GEOLOGICAL SURVEY OF WESTERN AUSTRALIA

REPORT 14

PROFESSIONAL PAPERS FOR 1983

DEPARTMENT OF MINES
WESTERN AUSTRALIA

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CONTENTS

THE HYDROGEOLOGY OF LAKE MARIGINIUP, PERTH, WESTERN AUSTRALIA

by J. Hall

ABSTRACT

Lake Mariginiup is one of a series of round or oval lakes and swamps on the Swan Coastal Plain near Perth. The lake is up to 2 m deep and occupies an area of 1.6 x $10^6 m^2$, with reeds and bullrushes occupying the periphery. About 70% of the lake bed is covered with diatomaceous lake deposits up to 2 m thick. The remainder is covered with carbonaceous sand. These lake deposits rest on older Quaternary sands which unconformably overlie Late Cretaceous sedimentary rocks. The lake is in hydraulic connection with a regional unconfined groundwater flow system (Gnangara Mound); and variations in water table levels, together with rainfall and evapotranspiration, affect the level of water in the lake. Groundwater inflow takes place on the eastern side of the lake, through sandy lake deposits, and outflow takes place on the western side of the lake, through lake sediments. Between May 1979 and May 1980 groundwater inflow was estimated to be 1.3 x $10^6 m^3$; and rainfall over the lake, 1.1 x $10^6 m^3$. Evapotranspiration was estimated to be 2.2 x $10^6 m^3$; and outflow to groundwater (derived by difference), 0.2 x $10^6 m^3$. Outflowing water from the lake is more saline than groundwater inflow. Evapotranspiration accounts for about 92% of total water inputs to the lake.

INTRODUCTION

LOCATION AND TENURE

Lake Mariginiup is 25 km north of Perth, and 3 km northeast of Wanneroo (Fig. 1).

The lake is surrounded by freehold and Crown Land, the boundaries of which extend into the lake, and incorporate about 35 per cent of the lake area. A mineral claim for diatomite and peat is also registered for the area of the lake.

CLIMATE AND VEGETATION

The climate is Mediterranean, with hot, dry summers and mild, wet winters. The average annual rainfall is 840 mm, most of which falls during the winter months between April and October. The average daily maximum temperature ranges between 29.5°C, in February, and 17.0°C, in July. The average annual evaporation is 1 707 mm, which exceeds twice the rainfall. The evaporation is highest in January, with an average of 266 mm, and the lowest in June, with an average of 50 mm.

The vegetation in the central area of open water consists of algae and aquatic plants. A narrow band of reeds (*Cladium junceum*) borders the open water. A broad zone of reeds (*Cladium articulatum*) extends to the shores of the lake with a narrow zone of bullrushes (*Typha* spp.) and grasses at the periphery of the present shoreline (Fig. 2a). A few stands of high trees and scrub grow on areas of older, higher shorelines.

The bordering vegetation has been cleared for up to 100 m from the highest shoreline toward the middle of the lake. Most clearing has been done around the southern and eastern shores.

PHYSIOGRAPHY

Lake Maringiniup is one of a series of round or oval lakes and swamps situated in a northwesterly trending depression, the main part of which is near, or at the contact between the Bassendean and Spearwood Dune Systems (McArthur and Bettenay, 1974).

Within this depression, Lake Mariginiup occupies a shallow, circular basin, 1.5 km long and 1.3 km wide, with an elevation between 45 m and 50 m (AHD). The lake is bounded by relatively high sand ridges to the north, west and south, and by lower dunes to the east. A low saddle to the northeast of the lake separates it from a small depression occupied by an intermittent lake, known as Little Mariginiup Lake, which discharges, via a man-made drain, into the main lake during winter.

The areal extent of Lake Mariginiup varies annually as well as seasonally. The outer margin of the lake vegetation, which marks the present mean winter shoreline, is at an elevation of about 42 km (AHD). Within the winter shoreline the lake is approximately 1.3 km long and 1.1 km wide, covering an area of 1.6 km^2. Open water is restricted to an area of 1.4 km^2, while the remainder is vegetated by reeds and bullrushes.

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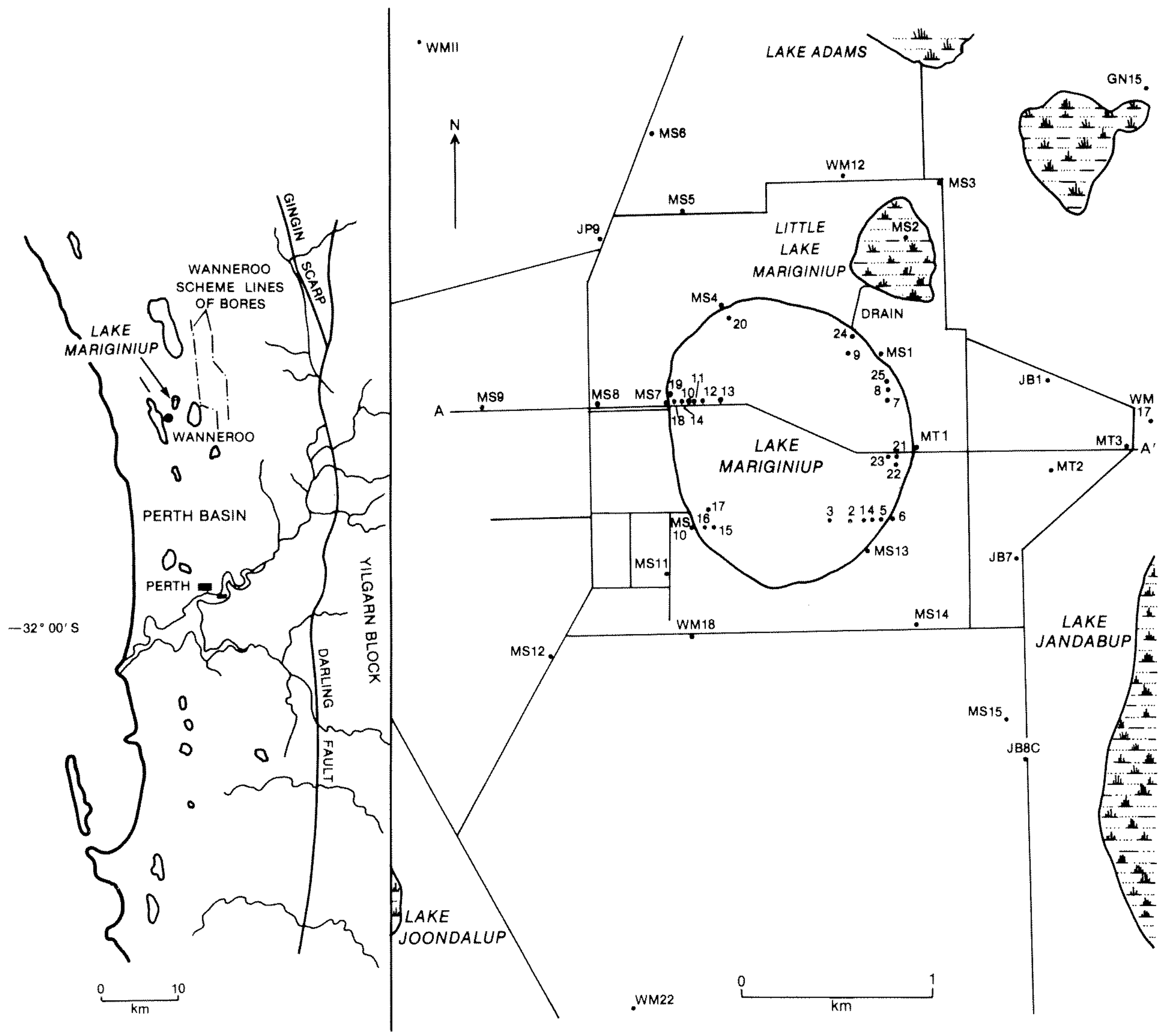

REFERENCE

MT1	MARIGINIUP DEEP, SHALLOW AND INTERMEDIATE BORES	GN15	GNANGARA BORE
MS4	MARIGINIUP SHALLOW BORE	JP9	JOONDALUP BORES
JB1	JANDABUP SHALLOW BORE	·22	AUGER/TRENCH SITE
WM11	WANNEROO MONITORING SHALLOW BORE	A—A′	LINE OF SECTIONS

GSWA 21089

Figure 1. Lake Mariginiup Locality Plan.

The lake bed is relatively flat except on the eastern side, where diatomaceous lake sediments form a mound which is exposed during periods of low water. Accurate bathymetric contours of the lake bed are unavailable, but poorly controlled contours, indicate that the depth of water in the lake probably varies from about 0.5 m to 2.0 m.

Approximately 70% of the lake bed is covered with diatomaceous sediments, and the periphery is covered with carbonaceous sand which extends beyond the present high-water mark in the lake. The extent of carbonaceous sand and the presence of ridges and benches on the eastern side of the lake indicate at least two periods of relatively stable and higher lake water levels in the past.

During summer, when lake water levels are low, springs with well-defined outflow channels and runnels develop along the shore and in the lake bed along the eastern side. These indicate that groundwater discharges into the lake along a broad seepage face. Small runnels also develop in the exposed lake sediments on the western side of the lake, between the zone of reeds and the area of open water, and indicate that a large quantity of water, temporarily held in the reeds and sediments, drains back into the lake as lake water levels decline.

Land owners have excavated three shallow sumps on the western and southwestern shores of the lake and these have permanent water.

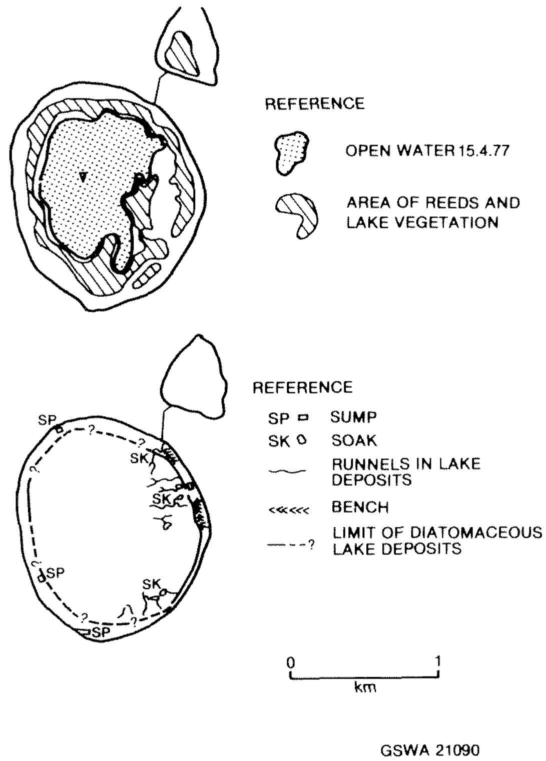

Figure 2. Physiographic features of Lake Mariginiup.

INVESTIGATION

An investigation programme to determine the hydrogeological environment of lakes likely to be affected by MWA (Metropolitan Water Authority) groundwater pumping schemes was proposed by Allen (1976). The present investigation of Lake Mariginiup is a modification of the original proposals.

OBSERVATION BORES

Twenty-four observation bores were drilled at 18 sites around Lake Mariginiup between November 1978 and February 1979. A proposed deep bore in the centre of the lake was not drilled owing to high costs and anticipated on-site difficulties. The bores were drilled by private contractors using cable-tool rigs, and ranged in depth from 9.0 m to 61.5 m, with an aggregate depth of 577.5 m. Details of these bores together with 12 MWA monitoring bores in the area are summarized in Table 1.

Deep, intermediate and shallow bores were drilled at three sites designated by the prefix "MT" (e.g. MT1D is the deep bore at site 1, MT1I is the intermediate bore at site 1). Shallow bores were drilled at the remaining sites, designated by the prefix "MS" (Fig. 1).

At the "MT" sites, the deep bore was drilled first. Lithological samples were taken at 2 m intervals and, after penetration of the underlying Cretaceous sediments, bottom-hole samples were collected for palaeontological examination. Each deep bore was geophysically logged (natural gamma ray) to aid in the definition of stratigraphic boundaries and the correlation of lithological samples.

One observation interval was selected near the bottom of each deep bore, at the base of the "superficial formations". Class 9, 80 mm PVC casing, with bottom cap, 3 m sump and 2 m slotted interval was run into each deep bore. The annulus was filled with very coarse sand to 0.5 m above the slotted interval, and then cement grouted back to the surface.

For each intermediate bore an interval near the middle of the "superficial formations" was selected and the slotted casing installed in the same manner as it was in the deep bores. The shallow bores were drilled to about 7 m below the water table. Class 9, 80 mm PVC casing with the bottom 7 m slotted and no sump was installed, and sand packed to the surface.

All bores were fitted at the surface with 152 mm protective steel casing set into a concrete base. This casing was fitted with either hinged caps or magnetic caps flush with ground level. Each bore was developed by bailing.

LAKE-DEPOSIT CORING

As part of the initial proposal (Allen, 1976, a series of cored holes was to be drilled in the lake bed. However, attempts at lake-bed coring at Lake Jandabup, in 1977, proved to be costly and mostly unsuccessful (Allen, 1979) and it was not attempted in Lake Mariginiup.

AUGERING AND TRENCHING

In March 1978, and October 1981, a number of auger holes were drilled and trenches excavated around the shores of the lake to determine groundwater movement and to examine lake sediments. (Fig. 1)

WATER SAMPLING

All bores were developed and on completion of development, in May 1979, the bores were pumped, using a portable submersible pump, and air-free samples were collected for standard chemical analysis. Water from the lake was sampled and analyzed at the same time (Table 2a). The bores were again sampled and analyzed in April 1980. (Table 2b).

All chemical analyses were made by the Government Chemical Laboratories.

WATER-LEVEL OBSERVATIONS

The natural surface and top of casing, for all bores, were levelled to the Australian Height Datum (AHD).

Synoptic groundwater levels have been monitored monthly in all observation bores since May 1979, and the results recorded in the MWA computerized groundwater record systems (GROWLS).

TABLE 1(a). SUMMARY OF CABLE TOOL DRILLING RESULTS—LAKE MARIGINIUP PROJECT BORES

Bore name	*Commenced*	*Completed*	*Natural surface (m AHD)*	*Steel casing (m AHD)*	*Total depth (m)*	*Slotted interval (m bns)*	*Water Level (m AHD) April 1980*	*Chloride (mg/L) April 1980*	*Status*	*Base superficial formations (m bns)*	*Comments*
MT1 D	27/11/78	4/12/78	45.359	45.359	52.0	45.0—47.0	42.90	101	Deep	48	?Poison Hill Greensand at base
MT1 I	4/12/78	6/12/78	45.285	45.285	27.0	22.0—24.0	42.87	123	Intermediate		—
MT1 S	6/12/78	7/12/78	45.254	45.254	9.0	2.0— 9.0	42.56	33	Shallow		—
MT2 D	3/1/79	12/1/79	49.520	50.102	61.5	54.0—56.0	43.57	106	Deep	59	?Poison Hill Greensand at base
MT2 I	12/1/79	16 1/79	49.490	50.068	29.0	24.0—26.0	43.52	90	Intermediate		—
MT2 S	16/1/79	18/1/79	49.430	50.038	18.0	11.0—19.0	43.46	88	Shallow		—
MT3 D	24/1/79	5/2/79	50.962	51.524	61.0	54.0—56.0	43.94	83	Deep	57.5	?Poison Hill Greensand at base
MT3 I	6/2/79	8/2/79	50.942	51.499	36.0	30.0—33.0	43.90	101	Intermediate		—
MT3 S	8/2/79	9/2/79	50.882	51.487	13.0	6.0—13.0	44.56	123	Shallow		—
MS1	7/12/78	18/12/78	44.366	44.366	9.0	2.0— 9.0	42.56	27	Shallow		—
MS2	8/12/78	9/12/78	44.083	44.083	9.0	2.0— 9.0	43.20	61	,,		—
MS3	9/12/78	11/12/78	53.731	53.731	19.0	8.5—19.0	43.61	111	,,		—
MS4	13/12/78	14/12/78	44.126	44.726	9.0	2.0— 9.0	41.46	300	,,		—
MS5	21/12/78	22/12/78	52.522	52.522	18.0	11.0—18.0	41.95	57	,,		—
MS6	31/1/79	1/2/79	54.361	54.370	22.5	15.5—22.5	40.25	114	,,		—
MS7	18/12/78	19/12/78	43.300	43.835	9.0	2.0— 9.0	41.00	525	,,		—
MS8	14/12/78	16/12/78	58.992	59.273	28.5	21.0—28.5	38.34	200	,,		—
MS9	16/12/78	18/12/78	59.369	60.098	28.5	21.5—28.5	38.16	52	,,		—
MS10	18/1/79	19/1/79	42.842	43.495	9.0	2.0— 9.0	41.09	374	,,		—
MS11	19/1/79	20/1/79	49.476	50.095	19.5	11.0—19.5	38.36	340	,,		—
MS12	23/1/79	29/1/79	66.616	67.218	38.5	31.0—38.0	37.86	na	,,		—
MS13	11/12/78	12/12/79	43.389	43.949	9.0	1.0— 9.0	42.31	47	,,		—
MS14	12/12/78	13/12/79	50.801	50.834	16.0	8.0—16.0	42.73	63	,,		—
MS15	22/1/79	24/1/79	62.995	63.230	26.5	19.2—26.5	43.19	35	,,		—

TABLE 1(b). SUMMARY OF DRILLING RESULTS—OTHER BORES USED IN STUDY

Bore name	*Commenced*	*Completed*	*Natural surface (m AHD)*	*Steel casing (m AHD)*	*Total depth (m)*	*Slotted interval (m bns)*	*Water Level (m AHD) April 1980*	*Chloride (mg/L) April 1980*	*Status*	*Base superficial formations (m bns)*	*Comments*
JB1	11/5/77	12/5/77	52.31	52.89	15.5	9.0—15.5	43.82		Shallow		—
JB7	9/7/77	11/7/77	57.06	57.63	22.0	10.0—22.0	43.50		,,		—
JB8C-WM21			55.79	55.80	na	na	43.22		,,		—
JB11C	14/4/77	15/4/77	63.18	63.22	20.0	18.0—20.0	na		,,		—
WM17			49.88	49.90	15.8	0—15.8	na		,,		—
GN15			49.79	53.03			46.61		,,		—
WM11			65.63	65.64	40.2	0—40.2	37.52		,,		—
WM12			56.43	56.50	25.6	0—25.6	43.02		,,		—
WM18			54.55	54.56	27.4	0—27.4	38.44		,,		—
WM22			65.48	65.52	41.4	0—41.4	37.42		,,		—
JP9			45.22	45.78			—		Fully penetrating	57.5	Poison Hill Greensand at base—used only for lithological information

bns = below natural surface; AHD = Australian Height Datum

TABLE 2(a). STANDARD ANALYSES OF GROUNDWATER AFTER BORE DEVELOPMENT (MAY 1979)

Bore	GCL Lab No.	pH	Turbidity (APHA units)	Colour (APHA units)	Odour	E.C. (mS/m @ 25° C)	Sat. Index (Langlier @ 20°)	TDS (E.C.) (a)	TDS (Calc) (a)	Free CO_2 (a)	Total Hardness (a) (b)	Total Alk (a) (b)	Ca (a)	Mg (a)	Na (a)	K (a)	CO_3 (a)	HCO_3 (a)	Cl (a)	SO_4 (a)	NO_3 (a)	SiO_2 (a)	B (a)	F (a)	Fe (a)	Mn (a)	As (a)	Cu (a)	Pb (a)	NH_4 (a) (c)	NO_3 (a) (c)	P (a)	Remarks (Pumped samples after . . .)
MT1 D	82419	5.6	2100	210	nil	43	-3.5	280	230	186	55	37	9	8	62	5	nil	45	103	9	<1	12	0.1	<0.1	2.2	0.4	<0.01	<0.02	<0.01	0.19	<0.02	0.15	2 hours @ 21.6 m³/d
MT1 I	82420	5.5	1400	280	H_2S	47	-3.8	300	250	171	57	27	8	9	68	4	nil	33	121	4	<1	11	0.11	<0.1	0.45	0.02	<0.01	<0.02	<0.08	0.37	<0.02	0.04	1½ hours 21.6 M³/d
MT1 S	82421	5.4	2.1	660	nil	17	-5.3	110	90	64	27	8	1	6	21	3	nil	10	32	16	<1	7	0.08	<0.1	0.28	<0.02	<0.01	0.06	0.02	0.20	0.06	0.51	as above
MT2 D	86632	5.8	190	220	H_2S	45	-3.0	290	240	155	59	49	12	7	60	4	nil	60	108	2	<1	14	0.1	<0.1	1.8	0.03	<0.01	<0.02	<0.01	0.42	<0.02	0.07	as above
MT2 I	82631	5.7	150	1120	nil	31	-4.3	200	170	80	21	20	2	4	52	1	nil	24	83	3	<1	9	0.13	<0.1	0.42	<0.02	<0.01	<0.02	<0.01	0.24	<0.02	0.05	as above
MT2 S	82630	5.5	300	740	nil	33	-4.2	210	180	164	40	26	3	8	47	1	nil	32	86	5	<1	9	0.09	<0.1	0.66	<0.02	<0.01	<0.02	0.01	0.24	<0.02	0.06	as above
MT3 D	82722	7.2	50	110	nil	67	-0.3	430	360	24	208	194	70	8	51	3	nil	236	90	1	<1	15	0.07	0.1	0.47	0.02	0.01	<0.02	<0.01	0.37	<0.02	0.07	2 hours @ 21.6 m³/d
MT3 I	82721	5.6	1800	130	H_2S	42	-3.4	270	220	176	56	35	11	7	57	4	nil	43	98	5	<1	12	0.09	<0.1	0.75	0.02	<0.01	<0.02	<0.01	0.40	<0.02	0.06	as above
MT3 S	82720	5.3	70	210	H_2S	46	-4.3	290	240	200	53	20	5	10	66	3	nil	24	120	7	<1	11	0.14	<0.1	0.57	<0.02	<0.01	0.02	<0.01	0.48	<0.02	0.02	as above
MS1	82516	4.9	4.9	500	nil	15	-5.6	100	80	176	21	7	2	4	19	1	nil	9	31	11	5	6	0.1	0.1	0.4	<0.02	<0.01	0.17	0.03	0.12	1.2	0.47	as above
MS2	82517	5.4	9.9	174	nil	32	-4.5	200	170	96	39	12	4	7	45	2	nil	15	79	13	<1	10	0.09	0.1	0.31	<0.02	<0.01	<0.02	0.01	0.13	<0.02	0.02	as above
MS3	82518	5.6	12	174	nil	33	-4.3	210	170	80	32	16	3	6	45	2	nil	19	79	10	2	9	0.09	<0.01	0.59	<0.02	<0.01	0.02	0.01	0.12	0.70	0.01	as above
MS4	82345	6.1	3.1	200	H_2S	123	-2.5	790	650	102	123	64	15	21	193	7	nil	78	313	35	<1	15	0.09	0.1	0.58	0.04	<0.01	0.05	<0.01	0.75	<0.02	0.04	as above
MS5	82515	5.6	17	163	nil	24	-4.2	150	130	75	47	15	4	9	26	1	nil	18	52	12	7	11	0.07	0.01	0.82	<0.02	<0.01	0.02	0.01	0.06	1.9	0.01	3 hours @ 14.4 m³/d
MS6	82514	5.8	95	310	nil	45	-3.5	290	240	95	48	30	6	8	69	2	nil	37	113	5	<1	10	0.07	0.01	0.52	0.02	<0.01	<0.02	0.01	0.22	<0.02	0.04	3 hours @ 14.4 m³/d
MS7	82346	5.9	0.5	150	nil	168	-2.4	1080	900	214	175	85	24	28	272	7	nil	103	469	20	<1	13	0.08	0.1	0.50	0.04	<0.01	0.02	<0.01	0.69	<0.02	0.05	2 hours @ 21.6 m³/d
MS8	82840	6.0	4.1	74	nil	97	-2.6	620	520	104	149	52	20	24	122	9	nil	63	205	67	22	15	0.08	0.1	0.75	0.02	<0.01	0.02	<0.01	1.11	4.3	0.01	6 hours @ 4.8 m³/d
MS9	82841	6.1	370	21	nil	36	-2.7	230	220	68	58	43	15	5	47	1	nil	52	51	26	27	15	0.05	<0.1	<0.05	<0.02	<0.01	0.02	<0.01	0.20	6.0	0.04	6 hours @ 4.8 m³/d
MS10	82417	5.8	0.8	170	H_2S	133	-2.7	850	690	240	141	76	17	24	207	9	nil	92	352	11	<1	9	0.11	<0.1	0.66	0.03	<0.01	<0.02	<0.01	0.70	<0.02	0.02	2 hours @ 21.6 m³/d
MS11	82513	5.8	18	158	nil	133	-2.9	850	680	209	131	66	13	24	202	9	nil	80	361	4	<1	13	0.13	0.0	0.49	0.02	0.01	<0.02	0.01	1.60	<0.02	0.02	3 hours @ 14.4 m³
MS12	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	not sampled
MS13	82723	5.7	99	140	nil	30	-3.6	190	170	108	38	27	7	5	42	2	nil	33	50	32	<1	12	0.08	0.1	0.47	0.02	<0.01	<0.02	<0.01	0.13	<0.02	0.05	2 hours @ 21.6 m³/d
MS14	82418	5.9	15	42	nil	45	-2.8	290	250	174	113	55	19	16	45	2	nil	67	73	39	8	8	0.06	0.1	0.09	0.02	<0.01	0.20	0.01	0.09	1.8	0.01	2 hours @ 21.6 m³/d
MS15	82842	5.3	2.9	89	nil	15	-4.8	100	90	70	12	7	1	3	22	1	nil	8	27	10	12	10	0.08	0.1	0.28	<0.02	<0.01	0.05	<0.01	0.02	2.6	0.01	6 hours @ 4.8 m³/d
LAKE MARIGINIUP	81244	6.8	5	170	nil	223	-1.3	1430	1160	35	245	110	32	40	329	17	nil	134	589	52	<1	13	0.03	0.01	0.11	0.02	0.01	0.02	0.01	0.14	0.02	0.07	Sample taken on southern shore of lake

(a) = mg/L (b) = as $CaCO_3$ (c) = Nitrogen as

TABLE 2(b). STANDARD ANALYSES OF GROUNDWATER (APRIL 1980)

Bore	GCL Lab No.	pH	Turbidity (APHA units)	Colour (APHA units)	Odour	E.C. (mS/m @ 25° C)	Sat. Index (Langlier @ 20°)	TDS (E.C.) (a)	TDS (Calc) (a)	Free CO_2 (a)	Total Hardness (a) (b)	Total Alk (a) (b)	Ca (a)	Mg (a)	Na (a)	K (a)	CO_3 (a)	HCO_3 (a)	Cl (a)	SO_4 (a)	NO_3 (a)	SiO_2 (a)	B (a)	F (a)	Fe (a)	Mn (a)	As (a)	Cu (a)	Pb (a)	NH_4 (a) (c)	NO_3 (a) (c)	P (a)	Remarks (Pumped samples after . . .)
MT1 S	81568	4.8	1.9	560	nil	18	-6.0	100	100	—	27	6	1	6	22	3	nil	7	33	21	< 1	7	0.05	< 0.1	0.24	< 0.02	<0.01	<0.02	0.01	0.16	0.02	0.20	
MT1 I	81569	5.0	180	76	nil	47	-4.3	260	240	—	58	21	10	8	65	2	nil	25	123	9	< 1	11	0.10	< 0.1	0.34	< 0.02	<0.01	<0.02	0.01	0.43	<0.02	0.10	
MT1 D	81570	5.1	380	65	nil	41	-4.1	230	210	—	51	27	9	7	59	2	nil	33	101	8	< 1	12	0.07	< 0.1	0.66	< 0.02	<0.01	<0.02	<0.01	0.33	0.02	0.06	
MT2 D	81653	5.4	64	96	nil	44	-3.6	240	230	271	57	34	13	6	60	3	nil	41	106	8	< 1	14	—	< 0.1	1.6	0.03	<0.01	<0.02	<0.01	0.32	<0.02	0.09	
MT2 I	81654	5.1	40	810	nil	34	-5.4	190	180	95	21	6	2	4	57	< 1	nil	7	90	11	< 1	8	—	< 0.1	0.29	< 0.02	<0.01	<0.02	<0.01	0.22	0.02	0.02	
MT2 S	81655	4.9	38	620	nil	33	-5.6	180	170	101	32	4	3	6	49	1	nil	5	88	10	< 1	9	—	< 0.1	0.35	< 0.02	<0.01	<0.02	<0.01	0.21	0.02	0.01	
MT3 S	81647	5.4	4.8	500	H_2S	46	-4.6	250	250	72	51	9	4	10	71	3	nil	11	123	19	< 1	10	—	< 0.1	0.56	< 0.02	<0.01	<0.02	<0.01	0.37	0.02	0.02	
MT3 I	81648	5.5	59	170	nil	40	-3.7	220	210	133	56	21	11	7	56	2	nil	25	101	9	< 1	12	—	< 0.1	0.40	< 0.02	<0.01	<0.02	<0.01	0.31	<0.02	0.02	
MT3 D	81649	6.7	1.6	35	nil	64	-0.7	350	350	79	214	199	71	9	46	3	nil	243	83	2	< 1	16	—	< 0.1	0.62	0.02	<0.01	<0.02	<0.01	0.30	<0.02	0.05	
MS1	81650	4.9	0.90	440	nil	14	-5.9	80	80	—	19	2	3	3	20	< 1	nil	2	27	14	4	6	—	< 0.1	0.28	< 0.02	<0.01	0.02	<0.01	0.07	1.1	0.42	
MS2	81651	5.0	1.4	150	nil	26	-5.3	140	140	—	28	6	3	5	38	1	nil	7	61	16	< 1	10	—	< 0.1	0.15	< 0.02	<0.01	<0.02	<0.01	0.08	<0.02	0.01	
MS3	81652	5.1	1.5	130	nil	43	-4.6	240	220	—	56	13	6	10	57	2	nil	16	111	14	4	9	—	< 0.1	0.52	< 0.02	<0.01	0.02	0.01	0.13	1.0	0.02	
MS4	81566	5.5	0.94	220	H_2S	118	-3.1	650	630	—	123	72	15	21	188	9	nil	88	300	34	< 1	14	0.05	< 0.1	0.38	< 0.02	<0.01	0.02	0.01	0.47	0.04	0.06	
MS5	81656	4.9	27	150	nil	26	-5.3	140	140	151	47	6	4	9	30	1	nil	7	57	13	11	11	—	< 0.1	0.64	< 0.02	<0.01	0.02	0.01	0.06	2.5	<0.01	
MS6	81443	5.0	5.6	300	nil	47	-4.4	300	240	—	50	25	8	68	67	2	nil	30	114	14	< 1	10	0.08	< 0.1	0.20	< 0.02	<0.01	<0.02	<0.01	0.23	< 0.02	0.01	
MS7	81510	5.5	0.65	290	H_2S	212	-2.7	1170	1030	—	185	102	25	30	327	9	nil	124	525	40	1	13	0.16	< 0.1	0.46	0.03	<0.01	<0.02	0.01	0.62	< 0.02	0.05	2 hours @ 14.4 m³/d
MS8	81681	5.6	28	100	nil	88	-3.1	480	480	246	118	49	16	19	121	6	nil	60	200	47	22	15	0.05	< 0.1	0.36	0.02	<0.01	0.04	0.01	0.89	4.9	<0.01	
MS9	81682	5.7	350	54	nil	39	-3.0	210	220	192	61	48	16	5	50	< 1	nil	58	52	26	31	15	0.05	< 0.1	0.46	< 0.02	<0.01	0.07	0.01	0.04	7.0	0.01	
MS10	81511	5.6	1.0	140	H_2S	144	-2.9	790	720	—	132	82	15	23	224	9	nil	100	374	12	< 1	8	0.11	< 0.1	0.24	0.03	<0.01	0.06	0.01	0.42	< 0.02	<0.01	2 hours @ 14.4 m³/d
MS11	81444	5.4	1.0	210	nil	132	-3.3	240	630	—	108	61	12	19	201	7	nil	74	340	7	< 1	11	0.12	< 0.1	0.38	0.02	<0.01	<0.02	<0.01	1.2	< 0.02	0.01	
MS12	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	not sampled
MS13	81567	5.3	6.5	93	nil	29	-4.4	160	170	—	37	17	5	6	41	2	nil	21	47	39	5	13	0.05	< 0.1	0.29	< 0.02	<0.01	<0.02	0.01	0.11	1.2	0.04	
MS14	81445	5.6	2.5	19	nil	42	-3.1	270	230	—	105	45	19	14	38	3	nil	55	63	47	6	9	0.05	< 0.1	0.05	< 0.02	<0.01	<0.02	<0.01	0.04	1.6	<0.01	
MS15	81683	5.1	33	46	nil	20	-5.0	110	110	—	23	11	3	4	27	< 1	nil	13	35	15	13	10	0.06	0.1	0.28	< 0.02	<0.01	0.12	0.01	0.05	3.3	0.01	2 hours @ 9.6 m³/d

GEOLOGY

SETTING

Lake Mariginiup is situated within the Perth Basin. It occupies a shallow, circular depression in the Quaternary "superficial formations", which unconformably overlie Late Cretaceous sedimentary rocks. The lake contains a thin sequence of lake deposits.

STRATIGRAPHY

The near surface stratigraphic sequence at Lake Mariginiup is summarized in Table 3.

TABLE 3. STRATIGRAPHIC SEQUENCE, LAKE MARIGINIUP

Age	Formation		Maximum thickness (m)	Lithology	Comments
Quaternary	"Superficial formations"	Lake Deposits	2	Diatomite, peat carbonaceous sand	
		? DISCONFORMITY			
			70	Sand; fine to coarse, minor ferruginous sand, clay and gravel	
UNCONFORMITY					
Late Cretaceous	?Poison Hill Greensand		4+	Glauconitic sandstone	Formation not definitely identified

CRETACEOUS

?Poison Hill Greensand

Glauconitic, silty sands were intersected at the base of bores MTD, MT2D and MT3D on the eastern side of the lake, and in bore JP9 on the northwestern side of the lake. Palaeontological evidence (Backhouse, 1979; Cockbain, 1979; Hooper, 1972) suggests that these sands belong to the Poison Hill Greensand (Fairbridge, 1953).

The formation is unconformably overlain by the "superficial formations". The total thickness of the formation is not known, but exceeds 4 m in bore MT1D.

QUATERNARY

"Superficial formations"

The "superficial formations" consist of sands and lake deposits.

Sands: The sands of the "superficial formations" consist of an unconsolidated sequence of shallow-marine and eolian deposits. The upper part predominantly consists of poorly to moderately sorted, fine to coarse sand containing minor clay and scattered heavy minerals and feldspar. The lower part consists of very poorly sorted, very fine to coarse sand, and gravel, ranging in thickness from about 4 m to 10 m and containing clay, heavy minerals, and scattered feldspar and phosphatic pebbles.

An irregular and discontinuous bed of variably cemented ferruginous sand, ("coffee rock") occurs at and below the water table on the eastern side of the lake.

The unconformity with the Poison Hill Greensand occurs at an elevation of between about -3 m (AHD) and -10 m (AHD), and the thickness of the "superficial formations" ranges from 45 m to 70 m depending on the topography.

Lake Deposits: At the periphery of the lake, the deposits consist of fine- to medium-grained carbonaceous sand and peat, up to 1 m thick, which grades into carbonaceous diatomite toward the central part of the lake. The maximum thickness of diatomite intersected while augering was 1.6 m, but it is probably about 2 m in the middle of the lake. The lake deposits disconformably overlie the sands of the "superficial formations".

A low mound of diatomite on the eastern side of the lake was probably formed by the prevailing westerly wind during periods when the level of the lake was very low.

STRUCTURE

A geological cross section through Lake Mariginiup is shown in Figure 3.

The Poison Hill Greensand is believed to be flat lying, but the unconformity with the overlying "superficial formations" is undulating.

The "superficial formations" are essentially flat lying, but the upper surface of the basal sands and gravels appears to parallel the unconformity surface.

The lake deposits form a thin veneer on the lake bed but are thicker in the middle and on the eastern side of the lake where the mound of diatomite occurs.

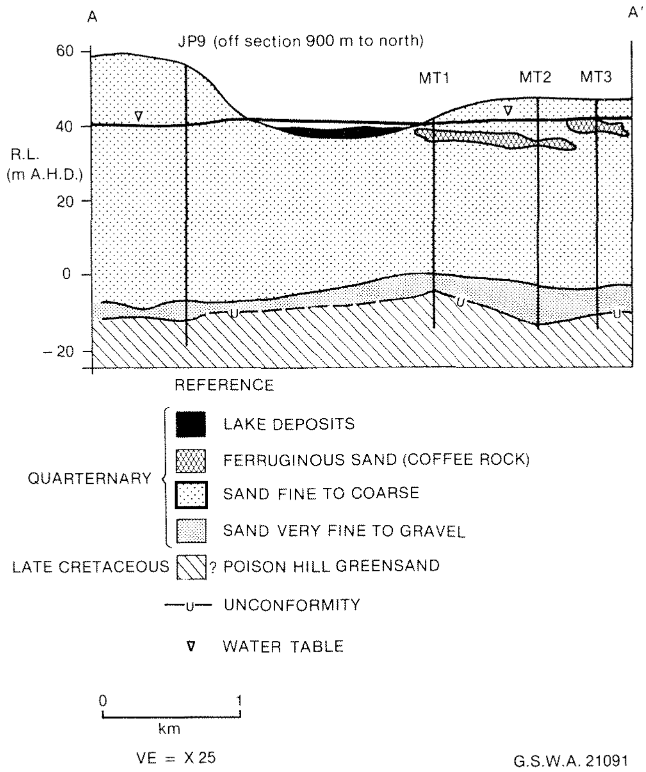

Figure 3. Geological Cross Section A-A' of Lake Mariginiup.

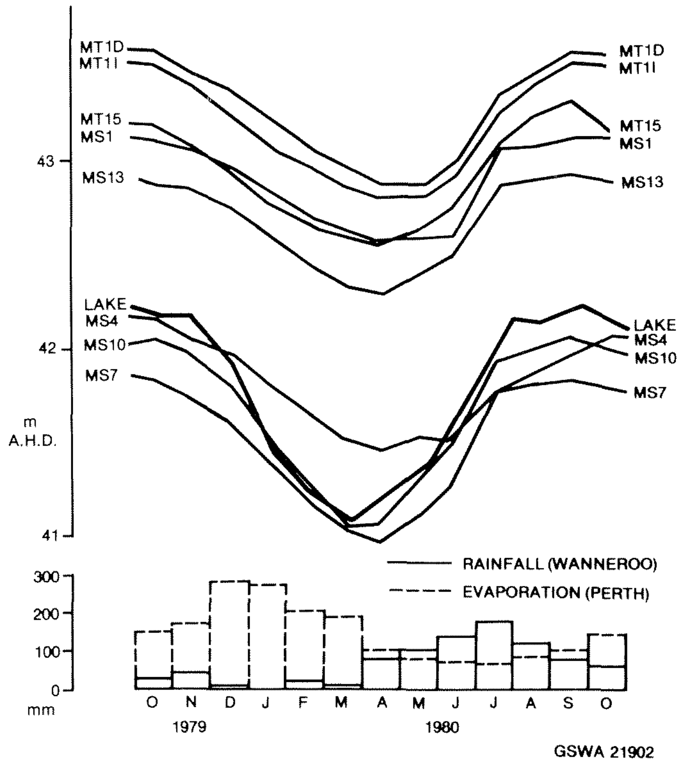

Figure 4. Hydrographs of selected bores and Lake Mariginiup.

HYDROGEOLOGY

LAKE—GROUNDWATER INTERACTION

Lake Mariginiup is in hydraulic connection with the unconfined groundwater which forms part of the Gnangara Mound regional flow system.

At present, the elevation of the water table on the eastern side of the lake is higher than that of the lake bed, so that groundwater inflow, together with rainfall, maintains the lake.

The elevation of the water table on the western side of the lake is lower than the water-level of the lake, so that some water not lost by evapotranspiration leaves the lake by outflow as groundwater.

Water also enters the lake via a narrow drain on the northeastern side of the lake from Little Mariginiup Lake.

VARIATIONS IN WATER TABLE AND LAKE WATER LEVELS

The water table and water-level of the lake vary in phase (Fig. 4), but the water-level of the lake responds to a greater extent to periods of high rainfall or evapotranspiration than does the water table.

The water table and water-level of the lake are highest at the end of winter (usually in October), and lowest at the end of summer (usually in April).

CONFIGURATION OF THE WATER TABLE

Water-table contours for the end of winter 1979 and the end of summer 1980 (Figs 5 and 6) show that the water table has a generally uniform gradient on the eastern side of the lake, except where flow converges toward the lake. On the western side of the lake, the water table has a generally flatter gradient than on the eastern side, except for a narrow zone adjacent to the lake which has a very steep gradient, and flow diverges outward, away from the lake.

The configuration of the water-table contours has the same general shape for summer and winter, but the greater seasonal response of water-levels of the lake than of the water table (refer to the previous section) affects the areas through which groundwater onflow to the lake and outflow to groundwater from the lake takes place, and thus the configuration of the water-table contours up and downstream from the lake. For example, in summer, when the water-level of the lake declines at a faster rate than the water table, groundwater inflow is diverted from a wider area than in winter, when the water-level of the lake rises at a faster rate than the water table.

The steep hydraulic gradient at the western margin of the lake (Figs 5 and 6) suggests that the aquifer is not homogeneous. A possible explanation is that the hydraulic conductivity of the "superficial formations" beneath the western side of the lake is less than that for the rest of the aquifer.

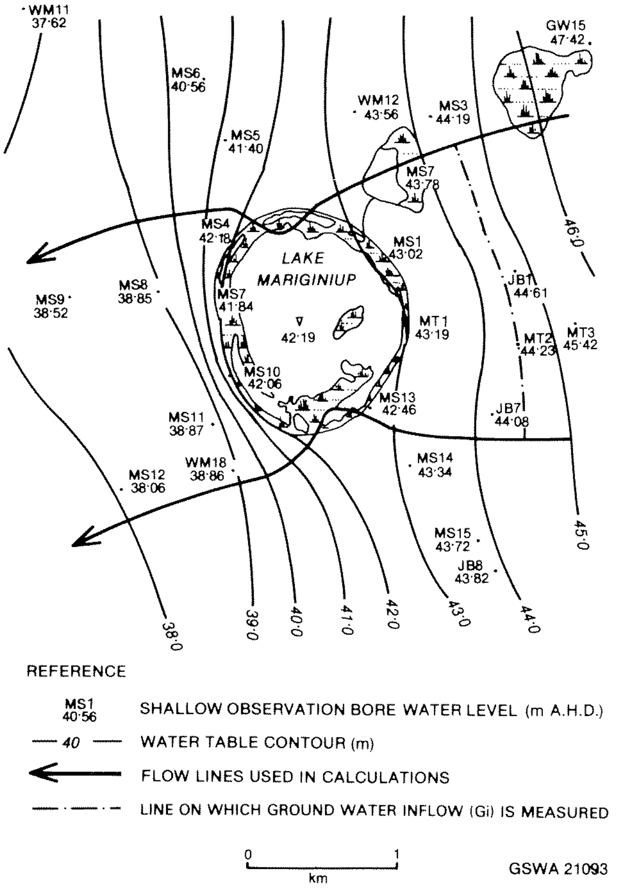

Figure 5. Water Table Contours of Lake Mariginiup for October 1979.

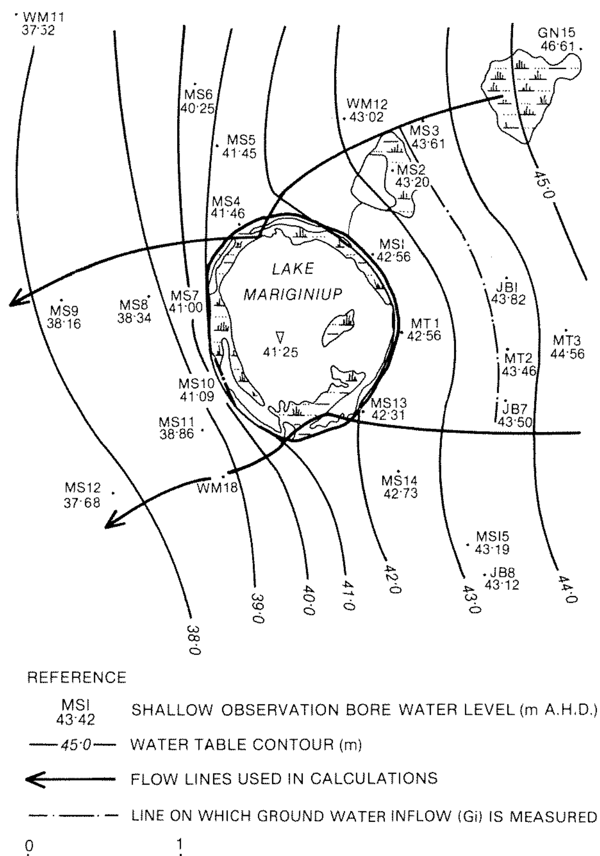

Figure 6. Water Table Contours of Lake Mariginiup for April 1980.

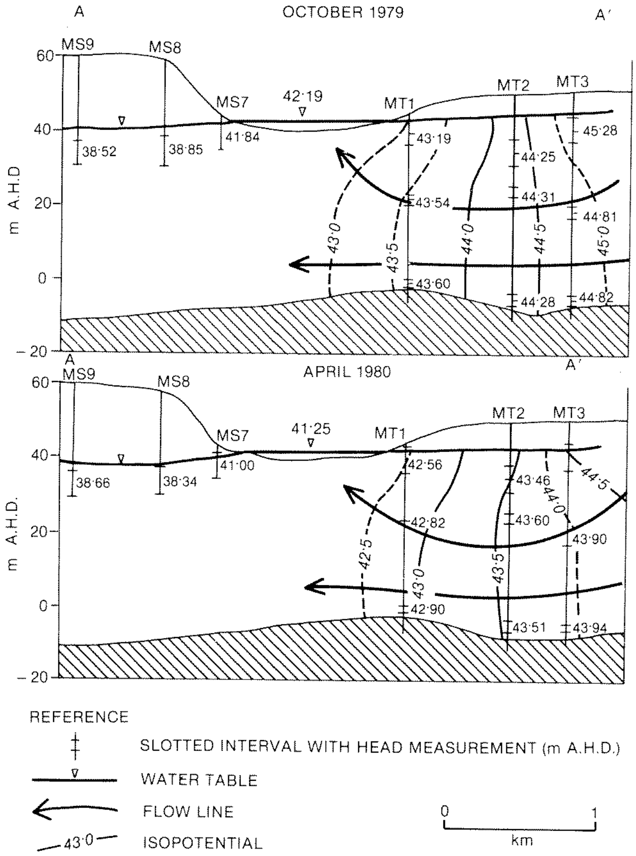

Figure 7. Hydrogeological Section A-A' of Lake Mariginiup.

GROUND FLOW SYSTEM OF THE LAKE

Sections, approximately parallel to groundwater flow and including shallow, intermediate and deep head measurements from three sites on the eastern side of the lake, are shown in Figure 7.

The sections show that groundwater enters the eastern side of the lake by upward flow from the upper half of the aquifer. This was confirmed in auger holes and trenches on the eastern side of the lake, which detected increasing heads with depth and, in some cases, small artesian flows. The groundwater inflow occurs along a broad seepage face marked by channels and runnels on the eastern shore.

Outflow occurs on the western side of the lake. It is inferred to be by downward flow, as indicated by decreasing heads with depth observed in auger holes and trenches. The thickness of the flow section is presumably less in summer than in winter, when outflow to groundwater is less, owing to higher evapotranspirative losses and less rainfall input.

CONFIGURATION OF SALINITY CONTOURS

Figure 8 shows the contours of salinity measured at the water table in March 1979 (Table 2). The configuration of the salinity contours demonstrates the

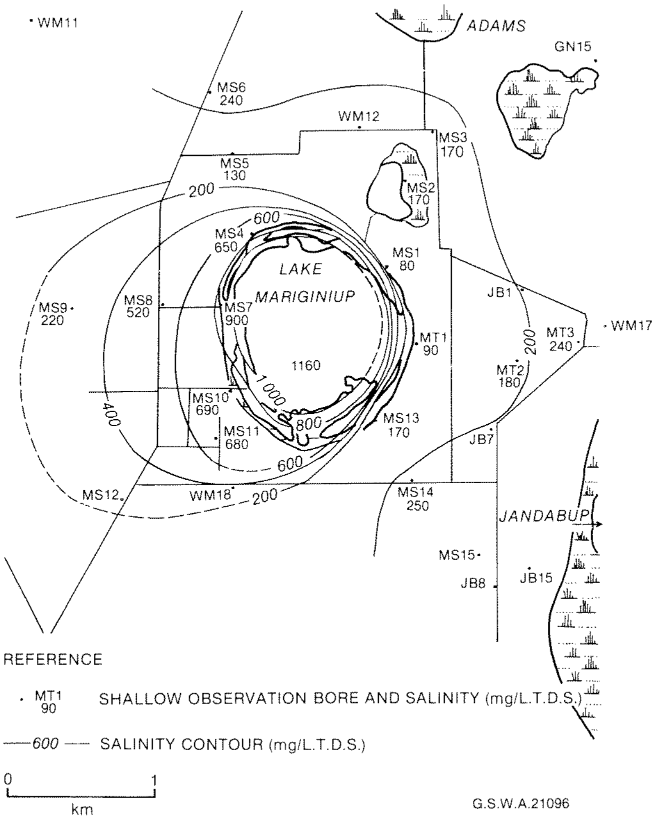

Figure 8. TDS Salinity Contours of Lake Mariginiup, May 1979.

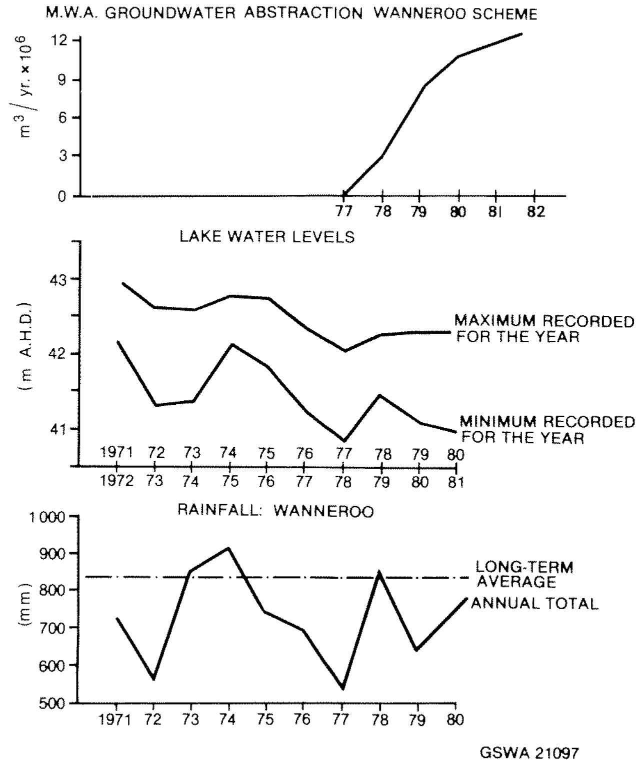

Figure 9. Rainfall, lake water levels and MWA groundwater abstraction, Lake Mariginiup, 1971-1980.

effects on the adjacent groundwater of evapotranspiration from the lake. Groundwater flowing into the lake, at the water table, is very fresh (less than 200 mg/L TDS), but the water in the lake and water outflowing from the lake is more saline (up to about 1 200 mg/L TDS) owing to the concentrating effect of high evapotranspirative losses from the lake. Outflow from the lake has a downward component so that the salinities at the water table will be somewhat lower (up to 700 mg/L TDS) than those at depth.

TRENDS IN LAKE WATER-LEVELS 1971-1980

Rainfall, maximum and minimum lake levels, and annual groundwater abstraction rates from the Wanneroo Scheme bores for the period 1971-1980 are shown in Figure 9. (Note that the time scale for the minimum lake levels is a year out of phase with the time scale for the maximum lake levels). In attempting to correlate rainfall with the lake levels it is necessary to consider the minimum lake levels following the rainfall period. These occur in the following year, usually March or April.

The maximum and minimum lake water-levels reflect the rainfall, but the variations in the maximum lake water levels are more subdued owing to the shape and depth of the lake.

Groundwater abstraction from the Wanneroo Scheme bores commenced in 1976, reaching full production in 1980 (approx. 12.2 x $10^6 m^3/a$).

Apart from a possible dampening of maximum and minimum lake level fluctuations, this abstraction has had no apparent effect on the lake water-levels.

For comparison, lake water-levels for Lakes Jandabup and Joondalup (Fig. 1) are shown in Figure 10. These show similar responses to those at Lake Mariginiup.

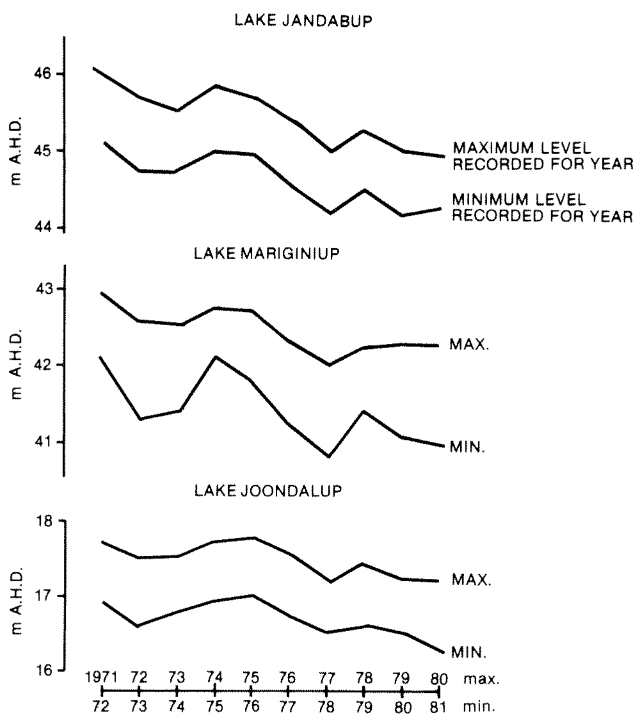

Figure 10. Hydrographs for Lakes Jandabup, Mariginiup and Joondalup, 1971-1980.

WATER BALANCE

A water balance is an accounting of all water entering and leaving a finite water system. For Lake Mariginiup the water balance can be expressed by the equation:

$$Gi + R + V_{79} = D + E + V_{80} \quad (1)$$

where Gi = groundwater inflow to the lake
R = rainfall on the lake
V_{79} = volume of the lake in 1979
D = outflow to groundwater, from the lake
E = evapotranspiration from the lake
V_{80} = volume of the lake in 1980

Groundwater abstraction by land owners in the immediate vicinity of Lake Mariginiup is not monitored and bore records are not complete, but aerial photography suggests that there are problably no more than ten private bores around the lake. Assuming that ten bores pump at an average of 200 m^3/d (probably an overestimate), the annual abstraction is less than $0.1 \times 10^6 m^3$. This figure is insignificant in the water balance so the effects of groundwater abstraction by private bores have been disregarded.

Annual flow via a drain from Little Mariginiup Lake was estimated to be less than $0.1 \times 10^6 m^3$ and has also been disregarded.

Outflow to groundwater and the volume of water in the lake in 1979 and 1980 cannot be estimated directly because of the lack of sufficient groundwater-head data on the western side of the lake and reliable bathymetric contours. However, by selecting a period when the lake water-levels at the beginning and end are essentially the same, there is negligible change in lake storage.

Equation (1) may then be rewritten as:

$$Gi + R = D + E \quad (2)$$

and outflow to groundwater may be estimated by difference.

Estimates of the components of the water balance have been evaluated for the period May 1979 to May 1980, when the lake water-levels were virtually the same at 41.421 m and 41.408 m (AHD) respectively.

GROUNDWATER INFLOW

Groundwater inflow to the lake may be estimated by using the form of the Darcy equation:

$$Q = KbIL \quad (3)$$

where Q = groundwater inflow (m^3/d)
K = hydraulic conductivity (m/d)
b = aquifer thickness (m)
I = hydraulic gradient (dimensionless)
L = width of flow section (m)

An estimate for Q is obtained by evaluating this equation for an end of winter and end of summer situation and averaging the values. Hydraulic conductivity is assumed to be a uniform 30 m/d on the eastern side of the lake. The other components of Equation (3) are measured on the water table maps and hydrogeological sections for October 1979 (end of summer) and April 1980 (end of winter). (Figs 5 to 7). They are measured on lines midway between the 44 m and 45 m contours (October 1979) and the 43 m and 44 m contours (April 1980), where the isopotentials are near vertical and groundwater flow is horizontal.

The aquifer thickness is taken as 27 m. Figure 7 shows that groundwater flow into the lake is from the upper half of the aquifer.

Hydraulic gradients were measured at a number of points between the flow lines bounding the flow section at the water table, and mean gradients of 2.3×10^{-3} and 1.8×10^{-3} were calculated for the end of winter and end of summer. However, as Figure 7 shows, the hydraulic gradients at the water table are steeper than those at depth. Analysis of the distribution of hydraulic gradients with depth indicates that the mean hydraulic gradient through the upper half of the aquifer is approximately 70% of that at the water table. The hydraulic gradients measured at the water table have therefore been adjusted to 1.7×10^{-3} and 1.3×10^{-3}.

The flow section widths (L) measured between the bounding flow lines are 2.1×10^3 m and 2.2×10^3 m for the end of winter and end of summer respectively.

The groundwater flows for the upper part of the aquifer through the line of section for October 1979 and April 1980 are:

$$Q_{(OCT)} = 30 \times 27 \times 1.7 \times 10^{-3} \times 2.1 \times 10^3$$
$$= 2.9 \times 10^3 m^3/d$$

and

$$Q_{(APR)} = 20 \times 27 \times 1.3 \times 10^{-3} \times 2.2 \times 10^3$$
$$= 2.3 \times 10^3 m^3/d$$

and the average daily flow is

$$Q = 2.6 \times 10^3 m^3/d$$

This value, however, underestimates the actual groundwater inflow to the lake because it takes no account of the increase in groundwater flow, due to net rainfall recharge to the shallow groundwater, between the line of measurement of the components of Equation (3) and the lake shore.

This was determined by applying the percentage net rainfall recharge to the rainfall for the period over the area between the lake shore, the line on which the components of groundwater inflow were calculated, and the bounding flow lines (Figs 5 and 6).

The percentage net rainfall recharge for the area was estimated from the ratio of the chloride concentration of rainfall to the chloride concentration of the groundwater at the water table. The mean annual chloride concentration of rainfall, in Perth, was 12.0 mg/L in 1973 and 10.7 mg/L in 1974 (Hingston and Gailitis, 1977). The average of these two values, 11.4 mg/L, is assumed to be the rainfall chlorinity for May 1979 to May 1980. The average chloride concentration of groundwater at the water table in the area under review was about 50 mg/L (Fig. 11) and the percentage net rainfall recharge was:

$$(11.4/50) \times 100 = 23\%$$

The rainfall for the period was 704 mm (Table 4) and the mean area was about $1.6 \times 10^6 m^2$ and therefore the net rainfall recharge to groundwater inflow was:

$$= 0.23 \times 0.704 \times 1.6 \times 10^6$$
$$= 0.3 \times 10^6 m^2$$

The groundwater inflowing to the lake for the period May 1979 to May 1980 becomes:

$$Gi = (2.6 \times 10^3 \times 366) + (0.3 \times 10^6)$$
$$= 1.3 \times 10^6 m^3$$

RAINFALL

The rainfall component (R) was estimated by applying the total rainfall for the period may 1979 to May 1980 (Table 4) to the area of the lake.

The area of the lake is approximately 1.6×10^6 m^2 so:

$$R = 0.704 \times 1.6 \times 10^6$$
$$= 1.1 \times 10^6 m^3$$

EVAPOTRANSPIRATION

Evapotranspiration from the lake is the sum of direct evaporation from the open water surface and transpiration by vegetation within the area of the lake. It is assumed to be 80% of the pan evaporation. Table 5 shows the pan evaporation for Perth.

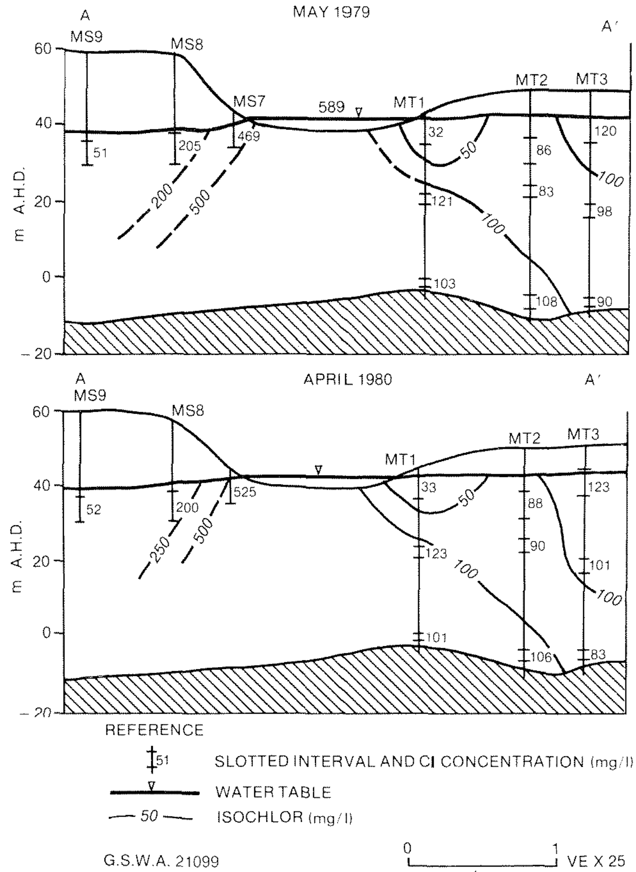

Figure 11. Isochlor Section A-A' of Lake Mariginiup.

The evapotranspiration (E) component was estimated by applying the pan evaporation for the period May 1979 to May 1980 to the area of the lake:

$$E = 1.749 \times \frac{80}{100} \times 1.6 \times 10^6$$

$$= 2.2 \times 10^6 m^3$$

BALANCE

By substituting the estimates for the components of the water balance into equation (2), the water balance is:

$$(1.3 + 1.1) \times 10^6 = (D + 2.2.) \times 10^6$$
$$2.4 \times 10^6 = (D + 2.2) \times 10^6$$

The outflow to groundwater (D), obtained by difference, is 0.2×10^6 m^3.

TABLE 4. RAINFALL AT WANNEROO

	1979							1980					
	June	July	Aug	Sept	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Total
Rainfall (mm)	154	126	112	41	28	5	4	1	13	3	77	100	704

TABLE 5. CLASS-A PAN EVAPORATION AT PERTH

	1979							1980					
	June	July	Aug	Sept	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Total
Evaporation (mm)	54	59	70	102	156	173	284	273	209	190	101	78	1 794

GROUNDWATER THROUGHFLOW

Groundwater throughflow (i.e. flowing beneath the lake) may be estimated by applying equation (3) to the bottom half of the aquifer. Hydraulic conductivity (K) and width of flow section (L) will be the same as for the groundwater inflow calculations; aquifer thickness is 27 m; and hydraulic gradients (I) are 1.2 x 10^{-3} and 1.0 x 10^{-3} for October 1979 and April 1980 (measured on Fig. 7).

Groundwater throughflow G_T for October 1979 and April 1980, then, is estimated to be:

$$G_{T\,(OCT)} = 30 \times 27 \times 1.2 \times 10^{-3} \times 2.1 \times 10^3$$
$$= 2.0 \times 10^3 \ m^3/d$$
$$\text{and } G_{T\,(APR)} = 30 \times 27 \times 1.0 \times 10^{-3} \times 2.2 \times 10^3$$
$$= 1.8 \times 10^{-3} \ m^3/d$$

with a mean value of:

$$G_T = 1.9 \times 10^3 \ m^3/d.$$

For the period May 1979 to May 1980 the throughflow is estimated to be:

$$G_T = 1.9 \times 10^3 \times 365$$
$$= 0.7 \times 10^6 \ m^3$$

The total groundwater flow beneath the eastern side of the lake (Gi + G_T) is

$$= (1.3 + 0.7) \times 10^6 \ m^3$$
$$= 2.0 \times 10^6 \ m^3$$

Net loss of water from the lake (R-E) is

$$= (1.1 - 2.2.) \times 10^6$$
$$= -1.1. \times 10^6 \ m^3$$

Therefore groundwater flow through the total aquifer thickness beneath the lake is reduced by

$$\frac{[1-(2.0-1.1)]}{2.0} \times 100 = 55\%$$

CHLORIDE MASS BALANCE

A chloride mass balance is an accounting for all chloride entering and leaving a groundwater system. The chloride in the components of the water balance (Equation 3) affect the mass of chloride in the system, except for evapotranspiration which only affects chloride concentration. The chloride mass balance may be expressed as:

$$\text{Gi}\,[Cl]_i + R\,[Cl]_R + V_{79}\,[Cl]_{L79} = D[Cl]_D + V_{80}\,[Cl]_{L80} \qquad (4)$$

where $[Cl]_i$ = chloride concentration of groundwater in flowing to the lake

$[Cl]_R$ = chloride concentration of rainfall

$[Cl]_{L79}$ = chloride concentration of the lake water in 1979

$[Cl]_D$ = chloride concentration of the groundwater outflowing from lake

$[Cl]_{L80}$ = chloride concentration of the lake water in 1980

Assuming that the mass of chloride in the lake was essentially the same for the beginning and end of the period, the last terms on either side of Equation (4) may be deleted, so that the chloride mass balance equation becomes:

$$\text{Gi}\,[Cl]_i + R[Cl]_R = D[Cl]_D \qquad (5)$$

If the chloride mass balance is calculated over the same period as the water balance, the results may be used to check the accuracy of the water balance.

GROUNDWATER INFLOW

Comparison of the isochlor sections (Fig. 11) with the hydrogeological sections (Fig. 7) indicates that the mean chloride concentration of groundwater inflowing to the lake is about 77 mg/L. The groundwater inflow for the period has been estimated to be 1.2 x $10^6 m^3$ so the mass of chloride entering the lake by groundwater inflow is:

$$\text{Gi}[Cl]_i = 1.3 \times 10^6 \times 77 = 100 \text{ t.}$$

RAINFALL

The mean chloride concentration of rainfall at Wanneroo is 11.4 mg/L and the rainfall input to the lake has been estimated to be 1.1 x $10^6 m^3$, so the mass of chloride entering the lake via rainfall is:

$$R[Cl]_R = 1.1 \times 10^6 \times 11.4 = 13 \text{ t.}$$

OUTFLOW TO GROUNDWATER

Assuming that the chloride concentrations in bore MS 7 (Fig. 11) reflect the concentrations of the water outflowing from the lake to groundwater, a mean value of 497 mg/L is used. Outflow to groundwater was estimated by difference to be 0.2 x 10^6 m^3, so the mass of chloride leaving the lake is:

$$D[Cl]_D = 0.2 \times 10^6 \times 497 = 99 \text{ t.}$$

MASS BALANCE

If the estimates of the chloride mass balance are substituted into Equation (5), the balance becomes:

$$100 + 13 \text{ t} \approx 99 \text{ t.}$$
$$113 \text{ t} \approx 99 \text{ t.}$$

This is a fair balance considering the quality of data used and the assumptions made, and tends to confirm that the water balance estimates are of the right order.

CONCLUSIONS

The water balance is based on limited data, mainly owing to the lack of adequate bores downstream of the lake, and many assumptions. However, the chloride mass balance tends to confirm that the water balance is probably of the right order.

Lake Mariginiup is mostly maintained by groundwater inflow from the upper half of the aquifer and by rainfall.

Outflow to groundwater from the lake is retarded by lake sediments and evapotranspiration removes about 92% of water input to the lake. Consequently the salinity of the lake is considerably higher than that of the groundwater and the outflow forms a plume of saline water downstream of the lake.

The lake is a groundwater sink. The groundwater flow beneath the "superficial formations" is reduced by 55%.

Groundwater abstraction from the Wanneroo Scheme bores has had no apparent effect on the lake. This is not surprising since the nearest production bore is over 3 km to the east and all production bores are screened against the bottom third of the aquifer, intercepting groundwater which flows beneath the lake. However, large-scale abstraction of groundwater from the upper half of the aquifer by private bores in the vicinity, and upstream, of Lake Mariginiup could reduce the amount of groundwater flowing into the lake and thus affect lake water levels.

RECOMMENDATIONS

To improve the water balance, more precise values for outflow, lake storage and their respective chlorinities are needed. However, of more importance are the long-term effects, if any, of groundwater abstraction on the lake, and the following recommendations are made:

(a) Bathymetric contours of the lake should be improved and the present form of monitoring of the lake water-level be replaced by a permanent staff, located near the middle of the lake. This would enable the production of a depth/area/volume rating chart and more accurate measurements of lake water levels.

(b) Some estimate of groundwater abstraction by landowners near the lake should be made.

(c) A rainfall gauge should be placed on the lake shore and measured monthly.

(d) Monitoring of groundwater levels should be maintained on a monthly basis.

(e) Aerial photography should be implemented on an annual basis to monitor broad changes in lake physiography and land use in the vicinity of the lake.

(f) Some measurement of the water contribution to Lake Mariginiup via the drain from Little Lake Mariginiup should be made.

(g) If the above recommendations are adopted, all monitoring should be reviewed annually to determine the likely causes of any changes in the lake.

(h) In all future investigations of this type, intermediate and deep bores should be installed on both the upstream and downstream sides of the lake in question.

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CARBONIFEROUS OF WESTERN AUSTRALIA—A REVIEW

by A. E. Cockbain

ABSTRACT

Rocks of Carboniferous age (the base of which is taken at the base of Tournaisian Tnlb (Fig. 2) and the top at the base of microfloral assemblage Unit II) occur in the Carnarvon, Canning and Bonaparte Basins in Western Australia. Sedimentation was predominantly siliciclastic with a small amount of carbonate, particularly in the Bonaparte Basin. Early Carboniferous sediments were deposited under near-tropical conditions and cooling took place gradually, culminating in glacial conditions in Late Carboniferous (and continuing into Early Permain) times. There seems to have been a widespread period of non-deposition in the mid Carboniferous although sedimentation was continuous in the basinal parts of the Bonaparte Basin and probably in the deeper parts of the Fitzroy Trough.

INTRODUCTION

Carboniferous rocks make up only a small proportion of the Phanerozoic rocks of Western Australia. They crop out in three sedimentary basins—the Carnarvon, Canning and Bonaparte Basins*—and are known in the subsurface of all three basins from oil-exploration wells (Fig. 1). The Carboniferous rocks are mainly marine clastics and carbonates with some continental clastics. There are no coal deposits in the Carboniferous System in Western Australia.

The Carboniferous is taken to include the Tournaisian, Visean, Namurian, Westphalian and Stephanian Series and to extend from 280 to 345 million years ago. The Strunian Stage (Tnla) is here considered latest Devonian. Correlation of the rocks in Western Australia with the European series is based mainly on the brachiopod, conodont and foraminifer faunas and on the microfloras (Druce,1969; Kemp and others, 1977; Mamet and Belford, 1968; Nicoll and Druce, 1979; Playford,1971, 1976; Roberts, 1971; Thomas, 1971). No isotopic age determinations are available on Carboniferous rocks from Western Australia.

The boundary between the Carboniferous and Permian Systems has been traditionally taken as the unconformity at the base of the glacial sequence in Western Australia, that is at the base of the Lyons Formation, Grant Group and Keep Inlet Formation in the Carnarvon, Canning and Bonaparte Basins respectively. However, recent work indicates that the major break is not at the base of the Permian but is within the Upper Carboniferous.

*The Geological Survey of Western Australia has dropped the term "Gulf" from the name of this basin, thus bringing it in line with those other basins (e.g. Ord, Bremer) in which the geographic descriptor is omitted.

BIOSTRATIGRAPHY

The correlation of some of the biostratigraphic zones used in Western Australia is shown in Figure 2. Zonal schemes using conodonts have been erected in the Bonaparte Basin (Druce, 1969; modified by Nicoll and Jones, pers. comm., 1981) and Canning Basin (Nicoll and Druce, 1979) and ostracod zones have also been erected in the Bonaparte Basin by Nicoll and Jones (pers. comm., 1981). A number of brachiopod assemblages were established by Thomas (1971) in the Carnarvon, Canning and Bonaparte Basins and Roberts (1971) published a brachiopod zonation based on Bonaparte Basin material. Roberts' scheme is followed here with two modifications. The *Anthracospirifer milliganensis* zone and the *Echinoconchus gradatus* fauna are combined into one zone (*milliganesis—gradatus* zone); and the interzone between the *pauciplicatus* zone and this new zone is eliminated (see discussion under Burvill Formation). Two of the four species comprising the *gradatus* fauna occur in the *milliganesis* zone in the Burvill Formation and the other two species occur in the Waggon Creek Formation which is correlated with the Burvill Formation (Roberts, 1971). Consequently it seems that the *gradatus* fauna cannot be satisfactorily separated from the *milliganesis* zone.

The microfloral assemblages, described mainly from the Canning and Bonaparte Basins, have been synthesized by Kemp and others (1977). They provide additional evidence for amalgamating the *milliganensis* zone and the *gradatus* fauna. In Bonaparte 1, core 2, the *largus* microflora is associated with the *gradatus* fauna while in the lower cores 4, 5 and 6 it occurs with *milliganesis*-zone brachiopods However, in Kulshill 1 the brachiopods are associated with younger microfloras; the *gradatus* fauna in core 23 occurs with the *ybertii* microflora

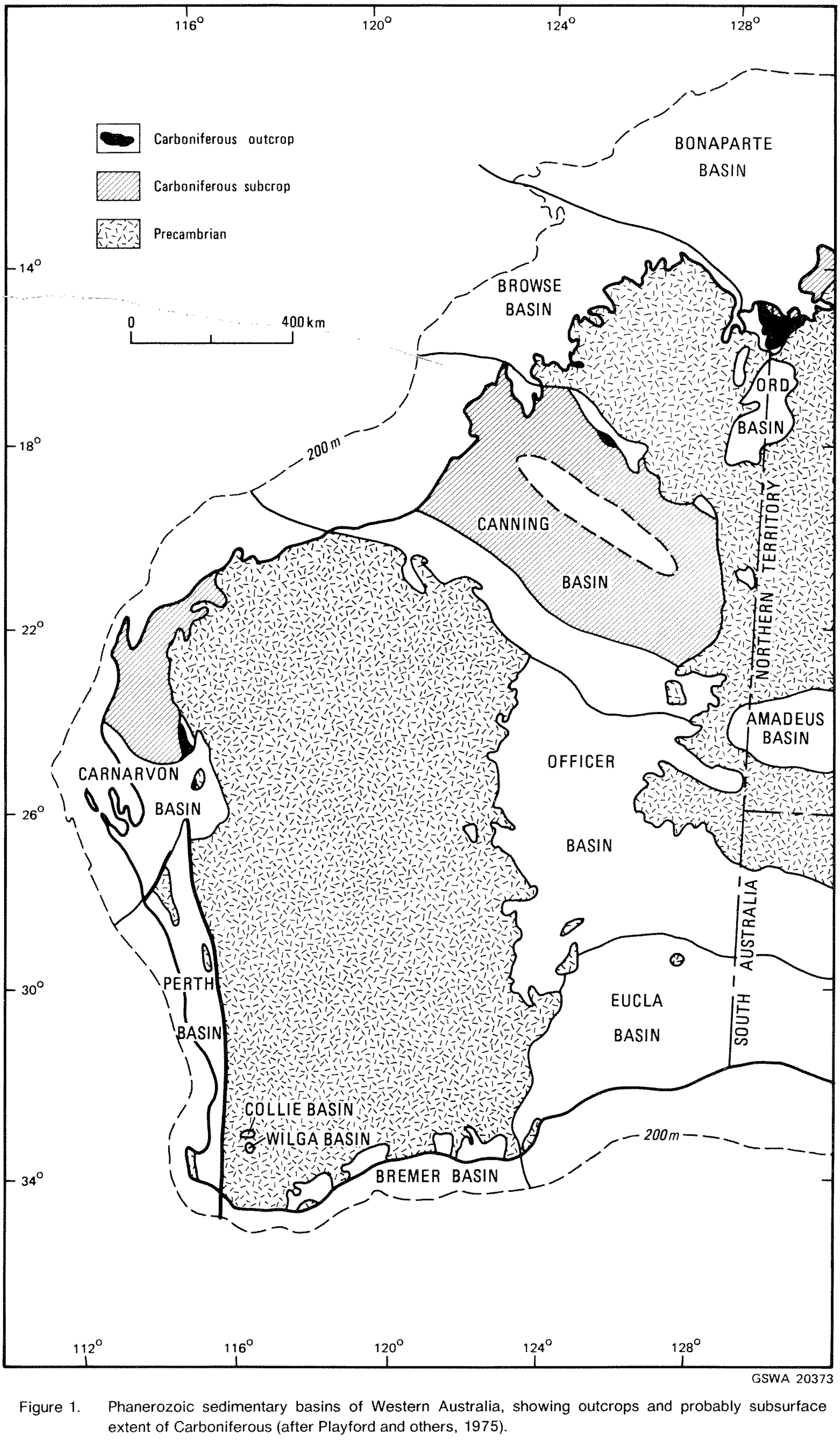

Figure 1. Phanerozoic sedimentary basins of Western Australia, showing outcrops and probably subsurface extent of Carboniferous (after Playford and others, 1975).

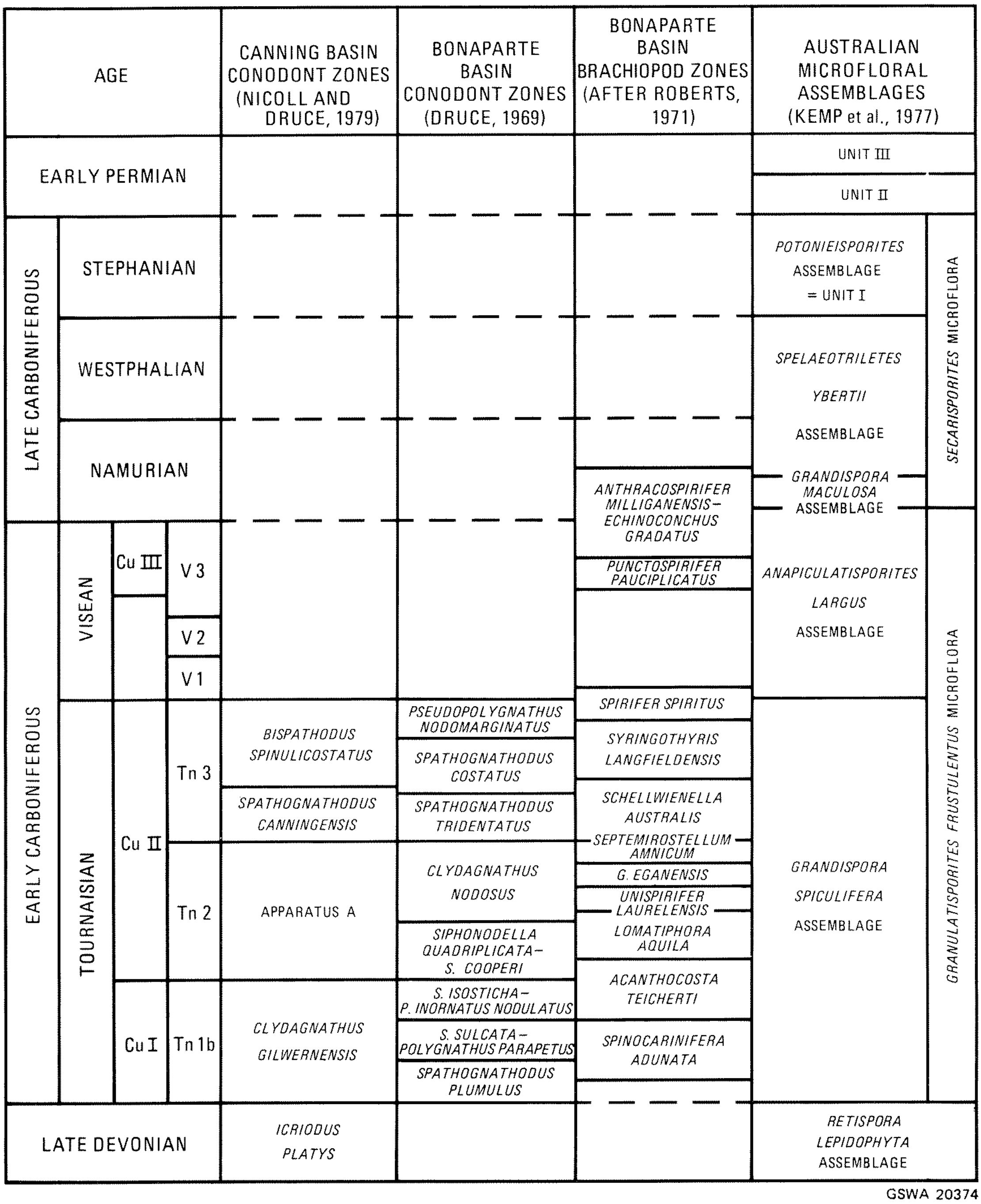

Figure 2. Correlation of conodont, brachiopod and mnicrofloral zonations in Western Australia.

and the deeper core 27 contains *milliganensis*-zone brachiopods and the *maculosa* microflora. Assuming that the zone boundaries are synchronous, the brachiopod and microfloral zones are correlated in Figure 2. This correlation supports Kemp and others' (1977) suggestion (attributed by them to Bischoff *in* Jones and others, 1973) that the upper limit of the *largus* microflora is early Namurian.

At various times the Carboniferous-Permian boundary has been drawn at the base of Unit I (Helby, 1969), the base of Unit II (Evans, 1969), or at the base Unit III (Balme, 1980). While this is not the place to review this problem (*see for example*, Kemp and others, 1977; Waterhouse, 1976; Archbold, 1982), the base of Unit II is here taken to coincide with the Carboniferous-Permian boundary; this has the additional advantage of retaining the bulk of the "Permian glacial sequence" in the Permian.

CARNARVON BASIN

Rocks of Carboniferous age crop out along the eastern margin of the Carnarvon Basin and have been encountered in a number of exploratory wells drilled in the northern part of the basin (Fig. 3). Four formations are recognised in the Williambury-Moogooree area

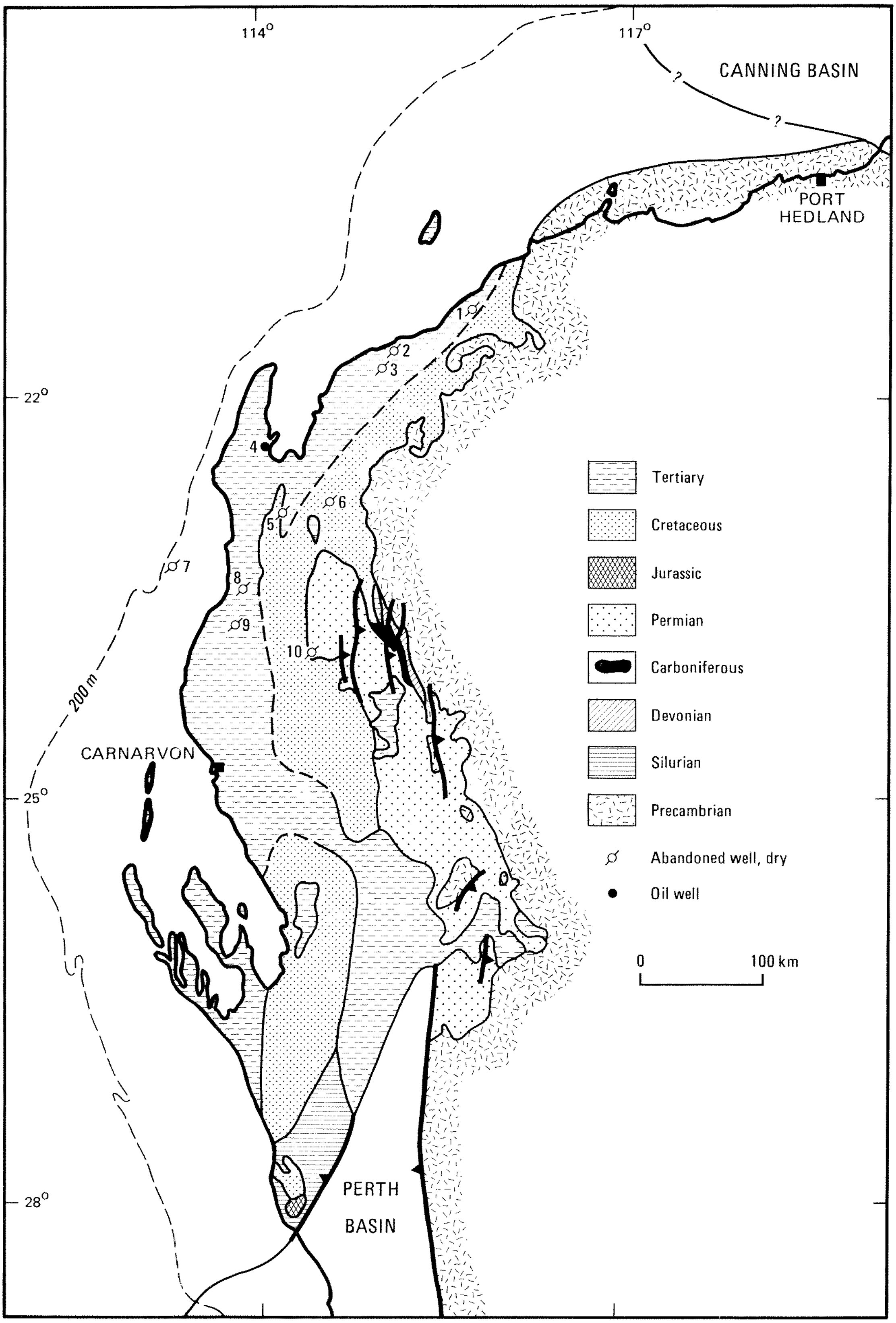

Figure 3. Carnarvon basin, showing solid geology and position of principal oil-exploration wells encountering Carboniferous rocks (after Playford and others, 1975). 1. Yarraloola 1; 2. Cane River 1; 3. Minderoo 1; 4. Rough Range 1; 5. Remarkable Hill 1; 6. East Marilla 1; 7. Pendock 1; 8. Warroora 1; 9. Gnaraloo 1; 10. Quail 1.

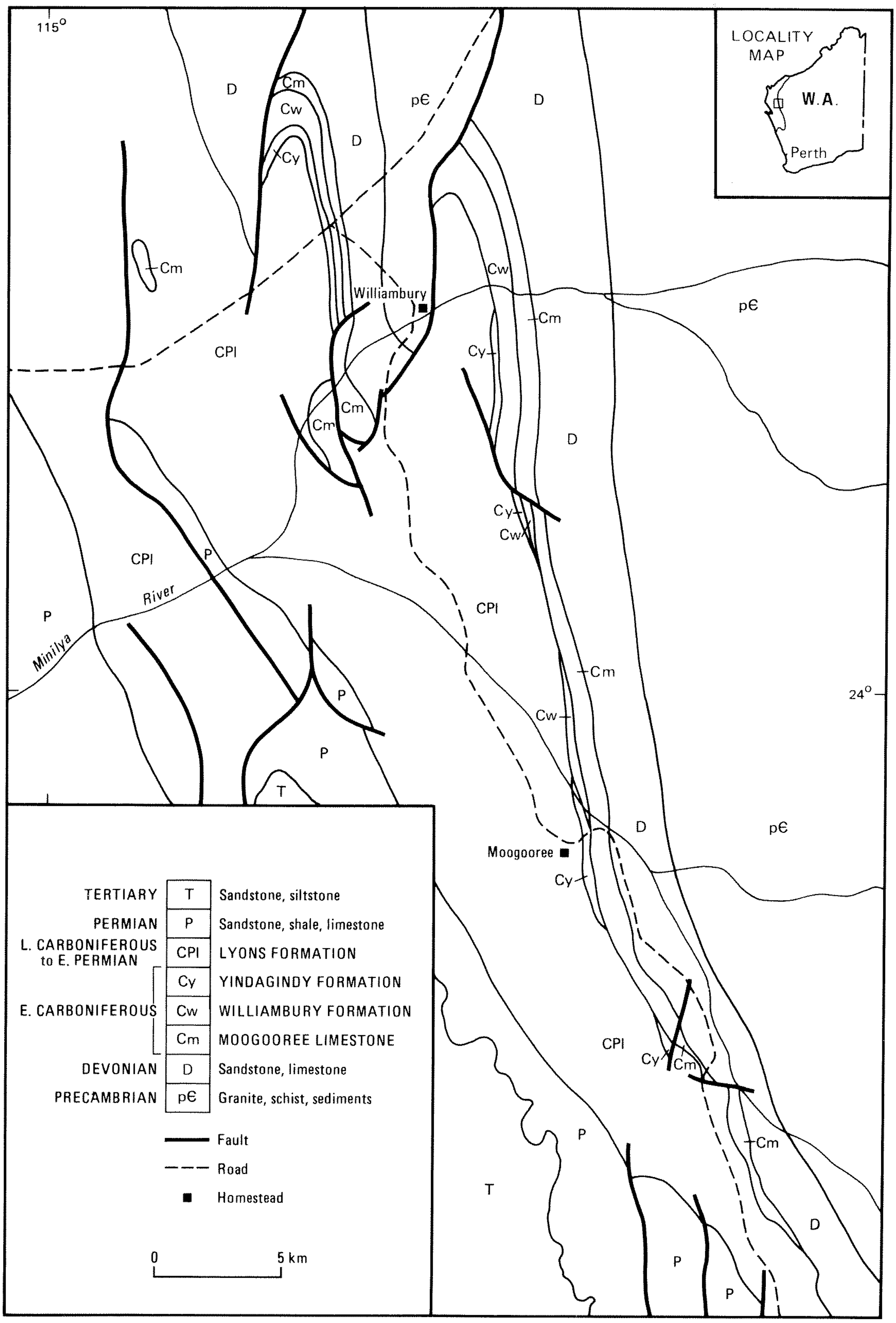

Figure 4. Carnarvon Basin, solid geology of the Williambury Moogooree area (after Hocking, Williams, Lavaring and Moore, 1983; and Hocking, Williams, Moore, Denman and Lavaring, 1983).

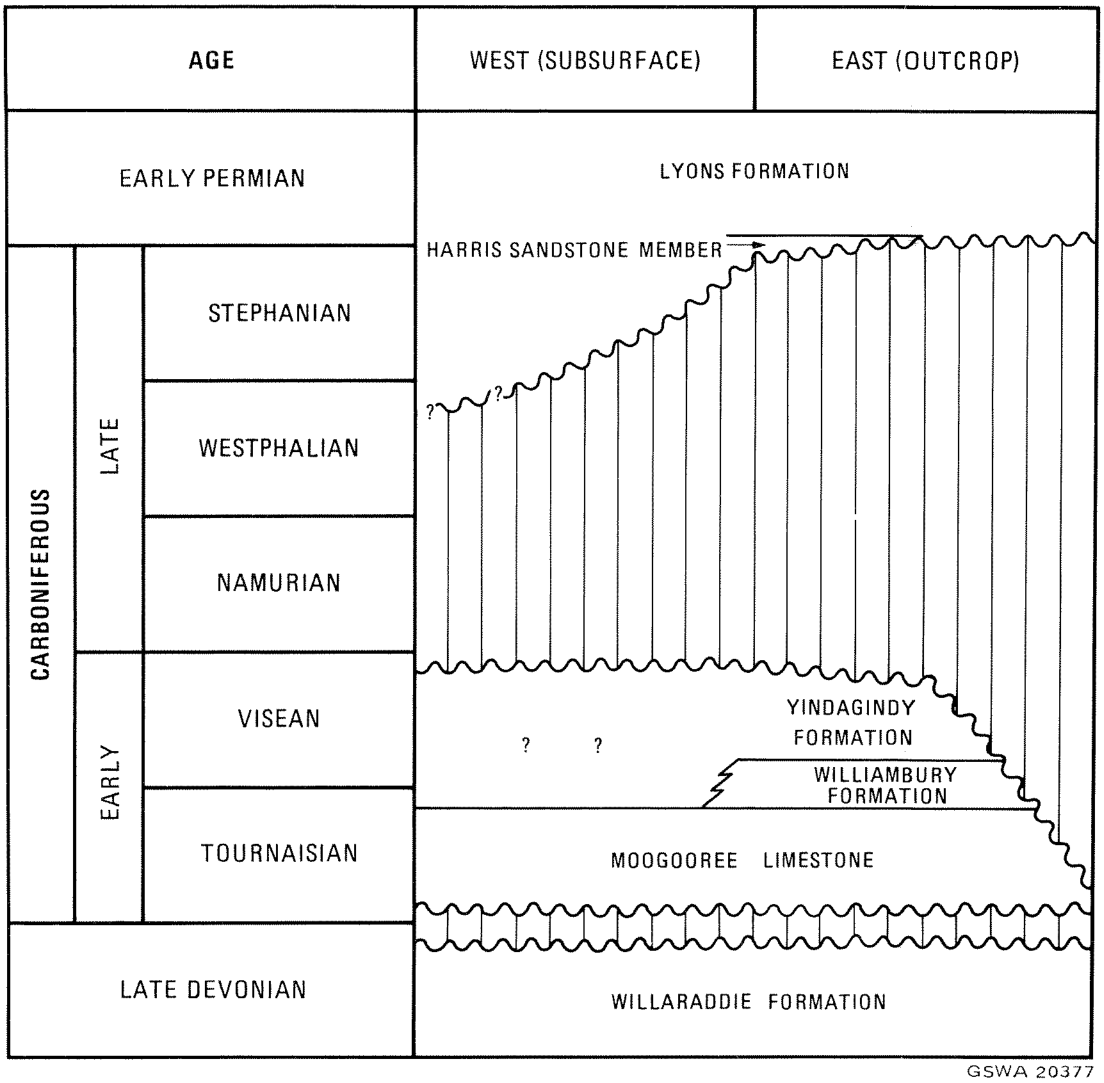

Figure 5. Carnarvon basin, stratigraphic correlation chart.

(Fig. 4), some 180 km northeast of Carnarvon. They are (in ascending order) the Moogooree Limestone, Williambury Formation, Yindagindy Formation and the Lyons Formation which continues into the Early Permian (Fig. 5). Only the Moogooree Limestone and the Lyons Formation can be confidently recognized in the subsurface. The maximum thickness of Carboniferous rocks exceeds 1 000m.

STRATIGRAPHY

Moogooree Limestone

The Moogooree Limestone (Teichert, 1949; Condon, 1954) is a sequence of calcarenite with some quartz sandstone and dolomite which overlies the Devonian Willaraddie Formation (possibly disconformably) and is overlain conformably by the Williambury Formation or unconformably by the Lyons Formation. The type section is 4 km northwest of Gneudna Well and is 326 m thick. The formation is exposed in a series of fault-controlled synclines around Williambury homestead and extends some distance south of Moogooree homestead. The unit is widespread in the subsurface of the northern Carnarvon Basin, and occurs in Pendock 1, Quail 1 and East Marrilla 1. The Moogooree Limestone was deposited in a near-shore shallow-marine environment (Lavaring, 1979) which at times may have been evaporitic (Radke an Nicoll, 1981).

The formation is richly fossiliferous. In the upper part of the unit the fossils are silicified and chert nodules are common. The most abundant fossils are brachiopods including *Prospira* aff. *laurelensis*, *Unispirifer fluctuosus*, *Syringothyris spissus* and *Rhipidomella michelini?* (Glenister, 1956; Thomas, 1962b, 1971; Veevers, 1959b; Lavering, 1979). Other fossils include the corals *Syringopora* and *Palaeacis*, ostracods, bivalves, gastropods, bryozoans, crinoid columnals and stromatolites. Thomas (1971) considered the fauna to belong to his *Unispirifer fluctuosus* assemblage and to be of Tournaisian age. This assemblage broadly correlates with *laurelensis*, *eqanensis* and *amnicum* zones of the Bonaparte Basin (Roberts, 1971) which on conodont evidence are of middle Tournaisian age (Druce, 1969).

Williambury Formation

The Williambury Formation ("Williambury Sandstone" of Teichert, 1949; emended Condon, 1954) is a sequence of sandstone, commonly conglomeratic, and siltstone which lies conformably between the Moogooree Limestone below and the Yindagindy Formation above. In some areas it is overlain unconformably by the Lyons Formation. The unit is poorly exposed between Williambury and Moogooree homesteads. The type section is in Gneudna paddock, 4 km southeast of Williambury homestead and is 235 m thick. Condon states that the maximum thickness is 338 m.

The Williambury Formation is of fluvial origin and formed as an alluvial fan centred on the Williambury area (Hocking, Williams, Lavaring and Moore, 1983). No fossils are known from the formation but it is probably Lower Carboniferous from its stratigraphic position.

Yindagindy Formation

The Yindagindy Formation ("Yindagindi Limestone" of Teichert, 1950; emended Condon, 1954) is a unit of coarse-to medium-grained sandstone with thin interbeds of oolitic and algal limestone which conformably overlies the Williambury Formation and is overlain disconformably by the Lyons Formation. The formation is named after Yindagindy Creek and the type section is 4 km west of Williambury homestead. This section is 76 m thick and the maximum thickness of the unit is about 115 m (Condon, 1965). The formation crops out between Williambury and Moogooree homesteads, generally with very poor exposure.

Some of the limestone beds are richly fossiliferous but the shells are difficult to extract. The fossils identified include brachiopods (e.g. *Camarotoechia* sp. and *Composita variabilis;* Thomas, 1962b, 1971), ostracods, bryozoans, gastopods, serpulid worms, crinoid columnals and stromatolites (Thomas, 1962b). Thomas (1971) considers that the brachiopods suggest a Visean age. However, in the Bonaparte and Canning Basins the highest unit beneath the major unconformity extends into the early Namurian and the Yindagindy Formation could be Visean to early Namurian in age. The unit was deposited in shallow-marine conditions with the sandy interbeds suggesting intermittent terrigenous sedimentation (Hocking, Williams, Lavaring and Moore, 1983).

Lyons Formation

The Lyons Formation ("Lyons Conglomerate" of Maitland, 1912; "Lyons Group" of Teichert, 1950; emended to Lyons Formation by van de Graaff and others, 1977) is a unit of feldspathic wacke with minor siltstone, conglomerate, limestone, calcareous sandstone, tillite and varved shale. The type section is on the north side of the Wyndham River, some 200 km east of Carnarvon and is about 1 125 m thick (Condon, 1967). The formation unconformably overlies the underlying Carboniferous and older rocks and is conformably overlain by the Callytharra or Carrandibby Formations. It crops out extensively in the eastern Carnarvon Basin and occurs in the subsurface of the northern part of the basin. The maximum known thickness is more than 2 300 m in Remarkable Hill 1. This figure includes the 850 m of "Permo-Carboniferous" clastic sediments originally excluded from the Lyons Formation by Berven (1969). At the base of the formation there is a discontinuous basal sandstone, the Harris Sandstone Member (here taken to include the Austin Member) which locally contains boulder beds, ice-drag striae and lepidodendroid plant remains.

The Lyons Formation was deposited in marine and fluvial environments under glacial conditions. It contains microfloras belonging to Unit I (in East Marilla 1; Dolby in Osborne, 1972) and Unit II (Kemp and others, 1977) and is of Late Carboniferous and Early Permian age. The macrofossils are all Sakmarian (Dickins and Thomas, 1959). Archbold (1982) considers the upper part of the formation to be early Sakmarian (Tastubian) in age, rather than late Asselian as suggested by Waterhouse (1976).

Subsurface Carboniferous rocks

Clastic rocks of Carboniferous or probable Carboniferous age have been encountered in Quail 1, Gnaraloo 1, Warroora 1, Rough Range 1, Minderoo 1, Cane River 1 and Yarraloola 1 drilled in the northern Carnarvon Basin. Palaeontological control is very poor.

The sequence in Quail 1 may be taken as typical (Pearson,1964). Here 67 m of siltstone rests on 285 m of sandstone. These rocks are underlain (probably conformably) by the Moogooree Limestone and are overlain (unconformably?) by the Lyons Formation and may correlate with clastic sediments in Rough Range 1, Warroora 1 and Minderoo 1. The microflora in Quail 1 is probably of Early Carboniferous age (B. E. Balme, pers. comm., 1981). In Minderoo 1 the section was originally dated as Late Carboniferous (Edgell *in* Johnstone and others, 1963) on the basis of a microflora that is now thought to be late Early Carboniferous in age (Balme *in* Hematite Petroleum Proprietary Ltd, 1972). The clastic unit in these four wells probably correlates with the Williambury and Yindagindy Formations.

STRUCTURE

The carboniferous rocks are exposed in three north-trending belts in the latitude of Williambury. The threefold repetition is interpreted as being due to faulting (Condon, 1954; Teichert, 1957; Thomas, 1962b).

GEOLOGICAL HISTORY

A shallow sea transgressed the northern Carnarvon Basin in the Early Carboniferous. Terrigenous input was small and carbonate rocks (Moogooree Limestone) were deposited. The sea was extremely shallow in the east, and near-shore (at times intertidal) sedimentation took place. Very little clastic sedimentation occurred at first but somewhat later an alluvial fan (Williambury Formation) formed in the Williambury area. Towards the close of the Early Carboniferous, mixed carbonate and clastic deposition was re-established (Yindagindy Formation) with intertidal and terrigenous sedimentation in the east and shallow-water clastics probably extending over most of the northern Carnarvon Basin. The sea withdrew, at least from the east part of the basin, for much of the Late Carboniferous. It returned in the late Late Carboniferous with glacial-marine sedimentation (Lyons Formation) which continued into the Permian.

CANNING BASIN

In the Canning Basin Carboniferous rocks crop out on the Lennard Shelf and occur in the subsurface in the Fitzroy Trough, Kidson Sub-basin and Lennard Shelf (Fig. 6). The Carboniferous sequence consists of the Fairfield Group, which extends into the Upper Devonian, the Anderson Formation, and the Grant Group, which passes up into the Lower Permian (Fig. 7). In parts of the Fitzroy Trough deposition was probably continuous but in most sections there is an unconformity at the base of the Grant Group. The exposed Carboniferous rocks are assigned to the Fairfield Group; the Anderson Formation and the Carboniferous part of the Grant Group do not crop out. The solid geology of the main areas of Carboniferous exposure is shown in Figure 8.

STRATIGRAPHY

Fairfield Group

The Fairfield Group ("Fairfield Formation" of Playford and Lowry, 1966; raised to group status by Druce and Radke, 1979), consisting of limestone, siltstone, shale and sandstone, overlies the Devonian reef complexes (apparently conformably) on the Lennard Shelf or the Luluigui Formation in the Fitzroy Trough and is overlain, probably conformably, by the Anderson Formation. Company seismic data from the Lennard Shelf indicate the possibility of an angular unconformity within or at the top or the Fairfield Group. The unit straddles the Devonian-Carboniferous boundary and comprises the Gumhole Formation at the base, the Yellow Drum Sandstone, and the Laurel Formation. The geology of the group has been summarized by Druce and Radke (1979) who also discuss the complex history of the term "Fairfield".

Gumhole Formation

The Gumhole Formation (Druce and Radke, 1979) consists of limestone, siltstone, shale and sandstone with minor dolomite. The type section is on the Great Northern Highway 19 km west-northwest of Fitzroy Crossing and 1.5 km southeast of Gumhole bore and is 70 m thick. The formation is over 200 m thick in the Fitzroy Trough. The Gumhole Formation conformably overlies the Nullara Limestone in the Horseshoe Range, at Red Bluffs, and in many wells on the Lennard Shelf; elsewhere it probably overlies the Luluigui Formation. The formation crops out in the Fairfield Valley, Oscar Range, Horseshoe Range, and Red Bluffs area, and occurs in the subsurface in the Lennard Shelf and Fitzroy Trough.

The Gumhole Formation contains a rich fauna of brachiopods (Veevers, 1959a), bryozoans (Ross, 1961), corals (Hill and Jell, 1971), condonts (Nicoll and Druce, 1979), ostracods (Jones, *in* Veevers and Wells, 1961) and spores (Balme and Hassell, 1962; Playford, 1976). The conodonts and spores suggest a late Famennian age (Druce and Radke, 1979), do V and do VI according to Nicoll and Druce (1979).

Yellow Drum Sandstone

The Yellow Drum Sandstone (Druce and Radke, 1979) is a unit of calcareous sandstone and silty dolomite with minor shale that lies stratigraphically between the Gumhole Formation and the Laurel Formation, probably conformably although the upper and lower contacts are nowhere exposed. The type section is 1 km northeast of Yellow Drum bore (25 km west-northwest of Fitzroy Crossing) where the formation is about 70 m thick (including 69 m penetrated in BMR Noonkanbah 4 drilled at the type section). The maximum thickness is 327 m in Napier 1 between 333 and 660 m. The Yellow Drum Sandstone outcrops mainly in the area of the type section but occurs extensively in the subsurface of the Lennard Shelf and Fitzroy Trough.

The biota consists predominantly of microfossils-ostracods, conodonts and spores. Conodont evidence suggests that the formation is latest Famennian (do VI, Tnla) to earliest Tournaisian (Tnlb) in age (Nicoll and Druce, 1979). The unit is diachronous, being almost entirely Famennian in Napier 1 and Tournaisian (Tnlb)in Meda 1, both on the Lennard Shelf and straddling the Famennian-Tournaisian boundary in the Fitzroy Trough (Druce and Radke, 1979).

Laurel Formation

The Laurel Formation ("Laurel Beds" of Thomas, 1957; Laurel Formation of Thomas, 1959) consists of interbedded limestone, shale, siltstone, sandstone and minor dolomite.The type area is near Twelve Mile

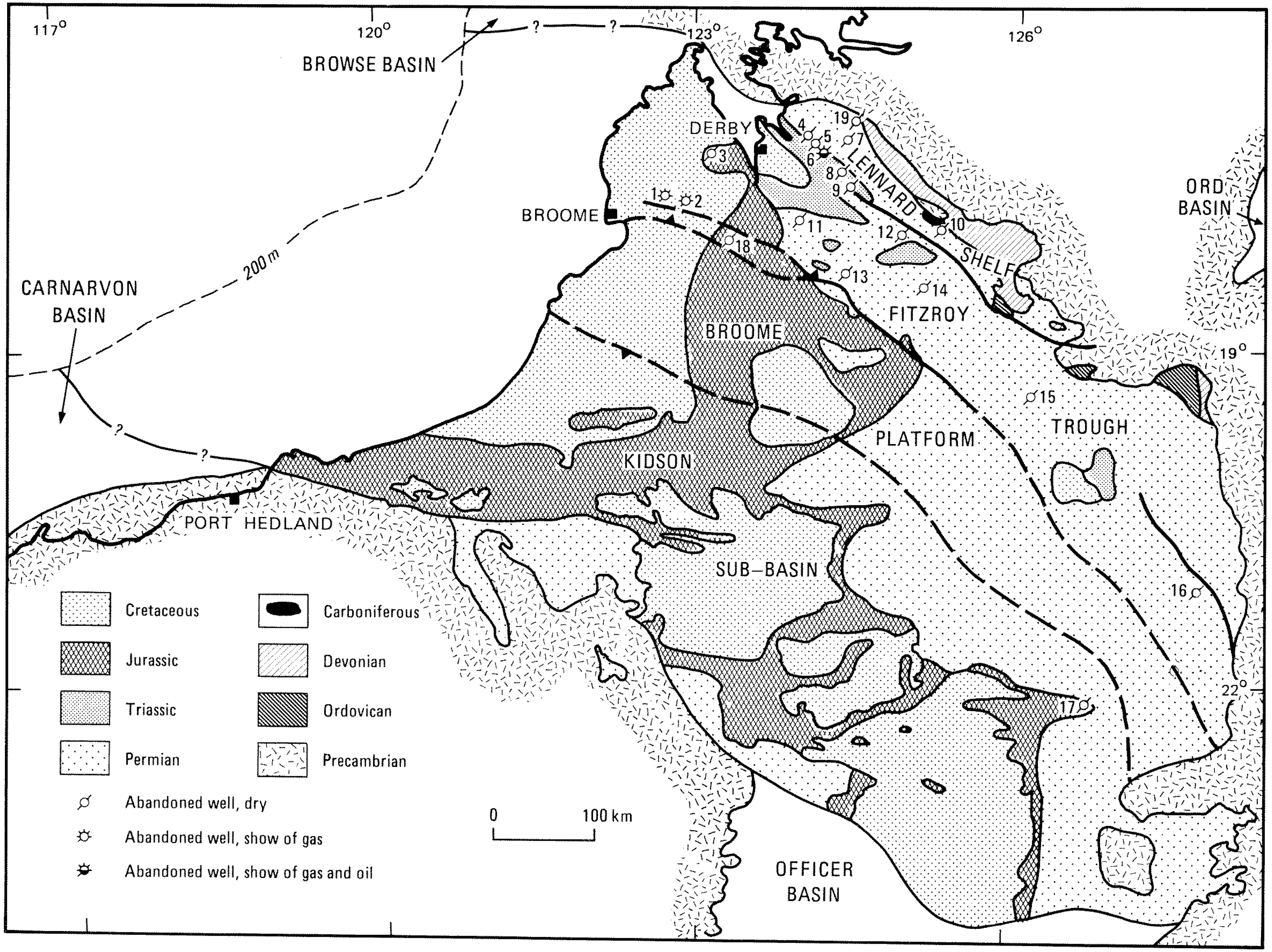

Figure 6. Canning Basin, showing solid geology and position of principal oil-exploration wells encountering Carboniferous rocks (after Playford and others, 1975). 1. Barlee 1; 2. Yulleroo 1; 3. Fraser River 1; 4. May River 1; 5. Langoora 1; 6. Meda 1 and 2; 7. Hawkstone Peak 1; 8. Blackstone 1; 9. Sisters 1; 10. BMR Noonkanbah 2 (formerly Laurel Downs 2); 11. Grant Range 1; 12. Mount Hardman 1; 13. Nerrima 1; 14. St George Range 1; 15. Lake Betty 1; 16. Point Moody 1; 17. Wilson Cliffs 1; 18. Logue 1; 19. Napier 1.

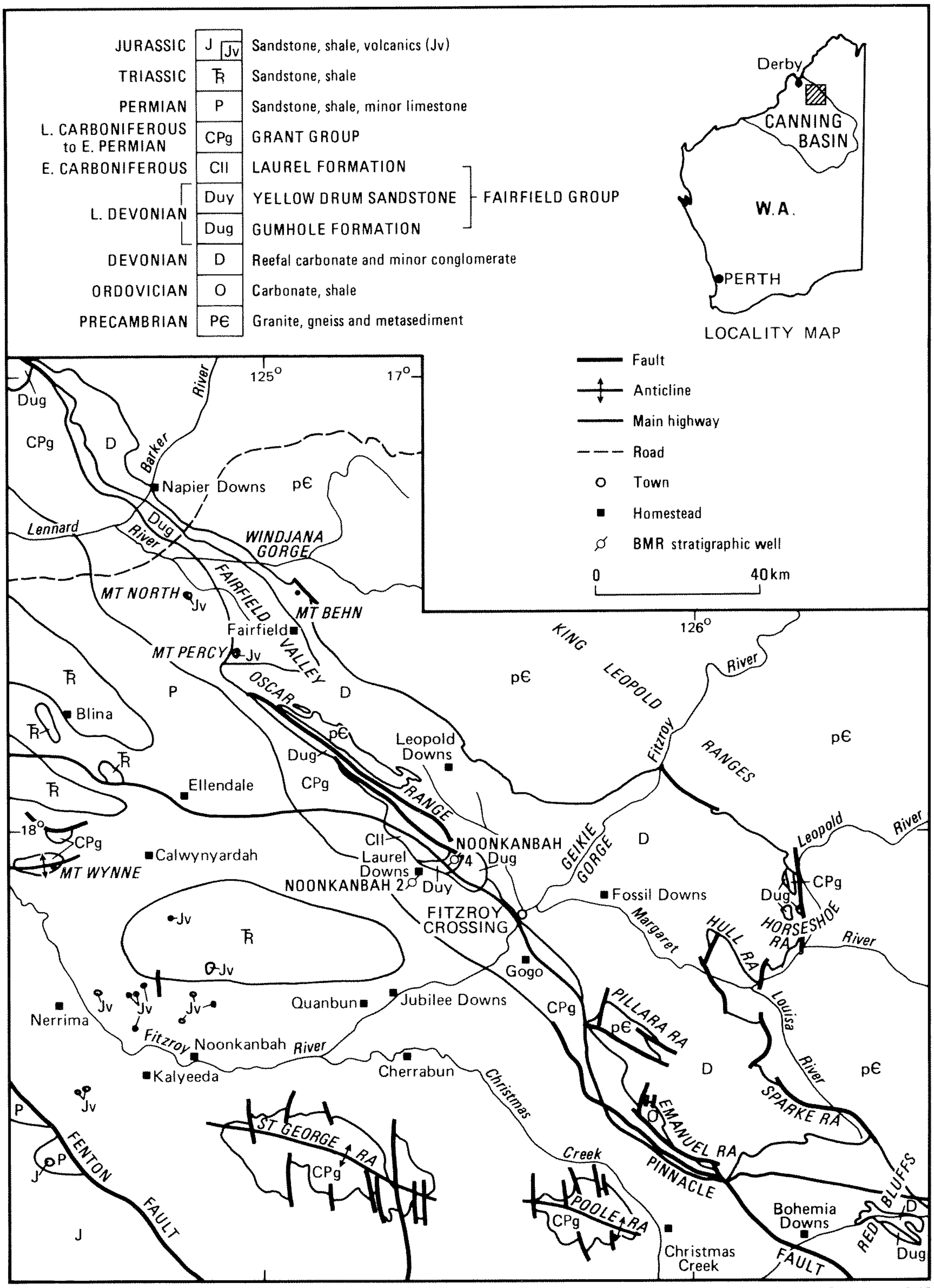

Figure 7. Canning Basin, solid geology of part of the northern Canning Basin (after Playford and Lowry, 1966; Druce and Radke, 1979).

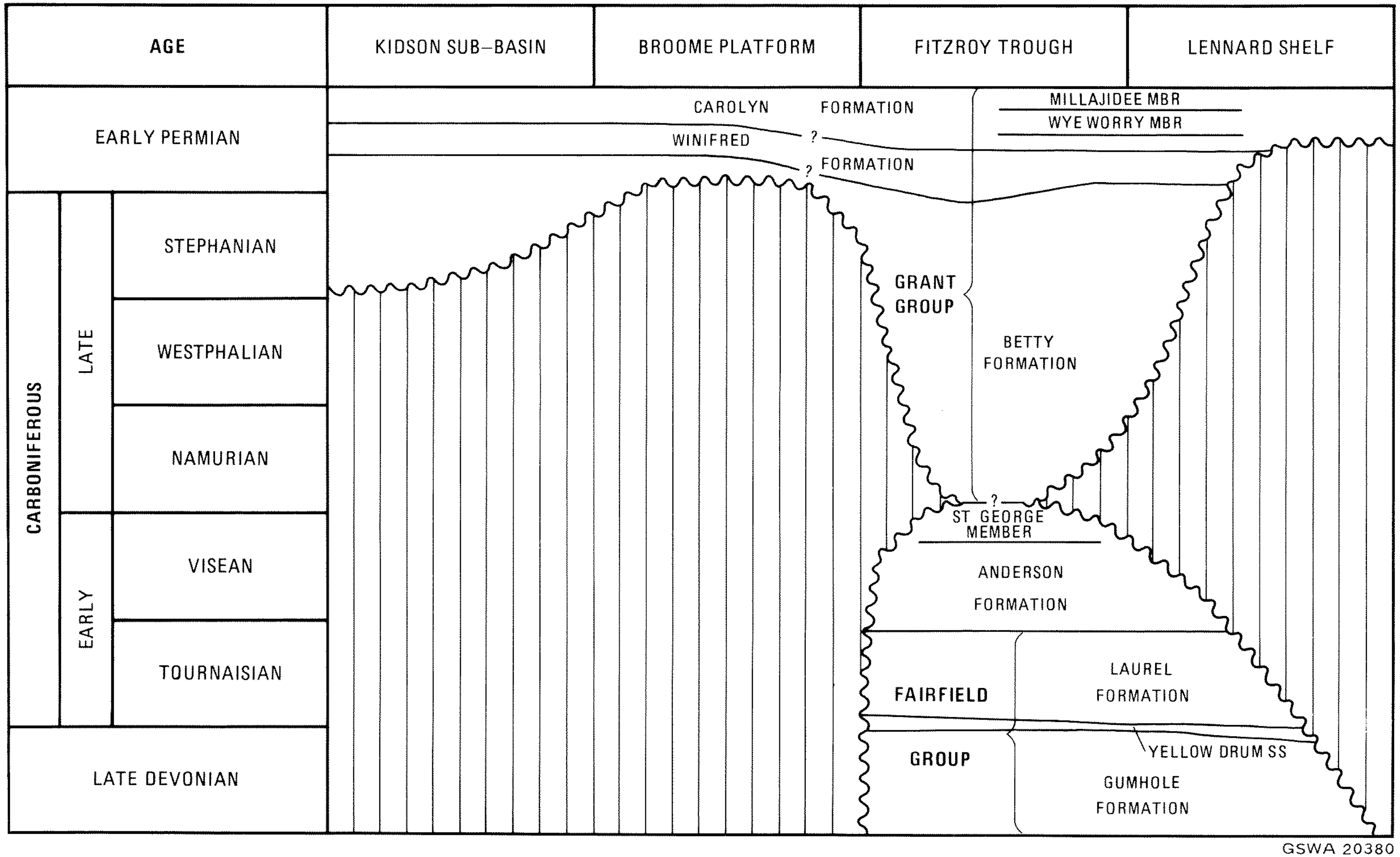

Figure 8. Canning Basin, stratigraphic correlation chart.

Bore, some 45 km northwest of Fitzroy Crossing, and the formation is named after the nearby Laurel Downs pastoral station. Thomas (1959) chose two type section near Twelve Mile Bore and estimated the total thickness to be 455 m. However, Druce and Radke (1979, p.24) measured 386 m but consider the 357 m penetrated in BMR Noonkanbah 2 as a "... more realistic figure than either of the thicknesses measured from surface outcrop". The formation conformably overlies the Yellow Drum Sandstone and is conformably overlain by the Anderson Formation in the subsurface.

The Laurel Formation is predominantly calcareous in the lower part and shaly in the upper part, and on this basis two unnamed members may be recognized.The lower limestone member has its maximum thickness (189 m) in BMR Noonkanbah 2 between 244 and 433 m, and the upper shale member reaches its greatest thickness (667 m) in Lake Betty 1 from 1 808 to 2 475 m. The formation outcrops mainly in front of the Oscar Range and is widespread in the subsurface of the Lennard Shelf and Fitzroy Trough.

Numerous fossils occur in the Laurel Formation including brachiopods (Veevers, 1959b; Thomas, 1971) which suggest a correlation with the *teicherti, aquila, laurelensis, eganensis* and *amnicum* zones of the Bonaparte Basin (Roberts, 1971), conodonts (Nicoll and Druce, 1979), spores of the *Grandispora spiculifera* Assemblage (Playford, 1976; Kemp and others, 1977); bryozoans, corals, bivalves, ostracods and fish also occur (Druce and Radke, 1979). The Laurel Formation conodonts are assigned to the upper part of the *Clydagnathus gilwernensis*, Apparatus A, *Spathognathodus canningensis* and *Bispathodus spinulicostatus* assemblage zones which are Tournaisian (Tnlb to Tn3) in age (Nicoll and Druce, 1979).

Anderson Formation

The Anderson Formation (McWhae and others, 1958) is a subsurface unit of interbedded sandstone, siltstone and shale, with minor limestone, dolomite and anhydrite, which overlies the Fairfield Group, probably conformably, and is overlain, usually unconformably, by the Grant Group. The type section is in Grant Range 1 from 2 408 to 3 936 m (total depth), a thickness of 1 528 m, and the formation is named after Mount Anderson. The formation occurs in the Fitzroy Trough and on the Lennard Shelf.

In Yulleroo 1, the Anderson Formation occurs from 1 060 to 3 409 m, which represents the maximum known thickness (2 349 m). Between 1 060 and 1 871 m Bischoff (1968) recognized a "Unit B". This is a continental unit of multicoloured red and green shale and kaolinitic sandstone with a minor amount of coal, which is present in several wells drilled in the Fitzroy Trough. It was encountered in St George Range 1 between 1 518 and 2 487 m where it was named the "St George Formation" by Shannon and Henderson (1966) and was considered to unconformably overlie the Anderson Formation. However, regionally the base of the unit appears to be conformable and it is best to regard the multicoloured beds as a member of the Anderson Formation confined to the Fitzroy Trough.

The Anderson Formation was deposited under paralic to continental conditions. The fossils include foraminifers, fish remains, bivalves, conchostracans, ostracods, conodonts and plant microfossils. Conodonts in Yulleroo 1 suggest a ?Tournaisian to Visean age for the formation. On palynological evidence, Dolby (*in* Broad and McDermott, 1974) dates samples from the Anderson Formation as ?late Tournaisian to Visean. The bulk of the formation is probably Visean. The lower part may extend into the Tournaisian and interfinger with the Laurel Formation. In Yulleroo 1 the formation is overlain by the Betty Formation ("Unit A" of Bischoff, 1968), the lowermost beds of which contain late Visean to early Namurian conodonts (Bischoff, 1968). Sedimentation may, therefore, have been continuous in the deeper parts of the Fitzroy Trough.

Grant Group

The Grant Group ("Grant Range Beds" of Talbot, *in* Blatchford, 1927 and Woolnough, 1933; "Grant Formation" of Guppy and others, 1958; raised to group status by Crowe and Towner, 1976b) is a unit of sandstone, conglomerate, siltstone, shale, tillite and minor varved shale which overlies the Anderson Formation, usually unconformably, and is overlain with minor disconformity by the Poole Sandstone. The group is Namurian to Sakmarian in age.

The Grant Group is widely distributed through the Canning Basin and has been encountered in most exploratory wells drilled to date. The base is one of the strongest and most widespread unconformities in the basin; and, in different areas, the group rests on Carboniferous, Devonian, Ordovician, or Precambrian rocks. Three units have been recognized in the subsurface in oil-exploration wells. These units were given formation status by Crowe and Towner (1976b) and in Table 1 their nomenclature is correlated with the informal units of other authors.

TABLE 1:
NOMENCLATURE OF UNITS
OF THE GRANT GROUP

Young and O'Shaugnessy, 1973	Crowe and Towner, 1976a	Crowe and Towner, 1976b
"Binda Member"	"Upper Sandstone Unit"	Carolyn Formation
"Dora Shale Member"	"Middle Shale Unit"	Winifred Formation
"Cuncudgerie Sandstone Member"	"Lower Sandstone Unit"	Betty Formation

The Carboniferous-Permian boundary corresponds approximately to the top of the Betty Formation.

Betty Formation

The Betty Formation (Crowe and Towner, 1976b) consists of medium- to coarse-grained sandstone with minor mudstone and conglomerate. It is rarely glauconitic or carbonaceous and frequently contains lithic fragments. The type section is in Lake Betty 1 between 1 058 and 1 657 m where it is conformably overlain by the Winifred Formation and disconformably overlies the Laurel Formation. It is thickest in the Fitzroy Trough where it reaches 1 713 m in Grant Range 1, and occurs throughout the Kidson Sub-basin. Towards the margin of the Canning Basin the Betty Formation is overlapped by higher units of the Grant Group and is not known to crop out. The sequence is probably glacial-marine in origin, at least in part.

Marine fossil fragments occur in Wilson Cliffs 1 (Creevey, 1969); and conodonts, ostracods, and foraminifers occur in Yulleroo 1 in "Carboniferous Unit A", which is here placed in the Betty Formation (Bischoff, 1968). Elsewhere, only palynomorphs are recorded from the formation. In the Fitzroy Trough, the base of the formation contains late Visean to early Namurian conodonts in Yulleroo 1 (Bischoff, 1968); and the Namurian *Grandispora maculosa* Assemblage is present in Fraser River 1 and Grant Range 1 (Playford and Powis, 1979). The oldest Betty Formation on the Lennard Shelf is in Blackstone 1 where the *Spelaeotriletes ybertii* Assemblage is recorded (Playford and Powis, 1979). On the Broome Platform (e.g. McLarty 1, Crosslands 2), Unit I is absent; and the oldest rocks are Unit II in age (Powis, 1979). The upper limit of the Betty Formation is diachronous. In Lake Betty 1 (Williams and Dolby, *in* Crank, 1972), Blackstone 1 (Balme, *in* Johnson, 1968) and Kidson 1 (Powis, 1979), the highest strata contain a Unit II microflora. In Logue 1, a sample just above the base of the overlying Winifred Formation is assigned to Unit I by Dolby (*in* Meath and Scott, 1972). Hence the top of the Betty Formation crosses the Carboniferous-Permain boundary.

Winifred Formation

The Winifred Formation (Crowe and Towner, 1976b) is a sequence of dark-grey mudstone with minor fine-grained sandstone and lenses of coal and limestone. The type section is at lat. 22°52′40″S, long.123°36′20″E near Lake Winifred and is 19 m thick. The base is not exposed and the unit is unconformably overlain by the (Cretaceous) Anketell Sandstone. At a nearby outcrop the formation lies conformably between the Paterson Formation below and the Carolyn Formation above. Crowe and Towner (1976b) nominated a subsurface reference section in Kidson 1 between 815 and 1 070 m. The formation is widespread through the basin and is usually readily recognizable in the subsurface.

Unidentified bryozoan, crinoid and echinoid fragments occur in the formation in Sahara 1 (Singleton, 1965) and trace fossils are found at the type section (Crowe and Towner, 1976b). Spores and pollen from oil-exploration wells indicate that the formation contains microfloral assemblages from Units I, II, and III. In the central part of the Fitzroy Trough (Logue 1), the formation correlates with Units I and II (Dolby *in* Meath and Scott, 1972). In other parts of the Trough, for example Lake Betty 1 (Williams and Dolby *in* Crank, 1972), and on the Lennard Shelf, for example Blackstone 1 (Balme *in* Johnson, 1968), only Unit II microfloras are recorded. However, in the Kidson Sub-basin (Kidson 1) and on the Broome Platform (McLarty 1, Crossland 2) the Winifred Formation contains Units II and III microfloral assemblages (Powis, 1979) and is younger than in the Fitzroy Trough and Lennard Shelf. It is not clear whether the Winifred Formation in the Fitzroy Trough and Lennard Shelf is the same mappable unit as that in the Kidson Sub-basin and Broome Platform or is a different, slightly older unit. For the present the formation is retained as a diachronous unit.

Carolyn Formation

The Carolyn Formation (Crowe and Towner, 1976b) is a unit of meduim-grained quartz arenite and quartz wacke with some shale and conglomerate. The type section is at the western end of the St George Range near Carolyn Valley and consists of two sections with an aggregate thickness of about 140 m. The formation is overlain uncomformably by the Poole Sandstone and rests conformably on the Winifred Formation, although the contact is nowhere exposed. The maximum thickness is 415 m in Logue 1.

The Carolyn Formation occurs throughout the Canning Basin except where removed by erosion. It is the only unit of the Grant Group that outcrops on the northern and eastern parts of the basin and all the rocks in this region formerly mapped as "Grant Formation" are now placed in the Carolyn Formation.

Crowe and Towner (1976a) have recognized two members at the top of the formation, the Wye Worry Member of tillitic conglomerate and siltstone and claystone, sometimes varved, and the overlying Millajiddee Member which is mainly sandstone.

Marine shells of Sakmarian age, together with trace fossils, occur in the Wye Worry Member (Crowe and Towner, 1976a; Dickins and others, 1977). Plant fragments including *Glossopteris* are known, and palynomorphs of Units II and III have been recovered from subsurface samples, for example in Blackstone 1 (Balme *in* Johnson, 1968). On the basis of these plant microfossils the formation is dated as Sakmarian.

Sandstone at Red Bluffs

A sandstone unit in the Red Bluffs area (referred to as the Bohemia Downs area by Playford and others, 1975) unconformably overlies the Gumhole Formation and the Nullara Limestone and was mapped as Permian "Grant Formation" by Playford and Lowry (1966). However, Veevers and others (1967) believe that it is of Early Carboniferous age, based on its stratigraphic position above Famennian rocks and on the occurrence of *Leptophloem australe*, which is supposed to be restricted to the Famennian and Tournaisian (Hill and Woods, 1964). B.E. Balme (pers. comm., 1981) points out that in South Africa the species is known from the Dwyka Tillite (*see* Plumstead, 1969 *for details*) which Truswell (1980) correlates with Stage 2 of the eastern Australian palynological zonation; this broadly equates with Unit II which is Early Permian. It is possible that the unit should be correlated with the Anderson Formation but in mapping, it has not been possible to separate the outcrops from those included in the Carolyn Formation. Provisionally, the Red Bluffs outcrops are considered to belong to the Grant Group.

STRUCTURE

The Canning Basin is not strongly faulted and the principal faults are those bounding the Fitzroy Trough. These are the Pinnacle Fault System on the northeast and the Fenton Fault on the southwest, both of which have a throw of up to 6 000 m. The Carboniferous rocks outside the Fitzroy Trough are comparatively undeformed. In the trough, they are folded into a series of en echelon anticlines trending at about 100°—Mt Wynne, St George Range and Poole Range Anticlines are examples of these folds. Most of the anticlines are cut by north-trending normal faults which are usually small, although some have displacements of as much as 300 m. The folds and their associated faults are believed to have resulted from regional right-lateral wrenching movements (Rattigan, 1967; Smith,1968) and the main period of folding may have been in the Jurassic.

GEOLOGICAL HISTORY

A considerable thickness of sediments was deposited in the Fitzroy Trough where there was contemporaneous movement along the bounding faults throughout the Carboniferous. Marine sedimentation, probably under tropical conditions, continued from the Late Devonian into the Early Carboniferous in both the Fitzroy Trough and Lennard Shelf (Fairfield Group) with a deepening of the trough in the early Tournaisian. Shallowing took place in the Visean with conditions gradually changing to paralic (lower part of Anderson Formation) and continental (St George Member). Deposition probably continued into the Namurian in the trough; and estuarine and intermittent marine conditions were re-established (lowest part of Betty Formation). On the Lennard Shelf and Broome Platform, erosion took place. The trough continued to subside, and sedimentation occurred on the edge of the Lennard Shelf and spread over the Kidson Sub-basin. Glacial conditions were established, both marine and continental (Betty Formation) in the Late Carboniferous, probably waned in the earliest Permian (Winifred Formation), and returned in the late Early Permian (Carolyn Formation).

BONAPARTE BASIN

Carboniferous rocks are exposed in the southeast and centre of the onshore part of the Bonaparte Basin in Western Australia (Figs 9, 10). The basin consists of a northern basinal area and a southern marginal shelf area which is divided into eastern and western parts by the Precambrian Pincombe Range Inlier. The basinal sequence consists (in ascending order) of the Bonaparte, Tanmurra and Keep Inlet Formations. On the western shelf the sequence includes the Milligans Formation, Utting Calcarenite, Waggon Creek Formation, Burvill Formation, Point Spring Sandstone and Keep Inlet Formation. The most complete Carboniferous section is on the eastern shelf and comprises (from the Base) the Burt Range Formation, Enga Sandstone, Septimus Limestone, Zimmerman Sandstone, Milligans Formation, Burvill Formation, Point Spring Sandstone and Keep Inlet Formation (Fig. 11). A solid geology sketch map is given in Figure 10. The total thickness of Carboniferous sediments in the basin exceeds 2 000 m.

STRATIGRAPHY

Burt Range Formation

The Burt Range Formation ("Burt Range Series" of Matheson and Teichert, 1948; "Burt Range Limestone" of Noakes and others, 1952, amended Veevers and Roberts, 1968) consists of olive-grey to olive-brown calcarenite grading to sandy calcarenite and sandstone in the upper part. It crops out on the shelf area east of the Pincombe Range inlier where it

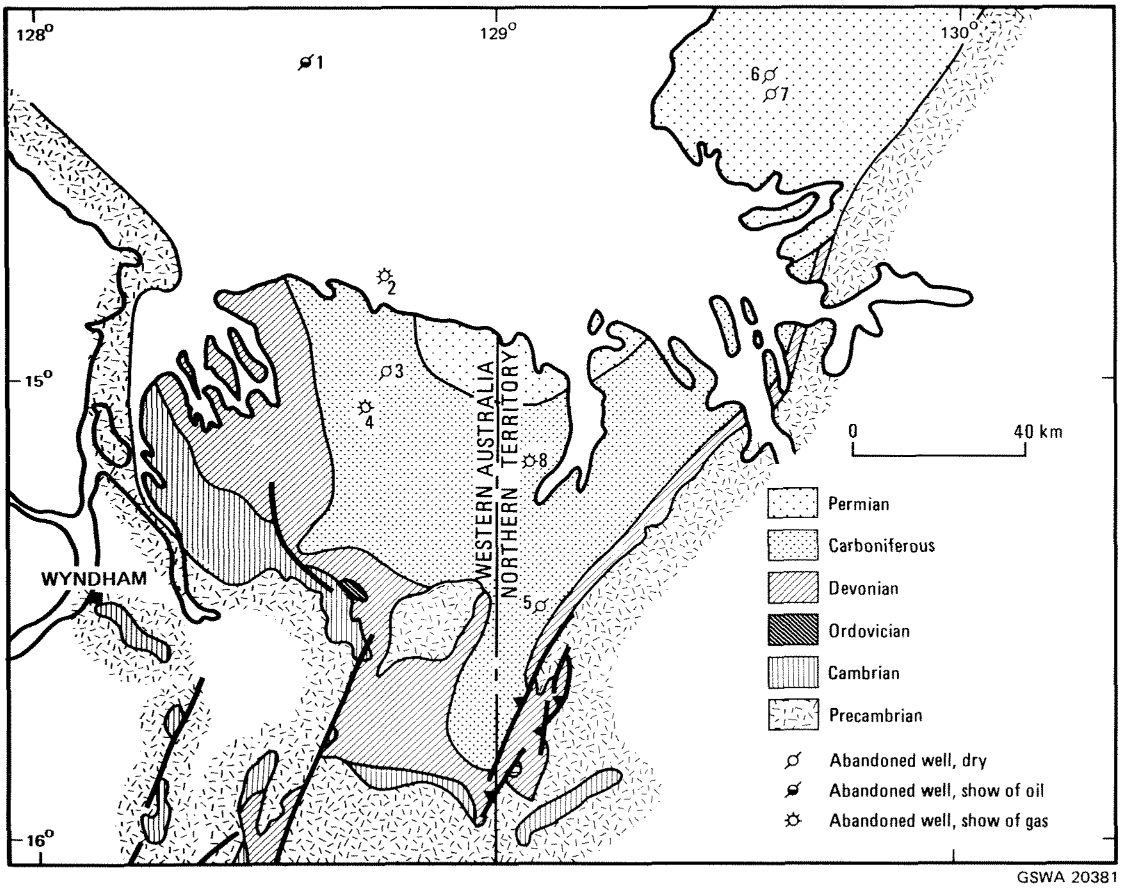

Figure 9. Bonaparte Basin, showing solid geology and position of principal oil-exploration wells encountering Carboniferous rocks (after Playford and others., 1975). 1. Lacrosse 1; 2. Pelican island 1; 3. Bonaparte 1; 4. Bonaparte 2; 5. Spirit Hill 1; 6. Kulshill 1; 7. Kulshill 2; 8. Keep River 1.

overlies the Buttons Formation disconformably and is overlain conformably by the Enga Sandstone or unconformably by the Keep Inlet Formation. The type section is situated 6 km west-southwest of Milligans Hills where the formation is 290 m thick with the top not exposed. The total thickness of the unit is believed to be about 460 m. The "Spirit Hill Limestone" of Traves (1955) was included in the Burt Range Formation by Veevers and Roberts (1968).

The formation is dated as early to middle Tournaisian, based on conodonts (Druce, 1969). Brachiopods (Roberts, 1971; Thomas, 1971) are abundant in the formation and Roberts (1971) recognized the *adunata, teicherti, aquila*, and *laurelensis* zones and part of the *eganensis* zone in the unit.

Enga Sandstone

The Enga Sandstone (Traves, 1955) is a unit of quartz sandstone with minor interbedded calcarenite, which lies conformably between the Burt Range Formation below and the Septimus Limestone above. The type section is near the middle of Enga Ridge and is 148 m thick. The top of the formation is not exposed in this section and its total thickness is estimated to be 158 m. The unit crops out at Burt Range and in the area immediately to the north.

The Enga Sandstone is dated as Tournaisian on the basis of the conodont fauna (Druce, 1969). Brachiopods (Roberts, 1971; Thomas, 1971), of the *eganensis* and *amnicum* zones, foraminifers (Mamet and Belford, 1968), gastropods and bivalves (Thomas, 1962a), bryozoans and trilobites also occur.

Septimus Limestone

The Septimus Limestone ("Mount Septimus Limestone" of Noakes and others, 1952, emended Traves, 1955) is a unit of olive-grey to brown calcarenite, sandy calcarenite and sandstone. The type section, 180 m thick, is on the northwestern flank of Mount Septimus. The formation, which crops out in the Mount Septimus—Spirit Hill area, overlies the Enga Sandstone with apparent conformity in the type area, although a slight disconformity may exist between them in some localities. The Septimus Limestone is overlain conformably by the Zimmermann Sandstone or unconformably by the Keep Inlet Formation.

The unit contains conodonts which indicate a Tournaisian age (Druce, 1969). Thomas (1962a, 1971) and Roberts (1971) have recognized a number of brachiopod assemblages in the formation. Thomas's (1962a) four faunal assemblages are now placed in the *australis* and *langfieldensis* zones and the lower part of the *spiritus* zone by Roberts (1971).

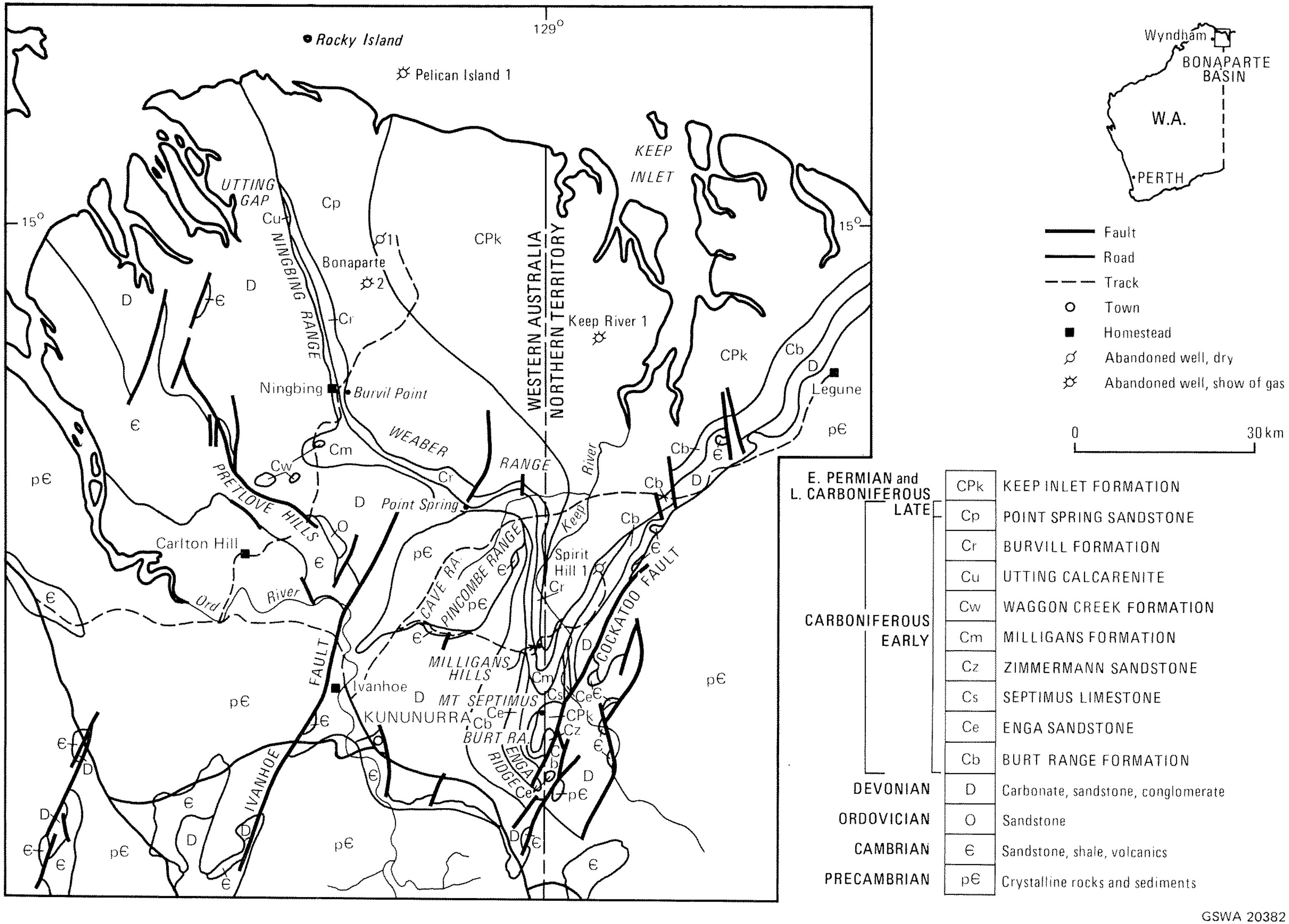

Figure 10. Bonaparte Basin, solid geology of the Carboniferous (after Veevers and Roberts, 1968; A. J. Mory and G. M. Beere pers. comm. 1982).

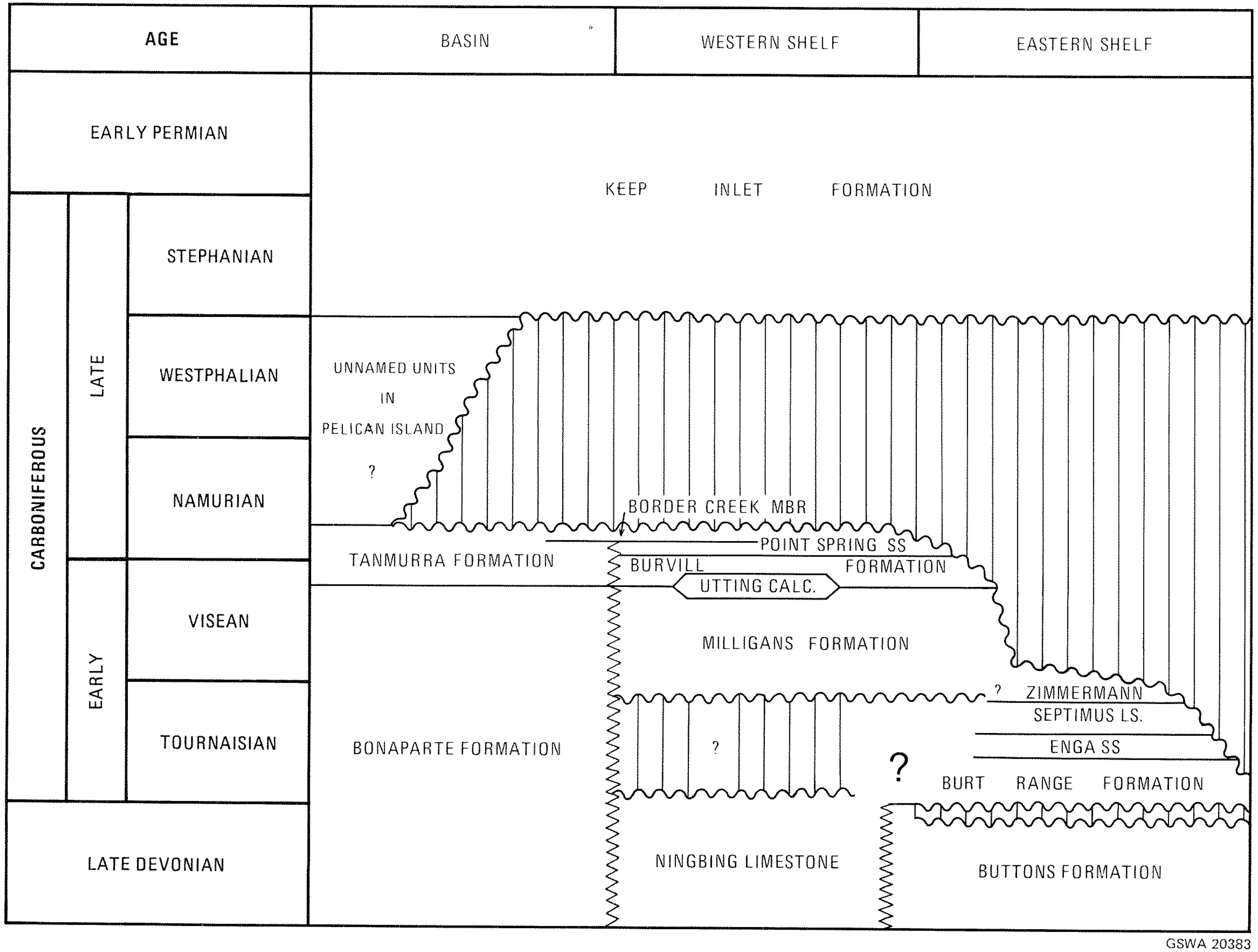

Figure 11. Bonaparte Basin, stratigraphic correlation chart.

Zimmermann Sandstone

The Zimmermann Sandstone (Veevers and Roberts, 1968) is a unit of brown to white quartz sandstone which overlies the Septimus Limestone and is overlain disconformably by the Keep Inlet Formation. The type section is at Mount Zimmermann and is 140 m thick. The formation is known only from the southern Burt Range area

The Zimmermann Sandstone is dated as late Tournaisian to early Visean by Veevers and Roberts (1968) on the basis of the brachiopod fauna, which is assigned to the *spiritus* zone (Roberts, 1971). The fauna also includes bivalves, bryozoans, and the trace fossil, *Rhizocorallium*."

Milligans Formation

The Milligans Formation ("Milligans Beds" of Thomas, in Hare and Associates, 1961) is a unit of silty shale and siltstone which rests uncomfortably on the Burt Range Formation (in Spirit Hill 1) and is overlain probably conformably by the Burvill Formation or unconformably by the Keep Inlet Formation in the Spirit Hill area. The unit is poorly exposed around the basinward edge of the shelf area and occurs in the subsurface in the Spirit Hill and Milligans wells. The formation is a lateral equivalent of the upper part of the Bonaparte Formation. The type section is in Milligans 1 between 44 and 155 m (base not reached). In Spirit Hill 1, the unit is 252 m thick and the top is eroded. On the western shelf the formation is present (conformably?) below the Burvill Formation and (unconformably?) above the Ningbing Limestone.

Fossils in the Milligans Formation include ostracods, foraminifers, rare conodonts, echinoid fragments, bryozoans, fish scales and plant spores. The spores belong to the *Anapiculatisporites largus* assemblage (Playford, 1971: Kemp and others, 1977). The microfossils in the formation in Milligans 2 also occur in core 9 in Bonaparte 1 which is of Visean age (Veevers and Roberts, 1968); conodonts from the base of the unit are late Tournaisian in age (P. Jones and R. Nicoll, pers. comm., 1981).

Utting Calcarenite

The Utting Calcarenite (Veevers and Roberts, 1968 is a unit of grey and yellow-brown calcarenite and shale which is conformably overlain by the Burvill Formation. The nature of the contact with the adjacent Ningbing Limestone is indefinite but is probably faulted. The type section (61 m thick) and principal exposures are at Utting Gap and the total thickness of the formation is probably about 120 m. The unit is of very limited extent and is laterally equivalent to the upper part of the Milligans Formation and possibly to the lower part of the Burvill Formation.

The Utting Calcarenite contains a rich fauna of conodonts (Druce, 1969), foraminifers (Mamet and Belford, 1968), brachiopods of the *pauciplicatus* zone (Roberts, 1971), corals, echinoids, trilobites, ostracods and sharks. The foraminifers suggest an early Late Visean age for the unit (Mamet and Belford, 1968).

Burvill Formation

The Burvill Formation ("Burvill Beds" of Veevers and Roberts, 1968) is a sequence of sandstone, shale and interbedded sandy limestone that lies conformably between the Utting Calcarenite or Milligans Formation below and the Point Spring Sandstone above. The formation is exposed along the Weaber Range and at Milligans Hills. The type section is at Burvill Point 1, ½ km east of Ningbing homestead, and is 40 m thick. The thickest measured section is 85 m thick.

The formation contains a rich fossil fauna of brachiopods, gastropods, foraminifers and conodonts. The brachiopods belong to the *milliganensis-gradatus* zone which is of late Visean to early Namurian age. The conodont fauna suggests the same age (Druce, 1969) but Mamet and Belford (1968) regard the foraminifer fauna as probably latest Visean. Jones and Nicoll (pers. comm., 1981) state that the conodont faunas from the type Utting Calcarenite and from the Burvill Formation near Point Spring are time-equivalent. This could suggest that the Burvill Formation is slightly older near Point Spring than at Utting Gap. Certainly from their stratigraphic position the two formations are very close in age and there seems to be no good reason for the interzone which Roberts (1971) shows between the *pauciplicatus* and the *milliganensis-gradatus* zones.

Waggon Creek formation

The Waggon Creek Formation (Veevers and Roberts, 1968; emended Beere, 1984) consists of a lower unit of conglomerate (containing dolomite blocks), pebbly sandstone, and sandstone, overlain by a unit of sandstone and shale. The formation rests with angular unconformity on the Devonian Cockatoo Sandstone. Veevers and Roberts (1968) mapped the upper unit as Point Spring Sandstone but Beere (1984) considers it to be part of the Waggon Creek Formation. The unit is known only from the Waggon Creek area on the north side of the Pretlove Hills and is 120 m thick. It contains brachiopods of the *milliganensis-gradatus* zone and correlates with the Burvill Formation and Point Spring Sandstone.

In the same area, Veevers and Roberts (1968) recognize an unnamed breccia having the same lithology and stratigraphic position as the Waggon Creek Formation. Roberts (1971) records an *amnicum* zone brachiopod fauna from these beds and Druce (1969) found abraded conodonts referred to *Clydagnathus cavusformis*, both of which suggest a mid-Tournaisian age. The relationship of this breccia to the Waggon Creek Formation is not clear. They may be different units as stated by Veevers and Roberts (1968) or they may be the same unit of late Visean to early Namurian age with reworked Tournaisian fossils as proposed by Beere (1984).

Point Spring Sandstone

The Point Spring Sandstone (Noakes and others, 1952; emended Veevers and Roberts, 1968; emended herein) is a unit of quartz sandstone, pebbly quartz sandstone, conglomerate and siltstone which conformably overlies the Burvill Formation and is overlain disconformably by the Keep Inlet Formation. The type section is 6.4 km northeast of Point Spring in the Weaber Range and is 236 m thick. The maximum known thickness is 383 m.

As emended, the formation includes the "Border Creek Formation" of Veevers and Roberts (1968) which is here considered a member of the Point Spring Sandstone and Tanmurra Formation. These authors considered that the base of the Border Creek Member was a major disconformity. However, recent mapping (Beere and Mory, pers. comm., 1981) suggests that the Border Creek Member is the conglomeratic upper part of the Point Spring Sandstone with local channelling at its base which is diachronous. Veevers and Roberts (1968) have already pointed out the similarity between the quartz sandstone of the two units and the rapid lateral lithological change in the Point Spring Sandstone. The formation occurs only in the Weaber Range area and is known in the subsurface in Bonaparte 2. Outcrops mapped as Border Creek Formation by Veevers and Roberts (1968) in the Burt Range and Spirit Hill areas are now considered to be Keep Inlet Formation (Beere, pers. comm., 1981).

Fossils in the Point Spring Sandstone include brachiopods, bivalves, trace fossils, plant remains and spores and pollen. The brachiopods belong to the *milliganensis-gradatus* zone and the fauna extends up to the base of the Border Creek Member which contains only *Phyllotheca*-like plants (Veevers and Roberts, 1968). The Border Creek Member in Bonaparte 2 contains an *Anapiculatisporites largus* Assemblage microflora (Kemp and others, 1977). The fossil evidence suggests that the formation is late Visean to early Namurian in age.

The Point Spring Sandstone and Burvill Formation correlate with the Tanmurra Formation which is the more open-marine basinal equivalent of these two shelf units. The Border Creek Member is probably a fluvial deposit and does not extend very far into the basin.

Bonaparte Formation

The Bonaparte Formation ("Bonaparte Beds" of Veevers and Roberts, 1968) is a subsurface unit of dark shale, siltstone and sandstone of Late Devonian and Early Carboniferous age which underlies the Tanmurra Formation in wells drilled in the northern part of the onshore part of the Bonaparte Basin. The type section is in Bonaparte 1 between 497 and 3 209 m (total depth), a thickness of 2 712 m, although the base of the formation had not been reached at the total depth of the well.

The Bonaparte Formation is the basinal equivalent of the Devonian and Early Carboniferous carbonates and sandstones deposited on the marginal shelf of the basin. In Bonaparte 1 the highest Devonian sample is core 28 at 2 533 m which contains the bivalve *Buchiola*. The overlying Early Carboniferous section is dated as Tournaisian and Visean on the basis of foraminifers (Mamet and Belford, 1968). Ostracods, brachiopods of the *milliganensis-gradatus* zone, corals and bryozoans also occur (Belford, Jones and Roberts, *in* Le Blanc, 1964). Playford (1971) has described spores, mainly from the upper part of the formation. They belong to the *Anapiculatisporites largus* Assemblage (Kemp and others, 1977).

Tanmurra Formation

The Tanmurra Formation (Le Blanc, 1964; Veevers and Roberts, 1968) is a subsurface unit of limestone, dolomite and sandstone which conformably overlies the Bonaparte Formation and is overlain conformably by an unnamed unit in Pelican Island 1 or disconformably by the Keep Inlet Formation in Bonaparte 1. The type section is from 194 to 497 m in Bonaparte 1.

The Tanmurra Formation is dated as late Visean to early Namurian in age on foraminifers, the boundary between the Lower and Upper Carboniferous occurring between 204 and 207 m in the type section (Mamet and Belford, 1968). The type section has yielded brachiopods of the *milliganensis-gradatus* zone (Roberts, 1971). Spores of the *Anapiculatisporites largus* Assemblage occur in the formation (Kemp and others, 1977).

Keep Inlet Formation

The Keep Inlet Formation ("Keep Inlet Beds" of Glover and others, 1955; Veevers and Roberts, 1968) is a sequence of calcareous sandstone, lithic quartz sandstone, siltstone and conglomerate. The type locality is 4 km northeast of Cleanskin bore, which is west of Keep Inlet. Exposures are poor in this area, and the stratigraphic relationships have been determined from borehole information. The maximum known thickness is 176 m in Bonaparte 1, where the formation rests disconformably on the Tanmurra Formation. On the shelf in the Burt Range area the formation (formerly mapped as "Border Creek Formation") oversteps all units down to the Septimus Limestone. The conglomerate in the type section contains pebbles and boulders of various rock types (Veevers and Roberts, 1968) and the unit is considered to be glacial-marine in origin.

The only macrofossil from the Keep Inlet Formation is one specimen of *Strophalosia* sp. (Veevers and Roberts, 1968). A Unit I microfloral assemblage occurs in the formation in Bonaparte 1 and Pelican Island 1 (where the unit was named "Border Creek Formation") and the age is Late Carboniferous (Kemp and others, 1977).

The Keep Inlet Formation occurs in Keep River 1 and Kulshill 1, where it was called the Kulshill Formation. In Kulshill 1, it extends into the Early Permian and is conformably overlain by the Sugarloaf Formation (Kemp and others, 1977). According to Laws and Brown (1976) the Kulshill Formation is overlain by the Fossil Head Formation. All three formations (Kulshill, Sugarloaf and Fossil Head) are placed in the Port Keats Group and clearly there is a need to clarify the stratigraphic nomenclature of these units.

CARBONIFEROUS STRATA IN THE OFFSHORE BONAPARTE BASIN.

Kemp and others, (1977) correlate, a number of stratigraphic units belonging to the *Spelaeotriletes ybertii* and *Grandispora maculosa* Assemblages in the offshore Bonaparte Basin. In Pelican Island 1 the strata are unnamed interbedded sandstone and shale with minor limestone lying, probably conformably, between the overlying Keep Inlet Formation and the underlying Tanmurra Formation. In Lacrosse 1 they are unnamed sandstone and siltstone, and the "Medusa Beds" of limestone and calcareous sandstone. In Kulshill 1 the strata were named "Tanmurra Formation" and "Bonaparte Formation" but these assignments are questionable as both these formations are older.

While the relationship of these units is not clear and their nomenclature has not been formalized, they represent (at least in Pelican Island 1) a sequence deposited during the time when erosion took place further south. The unconformity between the Keep Inlet Formation and the Tanmurra Formation (seen in Bonaparte 1) disappears to the north where deposition in the basin was continuous throughout the Late Carboniferous and into the Early Permian.

STRUCTURE

The Carboniferous rocks in the Bonaparte Basin crop out as basinward-dipping strata except on the eastern shelf in the Burt Range and Spirit Hill areas wher they are gently folded along north- to northeast-trending axes. Faulting is common in the southeast part of the basin where the Cockatoo Fault system occurs.

GEOLOGICAL HISTORY

Deposition of fine-grained sediments (Bonaparte Formation) continued uninterruptedly from Late Devonian into Early Carboniferous in the basinal part of the Bonaparte Basin. Reef development (Ningbing Limestone) possibly continued into the Tournaisian on the western shelf; on the eastern shelf there may have been a short break in sedimentation at the end of the Devonian (Buttons Formation—Burt Range Formation boundary). A shallow sea covered the eastern shelf in the Tournaisian and Visean, and carbonate (Burt Range Formation, Septimus Limestone) and clastic (Enga and Zimmermann Sandstones) deposition took place. The sea may have withdrawn from the western shelf in the mid-Tournaisian and returned depositing shales (Milligans Formation) which spread on to the eastern shelf. Gradual shallowing of the shelf area (both east and west) occurred with the deposition of shallow near-shore sand (Burvill Formation, Point Spring Sandstone) culminating in fluvial sand and pebbles (Border Creek Member). Local areas of carbonate (Utting Calcarenite) and breccia (Waggon Creek Formation) occurred on the western shelf. Uplift and erosion took place on the shelf in early Namurian time. Further offshore, in the basinal area, sedimentation was continuous throughout the Carboniferous (Bonaparte and Tanmurra Formations, unnamed units in Pelican Island 1). In the Late Carboniferous glacial-marine sedimentation (Keep Inlet Formation) extended over the whole basin as the sea spread on to the shelf area once more.

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PALAEOZOIC STRATIGRAPHY OF THE ORD BASIN, WESTERN AUSTRALIA AND NORTHERN TERRITORY

by A. J. Mory and G. M. Beere

ABSTRACT

The stratigraphy of the Ord Basin is reviewed in the light of field work carried out in 1982. Three major subdivisions of the Palaeozoic sequence are made, based on lithological changes and differences in age. In ascending order these are: the Antrim Plateau Volcanics-?Early Cambrian; the Goose Hole Group (new name)-Cambrian; and the Mahony Group (new name)-?Devonian.

The Goose Hole Group incorporates all sediments considered to be Cambrian in age and consists of the Negri Subgroup and Elder Subgroup. In the Negri Subgroup (formerly Negri Group) the Panton Formation is expanded to include the Hudson Formation. The Elder Subgroup comprises the Eagle Hawk and Overland Sandstones (new names) which, together with the overlying Mahony Group, were previously included in the ?Devonian Elder Sandstone.

The Mahony Group is considered to be Devonian in age, based on a lithological correlation with the Cockatoo Formation in the Bonaparte Basin. The group is divided into the Glass Hill Sandstone, Buchanan Sandstone and Boll Conglomerate (new names). The latter two units are, in part, laterally equivalent to the former.

INTRODUCTION

The Ord Basin was the name given by Gentilli and Fairbridge (1951) to the physiographic unit encompassing the area dissected by the middle course of the Ord River and its tributaries. Subsequently, McWhae and others (1958) defined the Ord Basin, in a geological context, as the area underlain by Cambrian volcanics and clastics south of the Bonaparte Basin.

This paper presents a summary of the Palaeozoic stratigraphy of the Ord Basin in the East Kimberley region of Western Australia and in the Northern Territory (Figs 1 and 2) based on field work carried out in 1982. The depositional environments of the units described in this paper will be discussed on a Geological Survey of Western Australia report on the Bonaparte and Ord Basins presently in preparation. The Ord Basin lies within the following 1:250 000 geological sheets: Dixon Range, Gordon Downs, and Lissadell (W.A.), and Limbunya and Waterloo (N.T.)

PREVIOUS INVESTIGATIONS

The first geological investigation in the basin was that of Hardman (1885). Although his geological map is reasonably accurate, especially considering the reconnaisance nature of his work, Hardman recognized only two units above the Antrim Plateau Volcanics and this makes correlation with modern units difficult (Fig. 3).

Subsequent surveys by Jack (1906), Blatchford (1922, 1927), Mahony (1922) and Wade (1924) relied substantially on Hardman's work. Mahony (1922) proposed the "Negri series", and "Mount Elder sandstones" and through his fossil collection the Cambrian age of the "Negri series" first suggested by Foord (1890), was confirmed by Chapman (1924).

Matheson and Teichert (1948) concentrated on the basaltic rocks (Antrim Plateau Volcanics) and "Negri Series" in the Ord Basin. The seven informal units they proposed for the "Negri Series" formed the basis of the formal nomenclature of Traves (1955).

The results of Bureau of Mineral Resources (BMR) mapping of the Western Australian portion of the Ord Basin in 1962/63 have so far been published only in map form (Dow and Gemuts, 1967; Plumb, 1968). The unpublished account in Dow and others (1964) has recently been revised in another unpublished BMR record (Dow, 1980). The results of BMR mapping in the Northern Territory portion of the Ord Basin were reported in Sweet and others (1974). Differences in the interpretation of the sequence across the N.T./W.A. border by these two parties have been resolved by the work reported upon herein and are summarized in Figure 3 (columns 6-8).

Reviews of the Ord Basin sequence by McWhae and others, (1958), Playford and others (1975) and Jones (1976) are based on Traves (1955) and the unpublished work of Dow and others (1964).

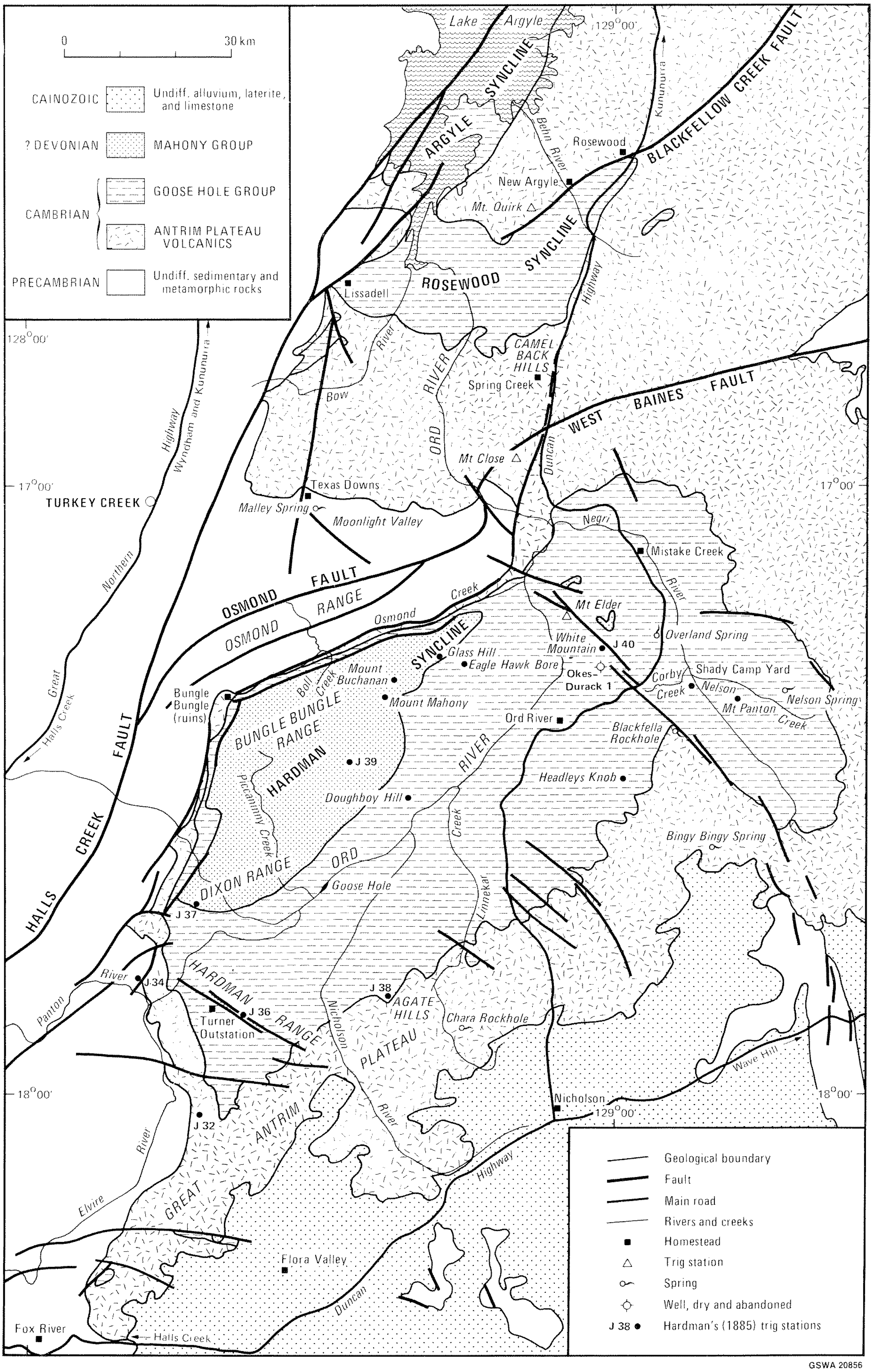

Figure 1. Locality map, Ord Basin.

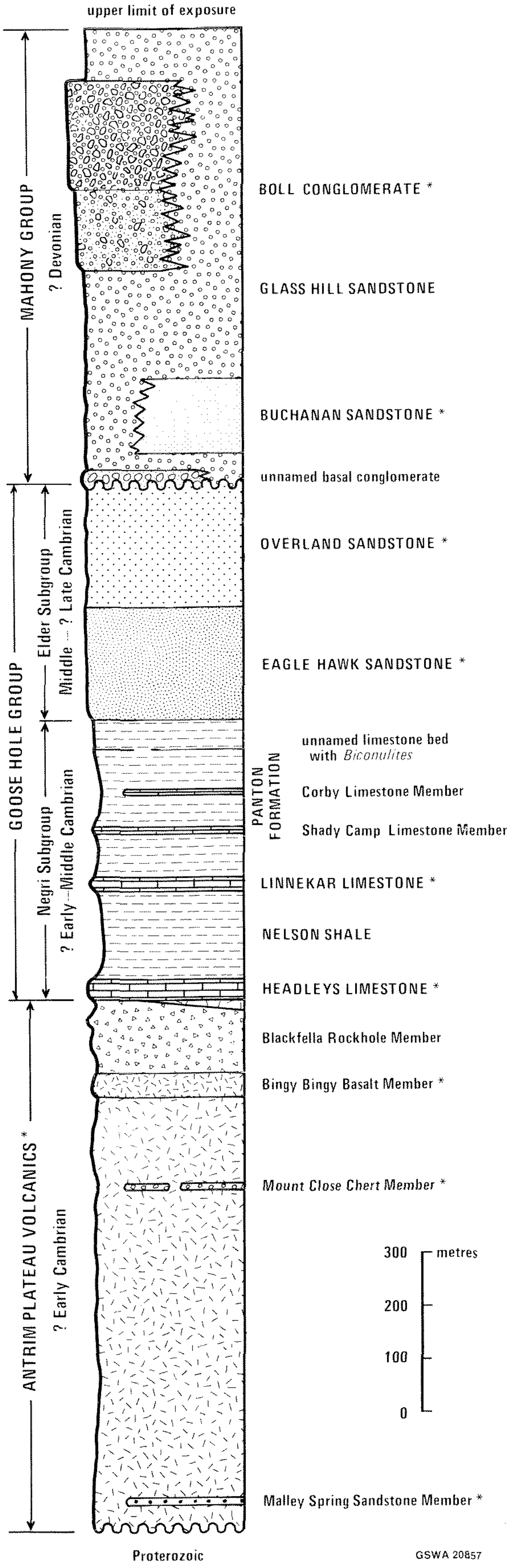

Figure 2. Generalized stratigraphic column showing the Ord Basin sequence. Thicknesses marked * are from type sections, otherwise the maximum thickness is shown. (See text for description of lithologies.)

STRATIGRAPHY

Three major subdivisions of the Palaeozoic sequence in the Ord Basin are proposed herein. These divisions correspond to significant lithological changes and differences in age (Fig. 2). In descending order these are:

Mahony Group (new name)—

yellow and white friable sandstone and conglomerate of probable Devonian age.

Goose Hole Group (new name)—

red, grey and white siltstones, arkosic sandstones and minor limestones of Cambrian age.

Antrim Plateau Volcanics—

basalt, agglomerate, minor chert, sandstone and siltstone of probable Early Cambrian age.

ANTRIM PLATEAU VOLCANICS

Definition: Hardman (1885) named the dissected basaltic hills east of the Elvire River the Great Antrim Plateau. David (1932) was, however, the first to use the name Antrim Plateau Basalts. Traves (1955) modified the name to Antrim Plateau Volcanics because, although dominated by basalt, the sequence also includes minor agglomerate and rare sandstone, tuff, siltstone and chert. Although Traves did not designate a type section, all subsequent workers have considered the Great Antrim Plateau to be the type area. Playford and others (1975, p.398) suggested that the section west of Hardman's trig. station J37 "may be taken as the type section." This designation is presumably based on the two sections shown by Dow and Gemuts (1967) on the Dixon Range sheet. As Dow (1980, p.3) later stated that "these sections have been shown, incorrectly, as type sections . . . " Playford and others' suggestion is not followed here. The only well-exposed section in the Great Antrim Plateau with both the upper and lower contacts visible is an east-west transect through Hardman's (1885) trig. station J32, 20 km south of "Turner". However, ground access to that section is only via "Fox River" and is difficult. Consequently the type section is here designated to be in the "Spring Creek" area where the diverse lithologies which make up this unit are better shown and where access, via the Duncan Highway, is much easier than at J32. The type section extends 12 km north from the eastern end of Moonlight Valley at 17°02′20″S and 128°44′25″E (base), to the base of the Blackfella Rockhole Member at 16°36′25″S and 128°45′50″E, and from there 24 km along strike to the Camel Back Hills at 16°45′55″S and 128°54′00″E, finishing 2.5 km NNW at the base of the Headleys Limestone at 16°44′50″S and 128°53′30″E.

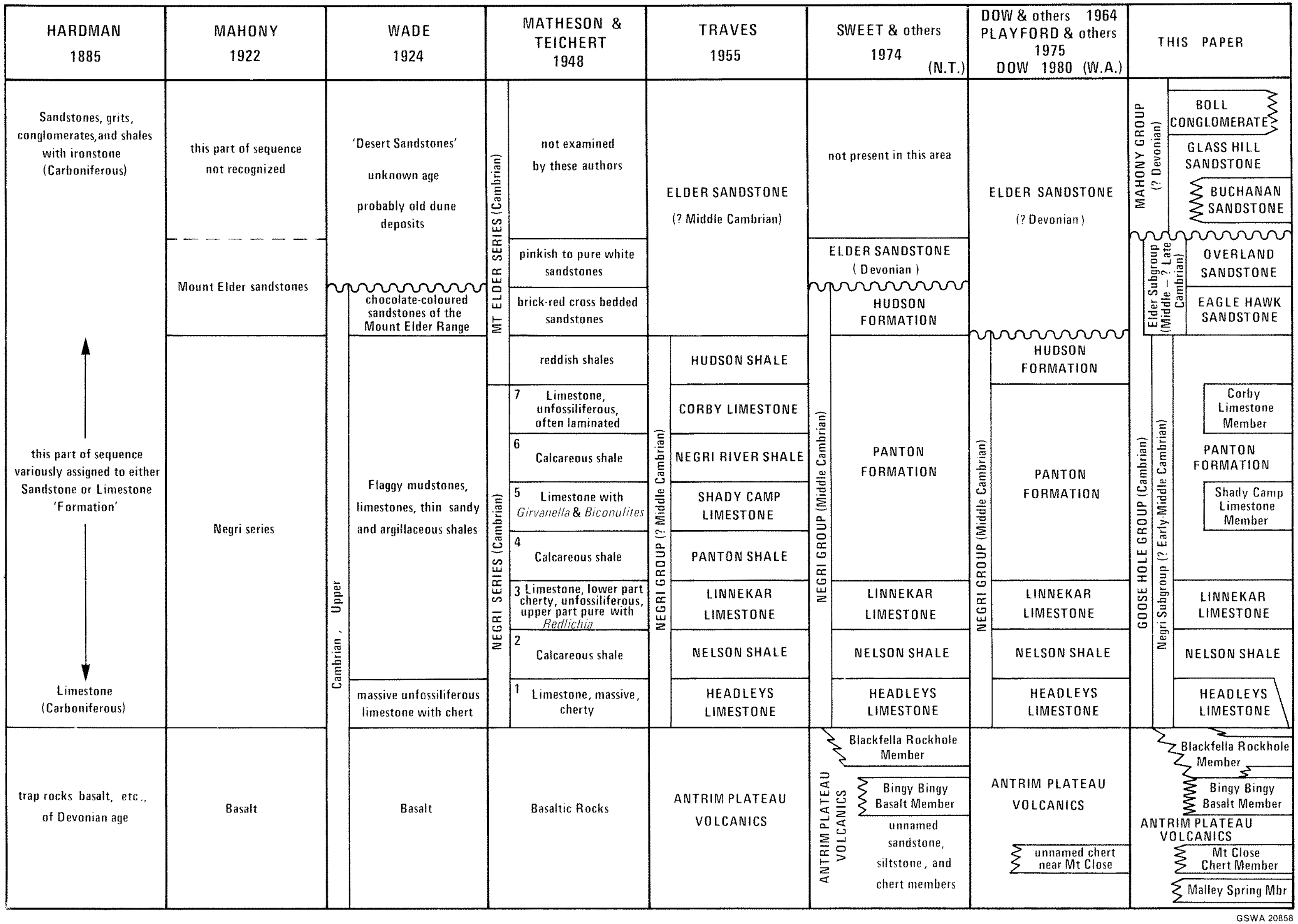

Figure 3. Correlation of previous stratigraphic nomenclature, Ord Basin.

Stratigraphic relationships: The Antrim Plateau Volcanics unconformably overlies Proterozoic sediments, locally with a marked angular relationship, and is conformably overlain by the Goose Hole Group.

Distribution and thickness: The Antrim Plateau Volcanics outcrops continuously around the edge of the Ord Basin and is also found in outliers west of the Halls Creek Fault, in the Bonaparte Basin, and along the western margins of the Wiso and Daly River Basins in the Northern Territory. Generally, the thickness of the unit ranges from 237 to 1 100 m in the Ord Basin and is thinnest to the north and east. The type section is approximately 1 000 m thick.

Fossils and age: The only fossils recovered from the Antrim Plateau Volcanics are the stromatolites *Conophyton basalticum* and *Conophyton* cf. *gaubitza* (Walter, 1972), from the Mount Close Chert Member, and these are not diagnostic of a particular age. The age of this unit, then, must be determined from stratigraphic relationships. The 650 m.y. radiometric date determined by Bofinger from the underlying Albert Edward Group (Dow, 1980), and the Middle Cambrian age of the overlying Goose Hole Group (Opik, 1967) indicate an Early Cambrian age.

Members: Four named members within the Antrim Plateau Volcanics are recognized. These are (in ascending order): the Malley Spring Member (new name), the Mount Close Chert Member (new name), the Bingy Bingy Basalt Member and the Blackfella Rockhole Member (amended).

The Malley Spring Member consists of 3 to 5 m of white to grey siltstone and medium- to coarse-grained sandstone exposed approximately 50 m above the base of the Antrim Plateau Volcanics in Moonlight Valley over a distance of 25 km from its eastern end. The type section is at the eastern end of Moonlight Valley (at 17°02′50″S and 128°43′05″E) and consists predominantly of siltstone. Sandstone becomes more common to the west.

The Mount Close Chert Member is named after Mount Close and consists of up to 5 m of laminated chert with intraformational breccias and non-branching, conical, columnar stromatolites up to 12 cm in diameter. The stromatolites are similar to those illustrated by Sweet and others (1974, figs 42 and 43) from the Wave Hill sheet. The type section is 2.5 km northeast of Mount Close by the side of the Duncan Highway at 16°56′25″S and 128°51′30″E. In this section the Mount Close Chert is 650 m above the base of the Antrim Plateau Volcanics.

Apart from the type area this unit is also recognized west of J37 and, questionably, at Mount Wittenoom on the Gordon Downs sheet.

The Bingy Bingy Basalt Member was originally defined in the Northern Territory by Sweet and others (1974, p.100). It is a massive, fine- to medium-grained glomeroporphyritic basalt with clots of plagioclase crystals up to 0.5 cm in diameter. The type section is 0.7 km north of Bingy Bingy Springs at 17°32′S and 129°15′E.

This member is present only betweeen Agate Hill and the Negri River at the southeastern end of the Ord Basin, and as lenses near "Spring Creek". At the type section it is 40 m thick. Near Charra Rockhole, the Bingy Bingy Basalt Member reaches a maximum thickness of 130 m, while at "Spring Creek" it is 20 m thick. The Bingy Bingy Basalt Member is presumably concordant with underlying and overlying basalt flows.

The Blackfella Rockhole Member was originally defined by Sweet and others (1974, p.99) in the Northern Territory and can be traced into Western Australia. This member, which constitutes the uppermost part of the Antrim Plateau Volcanics, is present in all sections of the formation and consists predominantly of fine-grained to very fine-grained basalt, agglomerate and minor amygdaloidal basalt, sandstone and siltstone. Sweet and others (1974) defined the base of this member as being the lowest agglomerate horizon. This definition is here amended to include the fine-grained basalt below the agglomerate which is virtually identical to that above. The abrupt upper contacts and gradational lower contacts of the agglomerate with the fine-grained basalt suggest that the agglomerate represents the upper part of single lava flows and that the fine-grained basalt and agglomerate should be included within the one unit.

The type section of this member is at Blackfella Rockhole (17°23′S, 129°06′E) in the Northern Territory, where 71 m of this unit is exposed. The base of the Blackfella Rockhole Member, as defined here, is not exposed at the type section and so the section east of Linnekar Creek near Charra Rockhole is suggested as a reference section—this is the thickest known section (140 m).

The Blackfella Rockhole Member has a lower conformable contact with the Bingy Bingy Basalt Member south and east of Ord River station, whereas to the north up to 50 m of undifferentiated basalt separates the two members. The upper contact is with the basal limestone of shale of the Goose Hole Group, except between Agate Hill and Headleys Knob, where up to 20 m of fine-grained, flow-banded, massive basalt is present above the Blackfella Rockhole Member.

GOOSE HOLE GROUP

Definition: The Goose Hole Group (new name) is named after Goose Hole on the Ord River and incorporates all sediments considered to be of Cambrian age within the Ord Basin. Thus defined, the Goose

Hole Group is composed of the Negri Subgroup (Negri Group of Traves, 1955) and the Elder Subgroup (previously the ?Devonian Elder Sandstone, in part).

The units which make up the Goose Hole Group are (in descending order):

Elder Subgroup (c)	Overland Sandstone (a)	- white clayey arkose.
	Eagle Hawk Sandstone (a)	- red, micaceous, arkosic sandstone.
Negri Subgroup (c)	Panton Formation (b)	- grey-purple siltstones, arkosic sandstones, & minor limestone members.
	Linnekar Limestone	- flaggy grey limestone & siltstone.
	Nelson Shale	- red-purple siltstone & flaggy arkosic sandstones.
	Headleys Limestone	- grey laminated limestone with chert.

(a) new name (b) amended (c) change in status

Distribution and thickness: The Goose Hole Group outcrops in three post-depositional synclines named the Hardman, Rosewood and Argyle Basins by Matheson & Teichert (1948). These were subsequently designated as sub-basins by McWhae and others (1958). The Hardman Syncline was referred to as the Ord River Basin by Wade (1924, p.26). However the Ord Basin (McWhae and others, 1958) has become well established in the literature and the name Hardman is retained for the syncline between Osmond Range and the Great Antrim Plateau.

The thickest sections of the Goose Hole Group are preserved between the hills north of White Mountain and the junction of the Ord and Negri Rivers—there the Goose Hole Group is approximately 700 m thick. Within the Hardman Syncline, the thinnest sections (500 m), in the vicinity of Linnekar Creek-Doughboy Hill, are chiefly the result of thinning of the Eagle Hawk Sandstone and the removal, by erosion, of most of the Overland Sandstone. In the Argyle Syncline the Goose Hole Group is inferred to be 180 m thick. Exposures of other than the resistant limestone units within the group are exceptionally rare in the Argyle and Rosewood Synclines.

NEGRI SUBGROUP

The Negri Subgroup was originally called the "Negri series" by Mahony (1922) and refers to the "limestone, mudstone, and shale" which overlie the basalts. It is thus equivalent to the Negri Group of Traves (1955) and Dow (1980). Two publications have adopted different definitions: Matheson and Teichert (1948), who excluded the siltstones above the highest limestone they recognized (Corby Limestone of Traves, 1955); and Sweet and others (1974), who included the red feldspathic sandstones and arkose (their Hudson Formation which is equivalent to the Eagle Hawk Sandstone of this paper) above the siltstone sequence.

HEADLEYS LIMESTONE

Definition: The Headleys Limestone is the basal unit of the Negri Subgroup in the Hardman and Rosewood Synclines. It appears to be absent in the Argyle Syncline. The limestone consists of approximately 40 m of grey, massive, or laminated micrite with chert nodules common in the lower half of the unit. Traves (1955,p.37) stated that "the type locality is at Headleys Knob 11 miles [18 km] southeast of Ord River Homestead". This renders invalid the "type section" proposed by Dow (1980, p.8) to be 3 km south of Dixon Range. At Headleys Knob dips are low (less than 2°), and so to include the base and top of the unit, the type section here defined runs 4.5 km in a NNW traverse across Headleys Knob from 17°30′25″S and 129°01′25″E (base) to 17°27′25″S and 129°00′25″E (top).

Stratigraphic relationships: The lower contact of the Headleys Limestone with the Antrim Plateau Volcanics is sharp but concordant. Ferruginization and silicification of the basalts below the limestone are possibly recent events, not necessarily the result of erosion or weathering prior to deposition of the Headleys Limestone as suggested by Sweet and others (1974) and Dow (1980). The contact with the overlying Nelson Shale is similarly abrupt and is presumably conformable.

Fossils and age: The only fossils recovered from the Headleys Limestone are simple, non-branching stromatolites which have not been described. Based on the stratigraphic position of this unit, Traves (1955) suggested an uppermost Lower Cambrian or basal Middle Cambrian age.

Nelson Shale

Definition: The Nelson Shale is a purple siltstone with thin beds of fine arkosic sandstone and rare laminated micrite. The unit is named after Nelson Springs where exposures are incomplete (Traves, 1955). In the absence of a completely exposed section, the exposures between Nelson Spring and Mount Panton in the Northern Territory (from 17°19′20″S, 129°19′15″E (base) to 17°21′10″S, 129°13′20″E (top)) are, as the nominal area, here designated as the type section.

Stratigraphic relationships: In the Argyle Syncline the Nelson Shale rests directly on the Antrim Plateau Volcanics, but the only section in which the Nelson Shale was exposed is now covered by the waters of Lake Argyle. Elsewhere in the Ord Basin the Nelson

Shale conformably overlies the Headleys Limestone. The upper contact with the Linnekar Limestone is sharp but apparently conformable.

Distribution and thickness: The Nelson Shale occurs in the Hardman, Rosewood and Argyle Synclines, but inthe latter two synclines outcrop is practically non-existent. In the Hardman Syncline, the Nelson Shale is between 100 and 183 m thick, with the greatest thickness being recorded in Okes-Durack 1.

Fossils and age: Girvanella is the only fossil recorded from the Nelson Shale: The Middle Cambrian age suggested by Traves (1955) is based on the stratigraphic position below the fossiliferous Linnekar Limestone.

Linnekar Limestone

Definition: The Linnekar Limestone consists of flaggy limestone and shale lying conformably between the Nelson Shale below and the Panton Formation above. The unit was named by Traves (1955) after Linnekar Creek [Linacre on Hardman's (1885) and Matheson and Teichert's (1948) maps]. Playford and others (1975) and Dow (1980) consider the type section to be near the junction of Linnekar and Brook Creeks (at 17°32′50″S and 128°42′35″E).

The Linnekar Limestone is subdivided into three unnamed members: basal laminated micrite with chert nodules and circular, non-columnar, flat-laminated, domed stromatolites up to 1.5 m in diameter; middle-grey to olive trilobitic shale; and upper fossiliferous flaggy limestone and shale with rare turnbinate, non-branching, columnar stromatolites.

The first known occurrence of galena in the East Kimberley region was discovered in this unit by Hardman (1885) in the bed of the Elvire River, 10 km north of "Turner".

Distribution and thickness: The Linnekar Limestone is present in all three synclines in the Ord Basin. In the Hardman Syncline the limestone is between 8 and 21 m thick with 18 m, measured at the type section, being close to the mean.

Fossils and age: The presence of *Redlichia forresti* in the middle and upper parts of the Linnekar Limestone suggests an early Middle Cambrian age (Opik, 1967). *Redlichia* is also known from the Early Cambrian in South Australia (Jell, 1983), which suggests that the unit could be somewhat older. Besides *Redlichia* the only other fossils known in this unit are *Girvanella, Biconulites hardmanni* and unnamed stromatolites.

PANTON FORMATION

Definition: The Panton Formation (amended), as defined herein, is the sequence of purple siltstone, flaggy, chocolate-brown arkosic sandstone and minor limestone which conformably overlies the Linnekar Limestone. Thus, as in Sweet and others (1974), the Panton Formation incorporates (in ascending order) the Panton Shale, Shady Camp Limestone, Negri River Shale, Corby Limestone and Hudson Shale of Traves (1955). The Hudson Formation cannot be distinguished from the Panton Formation (as defined by Dow, 1980) in the absence of the intervening lenticular Corby Limestone: the flaggy chocolate-brown sandstone and the siltstone which Dow (1980) claimed characterized the Hudson Formation are also common below the Corby Limestone. The Hudson Formation of Sweet and others (1974) is equivalent to the Eagle Hawk Sandstone of this paper.

Dow and others (1964) stated that the type locality for the Panton Formation is the Mount Panton—White Mountain area but did not designate a specific type section. Playford and others (1975) designated Mount Panton in the Northern Territory (17°16′15″S and 129°12′40″E) as the type section. This designation has priority over the subsequent designation of Dow (1980) who suggested the section in Hudson Creek on the southwest flank of White Mountain. As the uppermost part of the Panton Formation (the Hudson Formation as defined by Dow, 1980) is not present at Mount Panton, the section in Hudson Creek (from 17°15′30″S, 128°57′30″E (base to 17°15′35″S, 128°57′50″E (top)) is here proposed as a reference section for the Panton Formation.The correlation between Mount Panton and White Mountain was demonstrated by Dow (1980, Fig. 4.)

Distribution and thickness: The greatest thickness through the Panton Formation was recorded in the reference section at White Mountain (308 m). At the type section (Mount Panton) only 85 m is exposed but this section is incomplete. In the Hardman Syncline the thinnest section, at 105 m, was measured in Osmond Creek 40 km west of White Mountain, a thickness comparable to that estimated in the Argyle Syncline.

Fossils and age: The most fossiliferous part of the Ord Basin sequence is the shale and limestone succession between the Shady Camp and Corby Limestone Members, especially in the Mount Panton area. The assemblage of *Redlichia* and *Xystridura* from this part of the Panton Formation suggests a Middle Cambrian age (Opik, 1967; Jell, 1983). Also reported from the Panton Formation are the brachiopods *Billingsella* and *Winanella*, girvanellids, stromatolites, and a small conical shell of uncertain affinities, *Biconulites hardmani* (Traves, 1955). In the collection made by Mahony (1922), presumably from the Panton Formation, Chapman (1924) identified "a stromatoporoid, foraminifera, (encrusting), radiolaria, ostracoda and archeocyathina". This collection has been reassessed by P. Jell (written comm., 1982), who identified these fossils as a stromatolite, pisolites, carbonate pellets, *Biconulites* fragments and cone-in-cone *Biconulites* respectively.

Members: Two members within the Panton Formation can be recognized: the Shady Camp Limestone, and the Corby Limestone of Traves (1955). Both limestones lens out to the west in the Hardman Syncline and cannot be traced beyond 128°35′E.

The Shady Camp Limestone Member consists of up to 4 m of oncolitic, fossiliferous limestone with abundant *Biconulites* and trilobites. The type section is here designated as Mount Panton, where this member occurs 41 m above the base of the Panton Formation. The Shady Camp Limestone may occur between 30 m (in the Argyle Basin) and 92 m (at White Mountain) above the base of the Panton Formation.

The Corby Limestone Member consists of up to 4 m of massive to laminated, unfossiliferous micrite with chert nodules, and is similar to the Headleys Limestone. The type section is here designated as Hudson Creek, 160 m above the base of the Panton Formation (at 17°15′35″S and 128°57′20″E). At Mount Panton, where the limestone is 83 m above the base of the Panton Formation, the overlying shales are not exposed. In the Hardman Syncline the top of the Corby Limestone may be as little as 55 m above the base of the Panton Formation (estimated in Osmond Creek, 40 km west of White Mountain)—a thickness comparable to that in the Argyle Syncline (40 m).

ELDER SUBGROUP

Mahony (1922) originally named the "Mount Elder sandstones" after the "red sandstone . . . which forms a conspicuous chain of hills of which Mount Elder is the most conspicuous point." Subsequent workers have applied the name "Elder" to the entire arenaceous sequence above the Panton Formation, as defined herein, and in doing so have grouped together sandstones here considered to be of probable Cambrian age (Elder Subgroup) and sandstones of probable Devonian age (Mahony Group). Since Mahony (1922) named his "Mount Elder sandstones" after Mount Elder, Playford and others (1975) and Dow (1980) considered the type area to be in the immediate vicinity of Mount Elder, that is, in the White Mountain Hills. In this area two sandstone units may be recognized: a lower red, micaceous, arkosic sandstone and an overlying white, clayey arkose, here named the Eagle Hawk Sandstone and Overland Sandstone (new names) respectively, and together comprising the Elder Subgroup. The Elder Subgroup has a contact with the underlying Panton Formation which, although generally gradational, may also be abrupt. Diagnostic fossils have yet to be recovered from these two sandstone units. The gradational contacts suggest that a significant hiatus between them, or with the underlying Negri Subgroup, is unlikely.

EAGLE HAWK SANDSTONE

Definition: The Eagle Hawk Sandstone (new name) is named after Eagle Hawk bore and consists of thin-to medium-bedded, festoon cross-bedded and current-lineated, fine-grained, red, micaceous, arkosic sandstone and minor red siltstone. The type section is here designated as Hudson Creek on the southwestern flank of White Mountain from 17°17′45″S and 128°58′45″E (base) to 17°15′30″S and 128°57′30″E (top), since this is the most accessible section where both the upper and lower contacts are exposed.

Stratigraphic relationships: The Eagle Hawk Sandstone overlies, generally with a transitional contact, the flaggy, ripple-cross-laminated, arkosic sandstone and siltstone characteristic of the uppermost part of the Panton Formation. The Eagle Hawk Sandstone is readily distinguished from the white, clayey arkose of the conformably overlying Overland Sandstone, although the contact is gradational. Where the Overland Sandstone has been removed by erosion, the Eagle Hawk Sandstone is easily distinguished from the pebbly quartz sandstones of the disconformably overlying Mahony Group.

Distrubution and thickness: Outcrop of Eagle Hawk Sandstone is primarily in the Hardman Syncline. Red sandstones, poorly exposed in the centre of the Argyle Syncline, are also referred to this unit. In the type section at Hudson Creek, the unit reaches a maximum thickness of 210 m. At Doughboy Hill, 80 m (estimated from air photographs) of Eagle Hawk Sandstone is present, and is comparable to the thickness (75 m) determined by Sweet and others (1974) in the Northern Territory for their Hudson Formation (the Eagle Hawk Sandstone of this paper). These are the thinnest non-eroded and non-faulted sections in this unit.

Fossils and age: The only fossils discovered in the Eagle Hawk Sandstone are trilobite tracks which are not age-diagnostic. The Middle Cambrian age here suggested for this unit is based on its stratigraphic position above the fossiliferous Negri Subgroup and on a lithological correlation with the Middle Cambrian Hart Spring Sandstone in the Bonaparte Basin, a correlation previous suggested by Traves (1955) for his Elder Sandstone.

OVERLAND SANDSTONE

Definition: The Overland Sandstone (new name) consists of white, medium- to fine-grained, clayey lithic arkose which overlies the Eagle Hawk Sandstone with a gradational lower contact and which is disconformably overlain by pebbly sandstone and conglomerate of the Mahony Group. The sandstone is named after Overland Spring, 7 km southeast of the type section which runs from 17°09'15"S and 129°00'15"E (base) to 17°10'55"S and 128°59'50"E (top) and is the thickest known section (230 m) in which both the underlying and overlying units are exposed. The Overland Sandstone was formerly included in the Elder Sandstone (Fig. 3) together with the underlying Eagle Hawk Sandstone and overlying Mahony Group.

Distrubution and thickness The chief area of outcrop of the Overland Sandstone is the range of hills north of White Mountain where thicknesses of up to 230 m have been estimated from aerial photographs, although only the uppermost 50 m is well exposed. The unit thins to the southwest, being approximately 10 m thick near Mount Buchanan and Doughboy Hill, and pinches out just north of where the Ord River cuts through Dixon Range. This thinning appears to be the results of erosion prior to the deposition of the Mahony Group. The overland Sandstone is not known in the Rosewood or Argyle Synclines.

White Mountain takes its name from the brilliant white cliff of Overland Sandstone near its summit. The name "White Mountain" has, however, previously been applied to the Tertiary limestone which caps this hill (Matheson and Teichert, 1948).

MAHONY GROUP

Definition: The Mahony Group (new name), named after Mount Mahony, is the white and yellow quartz sandstone, pebbly sandstone, and conglomerate which disconformably overlies the Goose Hole Group. The group is divided into three formations of which two, the Boll Conglomerate and Buchanan Sandstone (new names), interfinger with and are laterally equivalent to parts of the third—the Glass Hill Sandstone (new name).

Distribution and thickness: The Mahony Group outcrops chiefly in the centre of the Hardman Syncline in the triangular area between Bungle Bungle Outcamp, Dixon Range and Mount Buchanan. Small inliers also occur at Hardman Range, west-northwest of "Flora Valley" and north of White Mountain. The maximum thickness in Bungle Bungle Range, estimated perpendicular to the syclinal axis, is between 750 and 1 000 m. This thickness may, nevertheless, be considered as a minimum as the upper parts of this group have been removed by recent erosion.

Fossils and age: The only fossils recovered from the Mahony Group are several species of bivalve from north of White Mountain and these are too poorly preserved to determine the age of the Group. The Mahony Group is here considered to be of probable Devonian age, based on a lithological correlation with the Frasnian Cockatoo Formation in the Bonaparte Basin. Previously the Mahony Group was included in the Elder Sandstone and assigned a Cambrian age by Matheson and Teichert (1948) and Traves (1955) on the basis of the gradational lower contact of the Eagle Hawk Sandstone with the Negri Subgroup. Dow and others (1964) and Dow (1980), however, suggested a Devonian age, based on a correlation of the basal conglomerates at Dixon Range with the Ragged Range Conglomerate Member in the Bonaparte Basin. At Dixon Range the Eagle Hawk Sandstone below the disconformity was mapped as Hudson Formation by these workers.

GLASS HILL SANDSTONE

Definition: The Glass Hill Sandstone (new name) consists of pebbly quartz sandstone, rare siltstone, and a lenticular basal conglomerate. The sandstones chiefly consist of medium- to fine-grained, well-sorted, friable, clean quartz sandstone in which pebbles generally make up less than 1% and rarely greater than 5% of the bulk. The type section is at the south end of Dixon Range, near Hardmans' (1885) trig. station J37, beginning at a small breakaway at 17°41'45"S and 128°16'50"E and continuing 1 km NE to the top of the range. In this section approximately 200 m is exposed. Thicker sections of up to 850 m have been estimated to the north on the eastern side of Bungle Bungle Range.

Stratigraphic relationships: The basal lithologies disconformably overlie either the Eagle Hawk Sandstone or Overland Sandstone of the Elder Subgroup. The uppermost part of this unit is exposed at the top of the range southeast of Bungle Bungle Outcamp where younger strata are unknown. The Glass Hill Sandstone has conformable contacts with and is in part equivalent to the two other formations of the Mahony Group—the Buchanan Sandstone and Boll Conglomerate.

BUCHANAN SANDSTONE

Definition: The Buchanan Sandstone (new name) is a clean, well-sorted, medium-grained quartz sandstone in which well-rounded, frosted grains are common. The type section begins at the base of a hill 5.5 km south of Mount Buchanan (17°22'20"S and 128°36'50"E) and continues northwest across a line of hills for 1.5 km. As dips are low in this area and the upper parts of the cliff exposures are largely inaccessable a reference section is proposed. This reference section lies 3 km southwest of the type section and extends from 17°23'25"S and 128°35'00"E (base) to 17°23'05"S and 128°33'50"E (top).

Stratigraphic relationships: Although the lower contact is nowhere exposed, it is known that the lateral and upper contacts of the Buchanan Sandstone interfinger with the Glass Hill Sandstone.

Distribution and thickness: The Buchanan Sandstone outcrops along the eastern margin of Bungle Bungle Range. The unit is appropriately 130 m thick in the type area but the lower contact is not exposed. The total thickness of the Buchanan Sandstone is, however, unlikely to exceed 140 m.

BOLL CONGLOMERATE

Definition: The Boll Conglomerate (new name) consists of a lower unit of conglomerate, pebbly sandstone and sandstone in which clasts rarely exceed 18 cm in diameter, and an upper, cliff-forming, massive conglomerate with quartzite boulders up to 0.9 m in diameter. The type section is the west wall of the north-directed gorge 3 km east-southeast of Bungle Bungle Outcamp at 17°22′10″S and 128°22′10″E.

Stratigraphic relationships: The lower contact of the Boll Conglomerate with the Glass Hill Sandstone is abrupt but concordant. Laterally, the conglomerate appears to interfinger with the Glass Hill Sandstone. The upper contact of the Boll Conglomerate is not accessible but on airphotos it appears to be overlain by Glass Hill Sandstone.

Distribution and thickness: The Boll Conglomerate outcrops in a narrow belt running 10 km east-northeast from Bungle Bungle Outcamp along the southern side of Red Rock Creek. A thin tongue also extends 8 km south along the top of the range to the head of Piccaninny Creek. In the type section the Boll Conglomerate is approximately 350 m thick.

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STRUCTURAL AND STRATIGRAPHIC RELATIONSHIPS IN THE ARCHAEAN GRANITE-GREENSTONE TERRAIN AROUND CUE WESTERN AUSTRALIA

by K. P. Watkins and I. M. Tyler

ABSTRACT

Recent mapping of Archaean granite-greenstone terrain in the Cue region of the Yilgarn Block has outlined two tectonic domains in the supracrustal (*i.e.* greenstone belt) rocks. One is characterized by isoclinal folding and intense deformation; the other by open to close folding, weak deformation, and preservation of primary lithological features. One of two alternative models could account for these domains: the domains represent supracrustal sequences of different ages separated by an unconformity, the intense deformation recorded in one domain predating deposition of rocks in the other; or the domains represent different tectonic levels (and regimes) within one stratigraphic sequence. In both models, the weakly deformed domain contains rocks which are younger than those in the other domain.

Large enclaves of banded gneiss occur in the granitoids; these are structurally more complex than the supracrustals and are probably remnants of a sialic basement to the granite-greenstone system. Two phases of granitoid intrude the gneisses and supracrustals: a volumetrically dominant early phase which is recrystallized and has foliated margins with the greenstone belts; and a later, post-tectonic phase which truncates all lithologic and tectonic trends and retains igneous mineral assemblages and textures.

INTRODUCTION

Cue is situated near the centre of a large tract of supracrustal rocks (greenstones), stretching from Meekatharra in the northeast to Perenjori in the southwest, which, together with the associated granitoids, forms the Murchison Province of the Archaean Yilgarn craton (Fig. 1). The geological framework of the Cue region is typical of the Murchison Province; isolated, irregularly shaped outcrops of greenstones occur in a much more voluminous granitoid terrain.

Recent mapping in this region has revealed the presence of distinct tectonic domains of differing styles and intensities of deformation in the greenstone belts. Large areas of banded granitoid gneisses which bear similarities to those in the Western Gneiss Terrain have been identified. In addition, two phases of intrusive granitoids have been recognized. Figure 2 is a simplified lithological interpretation map of the region; outcrop extent (particularly of the granitoids) is partly based on photo-interpretation, aeromagnetic data (Waller and Beattie, 1971), landsat scenes 120-079 CUE and 119-079 SANDSTONE (W), and previous geological mapping (de la Hunty, 1973). Structural information relating to this map is given in Figure 3.

In the following sections, the major structural relationships in the greenstones, gneisses, and granitoids are described, and models are proposed which may explain these. The stratigraphic implications of the models are discussed.

STRUCTURAL RELATIONSHIPS

Supracrustal rocks (greenstone belts)

Two distinct tectonic domains have been mapped in the greenstone belts (Fig. 3). The largest, domain A, is characterized by intense deformation and penetrative fabrics associated with isoclinal folding. In this domain, strain is often taken up in discrete shear zones (a few metres to a few hundred metres in width) in which primary lithologies become schistose; for example, komatiitic basalt becomes tremolite-chlorite-talc schist, and tholeiitic basalt becomes amphibolite (although other protoliths are possible for these rocks). Shear may also occur along anastomosing planes, between which primary textures are preserved (Fig. 4). Thick, competent units of gabbro and dolerite, and some tholeiitic piles, often have well-preserved interiors with relic igneous textures; however, the margins are generally sheared to amphibolite. Only one megascopic isocline, that at Tuckabianna (Fig. 3), has been mapped in domain A. This fold has an exposed (minimum) amplitude of 15 km; aeromagnetic data (Bureau of Mineral Resources Cue 1:250 000 aeromagnetic/radiometric sheet) indicate that the southeast limb extends a further 12 km southwest under Lake Austin. It is likely that other large isoclines exist in the area, but the exposure is such that these are obscured. Minor, tight to isoclinal folds are ubiquitous in the thin (1-5 m) units of banded iron-formation in this domain (Fig. 5); the foliation is axial planar to these folds. Minor folds are rare in other lithologies, which tend to be massive or schistose and to lack marker horizons.

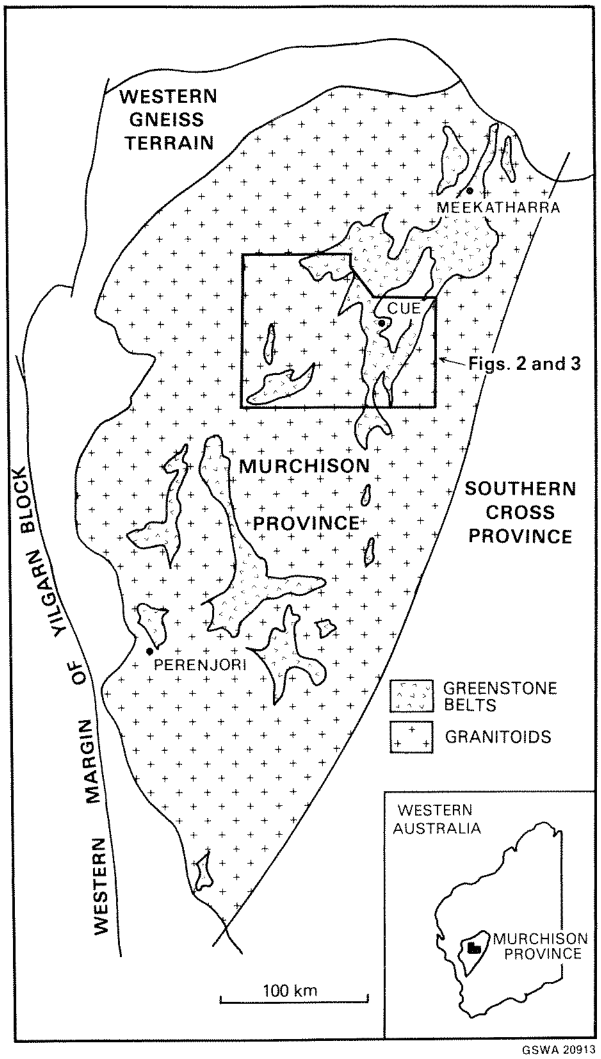

Figure 1. Distribution of granitoids and greenstone belts in the Murchison Province of the Yilgarn Block. Area of Figures 2 and 3 is indicated.

Domain B is characterized by megascopic open-to-close folding, *cf.* Ramsay (1967, p.349); minor folding is rare; axial-planar cleavages are developed only sporadically; and primary lithological features and textures are preserved. The outcrops at Ryansville and Dalgaranga (Figs 2 and 3) are assigned to this domain. At Ryansville, a sequence of komatiite, komatiitic basalt—terminology after Arndt and Nesbit (1982)—and tholeiite 3.5 km thick, is folded into an open syncline plunging about 70° southwest (Fig. 3). There is little internal deformation, and an axial-planar cleavage is not developed. An example of a well-preserved pillow lava from this sequence is shown in Figure 6. In the Dalgaranga region, a close syncline plunges about 65° towards the northeast (Fig. 3); this fold tightens towards the southwest. Structural elements (foliations etc.) in domain A swing around this fold (Fig. 3), *see* southwest portion of Cue 1:250 000 sheet (de la Hunty, 1973), and southeast portion of Murgoo 1:250 000 sheet (Baxter, 1974). Primary textures are preserved in dacitic tuffs and coarse-grained volcaniclastic sediments in the core of this fold (Fig. 7). Axial-planar cleavage is developed only towards the southwest, where the fold tightens; no minor folding has been found despite the abundance of planar marker horizons. A thick, multiple gabbro sill intruded and substantially thickened the sequence prior to folding. As a consequence of this intrusion, less competent pelitic sediments adjacent to the gabbro have been attenuated and statically metamorphosed to amphibolite facies—andalusite (-cordierite-garnet)-bearing assemblages.

A further two phases of folding and associated spaced cleavage affected the greenstone belts; but in order to preserve clarity, these are not marked on Figure 3. A phase of open folding with north- to north-northeast-trending axes occurs most noticeably between Tuckabianna and Ryansville, although a closure also occurs southeast of Moyagee. A cleavage associated with this folding cuts across the axial plane of the Ryansville syncline. A later phase of open folding with northwest-trending axes and an associated, spaced cleavage is seen in most of the greenstone belts. The axial plane of the Dalgaranga syncline, and the foliation in domain A, immediately to the south are folded by this phase. A phase of late, large-scale faulting, which trends north to northeast, dissects the Moyagee-Tuckabianna belt (Figs 2 and 3).

Banded gneiss

The gneisses are compositionally banded plagioclase-microcline-quartz-biotite rocks, which record a complex deformation history. They occur mainly along the western margin of the area as large, fairly coherent units and also as smaller "rafts" in the granitoids throughout the area (Fig. 2). The banding is folded, and a variety of fold styles are seen (Figs. 8, 9, and 10); an isoclinal phase is the most commonly observed. The main trend of the banding is parallel to tectonic trends in domain A of adjacent greenstone belts and in foliated granitoids.

The isoclinal fold phase commonly seen in the banded gneiss is thought to be the same as that seen in domain A. Axial planes of minor isoclinal folds in the Warda Warra greenstone belt (Figs 2 and 3) are approximately co-planar with those of isoclines in the banded gneiss to the west. The only evidence for a phase of deformation predating the isoclinal folding in the greenstones is an early lineation of unknown significance recorded at only one locality in the Tuckabianna fold. Extensive deformation must have occurred in the gneiss prior to isoclinal folding in order to develop the pronounced banding, *cf.* Myers (1978). Thus it is likely the banded gneiss predates the supracrustal rocks and forms an early sialic basement.

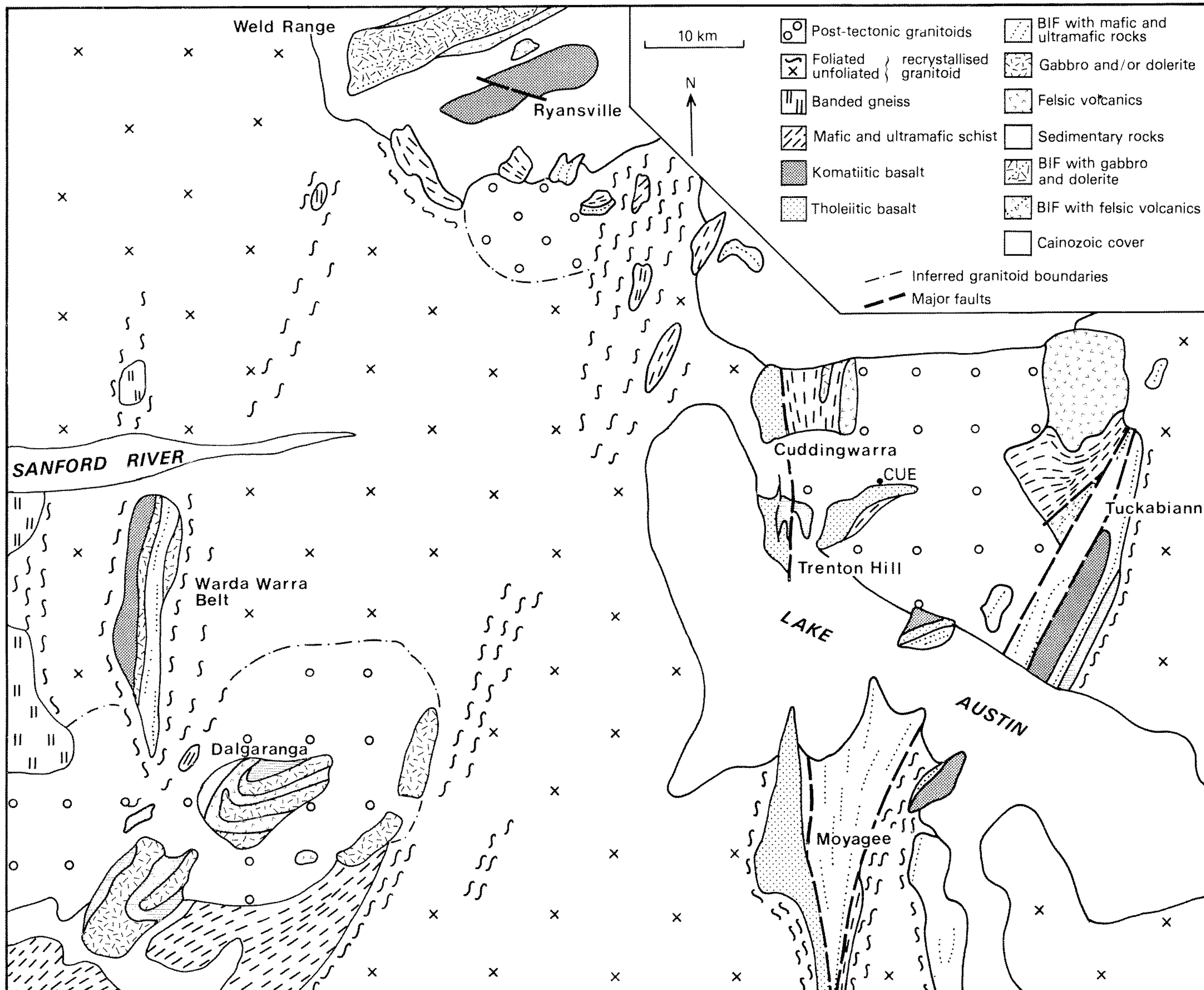

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Figure 2. Simplified lithological interpretation map of the Cue region. Interpretation is impossible beneath some extensive areas of Cainozoic cover.

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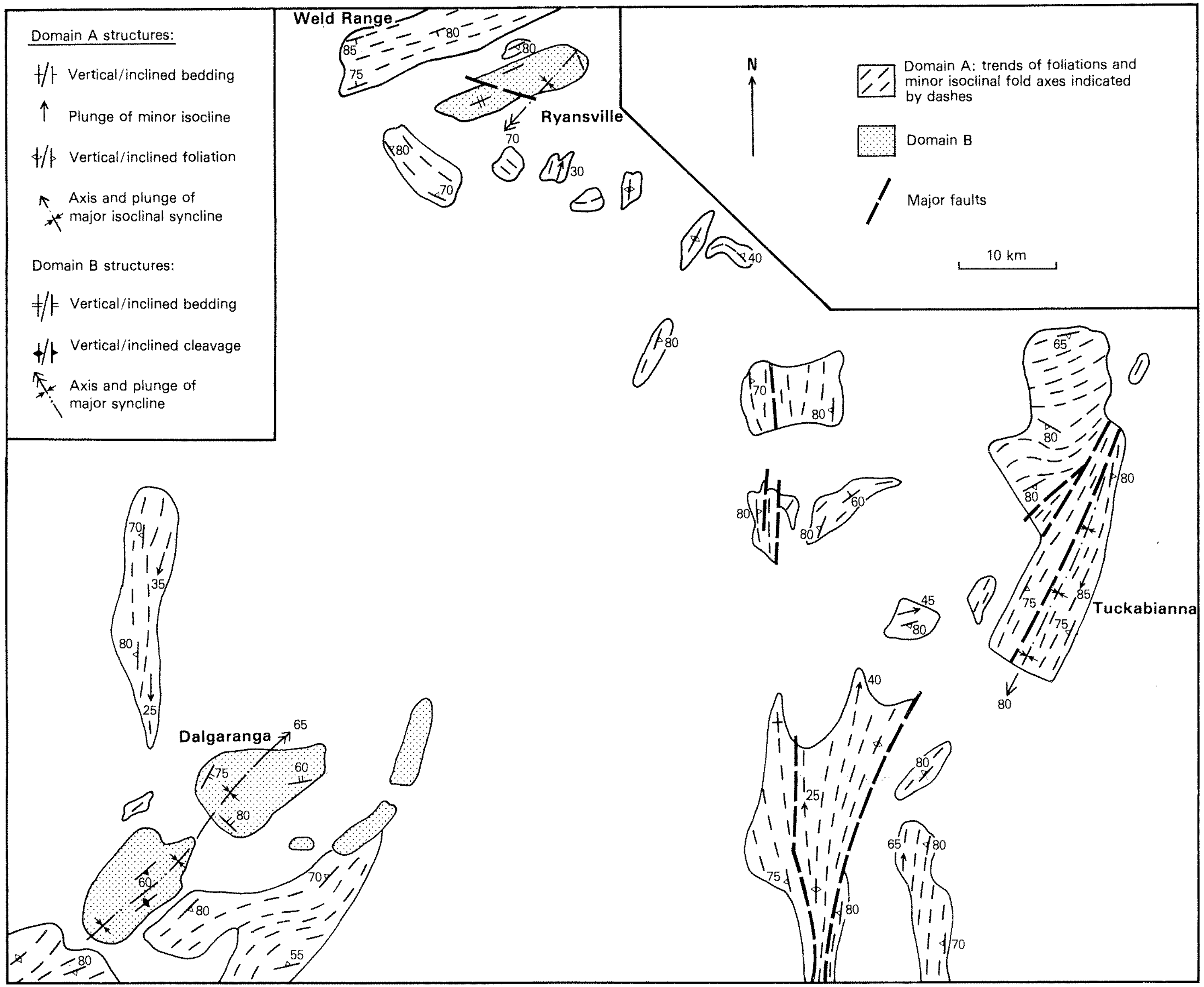

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Figure 3. Structural map of the supracrustal rocks in the Cue region. Two, late, open phases of folding, with north and northwest axial trends, are omitted for clarity. Domain A is characterized by isoclinal folding and intense deformation. Domain B is characterized by open to close folding, weak deformation and preservation of primary lithological features.

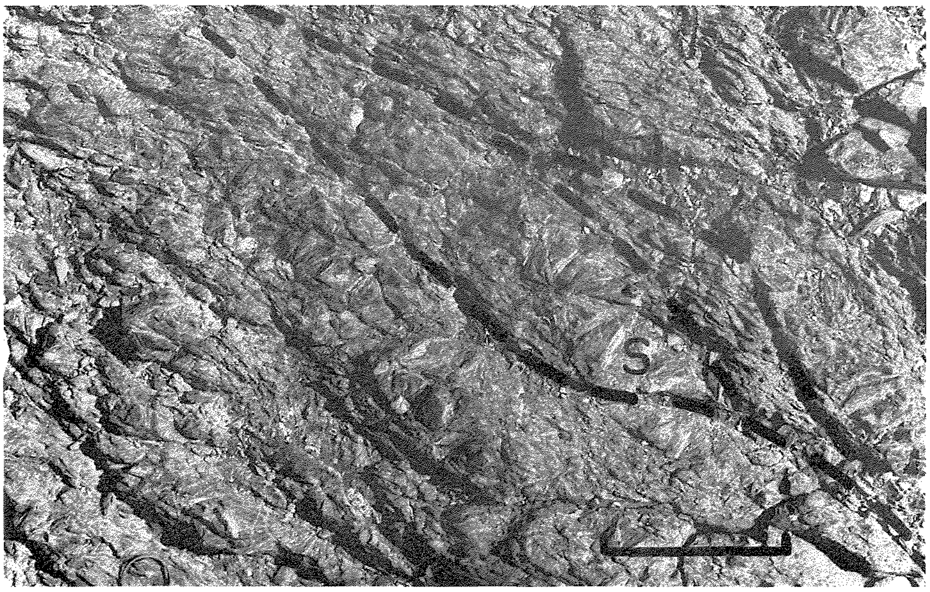

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Figure 4. Anastamosing shears composed of tremolite-chlorite-talc schist in a komatiitic basalt (examples are indicated by dashed lines); original spinifex texture is preserved in lenses between the shears (*e.g.* at S). The scale bar is 5 cm.

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Figure 5. Minor, tight to isoclinal folding in a banded iron-formation in domain A. The hammer shaft is 30 cm long. GSWA

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Figure 6. Weakly deformed komatiitic pillow basalt at Ryansville, stratigraphic younging is towards the top of the photograph. A fracture cleavage (parallel to the short dimension of the photograph) cuts across the pillows and is axial planar to the late north-trending open phase of folding (see text).

GSWA 20919

Figure 7. Undeformed coarse-grained (conglomeratic) volcaniclastic derived from underlying dacites at Dalgaranga. Bedding runs from bottom left to top right of the photograph. The lens cap is 5 cm in diameter.

GSWA 20920

Figure 8. Tight to isoclinal fold of banding in granitoid gneiss.

GSWA 20922

Figure 10. Complex folding of banding in granitoid gneiss.

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Figure 9. Open to close folding of banding in granitoid gneiss.

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Figure 11. Strongly foliated and compositionally banded recrystallized granitoid, 300 m from a greenstone belt contact.

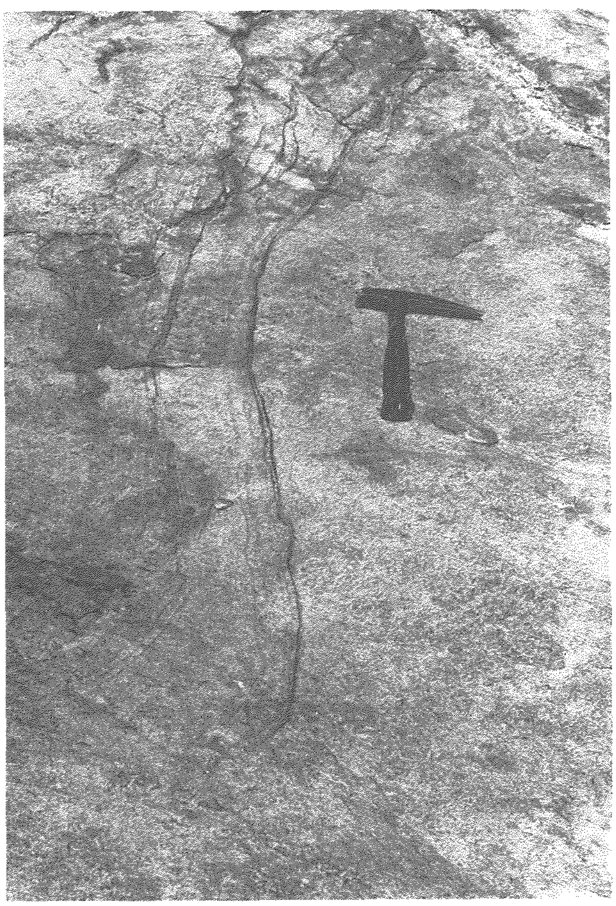

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Figure 12. "Ghost banding" in an unfoliated recrystallized granitoid; this is a recrystallized and partially absorbed xenolith of banded gneiss.

Intrusive granitoids

Two types of intrusive granitoid are recognized, a recrystallized and partly foliated phase and a post-tectonic phase. The recrystallized granitoids are the most voluminous (Fig. 2) and are adamellitic in composition. The margins of these granitoids adjacent to greenstone and gneiss belts are strongly foliated and often compositionally banded (Fig. 11). The foliation trends are generally parallel to the margins of the greenstone and gneiss belts, which themselves are often parallel to tectonic trends within these belts. The granitoid-greenstone contacts are generally zones of mixed and foliated granitoid and greenstone rocks ranging from a few tens of metres to a kilometre wide. The intensity of foliation and degree of compositional banding decrease rapidly away from the margins, and for the greater part of their outcrop, these granitoids are unfoliated. Both the foliated and unfoliated portions are thoroughly recrystallized and exhibit granoblastic metamorphic textures; original phenocrysts in unfoliated porphyritic varieties are pseudomorphed by microcline porphyroblasts. Metamorphic recrystallization outlasted deformation in foliated types. Several linear and apparently discontinuous zones of moderate to high strain trend north-northeast across these granitoids (Fig. 2); the significance of these is unknown. Throughout large areas of recrystallized granitoid adjacent to the gneiss outcrops, patches of "ghost banding", which are remnants of absorbed xenoliths of gneiss, are common (Fig. 12).

The post-tectonic granitoids occur as relatively small bodies which range in composition from adamellite to tonalite. They truncate tectonic and lithological trends in the country rocks (Figs. 2 and 3). Original igneous mineral assemblages and hypidiomorphic granular texture are preserved; the plagioclase characteristically exhibits complex oscillatory zoning.

TECTONO-STRATIGRAPHIC MODELS

Contacts between domains A and B in the supracrustal rocks are not exposed; therefore, relationships between these domains cannot be determined with certainty. However, it is possible to account for the sharply contrasting deformation styles in these domains by invoking one of two alternative models.

(i) The domains represent supracrustal sequences of different ages and are separated by an unconformity; the isoclinal fold phase recorded in domain A predates deposition of rocks in domain B (Fig. 13). In the Dalgaranga region, the lower (domain A) and upper (domain B) sequences may be broadly concordant. However, in the Weld Range — Ryansville area the situation is more complex. Trends in the lower sequence are not consistent to the south and southwest of Ryansville, where the rocks have been affected by the phase of late folding and are surrounded by granitoids (Fig. 2). A major unconformity of the type proposed here has been identified in the Warriedar greenstone belt in the southern Murchison Province (Baxter and others, 1980; Lipple and others, 1980; Seccombe and Frater, 1981, Fig. 1).

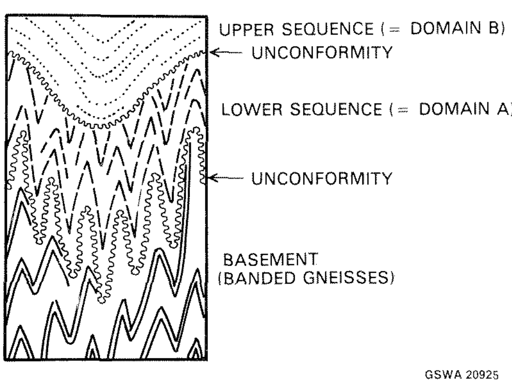

Figure 13. Schematic representation of tectono-stratigraphic model I.

(ii) The deformation in both domains is essentially contemporaneous, and they represent different tectonic levels within a single heterogeneously deformed supracrustal sequence (Fig. 14). They may be separated by a tectonic dislocation within this sequence, as indicated by the dashed line in Figure 14, or there may be a sharp gradation in deformation intensity between the two domains. Hickman (1975) and Hickman and Lipple (1975, Fig. 6) have described similar structures in the Pilbara Block, although no documented examples exist in the Yilgarn Block.

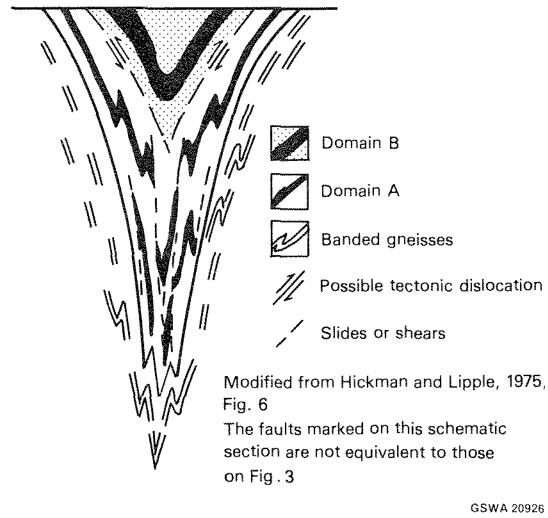

Figure 14. Tectono-stratigraphic model II.

In each of the models described above, it is possible to divide the supracrustal rocks into a lower (domain A) and an upper (domain B) sequence separated by the domain boundary. In the second model (II) this boundary has much less stratigraphic significance than in the first (I); however, it still separates rocks higher in the supracrustal sequence from those lower in it. In either case a sialic basement of branded gneiss is likely (Figs 13 and 14).

The relative age of the recrystallized granitoids is uncertain. Marginal foliations in these granitoids cut across fabrics associated with isoclinal folding in the greenstone belts in some parts of the area—*e.g.* at the western margin of the greenstone belt southeast of Moyagee (Figs 2 and 3). Thus, if the first model is correct, the foliations in the recrystallized granitoids probably formed during the open-close fold phase. If the second model is correct, these foliations probably formed during the phase of heterogeneous deformation of greenstone belts (isoclinal and open-close folding): however, a problem is posed here by the cross-cutting foliations mentioned above. The two later fold phases (north and northwest axial trends) are of insufficient intensity to impart a strong foliation to the recrystallized granitoids. The second model seems less likely in the light of these relationships. The intrusive granitoids are omitted from Figures 13 and 14 for simplicity.

SUPRACRUSTAL ROCK TYPES

In domain A (lower sequence) there appear to be three major mafic units (Fig. 2): a dominantly komatiitic unit; a *dominantly tholeiitic unit*; and a unit containing abundant banded iron-formation with intervening mafic and ultramafic rocks. Where deformation is particularly intense and protoliths are thus difficult to recognize, rocks are assigned to a unit of "mafic and ultramafic schist" on Figure 2. Facing evidence (differentiation in conformable layered gabbro sills) in the Moyagee and Tuckabianna areas indicates that the komatiitic unit overlies the unit with common banded iron-formation. The stratigraphic position of the tholeiitic unit is unknown. Felsic volcanics and sediments are much less common than mafic rocks in the lower sequence. The only major occurences are at Eelya Hill, Cuddingwarra, west of Moyagee, Trenton Hill, and the Weld Range. At Eelya Hill, an 8 km-thick (possibly tectonically thickened) sequence of strongly sheared rhyodacitic schists with minor metasediments is exposed. Sheared and disrupted mafic to ultramafic sills occur throughout the succession. In the Cuddingwarra area, a 1 km-thick unit of strongly sheared rhyodacitic schist occurs. Ten kilometres west of Moyagee, a 0.5 km-thick unit of sheared, poorly sorted, coarse sandstone occurs within an extensive mafic sequence. This unit is thought to extend northwards across Lake Austin to the Trenton Hill area, where sandstone units are exposed as pavements at the margin of the salt lake. These occurrences indicate that an exposed sialic terrain existed during development of the lower greenstone sequence. It is unlikely that this terrain was volcanic, as many of the sandstone units comprise about 70% large (up to 3 mm, average 1 mm) quartz grains with occasional quartz pebbles up to a few centimetres across; the matrix is a mixture of fine-grained quartz, sericite and kaolin (after feldspar). In the Weld Range, thick (up to 150 m) jaspilitic banded iron-formation, fine-grained felsic volcanics and volcaniclastics, and minor mafic volcanics are intruded by multiple dolerite sheets with little apparent disruption of bedding.

In domain B (upper sequence), the major mafic unit is a sequence of Komatiitic and tholeiitic volcanics, 3.5 km thick, which occurs at Ryansville. The unit is dominantly basaltic with pillowed flows and breccias (hyaloclastites): an ultramafic komatiite with bladed spinifex texture forms the base. The volcanics differ from komatiitic rocks in the lower sequence in that pillows and breccias are common. At Dalgaranga, three major depositional units occur. A 1.5 km-thick unit of pelitic and psammitic rocks, within which thin, banded iron-formation and mafic units are common, occurs at the base of the sequence. A unit of finely layered dacite tuff about 1 km thick overlies this. A coarse-grained, poorly sorted volcaniclastic unit, 300 to 400 m thick, derived directly from the underlying dacites, marks the top of the sequence. A thick, multiple gabbro sill intrudes the sequence (Fig. 2).

The fact that at least 50% of the basaltic rocks in the Cue region are komatiitic is significant, as it had previously been thought that this rock type was rare in the Murchison Province; this has been used as a major lithological criterion in distinguishing the province from the neighbouring Southern Cross Province, in which komatiitic rocks are common (Gee and others, 1981). Komatiitic rocks were not identified during the original GSWA mapping from 1966 to 1968 (de la Hunty, 1973) because these rocks were first documented in southern Africa in 1969 (Viljoen and Viljoen, 1969a, 1969b) and later in Western Australia (Nesbitt, 1971).

PREVIOUS STRATIGRAPHY

De la Hunty (1973) proposed a sequence of three litho-stratigraphic formations in the Cue region, the Moyagee, Cuddingwarra, and Ryansville Formations (Table 1). It seems unlikely, from recent mapping by the authors, that this scheme is tenable. For example, all the felsic volcanics and sedimentary rocks (with the exception of banded iron-formation and chert units) were grouped in one formation—the Cuddingwarra. Yet, from the preceding discussion it is evident that the occurrence of felsic and sedimentary lithologies span the chronological development of both lower (domain A) and upper (domain B) sequences. De la Hunty (1973) thought the Ryansville outcrop represented part of an upper mafic sequence; this is compatible with the models presented here; however, he also thought the rocks in the Tuckabianna fold were equivalent to these. No evidence is given by de la Hunty (1973) to support a major threefold division.

TABLE 1. De La HUNTY'S (1973) CUE STRATIGRAPHY

Name	*Description*
Ryansville Formation	Mafic extrusive and intrusive rocks; thickness=1524 m (5 000 ft)
Cuddingwarra Formation	Felsic volcanics and sediments minor mafic rocks, banded iron-formation; thickness=2 438 m (8 000 ft)
Moyagee Formation	Mafic and ultramafic extrusive and intrusive rocks, banded iron-formation, minor sediments; base of formation intruded by granite; thickness=4 267 m (14 000 ft)

CONCLUSIONS

(a) Two tectonic domains can be distinguished in the supracrustal rocks. It is likely that these have stratigraphic significance and represent older and younger sequences.

(b) Occurrences of complexly deformed, banded, granitoid gneiss probably represent remnants of a sialic basement to the granite-greenstone system.

(c) Two phases of granitoid intrusion are recognized: an early and volumetrically dominant phase, now recrystallized and partly foliated; and a later postectonic phase which retains original igneous features.

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THE FRASER COMPLEX—A MAJOR LAYERED INTRUSION IN WESTERN AUSTRALIA

by John S. Myers

ABSTRACT

A survey of the Fraser Complex shows that most of the pyroxene granulites are derived from basic igneous rocks and form part of a major layered intrusion. No evidence was found to substantiate the interpretation of Wilson (1969a), supported by Tyrwhitt and Orridge (1975), and Bunting and others (1976), that most of the pyroxene granulites represent metamorphosed volcanic rocks.

The Fraser Complex is redefined to include only the layered basic plutonic rocks, and is divided into five tectonostratigraphic units. Contacts of the units are zones of intense deformation, and the basic plutonic rocks are tectonically interleaved with metasedimentary rocks mainly quartzite and gneiss. All these rocks have been intruded by sheets of porphyritic granite and pegmatite, strongly deformed, and metamorphosed to granulite facies. Deformation was heterogeneous, and parts of the Fraser Complex preserve undeformed igneous textures and igneous cumulate mineral assemblages.

INTRODUCTION

The Fraser Complex is part of the Albany-Fraser Province (Doepel, 1975; Gee, 1979), a Proterozoic mobile belt, over 1 100 km long and 200 km wide, which extends along the southern and southeastern margin of the Archaean Yilgarn Block (Fig. 1). The mobile belt consists of strongly deformed ortho- and paragneisses of amphibolite or granulite facies, mafic dykes, and granite intrusions. The structures and intrusions of the mobile belt cut sharply across structures of the Yilgarn Block and, in the region west of Hopetoun, high-grade metasedimentary rocks are thrust northwestwards from the mobile belt over lower grade metasedimentary rocks on the edge of the Yilgarn Block.

Rocks from the Albany-Fraser Province have given Rb-Sr ages in the range 1.8-1.1 Ga (Turek and Stephenson, 1966; Arriens and Lambert, 1969; Doepel, 1975; Bunting and others, 1976; Thom and others, 1981). Most of these ages fall into 3 groups which appear to be related to the following sequence of events:

(a) An early metamorphic and tectonic episode 1.8-1.6 Ga ago involving gneisses, granitoid rocks, metasedimentary rocks, and the Fraser complex. This is the first metamorphic and tectonic episode recognized in the Mount Barren metasedimentary rocks (Thom and others, 1981), and appears to be associated with tectonic interleaving, at a deep crustal level, of stratigraphic units within the Fraser Complex and Mount Barren Group.

(b) A high-grade (granulite facies) metamorphic and tectonic episode 1.3-1.2 Ga ago associated with the emplacement of porphyritic granite sheets, and the folding of foliation and layering of the Fraser Complex.

(c) An episode of granite intrusion, low-grade metamorphism, and high-level deformation about 1.1 Ga ago.

The name Fraser Complex was proposed by Doepel (1973) for the belt of pyroxene granulites, about 32 km wide and over 185 km long, which extends from the Fraser Range, east of Norseman, to Zanthus (Fig. 1). The outcrop of these rocks is shown on the 1:250 000 maps, Zanthus (Doepel and Lowry, 1970a), Balladonia (Doepel and Lowry, 1970b), Norseman (Doepel, 1973), and Widgiemooltha (Sofoulis, 1966). The northeastern extent of these rocks, which is mainly below younger cover, was outlined by Bunting and others (1976). The Fraser Complex, defined by Doepel (1973), consists of basic pyroxene granulites which contain veins and bands of acid granulite.

Doepel (1973) and Wilson (1969a, 1969b) recognized "remnants of gabbro" within the pyroxene granulites. Wilson (1969b) mapped some of these gabbro remnants as a flat-lying sheet, which he called the "East Fraser Gabbro" and considered to be intrusive into the more widespread pyroxene granulites. He described the petrology of the "East Fraser Gabbro" from outcrops along the Eyre Highway; he also observed that the marginal zones of this gabbro were metamorphosed to granulite facies, and speculated that some of the homogeneous pyroxene

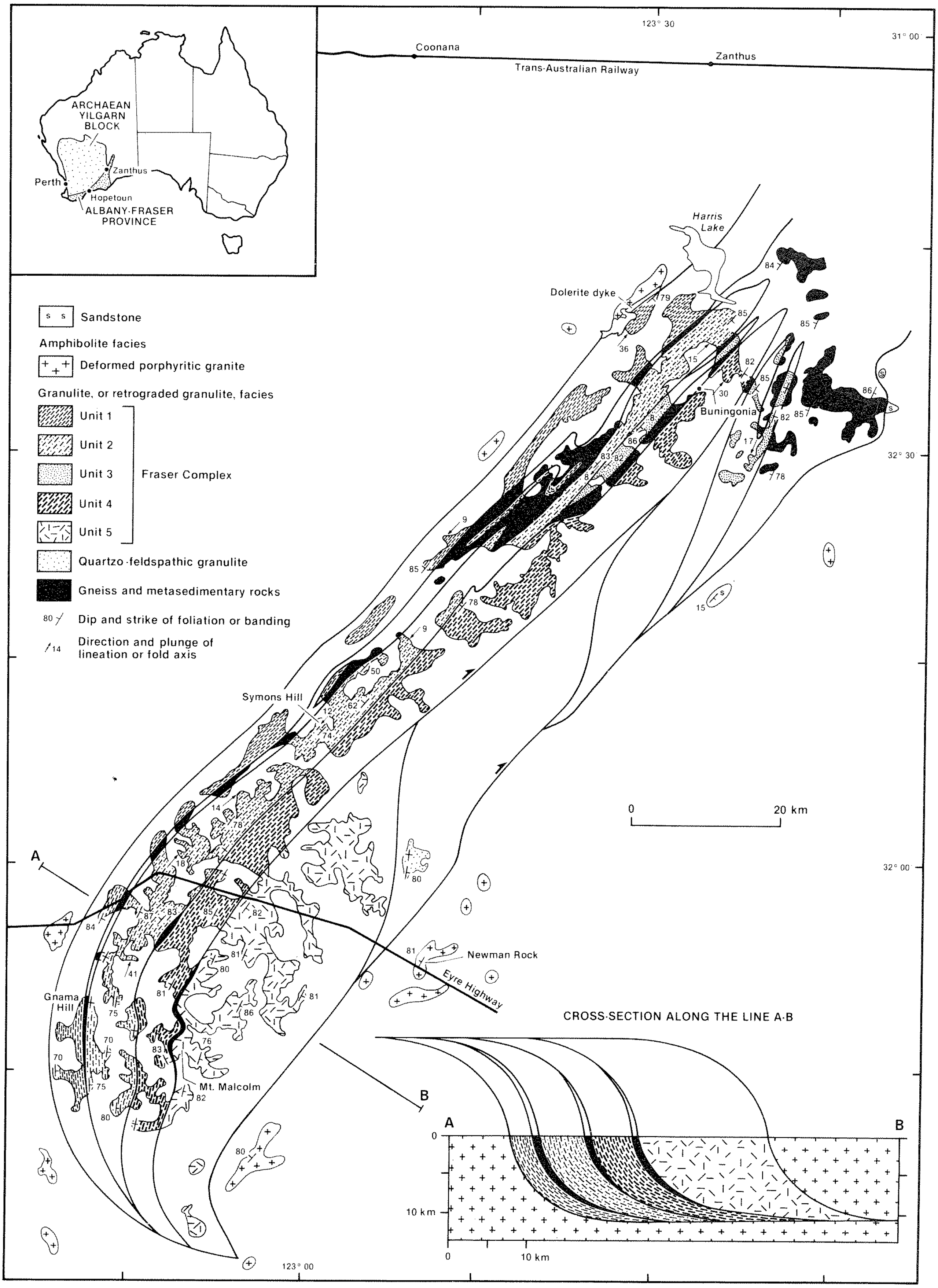


Figure 1. Simplified geological outcrop map and schematic cross-section of the Fraser Complex.

granulites could represent metamorphic equivalents of similar intrusive gabbroic rocks. However, he considered that most of the granulites were derived from metamorphosed volcanic rocks, a view supported by Bunting and others (1976), but contested by Doepel (1973). Wilson (1969b) interpreted pyroxenite layers as the "decayed tops of basalt flows"; concentric jointing as pillow-lava structures; ellipsoids of quartz and feldspar as relic amygdules; and thin quartzo-feldspathic layers as either sandy intercalations in a mafic volcanic pile, acid volcanic rocks, minor intrusions, or metasomatized basic rocks.

Newmont Pty Ltd explored the Fraser Complex for nickel between 1965-70 but no significant mineral deposits were found (Tyrwhitt and Orridge, 1975). The work included an aeromagnetic survey, geological mapping, regional soil sampling, and shallow reconnaissance drilling.

The following account and map (Fig. 1) are based on 21 days of reconnaissance mapping of the Fraser Complex during 1982-83. The Fraser Complex is here redefined to include only the metamorphosed basic igneous rocks, and to exclude intercalated quartzo-feldspathic gneisses, metasedimentary rocks, and deformed porphyritic granites and pegmatites, which, together with the basic igneous rocks, are part of the Albany-Fraser province (Doepel, 1975).

FRASER COMPLEX

The Fraser Complex outcrops in a belt 30 km wide and 200 km long, and can be traced as large positive gravity and magnetic anomalies for a further 250 km to the northeast. It lies within, and is parallel to, the regional trend of the Albany-Fraser Province (Bunting and others, 1976, Fig. 63). The structure is superficially simple: The complex, and all layers and foliations within it, strike northeast and are vertical or dip steeply to the northwest or southeast. Lineations and fold axes plunge northeast or southwest.

The Fraser Complex can be divided into five major tectonostratigraphic units (Fig. 1), which are described below from northwest to southeast.

Unit 1

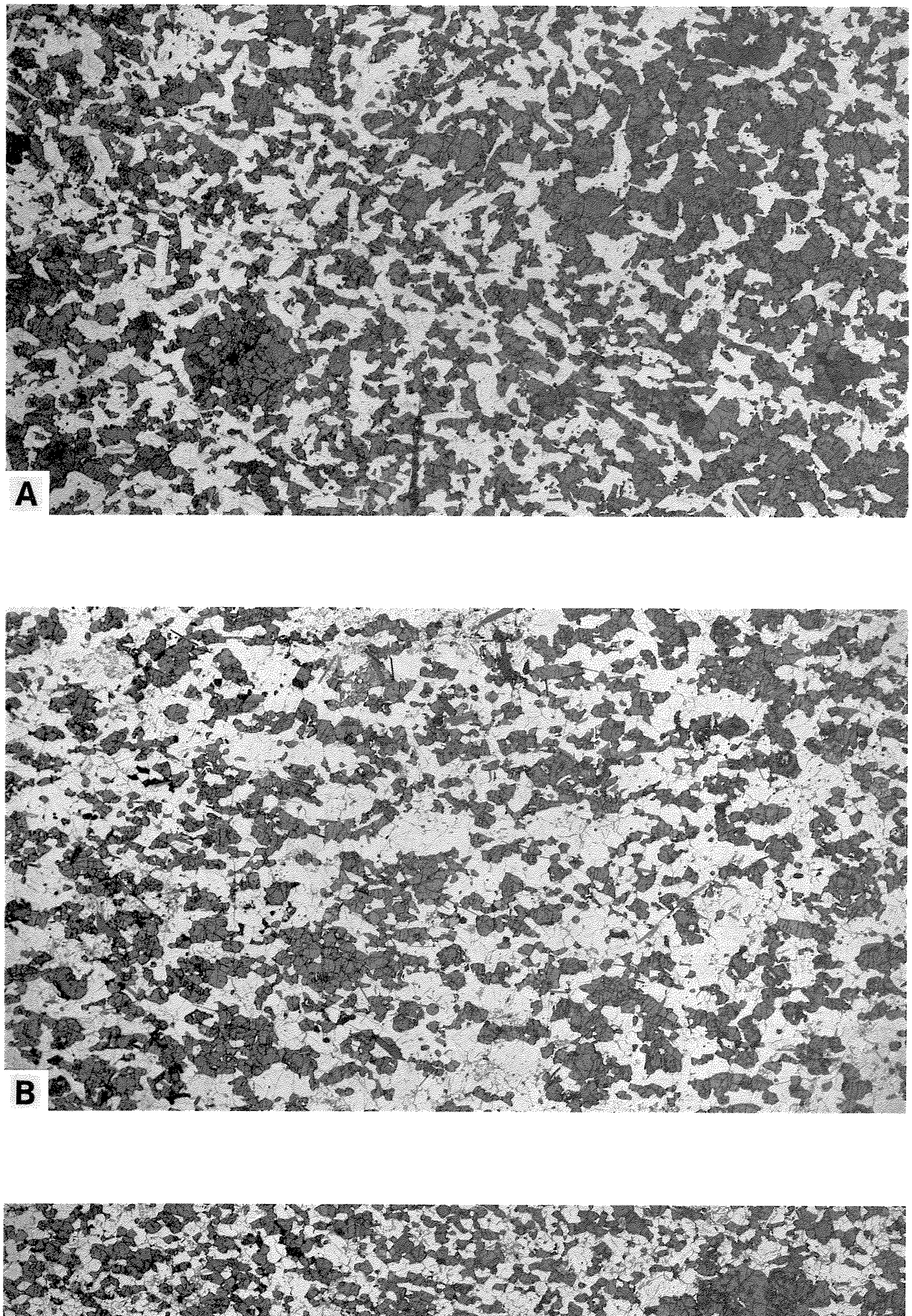
Unit 1 forms the western part of the Fraser Complex and is a steeply east-dipping sheet between 3 and 6 km thick (Fig. 1). It consists mainly of garnet amphibolite (interpreted as garnet metagabbro) and thin layers of metamorphosed ultramafic rocks, melanogabbro, and anorthosite. All these rocks are strongly deformed and show macroscopic foliation and (locally) minor folds, but they generally have coarse-grained, post-tectonic metamorphic textures. Most of the garnet amphibolites consist of plagioclase, green hornblende, garnet, and epidote. Relic igneous textures are only locally preserved, but show that the amphibolite is derived from metagabbro (probably olivine, pyroxene, plagioclase cumulate). The leucogabbro and anorthosite layers consist mainly of plagioclase, clinopyroxene, and orthopyroxene (with or without garnet), which form post-tectonic, granoblastic textures and may represent mainly plagioclase cumulate igneous layers. Ultramafic layers consist mainly of metamorphic-textured orthopyroxene and green spinel, and may be derived from orthopyroxene(-olivine) cululates. Newmont Pty Ltd discovered a sheet of pyroxenite and serpentinized dunite (2 500 m long and up to 250 m wide) by drilling within this unit (Tyrwhitt and Orridge, 1975).

Unit 2

Unit 2 forms a sub-vertical sheet between 2 and 6 km thick, east of unit 1, and mainly comprises basic pyroxene granulite—the rock type previously mapped as forming most of the Fraser Complex. It is generally medium grained, has a granoblastic texture, and contains about 60% plagioclase, 20% orthopyroxene, 15% clinopyroxene, 3% biotite, and 2% magnetite. Some of the minerals are elongate and are aligned parallel to the macroscopic foliation. Relic igneous texture is locally preserved and shows that most of these rocks are derived from gabbros or norites. Many outcrops show relic igneous layering, and two kinds occur: compositional layering ranging from metamorphosed melanogabbro to gabbro and leucogabbro; and size-controlled layering marked by variations in the grain sizes of relic cumulus pyroxenes. No relic, size- or mineral-graded layering was observed; each layer is generally uniform and has symmetrical boundaries, which may be either gradational or sharp. The rocks are generally strongly deformed, and relic igneous layering forms tight or isoclinal folds (such as on Symons Hill). The main macroscopic fabric is axial planar to these folds and is sub-parallel to the long limbs of the folds. Most relic igneous layering is therefore locally transposed into the main tectonic fabric.

Unit 3

Unit 3 occurs to the south and southeast of Harris Lake as a tightly folded, steeply dipping slab between 1 and 2 km thick, (Fig. 1). It is enclosed by quartzite and gneiss except in the southwest where it appears to be in contact with unit 2. It consists mainly of metamorphosed leucogabbro, layers of anorthosite, and minor layers of gabbro and melanogabbro. The contacts between the layers are either sharp or gradational, but no asymmetric graded layering was observed. Most rocks are strongly deformed and show macroscopic foliation parallel to compositional layering; but recrystallization outlasted deformation, and the rocks generally show medium-grained equigranular metamorphic textures.



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Unit 4

Unit 4 occurs as a sub-vertical sheet between 5 and 6 km thick. Along most of its outcrop, it lies adjacent to, or is separated by a thin layer of quartzite from, unit 2 to the west (Fig. 1). South of Harris Lake, it is separated from unit 2 by unit 3 and a thick layer of quartzite. The main rock type and features of unit 4 are similar to those of unit 2.

Unit 5

Unit 5 forms the eastern part of the Fraser Complex and outcrops as a steeply east-dipping sheet up to 16 km thick. Its eastern margin is not exposed. It consists mainly of gabbro and metagabbro with a little-deformed relic igneous texture. Igneous minerals are widely preserved and consist of cumulus orthopyroxene, plagioclase, green spinel, and clinopyroxene; and intercumulus plagioclase. Olivine occurs as remnants within cumulus orthopyroxene, and green spinel also occurs as an exsolution product within orthopyroxene. Biotite generally occurs as an alteration product of orthopyroxene. In many cases, the igneous minerals are clouded. The "eucrite" described by Wilson (1969b) is part of this unit; but it is mainly an orthopyroxene cumulate with intercumulus plagioclase and not a typical rock type.

Gradational sequences can locally be seen between these rocks and strongly deformed metagabbros, similar to those which comprise most of unit 2, in which igneous textures are extensively obliterated (Figs 2A, B, C). These local zones of strong deformation are parallel to the strike of the whole unit and the main foliation of the other units; and some of them enclose lenticular bodies of little-deformed or undeformed metagabbro.

Geochemistry

Thirty-two whole rock samples were analyzed for major and trace elements by XRF to determine the composition of the main rock types and any igneous differentiation trends which might provide evidence of the original "way-up" of the major stratigraphic units. Representative analyses are given in Table 1, and igneous differentiation trends of stratigraphic units are outlined on an AFM diagram (Fig. 3). The whole suite of analyzed samples shows a clearly defined trend towards iron enrichment, which is broadly similar to that shown by other major layered intrusions that crystallized from basaltic magma (Fig. 3).

The rocks of unit 5 plot as a group and have generally lower FeO/MgO ratios than those from other units, and they therefore appear to be derived from a less evolved magma. If the five main stratigraphic units represent parts of a single-layered intrusion derived from one body of magma, then unit 5 would be interpreted as the lowest structural unit. Rocks from unit 3 plot in a group and show the highest FeO/MgO ratios; these rocks appear, therefore, to be derived from the most evolved magma. Rocks from units 1, 2 and 4 plot as a group with an intermediate position between, and partly overlapping with, units 3 and 5.

The different tectonostratigraphic units could have been derived from different magmas, but if all five units are assumed to be the crystallization products of a single body of magma, then the FeO/MgO ratios suggest that the original order of crystallization and deposition was firstly unit 5, then 4, units 1 and 2, and lastly unit 3.

ASSOCIATED ROCKS

Gneiss

Gneiss occurs interlayered with quartzite and metamorphosed leucogabbro in the northeastern part of the Fraser Complex (Fig. 1). It shows banding defined by variations in concentrations of mafic minerals (mainly biotite) and by alternations of quartzofeldspathic and amphibolite layers. It is strongly deformed, fine to medium grained, and shows a greater variety of rock types and a more complicated geological history than the recrystallized granites described later.

Metasedimentary rocks

Metasedimentary rocks are interlayered with metamorphosed basic igneous rocks within the Fraser Complex (Fig. 1). They occur as thin layers of great

Figure 2. Progressive stages in the deformation and recrystallization of gabbro of unit 5 on Mount Malcolm. The sections shown are 25 mm wide.

A. Undeformed, partly recrystallized gabbro with laths of igneous plagioclase (An_{75}) set in large poikilitic orthopyroxene and clinopyroxene crystals. Olivine is partly altered to iddingsite. (GSWA sample 74644).

B. Weakly deformed, mainly recrystallized gabbro with relic plagioclase laths recrystallized to granular mosaics (An_{65}), and mainly metamorphic orthopyroxene, clinopyroxene and biotite. (GSWA sample 74641).

C. Strongly deformed and completely recrystallized gabbro with granoblastic polygonal texture of plagioclase (An_{55}), orthopyroxene and clinopyroxene. The upper portion is less deformed than the lower portion and shows aggregates of granoblastic plagioclase defining relic igneous laths, tectonically rotated into parallelism (GSWA sample 74643).

TABLE 1. MODAL AND CHEMICAL ANALYSES OF METAGABBRO SAMPLES FROM THE FRASER COMPLEX

GSWA Sample No.	Unit 1 74626	Unit 2 74615	Unit 3 74704	Unit 4 74665	Unit 5 74652
		Minerals (%)			
K-feldspar	2	7	2	2	
Plagioclase	60	58	62	62	65
Olivine					3
Orthopyroxene	19	17	17	17	12
Clinopyroxene	16	13	17	17	11
Hornblende	1				4
Biotite		1			3
Opaques	2	4	2	2	2
		Major oxides (wt%)			
SiO_2	50.09	50.00	53.77	50.95	49.21
TiO_2	1.35	1.50	1.63	1.04	0.99
Al_2O_3	15.75	15.83	14.00	16.16	17.72
Fe_2O_3	1.41	1.63	2.05	1.83	1.37
FeO	10.00	9.34	10.30	8.67	8.59
MnO	0.30	0.25	0.28	0.21	0.18
MgO	7.17	7.68	4.58	7.78	8.78
CaO	9.54	9.67	7.95	9.87	9.86
Na_2O	2.45	2.37	2.71	2.49	2.40
K_2O	0.53	0.69	1.35	0.39	0.46
P_2O_5	0.16	0.30	0.40	0.11	0.10
H_2O^-	0.60	0.59	0.49	0.46	0.59
H_2O^+	0.23	0.16	0.12	0.16	0.14
CO_2	0.07	0.05	0.03	0.18	0.15
Total	**99.65**	**100.06**	**99.66**	**100.16**	**100.53**
		Trace elements (ppm)			
Li	17	18	16	10	5
Rb	11	10	33	5	12
Sr	182	232	158	192	181
Ba	157	299	508	204	199
Zr	84	120	155	69	84
La	26	39	27	10	15
Ce	42	61	85	25	27
Sc	38	32	34	37	30
V	211	187	185	207	181
Cr	285	271	48	297	254
Co	55	65	55	75	70
Ni	40	85	20	75	150
Cu	100	120	110	140	130
Zn	94	105	155	96	81
Ga	15	14	17	15	13

lateral extent, and are especially prominent between the five major units of the Fraser Complex. They also form a thicker sequence to the east and southeast of Harris Lake. Most of these metasedimentary rocks are coarse-grained quartzites, with minor amounts of pelitic and semi-pelitic rocks comprising quartz, sillimanite, mica, feldspar, garnet, and amphibole, and thinly banded quartz-magnetite rocks. They are strongly deformed and show mainly post-tectonic mineral assemblages of granulite or amphibolite facies.

Recrystallized granite and pegmatite

Recrystallized granite and pegmatite occur throughout the Fraser Complex as thin sheets parallel or sub-parallel to the main foliation of the complex. Most granites are coarse grained, porphyritic, and contain hypersthene and biotite. Some granites and most pegmatites contain garnet, and some pegmatites are zoned and have aplitic margins.

The granite and pegmatite cut across relic igneous layering of the Fraser Complex and post-date an episode of strong deformation of the Fraser Complex, metasedimentary rocks, and gneisses. They were subsequently strongly deformed together with these rocks and now possess the same main tectonic fabrics as the Fraser Complex. The zoned margins of some pegmatites and the discordant contacts of some granites suggest that these granitic rocks were intruded as sheets into the other rocks before or during the deformation which formed the main tectonic fabrics of the Fraser Complex.

Quartzo-feldspathic granulite

A fine-grained, uniform, hypersthene-bearing quartzo-feldspathic granulite occurs to the southeast of the Fraser Complex (Fig. 1). The rock shows a mineral lineation and a pronounced foliation parallel to the main fabrics of the Fraser Complex. The origin of the rock is not clear, but it may have been a granitoid plutonic rock.

Sandstone

Sandstone, which is undeformed and not obviously metamorphosed, occurs to the northeast of the Fraser Complex (Fig. 1). It shows a greater variety of composition and grain size than the deformed and

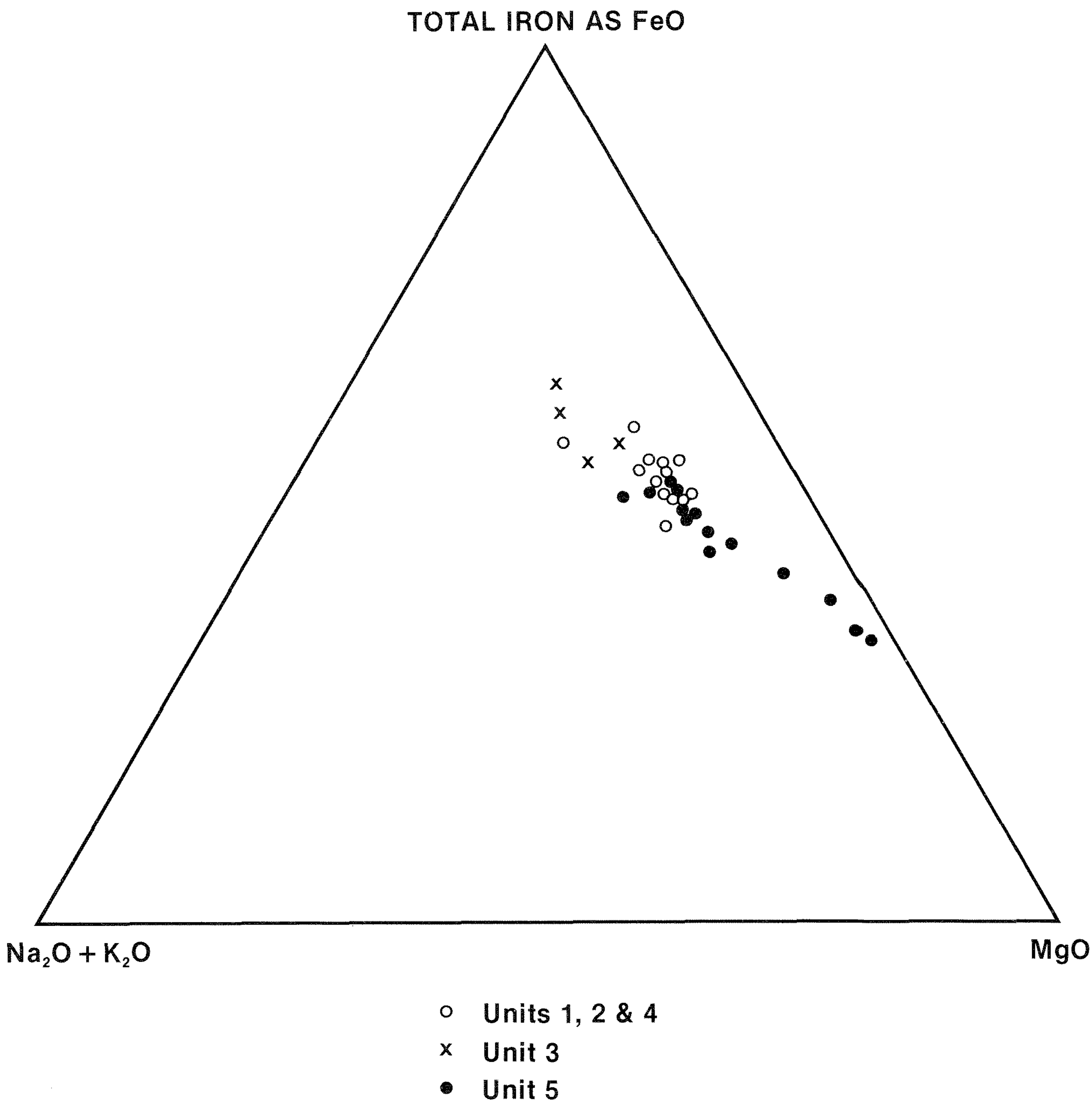

Figure 3. A-F-M diagram showing a linear trend towards iron enrichment.

metamorphosed quartzite (which is mainly orthoquartzite), and it generally contains feldspar and is arkosic. It appears to post-date the main deformation and metamorphism of the Fraser Complex.

TECTONIZED BOUNDARIES OF THE FRASER COMPLEX

The boundaries of the Fraser Complex are concordant with the main regional foliations and are zones of intense deformation.

The northwest margin of the Fraser Complex was only seen exposed to the southwest of Harris Lake (Fig. 1). There, rocks of the Fraser Complex occur as lenses and pods of gabbro in flaggy quartzo-feldspathic gneiss and as thin layers of amphibolite tectonically interleaved with quartzo-feldspathic rocks forming amphibolite-banded gneiss or schist. All the rocks are intensely deformed in a zone 2 km wide; they are finely banded, fine-grained, schistose or mylonitic, and have been recrystallized in the epidote-amphibolite or garnet-amphibolite facies. They show complex deformation, and the schistosity is locally folded into small folds which plunge at moderate angles to the northeast.

These rocks are cut by an undeformed, vertical dolerite dyke, 500 m wide (Fig. 1), which lies along strike from mapped outcrops of the Binneringie dyke on the Widgiemooltha sheet to the southwest (Sofoulis, 1966). The dyke cuts—and encloses xenoliths of—strongly deformed flaggy gneiss, amphibolite, and augen granite. It is cut by sheets of muscovite-bearing granite, pegmatite, and aplite, which show a lineation plunging moderately to the northeast, similar in orientation to late lineations and fold axes in the flaggy gneiss. To the northwest, the flaggy gneiss is suceeded by strongly deformed porphyritic granite with a pronounced foliation parallel to that of the flaggy gneiss, then by less-deformed porphyritic granite and coarse, even-grained, granite.

The westernmost outcrops of the Fraser Complex along the Eyre Highway also show intense deformation of strips of garnet amphibolite in quartzo-feldspathic gneiss, and the intensity of this deformation may reflect the proximity of these outcrops to the northwestern margin of the Fraser Complex. Much of the quartzo-feldspathic gneiss at this locality appears to be derived from rocks intrusive into the garnet amphibolite, which occurs as agmatite. The rocks were strongly deformed together and are locally fine grained and mylonitic. The deformation was associated with retrogression from granulite to garnet-amphibolite facies, and recrystallization generally outlasted the deformation.

The northwestern margin of the Fraser Complex was thus a zone of intense ductile deformation and retrogression from granulite to amphibolite facies, but the name Fraser Fault introduced for this boundary by Wilson (1958) is misleading. Most of the deformation was ductile and occurred at a deep crustal level.

The southeastern margin of the Fraser Complex is not exposed. The fine-grained quartzo-feldspathic granulites to the east of unit 5 show similar tectonic and metamorphic fabrics to those in the Fraser Complex and appear to have shared the same episode of deformation and granulite-facies metamorphism. However, in contrast, the adjacent rocks further to the southeast, which are best exposed at Newman Rock (Fig. 1), do not appear to have been metamorphosed to granulite facies. They consist of metasedimentary rocks intruded by a moderately deformed porphyritic granite, all metamorphosed to amphibolite facies. Xenoliths of metasedimentary rocks within the granite are aligned parallel to a weakly developed foliation at a high angle to the trend of the Fraser Complex. This foliation is cut by sinistral ductile shear zones in which mylonites formed in amphibolite facies parallel to the main (granulite facies) foliation of the Fraser Complex. Therefore all the rocks appear to have shared the same episode of deformation but at different metamorphic grades. The porphyritic granite at Newman Rock does not appear to have been metamorphosed above amphibolite facies, and yet its oldest tectonic fabrics pre-date the main foliation and associated granulite—facies metamorphism of the Fraser Complex. There may therefore be a major tectonic dislocation between the rocks in the vicinity of Newman Rock and the quartzo-feldspathic granulites to the northwest.

DISCUSSION AND CONCLUSIONS

The igneous composition of the Fraser Complex; the widespread presentation of relic igneous textures and, in unit 5, of igneous minerals with cumulus textures; and the layering of metamorphosed anorthosite, leucogabbro, gabbro, melanogabbro, and ultramafic rocks indicate that the Fraser Complex is part of a major layered basic intrusion.

The absence of basic volcanic rocks and sheeted-dyke complexes suggests that the intrusion was not the lower part of a slice of oceanic crust, unless it was detached from such crust.

The intimate association of gneiss and metasedimentary rocks suggests that the Fraser Complex was either intruded into, or tectonically interleaved with sialic crust

The great extent of the relatively very thin layers of metasedimentary rocks and their occurrence only between, and not within, the major tectonostratigraphic units of the Fraser Complex, suggest that the metasedimentary rocks are not thin rafts of host rocks spalled off during the intrusion of the Fraser Complex (unless the Complex was intruded as a number of sill-like bodies). Rather, the relative size, shape, and position of the metasedimentary layers, combined with the locally intense heterogeneous deformation concentrated along these layers and the immediately adjacent igneous rocks suggest that the Fraser Complex was disrupted by sub-horizontal movements, and was tectonically interleaved with metasedimentary rocks and associated quartzo-feldspathic gneiss by thrusting. The total apparent thickness of the Fraser Complex is unusually great, and so zones of dislocation may also occur within the major units, by which originally thinner units may be stacked as thrust slices.

The ductile nature of the deformation along the boundaries of rock units suggests that tectonic interleaving occurred at depth in the crust (Fig. 4). At the surface, these thrust contacts are now steeply dipping, but the interpretation of gravity data by Everingham (1964) that the pyroxene granulites extend to a depth of only 10 km, suggests that at depth the thrusts are listric (Fig. 1, Section A-B). The overall structure of the Fraser Complex and associated rocks shown by the map (Fig. 1) does not support its interpretation by Gee (1979) as a large synform.

The lenticular bodies of ultramafic rocks and associated norites discovered by Newmont Pty Ltd within and along the southeastern margin of unit 1

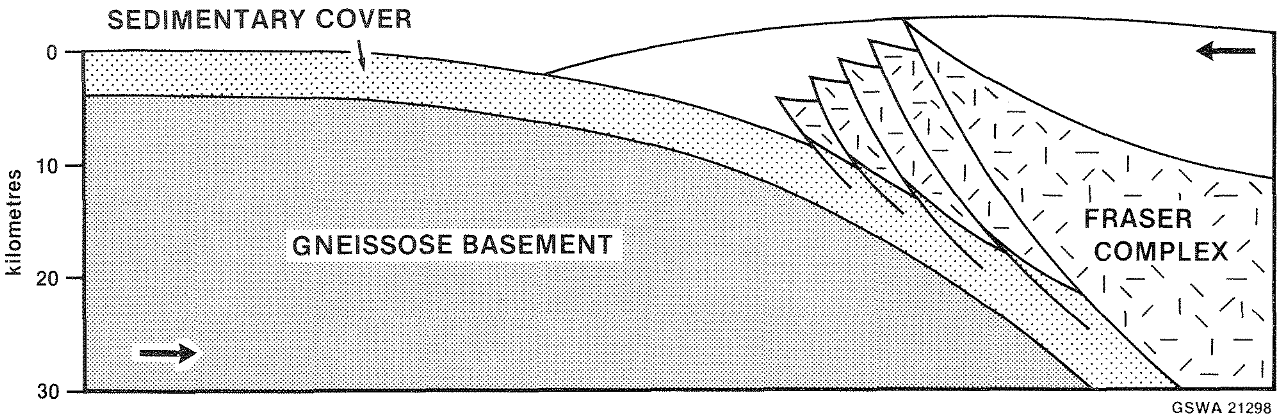

Figure 4. Diagrammatic section showing a mechanism for the tectonic interleaving of the Fraser Complex with metasedimentary rocks and gneisses (?1.8-1.6 Ga ago).

could be tectonically emplaced slivers rather than *in situ* intrusions or cumulate layers. The mapped example presented by Tyrhwitt and Orridge (1975, Fig. 2) shows that these ultramafic bodies and associated norite occur in a complex zone of interleaved quartzites and basic pyroxene granulites.

The Fraser Complex and interleaved gneiss and metasedimentary rocks were subsequently intruded by sheets of porphyritic granite and pegmatite and were all strongly deformed together. They developed tectonic fabrics in granulite facies which indicate that the deformation occurred at a deep crustal level. Metamorphism outlasted the deformation, but post-tectonic mineral assemblages are also mainly in granulite facies.

The northwestern margin of the Fraser Complex continued to be a zone of tectonic activity; intense deformation in this zone was accompanied by retrogression to amphibolite facies and led to the formation of flaggy gneiss, schist, and mylonite, probably associated with movement of these rocks to a higher crustal level. The adjacent, older, granite complexes both to the northwest, west of Harris Lake, and to the southeast, at Newman Rock, do not appear to have been metamorphosed above granulite facies. This suggests that the Fraser Complex, together with the interleaved gneiss, metasedimentary rocks, and quartzo-feldspathic granulite, was emplaced as a large, composite, tectonic slice from a deep crustal level, where granulite assemblages were stable, into amphibolite facies rocks at a higher crustal level.

The exposed Fraser Complex may represent only part of a major layered intrusion. It has been severely disrupted and distorted by deformation, and its original location and thickness are unknown. The maximum age of the Fraser Complex is also unknown. A Rb-Sr whole-rock isochron of 1 328 ± 12 Ma determined by Arriens and Lambert (1969) from granulite-facies rocks of the Fraser Range is mainly influenced by granitoid rocks which are intrusive into the basic igneous rocks of the Fraser Complex.

Although the outcrops of the Fraser Complex may represent only part of the layered sequence of a major intrusion, they are an important suite of rocks. They are a major component of the Albany-Fraser Province, and the interpretation of their geological history is therefore of considerable importance in understanding the tectonic evolution of this mobile belt. In additon, the Fraser Complex presents a sizeable exploration target for chromite and platinum.

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A RUBIDIUM-STRONTIUM DATE FROM FELSIC VOLCANICS WITHIN THE MOUNT ROE BASALT OF THE WYLOO DOME

by J. R. de Laeter*, D. B. Seymour, and W. G. Libby

ABSTRACT

A thin felsic volcanic unit within predominantly basic volcanic rock in the Mount Roe Basalt, low in the Fortescue Group of the Hamersley Basin at Wyloo Dome, has yielded a Rb-Sr whole-rock date of 2 032 $\pm$148 Ma with an R_i of 0.748 8 $\pm$0.009 4. Comparison with other dating in the Hamersley Basin, a large initial $^{87}Sr/^{86}Sr$ ratio, and low-grade metamorphism of the rock, collectively suggest that this is a metamorphic age.

INTRODUCTION

Various attempts at direct dating of the Fortescue Group have encountered difficulty, in part due to the low enrichment of Rb in the mafic volcanic rocks which are the predominant lithology in the unit. In the course of re-mapping the central part of the Wyloo 1:250 000 sheet (D. B. Seymour, A. M. Thorne and D. F. Blight), a suite of felsic volcanic rocks was found in the lower part of the Fortescue Group. The presence of a felsic unit provided an opportunity to study material with a more favourable composition for isotopic geochronology. In fact, the resulting Rb-Sr isochron date (2 032 Ma) seems to be young beyond the error limits of the dating, based on associated basic rocks. Together with a large mean square of weighted deviates and a large initial ratio, the young date seems best considered a metamorphic age.

LOCAL GEOLOGY AND PETROGRAPHY

Dipping gently southward off Archaean rocks of the Pilbara Block is a supracrustal sequence, the Mount Bruce Supergroup, which is unconformably overlain to the south by the Wyloo Group, of mixed sedimentary lithology. Because of erosion associated with the unconformity, the Wyloo Group may overlie any group of the Mount Bruce Supergroup, but it commonly rests on the dominantly ferruginous Hamersley Group, which in turn overlies mafic volcanic rocks of the Fortescue Group—the basal group of the Mount Bruce Supergroup. The Fortescue Group rests on Archaean plutonic and metamorphic rocks of the Pilbara Block. In the area of this study, the Mount Roe Basalt, which is the basal formation of the Fortescue Group, is exposed in the Wyloo Dome where the formation rests on the Metawandy Granite, presumed to be an inlier of Pilbara Archaean basement. The unconformity is marked discontinuously by thin sequences of arkose and arkosic conglomerate.

Figure 1 is a map and geological column based on mapping by the Geological Survey in 1980. The previously undivided lower part (Daniels, 1970) of the Fortescue Group, below the Mount Jope Volcanics, has now been sub-divided as shown in the column, based on correlations with the Mount Roe Basalt and the Hardey Sandstone elsewhere in the southern part of the Hamersley Basin.

The Metawandy Granite has been dated by Rb-Sr methods as "younger Archaean" (Riley, 1978). However, the reliability of this date may be in some doubt due to the general deformation (often associated with partial recrystallization) and sericitic alteration observed in the granite, both in the field and in thin section.

Thin felsic volcanic sequences occur within the dominantly basaltic sequence of the Mount Roe Basalt (Blight, 1985) below a prominent quartz arenite sequence comprising the Hardey Sandstone (Fig. 1). At the sampling site, a sequence approximately 30 m of felsic material rests on basalts. The felsic material contains a thin basal, pebbly arkose, overlain by delicately layered, very fine-grained felsic ash-fall tuff and dacitic crystal tuff containing accretionary lapilli. The tuff units are demonstrably part of the normal stratigraphic sequence, and are not intrusive.

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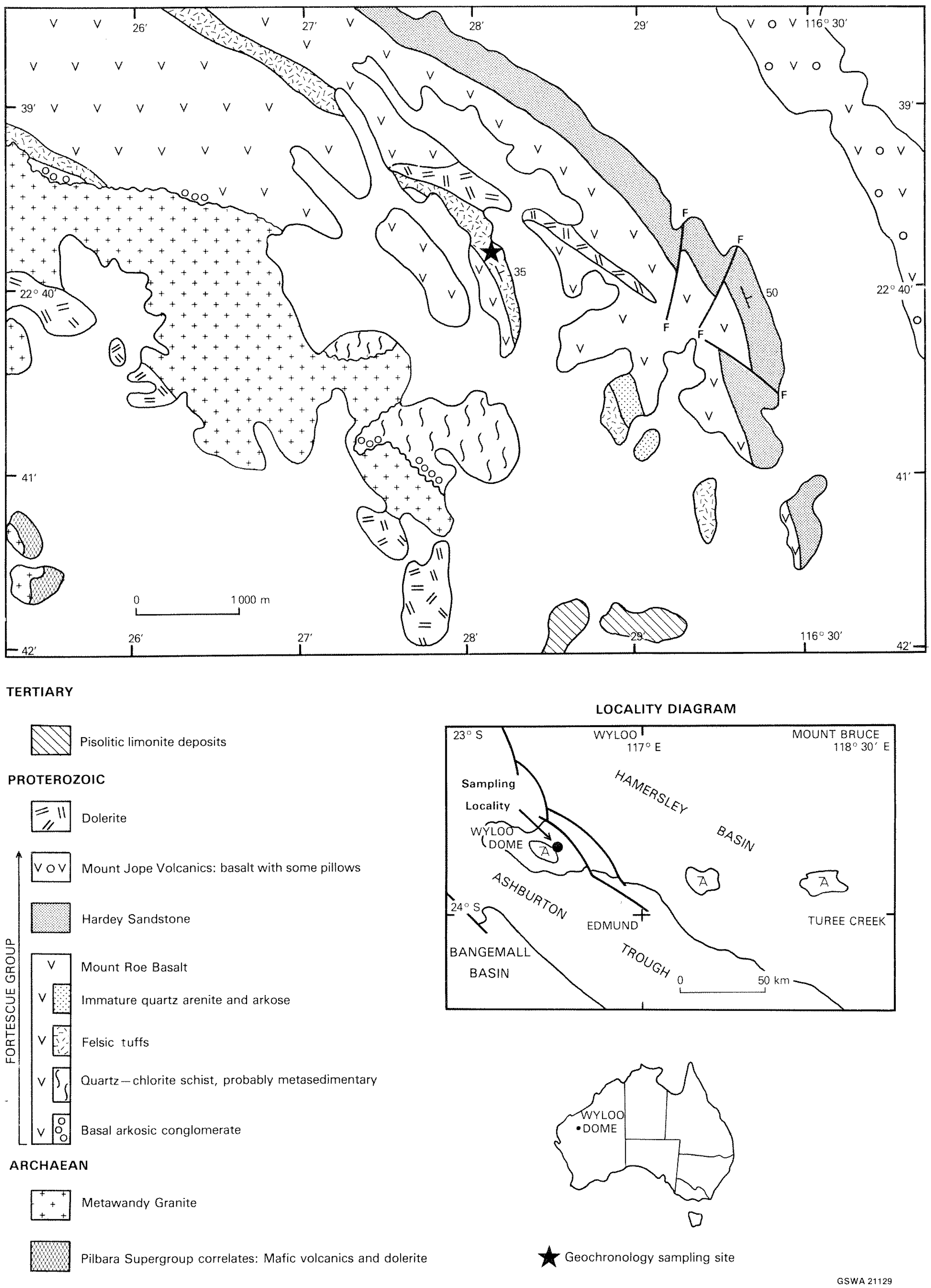

Figure 1. Geology and sample locations.

Crystals and crystal fragments averaging 0.1 mm, but up to 1 mm, in diameter comprise about 10% of the tuffs. Thicker layers ($>$ 1 mm) in the layered tuffs are commonly graded from coarse to fine upwards. Some of these rocks contain scattered pseudomorphs (now iron oxides) after a platy mineral. The pseudomorphs are subhedral and considerably larger (up to 5 mm x 2 mm) than other crystal and vitric fragments in the rock, and are possibly after a diagenetic mineral.

The accretionary lapilli range up to 10 mm x 5 mm in size. They have a concentric structure, with a margin of very fine tuff surrounding a series of concentric zones of fine tuff, in turn enclosing tuff with the same texture and composition as the interlapilli matrix (commonly including crystal fragments). The lapilli appear to have the same composition as the remainder of the rock.

Eleven samples of the felsic tuffs (GSWA samples 69338 to 69348) were collected through the sequence. The distribution of rock types is:

69338-69342 (5 samples): layered felsic tuff
69343-69344 (2 samples): layered felsic tuff with large crystal pseudomorphs
69345-69348 (4 samples): accretionary lapilli tuff

Petrographically, all samples consist of medium to fine grains of quartz and alkali feldspar set in a finely microgranular, devitrified, glassy matrix with relic shards. The mineral assemblage of the matrix is dominated by felsic components which were not specifically identified, and includes abundant sericite and somewhat coarser biotite or chloritized biotite. Biotite is less abundant in the layered tuff (samples 69338-69341) than in the other units. A group of minerals of obscure identity, ranging in grain size between that of the matrix and of the phenocrysts, may include zircon and an unusual chlorite.

Carbonate is present in most samples and is abundant in a few, namely 69341, 69342 and 69344. Of these samples, 69342 and 69344 are the two which do not plot on the isochron. If strontium were introduced with other alkaline earths accompanying carbonatization, the $^{87}Rb/^{86}Sr$ ratio would be reduced, causing the observed migration of these points away from the isochron.

Devitrified relic glass shards are abundant in most samples and readily visible in plain light. In cross-polarized light, the shards are obscured by polygranular devitrification.

The matrix of all samples seems to be recrystallized. This is most clearly demonstrated by the crystallinity of former glass shards, but, in addition, both ultra-fine sericite and somewhat coarser biotite have a static metamorphic, hornfelsic aspect. It is not clear whether recrystallization was associated with devitrification during cooling, or was consequent on later metamorphic heating. Biotite texture supports the latter interpretation.

All samples, with the possible exception of 69343, carry small amounts of pyrite; 69344 has a second metallic sulphide and 69342 may contain a small amount of sphalerite.

Alteration due to weathering or other low-grade processes is minimal.

The mineral assemblages can be summarized as:

quartz-albite-sericite(-biotite) (-carbonate): samples 69338-69344

quartz-albite-sericite-biotite-sphene: samples 69345-69348 (accretionary lapilli tuff)

Amongst other samples taken in the general vicinity of the geochronology sampling site, only one shows petrographic evidence of higher grade metamorphism. This is a dacitic accretionary lapilli tuff (GSWA sample 69306) taken from a point 600 m to the northwest, in close proximity to the margin of a younger intrusive body of dolerite (Fig. 1). The mineral assemblage of this rock is quartz, albite, K-feldspar, chlorite, actinolite, Fe/Ti oxides, and epidote. In the context of regional metamorphism, this assemblage would normally be taken to indicate greenschist facies. However, the same assemblage is consistent with the albite-epidote-hornfels facies of contact metamorphism (Turner, 1981, p.204), and this interpretation is preferred here, due to the presence of the dolerite and to the lack of actinolite and epidote in the other samples. Sample 69306 was not included in the geochronological analyses.

ANALYTICAL METHODS

The samples were prepared mechanically at the Geological Survey of Western Australia and analyzed in the School of Physics and Geosciences, Western Australian Institute of Technology. The methods of analysis are essentially as reported by de Laeter and others (1981). The value of $^{87}Sr/^{86}Sr$ for the NBS 987 standard measured during this project was 0.710 2 $\pm$0.000 1, normalized to a $^{88}Sr/^{86}Sr$ value of 8.375 2.

Measured Rb and Sr values and Rb/Sr ratios, as determined by X-ray fluorescence spectrometry, are listed with mass spectrometric determinations in Table 1. Errors accompanying the data are at the 95% confidence level. We believe the values of Rb and Sr are accurate to $\pm$7%; however, the measured Rb/Sr ratios may not correspond precisely with ratios which would be derived from the separate Rb and Sr values listed.

TABLE 1. WYLOO DOME ANALYTICAL RESULTS

Sample	Rb (ppm)	Sr (ppm)	Rb/Sr	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
69347	72	80	0.902 ± 0.009	2.63 ± 0.02	0.82550 ± 0.00017
69344 (a)	66	59	1.12 ± 0.01	3.35 ± 0.03	0.90375 ± 0.00012
69346	85	76	1.13 ± 0.01	3.31 ± 0.03	0.84971 ± 0.00014
69340	35	27	1.30 ± 0.01	3.80 ± 0.04	0.85445 ± 0.00016
69345	95	66	1.45 ± 0.02	4.26 ± 0.04	0.87818 ± 0.00022
69341	52	32	1.60 ± 0.02	4.70 ± 0.05	0.88634 ± 0.00024
69348	112	67	1.68 ± 0.02	4.93 ± 0.05	0.89052 ± 0.00025
69338	78	33	2.36 ± 0.02	6.97 ± 0.07	0.94755 ± 0.00016
69339	108	38	2.87 ± 0.03	8.52 ± 0.08	0.99616 ± 0.00025
69343	106	33	3.21 ± 0.03	9.57 ± 0.09	1.03801 ± 0.00031
69342 (a)	168	47	3.58 ± 0.04	10.7 ± 0.1	1.10751 ± 0.00016

(a) Omitted from isochron

RESULTS

Analytical results are plotted on the isochron, Figure 2. Samples 69342 and 69344 fall at some distance from an isochron fitted through the remaining points. The results from these samples are omitted from age calculations. These samples are two of the three samples containing substantial carbonate, and may have been open to input of strontium.

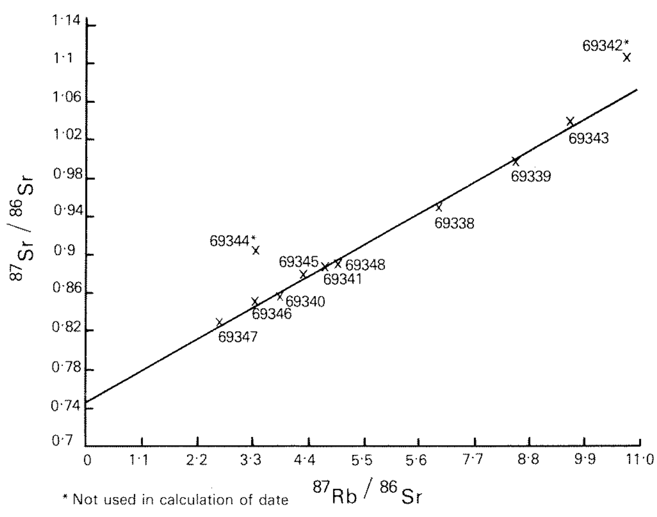

Figure 2. Rubidium-strontium whole-rock isochrons.

The analytical data from the accepted samples have been regressed using the least squares programme of McIntyre and others (1966). The value of 1.42×10^{-11} a^{-1} was used for the decay constant of ^{87}Rb (Steiger and Jäger, 1977). The nine accepted samples yield a model 1 date of 2 041 ±26 Ma, with an initial $^{87}Sr/^{86}Sr$ ratio (R_i) of 0.748 3 ±0.001 6 and a mean square of weighted deviates (MSWD) of 34.6. However the preferred model 4 date is 2 032 +148 Ma with an R_i of 0.748 8 ±0.009 4.

DISCUSSION

Isotopic geochronology of the Fortescue Group gives widely disparate ages from approximately 2 750 Ma by U-Pb (zircon) and Rb-Sr techniques on the Spinaway and Bamboo Creek porphyries (Trendall, 1975) to about 1 500 Ma by Rb-Sr (Blockley and others, 1980) in the Newman area. As is usually the case, U-Pb zircon dates are older than associated Rb-Sr dates. U-Pb dates of 2 490 Ma (Compston and others, 1981) and 2 768 Ma (Trendall, 1983) come from the Fortescue Group and 2470 Ma (Compston and others, 1981) from the overlying Woongarra Volcanics. Likewise samples of galena from the Fortescue Group yield t_{76} Pb model ages of 2 702 Ma and 2743 Ma (Richards and others, 1981). Ages from the Rb-Sr method vary widely, including whole-rock isochron dates of 2 100 to 2 300 Ma (de Laeter and Trendall, in prep.) and 1 487 Ma (Blockley and others 1980). A further Rb-Sr analysis of shale within the Hardey Sandstone in the Fortescue Group yields an age of 2 707 Ma (Hickman and de Laeter, 1977), but probably represents the age of source material. Trendall (1983) has presented a thorough review of radiometric dating in the Hamersley Basin.

While inherited or displaced material may give spuriously old zircon and galena ages, Rb-Sr dates are susceptible to updating by isotopic homogenization which may occur during metamorphic reheating subsequent to emplacement. Thus, on these considerations alone, the Fortescue Group would seem to be bracketed between 2 470 Ma and about 2 250 Ma. On closer analysis of individual dates, Trendall (1983) brackets the base of the Fortescue Group between 2 775 Ma and 2 250 Ma. In either case, the 2 040 Ma date generated by the current study would seem to be anomalously young in comparison with other dating.

Several features of the suite suggest that metamorphic updating may be responsible for the anomalously young age. Both the large mean square of weighted deviates (MSWD), which is 34.6, and the selection of model 4 (variation in both initial ratio and slope) as the most appropriate model suggest disturbance of the isochron. The high initial $^{87}Sr/^{86}Sr$ ratio is consistent either with derivation of magma from an older crustal rock or with metamorphic resetting. The general lack of substantial metamorphic recrystallization in samples studied under the microscope is evidence against metamorphic resetting.

Strontium evolution analysis of the present data suggests a mantle derivation age of 2 619 Ma, assuming single-stage evolution and an isochemical system. This analysis does not distinguish between a metamorphosed rock derived directly from the mantle and an unmetamorphosed rock melted from pre-existing crustal material. However, the strontium evolution date is close to the dates recorded by U-Pb and Pb-Pb techniques and thus, in supporting the Pb dates, supports a metamorphic origin for the isochron date.

CONCLUSIONS

The felsic tuff in the Mount Roe Basalt probably was erupted somewhat prior to the strontium evolution date of 2 619 Ma. Local metamorphism affected some samples in the area and may have provided the energy to reset the Rb-Sr isochron at about 2 032 Ma.

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RUBIDIUM-STRONTIUM BIOTITE DATES IN THE GASCOYNE PROVINCE, WESTERN AUSTRALIA

by W. G. Libby and J. R. de Laeter*

ABSTRACT

New Rb-Sr whole-rock isochrons and two-point biotite and whole-rock isochrons suggest that the Gascoyne Province was generated at about 1.6 Ga, in part at least from a basement of Yilgarn rocks. Continuation of the 1.6 Ga event into the Yilgarn Block suggests that the two provinces have not been substantially displaced relative to one another since that date. Later activity, a thermal event or uplift at 0.8 Ga, is suggested by two-point biotite and whole-rock isochrons. This activity may have been greater, or later, near the west-central part of the province, where biotite dates are younger, than at the northern and southern margins of the province.

INTRODUCTION

The Gascoyne Province, a triangular area of gneiss, granitoid, and metasediment, lies north of the Archaean Yilgarn Block, between latitudes 23° and 26° South (Fig. 1). The transition from Yilgarn Block into Gascoyne Province is initially expressed by tectonic reworking, the appearance of over-printed fabrics, and further to the north, by the appearance of metamorphosed sediments and intrusive granitoids. It is distinguished from the Yilgarn Block by more intense deformation, younger radiometric ages, and predominantly east-west tectonic trends.

To the east, the Gascoyne Province is covered by weakly metamorphosed sedimentary rocks of the Bangemall Group (approximately 1.0 Ga. old), and to the west by Phanerozoic sediments of the Carnarvon Basin. Isotopically, transition with the Yilgarn Block is seen in relics of Yilgarn whole-rock Rb-Sr dates as far north as Dunawah Well, 110 km into the Gascoyne Province, and by Sm-Nd model ages of considerable antiquity that extend into the Gascoyne Province (Fletcher and others, 1983).

Three previous Rb-Sr studies have dealt with the geochronology of the area. De Laeter (1976) published whole-rock and mineral data on samples collected from the western part of the Gascoyne Province by J. Daniels during 1970-71. Williams, Elias, and de Laeter (1978) added dates from the eastern part of the province. Libby and de Laeter, (1979) after a study of the distribution of biotite dates near Perth, drew attention to the similarity between the westward decrease of biotite dates in the Gascoyne area, the Perth area, and the southwestern part of the state near Donnybrook.

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The present study records new biotite and whole-rock dates from the northwestern corner of the Gascoyne Province (Wyloo sheet), further Rb-Sr (whole-rock and biotite) dates from the central part of the Gacoyne Block, and two biotite dates from an adjacent part of the Yilgarn Block. The biotite dates (about 1.6 Ga) from Yilgarn rocks adjacent to the Gascoyne Province are about the same as whole-rock dates within the Gascoyne Province. The event which reset whole-rock systems in the Gascoyne Province appears to have extended, at lower temperature, into adjacent parts of the Yilgarn Block.

Although biotite dates become progressively younger to the west, trend-surface studies incorporating new data from the northwestern corner of the block indicate that the gradient of this decease is less than that shown by Libby and de Laeter (1979) and much less than the gradient in the Yilgarn Block near Perth.

Sample material was blasted from three exposures at Errabiddy, and from two to four samples from each shot point were used to construct a local isochron. These isochrons, which are apparently younger than associated Rb-Sr biotite dates, are tentatively attributed to low-temperature resetting consequent on saussuritization of plagioclase.

ANALYTICAL PROCEDURES

The methods of analysis used during this study are essentially the same as reported by de Laeter and others (1981). The value of $^{87}Sr/^{86}Sr$ for the NBS 987 standard measured during this project was 0.7102± 0.0001, normalized to a $^{88}Sr/^{86}Sr$ value of 8.3752.

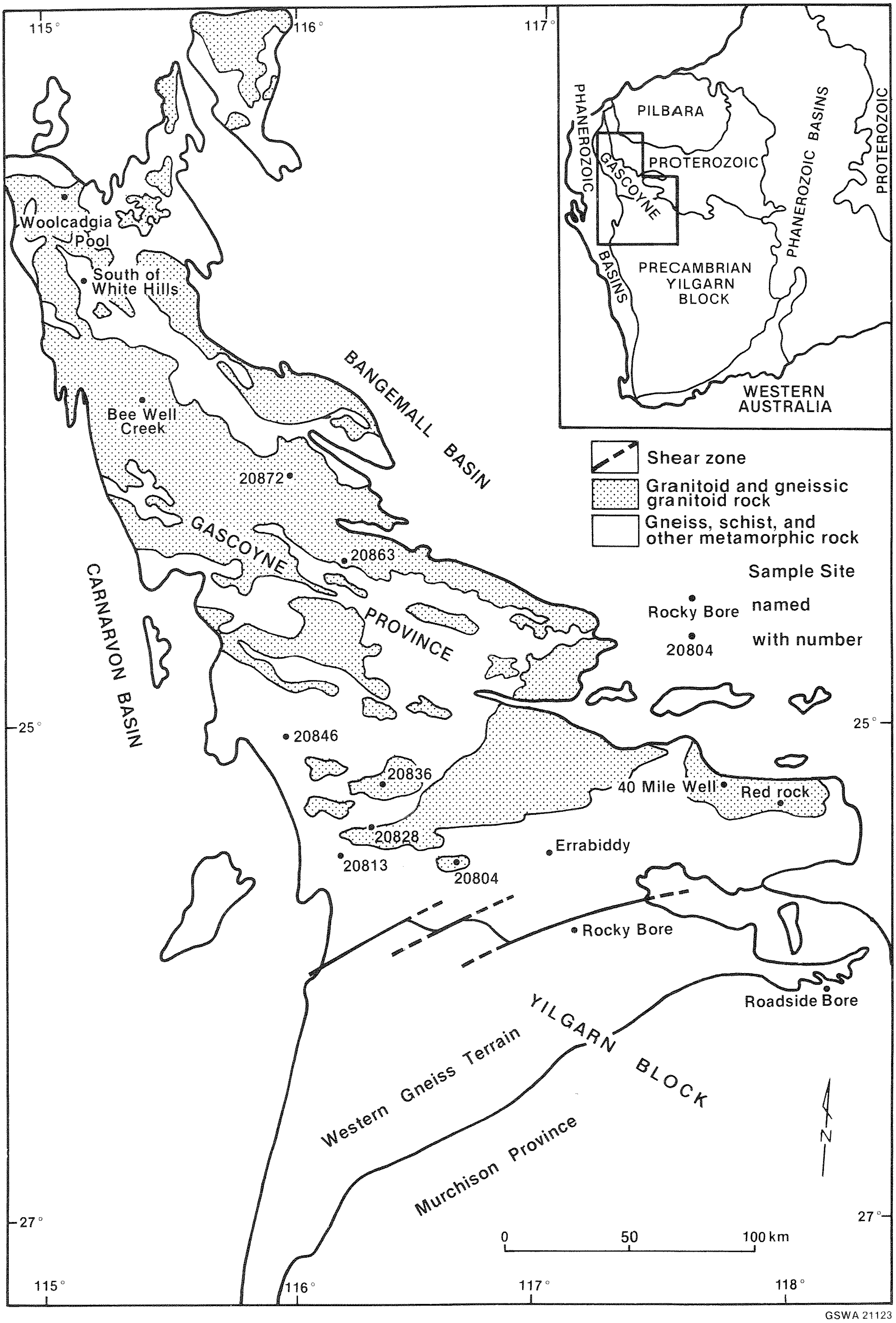

Figure 1. Locality and generalized lithological index to the Gascoyne Province.

Measured Rb and Sr values and Rb/Sr ratios, determined by x-ray fluorescence spectrometry, are listed with mass-spectrometric determinations of $^{87}Sr/^{86}Sr$ in Table 1. Errors accompanying these data are at the 95 per cent confidence level. We believe the values of Rb and Sr are accurate to ±7 per cent; however, the measured Rb/Sr ratios may not correspond precisely with ratios which would be derived from the separate Sr and Rb values listed. A decay constant of $1.42 \times 10^{-11} a^{-1}$ has been used for ^{87}Rb (Steiger and Jäger, 1977). Regression of the whole-rock data has been carried out using the least-squares fit of McIntyre and others (1968)

RESULTS FROM ERRABIDDY, ROCKY BORE, AND ROADSIDE BORE

New analytical data on biotite from Errabiddy, Rocky Bore, and Roadside Bore are listed in Table 1 together with data from the whole-rock splits of the same samples which were analyzed and published earlier (Williams and others, 1978).

Dates generated from 2-point isochrons on mineral and whole-rock pairs are listed in Table 2. The data from Errabiddy are plotted on an isochron-type ($^{87}Sr/^{86}Sr$ *vs* $^{87}Rb/^{86}Sr$) diagram in Figure 2.

TABLE 1. ANALYTICAL DATA, ERRABIDDY, ROCKY BORE, ROADSIDE BORE, WINNING POOL

Sample	*Rb*	*Sr*	*Rb/Sr*	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
			ERRABIDDY		
47046	185	125	1.51 ± 0.02	4.40 ± 0.50	0.83217 ± 0.00053
47047 w.r.	173	187	0.93 ± 0.01	2.71 ± 0.03	0.81427 ± 0.00034
47047 b.t.	799	9	87 ± 1.0	354 ± 4	4.9006 ± 0.0006
47048	110	140	0.79 ± 0.01	2.31 ± 0.02	0.80994 ± 0.00037
47049	61	140	0.44 ± 0.01	1.28 ± 0.02	0.80836 ± 0.00051
47050	75	110	0.68 ± 0.01	1.98 ± 0.02	0.81704 ± 0.00048
47051 w.r.	190	149	1.28 ± 0.01	3.76 ± 0.04	0.83864 ± 0.00048
47051 b.t.	653	8	82.0 ± 0.8	319 ± 3	4.2571 ± 0.0004
47052 w.r.	372	75	4.87 ± 0.05	14.4 ± 0.1	0.94290 ± 0.00058
47052 b.t.	612	13	48.0 ± 0.05	166 ± 2	2.7503 ± 0.0002
47053 w.r.	187	146	1.27 ± 0.01	3.75 ± 0.04	0.87251 ± 0.00051
47053 b.t.	849	35	24.4 ± 0.3	78 ± 1	1.7707 ± 0.0002
			23.6 ±	75 ±	1.7450 ±
47054 w.r.	232	131	1.91 ± 0.02	5.61 ± 0.06	0.89331 ± 0.00050
47054 b.t.	955	8	123 ± 2	625 ± 6	8.4862 ± 0.0009
			ROCKY BORE		
47040 w.r.	190	145	1.30 ± 0.01	3.81 ± 0.04	0.85230 ± 0.00038
47040 b.t.	979	23	42.2 ± 0.4	166 ± 2	4.4491 ± 0.0003
			ROADSIDE BORE		
47075 w.r.	110	150	0.74 ± 0.01	2.15 ± 0.02	0.78208 ± 0.00039
47047 b.t.	840	24	35.1 ± 0.4	134 ± 1	3.9900 ± 0.0006
			WINNING POOL		
60741 w.r.	128	502	0.256 ± 0.003	0.741 ± 0.007	0.73736 ± 0.00031
60741 b.t.	996	15	66.7 ± 0.7	245 ± 2	3.5129 ± 0.00084
60743 w.r. (a)	390	1 180	0.333 ± 0.003	0.96 ± 0.01	0.72851 ± 0.00041
60740 w.r.	230	447	0.514 ± 0.005	1.49 ± 0.01	0.74927 ± 0.00035
60740 b.t.	1 075	24	44.3 ± 0.4	149 ± 2	2.42187 ± 0.00093
60745 w.r. (a)	595	501	1.19 ± 0.01	3.46 ± 0.03	0.77512 ± 0.00029
60748 w.r.	532	418	1.27 ± 0.01	3.70 ± 0.04	0.80424 ± 0.00016
60744 w.r.	373	280	1.35 ± 0.01	3.91 ± 0.04	0.80557 ± 0.00025
60744 b.t.	1 507	18	84.4 ± 0.8	31 ± 3	4.3758 ± 0.00098
60749 w.r.	459	318	1.46 ± 0.01	4.26 ± 0.04	0.81237 ± 0.00041
60749 b.t.	1 281	25	51 ± 0.05	179 ± 2	2.8912 ± 0.00051
60742 w.r.	464	303	1.53 ± 0.02	4.46 ± 0.04	0.81372 ± 0.00028
60742 b.t.	1 525	24	63.1 ± 0.6	229 ± 2	3.3300 ± 0.00077
60746 w.r.	740	162	4.59 ± 0.05	13.67 ± 0.1	1.03038 ± 0.00025
60747 w.r.	710	100	7.13 ± 0.07	21.54 ± 0.2	1.18061 ± 0.00038

(a) Not included on isochron

The dates from Errabiddy biotite and whole-rock pairs are 814, 760, 835, 847, and 858 Ma, and are typical of biotite dates determined earlier from the Gascoyne Province (de Laeter, 1976). On the other hand, dates from Rocky Bore (1 545 Ma) and Roadside Bore (1 695 Ma), in the Yilgarn Block, adjacent to the Gascoyne province, are substantially older than Gascoyne dates but younger than biotite dates from more central parts of the Yilgarn Block (Libby and de Laeter, 1979).

Whole-rock analyses of samples from Errabiddy station plotted by Williams and others (1978) clearly failed to define a single isochron (Figure 3). When samples were sorted into sets, each set from a single shot point, data within each set plotted along a line (Fig. 2). The line defined by each set was sub-parallel to the line from each other set, and encouraged the authors to believe that these lines may represent isochrons. The dates generated by this means ranged from 725 Ma to 783 Ma and had initial $^{87}Sr/^{86}Sr$ ratios that averaged about 0.805. These dates were slightly younger than preliminary biotite and whole-rock two-point isochron dates from the same sample.

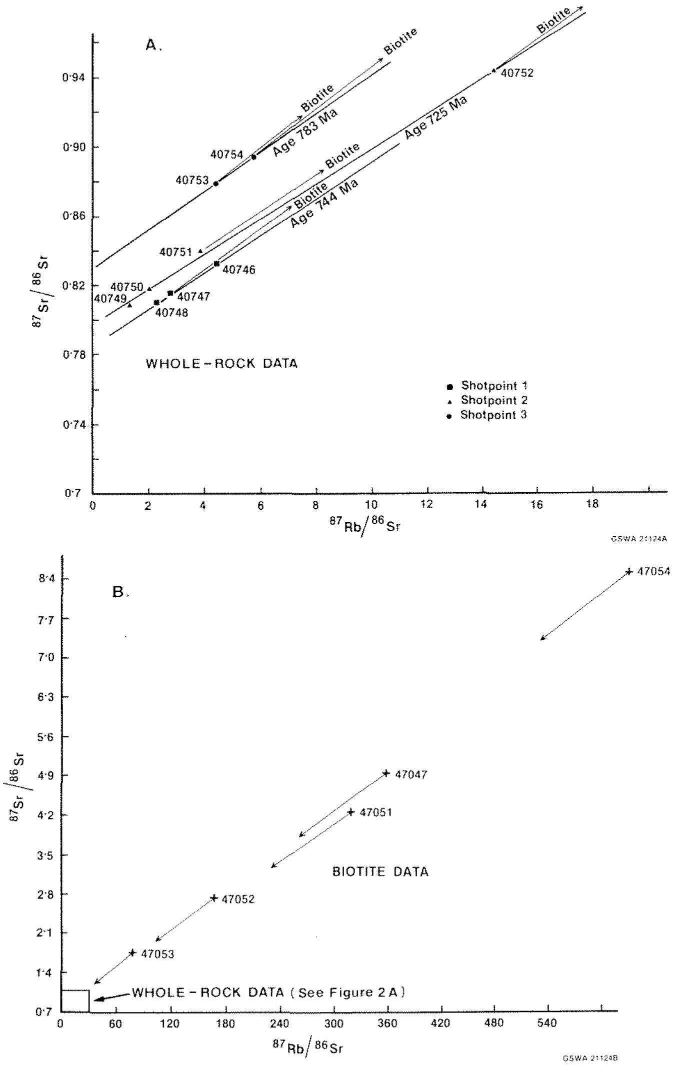

Figure 2. A—Isochrons, whole-rock data from Errabiddy
B—Isochrons, biotite and whole-rock data from Errabiddy.

To check the apparent anomaly of mineral dates which are older than associated whole-rock dates, biotite from two samples (47052 and 47053) was re-analyzed, and biotite was separated from a further three of the original samples (samples 47047, 47051, and 47054). These five new analyses, together with analyses of associated whole rocks (Table 2), confirm that the apparent biotite and whole-rock dates are older than the associated whole-rock isochrons. The biotite date from sample 47047 is 814 Ma, whereas the whole-rock date from samples 47046, 47047, and 47048 (shot point 1) is 744 Ma. The biotite date from sample 47051 is 760 Ma and from 47052 is 835 Ma, whereas the whole-rock date from samples 47049, 47050, 47051 and 47052 (shot point 2) is 725 Ma. The biotite date from sample 47053 is 847 Ma and from 47054 is 858 Ma, whereas the whole-rock date from these samples (shot point 3) is 783 Ma (Table 2 and Fig 3).

None of the sets of whole-rock values contains sufficient elements to properly evaluate any of the whole-rock isochrons. In the single case where four points are available (shot point 2, samples 47049 to 47052), the points are not strictly collinear, and the resulting curve is concave downward (Figure 3).

Available data seem inadequate to establish a reason for the whole-rock dates to be younger than associated biotite dates. Explanations can be suggested but have not been tested on the material at hand. It seems possible that the obvious extensive saussuritization of plagioclase was within, or at least continued into, the temperature range between 200°C and 300°C; as a result Sr in plagioclase was mobilized, but Sr in biotite was trapped above its closure temperature. The whole-rock isochron was rotated to a younger apparent date while biotite retained its earlier values.

If this model is correct, then the biotite ages an whole-rock isochron dates may represent dates of points in the cooling history of the body rather than dates of discrete events.

In any case, there is no reason to believe that the biotite at Errabiddy has responded to conditions which were different in principle to those at the other biotite localities in the Gascoyne Province.

WINNING POOL

Eleven samples from three localities on the Winning Pool 1:250 000 sheet were collected. Samples 60740 to 60748 are from Woolcadgia Pool, sample 60749 is from a locality south of White Hills and 60750 is from Bee Well Creek. Bee Well Creek is 100 km south of Woolcadgia Pool, and the locality south of White Hills lies between the two.

TABLE 2. Rb-Sr WHOLE-ROCK AND BIOTITE DATES FROM THE GASCOYNE PROVINCE AND VICINITY

Locality	Sample Number (a)	Biotite date (b)	Initial ratio (c)	Comments
				ROBINSON RANGE SHEET
Errabiddy	47046			
Shot point 1	47047	814	0.7827	Whole-rock isochron date = 744 Ma, R_i = 0.7854 (e)
	47048			
Shot point 2	47049			Whole-rock isochron date = 725 Ma, R_i = 0.7973 (e)
	47050			
	47051	760	0.7978	
	47052	835	0.7712	
Shot point 3	47053	847	0.8271	Whole-rock isochron date = 783 Ma, R_i = 0.8305 (e)
	47054	858	0.8245	
Rocky Bore	47040	1 545		Whole-rock isochron date = 2 603 ± 149 Ma, R_i = 0.7095 ± 0.0065 (e)
Roadside Bore	47075	1 695		Whole-rock isochron date = 2 461 ± 92 Ma R_i = 0.7058 ± 0.0021 (e)
				WINNING POOL SHEET
Woolcadgia Pool	60740	794	0.7323	Model 1 whole-rock isochron date: 1 532 ± 13 Ma with an initial $^{87}Sr/^{86}Sr$ of 0.7192 ± 0.0006 MSWD is 41.6. The preferred model 4 date is 1 529 ± 85 Ma with an R_i of 0.7194 ± 0.0040
	60741	797	0.7289	
	60742	785	0.7637	
	(d) 60743			
	60744	766	0.7626	
	(d) 60741			
	60746			
	60747			
	60748			
S. of White Hills	60749	831	0.7616	

(a) Samples for which isotopic data is available
(b) Biotite and whole-rock, 2-point isochron dates
(c) Initial $^{87}Sr/^{86}Sr$ ratio on the biotite—whole-rock join
(d) Not used in determination of whole-rock isochron date
(e) Whole-rock dates after Williams and others, 1978. Initial ratios are approximate, calculated from data of Williams and others, 1978.

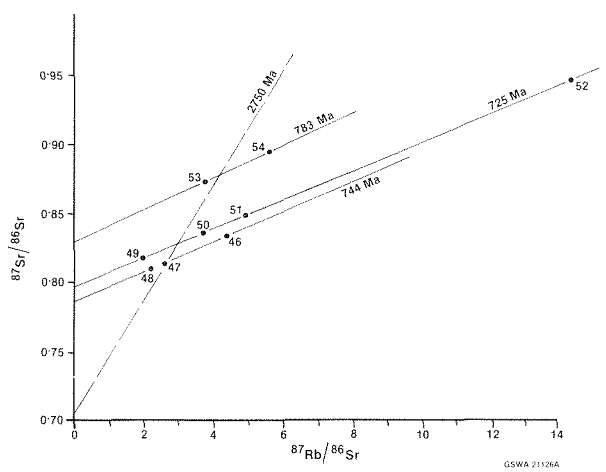

Figure 3. Whole-rock isochrons from individual shotpoints at Errabiddy, after Williams and others, 1978. The first three digits of sample numbers have been omitted for clarity. Samples 47046, -47 and -48 are from one shotpoint; samples 47049, -50, -51 and -52 are from a second shotpoint, and samples 47053, and -54 are from a third.

Analytical results for ten whole-rock samples and for biotite splits from five of these samples are listed in Table 1. Dates derived from a whole-rock isochron and from biotite and whole-rock pairs are shown in Table 2. The same data are represented graphically on a $^{87}Rb/^{86}Sr$ vs $^{87}Sr/^{86}Sr$ isochron plot in Figure 4A. Vectors from five whole-rock points are directed toward associated biotite plots which lie off the diagram. Figure 4B is similar, but at a smaller scale, and shows the biotite data.

The whole-rock data fail to establish an isochron; however, if three points, 60743, 60745, and 60750 are discarded, a low-precision isochron is formed by the remaining points at 1 532 ± 0.0006 Ma and a mean square of weighted deviates (MSWD) of 41.6. The prefered model 4 date is 1 529 ± 85 Ma with an R_i of 0.7194 ± 0.0040.

Five biotite and whole-rock dates are 794, 797, 785, 766 and 831 Ma yielding a mean biotite date of 795 Ma.

The three discarded, anomalous, whole-rock data points are reasonably collinear by themselves and define a reasonable date of 1 390 Ma and a reasonable R_i of about 0.707. However, neither petrographic nor field chatacteristics provide a clear reason to believe that this line is a true isochron, recording an event distinct from that of the accepted isochron. Several of the samples (60746, 60747, 60748, and 60750) are described as late phases (pegmatite or late granitic dykes); however, only one of these is in the isotopically anomalous group. Although another anomalous sample (60745) is described as a xenolith from a late dyke, data from a sample of the dyke itself

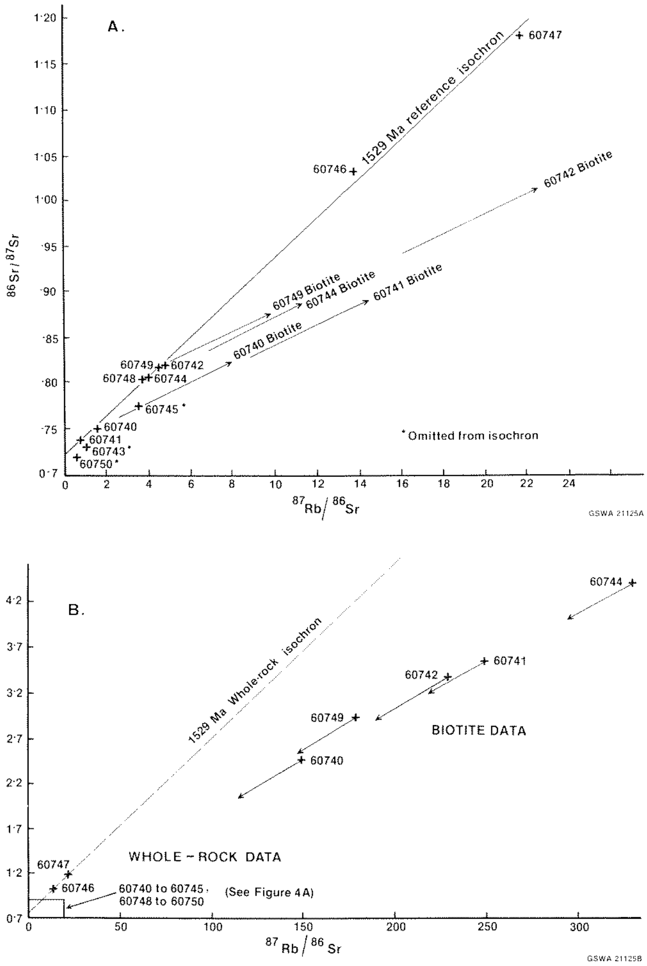

Figure 4. A—Isochron, whole-rock data from the Winning Pool sheet.
B—Isochrons, biotite and whole-rock data from the Winning Pool sheet.

plot near the accepted isochron. Two of the anomalous samples have distinct penetrative deformation but the third does not. Thus there is no external reason to assign separate events to the main isochron and the "isochron" defined by the three anomalous samples.

The scatter of data from both "accepted" and "anomalous" groups, together with variable deformation in the rock and relatively young biotite dates, indicates that the rocks have experienced heating after the event identified by the isochron date. Furthermore, the high R_i (0.7194) shows that the rock was not derived from the mantle at the isochron date. If single-stage evolution, isochemical processes, and representative sampling are assumed, then a mantle derivation date of 1 706 Ma is suggested by strontium-evolution analysis.

In contrast to the disturbed character of the whole-rock isochron, biotite dates at 794, 797, 785, 766 and 831 Ma are reasonably consistent, the more anomalous sample (60749, at 831 Ma) is from south of White Hills, some 35 km from Woolcadgia Pool, where the other samples were collected. The mean of these dates is well within one standard deviation of the mean of all Gascoyne biotite dates.

DISCUSSION AND CONCLUSIONS

Both whole-rock data from the Gascoyne Province and biotite data from adjacent parts of the Yilgarn Block (Roadside Bore and Rocky Bore) yield dates of approximately 1.6 Ga. The 1.6 Ga biotite dates in the Yilgarn Block can reasonably be attributed to marginal effect of the main plutonic event of that age in the Gascoyne Province. This conclusion may provide an argument that the relative positions of the Yilgarn Block and Gascoyne Province have remained effectively fixed since the 1.6 Ga event.

The further analyses of biotite from Errabiddy confirm earlier observations that the biotite dates are younger than associated local apparent whole-rock dates. This situation may be brought about by the updating of the whole-rock isochron by mobilization of Sr in plagioclase during low-temperature saussuritization.

The first-order trend surface on biotite Rb-Sr dates (Fig. 6) which includes recent data has much the same trend as a similar surface published earlier (Libby and de Laeter, 1979), but the gradient of the surface has been substantially reduced by the new data. A first-order trend surface is generated by fitting the data to a linear (first-order) equation. The surface formed is a plane. A section with individual values projected to a line normal to trend surface isopleths (Fig. 5) shows that the scatter of values at most localities exceeds the entire range of the surface within the area mapped.

A second-order surface (Fig. 7) is likewise poorly fitted, but suggests that dates near the centre of the basin may be younger than at the north and south margins of the basin. A second-order surface is generated by fitting the data to a quadratic (second-order) equation. Horizontal sections through the surface so formed are parabolas.

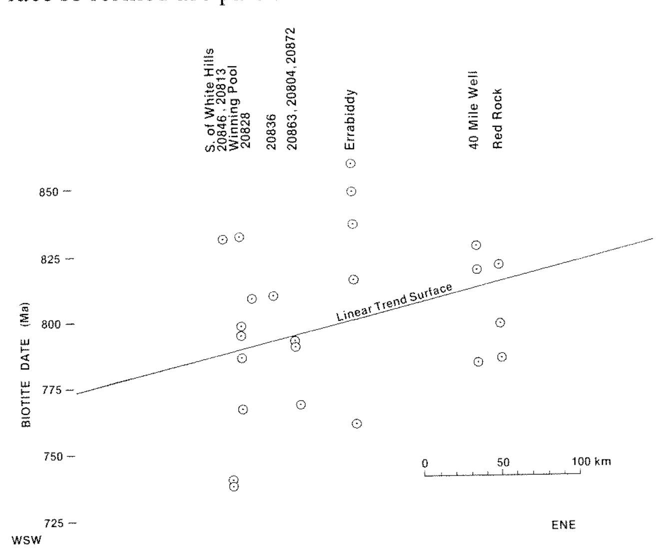

Figure 5. Individual sample values projected onto a section normal to the first-order trend surface.

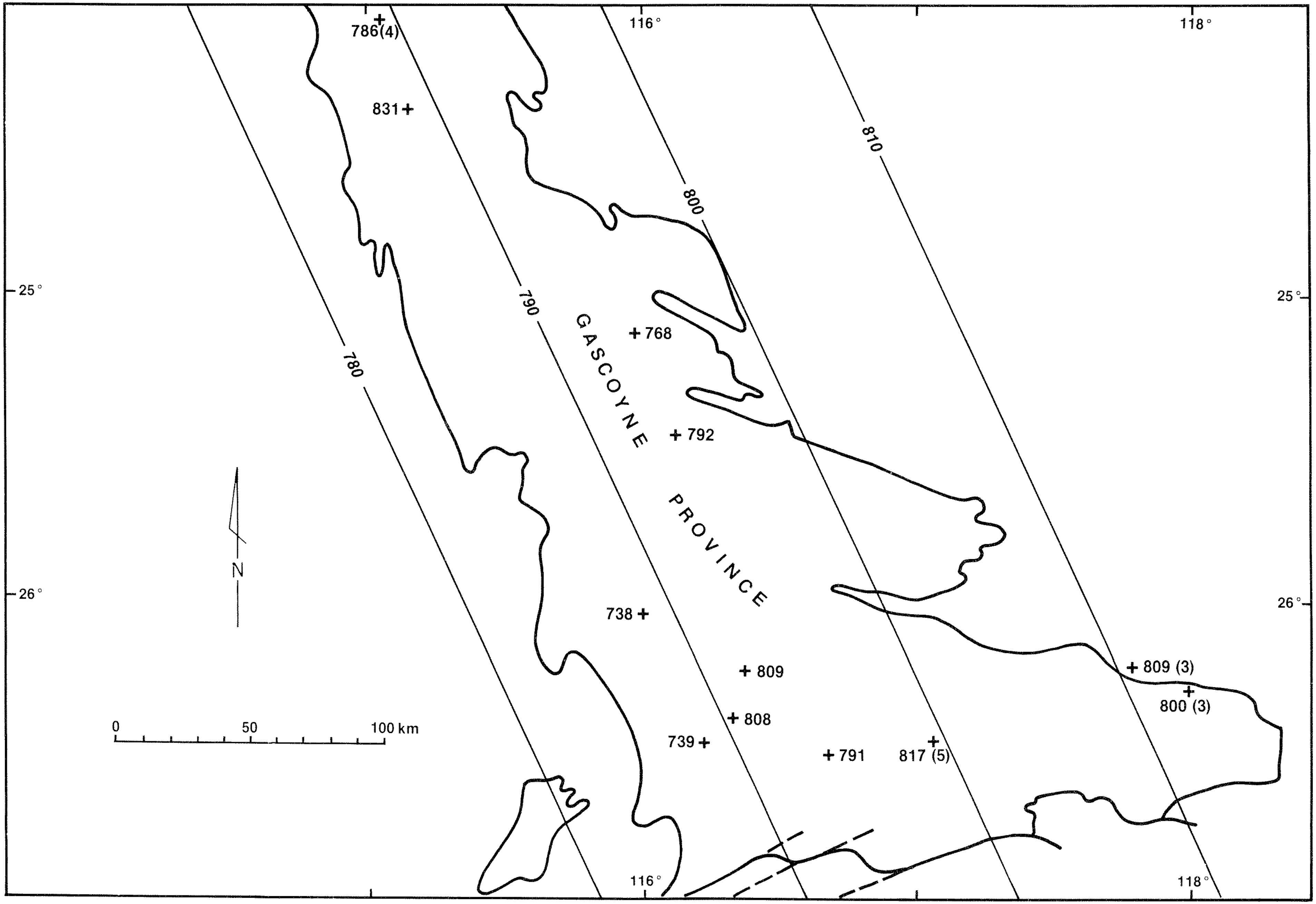

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Figure 6. First-order trend surface on biotite dates from the Gascoyne Province. Multiple dates from a single locality have been averaged from the number of determinations shown in curved brackets. Values are in millions of years.

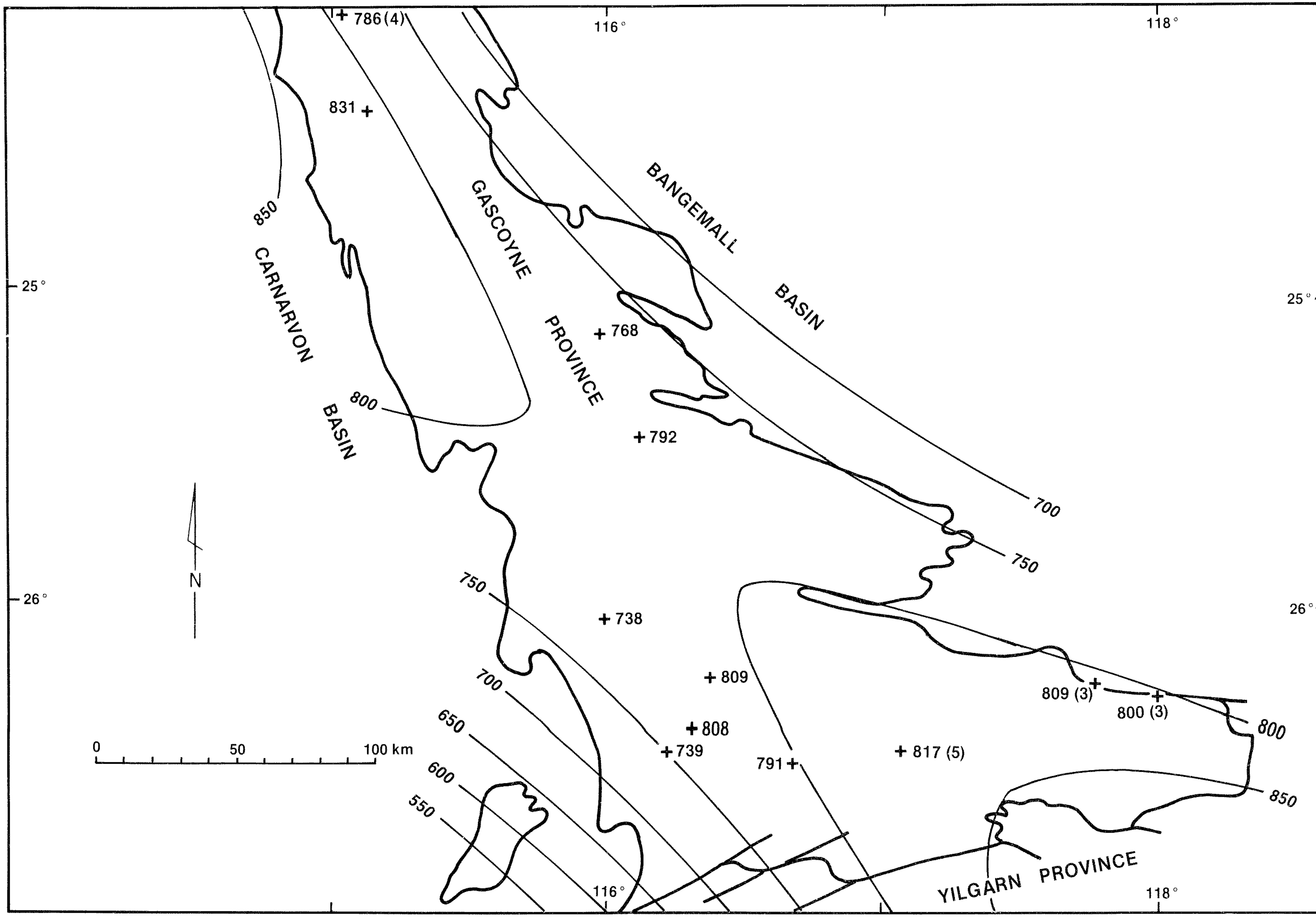

Figure 7. Second-order trend surface on biotite dates from the Gascoyne Province. Multiple dates from a single locality have been averaged from the number of determinations shown in curved brackets. Values are in millions of years.

ACKNOWLEDGEMENTS

The authors would like to thank Mr D. J. Hosie who provided technical assistance to the project.

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UPWARD-SHALLOWING SEQUENCES IN THE PRECAMBRIAN DUCK CREEK DOLOMITE WESTERN AUSTRALIA

by A. M. Thorne

ABSTRACT

Nine major sedimentary facies are recognized in the 2.0 to 1.7 Ga old Duck Creek Dolomite, which outcrops at Duck Creek Gorge in the western Pilbara. Markov Chain analysis demonstrates that these facies occur in repeated upward-shallowing sabkha-type sequences, among which offshore-barrier, lagoonal, intertidal, and supratidal deposits are recognized. These Proterozoic sequences are closely similar to the transgressive-regressive sequence of lithologies preserved in the Holocene sediment profile of Abu Dhabi, Arabian Gulf. A comparison of sequences from the upper and lower portions of the measured section at Duck Creek Gorge indicates that the coastal lagoonal complex became better developed with time. Small, branching-columnar stromatolites and large, branching-columnar stromatolites in the upper part of the section are interpreted as subtidal to shallow-intertidal forms. These, and tiny aborescent stromatolites that occur in the supratidal lithologies, are closely comparable to stromatolites reported from the 2.2 to 1.8 Ga old Rocknest Formation of Canada.

INTRODUCTION

Stratigraphy

The Duck Creek Dolomite is part of the Wyloo Group, an association of clastic carbonate and intermediate to basic volcanic rocks that outcrops in the Western Pilbara (Fig. 1). the Wyloo Group unconformably overlies the Fortescue, Hamersley, and Turee Creek Groups, and has an age of 2.0 to 1.7 Ga (Gee, 1980).

Pevious Work

No systematic studies of the sedimentology of the Duck Creek Dolomite have previously been undertaken. Daniels (1970) measured the succession at Duck Creek Gorge, recording colour, nature of bedding, and variation in abundance of stromatolites. Most interest in the Duck Creek Dolomite has concentrated on palaeontology. Edgell (1964), Grey (1979, 1982), Preiss (1977), and Walter (1972) studied stromatolites; and Knoll and Barghoorn (1975) recorded a gunflint-type flora from cherts in the upper levels of the formation.

Terminology

Intraclast: Fragment of penecontemporaneous, usually weakly consolidated, carbonate sediment that has been torn up and redeposited by current action.

Oncoid: A concentrically laminated sedimentary clast, which generally comprises alternating layers of blue-green algae and trapped carbonate sediment.

Grainstone: A mud-free grain-supported sedimentary carbonate rock.

Packstone: A grain-supported sedimentary carbonate rock which has a matrix of carbonate mud.

Fenestrae: Early diagenetic voids (larger than the grain-supported interstices) in the framework of a sedimentary rock. Laminoid fenestrae are elongate voids aligned parallel to the bedding.

Synoptic relief: In stromatolites, the relief of a stromatolite above its substrate at an instant of time during formation of the stromatolite.

FACIES DESCRIPTIONS

The succession of dolomites exposed at Duck Creek Gorge (lat. 22°28′39″ S, long. 116°19′25″ E) was examined in detail during the course of 1:250 000-scale remapping of the Wyloo area. The information presented in this paper was obtained by logging 220 m of particularly well-exposed dolomite at the eastern end of the gorge. In this part of the succession nine facies were recognized.

Facies A—intraclast grainstone

This facies outcrops in laterally persistent, tabular beds 5 to 100 cm thick (Fig. 2A). The upper and lower bedding surfaces are sharply defined, and the lower contact usually shows evidence of erosion. Little evidence of cross-stratification is preserved within these grain-stones, which may contain clasts up to 1.5 m across.

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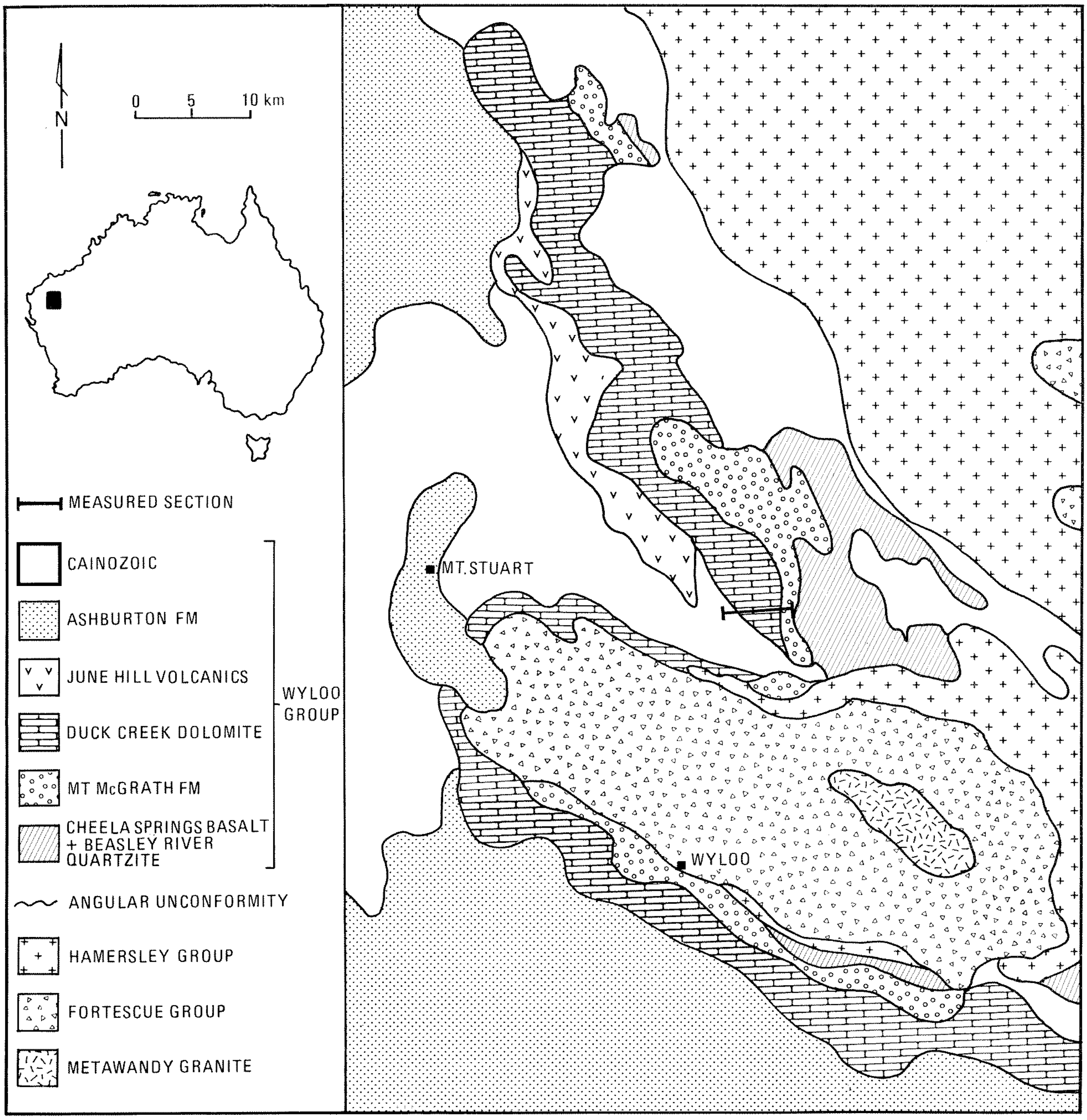


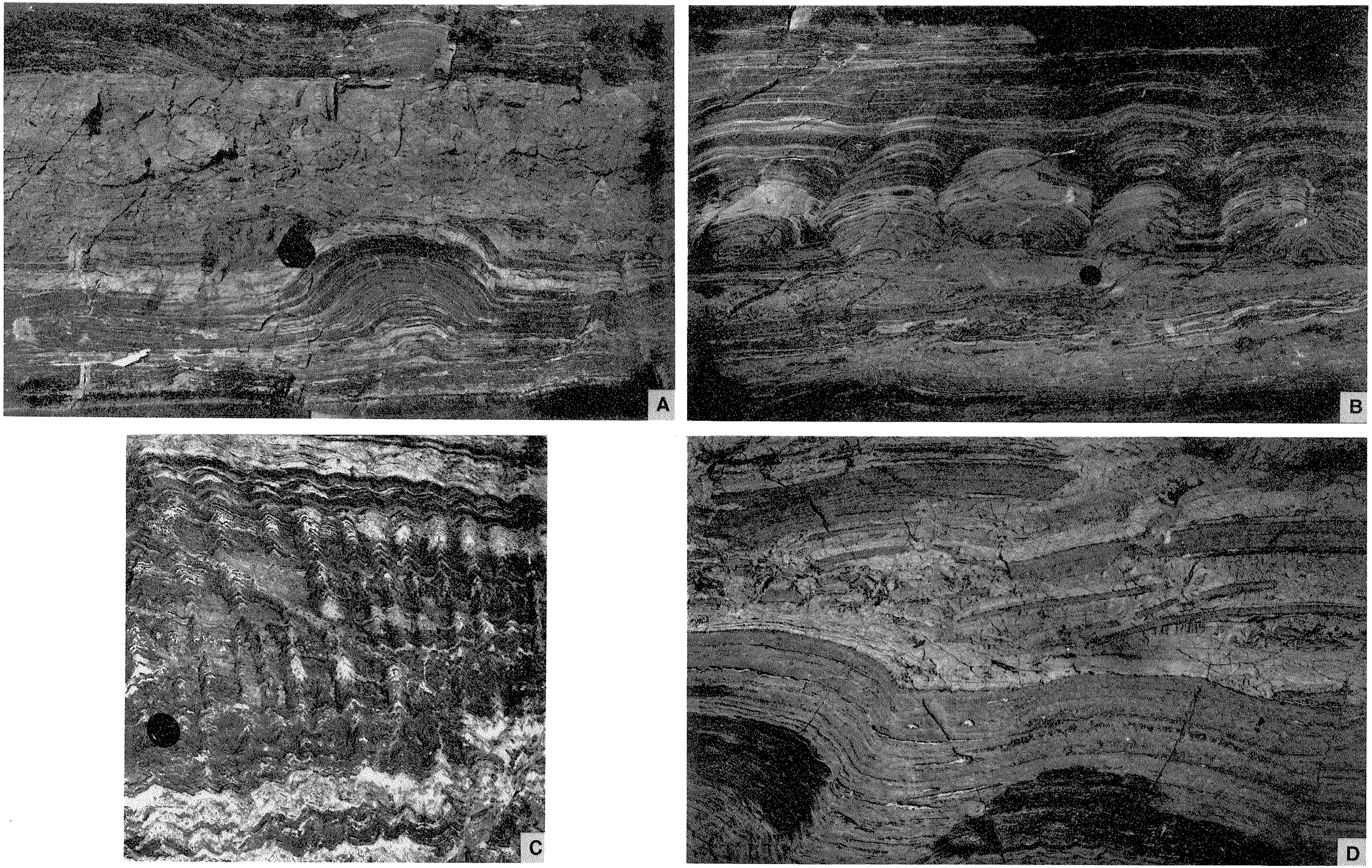
Figure 1. A geological sketch map of part of the western Pilbara, Western Australia.

Figure 2. A—A 25 cm-thick, erosively based intraclast grainstone overlying intertidal and supratidal dolomites. The grainstones, which rarely display any cross-stratification, are poorly sorted and may contain clasts up to 1.5 m across. This facies is interpreted as a transgressive barrier deposit.

B—Lagoonal, nucleated domical stromatolites overlying an intraclast grainstone containing a large imbricate clast (dipping gently to the left). The domical stromatolites are nucleated on pebble-sized oncoids and intraclasts, and the synoptic relief of the domes decreases upwards until they merge into the overlying intertidal laminated dolomite facies.

C—Highly silicified cuspate stromatolites showing the change to a convex form when traced upwards. Individual laminae either maintain constant thickness or vary in thickness when traced from peak to trough. These stromatolites were ridge-like and probably grew in a high-intertidal to supratidal environment.

D—High-intertidal to low-supratidal low domical stromatolites overlain by a very poorly sorted intraclast grainstone. The broad domes frequently comprise 1 to 5 cm layers of tiny aborescent stromatolites (field of view 1.5 m across).



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In thin section, the intraclasts are seen to be very poorly sorted and angular to well rounded. They are derived from pre-existing stromatolitic and grainstone deposits. The clasts are cemented by an early generation of fine- to medium-grained (20 to 200 µm) dolomite, and by later generations of coarse-grained subhedral dolomite and large, equant to bladed quartz crystals.

Facies B—nucleated domical stromatolites

Domical stromatolites nucleated on tabular oncoids and intraclasts are abundant within the Duck Creek succession (Fig. 2B). Usually the oncoid or intraclast nucleii are found as isolated clasts and give rise to spaced domical stromatolites whose synoptic relief decreases from about 10 cm at the base to zero at the top of the dome. The nucleii, particularly the oncoids, may occur in clusters; in which case, they form the cores of large, isolated domes up to 1.5 m across.

Facies C—laminated dolomite

This facies is characterized by planar to gently undulatory laminations and abundant laminoid or irregular fenestrae—now infilled by sparry dolomite, chalcedony, and coarsely crystalline quartz (Fig. 2B). Some laminae are disrupted by penecontemporaneous shrinkage cracks oriented normal to the bedding. Petrographically, two distinct types of laminae can be recognized. One type is marked by variations in the average size (5 to 250 µm) of the dolomite crystals. The finest grained laminae, in particular, are discontinuous and irregularly undulating. The second type of lamination is formed by sand-sized intraclasts that occur as laterally continuous layers 1 mm to 1 cm thick.

Facies D—cuspate stromatolites

Stromatolites in this facies display an irregular cuspate form with a synoptic relief of 1 to 3 cm, and a distance between peaks of 2 to 7 cm (Fig. 2C). Individual laminae either maintain a constant thickness or vary in thickness when traced from peak to trough. The stromatolite crests may merge upwards into convex forms. This stromatolite was probably ridge-like.

Facies E—low-domical stromatolites

These stromatolites show a gentle domical cross-section, 20 to 60 cm wide, with a synoptic relief of 3 to 10 cm (Fig. 2D). They have nucleated upon irregularities on the underlying substrate and exhibit two distinct forms of internal structure. One form is identical to the internal structure of the laminated-dolomite facies; the other form comprises 1 to 5 cm-thick layers of tiny aborescent stromatolites with close affinities to *Asperia* Semikhatov 1978 (Kathleen Grey, pers. comm., 1983).

Facies F—disrupted-domical stromatolites

These are bulbous stromatolite domes with relief up to 10 cm (Fig. 3A). The domes are characterized by having a core in which the original laminae are disrupted, and either partially or completely replaced by silica or coarsely crystalline dolomite. Many domes show evidence of penecontemporaneous erosion.

Facies G—tepee structures

In this facies, dolomite laminae have an irregular cuspate cross-section with an amplitude of a few tens of centimetres and wavelengths of up to 2.5 m (Fig. 3B). The anticlinal portions of the tepee are frequently brecciated and infilled by coarse detritus. The fact that the tepees are gradually overlapped by successively younger sediments, and often show evidence of penecontemporaneous erosion, demonstrates their synsedimentary origin.

Facies H—small, branching-columnar stromatolites

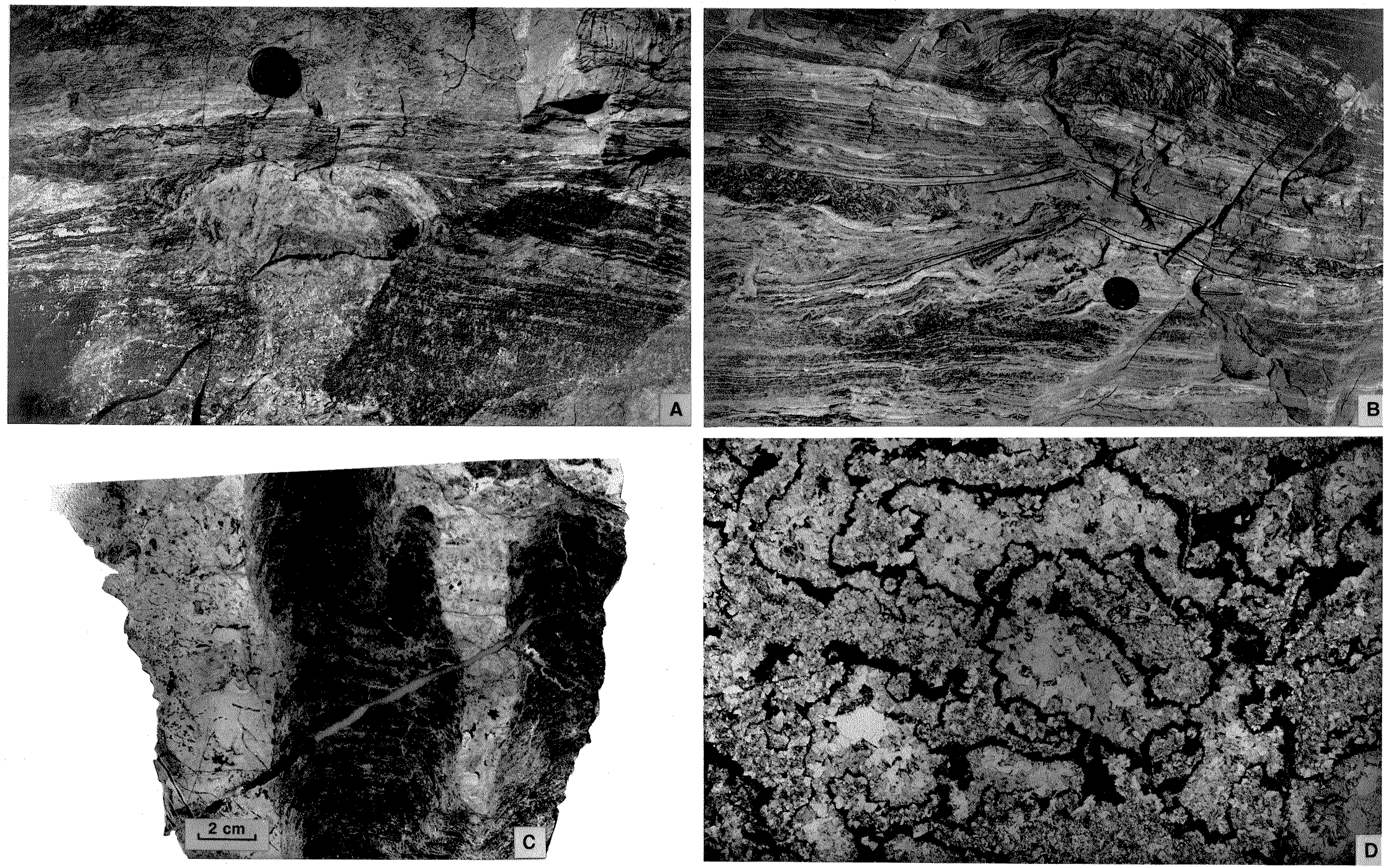
These stromatolites (*Pilbaria perplexa* Walter 1972) form continuous tabular biostromes 10 to 25 cm thick (Fig. 3C). The stromatolites are 2 to 5 cm wide, up to 25 cm high, and display an irregular style

Figure 3. A—The disrupted domical facies (supratidal). The lamination in the core of the small dome has been disrupted and is now replaced by coarsely crystalline dolomite and silica. This structure is closely similar to anhydrite diapirs reported from recent sabkha sediments of the Arabian Gulf. The planation surface truncating the dome may represent the effects of penecontemporaneous wind erosion.

B—The tepee facies (some of the bedding has been emphasised for clarity). The anticlinial portion shown here represents the buckled margins of broad polygonal structures formed by the repeated C—Part of the small branching-colomnar-stromatolite facies (lagoonal) showing the nature of the as a supratidal deposit.

C—Part of the small-branching-colomnar-stromatolite facies (lagoonal) showing the nature of the stromatolite morphology and internal structure. The stromatolites are the *Pilbaria perplexa* of Walter (1972). The interbranch areas are infilled by dolomitic intraclast packstone and grainstone.

D—Coarsely crystalline, subhedral to anhedral dolomite spar interpreted as a replacement after former anhydrite. The anhydrite is believed to have grown displacively within the original sediment, squeezing it aside until only vestiges remain. These vestiges are now represented by the thin, dark, finely crystalline dolomite walls.



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of branching in which individual branches do not maintain constant thickness. In addition, most stromatolite margins are embayed; and the internal laminae are steeply convex and may give rise to distinct walls. The area between branches is dolomitic intraclast grainstone or intraclast packstone, and is generally silicified.

Facies I—large, branching-columnar stromatolites

In this facies, the stromatolites occur in tabular units 50 to 200 cm thick (Fig. 3D). Stromatolite columns are usually 5 to 10 cm wide and sub-cylindrical in form with embayed walls. Some columns are coalesced or bridged, and most forms probably had a depositional relief of 5 to 25 cm. Unsilicified inter-column areas are infilled by dolomitic intraclast grainstones or intraclast packstones.

Fenestral dolomites

A feature commonly associated with facies C, D, E, F, and G, is the presence of irregular fenestrae, 1 to 20 cm wide, that cut across, disrupt, or merge with the original sedimentary layering. The fenestrae may contain disrupted relics of the original laminated dolomite, as shown in Figure 3D, or they may be composed entirely of coarsely crystalline, subhedral-to-anhedral dolomite cement; bladed to equant, subhedral-to-anhedral replacement dolomite; or coarsely crystalline quartz and chalcedony. The latter may show either length-slow or length-fast optical properties.

SEQUENCE ANALYSIS

An important consequence of Walther's Law of Facies-see Middleton (1973)—is that facies occurring in conformable vertical sequences were formed in laterally adjacent environments of deposition. This relationship provides the basis for interpreting the succession of facies observed at Duck Creek Gorge. Because of the large number of facies transitions recorded at this locality, the logged data were examined statistically to determine the most significant facies associations. The technique adopted for this purpose is based upon the embedded Markov Chain method (Powers and Easterling, 1982).

A sequence of lithologies is said to have a Markov property if it can be demonstrated that the occurrence of a particular facies is dependent upon the nature of the underlying facies—*e.g.* in a sequence of sediments it may be observed that a particular facies, say facies "P", tends to be followed by facies "Q" but not by facies "R" or "S". The transitions in a sequence of lithologies can be summarized in a matrix of one-step transitions; in Table 1 for example, the row A, column B entry in the matrix is the number of transitions

TABLE 1. Observed transition frequencies (Lower part of measured section).

	S	A	B	C	D	E	F	G	RT
S	0	6	9	3	0	0	0	0	18
A	12	0	4	8	1	0	1	0	26
B	0	4	0	10	0	3	1	0	18
C	3	8	3	0	1	5	5	2	27
D	1	1	0	2	0	0	0	0	4
E	1	2	0	3	2	0	1	0	9
F	1	3	1	3	0	0	0	0	8
G	0	2	0	0	0	0	0	0	2
CT	18	26	17	29	4	8	8	2	

S = Sharp contact
A = Intraclast - grainstone facies
B = Nucleated - domical - stromatolite facies
C = Laminated - dolomite facies
D = Cuspate - stromatolite facies
E = Low - domical - stromatolite facies
F = Disrupted - domical - stromatolite facies
G = Tepee facies
H = Small branching – columnar – stromatolite facies
I = Large branching – columnar – stromatolite facies
CT = Column total
RT = Row total

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TABLE 2. Estimated expected transition frequencies (Lower part of measured section). See Table 1 for key.

	S	A	B	C	D	E	F	G	RT
S	0.0	5.3	3.1	6.1	0.63	1.3	1.3	0.3	18
A	5.3	0.0	5.0	9.9	1.0	2.1	2.1	0.55	26
B	3.2	5.2	0.0	6.0	0.6	1.3	1.3	0.3	18
C	5.8	9.4	5.4	0.0	1.1	2.3	2.3	0.6	27
D	0.6	1.0	0.6	1.2	0.0	0.26	0.25	0.06	4
E	1.5	2.4	1.4	2.8	0.3	0.0	0.69	0.14	9
F	1.3	2.1	1.2	2.4	0.3	0.6	0.0	0.1	8
G	0.3	0.5	0.3	0.6	0.6	0.16	0.13	0.0	2
CT	18	26	17	29	4	8	8	2	

See Table 1 for key

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TABLE 3. Observed minus expected probabilities (Lower part of measured section). Only positive values shown. See Table 1 for key.

	S	A	B	C	D	E	F	G
S	0	0.04	0.43	0.17	-	-	-	-
A	0.26	-	0.04	0.08	-	-	-	-
B	-	0.07	-	0.23	-	0.1	-	-
C	0.1	0.05	0.09	-	-	0.1	0.11	0.05
D	0.1	-	-	0.2	-	-	-	-
E	0.06	0.05	-	0.03	0.19	-	0 04	-
F	0.03	0.12	0.02	0.08	-	-	-	-
G	-	0.75	-	-	-	-	-	-

See Table 1 for key

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from lithology A upwards into lithology B (this occurred 4 times). Because transitions from one facies into another bed of the same facies are not recorded, the matrix has the characteristic that the main diagonal frequencies must be zero. These values are referred to as structural zeros, and the sequence itself is termed an embedded Markov Chain.

More tests for a Markov property, *e.g.* Gingerich (1969) and Miall (1973), establish the Null Hypothesis that the sequence of facies transitions observed is random, a hypothesis which is tested using a Chi-squared statistic against an "independent trials" matrix. However, because matrices containing structural zeros cannot result from simple independent random processes, the above test is inappropriate. In particular, the "independent trials" (or expected transition frequencies) should be calculated taking into account the presence of the structural zeros. This problem can be resolved by various statistical techniques (Carr, 1982; Powers and Easterling, 1982). The method adopted for this work is based upon that of Powers and Easterling (1982) and employs a model of quasi-independence to overcome the problem of structural zeros in the transition matrix.

The logged section at Duck Creek Gorge can be divided into two halves on the basis of the distribution of facies types. In the following analysis, the presence of five small gaps in the complete sequence was ignored, although it is recognized that they probably add a small random component to the sequence of facies. It should also be noted that sharp facies contacts are shown in the transition matrices (Tables 1-3). All other contacts are gradational except for the base of the intraclast-grainstones facies, which is always sharp and erosive.

Lower half of logged section

Table 1 shows the observed facies transitions in the lower part of the Duck Creek Gorge section. Table 2 gives an estimate of the expected transition frequencies using a model of quasi-independence. Testing the fit of the quasi-independent model yields $\chi^2 = 73.5$ a value which would be expected less than once in 200 times (with forty-one degrees of freedom) if a Markov process were not operating. There is thus considerable evidence against the assumption of quasi-independence, and the facies associations in this part of the succession are shown to have a strong Markov property.

We can now ascertain which of the facies transitions are the most significant. This is achieved by converting the observed and expected frequency data in the matrices into probabilities (these values are obtained by dividing the value of a particular element in the matrix by the appropriate row total). The expected probabilities are then subtracted from the observed probabilities, and the highest positive values that remain (those greater than 0.1) constitute the most significant transitions (Table 3).

The most significant transitions are summarized in Figure 4, where it can be seen that the most preferred sequence, repeated throughout this lower part of the section, is one which shows an intraclast-grainstone base, overlain by a sharp contact, which is, in turn, succeeded by nucleated domical stromatolites and then by the laminated-dolomite facies. From here the sequence passes upwards into one or several of the following facies, tepee, disrupted-dome, cuspate-stromatolite or broad-domical-stromatolite before being overlain by the next erosively based intraclast grainstone. This sequence, designated "Preferred Sequence 1", is shown diagrammatically in Figure 5.

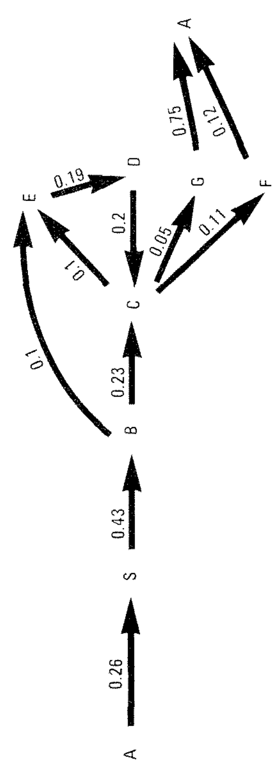


Figure 4. Facies-relationship diagram showing the most significant transitions (lower part of measured section) *See* Table 1 for key.

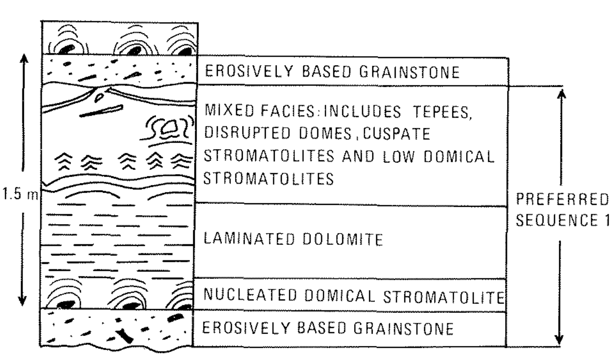


Figure 5. Preferred Sequence 1—lower part of measured section. Subdivisions of sequence not drawn to scale.

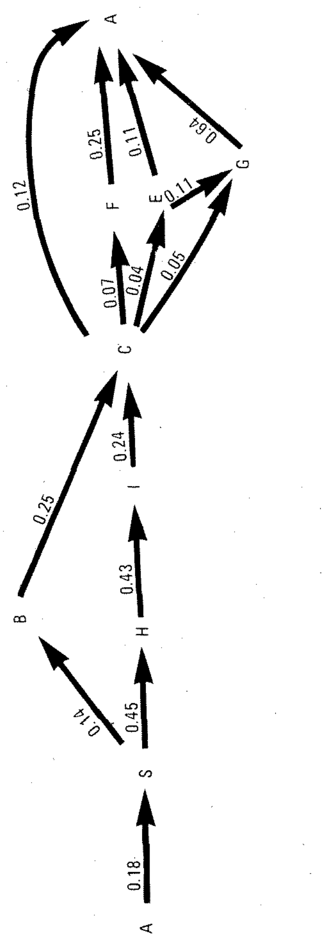


Figure 6. Facies-relationship diagram showing most significant transitions (upper part of measured section). *See* Table 1 for key.

Upper half of logged section

The uppermost part of the measured section was also analyzed by the statistical method outlined earlier. The strong Markov tendency present in the lower part of the section is even more pronounced in the upper part. Testing the fit of the quasi-independent model yields $\chi^2 = 156$, a value which could be expected less than once in 1000 times (with 55 degrees of freedom) if a Markov process were not operating. The more significant transitions in the upper part of the section are shown in Figure 6, and the most preferred repeated sequence is characterized by a basal intraclast grainstone overlain by a sharp contact, which is followed, in most cases, by small, branching-columnar stromatolites then by large, branching-columnar stromatolites, which are overlain by laminated dolomite. From here the sequence passes back into the erosive grainstone either directly or through the tepee, low-domical-stromatolite, or disrupted-domical-stromatolite facies. The complete sequence, designated "Preferred Sequence 2", is shown diagrammatically in Figure 7.

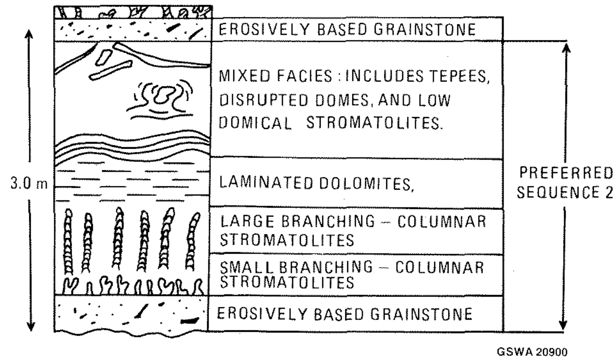


Figure 7. Preferred Sequence 2 (upper part of measured section). Subdivisions of sequence not drawn to scale.

A RECENT ANALOGUE

The various facies and facies sequences outlined in the previous sections are most easily interpreted by comparing them with the coastal deposits of the Trucial Coast, Arabian Gulf.

The morphology of Holocene sediments of the Trucial Coast was studied by numerous workers during the 1960's and 1970's; the results of many of these studies were presented by Purser (1973) and summarized by Till (1978). Briefly, the coastal strip, which is up to 40 km wide in the vicinity of Abu Dhabi, has a complex outer barrier zone of offshore islands, channels and tidal deltas, separated from the mainland by a lagoonal system. Landward of the lagoonal complex is a zone of intertidal algal flats which pass imperceptibly into a dessicated supratidal plain or sabkha. The sediments associated with the outer barrier system are composed of ooids, skeletal sand, and lesser amounts of coral/algal reef material; but the lagoon is characterized by muddy sands and coarser detrius in the lagoonal channels. The intertidal flats are colonized by blue-green algae, which trap sediment and show a variety of algal-mat forms, the morphology of which is controlled by the frequency of storm-driven flooding events. The sabkha surface contains carbonate sediments deposited by exceptional storm events. However, the characteristic feature of the sabkha is the occurence of early diagenetic nodular anhydrite, gypsum, celestite, and ephemeral halite. The growth of anhydrite layers within the sediment (they may constitute half the thickness of the supratidal sediment profile) results in a raising of the sabkha surface: this is balanced by wind deflation and erosion caused by storm flooding.

Stratigraphy of Holocene deposits

The lateral distribution of the surface environments described above is mirrored in the Holocene sediment profile of the coastal sabkha zone; this distribution is due to a marine transgression and regression which occurred during the past 5 000 years. During the initial Holocene transgression, sedimentation rates were too slow to keep pace with the retreat of the shoreline,

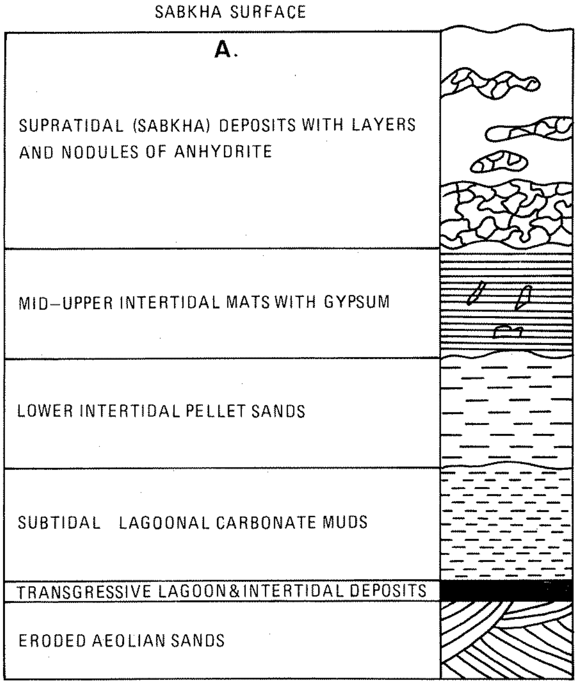
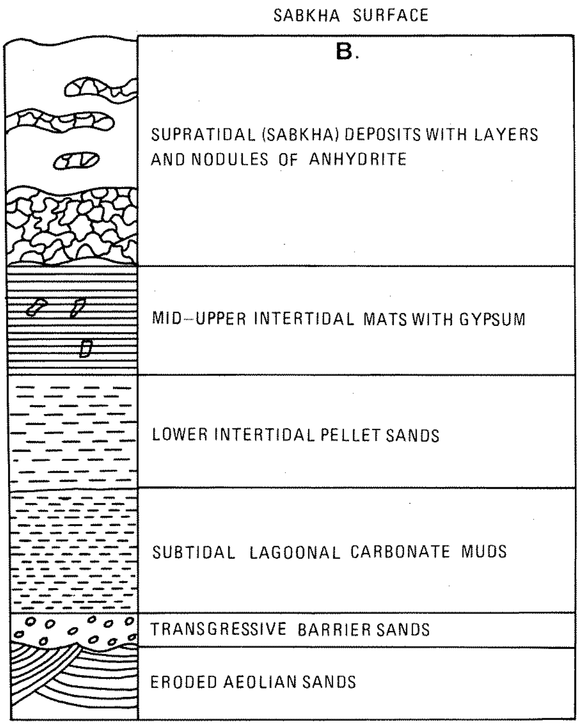


Figure 8. A—Diagrammatic representation of the upward-shallowing sabkha sequence from Abu Dhabi (modified from Till, 1978).



B—Diagrammatic representation of an "ideal" upward-shallowing sabkha sequence.

and only a 10 cm sequence of transgressive subtidal and intertidal sediments is preserved. Subsequent regression, caused mainly by sediment offlap, together with a 1.2 m drop in sea level, has resulted in the deposition, above the transgressive deposits, of up to 4 m of regressive subtidal, intertidal, and supratidal sediments (Fig. 8A). This profile represents one type of upward-shallowing sequence (James, 1979) and could readily have been predicted by applying Walther's Law of Facies to the association of sedimentary environments displayed in this area.

The sequence of lithologies shown in Figure 8A is to some degree incomplete because it does not record the presence of the offshore-barrier deposits. During the Holocene transgression, these sands did not advance as far as the present day sabkha zone. Had this transgression migrated far enough inland, the sabkha profile would be expected to show the complete sequence of barrier, lagoonal, intertidal, and supratidal deposits shown in Figure 8B.

PALAEOENVIRONMENTAL INTERPRETATION

Preferred Sequence 1

Figure 9 compares the ideal sabkha sequence from the Trucial Coast with Preferred Sequence 1 from Duck Creek Gorge. The erosively based grainstone that occurs at the bottom of Preferred Sequence 1 is equivalent to the ooid and skeletal-sand horizon of the ideal sabkha sequence, and represents the remnants of a transgressive offshore barrier system. The largest clasts occurring within the Precambrian intraclast grainstone are derived from (at least partially) lithified lagoonal, intertidal, and supratidal deposits. As discussed by Kraft (1971), a thin transgressive record, such as occurs in both the Duck Creek Gorge and Trucial Coast sequences, points to a slow rise in relative sea level, because such a slow rise allows shoreface erosion to remove most of the retreating shoreline sediments soon after deposition.

The thin (1 to 5 cm) layers of intraclast grainstone that occur randomly within the sequence are interpreted as storm deposits. Park (1976) notes the storm sedimentation is probably the single most important depositional process in the intertidal zone of the Trucial Coast; sediment layers greater than about 1 cm in thickness are attributed to storm activity which was sustained for several days.

The nucleated-domical-stromatolite facies is interpreted as a subtidal lagoonal deposit despite the fact that it apparently has no exact counterpart in the recent sediments of the Arabian Gulf. Kinsman and Park (1976) record small, gelatinous, isolated domes associated with bioclastic muddy sands in some Trucial Coast lagoons. Those stromatolites, however, differ in detailed morphology and internal structure from the domes at Duck Creek Gorge. More closely analogous to the Precambrian domical stromatolites are the subtidal forms reported by Gebelein (1976), from Florida, the Bahamas, and Bermuda, where the stromatolites grow up to 10 cm high and 30 cm wide and are associated, particularly in the moderately agitated environments, with oncoids.

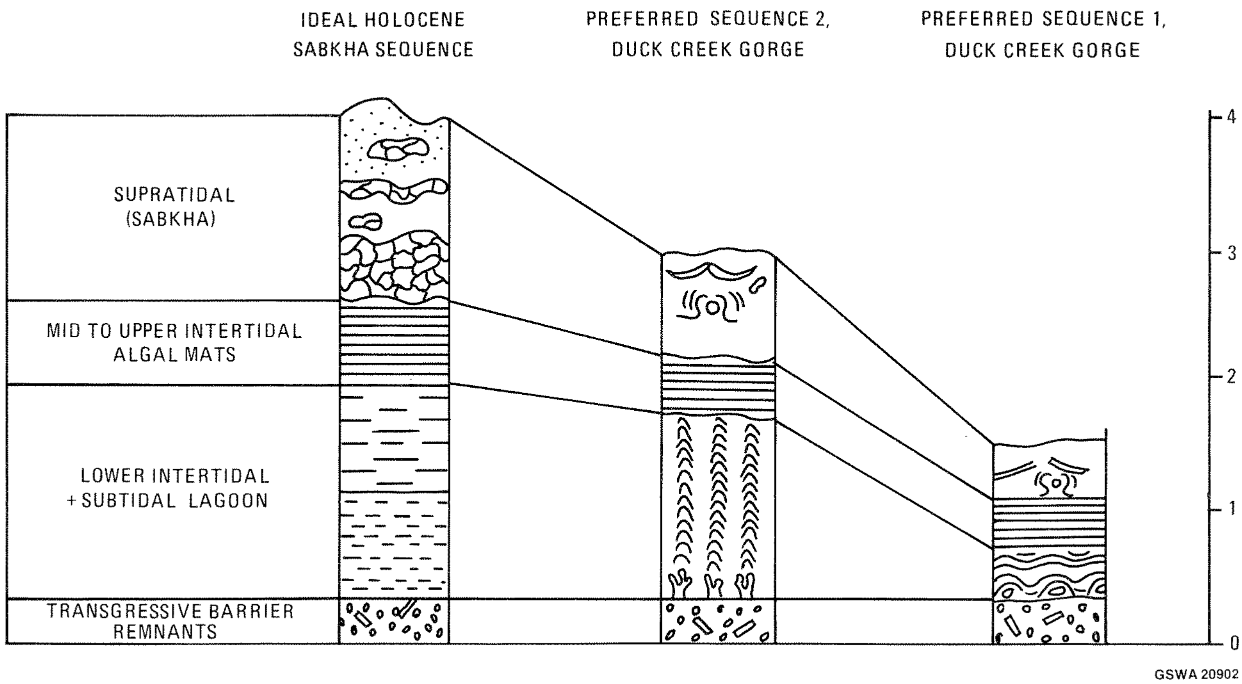


Figure 9. A comparison of lithologies and thicknesses between an ideal Holocene sabkha sequence and Preferred Sequences 1 and 2 from Duck Creek Gorge.

The laminated-dolomite facies, which overlies the nucleated-domical-stromatolite facies, is closely analogous to the laminated middle intertidal sediments of the ideal sabkha sequence. Park (1976) describes how these Recent laminites consist of alternations of blue-green algae and layers of trapped sediment, many of the latter being deposited during storm events. Rotting of the algal layers gives rise to the development of irregular laminoid fenestrae; however, if the decay is accompanied by dehydration, compaction, and lithification, thin layers of finely crystalline carbonate would result instead (Park, 1977).

The low-domical-stromatolite, cuspate-stromatolite, disrupted-domical-stromatolite, and tepee facies, are grouped together as a zone of mixed facies equivalent to the high-intertidal and supratidal portion of the ideal sabkha sequence.

The cuspate stromatolites and low domical stromatolites have no precise analogue in the Recent sediments of the Arabian Gulf. The cuspate stromatolites are loosely comparable with the pinnacle algal mat described by Park (1976) from high-intertidal environments on the Trucial Coast, though these Recent stromatolites apparently do not develop a ridge-like morphology. As described previously, the internal structure of the low domical stromatolites is often closely similar to that of the laminated dolomite facies, which is interpreted as an intertidal deposit. These domes, which occur above the laminated dolomites in the preferred sequence, are therefore likely to be a high-intertidal form. Low domes whose internal structure is characterized by layers of tiny aborescent stromatolites are more problematical. These forms possibly represent stromatolite growth in laterally restricted, hypersaline pools on the very high-intertidal of supratidal surfaces.

The tepee structures, diagnostic of facies G, represent the buckled margins of large polygonal structures formed in the supratidal zone by the dessication, wetting, cementation, and mechanical fracturing of sediment layers (Assereto and Kendall, 1977).

Relic evaporite textures and mineralogies

A characteristic feature of modern sabkha sequences is the association of evaporite minerals, notably gypsum and anhydrite, with dolomitized algal-laminated sediments. The diagenesis of these evaporite minerals can be particularly complex; they are commonly subject to recrystallization and solution, and replacement by a variety of minerals. No occurrences of evaporite minerals have yet been found in the Duck Creek Dolomite; however, evidence in the form of relic textures and replacement minerals points to their former widespread development.

Structures closely comparable with the disrupted domical stromatolites of facies F are reported to be forming just below Recent sabkha surfaces and result from the displacive growth of anhydrite within the sediment; this often leads to the formation of small anhydrite diapirs (Shearman 1980).

As described earlier, facies C, D, E, F, and G (all interpreted as being intertidal or supratidal in origin) often contain irregular fenestrae (1 to 20 cm wide) that cut across, merge into, or disrupt the sedimentary layering. The internal structure of some of those fenestrae (fig. 3E) is very similar to the chicken-wire texture resulting from the growth of anhydrite (often as a replacement after gypsum) in Recent sabkha profiles. In many of the dolomites from Duck Creek Gorge, the original evaporite mineral has apparently undergone *in situ* replacement by anhedral dolomite spar. In other cases, the replacement dolomite occurs as small rosettes or layers of subhedral bladed crystals. The large anhedral to subhedral quartz crystals and chalcedony that infill many fenestrae are commonly found as replacements after evaporites in ancient sabkha sequences. Chalcedony that is optically length slow is, in particular, regarded as a reliable indicator of original evaporite minerals (Folk and Pittman, 1976).

The above comparison between Preferred Sequence 1 from the lower part of the Duck Creek Gorge section and the ideal sabkha sequence shows that they are similar, both in the types of lithology present and in the vertical distribution of these lithologies. From this evidence, the lower part of the Duck Creek Gorge section can be interpreted as comprising numerous upward-shallowing sequences caused by repeated marine transgressions and regressions across a sabkha-type coastline.

Preferred Sequence 2

Preferred Sequences 1 and 2, obtained from the lower and upper parts of the Duck Creek Gorge section respectively, are similar both in the type of lithology and in the vertical arrangement of these lithologies. On this basis, Preferred Sequence 2 is also interpreted as being an upward-shallowing sabkha-type sequence.

A characteristic feature of Preferred Sequence 2 is the presence of layers of small branching-columnar stromatolites and large, branching-columnar stromatolites, which occur between the basal intraclast-grainstone facies and laminated-dolomite facies—*i.e.* in place of the nucleated-domical-stromatolite facies. Small branching-columnar and large, branching-colunmar stromatolites, comparable to those found at Duck Creek Gorge, have not been recorded in Recent sedimentary environments. Columnar stromatolites presently found in subtidal settings at Shark Bay (Playford and Cockbain, 1976) are loosely comparable to the Duck Creek Gorge examples in their general form but differ markedly in details of internal and external structure.

In the absence of any suitable modern analogue, we have to rely on the position of the small branching-columnar and large, branching-columnar-stromatolites in the upward-shallowing sequence to provide evidence concerning their original environment of deposition. As shown in Figures 7 and 9, the stromatolitic facies in question are found above the intraclast-grainstone facies (the remnants of the transgressive barrier deposit) and below the intertidal laminated-dolomite facies. This would indicate that the small, branching-columnar-stromatolite facies is a subtidal lagoonal deposit whereas the large, branching-columnar-stromatolite facies was formed in a subtidal-lagoonal to shallow-intertidal setting.

DISCUSSION

As shown in Figure 9, Preferred Sequence 2 and the ideal sabkha sequence from the Trucial Coast are of comparable thickness; the slightly greater development of the Recent sequence is attributed to the fact that this profile has undergone little compaction. Preferred Sequence 1 is, on average, 1.5 m thick, approximately half the thickness of the other two sequences, and most of the difference is due to the poor development of the lagaoonal to lower intertidal portion of this sequence. The thickness of the subtidal-lagoonal portion of the upward-shallowing sequence (prior to compaction), largely reflects the depth of the original lagoon. This indicates that the lagoonal complex, poorly developed in the early history of this shoreline, became deeper and probably wider with time. This environment, with its higher wave energies (and possible tidal channels), was apparently a favourable setting for the growth of the small, branching-columnar and the large, branching-columnar stromatolites found in the upper part of the measured section.

In many coastal sabkhas of the Abu Dhabi region, zones of nodular anhydrite make up at least half the thickness of the supratidal deposits (Shearman, 1980). Relic evaporite textures present in the Duck Creek Dolomite, however, form only 10 to 15% of the supratidal lithologies. Even allowing for the fact that some of the evaporites originally present in the Duck Creek Dolomite may have vanished without trace during erosion or diagenesis, it is unlikely that they were ever quite as extensive as they are in the present-day coastal sabkhas of the Trucial States. This would indicate that a cooler, possibly more humid climate than presently occurs in the Arabian Gulf, prevailed in the Pilbara region during deposition of the Duck Creek Dolomite. No evaporite relics have yet been recorded from the subtidal portions of the Precambrian upward-shallowing sequences. This suggests that the ancient sabkha-type coastline bordered on an ocean whose salinity was equal to or slightly higher than that of presently normal marine waters (Kendall, 1979).

No satisfactory explanation can be given to account for the history of repeated marine transgressions recorded in the Duck Creek Gorge section. Kendall (1979) and White (1981) discuss possible mechanisms for producing repeated transgressions in other carbonate successions—*e.g.* tectonic subsidence or sea level changes induced by glaciations or global tectonism. Any one of these mechanisms acting alone, or in combination, could account for the repetition of upward-shallowing sequences observed in the Duck Creek Dolomite.

Comparisons with other ancient examples of upward-shallowing sequences

Many examples of upward-shallowing sabkha-type sequences are preserved in the geological record, and most are recorded from Phanerozoic successions—*see* summary by Till (1978). Published accounts giving details of this type of sequence in Precambrian rocks are less common. Hoffman (1976) describes the occurrence of almost 200 upward-shallowing sequences in the 2.2 to 1.8 Ga old Rocknest Formation of the Canadian Shield. These sequences differ from those described in this study in that the regressive deposits grade from offshore marine to intertidal sand flats without any intervening barrier or lagoonal sediments. In addition, Hoffman (1976) does not record the presence of relic evaporite textures in any supratidal deposits. Tiny aborescent stromatolites, similar in form to those occurring in the supratidal horizons of Duck Creek George succession, are recorded in equivalent facies in the Rocknest dolomites. Columnar stromatolites closely comparable to those described in this study are, in the Canadian succession, restricted to the low-intertidal facies. White (1981) reports on upward-shallowing sequences in the 1.45 Ga old Altyn Formation of Montana, U.S.A. As with the Rocknest Formation, these sequences do not record any barrier or lagoonal facies; they do, however, contain replacement evaporite textures and minerals in the supratidal deposits.

CONCLUSIONS

This work provides further evidence for the existence of upward-shallowing sabkha-type sequences in the Precambrian sedimentary succession. The association of particular stromatolite morphologies with distinct environments of deposition points to a strong environmental control on their gross morphology. Detailed features of morphology and internal structure, however, might well be the result of biological factors. The "sequence approach" to interpretation used in this study may provide a rigorous method of demonstrating possible evolutionary trends in stromatolites, because forms of differing age and varied location that grew in similar environments of deposition can now be compared.

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STROMATOLITES IN THE PROTEROZOIC DUCK CREEK DOLOMITE WESTERN AUSTRALIA

by Kathleen Grey

ABSTRACT

Stromatolites are abundant in the 2.0 Ga Duck Creek Dolomite, Wyloo Group, Western Australia, although the taxonomic diversity is low. Upward-shallowing sequences occur in the lower part of the dolomite and contain a new form, *Asperia ashburtonia*, in addition to *Pilbaria perplexa* Walter 1972 and *P.* cf. *perplexa*, as well as domical and stratiform stromatolites not given taxonomic status. Each distinct taxa or morphology characterizes a particular facies in the upward-shallowing sequences. An exhaustive summary of the occurrences of stromatolites in the lower part of the dolomite is given.

Stromatolites in the upper part of the Duck Creek Dolomite are less well known and further systematic studies are required. However, the taxa which occur are different from those in the upward-shallowing sequences.

INTRODUCTION

During re-mapping of the Wyloo 1:250 000 sheet (Seymour and others, in prep.), several stromatolitic dolomite samples from the Wyloo Group were collected for identification and palaeoenvironmental intrepretation. Some of these samples consisted of a new form which apparently occurred together with a previously described form in repeated sedimentary sequences. A more comprehensive sampling programme was initiated to obtain additional material and to investigate the possible relationships between stromatolite taxa and environment of deposition.

Stromatolites are abundant in the Duck Creek Dolomite, but few forms are represented. They were first reported from the Wyloo Group by Halligan & Daniels (1964) and Edgell (1964), but the first major systematic study was that by Walter (1972). He described *Pilbara perplexa, Patomia* f. indet.; and an "Unnamed Stromatolite" that was previously incorrectly identified as *Collenia australasica* (Howchin) by Edgell (1964). Another new form, *Asperia ashburtonia,* is described below, and the distribution of stromatolite taxa is documented.

GEOLOGICAL SETTING AND AGE

The Duck Creek Dolomite, a unit in the upper part of the Wyloo Group, crops out extensively (Fig. 1) in the Western Pilbara (Halligan & Daniels, 1964; Trendall, 1975). It conformably overlies the Mount McGrath Formation, and is overlain by June Hill Volcanics or the Ashburton Formation. It is over 1 000 m thick, but stromatolites occur mainly in the lower 300 m. Thorne (1985) carried out a Markhov Chain analyses on the lower 220 m of the dolomite, where it is well exposed in Duck Creek Gorge (Fig. 1), and recognized nine major sedimentary facies. The facies occur in repeated upward-shallowing sequences, and the distribution of stromatolite taxa is closely linked to the type of facies (Grey & Thorne, in prep).

The precise age of the Duck Creek Dolomite is not known, but Gee (1980, Fig. 1) favoured an age of approximately 2.0 Ga, and suggested correlation with the Glengarry Group in the western part of the Nabberu Basin. Clasts from the Woongarra Volcanics (in the underlying Mount Bruce Supergroup) occur at the base of the Wyloo Group and provide a lower age limit, based on a U-Pb zircon age of 2 470± 30 Ma (Compston and others, 1981). West of Mount Stuart (outside the area shown on the map) the overlying Ashburton Formation is intruded by the Boolaloo Granodiorite, for which Leggo and others (1965) obtained a Rb/Sr mineral isochron of 1.68 Ga. Rb/Sr whole-rock isochron dates from the Wyloo Group of *1 977 ±165 Ma from "acid igneous rock" (Compston and Arriens, 1968) and *1 811 Ma from "tuffaceous siltstone" (Leggo and others, 1965) are probably less reliable (Seymour, pers. comm. 1984) although they fall within the upper and lower limits referred to above.

STROMATOLITE OCCURRENCES

Localities referred to in the text below are listed in Table 1 and shown in Figure 1, and are identified by GSWA fossil locality numbers, in which the relevent 1:250 000 geological sheet is given a three-letter code,

*recalculated values using ^{87}Rb decay constant = $1.42 \times 10^{-11} a^{-1}$

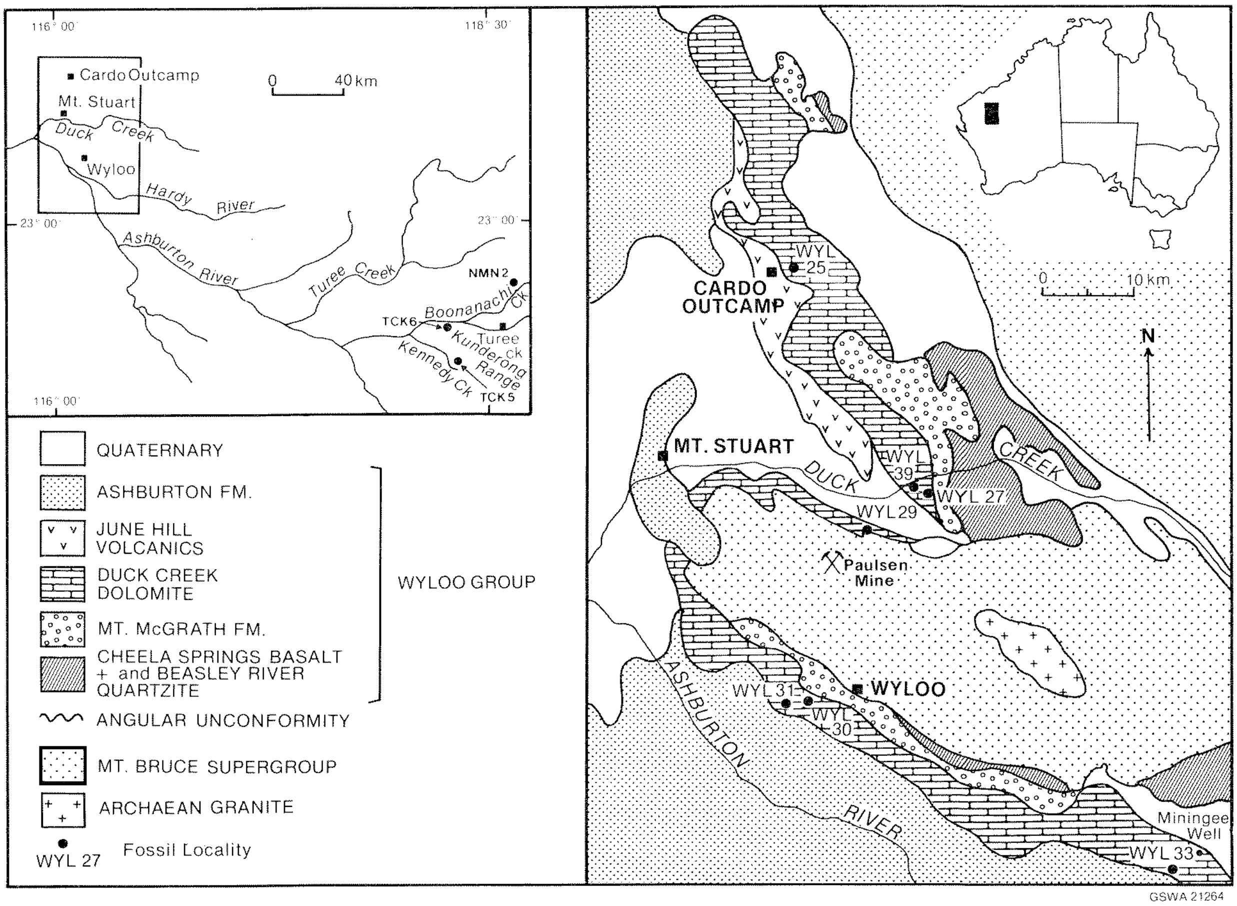

Figure 1. Part of the Western Pilbara, W.A., showing fossil localities referred to in the text, and a detailed geological sketch map of the Wyloo area.

e.g. WYL 31 for Wyloo, location 31. The fossils themselves (including holotypes) are stored either in the GSWA collection (prefix F) or in the Geology Department Collection, University of Adelaide (prefix S).

Pilbaria perplexa Walter 1972 and the new form *Asperia ashburtonia* both occur in the lower part of the Duck Creek Dolomite, in different facies of the upward-shallowing sequences. (Grey & Thorne, in prep.). *Patomia* f. indet. is known only as a single specimen, possibly from the upper part of the formation. Details are also given for other stromatolites not yet assigned to taxa, either because of poor preservation or lack of material.

Stromatolites in the lower part of the Duck Creek Dolomite are best exposed in the Duck Creek Gorge section (WYL 39 in Fig. 1). Thorne (1985) has logged in detail the lower 300 m of the section which contains the upward-shallowing sequences, and Grey and Thorne (in prep.) discuss the significance of the distribution of stromatolite taxa in the sequences described by Thorne (1985).

Upward-shallowing sequences are approximately 1.5 m thick in the lower part of the section, Sequence 1 of Thorne (1985), and consist of an erosively based grainstone, a nucleated-domical-stromatolite facies, a laminated-dolomite facies, and a mixed facies which includes tepees, disrupted domes, cuspate stromatolites, and low-domical stromatolites. The sequence is overlain by the next erosively based grainstone. Sequences in the upper part of the measured section are usually about 3 m thick; and, in the preferred sequence, Sequence 2 of Thorne (1985), the nucleated-domical-stromatolite facies is missing, and is replaced by a small, branching-columnar-stromatolite facies and a large, branching-columnar-stromatolite facies.

Of the stromatolites which occur in the gorge section, only three types have been accorded taxonomic status. These are *P. perplexa*, *P.* cf. *perplexa* and *A. ashburtonia*. Nucleated-domical stromatolites; oncolites, which occasionally occur in the grainstone facies, and stratiform stromatolites, which comprise much of the laminated-dolomite facies, do not have characteristics sufficiently distinctive for systematic treatment.

The low-domical stromatolites (Fig. 2) in the mixed facies consist of closely packed columns of *Asperia ashburtonia*, which, as discussed by Grey (1984) and Grey and Thorne (in prep.), most probably grew in subaerial conditions subject to intermittent periods of submergence, possibly seasonal in origin. This is consistent with the high-intertidal to supratidal environment postulated by Thorne (in prep) from the

TABLE 1. DETAILS OF STROMATOLITE LOCALITIES

Fossil locality number	*Sample No.*	*Identification*	*Locality description*	*Map*	*Latitude and longitude*
NMN 2	F 12754	New group and form (previously *Patomia* f. indet)	near Boonoonachi Creek 15km north northeast of Turee Creek homestead	Newman	23°28'15"S 118°37'11"E
TCK 5	F 8134	*Patomia* f. indet. (name requires revision)	(from Walter, 1972) "26 miles east-northeast of Mount Bresnahan"; near Kennedy Creek (locality position doubtful D. B. Seymour pers. comm., 1984)	Turee Creek	23°44'09"S 118°18'48"E
TCK 6	F 46625	unnamed form	3 km west of west end of Kunderong Range	Turee	23°35'37"S 118°26'10"E
WYL 25	F 46623 F 46624	*Pilbaria perplexa* *Asperia ashburtonia*	6.4 km east Cardo Outcamp	Wyloo	22°16'30"S 116°12'46"E
WYL 27	S 206	*Pilbaria perplexa* (holotype)	(from Walter, 1972) "20 miles east of Mount Stuart homestead . . . beside the track" (the track is an old one, about 2 km south of Duck Creek Gorge M. R. Walter, pers. comm., 1983)	Wyloo	22°29'07"S 116°20'04"E
WYL 29	F 46610 F 46611 F 46612 F 46613 F 46614 F 46620 F 46621	*Pilbaria perplexa* *Pilbaria perplexa* *Pilbaria perplexa* *Pilbaria perplexa* *Asperia ashburtonia* *Asperia ashburtonia* (holotype) *Asperia ashburtonia*	5 km north east of Paulsen mine	Wyloo	22°32'31"S 116°15'46"E
WYL 30	F 9980 F 46622	*Pilbaria perplexa* *Pilbaria perplexa*	(from Walter, 1972) "between Wyloo and the homestead" (The locality is in fact between Wyloo hst and the woolshed, 5.5 km south-west of Wyloo homestead)	Wyloo	22°43'45"S 116°11'13"E
WYL 31	—	*Pilbaria perplexa*	9.9 km west-southwest of Wyloo homestead	Wyloo	22°43'30"S 116°08'00"E
WYL 33	F 12519	unnamed columnar stromatolite	4.5 km southwest of Miningee Well	Wyloo	22°52'36"S 116°36'21"E
WYL 39	F 46615,16 F 46617,18 F 46619	*Pilbaria perplexa* *Asperia ashburtonia* *Pilbaria* cf. *perplexa*	Duck Creek Gorge	Wyloo	22°28'53"S 116°18'40"E

sedimentological evidence. Bioherms in the small branching-columnar-stromatolite facies are formed by *Pilbaria perplexa* (Fig. 3) which probably grew in shallow lagoons with rare periods of emergence (Grey and Thorne, in prep.) The large branching-columnar-stromatolite facies consists of *Pilbaria* cf. *perplexa* (Fig. 4), which most probably developed in a higher energy, intertidal environment. These conclusions, based on interpretation of stromatolite morphology (Grey and Thorne, in prep.) are consistent with Thorne's interpretaions based on sedimentological data.

Elsewhere in the Duck Creek Dolomite, upward-shallowing sequences are less readily recognizable because outcrop is poorer. Walter (1972) described *P. perplexa* (F 9980) from near Hardy Junction (Fig. 1, WYL 30). Preservation is poor, but *P. perplexa* can be recognized in outcrop. *A. ashburtonia* is absent. A large bioherm of *P. perplexa* crops out 4 km further west of the above locality (Fig. 1, WYL 31). Here, the rocks are steeply dipping and strongly cleaved. The stromatolite columns have been deformed (Fig. 5), and their cross-sections are oval. The laminations parallel to the direction of compression are steeply convex, and column interspaces are almost totally destroyed. *A. ashburtonia* is absent and the relationship of the bioherm to strata above and below cannot be determined.

At the type locality of *P. perplexa* (Fig. 1, WYL 27) Walter (1972) reported "tabular biostromes about 0.5 m thick and much thicker (several metres) beds of unknown shape". This mode of occurrence seems similar to the *P. perplexa* horizons in Duck Creek Gorge; however, he did not report any small, digitate stromatolites (here named *ashburtonia*) from this locality and the area was not visited during the present study.

GSWA 21265

Figure 2. Part of a low-domical bioherm consisting of closely packed, silicified columns of *Asperia ashburtonia* new form—Duck Creek Gorge, WYL 39.

GSWA 21266

Figure 3. Part of a column of *Pilbaria perplexa*, showing typical niche and pocket development. Polished slab of GSWA F 46610—5 km north-east of Paulsen Mine, WYL 29.

At the type locality of *A. ashburtonia* (Fig. 1, WYL 29), the Duck Creek Dolomite overlies the Mount McGrath Formation, and is fault-bounded by the Brockman Iron Formation of the older Hamersley Group. Preservation of the stromatolites is particularly good and collecting is easy from the rubbly outcrop. *P. perplexa* is abundant at this locality, although only two horizons of *P. perplexa* separated by one of *Asperia ashburtonia* can be recognized. Flat-laminated dolomite layers also occur, but detailed sedimentological relationships cannot be determined because of the rubbly nature of the outcrop.

The Duck Creek Dolomite crops out patchily in a series of low ridges parallel to the track for about 2 km east of the *A. ashburtonia* type locality. *A. ashburtonia* does not occur, although *P. perplexa* is common at some outcrops, but is nowhere well exposed or well preserved. The "Unnamed Stromatolite" of Walter (1972, p. 177), which in this paper is placed in *P. perplexa*, was collected from somewhere near here, although the precise locality could not be relocated.

North of Duck Creek, well-preserved stromatolites occur at only one locality (Fig. 1, WYL 25), where poorly outcropping dolomite at the base of a ridge north of the track contains two *P. perplexa* horizons separated by an *A. ashburtonia* horizon. The stromatolites form a series of low benches; but because of the sporadic outcrop, the nature of the sediments separating the biostromes cannot be determined. Both stratiform and domical stromatolites are present, suggesting that the sequence, though not as extensively developed, is similar to the Duck Creek gorge section.

The stromatolite taxa which occur in the lower part of the Duck Creek Dolomite do not seem to be present in the upper part of the formation. Little is known about the distribution of stromatolites above the upward-shallowing sequences, and all of them require systematic study. Walter (1972) described *Patomia* f. indet. (F8134) probably from the upper part of the Duck Creek Dolomite (Fig. 1, TCK 5). He stressed the tentative basis on which the specimen was placed in *Patomia*. The degree of curvature of the laminae is more variable, the column margins are more ragged,

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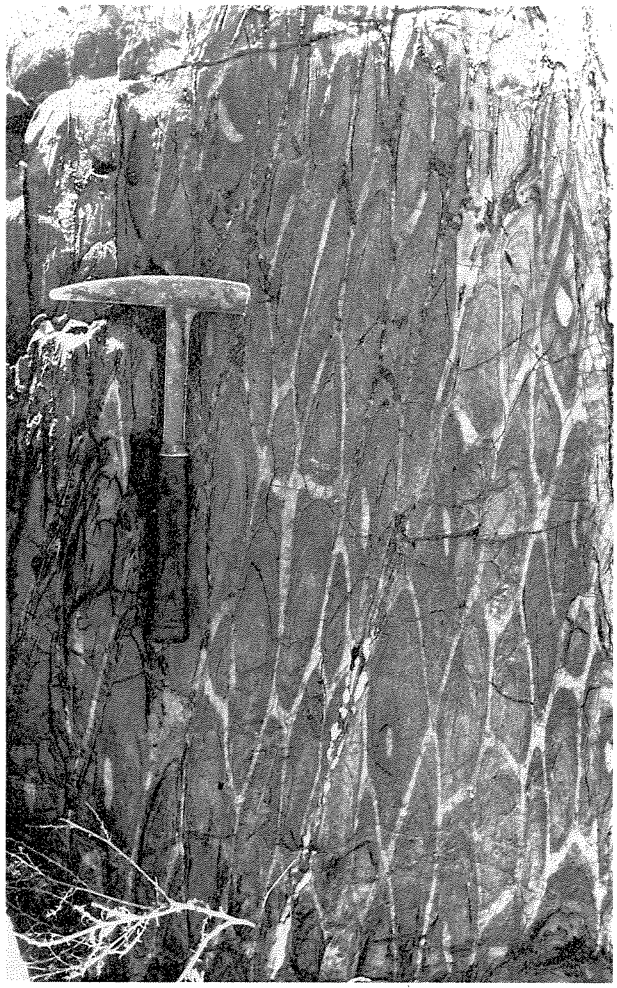
GSWA 21267

Figure 4. *Pilbaria perplexa* biostrome in the lagoonal facies, overlain by the larger, and more irregular, *Pilbaria* cf. *perplexa*. This in turn is succeeded by intertidal laminated dolomite facies —Duck Creek Gorge, WYL 39.



GSWA 21268

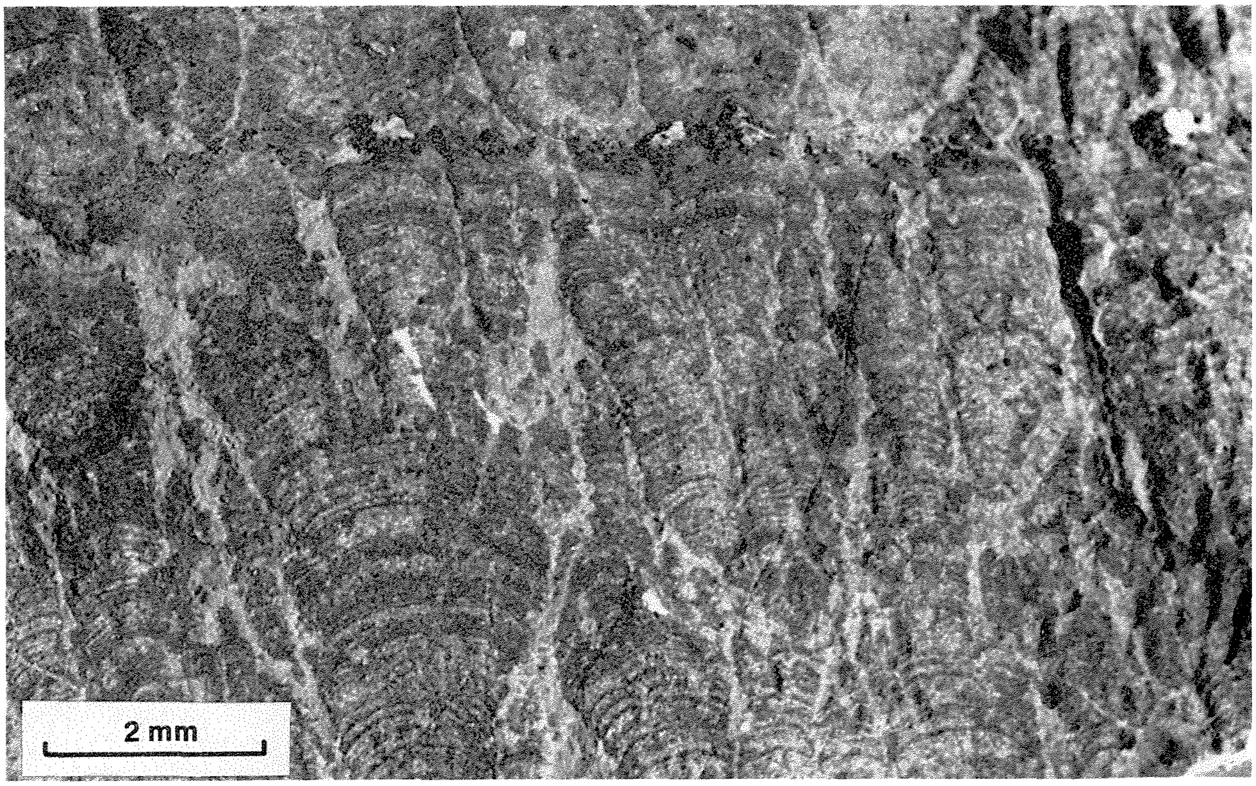
Figure 5. Deformed bioherm of *Pilbaria perplexa* showing tangential sections of columns. Note the reduced intercolumnar area —From 9.9 km west-south-west of Wyloo Homestead, WYL 31.

and the branching pattern more divergent than in *Patomia*. Some of the features of this specimen are also found in several groups erected since the publication of Walter's description. These include *Sundia* Butin 1966 and *Tibia* Bertrand-Safati and Eriksson 1977. Identification of the Duck Creek Dolomite form is not possible until better preserved material is available; however, only laminated stromatolitic dolomite has since been found at the reported locality (D. B. Seymour pers. comm. 1984).

Other stromatolite localities have been recorded from the (possibly) upper Duck Creek Dolomite (Walter, 1972; Grey, 1979), but many remain unsampled. Three have been the subject of unpublished reports (Grey, 1981 a, b, c) but none of the stromatolites can be placed in existing taxa. A small fragment (F 12519) from one locality (Fig. 1, WYL 33), resembles undescribed and poorly preserved stromatolites from the Glengarry Group (Grey, 1981a). A second (Fig. 1, NMN 2) was at first (Grey, 1981b) placed in *Patomia* f. indet. on the basis of one small sample (F 12754). Additional material suggests that the stromatolite belongs to a new group and form that are awaiting description. The stratigraphic position of the dolomite unit is uncertain. It may be Duck Creek Dolomite or possibly Mount McGrath Formation, but it could also belong to the Turee Creek Group. The third stromatolite (Grey, 1981c, F 46625) is also from a unit of uncertain stratigraphic position (Fig. 1, TCK 6). It is a small digitate form, still awaiting description, but differs in several respects from *Asperia ashburtonia*. Studies are continuing on all three stromatolites.

CONCLUSIONS

This study has shown that stromatolites in the Duck Creek Dolomite are restricted to relatively few taxa. *Pilbaria perplexa*, *Pilbaria* cf. *perplexa* and *Asperia ashburtonia* are common in the lower part, where upward-shallowing sequences are well developed (Thorne, 1985). The relationship between the distribution of stromatolite taxa and sedimentary environment can best be studied in the well-exposed



GSWA 21269

Figure 6. Weathered surface of *Asperia ashburtonia* showing branching pattern and banded lamination—From 5 km north-east of Paulsen mine, WYL 29.

section in Duck Creek Gorge. Here it can be shown that each of the three taxa is restricted to a specific facies (Grey and Thorne, in prep). Elsewhere the association between stromatolite taxa and depositional environment cannot be as clearly shown. *P. perplexa* has been reported from more localities than the other two taxa, but this could, in part, be due to problems of recognition in the field, particularly in the case of *A. aspera*. Stromatolites belonging to different taxa occur in the upper part of the Duck Creek Dolomite, but are poorly known.

If the rather dubious identification of the middle to late Proterozoic group *Patomia* is discounted, as discussed above, the occurrence of the other taxa is consistent with known time distributions overseas (Grey, 1984; Grey and Thorne, in prep.). In particular *P. perplexa* and *Asperia* indicate an early Proterozoic age of between about 1.8 and 2.1 Ga.

SYSTEMATIC DESCRIPTIONS

Asperia Semikhatov 1978

Type form

Asperia aspera Semikhatov 1978, Rocknest Formation (Unit 8), Epworth Group, Asiak River Basin, Canadian Shield.

Diagnosis

"Small (usual diameter 3-8 mm, usual height 10-30 mm), tightly contiguous, digitate, subcylindrical, branching and coalescing wall-less columns, associated in massive segments by somewhat partial transitional bridges and (or) covered by relatively thick (several millimetres) extended bundles of finely crimped laminae. Branching is frequent, by means of single successive breakup of the mother column into subparallel or very gently divergent thinner columns (passive branching), lateral surface with fine wrinkles and cornices. Good inheritance of lamination" (Translated from Semikhatov, 1978, p. 120-121).

Content

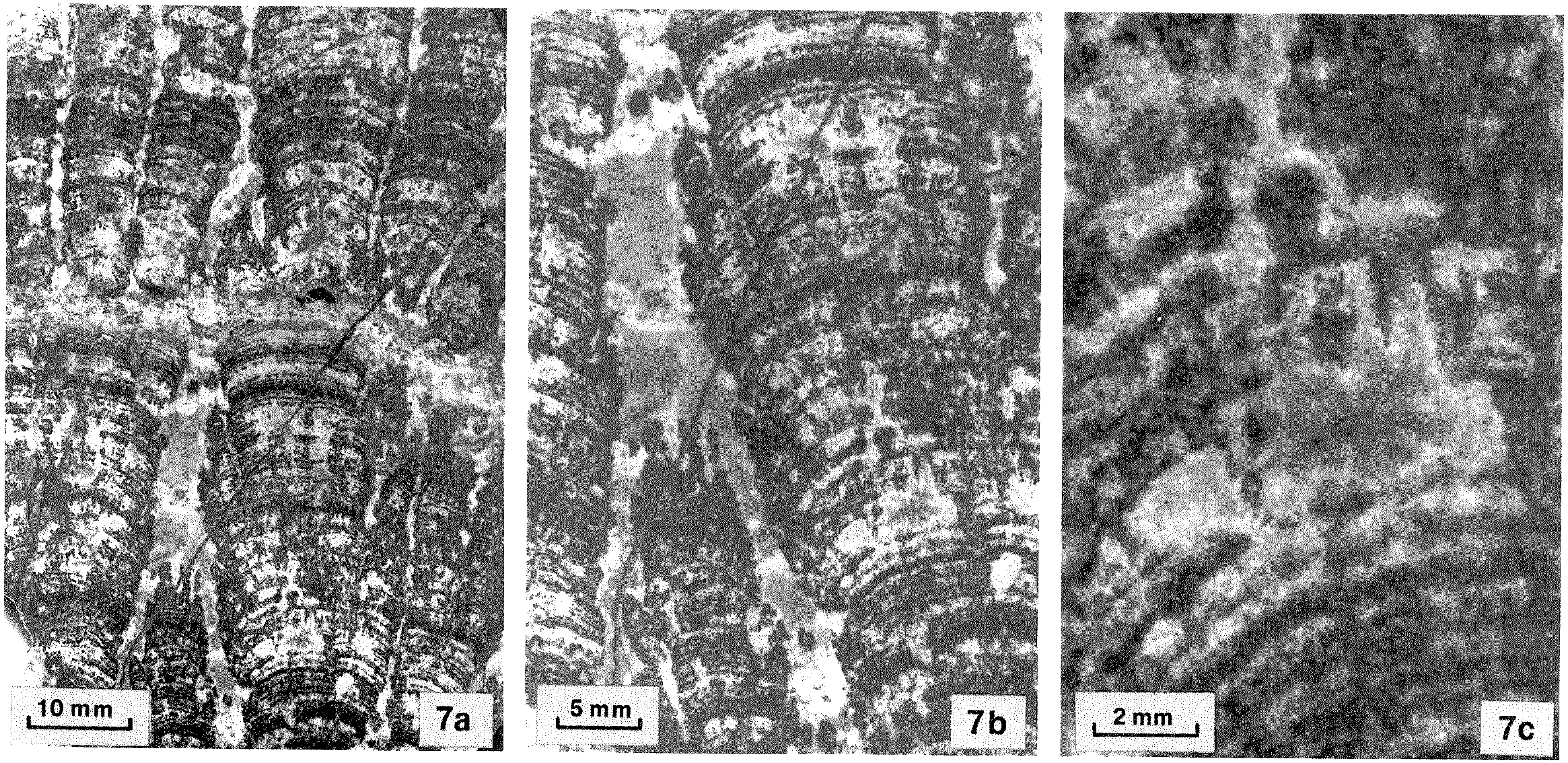
Asperia aspera Semikhatov 1978. *Asperia ashburtonia* form nov.

Remarks

Semikhatov (1978, p. 120) placed the small digitate stromatolites reported (but not formally described) by Donaldson (1963) into synonomy with *Asperia*, and also indicated that forms from Canada (Belcher Group) and Afghansistan should be placed in *Asperia*. However, none of these stromatolites have been described at form level; and, while there is considerable justification for assigning many of the informally documented occurrences listed in Table 2 to the group *Asperia*, this should only be done as part of a thorough systematic study.

Distribution and age

Rocknest Formation, Epworth Group, Great Slave Lake area, northern Canada, age approximately 1.9 Ga; Duck Creek Dolomite, Wyloo Group, western Pilbara, Western Australia age approximately 2.0 Ga.



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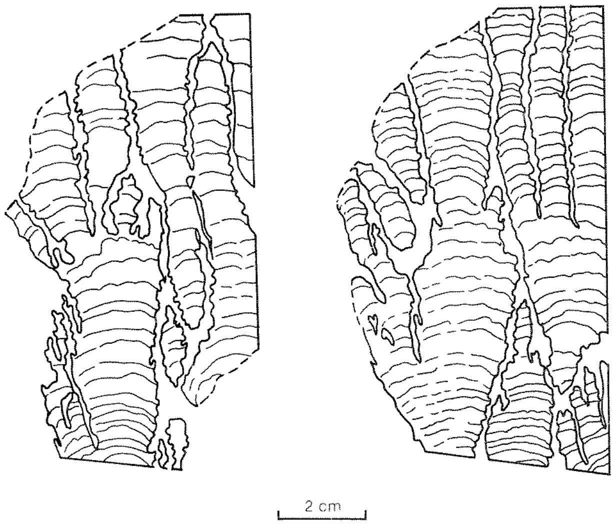

Figure 8. Holotype of *Asperia ashburtonia* new form, GSWA F 46620; laminae profiles drawn from polished slabs.

Asperia ashburtonia new form

(Figs 2, 6-8)

Material

Holotype: GSWA F 46620, from WYL 29 (Fig. 1)

Paratypes: F 46614 from the type locality, F 46617 and F 46618 from WYL 39, F 46624 from WYL 25, all from Wyloo 1:250 000 scale map area (Fig. 1).

Derivation of name

From the Ashburton River near the type locality.

Diagnosis

Branching, columnar stromatolites form large domes without well-defined fascicles (Grey, 1984), with erect, slender, subparallel to slightly divergent, closely spaced columns and beta-style branching. Columns arise from a stratiform base. Branching into two, three, or often more, columns of equal width is common. Some columns show a gradual upwards increase in width. Laterally occurring daughter columns are occasionally present. Walls are not developed. Column margins vary from scalloped to smooth, Laminae are flat to very gently convex and smoothly curved. Slight flexures appear prior to branching. Microstructure is distinctly banded; laminae of similar fabric and comparative thicknesses occur in adjacent columns.

Description

Outcrops: The type locality and other outcrops are described in the section of this paper dealing with the Duck Creek Dolomite. The base of the *Asperia ashburtonia* horizon at the type locality is poorly exposed. Columns develop directly from a stratiform horizon and form large contiguous bioherms up to 1 m in diameter. Tops of bioherms are eroded.

Mode of occurrence: Bioherms are domical, formed from closely-packed columns arising from a basal layer of stratiform laminae. Bioherms are approximately 30 cm high, and show very little variation throughout.

Branching habit: Branching is multifurcate, or occasionally bifurcate. Daughter columns are of equal width; they may gradually widen or taper upwards. Adjacent branches sometimes interfere with the growth pattern of one another. In general, branches are divergent, but where they are closely spaced, columns are usually parallel. Column interspaces are of more or less uniform width (1 to 3 mm).

Column shape and margin structure: Individual columns are usually cylindrical, but many widen gradually and in some cases, double in width in about 10 mm of height. Columns rounded to lobate in cross-section. Columns slender, erect and only rarely constringed. Wall is not present. Column margins are smooth to irregular with a scalloped appearance. Bridging occurs rarely. Columns vary in height, but many are up to 20 cm. Diameters vary from 10 to 15 cm.

Lamina shape: Lamina profiles are smooth and gently convex, with slight development of flexures before branching. Individual laminae can be traced as distinctive horizons across many adjacent columns. The serial development of laminae shows a high degree of inheritance.

Microstructure and texture: Microstructure and texture are distinctly banded with well-defined light and dark laminae. Individual laminae are of constant thickness across the column width, but the thickness of successive laminae varies considerably from only a few micrometres up to 3 mm. A secondary banding pattern, in which macrobands, 1 to 2 mm thick, consist of finer laminations, can usually be detected. The overall colour of the macrolaminae is determined by the thickness of the component microlaminae. Thickness of microlaminae is difficult to determine because of poor preservation. Both light and dark laminae are approximately 250 µm thick. Light laminae are homogeneous throughout, but dark laminae show a gradation in colour intensity from relatively light at the base to almost opaque at the top. The upper boundary is usually well defined, but where recrystallization has disrupted the laminae, micro-flaring occurs, giving the boundary a ragged appearance.

The original nature of the microstructure cannot be determined because of recrystallization. Grains are approximately 250 µm in diameter and equigranular.

Interspace filling: Interspaces are filled with micritic dolomite. Laminations cannot be recognized in the interspace fillings. Details of microstructure have been destroyed by recrystallization.

Secondary alteration: Extensive recrystallization and dolomitization has altered much of the original fabric. Only a coarse mosaic of secondary grains is present. At Duck Creek Gorge secondary silicification has been widespread. Interspace areas have been replaced by a coarse mosaic of quartz crystals. Extensive silicified patches occur in the columns, often destroying laminations.

Comparisons

Both *Asperia ashburtonia* and *Asperia aspera* Semikhatov 1978 are very similar in shape and fabric, and show a wide diversity in morphology. Features characteristic of one form may occasionally be present in the other. The principal differences are that the laminae in *Aperia aspera* are flatter and more wrinkled than in

Figure 7. A.—Holotype of *Asperia ashburtonia* new form, GSWA F 46620. Polished slab showing branching pattern and lamination—From 5 km north-east of Paulsen mine, WYL 29.

B—Detail of part of Fig. 7A showing banded microstructure.

C—Detail of part of Fig. 7B showing microstructure and effect of recrystallization.

Asperia ashburtonia, and are slightly more complex because of the development of double-layered dark laminae (Semikhatov, 1978, Pl.XIII, Fig. 2). Branching is more frequent in *A. ashburtonia*, and is more commonly multifurcate. These differences are relatively minor ones, and it may be that the two forms are synonymous. It will probably be necessary to resort to statistical methods (Hofmann, 1977) to confirm that the forms can be differentiated. However, from non-morphometric comparisons of type material, the distinctions appear to be valid.

A. ashburtonia differs from the digitate stromatolites in the Denault Formation (Donaldson, 1963; Semikhatov, 1978) in having multifurcate (rather than bifurcate) branching, flatter lamitations and broader columns. Anastomosing of columns is common. The columns of digitate stromatolites from the Belcher Group (Hofmann, 1977; Semikhatov, 1978) are shorter and stubbier than those in *A. ashburtonia*. In the Hornby Bay Group (Baragar and Donaldson, 1974) the digitate stromatolites have much narrower columns with very irregular margins, and more frequent and abundant branching than in *A. ashburtonia*.

In *Yelma digitata* Grey (1984), the columns are narrower and linked together in fascicles. Branching in *Y. digitata* is sometimes more widely divergent than in *A. ashburtonia*. Laminae in *Y. digitata* are thicker, less constant in thickness, and they are of three kinds, rather than two as in *A. ashburtonia*.

Katernia africana Cloud & Semikhatov 1969 and *Katernia perlina* Bertrand-Sarfati & Eriksson 1977 both have more divergent branching than *A. ashburtonia*, and their microstructure is filmy rather than banded. *Lenia jacutica* Dolnik (in Dolnik and Vorontsova 1971) has bifurcating and lateral branching, rather than multifurcate branching, and has more irregular columns.

Remarks

The small columns and close spacing make serial reconstruction of *A. ashburtonia* difficult, and only laminae tracings (Fig. 10) are shown. Visual comparison of digitate stromatolites is obviously a very subjective method, and in order to delimit the subtle differences which occur between forms, it will be necessary to resort to statistical methods. Hofmann (1977) described a computer assisted method of morphometric analysis which he applied to the digitate stromatolites from the McLeary formation (Hofmann assigned these to *Lenia* f.). The methods involved are relatively sophisticated, but would probably provide a suitable means for distinguishing between the various asperiform stromatolites in future studies.

Distribution and age

Duck Creek Dolomite, Wyloo Group, Western Australia, age approximately 2.0 Ga.

Pilbaria Walter 1972

Type form

Pilbaria perplexa Walter 1972, Duck Creek Dolomite, Wyloo Group, western Pilbara, Western Australia.

Diagnosis

As in Walter, 1972, p. 167.

Content

Pilbaria perplexa Walter 1972, *Pilbara boetsapia* and *Pilbaria inzeriaformis* Bertrand-Sarfati and Ericksson 1977, *Pilbaria deverella* Grey 1984, *Pilbaria* cf. *perplexa* Walter in Zhu Shixing (1982), cf. *Pilbaria perplexa* Walter in Crick & others (1980).

Distribution and age

Duck Creek Dolomite, Wyloo Group, Western Australia, age approximately 2.0 Ga; Schmidtsdrift Formation, Transvaal Group, South Africa, age approximately 2.2 Ga or older; Koolpin Formation, Pine Creek Geosyncline, central Australia, age between 2.0 to 2.3 Ga; Rocknest Formation, Epworth Group, Great Slave Lake area, northern Canada, age approximately 1.9 Ga; Assemblage 2, Hutuo Group, China, age approximately 1.9 Ga; Earaheedy Group, Nabberu Basin, Western Australia, age approximately 1.7 Ga.

Pilbaria perplexa Walter 1972

(Figs. 3, 4, 5)

1964, *Collenia australasica* Edgell, p. 244, Pl. 5.

1972, *Pilbaria perplexa* Walter, p. 167, Pl. 4, Fig. 4, Pl. 29, Figs 2-7.

1972, Unnamed stromatolite Walter, p. 177, Pl. 28, Fig. 4.

1978, *Pilbaria perplexa* Walter; Semikhatov, p. 141, Pl. XXIII, Figs. 1-3.

Type specimen

S206, Department of Geology and Mineralogy, University of Adelaide, South Australia. Duck Creek Dolomite (WYL 27, Fig 1).

Diagnosis

As in Walter 1972, p. 167.

Remarks

From examination of the sample (F 5015) referred to as "Unnamed Stromatolite" by Walter (1972), and comparison with material in outcrop, the poorly preserved columns are within the range of diversity shown by *Pilbaria perplexa* and consist of the small, uppermost parts of *P. perplexa* columns.

Distribution

Duck Creek Dolomite, Wyloo Group, Western Australia, age approximately 2.0 Ga. Rocknest Formation, Epworth Group, Great Slave area, northern Canada, age approximately 1.9 Ga.

Pilbaria cf. **perplexa** Walter 1972

(Fig. 8)

Remarks

Branching-columnar stromatolites that consist of large columns with niches. Columns are erect with irregular margins. Branching into two columns is infrequent. Laminae are gently to steeply convex, with numerous micro-cross laminations. Details of microstructure and fabric cannot be determined because of poor preservation.

These stromatolites are larger than *Pilbaria perplexa* and have less frequent branching and more irregular laminations with numerous micro-crosslaminations. The poor preservation precludes assignment to *P. perplexa*, but the presence of niches indicates that the stromatolites belong to *Pilbaria*.

Distribution and age

Intertidal facies, Duck Creek Dolomite, Wyloo Group, Western Australia, age approximately 2.0 Ga.

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STRATABOUND AXINITE IN THE WEELI WOLLI FORMATION AND ITS OCCURRENCE IN RELATED DOLERITES

by R. Davy and M. Pryce*

ABSTRACT

A member of the axinite group (a boron-bearing alumino silicate series) has been identified in strata-bound layers in iron-formation of the Weeli Wolli Formation and is also present in intercalated dolerite sills.

The axinite member in both iron-formation and dolerite is strictly the mineral ferroaxinite, though that in the dolerite contains a slightly greater proportion of magnanese. The habits are also different: axinite in the iron-formation is idioblastic and contains abundant inclusions whereas that is the dolerite is porphyroblastic-irregular, with no inclusions. In both cases it is incorporated into the body of the rock rather than occurring as discrete veins.

It is inferred that axinite formed during burial metamorphism at temperatures of about 250°C and a pressure of 200-500 MPa. The origin of the boron is less certain; it has probably been integral to the dolerite magma, being carried to the sediments by fumerolic activity, or, less likely, it may have been derived from pre-existing detrital tourmaline within the sediment.

INTRODUCTION

During logging of a diamond-drill core through the Weeli Wolli Formation and its associated dolerite skills, grey-white 'rhombic' grains were noted in specific bands. The character and position of these grains initially suggested pseudomorphs of former evaporite minerals (Fig. 1). However, microscope and X-ray diffraction studies showed that the mineral was in the axinite group, a boron-bearing Ca-Fe-Mg-Mn alumino silicate of general formula $HCa_2(Mn,Fe)BA1_2Si_4O_{15}OH$.

Though axinite had been recognized previously in the dolerite sills of this core (de Laeter and others, 1974) this is the first time that it has been seen as an apparently integral part of an iron-formation. A literature search has shown that, though axinite is well known in metamorphic contact deposits or in hydrothermal veins, it has been described rarely as a rock-forming mineral in a strata-bound environment. Opportunity has been taken, therefore, to document this occurrence and to compare the axinites in the dolerite with those in the iron-formation.

LOCATION OF THE CORE

The Weeli Wolli Formation is one of eight constituent formations of the (Precambrian) Hamersley Group in the northwest of Western Australia. The formation's regional characteristics have been described by Trendall and Blockley (1970). The Weeli

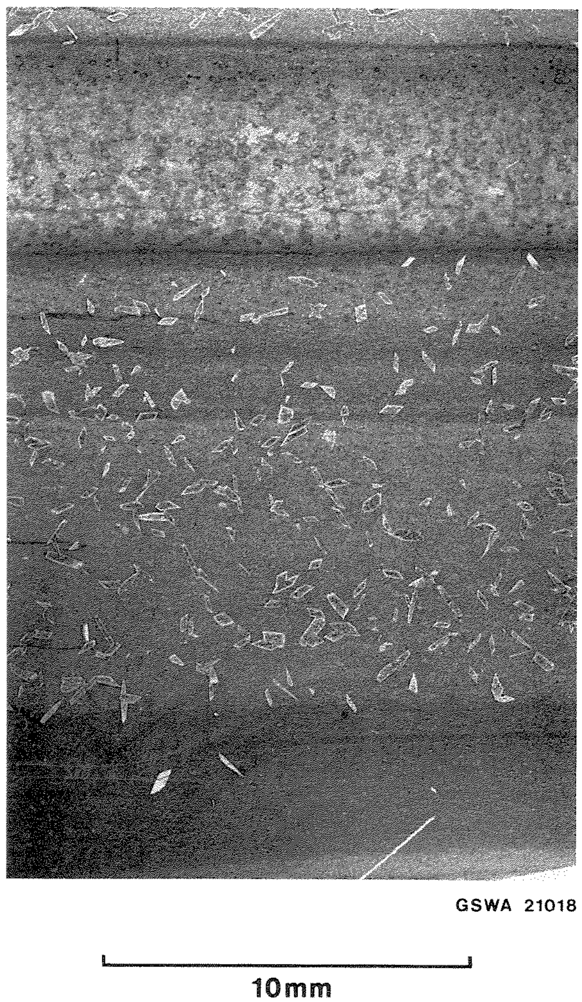

Figure 1. Photograph of core at 175.6 m showing randomly oriented axinite grains concentrated in specific bands.

*Government Chemical Laboratories.

Wolli Formation is considered to have an age of 2.3 Ga (minimum) based on a Rb-Sr model age on a sill within the formation (Trendall, 1983). The Weeli Wolli Formation is less well exposed than other Hamersley Group formations, and DDH WW1 was drilled for stratigraphic purposes at 22°17'30"S and 118°24'00"E, about 44 km south of Wittenoom. The hole, angled at 60° to the horizontal, was collared in surficial material and passed through weathered 'dolerite and clay'. Recognizable iron-formation was reached at 27.1 m, and fresh iron-formation at 38.7 m, (down-hole distances from the collar to the position in the core). The hole terminated at a core length of 272.3 m, remaining in Weeli Wolli Formation throughout. A simplified log of the core, showing the location of the samples, is given in Figure 2. Axinite is reported from the two lower sills (in the cored hole) and from the metamorphosed sedimentary rocks between them. These sills are subsequently referred to as 'upper' and 'lower' respectively.

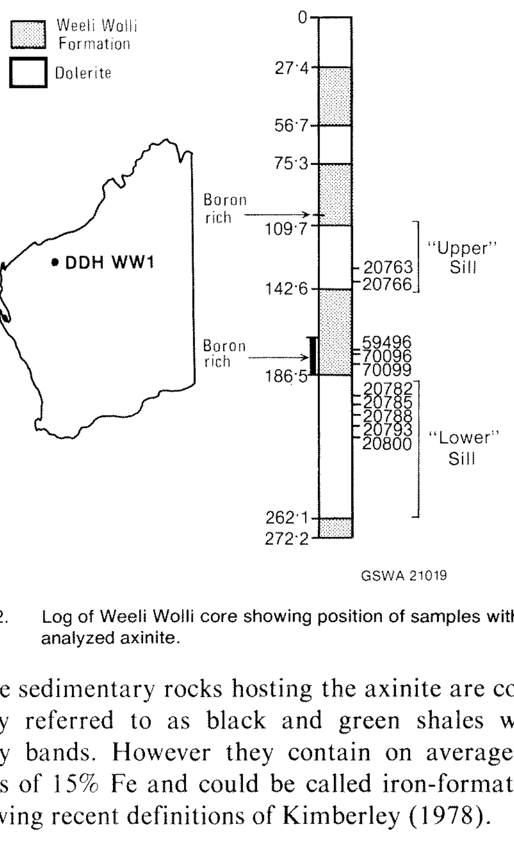

Figure 2. Log of Weeli Wolli core showing position of samples with analyzed axinite.

The sedimentary rocks hosting the axinite are commonly referred to as black and green shales with cherty bands. However they contain on average in excess of 15% Fe and could be called iron-formation following recent definitions of Kimberley (1978).

METHODS OF STUDY

Thin and polished sections of various parts of the core were prepared and studied. Some core samples which had been described by de Laeter and others (1974) were also included.

Identification of axinite was made optically and confirmed by X-ray diffraction analysis. The compositions of both axinite and epidote were determined by electron-probe microanalysis at the CSIRO Division of Mineralogy, Floreat, W. A. Boron and H_2O were not determined except in bulk samples of core. The small amount of each sample available and the nature of the axinite precluded separation of the latter by heavy liquids.

DISTRIBUTION IN THE CORE

Traces of axinite were found in the upper dolerite. It is present in stratigraphically contained bands at specific positions within the iron-formation, especially at 175.6 m and 178.0 m, and is present as a common accessory mineral within the lower dolerite. The mode of occurrence of axinite is virtually the same in both dolerites but it has a quite different habit in the iron-formation.

Axinite in the iron-formation

The greatest concentration of the axinite-bearing bands occurs between 175.6 and 175.8 m, approximately 11 m above the 'lower' sill. However axinite-bearing bands have been found down to 180 m. Other boron-rich zones contain no detectable axinite (see later discussion).

The axinite occurs as euhedral to subhedral, wedge-shaped porphyroblasts of random orientation set in ferrostilpnomelane and minor chorite, in stratigraphically controlled layers up to 2.5 cm thick (Fig. 3). The maximum size of the axinite is about 1.2 mm long but most grains range between 0.5 and 0.8 mm long. Adjacent to some of the axinite-rich layers, in which axinite comprises 10-15% of the layer, are zones with lesser axinite ($<5\%$). In those layers with small amounts of axinite, the axinite is consistently less well formed (subhedral to irregular) and is smaller (averaging 0.3 mm in length). Axinite is crowded with fine inclusions mainly of sphene or epidote, and also some stilpnomelane and quartz. A thin rim on the axinite crystals, some 5-10 µm thick, has enabled accurate probe analyses to be made.

The wedge-shaped axinite grains of the iron-formation contrast sharply with those in the dolerite, by virtue of their well-formed shape and the large proportions of contained inclusions.

The phyllosilicates in which the axinite has developed are very fine-grained ($<$5-10 µm) with random orientation. In a few places, local recrystallization has produced small rosettes (up to 25 µm in diameter) of radiating ferrostilpnomelane.

Axinite in the dolerite

The petrography of the dolerite has been described by de Laeter and others (1974). Both dolerite sills are now largely composed of secondary minerals. Thus, most primary pyroxene has been converted to actinolitic amphibole, and calcic plagioclase is altered to albitic feldspar. Exsolved calcium is now

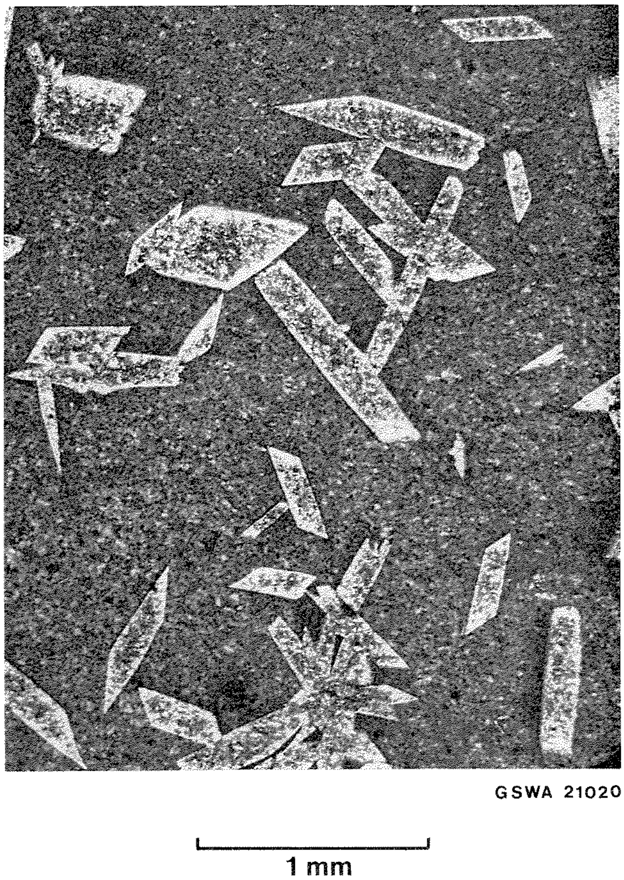

Figure 3. Photomicrograph of axinite porphyroblasts within iron-rich phyllosilicate of the Weeli Wolli Formation. The photomicrograph shows the euhedral shapes, clear rims, and inclusions of the axinite. Sample 59496.

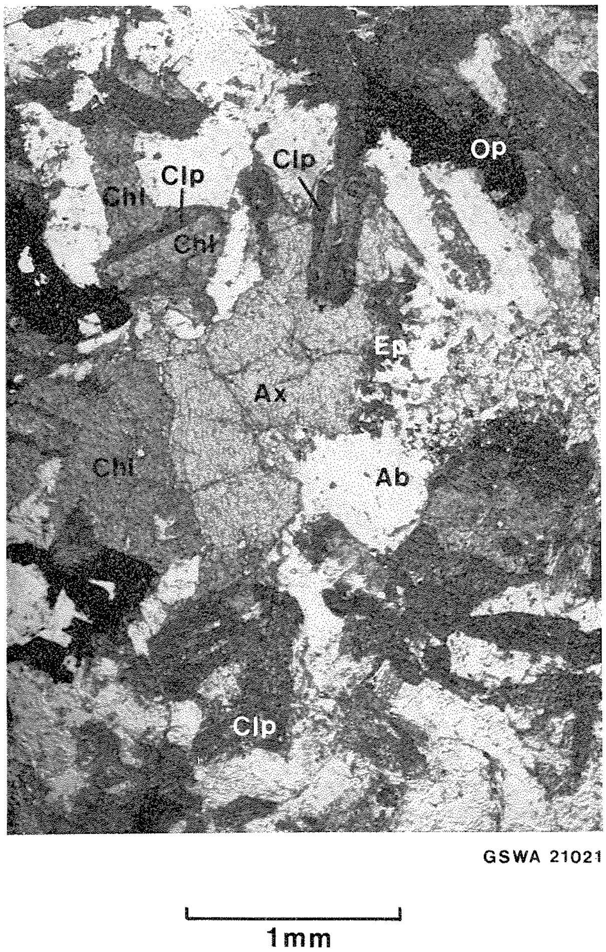

Figure 4. An axinite grain within the lower dolerite at 202.7 m. Adjacent minerals include albite (Ab-white), chlorite (Chl-dark grey), opaques (Op-leucoxene after ilmenite), pyroxene (Clp-very dark grey) and epidote (Ep-small dark grains). Plane-polarized light. Sample 20785.

represented by epidote, carbonate and axinite. Alteration of ferromagnesian minerals has progressed as far as chlorate and ferrostilpnomelane in places. Relict primary ophitic texture is, however, still recognisable. Interstitial pockets between former large grains of pyroxene and plagioclase are infilled with various assemblages of quartz, K-feldspar, apatite, zoisite, chlorite, carbonate and axinite. Within this interstitial material, axinite occurs as irregular grains 0.5-2 mm in diameter, commonly in association with granular quartz or epidote (Fig. 4); it does not occur where calcite is present. Though these interstitial minerals appear to have developed at a late stage, axinite does not occur in the last vein-phase which contains quartz, carbonate and epidote. Some of these veins cut preexisting axinite grains. Axinite in the dolerite is free of inclusions and is not zoned.

Axinite is much more abundant in the lower sill than in the upper sill.

CHREMICAL COMPOSITION OF THE AXINITE

The mineral is a low-manganese ferroaxinite, with a variable composition. The ferroaxinite from the dolerite is consistently more manganiferous than that of the iron-formation, and the upper-sill axinite contains more manganese than that of the lower sill (Table 1).

Axinite in iron-formation contains an extremely small content of manganese but is inherently more variable in composition (Table 1). This may be due, in part, to the micro-inclusions in the axinite.

In the lower sill, axinite composition changes with depth (Table 1), and probably reflects systematic changes of the composition of the dolerite as a whole (*cf.* de Laeter and others, 1974). A few axinite grains in the iron-formation contain traces of TiO_2, K_2O and Na_2O, whilst traces of Cl, $SO_3$and Na_2O have been detected in some grains from the dolerite.

DISCUSSION

Metamorphic Grade of the Core

The overall metamorphic grade of the rocks in the region suggested by Smith and others (1982) to be in the prehnite-pumpellyite-epidote facies of regional (burial) metamorphism. However, neither prehnite nor pumpellyite are known from the core. Comparison of analyses of epidote in dolerite (Table 2) with those cited by Smith and others (1982, Table 4, p. 94)

TABLE 1. COMPOSITION OF REPRESENTATIVE AXINITES FROM CORE WW1 ELECTRON-PROBE ANALYSIS

	Upper dolerite	*Weeli Wolli Formation*				*Lower dolerite*			
Sample	20766	59496	70096A	70096B	70099	20782	20788	20793	20800A
Depth (m)	137.2	175.6	175.7	175.72	180.0	198.1	206.2	213.4	224.0
Points counted	5	1	2	2	2	3	2	2	2
%									
SiO_2	42.36	41.91	42.44	42.47	44.04	42.55	42.55	41.67	42.69
Al_2O_3	17.28	15.64	17.55	17.73	18.02	17.24	17.26	17.30	17.66
FeO (a)	7.88	10.20	11.51	10.68	11.42	9.92	9.11	8.15	7.69
MnO	3.39	0.91	1.27	1.60	1.45	2.71	2.50	2.89	2.93
MgO	1.30	0.71	1.02	1.38	0.74	0.70	0.98	1.09	1.42
CaO	19.71	21.83	19.93	19.34	20.58	19.52	19.61	19.22	19.52
TiO_2		5.23							
				Cation proportions (cations = 18)					
Si	8.01	7.72	7.90	7.93	8.00	8.04	8.06	8.02	8.06
Al	3.84	3.38	3.86	3.89	3.85	3.83	3.84	3.92	3.92
Fe	1.25	1.57	1.79	1.67	1.73	1.56	1.44	1.31	1.21
Mn	0.54	0.14	0.20	0.25	0.24	0.43	0.40	0.47	0.46
Mg	0.36	0.19	0.28	0.39	0.22	0.19	0.27	0.31	0.40
Ca	3.99	4.30	3.97	3.87	3.95	3.94	3.98	3.96	3.94
Ti		0.70							

NOTES:

(1) The total number of axinite grains analyzed (and used in figure 5) are; upper dolerite—1, iron-formation—22, lower dolerite—12.

(2) Cation percentages have been calculated assuming that Ca + Fe + Mn + Mg + Al + Si = 18. It is clear that a small proportion of the iron is present as Fe_2O_3, though this cannot be determined from the electron probe analysis.

(3) Sequential analyses along traverses across grains did not reveal clearly recognisable zoning although slight compositional variations were apparent.

(a) Total Iron as FeO

TABLE 2. COMPOSITION OF REPRESENTATIVE EPIDOTE GRAINS FROM DOLERITE IN CORE WW1, AND COMPARISON WITH EPIDOTE DATA FROM SMITH AND OTHERS (1982)

	This Study					*Smith and Others, 1982, p. 94*			
	Upper dolerite		*Lower dolerite*			*Zone II*		*Zone III*	
Sample	20763	20766	20785	20786	20793	10077	10064	10935A	10920
Points counted	2	2	3	2	3				
%									
SiO_2	37.82	37.65	37.26	37.48	37.12	37.35	38.08	38.02	38.41
Al_2O_3	23.10	22.23	21.29	21.22	22.09	22.41	22.96	24.01	26.17
FeO (a)	13.10	13.84	14.23	14.40	13.52	(b) 13.38	12.76	10.41	8.25
CaO	23.61	23.52	23.15	23.00	23.03	21.45	22.58	23.65	24.04
V_2O_5	0.24		0.39	0.33					
MgO						1.22	0.95		
			Cation proportions (assuming Si + Al + Fe + Ca = 8)						
Si	3.00	3.00	3.00	3.00	3.01	3.00	3.01	3.03	3.01
Al	2.14	2.10	2.03	2.03	2.08	2.12	2.14	2.25	2.42
Fe	0.86	0.90	0.97	0.97	0.92	0.89	0.84	0.69	0.54
Ca	2.00	2.00	1.98	1.98	1.99	1.85	1.91	2.02	2.02
V	0.01		0.02	0.02					
Mg						0.14	0.11		

(a) Total Iron as FeO (b) Fe_2O_3 converted to FeO for comparison

suggests that the metamorphic grade lies between their prehnite-pumpellyite-epidote zone (ZII) and their prehnite-pumpellyite-epidote-actinolite zone (ZIII). Amphibole in the dolerite suggests that the rocks should properly be in zone ZIII.

Constraints on the temperature of formation of the axinite-bearing rocks have been suggested by a number of workers. Liou (1971) determined the upper stability limit of prehnite as 403°C at 300 MPa pressure. Similarly, Seki (1972) determined a lower limit for the stability of epidote as 220 ±50°C for pressure of 100-500 MPa. Ayres (1972) postulated metamorphic temperatures in the range 205°-325°C for the Brockman Iron Formation, based on the depth of burial and an estimation of thermal gradients. Becker and Clayton (1976) concluded, from oxygen isotope studies, that the Dales Gorge Member of the

Brockman Iron Formation (some 500 m below the Weeli Wolli Formation) had experienced a maximum temperature in the range 270-310°C.

There is no indication in the core of drill hole WW1, of any metamorphic changes other than those caused by burial to a depth of about 5 to 6 km (*cf.* Smith and others, 1982), except, in the immediate vicinity of the dolerite sills, where contact metamorphism has produced reddening of the core over approximately 0.5 m.

The iron-formation of WW1 contains, at other intervals, both minnesotaite and greenalite. Greenalite is considered by some workers (*e.g.* French 1973) as an early diagenetic mineral, and minnesotaite as a slightly higher temperature mineral. Grubb (1971) experimentally converted greenalite to minnesotaite at about 150°C, but found that greenalite and minnesotaite could still coexist at 250°C and 100 MPa pressure. Melnik and Siroshtan (1973), however, calculate that minnesotaite decomposes to grunerite at temperatures of 250-280°C at 200-500 MPa pressures.

The implication from these observations is that the axinite in the Weeli Wolli core was formed at temperatures close to 250°C, and at pressures in the range 100-200 MPa.

Axinites from other regionally metamorphosed terrains

Axinite schists have been reported from Sambagawa (Kojima, 1944) and Sangun (Nureki, 1967), both in Japan. These occurrences of axinite schists are associated with mafic schists, and the axinite is part of an assemblage which includes stilpnomelane, (chlorite) and epidote. Axinite has also been noted in metasediments at four New Zealand localities. Near Dansey Pass, axinite occurs within 'stilpnomelane-chlorite-epidote laminae in a fine-grained hematitic chert' (Pringle and Kawachi, 1980, p 1121). Other axinites were found by these workers within pumpellyite-bearing metacherts.

In all the above occurrences, the axinite crystals are porphyroblastic, subhedral to euhedral grains, and contain masses of minute inclusions of quartz, hematite and, possibly sphene. Nureki (1967) notes the random orientation and the clean rims of the Sangun axinite.

It is clear that axinite forms readily in low-grade, regional-metamorphic environments where temperature and pressure conditions are appropriate, and there is a source of boron. Axinite may therefore prove to be much more common in this type of environment than is currently recorded.

Axinite has been reported from other regionally metamorphosed, mainly mafic, igneous rocks, but in the majority of cases it is present in veins rather than as an integral component of the rock (*e.g.* Ozaki 1969, 1972).

Composition compared with other axinites

A ternary diagram showing the position of the Weeli Wolli axinites in relation to some other published data is included as Figure 5.

The axinite is highly ferroan, as is axinite in monzonitic rock reported by Hietanen and Erd (1978). However the axinite from the regionally metamorphosed rocks in Japan are manganese rich, as are some of those from New Zealand. Most documented ferroaxinites are found as vein minerals (Ozaki, 1970; Hietanen and Erd, 1978; Benjamin, 1968; Simonen and Wiik, 1952; and Deer, Howie and Zussman, 1962).

A relationship between the composition of axinite and that of the surrounding host rocks does not appear to have been established by previous workers. However, the high iron and low manganese content most likely reflect the original composition of the rock. The Weeli Wolli Formation at least contains high concentrations of iron (21.9% Fe or 28.1% FeO in the main axinite-rich zone) and relatively low MnO (an average of 0.11% in the entire formation, 0.23% in the vicinity of the main axinite-rich bands).

Origin of boron in the iron-formation

The Weeli Wolli occurrences of axinite suggest that the iron-formation must have contained high primary boron concentrations in particular layers. This is an unusual feature since the normal trace-element level in iron-formations is very low (Davy, 1983). It seems likely that boron was precipitated from solution on clay minerals of the illite or hydromica group, possibly through the agency of magnesium-hydroxy clusters (brucite-type). Boron could also have been coprecipitated with hydroxy-iron or hydroxy-aluminium minerals (Sims and Bingham, 1967, 1968a, 1968b; Rhoades and others, 1970). Boron retention by micas and ferro-aluminous compounds lessens with aging (Sims and Bingham, 1968a), and it seems probable that the boron, now found in axinite porphyroblasts in the iron-formation, was expelled from its primary position either during compaction and dewatering, or during the metamorphism which changed the original phyllosilicates to ferro-stilpnomelane.

The formation of axinite is considered to post-date the dewatering process, since no compaction features are evident. Growth has been a two-stage process. An initial, fast developing period, in which the axinite surrounded its present inclusions, has been followed by a later period of slower growth in which extraneous material was displaced.

Origin of axinite in the dolerite

Boron may have been inherent within the dolerite even though axinite is a relatively late-formed mineral. Boron did not take part in the final metasomatic

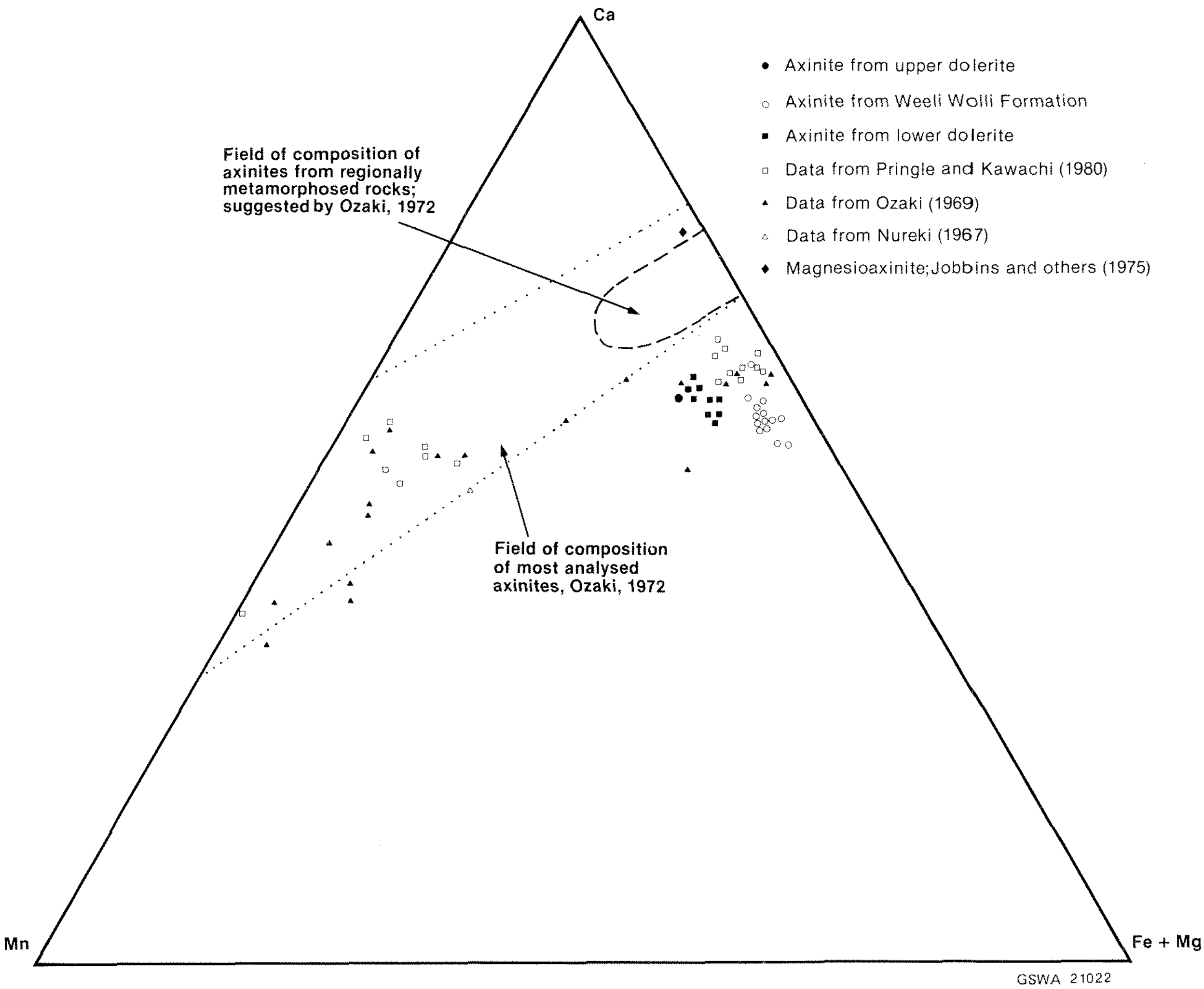

Figure 5. Comparison of axinites from core WW1 with other published values Diagram of atomic ratios, based on Ozaki, 1969.

activity which produced the quartz-carbonate (-epidote) veins. However, axinite was formed during a retrogressive metamorphic phase (with Ca coming from plagioclase, and with Mn, Fe and Mg from pyroxene) along with epidote and calcite. Its habit is porphyroblastic but not idioblastic; it shows no evidence of strain, and has no inclusions.

There is no evidence to suggest that boron has been introduced to the dolerite.

Origin and petrogenetic significance of boron

Boron is not universal in the Weeli Wolli Formation. The average boron content, calculted as a geometric mean over the whole length of the fresh iron-formation, is 10 ppm. Within this length there are two main boron-rich zones, namely between 167.4 m and 186.5 m and between 103.6 and 104.6 m. The average boron concentration for these zones is in the range 70-75 ppm, but in axinite-rich layers boron values are much greater, viz. 975 ppm over 20 cm at 175.7 $\pm$ .1 m, and 189 ppm over 10 cm at 171.6 m. The 19.1 m thick boron-rich zone immediately overlies the lower dolerite, but the upper zone, though it overlies the upper sill is separated from it by 5 m of iron-formation with background values of $<$10 ppm boron.

The close spatial association of the boron-rich part of the iron-formation with the dolerite sills suggests a common origin for the boron. The boron may ultimately have been derived from sea water, the dolerite magma or some hydrothermal source unrelated to the magma but part of a larger metamorphic event.

The stratabound mode of occurrence of the axinite in the iron-formation suggests that, if the boron were derived from sea water, a pulsed inflow of boron-rich water is necessary. Alternations of pulses of boron-enriched water with ordinary, essentially boron-free sea water would account for the layered distribution of the axinite, assuming that other constraints were kept relatively constant. The close proximity of the boron-rich sediments to the dolerite suggests that the dolerite was the ultimate source of the boron. It is postulated that after intrusion, late hydrothermal or fumarolic activity, related to the magma, added fluids containing boron to the sediments overlying the sills. The observed distribution of that element is best accounted for by periodic outbursts of solfataric or fumarolic activity alternating with quiescent periods. If this hypothesis is correct, it implies that the magma emplacement was barely intrusive. The dolerite may well have been emplaced into wet, unconsolidated sediments. It also implies that the dolerite and iron-formation are essentially contemporaneous. Some boron was retained in the dolerite, the remainder trapped in the phyllosilicates in the iron-formation.

Much of the fumarolic activity would be caused by the contact of magma at 1200°C with wet sediments, producing local boiling (the critical point for ordinary sea water is 408°C at 30.4 MPa pressure). Magmatic fluids supplied boron, but steam from pore waters was probably the main carrier to the sediments. Some of the steam generated may have circulated back through the dolerite and caused further leaching of boron. Movement of steam through the dolerite could have occurred along contraction cracks, and may account for Trendall's recent observation of a patchy distribution of fresh and altered dolerite and the presence of two chlorite species (Trendall and de Laeter, in prep.).

A more modern illustration of the effects of intrusion into wet sediments has been provided by Einsele and others (1980) in the Gulf of California.

This hypothesis of the origin of the boron is at variance with that of de Laeter and others (1974) who suggested that the dolerite magma assimilated some iron-formation and shale during intrusion and then acquired much of its present mineral composition by duteric activity. Assimilation of sediments could account for the high K_2O content (2.7%) of the dolerite, and de Laeter and others (1974) suggested that the boron was introduced into the dolerite from the incorporated sediment, possibly as detrital tourmaline (tourmaline had been reported earlier by Trendall and Blockley, 1970, in shales of the Brockman Iron Formation). This mechanism is considered unlikely because there is no sign of other detrital material, and, more particularly, because tourmaline once formed is resistant to later modification.

Whatever the origin of the boron the axinite seems to be of late origin, and related to the later metamorphic events.

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