EXPLANATORY
NOTES

GOVERNMENT OF
WESTERN AUSTRALIA

LISSADELL
1:250 000 SHEET
WESTERN AUSTRALIA

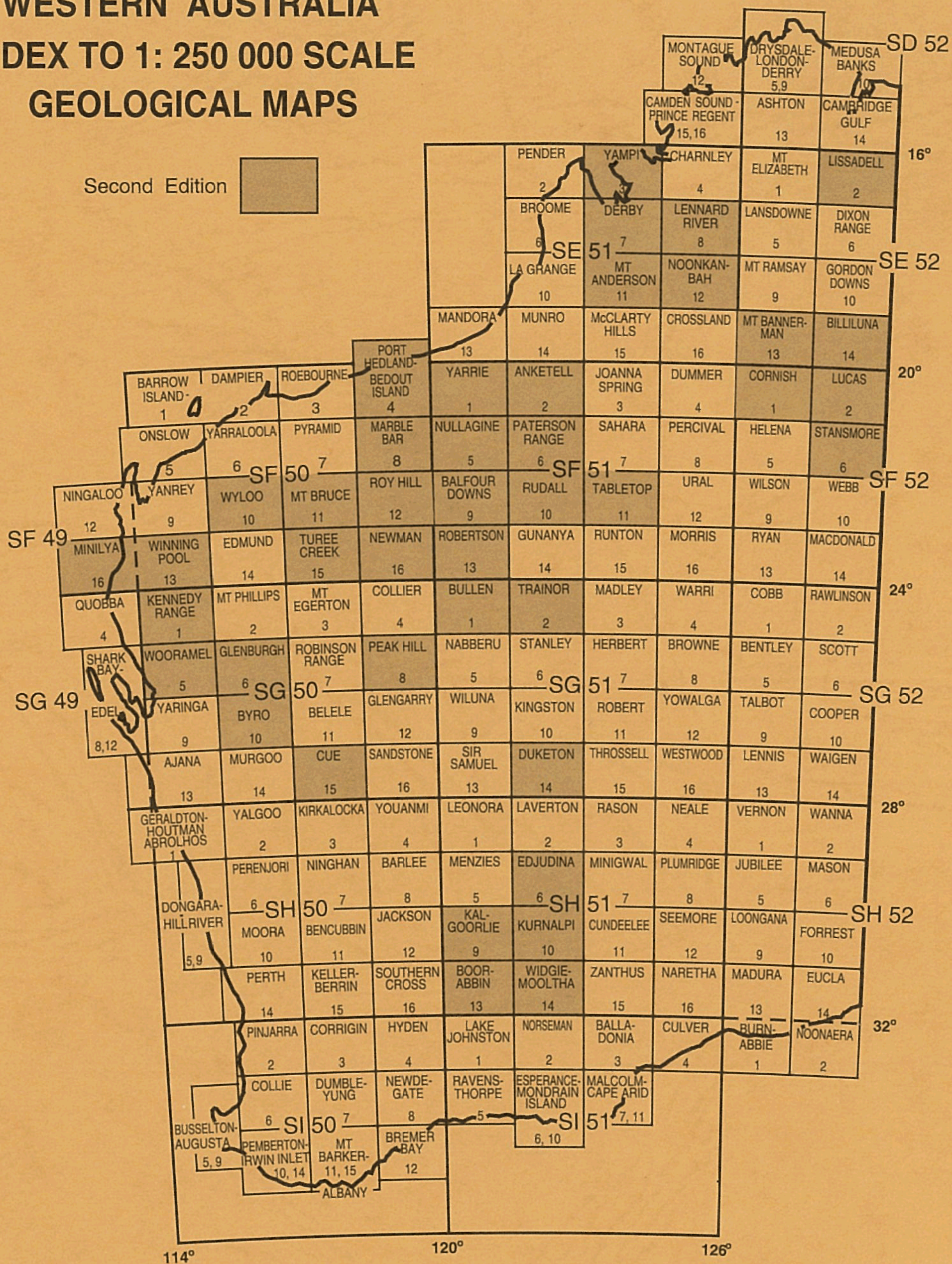
SECOND EDITION

SHEET SE 52-2 INTERNATIONAL INDEX

GEOLOGICAL SURVEY OF WESTERN AUSTRALIA
DEPARTMENT OF MINERALS AND ENERGY

WESTERN AUSTRALIA
INDEX TO 1: 250 000 SCALE
GEOLOGICAL MAPS

Second Edition



GEOLOGICAL SURVEY OF WESTERN AUSTRALIA
1:250 000 GEOLOGICAL SERIES—EXPLANATORY NOTES

LISSADELL

WESTERN AUSTRALIA

SECOND EDITION

SHEET SE 52-2 INTERNATIONAL INDEX

by
A. M. THORNE, S. SHEPPARD, AND I. M. TYLER

Perth, Western Australia 1999

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Explanatory Notes on the Lissadell 1:250 000 Geological Sheet, Western Australia (Second Edition)

by A. M. Thorne, S. Sheppard, and I. M. Tyler

ABSTRACT

The LISSADELL 1:250 000 map sheet (SE 52-2) lies mainly within the Halls Creek Orogen of the east Kimberley region of Western Australia, a major northeasterly trending orogenic belt that developed in the Palaeoproterozoic and continued to be the focus of crustal activity into the Phanerozoic. The Halls Creek Orogen formed in the Palaeoproterozoic between the Kimberley Craton to the northwest, and a composite Archaean craton to the east.

The oldest rocks on LISSADELL are part of the Palaeoproterozoic Lamboo Complex. In the Western zone of the Lamboo Complex, the c. 1870 Ma turbidites of the Marboo Formation were deformed and metamorphosed during the Hooper Orogeny, before being overlain by felsic volcanics of the Whitewater Volcanics. Subvolcanic porphyries, and granitoid and gabbro of the 1865–1850 Ma Paperbark supersuite then intruded the Marboo Formation and Whitewater Volcanics. In the Central zone of the Lamboo Complex, sedimentary and mafic volcanic protoliths to the Tickalara Metamorphics were deposited at c. 1865 Ma. At c. 1850 Ma they were intruded by sheets of tonalite and granodiorite and metamorphosed at high grade.

The Western and Central zones of the Lamboo Complex on LISSADELL were intruded by granitoid and gabbro of the Sally Downs supersuite at 1835–1805 Ma. At the same time, the rocks of the Central zone were deformed and metamorphosed during the Halls Creek Orogeny. This event also affects rocks of the Eastern zone of the Lamboo Complex, and reflects the suturing of the Kimberley Craton and a composite Archaean craton at c. 1820 Ma. During the early stages of the Halls Creek Orogeny, fluvial and shallow-marine clastic sediments of the Speewah Group were deposited in the Speewah Basin on the Western zone of the Lamboo Complex and the now unexposed Kimberley Craton.

At c. 1800 Ma, clastic sediments and basalt of the Texas Downs and Revolver Creek Formations were deposited in the Texas Downs and Revolver Creek Basins in the eastern part of LISSADELL. These sedimentary rocks are lithologically similar to the Kimberley Group that was deposited in the Kimberley Basin to the west. Palaeoproterozoic to Mesoproterozoic rocks of the Bastion Group unconformably overlie the Kimberley Group. Mesoproterozoic siliciclastic sedimentary rocks of the Carr Boyd Group unconformably overlie the Lamboo Complex and Revolver Creek Basin. The Carr Boyd Group was intruded at c. 1180 Ma by the diamondiferous Argyle (AK1) lamproite pipe. Large-scale, north-northeasterly trending sinistral strike-slip faults developed during the Yampi Orogeny at c. 1000 Ma.

During the Mesoproterozoic to Neoproterozoic, siliciclastic sediments were deposited in the Victoria River Basin on the eastern part of LISSADELL. Glacigene rocks of the c. 610 Ma

Duerdin and Albert Edward Groups of the Wolfe Creek Basin unconformably overlie older sedimentary rocks and the Lamboo Complex. Reactivation of strike-slip faults, and associated uplift and erosion, during the c. 560 Ma King Leopold Orogeny was followed in the Early Cambrian by eruption of the Antrim Plateau Volcanics in the Ord and Bonaparte Basins. The Antrim Plateau Volcanics is overlain by Cambrian siliciclastic and carbonate rocks of the Goose Hole and Carlton Groups; the latter is succeeded by Late Devonian coarse-grained siliciclastic rocks of the Cockatoo Group. Sedimentation in the Late Devonian was probably controlled by strike-slip faulting during the c. 400 to 300 Ma Alice Springs Orogeny.

Alluvial gravels, ranging in age from Miocene to the present day, host diamond deposits in creeks that drain the Argyle (AK1) lamproite pipe.

KEYWORDS: Halls Creek Orogen, Lamboo Complex, Speewah Basin, Texas Downs Basin, Revolver Creek Basin, Carr Boyd Basin, Victoria River Basin, Wolfe Creek Basin, Ord Basin, Bonaparte Basin, Marboo Formation, Paperbark supersuite, Tickalara Metamorphics, Sally Downs supersuite, Antrim Plateau Volcanics, Argyle lamproite pipe, diamonds.

INTRODUCTION

LOCATION, ACCESS, AND PREVIOUS WORK

The LISSADELL* 1:250 000 map sheet (SE 52-2) is bounded by latitudes 16°00' and 17°00'S and longitudes 127°30' and 129°00'E, and lies within the east Kimberley region of Western Australia.

The main commercial activities on LISSADELL are diamond mining at the Argyle mine, and cattle grazing for beef. The map sheet includes the Lissadell, Bow River, and Doon Doon pastoral leases, and part of El Questro and Bedford Downs. The Bow River and Lissadell Homesteads in the southern part of the sheet, and the El Questro Homestead along the northern edge of the sheet, are permanently occupied, as are aboriginal settlements at Doon Doon and Glen Hills in the centre of the map sheet. Lake Argyle covers much of the northeastern part of the sheet.

The sealed Great Northern Highway crosses the centre of the sheet, linking Kununurra (60 km by road to the north) and Halls Creek (150 km to the south). The unsealed Duncan Highway traverses the eastern edge of LISSADELL. All-weather graded roads provide access to the Argyle diamond mine in the southeastern part of the sheet. Access within the rest of the sheet is via graded roads and station tracks, most of which are accessible only during the dry season.

Geological investigations prior to 1968 are covered in the Explanatory Notes for the first edition LISSADELL 1:250 000 geological sheet (Plumb, 1968). More recent work is referred to in these second edition Notes.

The present survey continues the remapping of the King Leopold and Halls Creek Orogens by the Geological Survey of Western Australia (GSWA) that commenced in 1986. Field work on LISSADELL was carried out in 1994 and 1995 using 1:40 000 black and white

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* Capitalized names refer to standard map sheets. Where 1:100 000 and 1:250 000 sheets have the same name, the 1:250 000 sheet is implied unless otherwise indicated.

aerial photography flown by the Western Australian Department of Lands Administration (DOLA). The mapping forms part of a joint project with the Australian Geological Survey Organisation (AGSO), carried out as part of the National Geoscience Mapping Accord (NGMA) Kimberley–Arunta project.

PHYSIOGRAPHY, VEGETATION, AND CLIMATE

LISSADELL contains several physiographic provinces, namely the Kimberley Plateau, Kimberley Foreland, Lamboo Hills, Ord Plains, and Cambridge Gulf Lowlands (Plumb, 1968). The physiographic provinces of LISSADELL are shown in Figure 1.

The Kimberley Plateau along the western edge of the sheet is underlain by gently dipping sedimentary rocks of the Kimberley and Bastion Basins.

The Kimberley Foreland Province consists of the O'Donnell and Saw Ranges flanking the Kimberley Plateau, and the Ragged, Carr Boyd, and Osmand–Albert Edward Ranges in the eastern half of LISSADELL. The O'Donnell and Saw Ranges consist of a rugged,

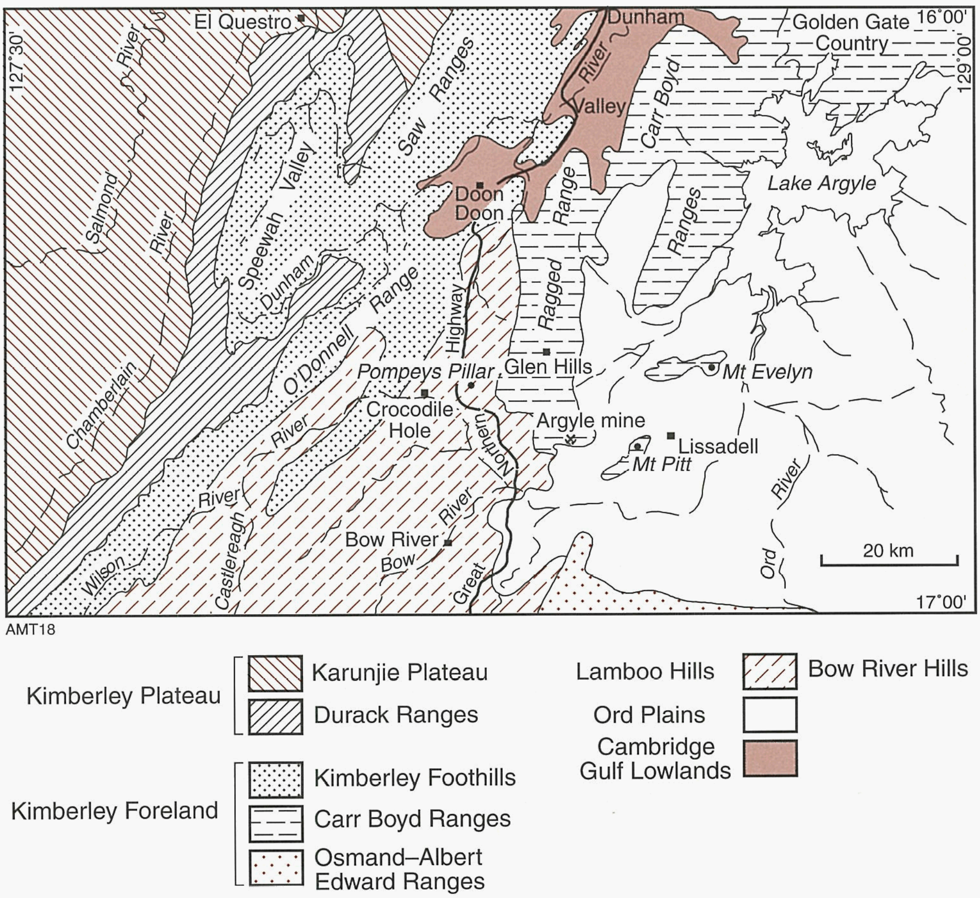


Figure 1. Physiographic and drainage sketch map of LISSADELL (after Plumb, 1968)

highly dissected topography with local relief of up to 300 m. The western edge of the Carr Boyd Range is a high scarp that slopes eastwards in a series of sandstone cuestas (Plumb, 1968). The Carr Boyd and Ragged Ranges are very rugged and relief is up to 500 m. The Osmand–Albert Edward Ranges in the southeast corner consist of high hogbacks and cuestas with a feathered dendritic drainage and relief of about 200 m (Plumb, 1968).

The Lamboo Hills Province is composed of rugged, chaotic hills covered with boulders and tors and having a maximum relief of about 150 m. The province is underlain by granitic and high-grade metamorphic rocks.

The Ord Plains Province covers much of the eastern part of LISSADELL and consists of widespread, low-lying plains developed over poorly exposed Cambrian basalts and carbonate rocks (Plumb, 1968). The Cambridge Gulf Lowlands are confined to the vicinity of the Dunham River in the north of the sheet.

The vegetation of the Kimberley region is described by Beard (1979). The Kimberley Plateau supports curly spinifex with a low eucalypt overstory. Metasedimentary rocks of the Kimberley Foreland Province generally support a tree steppe of snappy gum over hard spinifex or a sparse tree steppe of snappy gum over soft spinifex. Valleys underlain by dolerite carry high-grass savanna woodland with grey box and bloodwood over white grass, cane grass, and blue grass.

Felsic rocks of the Lamboo Hills Province usually support a low savanna woodland of snappy gum over curly spinifex and cane grass. Felsic to intermediate porphyry of the O'Donnell Range supports a low-tree savanna of snappy gum over cane grass. The Ord Plains Province is covered by short-grass plains with scattered snappy gum or a low-tree savanna of cabbage gum and silver-leaved box.

LISSADELL has a semi-arid monsoonal climate with an average annual rainfall of between 550 and 750 mm. Almost all the rain falls between November and April as the result of thunderstorms and cyclones. During the wet season the days are hot to very hot (average daily maxima of about 38°C) and humid, whereas in the dry season the days are warm to hot and dry, with the coolest month being July (average maximum of 27°C). The annual evaporation rate is between 2000 and 2500 mm (Beard, 1979). All watercourses are intermittent, but some creeks contain waterholes that persist until late into the dry season.

REGIONAL GEOLOGICAL SETTING

The major tectonic units on LISSADELL are shown in Figures 2 and 3. A summary of the geological history of LISSADELL is presented in Table 1 and the stratigraphy of the Palaeoproterozoic to Palaeozoic sedimentary rocks is summarized in Table 2. Most of the map sheet lies within the Halls Creek Orogen, which formed initially in the Palaeoproterozoic between the Kimberley Craton to the northwest, and a composite Archaean craton to the east (Tyler et al., 1995). Crystalline rocks in the centre of the sheet are part of the 1920–1790 Ma Lamboo Complex (Dow and Gemuts, 1969; Griffin and Grey, 1990; Page and Sun, 1994; Tyler et al., 1995). The complex is a north-northeasterly trending belt of Palaeoproterozoic igneous and low- to high-grade metamorphic rocks. In the western part of the sheet area the complex is unconformably overlain by the c. 1835 Ma Speewah Basin and the c. 1800 Ma Kimberley Basin (Dow and Gemuts, 1969; Plumb and Gemuts, 1976; Plumb et al., 1985; Page and Sun, 1994; Tyler et al., 1995). In the eastern part of the sheet, crystalline rocks of the Lamboo Complex are unconformably overlain by, or in tectonic contact with, Palaeoproterozoic to Devonian quartz-rich sedimentary rocks and basalt (Dow and Gemuts, 1969; Plumb and Gemuts, 1976; Plumb et al., 1985; Mory and Beere, 1988; Thorne and Tyler, 1996).

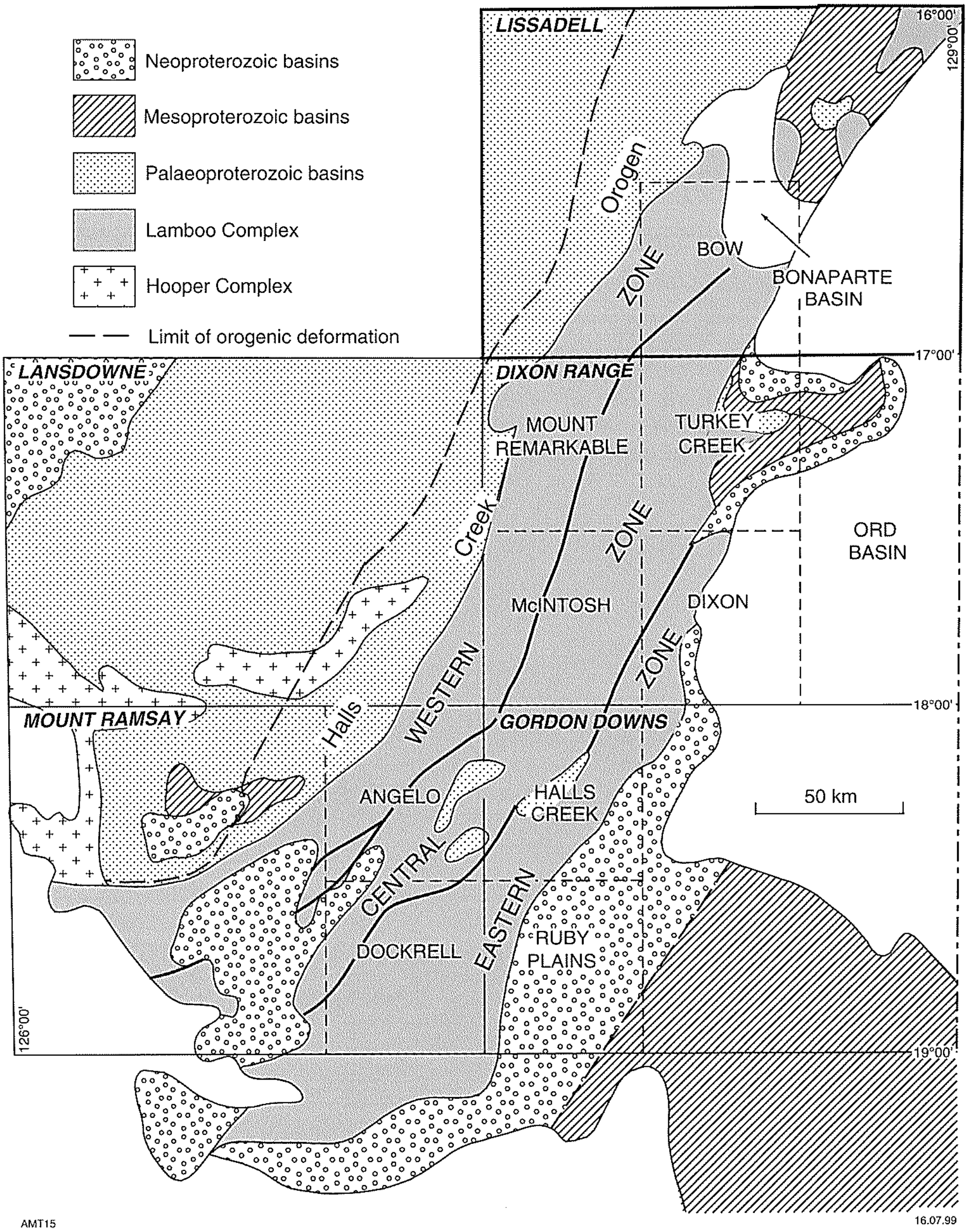
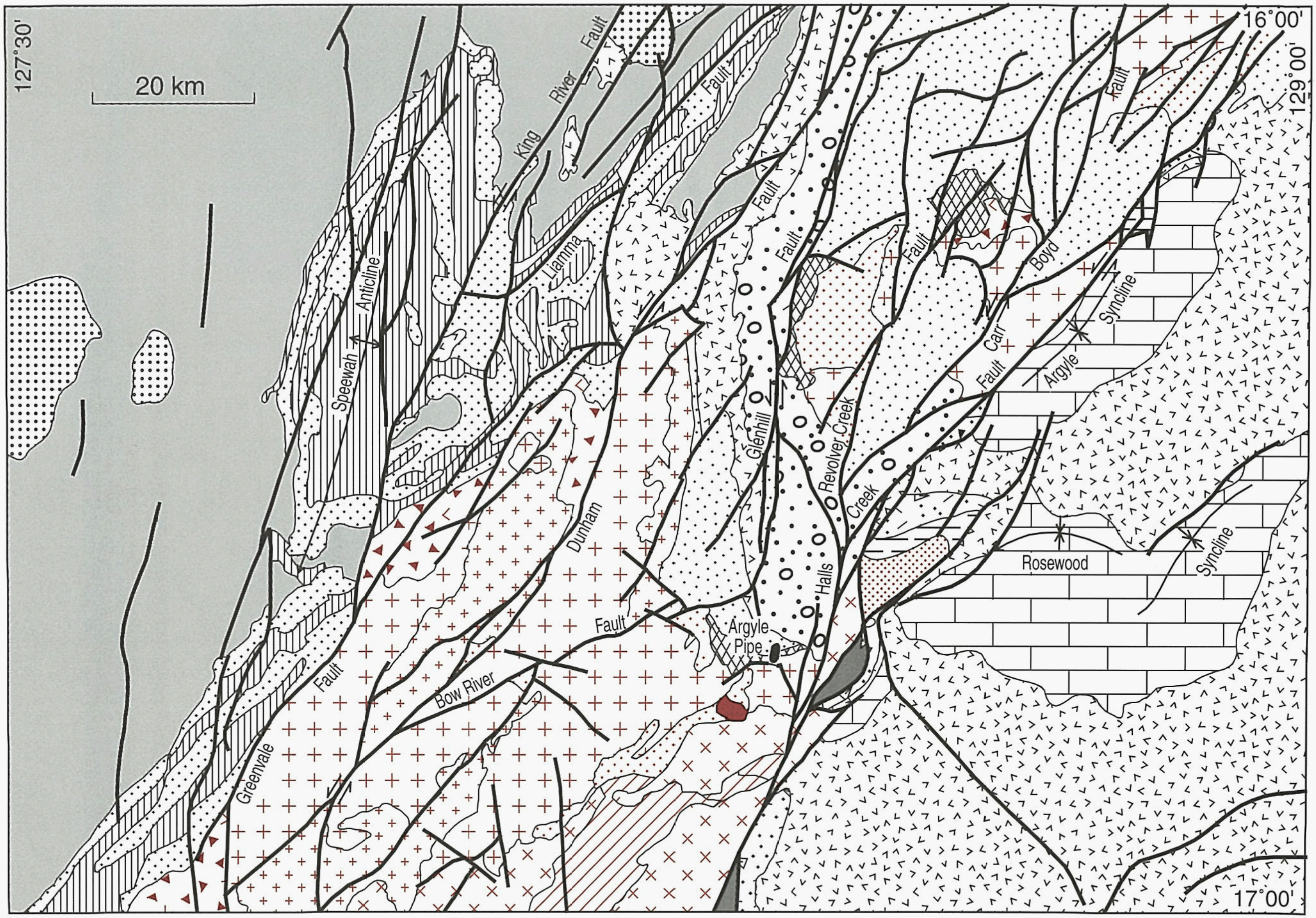


Figure 2. Location of published 1:100 000 and 1:250 000 map sheets in the east Kimberley and their relationship to tectonic zones in the Lamboo Complex

Earlier models for the formation of the Halls Creek Orogen in the Palaeoproterozoic, and other belts of similar age in northern Australia, proposed extension and crustal thinning followed by convergence without subduction of oceanic crust (Hancock and Rutland, 1984; Etheridge et al., 1987; Wyborn, 1988). However, Ogasawara (1988) noted that the



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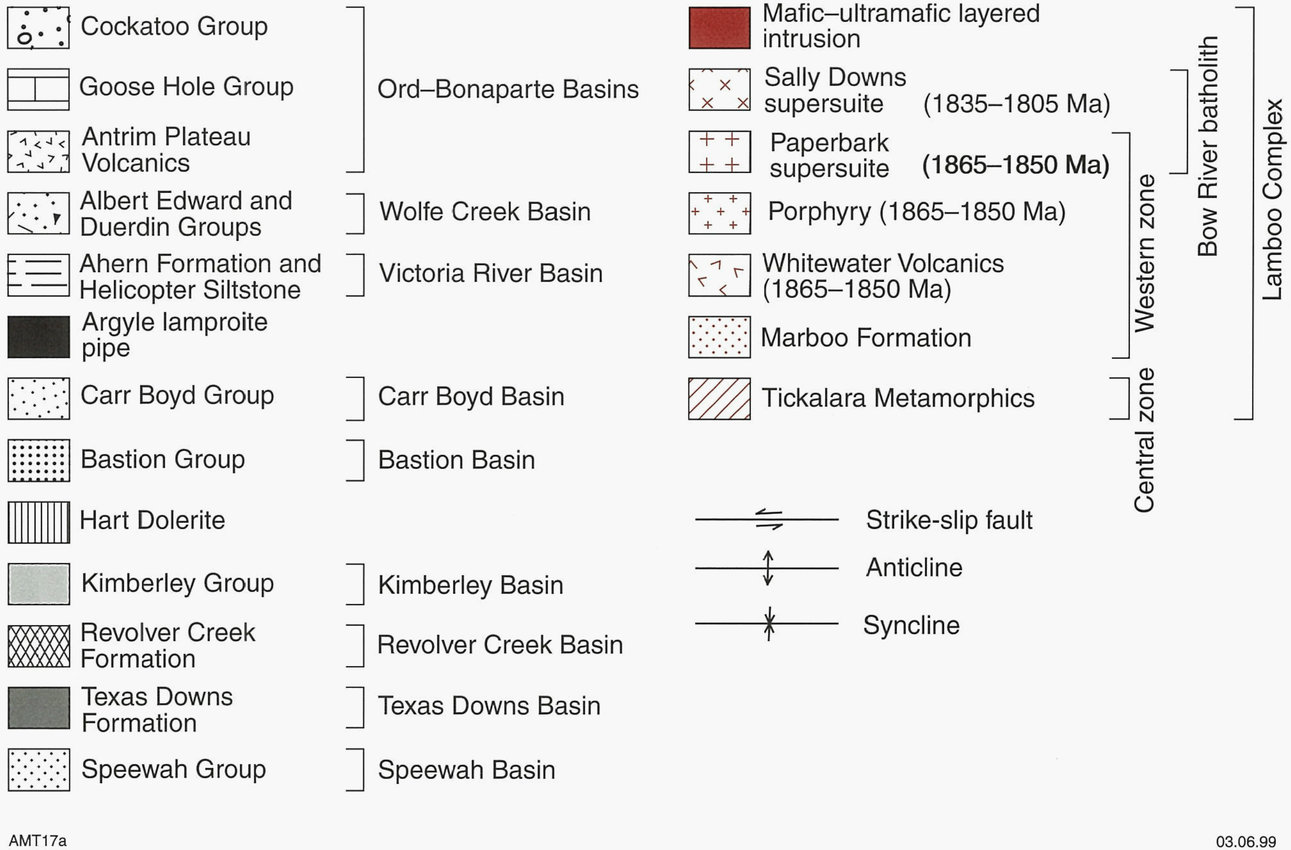

Figure 3. Simplified geological map of Lissadell

Table 1. Summary of the geological history of LISSADELL

Age (Ma)	*Speewah and Kimberley Basins*	*Lamboo Complex* *Western zone*	*Lamboo Complex* *Central zone*
>2500–?1950	Unexposed Kimberley Craton		
c. 1870 Ma		Deposition of the Marboo Formation	
1865–1850		.... HOOPER OROGENY	....
c. 1865		D_1 in Western zone	Deposition of the protolith of the Tickalara Metamorphic Intrusion of layered mafic–ultramafic intrusion parallel to compositional layering
1865–1850		Deposition of Whitewater Volcanics Intrusion of Greenvale and Castlereagh Hill porphyries Intrusion of granitoid and gabbro of the Paperbark supersuite D_2 in the Western zone	Early thrusting (D_1) in Central zone Medium- to high-grade metamorphism (M_1) Intrusion of tonalite–trondjhemmite–granodiorite sheets
c. 1850			Easterly directed thrusting–large-scale recumbent isoclinal D_2 folds
c. 1845			Peak low to moderate P/high T metamorphism (M_2) — syn- to post-D_2
1835–1800	....	HALLS CREEK OROGENY	....
1835–1830	Deposition of the Speewah Group in the Speewah Basin	Uplift and erosion	Intrusion of Mabel Downs Tonalite Formation of D_3 folds, and prominent foliation in Mabel Downs Tonalite, synchronous with epidote–amphibolite facies metamorphism (M_3)
c. 1830		Intrusion of Salt Lick Creek layered mafic–ultramafic intrusion	
1835–1805		Intrusion of other Sally Downs supersuite granitoid and gabbro	
1830–1800			Tight, upright, northeasterly plunging D_4 folds. Activation of Highway Shear Zone under upper greenschist to lower amphibolite facies conditions (M_4)
1827–1790	Deposition of Kimberley Group	Deposition of the Texas Downs and Revolver Creek Formations and ?Kimberley Group	
1800–1790	Intrusion of Hart Dolerite sills into Speewah and Kimberley Groups		
1790–1750		Folding, uplift and erosion	

Table 1. (continued)

Age (Ma)	Speewah and Kimberley Basins	Lamboo Complex: Western zone	Lamboo Complex: Central zone
1750–1550		Deposition of Bastion Group	
1550–1250		Folding, uplift and erosion	
c.1250		Deposition of the Carr Boyd Group	
c.1200		Intrusion of Argyle lamproite and dolerite dykes	
c. 1000	.... YAMPI OROGENY		
	Large-scale, sinistral strike-slip faulting and associated thrusting (D_5) during transpressional event. Greenschist-facies metamorphism (M_5)		
c. 815		Intrusion of Bow Hill lamprophyre dykes	
c. 800		Deposition of the Ahern Formation and Helicopter Siltstone	
c. 610		Deposition of the Duerdin and Albert Edward Groups	
c. 560	.... KING LEOPOLD OROGENY		
		Reactivation of major faults (D_6).	
c. 540		Eruption of the Antrim Plateau Volcanics	
c.525		Deposition of the Goose Hole and Carlton Groups	
c. 400–300	.... ALICE SPRINGS OROGENY		
c. 370		Reactivation of major faults (D_7)	
		Deposition of the Cockatoo Group	
c. 250 to present		Uplift and erosion	

chemistry of tonalites in the Halls Creek Orogen is similar to that of tonalites formed by partial melting of basaltic rock above Phanerozoic subduction zones, and suggested that the Halls Creek Orogen may represent the site of a Palaeoproterozoic convergent margin. More recently, Griffin and Tyler (1992a), Griffin et al. (1994), Tyler et al. (1995) and Sheppard et al. (1995) have argued that the Halls Creek Orogen shares many features with convergent Phanerozoic plate margins associated with subduction of oceanic crust.

Pre-, syn- and post-collisional granites of the Sally Downs supersuite were emplaced at the same time as, or after, the Speewah Group was deposited to the northwest (Tyler et al., 1995; Page and Sun, 1994). Quartz-rich sedimentary rocks and basaltic volcanic rocks that may correlate with the Kimberley Basin (c. 1800 Ma) extended across the Lamboo Complex, and were deposited in the Revolver Creek and Texas Downs Basins (Dow and Gemuts, 1969; Plumb and Gemuts, 1976; Plumb et al., 1985; Tyler et al., 1995; this publication, p. 35–37). The sedimentary rocks of the Kimberley Basin are overlain by those of the ?Mesoproterozoic Bastion Basin (Dow and Gemuts, 1969; Plumb and Gemuts, 1976; Plumb et al., 1985; Tyler et al., 1998b).

Large-scale sinistral strike-slip faulting took place in the Halls Creek Orogen during the Yampi Orogeny between c. 1400 and 1000 Ma. Deformation was accompanied by low-

Table 2. Stratigraphy of the Palaeoproterozoic to Palaeozoic sedimentary rocks on LISSADELL

Basin	*Group*	*Formation*	*Thickness (m)*	*Lithology*
BONAPARTE BASIN	**Cockatoo Group**	Galloping Creek Formation (*DCg*)	1 600	Feldspathic sandstone, pebbly sandstone, pebble to boulder conglomerate
		Kellys Knob Sandstone (*DCk)*	70–350	Quartz sandstone
		Cyril Sandstone (*DCy*)	180	Quartz sandstone, pebbly sandstone, conglomerate and siltstone
		Ragged Range Conglomerate (*DCr*)	300	Pebble to boulder conglomerate and sandstone
	~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ *unconformity/disconformity* ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~			
	Carlton Group	Tarrara Formation (*€Ct*)	60	Siltstone, sandstone and dolomite
	~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ *unconformity/disconformity* ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~			
ORD BASIN	**Goose Hole Group**	Eagle Hawk Sandstone (*€Ge)*	210	Fine-grained micaceous sandstone and limestone
		Panton Formation (*€Gp*)	100–300	Siltstone, mudstone and minor sandstone
		Corby Limestone Member (*€Gpc)*		Limestone and chert
		Shady Camp Well Limestone Member (*€Gps)*		Fossiliferous limestone
		Linnekar Limestone (*€Gl)*	10–25	Laminated fossiliferous limestone and siltstone
		Nelson Shale (*€Gn*)	100–180	Purple siltstone, minor sandstone
		Headleys Limestone (*€Gh*)	35–50	Grey, massive to laminated limestone, minor chert and stromatolites
ORD AND BONAPARTE BASINS		Antrim Plateau Volcanics (*€a*)	200–1 000	Massive or amygdaloidal basalt, minor chert and medium-grained sandstone
		Mount Close Chert Member (*€ac*)	5	Stromatolitic chert and chert breccia
	~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ *unconformity/disconformity* ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~			
WOLFE CREEK BASIN	**Albert Edward Group**	Mount Forster Sandstone *PLo*)	40	Quartz sandstone, pebbly sandstone and conglomerate
	Duerdin Group	Ranford Formation (*PEo*)	600	Siltstone, quartz sandstone, dolomitic sandstone

Table 2. (continued)

Basin	*Group*	*Formation*	*Thickness (m)*	*Lithology*
		Moonlight Valley Tillite (*ⱣEm*)	280	Massive, matrix supported pebble to boulder conglomerate and sandstone
	~ ~ ~ *unconformity/disconformity* ~ ~ ~			
VICTORIA RIVER BASIN		Helicopter Siltstone (*Ᵽh*)	330	Micaceous siltstone and quartz sandstone
		Ahern Formation (*Ᵽf*)	200	Quartz sandstone, pebbly sandstone and conglomerate
	~ ~ ~ *unconformity/disconformity* ~ ~ ~			
CARR BOYD BASIN	**Carr Boyd Group**	Stonewall Sandstone (*ⱣCs*)	750	Quartz sandstone, pebbly sandstone and siltstone
		Glenhill Formation (*ⱣCg*)	1 300	Micaceous siltstone and quartz sandstone
		Lissadell Formation (*ⱣCl*)	1 200	Quartz sandstone and siltstone
		Golden Gate Siltstone (*ⱣCd*)	50–450	Chloritic and carbonaceous siltstone, quartz sandstone and ferruginous sandstone
		Hensman Sandstone (*ⱣCh*)	120	Quartz sandstone
	~ ~ ~ *unconformity/disconformity* ~ ~ ~			
BASTION BASIN	**Bastion Group**	Cockburn Sandstone (*ⱣBc*)	60	Quartz sandstone and pebbly sandstone
		Wyndham Shale (*ⱣBw*)	700	Siltstone, mudstone, micaceous sandstone
		Mendena Formation (*ⱣBm*)	110	Quartz sandstone, and micaceous sandstone, and siltstone
	~ ~ ~ *unconformity/disconformity* ~ ~ ~			
KIMBERLEY BASIN	**Kimberley Group**	Pentacost Sandstone (*ⱣKp*)	1 100	Quartz sandstone, pebbly sandstone and siltstone
		Elgee Siltstone (*ⱣKe*)	200	Siltstone, mudstone, dolomite, and sandstone
		Teronis Member (*ⱣKet*)		Stromatolitic dolomite, siltstone and sandstone
		Warton Sandstone (*ⱣKw*)	220	Massive quartz sandstone and feldspathic sandstone
		Carson Volcanics (*ⱣKc*)	50–220	Basalt and basaltic volcaniclastic rock interbedded with sandstone and mudstone

Table 2. (continued)

Basin	*Group*	*Formation*	*Thickness (m)*	*Lithology*
		King Leopold Sandstone (*ⱣKl*)	700	Quartz sandstone and pebbly quartz sandstone
	~ ~	~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ *unconformity/disconformity*	~ ~ ~ ~ ~ ~ ~ ~ ~ ~	~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~
REVOLVER CREEK BASIN		Revolver Creek Formation (*Ᵽr*)	1 200	Lithic and feldspathic quartz sandstone, siltstone, basalt, and dolerite
	~ ~	~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ *unconformity/disconformity*	~ ~ ~ ~ ~ ~ ~ ~ ~ ~	~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~
TEXAS DOWNS BASIN		Texas Downs Formation (*Ᵽx*)	1 000	Lithic quartz sandstone, siltstone, conglomer ate and basalt
	~ ~	~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ *unconformity/disconformity*	~ ~ ~ ~ ~ ~ ~ ~ ~ ~	~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~ ~
SPEEWAH BASIN	**Speewah Group**	Bedford Sandstone (*ⱣSb*)	300	Quartz sandstone
		Luman Siltstone (*ⱣSl*)	75	Siltstone, mudstone, thin-bedded sandstone
		Lansdowne Arkose (*ⱣSo*)	400	Feldspathic sandstone, and micaceous siltstone
		Valentine Siltstone (*ⱣSv*)	75	Micaceous and chloritic siltstone, quartz sandstone and volcaniclastic sandstone
		Tunganary Formation (*ⱣSt*)	290	Quartz sandstone and feldspathic quartz sandstone and siltstone
		O'Donnell Formation (*ⱣSn*)	260	Quartz and lithic quartz sandstone, siltstone and conglomerate

to medium-grade metamorphism, and established a pattern of north-northeasterly trending synthetic sinistral faults, and east-northeasterly trending antithetic dextral faults (White and Muir, 1989; Tyler et al., 1995; Thorne and Tyler, 1996). Deposition of quartz-rich sedimentary rocks took place in the Carr Boyd Basin in the northeastern part of the sheet (Dow and Gemuts, 1969; Plumb and Gemuts, 1976; Plumb et al., 1985; Thorne and Tyler, 1996) and was followed by intrusion of the Argyle lamproite pipe at c. 1200 Ma (Jaques et al., 1986; Pidgeon et al., 1989).

The Neoproterozoic Ahern Formation and Helicopter Siltstone in the eastern part of LISSADELL are correlated with units in the Victoria River Basin and probably represent Supersequence 1 of the Centralian Superbasin (Thorne and Tyler, 1996; this publication, p. 48–49). The Duerdin Group, which includes glacigene rocks, and the overlying Albert Edward Group were deposited in the Wolfe Creek Basin and are correlated with Supersequence 3 of the Centralian Superbasin, equivalent to the c. 610 Ma Marinoan glaciation (Grey and Corkeron, 1998). Deposition of the Neoproterozoic sedimentary rocks was followed by the King Leopold Orogeny at c. 560 Ma, which produced thrusting and sinistral strike-slip faulting within the Halls Creek Orogen (Tyler and Griffin, 1990; Tyler et al., 1991; Shaw et al., 1992).

The Ord Basin, which overlies the eastern part of the Lamboo Complex, was initiated in the latest Proterozoic to early Cambrian with the extrusion of widespread flood basalts (Mory and Beere, 1988). Late Devonian alluvial fans developed in response to strike-slip movements along the Halls Creek fault system (Mory and Beere, 1988; Thorne and Tyler, 1996). Later sinistral strike-slip deformation may be related to the c. 400 to 300 Ma Alice Springs Orogeny in central Australia (Thorne and Tyler, 1996).

PALAEOPROTEROZOIC LAMBOO COMPLEX

Hancock and Rutland (1984) divided the Lamboo Complex into four zones. Griffin and Tyler (1992b) amalgamated zones II and III, and Tyler et al. (1994, 1995) subsequently modified their zone boundaries. The three zones (Western, Central, and Eastern) are parallel to the length of the complex, and are separated by major fault systems. Only the Western and Central zones are present on LISSADELL. The difficulty in correlating stratigraphic units and structural events across the zone boundaries in the Lamboo Complex suggests that they represent tectono-stratigraphic terranes (Tyler et al., 1995).

The Western zone of the Lamboo Complex is a continuation of the Hooper Complex in the King Leopold Orogen of the west Kimberley region (Griffin et al., 1994; Tyler et al., 1995). The Western zone is composed of low- to high-grade metasedimentary rocks of the Marboo Formation (c. 1870 Ma; Tyler et al., in press) unconformably overlain by felsic volcanic rocks of the Whitewater Volcanics dated at c. 1855 Ma (Page and Sun, 1994). The metasedimentary and volcanic rocks were intruded by granitoid and gabbro of the Paperbark supersuite, as well as cogenetic subvolcanic porphyries at 1865–1850 Ma (Tyler et al., 1995; Griffin et al., in prep.). Deformation and metamorphism of the Marboo Formation, together with the extensive magmatism, took place during the 1865–1850 Ma Hooper Orogeny (Tyler and Page, 1996).

The Central zone is dominated by medium- to high-grade metasedimentary and meta-igneous rocks of the Tickalara Metamorphics, which were deformed and metamorphosed between c. 1865 and 1856 Ma and at 1850–1845 Ma (Page and Sun, 1994; Tyler and Page, 1996; Bodorkos et al., 1998). In the southern part of the Central zone, low-grade metasedimentary and mafic and felsic metavolcanic rocks of the Koongie Park Formation were deposited on the Tickalara Metamorphics at 1845–1840 Ma. The Central zone was intruded by large volumes of granitoid and associated gabbro of the Sally Downs supersuite during the Halls Creek Orogeny at 1835–1805 Ma (Tyler and Page, 1996).

The Eastern zone is composed of low-grade metasedimentary and metavolcanic rocks of the Halls Creek Group, which unconformably overlie domal culminations of granitoid and volcanic rock dated at 1920–1900 Ma (Griffin and Tyler, 1992b; Page and Sun, 1994; Blake et al., 1998). Alkaline volcanic rocks of the Butchers Gully Member (Griffin and Tyler, 1992b) and Maude Headley Volcanic Member in the middle to upper part of the sequence were erupted between c. 1857 and 1848 Ma (Blake et al., 1998). The Halls Creek Group was deformed and metamorphosed before being intruded by 1820–1810 Ma granitoid of the Sally Downs supersuite at the southern end of the Lamboo Complex.

WESTERN ZONE

Marboo Formation (*Ɇm, Ɇmh*)

The Marboo Formation outcrops within granitoid rocks in the central part of LISSADELL. To the east of Bow River Homestead the formation is faulted against the Tickalara Metamorphics. Previously these rocks were placed either within the Tickalara Metamorphics, or designated as undifferrentiated Halls Creek Group (Plumb, 1968). However, although the rocks are lithologically similar to the Olympio Formation of the Halls Creek Group, they were deformed prior to being intruded by granitoids that have been dated at c. 1860 Ma (Page et al., in prep.). In contrast, the Olympio Formation was still being deposited after c. 1847 Ma (Blake et al., 1998). Griffin and Tyler (1992b) correlated the turbiditic metasedimentary rocks in the Western zone of the Lamboo Complex with the Marboo Formation in the Hooper Complex of the west Kimberley region. The Marboo Formation in the Hooper Complex was also deformed and metamorphosed prior to the intrusion of granitoid plutons at c. 1865 Ma. The youngest detrital zircons in the Marboo Formation indicate a maximum age of c. 1872 Ma (Tyler et al., in press). The rocks that are included in the Marboo Formation on LISSADELL are typically low-grade, except where they have been metamorphosed (*Ɇmh*) within the contact aureoles of granitoid plutons of the Sally Downs supersuite.

The rocks that make up the Marboo Formation on LISSADELL (*Ɇm*) typically consist of deformed and weakly metamorphosed interbedded mudstone, siltstone, greywacke, lithic greywacke, and lithic quartz wacke. The top of the unit is not seen, and no base or basement has been recognized. Griffin et al. (1993) estimated the thickness of the Marboo Formation on LENNARD RIVER to be in excess of 7 km, although this may have been tectonically thickened by layer-parallel shearing and isoclinal folding.

The thinly-bedded nature of the Marboo Formation rocks on LISSADELL is consistent with them forming part of the upper part of the unit (Griffin et al., 1993). The sandstone beds are typically up to 1 m thick, and develop graded bedding, passing up into siltstones and mudstones. They are interpreted as representing deposition by turbidity currents (e.g. Walker, 1984), with partial preservation of Bouma cycles (units ADE). Palaeocurrent data have not been obtained from LISSADELL, but Hancock (1991) recorded palaeocurrent data on LANSDOWNE that indicated a depositional slope from the north and northeast. This contrasts with data from the Olympio Formation, where depositional currrents came from the northwest (Hancock, 1991).

The arenite units consist of subangular to subrounded clasts of quartz, plagioclase, K-feldspar, and quartzofeldspathic lithic fragments within a recrystallized matrix. According to Hancock (1991) the provenance of the arenite component of the Marboo

Formation was a 'mature quartzose to quartzo-feldspathic hinterland with some acid volcanic material interlayered with or overlying granite'. Sensitive High-Resolution Ion Microprobe (SHRIMP) U–Pb dating of detrital zircons from a low-grade wacke in the Hooper Complex, indicates that the provenance terrain contributed zircons dated at about 2500–2400 Ma, 2300 Ma, 2200–2050 Ma, 1970 Ma and 1910 Ma (Tyler et al., in press).

Whitewater Volcanics (*Ƀw*)

The Whitewater Volcanics outcrops as a belt about 600 km long and up to 30 km wide that extends from the western end of the King Leopold Orogen to north of Lake Argyle in the Halls Creek Orogen. Page and Hancock (1988) obtained a conventional U–Pb zircon age of 1850 ± 5 Ma for the Whitewater Volcanics from a sample about 20 km south-southwest of Doon Doon Homestead. Griffin et al. (in prep.) re-analyzed the same sample and obtained a SHRIMP U–Pb zircon age of 1857 ± 4 Ma. This age is identical to that obtained from a sample of the Whitewater Volcanics in the Hooper Complex. In the Hooper and Lamboo Complexes, the Whitewater Volcanics is intruded by 1865–1850 Ma granitoids of the Paperbark supersuite. However, the Whitewater Volcanics grades into compositionally identical high-level porphyry intrusions and coarse-grained granitoids, suggesting that the three rock types are broadly coeval and cogenetic (Dow et al., 1964; Gellatly et al., 1975; Griffin et al., 1993; Griffin et al., in prep.).

On LISSADELL, most of the Whitewater Volcanics outcrops adjacent to the Greenvale Fault in the southern part of the sheet, where this unit forms rugged dissected hills with about 100 m relief. In addition, about 15 km^2 of Whitewater Volcanics is exposed in a small outlier on the western edge of Lake Argyle. Fine-grained feldspar porphyry adjacent to the Dunham Fault in the O'Donnell Range, previously shown as Whitewater Volcanics (Plumb et al., 1968), has been reassigned to the Castlereagh Hill Porphyry as it lacks evidence of an extrusive origin.

The Whitewater Volcanics is unconformably overlain by the O'Donnell Formation, which forms the base of the Speewah Group. A strip of Whitewater Volcanics east of Aida Vale Mill is intruded by coarse-grained porphyritic granite of the Gordons Gorge Granite of the Paperbark supersuite.

On LISSADELL, most of the Whitewater Volcanics consists of dacitic to rhyolitic ignimbrites (equivalent to the 'porphyries' of Gellatly et al., 1975), minor lava flows, and lapilli tuff. The ignimbrites are crystal-rich and also contain several percent of fine-grained volcanic lithic fragments in a micrographyric groundmass. Columnar jointing is well developed in crystal-poor ignimbrites about 8 km southwest of the Dunham Jump-Up. A pyroclastic origin is suggested by the presence of faint, recrystallized pumice fragments and abundant lithic fragments, which are recognizable in outcrop. The columnar jointing and microgranophyric groundmass indicate that these rocks may have formed thick ignimbrite flows (McPhie et al., 1993). Interbeddded with the volcanic rocks are minor granule conglomerate, volcanic sandstone, and volcanic lithic breccia. East of Aida Vale Mill, a thin strip of Whitewater Volcanics consists of recrystallized, crystal-rich ignimbrites.

In contrast to the extensive exposures on DIXON RANGE and elsewhere on LISSADELL, the Whitewater Volcanics that outcrops on the western edge of Lake Argyle is dominated by volcanogenic sedimentary rocks. The rocks consist of medium- to coarse-grained, crystal-rich sandstone, siltstone, phyllite, and local cross-stratified, granular, and pebbly quartz sandstone.

Castlereagh Hill Porphyry (*Epc*)

The Castlereagh Hill Porphyry of Dow and Gemuts (1969) and Gemuts (1971) is composed of two distinct mappable units: a porphyry with medium to coarse phenocrysts of quartz, K-feldspar and plagioclase, and a plagioclase microgranodiorite porphyry. The name Castlereagh Hill Porphyry is retained for the microgranodiorite, which is exposed around the type area at Castlereagh Hill on LISSADELL. The strongly porphyritic rock unit is redefined as the Greenvale Porphyry, which is mainly exposed to the southwest on the MOUNT REMARKABLE 1:100 000 map sheet (Sheppard et al., 1997b).

The Castlereagh Hill Porphyry has not been dated but it is likely to be coeval with the Whitewater Volcanics and granitoids of the Paperbark supersuite. This porphyry strongly resembles the quartz–feldspar porphyry that constitutes much of the Whitewater Volcanics, and it has a texture and mineralogy identical to that of the Crooked Creek Granite, differing only in its finer grainsize. There are also strong similarities in whole-rock chemistry between the Whitewater Volcanics, the high-level porphyry intrusions, and the granitoids from the Paperbark supersuite, suggesting that the three units are cogenetic (Griffin et al., in prep.).

The Castlereagh Hill Porphyry outcrops in two north-northeasterly trending belts around Castlereagh Hill and in the O'Donnell Range, and underneath the scarp along the western edge of the Ragged Range. In the O'Donnell Range the porphyry forms a dissected plateau about 200–250 m above the surrounding granite country. Elsewhere the porphyry forms rounded or steep hills with a smoother pattern than that of the granitoids on aerial photographs and Landsat TM images. The Castlereagh Hill Porphyry also outcrops in a small (~2 km^2) outlier southwest of the Carr Boyd lead–zinc–silver occurrence on the western edge of Lake Argyle.

The bulk of the Castlereagh Hill Porphyry consists of dark-grey, massive microgranodiorite with fine to medium phenocrysts of plagioclase. Locally (e.g. 2–3 km southeast of Mount Lookout), the unit may coarsen, having a fine- to medium-grained groundmass and phenocrysts up to nearly 7 mm long. These coarser grained parts of the Castlereagh Hill Porphyry strongly resemble the Crooked Creek Granite. The exposure underneath the Ragged Range is a sheet-like intrusion which grades from a medium-grained porphyry at the base through a fine grained porphyry to a black, very fine grained porphyry at the top (Gemuts, 1971). Most samples are moderately to strongly altered, and micrographic and myrmekitic intergrowths between quartz and feldspars are widespread. Fine-grained, aphyric mafic inclusions are generally sparse but widespread.

The contact between the Castlereagh Hill Porphyry and the Whitewater Volcanics was not observed on LISSADELL. The relationship between the porphyry and individual plutons of the Paperbark supersuite on Bow (1:100 000) is not known, and faulting has complicated many of the contacts. However, 5 km south of Pompeys Pillar, just north of the highway, medium-grained granodiorite of the Crooked Creek Granite contains numerous inclusions of Castlereagh Hill Porphyry over an area of about 40 m^2. The inclusions are up to one metre in diameter, are rounded, and contacts with the host granodiorite vary from curvilinear to weakly cuspate. The rounded shape of the inclusions, and the cuspate margins of some inclusions, suggest that the porphyry was not rigid when included in the granodiorite. The inclusion shapes and textures are consistent with either the granodiorite having intruded incompletely solidified porphyry, or porphyry having intruded partly crystallized granodiorite (i.e. the inclusions represent a swarm). Thus, textural and mineralogical similarities between the Castlereagh Hill Porphyry and the Crooked Creek Granite suggest that the two units are related, and therefore they may not be separated by a substantial time gap.

Most samples contain 15–25% phenocrysts of subhedral plagioclase some 1–2 mm long, minor pseudomorphs after hornblende and biotite, and rare quartz and K-feldspar, in a fine- or very fine grained groundmass with a granular texture. Scattered microphenocrysts of opaque minerals are also present. The groundmass is composed of quartz, plagioclase, K-feldspar, chlorite, sericite, epidote, and accessory apatite and zircon. Micrographic and granophyric intergrowths between quartz and K-feldspar, as well as myrmekitic intergrowths between quartz and plagioclase, are widespread. Most samples from the Castlereagh Hill Porphyry are moderately to strongly altered. Microcrystalline sericite and calcite (or clinozoisite) replace phenocryst and groundmass plagioclase. Hornblende is pseudomorphed by chlorite, quartz and epidote, and chlorite and quartz replace biotite.

Greenvale Porphyry (*Ꞓpg*)

The Greenvale Porphyry is a massive, dark-grey quartz–feldspar porphyry with medium to coarse phenocrysts of quartz, sanidine, and plagioclase. This unit outcrops extensively in the northern part of MOUNT REMARKABLE (1:100 000), but on LISSADELL is restricted to about 15 km^2 west of Castlereagh Creek on the southern edge of the sheet. In addition, there are several small exposures (each <0.5 km^2) 7 km east-southeast of Beefwood Yard, 2 km south of Bow Hill, and 7 km west of Bow Hill. All three of the smaller occurrences are located at the margin of medium- or coarse-grained porphyritic granite intrusions. The porphyry is composed of 20–30% phenocrysts of subhedral plagioclase up to 5 mm long, oblate quartz 5–10 mm long, and rounded K-feldspar up to 20 mm in diameter in a groundmass of fine-grained quartz, plagioclase, microcline, and altered ferromagnesian minerals.

A sample of the Greenvale Porphyry from MOUNT REMARKABLE (1:100 000) yielded a SHRIMP U–Pb zircon age of 1855 ± 4 Ma (Griffin et al., in prep.). This age is indistinguishable from that of the Mount Nyulasy Granite, with which the porphyry shares some textural features. The age of the Greenvale Porphyry is also inseparable from that of the Whitewater Volcanics, and other high-level porphyry intrusions and coarse-grained granitoids in the Hooper Complex and Western zone of the Lamboo Complex (Griffin et al., in prep.).

Contacts between the porphyry and Mount Nyulasy Granite are sharp, and west of Bow Hill, granite close to the contact contains a few inclusions of porphyry. The large, rounded sanidine phenocrysts and oval-shaped quartz phenocrysts in the Greenvale Porphyry are similar to those in the Mount Nyulasy Granite. At a granite quarry 10 km north of Pompeys Pillar, the Mount Nyulasy Granite contains inclusions of a rock type strongly resembling the Greenvale Porphyry. A dyke of Greenvale Porphyry intrudes the Dinner Creek Tonalite about 3.5 km west-southwest of the Argyle diamond mine.

Paperbark supersuite

Geochronology (U–Pb SHRIMP) indicates that the 'Bow River Granite' of Dow and Gemuts (1969) and the 'Bow River Granitoid Suite' of Ogasawara (1988) contain a wide range of granitoid and gabbro compositions ranging in age from 1865–1800 Ma. The granitoid and gabbro are subdivided into the 1865–1850 Ma Paperbark supersuite and the 1835–1805 Ma Sally Downs supersuite. Collectively, the two supersuites constitute the Bow River batholith.

The Paperbark supersuite consists of a belt of 1865–1850 Ma I-type (Chappell and White, 1974) granitoid and subordinate gabbro plutons that intruded the Western zone of the Lamboo Complex and the Hooper Complex in the west Kimberley. Although the gabbros constitute a magmatic event separate from that of the granitoids, they are so closely associated in space and time with the granitoids that they are included in the supersuite (Sheppard et al., 1995). Field relationships indicative of mafic–felsic magma mingling are widespread in the Paperbark supersuite (Blake and Hoatson, 1993; Sheppard, 1996).

Biotite gabbro (ᵽgPob)

On LISSADELL, massive biotite-rich gabbros of 1865–1850 Ma age are restricted to several isolated outcrops <1 km^2 surrounded by granitoid. Rock types consist of fine-grained biotite gabbro with xenocrysts of quartz and feldspar, and medium- to coarse-grained biotite gabbro. Farther south, on MOUNT REMARKABLE (1:100 000), Sheppard et al. (1997b) showed that quartz and feldspar xenocrysts in biotite gabbro were the product of mingling between mafic and felsic magmas.

Mingled gabbro–granite (ᵽgPog)

In the Carr Boyd Ranges, about 13 km north of Mount Chambers, strongly weathered, mingled gabbro–granite is exposed in an inlier less than 1 km^2 within the Hensman Sandstone.

Beefwood Yard Granite (ᵽgPby)

The Beefwood Yard Granite outcrops to the east and northeast of Beefwood Yard, after which the intrusion is named. The granite is elongate in a northeast direction and about 20 km long by some 5 km wide. The intrusion forms very rugged hills with up to 150 m relief that are covered in boulders and tors. The Beefwood Yard Granite is composed of medium- and coarse-grained porphyritic biotite monzogranite. The rocks typically consist of 20–30% rounded microperthite phenocrysts up to 2.5 cm in diameter and less than 10% oblate quartz phenocrysts in a groundmass of quartz, plagioclase, microperthite, and biotite. Thin sections show a moderate degree of recrystallization of quartz and microperthite, and alteration of plagioclase and biotite.

The contact between the Beefwood Yard Granite and the Crooked Creek Granite is, in part, marked by a strip of Greenvale Porphyry. The porphyry contains microperthite and quartz phenocrysts similar to those in the Beefwood Yard Granite, and the porphyry may represent an early marginal phase of the granite. The contact between the Beefwood Yard Granite and the Castlereagh Hill Porphyry is typically strongly sheared.

Crooked Creek Granite (ᵽgPcc)

The Crooked Creek Granite shows a strong spatial relationship with the Castlereagh Hill Porphyry, and was included in the porphyry by Plumb (1968) and Dow and Gemuts (1969). The two rock types are mineralogically and texturally similar, the main difference being that the Crooked Creek Granite is coarser grained.

The bulk of the Crooked Creek Granite outcrops as an elongate (~20 km long by 5–7 km wide), northeast-trending intrusion extending from west of Fish Hole Yard to north of Bow Hill. This unit also outcrops discontinuously from southeast of Pompeys Pillar to the Dunham Jump-Up, between the Castlereagh Hill Porphyry and Mount Nyulasy Granite. Further exposures are present north of Lake Argyle, between Spillway Creek and Matchbox Creek. The intrusion outcrops as rounded, bouldery hills with a maximum relief of about

120 m. The Crooked Creek Granite is composed of medium-grained biotite granodiorite and monzogranite. The rocks typically have about 10–20% phenocrysts of euhedral plagioclase in a granular groundmass of plagioclase, quartz, microperthite, biotite, and amphibole, with accessory iron oxides, apatite, zircon, and secondary schorl. Most samples are moderately to strongly altered.

Contacts between the Crooked Creek Granite and other rock units in the south of the map sheet area are commonly complicated by faulting and shearing. About 7 km southwest of Bow Hill rare veins of coarse-grained, porphyritic biotite syenogranite of the Mount Nyulasy Granite cut even-textured granodiorite of the Crooked Creek Granite. Field relationships between the Crooked Creek Granite and the Castlereagh Hill Porphyry 5 km south of Pompeys Pillar suggest that the granite and porphyry are in part coeval. The mineralogical and textural similarities of the two units suggest that they are related.

Dinner Creek Tonalite (PgPdc)

The main mass of the Dinner Creek Tonalite outcrops over about 60 km^2 southwest of the Argyle diamond mine, with a small plug or remnant outcropping about 2 km west-southwest of Bow River Homestead. The intrusion is poorly exposed, with outcrops generally consisting of scattered boulders and isolated low hills amongst colluvium. The Dinner Creek Tonalite consists of fine- to medium-grained, massive hypersthene–biotite–amphibole tonalite.

The Dinner Creek Tonalite intruded and contact metamorphosed the Marboo Formation and is intruded by the Mount Nyulasy Granite at numerous localities; for example 1.5 km west of Bow River Homestead, and south of Mount Nyulasy. The tonalite has not been dated, but the Mount Nyulasy Granite has a SHRIMP U–Pb zircon age of 1859 ± 3 Ma. About 1 km west of Wesley Yard the Dinner Creek Tonalite is intruded by medium-grained, porphyritic biotite monzogranite of the Wesley Yard Monzogranite. About 3.5 km west-southwest of the Argyle diamond mine the tonalite is intruded by a dyke of Greenvale Porphyry.

Rocks in the Dinner Creek Tonalite vary from fine grained and weakly porphyritic to medium grained (2–3 mm) and even textured. Phenocrysts consist of subhedral plagioclase and hypersthene. The groundmass is composed of subhedral to anhedral granular plagioclase, quartz, biotite, actinolite, and rare microcline. Accessory minerals include opaque minerals, apatite, zircon, and allanite. Alteration and recrystallization are a feature of most samples.

Inclusions of rounded to subangular, very dark grey, fine-grained tonalite 2–5 cm in diameter are widespread and may be locally abundant. Angular inclusions of metasedimentary rocks are locally abundant.

Gordons Gorge Granite (PgPgg)

The Gordons Gorge Granite, named after Gordons Gorge, is a northeast-trending intrusion about 50 km long and 5–10 km wide that extends from near the southern edge of the sheet to north of Mount Lookout. This granite forms rugged, bouldery hills with a maximum relief of about 100 m in the south, and scattered hills and tors amongst colluvium in the north. The intrusion is composed of massive, coarse-grained porphyritic biotite(–hornblende) monzogranite and syenogranite with a rapakivi-like texture. Microperthite phenocrysts may reach 3 cm in diameter, and most contain abundant inclusions of plagioclase, quartz, and ferromagnesian minerals.

The Gordons Gorge Granite intruded the Whitewater Volcanics east of Aida Vale Mill, and strongly sheared, medium-grained, even-textured biotite granitoid (*ȒgPe*) northeast of Aida Vale Mill. The relationship of the Gordons Gorge Granite to the Castlereagh Hill Porphyry is uncertain, but about 2 km south of Jacks Yard weathered pavements show veins of a microgranodiorite, similar in appearance to the Castlereagh Hill Porphyry, in the Gordons Gorge Granite. The contact between the Gordons Gorge Granite and the Neil Creek Monzogranite is exposed beside a track about 14 km southwest of Beefwood Yard. Close to the contact the Gordons Gorge Granite contains numerous rounded inclusions of the Neil Creek Monzogranite. Several textures suggest that the magmas forming the two granite intrusions may have mingled: firstly, rare veins of each granite type cut the other phase; secondly, inclusions of the Neil Creek Monzogranite are all rounded, some with crenulate margins; and thirdly, as the contact is approached, the Neil Creek Monzogranite contains progressively more K-feldspar and quartz phenocrysts, similar in appearance to those in the Gordons Gorge Granite.

Matchbox Granite (ȒgPmt)

The Matchbox Granite is named after Matchbox Creek in the Golden Gate Country north of Lake Argyle. The granite, which outcrops over more than 50 km^2 on LISSADELL and also extends to CAMBRIDGE GULF, forms scattered boulders and tors amongst extensive colluvium. The Matchbox Granite is composed of massive, medium- to coarse-grained, porphyritic biotite monzogranite and syenogranite. The age of this unit is not known.

The Matchbox Granite intruded low-grade metasedimentary and minor mafic meta-igneous rock of the Marboo Formation. The contact between the Matchbox Granite and Castlereagh Hill Porphyry is faulted. North of the Lake Argyle Road between Matchbox Creek and Stonewall Creek, scattered exposures show extensive veins and dykes of a fine- to medium-grained, porphyritic biotite granitoid in the Matchbox Granite. The finer granitoid contains phenocrysts of rounded K-feldspar up to 2 cm in diameter, oblate quartz and minor subhedral plagioclase. This rock type strongly resembles the Greenvale Porphyry, although the groundmass in the former is coarser grained.

Mount Nyulasy Granite (ȒgPmn)

The Mount Nyulasy Granite is a large intrusion (>250 km^2) that extends from southwest of the Bow River Homestead northeast to Mount Nyulasy and northwards to the Dunham Jump-Up. The intrusion has a SHRIMP U–Pb zircon age of 1859 ± 3 Ma (Page et al., in prep.). The southern part of the intrusion outcrops as rugged bouldery hills with about 200 m relief, and as prominent whalebacks. Farther north, the granite forms isolated bouldery hills developed in an extensive colluvial plain. The Mount Nyulasy Granite is composed of massive, coarse-grained porphyritic biotite monzogranite and syenogranite with a rapakivi-like texture. Near major fault zones the intrusion is moderately to strongly foliated. In the southwest corner of the intrusion, and locally along its eastern margin, medium- to coarse-grained, even-textured or weakly porphyritic monzogranite predominates.

The Mount Nyulasy Granite intruded hornfels of the Marboo Formation southwest of Bow River Homestead and northeast of Mount Nyulasy. The granite intruded the Dinner Creek Tonalite at several localities, and the Crooked Creek Granite 6 km southwest of Bow Hill.

Most samples from the intrusion contain 20–30% or more phenocrysts of rounded plagioclase 1–3 cm (rarely up to 4 cm) in diameter, some of which have narrow rims of fine-grained sodic plagioclase in a checkerboard pattern. Grey quartz forms rounded or oblate crystals up to 1 cm long. The groundmass consists of quartz, microperthite, and subhedral plagioclase, with subordinate fine-grained, dark-brown biotite commonly in clots up to 5 mm in diameter. Where fresh, plagioclase displays strong oscillatory zoning. Accessories include opaque minerals, apatite, and zircon. Southwest of Pompeys Pillar, the intrusion contains widely dispersed clots of tourmaline up to several centimetres in diameter and, locally, pods of tourmaline–quartz–feldspar up to 20 cm long.

The Mount Nyulasy Granite contains sparse, lenticular to angular, fine-grained biotite-rich inclusions up to 4 cm long. However, between the Great Northern Highway and the western edge of the Ragged Range, porphyry inclusions are abundant. In the quarry 10 km north of Pompeys Pillar the granite contains two types of rounded porphyry inclusions typically 10 cm or more in diameter. The first type is of feldspar(–quartz) porphyry with less than 10% phenocrysts of microperthite and minor quartz and biotite in a black microcrystalline groundmass. The second type, which resembles the Greenvale Porphyry, consists of feldspar–quartz(–biotite) porphyry with about 30% phenocrysts of rounded K-feldspar, oblate quartz, subhedral plagioclase, and minor biotite.

Neil Creek Monzogranite (ⱣgPnc)

The Neil Creek Monzogranite is a north-northeasterly trending intrusion adjacent to the Greenvale Fault in the southwest corner of the map sheet. This unit outcrops over about 40–50 km^2 as low bouldery rises and hills with up to 60 m relief, and extensive pavements. The Neil Creek Monzogranite is composed of fine- to medium-grained, weakly porphyritic biotite monzogranite.

Contacts between the Neil Creek Monzogranite and the Whitewater Volcanics and Castlereagh Hill Porphyry are complicated by faulting. Field relationships consistent with magma mingling suggest that the Gordons Gorge Granite and Neil Creek Monzogranite are coeval.

Medium-grained, even-textured biotite granitoid (ⱣgPe)

Massive, medium-grained, even-textured biotite monzogranite and syenogranite form a series of discontinuous exposures south and east of Bow Hill. The granitoid outcrops in a roughly northeast-trending zone about 11 km long and 1–2 km wide. Between the Wilson River and the Great Northern Highway the granitoid is more leucocratic and coarser grained and contains numerous bundles of tourmaline, up to 4 cm long, disseminated throughout the rock. Tourmaline-bearing, even-textured biotite syenogranite also occurs as a fault slice east of Aida Vale Mill. Adjacent to the Greenvale Fault and its splays, the syenogranite is mylonitic. Sericitization and some recrystallization characterize all tourmaline-bearing samples.

Even-textured, leucocratic biotite monzogranite also outcrops along the southern edge of the sheet area, west of Castlereagh Creek, and between the Ragged Range and Lake Argyle in the centre of the map sheet.

About 3 km southwest of Bow Hill the contact between even-textured granitoid and the Mount Nyulasy Granite is exposed, but at neither locality are veins or inclusions of one phase present in the other. Northeast of Aida Vale Mill, coarse-grained porphyritic monzogranite of the Gordons Gorge Granite extensively veins a mylonitic, even-textured biotite granitoid.

Medium- to coarse-grained, porphyritic biotite granitoid (ÞgPp)

Medium- to coarse-grained, porphyritic biotite granitoid outcrops along the western edge of Lake Argyle, where it is associated with volcaniclastic rocks of the Whitewater Volcanics. The granitoid contains rounded phenocrysts of K-feldspar up to about 3 cm in diameter. This granitoid also contains inclusions of a porphyritic microgranite and is associated with a porphyry with medium phenocrysts of oval-shaped quartz. The relationship between the granitoid and porphyry is unknown.

CENTRAL ZONE

Tickalara Metamorphics

The Tickalara Metamorphics on LISSADELL consists of high-grade migmatitic pelitic gneiss with rafts of psammitic gneiss, calc-silicate, and mafic granulite (*ÞmTpn*), marble and calc-silicate (*ÞmTpc*), mafic granulite (*ÞmTon*), and amphibolite (*ÞmToa, ÞmTa*), together with mafic and felsic sheet-like meta-igneous bodies (*ÞmTgk, ÞmTgd*). These rocks outcrop in the southern part of LISSADELL.

The rocks forming the Tickalara Metamorphics on LISSADELL show compositional layering, but they do not preserve features that can be interpreted as primary. To the south and southeast, on the MCINTOSH and DIXON 1:100 000 map sheets, high-grade Tickalara Metamorphics can be traced into medium-grade metasedimentary rocks and meta-igneous rocks that do preserve sedimentary and volcanic structures (Plumb et al., 1985; Allen, 1986; Tyler et al., in prep.a,b). On LISSADELL, most of the Tickalara Metamorphics rocks are pelitic metasedimentary and belong to a sequence which, at medium-grade, is interpreted as turbiditic clastic sedimentary rocks (Tyler et al., in prep.a). A sequence which is interpreted as metamorphosed interbedded mafic volcanic rocks, clastic sedimentary rocks, calcareous sedimentary rocks, and carbonates (Tyler et al., in prep.b) outcrops along the southeastern margin of the Mabel Downs Tonalite on TURKEY CREEK (1:100 000) and extends onto the southern part of LISSADELL.

Both the metasedimentary and mafic-dominated sequences contain metamorphosed sheet-like mafic bodies, concordant with layering, and were intruded also by sheet-like granitoid intrusions that were subsequently metamorphosed. These include the Dougalls Tonalite (Ogasawara, 1988; Sheppard et al., 1995), and the Corkwood Tonalite (Tyler et al., in prep.a).

A depositional age for the protoliths of the Tickalara Metamorphics has not been obtained. Nevertheless, detrital zircons from within both medium-grade and high-grade rocks of the turbiditic metasedimentary sequence indicate a maximum age of c. 1865 Ma (Page and Sun, 1994; Page et al., 1995; Bodorkos et al., 1998). A minimum age is provided by the Rose Bore Granite, a deformed and metamorphosed intrusive granitic sheet exposed on MCINTOSH, which has an igneous intrusive age of 1863 ± 3 Ma (Page and Sun, 1994). The stratigraphic relationship between the turbiditic sequence and the mafic-dominated sequence is not clear. However, a minimum age for the deposition of the mafic-dominated sequence is provided by the Panton layered mafic–ultramafic intrusion, dated at 1856 ± 2 Ma (Page et al., 1995), which is intruded into it on MCINTOSH (1:100 000).

Previous workers have assumed the Tickalara Metamorphics to be the metamorphosed medium- to high-grade equivalent of the Halls Creek Group (Dow and Gemuts, 1969; Hancock and Rutland, 1984; Plumb et al. 1985; Allen, 1986), which is exposed within the eastern zone of the Lamboo Complex (Griffin and Tyler, 1992b; Tyler et al., 1995). However, a felsic unit within the Biscay Formation in the lower part of the Halls Creek

Group, has been dated at 1880 ± 3 Ma (Page and Sun, 1994). In the upper part of the Halls Creek Group, felsic and alkali volcanic units were erupted between 1857 and 1848 Ma (Blake et al., 1998). The conformably overlying turbiditic sedimentary rocks contain detrital zircons as young as c. 1847 Ma (Blake et al., 1998). Deformation and metamorphism of the Tickalara Metamorphics took place between c. 1863 and 1850 Ma (Tyler and Page, 1996), and occurred, therefore, while the upper part of the Halls Creek Group was still being deposited. Allen (1986) suggested that quartzose metasedimentary rocks that apparently lie within the Tickalara Metamorphics on the DIXON and TURKEY CREEK 1:100 000 sheets were of Saunders Creek Formation at the base of the Halls Creek Group. However, these outcrops are now considered to be low-strain pods of c. 1800 Ma Red Rock Formation sandstones, strung out within mylonitic rocks along the Halls Creek and Alice Downs Faults (Tyler et al., 1995, 1997, 1998a).

Tyler et al. (1995) concluded that the differences between the Tickalara Metamorphics and the Halls Creek Group were consistent with their deposition in geographically separate tectono-stratigraphic terranes that were juxtaposed during subsequent tectonism.

The petrography and metamorphic history of the Tickalara Metamorphics will be described under the **Hooper Orogeny** and **Halls Creek Orogeny** sections below.

HOOPER OROGENY

The Hooper Orogeny was recognized first in the Hooper Complex of the King Leopold Orogen (Tyler and Griffin, 1993; Griffin et al., 1993; Griffin and Tyler, in prep.), and took place between c. 1870 and 1850 Ma. Tyler et al. (1995) noted the similarities in the geological evolution of the Hooper Complex and the western zone of the Lamboo Complex. Rocks of the Marboo Formation in the Western zone, and the Tickalara Metamorphics within the Central zone, are affected by two early phases of deformation. In the Tickalara Metamorphics, these were interpreted as being related to large-scale easterly directed thrusting (Hancock and Rutland, 1984). More recent zircon and monazite U–Pb ages (Bodorkos et al., 1998) suggest that the tectonic history of the Central zone may be more complex than that recognized by Tyler and Page (1996).

The first deformation (D_1) in the Western zone took place between c. 1870 Ma, the age of the youngest detrital zircons in the Marboo Formation, and c. 1865 Ma, the age of intrusion of the oldest Paperbark supersuite intrusions in the Hooper Complex (Tyler et al., 1995; Tyler et al., in press; Griffin et al., in prep.). The second deformation (D_2) and accompanying metamorphism (M_2) occurred synchronously with the intrusion of the Paperbark supersuite between 1865 and 1850 Ma (Griffin et al., 1993; Tyler et al., 1995; Tyler et al., in press).

In the Central zone a maximum age for the first deformation is provided by the c. 1863 Ma Rose Bore Granite on MCINTOSH (1:100 000), which pre-dates D_1 (Tyler and Page, 1996). A minimum age for D_1 is provided by the c. 1850 Ma granitoids of the Dougalls suite, which post-dated D_1 but pre-dated D_2 (Tyler and Page, 1996). In the Central zone, D_1 bears time constraints similar to D_2 in the Western zone (Table 1).

Sheppard et al. (in press) suggested that the metasedimentary and metabasaltic rocks of the Tickalara Metamorphics in the Central zone represent a c. 1865 Ma intra-oceanic island arc. The Hooper Orogeny may correspond to the accretion of the island arc to the edge of the Kimberley Craton (Myers et al., 1996; Sheppard et al., 1997a). From the relative position of the zones within the Lamboo Complex, subduction of oceanic crust at this time was probably to the southeast (Sheppard et al., in prep.).

Deformation and metamorphism in the Western zone

Where they are unaffected by later contact metamorphism, Marboo Formation rocks on LISSADELL are typically low-grade with mineral assemblages consistent with metamorphism under lower to middle greenschist facies conditions. Phyllitic pelitic and semi-pelitic rocks, representing metamorphosed mudstones and siltstones, consist of quartz, plagioclase, K-feldspar, sericite, chlorite, fine biotite, and iron oxides.

The rocks have been deformed into open-to-tight, upright, small- to medium-scale folds that have subhorizontal to moderately inclined axes. A cleavage is commonly developed parallel to the axial surfaces of these folds (S_2). Large-scale fold closures are not recognized, although this may reflect the subhorizontal nature of the plunge, as larger scale younging reversals are apparent.

Hancock (1991) recognized similar upright medium-scale folds in Marboo Formation rocks on LANSDOWNE. However, downward-facing structures were recognized locally, and Hancock concluded that the upright folds were a second generation (D_2) refolding of an earlier recumbent fold phase (D_1). No D_1 fold closures were identified. Downward-facing strata have not been recognized in the Marboo Formation on LISSADELL, and evidence for D_1 is restricted to the occurrence of a weak layer-parallel foliation (S_1) in some thin sections.

The first deformation and metamorphic event in the Central zone

The oldest deformation that is recognized in the Tickalara Metamorphics on LISSADELL (D_1) produced a pervasive layer-parallel foliation or gneissic banding (S_1) suggesting that the rocks reached a medium to high metamorphic grade during D_1. No D_1 folds were recognized.

The second deformation and metamorphic event in the Central zone

Low to moderate pressure – high temperature metamorphism in the Tickalara Metamorphics has been interpreted as reaching its peak during and after the second deformation and has been previously attributed to the Hooper Orogeny (Griffin and Tyler, 1992b; Tyler et al., 1995; Tyler and Page, 1996). Page and Hancock (1988), using conventional U–Pb zircon dating techniques, obtained an age of 1851 ± 1 Ma from migmatitic pelitic gneiss on the TURKEY CREEK 1:100 000 sheet (recalculated by Page and Sun, 1994, assuming zero-age Pb loss), which they interpreted as the age of peak metamorphism. This date was confirmed by a SHRIMP U–Pb zircon age of 1852 ± 2 Ma from the same rock (Page and Sun, 1994). However, Bodorkos et al. (1998) have obtained a SHRIMP U–Pb zircon age 1845 ± 4 Ma and a conventional U–Pb monazite age of 1845 ± 3 Ma from migmatitic pelitic rocks in the southern part of TURKEY CREEK (1:100 000). These ages suggest that the M_2 event in the Central zone continued after the end of the Hooper Orogeny in the Western zone.

The sheet-like tonalite intrusions of the Dougalls suite intruded the Tickalara Metamorphics at c. 1850 Ma, before and during high-grade metamorphism. The Dougalls suite resembles Phanerozoic tonalites and trondhjemites found in island arcs, or along continental margins related to subduction or subsidiary back-arc spreading (Sheppard et al., 1997a). Metamorphism may reflect an underlying magmatic heat source (Thornett, 1987), possibly related to igneous underplating along an active plate margin (Tyler and Page, 1996).

The second deformation (D_2) has produced extensive small-scale folding of S_1. Folds are tight to isoclinal and at lower grades may produce an axial-planar crenulation cleavage, which is generally oriented subparallel to S_1.

All rocks of the Tickalara Metamorphics on LISSADELL lie within Zone C of Gemuts (1971, fig. 6). Grade varies from upper amphibolite facies at the southeastern margin of the Mabel Downs Tonalite, to granulite facies to its northwest. A coarse gneissic banding is developed parallel to layering in the metasedimentary rocks. Anatectic melts are seen to cross-cut D_2 fabrics, and peak metamorphism is interpreted as post-dating D_2 in the highest grade areas. Plumb et al. (1985) and Thornett (1987) suggested that the metamorphic peak occurred synchronously with D_3. However pre-D_3 mafic dykes either cut across migmatitic structures, or have been interpreted as producing localized re-melting (Oliver et al., 1998), whereas the pre- to syn-D_3 Mabel Downs Tonalite (Griffin and Tyler, 1992b; Tyler and Page, 1996) has given a SHRIMP U–Pb zircon date of 1832 ± 3 Ma, which is about 10 m.y. younger than the age of peak metamorphism.

Blake and Hoatson (1993) and Blake (1994) suggested that the high-grade metamorphism affecting the Tickalara Metamorphics was of contact type related to the emplacement of voluminous gabbro and granite intrusions. Metamorphic grade was described as decreasing away from 'large, post-tectonic mafic intrusions', with migmatites merging into lower grade hornfelses. Such variations in grade or texture were not observed during mapping of the Tickalara Metamorphics on LISSADELL as part of the present survey, nor had they been observed in the highest grade Tickalara Metamorphics by previous workers (Gemuts, 1971; Thornett, 1987). High-grade migmatitic hornfelses are developed within Tickalara Metamorphics adjacent to cross-cutting granitoid and mafic intrusions belonging to the 1835–1805 Ma Sally Downs supersuite (see **Contact metamorphism**). These are characterized by low pressure andalusite–cordierite–K-feldspar assemblages in pelitic rocks, and are superimposed on the older, higher pressure garnet–cordierite–sillimanite–K-feldspar assemblages characteristic of the regional M_2 event.

Metamorphism of sedimentary and mafic volcanic rocks (ꞒmTpn, ꞒmTpc, ꞒmTa)

The Tickalara Metamorphics at the southeastern margin of the Mabel Downs Tonalite belongs to the mafic-dominated sequence of Sheppard et al. (1997b), with migmatitic pelitic rocks and amphibolite intruded by metamorphosed mafic and felsic sheet-like bodies. In the migmatitic pelitic rocks, stromatic (layered) leucosomes have developed parallel to layering (migmatite nomenclature follows Mehnert, 1968 and Ashworth, 1985). Mineral assemblages and textures have been modified by later contact metamorphism by the Mabel Downs Tonalite.

Marble and calc-silicate (*ꞒmTpc*) on LISSADELL are present in a fault slice within the Halls Creek Fault and within amphibolite to the northwest of Lissadell Homestead.

To the northwest of the Mabel Downs Tonalite the Tickalara Metamorphics belongs to the turbiditic sedimentary sequence, and is within the 'granulite zone' of Plumb et al. (1985). The metamorphic rocks are made up predominantly of coarse-grained migmatitic pelitic gneisses containing rafts of psammitic gneiss, calc-silicate, and mafic granulite (*ꞒmTpn*) (Gemuts, 1971; Thornett, 1983, 1987). Mineral assemblages include: cordierite–garnet–biotite–K-feldspar–plagioclase–quartz in rocks exposed around the Bow River copper–nickel prospect, between Turkey Creek and Blackfellow Creek, representative of a garnet–cordierite–K-feldspar zone. A gneissic banding is defined by the alignment of biotite and sillimanite. Garnet occurs as large skeletal porphyroblasts up to 2 cm in diameter, intergrown with quartz and cordierite. Cordierite may also occur as porphyroblasts, or as groundmass crystals. K-feldspar is typically perthitic whereas plagioclase is antiperthitic.

Grade within the migmatitic rocks appears to decrease to the northwest of the Great Northern Highway where garnet is generally absent and cordierite–sillimanite–biotite–K-feldspar–plagioclase–quartz is the dominant assemblage representative of a cordierite–K-feldspar zone. Muscovite is present in some samples intergrown with biotite and cordierite.

The metasedimentary gneisses display structures typical of large-scale anatexis (partial melting) in situ with stromatic (layered), phlebitic (vein), and schollen (raft) structures present, reflecting increasing degrees of partial melting (Mehnert, 1968; Brown, 1994). Rocks showing very high degrees of partial melting (diatexites) contain numerous angular to rounded rafts of restite (refractory material), including calc-silicate, psammite, and mafic granulite (e.g. Gemuts, 1971, plate 11, fig. 2), which can be up to several metres in length.

Thornett (1983, 1987) regarded the onset of large-scale anatexis as being controlled by the reaction

biotite + cordierite + quartz + plagioclase = garnet + K-feldspar + melt,

which is characteristic of Al-poor, Fe–Mg-rich bulk compositions (Thompson, 1982). However, garnet is absent from diatexites within the cordierite–K-feldspar zone and the onset of anatexis was probably controlled by the lower grade reaction

biotite + sillimanite + quartz = cordierite + K-feldspar + melt.

Thornett (1983, 1987) estimated pressures of about 400 MPa at 840°C, increasing to 550 MPa towards the Highway Shear Zone on TURKEY CREEK (1:100 000).

Amphibolite and mafic granulite (ɆmToa, ɆmTon)

The Tickalara Metamorphics includes both mafic and felsic metamorphosed sheet-like bodies, which probably represent original igneous intrusions. The oldest of these are mafic bodies that lie parallel to layering, and pre-date at least D_2 and possibly D_1. Mafic granulite (*ɆmTon*), associated with pods and lenses of chlorite–amphibole rock, is present within the migmatitic gneisses to the northwest of the Mabel Downs Tonalite. The rocks here are medium to coarse grained and typically have polygonal to amoeboid granoblastic textures. They consist of clinopyroxene, orthopyroxene, brown hornblende (typical of the granulite facies; Winkler, 1976, p. 256), plagioclase (andesine–labradorite), quartz, and iron oxide. They are interpreted as the remnants of an early layered mafic–ultramafic intrusion which has been deformed and metamorphosed under granulite facies conditions, and are similar to the rocks forming the Norton intrusion on the TURKEY CREEK 1:100 000 sheet (Hoatson, 1995; Tyler et al., in prep.c).

Smaller pods, lenses, and layers of mafic granulite are present throughout the Tickalara Metamorphics on LISSADELL.

In some outcrops, trains of pods up to 0.5 m across of a massive, fine- to medium-grained intermediate rock are associated with patches of leucosome within the migmatites. The rock consists of biotite, epidote, and quartz, with minor amounts of amphibole, chlorite, titanite, and iron oxide. Intergrowths of epidote, chlorite, and plagioclase pseudomorph plagioclase phenocrysts. According to Oliver et al. (1998) the rock may represent synmetamorphic, dyke-like intrusions into migmatite that was probably not yet frozen, triggering remelting and the formation of irregular sheaths of leucosome around the intrusions, veins into the metasedimentary country rocks, and back veining into the intrusion.

More continuous lenses and folded dykes of amphibolite (*ȆmToa*) are present throughout the migmatitic gneisses. These bodies are typically foliated and consist of pale-green hornblende, plagioclase (andesine), quartz, and epidote, with minor amounts of biotite, and iron oxides rimmed by titanite. They are lower grade than the mafic granulites and locally cut across migmatite structures, and are interpreted as a suite of post-M_2 mafic dykes.

Outcrops of foliated medium-grained, clinopyroxene-bearing amphibolite lie to the north and west of the Pitt Range. These rocks have a polygonal granoblastic texture which, together with the presence of brown hornblende, suggests that they may have been metamorphosed under granulite facies conditions. The plagioclase is andesine.

Metamorphosed granitoid rocks (ȆmTgk, ȆmTgd)

Sheet-like granitoid intrusions belonging to the Dougalls suite (Ogasawara, 1988; Sheppard et al., 1995; Sheppard et al., in prep.) outcrop within the Tickalara Metamorphics towards the southern margin of LISSADELL. They were metamorphosed and recrystallized during M_2, and developed an S_2 fabric. The intrusions have therefore been included within the Tickalara Metamorphics.The Dougalls Tonalite has given a SHRIMP U–Pb zircon age of 1848 ± 2 Ma (Tyler and Page, 1996), and intrusion must have taken place pre- to syn-D_2/M_2.

The Corkwood Tonalite (*ȆmTgk*) forms a 100 to 150 m-wide, steeply northwesterly dipping concordant sheet intruded into amphibolite at the eastern margin of the Mabel Downs Tonalite. The tonalite is medium-grained, has been foliated and recrystallized, and consists of biotite, plagioclase (andesine), and quartz, with minor amounts of epidote, calcite, apatite, zircon, and iron oxide.

The Dougalls Tonalite (*ȆmTgd*) consists mainly of medium-grained foliated and recrystallized biotite–hornblende or biotite–hypersthene metatonalite to metagranodiorite, together with sheets of more leucocratic biotite metamonzogranite. The main area of outcrop is to the west of the Great Northern Highway, 4.5 km north of where it crosses the southern margin of LISSADELL. Here the intrusion is complex, with a number of sheets present that are separated locally by patches and screens of migmatitic gneiss. Two mineral assemblages are recorded: biotite, hypersthene, plagioclase, K-feldspar, and quartz; or biotite, blue-green edenitic amphibole, epidote, plagioclase, K-feldspar, and quartz. In both assemblages the plagioclase is antiperthitic andesine, whereas the K-feldspar is perthitic microcline and may be altered to myrmekite. Minor amounts of titanite, allanite, zircon, and iron oxides are also present.

Relict clinopyroxene can occur within the amphibole-bearing assemblage, where it is rimmed by amphibole. Intergrowths of amphibole, green biotite, epidote, and quartz up to 5 mm in length pseudomorph what were probably magmatic ?pyroxene phenocrysts. Sheppard et al. (1995) also suggested a magmatic origin for orthopyroxene, which survived as a result of equilibration following intrusion into country rocks that were at granulite facies grade.

PALAEOPROTEROZOIC SALLY DOWNS SUPERSUITE

The Sally Downs supersuite includes all 1835–1805 Ma granitoid and massive gabbro in the Lamboo Complex, the bulk of which outcrops in the Central zone. Like the Paperbark supersuite, evidence for coeval mafic and felsic magmas is widespread in the Sally Downs supersuite (Blake and Hoatson, 1993; Sheppard, 1996).

Biotite gabbro (*₽gSob*)

Biotite gabbro is restricted to several small (<0.10 km^2) exposures east of the Great Northern Highway in the southern part of LISSADELL. However, the presence of extensive black soil plains in this part of the sheet suggests that biotite gabbro underlies much of this area. The contact relationships of the biotite gabbro are not known. On LISSADELL, the biotite gabbro unit includes subophitic olivine norite and metagabbro. Oikocrysts of pyroxene (or hornblende) may reach 1 cm in diameter. Biotite, which replaces hypersthene, forms up to 5% of some samples.

Mingled gabbro–granite (*₽gSog*)

On LISSADELL, mingled gabbro–granite outcrops along the Great Northern Highway at the southern edge of the sheet area. Rock types consist of fine-grained biotite gabbro and quartz diorite veined by biotite monzogranite, biotite-bearing pegmatite, and quartz. Mingled gabbro–granite in the Sally Downs supersuite on MOUNT REMARKABLE (1:100 000) is fully described by Sheppard et al. (1997b).

Kevins Dam Monzogranite (*₽gSkd*)

The Kevins Dam Monzogranite is a large (>300 km^2) intrusion, which outcrops mainly in the northeastern part of MOUNT REMARKABLE (1:100 000) to the southwest. On LISSADELL it is restricted to about 15–20 km^2 on the southern edge of the sheet area around Castlereagh Hill. The Kevins Dam Monzogranite is composed of pale-grey or pinkish-grey, medium-grained, even-textured or weakly porphyritic biotite monzogranite. This unit intruded the Castlereagh Hill Porphyry about 2 km southwest of Castlereagh Hill (at 16°59'45"S, 127°56'39"E).

Mabel Downs Tonalite (*₽gSmd*)

The Mabel Downs Tonalite (formerly Mabel Downs Granodiorite of Dow and Gemuts, 1969) is a northeasterly trending sheet-like intrusion, about 75 km long and several kilometres thick, the bulk of which outcrops on the TURKEY CREEK 1:100 000 sheet (Tyler et al., in prep.a). The intrusion is strongly foliated and was apparently emplaced in a shear zone. A U–Pb SHRIMP age of 1832 ± 3 Ma was obtained for a sample 12 km south-southwest of Warmun (Page et al., in prep.).

Only the northern end of the Mabel Downs Tonalite outcrops on LISSADELL, between Fargoo Creek and Turkey Creek at the southern end of the map sheet. This unit consists of moderately to strongly foliated, medium- to fine-grained hornblende–biotite tonalite and quartz diorite with abundant angular inclusions of fine-grained mafic rock. Along its eastern edge the Mabel Downs Tonalite intruded amphibolite, calc-silicate and metatonalite of the Tickalara Metamorphics. In places the tonalite is tectonically interleaved with the older metamorphic rocks. Beside Turkey Creek, a 20–50 cm-wide dyke of medium-grained, weakly porphyritic biotite granodiorite, possibly belonging to the Violet Valley Tonalite (*ЕgSvm*), cuts the Mabel Downs Tonalite. The tonalite is also cut by numerous pegmatite veins and dykes.

Angular, fine-grained mafic inclusions are ubiquitous in the Mabel Downs Tonalite. They are elongate within the foliation, and are typically 5–10 cm wide and 10–50 cm long, although inclusions over 1 m long are common. They may be of two types: fine-grained, even-textured biotite-rich inclusions; and fine- to medium-grained, weakly porphyritic, quartz diorite and tonalite inclusions.

Maggotty Springs Monzogranite (*ЕgSms*)

The Magotty Springs Monzogranite outcrops as a narrow sheet-like body intruding the Tickalara Metamorphics to the west of the Great Northern Highway, towards the southern margin of the map sheet, where it is adjacent to outcrops of mingled gabbro–granite (*ЕgSob*). The unit consists of coarse-grained, weakly porphyritic biotite monzogranite, and medium-grained, even-textured biotite granodiorite.

The monzogranite contains some hornblende as fine-grained intergrowths apparently replacing an earlier phase intergrown with biotite. Epidote is also present, as are muscovite and myrmekitic intergrowths that replace K-feldspar. Muscovite intergrown with epidote and calcite replaces feldspar in the granodiorite. Apatite, zircon, and an iron oxide phase are present in both rock types.

Violet Valley Tonalite (*ЕgSvv, ЕgSvm*)

The Violet Valley Tonalite (Dow and Gemuts, 1969) is a large, northeast-trending, sheet-like intrusion composed of several different phases. The intrusion is about 50 km long and between 1.5 and 7 km wide. The northern two-thirds of the intrusion outcrops on LISSADELL, with the remainder on DIXON RANGE to the south. The Violet Valley Tonalite forms rounded, bouldery hills with a relief of about 100 m.

The bulk of the Violet Valley Tonalite is composed of massive, fine- to medium-grained, weakly porphyritic biotite granodiorite and minor tonalite (*ЕgSvv*). However, much of the northern half of the intrusion is composed of foliated, medium-grained, weakly porphyritic biotite granodiorite and biotite–hornblende tonalite (*ЕgSvm*). Gemuts (1971) previously assigned much of this unit (his 'orthopyroxene-rich granodiorite', p. 54) to the Mabel Downs Tonalite. In the field the two units constituting the Violet Valley Tonalite can be difficult to differentiate from each other, but the medium-grained granodiorite (*ЕgSvm*) is slightly more felsic that the somewhat finer grained *ЕgSvv*, and is generally foliated. On aerial photographs the fine- to medium-grained granodiorite (*ЕgSvv*) is distinguished from other granites by its dark-grey to black pattern.

The Violet Valley Tonalite intruded the Tickalara Metamorphics, although shearing complicates the contacts locally. Angular inclusions of migmatite are common close to the margins of the various intrusions. The Violet Valley Tonalite is cut by widespread veins and dykes of fine- to medium-grained, even-textured leucocratic biotite granite. Immediately southeast of where the Great Northern Highway crosses the Bow River, foliated medium-grained granodiorite (*ЕgSvm*) is intruded by a fine-grained, inclusion-rich porphyritic tonalite of limited extent. The tonalite is finer grained and more mafic than the granodiorite, but otherwise has the same mineral assemblage, suggesting that it may be another intrusive phase of the Violet Valley Tonalite.

Fine- to medium-grained, weakly porphyritic biotite granodiorite and minor tonalite (ꝒgSvv)

The weakly porphyritic biotite granodiorite and minor tonalite phase contains plagioclase phenocrysts up to 4 mm long that compose less than 5% of most samples. The groundmass consists of subhedral plagioclase, anhedral quartz, biotite, and minor microperthite and epidote. Traces of a green hornblende are also present in some samples. Phenocryst and groundmass plagioclase has mottled cores and strong oscillatory zoning. Accessory minerals include titanite, apatite, zircon, and allanite. Epidote is associated with alteration of plagioclase, pseudomorphs of magnetite, and forms very fine grained granular crystals along the grain boundaries of biotite.

Foliated, medium-grained, weakly porphyritic biotite granodiorite and biotite–hornblende tonalite (ꝒgSvm)

The foliated, medium-grained granodiorite resembles the Mabel Downs Tonalite; both are foliated and have a similar pitted texture on weathered surfaces, but the granodiorite of the Violet Valley Tonalite is more felsic, and contains fewer mafic inclusions than the Mabel Downs Tonalite.

The foliated, medium-grained granodiorite consists of several intrusions that extend from southwest of Radigans Yard to the Bow River. The intrusions range from small plugs ($\leq$0.1 km^2) up to a stock that outcrops over 25 km^2 southwest of Radigans Yard. Around Bow River Homestead and down to the southern edge of LISSADELL, biotite granodiorite is the sole rock type, whereas around the Great Northern Highway and across to the Bow River, only biotite–hornblende tonalite is present.

The granodiorite and tonalite are even textured, medium grained, and subhedral granular, with weakly porphyritic varieties present locally. The tonalite is composed of plagioclase (andesine), quartz, biotite, hornblende, and epidote, with accessory titanite, apatite, zircon, and rare allanite. The granodiorite has only traces of hornblende, but it contains more quartz, small amounts of perthite, and allanite is a prominent accessory. The hornblende (a=greenish yellow, b=blue–green, g=olive) is probably hastingsite. Hornblende is partly replaced by intergrowths of biotite and epidote. Epidote and titanite pseudomorphs of magnetite and ilmenite respectively are associated with alteration of plagioclase. Allanite crystals are commonly rimmed by epidote.

Wesley Yard Monzogranite (*ꝒgSwy*)

The Wesley Yard Monzogranite is composed of medium- to coarse-grained, variably porphyritic biotite monzogranite and minor granodiorite. This unit is distinguished from the Mount Nyulasy Granite by a finer grainsize, fewer phenocrysts, the presence of tabular and rounded K-feldspar phenocrysts, and widespread mafic inclusions. Most of the intrusion is weakly foliated, but along the southeast margin it may be strongly foliated.

About 4 km south-southeast of Wesley Yard, the Wesley Yard Monzogranite veins and encloses mafic granulite of the Tickalara Metamorphics. About 1 km west of Wesley Yard the monzogranite intrudes dark-grey, medium-grained tonalite of the Dinner Creek Tonalite. Locally along its southeast margin, contacts between the Wesley Yard Monzogranite and Violet Valley Tonalite are marked by shearing.

Most samples of the Wesley Yard Monzogranite have microperthite phenocrysts up to 2 cm long in an anhedral granular groundmass of medium-grained andesine, quartz, biotite, and microperthite, with accessory magnetite (now epidote), ilmenite (now titanite), apatite, and zircon. Some samples contain trace amounts of blue-green hastingsite. In addition to angular inclusions of country rock near its margins, the Wesley Yard Monzogranite contains widespread mafic inclusions.

Medium- to coarse-grained porphyritic granite (*ɆgSp*)

Medium- to coarse-grained porphyritic monzogranite and syenogranite outcrops over about 15 km^2 along the southern edge of the map sheet. The granite outcrops much more extensively on DIXON RANGE to the south. Owing to the inaccessible nature of the unit it has not been formally defined. It is generally characterized by a more subdued response on radiometric images relative to the Kevins Dam Monzogranite. On DIXON RANGE the porphyritic granitoid intruded fine- to medium-grained granodiorite of the Violet Valley Tonalite. The porphyritic granitoid contains 30% or more rounded phenocrysts of microcline up to 2.5 cm in diameter.

Contact metamorphism

High-grade contact aureoles are developed in both the Marboo Formation (*Ɇmh*) and the Tickalara Metamorphics (*ɆmTh*) adjacent to the granitoids and mafic intrusions of the Sally Downs supersuite (Thornett, 1987; Sheppard et al., 1997b; Tyler et al., in prep.a,b). In Marboo Formation rocks on LISSADELL, biotite–muscovite(–cordierite)–quartz hornfelses grade into migmatitic cordierite–andalusite/sillimanite–biotite–muscovite–K-feldspar–plagioclase–quartz hornfelses at the contacts. The migmatitic hornfelses are well exposed where the Great Northern Highway crosses the Bow River. Here they are medium grained, and typically display chaotic schollen (raft) structures, characteristic of anatexis in contact aureoles (Pattison and Harte, 1988). They are distinctively 'spotted' with cordierite porphyroblasts reaching up to 1 cm in diameter, which are sieved with inclusions of fine biotite, quartz, and an opaque mineral. The lower grade hornfelses are well exposed along the highway to the south, and are spotted with pinitized poikiloblasts of cordierite up to 3 mm across. Garnet is not present in the hornfelses, and the evidence of melting in the andalusite-bearing rocks implies very low pressure metamorphism (<100 MPa, Thompson, 1982, fig. 8), considerably lower than peak M_2 conditions.

In the Tickalara Metamorphics, overprinting of the regional M_2 metamorphism by contact metamorphism typically produces recrystallization near the contacts and the development of hornfelsic textures. Where melting has occurred the regional textures are completely destroyed. Andalusite and sillimanite may be present, but garnet is absent.

PALAEOPROTEROZOIC SALT LICK CREEK LAYERED INTRUSION (*Ɇa*)

The Salt Lick Creek layered intrusion is shown on the accompanying map as 'undivided layered mafic–ultramafic intrusions'. The intrusion is an oval-shaped body about 7 km^2 in area that is centred about 6 km south-southwest of the Argyle diamond mine. The Salt Lick Creek intrusion invaded low-grade metasedimentary rocks of the Marboo Formation. Wilkinson et al. (1975) record the Salt Lick Creek intrusion being locally veined by granitoid; the granitoid probably belongs to the Violet Valley Tonalite of the 1835–1805 Ma Sally Downs supersuite.

Wilkinson et al. (1975) divided the intrusion into a Basal Zone, about 360 m thick, and a Main Zone, 600 m or more in thickness. The Basal Zone is composed of olivine–plagioclase cumulates, mainly troctolite, olivine gabbronorite, anorthositic gabbro and plagioclase-bearing dunite. The Main Zone consists of plagioclase–orthopyroxene cumulates, namely, anorthositic gabbro, anorthosite, norite, and gabbronorite. Mild rhythmic layering is present in the three members constituting the Basal Zone. Samples from the intrusion show widespread serpentinization of olivine, and patchy alteration of orthopyroxene to serpentine and of plagioclase to zoisite and prehnite.

THE HALLS CREEK OROGENY

Deformation and metamorphism during the Halls Creek Orogeny (Tyler and Page, 1996) have affected rocks of the Tickalara Metamorphics in the central zone, and the synorogenic granitoids and intrusive mafic rocks of the Sally Downs supersuite. Two phases of deformation have taken place with the oldest (D_3) having affected the Tickalara Metamorphics and the Mabel Downs Tonalite. This deformation did not affect the Sally Downs Tonalite on the McIntosh 1:100 000 sheet (Tyler et al., in prep.b) or the biotite gabbro intrusions, including those on Lissadell, and must therefore have taken place between c.1830 Ma (the age of the Mabel Downs Tonalite) and c. 1820 Ma (the age of the Sally Downs Tonalite). The youngest deformation (D_4) post-dates the Sally Downs Tonalite on McIntosh (1:100 000). An upper age limit for D_4 is not well defined, although from relationships on the Turkey Creek and Dixon 1:100 000 sheets to the west (Tyler, in prep.a,c) it appears to pre-date the Red Rock Formation, probably deposited some time between c. 1820 and c. 1790 Ma. The Halls Creek Orogeny may be the result of a collision between the combined central and western zones, and the eastern zone, representing final suturing of the Kimberley Craton onto the North Australian Craton to the east by c. 1800 Ma (Tyler and Page, 1996).

The third deformation and metamorphism event in the Central zone

The D_3 deformation has produced small- to large-scale, close to isoclinal folds. On Lissadell these are best developed within the Tickalara Metamorphics to the west of the Great Northern Highway, where folds of layering and of amphibolite dykes can be seen. However, owing to the lack of persistent markers, the folds can not be followed for any distance and the regional-scale structure on Lissadell is not clear. On the Mount Remarkable and Turkey Creek 1:100 000 sheets, large-scale D_3 folds were picked out by mafic granulite layers (Sheppard et al., 1997b; Tyler et al., in prep.a). The folds were refolded by D_4 and their original pre-D_4 orientation was north-northwesterly, plunging to the south-southeast. A weak crenulation cleavage may be developed axial planar to the D_3 folds.

The Mabel Downs Tonalite has been recrystallized under epidote–amphibolite facies metamorphic conditions with the minerals epidote, hornblende, biotite, plagioclase (oligoclase–andesine), quartz, and titanite present. The prominent foliation that dips moderately to steeply to the northwest within the Mabel Downs Tonalite is regarded as S_3, having formed at the same time as epidote–amphibolite facies metamorphism (M_3).

The fourth deformation and metamorphism event in the Central zone

Northeasterly oriented veins and dykes of granitic material intrude post-D_3 granitoid and biotite gabbro, and are strongly foliated. The foliation is regarded as S_4 and the rocks have been recrystallized under upper greenschist to lower amphibolite facies conditions, with the minerals biotite, garnet, epidote, plagioclase, K-feldspar, and quartz showing blastomylonitic textures. Shear-sense indicators, including tails on feldspar porphyroclasts and S–C fabrics, indicate transpressive sinistral and northwest-side-up movement.

On the MOUNT REMARKABLE and TURKEY CREEK 1:100 000 sheets, steeply northeasterly plunging open-to-tight folds have refolded the D_3 folds to produce Type 3 hooked fold-interference patterns (Sheppard et al., 1997b; Tyler et al., in prep.a). Locally, folding is developed in a similar orientation on LISSADELL refolding D_3 folds. A crenulation cleavage (S_4) that is parallel to the axial surfaces of these folds may be present.

PALAEOPROTEROZOIC SPEEWAH BASIN

Previously, the Speewah Group and the overlying Kimberley Group have been regarded as having been deposited within a single 'Kimberley Basin'(e.g. Dow and Gemuts, 1969; Plumb and Gemuts, 1976). However, it has become apparent that there is a significant unconformity or disconformity between the two groups (Griffin et al., 1993), and that they had different palaeogeographies, representing two discrete depositional basins in different tectonic settings. The Speewah Group was deposited in the the Speewah Basin, and was restricted to the west of the Greenvale and Dunham Faults. The term 'Kimberley Basin' is restricted now to the depositional basin of the Kimberley Group, which may have extended farther to the east.

A SHRIMP U–Pb zircon age of 1834 ± 3 Ma has been obtained from a felsic volcanic unit within the Valentine Siltstone of the Speewah Group on LISSADELL (Page and Sun, 1994), indicating that deposition took place during the early stages of the Halls Creek Orogeny. Plumb et al. (1981) interpreted the Speewah Group as a transgressive–regressive cycle with fluvial sands passing into alternating or interfingering fluvial and shallow-marine facies and then back into fluvial sands. Palaeocurrent direction indicators determined from cross-bedding in the Speewah Group (Gellatly et al., 1970) suggest that sediment was derived from the elevated and tectonically active Lamboo Complex to the northeast and east, and then transported along a fault-bounded trough.

A SHRIMP U–Pb age of c. 1790 Ma obtained from a granophyric unit of the Hart Dolerite on LENNARD RIVER places a minimum age on the deposition of the Kimberley Group (Page, R. W., 1995, pers. comm.). The Kimberley Group was probably deposited in a similar regressive–transgressive shallow-marine to fluvial environment as the Speewah Group; palaeocurrents indicate sediment transport was from the north and north-northwest (Gellatly et al., 1970). Deposition probably post-dated the Halls Creek Orogeny.

Gellatly et al. (1975) defined the boundary between the Speewah and Kimberley Groups as the conformable upper contact of the Luman Siltstone. However, on LANSDOWNE, Williams (1969, unpublished report, quoted in Griffin et al., 1993) identified an unconformity, marked by a conglomerate bed, within the 'King Leopold Sandstone'. The conglomerate was recognized throughout the East Kimberley by Dow et al. (1964) and Gellatly et al. (1970, 1975). The boundary between the two groups is now defined as the base of the unconformity, and that part of the old 'King Leopold Sandstone' beneath the unconformity is now named the Bedford Sandstone (Griffin et al., 1993). The King Leopold Sandstone is the basal unit of the Kimberley Group, and the Bedford Sandstone the uppermost unit of the Speewah Group (Griffin et al., 1993, fig. 5).

SPEEWAH GROUP

Folded and faulted sedimentary rocks of the Speewah Group outcrop over a wide area in the northwestern part of LISSADELL. Here, the succession consists of (in ascending order)

O'Donnell Formation, Tunganary Formation, Valentine Siltstone, Lansdowne Arkose, Luman Siltstone, and Bedford Sandstone. The Speewah Group has an estimated total thickness of 1500 m and is intruded extensively by the Hart Dolerite.

O'Donnell Formation (*ᵽSn*)

On LISSADELL, the O'Donnell Formation unconformably overlies the Whitewater Volcanics with a slight angular discordance, whereas the contact with the overlying Tunganary Formation is conformable. The formation is about 260 m thick and consists of a lower quartz sandstone and siltstone member outcropping as prominent strike ridges and cuestas, and an upper siltstone and thin-bedded sandstone member, which in places has been eroded to form a narrow valley.

The lower sandstone member of the O'Donnell Formation is composed of white, pale-grey or pale-purple, medium- to very coarse grained, thin- to thick-bedded quartz sandstone and lithic quartz sandstone, interbedded with minor siltstone. The proportion of fine-grained rocks increases upwards, forming a transitional contact with the overlying siltstone-dominated member. The thicker sandstones may show cross-stratification whereas thinner beds are either parallel-laminated or ripple cross-laminated. Locally, the base of the lower sandstone member is marked by a pebbly quartz sandstone or conglomerate unit up to 5 m thick.

Siltstones in the upper member are typically green-grey in colour and micaceous. Most are parallel-laminated with cross-lamination in places. The interbedded sandstones are thin bedded, fine to medium grained and quartz-rich. Some sandstone beds are normally graded. Internal structure generally consists of parallel-lamination or cross-lamination, either separately or in combination.

Tunganary Formation (*ᵽSt*)

The Tunganary Formation conformably overlies the O'Donnell Formation, and is itself conformably overlain by the Valentine Siltstone. The Tunganary Formation is about 290 m thick and consists of lower and upper sandstone units separated by a prominent siltstone.

The sandstone-dominated units consists of fine- to very coarse grained, thin- to thick-bedded feldspathic sandstone and lithic quartz sandstone, with thin interbeds of quartz sandstone, pebbly lithic sandstone and siltstone. The beds display a wide range of sedimentary structures including isolated or stacked sets of small- to medium-scale trough cross-strata, planar or undulatory parallel lamination, scour and fill, and ripple cross-lamination. In addition, symmetric or asymmetric ripples are preserved on some bedding surfaces. Sparse palaeocurrent data from trough cross-stratified units in the upper part of the formation suggest that the principal sediment transport direction was towards the north.

The central siltstone unit comprises planar or undulatory parallel-laminated siltstone interbedded with thin- to medium-bedded quartz sandstone.

Valentine Siltstone (*ᵽSv*)

The Valentine Siltstone is a poorly exposed unit, which is conformable with the underlying Tunganary Formation. The siltstone is about 75 m thick over much of LISSADELL and is intruded extensively by the Hart Dolerite. The Valentine Siltstone is composed of green to grey, thinly bedded, siltstone and sandstone, with thin interbeds of dacite to rhyolite tuff, and tuffaceous siltstone. Siltstones and sandstones are typically parallel laminated or ripple cross-laminated, although thicker bedded sandstones may show trough cross-stratification or undulatory lamination. A black, crystal-poor rhyolite tuff locally marks the base of the formation on LISSADELL. This volcanic rock was dated at 1834 ± 3 Ma by Page and Sun (1994).

Lansdowne Arkose (*ⱣSo, ⱣSoa*)

The Lansdowne Arkose, which conformably overlies the Valentine Siltstone, typically forms low, finely terraced escarpments giving rise to a ribbed appearance on aerial photographs. The formation is extensively intruded by the Hart Dolerite over much of the map sheet. The succession is about 400 m thick and consists of pink to pale-purple, buff and pale-grey, medium- to coarse-grained feldspathic sandstone and arkose, with interbedded fine- to coarse-grained quartz sandstone and micaceous siltstone. Throughout most of the map sheet the formation shows a threefold subdivision into lower and upper sandstone members separated by a mixed siltstone and sandstone unit (*ⱣSoa*). The lower and upper sandstone units are dominated by small- to large-scale, high- or low-angle trough cross-stratification, present in single or multiple-stacked sets. Bedding surfaces are marked locally by straight-crested, symmetrical or asymmetrical ripples and megaripples. Palaeocurrent data from medium- to large-scale cross-stratification indicate that sediment transport was principally from the northeast and east and lend support to the earlier findings of Gellatly et al. (1970) and Plumb and Gemuts (1976). Local bipolar palaeocurrent trends recorded from small-scale cross-stratification in beds near the top of the lower sandstone member are thought to reflect the influence of marine reworking at this level.

Luman Siltstone (*ⱣSl*)

The Luman Siltstone is conformable with the underlying Lansdowne Arkose. This siltstone is a recessive unit about 75 m thick, consisting of purple-grey to green-grey siltstone and shale, with thin interbeds of feldspathic sandstone.

Bedford Sandstone (*ⱣSb*)

On the first edition of DIXON RANGE (Dow and Gemuts, 1967) and on LANSDOWNE (Gellatly et al., 1975) the recently defined Bedford Sandstone (Griffin et al., 1993) was included in the King Leopold Sandstone. The Bedford Sandstone is conformable with the underlying Luman Siltstone, and its contact with the overlying King Leopold Sandstone in the east Kimberley region is apparently disconformable. The Bedford Sandstone is about 300 m thick and is composed almost entirely of cliff-forming, white, pale-pink and buff, medium- to coarse-grained quartz sandstone with abundant cross-bedding.

PALAEOPROTEROZOIC TEXAS DOWNS BASIN

The Texas Downs Basin (Tyler et al., 1997) corresponds to the present day outcrop of the Texas Downs Formation, which is exposed in isolated fault slices immediately to the east of the Halls Creek Fault on the southeastern part of the map sheet and includes units mapped either as Red Rock Beds or Mount Parker Sandstone on the first edition of LISSADELL (Plumb, 1968). The lower contact of the Texas Downs Formation is not exposed on LISSADELL, but on neighbouring TURKEY CREEK (1:00 000) the formation rests with angular unconformity on either the Red Rock Formation or the McHale Granodiorite of the Lamboo Complex (Tyler et al., in prep.a). The upper contact of the Texas Downs Formation is unconformably overlain by the Mount Parker Formation.

The age of the Texas Downs Formation is very poorly constrained. The older age limit is fixed by a U–Pb SHRIMP zircon date of 1827 ± 3 Ma from the McHale Granodiorite (Page et al., in prep.), which is unconformably overlain by the Texas Downs Formation on the TURKEY CREEK 1:100 000 sheet (Tyler et al., in prep.a). The younger age limit is speculative since the age of the Mount Parker Formation and the immediately overlying succession is unknown. On purely lithological grounds, there is a broad similarity between the Texas Downs, Red Rock, and Revolver Creek Formations along the eastern margin of the Halls Creek Orogen and the lower part of the Kimberley Group farther to the west. Although there is little evidence at present to suggest that all these units are direct correlatives, they may be broadly coeval, and were possibly deposited as a result of similar regional tectonic controls. Such a relationship would suggest the younger age limit for these units is c. 1790 Ma, this being the age of the Hart Dolerite which intrudes the Kimberley Group on LENNARD RIVER (Page, R. W., 1995, pers. comm.).

TEXAS DOWNS FORMATION (*Ex, Exa*)

The Texas Downs Formation is exposed in three fault-bounded slices in the southeastern part of LISSADELL between 17°00'00"S, 128°19'24"E and 16°41'45"S, 128°29'19"E. The formation has an estimated minimum thickness of 1 km and consists of sandstone, siltstone, and minor conglomerate. Basaltic units lie interbedded with the sedimentary rocks on TURKEY CREEK (1:100 000); however, these volcanic rocks have not been recognized on LISSADELL.

In the Pitt Range, the Texas Downs Formation can be subdivided into lower and upper sandstone units that are separated by an argillaceous member (*Exa*). The sandstone units consist of fine- to coarse-grained quartz sandstone, feldspathic quartz sandstone, and lithic quartz sandstone. These are interbedded with lesser amounts of siltstone, pebbly sandstone and conglomerate. Sandstones are thin to thick bedded and display a variety of stratification types. Thin- to medium-bedded sandstones show a predominance of parallel-lamination, together with current- and wave-ripple cross-lamination. Medium- to thick-bedded sandstone and pebbly sandstone show planar parallel to undulatory lamination alternating with sets of trough cross-stratification. Many sandstone bedding surfaces display wave- or current-ripple bedforms. Limited palaeocurrent data from the sandstone units suggest a bipolar current regime with flow towards the north-northeast and south-southwest. Parallel-stratified pebble- to cobble-conglomerate is exposed in the upper sandstone unit. Clasts are well rounded and consist largely of silicified sandstone and vein quartz.

PALAEOPROTEROZOIC REVOLVER CREEK BASIN

The Revolver Creek Basin equates to the current outcrop of the Revolver Creek Formation and is confined to the east central part of LISSADELL, where it is exposed in three small areas between the Glenhill – Bow River and Carr Boyd – Halls Creek fault systems. The Revolver Creek Formation was defined by Dow et al. (1964) as a sequence of sandstone, siltstones, and basic volcanic rocks that underlies the Hensman Sandstone of the Carr Boyd Group with angular unconformity. The Revolver Creek Formation unconformably overlies rocks of the Lamboo Complex including the Marboo Formation, Whitewater Volcanics, and the Castlereagh Hill Porphyry. These relationships indicate that the maximum age for the Revolver Creek Formation is around 1855 Ma, this being the younger age limit for the Castlereagh Hill Porphyry. The minimum age for the Revolver Creek Formation is about 1200 Ma, based on Rb–Sr whole rock and K–Ar phlogopite data from the Argyle lamproite, which intrudes the Carr Boyd Group (Pidgeon et al., 1989; Boxer et al., 1989). As noted earlier in the description of the Texas Downs Formation, there is a broad lithological similarity between the Texas Downs, Red Rock, and Revolver Creek Formations, and the lower part of the Kimberley Basin succession. Although these units are unlikely to be direct correlatives, they may be broadly coeval with a minimum age of around 1800 Ma.

REVOLVER CREEK FORMATION (*Ɇr, Ɇrb*)

On LISSADELL, the Revolver Creek Formation is exposed near the Argyle mine site around 16°43'22"S, 128°23'24"E, east of the Ragged Range around 16°22'52"S, 128°24'36"E, and west of Ulysses Bay around 16°15'00"S, 128°36'00"E. West of the Argyle mine site the formation unconformably overlies metasedimentary rocks of the Marboo Formation and granitoids of the Paperbark supersuite, and is overlain unconformably by the Hensman Sandstone. The succession has a minimum thickness of 1200 m and consists of a 20 m-thick lower sandstone and conglomerate unit overlain by a 700 m-thick composite member consisting of dolerite and basalt (*Ɇrb*) interbedded with siltstone and sandstone. This central member is in turn overlain by about 500 m of interbedded siltstone and sandstone.

The base of the lower sandstone unit is marked by a thin, lenticular cobble conglomerate containing rounded clasts of vein quartz, silicified sandstone, chert, schist, and jaspillite. Palaeocurrents from the overlying trough cross-stratified lithic quartz sandstones indicate a derivation from the northeast.

The top of the lower sandstone unit is sharply overlain by a chocolate-coloured siltstone and sandstone that marks the base of the middle composite member. This middle member consists of massive or amygdaloidal basalt and dolerite, interbedded with subordinate thin- to thick-bedded siltstone and sandstone. Basalt flows are up to 35 m thick and are seen in thin section to be strongly altered (Dow et al., 1964). The ferromagnesian minerals, probably pyroxene, are altered to chlorite and the feldspar to albite and calcite. Spinel is a common accessory mineral. The sandstones are generally cream or pink, fine- to coarse-grained feldspathic quartz sandstones or arkose. Locally, however, they are coarse grained or pebbly and contain layers of quartz amygdale clasts. The internal structure of the sandstones is variable and consists of combinations of low-angle cross-stratification, undulatory and planar parallel stratification, and ripple cross-lamination. Wave-ripple bedforms are preserved on some sandstone surfaces.

The upper unit of the Revolver Creek Formation contains a higher proportion of purple or chocolate micaceous siltstone than the lower and middle parts of the succession. These argillaceous rocks are interlayered with very thin to thick-bedded feldspathic quartz sandstone, arkose, and lithic quartz sandstone The siltstone units, which are often poorly exposed, show a combination of parallel lamination and current- or wave-ripple cross-lamination. Sandstones are fine to coarse grained and are present as thin interbeds or medium- to thick-bedded, amalgamated units ranging in thickness up to 25 m. Many beds contain abundant siltstone clasts. Thin-bedded sandstones often show sharp erosional bases, local normal grading, and a combination of parallel lamination and current- or wave-ripple cross-lamination. The thicker sandstones are dominated by high- or low-angle trough cross-stratification, planar or undulatory stratification and both wave- and current-ripple cross-lamination. Palaeocurrent data, taken from units that display stacked sets of trough cross-stratification, suggest that sediment transport was mostly from north to south.

West of Ulysses Bay, the Revolver Creek Formation is dominated by medium- to coarse-grained sandstone with minor basalt, siltstone, and conglomerate. Palaeocurrent data indicate that sediment transport was towards the west in the north eastern part of the outcrop and towards the north and east along its southern margin.

PALAEOPROTEROZOIC KIMBERLEY BASIN

Sedimentary rocks deposited in the Kimberley Basin, and which make up the Kimberley Group, are exposed over a large area in the western part of LISSADELL. Here, they lie disconformably on the Speewah Group and are unconformably overlain by the Bastion Group. The Kimberley Group has been described previously by Dow et al. (1964), Plumb (1968), Dow and Gemuts (1969), Gellaty et al. (1970), Plumb and Gemuts (1976), Plumb et al. (1981, 1985), and Griffin et al. (1993).

A lower limit for the age of the Kimberley Group is provided by the c. 1834 Ma age of the Speewah Group. An upper age limit is provided by a SHRIMP U–Pb zircon age obtained from granophyre from the Hart Dolerite, which was intruded into the Kimberley Group at c. 1790 Ma (Page, R. W., 1995, pers. comm.).

The Kimberley Group has been interpreted by Plumb et al. (1981) as having been deposited within a broad, semi-enclosed, shallow marine basin. Palaeocurrents were from the north-northwest (Gellatly et al., 1970).

KIMBERLEY GROUP

King Leopold Sandstone (*ЕKl*)

The King Leopold Sandstone is the basal unit of the Kimberley Group and lies disconformably upon the Bedford Sandstone of the Speewah Group. The succession is about 700 m thick and consists of white, blocky to massive, thick-bedded, medium- to coarse-grained quartz sandstone. Pebble bands are present, and the internal structure of most beds is dominated by stacked sets of medium- to large-scale trough cross-stratification. Sparse palaeocurrent data from the north central part of the map sheet indicates sediment transport to the south-southwest. Gellatly et al., (1970) recorded a dominant southerly palaeoflow for this part of the Kimberley Group throughout the Kimberley Basin.

Carson Volcanics (*ЕKc*)

The Carson Volcanics (Dow et al., 1964) conformably overlies the King Leopold Sandstone and is conformably overlain by the Warton Sandstone. This unit consists of 50–220 m of interlayered basalt, quartz sandstone, feldspathic sandstone, siltstone, and chert. Basalts are green to black and are massive or amygdaloidal. Sandstones are commonly cross-stratified and may contain clay pellets.

Warton Sandstone (*ЕKw*)

The Warton Sandstone, which lies conformably on the Carson Volcanics, consists of about 220 m of white to purple, massive thin- to thick-bedded, fine- to coarse-grained quartz sandstone and minor feldspathic sandstone. Siltstone and flaggy sandstone characterize both the top and bottom of the unit. Internal structure is dominated by medium- to large-scale trough cross-stratification and horizontal planar stratification. Paleocurrent data from the north central part of the map sheet show sediment transport was towards the south-southeast. This trend is consistent with data recorded from the Warton Sandstone elsewhere in the Kimberley Basin by Gellatly et al. (1970).

Elgee Siltstone (*ᵽKe, ᵽKet*)

The Elgee Siltstone lies conformably on the Warton Sandstone and consists of about 200 m of red-brown siltstone interbedded with thin quartz sandstone layers. A dolomitic unit, the Teronis Member (*ᵽKet*), is found locally at the base of the formation and consists of up to 30 m of dolomite, sandy dolomite, micaceous siltstone, and sandstone. Some dolomite beds contain abundant stromatolites. Minor copper mineralization is associated with calcareous sandstone and siltstone layers within the Elgee Siltstone (Marston, 1979).

Pentecost Sandstone (*ᵽKp, ᵽKpl, ᵽKpm, ᵽKpu*)

The Pentecost Sandstone (*ᵽKp*), the uppermost unit of the Kimberley Group, lies conformably on the Elgee Siltstone and is unconformably overlain by the Bastion Group. This sandstone is about 1100 m thick on Lissadell and can be divided into three units, forming the lower, middle, and upper parts of the formation.

The lower part (*ᵽKpl*) consists of medium- to coarse-grained, thinly bedded to laminated quartz sandstone. The middle division (*ᵽKpm*) comprises fine- to coarse-grained planar-stratified or cross-stratified quartz sandstone and siltstone. The upper part (*ᵽKpu*) consists of massive and blocky, trough cross-bedded, coarse-grained, quartz sandstone, and pebbly sandstone.

Palaeocurrent data from the middle division of the Pentecost Sandstone indicate that sediment transport was towards the south-southeast, a trend that was noted earlier by Gellatly et al. (1970).

PALAEOPROTEROZOIC HART DOLERITE (*ᵽdh, ᵽdhg*)

The Hart Dolerite consists of a series of massive dolerite sills (*ᵽdh*) and less extensive granophyre (*ᵽdhg*) that intrude the Speewah Group and the lower part of the Kimberley Group. The sills have a combined thickness of up to 3 km (Plumb and Gemuts, 1976). The Hart Dolerite underlies an area of about 160 000 km^2, and has an estimated volume of 250 000 km^3 (Griffin et al., 1993). Harms (1959) was the first to recognize the intrusive nature of the unit and named it the Hart Dolerite. An excellent description of the Hart Dolerite on Lansdowne is provided by Gellatly et al. (1975), and Alvin (1993) has studied the Hart Dolerite in the Speewah Valley in some detail. A SHRIMP U–Pb zircon age of c. 1790 Ma was obtained for the Hart Dolerite (Page, R. W., 1995, pers. comm.).

DOLERITE (*ᵽdh*)

On Lissadell the Hart Dolerite is well exposed in the Speewah Valley, where it forms the core of a large dome about 50 km long and 25 km wide, and along the Bedford Stock Route between the Nellie Range and the Dunham Pilot Dam. In the southwest of the sheet, the sills are commonly exposed as low, rounded bouldery hills, or treeless stony black-soil plains. In the Speewah Valley the dolerite forms rounded, bouldery hills with up to 100 m relief.

Along the Bedford Stock Route, thick sills of the Hart Dolerite intrude the upper part of the Tunganary Formation, the base of the overlying Valentine Siltstone, the base of the Luman Siltstone, and the contact between the Luman Siltstone and the overlying Bedford Sandstone. In the Speewah Valley, thick sills intrude the base of the Valentine Siltstone and along the contact of the Luman Siltstone and overlying Bedford Sandstone. Thinner sills intrude the Lansdowne Arkose, and dykes cut the King Leopold Sandstone. The marked preference of the Hart Dolerite for siltstone beds was also noted by Gellatly et al. (1975) on LANSDOWNE. The dolerite contains rafts of sedimentary rock up to several kilometres across.

The dolerite unit is composed of fine- to medium-grained dolerite, quartz dolerite and gabbro. Alvin (1993) and Gellatly et al. (1975) both report the presence of olivine dolerite. Adjacent to sedimentary rocks, the Hart Dolerite is commonly chilled. Sharp internal contacts between dolerite types indicate that the thicker sills are composite bodies. In the Speewah Valley, there is a contact between fine-grained and medium-grained quartz dolerite about 14 km north of the Speewah Yards, and pegmatitic dolerite intrudes quartz dolerite about 7 km west of the Speewah Yards. An igneous flow lamination defined by alignment of plagioclase crystals or by small lens-shaped leucogabbro inclusions, is sporadically developed. Alvin (1993) also reported a macroscopic layering of plagioclase-rich and mafic-rich dolerite.

The most common rock type is fine- to medium-grained dolerite dominated by subhedral plagioclase crystals and interstitial anhedral crystals of clinopyroxene and minor orthopyroxene. Plagioclase crystals have weak normal zoning with compositions around An_{50-70}. Clinopyroxene crystals display very fine lamellar and herringbone exsolution of orthopyroxene. Fe–Ti oxides constitute about 3–10% of the rocks, and interstitial quartz and granophyric intergrowths of quartz and K-feldspar form less than 2%.

Quartz dolerite is medium grained, and contains more granophyric intergrowths of quartz and K-feldspar (5–15%) and less pyroxene and Fe–Ti oxide than the dolerite. Plagioclase compositions are typically around An_{40-60}. Gabbro and gabbronorite are medium to coarse grained with a subhedral granular texture. They have a mineralogy and composition similar to that of the fine- to medium-grained dolerite.

All dolerite and gabbro samples have weak to moderate alteration of plagioclase to sericite and clinozoisite, and of pyroxenes to green biotite and minor amounts of green hornblende.

The fine- to medium-grained dolerite and quartz dolerite locally contain numerous inclusions of fine-grained leucogabbro. The inclusions are typically 2–10 cm long (although they may reach about 20 cm), and are either lens-shaped or blebby. In thin section they consist of about 80% or more of plagioclase with euhedral Fe–Ti oxide inclusions, and interstitial clinopyroxene. Interstitial granophyric intergrowths are a minor component of the rocks. Plagioclase is typically strongly altered to prehnite, or sericite and clinozoisite.

GRANOPHYRE (*Ɇdhg*)

The granophyre is developed extensively in the Speewah Valley along the upper contact of the thick sill at the base of the Valentine Siltstone. This rock typically forms cuestas with a relief of about 80–100 m. On aerial photographs the granophyre has a paler tone and a smoother appearance than the dolerite.

On an outcrop scale, the granophyre is a heterogeneous unit showing a range of textures from fine to medium grained to pegmatitic. Most of the unit is medium grained. Pegmatitic granophyre is a volumetrically minor component, generally forming patches or veins cutting the medium-grained granophyre. Fine- to medium-grained granophyre is locally abundant and may contain inclusions of the medium-grained variety. Variolitic cavities are widely developed in the granophyre. Alvin (1993) suggests that the granophyre in the Speewah Valley area is divisible into a lower mafic unit and an upper felsic unit. Most samples taken during this study are similar to descriptions of Alvin's mafic unit.

The rock is composed of 2–4 mm granophyric and micrographic intergrowths of quartz and K-feldspar with subordinate sodic plagioclase. Clinopyroxene and minor green hornblende constitute roughly 5–20% of the rock, and apatite is a common accessory mineral in all samples. Rare samples contain abundant discrete quartz and K-feldspar crystals with no intergrowths. All samples show a dusting of K-feldspar and moderate to strong alteration of plagioclase and clinopyroxene.

The granophyre contains inclusions of sedimentary rock and, in places, abundant lenticular inclusions of dolerite up to about 5 cm long. The dolerite inclusions consist of sericite and epidote after plagioclase, and chlorite and actinolite after pyroxene.

Relationship of the dolerite and granophyre

Field relationships indicate that the granophyre is unlikely to be related to the dolerite by crystal fractionation in situ. There is a range of rock types from olivine dolerite to granophyre, but transitional rock types between dolerite and granophyre are a very minor component of the unit (Gellatly et al., 1975). Although a contact between dolerite and granophyre was not located, medium-grained quartz dolerite is commonly present to within about 5 m of the granophyre. The presence of altered dolerite inclusions in the granophyre (Alvin, 1993), and granophyre dykes intruding dolerite on LANSDOWNE (Gellatly et al., 1975) and in the Speewah Valley (Alvin, M. P., 1995, pers. comm.), indicate that the granophyre is an intrusive body which is separate from the dolerite. Eupene (1970, quoted in Alvin, 1993) suggested that the granophyre intruded by filter-press action from the central part of the sill after the carapace had solidified.

YUNGUL CARBONATITE (*ꞒcCy*)

The Yungul Carbonatite outcrops about 2 km northwest of the old Speewah Yards in the Speewah Valley. The unit has been described by Alvin (1993) and was briefly examined during the course of this work. The age of the carbonatite is not known, but it intruded the Hart Dolerite and the Speewah Group.

The carbonatite consists of a north-trending dyke up to 15 m wide and at least 500 m long that forms a prominent ridge up to 50 m high. In outcrop the carbonatite is pale brown and blocky, and near its margins contains numerous variably carbonated inclusions of Hart Dolerite and quartzose sedimentary rocks of the Speewah Group. The core of the dyke is composed almost entirely of massive calcite with accessory apatite and anatase (Alvin, 1993). The dyke is surrounded by a zone of fenitization some 5–10 m wide. The accessory mineral assemblage and associated fenitization suggest that the rock is a magmatic carbonatite rather than a carbonate vein. Petrography and whole-rock analyses (Alvin, 1993) indicate that the rock is a calciocarbonatite (Woolley and Kempe, 1989). However, the Yungul Carbonatite is much less enriched in rare earth elements, particularly in light rare earth elements, than most calciocarbonatites (Alvin 1993, fig. 6.7c).

PALAEOPROTEROZOIC TO MESOPROTEROZOIC BASTION BASIN

The Bastion Basin corresponds to the present outcrop of the Bastion Group. These rocks unconformably overlie the Kimberley Group and are themselves unconformably overlain by the Antrim Plateau Volcanics. The Bastion Group was originally defined by Dow et al. (1964) and subsequent descriptions have been given by Plumb (1968), Dow and Gemuts (1969), Plumb and Veevers (1971), Plumb and Gemuts (1976), and Plumb et al. (1985). These workers regarded the contact between the Bastion Group and the underlying Kimberley Group as conformable. However, in the north central part of LISSADELL, the Bastion Group truncates the upper member of the Pentecost Sandstone.

The age of the Bastion Group is uncertain, but the older age limit is about 1790 Ma, the age of the Hart Dolerite which intrudes the Kimberley Group. The younger age limit is about 540 Ma, the inferred age of the Antrim Plateau Volcanics. Limited isotopic dating and correlation with other units in the Kimberley suggest a late Palaeoproterozoic to early Mesoproterozoic age for the Bastion Group. Bofinger (1967) obtained a whole-rock Rb–Sr isochron of 1789 ± 58 Ma from the Wynham Shale, whereas proposed correlation with the Birrindudu Group of the Northern Territory would indicate an age of about 1560 Ma (Tyler et al., 1998b).

BASTION GROUP

The Bastion Group is exposed in two areas, the central west and north central parts of LISSADELL. A relatively complete section through the stratigraphy is present in the latter area and comprises, in ascending order, Mendena Formation, Wyndham Shale, and Cockburn Sandstone.

Mendena Formation (*ⱣBm*)

The Mendena Formation unconformably overlies the Pentecost Sandstone and is about 110 m thick on neighbouring CAMBRIDGE GULF. The formation consists of thin- to thick-bedded quartz arenite, lithic quartz sandstone, micaceous sandstone, pebbly sandstone, and minor siltstone and carbonate rock. Internal structure consists largely of medium- to large-scale, low- or high-angle trough cross-stratification and planar horizontal stratification. Ripple cross-lamination is present in the finer grained beds. Paleocurrent data from trough cross-stratified units indicate a southerly directed palaeoflow.

Wyndham Shale (*ⱣBw*)

The Wyndham Shale, which conformably overlies the Mendena Formation, is poorly exposed in the north central part LISSADELL. The formation is 700 m thick on CAMBRIDGE GULF and consists of green or grey siltstone with minor fine-grained micaceous sandstone and calcareous quartz sandstone. Internal structure of the siltstones and sandstones consists largely of planar lamination or ripple cross-lamination. Mudstone pebble clasts are present in some sandstones.

Cockburn Sandstone (*ⱣBc*)

The Cockburn Sandstone is the topmost unit of the Bastion Group and conformably overlies the Wyndham Shale. The upper part of the formation is not exposed in the map sheet area, but on CAMBRIDGE GULF this unit is less than 60 m thick and consists largely of fine- to coarse-grained micaceous quartz sandstone and pebbly sandstone. Sandstone beds are characterized by a varied internal structure consisting of small- to large-scale, high- or low-angle trough cross-stratification, planar tabular cross-stratification, parallel planar to undulatory stratification and ripple cross-lamination. Straight-crested symmetric and asymmetric ripple bedforms are preserved on many bedding surfaces. Paleocurrent data suggest a complex current regime with flow directed mainly towards the northwest and northeast.

MESOPROTEROZOIC CARR BOYD BASIN

The Carr Boyd Group is the only stratigraphic unit within the Carr Boyd Basin and is exposed in the Carr Boyd and Pincombe Ranges in the northeastern part of the Halls Creek Orogen. These rocks were first described by Dow et al. (1964), and subsequently by Dow and Gemuts (1969), Plumb (1968), Plumb and Veevers (1971), Plumb and Gemuts (1976), and Thorne and Tyler (1996). The Carr Boyd Group unconformably overlies Palaeoproterozoic metasedimentary and igneous rocks of the Lamboo Complex and Revolver Creek Basin, and is in turn overlain unconformably by Neoproterozoic glacial deposits of the Duerdin Group. No single complete section of the Carr Boyd Group is known because faulting has disrupted the succession extensively. However, six formations which together total some 4.4 km in thickness are recognized. These are, in ascending order, Hensman Sandstone, Golden Gate Siltstone, Lissadell Formation, Glenhill Formation, Stonewall Sandstone, and Pincombe Formation. All formations are dominated by siliciclastic sedimentary rocks comprising fine- to coarse-grained quartz sandstone, argillite, and minor lithic sandstone and conglomerate. These rocks have been interpreted as alluvial-fan, fluvial and shallow-marine shelf deposits which were laid down in an active strike-slip setting (Plumb et al., 1985). This view is not supported by the work of Thorne and Tyler (1996), who proposed that sedimentation took place in a deltaic to shallow-marine setting and that most of the post-Palaeoproterozoic sinistral faulting in the Halls Creek Orogen occurred after deposition of the Carr Boyd Group.

The age of the Carr Boyd Group is loosely constrained by Rb–Sr whole-rock dates of 1158 ± 123 Ma, 1057 ± 80 Ma, and 891 ± 149 Ma obtained from shales within the Golden Gate Siltstone, Glenhill Formation, and Pincombe Formation respectively (Bofinger, 1967, recalculated by Plumb et al., 1981). However, these ages are younger than the values of 1178 ± 47 Ma (Rb–Sr whole rock) or 1238 ± 12 Ma (K–Ar, phlogopite) reported from the Argyle lamproite diatreme (Pidgeon et al., 1989), which post-dates the Lissadell Formation (Boxer et al., 1989).

The Stonewall Sandstone and Pincombe Formation of the upper Carr Boyd Group are correlated directly with Lalngang Sandstone and Legune Formation (Fig. 4) in the upper part of the Fitzmaurice Group of the Northern Territory (Sweet, 1977; Plumb and Gemuts, 1976). Older formations in the Carr Boyd and Fitzmaurice Groups cannot be correlated directly, although correlations have been made on the basis of inferred relations with the Angalarri Siltstone (Auvergne Group) in the Victoria River Basin (Sweet, 1977; Plumb and Gemuts, 1976; Plumb et al., 1985). However, limited isotopic age data from the Angalarri Siltstone (Webb and Page, 1977) and comparison with the Centralian Superbasin succession (Walter et al., 1995) suggest that the Auvergne Group and its Osmand Range equivalents (Ahern Formation and Helicopter Siltstone) are younger than the Carr Boyd Group.

CARR BOYD GROUP

Most of the Carr Boyd Group stratigraphy, up to and including the lower part of the Stonewall Sandstone, is exposed on LISSADELL. These rocks outcrop in the central and

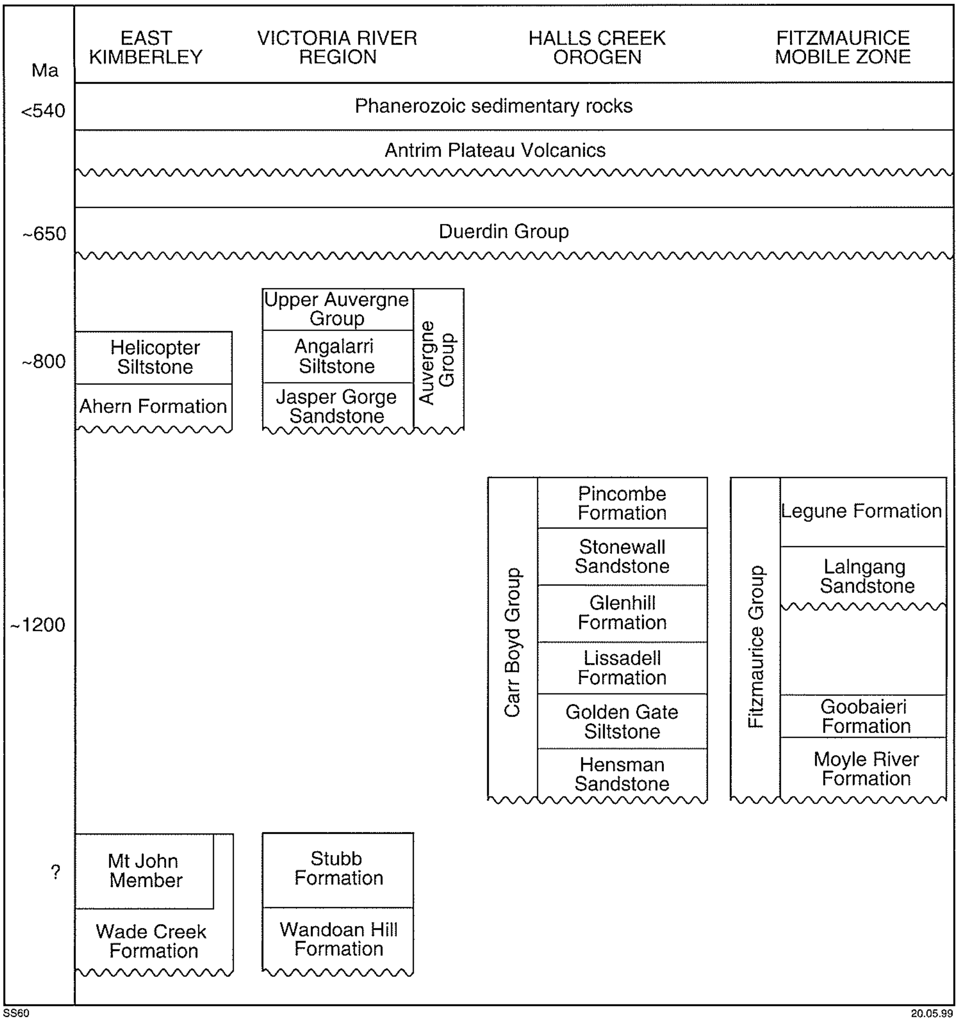

Figure 4. Regional correlation of Mesoproterozoic to Phanerozoic rocks in the east Kimberley region

northeastern part of the sheet area, where they unconformably overlie the Marboo Formation, Whitewater Volcanics, granitoids of the Paperbark and Sally Downs supersuites, Castlereagh Hill Porphyry, and Revolver Creek Formation.

Previous workers recorded the presence of significant angular and erosional unconformities at the bases of the Lissadell and Glenhill Formations (Dow et al., 1964; Dow and Gemuts, 1969; Plumb and Gemuts, 1976). Our mapping on LISSADELL has failed to confirm the presence of these major stratigraphic breaks within the succession.

The discordant boundary relationships and anomalous thickness variations can be more readily attributed to post-Carr Boyd Group tectonism. Specific examples include the areas west of the Argyle mine site, around 16°39'23"S, 128°16'41"E and 16°42'59"S, 128°22'51"E, where the Lissadell Formation is shown to unconformably overlie the Golden

Gate Siltstone, Hensman Sandstone, Revolver Creek Formation and Lamboo Complex on the first edition of LISSADELL (Plumb et al., 1968). In the first area, the apparent unconformable relationship is the result of faulting associated with the Glenhill – Bow River fault system, and the Carr Boyd Group, including the lower part of the Lissadell Formation, is cut by a set of northwest- and northeast-trending fractures which juxtaposes these rocks against the older Lamboo Complex. In the second area, the apparent truncation of Hensman Sandstone and Golden Gate Siltstone by the Lissadell Formation was not confirmed by our mapping, which showed that a complete section of the Hensman Formation was observed throughout this area. The overlying Golden Gate Siltstone was also recorded; however, its thickness here (50 m) is considerably less than the 180–450 m recorded elsewhere on the map sheet. It is unclear whether this variation is the result of localized erosion prior to deposition of the Lissadell Formation, as the contact between it and the Golden Gate Siltstone appears to be conformable and transitional.

Thorne and Tyler (1996) interpreted the depositional environment of the Carr Boyd Group as a braided delta complex, which prograded northward and northeastward over a low-gradient, wave-influenced shallow-marine shelf. The fine grain size and compositional maturity of the sediment was taken as evidence that the source area probably lay several hundred kilometres to the south and southwest. Sediment was transported to the shelf by a system of shallow braided channels and redistributed at the delta front by waves and longshore currents.

Hensman Sandstone (*ȻCh*)

The Hensman Sandstone (Dow et al., 1964) is the lowermost formation in the Carr Boyd Group and has a thickness of 120 m on LISSADELL. It rests with angular unconformity on rocks of the Lamboo Complex and the Revolver Creek Formation and is conformably overlain by the Golden Gate Siltstone. Much of the formation consists of fine- to coarse-grained quartz arenite, with minor siltstone and stratified conglomerate. Upper parts of the formation are commonly ferruginous. Most exposures of the Hensman Formation are massive, and show little detail of the internal structure. In areas where silicification is less intense, the formation is thin to thick bedded and is characterized by an assemblage of sedimentary structures which include medium-scale trough cross-stratification, scour-and-fill structures, planar to undulatory stratification, and both wave- and current-ripple cross-lamination. Ripple bedforms are preserved on many bedding surfaces.

Golden Gate Siltstone (*ȻCg*)

The Golden Gate Siltstone (Dow et al., 1964) has a conformable, transitional contact with the underlying Hensman Sandstone and an apparently conformable, transitional boundary with the Lissadell Formation above. The formation has a measured thickness of 50 m at the Argyle mine site, but estimates from aerial photographs suggest that this value increases to about 450 m in the area west of Lake Argyle. Most of the stratigraphy consists of interbedded siltstone and sandstone, the latter being more abundant in lower and upper parts of the succession. The base of the formation is marked by a thin green siltstone and sandstone, which is overlain by a 7–10 m-thick ferruginous sandstone and sandy ironstone unit. This ferruginous interval forms the host rock to the Pompeys Pillar and Matsu iron prospects. The remainder of the formation comprises interbedded green argillite and thin- to medium-bedded fine- to medium-grained quartz sandstone. Sandstone beds are generally cross-laminated and may have wave- or current-ripple bedforms on their upper surfaces.

Lissadell Formation (*ᵽCl, ᵽCla*)

The Lissadell Formation is about 1200 m thick on LISSADELL and is conformably overlain by the Glenhill Formation. The Lissadell Formation consists of quartzarenite and lithic quartz sandstone, interbedded with green or purple micaceous siltstone. Three major sandstone-dominated units (*ᵽCl*) and three siltstone-dominated units (*ᵽCla*) are recognized. Sandstone-dominated units are commonly massive and silicified and consist of thin- to very thick bedded, fine- to coarse-grained sandstone interbedded with subordinate siltstone. Thin- to medium-bedded sandstones show wave- and current-ripple cross-lamination, parallel planar and undulatory stratification and small-scale trough cross-stratification. Thicker bedded sandstones consist of amalgamated, tabular to lenticular layers displaying medium- to large-scale trough and tabular cross-stratification and parallel stratification. Palaeocurrent data suggest much of the formation was deposited in a complex current regime in which the principal flow directions were towards the west, northwest, and east.

Siltstone-dominated units are parallel planar laminated to undulatory laminated and contain abundant wave- and current-ripple bedforms locally.

Glenhill Formation (*ᵽCg, ᵽCgs, ᵽCga*)

The Glenhill Formation conformably overlies the Lissadell Formation and is unconformably overlain by the Stonewall Sandstone. The succession consists of an estimated 1300 m of interbedded siltstone and sandstone and contains minor conglomerate locally. A 40–50 m-thick sandstone-dominated unit (*ᵽCgs*) forms the lower part of the formation and consists of amalgamated, medium- to thick-bedded quartz sandstone with thin, discontinuous mudstone pebble layers. The internal structure of this unit comprises alternations of stacked medium-scale trough and tabular cross strata, planar and undulatory parallel stratification, and wave- and current-ripple cross-lamination.

The remainder of the Glenhill Formation consists of laminated, green to purple weathering, micaceous siltstone interbedded with thin- to thick-bedded micaceous quartz sandstone and glauconitic sandstone. The sandstones may be present as isolated beds, or form amalgamated units in which individual sandstone beds are separated by thin siltstone layers. Most of the thinner sandstone beds are either parallel laminated or ripple cross-laminated; thicker beds are also trough cross-stratified. Siltstone units that contain only minor amounts of sandstone (*ᵽCga*) are present locally in lower and upper parts of the succession.

Palaeocurrent data from the trough cross-stratified sandstone units are variable and indicate that local transport was principally towards the northeast, with local reworking towards the southwest and southeast.

Stonewall Sandstone (*ᵽCs*)

The Stonewall Sandstone is about 750 m thick, but only the lower part of the formation is exposed on LISSADELL. This unit consists of medium- to very thick bedded, fine- to very coarse grained quartz sandstone and pebbly sandstone interbedded with minor siltstone. Medium- to very coarse grained sandstone and pebbly sandstone show abundant low- to high-angle trough and planar cross-stratification interlayered with parallel planar lamination. Palaeocurrent data from these units show a strong southerly to northerly trend. Fine-grained sandstones are generally parallel planar laminated or ripple cross-laminated.

ARGYLE (AK1) LAMPROITE PIPE (*ƐiA*)

Various aspects of the geology of the diamondiferous Argyle (AK1) lamproite pipe have been comprehensively described in a number of publications. Atkinson et al. (1984) described the discovery of the pipe, Jaques et al. (1986) documented its geology, mineralogy and geochemistry, and subsequent publications have dealt with the geology and volcanology (Boxer et al., 1989), mineralogy and petrology (Jaques et al., 1989a), geochemistry (Jaques et al., 1989b), and geochronology (Pidgeon et al., 1989) of the pipe. The following summary of the geology of the pipe is taken from the above publications, particularly Jaques et al. (1986) and Boxer et al. (1989).

The Argyle (AK1) lamproite pipe intruded the eastern end of the Ragged Range and is host to the world's largest diamond mine. The pipe is composed of various volcaniclastic rocks and subordinate dykes of olivine lamproite. Pidgeon et al. (1989) dated the pipe at 1177 ± 47 Ma based on a combined whole-rock/phlogopite Rb–Sr isochron. At surface, the Argyle pipe is about 2 km long and 150–500 m wide. The pipe is elongate in a north–south direction, and contains a lobe at its northern end. Part of the elongation is primary, but this has been exaggerated by post-intrusion faulting. The pipe intruded the Revolver Creek Formation, as well as the Hensman Sandstone, Golden Gate Siltstone, and Lissadell Formation of the Carr Boyd Group. Angular inclusions of granite and gabbro of the Lamboo Complex are common within the pipe.

Volcaniclastic rocks within the pipe have been subdivided into two groups; 'sandy' and 'non-sandy' tuffs. The sandy tuff is composed of clasts of juvenile lamproite country rock in a matrix of recrystallized lamproite ash and ash-sized quartz fragments. The sandy tuff varies from very thinly to thickly bedded, and cross-bedding is widely developed. Slump and de-watering structures are locally present. Individual beds are poorly sorted and display normal and reverse grading. The non-sandy tuffs consist of massive to very poorly bedded vitric tuffs (hyalotuffs of Jaques et al., 1986, and others), hyaloclastites and autobrecciated lamproite flows. The percentage of clasts in the non-sandy tuffs is much greater than in the sandy tuffs.

MESOPROTEROZOIC YAMPI OROGENY

Tyler and Griffin (1990) identified a deformational event in the King Leopold Orogen, in the west Kimberley region, that produced large-scale shearing in the crystalline rocks of the Hooper Complex, together with northeasterly directed folding and thrusting along the southwestern margin of the Kimberley Basin. Deformation was accompanied by medium-grade metamorphism. This event was referred to as the Yampi Orogeny by Tyler and Griffin (1993) and Griffin et al. (1993).

The age of deformation is poorly constrained, and is inferred to have taken place after intrusion of the Hart Dolerite into the Kimberley Group at c. 1790 Ma (Page, R.W., 1995, pers. comm.), but before deposition of the Neoproterozoic glacigene rocks of the Mount House Group. Tyler and Griffin (1990) suggested that deformation and metamorphism might be linked to c. 1300 Ma events in the Paterson Orogen to the southwest. Shaw et al. (1992) obtained K–Ar ages from sheared granitoid rocks from the Hooper Complex that placed age limits of between 1475 ± 12 Ma and 999 ± 9 Ma on the Yampi Orogeny.

In the Halls Creek Orogen, large-scale, north-northeasterly trending, sinistral strike-slip faults and easterly trending dextral faults have developed after the deposition of the Kimberley Basin (Dow and Gemuts, 1969; Plumb and Gemuts, 1976; Tyler et al., 1995; Thorne and Tyler, 1996), affecting rocks as young as Devonian. However, Dow and

Gemuts (1969) noted that younger rocks showed smaller displacements than older rocks, suggesting that the faults have long, complex histories.

Tyler et al. (1995) suggested that the pattern of strike-slip faulting was controlled by major northeasterly trending structures that developed during the Palaeoproterozoic, and whose position is now marked by the zone boundaries within the Lamboo Complex. The current fault pattern developed first as ductile structures in the Mesoproterozoic, accompanying the northeasterly directed folding and thrusting developed during the Yampi Orogeny in the King Leopold Orogen. Later reactivations of the faults were more brittle. Griffin and Tyler (1992b) referred to the initial, more ductile deformation as D_5, and on LISSADELL foliations that have been attributed to D_5 have been recognized only along the Dunham Fault.

The unconformity between the Revolver Creek Formation and the Carr Boyd Group indicates that tectonism began before c. 1200 Ma. Post-Carr Boyd Group tectonism is indicated by the unconformity at the base of the c. 800 Ma Ahern Formation (see below).

BOW HILL LAMPROPHYRE DYKES (*ElB*)

The Bow Hill lamprophyre dykes form a north-northeasterly trending en echelon swarm about 19 km long (Jaques et al., 1986). The dykes range in width from a few centimetres to 13 m, and are up to 2 km long. On aerial photographs, they have a similar appearance to dolerite dykes, but the two rock types have a different orientation. Pidgeon et al. (1989) determined a K–Ar age of 815 ± 20 Ma for three samples of the Bow Hill dykes. The dykes intruded the Mount Nyulasy Granite of the Paperbark supersuite, which may be fenitized to a distance of 10 m from the dykes. The dykes are composed of ultramafic olivine–phlogopite lamprophyre and garnet–phlogopite pegmatitic lamprophyre. Fine-grained ultramafic rocks are present as selvages to many dykes. Fielding and Jaques (1989) showed that the dykes share many features of ultramafic lamprophyres (Rock, 1991) and, like them, probably have affinities with carbonatites.

MESOPROTEROZOIC TO NEOPROTEROZOIC VICTORIA RIVER BASIN

In the east Kimberley region, the Wade Creek Formation, Ahern Formation, and Helicopter Siltstone are correlated with units of the Victoria River Basin in the Northern Territory (Dow and Gemuts, 1969; Sweet, 1977; Plumb et al., 1985; Thorne and Tyler, 1996; Tyler et al., 1997). Dow et al. (1964) divided this part of the Kimberley succession into four stratigraphic units; in ascending order these are Wade Creek Sandstone, Mount John Shale, Boll Sandstone, and Helicopter Siltstone. This subdivision was later modified by Dow and Gemuts (1967, 1969) who included all units below the Helicopter Siltstone within the Wade Creek Sandstone, while retaining the Mount John Shale as a member within this formation. Mapping by Tyler et al. (1997, 1998c) supports the initial subdivision of Dow et al. (1964), who recognized an unconformity between the Mount John Shale and the overlying succession. The sandstone unit which unconformably overlies the Mount John Shale and which was previously referred to as the Boll Sandstone was renamed the Ahern Formation (Tyler et al., 1997) to avoid confusion with the Boll Conglomerate (Devonian) on DIXON RANGE.

Correlation between the east Kimberley and Victoria River successions (Fig. 4) is based on the strong similarity between the Ahern Formation and the Jasper Gorge Sandstone

(Auvergne Group), and also between the Helicopter Siltstone and the Angalarri Siltstone. It is also likely that the lower part of the redefined Wade Creek Formation and the Mount John Member are equivalent to the Wondoan Hill and Stubb Formations respectively (Plumb and Gemuts, 1976; Sweet, 1977).

There are few reliable age data for the Victoria River succession. The Mount John Member yielded a Rb–Sr whole-rock age of 1128 ± 110 Ma (Dow and Gemuts, 1969) but the same method yielded ages of 1431 ± 440 Ma and 1347 ± 150 Ma respectively from the Wondoan Hill and Stubb Formations (Webb and Page, 1977). The Rb–Sr whole-rock age of 838 ± 80 Ma (Webb and Page, 1977) from the Angalarri Siltstone provides further evidence of a significant time break at the unconformity below the Ahern Formation. Based on this date, its stratigraphic position immediately below the Duerdin Group and its overall lithological makeup, the Auvergne Group (and its east Kimberley equivalents) may correlate with Supersequence 1 of the Centralian Superbasin (Walter et al., 1995).

AHERN FORMATION (*Ɇf*)

On LISSADELL, the Ahern Formation is confined to four small outcrops in the eastern part of the map sheet. The most complete exposure is found in the Evelyn Range around 16°36'00"S, 128°36'00"E. Here, the formation unconformably overlies Lamboo Complex rocks and is either conformably overlain by the Helicopter Siltstone or unconformably overlain by the Moonlight Valley Tillite. The succession is about 200 m thick and consists of thin- to very thick bedded, fine- to very coarse grained lithic quartz sandstone interbedded with minor pebbly sandstone, conglomerate, and siltstone. Coarse-grained rocks predominate in lower and middle parts of the formation, whereas upper levels have a higher proportion of siltstone and fine-grained sandstone. Medium- to thick-bedded sandstones show stacked trough cross-stratification and parallel planar to undulatory stratification. Bifurcating, straight-crested ripples are preserved on the tops of some beds. Several beds in the middle part of the formation show evidence of extensive soft-sediment deformation.

HELICOPTER SILTSTONE (*Ɇh*)

The Helicopter Siltstone, about 330 m thick in the Evelyn Range, conformably overlies the Ahern Formation and is unconformably overlain by the Moonlight Valley Tillite. The formation consists of laminated micaceous green siltstone, interlayered with thin, tabular to lenticular beds of micaceous quartz sandstone. Sandstone is locally dominant in the middle to upper parts of the stratigraphy. These beds commonly fine upwards from a medium-grained sandstone base to coarse-grained siltstone near the top. The lower parts of these beds commonly show parallel planar to undulatory lamination, whereas upper levels are ripple cross-laminated.

NEOPROTEROZOIC WOLFE CREEK BASIN

Neoproterozoic glacigene rocks of the Wolfe Creek Basin outcrop extensively throughout the east Kimberley region (Dow and Gemuts, 1969; Coates and Preiss, 1980; Plumb, 1981). On LISSADELL these rocks are represented by the Duerdin Group and the overlying Albert Edward Group. The Duerdin Group unconformably overlies units of the Victoria River and Carr Boyd Basins, and also Palaeoproterozoic rocks of the Lamboo Complex; both the Duerdin and Albert Edward Groups are unconformably overlain by the Antrim Plateau Volcanics.

Coates and Preiss (1980) correlated the combined Duerdin and Albert Edward Groups with the Louisa Downs Group on MOUNT RAMSAY, and equated both successions with the Marinoan glaciation in South Australia, then dated at c. 670 Ma. This was consistent with an age of c. 660 Ma for deposition of the McAlly Shale (Louisa Downs Group) interpreted from the Rb–Sr data of Bofinger (1967) by Coates and Preiss (1980).

Plumb (1996) disputed this correlation and preferred to equate the Albert Edward Group with the Louisa Downs Group and the Duerdin Group with the Kuniandi Group. He regarded the Louisa Downs Group as the product of a local, mountain glaciation that is not known elsewhere in Australia, but which has been identified just below the base of the Cambrian elsewhere in the world. Recent identification of the stromatolite *Tungussia julia* from carbonate near the base of the group confirms this correlation, and suggests that the glacigene rocks of the Kimberley region all belong to the c. 610 Ma Supersequence 3 of the Centralian Basin (Walter et al., 1995; Corkeron et al.; 1996; Walter and Veevers, 1997; Grey and Corkeron, 1998).

DUERDIN GROUP

Only the upper part of the Duerdin Group, consisting of the Moonlight Valley Tillite and the overlying Ranford Formation, is exposed on LISSADELL. Both formations occur in scattered outcrops close to the line of the Halls Creek Fault.

Moonlight Valley Tillite (*ĐEm*)

The Moonlight Valley Tillite unconformably overlies the Helicopter Siltstone and underlying rocks and has an estimated thickness of 280 m. The tillite consists of a massive, unsorted, pebble- to boulder-conglomerate containing angular to well-rounded clasts of sandstone, granitoid, laminated dolostone, dolorudite, jaspilite, basalt, and dolerite. Many of the clasts are polished and striated. Large sandstone blocks, measuring tens of metres across, are present within the tillite to the north of the Evelyn Range, around 16°35'21"S, 128°31'00"E. Plumb (1968) records a laminated, thinly bedded pink and green dolomite above the tillite on LISSADELL.

Dow and Gemuts (1969) suggested that the tillite was a subaqueous deposit from a grounded ice sheet. Plumb (1981; 1993, pers. comm.) pointed out that the tillite interfingers with fluvioglacial outwash material to the north and suggested that the deposits were derived from a widespread sheet glaciation in which the major ice movement was from the northeast. Terrestrial conditions prevailed in the north but gave way to marine in the south.

Ranford Formation (*ĐEo*)

The Ranford Formation conformably overlies the Moonlight Valley Tillite and is itself either unconformably overlain by the Mount Forster Sandstone, or unconformably overlain by the Antrim Plateau Volcanics. The Ranford Formation is up to 600 m thick on LISSADELL and consists of grey to chocolate-coloured micaecous siltstone interbedded with medium- to very thick bedded, poorly sorted lithic quartz sandstone, dolomitic quartz sandstone, and minor dolostone. The upper part of the formation contains local development of 'zebra stone' near Lake Argyle. Very thick bedded sandstones are massive, except for local scour-and-fill and dish structures; thin- to thick-bedded sandstones are commonly graded and may show flute marks or tool-mark lineations on their lower surfaces. Current ripple bedforms are sometimes preserved on the upper bedding surfaces. Palaeocurrent trends are from north to south.

Early reports of fossil 'jellyfish' or 'medusoids' from the Ranford Formation (Dunnet, 1965) have not been supported by more recent work, which indicates that they are of inorganic origin (Cloud, 1968; Grey, 1981).

ALBERT EDWARD GROUP

This group was originally defined by Smith (1963) and later modified by Dow et al. (1964) and Dow and Gemuts (1969) to include six formations: Mount Forster Sandstone, Elvire Formation, Boonal Dolomite, Timperley Shale, Nyuless Sandstone and Flat Rock Formation. Of these, only the Mount Forster Sandstone outcrops on LISSADELL.

Mount Forster Sandstone (*ƐLo*)

The Mount Forster Sandstone outcrops in the south central part of the map sheet, around 16°52'00"S, 128°24'30"E. Here, the sandstone unconformably overlies the Ranford Formation and is itself unconformably overlain by the Antrim Plateau Volcanics. The formation is about 40 m thick and consists of medium- to very thick-bedded, chocolate-weathering, medium- to coarse-grained quartz sandstone. Internal structure is dominated by stacked sets of medium- to large-scale trough cross-stratification. Discontinuous mudstone pebble lags lie on the bases of some troughed cosets, whereas coset tops may show straight-crested symmetrical ripples. Paleocurrent data indicate that most sediment transport was towards the south-southwest.

GRANITOID (*ge*), PEGMATITE (*p*), AND DOLERITE (*d*) DYKES

Most of the granitoids in both the Bow River and Sally Downs supersuites on LISSADELL are cut by dykes of fine- to medium-grained, even-textured biotite granitoid (*ge*) up to about 2 m wide. The dykes commonly strike between northeast and east-southeast.

Numerous east-northeasterly to northeasterly trending pegmatite dykes (*p*) cut the Radigans Yard Granodiorite west of Radigans Yard. Locally abundant pegmatite dykes, which are too small to be shown on the map, also cut the Mabel Downs Tonalite. They may have a wider distribution, as some of the dykes marked on the map as granite may be pegmatite. Much of the granite country is inaccessible and the two types of dyke have a similar appearance on aerial photographs. Some of the dykes are zoned with a core of fine-grained biotite monzogranite and selvages of pegmatite. Southeast of No. 4 Bore, the pegmatite dykes commonly contain a few percent of coarse crystals of magnetite.

Dolerite dykes (*d*) are widespread on LISSADELL, but most are concentrated in three swarms; northwest of Bow River Homestead, southwest of Pompeys Pillar, and around Jacks Yard. Within each of the swarms the dykes generally have a consistent orientation. The dykes northwest of Bow River Homestead strike roughly north, the dykes near Pompeys Pillar strike northwest, and the dykes around Jacks Yard strike east. There are no discernable differences between the swarms in mineralogy or texture of the dykes. Individual dykes are up to four kilometres long, and most dykes are about 1–5 m wide. They are composed of sparse psuedomorphs of actinolite and chlorite after microphenocrysts of pyroxene or olivine, in a groundmass of very fine grained actinolite, chlorite, albite, epidote, calcite, partly decomposed Fe–Ti oxides, and interstitial quartz.

NEOPROTEROZOIC KING LEOPOLD OROGENY

The King Leopold Orogeny (Tyler and Griffin, 1993; Griffin et al., 1993) produced extensive, well-exposed, west-northwesterly trending folding and thrusting in the King Leopold Ranges, along the southwestern margin of the Kimberley and Speewah Basins (Griffin and Myers, 1988; Tyler and Griffin, 1990), together with the reactivation of shear zones in the Hooper Complex (Tyler et al., 1991; Shaw et al., 1992). Deformation affected Neoproterozoic glacigene rocks and Shaw et al. (1992) obtained K–Ar ages of c. 560 Ma from reactivated shear zones and interpreted this date as the age of deformation. Coates and Preiss (1980) and Plumb (1981) recalculated and reinterpreted the data from Bofinger (1967) and reported Rb–Sr ages of 568 Ma and 576 ± 80 Ma respectively from the McAlly Shale of the Louisa Downs Group. These ages were interpreted as reflecting a metamorphic, cleavage-forming event, which was correlated by Shaw et al. (1992) with the King Leopold Orogeny. Thrusting in the west Kimberley region was linked to sinistral strike-slip faulting in the east Kimberley (Tyler and Griffin, 1990; Tyler et al., 1991). Deformation occurred at about the same time as the Paterson Orogeny in the eastern Pilbara and the Petermann Ranges Orogeny in central Australia (Myers et al., 1996).

Small-scale structures that can be attributed to the King Leopold Orogeny (D_6) have not been recognized on LISSADELL, but uplift and erosion at that time is indicated by the unconformity at the base of the early Cambrian Antrim Plateau Volcanics.

PHANEROZOIC ORD AND BONAPARTE BASINS

Phanerozoic sedimentary and volcanic rocks belonging to the Ord and Bonaparte Basins are exposed in the eastern part of LISSADELL. The Ord Basin lies to the east of the Halls Creek Fault and contains rocks of Early Cambrian to Late Devonian age, whereas the Bonaparte Basin lies to the north of the Halls Creek Fault and contains rocks which range in age from Early Cambrian to Cainozoic. Only one formal stratigraphic unit, the Antrim Plateau Volcanics, is recorded in both basins.

The following description of the Ord and Bonaparte Basins is based upon the work of Mory and Beere (1988).

Cambrian Antrim Plateau Volcanics (*Ꞓa, Ꞓac*)

The Antrim Plateau Volcanics (Traves, 1955; Mory and Beere, 1988) covers a large area in the southeastern part of LISSADELL and is also exposed in scattered outcrops between Glen Hills and Doon Doon Homesteads and the Speewah Valley. The formation is about 200 m thick in the north and unconformably overlies Palaeoproterozoic to Mesoproterozoic rocks, whereas in the southeast it is about 1000 m thick and unconformable on the Neoproterozoic Duerdin and Albert Edward Groups. The Antrim Plateau Volcanics is conformably overlain by the Headleys Limestone in the Ord Basin and by the Tarrara Formation in the Bonaparte Basin.

The Antrim Plateau Volcanics consists predominantly of massive to amygdaloidal, aphyric to porphyritic basalt and basaltic breccia interbedded with minor sedimentary rock. The Antrim Plateau Volcanics includes a 5 m-thick laminated, stromatolitic chert and chert breccia unit, the Mount Close Chert Member (*Ꞓac*), in the southeastern part of the map sheet. Most basalt is fine to medium grained and consists of plagioclase, clinopyroxene and opaques, within a matrix of devitrified glass or quartz–feldspar residuum. Amygdales are commonly filled with quartz, chert, calcite, chlorite, prehnite or pumpellyite. A 5 m-thick unit of medium-grained quartz sandstone showing large-scale, low-angle cross-stratification, lies interbedded with the basalt flows south of Glen Hills Homestead.

ORD BASIN

Cambrian Goose Hole Group (*€G*)

In the Ord Basin, the Antrim Plateau Volcanics is overlain by a thick succession of Cambrian siliciclastic and carbonate rocks which Mory and Beere (1985) named the Goose Hole Group. This succession is subdivided into the Negri Subgroup and the overlying Elder Subgroup. The Headleys Limestone is the lowest formation in the Negri Subgroup and is conformably overlain by the Nelson Shale. However, in the vicinity of Lissadell Hill the Headleys Limestone is also succeeded by a locally developed unnamed unit consisting of lithic sandstone, pebbly sandstone and minor conglomerate (*€G*).

Negri Subgroup

Headleys Limestone (€Gh)

The Headleys Limestone (Traves, 1955; Mory and Beere, 1988) is 35 to 50 m thick and has a sharp, concordant contact with the underlying Antrim Plateau Volcanics. The contact with the overlying Nelson Shale is also abrupt and apparently conformable. The formation consists of grey, laminated or massive micritic limestone with chert nodules common in the more massive sections. Pustular microbial fabrics, intraformational breccias and simple stromatolites are present in addition to the ubiquitous laminated microbial micrites.

Nelson Shale (€Gn)

The Nelson Shale is 100 to 180 m thick and conformably overlies the Headleys Limestone, although Mory and Beere (1988) note that it directly overlies Antrim Plateau Volcanics in the Argyle Syncline. The formation consists of purple siltstone with thin beds of sandstone and micrite (Mory and Beere, 1988).

Linnekar Limestone (€Gl)

The Linnekar Limestone lies conformably between the Nelson Shale below and the Panton Formation above, is 10 to 25 m thick, and consists of laminated micritic limestone, calcareous shale, shale, and chert. This unit contains a variety of fossils including *Redlichia forresti*, *Girvanella* and *Biconulites hardmani*, and unnamed stromatolites (Mory and Beere, 1988).

Panton Formation (€Gp, €Gps, €Gpc)

The Panton Formation is a unit of purple siltstone and mudstone interbedded with minor sandstone and limestone which lies conformably between the underlying Linnekar Limestone and the overlying Elder Subgroup. The formation varies in thickness from 100 to 300 m and contains two carbonate members, the lower Shady Camp Limestone Member (*€Gps*) and the upper Corby Limestone Member (*€Gpc*). Both units are less than 5 m thick. The lower member consists of oncolitic fossiliferous limestone, whereas the upper member consists of laminated and massive unfossiliferous micrite and chert.

Elder Subgroup

Eagle Hawk Sandstone (€Ge)

The Eagle Hawk Sandstone has been reported from the core of the Argyle Syncline (Mory and Beere, 1988) but is not shown as outcrop on the accompanying map. In the type area on DIXON RANGE the formation has a maximum thickness of 210 m and consists of fine-grained micaceous sandstone and minor siltstone and mudstone. The unit labelled *€Ge* in the Ragged Range belongs to the Hart Spring Sandstone of the Bonaparte Basin and is believed to be a correlative of the Eagle Hawk Sandstone (Mory and Beere, 1988).

BONAPARTE BASIN

Cambrian and Ordovician Carlton Group

The Carlton Group (Mory and Beere, 1988) comprises the Cambrian and Ordovician succession in the Bonaparte Basin that conformably overlies the Antrim Plateau Volcanics and is disconformably overlain by the Devonian Cockatoo Group. Six formations make up the Carlton Group; however, only the lowermost unit, the Tarrara Formation, and the Hart Spring Sandstone (see Eagle Hawk Sandstone) are present on LISSADELL.

Tarrara Formation (€Ct)

The Tarrara Formation is exposed in the southern part of the Ragged Range where it is about 60 m thick. It has an abrupt but apparently conformable contact with the underlying Antrim Plateau Volcanics and is disconformably overlain by the Devonian Ragged Range Conglomerate. The formation consists of grey, laminated micaceous siltstone with subordinate quartz sandstone and fossiliferous dolomite.

Devonian Cockatoo Group (*DC*)

The Upper Devonian (Frasnian) succession in the Bonaparte Basin is referred to as the Cockatoo Group and is subdivided into ten formations (Mory and Beere, 1988). Four of these units, Ragged Range Conglomerate, Cyril Sandstone, Kellys Knob Sandstone, and Galloping Creek Formation are exposed on LISSADELL. In addition, a number of small outliers of unassigned Cockatoo Group sandstone and conglomerate (*DC*) are present near the northern boundary of the map sheet.

The Cockatoo Group is interpreted as a mixed alluvial-fan–fluvial–eolian–shallow-marine succession, deposited in response to strike-slip movement along the Glenhill – Revolver Creek – Carr Boyd fault system (Mory and Beere, 1988; Thorne and Tyler, 1996).

Ragged Range Conglomerate (DCr)

The Ragged Range Conglomerate rests with angular unconformity on Cambrian or Precambrian rocks in the Ragged Range and is conformably overlain by the Cyril Sandstone. It is about 300 m thick in the type area and consists of pebble to boulder conglomerate, pebbly sandstone and fine- to very coarse grained sandstone (Mory and Beere, 1988). Conglomerate is polymodal and clast supported, or more rarely, matrix supported. Massive bedding is dominant with rare trough cross-stratification and parallel planar stratification. Sediment transport was towards the west and north.

Cyril Sandstone (DCy)

The Cyril Sandstone lies conformably between the Ragged Range Conglomerate below and the Kellys Knob Sandstone above. Mory and Beere (1988) record a thickness of 180 m for the formation in the Ragged Range, where it consists largely of fine- to medium-grained quartz sandstone with pebbly sandstone, conglomerate, and siltstone. Internal structure consists mainly of small- to medium-scale trough cross-stratification with lesser amounts of parallel planar stratification and planar tabular cross-stratification.

Kellys Knob Sandstone (DCk)

The Kellys Knob Sandstone is the uppermost formation of the Cockatoo Group on LISSADELL. An incomplete section through the formation is exposed in the northern part of the Ragged Range, where it consists of fine- to medium-grained quartz sandstone. Internal structure is dominated by medium- to very large scale planar cross-stratification with local parallel stratification. The Kellys Knob Sandstone is 70 to 350 m thick elsewhere in the Bonaparte Basin

Galloping Creek Formation (DCg)

On LISSADELL, the Galloping Creek Formation (Beere and Mory, 1986) unconformably overlies Proterozoic basement rocks southeast of the Ragged Range. The type section for the formation lies 10.5 km south of Glen Hills Homestead. Here, the formation is about 1.6 km thick and consists of thick sequences dominated by conglomerate, pebbly sandstone, and sandstone. The conglomeratic intervals consist of very thick bedded conglomerate with minor sandstone beds and lenses. Conglomerates are generally clast-supported and contain well-rounded pebbles and boulders of quartz sandstone with minor granitoid and volcanic rock. Sandstones and pebbly sandstones contain a similar suite of clast types with a minor dolomitic component. Massive bedding is common in all conglomeratic sequences, with minor planar and trough cross-stratification and parallel stratification. Pebbly sandstone and sandstone beds are feldspathic and exhibit trough and planar cross-stratification, parallel stratification and minor ripple cross-lamination. Palaeocurrent data from the Galloping Creek Formation indicate that sediment transport was principally towards the east and northeast.

PALAEOZOIC ALICE SPRINGS OROGENY

The youngest phase of major faulting and folding in the east Kimberley region is attributed to the Late Devonian to Carboniferous (400–300 Ma) Alice Springs Orogeny (Shaw et al., 1992; Tyler et al., 1995; Thorne and Tyler, 1996). East of the Halls Creek Fault, Cambrian rocks of the Antrim Plateau Volcanics and the Goose Hole Group are folded into large-scale, doubly plunging, northeast- to east-trending open synclines and anticlines. These folds are cut by northeast- and northwest-trending splays from the Halls Creek Fault. West of the Halls Creek Fault, Cambrian to Upper Devonian rocks of the Bonaparte Basin are tilted and deformed into open, upright flexures with curvilinear axes locally (e.g. east of the Ragged Range at 16°34'41"S, 128°24'17"E). These folds are cut by an array of steeply dipping, northeast-trending sinistral wrench faults and associated synthetic and antithetic fractures which form part of the Halls Creek – Carr Boyd – Glenhill fault system.

Late Devonian alluvial fan and braided fluvial deposits in the east Kimberley region were deposited in response to strike-slip movement along the Halls Creek fault system (Mory and Beere, 1988; Mory, 1990; Thorne and Tyler, 1996). Thorne and Tyler (1996) noted that most of the Frasnian sedimentary sub-basins have the structural characteristics of either stepover or transpressional strike-slip basins and were formed as a result of major sinistral movements along the Halls Creek and Dunham–Ivanhoe fault systems.

In the southwestern part of LISSADELL, the Dunham Fault was reactivated during the Alice Springs Orogeny. Both this and the Bow River Fault are marked by lines of quartz veins (*q*).

CAINOZOIC SURFICIAL DEPOSITS

Semi-consolidated slope deposits and scree (*Czc*) outcrop below scarps and in valleys where erosion has produced terraces above the current alluvium-filled stream channels. Colluvial sand and gravel (*Czcv*) forms valley-fill deposits above the present day drainage channels. Semi-consolidated and unconsolidated silt, sand, and gravel (*Czs*) covers valley floors and large plains and is associated locally with older, dissected alluvial sand, gravel, and silt (*Czsa*). Treeless black-soil plains (*Czb*) cover broad floodplains adjacent to the larger creeks. These plains consist of black and dark grey-brown soils and cracking clays, and are developed extensively over Phanerozoic rocks and the Hart Dolerite. Calcrete (*Czk*) is associated with colluvium and alluvium locally, whereas sandplain deposits (*Czn*) are developed close to Phanerozoic and Precambrian sandstone outcrops

Alluvium (*Qa*) consisting of unconsolidated silt, sand, and gravel lies along present day drainage channels.

ECONOMIC GEOLOGY

The following descriptions of mineral deposits and occurrences on LISSADELL are based on a summary of both published information and that contained within open-file WAMEX reports compiled by Sanders (1999), covering the whole of the Halls Creek Orogen.

COPPER

Copper mineralization has been reported from several Precambrian and Phanerozoic igneous and sedimentary rock units on LISSADELL, although none has yet proved to be economic.

Disseminated, vein, and minor massive copper–nickel sulfide mineralization is hosted by a small, metamorphosed layered intrusion at the Bow River prospect, about 18 km south-southwest of the Argyle diamond mine. The intrusion is about 6.5 km long and strikes northeast. Layering dips steeply to the north. The intrusion is part of the Tickalara Metamorphics, and was metamorphosed to granulite facies during D_2/M_2 in the Central zone. The dominant rock type was originally a norite, but the intrusion varies from leucogabbroic to troctolitic in composition.

The mineralized part of the intrusion is roughly lenticular measuring some 900 by 300 m. The host rock type is a mafic granulite derived from norite. Retrogressive amphibole–chlorite alteration is extensive. At the surface, mineralization consists of disseminated limonite, accompanied in some areas by malachite and chrysocolla stains and disseminated grains. Drilling by Pickands Mather International indicated an average grade of 0.20% copper and 0.15% nickel. Sulfide mineralization in drillholes consists of: (i) disseminated pyrrhotite and minor chalcopyrite, (ii) rare, thin bands of massive pyrrhotite and minor chalcopyrite, (iii) pyrrhotite and chalcopyrite as veins and stringers along the margins of younger dykes, and (iv) shear- and fracture-hosted pyrrhotite and chalcopyrite. Pentlandite forms flames in the pyrrhotite.

Australian Anglo American concluded from their work that the mineralization at Bow River appears to be low-grade, primary, disseminated sulfide. Metamorphism and deformation have produced patchy remobilization of the sulfide into small, richer, disseminated and massive pockets.

The Durack Ranges Prospect is hosted by the Carson Volcanics of the Kimberley Group. Copper mineralization has been reported from a variety of rock types and along several stratigraphic horizons but most deposits are commercially insignificant. One fault-related occurrence hosts an inferred resource of 1.84 Mt at 0.51% copper.

Harms (1959) and Plumb (1968) recorded copper staining on joints in Pentecost Sandstone at Campbellmerry and copper mineral veinlets in Mendena Formation at Plants. Minor copper mineralization is also associated with calcareous sandstone and siltstone layers within the Elgee Siltstone in the western part of LISSADELL (Marston, 1979).

In the Speewah Valley, area several occurences of copper mineralization are associated with the Hart Dolerite. At Martins Duracks, lead–silver–copper–gold mineralization is in a flat-dipping quartz reef hosted by granophyric and doleritic rocks of the Hart Dolerite. This deposit is thought to be same as the Martins prospect, located at 16°12'00"S, 127°58'00"E on the first edition of LISSADELL (Plumb et al., 1968) and subsequently described by Blockley (1971). The outcropping quartz veining strikes northwest at 145°, dips west at around 10–15° and can be traced on the surface for a distance of approximately 100 m. The mineralization consists of galena, cerussite, malachite, and azurite, with some native silver distributed in patches (Blockley, 1971). Pyrite and arsenopyrite are also present. The quartz reef is soil covered to the north and south.

Mineralization at the nearby Calamondah prospect is also in a flat-lying quartz reef within granophyric rocks of the Hart Dolerite. The quartz reef strikes northwest at 150° and dips to the east at 15°. The reef has a thickness of 1.5–6.0 m and can be traced along strike for 240 m. Analytical results returned a range in silver values from 1.0 to 13 ppm, copper values ranged from 13 to 1750 ppm, and lead values ranged from 20 to 350 ppm.

Minor copper mineralization has been reported from the Cambrian Antrim Plateau Volcanics and the overlying Headleys Limestone (Dow et al., 1964; Plumb, 1968; Marston, 1979). Amygdales and geodes in the Antrim Plateau Volcanics contain quartz, chalcedony, calcite, prehnite, celadonite, zeolite group minerals, bituminous material, and minor amounts of malachite, chalcocite, cuprite, and native copper. Copper minerals may also be present as disseminations and thin lamellae within the basalts, particularly in amygdaloidal or agglomeratic volcanic rocks (Marston, 1979). In 1965, Pickands Mather International outlined a 5.5 km-long stream-sediment copper anomaly coincident with the contact between the Headleys Limestone and the Antrim Plateau Volcanics at Discovery Yard (Bore). This anomaly was recognized as having potential similar to several reported rich but thin, discontinuous occurrences on the same stratigraphic horizon in localities to the east. Chalcocite and malachite 'are associated with' the Headleys Limestone and have a maximum thickness of 0.3 m. The extent of mineralization was observed to be 30 by 10 m. Plumb (1968) reported that disseminated and nodular chalcocite and associated nodular carbonates have been recorded from the base of the Headleys Limestone at Rosewood Wall.

Minor amounts of copper carbonate are present in joints in calcareous rocks of the Cambrian Blatchford Formation (Hart Spring Sandstone) in the northern Ragged Range (Plumb,1968) on CAMBRIDGE GULF.

DIAMOND

The Argyle mine is the largest diamond producer in the world, although most of the diamonds are of the 'cheap gem' and 'industrial' types (Jaques et al., 1986; Lewis, 1990). Total production to July 1997 from the Argyle mine is 441 Mct at a value of A$5.280 billion. The diamonds are hosted by the Argyle (AK1) lamproite pipe (*ℙiA*). The majority of the diamonds are hosted by 'sandy tuffs' (Lewis, 1990).

Production from the Argyle pipe began in 1985, when the mine produced 7 Mct (Lewis, 1990). In 1986 production was just under 30 Mct. Production rose gradually to a peak of just over 40 Mct in 1994 and in 1997 was 38.6 Mct (Argyle Diamonds, 1998). In 1996 Argyle contained openpit reserves of about 70 Mt at 3.3 ct/tonne for a total of about 230 Mct (Preston, 1997). In June 1998 the joint venture partners, Ashton Mining Ltd and Rio Tinto Ltd, announced plans to expand the openpit and extend the mine life beyond 2003 (Flint and Abeysinghe, 1998).

Alluvial diamond deposits of economic grade are also present in the vicinity of the Argyle mine; streams that drain the lamproite pipe deposited the diamonds (Deakin et al., 1986). The Limestone Creek deposits east of Argyle consist of a dissected piedmont fan up to 3.5 m thick of Pliocene age. Modern gravels are derived from erosion of the piedmont fan. The Smoke Creek deposits north and northeast of Argyle are contained in modern floodplain gravels and older terrace gravels. The alluvial diamonds at Limestone Creek and Smoke Creek were deposited rapidly during flash flooding. As a result, the grades are similar to that of the Argyle pipe.

Mining of the Upper Smoke Creek and Limestone Creek deposits by Argyle Diamond Mines Pty Ltd is complete. These deposits produced over 17 Mct between 1982 and 1985 (Lewis, 1990). The Lower Smoke Creek deposit has produced, on average, about 1.5 Mct annually since 1989 (Argyle Diamonds, 1998). The nearby Bow River alluvial deposit was worked by Normandy Mining Ltd between 1985 and 1996 and produced 7.37 Mct during this time.

Small quantities of diamond have been recovered from lamprophyres of the Lissadell Road dykes, located just to the southwest of the Argyle Pipe at 16°43'00"S, 128°18'24"E. The lamprophyre is present in a discontinuous, en echelon series of narrow veins up to 10 m in length

The Bow Hill dykes comprise a suite of micaceous, andradite-bearing, ultramafic–mafic lamprophyres emplaced in granite rocks associated with the 1865–1850 Ma Paperbark supersuite. Eight individual dykes, ranging in width from a few centimetres to 13 m, and up to 2 km long, are emplaced en echelon over a strike length of 19 km. They are considered to be mica lamprophyres of carbonatitic affinity (Jacques et al., 1986). No diamonds have been reported from these lamprophyres.

At Maude Creek, a 1.5 m-wide kimberlite dyke intrudes the Hart Dolerite, which itself is intrusive into the Speewah Group. The highly micaceous rock has not been dated, but the similarity to other examples in the North Kimberley Province suggests a Late Proterozoic age of about 800 Ma. Small quantities of diamonds have been recovered from these kimberlites (Jaques et al., 1986).

FLUORITE

Fluorite mineralization on LISSADELL has been reported from the Lamboo Complex and parts of the Hart Dolerite and Speewah Group.

Fluorite was discovered at Archie Creek by Minatome in the mid-1970s during exploration for uranium. The geology of the prospect is dominated by Whitewater Volcanics rocks, which lie adjacent to granitoid of the Paperbark supersuite that contains interfingering mafic intrusive rocks. Fluorite is present as an accessory mineral in some of the porphyritic rhyolite and rhyolite tuff, particularly in zones of higher (thorium) radioactivity in close proximity to the contact with the granitoid. Fluorite is present as a minor, accessory, quartz-replacive mineral in the xenolithic lava, and also as large, dark-purple sheared crystals in the rhyolitic tuff horizons. Minatome obtained a best assay of 1.8% CaF_2 in 'mineralized rhyolite'. Another sample described as '?volcanogenic sediment' returned 3000 ppm fluorine, 15 ppm copper, 10 ppm lead, 40 ppm zinc, 16 ppm uranium, and 40 ppm thorium.

Fluorite was first recorded in the Speewah Valley area in 1905 and the Speewah fluorite deposit was discovered in 1972 by Great Boulder. Fluorite–barium–copper–lead mineralization is associated with subvertical, north-northeasterly and northerly trending faults that cut the Speewah Group and doleritic to granophyric rocks of the Hart Dolerite. The faults are commonly filled with massive to laminated quartz and locally carry significant fluorite, barite, galena, specularite, chalcopyrite, and sphalerite. Substantial fluorite mineralization is developed in two areas, referred to as the 'main'and 'west ridge' zones of mineralization.

The 'main' zone is made up of three en echelon, north-northeasterly trending faults that have been traced for 8 km. Fluorite mineralization is recorded on all three faults, and subsidiary structures, over a total strike length of 5 km. The vast bulk of the resource is present over a 2 km strike length of multiple fluorite lodes, termed the 'ABC' zone. Here,the mineralization lies in subvertical, stockwork and cross veins and consists of variable amounts of fluorite, quartz, and, less commonly, barite. The fluorite is grey or white in outcrop, but typically pale green where fresh. The veins exhibit symmetrical lamination, colloform and crustiform banding, and spider vein brecciation that suggest an upper-crustal depositional environment (Alvin, 1993). The ABC deposit has a measured resource of 1.87 Mt at 25.8% CaF_2.

The 'west ridge'mineralization is located on the crest of a north-northwesterly trending ridge, approximately 5.6 km southwest of the ABC zone. The full extent of the mineralization has not been traced along strike. The host rock is described by Great Boulder as a 'breccia-shear zone invaded by fluorite-bearing granophyre'. Sporadic and relatively minor fluorite(–barite) mineralization is associated with irregular cross-cutting quartz–carbonate–fluorite segregations, veins, veinlets, and cavity fill. Open-space filling textures and quartz pseudomorphs after bladed calcite are common. Mineralization occurs preferentially in the altered doleritic material but is also found in the quartzite.

GEMSTONE

Ornamental 'zebra stone' has been quarried from the Ranford Formation in the vicinity of Lake Argyle.

GOLD

The Jailhouse Creek prospect consists of thin and widely spaced, weakly mineralized quartz veins in rocks of the Whitewater Volcanics in the south central part of LISSADELL.

The location of the prospect on the map is the central stream-sediment anomaly. Weakly anomalous gold is associated with north-, northeast- and east-trending quartz veins in quartz–feldspar porphyry adjacent to the Dunham Fault. Quartz veins are locally accompanied by intense silicification of the volcanic rocks. Minor base metal mineralization is present in several of the quartz veins, but not necessarily associated with anomalous gold contents.

IRON

Northwest of the Argyle diamond mine, beds of low-grade, siliceous hematite up to 10 m thick outcrop over a strike length of 16 km (Dow and Gemuts, 1969). The two main areas of massive hematite, Pompeys Pillar and Matsu, are separated and displaced by the Bow River Fault. The deposits are hosted in the lower part of the Golden Gate Siltstone of the Carr Boyd Group. The deposits consist of massive sandy hematite, and hematitic sandstone, with interbedded ferruginous shale and quartz sandstone. The massive hematite ore contains cross-bedding, indicating a clastic origin for the deposits (Dow and Gemuts, 1969).

The Pompeys Pillar deposit (also known as the 'western' deposit) is nearly 3 km long, 4–10 m thick, and up to 120 m wide (MacLeod, 1963). The Matsu deposit (also known as the 'eastern' deposit) is about 6 km long and 10 m thick. Although individual beds contain up to 60% iron, numerous interbeds of sandstone and shale reduce the grade of ore that could be mined to about 50% iron (Dow and Gemuts, 1969). The Pompeys Pillar and Matsu deposits together contain an estimated resource of 15 Mt at about 64% iron (Preston, 1997).

LEAD

A number of small lead occurrences have been reported from the map sheet area (Dow et al., 1964; Plumb, 1968; Blockley, 1971; Sanders, 1999). The largest of these are at Martins Duracks (see **Copper**) in the Speewah Valley, and Carr Boyd and Boxers at Lake Argyle.

The Carr Boyd lead–zinc–silver occurrence is located 500 m from the shores of Lake Argyle and was discovered by Pickands Mather International in 1965 following field investigation of an extensive stream-sediment lead anomaly. Mineralization is hosted by sedimentary rocks of the Golden Gate Siltstone (Carr Boyd Group). The succession is exposed in a northerly to north-northeasterly trending anticline, the eastern limb of which is truncated by the Carr Boyd Fault. Scattered lead–zinc–silver mineralization is found over a north-trending, 4 by 1 km zone and is associated with subvertical northeast- and northwest-trending cross faults, with minor normal and reverse displacements. Lead mineralization is typically present as small, widely spaced, north-trending mineralized breccia zones and consists of disseminated low-grade galena and yellow lead oxide. The mineralized breccia zones are essentially confined to competent sandstone units and are characterized by angular brecciated sandstone fragments cemented with limonitic material, commonly showing well-developed cubic boxwork. There is generally some vuggy quartz veining. Many of the breccia zones fade into closely spaced joint sets or merge into barren quartz veins. Samples from the breccia zones assayed from 0.05 to 30% lead and averaged about 1%: most values were from 0.1 to 8% lead. Zinc values up to 1.0% were recorded, but were typically much lower. The highest silver value was 36 ppm (average 15 ppm). Samples of apparently barren sandstone adjacent to breccia zones returned values of less than 0.02% lead.

At the Boxers prospect galena, fluorite, and barite are present in a quartz vein 25–38 cm wide and about 6 m long, striking at 230° in coarse-grained granite of the Paperbark supersuite (Blockley, 1971). This occurrence is no longer accessible due to inundation following construction of the Ord River Dam.

TIN

Detrital tin has been recorded from the Castlereagh prospect in the southwest corner of the map sheet. The deposits are associated with the Beefwood Yard Granite of the Paperbark supersuite and the Castlereagh Hill Porphyry. Both the granite and the porphyry contain cassiterite and scheelite in subeconomic proportions.

URANIUM

Small uranium occurrences have been reported in rocks of the Lamboo Complex, the Carr Boyd Group, and the Cockatoo Group (Plumb, 1968; Dow and Gemuts, 1969; Carter, 1976)

The Frog (Dunham Hill) prospect is located in Whitewater Volcanics, 1.7 km west of the Great Northern Highway. The area was explored by Minatome–Aquitaine in the period 1977–78. Uranium–fluorite mineralization lies in and adjacent to the Dunham Fault Zone and is related to zones of hematitization, most of which appear to be structurally controlled. Mineralization is reported from two principal areas, termed the Frog prospect and the 'Western Area'.

At the Frog prospect, mineralization is associated with a radiometric anomaly that trends at 040°, coincident with the faulted contact between a cream–grey porphyritic rhyolite and ferruginous, sericitic felsic agglomerate. Mineralization consists of minor yellow and green secondary uranium minerals coating silicified joint planes. Several minor anomalous zones, with maximum strikes of 30 m and widths up to 2 m, are located within 500 m of the Dunham Fault. Geological mapping indicates that the radiometric anomalies are spatially associated with zones of shearing. Maximum non-coincident assays were 0.32% ppm uranium, 110 ppm thorium, and 1440 ppm fluorine.

The 'Western Area' comprises a group of radiometric anomalies and is located approximately 1 km west of the Frog prospect. The highest assays of 720 ppm uranium, 35 ppm thorium, 4800 ppm fluorine, and 225 ppm copper were obtained from areas with strong iron staining and fluorite mineralization.

The Dunham River prospect (Dow and Gemuts, 1969) lies 8 km south of Doon Doon Homestead in granite of the Paperbark supersuite. Here, autunite coats joints in a dyke cutting the granite but only a very small patch of mineralization was discovered.

Minor uranium mineralization has been recorded from the Carr Boyd Group to the west of Lake Argyle at 16°11'07"S, 128°35'31"E. Carter (1976) notes that secondary uranium minerals including phosphuranylite and torbernite coat joints and fractures in the Hensman Sandstone, close to the Revolver Creek Fault.

Exploration of the Upper Devonian Galloping Creek Formation has revealed traces of uranium enrichment. East of the Ragged Range, around 16°25'48"S, 128°24'23"E, secondary uranium mineralization, represented by carnotite, is restricted to the top layer of unconsolidated, iron-rich conglomerate and sandstone. This occurrence is considered to have no commercial potential (Carter, 1976).

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Appendix

Gazetteer of localities

Locality	*Latitude (S)*	*Longitude (E)*	*AMG coordinates* *Easting*	*Northing*
Aida Vale Mill	16°43'52"	127°50'42"	376900	8149800
Archie Creek prospect	16°37'47"	127°56'41"	387448	8161103
Argyle mine	16°42'59"	128°23'41"	435500	8151700
Beefwood Yard	16°51'26"	127°52'38"	380400	8135900
Bow Hill	16°41'42"	128°12'43"	416000	8154000
Bow River	16°53'37"	128°10'25"	412000	8132000
Bow River Homestead	16°52'13"	128°11'03"	413100	8134600
Bow River mine	16°37'32"	128°36'22"	458000	8161800
Bow River prospect	16°52'15"	128°19'54"	428810	8134606
Boxers prospect	16°17'30"	128°40'00"	464388	8198766
Campbellmerry prospect	16°10'35"	127°35'15"	348994	8211028
Carr Boyd prospect	16°12'29"	128°41'35"	467193	8208019
Castlereagh Creek	16°54'22"	127°52'07"	379500	8130500
Castlereagh Hill	16°58'40"	127°57'17"	388700	8122600
Castlereagh prospect	16°53'29"	127°52'24"	379998	8132110
Crooked Creek	16°49'33"	128°08'28"	408500	8139500
Discovery Yard (Bore) prospect	16°43'07"	128°35'00"	455582	8151520
Doon Doon Homestead	16°18'33"	128°14'33"	419100	8196700
Dunham Jump-Up	16°22'40"	128°13'05"	416500	8189100
Dunham Pilot Dam	16°01'36"	128°22'25"	433000	8228000
Durack Ranges prospect	16°44'47"	127°39'47"	357489	8148014
El Questro Homestead	16°00'28"	127°58'49"	390900	8229900
Evelyn Range	16°35'47"	128°31'52"	450000	8165000
Fargoo Creek	16°58'32"	128°21'10"	431100	8123000
Fish Hole Yard	16°49'29"	128°03'18"	399300	8139600
Frog (Dunham Hill) prospect	16°27'02"	128°12'03"	414687	8181050
Glen Hills	16°32'37"	128°21'18"	431200	8170800
Gordons Gorge	16°44'42"	127°52'14"	379600	8148300
Jacks Yard	16°43'35"	127°55'30"	385400	8150400
Jailhouse Creek prospect	16°31'20"	128°10'44"	412376	8173113
Lake Argyle	16°16'50"	128°47'38"	478000	8200000
Lissadell Hill	16°48'09"	128°26'50"	441100	8142200
Lissadell Homestead	16°45'59"	128°33'01"	452100	8146200
Martins Duracks prospect	16°13'43"	127°59'25"	392086	8205504
Matchbox Creek	16°02'25"	128°49'00"	480400	8226600
Matsu prospect	16°41'43"	128°19'49"	428600	8154000
Maude Creek prospect	16°45'00"	127°46'00"	368538	8147686
Mount Chambers	16°29'40"	128°28'14"	443500	8176300
Mount Lookout	16°39'52"	128°00'45"	394700	8157300
Mount Nyulasy	16°45'05"	128°16'43"	423100	8147800
Neil Creek	16°56'33"	127°47'46"	371800	8126400
Nellie Range	16°55'05"	127°38'52"	356000	8129000
No. 4 Bore	16°55'27"	128°17'14"	424100	8128700
O'Donnell Range	16°34'38"	128°06'00"	404000	8167000
Pitt Range	16°45'00"	128°27'54"	443000	8148000
Pompeys Pillar	16°36'56"	128°12'55"	416300	8162800
Pompeys Pillar prospect	16°37'10"	128°15'47"	421400	8162400
Radigans Yard	16°56'33"	128°08'13"	408100	8126600
Ragged Range	16°19'29"	128°21'14"	431000	8195000
Rosewood Wall	16°28'40"	128°58'20"	390296	8177928
Speewah Valley	16°22'07"	127°56'31"	387000	8190000
Speewah Yards	16°26'47"	127°57'17"	388400	8181400
Spillway Creek	16°01'39"	128°46'46"	476400	8228000
Stonewall Creek	16°01'04"	128°52'26"	486500	8229100
Turkey Creek	16°56'28"	128°16'36"	423000	8126800
Ulysses Bay	16°16'17"	128°39'47"	464000	8201000
Wesley Yard	16°48'11"	128°20'28"	429800	8142100
Wilson River	16°41'52"	127°59'23"	392300	8153600

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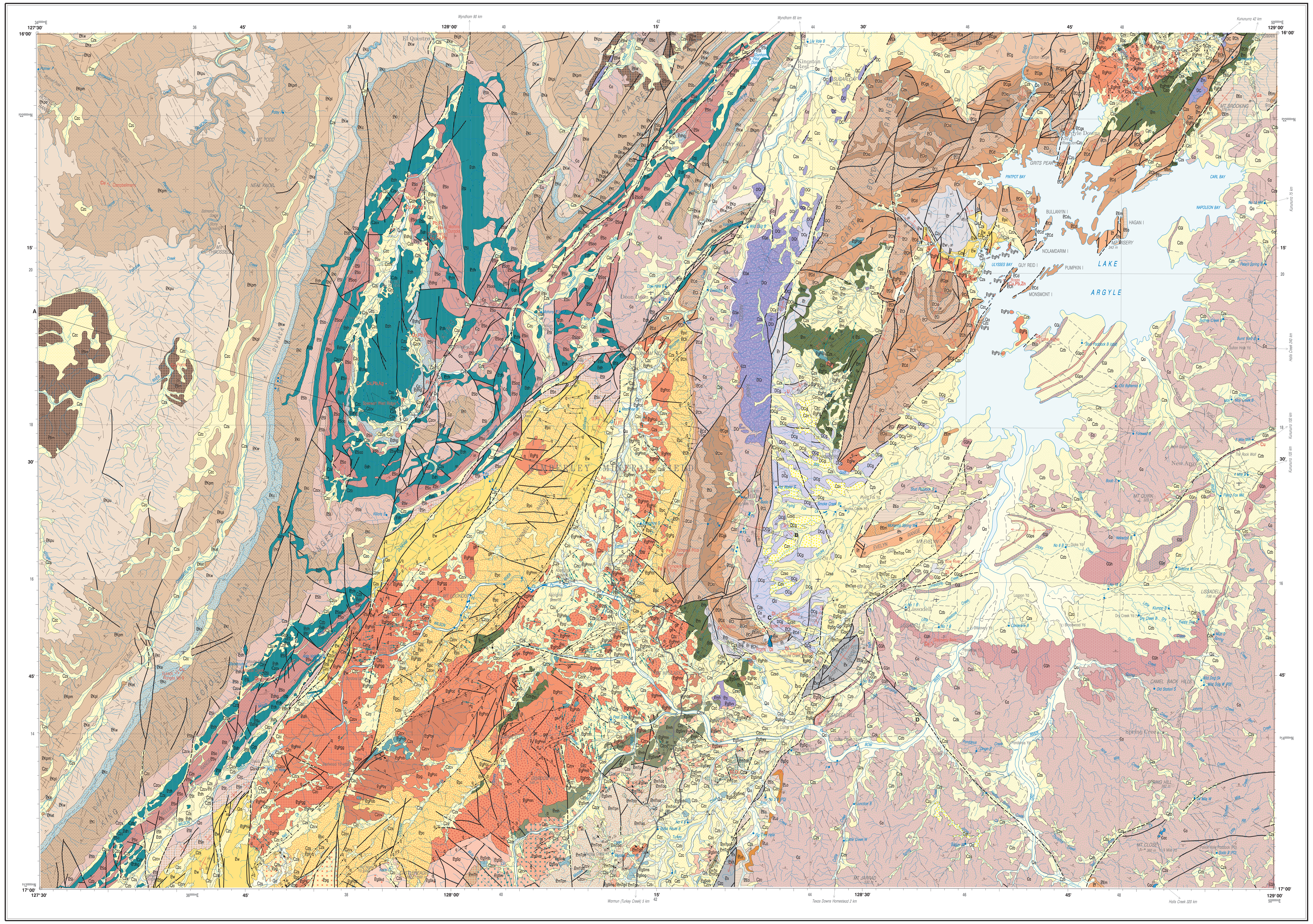

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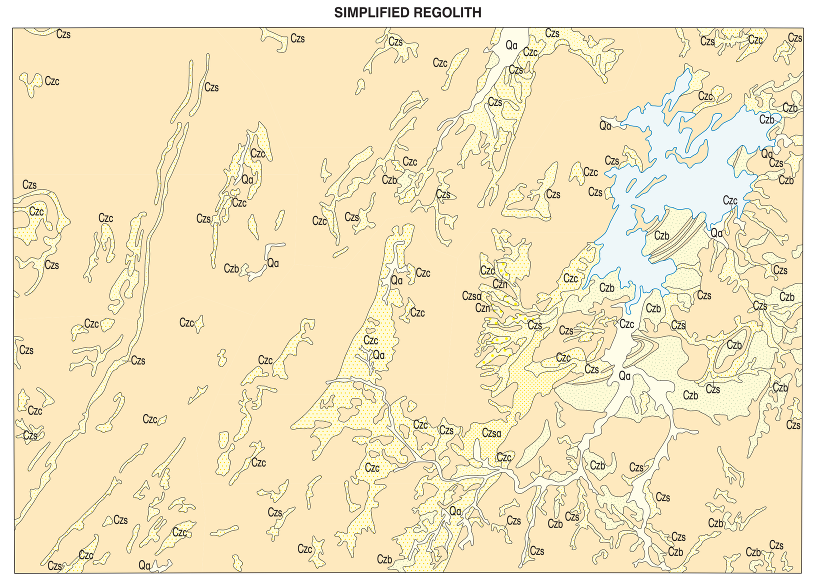

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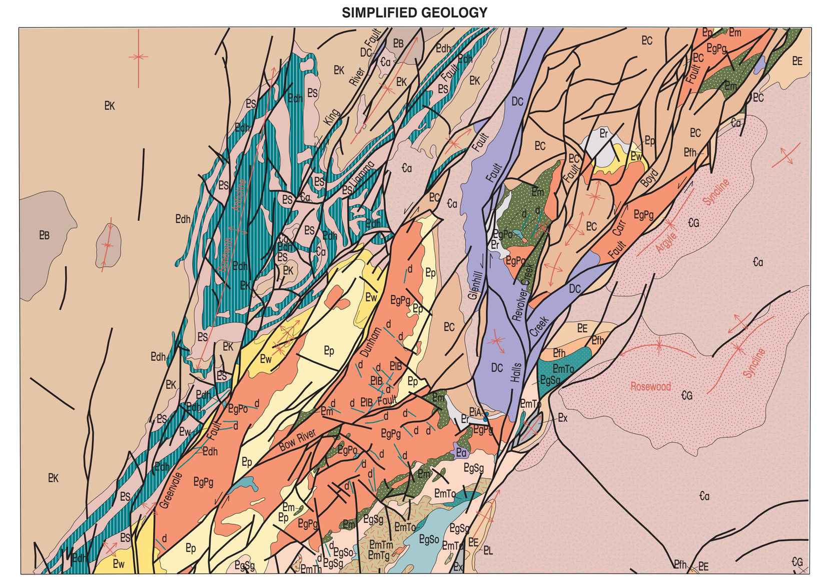
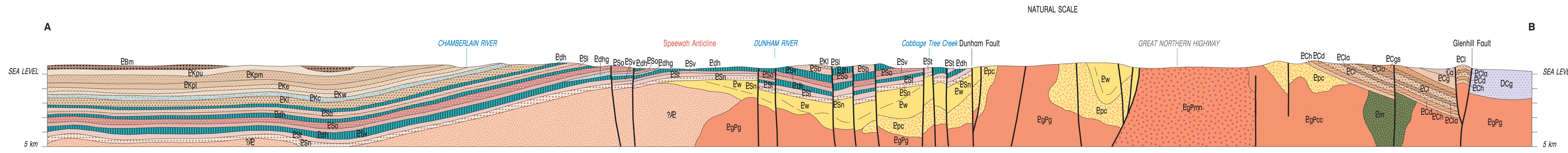
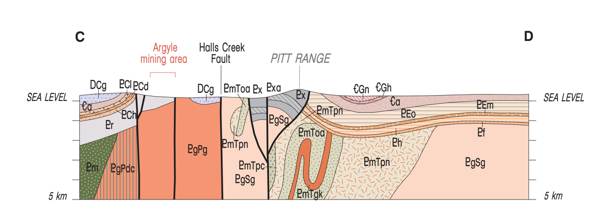


DIAGRAMMATIC SECTIONS

NATURAL SCALE





ƐcCy Yungul carbonatite: calcite–apatite–anatase rock with abundant inclusions of sedimentary rock and dolerite

c. 1790 Ma

Ɛdh HART DOLERITE: dark grey dolerite
Ɛdhg Pink to grey granophyre

SYMBOLS

Geological boundary
- exposed
- concealed

Structural symbols are labelled according to the sequence of deformation events, where known
- Alice Springs Orogeny (400–300 Ma)
  - D_7
- King Leopold Orogeny (c. 560 Ma) (not present on map)
  - D_6
- Yampi Orogeny (c. 1000 Ma)
  - D_5
- Halls Creek Orogeny (1835–1805 Ma)
  - D_4
  - D_3
- Hooper Orogeny (1865–1850 Ma)
  - D_2
  - D_1

Fault
- exposed
- concealed
- thrust

Fold, showing trend and plunge direction
- anticline; exposed
- syncline; exposed, concealed
- overturned anticline; exposed

Small-scale fold axial surface, showing strike and dip
- inclined
- vertical

Small-scale fold axis, showing trend and plunge
- minor anticline;
- minor syncline;
- M-vergence
- S-vergence

Bedding, showing strike and dip
- inclined
- vertical
- horizontal
- overturned
- strike and dip estimated from aerial photography
  - 0–15°
  - 15–45°
  - 45–90°
  - horizontal
  - trend of bedding

Way-up indicator
- sedimentary structure

Foliation, showing strike and dip
- inclined
- vertical
- trend of foliation

Gneissic banding, showing strike and dip
- inclined
- vertical

Cleavage, showing strike and dip
- inclined
- vertical

Crenulation cleavage, showing strike and dip
- inclined

Lineation, showing trend and plunge direction
- stretching lineation; inclined
- bedding–cleavage intersection; inclined
- axis of crenulation; inclined

Airphoto lineament
- structural trend

Highway with national route marker

Formed road

Track

Homestead — Lissadell

Yard — Yd

Building

Microwave repeater station

Reserve boundary

Horizontal control; minor

Watercourse; permanent, ephemeral, pool — P

Rockhole, spring, soak — RH, S, Sk

Bore, well — B, W

Windpump

Dam, tank — D, T

Abandoned, position doubtful — (abd) (PD)

Major opencut — Argyle

Opencut, abandoned

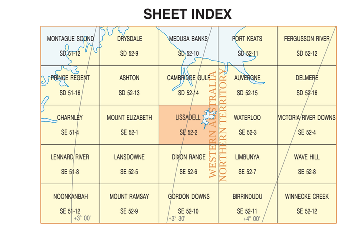
Alluvial workings

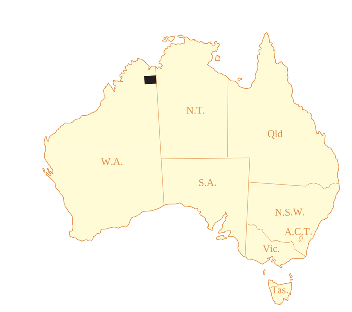
Prospect

Mining area

Mineral occurrence
- Barite — Ba
- Copper — Cu
- Diamond — Di
- Fluorite — Fl
- Gemstone (zebra stone) — Gs
- Gold — Au
- Iron — Fe
- Lead — Pb
- Lithium — Li
- Nickel — Ni
- Silver — Ag
- Tantalum — Ta
- Tin — Sn
- Uranium — U

SHEET INDEX





Geology by A. M. Thorne, S. Sheppard, and I. M. Tyler 1994–1995

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Edited by C. Strong and G. Loan

Cartography by P. Taylor and B. Williams

Topography from the Department of Land Administration Sheet SE 52-2, with modifications from geological field survey

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GEOLOGICAL SURVEY OF WESTERN AUSTRALIA
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SCALE 1:250 000

5000 Metres 0 5 10 15 20 25 Kilometres

UNIVERSAL TRANSVERSE MERCATOR PROJECTION
HORIZONTAL DATUM: GEOCENTRIC DATUM OF AUSTRALIA 1994
VERTICAL DATUM: AUSTRALIAN HEIGHT DATUM

Grid lines indicate 20 000 metre interval of the Map Grid Australia Zone 52

The Map Grid Australia (MGA) is based on the Geocentric Datum of Australia 1994 (GDA94)
GDA94 positions are compatible within one metre of the datum WGS84 positions

Reference points to align maps based on the previous datum, AGD84, have been placed near the map corners

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