GEOSCIENCE AUSTRALIA RECORD 2006/13

REGIONAL GEOLOGY AND METALLOGENY OF THE EASTERN AILERON AND IRINDINA PROVINCES: A FIELD GUIDE

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Department of Industry, Tourism and Resources

Minister for Industry, Tourism and Resources: The Hon. Ian Macfarlane, MP
Parliamentary Secretary: The Hon. Bob Baldwin, MP
Secretary: Mark Paterson

ISSN: 1448-2177
ISBN: 1 920871 86 1

GeoCat No: 64220

Bibliographic reference: Huston, D.L., Maidment, D., and Hussey, K., 2006. *Regional geology and metallogeny of the eastern Aileron and Irindina Provinces: a field guide*. Geoscience Australia, Record 2006/13.

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1. INTRODUCTION

The last five to seven years have seen a major revision in our understanding of the geology of central Australia, with advancements in the field of geochronology redefining the geological framework and history of this area. Prior to about 1995, the eastern Arunta region was thought to consist of moderate to high grade metamorphic Palaeoproterozoic rocks with a complex geological history. This basement was interpreted to be overlain by low grade, mainly sedimentary rocks of the Georgina and Amadeus Basins, which are part of the extensive Neoproterozoic to late Palaeozoic Centralian Superbasin (Walter et al., 1995). However, workers at Adelaide, La Trobe, Australian National and Monash Universities found that rocks of the Harts Range Group (also referred to as the Irindina Supracrustal Assemblage) have a Neoproterozoic to early Palaeozoic depositional age, and were intensely metamophosed and deformed during the Ordovician (e.g. Miller et al., 1998, Mawby et al., 1999). Subsequent work by the Northern Territory Geological Survey (NTGS) and Geoscience Australia (GA) indicates that the Palaeoproterozoic part of the Arunta region can itself be divided into two discrete provinces: (1) the older (1840-1760 Ma) Aileron Province, and (2) the younger (1680-1610 Ma) Warumpi Province. Further work by the two geological surveys and the Australian National University suggests that in the eastern Arunta region, the Aileron Province can be divided into two sequences, the older (1820-1800 Ma) Strangways Metamorphic Complex and the younger (~1760 Ma) Oonagalabi Assemblage. Definition of these sequences became possible only because of the availability of a critical mass of geochronological data.

One of the purposes of this field trip is to illustrate the characteristics of the Palaeoproterozoic basement and the overlying Neoproterozoic to Palaeozoic rocks, with emphasis on the role of modern geochronology in resolving the newly defined terrains. The rocks of the eastern Arunta region have also undergone multiple deformation and metamorphic events that are now being unravelled. This complex history will also be illustrated during the excursion. The second aspect that will be illustrated in this field trip is the metallogeny of the eastern Arunta region. The region is characterised by a large variety of mineral deposits, including lode gold, volcanic-hosted massive sulphide (VHMS), carbonate replacement Zn-Cu, iron-oxide Cu-Au (IOCG), skarn W-(Mo-Cu-Au), carbonatite and aeolian/alluvial garnet deposits. However, most of the known deposits are small, with only the Molyhil W(-Mo-Cu-Au?) deposit, the Mud Tank vermiculite and the White Range Au deposits mined in recent times. Potentially the most economically important deposits are the garnet deposits and the Nolans Bore rare-earth-element (REE)-phosphate-U deposit. This excursion will visit examples of all of these deposit types.

During this excursion we will be travelling on pastoral leases, so gates should be left as found, and the property of the landholder must be respected. In deference to the traditional owners, hammering and sampling should be kept to a minimum, and, at some stops, hammers will be banned.

2. GEOLOGY AND EVOLUTION OF THE EASTERN ARUNTA

The Arunta region, which comprises the Aileron, Warumpi and Irindina Provinces (or Harts Range Group: Fig. 1), is one of the most geologically complex areas in Australia.

2A. SEDIMENTARY PACKAGES

Scrimgeour (2003: Table 1) identified nine metasedimentary packages and twelve discrete overprinting tectonothermal events in the Arunta region, indicating this is an important geological building block of central Australia. Four of Scrimgeour's (2003) sedimentary packages (the Ongeva, Cadney and Ledan packages and the Harts Range Group) are present in the area covered by this excursion (Fig. 2), and a fifth, the ~1840 Ma Lander package, is probably present. Although this last package has not been definitively identified in the eastern Arunta region, a correlative package may be present, as ~1812 Ma mafic igneous bodies (Claoue-Long and Hoatson, 2005) intrude parts of the Strangways Metamorphic Complex, and the Rankins Cu-Zn prospect has a Pb isotope model age of ~1822 Ma (Hussey et al., 2005). In the central and northern Aileron Province, the Lander package consists mainly of turbiditic pelites and psammites (Scrimgeour, 2003). The temporal relationships of the Ongeva, Cadney and Ledan packages and the Harts Range Group are shown relative to tectonothermal events in Table 1, and the geology of these units is described below.

Table 1. Depositional and tectonothermal events of the Arunta region.

Age (Ma)	**Warumpi Province**	**North Australia Craton—Aileron Province**				**Irindina Province**
		Central	Northern	Mt Hay-Mt Chapple	Eastern	
450-300	**Alice Springs Orogeny**					
480-460						**Larapinta Event**
520						**Stanovos Event**
850-500						Harts Range Group
1150-1130	*Teapot Event*			*Teapot Event*		
1590-1570	**Chewings Orogeny**					
1610-1600	*Ormiston Event*					
1620-1610	Iwupataka package					
1640-1630	*Liebig Orogeny*					
1660-1650	Yaya package					
1680-1660	Argilke Igneous Event					
1690-1670	Madderns package					
~1690				*Unnamed event*		
1730-1710		?	**Strangways Event**			
1770-1740				*Inkamulla Igneous Event*		
1770-1740			Ledan package		Ledan, Cadney packages	
1780-1770		**Yambah Event**				
~1780			Reynolds package			
1810-1790			Ongeva package	?	Ongeva package	
1820-1800		?	**Stafford Event**	?		
1865-1820		Lander package	Lander package	Lander package		

Shaded boxes indicate tectonothermal events. Bold names indicate major tectonothermal events, whereas italicised names indicate minor or local tectonothermal events.

2A.1. Ongeva package

The main supracrustal package in the eastern Arunta region, the Ongeva package, includes the lower part of the Strangways Metamorphic Complex (Lower and middle SMC on Fig. 2; excluding the Cadney Metamorphics) and the Bonya Schist, Deep Bore Metamorphics, Cacklebery Metamorphics, Kanandra Granulite and Mount Bleechmore Granulite further to the east. Geochronological data from the Strangways Metamorphic Complex, Bonya Schist and Deep Bore Metamorphics indicate ages between 1810 and 1800 Ma (Scrimgeour, 2003; Hussey et al., 2005) for this package. Lithologically, the Ongeva package consists of metapelitic and metapsammitic rocks with subordinate calc-silicate, marble, and felsic and mafic orthogneiss (Scrimgeour, 2003).

2A.2. Cadney package

Scrimgeour (2003) interpreted the Cadney package (Upperm SMC on Fig. 2), which includes marbles and calc-silicates of the Cadney Metamorphics, to have an age of 1780-1760 Ma. However, the age of this unit is poorly known, with its age constrained between ~1800 and ~1730 Ma by the underlying Strangways Metamorphic Complex and the overprinting ~1730 Ma Strangways metamorphic event. It is possible that the Cadney package may have been deposited shortly after the Ongeva package, with a depositional age of ~1800 Ma.

2A.3. Ledan package

The ~1770-1730 Ma Ledan package includes pelitic and psammitic metasediments that uncomformably overlie the Strangways Metamorphic Complex (Scrimgeour, 2003; Maidment et al., 2005). This package is interpreted to contain the Oonagalabi assemblage, which hosts the Oonagalabi deposit. Recent geochronological studies identified a single zircon population age of 1765 ± 4 Ma (Hussey et al., 2005), which was interpreted as a significant volcaniclastic component, implying that this age closely approximates the depositional age of the Oonagalagi assemblage.

2A.4. Harts Range Group

The Harts Range Group comprises a complex assemblage of granite gneiss, marble, calc-silicate, amphibolite, psammites and pelites (Table 2) that have been metamorphosed to upper amphibolite- to granulite-facies. Detrital zircon data from these rocks (Fig. 3) indicate that they are the high-grade metamorphic equivalents of sedimentary rocks in the adjacent Amadeus and Georgina basins (Buick et al., 2001, 2005; Maidment, 2005). Comparison of detrital zircon data from the high-grade metamorphic and unmetamorphosed successions indicates that the Harts Range Group was deposited between ~850 Ma and ~500 Ma (Fig. 4).

2B. TECTONOTHERMAL EVENTS

Of the twelve tectonothermal events that Scrimgeour (2003) interpreted to have affected the Arunta region (Table 1), seven (Stafford [1810-1790 Ma], Yambah [1780-1770 Ma], Inkamulla [1770-1740 Ma], Strangways [1730-1700 Ma], Teapot [1150-1130 Ma], Larapinta [480-460 Ma] and Alice Springs [450-300 Ma]) are recognised in the eastern Arunta region. Two of these events (Strangways and Larapinta) are dominated by extensive high-grade (upper amphibolite to granulite) metamorphism, with minimal magmatism, in the eastern Arunta region. Localised, low- to high-grade metamorphism, deformation and retrogression are associated with the 450-300 Ma Alice Springs Orogeny. The remainder are dominated by magmatism. In addition, the Mud Tank carbonatite was emplaced at ~732 Ma (Black and Gulson, 1978), and Maidment (2005) has recognised localised felsic magmatism of the ~520 Ma Stanovos event in the southeastern Harts Range Group.

2B.1. Stafford Event (1820-1800 Ma)

The oldest tectonothermal event affecting the eastern Arunta region, the ~1810-1790 Ma Stafford Event, involved the intrusion of mafic and felsic bodies and the deposition of extensive felsic volcaniclastic units in the Ongeva package, including the Strangways Metamorphic Complex.

2B.2. Yambah Event (1780-1770 Ma)

The Yambah Event, previously known as the Early Strangways Event, comprises the emplacement mostly of felsic, with lesser mafic, bodies. Although dominated by magmatism, this event was accompanied by local high grade metamorphism in the eastern Strangways Range and at Mount Hay further to the west (Claoue-Long and Hoatson, 2005). Zhao and Bennett (1995) documented extensive felsic magmatism of this age, and Zhao and McCulloch (1995) interpreted this magmatism to have formed by the melting of sources that had been produced in an earlier subduction environment.

Table 2. Lithostratography of the Harts Range Group and Entia Dome.

Unit	Description
Harts Range Group	
Brady Gneiss	Forms the structurally uppermost unit of the exposed Harts Range Group. Comprised of a lower unit dominated by garnet-bearing muscovite-biotite ± sillimanite schist and gneiss, with minor amphibolite and psammite; and an upper unit dominated by clinozoisite-hornblende-clinopyroxene-scapolite-bearing calc-silicate rock, with minor metapelitic and psammitic layers.
Irindina Gneiss	Relatively thick sequence of garnet-biotite-plagioclase ± sillimanite schist and gneiss, with minor layers of marble, calc-silicate and quartzite. Migmatitic in places, containing coarse-grained garnet up to ~15 cm in size. Intercalated with amphibolite of the Harts Range Meta-Igneous Complex.
Harts Range Meta-Igneous Complex (HRMIC)	Consists of numerous units of layer parallel amphibolite ranging from <1 cm to ~2 km thick within the Irindina Gneiss. Migmatitic in places with coarse-grained garnet ± clinopyroxene-bearing leucosomes. The presence of detrital zircon in the amphibolite and its intimate interlayering with metasediments indicate that many of these units are volcanics or volcaniclastics.
Naringa Gneiss	Calcareous unit that crops out in the western Harts Range, consisting of quartzite, hornblende-diopside-scapolite calc-silicate, marble, calcareous quartzofeldspathic gneiss and biotite gneiss. Occupies the structurally lowest position of the Harts Range Group.
Stanovos Gneiss	Crops out in the southeastern Harts Range, consisting of a lower calcareous unit of quartzite, grossular-diopside-wollastonite marble, calc-silicate and minor metapelite, and an upper unit dominated by garnet-poor metapelite with amphibolite, quartzofeldspathic gneiss and quartzite. Lower unit is a possible equivalent of the Naringa Gneiss.
Entia Dome	
Bruna Gneiss	Flat-lying, sheet-like layer of megacrystic granitic gneiss 50-400 m thick separating the Harts Range Group from the underlying Entia Gneiss Complex. Consists of quartz-biotite-plagioclase-K-feldspar-hornblende, with K-feldspar augen up to 5 cm in size. Commonly mylonitic, but locally contains domains preserving primary igneous textures.
Entia Gneiss Complex	Structurally lowest unit of the Entia Dome, cropping out in the core. Dominated by orthogneiss, ranging from ultramafic to granitic in composition, intruding a volumetrically minor supracrustal sequence minor calc-silicate, quartzite, aluminous metapelite and amphibolite.

2B.3. Inkamulla Event (1770-1740 Ma)

In the very southeastern part of the Arunta region, rocks of the Ledan package have been intruded by 1760-1740 Ma granitoids of the Inkamulla Igneous Event. The volcaniclastic rocks in the Oonagalabi assemblage may be early products of this event. This event is characterised by voluminous felsic and minor mafic magmatism in the western Strangways Range and Entia Dome, extending to the far east on the Tobermory 1:250 000 sheet. This event does not appear to have been associated with metamorphism (Scrimgeour, 2003). Geochemical studies by Zhao and McCulloch (1993) indicate that some Inkamulla granites appear to be related directly to a subduction event.

2B.4. Strangways Event (1740-1690 Ma)

Although in the southeastern Arunta, the Strangways Event (formerly Late Strangways Event) is dominated by granulite-facies metamorphism accompanied locally by partial melting (Scrimgeour, 2003), 1730-1700 Ma granite bodies are present through much of the northern Arunta and into the southern part of the Tennant region (Davenport Province). In the Strangways Range, the Wuluma Granite (Lafrance et al., 1995) is related to partial melting during high-grade metamorphism. The complex Strangways Event is the dominant tectonothermal event in the eastern Aileron Province

2B.5. Teapot Event (1150-1130 Ma)

The Mordor Igneous Complex is the main manifestation of the 1150-1130 Ma Teapot Event in the eastern Arunta region. This event, which affected the southern part of the Arunta region, also resulted in the emplacement of granite and isotopic resetting (Scrimgeour, 2003).

2B.6. Mud Tank carbonatite (~730 Ma)

The Mud Tank carbonatite, located ~10 km east of Gemtree, and related rocks intruded at ~732 Ma (Black and Gulson, 1978). Other carbonatites in the region are likely of similar age, although no other manifestation of this event is known.

2B.7. Stanovos Event (520 Ma)

The intrusion of felsic intrusives and migmatisation of the lower parts of the Harts Range Group at ~520 Ma record the effects of a tectonothermal event in the Harts Range region, termed the Stanovos Event (Maidment et al., 2004; Maidment, 2005). Many of the felsic intrusives have gradational contacts with migmatitic metasedimentary host rocks and limited dating of inherited zircon cores shows a similar range of ages to those in the the latest Neoproterozoic/Early Cambrian host rocks. These features suggest that the felsic intrusives formed, at least in part, by partial melting of ~545 Ma sediments at ≥520 Ma, implying rapid burial of sedimentary rocks within a palaeogeographic setting that was dominantly shallow marine in character. The ~520 Ma felsic intrusives show evidence of magma mingling with mafic intrusives and form part of a bimodal igneous complex. The bimodal magmatism and burial of sediments within a persistently marine setting imply an extensional driver for this tectonism (Maidment, 2005).

2B.8. Larapinta Event (480-460 Ma)

Metamorphsim of the Harts Range Group at up to granulite-facies conditions (~800 ºC, 1.0-1.2 Gpa – 30-35 km) took place during the 480-460 Ma Larapinta Event. Metamorphism was associated with the formation of a shallowly dipping foliation, N-S to NNE-SSW trending mineral lineations and the intrusion of mafic dykes. Peak metamorphism at ~480 Ma coincided with the formation of a large marine embayment across central Australia, the Larapintine Sea, and the deposition of fine-grained sediments in a depocentre above the locus of metamorphism. These observations indicate that metamorphism occurred within an extensional setting. This implies that burial of Cambrian sediments to 30-35 km occurred by progressive sedimentation within an exceptionally deep sub-basin, in what amounts to an extreme form of burial metamorphism (Hand et al., 1999a,b; Maidment et al., 2004; Maidment, 2005). As improbable as this interpretation may sound at first, there are several rift basins known worldwide with seismically imaged depths of 25-30 km, demonstrating that deep sediment burial does not necessarily require a compressional driver.

2B.9. Alice Springs Orogeny (450-300 Ma)

The Alice Springs Orogeny was a major period of N-S to NE-SW intraplate directed shortening between ~450 and 300 Ma that affected much of central Australia. Tectonism during this period was distinctly episodic, with phases of movement recognised by changes in patterns of sedimentation in the Amadeus and Georgina basins (Bradshaw and Evans, 1988; Shaw, 1991), K-Ar and Rb-Sr cooling ages in the eastern Arunta region (Dunlap, 1997; Dunlap and Teyssier, 1995; Dunlap et al., 1995), Sm-Nd dating of shear zones (Bendall et al., 1998) and metamorphic zircon and monazite in the areas of highest metamorphic grade (Hand et al., 1999a; Maidment, 2005; Maidment et al., 2005). The effects of the Alice Springs Orogeny appear to have been most intense in the eastern Arunta region, where tectonism was expressed as amphibolite-facies metamorphism, felsic magmatism and S- to SW-verging folding and thrusting. It was at this time that the Harts Range Group was exhumed and thrust southwards over Palaeoproterozoic basement.

3. MINERALISATION

Mineral deposits in the Arunta region vary in commodity, style and age. This excursion visits a variety of mineral deposits with ages ranging from ~1800 Ma to those formed in the Cainozoic. Although base-metal and gold deposits in the Arunta are relatively widespread and geologically interesting, these deposits have been economically insignificant. The economically most important deposits are industrial minerals: vermiculite associated with the weathered rocks in the Mud Tank carbonatite complex, and garnet-amphibole-rich sands concentrated by aeolian and alluvial processes to the north of the Harts Ranges. In addition to these deposits, the most significant recent discovery in the Aileron Province is the Nolans Bore REE deposit, which is developing into a world class deposit. With the exception of Mud Tank vermiculite deposits, this excursion visits examples of all geologically or economically significant mineral deposits in the eastern Arunta region.

The oldest deposits in the eastern Arunta are base-metal and gold deposits hosted by the Strangways Metamorphic Complex, Bonya Schist and Cadney Metamorphics. These deposits were grouped together by Warren and Shaw (1985) into their Oonagalabi-type and interpreted as VHMS deposits. Our recent work, however, suggests that there are systematic differences within Warren and Shaw's (1985) Oonagalabi-type. Although most of these deposits are associated with marble lenses, our data suggest that there are important differences in the age of the host rock, the abundance of magnetite, metal assemblages and the composition of the dominant alteration assemblage. Hence, we have divided the Oonagalabi-type of Warren and Shaw (1985) into three sub-types: (1) the Utnalanama-type, which we interpret as VHMS deposits, (2) the Johnnies-type, which we interpret as IOCG deposits, and (3) the re-defined Oonagalabi-type, which we interpret as either carbonate-replacement or VHMS deposits. Table 3 summarises the characteristics that distinguish these three groups.

Table 3. Characteristics of Palaeoproterozoic Zn-Cu-Pb(Ag-Au) deposits in the eastern Arunta.

Type	Metal assemblage	Other elements	Host	Alteration assemblages	Interpreted age (Ma)
Utnalanama	Mineralised marble: Zn-Pb-Cu(Ag-Au) Calc-silicate: Pb-Zn	Mineralised marble: Bi-Cd Calc-silicate: Sn, HFSE, REE	Marble and calc-silicate after carbonate rocks.	Quartz-cordierite± orthopyroxene rock > massive amphibole± spinel±clinopyroxene rock. Both are concentrated in the footwall to mineralised marble lens.	1810-1800 (age of host); calc-silicate may be younger
Johnnie's	Lode rock: Cu-Pb(Zn-Ag-Au) Footwall garnetiferous zone: Au(Cu)	Lode rock: Mn-Ca-HFSE-REE Footwall garnetiferous zone: Bi±Mo	Lode rock: magnetite-diopside-amphibole± quartz rock (after marble). Footwall garnetiferous zone: Quartz-biotite-garnet±magneite gneiss.	Quartz-biotite-garnet gneiss in structural footwall to lode rock.	1795-1770 (Pb isotope model age)
Oonagalabi	Zn-Cu-Pb(Ag-Au)	Bi	Marble → calc-silicate → massive anthophyllite schist.	Quartz-garnet rock symmetrically developed about host marble lens.	1765 (?) (age of host)

Utnalanama-type deposits, which constitute the majority of known Palaeoproterzoic deposits in the Strangways Metamorphic Complex, are Zn-Pb-Cu(Ag-Au) deposits characterised by the extensive development of asymmetric alteration zones dominated by quartz-cordierite ± orthopyroxene ± biotite ± orthoamphibole ± garnet gneiss. Feldspar is typically absent in these rock types. Most of the quartz-cordierite rocks lack magnetite and have a very low magnetic susceptibility. Magnetite is not a major component of the ores or alteration assemblage. However, localised magnetite rich zones do occur. Geochemical analyses suggest that prior to metamorphism the quartz-cordierite rocks had a quartz-chlorite ± muscovite/illite protolith, which we interpret to be the alteration assemblage associated with mineralisation. Other minerals inferred to be present in the protolith alteration assemblages at these deposits include talc, tremolite and carbonate. These rocks are Mg-rich with (Mg/Mg+Fe) values typically between 0.6 and 1.0. The altered rocks have $\delta^{18}O_{whole\ rock}$ between 1.8 and 7.0‰ (most values between 2.5 and 4.8‰), consistent with formation through interaction with high temperature, evolved seawater. These deposits have 100Zn/(Zn+Pb) values mainly between 60 and 75. The presence of asymmetric proto-quartz-chlorite alteration zones formed via interaction with heated seawater and the 100Zn/(Zn+Pb) values are characteristic of VHMS deposits (Hussey et al., 2005).

Johnnies-type deposits, which include Johnnies Reward and Gumtree in the Strangways Metamorphic Complex and the base-metal-Au deposits of the Jervois district in the Bonya Schist further to the east, are Cu-Au(Pb-Zn-Ag) deposits characterised by a close association with abundant magnetite, and an asymmetric quartz-biotite-garnet ± feldspar alteration assemblage. These deposits are closely associated with magnetite, either in a magnetite-diopside ± amphibole skarn assemblage (e.g. Johnnies Reward) or in an iron formation (amphibole-quartz-magnetite rocks, e.g. Gumtree). The host rocks are considerably more Fe rich than the Utnalanama-type deposits, with (Mg/Mg+Fe) values usually between 0.3 and 0.6. Although base-metals are most concentrated in magnetite-rich zones, Au is concentrated in the structural footwall of these deposits. Johnnies-type deposits are characterised by highly variable 100Zn/(Zn+Pb), with Pb concentrations generally greater than Zn abundances. Gold values are typically one or two orders of magnitude higher in the Johnnies-type than the Utnalanama-type. Moreover, at Johnnies Reward, Mn and some high field strength elements (HFSE) and REE are highly enriched in places within the lode. Based on these characteristics, Johnnies-type deposits are more likely to be IOCG deposits rather than VHMS deposits (Hussey et al., 2005).

Oonagalabi-type deposits, which are represented by the Oonagalabi deposit and two nearby prospects, are hosted by the ~1765 Ma Oonagalabi assemblage of the Ledan package. Like the Utnalanama-type deposits, Oonagalabi-type deposits are not associated with abundant magnetite, are characterised by a Zn-Cu-Pb(Ag-Au) metal assemblage and have high 100Zn/(Zn+Pb) ratios (87 at Oonagalabi). However, unlike the Utnalanama-type deposits, the main alteration assemblage outside of the host marble is a quartz-garnet-feldspar rock: quartz-cordierite gneiss is rare. Carbonate in the ore host is progressive replaced by calc-silicate and then massive anthophyllite rock. All three rock types are mineralised. These characteristics are most consistent with a carbonate replacement origin, although a VHMS origin cannot be ruled out (Hussey et al., 2005). If Oonagalabi-type deposits are carbonate replacement deposits, the 1760-1740 Ma Inkamulla igneous suite may have been involved in their formation.

The Molyhil Mo-W skarn deposit, which is located just to the north of the Delny-Sainthill Shear zone on the Huckitta 1:250000 sheet (SF53-11), is hosted by unnamed metasedimentary rocks that are present as rafts within the Marshall Granite (Freeman, 1990). Hornblende from exoskarn associated with the deposit yielded an ^{40}Ar-^{39}Ar plateau age of 1702 ± 5 Ma (G. Fraser, pers. comm., 2003). Given uncertainties associated with ^{40}Ar-^{39}Ar ages relative to U-Pb ages (c.f. Min et al., 2000), the ^{40}Ar-^{39}Ar data constrain the age of exoskarn (and mineralisation) to 1720-1700 Ma, which is consistent with the age of granites elsewhere in the area (e.g. Mt Swan Granite at c. 1713 Ma: Zhao and Bennet, 1995).

The 1132 ± 5 Ma (Claoue-Long and Hoatson, 2005) Mordor Igneous Complex hosts orthomagmatic PGE-Au-Cu-Ni prospects associated with ultramafic rocks in this alkaline igneous suite. These prospects are the only known deposits of this type in the Arunta region. The Mud Tank carbonatite, which has been dated at 732 ± 5 Ma (Black and Gulson, 1978), hosts gem quality zircon. However, vermiculite deposits, which formed from the weathering of biotite, are economically the most important deposits in the eastern Arunta, with 69,693 tonnes of vermiculite products sold between 1995 and December 2003, with an estimated value of A$18-25M. Another 30,000 tonnes or more of refined vermiculite product can be extracted from the current stock pile and open cut. In addition, two more open cuts of comparable size planned are within the current mine lease.

The undated Nolans Bore REE-phosphate-U deposit consists of a series of REE-bearing fluoro-apatite veins that are hosted mostly by granite gneiss in the southeasternmost Reynolds Ranges near Aileron. This deposit is the most significant REE deposit in Australia and is comparable in size and grade to other large REE deposits around the world. More details of this deposit are provided in section 4W.

Garnet-rich para-amphibolites of the Harts Range Group are the source of the other major industrial mineral deposit in the eastern Arunta region. Rivers draining the Harts Range and aeolian processes have concentrated garnet and hornblende into potentially economic deposits along the northern margin of the ranges. Although the dunes were last stabilised about 20,000 years ago, alluvial processes currently concentrate garnet and hornblende in sands of creeks draining the ranges.

4. EXCURSION STOPS

Over the three day period of this excursion, a total of 19 excursion stops are planned. Most should take between 30 minutes and one hour, although stops at certain prospects will take up to 3-4 hours. Although most stops are designed to minimise the amount of walking, walks of 2 km at the Johnnie's Reward (Stops 4.2 and 4.3) and 1 km at the Edwards Creek prospect (Stop 4.4) are required. Consequently, good quality boots, a hat, other sun protection and drinks are required. To minimise the impact on the rocks we ask that hammers be used sparingly, particularly as some exposures visited may have significance to traditional land-owners. We are visiting the stops on this excursion with permission from leaseholders, both pastoral and exploration, and, in some cases, traditional owners. Gates should be left as found, and the property of the landholder must be respected. Figure 1 shows the location of stops and the main roads upon which we will travel.

4A. STOP 1.1. PALAEOPROTERZOIC-NEOPROTEROZOIC UNCONFORMITY AT HEAVITEE GAP

The Heavitree Gap contains the type-section for the Neoproterozoic Heavitree Quartzite. At this location, the unconformity is exposed between the Palaeoproterozoic Sadadeen Gneiss, which forms part of the Hayes Metamorphic Complex, and the Neoproterozoic Heavitree Quartzite, which is the basal unit of the Amadeus Basin. Clarke (1974) divided the fluvial to shallow marine Heavitree Quartzite into the Undoolya Siltstone, Temple Bar Sandstone, Fenn Gap Conglomerate and Blatherskite Members and suggested an east or northeast provenance. The unconformity is marked by the contact between coarsely feldspar-phyric, foliated biotite granite (Sadadeen Gneiss) and a 16 m thick, buff to red mudstone/siltstone (Fig. 5) with minor fine-grained sandstone lenses up to 1 m across and 0.3 m thick in the upper part. These sandstone lenses are inferred to be sand-filled channels (Clarke, 1974). This mudstone unit, the Undoolya Siltstone Member of the Heavitree Quartzite, is only locally developed at the unconformity (Clarke, 1974). Elsewhere the unconformity is marked by conglomeratic and locally arkosic units of the upward fining Temple Bar Sandstone Member.

The basement Sadadeen Gneiss, although undated, is intruded by, or has gradational contacts, with the Alice Springs Granite, which has been dated at 1752 ± 11 Ma (Zhao and Bennett, 1995). This suggests that the Sadadeen Gneiss and the Alice Springs Granite were both intruded during the Inkamulla Igneous Event. These two units make up many of the hills within the town of Alice Springs. The fabric in the Sadadeen Gneiss, which is defined by biotite, is parallel to bedding within the Heavitree Quartzite, suggesting that the fabric was flat lying as sedimentation in the Amadeus Basin initiated.

The age of the Heavitree Quartzite is poorly constrained. Correlations with other units of the Centralian Superbasin suggest that the Heavitree Quartzite was deposited about ~840 Ma (e.g. Walter et al., 1995). The Heavitree Quartzite unconformably overlies 1080 Ma Stuart Pass Dolerite (Zhao and McCulloch, 1993), and is conformably overlain by the Bitter Springs Formation which contains Late Riphean (950-680? Ma) aged fossils (Walter, 1972; Shergold et al., 1991). Recent detrital zircon studies indicate a maximum depositional age of ~1050 Ma (D. Maidment, unpub. data), which allows for the possibility that it was deposited significantly earlier than 840 Ma.

The drive eastwards along the Ross River Road provides spectacular views of the Heavitree Quartzite, which forms prominent ridges, and younger Neoproterozoic and Cambrian sediementary units of the Amadeus Basin.

4B. STOP 1.2. ARUMBERA SANDSTONE

The Ross River has cut an excellent section (Fig. 6) that exposes many of the units that make up the Ooraminna Sub-basin in the northeastern part of the Amadeus Basin. At this and the next stops we examine units that are the approximate temporal equivalents of high grade metamorphic rocks of the upper Stanovos Gneiss that we will see at stop 2.4. At this stop, we examine the Arumbera Sandstone (Fig. 7), a cyclical red-brown mudstone and sandstone unit consisting of four upward coarsening members spanning the Neoproterozoic-Cambrian boundary. The lower two members of the Arumbera Sandstone contain a variety of non-skeletal metazoan fossils that have been deposited in the Neoproterozoic, contemporaneous with the Ediacara fauna in the Adelaide Geosyncline (Walter et al., 1989). The upper two members of the Arumbera Sandstone contain a wealth of trace fossils and were deposited in the Early Cambrian. At this stop, the upper sequence of the lower Cambrian Arumbera Sandstone passes upwards from massive, fine- to medium-grained, dirty sandstone into laminated shales, siltstone and rare carbonates. Trace fossils are visible in several loose boulders. Additional information about this unit can be found in Freeman et al. (1987), Walter et al. (1989), Shergold et al. (1991), and Kennard and Lindsay (1991).

Dating of detrital zircons in the Arumbera Sandstone at this locality (D. Maidment, unpub. data) and at Ellery Creek, 80 km west of Alice Springs (Buick et al., 2005) yields an age spectrum dominated by grains between ~1000 and 1200 Ma (Fig. 4). This is interpreted to reflect uplift of the Musgrave Block to the south of the Amadeus Basin during the Petermann Orogeny between ~570 and 530 Ma, which exposed voluminous felsic intrusives of this age.

4C. STOP 1.3. TODD RIVER DOLOMITE

The lower Cambrian Todd River Dolomite conformably overlies the Arumbera Sandstone (Fig. 6). Four of the six lithofacies developed within the Todd River Dolomite are apparent at this stop (Kennard, 1991). The basal unit (lithofacies 1), which crops out relatively poorly, consists mainly of dolostone with stromatolitic bioherms, and is interpreted as a very shallow subtidal and intertidal mixed siliciclastic-carbonate flat. The second unit (lithofacies 3) consists mainly of fine-grained sandstone with lesser dolostone, which are interpreted as a series of barrier bars that have transgressed the underlying carbonate flat. The third unit (lithofacies 4) consists mainly of dolostone with archaeocythans and possible stromatolites. Two horizons with large bioherms are well exposed at the southern edge of the water pool. This lithofacies is interpreted to have formed in a high energy, intermittently emergent, shallow marine environment. The uppermost unit (lithofacies 6) consists of microbial boundstone (stromatolites) and massive dolo-mudstones, and is interpreted to have formed within a cyanobacterial mud flat. Additional information about this unit and the Ross River section can be found in Freeman et al. (1987), Kennard (1991), and Kennard and Lindsay (1991).

4D. STOP 1.4. ROSS RIVER SYNCLINE

This stop is situated in the core of the Ross River Syncline and provides a view of the Cambro-Ordovician Pacoota Sandstone in the hillside to the west (Fig. 8). The Pacoota Sandstone is a widespread, dominantly marine siliciclastic unit that forms the principal hydrocarbon reservoir in the Amadeus Basin. Detrital zircon ages from the second lowest sandstone unit are dominated by grains between ~ 950 and 1200 Ma, with a significant population of grains between ~500 and 600 Ma. The 500-600 Ma grains have no apparent central Australian source and appear to reflect the development of a marine embayment into central Australia that allowed detritus from the Pacific Gondwana margin to move westwards into the central Australian region.

4E. STOP 1.5. ARLTUNGA POLICE STATION

Restored stone ruins at this site are the remains of the old Arltunga police station and gaol, which were constructed in 1912. Additional ruins, which can be reached by following a walking track about 1 km to the north-northwest, are present at the government battery and cyanide works, which were operational between 1898 and 1934 (Mackie, 1986). Alluvial gold was first discovered in 1887 at Paddy Rockhole, with reef mining beginning prior to 1890. The opening of the government battery coincided with discovery of the White Range veins (Fig. 9), which produced 81% of the 15,396 ounces (0.479 tonnes) produced in the Arltunga goldfield prior to 1984 (Mackie, 1986). After 1989 an additional 2.098 tonnes of Au were produced from the White Range veins (to December 2003: Ahmad et al., 1999; P. Lamuri, pers. comm., 2004). Historical mining around Arltunga was most active between 1896 and 1913, with peak production in 1903 corresponding to the peak in population (Mackie, 1986).

The Arltunga goldfield occurs within the Alice Springs Arltunga Nappe Complex (~430-300 Ma: Stewart, 1971; Dunlap et al., 1995). The Arltunga Nappe Complex was formed when a south-directed thrusting event resulted in the current structural duplex arrangement of the Amadeus Basin and the Palaeoproterozoic basement. The deposits are hosted both by the Palaeoproterozoic basement and by the Neoproterozoic Heavitree Quartzite (Fig. 9), and are inferred to have an age of ~300-290 Ma (late Alice Springs Orogeny) based on structural relationships of Au-bearing veins and ^{39}Ar-^{40}Ar ages of white micas (J. Dunlap, pers. comm., 2002). Palaeoproterozoic rocks that host the deposits (Fig. 3) include the Cadney Metamorphics (marble and calc-silicates), the Hillsoak Bore Metamorphics (predominantly metasediments, including calcareous units and rare marbles, and amphibolites), the Cavenagh Metamorphics (mainly metasediments, including calcareous units, and quartzofeldspathic gneiss with minor iron formation) and the Atnarpa

Igneous Complex (retrogressed tonalitic gneiss: Mackie, 1986). Of these units, only the Atnarpa Igneous Complex has been reliably dated at ~1770 Ma (Zhao and Bennett, 1995).

4F. STOP 1.6. MACDONNELL RANGE REEF

The Macdonnell Range reef, which can be traced along strike for about 300 m, was mined from several small shafts and diggings along its easternmost extent. In addition, alluvial diggings worked a small creek that drained the eastern end of the reef. The reef has a sinuous outcrop pattern with a broadly east-west strike and a shallow (15°) northerly dip (Mackie, 1986). The vein orientation is subparallel to that of the foliation developed in the enclosing metasedimentary rocks, which are calcareous in vicinity of the workings (Mackie, 1986). The vein varies in thickness from 0.2 to 0.5 m and consists of white quartz that is variably iron stained and locally contains malachite. A thin, phyllosilicate-rich alteration zone is locally developed on the hanging wall, and Mackie (1986) reported a sheared, graphitic alteration zone ("black plumbaginous slate") along the footwall of the vein. This reef produced a total of 248 ounces of gold from 353 tons of ore between 1896 and 1908 with an average grade of 0.70 oz/ton or 21.8 g/t (Mackie, 1986).

Fluid inclusions from the White Range veins, to which this reef is probably related, homogenise to both the liquid and vapour phase or show critical behaviour. Homogenisation temperatures vary between 280 and 325°C, with most between 315 and 325°C. The inclusions are of low salinity (0-3.5 eq wt % NaCl) and contain significant CO_2, but no other gases (A. Wygralak, pers. comm., 2004). The fluid inclusion data are consistent with phase separation, and the compositions are typical of lode-gold or distal intrusion-related gold ore fluids.

4G. STOP 1.7. HEAVITREE QUARTZITE, MORDOR POUND

This stop along the track provides a view of the Heavitree Quartzite, which forms the ridges around the margin of the Mordor Pound (Fig. 10). The Heavitree Quartzite forms a sheet that was thrust across the Mesoproterozoic rocks of the Mordor Igneous Complex, presumably during the ~450-300 Ma Alice Springs Orogeny. The view of the eastern wall of the compound at this point shows a large-scale recumbent fold (Fig. 11) that formed during the south-directed emplacement of the quartzite sheet.

4H. STOP 1.8. SYENITE, MORDOR IGNEOUS COMPLEX

The Mordor Igneous Complex is a plug-like alkaline body that forms a 6 by 6 km body inside the Mordor Pound. The complex intrudes Palaeoproterozoic schist, gneiss and amphibolite that form the floor of the pound (Fig. 10). The walls to the pound are capped by Neoproterozoic Heavitree Quartzite (Hoatson and Stewart, 2001). This complex has been dated at 1132 ± 5 Ma using zircons from a phlogopite-plagioclase-clinopyroxene pyroxenite plug (Claoué-Long and Hoatson, 2005).

The stop is located approximately 100 m north of the road. Mordor syenite generally crops out poorly, typically forming sandy plains dominated by feldspar. At this location a number of low bouldery outcrops are present. The syenite is homogeneous and coarse-grained with minor phlogopite and rare plagioclase. Hoatson and Stewart (2001) suggest that the strong alignment of feldspar characteristic of the syenite and other felsic units of the Mordor Igneous Complex is the consequence of flow lamination.

4I. STOP 1.9. MITHRIL PGE-AU PROSPECT

The Mithril PGE-Au prospect is an orthomagmatic deposit hosted by the layered ultramafic phase of the Mordor Igneous Complex (Fig. 12). Three diamond drill holes have intersected mineralisation at this locality, with the best intersection being 2 m grading 1.1 g/t Pt+Pd+Au within an 8 m interval assaying 0.67 g/t Pt+Pd+Au. The PGE mineralisation is associated with disseminated sulphides (chalcopyrite > pyrrhotite ~ pentlandite) within a pyroxenite immediately underlying a olivine-bearing unit (Tanami Gold Ltd. Announcement to Australian Stock Exchange, 20 December 2002). These ultramafic rocks typically contain phlogopite, with some of the olivine-bearing rocks containing up to 40% phlogopite. Tanami Gold Ltd (announcement to Australian Stock Exchange, 20 December 2002) suggest that the PGE mineralisation formed by the injection of a primitive (i.e. olivine-bearing) magma into a differentiating magma chamber.

4J. STOP 2.1. FLORENCE-MULLER SHEAR ZONE AT FLORENCE CREEK

At this locality the track crosses the Florence-Muller Shear Zone, a structure that juxtaposes the Harts Range Group to the north against the Strangways Metamorphic Complex to the south (Fig. 2). The Harts Range Group was pervasively metamorphosed at granulite-facies conditions during the Early Ordovician Larapinta Event, whilst adjacent rocks in the Strangways Metamorphic Complex preserve only sporadic evidence of minor isotopic disturbance at this time. The Florence-Muller Shear Zone thus forms a major boundary separating these two domains in the eastern Arunta region.

The Florence-Muller Shear zone dips gently northwards, is up to 2 km thick and is developed within interlayered mafic, felsic and metasedimentary gneisses of the Strangways Metamorphic Complex (Fig. 13). Mafic assemblages consist of hornblende-garnet-plagioclase-quartz ± clinopyroxene. In metapelitic rocks the shear fabric is defined by biotite-sillimanite-K-feldspar-quartz-plagioclase. A north-dipping stretching lineation is well-developed and the majority of shear sense indicators indicate top-to-south movement. The presence of some shear sense indicators with an opposite sense of movement might reflect an earlier phase of top-to-north normal movement, as suggested by James and Ding (1988) and Ding and James (1989).

A maximum age for thrusting in the Florence Muller Shear Zone is constrained by the ~480-460 Ma age of the Larapinta Event in the hangingwall rocks. A ^{40}Ar-^{39}Ar hornblende age of 400 Ma obtained from the hangingwall further to the southeast near Mout Ruby (Dunlap et al., 1995) indicates that the hangingwall cooled through ~500 ºC at this time. Since the shear fabrics in the Florence-Muller Shear Zone are up to upper amphibolite-facies metamorphic grade, the ~400 Ma cooling age provides a minimum age for the movement, thus constraining south-directed thrusting to between ~460 and ~400 Ma, the earliest stages of the Alice Springs Orogeny.

4K. STOP 2.2. IRINDINA GNEISS AT LIZZY CREEK

The Irindina Gneiss of the Harts Range Group is exposed in Lizzy Creek, 10.5 km east of Florence Creek. The Irindina Gneiss consists of a relatively thick succession of metapelite, marble, calc-silicate and minor quartzite (Fig. 14). It contains numerous layer-parallel units of metabasite up to 2 km thick, which are collectively termed the Harts Range Meta-Igneous Complex (Sivell and Foden, 1985). Detrital zircon data from the Irindina Gneiss ~7 km to the east of Lizzie Creek, and at Mallee Bore to the north of the Harts Range contain populations of relict detrital zircons with ages ranging as low as 520-510 Ma. Comparison of the detrital zircon age spectra with rocks of known depositional age in the adjacent Amadeus and Georgina basins suggests that the Irindina Gneiss was deposited during the Early to Middle Cambrian, at a similar time to the shallow marine sandstones and carbonates seen in Ross River Gorge on Day 1 stops 1.2 to 1.4). The more pelitic nature of the Irindina Gneiss is consistent with the interpretation that the Irindina Gneiss and associated mafic volcanics were deposited within a deeper extensional sub-basin between the Amadeus and Georgina basins.

Peak metamorphic assemblages in the Harts Range Group formed at ~800 ºC and 10-12 kbar (Miller et al., 1998; Mawby et al., 1999), but are rarely preserved due to the overprinting effects of later retrograde tectonism during the Larapinta Event. South of the track crossing, peak metamorphic metabasite is preserved within one of these low-strain domains. The metabasite consists of coarse-grained migmatitic amphibolite with garnet- and hornblende-bearing partial melts. North of the track metapelite and metabasite are overprinted by an intense flat-lying foliation that characterises the retrograde phase of the Larapinta Event. Layers of metabasite within the metasedimentary rocks are boudinaged and the metapelite contains a variably developed N-S oriented stretching lineation.

Around 300 m north of the track a small cliff face on the eastern side of the creek exposes representative lithologies of the Irindina Gneiss. A shallowly-dipping layer of coarse-grained marble is interbedded with garnet-biotite-sillimanite-plagioclase-quartz metapelite and calc-silicate rock. The marble contains blocks of metapelite and calc-silicate which have been incorporated within the more ductile marble during deformation. The intense, shallowly-dipping layer-parallel foliation is typical of that formed during the later retrograde phase of the Larapinta Event, and is one of the lines of evidence that supports an extensional setting for this early Palaeozoic tectonism.

4L. STOP 2.3. CONTACT BETWEEN ENTIA GNEISS AND BRUNA GNEISS

The Palaeoproterozoic Entia Gneiss Complex forms basement to the Neoproterozoic to Cambrian Harts Range Group. It crops out in the core of a complex domal structure, the Entia Dome, which formed during the later stages of the Alice

Springs Orogeny (Maidment, 2005). The Entia Gneiss Complex is dominated by felsic and mafic orthogneiss, with minor supracrustal rocks, which include metapelite, calc-silicate rock, marble and quartzite. SHRIMP U-Pb ages of detrital zircons from a calc-silicate unit in the centre of the Entia Dome is dominated by grains with ages of ~2500 Ma, with no younger Palaeoproterozoic grains (D. Maidment, unpubl. data).

Tonalitic to granodioritic intrusives within the Entia Gneiss Complex have high Na_2O, Na/K, Sr, K/Rb and Sr/Y, with relatively low K_2O, Rb, Rb/Sr, Th, U, REE, Nb and Y, similar geochemical features to calc-alkaline suites in modern continental arcs (Foden et al., 1988; Zhao and McCulloch, 1995). They are also associated with amphibolites which have been interpreted as arc-type metatholeiites (Sivell, 1988). The arc-like granitoids have SHRIMP ages of 1762 ± 3 Ma and 1773 ± 4 Ma (Maidment et al., 2005) and form part of the Inkamulla Igneous Event of Scrimgeour (2003). Conventional zircon dating of intrusives from the Entia Gneiss Complex has yielded ages between ~1765 and ~1730 Ma (Cooper et al., 1988; Foden et al., 1995).

The Entia Gneiss Complex is separated from the overlying Harts Range Group by the Bruna Gneiss, a sheet of megacrystic felsic gneiss between 50 and 400 m thick. The Bruna Gneiss is geochemically distinct from the underlying Entia Gneiss Complex, with A-type geochemical affinities (Foden et al., 1988), and appears to have been emplaced at shallower crustal levels (Foden et al., 1988; Hand et al., 1999b). The Bruna Gneiss has a conventional zircon age of 1747 ± 2 Ma (Mortimer et al., 1987; Cooper et al., 1988), indistinguishable from that of similar megacrystic gneisses that have been dated elsewhere in the Harts Range (Maidment, 2005). The Bruna Gneiss is typically strongly foliated, and has intensely mylonitic upper and lower contacts. This zone of high-strain has been termed the Bruna Detachment Zone (Mawby et al., 1999). Ding and James (1985, 1989) and James and Ding (1988) considered that the Bruna Gneiss was emplaced syntectonically along the Bruna Detachment Zone, and that its 1747 Ma age thus defined the minimum depositional age of the Harts Range Group and the juxtaposition of the basement and cover sequences. Despite the absence of a clear intrusive contact with the Harts Range Group, this interpretation was adopted by most other workers in the area and was only challenged recently when the first detrital zircon age data were collected.

At this locality (Fig. 15) the track runs along the contact between the Entia Gneiss Complex and the Bruna Gneiss. Equigranular quartzofeldspathic gneiss of the Entia Gneiss Complex crops out to the north of the track, and is structurally overlain by the Bruna Gneiss to the south. The gneisses in this area contain a pervasive shallowly-dipping foliation, which elsewhere in the Entia Dome has been shown to have formed during the Alice Springs Orogeny (Hand et al., 1999a; Maidment, 2005). This foliation is deformed by isoclinal recumbent folds with fold axes parallel to a N-S lineation defined by elongate quartz aggregates. On the southern side of the road biotite-K-feldspar-quartz-plagioclase-hornblende gneiss of the Bruna Gneiss forms hills up to a couple of hundred metres in height. Given the relatively shallow dip of the unit, the height of the hill gives a minimum thickness for the Bruna Gneiss in this area.

4M. STOP 2.4. UPPER STANOVOS GNEISS

The Stanovos Gneiss structurally underlies the Irindina Gneiss and is the lowest recognised unit in the Harts Range Group. It consists of two lithological associations: 1) a lower calcareous unit of quartzite, grossular-diopside-wollastonite marble, calc-silicate and minor metapelite; and 2) an upper unit dominated by garnet-poor biotite gneiss with amphibolite, quartzofeldspathic gneiss and quartzite. The Stanovos Gneiss was considered by Shaw et al. (1982) to be a member of the Irindina Gneiss, but reinterpretation of this relationship in light of new detrital zircon dating suggests that the Stanovos Gneiss is older than the Irindina Gneiss (Maidment, 2005).

A small hill adjacent to the track in the Stanovos Valley consists of a biotite-rich quartzofeldspathic gneiss and amphibolite with minor calc-silicate and quartzite, typical of the upper Stanovos Gneiss. The sequence is tightly folded and the amphibolite layers are attenuated and boudinaged. Detrital zircon ages from the upper Stanovos Gneiss ~7 km to the southeast are dominated by grains between ~ 1000 and 1200 Ma with a single analysis at ~630 Ma, a similar result to that obtained from the Early Cambrian Arumbera Sandstone at Stop 1.2 (Fig 4). The upper Stanovos Gneiss is interpreted as an approximate correlative of the Arumbera Sandstone, deposited in the latest Neoproterozoic to Early Cambrian sourcing sediment from the Musgrave Inlier which was uplifted during the Petermann Orogeny at 560-530 Ma (Edgoose et al., 2003). The generally more pelitic character of the upper Stanovos Gneiss suggests that it was deposited in deeper water than similarly-aged sediments in the Amadeus Basin. The widespread mafic units within the upper Stanovos Gneiss are absent in the adjacent basins, consistent with the interpretation that the upper Stanovos Gneiss was deposited within a rift-related sub-basin between the Amadeus and Georgina basins (Maidment, 2005).

A perspective of the local geology can be gained from the top of the hill at this stop (Fig. 16). The view to the NNW along the Stanovos Valley (Fig. 17) shows a line of hills along the eastern (right hand) side of the valley, informally referred to as Indiana Walls (Fig. 18). These consist of biotite-rich gneiss, amphibolite and a ridge-forming unit of quartzofeldspathic gneiss within the upper Stanvos Gneiss. The hills on the western side of the valley have a similar range of lithologies, and in localised lower-strain domains, show comagmatic textures and evidence of magma mingling between felsic and mafic intrusives. The hill to the north in the central part of the valley consists of biotite-rich quartzofeldspathic gneiss of the upper Stanovos Gneiss overlain by quartzite of the lower Stanovos Gneiss. This inversion of stratigraphy is interpreted to be a result of SW-directed thrusting, which possibly also caused repetition of the quartzofeldspathic units on each side of the valley.

4N. STOP 2.5. TRANSECT THROUGH UPPER STANOVOS GNEISS AND INDIANA WALLS GRANITE

The prominent ridge-forming unit of quartzofeldspathic gneiss in the Indiana Walls ridge (Fig. 18) crops out adjacent to the track heading to Indiana Station, towards the southern end of the range. The quartzofeldspathic gneiss is underlain by the lower Stanovos Gneiss, which consists of quartzite overlain by marble and calc-silicate. Detrital zircon data from quartzite of the lower Stanovos Gneiss has a markedly different age spectrum to that obtained from the upper Stanovos Gneiss. Although both units contain significant proportions of 1000-1200 Ma ages, the lower Stanovos Gneiss contains a large proportion of ages between ~1350 Ma and ~1900 Ma, which are poorly represented in the upper Stanovos Gneiss (Fig. 4). There were no grains found younger than ~1000 Ma in the lower Stanovos Gneiss, while the upper unit contains a small, but significant proportion of younger grains to ~630 Ma. The detrital zircon age spectrum of the lower Stanovos Gneiss is similar to those obtained from early to mid-Neoproterozoic sedimentary rocks in the Amadeus and Georgina basins, and it is possible that the difference in provenance between the upper and lower units of the Stanovos Gneiss reflects a significant difference in depositional age of the protoliths. If the lower Stanovos Gneiss is in fact Neoproterozoic in age, it is interesting to note that its quartzite/carbonate lithological association matches that of the Heavitree Quartzite/Bitter Springs units in the Amadeus Basin. This raises the possibility that these packages might be correlatives, with the lower Stanovos Gneiss representing sedimentation that took place in the former Centralian Superbasin before rifting commenced in the Harts Range region.

The lower Stanovos Gneiss is structurally overlain by biotite-rich, megacrystic quartzofeldspathic gneiss of the Indiana Walls Granite (Fig. 19). A sample from this locality yielded a SHRIMP U-Pb zircon age of 523 ± 4 Ma (Maidment, 2005), an age at least 40 million years older than metamorphism associated with the Larapinta Event. The homogeneity of the gneiss and the similarity of its age to other demonstrably intrusive granites within the Stanovos Gneiss are consistent with the interpretation that the protolith to the gneiss was intrusive rather than volcanic. Limited data for inherited zircon in the granite yields ages similar to those in the host Stanovos Gneiss. This fact, coupled with the observation that 520 Ma granite often has gradational contacts with adjacent migmatite suggests that the granite was derived from partial melting of the upper Stanovos Gneiss. The ~545 Ma age for the upper Stanovos Gneiss indicates that rapid burial of sediment took place during the Early Cambrian, coeval with mafic magmatism of the Harts Range Meta-Igneous Complex. Partial melting of the Harts Range Group appears to have taken place at the base of an extensional basin, coeval with deposition of the overlying Irindina Gneiss.

4O. STOP 3.1. INTERLAYERED MARBLE AND CALC-SILICATE ROCKS

This exposure (Fig. 20), which displays the possible protoliths to the host rocks of the Molyhil Mo-W skarn deposit, consists of calc-silicates, diopside marble and banded, andradite-pyroxene-scapolite marble.

4P. STOP 3.2. MOLYHIL MO-W SKARN

The Molyhil deposit (Fig. 21) is the most significant granite-related deposit in the Aileron Province. 20,000 tonnes of ore were mined from the current open cut between 1974 and 1976, yielding 100 tonnes of concentrate grading 70% WO_3 (Barraclough, 1979; Freeman, 1990; MoS_2 was not extracted). Thor Mining PLC, the current lease-holders, have defined a JORC-compliant mineral resource of 2.38 Mt grading 0.54% WO_3 and 0.26% MoS_2. This resource is split between the larger Southern and smaller Yacht Club orebodies, which are localised within metasedimentary roof pendants within granite texturally similar to the Marshall Granite (e.g. Freeman, 1990; Shaw et al., 1984).

These roof pendants are composed of variably chloritised biotite quartzo-feldspathic paragneiss and skarn, which has probably replaced dirty carbonate rocks (as indicated by chemical composition of the skarn assemblage). Although layering in both the gneiss and skarn, which most likely reflects bedding, generally strikes north-south, with sub-vertical dips (Fig. 21), much of the layering in the southwest corner of the open cut strikes east-southeast and dips moderately (~45°) to the south. These relationships may indicate south-southwest plunging folds, consistent with the regional structure described by Barraclough (1979).

The Marshall Granite at the Molyhil deposit is a medium grained biotite granite that has been chloritised and sericitised ("green" granite) along the southern and walls of the open cut, and K-feldspar altered ("pink" granite) along the northern and western wall of the open cut. Geochemical analyses indicate that relative to the "pink" granite, the green granite has gained Fe, Mg, Sb, Be and volatiles (e.g. H_2O), but lost Ca, Na, Sr, Cu and Pb.

Barraclough (1979) described three skarn assemblages at the Molyhil deposit: (1) granitoid endoskarn, (2) unmineralised (or "banded or "mixed" hornfels) calc-silicate exoskarn, and (3) "ore zone" calc-silicate exoskarn. Granitoid endoskarn occurs as veins and pods within the host granite and consists mostly of K-feldspar and hornblende with varying amounts of quartz, calcite, biotite, molybdenite, magnetite and scheelite.

Unmineralised skarn generally separates the mineralised calc-silicate from the granite. The banded variety contains alternating diopside-rich and garnet-rich bands. This rock type also contains accessory garnet, quartz, biotite and epidote, and locally contains minor magnetite, pyrite and molybdenite. The mixed variety comprises a mixture of garnet, pyroxene, epidote and calcite; it lacks opaque minerals (Barraclough, 1979).

Mineralised skarn is Fe-rich and generally dominated by magnetite. Other minerals present include pyrite, pyroxene, garnet, amphibole, scheelite, molybdenite, chalcopyrite and quartz (Barraclough, 1979; this study). Geochemically, the mineralised skarn is enriched in Fe, Cu, S, Sn and Y in addition to W and Mo. Relative to unmineralised skarn, mineralised skarn is depleted in K and F.

Site A. Calc-silicate rocks replacing the host unit to the Yacht Club orebody (Fig. 22). These banded rocks (after original bedding) consist of garnet-pyroxene skarn that replaced original "dirty" carbonate rock.

Site B. "Pink" granite in contact with calc-silicate rocks lateral to the Yacht Club orebody. This granite contains coarse-grained K-feldspar-rich segregations (endoskarn) and grades to the south at this site into chloritic "green" granite.

Site C. Very coarse-grained hornblende in a narrow zone along face. Higher up the face and in drill core, very coarse-grained K-feldspar and calcite are associated with the hornblende. In drill core, molybdenite and quartz are locally associated with the hornblende-K-feldspar assemblage. The hornblende yielded the ^{40}Ar-^{39}Ar age of ~1700 Ma (G. Fraser, pers. comm., 2004), which is taken as the age of mineralisation. This hornblende-K-feldspar-calcite assemblage is interpreted to be endoskarn.

Site D. Intrusive contact of "pink" granite into chloritic quartzo-feldspathic gneiss. The gneiss locally contains 0.5-2 mm magnetite porphyroblasts and appears to be localised along the western face of the open cut where it is extensively intruded by granite. Late quartz veins with hornblende-K-feldspar selvages cut the granite and the gneiss. Other veins at this site include carbonate and quartz-carbonate-fluorite veins.

Site E. "Pink" granite vein cutting calc-silicate rock.

Site F. Magnetite skarn boulders containing coarse-grained rosettes of molybdenite and coarse-grained scheelite. The coarse-grained character of the ore minerals has led to difficulties in estimating true ore grades. During production in the late 1970s, the ore grade was found to be up to 0.5% (combined WO_3 and MoS_2) higher than grades estimated from percussion drilling (Freeman, 1990). To overcome this problem, Thor Mining PLC drove three shafts into the Southern orebody and used analyses for these shafts to constrain the present JORC-compliant mineral resources.

Site G. "Green" chloritically altered biotite granite. This alteration facies dominates the southern and eastern walls of the open cut.

Site H. Boulder, possibly in-place, of banded, light green pyroxene-garnet skarn. Other boulders various calc-silicate assemblages, with one showing magnetite-molybdenite skarn against unmineralised skarn.

4Q. STOP 3.3. GARNETIFEROUS SAND DUNES

Low stabilised sand dunes at this location form part of a reserve (proven and probable) containing 2.3 Mt of garnet and 12 Mt of alumino-magnesio-hornblende as defined by Olympia Resources Ltd. In addition to the stabilised sand dunes present at this site, the reserves also include buried alluvial sands in the Plenty River floodplain and in channels of Aturga Creek, Plenty River and Ongeva Creek. Additional information regarding this project can be found at http://www.olympiaresources.com.au.

4R. STOP 3.4. RIDDOCK AMPHIBOLITE AT ATURGA CREEK

This water-washed outcrop is one of the best exposures of the Riddock Amphibolite, which forms part of the Harts Range Meta-Igneous Complex. It consists of garnetiferous para-amphibolite interlayered with garnet quartzo-feldspathic gneiss, biotite gneiss, garnet-biotite gneiss, sillimanite gneiss and plagioclase-rich gneiss (Hoatson and Stewart, 2001). More competant layers within the amphibolite have been boudinaged within a strong layer parallel foliation (Fig. 23). This foliation is folded by tight east-west trending folds (Fig. 24) which are typical of the deformational style in the northern Harts Range. This folding is interpreted to have taken place during N-S shortening associated with the Alice Springs Orogeny.

Claoué-Long and Hoatson (2005) report a maximum depositional age for this unit of 734 ± 44 Ma based on the analysis of zircon cores. Other volcaniclastic units within the Harts Range Meta-Igneous Complex have yielded detrital zircon spectra indicating deposition during the Early to Middle Cambrian (Buick et al., 1999; Maidment, 2005). Zircon rims have an age of 461 ± 6 Ma, which record high grade metamorphism associated with the Larapinta Event. The detrital zircon data for the Harts Range Meta-Igneous Complex suggest that the most likely chronostratigraphic correlatives to this unit are the carbonates and calc-silicates of the Early to Middle Cambrian succession of the Amadeus and Georgina basins, which was seen in Ross River Gorge on Day 1. Alluvial processes in Aturga Creek drainage catchment have produced sand highly enriched in garnet and hornblende. North of the ranges these alluvial sands form part of a potentially world-class, economic, heavy minerals sand deposit (see stop 3.3 above).

4S. STOP 3.5. OONAGALABI ZN-CU DEPOSIT

Studies by Hussey et al. (2005) indicate that the Oonagalabi deposit and several small prospects (Silver City and Silverado) to the southwest define a separate type of deposit, different in genesis to other base-metal prospects in the Strangways Metamorpic Complex. The host to this unit, the Bungintina Metamorphics (part of the Oonagalabi assemblage of the Ledan package), has an age of 1765 ± 4 Ma, at least 35 million years younger than the Ongenva package that hosts Utnalanama-type deposits at Edwards Creek (stop 4.4), Harry Creek, Coles Hill and Utnalanama.

The deposit is hosted by a metasomatised carbonate lens that can be traced at least 4 km along strike. This marble lens is within a sequence dominated by biotite-quartzofeldspathic gneiss with lesser concordant to discordant amphibolite (mafic sills and dykes) lenses. This sequence was interpreted by Hussey et al. (2005) to be dominated by pelitic and psammo-pelitic rocks, although the nearly unimodal population of zircon ages from these rocks indicates volcaniclastic derivation for the protolith. Hussey et al. (2005) subdivided the biotite quartzofeldspathic gneiss into two units (units A and B) based on the abundance of K-feldspar porphyroblasts.

Low grade Cu-Zn(Pb-Ag-Au) zones at Oonagalabi are hosted by marble and its metasomatised equivalents, massive orthoamphibole rock and diopside-rich calc-silicate rocks. All three of these rock types are characterised by low Al_2O_3 and Zr, confirming that the orthoamphibole and calc-silicate rocks formed by replacement of a marble protolith. Chalcopyrite, sphalerite and pyrrhotite, along with minor galena and pyrite, are disseminated mostly through the marble and orthoamphible rock at levels up to a few percent (Hussey et al., 2005). These minerals are commonly preserved in marble exposures at surface. Geochemical analyses that the pre-metamorphic alteration assemblage for the massive orthoamphibole rock was likely to have been talc-quartz, with variable amounts of carbonate.

Outside of the metasomatised marble lens, alteration zones are restricted, rarely extending more that 10 m from mineralised rock. Quartz-garnet rock is by far the dominant assemblage, with cordierite-bearing assemblages restricted to one small exposure. Geochemical analyses suggest that the protolith to the quartz-garnet rock was part of the psammo-pelitic assemblage that hosts the marble lens and that the pre-metamorphic assemblage was quartz-chlorite (Hussey et al., 2005).

Site A. The rock unit exposed at this site (unit B) structurally underlies the host rock to the Oonagalabi deposit and is characterised by large (to 30 mm) K-feldspar porphyroblasts that range in abundance from 1 up to 3%. These feldspar porphyroblasts are set in a fine-grained, gneissic quartz-K-feldspar-biotite±garnet matrix interpreted to have a pelitic protolith. Other, less abundant, rock types within unit B include mafic granulite (~20%), granite gneiss and garnet-rich calc-silicate rocks. Unit B grades into unit A by a decrease in the abundance of K-feldspar porphyroblasts.

Site B. In contrast to the previous site, the rock at this site lacks the characteristic K-feldspar porphyroblasts. This rock (unit A) comprises layered to homogeneous, fine-grained biotite-quartzofeldspathic gneiss. This rock unit appears to overlie and underlie the host unit to the Oonagalabi deposit, and contains significant lenses of mafic granulite.

Site C. Weakly mineralised diopside-rich calc-silicate rocks, exposed at this site, form a subordinate rock type to the mineralised unit at Oonagalabi. These rocks are interpreted to have replaced original marble and are themselves replaced by massive orthoamphibole rock.

Site D. Massive to foliated orthoamphibole rock exposed along this ridge is the main host to mineralisation at the Oonagalabi deposit. This rock is dominated by anthophyllite with lesser gedrite. It also contains minor amounts of chlorite and phlogopite as well as a number of trace minerals. Although sphalerite, chalcopyrite and pyrite are locally preserved at surface, evidence for mineralisation in this unit is mostly confined to sulphide boxwork and local malachite staining. The abundance of massive orthoamphibole rock distinguishes Oonagalabi from other base-metal prospects in the eastern Arunta.

Site E. At this site, poorly exposed quartz-garnet rock ("garnet quartzite") appears to grade into biotite-quartzofeldsapthic gneiss (unit A). The quartz-garnet rock is the main wall-rock alteration assemblage at the Oonagalabi deposit. This assemblage, and the lack of cordierite-bearing alteration assemblages, sets the Oonagalabi-type deposits apart from other base-metal deposits in the eastern Arunta area. Moreover, quartz-garnet rock at Oonagalabi is not as extensive as the alteration assemblages present at Utnalanama-type deposits.

Site F. Calc-silcate rock with pods of remnant forsterite marble. Elsewhere in the mineralised unit at Oonagalabi, marble contains nodules of calc-silicate (Fig. 26).

Site G. Mafic amphibolite exposed at this site is typical of this rock type throughout the Oonagalabi deposit. This rock type typically forms concordant to discordant bodies interpreted as sills. As these bodies are not altered, they are interpreted to have been emplaced after mineralisation.

Site H. Forsterite marble at this site is typical of variably mineralised marble at Oonagalabi. In addition to forsterite, this unit typically contains humite group minerals, spinel, phlogopite and clinopyroxene. In addition, this rock type can contain up to several percent sulphides, including sphalerite, chalcopyrite and pyrite.

4T. STOP 4.1. PINNACLES CU DISTRICT

The Pinnacles Cu district, which is located 2.3 km east of the Johnnies Reward prospect, was discovered in 1889 and produced a total of 248 tonnes of ore averaging 12.4% Cu when mining ceased in 1968 (Mackie, 2002). The district is hosted by marble of the upper Cadney Metamorphics. The visited prospect (Fig. 27) is hosted by interlayered phlogopite-diopside marble and scapolite marble. The mineralisation occurs in a shallowly dipping, brittle-style vein comprised of quartz and siderite(?) that is oriented sub-parallel to bedding within the host marbles. At depth the main Cu mineral is chalcopyrite, but most ore was extracted from the oxidised zone where malachite, chalcocite and bornite are the main ore minerals. In addition to Cu, these deposits also contain significant Au, Ag and Bi (Warren et al., 1974). We interpret these veins to have formed during the Alice Springs Event (380-300 Ma) based on the brittle character of the veins.

4U. STOP 4.2. UPPER CADNEY METAMORPHICS

This stop is within the marbles and calc-silicate rocks typical of the upper part of the Cadney Metamorphics, which we interpret as being near the base of the uppermost unit within the Strangways Metamorphic Complex. Although this unit has not been dated, we infer that it has an age of ~1800 Ma based on its (conformable?) relationship with the underlying units of the Strangways Metamorphic Complex which have been dated at ~1805 Ma at several locations (see stop 4.5 below). The stop is dominated by clinopyroxene-bearing and scapolite-bearing marble, with the clinopyroxene and scapolite forming in separate layers (probable beds). Minor sillimanite-garnet-biotite metapelitic schists are interlayered with the marbles, and calc-silicate rocks are commonly found along the contact between marble and metapelitic rocks.

4V. STOP 4.3. JOHNNIES REWARD PROSPECT

We interpret the Johnnies Reward prospect as an IOCG deposit, not a VHMS deposit as interpreted previously (e.g. Warren and Shaw, 1985). Johnnies-type deposits differ from Utnalanama-type VHMS deposits in terms of the host, metal assemblages and zonation, associated alteration assemblages and apparent age (Table 3). Other Johnnies-type deposits include Cu-Au deposits in the Jervois mining field and possibly the Gumtree prospect. This stop involves a traverse across the Johnnies Reward prospect from hanging wall into the footwall (Fig. 28) to illustrate characteristics of Johnnies-type deposits. The hanging wall to this deposit (site A) comprises migmatitic biotite quartzo-feldspathic gneiss (probable metasediments) with local bodies of mafic granulite that are interpreted as mafic sills.

The lode unit (site B) consists of an apparently stratiform body of diopside-tremolite-magnetite rock that extends about 200 m along strike and is up to 50 m wide (Chuck, 1984a,b). The presence of isolated, small bodies of forsterite marble at surface and in drill core suggests that the lode rock replaced a carbonate lens. At surface, tremolite is the main mineral other than magnetite, and quartz and malachite staining are locally present. The lode rock is characterised at depth by a Cu-Pb(Zn-Ag-Au) assemblage, with pyrite, chalcopyrite, galena and sphalerite being the main minerals. Unweathered lode rock is characterised by highly variable 100Zn/(Zn+Pb) ratios with an average of 30. The lode rock is characterised by an overall enrichment in Fe and Mg. From structural bottom to top the metal zonation appears to be: Au(Cu-Bi-S±Mo) → Cu-Pb-S(Zn-Ag-Au) → Pb-Mn(Cu-S±Ca) → REEs-HFSEs → Ca. The basal Au enrichment is present in the footwall rocks which can be seen at site C.

The footwall to the lode rock at Johnnies Reward (site C) is characterised by quartz-garnet-biotite-feldspar gneiss with minor magnetite, spinel and orthopyroxene. Chalcopyrite and pyrite are present at depth (Chuck, 1984a,b). The most significant Au grades (to 10 g/t) are present at depth within this zone, with one drill hole (E058-002) having a 50 m intersection grading 1.83 g/t that ended in grade. At surface these garnet-rich rocks have also returned significant Au grades. Further into the footwall (location D), the amounts of garnet decreases and feldspar increases.

4W. STOP 4.4. EDWARDS CREEK PROSPECT

The Edwards Creek prospect (Fig. 29) is probably the best exposed VHMS (Utnalanama-type) deposit in the Strangways Metamorphic Complex. The mineralised zone is hosted by a marble that occurs within a sequence that contains metamorphosed volcaniclastic rocks. This marble lens is up to 8 m thick and can be trace along strike for about 700 m. The main unit in this lens is forsterite marble with up to 5% disseminated pyrite, sphalerite, galena and chalcopyrite.

The marble is underlain by quartz-cordierite ± orthopyroxene ± phlogopite ± orthoamphibole ± sillimanite gneiss which we interpret to be metamorphosed quartz-chlorite altered volcaniclastic rocks. Other, less common, altered rocks include amphibole-pyroxene, massive cordierite and amphibole-spinel rocks. Whole rock $\delta^{18}O$ compositions of this and related altered rocks range between 1.8 and 5.8‰, which is most consistent with alteration by heated seawater.

A sample of quartz-cordierite-orthopyroxene-biotite rock yielded a simple zircon population with an age of 1802 ± 5 Ma, which is most consistent with a volcaniclastic origin for the rock. Analyses of host rocks for the Harry Creek (1801

± 3 Ma) and the Utnalanama (1810 ± 4 Ma) prospects are similar, suggesting that the rock packages that host Utnalanama-type deposits are correlative and have a probable volcaniclastic origin (Hussey et al., 2003).

The hanging wall is dominated by migmatitic metapelites, and granite gneiss that locally has an intrusive contact with the marble lens. We interpret this granitic rock to be an accumulation of partial melt that formed during peak metamorphism. Remnant lenses of biotite-rich, well-banded quartzo-feldspathic gneiss probably reflect the original rock that overlaid the marble lens. The protolith to this rock probably was sedimentary or volcaniclastic in origin. These remnant lenses have not been altered, indicating an asymmetric alteration pattern consistent with VHMS deposits. A series of mafic granulite lenses, which we interpret as mafic sills, are also present in the hanging wall. The distribution of the mafic sills and the marble lens suggests that the Edwards Creek deposit is located within a syncline. Figure 29 shows the locations of sites visited at this stop.

Site A. The variability in the quartz-cordierite gneiss that characterises the footwall alteration zone is apparent at this site. A sample from this site yielded a SHRIMP U-Pb zircon age of 1802 ± 5 Ma with a single igneous zircon population. As the single population and igneous character of the zircons imply a probable volcaniclastic origin for the population (Hussey et al., 2005), and this age is taken as the true age of the host sequence to the deposit. This age was used to pin the Pb isotope evolution model for the Strangways Metamorphic Complex. Zircon overgrowths from this sample returned an age of 1716 ± 3 Ma (Hussey et al., 2005), implying that these rocks have been overprinted by high-grade, Strangways-aged metamorphism.

Site B. This site contains a variety of rock types, the most unusual being a quartz-cordierite-orthoamphibole rock with a relic coarse clastic texture. Gedrite-hercynite-magnetite-phlogopite rock nearby is inferred to be metamorphosed chloritite that either replaced the host or formed as veins.

Site C. Calc-silicates (quartz-diopside-anorthite rock) at this and other locations have replaced marble. In addition to being weakly mineralised in Pb (200-1000 ppm) and Zn (100-1900 ppm), these rocks are also enriched in Bi (to 18 ppm), Sn (to 100 ppm), U (to 100 ppm), Th (to 70 ppm), Y (to 420 ppm) and REEs (totals to 1100 ppm). We interpret this mineralising event to be a separate event from the VHMS event based on this elemental assemblage, which contrasts with that associated with mineralised marble (see below). This event may relate to the Yambah Igneous Event (~1780 Ma).

Site D. The banded, biotite-rich quartzo-feldspathic gneiss at this site is interpreted to be the unit that overlies the mineralised zone. The protolith to this rock was most likely a psammopelitic or volcaniclastic rock. The more massive granitic gneiss is interpreted as a melt generated during high grade metamorphism. Mafic granulite bodies are interpreted as mafic sills.

Site E. Malachite staining is present at the base of the siliceous cap exposed at this site. This siliceous cap is interpreted as a Tertiary weathering feature. Drilling of an induced polarisation anomaly to the north of this site yielded highly gossanous, vuggy material that assayed up to 7.1% Cu and 3.3% Zn. Surface samples of the siliceous rock all contain significant mineralisation, with maximum grades of 0.6% Pb, 1.3% Zn and 0.6% Cu. Along the base of the siliceous ridge to the south, coarse-grained amphibole±spinel rock is present.

Site F. The banded, forsterite marble at this site is typical of the mineralised zone at Edwards Creek. The marble contains up to several percent disseminated and vein galena, sphalerite, pyrite and chalcopyrite. Our analyses of the marble indicate grades up to 0.67% Pb, 3.29% Zn and 0.80% Cu, and analyses of marble in diamond drill core indicate grades up to 0.57% Pb, 1.00% Zn and 0.55% Cu. Other anomalous elements include Cd (to 137 ppm) and Bi (to 20 ppm). The Sn, HFSE and REE anomalies characteristic of mineralised calc-silicate rocks are not present in mineralised marble, which is consistent with two separate mineralising pulses.

Site G. At this site the replacement of marble by siliceous rock is well illustrated. The fabric in the marble is a granulite facies Strangways fabric and silicification clearly truncates the fabric.

Site H. The poorly outcropping black amphibolite between the marble lens and the footwall quartz-cordierite gneiss at this site assayed 11.68% Zn (much to the surprise and consternation of GA's geochemistry lab).

Subsequent investigation of this rock indicated that it is comprised almost entirely of hornblende and gahnite. This rock can be traced for about 150 m along the contact between the mineralised marble lens and the quartz-cordierite gneiss, and it varies in width between 0.5 and 5 m. In addition to Zn this rock is enriched in Bi (to 0.48%), Cu (to 0.26%), Pb (to 0.21%), Sn (to 70 ppm), Th (to 40 ppm) and total REEs (~500 ppm). The sulphur values are exceptionally low, with a maximum of only 400 ppm. The origin of this rock is problematic; it may represent original VHMS-related mineralisation that was subsequently metasomatised during the later Yambah or Strangways events.

Site I. At this site an Alice Springs-aged (385 Ma age from monazite intergrown with biotite-chlorite: Ballèvre et al., 1999) shear has retrogressed the quartz-cordierite rock to a chlorite-biotite-staurolite-kyanite schist.

Site J. The massive, coarse-grained gedrite-clinoproxene rock at this site occurs mainly in the southern part of the Edwards Creek deposit. Locally the rock has been retrogressed to a talc-chlorite assemblage. We interpret that it occurs in the footwall to the mineralised marble lens and represents a metamorphosed quartz-talc alteration assemblage.

4X. STOP 5.1. NOLANS BORE DEPOSIT

The Nolans Bore REE-P deposit, which is located in the southeastern part of the Reynolds Range (Fig. 30), contains a mineral resource of 18.6 million tonnes at 3.1% REE oxides, 14% P_2O_5 and 0.47 lb/tonne U_3O_8 (Goulevitch, 2006). The deposit consists of a series of REE-bearing fluoro-apatite veins that are hosted mostly by granite gneiss (Fig. 31). The prospect area is dominated by the Lander Rock beds, which locally comprise schist, phyllite, andalusite hornfels, garnet-cordierite-biotite-quartz granofels, sillimanite-biotite-cordierite-orthoclase granofels and tourmaline-bearing quartzite (Stewart, 1981). These rocks have been intruded by granites (now granite gneisses) that have been correlated with the Boothby Orthogneiss and the Napperby Gneiss, the latter of which has been a SHRIMP U-Pb zircon age of 1778 ± 8 Ma (Collins and Williams, 1995). Much of the prospect is covered by alluvial sand and gravel, up to 4.5 m in thickness (Goulevitch, 2005).

The deposit consists mostly of a series of east-northeast trending and steeply (65-90°) north dipping fluorapatite veins or dykes that are emplaced in strongly kaolinised rock with a granitic gneiss protolith. The veins very in thickness from 0.3 to 25 m and are concentrated in two zones, the north zone and the south zone. The north zone is at least 900 metres long and up to 350 metres wide, whereas the south zone is up to 700 metres wide and at least 1000 metres long (Goulevitch, 2006). The established mineral resource is defined mainly in the north zone.

Stephens (2005) has inferred that both the north and south mineralised zones have a north-northeast-trending sigmoidal shape which is consistent with emplacement in dilatant zone during the development of sinistral west-northwest-trending shears (Fig. 31). Similarly oriented shears in the area (Fig. 30) have been dated at ~335 Ma (Cartwright et al., 1999), raising the possibility that the apatite veins were emplaced during the late stage of the Alice Springs Orogeny. However, the geometric interpretation of Stephens (2005) is speculative and requires further testing. J Goulevitch (pers. comm., 2005) suggests that the apatite veins formed within a swarm of dilatent fractures.

Four styles of REE mineralisation have been recognised at Nolans Bore: (1) massive fluorapatite veins that typically contain 4-6% REE oxides and constitute most of the defined resource, (2) very high-grade (7-10% REE oxide) zones found in apatite-poor kaolinitic zones outside of the veins, (3) apatite-allanite-epidote zones hosted by calc-silicates, and (4) low-grade stockwork zones in gneiss and kaolinised rock adjacent to the veins and mylonite zones (Goulevitch, 2005). The main difference between the ore types is the REE/P ratio, which is significantly higher in the high-grade cheralite-dominated zones outside of the main fluorapatite veins. In the veins, the apatite is generally fine-grained, although local zones are coarse-grained with grains up to 15 mm. Locally the apatite has been brecciated, with coarse clasts (to 15 mm) in a fine-grained matrix. Goulevitch (2005) interprets that these clastic textures indicate local post depositional brecciation, but note that the apatite veins do not display a shear fabric.

Mineralogically, most of the REE are hosted by cheralite, a phosphate deficient REE mineral. Only about 30-35% of the REE is hosted in the crystal structure of apatite. Texturally, cheralite fills microfractures and microveins within the apatite (Goulevitch, 2005). Arafura Resources geologists will lead the excursion to the Nolans Bore deposit, commencing at the Aileron Roadhouse.

5. ACKNOWLEDGMENTS

This field guide is the result of research collaboration between Geoscience Australia, the Northern Territory Geological Survey, the Research School of Earth Sciences, Australian National University, and the Department of Geology, Adelaide University. We thank Max Frater, Ian Scrimgeour (NTGS), Ian Williams (ANU) and Martin Hand (Adelaide University) for discussions and input during the research program. Anthony Budd, David Champion (Geoscience Australia) and Ian Scrimgeour are thanked for their comments on an early draft of this excursion guide. Arafura Resources NL are thanked for providing both access and a guide for the Nolans Bore deposit. We also thank the pastoral landholders and exploration leaseholders for access to ground.

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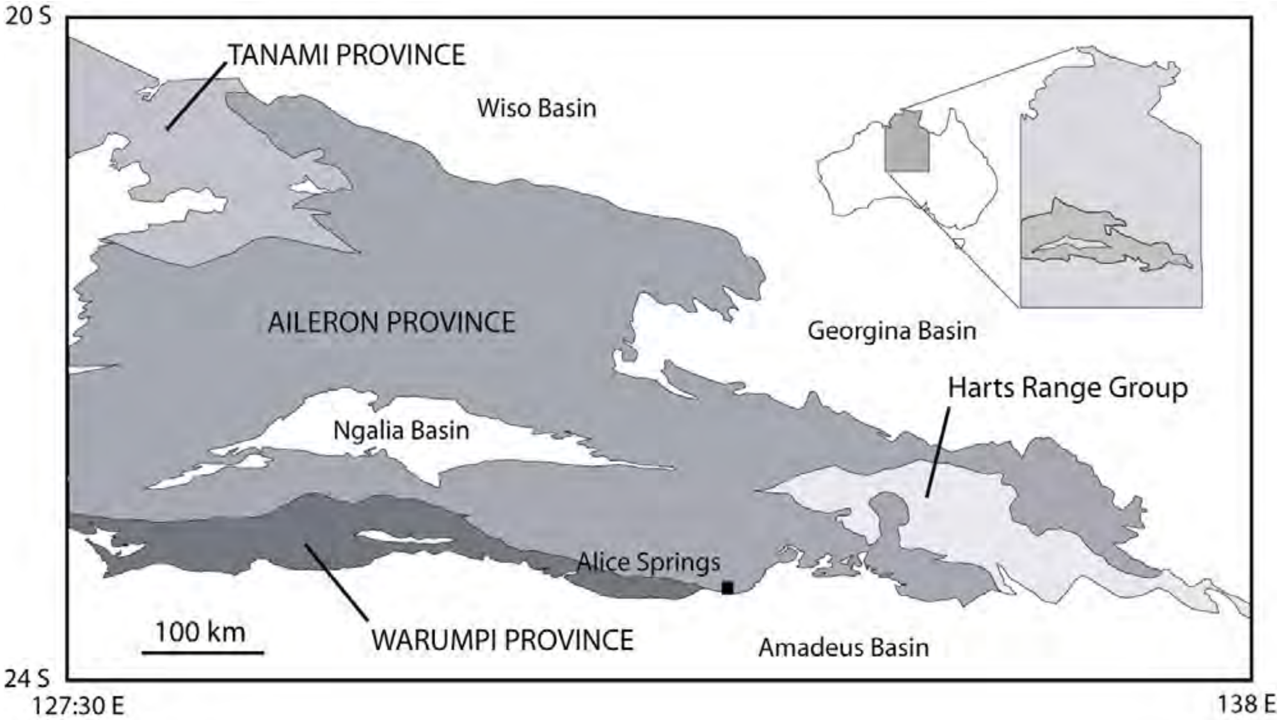

Figure 1. Major crustal elements of the Arunta region.

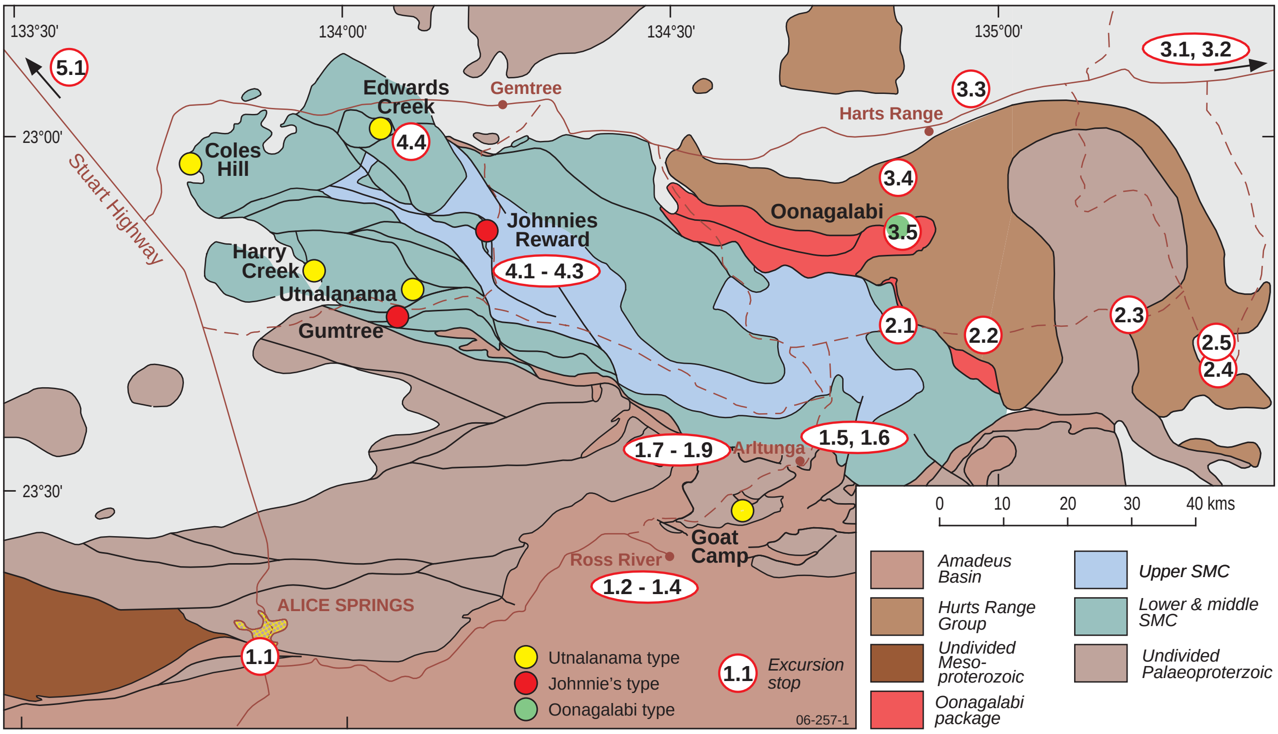

Figure 2. Generalised geology of the eastern Arunta region showing the location of excursion stops.

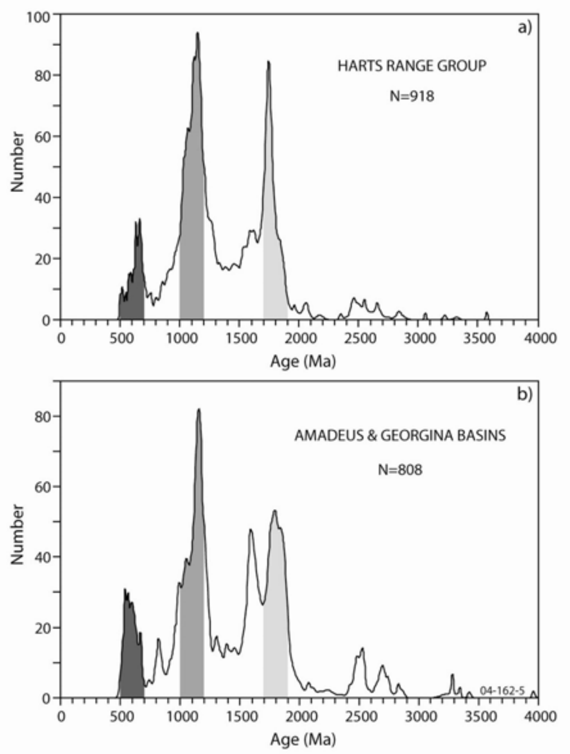

Figure 3. Combined detrital zircon data for (a) metasediments of the Harts Range Group and (b) Neoproterozoic to Early Ordovician sedimentary rocks of the southern Georgina and northeastern Amadeus basins, showing the close similarity in provenance between the two sequences. Data from Maidment (2005), Buick et al. (2001, 2005) and Zhao et al. (1992).

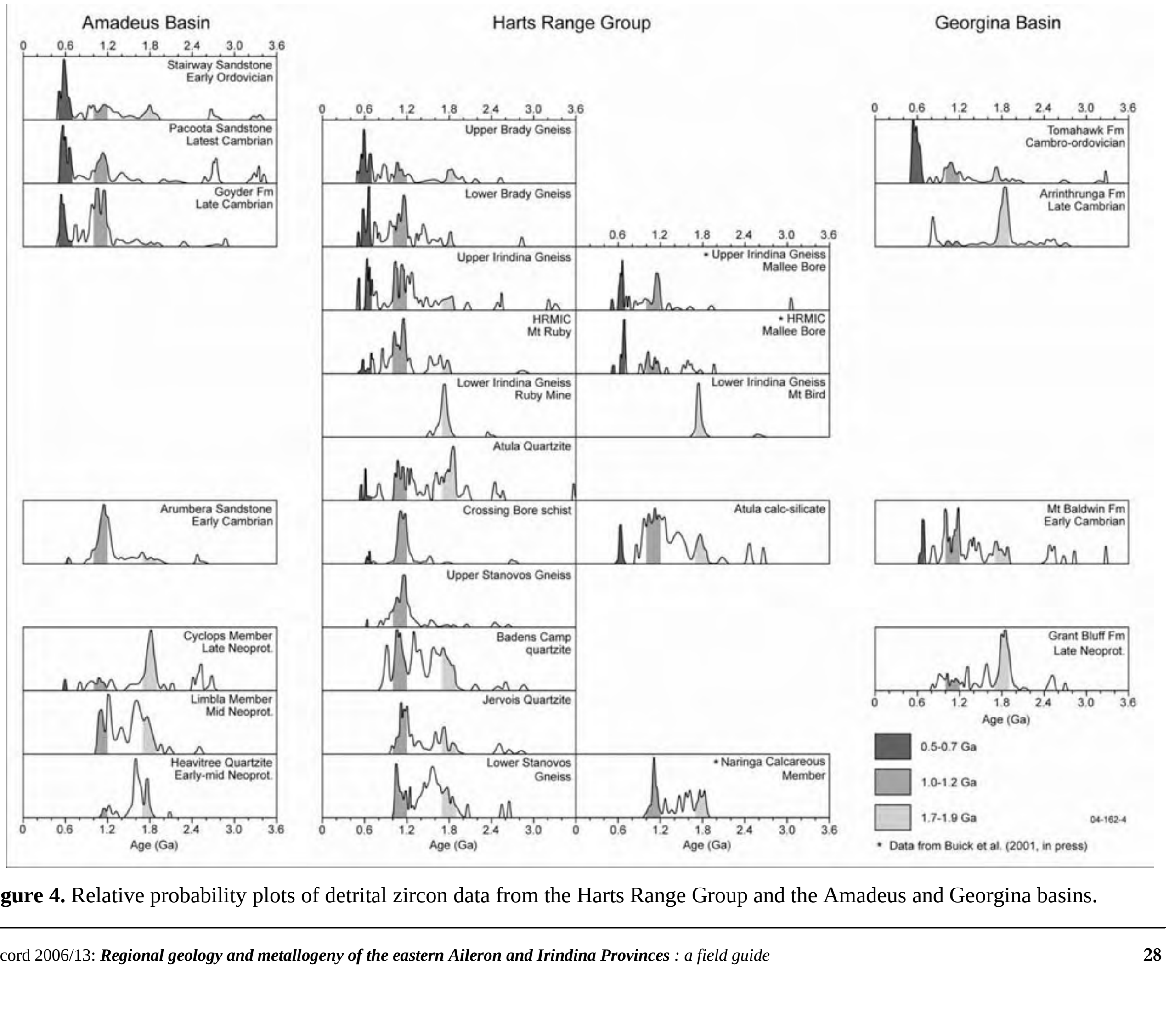

Figure 4. Relative probability plots of detrital zircon data from the Harts Range Group and the Amadeus and Georgina basins.

Figure 5. The early Neoproterozoic Heavitree Quartzite unconformably overlying Palaeoproterozoic Sadadeen Gneiss at Heavitree Gap.

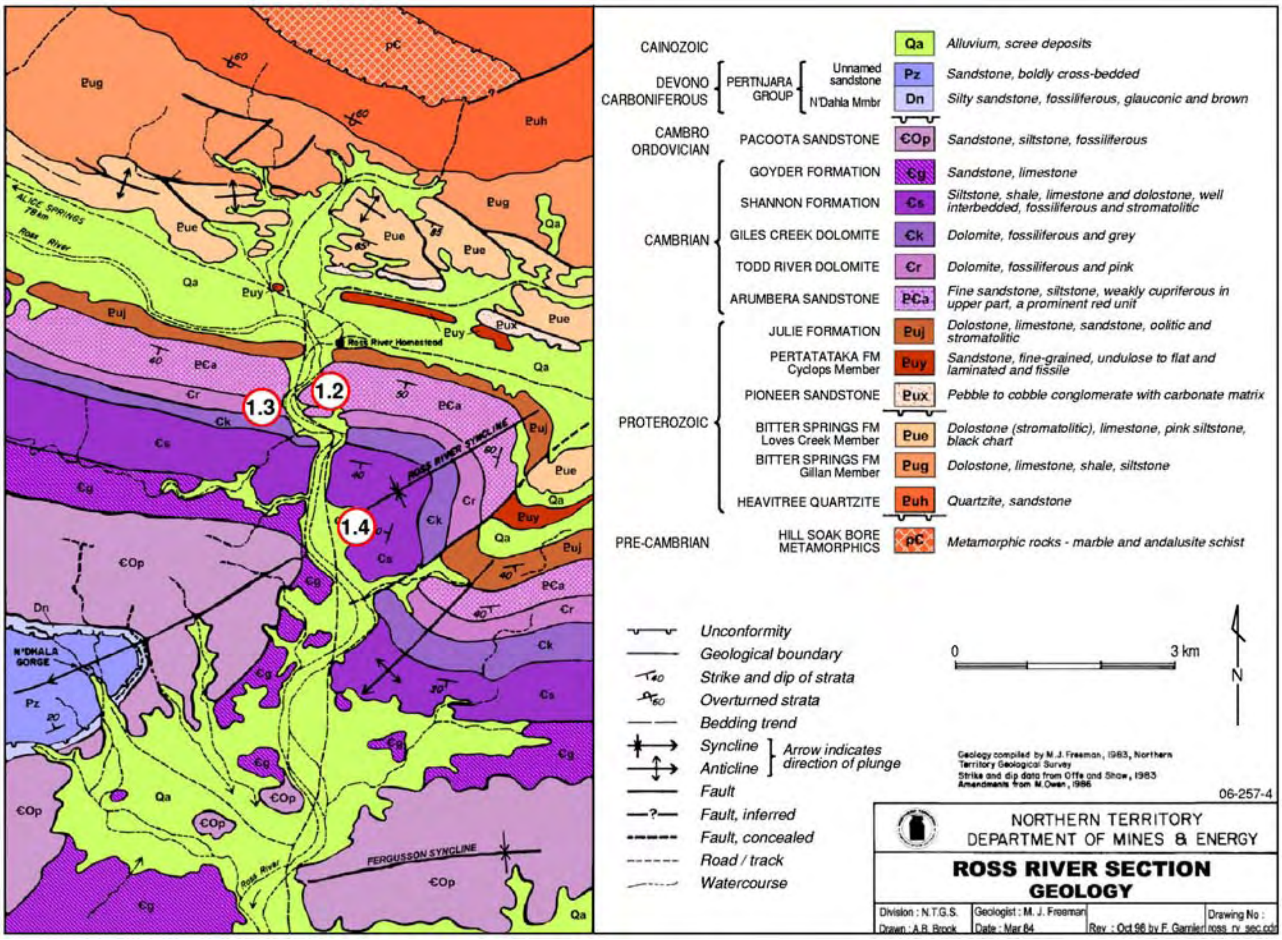

Figure 6. Geology of the Ross River Gorge showing the location of excursion stops 1.3 and 1.4 (modified after Freeman et al., 1987).

Figure 7. The latest Neoproterozoic to Early Cambrian Arumbera Sandstone at Ross River Gorge.

Figure 8. The Cambro-Ordovician Pacoota Sandstone in the core of the Ross River Syncline.

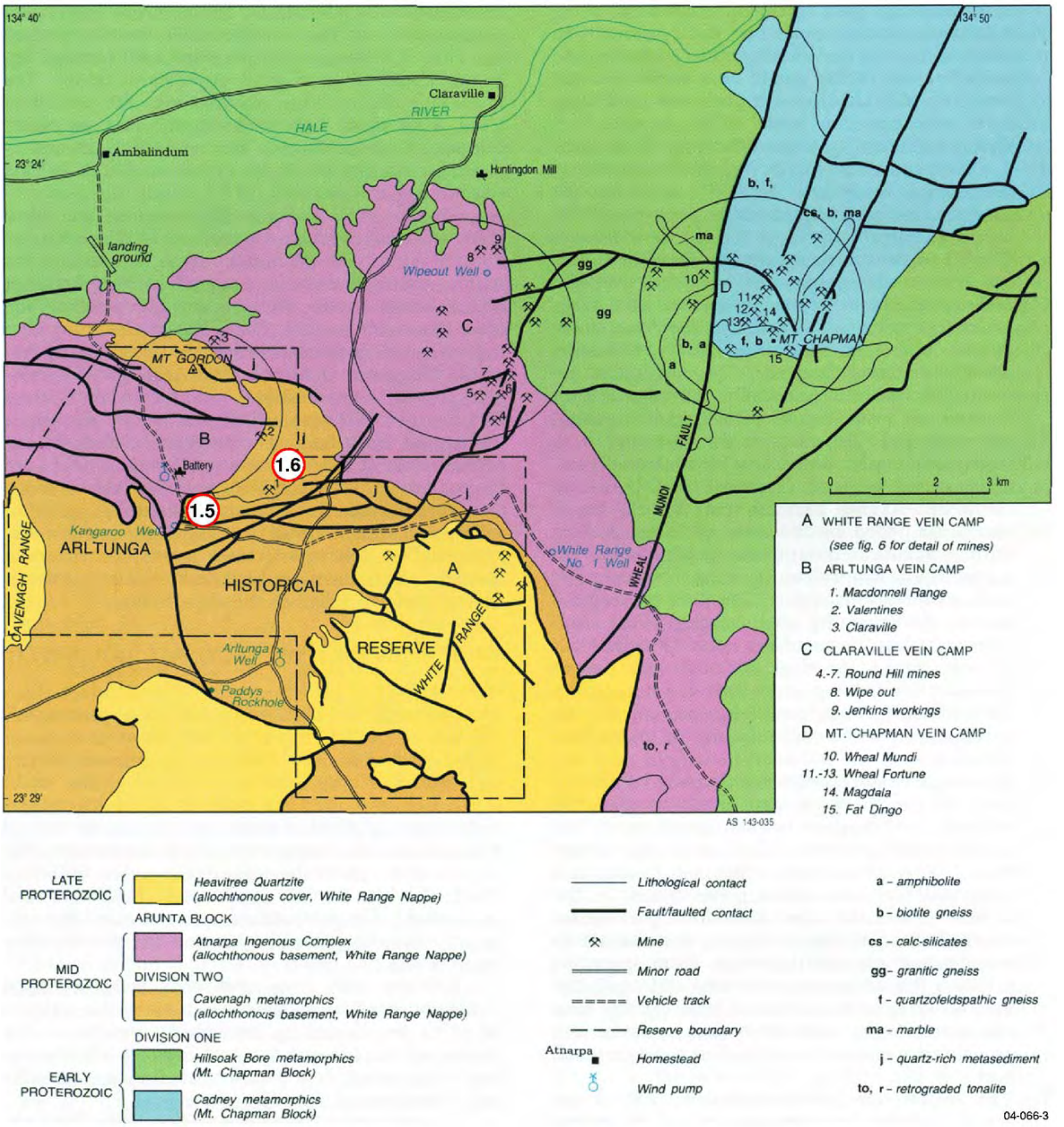

Figure 9. Geology of the Arltunga area showing the locations of excursion stops 1.5 and 1.6 (modified after Mackie, 1986).

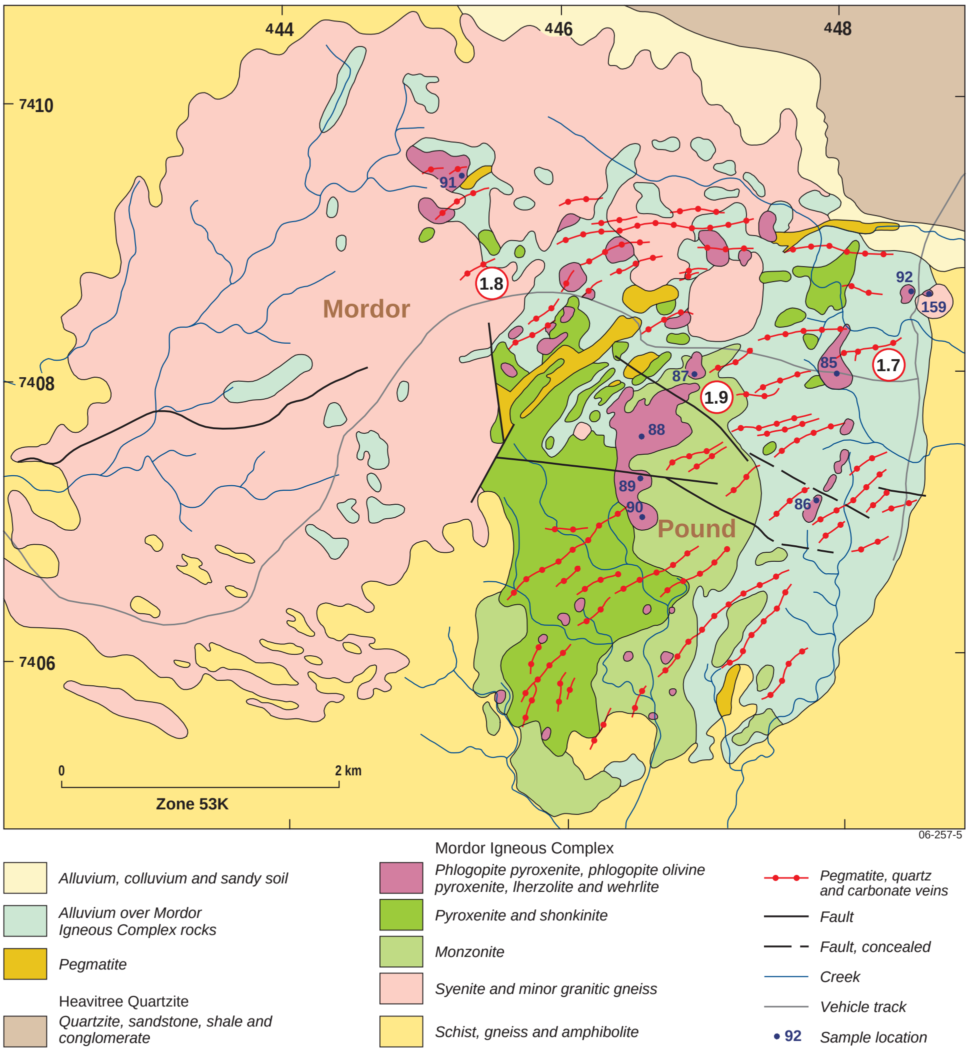

Figure 10. Geology of the Mordor Igneous Complex showing the locations of excursion stops 1.7 and 1.8 (modified after Hoatson and Stewart, 2001).

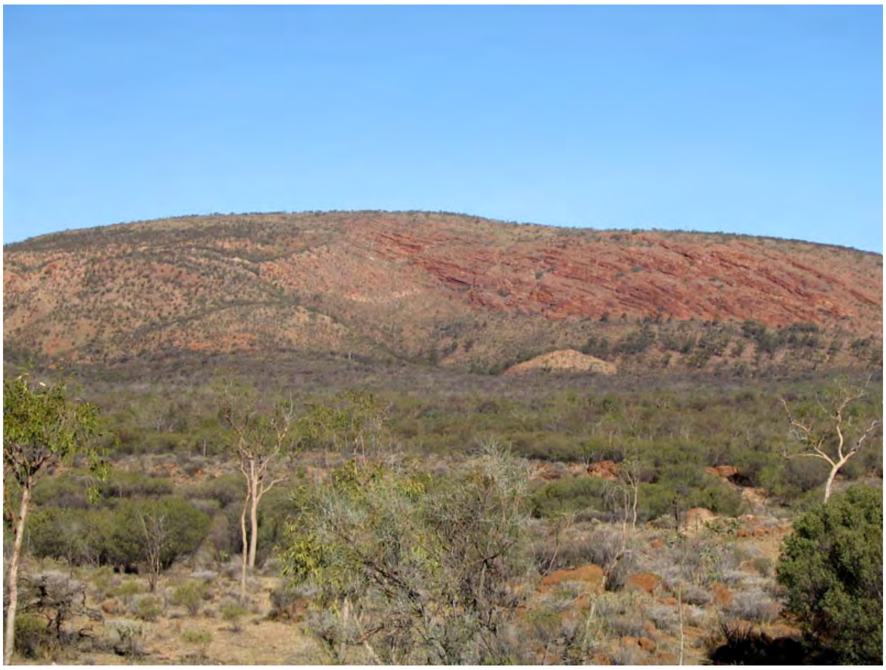

Figure 11. Large-scale recumbent folding of the Heavitree Quartzite, which was thrust over Mesoproterozoic rocks of the Mordor Igneous Complex during the Alice Springs Orogeny.

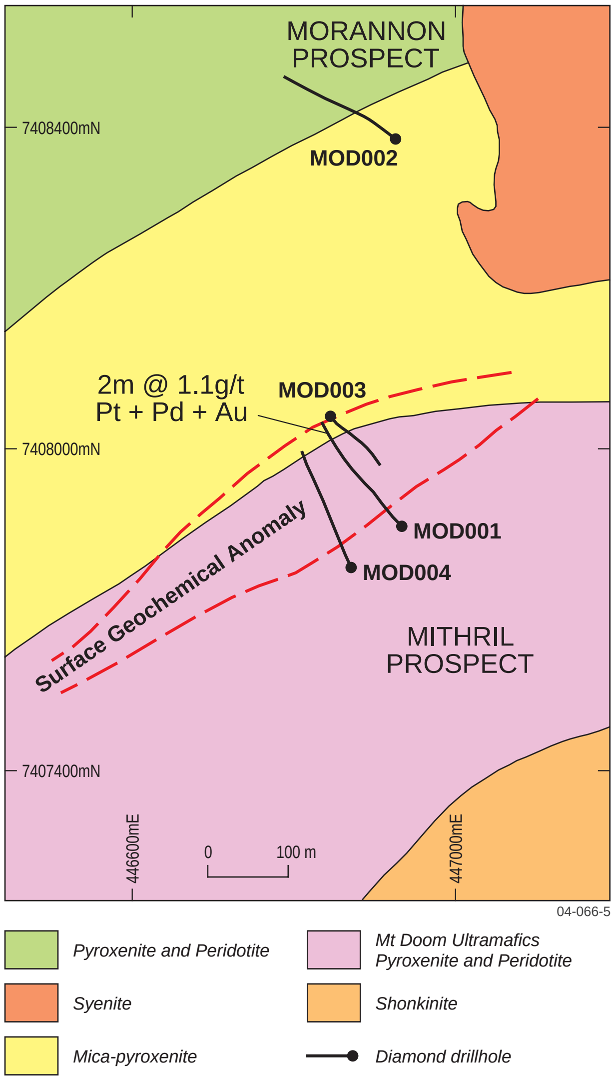

Figure 12. Surface geology of the Mithril prospect (stop 1.8: modified after Tanami Gold Ltd announcement to the Australian Stock Exchange, 20 December 2002).

Figure 13. Mylonitic mafic and felsic gneiss of the Strangways Metamorphic Complex in the Florence-Muller Shear Zone at Florence Creek.

Figure 14. Metapelite, marble and calc-silicate rock of the Irindina Gneiss in Lizzy Creek.

Figure 15. Shallowly dipping quartzofeldspathic gneiss of the Entia Gneiss Complex (foreground) structurally overlain by the megacrystic Bruna Gneiss (background).

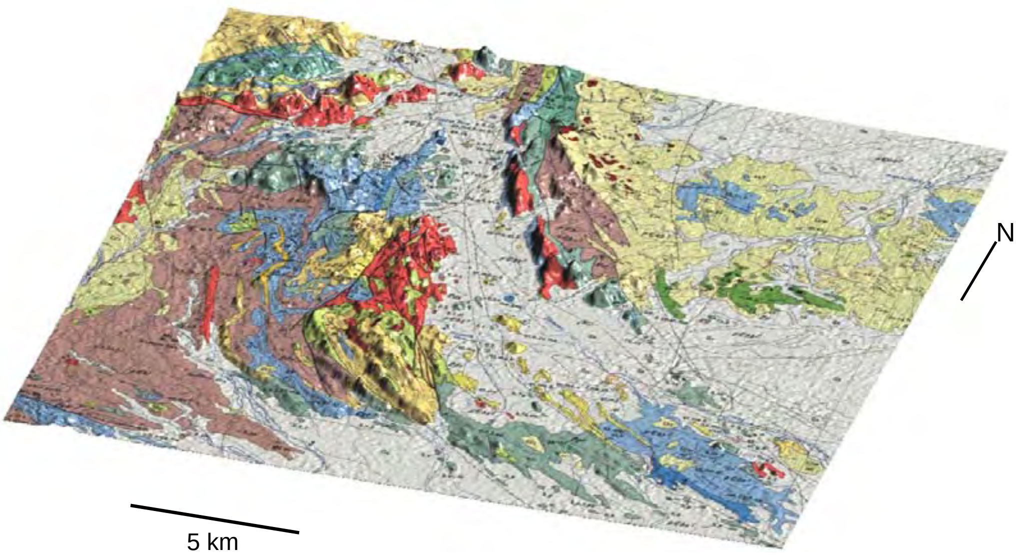

Figure 16. Perspective view of the Stanovos Valley in the southeastern Harts Range (1:100 000 scale geology draped over digital elevation model). Red and yellow ridge-forming units on either side of the valley are megacrystic gneiss and equigranular quartzofeldspathic gneiss respectively. Geology from Shaw et al. (1990), digital elevation data from Geoscience Australia.

Figure 17. View to the NNW from stop 2.4 along the Stanovos Valley.

Figure 18. Indiana Walls, showing the prominent ridge-forming unit of the Indiana Walls Granite.

Figure 19. Megacrystic biotite-rich quartzofeldspathic gneiss of the 524 ± 4 Ma Indiana Walls Granite.

Figure 20. Interlayered marble and calc-silicate rock that is probably similar to the sequence that hosts the Molyhil deposit.

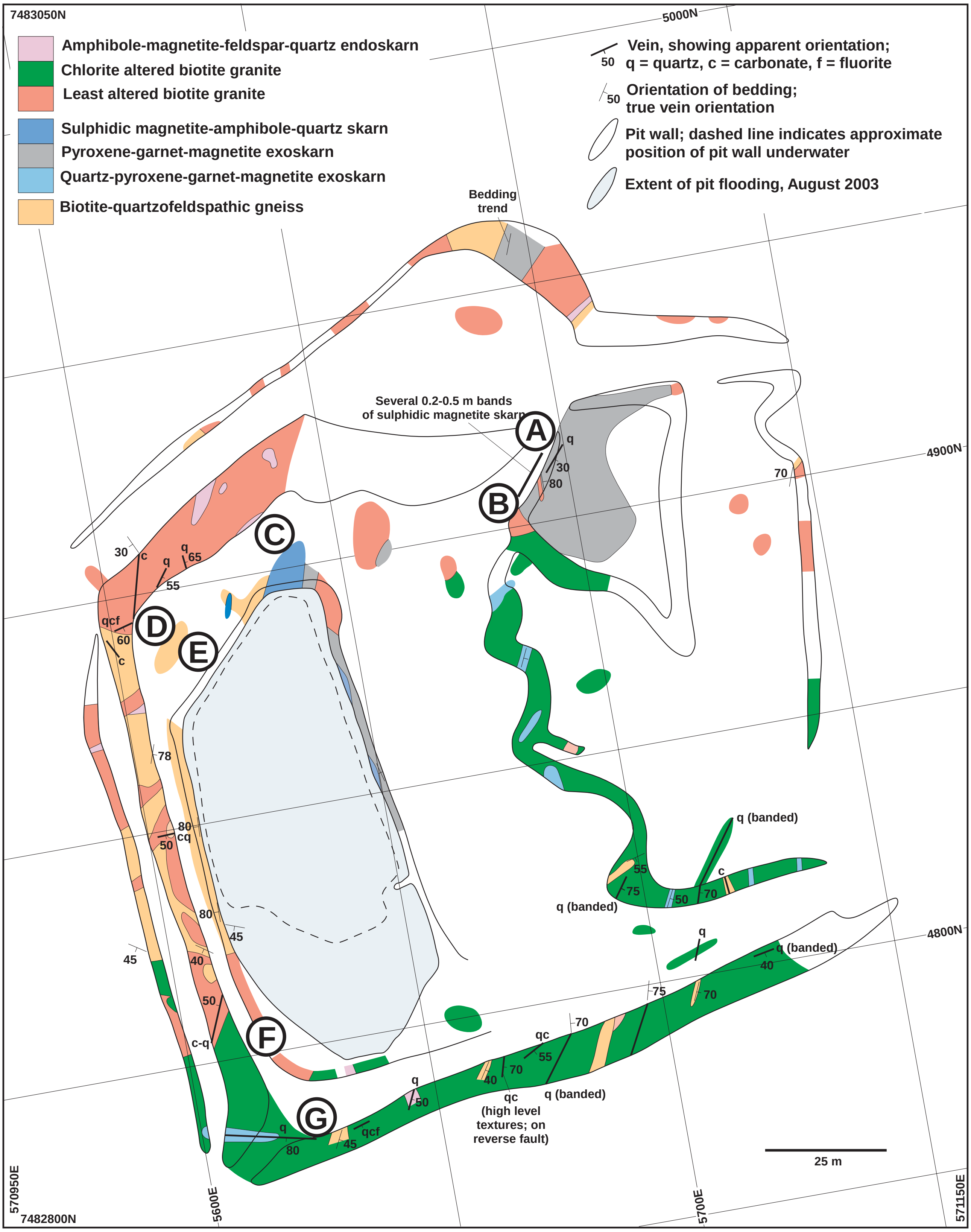

Figure 21. Geological pit map of the Molyhil prospect showing locations of sites visited at stop 3.2.

Figure 22. Banded calc-silicate rocks that form host unit to Yacht Club orebody at depth.

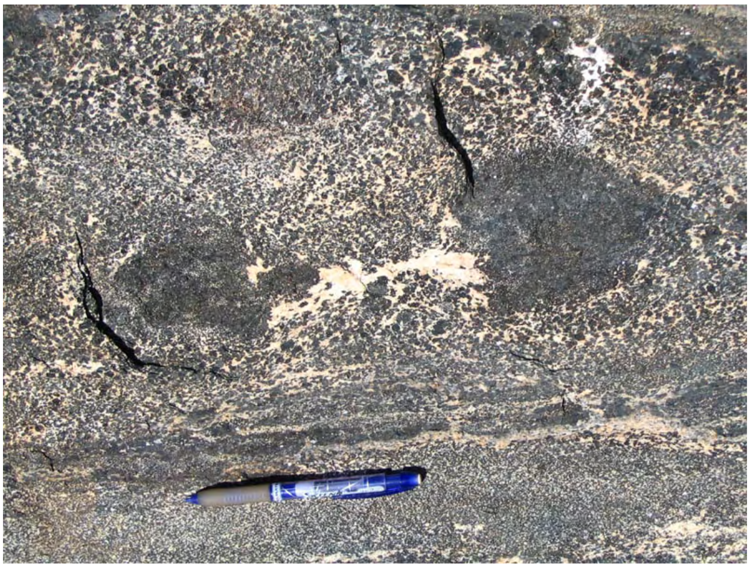

Figure 23. Boudinaged amphibolite layer within the Riddock Aphibolite, Aturga Creek.

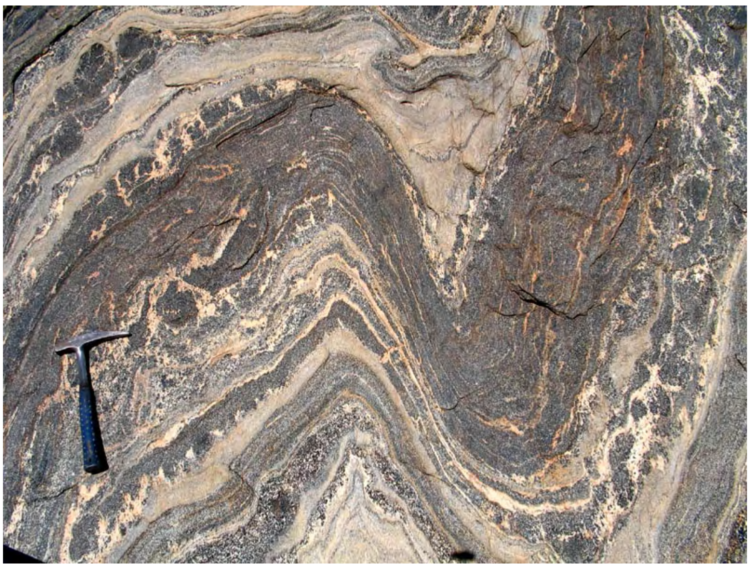

Figure 24. East-west trending folding of the Riddock Amphibolite, Aturga Creek.

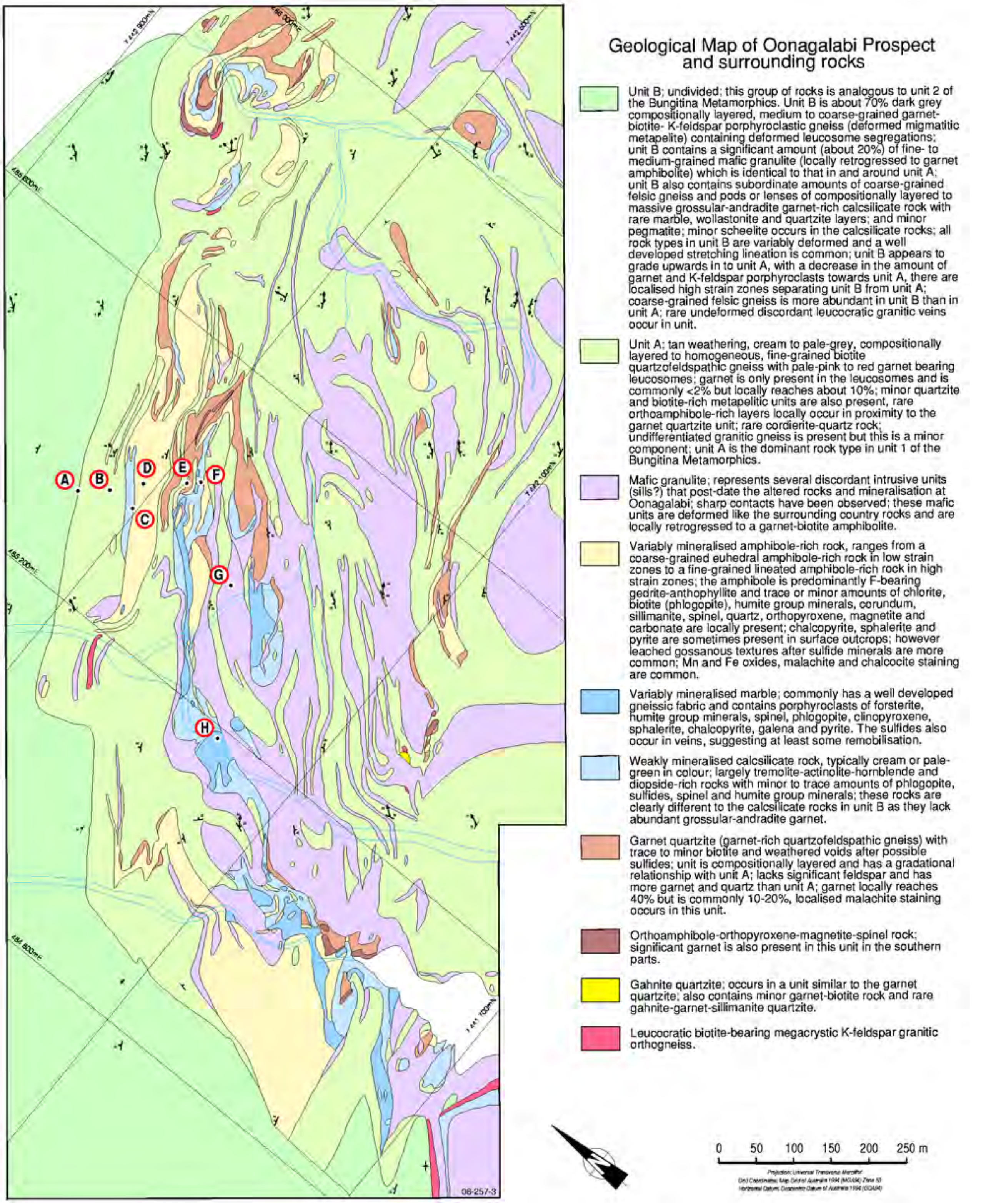

Figure 25. Geology of the Oonagalabi deposit showing the location of sites visited at stop 3.5 (modified after Hussey et al., 2005).

Figure 26. Calc-silicate nodules in marble, Oonagalabi prospect.

Figure 27. Mine workings at the Pinnacles Cu district.

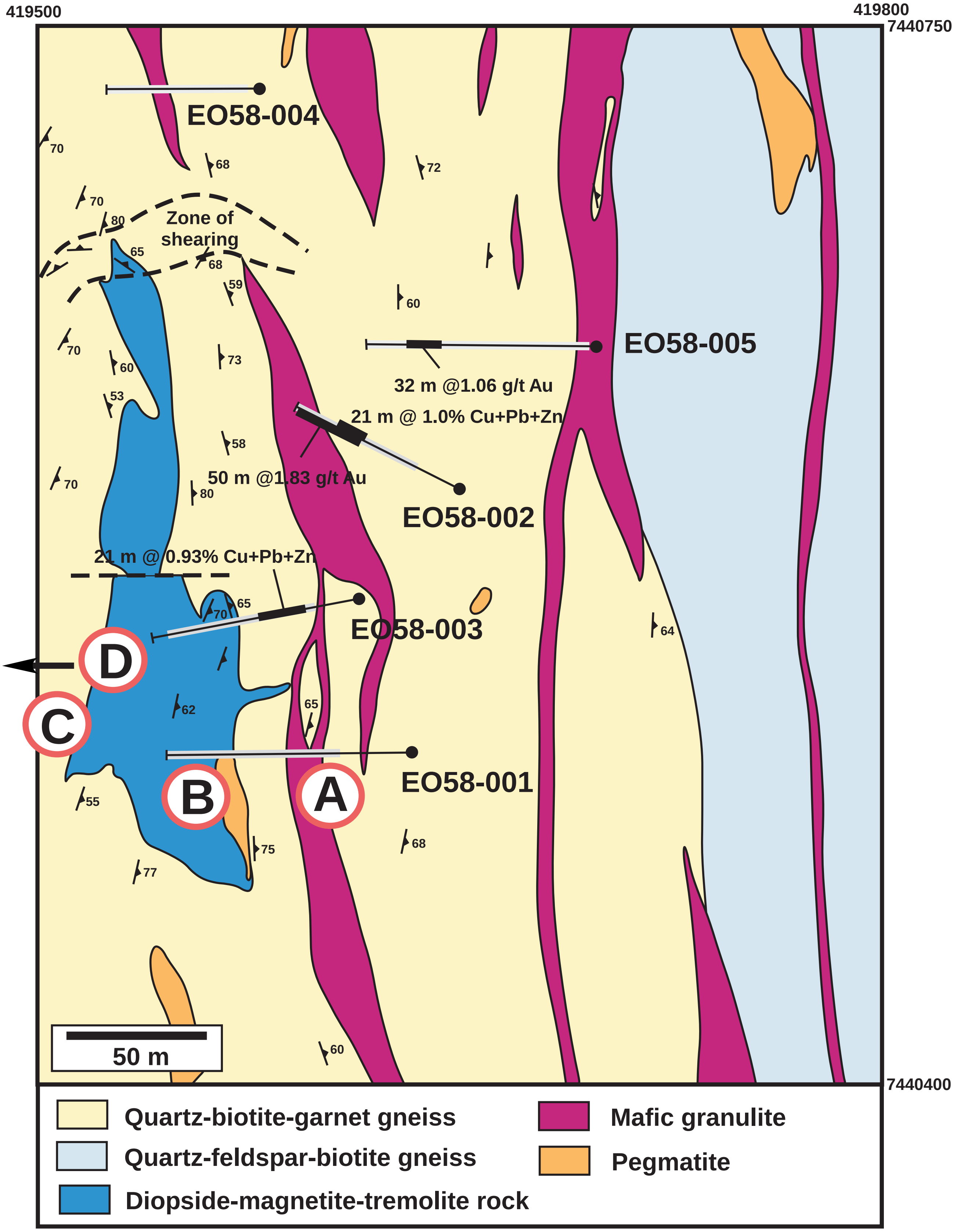

Figure 28. Geology of the Johnnie's Reward prospect showing sites from the traverse for stop 4.3 (modified after Chuck, 1984b).

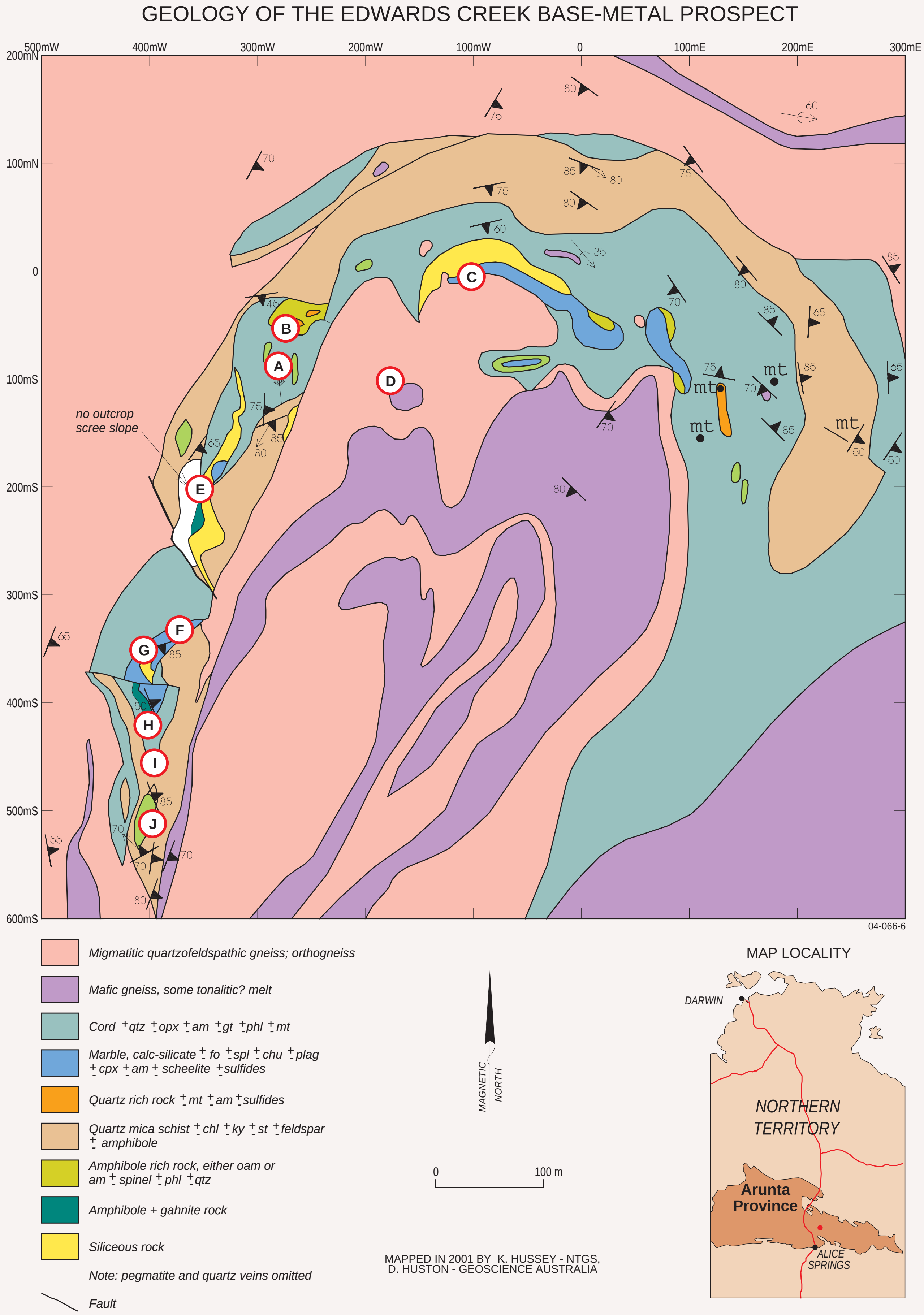

Figure 29. Geology of the Edwards Creek deposit, showing sites from the traverse at stop 4.4.

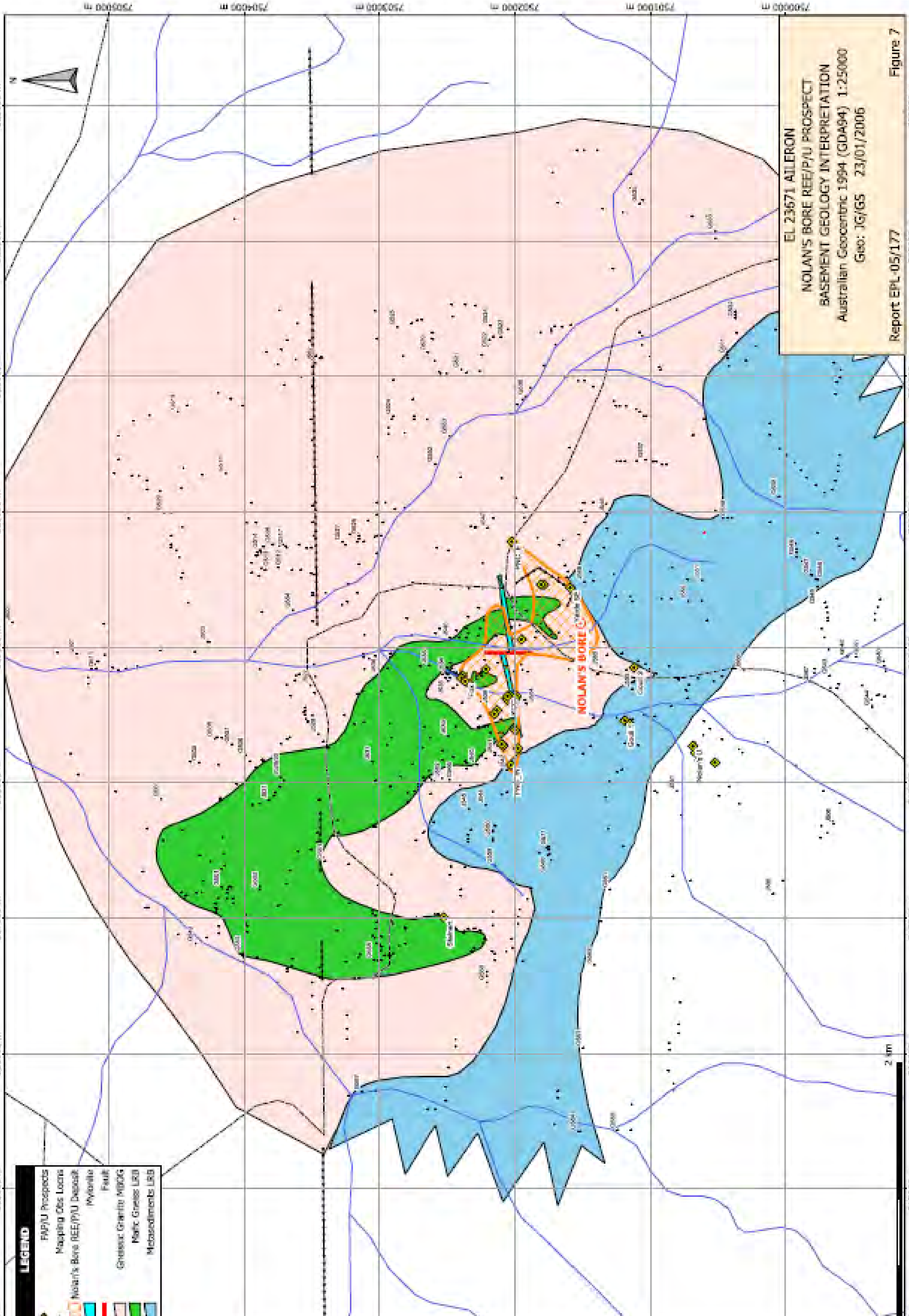

Figure 30. Geology of the souteastern Reynolds Range (courtesy Arafura Resources NL)

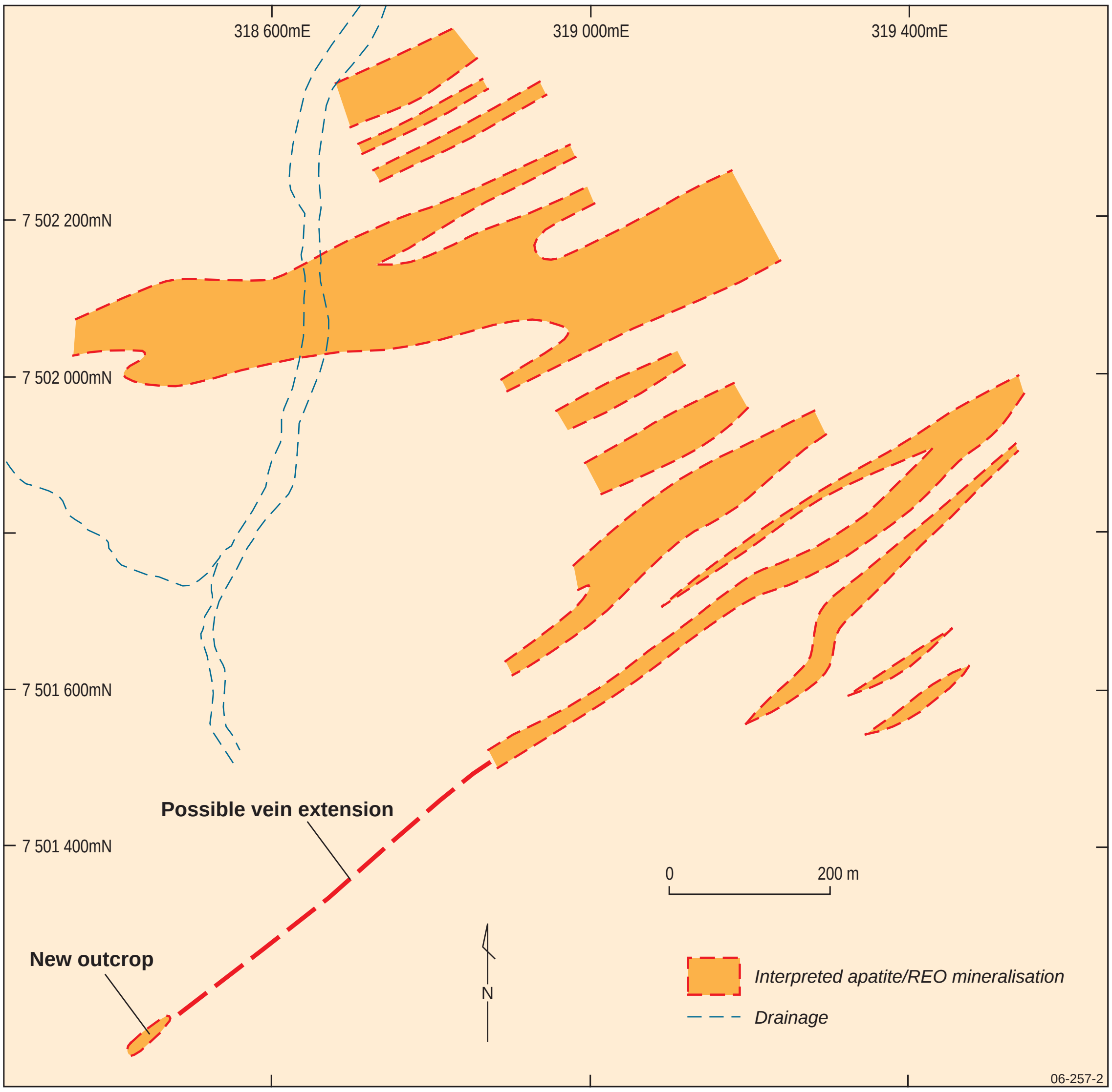

Figure 31. Geology of the Nolans Bore deposit (courtesy Arafura Resources NL).