006)
Saddle Ridge	-13.55	132.57	Small	Koolpin Fm	Tabular	Unconf-related	78			Lally and Bajwah (2006)
Scinto 5	-13.53	132.55	Small	Koolpin Fm	Tabular	Unconf-related	3			Lally and Bajwah (2006)
Scinto 6	-13.52	132.56	Small	Coronation Sst	Tabular	Superf	22			Lally and Bajwah (2006)
Skull	-13.54	132.57	Small	Koolpin Fm	Tabular	Unconf-related	3			Lally and Bajwah (2006)
Sleisbeck	-13.78	132.83	Small	Mount Bonnie Fm	Tabular	Superf	3			Lally and Bajwah (2006)
Twin	-13.58	132.03	Medium	Masson Fm	Tabular	Vein-type		0.824	0.031	Atom Energy (2008)
Whites	-12.99	131.01	Medium	Whites Fm	Tabular	Unconf-related	1069			Lally and Bajwah (2006)

Table 5.7 Summary data on some significant PCO uranium deposits. Abbreviations: Fm = Formation; Sst = Sandstone; Ck = Creek; Sgp = Subgroup; Superf = Superficial enrichment. Resources in italics are non-JORC compliant.

corresponds with the regional S_2 foliation (Johnston 1984). Quartz-feldspar-tourmaline pegmatite bodies intruded the mine succession and U-Pb monazite geochronology on one such intrusion has yielded an age of 1847 ± 1 Ma (Annesley *et al* 2002).

Primary uranium mineralisation in both orebodies includes uraninite or pitchblende, some coffinite and minor brannerite and curite. Secondary uranium minerals include saleeite, sklodowskite, torbernite and kasolite. Native gold has been reported from both Ranger 1 and 3, and is associated with uraninite. One sample of uranium ore from Ranger 3 that assayed 0.3% U_3O_8 also contained 1100 ppb Pt and 980 ppb Pd (Lally and Bajwah 2006).

Open-cut mining of the Ranger 1 deposit between 1980–1994 produced 18.036 Mt at an average grade of 0.338% U_3O_8. Mining of the Ranger 3 deposit, located about 1.5 km north of Ranger 1, started in mid 1997 at the rate of about 5000+ t of U_3O_8 per year.

On 29 January 2010, ERA reported the total JORC-compliant resource at Ranger 3 to be 127.23 Mt of ore averaging 0.09% U_3O_8, and total stockpiled, proved and probable reserves as 27.69 Mt at 0.14% U_3O_8. At Ranger 3, recent drilling has outlined an orebody to the east of the current open cut, which has been termed *'Ranger 3 Deeps'*. The resources given above include 34 000 t of uranium oxide from this orebody. The Ranger 3 Deeps exploration target has further potential, as it remains open to the north (Energy Resources 2010). ERA is considering placing an exploration decline to allow them to conduct closely spaced underground exploration drilling, in order to further evaluate the extent and continuity of the Ranger 3 Deeps orebody.

Ranger 1 is within a basin shaped structure formed by the thinning of LMS carbonate rocks, relative to UMS rocks (**Figure 5.56**). The LMS in the synformal hinge is predominantly chert, which has resulted from silica replacement of original carbonate sediments (Eupene *et al* 1975). UMS rocks within the synform are extensively brecciated, particularly near the UMS–LMS contact, and the contact itself is folded.

Dark green and light green chlorite-altered pegmatites and dolerite dykes intrude the mine succession. Pegmatites may be weakly mineralised; they are affected by ore-related

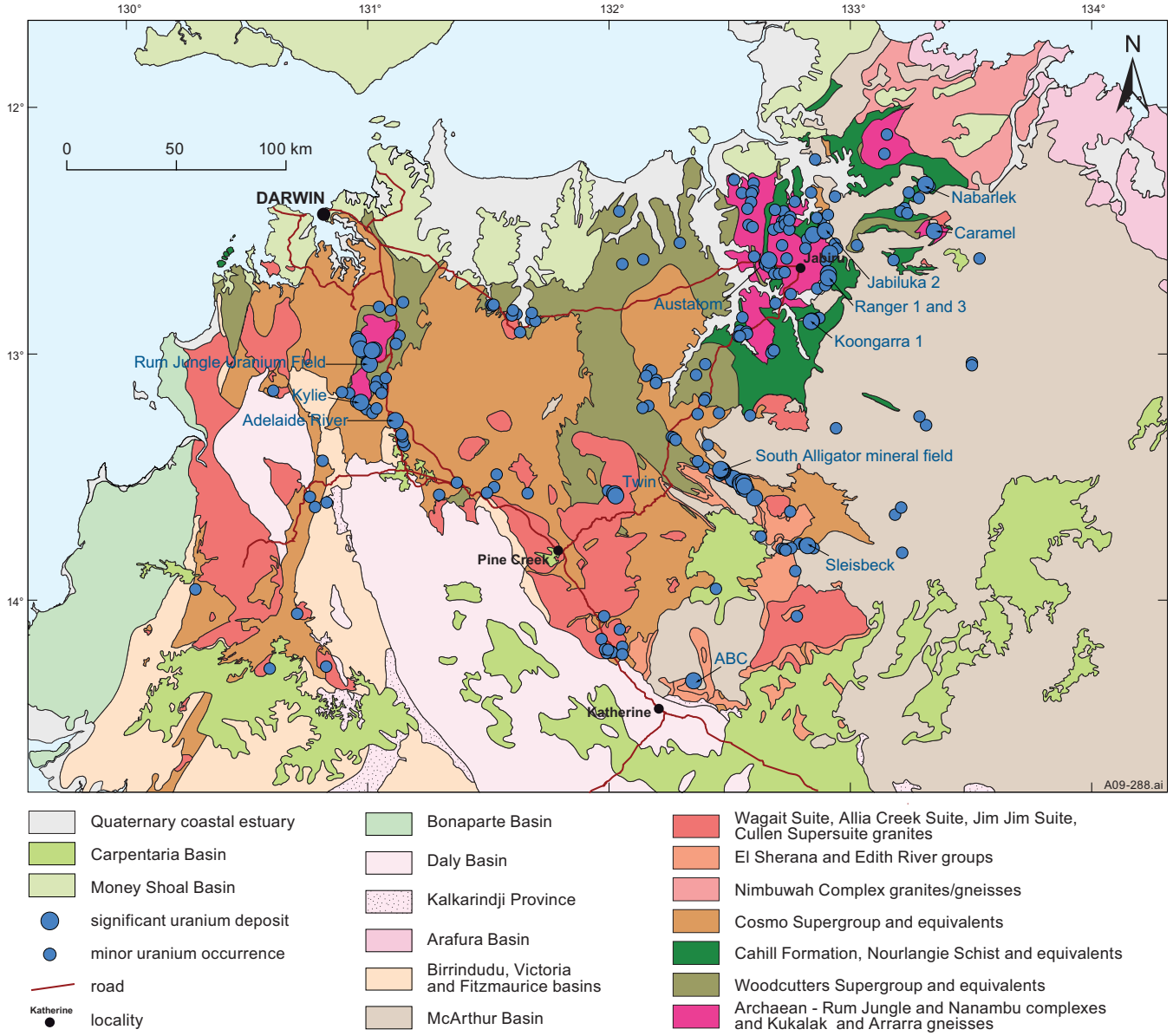


Figure 5.53. Distribution of uranium occurrences in PCO. Data from NTGS MODAT database. Significant deposits are labelled. Geology simplified from 1:2.5M geological map of NT (Ahmad and Scrimgeour 2006).

alteration assemblages and are offset by syn-mineralisation faults (Lally and Bajwah 2006). Dolerite dykes were intruded along faults associated with primary uranium mineralisation (Kendall 1990).

Ranger 3 has a broad 'ramp-and-flat' geometry at the LMS–UMS contact, which is steeper on the western and eastern sides and sub-horizontal beneath the main orebody (**Figure 5.57**). The FWS–LMS contact also steepens to the

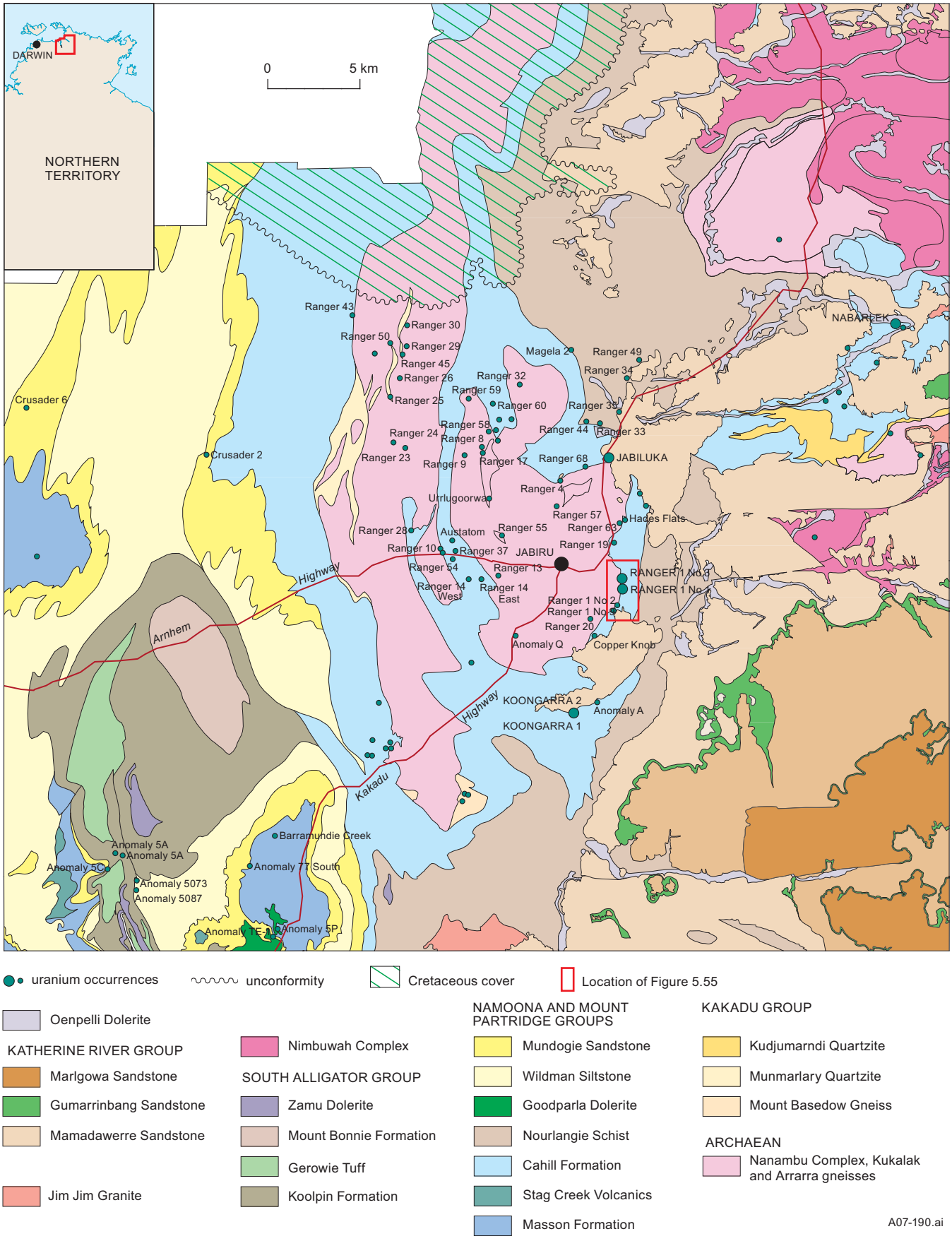


Figure 5.54. Distribution of uranium occurrences in East Alligator River uranium field. Geology simplified from 1:2.5M geological map of NT (modified from Lally and Bajwah 2006).

west, but the 'flat' beneath the orebody is less pronounced than the LMS–UMS contact. In the western wall of the pit, a faulted outlier of haematite- and chlorite-altered and brecciated Mamadewerre Sandstone (Katherine River Group) occurs between the LMS and UMS. The eastern side of this outlier is a reverse fault, with UMS in the hangingwall, and can be traced into the UMS–LMS contact down-dip. Locally steeply dipping beds within the sandstone (80–90°) indicate that folding was associated with the reverse faulting (Lally and Bajwah 2006).

Mineralisation is almost entirely within the UMS, close to the contact with the LMS, but at depth to the east (Ranger 3 Deeps), this high-grade zone lies further above the contact (Lally and Bajwah 2006). Above the flat portion of the contact, there is a broader zone of mineralisation within the UMS that is associated with reverse faults. Two pods of uranium mineralisation occur within the LMS, well below the UMS contact, within chlorite veins infilling brecciated carbonate rocks. One high-grade pod of mineralisation in the UMS occurs in the southern part of the pit within a portion of the steeply dipping limb of a large-scale kink fold. The geometry of the kink fold is consistent with reverse movement along the fault and LMS–UMS contact (Lally and Bajwah 2006). Pegmatites that intrude the succession are barren or very weakly mineralised and are commonly strongly altered to massive quartz-chlorite or chlorite rock.

Lally and Bajwah (2006) interpreted the LMS–UMS contact as a fault surface in a west-directed thrust system. Dilation and brecciation is most intense along the low-angle 'flat' in the geometry of the fault plane. Changes in the geometry of the LMS–UMS contact, due to pre-mineralisation folding, faulting and LMS carbonate dissolution, controlled the fracturing of the UMS and therefore focused mineralising fluids to form the orebody.

Nabarlek

Nabarlek is located about 62 km northeast of Jabiru and was discovered by Queensland Mines Ltd in 1970 from follow-up exploration of an intense airborne radiometric anomaly. In 1971, ore reserves were stated as 402 000 t at 2.37% U_3O_8 for a total of 9527.4 t U_3O_8 (Anthony 1975). Open-cut mining commenced in June 1979 and was completed in 4 months and 11 days. During this period, 546 437 t of ore averaging 1.84% U_3O_8 and 157 000 t of below-grade ore at 0.05% U_3O_8 was stockpiled (McKay and Meizitis 2001). Processing of the stockpile commenced in June 1980 and continued until 1988, for a total production of 11 084 t U_3O_8. Uranium Equities Ltd is currently exploring the Nabarlek Uranium Lease and surrounding region with some success. The best RC drilling result, from 250 m south of the Nabarlek pit, has been reported as 9.5 m at 0.069% U_3O_8 from 68.5 m. This mineralisation is along the projected southern trend of the Nabarlek Shear Zone (Uranium Equities 2009).

Mineralisation is within altered metapelitic rocks that were previously included within the Myra Falls Metamorphics, but may be regarded as a possible stratigraphic equivalent of the Nouralangie Schist (**Figures 5.58** and **5.59**). The mine succession comprises

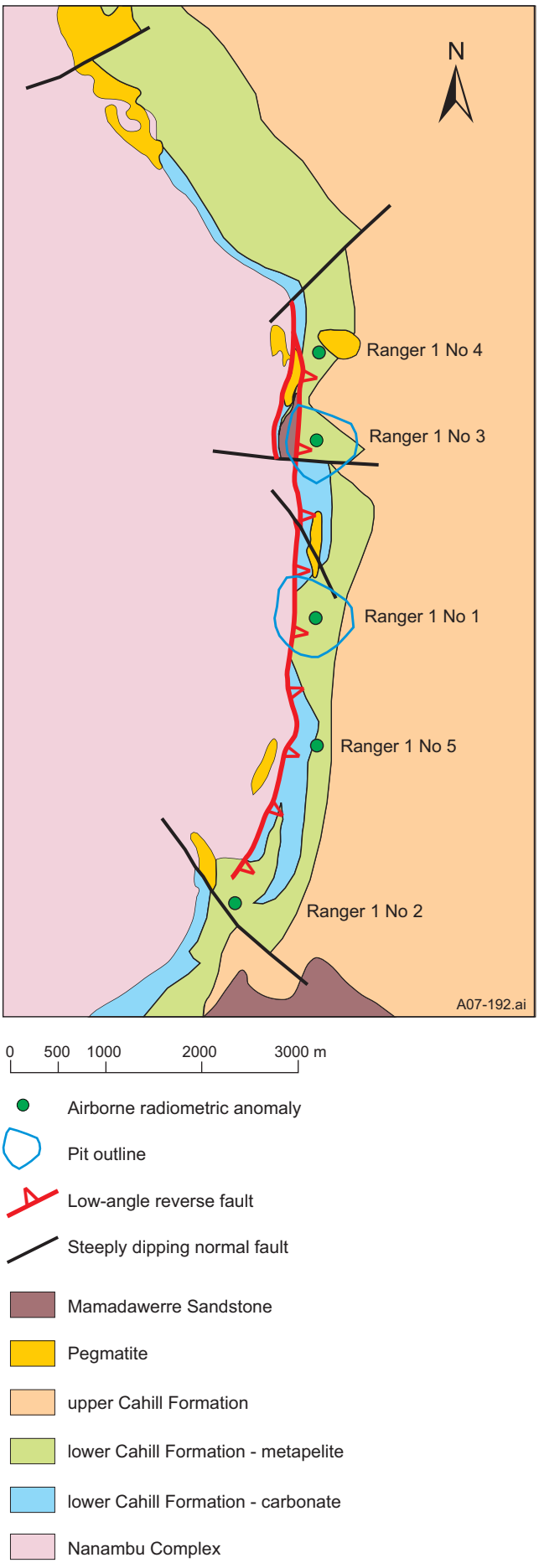


Figure 5.55. Geology of area around Ranger 1 uranium deposits (modified from Eupene *et al* 1975). Location shown in **Figure 5.54**.

about 200 m of muscovite-quartz-biotite schist ('Lower Schist') which is overlain by 100 m of interlayered hornblende amphibolite and muscovite-biotite-quartz-schist ('Footwall Amphibolite'), which is in turn overlain by biotite-muscovite-quartz and feldspar schist ('Hanging Wall Schist'). Rare quartz-feldspar-muscovite-tourmaline pegmatite veins intrude the schist and amphibolite units. The Nabarlek Granite has been intersected 450 m below the orebody and has a faulted contact with the schist succession (Wilde and Wall 1987). Although now eroded from above the deposit, sandstone of the Kombolgie Subgroup forms escarpments 1.5 km to the north and south and, from extrapolation, the unconformity would have been about 50 m above the current ground surface (Lally and Bajwah 2006).

A gabbro sill (220–250 m thick) intrudes the succession and dips gently to the south into a regional-scale west-southwest-plunging, open synformal hinge. This sill has

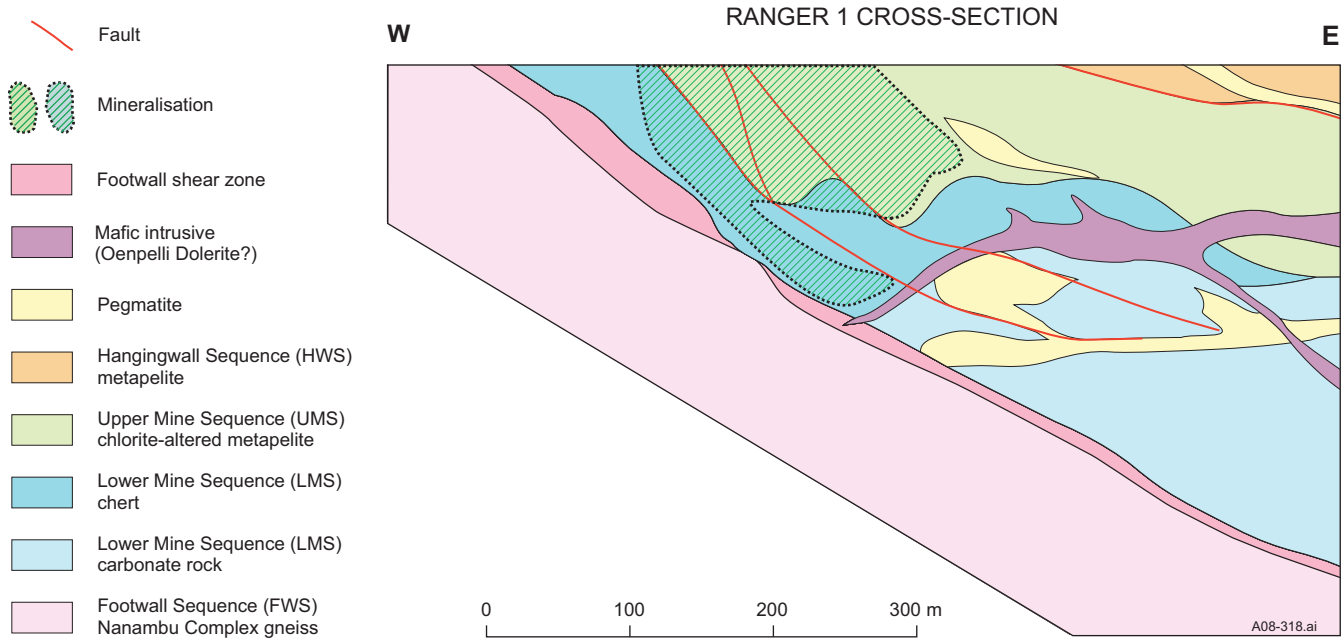


Figure 5.56. East–west geological cross-section through Ranger 1 orebody, showing outline of ore-grade mineralisation [modified from Savory (1994) and Kendall (1990)].

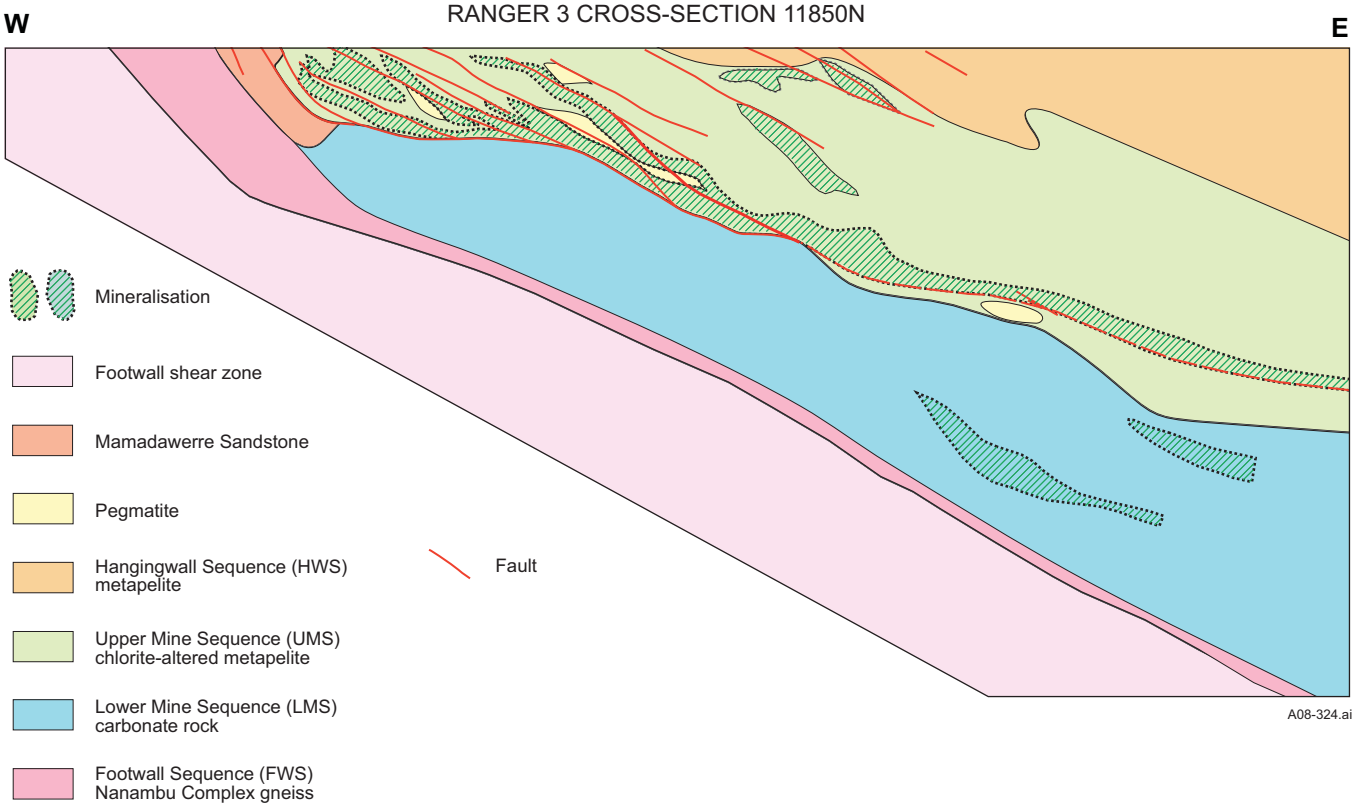


Figure 5.57. East–west geological cross-section through Ranger 3 orebody, showing outline of ore-grade mineralisation (after Lally and Bajwah 2006).

been correlated with Oenpelli Dolerite. U-Pb baddelyite and hornblende K-Ar dates on this sill indicate an emplacement age of 1725–1720 Ma (Polito *et al* 2004).

Ductile deformation, associated with the ca 1865 Ma Nimbuwah Event, predated intrusion of the Nabarlek Granite at 1818 Ma and was followed by brittle reverse faulting (Johnston 1984, Lally and Bajwah 2006). These faults were initiated soon after, or during intrusion of the Nabarlek Granite and were reactivated at various times, including during mineralisation (Polito *et al* 2004).

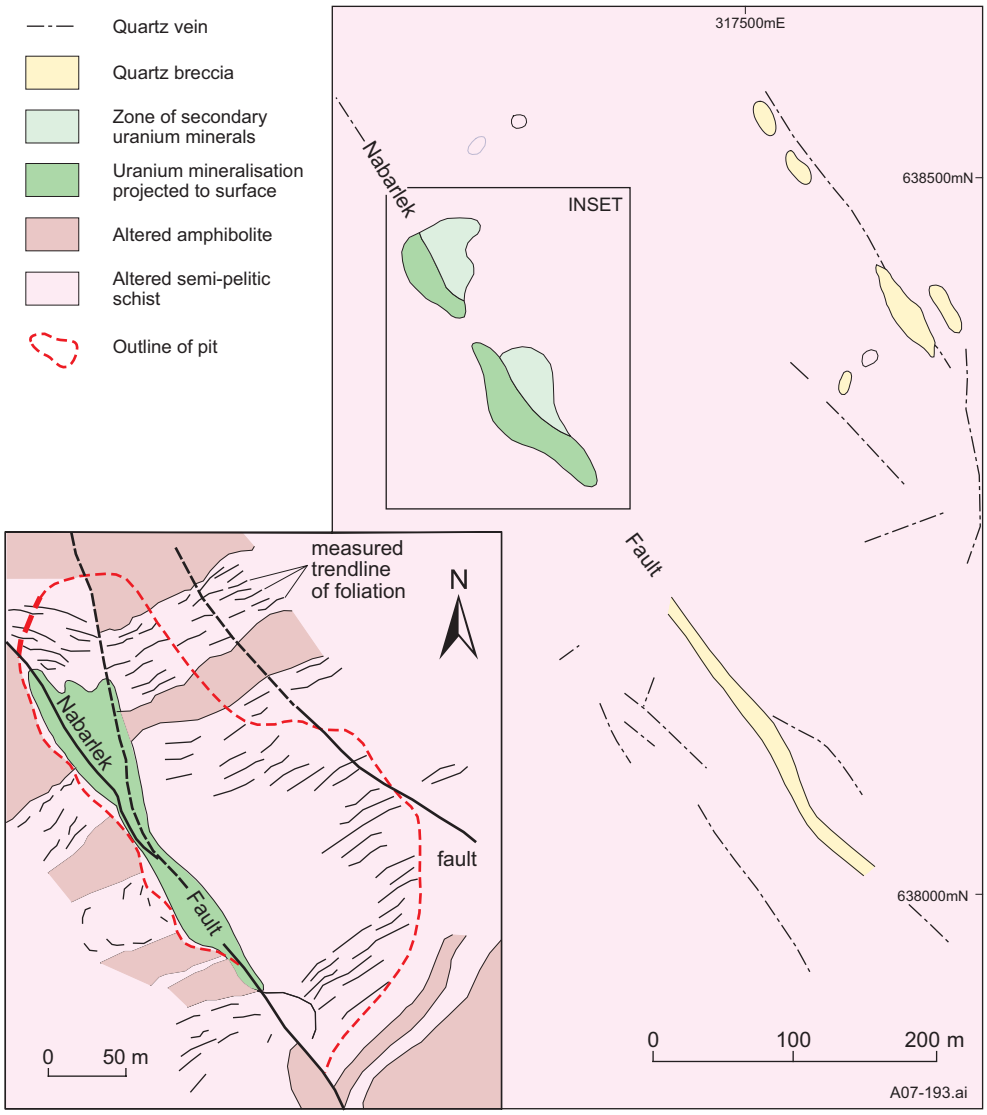


Figure 5.58. Geology of Nabarlek deposit (modified from Polito *et al* 2004).

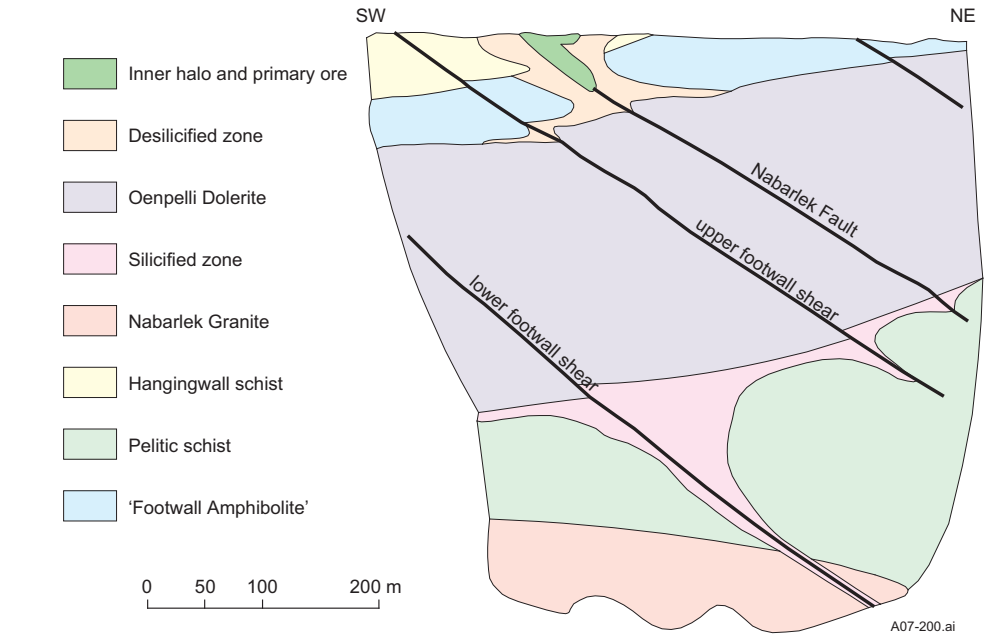


Figure 5.59. Cross-section through centre of Nabarlek orebody (modified from Polito *et al* 2004).

Individual reverse faults are defined by breccia zones up to 15 m wide, intense veining and hydraulic fracturing.

Mineralisation is within the 'Footwall Amphibolite' and 'Hanging Wall Schist', and is associated with the Nabarlek Fault breccia. It comprises high-grade ore (>1.0% U_3O_8) within an envelope of low-grade ore (0.15% U_3O_8). The low grade envelope extends for several metres, with ore occurring in veins and disseminations in intensively altered wall rocks (Wilde and Wall 1987). Grades of up to 75% U_3O_8 have been reported in the breccia-cementing ore (Anthony 1975). Mineralisation in the Oenpelli Dolerite is negligible, possibly due to its impermeability relative to the overlying schist units (Wilde and Wall 1987). The orebody is irregular, wider and shallower in the northwest and narrower and deeper in the southeast. This was interpreted as a function of the depth to the impermeable Oenpelli Dolerite, which dips towards the southeast (Wilde and Wall 1987).

Primary ore comprises uraninite and rare brannerite in a gangue of magnesian chlorite, anatase with rare haematite, apatite and sericite-phengite. Secondary near-surface minerals include sklodowskite, rutherfordine, kasolite and curite in a matrix of illite, kaolinite, anatase and haematite.

The host rocks have been altered in an outer halo that extends for at least 1 km perpendicular to the plane of the Nabarlek fault and an inner halo that includes the main orebody. The outer halo is characterised by retrograde replacement of metamorphic biotite, feldspar and hornblende by Fe-chlorite and minor sericite. Coincident silicification and dolomite veining in the outer halo was caused by intrusion of the Oenpelli Dolerite (Polito *et al* 2004). The inner alteration halo is characterised by phengite, haematite, Fe-chlorite and the complete absence of quartz (de-silicification). Chlorite and phengite increase towards the ore zone, with phengite becoming more dominant closer to ore (Polito *et al* 2004). Chlorite, quartz and sulfides (mostly galena and chalcopyrite) overprint the ore alteration assemblage.

Jabiluka 1 and 2

These two adjacent deposits are located 20 km north of Jabiru and were discovered in the early 1970s by Pancontinental Mining Ltd during a ground radiometric survey. Drilling defined a resource of 1.3 Mt of ore at 0.25% U_3O_8 at Jabiluka 1 and 52 Mt at 0.39% U_3O_8 at Jabiluka 2 (Hancock et *al* 1990).

In 1991, the Jabiluka leases were transferred to Energy Resources of Australia Ltd (ERA). Construction of a decline and cross-cut, followed by underground resource drilling, was completed in 1999. Mineral resources for Jabiluka 2 have been recently revised. The total JORC-compliant reserves are stated to be 13.8 Mt at an average grade of 0.49% U_3O_8 and total resources (excluding reserves) are given as 29.24 Mt averaging 0.48% U_3O_8 (Energy Resources of Australia, 2010). In 2003, ERA attempted to mine the Jabiluka ore and constructed a decline, but the full consent of the aboriginal traditional owners was not forthcoming and later that year, the decline was rehabilitated.

Like Ranger, Jabiluka 1 and 2 are hosted by the lower Cahill Formation, which is unconformably overlain by the Mamadewerre Sandstone (**Figure 5.60**). The former

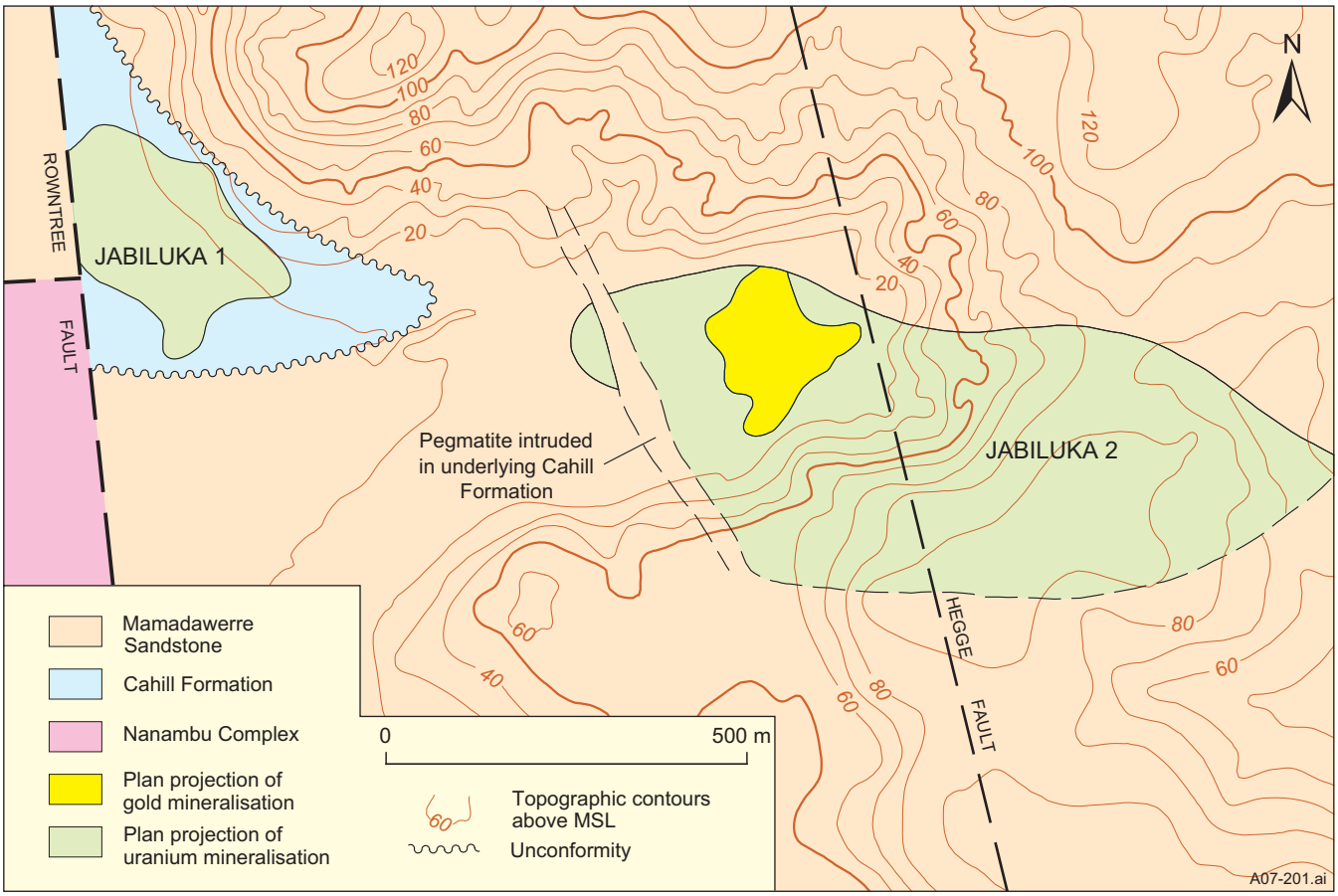


Figure 5.60. Geological plan of Jabiluka uranium deposits (modified from Hancock *et al* 1990).

comprises a bottom-to-top succession of amphibolite, pelitic schist and carbonate rocks; this is the reverse of the regional succession and is therefore interpreted to be overturned (Hancock *et al* 1990). Archaean basement was not intersected in drilling beneath the deposits, but is likely to be present at depth.

Pancontinental geologists divided the deposit succession into nine units (**Figure 5.61**). These are, from bottom to top: amphibolite, lower schist series, dolomite magnesite complex, footwall schist series, mine sequence, hanging wall schist series, graphite schist and upper schist. All units, except the amphibolite, lower schist and upper schist series, contain uranium mineralisation (Rowntree and Mosher 1976). These units were later re-named by ERA geologists to 'Stratigraphic Assay Levels' (SAL), on the basis of muscovite content (**Table 5.8**). Uranium mineralisation occurs mostly within SAL 4 and SAL 8 in Jabiluka 2 and exclusively within SAL 4 at Jabiluka 1. Other units contain smaller amounts of generally patchy mineralisation at Jabiluka 2, mostly near the contacts with mineralised units (Hancock *et al* 1990). The succession is intruded by a number of pegmatites which postdate the schistosity and are barren of mineralisation.

The unconformity at the base of the Mamadewerre Sandstone is gently dipping over the northern part of Jabiluka 2, but becomes steeper to the south. It was previously described as a tectonised contact by earlier workers (eg Rowntree and Mosher 1976), but observations made from more extensive drill core logging and mapping in the decline do not appear to support this interpretation (Lally and Bajwah 2006).

Rowntree and Mosher (1976) and Hancock *et al* (1990) considered that the uranium mineralisation is stratabound. However, subsequent observations in underground workings and drill core suggest that mineralisation is controlled by fracture and breccia zones related to graphitic shears (Energy Resources of Australia 2000). The primary uranium minerals are uraninite and pitchblende, with minor coffinite, brannerite and organo-uranium minerals (Binns *et al* 1980a). Secondary uranium minerals occur in the weathering zone at Jabiluka 1 and include autunite, sklodowskite and saleeite (Rowntree and Mosher 1976). Sulfides include pyrite, chalcopyrite, rare chalcocite, covellite and galena.

Alteration comprises an outer zone, within which there was pseudomorphous replacement of metamorphic biotite and feldspar by Fe-chlorite and white mica respectively, and an intensive inner zone that occurs within 100 m of the ore at Jabiluka 2. Binns *et al* (1980a) referred to the outer zone as a 'retrograde metamorphic halo' and noted that quartz, metamorphic muscovite and graphite was unaltered. Disseminated uraninite occurs in cleavage planes and as fractures in chlorite in this zone (Hancock *et al* 1990). Wilde (1988) defined several mineralogical changes in the inner alteration zone that occurred with increasing proximity to ore breccias: 100 m from ore, chlorite and white-mica replaced hornblende; 50 m from ore, chlorite pseudomorphically replaced garnet, and; less than 50 m from ore, chert replaced dolomite and magnesite marble. Immediately adjacent to the ore breccias, very fine-grained Mg-chlorite and phengite replaced all minerals, including quartz, but some relic metamorphic muscovite remained (Binns *et al* 1980a, Wilde 1988).

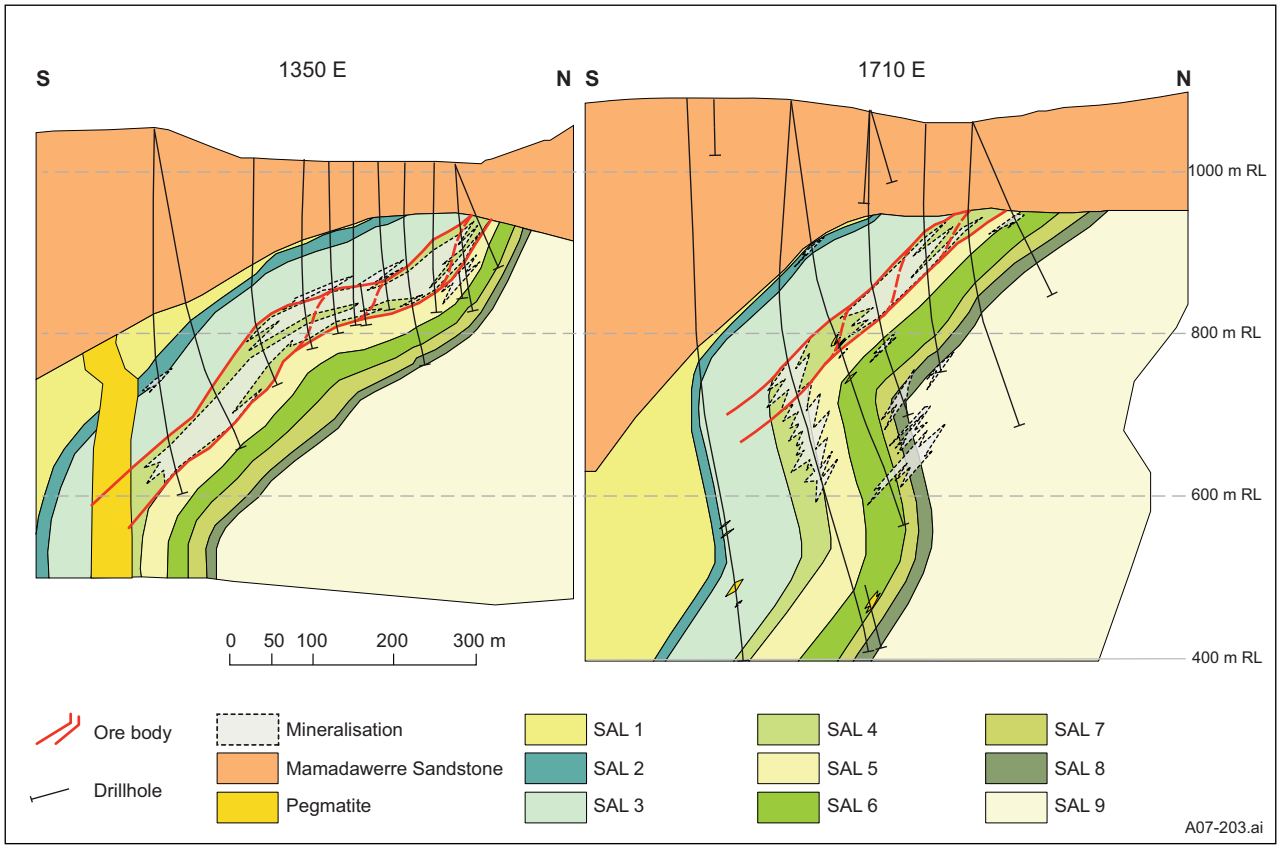


Figure 5.61. South–north cross-sections though Jabiluka 2, showing extent of mineralisation and interpreted ore-controlling structures (modified from Hancock *et al* 1990). The succession is divided into 9 'stratigraphic assay levels' (SAL) on the basis of mineralogy and uranium concentration (for explanation see **Table 5.8**). Uranium mineralisation occurs mostly within SAL 4 and SAL 8.

Koongarra 1 and 2

Noranda Australia Ltd discovered these two adjacent deposits in 1970, as a result of follow-up ground exploration of an airborne radiometric anomaly. Drilling defined a northwest-striking zone, over 650 m in length that contained two ore lenses, named Koongarra 1 and Koongarra 2 (**Figure 5.62**). The deposits were sold to Denison Mines Australia Ltd in 1980 and in the same year, the mine lease area was excised from Kakadu National Park. Cogema Australia Pty Ltd acquired 70% of the project in 1993 and the remaining 30% in 1995 and are the current owners. The total resource at Koongarra is estimated to be 0.812 Mt of ore at 1.102% U_3O_8 (AREVA 2008). An additional gold reserve at Koongarra 1 was also delineated at 1.04 Mt of ore at 3 g/t Au (Snelling 1990).

The orebodies are hosted within the lower Cahill Formation, which comprises quartz-chlorite schist, mica-quartz-feldspar schist, garnet-mica-quartz schist, graphite-mica-quartz schist and amphibolite. These rocks are in reverse-faulted contact with the Mamadewerre Sandstone (**Figure 5.62**). Most uranium mineralisation is within quartz-chlorite schist (± illite, garnet, sillimanite and muscovite), which also contains well banded siliceous lenses up to 5 m thick. The latter are interpreted as silicified carbonate (Snelling 1990). Pegmatite veins are common.

The primary mineralisation comprises uraninite, which occurs in hard crystalline form, or as a sooty amorphous mass. Pyrite, along with traces of galena and chalcopyrite, is present in high-grade portions, but only in minor amounts. The bulk of the ore is disseminated uraninite occurring as thin wisps, streaks and discontinuous strings, parallel to and crosscutting S_2, and parallel to phyllosilicate cleavage planes. Oxidation and alteration of uraninite within the primary ore zone has produced a variety of secondary uranium minerals dominated by the uranyl silicates, kasolite, sklodowskite and uranophane. These minerals occur mainly as *in situ* replacements of uraninite, which is most pronounced near the breccia zone of the Koongarra Fault. Rare grains of native gold, clausthalite, gersdorffite-cobaltite and mackinawite also occur (Snelling 1990). The orebody consists of a number of parallel lenses, up to 15 m thick, which coalesce towards the surface and become narrower with depth (**Figure 5.62**). The high-grade mineralisation occurs immediately beneath the graphite-quartz-chlorite schist.

An outer alteration halo extends up to 1.5 km perpendicular to the reverse fault and includes replacement of biotite by chlorite, rutile and quartz, and of feldspar by sericite. Metamorphic muscovite, garnet, tourmaline, magnetite, pyrite and apatite are preserved. Silicification of fractured Mamadewerre Sandstone in the footwall of the Koongarra fault is also associated with this alteration. An inner alteration halo occurs less than 50 m from primary ore. It involves replacement of quartz by chlorite and phengite, and of garnet by chlorite.

Other significant occurrences.

There are a number of other uranium prospects in the East Alligator River area. These are briefly described below. A complete listing is provided in MODAT.

SAL unit – North Ltd succession	Pancontinental succession	Schist composition	Thickness	Mineralisation
1	USS – Upper Schist Series	CM, CM-I, CIM	0–20 m, usually <5 m	Never mineralised
2	UGS– Upper Graphite series	CG (20 m) CIM (5 m) CG/CI (2–4 m)	15–25 m	U throughout unit, more concentrated near upper and lower boundaries. Grade variable and discontinuous
3	HWSS – Hangingwall Schist Series	CM (2 m) CMI-CM-CI (15 m) CM (8–15 m) CI (4–12 m) MC (5 m) CIM CI (3–6 m)	Highly variable, 40–70 m true thickness	Thin graphite bands can be weakly mineralised, local economic Au. 4–12 m thick CI band can be weakly mineralised.
4	MMS – Main Mine Series	CG (3 m) CI (3–6 m) CIM (10–20 m) Bx (up to 5 m) C-G (12–30 m) CM (0.3–2 m) C-G (1–4 m)	Highly variable, 20–60 m true thickness	Contains 70% of global resource. Top 6–12 m contains U and rarely Au-Pd, not brecciated. Chlorite-rich CIM unit commonly brecciated and well mineralised. BX mineralised with U and Au. C-G units can be well mineralised
5	FWS – Footwall Schist Series	CM/CMI/MC (up to 24 m) CMI-CI	10–30 m	Pitchblende veinlets, < 2 mm wide, cut schist randomly
6	LMS1 – Lower Mine Series 1	CI, CI(B)	10–80 m	U throughout unit, generally in upper 1/4, no Au
7	LSS1 – Lower Schist Series 1	CMI, CM, MC	5–30 m	Minor mineralisation in more chlorite- or sericite-rich sections
8	LMS2 – Lower Mine Series 2	C/CI, C, CI	Highly variable, 2–80 m	Can be mineralised throughout, most in upper 2/3, can be high grade and continuous. No Au
9	LSS2 – Lower Schist Series 2	MC, A, CM, CMI	Minimum 400 m	Rare mineralisation

Table 5.8. Subdivision of lithological units in mine sequence at Jabiluka (after Lally and Bajwah 2006). Rock type key: B = biotite, Bx = breccia, C = chlorite, G = graphite, I = sericite, M = muscovite; presence of quartz is always assumed. Order of letters in code gives decreasing order of abundance of mineral species. SAL = Stratigraphic Assay Level (from Gustafson and Curtis 1983, Poilto *et al* 2005).

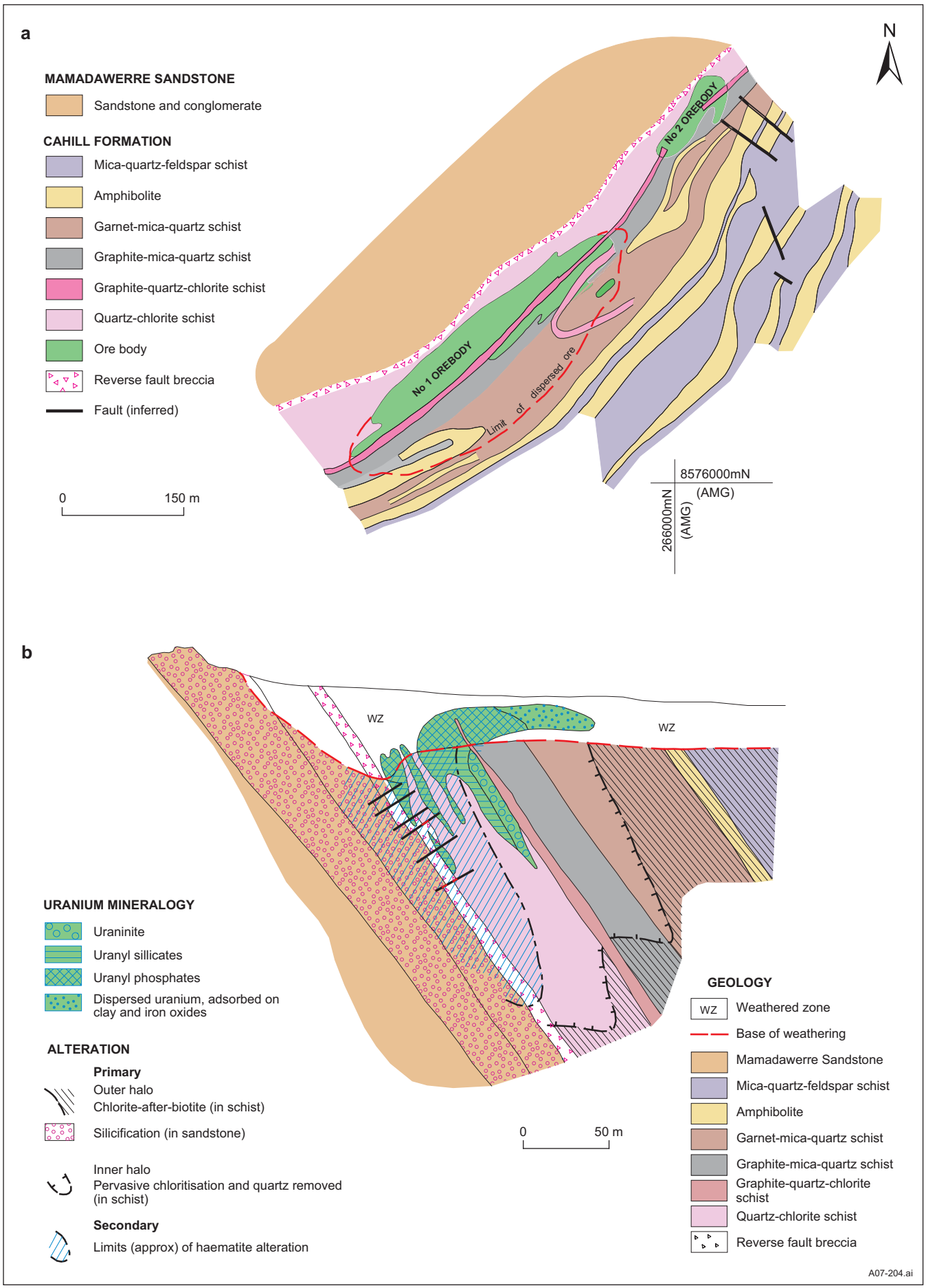


Figure 5.62. (**a**) Geological plan and (**b**) cross-section of Koongarra uranium deposits. Geology and outline of orebodies are projected to surface (modified from Snelling 1990).

Ranger 68 was discovered by Geopeko Ltd in 1976. It is hosted in the lower Cahill Formation, near the contact with the Nanambu Complex. Resources were estimated at 1.5 Mt of ore at 0.357% U_3O_8, using a cut-off of 0.1% U_3O_8 (Browne 1990). The primary uranium mineral is sooty pitchblende within intensely chloritised breccia zones. Copper is also locally associated with high uranium grades, occurring as sooty chalcocite. The best intersections were 5.75 m at 1.1% Cu in drillhole S/1 and 12 m at 0.26% Cu in hole S/4.

The *Caramal* prospect is located 22 km south of Nabarlek, adjacent to an escarpment of Mamadawerre Sandstone. It was discovered in 1969 by Queensland Mines Ltd during ground follow-up exploration of an airborne radiometric survey. A 'resource' of 2500 t U_3O_8 (0.758 Mt ore at 0.33% U_3O_8) was reported in McKay and Miezitis (2001). Mineralisation occurs within altered metasedimentary schist and carbonate rocks. Primary uranium mineralisation occurs within an elongate zone, about 80 m wide in plan view, which trends northeast. The schist is variably altered and the most extensive alteration involves the replacement of biotite by chlorite and feldspar by sericite, identical to the 'outer alteration halo' described from other deposits of the East Alligator River region (McKay and Miezitis 2001).

At the *Hades Flat* prospect, mineralisation is in chlorite-feldspar schist, graphite schist and dolostone of the lower Cahill Formation. A resource of 726 t U_3O_8 was quoted in the draft Environmental Impact Statement for Jabiluka (McKay and Miezitis 2001). Pitchblende is the main mineral and occurs in fractures and breccia zones in chlorite schist and in chloritised breccia zones at the faulted contact with the Nanambu Complex. Minor secondary uranium mineralisation occurs in the weathering zone above the prospect (Hegge *et al* 1980).

The *Austatom* prospect is hosted within dolostone and schist of the lower Cahill Formation. A contact between this unit and the Nanambu Complex is 800 m east of the prospect. The best auger hole intersections reported by the Australian Atomic Energy Commission (AAEC) were 1.5 m at 0.48% U_3O_8 and 6 m at 0.21% U_3O_8 (Needham 1982). Anomalous copper, lead, nickel, cobalt and manganese were also recorded. A resource of about 10 000 t U_3O_8 has been quoted for Austatom in the past (Needham 1988a), but is not considered reliable (Lally and Bajwah 2006).

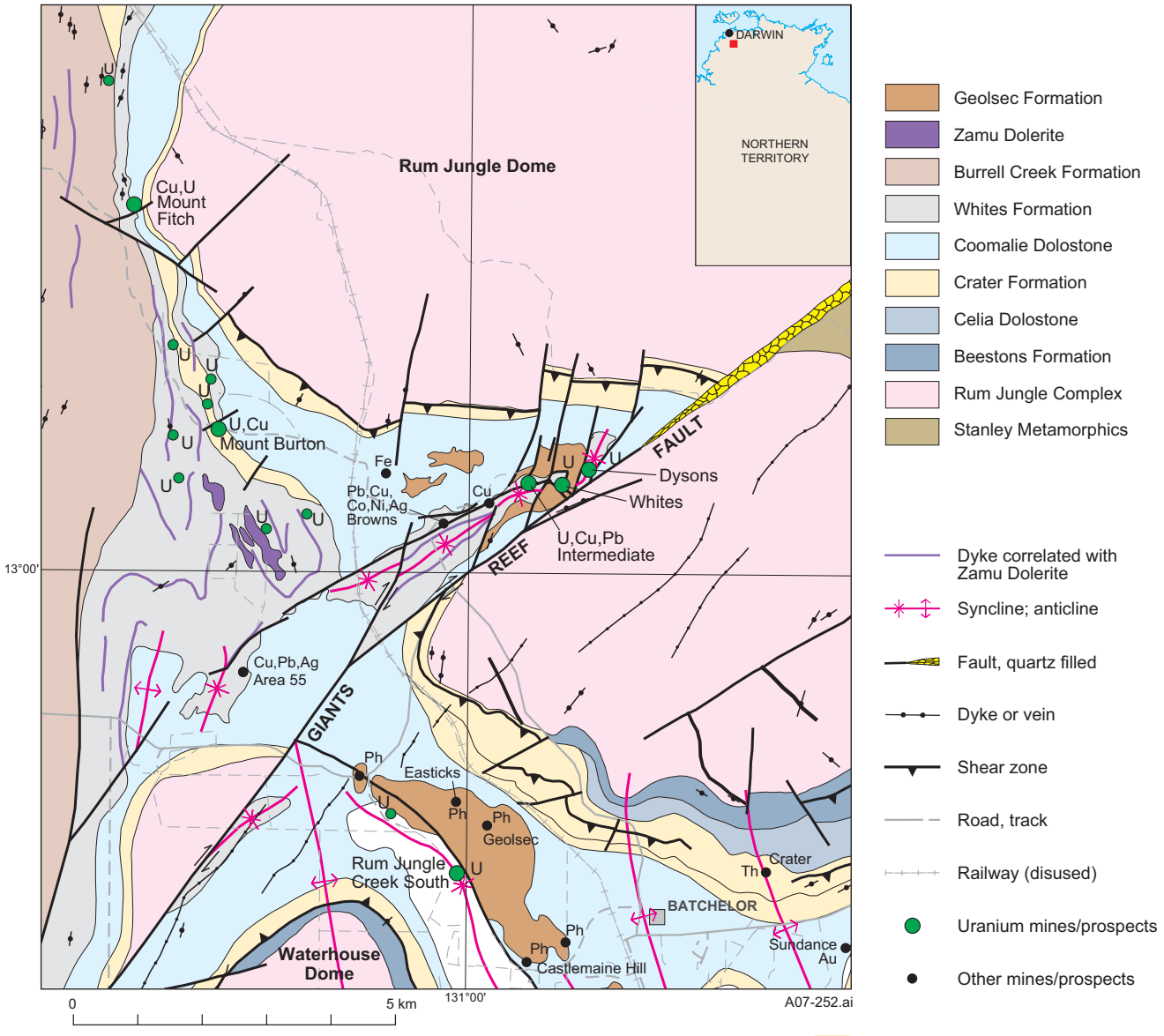


Figure 5.63. Geology of the Rum Jungle Mineral Field showing uranium and other mineral occurrences (modified from Lally 2002).

Rum Jungle Mineral Field

In 1949, JM White found secondary uranium minerals near Rum Jungle Siding, at the site of the now mined-out Whites deposit. Subsequent exploration by BMR led to the discovery of Whites and the nearby Dysons, Intermediate, Rum Jungle Creek South, Mount Burton and Mount Fitch deposits, as well as several other smaller occurrences (**Figure 5.63**). Mining of these deposits was carried out between 1954–1970 by Territory Enterprises Pty Ltd (TEP), a wholly owned subsidiary of Rio Tinto Australia Ltd. Other companies active in the area in 1970s and 1980s included Aquitaine Australia Minerals Pty Ltd, Queensland Mines Ltd, Mobil Energy Minerals Australia Inc, Marathon Petroleum Company, Rio Tinto Australia Ltd, Uranerz Australia Ltd and Afmeco Mining Ltd. Following a recent surge in uranium prices, the Rum Jungle Mineral Field is currently being explored by a number of uranium explorers.

With the exception of Rum Jungle Creek South, the major Rum Jungle Mineral Field uranium deposits occur within a triangular area known as the Embayment (**Figure 5.63**). Its southeastern boundary is defined by the Giants Reef Fault and the northern boundary by a series of east-trending ridges of the Crater Formation. Five deposits are known from this area; from east to west, these are: Dysons (U), Whites East (U), Whites (U, Pb, Cu, Co), Intermediate (U, Cu) and Browns (Pb, Cu, Ni, Co, Zn). They are spaced 0.5 to 1 km apart and lie on a northeast-trending shear zone. These deposits (and the shear zone) are located on the northern limb of a northeast-trending, southwest-plunging asymmetric syncline, with a shallowly dipping northern limb and a steeply dipping southern limb. North- to northeast-trending faults cut across the area and the Dysons, Whites and Intermediate deposits are located at the intersection of the shear zone and north-trending faults. All deposits are within the Whites Formation, close to the contact with the underlying Coomalie Dolostone. Most deposits are overlain by quartz-haematite breccia assigned to the phosphate-rich Geolsec Formation, which is overlain by the Depot Creek Sandstone. Carson *et al* (2010) obtained a SHRIMP U-Pb zircon age of 1837 ± 15 Ma for the Depot Creek Sandstone, interpreted as a conservative maximum age of deposition. Thus it may represent, in part, a stratigraphic equivalent of the Katherine River Group in the East Alligator River region.

Rum Jungle Creek South and Dysons are the only deposits without reported base metals sulfides. All other deposits in the Embayment area contain associated copper minerals, and Ni and Co may also be present, as in the Browns, Intermediate and Mount Fitch deposits.

Rum Jungle Creek South

Rum Jungle Creek South is hosted by carbonaceous shale of the Whites Formation, just above the contact with the underlying Coomalie Dolostone. The deposit lies to the southwest of an outlier of the Geolsec Formation that hosts small phosphate occurrences. The uranium orebody is located within the hinge zone of a doubly-plunging syncline that trends 315° (**Figure 5.64**) within chloritic and carbonaceous shale of the Whites Formation (Berkman 1964). The orebody is 245 m long, 61 m wide and 80 m deep. Open-cut mining during 1961–1963 produced a total of 663 500 t of ore averaging 0.43% U_3O_8 and a further 115 800 t averaging 0.065% U_3O_8 was stockpiled (Berkman 1968). Mineralisation is preferentially located in chlorite schist. Towards the northeast, the Whites Formation and Coomalie Dolostone are faulted against the Geolsec Formation, which lies directly on the Coomalie Dolostone, suggesting a period of erosion and uplift prior to

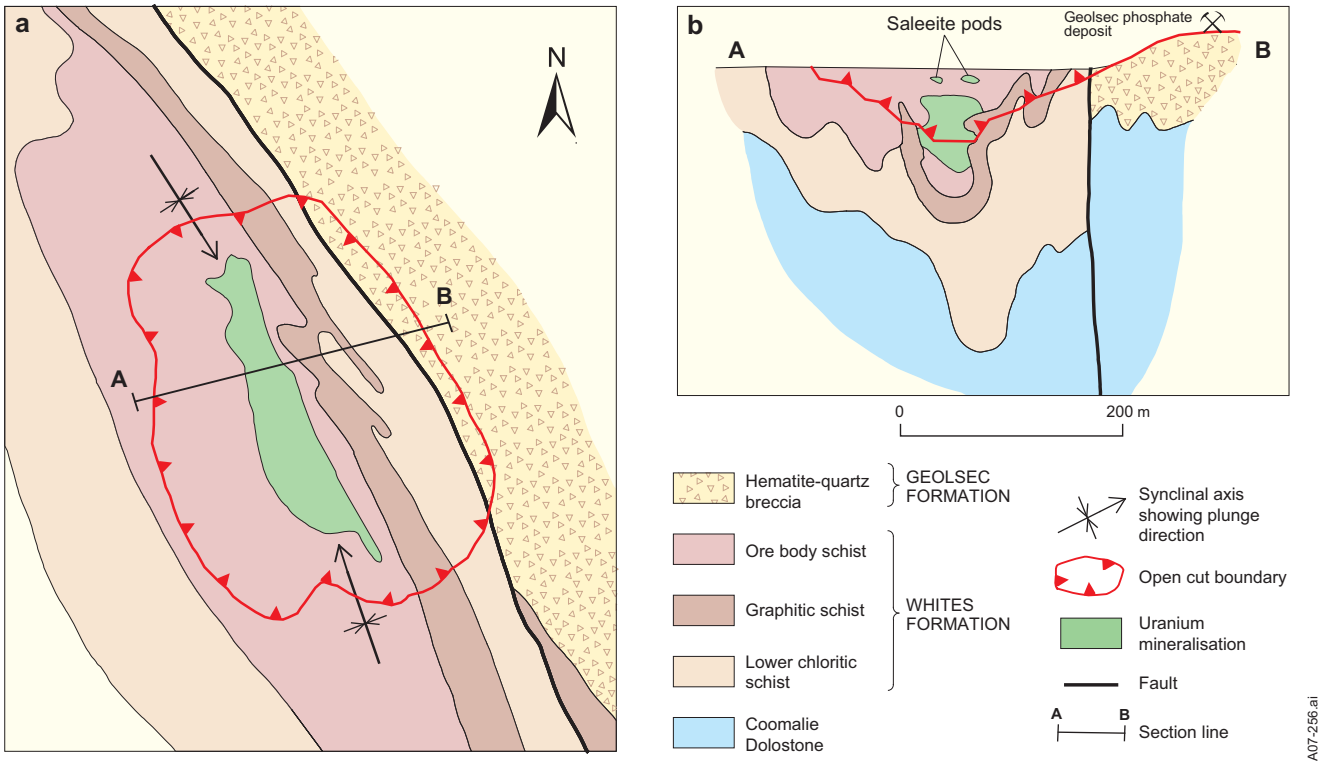


Figure 5.64. (**a**) Geological plan and (**b**) cross-section of Rum Jungle Creek South deposit (modified from Berkman 1968).

deposition of the Geolsec Formation. Primary ore consists of pitchblende, occurring as fine sooty coatings on foliation planes and joints, and as thicker veins near the contact between chlorite schist and black slate. Apart from pyrite and trace amounts of sphalerite, no sulfides were associated with the uranium mineralisation. The weathered zone to 30 m depth was almost devoid of mineralisation, except for three small near-surface pods of saleeite that were the source of surface radiometric anomalies over the deposit.

Dysons

The Dysons orebody is 60 m long, 8 m wide and 100 m deep, and is hosted within a narrow zone of strongly sheared carbonaceous shale of the Whites Formation at its contact with the Coomalie Dolostone. Open-cut mining during 1957–1958 produced a total of 157 000 t of ore at 0.34% U_3O_8 (Berkman 1968, Fraser 1980). The Giants Reef Fault occurs about 200 m to the south of the mine and four north-northeast-trending faults were mapped in the open cut (**Figure 5.65**). The orebody is partly covered by the Geolsec Formation, which unconformably overlies the Coomalie Dolostone. Saleeite is the predominant uranium mineral to 40 m depth, but pitchblende occurs in fresh carbonaceous slate from 25 m downward. Sklodowskite and uranosphaerite are also present in the near-surface weathered zone (Berkman 1968). There is no reported base metals mineralisation and the only sulfide present is pyrite.

Whites

The Whites orebody comprises disseminated uranium and copper mineralisation in a steeply south-dipping zone, 15–25 m wide and about 120 m long. Underground mining in 1953 was followed by open-cut mining during 1954–1958, producing a total of 396 000 t ore averaging 0.27% U_3O_8 and 2.7% Cu. In addition, 295 000 t of copper-cobalt ore averaging 2.8% Cu and 0.3% Co was also treated (Berkman 1968). Mineralisation is hosted by black shale of the Whites Formation, near to its contact with the Coomalie Dolostone (**Figure 5.66**). It is confined to a south-dipping zone, bordered to the west by a northeast-striking, south-dipping fault and to the east by graphitic shale. The orebody can be divided, from footwall to hangingwall, into a U-Cu zone, a Cu-Co zone, a Co-Ni zone and a Pb-Co zone. These zones occasionally overlap and are sometimes separated by a band of barren material (Berkman 1968).

In the U-Cu zone, the common ore minerals are pitchblende and associated secondary uranium minerals. In the Cu-Co zone, the main minerals are chalcopyrite, chalcocite, digenite, bornite and covellite. In the Co-Ni zone, the common minerals are linnaeite, carrollite, and gersdorffite. In the Pb-Co zone, galena and linnaeite are the common minerals. Pitchblende is known to continue to a depth of 180 m, and the Cu-Co and Pb-Co zones continue to more than 300 m depth, but become very narrow.

Whites Extended and Whites East

These deposits are situated midway between the Dysons and Whites deposits. Mineralisation is within pyritic carbonaceous shale of the Whites Formation, extending into the Haematite Quartz Breccia (HQB). The Whites Extended mine produced 100 tonnes of ore at 0.185% U_3O_8 (Berkman 1968). Only secondary uranium minerals were present, occurring as narrow high-grade intersections in slate.

Other uranium occurrences in the Rum Jungle Mineral Field

At the *Mount Burton* prospect, costeaning, geological mapping, geophysical surveys and drilling by TEP and BMR were conducted during the 1960s. Small-scale open-cut mining of this deposit occurred in 1958, producing 6096 t ore averaging 0.21% U_3O_8 and 1.04% Cu, and an additional 2440 t ore at 0.07% U_3O_8 and 1420 t ore at 2.66% Cu was stockpiled (Berkman 1968). The old open cut reached a depth of about 30 m. The uranium-copper ore is within

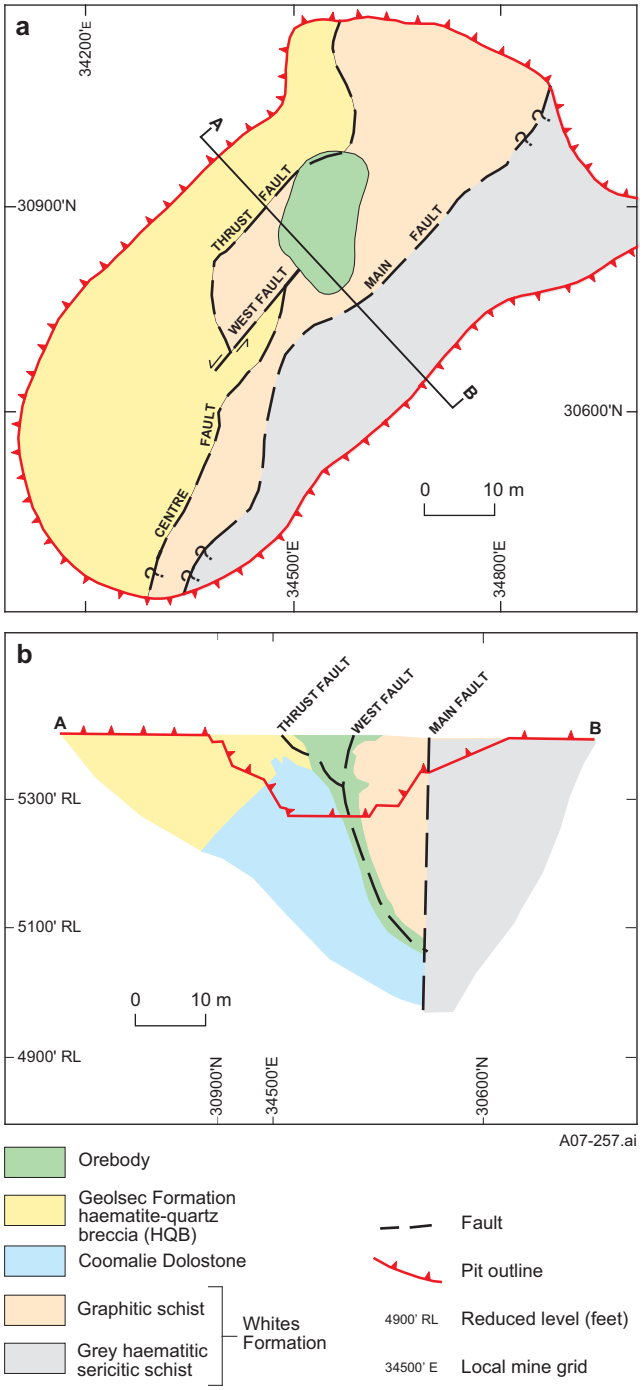


Figure 5.65. (**a**) Geological plan and (**b**) cross-section of Dysons deposit (modified from Berkman 1968).

black shale and quartzite of the Whites Formation, close to the contact with the underlying Coomalie Dolostone. This contact dips at about 40° to the west and steepens at depth. The orebody is restricted to the crest of an anticline. Ore grade diminishes rapidly below a depth of about 30 m. Compass Resources NL currently holds titles over the area and commenced exploration in 1990, but early drillholes failed to intersect any mineralisation.

The *Mount Fitch South* deposit contained a resource of 3.5 Mt of ore at 0.042% U_3O_8 and 290 000 t ore at 0.65% Cu (Berkman and Fraser 1980). It was explored by TEP during 1952–1970 and by Uranerz Australia Pty Ltd in the late 1970s (Pagel *et al* 1984). The leases are currently held by Compass Resources NL, who has estimated a JORC-compliant Resource of 5.03 Mt at 0.03% U_3O_8 (Compass Resources 2008). Uranium mineralisation is present in both the Coomalie Dolostone and Whites Formation.

The *Kylie, SE Kylie* and *Spring Creek* prospects occur within the Coomalie Dolostone along the southern edge of the Waterhouse Dome. In this area, the Coomalie Dolostone contains numerous lenses of graphite, chlorite, tremolite, tourmaline and biotite-rich metapelite at various levels (Pagel *et al* 1984, Williams 1996).

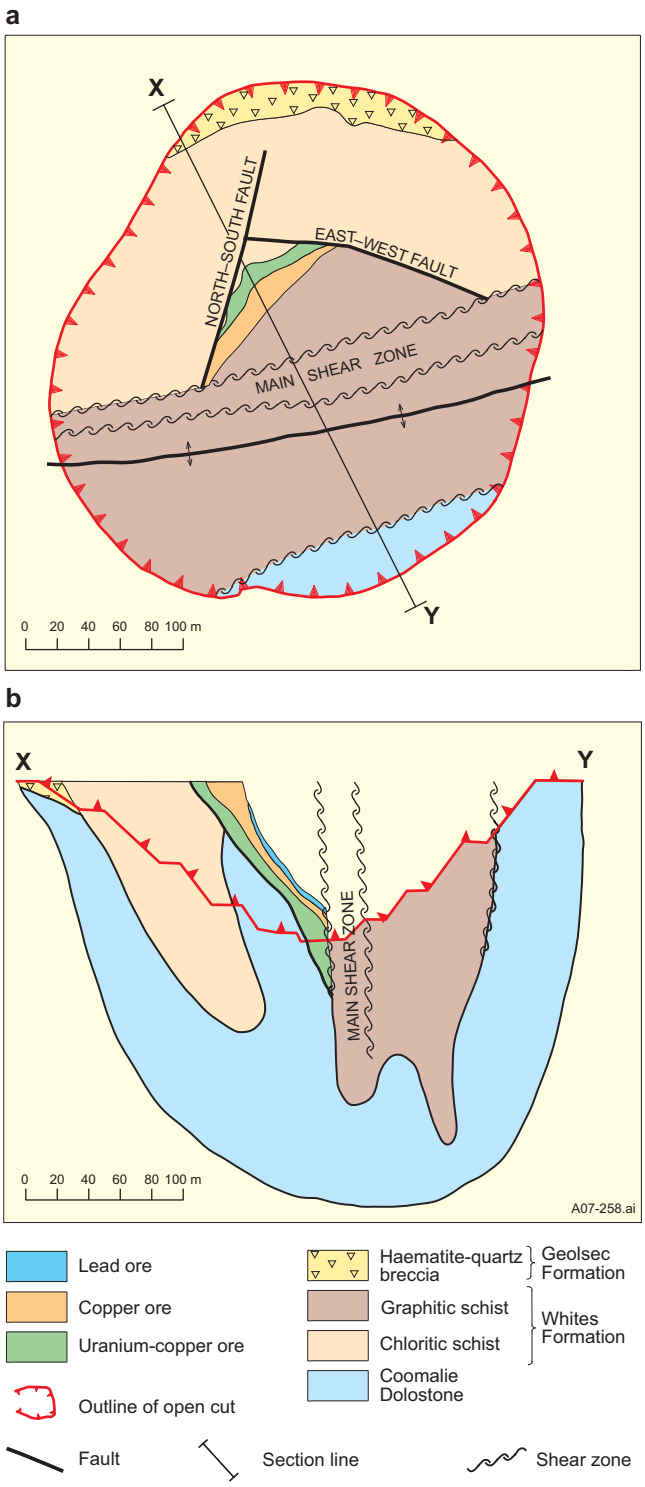


Figure 5.66. (**a**) Geological plan and (**b**) cross-section of Whites deposit (modified from Berkman 1968).

South Alligator Valley Mineral Field

A total of 13 small deposits and 15 uranium prospects (**Figure 5.67**) are known from this area (McKay and Miezitis 2001). Thirteen of these deposits were mined by United Uranium NL between 1955 and 1964, producing some 875 t of U_3O_8 from 146 500 t of ore averaging 0.12–2.5% U_3O_8 (Fisher 1968). The bulk of production was from underground operations, with a few small open cuts. From 1956–1959, ore was trucked to Rum Jungle for processing. In 1959, a gold processing plant at Moline was converted to treat uranium ore by solvent extraction. A small treatment plant also operated at Rockhole during 1959–1962. Gold was also produced as a by-product of uranium mining, and from the re-treatment of uranium tailings. Uranium exploration in the area increased again in the early 1970s and identified many prospects and anomalies, but no economic deposits were found. Further exploration by the Coronation Hill Joint Venture (CHJV), from 1984 to 1989, defined a gold-platinum-palladium resource at Coronation Hill.

During 1987–1990, BMR investigated various aspects of the geology and mineralisation of an area known as the Conservation Zone. In 1991, the final report from these studies was presented to the government. It was decided that mineral exploration and mining would not be permitted within the Conservation Zone, which was subsequently included within Stage 3 of Kakadu National Park.

Uranium deposits in the South Alligator Valley are located within the major northwest-trending dextral strike-slip Rockhole-Palette Fault System, at or close to the unconformity between the El Sherana and South Alligator groups. Sandstones of the Mamadawerre Sandstone of the Katherine River Group are present on both sides of the South Alligator Valley. Almost all of the mineralisation is located within fault zones containing either a contractional jog (eg Rockhole mines), or a dilational jog (eg Palette), producing sub-horizontal ribbon-like and sub-vertical pipe-like bodies, respectively (Valenta 1991). Primary uranium mineralisation occurs in fractured, cherty ferruginous siltstone of the Koolpin Formation, near the unconformity with the Coronation Sandstone. The Coronation Hill, Sleisbeck and Anomaly 2J deposits have different host rocks. Orebodies are either sub-horizontal and ribbon-shaped (El Sherana, Saddle Ridge), or subvertical and pipe-shaped (Palette, Skull, Coronation Hill), depending on the geometry of the dilational zone. Alteration related to mineralisation is characterised by muscovite-chlorite ± haematite ± kaolinite ± biotite (Wyborn *et al* 1990a, b, Lally and Bajwah 2006).

Rockhole Group

The Rockhole group of mines (*Rockhole No 1* and *No 2*, *O'Dwyers*, *Sterrets No 1* and *No 2)* are located about 6 km northwest of El Sherana Camp and have produced 152 t U_3O_8 from 13 418 t of ore averaging 1.1% U_3O_8 (Taylor 1968, Lally and Bajwah 2006).

The mineralised lodes strike northwest (300°) and dip steeply (70°) to the southwest. These lodes occupy dilatant sites along reverse faults, where carbonaceous shale and cherty ferruginous shale of the Koolpin Formation form the hangingwall, and quartz sandstone (Coronation Sandstone) and altered felsic volcanic rocks (Pul Pul Rhyolite) comprise the footwall.

The primary ore mineral, pitchblende, occurs in two forms: an early massive type and later cross-cutting veinlets that are associated with clausthalite and sulfides (Threadgold 1960). Secondary uranium minerals are rare. Minor clausthalite, eskebornite, pyrite, marcasite and chalcopyrite occur, along with native gold. Trace element analyses of ore and wallrocks show strong associations between uranium and vanadium, copper, nickel, cobalt and gallium. Carbon values in the Koolpin Formation shale are erratic and independent of uranium (Ayres and Eadington 1975).

Valenta (1991) interpreted the main fault at Rockhole to be a restraining bend on the larger Rockhole-El Sherana-Palette dextral strike-slip fault system, resulting in vertical extension and the formation of dilational zones near gently dipping structures. These dilational zones were the focus of mineralising fluids, and this model explains the geometry of the ore lodes in relation to faulting.

El Sherana

El Sherana and *El Sherana West* were mined by small open-cut and underground operations over a strike length of 400 m and a vertical range of 180 m (Taylor 1968). The total production was 226 t U_3O_8 and 0.55 t Au from El Sherana, and 185 t U_3O_8 and 0.007 t Au from El Sherana West (Lally and Bajwah 2006).

Mineralisation is associated with fault zones and brittle fracturing within cherty ferruginous shale, and carbonaceous shale of the Koolpin Formation. Minor mineralisation also occurred within the Coronation Sandstone. The two dominant fault orientations are steeply

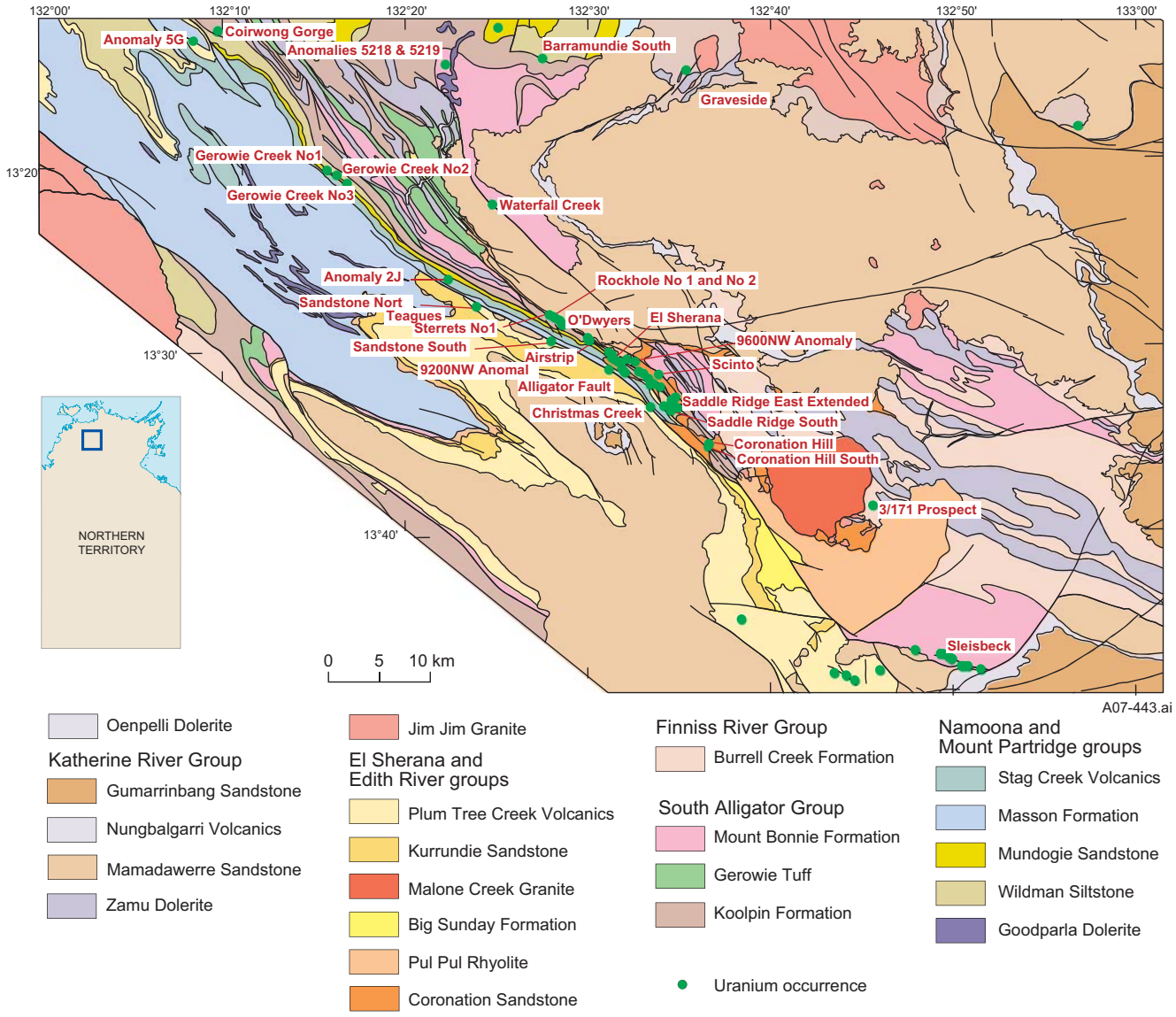


Figure 5.67. Geology and uranium occurrences of South Alligator Valley Mineral Field (modified from Needham 1988b).

southwest-dipping normal faults and later, moderate to shallowly southwest-dipping reverse faults (Valenta 1991). Faulting created a sheared and brecciated contact between carbonaceous and cherty ferruginous shale that was the focus of mineralisation.

Pitchblende is the main uranium mineral. Native gold and minor galena, anglesite, clausthalite, pyrite, marcasite, cobalt-nickel arsenides, nickel selenide and copper sulfides also present (Threadgold 1960, Ayres and Eadington 1975).

Coronation Hill

Two significant orebodies (uranium-gold and gold-PGE) are located at Coronation Hill, but only the uranium-gold orebody has been mined. From 1961–1963, this deposit produced 75 t U_3O_8 from 28 850 t of ore averaging 0.26% U_3O_8 (Foy 1975). Gold is associated with uranium mineralisation and an average grade of 6.1 g/t Au was reported during early reserve definition work (Prichard 1960), but gold production figures are unknown. BMR calculated a high-grade uranium-gold resource at a depth of 120 m below the old open cut of 344 170 t averaging 0.537% U_3O_8 (1850 t U_3O_8) and 9.95 g/t Au (McKay 1990, Lally and Bajwah 2006).

Uranium-gold mineralisation occurs in a vertical pipe-like body about 20 m across (**Figures 5.41**, **5.42**). Mineralisation is associated with faulted blocks of carbonaceous shale (Koolpin Formation) near the base of the pit and with carbonaceous shale clasts in debris-flow conglomerate (Needham 1987). Mineralisation consists of disseminated and sooty pitchblende with minor secondary uranium minerals at the surface. Veinlets and disseminations of native gold occur in association with pitchblende and as gold-only shoots (Needham 1987).

The Coronation Hill gold-PGE orebody (see **Gold in association with platinum group elements**) is located to the east of the mined uranium-gold orebody. It occurs in a broadly north-northwest-trending zone in fractured and sheared volcaniclastic rock, quartz-feldspar porphyry, sedimentary breccia and diorite (Carville *et al* 1990). The stratigraphic position of this succession is uncertain but Ferenczi and Sweet (2005) considered it to be part of the Coronation Sandstone.

Other uranium deposits in the South Alligator Valley Mineral Field

The *Koolpin Creek* deposit has produced 2327 t of ore averaging 0.13% U_3O_8 from a small open cut. The orebody is hosted within a northwest-trending shear zone in black shale and phyllite of the Koolpin Formation.

The *Scinto 5* and *6* deposits have produced 2.7 t and 21.7 t of U_3O_8, respectively. *Scinto 5* is hosted in steeply southwest-dipping shale and phyllite of the Koolpin Formation. Uranium mineralisation consists mainly of torbernite, with minor pitchblende (Stuart-Smith *et al* 1988). Mineralisation at *Scinto 6* is hosted in sheared and altered rhyolite occurring within the Coronation Sandstone. Ore consists of autunite, torbernite and saleeite in a subvertical shear zone and adjacent joints (Stuart-Smith *et al* 1988).

At the *Palette* deposit, mineralisation occurs at the unconformable contact between cherty ferruginous shale and carbonaceous shale of the Koolpin Formation and altered sandstone of the Coronation Sandstone. It produced 124 t U_3O_8 at a grade of 2.5% U_3O_8. Pitchblende occurs as veins and sooty coatings in shears and fractures, and high-grade ore shoots are composed of massive veins and nodular pitchblende (Taylor 1968, Stuart-Smith *et al* 1988). Native gold occurs as veinlets, shrinkage cracks and disseminations in the pitchblende (Taylor 1968). Minor amounts of galena-clausthalite, coloradoite, pyrite and marcasite are also associated with the ore (Ayres and Eadington 1975). Fine haematite, sericite and minor tourmaline are present in wall rocks.

The *Skull* mine produced 531 t of ore at a grade of 0.55% U_3O_8 (Foy 1975). Mineralisation is within a zone of northeast-trending fractures, associated with a north-northeast-trending dextral strike-slip fault, at the contact between the Koolpin Formation and Mamadawerre Sandstone (Valenta 1991). Pitchblende is present as veinlets and as a sooty variety within carbonaceous shale of the Koolpin Formation and was associated with minor gold and copper sulfides (Ayres and Eadington 1975).

The *Saddle Ridge* deposit produced 78 t at a grade of 0.24% U_3O_8 (Foy 1975). Mineralisation is comprised entirely of secondary uranium minerals, mainly metatorbernite, which is hosted by bleached shale of the Koolpin Formation. Shepherd and Grenning (1961) reported occurrences of sooty pitchblende in drillholes below the main orebody. No gold was recovered from ore in the open cut (Lally and Bajwah 2006).

Sleisbeck is an isolated occurrence, located 30 km southeast of Coronation Hill in the headwaters of the Katherine River. A small open-cut operation extracted 637 t of ore averaging 0.34% U_3O_8 in 1956. Mineralisation is within the Mount Bonnie Formation (formerly Kapalga Formation in this area; Ferenczi and Sweet 2005), adjacent to a major reverse fault, with the Mamadawerre Sandstone and Scinto Breccia in its footwall. Pitchblende occurs in rocks at depth, but most of the ore consists of uraniferous phosphatic minerals near the surface. Minor copper, nickel and arsenic minerals have also been recorded (Stuart-Smith *et al* 1988).

Geochronology of unconformity-type uranium deposits

Results of geochronological investigations of the major deposits in the East Alligator River area are summarised in Lally and Bajwah (2006). Pegmatite at Ranger 1 intruded at 1847 Ma (Annesley *et al* 2002). Pre-mineralisation retrograde metamorphism during the period 1810–1750 Ma is recorded by the resetting of Rb-Sr, Ar-Ar and K-Ar systems, in samples of unaltered rocks from Jabiluka and Nabarlek (Page *et al* 1980, Riley *et al* 1980, Polito and Kyser 2002). Polito and Kyser (2002) dated retrograde Fe-rich chlorite in the outer alteration zone at Nabarlek at 1750 Ma, which they interpreted as being due to early alteration along faults by fluid inflow from overlying aquifers in the Mamadawerre Sandstone. This 1750 Ma event is within error of the age of primary uranium mineralisation at Ranger 1 (1737 ± 20 Ma), determined from U-Pb dating of uraninite by Ludwig *et al* (1987). Ar-Ar ages of 1696–1715 Ma from sericite in altered schist at Nabarlek (Polito and Kyser 2002) were interpreted as being due to

resetting by localised heating during the intrusion of a sill of Oenpelli Dolerite at 1720 Ma beneath the deposit. Estimates of U-Pb uraninite ages of 1700 Ma and 1800–1700 Ma were obtained at Ranger and Koongarra, respectively, by Hills and Richards (1976). Rb-Sr, U-Pb, Pb-Pb and Sm-Nd studies of uraninite and inner-alteration-zone minerals in Jabiluka and Nabarlek produced consistent ages of primary mineralisation in the range 1650–1610 Ma (Page *et al* 1980, Gulson and Mizon 1980, Riley *et al* 1980, Maas 1989, Polito and Kyser 2002). Post-mineralisation remobilisation of uraninite at 1380 Ma and 1100 Ma is recorded by U-Pb, Pb-Pb and Rb-Sr systems in Jabiluka and Nabarlek (Gulson and Mizon 1980, Polito *et al* 2004). Younger remobilisation of uranium mineralisation at 900 Ma is indicated by U-Pb ages of uraninite from Ranger, Jabiluka, Nabarlek and Koongarra, as well as from deposits in the South Alligator River area (Hills and Richards 1972).

For the Rum Jungle Mineral Field, dating of pitchblende from the Kylie and Whites deposits has provided inconclusive results. U-Pb age determinations on whole-rock and mineral separates (pitchblende) of uraniferous ores from the Kylie deposit are given in Von Pechmann (1992). Whole rock samples yielded an age of 1627 ± 45 Ma. Mineral separate samples and two whole-rock samples were used to construct a discordia, the upper intercept of which gave an age of 1775 ± 175 Ma. Sm/Nd investigations on the Kylie ore gave an age of 469 ± 31 Ma, but the significance of this age is not clear (Von Pechmann 1992). It might indicate uranium mobilisation, coinciding with outpouring of the early Cambrian Kalkarindji Suite over most of the North Australian Craton (see **Kalkarindji Province**).

A single sample of pitchblende from the Whites deposit has yielded a wide range of ages from 1015 to 560 Ma (Fraser 1975). At the Whites East deposit, several whole-rock uranium-bearing samples were analysed and three samples provided a poorly defined discordia, with a concordia intercept at about 2200 Ma. Richards (1963) reported on the model lead ages of galena from the Whites and Browns deposits of 1520 Ma. A similar age has been obtained for several galena samples from the Browns deposit (McCready *et al* 2004). These samples included both coarse-grained vein and disseminated fine-grained galena. There is no distinction in the lead isotope values between these two galena types.

Genesis of unconformity-type uranium deposits

Several models have been proposed to explain the genesis of uranium deposits of the Rum Jungle area. Condon and Walpole (1955), Fraser (1975) and Berkman and Fraser (1980) suggested that they formed from syngenetic uranium precipitation with subsequent local remobilisation. This model is based on strong stratigraphical controls on the disposition of the major orebodies. Roberts (1960) studied the texture and mineralogy of the Whites orebody and proposed that the presence of tourmaline and quartz in the gangue suggest a hydrothermal origin for mineralisation that was related to the intrusion of the Rum Jungle granite (Rum Jungle Complex). The latter was then considered to be an intrusive, but has subsequently proved to be unconformable basement to the PCO succession. However, late, cross-cutting quartz-tourmaline veins, pegmatitic bodies and dolerite dykes, both within the Rum Jungle Complex and the overlying sedimentary rocks, testify to younger intrusive activity in the region. Condon and Walpole (1955) and Berkman (1968) suggested that carbonaceous shale within the Whites and Koolpin formations, in the Rum Jungle and South Alligator fields, respectively, acted as traps for fixing uranium during sedimentation and that the orebodies formed principally as a result of remobilisation of this syngenetic mineralisation during the Nimbuwah Event. Crick and Muir (1980) proposed that uranium was concentrated during the deposition of evaporitic sediments of the Coomalie Dolostone, and lower Cahill and Koolpin formations.

Eupene *et al* (1975) and Eupene (1980) proposed a hypogene model to explain the genesis of the uranium deposits of the Alligator River area, deriving the uranium from depth. They observed that the uranium mineralisation is intimately associated with Mg metasomatism, which produced the Mg-rich chlorite in the orebody and magnesitised the footwall dolomitic lithologies. The orebodies are located within zones of brecciation. Footwall carbonate rocks were considered to be dewatering channelways, both during metamorphism and metasomatism/mineralisation. During dewatering, clay particles within the carbonate rocks adsorbed the uranium and stored it to be released later (post-Kombolgie Subgroup), due to sediment loading. These uranium-rich fluids moved upward and deposited uranium in the overlying Cahill Formation.

Ferguson *et al* (1980) proposed that uranium was derived from the carbonaceous shale and was trapped in pre-Kombolgie dolines in a karstic topography. Psammitic and pelitic sediments filled these dolines. Insoluble Mg-rich clays and carbonaceous material were washed into these cavities and uranium was absorbed onto the clays. This was followed by the deposition of the Kombolgie Subgroup, which sealed the deposits.

Binns *et al* (1980b) conducted petrological, mineralogical and isotopic studies at the Jabiluka deposit and concluded that the hydrothermal fluid had a high initial temperature of 300–350°C. They noted that the orebody has a large Mg-chlorite-dominated alteration halo extending into the unconformably overlying Kombolgie Subgroup. Ore minerals show replacive relationships and there is a general lack of open-space-filling textures. Their model involves a uranium source in the underlying Archaean granite gneiss. Radiogenic heat from the granitic bodies resulted in a large fluid convection cell extending into the overlying Kombolgie Subgroup. Epeirogenic movement associated with the deposition of the Kombolgie Subgroup, or its fracturing, caused the initiation of fluid convection cells. The uranium deposits were interpreted to have formed at the upwelling zone.

Wilde (1988) and Wilde *et al* (1989a, b), on the basis of alteration mineralogy, suggested that uranium mineralisation resulted from descending hot (150–200°C) fluids sourced from the Kombolgie Subgroup into relatively reduced schist (Cahill Formation and equivalents). Supporting evidence is provided by the abundance of haematite, anatase and magnesium chlorite, indicating high-temperature fluids. The orebodies are within zones of desilicification, which

implies that the fluid was heating and that silica was removed. High temperatures are also supported by fluid inclusion studies by Ypma and Fuzikawa (1980) and Wilde *et al* (1989b). Polito *et al* (2004) provided additional support for this model, using alteration and geochronological and stable isotope data from the *Nabarlek* deposit. Ar-Ar dating of pre-ore sericite from the alteration envelope provided an age of ca 1700 Ma, and uraninite was dated at 1640 Ma. Reactivation of the Nabarlek Fault at 1360 Ma, 1000 Ma and 900 Ma was proposed to have reset uraninite ages.

Solomon and Groves (1994) synthesised fluid inclusion, stable isotope, mineralogical and alteration data, and proposed a fluid-flow model, initiated by radiogenic heat generation within the Archaean granitic basement. Fluid flow was strongly focused along faults and shear zones. Small-scale fluid convection during the first stage transferred some uranium from basement rocks into the cover sandstone (**Figure 5.68a**). In the second stage, large-scale fluid convection cells transferred the uranium along well defined pathways (faults, shear zones) to the site of uranium orebodies within the Cahill Formation (**Figure 5.68b**). Lally and Bajwah (2006) considered that the first stage involved in this model seems to be unnecessary, as uranium would continue to be leached from felsic basement rocks during the second stage. According to the model, deposits were connected with a surface brine reservoir by permeable fault zones, a scenario that appears unlikely at Jabiluka, where no large-scale fault has been identified in the overlying sandstone and at Nabarlek, where the total displacement on the basement fault is only about 20 m. Convection cells extending through the entire thickness of the cover sandstone succession are also unlikely, because aquicludes would develop in well sorted, fine-grained quartz sandstone interbeds during diagenesis (Polito *et al* 2004).

Ahmad *et al* (2006) proposed a model, in which uranium and other metals were leached by oxygenated groundwater from the uranium-enriched first-cycle basal sandstone succession (Kakadu Group equivalents) and Archaean basement. These fluids were then raised to the surface via faults and shears. Reduction by carbonaceous lithologies (eg Cahill, Koolpin, Whites formations) was essentially the main cause of uranium precipitation. Heat generation due to radioactive decay in the Archaean basement rocks may also have resulted in long-lasting convection cells.

Associated with the uranium mobilisation and precipitation is the widespread magnesitisation of the footwall dolostone. Magnesitisation of dolomite can occur either from an increase in temperature, or an increase in the Mg concentrations of the fluid. Both are viable mechanisms: heat generation might be due to radiogenic decay in the granitic rocks and an increase in Mg concentrations might be due to sericitisation of ferromagnesian minerals, such as chlorite, biotite and amphibole. Magnesitisation of dolomite results in net volume loss and consequent brecciation, which is common to most deposits (Ahmad *et al* 2006).

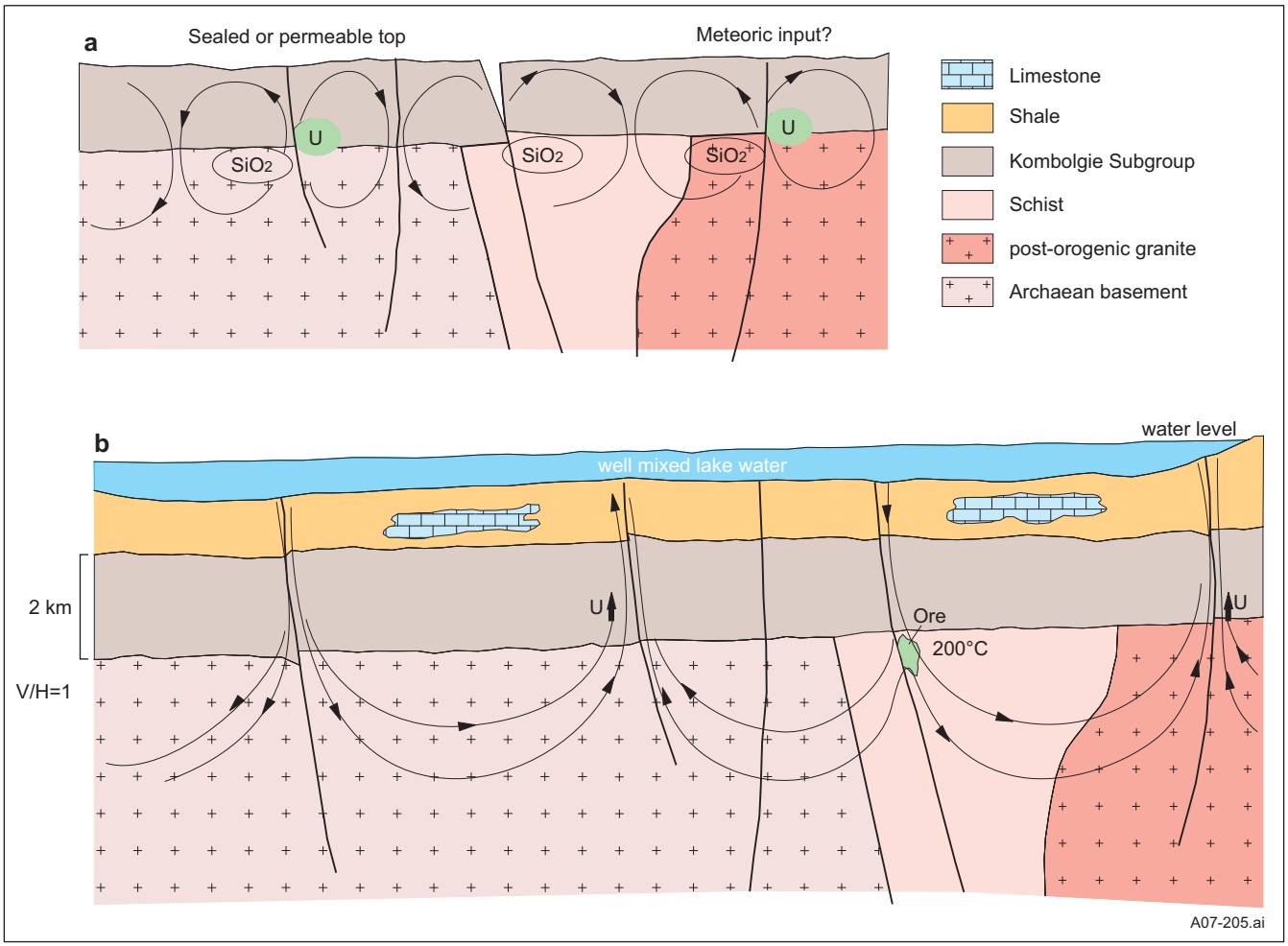


Figure 5.68. Model for the formation of unconformity-related uranium deposits in Australia (after Solomon and Groves 1994). See text for detailed explanation.

Vein-type uranium deposits

Small vein-type uranium deposits within the PCO are hosted by a variety of rock types and all are associated with narrow, steeply-dipping brittle fault or fracture zones. Quartz and haematite are common gangue minerals and minor copper mineralisation may also be present. Some of the larger vein type deposits were mined, yielding a total of about 20 t U_3O_8. Some of the more significant occurrences are briefly described below.

Adelaide River mine

Adelaide River mine is located 3.5 km northwest of Adelaide River township. The deposit was discovered in 1954 and worked until 1957, for a total production of 3447.4 t ore at an average grade of 0.5% U_3O_8. The ore was trucked to the Rum Jungle plant. Between 1959 and 1960, BMR drilled four diamond holes and calculated the remaining reserves to be 6350.4 t ore averaging 0.32% U_3O_8 (Plumb 1960, Fander 1981). Urangesellschaft Australia Pty Ltd carried out electromagnetic, magnetic and radiometric surveys over the Adelaide River mine area during 1979–1981 (Fander 1981). Uranium mineralisation is hosted within well bedded siltstone, greywacke and conglomerate of the Burrell Creek Formation. Four ore zones, the *Black*, *White*, *Brown* and *Orange* lodes, were identified. The largest of these is the Black Lode, which produced most of the ore, and it occupies a fault that dips 70° towards east with a reverse (east-side-up) movement sense (Walpole 1957). Mineralisation occurs only where the fault intersects a 15 m-thick bed of coarse-grained greywacke. The ore lode plunges 45° south and has an average width of approximately 1.5 m. Pitchblende is disseminated in country rock adjacent to the fault and forms coatings on joint and fracture surfaces associated with quartz veining. Accessory minerals include abundant pyrite and chalcopyrite, with lesser arsenopyrite, marcasite and linnaeite (Plumb 1960).

The Adelaide River mine lease is currently being explored by a joint venture of Royal Resources Ltd and Aldershot Resources Ltd. Drilling in 2007 intercepted 7.1 m at 0.35% U_3O_8 (Royal Resources Ltd 2009), including 4 m at 0.72% U_3O_8. The mineralised zone was also elevated in cobalt, copper and nickel. Surface sampling has returned zones in excess of 0.1% U_3O_8, and grab samples have returned up to 4.0% U_3O_8 (Royal Resources Ltd 2009).

George Creek

Uranium mineralisation at George Creek was discovered in 1954. From 1954–1960, BMR drilled seven diamond holes and carried out some underground exploration by a 39 m-deep shaft with 60 m of drives and cross-cuts. A total of 120 t ore averaging 0.26% U_3O_8 was extracted during the underground exploration phase and this was sent to Rum Jungle for processing (Arkin and Walpole 1960). Exploration in the 1970s by Central Pacific Minerals NL and Urangesellschaft Australia Pty Ltd failed to find any further indications of economic mineralisation (Fander 1981). Mineralisation is within the Burrell Creek Formation and comprises pitchblende, coating joints and fractures, associated with pyrite- and chalcopyrite-bearing quartz veins. The prospect lies on the western limb of a north-plunging anticline and bedding dips 30° towards 280°. Subvertical northwest-trending faults and related fractures control mineralisation and the exploration drives followed one such structure. Pods and thin stringers of pitchblende occur along the fault elsewhere. Mineralisation is at the intersection of a fault and coarse-grained greywacke beds, which are interbedded with barren siltstone beds. The primary ore consists of pitchblende; torbernite, metatorbernite and autunite are present in the surface oxidation zone to 25 m depth.

The Royal Resources Ltd/Aldershot Resources Ltd joint venture is currently exploring at George Creek and has drilled two holes: the first intersected the old BMR drive at 25 m; the second confirmed the grade of the mineralisation, with diamond drillhole GCDDH009 intersecting 1 m at 0.48% U_3O_8 (10.6 lb/t) from 21.93 m (Royal Resources Ltd 2009).

Fleur de Lys

This prospect is located 2 km northwest of the Cosmo Howley gold mine and was discovered in 1953. It produced 170 t ore averaging 0.12% U_3O_8 during 1954–1955 (Crohn 1968). Later radiometric and stream sediment surveys by United Uranium NL failed to discover any economic mineralisation (Newton 1968). Mineralisation is within siltstone, tuff, mudstone and phyllite of the Gerowie Tuff, in bedding-parallel shears and joints (Firman 1955, 1956). At the surface, torbernite, malachite, azurite and cuprite occur as veins and coatings on joints and bedding planes within the strongly sheared shale unit (Thundelarra 2009). The primary ore consists of uraninite, pyrite, arsenopyrite, chalcopyrite and bismuthinite in narrow veins, with varying amounts of quartz and sericite. Pitchblende appears to have formed after pyrite and arsenopyrite, but before chalcopyrite and bismuthinite (McAndrew 1954). Torbernite, malachite, azurite and cuprite are present as coatings on bedding planes, shears and joints in the oxidised zone (Firman 1955).

Thundelarra Exploration Ltd is currently testing the prospect. A six-hole RC drilling program was completed by this company, but only one hole, 08PCRC002, intersected significant uranium returning an assay of 2608 ppm U_3O_8 from 57–58 m (Thundelarra 2009).

Twin and Dam

The Twin and Dam prospects are located 36 km north-northeast of Pine Creek township and were discovered by Total Mining Australia Pty Ltd in 1983. Drilling defined two main mineralised zones and several minor occurrences within an area about 400 m wide and 1800 m long (Earthrowl *et al* 1988). Total (proven and probable) ore reserve estimates were 256 t ore at 0.1187% U_3O_8 for Twin and 341 t ore at 0.13% U_3O_8 for Dam (Berthault and Harrop 1990). Structurally controlled mineralisation is within carbonaceous and chloritic shale, chert and dolostone of the Masson Formation, at the margin of the Cullen batholith. Orebodies are composed of concentrations of narrow, closely spaced veins of uraninite, within and on the margins of the faults, and are associated with chloritic, sericitic and haematitic alteration. Secondary mineralisation is developed in the oxidised zone at Twin and consists of bassetite, coffinite and meta-autunite.

Based on recent field exploration, drilling and information collected from its 2007 drilling program, Atom Energy Ltd announced a JORC-compliant inferred resource of 585 000 t at 286 ppm U_3O_8 for Dam and 824 000 t at 316 ppm U_3O_8 for Twin, at a cut-off grade of 100 ppm U_3O_8 (Atom Energy Ltd, ASX announcement, 26 March 2008).

Other vein-type occurrences

There are several other vein-type uranium occurrences within the PCO. A full record is within the MODAT database.

The *ABC* prospect is located 17.5 km northeast of Katherine and has an estimated resource of 1073 t ore at 0.4% U_3O_8 (Gardner *et al* 1955). The mineralisation is hosted within northeast-dipping mafic volcanic rocks of the McAddens Creek Volcanic Member, which lies between Mamadawerre and Gumarrirnbang sandstones of the lower Katherine River Group.

Several uranium occurrences are in the vicinity of the old Edith River railway siding, 40 km northwest of Katherine. These occurrences are in steeply dipping north-northwest-trending shear zones, within greisenised Tennysons Leucogranite and include *YMCA, Yenberrie,* and *Tennysons 1* to *6* (Firman 1954, Gardner 1953 a, b).

At the *Thunderball* Prospect, near Hayes Creek, a number of holes drilled by Thundelarra Exploration Ltd in 2009–2010 have intersected significant uranium mineralisation within the Mount Bonnie Formation. Mineralisation is located along a northerly-trending anticlinal axis and consists of veins and disseminations of uraninite (pitchblende) hosted within a folded brittle ductile shear that appears to plunge approximately 40º to the north. The highest grade mineralisation occurs at or near the hinge zone of the interpreted anticlinal structure (Thundelarra Exploration website; http://www.thundelarra.com).

Genesis of vein-type uranium deposits

There are no detailed models on the genesis of vein-type uranium occurrences. Bajwah (1994) considered that these deposits are related to late-stage hydrothermal activity associated with intrusion of granitoids. However, Lally and Bajwah (2006) were of the opinion that deposits such as Adelaide River and George Creek are an appreciable distance from any granite contact and mineralisation at Twin and Dam overprints contact metamorphism of the host rocks. The ABC deposit certainly cannot be related to granite intrusion, since it is hosted by volcanics in the Kombolgie Subgroup. McKay and Miezitis (2001) suggested that the vein-type deposits in the PCO were formed in a similar manner to unconformity-related deposits, by oxidising fluids descending along structures from overlying platform cover sandstones.

Surficial uranium occurrences

There are numerous surficial uranium occurrences in the PCO. The majority of these are within Kakadu National Park. These occurrences are largely radiometric anomalies within either Cenozoic laterites, exposed palaeoweathering surfaces, deeply weathered Palaeoproterozoic metasedimentary rocks, fractured or altered volcanics or black soils.

Copper

Copper mineralisation was discovered in the Rum Jungle area in 1869 and was followed by the discovery, in 1872, of the *Copperfield* mine near Pine Creek. With the exception of uranium-copper deposits in the Rum Jungle region, which were discovered between 1949 and 1965, most of the copper occurrences in the PCO were known by late 1800. There are currently 141 mineral occurrences in the PCO with copper as the major commodity (**Figure 5.69**). Most of these are economically insignificant and none is currently producing. A summary of base metals occurrences, including copper, is provided in **Table 5.9.** Previous references to copper deposits in the PCO include Crohn (1968), Stuart-Smith (1985), Ferenczi (1990a–d) and Ahmad *et al* (1993). The following description relies on these previous studies.

Major Cu occurrences are located in the Rum Jungle and Mount Davis areas. Most occurrences are vein-type and resulted from hydrothermal fluids possibly associated with the granitic activity (eg *Mount Davis*, *Mount Diamond*, *Coronet Hill*). Another group is associated with carbonaceous shale, eg Rum Jungle area (*Intermediate* mine), and could be classified as stratabound and may represent remobilisation of syngenetically-deposited copper mineralisation. In addition to copper, many of these deposits also contain Pb, Zn, Ni and Co (eg *Browns* deposit) and are discussed under polymetallic deposits. Volcanigenic copper mineralisation occurs within the Warrs Volcanic Member of the Burrell Creek Formation near the Daly River Settlement.

Granite-related copper veins

Granite-related Cu veins are hosted by the Burrell Creek and Koolpin formations, but some are within Cullen Supersuite granitoids. Most lodes trend northwest or north-northwest or north-northeast. Paragenetically, compared to other mineralised veins in the PCO, they appear to be younger than the Au-quartz veins, overlap the Sn-quartz veins and are probably earlier than, or synchronous with the Pb-Zn-Ag-quartz veins. Near-surface mineralisation commonly includes malachite, azurite and chalcocite and a number of other oxidised copper minerals. Covellite and bornite are common in the supergene zone. Pyrite, chalcopyrite and minor arsenopyrite are the dominant primary sulfides.

Copperfield

This deposit is located about 5.5 km southwest of Pine Creek township and mining produced some 3450 t of ore averaging ca 25% Cu from 1875–1917 (**Table 5.9**). A remaining resource was estimated as 90 000 t at 6.1% Cu and 185 g/t Ag (United Uranium NL 1967). Mineralised veins are hosted within hornfelsed greywacke and phyllite of the Burrell Creek Formation. The quartz-filled breccia lode strikes 335° and dips 50–60° west. At the surface, the lode averages about 1 m in width and it can be traced for

about 200 m along strike. Pyrite and chalcopyrite, present as breccia fill, constitute the main primary ore minerals, with lesser chalcocite. Covellite and bornite are common minerals in the supergene zone, and malachite and azurite are present in oxidised near-surface ore.

Mount Ellison

The *Mount Ellison* mine is located about 2.7 km north of Ban Ban Springs homestead. During 1891–1911, it produced some 3300 t of copper-bismuth ore averaging about 20% Cu and 0.1% Bi (Crohn 1968). The lodes are hosted within micaceous mudstone, carbonaceous mudstone and tourmalinite of the lower Koolpin Formation. In the mine area, these rocks strike 305° and dip 40–60° northeast. Two copper-bearing quartz lodes have been worked. In the secondary enrichment zone, that extending to a depth from 23–60 m, the main minerals are cuprite, chalcocite and bornite. The primary ore minerals include chalcopyrite, pyrite, linnaeite (Co_3S_4), aikinite ($PbCuBiS_3$), witchenite (Cu_3BiS_3) and native bismuth (Lawrence 1963).

Mount Davis area

There are 16 recorded vein-type copper occurrences hosted by the Mount Davis Granite. These are typified by north- to northwest-trending Cu-As-Fe sulfide-bearing quartz veins and chloritic shear zones within medium to coarse, equigranular biotite granite. The largest, *Mount Davis*, deposit produced 844 t of ore averaging about 12% Cu (Walpole 1962, Shields 1965). The mineralised lode at this locality strikes north and is up to 1.5 m wide. Primary mineralisation consists of massive arsenopyrite, chalcopyrite, minor pyrite, galena and trace molybdenite. Quartz and chlorite are the main gangue minerals with minor sericite. The granitic wallrocks are commonly sheared and chloritised. Other significant occurrences in this area are *Mount Davis East*, *Ranfords Shaft*, *Foleys Shaft*, *Edwards and Connellys*, *Bell and Witherdens*, *Foley and Cahills* and *Beirnes*.

Mount Diamond

This deposit was worked intermittently from 1898–1930, producing 762 t of ore averaging about 7% Cu. The

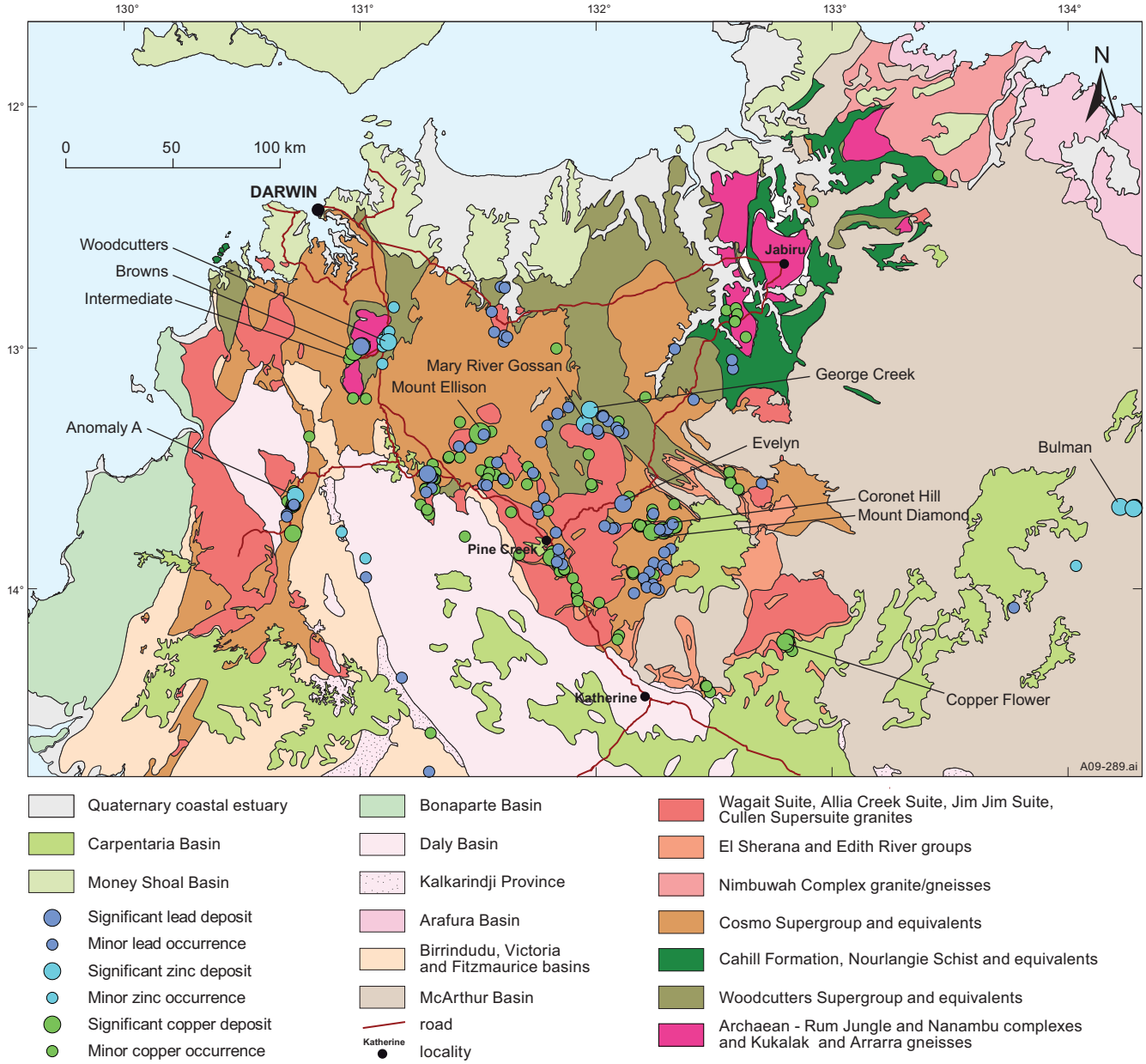


Figure 5.69. Base metals occurrences of the PCO. Data from NTGS MODAT database. Geology simplified from 1:2.5M geological map of NT (Ahmad and Scrimgeour 2006).

mine was re-opened between 1970–1973 and produced 2453 t Cu, 3.73 t Ag and 43 t Bi and 10.6 kg Au from 51 000 t of ore (Shields 1974). A remaining indicated resource of 150 000 t averaging 5% Cu exists to a depth of 100 m (Taylor 1973). Host rocks are hornfelsed greywacke and phyllite of the Burrell Creek Formation. Mineralisation is within a quartz vein that has a strike length of 2600 m and is about 1 m wide in the mineralised sections (Taylor and Cox 1969). Primary mineralisation consists of chalcopyrite and minor pyrite, arsenopyrite, galenobismutite ($PbBi_2S_4$), bismutite (Bi_2S_3), sphalerite and rare native gold. Secondary ore minerals include chalcocite, covellite, malachite, scorodite, cuprite and azurite. Gangue minerals include quartz, chlorite, magnetite, sericite, calcite and siderite (Taylor and Cox 1969).

Common name	Lat	Long	Size	Shape	Commodities	Rock unit	Resource	Grade	Production	Reference
Anomaly A	-13.61	130.72	Small	Tabular	Zn-Pb-Cu	Burrell Ck Fm	0.3 Mt	12% Zn		Woodland (1976)
Area 55	-13.01	130.97	Medium	Tabular	Cu-Co-Ni	Whites Fm	*6.6 Mt*	*0.83% Cu, 0.17% Co, 0.17% Ni*		Compass Resources (2006)
Browns and Browns East	-12.99	131.00	Large	Tabular	Pb-Cu-Co-Ni	Whites Fm	*62.34 Mt*	*0.52% Cu, 3.10% Pb, 0.1% Co, 0.09 Ni*		Compass Resources (2008)
Copperfield	-13.87	131.81	Small	Sheet	Cu	Burrell Ck Fm	0.09 Mt	6.1% Cu	3450 t at 25% Cu	United Uranium (1967)
Coronet Hill	-13.73	132.33	Medium	Sheet	Cu-Au-Pb	Burrell Ck Fm	0.05 Mt	1.6% Cu, 131 g/t Ag	252 t at 21.6% Cu	Butler (1993a)
Daly River	-13.67	130.70	Small	Sheet	Cu	Burrell Ck Fm			6000 t at 20% Cu	Hossfeld (1937)
Evelyn	-13.65	132.12	Small	Tabular	Pb-Cd-Au-Zn	Koolpin Fm	0.007 Mt	3.7% Zn, 6.7% Pb, 343 g/t Ag	0.08 Mt at 5.5% Pb, 7.5% Zn	Ferenczi and Sweet (2005)
Flora Belle	-13.66	131.75	Small	Sheet	Pb	Burrell Ck Fm	0.085 Mt	4% Pb	390 t at 6.45% Pb	Ahmad *et al* (1993)
George Creek Zn	-13.25	131.98	Small	Tabular	Zn-Pb	Masson Fm	0.05 Mt	4% Zn, 2% Pb		Ahmad *et al* (1993)
Intermediate (Browns East)	-12.99	131.00	Small	Tabular	Pb-Cu-Co-Ni	Whites Fm	*14.6 Mt*	*1.06% Cu, 1.09% Pb, 0.15% Ni, 0.14% Co*	0.72 Mt at 2.2% Cu	Compass Resources (2008)
Iron Blow	-13.51	131.55	Medium	Tabular	Zn-Cu-Au-Pb	Mount Bonnie Fm	*3.174 Mt*	*3.28% Zn, 0.76% Pb, 0.19% Cu, 100 g/t Ag, 2.08 g/t Au*	0.001 Mt at 9 g/t Au (oxidised ore); 0.002 Mt at 7 g/t Au (sulfide ore)	Crocodile Gold (2009)
Mary River Gossan	-13.31	131.95	Medium	Sheet	Zn-Cu-Pb	Mundogie Sst	0.91 Mt	9.7% Zn, 0.5% Pb		Darby (1985)
Mount Bonnie	-13.54	131.55	Small	Tabular	Zn-Cu-Au-Pb	Mount Bonnie Fm	0.65 Mt	9% Zn, 2% Pb, 0.5% Cu, 1.7g/t Au, 280 g/t Ag	0.11 Mt at 7 g/t Au	Northern Gold (2005)
Mount Ellison	-13.35	131.50	Small	Sheet	Cu-Bi	Koolpin Fm			3300 t at 20% Cu	Crohn (1968)
Mount Fitch	-12.95	130.95	Medium	Tabular	Cu-Co-Ni-U	Whites Fm	*5.3 Mt*	*0.36% Cu, 0.09% Co, 0.12% Ni*		Compass Resources (2008)
Mount Davis	-13.76	132.29	Small	Sheet	Cu-Pb	Mount Davis Granite			844 t at 12% Cu	Shields (1965)
Mount Diamond	-13.76	132.24	Small	Sheet	Cu-Bi-Au	Burrell Creek Fm	0.15 Mt	5% Cu	2453 t Cu, 3.73 t Ag, 43 t Bi	Shield (1974)
Woodcutters	-12.97	131.11	Medium	Tabular	Zn-Pb-Sb	Whites Fm			4.651 Mt at 12.28% Zn, 5.6% Pb, 0.73% Sb, 87 g/t Ag	Ahmad *et al* (2006)

Table 5.9. Summary data on significant PCO base metals deposits. Resources in italics are JORC compliant. Abbreviations: Lat = latitude; Long = longitude; Fm = Formation; Sst = Sandstone; Ck = Creek.

Coronet Hill

The Coronet Hill group of mines are hosted by the Mount Bonnie Formation adjacent to the Coronet Hill Fault, some 2.5 km east of the Mount Davis Granite. The main copper-bearing quartz lode contains significant amount of arsenopyrite, along with pyrite, galena, covellite, bornite, sphalerite, bournonite, pyrrhotite, marcasite, tetrahedrite and gold (Ruxton and Shields 1962). Drilling by United Uranium NL outlined a remaining resource of 51 000 t at 1.6% Cu and 131g/t Ag (Butler 1993a).

Other vein-type copper deposits

There are a number of small cupriferous quartz vein/shear-related occurrences within the PCO. A full description of these is not included here; more details are available in MODAT. At the *Enterprise 2* prospect, mineralisation is within hornfelsed Burrell Creek Formation and is confined to a north-northwest-striking, 0.5 m-wide quartz vein that is discontinuously exposed over a distance of 330 m. The *Burrundie Copper* prospect is hosted by the middle Koolpin Formation and is contained within vein quartz breccia. Pyrite, chalcopyrite and minor sphalerite have been recorded in drill core. At *The Jar* copper prospect, mineralisation is within a north-northwest-trending shear zone, which is about 1 m wide and can be traced for over 700 m. The host rocks are greywacke and shale of the Burrell Creek Formation. The *Phillip Greets* copper prospect produced some 363 t of copper ore averaging 25–30% Cu. The mineralisation comprises secondary copper minerals, associated with a north-northwest-trending fault, which has been worked over a strike length in excess of 300 m. The mineralised lode lies within greywacke and shale of the Mount Bonnie Formation. Mineralisation at the *Fullhand* mine is hosted by the Mount Bonnie Formation. Two lodes are present: a north-northeast-trending copper lode and an east-northeast-trending copper-lead lode. The former occupies a 2 m-wide shear zone and can be traced for about 30 m. It is 0.3–1 m thick and contains malachite, azurite and rare torbernite. The copper-lead lode contains galena, cerussite and chalcopyrite, associated with a carbonaceous gangue.

Copper veins of volcanigenic affiliation

A number of copper occurrences in an area 9–18 km north of the Daly River Mission are tentatively placed in this group.



Figure 5.70. Old workings following the lode at Daly River Copper mine (52L 683049mE 8488092mN).

These occurrences are present in chloritised shear zones and carry few, or no quartz veins. The deposits are hosted by the Warrs Volcanic Member or by adjacent schist and phyllite of the Burrell Creek Formation. The lodes usually trend north-northeast and appear to be relatively narrow and conformable with the cleavage or bedding of the enclosing strata. Moderate to intense chloritisation is a common alteration feature of these deposits. Magnetite is commonly disseminated within the chloritised zone. Quartz-carbonate-chlorite and sericite-chlorite alteration assemblages are prominent in some areas. The oxidised material, which has been mined, contains malachite, azurite and chalcocite. Primary ore minerals comprise pyrite, chalcopyrite, marcasite, arsenopyrite and minor galena and sphalerite.

Daly River copper mine

The *Daly River* copper mine (**Figure 5.70**) was discovered around 1884 and worked intermittently until 1918. It produced some 6000 t of concentrate averaging about 20% Cu (Hossfeld 1937). The lodes have been explored by drilling, but no economic resources have been outlined (Thomas 1957, Woodland and Cotton 1977).

The mine is hosted within a tongue of the Burrell Creek Formation, between the Warrs Volcanic Member to the east and the Mulluk Mulluk Volcanics to the west (**Figure 5.71**). The strata strike north-northwest and dip steeply southwest. The mine is close to the contact with the overlying Warrs Volcanic Member. The host rocks are volcaniclastic greywacke and minor siltstone of the Burrell Creek Formation. At least three sulfide-bearing zones are associated with bedding-parallel, 018°-trending, near-vertical shears. Surface mineralisation consists of malachite and azurite, with chalcocite and chalcopyrite at depth. Diamond drillhole DDH A4 intersected the main lode 280 m below the surface, where it consisted of sphalerite and chalcopyrite disseminations and stringers, with lesser pyrite and marcasite, within a quartz chlorite schist host (Berger 1973). There are no massive sulfide lenses. Maddocks (1979) pointed out similarities between the mineralisation at the Daly River copper mine and stringer zone mineralisation in volcanigenic massive sulfide deposits.

De Ross (NTGS unpublished data) analysed 43 core samples from the deposits in the Daly River area. Three of the samples were from the Daly River copper mine and these yielded 0.23, 0.20 and 0.02 g/t Au. Samples from other deposits gave similar values. One sample from *Anomaly A* gave an Atomic Absorption Spectroscopy value of 9.2 g/t Au, which when checked by Fire Assay, returned 2.2 g/t Au.

Empire copper mine

The Empire copper mine is located some 11 km north of the Daly River Mission. It was worked from 1904–1907 and produced a total of about 5.5 t ore, averaging better than 25% Cu, and 138 t of low-grade ore (<10% Cu) from two shafts and an open cut (Thomas 1958). Samples of gossanous ore from the dump have given assays of up to 13.7% Cu, 3 g/t Au and 20 g/t Ag (Hossfeld 1937). Subsequent exploration, conducted by Western Nuclear Pty Ltd, included geochemical, magnetic and IP surveys, and diamond drilling (Woodland and Cotton 1977). Mineralisation is hosted by interbedded volcanic rocks and phyllite of the Warrs

Volcanic Member. Drillhole A9 intersected a succession of predominantly intermediate to basic lava flows and breccia zones, which contain fragments of chert, siltstone or volcanic rocks within a calcite matrix (Berger 1973).

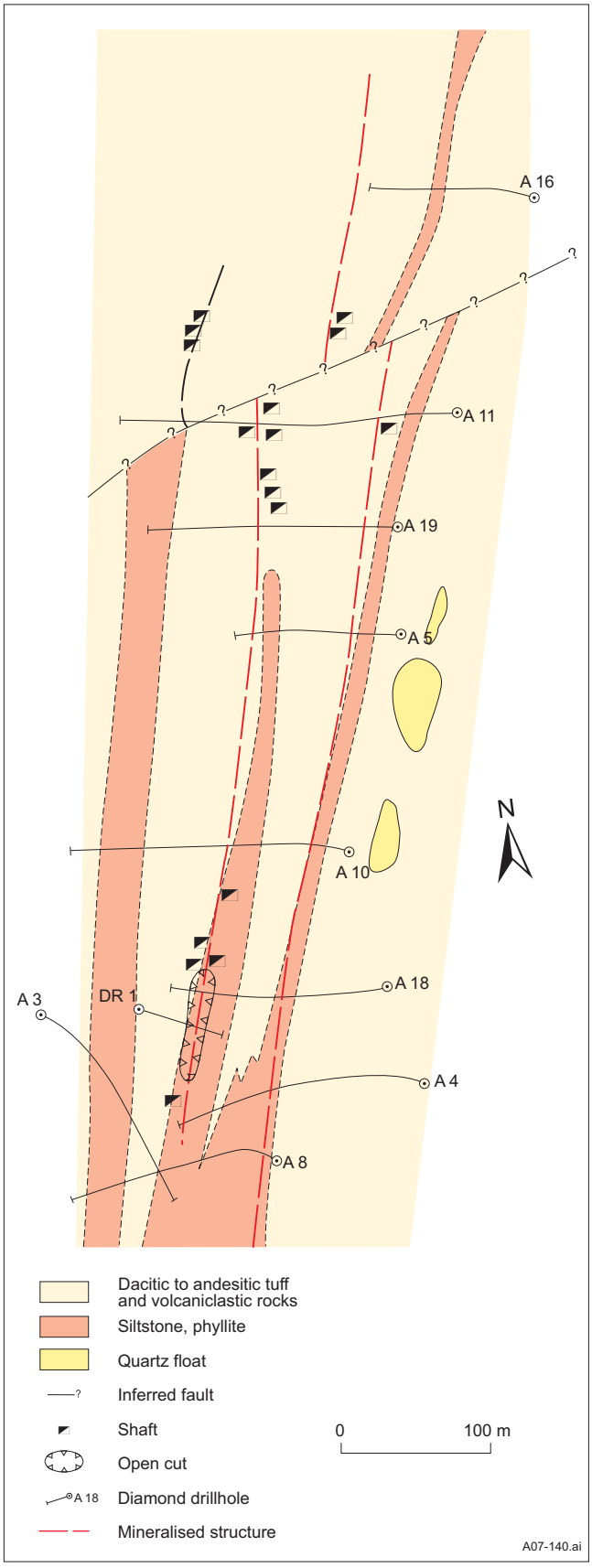


Figure 5.71. Surface geological map of the Daly River mine area (modified from Berger 1973).

Wheal Danks prospects

The *Wheal Danks, Wheal Danks South* and *Wheal Danks North* prospects are about 12 km north of the Daly River Mission. Wheal Danks is the largest amongst these and this deposit has been worked with a 60 m-long adit and several shafts. Recorded production is about 500 t of hand-sorted ore, averaging about 28% Cu (Crohn 1968, Ferenczi 1990d). The Wheal Danks prospects are hosted within chlorite-sericite schist of the Burrell Creek Formation, close to the contact with the overlying Warrs Volcanic Member. Secondary copper minerals are present at the surface. Hossfeld (1937) sampled lode and dump material, which assayed 2–4% Cu and 1.5–4.6 g/t Au.

Other copper occurrences of volcanigenic affiliation

Copper mineralisation at the *Wallaby* mine consists of en echelon quartz-haematite lenses. Malachite is the only ore mineral visible at the surface, forming stains on ferruginous sugary vein quartz. Sub-surface drilling has intersected traces of pyrite and chalcopyrite, associated with calcite veinlets or cleavage planes (Berger 1973). Small copper occurrences have been worked by shafts and open cuts in the *Anomaly C* area (*Hill 3, Hill 4* and three unnamed occurrences). Malachite is the principal ore mineral and it occurs as fracture fillings, with or without quartz veining. These occurrences are hosted by the Warrs Volcanic Member. *Occurrence No 231* is hosted by chloritic schist of the Burrell Creek Formation and represents the northernmost occurrences of the same mineralisation style as is present at the Daly River copper mine. Quartz and tourmaline veins are common in the vicinity of this occurrence, but they are not mineralised (Ahmad *et al* 1993).

Polymetallic deposits

A number of mineral commodities are closely associated in this class of deposits. A broad subdivision is as follows: Pb-Zn-Cu; Pb-Cu-Zn-Ni-Co ± U; and Zn-Cu-Pb-Ag-Au. On the basis of orebody characteristics, these can be further subdivided into: (1) veins; (2) shale-hosted stratabound/stratiform deposits; (3) volcanigenic massive sulfide deposits; and (4) skarns.

Lead-zinc-silver mineralisation in the PCO was first discovered at the Mount Evelyn mine in 1886 and was soon followed by a number of other discoveries. However, the largest Pb-Zn-Ag deposit (Woodcutters mine) was discovered in 1966 and was mined between 1985 and 1999. The largest deposit is the Browns prospect. Mining of the oxidised ore from this deposit commenced in the second half of 2008. but was suspended in early 2009.

Lead-zinc-silver veins

Most Pb-Zn-Ag occurrences are associated with the Pine Creek Shear Zone and the southern extension of the Mount Shoobridge Fault. However, the largest deposit (Woodcutters mine) is hosted within a north-trending fault system at the eastern flank of the Rum Jungle Dome.

Exploration and mining details of a number of these occurrences are given in Crohn (1968), Stuart-Smith (1985), Ahmad *et al* (1993) and Ferenczi (1990a–d). A large part

of the following information is based on these previous studies. Galena and sphalerite are the main ore minerals of this deposit type and they are associated with pyrite, chalcopyrite and minor arsenopyrite. Minor silver sulfosalts are present, but the bulk of the silver is contained within galena. Quartz, chlorite and carbonates constitute the main gangue minerals. Contacts with wallrocks are sharp and the alteration fringes include chlorite, sericite and carbonates.

Woodcutters

Woodcutters deposit is located 65 km south of Darwin alongside the Stuart Highway. In 1966, BMR conducted a soil geochemical survey and identified six lead-zinc anomalies, designated L1–L6. Primary lead-zinc mineralisation was intersected by BMR drilling at the L5 anomaly, which was later named 'Woodcutters'. In 1968, the area was acquired by Peko-Walsend Ltd. Evaluation of the Woodcutters deposit during 1968–1971 outlined 1.2 Mt of ore at 12.9% Zn, 7.5% Pb and 154 g/t Ag (Taube 1984). In 1984, a consortium comprising Nicron Resources Ltd, Petrocarb Exploration NL, Lachlan Resources NL and Aztec Mining Company Ltd acquired the leases and commenced mining from an open cut in 1985 at the rate of 360 000 tpa. Normandy Mining Ltd acquired the mine in 1994 and spent $33.7 million to upgrade the mining and processing facilities to produce at a rate of 540 000 tpa. Mining continued until March 1999, when operations ceased, after producing a total of 4.65 Mt ore averaging 12.28% Zn, 5.65% Pb and 87 g/t Ag.

Mineralisation at Woodcutters is essentially discordant and is hosted within the Whites Formation. In comparison to the uranium and polymetallic deposits of the embayment area, which lie closes the Whites Formation-Coomalie Dolostone contact, the Woodcutters mineralisation spans from this contact to about 700 m vertically up the succession (**Figure 5.72**). In the mine area, the Whites Formation is divided into Lower, Middle and Upper sequences (**Figure 5.73**). The Lower Sequence (ca 300m thick) comprises alternating, interbedded grey dololutite and shale and relatively homogeneous dolomitic black shale units. The Middle Sequence (ca 100m thick) comprises finely bedded black shale. The Upper Sequence (>200m thick) comprises spotted black shale with minor dololutite. The Lower and Middle sequences contain a number of up to 2 cm-thick interbeds of light grey waxy slate, described as tuff marker

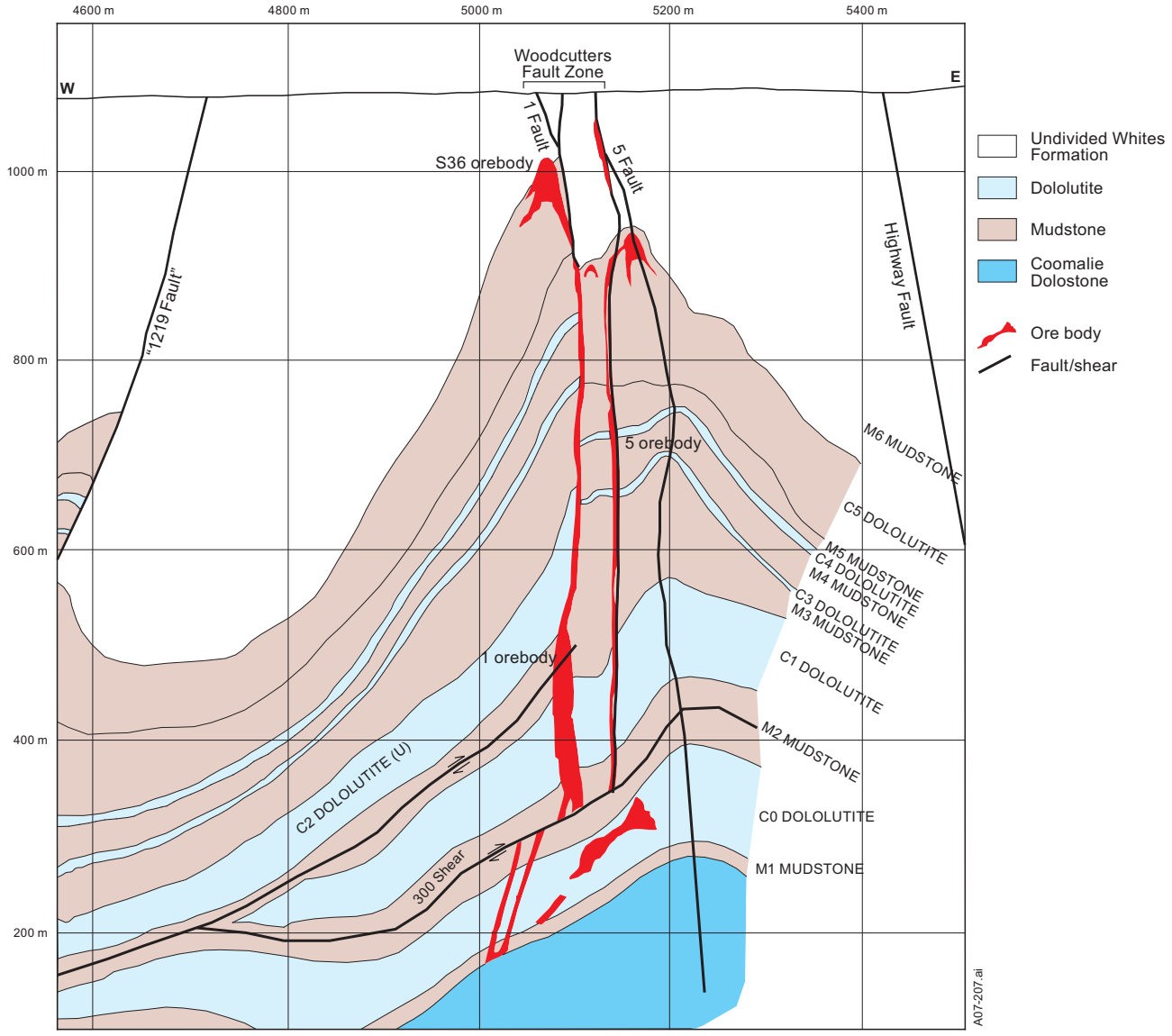


Figure 5.72. Geological cross-section of Woodcutters deposit (modified from Goulevitch 1997).

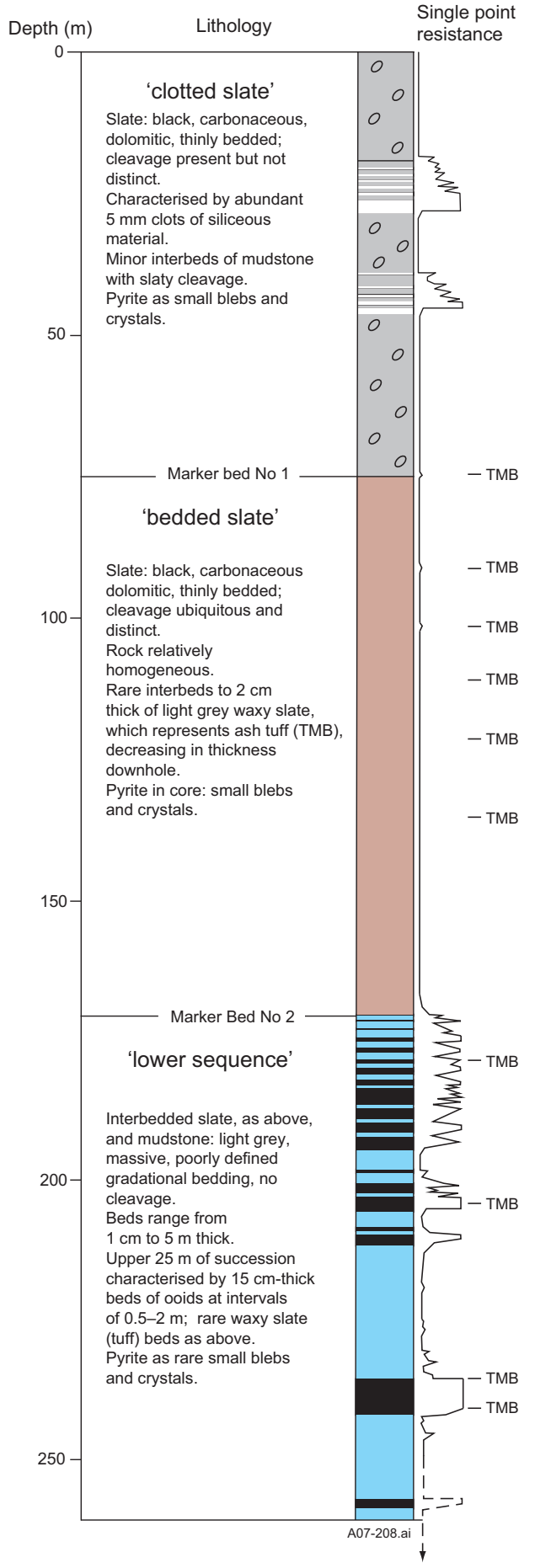


Figure 5.73. Stratigraphic column and single point resistance log showing lithological subdivision of Whites Formation at Woodcutters (after Taube 1984). TMB = tuff marker bed.

beds. These beds consist of fine crystals of microcline, shards of quartz and rare zircon in a very fine recrystallised ashy matrix (Taube 1984). The *L1* and *L6* prospects, located to the north of Woodcutters, lie higher up in the succession and mainly consist of black shale with minor dololutite (**Figure 5.74**). The total estimated thickness of the Whites Formation in the Woodcutters area is about 1300 m. Amphibolite sills marks the top of the succession, which is then followed by the Wildman Siltstone (Taube 1984, 1990).

Three types of intrusives are present in the mine succession. Amphibolite sills occurring at the top of the Whites Formation are composite bodies ranging from metadolerite to metagranophyre; these have been altered to a chlorite-biotite-bearing assemblage (Fleming *et al* 1994). Two types of monchiquitic lamprophyric dykes (dolomitised and calcitised dykes) are also present (Smolonogov 1988). These are described as low-temperature intrusives, formed by hydrous volatile-rich melts. Muscovite from the dolomitised dyke has been dated by the K-Ar method at 1728 ± 17 Ma. This age is about 100 Ma younger than the late orogenic granites within the PCO and about 140 Ma younger than the Nimbuwah Event. This compares well with the conventional U-Pb baddelyite age of 1723 ± 6 Ma for the Oenpelli Dolerite (OZCHRON database). Recent dating of muscovite (Ar-Ar), tantalite and cassiterite (Pb-U) from the Bynoe pegmatite belt also falls within this range (Frater 2005).

The conformably underlying Coomalie Dolostone is a ca 250 m-thick succession of white crystalline dolostone, magnesite, stromatolitic dolostone and massive fine-grained dolostone.

The orebodies lie along the crest of a north-trending, upright, F_2 doubly-plunging anticline (Woodcutters Anticline) and follow sub-parallel, near-vertical transpressional faults, collectively termed the Woodcutters Fault Zone (**Figure 5.72**). The anticline continues north for a known strike length of about 8 km and is offset by dextral movement on the Giants Reef Fault. Plunge reversal occurs in the vicinity of the Woodcutters mine. The anticline bifurcates into two in the mine area. A well defined S_2 slaty cleavage parallel to the F_2 fold axis is present at all scales (Fleming *et al* 1994).

The Woodcutters Fault Zone consists of a 300 m-wide zone of north-trending, steeply dipping subparallel faults. These faults host a substantial portion of the mineralisation and are numbered from west to east as 6, 4, 2, 1, 3, 5, 7 and 9. Most mineralisation is within the 1, 3 and 5 structures.

Late, northwest-trending, steeply southwest-dipping cross-faults have minor sinistral movement and cut the earlier structures. These faults were described as being unfilled shears, but may have graphite-coated slickensided surfaces, thin quartz-carbonate veins and intense cleavage development; they are possibly the precursors of structures intruded by carbonate dykes (Fleming *et al* 1994).

Ore comprises sulfide veins and replacement bodies close to fault structures. Two mineralisation styles are known to be present: sheeted veins associated with faults, and replaced dololutite adjacent to faults. The orebodies transgress the Woodcutters Anticline and are lensoidal in shape; these lenses widen in dololutite and are narrower in black shale.

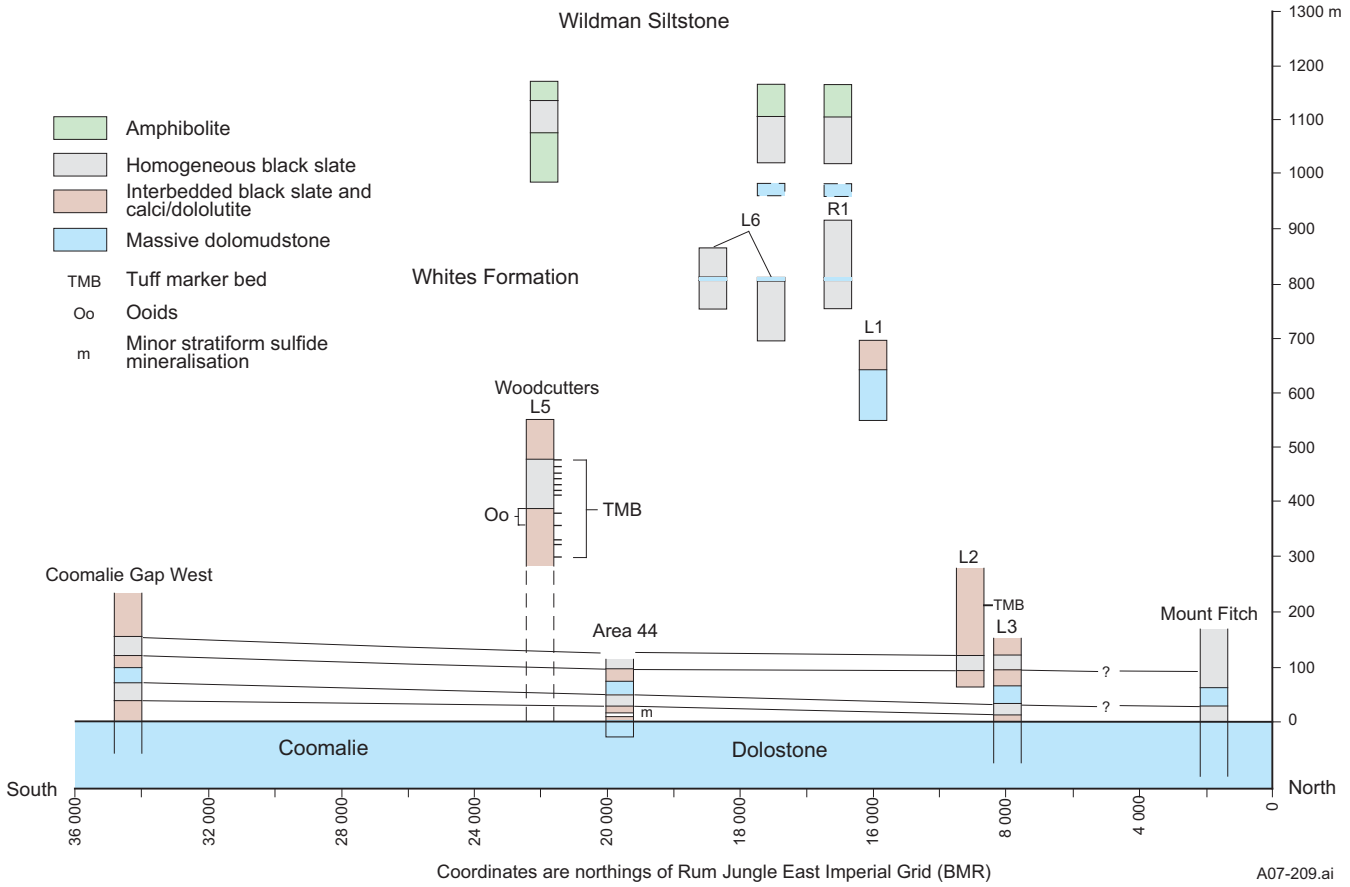


Figure 5.74. Location of Woodcutters area deposits with respect to local stratigraphic succession (modified from Taube 1984). L1–L5 and R1 represent sections associated with geochemical or geophysical anomalies. Stratigraphic columns from each prospect are plotted on local grid (horizontal axis). Vertical axis represents distance upward from Coomalie Dolostone/Whites Formation contact.

Sulfides constitute about 60% of the ore and more then 20 sulfide mineral phases have been identified. Sulfide minerals are coarse and individual galena crystals may be up to 1 cm in size. The main ore-bearing minerals are sphalerite, galena, boulangerite and falkmanite. Pyrite, sphalerite, galena and arsenopyrite are present in most samples followed by sulfosalts; other sulfides, including chalcopyrite and pyrrhotite, are subordinate (Roberts 1973). Just (1990) reported freibergite, gersdorffite, cobaltite and ullmannite. Anomalous levels of Au, Ni, Co, In and Sn have also been reported (Goulevitch and Butler 1998). Native gold was is present in some specimens (England 1997). Stannite is present as fine veins in sphalerite and trace amounts of cassiterite have also been reported (Just 1990).

The main gangue minerals are dolomite, quartz, tourmaline, calcite and apatite (Roberts 1973). Tourmaline is widespread in the ore and it is identified as a magnesian dravite variety, suggesting a sedimentary rather than an igneous parentage (England 1997). Magnesite has not been reported from thin section studies, but it is likely to be present and has probably been misidentified as dolomite.

Ahmad *et al* (2006) constructed a simplified paragenetic sequence, which is given in **Figure 5.75**. Three generations of ore mineral precipitation have been identified, with pyrite forming in all the stages. Because gold is within arsenopyrite, it is assumed that it formed early. Sphalerite formation is ascribed to stage II and, because it is veined and replaced by galena, the latter is placed in stage III. Stannite is seen as veining sphalerite and has been placed

Mineral/stage	I	II	III
Pyrite	----------	----------	----------
Arsenopyrite	----------	-----	
Sphalerite		----------	
Galena			----------
Lead sulfosalts			----------
Tetrahedrite			----------
Chalcopyrite			----------
Stannite			-----
Cassiterite			-----
Gold	----------	-----	

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Figure 5.75. Mineral paragenetic sequence for Woodcutters deposit (after Ahmad *et al* 2006).

in stage III. Of the gangue minerals, dolomite and quartz probably precipitated throughout the sequence. In the veins, quartz appears to be later than dolomite. Layered pyrite and tourmaline are apparently syn-sedimentary and must have formed earlier than the stage I coarse pyrite.

Woodcutters: Fluid inclusions

Fluid inclusions in sphalerite contain saline fluid + salt + CO_2 (vapour ± liquid). These inclusions yielded homogenisation temperatures in the range of 220–294°C and have salinity values in excess of 26% eq NaCl. Fluid inclusions in quartz have saline fluid + CO_2 (vapour ± liquid) ± salt. These inclusions homogenised in the temperature range 113–187°C and have salinity values ranging from 0.4–38.5% eq NaCl. Fluid inclusions in magnesite are simple saline fluid + H_2O

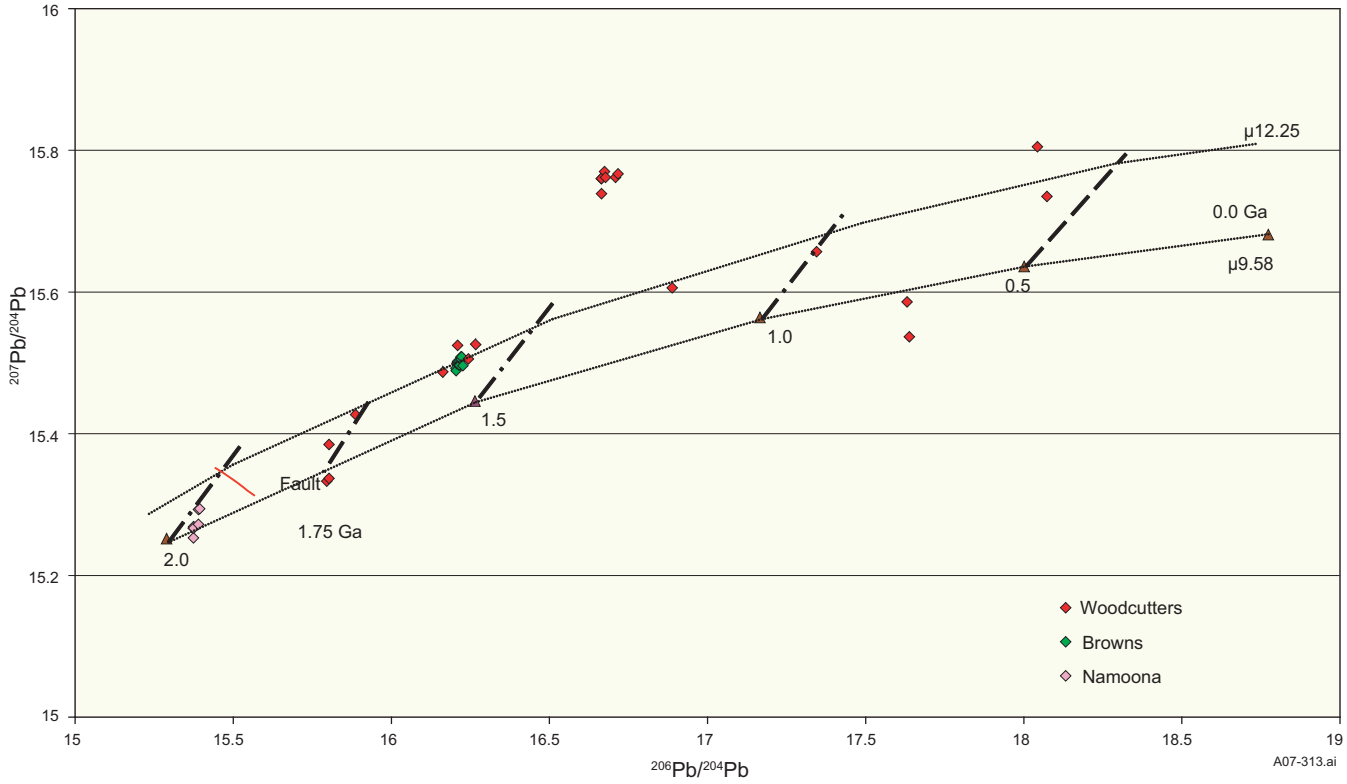


Figure 5.76. Lead isotope data for Woodcutters, Browns and Namoona deposits (data courtesy of Graham Carr, CSIRO, pers comm 2001).

(vapour) and may or may not have a salt and no CO_2. These inclusions provide homogenisation temperatures in the range 109–360°C and salinity values from 3.8–38.5% eq NaCl.

Woodcutters: Sulfur isotopes

Ahmad *et al* (1993) summarised sulfur isotope analyses of galena and sphalerite. $\delta^{34}S$ on pyrite is in the range 5.5–22.5‰ (mean 10.66‰), on sphalerite 5.5–12‰ (mean 8.5‰), and on galena 4.5–10.5‰ (mean 7.53‰). The sulfur isotope data is not compatible with common magmatic values of 0 ± 3%, given by Ohmoto and Rye (1979). However, they are compatible with connate waters and are similar to values from Mississippi Valley-type deposits.

Only one sample that contained coexisting galena and sphalerite was analysed and this gave $\delta^{34}S$ values of 7 and 9‰, respectively, indicating a temperature of 328°C (Ahmad *et al* 1993), which is in accordance with fluid inclusion temperatures obtained on the sphalerite. The mean of all analyses shows that $\delta^{34}S$ pyrite > sphalerite > galena and indicates equilibrium precipitation at a high temperature.

Woodcutters: Lead isotopes

Galena separates and whole-rock samples from Woodcutters were analysed for lead isotopes by CSIRO and analyses were supplied by Graham Carr (CSIRO, pers comm 2001). Lead isotope ratios for the Namoona, Woodcutters and Browns deposits are plotted in **Figure 5.76**. The oldest galenas from the Namoona deposit provide a reasonable age of ca 2000 Ma. The data from Woodcutters has a large spread, with a cluster of values well outside the growth curve indicating the incorporation of radiogenic lead from older crust. The oldest galenas from Woodcutters give an age of 1775 Ma. Galenas from the Browns deposit form a tight cluster of values right on the growth curve and provide an age of 1660 Ma. Some Woodcutters galenas are also within this cluster. Overall, the lead isotope values from the Woodcutters deposit are anomalous and have a wide spread of values, indicating crustal derivation and a complex history of lead growth.

Woodcutters: Genesis

Proposed genetic models for Woodcutters range from syndiagenetic to epigenetic to remobilised syngenetic. Roberts (1973) considered that during dolomitisation, base metals were released to pore solutions and remained in solution as organic complexes during lithification. During folding, ore solutions were transported to fractures and metal sulfide precipitation occurred when the organic complexes became unstable. Taube (1984) considered that Woodcutters mineralisation represented a late-stage manifestation of stratiform or stratabound mineralisation. Smolonogov (1988) postulated that the orebody geometry and bedding replacement, as well as interbedded tourmalinite and sulfides, could be interpreted as indicating an exhalative-type origin for some of the mineralisation. A basin dewatering model was proposed, in which hydrothermal fluid moved up the gradient towards the basin margin and channeled up through growth faults to form the Woodcutters deposit. Giles and Marshall (1998) also proposed a basin dewatering model and considered that the ore was late diagenetic in origin, with basin-derived metals and basement-derived sulfur. This mineralisation was remobilised later by fault reactivation. Fleming *et al* (1994) considered that the mineralisation was epigenetic. It formed in response to thermochemical sulfate reduction, brought about by hydrocarbons that were generated by reaction of hot fluids with carbonaceous matter within the Whites Formation. They interpreted that the mineralisation

postdated north-trending faults and lamprophyric dykes. Goulevitch (1997) synthesised and interpreted the drilling data from Woodcutters. His observations suggested that the 300 shear is a thrust fault, above which strong deformation during east–west compression resulted in the formation of the Woodcutters Anticline. Pre-existing mineralisation was remobilised into fault structures subparallel to the anticline. Below the 300 shear, the mineralisation was not remobilised and remained as stratiform replacement style.

Ahmad *et al* (2006) synthesised and evaluated structural, mineralogical, fluid inclusion, geochronological, sulfur and lead isotope data. They proposed that the Woodcutters mineralisation is essentially epigenetic. The deposit is located along the basin margin on the side of a palaeohigh. **Figure 5.77** provides a generalised view of this setting and its genetic implications. A possible origin involves basin shortening, during and after the ca 1865 Ma Nimbuwah Event, resulting in upward migration of metal-enriched formational waters towards the basin margin. The metal-bearing fluid encountered a major fracture system (Woodcutters Fault Zone) and rose upwards. Reaction with reduced wall rocks, as well as an increase in pH (due to reaction with carbonate), is probably the likely cause of ore mineral precipitation.

Flora Belle

The Flora Belle mine is situated about 20 km northwest of Pine Creek township and has been worked by numerous shafts down to the water table. It was first worked in the late 1880s, when 20 t of lead and 36 kg Ag were produced from 135 t of ore averaging 14.8% Pb and 260 g/t Ag (Ferenczi 1990c). The last known production in 1978 involved the extraction of 255 t of ore material from the dump, yielding 5 t Pb, 8 kg Ag and 55 g Au (Nyunt 1985). A mine feasibility study indicated reserves of 15 000 t of stockpiled material (4% Pb, 400 g/t Ag) and 70 000 t of *in situ* ore (Nyunt 1985).

The mine lies within interbedded chloritic phyllite and metagreywacke of the Burrell Creek Formation, about 3 km northeast of the contact with the McMinns Bluff Granite. The host rock dips 70–85° west and occupies the limb of a tight south-plunging anticline, the axis of which lies some 650 m to the east. Mineralisation is contained in a tabular, northwest-trending near-vertical vein, about 600 m long and 1 m wide. The surface expression of the lode is represented by a discontinuous quartz-bearing gossan, 0.5–2 m wide. Primary ore consists of coarse galena, tetrahedrite, pyrite, arsenopyrite, chalcopyrite, marcasite and pyrargyrite. The gangue minerals are quartz, siderite and chlorite.

Mary River gossan

This prospect (also known as *Gubberah Gossan*) was discovered by BMR in 1966. Between 1967 and 1977, geological, geochemical and geophysical surveys, and diamond drilling were undertaken by BMR and NTGS (Duckworth 1969, Shields 1969, Watts 1969, Daly 1971, 1975, Williams 1971, Bullock 1972, Michail 1974, Hone and Major 1978, Shields and Willis 1978). The prospect is within the Kakadu National Park and there has been no exploration for the last three decades. Darby (1985) compiled the results of these studies and estimated a possible resource of 0.9 Mt averaging 9.7% Zn and 0.5% Pb. Mineralisation is hosted within medium to coarse quartz-lithic sandstone and minor carbonaceous slate of the Mundogie Sandstone. The gossan occupies a discordant, northeast-trending quartz-filled shear about 1 m wide, 280 m long and dipping 70–80° to the northwest. In the richest mineralised portion, the surface expression of the lode consists of vein quartz with iron oxide boxwork after sulfides. Within the primary zone, sphalerite is the dominant ore mineral. Other minerals are pyrite, galena and rare stannite. Quartz is the predominant gangue mineral, but minor dolomite and sericite are also present. A large geochemical anomaly, termed the *Mary River South Anomaly*, is present about 1 km south of Mary River Gossan. A diamond drillhole was sunk to test this anomaly in 1971 (Daly 1971), but it did not fully intersect the projected target, because it was drilled down dip from the footwall side.

Namoona

The Namoona and *Namoona South* Pb-Zn prospects are located within Kakadu National Park, about 60 km northeast of Pine Creek Township, and were discovered in 1954 by Enterprise Exploration Company Ltd (Patterson 1959, Ikstrums 1979, Ikstrums and Steemson 1980, Cook 1981). Initial drill intersections encountered high-grade Pb-Zn-bearing quartz and carbonate veins in breccia zones. Follow-up infill drilling indicated narrow (1–3 m) poddy mineralisation of variable grade, associated within a rhyolitic tuff bed (Cook 1981).

Ore lenses are within a succession of dolomitic graphitic shale, carbonaceous pyritic shale, rhyolitic tuff, greywacke and tuffaceous shale of the Masson Formation. Ore comprises fine galena and minor amounts of tetrahedrite, sphalerite, pyrite and quartz. This mineralisation may represent the remobilisation of stratiform sulfides.

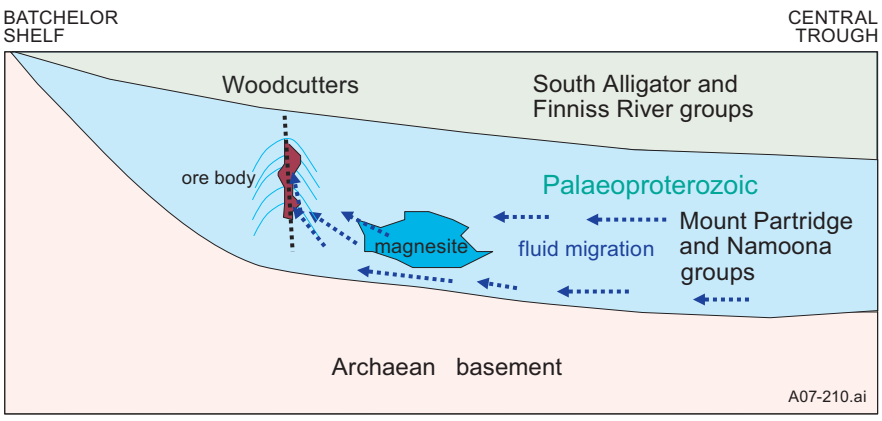


Figure 5.77. Schematic model for genesis of Woodcutters-style deposits (after Ahmad *et al* 2006). Basin shortening during and subsequent to Nimbuwah Event leads to migration of metal-enriched connate waters towards basin edge; mineralised fluids rise upon encountering a discontinuity, such as the Woodcutters Fault Zone; metal precipitation may be due to reduction by reaction with reduced rocks, or to pH increase by reaction with carbonate rocks; it is possible that metals were released due to dolomitisation of limestone or magnesitisation.

Other lead-zinc-silver veins

Other small notable occurrences in the Pine Creek Shear Zone include *Lucknow, Mount Wigley, Basin-6, Union Reefs Pb-Ag, Union Extended Pb-Ag, Jensens* and *Quests*. Galena is the main mineral in all these deposits and is associated with economically insignificant sphalerite. The mineralised veins are essentially similar to those at *Flora Belle*. Galena-bearing quartz veins are also present in the *Enterprise* gold mine and cross-cut the gold-quartz veins.

The *George Creek* Zn-Pb prospect was discovered by CRA Exploration Pty Ltd during a regional soil geochemical survey (Wills 1978). Surface mapping and sampling revealed three discontinuous gossanous intervals, anomalous in Pb and Zn, within moderately dipping shale and sandstone assigned to the Masson Formation. Subsurface drilling intersected significant intervals of basic volcanic rocks (probably Stag Creek Volcanics), which host the mineralisation. The drilling results suggest a possible resource of 50 000 t averaging 4% Zn, 2% Pb and 130 g/t Ag.

The *McCarthys* silver-lead mine is located in tightly folded, hornfelsed shale of the Wildman Siltstone, on the northern limb of a northwest-trending anticline. Some 580 t of concentrate averaging 69% Pb and 308 g/t Ag was extracted from two sets of shallow workings between 1913 and 1929 (Bagas 1983).

Lead-copper-zinc-nickel-cobalt

Significant polymetallic deposits of this type occur within the Rum Jungle area. They include the Intermediate, Browns, Area 55, Mount Fitch and Mount Burton prospects. These have been described by McCready *et al* (2004) and Ahmad *et al* (2006), and the following summary is based on their work. Two representative sediment-hosted stratabound deposits, the *Iron Blow* and *Mount Bonnie* mines, are described above in **Polymetallic gold deposits**. Both of these deposits contain a sub-economic resource of zinc and lead, but only gold and silver have been recovered from recent mining efforts.

Intermediate

The Intermediate deposit is hosted in a breccia of carbonaceous and talcose shale, enveloped by strongly sheared and contorted shale of the Whites Formation, near the contact with the underlying Coomalie Dolostone. The orebody is 150 m long, 40 m wide and extends to a depth of 350 m (Berkman 1968, Fraser 1980). This deposit was mined during 1964–1965 by an open cut that reached a depth of 70 m. The total ore mined was 395 000 t at 2.7% Cu and 367 000 t at 1.79% Cu. Pitchblende was a very minor component of the predominantly copper ore.

Browns

The Browns deposit lies at the southwestern extremity of a shear zone connecting the Dysons, Whites, Whites East and Intermediate deposits (**Figure 5.63**). It was discovered in 1954 by Territory Enterprises Pty Ltd (TEP). Subsequent exploration outlined an orebody up to 700 m long, 50 m wide and persisting to a depth of 450 m. In 1985, CRA Exploration Pty Ltd calculated total resources at 20 Mt averaging 5.6% Pb, 0.3% Zn, 0.19% Cu, 0.1% Co and 0.145% Ni (Allnutt 1985). The deposit is currently owned by a joint venture between Compass Resources NL (90%) and Guardian Resources Pty Ltd (10%). Resources at Browns are divided into oxide and sulfide orebodies. In October 2008, Compass Resources announced a JORC-compliant sulfide resource of 45.1 Mt averaging 0.355 Cu, 3.745 Pb, 0.73% Zn, 0.09% Co and 0.07% Ni. The oxide resource at the deposit is given as 2.64 Mt averaging 1.02% Cu, 0.12% Co and 0.1% Ni. Mineralisation extends eastward for approximately one kilometre along strike from the eastern margin of the Browns resource to the eastern margin of the historical Whites open cut. This eastward extension is referred to as *Browns East* and it contains a JORC-compliant total resource of 14.6 Mt averaging 1.06% Cu, 1.09% Pb, 0.21% Zn, 0.14% Co and 0.15% Ni. A separate oxide resource at Browns East has not been released (Compass Resources 2008). Mining of the oxidised ore from the Browns deposit commenced by open cut in mid-2008 and stripping of the first copper cathode occurred in September 2008. Shortly afterward, Compass Resources went into voluntary administration on 29 January 2009 and production was thereafter suspended.

The geological setting of the Browns deposit is similar to other Embayment area deposits (see **Unconformity-related uranium deposits**). It is situated on the northern limb of a regional F_2 syncline (**Figure 5.63**). Mineralisation is within carbonaceous shale of the Whites Formation (**Figures 5.78**, **5.79**), close to the



Figure 5.78. Steeply south-dipping carbonaceous shale of Whites Formation in Browns feasibility pit.



Figure 5.79. Sheared black carbonaceous shale coated with fine galena in Browns feasibility pit.

contact with the Coomalie Dolostone (**Figure 5.80**). At depth, the sheet-like orebody dips steeply (ca 80°) to the south, but it rolls into shallower dips (ca 45°) near the surface. Continuity of the orebody at depth has been established by drilling to a depth of 380 m (Compass Resources 2001).

In places, the deposit is vertically zoned, with Cu stratigraphically below Pb, while Co-Ni overlaps both of these zones. However, in many cases, the zonation is either more complex or is reversed, with Pb below Cu, without any evidence of stratigraphic inversion.

The regional metamorphic grade is low- to sub-greenschist facies and the main minerals are Fe-Mg chlorite, sericite, graphite and quartz. Andalusite porphyroblasts are common in the hangingwall of the lode and these are totally retrogressed to sericite-quartz assemblages.

Thin and polished sections of core and surface samples from the Browns deposit were examined by McCready *et al* (2004). **Figure 5.81** provides details of the mineralogy and textural relationships amongst ore minerals. Pyrite, galena, chalcopyrite and siegenite are the main sulfides. These generally occur as fine crystals, typically less than 50 microns across, or as coarse mono- or polymineralic patches of >1 cm. The latter are observed to cross-cut both slaty and crenulation cleavages at a variety of angles.

Siegenite, $(Co, Ni)_3S_4$, is the main carrier of nickel and cobalt, and also contains minor Fe (<2 wt%). This mineral is primarily present as small (<50 μm) anhedral aggregates in fine-grained, polymineralic grains, generally as inclusions within galena. It is also present as small bedding-parallel flaser-like grains. Co-Ni atomic ratios of the siegenite vary from 0.94 to 1.68 and average 1.25. Other minor minerals include digenite, bornite, plumbogummite, arsenopyrite, gersdorffite and wittichenite. Secondary minerals (malachite, cerussite, pyromorphite and minor chrysocolla) are present in surface samples.

Quartz is the principal gangue mineral. It is closely associated with sulfide minerals, either as small microcrystalline aggregates, individual euhedral crystals within sulfides, or small veinlets that cross-cut the host rock. In addition to quartz, subordinate amounts of carbonate may be present in quartz-dominant veins or, less commonly, may form separate veinlets.

Browns: Sulfur isotopes

Sulfur isotope measurement on pyrite and base metals sulfides from the Browns deposit have been provided by Bone (1985), T Donnelly (CSIRO, unpublished data 1992) and McCready *et al* (2004). There is no difference between the ore zone and non-ore zone sulfides and both have a $\delta^{34}S$ range of +14 to –5‰, whereas diagenetic pyrite from unmineralised rocks from the Crater and Whites formations are in the range +6.5 to +19.4‰.

There is no systematic variation in sulfur isotopes amongst the various sulfide species, indicating disequilibrium conditions. Ahmad *et al* (2006) have argued that the $\delta^{34}S$ range obtained for sulfides in the Browns deposit could have resulted either from biogenic and/or abiogenic reduction of seawater sulfate by sulfur-reducing bacteria or carbonaceous matter, respectively.

Browns: Lead isotopes

A discussion on the lead isotopes at Browns is included in the section on the Woodcutters deposit (see **Lead-zinc-silver veins**). Samples from the Browns deposit form a tight cluster and give an age of ca 1.6 Ga. This coincides with the age of uranium mineralisation in the PCO and of sediment-hosted base metals deposits of the Carpentaria zinc belt in the eastern Northern Territory and western Queensland. The significance of this coincidence is not clear.

All lead isotope ratios from the Browns deposit are extremely homogeneous. These ratios show that there is no difference between fine-grained bedding-parallel galena (average 16.215 ± 0.01 $^{206}Pb/^{204}Pb$, 15.497 ± 0.014 $^{207}Pb/^{204}Pb$ and 35.891 ± 0.042 $^{208}Pb/^{204}Pb$) and coarse cross-cutting galena (average 16.212 ± 0.009 $^{206}Pb/^{204}Pb$, 15.501 ± 0.14 $^{207}Pb/^{204}Pb$, 35.902 ± 0.042 $^{208}Pb/^{204}Pb$).

Browns: Uranium–base metals relationships

Uranium minerals at Browns are rare and are only observed either in association with secondary minerals, or within post-metamorphic organic matter. The latter either encloses micron-scale inclusions of uraninite that locally may form a rim around larger (50 μm) inclusions of siegenite, or uranium is present as an unresolved urano-organic complex. Other workers (Paterson *et al* 1984, Roberts 1960, Spratt 1965) described uraninite from Browns East occurring either as small spherules or disseminations, or as fracture-filling veinlets that locally show a colloform texture.

Roberts (1960) deduced from petrographic studies of ore samples from the Whites deposit that uraninite and pyrite mineralisation preceded a period of shearing, which was later followed by the introduction of Cu, Co, and Pb sulfides. This model only recognises one phase of base metals mineralisation, which took place after deformation. However, McCready *et al* (2004) presented evidence for

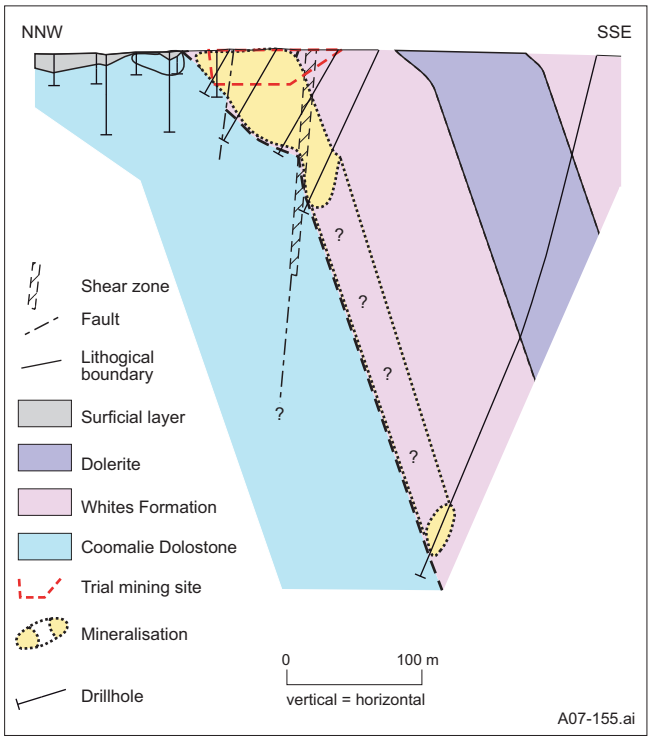


Figure 5.80. Geological cross-section of Browns deposit (after McCready *et al* 2004).

two stages of base metals mineralisation, in which the first stage was syndiagenetic and was later remobilised during the second stage to form stringers and veinlets. It is probable that no genetic link exists between uranium and base metals and that organic-rich shale has provided a suitable environment for uranium deposition (McCready *et al* 2004).

Browns: Genesis

In a study of the nearby Whites deposit, Roberts (1960) considered that the uranium mineralisation was paragenetically earlier than the base metals sulfides and was associated with the intrusion of granite. Thomas and Whitcher (1965) considered a syn-sedimentary origin for the Browns orebody. Recent mineralogical, sulfur and lead isotope studies by McCready *et al* (2004) identified at least two generations of sulfides: (a) bedding/foliation-parallel fine-grained sulfides associated with sedimentary organic matter; and (b) coarse polymetallic discordant veins. The fine-grained sulfides were interpreted as diagenetic in origin and this was supported by the sulfur isotope data, which indicated the involvement of sea water sulfate and its reduction by bacterial activity or organic matter. Later vein sulfides were suggested to have been formed by remobilisation.

Area 55

Area 55 prospect was discovered by BMR during an airborne radiometric survey in 1952. Subsequent costeaning, ground

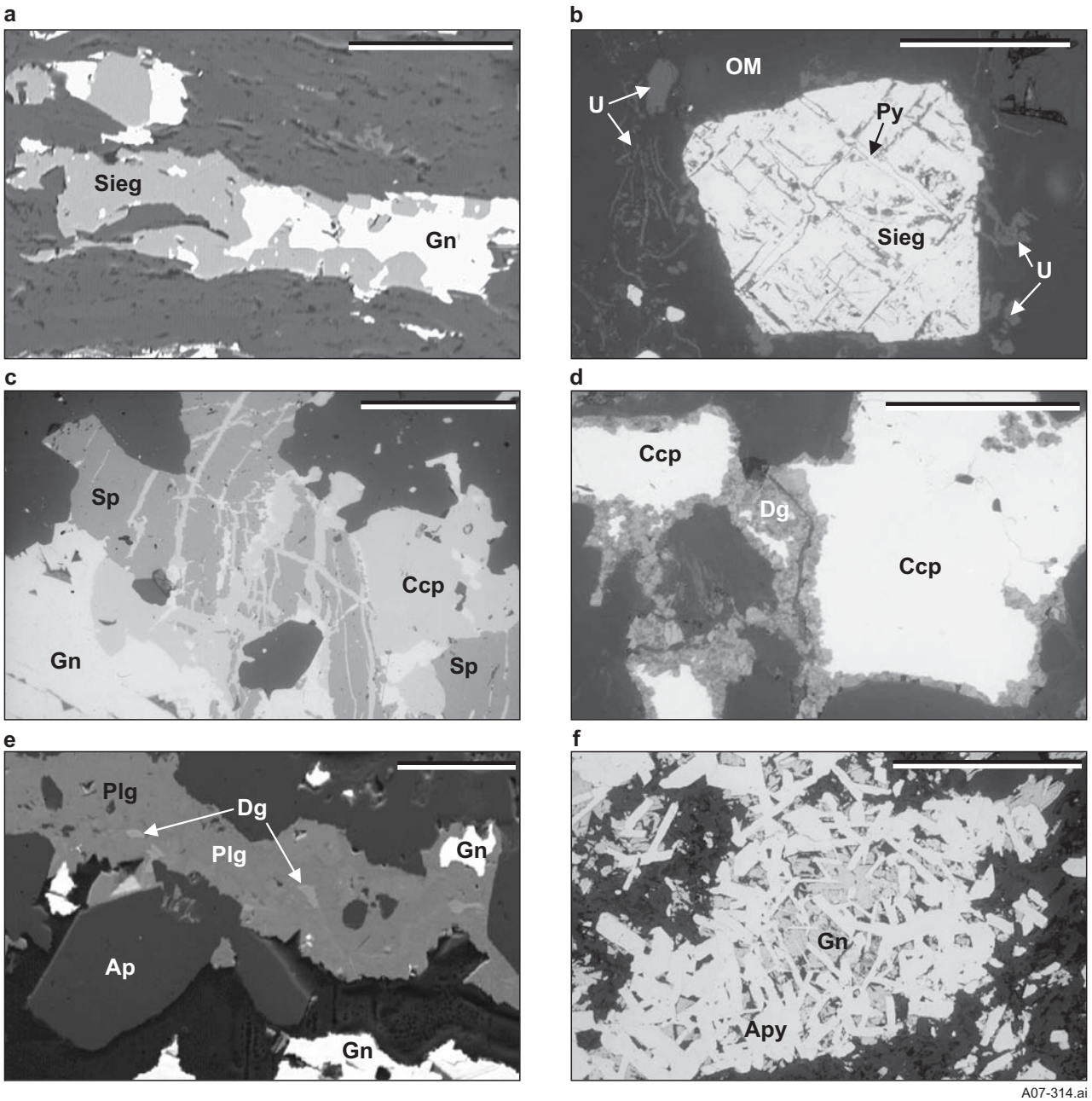


Figure 5.81. Photomicrographs and back-scattered electron (BSE) images of mineralised rocks from Browns deposit (after McCready *et al* 2004) . (**a**) Fine-grained siegenite (Sieg) intergrown with galena (Gn); scale bar is 20 mm; BSE image. (**b**) Siegenite grain (Sieg) with pyrite (Py) laths in organic matter containing uraninite (U); scale bar is 100 mm; reflected plane-polarised light. (**c**) Sphalerite (Sp) veined by chalcopyrite within galena (Gn); scale bar is 200 mm; reflected plane-polarised light. (**d**) Digenite-group minerals (Dg) mantling chalcopyrite (Ccp); scale bar is 100 mm; reflected plane-polarised light. (**e**) Ca-bearing plumbogummite (Plg) vein, with inclusions of digenite-group minerals (Dg) and galena (Gn), intergrown with apatite (Ap); scale bar is 50 mm; BSE image. (**f**) Trellis of arsenopyrite (Apy) laths in galena (Gn) matrix; scale bar is 1 mm; reflected plane-polarised light.

scintillometer surveys, geological mapping and drilling located secondary lead and copper mineralisation. Spratt (1965) provided the first resource estimate at Area 55 of about 1 Mt averaging 0.9% Cu and 4.5% Pb to a depth of 66 m. The next phase of base metals exploration in the Rum Jungle area commenced in 1990, when Compass Resources NL were granted tenements over Area 55. In 1993, Compass Resources reached a joint venture agreement with Acacia Resources Ltd, which undertook extensive exploration, including geochemical and geophysical surveys, drilling and metallurgical testing (Capp *et al* 1994). Compass Resources subsequently conducted further drilling and metallurgical testing. In 2004, the indicated resource at Area 55 was 6.6 Mt averaging 0.83% Cu, 0.17% Ni and 0.17% Co (Compass Resources 2006).

The Area 55 orebody is hosted within sericite-biotite-talc schist at the contact between the Coomalie Dolostone and Whites Formation. Minor copper mineralisation is also present in the dolostone. The Whites Formation at Area 55 comprises quartz mica schist and silicified dololutite overlain by grey shale. Mineralisation is confined to the basal mica schist unit, which is about 100 m in thickness. Structurally, the orebody is located on the eastern limb of a north-northeast-plunging syncline.

AMDEL report G831900G/94 (in Capp *et al* 1994) provides electron microprobe analyses and details of the mineralogy of three composite samples, two from secondary ore and one from primary ore. The ore minerals comprise pyrite, chalcopyrite, chalcocite, bornite and Co-Ni-Cu sulfides, mainly siegenite, carrollite and linnaeite. These sulfides range in size from 0.1–1 mm. Pyrite tends to form euhedral crystals, but other sulfides typically form anhedral grains and aggregates. Siegenite, carrollite and linnaeite generally form rims around the copper sulfides (chalcopyrite, bornite and chalcocite). Rare native copper was observed in one sample. Gangue minerals include chlorite, quartz, muscovite, talc and feldspar. Other minor minerals in the primary ore include calcite and feldspar. Secondary ore contains siderite, feldspar, alunite and rutile. Chlorite is present in two varieties: coloured, highly pleochroic chlorite with high birefringence; and colourless pale green chlorite with low birefringence. Electron microprobe analyses indicate that the coloured variety averages 7.69% Fe and 21.6% Mg, whereas the colourless variety averages 3.96% Fe and 28.19% Mg (Capp *et al* 1994). Electron microprobe analyses shows that both varieties of chlorite contain significant Co, Ni and Cu. It has been estimated that in the primary ore, >90% of Cu, Co and Ni are in sulfides, whereas in the oxidised ore, ca 50–60% of Cu, Co and Ni are in phyllosilicates.

Other polymetallic deposits

The *Mount Fitch* copper prospect lies to the east of the Mount Fitch South uranium deposit, which is described in **Rum Jungle Mineral Field**. The copper + lead + cobalt + nickel ± uranium mineralisation is at the base of the Whites Formation. Compass Resources NL has conducted an extensive drilling campaign and has estimated a JORC-compliant resource at this deposit of 5.3 Mt averaging 0.36% Cu, 0.09% Co and 0.12% Ni (Compass Resources 2008). The majority of the deposit lies within the Coomalie Dolostone rather than at the contact between the dolostone and the Whites Formation.

The *Mount Fitch North* prospect is located about 2 km north of the Mount Fitch South deposit. Acacia Resources conducted a soil sampling program and identified a Cu-Ni-Co-Mo-Bi anomaly in 1993–1994. Subsequent drilling by Compass Resources located primary Cu and Pb, and minor Co and Ni mineralisation. Several holes have intersected in excess of 10 m of greater than 2% copper, dominantly as chalcopyrite. Mineralisation is at the base of the Whites Formation in a setting similar to that of the other polymetallic deposits of the area.

At the *Mount Burton* deposit, uranium-copper mineralisation is within black shale and quartzite of the Whites Formation, close to the contact with the underlying Coomalie Dolostone. The orebody is restricted to the crest of an anticline. The ore grade diminishes rapidly below a depth of about 30 m. Ore minerals in the oxidised zone include torbernite, malachite, minor chalcocite and native copper. Pitchblende, chalcopyrite and pyrite are present in the primary ore (Berkman 1968, Compass Resources 1990).

Volcanigenic massive sulfide deposits

In 1966, BMR conducted an aeromagnetic survey in the Daly River area that revealed several anomalies unrelated to previously known base metals occurrences in this area (Shelley 1969). Drilling of these anomalies by Le Nickel Australia Pty Ltd revealed the presence of massive sulfide mineralisation (Woodland and Cotton 1977). The mineralisation consists predominantly of pyrite, pyrrhotite and sphalerite, with relatively minor galena and chalcopyrite. These minerals occur in stratiform lenses at the Anomaly A, Anomaly B and Warrs prospects. At the other prospects, mineralisation is partly associated with shears, fractures and stringers, and may be discordant or concordant.

Anomaly A and Anomaly B

Anomaly A is a sub-surface zinc deposit, located about 17 km north of the Daly River Mission. It is hosted within a sub-vertical, tightly folded, predominantly volcanic succession (about 500 m thick) of the Warrs Volcanic Member. The base of the volcanic pile is rhyolitic and it becomes increasingly basic upwards. Dacite, rhyolite, felsic tuff, andesite, basalt and minor tuffaceous phyllite, schist and agglomerate are the most common rock types intersected in drillholes. Possible pillow-lava structures have been identified in basaltic rocks (Berger 1973). Drilling intersections (**Figure 5.82**, **5.83**) have delineated eight discrete semi-conformable to conformable stacked lenses of banded massive sulfides, which strike 330°, dip 75–80° southwest and plunge about 50° northwest (Woodland 1976). Of the eight sulfide lenses outlined, reserves for the three largest (lenses 7, 2, and 3) have been calculated to a depth of 240 m and stand at 300 000 t, averaging 12.0% Zn (Woodland 1976). The primary sulfide mineral assemblage comprises pyrite, pyrrhotite, Fe-rich sphalerite, and minor chalcopyrite and galena. Traces of tetrahedrite-tennantite and native silver are present. Patchy marcasite replaces pyrrhotite in places. The ores are metamorphosed and have a distinct banded

appearance, which is, at least in part, due to metamorphic recrystallisation and segregation. Fine disseminated magnetite is common in the chloritic alteration envelopes of the ore lenses, causing the aeromagnetic anomalies that led to the discovery of this deposit. The gangue mineralogy is dominated by chlorite, followed by quartz, calcite, sericite, and minor albite, epidote, phlogopite and actinolite. Malachite, azurite, haematite and limonite are present in gossanous outcrops.

Anomaly B is present in the Warrs Volcanic Member and is essentially similar to that at Anomaly A. Drilling has intersected several narrow intervals of low-grade zinc

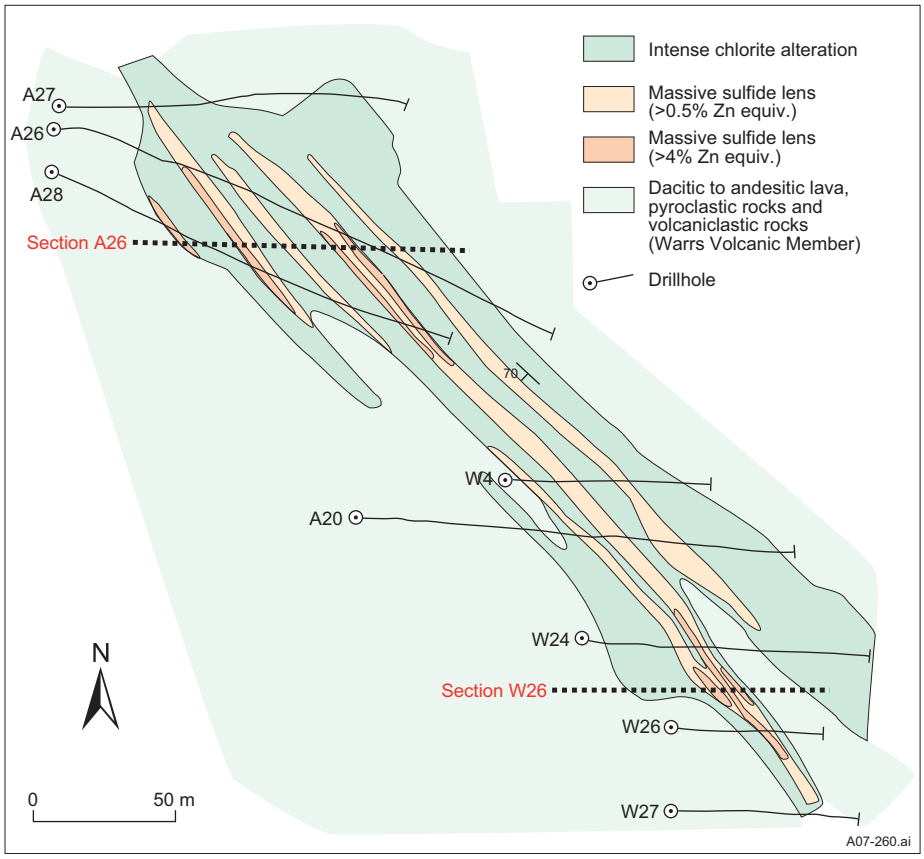


Figure 5.82. Geological plan at -150 m level at Anomaly A (after Ferenczi 2004).

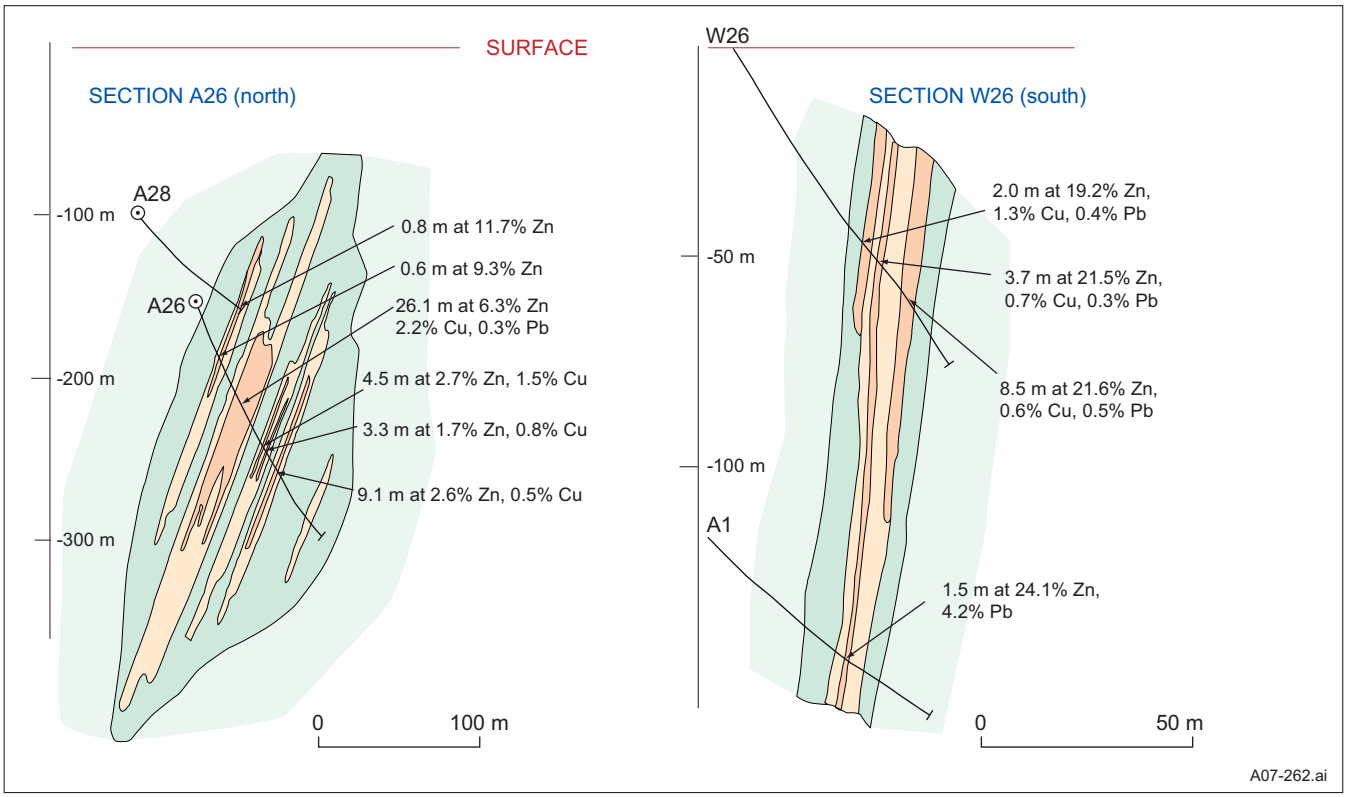


Figure 5.83. Geological cross-sections of Anomaly A deposit with assay values (modified after Ferenczi 2004 and Outback Metals 2008). See **Figure 5.82** for locations of sections and legend.

mineralisation. The best intersection was in DDH W8 (32.3 m averaging 3.4% Zn), where sphalerite and pyrite were present in sericitic stringers within chloritic schist. Minor low-grade (1.6% Zn) mineralisation was also intersected in DDH A6 as massive sulfides within andesitic tuff (Woodland 1976).

Skarns

Known skarn-related lead-zinc-silver deposits are located within marble and calc-silicate hornfels of the Koolpin Formation, adjacent to the Allamber Springs Granite, some 35 km northeast of Pine Creek township.

The *Evelyn* lead-zinc-silver mine was initially worked between 1886 and 1889, producing 610 t Pb and 2.8 t Ag (Hossfeld *et al* 1937). Mining was intermittent between 1894 and 1948, producing 4149 t of ore (Bagas 1983). United Uranium NL re-opened the mine in 1966, producing 81 554 t of ore averaging 5.5% Pb, 7.5% Zn and 276 g/t Ag until mine closure in August 1970 (Ferenczi and Sweet 2005). Some 54 t of cadmium and 56.6 kg of gold were also extracted as by-products (Bagas 1983). A remaining indicated resource of 7420 t at 6.7% Pb, 3.7% Zn and 343 g/t Ag was reported at the time of mine closure (Butler 1993b).

The workings are within a thickly bedded marble at the base of the Koolpin Formation. This marble unit occupies the hinge area of a faulted, east–west-trending asymmetric anticline. The lodes trend 340–010°, dip about 80° to the east, and are up to 4.5 m wide and 80 m long. These lodes are widest at their northern end, narrowing southwards, whereas at depth, they are known to pinch and swell (Taube 1966). The host rocks strike westerly (290°) and dip 50° to the north. Faulting is common in the vicinity of the workings and appears to control the orientation and strike extent of the orebodies. Primary mineralisation consists of cadmium-bearing sphalerite, galena, pyrite, pyrrhotite, minor chalcopyrite, arsenopyrite and rare pentlandite (Garth 1970).

North Evelyn and *NW Evelyn* are located 400 and 500 m northwest of the Evelyn mine, respectively. About 500 t of high-grade silver-lead ore was extracted from each open cut by United Uranium NL in 1967–68. The rich pods were 0.5–1 m wide and continuous over a strike length of 40 m. Weathered shale and calc-silicate hornfels formed the wallrocks (Taube 1966). A combined remaining resource of 38 100 t at 3.7% Pb and 193 g/t Ag has been estimated for the deposits (Williams 1998).

Tin and tantalum

Tin in the PCO is present either as a single commodity, or in association with tantalum. The deposits occur as quartz veins, quartz-sulfide veins, tourmaline-quartz veins, pegmatites and greisens. Significant tin deposits are known from the Mount Wells, Mount Masson, Hayes Creek, Jessops, Mount Harris, Maranboy and Horseshoe areas. Tin-tantalum-bearing pegmatites are present in a belt extending from Bynoe Harbour to Mount Tolmer. Another cluster of these pegmatites is in the Mount Shoobridge area (**Figure 5.84**). All occurrences are within the contact aureole of granites and most are hosted within the Burrell Creek Formation. Summaries of the more significant Sn-W-Mo deposits and Sn-Ta-Nb pegmatite deposits of the PCO are in **Tables 5.10** and **5.11**, respectively.

Tin mineralisation was discovered in 1879 at Mount Wells and was soon followed by discoveries of tin-tantalum-bearing pegmatites in the Bynoe Harbour area. Over one hundred Sn veins and more than 150 Sn ± Ta-bearing pegmatite bodies are now known. The latter are concentrated in the Mount Shoobridge and Bynoe areas. Previous studies on these deposits are limited to surface and underground sampling, production, ore reserves, grade, lode structure and mineralogy (Summers 1957, Crohn 1968, Needham *et al* 1980, Stuart-Smith 1985, Frater 2005). Ahmad (1991a, b) has elaborated on the genesis of tin mineralisation in the Mount Wells area and Sn-Ta pegmatites of the Bynoe area. On the basis of shape and mineralogy, the tin and tin-tantalum occurrences of the PCO can be divided into the following classes:

- Sn-quartz veins with minor sulfides.
- Sn-sulfide veins with up to 80 volume % sulfides.
- Sn-tourmaline-quartz veins.
- Sn-Ta pegmatites.
- Greisens.
- Alluvial deposits.

Sn-quartz veins with minor sulfides

These veins, hosted by sedimentary rocks of the South Alligator or Finniss River groups, generally trend north-northeast and have sub-vertical dips. Cassiterite is the principal economic mineral and it is associated with pyrite, arsenopyrite, chalcopyrite, minor wolframite and molybdenite. Traces of gold are present in some veins. Quartz is the principal gangue mineral. The more significant Sn-quartz vein deposits of the PCO are summarised in **Table 5.10**.

Mount Wells

The Mount Wells mine is located some 38 km north-northeast of Pine Creek township and was worked intermittently from 1879–1929 for a recorded production of 1555 t of SnO_2 concentrate from ore, which had an average grade of 1% Sn. In 1917, 7.1 t of hand-sorted copper ore averaging 37% Cu was also produced (Crohn 1968, Mookhey 1971). Investigations in 1957–1958 and 1966–1977 included nearly 2000 m of diamond drilling, 101 m of exploratory underground development and geological mapping. Mookhey (1971) calculated probable reserves at 800 000 t averaging 1.2% Sn. Robinson (1986) recalculated the reserves and reported 361 290 t of probable ore at 1.46% Sn and 375 020 t possible ore at 1.3% Sn. Some lodes contain up to 0.5% Cu. Additional percussion holes were drilled by Territory Resources Ltd in 1988–1989 to better define these reserves. The current underground ore resource is stated to be 737 000 t at 1.38% Sn and there is an open-cut resource of 400 000t at 0.4% Sn. The deposit also contains a target copper resource of 950 000–971 000 t at 1.0–1.5% Cu (Outback Metals 2008).

Six lodes have been identified at Mount Wells and represent tension-fill quartz veins within siltstone and

greywacke of the Burrell Creek Formation (**Figure 5.85**). The width of the lodes ranges from 0.5–5 m, averaging about 2 m. The lodes are arranged in an en echelon pattern. Contacts with the host rock are sharp and there is no evidence of displacement or movement, although the lodes near the contact are more fractured than in the centre. Branching of lodes into three or four narrow veins is usual and, towards their northern and southern ends, they break into several millimetre-thin stringers.

Cassiterite occurs as single coarse crystals or as aggregates, usually along the hangingwall of most lodes. The centre of the lodes is generally barren and the footwall is marked by haematite (oxidised sulfides)-quartz breccia, which also contains cassiterite (Robinson 1986). This breccia zone was also seen to jump from footwall to hangingwall in the No 2 and No 3 levels, and may represent late-stage sulfide-rich mineralisation. In the primary ore, pyrite, chalcopyrite, arsenopyrite, and pyrrhotite are common. Rare patches of wolframite and molybdenite are also present. Common gangue minerals are quartz, muscovite, tourmaline, feldspar and chlorite. Within the oxidised zone, which extends down to 50 m, the lode material comprises quartz, haematite, limonite, cassiterite, malachite, chalcocite, bornite, covellite and scorodite.

Diamond drillholes have indicated lode continuity to a depth of about 200 m and have intersected a cupola of greisenised biotite-muscovite granite, at a depth between 150–200 m. Most lodes apparently do not continue into the granite, but minor quartz-cassiterite veins are present within the granite and one such vein has been correlated with the Main Lode (Mookhey 1971).

Hayes Creek

This mine has produced about 156 t of cassiterite concentrate (Dunn 1963, Crohn 1968, Stuart-Smith 1985). The Sn-quartz lodes are similar to those at Mount Wells and can be discontinuously traced for a distance of about 0.5 km. These lodes are generally concordant to the bedding and dip steeply west, but some discordant veins are also present. The

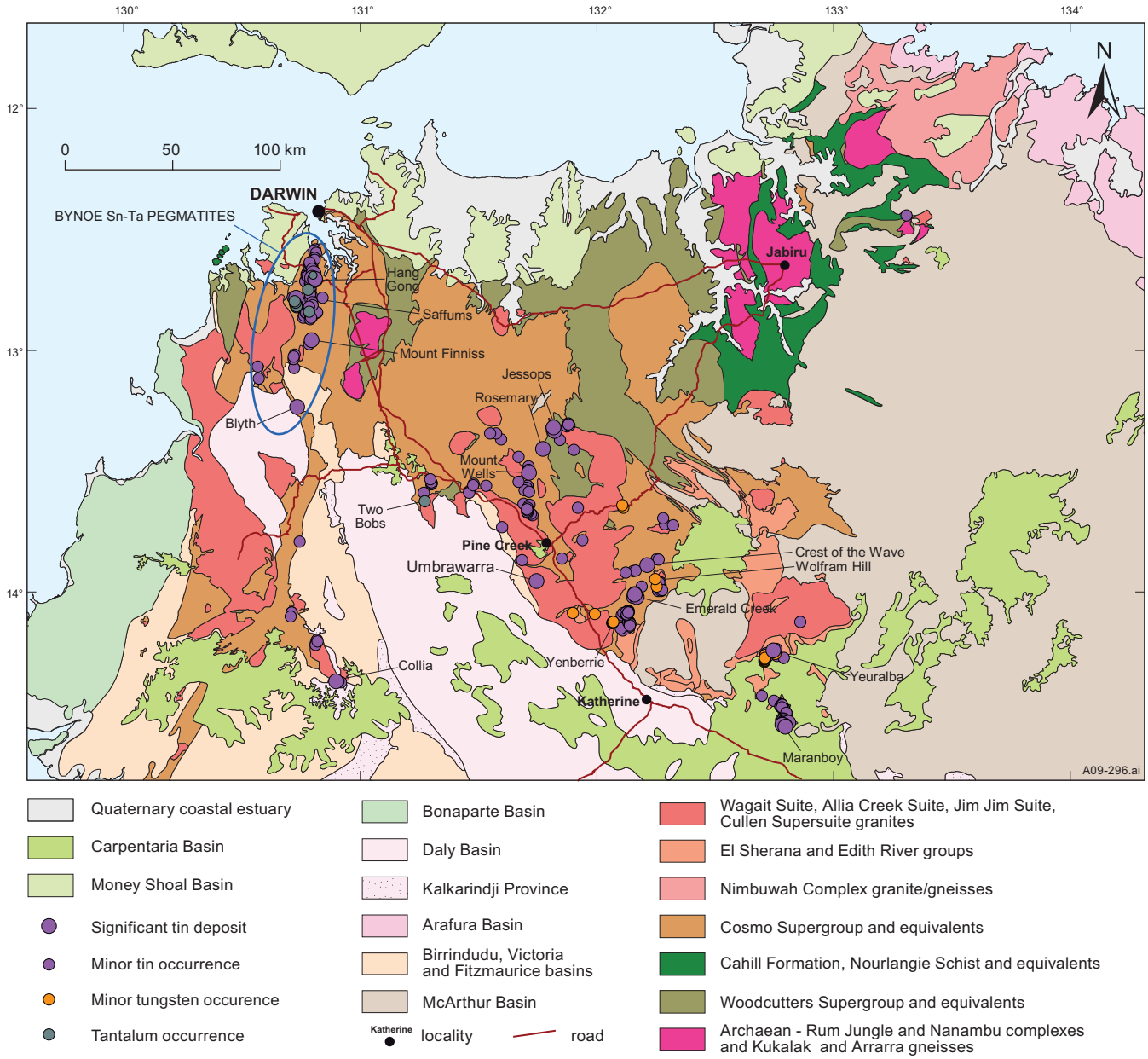


Figure 5.84. Location of tin, tantalum, tungsten and molybdenum deposits of the PCO. Data from NTGS MODAT database. Significant deposits are labelled.

cassiterite is coarse grained, with individual crystals up to 2 cm across. Iron oxides (gossanised sulfides) are common, but unlike Mount Wells, there are no secondary copper minerals. The width of the lodes ranges from 0.5–3 m and, as at the Mount Wells mine, the contacts with the host rock are sharp.

Jimmys Knob

The Jimmys Knob tin mine (also known as *Deans Camp*) is located some 31 km northwest of Pine Creek township and was initially worked during the late 1880s (Jensen *et al* 1916), but no production records are available. Production between 1964 and 1977 yielded 1.40 t of SnO_2 concentrate (Bagas 1981). High-grade mineralisation is restricted to narrow sub-vertical lodes with little persistence below 20 m under the existing surface workings (Newton 1980). In 1978–1979, Geopeko Ltd carried out an extensive exploration program, which included 1313 m of diamond drilling. The results were disappointing, but did show broad zones of higher-than-background Sn in acid intrusive rocks (Goulevitch 1979, 1980a). In 1983, CRA Exploration Pty Ltd conducted rock chip and soil geochemical surveys

Common name	Lat	Long	Shape	Model	Commodities	Rock unit	Resource	Grade	Production concentrate	Reference
Barretts	-13.55	131.3	Tabular	Pegmatite	Sn	Burrell Creek Fm			117 t SnO_2	Crohn (1968)
Billycan	-13.32	131.82	Sheet	Sn-sulfide vein	Sn	Wildman Siltstone			38 t SnO_2	Wygralak (1983)
Crest of the Wave	-13.89	132.22	Sheet	Sn-sulfide vein	Sn	Burrell Creek Fm			155 t SnO_2	Bagas (1983)
Emerald Creek alluvials	-14.01	132.16	Irregular	Placer	Sn	Czs			243 t SnO_2	Kruse *et al* (1994)
Hayes Creek	-13.58	131.46	Sheet	Sn-Qtz vein	Sn	Mount Bonnie Fm			156 t	Crohn (1968)
Hidden Valley field	-13.99	132.27	Sheet	Sn-sulfide vein	Sn	Burrell Creek Fm			51 t SnO_2	Crohn (1968)
Horners Creek alluvials	-13.49	131.71	Irregular	Placer	Sn, Au	Czs			47 t SnO_2	Ahmad *et al* (1993)
Horseshoe field	-14.10	132.13	Irregular	Sn-sulfide vein	Sn	Czs			660 t SnO_2	Kruse *et al* (1994)
Jessops	-13.32	131.82	Sheet	Sn-sulfide vein	Sn	Wildman Siltstone			193 t SnO_2	Ahmad *et al* (1993)
Jimmys Knob	-13.57	131.7	Sheet	Sn-Qtz vein	Sn	Mount Bonnie Fm			1.4 t SnO_2	Ahmad *et al* (1993)
Margaret	-13.30	131.88	Sheet	Sn-sulfide vein	Sn	Mundogie Sandstone			62 t SnO_2	Crohn (1968)
Maranboy field	-14.48	132.79	Sheet	Sn-tur-qtz vein	Sn	Tollis Fm	170 000 t	1.06% Sn	1283 t SnO_2	Crohn (1968)
Mount Masson	-13.33	131.82	Sheet	Sn-sulfide vein	Sn	Wildman Siltstone			31.5 t SnO_2	Crohn (1968)
Morris field	-14.13	132.14	Sheet	Sn-sulfide vein	Sn	Tollis Fm			180 t SnO_2	Crohn (1968)
Mount Wells	-13.50	131.71	Sheet	Sn-Qtz vein	Sn, Cu	Burrell Creek Fm	1.13 Mt	1.03% Sn	1555 t SnO_2	Outback Metals (2008)
Mount Shoobridge	-13.53	131.29	Tabular	Pegmatite	Sn	Burrell Creek Fm			145 t SnO_2	Crohn (1968)
Rosemary	-13.41	131.77	Sheet	Sn-Qtz vein	Sn	Gerowie Tuff	80 000 t	0.68% Sn	26 t SnO_2	Wygralak (1983)
Umbrawarra	-13.95	131.75	Irregular	Placer	Sn	Czs			267 t SnO_2	Ahmad *et al* (1993)
Wandie Creek	-13.92	132.12	Irregular	Placer	Sn	Czs	8000 t	220 g/t Sn		Ferenczi and Sweet (2005)
Wolfram Hill	-13.94	132.25	Tabular	Pegmatite	W, Cu, Sn	Burrell Creek Fm			740 t WO_4, 65 t SnO_2	Ferenczi and Sweet (2005)
Yenberrie	-14.12	132.07	Sheet	Geisen	W, Mo, Bi, Sn	Yenberrie Leucogr			161 t WO_4	Crohn (1968)
Yeuralba field	-14.27	132.8	Sheet	Wol ± Sn ± Mo-qtz vein	Sn	Tollis Fm			19.6 t WO_4, 8.2 t SnO_2	Crohn (1968)

Table 5.10. Summary data on significant PCO Sn-W-Mo deposits. Note that all resources given in the table are non-JORC compliant. Abbreviations: Lat = latitude; Long = longitude; Fm = Formation; Leucogr = Leucogranite; Mo = molybdenum; Qtz = quartz; Tur = tourmaline; Wol = wolframite.

Common name	Lat	Long	Shape	Rock unit	Resource	Grade	Production concentrate
Observation Hill group					1 761 870 t pegmatite 1 084 000 t alluvial	0.0189% SnO_2, 0.0081% Ta_2O_5 0.0187% SnO_2, 0.0047% Ta_2O_5	101 t SnO_2 45 t Ta_2O_5
Bells Mona	-12.71	130.76	Tabular	Burrell Creek Fm	9000 t	0.0 446% SnO_2 and 0.0011% Ta_2O_5	95 t SnO_2
Booths	-12.70	130.80	Tabular	Burrell Creek Fm	62 900 t	0.012% SnO_2, 0.0039% Ta_2O_5	NSR
BP 6	-12.65	130.78	Tabular	Burrell Creek Fm	1320 t	167 g/t SnO_2	
Grants	-12.67	130.78	Tabular	Burrell Creek Fm	0.124 Mt	0.0209% SnO_2 and 0.0015% Ta_2O_5	3.15 t SnO_2
Hang Gong	-12.67	130.79	Tabular	Burrell Creek Fm	0.222 Mt	0.0083% SnO_2, 0.0059% Ta_2O_5	200 t SnO_2
Hendersons	-12.68	130.80	Tabular	Burrell Creek Fm	50 000 t	0.041% SnO_2, 0.002% Ta_2O_5	NSR
Johnstones	-12.67	130.78	Tabular	Burrell Creek Fm	0.034 Mt	0.0129% SnO_2, 0.0025% Ta_2O_5	NSR
Lees	-12.70	130.79	Tabular	Burrell Creek Fm	0.250 Mt	0.06% SnO_2, 0.01% Ta_2O_5	NSR
Yan Yam	-12.69	130.80	Tabular	Burrell Creek Fm	40 000 t	0.05% SnO_2, 0.026% Ta_2O_5	NSR
Leviathan group					14 4570 t pegmatite 42 000 t alluvial	0.0104% SnO_2, 0.0085% Ta_2O_5 0.0212% SnO_2, 0.0032% Ta_2O_5	
Angers	-12.81	130.72	Tabular	Burrell Creek Fm	0.15 Mt	0.0280% Ta_2O_5, 0.0798% SnO_2	NSR
Leviathan	-12.79	130.72	Tabular	Burrell Creek Fm	3000 m^3	0.1% SnO_2, 0.04% Ta_2O_5	170 t SnO_2, 19 t Ta_2O_5
Mackas Reward	-12.80	130.72	Tabular	Burrell Creek Fm	6920 t	0.0112% SnO_2, 0.0175 Ta_2O_5	
North Phoenix	-12.79	130.72	Tabular	Burrell Creek Fm	1160 t	0.04% SnO_2, 0.003% Ta_2O_5	
Old Buck	-12.78	130.72	Tabular	Burrell Creek Fm	4000 m^3	0.4% SnO_2, 0.1% Ta_2O_5	NSR
River Annie pegmatite group					205 700 t pegmatite 59 000 t alluvial	0.0249% SnO_2, 0.0067% Ta_2O_5 0.0297% SnO_2, 0.0028% Ta_2O_5	
Bilatos (Pickets)	-12.85	130.76	Tabular	Burrell Creek Fm	10 000 m^3	0.04% SnO_2, 0.008% Ta_2O_5	NSR
Fred	-12.85	130.77	Tabular	Burrell Creek Fm	5000 m^3	0.02% SnO_2, 0.06% Ta_2O_5	
Martins	-12.86	130.79	Tabular	Burrell Creek Fm	5000 m^3	0.06% SnO_2, 0.008% Ta_2O_5	NSR
Saffums 1	-12.86	130.75	Tabular	Burrell Creek Fm	10 000 m^3	0.30 kg/m^3 SnO_2, 0.08 kg/m^3 Ta_2O_5	15.62 t Ta_2O_5, 17.68 t SnO_2
Sandras	-12.87	130.78	Tabular	Burrell Creek Fm	5000 m^3	0.4 kg/m^3 SnO_2, 0.03 kg/m^3 Ta_2O_5	NSR
Turners	-12.86	130.79	Tabular	Burrell Creek Fm	1000 m^3	0.01% SnO_2, 0.01% Ta_2O_5	NSR
Kings Table pegmatite group					43640 t	0.0099% SnO_2, 0.0038% Ta_2O_5	
Kings Table	-12.58	130.80	Tabular	Burrell Creek Fm	25350 t	0.0039% SnO_2, 0.0047%Ta_2O_5	NSR
Jewellers	-12.62	130.80	Tabular	Burrell Creek Fm	8380 t	0.0119% SnO_2, 0.0029% Ta_2O_5	NSR
Jewellers Extended	-12.63	130.79	Tabular	Burrell Creek Fm	2600 t	0.0139% SnO_2, 0.0021% Ta_2O_5	
Perseverance	-12.62	130.80	Tabular	Burrell Creek Fm	7310 t	0.0269% SnO_2, 0.0026% Ta_2O_5	
Muldiva, Buldiva and Collia	-14.20	130.81	Tabular	Burrell Creek Fm Soldiers Creek Granite			45.7 t SnO_2
Miscellaneous deposits							
Mount Finniss	-12.95	130.79	Tabular	Burrell Creek Fm	160 000 t pegmatite	0.016% SnO_2,, 0.0035% Ta_2O_5	165 t SnO_2, 55 t Ta_2O_5
Bamboo Creek	-13.07	130.71	Irregular	Burrell Creek Fm			46 t SnO_2 50 kg Ta_2O_5
Goodwill Extended	-13.02	130.72	Irregular	Burrell Creek Fm			56 t SnO_2 + Ta_2O_5
Blyth	-13.23	130.73	Tabular	Burrell Creek Fm			76 t SnO_2
Labelle	-13.11	130.56	Irregular	Welltree Metamorphics	250 000 t	0.007% SnO_2, 0.012% Ta_2O_5	

Table 5.11. Summary data on significant Sn-Ta-Nb bearing pegmatites of Litchfield Domain of PCO (data from Frater 2005). Note that all resources given are non-JORC compliant. SnO_2 and Ta_2O_5 represent cassiterite and tantalite concentrates, respectively. Abbreviations: Lat = latitude; Long = longitude; Fm = Formation; NSR = Not separately recorded.

aimed at identifying a low-grade large-tonnage resource, but the results were discouraging (Harvey 1983).

Mineralisation is within quartz-filled fractures at, or close to, the contact between carbonaceous shale, greywacke and siltstone of the Mount Bonnie Formation and a predominantly quartz-syenite intrusive rock. The lodes range in width from a few centimetres to about one meter and are composed of quartz and iron oxides (oxidised sulfides). Cassiterite is coarse and is erratically distributed, but some fine cassiterite is also present.

Other Sn-quartz veins occurrences

At the *Mavis* mine, bedding-parallel quartz cassiterite veins, 0.2–1 m thick and trending 310°, can be traced for a distance of 200 m. The cassiterite is coarse and pink-grey, and crystals up to 10 mm across are common. At the *Lewis* prospect, cassiterite, malachite and galena are present in a north-trending quartz vein within greywacke and siltstone of the Burrell Creek Formation. At the *Horseshoe* mine, thin (0.1–0.3 m) cassiterite quartz veins trend northwest and cover an area of 50 x 100 m in carbonaceous shale of the upper Koolpin Formation. At the *Teacup* mine, a cassiterite-quartz-muscovite vein trends northeast and is hosted by a basic dyke. At the *Mundic* mine, cassiterite-quartz veins up to 1 m thick, trend east and are exposed over a strike length of about 200 m. At the *Sneddens Creek* mine, mineralisation is within northwest-striking, steeply-dipping stringers. At the *Rosemary* mine, a haematite-quartz mineralised lode trends 320°, dips 75° east and extends over a length of about 400 m. This mine has been tested by drilling and underground development, and has produced 26.8 t of SnO_2 concentrate. Reserves are estimated to be about 80 000 t of ore at 0.68% Sn (Newton 1979).

Sn sulfide veins

These veins are hosted by sedimentary rocks of the Mount Partridge Group. In the oxidised zone, which extends down to about 50 m, the lodes are composed of up to 80 volume % iron oxides and contain angular fragments of vein quartz and host rocks. The composite term haematite-quartz-breccia has been used to describe these lodes. The cassiterite is very fine-grained and individual grains are coated with a thin film of iron oxide, usually limonite. Within the oxidised zone, the mineralogy is dominated by iron oxides, including haematite, limonite and goethite. The primary ore is composed of pyrite, and arsenopyrite and quartz. The more significant Sn-sulfide vein deposits of the PCO are summarised in **Table 5.10**.

Jessops and Billycan

The Jessops and adjoining Billycan (southern extensions of the Jessops lode) mines are located about 56 km north of Pine Creek township and were discovered in 1957. They were worked until 1972, for a total production of 193 t SnO_2 concentrate from the *Jessops* mine and 38 t SnO_2 concentrate from the *Billycan* mine (Wygralak 1983). Mine workings included a large open cut, shaft, adit and several costeans. The orebody has a strike length of about 365 m, trends 350°, dips about 60° west and is 1–1.5 m thick.

Hays (1960) described the nature of the near-surface oxidised portion of the lodes. The lode material consists of angular fragments of vein quartz and country rocks, ranging in size from small grains to large boulders, and cemented by iron oxides, mostly limonite and minor haematite. Cassiterite is generally not visible to the naked eye, except in very rich ore, and grains range in size from 0.06–1.5 mm. Crohn (1968) suggested that the haematite-quartz breccia

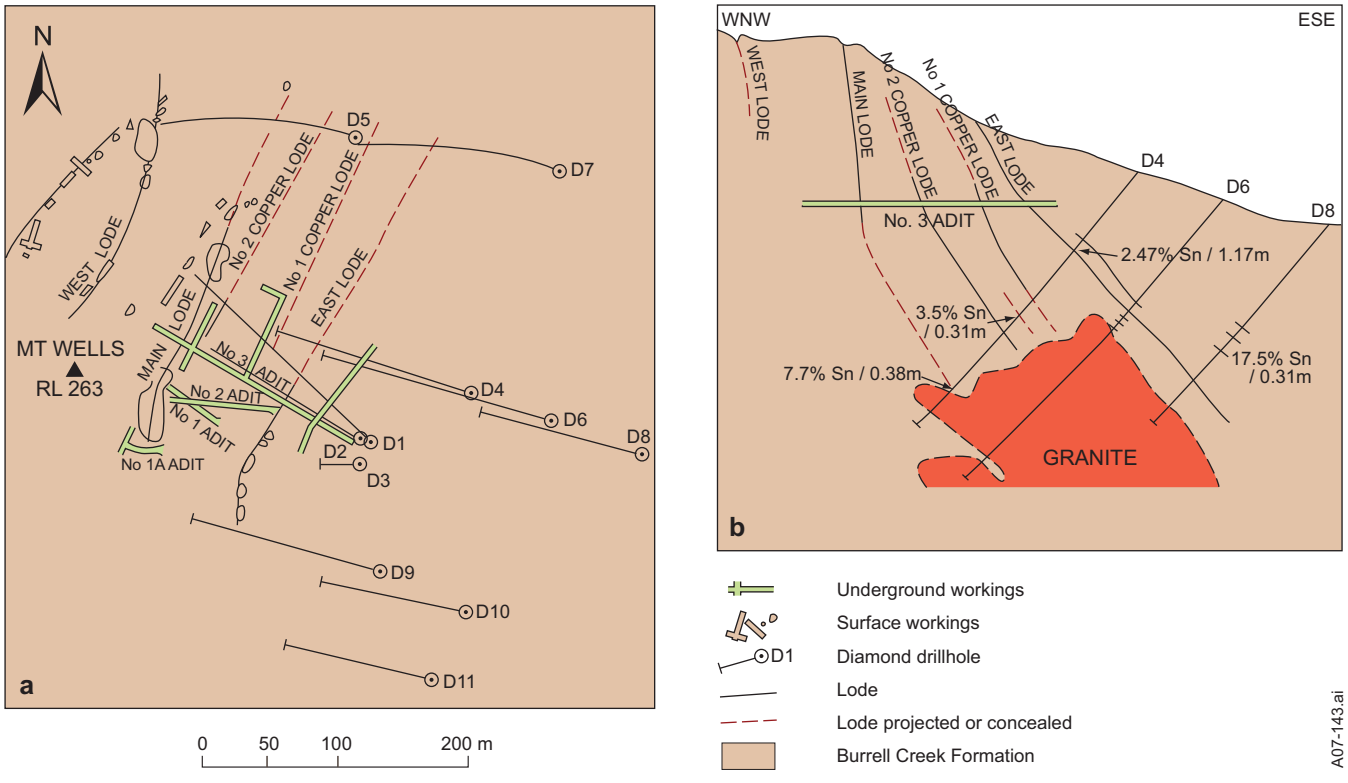


Figure 5.85. Geological plan and cross-section of Mount Wells deposit (modified from Mookhey 1971).

was formed by collapse of the original sulfide-rich lode after a decrease in volume resulting from oxidation and leaching. This was confirmed in 1981–1982, after diamond drilling conducted by CRA Exploration Pty Ltd, which showed that primary lode consists of pyrite, arsenopyrite and quartz. This drilling also yielded anomalous values of Sn, Au, Ag, As and base metals. The best gold values ranged from 0.25–5.1 g/t. However, follow-up drilling and sampling in 1988 yielded gold values that were generally less than 0.5 g/t (Orridge 1988).

Mount Masson

The Mount Masson mine is located about 2 km southeast of the Jessops mine and was intermittently worked between 1942 and 1968, for a total production of 31.5 t SnO_2 concentrate from 2916 t of ore (Crohn 1968). The main lode outcrops on the southern flank of Mount Masson over a strike length of 120 m. A second parallel lode is located some 60 m further east. Both lodes trend 350°, dip steeply west, and are situated on the same lineament that contains the Jessops lode. The width of the lode ranges from a few centimetres to about one metre.

The near-surface mineralogy and lode characteristics are similar to those of the Jessops lode. No deep holes have been drilled. Some shallow-dipping to sub-horizontal quartz veins containing trace sulfides are present in the adjacent wallrock. A few of these stringers cut across the haematite-quartz-breccia lodes and are therefore apparently younger (Ahmad *et al* 1993).

Hidden Valley tinfield

This field is located about 50 km southeast of Pine Creek township and was mined between 1905 and 1955, for about 51 t of cassiterite concentrate (Kleeman 1938, Walpole 1962, Bagas 1983). Mineralised veins and stockworks are contained within metagreywacke and phyllite of the Burrell Creek Formation. The mineralogy of the lodes comprises goethite, haematite, minor vein quartz and trace malachite. Cassiterite is fine to medium, with a grain size ranging to 3 mm in diameter. Vein quartz is insignificant in the lode material, but late, barren massive quartz veins, up to 2 m wide, crosscut the mineralised lodes at a few places.

Horseshoe, Morris and Boyling tinfields

These tinfields lie immediately northeast of Mount Todd and were described by Jensen (1919), AGGSNA (1937) and Rattigan and Clark (1955). Recorded production (as cassiterite concentrate) for the period 1906–1921 is reported at 660 t from the Horseshoe and 180 t from the Morris tinfields (Crohn 1968). There are no production records available for the early 1980s, but field observations suggest that more than 100 t of cassiterite concentrate was obtained. Although 40 mineral occurrences have been located within these tinfields, most of the production is from the *Morris, Boyling, Doris, Marie, Irwin* and *Scotchman* mines.

Mineralisation is associated with up to 20 cm-wide, en echelon sheeted veins, comprising haematite, goethite and quartz; minor malachite and tourmaline are present in some veins. Cassiterite is present as small (up to 2 mm in size) euhedral black crystals, often forming almost monomineralic stringers, but most commonly, it is associated with iron oxides, especially goethite. Cassiterite is also associated with 0.5–1 m-thick veins of white, massive quartz, which outcrop as pods, often capping the tops of hills.

Other occurrences

The *Big Drum* and *Big Julie* mines represent southward extensions of the Mount Masson lodes. Both mines are in haematite-quartz-breccia. Sampling at the *Big Julie* mine (Newton 1977) also yielded up to 22 g/t Au. At the *Mount Harris* leases, (*Nelson 1*, *Nelson 2*, *Margaret*, *Buffalo*), the lodes have a smaller strike length and are discontinuous. They are composed of haematite-quartz breccia similar to that described earlier. The *Margaret* mine is the biggest and has produced some 62 t of SnO_2 concentrate (McQueen 1956), but this figure probably also includes production from *Nelson 2*. Mineralised veins and stockworks are contained within metagreywacke and phyllite of the Burrell Creek Formation. The mineralogy of the lodes comprises goethite, haematite, minor vein quartz and trace malachite. Cassiterite is fine to medium, with a grain size ranging to 3 mm in diameter. The *Crest of the Wave* mine was worked intermittently between 1907 and 1923, briefly re-opened in 1934 and was worked again in 1980–1981. Total recorded production was reported to be 155 t of cassiterite concentrate from 3027 t of ore (Bagas 1983). The lodes are mineralogically similar to those in the Hidden Valley tinfield. Iron oxides (goethite, haematite) are predominant, followed by quartz, minor tourmaline and fine muscovite. Cassiterite is present as well formed crystals up to 3 mm in size. Occasional stringers of almost monomineralic cassiterite have been noted. The primary ore consists of cassiterite and quartz with minor arsenopyrite and pyrite (Kleeman 1938).

Tin-tourmaline-quartz veins

These veins are hosted by contact metamorphosed sedimentary rocks of the Tollis Formation. Cassiterite is contained in fine-grained, 2–8 m-wide quartz-tourmaline lodes. Minor sulfides, including chalcopyrite and pyrrhotite, as well as native bismuth, are present. Some lodes adjacent to the granite contact also contain mica. All the occurrences are within the Maranboy tinfield, which is briefly described below. The more significant Sn-tourmaline-quartz vein deposits of the PCO are summarised in **Table 5.10**.

Maranboy tinfield

The Maranboy tinfield lies 65 km southeast of Katherine and was discovered in 1913. It was worked intermittently from 1913 until at least 1961. It produced 1280 t of cassiterite concentrate between 1913 and 1952 (Crohn 1968) from about 50 000 t of ore. Walpole (1958) assessed the remaining reserves to a depth of 43 m (the depth of deepest workings) as being 170 000 t of ore averaging 1.06% Sn.

Outback Metals (2008) considered that the Maranboy tinfield has a potential resource of 722 000 t of ore at 2.20% Sn.

The field is located within a small inlier of the Tollis Formation, which is locally intruded and hornfelsed by

the Maranboy Porphyry and Yeuralba Granite. The major structure of the field is a 340°-trending, southeast-plunging anticline.

Most of the tin is contained in fine-grained tourmalinite lodes, which are structurally controlled by two systems of strike-slip fractures striking 290° and 340°. Lodes following the first system (290°) are richer in cassiterite. Along strike, the lodes are divided into several parts that have been dislocated by post-ore faults striking northeast to north-northeast. The lodes are hosted by hornfels and are composed of tourmaline and quartz. Two lines of lodes, the *Stannum King Lode* (including *Ibis, Stannum King, Red Cross* and *Progress* prospects) and the *Main Lode* (including *Star of the West, Osman, Anaconda, Ray, Eureka, Bull* and *Southern Claim* prospects), situated about 1 km apart are known. The Main Lode has a total strike length of about 4 km and an average width of 2–8 m, whereas the Stannum King Lode extends for a strike length of about 2 km.

Other types of lodes, known from the northern part of the field, are associated with quartz-mica greisens, developed along the contact between the sedimentary rocks and the Yeuralba Granite. In both types, the only economical mineral is cassiterite, which occurs most commonly in groups of small stringers in joints and zones of brecciation. Less frequently, it is finely disseminated. Other minerals present include chalcopyrite, pyrrhotite and bismuth.

Sn-Ta pegmatites

Sn-Ta-bearing pegmatites are present in a north-trending up-to-10 km-wide belt, extending from Bynoe Harbour to the Wingate Mountains. Another cluster of pegmatites is present in the Mount Shoobridge area. The northern part (**Figure 5.86**) of the Bynoe belt is the most prospective area and over 90 mineralised pegmatite bodies have been identified in this region, including the *Mount Finniss* mine, the largest Sn-Ta producer of the Northern Territory. Production, resources and grades of significant deposits are given in **Table 5.11**.

Published information on these pegmatites is scanty. Summers (1957) gave a brief account of 34 pegmatites in the Bynoe belt and provided sketch maps of some occurrences. Fisher (1964) and Burke (1987) provided some details on the pegmatites in the Mount Shoobridge area. Pietsch and Clayton (1990) gave a generalised account of pegmatites in the Bynoe belt. Ahmad (1991a, 1993) provided results on fluid inclusion, stable isotope and geochemical studies. More recently, Frater (2005) studied the geochemistry, structure and geochronology of the pegmatites. Exploration of these pegmatite bodies has been conducted by Greenbushes Ltd (1980s), Corporate Developments Pty Ltd (1990s) and more recently (2001–2002), by Julia Corporation Ltd. These exploration activities are summarised in Frater (2005).

The pegmatite bodies are linear or lensoidal, and up to 10 m wide and 250 m long. Most are poorly exposed and extensively altered. Usually, the feldspar is completely decomposed to white kaolinitic clay and has been removed by erosion, leaving behind rubbly or bouldery remnants of white massive quartz. Mine workings are confined to a depth of less than 25 m in heavily altered material.

Quartz, muscovite and feldspar are the most common gangue minerals and their distribution suggests a crude zonation, in which a border zone of quartz + muscovite is followed by an intermediate zone of feldspar + muscovite + quartz. A massive quartz core is seen in some pegmatites.

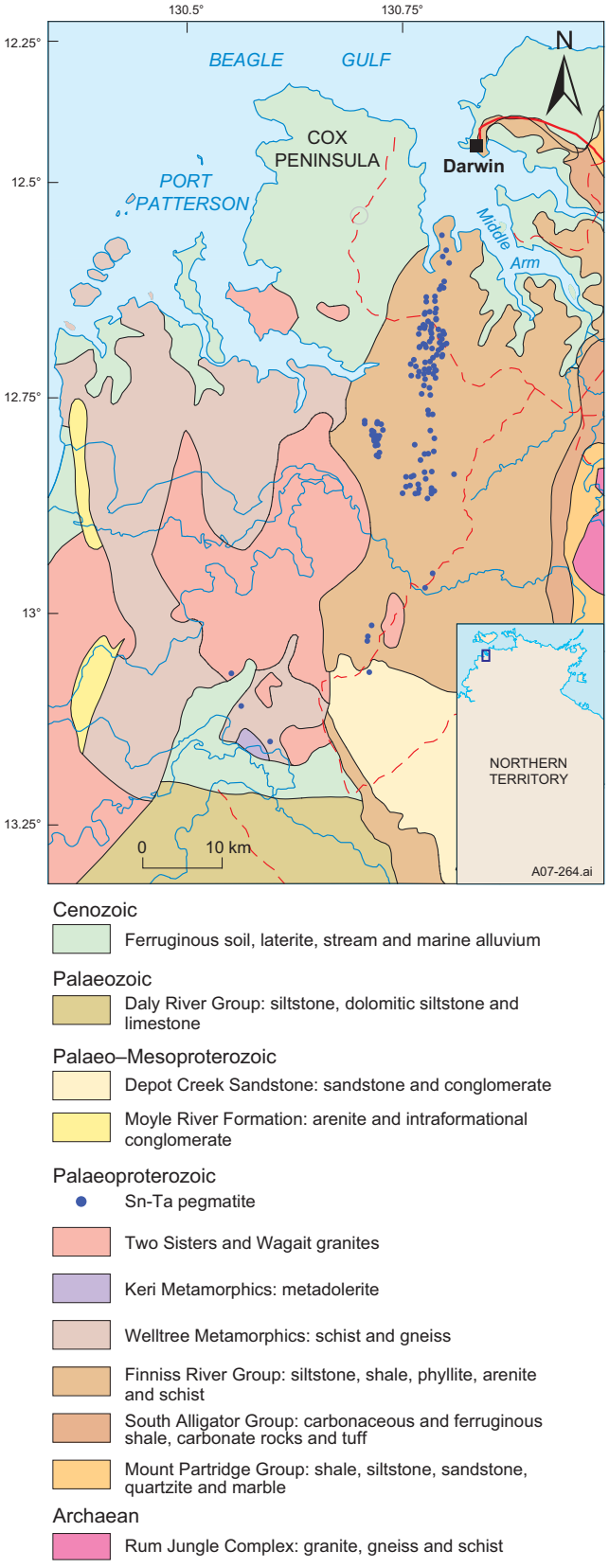


Figure 5.86. General geological setting of northern portion of Bynoe pegmatite belt, western PCO.

Ore minerals include cassiterite, tantalite and columbite, occurring as coarse crystals or aggregates of crystals (**Figures 5.87a**, **b**, **c**). Minor minerals include amblygonite, montebrasite, wordite, augelite, rutile, ilmenite, magnetite, zircon and tourmaline. Beryl is present in significant proportions at the Labelle pegmatite.

The wallrocks are siltstone and greywacke of the Burrell Creek Formation, which are regionally metamorphosed to middle to upper greenschist facies. Granitoids adjacent to the Bynoe belt are typically S-type (eg Two Sisters Granite), whereas those in the Shoobridge area display I-type characteristics. The adjoining sedimentary rocks are contact-metamorphosed to albite-epidote hornfels facies. Some significant deposits are described below.

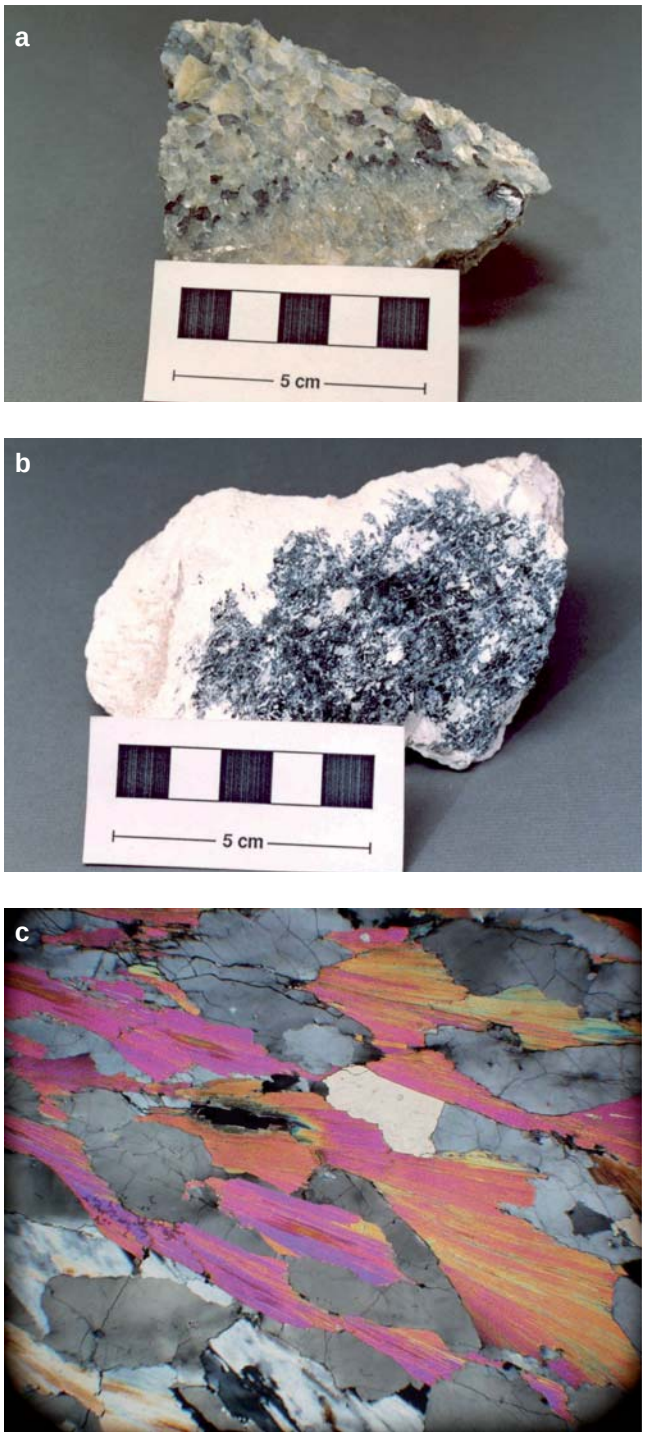


Figure 5.87. Mineral composition and textures of various pegmatite zones. (**a**) Cassiterite (black) quartz and muscovite in border zone (Mount Finniss mine). (**b**) Tantalite and kaolinised feldspar in intermediate zone (Yan Yam deposit). (**c**) Preferred orientation of books of muscovite and elongate prismatic grains of quartz, wall zone (Saffums 1 pegmatite, crossed-polarised light, width of view 5 mm).

Mount Finniss

Tin-tantalum mineralisation at Mount Finniss was discovered in 1886, but little work was done until 1906, when 1.5 t of tantalite concentrate was produced (Summers 1957). Mining resumed in 1925 and continued uninterrupted until 1951, with a total production of about 20 t of cassiterite and 12.26 t of tantalite concentrate. The mine was re-opened in 1980 and by 1989, approximately 150 000 t of pegmatite had been mined. In addition, 100 000 t of alluvial material was treated, producing 145.643 tonnes of Sn concentrate and 53.152 tonnes of Ta concentrate. Proved/probable reserves down to 30 m depth were estimated to be 160 000 t averaging 160 ppm Sn, 35 ppm Ta and 70 ppm Nb (Nicholson 1988). Parts of these reserves have been subsequently mined.

Pegmatite bodies occur as three irregular interconnected lenses (**Figure 5.88**) over a length of about 200 m. These lenses have a general north-northeast trend. The dip changes from sub-vertical in the south, to about 30° east in the center, to sub-horizontal in the north. Pinching and swelling is common and the maximum true thickness is about 30 m.

The enclosing sedimentary rocks are contact-metamorphosed greywacke and siltstone of the Burrell Creek Formation. Tourmaline laths up to 5 mm long are developed along the pegmatite–wallrock contact. Blue-white quartz veins, similar to the Au-quartz veins, are present in the wallrock and in places, these are terminated at the wallrock–pegmatite contact, suggesting their earlier formation in the paragenetic sequence.

Zonation within the pegmatite bodies consists of (from the margins inwards): an outer border, up to 5 cm thick, consisting of fine-grained quartz + muscovite; a wall zone (up to 30 mm thick) of coarse, quartz + mica; an intermediate zone of feldspar + mica + quartz; an inner core of massive quartz.

Mineralisation consists of coarse cassiterite and tantalite-columbite crystals mainly in the intermediate zone. Some very large crystals are also present in the border zone.

At the Mount Finniss mine, and also in all the Sn-Ta-bearing pegmatites in the Bynoe belt, kaolinisation of feldspar is ubiquitous (**Figure 5.89**) and has rendered the ore amenable to economic extraction and processing at grades far lower than those required to support a hard-rock operation. The ore is mined by using an excavator and without any need for drilling and blasting. As most mineralisation is disseminated throughout the intermediate kaolin-rich zone, a crushing and grinding circuit is also not required (Pietsch and Clayton 1990).

Goodwill Extended

The Goodwill Extended (also known as *Walkers Creek*) mine was probably discovered and mined at the same time as the Mount Finniss mine, but no record of production exists until the year 1955, when 0.32 t concentrate was produced. The mine was re-opened in the early 1980s and produced about 56 t of cassiterite-tantalite concentrate from open-cut mining. The open cut is about 150 m long, 50 m

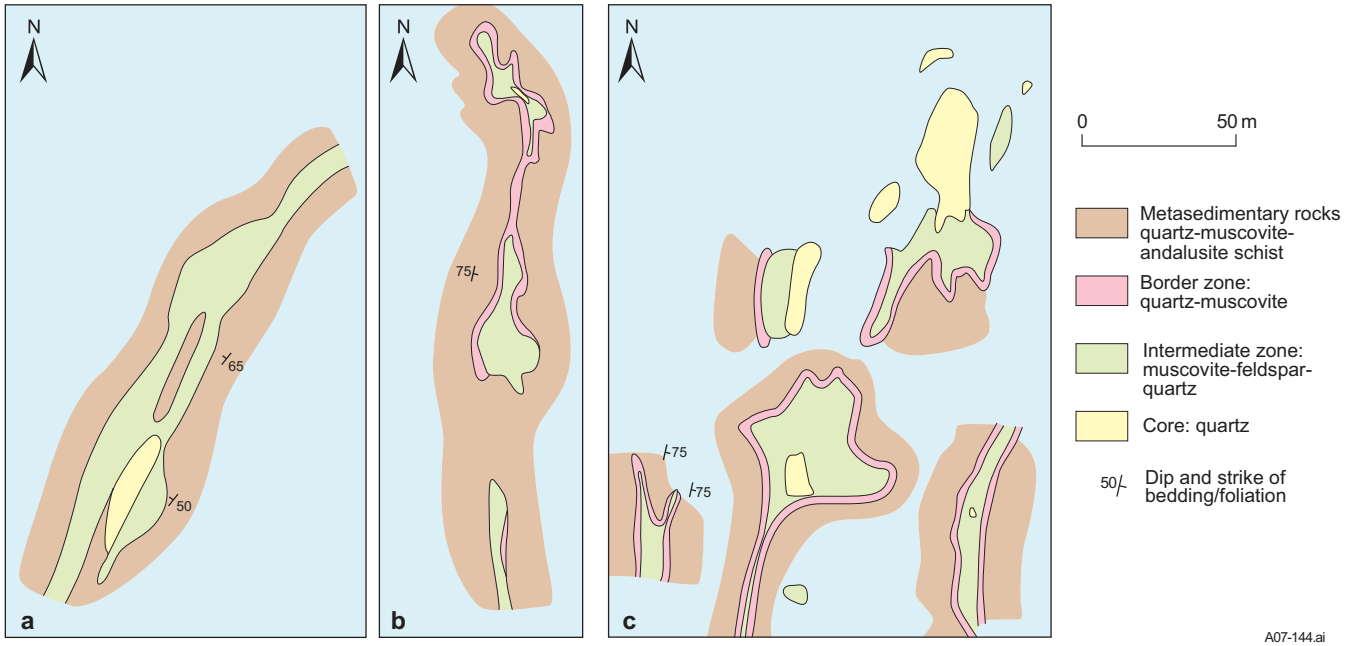


Figure 5.88. Sketch maps of (**a**) Labelle, (**b**) Goodwill Extended and (**c**) Mount Finniss Sn-Ta-bearing pegmatites (after Frater 2005).

wide and 30 m deep. The pegmatite that has been mined trends north and pinches out at both ends of the open cut. In the centre, the width of the pegmatite is about 15 m.

Several other pegmatite lenses are present outside the area of the open cut. All of these pegmatites trend north-northeast or north and have near-vertical dips. The overall length of the zone containing the pegmatite lenses is about 250 m.

The host rocks, siltstone and greywacke of the Burrell Creek Formation, have been contact-metamorphosed to quartz-muscovite-andalusite schist. Contacts with the pegmatites are sharp, and zoning, as described previously, is prominent. The quartz core is not as pronounced as at the Mount Finniss mine, although some old workings at the northern end do contain massive quartz.



Figure 5.89. Extensively kaolinised intermediate zone of Mount Finniss pegmatite.

Hang Gong

This deposit is located on the northern side of Mandorah Road, 2.3 km northwest from Observation Hill. Between 1903 and 1956, it produced about 190 t of tin concentrate (Crohn 1968). Between 1981 and 1982, a total of 11.06 t of concentrate, with an average grade of 57.9% SnO_2 and 9.3% Ta_2O_5, was produced (Frater 2005). It was further explored and worked in 1989 by Greenbushes Tin Ltd. Total production from this and a group of other deposits (*Lees Extended*, *Yan Yams*, *Booth Extended* and *Henderson West*) is given as 4.213 t of tin concentrate and 3.598 t of tantalite concentrate (Frater 2005), but separate production figures have not been provided for the individual deposits. In 1988, Greenbushes Tin Ltd estimated an ore reserve to 20 m at Hang Gong of 0.222 Mt at 83 g/t SnO_2 and 59 g/t Ta_2O_5 recovered plant grade. In 1997, Fieldcorp Pty Ltd conducted a trial mining operation, during which 50 144 t of ore was put through the plant. Indurated ore resulted in lowered recovery grades from the plant, which had no crushing circuit (Frater 2005).

The pegmatite is zoned and is predominantly composed of muscovite, kaolinite and quartz (intermediate zone). Some pods of blue-grey quartz are present and range up to 2 m in size. In partly kaolinised zones, relatively unaltered white feldspar (orthoclase) occurs. The host rock is andalusite-mica schist of the Burrell Creek Formation.

The main pegmatite is approximately 390 m long and was mined over a maximum width of 60 m. In the south, it is a horizontal sill that plunges at a shallow angle to the north. The pegmatite strikes 350° and continues under alluvial cover. The *Hang Gong* pegmatite is conformable with the strike of the enclosing rocks, but is discordant with the near-vertical dip. It has various muscovite- and kaolinite-rich zones, the former often associated with high tantalum grades (Frater 2005).

Leviathan

This pegmatite is located on the eastern bank of Leviathan Creek and was discovered in 1886. It has a total recorded production of 170 t of cassiterite concentrate, all of which was obtained before 1909. Workings originally consisted of two open cuts about 200 m apart, but these are now largely collapsed. During 1986–1987, Greenbushes Tin NL carried out extensive costeaning in the Leviathan Creek area and in the vicinity of the mine.

In 1993, Corporate Developments Pty Ltd estimated an indicated resource of 1000 m^3 (potential resource of 3000 m^3) at 1000 g/t SnO_2 and 400 g/t Ta_2O_5 for Leviathan (Bluck 1993), but it is unclear whether or not this included all of the pegmatites (Frater 2005). In 2001, Julia Corporation Ltd excavated four costeans and drilled 17 RC holes for a total of 822 m (Horn 2001). Costeaning confirmed the presence of three pegmatites striking approximately 030° and dipping to the southeast at *Leviathan North* (original mining site).

The pegmatite forms a concordant body within the Burrell Creek Formation and is zoned. It is possible that mineralisation was mainly concentrated close to, or in the border and wall zones, as these appear to have been selectively mined (Summers 1957). The pegmatite body is about 200 m long and its maximum width is about 20 m.

Saffums area

A number of occurrences are located about 20 km south of Observation Hill. Talmina Trading Pty Ltd (Talmina) conducted considerable mining and exploration activity in this area during 1981–1985. However, in 1985, Talmina was placed in receivership and in 1986, Corporate Development Pty Ltd purchased the plant and tenements from the receivers. In 1988, exploration and mining rights were assigned by Corporate Development to Brevcorp Pty Ltd (Outback Minerals 2008). The most productive mines were *Saffums 1*, *Saffums 2*, *Bilatos* and *Sandras*. Between 1983–1985, an alluvial treatment plant, located at Saffums 1, processed 22 600 t of eluvium from this area for an average recovery of 0.47 kg/t Ta_2O_5 concentrate and 0.34 kg/t SnO_2 concentrate. Between 1988 and 1990, Brevcorp Pty Ltd expanded the plant and apparently treated some 20 000 to 30 000 m^3 of eluvium, producing 5 t of tantalite and 10 t of cassiterite concentrate (Robinson 1993). Corporate Developments Pty Ltd calculated a resource (to 20 m) of 10 000 m^3 of 0.08 kg/m^3 Ta_2O_5 and 0.30 kg/m^3 SnO_2 at Saffums 1 (Robinson 1993).

The pegmatite bodies are emplaced in contact-metamorphosed Burrell Creek Formation, which comprises sericitic and graphitic schist. The graphitic schist is unique to this area and does not occur in the vicinity of pegmatites from the Observation Hill or Leviathan Creek areas.

Detail mineralogical studies (Friedrich and Jutz 1984, Friedrich and Wiechowoski 1984) have shown that the pegmatite contains quartz, muscovite, feldspar, kaolinite, chlorite, tourmaline, barite, amblygonite, zircon, tantalite-columbite, cassiterite, rutile, ilmenite, magnetite, haematite and limonite. Microprobe studies on Saffums 1 tantalite-columbite concentrate samples indicate that Ta_2O_5 varies between 35% and 80.5%, and Nb_2O_5 from 45.4% to 4.8%. The tantalite contains variable FeO from 1% to 13.1% and MnO from 1.2% to 12.5%. The SnO_2 content ranges from 0.1% in Nb-rich members to 11.5% in Ta-rich samples. The composition of the cassiterite at Saffums 1 is in the range 95.4–100% SnO_2, with up to 4.2% Ta_2O_5, 0.6% Nb_2O_5, 0.6% FeO and 0.14% MnO. Small grains of tantalite have been identified as inclusions in cassiterite grains.

The *Saffums 2* pegmatite is located 700 m north of Saffums 1 and 300 m west of Bilatos. Production from Saffums 2 is known to have contributed to Talmina Trading's output of 1364 kg of tantalite and 937 kg of Sn concentrate, recorded in 1995 from its combined Finniss River operations. A wall-rock sample from Saffums 2 has assayed 11.4 g/t gold (Wallen-Teluk 1988).

The *Sandras* pegmatite is located approximately 3 km east-southeast from Saffums 1 and has contributed to Talmina Trading's total output, as is the case with Saffums 2. The pegmatite body was strongly kaolinised, with tantalite identified in quartz-mica zones and in kaolinite on the margins of these zones (Friedrich and Jutz 1984). A resource to 10 m depth of 5000 m^3 at 0.03 kg/m^3 Ta_2O_5 and 0.40 kg/m^3 SnO_2 has been calculated from a 1991 sampling program by Corporate Developments Pty Ltd (Robinson 1993). The country rock on the eastern contact is carbonaceous schist containing numerous quartz veins. The country-rock to the west is more complex, with outcrop consisting of alternating andalusite-mica schist and greisenised pegmatite.

Labelle

J Walton discovered the *Labelle* pegmatite in 1984. This lensoidal pegmatite is 5–30 m thick and over 100 m long. At the surface, it dips steeply to the west. It is kaolinised and has a massive quartz core. In 1987, a 3000 t bulk sample assayed 270 ppm Ta, 80 ppm Sn and up to 1% Be. Ta and Sn were reported to be elevated at the outer contacts and Be towards the centre. The country rock consists of carbonaceous slate and meta-sandstone and some low-grade Sn values were recorded in these sedimentary rocks adjacent to the pegmatite. A preliminary resource to 50 m depth of 250 000 t averaging 120 ppm Ta and 75 ppm Sn has been defined (Frater 2005).

Annie

The *Annie* pegmatite is situated about 200 m north of the Annie River, approximately 8 km south of the Charlotte River crossing on the Fog Bay road. The pegmatite strikes northeast and extends for over 400 m with a width of between 5 and 27 m. The most prominent feature is two quartz 'blows' (core zone), one in the north and the second in the central portion of the prospect. Cassiterite is a main ore mineral and forms black thin plates or tabular crystals to 2 cm. Eleven tonnes of tantalite concentrate and 28 t of tin were produced between 1995 and 1997, and a further 69 t of combined Ta-Sn concentrate was produced in 1997–1999. Auger drilling has indicated a resource in the order of 98 000 t averaging 156 g/t SnO_2 (Frater 2005).

Mount Shoobridge field

This field is located about 10 km west of the Cosmo Howley gold mine and is the site of the first discovery of tin in the Northern Territory by Mr George Barrett in 1882. The Mount Shoobridge field includes the *Barretts, Mount Shoobridg*e, *Chinese, Halls Creek* and *Plateau Point* occurrences (**Table 5.10**). Recorded past production is from *Barrett* (117 t concentrate) and *Mount Shoobridge* (145 t concentrate).

The pegmatite bodies intrude quartz-mica schist, siltstone and fine-grained greywacke of the Burrell Creek and Mount Bonnie formations. The Shoobridge Granite is considered to be the parent granite to the pegmatites (Frater

2005). This is a small concentrically-zoned granitoid, consisting of a central leucogranite, passing outward through biotite-hornblende granite to a peripheral quartz-monzodiorite. The pegmatite forms discordant dykes and veins up to 10 m wide and 100 m long, concentrated along the axes of the Mount Shoobridge and adjacent anticlines.

Other Sn-Ta pegmatite occurrences

The *Muldiva, Buldiva* and *Collia* prospects occur in inaccessible hill country on the upper tributaries of Soldiers Creek (**Table 5.11**). The Soldiers Creek Granite outcrops 400 m to the west of Buldiva. The recorded production from these prospects, between 1923 and 1967, is 45.7 t of tin concentrate (Frater 2005). Some eluvial tin was also produced from gravel beds at the base of Cretaceous strata overlying Burrell Creek Formation schist and quartzite, on the north side of the main branch of Buldiva Creek.

Fluid inclusions

Ahmad (1995) studied fluid inclusions in quartz from Sn-Ta-bearing pegmatites, mostly from the Bynoe belt. This study showed that early Type A ($CO_2 + H_2O \pm CH_4$) inclusions were trapped at the H_2O-CO_2 solvus at a pressure of ca 1 kbar, a temperature of ca 300°C (range 240–328°C) and a salinity ca 6 wt% eq NaCl. Pressure-salinity-corrected temperatures on Type B (H_2O + ca 20% vapour), C (H_2O + <15% vapour) and D (H_2O + halite + vapour) inclusions also fall within the range of Type A inclusions. Wygralak and Mernagh (2006) reported a similar fluid inclusion population in quartz from the Muldiva and Buldiva prospects.

Stable isotopes

Oxygen and hydrogen isotope data (Ahmad 1995) show that kaolinite was formed in isotopic equilibrium with meteoric waters. Fluid inclusion waters from core-zone quartz show an enrichment in deuterium, suggesting a metamorphic influence. Isotope values on muscovite are consistent with a magmatic origin. It is suggested that the pegmatites were derived from the post-orogenic phase of the Two Sisters Granite. Precipitation of cassiterite took place at about 300°C from an aqueous fluid, largely as a result of an increase in pH, due to feldspar alteration (Ahmad 1995).

Geochronology

Frater (2005) provided geochronological data on samples of tantalite, cassiterite and muscovite from the Bynoe pegmatites, which were analysed by PD Kinny at Curtin University, Western Australia. The tantalite and cassiterite samples were U-Pb dated by SHRIMP ion microprobe (SIMS) analysis, whereas the muscovite was Ar-Ar dated by the Western Australian Argon Isotope Facility.

The most precise ages are given by tantalite from the Angers pegmatite at 1723.5 ± 5.5 Ma and from the Saffums pegmatite at 1733 ± 13 Ma, indicating that mineralisation probably occurred between 1740 Ma and 1720 Ma. The $^{40}Ar/^{39}Ar$ muscovite ages have a much wider range, 1735–1680 Ma, which suggests that the cooling rates were very slow, or that fluid activity was prolonged until well after mineralisation. Wygralak and Mernagh (2006) reported on the $^{40}Ar/^{39}Ar$ ages of muscovite associated with tin- and gold-bearing veins from the southern Litchfield Domain. They showed that veins that are spatially related to the Allia Creek Granite are relatively old and in the range 1742–1726 Ma, whereas veins associated with the Soldiers Creek Granite are relatively young and in the range 1707–1703 Ma. These ages are significantly younger than, and incompatible with the 1806 Ma age (U-Pb SHRIMP) of the Allia Creek Granite and 1862 Ma age (U-Pb SHRIMP) of the Two Sisters Granite (Worden *et al* 2008b), generally recognised as the parents to the Bynoe field pegmatites.

Greisens

Sn is present in greisens associated with the Wolfram Hill and Wandie granites. The main ore minerals are cassiterite with minor wolframite. Malachite has been noted in some places. The gangue minerals are quartz, muscovite, K-feldspar and epidote. Production from this group is minor.

Alluvial deposits

Most vein-type tin deposits and tin tantalum-bearing pegmatites are associated with eluvial/alluvial concentrations. Most of these occurrences are small and insignificant. Major workings are in the *Mary River Camp*, *Wandie Creek*, *Emerald Creek*, *Morris* and *Umbrawarra* areas.

Tungsten and molybdenum

Tungsten deposits in the PCO are located in the Wolfram Hill, Yenberrie and Yeuralba fields (**Table 5.10**, **Figure 5.84**) and have produced about 950 t of wolframite concentrate, of which 740 t has been produced from the Wolfram Hill mine. These deposits occur as wolframite ± Sn ± Mo-quartz veins or greisens, and are associated with the Wolfram Hill, Yenberrie or Yeuralba granites of the Cullen Supersuite.

The *Wolfram Hill* deposit was discovered in 1900 (Jones 1987) and was worked intermittently between 1900 and 1953, for a recorded production of 740 t wolfram concentrate, 258 t of copper concentrate and 65 t of cassiterite concentrate (Bagas 1983). This deposit has been described in a number of previous studies (Jensen 1919, Kleeman 1938, Crohn 1968, Bagas 1983). Two prominent lines of reefs, separated by 300 m of barren hornfels, have been mined via two shafts, an adit and several small open-cuts. Wolframite, cassiterite and chalcopyrite are the main ore minerals, with small amounts of pyrite and arsenopyrite. In some veins, coarse muscovite, biotite and quartz are present near the vein edge, whereas the central part is almost barren massive quartz. The 'main reef' extends for about 300 m along strike and is mainly concordant with the enclosing rocks. High-grade ore shoots lie in zones of discordance.

The *Yeuralba mineral field* is located 55 km east-northeast of Katherine and was discovered in 1924. It was worked for W and Sn intermittently until 1954, for a total recorded production of 19.6 t W, 8.2 t Sn and 110 g Au (Randal 1963). Walpole and Drew (1953) listed 17 separate mineralised occurrences. Most of these are located along the western greisenised contact between the Yeuralba

Granite and hornfelsed greywacke and siltstone of the Tollis Formation. The greisen consists of saccharoidal quartz, muscovite, wolframite, cassiterite, scheelite, fluorite and varying amounts of tourmaline. Wolframite occurs as subhedral or anhedral coarse grains, as well as bismuth, copper minerals and some gold.

The *Yenberrie mineral field* (W, Mo, and Bi) was discovered in 1912 and was exploited until 1930. It was reopened later and briefly worked between 1951 and 1953. Crohn (1968) gave a total production of 161 t of wolframite, 127 kg of molybdenite and a small quantity of bismuth. Randal (1963) gave a much larger production of wolframite (332 t). The production was derived from nine lodes, consisting of greisenised quartz-aplite dykes, hosted by the Yenberrie Granite. Most mineralisation is confined to veins of coarse white quartz, containing disseminated wolframite grains (up to 5 cm in size) and minor molybdenite. Some wolframite also occurs in greisenised aplite and in irregular mica patches in granite that are not associated with any fissures or shears.

Iron ore

Significant ironstone deposits in the PCO occur in the Mount Bundey and Frances Creek areas (**Table 5.12**, **Figure 5.90**). Small surficial concentrations are also known from elsewhere. Ferenczi (2001) provided a comprehensive account of these deposits.

Mount Bundey

The Mount Bundey ironstone occurrence was identified by BMR in 1958. Government diamond drilling in 1962 and 1964 established a resource of 1.5 Mt averaging 64% Fe, 0.05% P and 0.07% S (Dunn 1964). The deposit was mined between 1968 and 1971 producing 843 063 t of ore averaging 63.4% Fe, 0.057% P and 0.11% S (Ryan 1975). Mining operations ceased when the level of sulfur in the ore exceeded market specifications. The concordant, tabular ironstone bodies are hosted within a roof pendent of near-vertical metasedimentary rocks of the Wildman Siltstone within the Mount Goyder Syenite. The ironstone bodies consist of massive medium to very coarse (0.2–10 mm) martite, with varying amounts of haematite, goethite and quartz near the surface. Below the zone of oxidation (25 m), a martite-magnetite-pyrite association, which contains minor chalcopyrite (<1%), is present. Pyrite averages 2–5%, but may constitute of up to 20% of the ore in places. Vugs are common in upper parts of the martite ore and probably represent voids after leached-out sulfides. Ryan (1975) suggested that the iron ore (magnetite-pyrite) lodes were essentially formed by the remobilisation of sedimentary iron within the metasedimentary rocks during intrusion of the Mount Goyder Syenite. The magnetite-pyrite ore was subsequently 'upgraded' during a period of supergene enrichment to produce the massive vuggy martite ore.

Common name	Lat	Long	Shape	Rock unit	Resource (Mt)	Resource (%Fe)	Production (Mt)	Production (%Fe)	Reference
Beestons	-12.89	130.96	Irregular	Whites Fm	*1.5*	*59*			Pritchard (1969)
Daly River Road	-13.47	131.12	Tabular	Petrel Fm	*0.9*	*50*			Sullivan (1970)
Frances Creek East	-13.57	131.97	Tabular	Masson Fm	*1.5*	*>50*			Ferenczi (2001)
Helene 1	-13.58	131.85	Tabular	Wildman Sltst	*0.065*	*62.9*			Ferenczi (2001
Helene 11	-13.59	131.85	Tabular	Wildman Sltst	0.25	57.1			Territory Resources (2010b)
Helene 2	-13.59	131.85	Tabular	Wildman Sltst	*0.03*	*62.3*			Ferenczi (2001)
Helene 3	-13.59	131.85	Tabular	Wildman Sltst	0.315	59.4	0.3	64	Territory Resources (2010b)
Helene 4	-13.59	131.85	Tabular	Wildman Sltst	*0.025*	*60.2*	0.12	60	Ferenczi (2001)
Helene 5/6/7	-13.61	131.85	Tabular	Wildman Sltst	5.21	58.8	6.1	64	Territory Resources (2010b)
Helene 9	-13.61	131.85	Tabular	Wildman Sltst	0.189	59.6	0.15	60	Territory Resources (2010b)
Jasmine Centre	-13.58	131.87	Tabular	Wildman Sltst	0.971	59.8	0.11	62	Territory Resources (2010b)
Jasmine East	-13.58	131.88	Tabular	Wildman Sltst	0.196	59.0			Territory Resources (2010b)
Jasmine West	-13.57	131.87	Tabular	Wildman Sltst	*0.065*	*63.3*	0.08	61	Ferenczi (2001)
Millers	-13.44	131.81	Tabular	Wildman Sltst	1.28	53.2			Territory Resources (2010b)
Mount Bundey	-12.85	131.59	Tabular	Whites Fm			0.84	63	Ferenczi (2001)
Ochre Hill	-13.53	131.86	Tabular	Wildman Sltst	0.114	58.1			Territory Resources (2010b)
Thelma-Rosemary	-13.58	131.87	Tabular	Wildman Sltst	0.808	57.8			Territory Resources (2010b)
Rum Jungle Iron Ore	-12.99	130.99	Irregular	Whites Fm	0.280	60.8			Territory Iron (2006a)
Saddle 1	-13.52	131.85	Tabular	Wildman Sltst	*0.217*	*57.9*			Ferenczi (2001)
Saddle East	-13.51	131.85	Tabular	Wildman Sltst	0.101	58.7			Territory Resources (2010b)
Saddle Extended	-13.49	131.83	Tabular	Wildman Sltst	*0.229*	*61.5*			Ferenczi (2001)
Thelma 1	-13.58	131.86	Tabular	Wildman Sltst	0.326	58.4	0.35	64	Territory Resources (2010b)
Yarram	-12.99	130.98	Sheet	Coomalie Dol	0.340	60.2			Territory Iron (2006b)

Table 5.12. Summary data on significant PCO iron ore deposits. Resources in italics are non-JORC compliant. Production given is historical to 1974. During current mining operation (2007–June 2010) at Frances Creek field, some 4.16 Mt of iron ore has been produced by Territory Resources, mostly from Helene 5/6/7 deposits, with minor component from Ochre Hill and Thelma-Rosemary deposits. Abbreviations: Lat = latitude; Long = longitude; Fm = Formation; Dol = Dolostone; Sltst = Siltstone; Prod = Production.

Frances Creek

The Frances Creek iron field was discovered in 1961. A pre-mining resource of 5.14 Mt averaging 64% Fe was estimated for the southern part of the field (Elliot 1965). Mining by Frances Creek Iron Mining Corp Pty Ltd between 1966–1974 produced some 7.98 Mt of ore averaging 59% Fe (Bagas 1981). There are some fifty named iron ore occurrences (**Figure 5.91**). Most of the iron ore was obtained from the *Helene 6/7* pit (**Figure 5.92**). Other pits included *Thelma 1* and *2*, *Rosemary*, *Jasmine West*, *Jasmine Central*, *Helene 10, 9, 8, 5, 4, 3* and *2*. This first phase of mining operations ceased in July 1974. The current operator, Territory Resources Ltd (formerly Territory Iron Ltd), has carried out a comprehensive RC/RAB drilling program since 2005. This company commenced mining in April 2007 with the first shipment to China in September 2007. Current (JORC-compliant) indicated and inferred resources (inclusive of reserves) for the field are given as 8.585 Mt at 59.2% Fe. The largest resources are at the Helene 5/6/7 deposit and have been stated to be 5.21 Mt at 58% Fe (Territory Resources 2010b). Production from this deposit from September 2007–June 2010 was given as 4.16 Mt of ore (Territory Resources 2010a).

Ore consists of massive, fine, micaceous to bladed haematite, containing varying amounts of shale fragments and quartz grains. The orebodies are hosted by the Wildman Siltstone. Crohn (1968) considered that the deposits formed as a result of supergene enrichment of pyritic shale breccia. A more recent review of mining data (geological plans and cross-sections), along with petrological studies and field inspections of the open pits (Pontifex 2000), suggested that the iron ore mineralisation is the result of hydrothermal remobilisation (Bowden 2000). In this model, iron oxides from ferruginous banded shale in the Wildman Siltstone were concentrated into favourable structural sites that developed prior to the intrusion of the Allamber Springs Granite.

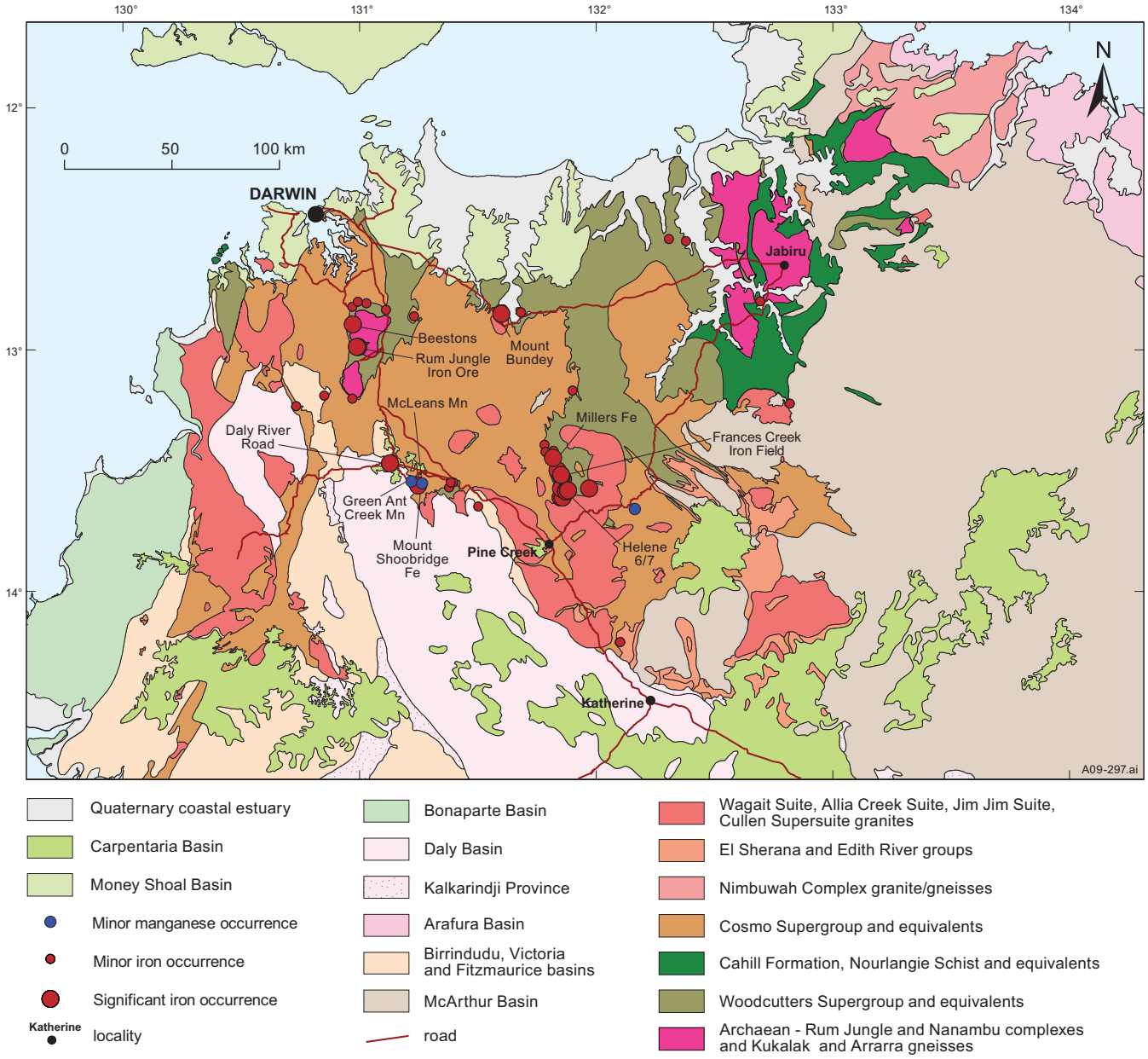


Figure 5.90. Location of iron ore and manganese deposits of the PCO. Data from NTGS MODAT database. Significant deposits are labelled.

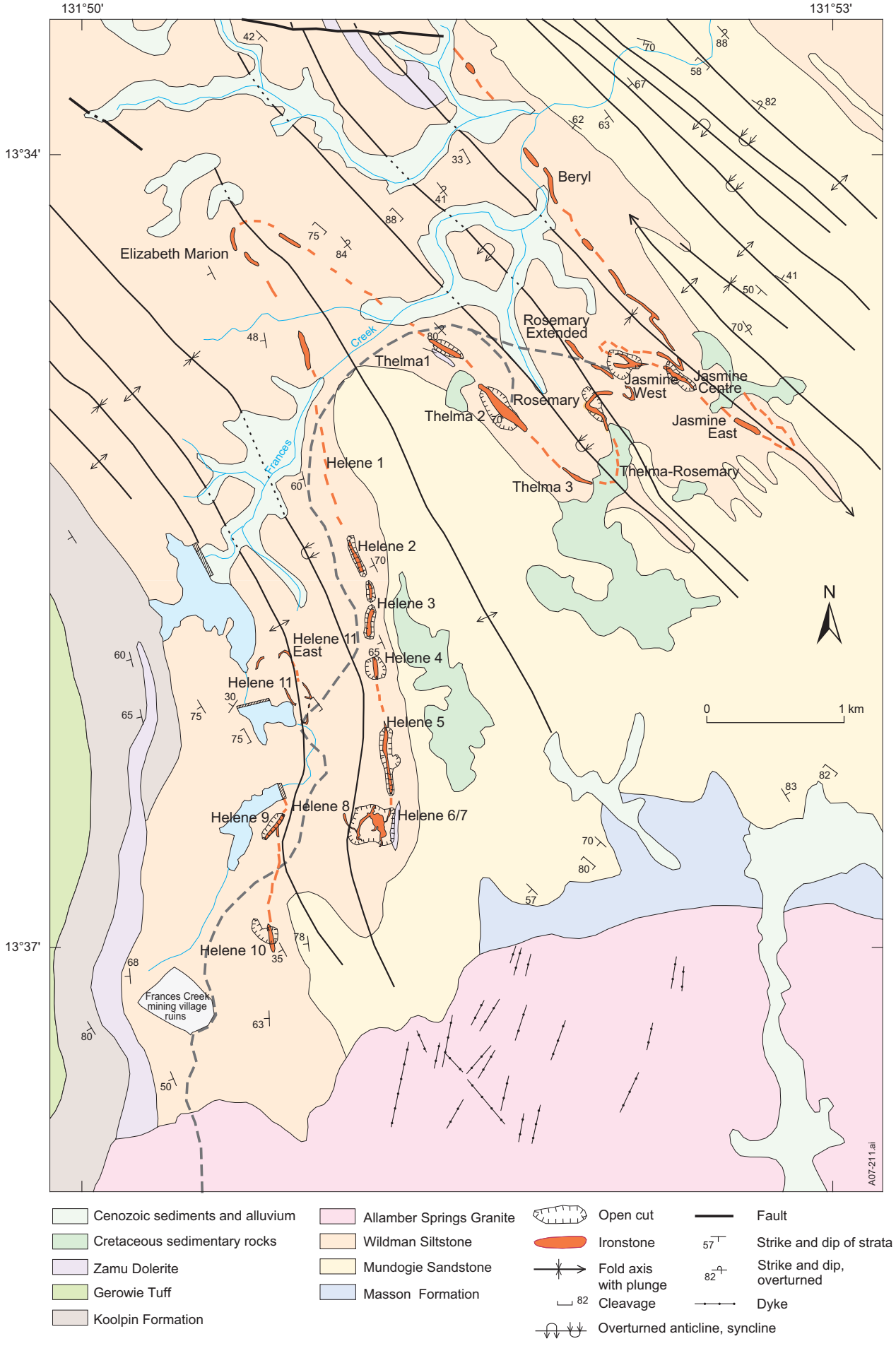


Figure 5.91. Geological map of southern part of Frances Creek iron ore field, with significant deposits labelled (modified from Stuart-Smith *et al* 1987).



Figure 5.92. Territory Resources Ltd mining operations in the Frances Creek iron field. Satellite photo courtesy of Andrew Jettner.

Other small ironstone occurrences in the PCO probably resulted from surficial enrichment and these include *Rum Jungle, Beestons, Howley* and *Daly River Road*.

Industrial minerals and rocks

The PCO contains a variety of industrial minerals, including significant deposits of magnesite and phosphate (**Table 5.13**). Minor occurrences of barite, corundum, fluorspar, kyanite, graphite and ochre are also known from the region. **Figure 5.92** provides the location of these deposits.

Magnesite

Magnesite was first noted in the Coomalie Dolostone at the Mount Fitch uranium deposit in the Rum Jungle Mineral Field (Prichard 1975). Between 1978 and 1982, BHP conducted extensive exploration for magnesite and identified the Celia and Coomalie magnesite deposits. In 1980, beneficiation tests indicated that a low-silica product could be obtained that might be suitable for refractory-grade magnesite. Between 1990 and 1993, Aztec Mining Company Ltd carried out drilling over the *Area 44, Celia* and *Huandot* deposits. The company identified a total resource of 5.08 Mt magnesite at Huandot and 0.5 Mt down to 40 m depth at the Celia deposit (Pima Mining 2001, Grove 1992). In 1998, Mount Grace Resources NL commenced exploration in the area surrounding the *Coomalie* magnesite prospect. This work included closely spaced drilling that outlined a JORC-compliant resource of 16.6 Mt averaging 43.23% MgO at the *Winchester* deposit. The resource was considered sufficient for over 50 years of production (Korab Resources Ltd 2010). Magnesite is known through much of the extent of the Coomalie and Celia dolostones, including the footwall of the *Browns* and *Mount Fitch* deposits.

The magnesite deposits in the Rum Jungle area are discussed in detail by Bone (1985) and Ahmad *et al* (2006). They represent stratabound occurrences within the Celia and Coomalie dolostones (**Figure 5.93**). Both dolomite and magnesite are present as interbedded irregular to lenticular bodies. At the Celia, Huandot, Area 44 and Winchester deposits, magnesite, talc and dolomite are the main minerals. Trace pyrite is noted in places. High-grade magnesite bodies are irregular in shape. Magnesite varies from dark brown to grey to pure white, and possibly reflects variations in trace iron oxide contents. Magnesite is present as coarse, up to 1 cm size, euhedral crystals in two dominant morphologies: rhombohedral and bladed (Bone 1983). Examination of core and open pit samples indicates that the rhombohedral magnesite is apparently paragenetically later and is usually present as veins or vug fills. Dolomite is present in distinct bands or irregular bodies. It is generally finer grained than and may also be intergrown with magnesite. Talc is present as fine-grained irregular masses within magnesite or as a predominantly talc rock. Other minerals include quartz, chlorite, iron oxides (haematite), apatite and minor tourmaline.

Stable isotopes

Aharon (1988) and Ahmad *et al* (2006) discussed carbon and oxygen isotope data. Magnesite from the Coomalie and Celia deposits is depleted in ^{18}O relative to the associated dolomite. This suggests isotopic disequilibrium or precipitation by two different fluids. The ^{18}O values of recent sabkha magnesite are distinctly heavier than those of the Rum Jungle magnesite. Carbon isotope values show that magnesite is enriched in ^{13}C in comparison to dolomite.

Fluid inclusions

Bone (1983, 1985) examined the fluid inclusions in magnesite from the Celia and Coomalie deposits. On the basis of the phases present, five types of inclusions were identified: (1) liquid only; (2) liquid + vapour (most common); (3) liquid + vapour+ daughter mineral(s); (4) two immiscible liquids; (5) immiscible liquids + vapour. The latter two types are seen in many of the deposits of the PCO (Ahmad *et al* 1993). Solid phases in type 3 inclusions were haematite, dawsonite, halite and unidentified minerals, which did not dissolve upon heating. Solid dissolution temperatures were 50–100°C higher than vapour phase homogenisation temperatures, indicating that the inclusions trapped a saturated fluid. Bone (1983, 1985) concluded that temperature was a major control on the two distinctive morphological forms. The

Common name	Lat	Long	Size	Shape	Commodity	Model	Overall style	Resource	Grade	Reference
Tomahawk	-13.45	131.42	Occurrence	Tabular	Fluorite	Vein	Multi-vein			Chaku (1974)
Area 44	-12.98	131.09	Small	Unknown	Magnesite	Metas	Massive			Butler (1994)
Celia	-13.03	131.06	Medium	Tabular	Magnesite	Metas	Massive	0.5 Mt		Grove (1992)
Coomalie (Winchester)	-13.05	131.06	Medium	Tabular	Magnesite	Metas	Massive	16.6 Mt	43.23% MgO	Korab Resources (2010)
Huandot	-13.00	131.10	Medium	Tabular	Magnesite	Metas	Massive	5.08 Mt		Grove (1992)
Area 4	-13.05	131.01	Medium	Tabular	Phosphate	Sed	Dissem	0.1 Mt	10% P_2O_5	Pritchard *et al* (1966)
Easticks	-13.03	131.00	Medium	Tabular	Phosphate	Sed	Dissem	0.04 Mt	12.2% P_2O_5	Pritchard *et al* (1966)
Geolsec	-13.04	131.00	Medium	Tabular	Phosphate	Sed	Dissem	1.01 M t	12.5% P_2O_5	Pritchard *et al* (1966)
Zeta	-13.03	130.98	Small	Irregular	Phosphate	Sed	Dissem	0.06 Mt	7% P_2O_5	Pritchard *et al* (1966)
Crater	-13.03	131.09	Occurrence	Irregular	Thorium	Palaeop	Dissem			Rhodes (1960)

Table 5.13. Summary data on significant PCO industrial mineral deposits. All resources are non-JORC compliant, except for Coomalie magnesite deposit. Abbreviations: Lat = latitude; Long = longitude; Dissem = Disseminated; Metas = metasomatic; Sed = sedimentary; Palaeop = Palaeoplacer.

rhombohedral form crystallised at temperatures of less than 150°C, whereas the discoidal form crystallised at higher temperatures. Homogenisation temperatures ranged from 100–225°C and freezing point depression temperatures from 0 to less than -40°C. Homogenisation of CO_2 liquid and CO_2 vapour occurred within a narrow range of 28–29°C, indicating a density of about 0.65 g/cm^3. Using this density value and an average homogenisation temperature of 150°C, the intersecting CO_2-H_2O isochors method provides an approximate pressure of about 600 bar, equal to a depth of 2.2 km (lithostatic), or about 6 km (hydrostatic). This is similar to the values obtained for a number of vein-type deposits in the PCO (Ahmad *et al* 1993).

Genesis

Bone (1985) considered that the precursor to the crystalline magnesite was sedimentary magnesite precipitated in a sabkha environment. She quoted stromatolites, tepee structures, low Sr and the interbedded nature of the magnesite as strong evidence of a sedimentary origin. Magnesite is known to occur in sabkhas; however, it generally forms only a small percentage of sabkha sediment, which typically comprises gypsum, aragonite and dolomite. The latter predominantly results from dolomitisation of aragonite (Pohl 1989). In addition, oxygen isotope data on magnesite (see above) does not support a sabkha-type environment and $\delta^{18}O$ values are distinctly lighter that those deposited in sabkha-type setting.

Morteani *et al* (1983) considered that magnesite formed as a result of metasomatic replacement of a carbonate precursor, due to an increase in temperature or Mg ion supply. If temperature is kept constant, magnesite formation from a dolomite precursor requires equilibration with fluids having low Ca/Ca+Mg ratios. An alternative is

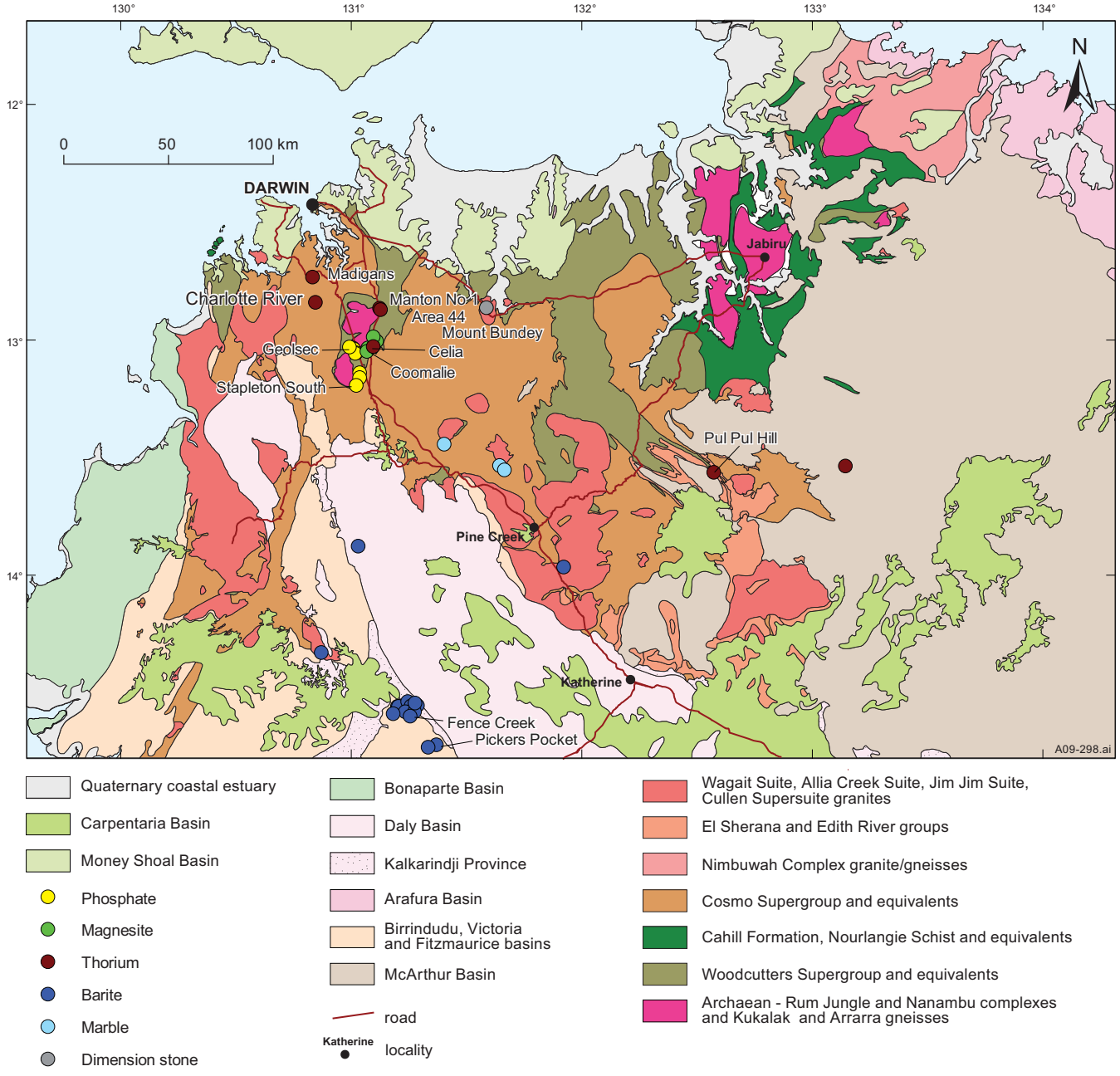


Figure 5.92. Industrial mineral occurrences in the PCO. Data from NTGS MODAT database. Significant deposits are labelled.

to increase the temperature to about 250°C and maintain a high Ca/Ca+Mg ratio.

The coarsely crystalline nature, as well as the presence of several generations of magnesite and ubiquitous brecciation, suggests post-diagenetic recrystallisation. This recrystallisation and metasomatic replacement of sedimentary dolomite by magnesite is likely to have occurred at temperatures between 150 and 300°C in the presence of fluids with a high Mg concentration. The latter may have been released during extensive sericitisation of the PCO rocks (Ahmad *et al* 2006).

Phosphate rock

Eight phosphate rock occurrences are known from the Rum Jungle area and are within the Geolsec Formation. Similar occurrences are also recorded from the South Alligator Valley within the Scinto Breccia (Prichard *et al* 1963, Crohn 1968, Freidman and Grotzinger 1991) and from the East Alligator River district, as apatite-bearing siltstone at the base of the Kombolgie Subgroup (Afmeco Mining 1998).

Total reserves for all the phosphate occurrences in the Rum Jungle area were stated by Crohn (1968) to be 4 Mt at 10% P_2O_5 with 1 Mt located at the *Geolsec* deposit. This deposit is currently being explored by Korab Resources Ltd. This company reported drilling that has yielded encouraging intercepts. Hole DDH6, drilled through the northern section of the Geolsec deposit, intercepted 6.9 m at an average grade of 32.7% P_2O_5 from 7.1 m to 14 m. Hole DDH1, drilled 80 m to the north of the Geolsec deposit, intercepted 12.5 m at an average grade of 20.36% P_2O_5 from 7.6 m to 20.1 m. Korab Resources, through its wholly owned subsidiary, Geolsec Phosphate Operations Ltd, is looking to commence production in 2011.

The phosphate occurrences of the Rum Jungle area have been described in Ruxton (1961), Walpole (1961), Pritchard *et al* (1963, 1966), Pritchard (1964), Ahmad *et al* (1993) and Ahmad *et al* (2006). Most of the deposits are located along the edges of a ridge-forming outlier of haematite-quartz-breccia, sandstone, fine-grained apatite rock and apatite breccia of the Geolsec Formation, situated about 2 km west of Batchelor (Castlemaine Hill). The stratigraphic age of the Geolsec Formation is unconfirmed. Previous workers have correlated this unit with the late Palaeoproterozoic Depot Creek Sandstone, but it may be older.

Phosphate mineralisation is within fine-grained lilac or red apatite-rich mudstone and its brecciated facies. The haematite quartz breccia and ferruginous sandstone are virtually devoid of any phosphate mineralisation. Phosphate-rich mudstone and breccia is restricted and overlies a ridge of Coomalie Dolostone. The rock commonly contains vuggy and open cavities, giving it the appearance of weathered vesicular volcanic rock. Apatite is the major constituent of the phosphate rock. Two varieties of apatite are seen: cloudy turbid apatite crystals; and relatively clear later overgrowths and veins. The clear variety commonly lines vugs and cavities. Haematite is present in growth zones within apatite and both minerals apparently crystallised together. Other minor phosphate minerals are millisite ($Na_2O.2CaO.6Al_2O_3.4P_2O_5.15H_2O$), variscite-sterngite ($AlPO_4.2H_2O$-$FePO_4.2H_2O$ and wavellite ($Al_3(OH)_3(PO_4)_2.5H_2O$). These minerals probably formed due to weathering. Other minor minerals present include kaolinite, sericite, illite, haematite goethite, tourmaline, epidote, anatase and zircon.

Howard and Hough (1979) provided major and selected trace element data on the phosphorite deposits in the Georgina Basin. The major element characteristics are essentially similar to those of the Geolsec deposit. Compared with Geolsec, the Wonarah deposit of the Georgina Basin is significantly enriched in base metals (Cu, Pb, Zn and Ni), which may have been sourced either from basinal fluids

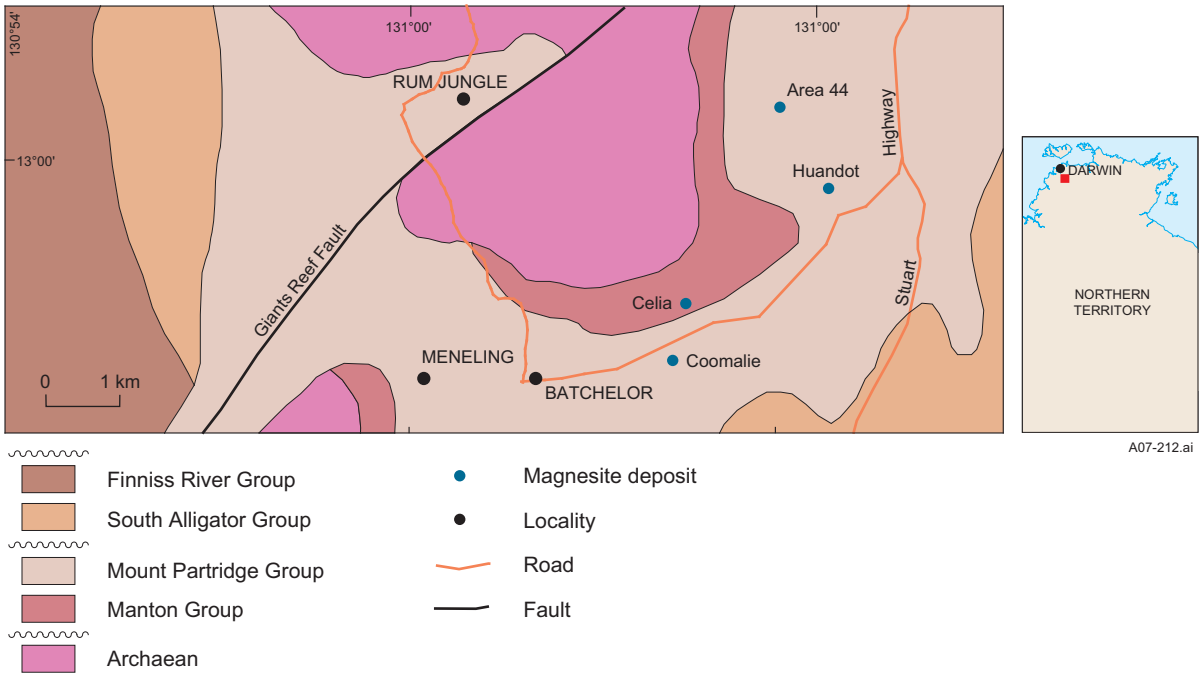


Figure 5.93. Location and geological setting of magnesite occurrences of Rum Jungle area. Data from NTGS MODAT database. Significant deposits are labelled.

or from the underlying Antrim Plateau Volcanics of the Kalkarindji Province.

Genesis

Various hypotheses have been put forward for the origin of HQB and phosphate rock in the Rum Jungle area. These include hydrothermal replacement (Paterson *et al* 1984), residual phosphate concentration as a result of weathering of Coomalie Dolostone (Pritchard *et al* 1966, Crohn 1968) and syn-sedimentary phosphate accumulations (Pritchard *et al* 1963, McClennan and Saavedra 1986, Ahmad *et al* 2006).

The hydrothermal model involves the formation of fluorapatite from metasomatic fluids initiated by movements along the Giants Reef Fault. These fluids were concentrated into zones of high permeability.

The secondary enrichment model involves the weathering of phosphatic dolomite and the concentration of phosphates by diagenetic processes. However, chemical analyses of the underlying Coomalie Dolostone indicate that it contains on the average about 0.025% P_2O_5 (range 0.01–0.09%) compared with the crustal abundance of 0.1 ppm (Ahmad *et al* 2006). The residual enrichment process is therefore an unlikely mechanism for the creation of phosphate deposits.

Compared to average phosphorite (Altschuler 1980), phosphatic rocks from the Geolsec deposit are depleted in Co, Cu, Ni, Pb, Se, Sr, La, Y and Zr (Ahmad *et al* 2006). The total REE concentration in the phosphatic rocks within the Geolsec deposit ranges from 18–370 ppm and averages 53 ppm; the lower value is for weakly phosphatic sandstone. PAAS-normalised REE patterns for the Geolsec phosphate do not show a Ce anomaly, indicating a non-marine environment of deposition. There is depletion of HREE and LREE, giving hat-shaped concave-down patterns. These patterns are seen in a few other phosphate deposits and are considered to be due to deposition/diagenesis in an estuarine environment or to scavenging of REE by ferric oxyhydroxides (Jarvis *et al* 1994). Such scavenging is considered to be a characteristic of pre-Cretaceous seawater (Grandjean-Lecuyer *et al* (1993). A fault-controlled estuarine depositional environment is most likely for the Geolsec Formation. The absence of a Ce anomaly further indicates a fresh water influence (Ahmad *et al* 2006).

The sedimentary model accounts for many of the characteristics of the phosphate rocks. These include a sedimentary succession comprising coarse matrix supported breccia, sandstone and fine-grained phosphatic rocks. Low trace element and REE concentrations and the major element geochemistry would favour deposition by sedimentary processes under estuarine conditions.

Other commodities

Barite veins occur in the McCarthys Granite, which is approximately 19 km southeast of Pine Creek. A resource for the largest vein has been estimated to be 900 t barite (Willis and Newton 1975). A *corundum* occurrence has been reported from about 10 km northeast of Mount Holder in GREENWOOD. Corundum is present as boulders up to 1 m in diameter and is presumed to have been derived from a subjacent body within the Hermit Creek Volcanics. A grab sample from this locality assayed 81.5% Al_2O_3, 9.5% SiO_2 and 0.5% Fe_2O_3 (Dundas *et al* 1987). *Fluorite* veins occur within the Koolpin Formation at the *Olive* prospect, about 3 km northeast of Ban Ban Station (Willis 1972). Minor *graphite* occurrences are known from the Sweets Member of the Welltree Metamorphics (Pietsch 1986b) and from the Koolpin Formation in the Golden Dyke Dome area. About 2 t of graphite was extracted from the latter occurrence (Crohn 1968). The *lithium*-bearing fluorophosphate mineral, amblygonite, has been reported from the *Pickets* Sn-Ta pegmatite in the Bynoe area and an analysed sample has returned 47% P_2O_5, 35% Al_2O_3 and 7.9% Li (Crohn 1968). The presence of amblygonite in this Bynoe pegmatite is significant in view of the current emerging demand for lithium, and further investigations may be warranted.

Extractives

Extractive materials in the PCO include fine and coarse sand, natural gravel, crushed rock, porcellanite, clay, soil and dimension stone. The following extracts are based on the Doyle (2001).

Coarse sand

Coarse sand lenses occur within the Howard Sand Member of the Cretaceous Darwin Formation. Drilling in the Howard Springs and Koolpinyah areas has outlined large buried deposits of clay-matrix quartz sandstone, which has been weathered to produce subrounded, mature medium to coarse sand. The *Boral* sand pit near Howard Springs has been dredged to 20–25 m below the surface and a total recoverable inferred resource at this site is estimated at 0.84 Mt. At the *AN47* sand pit, located on the eastern side of Howard River, the total inferred recoverable resource is estimated at 2.22 Mt. The *Mary River* sand deposit is located on the eastern side of Mary River on Annaburroo Station. In September 1999, the Mary River stockpile was noted to contain approximately 2000 m^3 of clean sand with minimal silt and clay. Coarse sand is also mined from an area near *Douglas Hot Springs* and about 6500 m^3 per year of this material is trucked 200 km to a depot in Coolalinga. The material is washed to remove fines and the product is then used for concrete aggregate. Doyle (2001) identified seven other prospects as potential sand deposits in the Howard River, Shoal Bay and Gunn Point areas.

Fine sand

Fine sand resources occur in current drainage areas and low-lying swamps. These sands have been alluvially derived from a number of sources including the Acacia Gap Quartzite Member, sandy interbeds within the Wildman Siltstone, sandy clay and claystone within the Bathurst Island Formation and sandy facies at the base of the Burrell Creek Formation. Fine sand generally contains low clay and silt (<10%). The sand layer averages 1 m in thickness and can lie either on top of Cenozoic laterite or in channels where the laterite has been etched away. The principle sources of fine sand are Palaeoproterozoic

rocks, specifically the Acacia Gap Quartzite Member and Burrell Creek Formation in the Darwin area. The total *in situ* resource estimate, based on an average sand depth of 0.5 m, is 11 million m^3 (Doyle 2001).

Gravel

Two varieties of gravel are mined from the Darwin area. Natural gravel is mostly sourced from lateritic duricrust by shallow pits for use as roadbase, fill material and landscaping gravel. Crushed rock gravel is produced from hard rock quarries, where material is extracted by drill and blast methods before crushing and screening into various sized aggregates and other products. A total of six operating hard-rock quarries are supplying material (quartzite, hornfels, syenite) to local markets. Three quarries, located in the *Yarrawonga* area near Palmerston, are in quartzite. Two quarries, located at *Mount Bundey*, are in hornfels and syenite. *Mauries Quarry*, located at Acacia to the south of Darwin, also quarries quartzite. The total estimated resources of these deposits is over 20 Mt (Doyle 2001). There are also a number of abandoned and disused quarries, including the *Batchelor* and *Darwin River* quarries.

Dimension stone

Two types of dimension stone are known from the Darwin region: Palaeoproterozoic Mount Goyder Syenite (trade name: Darwin Brown, **Figure 5.94**) and Cretaceous porcellanite. Darwin Brown is extracted from the *Halkitis Quarry* at Mount Bundey. Raw slabs are exported interstate and overseas, where they are cut and polished. Darwin Brown dimension stone can be seen in various places around Darwin City, for example, in the entrances to the Paspalis Centrepoint Building from the Smith Street Mall and the NT Parliament House. *Porcellanite* is a term used for silicified mottled claystone that occurs in the siliceous zone beneath the lateritic duricrust in the Darwin area. The parent unit is the lowermost Darwin Formation of the Bathurst Island Group. Porcellanite has been used as building stone in Darwin since 1879, notable examples include the Police Station and Courthouse, Town Hall (left in ruins after Cyclone Tracy in 1974), Browns Mart, Victoria Hotel, Anglican Church, Lyons Cottage and St Marys Cathedral. Abandoned porcellanite quarries are located at *Larrakeyah* and *Stuart Park*. Porcellanite is mined intermittently from a quarry at *Koolpinyah*.



Figure 5.94. Block of Mount Goyder Syenite (trade name: Darwin Brown) ready for export.

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