EXPLANATORY NOTES

GOVERNMENT OF WESTERN AUSTRALIA

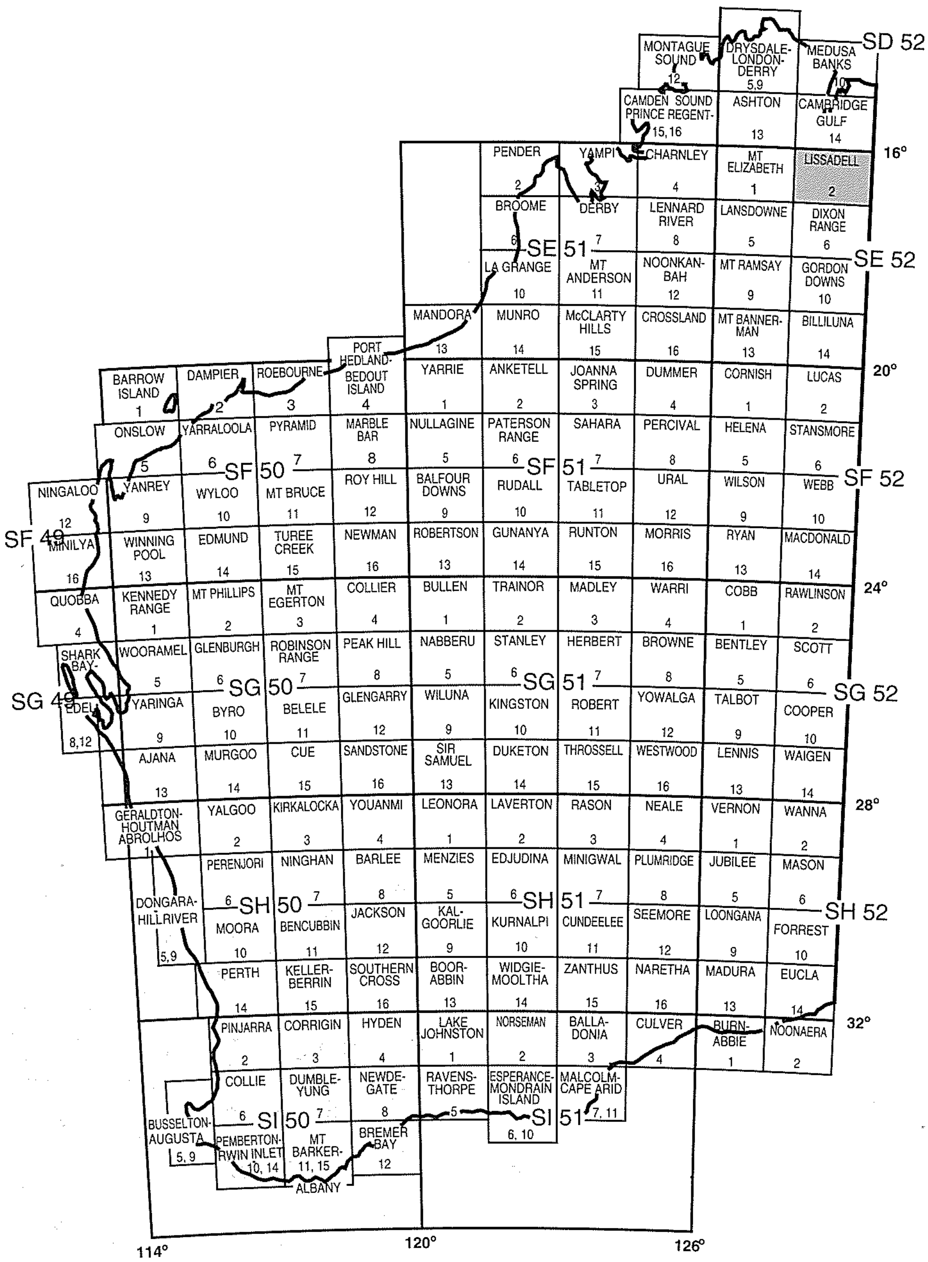
GEOLOGY OF THE BOW 1:100 000 SHEET

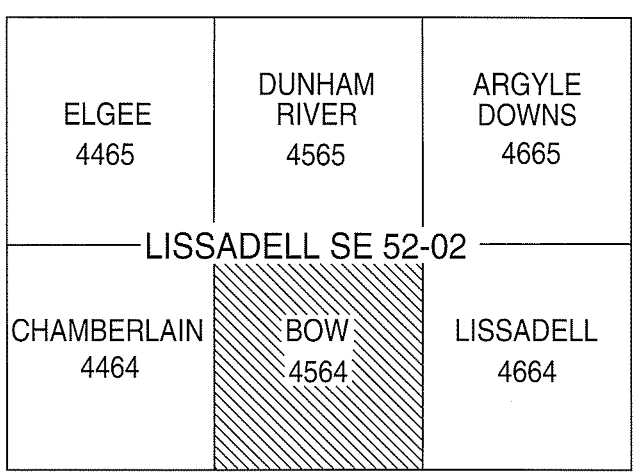
by S. Sheppard, A. M. Thorne, and I. M. Tyler

1:100 000 GEOLOGICAL SERIES

GEOLOGICAL SURVEY OF WESTERN AUSTRALIA

DEPARTMENT OF MINERALS AND ENERGY





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Cover photograph:
Mesoproterozoic sedimentary rocks of the Carr Boyd Group along the western edge of the Ragged Range, in the northeastern part of the BOW 1:100 000 sheet.

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Geology of the Bow 1:100 000 sheet

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Abstract

The Bow 1:100 000 map sheet (SE 52-2, 4564) lies entirely within the Halls Creek Orogen, a major northeasterly trending orogenic belt developed in the Palaeoproterozoic to Phanerozoic rocks of the east Kimberley region of Western Australia. The Halls Creek Orogen initially formed in the Palaeoproterozoic, between the Kimberley Craton to the northwest and the North Australian Craton to the east.

The oldest rocks on Bow 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 volcanic rocks 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 Bow 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 affected rocks of the Eastern zone of the Lamboo Complex, and reflected the suturing of the Kimberley and North Australian Cratons 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 Kimberley Craton and Western zone of the Lamboo Complex.

At c. 1800 Ma clastic sediments and basalt of the Texas Downs and Revolver Creek Formations were deposited, respectively in the Texas Downs and Revolver Creek Basins on the eastern part of Bow. These sedimentary rocks are lithologically similar to the Kimberley Group to the west. 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 edge of Bow. 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 are overlain by Cambrian siliciclastic and carbonate rocks of the Ord Basin, and Late Devonian coarse-grained siliciclastic rocks of the Bonaparte Basin. 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 draining 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 Bow* 1:100 000 geological sheet (SE 52-2, 4564) is bounded by latitudes 17°00' and 16°30'S and longitudes 128°00' and 128°30'E. The map sheet lies within the LISSADELL 1:250 000 map sheet in the east Kimberley region of Western Australia.

The main commercial activities on Bow are diamond mining at the Argyle mine, and cattle grazing for beef. The Lissadell, Bow River, and Doon Doon pastoral leases cover the map sheet. The Bow River Homestead in the southwestern part of the sheet is permanently occupied, as are aboriginal settlements at Glen Hills and Crocodile Hole in the northern part of the sheet area.

The sealed Great Northern Highway traverses the centre of the sheet, linking Kununurra 140 km to the north and Halls Creek 150 km to the south. Access within the rest of the sheet is via graded roads and station tracks, most of which are accessible only during the dry season. All-weather access to the Argyle diamond mine in the eastern part of the sheet, and the Bow River diamond mine on LISSADELL, is provided by a graded road.

Geological investigations prior to 1968 on Bow are covered in the explanatory notes for the first edition LISSADELL 1:250 000 geological sheet (Plumb, 1968). More recent work is referred to as appropriate in the following 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. Fieldwork was carried out in 1994 and 1995 using 1:40 000 black and white 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 Kimberley–Arunta project.

Physiography, vegetation, and climate

The physiographic units on Bow are shown on Figure 1. Most of Bow lies within the Lamboo Hills Province of the Ordland Division and the Kimberley Foreland Province of the North Kimberley Division (Plumb, 1968; Beard, 1979). The northeast edge of the sheet lies within the Ord Plains Province of the Ordland Division.

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.

On Bow, the Kimberley Foreland Province consists of the Kimberley Foothills and Carr Boyd Ranges in the north, and the Osmand – Albert Edward Ranges in the southeast. The O'Donnell Range consists of rugged, highly dissected topography with local relief of up to 200 m. The western edge of the Carr Boyd Ranges is a high scarp which slopes eastwards in a series of sandstone cuestas (Plumb, 1968). The 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 Ord Plains Province consists of widespread, low-lying plains developed over poorly exposed Cambrian basalts and carbonate (Plumb, 1968).

The vegetation of the Kimberley region is described by Beard (1979). Felsic rocks of the Lamboo Hills Province usually support a low savanna woodland of snappy gum over curly spinifex and cane grass. Meta-sedimentary rocks of the Kimberley Foreland Province commonly support a tree steppe of snappy gum over hard spinifex or a sparse-tree steppe of snappy gum over soft spinifex. 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.

Bow has a semi-arid monsoonal climate with an average annual rainfall of 550 to 700 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 maximum 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 per annum (Beard, 1979), so that all water-courses are intermittent. Some creeks contain waterholes that persist until late in the dry season.

Regional geological setting

The regional geological setting of Bow is shown in Figure 2, and the major tectonic units on Bow are shown in Figure 3. A summary of the geological history of Bow is presented in Table 1. The stratigraphy of the Palaeo-proterozoic to Palaeozoic sedimentary rocks on Bow is summarized in Table 2. The map sheet lies within the Halls Creek Orogen, which formed in the Palaeo-proterozoic between the Kimberley Craton to the northwest and a composite Archaean craton to the east (Tyler et al., 1995). The Halls Creek Orogen was the locus of repeated fault reactivation and deformation during the Mesoproterozoic, Neoproterozoic, and Phanerozoic. Most of Bow is underlain by crystalline rocks that are part of the c. 1920–1780 Ma Lamboo Complex (Dow and Gemuts, 1969; Griffin and Grey, 1990; Page and Sun, 1994; Tyler et al., 1995), which is the oldest component of the orogen. The complex is a north-northwesterly trending belt of Palaeoproterozoic igneous and low- to

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

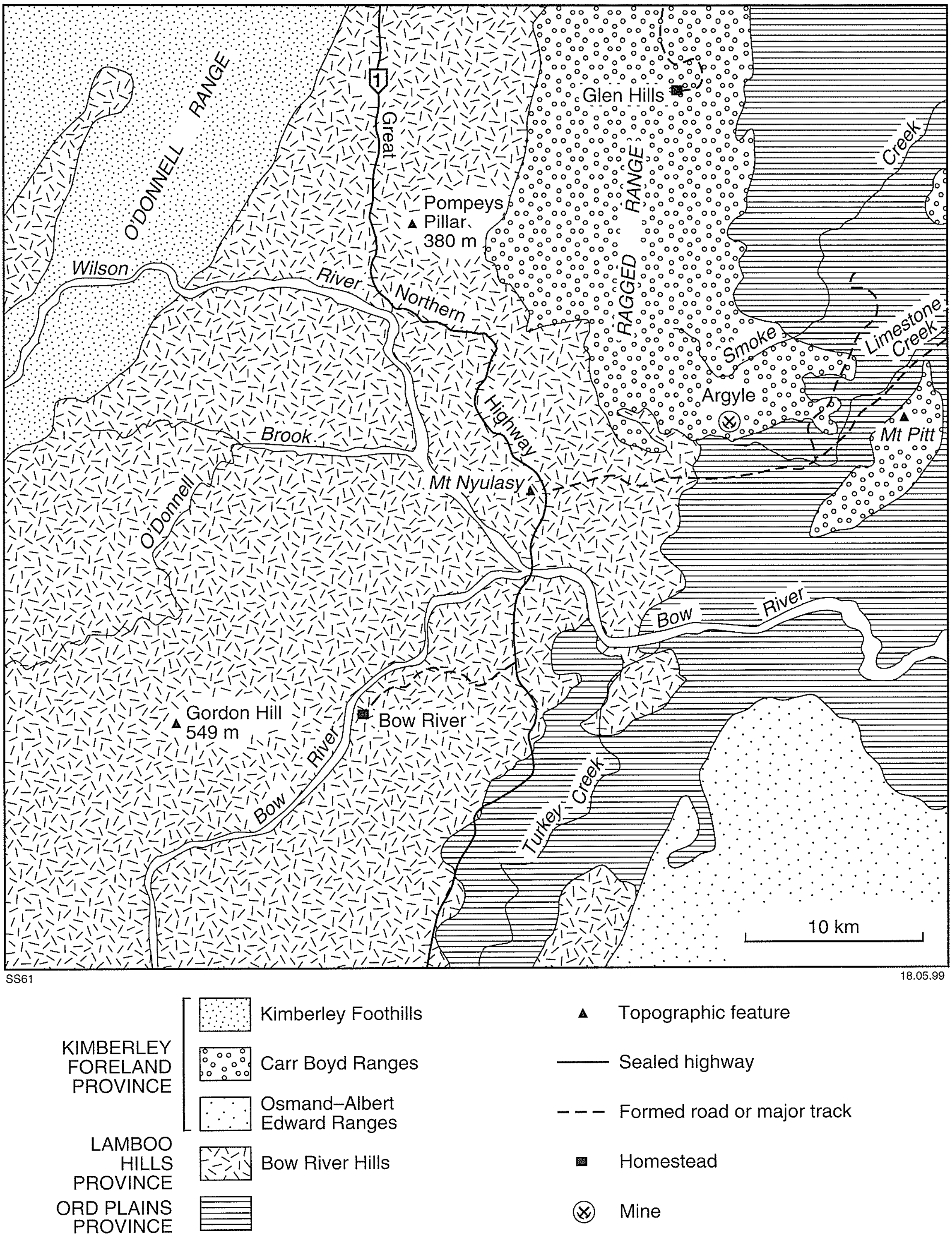

Figure 1. Physiographic and drainage sketch map of Bow

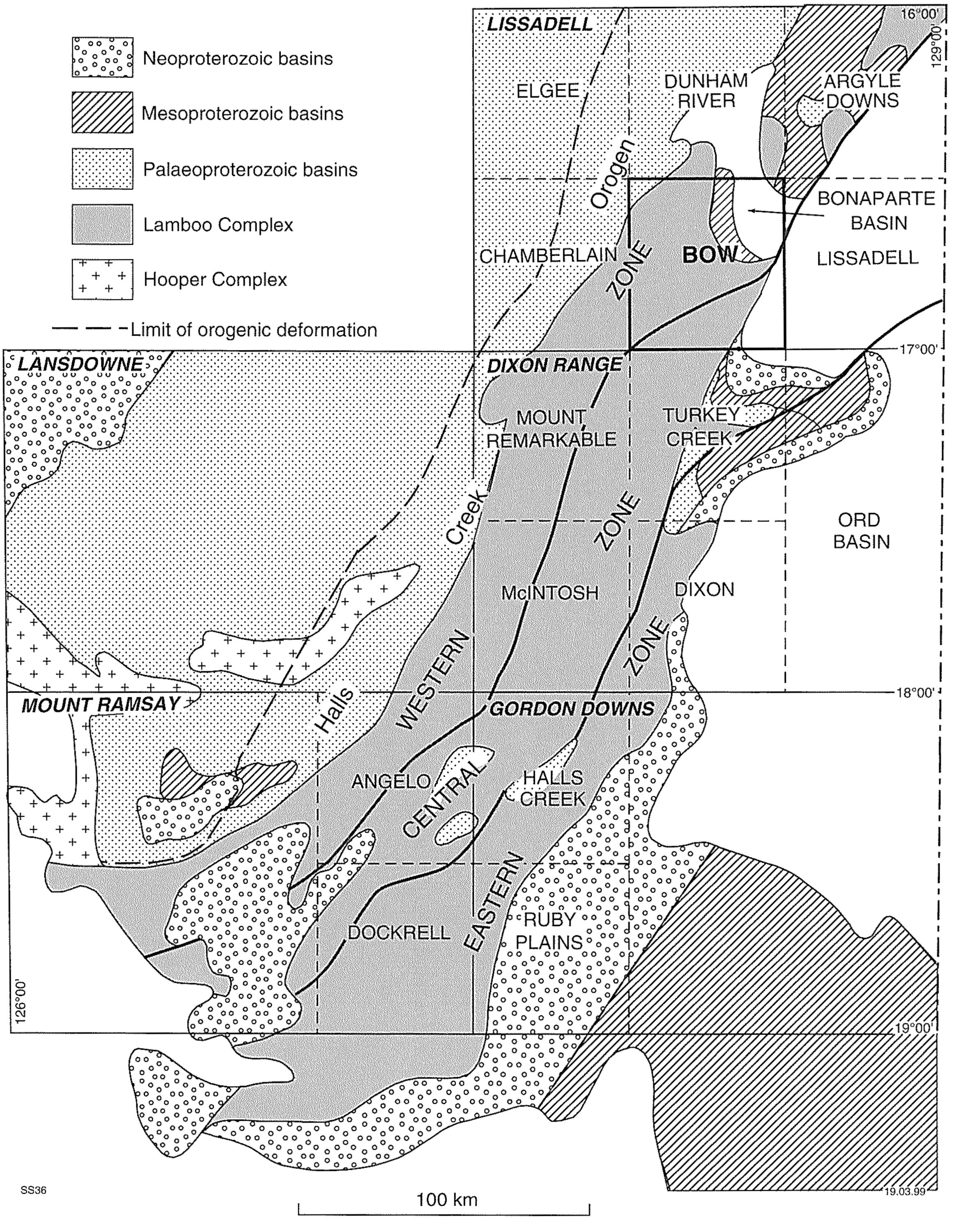

Figure 2. Location of published 1:100 000 and 1:250 000 map sheets in the east Kimberley region and their relationship to tectonic zones in the Lamboo Complex

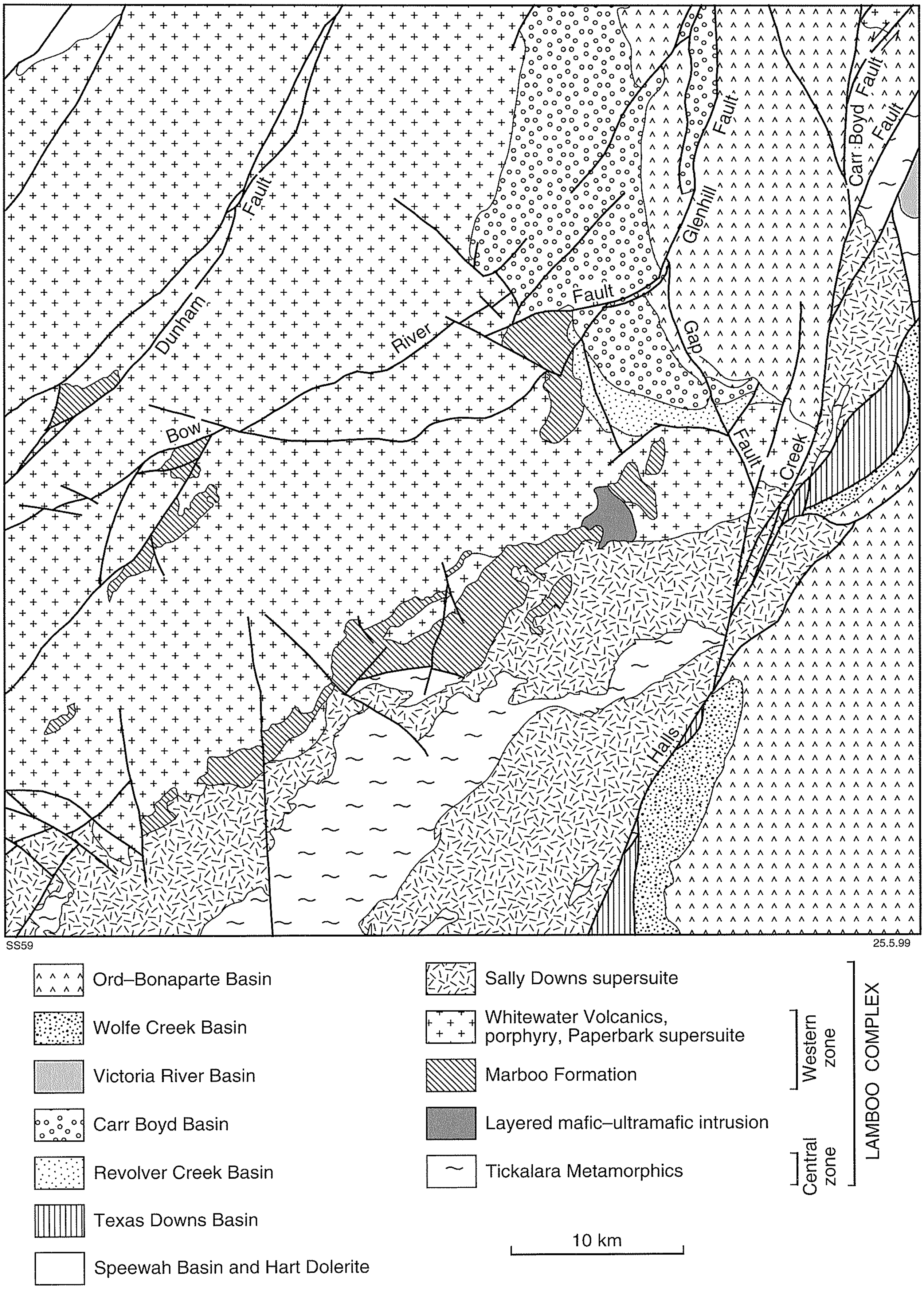

Figure 3. Simplified geological map of Bow

Table 1. Summary of the geological history of Bow

Age (Ma)	*Speewah Basin*	*Lamboo Complex*	
		Western zone	*Central zone*
>2500–1950?	Unexposed Kimberley Craton		
c. 1870		Deposition of the Marboo Formation	
1865–1850		HOOPER OROGENY	
c. 1865		D_1 in Western zone	Deposition of the protolith of the Tickalara Metamorphics 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–trondhjemite–granodiorite sheets
c. 1850			Easterly directed thrusting — large-scale recumbent isoclinal D_2 folds. 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)
1835–1805		Intrusion of other Sally Downs supersuite granitoid and gabbro	
1830–1800			Tight, upright, northeasterly plunging D_4 folds refold D_3 folds Activation of Highway Shear Zone under upper greenschist to lower amphibolite facies conditions (M_4)
1827–1790		Deposition of the Texas Downs and Revolver Creek Formations?	
1800–1790	Intrusion of Hart Dolerite sills into the Speewah Group		
1790–1250		Folding, uplift and erosion	
1250		Deposition of the Carr Boyd Group	
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 Group and Albert Edward Group	
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 Group	
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	

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

Basin	*Group*	*Formation*	*Thickness (m)*	*Lithology*
Bonaparte Basin	Cockatoo Group	Galloping Creek Formation (*Dcg*)	1 600	Feldspathic sandstone, pebbly sandstone, pebble- to boulder-conglomerate
		unconformity/disconformity		
Ord Basin	Goose Hole Group	Nelson Shale (*ЄGn*)	100–180	Purple siltstone, minor sandstone (section only)
		Headleys Limestone (*ЄGh*)	35–50	Grey, massive to laminated limestone, minor chert, and stromatolites
Ord and Bonaparte Basins		Antrim Plateau Volcanics (*Єa*)	600–1 000	Massive or amygdaloidal basalt, minor medium-grained sandstone
		unconformity/disconformity		
Wolfe Creek Basin	Albert Edward Group	Mount Forster Sandstone (*ᵽLo*)	40	Quartz sandstone, pebbly sandstone, and conglomerate
	Duerdin Group	Ranford Formation (*ᵽEo*)	600	Siltstone, quartz sandstone, dolomitic sandstone
		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	Glenhill Formation (*ᵽcg*)	300	Micaceous siltstone and quartz sandstone
		Lissadell Formation (*ᵽcl*)	1200	Quartz sandstone and siltstone
		Golden Gate Siltstone (*ᵽcd*)	50–60	Chloritic and carbonaceous siltstone, quartz sandstone, and ferruginous sandstone
		Hensman Sandstone (*ᵽch*)	120	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, conglomerate, and basalt
		unconformity/disconformity		
Speewah Basin	Speewah Group	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

high-grade metamorphic rocks unconformably overlain by the c. 1835 Speewah Basin and 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 are 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).

Earlier models for the formation of the Lamboo Complex, and other belts of similar age in northern Australia, proposed extension and crustal thinning, then convergence without subduction of oceanic crust (Hancock and Rutland, 1984; Etheridge et al., 1987; Wyborn, 1988). However, Ogasawara (1988) noted that the chemistry of tonalites in the Lamboo Complex 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; 1997a) have argued that the Lamboo Complex shares many features with convergent Phanerozoic plate margins associated with subduction of oceanic crust.

Before and after collision of the Kimberley and North Australian Cratons, plutons of granitoid and gabbro intruded the Lamboo Complex at the same time as sedimentary rocks in the Speewah Basin were being deposited to the west (Tyler et al., 1995; Page and Sun, 1994). Sedimentary rocks equivalent to the Kimberley Group (c. 1800 Ma) extended across the Lamboo Complex (Dow and Gemuts, 1969; Plumb and Gemuts, 1976; Plumb et al., 1985; Tyler et al., 1995).

Large-scale sinistral strike-slip faulting took place in the Halls Creek Orogen during the Mesoproterozoic Yampi Orogeny (c. 1400–1000 Ma). During the orogeny, a pattern of north-northeasterly trending synthetic sinistral faults, and east-northeasterly trending antithetic dextral faults was established (Tyler et al., 1995; Thorne and Tyler, 1996). Deformation was accompanied by low- to medium-grade metamorphism. 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., 1980).

The Neoproterozoic Ahern Formation and Helicopter Siltstone in the eastern part of LISSADELL are correlated with units in the Victoria River Basin and represent Supersequence 1 of the Centralian Superbasin (Thorne and Tyler, 1996; this publication, p. 27). 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). Their deposition 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 of the Antrim Plateau Volcanics (Mory and Beere, 1988). Late Devonian alluvial fans and braided rivers 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

Introduction

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 Bow (Figs 2 and 3). 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 turbiditic 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 deformed and metamorphosed, and were extensively intruded by granitoid and gabbro of the Paperbark supersuite, as well as by subvolcanic porphyries during the Hooper Orogeny between c. 1865 and 1850 Ma (Tyler and Page, 1996; Griffin et al., in prep.).

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. Large volumes of granitoid and gabbro of the Sally Downs supersuite intruded the Central zone during the Halls Creek Orogeny at 1835–1805 Ma.

The Eastern zone is composed of low-grade metasedimentary and metavolcanic rocks of the Halls Creek Group, which unconformably overlie domal culminations of 1920 to 1900 Ma granitoid and associated volcanic rock. Alkaline volcanic rocks of the Butchers Gully Member (Griffin and Tyler, 1992b) and Maude Headley Volcanic Member (Blake et al., 1998) in the middle and upper part of the sequence were erupted between c. 1857 and 1848 Ma (Blake et al., 1998). A series of dolerite sills, known collectively as the Woodward Dolerite, intruded the Halls Creek Group before both were deformed and metamorphosed. Following deformation and metamorphism, the Halls Creek Group and Woodward Dolerite were intruded by c. 1820–1810 Ma granitoid of the Sally Downs supersuite at the southern end of the Lamboo Complex.

Western zone

Marboo Formation (*Ꞓm, Ꞓmh*)

Rocks of the Marboo Formation on Bow were previously placed within the Tickalara Metamorphics or designated as undifferentiated Halls Creek Group (Plumb, 1968). Although the rocks are lithologically similar to the Olympio Formation of the Halls Creek Group, the Marboo Form-ation was deformed before being intruded by granitoids dated at c. 1860 to 1850 Ma. 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. The Marboo Formation in the Hooper Complex was deformed and metamorphosed prior to the intrusion of granitoid plutons at c. 1865 to 1850 Ma. The youngest detrital zircons in the Marboo Formation indicate a maximum age of c. 1872 Ma (Tyler et al., in press).

The Marboo Formation outcrops within granitoids in the central part of Bow. To the east of Bow River Homestead, the Marboo Formation is faulted against the Tickalara Metamorphics.The rocks that are included in the Marboo Formation (*Ꞓm*) are typically low grade, except where they have been metamorphosed within the contact aureoles of granitoid plutons of the Sally Downs supersuite (*Ꞓmh,* see below). The rocks that are not hornfelsed 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 was recognized. Griffin et al. (1993) estimated the thickness of the Marboo Formation on the LENNARD RIVER 1:250 000 sheet 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 Bow is consistent with them belonging to the upper part of the unit (Griffin et al., 1993). The sandstone beds are typically up to 1 m thick, and show graded bedding, passing up into siltstone and mudstone. They are interpreted as turbidity current deposits (e.g. Walker, 1984), with partial preservation of Bouma cycles (units ADE). Palaeocurrent data have not been obtained from Bow, but Hancock (1991) recorded palaeocurrent data on LANSDOWNE (1:250 000) that indicated a depositional slope from the north and northeast.

The arenite units consist of subangular to subrounded clasts of quartz, plagioclase, K-feldspar, and quartzo-feldspathic 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 (*Ꞓw*)

The Whitewater Volcanics are exposed as a belt about 600 km long and up to 30 km wide that extends from the western end of the Hooper Complex to the northern end of the Western zone of the Lamboo Complex. Page and Hancock (1988) obtained a conventional U–Pb zircon age of 1850 ± 5 Ma from a sample of the Whitewater Volcanics just north of Bow. Griffin et al. (in prep.) re-analyzed the same sample using SHRIMP U–Pb in zircon, and obtained an age of 1854 ± 5 Ma. This age is identical to that obtained from a sample of the Whitewater Volcanics in the Hooper Complex by Griffin et al. (in prep.). In the Hooper and Lamboo Complexes, the Whitewater Volcanics is intruded by granitoids ranging from about 1865 to 1850 Ma. 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 (Griffin et al., in prep.).

The Whitewater Volcanics is composed of dacitic to rhyolitic ignimbrite and quartz–feldspar porphyry, lesser lava flows, and minor lapilli tuff and volcaniclastic rock (Dow and Gemuts, 1969; Gellatly et al., 1975; Sheppard et al., 1997b). The large volume of ignimbrite, in conjunction with a paucity of non-volcanic sedimentary rocks and well-bedded volcaniclastic sedimentary rocks, indicates a largely subaerial environment.

On Bow the Whitewater Volcanics is restricted to the extreme northwestern corner of the map sheet, where this unit forms rugged dissected hills with about 100 m relief. On the first edition LISSADELL 1:250 000 sheet (Plumb et al., 1968) the Whitewater Volcanics is shown as forming a mass about 5–10 km wide and 50 km long adjacent to the Dunham Fault in the O'Donnell Range. Most of this area consists of a massive, fine-grained feldspar porphyry, very similar in appearance to the Castlereagh Hill Porphyry (Plumb, 1968). The rocks have a pattern similar to the Castlereagh Hill Porphyry on aerial photographs and Landsat TM images, and there is no evidence for an extrusive origin. Therefore, most of the Whitewater Volcanics adjacent to the Dunham Fault is reassigned to the Castlereagh Hill Porphyry.

The Whitewater Volcanics at the northern end of the O'Donnell Range shows well-defined layering, and rock types include quartz–feldspar porphyry, columnar-jointed, welded ignimbrite with strongly flattened pumice fragments, volcanic breccia ('agglomerate' on the map legend), lapilli tuff, and various fine-grained quartz-rich sedimentary rocks (Plumb, 1968). The quartz–feldspar porphyry is thick bedded, massive and fine to medium grained, and strongly resembles much of the Castlereagh Hill Porphyry. Some of the quartz–feldspar porphyry in the Whitewater Volcanics is probably intrusive (Dow and Gemuts, 1969).

On the LISSADELL 1:250 000 sheet the Whitewater Volcanics is unconformably overlain by the O'Donnell Formation, which defines the base of the Speewah Group. A strip of Whitewater Volcanics east of Aida Vale Mill on LISSADELL (about 15 km west of Bow) is intruded by coarse-grained porphyritic granite.

Castlereagh Hill Porphyry (*Ɇpc*)

The Castlereagh Hill Porphyry of Dow and Gemuts (1969) and Gemuts (1971) is composed of two distinctly mappable units: porphyry with medium to coarse phenocrysts of quartz, K-feldspar and plagioclase, and 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 exposed mainly to the southwest on MOUNT REMARKABLE (Sheppard et al., 1997b).

The Castlereagh Hill Porphyry has not been dated. However, it strongly resembles the quartz–feldspar porphyry constituting much of the Whitewater Volcanics, and it also has an identical texture and mineralogy to that of the Crooked Creek Granite of the 1865–1850 Ma Paperbark supersuite, differing only in its finer grain size. Also, strong similarities in whole-rock chemistry between the Whitewater Volcanics, high-level porphyry intrusions, and granitoids of the Paperbark supersuite, suggest that the three units are cogenetic (Griffin et al., in prep.).

On Bow, the Castlereagh Hill Porphyry outcrops mainly in two areas: a north-northeasterly trending belt, 5–10 km wide by 30 km long, in the O'Donnell Range; and a northerly trending belt, 3 km wide by 20 km long, underneath the scarp along the western edge of the Ragged Range. About 35 km^2 of porphyry that outcrops southwest of Fish Hole Yard (AMG 993396)* is part of a much larger mass extending to the southwest. In the O'Donnell Range the porphyry forms a dissected plateau about 200–250 m above the surrounding granitoid country. Elsewhere, the porphyry forms rounded or steep hills with a smoother pattern on aerial photographs and Landsat TM images than the granitoids.

The bulk of the Castlereagh Hill Porphyry consists of dark-grey, massive microgranodiorite with fine- to medium-grained phenocrysts of plagioclase. Locally (e.g. 2–3 km southeast of Mount Lookout), the unit may coarsen to a fine- to medium-grained groundmass with phenocrysts up to 7 mm long. These coarser grained parts of the Castlereagh Hill Porphyry strongly resemble the Crooked Creek Granite of the Paperbark supersuite.

The contact between the Castlereagh Hill Porphyry and the Whitewater Volcanics was not observed on Bow. The relationship between the porphyry and individual plutons of the Paperbark supersuite on Bow is not known, and faulting has complicated many of the contacts. However, 5 km south of Pompeys Pillar, just north of the highway (AMG 163580), 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). The 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 from the Castlereagh Hill Porphyry are moderately to strongly altered. Microcrystalline sericite and calcite (or clinozoisite) replaces phenocryst and groundmass plagioclase. Hornblende is pseudomorphed by chlorite, quartz and epidote, and chlorite and quartz replace biotite. 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.

Fine-grained, aphyric mafic inclusions are widespread, although only locally abundant. The inclusions are rounded or elliptical, and generally 1–4 cm in diameter. They have a subhedral granular igneous texture. The inclusions are composed of plagioclase, quartz, chlorite, epidote, K-feldspar, and accessory apatite, opaque minerals, titanite, and zircon.

Greenvale Porphyry (*Ɇpg*)

The Greenvale Porphyry is a massive, dark-grey quartz–feldspar porphyry with medium to coarse phenocrysts of quartz, sanidine, and plagioclase. On Bow, this unit is restricted to two small exposures, 2 km south of Bow Hill and 7 km west of Bow Hill, each of which is less than 0.1 km^2 in area, but it outcrops extensively on the northern part of MOUNT REMARKABLE to the southwest. The occurrence south of Bow Hill is located along the margin

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

of the Mount Nyulasy Granite (AMG 145521), whereas the other is situated between the Mount Nyulasy Granite and a large pendant of coarse-grained gabbro (AMG 087546).

A sample of the Greenvale Porphyry from MOUNT REMARKABLE (AMG 722013) yielded a SHRIMP U–Pb zircon age of 1855 ± 2 Ma (Griffin et al., in prep.). This age is indistinguishable from that of the Mount Nyulasy Granite (see below), with which the porphyry shares some textural features. The age of the Greenvale Porphyry is also inseparable from that of the Whitewater Volcanics (see above) 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, but the porphyry has a much finer grained groundmass. At a granite quarry 10 km north of Pompeys Pillar (AMG 165728), the Mount Nyulasy Granite contains inclusions of a rock type resembling the Greenvale Porphyry. A dyke of Greenvale Porphyry intruded the Dinner Creek Tonalite about 3.5 km west-southwest of the Argyle diamond mine.

Porphyry samples are composed of 20–30% phenocrysts of white, subhedral plagioclase up to 5 mm long, oblate quartz 5–10 mm long, and rounded sanidine up to 20 mm in diameter. Quartz phenocrysts contain some groundmass inclusions, and sanidine has a patchwork of small plagioclase, groundmass and quartz inclusions. The sanidine phenocrysts typically have a narrow rim of plagioclase. The groundmass consists of fine-grained quartz, plagioclase, microcline, biotite (commonly altered to chlorite), hornblende (pseudomorphed by chlorite and epidote), and accessory opaque minerals, apatite, and zircon. Plagioclase phenocrysts are strongly altered to very fine grained sericite and clinozoisite. This alteration is typically accompanied by partial replacement of magnetite by epidote. Micrographic and granophyric intergrowths between quartz and K-feldspar, and myrmekitic intergrowths between quartz and plagioclase are abundant in the groundmass.

Paperbark supersuite

SHRIMP U–Pb geochronology indicates that the 'Bow River Granite' of Dow and Gemuts (1969) and the 'Bow River Granitoid Suite' of Ogasawara (1988) contains a wide range of granitoid compositions ranging in age from 1865 to 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, along with granitoids in the Hooper Complex, 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, which intruded the Western zone of the Lamboo Complex and the Hooper Complex in the west Kimberley region. Although the tholeiitic magma event was 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 in the Paperbark supersuite are common farther to the south on McINTOSH and MOUNT REMARKABLE (Blake and Hoatson, 1993; Sheppard, 1996).

The Paperbark supersuite on Bow contains several newly defined intrusions; the Beefwood Yard Granite, Crooked Creek Granite, Dinner Creek Tonalite, Gordons Gorge Granite, and the Mount Nyulasy Granite.

Biotite gabbro (ꞒgPob)

On Bow, massive biotite-rich gabbros of 1865–1850 Ma age are restricted to several isolated outcrops (each <1 km^2) surrounded by granitoid. Rock types consist of fine-grained biotite gabbro with xenocrysts of quartz and feldspar (e.g. AMG 136515), and medium- to coarse-grained biotite gabbro. Farther south on MOUNT REMARKABLE, Sheppard et al. (1997b) showed that quartz and feldspar xenocrysts in biotite gabbro were the product of mingling between mafic and felsic magmas.

Beefwood Yard Granite (ꞒgPby)

Only about 10 km^2 of the Beefwood Yard Granite outcrops on the western edge of Bow. The main part of the intrusion extends to the southwest onto CHAMBERLAIN. The intrusion is composed of massive, medium-grained, porphyritic biotite monzogranite and syenogranite.

Crooked Creek Granite (ꞒgPcc)

The Crooked Creek Granite is named after Crooked Creek, which flows into Bow River about 2 km north of the Bow River Homestead. The intrusion has not been dated, but it is confined to the Western zone, and thus it is assigned to the Paperbark supersuite. The bulk of the Crooked Creek Granite outcrops as an elongate (~20 km long by 5–7 km wide), northeasterly trending intrusion extending from west of Fish Hole Yard to north of Bow Hill in the centre of the sheet. This unit also outcrops discontinuously between the Castlereagh Hill Porphyry and Mount Nyulasy Granite in the northern part of the map sheet. 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 with about 15–20% phenocrysts of euhedral plagioclase.

Contacts between the Crooked Creek Granite and other rock units in the southwest of the map sheet are commonly complicated by faulting. About 6 km north-northwest of Pear Tree Bore (AMG 105481), 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 granitoid and porphyry are broadly coeval (see above).

The mineralogical and textural similarities of the two units suggest that they are related.

Most samples of the Crooked Creek Granite are medium grained (~2 mm average grainsize), but fine- to medium-grained rocks are also present. The rocks contain about 10–15% phenocrysts of subhedral plagioclase up to 4 mm long in a granular groundmass of plagioclase, quartz, microperthite, biotite, and amphibole with accessory opaque minerals, apatite, zircon, and secondary schorl. In some samples larger microperthite crystals contain numerous inclusions of plagioclase and quartz. Where they are weakly altered, the plagioclase phenocrysts display simple normal zoning. However, in most samples plagioclase is moderately to strongly altered to sericite and clinozoisite. In addition, biotite is commonly partly replaced by chlorite, and amphibole is pseudomorphed by chlorite and epidote. Magnetite and ilmenite are partly replaced by epidote and sphene respectively, in association with alteration of plagioclase. Very fine grained clots 1–2 mm in diameter, composed of quartz, green-brown biotite and minor epidote, are common in some samples.

Dinner Creek Tonalite (ꞒgPdc)

The Dinner Creek Tonalite is named after Dinner Creek, which flows into Bow River near Lissadell Hill in the east of the map sheet. The main mass of the intrusion outcrops over about 60 km^2 in the centre of the sheet, with a small plug or remnant outcropping about 2 km west-southwest of Bow River Homestead. It is a poorly exposed intrusion, in which outcrops commonly consist of scattered boulders and isolated low hills amongst colluvial plains.

The Dinner Creek Tonalite has not been dated, but it is intruded by the Mount Nyulasy Granite, which has a SHRIMP U–Pb zircon age of 1859 ± 3 Ma (see below). The Dinner Creek Tonalite consists of fine- to medium-grained, massive hypersthene–biotite–amphibole tonalite. Small inclusions of fine-grained tonalite are widespread and locally abundant, for example about 3–4 km southwest of the Argyle diamond mine. Angular inclusions of metasedimentary rocks are locally abundant.

The Dinner Creek Tonalite intruded and hornfelsed the Marboo Formation. The Dinner Creek Tonalite is intruded in several places by porphyritic monzogranite of the Mount Nyulasy Granite, namely 1.5 km west and 6 km northeast of Bow River Homestead (AMG 116342 and 193408) and south and east of Mount Nyulasy (AMG 221460 and 245475). About 1 km west of Wesley Yard (AMG 290421), 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 mine (AMG 319499) the Dinner Creek 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. Phenocryst and groundmass plagioclase have mottled calcic cores commonly separated from narrow sodic rims by a region of strong oscillatory zoning. The groundmass consists of subhedral to anhedral granular plagioclase, quartz (~20%), biotite (10–15%), actinolite (~5%), and rare microcline. Rare examples of antiperthitic textures are present in some groundmass plagioclase crystals. Accessory minerals include opaque minerals, apatite, zircon, and allanite.

Plagioclase crystals are moderately to strongly altered to very fine grained clinozoisite and minor sericite. Associated with this alterati0Oon is replacement of ?magnetite by epidote. Hypersthene in most samples is pseudomorphed by a weakly pleochroic, very pale green tremolite–actinolite. Where it is well preserved, hypersthene is rimmed by actinolite and shows incipient replacement by biotite along fractures. Biotite is reddish- or orange-brown and typically shows exsolution of ?Fe–Ti oxide needles and granular epidote along grain boundaries.

The prominent mafic inclusions in the Dinner Creek Tonalite are of fine-grained, weakly porphyritic and even-textured tonalite. The inclusions are very dark grey, typically rounded to subangular, and 2–5 cm in diameter. The inclusions are commonly identical to the host rock, except for their finer grain size and metamorphic textures. They probably represent fragments of early crystallized cognate material subsequently entrained into the magma.

Gordons Gorge Granite (ꞒgPgg)

This intrusion is named after Gordons Gorge on CHAMBERLAIN, where the bulk of the intrusion outcrops. On Bow, the Gordons Gorge Granite outcrops over no more than 15–20 km^2 along the western edge of the sheet north of Mount Lookout. The intrusion consists of massive, coarse-grained porphyritic biotite(–hornblende) monzogranite and syenogranite with a rapakivi-like texture.

Mount Nyulasy Granite (ꞒgPmn)

The Mount Nyulasy Granite is a large intrusion (>250 km^2) that extends northeast from the south-western corner of the map sheet across to Mount Nyulasy and northwards to the northern edge of the sheet. The intrusion has a SHRIMP U–Pb zircon age of 1859 ± 3 Ma. In the southwest of the sheet, the granite outcrops as rugged bouldery hills with about 200 m relief and prominent whalebacks. However, in the centre and north of Bow, the intrusion 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 may be moderately to strongly foliated. In the south-western 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 unit intruded the Dinner Creek Tonalite at several localities (see above), and it also intruded the Crooked Creek Granite 6 km southwest of Bow Hill (see above).

Most samples from the intrusion contain 20–30% or more phenocrysts of rounded microperthite 1–3 cm (rarely up to 4 cm) in diameter, some of which have narrow rims of fine-grained sodic plagioclase in a chequerboard 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. Phenocryst and groundmass microperthite contains numerous inclusions of plagioclase, biotite, and quartz. Where fresh, plagioclase displays strong oscillatory zoning. Accessories include opaque minerals, apatite, and zircon.

In most samples plagioclase is moderately to strongly altered to sericite and clinozoisite, biotite is partly replaced by chlorite, and ?magnetite is replaced by epidote. Micrographic and granophyric textures between quartz and microperthite crystals are widespread. 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. Along the margins of the intrusion, inclusions of country rock are abundant.

The Mount Nyulasy Granite contains numerous porphyry inclusions between the Great Northern Highway and the western edge of the Ragged Range. In the quarry 10 km north of Pompeys Pillar (AMG 164729), the granite contains two types of rounded porphyry inclusions typically 10 cm or more in diameter. Type (i) inclusions are of feldspar(–quartz) porphyry with less than 10% phenocrysts of microperthite and minor quartz and biotite in a black microcrystalline groundmass. Type (ii) inclusions consist of feldspar–quartz(–biotite) porphyry with about 30% phenocrysts of rounded K-feldspar, oblate quartz, subhedral plagioclase and minor biotite. The phenocrysts are of a size similar to those in the host granite. Type (i) inclusions are enclosed within type (ii) inclusions in addition to the granite. In detail, contacts between host granite and type (ii) inclusions may be highly irregular. The type (ii) inclusions bear a striking resemblance to the Greenvale Porphyry.

Medium-grained, even-textured biotite monzogranite (ᏢgPe)

Massive, medium-grained, even-textured biotite monzogranite forms a series of discontinuous exposures south and east of Bow Hill. The monzogranite outcrops in a roughly northeasterly trending zone about 11 km long and 1–2 km wide. The rock contains very fine grained clots of biotite up to 5 mm in diameter. Rounded to oblate, fine-grained mafic inclusions 3–10 cm in diameter are widespread. Between the Wilson River and the Great Northern Highway the granite is more leucocratic and coarser grained and contains numerous bundles of tourmaline up to 4 cm long disseminated through the rock. Fine-grained mafic inclusions are rare.

About 3 km southwest of Bow Hill (AMG 138514 and 141510) the contact between the even-textured monzogranite and the Mount Nyulasy Granite is exposed. At neither locality are veins or inclusions of one phase present in the other, so that the relationship between the two granitoids is not known.

Central zone

Tickalara Metamorphics

The Tickalara Metamorphics outcrops in the southern part of Bow, within the Central zone of the Lamboo Complex. This unit consists of high-grade migmatitic pelitic gneiss with rafts of psammitic gneiss, calc-silicate and mafic granulite (*ᏢmTpn*), marble and calc-silicate (*ᏢmTpc*), mafic granulite (*ᏢmTon*), and amphibolite (*ᏢmToa*), together with mafic and felsic sheet-like meta-igneous bodies (*ᏢmTgk, ᏢmTgd*).

The rocks forming the Tickalara Metamorphics on Bow show compositional layering, although they do not preserve features that can be interpreted as primary. To the south and southeast, on McIntosh and Dixon, 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 Bow, most of the Tickalara Metamorphics consists of pelitic metasedimentary rocks and belongs 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, carbonates and calc-silicates (Tyler et al., in prep.b) outcrops along the southeastern margin of the Mabel Downs Tonalite on Turkey Creek and extends onto the southern part of Bow.

Both the metasedimentary and the 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.c).

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 has intruded into it on McIntosh.

Previous workers have assumed the Tickalara Metamorphics to be the metamorphosed medium- to high-grade equivalents 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 are apparently within the Tickalara Metamorphics on DIXON and TURKEY CREEK were part of the 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; Tyler et al., in prep.b).

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 is described under **Hooper Orogeny** and **The 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 press.), 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). 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, which pre-dates D_1 (Tyler and Page, 1996). A minimum age for D_1 is provided by the c. 1850 Ma age for the granitoids of the Dougalls suite, which post-dated D_1 but pre-dated D_2 (Tyler and Page, 1996). D_1 in the Central zone has similar time constraints on it as 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 Bow 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 mudstone and siltstone, 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 usually 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. Large-scale younging reversals across strike are apparent.

Hancock (1991) recognized similar upright medium-scale folds in Marboo Formation rocks on LANSDOWNE (1:250 000). However, downward-facing structures were recognized locally, and Hancock concluded that the upright folds were a second generation (D_2) refolding an earlier recumbent fold phase (D_1). D_1 fold closures were not identified. Downward-facing strata have not been recognized in the Marboo Formation on Bow, 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 metamorphism in the Central zone (D_1/M_1)

The oldest deformation that is recognized in the Tickalara Metamorphics on Bow (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 metamorphism in the Central zone (D_2/M_2)

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 TURKEY CREEK (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) obtained a SHRIMP U–Pb zircon age of 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. 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 commonly oriented subparallel to S_1.

All rocks of the Tickalara Metamorphics on Bow 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 in the 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 remelting (Oliver et al., 1998), and 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 some 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 the present mapping of the Tickalara Metamorphics on Bow, nor had they been observed in the highest grade Tickalara Metamorphics by previous workers (Gemuts, 1971; Thornett, 1987). As will be discussed below, 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. These are characterized by low-pressure andalusite–cordierite–K-feldspar assemblages in pelitic rocks, and are superimposed onto the older, higher pressure garnet–cordierite–sillimanite–K-feldspar assemblages characteristic of the regional M_2 event.

Metasedimentary and mafic volcanic rocks (ĿmTpn, ĿmTpc, ĿmTa)

The Tickalara Metamorphics at the southeastern margin of the Mabel Downs Tonalite belong 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 rock (*ĿmTpc*) on Bow is present only in a fault slice within the Halls Creek Fault.

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 rock, and mafic granulite (*ĿmTpn;* Gemuts, 1971; Thornett, 1983, 1987). Mineral assemblages include cordierite–garnet–biotite–K-feldspar–plagioclase–quartz in rocks exposed 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 forms large skeletal porphyroblasts up to 2 cm in diameter, intergrown with quartz and cordierite. Cordierite may also be present as porphyroblasts, or as groundmass crystals. K-feldspar is typically perthitic, whereas plagioclase is antiperthitic.

Grade within the migmatitic rocks appears to decrease northwest of the Great Northern Highway, where garnet is typically 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 rock, 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 800°C, increasing to 550 MPa towards the Highway Shear Zone on TURKEY CREEK.

Amphibolite and mafic granulite (ꞒmTOa, ꞒmTOn)

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 occur parallel to layering, and pre-date at least D_2 and possibly D_1. Mafic granulite (*ꞒmTon*), 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 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 that has been deformed and metamorphosed under granulite facies conditions, and are similar to the rocks forming the Norton intrusion on TURKEY CREEK (Hoatson, 1995; Tyler et al., in prep.c).

Smaller pods, lenses, and layers of mafic granulite occur throughout the Tickalara Metamorphics on Bow.

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 (*ꞒmToa*) also occur 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 of 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 occur to the 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 (ꞒmTgk,ꞒmTgd)

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 Bow. 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 been pre- to syn-D_2/M_2.

The Corkwood Tonalite (*ꞒmTgk*) forms a 100 to 150 m-wide, steeply northwesterly dipping concordant sheet intruded into amphibolite at the eastern margin of the Mabel Downs Tonalite. It is medium grained, has been foliated and recrystallized, and consists of biotite (20%), andesine (40%), and quartz (35%), with minor amounts of epidote, calcite, apatite, zircon, and iron oxide.

The Dougalls Tonalite (*ꞒmTgd*) 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 Bow. Here the intrusion is complex with a number of sheets present, which are separated locally by patches and screens of migmatitic gneiss. Two mineral assemblages occur, consisting of 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, while 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 original 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 plutons in the Lamboo Complex, the bulk of which outcrop 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). On Bow, the Sally Downs supersuite consists of the Mable Downs Tonalite, Maggoty Springs Monzogranite (new name), Violet Valley Tonalite, and Wesley Yard Monzogranite (new name).

Biotite gabbro (*Ꝑgsob*)

Biotite gabbro is restricted to several small (<0.10 km^2 in total) exposures east of the Great Northern Highway in the southern part of the sheet. 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 Bow, biotite gabbro is composed of subophitic olivine norite and metagabbro. Olivine is enclosed by a narrow rim of hypersthene, which is in turn rimmed by green or brown hornblende. Hornblende replacement of hypersthene crystals ranges from incipient to complete. Oikocrysts of pyroxene (or hornblende) may reach 1 cm in diameter. Biotite, which replaces hypersthene, constitutes up to 5% of some samples. Biotite probably formed via a reaction between hypersthene crystals and melt (Naney, 1983), and hypersthene rims on olivine are a reaction product of olivine with increasingly siliceous melt. Green and brown hornblende is a metamorphic overprint of the igneous assemblages.

Mingled gabbro–granite (*Ꝑgsog*)

On Bow, mingled gabbro–granite outcrops along the Great Northern Highway at the southern edge of the sheet. 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 is described by Sheppard et al. (1997b).

Mabel Downs Tonalite (*Ꝑgsmd*)

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 Turkey Creek. The tonalite was emplaced into a shear zone and thus most of the intrusion is strongly foliated. A U–Pb SHRIMP age of 1832 ± 3 Ma was obtained for a sample 12 km south-southwest of Warmun.

Only the northern end of the Mabel Downs Tonalite outcrops on Bow, between Fargoo Creek and Turkey Creek at the southern end of the map sheet. The unit consists of moderately to strongly foliated, medium- to fine-grained hornblende–biotite tonalite 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. The contact is complicated by shears, and slices of the tonalite up to 20 or 30 m wide are tectonically interleaved with the metamorphic rocks. Beside Turkey Creek (AMG 269297), the Mabel Downs Tonalite is cut by a 20–50 cm-wide dyke of medium-grained, weakly porphyritic biotite granodiorite, possibly belonging to the Radigans Yard Granodiorite. The tonalite is also cut by numerous pegmatite veins and dykes, some of which are zoned from a biotite microgranite core to pegmatite at the margins.

Samples of the Mabel Downs Tonalite are composed of about 50–60% andesine (~An_{45}), 20–10% quartz, about 10% biotite, 10–15% amphibole, and minor secondary epidote. Accessory minerals consist of opaques, apatite, secondary titanite, and zircon. Plagioclase displays weak normal and oscillatory zoning, and contains numerous tiny inclusions of epidote and acicular microlites of an unidentified mineral. Epidote and titanite are replacement products of magnetite and ilmenite respectively. Igneous hornblende is replaced by pseudomorphs of a blue-green ?hastingsitic amphibole and minor epidote, or biotite and epidote. Similar mineralogical changes were noted by Wyborn and Page (1983) in granitoids metamorphosed to amphibolite facies in the Mount Isa Inlier.

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 exceeding 1 m in length 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. The latter type of inclusions has a similar texture, and identical mineral assemblage, to the host tonalite. The inclusions contain plagioclase phenocrysts similar to crystals from the host rock, have more amphibole, biotite and opaque minerals, and less quartz (typically <5%). Their angular shape and mineralogy suggest that they represent early crystallized cognate material, possibly from a magma chamber at depth, subsequently entrained into the main magma.

Magotty Springs Monzogranite (*Ꝑgsms*)

The Magotty Springs Monzogranite outcrops as a narrow sheet-like body, less than 1 km wide, west of the Great Northern Highway near the southern margin of the map sheet. It is correlated with more extensive sheet-like bodies of similar composition and field relationships to the south on Turkey Creek (Tyler et al., in prep.c). The Magotty Springs Monzogranite intruded the Tickalara Metamorphics. Contacts between the Magotty Springs Monzogranite and biotite gabbro (*Ꝑgsob*) and mingled gabbro–granite (*Ꝑgsog*) of the Sally Downs supersuite on Bow and Turkey Creek are commonly gradational over a distance of several metres, and are marked by extensive mingling of the two phases, with veins of each penetrating the other. Similar relationships are displayed by granitoid and gabbro plutons elsewhere in the Sally Downs supersuite, and in the Paperbark supersuite (Blake and Hoatson, 1993; Sheppard, 1996; Sheppard et al., 1997b), and are interpreted to indicate that the granitic and gabbroic magmas were broadly coeval.

The Magotty Springs Monzogranite consists of coarse-grained, porphyritic biotite monzogranite, and medium-grained, even-textured hornblende–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, which replace K-feldspar. Muscovite intergrown with epidote and calcite may replace feldspar in the granodiorite. Apatite, zircon and an iron oxide phase occur in both rock types.

Violet Valley Tonalite (*Ꞓgsvv, Ꞓgsvm*)

The Violet Valley Tonalite (Dow and Gemuts, 1969) is a large, northeasterly 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 outcrop on Bow, with the remainder on TURKEY CREEK. 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 (*Ꞓgsvv*). Much of the northern half of the intrusion is composed of foliated, medium-grained, weakly porphyritic biotite granodiorite and biotite–hornblende tonalite (*Ꞓgsvm*). Gemuts (1971) previously assigned much of this unit to the Mabel Downs Tonalite (his 'orthopyroxene-rich granodiorite', p. 54). In the field the two units can be difficult to distinguish, but the medium-grained granodiorite (*Ꞓgsvm*) is slightly more felsic and is typically foliated. On aerial photographs the fine- to medium-grained granodiorite (*Ꞓgsvv*) is distinguished from other granites by its dark grey to black colour.

The Violet Valley Tonalite intruded the Tickalara Metamorphics, although locally, shearing complicates the contacts. Angular fragments of migmatite are common close to the margins of the various intrusions. On TURKEY CREEK, undivided, medium- to coarse-grained porphyritic granite of the Sally Downs supersuite may have intruded, or mingled with, the Violet Valley Tonalite (Tyler et al., in prep.c). 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 (*Ꞓgsvm*) 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 (Ꞓgsvv)

Plagioclase phenocrysts up to 4 mm long constitute 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 prominent 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 (Ꞓgsvm)

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 up to Bow River. The intrusions range from small plugs each no more than 0.1 km^2 in area up to a stock that outcrops over 25 km^2 southwest of Radigans Yard. The granodiorite and tonalite in the Radigans Yard Granodiorite probably compose geographically distinct intrusive phases. Around Bow River Homestead and down to the southern edge of Bow, 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.

Rounded to lenticular, fine-grained mafic inclusions about 5–20 cm long are widespread, but only locally abundant. The inclusions may either be aphyric with 4–5 mm-wide biotite-rich rims, or they may contain about 20% phenocrysts of rounded quartz and subhedral plagioclase in a biotite-rich groundmass.

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. In both rock types, plagioclase has mottled and strongly oscillatory zoned cores with narrow, normally zoned rims. The hornblende (α=greenish yellow, β=blue-green, γ=olive) is probably a hastingsitic amphibole. Hornblende is partly replaced by intergrowths of biotite and epidote. Epidote and titanite psuedomorphs, of magnetite and ilmenite respectively, are associated with alteration of plagioclase. Allanite crystals are commonly rimmed by epidote.

Wesley Yard Monzogranite (*Ꞓgswy*)

The Wesley Yard Monzogranite is composed of medium- to coarse-grained, variably porphyritic biotite monzogranite and minor granodiorite. The unit is distinguished from the Mount Nyulasy Granite by its finer grain size, less porphyritic nature, lack of a rapakivi-like texture, and the presence of widespread mafic inclusions. Most of the intrusion is weakly foliated, but along the southeastern margin it is locally strongly foliated. The monzonite outcrops in the centre of the map sheet as a northeasterly trending intrusion about 10 km long by 5 km wide. The unit forms isolated, low rounded hills, and scattered boulders and tors amongst colluvium and alluvium.

The Wesley Yard Monzogranite is one of several intrusions of the Sally Downs supersuite on Bow that intruded across the boundary between the Central and Western zones. About 4 km south-southeast of Wesley Yard (AMG 311381), the Wesley Yard Monzogranite contains veined and enclosed mafic granulite of the Tickalara Metamorphics. About 1 km west of Wesley Yard (AMG 290421), the monzonite intruded dark-grey, medium-grained tonalite of the Dinner Creek Tonalite. Along its southeastern margin, contacts between the Wesley Yard Monzogranite and Radigans Yard Granodiorite are marked by shearing (AMG 258369).

Microperthite phenocrysts up to 2 cm long form about 15–20% of the rock, in an anhedral granular groundmass of medium-grained plagioclase (An_{35-40}), quartz, biotite and microperthite, and accessory magnetite (now epidote), ilmenite (now titanite), apatite, and zircon. Plagioclase crystals show mottling and weak oscillatory zoning. Much of the quartz is partially recrystallized to fine-grained granoblastic aggregates. Dark-brown biotite contains numerous inclusions of accessory minerals, and its grain boundaries are lined with very fine grained, granular epidote. Trace amounts of a blue-green ?hastingsitic amphibole may be present. Some epidote may be associated with replacement of amphibole by biotite.

In addition to angular inclusions of country rock near its margins, the Wesley Yard Monzogranite contains widespread mafic inclusions. Most of the inclusions are oblate, 5–20 cm long, and elongate in the foliation. Many contain a 2–3 mm-wide biotite-rich rim. The inclusions vary from aphyric to strongly porphyritic; the latter contain up to 20% phenocrysts of quartz and feldspar similar in appearance to that in the host rock. Most of the inclusions are quartz monzodiorite in composition. The inclusions have the same mineralogy as the host granite, but are more mafic (quartz constitutes 10–20% of the inclusions). Quartz crystals in some inclusions contain abundant acicular and prismatic apatite inclusions.

Medium- to coarse-grained porphyritic granitoid (*Pgsp*)

Medium- to coarse-grained porphyritic monzogranite and syenogranite outcrop over about 15 km^2 in the far southwestern corner of the map sheet. This unit outcrops much more extensively along the eastern edge of MOUNT REMARKABLE and the western edge of TURKEY CREEK. Owing to the inaccessible nature of the unit, it has not been formally defined. It is commonly characterized by a more subdued response on radiometric images relative to the Kevins Dam Monzogranite, which outcrops to the west on LISSADELL (1:250 000). On TURKEY CREEK the porphyritic granite 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 (*Pmh*) and the Tickalara Metamorphics (*PmTh*) adjacent to the granitoids and mafic intrusions of the Sally Downs supersuite (Thornett, 1987; Sheppard et al., 1997b; Tyler et al., in prep.a,c). In Marboo Formation rocks on Bow, 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 1 cm in diameter, which have a sieved texture with inclusions of fine biotite, quartz, and opaques. The lower grade hornfelses are well exposed along the highway to the south, and are spotted also with pinitized poikiloblasts of cordierite that are up to 3 mm across. Garnet is not present in the hornfelses and the occurrence of melting in 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 with the development of hornfelsic textures. Where melting has occurred, the M_2 textures are completely destroyed. Andalusite and sillimanite may be present, but garnet is absent.

Palaeoproterozoic Salt Lick Creek layered intrusion (*Pa*)

The Salt Lick Creek layered intrusion is shown on the accompanying map as ‘undivided layered mafic–ultramafic intrusions’. It 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 gabbro-norite, 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 composing 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 have affected rocks of the Tickalara Metamorphics in the Central zone, and the synorogenic granitoids and intrusive mafic rocks of the Sally Downs supersuite (see below). The third deformation (D_3) affected the Tickalara Metamorphics and the Mabel Downs Tonalite. It did not affect the Sally Downs Tonalite on McIntosh (Tyler et al., in prep.a) or the biotite gabbro intrusions, including those on Bow, 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, Tyler and Page, 1996). The youngest deformation (D_4) post-dates the Sally Downs Tonalite on McIntosh. An upper age limit for D_4 is not well defined, although from relationships on Turkey Creek and Dixon to the south (Tyler et al., in prep.b,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.

The third deformation and metamorphism in the Central zone (D_3/M_3)

The D_3 deformation has produced small- to large-scale, close to isoclinal folds. On Bow 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 cannot be followed for any distance and the regional-scale structure on Bow is not clear. On Mount Remarkable and Turkey Creek, large-scale D_3 folds are picked out by mafic granulite layers (Sheppard et al., 1997b; Tyler et al., in prep.c). 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, forming synchronously with epidote–amphibolite facies metamorphism (M_3).

The fourth deformation and metamorphism in the Central zone (D_4/M_4)

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, showing blastomylonitic textures. The minerals biotite, garnet, epidote, plagioclase, K-feldspar, and quartz are present. Shear sense indicators, including tails on feldspar porphyroclasts and S–C fabrics, indicate transpressive sinistral and northwest-side-up movement.

On Mount Remarkable and Turkey Creek, steeply northeasterly plunging, open-to-tight folds have refolded the D_3 folds to produce Type 3 hooked fold interference patterns (Ramsay and Huber, 1986; Sheppard et al., 1997b; Tyler et al., in prep.c). Locally, folding is developed in a similar orientation on Bow refolding D_3 folds. A crenulation cleavage (S_4) that is parallel to the axial surfaces of these folds is locally present.

Palaeoproterozoic Speewah Basin

Previously, the Speewah Group and the overlying Kimberley Group have been regarded as being deposited within a single 'Kimberley Basin' (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 dissimilar palaeogeographies, representing two different depositional basins in disparate tectonic settings. The Speewah Group was deposited in the Speewah Basin, and was restricted to the west of the Greenvale Fault. 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 the Lissadell 1:250 000 sheet (Page and Sun, 1994), indicating that deposition took place synchronously with 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 (1:250 000) 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; however, 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, the boundary between the two groups is now defined as the base of an unconformity, recognized within the 'King Leopold Sandstone' (Griffin et al., 1993). That part of the old 'King Leopold Sandstone' beneath the unconformity is now named the Bedford Sandstone. 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 Basin outcrop in the extreme northwestern part of Bow. Here, only the lower to middle part of the Speewah Group is exposed, and consists in ascending order of the O'Donnell Formation, Tunganary Formation, Valentine Siltstone, and Lansdowne Arkose. This succession is intruded extensively by the Hart Dolerite. On adjacent parts of LISSADELL, the Speewah Group has an estimated total thickness of 1500 m (Thorne et al., in prep.).

O'Donnell Formation (*Psn*)

On Bow, the O'Donnell Formation unconformably overlies the Whitewater Volcanics with a slight angular discordance, whereas the contact with the overlying Tunganary Formation is conformable. Plumb (1968) estimates the thickness of the O'Donnell Formation on adjacent parts of LISSADELL to be about 260 m. The formation 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 and micaceous. Most are parallel-laminated with cross-lamination occuring in places. The interbedded sandstones are thin bedded, fine to medium grained and quartz rich. Some sandstone beds are normally graded. Internal structure typically consists of parallel-lamination or cross-lamination, occurring either separately or in combination.

Tunganary Formation (*Pst*)

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 consist 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.

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

Valentine Siltstone (*Psv*)

The Valentine Siltstone is a poorly exposed unit that is conformable with the underlying Tunganary Formation. It 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, parallel-laminated or thinly bedded, siltstone and sandstone, with thin interbeds of dacite to rhyolite tuff and tuffaceous siltstone. A black, crystal-poor rhyolite tuff locally marks the base of the formation elsewhere on LISSADELL. This volcanic rock was dated at 1834 ± 3 Ma by Page and Sun (1994).

Lansdowne Arkose (*Pso*)

The Lansdowne Arkose conformably overlies the Valentine Siltstone, and typically forms low, finely terraced escarpments giving rise to a ribbed appearance on aerial photographs. On adjacent parts of LISSADELL the Lansdowne Arkose is about 400 m thick, whereas only the lower part of the formation outcrops on BOW.

The Lansdowne Arkose is composed of pink to pale-purple, buff and pale-grey, medium- to coarse-grained feldspathic sandstone and arkose, with interbedded medium- to coarse-grained quartz sandstone, and micaceous siltstone and shale. The formation is dominated by small- to medium-scale, trough cross-stratification, occurring as single or vertically stacked sets.

Palaeoproterozoic Texas Downs Basin

The Texas Downs Basin (Tyler et al., in prep.c) corresponds to the present-day outcrop of the Texas Downs Formation. It is exposed in isolated fault slices immediately to the east of the Halls Creek Fault on the southeastern part of Bow, and includes units mapped either as Red Rock Beds or Mount Parker Sandstone on the first edition 1:250 000 LISSADELL sheet (Plumb, 1968). The lower contact of the Texas Downs Formation is not exposed on Bow; however, on neighbouring TURKEY

CREEK 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.c). 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 TURKEY CREEK (Tyler et al., in prep.c). The younger age limit is speculative, as 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 that 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 the LENNARD RIVER 1:250 000 sheet (Page, R. W., 1995, pers. comm.).

Texas Downs Formation (*Ṗx*)

The Texas Downs Formation is exposed in three fault-bounded slices (between AMG 280202 and AMG 455540) in the southeastern part of Bow. The formation has an estimated minimum thickness of 1 km and consists of sandstone, siltstone, and minor conglomerate. On TURKEY CREEK, basaltic units are interbedded with the sedimentary rocks; however, these volcanic rocks have not been recognized on Bow.

In the Pitt Range (AMG 420480), the Texas Downs Formation can be subdivided into lower and upper sandstone units which are separated by an argillaceous member (*Ṗxa*). 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 suggests 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 eastern part of LISSADELL, where it is exposed in three small areas between the Glenhill and Carr Boyd Faults. 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 an angular unconformity. The formation unconformably overlies rocks of the Lamboo Complex including the Marboo Formation, the Whitewater Volcanics, the Castlereagh Hill Porphyry, and granitoids of the Paperbark supersuite. These relationships indicate that the maximum age for the Revolver Creek Formation is around 1855 Ma, this being the age of the Whitewater Volcanics and 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 all these units are unlikely to be direct correlatives, they may be broadly coeval with a minimum depositional age of c. 1800 Ma.

Revolver Creek Formation (*Ṗr*)

On Bow, the Revolver Creek Formation is exposed immediately west of the Argyle mine site (AMG 290510). In this area 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 (*Ṗrb*) 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 jaspilite. 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 composite 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 to be strongly altered in thin section (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 commonly cream or pink, fine- to coarse-grained feldspathic quartz sandstone 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 commonly poorly exposed, show a combination of parallel lamination and current or wave-ripple cross-lamination. Sandstones are fine to coarse grained and occur either 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 commonly 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.

Palaeoproterozoic Hart Dolerite (*Pdh*)

The Hart Dolerite consists of a series of massive dolerite sills and less extensive granophyre that intruded the Speewah Group and the lower part of the Kimberley Group. The sills have a combined thickness of up to 3000 m (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 the LANSDOWNE 1:250 000 sheet is provided by Gellatly et al. (1975), and Alvin (1993) has studied the Hart Dolerite in the Speewah Valley on the LISSADELL 1:250 000 sheet. A SHRIMP U–Pb zircon age of c. 1790 Ma was obtained for the Hart Dolerite (Page, R.W., 1995, pers. comm.).

On BOW, the Hart Dolerite is exposed in the northwestern part of the map sheet, where it intruded the base of the Valentine Siltstone and siltstone beds in the upper part of the Landsdowne Arkose. The granophyre, which is associated with the dolerite elsewhere on LISSADELL, was not observed on BOW.

On adjacent parts of LISSADELL, the dolerite unit is composed of fine- to medium-grained dolerite, quartz dolerite and gabbro, although Alvin (1993) also reported the presence of olivine dolerite. Adjacent to sedimentary rocks, the Hart Dolerite is typically chilled. Sharp internal contacts between dolerite types indicate that the thicker sills are composite bodies. Thorne et al. (in prep.) recorded the presence of a contact between fine-grained and medium-grained quartz dolerite in the Speewah Valley, 15 km northwest of Bow. They also noted that a pegmatitic dolerite intruded quartz dolerite in this area. An igneous flow lamination, defined by alignment of plagioclase crystals, or locally 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 of between An_{50} and An_{70}. Clinopyroxene crystals display very fine lamellar and herringbone exsolution of orthopyroxene. Some 3–10% of the rocks consists of Fe–Ti oxides, and interstitial quartz and granophyric intergrowths of quartz and K-feldspar constitute 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 oxides 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 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 in length), 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.

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 described first 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 about 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 the 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

Only the lower and middle parts of the Carr Boyd Group, up to and including the Glenhill Formation, are exposed on Bow. These rocks outcrop in the central northern part of the map sheet where they unconformably overlie the Marboo Formation, granitoids of the Paperbark supersuite, Castlereagh Hill Porphyry, and the Revolver Creek Formation.

Previous workers recorded the presence of 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 Bow has failed to confirm the presence of these major stratigraphic breaks, indicating, rather, that the discordant boundary relationships and anomalous thickness variations can be more readily attributed to post-Carr Boyd Group tectonism. Specific examples include the areas 10 km north of Mount Nyulasy (AMG 230583) and immediately west of the Argyle mine (AMG 340517), 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 the 1:250 000 LISSADELL map sheet (Plumb et al., 1968). In the first case, the apparent unconformable relationship is the result of faulting associated with the Glenhill – Bow River Fault system. Here, the Carr Boyd Group, including the lower part of the Lissadell Formation, is cut by a set of northwesterly and north-easterly trending fractures which juxtapose these rocks against the older Lamboo Complex. In the second example, 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 about one-third of that 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 this formation and the Golden Gate Siltstone appears to be conformable and transitional.

Hensman Sandstone (*Ꝑch*)

The Hensman Sandstone (Dow et al., 1964) is the lowermost formation in the Carr Boyd Group and has a thickness of 120 m on Bow. The unit 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 on Bow are massive, and show little detail of the internal structure. On the adjacent DUNHAM RIVER sheet 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 (*Ꝑcg*)

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 160 m in the

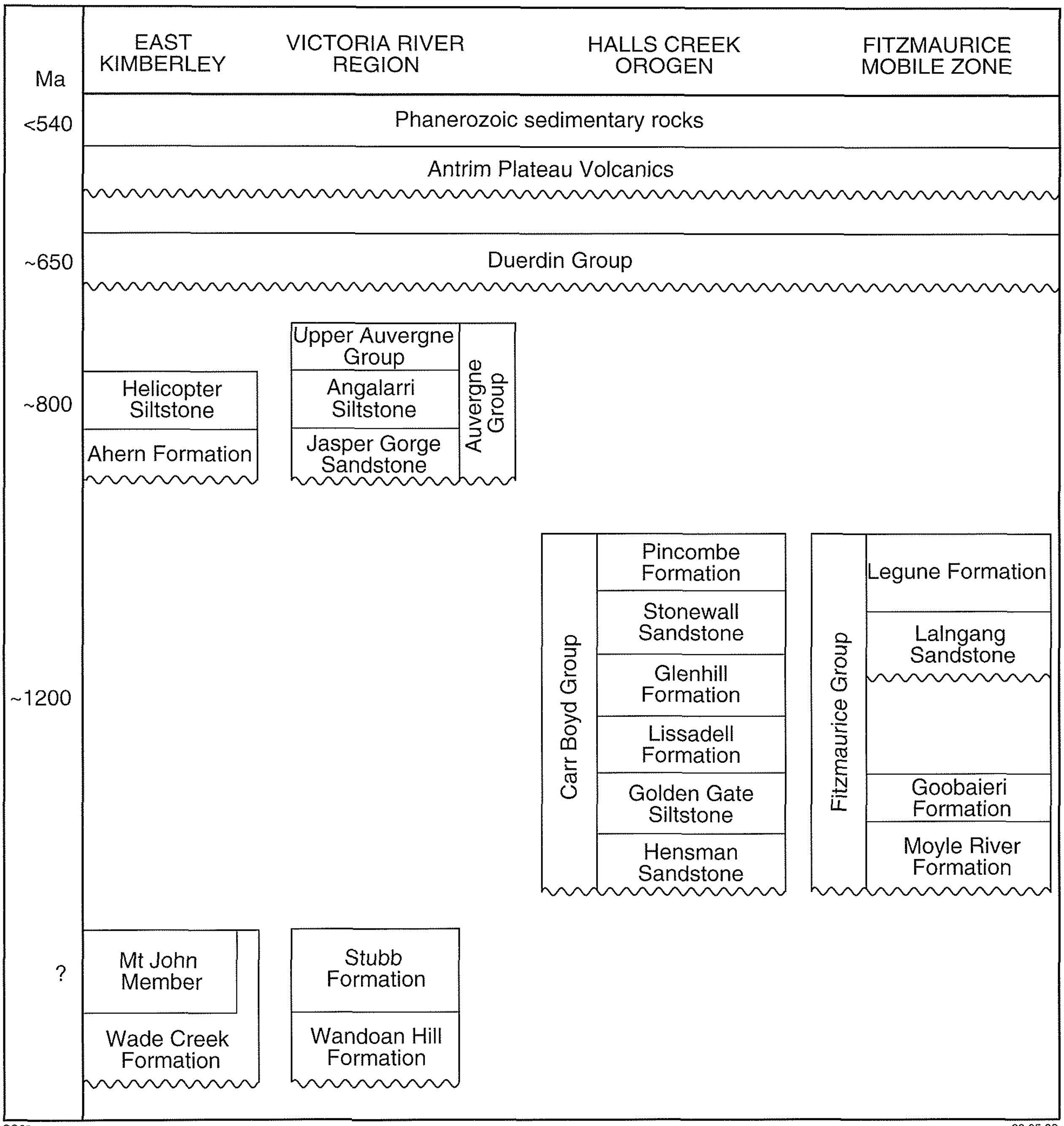

Figure 4. Regional correlation of Mesoproterozoic to Phanerozoic rocks in the east Kimberley region

vicinity of the Pompeys Pillar (Sam) iron prospect (AMG 221635). Most of the formation 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 (around AMG 220620 and AMG 287540 respectively). The remainder of the formation comprises interbedded green argillite and thin- to medium-bedded, fine- to medium-grained quartz sandstone. Sandstone beds are commonly cross-laminated and may have wave- or current-ripple bedforms on their upper surfaces.

Lissadell Formation (*ECl*)

The Lissadell Formation is about 1200 m thick on Bow and is conformably overlain by the Glenhill Formation. The unit consists of quartz arenite and lithic quartz sandstone, interbedded with green or purple micaceous siltstone. Two major sandstone (*ECl*) and two siltstone-dominated units (*ECla*) are present in the area south of

Glen Hills Homestead (AMG 311708). 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.

Glenhill Formation (*Ꞓcg*)

The Glenhill Formation is the highest formation in the Carr Boyd Group exposed on Bow. The unit conformably overlies the Lissadell Formation and is unconformably overlain by the Antrim Plateau Volcanics. The succession consists of an estimated 300 m of siltstone and sandstone and contains minor conglomerate locally. A 40–50 m-thick sandstone-dominated unit (*Ꞓcgs*) 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 comprises alternations of stacked medium-scale trough and tabular cross-strata, planar and undulatory parallel stratification, and wave- and current-ripple cross-lamination. Palaeocurrent data from the trough cross-stratified units are variable and indicate that local transport was towards the southwest, southeast or northeast.

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 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.

Argyle (AK1) lamproite pipe (*Ꞓi*A)

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 is distilled from the above publications.

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 Mesoproterozoic 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 dewatering 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 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 in age. 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 is interpreted to have 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 Bow, 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 prior to c. 1200 Ma. Post-Carr Boyd Group tectonism is indicated by the unconformity at the base of the c. 800 Ma Ahern Formation (see **Ahern Formation**).

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 up to 13 m, and in length up to 2 km. 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, 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) show that the dykes share many features of ultramafic lamprophyres (Rock, 1991) and 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., in prep.c). Dow et al. (1964) divided this part of the Kimberley succession into four stratigraphic units (in ascending order): 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, 1998, in prep.c) 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 was previously referred to as the Boll Sandstone, was renamed the Ahern Formation by Tyler et al. (1997) to avoid confusion with the Devonian Boll Conglomerate on DIXON RANGE (1:250 000).

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) although 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 on 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 (*Ef*)

On Bow, the Ahern Formation is confined to a small faulted outcrop on the eastern edge of the map sheet (around AMG 460640). Immediately to the east of this locality, the formation unconformably overlies Lamboo Complex rocks and is conformably overlain by the Helicopter Siltstone. Here, the Ahern Formation 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 (*Eh*)

The Helicopter Siltstone is exposed in a small area about 4 km south-southeast of Smoke Creek Yard (around AMG 460650) and is about 330 m thick on the adjoining part of the LISSADELL 1:250 000 sheet. This unit 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 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 Bow 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 as well as 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.

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 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 Bow. The tillite is exposed only in three small outcrops; near the Upper Smoke Creek diamond deposit (AMG 330540), on the eastern edge of the map sheet (AMG 465658) and at Lissadell Hill (AMG 410428), whereas the Ranford Formation outcrops in the southeast of the map sheet between Fargoo Creek (AMG 312260) and the Pitt Range (AMG 440500).

Moonlight Valley Tillite (*ꝑEm*)

The Moonlight Valley Tillite unconformably overlies the Helicopter Siltstone and Carr Boyd Group and has an estimated thickness of 280 m on the adjacent part of the LISSADELL 1:250 000 sheet. The unit 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, occur within the tillite about 4 km south-southeast of Smoke Creek Yard (east of AMG 465658).

Dow and Gemuts (1969) suggested that the tillite was a subaqueous deposit from a grounded ice sheet. Plumb (1981, 1993, pers. comm.) points out that the tillite interfingers with fluvioglacial outwash material to the north and suggests 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 (*ꝑEo*)

The Ranford Formation conformably overlies the Moonlight Valley Tillite and is itself unconformably overlain by either the Albert Edward Group or the Antrim Plateau Volcanics. It is about 600 m thick elsewhere on LISSADELL (1:250 000) and consists of grey to chocolate-coloured micaceous siltstone interbedded with medium- to very thick bedded, poorly sorted lithic quartz sandstone, dolomitic quartz sandstone, and minor dolostone. 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.

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 Bow.

Mount Forster Sandstone (*ꝑLo*)

The Mount Forster Sandstone outcrops in the south-eastern part of the map sheet, around Fargoo and Gate Creeks. Here, this unit 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 occur on the bases of some troughed cosets, whereas coset tops may show straight-crested symmetrical ripples. Palaeocurrent data indicate most sediment transport was towards the south-southwest.

Granitoid (*ge*), pegmatite (*p*), and dolerite (*d*) dykes

Most of the granitoids in both the Paperbark and Sally Downs supersuites 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 Violet Valley Tonalite 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 dykes 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 (AMG 240287), the pegmatite dykes commonly contain a few percent coarse crystals of magnetite.

Dolerite dykes (*d*) are widespread on Bow, but most are concentrated in two swarms, one northwest of Bow River Homestead and the other southwest of Pompeys Pillar. Dykes in these two swarms strike roughly north and northwest respectively. Most dykes are about 1–5 m wide with individual dykes reaching 4 km in length. They are composed of a few percent pseudomorphs 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 Basin (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 Rb–Sr data from Bofinger (1967) and reported 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 region (Tyler and Griffin, 1990; Tyler et al., 1991). Deformation occurred at about the same time as the Paterson Orogeny at the eastern edge of the Pilbara Craton, 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 Bow, but uplift and erosion at that time is indicated by the unconformity at the base of the early Cambrian Antrim Plateau Volcanics (see **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 Bow. 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 that 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).

Antrim Plateau Volcanics (*Ꞓa*)

The Antrim Plateau Volcanics (Traves, 1955; Mory and Beere, 1988) of Cambrian age is exposed around Glen Hills Homestead (AMG 311708) and also covers a large area in the southeastern part of the map sheet between Fargoo Creek (AMG 312260) and Cattle Creek (AMG 450298). Near Glen Hills Homestead, the formation is about 600 m thick and unconformably overlies the Carr Boyd and Duerdin Groups, whereas in the southeast it is about 1000 m thick and is unconformable on the Duerdin and Albert Edward Groups. The upper contact is not seen in the Glen Hills area; in the southeastern part of the map sheet it is conformably overlain by the Headleys Limestone.

The Antrim Plateau Volcanics consists predominantly of massive to amygdaloidal, aphyric to porphyritic basalt and basaltic breccia interbedded with minor sedimentary rock. 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 (*Ꞓas*) showing large-scale, low-angle cross-stratification is interbedded with the basalt flows south of Glen Hills Homestead.

Ord Basin

Goose Hole Group

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 the only formally named unit of the Goose Hole Group exposed on Bow. Elsewhere in the Ord Basin the Headleys Limestone is overlain by the Nelson Shale; however, in the vicinity of Lissadell Hill (AMG 413423) this formation is also succeeded by an unnamed unit consisting of lithic sandstone, pebbly sandstone, and minor conglomerate (*ЄG*).

Negri Subgroup

Headleys Limestone (ЄGh)

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 presumably 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 (ЄGn)

The Nelson Shale does not outcrop on Bow, but it is thought to underlie a thin cover of Cainozoic black soil immediately northeast of Lissadell Hill. Elsewhere in the Ord Basin it is 100 to 180 m thick and consists of purple siltstone with thin beds of sandstone and micrite (Mory and Beere, 1988).

Bonaparte Basin

Cockatoo Group

The Late Devonian (Frasnian) succession in the Bonaparte Basin is referred to as the Cockatoo Group and is subdivided into ten formations (Mory and Beere, 1988). Of these only one, the Galloping Creek Formation, is exposed on Bow.

Galloping Creek Formation (Dcg)

On Bow, the Galloping Creek Formation (Beere and Mory, 1986) unconformably overlies Proterozoic basement rocks in the northeastern corner of the map sheet. The type section for the formation is 10.5 km south of Glen Hills Homestead. Here, the formation is about 1.6 km thick and consists of thick successions dominated by conglomerate, pebbly sandstone, and sandstone. The conglomeratic intervals consist of very thick bedded conglomerate with minor sandstone beds and lenses. Conglomerates are commonly 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 successions, with minor planar and troughed 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.

The Galloping Creek Formation is interpreted as an alluvial fan deposited in response to strike-slip movement along the Glenhill – Revolver Creek – Carr Boyd fault system (Mory and Beere, 1988; Thorne and Tyler, 1996).

Palaeozoic
Alice Springs Orogeny

The youngest phase of major faulting and folding (D_7) in the east Kimberley region is attributed to the Upper 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, northeasterly to easterly trending open synclines and anticlines. These folds are cut by northeasterly and northwesterly trending splays from the Halls Creek Fault. West of the Halls Creek Fault, the Antrim Plateau Volcanics and Late Devonian Galloping Creek Formation are tilted and deformed into open, upright flexures with curvilinear axes locally (e.g. AMG 365670). 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 and Carr Boyd – Glenhill fault systems.

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 western part of Bow, 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 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 largely developed over the Antrim Plateau Volcanics and Nelson Shale.

Alluvium (*Qa*), consisting of unconsolidated silt, sand and gravel, lies along present-day drainage channels. Unconsolidated scree (*Qc*) is present adjacent to the Halls Creek Fault scarp.

Economic geology

The following descriptions of mineral deposits and occurrences on Bow are based on a summary compiled by Sanders (1999), covering the whole of the east Kimberley region, of both published information and that contained within GSWA's Western Australian Mineral Exploration (WAMEX) database open-file reports.

Copper–nickel

Disseminated, vein and minor massive copper–nickel sulfide mineralization is hosted by a small, metamorphosed layered intrusion at the Bow River prospect (AMG 288346) 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 in composition from leucogabbroic to troctolitic.

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 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% Cu and 0.15% Ni. 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.

Work by Australian Anglo American concluded 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.

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 Argyle mine was 441 Mct at a value of $5.280 billion. The diamonds are hosted by the Argyle (AK1) lamproite pipe, the geology of which is discussed above. The majority of the diamonds are hosted by 'sandy tuffs' (Lewis, 1990).

Production from the Argyle pipe began in 1985, in which year 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 of Pliocene age up to 3.5 m thick. 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).

Gold

The Jailhouse Creek gold prospect (AMG 124731) consists of thin and widely spaced, weakly mineralized quartz veins in rocks of the Whitewater Volcanics near the northern edge of Bow. Several gold anomalies were detected in this area. The location of the prospect on the map is that of the stream-sediment anomaly in the centre of the mineralized area. Weakly anomalous gold is associated with northerly, northeasterly and easterly 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. While individual beds contain up to 60% Fe, numerous interbeds of sandstone and shale reduce the grade of ore that could be mined to about 50% Fe (Dow and Gemuts, 1969). The Pompeys Pillar and Matsu deposits together contain an estimated resource of 15 Mt at about 64% Fe (Preston, 1997).

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BOW

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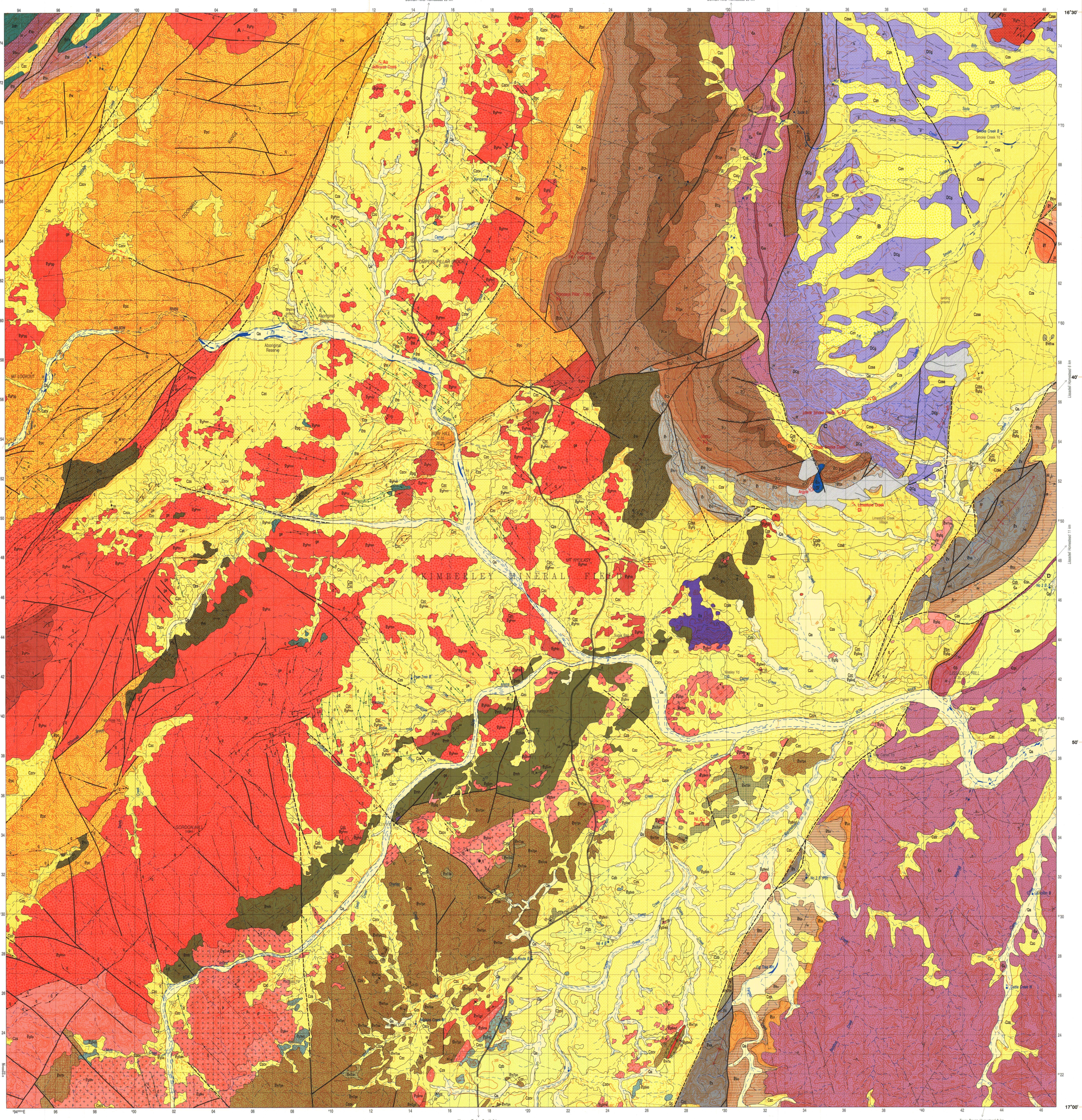

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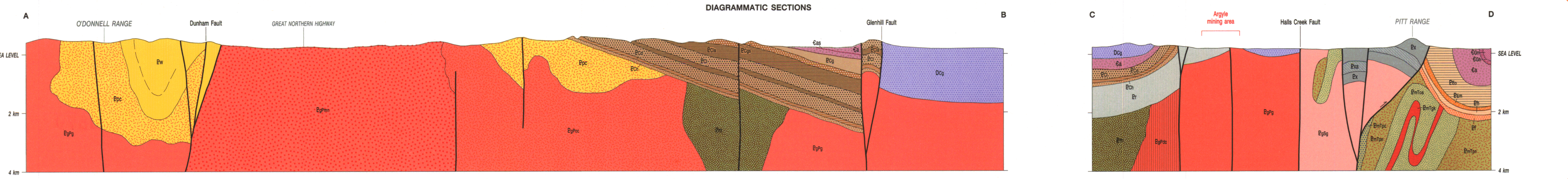


Legend

PHANEROZOIC — CAINOZOIC — QUATERNARY

Qa Alluvium—unconsolidated clay, silt, sand, and gravel in channels and floodplains
Qc Colluvium—unconsolidated rock fragments in soil; scree deposits

Czb Black soil—clay and silt
Czc Colluvium and rubble—sand, gravel, conglomerate, and sedimentary breccia forming sheetwash fans and scree deposits
Czcv Colluvium and alluvium—clay, silt, sand, and gravel as dissected, partly consolidated valley-fill deposits
Czn Unconsolidated and partly consolidated sandplain deposits
Czs Colluvium and alluvium—partly consolidated clay, silt, sand, and gravel adjacent to drainage
Czsa Alluvium and colluvium—dissected sheets of partly consolidated silt and sand; pebble, cobble, and boulder conglomerate

q Quartz veins, of various ages; youngest generation postdates Devonian

PALAEOZOIC — LATE DEVONIAN — FRASNIAN — Cockatoo Group (BONAPARTE BASIN)

Đcg **GALLOPING CREEK FORMATION:** fine- to very coarse-grained feldspathic sandstone, and pebble to boulder conglomerate

MIDDLE CAMBRIAN — ORDIAN — Goose Hole Group — Negri Subgroup (ORD BASIN)

ЄG Undivided, fine- to coarse-grained sandstone, pebbly sandstone, and pebble conglomerate
ЄGn **NELSON SHALE:** purple siltstone and mudstone; minor sandstone (section only)
ЄGh **HEADLEYS LIMESTONE:** grey, massive to laminated limestone; minor chert and stromatolites locally

EARLY CAMBRIAN (ORD AND BONAPARTE BASINS)

Єa **ANTRIM PLATEAU VOLCANICS:** massive or amygdaloidal, locally porphyritic basalt, and basaltic breccia
Єas Medium-grained quartz sandstone

d Dolerite dyke
ge Fine- to medium-grained, even-textured biotite granitoid dyke
p Pegmatite dykes and sheets

PROTEROZOIC

Albert Edward Group (WOLFE CREEK BASIN)

ꝐLo **MOUNT FORSTER SANDSTONE:** fine- to medium-grained quartz sandstone; minor pebbly sandstone and conglomerate

Duerdin Group — c. 610 Ma

ꝐEo **RANFORD FORMATION:** thin-bedded siltstone and thin- to thick-bedded quartz sandstone, dolomitic quartz sandstone, and dolomite
ꝐEm **MOONLIGHT VALLEY TILLITE:** massive, matrix-supported, polymictic pebble to boulder conglomerate, and poorly sorted, massive sandstone; larger clasts in conglomerate are polished and striated locally

VICTORIA RIVER BASIN — c. ?800 Ma

Ꝑh **HELICOPTER SILTSTONE:** green and grey, micaceous siltstone, fine- to medium-grained quartz sandstone, and mudstone
Ꝑf **AHERN FORMATION:** fine- to coarse-grained quartz sandstone, pebbly sandstone, conglomerate, and siltstone

c. 815 Ma — ꝐlB **Bow Hill lamprophyre dykes:** olivine-phlogopite lamprophyre; garnet-phlogopite lamprophyre

c. 1200 Ma — ꝐlA **Argyle lamproite pipe:** volcaniclastic olivine lamproite intruded by olivine-phlogopite lamproite dykes

Carr Boyd Group (CARR BOYD BASIN)

ꝐCg **GLENHILL FORMATION:** micaceous siltstone, fine- to coarse-grained quartz sandstone, and mudstone
ꝐCgs Fine- to coarse-grained quartz sandstone
ꝐCl **LISSADELL FORMATION:** fine- to coarse-grained quartz sandstone; subordinate micaceous siltstone and mudstone
ꝐCla Siltstone and mudstone; thin- to thick-bedded quartz sandstone
ꝐCd **GOLDEN GATE SILTSTONE:** siltstone and mudstone; either carbonaceous, chloritic, or micaceous; thin- to medium-bedded ferruginous sandstone, and quartz sandstone
ꝐCh **HENSMAN SANDSTONE:** fine- to coarse-grained quartz sandstone; heavily silicified

c. 1790 Ma — Ꝑdh **HART DOLERITE:** dark grey dolerite

REVOLVER CREEK BASIN

Ꝑr **REVOLVER CREEK FORMATION:** fine- to coarse-grained quartz sandstone, lithic quartz sandstone, feldspathic quartz sandstone, and carbonaceous siltstone and mudstone; minor conglomerate
Ꝑrb Dolerite, basalt, and basaltic breccia; minor lithic sandstone, siltstone, and mudstone

TEXAS DOWNS BASIN

Ꝑx **TEXAS DOWNS FORMATION:** fine- to coarse-grained lithic quartz sandstone, pebbly sandstone, conglomerate, siltstone, mudstone, basalt, and basaltic breccia
Ꝑxa Mudstone and siltstone; thin- to thick-bedded quartz sandstone

Speewah Group — c. 1835 Ma (SPEEWAH BASIN)

ꝐSo **LANSDOWNE ARKOSE:** thin- to thick-bedded feldspathic quartz sandstone; minor micaceous siltstone and mudstone
ꝐSv **VALENTINE SILTSTONE:** siltstone and mudstone; commonly micaceous or chloritic; minor thin-bedded feldspathic quartz sandstone and rhyolitic volcaniclastic sandstone
ꝐSt **TUNGANARY FORMATION:** quartz sandstone, feldspathic quartz sandstone, and subordinate siltstone and mudstone
ꝐSn **O'DONNELL FORMATION:** fine- to coarse-grained quartz sandstone and lithic quartz sandstone, pebbly sandstone, chloritic siltstone and mudstone, and granule to pebble conglomerate

Ꝑa Undivided layered mafic-ultramafic intrusions; includes olivine norite, gabbronorite, and minor anorthosite

Sally Downs supersuite — c. 1835–1805 Ma (Bow River batholith)

ꝐgSg Undivided granitoid rock
ꝐgSmd **MABEL DOWNS TONALITE:** medium- to fine-grained, foliated, hornblende-biotite tonalite; extensively recrystallized; abundant flattened inclusions of mafic rock
ꝐgSms **MAGOTTY SPRINGS MONZOGRANITE:** medium- to coarse-grained, porphyritic biotite monzogranite; medium-grained, even-textured hornblende-biotite granodiorite
ꝐgSvv **VIOLET VALLEY TONALITE:** fine- to medium-grained, weakly porphyritic biotite granodiorite and biotite-hornblende tonalite; weakly to moderately foliated
ꝐgSvm Medium-grained, weakly porphyritic biotite granodiorite and biotite-hornblende tonalite; weakly to moderately foliated
ꝐgSwy **WESLEY YARD MONZOGRANITE:** massive, medium- to coarse-grained, porphyritic biotite monzogranite; minor granodiorite
ꝐgSp Medium- to coarse-grained, porphyritic monzogranite and syenogranite
ꝐgSob Biotite-bearing norite, gabbronorite, and minor gabbro
ꝐgSog Biotite-bearing norite, gabbronorite, and minor gabbro with veins of intermediate hybrid rock; mafic and intermediate hybrid rock cut by abundant veins of coarse-grained porphyritic tonalite and granodiorite

Paperbark supersuite — c. 1865–1850 Ma

ꝐgPg Undivided granitoid rock
ꝐgPby **BEEFWOOD YARD GRANITE:** medium- and coarse-grained, porphyritic biotite monzogranite
ꝐgPcc **CROOKED CREEK GRANITE:** medium-grained, plagioclase-phyric, biotite granodiorite and monzogranite
ꝐgPdc **DINNER CREEK TONALITE:** fine- to medium-grained, massive or weakly foliated hypersthene-biotite-amphibole tonalite; abundant inclusions of fine-grained biotite-amphibole tonalite; commonly extensively recrystallized
ꝐgPgg **GORDONS GORGE GRANITE:** massive, coarse-grained, porphyritic biotite(-hornblende) monzogranite and syenogranite with a rapakivi-like texture
ꝐgPmn **MOUNT NYULASY GRANITE:** massive, coarse-grained, porphyritic biotite monzogranite and syenogranite with a rapakivi-like texture; subordinate medium- to coarse-grained, even-textured monzogranite
ꝐgPe Medium-grained, even-textured biotite monzogranite
ꝐgPob Biotite gabbro

c. 1865–1850 Ma

Ꝑpg **GREENVALE PORPHYRY:** massive, dark grey quartz-feldspar micromonzogranite porphyry; medium to coarse phenocrysts of quartz, K-feldspar, and plagioclase
Ꝑpc **CASTLEREAGH HILL PORPHYRY:** dark grey, massive, fine-grained feldspar microgranodiorite porphyry; medium phenocrysts of plagioclase; commonly strongly recrystallized

c. 1865–1850 Ma — Ꝑw **WHITEWATER VOLCANICS:** thick-bedded, massive, fine- to medium-grained quartz-feldspar rhyolite to dacite porphyry and welded ignimbrite; minor agglomerate, lapilli tuff, and volcaniclastic sedimentary rocks

c. 1870 Ma (Western zone, Lamboo Complex)

Ꝑm **MARBOO FORMATION:** low-grade, turbiditic metasandstone and quartz-plagioclase-muscovite-biotite phyllite; quartz-muscovite-biotite(-cordierite) hornfels adjacent to granitoid intrusions
Ꝑmh Quartz-K-feldspar-plagioclase-muscovite-biotite-cordierite-andalusite hornfels and hornfelsic schollen (raft) migmatite

Tickalara Metamorphics (Central zone)

ꝐmTh Quartz-K-feldspar-plagioclase-biotite-cordierite-andalusite(-muscovite) hornfels and hornfelsic schollen (raft) migmatite

c. 1850 Ma

ꝐmTgd **DOUGALLS TONALITE:** foliated and metamorphosed hypersthene-hornblende-biotite tonalite; abundant pyroxene amphibolite inclusions, and inclusions of country rock
ꝐmTgk **CORKWOOD TONALITE:** foliated and metamorphosed biotite(-hornblende) tonalite and granodiorite; locally garnet-bearing

ꝐmToa Amphibolite; relict gabbroic textures
ꝐmTon Clinopyroxene-orthopyroxene-plagioclase-quartz(-hornblende-biotite) mafic granulite; local relict gabbroic textures

c. 1865 Ma

ꝐmTa Amphibolite; interlayered with pelite and psammite, banded iron-formation, calc-silicate rock, and marble
ꝐmTpc Calc-silicate rock and marble; minor banded iron-formation, quartzite, and amphibolite
ꝐmTpn Biotite-plagioclase-quartz(-K-feldspar-sillimanite-garnet-cordierite-spinel) migmatitic pelitic gneiss; numerous layers, pods, and angular to rounded schollen (rafts) of mafic granulite, psammitic gneiss, and calc-silicate rock; intruded by discontinuous syn- to post-metamorphic mafic to intermediate dykes

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 (560 Ma) (not present on map) — D_6
- Yampi Orogeny (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

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

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

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

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

Igneous layering
- trend of layering

Way-up indicator
- sedimentary structure

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

Gneissic banding, showing strike and dip
- inclined
- vertical
- trend of gneissic banding

Cleavage, showing strike and dip
- inclined

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
Fence, generally with track
Homestead — Bow River
Yard — Yd
Building
Mircowave repeater station
Reserve boundary
Horizontal control, minor
Breakaway
Ridge
Contour line, 20 metre interval

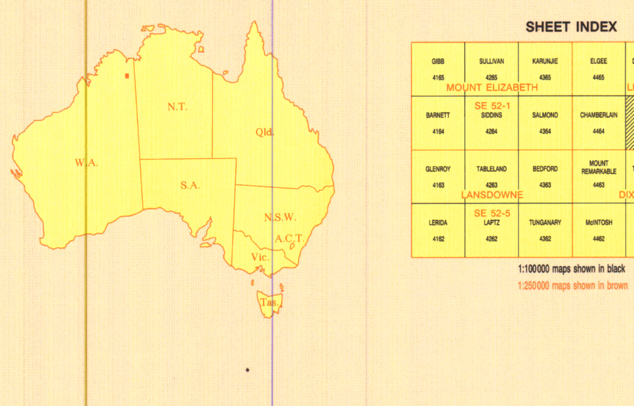
Watercourse, ephemeral; pool — P
Waterhole, spring — WH, S
Bore, well — B, W
Windpump
Dam, tank — D, T
Abandoned, position doubtful — (abd), (PD)

Major opencut — Argyle
Quarry
Alluvial workings
Prospect
Mining area

Mineral occurrence
- Copper — Cu
- Diamond — Di
- Gold — Au
- Gravel — Gr
- Iron — Fe
- Nickel — Ni

SIMPLIFIED GEOLOGY

SHEET INDEX



GOVERNMENT OF WESTERN AUSTRALIA
HON. NORMAN MOORE, M.L.C.
MINISTER FOR MINES

DEPARTMENT OF MINERALS AND ENERGY
K.R. PERRY, DIRECTOR GENERAL

NGMA
PRODUCT OF THE NATIONAL GEOSCIENCE MAPPING ACCORD

GEOLOGICAL SURVEY OF WESTERN AUSTRALIA
PIETRO GUJ, DIRECTOR

SCALE 1:100 000

1000 0 1 2 3 4 5 6 7 8 9 10
Metres — Kilometres

TRANSVERSE MERCATOR PROJECTION
Grid lines indicate 1000 metre interval of the Australian Map Grid Zone 52

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BOW

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