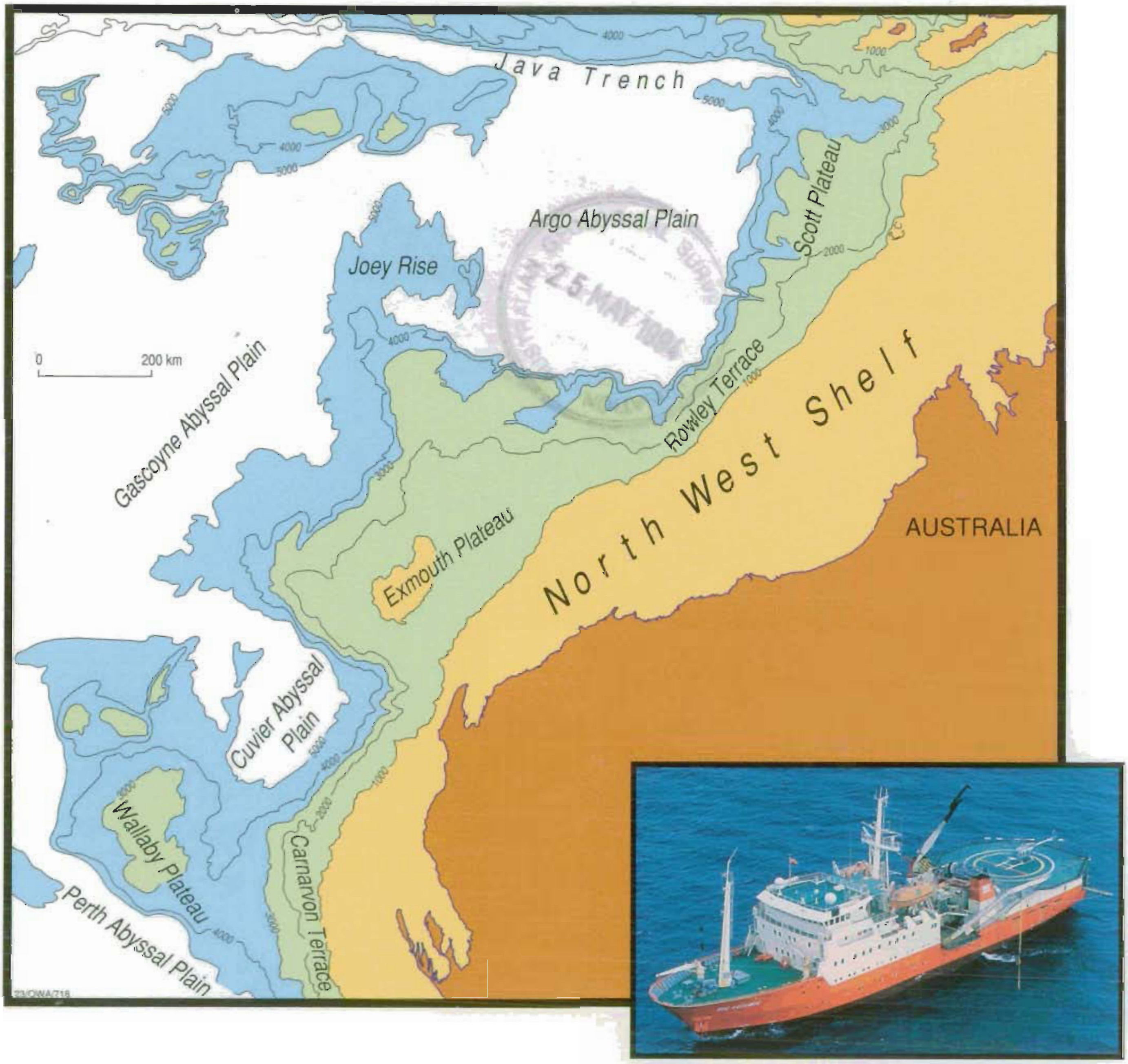
AGSO JOURNAL
OF AUSTRALIAN GEOLOGY & GEOPHYSICS



THEMATIC ISSUE: GEOLOGY OF THE OUTER NORTH WEST SHELF, AUSTRALIA

VOLUME 15, NUMBER 1
1994

AGSO Journal of Australian Geology & Geophysics

Policy

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Abstract and key words

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AGSO JOURNAL

OF AUSTRALIAN GEOLOGY & GEOPHYSICS

VOLUME 15, NUMBER 1, 1994

Thematic issue: Geology of the outer North West Shelf, Australia*

CONTENTS

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Front-cover illustration: **Morphology of the sea bed off northwest Australia, showing the North West Shelf. Map prepared by AGSO's Cartographic Services Unit from general Bathymetric Chart of the Oceans at a scale of 1:10 000 000, published by the International Hydrographic Organisation & the Intergovernmental Oceanographic Commission. Contours in metres.**

Lower right corner: **The AGSO research vessel *Rig Seismic* which gathered the data presented in this issue of the *AGSO Journal*. This vessel was built in Norway in 1982; has dynamic positioning capability; and has a length of 72.5 m and a gross registered tonnage of 1545 tonnes. It has been used by BMR/AGSO since October 1984.**

Thematic issue: Geology of the outer North West Shelf, Australia

Preface

The North West Shelf is a major petroleum province, and an understanding of its outer margin is important to petroleum exploration companies. This thematic issue of the AGSO Journal documents a variety of studies based largely on Mesozoic sedimentary rocks and seismic profiles acquired on two cruises of AGSO's research vessel *Rig Seismic* along the shelf's outer margin. The cruises were designed to dredge targets in canyons cut into the continental slope, in order to provide a better idea of the ages, lithologies and palaeoenvironments of the strata cropping out. The sampling sites were tied to existing reflection seismic profiles and new profiles acquired during this program, and the results have been correlated through the seismic profiles with stratigraphic information from petroleum exploration wells.

Samples of rocks recovered from the cruise were provided to palaeontologists, sedimentologists and petrologists, both inside and outside AGSO. The result is eight papers: six on palaeontology, one on sedimentary petrology, and one on igneous petrology. In addition, there are three articles on the results of seismic reflection profiling using AGSO, ODP and Soviet data, one general paper on the enigmatic Wallaby Plateau, and a synthesis of all the results. These papers represent a considerable extension of our knowledge of the region, as well as of the history of this part of Tethys and Gondwana.

Neville Exon

AGSO Journal of Australian Geology & Geophysics, 15/1, 3–10

An introduction to the geology of the outer margin of Australia's North West Shelf

Neville Exon[1]

The North West Shelf is a geographic province that extends about 2400 km along the northwest margin of Australia out at least as far as the 2000 m isobath, with an average width of 300 km (Fig. 1). According to Purcell & Purcell (1988), its limits are near Darwin in the northeast and near Exmouth Gulf in the southwest, and its area is some 720 000 km^2. It includes the Scott and Exmouth Plateaus and overlies four major Phanerozoic sedimentary basins (from north to south: Bonaparte, Browse, offshore Canning, and Carnarvon Basins) made up of superimposed

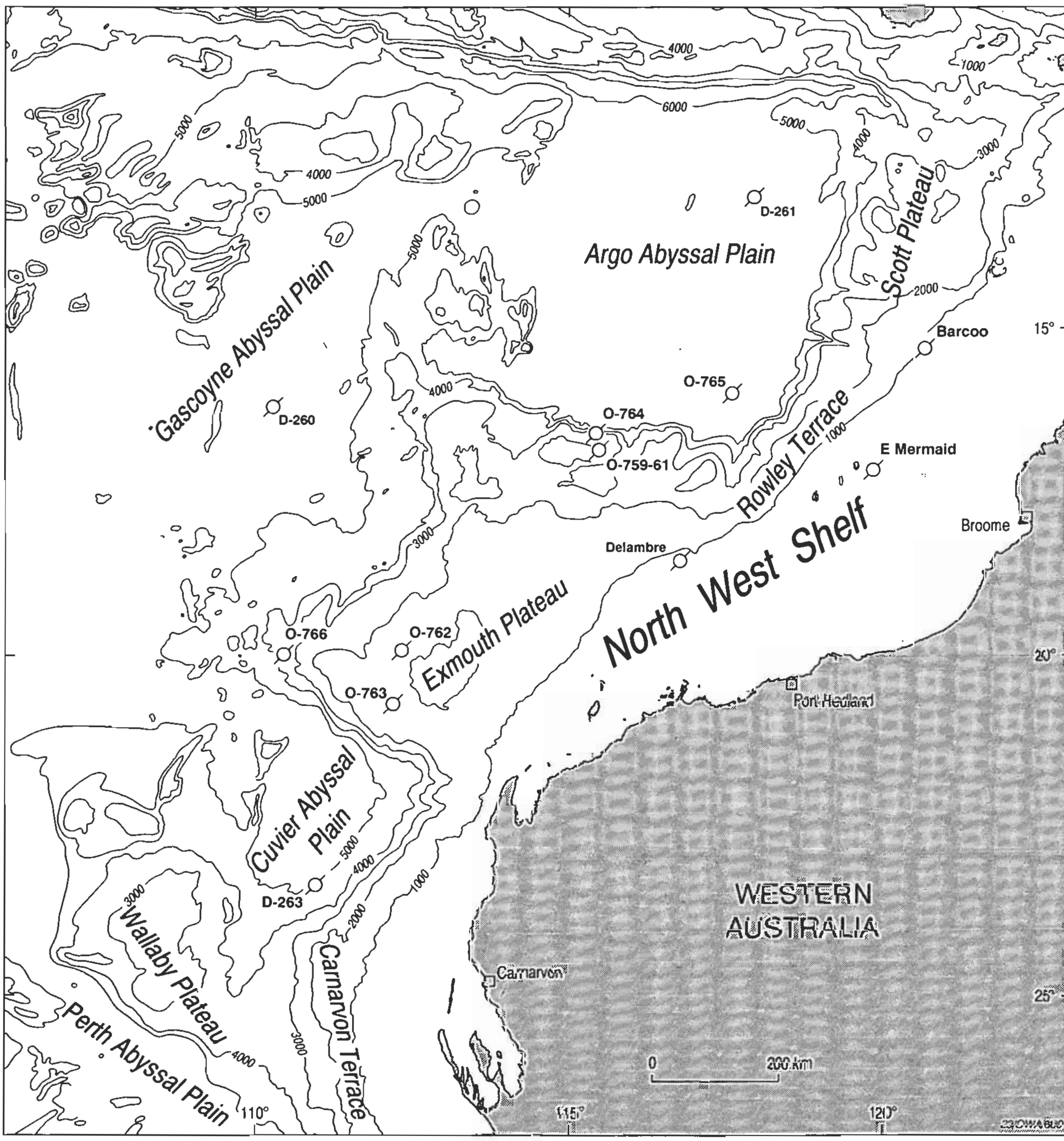

Figure 1. Bathymetric map of the North West Shelf Region. Isobaths at 1000 m intervals, and showing the main structural features discussed in the text. Shows Deep Sea Drilling Project (DSDP) Sites D260, 261 & 263 (Veevers et al., 1974); ODP Sites O759–766; petroleum exploration wells Barcoo No. 1, East Mermaid No. 1 & Delambre No. 1.

[1] Marine Geoscience and Petroleum Geology Program, Australian Geological Survey Organisation, GPO Box 378, Canberra 2601, Australia.

sequences with shelf-wide unconformities between them, and more than 15 km thick in places. Permian and thick Mesozoic sequences were laid down in the Westralian Superbasin on the southern edge of the Tethyan Ocean and the northern shores of Gondwana (Yeates et al., 1987). The Mesozoic sequence is a major producer of oil and gas. When Gondwana broke up in the Late Jurassic and Early Cretaceous, and abyssal plains formed to the west, the North West Shelf gradually took up its present configuration as the result of subsidence and deposition. Thick wedges of Cainozoic carbonates have formed from warm-water calcareous organisms.

This thematic issue of the *AGSO Journal* documents a variety of studies based on Mesozoic sedimentary rocks and seismic profiles acquired on two cruises of the Australian Bureau of Mineral Resources' (BMR's) research vessel RV *Rig Seismic* along the outer margin of the North West Shelf in 1990, and synthesises the results. [Since then BMR has become the Australian Geological Survey Organisation (AGSO)]. The two surveys were designed to dredge most of the previously undredged targets in canyons cut into the continental slope, in order to provide a better idea of the ages, lithologies and palaeoenvironments of the strata cropping out in those canyons. The sampling sites were tied to existing reflection seismic profiles and new profiles acquired during this program, and the results have been correlated through the seismic profiles with stratigraphic information from petroleum exploration wells. In this way, a better understanding of the geology of the outer shelf and continental slope has been developed. This study builds on the results of some earlier studies, discussed briefly below.

Three major regional studies of the outer North West Shelf, carried out by BMR in the 1970s, were based largely on reflection seismic profiles, with lesser input from magnetic and gravity profiles, and from petroleum exploration wells farther inshore. The first study, of the Exmouth Plateau, started in 1974 and was finalised in the major synthesis of Exon & Willcox (1980). A study of the Carnarvon Terrace and Wallaby Plateau was published by Symonds & Cameron (1977). The third study was of the Scott Plateau and Rowley Terrace (Stagg & Exon, 1981).

The German Bundesanstalt für Geowissenschaften und Rohstoffe (BGR), in conjunction with BMR, carried out two major sampling cruises in the region in the late 1970s. The first, Cruise VA-16, used RV *Valdivia* for seismic profiling and sampling in the region of the Scott Plateau and the adjacent Argo Abyssal Plain (Hinz et al. 1978). The data set included 1700 km of multichannel profiles, 13 dredges, and two piston cores. The results showed that basic volcanism and shallow-marine sedimentation preceded Callovian breakup, and that post-breakup sedimentation included shallow-marine Upper Jurassic detrital sediments, and Cretaceous and Cainozoic carbonates.

A second BGR cruise (SO-8) used RV *Sonne* for minor seismic profiling and extensive sampling of the Exmouth and Wallaby Plateaus (von Stackelberg et al., 1980). The lithofacies scheme established on this cruise has been used for all later sampling cruises in the region. Pre-Quaternary rocks were obtained (mostly by dredging) from 29 stations on the Exmouth Plateau and 12 stations on the Wallaby Plateau and Cuvier Abyssal Plain, in water depths of 1800–5340 m. On the Exmouth Plateau, the rocks recovered included rift volcanics from the Triassic–Jurassic boundary, and sedimentary sequences of Upper Triassic and Lower Jurassic shallow-marine limestones and siliciclastic sediments, Middle Jurassic coal measures, post-breakup Lower Cretaceous shallow-marine mudstones, and Upper Cretaceous and Cainozoic carbonates laid down in increasing water depths with time. On the Wallaby Plateau, a layered sequence below the Neocomian breakup unconformity proved to consist of basalts and very varied volcaniclastic rocks, as did the Sonne Ridge on the Cuvier Abyssal Plain.

The results of the two BGR cruises and the BMR seismic studies were combined in a review of the sedimentary and volcanic evolution of the northwest Australian continental margin by von Rad & Exon (1983). A complementary study, dealing with the mode of formation of the margins of the Exmouth Plateau, was published by Exon et al. (1982).

In the late 1970s and early 1980s, during a major petroleum exploration program on the Exmouth Plateau, a great deal of multichannel seismic data was acquired and ten deepwater wells were drilled. All the data and much interpretive work are now in the public domain. Barber (1988) published a major overview of results for the central plateau.

Three commercial wells on the North West Shelf are of key significance for this volume: Delambre No. 1 in the northern Carnarvon Basin (Woodside Petroleum Development, 1981); East Mermaid No. 1 (Shell Development Australia, 1973); and Barcoo No. 1 (Woodside Petroleum Development, 1980) in the Rowley Sub-basin of the Canning Basin (locations in Fig. 1). These three wells all drilled thick Cainozoic, Cretaceous and Jurassic sequences, and Delambre No. 1 and Barcoo No. 1 also drilled the Upper Triassic. Shell East Mermaid No. 1 was drilled in a water depth of 386 m, Woodside Delambre No. 1 well in a water depth of 894 m, and Woodside Barcoo No. 1 in a water depth of 731 m. None had any significant shows of oil or gas. Figures 2 to 4 are drawn from the company well-completion reports that are now on open file, with additional palaeontological and palaeoenvironmental input from AGSO palaeontologists G. Chaproniere and S. Shafik.

In 1985, BGR and BMR scientists prepared a proposal for Ocean Drilling Program (ODP) drilling on the Exmouth Plateau. To strengthen this proposal, BMR Cruise 56 was carried out on the northern plateau, using RV *Rig Seismic*. It obtained 2100 km of regional seismic profiles, 550 km of site survey profiles, and 16 dredge hauls of rocks in water depths of 2000–5600 m (Exon & Williamson, 1988). The rocks recovered were quite similar to those from the *Sonne* cruise, and included Triassic volcanics, Triassic and Lower Jurassic shelf carbonates, Middle Jurassic coal measures, Cretaceous detrital sediments, and Cainozoic chalks and oozes. The sedimentology of these sediments was described in detail by von Rad et al. (1990).

In 1988, ODP fully cored wells at eight sites on and near the Exmouth Plateau (Fig. 1), providing reference sections for the Cainozoic, Cretaceous and Upper Triassic sequences, and discovering the first Triassic reefs in Australia (Williamson et al., 1989). Leg 122 cored 3370 m of sedimentary section at six sites on the Exmouth Plateau, and the results are described in Haq et al. (1990), and further described and synthesised by von Rad et al.

(1992a, b). Leg 123 cored 1389 m of Cainozoic and Cretaceous sedimentary section and 333 m of oceanic crust at two sites (Fig. 1), one each on the Argo and Gascoyne Abyssal Plains (Gradstein et al. 1990, 1992).

The four ODP core sites on Wombat Plateau give a great deal of detailed information on the Upper Triassic sequences (Haq et al., 1990; von Rad et al., 1992b). A composite diagram of the thickest sequences from these sites (Fig. 5), at that time located much farther offshore than the sites of the three exploration wells, shows how the fluvio-deltaic and deltaic detrital sedimentation of the Carnian and Norian gave way to reefal and lagoonal carbonate sedimentation in the Rhaetian, as the detrital supply waned.

The first survey discussed in this issue, BMR Cruise 95, was designed to follow up the discovery of Triassic reefs

STRATIGRAPHIC TABLE : DELAMBRE 1

Latitude: 18°31'05"S Longitude: 116°41'48"E RT 10m

			Age	Lithology	Depth (m)	Thickness (m)	Reflector	Two-way time (s)
					10 (Sea level)	10		
					894 (Sea bed)	884		
			NOT SAMPLED		1237	343		
CAINOZOIC	NEOGENE		LATE PLIOCENE	Upper slope foramininferal argillaceous calcilutite	1296	59		
CAINOZOIC	NEOGENE		EARLY PLIOCENE	(as above)	1470	174		
CAINOZOIC	NEOGENE		LATE MIOCENE	Upper slope claystone and calcisiltite	1721	251		
CAINOZOIC	NEOGENE		MIDDLE MIOCENE	Upper slope calcareous claystone, marl and calcilutite	~1820~	109	A_0	2.17
CAINOZOIC	NEOGENE		EARLY MIOCENE	Upper slope argillaceous calcilutite	1857	37		
CAINOZOIC	PALAEOGENE		LATE OLIGOCENE	(as above)	~1869~	12	A_1	2.18
CAINOZOIC	PALAEOGENE		MIDDLE EOCENE	Outer shelf to upper slope argillaceous calcilutite	1916	47		
CAINOZOIC	PALAEOGENE		EARLY EOCENE	(as above)	~1991~	75	A_2	
CAINOZOIC	PALAEOGENE		LATE PALEOCENE	Outer shelf argillaceous calcilutite and calcarous claystone	2060	69		
CAINOZOIC	PALAEOGENE		PALEOCENE UNDIFFERENTIATED	(as above)	2066	6		
CAINOZOIC	PALAEOGENE		EARLY PALEOCENE	Shelf claystone	2069	3	B	2.34
MESOZOIC	CRETACEOUS	LATE	LATE MAASTRICHTIAN	Outer shelf marl	2101	32		
MESOZOIC	CRETACEOUS	LATE	EARLY MAASTRICHTIAN	Outer shelf marl	?2120	19		
MESOZOIC	CRETACEOUS	LATE	LATE CAMPANIAN	(as above)	2156	36		
MESOZOIC	CRETACEOUS	LATE	INDETERMINABLE	Outer shelf marl and calcilutite	2169	13		
MESOZOIC	CRETACEOUS	LATE	LATE SANTONIAN	(as above)	?2193	24		
MESOZOIC	CRETACEOUS	LATE	EARLY SANTONIAN	(as above)	2200	7		
MESOZOIC	CRETACEOUS	LATE	CONIACIAN	(as above)	2222	22		
MESOZOIC	CRETACEOUS	LATE	? TURONIAN	Inner shelf calcilutite	2240	18	C	2.46
MESOZOIC	CRETACEOUS	LATE	LATE CENOMANIAN	Outer shelf to upper slope marl and calcilutite	?2245	5		
MESOZOIC	CRETACEOUS	EARLY	EARLIEST CENOMANIAN TO LATE ALBIAN	Outer shelf to upper slope marl	2253	8		
MESOZOIC	CRETACEOUS	EARLY	MIDDLE TO EARLY ALBIAN	Outer shelf marl	2286	33		
MESOZOIC	CRETACEOUS	EARLY	INDETERMINABLE	Shelf claystone with siliceous plankton	2288	2		
MESOZOIC	CRETACEOUS	EARLY	BARREMIAN	(as above)	~2318~	30	D	2.51
MESOZOIC	CRETACEOUS	EARLY	BERRIASIAN	Inner shelf sandstone and claystone	~2346~	28	E	2.53
MESOZOIC	JURASSIC	MID	BATHONIAN TO BAJOCIAN	Inner shelf and paralic sandstone, siltstone, claystone and minor coal	3353	1007		
MESOZOIC	JURASSIC	EARLY	AALENIAN TO TOARCIAN	Inner shelf claystone; minor sandstone and dolomite	3964	611		
MESOZOIC	JURASSIC	EARLY	PLIENSBACHIAN TO HETTANGIAN	Inner shelf claystone, limestone and sandstone	4287	323	F	3.71
MESOZOIC	TRIASSIC	LATE	RHAETIAN	Inner shelf claystone and sandstone; minor dolomite	4604	317		
MESOZOIC	TRIASSIC	LATE	NORIAN	(as above)	4665	61		
MESOZOIC	TRIASSIC	LATE	NORIAN TO CARNIAN	Inner shelf and paralic claystone, sandstone, minor coal	5264	599		
MESOZOIC	TRIASSIC	LATE	CARNIAN	(as above)	TD 5495	231		

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Figure 2. Stratigraphic table of Woodside Delambre No. 1 petroleum exploration well. Location in Figure 1. Modified from company well completion report (Woodside, 1981).

on ODP Leg 122 (Exon & Ramsay, 1990). The survey, over areas of the outer Rowley Terrace (Canning Basin) and northern Exmouth Plateau (Carnarvon Basin), involved both reflection seismic profiling and dredging. The conventional multichannel seismic program, 1750 km of profiling, consisted largely of long lines connecting three exploration wells to the abyssal plain and to ODP Sites 761 and 764 on the northernmost Exmouth Plateau. Another 255 km of high-resolution seismic profiles, in a tightly spaced grid including the two ODP sites, investigated the seismic character and distribution of the Upper Triassic reef complexes drilled.

The remainder of Cruise 95 was devoted to the successful dredging of 13 sites on the outer slope of the Rowley Terrace, in water depths of 3100–5325 m. Mesozoic volcanic rocks were predominant in the north and sedimentary rocks in the south. The sedimentary rocks included Upper Triassic and Lower Jurassic limestones, Upper Triassic to Middle Jurassic detrital sediments, Cretaceous mudstones and limestones, and Cainozoic limestones and chalks, much like those recovered on the *Sonne* cruise farther to the southeast. The dredge results from each of the stations are summarised in Table 1.

Table 1. Dredge hauls from BMR Cruise 95.

Location & water depth (m)	*Dredge number and description of main Mesozoic sediment groups*
15° 01′S 118° 58′E 4650–3800	1: ?L Triasic–M Jurasssic volcaniclastics, breccias, basic & intermediate volcanics
14° 46′S 119° 05′E 4000–3100	2: L Jurassic volcanic conglomerate, basaltic & andesitic volcanics
15° 09′S 118° 58′E 4900–4400	3: Possibly slumped ?Cretaceous mudstone
15° 22′S 118° 58′E 4200–3750	4: E–M Jurassic carb mudstone; M–L Jurassic quartz sandstone; ?M–L Jurassic ferruginous sediments; E Cretaceous radiolarian chalks & marls; L Cretaceous chalks & marls
15° 23′S 118° 58′E 3850–3330	5: ?L Triassic-Jurassic tuffaceous sandstone & volcanics; Jurassic quartz sandstone; M Jurassic carb mudstone; ?M–L Jurassic ferruginous sediments
16° 01′S 118° 46′S 3400–3100	6: L Triassic–E Jurassic tuffaceous mudstone; E–M Jurassic carb mudstone & volcaniclastics; ?M–L Jurassic ferruginous sediments
16° 18′S 118° 23′E 4530–3900	7: L.Triassic shelf carbonates including coral boundstone; ?L Triassic–Jurassic hyaloclastite & sub-arkose; E Jurassic mudstone with shelf carbonates; M Jurassic coal measures; ?M–L Jurassic ferruginous sediments
16° 19′S 118° 23′E 4400–4230	8: ?LTriassic-Jurassic volcanic sandstone; E Jurassic shelf carbonates; ?M–L Jurassic ferruginous sediments; E Cretaceous radiolarian mudstone
16° 19′S 118° 25′E 4200–3350	9: L Triassic shelf carbonates; E Jurassic mudstone; ?M Jurassic carb mudstone; ?M–L Jurassic ferruginous sediments; Jurassic quartz sandstone
16° 24′S 118° 10′E 5325–4090	10: L Triassic shelf carbonates including algal boundstone; ?L Triassic–Jurassic volcaniclastics, basic & intermediate volcanics; ?M–L Jurassic ferruginous sediments; Jurassic quartz sandstone
16° 26′S 118° 11′E 4000–3300	11: ?L Trias carb mudstone–v.f. sandstone; E–M Jurassic carb mudstone; ?M–L Jurassic ferruginous sediments
16° 29′S 118° 01′E 4470–3625	12: L Triassic shelf carbonates; ?L Triassic–Jurassic volcaniclastics & volcanics; ?Jurassic coal measures
16° 29′S 118° 10′E 3480–3100	13: ?M–L Jurassic ferruginous sediments; ?Jurassic coal measures

"carb" = carbonaceous. Descriptions modified from those of U. von Rad and U. Röhl of the Bundesanstalt für Geowissenschaften und Rohstoffe, Hannover, Germany, in Colwell et al. (1994 —this issue).

BMR Cruise 96 was a sampling survey extending from the Wallaby Plateau in the south to the Scott Plateau in the north, in water depths of 1995–5612 m (Colwell et al., 1990). Sampling was predominantly deep-water dredging of old strata, with a few cores taken in Quaternary sediments near the Rowley Shoals. Thirty-eight dredge stations were occupied, two on the Wallaby Plateau, eleven on the Carnarvon Terrace, seven on the Exmouth Plateau, eight up the eastern side of the Swan Canyon or nearby on the southernmost Rowley Terrace, and nine on the Scott Plateau. A great variety of Mesozoic volcanic and sedimentary rocks was recovered, as well as Cainozoic limestone and chalk.

Because it was clear that the rocks recovered from BMR Cruises 95 and 96 could provide a great deal of information on the development of the outer North West Shelf, samples were provided to palaeontologists, sedimentologists, and petrologists, both inside and outside AGSO. The result is eight papers in this issue: six on palaeontology, one on sedimentary petrology, and one on igneous petrology. In addition, there are three papers on the results of seismic reflection profiling (using AGSO, ODP and Soviet data), one general paper on the enigmatic Wallaby Plateau, and a synthesis of all the results. These thirteen papers represent a considerable extension of our knowledge of the region, and of the history of this part of Tethys and Gondwana. (Note that the 1994 papers mentioned below are all contained in this issue.)

Three papers, together with this introduction, provide the basic geological framework on which the other papers build. The sediment petrology paper by Colwell (AGSO), von Rad & Röhl (BGR), and Kristan-Tollmann (Vienna) (1994) provides detailed information on all the sedimentary and volcaniclastic rocks dredged on BMR Cruises 95 and 96. The paper on igneous petrology by Crawford (Tasmania) & von Rad (BGR) (1994) does the same for the dredged volcanic rocks. The seismic interpretation paper by Ramsay & Exon (AGSO) (1994) integrates the new and old seismic profiles on the northern Exmouth Plateau, to give a better understanding of its structure and seismic stratigraphy.

Six papers deal with the palaeontology of the rocks and with inferred palaeoenvironments. A paper on the calcareous nannofossils of the Mesozoic seqences of the Rowley Terrace by Shafik (AGSO) (1994) provides ages for marine sequences and evidence of changes in surface water temperatures. A paper by Burger (AGSO) (1994) on Mesozoic spores, pollen and microplankton from dredges from the whole area provides ages and useful environmental information on both marine and non-marine sequences. Late Triassic conodonts and microplankton from marine dredge material and from commercial and ODP wells are discussed by Nicoll & Foster (AGSO) (1994), and the study has led to a better palynological zonation for the Australian Triassic. Late Triassic and

STRATIGRAPHIC TABLE : EAST MERMAID 1

Latitude: 17°10'0.1"S Longitude: 119°49'21.2" E RT 11m

Era	Period	Epoch	Age	Lithology	Depth (m)	Thickness (m)	Reflector	Two-way time (s)
				Sea level	11			
				Sea bed	386	375		
			NOT SAMPLED		509	123		
CAINOZOIC	NEOGENE		PLIOCENE	Outer shelf marl	563	54	AP	0.65
CAINOZOIC	NEOGENE		MIDDLE MIOCENE	Inner shelf calcarenite, minor marl	1011	448	A_0	0.99
CAINOZOIC	NEOGENE		MIDDLE MIOCENE	Inner shelf calcarenite, dolomitised toward base	1203	192		
CAINOZOIC	NEOGENE		LOWER MIOCENE	Inner shelf calcarenite, dolomitised	1262	59		
CAINOZOIC	PALAEOGENE		LOWER MIOCENE TO UPPER OLIGOCENE	Inner or middle shelf calcarenite, minor calcareous siltstone	1331	69	A_1	1.20
CAINOZOIC	PALAEOGENE		UPPER TO MIDDLE EOCENE	Middle to outer shelf calcarenite, calcareous siltstone and minor chert	1524	193		
CAINOZOIC	PALAEOGENE		LOWER EOCENE	Outer shelf calcilutite	1597	73	A_2	1.34
CAINOZOIC	PALAEOGENE		UPPER PALEOCENE	Outer shelf calcareous claystone	1627	30	B	1.37
MESOZOIC	CRETACEOUS	LATE	EARLY MAASTRICHTIAN	Inner shelf calcareous claystone	1704	77		
MESOZOIC	CRETACEOUS	LATE	LOWER CAMPANIAN TO CONIACIAN	Outer shelf to upper slope calcilutite, minor marl	1767	63	C	1.46
MESOZOIC	CRETACEOUS		LOWER CENOMANIAN TO UPPER ALBIAN	Outer shelf marl and calcareous claystone	1792	25		
MESOZOIC	CRETACEOUS	EARLY	ALBIAN	Middle shelf marl to calcareous claystone	1938	146		
MESOZOIC	CRETACEOUS	EARLY	APTIAN	Middle to inner shelf claystone and siltstone, minor calcareous claystone and limestone	2301	363		
MESOZOIC	CRETACEOUS	EARLY	UPPER NEOCOMIAN	Inner shelf silty claystone and siltstone	2353	52	D	1.90
MESOZOIC	CRETACEOUS	EARLY	BERRIASIAN	Marine deltaic siltstone	2883	530	Y	2.20
MESOZOIC	JURASSIC	LATE	KIMMERIDGIAN TO CALLOVIAN	Marine deltaic sandstone and siltstone	2902	19	E	2.21
MESOZOIC	JURASSIC	MID	BATHONIAN TO BAJOCIAN	Fluviodeltaic sandstone, siltstone, shale and coal	3624	722		
MESOZOIC	JURASSIC	EARLY	TOARCIAN TO PLIENSBACHIAN	Deltaic shale, siltstone and sandstone, minor coal; increasingly marine downward	3916	292		
MESOZOIC	JURASSIC	EARLY	HETTANGIAN	Inner to middle shelf shale, siltstone and sandstone	TD 4068	152		2.78

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Figure 3. Stratigraphic table of Shell East Mermaid No. 1 petroleum exploration well. Location in Figure 1. Modified from company well completion report (Shell, 1973).

Early Jurassic bivalves from shelf carbonates from two sites on the Rowley Terrace outer margin are identified by Grant-Mackie (Auckland) (1994). Late Triassic corals and sponges from shelf carbonates from two Rowley Terrace sites (one in common with a bivalve site) are described by Stanley (Montana) (1994), and provide the first concrete evidence that Triassic reefs could exist in the Canning Basin. Campbell (Dunedin) (1994) describes Late Triassic brachiopods from shelf carbonates from a single site on the Rowley Terrace.

Four miscellaneous papers complete the volume. A paper by Colwell and Symonds (AGSO) and Crawford (Tasmania) (1994) on the geology of the Cretaceous Wallaby Plateau, and its relationship to other igneous bodies of the western Australian margin, provides a better understanding of volcanism associated with Gondwanan breakup. A paper by Buffler (Austin, Texas) (1994), on a seismic tie line on the Argo Abyssal Plain between ODP Site 765 and DSDP Site 261, provides an insight into the history of the eastern plain since it formed following Late Jurassic breakup. A complementary paper by an Indian and Russian team headed by Rao (Goa) (1994) reports the results of several Soviet geophysical cruises over the Argo Abyssal Plain. The resultant maps give an understanding of the development of the sequences below most of the plain. The final paper by Exon & Colwell (AGSO) (1994) synthesises all the results in the form of a geological history of the outer North West Shelf.

Acknowledgements

This introduction has profited from editorial comments and suggestions from Barry Willcox and Jim Colwell (AGSO), and Peter Purcell (P&R Geological Consultants, Perth).

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STRATIGRAPHIC TABLE : BARCOO 1

Latitude: 15°20'37"S Longitude: 120°38'12.2"E RT 11m

Era	Period	Epoch	Age	Lithology	Depth (m)	Thickness (m)	Reflector	Two-way time (s)
				Sea level	11	720		
				Sea bed	731			
			NOT SAMPLED			394		
CAINOZOIC	NEOGENE		MIDDLE MIOCENE	Shelf skeletal calcarenite; minor calcisiltite	1125	580		
CAINOZOIC	NEOGENE		UPPERMOST EARLY MIOCENE	Shelf calcarenite, calcisiltite and calcilutite	1705	517		
CAINOZOIC	NEOGENE		EARLY MIOCENE	Upper slope marl; minor calcilutite	2222	63	A_0	1.99
CAINOZOIC	PALAEOGENE		LATE OLIGOCENE	Upper slope marl; minor calcilutite	?2285	58		
CAINOZOIC	PALAEOGENE		LATE EOCENE	Upper slope calcilutite	2343	12	A_1	2.07
CAINOZOIC	PALAEOGENE		INDETERMINABLE	Upper slope calcilutite	2355	45		
CAINOZOIC	PALAEOGENE		MIDDLE EOCENE	Upper slope calcilutite, partly recrystallised	2400	30		
CAINOZOIC	PALAEOGENE		INDETERMINABLE	Upper slope calcilutite, partly recrystallised; and calcareous claystone	2430	27		
CAINOZOIC	PALAEOGENE		EARLY EOCENE	Upper slope calcilutite, partly recrystallised; and calcareous claystone	2457	73		
CAINOZOIC	PALAEOGENE		LATE PALEOCENE	Upper slope calcilutite and calcareous claystone	2530	12		
CAINOZOIC	PALAEOGENE		LATE PALEOCENE	Outer shelf argillaceous calcilutite, claystone and marl	2538	100	A_2	2.17
CAINOZOIC	PALAEOGENE		MIDDLE PALEOCENE	Outer shelf argillaceous calcilutite	?2638	37		
CAINOZOIC	PALAEOGENE		EARLY PALEOCENE	Upper slope marl, minor calcilutite	2675	27		
MESOZOIC	CRETACEOUS	LATE	LATE MAASTRICHTIAN	Upper shelf marl, minor calcilutite	2702	47	B	2.28
MESOZOIC	CRETACEOUS	LATE	LATE CAMPANIAN	Outer shelf marl; increasing calcilutite downward	2749	217		
MESOZOIC	CRETACEOUS	LATE	EARLY CAMPANIAN	Outer shelf marl; increasing calcilutite downward	2966	1		
MESOZOIC	CRETACEOUS	LATE	SANTONIAN	Upper slope to outer shelf calcilutite, marl and calcareous claystone	2967	34		
MESOZOIC	CRETACEOUS	LATE	CONIACIAN	Upper slope calcareous claystone	3001	5		
MESOZOIC	CRETACEOUS	LATE	TURONIAN	Upper slope calcilutite	3006	1		
MESOZOIC	CRETACEOUS	LATE	PROBABLE MIDDLE CENOMANIAN	Outer shelf claystone	3007	47	C	2.47
MESOZOIC	CRETACEOUS	EARLY	EARLIEST CENOMANIAN TO LATEST ALBIAN	Shelf claystone	3054	83		
MESOZOIC	CRETACEOUS	EARLY	LATE TO MIDDLE ALBIAN	Shelf claystone	3137	37		
MESOZOIC	CRETACEOUS	EARLY	ALBIAN UNDIFFERENTIATED	Shelf claystone	3174	156		
MESOZOIC	CRETACEOUS	EARLY	EARLY ALBIAN TO LATE APTIAN	Upper slope claystone; clayey siltstone below 3565m; siliceous plankton	3330	323		
MESOZOIC	CRETACEOUS	EARLY	EARLY APTIAN TO LATE BARREMIAN	Inner shelf claystone	3653	59		
MESOZOIC	CRETACEOUS	EARLY	BARREMIAN TO VALANGINIAN	Inner shelf claystone; marl below 3770m	?3712	84		
MESOZOIC	JURASSIC	LATE	OXFORDIAN TO LATE CALLOVIAN	Shelf claystone; minor sandstone	3796	20	D	3.01
MESOZOIC	JURASSIC	LATE	CALLOVIAN	Shelf claystone	3816	13		
MESOZOIC	JURASSIC	MID	BAJOCIAN TO AALENIAN	Inner shelf sandstone, claystone	3829	276	E	3.02
MESOZOIC	JURASSIC	EARLY	TOARCIAN	Inner shelf to paralic claystone, minor sandstone	4105	178		
MESOZOIC	JURASSIC	EARLY	PLIENSBACHIAN TO HETTANGIAN	Inner shelf claystone, calcareous claystone, minor limestone	4283	465		
MESOZOIC	TRIAS	LATE	RHAETIAN	Inner shelf claystone	?4748	97	F	3.48
MESOZOIC	TRIAS	LATE	RHAETIAN TO NORIAN	Shelf claystone, calcareous siltstone; recrystallised limestone below 4930m; minor acid intrusives at 4945m	?4845	264		

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Figure 4. Stratigraphic table of Woodside Barcoo No. 1 petroleum exploration well. Location in Figure 1. Modified from company well completion report (Woodside, 1980).

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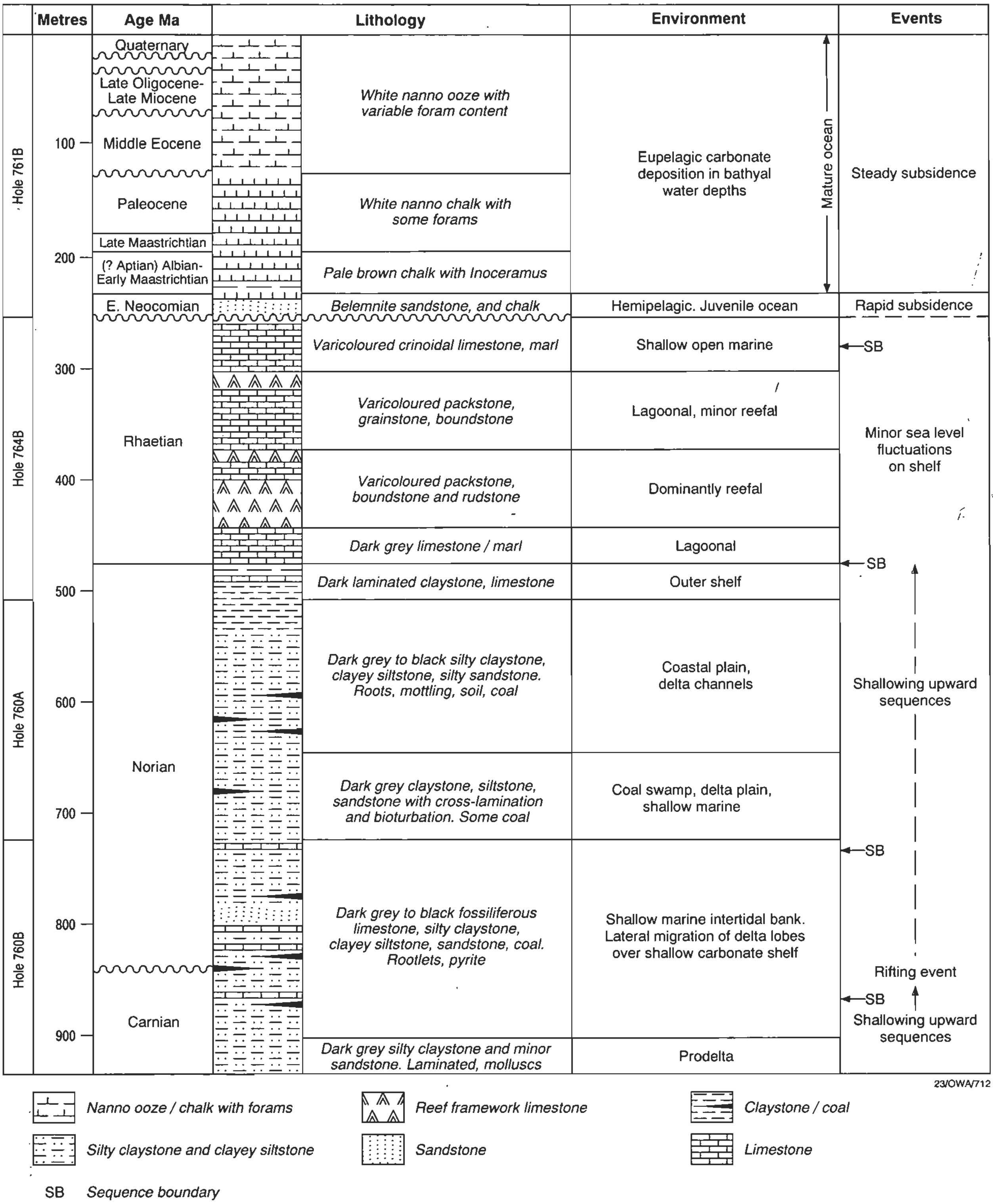

Figure 5. Composite stratigraphic diagram from Wombat Plateau ODP Sites 759, 760, 761 & 764. Locations in Figure 1. Shows maximum thicknesses and representative facies of the most studied Triassic sequence on the North West Shelf, and a 40 m.y. unconformity between the Triassic and Cretaceous. After von Rad et al. (1992).

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Mesozoic sedimentary and volcaniclastic rocks dredged from the northern Exmouth Plateau and Rowley Terrace, offshore northwest Australia

J.B. Colwell[1], U. Röhl[2], U. von Rad[2] & E. Kristan-Tollmann[3]

An extensive program of dredging was undertaken by AGSO in 1990 on the northern Exmouth Plateau and Rowley Terrace margin, mainly to provide geological control for seismic interpretations. Forty-two major Mesozoic lithofacies types were recovered. The rocks indicate that:

- During the Late Triassic (Norian–Rhaetian), the Rowley Terrace margin was, like the adjacent northern Exmouth Plateau, host to shelf carbonates, including reef and peri-reefal deposits. Volcanics were emplaced along the margin (particularly in the north) during the Late Triassic–Mid Jurassic, probably as a result of the commencement of rifting between Australia and Greater India. The volcanics were laid down partly or reworked into shallow water, producing hyaloclastites and tuffaceous mudstones with marine microfossils. By the Early Jurassic, carbonate deposition was restricted to relatively small areas with most of the outer Rowley Terrace being covered by thick, fluvial–paralic siliciclastics with marine interbeds. These deposits continued to be deposited until the early Late Jurassic. They were subjected in the ?Late Jurassic to subaerial exposure under arid conditions resulting in cementation by Fe-oxides. With the early Late Jurassic (Oxfordian–Callovian) breakup and formation of the Argo Abyssal Plain, the margin started to subside and thin, shallow-marine, Late Jurassic–Early Cretaceous sediments were deposited. These were followed, with continuing margin subsidence, by progressively deeper-water deposits, including Cretaceous and younger hemipelagic–eupelagic claystones, chalks, and marls.
- Reefs grew in the area of the outer part of the Rowley Terrace during the Late Triassic. The potential therefore exists for Upper Triassic, reefal petroleum reservoirs to be present within the main part of the Rowley Sub-basin to the east, if reefs in that area were not prevented from growing by sand influx from rivers. Similarly, Lower Jurassic shallow-water platform carbonates, which are present on Triassic horst blocks beneath the central northern part of the Exmouth Plateau, could act as petroleum reservoirs if porosity is preserved.
- During the Late Triassic and Early Jurassic there is a clear similarity between the facies and associated foraminiferal and ostracod microfaunas of the northern Exmouth Plateau/Rowley Terrace area and those of other southern Tethyan margins, including the Northern Calcareous Alps. This indicates that broadly similar depositional conditions existed along much of Neo-Tethys at that time.

Introduction

In recent years, a considerable amount of effort has been directed at studying the geology and tectonic development of the Exmouth Plateau, largely as a result of drilling undertaken in 1988 by the Ocean Drilling Program (ODP) Legs 122 and 123 (Fig. 1; Haq et al., 1990; Gradstein, et al., 1990). One of the major outcomes of the ODP drilling was the recognition for the first time of Triassic reefs in Australia, a potential new exploration play on the outer North West Shelf (Williamson et al., 1989; Exon et al., 1989, 1991; von Rad et al., 1992c; Röhl et al., 1992). This discovery prompted the Australian Geological Survey Organisation (AGSO; formerly the Bureau of Mineral Resources, BMR) to undertake additional dredging and seismic profiling in the region in 1990 (AGSO Surveys 95 and 96). This work was aimed at better defining the extent and seismic character of the reef complexes, tracing them, if posssible, back to NW Shelf commercial wells, and to further refine the Mesozoic stratigraphy and, consequently, the seismic interpretation of the region (Exon et al., 1990; Colwell et al., 1990; Ramsay & Exon, 1994 — this issue).

This paper synthesizes the results of the *Rig Seismic*'s 1990 dredging on the northern Exmouth Plateau and the adjacent Rowley Terrace (Fig. 2). General analysis of thin-sections was undertaken by J. Colwell (AGSO), U. Röhl (Bundesanstalt für Geowissenschaften und Rohstoffe, BGR), and U. von Rad (BGR); microfacies analysis of carbonates by U. Röhl (BGR); foraminiferal micropalaeontology by E. Kristan-Tollmann (University of Vienna), D. Lynch (University of Western Australia), and P. Wells (Australian National University); nannofossils by S. Shafik (AGSO); palynomorphs by D. Burger (AGSO); conodonts by R. Nicoll (AGSO); bivalves by J. Grant-Mackie (University of Auckland); brachiopods by D. Campbell (University of Otago); corals by G. Stanley (University of Montana); and igneous petrology by A. Crawford (University of Tasmania). Details of many of these specialist studies are given in other papers here, namely: Shafik; Burger; Nicoll & Foster; Grant-Mackie; Campbell; Stanley; and Crawford & von Rad (all 1994 — this issue). Details of the work of Lynch and Wells are given in Exon et al. (1990) and Colwell et al. (1990), respectively. Ramsay & Exon (1994 — this issue) use the geological information in this paper as a major control on their interpretation of seismic data in the region.

The environmental interpretations given here are based primarily upon palaeontological evidence. Where units are unfossiliferous, sedimentological criteria have been used as appropriate.

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Geological background

Northwest Australia is characterised by an old, sediment-starved continental margin. This margin has undergone a complex syn-rift and post-breakup history which has produced a range of geologic settings and physiographic features.

Northern Exmouth Plateau

The Exmouth Plateau (Figs 1 and 2) forms part of the northern Carnarvon Basin (Exon & Willcox, 1980; Hocking et al., 1987; Cockbain, 1989). The generalised stratigraphy of the northern part of the plateau is shown in Figure 3, based upon a study of North West Shelf wells and seismic profiles (Willcox & Exon, 1976; Exon & Willcox, 1978, 1980: Ramsay & Exon, 1994 — this

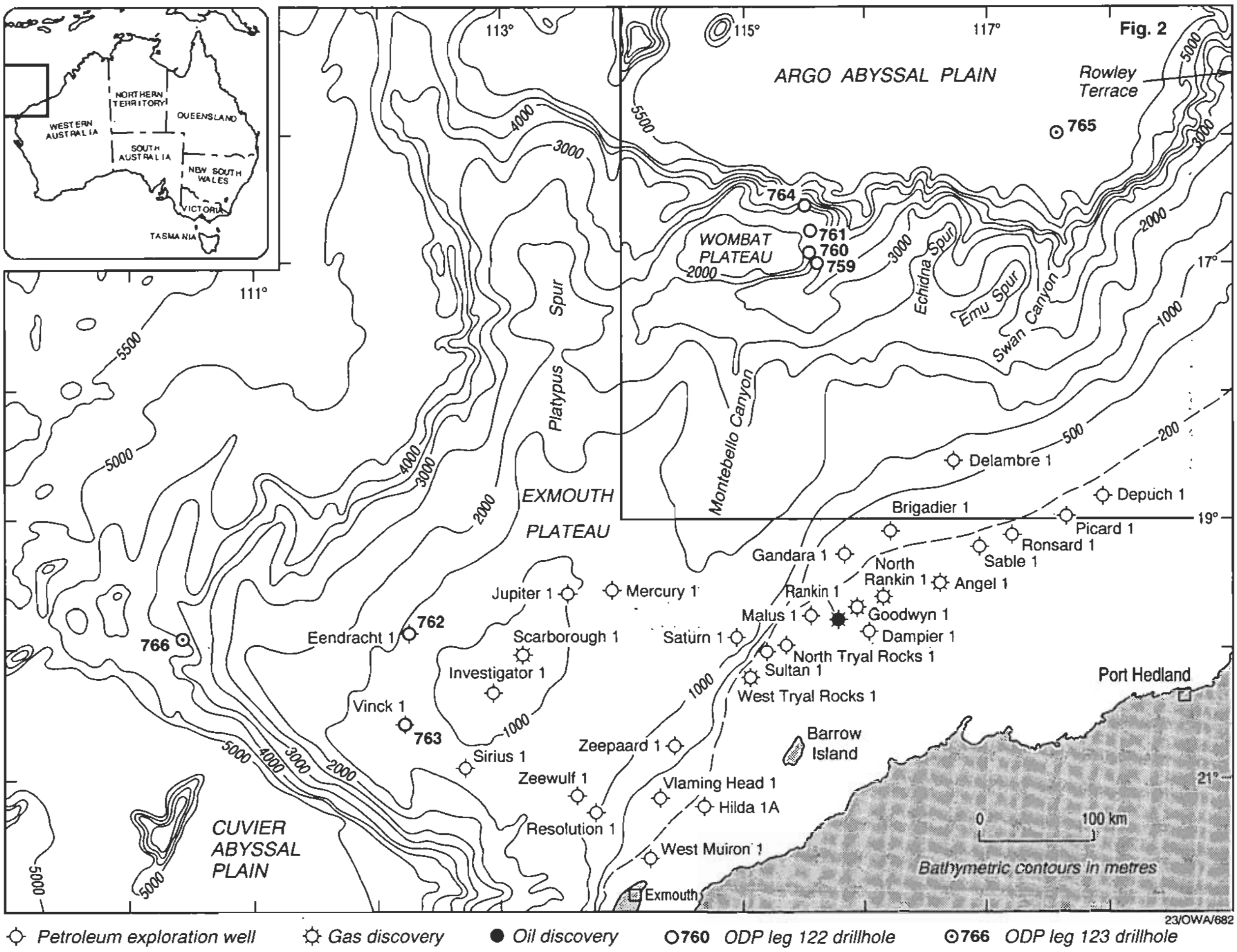


Figure 1. Location map (after Exon et al., 1992). Detailed location shown in Figure 2.

issue), dredging on the plateau margin (von Stackelberg et al., 1980; von Rad & Exon, 1983; von Rad et al., 1990; this paper), and drilling during ODP Leg 122 (Haq et al., 1990; von Rad et al., 1992c). The northern part of the plateau is downwarped along the North Exmouth Hinge Zone and cut by major northeast and east-trending, steeply dipping faults into a series of horsts, grabens and half-grabens (Wombat Plateau, Montebello Canyon, Emu and Echidna Spurs, Swan Canyon, etc.; Fig. 1).

The oldest rocks drilled or dredged in the area are Upper Triassic (Carnian) siltstones and sandstones overlying a thick Permo-Triassic section (Exon et al., 1992). Extensive dredging by R/V *Sonne* in 1979 (cruise SO–08) and R/V *Rig Seismic* in 1986 (Survey 56) along the western side of the Swan Canyon and flanks of the Cygnet Canyon (Figs 2 and 4) recovered a range of Upper Triassic carbonates and Lower to Middle Jurassic 'coal measures'. These are overlain by marginal-marine to marine Lower Cretaceous claystones, and then by deeper-water Upper Cretaceous to Cainozoic marls, chalks and oozes (von Stackelberg et al., 1980; von Rad et al., 1990; Kristan-Tollmann & Gramann, 1992). This contrasts with the horst areas, such as the Wombat Plateau where ODP drilling has shown that Upper Triassic (Carnian and Norian) deltaic siltstones and mudstones are overlain by Rhaetian shallow-water limestones (including reefs; Röhl et al., 1992; Kristan-Tollmann & Gramann, 1992). These are overlain in turn above the post-rift unconformity by a thin lower Neocomian sand, a very condensed Cretaceous chalk sequence, and Cainozoic chalks and oozes. Jurassic sediments were eroded after Callovian/Oxfordian times on highs, such as the Wombat Plateau (Exon et al., 1992; von Rad et al., 1992a,b).

The objective of the present work was to further refine the stratigraphy of the northern part of the plateau, particularly in the vicinity of the Swan Canyon adjacent to the Rowley Terrace margin (Survey 96 sites 12–18, Fig. 2). In addition, two samples of basement were taken on the western flank of the Wombat Plateau (Survey 96 sites 31–33), and two (Survey 96 sites 29 & 30) on the central northern part of the plateau over what had been interpreted on AGSO seismic data to be possible Jurassic "reefs" sitting on Triassic horst blocks (Exon, 1990).

Rowley Terrace

The Rowley Terrace, which adjoins the northern Exmouth Plateau margin (Fig. 2), is underlain by the Rowley Sub-basin. This sub-basin connects the Beagle Sub-basin of the northern Carnarvon Basin in the south to the Browse Basin in the north, although the basin boundaries are ill-defined (Stagg & Exon, 1981; Horstman & Purcell, 1988). Interpretation of regional seismic lines from this area, including recent AGSO data tied to Barcoo–1 and East Mermaid–1 exploration wells (Fig. 2; Ramsay & Exon, 1994 — this issue), indicates that: (1) the geology is broadly similar to that of the northern Exmouth Plateau

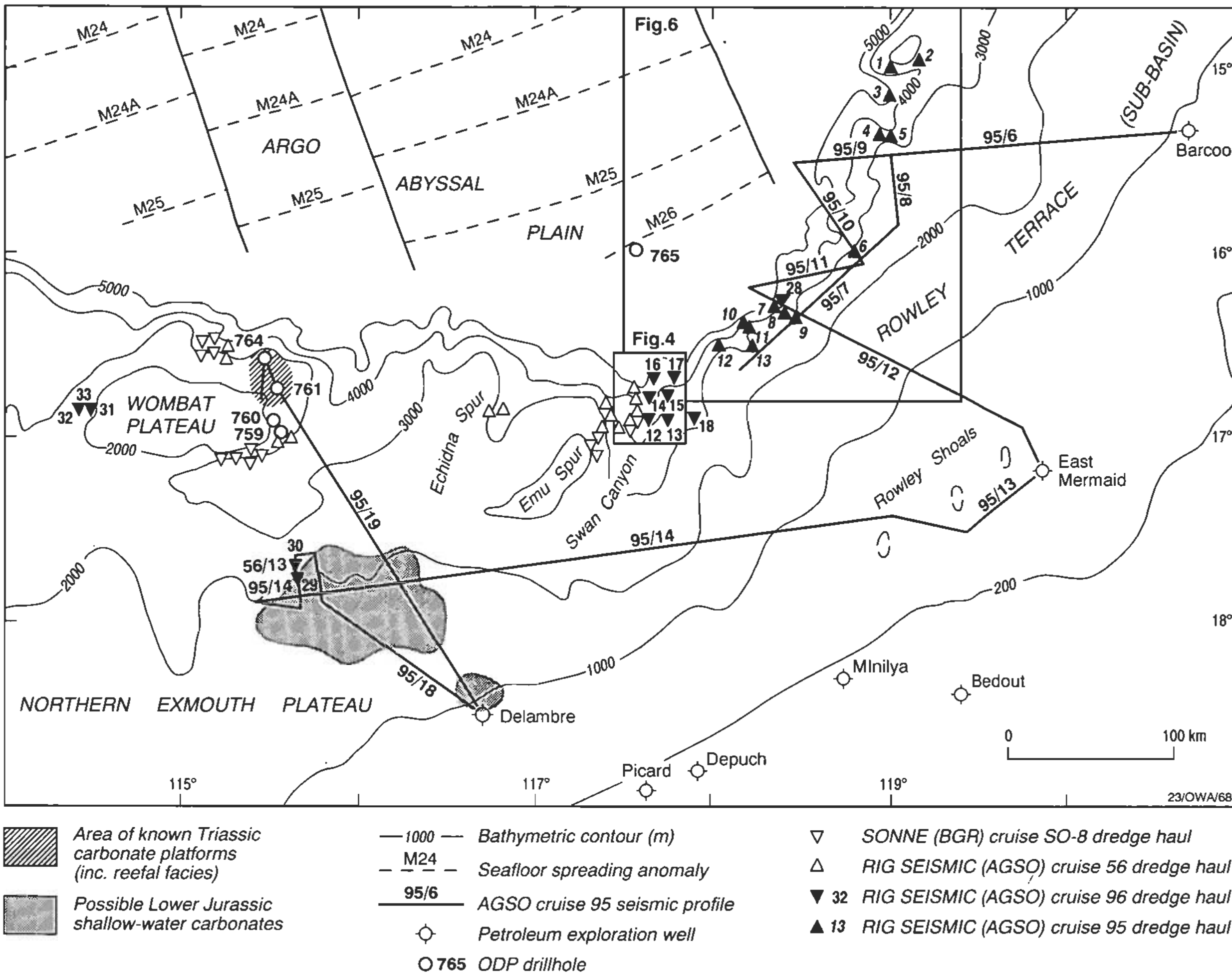


Figure 2. Map of the northern Exmouth Plateau–Rowley Terrace margin showing the location of Survey 95 and 96 dredge hauls. Survey 95 seismic lines, previous (*Sonne* SO-8 and BMR Survey 56; von Stackelberg et al., 1980; von Rad et al., 1990) dredge sites, ODP drill holes, and selected petroleum exploration wells are also shown.

area, and (2) sediments of presumed Triassic–Middle Jurassic age are widely exposed in canyon systems along the margin of the Rowley Terrace.

The present study was aimed at providing the geological control to 'ground-truth' these seismic interpretations as well as to add to the regional stratigraphic picture. Thirteen dredge hauls (1–13) were taken during AGSO Survey 95 and one (28) during the subsequent Survey 96 in outer-margin canyons (Fig. 2); all were successful in recovering material. The dredge sites were located on existing seismic lines ranging from 1971 Shell *Petrel* regional data to AGSO's 1990, high-quality *Rig Seismic* data.

Dredging results

All samples were taken from R/V *Rig Seismic* using a chain-bag dredge with attached pipe dredge. Full information on the sample locations and techniques are contained in the respective cruise reports: Exon et al. (1990) for Survey 95, and Colwell et al. (1990) for Survey 96.

The rock types recovered are detailed in Table 1 in approximate stratigraphic order. Age control is based on a variety of macro- or microfossil groups (see Table 2). The rocks range in age from Neogene (chalks and manganese crusts) to Late Triassic (shallow-water carbonates, and volcanics). Lithofacies are designated by letters and numbers (e.g. E3) as was first done by von Stackelberg et al. (1980). Although common in the dredge hauls (see Figs 5, 7 and 8), Cainozoic deposits (Quaternary muddy sands, muds and oozes; and Tertiary chalks and marls) are not described in detail here which concentrates on the more-important, Mesozoic section. Details of the benthic foraminiferal faunas present in some important Late Triassic–Early Jurassic carbonates are given in the Appendix.

Eastern side of the Swan Canyon

Six dredge hauls were taken from the eastern wall of the Swan Canyon, and one farther to the east, in areas across the canyon from the slopes dredged by the R/V *Sonne* and *Rig Seismic* in 1979 and 1986 (Fig. 4). Dredges 96/12 and 13 were located in water depths of 4950–-3600 m; 14 and 15 from the central part of the canyon in water depths of 5250–3875 m; and 16 and 17 from the northern (seaward) part of the canyon 5610–3910 m below sea level. Dredge 96/18 was located approximately 20 km east of dredge 13, on the slope leading to the canyon, in 3360–2980 m of water.

Rocks recovered include Upper Cretaceous (mid Campanian–late Maastrichtian) calcareous mudstones (Lithofacies E3 of Table 1); radiolarite (E4) and radiolarian

Table 1. Details of lithofacies.

Litho-facies	*Lithology*	*Age**			*Description*	*Depositional environment†*	*Samples§*	*Loca-tion#*	*Equiv. microfacies Röhl et al. (1992)‡*
G1	Mn crusts	Neogene		(Neogene)	Variable morphology and thickness	Open marine	95:1/7, 2/3, 12/19 96:14/7,29/6, 30/2	RT SC, EP	—
E1	Chalk & calc. mudstone	Neogene		Early–Late Miocene (1,2,3)	White-gr. orange, mod. soft-soft, common borings & burrows, foraminiferal, eupelagic-hemipelagic	Open marine	95:1p,4/1E,4/5, 4P,11/8,13/M1b 96:18/3,29/1, 29/2,29/3,29/4, 29/5,32/2,33/2, 33/3,33/4	RT SC, EP, WP	—
E2	Marl & chalk	Paleog.		Late Paleocene–Mid. Eocene (1,2)	White-grey, soft, commonly bored, 13/2: radiolaria bearing	Open marine	95:13/1,13/P1, 13/M1a 96:15/3,15/5, 13/2,13/9	RT SC	—
E3	Chalk & calcareous mudstone	Cretaceous	Late	Mid. Campanian — Late Maastrichtian (1,3)	Grey. orange-yell. grey, mod. soft-soft, commonly bored & burrowed 14/4: trace of radiolaria, very slightly silicified in places	Open marine	95:4/1B 96:13/7,14/4, 14/9	RT SC	—
E4	Radiolarite & silicified radiolarian mudstone	Cretaceous	Early	(Early Cretaceous)	Mod. yellow. br. — yellow, hard	Open (?shelfal) marine	95:8/5A 96:12/3	RT SC	—
E5	Radiolarian chalk — calc. mudstone	Cretaceous	Early	Valangin.–early Aptian (1,3)	Lt. ol. grey -yell. grey, commonly bored, variable carbonate content	Open marine	95:4/1C,4/4 96:13/3,13/5, 14/8,15/1,15/6	RT SC	—
C1	Coquinite (clayey bioclastic rudstone)	Cretaceous	Early	Late Berriasian–early Barremian (1)	Dusky yellow-mod. red. brown, *Inoceramus* plates dominant, poorly sorted, <20% iron oxides, clays & silt size quartz	Shallow marine, ? minor subaerial exposure	96:13/1,13/8	SC	—
H1	Calcite-cemented, volcaniclastic conglomerate	Jurassic	Late	Oxfordian (3)	Highly altered, very poorly sorted, >60% altered volc. rock frags., 5% crinoid, echinoid & bivalve frags.	Shallow marine	95:2/2	RT	—
B1	Boxstones, ferruginous crusts & concretions	Jurassic		(Mid–Late Jurassic)	Dark brown, hard	Subaerial, arid	95:(4/15,5/1,5/4, 6/2,7/7,8/2,8/6, 9/9,10/9,11/9, 13/7) 96:(17/15),18/1	RT SC	—
B2	Ferruginous quartz sandstone	Jurassic		(Mid–Late Jurassic)	Reddish brown-br. black, variable sorting, v.f.-c. grained, cemented with iron oxides	Fluvial/paralic, later subaerial exposure	95:(4/12,4/13, 4/14,13/3) 96:14/12,17/7	RT SC	—
B3	Ferruginous mudstones	Jurassic		(Mid–Late Jurassic)	Grey-mod. br., hard, 20% fine silt, mica in places	Fluvial/paralic, later subaerial exposure	95:(5/3) 96:28/13	SC RT	—
A1	Coal	Jurassic		(Jurassic)	Black, vitreous	Paralic	95:(12/1)	RT	—

Litho-facies	Lithology	Age*		Description	Depositional environment†	Samples§	Loca-tion#	Equiv. microfacies Röhl et al. (1992)‡
A2	Carbonaceous, clayey siltstones & silty claystones		Toarcian-Kimmeridgian (1,3,5,10)	Black-grey, carbonaceous, mainly qtz silt with minor feldspar, trace mica-micaceous, commonly bored, iron oxide staining and crusts in places, some sandstone(A3) stringers. Cherty lithics in 28/11, 28/11A & 28/11B. Minor secondary carbonate in 28/15 & 28/16. Nannofossils (marine) in 6/3,6/4,7/8 and 28/16.	Fluvial/paralic (flood plain, marsh,swamp, delta etc.) with marine incursions in places. Marine post-Callovian. Minor subaerial exposure.	95:4/16A,5/5A, 5/9A,6/1A,6/3, 6/4,7/3,7/8,9/4A, 11/2A,11/3A,11/7A, (4/3,4/17,5/7,5/11, 5/14-17,7/3,8/1,9/5, 11/5,11/6,12/2, 12/3,13/4,13/6) 96:12/1,12/4, 12/5,13/4,14/5,14/6, 14/10,15/2,16/2,17/4, 17/5,17/6,17/9,17/19, 18/5,18/6,28/11,28/ 11A,28/11B,28/15, 28/16	RT RT, SC	—
A3	Very fine-medium grained,quartzose sandstone(minor greywacke)	Jurassic	Mid-Late Jurassic (1,3)	Lt. olive grey-dk. yell. orange, gen. well sorted, micaceous in places, minor secondary carbonate, variable amounts of clay matrix and iron oxides, framework is mainly qtz with minor feldspar and mica (and trace of lithics), minor organics, common interbeds of carbonaceous mudstones. X-bedded in places.12/5: med.-coarse grained.	Fluvial/paralic (flood plain). Minor marine component (sample 4/7).	95:4/7,4/11A, 7/2A,7/5A,7/6 (4/2,4/6,4/8,4/9,5/1, 5/8,10/2,12/4,12/5, 13/2) 96:12/2,14/1, 14/2,16/1,17/2,17/3, 17/10,17/11,17/12, 18/2	RT SC	—
A4	Coarse-v. coarse grained, quartzose sandstone		(Jurassic)	Med. grey-lt. brown, poorly-very poorly sorted, generally mature, mainly quartz, trace feldspar, chert and lithics, matrix composed of clays and Si cement 5/2: ferruginous	Fluvial	95:4/10A, (5/2,7/1) 96:17/8	RT SC	—
A5	Ortho-quartzite		(Jurassic)	Lt. br. grey-dk. grey. br., compacted, qtz overgrowths, very mature, coarse-v. coarse grained. Water-worn pebbles	?Fluvial	95:10/1A	RT	
A6	Calcite-cemented, quartz sandstone		(Jurassic)	Very fine-fine grained, 50% calcite cement, 5% feldspar, ~3%mica, 5% glauconite in 7/10B	Paralic (transitional marine)	95:7/10B,9/1A	RT	
C2	Mudstone with thin packstone layers		Toarcian (5)	Dark grey-black, contains echinoderm — mollusc packstone layers, trace of iron oxides and pyrite	Paralic (paludal) — shallow marine	95:7/14A,?9/2C	RT	11 (coarse layers)
C3	Bivalve-rich wacke-packstone (packed biomicrite)	Early	Hettangian — lower Sinemurian (4)	Grey. orange-dk. yell. orange, iron oxide stained, poorly sorted, allochems mainly recrystallised bivalve frags., minor echinoid debris and forams, mainly micritic matrix, minor spar, trace of quartz and feldspar.	Shallow marine (shelf)	96:30/1,30/1A, 30/1B,30/1C	EP	?
C4	Echinodermal packstone–grainstone ('encrinite')		?Liassic (4)	Lt. grey-brown, approx. 10% quartz, 50% calcite cement, 10% volcanic lithics, 20% echinoderm debris, 5% mollusc frags, trace of foraminifera.	Shallow marine (shelf)	95:7/12A	RT	11
C5	Bivalve-rich, skeletal wackest.		Early Jurassic (9)	Brownish yellow, poorly sorted	Shallow marine (shelf)	95:8/4	RT	7

Litho-facies	Lithology	Age*		Description	Depositional environment†	Samples§	Location#	Equiv. microfacies Röhl et al. (1992)‡
A7	Calcite-cemented, sub-arkosic sandstone	Late Triassic–Jurassic	(Late Triassic–Jurassic)	Med. coarse-very coarse grained, alternates with very fine grained quatz sst., up to 10% feldspar, minor glauconite, trace of lithics	Paralic (transitional marine)	95:7/12B	RT	—
A8	Sub-arkosic sandstone/ litharenite		Rhaetian-earliest Jurassic (5)	Med. light grey-light grey, fine-coarse grained, poorly sorted in places. Up to 50% quartz, 30% feldspar & 30% cherty lithics & volcs. in framework. Trace mica. Matrix: quartz cement dominant, minor clays	Paralic (transitional marine)	96:28/10,28/12, 28/14	RT	—
A9/ ?D	Carbonaceous mudstones-fine & v.fine sst.	Late Triassic	Rhaetian (5)	Greyish brown-dk. grey, laminated, carbonaceous, slightly micaceous, framework mainly v.f.-f. grained quartz, minor feldspar, volc. rock frags., and mica. Trace of carbonate.	Restricted marine	95:7/4A,7/11A, 10/3A,10/4A,10/4B, 10/4C,11/2A,(?3/2, 10/5,11/1)	RT	—
C6	Dolomitic/sideritic limestone		(Late Triassic)	Grey-reddish brown,minor silt-sized quartz. 28/8: virtually cryptocrystalline with a few 'ghosts' of skeletal material	?Lagoonal-shallow marine	95:7/16A 96:28/8	RT	?1
C7	Calcrete		(Late Triassic)	Grey. orange-pink, hard, minor iron oxide staining, 20% skeletal debris	Shallow marine with subaerial exposure	96:28/5	RT	?
C8	Mudstone-wackest.		(LateNorian) — Rhaetian (4)	White-grey. brown, minor quartz, fossil 'ghosts', dolomitized in places	Shallow marine-lagoonal	95:10/10A, 10/10B,9/8V	RT	?3
C9	Peloidal wackestone -mudstone		Late Triassic (6)	Brown, ferruginous, brachs., partly dissolved crinoids, peloids, forams, 20% clays, 50% micrite, 10% quartz	Restricted lagoonal	95:7/13A	RT	4a
C10	Iron-oxide-rich packstone		(Late Triassic)	Pale yell. brown, 15-20% iron-oxide-rich ooids (carbonate replacement), shallow-water allochems cemented by microspar cement. Trace of rounded volc. grains, quartz and forams	Shallow-water shoal with subsequent subaerial exposure	96:28/1A	RT	?
C11	Echinodermal (crinoid-rich) wacke-to packstone		M. Norian–Rhaetian (7)	Yell. - light br. grey, med. - coarse grained, abundant crinoid and echinoid fragments (crinoid-rich biosparite)	Shallow marine	95:10/6A,10/6D, 10/6W 96:28/2a,28/2b,28/6	RT	6
C12	Skeletal wackestone -floatstone		Norian-Rhaetian (4,7,8)	Grey-yell. grey, slightly recrystallised, contains mollusc debris, criniod/echinoid pieces, minor forams. Minor dolomitized ooids in 12/9V (mainly packed biomicrites-biosparites)	Lagoonal-shallow marine	95:10/8B, 10/8V,12/9V,12/9C, 12/10A,12/10D 96:28/1b,28/1c,28/1d	RT	7
C13	Peloidal wackestone -mudstone		Rhaetian (7)	Yellowish grey, massive, 10% skeletal material inc. echinoderm and bivalve fragments and forams(fossiliferous micrite)	Lagoonal-shallow marine	96:28/1a	RT	4a
C14	Peloidal grainstone -packstone		(Late Triassic)	Dk. grey, includes echinoderm fragments, quartz and forams. Recrystallised.	Shallow marine	95:7/14B	RT	9
C15	Skeletal packstone		Late Norian (4)	Dk. yellowish brown-grey, medium-fine grained, mainly echinoderm, foram and mollusc fragments, minor volcaniclastics and quartz. (biosparite-biomicrite)	Shallow marine	95:9/6B 96:28/3a,28/3b,28/3c	RT	10
C16	Oolitic grainstone		(Late Triassic)	Lt. yell. brown, well sorted ooids with quartz nuclei	Shoal	95:9/7V	RT	17
C17	Grainst.-rudstone		Norian-Rhaetian (4,9)	Pale yell., med. -coarse grained, composed mostly of bivalve and crinoid fragments, forams and dolomitized ooids. Poorly sorted.	Shoal	95;12/9W,12/11, 12/11V	RT	22

Litho-facies	Lithology	Age*		Description	Depositional environment†	Samples§	Location#	Equiv. microfacies Röhl et al. (1992)‡
C18	Algal boundstone	L. Tr.	(Late Triassic)	Yellowish white	Restricted lagoonal	95:10/6V,10/8A	RT	23
C19	Coral bounstone	L. Tr.	Norian–Rhaetian (4)	Grey-light grey, contains hexacorals, minor mollusc debris, ostracods and forams. Recrystallised, no terrigenous material. Grades into wackestone.	Reef	95:7/15A,7/15B, 7/15V	RT	24/25
H2	Calcite-cemented lapillist.-hyaloclastite	(Late Triassic–Jurassic)	(Late Triassic–Jurassic)	Commonly graded, calcite cemented, particles from fine sand to pebble size. Volcanic fragments highly altered, (iron oxides, chlorite & zeolites). Largely vitric (pumice), also some vesicular basalt	?Shallow marine	95:6/6A,6/6C1, 6/6C2,6/6B,7/10A 96:28/7,28/9a,28/9b, 28/9c	RT	—
H3	Volcaniclastic (altered hyaloclastite) - basaltic breccia	(Late Triassic–Jurassic)	(Late Triassic–Jurassic)	Brown-lt. br.-greenish grey, totally altered (chloritised,smectitised,ferruginised), mainly vesicular basalt, some altered glass. 1/6: contains secondary calcite.	?	95:1/2A,1/6	RT	—
H4	Volcaniclastic litharenite	(Late Triassic–Jurassic)	(Late Triassic–Jurassic)	Dk. yell. orange - pale brown, mixture of lithics(cherty & volc. grains), and highly altered glauconite/limonite-rich grains, ?phosphate, well developed iron oxide meniscus cements, trace of calcite cement	?	95:(1/5) 96:17/1A,17/1	RT SC	—
H5	Volcaniclastic/ tuffaceous sandstone	(Late Triassic–Jurassic)	(Late Triassic–Jurassic)	Grey-dk. grey-olive, fine grained, up to 20% quartz.	?	95:5/12,8/3A, (1/3,5/10,5/13)	RT	—
H6	Tuffaceous mudstone	(Late Triassic–Jurassic)	Rhaetian–Toarcian (3)	Grey green,slightly calcareous, ?reworked tuff, bioturbated	?marine	95:6/5	RT	—
H7	Volcaniclastic - sedimentary breccia	(Late Triassic–Jurassic)	(Late Triassic–Jurassic)	Extremely heterogeneous. Contains metamorphosed 'basalt', silty claystone, cherty sediments and carbonate rock fragments and fossils	?	95:10/11A, (?12/17)	RT	—
H8	Vesicular basalt/ aphyric andesite/ trachyandesite	(Late Triassic–Jurassic)	(Late Triassic–Jurassic)	Commonly extensively altered to iron oxides	—	95:1/4A,1/4B, 2/1A,5/6A,10/12A, 10/13A10/14A, 10/15A(10/16, ?12/14,12/15) 96:32/3,33/1	RT WP	—

* Ages in brackets are estimates only. Ages determined from: 1: nannofossils (Shafik), 2: foraminifera (Wells), 3: foraminfera (Lynch), 4: foraminifera (Kristan-Tollmann), 5: palynomorphs (Burger), 6: brachiopods (Campbell), 7: conodonts(Nicoll), 8: corals (Stanley), 9: pelecypods (Grant-Mackie), 10: foraminifera (Apthorpe, pers. comm., 1993). Age details either in papers in this issue (Shafik, Burger, Campbell, Nicoll & Foster, Stanley, and Grant-Mackie; all 1994) or in the respective cruise reports — Exon et al. (1990) for Survey 95, and Colwell et al. (1990) for Survey 96.

† Environmental interpretations based primarily upon palaeontological evidence. Where units are unfossiliferous, sedimentological criteria have been used as appropriate.

§ Sample nomenclature: XX:Y/Z where XX is the cruise number (95 or 96), Y is the dredge number, Z is the rock type within the dredge haul, and the suffix A, B, C etc. identifies a particular specimen.
Underlined samples have been dated (see Table 2).
Samples in brackets have been assigned to the lithofacies on the basis of hand specimen only.

\# Location: RT: Rowley Terrace, SC: eastern side of the Swan Canyon, EP: northeastern Exmouth Plateau, WP: western margin of Wombat Plateau.

‡ See Table 3.

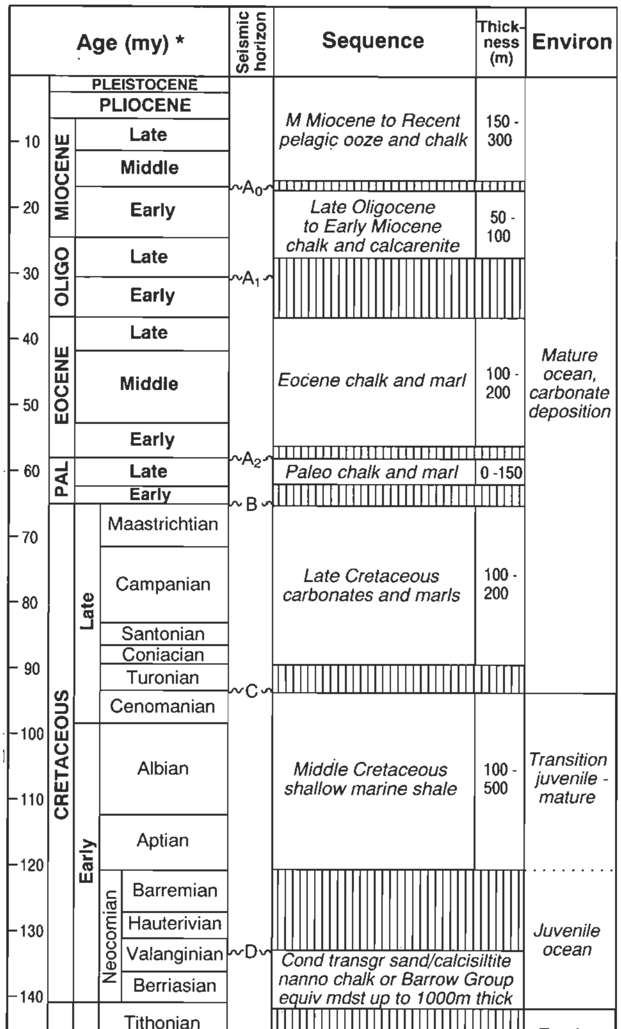
Table 2. Palaeontological Ages.

Sample§	*Age**	*Lithofacies#*	*Sample§*	*Age**	*Lithofacies#*
		EASTERN FLANK OF SWAN CANYON			
96:12/1	Oxfordian.-(Kimmer.) (5)	A2	**96**:14/8	Valanginian (1)	E5
12/3	Mesozoic (1)	E4	14/9	M. Camp.-L. Maast. (1)	E3
13/1	Mesozoic (1)	C1	14/10	Bajocian (5)	A2
13/2	Early Eocene (1,2)	E2	15/1	Valanginian (1)	E5
13/3	Valanginian (1)	E5	15/2	Oxford.-(Kimmer.) (5)	A2
13/4	Oxfordian-Kimmer. (5)	A2	15/3	L. Paleocene (1)	E2
13/5A,B	Hauterivian-E.Aptian.(1)	E5	15/5	L. Paleocene (1)	E2
13/7	M. Camp.-L. Maast. (1)	E3	15/6	Valanginian (1)	E5
13/8	L. Ryazan.-E. Barrem. (1)	C1	16/2	Bajocian (5)	A2
13/9	M. Eocene (1,2)	E2	17/5	Bajocian (5)	A2
14/4	M.-L. Campanian (1)	E3	17/9	Bajocian (5)	A2
14/5	Oxford.-(Kimmer.) (5)	A2	18/3	M. Miocene (1,2)	E1
		ROWLEY TERRACE			
95:1	L. Miocene (1)	E1	**95**:10/4A	Rhaetian (5)	A9/?D
2/2	Oxfordian (3)	H1	10/6D	M. Norian (7)	C11
4/1B	L. Campanian (1)	E3	10/8V	Norian-Rhaetian (4)	C15
4/1B	Cretaceous (3)	E3	10/10B	(L. Norian)-Rhaet. (4)	C8
4/1C	L.Barrem.-E.Aptian (1)	E5	11/3A	Aalen.-Bajocian (5)	A2
4/1E	Late E. Miocene (1)	E1	11/7A	Aalen.-Bajocian (5)	A2
4/4	L.Barrem.-E.Aptian (1,3)	E5	11/8	E.-M. Miocene (1,3)	E1
4/5	M. Miocene (1,3)	E1	12/6C	?M. Norian (7)	?
4/7	Callovian-Oxford. (3)	A3	12/9	?M.-L. Triassic (?)	C12
4/7	Bajoc.-Barrem. (1)	A3	12/9C	Middle Norian (7)	C12
4/16A	Aal.-Bajocian (5)	A2	12/9V	Norian-Rhaetian (4)	C12
4/p	L. Miocene (1)	E1	12/10A	Norian-Rhaetian (8)	C12
5/5A	Bajoc.-Bathon. (5)	A2	12/10D	M. Norian (7)	C12
5/9A	Bathonian (5)	A2	12/11	Norian-Rhaetian (9)	C17
6/1A	Toarcian (5)	A2	12/11V	Norian-Rhaetian (4)	C17
6/3	E.Toarc.(1) L. Toarc.(3)	A2	13/1	L. Paleocene (1)	E2
6/4	E. Toarcian(1)	A2	13/M1a	Paleocene (1)	E2
6/5	Rhaet.-Toarc. (3)	H6	13/M1b	M. Miocene (1)	E1
7/4A	Rhaetian (5)	A9/?D	13/P1	Late Paleocene (1)	E1
7/8	Bajo.-earliest Bath. (1,10)	A2	**96**:28/1	Rhaetian (7)	C12/13
7/11A	Rhaetian (5)	A9/?D	28/2	Rhaetian (7)	C11
7/13A-B	Late Triassic (6)	C9	28/6	Rhaetian (7)	C11
7/14A	Toarcian (5)	C2	28/11B	M. Jurassic (5)	A2
7/15B	Norian-Rhaetian (4)	C19	28/14	Rhaet.-earliest Jurass.(5)	A8
8/4	Early Jurassic (9)	C5	28/15	Bajocian (5)	A2
9/2C	Toarcian (5)	C2	28/16	E. Bajocian (1,2)	A2
9/6B	L. Norian (4)	C15			
		CENTRAL NORTHERN EXMOUTH PLATEAU			
96:29/1	Early M. Miocene (1,2)	E1	**96**:29/4	Early Miocene (1,2)	E1
29/2	M. Miocene (1,2)	E1	29/7	Mix Plio. & Mio. (1,2)	—
29/3	Early Miocene (1,2)	E1	30/1	Hettan.-lower Sine. (4)	C3
		WESTERN MARGIN OF WOMBAT PLATEAU			
96:32/2	M. Miocene (2)	E1	**96**:33/3	M. Miocene (1,2)	E1
33/2	Early M. Miocene (1,2)	E1	33/4	M. Miocene (1,2)	E1

§ Sample nomenclature XX:Y/Z where XX is the cruise number (95 or 96), Y is the dredge number, and Z is the rock type within the dredge haul.

* Ages from: 1: nannofossils (Shafik), 2:foraminifera (Wells), 3: foraminifera (Lynch), 4: foraminifera (Kristan-Tollmann), 5: palynomorphs (Burger), 6: brachiopods (Campbell), 7: conodonts (Nicoll), 8: corals (Stanley), 9: pelecypods (Grant-Mackie), 10: foraminfera (Apthorpe, pers. comm., 1993). Age details either in papers in this issue (Shafik, Burger, Campbell, Nicoll & Foster, Stanley, and Grant-Mackie; all 1994) or in the respective cruise reports — Exon et al. (1990) for Survey 95, and Colwell et al. (1990) for Survey 96.

See Table 1.



* *Tertiary timescale from Truswell et al. (1989); Mesozoic - from Gradstein et al. (in press)*

Figure 3. Generalised stratigraphy of the northern Exmouth Plateau (north of about 18°S, excluding the Wombat Plateau) and Rowley Terrace margin. Partly after Exon et al. (1982, 1992). Vertical hatching indicates absent or condensed section.

chalks and mudstones of probable Early Cretaceous age (E5); Middle–Upper Jurassic carbonaceous mudstones (A2); and Jurassic quartzose sandstones (A3 and A4) (Fig. 5). These rocks are similar to the sequences dredged by the *Sonne* and *Rig Seismic* from the canyon's western wall (von Rad et al., 1990). Although Upper Triassic shallow-water carbonates have been dredged from the nearby Cygnet Canyon (Kristan-Tollmann & Gramann, 1992), no Triassic rocks have been recovered from either the western or eastern walls of the Swan Canyon, probably due to their displacement by complex faulting.

Rowley Terrace margin

Dredge hauls 95/1 to 13 and 96/28 were taken in canyons on the margin of the Rowley Terrace (Figs 2 and 6). These canyons, named by Exon et al. (1990), incise deeply the continental slope and therefore present the potential opportunity to sample a large part of the margin's geology.

Dredges 1–5 were located on the northern part of the terrace in the major, Taipan/Copperhead Canyon system (Fig. 6). Dredges 1, 3, 4 and 5 were located on Shell 'Petrel' seismic line N206 in water depths of 4900–3330 m, whereas dredge 2 was sited just south of BMR profile 17/086 on the western slope of the northern arm of Taipan Canyon in water depths of 4000–3100 m. A wide range of rocks were recovered (Fig. 7A–C) including: Lower Cretaceous radiolarian chalks and mudstones (E5); ironstones and ferruginous sandstones and mudstones (B1–B3); Lower–Middle Jurassic carbonaceous mudstones (A2); quartzose sandstones (A3 and A4); and a variety

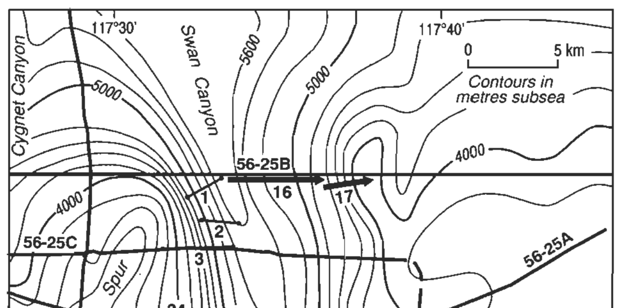


Figure 4. Detailed bathymetry of the Swan Canyon area showing the location of seismic lines, and Survey 96 and older dredge hauls.

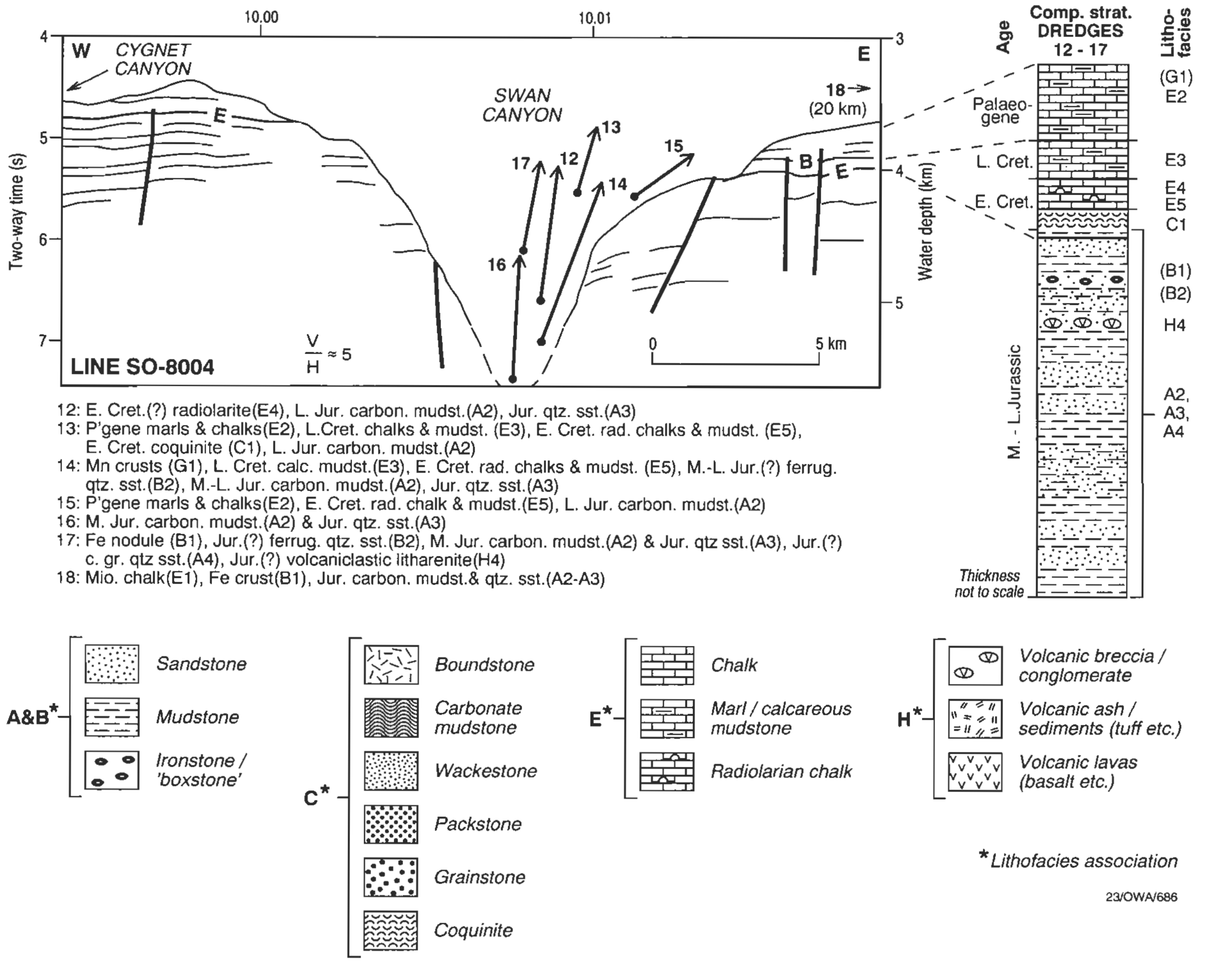


Figure 5. Line drawing of part of seismic line SO–8004 showing the location, results and composite stratigraphy of Survey 96 dredge hauls in the Swan Canyon area. Dredge hauls 12, 13, 16 and 17 projected onto the line (see Fig. 4 for relative locations).

of volcaniclastics (H3 and H4), basaltic breccias (H5), and basaltic and andesitic lavas (H8).

Dredge 6 was taken on the central part of the Rowley Terrace in 3400–3100 m of water, on Shell 'Petrel' line N206 and Survey 95 line 10 (Figs 6 and 7D). This large haul yielded ironstones and 'boxstones'; Lower–Middle Jurassic carbonaceous mudstones (A2), possibly including some originally evaporite components; lapillistone–hyaloclastite (H2); and an Upper Triassic–Lower Jurassic tuffaceous mudstone (H6). The bulk of the material is of volcanic origin.

Four dredge hauls (7, 8, 9 and 28) were recovered from the upper part of the Mermaid Canyon on BMR seismic lines 17/091 and 95/12 in water depths of 4530–3350 m (Fig. 6). These dredges extensively sampled the Upper Triassic–Middle Jurassic section (Fig. 7E). Rocks recovered include: Middle Jurassic carbonaceous mudstones (A2); quartzose sandstones (A3,A4 and A6); Lower Jurassic encrinite and skeletal wackestones (C4 and C5); Upper Triassic–lowermost Jurassic sub-arkosic sandstones and litharenites (A7 and A8); a calcite-cemented lapillistone/hyaloclastite (H2); and a wide variety of Upper Triassic shallow-water carbonates including reefal, shoal and lagoonal facies (C6–C16 and C19). Of particular importance is the discovery of the coral boundstone (lithofacies C19) in dredge 7 as this confirms the extension of Upper Triassic reefs from the Wombat Plateau to at least the central Rowley Terrace. The presence of encrusting colonial corals in a skeletal limestone (C12) in dredge 12 (Stanley, 1994 — this issue) also suggests the possibilty of reefs in the vicinity.

The southern-most dredge hauls, 10-13, were taken in a water depth of 5325–3100 m within the Clerke Canyon on two seismic lines, Shell 'Petrel' line N207 and Gulf 'Gulfrex' AU-26 (Fig. 6). Hauls 10, 11 and 13 were sited on the eastern flank of the canyon, and 12 on the western flank. The dredges recovered a varied suite including: ferruginous 'boxstones' and crusts (B1); Jurassic carbonaceous mudstones and quartzose sandstones with variable amounts of iron-oxide cementation (A2, A3 & A5); Upper Triassic algal boundstone, mudstones, skeletal wackestones, packstones and grainstones (C8, C11, C12 & C18); and volcaniclastic breccia and basaltic and andesitic lavas (H7 & H8). Dredges 10 and 12 recovered Triassic and Jurassic rocks, whereas the shallower 11 and 13 were confined totally to the Jurassic and younger section (Fig. 7F, G). The common occurrence of ferruginous 'boxstones' and iron-oxide cements indicates a significant period of subaerial exposure and weathering along this margin, probably in the Middle or Late Jurassic.

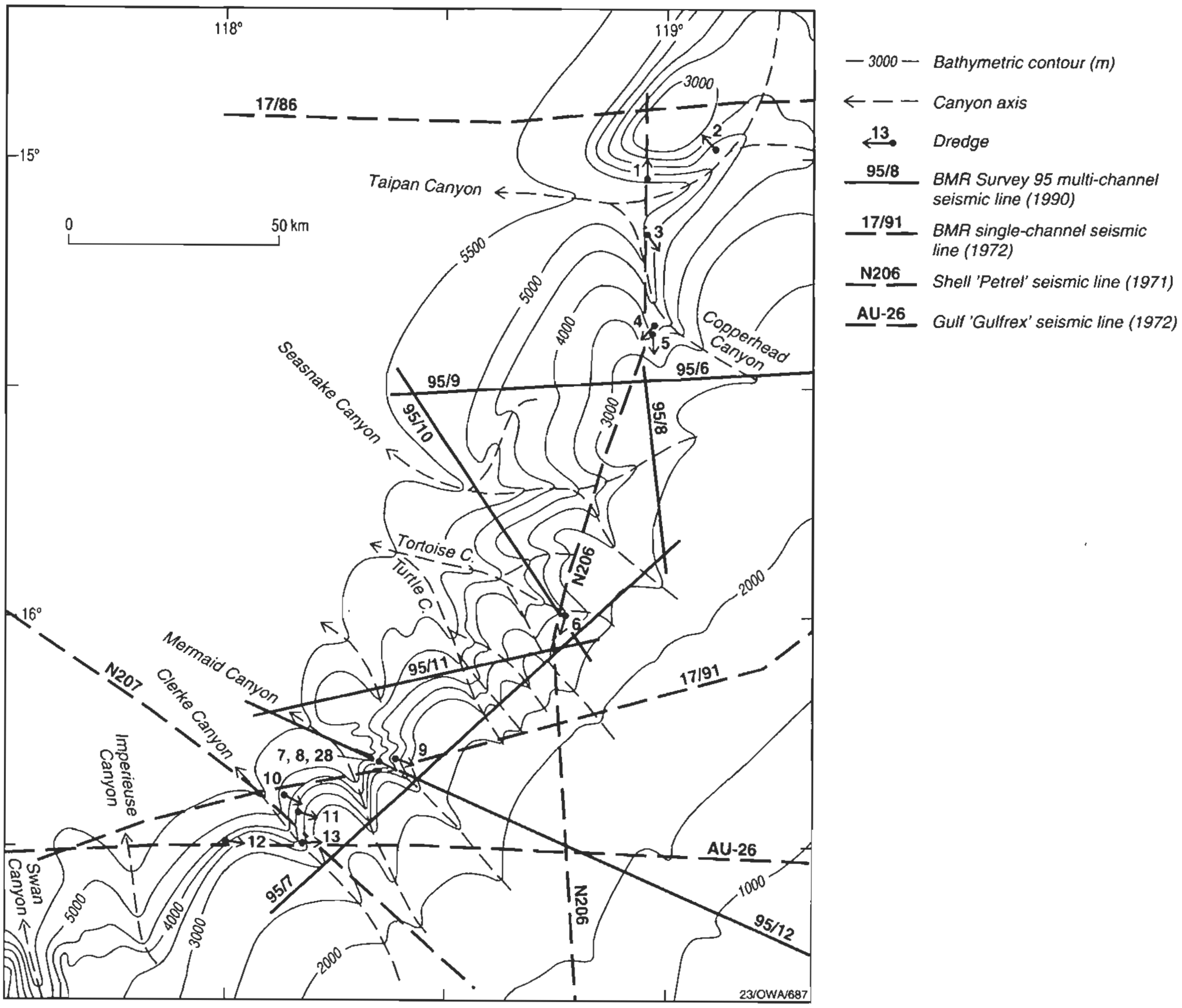


Figure 6. Location of Survey 95 (1–13) and Survey 96 (28) dredge hauls on the Rowley Terrace margin in relation to key regional seismic lines.

Central northern Exmouth Plateau

Two dredge hauls were attempted during Survey 96 to sample what had been interpreted on AGSO and industry seismic data, on the basis of regional correlations, to be Jurassic buildups (?reefs) developed on Triassic horst blocks (Exon, 1990; Fig. 8); however, recent detailed interpretations (Ramsay & Exon, 1994–this issue) show that the "buildups" are probably erosional remnants rather than depositional features. In one location, one of the "buildups" is exposed at the seafloor. This was a target for dredging during Survey 96. Two hauls (29 and 30) were run in a southerly direction from approximately 2300 to 2000 m of water (Fig. 8). The first recovered purely Neogene deposits (E1 and G1); the second (30) succeeded in sampling the target section. It recovered a hard, yellowish-orange, iron-oxide stained, bivalve-rich, Early Jurassic (Hettagian–lower Sinemurian) skeletal wacke- to packstone (C3) of tidal/shallow-marine bank origin.

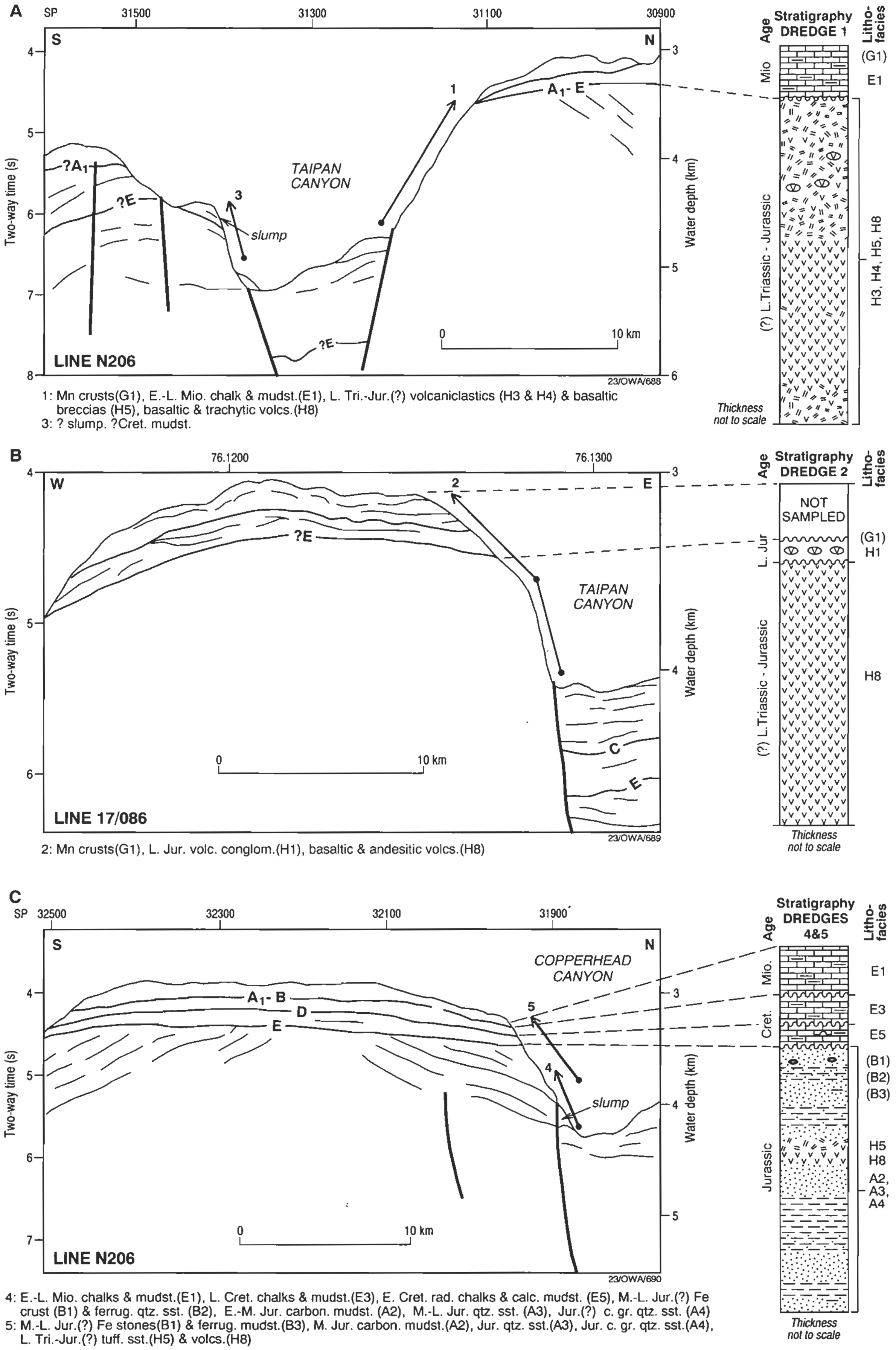
Western Wombat Plateau

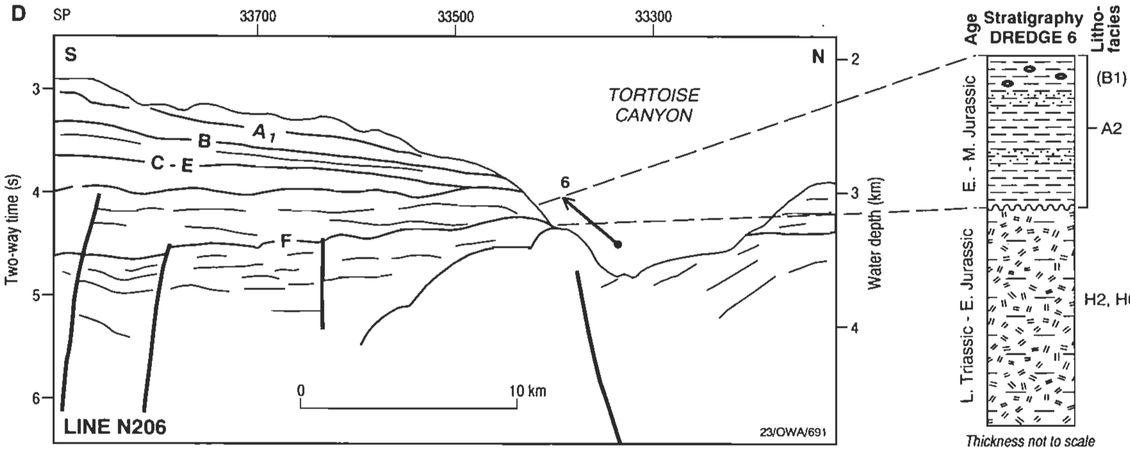
Regional seismic data (notably seismic line BMR 17/079) suggest that the Wombat Plateau's western margin is dominated by volcanic rock (von Rad et al., 1990). In an attempt to confirm this interpretation and obtain samples for petrographic and geochemical work, three dredge hauls (31–33) were attempted during Survey 96 on the western flank of the plateau in 3100–2310 m of water (Fig. 2). Only one (33) was successful. It recovered Middle Miocene chalks (E1) and a highly altered, moderately vesicular basalt (H8).

Lithofacies

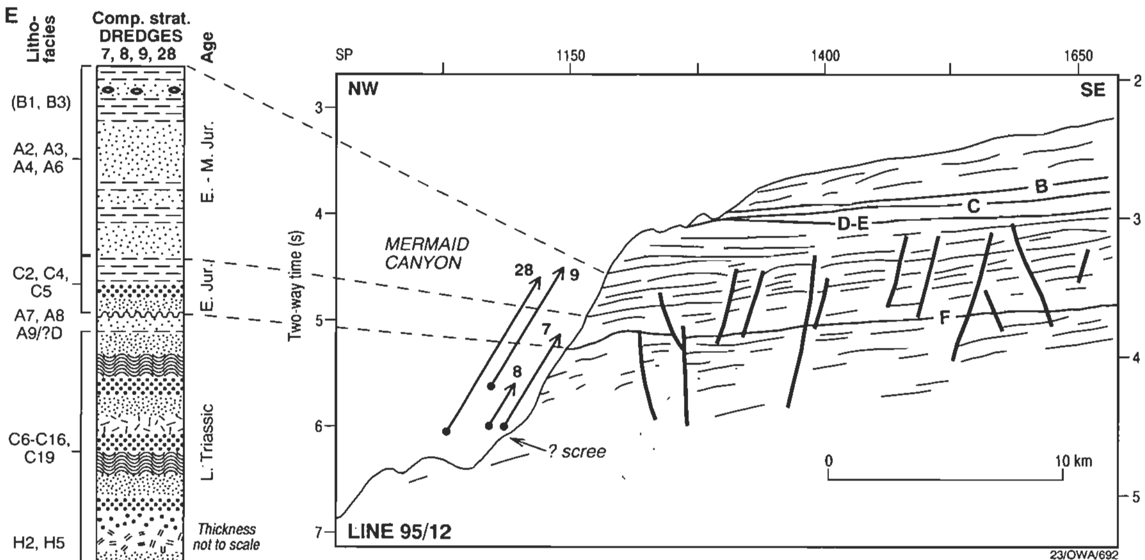
Adopting the approach used in the description of the *Sonne* SO–08 and AGSO Survey 56 dredge samples from the northern Exmouth Plateau (von Stackelberg et al., 1980; von Rad & Exon, 1983; von Rad et al., 1990), the rock types recovered during Surveys 95 and 96 can be grouped into a number of major lithofacies associations. These are:

- A. Siliciclastic association,
- B. Ferruginous association,
- C. Shallow-water carbonates,
- E. Hemipelagic to eupelagic calcareous mudstones and chalks, and
- H. Volcanic and volcaniclastic rocks.


A
SP
31500
31300
31100
30900
S
N
A1- E
?A1
?E
TAIPAN
CANYON
slump
?E
Two-way time (s)
Water depth (km)
0
10 km
LINE N206
23/OWA/688
1: Mn crusts(G1), E.-L. Mio. chalk & mudst.(E1), L. Tri.-Jur.(?) volcaniclastics (H3 & H4) & basaltic breccias (H5), basaltic & trachytic volcs.(H8)
3: ? slump. ?Cret. mudst.
Age
Stratigraphy
DREDGE 1
Litho-facies
Mio
(G1)
E1
(?) L.Triassic - Jurassic
H3, H4, H5, H8
Thickness
not to scale
B
76.1200
76.1300
W
E
?E
TAIPAN
CANYON
C
E
Two-way time (s)
Water depth (km)
0
10 km
LINE 17/086
23/OWA/689
2: Mn crusts(G1), L. Jur. volc. conglom.(H1), basaltic & andesitic volcs.(H8)
Age
Stratigraphy
DREDGE 2
Litho-facies
NOT
SAMPLED
L. Jur
(G1)
H1
(?) L.Triassic - Jurassic
H8
Thickness
not to scale
C
SP 32500
32300
32100
31900
S
N
COPPERHEAD
CANYON
A1- B
D
E
slump
Two-way time (s)
Water depth (km)
0
10 km
LINE N206
23/OWA/690
4: E.-L. Mio. chalks & mudst.(E1), L. Cret. chalks & mudst.(E3), E. Cret. rad. chalks & calc. mudst. (E5), M.-L. Jur.(?) Fe crust (B1) & ferrug. qtz. sst. (B2), E.-M. Jur. carbon. mudst. (A2), M.-L. Jur. qtz. sst. (A3), Jur.(?) c. gr. qtz. sst. (A4)
5: M.-L. Jur.(?) Fe stones(B1) & ferrug. mudst.(B3), M. Jur. carbon. mudst.(A2), Jur. qtz. sst.(A3), Jur. c. gr. qtz. sst.(A4), L. Tri.-Jur.(?) tuff. sst.(H5) & volcs.(H8)
Age
Stratigraphy
DREDGES
4&5
Litho-facies
Mio.
Cret.
Jurassic
E1
E3
E5
(B1)
(B2)
(B3)
H5
H8
A2,
A3,
A4
Thickness
not to scale



6: M.-L.Jur.(?) Fe stones & 'boxstones'(B1), E.-M. Jur. carbon. mudst.(A2), graded lapillist.-hyaloclastite(H2), L.Tri.-E. Jur. tuff. mudst.(H6)

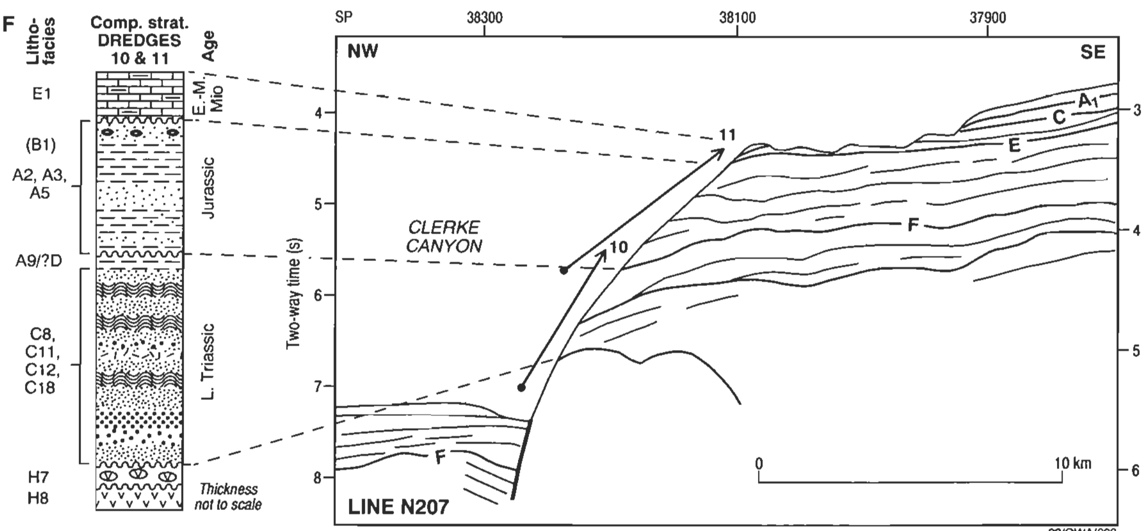


7: M.-L. Jur.(?) Fe 'boxstones'(B1), M. Jur. carbon. mudst.(A2), Jur. qtz. sst.(A3, A4 & A6), E. Jur. mudst. with calc. packst. layers(C2), E. Jur. 'encrinite'(C4), L.Tri.-Jur.(?) calc.-cement. sub-arkose(A7), L.Tri. carb. mudst.-sst.(A9/?D), L.Tri.(?) dol. lst. (C6), L. Tri. pel. wackest.(C9), L. Tri.(?) pel. grainst.-packst. (C14), L. Tri. coral boundst.(C19), L. Tri.-Jur.(?) calc.-cement. hyaloclastite(H2)

8: E. Cret.(?) silici. radio. mudst.(E4), M.-L. Jur.(?) Fe 'boxstones'(B1), E.-M. Jur.(?) carbon. mudst.(A2), E.Jur. skel. wackest.(C5), L.Tri.-Jur.(?) volc. sst.(H5)

9: M.-L. Jur.(?) Fe 'boxstones'(B1), M. Jur.(?) carbon. mudst.(A2), E. Jur. mudst. with calc. layers(C2), Jur. calc.-cement. qtz. sst.(A6), L. Tri.(?) calc. mudst.-wackest.(C8), L.Tri. skel. packst.(C15), L. Tri.(?) oolitic grainst.(C16)

28: M.-L. Jur.(?) ferrug. mudst.(B3), M. Jur. carb. mudst.(A2), L. Tri.-earliest Jur. sub-ark. sst./litharenite(A8), L. Tri.(?) sideritic lst.(C6), calcrete(C7), Fe-rich packst.(C10), L.Tri. crinoid-rich wacke-packst.(C11), L.Tri. skel. wackest.(C12), L.Tri. pel. wackest.-mudst.(C13), L.Tri.(?) skel. packst.(C15), L.Tri.-Jur.(?) graded lapillist.-hyaloclastite(H2)



10: M.-L. Jur. (?) Fe 'boxstones' & crusts (B1), Jur. qtz. sst. (A3), Jur.(?) orthq'tzite (A5), L.Tri. carbon. mudst. (A9/?D), L. Tri. mudst.-wackest. (C8), L. Tri. echin. wacke-packst.(C11), L. Tri. skel. wackest.(C12), L. Tri.(?) algal boundst.(C18), L.Tri.-Jur.(?) volc. sed. breccia(H7), basalt & trachyandesite(H8)

11: E.-M. Mio. calc. mudst. (E1), M.-L. Jur. (?) Fe 'boxstones' & crusts (B1), E.-M. Jur. carbon. mudst. (A2), L. Tri. (?) carbon. mudst.-v.f. sst. (A9/?D)

The siliciclastic association includes the 'coal measures' association first described in the *Sonne* SO–08 samples by von Stackelberg et al. (1980). Compared to the *Sonne* SO–08 and AGSO Survey 56 rock suites, our material contains a much greater diversity of Triassic–Lower Jurassic shallow-water carbonates (including a coral boundstone), a greater diversity of clastics in the 'A' association, and more types of volcaniclastics in the 'H' association.

Upper Triassic & Jurassic siliciclastic sequence (Lithofacies: A1 to A9)

Rocks of this association were dredged on the eastern side of the Swan Canyon and along the Rowley Terrace margin. They comprise a heterogeneous suite of mainly paralic, carbonaceous, mica-bearing, clayey siltstones to silty claystones, coal, and quartzose (30--50% quartz) and sub-arkosic (5–10% feldspar) sandstones, mostly with clayey matrix, but sometimes also with secondary calcite cement. Marine components occur in places, indicating marine incursions (see Shafik, 1994–this issue), and a general change to more-widespread marine conditions following the Mid Jurassic. Ages range from Late Triassic (Rhaetian, A9; Table 1) to Late Jurassic (Oxfordian–Kimmeridgian, A2).

The carbonaceous siltstones and claystones (facies A2 and A9) are commonly interbedded with very fine to medium-grained quartz sandstones (A3) which in some cases are cross-bedded. Thin coal (A1) stringers and laminated carbonaceous silty claystones were recovered from dredge hauls 95/5, 95/6, 95/7 and 95/12, on the Rowley Terrace margin (Fig. 6). They were probably deposited in a subaerial, deltaic/swamp environment.

The coarse to very coarse-grained quartz sandstones of facies A4 and A5 (Table 1) are mineralogically very mature, moderately to poorly sorted, and probably deposited in a fluviodeltaic environment. They contrast mineralogically with the sub-arkosic sandstones and litharenites of facies A6 and A7, which contain up to 10% feldspar and some rock fragments (mainly chert and altered volcanics). In some cases (e.g. sample 95:7/10), these latter rocks have trace amounts of glauconite, indicating deposition under (marginal) shallow-marine conditions.

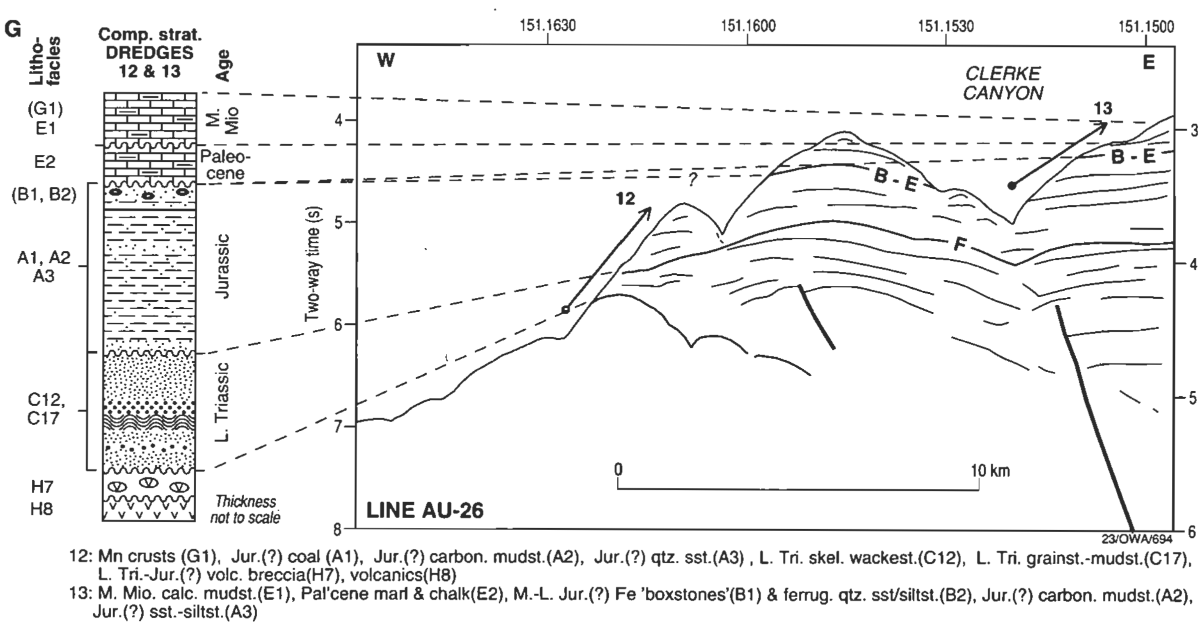
Secondary minerals include calcite (e.g. facies A6), pyrite (usually as nodules or concretions), and iron oxides. Late-stage impregnation by iron oxides under arid conditions transformed some of the clastic rocks of the 'A' association into ferruginous rocks of the 'B' association.

Ferruginous association (Lithofacies B1–B3)

This association is characterised by the presence of substantial quantities of secondary iron oxides. It has been recovered from the Rowley Terrace margin and Swan Canyon as three major lithofacies: ferruginous crusts, concretions and 'boxstones'; ferruginous quartz sandstones; and ferruginous mudstones (Table 1). As noted by von Stackelberg et al. (1980), these rocks appear to be subaerially weathered equivalents of the siliciclastic association.

Shallow-water carbonate association (Lithofacies C1–C19; Plates 1 to 3)

Rocks of this highly diverse association were recovered mainly from the Rowley Terrace canyons. They typically range in age from Late Triassic (Norian) to Early Jurassic (Toarcian). In general, they can be correlated with rocks described by von Stackelberg et al. (1980) and von Rad et al. (1990) from canyons on the northern Exmouth Plateau margin, and with microfacies described by Röhl et al. (1992; Table 3) from the Wombat Plateau ODP drill



12: Mn crusts (G1), Jur.(?) coal (A1), Jur.(?) carbon. mudst.(A2), Jur.(?) qtz. sst.(A3), L. Tri. skel. wackest.(C12), L. Tri. grainst.-mudst.(C17), L. Tri.-Jur.(?) volc. breccia(H7), volcanics(H8)
13: M. Mio. calc. mudst.(E1), Pal'cene marl & chalk(E2), M.-L. Jur.(?) Fe 'boxstones'(B1) & ferrug. qtz. sst/siltst.(B2), Jur.(?) carbon. mudst.(A2), Jur.(?) sst.-siltst.(A3)

Figure 7. *(above and two previous pages)* **Line drawings of Rowley Terrace seismic sections showing AGSO Survey 95 and 96 dredge locations, results and composite stratigraphic columns. (A): northern-most; (G): southern-most (see Fig. 6). Key to lithological symbols given on Figure 5. Note the abundance of volcanic rocks on the northern part of the Rowley Terrace (A & B), the thick Jurassic section overlying the regional 'F' top Triassic unconformity (C–G), and the widespread occurrence of Upper Triassic carbonates (E–G).**

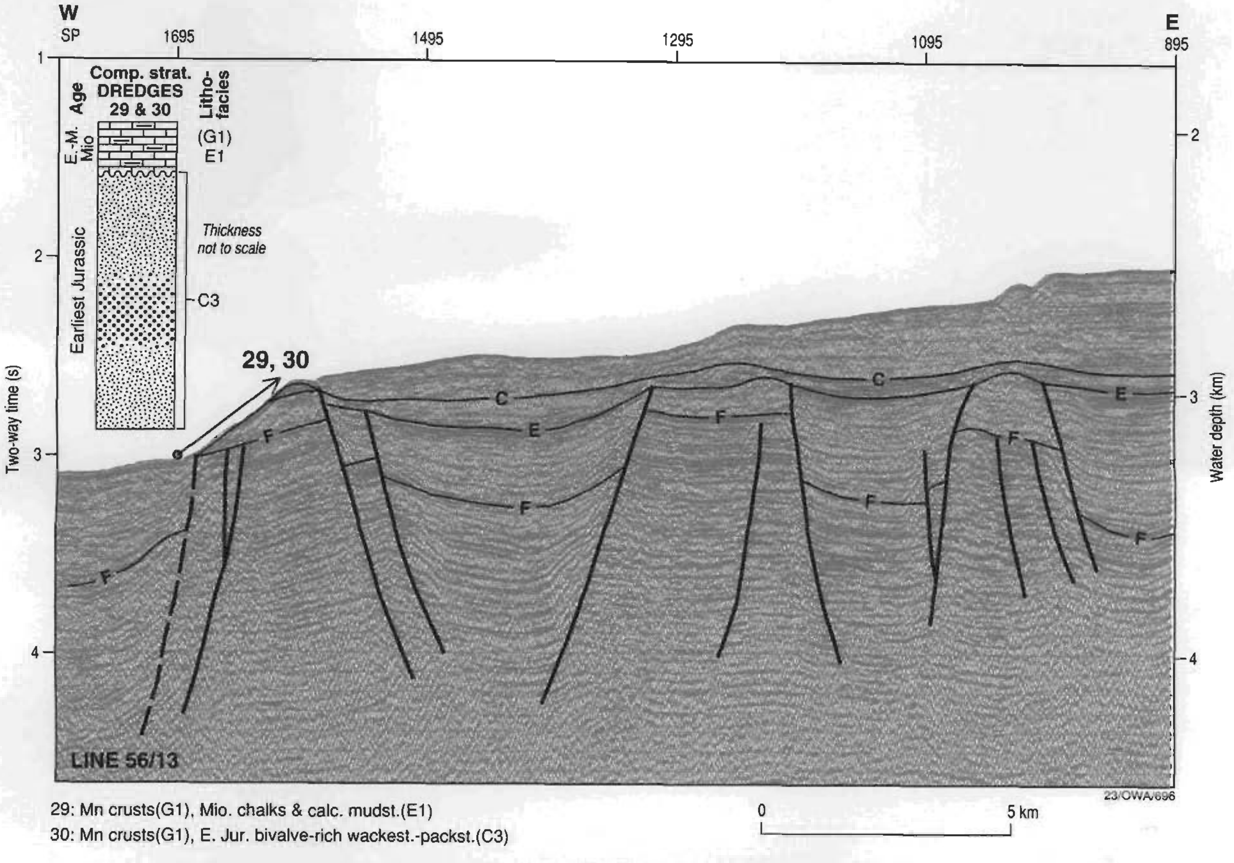


Figure 8. Part of seismic line 56/13 showing the location of dredge hauls 29 and 30 in relation to the Mesozoic section beneath the central northern Exmouth Plateau. Also indicated are the dredge results. Seismic interpretation from Ramsay & Exon (1994–this issue). Key to lithological symbols given on Figure 5.

holes. Shelf carbonates of similar age are known from exploration wells located on other parts of the Exmouth Plateau (e.g. Rhaetian marls in Sirius 1, Investigator 1, Vinck 1, and Eendracht 1; Norian limestone in Jupiter 1 and Mercury 1; and Hettangian limestone in Saturn 1; Barber, 1982, 1988; Williamson et al., 1989).

The Lower Cretaceous (late Berriasian–early Barremian) lithofacies C1 consists of a dusky yellow, iron-oxide stained, poorly sorted, coquinite (rudstone) composed almost entirely of *Inoceramus* plates. It was deposited under shallow-marine conditions immediately prior to the rapid, post-breakup subsidence of the margin in the mid- to late Early Cretaceous (von Rad & Bralower, 1992).

The great bulk of the 'C' association is made up of Late Triassic to Early Jurassic limestone types. According to the Dunham (1962) classification, we identified wackestones, packstones, grainstones, floatstones, and boundstones (Plates 1 to 3). Eighteen lithofacies, C2 to C19, were distinguished (Table 1). Peloidal wackestones with partly silicified shell debris and minor foraminifers (C9) and partly bioturbated algal boundstones with serpulids (C18) are interpreted as having been deposited under restricted lagoonal conditions. Skeletal wackestones to floatstones, and packstones with recrystallized mollusc debris, some foraminifers and ostracods (C8, C12, C13), are classified as lagoonal to shallow-marine facies. A range of shallow-marine conditions with distinct terrigenous input is represented by peloidal wackestones, peloidal pack- to grainstones with some echinoid fragments, and mud- to wackestones with crinoids, foraminifers, and mollusc fragments (C2 to C7, C11, C14, C15). Packstones–grainstones with ooids and some iron staining (C10, C6, C17) are probably shoal facies; reefs are documented by coral boundstones (C19).

In general, the Lower Jurassic rocks were laid down under more-open-water, shallow-marine conditions than the Upper Triassic section. Reefal deposits are restricted to lithofacies C19, which consists of a Norian–Rhaetian coral boundstone (Plate 1(2)). However, reef-derived material is present in at least one other facies, C12 (see Stanley, 1994–this issue).

Diagenetically, rocks of the 'C' association have commonly undergone dissolution of fossil fragments, micritization of shells, recrystallization, and silicification. Marine cementation mainly by low-magnesium calcite, occurs in different generations (early fringing cements, later blocky calcite cements; Plates 1 to 3). Echinoid fragments are commonly surrounded by syntaxial rim cement. Rocks of facies C6 have been extensively replaced by siderite or dolomite. Facies C7 is a calcrete. In facies C10, which is an iron-oxide rich packstone, original calcite ooids have been replaced by iron oxides.

Hemipelagic–eupelagic calcareous mudstones and chalk (Lithofacies E1–E5)

This association is characterised by hemipelagic calcareous mudstones, and eupelagic chalks and marls, ranging in age from Early Cretaceous (Late Berriasian) to Late

Table 3. Carbonate microfacies types, facies units and facies zones identified in ODP drillholes on the northern Exmouth Plateau (after Röhl et al., 1992).

Detritus-mud facies Foraminifers-detritus facies	1. Mudstone 2. Bioturbated wackestone 3. Foraminiferal wackestone a. *Triasina* predominant b. *Involutinidae* predominant c. *Duostominidae* predominant 4. Wackestones s.l. a. peloidal wackestone b. coated grain wackestone 5. Ostracod -rich wackestone 6. Echinodermal wackestone 7. Skeletal wackestone	(Estuarine) Lagoonal facies (Shelf)	
Foraminifers-calcareous algae and crinoid facies	8. Foraminiferal packstone a. *Involutinidae* predominant b. *Triasina* predominant 9. Peloidal packstone	Lagoonal	Transition facies
	10. Skeletal packstone (to grainstone-partly grapestone facies) 11. Echinodermal packstone 12. Codiacean wackestone to packstone	open-marine	
	13. Skeletal floatsone 14. Coral/sponge floatstone	Reefal debris	
Oncolitic-oolitic facies	15. Foraminiferal grainstone 16. Coated grain grainstone 17. Oolitic grainstone 18. Oncolitic grainstone 19. Dasycladacean grainstone 20. Codiacean grainstone 21. Peloidal grainstone 22. Skeletal grainstone s.l. to rudstone	Carbonate sand-shoal facies	
Biolithite facies	23. Algal bindstone 24. Boundstone a. coral- b. sponge- c. sponge/hydrozoan 25. Boundstones and framestone	Reefal facies	

Miocene. These rocks were recovered throughout the study area and form an open-marine pelagic 'cap' above the Jurassic shallow-marine and paralic deposits.

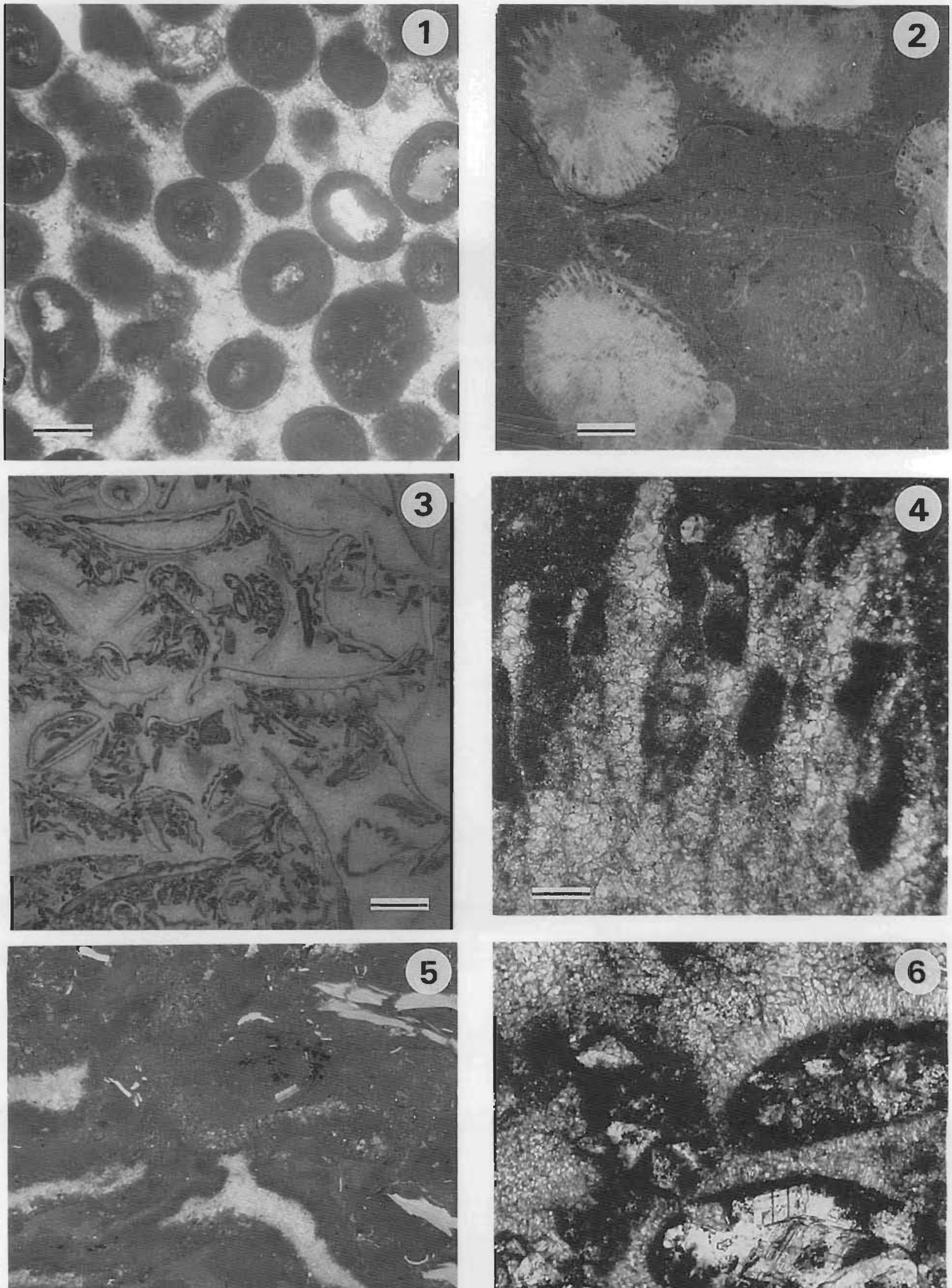
Five major lithofacies types can be recognised (Table 1). These range from soft, white, foraminiferal chalks of facies E1 to yellowish-grey to grey, hard radiolarites and radiolarian mudstones and chalks of facies E4 and E5. In general, the Cretaceous chalks contain more radiolaria and are more silicified (to porcellanites and/or quartz chert) than the Tertiary chalks. Radiolaria or radiolarian 'ghosts' (replacement by secondary silica) are particularly abundant in facies E4, which is interpreted to be a stratigraphic equivalent to the Windalia Radiolarite of the Carnarvon Basin (Condon et al., 1956).

Volcanic and volcaniclastic rocks (Lithofacies H1 to H8; Plate 4)

Volcanic and volcaniclastic rocks were dredged from the Rowley Terrace margin and the western flank of the Wombat Plateau. Eight major lithofacies types have been recognised (Table 1).

PLATE 1. NORIAN–RHAETIAN SHALLOW-WATER LIMESTONES

1. Oolitic grainstone showing two cement generations (early fringing, later blocky calcite cement), oomoldic porosity. Lithofacies C16, Sample 95:9/7V, Rowley Terrace margin, scale 30 µm.
2. Coral boundstone. Lithofacies C19, Norian–Rhaetian, Sample 95:7/15V, Rowley Terrace margin, scale 2.5 mm.
3. Bioclastic float- to rudstone, poorly sorted, large shells are recrystallised, ooids and smaller shell fragments are concentrated below the larger shells ("umbrella effect"). Lithofacies C12, Norian–Rhaetian, Sample 95:12/9V, Rowley Terrace margin, scale 7 mm.
4. Cross section of coral fragment, note recrystallisation of original structure. Lithofacies C19, Norian–Rhaetian, Sample 95:7/15A, Rowley Terrace margin, scale 300 µm.
5. Algal mat facies, calcrete horizons. Note irregular algal fabric. Lithofacies C18, Sample 95:10/6V, Rowley Terrace margin, scale 5 mm.
6. Dolomitized foraminifer (Aulotortus sp.?) and ?ooids in a bioclastic grainstone. Note relatively thick fringing cements and recrystallised secondary cement. Lithofacies C17, Norian–Rhaetian, Sample 95:12/9W, Rowley Terrace margin, scale 75 µm.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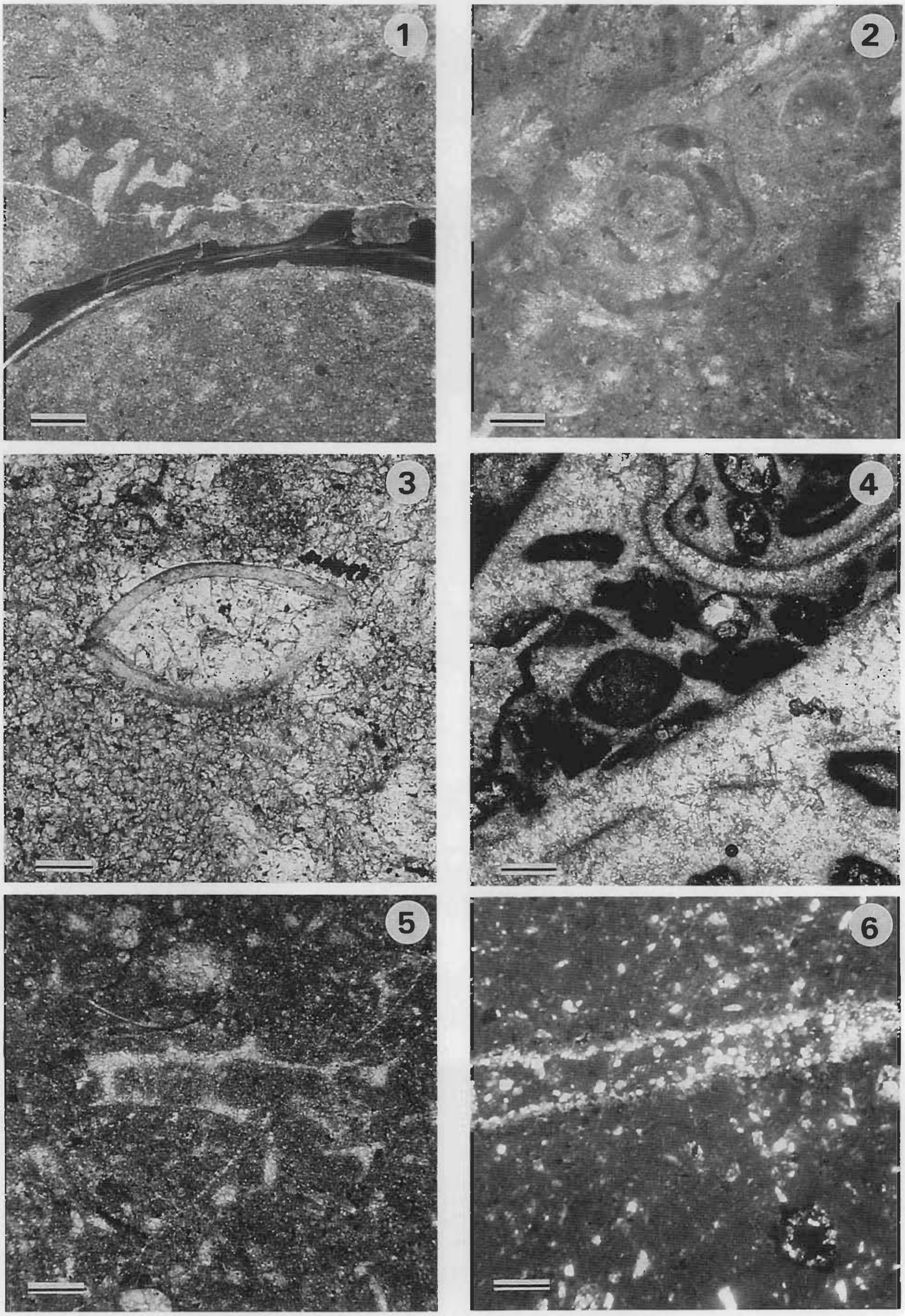
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Lithofacies H1 consists of highly altered, calcite-cemented volcaniclastic conglomerates. The rocks are very poorly sorted, with predominantly altered basaltic rock fragments, and include shallow-marine components such as echinoid, crinoid and mollusc fragments. A Late Jurassic (Oxfordian) age has been determined for one sample using foraminifera (Lynch *in* Exon et al., 1990). The rocks contain well-rounded pebbles in a fine to medium sand matrix, and were probably deposited under shallow-marine conditions on top of an eroded volcanic basement.

Hyaloclastites and lapillistones were dredged from the Rowley Terrace (lithofacies H2 and H3; Table 1). The main components were originally vesicular glass shards, but were later highly altered into iron oxides, smectites, chlorite and zeolites. In the case of H2, the graded volcaniclastic rock fagments are commonly cemented by secondary calcite. Submarine quenching of basaltic to andesitic lavas and thermal shock resulted in spalling of pillow rinds and further granulation into hyaloclastite glass sands. Hyaloclastite streams were probably deposited near active submarine volcanoes (seamounts) or moved down their steep slopes, to be redeposited and mixed with autochthonous shallow-water carbonates or siliciclastic–carbonate deposits.

Lithofacies H4 to H7 consist of a series of volcaniclastic breccias, sandstones and mudstones composed of mixtures of volcanic-derived and other terrigenous (quartz, feldspar, chert, and sedimentary) components. Many of the volcanogenic components have undergone extensive alteration to iron oxides, clays and/or zeolites. In the case of facies H7, 'basalt' fragments have undergone low-level, greenschist metamorphism.

A series of basaltic, trachytic and andesitic lavas constitute lithofacies H8. These volcanic rocks, of 'basement' along much of the margin, are described in detail by Crawford & von Rad (1994–this issue).

Stratigraphy

One of the principal objectives of the 1990-sampling program was to provide geological ground-truth to assist with the interpretation of seismic reflectors in the region. In general, this control has been provided, although in several cases some ambiguity exists because of the large depth range of the dredge hauls. In a number of cases (e.g. AGSO Survey 95 dredges 4 and 8), scree or slump material at the base of a slope was sampled rather than exposed rocks.

The dredge results, together with ties to ODP and commercial exploration wells, are used by Ramsay & Exon (1994–this issue) to identify and date the key seismic reflectors in the region and to characterise the intervening sequences. Overall, the following regional reflectors can be mapped; these are partly similar to those originally proposed for the Exmouth Plateau by Exon & Willcox (1980):

- A_0: Mid Miocene
- A_1: Mid Oligocene
- A_2: Base Eocene
- B: Cretaceous–Tertiary boundary
- C: Base Upper Cretaceous carbonates
- D: Valanginian unconformity
- E: Upper Jurassic (Oxfordian–Callovian) northern Exmouth Plateau — Argo Abyssal Plain 'breakup' unconformity
- F: Top Triassic

The correlation of the dredge results with the seismic horizons is shown in Figures 5, 7 and 8.

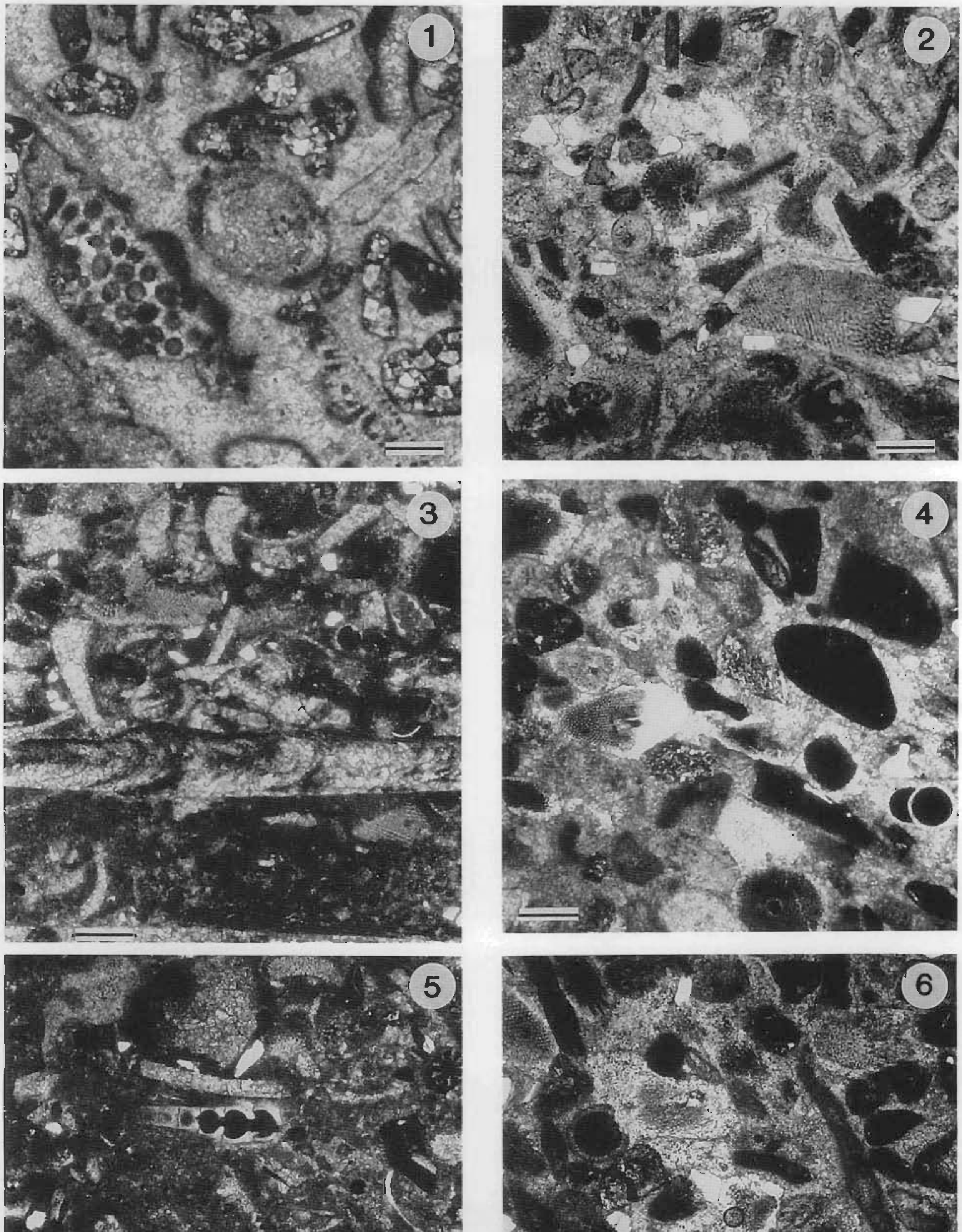
Except for the two dredge hauls taken on the central northern Exmouth Plateau and the three on the western flank of the Wombat Plateau, the dredged material comes from sequences which form the outer part of the Rowley Sub-basin, part of the offshore Canning Basin. This sub-basin lies mainly beneath the physiographic Rowley Terrace, is sparsely explored, and is known from exploration wells (Barcoo–1 and East Mermaid–1) to contain thick Jurassic, Cretaceous and Cainozoic sequences overlying a thick Triassic sequence (320+ m of Late Triassic in Barcoo–1; Woodside, 1980; Shell, 1973; Forrest & Horstman, 1986; Horstman & Purcell, 1988). The sub-basin extends down the eastern side of the north Exmouth Plateau's Swan Canyon, where it appears to be separated from the Dampier Sub-basin of the northern Carnarvon Basin by a series of down-to-the-west faults, the North Turtle Hinge (Forrest & Horstman, 1986). The sub-basin's stratigraphic succession is unnamed.

Our dredge results show that in the outer part of the Rowley Sub-basin:

(i) Volcanics (volcaniclastics and basaltic–trachytic lavas) are common along the margin, particularly in the north where they constitute much of the lower continental slope. The volcanics are undoubtedly associated with continental rifting and the subsequent formation of the Argo Abyssal Plain. Commonly they are overlain by the Callovian–Oxfordian (Middle–Late Jurassic) 'E' unconformity. Precise ages of the volcanics are unknown. However, a tuffaceous mudstone from dredge 6 on the central part of the Rowley Terrace margin has been dated using foraminifera as Rhaetian–Toarcian (Latest Triassic–Early Jurassic;

PLATE 2. NORIAN–RHAETIAN SHALLOW-WATER LIMESTONES

1. Sessile foraminifer (?*Alpinophragmium perforatum* FLÜGEL) on phosphatised shell fragment within coral boundstone. Lithofacies C19, Norian–Rhaetian, Sample 95:7/15V, Rowley Terrace margin, scale 300 µm.
2. Bioclastic wackestone with foraminifers (?*Aulotortus* sp.). Lithofacies C12, Norian–Rhaetian, Sample 95:10/8V, Rowley Terrrace margin, scale 300 µm.
3. Ostracod in a recrystallised, dolomitised limestone containing microbially induced framboidal pyrite. Note the different crystal size inside the ostracod to that outside. Lithofacies C19, Norian–Rhaetian, Sample 95:7/15B, Rowley Terrace margin, scale 75 µm.
4. Poorly sorted bioclastic rudstone containing big (cm-scale) shells, foraminifers and dolomitised ooids. Note that the outer parts of the shell fragments are micritised, the original shell structure is recrystallised to neomorph spar, and small components exhibit interpenetration due to early compaction. Lithofacies C17, Norian–Rhaetian, Sample 95:12/9W, Rowley Terrace margin, scale 300 µm.
5. ?*Problematikum 2* SENOWBARI-DARYAN (1980) in bioclastic wackestone with shell fragments and framboidal pyrite. Lithofacies C19, ?Rhaetian, Sample 95:7/15A, Rowley Terrace margin, scale 300 µm.
6. Silicified shell fragment in wacke- to mudstone of Late Triassic age. Lithofacies C9, sample 95:7/13A, Rowley Terrace margin, scale 300 µm.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Lynch *in* Exon et al., 1990). Also, a Late Triassic–? Middle Jurassic age is supported by the presence of volcanic components in some of the dated limestones of the 'C' lithofacies association (e.g. Facies C4), and in sub-arkosic sandstones and litharenites of facies A8 (Table 1). The only reliable K/Ar ages in the region are from rhyolites/trachytes on the northern Wombat Plateau dredged from near the 'early rift, "F" unconformity' on *Sonne* Cruise SO–08 which yielded ages ranging from 190–213 Ma (Latest Triassic–Early Jurassic; Kreuzer, *in* von Rad & Exon, 1983).

(ii) A thick (800+ m), Late Triassic, shallow-water carbonate-rich sequence dominates much of the lower continental slope, particularly on the southern part of the Terrace. Rocks present include reefal and peri-reefal deposits, for example Facies C19 and C18 of Table 1. Both of the existing exploration wells in the Rowley Sub-basin failed to significantly test the Upper Triassic section, bottoming in the case of East Mermaid–1 in Early Jurassic (Hettangian) shale, siltstone and sandstone, and in Barcoo–1 in Late Triassic (Norian–Rhaetian) claystones, sandstones, siltstones, recrystallised limestones, and minor acid intrusives (Shell, 1973; Woodside, 1980). This leaves open the question as to whether significant Triassic reefal limestones, which might be potential petroleum reservoirs, occur in the main part of the Rowley Sub-basin east of the dredged margin.

(iii) Along much of the outer Rowley Terrace, a thick Jurassic section overlies the regional 'F', top-Triassic unconformity (Fig. 7). This section is dominated by generally fluviodeltaic to paralic, interbedded quartzose sandstones and carbonaceous mudstones of mainly Middle Jurassic age. In places (dredges 95/7–9), the section includes, at its base, Lower Jurassic shallow-marine (shelf) carbonates. Iron-oxide staining and ferruginous crusts and nodules are common, indicating a period of subaerial exposure, probably in the Middle to Late Jurassic, possibly related to uplift associated with the Oxfordian–Callovian 'breakup' along this section of the margin. On the eastern side of the Swan Canyon, the Jurassic section is dominated by fluvial to paralic Mid–Upper Jurassic carbonaceous clayey siltstones and silty claystones, and quartzose sandstones. Although the Jurassic section in both areas is largely of fluvial to paralic origin, Shafik (1994–this issue) cites evidence in the nannofossil record of two major marine incursions, one in the early Toarcian and the other in the early Bajocian.

(iv) The Late Jurassic section is typically overlain above the Oxfordian–Callovian 'E' unconformity by Cretaceous radiolarian chalks, radiolarian porcellanites, and calcareous mudstones or, where the Cretaceous section is missing, by Tertiary chalks and marls (Fig. 7). In the case of the eastern wall of the Swan Canyon, a thin, Oxfordian–Kimmeridgian, shallow-marine mudstone immediately overlies the 'E' unconformity. This is in turn overlain by an Early Cretaceous, shallow-marine, *Inoceramus*-dominated coquinite passing upwards into deeper-water deposits (Fig. 6).

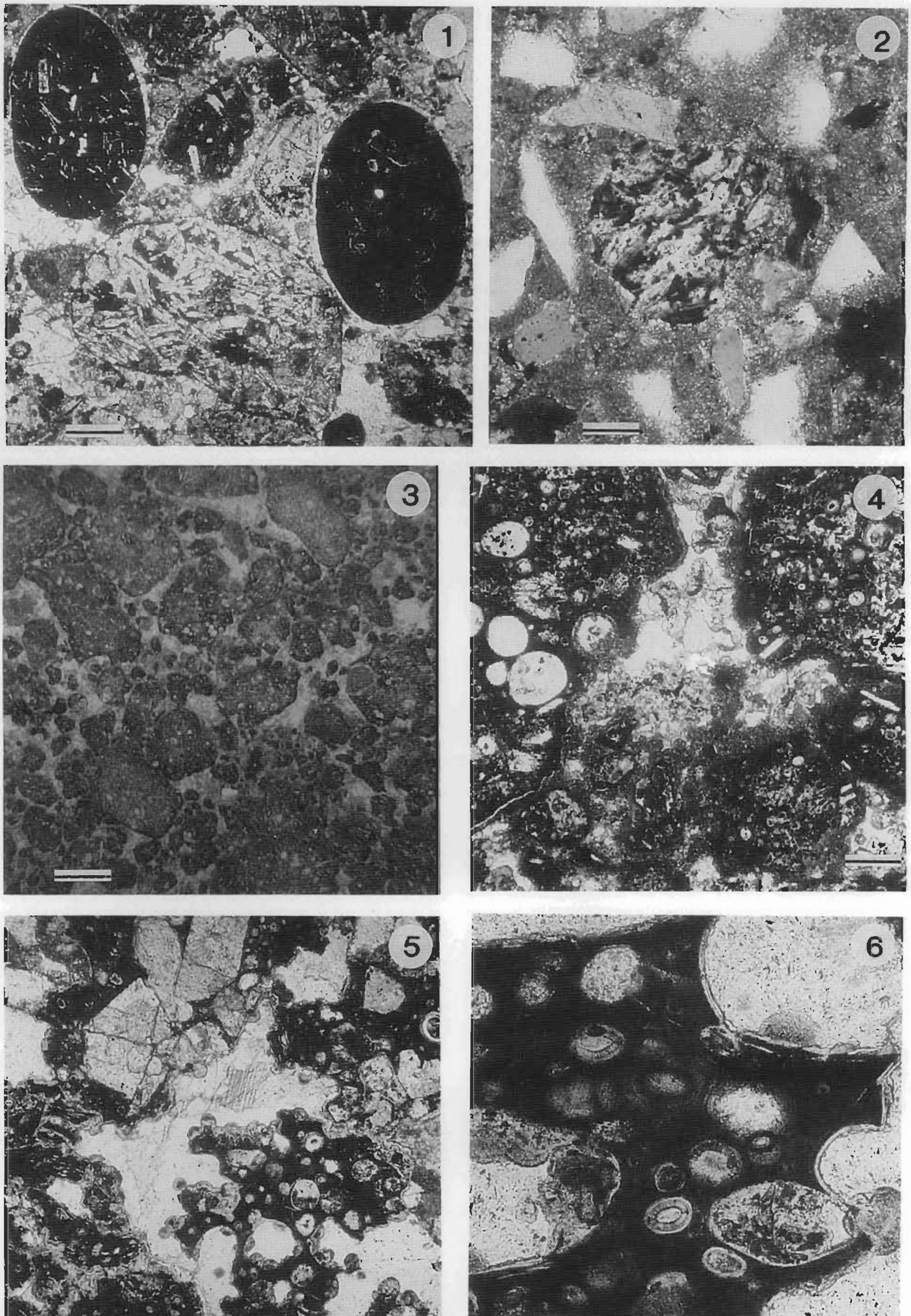
Discussion

During the Triassic to Jurassic the northwestern margin of Australia formed part of the southern shore of Tethys (Sengör, 1985; Audley-Charles, 1988; Görür & Sengör, 1992). One of the principal outcomes of the recent AGSO dredging and earlier work (BMR & BGR dredging, and ODP drilling) has been the information provided on the facies relationships which existed along the Tethyan margin during the Late Triassic and earliest Jurassic. This has wider implications because it has been known for some time (e.g. Kristan-Tollmann & Tollmann, 1981, 1982, 1983; Kristan-Tollmann, 1987, 1988a,b,c) that the macro- and microfossils of planktonic, benthonic and sessile biotopes can be correlated over the whole Tethys realm during the Triassic. For example, the ammonite-rich Hallstätter Kalk of Triassic age can be correlated from the Alps to the Himalayas and Timor, and the Rhaetian Dachsteinkalk/Oberrhaet/Riffkalk facies which is found in the Alps, can also be correlated with rocks on the Wombat Plateau, Timor, Papua-New Guinea and Japan (Pigram & Panggabean, 1984; Struckmeyer et al., 1990; Kristan-Tollmann, 1990, 1991).

In addition, recent studies on ODP cores have shown that two global Late Triassic (Rhaetian) sealevel events are documented by the litho- and biofacies and by the log character of ODP Sites 761 and 764 on the Wombat Plateau (Röhl et al., 1992; von Rad et al., 1992a,b; Kristan–Tollmann & Gramann, 1992): a sequence boundary at the Norian/Rhaetian boundary and a sequence boundary inside the upper part of the Rhaetian. These sealevel changes were probably caused by major plate-tectonic reorganisations. The Rhaetian facies of the Wombat Plateau shows a striking similarity to that of the western Tethys (Kristan-Tollmann & Gramann, 1992). As in the Wombat Plateau, the base of the Rhaetian in the western, northern and southern Alps is characterised by significant upward-decreasing terrigenous input which may be explained by global sealevel changes, i.e. a transgressive system tract (Dumont, 1992). In the Northern Calcareous Alps, the limestone and marly shale facies of the Kössen marls can be compared to the marl/limestone alternations within the 'lower Rhaetian' section of Site 761. Also, the northern Alpine 'Upper Rhaetian Reefal Limestone (Oberrhaetriffkalk) can be compared with the reefal facies in Site 764. In addition, at the Triassic/Jurassic boundary, the European margin of the Ligurian Tethys exposed in the western Alps of France shows strong facies similarities with the

PLATE 3. UPPER TRIASSIC–LOWER LIASSIC SHALLOW-WATER LIMESTONES

1. Bioclastic rudstone with dolomitised foraminifers, calacareous algae fragments, and micritized shell pieces. Pore space between components almost completely infilled with cement. Lithofacies C17, Norian–Rhaetian, Sample 95:12/9W, southern Rowley Terrace margin, scale 300 μm.
2. Echinodermal packstone. Note quartz grains and high proportion of ecinoderm fragments with syntaxial cements. Lithofacies C11, Norian–Rhaetian, Sample 96:28/6, Rowley Terrace margin, scale 300 μm.
3. Skeletal packstone with pelecypod fragments (partly with original internal structure), foraminifers, echinoderm fragments, and quartz grains. Lithofacies C3, Hettangian–lower Sinemurian, Sample 96:30/1, central northern Exmouth Plateau, scale 300 μm.
4. Skeletal packstone with partly dissolved echinoderm fragments, foraminifers, and minor volcaniclastics. Lithofacies C15, late Norian, Sample 95:9/6B, Rowley Terrace margin, scale 300 μm.
5. As for 3, but including crinoid fragments. Scale 300 μm.
6. As for 4, but with syntaxial cements surrounding echinoderm fragments. Scale 300 μm.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east Gondwanan margin facies (Dumont & Röhl, 1992).

The AGSO Surveys 95 and 96 dredging has shown that the Upper Triassic shallow-water carbonate lithofacies types described by Röhl et al. (1992) from ODP holes on the Wombat Plateau, can be recognised elsewhere on the margin. These rocks include carbonates of restricted lagoonal (algal boundstones, peloidal wackestone), of open-lagoonal to shallow-marine facies (skeletal pack-, float-, mudstone), of oolitic shoal facies (oolitic grainstone, grain- to mudstone), and of reefal facies (coral boundstone — see Stanley, 1994–this issue), as well as mixed shallow-marine carbonate–siliciclastic rocks.

Our dredging shows that a variety of Jurassic sediments is present in the Swan Canyon and Rowley Terrace areas. However, ODP drilling has indicated that on the Wombat Plateau horst, Upper Triassic rocks are directly overlain by Cretaceous strata. This is despite the fact that Quilty (1981) reported a Sinemurian (Early Jurassic) age for an encrinitic biomicrite from *Sonne* dredge SO8-61KD on the northern slope of the plateau very close to ODP Site 764 (see Haq et al., 1990, p. 368–369). This apparent contradiction has recently been resolved by Kristan-Tollmann & Gramann (1992) using other microfossils and slightly extended foraminiferal ranges, who have re-investigated the *Sonne* material and suggested that it is Late Triassic (Rhaetian) in age, as well as noting its similarity to European faunas. Kristan-Tollmann & Colwell (1992), however, state that the presence of *Ptychobairdia hettangica* (DONZE) in sample SO8–62KD could also be interpreted as including an earliest Jurassic age, because this ostracod has been described from the Hettangian of France.

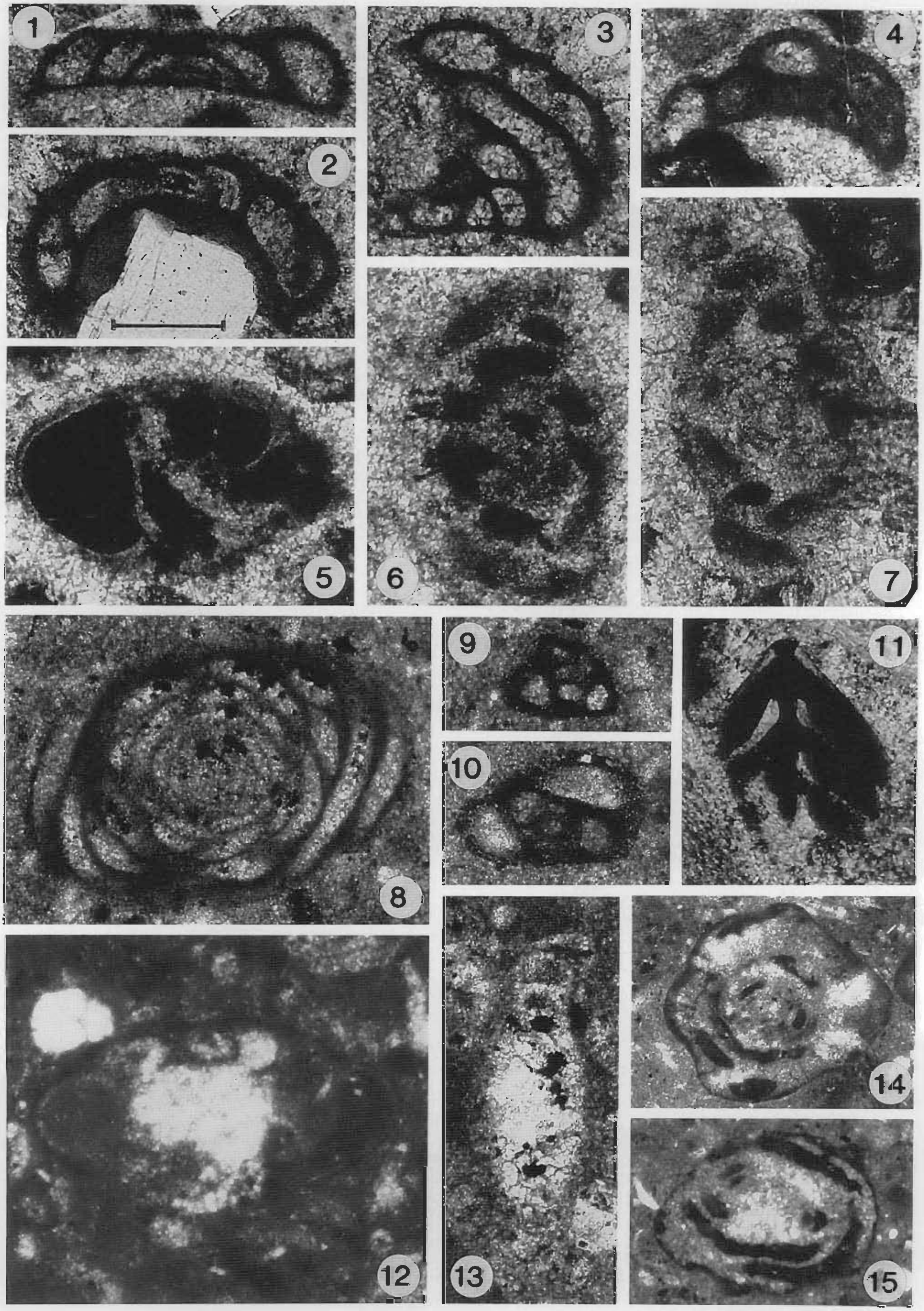
The skeletal (mollusc–echinoderm–foraminifer) wacke- to packstone recovered from the central northern Exmouth Plateau (sample 96:30/1, Facies C3) is an exciting discovery because of its early Liassic age and its great similarity to rocks in the Northern Calcareous Alps and on other parts of the Tethyan margin, including Timor (Kristan-Tollmann & Colwell, 1992). The rock is very similar to the Enzesfelder Limestone of Austria, a well-bedded carbonate mudstone locally rich in ammonites and bivalves. This formation, which has been dated by ammonites as early Hettangian to early Sinemurian, contains a very typical foraminiferal fauna of early Jurassic age, notably *Involutina liassica* JONES, 1853, and *Involutina turgida* KRISTAN, 1957, as well as *Trocholina granosa* FRENTZEN, 1941 and *Trocholina intermedia* FRENTZEN, 1941. All of these taxa also occur in the Rhaetian (e.g. in the Zlambach Marls of Austria), but the abundance of *Involutina,* especially of *Involutina turgida,* is an important characteristic of the lowermost Liassic Alpine carbonates, such as the Adneth and Enzesfelder Limestones. The great similarity in foraminiferal microfaunas and microfacies between the widely separated areas (Northern Alps and northern Exmouth Plateau) strongly indicates, yet again, a conspicuously similar facies evolution in the Tethys during Late Triassic/Early Jurassic times over distances of up to 15 000 km.

Acknowledgements

We wish to acknowledge the valuable contribution made by the scientists, technicians and crew who participated on *Rig Seismic* Surveys 95 and 96. N. Exon, M. Apthorpe, and anonymous referees are thanked for their comments on the manuscript. Figures were drafted by M. Huber of the AGSO Drawing Office.

PLATE 4. VOLCANICLASTIC ROCKS

1. Volcaniclastic breccia with rounded pebbles of altered basalt, palagonite and quartz cemented by calcite. Lithofacies H3, sample 95:1/2A, Rowley Terrace margin, scale 740 μm.
2. Volcanic grain in a well-sorted, quartz-rich greywacke. Lithofacies A3, Sample 95:4/7A, Rowley Terrace margin, scale 75 μm.
3. Greenish grey, calcite-cemented tuff to lapillistone. Lithofacies H2, Sample 95:6/6A, Rowley Terrace margin, scale 300 μm.
4. Altered, vesicular, volcanic grains in lapillistone. Contains micrite, smectite and zeolites. Lithofacies H2, sample 95:6/6A, Rowley Terrace margin, scale 300 μm.
5. Hyaloclastite with vesicular glass shards and calcite cement. Lithofacies H2, Sample 95:7/10A, Rowley Terrace margin, scale 300 μm.
6. Vesicular, altered, glass shard in quartz-rich, echinoderm grainstone. Lithofacies C4, Sample 95:7/12A, Rowley Terrace margin, scale 75 μm.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Appendix: Tethyan foraminiferal faunas

E. Kristan-Tollmann

(A) Upper Triassic and Lower Jurrassic faunas

Sample 95:7/15V (Rowley Terrace)

Thin sections from this sample contain only the sessile foraminifer *Alpinophragmium perforatum* FLÜGEL [Plate 6 (5)]. *A. perforatum* is an index fossil for the biolithite facies of the central reef. This species is now well known from throughout the Tethys realm, extending from the Norian–Rhaetian reefal areas of the Northern Calcareous Alps in western Tethys to the Dachstein reefal limestone on Kyushu Island in Japan at its eastern end (Kristan-Tollmann, 1991).

Sample 95:9/6B (Rowley Terrace)

The following foraminifera were recognised in this sample:

Ammovertella polygyra KRISTAN-TOLLMANN
Glomospirella friedli KRISTAN-TOLLMANN [Plate 5 (6,7)]
Verneuilinoides mauritii (TERQUEM) [Plate 7 (3)]
Planiinvoluta carinata LEISCHNER [Plate 5 (1)]
Planiinvoluta deflexa LEISCHNER [Plate 5 (2,4)]
Planiinvoluta multitabulata KRISTAN-TOLLMANN [Plate 5 (3)]
Lenticulina sp.
Frondicularia rhaetica KRISTAN-TOLLMANN [Plate 5 (11)]
Auloconus permodiscoides (OBERHAUSER) [Plate 7 (4)]
Variostoma helictum (TAPPAN) [Plate 5 (5)]

This comparatively plentiful fauna is a compound association. Whilst *Glomospirella friedli* is typical of the backreef environment, especially of the lagoon, the sessile taxa *Planiinvoluta* and *Verneuilinoides mauritii* lived during the Late Triassic mainly in the reef and sill regions. All other species are typical inhabitants of the fore-reef. This suggests an environment near a patch reef close to both the basin and lagoon.

Variostoma helictum is very rare in the sample, being present in only one thin section [Plate 5 (5)]. It is typical of a sill setting in an open basin. On the basis of this species, the sample can be dated as Sevatian (Late Norian). The species is present throughout the Tethys realm including the Northern Calcareous Alps, Austria; Kotel, Bulgaria; Waliabad, Iran; Sichuan and Yunnan, China; Timor, Indonesia; and northeast Alaska, USA (Kristan-Tollmann, 1988a, p. 249).

Sample 95:10/8V (Rowley Terrace)

This sample contains the following foraminifera:

Glomospira sp.
Glomospirella hoae KRISTAN-TOLLMANN [Plate 5(8)]
Agathammina austroalpina KRISTAN-TOLLMANN & TOLLMANN [Plate 5(9,10)]
Angulodiscus tumidus KRISTAN-TOLLMANN [Plate5(13)]
Aulotortus praegaschei (ZANINETTI) [Plate 5(14,15)]
Diplotremina placklesiana KRISTAN-TOLLMANN [Plate 5(12)]

All of these species are known from many sites within the Tethys realm. Together, they indicate a Norian to Rhaetian age.

Until now, *Diplotremina placklesiana* was known from both the western and middle part of Tethys, from the Northern Calcareous Alps in Austria and from Esfahan, Iran, and also from the east of the Panthalassa, from Oregon and Nevada, USA (see Kristan-Tollmann, 1988a, text-fig. 2). It is confirmed here for the first time from the southeastern part of Tethys.

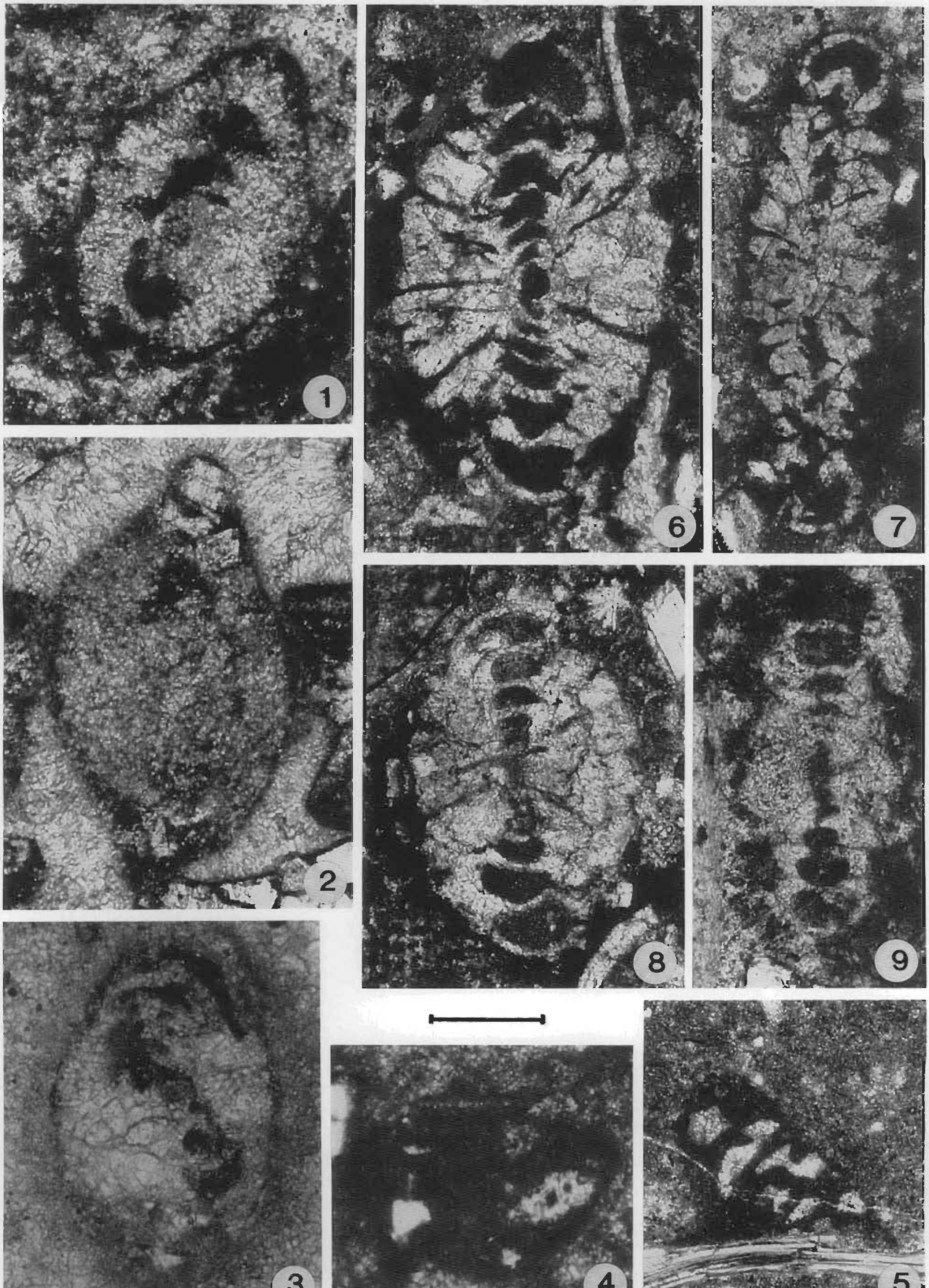
Sample 95:10/10B (Rowley Terrace)

This sample contains the following foraminifera:

Angulodiscus communis KRISTAN
Aulotortus sinuosus WEYNSCHENK [Plate 6(3)]
Triasina oberhauseri ?
Variostoma cochlea KRISTAN-TOLLMANN [Plate 6(4)]

Although there is only one typical section of *V. cochlea*

PLATE 5. FORAMINIFERA FROM UPPER TRIASSIC LIMESTONES, ROWLEY TERRACE
1–7, 11: Dredge sample 95:9/6B; age: late Norian (Sevatian).
8–10, 12–15: Dredge sample 95:10/8V; age: Norian–Rhaetian.
Scale bar indicates 600 µm for 14 and 15, 200 µm for all others.
1. *Planiinvoluta carinata* LEISCHNER, 1961.
2,4. *Planiinvoluta deflexa* LEISCHNER, 1961.
3. *Planiinvoluta multitabulata* KRISTAN-TOLLMANN, 1990.
5. *Variostoma helictum* (TAPPAN, 1951).
6,7. *Glomospirella friedli* KRISTAN-TOLLMANN, 1962.
8. *Glomospirella hoae* KRISTAN-TOLLMANN, 1970.
9,10. *Agathammina austroalpina* KRISTAN-TOLLMANN & TOLLMANN, 1964.
11. *Frondicularia rhaetica* KRISTAN-TOLLMANN, 1964.
12. *Diplotremina placklesiana* KRISTAN-TOLLMANN, 1960.
13. *Angulodiscus tumidus* KRISTAN-TOLLMANN, 1964.
14,15. *Aulotortus praegaschei* (ZANINETTI, 1976).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and it is a juvenile, this marker fossil places the sample in the Rhaetian.

Sample 95:12/9V (Rowley Terrace)

This sample contains *Angulodiscus communis* KRISTAN [Plate 6(2)] and is therefore of Norian–Rhaetian age.

Sample 95:12/11V (Rowley Terrace)

Angulodiscus communis KRISTAN [Plate 6(1) and Plate 7(5–10)] is relatively abundant in this sample. It is mostly poorly preserved, recrystallized and micritized. This species is present in the Norian and Rhaetian throughout Tethys, and is typical of a lagoonal environment.

Sample 96:30/1 (NE Exmouth Plateau)

The sample contain the following foraminifera:

Involutina liassica (JONES, 1853)
Involutina turgida KRISTAN, 1957
Trocholina granosa FRENTZEN, 1941
Trocholina intermedia FRENTZEN, 1941

This assemblage, together with the microfacies (foraminifera–echinoderm–mollusc-rich microfauna in a micritic limestone) and the lithofacies (yellow-ochre, iron-oxide stained, hard limestone), corresponds in all details with the Enzesfelder Limestone (lower Liassic: Hettangian–lower Sinemurian), a very characteristic formation, known from its type locality and other sites in the Northern Calcareous Alps as well as from Timor (for more details see Kristan-Tollmann & Colwell, 1992).

(B) Discussion of some Upper Triassic Foraminifera

Only quotations which are relevant for the stratigraphy, palaeogeography and palaeoecology are cited.

***Verneuilinoides mauritii* (TERQUEM, 1866)**

Plate 7 (3)

1866 *Verneuilina mauritii*, Terq. — TERQUEM, p. 448, pl. 18, fig. 18.

1936 *Verneuilina mauritii*, Terq. — FRANKE, p. 126, pl. 12, figs. 22, 23.

1962 *Verneuilinoides mauritii* (Terq.) — TRIFONOVA, p. 153, pl. 3, figs. 6,7.

1988 *Verneuilinoides mauritii* (Terquem, 1866) — KRISTAN-TOLLMANN, p. 248, fig. 1, Nos. 18-22.

1990 *Verneuilinoides mauritii* (TERQUEM, 1866) — KRISTAN-TOLLMANN, text fig. 8, no. 17; pl. 6, fig. 5; pl. 7, fig. 1.

Remarks: *V. mauritii* was first described from the Liassic of France and Germany. Trifonova (1962) described the species from limestones (Hallstätter Kalk?) of Carnian and Norian age from Kotel in the Eastern Balkan Mountains in Bulgaria. Since then, the species has been recorded from Upper Triassic limestone throughout the Tethys, e.g. from the Upper Carnian Hallstatt Limestone from the Bihati Brook near Baun in Timor, from the Norian Hallstatt Limestone of Berchtesgaden in the Northern Calcareous Alps, Bavaria (Kristan-Tollmann, 1988a), as well as from the Rhaetian reef limestone of Gurumugl in Central Papua New Guinea (Kristan-Tollmann, 1988a, 1990).

***Planiinvoluta multitabulata* KRISTAN-TOLLMANN, 1990**

Plate 5 (3)

1983 *Planiinvoluta deflexa* LEISCHNER, 1961 — SALAJ, BORZA & SAMUEL, p. 105 f, pl. 62, pars: fig. 4.

1990 *Planiinvoluta multitabulata* n. sp. — KRISTAN-TOLLMANN, p. 232, text fig. 11, no. 4; pl. 4, figs. 3–6.

Remarks: *P. multitabulata* has the following characteristics: the planispiral, coiled, initial part is made up of five to six slowly increasing whorls; the last uncoiled part of the tube grows over one, two, or rarely three whorls. The outer rim of the tube is angular on the base.

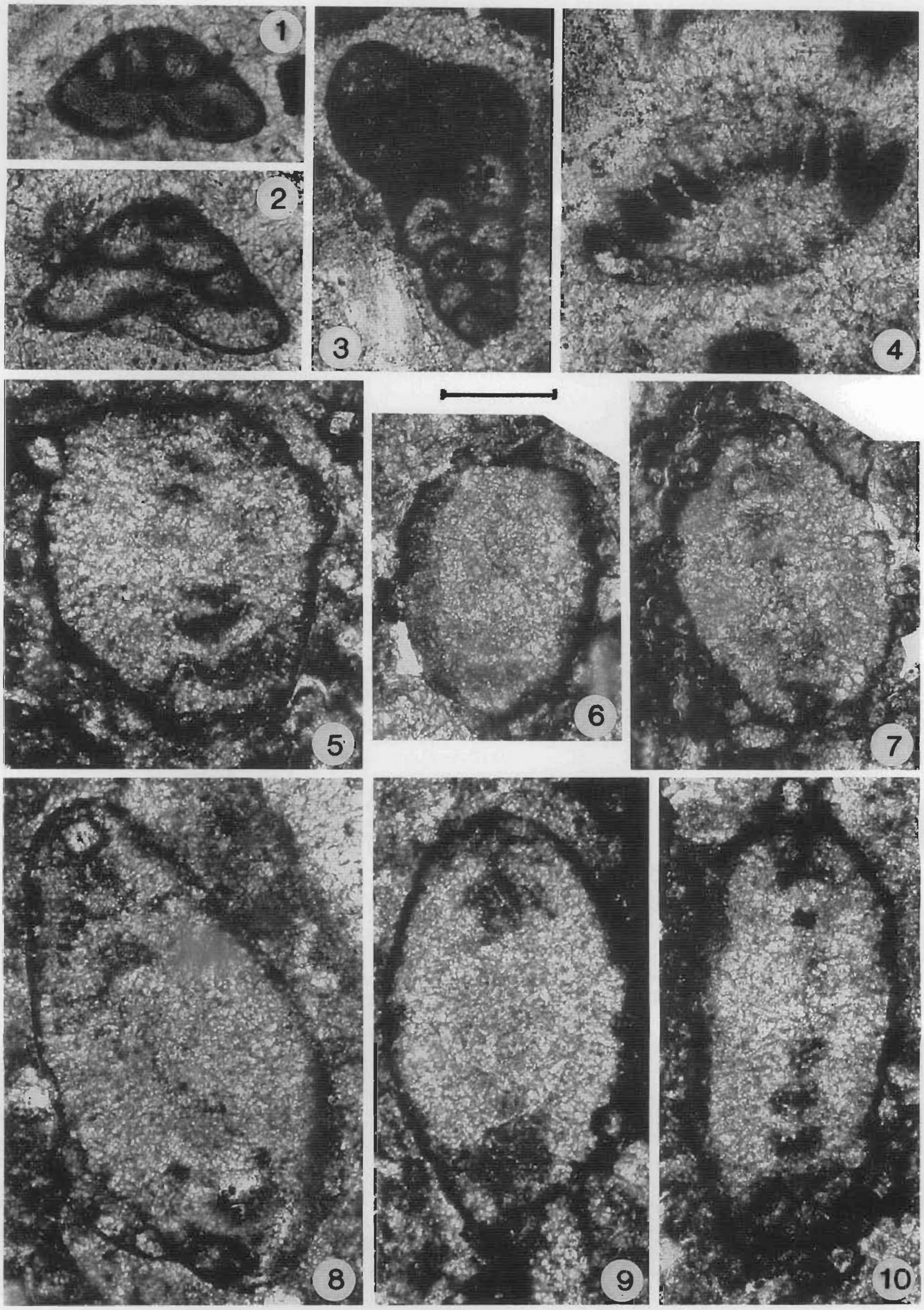
This species has been recorded from the Rhaetian of both the western part of the Tethys (Western Carpathian Mountains, Slovakia, Europe), as well as the eastern part (Gurumugl reef, Centrai Highlands of Papua New Guinea). The species is very rare in dredge sample 95:9/6B, an Upper Norian (Sevatian) limestone, from the Rowley Terrace.

***Frondicularia rhaetica* KRISTAN-TOLLMANN, 1964**

Plate 5 (11)

1964 *Frondicularia rhaetica* n. sp. — KRISTAN-TOLLMANN, p. 146f, pl. 32, figs. 1–8.

PLATE 6. FORAMINIFERA FROM UPPER TRIASSIC TO LOWERMOST LIASSIC LIMESTONES, ROWLEY TERRACE
Scale bar indicates 600 μm for 5, 200 μm for all others.
1. *Angulodiscus communis* KRISTAN, 1957.
Dredge sample 95:12/11V; Norian–Rhaetian.
2. *Angulodiscus communis* KRISTAN, 1957.
Dredge sample 95:12/9V; Norian–Rhaetian.
3. *Aulotortus sinuosus* WEYNSCHENK, 1956.
Dredge sample 95:10/10B; Rhaetian.
4. *Variostoma cochlea* KRISTAN-TOLLMANN, 1960 (juvenile).
Dredge sample 95:10/10B; Rhaetian.
5. *Alpinophragmium perforatum* FLUGEL, 1967.
Dredge sample 95:7/15V; Norian–Rhaetian.
6,8. *Involutina turgida* KRISTAN, 1957.
6, somewhat depressed. Dredge sample 96:30/1; lowermost Liassic (Hettangian–lower Sinemurian).
7,8. *Involutina liassica* (JONES, 1853).
7, depressed. Dredge sample 96:30/1; lowermost Liassic (Hettangian–lower Sinemurian).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1984 *Frondicularia otamitaensis* n.sp. — STRONG, p. 21, pl. 2, figs. 40–41.

1984 *Frondicularia rhaetica* Kristan-Tollmann — STRONG, p. 21, pl. 2, figs. 42-46; pl. 8, figs. 172, 174.

1990 *Frondicularia rhaetica* KRISTAN-TOLLMANN, 1964 — KRISTAN -TOLLMANN, text fig. 10 , no. 9, 10; pl. 9, fig. 1.

1991 *Frondicularia rhaetica* KRISTAN-TOLLMANN, 1964 — KRISTAN-TOLLMANN in KRISTAN-TOLLMANN, LOBITZER & SOLTI, p. 164, pl. 6, figs. 11–14.

1992 *Frondicularia rhaetica* Kristan-Tollmann, 1964 — KRISTAN-TOLLMANN & GRAMANN, p. 467, pl. 2, fig. 1.

Remarks: This very characteristic species of *Frondicularia* is present in all Rhaetian marls in the Northern Calcareous Alps (Eastern Alps); for example in the Zlambach Marls (Kristan-Tollmann, 1964) and Kössener Schichten (Kristan-Tollmann, 1991). It is also distributed in other reefal limestones (e.g. Gurumugl, Central Highlands of Papua New Guinea; Kristan-Tollmann, 1990). Its Tethys-wide distribution is proved, including the region's southeastern margin (New Zealand; Strong 1984).

Recently, this species was also found in three sites on the Exmouth Plateau off northwest Australia: in whitish, Rhaetian, crinoidal marls from the Wombat Plateau (Kristan-Tollmann & Gramann, 1992); in Rhaetian marls in Investigator-1 and Vinck-1 exploration wells from the southwestern Exmouth Plateau (Kristan-Tollmann, unpublished); and in Upper Norian (Sevatian) limestones from the Rowley Terrace [this paper, Plate 5(11)].

***Aulotortus praegaschei* (ZANINETTI, 1976)**

Plate 5 (14,15)

1969 *Involutina gaschei* (KOEHN-ZANINETTI et BRÖNNIMANN) *praegaschei*, n. subsp. — KOEHN-ZANINETTI, p. 130, fig. 39.

1976 *Involutina gaschei praegaschei* Koehn-Zaninetti, 1968 — ZANINETTI, p. 161, pl. 14, figs. 17, 18 and 22; pl. 15, figs. 17–21.

1983 *Angulodiscus gaschei praegaschei* (KOEHN-ZANINETTI, 1968) — SALAJ, BORZA & SAMUEL, p. 145, pl. 11, fig. 3c; pl. 67, figs. 1–6; pl. 68, figs. 1, 3, 5, 6; pl. 69, figs. 1, 3, 4.

1991 *Aulotortus praegaschei* (Koehn-Zaninetti) — HO & NORLING, p. 29, pl. 2, fig. 10.

Remarks: The holotype of *A. praegaschei* was established only in 1976 by Zaninetti in the comment p. 236 of pl. 14, fig. 17. Therefore, this taxon was a *nomen nudum* in earlier papers.

A. praegaschei has a wide distribution across the whole of Tethys in the Ladinian to Norian. The species is well documented with characteristic sections by Salaj et al. (1983, pl. 67). These thick-shelled specimens are common in the Norian, Furmanec Limestone of the West Carpathians (Slovakia, Europe). They correspond very well with our material from the Rowley Terrace.

***Variostoma cochlea* KRISTAN-TOLLMANN, 1960**

Plate 6(4)

1960 *Variostoma cochlea* nov.gen.nov.spec. — KRISTAN-TOLLMANN, p. 63, pl. 12, fig. 6; pl. 13, figs. 1–12; pl. 14, fig. 5.

1964 *Variostoma cochlea* KRISTAN, 1960 — KRISTAN-TOLLMANN, p. 49f, pl. 39, figs. 3–5.

1977 *Variostoma cochlea* Kristan-Tollmann — HO & HU, p. 21, pl. 5, figs. 1, 2.

1978 *Variostoma cochlea* Kristan-Tollmann — MIRAUTA & GHEORGHIAN, p. 143, pl. 10, fig. 3.

1983 [NON] *Variostoma cochlea* KRISTAN-TOLLMANN, 1960 — SALAJ, BORZA & SAMUEL, p. 154, pl. 86, figs. 7, 8.

1986a *Variostoma cochlea* KRISTAN-TOLLMANN — KRISTAN-TOLLMANN, pl. 2, fig. 5.

1986b *Variostoma cochlea* KRISTAN-TOLLMANN — KRISTAN-TOLLMANN, p. 205, text fig. 1, nos. 9–14.

1988 *Variostoma cochlea* Kristan-Tollmann, (1960) — KRISTAN-TOLLMANN, p. 248, pl. 1, figs. 1-10; pl. 2, figs. 11–13.

1990 *Variostoma cochlea* KRISTAN-TOLLMANN, 1960 — KRISTAN-TOLLMANN, p. 242f, textfig. 15, nos. 1-16; pl. 14, figs. 1–6; pl. 15, fig. 5.

1991 *Variostoma cochlea* Kristan-Tollmann — KRISTAN-TOLLMANN, pl. 4, figs. 3, 7, 8.

Remarks: *V. cochlea* is a typical Rhaetian, Tethyan foraminiferid of basin and fore-reef regions. It is absent in lagoonal sediments. Its Tethys-wide distribution is

PLATE 7. FORAMINIFERA FROM NORIAN–RHAETIAN LIMESTONES, ROWLEY TERRACE
Scale bar indicates 200 μm.
1,2. *Tetrataxis inflata* KRISTAN, 1957.
Dredge sample 96:28/2a.
3. *Verneuilinoides mauritii* (TERQUEM, 1866).
Dredge sample 95:9/6B.
4. *Auloconus permodiscoides* (OBERHAUSER, 1964).
Dredge sample 95:9/6B.
5–10. *Angulodiscus communis* KRISTAN, 1957.
Dredge sample 95:12/11V. The figures display different stages of preservation.While 5 and 8 indicate lamination on the left, the other specimens are totally recrystallised. The outer wall is mainly micritised. The last whorl can be damaged (7, 9) or broken (5 bottom, 10) or the specimens partly corroded (5 top, 6). Often the lateral wall is weathered (10). The alterations, particularly of the *Angulodiscus* carapaces, indicate extended diagenesis in a very shallow-water environment.

documented in the latest review (Kristan-Tollmann, 1988a, p. 249, text fig. 2), with some more sites now identified in the eastern realm: apart from both the reefal limestone and the crinoidal limestone (the so-called Kuta Limestone) in the Central Highlands of Papua New Guinea (Kristan-Tollmann (1986, 1990), *V. cochlea* can also be identified in a crinoidal limestone of a Dachstein Limestone-block of Tethyan origin in the Sambosan Chichibu Zone of Kyushu Island, Japan (Kristan-Tollmann, 1991, p. 39, pl. 4, figs. 3, 7, 8) which appears to have been deposited in a high-energy environment in the fore-reef.

The very rare *V. cochlea* in dredge sample 95:10/10B seem to have floated into a backreef environment. The abundant coproliths there, as well as the other foraminifer species of the genera *Angulodiscus, Aulotortus* and *Triasina,* suggest a lagoonal environment.

Abundant *V. cochlea* occurs in the Rhaetian marls of the four wells Sirius-1, Vinck-1, Investigator-1, and Eendracht-1 from the southwestern Exmouth Plateau (Kristan-Tollmann, unpublished).

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The petrology, geochemistry and implications of basalts dredged from the Rowley Terrace–Scott Plateau and Exmouth Plateau margins, northwestern Australia

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We report major and trace-element compositions of 14 basalts and dolerites dredged from seven sites along the outer rifted margin of northwestern Australia. Lavas from the margin of the Scott Plateau–Rowley Terrace, of interpreted Callovian–Oxfordian age, are evolved basalts and ferrobasalts with Zr/Nb from 5–17, and LREE-enriched REE patterns [(La/Yb)N from 3.4 to 8], and are T-MORB transitional to more P-MORB from the SW Indian Ridge and the Red Sea–Afar rift zone. They are significantly different from the unusually depleted N-MORB drilled less than 100 km farther west on the Argo Abyssal Plain at ODP Site 765. T- to P-MORB were also dredged at the foot of the southwestern corner of the Exmouth Plateau. This dredge site is only 80 km from ODP Site 766, which also yielded N-MORB. Basalts very similar to those drilled at ODP Site 766 were also dredged at a site only 20 km east of Site 766.

The evolved, mainly ferrobasaltic nature of the Rowley Terrace–Scott Plateau margin basalts, their T- to P-MORB compositional signatures, and their close spatial association with strongly depleted N-MORB resemble in many ways the lava pile formed during the last 4 m.y. along the southern Red Sea–Gulf of Aden–Asal–Afar region. Although the latter region is associated with the effects of the long-lived (~40 m.y.) Afar plume, a plume origin for the northwest Australian margin basalts is less easily demonstrated. Problems with the plume hypothesis for this region include (1) the remarkably rapid change from 'plume-influenced' MORB along the Rowley Terrace margin to depleted N-MORB at the foot of this margin at the eastern end of the Argo Abyssal Plain, requiring a sudden switch-off of the hypothetical plum; (2) the apparent recurrence of this same pattern of T-MORB to depleted N-MORB basalts 400 km farther south on the same margin some 20–25 m.y. later (Valanginian); and (3) the occurrence of perhaps more convincingly plume-related basalts that constitute the Wallaby Plateau, which formed as part of interpreted plume head eruptions of the Kerguelen plume around 115 Ma, and were apparently erupted onto Cuvier Abyssal Plain oceanic crust only probably 30–40 m.y. old.

Introduction

Rifting of continental crust to produce a new ocean has been considered to be either essentially amagmatic, in which case a non-volcanic passive margin results, or to involve intense basaltic magmatic activity, producing a volcanic passive margin. White & McKenzie (1989) associated volcanic passive margins with massive melt production resulting from decompression melting of a megaplume during attenuation and rifting of thinned continental crust. As one example of this, they argued that the evidence for considerable magmatism along the west Australian margin (Fig. 1), including seaward-dipping reflectors, massive volcanic plateaux, and seismic evidence for broad-scale magmatic underplating, suggested that a mantle plume ascended beneath this margin close to the time of breakup.

Breakup commenced along the volcanic passive continental margin of northwestern Australia around 155 Ma in the north, and around 20–25 Ma later farther south, in what is now the Cuvier, Gascoyne and Perth Abyssal plains (Exon et al., 1982; Powell et al., 1988; Veevers & Li, 1991; Veevers et al., 1991). This rifting began when the 'Argo Land' continental block rifted off northwestern Australia in the Callovian–Oxfordian (~155 Ma) and seafloor spreading produced oceanic crust from anomaly M26 until anomaly M16 (Berriasian). Celadonite cementing a basaltic breccia directly overlying basaltic basement at ODP Site 765 at the foot of Rowley Terrace (on the oldest oceanic crust of Argo Abyssal Plain) yielded a K–Ar date of 155.3 ± 3.4 Ma (Ludden, 1992), and provides a minimum breakup age for this region. Farther south, breakup along the Cuvier and Gascoyne Abyssal Plains began in the Valanginian, around M10 (130 Ma) to M11 (132 Ma), and around anomaly M9 (129 Ma) in the Perth Abyssal Plain. Spreading continued until anomaly M4–5 time (127 Ma), when the ridge jumped westwards, abandoning the ridge system in the Cuvier and Gascoyne Abyssal Plains (Fig. 1).

Magmatism significantly preceding breakup along this margin is known from evolved trachytic to rhyolitic lavas of mainly Rhaetian to Early Jurassic age (213–190 Ma) drilled in several Browse Basin wells landward of Scott Plateau, from the margins of the Wombat Plateau, and from the southwestern margin of the Exmouth Plateau (von Rad et al., 1992a, b; Exon & Buffler, 1992; von Stackelberg et al., 1980). These eruptions accompanied a widespread intracratonic rift phase along this margin, and were probably erupted in shallow water to subaerial conditions (von Rad & Exon, 1983; von Rad et al., 1990). Basalt flows alternate with earliest Jurassic shallow-water limestone in Scott Plateau well Scott Reef No. 1 (von Rad et al., 1992a, b). Little is known of the geochemistry of these early Jurassic volcanics.

More widely represented and voluminous than the pre-breakup Late Triassic to Early Jurassic volcanics on this margin are extensive basalts of Oxfordian–Callovian age, that Ramsay & Exon (1994 — this issue) show to cover an area of at least 25 000 km^2 beneath the northwestern Rowley Terrace. This same magmatic phase is observed as packets of seaward-dipping reflectors of interpreted Oxfordian–Callovian on the northeastern margin of the Argo Abyssal Plain adjacent to the Scott Plateau (Hinz et al., 1978). Similar reflectors, and a well-defined underplated crustal layer with velocity 7.2–7.3 km/s, occur along the western and southern margin of the Exmouth Plateau and adjacent to the Cuvier Abyssal Plain (Mutter et al., 1989; Hopper et al., 1992). The latter authors suggested that the Cuvier margin reflectors formed shortly after the Valanginian–Hauterivian breakup in this region.

During cruises 95 and 96 of the *Rig Seismic* to the northwestern Australian margin, samples of the Oxfordian–Callovian basalts were dredged from along the

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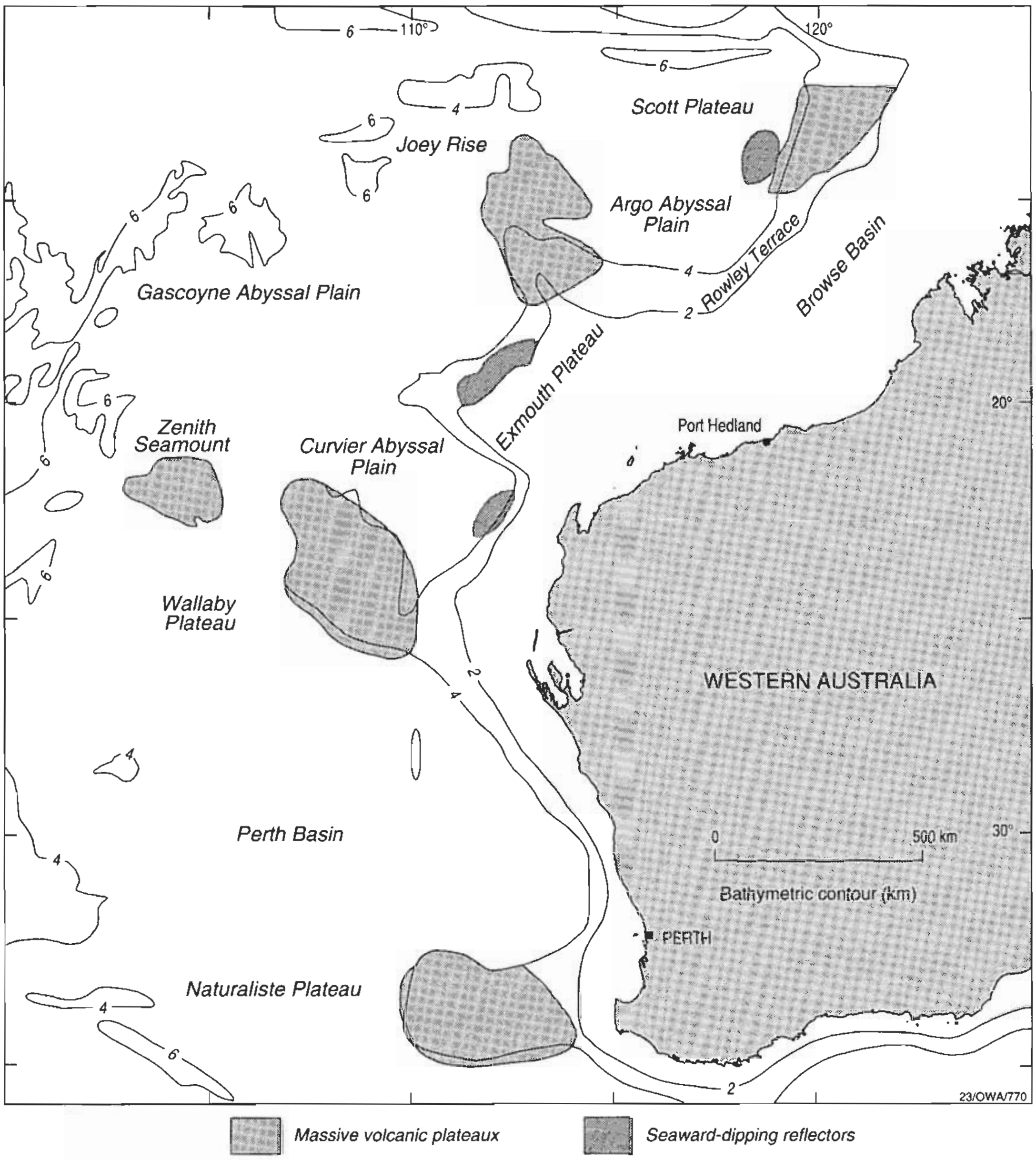

Figure 1. Map showing the the distribution of major volcanic provinces along the northwestern rifted margin of Australia, including the Exmouth and Scott Plateaux, Argo, Cuvier and Gascoyne Abyssal Plains, Wallaby Plateau and the Naturaliste Plateau and Perth Basin. Note massive volcanic plateaux argued by White and McKenzie (1989) to be related to inception of a mantle plume in this region. Note also regions of seaward-dipping reflectors. Modified from White and McKenzie (1989).

Rowley Terrace margin and the southwestern margin of the Exmouth Plateau. In this paper, we present a preliminary account of their major and trace-element geochemical features, and geodynamic implications. More detailed geochemical work, including Nd-Sr-Pb isotopic studies, are in progress (J.J. Mahoney & A.J. Crawford, in prep.).

Sampling

Precise dredge site locations are provided in the cruise final reports of AGSO (BMR) cruises 95 (Exon & Ramsay, 1990) and 96 (Colwell et al., 1990) to northwestern Australia. Those dredges which sampled volcanic rocks are described below, and dredge localities are shown on Figure 2.

Dredge 95/01: Northern slope of Taipan Canyon at water depths of 4650–3800 m. Seismic interpretation suggests that it should have recovered pre-Oxfordian rocks, and it sampled altered basaltic lavas, claystones of probable Jurassic age, a possibly Triassic red, calcite-veined claystone, and some Late Miocene and younger chalks and oozes. The basalts are mainly quite strongly altered. Most are quite evolved augite+plagioclase-phyric basaltic lavas, but 95/01-4A is a vesicular, sparsely plagioclase-phyric basalt with zeolite–smectite (chlorite) alteration; 95/01–6 is an oxidized and carbonate-altered augite+plagioclase-phyric basaltic lava breccia, and 95/01–2A is a strongly altered sparsely plagioclase-phyric hyaloclastite. Only relatively weakly-altered sample 95/0–4B was selected for analysis; it is a moderately augite+plagioclase-phyric basalt, with altered olivine microphenocrysts and common elongate ilmenite needles in a groundmass that includes slightly altered devitrified glassy mesostasis.

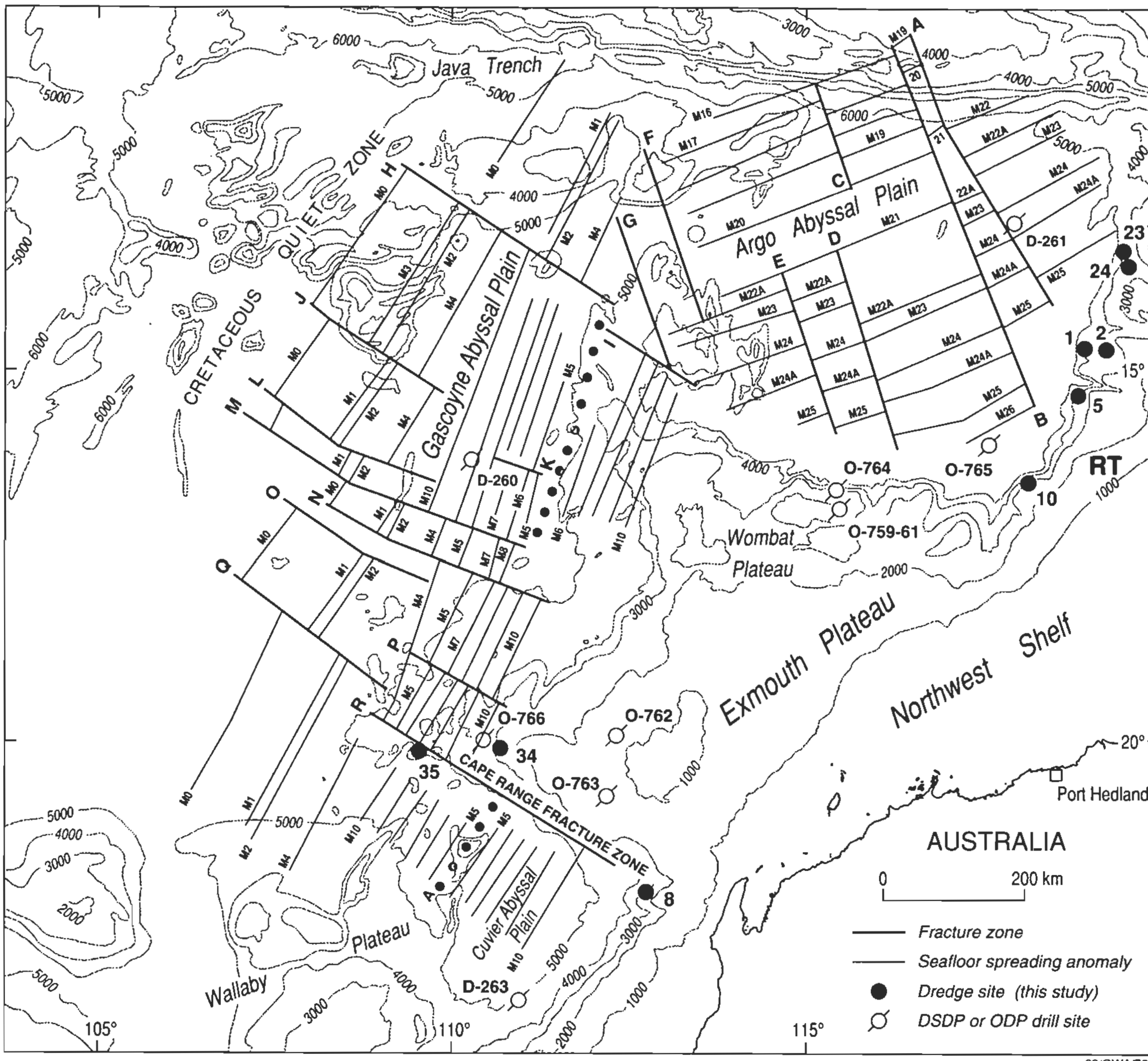

Figure 2. Map of the eastern Indian Ocean and northwestern continental margin of Australia showing Exmouth Plateau, Rowley Terrace (RT), magnetic lineations on the oceanic crust, and locations of dredge sites 95/01, 95/02, 95/05, 95/10, 96/08, 96/23, 96/24, 96/34 and 96/35. Also shown are locations of ODP Site 765 and 766. (Modified from von Rad et al., 1992.)

Dredge 95/02: Western slope of the northern arm of Taipan Canyon in water depths of 4000–3100 m. Seismic data suggest that it should have sampled similar pre-Oxfordian rocks to dredge 95/01. The dredge consisted largely of shallow-marine Oxfordian volcaniclastic grits with shelly fossil remains, minor altered basaltic volcanics, and minor foram ooze. Sample 95/02–1A is a rather altered augite+plagioclase-phyric evolved basaltic lava with altered olivine microphenocrysts, and it is essentially identical petrographically to 95/01–4B. Sample 95/02–2A is a calcite-cemented volcaniclastic grit with abundant dolerite clasts to at least 5 mm across, and common altered, devitrified glassy clasts.

Dredge 95/05: Northwest slope of Copperhead Canyon in 3850–3330-m water depth, across a section interpreted from seismic data to be mainly Jurassic–Cretaceous rocks. Palaeontological ages are Cretaceous (Colwell et al., 1994a, Table 2 — this issue). A few altered lavas were recovered in this dredge, which was dominated by ferruginous sandstones and ironstone breccias, and less abundant grey siltstones. The only sample selected for further study, 95/05–6A, is a grey, aphyric trachybasaltic to andesitic lava composed of an intersertal intergrowth of fresh plagioclase and slightly altered, granular augite and abundant FeTi oxides, with minor altered glassy mesostasis. One volcaniclastic sandstone also contained ~20 modal% terrigenous quartz clasts.

Dredge 95/10: Eastern slope of Clerke Canyon in water depths of 5325–4090 m. The dredge, which by inference from seismic data traversed a Late Triassic section, was dominated by a diverse haul of detrital marine sedimentary rocks, with common macrofossil-rich carbonates and around 30% of altered grey volcanics. Norian–Rhaetian ages have been determined palaeontologically (Colwell et al., 1994a, Table 2 — this issue). Most lavas are trachytic-textured sparsely plagioclase-phyric or aphyric basalts or trachybasalts, and two of these (95/10–14A and –15A) have been selected for detailed study; 14A has altered augite in the groundmass, whereas augite is fresh in 95/10–15A.

Dredge 96/08: Southeasternmost flank of Exmouth Plateau in water depths of 3550–3180 m. Sandstones and mudstones make up most of the dredge, but a few altered lavas were also recovered. The only rock selected for study is 96/08–10A, a microgabbroic shallow intrusive rock composed of intergrown fresh augite and calcic plagioclase, with interstitial FeTi oxide and minor chloritic alteration products.

Dredges 96/23 and 96/24: These dredge sites are from the flank of the Scott Plateau about 200 km north of dredge 95/01, and the age of one dredged rock is Toarcian (uppermost Lower Jurassic). Sample 96/23–4 is a moderately altered very sparsely plagioclase-phyric basalt with an intersertal groundmass composed of fresh augite, plagioclase, FeTi oxides and altered glass. Sample 96/24–1 is a moderately altered sparsely plagioclase-phyric basalt similar to 96/23–4.

Dredge 96/34: Southwestern corner of the Exmouth Plateau at 4000 m water depth, only 20 km east of ODP Site 766 located at the geophysical ocean–continent transition. The dredged sequence is interpreted to be of Valanginian age, based on the strong similarity of the dredged lavas to those drilled at ODP Site 766 just a short distance to the west, which occur within a late Valanginian (~134 Ma) sequence. Four samples were selected for detailed study. Sample 96/34–1 is a vesicular sparsely plagioclase-phyric trachybasalt with moderate smectite alteration. Sample 96/34–2 is a slightly altered sparsely plagioclase-phyric microlitic glassy basalt, and samples 96/34–3 and –4 are respectively finer and coarser-grained dolerites, with quite common iddingsite+calcite-altered olivine phenocrysts and either fresh or clay+zeolite-altered clusters of plagioclase phenocrysts in a holocrystalline, ophitic-textured groundmass composed of pinkish augite and fresh plagioclase, with subordinate interstitial FeTi oxides.

Dredge 96/35: Located 80 km west-southwest of ODP Site 766 on the southwestern corner of the Exmouth Plateau, at the foot of the slope at about 4000 m water depth. Dredged rocks are interpreted to be of Valanginian age (Colwell et al., 1990).The single sample selected for analysis is a strongly altered plagioclase-rich evolved dolerite, with abundant equant large FeTi oxides and altered augite; chlorite–smectite alteration is common, and veinlets of red-brown manganiferous material are not uncommon.

Whole rock geochemistry

Fourteen least-altered samples were selected for detailed geochemical study, and have been analyzed by XRF spectrometry at the Department of Geology, University of Tasmania, for major and trace elements (Table 1); nine of these were analyzed for rare earth elements (REE), Ta, Hf and Th using instrumental neutron activation analysis at Becquerel Laboratories (Sydney). As all samples were slightly to strongly altered, all analyses have been recalculated to 100% volatile-free, and loss on ignition is reported together with original totals below the recalculated major element analyses in Table 1. The high and variable ignition losses of the analyzed samples preclude evaluation of magmatic affinities using norms, and emphasis is placed on those elements (Ti, Zr, Y, Nb, Hf, Th, REE) considered to be immobile during low-grade alteration of basaltic rocks.

Compositions of the analyzed samples are shown as Harker diagrams (Fig. 3), with important oxides and trace elements plotted against the fractionation index FeO*/MgO. Two compositional groups are distinguished on the basis of trace element contents and REE patterns (see below): Group 1 comprises samples from dredges 1 2, 5, 10, 23 and 24, all from along the Rowley Terrace margin, and also samples 96/8–10A and 96/35–4 from the southeastern and southwestern corners, respectively, of the Exmouth Plateau; whereas Group 2 is represented by the four analyzed samples from dredge 34 from near the base of the southwestern escarpment of the Exmouth Plateau, 80 km east-northeast of dredge site 96/35.

Rowley Terrace Margin lavas and dredges 96/8 and 96/35

The Rowley Terrace margin (herein RT) and southern Exmouth Plateau dredged lavas and dolerites are of evolved basaltic–ferrobasaltic compositions with FeO*/MgO values greater than 1.78, and correspondingly high TiO_2 (3.0–6.0%) and P_2O_5 (0.36–1.6%) contents. In general, these rocks fall within the compositional fields defined by basalts dredged from the SW Indian Ocean ridges (Le Roex et al., 1983; Melson et al., 1977), an exception being the depletion of CaO and strong enrichment of Na_2O in 95/10–14B, which are obviously secondary alteration effects. The K_2O (and related elements Rb and Ba) contents of the RT lavas vary in an irregular manner, although still broadly following the trend defined by the Indian Ocean MORB; this possibly reflects the significant alteration of these rocks, as shown by their high and variable loss on ignition values (3.1–8.2%).

SW Exmouth Plateau

The four analyzed rocks from dredge 96/34 are less evolved, significantly less enriched in TiO_2, P_2O_5 and Na_2O, and have higher CaO contents at any value of FeO*/MgO than the RT margin samples. As for the RT lavas, the K_2O contents of dredge 96/34 rocks are high and rather variable (0.4–1.8%), posssibly reflecting alteration.

Strongly altered basalt 96/35–4 has very high LOI (13.5%), and shows massive enrichment of MnO (reflected petrographically in the amorphous brown–black manganiferous oxide–hydroxide material throughout the matrix of this basalt), with coupled enrichment of Ni (1000 ppm) and Ce and Pb (520 and 510 ppm, respectively, measured by XRF), and a pronounced depletion of CaO (2.7%). Despite this strong seafloor alteration, the immobile elements show that it is a ferrobasalt with high TiO_2 (3.7%), V (695 ppm), P_2O_5 (0.65%), Zr (476 ppm), and Nb (28.7 ppm), and is very similar to the ferrobasalts from farther north along the RT margin. An important point to note is that Th is remarkably enriched to 37 ppm (measured by XRF) in this sample; this is discussed further on.

Significant compositional differences are shown by the high field strength element (HFSE; e.g. Ti, Zr, Y, Nb, Hf) contents and ratios of the RT lavas and 96/35–4 relative to the rocks from SW Exmouth Plateau dredge 96/34. This is shown, for example, on the Zr/Nb versus Zr/Y(Fig. 4a) and (La/Yb)N (Fig. 4b) diagrams in which the 96/34 rocks fall at notably higher Zr/Nb values and lower Zr/Y and (La/Yb)N values than the RT margin and 96/35 lavas. This difference is also reflected in their respective rare earth element (REE) patterns (Fig. 5), with the dredge 96/34 rocks having broad, LREE-depleted patterns with slight HREE depletion, contrasting with the

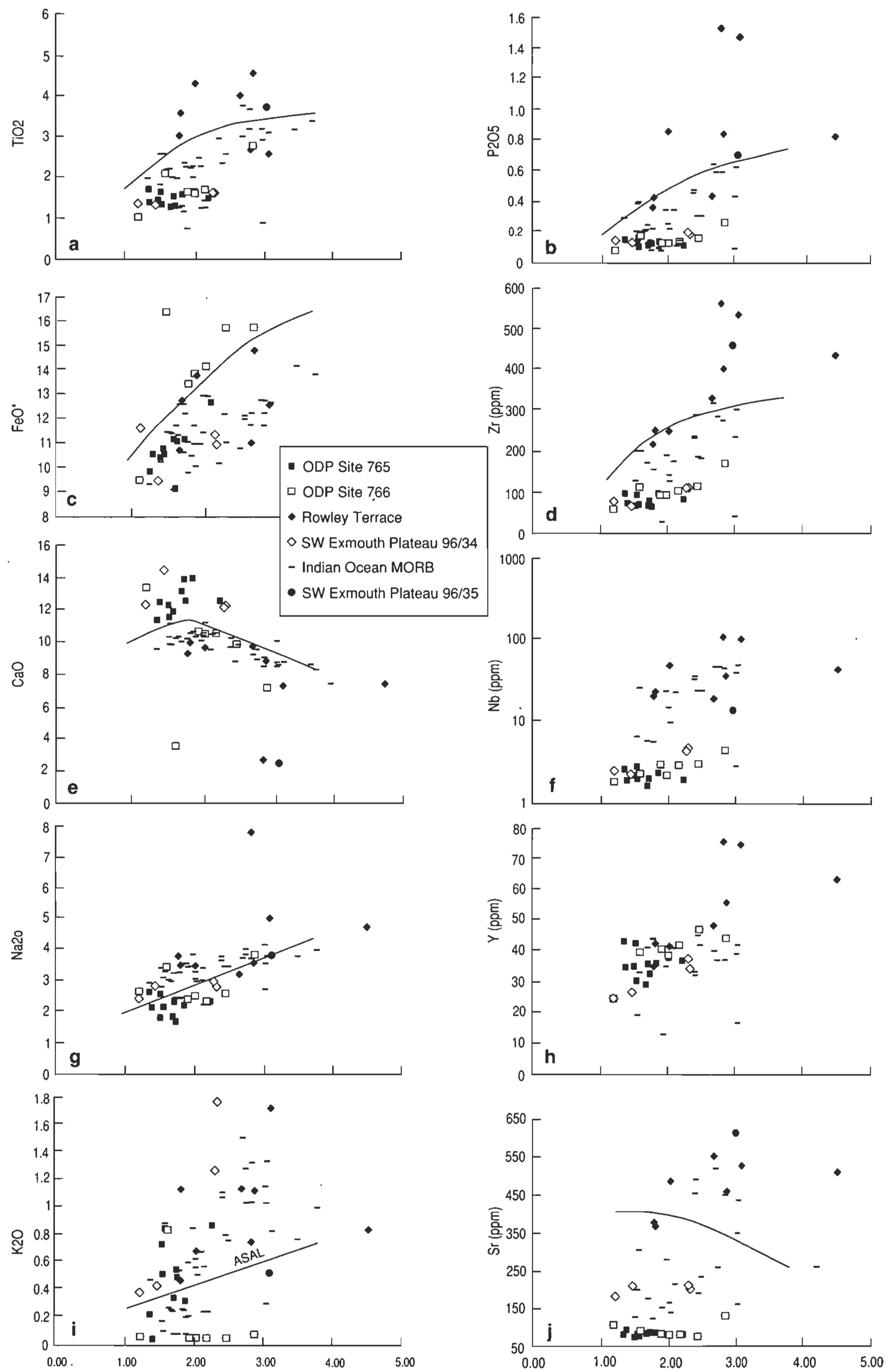

Figure 3. Harker diagrams for dredged samples from the northwestern Australian margin. Also shown are selected basaltic glasses from the Indian Ocean (Melson et al., 1977) and basalts from the plume-affected SW Indian Ridge (Le Roex et al., 1983; 1989). A best-fit line for data for the Asal Rift in the Horn of Africa region is also shown (from Piccirillo et al., 1979).

Table 1: Wholerock analyses of dredged basalts from the Rowley Terrace and SW Exmouth Plateau margin. LOI = Loss on Ignition; BDL = below detection limit.

Identification	*95/01–4B*	*95/05–6A*	*95/10–13A*	*95/10–14B*	*95/10–15A*	*96/8–10A*	*96/23–4*	*96/24–1*	*96/34–1*	*96/34–2*	*96/34–3*	*96/34–4*	*96/35–4*
SiO_2	51.66	51.50	46.73	54.08	52.28	48.58	49.76	49.30	50.49	51.66	47.36	50.70	42.45
TiO_2	3.03	6.00	4.33	2.67	2.55	4.02	4.57	3.57	1.60	1.61	1.34	1.31	3.72
Al_2O_3	17.06	19.46	16.55	17.70	17.47	15.91	14.49	15.69	17.32	17.96	16.06	17.99	14.47
Fe_2O_3	11.79	9.97	15.29	12.12	13.84	13.70	16.35	14.03	12.03	12.47	12.83	10.43	19.13
MnO	0.17	0.09	0.25	0.23	0.31	0.19	0.28	0.20	0.19	0.21	0.32	0.17	6.25
MgO	5.97	1.99	6.84	3.88	4.03	4.63	5.16	7.00	4.69	4.92	9.56	6.46	6.29
CaO	9.22	7.38	9.64	2.55	7.28	9.71	8.85	9.95	12.17	11.99	12.23	14.36	2.74
Na_2O	3.76	4.68	3.45	7.77	4.97	3.17	3.54	3.45	2.76	2.94	2.41	2.78	3.72
K_2O	1.14	0.84	0.69	0.75	1.73	1.15	1.13	0.46	1.77	1.27	0.38	0.43	0.54
P_2O_5	0.36	0.82	0.85	1.52	1.46	0.44	0.84	0.43	0.19	0.20	0.14	0.13	0.65
LOI	4.64	5.85	4.13	5.15	3.13	6.75	3.82	4.65	5.06	3.57	5.86	3.93	13.46
FeO*/MgO	1.78	4.51	2.01	2.81	3.09	2.66	2.85	1.80	2.31	2.28	1.21	1.45	2.74
Ni	80	15	61	10	8	10	37	82	64	74	173	138	1178
Cr	222	39	131	2	2	3	7	274	351	345	365	359	331
V	349	539	405	93	93	433	363	361	340	366	320	315	695
Sc	45	55	45	25	22	34	32	42	45	47	36	40	47
Zr	219	434	248	563	536	329	402	250	110	109	76	72	476
Nb	20.8	44.0	48.8	108.1	101.9	18.9	36.2	23.7	4.8	4.4	2.5	2.3	28.7
Y	35	63	42	75	74	48	56	43	34	37	25	26	114
Sr	375	507	484	932	523	547	457	362	201	212	182	209	607
Rb	37	9	11	18	37	50	27	10	36	27	8	9	16
Ba	182	285	227	139	625	329	274	126	54	43	38	27	497
La	14.20	31.36	29.94		69.60	27.21	30.92	16.50		6.36		3.22	
Ce	35.07	76.70	70.91		159.53	64.14	75.94	40.43		17.05		9.03	
Nd	22.07	47.53	40.54		86.87	39.23	48.08	28.41		12.78		7.87	
Sm	6.34	12.78	9.89		17.70	9.68	12.67	7.74		4.02		2.88	
Eu	2.17	4.45	3.43		5.88	2.97	4.38	2.71		1.58		1.15	
Tb	1.08	1.99	1.42		2.56	1.44	1.99	1.46		0.92		0.67	
Ho	1.27	2.13	1.56		2.51	1.75	2.16	1.75		1.22		0.90	
Yb	2.25	3.93	3.22		5.74	3.93	3.52	3.22		2.98		2.00	
Lu	0.27	0.55	0.42		0.85	0.52	0.46	0.39		0.38		0.27	
Hf	5.02	10.30	5.02		11.04	7.34	9.24	6.15		2.53		1.79	
Ta	1.20	3.07	3.17		6.80	1.53	3.07	1.63		0.67		0.66	
Th	1.44	3.32	2.10		5.38	7.48	2.26	1.72		0.83			
(La/Sm)N	1.37	1.50	1.85		2.40	1.71	1.49	1.30		0.96		0.68	
(La/Yb)N	4.17	5.26	6.13		8.01	4.57	5.80	3.38		1.41		1.06	
Zr/Nb	10.53	9.85	5.08	5.21	5.26	17.40	11.12	10.55	22.95	25.00	30.43	31.43	

strongly LREE-enriched RT margin basalts. REE were not determined by INAA for sample 96/35–4; however, XRF analyses yielded the following values: La 134 ppm, Ce 520 ppm and Nd 142 ppm. These data indicate a strongly LREE-enriched pattern for this sample, with a prominent positive Ce anomaly resulting from oxidative seafloor alteration. The significance of these differences in HFSE and REE contents between the 96/34 rocks and the evolved, LREE-enriched basalts and ferrobasalts from the RT margin and dredge site 96/35 is discussed below.

Affinities of the dredged basalts

In the Harker diagrams in Figure 3, the dredged basalts from the northwestern margin of Australia are compared with basalts drilled at ODP Sites 765 and 766 in the same region, and with a selected suite of basalts and basaltic glasses from Indian Ocean spreading centres (Le Roex et al., 1983; 1989; Melson et al., 1977). Also shown

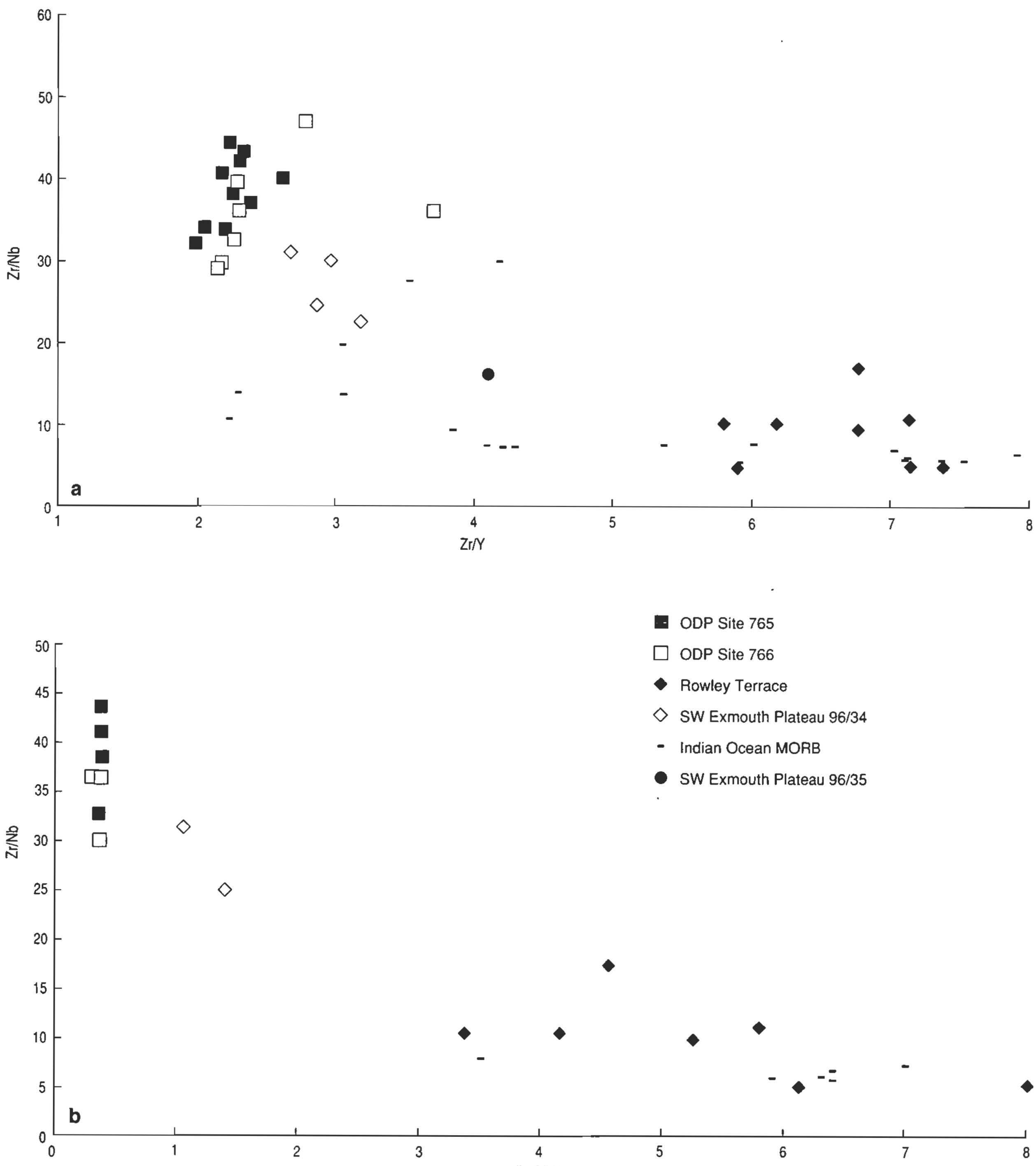

Figure 4. Plots of Zr/Nb versus (a) Zr/Y and (b) (La/Yb)N for basalts from Rowley Terrace and the southwestern corner of the Exmouth Plateau, from ODP Sites 765 and 766, selected basaltic glasses from the Indian Ocean (Melson et al., 1977) and basalts from the plume-affected SW Indian Ridge (Le Roex et al., 1983, 1989).

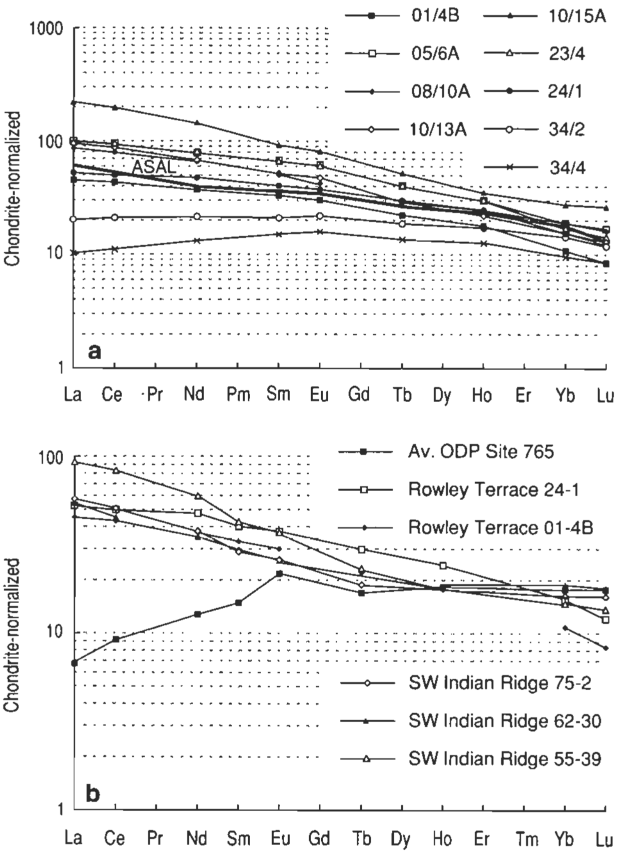

Figure 5. Chondrite-normalized REE patterns for (a) basalts from the SW Exmouth Plateau (34/2 and 34/4) and the Rowley Terrace; also shown is an average pattern for the Asal rift ferro-tholeiites from the Afar region on the Horn of Africa (Schilling et al., 1992), and (b) representative Rowley Terrace least-evolved basalt REE patterns compared with an average pattern for the depleted N-MORB drilled at ODP Site 765 (Ludden and Dionne, 1992), and three representative patterns for T- to P-type tholeiitic basalts from the SW Indian Ridge (Le Roex et al., 1983).

is a best-fit line for the well-defined magmatic trends (Piccirillo et al., 1979) for modern basalts from the Asal Rift in the Afar triangle, adjacent to the southernmost Red Sea. Selection of the SW Indian Ridge basalts for plotting on these Harker diagrams, many of which are from spreading ridge segments along the SW Indian Ridge that are significantly influenced by the Bouvet mantle plume (Le Roex et al., 1983, 1989), was deliberately biased towards more fractionated lavas that are comparable in terms of extents of fractionation with the evolved samples that form the focus of this study. Important points to be noted from the Harker diagrams are that the RT margin basalts are more strongly enriched in TiO_2, P_2O_5 and Zr at any FeO*/MgO value than even plume-related basalts from the SW Indian Ridge, although other major elements fall within the fields defined by the Indian Ocean MORB. Also, the SW Exmouth Plateau ferrobasalt 96/35–4 is directly comparable with the RT ferrobasalts, and notably unlike the N-MORB lavas drilled at nearby ODP Site 766. Compared with the drilled basaltic lavas and sills from ODP Sites 765 and 766, the RT lavas and 96/35–4 are strongly enriched in TiO_2, P_2O_5, Zr, Nb and especially Sr, and depleted in CaO.

Key element ratios unaffected by fractionation, such as Zr/Nb, La/Nb, Ta/La and Th/La are useful in distinguishing the affinities of these lavas. According to the classification scheme for oceanic basalts (MORB) proposed by Le Roex et al. (1983), normal (N-) MORB have Zr/Nb> 17, Y/Nb> 8 and chondrite-normalized La/Yb values (herein (La/Yb)N) <1.1. Enriched, or plume (P-) MORB have Zr/Nb< 8, Y/Nb values from 0.9–1.2, and (La/Yb)N> 4.8. Compositions falling between these extremes are classified as transitional (T-) MORB.

On the basis of their low Zr/Nb (~5) and Y/Nb (0.69–0.85) values and strongly LREE-enriched REE patterns (Figs 4, 5), RT basalts from dredge 95/10 are P-MORB. The remainder of the Group 1 basalts (Zr/Nb 9 to 17, Y/Nb 1.4 to 2.5, and (La/Yb)N values from 3.4 to 4.1 are T-MORB, as is the altered ferrobasalt 96/35–4. Two REE patterns of the least evolved analyzed RT basalts are compared in Figure 5 with patterns for plume-influenced basalts from the SW Indian Ridge, and an average pattern for the Asal Rift basalts (Schilling et al., 1991), and a strong similarity is evident. In contrast, Group 2 basalts, with Zr/Nb from 23 to 31, Y/Nb from 7 to 11 and (La/Yb)N values of 1.06 and 1.4 for the two analyzed

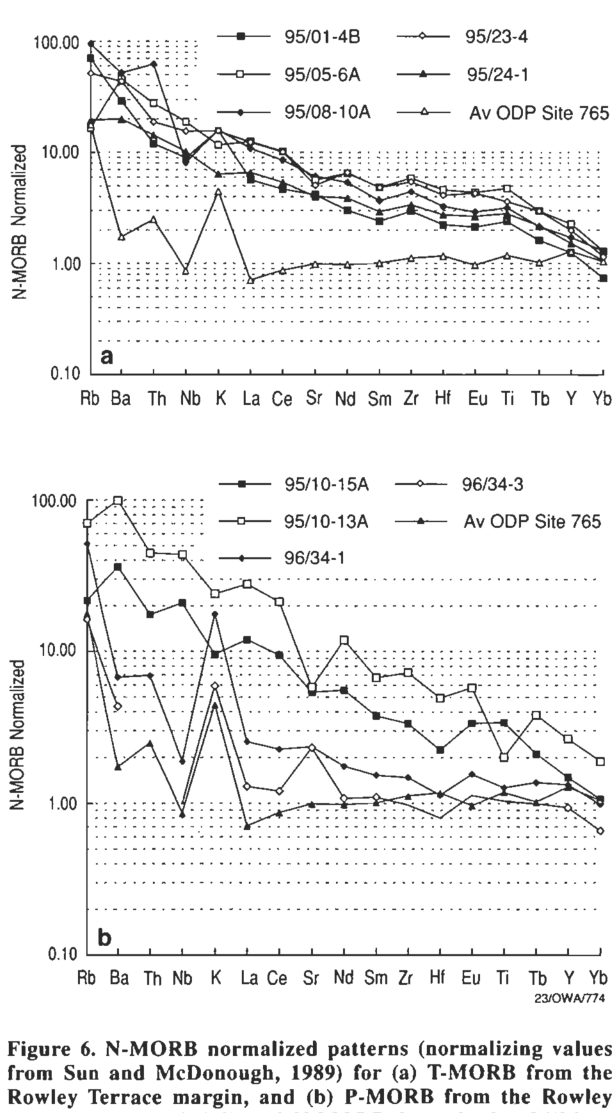

Figure 6. N-MORB normalized patterns (normalizing values from Sun and McDonough, 1989) for (a) T-MORB from the Rowley Terrace margin, and (b) P-MORB from the Rowley Terrace margin (95/10) and N-MORB from dredge 96/34 on the southwestern corner of the Exmouth Plateau. In both diagrams, the average ODP Site 765 N-MORB are shown for comparison.

samples, are best classified as N-MORB. Figure 6 shows N-MORB-normalized element variation patterns for analyzed samples from each of the above groups, a pattern for the average N-MORB drilled at ODP Site 765, and an average pattern for the Asal Rift basalts (Piccirillo et al., 1979; Schilling et al., 1991). Whereas the P- and T-MORB from the RT region are clearly distinguished from the Site 765 N-MORB on the basis of their pronounced enrichment in LREE and incompatible elements, those N-MORB dredged at 96/34 are very similar to the Site 765 basalts, the only difference being slightly more depleted LREE in the latter. This similarity extends to identical anomalous enrichments in Rb, Th and K in the 96/34 lavas and those from ODP Site 765; the same style of anomalous enrichment is evident in RT basalt 96/08–10A and highly altered ferrobasalt 96/35–4. Without isotopic data, it is difficult to determine with confidence whether these unusual patterns reflect some crustal contamination (or involvement of crustally contaminated subcontinental mantle lithosphere), or whether they are simply alteration effects. Although Th is regarded as an immobile element during seafloor alteration of basalts, recent work by Bienvenu et al. (1990) showed clearly that Th is strongly enriched in the alteration products of glassy oceanic basalts. In this light, and the fact that anomalously high K, Rb and Th contents occur in the most altered samples (as based on petrographic examination and LOI contents), we suggest that the positive spikes for Rb, K and Th in the patterns shown in Figure 6 are alteration-related.

We draw attention to the strong compositional similarity between the RT ferrobasalts and samples 96/8–10A and 96/35–4 from the southeastern and southwestern corners, respectively, of the Exmouth Plateau, some 400 km farther south. In a similar manner that the RT P- and T-MORB are located close to the Oxfordian N-MORB drilled at ODP Site 765, so is 96/8–10A (dredged from the southeastern margin of the Exmouth Plateau, southeast of dredge 35) and 96/35–4 (from less than 80 km from the Valanginian N-MORB drilled at ODP Site 766). Either the southern Exmouth margin samples are from an extensive continuation of the RT lava pile stepped some 400 km farther south and 400 km west (see Fig. 1), or more plausibly perhaps, the same magma generation processes led to production of P- and T-MORB followed by N-MORB during the Valanginian along the Cuvier Abyssal Plain margin, and during the Oxfordian some 30 m.y. earlier on the RT margin.

Discussion

The extensive Oxfordian–Callovian basaltic sequences along the Rowley Terrace margin of northwestern Australia are composed of quite strongly fractionated basalts and ferrobasalts that vary in key element signatures from T-MORB to P-MORB, with the former being dominant if the dredge volumes of each magma type are representative of their proportions in the volcanic pile. These basalts were produced immediately prior to breakup along this margin, and magma compositions changed rapidly at breakup towards the strongly depleted N-MORB compositions drilled at ODP Site 765, less than 100 km to the west.

This rapid change in magma composition accompanying break-up is clearly shown by plots such as Zr/Nb versus Zr/Y (Fig. 4a) and Zr/Nb versus $(La/Y)_N$ (Fig. 4b). A simple explanation of the data is mixing between melts derived from a depleted asthenospheric N-MORB mantle source, and melts derived from a more enriched 'plume-type' source (e.g. Le Roex et al., 1983, 1989). Earlier-formed melts related to the pre-breakup magmatism are derived from a source with more 'plume' component [i.e. high $(La/Y)_N$, low Zr/Nb], which is 'switched off' or absent by the time the Site 765 N-MORB are erupted.

A mantle plume origin for the NW Australian margin?

The pronounced similarity in the compositional characteristics of the RT margin lavas with the plume-affected T- and P-MORB from the SW Indian Ridge supports the claim that a mantle plume may have been involved in the lithospheric stretching, rifting and breakup along this margin. The low $(La/Nb)_N$ values for the RT lavas (<1) are unlike those (>1.5) for basalts in which involvement of subcontinental lithospheric mantle is postulated (Mahoney et al., in press; Colwell et al., 1994b — this issue). Thus, if a plume did generate the RT basalts, it apparently did not interact with the subcontinental lithosphere along this margin, but yielded the RT basalt parental magmas by direct partial melting of the plume (plus entrained asthenosphere?) itself. However, a key point is that *by the time of breakup*, voluminous basaltic magmatism such as that sampled at ODP Site 765 and 766 shows no plume signature at all. Rather, it bears the signature of strongly depleted N-MORB that appear to have been generated by relatively high degrees of partial melting compared to even typical N-MORB (Ludden & Dionne, 1992). No apparent plume trace extends from this margin, and the plume seems to have abruptly shut down at this time.

An alternative to the plume scenario for the generation of the Late Triassic–Jurassic basalts along the northwestern Australian margin involves passive mantle ascent and melting, in response to lithospheric stretching and extension in this area starting around 155 Ma, and jumping southward to commence at ~130 Ma on the Cuvier margin.. In this model, which incorporates aspects of the Mutter et al. (1988) and Hopper et al. (1992) models, early magmatism derives from the 'low melting temperature' enriched component in the asthenospheric mantle, possibly present as veins in more refractory depleted mantle (Altherr et al., 1990). The binary mixing (two components) evident from trace element data for the RT basalts would involve depleted asthenospheric mantle as one end-member, and the low-melting veins as the other. Once strong extension begins, a short burst of more rapid upwelling and convection leads to more extensive melting of (residual, refractory?) mantle, producing strongly LREE-depleted N-MORB (Mutter et al., 1988), before magmatism settles into the steady-state N-MORB generation mode recorded by the magnetic anomalies in both the Argo and Cuvier–Gascoyne–Perth abyssal plains. This scenario broadly supports the model proposed by Mutter et al. (1988) and Hopper et al. (1992) in which the earliest stages of breakup along this margin are characterized by rapid rifting and extension, leading to establishment of strong lateral temperature gradients in the subjacent upper mantle. This forced small-scale convection, increased mantle upwelling and increased amounts of partial melting, and a significant increase in the volume of magma generated during this decompression by partial melting. The 'more depleted than normal' N-MORB basalts at both ODP Sites 765 and 766 have more extreme compositional characteristics [higher CaO/Na_2O, lower $(La/Sm)_N$, and Ti/Zr values (av. 110) than in typical SW Indian Ridge or

Mid-Indian triple junction MORB] expected of relatively high-degree melts of an N-MORB mantle source, as pointed out by Ludden & Dionne (1992).

Red Sea–Gulf of Aden–Afar analogy

A very similar array of basalt compositions to that documented along the northwestern Australian margin is recorded from the Red Sea–Gulf of Aden–Gulf of Tajoura region. In this area, an extended period of crustal attenuation and rifting was accompanied by extensive flood basalt magmatism around 20–25 Ma (Sebai et al., 1991); these flood basalts are correlated by White & McKenzie (1989) with arrival beneath the region of a plume head on the order of 1000 km across. Volcanism waned from 20 Ma until around 4.5 Ma, when renewed extension and an accompanying burst of basaltic magmatism led to breakup and the opening of the Red Sea and the western end of the Gulf of Aden, both of which are still actively spreading, and erupting P-, through T-, to N-MORB where extension is greatest.

Considering only the latter stages (last 10 m.y.) of crustal extension and magmatism in the Horn of Africa region, some impressive similarities with the Late Jurassic–Early Cretaceous northwestern Australian margin are evident. The dominant basalt compositional variant erupted in the southern Red Sea–Gulf of Tajoura region since 4 Ma is T-MORB, many of which are transitional to E-MORB, with Zr/Nb ranging from 11.2 to 7.0, averaging 8.1, and (La/Yb)N in the range 1.3 to 5.3, averaging around 3 (Barrat et al., 1990; Schilling et al., 1992). However, in the axial valley of the Red Sea, typical N-MORB basalts are recorded, and significantly, some are strikingly depleted in LREE, akin to the ODP Site 765 and 766 basalts from the northwestern Australian margin. Basalts from the extensive on-land ridge segment of the Asal (Ardoukoba) rift zone show pronounced compositional similarities to the RT margin basalts (Figs 3, 4, 5), and are dominated by evolved basalts and ferrobasalts with high TiO_2, P_2O_5 and HFSE abundances, and moderately to strongly LREE-enriched REE patterns (Barrat et al., 1990; Schilling et al., 1992).

Schilling et al. (1992) have presented a case, based on extensive trace element and isotopic studies of basalts from the Red Sea–Afar area, that production of the post-40 Ma magmatism in this region involved both passive lithospheric extension (60–40 Ma) and impingement and flattening of a torus-like mantle plume on the base of the lithosphere around 30 Ma. Prior to plume arrival, diffuse lithospheric extension was accompanied by only minor magmatic activity, derived from the asthenosphere. Arrival of the plume at the base of the lithosphere, and flattening of the plume head, produced the Ethiopian–Yemen plateau basalts, which appear to require involvement of some component from the continental lithospheric mantle incorporated into the plume during thermal erosion along the base of the lithosphere. The tail of the plume is now dragged northwards by movement in that direction of the African plate, and asthenospheric mantle incorporated into the plume centre provides N-MORB basalts to the developing Red Sea and Gulf of Aden rifts, while the plume conduit itself provides alkaline magmas that form small islands in the southern Red Sea. This clearly requires important lateral and vertical heterogeneity in the Afar plume at the present time.

For the northwestern Australian margin, an extensive plateau basalt formation correlated with the Ethiopian flood basalts either remains undiscovered, or does not exist. Strong geological evidence exists for crustal stretching and block faulting along this margin before onset of the main Callovian–Oxfordian magmatic event recorded by the RT basalts. The compositions and significance of the limited volumes of Late Triassic–Jurassic volcanics accompanying this block faulting are still to be determined. The main magmatic event at around 155 Ma along this margin involves a significant 'plume' input, and shows no evidence of input from thermally eroded subcontinental lithospheric mantle. It might be argued that the voluminous RT basalts are the northwest Australian analogue of the Ethiopian plateau basalts, and record the arrival beneath the region of the plume head. However, the rapid (<5 Ma) switch from 'plume'-dominated basalts to depleted N-MORB (ODP Site 765) demands, as for the current Afar setting, massive heterogeneities in the plume, and in the case of northwestern Australia, the plume appears to have shut off abruptly at this stage.

A second complicating factor with the Afar plume analogy for northwestern Australia's Mesozoic magmatic evolution is that the sequence of magmatism that produced the RT lavas followed by highly depleted MORB at Site 765 around 155 Ma appears to have repeated itself along the Cuvier margin 500 km to the south at around 130 Ma. Furthermore, as discussed in Colwell et al. (1994b — this issue), it appears that the flood basalts of the Wallaby Plateau were superimposed, possibly around 115 Ma, on relatively new oceanic crust of the Cuvier Abyssal Plain following impact of the massive Kerguelen plume head on the base of the lithosphere in this region (and extending for more than 2000 km southwards and to the west). This represents a complicated (albeit hypothetical) scenario, in which two major mantle plumes affect the northwestern Australian margin within 40 m.y. The first evolved rapidly into a setting in which strongly depleted N-MORB are produced, then typical N-MORB, and plume-tail related alkali basalts are unknown; the second shows no such evolution, and evolved into the well-mapped Kerguelen plume.

Isotope studies on the northwest Australian margin basalts described herein are still in progress, and should better constrain the nature and composition of the source of these extensive lavas. If a mantle plume was not involved in the genesis of these basalts, we need to investigate a problem common to all models denying the role of a mantle plume in rifting and breakup leading to ocean opening, namely, why the RT (and other continental rift) lavas bear a compositional imprint essentially identical in major and trace element terms to plume-affected MORB. Future ODP drilling of massive seaward-dipping reflector sequences along the north Atlantic margin may provide key evidence to choose between these models.

Conclusions

Callovian–Oxfordian (~155 Ma) basalts form a massive welt along the Rowley Terrace–Scott Plateau margin of northwestern Australia. They are evolved basalts and ferrobasalts with affinities extending from T-MORB through to plume-influenced P-MORB, and on available evidence show no sign of involvement of the subcontinental lithospheric mantle in their petrogenesis. A rapid change to strongly depleted N-MORB is recorded along this margin by the basalts recovered at ODP Site 765 only 100 km west of the Rowley Terrace margin.

A single dredge at the foot of the southwestern corner of the Exmouth Plateau yielded highly altered ferrobasalts with immobile element signatures identical to those of the Rowley Terrace margin basalts 500 km farther north. Some 80 km farther east, and still on the foot of the southwestern Exmouth Plateau scarp, one dredge station, and also ODP Site 766 produced N-MORB basalts and sills of interpreted Valanginian age (~130 Ma).

Two quite different classes of models for the development of this continental margin might be considered. The first class, as proposed by White & McKenzie (1989) involves, as a major 'enforcer', a mantle plume that probably affected more than 1000 km of this margin. Problems with this model are the apparent absence of plume head-related plateau flood basalts, the rapid switch-off of the plume shortly after generation of the RT basalts, and the jump to strongly depleted N-MORB compositions during earliest breakup magmatism. A second class of model does not involve a mantle plume, and implies that lithospheric extension leads asthenospheric mantle to passively ascend beneath the developing rift. Earliest erupted lavas are hybrids between an enriched, easily fusible component probably present as veins, and the more typical depleted asthenospheric reservoir that hosts these veins. The convective partial melting model of Mutter et al. (1988) and Hopper et al. (1992) can be accommodated within the latter model. In addition, recent documentation by Holbrook & Kelemen (1993) of extensive seaward-dipping reflectors along the Atlantic coast of USA, with no associated plume trace, suggest that mantle plumes may not be required to produce the extensive basaltic magmatism associated with several of the Earth's major volcanic passive margins.

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Structure and tectonic history of the northern Exmouth Plateau and Rowley Terrace: outer North West Shelf

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A series of regional seismic reflection lines has been recorded, which link three key exploration wells on the northern Exmouth Plateau and Rowley Terrace to the lower continental slope; this slope has also been extensively dredged. This study primarily examines these new data, combined with all existing seismic, well and dredge data for some 200 000 km^2 of the northwest Australian margin, to come up with an improved structural and stratigraphic interpretation, and hence a better understanding of geological history. The study area lies between the shallow North West Shelf and the Argo Abyssal Plain and is underlain by continental crust. Water depths increase slowly seaward from 500 to 3000 m and then rapidly to 5500 m at the edge of the abyssal plain.

The northern Exmouth Plateau and Rowley Terrace margin is underlain by thinned continental crust, and contains 5000–10 000 m of Permian and younger sediments. The sequences we analyse here are the top 1000 m of Upper Triassic sedimentary rocks, 1000–2000 m of Jurassic sedimentary and volcanic rocks, 500–1000 m of Cretaceous sedimentary rocks and 500–2000 m of Cainozoic sedimentary rocks. Triassic sediments were laid down as a fluvio-deltaic blanket across this part of Gondwana, which had continental crust extending over the area presently occupied by the Argo Abyssal Plain. Coral reefs existed near the open ocean, but we have no unequivocal seismic evidence of them except on the Wombat Plateau. Toward the end of the Triassic a period of faulting on the Exmouth Plateau led to the formation of the F unconformity, but the Rowley Terrace remained undisturbed. In the Early Jurassic, a thick sequence of marine clastics with some carbonates was deposited, which, after a regression, was overlain unconformably near the sea by a thick Middle Jurassic coal measure sequence.

At the end of the Middle Jurassic, there was a period of thermal uplift, normal and strike-slip faulting, volcanism and erosion over a zone within 100–150 km of the future abyssal plain, making the widespread E unconformity particularly angular in this area. This tectonism culminated in breakup in the Callovian–Oxfordian, and the "Argo Landmass" drifted northwestward leaving oceanic crust behind. Continental volcanics are interpreted to cover about 25 000 km^2 of the northwestern Rowley Terrace below the E unconformity. An Upper Jurassic to Berriasian deltaic sequence was commonly confined to structural lows, suggesting that rapid subsidence of the margin commenced later, accompanied by a marine transgression that led to the formation of the Valanginian D unconformity. Rapid subsidence of oceanic crust led to the collapse of the continental margin along faults in a zone 30–40 km wide, with the continent–ocean boundary positioned somewhat shallower than the abyssal plain, and about 10–20 km seaward of the magnetic anomaly that previous authors used to locate it. Maximum vertical displacement of the Mesozoic sequences across this collapsed zone is now about 2000 m.

No major tectonic events occurred on the continental margin after the Valanginian, with steady subsidence leading to a general deepening of environments. Continental movement northward led to an increasing proportion of biogenic carbonate in the sediments. Much of the area was in bathyal depths by the Paleocene, but thereafter increased carbonate production and oceanward progradation led to a shallowing of water with time in many areas.

Introduction

The northern Exmouth Plateau and the Rowley Terrace both lie in water depths largely in the range 1000–3000 m (Fig. 1). They flank the Argo Abyssal Plain where the water is about 5700 m deep, and merge landward into the North West Shelf. They are contiguous areas of continental crust and have similar geological histories, their Mesozoic history being of most interest. The plateau consists of rifted and deeply subsided continental crust, with a Phanerozoic sedimentary sequence about 10 km thick. Both the plateau and the terrace have steep outer margins which are heavily faulted and cut by canyons.

The Exmouth Plateau is a marginal plateau with water depths of 800–4000 m, and the area shallower than 2000 m is approximately 150 000 km^2. It is separated from the North West Shelf by a bathymetric depression, and is bounded to the north, west, and south by oceanic crust. The Rowley Terrace lies between the Rowley Shoals and the Argo Abyssal Plain, and is limited to the north by Bowers Canyon (about 14°S), and to the southwest by the Swan Canyon. Water depths range from 300 to 3000 m and the area of the terrace is about 200 000 km^2.

There have been seismic reflection surveys of both areas by petroleum exploration companies, the Bureau of Mineral Resources (BMR), and its successor, the Australian Geological Survey Organisation (AGSO). The early seismic and other geophysical data were interpreted and reviewed by Exon & Willcox (1978, 1980) and Exon et al. (1982) for the Exmouth Plateau, and by Stagg & Exon (1981) for the Rowley Terrace. A composite structure map of the base of the Upper Cretaceous carbonate sequence, drawn from these authors, is shown in Figure 2. More recent seismic cruises are those reported by Exon et al. (1988), and Exon & Ramsay (1990).

There have been a number of sampling cruises to dredge the steep slopes in the study area, and the results have been reported by von Stackelberg et al. (1980), von Rad & Exon (1983), Exon et al. (1988), von Rad et al. (1990), Exon & Ramsay (1990), Colwell & Graham (1990), and Colwell et al. (1994 — this issue). More than half of the recovered samples contained Jurassic and Triassic pre-breakup shallow-water sediments. This work provided the vital rocks with which the interpretations of seismic profiles could be controlled, and a firmly based geological history has been developed. In 1988, during ODP Leg 122, four continuously cored holes were drilled on the Wombat Plateau, a northern subplateau of the Exmouth Plateau: Sites 759, 760, 761 and 764 (Fig. 1). These four holes all intersected Late Triassic sedimentary rocks directly beneath an unconformity spanning the entire Jurassic: carbonates, including reef complexes (the first of this age found in Australia), and detrital low-energy paralic to fluvio-deltaic sediments (Haq et al., 1990; von Rad et al., 1992).

No petroleum exploration leases are held in either area

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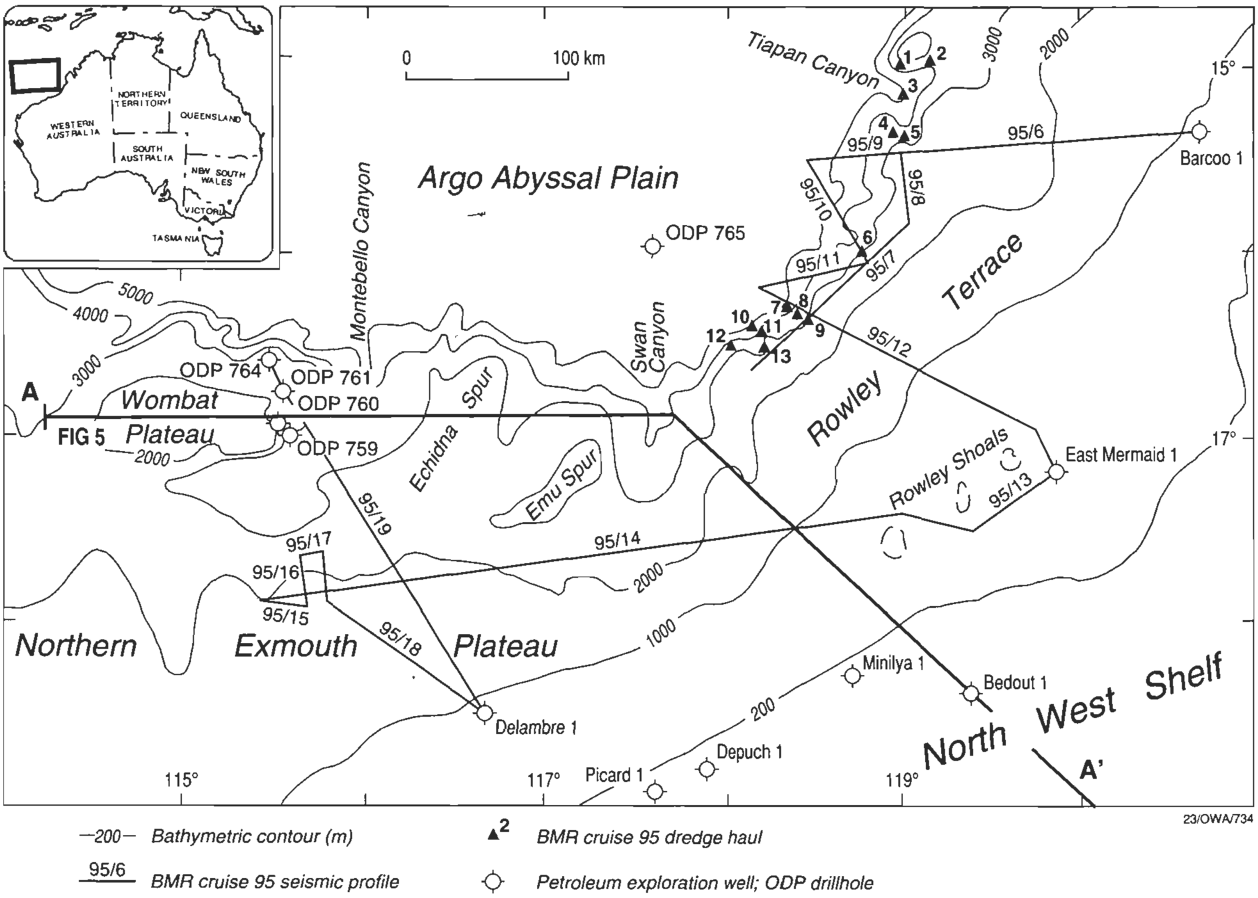

Figure 1. Bathymetric map of study area. Shows locations of exploration wells, ODP drill-holes, and BMR Cruise 95 seismic profiles and dredges. Also shows location of cross-section depicted in Figure 5.

at present. Petroleum exploration started on the northern Exmouth Plateau with multichannel seismic group shoots carried out by Geophysical Services International in 1976 and 1977, some of which were interpreted by Wright & Wheatley (1979). A lease in the northern part of the plateau was held by a consortium headed by Hudbay Exploration Company. Delambre No. 1 (Fig. 1) was dry, and is the only well of relevance to the present study. There has been little petroleum exploration in the Rowley Sub-basin (Fig. 2), a thickening wedge of sediments which underlies the deep-water, oceanward part of the Rowley Terrace. Two dry wells have been drilled: East Mermaid No. 1 and Barcoo No. 1 (Fig. 2).

Both the northern Exmouth Plateau and the Rowley Terrace are underlain by Cainozoic, Cretaceous, Jurassic and Triassic sedimentary sequences, with major unconformities of latest Triassic and approximately Callovian age. The northern Exmouth Plateau is structurally complex, with a series of horsts and grabens formed by northeast to easterly trending faults. The Wombat Plateau is dominated by a thick Triassic sequence, but the spurs farther east are dominated by thick Jurassic sequences. The ENE-trending Rowley Sub-basin lies beneath the Rowley Terrace; the central part of the sub-basin is structurally simple but its margins are complex, being both faulted and gently folded, especially beneath the Cretaceous sequence.

The adjacent Argo Abyssal Plain (Figs 1 and 3) is about 5700 m deep and has been extensively surveyed geophysically (Larson, 1975; Heirtzler et al., 1978; Veevers et al., 1985; and Fullerton et al., 1989). It was drilled during DSDP Leg 27 at site 261 (Veevers et al., 1974) and on ODP Leg 123, at Site 765 (Ludden et al., 1990; Gradstein et al., 1992). It is floored by oceanic crust that formed in Late Jurassic and earliest Cretaceous times, after this part of Gondwana broke up, and has about 1000 m of sediment, largely Cretaceous and Neogene in age (see Buffler, 1994 — this issue).

Present survey

Following the reporting of the discovery of Triassic reefs and their significance as pointing to possible new petroleum exploration targets on the North West Shelf (Williamson et al., 1989), BMR decided to follow up with a research cruise designed to evaluate the seismic character of the reefs on the Wombat Plateau. The cruise would also investigate whether reefs were present in shallower water landward of the Wombat Plateau (where the water depth is around 2500 m — clearly too deep to be of interest to petroleum explorers). The objectives of that cruise, BMR Cruise 95, as reported by Exon & Ramsay (1990), were to:

- Map the extent of any facies changes within the Triassic/Jurassic carbonates in and south of Wombat Plateau, and in the Rowley Sub-basin of the south-western Canning Basin, using seismic profiling and dredging techniques.

- Define the extent, character and petroleum potential of any Triassic or Jurassic reefs found.

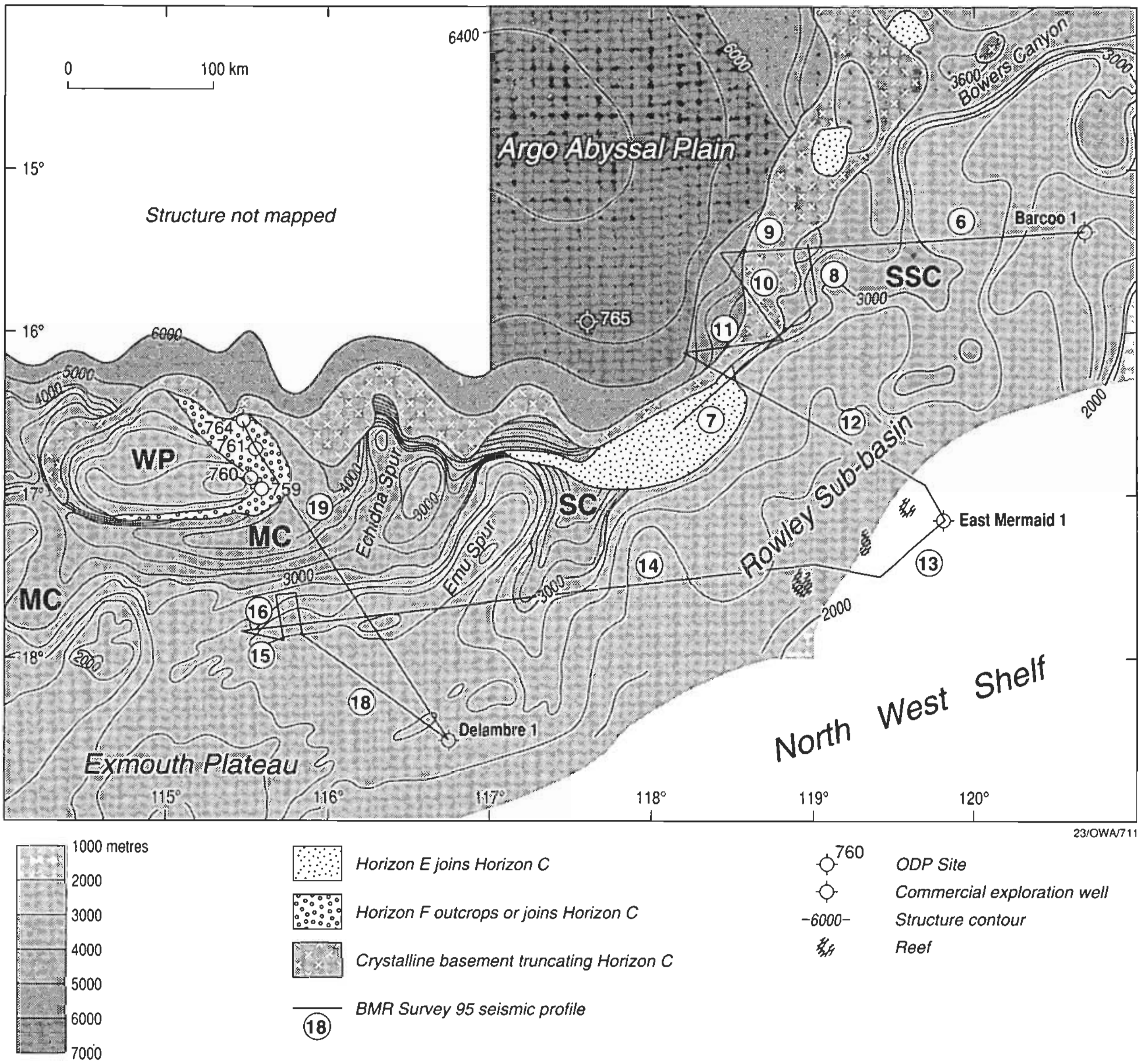

Figure 2. Structure contour map of Horizon C. Compiled from the equivalent maps of Exon & Willcox (1980) west of 118° E, and Stagg & Exon (1981) east of 118° E. Contours in intervals of 200 m above 4000 m and below 6000 m. WP = Wombat Plateau, MC = Montebello Canyon, SC = Swan Canyon, SSC = Seasnake, Turtle & Tortoise Canyons.

- Use the cruise results, along with other data, to help assess the petroleum prospects of Mesozoic reef plays elsewhere on the North West Shelf.
- Better define the geological history of the northern Exmouth Plateau and the adjacent southwest Canning Basin.

The aim of the present paper is to present the results of the regional seismic survey carried out to that end, making use of pre-existing seismic data, information from seabed samples gathered on this and earlier cruises, and data from exploration wells and ODP drill-holes. The extent of the seismic coverage is shown in Figure 4, and the sampling results are reported in detail by Colwell et al. (1994 — this issue). Other results of the research cruise are covered elsewhere in this issue.

Acknowledgments

We wish to thank all those involved in carrying out the cruise: the captain and crew of the R.V. *Rig Seismic*, and AGSO's on-board scientific and technical contingent. Barry Willcox helped us when the seismic interpretation needed another opinion. George Chaproniere and Samir Shafik helped sort out the biostratigraphy and palaeoenvironments in the three exploration wells in the area. Jim Colwell reviewed an early version of the paper; Dick Buffler and Tony Cockbain reviewed a more mature version — all made useful comments, for which we are grateful. We are also indebted to Andrew Warnes who processed the seismic data at AGSO.

Regional setting

Physiographically, the area is quite complex (Fig. 1), with the Argo Abyssal Plain in the northwest separated, by a remarkably steep continental slope, from the Exmouth Plateau in the south and the Rowley Terrace in the east. The continental slope is cut by faults and canyons, and overall it trends east–west along the Exmouth Plateau margin and northeast–southwest along the Rowley Terrace margin. The average water depth on the Argo Abyssal Plain is about 5700 m; the steep continental slope lies between 3000 m and 5500 m; the Exmouth Plateau is 1000–3000 m deep, and the Rowley Terrace is 300–3000 m deep. From west to east along the continental margin, the major bathymetric features are the Wombat Plateau, Montebello Canyon, Echidna and Emu Spurs, Swan Canyon, Taipan Canyon and Bowers Canyon. Many of the canyons — especially Montebello, Swan and Taipan Canyons — are strongly controlled by major faults.

The region was part of the Westralian Superbasin (Yeates et al., 1987) on the southern margin of the Tethyan Ocean in the Mesozoic, and was subject to the period of Gondwanan breakup in the Late Jurassic when the Argo Abyssal Plain was formed. Thus, geological similarities through the region tend to outweigh differences. However, the sedimentary rocks beneath the Exmouth Plateau are regarded as part of the Carnarvon Basin, and those beneath the Rowley Terrace as part of the outer (Mesozoic) Canning Basin. Structurally, there is a great difference between the strongly faulted northern margin of the Exmouth Plateau, and the relatively undeformed Canning Basin, beneath the Rowley Terrace (Fig. 5 and Plate 1C — in pocket: line 95/13–14). Even the structure of the Late Cretaceous sequence, laid down well after most faulting had ended, closely reflects the deeper features. Thus, the major highs and lows along the Exmouth Plateau margin, fault-bounded at depth, are clearly visible. The old structure is still preserved at this level because Cretaceous sedimentation was pelagic and hence did not fill the lows completely first, as detrital sedimentation would have done.

Main seismic horizons and sequences

Many of the regional seismic horizons and sequences defined by Exon & Willcox (1980) are used here (Table 1), but there are some changes, and the definitions are now much tighter. The A1 disconformity apparent in the fossil record obtained from the drill cores (Fig. 6) is not apparent in the seismic data. Several locally important reflectors are not listed in Table 1, but are discussed later in this paper.

The sedimentary megasequences of most relevance to this paper are:

- Late Triassic deltaic sediments and carbonates that are kilometres thick everywhere (seismic sequence F–G: Cockbain, 1989, megasequence Mz 1).
- Early and Middle Jurassic coal measures and carbonates that are 1–2 km thick in most of this area, except in the west and especially on the Wombat Plateau where they appear to be absent (E–F: Mz 2).
- Late Jurassic and earliest Cretaceous deltaic sediments that vary markedly in thickness from 0.5 km to absent (D–E: Mz 3 + part Mz 4).
- Early to Middle Cretaceous shallow-marine detrital sediments including marls that vary from 1 km thick to very thin or absent (C–D: part Mz 4).
- Late Cretaceous shelf carbonates that are generally less than 500 m thick (B–C: Mz 5).
- Cainozoic shelf to bathyal carbonates that are generally 0.5–2 km thick (Seabed–B: Cz 1–4).

The regional geology of the continental margin is well summarised in Figure 5, a simplified line drawing of a seismic profile extending from the northern Exmouth Plateau (west) into the Canning Basin (east) along the margin (Fig. 1). The consistent Jurassic (E–F) thickness indicates that most of the faulting along the margin happened in the Late Jurassic, except on the Wombat Plateau where it possibly may have happened earlier (non-deposition rather than erosion of E–F sequence). The Late Jurassic and earliest Cretaceous sequence (D–E), and the Early to Middle Cretaceous sequence (C–D), are much thicker beneath the Rowley Terrace than beneath the northern Exmouth Plateau. The Late Cretaceous and Cainozoic carbonates are generally thickest in water less than 1000 m deep.

The broad tectonic history of the region has been revealed by well, seabed sampling and seismic data on the continental margin, and by well data and magnetic lineations on the abyssal plain. It can be summarised as follows:

- Stretching and thinning of Gondwana crust in the Permo-Carboniferous.
- Thick deposition in a rapidly subsiding basin in Late

Table 1. Regional seismic horizons and seismic sequences.

Seismic horizon	*Typical characteristics of horizons & sequences where thick sequences occur*	*Common age range*
AP	Unconformity between prograding/downlapping thickly layered to acoustically transparent sequence above, and parallel stratified sequence below	Late Miocene to Early Pliocene
A0	Unconformity between prograding/downlapping well stratified beds above, and gently truncated less-stratified beds below	Late early to early Middle Miocene
A1	Conformable reflector separating stratified beds above, from sequence with hummocky stratiforms below	Late Eocene to Early Oligocene
B	Conformable strong reflector within hummocky clinoforms; those below have more angularity than those above	Late Maastrichtian to Early Paleocene
C	Generally conformable reflector at base of hummocky clinoforms; onlaps older sequence in places	Cenomanian to Turonian
D	Unconformity between well-bedded generally parallel but occasionally onlapping sequence above, and gently truncated moderately stratified sequence below	Tithonian to Hauterivian
E	Unconformity between onlapping moderately stratified sequence above, and truncated well-bedded highly reflective sequence below	Callovian to Oxfordian
F	Low frequency reflector at base of well-bedded sequence; truncation of underlying sequence in places	Late Rhaetian to early Hettangian

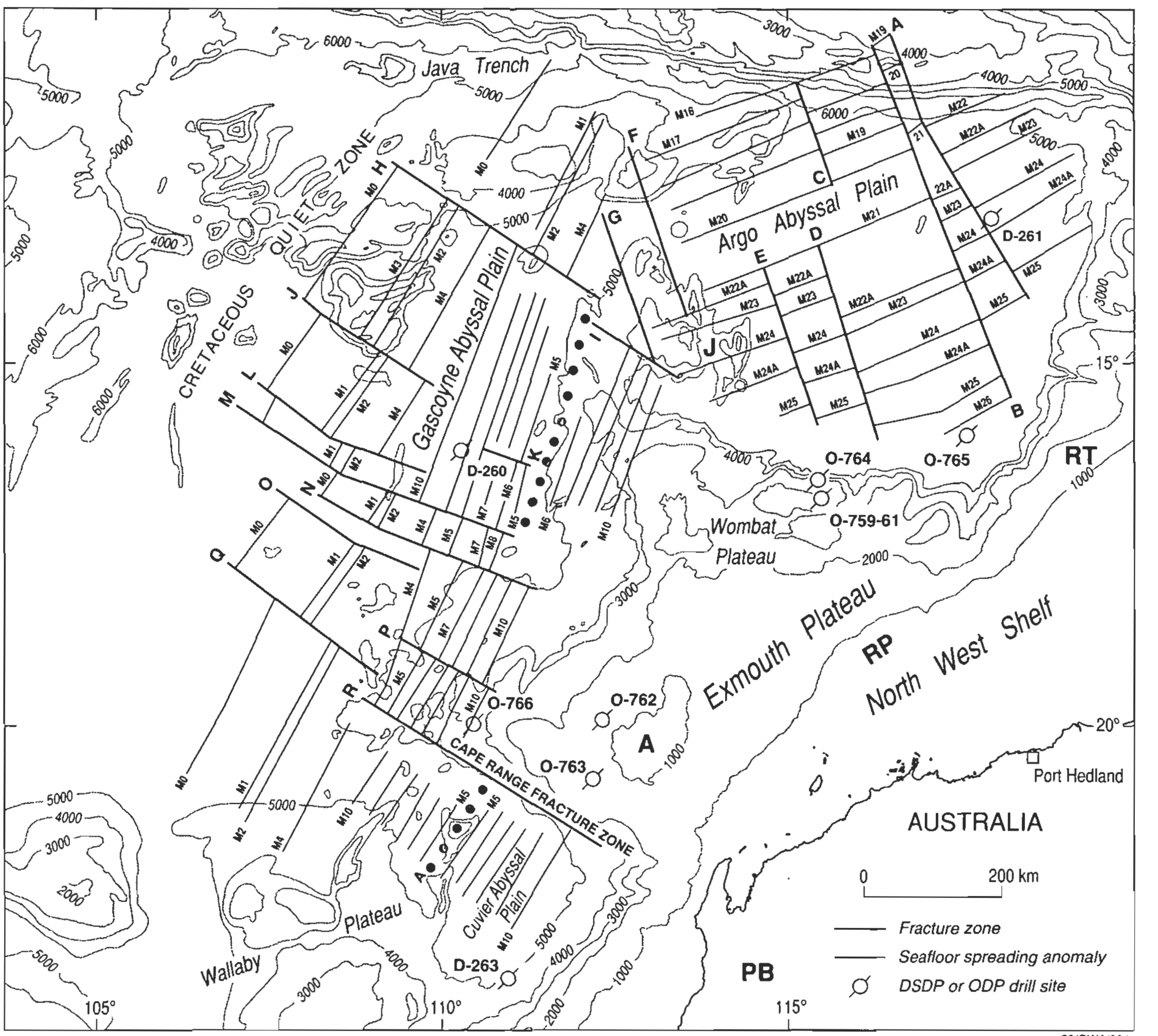

Figure 3. Map of northwest Australian abyssal plain magnetic anomalies. Anomalies after Fullerton et al. (1989). Shows ODP Sites 759–766 (Legs 122 & 123) and DSDP Sites 260, 261 and 263. A = Exmouth Plateau Arch, J = Joey Rise, PB = Pilbara Block, RP = Rankin Platform, RT = Rowley Terrace.

Triassic to Middle Jurassic times, with both normal and transcurrent faulting, striking northeast overall.

- Major uplift and probably largely normal fault movement on the old faults in the Middle Jurassic, just prior to Callovian–Oxfordian continental breakup.
- Post-breakup subsidence of the margin as the spreading axes moved northwestward and the abyssal plain formed.

Stratigraphic sequences and seismic horizons

The aim of this section is to outline the stratigraphic evidence on which the seismic interpretations in this paper are based, and to summarise the relationship between the stratigraphic and seismic sequences. The stratigraphic evidence comes from three petroleum exploration wells, four ODP sites, and a great number of dredge hauls.

Well sequences and unconformities

The three petroleum exploration wells drilled in the region are Woodside Delambre No. 1, in the northern Carnarvon Basin, and Shell East Mermaid No. 1 and Woodside Barcoo No. 1, in the Rowley Sub-basin of the Canning Basin (locations in Fig. 1). These three wells all drilled thick Cainozoic, Cretaceous and Jurassic sequences; and Delambre No. 1 and Barcoo No. 1 also penetrated the Late Triassic (Fig. 6). East Mermaid No. 1 was drilled in 1973 in a water depth of 386 m, Delambre No. 1 well in 1981 in a water depth of 894 m, and Barcoo No. 1 in 1982 in a water depth of 731 m. None had any significant shows of oil or gas [Shell Development (Australia) Pty. Ltd., 1973; Woodside Petroleum Development Pty. Ltd., 1980, 1981].

In Delambre No. 1 (see fig. 2 in Exon, 1994 — this issue), the Cainozoic sequence of shelf and slope carbonates is 1175 m thick. The Cretaceous shelf sequence, grading upward from clastics to calcilutite, is 277 m thick. The Jurassic sequence, largely inner shelf and paralic clastics, is 1941 m thick. The Late Triassic sequence of inner shelf and paralic clastics is 1208+ m thick. Five hiatuses/unconformities are apparent, based on palaeontological evidence.

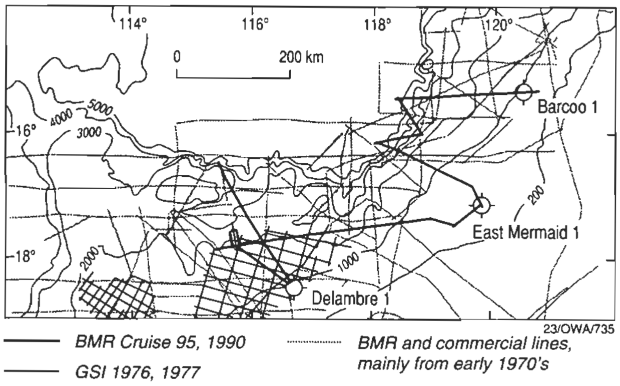

Figure 4. Map of seismic profiles used in this study.

In East Mermaid No. 1 (see fig. 3 in Exon, 1994 — this issue), the Cainozoic sequence of shelf carbonates is 1281 m thick. The Cretaceous sequence grades up from Berriasian deltaic sediments to shelf clastics, marl and calcilutite, and is 1256 m thick. The Jurassic sequence, of fluvio-deltaic, marine deltaic and shallow-marine clastics, is 1185+ m thick. Nine hiatuses/unconformities were identified, again based on palaeontological evidence.

In Barcoo No. 1 (see fig. 4 in Exon, 1994 — this issue), the Cainozoic sequence, largely outer shelf and upper slope calcilutite, is 1971 m thick. The Cretaceous sequence of shelf claystone, grading up into outer shelf calcilutite, is 1094 m thick. The Jurassic sequence, mostly shelf claystone, is about 952 m thick. The Late Triassic sequence, largely shelf claystone and calcareous siltstone, is about 361+ m thick. In this case, seven hiatuses/unconformities are apparent, once again based on palaeontological evidence.

A time-stratigraphic comparison of the three wells (Fig. 6), indicates that five time breaks are present in all three wells, and more occur in the eastern wells. All five can be related to prominent seismic horizons (Table 1), where a seismic line crosses a well (and these seismic horizons can be mapped throughout the region). The diagram indicates that the Triassic to Middle Jurassic sequences are generally similar claystone and sandstone, with minor limestone, but the East Mermaid sequence appears to be less marine. There is a Late Jurassic to Berriasian sequence of thick deltaic clastics in East Mermaid well, and thin shelf marine clastics covering a much shorter time interval in the other wells. The Cretaceous sequence above the Berriasian is quite similar in character in all three wells, with shelf claystone being replaced by limestone with time. However, it is much thinner in Delambre than in the other wells. Within the Cainozoic, the various sequences vary greatly in thickness. In Delambre well, the sequence is dominated by calcilutite, suggesting that the sea was always fairly deep, but in the other two wells calcilutite gives way to calcarenite with time, suggesting a shallowing.

The four fully cored ODP sites on Wombat Plateau give a great deal of information on the Late Triassic sequences (Haq et al., 1990; von Rad et al., 1992). A composite diagram of the thickest sequences from these sites (see fig. 5 in Exon, 1994 — this issue), at that time located much farther offshore than the sites of the three exploration wells, shows how the fluvio-deltaic and deltaic detrital sedimentation of the Carnian and Norian gave way to reefal and lagoonal carbonate sedimentation in the Rhaetian, as the terrigenous supply waned.

Dredge information

A large amount of dredge information is available from the area (see for example von Rad et al., 1990; Colwell et al., 1994 — this issue), and is considered in this interpretation, but the key dredge hauls are the thirteen located on or near the seismic profiles recorded during BMR Cruise 95. These dredge hauls, mostly obtained on the same cruise, are related geographically to the seismic profiles in Figure 7. The results, summarised in Table 1

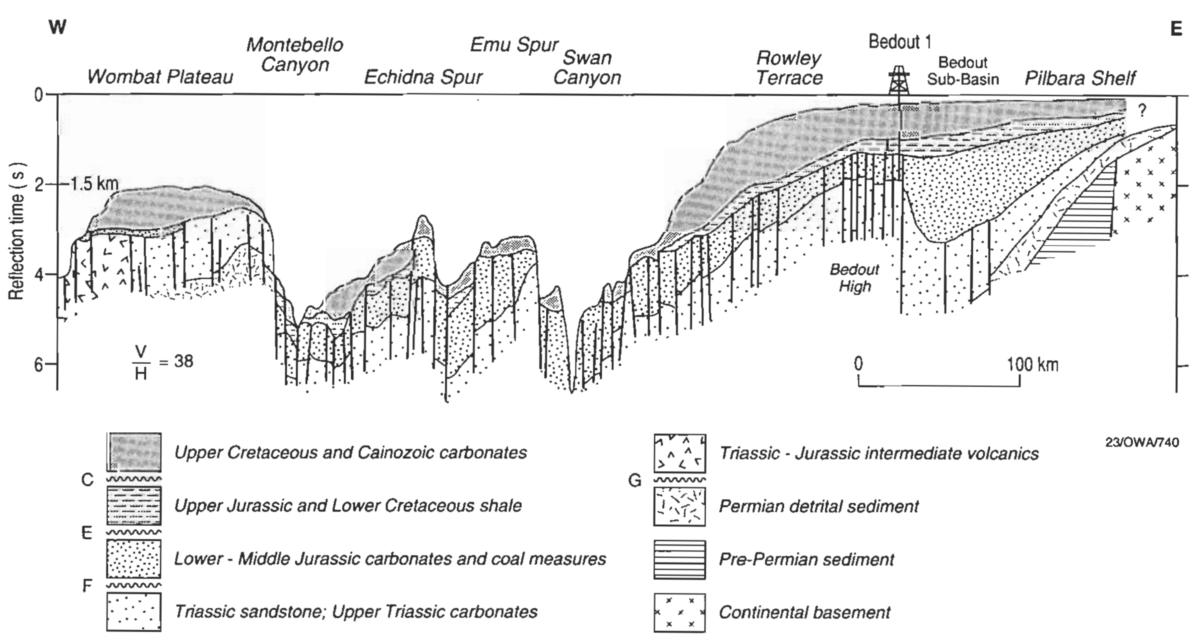

Figure 5. Geological cross-section along northern margin of Exmouth Plateau and extending across Rowley Terrace. Drawn from BMR seismic profile 17–079 and other information; shows structural and bathymetric features. Location shown in Figure 1.

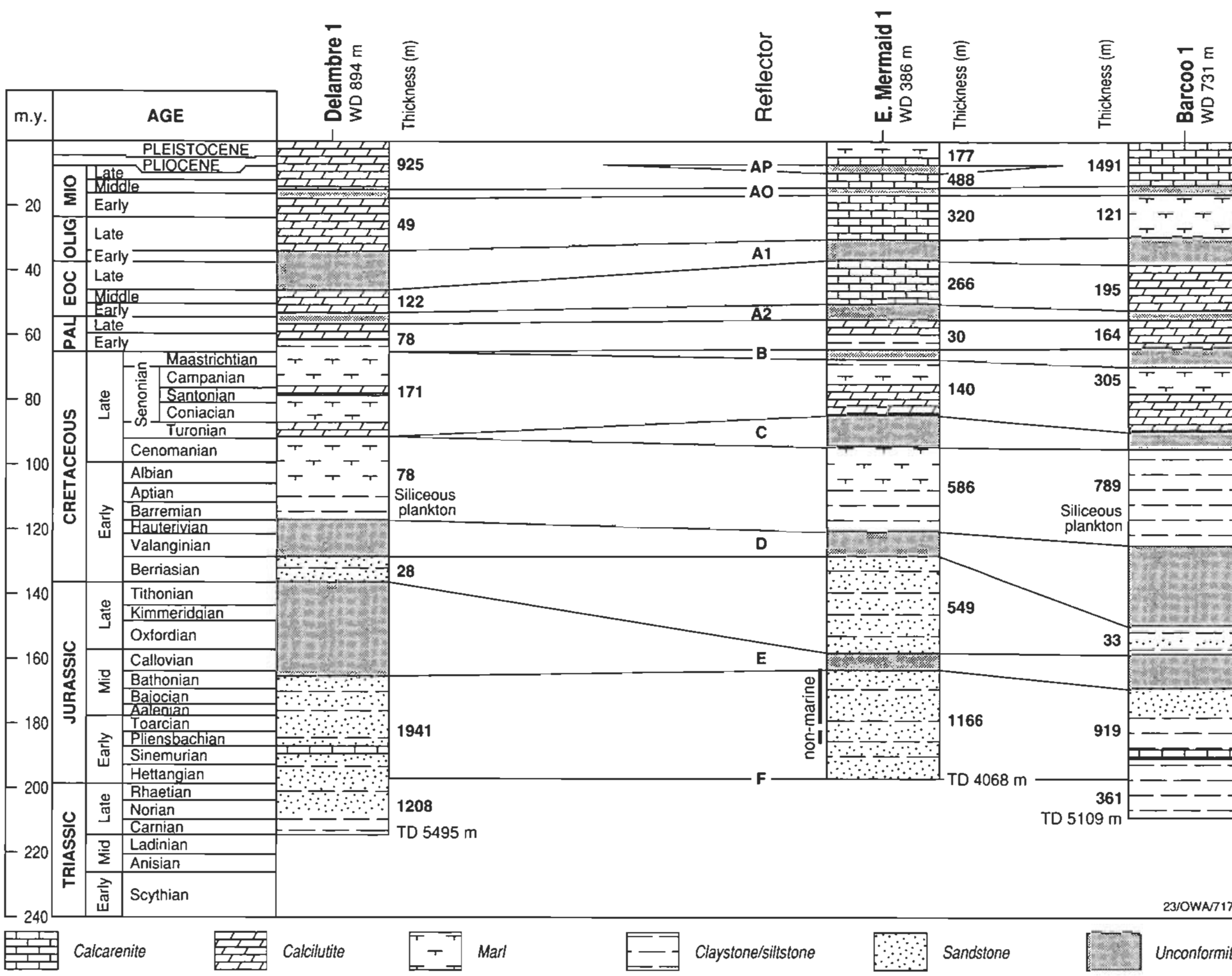

Figure 6. Time-stratigraphic diagram linking Delambre, East Mermaid and Barcoo exploration wells. Note the time breaks apparent in the wells and their relationship to seismic reflectors. Prepared with modifications from well completion reports of Shell Development Australia (1973) and Woodside Petroleum Development (1980, 1981).

in Exon et al. (1994 — this issue), provide critical lithological, age and palaeoenvironmental information about the Mesozoic sedimentary rocks along the outer margin of the Carnarvon and Canning Basins, that is complementary to the well data described above.

Seismic interpretations

Line-drawings of the longer lines recorded on BMR Cruise 95 at a horizontal scale of 1:435 000 are shown in Plate 1 A to E — in pocket; the locations of the lines are shown in Figure 1. The thicknesses quoted in the following section are based on well control, where available, and calculated using average seismic interval velocities elsewhere.

Profile 95/06–9 (Plate 1A) extends from Woodside Barcoo No. 1 well westward to the abyssal plain. The Neogene carbonate sequence (seabed–A1; in the following descriptions 'Neogene' is taken to include upper Oligocene sediments) consists largely of calcarenite in the well, and forms a wedge up to about 1500 m thick which progrades about 150 km westward across the sub-basin, thinning and becoming more bathyal to end as a sequence a few hundred metres thick. The Palaeogene carbonate sequence (A1–B) consists largely of calcilutite in the well, and forms a well-bedded sequence 300–350 m thick across most of the sub-basin, but pinches out near the continental margin. The Upper Cretaceous carbonate sequence (B–C) is well-bedded with small-scale diffractions common, consists of calcilutite and marl in the well, and is generally about 200 m thick. The Lower Cretaceous claystone sequence (C–D/E) is well-bedded and generally 500–800 m thick, also pinching out near the continental margin. Interpretation of these horizons between the well and shot point (SP) 1400 is made difficult by canyons which are still forming today, as evidenced by the rough seabed. The earliest indication of canyoning is a distinct notch in the C horizon, just west of Barcoo No. 1 well, but landward of the present day canyons.

The D–E sequence is present only on parts of this line, and reaches a maximum thickness of about 1000 m in two depocentres on the outer terrace. In other places, as at Barcoo No. 1 well, the D unconformity has truncated the E unconformity. Where present, the D–E sequence appears to be relatively transparent, with generally flat-lying, weak reflectors. The Jurassic, largely claystone, sequence beneath the D or E unconformity is visible all along the line, although its base is not always clearly defined beyond about SP 4000. From SP 2870 to 6500 (a distance of about 90 km) an inferred volcanic sequence (E–V) of presumed Middle Jurassic age overlies Jurassic sediments. The volcanic sequence is characterised by irregular reflective layers, and attenuates the acoustic energy, thereby reducing the continuity in the underlying sediments. It appears to have originated from a feeder or feeders along one or more of the significant normal faults depicted between SP 5700 and 6500, as it thins landwards from these faults. The maximum thickness of the volcanic sequence is about 500 m. Seaward of SP 6500, the Jurassic sequence becomes markedly faulted and folded, but with very little extension evident, and some of the faults clearly indicate strike-slip movement by the geometrical disposition of reflectors on either side. This sequence is interpreted to persist to about SP 1500 on 95/09, beyond which is a zone of indeterminate character which gives way to oceanic crust beyond SP 1600. The abyssal plain has sedimentary cover about 500 m thick on this line.

The top Triassic (F) reflector can be followed seawards from Barcoo No. 1 well, although it becomes less obvious below the interpreted volcanic sequence. The Triassic is again identified beneath the highly structured Jurassic sequence from SP 6900 onwards, and may be followed, gently dipping to a depth of 8 sec TWT, to the continent/ocean boundary.

Faulting is only present west of SP 5700, and appears to be of late Middle or early Late Jurassic age. The marginal high, west of SP 6500, was eroded at that time. A poorly imaged G reflector is mapped between SP 300 and 900 on line 95/09. The significance of this event is not obvious, and it could it be an artefact (sideswipe) since it does not appear to conform to the faulting pattern in the Triassic/Jurassic. If it is real, it could represent a detachment surface.

Profile 95/12 (Plate 1B) extends from Shell East Mermaid No. 1 well westward to the abyssal plain. The seismic line-drawing shows, like profile 95/06-9, a thick wedge of prograding Neogene sediments (seabed–A1) that are dominantly calcarenites in the well, a relatively thin Palaeogene sequence (A1–B) largely of calcarenites and calcilutite, and a very uniform and thin Upper Cretaceous sequence (B–C) of calcareous claystone and calcilutite. Probable drowned reefs are visible within the Miocene sequence at SP 4900 and especially at SP 5450, which is in line with the Rowley Shoals reefs. The Lower Cretaceous shelf mudstone sequence (C–D), 600 m thick in the well, onlaps the pre-Hauterivian sequence west of SP 3300 and thins to nothing at the top of the steep continental slope. The Upper Jurassic deltaic detrital sequence (D–E) downlaps the E–F sequence westward between SP 3200 and 3900, and west of SP 3200 it onlaps the underlying sequence, finally pinching out at SP 1900.

The Lower and Middle Jurassic sequence (E–F) consists of shelf and deltaic shale and sandstone in the well, and downlaps the F (top Triassic) unconformity westward. Aided by dredging results, we have interpreted a facies change at reflector E1 within the seaward part of the Jurassic sequence (E–F), separating strongly reflecting coal measures above from relatively more transparent carbonates below. The coal measure sequence (E–E1) thickens rapidly westwards, onlaps the underlying sequence (E1–F) in places, and has been eroded in the upper Middle or lower Late Jurassic (horizon E) and earliest Cretaceous (horizon D).

East Mermaid No. 1 well did not penetrate Triassic rocks, bottoming in lowest Jurassic, but what we interpret to be the F unconformity can be traced seawards from just below the well, showing very little structuring until about SP 2200. From here on, normal faults, of less than 100 m displacement, occur in the well-bedded Triassic and Jurassic sequences, normally terminating at the E horizon. An intra-Triassic horizon (F1) can also be traced northwestwards from the well. Whether it is the same horizon as either of the intra-Triassic levels correlated by drilling on the Wombat Plateau portion of line 95/19 is not known. At about SP 1100, the F horizon drops over 2000 m on a large normal fault, marking the steep slope down to the abyssal plain. On the down-thrown side of this fault, the F horizon can be traced as far as SP 800, where it disappears under an interpreted volcanic flow. Whether this flow marks the continent/ocean boundary, or whether intruded sedimentary rocks persist westwards cannot be distinguished. Oceanic crust is present at the end of the line, underlying a maximum of about 500 m of sediments.

Profile 95/13-14 (Plate 1C) provides a major transect from the Rowley Terrace westward onto the Exmouth Plateau; the line also ties East Mermaid No. 1 well to the 1976 GSI regional seismic lines (Fig. 4). That portion of the profile between East Mermaid No. 1 well and about SP 4600 corresponds with the Rowley Terrace, and shows a thick (up to 1500 m) prograding wedge of Neogene sediment (seabed–A1), overlying a more uniform thickness (100–300 m) of Palaeogene sediment (A1–B). A uniform, thin Upper Cretaceous sequence (B–C) underlies the Palaeogene, which in turn is underlain by a wedge of Lower Cretaceous sediments (C–D), thinning westwards from more than 500 m to less than 100 m. Another wedge (D–E) of Upper Jurassic to Berriasian sediments onlaps the Middle Jurassic sequence westwards, pinching out at SP 1750. The Lower to Middle Jurassic sequence (E–F), unlike the younger sequences above, thickens westward, from nearly 1000 m at the well, to more than 2000 m. The top of the E–F sequence has been eroded in Middle Jurassic times west of SP 1750. The majority of faulting occurs west of SP 1650 and consists of a mixture of normal and strike-slip motion. Most faults do not extend upwards past the D horizon, except in the east, where there are two faults which appear to be Pliocene in age (like the single fault visible near the well on profile 95/12). The F horizon can be followed only with difficulty and not at all beyond SP 3350 over this section of the line, and the F1 horizon, clearly visible in profile 95/12, cannot be distinguished here. The Swan Canyon, between SP 4600 and 5400 on this profile, marks the boundary between the Rowley Terrace and the Exmouth Plateau, and the Jurassic blocks step down about 1000 m to the west beneath the Swan Canyon and its feeders. This and other canyons in the same general area may be controlled by northwest-trending strike-slip faults.

The part of the profile from SP 5400 westwards transects the northwestern part of the Exmouth Plateau near and parallel to the North Exmouth Hinge Zone (NEHZ), and shows what could be regarded as a typical Exmouth Plateau sequence. We see dipping and folded Triassic and Lower to Middle Jurassic rocks cut by numerous normal faults, giving rise to a complex of northeast-trending horsts and grabens, which terminate at the E, D, or sometimes C horizon. The F horizon becomes visible again from about SP 8000 onwards, and is seen to rise considerably towards the western end of the line. From SP 9400 to the end of the line, horizon F1 is also intermittently traceable.

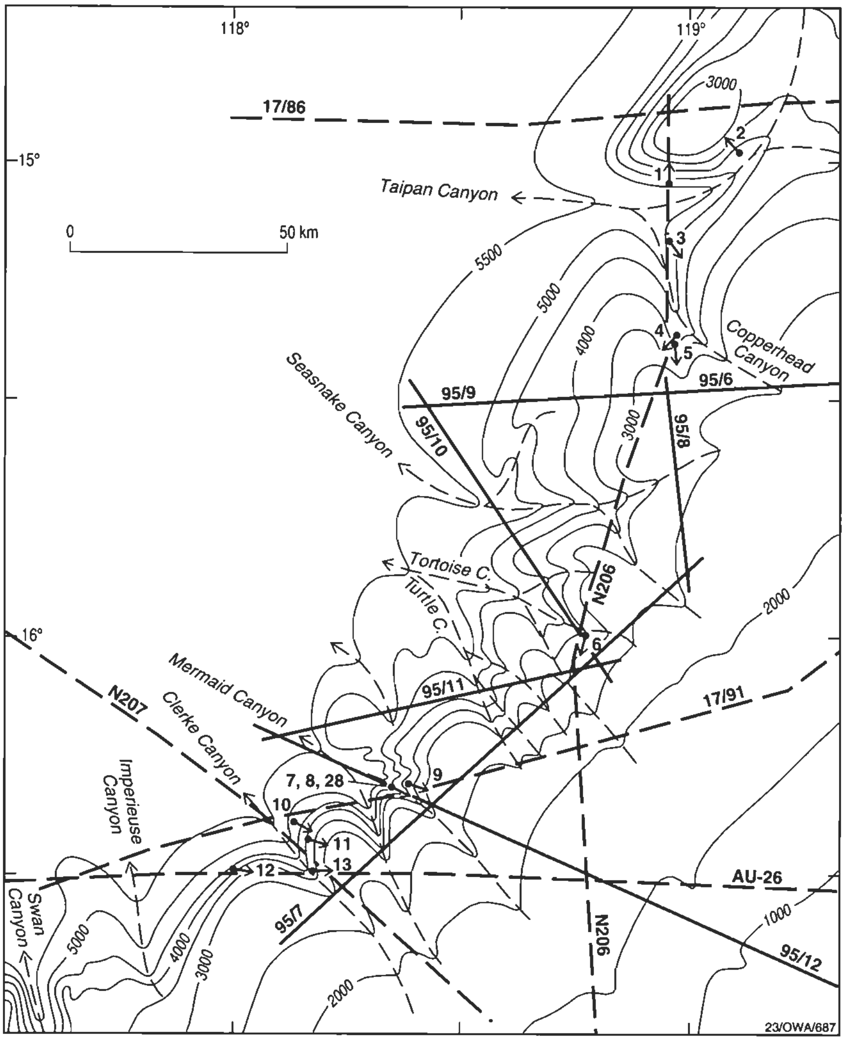

Figure 7. Bathymetric map of the Rowley Terrace. Shows canyons and relationship of BMR Cruise 95 dredge hauls to seismic profiles.

The Neogene sequence is highly canyoned and varies in thickness from around 150 to 600 m. The Palaeogene sequence retains its thickness overall, at between 100 and 200 m. The Upper Cretaceous is also generally 100 to 200 m thick, but onlaps some of the major horst blocks, on which it was not deposited. The Lower Cretaceous shallow-marine detrital sequence (C–D) reaches a maximum thickness of about 500 m, onlaps the older sequences, and is gently truncated by the Upper Cretaceous in places. This suggests that this sequence filled pre-existing depressions, and never covered the entire area. Similarly, the Upper Jurassic/lowest Cretaceous detrital sequence (D–E) reaches a maximum thickness of about 300 m, and onlaps the Lower to Middle Jurassic sequence, filling depressions in this surface. Its distribution is even more restricted than the C–D sequence. The Lower to Middle Jurassic sequence (E–F) is seen to be thinnest over structural highs (it is reduced to no more than 200 m at around SP 11250), and to thicken into the major structural low centred at SP 6500, where it is estimated to be more than 2000 m thick. The faulting on this part of the line mainly terminates at the E horizon, showing that the major period of tectonism was during late Middle or early Late Jurassic times. Occasionally, faults penetrate to the D level, perhaps showing some rejuvenation of faulting in the earliest Cretaceous.

Profile 95/19 (Plate 1E) extends northward 260 km from Woodside Delambre No. 1 well, across the outer North West Shelf, down from the North Exmouth Hinge Zone (at SP 2450) into the Wombat half-graben and up onto the Wombat Plateau, passing through ODP sites 761 and 764. The line-drawing shows a complex series of horsts and grabens with Triassic and Jurassic sediments, with most of the faults terminating at the E horizon. Between the well and the NEHZ, the top Triassic (F) horizon remains around 4 s TWT, and consequently the Jurassic sequence maintains a fairly constant thickness of around 1800 m. Between the NEHZ and the Wombat Half-graben, there is more intense faulting showing both slightly extensional and compressional features, and the F level falls gradually by around 2000 m. The southern flank of the Wombat Plateau is formed by a massive fault, north of which the F horizon rises to its highest point on the entire line.

The Neogene carbonate sequence (seabed–A1) consists of calcareous claystone and calcilutite in Delambre No. 1 and nannofossil ooze in the ODP sites. It forms a well-bedded, northerly prograding wedge up to 1000 m thick on the North West Shelf, but is generally much thinner and more poorly bedded elsewhere. The Palaeogene sequence (A1–B) consists of calcilutite and calcareous claystone in Delambre No. 1, and nanno chalk grading to ooze at the ODP sites. It is well-bedded and parallel bedded, and up to 300 m thick on the North West Shelf,

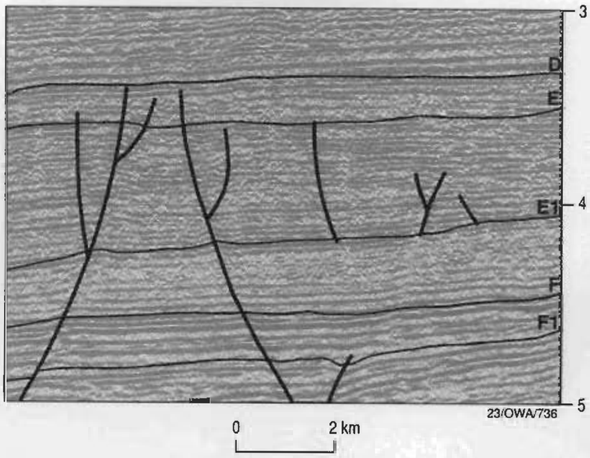

Figure 8. Contrasting seismic sequences within the Lower to Middle Jurassic megasequence (E–F) in the outer Rowley Sub-basin. The strongly bedded upper sequence (E–E1) is interpreted as coal measures and the less well bedded lower sequence (E1–F) as shallow marine carbonates and detrital sediments. (Profile 95/12, SPs 2100–2400.)

but more disturbed and thinner elsewhere. The Upper Cretaceous sequence (B-C) consists of marl and calcilutite in Delambre No. 1, and of chalk at the ODP sites. It is well-bedded with small-scale diffractions on the North West Shelf where it is up to 300 m thick, but elsewhere is more disturbed and thinner. Northwest of SP 3600, the Cainozoic and Cretaceous sequences are irregularly deposited and eroded, and therefore highly variable in thickness. The C unconformity is a major regional feature, which is almost planar on the Wombat Plateau where it truncates all sequences above the top Triassic.

The Lower Cretaceous (C–D) sequence, consisting of claystone, sandstone and marl in Delambre No. 1, has filled structural depressions, onlaps the older sequences on highs, and reaches a maximum thickness of about 200 m. In a similar manner, the Upper Jurassic to Berriasian (D–E) sequence is restricted to depositional lows, onlaps adjacent highs and has a maximum thickness of about 500 m. On the Wombat Plateau, the C–E sequence is represented by a very thin transgressive sand in ODP Site 761, sitting directly on Triassic rocks and grading rapidly upward to chalk.

The Lower and Middle Jurassic (E–F) sequence consists of shallow-marine sandstone, siltstone, claystone and a little limestone in Delambre No. 1 well, and is absent along this profile on the Wombat Plateau. Dredging farther east on the northern Exmouth Plateau margin shows it to consist of paralic detrital sediments and shelf carbonates. This well-bedded sequence is about 1900 m thick in Delambre No. 1 well, and maintains this thickness, on average, as far as the NEHZ. It becomes more complexly faulted beyond the NEHZ, and gradually thins as it drops into the Wombat Half-graben. The upper surface of this sequence has been slightly eroded in late Middle Jurassic times, but nowhere has it been planated, as it is farther east in places under the Rowley Terrace. This sequence is entirely absent on the Wombat Plateau. Within this sequence we interpret an isolated bright reflector, around SP 4300, as a volcanic intrusion.

In Delambre No. 1 well, the pre-F sequence consists of 1200 m of Upper Triassic deltaic claystone and sandstone. At the ODP sites, it consists of deltaic Carnian and Norian sediments with a few interbedded carbonates, and Rhaetian carbonates including reefs. On the Wombat Plateau, two intra-Triassic reflectors (F2 and F3) are mapped. From correlations with the ODP sites, these are identified as the Rhaetian/Norian and Norian/Carnian boundaries, respectively. Under the Wombat half-graben, between SP 4000 and 5100, a weak but persistent reflector (G) is recognised. This may represent the top Permian or an intra-Triassic reflector.

The whole of the line is subject to intense faulting trending northeast or east-northeast, of both normal and strike-slip character. Displacement on the faults is generally of the order of several hundred metres, at both top Triassic (F) and upper Middle Jurassic (E) levels, but is substantially greater on some faults (e.g. at SP 500 and at SP 4100), where the displacement at top Triassic is greater than at upper Middle Jurassic level. In general, the faulting does not penetrate beyond the E level, suggesting that tectonism culminated at this time.

Profile 95/18 (Plate 1D) provides a tie between profile 95/14 near the North Exmouth Hinge Zone (NEHZ) and Delambre No. 1 well, and also cuts across several of the 1976 GSI regional seismic lines (Fig. 4). This line is oriented northwestward and is very similar to that part of the previous northerly oriented profile (95/19) between the well and the NEHZ. Seaward from Delambre No. 1, the Jurassic sequence maintains a fairly even thickness of 1600–1800 m, with the top Triassic level lying at around 4 seconds TWT. Towards the margin of the Exmouth Plateau, between SP 250 and 1000, block faulting becomes more intense, with the top Triassic marker rising to less than 3 secs TWT, and the overlying Jurassic section being reduced to probably less than 500 m. On a similar high block, which crops out at the seabed a few kilometres to the west, the youngest section dredged was an earliest Jurassic, shallow-water carbonate mound facies (Kristan-Tollmann & Colwell, 1992). This confirms that much of the Lower and Middle Jurassic is missing on these high blocks. The similarity of the E-F sequence in both horsts and grabens suggests that most of the faulting was post-depositional, and hence of late Middle or early Late Jurassic age. The Triassic and Jurassic sections on this part of the profile are gently folded in part, and a regional high occurs centred on SP 550, similar to that seen on profile 95/14 at about the same location.

Overlying the faulted Mesozoic section are sequences D–E and C–D, which are very similar in composition and distribution as in profile 95/19, described above. Overlying these is a Cretaceous marine carbonate transgressive sequence of nearly constant thickness (B–C), which in turn is overlain by a 200 m thick layer of Palaeogene (A1–B). The uppermost layer is a prograding wedge of Neogene (seabed-A1) sediment, with a maximum thickness of about 1000 m, that thins rapidly towards the plateau margin.

Discussion

Within this study area, we observe considerable differences in the structure and style of the continental margin, reflecting in part the differing stress fields that operated, particularly around the times of continental break-up. The Argo oceanic crust first started forming around Callovian time, with a continental fragment, the 'Argo Landmass' of von Rad et al. (1992), moving off in a NNW direction

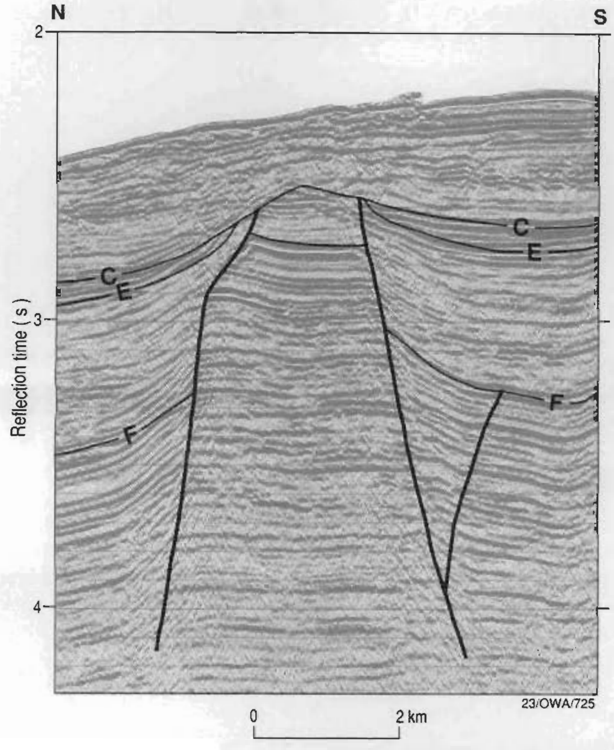

Figure 9. Lower Jurassic sequences atop fault blocks of Triassic sediments on the northern Exmouth Plateau. These sequences have been dredged at one locality where they proved to consist of carbonate banks. An unsolved question is whether the carbonates formed as buildups only on pre-existing horsts, or whether they formed as banks on an even surface that was later faulted. Reefs, which are more prospective as petroleum reservoirs, are more likely to be present in the former case than in the latter. (Profile 95/16, SPs 370–570.)

to be accreted finally to Laurasia. On the other sides of the Exmouth Plateau, the initiation of the Gascoyne and Cuvier oceanic crust dates from around Valanginian time, with that piece of Gondwana having moved off in a more WNW direction, according to the interpretation of the magnetic anomalies (Fig. 3). It is therefore not surprising that the Exmouth Plateau, caught between these two major tectonic events, shows very intense deformation.

In addition to the lines described in the previous section, we have reviewed all the other relevant seismic data within our area of interest. These lines are shown in Figure 4. In this discussion, we concentrate on new results coming from our studies.

Continental basement

Stagg & Exon (1981) mapped continental basement along the outer margins of the study area, and this was incorporated in a map of the tectonic elements of the North West Shelf (Stagg, 1993). Stagg & Exon, in their plates 16, 24 and 25, showed a structural high of Precambrian and Palaeozoic rocks below and near the Scott Plateau, 0–3 km below the seabed and beneath the Callovian breakup unconformity. This extends into the northern part of our study area. They also showed a basement high of pre-Jurassic age just east of the Swan Canyon. Stagg (1993) re-interpreted this basement high to be of volcanic origin, based on seismic character. Stagg & Exon (1981) interpreted seismically opaque rocks beneath the Scott Plateau as acoustic basement, which they equated with plutonic and metamorphic rocks like those of the Kimberley Block to the east. Such basement rocks have also been drilled in offshore wells along the eastern margins of the Browse and Canning Basins. Stagg & Exon (1981) indicated that continental basement should crop out at a number of places on the margins of the plateau, where it could be dredged. Three subsequent dredging programs along the margin have yielded basic and intermediate volcanic rocks, believed to be interbedded with Mesozoic sediments, but no plutonic or metamorphic basement rocks (von Stackelberg et al., 1980; Exon & Ramsay, 1990; Colwell & Graham, 1990). Our higher quality line 95/06 shows the clearest evidence for a volcanic flow sequence, apparently emanating from one of a number of faults just east of the uplifted marginal block. We assume that this is typical of the appearance of the 'basement high', as previously interpreted on the older seismic sections. In conjunction with the dredge results, this indicates that continental basement is far deeper than was interpreted by Stagg & Exon (1981).

Triassic sequence

Well and dredge information discussed earlier in this paper prove that Triassic sedimentary rocks are widespread in the area. On the basis of our seismic lines, we believe that these rocks exist everywhere on the continental margin (Plate 1 — in pocket). Our interpretation of this sequence indicates a number of unconformities within it: ODP drilling on the Wombat Plateau confirms that our reflectors F2 and F3 are unconformities within the Upper Triassic sequence. Most of the faulting affecting the Triassic sequence apparently occurred in the Jurassic. It is also clear that faulting was much more intense and widespread on the northern Exmouth Plateau and the northwestern edge of the Rowley Sub-basin than elsewhere. We have no seismic evidence for the presence of reefal buildups in the Triassic, except on the Wombat Plateau. However, Stanley (1994 — this issue) has identified Late Triassic reef-forming corals from two dredge hauls (95DR/12 and 95DR/7, Fig. 1) along the outer edge of the Rowley Terrace.

Lower and Middle Jurassic sequence

In the three wells in the south and east of the study area, the Lower and Middle Jurassic sequence is 1000 to 2000 m thick (Figs 1 and 6). However, on the Wombat Plateau, ODP drilling and dredging has not recovered any indisputably Jurassic rocks. Our seismic lines show that the thickness of the equivalent E–F seismic sequence on the inner Rowley Terrace is of the order of 1000 m, thickening to at least 2000 m along the western margin of the terrace. This thickening persists westwards onto the northern Exmouth Plateau (Fig. 5; Plate 1C, line 95/13–14), except above a Triassic high south of the eastern Wombat Plateau, where it is 200–1000 m thick. Dredge evidence suggests that the thinning is caused by non-deposition or erosion of the Middle Jurassic sequence (Kristan-Tollmann & Colwell, 1992). The Wombat Plateau, a unique high block within the study area, has no Jurassic rocks preserved. One possible cause would involve volcanic underplating, in order to thicken the crust, and then keep it high during the Mesozoic.

Dredge information shows that there is a thick Lower Jurassic shelf carbonate sequence along the outer continental margin, overlain by a thick Middle Jurassic coal-measure sequence (reviewed by Colwell et al., 1994 — this issue). We have two seismic facies on the western end of line 95/12 (Plate 1B and Fig. 8) that appear to correspond to these two lithofacies. The lower seismic facies is generally well-bedded, but with some poorly reflecting intervals which give the facies a moderately transparent look overall. The upper seismic facies is more strongly reflecting and onlaps the lower sequence to the east. It is cut by numerous faults, many of which are confined to this sequence. We interpret the lower sequence as consisting of relatively homogeneous shelf carbonates and marls, and the upper sequence as consisting of detrital sediments and coal. The small-scale faulting within the upper sequence is characteristic of coal measures, and forms during dewatering and compaction.

Exon et al. (1991) identified presumed large carbonate buildups on horsts on the northern Exmouth Plateau in the Lower and Middle Jurassic sequence, using older seismic data and early versions of our Cruise 95 data. Dredging of one of these buildups yielded Early Jurassic limestone of carbonate bank facies (Kristan-Tollmann & Colwell, 1992). From our seismic data (Fig. 9), we cannot be sure whether reefal buildups formed on high blocks, or alternatively, whether limestone banks were widespread on a flat Triassic surface, and subsequently cut by faulting into horsts and grabens. Reefal buildups on high blocks would probably be more porous and permeable than carbonate banks, and hence more likely to be petroleum reservoirs.

The Lower and Middle Jurassic sequence, like the Triassic sequence, is strongly faulted on the northern Exmouth Plateau and on the northwestern margin of the Rowley Sub-basin.

Tectonism: late Early to early Late Jurassic

Seismic evidence from most of the southern North West Shelf shows that there was a major period of tectonism at or near Callovian time, represented most clearly by a major unconformity (E in our terminology). This unconformity has been dated palaeontologically in the three wells in our area (Fig. 6). In the eastern wells, it covers a relatively short time interval, but in Delambre No. 1 it covers the entire Late Jurassic. We presume this unconformity is related to thermal uplift preceding the breakup of Gondwana that formed the Argo Abyssal Plain.

The areas of most uplift were around the continental margin and were extensively faulted, the faults showing both strike-slip and normal movement. These areas were then eroded and the relief across the faults was reduced. Some areas were bevelled by wave action but, in others, fault blocks were not completely eroded. The bevelled area is about 50 km wide, lying landward of the steep continental slope, along the western Rowley Terrace (Fig. 10), and along the high blocks on the northernmost

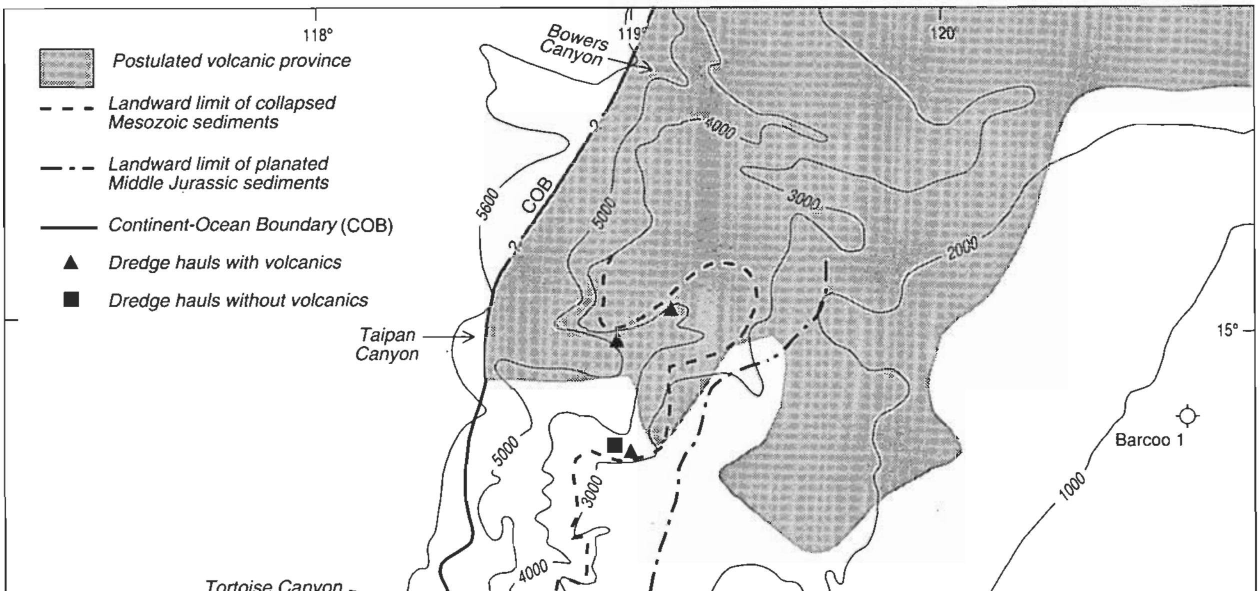

Figure 10. Map of the Rowley Sub-basin. Shows continent-ocean boundary (COB), zone of collapsed Mesozoic sediments between COB and high-standing Mesozoic sequence of the continental margin, and zone of remarkably planated Middle Jurassic sediments (eastward this planation becomes imperceptible as the angular relationship at the E unconformity disappears). The map also shows the presence or absence of Mesozoic volcanics in dredge hauls, and the extent of Middle Jurassic volcanics as interpreted from seismic sections.

Exmouth Plateau. The unconformity, where well developed, separates a lower well-bedded, highly reflective sequence which is frequently truncated, from an upper onlapping, moderately stratified sequence.

Dredges along the margin, taken on BMR Cruises 95 and 96, show that some areas contain abundant volcanic rocks, while others none or very few (Colwell et al., 1994 — this issue). Summary information for these dredges is shown in Figure 10. North of Taipan Canyon, volcanic rocks are common (e.g. Hinz et al., 1978); between Taipan and Clerke Canyon, they are present only in subsidiary amounts. They have not been recovered from the Swan Canyon (von Stackelberg et al., 1980; Exon et al., 1988). We interpret volcanic rocks on our line 95/06, where the volcanic sequence appears as a rough-surfaced, highly-reflective layer with discontinuous reflectors (Fig. 11). The sequence reaches a maximum thickness of about 500 m, and extends about 90 km, thinning to the east (Plate 1A). The layer is cut by faults which terminate in the overlying sequence (D–E). There is still some bedding apparent in places below this layer, but in general it is more obvious where the volcanic layer is not present. We interpret this layer to represent flows that have fed up faults; these faults have continued to move for at least a short time after the flows were laid down. We would assume that volcanic rocks dredged at sites 95/1 and 95/2 (Figs 1 and 10) belong to this interpreted sequence.

Using the characteristics best seen in line 95/06, we have generalised a map showing where it appears that volcanic flows and volcaniclastics underlie the E unconformity.

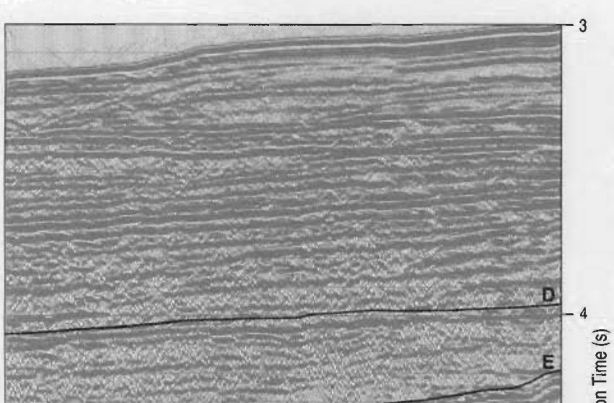

Figure 11. Typical expression of interpreted Middle Jurassic flows and volcaniclastics (E–V). Note rough top surface, discontinuous high amplitude reflectors, and suppression of reflections in underlying Mesozoic sediments. (Profile 95/06, SPs 5100–5400.)

As the older lines are generally of poorer quality, we have mapped the volcanic layer on the basis of a rough appearance and a suppression of underlying reflectors; this is by no means a definitive interpretation. The area mapped extends eastwards from the continent/ocean boundary roughly 150 km, and covers an area of about 25 000 km^2 on the northwestern edge of the Rowley Sub-basin (Fig. 10). On the evidence of line 95/06, we consider it probable that most of the vents were in the west in the uplifted area, and volcanics and volcaniclastics flowed down the slope to the east, thinning in that direction. It is possible that the volcanic flows are interbedded with normal detrital layers in places. Seismic stratigraphic evidence suggests that the age of the flows is immediately pre-breakup, i.e. Callovian–Oxfordian. One dredged calcite-cemented volcaniclastic grit (95DR/02) has been dated as Oxfordian by David Lynch (*in* Exon & Ramsay, 1990). Crawford & von Rad (1994 — this issue) date dredged basalt samples from the Rowley Terrace margin as Callovian–Oxfordian. They find the composition of the samples to be remarkably consistent, being that of mid-ocean ridge basalt modified by penetrating continental crust. Such rocks could well be late rift volcanics. The same authors also find a rapid change in magma composition to the rocks recovered from drilling at ODP Site 765, only about 100 km to the west, on the Argo Abyssal Plain.

Callovian–Oxfordian breakup

Studies of the Argo Abyssal Plain show that breakup occurred at Callovian–Oxfordian time. The age of the oldest oceanic crust was dated as 155 Ma ('Callovian' in the Haq et al. (1987) time scale) on a basaltic hyaloclastite directly overlying basaltic basement in ODP Site 765 (Gradstein et al., 1992). The oldest palaeontologically datable overlying sediments are of Tithonian age. The oldest magnetic anomaly was identified as M26 (Oxfordian) by Fullerton et al. (1989), and by Sager et al. (1992), in agreement with the drilling results. Thus, the age of breakup as dated on the abyssal plain corresponds well with the dating of the E unconformity on the continental margin, indicating that the term 'breakup unconformity' is appropriate.

Our interpretation of the position of the continent–ocean boundary (COB) is shown in Figure 10. This has been deduced entirely from study of seismic sections which go right down onto the abyssal plain. We have identified the outermost position of what we regard as sedimentary reflectors which look similar to those on the high-standing part of the continental margin. The position mapped is imprecise in most cases, since the quality of the reflectors is always reduced under the steeper and deeper parts of the margin as compared to the case in the high blocks. Stagg (1993) showed the COB to be of a similar orientation but some 10 to 20 km landwards of that shown in Figure 10, using the interpretation of Veevers et al. (1985), which was based on the recognition of a prominent positive magnetic anomaly along the margin. This anomaly was interpreted to be caused by a complex of rift-related dykes. Recent dredging (see for example Fig. 10) proves that continental rocks are present seaward of Veevers et al's COB.

Late Jurassic to earliest Cretaceous sequence

In the three wells in the south and east of the study area, the Late Jurassic to Berriasian sequence (earliest Cretaceous) is 30 to 500 m thick (Figs 1 and 6). On the

Wombat Plateau, ODP drilling and dredging revealed no, or very thin, sediments of this age. Our mapping of the equivalent D–E seismic sequence shows that its occurrence is areally variable and that it is commonly preserved in structural lows (Plate 1 — in pocket). It onlaps the bounding highs, showing that its deposition was governed by the pre-existing structure. The well evidence (Exon, 1994 — this issue) indicates that it consists of deltaic and shallow-marine detrital sediments, laid down as the margin started to subside following breakup.

Post-breakup history

Evidence from the ODP sites on the Wombat Plateau indicates that the marine transgression related to breakup did not occur until the Berriasian (von Rad et al., 1992), suggesting that the D–E sequence was deposited before rapid subsidence commenced, as the continental margin and the oceanic crust cooled and sank. This evidence is supported by the fact that the D–E sequence is shallow marine and largely restricted to structural lows. This apparent delay in the onset of subsidence following breakup may be explained by volcanic underplating and buttressing of the margin. Presumably, the associated deep heat source must have decayed only slowly. Marine erosion appears to be represented by the Valanginian D unconformity, and the overlying sediments are holomarine, laid down as the margin subsided. The most spectacular result of the cooling and subsidence of the oceanic crust was the formation of the collapsed zone along the continental margin between high-standing continental material and the COB. This zone averages 30 to 40 km wide (Fig. 10) and the total vertical displacement of the pre-breakup sequence is of the order of 2000 m (e.g. west of SP 7500 on line 95/06-09, Plate 1A). The collapse occurred on a series of faults, many of which probably pre-dated

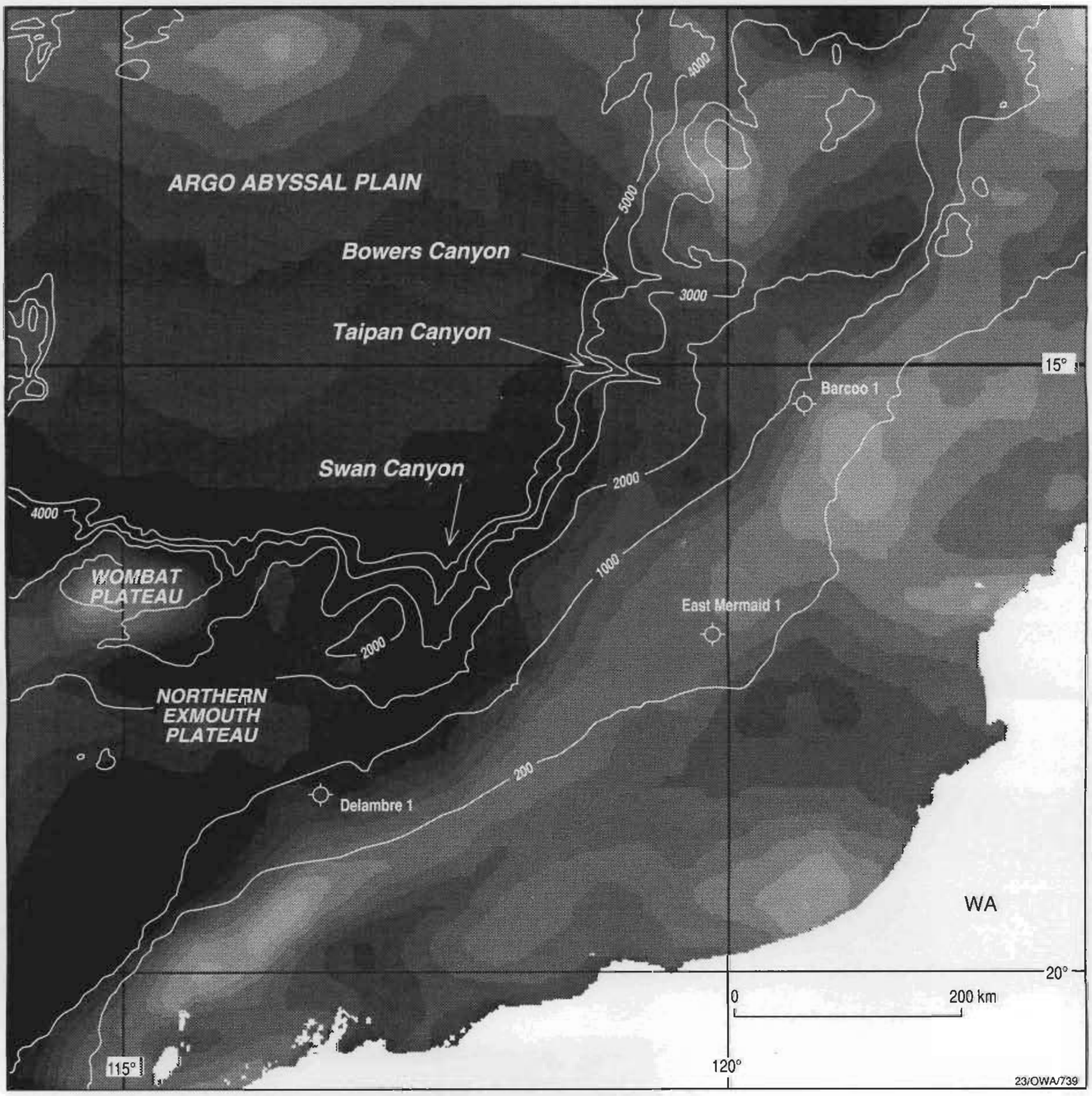

Figure 12. SEASAT gravity map of the northern Exmouth Plateau and Rowley Terrace. Note highs and lows corresponding to major features on northern Exmouth Plateau, east-west depression north of the Exmouth Plateau corresponding to low in oceanic crust, and northeast-trending depression on northwestern Rowley Terrace corresponding to structural low.

collapse. The steep outer margin is extensively cut by canyons, some controlled by fault structures. Evidence of Early Cretaceous and Cainozoic calcareous pelagic turbidites from ODP Site 765 on the Argo Abyssal Plain indicates that the canyons were active periodically from Berriasian times (Gradstein et al., 1992).

On the continental margin landward of the collapsed zone, there was little tectonic movement in post-Valanginian times (post-D unconformity). The general structure of the region is summarised in the map showing the structure of the C unconformity (Fig. 2), and in the SEASAT gravity imagery (Fig. 12). Figure 2 shows a general deepening of the C horizon towards the continental margin of more than 1000 m, and the strong influence of fault-controlled canyons on the structure of the outer margin. This fault control is most marked on the northern Exmouth Plateau, where the trend of faults is generally NNE or E. The SEASAT gravity map (Fig. 12) shows the main structural features of the area. Along the northern margin of the Exmouth Plateau, the structural lows and highs are very obvious; the long gravity depression north of the plateau corresponds to a depression in oceanic crust of about 500 m (Rao et al., 1994 — this issue), and was apparently caused by subsidence of the continental margin and its effect on the adjacent coupled oceanic crust. An apparently anomalously low area is depicted on the roughly N–S continental margin in the vicinity of Bowers and Taipan Canyons. This low is a continuation of a northeast-trending structural low mapped between the Scott Plateau and the North West Shelf by Stagg & Exon (1981) from seismic data. It corresponds to the northern part of the Rowley Sub-basin and to a bathymetric depression (Fig. 7) and structural depression (Fig. 2).

Following formation of the Valanginian unconformity, shallow-marine detrital sediments and marls (C–D sequence) were laid down over most of the area with a thickness of about 200 to 800 m, with the exception of the highly structured parts of the northern Exmouth Plateau, where they onlap structural highs and are absent in places (Plate 1P). The transition to carbonate sedimentation in the Late Cretaceous led to the deposition of two similar megasequences of shelf and upper slope carbonates in the Late Cretaceous and Paleocene–Eocene (B–C and A1–B, respectively). Each of the sequences averages 100 to 200 m thick except along the steep outer margin, where they are commonly absent because of non-deposition or erosion (Plate 1). By the Oligocene, the margin had subsided considerably and deep-water pelagic carbonate sediments were being deposited, particularly on the outer margin. However, nearer land, the productivity of calcareous organisms was high, and prograding wedges of shelf carbonates built outward into deeper water. We have mapped three sequences (Plate 1): Oligocene, Early to Middle Miocene, and Late Miocene to Holocene (A0–A1, AP–A0, seabed-AP, respectively). In deeper water, the combined thickness of the three sequences is less than 500 m; on the shelf, the three prograding sequences usually total 1000 to 2000 m.

Petroleum potential

Our mapping has not been on a sufficiently detailed grid to discuss individual petroleum prospects. As elsewhere on the North West Shelf, the best potential source rocks are probably in organic-rich parts of the Jurassic sequence, in particular the Middle Jurassic coal measures and Upper Jurassic deltaic sediments. A possible additional source sequence would be lagoonal carbonates in the Upper Triassic and Lower Jurassic along the outer margin. Possible reservoir rocks include Upper Triassic and Jurassic sandstones over the entire area, and Upper Triassic and Jurassic carbonates along the outer margin. Seals would be provided by Mesozoic shales and mudstones. Large fault blocks on the Exmouth Plateau are the most inviting exploration targets (such as the block drilled in Delambre No. 1), but unfortunately many lie in water more than 1000 m deep. Similar fault blocks in the Rowley Sub-basin are confined to very deep water (e.g. Plate 1A; 3000 m water depth on line 95/06–09). In shallower water in the Rowley Sub-basin, there appears to be little faulting, and explorers would therefore have to rely on broad regional highs as targets. However, the relative paucity of modern seismic data and of exploration wells means that this area is still considerably underexplored, and should not be classified as unprospective at this stage.

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Significance of calcareous nannofossil-bearing Jurassic and Cretaceous sediments on the Rowley Terrace, offshore northwest Australia

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Examination of dredge hauls taken from the Rowley Terrace, northwest Australia, has revealed the occurrence of several Mesozoic nannofossil-bearing levels. Dating has been satisfactorily achieved even though some key species important for subdividing Mesozoic rocks elsewhere are either missing or have a different stratigraphic significance in northwest Australia.

The Jurassic levels have a special significance, being within the paralic pre-breakup sequence which was deposited during the rifting of Gondwana. Two distinct nannofloras are recorded: one contains *Discorhabdus striatus* and *Carnolithus superbus* which suggest an early Toarcian age (between 183.5 and 182 Ma); and the other is dominated by the key species *Lotharingius contractus* and forms transitional between *L. contractus* and *Watznaueria britannica* suggesting an early Bajocian age (about 170.5 Ma). The older, seemingly more limited spatially, probably represents a marine ingression related to tectonically controlled pulse(s) of increased subsidence. The early Bajocian nannoflora evidently is a part of a major transgression; it is similar to another found in outcrops of the ammonite-rich Newmarracarra Limestone in the onshore Perth Basin to the south. This horizon correlates with a global eustatic rise in sealevel, more clearly than does the horizon bearing the early Toarcian nannofossils.

A notable change in the style of sedimentation to open-marine conditions in post-breakup times is indicated by several rich Cretaceous nannofloras; the oldest is Valanginian. The Valanginian nannoflora contains both Austral/Boreal and Tethyan elements (*Crucibiscutum salebrosum, Cruciellipsis cuvillieri, Microstaurus chiastius, Speetonia colligata*, and *Tubodiscus verenae*) suggesting a possible surface-water connection between the Early Cretaceous juvenile ocean northwest of Australia and the southern Tethyan ocean. The younger late Barremian/early Aptian nannoflora does not unequivocally show such a connection. Details of the Late Cretaceous nannofloras are consistent with the conclusion that distinct 'nannoprovinces' came into existence during the later part of the Senonian in the maturing ocean west and north of Australia. They include two combinations of stenothermal species (either *Biscutum coronum, Ceratolithoides aculeus, Kamptnerius magnificus*, and *Quadrum gothicum*; or *Petrarhabdus copulatus, Quadrum trifidum, Q. gothicum, Cribrocorona gallica*, and *Ceratolithoides aculeus*) suggesting locations well within the Extratropical Nannoprovince during the mid and late Campanian. Coeval nannofloras from the Giralia Anticline of the Carnarvon Basin to the south suggest location close to the southern limit of the Extratropical Nannoprovince, whereas those from the Papuan Basin to the north are indicative of the Tropical Nannoprovince.

Introduction

This study is based on samples obtained on board the R/V *Rig Seismic* during BMR Surveys 95 and 96 (Exon & Ramsay, 1990; Colwell et al., 1990) when dredging in the offshore Canning Basin of northwest Australia took place. Sites dredged are located on the outer slopes of the Rowley Terrace and the eastern side of the Swan Canyon (Fig. 1) in the Rowley Sub-basin. This sub-basin forms part of the eastern margin of the Argo Abyssal Plain. To the southwest, the northern Exmouth Plateau forms the southern margin of the Argo Abyssal Plain. Both the Rowley Terrace and the northern Exmouth Plateau have steep outer margins (deep canyons and steep slopes, most suited for dredging), and on their landward side they merge into the North West Shelf (Fig. 1). Compared with the Rowley Terrace, dredging on the northern Exmouth Plateau has been more extensive — during several surveys involving the R/V *Sonne* and R/V *Rig Seismic*. More geological data have come from the plateau because of recent Ocean Drilling Program (ODP) drilling (Haq et al., 1990; von Rad et al., 1992). The Swan Canyon, at the junction of the eastern and southern margins of the Argo Abyssal Plain (Fig. 1), links both the northern Exmouth Plateau and the Rowley Sub-basin.

The Rowley Sub-basin and the northern Exmouth Plateau seemingly have similar geological histories, the Mesozoic being the most eventful. The lithologies of the Mesozoic samples dredged from the Rowley Sub-basin (see Exon & Ramsay, 1990; Colwell et al., 1990; Colwell et al. 1994–this issue) and the northern Exmouth Plateau (e.g., von Stackelberg et al., 1980; Exon et al., 1988) reflect the development of these areas during the rifting of Gondwana and the following interval, leading to the formation of mature ocean northwest of Australia. Paralic sedimentation, characterising the rifting phase, began during the Late Triassic and continued on until late in the Jurassic; a largely fluviodeltaic to marginal marine regime prevailed during much of the Jurassic in the Rowley Sub-basin. Subsequent to the breakup, deposition of calcareous mud, consistent with a juvenile ocean stage, took place during most of the Early Cretaceous. The onset of shelf carbonate accumulation, suggesting a mature ocean, occurred during the early Late Cretaceous. These three sedimentation regimes were apparently discrete, as indicated by the presence of bounding unconformities in the stratigraphic sections revealed in nearby oil wells and ODP holes. The unconformities are major, as they coincide with regional seismic reflectors (see, e.g., Exon & Ramsay, 1990). Their precise age-limits vary in different sections, being generally Late Jurassic (mainly Callovian/Oxfordian — E unconformity), close to the Jurassic–Cretaceous boundary (Tithonian–Hauterivian — D unconformity), and mid Cretaceous (Turonian — C disconformity); the oldest unconformity is often referred to as the 'breakup' or the main unconformity. Within the post-breakup sequence, the mid Cretaceous disconformity coincides with the base of the fully marine Cretaceous carbonates.

Rich calcareous foraminiferal assemblages seem to be absent from the pre-breakup sequence in the Rowley Sub-basin, but not so in the younger sequence (see below). Only *sparse* benthic foraminiferids have been recorded at few levels in the pre-breakup sequence on the northern Exmouth Plateau (see Quilty, 1981, 1990). Except for a recently published abstract (Shafik, 1993a), Jurassic calcareous nannofossils are not known from the Rowley Sub-basin. However, there were some indications of their presence in the Swan Canyon area, because the Cainozoic section of ODP Site 765 in the southern Argo Abyssal Plain contains black claystone clasts which yielded nannofossil assemblages indicative of Middle or Middle to Late

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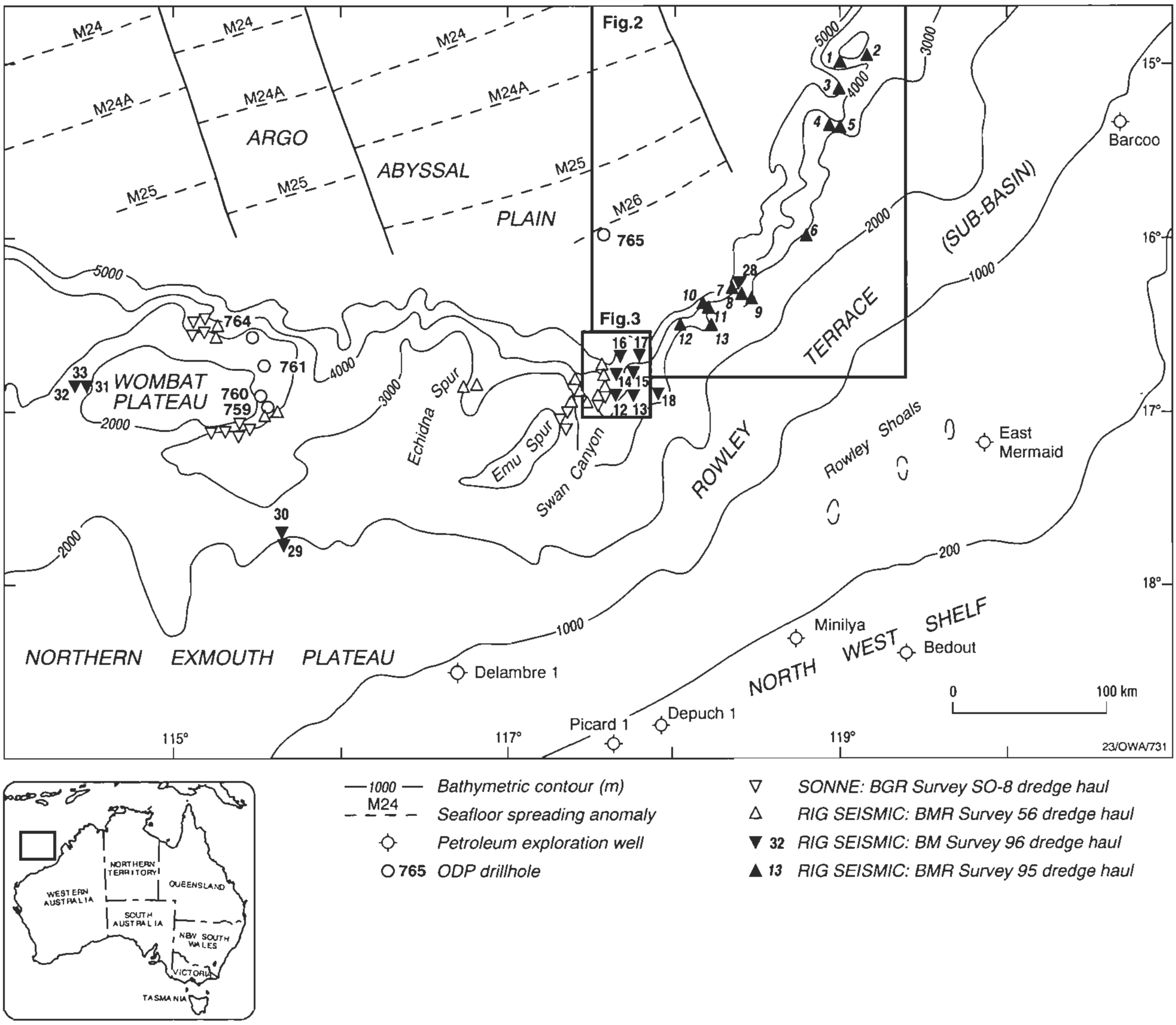

Figure 1. Map of North West Australia showing the southern Argo Abyssal Plain, northern Exmouth Plateau and Rowley Terrace. Boxes show locations of Figures 2 and 3.

Jurassic age (Ludden et al., 1990). These nannofossil-bearing Jurassic clasts were transported into the Argo Abyssal Plain probably from the Swan Canyon area. Cretaceous nannofossils have been known from the Exmouth Plateau either from dredge samples (see Shafik *in* von Stackelberg et al., 1980) or ODP cores (see von Rad et al., 1992), but not from the Rowley Sub-basin. In the Argo Abyssal Plain, Late Jurassic and Cretaceous sections have been sampled and their nannofossils studied (Gradstein et al., 1992).

The time scale used throughout this paper is that of Haq et al. (1987).

Mesozoic stratigraphy in the Rowley Sub-Basin and nearby areas

It is difficult to construct the stratigraphy of the Mesozoic of the areas dredged in the Rowley Sub-basin from the recovered samples alone, with many being undateable (see Exon & Ramsay, 1990; Colwell et al., 1990). Nevertheless, a brief statement on the major lithofacies sampled is given below; this is preceded by a discussion on relevant nearby stratigraphic successions in the sub-basin.

Several stratigraphic sections are known from drilling near our dredge sites. These are either oil wells in the Rowley Sub-basin itself (landward of our dredge sites), or ODP holes in nearby areas of the Exmouth Plateau and the Argo Abyssal Plain (Fig. 1). In addition, dredging on the northern Exmouth Plateau has contributed significantly to our knowledge of the stratigraphy of that region (see, e.g., von Stackelberg et al., 1980; Exon et al., 1988; von Rad et al., 1990).

Rowley Sub-basin wells

Only two wells within the Rowley Sub-basin, Shell East Mermaid No. 1 and Woodside Barcoo No. 1 (Fig. 1), are close enough to the areas dredged to be relevant (see stratigraphic summary figs 3 and 4 in Exon 1994 — this issue). The nannofossil biostratigraphy of these wells has not been worked out, although data about their palynology and foraminiferal successions have been compiled in unpublished reports.

Stratigraphic successions. The Jurassic in East Mermaid No. 1 is represented by a regressive facies of more than 1000 m of sandstone, shale with minor siltstone and coal. The Jurassic in Barcoo No. 1 is slightly thinner and is truncated by the breakup unconformity; the Bathonian is unrepresented. The main lithology is claystone with minor sandstone and siltstone; above the breakup unconformity, the claystone includes traces of calcilutite.

Immediately above the breakup unconformity in East Mermaid No. 1, there is a Neocomian to Cenomanian transgressive sequence, more than 1000 m thick, of siltstone grading upward into claystone and finally marl. Its top (the Cenomanian marl) is separated from a thin section of Upper Cretaceous outer shelf and bathyal calcilutite by the clearly delineated mid Cretaceous unconformity. In Barcoo No. 1, the approximately 1000 m thick Berriasian–Campanian sequence is divided into three parts by an intra-Aptian unconformity and the mid Cretaceous unconformity. The latter unconformity separates two main lithologies: claystone below and calcilutite above.

Depositional environments. During the Jurassic and Early Cretaceous, conditions were much less marine than during the remainder of the Cretaceous and during the Cainozoic. (The Upper Cretaceous carbonates and Cainozoic sediments in both the East Mermaid and Barcoo wells were deposited under fully marine conditions as indicated by the abundance of planktic foraminiferids and also by lithofacies.) The palynological evidence suggests that the Jurassic sequence in East Mermaid No. 1 was deposited mainly in deltaic environments. Such environments are unlikely to favour calcareous microplankton, so the Jurassic sequence in East Mermaid may not contain any significant calcareous planktic remains. In Barcoo No. 1, the intra-Aptian disconformity marks the lowest occurrence of planktic foraminiferids. Below it, the section contains only benthic forms. Palynological studies on the Jurassic sequence in this well indicate marine conditions, probably marginal (see **Discussion** below), during the deposition of its later part (Callovian and Oxfordian sediments).

Discussion. Two depositional megasequences are recognisable within the Mesozoic of Barcoo No. 1: a pre-breakup sequence of fluviodeltaic to marginal marine sediments of Triassic to Middle Jurassic age (deposited during the rifting of Gondwana); and a post-breakup marine transgressive sequence of Late Jurassic and Cretaceous age. The same is true for the Mesozoic of East Mermaid No. 1, although the post-breakup sequence seems to be younger there, being Early Cretaceous at its base.

Rich foraminiferal assemblages seem to be absent below the pre-breakup unconformity in both East Mermaid No. 1 and Barcoo No. 1, and planktic forms are altogether absent from below the Aptian in the latter well.

It is worth noting that *marine* conditions suggested in most palynological studies (e.g. Burger, 1994–this issue) do not necessarily mean *open marine* conditions or *good access to the open sea* based on the presence of calcareous nannofossils. In the absence of calcareous nannofossils, marine conditions based on palynomorphs invariably would not be true or fully marine, rather suggestive of some poor connection with sea.

Rowley Sub-basin dredges

As the rock types dredged from the Rowley Sub-basin are generally similar to those from the northern Exmouth Plateau, Exon & Ramsay (1990) adopted the lithofacies association scheme used first by von Stackelberg et al. (1980). The same practice is adopted in this study. The lithofacies, which include samples analysed here, are listed below; their ages are based partly on nannofossil data (see Shafik *in* Exon & Ramsay, 1990). They are

- *Lithofacies D.* Deltaic and marine mudstone association — Late Triassic to mid Cretaceous,
- *Lithofacies E.* Pelagic marls and chalks — Early Cretaceous to Pliocene, and
- *Lithofacies I.* Evaporitic association — Early Jurassic.

Colwell et al. (1994–this issue) included calcareous mudstones in Lithofacies E. This unnecessarily widens the concept of this type of lithofacies; Lithofacies E is better restricted to the pelagic carbonates as distinct from Lithofacies D of which the bulk is calcareous mudstones.

Other lithofacies, which on seismic evidence are from the pre-breakup sequence, lack calcareous nannofossils (Exon & Ramsay, 1990). These are

- *Lithofacies A.* Delta plain/coal measure association — Late Triassic to Middle Jurassic,
- *Lithofacies B.* Ferruginous association — Middle to Late Jurassic,
- *Lithofacies C.* Shallow-water carbonate association — Late Triassic to Middle Jurassic,
- *Lithofacies F.* Chert and orthoquartzite association — Jurassic to Early Cretaceous, and
- *Lithofacies H.* Volcanic and volcaniclastic rocks — Late Triassic to Middle Jurassic.

Colwell et al. (1994–this issue) included in Lithofacies C Early Cretaceous coquinite and shelly sandstone which have rare nannofossils.

Nearby oceanic sites (Exmouth Plateau and Argo Abyssal Plain)

A stratigraphic overview of the Mesozoic of areas on the northern Exmouth Plateau (near our dredge sites) revealed a significant difference in the Jurassic settings of the Swan Canyon area and the Wombat Plateau (a sub-plateau of the Exmouth Plateau, see Fig. 1). This is based on studies of rocks dredged from known seismic sequences (von Stackelberg et al., 1980; Exon et al., 1982; von Rad et al., 1983; Exon et al., 1988; von Rad et al., 1990) and on reports on several recently penetrated ODP holes (Haq et al., 1990; von Rad et al., 1992).

In the Swan Canyon area, there is a thick Jurassic sequence immediately underneath the main unconformity. This pre-breakup sequence consists largely of shallow-water sediments and coal measures. On the Wombat Plateau, at ODP Sites 759, 760, 761 and 764, the Jurassic is unrepresented; Upper Triassic sediments are directly overlain by Cretaceous deltaic sediments or carbonates. Either erosion exceeded deposition during the Jurassic in this part of the northern Exmouth Plateau, or there was post-depositional erosion of the Jurassic sequence (see fig. 32 in von Rad et al., 1992).

At ODP Site 765 in the Argo Abyssal Plain (Fig. 1), sedimentation began during the Tithonian, initially in an oxic environment on a newly formed ocean floor, at depths beneath the CCD (Dumoulin & Bown, 1992). This is consistent with the notion that the Argo Abyssal Plain was formed as a result of the Callovian/Oxfordian breakup event (E unconformity) clearly shown in seismic profiles of the outer slope of the Rowley Terrace (see Ramsay & Exon, 1994–this issue).

Calcareous nannofossil assemblages and their significance

This study considers only the Mesozoic calcareous nannofossils which are thought to be *in situ* in the Rowley Sub-basin dredges; some of the Tertiary material from the sub-basin contained rare, displaced Cretaceous nannofossils (Shafik *in* Exon & Ramsay, 1990; Colwell et al., 1990).

Detailed lithological descriptions and other relevant data (e.g. foraminiferal results) concerning the samples discussed in this study are given in Exon & Ramsay (1990), Colwell et al. (1990), and Colwell et al. (1994–this issue). Table 1 gives the location and water depth at those dredge stations yielding the samples studied herein.

Table 1. Location of dredge stations and water depths.

Dredge Station	Area	Lat. -- Long. Range	Water depth (m)
95DR/004	Southwest slope of Copperhead Canyon	15° 22.0'S - 118° 58.9'E 15° 22.5'S - 118° 57.9'E	4200 to 3750
95DR/006	Southwest slope of Tortoise Canyon	16° 0.133'S - 118° 0.00'E 16° 00.00'S - 118° 46.5'E	3400 to 3100
95DR/007	Notheast side of Mermaid Canyon	16° 18.9'S - 118° 22.2'E 16° 18.2'S - 118° 24.3'E	4530 to 3900
95DR/008	Notheast side of Mermaid Canyon	16° 19.1'S - 118° 22.9'E 16° 18.9'S - 118° 23.2'E	4400 to 4230
95DR/009	Notheast side of Mermaid Canyon	16° 19.0'S - 118° 23.8'E 16° 18.5'S - 118° 25.6'E	4200 to 3350
96DR/013	Notheast side of Mermaid Canyon	16° 50.02'S - 117° 38.45'E 16° 54.55'S - 117° 40.55'E	4150 to 3600
96DR/014	Notheast side of Mermaid Canyon	6° 54.40'S - 117° 35.22'E 16° 49.89'S - 117° 38.61'E	5250 to 4005
96DR/015	Notheast side of Mermaid Canyon	16° 50.01'S - 117° 37.98'E 16° 50.01'S - 117° 39.76'E	4200 to 3875

Jurassic assemblages

Four Jurassic calcareous nannofossil assemblages were identified from dredges collected from Tortoise and Mermaid Canyons (Fig. 2). In addition, a monospecific 'assemblage' of possible Jurassic age was found in a dredge from Copperhead Canyon. As discussed below, the rocks collected from these canyons, and other palaeontological and seismic evidence, indicate a thick, mostly paralic Upper Triassic–Jurassic sequence; the nannofossil-bearing horizons represent significant environmental changes.

Precision in dating of the recovered nannofossils has suffered somewhat because very little is known about the

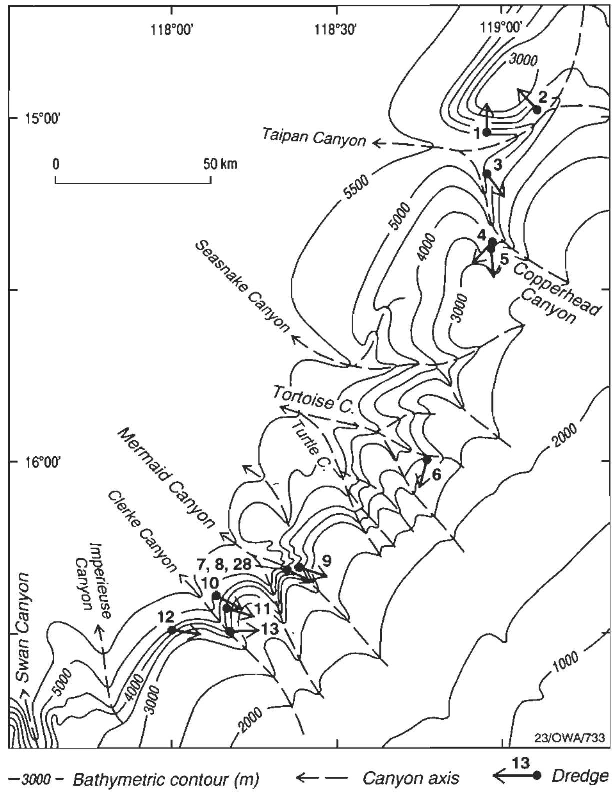

Figure 2. Location of dredge hauls on the Rowley Terrace.

Jurassic nannofossil biostratigraphic succession in the Australian region (in northwest Australia, the stratigraphic ranges of most key Jurassic species are virtually unknown), and also because the assemblages themselves are based on dredge samples (inherently lacking stratigraphic order) which have come from a largely paralic succession. Such assemblages may lack some of the species critical for precise age determination, and those present may have stratigraphic ranges different from those determined elsewhere. Even in the case where the stratigraphic order of samples is known, such as those from ODP Site 765 in the Argo Abyssal Plain, the nannofossil biostratigraphy of the Late Jurassic–Early Cretaceous sequence has been described as problematic (see Bown, 1992). Low diversities (associated with poor preservation and barren intervals) and the probability that certain key species are absent or rare, and have truncated ranges, contributed to the difficulty of biostratigraphic evaluation of the Jurassic–Early Cretaceous of Site 765.

Published Jurassic nannofossil zonal schemes (e.g. Barnard & Hay, 1974; Hamilton, 1979; 1982; Medd, 1982; Bown et al., 1988; Reale et al., 1992; Baldanza & Mattioli, 1992) are based on successions in northern and western Europe, and other (mainly Tethyan) areas in the Northern Hemisphere. Stratigraphic ranges of certain marker species in these two areas are not necessarily the same (see, e.g., Medd, 1982) essentially because of a differing biogeographic setting: boreal in northern Europe and Tethyan to the south (see, e.g., Cooper, 1989).

Table 2 summarises the Toarcian–early Bajocian parts of two published zonal schemes: Bown et al. (1988), based on material from southern England and southwest Germany (Sub-Tethyan and Boreal); and Baldanza & Mattioli (1992), based on sections in Italy (Tethyan). Several

Table 2. Pliens bachian to early Bajocian calcareous nannofossil zonations.

Stage	Sub-Tethyan / Boreal zones, subzones and defining events Bown et al. (1988)		Tethyan zones and some important events Baldanza & Mattioli (1992)		Stage
BAJOCIAN	Lower *S. speciosum* Zone		NOT ZONED		BAJOCIAN
		↑ *Stephanolithion speciosum*			
	Watznaueria britannica Zone				
		↑ *Watznaueria britannica*	*Watznaueria britannica* ↑		
AALENIAN	*Lotharingius contractus* Subzone			*Watznaueria manivitae* Zone	AALENIAN
		↑ ***Lotharingius contractus***	***Watznaueria manivitae*** ↑		
	Retecapsa incompta Subzone		***Carinolithus magharensis*** ↑	*Watznaueria barnesae* Zone	
TOARCIAN		↑ ***Retecapsa incompta***	***Retecapsa incompta*** ↑		TOARCIAN
			Discorhabdus criotus⊕ ↑		
	Discorhabdus striatus Zone		*Watznaueria barnesae* ↑		
		↑ ***Discorhabdus striatus***	***Discorhabdus striatus*** ↑	*Lotharingius contractus* Zone	
			Lotharingius contractus ↑		
	Carinolithus superbus Zone			*Discorhabdus ignotus* Zone	
		↑ ***Carinolithus superbus***	***Discorhabdus ignotus*** ↑		
	Crepidolithus cavus Subzone		***Carinolithus superbus*** ↑	*Lotharingius crucicentralis* Zone	
			Lotharingius crucicentralis ↑		
PLIENSBACHIAN		↑ ***Crepidolithus cavus***	***Lotharingius sigillatus*** ↑	upper *Lotharingius hauffii* Zone	

Stages are not to scale. All events are lowest occurrences. Species names in bold are those among the recorded species in the present study. ⊕ = This event is at about the same Toarcian level, within the ammonite *variabilis* Zone, in both zonations, but the underlying event, base *W. barnesae*, is early Bajocian in Bown et al. (1988).

bioevents are common to the two schemes. Most of the shared events, notably the earliest occurrences of *Carinolithus superbus, Discorhabdus striatus, Retecapsa incompta,* and *Watznaueria britannica,* are given similar ages in the two schemes. However, despite the seemingly minor differences between the two schemes, that of Bown et al. (1988) is more applicable to Australian material (see below). A notable difference between the two schemes is the level of the earliest occurrence of *Lotharingius contractus* in relation to other events. This event is critical to the biostratigraphic assignment and dating of the assemblages described herein.

Two similar assemblages were identified from dredge **95DR/06** collected from the southwest slope of Tortoise Canyon in water 3400 and 3100 m deep (Fig. 2).

- Several species of calcareous nannofossils were identified from sample 95DR/06–03, a yellowish-brown, heavily altered silty mudstone (Lithofacies D). These include *Biscutum novum, Carinolithus superbus, Crepidolithus crassus, Discorhabdus striatus, Lotharingius hauffii,* and *L. crucicentralis.*

- Rare, but apparently relatively more diversified, nannofossils were extracted from sample 95DR/06–04, a grey evaporitic mudstone with laminae of fibrous calcite (Lithofacies I). Most of the forms are small. Species identified include *Axopodorhabdus cylindratus, A. macrogranulatus, Biscutum novum, Carinolithus superbus, Crepidolithus crassus, Discorhabdus striatus, Lotharingius hauffii, L. sigillatus, Mitrolithus elegans,* and *Tubirhabdus patulus.*

These two assemblages are assignable to the *Discorhabdus striatus* Zone of Bown et al. (1988) (Table 2) based on the occurrence of the nominate species and *Carinolithus superbus* — in the absence of *Retecapsa incompta.* The *D. striatus* Zone is of long duration, extending from slightly below the top of the ammonite early Toarcian *falciferum* Zone to within the latest Toarcian *levesquei* Zone (see Bown et al., 1988). Based on the absence of *Discorhabdus criotus,* it is possible to narrow the assignment to the lower part of the zone. *Discorhabdus criotus* makes its earliest appearance within the ammonite *variabilis* Zone according to both the Bown et al. (1988) and the Baldanza & Mattioli (1992) schemes. This easily recognisable species is known from the Rowley Sub-basin, having been found in the younger Jurassic assemblages from Mermaid Canyon (described below). Its absence from the Tortoise Canyon assemblages is, therefore, likely to be biostratigraphic, i.e. these assemblages predate its appearance, having an age range between late *falciferum* Zone and mid *variabilis* Zone.

The longer interval from mid *falciferum* Zone to the top of the *variabilis* Zone is between about 184 and 181.5 Ma in Haq et al. (1987). Accordingly, the age of the Tortoise Canyon assemblages could be between 183.5 and 182 Ma. Biostratigraphic evaluation using the scheme of Baldanza & Mattioli (1992, see below), though not without problems, yielded a numeric age range not dissimilar from this one.

In the absence of *Watznaueria barnesae,* the occurrence of *Discorhabdus striatus* would also suggest assignment of the above assemblages to the upper part of the Toarcian *Lotharingius contractus* Zone of Baldanza & Mattioli (1992) (see Table 2). However, the index species (which is the name species) of the *L. contractus* Zone was not enountered, although it is present in the younger Jurassic assemblages from Mermaid Canyon (discussed below). And *Watznaueria barnesae* (the index species of the overlying zone — see Table 2) was not found in the younger assemblages from Mermaid Canyon either. The earliest occurrence of *W. barnesae*, which age is middle or late Toarcian in Italy (Baldanza & Mattioli,1992; Reale et al., 1992), is early Bajocian in NW Europe (Bown et al., 1988).

Assuming the likely case that the earliest occurrence of *Watznaueria barnesae* in northwest Australia is younger than the Toarcian — similar to, for example, in Portugal (Hamilton, 1979) or southern France (Erba, 1990) — a wider biostratigraphic assignment than the upper part of the *L. contractus* Zone of Baldanza & Mattioli (1992), would be more realistic. The Tortoise Canyon assemblages can be assigned to the upper *L. contractus*–lower *W. barnesae* zonal interval (in the scheme of Baldanza & Mattioli, 1992; see Table 2). This assignment is based on the presence of *Discorhabdus striatus*, considers the absence of *Discorhabdus criotus* a reliable biostratigraphic evidence (discussed above) and, at the same time, ignores the absence of both *W. barnesae* and *L. contractus*.

Baldanza & Mattioli (1992) correlated their entire *Lotharingius contractus* Zone within the ammonite *bifrons* Zone, and the lower part of their *Watznaueria barnesae* Zone (below the lowest occurrence of *Discorhabdus criotus*) with most of the ammonite *variabilis* Zone — a biostratigraphic interval estimated between about 183.5 and 181.5 Ma using the time scale of Haq et al. (1987). Accordingly, the numeric age of the Tortoise Canyon assemblages could be between about 183 and 181.5 Ma.

Considering both estimates of the age of the Tortoise Canyon assemblages reached above (essentially based on correlation of the lowest occurrences of *Discorhabdus striatus* and *D. criotus* to the ammonite zones in Bown et al., 1988; and Baldanza & Mattioli, 1992), the age range between 183.5 and 181.5 Ma is adopted.

Discussion. Inspection of the seismic profile at dredge station 95DR/06 (see fig. 7D in Colwell et al., 1994–this issue), supported by the rock types recovered, suggests a thick Upper Triassic and Jurassic sequence there. The dominant rock types recovered at this station (table 1 in Exon & Ramsay, 1990; see also Colwell et al., 1994–this issue) are: (a) unfossiliferous welded tuff to lapillistone (Lithofacies H, presumably Late Triassic to Middle Jurassic) slightly less than 60% of recovery; (b) unfossiliferous ironstone and boxstones with grey mudstone core (Lithofacies B, presumably Late Jurassic) 20%; and (c) dark-grey carbonaceous mudstone (Lithofacies A, with palynological evidence suggesting Middle Jurassic age) slightly less than 20%. The evidence suggests that these three dominant rock types were deposited under mostly paralic conditions; for example, palynological study of sample 96DR/06–1A (Lithofacies A), which lacks nannofossils, indicates a Toarcian paludal environment (Burger, 1994–this issue). The calcareous mudstone (Lithofacies D) which contains Jurassic nannofossils represents less than 1% of recovery at this station. Hence, it may be concluded that the sampled thick Upper Triassic–Jurassic sequence in Tortoise Canyon consists largely of paralic facies with very minor open-marine components.

Two similar assemblages were identified from dredges **96DR/28 and 95DR/07** collected from the northeast side of Mermaid Canyon (Fig. 2). They came from Lithofacies D.

- Poorly preserved, sparse calcareous nannofossils were recovered from sample 96DR/28–16, a soft, plastic, dark-grey mud/mudstone with some mica. Forms transitional between *Lotharingius contractus* and *Watznaueria britannica*, typical *Lotharingius contractus*, and *Watznaueria manivitae* predominate the recovered nannofossils. Others include *Axopodorhabdus macrogranulatus*, very rare (broken) *Carniolithus* sp., ?*Crucirhabdus primulus*, frequent *Discorhabdus striatus*, *Ethmorhabdus gallicus*, *Lotharingius crucicentralis*, *L. velatus*, *Mitrolithus elegans*, frequent *Retecapsa incompta*, common *Schizosphaerella punctulata*, *Vekshinella quariarculla*, and *Zygodiscus erectus*.
- A poorly preserved assemblage was extracted from sample 95DR/07–08, a puggy dark-grey, slightly calcareous clay. This assemblage is dominated by severely etched forms transitional between *Lotharingius contractus* and *Watznaueria britannica*, slightly better-preserved *Lotharingius contractus*, *Watznaueria manivitae*, *Schizosphaerella puntuclata*, and to a lesser degree of dominance by *Discorhabdus jungii*. Species of *Carinolithus* (including *C.* sp. cf. *C. superbus*: small central opening with wider distal rim) are common. Rare species include *Axopodorhabdus macrogranulatus*, *Biscutum intermedium, B. novum*, *Carinolithus magharensis, Discorhabdus criotus, D. striatus, Ethmorhabdus gallicus, Lotharingius velatus, Retacapsa incompta,* and *Tubirhabdus patulus*. Extremely rare species include *Crepidolithus* sp., *Crucirhabdus primulus, Diductius constans, Hexalithus noeliae*, and *Nodosella* sp.

Although the assemblage from 95DR/07–08 is notably more diversified than that from 96DR/28–16, the two assemblages are coeval because both have the same key species. They are assignable to the late Aalenian–early Bajocian *Lotharingius contractus* Subzone of Bown et al. (1988) on the occurrence of the name species. The presence of forms transitional between *Lotharingius contractus* and *Watznaueria britannica* restricts the assignment to the uppermost (early Bajocian) part of the subzone (i.e. a level close to the ammonite *discites/laeviuscula* zonal boundary in Bown et al., 1988 — about 170.5 Ma, using the scale of Haq et al., 1987). Typical *W. britannica* was not found, and assignment to the younger *Watznaueria britannica* Zone of Bown et al. (1988) is — in a strict application of the zonal definition — not possible.

The age of the earliest occurrence of the key species *Lotharingius contractus* is different in the two zonal schemes presented in Table 2. It is middle Toarcian (within the ammonite *bifrons* Zone) according to Baldanza & Mattioli (1992; see also Reale et al., 1992) — well below the level (*murchisonae* Zone) indicated in Bown et al. (1988); Toarcian assemblages from Tortoise Canyon (referable to the *L. contractus* Zone of Baldanza & Mattioli, 1992) lack *L. contractus* (discussed above). From the available data it is difficult to determine whether the difference in the two zonations is due to a differing concept of *L. contractus* or it is because the species appears earlier in Italy. Nevertheless, assignment of the two assemblages described from Mermaid Canyon to the uppermost (early Bajocian) part of the *L. contractus* Subzone of Bown et al. (1988), and thus to near the ammonite *discites/laeviuscula* zonal boundary, is confirmed by indirect evidence (discussed below). This came from the ammonite-rich Newmarracarra Limestone (onshore Perth Basin, Western Australia).

The same two assemblages (in 95DR/07–08 and 96DR/28–

16) can be assigned to the *Watznaueria manivitae* Zone in the scheme of Baldanza & Mattioli (1992) (Table 2), based on the presence of the name species and the absence of *Watznaueria britannica*. This zone is mostly Aalenian in age, but its uppermost part is early Bajocian, and assignment of the Mermaid Canyon assemblages to it is, therefore, not inconsistent with their age being early Bajocian as concluded above.

The absence of *Watznaueria barnesae*, in the presence of common *W. manivitae*, from the early Bajocian assemblages (in 95DR/07–08 and 96DR/28–16) is inconsistent with the zonal scheme of Baldanza & Mattioli (1992; see also the scheme of Reale et al., 1992). In these schemes, *W. barnesae* appears (in the Toarcian) well below the lowest occurrence of *W. manivitae*. Indeed, the earliest occurrence of *W. barnesae* is regarded by Reale et al. (1992, p. 61) as "a very important event", marking a drastic change in the nannofossil assemblages: the appearance of the long-ranging genus *Watznaueria* partly replacing the already well-established genus *Lotharingius*. On the other hand, the absence of *W. barnesae* is not inconsistent with assignment of the assemblages to the *Lotharingius contractus* Subzone of Bown et al. (1988). In the zonal scheme of Bown et al. (1988), the lowest occurrence of *W. barnesae* is about the same level as that of *W. britannica,* the latter defining the base of the overlying early Bajocian *Watznaueria britannica* Zone.

The presence of *Crucirhabdus primulus* may suggest some reworking from Upper Triassic/Lower Jurassic source(s). This species is restricted to the Upper Triassic and/or the Lower Jurassic in Great Britain according to Hamilton (1982). It disappears within the Pliensbachian (Barnard & Hay, 1974) or near the Pliensbachian/Toarcian boundary in southwest Germany (Crux, 1984), but Thierstein's (1976, fig. 7) generalised range chart shows the species ranging into the Bajocian.

It is worth noting that Burger (1994–this issue) gives evidence suggesting that Upper Triassic and Lower Jurassic sediments recovered from Mermaid Canyon (dredge 95DR/07) were deposited in restricted, brackish coastal marine palaeoenvironments.

Discussion. The nannofossil evidence in 96DR/28–16 and 95DR/07–08 suggests the existence of open-marine Middle Jurassic horizons at Mermaid Canyon, but the dominant lithologies and their lack of calcareous nannofossils, together with the seismic evidence (fig. 7E in Colwell et al., 1994–this issue) suggest the occurrence of a thick, fluviodeltaic to marginal marine Upper Triassic–Middle Jurassic sequence there. Dredging of the northeastern side of Mermaid Canyon at three stations during BMR Survey 95 (DR/07, DR/08 and DR/09) resulted in the recovery of huge amounts of rocks (see table 1 of Exon & Ramsay, 1990), with either Lithofacies type A or D predominating, followed by type C. The calcareous clay, bearing nannofossils (sample 95DR/07–08), represents less than 1% of rocks recovered in the dredge haul.

In addition to the four Jurassic assemblages discussed above, there is a monospecific nannofossil 'assemblage' of possible mid Jurassic age. This was extracted from sample **95DR/04-07**, a pale-olive quartz sandstone (Lithofacies D), dredged from Copperhead Canyon (Fig. 2) in water 4200 to 3750 m deep. The fossils are few and evidently represent a single species, namely *Watznaueria britannica*. This monospecific 'assemblage' is unlikely to be a result of dissolution, rather due to some ecological limitations (such as very shallow environment). The stratigraphic range of common *W. britannica* is Bajocian to Barremian; see also Grün & Zweili (1980) who gave the stratigraphic range of *W. britannica* as Toarcian–Campanian. *Watznaueria britannica*, without the association of other *Watznaueria* species including those which are equally ecologically robust and dissolution resistant (such as *W. barnesae*), is particularly suggestive of mid Jurassic age. The associated foraminiferids, though meagre, tend to support a Jurassic age assignment (see Lynch *in* Exon & Ramsay, 1990).

Both zonations in Table 2 agree that the earliest occurrence of *Watznaueria britannica* is early Bajocian in age. The maximum age of 95DR/04–07 is, therefore, early Bajocian, but certainly younger than the early Bajocian assemblages from Mermaid Canyon (samples 95DR/07–08 and 96DR/28–16) which predate the appearance of *W. britannica* (assigned to the *Lotharingius contractus* Subzone of Bown et al., 1988).

Jurassic marine incursions in the Rowley Sub-basin

All lines of evidence (drilling, dredging and seismic data) in the Rowley Sub-basin point to a thick, paralic pre-breakup sequence having been deposited in restricted shallow-water environments during the rifting phase. Environmental conditions were largely unsuitable for calcareous nannoplankton. However, this study suggests that open-sea conditions were established on at least two occasions, in the Toarcian and Bajocian.

Dredging of the pre-breakup sequence at most of the canyons on the outer Rowley Terrace (including the Tortoise and Mermaid Canyons) resulted in the recovery of mostly non-marine sediments, for which palynological ages are within the range of Late Triassic to Late Jurassic (see Burger, 1994–this issue). The record of reasonably rich early Toarcian and early Bajocian calcareous nannofossil assemblages from this sequence, at Tortoise and Mermaid Canyons respectively is, therefore, highly significant. It suggests a major environmental change: a shift to open-marine conditions on two occasions late in the rift history. The evidence discussed below suggests that the Bajocian change is widespread, and probably directly related to a major transgression caused by a global eustatic rise in sealevel. The earlier Toarcian change seems to have been more limited spatially. It is, therefore, best described as marine ingression rather than a transgression; the possible age range of the Tortoise Canyon assemblages (183.5 to 181.5 Ma) is marked by two eustatic rises in sealevel, at about 183.5 and 181.5 Ma, as shown in the eustatic curves of Haq et al. (1987). Nevertheless, it is likely that this ingression was caused by a local event rather than by either of these global sealevel rises. Tectonically controlled pulses of increased subsidence along the Rowley Terrace are not uncommon during the rifting phase, and their magnitude during the early Toarcian may have been just enough to allow a marine ingression to bring in calcareous nannoplankton.

Nannofossil evidence for either or both Jurassic marine incursions was searched for elsewhere on the Australian margin. The onshore Perth Basin (western margin) was thought to be the best candidate, because the Jurassic stratigraphic setting there is suitable and samples from the relevant unit were available. The Bajocian Newmarracarra Limestone — containing significant ammonite fauna and other mollusca, as well as brachiopods, ostracods and foraminiferids — is sandwiched between the Aalenian Bringo Shale or Colalura Sandstone (both lack open-marine fauna) and the Bajocian Kojarena Sandstone with its wood fragments and little molluscs (see Skwarko, 1974; Cockbain, 1990 and references therein for more details). Poorly preserved, rare calcareous nannofossils were recovered from two samples from the Newmarracarra Limestone in its type area near Geraldton. These are dominated by the index species *Lotharingius contractus*, and transitional forms between *L. contractus* and *Watznaueria britannica*, and included *Biscutum* sp., *Discorhabdus striatus*, *Lotharingius velatus*, and *Retecapsa incompta* (Shafik, unpublished data). A strong correlation with the early Bajocian nannoflora from Mermaid Canyon is evident. The same key species are found in the assemblages of both the Newmarracarra Limestone and Mermaid Canyon. An assignment of the Newmarracarra Limestone assemblage to the uppermost part of the *Lotharingius contractus* Subzone of Bown et al. (1988) logically follows.

According to Arkell & Playford (1954), the main Newmarracarra Limestone ammonite fauna is referable to the early Bajocian *sowerbyi* Zone. A more recent study of this fauna limited its correlation to the younger part of the *sowerbyi* Zone — the early Bajocian *laeviuscula* Zone (Hall, 1989). This confirms the early Bajocian age indicated by the nannofossils not only for the Newmarracarra Limestone, but also for the clay/mudstone unit(s) sampled twice (different surveys) at Mermaid Canyon (samples 95DR/07–8 and 96DR/28–16).

In southern England, the *sowerbyi* Zone includes the *discites* fauna in its lower part, and the *laeviuscula* Zone in its upper part. There, the uppermost part of the *Lotharingius contractus* Subzone of Bown et al. (1988) includes the top of the *discites* Zone and the very bottom of the *laeviuscula* Zone. As pointed out above, this level is about 170.5 Ma, which incidently coincides with a global eustatic rise in sealevel in the chart of Haq et al. (1987).

Accepting that the nannofossil assemblages from the clay/mudstone unit(s) at Mermaid Canyon (Rowley Subbasin) and the Newmarracarra Limestone in the Geraldton area (Perth Basin) are coeval, it will be evident that we are dealing with a widespread open-marine event — a major transgression. The cause of this transgression is likely to have been the same at these two widely-spaced locations — the global rise in sealevel at about 170.5 Ma.

Assuming the age of sample 95DR/04–07 is mid Jurassic, the presence of only *Watznaueria britannica* can be taken to indicate stressful conditions during the decline of the (Bajocian) transgression which earlier covered the site of Mermaid Canyon and the Geraldton area. Conditions at Copperhead Canyon then were probably too shallow to sustain a similarly diversified nannoflora as that which flourished earlier in the nearby Mermaid Canyon.

Early Cretaceous: pre-mid Aptian assemblages

Several Early Cretaceous calcareous nannofossil assemblages were identified from dredges collected from Swan and Copperhead Canyons. Details of these assemblages are given in Table 3. They were extracted from lithologies identifiable as Lithofacies D and E (see Table 3).

The Swan Canyon assemblages came from three sites on

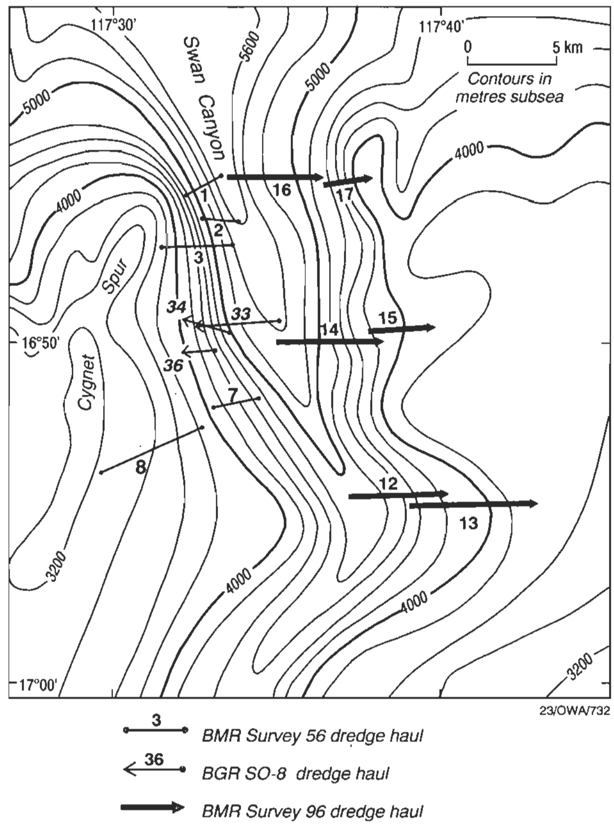

Figure 3. Bathymetric map of Swan Canyon showing the location of dredges from BMR Survey 96 and other surveys. Note that dredges studied here are from the eastern (Rowley Terrace) side of Swan Canyon.

the Rowley Terrace side of the canyon (Fig. 3). Those from dredges 96DR/13 and 96DR/15 are generally poorly preserved: in most cases, specimens show signs of severe etching, probably a result of prolonged exposure to cold bottom waters. Better preserved assemblages were extracted from dredge 96DR/14 and sample 96DR/15–06A. The Copperhead Canyon assemblages, from dredge 95DR/04 (Fig. 2), are moderately well preserved.

Dating of the assemblages relies heavily on determination of the species stratigraphic ranges outside the Australian region (Table 4), even though the little which is known about the Early Cretaceous nannofossil biostratigraphy northwest of Australia (ODP sites in the Argo Abyssal Plain and on the Exmouth Plateau; see Mutterlose, 1992; Bown, 1992; Bralower & Siesser, 1992) suggests that the stratigraphic ranges of some of the key species are substantially different there than elsewhere. This is particularly true for *Vagalapilla matalosa* and *Tegumentum striatum* (discussed below).

Most of the recently published Lower Cretaceous zonations consider the events of the earliest appearances of *Eprolithus floralis* and *Rhagodiscus angustus* as early (or middle) Aptian in age (see also Thierstein, 1976; Table 4). These two key species are absent from the assemblages detailed in Table 3. Since both species are distinct, easily identifiable and cosmopolitan in their geographic distribution, their absence is taken to indicate a minimum age limit on the assemblages of early Aptian.

All assemblages from Swan Canyon contain *Crucibiscutum salebrosum*, except for the 96DR/13–05B assemblage which includes *Vagalapilla matalosa* instead. The former species occurs in two groupings based on size. The measurements are defined by the length of the major axis of the ellipse describing the perimeter of the specimen; one group has values between three and four microns, and the other greater than five microns. Because of the potential biostratigraphic utility of either or both groups, in Table 3 *C. salebrosum* is differentiated into small and

Table 3. Distribution of calcareous nannofossils in Lower Cretaceous dredges from the Rowley Terrace, North West Australia.

Locality⇨	Swan Canyon						Copperhead Canyon	
Sample⇨	96DR/ 13-03	14-08A	15-06A	15-01	13-05A	13-05B	95DR/ 04-04	04-01C
Lithology⇨	Light grey & dusky yellow mudstone		Light grey mud	Light grey mudstones	Radiolarian Chalk		Light grey mudstone	Light grey marl
Species⇩ Preservation⇨	Moderate	Good	Good	Very poor	Moderate	Poor	Good	Good
Axopodorhabdus dietzmannii	Very rare	Rare	Frequent		Rare		Rare	Rare
Bidiscus rotatorius			Very rare			Frequent	Rare	Frequent
Biscutum constans	Frequent						Rare	Rare
Bukrylithus ambiguus							?	Rare
Chiastozygus litterarius							Very rare	Very rare
Conusphaera mexicana	Very rare	Very rare	Very rare					
Cretarhabdus conicus	Very rare	Rare	Rare	Frequent	Rare	Rare	Rare	Frequent
Cretarhabdus loriei					Rare			
Cretarhabdus surirellus	Very rare	Rare	Rare		Very rare	Rare	Rare	Rare
Crucibiscutum salebrosum (small)		Rare	Common	Rare	Frequent		Very rare	
Crucibiscutum salebrosum (large)	Frequent	Frequent	Frequent	Rare				
Cruciellipsis cuvillieri	Frequent	Common	Rare					
Cyclagelosphaera deflandrei	Rare	Common	Very rare	Rare	Very rare			
Cyclagelosphaera margerelii	Rare	Frequent	Rare	Frequent	Very rare	Very rare		
Diazomatholithus lehmanii	Rare	Rare	Frequent					Rare
Eiffellithus windii	Frequent	Rare						
Ethmolithus gallicus			Rare	Very rare			Rare	
Grantarhabdus coronadventus							Rare	Rare
Grantarhabdus medii	Rare	Rare	Frequent	Rare	Frequent	Rare	Rare	Rare
Haqius circumradiatus	Frequent	Rare	Rare	Very rare	Rare	Rare	Rare	Very rare
?Hemipodorhabdus gorkae							Rare	
Heterorhabdus sinuosus						Rare	Rare	Rare
Lapideacassis sp. aff. *L. cornuta*								Rare
Lithraphidites carniolensis	Rare	Frequent	Frequent	Very rare	Frequent	Frequent	Rare	Rare
Manivitella pemmatoidea	Rare	Frequent	Frequent	Rare	Frequent	Frequent		Rare
Micrantholithus sp. (fragments)	Rare							
Microstaurus chiastius	Rare	Rare	Frequent	Rare	Frequent	Frequent		Rare
Microstaurus quadratus		Very rare		Very rare	Rare	Very rare		
Parhabdolithus embergeri	Frequent	Common	Common	Rare	Abundant	Frequent	Very rare	Rare
Percivalia fenestrata	Very rare		Rare	Very rare		Very rare	Rare	Frequent
Perissocyclus noeliae					Frequent		Rare	Frequent
Pickelhaube furtiva			Very rare			?		
Polycostella senaria							Rare	
Retecapsa angustiforata			Rare			Very rare	Rare	Very rare
Rhagodiscus asper	Very rare	Rare		Rare	Rare		Rare	Rare
Rotelapillus laffittei	Rare	Rare	Frequent	Frequent		Rare	Rare	Rare
Rucinolithus terebrodentarius	Rare	Rare	Rare	Frequent	Rare			
Sollasites horticus		Rare						
Speetonia colligata	Frequent	Rare	Rare					
Tegumentum octiformis					Rare		Rare	Rare
Tegumentum stradneri					Frequent	Frequent		
Tegumentum striatum	Very rare				Frequent		Very rare	
Tranolithus gabalus (small)							Very rare	Frequent
Tubodiscus jurapelagicus	Rare	Rare						Rare
Tubodiscus verenae	Rare	Rare	Rare	Very rare				
Vagalapilla matalosa					Very rare	Frequent	Common	Abundant
Vekshinella spp.	Frequent		Frequent	Rare	Frequent			Rare
Viminites sp.							Rare	
Watznaueria barnesae	Abundant	Abundant	Abundant	Dominant	Dominant	Abundant	Abundant	Abundant
Watznaueria biporta			Rare					
Watznaueria britannica	Common	Rare	Very rare					
Watznaueria ovata	Rare	Frequent	Frequent	Common	Frequent	Frequent	Rare	Rare
Zygodiscus diplogrammus								Rare
Zygodiscus erectus	Frequent	Frequent	Frequent	Rare	Common	Frequent	Rare	Rare
Age determination⇨	Valanginian				? Hauterivian	late Barremian/early Aptian		

large forms; a cut-off point of five microns is adopted. Sample 96DR/13-05A contains both *C. salebrosum* and *V. matalosa*, but the latter species is represented by one specimen only, whereas *C. salebrosum* — represented by its small form only — dominates the fine fraction of the assemblage. The stratigraphic ranges of these two species do not usually overlap. The highest occurrence of *C. salebrosum* is early Barremian (in Britain and North Sea see Table 4), whereas the earliest occurrence of *V. matalosa* is late Barremian or early Aptian (in lower latitude areas: see Thierstein, 1976; Roth, 1983; Table 4). However, both Mutterlose (1992) and Bown (1992) recorded *V. matalosa* ranging down into the Valanginian, with *C. salebrosum*, in the Argo Abyssal Plain. This casts some doubt on the biostratigraphic value of both the highest occurrence of *C. salebrosum* (without being differentiated into its two forms) and the lowest occurrence of *V. matalosa*, and their use as means of dating. The data given in Table 3 seem to suggest that the small *C. salebrosum* disappeared later than the large form, but shortly after the appearance of *V. matalosa*.

The Valanginian age given to the Swan Canyon assemblages 96DR/13–03, 96DR/14–08A, 96DR/15–06A and 96DR/15–01 in Table 3 is based on the co-occurrence of *Tubodiscus verenae* and *Cyclagelosphaera deflandrei* (the NC3 Zone of Roth, 1983). The former species is restricted to the Valanginian according to several authors (see Table 4). Moreover, the presence of *Eiffellithus windii* in 96DR/13–03 and 96DR/14–08A confirms their Valanginian age, and the presence of *Speetonia colligata* in 96DR/13–03, 96DR/14–08A and 96DR/15–06A is not inconsistent with this age assignment.

The age of the 96DR/13–05A assemblage falls within the range early Valanginian to early Barremian (probably Hauterivian), considering the frequent occurrence of *Tegumentum striatum* and small *Crucibiscutum salebrosum*, as well as the very rare *Vagalapilla matalosa* and *Cyclagelosphaera deflandrei* (see Table 3). Most authors regard the lowest occurrence of *T. striatum* as early Hauterivian in age (Table 4), but Mutterlose's (1992) work on ODP cores from Sites 765 and 766 (in the Argo Abyssal Plain and on the western escarpment of the Exmouth Plateau, respectively) suggests that this species ranges down well within the Valanginian (probably close to the Berriasian/Valanginian boundary) in northwest Australia.

The Swan Canyon 96DR/13-05B assemblage and those from Copperhead Canyon (Table 3), contain *Vagalapilla matalosa* without the association of *Crucibiscutum salebrosum, Eprolithus floralis* or *Rhagodiscus angustus*. This suggests a late Barremian to early Aptian age (the NC6 Zone of Roth, 1983). The presence of *Tranolithus gabalus* (small) and *Chiasozygus litterarius* in the Copperhead Canyon assemblages strengthens this age assignment; some taxonomic vagueness regarding *C. litterarius* (Covington & Wise, 1987) notwithstanding. Thierstein (1973) indicated that *Tranolithus gabalus* occurs first in the upper Aptian–lower Albian, but this may only be true for the 'larger' form of this species.

Remark. The samples discussed above contain Early Cretaceous calcareous nannofossils, but lack planktic foraminiferids. Sample 96DR/13–03 contains rare agglutinated foraminiferids, but no calcareous benthic foraminiferids (Colwell et al., 1990), whereas the late Barremian — early Aptian samples (95DR/04–01C and 95DR/04–04) has abundant and well-preserved calcareous benthic foraminiferal assemblages (Lynch *in* Exon & Ramsay, 1990). The exclusion of planktic foraminiferids from the late Barremian–early Aptian samples was probably a result of intolerable environmental conditions such as high salinity, in an otherwise fully marine environment, as attested by the high diversity and abundance of the nannofossils.

NW Australia–Tethys Valanginian connection

The species *Crucibiscutum salebrosum*, which is common in the Valanginian assemblages from Swan Canyon (present study), is widely spread and common in the Neocomian of several oceanic sites northwest of Australia (Mutterlose, 1992). It is not known in low-latitude sections, and has been previously labelled as Boreal/Austral (see Crux, 1989; Table 4). Its geographic distribution in the Valanginian and Hauterivian shows a bipolar pattern (Mutterlose, 1992). Plate tectonic reconstructions indicate that the Rowley Terrace area was at palaeolatitudes around 50° during the Neocomian (Anomaly M11 see Johnson & Veevers *in* Veevers, 1984).

Based on the evidence of *Crucibiscutum salebrosum*, it is reasonable to conclude that the surface waters of the juvenile ocean being formed during the Neocomian northwest of Australia were cool for most of the time; such a cool surface-water regime matches the mid to high-latitude location of the Rowley Sub-basin indicated by plate reconstructions. Consistent with this conclusion is the observation that none of the species indicated by Thierstein (1976) as tropical (see Table 4) is present in the Swan Canyon assemblages (except for very rare *Conusphaera mexicana*). However, species regarded elsewhere as Tethyan are found co-occurring with *C. salebrosum* in the Valanginian assemblages from Swan Canyon (present study) and in Neocomian sediments at ODP Sites 765 in the southern Argo Abyssal Plain (near Swan Canyon) and 766 on the western escarpment of the Exmouth Plateau (see, Mutterlose, 1992; Bown, 1992).

Thus, the Tethyan species *Cruciellipsis cuvillieri, Speetonia colligata,* and *Tubodiscus verenae* (Crux, 1989; see Table 4), together with the suspected Tethyan *Cyclagelosphaera deflandrei* (Cooper, 1989) occur with *C. salebrosum*, albeit generally in much lesser abundance than *C. salebrosum*, in the Valanginian assemblages from Swan Canyon. Similarly, rare *Cruciellipsis cuvillieri, Microstaurus chiastius,* and *Speetonia colligata* occur in association with common *C. salebrosum* in the Neocomian of both Sites 765 and 766 (Mutterlose, 1992); at Site 765, Bown (1992) recorded *Tubodiscus verenae* in addition. Co-occurrences of *C. salebrosum* and *C. cuvillieri* have also been recorded in clasts from the Cainozoic section at these ODP sites (Ludden et al., 1990). At DSDP Site 261, in the northern Argo Abyssal Plain, Bown (1992) recorded rare *Cruciellipsis cuvillieri, Microstaurus chiastius,* and *Tubodiscus verenae* in association with sporadic common *C. salebrosum*, and argued that the Argo Basin occupied a position transitional between the Tethyan and Austral realms.

Very rare specimens of the warm-water *Conusphaera mexicana* were found not only in the Valanginian assemblages from Swan Canyon (Table 3), but also in the Lower Cretaceous of ODP Sites 765 and 766 (Ludden et al., 1990).

Table 4. Cretaceous calcareous nannafossil events.

Nannofossil Events❖	L.P.	Thierstein, 1976 -- Various regions: (mainly low-latitudes)	Perch-Nielsen, 1985 -- Various regions: (low & high-latitudes)	Taylor, 1982 -- United Kingdom	Roth, 1983 -- North Atlantic: Blake-Bahama Basin	Jakubwoski, 1987 -- North Sea	Applegate & Bergen, 1988 -- N. Atlantic off Iberia	Crux, 1989 -- NW Europe	Mutterlose, 1992 -- NW Australia
Base *Rhagodiscus angustus*		**middle Aptian4**	**early Aptian4**	**late Aptian**	**early Aptian4**	**mid Aptian**	late Aptian		**middle Aptian**
Base *Eprolithus floralis*		**middle Aptian4**	**early Aptian4**	middle Aptian	**early Aptian4**		**late Aptian**		**early Aptian**
Base ***Vagalapilla matalosa***		early Aptian			**late Barremian3**			**late Barremian**	
Base *Hayesites irregularis*	TR	**base Aptian**	**top Barremian**	late Albian	early Barremian⊕		**base Aptian**		**early Aptian**
Base ***Chiastozygus litterarius***		**latest Barremian**		late Hauterivian✱	**late Barremian3**				
Top *Nannoconus colomii*	TR	**latest Barremian**			late Barremian				
Top *Nannoconus abundans*	E			**late Barremian**		late Aptian			
Top *Nannoconus borealis*	E					**late Barremian**			
Top *Calcicalithina oblongata*	TR	**early Barremian**	**early Barremian**		late Barremian		**early Barremian✱**		
Top *Lithraphidites bollii*	TR	early Barremian	**late Hauterivian**						
Top ***Crucibiscutum salebrosum***	B/A		?Albian	early Barremian		**early Barremian**			
Base *Nannoconus abundans*	E			**late Hauterivian✱**				**early Barremian**	
Top *Tegulalithus septentrionalis*	B/A					**early Barremian2**			
Top ***Tegumentum striatum***	B/A		late Hauterivian3	latest Barremian			late Hauterivian2⊕		**early Hauterivian**
Top ***Speetonia colligata***	Te	late Hauterivian3	**late Hauterivian3**	early Hauterivian	early Barremian⊕	**early Barremian2**	**late Hauterivian2**		
Top ***Cruciellipsis cuvillieri***	Te	**late Hauterivian3**	late Hauterivian⊕	early Hauterivian3	**late Hauterivian**	**late Hauterivian**			**late Hauterivian**
Base *Lithraphidites bollii*	TR	**early Hauterivian**	**late Hauterivian⊕**				**early Hauterivian**		
Base *Tegulalithus septentrionalis*	B/A					late Hauterivian		**late Hauterivian**	
Top *Corollithion silvaradion*	B/A					**early Hauterivian**		**early Hauterivian**	
Top *Eprolithus antiquus*	B/A		middle Aptian					**early Hauterivian**	
Base *Eprolithus antiquus*	B/A		**early Hauterivian**				mid Hauterivian	**latest? Valanginian**	
Base ***Tegumentum striatum***	B/A		**early Hauterivian2**	**early Hauterivian3**	early Hauterivian		**early Hauterivian**	**late Valanginian**	**late Valanginian**
Base ***Cretarhabdus loriei***			**early Hauterivian2**						
Top ***Cyclagelosphaera deflandrei***		top Valanginian			top Valanginian				
Base *Nannoconus bucheri*	TR	early Hauterivian			late Valanginian✱		late Valanginian		
Top *Micrantholithus speetonensis*	E		**late Valanginian**	late Valanginian		**mid Valanginian**		**early? Valanginian**	
Top ***Tubodiscus verenae***	Te	late Valanginian		late Albian	**late Valanginian✱**		late Hauterivian⊕		
Top *Diadorhombus rectus*		late Valanginian2	**late Valanginian**				late Barremian✱		
Base *Parhabdolithus infinitus*					mid Valanginian				
Top *Hayesites wisei*	TR	late Valanginian2			early Valanginian		late Valanginian		
Base *Micrantholithus speetonensis*	E		**mid Valanginian1**	**early Valanginian**		latest Ryazanian		**early Valanginian**	
Base ***Eiffellithus windii***							**late Valanginian**		
Base ***Tubodiscus verenae***	Te	early Valanginian	**mid Valanginian1**	late Hauterivian	**early Valanginian2**		early Valanginian1		
Base *Calcicalithina oblongata*	TR	**early Valanginian**	**early Valanginian✱**		early Valanginian2		**early Valanginian1**		
Base *Diadorhombus rectus*	TR	early Valanginian	**early Valanginian✱**		early Valanginian2		early Valanginian1		
Base ***Speetonia colligata***	Te	base Valanginian	**top Berriasian**	**late Ryazanian2**	late Berriasian1	late Ryazanian1			
Base ***Crucibiscutum salebrosum***	B/A		base berrisian	**late Ryazanian2**	late Berriasian	late Ryazanian1			**Berriasian**
Base ***Percivalia fenestrata***		late Berriasian			late Berriasian				
Base ***Retecapsa angustiforata***		**late Berriasian**	late Berriasian	**late Ryazanian1**	late Berriasian	**late Ryazanian1**	**late Berriasian**		
Base *Retecapsa neocomiana*					**late Berriasian**				
Base ***Cruciellipsis cuvillieri***	Te	early Berriasian	**early Berriasian**	early Hauterivian	early Berriasian1	late Ryazanian1			**early Berriasian**
Base ***Lithraphidites carniolensis***		**base Berriasian1**	**base Berriasian**	late Ryazanian1					
Base *Nannoconus colomii*	TR	**base Berriasian1**			base Berriasian				

❖ = The sequential order of this set of events is not based on a single section, but is a compromise between sources cited; species in bold (not necessarily events) are those recorded in this study. L.P. = latitudinal preference: TR = tropical (Thierstein, 1976); Te = Tethyan, B/A = Boreal or Austral, E = endemic to northern high-latitude areas (Crux 1989). In all other columns: age in bold represents main events considered in reference cited; ✱ or ⊕ = sequential order of these events is reversed in the reference cited; events with similar numbers are about the same level according to the reference cited. Truncated range of *Cruciellipsis cuvillieri* and a younger age for the base of *Tubodiscus verenae* in the British sections than elsewhere are consistent with the species being Tethyan.

The association of several, usually rare, Tethyan species (*Cruciellipsis cuvillieri*, *Speetonia colligata*, *Tubodiscus verenae*, and *Microstaurus chiastius*), very rare *Conusphaera mexicana* and common *Crucibiscutum salebrosum*, in the Neocomian of the Exmouth Plateau–Swan Canyon–Argo Abyssal Plain region, suggests the introduction of warmer waters into the generally cool surface water regime of the juvenile ocean being formed northwest of Australia: a connection with the Tethys to the north is likely. The present study suggests that this connection developed during the Valanginian. The evidence for such connection in the younger Early Cretaceous assemblages from Swan and Copperhead Canyons is at best tenuous: *Microstaurus chiastius* is present in these assemblages but other Tethyan species, such as *Calcicalathina oblongata* and *Hayesites irregularis*, are lacking.

Co-occurrences of cool and warm water species have been recorded from Albian sediments in the south, on the Carnarvon Terrace, Perth Abyssal Plain and Naturaliste Plateau, western Australia (Shafik, 1993b with references therein), but this is more likely a result of the development of a nannoplankton province, the Extratropical Nannoprovince, in the maturing ocean west and north of Australia (Shafik, 1993b, in prep. — see Fig. 4).

Late Cretaceous: Campanian assemblages

Only two Late Cretaceous assemblages were recovered.

- A rich, well-preserved assemblage was recovered from sample 96DR/14–04, a yellowish-grey marl (Lithofacies E), dredged from the Rowley Terrace side of Swan Canyon (Fig. 3). Species identified include *Actinozygus regularis*, *Acuturris scotus*, *Ahmuellerella octoradiata*, *Arkhangelskiella specillata*, *Biscutum constans*, *B. coronum*, *Broinsonia parca*, *Ceratolithoides aculeus*, *Cretarhabdus surirellus*, *Cribrosphaerella ehrenbergii*, *Cylindralithus serratus*, *Eiffellithus eximius*, *E. turriseiffelii*, *Gartnerago obliquum*, *Grantarhabdus coronadventus*, *Heterorhabdus sinuosus*, *Kamptnerius magnificus*, *Lapideacassis* sp. aff. *L. cornuta*, *Lithastrinus grillii*, *Lithraphidites carniolensis*, *Manivitella pemmatoidea*, *Markalius inversus*, *Microrhabdulus belgicus*, *M. helicoideus*, *Micula concava*, *Micula staurophora*, *Parhabdolithus embergeri*, *Prediscosphaera cretacea*, *P. spinosa*, *P. stoveri*, *Quadrum gothicum*, *Reinhardtites anthophorus*, *R. levis*, *Rhagodiscus angustus*, *Rotelapillus laffittei*, *Watznaueria barnesae*, *W. biporta*, and *Zygodiscus bicrescenticus*.
- A diversified but poorly to moderately preserved assemblage was recovered from sample 95DR/04–01B, a white chalk (Lithofacies E), dredged from Copperhead Canyon (Fig. 2) in water 4200 to 3750 m deep. Species present included *Arkhangelskiella specillata*, *Broinsonia parca*, *Ceratolithoides aculeus*, *Chiastozygus litterarius*, *Cretarhabdus conicus*, *Cribrocorona gallica*, *Cribrosphaerella ehrenbergii*, *Cylindralithus serratus*, *Eiffellithus eximius*, *E. turriseiffelii*, *Heterorhabdus sinuosus*, *Lapideacassis* sp., *Manivitella pemmatoidea*, *Microrhabdulus belgicus*, *M. decoratus*, *Micula staurophora*, *Petrarhabdus copulatus*, *Prediscosphaera cretacea*, *P. majungae*, *P. spinosa*, *Quadrum gothicum*, *Q. trifidum*, *Reinhardtites levis*, *Rhagodiscus angustus*, several species of *Vekshinella*, and *Watznaueria barnesae*.

The age of the 96DR/14–04 assemblage is mid to late Campanian on account of the presence of *Broinsonia parca*, *Ceratolithoides aculeus*, *Quadrum gothicum*, and *Lithastrinus grillii*. The presence of *Eiffellithus eximius* and *Reinhardtites levis* in the same assemblage, and the absence of *Quadrum trifidum* and *Arkhangelskiella cymbiformis*, are consistent with this age assignment.

The presence of *Quadrum trifidum* in the 95DR/04–01B assemblage suggests a late Campanian to early Maastrichtian age. And the presence of both *Broinsonia parca* and *Eiffellithus eximius*, and the absence of *Arkhangelskiella cymbiformis*, favour a late Campanian age.

Remarks. The stratigraphic sequence in Copperhead Canyon at dredge station 95DR/04 (Fig. 2) includes horizons of Early and Late Cretaceous age. It is interesting to note that about 12 times as much Lower Cretaceous grey mudstone (95DR/04–04) was recovered compared to the younger Cretaceous white mud (95DR/04–01B), the former representing about 15% of the dredge haul. This suggests a much thicker and/or better exposed Lower Cretaceous sequence, and this suggestion is supported by the recovery of another Lower Cretaceous (grey calcareous mud, sample 95DR/04–01C) in the same dredge haul.

Campanian Extratropical Nannoprovince

Based primarily on the distribution and abundance of some stenothermal (with narrow temperature range tolerance) Late Cretaceous nannofossil species, Shafik (1990) was able to differentiate three biogeographic provinces during the late Campanian–late Maastrichtian interval in the Australian region. Figure 4 outlines these provinces during the latest Senonian and earlier — notwithstanding that the concept that the provinces' limits coincide with palaeolatitudes needs testing with more data points.

The combination of the species *Biscutum coronum*, *Ceratolithoides aculeus*, *Kamptnerius magnificus*, and *Quadrum gothicum*, noted in the mid to late Campanian assemblage of sample 96DR/14–04 from Swan Canyon, suggests a location well within the Extratropical Nannoprovince where mixing of surface waters from lower and higher latitudes occurred. The geographic distribution of both *C. aculeus* and *Q. gothicum* suggests their preference for mid to low latitudes. *Biscutum coronum*, first described from the Falkland Plateau (South Atlantic — see Wise & Wind, 1977), has subsequently been found in mainly high latitude areas such as the Perth Basin of Western Australia (see Shafik, 1990).

Similarly, the association of frequent *Petrarhabdus copulatus*, *Quadrum trifidum*, *Q. gothicum*, *Ceratolithoides aculeus*, and *Cribrocorona gallica*, which is noted in the late Campanian sample 95DR/04–1B from Copperhead Canyon, characterises areas well within the Extratropical Nannoprovince. Coeval assemblages with similar associations have been reported from the Giralia Anticline in the Carnarvon Basin of Western Australia — near the southern limit of the same province as indicated by Shafik (1993b). There, *P. copulatus* is more frequent, but the other species are less frequent — particularly *Cribrocorona gallica*, *Quadrum trifidum*, and *Q. gothicum* which are usually rare and sporadically distributed.

Coeval assemblages from the Papuan Basin, to the north, indicate the Tropical Nannoprovince (Shafik, 1990). These lack *Petrarhabdus copulatus*, but contain abundant *Quadrum trifidum*, *Q. gothicum*, *Ceratolithoides aculeus*, and *Cribrocorona gallica*.

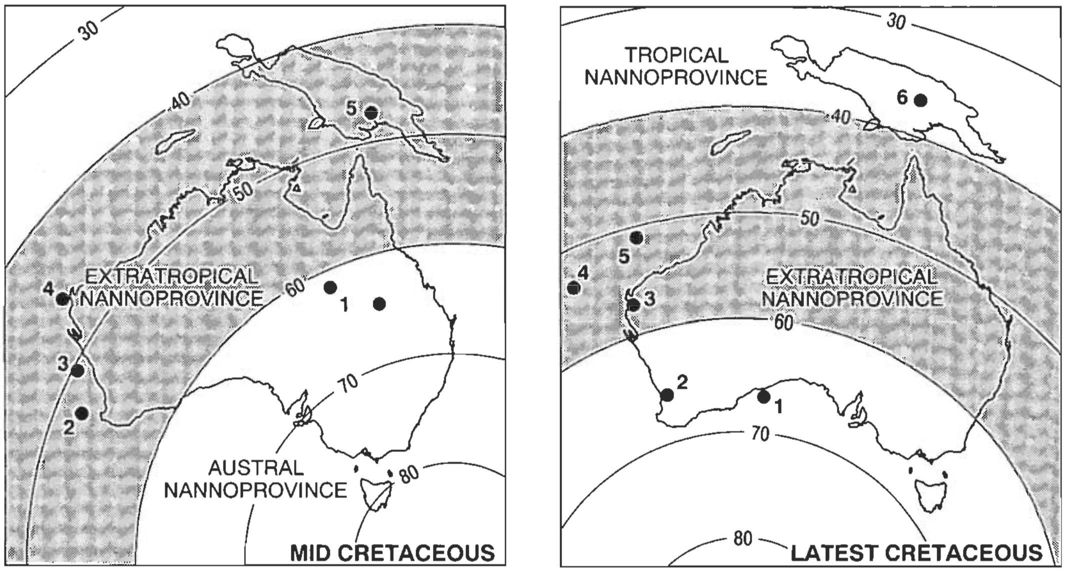

Figure 4. Nannoprovinces for the Australian region during mid and latest Cretaceous times (after Shafik, 1993b).

The foraminiferal assemblage recovered from sample 95DR/04–01B suggests cold-water influence (Lynch *in* Exon & Ramsay, 1990). This assemblage is dominated by *Rugoglobigerina* species, and contains a low proportion of keeled *Globotruncana*.

The concept of recognising the Extratropical Nannoprovince where cool and warm-water species co-existed, usually in equal but generally low abundances, is useful from a biostratigraphic point of view. This is particularly so if the biostratigraphically important species are stenothermal with narrow latitudinal preference (e.g. those with bipolar distribution). For example, Shafik (1990) showed that the relative order of appearance of the key late Maastrichtian *Nephrolithus frequens* and *Micula murus*, which co-occur in the Extratropical Nannoprovince, is reversed within the province: the warm-water *M. murus* appears first in the lower latitudes within the province, whereas the cool-water *N. frequens* appears first in the higher latitudes within the province. Evidence from ODP Leg 122 confirms this conclusion. At Site 761 on the Wombat Plateau (well within the Extratropical Nannoprovince — see Shafik, 1993b; Fig. 4), *N. frequens* first appears at the same level as *M. murus,* whereas at Site 762 on the Exmouth Plateau (closer to the southern limit of the Extratropical Nannoprovince) it lies much lower (see Bralower & Siesser, 1992).

Discussion

The continental mass north-westward of the Carnarvon and Canning Basins (which was later replaced by the Argo Abyssal Plain) remained continuous until Gondwana breakup in the late Oxfordian (Fullerton et al., 1989). This mass fractured into various microcontinents, now incorporated into Southeast Asia and the Himalayas. On the Australian side, the Rowley Terrace and the northern Exmouth Plateau now form parts of the eastern and southern margins of the Argo Abyssal Plain, respectively. Prior to the late Oxfordian and during the rifting phase (probably since the Late Triassic), these were covered by shallow water or were coastal, and maintained a balance between steady subsidence and shallow-water deposition. As a result, accumulation of several thousand metres of Upper Triassic deltaic sediments, Rhaetian carbonates, and Lower and Upper Jurassic detritus, shallow-water carbonates, and coal measures occurred on parts of the northern Exmouth Plateau (Exon et al., 1982) and in the Rowley Sub-basin (Exon & Ramsay, 1990); tilting and uplifting of the Wombat Plateau caused its Jurassic emergence (von Rad et al., 1992). As attested by the record of calcareous nannofossils in the present study, open-marine conditions were established occasionally during the Jurassic in the incipient Rowley Sub-basin. On two occasions, calcareous nannoplankton are shown to have gained access to the sub-basin, as indicated by the early Toarcian and early Bajocian nannofloras from Tortoise and Mermaid Canyons, respectively. The early Toarcian nannoflora appeared some time between 183.5 and 181.5 Ma, and is considered to represent a marine ingression because of its apparently limited spatial extent. Whereas the early Bajocian nannoflora, which appeared at about 170.5 Ma, is thought to indicate a major transgression because of its much wider geographic distribution. The early Bajocian nannoflora correlates well with a global sealevel rise at 170.5 Ma (see eustatic curves of Haq et al., 1987).

The Toarcian ingression is thought to be related to an episode of increased subsidence resulting from local tectonism, though it may have been coincident with a global sealevel rise at about 183.5 or 181.5 Ma (see eustatic curves of Haq et al., 1987). Its spatial extent is difficult to estimate because of the limited geographic extent of our samples in the Rowley Sub-basin. Evidence for the early Bajocian transgression has been found in the Geraldton area of the Perth Basin (Newmarracarra Limestone), but it is probably widespread on the Australian western margin. A monospecific nannofloral 'assemblage' from Copperhead Canyon, possibly mid Jurassic in age, indicated stressful conditions which possibly were induced by shoaling during a subsequent regression.

Horsts formed during the rifting phase were eroded late in the Jurassic (forming the main or breakup unconformity). Subsequently, steady subsidence led to the deposition of mainly pelagic sediments during the Early Cretaceous (in a juvenile ocean), and the Late Cretaceous and onward (in a mature ocean). Post-breakup fully-marine sedimentation occurred in the Rowley Sub-basin as early as perhaps the Valanginian on the basis of nannofossil assemblages recorded herein. The nannofossil evidence from the Swan Canyon area suggests that at about this time a connection between the Rowley Sub-basin and southern Tethys existed. But details of the younger late Barremian/early Aptian assemblages from Copperhead Canyon do not unequivocally show such a connection.

During the mid and latest Campanian, the water masses over the Rowley Sub-basin contained associations characteristic of a location well within the Extratropical Nannoprovince. The higher-latitude (southern) limit of this province probably coincided with the southern limit of the Carnarvon Basin. During the late Campanian, the Papuan Basin, to the north of the Rowley Sub-basin, lay within the Tropical Nannoprovince (Shafik, 1990, 1993b).

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Acuturris scotus (Risatti, 1973) Wind & Wise *in* Wise & Wind, 1977

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Arkhangelskiella cymbiformis Vekshina, 1959

Arkhangelskiella specillata Vekshina, 1959

Axopodorhabdus cylindratus (Noël, 1965) Wind & Wise, 1977

Axopodorhabdus dietzmannii (Reinhardt, 1965) Wind & Wise, 1983

Axopodorhabdus macrograngulatus (Prins, 1969) Shafik, n. comb. (Basionym: *Podorhabdus macrogranulatus* Prins, 1969, pl. 3, fig. 6.)

Bidiscus rotatorius Bukry, 1969

Biscutum Black *in* Black & Barnes, 1959

Biscutum constans (Górka, 1957) Black *in* Black & Barnes, 1959

Biscutum coronum Wind & Wise *in* Wise & Wind, 1977

Biscutum intermedium Bown, 1987

Biscutum novum (Goy, 1979) Bown, 1987

Broinsonia parca (Stradner, 1963) Bukry, 1969 partim

Bukrylithus ambiguus Black, 1971

Calcicalithina oblongata (Worsley, 1971) Thierstein, 1971

Carniolithus Prins *in* Grün & others, 1974

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Carinolithus superbus (Deflandre, 1954) Prins *in* Grün & others, 1974

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Corollithion silvaradion Filewicz & others *in* Wind & Wise, 1977

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Ethmorhabdus gallicus Noël, 1965

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Grantarhabdus coronadventis (Reinhardt, 1966) Grün *in* Grün & Allemann, 1975

Hayesites irregularis (Thierstein *in* Roth & Thierstein, 1972) Applegate & others *in* Covington &Wise, 1987

Hayesites wisei (Thierstein *in* Roth & Thierstein, 1972)

Haqius circumradiatus (Stover, 1966) Roth, 1978

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Hexalithus noeliae Loeblich & Tappan, 1966

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Lithraphidites carniolensis Deflandre, 1963

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Markalius inversus (Deflandre *in* Deflandre & Fert, 1954) Bramlette & Martini, 1964

Micrantholithus Deflandre, 1954

Micrantholithus speetonensis Perch-Nielsen, 1979

Microrhabdulus belgicus Hay & Towe, 1963

Microrhabdulus decoratus Deflandre, 1959

Microrhabdulus helicoideus Deflandre, 1959

Microstaurus chiastius (Worsley, 1971) Grün *in* Grün & Allemann, 1975

Microstaurus quadratus Black, 1971

Micula concava (Stradner *in* Martini & Stradner, 1960) Bukry, 1969

Micula murus (Martini, 1961) Bukry, 1973

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Mitrolithus elegans Deflandre, 1954

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Nannoconus colomii (de Lapparent, 1931) Kamptner, 1938

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Nodosella Prins, 1969 ex Rood & others, 1973

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Petrarhabdus copulatus (Deflandre, 1959) Wind & Wise *in* Wise, 1983

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Prediscosphaera spinosa (Bramlette & Martini, 1964) Gartner, 1968

Prediscosphaera stoveri (Perch-Nielsen, 1968) Shafik & Stradner, 1971

Quadrum gothicum (Deflandre, 1959) Prins & Perch-Nielsen *in* Manivit & others, 1977

Quadrum trifidum (Stradner *in* Stradner & Papp, 1961) Prins & Perch-Nielsen *in* Manivit & others, 1977

Reinhardtites anthophorus (Deflandre, 1959) Perch-Nielsen, 1968

Reinhardtites levis Prins & Sissingh *in* Sissingh, 1977

Retecapsa angustiforata Black, 1971

Retecapsa neocomiana Black, 1971

Retecapsa incompta Bown & Cooper, 1989

Rhagodiscus angustus (Stradner, 1963) Reinhardt, 1971

Rhagodiscus asper (Stradner, 1963) Reinhardt, 1967

Rotelapillus laffittei (Noël, 1957) Noël, 1973

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Tegumentum octiformis (Köthe) Crux, 1989

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Vagalapillla matalosa (Stover, 1966) Thierstein, 1973

Vekshinella Loeblich & Tappan, 1963 emended Gartner, 1968

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Viminites Black, 1975

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Watznaueria britannica (Stradner, 1963) Reinhardt, 1964

Watznaueria manivitae Bukry, 1973

Watznaueria ovata Bukry, 1973

Zygodiscus bicrescenticus (Stover, 1966) Wind & Wise *in* Wise & Wind, 1977

Zygodiscus diplogrammus (Deflandre *in* Deflandre & Fert, 1954) Gartner, 1968

Zygodiscus erectus (Deflandre *in* Deflandre & Fert, 1954) Manivit, 1971 emended Bralower, 1989

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Palynology of Mesozoic dredge samples from the North West Shelf

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Palynological examination was carried out on 34 sedimentary rock samples dredged from the offshore Rowley Terrace (Canning Basin), Scott and Exmouth Plateaux, and northern Carnarvon Terrace (Carnarvon Basin). The samples represent (1) an Upper Triassic (Rhaetian) to Middle Jurassic (Bathonian) sequence of paludal to restricted-marine character on the Rowley Terrace, with a sample gap in the Early Jurassic; (2) an Upper Jurassic (Oxfordian–Kimmeridgian) sequence of shallow-marine character on the Rowley Terrace; and (3) a Lower Cretaceous (Valanginian–Aptian) sequence of open-marine character on the Carnarvon Terrace. Most samples yielded rich to very rich and varied dinocyst and spore-pollen floras, and were dated with reasonable confidence. Long-term changes are detected in environments of deposition from different areas, and some of the changes are suggested to be of eustatic origin.

Introduction

The samples discussed here were dredged during 1990 from several offshore localities at various water depths by BMR Cruises 95 and 96 using *Rig Seismic*. Thirty-four dredge samples were selected for palynological examination. Sample locations are plotted in Figure 1, and details of samples are listed in Table 1 (Sample 96DR35/3, MFP-9356, from the western Exmouth Plateau yielded no microfossils and is not discussed). More sample information is given by Colwell et al. (1990; 1994 — this issue) and Exon & Ramsay (1990). The bulk of the samples comes from the Rowley Terrace (Canning Basin) and Carnarvon Terrace (Carnarvon Basin).

A preliminary palynological report of the samples was given by Burger (1991), and an updated summary on the basis of new insight into geological ages of palynological zonal intervals is given here. A fuller review of individual dinoflagellate and spore–pollen assemblages — too detailed to be included here — will be offered in a future publication.

The samples are sedimentary rock fragments of varying sizes. Splits of these samples were processed in the laboratory. The organic residue extracted was stained with safranin and mounted in PVA on permanent slides (coverslips 22x22 mm). The slides are stored in the Palynological Laboratory of the Australian Geological Survey Organisation (AGSO), formerly the Bureau of Mineral Resources, Geology & Geophysics (BMR), in Canberra.

The Mesozoic palynological sequence of the North West Shelf in Australia includes an abundant and varied dinoflagellate component. The author recovered extremely rich palynological assemblages from the dredge samples; many slides contain in excess of 20 000 specimens, and the documented microflora counts almost 200 genera and about 300 species. The biostratigraphic potential of these microfossils has been utilised for many years by palynologists engaged by ESSO Australia Ltd, Woodside Offshore Petroleum Pty Ltd, the Geological Survey of Western Australia, and by other organisations. Over the years regional biostratigraphic schemes, based on extensive (and often unpublished) in-house work were developed for parts of the Mesozoic in several Australian onshore and offshore sedimentary basins.

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Helby et al. (1987) compiled the first comprehensive Mesozoic biostratigraphy for Australia, being the result of study of numerous drillholes in Australia and Papua New Guinea by a number of palynologists. Their dinoflagellate scheme includes 5 Triassic, 12 Jurassic, and 18 Early Cretaceous zonal intervals; their spore–pollen scheme for Western Australia comprises 8 Triassic, 6 Jurassic, and 4 Early Cretaceous zones. The authors defined individual zonal intervals as Oppel, interval, range, or acme zones, such as defined by Hedberg (1976).

The assemblages from the dredge samples are readily accommodated within those zonal schemes, and biostratigraphic affinities of samples are set out in Table 2. In several instances, comparisons are made also with the detailed schemes proposed for the Jurassic (Filatoff, 1975) and Early Cretaceous (Backhouse, 1987, 1988) of the Perth Basin. For a few assemblages, age determinations based on spores and pollen disagree with those based on dinoflagellates. This is caused most likely by palynomorphs whose *in situ* status is uncertain. Their presence may usually be regarded as secondary (reworked), and in some instances may indicate that range limits given by Helby et al. (1987) need to be revised.

Biostratigraphic affinities of spore–pollen assemblages are not always obvious, in particular in those intervals of the sequence where zonal boundaries are defined by changes in relative abundance of selected taxa. Detailed studies of contemporaneous sequences in eastern Australia have demonstrated that such changes are useful for broad correlations but cannot always be applied everywhere, and often extend over time intervals which equal or even exceed that of an entire zone. On the North West Shelf, these problems arise chiefly in the Early and Middle Jurassic; and where necessary the author draws comparisons with the well-documented contemporaneous spore–pollen record from eastern Australia.

In the following discussion samples are grouped together by age and palynological zonal affinity. Representative species from the spore–pollen and the marine microphytoplankton floras are illustrated in Figures 9 to 11.

Rhaetian (Late Triassic): Rowley Terrace

Affinity, age, and palaeoenvironments

Four samples from the Rowley Terrace yielded spore–pollen assemblages which fall within the lower range-interval of the Circumpolles pollen group and predate the oldest known appearance of *Callialasporites dampieri* (Fig. 2). The scarce presence of Circumpolles indicates

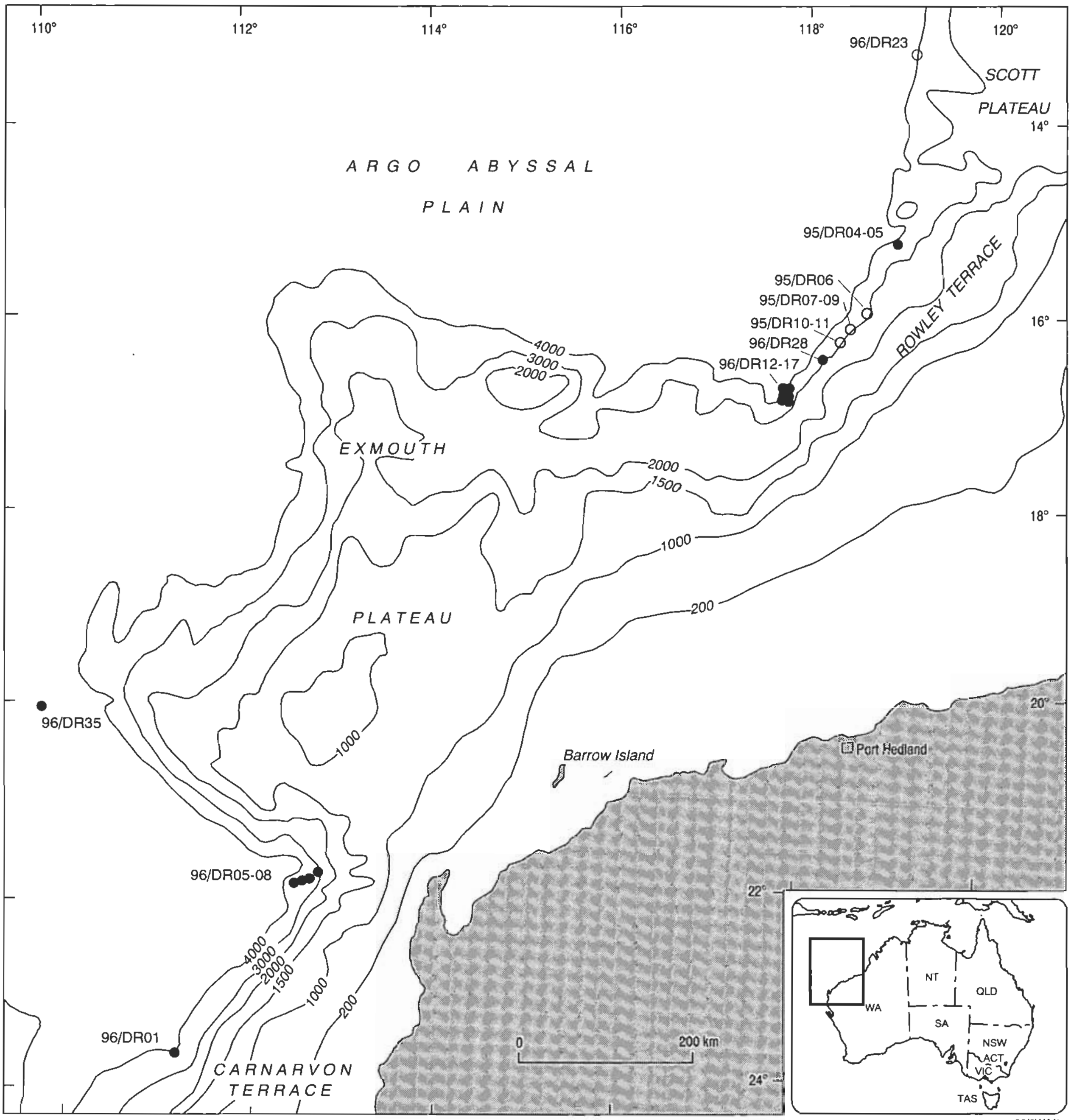

Figure 1. Study region and locations of dated dredge samples.

that the assemblages are all older than the *Corollina torosa* Zone, whose lower limit is placed at or near the Triassic–Jurassic boundary (Helby et al., 1987).

Sample	Geological age	Palaeoenvironments
Rowley Terrace		
1. 95/DR07/11A		Restricted marine
2. 95/DR10/4A	All samples	Restricted marine
3. 95/DR07/4A	dated Rhaetian	Restricted marine
4. 96DR28/14		Paludal

20/OWA/2

Figure 2. Details of Upper Triassic dredge samples.

Despite the low number of species recovered, the marine assemblages may be placed in the upper *"Shublikodinium"* Dinoflagellate Superzone. Burger (1991) dated Sample 2 as Rhaetian, although the spores and pollen, as they represent the *Samaropollenites speciosus–Minutosaccus crenulatus* zonal interval, indicated a pre-Rhaetian age. Fresh conodont evidence (Nicoll & Foster, 1994) now places the upper *M. crenulatus* Zone also within the Rhaetian.

As shown in Figure 2, the composition of the assemblages suggests paludal deposition for the oldest Sample 4, and restricted (coastal) marine for the other samples.

Toarcian (Early Jurassic)

Samples 1 to 3 were dredged relatively closely together

Table 1. List of dredge samples in MFP numerical order, and details of dredge sites.

Palynological number	BMR field number	Latitude South	Longitude East	Offshore Region	Water Depth
MFP- 9235	95/DR04/16A	15°22'	118°58'	Rowley Terrace	4200-3750 m
9236	95/DR05/5A	15°23'	118°58'	Rowley Terrace	3850-3300
9237	95/DR05/9A	15°23'	118°58'	Rowley Terrace	3850-3300
9238	95/DR06/1A	16°01'	118°46'	Rowley Terrace	3400-3100
9239	95/DR07/4A	16°18'	118°23'	Rowley Terrace	4530-3900
MFP- 9240	95/DR07/11A	16°18'	118°23'	Rowley Terrace	4530-3900 m
9241	95/DR07/14A	16°18'	118°23'	Rowley Terrace	4530-3900
9242	95/DR09/2C	16°19'	118°25'	Rowley Terrace	4200-3350
9243	95/DR10/4A	16°24'	118°10'	Rowley Terrace	5325-4090
9244	95/DR11/3A	16°26'	118°11'	Rowley Terrace	4000-3300
MFP- 9244	95/DR11/7A	16°26'	118°11'	Rowley Terrace	4000-3300 m
9246	96/DR01/3A	23°42'14"	111°16'30"	Northern Carnarvon Terrace	4125-3700
9247	96/DR05/5	21°52'07"	112°43'45"	Northern Carnarvon Terrace	4200-3150
9248	96/DR05/6	21°52'07"	112°43'45"	Northern Carnarvon Terrace	4200-3150
9249	96/DR06/2	21°52'06"	112°43'05"	Northern Carnarvon Terrace	4100-
MFP- 9250	96/DR07/7	21°52'40"	112°43'07"	Northern Carnarvon Terrace	3700- m
9251	96/DR07/12	21°52'40"	112°43'07"	Northern Carnarvon Terrace	3700-
9252	96/DR08/3	.21°50'56"	112°45'41"	Northern Carnarvon Terrace	3550-
9253	96/DR08/4	21°50'56"	112°45'41"	Northern Carnarvon Terrace	3550-
9254	96/DR08/5	21°50'56"	112°45'41"	Northern Carnarvon Terrace	3550-
MFP- 9255	96/DR12/1	16°54'30"	117°36'05"	Rowley Terrace	5000- m
9256	96/DR13/4	16°54'24"	117°38'27"	Rowley Terrace	4150-
9257	96/DR14/5	16°50'01"	117°35'13"	Rowley Terrace	5234-
9258	96/DR14/10	16°50'01"	117°35'13"	Rowley Terrace	5234-
9259	96/DR15/2	16°50'06"	117°37'59"	Rowley Terrace	4200-
MFP- 9260	96/DR16/2	16°44'05"	117°33'04"	Rowley Terrace	5612- m
9261	96/DR17/5	16°44'53"	117°35'54"	Rowley Terrace	4500-
9262	96/DR17/9	16°44'53"	117°35'54"	Rowley Terrace	4500-
9263	96/DR28/11B	16°18'30"	118°23'25"	Rowley Terrace	4530-
9264	96/DR28/14	16°18'30"	118°23'25"	Rowley Terrace	4530-
MFP- 9265	96/DR28/15	16°18'30"	118°23'25"	Rowley Terrace	4530- m
9266	96/DR23/2	13°26'10"	119°22'15"	Western Scott Plateau	5200-
9267	96/DR35/2	20°10'52"	109°54'51"	Western Exmouth Plateau	5025-
9356	96/DR35/3	20°10'52"	109°54'51"	Western Exmouth Plateau	5052-

20/OWA/9

from the Rowley Terrace and represent a narrow biostratigraphic interval. Samples 4 and 5 were dredged from other locations; their palynological assemblages are relatively poor and do not indicate a minimum age limit (Fig. 3).

Affinity and age

The spore–pollen assemblages from Samples 1 to 3 fall within the lower range-interval of *Callialasporites dampieri*, and predate the earliest appearance of *Antulsporites saevus*. This interval has been recognised in both eastern and western Australia (Fig. 4). Balme (1957) first defined it in the west as Microflora I, and Burger (1976) referred to it in northeastern Australia as Unit J2. The first appearance of *C. dampieri* is a significant biostratigraphic horizon in Australia; in western Australia it coincides with the lower limit of Filatoff's (1975) *Classopollis anasillos* Zone, and in northeastern Australia marks the lower limit of Unit PJ3 (Filatoff & Price, 1988) and the upper limit of the *Classopollis classoides* Zone (Reiser & Williams, 1969).

Sample	Geological age	Palaeoenvironments
Rowley Terrace		
1. 95/DR06/1A	Toarcian	Open (shallow) marine
2. 95/DR07/14A	Toarcian	Restricted marine
3. 95/DR09/2C	Toarcian	Paludal
Carnarvon Terrace		
4. 96DR06/2	Toarcian or younger	Uncertain
Scott Plateau		
5. 96DR23/2	Toarcian or younger	Paludal

20/OWA/3

Figure 3. Details of Lower Jurassic dredge samples.

Some difficulties arise with the correlation of various spore–pollen zones within this biostratigraphic interval. Helby et al. (1987) placed Filatoff's (1975) *Classopollis anasillos* Zone within the basal part of their own *Callialasporites turbatus* Zone. However, the boundaries of their zone are not sharply defined, whereas Filatoff's zone is readily identified. Species ranges in the Perth Basin given by Filatoff indicate that his zone largely coincides with the *Corollina torosa* Zone of Helby et al. (1987). The BMR dredge samples also suggest that the *Classopollis anasillos* Zone falls largely within the upper *Corollina torosa* Zone, but may overlap with the lower *Callialasporites turbatus* Zone (Burger, 1990b).

The evidence for the age of this interval is circumstantial, but reasonably consistent. Indirect spore–pollen and other

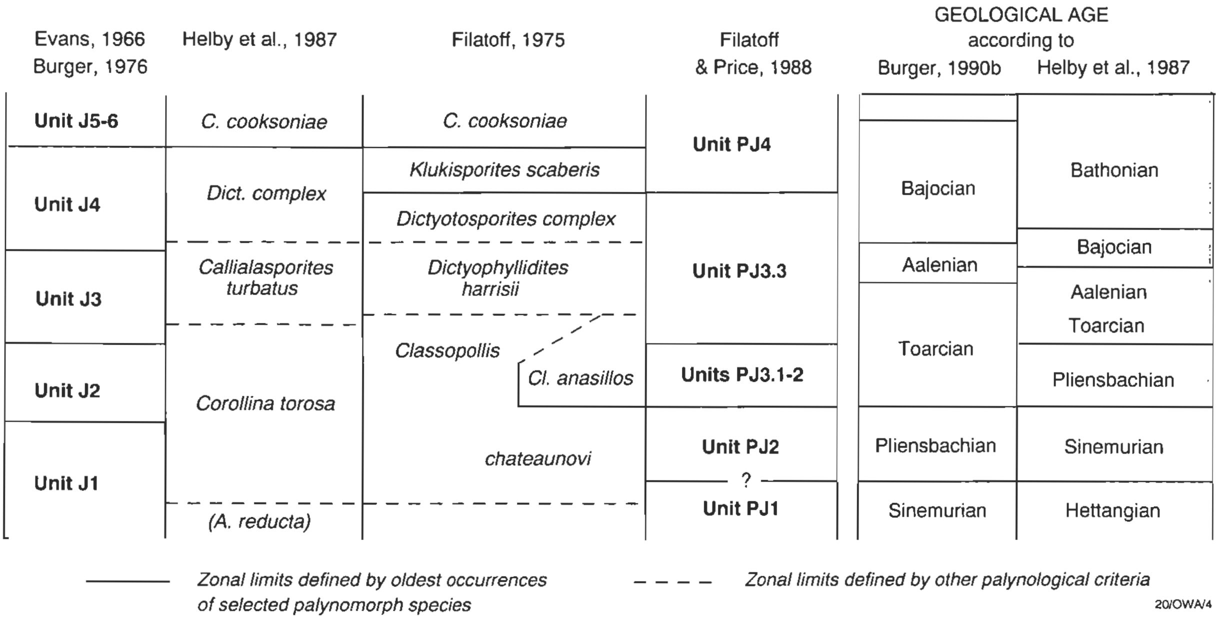

Figure 4. Early and Middle Jurassic spore-pollen zonal schemes for western and eastern Australia.

evidence suggest a Toarcian age for the Unit J2 interval in eastern Australia (McKellar, 1974; Exon & Burger, 1981; Filatoff & Price, 1988; Burger, 1990b), and this agrees with a Toarcian age which Filatoff (1975) gave for his *Classopollis anasillos* Zone. Samples 1 to 3 fall within the upper *Corollina torosa* Zone/Unit J2 and are thus dated as Toarcian.

The dinoflagellate record from this interval is extremely poor, and zonal assignments for Samples 1 and 2 are based chiefly on the presence of *Susadinium* sp. A, a form restricted to a narrow interval straddling the upper limit of the *Dapcodinium priscum* Zone. Helby et al. (1987) dated that zone chiefly as basal Jurassic, based on Toarcian occurrences of *Susadinium* in northern Europe.

Palaeoenvironments

The assemblages suggest that paludal conditions (Sample 3) preceded restricted-marine deposition on the Rowley Terrace (Samples 1 and 2). Restricted-marine environments occurred on the Carnarvon Terrace and paludal on the Scott Plateau, but neither instance can be dated with confidence.

Aalenian–Bajocian (Middle Jurassic): Rowley Terrace

Eight, and possibly nine samples have been dredged from the rock sequence falling within this time interval (Fig. 5). Geographically, Samples 2 to 4 are grouped comparatively close and Sample 1, located somewhat farther north, falls within the same time interval. Samples 5 to 8 are also geographically close, and Sample 9 from farther to the west yielded a very poor assemblage and may be of a different age.

Affinity and age

The spore–pollen assemblages recovered from Samples 1 to 8 fall within the interval bounded by the successive first appearances of *Antulsporites saevus* and *Contignisporites* spp. Time relations of various zones within that interval have not yet been fully resolved, and the assemblages are compared with the eastern Australian record, where various zonal boundaries are defined by first appearances of selected species (see Fig. 4).

Sample	Geological age	Palaeoenvironments
Rowley Terrace		
1. 95/DR04/16A	Aalenian-Bajocian	Paludal
2. 96/DR28/15	Bajocian	Paludal
3. 95/DR11/7A	Aalenian-Bajocian	Restricted marine
4. 95/DR11/3A	Aalenian-Bajocian	Restricted marine
5. 96DR14/10	Bajocian	Restricted marine
6. 96DR16/2	Bajocian	Restricted marine
7. 96DR17/5	Bajocian	Restricted marine
8. 96DR17/9	Bajocian	Paludal
Exmouth Plateau		
9. 96DR35/2	Toarcian or younger	Restricted marine

20/OWA/5

Figure 5. Details of Middle Jurassic dredge samples.

Burger (1976) subdivided this sequence in Queensland into Units J3 and J4 on the first appearance of *Camarozonosporites clivosus*, and those units can be recognised in several eastern Australian basins. Species ranges given by Filatoff (1975) suggest that in the Perth Basin Unit J3 corresponds with his upper *Classopollis chateaunovi* and *Dictyophyllidites harrisii* Zones, and that Unit J4 may be correlated broadly with his *Dictyotosporites complex–Klukisporites scaberis* zonal interval. Species ranges given by Helby et al. (1987) suggest that Units J3-4 coincides approximately with the interval which includes the uppermost *Corollina torosa* Zone and the *Callialasporites turbatus–Dictyotosporites complex* zonal interval of those authors.

These and earlier-mentioned pan-continental correlations are here proposed as a working hypothesis for zonal age determinations, but they need corroboration, as they are based on the unproven assumption that individual index species first appeared in the record at the same time in eastern and western Australia.

The *Callialasporites turbatus* Zone is dated Toarcian to early Bajocian, partly on foraminifera from the North West Shelf, and partly on dinoflagellates (see Helby et al., 1987). The *Dictyotosporites complex* Zone (*sensu* Filatoff 1975) and the partly overlapping *Dissiliodinium caddaense* Dinoflagellate Zone are associated with early Bajocian ammonites in the Perth Basin (Filatoff, 1975; Helby et al., 1987; Hall, 1989). On these data, Samples 1 to 4 are dated (late?) Aalenian to Bajocian.

Opinions differ as to the time at which *Contignisporites* spp. first appeared in Australia. McKellar (1974) favoured an (early) Bajocian age for that event, which is compatible with other evidence discussed in Burger (1990b, and in press). That age is here taken as a possible minimum age for the *Dictyotosporites complex* Zone (*sensu* Helby et al. 1987), and Samples 5 to 8 are dated as Bajocian. Sample 6 is dated on the spore–pollen evidence, as the age suggested for its marine assemblage may be too young. Sample 9 cannot be dated with confidence, and may be not older than Bathonian.

Palaeoenvironments

The dinoflagellate record is on average very poor. No systematic geographic pattern of palaeoenvironments is apparent. The assemblages from Samples 5 to 7 indicate restricted-marine environments; Sample 8 a paludal one; and Samples 3 and 4 represent restricted (coastal) marine environments. The paludal setting indicated for Samples 1 and 2 may coincide with an interval of poor marine record between the *Dissiliodinium caddaense* and *Caddasphaera halosa* Dinoflagellate Zones observed by Helby et al. (1987).

Bajocian–Kimmeridgian (Middle–Late Jurassic): Rowley Terrace

Six, and possibly seven samples were collected from the sequence falling in this time interval (Fig. 6). Samples 4 to 6 from the southernmost Rowley Terrace were dredged from a relatively small area (east side of Swan Canyon) and represent a narrow biostratigraphic interval. Samples 5 and 6, from farther north, are also geographically close, and are somewhat younger. Sample 7 yielded a very poor assemblage and cannot be dated with confidence.

Sample	Geological age	Palaeoenvironments
Rowley Terrace		
1. 95/DR12/1	Oxfordian-(Kimmeridgian)	Open marine
2. 96/DR13/4	Oxfordian-(Kimmeridgian)	Open marine
3. 96DR15/2	Oxfordian-(Kimmeridgian)	Restricted marine
4. 96DR14/5	Oxfordian-(Kimmeridgian)	Restricted marine
5. 95/DR05/5A	Bajocian-Bathonian	Paludal
6. 95/DR05/9A	Bathonian	Restricted marine
7. 96DR28/11B	Middle Jurassic	Restricted marine?

20/OWA/6

Figure 6. Details of Middle and Upper Jurassic dredge samples.

Affinity and age

The spore–pollen assemblages recovered from Samples 1 to 6 fall within the lower range-interval of the genus *Contignisporites*, and predate the first appearance of the genera *Cicatricosisporites* and *Ruffordiaspora*. This extended sequence interval is known in both eastern and western Australia, and is dated as Bajocian to Tithonian (see next section). It was first defined by Balme (1957) as Microflora IIa in the Perth and Canning Basins. Unfortunately, Samples 1 to 7 all yielded more or less restricted spore–pollen assemblages, which lacked crucial index species. The associated zones defined for this interval by Filatoff (1975), Backhouse (1978), and Helby et al. (1987) are therefore not recognised. The samples are dated chiefly on the marine evidence except Sample 5, which yielded only spores and pollen and is taken to represent the *Contignisporites cooksoniae* Zone (Burger, 1991).

Contemporaneous rock sequences in Western Australia and Papua New Guinea have yielded abundant dinoflagellate assemblages. Helby et al. (1987) dated marine zones from that interval on evidence from ammonites, belemnites, and benthonic foraminifera found in Western Australia and Papua New Guinea. Taking due account of various uncertain factors, they dated their *Wanaea spectabilis* Zone as Oxfordian, the lower limit of their *Wanaea digitata* Zone as (middle) Callovian, and their *Dingodinium jurassicum* Zone as Tithonian. On the basis of overseas evidence, Burger (1990b) suggested slightly older ages for various zones, and they are in keeping with independent (indirect) evidence for the age of associated spore–pollen zones in eastern Australia. The *W. digitata* Zone is taken here to be Bathonian, the *W. spectabilis* Zone Callovian, and the *D. jurassicum* Zone (late) Oxfordian to (early) Kimmeridgian.

From species ranges given by Helby et al. (1987), Sample 6 is associated with the *Energlynia indotata–Wanaea digitata* zonal interval. Sample 7 yielded a poor assemblage of uncertain affinity, and may be younger. The dinoflagellates from Samples 1 to 4 differ radically from those recovered from the previous intervals, and they are associated with the *Omatia montgomeryi–Dingodinium jurassicum* zonal interval. A large batch of newly incoming species indicates a sampling gap, which may represent the Late Jurassic Post Rift Unconformity (PRU) on the Exmouth Plateau, alluded to by von Rad et al. (1992). Sample 2 yielded both Middle and Late Jurassic elements, and the older elements are regarded as contamination, probably due to erosion associated with the PRU.

Palaeoenvironments

From the composition of their assemblages, Samples 5 and 6 are taken to represent Middle Jurassic (pre-breakup) coastal marine to paludal environments on the Rowley Terrace, and Samples 1 to 4 restricted to open and (immediately post-breakup) shallow-marine environments farther south. Sample 7 also indicates restricted-marine environments, but its stratigraphic position is uncertain.

Early Cretaceous: Carnarvon Terrace

This interval is represented by a series of eight samples, of which Samples 1 to 7 were dredged from a small area on the norther Carnarvon Terrace, and Sample 8 farther to the southwest (Fig. 7).

Sample	Geological age	Palaeoenvironments
Carnarvon Terrace		
1. 96DR01/3A	Aptian	Restricted marine
2. 96/DR08/4	Barremian	Open marine
3. 96DR08/3	(Hauterivian)-Barremian	Open marine
4. 96DR07/12	Hauterivian	Open marine
5. 96DR08/5	Hauterivian	Open marine
6. 96DR07/7	Valanginian-Hauterivian	Open marine
7. 96DR05/6	Valanginian-Hauterivian	Open marine
8. 96DR05/6	Valanginian	Open marine

20/OWA/7

Figure 7. Details of Lower Cretaceous dredge samples.

Affinity and age

All samples yielded spore–pollen assemblages which fall within the lower range-interval of *Cicatricosisporites/Ruffordiaspora* (Sample 96DR06/2 could not be dated, see Fig. 3). The first appearance of this group of spores is one of the most significant biostratigraphic marker horizons in the Australian Mesozoic. Balme (1957) first identified the sequence in question in Western Australia as Microflora IIb, and Backhouse (1978) redefined it in the Perth Basin as the *Biretisporites eneabbaensis* Zone. An upper limit for that zone was not defined until Backhouse (1988) established a new *Balmeiopsis limbata* Zone in the Perth Basin, of Neocomian and (early) Aptian age (Fig. 8).

Backhouse correlated the *Balmeiopsis limbata* Zone with the combined *Foraminisporis wonthaggiensis* and *Cyclosporites hughesii* zonal intervals in eastern Australia. Helby et al. (1987) recognised the *C. hughesii* Zone also in Western Australia, and this means that it partly overlaps with the *B. limbata* Zone. The zonal association of Samples 1 to 8 is not yet clear. None of the samples yielded *Foraminisporis asymmetricus* of the *C. hughesii* Zone, but Samples 3 and 4 yielded *Balmeiopsis robusta* of the *B. limbata* Zone. For the time being, this paper adopts a "truncated" version of the *B. limbata* Zone to avoid overlapping definitions, but this zone should be reviewed for the sake of a more accurate pan-Australian correlation.

The marine record indicates a large sampling gap at the Jurassic–Cretaceous boundary, which includes the absence of the Tithonian and Berriasian. Zonal association of the assemblages from Samples 1 to 8 can be determined within reasonably narrow limits (Table 2). There is general agreement on the ages of the Early Cretaceous zones. Helby et al. (1987) referred to ammonite evidence for an (early) Berriasian age of their *Batioladinium reticulatum* Zone, and noted the association of Hauterivian–Barremian foraminifera with the *Muderongia testudinaria* and *Muderongia australis* Zones. In northeastern Australia, the *Ovoidinium cinctum–Diconodinium davidii* zonal interval (as well as the *Cyclosporites hughesii* spore–pollen Zone) are associated with Aptian ammonites and pelecypods (Morgan, 1980a; Burger, 1980, 1990a).

Palaeoenvironments

The samples indicate that during the Early Cretaceous open (shallow) marine enviroments prevailed on the northern Carnarvon Terrace.

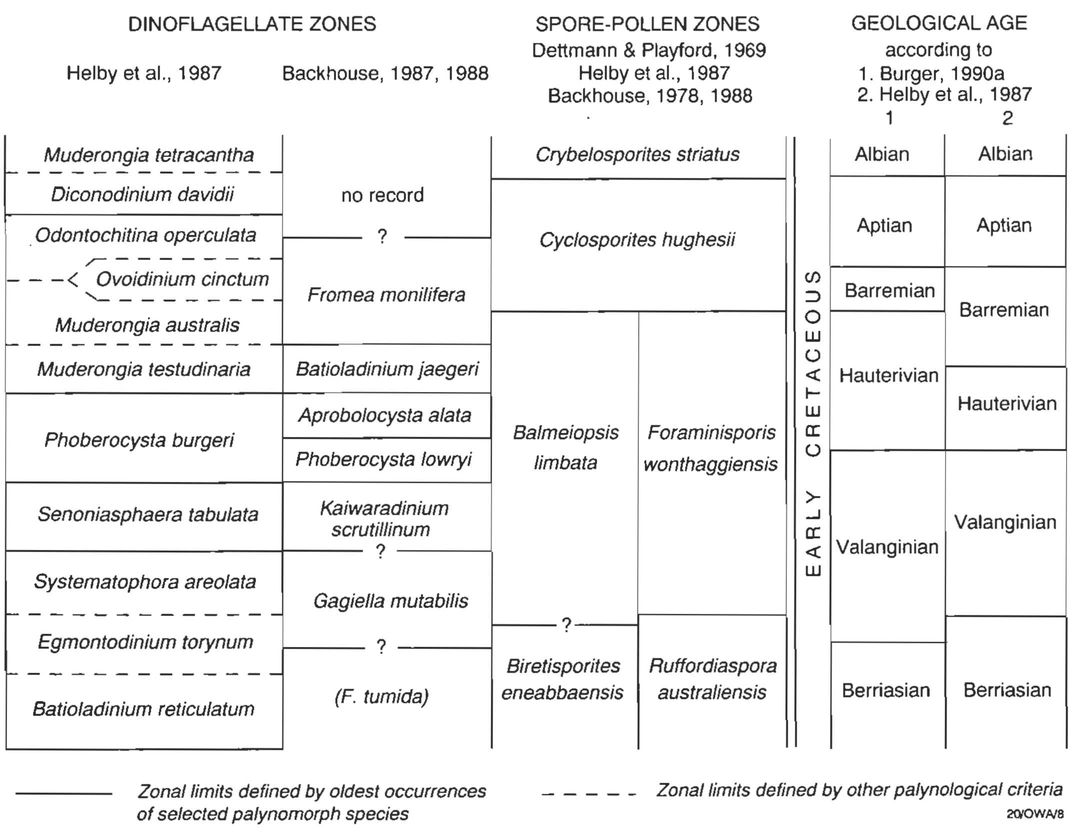

Figure 8. Early Cretaceous palynological zonal schemes for eastern and western Australia.

Table 2. Biostratigraphic affinity, geological age, and palaeoenvironments of dredged samples.

Palyn. number (MFP)	BMR field number	Associations to palynological zones: Spores and pollen grains	Associations to palynological zones: Dinoflagellates	Environm. of deposition	Estimated geological age
Carnarvon Terrace					
9246	96DR01/3A	*Cyclosporites hughesii?*	*Odontochitina operculata*	3	Aptian
9253	96DR08/4	*Balmeiopsis limb.* -lower *Cyclosp. hughesii*	*Muderongia australis*	4	Barremian
9252	96DR08/3	*Balmeiopsis limb.* -lower *Cyclosp. hughesii*	*Muderongia australis*	4	(Hauterivian)-Barremian
9251	96DR07/12	*Balmeiopsis limb.* -lower *Cyclosp. hughesii*	*Muderongia testudinaria-australis*	4	Hauterivian
9254	96DR08/5	*Biretisporites eneabb.-Balmeiopsis limbata*	*Muderongia testudinaria-australis*	4	Hauterivian
9250	96DR07/7	*Balmeiopsis limbata?*	*Phob. burgeri-Muderong. testudinaria*	4	Valanginian-Hauterivian
9247	96DR05/5	*Balmeiopsis limbata?*	*Batioladin. reticulatum-Senon. tabulata*	4	Valanginian-Hauterivian
9248	96DR05/6	upper *Biretisporites eneabbaensis?*	*Systematoph. areolata-Senon. tabulata*	4	Valanginian
9249	96DR06/2	upper *Corollina torosa (maximum)*	?	3-4?	?
Rowley Terrace					
9255	96DR12/1	*Murospora florida* (maximum)	upper *Om. montgom.-Ding. jurassicum*	4	Oxfordian-Kimmeridgian
9256	96DR13/4	*Murospora florida* (maximum)	upper *Om. montgom.-Ding. jurassicum*	4	Oxfordian-Kimmeridgian
9259	96DR15/2	?	upper *Om. montgom.-Ding. jurassicum*	3	Oxfordian-Kimmeridgian
9257	96DR14/5	*Murospora florida* (maximum)	*Omatia montgom.-Ding. jurassicum*	3	Oxfordian-Kimmeridgian
9260	96DR16/2	upper *Dictyotosporites complex*	*Energl. indolata-Wanaea digitata*	3	Bajocian
9258	96DR14/10	*Dictyotosporites complex*	*Caddasphaera halosa*	3	Bajocian
9261	96DR17/5	upper *Dictyotosporites complex*	*Dissiliodinium caddaense*	3	Bajocian
9262	96DR17/9	upper *Dictyotosporites complex*	?	2	Bajocian
9237	95/DR05/9A	*Contignisporites cooksoniae*	*Energl. indotata-Wanaea digitata*	3	Bathonian
9236	95/DR05/5A	*Contignisporites cooksoniae*	---	2	Bajocian-Bathonian
9235	95/DR04/16A	*Dictyotosporites complex*	?	2	Aalenian-Bajocian
9265	96DR28/15	*Dictyotosporites complex*	?	2	Bajocian
9245	95/DR11/7A	lower *Dictyotosporites complex*	*Dissilliodinium caddaense?*	3	Aalenian-Bajocian
9244	95/DR11/3A	*Callialasporites turbàtus*	*Dissilliodinium caddaense*	3	Aalenian-Bajocian
9238	95/DR06/1A	upper *Corollina torosa?*	upper *Dapcodinium priscum*	3-4	Toarcian
9241	95/DR07/14A	upper *Corollina torosa*	upper *Dapcodinium priscum*	3	Toarcian
9242	95/DR09/2C	upper *Corollina torosa*	?	2	Toarcian
9263	96DR28/11B	upper *Corollina torosa* (maximum)	*Rigaudella aemula* (maximum)	3?	Mid-Jurassic?
9239	95/DR07/4A	lower *Ashmoripollis reducta*	*Wanneria listeri-Heibergella balmei*	3	Rhaetian
9240	95/DR0711A	*Minutsaccus crenul.-Ashmoripoll. reducta*	lower *Dapcodinium priscum*	3	Rhaetian
9243	95/DR10/4A	*Samarop. speciosus-Minutosaccus crenul.?*	*Rhaetogonyaulax rhaetica*	3	Rhaetian
9264	96DR28/14	*Minutosaccus crenulatus*	?	2	Rhaetian
Exmouth Plateau					
9267	96DR35/2	upper *Corollina torosa* (maximum)	?	3	Maximum Toarcian
Scott Plateau					
9266	96DR23/2	upper *Corollina torosa* (maximum)	?	2	Maximum Toarcian

Environments of deposition: 1 - nonmarine; 2 - paludal; 3 - restricted marine; 4 - open marine

20/OWA/10

Palaeoenvironments (Table 2)

Late Triassic

Rowley Terrace

Sample 96DR28/14 was dredged from a sequence of fluvio–deltaic sandstone, siltstone, and limestone (Colwell et al. 1990; Colwell et al., 1994 — this issue), and indicates at most mildly brackish environments suggestive of a more inland (coastal) facies than indicated by the three samples mentioned below.

Samples 95/DR07/4A, 95/DR07/11A, and 95/DR10/4A were dredged from what Exon & Ramsay (1990) described as a sequence of shallow-marine and deltaic clastic sediments more than 1000 m thick, which is overlain by trachytes and rhyolites. They are regarded as slightly younger than Sample 96DR28/14 and indicate restricted brackish–saline conditions, possibly with only intermittent connection to open-marine or continental-shelf environments.

Early Jurassic

Rowley Terrace

Samples 95/DR06/1A, 95/DR07/14A, and 95/DR09/2C represent a sequence of alluvial-plain sandstone, siltstone, mudstone, and red-beds unconformably overlying sediments of Triassic age (Exon & Ramsay, 1990; Colwell et al., 1994 — this issue). They yielded assemblages which suggest brackish-coastal to restricted-marine environments, with intermittently opening corridors to the open ocean.

Scott and Exmouth Plateaux

Samples 96DR23/2 and 96DR35/2 from the western margins of the two plateaux could not be dated accurately, but if their estimated maximum Early Jurassic age is correct they were dredged, respectively, from alluvial plain and rifting paralic strata sequences (Colwell et al., 1990). Both samples yielded fragments of poorly preserved unidentified microplankton types, and this may indicate

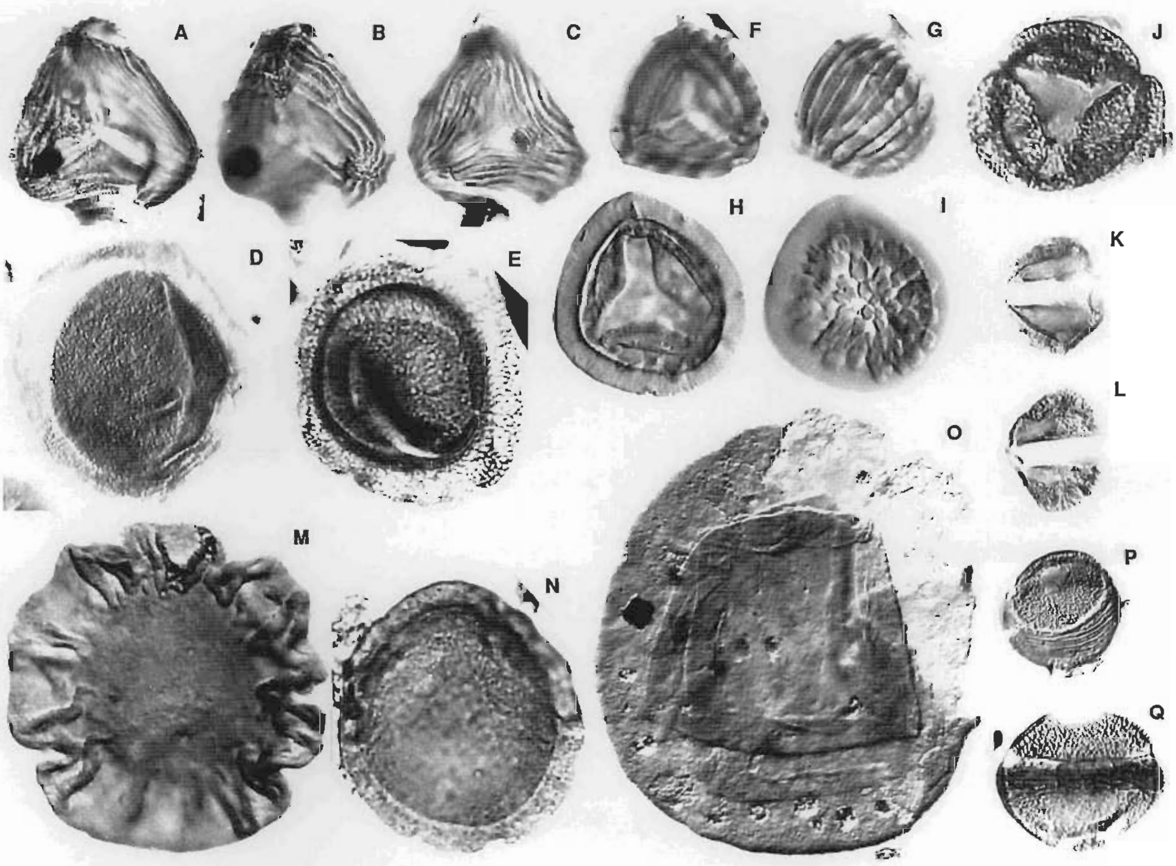

Figure 9. Selected Mesozoic zonal index spores and pollen grains (magnification 700x). Stage coordinates given for Zeiss Axioskop microscope no. 77110. A–C. *Ruffordiaspora australiensis* (Cookson) Dettmann & Clifford; CPC31081 (MFP9252–2, 928/066). D. *Dictyotosporites complex* Cookson & Dettmann; CPC31085 (MFP9265–1, 1053/218). E. *Crybelosporites stylosus* Dettmann; CPC31115 (MFP9250–2, 970/130). F, G. *Contignisporites cooksoniae* (Balme) Dettmann; CPC31110 (MFP9253–1, 913/116). H, I. *Antulsporites saevus* (Balme) Archangelsky & Gamerro; CPC31108 (MFP9261–1, 966/172). J. *Microcachryidites antarcticus* Cookson; CPC31152 (MFP9253–1, 1057/189). K, L. *Minutosaccus crenulatus* Dolby, in Dolby & Balme; CPC31142 (MFP9243–1, 935/228). M. *Callialasporites dampieri* (Balme) Dev; CPC31129 (MFP9261–1, 996/069). N. *Balmeiopsis robusta* Backhouse; CPC31138 (MFP9251–2, 963/082). O. *Callialasporites turbatus* (Balme) Schulz; CPC31130 (MFP9262–2, 1091/070). P. *Classopollis anasillos* Filatoff; CPC31158 (MFP9244–2, 930/080). Q. *Classopollis* cf. *C. chateaunovii* Reyre; CPC31155 (MFP9244–2, 898/222).

(one or more) shallow-marine incursions in the northern and western study area.

Middle Jurassic

Rowley Terrace

Samples 95/DR04/16A, 95/DR05/5A and 95/DR05/9A, 95/DR11/3A and 95/DR11/7A, and 96DR28/11B and 96DR28/15 were all dredged from the same alluvial and red-bed sequence referred to above, and indicate fluctuating, coastal-paludal to restricted-marine environments.

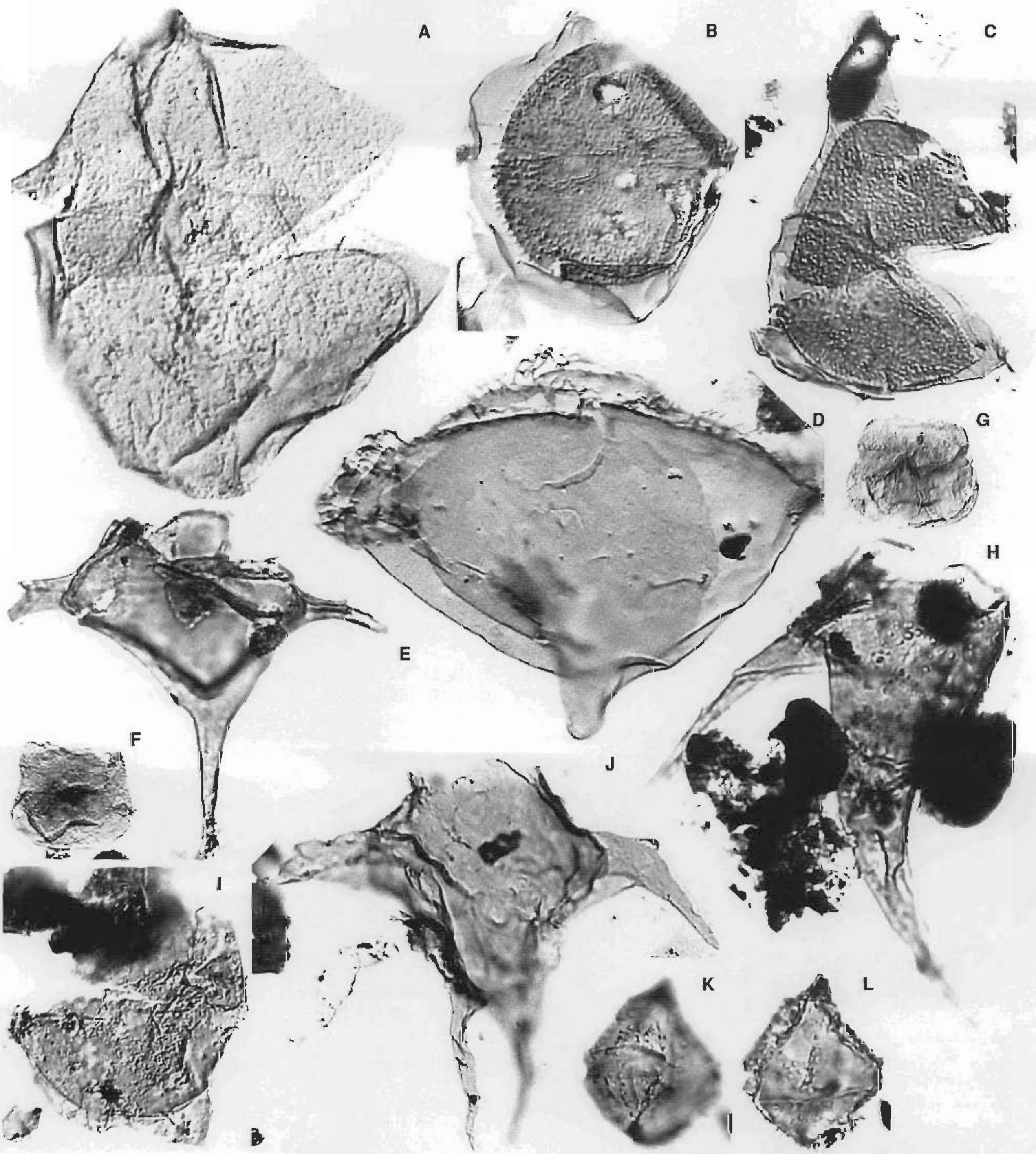

Figure 10. Selected Mesozoic zonal index dinoflagellate cysts (magnification 700x). Stage coordinates given for Zeiss Axioskop microscope no. 77110. A. *Dissiliodinium caddaense* (Filatoff) Stover & Helby; CPC31204 (MFP9256–2, 1073/029). B. *Dingodinium jurassicum* Cookson & Eisenack; CPC31197 (MFP9256–2, 1041/099). C. *Dingodinium cerviculum* Cookson & Eisenack; CPC31196 (MFP9254–1, 1089/148). D. *Energlynia indotata* (Drugg) Fensome; CPC31221 (MFP9237–1, 953/063). E. *Muderongia australis* Helby; CPC31245 (MFP9252–2, 1048/127). F,G. *Dapcodinium* cf. *D. priscum* Evitt; F. CPC31366 (MFP9239–1, 935/219); G. CPC31365 (MFP9239–1, 971/107). H. *Odontochitina operculata* (O. Wetzel) Deflandre & Cookson; CPC31251 (MFP9246–1, 970/165). I. *Ovoidinium cinctum* (Cookson & Eisenack) Davey; CPC31257 (MFP9246–1, 1030/137). J. *Muderongia testudinaria* Burger; CPC31239 (MFP9251–1, 934/076). K,L. *Chichaouadinium boydii* (Morgan) Bujak & Davies; CPC31184 (MFP9246–1, 936/100).

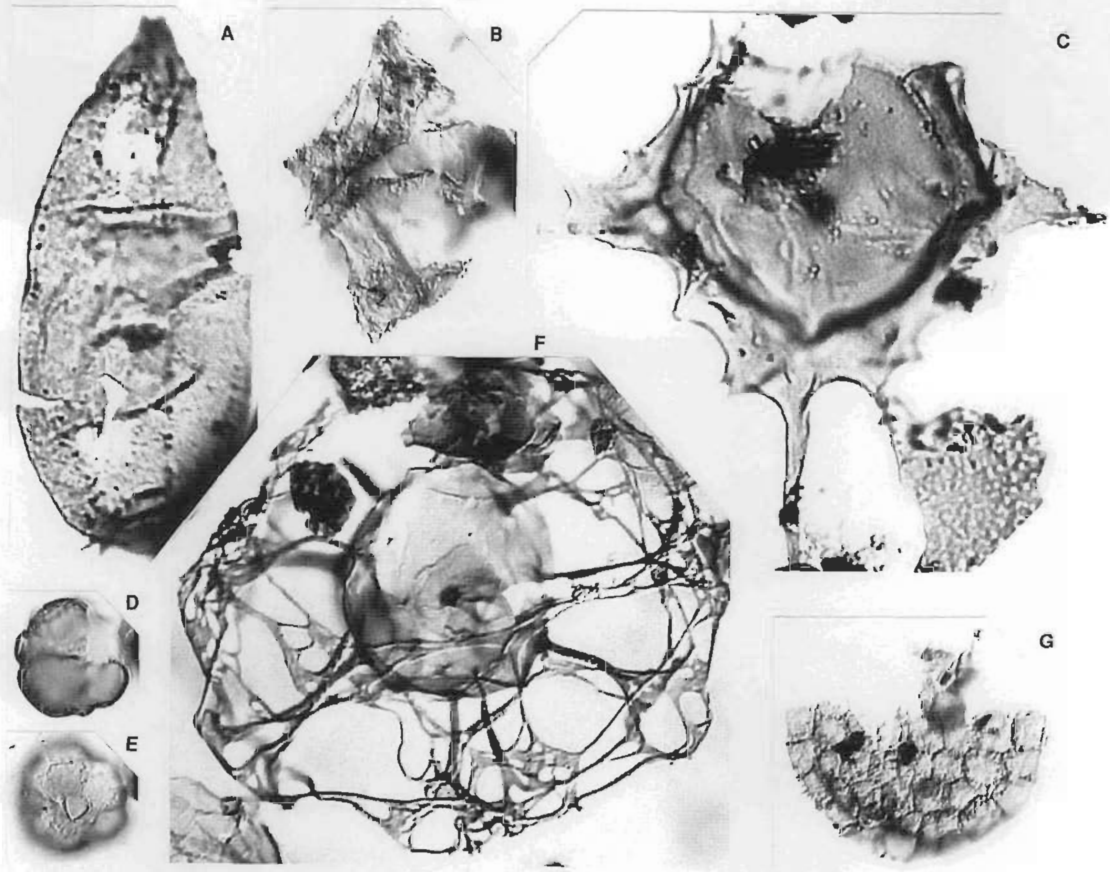

Figure 11. Selected Mesozoic zonal index dinoflagellate cysts (magnification 700x). Stage coordinates given for Zeiss Axioskop microscope no. 77110. A. *Omatia montgomeryi* Cookson & Eisenack; CPC31262 (MFP9252–1, 990/147). B. *Rhaetogonyaulax rhaetica* (Sarjeant) Loeblich & Loeblich; CPC31277 (MFP9243–1, 957/238). C. *Phoberocysta neocomica* (Gocht) Millioud; CPC31256 (MFP9253–2, 953/204). D.E. cf. *Susadinium* sp. A of Helby & others 1987; CPC31295 (MFP9241–2, 1020/084). F. *Rigaudella aemula* (Deflandre) Below; CPC31281 (MFP9256–2, 999/184). G. *Wanneria listeri* (Stover & Helby) Below; CPC31317 (MFP9239–1, 935/109).

The more southerly samples 96DR17/5 and 96DR17/9, 96DR14/10, and 96DR16/2 were dredged from a thick sequence at the east slope of Swan Canyon (lower Dingo Claystone). The samples fall within the same time-span as those from farther north, and from the comparatively richer yield of marine palynomorphs may perhaps represent a brief (restricted) marine incursion.

Late Jurassic

Rowley Terrace

Samples 96DR12/1, 96DR13/4, 96DR14/5, and 96DR15/2 were dredged from the east slope of Swan Canyon, where a post-breakup sequence of Upper Jurassic to Lower Cretaceous shale, sandstone, and mudstone (Dingo Claystone) rests unconformably on the Middle Jurassic strata (lower Dingo Claystone) mentioned above (von Rad et al., 1992; Colwell et al., 1990; Colwell et al., 1994 — this issue). The assemblages indicate (shallow) open-marine environments, and from the peculiar character of its assemblage Sample 15/2 might represent hypersaline (evaporative?) conditions.

Early Cretaceous

Carnarvon Terrace

Seven Neocomian samples (96DR05/5 and –/6, 96DR07/7 and –/12, 96DR08/3, –/4, and –/5), and one Aptian sample (96DR01/3A) from a location somewhat farther south, all represent the Barrow and lower Winning Groups or their western correlative intervals, and indicate open-marine (inland-sea or continental-shelf) environments.

Conclusions

This study agrees with Helby et al. (1987) and other studies that broad, but reasonably accurate, Jurassic and Cretaceous spore–pollen correlations can be made between eastern and Western Australia. More detailed information on the distribution of critical index species is awaited to solve several remaining problems argued in this paper, especially in the Early and Middle Jurassic.

Several papers have shown that comparisons of dinoflagellate records from the eastern and western Tethys Regions may yield good age determinations for the Australian Early Cretaceous (Morgan, 1980; Burger, 1982; and others). The ages here proposed for the dinoflagellate zones are determined by a similar comparison; they represent modifications of those given in Helby et al. (1987). More corroborative evidence is needed from study of new material.

Direct faunal data to support the palynological evidence from the BMR dredge samples are virtually absent. Some age determinations of the Rowley Terrace material (dredges 95/DR04 and 95/DR06) agreed reasonably well with the evidence from foraminifera and nannofossils. Age discrepancies are apparent from several other localities (dredge 95/DR07 and 96DR28 on the Rowley Terrace, dredge 96DR01 on the Carnarvon Terrace, and 96DR/14 and 96DR/15 on the northern Exmouth Plateau). Those discrepancies may probably be explained by the fact that palynological and faunal data are not always derived from the same rock fragment. It is clear that dredge samples may include rock fragments from several stratigraphic levels, due to gravity, submarine erosion, or other causes.

The palynological assemblages reflect long-term changes in marine influence on the western rim of the Australian Plate. Some are clearly related in part to the tectonic history. A possible eustatic origin for several changes is suggested here by observing the following similarities with the global sea-level curve of Haq et al. (1987): (1) the brackish to restricted-marine origin of the Rhaetian samples from the Rowley Terrace coincides with a low Rhaetian–Hettangian eustatic sealevel; (2) the Toarcian to Bathonian samples from the Rowley Terrace reflect less-restricted and in part open-marine conditions, which broadly coincide with a eustatic sealevel rise during the Middle Jurassic; and (3) the open-marine character of the Late Jurassic samples from the Rowley Terrace coincides with a maximum eustatic sealevel in the Kimmeridgian–Tithonian.

Sample density is too low to recognise sealevel movements which have been logged for the Neocomian of the Perth and Carnarvon Basins (Wiseman, 1979; Backhouse, 1988). It is noted that the Carnarvon Terrace assemblages reflect the generally high eustatic sealevel during the Early Cretaceous but — possibly because of sampling gaps — show no sign of a drop in sealevel during the Valanginian, but it has been registered in the Perth Basin and the Great Australian Basin in Queensland (Exon & Burger, 1981; Burger, 1986, 1989).

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Late Triassic conodont and palynomorph biostratigraphy and conodont thermal maturation, North West Shelf, Australia

Robert S. Nicoll[1] & Clinton B. Foster[1]

Late Triassic (Norian–Rhaetian) conodonts recovered from borecores and sea-bottom dredge samples on the North West Shelf off Western Australia are assigned to the *Metapolygnathus primitius, Epigondolella triangularis, E. spiculata, E. postera, E. bidentata, Misikella hernsteini*, and *M. posthernsteini* Zones. Based on previous studies, particularly from North America, these conodont zones can be used to tie with the standard Triassic ammonite zonation. The present record therefore provides the first set of chronologic anchor points for dating the co-occurring dinocyst assemblages and spore-pollen floras from the North West Shelf. Our conodont data show that the *Hebecysta* (al. *Heibergella) balmei, Rhaetogonyaulax* (al. *Shublikodinium) wigginsii*, and *Wanneria* (al. *Suessia) listeri* dinocyst zones are younger than suggested previously, and that some zonal ranges overlap. We conclude that further detailed palynological and conodont studies are urgently needed to resolve these problems and extend conodont age control into the early Late Triassic (Carnian) and the Middle Triassic. Key wells investigated include Ashmore Reef 1, Mt. Ashmore 1B, and Sahul Shoals 1.

Conodont thermal maturation data indicate a very low thermal gradient on the Ashmore Platform near the shelf margin, but a more normal thermal gradient on the Sahul Platform and a high heat flow in the onshore Bonaparte Basin.

Introduction

This study was initiated to provide age control for AGSO *Rig Seismic* sea-bottom dredge samples (Colwell et al., 1994 — this issue) collected along the continental slope margins of the Exmouth Plateau (Wombat Plateau) and the Rowley Terrace (Fig. 1). Following the recovery of several conodont faunas from slope scree samples, the study was extended to selected stratigraphically located core samples in the Ashmore Reef 1, Pokolbin 1, and Sahul Shoals 1 wells, and cuttings samples in the Mt. Ashmore 1B, and Troubadour 1 wells. The spore–pollen floras and dinocyst assemblages from these cores provided the initial biostratigraphic framework.

Late Triassic conodont faunas are reported here from 14 samples obtained from four dredge haul sites and 19 samples from five petroleum exploration wells on the North West Shelf (Fig. 1; Appendices 1, 2). They provide an alternative method of independently assessing sequences previously dated and correlated using dinocysts and spore–pollen. The conodonts also provide a link to the ammonoid zonations in western Cordilleran Canada (Orchard, 1983, 1991a, b) and the Tethyan of central Europe (Krystyn, 1973, 1980, 1988). A total of 1421 conodont elements were obtained from the 33 productive samples examined.

We discuss the conodont zones recognised and the consequent revision of ages of the dinocyst and spore–pollen zones resulting from the co-occurrence of these three groups.

Conodont faunas

Early Triassic conodonts were recognised from the Perth and Carnarvon Basins in the early 1970s by McTavish (1970, 1973, 1975), but the first Australian report of conodonts from the Late Triassic was from the North West Shelf by Jones & Nicoll (1985). In adjacent areas, Late Triassic conodont faunas are known from Timor (Nogami, 1968), Papua New Guinea (Skwarko et al., 1976) and from many localities in Sumatra and peninsular Malaysia (Metcalfe, 1989, 1990, 1992).

Conodonts are not abundant in most of the samples examined in this study (Appendices 1, 2), but they represent a consistent component in samples with other marine fauna, such as ostracods, foraminifers, and fish teeth and scales. Converted to a per kilogram basis, the productive samples ranged from 1 to 443 elements/kg and averaged 38 elements/kg. The two most productive samples, both from Sahul Shoals 1, core 4, (1886.4–1889.33 m) contained 421 and 337 elements, just over half of the total fauna. The smallest productive sample, of 80 g from the 2840–2885 m interval in Troubadour 1, yielded a single element. The generally low abundance presumably reflects relatively rapid sedimentation rates in the upper part of the Triassic sequence in this part of the North West Shelf. Conversely, the two most productive samples, both from Sahul Shoals 1, core 4, are indicative of slow sedimentation, characteristic of a condensed sequence.

Most of the samples have monospecific conodont faunas, in which elements of only one multi-element species have been recognised. In the 33 productive samples examined, only two had more than one species present, and those examples were of specimens transitional between evolutionally successional species. This extremely low faunal diversity is also observed in similar age sediments in western Canada (Orchard, 1983, 1991a, b), but faunas in Tethyan Europe tend to be more diverse (Krystyn, 1980, 1988). More extensive sampling might produce intervals with greater faunal diversity, but palaeoenvironmental factors, such as water temperature or salinity, may be the most important control on diversity.

The conodont biostratigraphy of the Late Triassic has been examined in detail in British Columbia (Orchard, 1983, 1991a, b), Austria (Krystyn, 1980, 1988), Greece (Krystyn, 1983), and Japan (Isozaki & Matsuda; 1982, 1983). Kozur (1980, 1989a) reviewed conodont zonations in Tethyan Europe, and Koike (1981) did the same for Japan. The conodont biostratigraphy developed in those studies, especially the conodont–ammonoid zonation ties demonstrated by Orchard (1983, 1991a, b) and Krystyn (1980, 1988), provide the basic conodont biostratigraphic framework for the, as yet, incompletely sampled well-sections examined in this report.

Orchard (1983) recognised eight conodont zones in the Norian to Rhaetian part of the Upper Triassic in western

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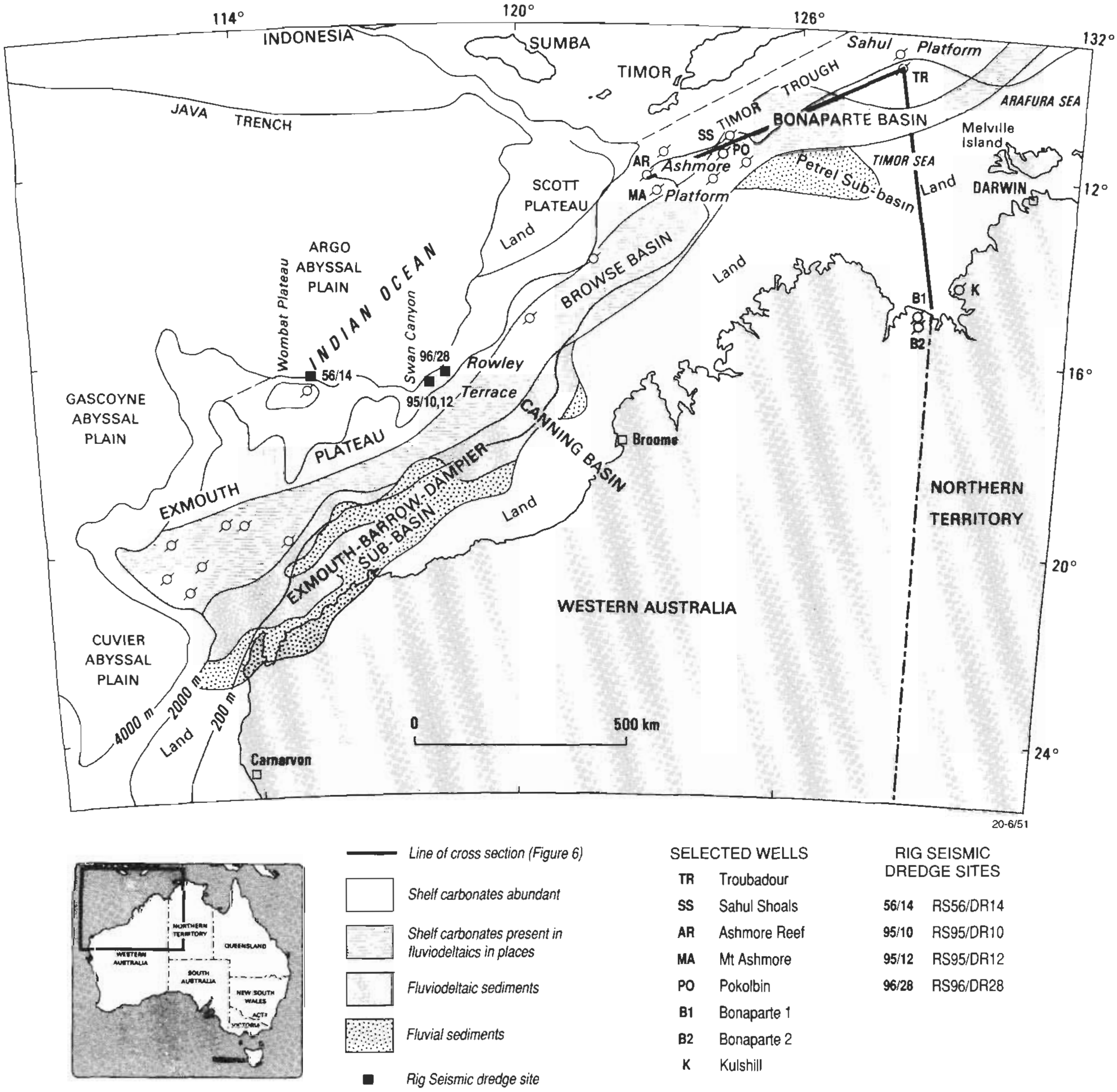

Figure 1. Locality map showing petroleum exploration wells and dredge sites from which conodont faunas have been obtained. AR–B2 indicates the line of section in Figure 6.

Canada. Later Orchard (1991a, b) recognised an additional two zones. Six of these zones are now recognised on the North West Shelf and the *Misikella* Zone of Orchard has been subdivided in this study. Orchard (1991) recognised a post-*Epigondolella* interval to which he applied the name *Misikella* (Fig. 3). This interval corresponds to the *Gondolella (Neogondolella) steinbergensis* to *Misikella posthernsteini* Zones of Krystyn (1980). Orchard (1991) did not subdivide his *Misikella* interval, but Krystyn (1988) recognised four subdivisions: a lower *Gondolella steinbergensis* Assemblage Zone, a *Oncodella paucidentata* Zone, the *Misikella rhaetica* Zone, and an upper *M. posthernsteini* Zone. In a similar zonation, Budurov & Sudar (1990) recognised a lower *M. hernsteini* Zone and an upper *M. posthernsteini* Zone. Kozur (1989) recognised three zones above *M. posthernsteini*, but none of the conodonts which typify these zones have yet been recognised in Australia or Canada.

Neogondolella steinbergensis is common in western Canada (Orchard, 1991b), but was not recovered in samples from this study. However, the almost total mutual exclusivity of *Misikella hernsteini* and *M. posthernsteini* in our samples indicates that in Australia the *Misikella* interval of Orchard (1991a, b) can be divided into a lower *M. hernsteini* Zone and an upper *M. posthernsteini* Zone (Fig. 3).

Conodont zonation — this study

Because the conodont faunas of the North West Shelf appear to be compositionally more similar to those of western Canada documented by Orchard (1983, 1991a, b) than those of Europe documented by Krystyn (1980, 1988), the basic zonal scheme from western Canada is followed here (Fig. 2). The Norian to Rhaetian zones are discussed in ascending order. Those zones appearing in

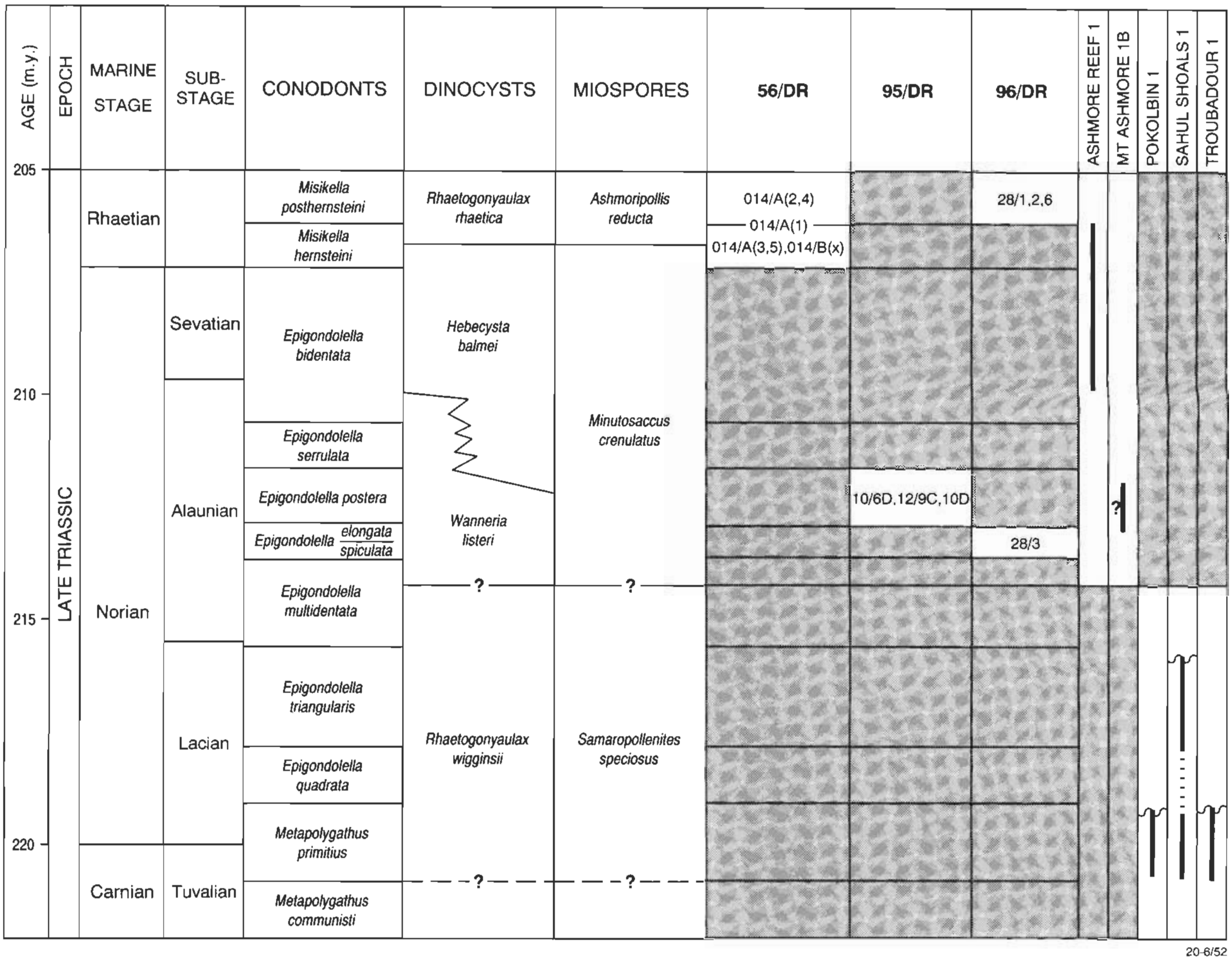

Figure 2. Revised Upper Triassic chronostratigraphy and biostratigraphy of the North West Shelf showing the relationship of the well and dredge samples to the chronostratigraphy.

bold face type have been found in our study.

***Metapolygnathus primitius* Zone** (= *Epigondolella primita* Zone of Orchard, 1983; see Orchard, 1991a, b): straddles the Carnian–Norian boundary and is identified in Sahul Shoals 1 (core 5, 2245–2254 m), Pokolbin 1 (1829.5–1835.5 m), and probably in Troubadour 1 (2840–2885 m).

STAGE	AMMONDID ZONES KRYSTYN 1988	CONODONT ZONES ORCHARD 1991 a	CONODONT ZONES KRYSTYN 1988	CONODONT ZONES THIS PAPER
Rhaetian + Sevatian	*Choristoceras marshi*	*Misikella interval*	*Misikella posthernsteini*	*Misikella posthernsteini*
			Misikella rhaetica	*Misikella hernsteini*
			Oncodella paucidentata	
	Vandaites stuerzenbaumi		*Gondolella steinbergensis*	
	Rhabdoceras suessi	*Epigondolella bidentata*	*Epigondolella bidentata*	*Epigondolella bidentata*

20-6/53

Figure 3. Comparison of the Rhaetian (Rhaetian + Sevatian) ammonoid and conodont zones of Krystyn (1988) with the conodont zones of Orchard (1991a) and the conodont zonation scheme used here.

The *Epigondolella quadrata* Zone(= *E. abneptus* subsp. A of Orchard, 1983; see Orchard, 1991b) is not recorded in our samples (see discussion below).

***Epigondolella triangularis* Zone** (Orchard, 1991a, b) is found only in Sahul Shoals 1, core 4 (1886.4–1889.3 m). This interval was identified initially as the *Epigondolella primitia* Zone by Jones & Nicoll (1985), and is revised here. The earlier determination was based on a fauna of only six elements, and only two of those elements preserved the posterior platform margin, the most distinctive part of *E. triangularis*. The large fauna recovered in the samples processed for this study demonstrated the full range of *E. triangularis* morphologies (see Orchard, 1983; Fig. 6), which include elements similar to those illustrated by Jones & Nicoll (1985).

Elements of the *Epigondolella multidentata* Zone (Orchard, 1983) were not recovered in this study (see discussion below).

***Epigondolella spiculata* Zone** (= *E.* n. sp. C of Orchard, 1983; see Orchard, 1991b): found only in *Rig Seismic* survey 96, dredge 28, sample 3. The sample is from the outer slope of the Rowley Terrace and represents the oldest conodont fauna obtained in any of the dredge samples.

The *Epigondolella elongata* Zone (see Orchard, 1991b) has not been identified on the North West Shelf (see discussion below).

***Epigondolella postera* Zone** (Orchard, 1983): recovered in *Rig Seismic* survey 95, dredge sites 10 (sample 6D) and 12 (samples 9C, 10D). These samples were obtained from the Swan Canyon area on the southern end of the Rowley Terrace (Fig. 1).

The *E. serrulata* Zone (= *E.* n. sp. D of Orchard, 1983; see Orchard, 1991b) has not been identified on the North West Shelf (see discussion below).

***Epigondolella bidentata* Zone** (Orchard, 1983): identified in Ashmore Reef 1 (core 26, 3676–3680 m) and, tentatively, in the cuttings samples from the Mt. Ashmore 1B (2415–2455 m).

***Misikella hernsteini* Zone** (Kozur, 1980): identified in Ashmore Reef 1 (cores 21 and 24, 2969–3364 m) and in *Rig Seismic* dredge samples from the northern edge of the Wombat Plateau (*Rig Seismic* survey 56, dredge site 014, samples A-3, A-5, B-x). The *M. hernsteini* Zone is defined by the presence of the eponymous species and the absence of elements of the genus *Epigondolella. M. posthernsteini* may be present in the zone.

***Misikella posthernsteini* Zone** (Kozur, 1980; Budurov & Sudar, 1990): the youngest conodont zone recognised in Triassic sequences of Australia and nearby areas. *M. posthernsteini* is found in western Canada (Orchard, 1991b), Europe (Krystyn, 1988), Japan (Isozaki & Matsuda, 1983), and has been reported from Papua New Guinea (Skwarko et al., 1976). Orchard (1991a, b) reported the co-occurrence of *M. posthernsteini* and the ammonoid *Choristoceras rhaeticum* in the Kunga Group and ascribes a latest Triassic, *Choristoceras crickmayi* Zone, age to the association. In our study, the zone occurs in Rig Seismic survey 96, dredge site 28 (samples 1, 2, 6) and R.S. survey 56, site 014 (samples A-2, A-4) .

Discussion. As noted above, four of the conodont zones recognised by Orchard (1983, 1991a, b) have thus far not been identified on the North West Shelf. Orchard (1991c) suggested that some of the zonal species, especially *Epigondolella multidentata* and *E. serrulata*, are only found in a restricted area of western Canada. If correct, these species would not be expected in the North West Shelf region.

However, it is also possible that their absence may be attributed to gaps in the present sampling. In the wells examined, the *Epigondolella serrulata* Zone is expected to occur below 3680 m (core 26) in Ashmore Reef 1, but conodonts were not recovered from core 27 (3787—3790 m), and no carbonate samples are available from a deeper part of the well, which reached a total depth of 3914 m. This same observation also applies to the *E. elongata* Zone.

If present, faunas of the *Epigondolella multidentata* Zone should occur above the *E. triangularis* Zone (core 4, 1886 m) in Sahul Shoals 1. However in this well, as in Pokolbin 1 and Troubadour 1 wells, younger rocks of either Jurassic or Cretaceous age immediately and unconformably overlie the fauna-bearing Triassic.

Following Orchard's zonation, the *Epigondolella gradata* Zone is predicted to occur in the Sahul Shoals 1 section between cores 4 and 5 (1890–2245 m), but no core material is available over that interval, and cuttings samples have yet to be studied.

In summary, there are insufficient data to assess if the remaining four conodont zones occur on the North West Shelf.

Spore–pollen floras and zonation

The first, and oldest Triassic spore–pollen assemblages to be dated independently by an associated marine fauna were described from the type section of the Kockatea Shale, Perth Basin, Western Australia by Balme (1963). The fauna, which includes ammonites and pelecypods, is of Early Triassic (Griesbachian–Smithian) age (Dickins & McTavish, 1963; McTavish & Dickins, 1974). Palynological assemblages from the Kockatea Shale were assigned by Dolby & Balme (1976) to their *Kraeuselisporites saeptatus* Zone. This was the oldest of five palynological assemblage zones defined by Dolby & Balme from the Triassic of the Carnarvon Basin, Western Australia. The succeeding *Triplexisporites* (al. *Tigrisporites*) *playfordii* Zone is also independently dated by associated conodonts of Smithian–Anisian (Early Triassic) age (see McTavish, 1973).

As noted by Dolby & Balme (1976), no independent faunal evidence to date the younger zones was available, but they suggested the following age relationships: *Staurosaccites quadrifidus* Zone, Anisian–Carnian; *Samaropollenites speciosus* Zone, Carnian; *Minutosaccus crenulatus* Zone, Carnian to ?Norian.

The Anisian age for the base of the *S. quadrifidus* Zone was deduced from an estimate of the minimum age of the underlying zone, and an assumption "of continuity of sedimentation in the Triassic sequence". Some supporting evidence for these age assignments was given using broadly defined spore–pollen ranges from the European Triassic, because these zones contain certain key species first described from the European sequences. Similar palynological criteria were also used to date the Triassic assemblages of eastern Australia (see De Jersey, 1975, for summary), although there are far fewer shared European species than in the Western Australian assemblages. However, age determinations based on ranges of European spore–pollen taxa, particularly as applied in eastern Australia, are "... not completely irresistible" (Dolby & Balme, 1976, p. 133). In summary, therefore, no rigorous control for the given ages of these assemblage-zones existed for the Middle and Late Triassic.

Moreover, later work (e.g. Brenner, 1992; Brenner et al., 1992; Bint & Helby, 1988) has suggested that distinctions between palynofloras of the *S. speciosus* Zone and the immediately overlying *M. crenulatus* Zone may result from facies controls on the parent floras, and therefore are not time significant. Evidence from this study also shows that differentiation between these zones is difficult. More taxonomic and quantitative studies of these assemblages are required urgently to resolve this problem.

Dinocyst floras and zonation

In their review of Australian Mesozoic palynological zones, Helby et al. (1987) defined six dinocyst zones in the Triassic of Western Australia, and correlated them with the spore–pollen zones of Dolby & Balme (1976).

Helby et al. (1987) also modified the zonal criteria of Dolby & Balme (1976), but complete descriptions of the assemblages have yet to be published.

Helby et al. (1987) revised the age assignments of the spore–pollen zones using evidence from dinocyst studies from Europe and North America (see Helby et al., 1987, p. 7), and the record of the first Late Triassic conodonts from associated sections in Western Australia (Jones & Nicoll, 1985). As noted above, only a single conodont sample was studied and the age of the fauna was considered to span the Carnian–Norian boundary. With this information, Helby et al. suggested that their newly described *Rhaetogonyaulax wigginsii* (dinocyst) interval zone, and its correlative spore–pollen zone (upper *S. speciosus*), were both of late Carnian age. Similarly, from comparative dinocysts from elsewhere in the Triassic, dated by independent faunal controls, they suggested the following ages: *Rhaetogonyaulax rhaetic* — Rhaetian; *Hebecysta (*al. *Heibergiella) balmei* — Norian; *Wanneria (*al. *Suessia) listeri* — Norian; *Sahulidinium ottii* — latest Anisian to Ladinian (Fig. 4).

Revised Upper Triassic biostratigraphic/chronostratigraphic chart

The result of this study is a significant revision of the biostratigraphic–chronologic relationships for the Upper Triassic of western Australia. The correlative relationship between the dinocyst and spore–pollen zonations remains essentially unchanged, but the ages of those zones are altered significantly (Fig. 4).

Analysis of conodont faunas and dinocyst and spore–pollen floras from samples in the Ashmore Reef 1, Mt. Ashmore 1B, Pokolbin 1, Sahul Shoals 1 and Troubadour 1 wells (Fig. 5, Appendix 1) form the basis for these revisions. The upper part of the section is preserved in Ashmore Reef 1 and Mt. Ashmore 1B. In the Pokolbin 1, Sahul Shoals 1 and Troubadour 1 wells, post-Triassic erosion has removed any Late Triassic (Rhaetian and upper Norian) sediments that may have been deposited.

The *R. rhaetica* dinocyst Zone and the *A. reducta* spore–pollen Zone are equated with the *Misikella posthernsteini* and the *M. hernsteini* conodont Zones. In Ashmore Reef 1, the *R. rhaetica* Zone is found between 2970 and 3362 m (Fig. 4). The *A. reducta* Zone is found between 2834 and 3231 m and the *M. hernsteini* Zone between 2969 and 3364 m. *M. posthernsteini* evolves from *M. hernsteini* and thus should be expected to co-occur with, and range higher than, *M. hernsteini*, as was noted in Europe (Krystyn, 1980). Where *M. posthernsteini* occurs without other conodont elements, as in some of the Wombat Plateau dredge samples (Fig. 2, Appendix 2), the samples are interpreted as equivalent to the "topmost Triassic *marshi* Subzone" (Krystyn, 1980, p. 80).

The *H. balmei* dinocyst Zone is correlated directly with the *M. crenulatus* spore–pollen Zone and the *Misikella hernsteini, Epigondolella bidentata, E. postera* and *E. spiculata* conodont Zones. In Ashmore Reef 1, the *H. balmei* Zone is reported at 3363 m in the bottom of core 24; the reference section for the zone ranges between 3523 and 3573 m (Helby et al., 1987) where it occurs with elements of the *M. crenulatus* Zone. In Mt. Ashmore 1B, *H. balmei* occurs between 2370–2458 m where it overlaps with *M. crenulatus* and *Epigondolella* aff. *E. postera* (cuttings: 2415–2455 m).

Correlation of the upper and lower boundaries of the dinocyst *W. listeri* Zone with the conodont zonation is hampered by a single data point for the upper boundary only. The extent of the zone is therefore constrained by the ranges of the adjacent dinocyst zones. In Ashmore Reef 1, the reference section for the *W. listeri* Zone occurs between 3612 and 3910 m where it is associated with *Epigondolella bidentata* (core 26, 3676–3679 m) and *M. crenulatus*. This overlaps with the range of the younger *H. balmei*, but is taken as the upper boundary for *W. listeri*, using the conodont zonation. Correlation between the lower boundary of *W. listeri* and the upper *E. multidentata* zone is based on stratigraphic position, namely younger than *E. triangularis* Zone, which is correlated with the immediately underlying *R. wigginsii* dinocyst zone (Fig. 4).

The *R. wigginsii* dinocyst Zone is correlated with the *S. speciosus* spore–pollen Zone (Helby et al., 1987) and the *Epigondolella triangularis* and *Metapolygnathus primitius* conodont Zones. The type *R. wigginsii* Zone is defined between 1883 and 2102 m in Sahul Shoals 1 (Helby et al., 1987). This range includes the occurrence of *Epigondolella triangularis* between 1886–1889 m in core 4 (Appendix 1) and with *S. speciosus* between 1795–2727 m. *Metapolygnathus primitius* occurs in Sahul Shoals 1 (core 5, 2245–2254 m), within the range of *S. speciosus*. In the Pokolbin 1 well, *M. primitius* occurs in core, between 1829–1835 m and within the range of *S. speciosus* between 1830–2220 m. In Troubadour 1, *M. primitius* occurs in cuttings between 2840–2885 m and *S. speciosus* is reported between 2783–2803 m.

The establishment of ties between conodont faunas and both dinocyst and spore–pollen assemblages now permits definition of the palynomorph zones on the North West Shelf in terms of the ammonoid-based stage terminology. Figures 2 and 3 show the effect of this new correlation in the younger age assignment of the palynomorph zones as compared with the correspondence suggested by Dolby & Balme (1976) and Helby et al. (1987). The co-occurrence of *M. posthernsteini, R. rhaetica*, and *A. reducta* Zones indicate a Rhaetian (post-Sevatian) *crickmayi/marshi* Zone age. It should be noted that both the palynological zones extend into the Early Jurassic (Helby et al., 1987), but conodonts are not found above the Triassic–Jurassic boundary. The *Epigondolella spiculata* through *E. bidentata* Zones, and the *H. balmei, W. listeri* and *M. crenulatus* Zone span the Middle Norian to early Rhaetian. The *Metapolygnathus primitius* to *E. triangularis, R. wigginsii* and *S. speciosus* Zones are of latest Carnian to Early Norian.

Conodont CAI thermal maturation data

Analysis of all of the conodont elements recovered indicate a low heat flow on the passive northwestern continental margin, but a more normal heat flow is indicated in the Sahul Platform area adjacent to Troubadour 1 and a high heat flow in the deep wells of the onshore Bonaparte Basin (Figs 1 and 6). All of the dredge samples from the Wombat Plateau and Rowley Terrace areas had conodont maturation values of CAI 1 (Epstein et al., 1977) indicating exposure to temperatures of no more than about 80°C. This would be in the range of the expected value with shallow burial of the Triassic section on the sediment starved continental margin. However, the CAI 1 values in Ashmore Reef 1, at depths of 3679 m (core 26), are much lower than the heat flow gradient for the Appalachian

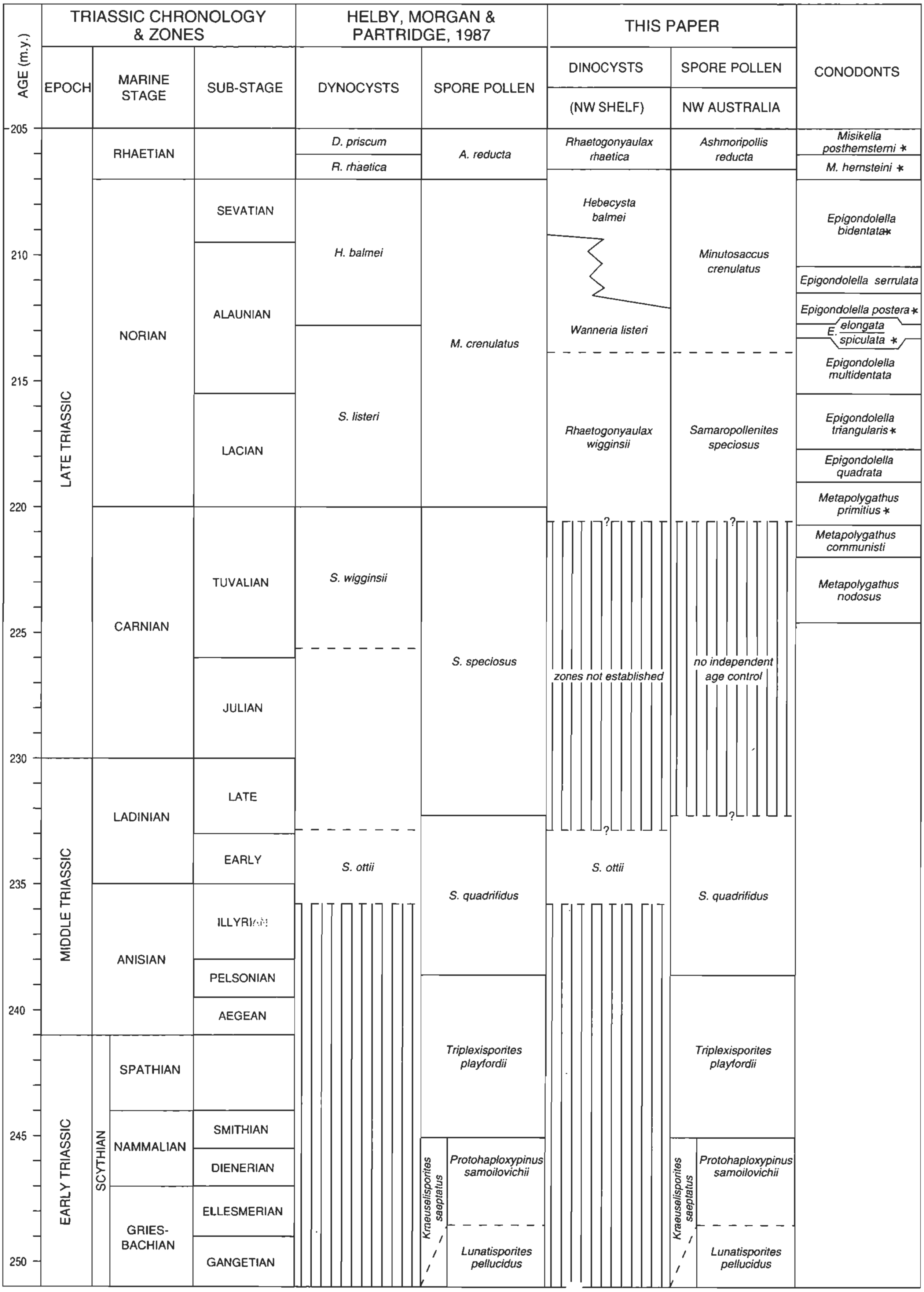

Figure 4. Triassic palynomorph and Late Triassic conodont zonations used here compared to the palynomorph zonal scheme by Helby et al. (1987). Differences in palynomorph zone names reflect recent nomenclatural changes and not changes in zone definition. Conodont zones for the pre-middle Carnian are not shown.

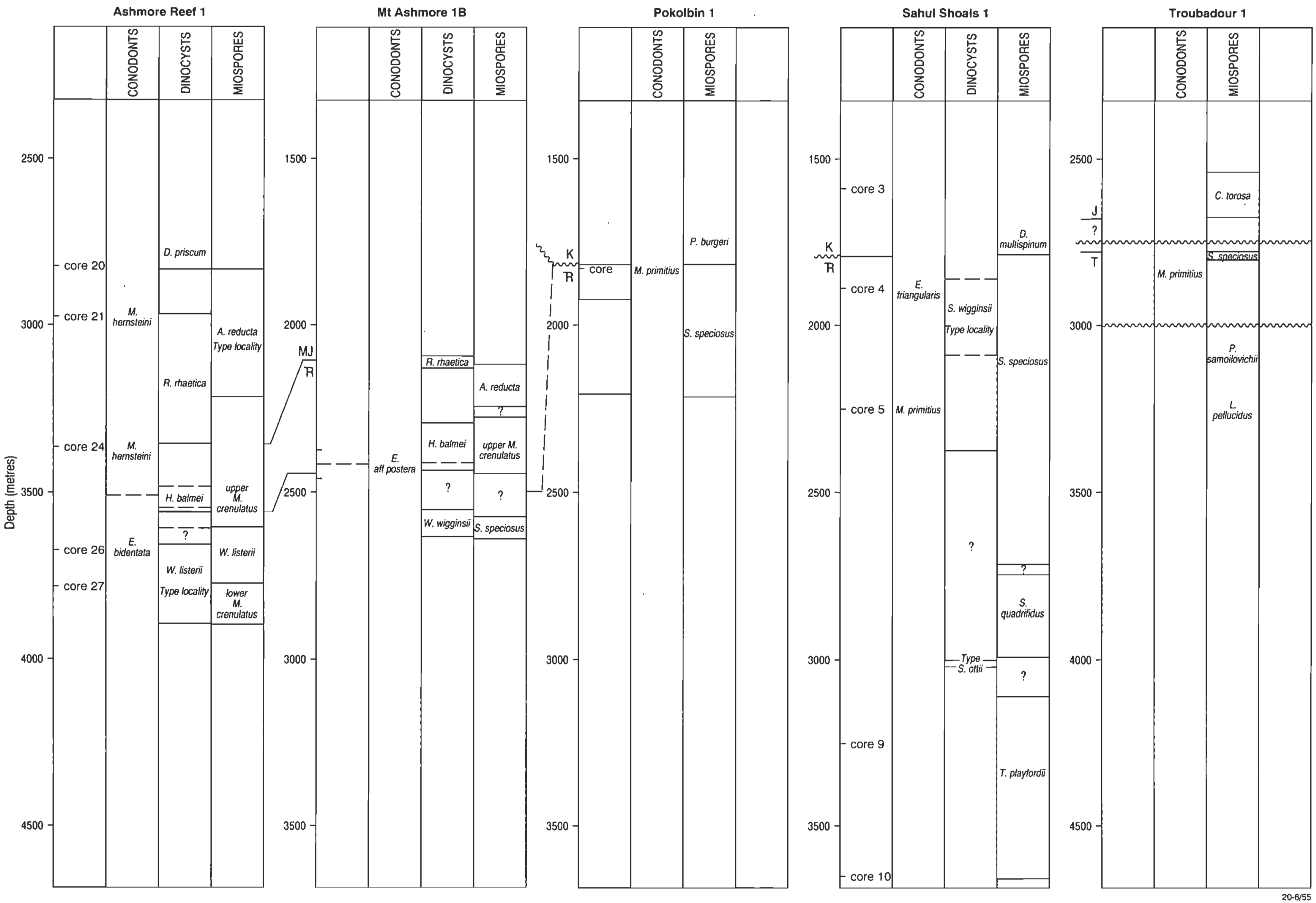

Figure 5. Cross-section showing Triassic conodont, dinocyst and spore–pollen zonations in the Ashmore Reef 1, Mt. Ashmore 1B, Polkolbin 1, Sahul Shoals 1, and Troubadour 1 wells.

area of eastern North America (Epstein et al., 1977) or for gradients determined in the cooler part (Kidson Sub-basin) of the Canning Basin (Nicoll & Gorter, 1984a, b).

The CAI 2 value in the interval 2840–2885 m in Troubadour 1 corresponds to a heat-flow gradient as found in the Appalachians (Epstein et al., 1977) or the Kidson Sub-basin of the Canning Basin.

Lower Carboniferous conodonts from cores in the Bonaparte 1 and 2 of the onshore Bonaparte Basin (Druce, 1969), have CAI values ranging from 1 to 2 (Fig. 6) and Upper Devonian samples from the base of Kulshill 1 (3970 m) have a CAI value of 5. CAI values in these three wells correspond closely to the high thermal gradient found in the Broome Arch area of the Canning Basin (Nicoll & Gorter, 1984a, b).

The conodont CAI values show a very similar maturation story to that demonstrated by Horstman (1988) using vitrinite reflectance in wells on the North West Shelf. It is possible that the regional variation in heat flow is the driving mechanism for the overpressurization observed by Horstman.

Conclusions

Seven of the eleven Late Triassic (Norian + Rhaetian) conodont zones established in the western Canadian succession by Orchard (1983, 1991a, b) are now recognised on the North West Shelf. These zones include our twofold division of Orchard's *Misikella* Zone. Given the sparseness of the present sampling, it is possible that the other zones would be recognised if a complete section were available for study. Orchard (1991c), however, suggested that the missing taxa may be geographically restricted to western Canada.

For the first time, independent age control is available to date the dinocyst and spore–pollen zones on the North West Shelf (Fig. 3). The *R. rhaetica, A. reducta* and *Misikella posthernsteini* Zones are Rhaetian (post-Sevatian). The *H. balmei, W. listeri, M. crenulatus,* and *Epigondolella spiculata* through *E. bidentata* Zones are Middle Norian (Alaunian to Sevatian). The *S. wigginsii* and *S. speciosus* Zones are early Norian in age, and may extend down into the Carnian. The *M. primitius* through *E. triangularis* Zones are latest Carnian to Early Norian.

The co-occurrence of conodonts faunas with dinocyst and spore–pollen assemblages allows a much improved correlation of the Upper Triassic of the North West Shelf with the ammonoid zonations of western Canada and Tethyan Europe.

Associated conodont faunas from Ashmore Reef 1 and Mt. Ashmore 1B show that part of the ranges of the *H. balmei* and *S. listeri* dinocyst zones overlap. At present there are limited published details on both the quantitative composition of these zones and their lithofacies associations: further studies are needed to assess the significance of areas of overlap.

Similarly, the lack of published detail on the precise composition of the spore–pollen zones *M. crenulatus* and *S. speciosus* is hampering the assessment of a more refined biostratigraphic utility of these zones. These zones might also reflect ecological associations and be, at least in part, time-equivalents.

Conodont taxonomic observations

This study is primarily biostratigraphic, but a few taxonomic observations are necessary for some of the species. Recognition of the full multi-element apparatus structure of the species considered here has not been attempted. For most of the samples (Table 1, Appendices 1, 2), too few ramiform elements were recovered to reveal the full range of element types. As noted by Kozur (1974, 1989b), the apparatus structure of all gondolellid conodonts is very similar. Thus, there would appear to be very little morphologic difference in the apparatus structure of *Epigondolella, Metapolygnathus* and *Misikella*, an observation borne out by the ramiform elements recovered in this study.

Table 1. Conodont element abundance. Summary abundance distribution of element types in conodont apparatuses. The S elements have not been separated into Sa, Sb, Sc and Sd types.

	Pa	*Pb*	*S+M*
Epigondolella bidentata Mosher, 1968	15	0	13
Epigondolella postera (Kozur & Mostler, 1971)	40	0	25
Epigondolella triangularis (Budurov, 1972)	690	6	68
Epigondolella spiculata Orchard, 1991b	6	0	1
Epigondolella sp. indet.	5	0	3
Metapolygnathus primitius (Mosher, 1970)	169	0	25
Misikella hernsteini Kozur & Mock, 1974	163	2	33
Misikella posthernsteini Kozur & Mock, 1974	97	0	60

Genus *Epigondolella* Mosher, 1968

Type species: *Polygnathus abneptis* Huckriede, 1958

Remarks. The recent suggested revisions of the genus *Epigondolella* by Kozur (1989b) have not been followed here because not enough material is available to make a reasoned decision as to the value of genera, such as *Mockina* Kozur. The taxonomic interpretations of Orchard (1983, 1991) are followed pending the availability of more material.

Epigondolella bidentata Mosher, 1968

Figs 7.1–7.2

Material studied. 28 elements (15–Pa, 13–S+M).

Remarks. The observation that many *Epigondolella* species may pass through a *bidentata* morphology stage (Orchard, 1983) was considered in this identification. However, most of the specimens in this study are large enough to be regarded as mature elements in which the *E. bidentata* bi-dentate morphology is diagnostic.

Epigondolella postera (Kozur & Mostler, 1971)

Figs 7.3–7.5

Material studied. 65 elements (40–Pa, 25 S+M).

Remarks. The specimens are similar to those illustrated by Krystyn (1980) and Orchard (1983, 1991a, b).

Epigondolella spiculata Orchard, 1991

Fig. 8.4

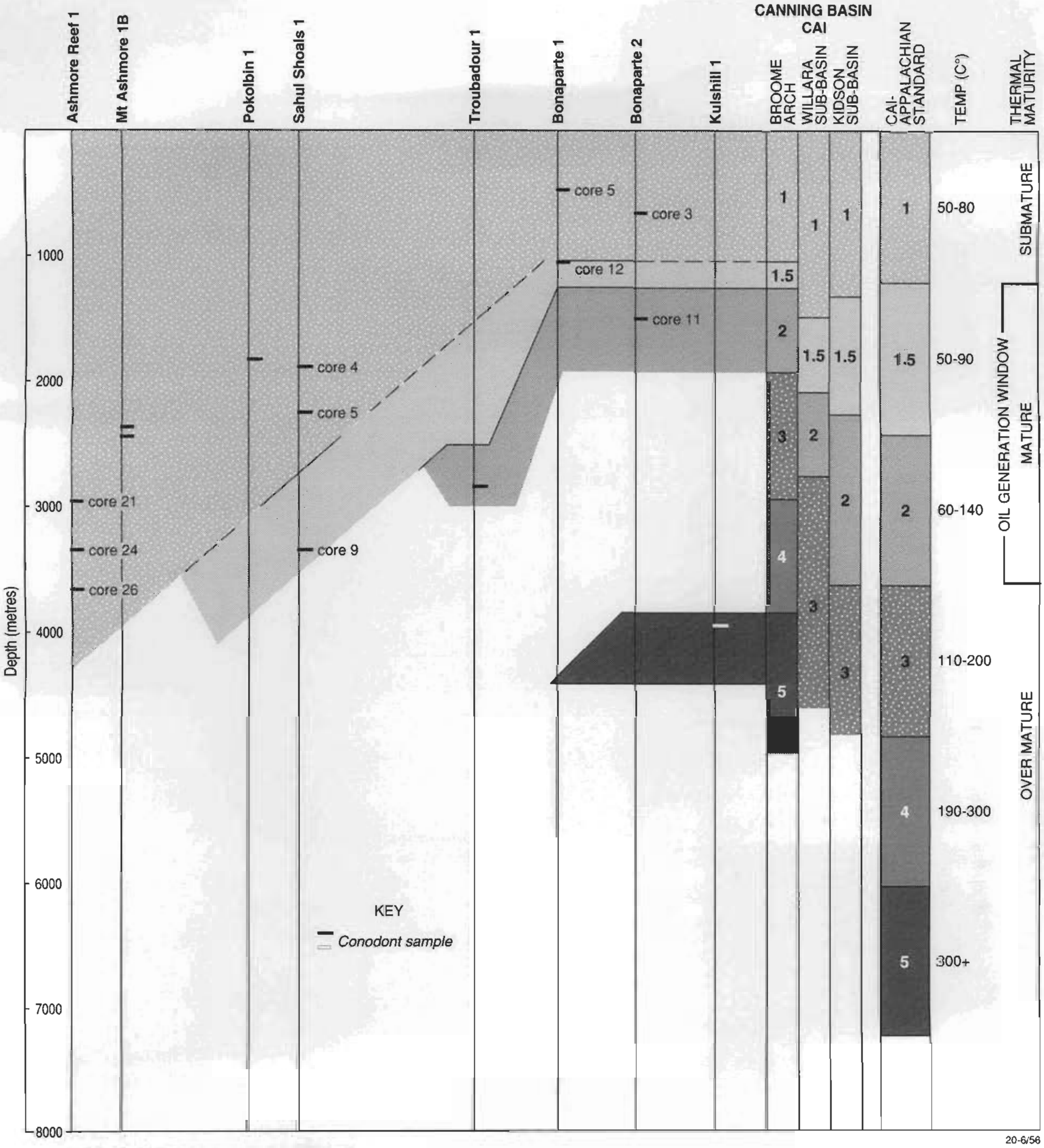

Figure 6. Plot of conodont colour alteration values (CAI) in wells examined in this study. Cross-section shows the apparent change in heat flow values from the cool outer shelf margin to the high heat flow regime of the onshore Bonaparte Basin. Conodonts in the Bonaparte wells are of Early Carboniferous age (Druce, 1969) and those from Kulshill 1 are of Late Devonian age. Comparison with the CAI intervals of the onshore Canning Basin (Nicoll & Gorter, 1984a,b) and the Appalachians (Epstein et al., 1977) is indicated.

Material studied. 7 elements (6–Pa, 1–S).

Remarks. The specimens are similar to those illustrated by Orchard (1983, 1991a, b).

Epigondolella triangularis (Budurov, 1972)

Figs 8.1–8.3

Material studied. 764 elements (690–Pa, 6–Pb, 68–S+M).

Remarks. Large specimens of *E. triangularis* have well-developed nodes on the posterior platform margin and compare closely with material illustrated by Budurov (1972, 1977), Budurov & Sudar (1990), and Orchard (1991).

Genus *Metapolygnathus* Hayashi, 1968

Type species. *Metapolygnathus communisti* Hayashi, 1968

Metapolygnathus primitius (Mosher, 1970)

Figs 9.1–9.3

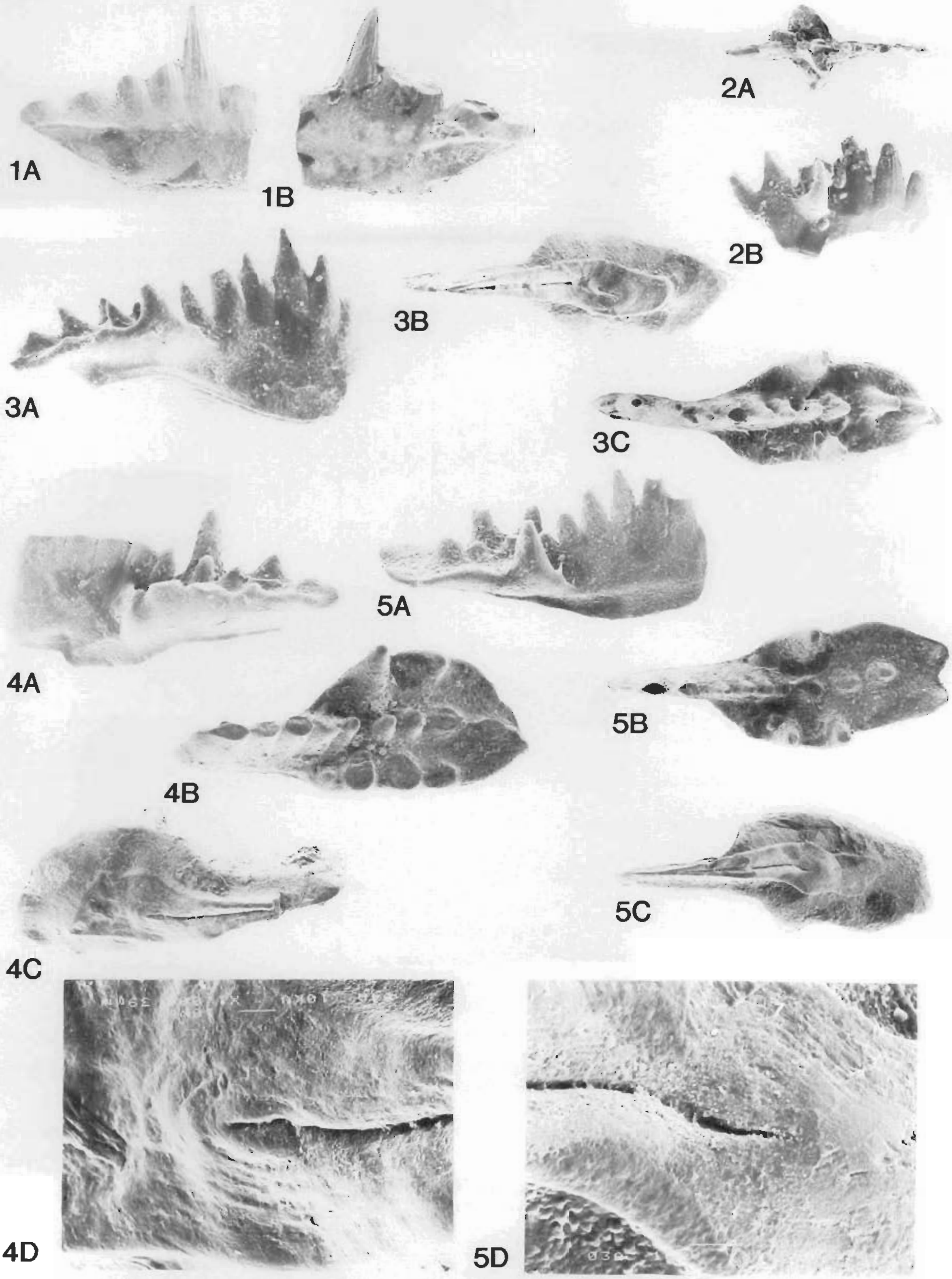

Figure 7. *Epigondolella bidentata* and *Epigondolella postera*, all Pa elements. Magnification as noted.
1. *Epigondolella bidentata* (CPC 31648)[Ashmore Reef 1, 12065-12070] ?left element (all X175); a, inner lateral view; b, outer lateral view.
2. *Epigondolella bidentata* (CPC 31649)[Ashmore Reef 1, 12065-12070] right element (all X175); a, oral view; b, outer lateral view.
3. *Epigondolella postera* (CPC 31650)[RS95-DR/12-10D] left element (all X140); a, inner lateral view; b, aboral view; c, oral view.
4. *Epigondolella postera* (CPC 31651)[RS95-DR/12-10D] right element (all X120, except 4d); a, inner lateral view; b, oral view; c, aboral view; d, enlargement (X600) of oral view showing basal cavity.
5. *Epigondolella postera* (CPC 31652)[RS95-DR/12-10D] left element (all X120, except 4d); a, inner lateral view; b, oral view; c, aboral view; d, enlargement (X850) of oral view showing basal cavity.

Material studied. 194 elements (169–Pa, 25–S+M).

Remarks. The Pa elements show a similar range of morphologic variation to that demonstrated by Orchard (1983). The Pa elements have a well-developed platform, but relatively poorly developed platform marginal nodes. The specimens illustrated by Jones & Nicoll (1985) as *Epigondolella primitia* are now assigned to *E. triangularis,* because they have more lateral node development than is typical of *Metapolygnathus*.

Genus *Misikella* Kozur & Mock, 1974a

Type species. *Misikella longidentata* Kozur & Mock, 1974a.

Remarks. Fåhræus & Ryley (1989) revised the definition of *Misikella,* recognising a tetramembrate apparatus of pectiniform and ramiform elements in the type species. They removed all of the other species previously assigned to the genus on the basis that they had recognised a bimembrate pectiniform–coniform apparatus structure. The new genus, *Axiothea*, included the species: *A. hernsteini* (Mostler, 1968), *A. koessenensis* (Mostler, 1978), *A. posthernsteini* (Kozur & Mock, 1974), and *A.* sp. cf. *A. rhaetica* (Mostler, 1978).

In this study, no coniform elements were recovered from any of the samples containing pectiniform elements of *Misikella (Axiothea*). However, in most of the samples there were ramiform elements similar to those illustrated as elements of the *Misikella longidentata* by Fåhræus & Ryley (1989). As these samples have no other pectiniform elements, it is concluded that the apparatus structure of *M. hernsteini* and *M. posthernsteini* includes ramiform elements and no coniform elements. Thus, the establishment of the genus *Axiothea* was not necessary for the pectiniform elements, but the genus is presumably still valid as a coniform element apparatus that has only been partly defined and apparently lacks pectiniform elements. Kozur (1989) mentioned a Rhaetian coniform genus, *Ziegliericonus* Kozur & Mock, which may be conspecific with the coniform elements attributed to *Axiothea* by Fåhræus & Ryley (1989).

Misikella hernsteini Kozur & Mock, 1974a

Figs 10.4–10.7

Material studied. 198 elements (163–Pa, 2–Pb, 33–S+M).

Remarks: Some of the Pa elements included here within *M. hernsteini* have a posterior margin that is rounded rather than pointed. These specimens are considered early stage transitions to *M. posthernsteini*.

Misikella posthernsteini Kozur & Mock, 1974b

Figs 10.1–10.3

Material studied. 157 elements (97–Pa, 60–S+M).

Remarks. A strict morphologic interpretation of the Pa element was used to distinguish *M. posthernsteini* from *M. hernsteini*. Only if the outline of the posterior margin of the Pa element was concave was the element considered to belong to *M. posthernsteini*. This is generally the same definition used by Isozaki & Matsuda (1983). Elements in which the posterior margin was flat were still considered to belong to *M. hernsteini*.

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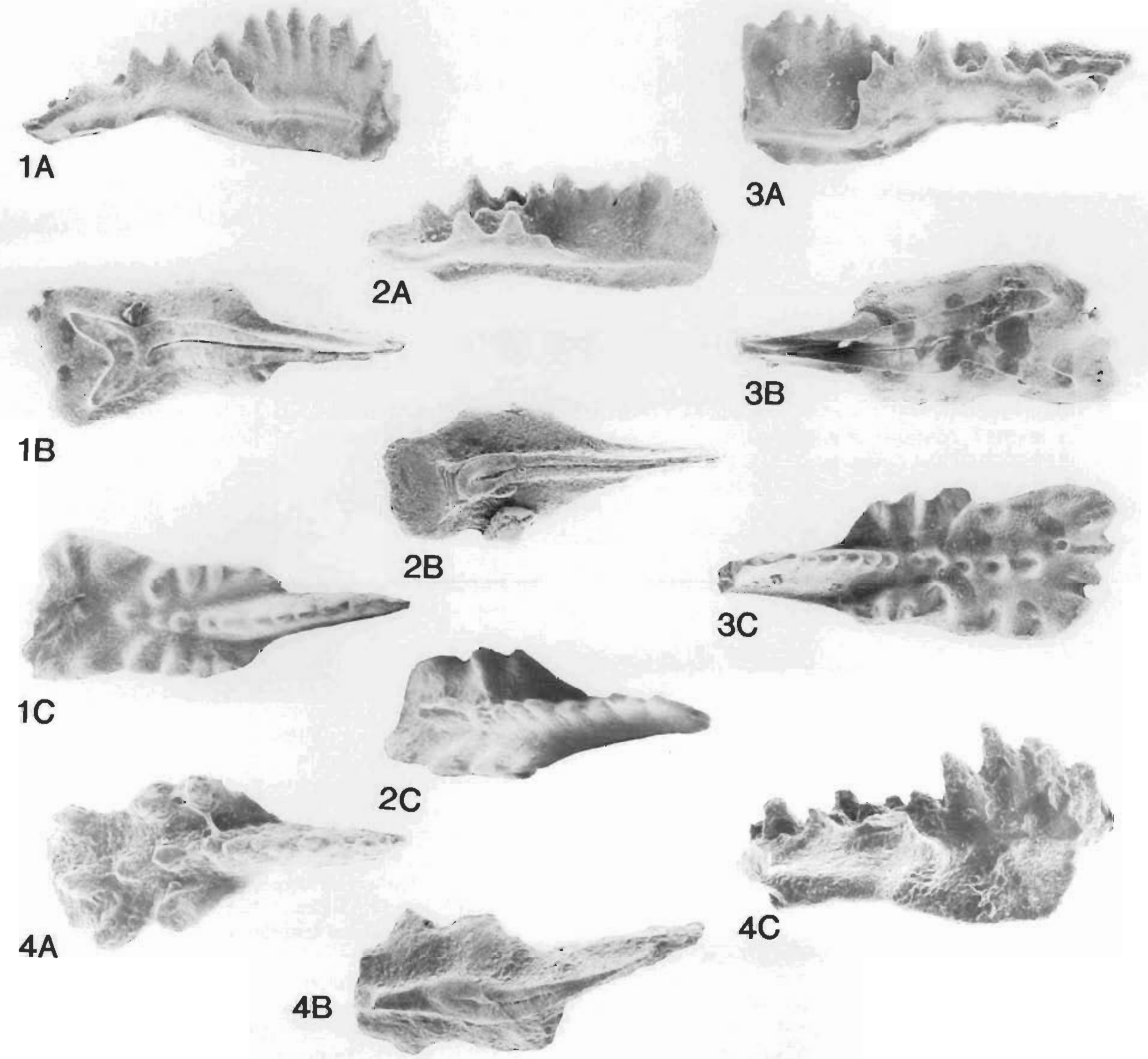

Figure 8. ***Epigondolella triangularis*** **and** ***Epigondolella spiculata*****, all Pa elements. Magnification as noted.**

1. *Epigondolella triangularis* (CPC 31653)[Sahul Shoals 1, 6187-6191] right element (all X70); a, outer lateral view; b, aboral view; c, oral view.
2. *Epigondolella triangularis* (CPC 31654)[Sahul Shoals 1, 6187-6191] right element (all X140); a, outer lateral view; b, aboral view; c, oral view.
3. *Epigondolella triangularis* (CPC 31655)[Sahul Shoals 1, 6187-6191] right element (X75); a, inner lateral view; b, aboral view; c, oral view.
4. *Epigondolella spiculata* (CPC 31656)[RS96-DR/28-3] left element (all X140); a, oral view; b, aboral view; c, inner lateral view.

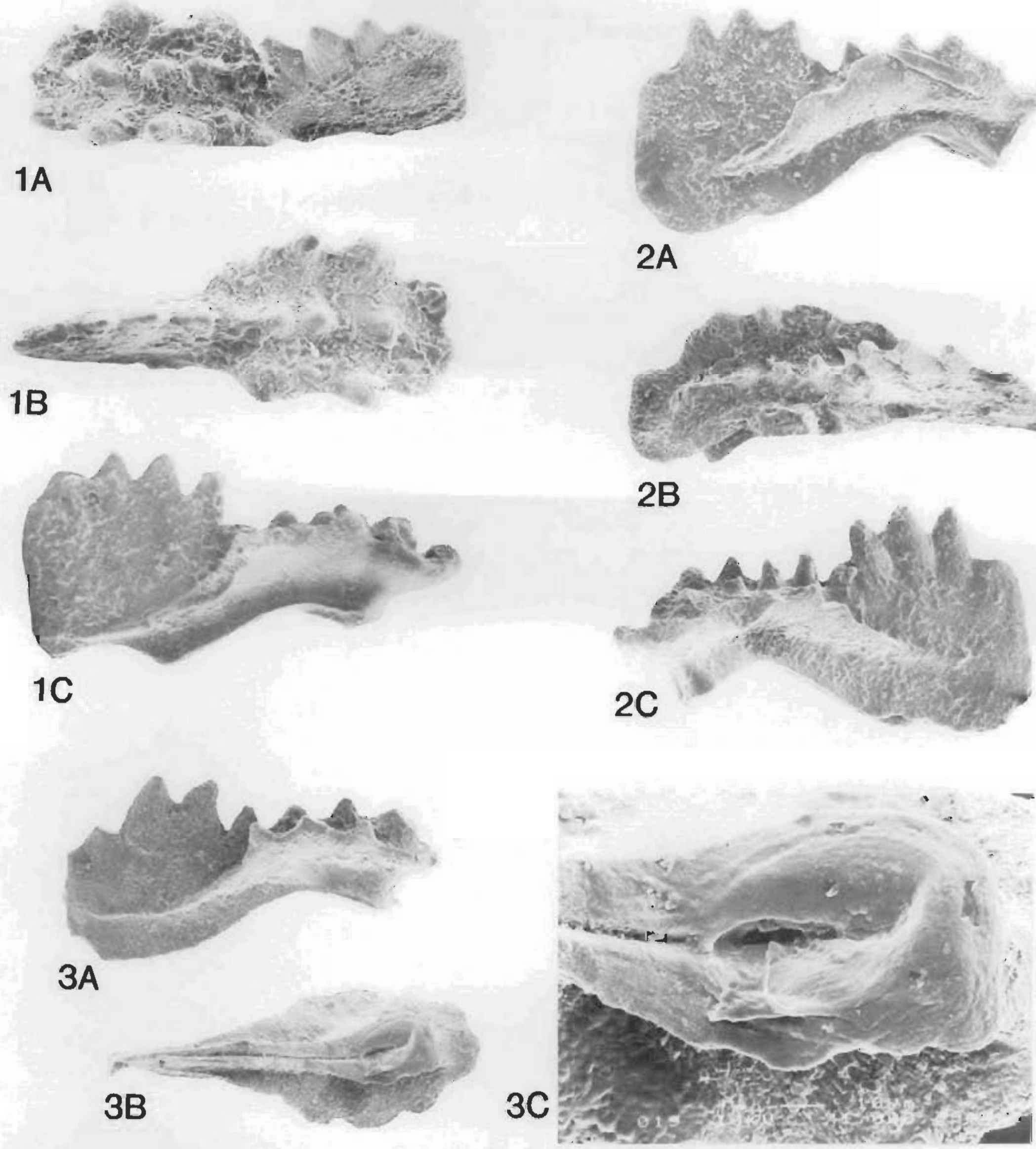

Figure 9. ***Metapolygnathus primitius*****, all Pa elements. All figures X140, except as noted.**
1. *Metapolygnathus primitius* (CPC 31657)[Sahul Shoals 1, 7376-7382] right element; a, oblique oral view; b, oral view; c, inner lateral view.
2. *Metapolygnathus primitius* (CPC 31658)[Sahul Shoals 1, 7376-7382] left element; a, outer lateral view; b, oral view; c, inner lateral view.
3. *Metapolygnathus primitius* (CPC 31659)[Sahul Shoals 1, 7376-7382] right element; a, inner lateral view; b, aboral view; c, enlargement (X500) of aboral view showing the basal cavity.

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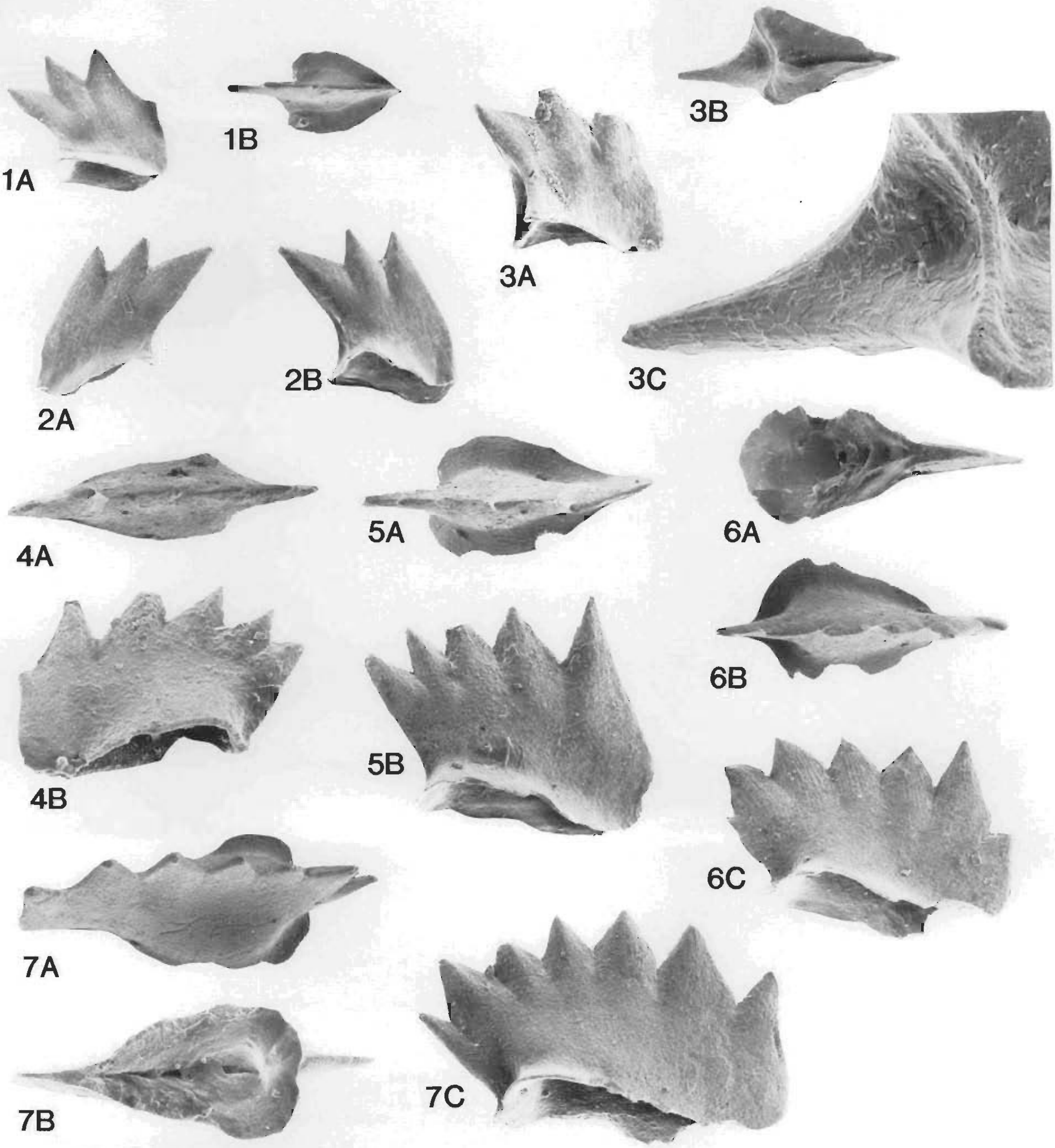

Figure 10. *Misikella posthernsteini* and *Misikella hernsteini*, all Pa elements. All figures X140, except as noted.
1. *Misikella posthernsteini* (CPC 31660)[RS56-DR/014 A(4)] right element; a, outer lateral view; b, oral view.
2. *Misikella posthernsteini* (CPC 31661)[RS56-DR/014 A(4)] ?right element; a, inner lateral view; b, outer lateral view.
6. *Misikella hernsteini* (CPC 31665)[RS56-DR/014 A(5)] right element; a, aboral view; b, oral view; c, outer lateral view.
7. *Misikella hernsteini* (CPC 31666)[RS56-DR/014 A(5)] right element; a, oral view; b, aboral view; c, outer lateral view.
3. *Misikella posthernsteini* (CPC 31662)[RS56-DR/014 A(4)] left element; a, inner lateral view; b, aboral view; c, enlargement (X550) of aboral view.
4. *Misikella hernsteini* (CPC 31663)[RS56-DR/014 A(5)] left element; a, oral view; b, outer lateral view.
5. *Misikella hernsteini* (CPC 31664)[RS56-DR/014 A(5)] right element; a, oral view; b, outer lateral view.

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Sample locality information

Well localities

	LATITUDE	LONGITUDE
Ashmore Reef 1	−12.180472	123.086278
Mt. Ashmore 1B	−12.560081	123.206144
Pokolbin 1	−11.523520	124.55254
Sahul Shoals 1	−11.427221	124.54723
Troubadour 1	−9.734394	128.123753

AGSO dredge localities

Rig Seismic dredge site localities, North West Shelf

RS56–DR/14	−16°31.3 to	115°26.5 to
	−16°34.0	115°32.7

Northern slope of the Wombat Plateau.

RS95–DR/10	−16°23.34 to	118°09.66 to
	−16°24.36	118°10.75

North of Clerke Canyon, western slope, Rowley Terrace.

RS95–DR/12	−16°29.16 to	118°59.55 to
	−16°28.76	118°03.64

South of Clerke Canyon, western slope, Rowley Terrace.

RS96–DR/28	−16°18.496 to	118°23.419 to
	−16°18.00	118°24.70

Western slope of Rowley Terrace.

Appendix 1. Conodont identification in petroleum exploration wells, North West Shelf

Conodont identifications in the Ashmore Reef 1, Mt. Ashmore 1B, Pokolbin 1, Sahul Shoals 1, and Troubadour 1 petroleum exploration wells, North West Shelf, offshore Western Australia.

Appendix 1.A. Conodont identifications from core samples from the Ashmore Reef 1 petroleum exploration well, North West Shelf, offshore Western Australia

Core	Depth	Conodont fauna	CAI
21	9741′ 0″–9743′ 9″	1 element	1
	(2969.82–2970.66 m)	1.00 kg	
	Misikella hernsteini (1 Pa)		
24	11024′ 0″–11030′ 0″	29 elements	1
	(3360.98–3362.81 m)	1.50 kg	
	Misikella hernsteini (29 Pa)		
	11030′ 0″–11035′ 0″	16 elements	1
	(3362.98–3364.33 m)	1.25 kg	
	Misikella hernsteini (14 Pa + 2 S)		
26	12060′ 0″–12065′ 1″	1 element	1
	(3676.83–3678.38 m)	1.10 kg	
	Epigondolella bidentata (1 Pa)		
	12065′ 1″–12070′ 0″	27 elements	1
	(3678.38–3679.88 m)	1.35 kg	
	Epigondolella bidentata (14 Pa + 13 S+M)		
27	12423′ 0″–12428′ 0″	0 elements	—
	(3787.50–3789.04 m)	1.20 kg	
	12428′ 0″–12433′ 0″	0 elements	—
	(3789.04–3790.55 m)	1.55 kg	

Appendix 1.B. Conodont identifications from cuttings samples from the Mt. Ashmore 1B petroleum exploration well, North West Shelf, offshore Western Australia

Core cuttings	Depth	Conodont fauna	CAI
	2370–2415 m	4 elements	1
	0.115 kg		
	?*Epigondolella* sp. indet. (2 P + 2 S)		
2415–2455 m		7 elements	1
0.125 kg			
	Epigondolella aff. *E. postera* (5 Pa + 2 S)		

Appendix 1.C. Conodont identifications from core samples in the Pokolbin 1 petroleum exploration well, North West Shelf, offshore Western Australia

1829.5–831.5 m		9 elements	1
1.90 kg			
	Metapolygnathus primitius (7 Pa + 2 S)		
1833.5–1835.5 m		26 elements	1
1.50 kg			
	Metapolygnathus primitius (20 Pa + 6 S)		
1836.6–1837.3 m		0 elements	—
0.85 kg			

Appendix 1.D. Conodont identification from core samples in Sahul Shoals 1 petroleum exploration well, North West Shelf, offshore Western Australia

Core 3	0 elements	—	
5195′ 0″ — 5198′ 1″			
(1583.84–1584.78 m)			
0.74 kg			
Core 4			
6187′ 5″ — 6191′ 11″		421 elements	1
(1886.4–1887.77 m)			
	Epigondolella triangularis (383 Pa + 4 Pb + 34 S+M)		
0.95 kg			
6191′ 11″ — 6197′ 0″		337 elements	1

(1887.77–1889.33 m)
Epigondolella triangularis (301 Pa + 2 Pb + 34 S+M)
1.00 kg

Core 5
7364′ 0″ — 7366′ 5.5″ 3 elements 1
(2245.12–22455.88 m)
Metapolygnathus primitius (1 Pa + 2 S+M)
0.48 kg

7366′ 5.5″ — 7371′ 11″ 28 elements 1
(2245.88–2247.53 m)
Metapolygnathus primitius (27 Pa + 1 S)
1.14 kg

7371′ 11″ — 7376′ 9.5″ 24 elements 1
(2247.53–2249.01 m)
Metapolygnathus primitius (23 Pa + 1 S)
0.94 kg

7376′ 9.5″ — 7382′ 7″ 71 elements 1
(2249.01–2250.79 m)
Metapolygnathus primitius (64 Pa + 7 S)
1.04 kg

7382′ 7″ — 7388′ 1.5″ 17 elements 1
(2250.79–2252.47 m)
Metapolygnathus primitius (12 Pa + 5 S+M)
1.09 kg

7388′ 1.5″ — 7393′ 0″ 15 elements 1
(2252.47–2253.96 m)
Metapolygnathus primitius (14 Pa + 1 S)
0.96 kg

Core 4
1883.2–1889.6 m 6 elements
Epigondolella triangularis (6 Pa)

Appendix 1.E. Conodont identifications from core samples in the Troubadour 1 petroleum exploration well, North West Shelf, offshore Western Australia

2840–2885 m 1 element 2
0.080 kg
Metapolygnathus cf. *M. primitius* (1 Pa)
2950–2955 m 0 elements
0.029 kg

3305–3330 m 0 elements
0.067 kg

Appendix 2. Conodont identification in AGSO — North West Shelf dredge samples

Conodont identifications in *Rig Seismic* dredge samples — from sites on the Wombat Plateau and margin of the Rowley Terrace, Northwest Shelf, offshore Western Australia.

Appendix 2.1. AGSO (BMR) Wombat Plateau dredge samples — Cruise 56

Conodont identifications in *Rig Seismic* dredge samples from site 014 on the northern margin of the Wombat Plateau.

56–DR/	WEIGHT (Kg)	CONODONTS	FISH
014 A(1)	2.85	25	+
	Misikella hernsteini (17 Pa + 1 Pb + 6 S)		
	Misikella posthernsteini (1 Pa)		
		sharp form	
014 A(2)	2.45	47	+
	Misikella posthernsteini (34 Pa + 13 S+M)		
		sharp form,	
		+ ramiform elements	
014 A(3)	2.75	13	+
	Misikella hernsteini (11 Pa + 2 S)		
		1 transition form — sharp	
014 A(4)	4.32	70	+
	Misikella posthernsteini (40 Pa + 38 S+M)		
		sharp form,	
		+ ramiform elements	
014 A(5)	2.63	111	+
	Misikella hernsteini (87 Pa + 1 Pb + 23 S+M)		
		forms early in transition to *posthernsteini*	
014 B(x)	5.00	4	+
	Misikella hernsteini (4 Pa)		
016 C(1)	1.04	0	+
	no conodonts		

Appendix 2.1. AGSO (BMR) Rowley Terrace dredge samples — Cruise 95

95/DR	WEIGHT	CONODONTS	FISH
10/6D	2.00	12	+
	Epigondolella postera (11 Pa + 1 S)		
12/6C	1.80	4	+
	Epigondolella sp. (3 Pa + 1 S)		
12/9C	2.00	20	+
	Epigondolella postera (6 Pa + 14 S)		
12/10D	2.00	26	+
	Epigondolella postera (18 Pa + 8 S+M)		

Appendix 2.2. AGSO (BMR) Rowley Terrace dredge samples — Cruise 96

96/DR	WEIGHT	CONODONTS	FISH

28/1 2.00 8 +
Misikella posthernsteini (3 Pa + 5 S+M)
round form,
+ ramiform elements

28/2 2.00 19 +
Misikella posthernsteini (15 Pa + 3 S)
sharp form,
+ ramiform elements

28/3 2.00 7 +
Epigondolella spiculata (6 Pa + 1 S)

28/6 2.00 6 +
Misikella posthernsteini (5 Pa + 1 S)
sharp form

AGSO Journal of Australian Geology & Geophysics, 15/1, 119–125

Mesozoic Bivalvia from Clerke and Mermaid Canyons, northwest Australian continental slope

J.A. Grant-Mackie[1]

Four sets of rock samples from two sites off the northwest Australian shelf in 3625–4480 m of water contain macrofaunas, mainly bivalves, of warm shallow-water origin. Mermaid Canyon (16° 19'S, 118° 23'E) provided many samples of oolitic calcarenite containing *Pseudopecten (Pseudopecten) dugong* n.sp., indicating an Early Jurassic age and Tethyan relationship. Three hand-specimens from the ridge forming the western edge of Clerke Canyon (16° 29'S, 118° 30'E) yielded a Norian coral–?*Lima*–oyster assemblage and the Norian–Rhaetian bivalve *Palaeocardita* aff. *globiformis* (Boettger). The latter shows relationship with southeast Asian (Indonesia–Vietnam–south China) forms.

Introduction

Four sets of rock samples from dredging by the Australian Geological Survey Organisation at two sites were received from Neville Exon for study. All are calcarenites with macrofossils, generally shelly, but in one case largely decalcified.

Fossils are mainly bivalves, which are reported on in detail, but other groups are also represented and are referred to below.

Bivalve descriptions are provided, followed by consideration of assemblages from each of the four sets of samples. Finally, brief biogeographic and palaeoecologic conclusions are offered. Type and figured specimens are held in the collections of AGSO, Canberra, and specimen numbers are those of the catalogue of that institution. Bivalve classification follows that of Vaught (1989).

Systematic palaeontology

Class Bivalvia Linné, 1758

Subclass Pteriomorphia Beurlen, 1944

Order Limoida Waller, 1978

Superfamily Limoidea Rafinesque, 1815

Family Limidae Rafinesque, 1815

Genus *Lima* Bruguière, 1797

Type species. (S.D., Cuvier, 1797): *Ostrea lima* Linné, 1758; Recent, Philippines Is.

?Lima sp. indet. Fig. 1A,B.

Material. A single valve (CPC 31431), lacking a significant portion of the ventrolateral region, and with most of the remainder of the shell exterior obscured by coral and a spongiomorph and the interior by what appears to be an oyster valve.

Description. The valve is moderately large (c. 90 x 50 mm), higher than long, of low inflation (c. 10 mm), with many (70–80?) flat-topped vertical-sided slightly wavey radial ribs seemingly of two orders and with narrower interstices; the low umbo and beak are flanked by small ears which bear weaker radials than on the disc. The hinge plate is large, triangular, with a long narrowly triangular deep median ligament pit on each side of which the hinge plate also shows a radial depression fading out ventrally.

Discussion. Although inadequately exposed, this valve is clearly a limid. In its sculpture and moderate shell thickness it is most like *Lima* s.s., although the shell material is thicker and the ligament pit is unusually narrow, as in *Ctenostreon*. It is too thick-shelled for *Acesta*, has much stronger sculpture than *Antiquilima*, and has a much higher cardinal area than *Ctenoides*. All other Mesozoic limids are much smaller. Too little is exposed to be able to attempt specific comparisons and even the generic location must be regarded as tentative.

Location and age. This shell is in sample 10A from dredge station 95/DR012, located from 16° 29.16'S, 118° 59.55'E to 16° 38.76'S, 118° 3.64'E on the ridge west of Clerke Canyon between 3625 and 4470 m depth. The associated coral gives a Norian age (see Stanley, 1994 — this issue).

Order Ostreoida Férussac, 1822

Suborder Ostreina Férussac, 1822

Superfamily Ostreoidea Rafinesque, 1815

?Ostreoidea gen. et sp. indet.

Material. A single valve embedded in matrix and covered on one surface by a limid valve.

Description. A moderately large valve (c. 80 x 55 mm) of oval outline, with a thick irregularly laminated shell and low inflation, but showing no internal or other diagnostic features.

Discussion. The thick irregularly laminated form of this shell is the only basis on which identification as a probable oyster can be made, for it lies largely within the rock with only the shell margins exposed.

Locality and age. As for ?*Lima* sp. (q.v.).

Family Gryphaeidae Vyalov, 1936.

Subfamily Gryphaeinae Vyalov, 1936.

Gryphaeinae gen. et sp. indet. Fig. 1C.

Material. One left valve (CPC 31433) in internal view,

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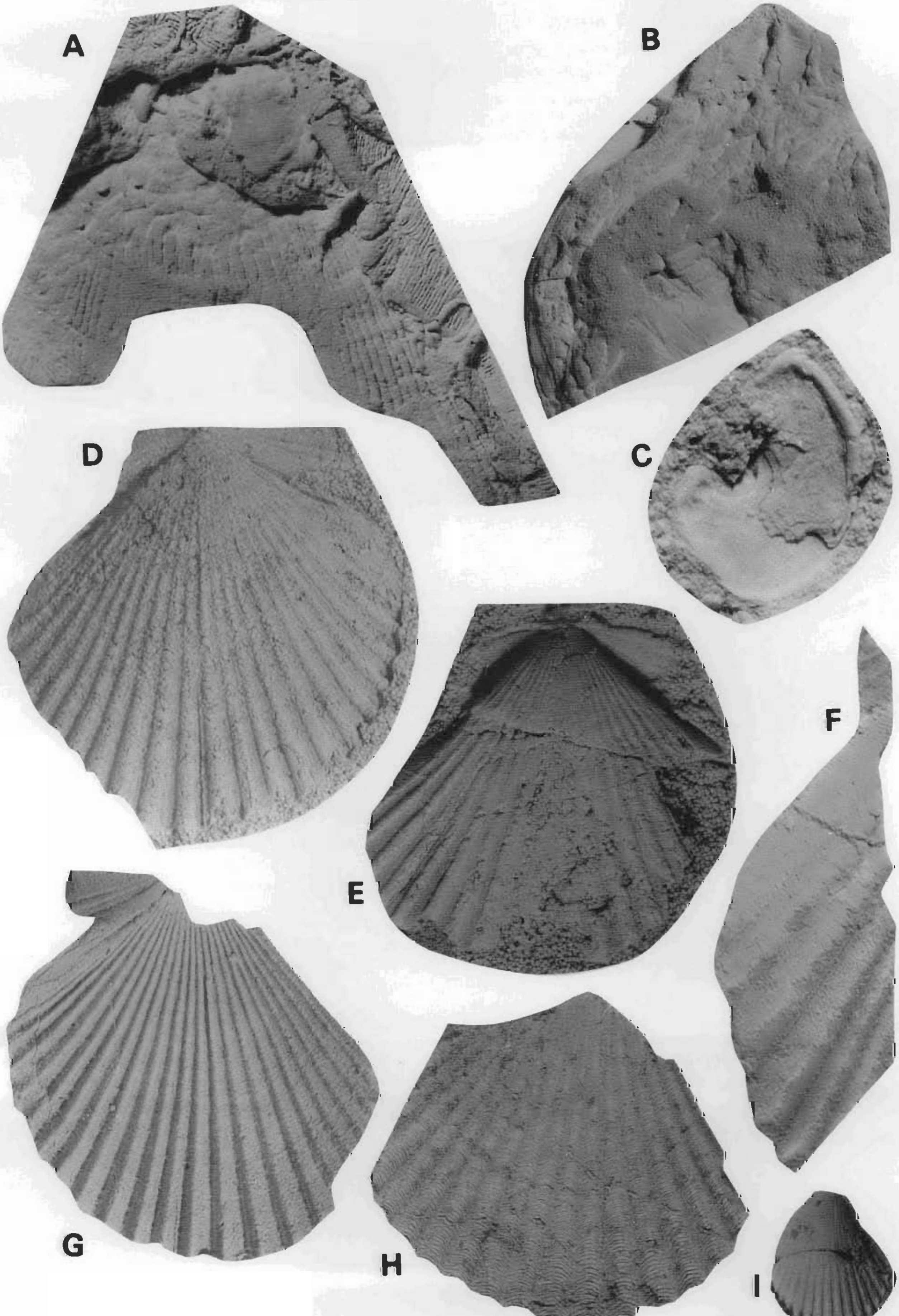
A
B
C
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I

rather exfoliated ventrally although outline preserved.

Description. Valve small (length = 6.8 mm, height = 9.3 mm), subpyriform to subquadrangular, flat with slightly raised anterior margin; shell material moderately thin; dorsal margin short, lacking any indication of chomata and without clear preservation of ligament pit; umbo low, insignificant; muscle scar partly obscured by matrix but apparently subcircular, located just dorsal of mid height and about one-third of length from posterior margin; exfoliation of inner shell layers ventrally obscures relationship between muscle scar surface and level of surrounding shell interior; no indication of presence of shell ornament.

Discussion. The shallow umbonal cavity, subcircular muscle scar and poorly developed ligamental area all point to location in the Gryphaeidae rather than Ostreidae. The valve cannot be a member of the Exogyrinae because of the short ligamental area. Lack of chomata indicates Gryphaeinae rather than Pycnodonteinae, as does the apparent lack of a commissural shelf, although at the size and state of preservation of this single specimen it would be unwise to assume too much. However, given the clear indication of an Early Jurassic age for the associated pectinid bivalves and the appearance of the Pycnodonteinae only in the Early Cretaceous (Stenzel, *in* Cox et al., 1971), allocation to the Gryphaeinae, which appeared in the Late Triassic, can be made with some confidence.

The valve cannot be readily allotted to a gryphaeine genus. It could be either a juvenile *Gryphaea* with large initial attachment area and with the radial posterior sulcus at least initially evanescent; or *Deltoideum* at an early stage of growth before significant lateral expansion of the ventral region. It is more feasibly a young *Liostrea* of a species lacking the sulcus, such as, for instance, the Rhaetian to Sinemurian type species, *L. hisingeri* (Nilsson) (see Stenzel, *in* Cox et al., 1971). It is much less likely to be a *Catinula* because of the absence of any indication of radial costation.

Locality and age. This valve comes from brownish-yellow calcarenite of dredge station 95/DR08, located from 16° 19.1'S, 118°22.9'E to 16° 18.9'S, 118° 23.2'E in 4400–4230 m in the Mermaid Canyon and, from the presence of *Pseudopecten* s.s. in the same collection, is believed to be of Early Jurassic age. Both *Gryhaea* and *Liostrea*, it should be noted, occur in strata of this age whereas *Deltoideum* is not known until the Bathonian (Stenzel, *in* Cox et al., 1971).

Suborder Pectinina Waller, 1978

Superfamily Pectinoidea Rafinesque, 1815

Family Pectinidae Rafinesque, 1815

Genus Pseudopecten Bayle, 1878

Type species (by monotypy). *Pectèn aequivalvis* J. Sowerby, 1816; Lower Jurassic, England.

Pseudopecten (Pseudopecten) dugong n. sp. Fig 1D–I.

Material. Holotype (No CPC 31434), internal and external casts of an almost complete right valve; 6 Paratypes (No CPC 31435–CPC 31440), internal and external casts of two valves, internal casts of two complete and one incomplete right valves, internal cast of one incomplete left valve; various other fragmentary casts of valves, all isolated.

Etymology. An allusion to the provenance of this species, Mermaid Canyon. The Latin "Sirenes" were the nearest approximation to mermaids of European seagoing mythology (and in French mermaid is "sirène"). It is widely believed that the myth of mermaids was based upon sightings of the Atlantic manatees and Indo-Pacific dugongs (mammalian Order Sirenia). The specific name is a noun in apposition.

Dimensions (in mm):

		Length	*Height*	*Inflation*
Holotype	RV	37	36	6
Para.	RV	c.36	c.34.5	c.5.5
Para.	?	c.38	c.36	c.3
Para.	RV	c.13	c.12.5	1.5

Diagnosis. *Pseudopecten* s.s. with umbonal angle of c. 90°, 21–25 radial ribs, fine comarginal ornament which tongues abapically in rib interspaces, and many short striations on dorsolateral margins.

Description. Valve of moderate size, suborbicular, height a little less than length; inflation low on all fragments, so shell probably equivalve or nearly so; subequilateral, only slightly produced posteroventrally (postumbonal length = 20 mm on valve 38 mm long); antero- and posterodorsal margins concave, forming umbonal angle of c. 90°; auricles well separated from disc, of moderate size, anterior larger and rounded anteriorly, posterior obliquely truncated; byssal notch about quarter length of anterior auricle, rounded apically, widening anteriorly; disc sculpture of 21–25 broad, angular radial plicae with narrower interspaces, both plicae and spaces crossed by fine, closely spaced (c. 5 mm), very low comarginal lamellae which lean over towards adjacent margin and run marginwards in rib interspaces (Fig. 1H); internally ribs expressed as rounded furrows; anterior auricle bears 3–4 radials of variable strength dorsal to shallowly concave zone of byssal notch; posterior auricle with 2–3 radials; disc margin scalloped laterally and ventrally; dorsolateral margins of disc curve almost vertically and bear many short striations normal to margin (Fig. 1F).

Discussion. The above characters fit best the genus *Pseudopecten*: the fossils lack the oblique spines found

Figure 1. A, B. ?*Lima* sp. (CPC31431), dredge station 95/DR012, west of Clerke Canyon; Norian. A. external view of partly exposed valve, showing also the scleractinian coral *Pamiroseris rectilamellosa* (Winkler), X2. B. internal view of sediment filled valve, showing partial outline and cardinal area with narrow ligament pit, X1. C. Gryphaeinae gen. et sp. indet. (CPC31433), dredge station 95/DR08, Mermaid Canyon, Early Jurassic; interior view of small partly exfoliated valve, X6. D–I. *Pseudopecten (Pseudopecten) dugong* n. sp., dredge station 95/DR08, Mermaid Canyon, Early Jurassic. D,G: holotype (CPC31434), right valve in internal and external view, X2; E: paratype right valve (CPC31437), internal cast, X2; F: paratype valve fragment (CPC31440) showing short vertical striae on dorsolateral margin, X3; H: paratype valve fragment (CPC 31435b), an external cast, showing well-preserved comarginal lamellae tonguing down rib interspaces, X2; I: paratype (CPC31436a), internal cast of a juvenile left valve, X2.

on some radials in the subgenus *Echinopecten* and fit rather the nominate subgenus. The type species, *P. (P.) equivalvis* (Sowerby), from the Lower Jurassic of Europe and the Americas, reaches a larger size than the present shells, has a larger apical angle, and although there is a similar number of radial plicae they generally lack the fine concentric sculpture of our shells. The European Lower and Middle Jurassic *P. (P.) dentatus* (Sowerby) has the same concentric sculpture, including the down-sulcus tonguing, and is of a similar size but has fewer ribs (16–19); *P. (P.) veyrasensis* (Dumortier) has even fewer ribs (13–15) but also shows the down-sulcal tonguing of the concentric ornament. Both these latter species are also said to have vertically striated disc flanks (Johnson, 1984), a feature visible in the present specimens (Fig. 1F) which are closest to *P. (P.) dentatus*.

Johnson (1984, p. 60) commented that within the subgenus two species groups could be distinguished on the "presence or absence of high, vertically striated disc flanks and comarginal striae which tongue down the sulci". Clearly, *P. dugong* possesses these features, along with *P. dentatus* and *P. veyrasensis* which he separated on the basis of rib frequency. He suggested, however, that the two may be part of a single polymorphic species with a bimodal plical frequency, since in other features the two are "virtually indistinguishable" (Johnson, 1984, p. 60). He also points, nevertheless, to a slight difference in stratigraphic range between the two species, and the evidence of rib frequency in the present collection must be seen as added evidence for the valid separation of the two European species.

This record is the first for the subgenus outside Europe and a few occurrences in North and South America, all in rocks of Early and Middle Jurassic age (Hettangian to Toarcian for *P. equivalvis*, Hettangian to Bajocian for *P. dentatus*, and Hettangian and Pliensbachian for *P. veyrasensis*).

Johnson (1984) concluded that *P. dentatus* and *P. veyrasensis* favoured high-temperature/low-turbidity environments, with shells being large and abundant in low palaeolatitudes and condensed sequences, sometimes associated with hermatypic corals. The occurrence of *P. dugong* in oolitic calcarenites indicates similarly a warm shallow mud-free environment.

Locality and age. *Pseudopecten (P.) dugong* is found in oolitic calcarenite with minor interbedded calcisiltite of dredge station 95/DR08, located from 16° 19.1'S, 118° 22.9'E to 16° 18.9'S, 118° 23.2'E in 4400–4230 m in the Mermaid Canyon. An Early Jurassic age is favoured, given that Middle Jurassic ages for other members of the subgenus are rare (Johnson, 1984).

Subclass Heterodonta Neumayr, 1884

Order Veneroida H & A Adams, 1856

Superfamily Carditoidea Fleming, 1820

Family Carditidae Fleming, 1828

Subfamily Palaeocarditinae Chavan, 1969

Genus *Palaeocardita* Conrad, 1867

Type species (O.D.). *Cardita austriaca* Hauer, 1853; Rhaetian, Austria.

Palaeocardita n. sp. aff. *P. globiformis* (Boettger, 1880). Fig 2A,B.

aff. 1880 *Cardita globiformis* Boettger, 38, pl. 1, fig. 21–22; pl. 2, fig. 12–16.

aff. 1910 *C. globiformis*, Wanner, 738.

aff. 1915 *Palaeocardita globiformis*, Jaworski, 115, pl. 45 (3), fig. 1–2.

aff. 1965 *P. globiformis*, Vu Khuc & others, 46, pl. 15, fig. 8–10.

Material. CPC 31432, an articulated individual lacking ventral and posterior margins due to erosion and with the surface somewhat corroded and partly obscured by dense fine calcarenite.

Description. Shell of medium size (c. 30 mm long, c. 28 mm high), thick-shelled, moderately inflated (1 valve = 13 mm), suborbicular, subequilateral, with small orthogyral beaks; ornament of 25 strong sharp-crested radial ribs with rounded interstices crossed by growth lines which produce very low ventrally projecting scales on rib crests spaced 1–2 mm; lunular area masked by matrix, but escutcheon relatively short and broad; maximum inflation at mid height a little posterior to

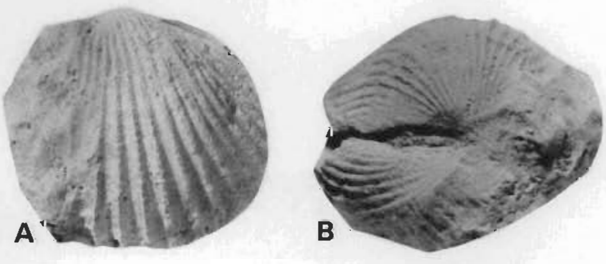

Figure 2. *Palaeocardita* n.sp. aff. *globiformis* (Boettger) (CPC31432), dredge station 95/DR012, west of Clerke Canyon, Norian-Rhaetian; A: left valve; B: dorsal view, left valve above, X2.

beaks; internal characters not seen.

Discussion. The external characters clearly indicate *Palaeocardita*. Most species of that genus are posteroventrally elongate (e.g. the type species, *P. austriaca* — Cox et al., 1969 — and the St Cassian *P. allasinazi, P. beneckei, P. crenata, P. pichleri* and *P. seelandica*; Zardini, 1981), laterally oval (e.g. the Southeast Asian *P. singularis* and the American *P. silberlingi* — e.g. Wen et al., 1976; Newton et al., 1987), or quadrangular (the widespread *P. trapezoidalis* and the Chinese *P. rhomboidalis* — e.g. Hudson & Jefferies, 1961; Wen et al., 1976); a few are subcircular and subequilateral (e.g. the southeast Asian *P. buruca, P. globiformis, P. langnongensis* and *P. mansuyi* — Hudson & Jefferies, 1961; Jaworski, 1915; Wen et al., 1976), but none of these has the high rib count of this specimen. *P. buruca, P. langnongensis* and *P. mansuyi* are also much more strongly inflated. *P. globiformis* s.s. has more ribs than these three species, at 18–20, inflation comparable with that of our specimen, and a very slightly prosogyral beak. The Yunnan *P. globiformis healeyae* (Reed, 1927) has fewer ribs than the nominate form. *P. pichleri* has 24 ribs, but is much more inflated and has a very short hinge; *P. crenata* has 25 ribs but low inflation and a concave posterior flank; *P. silberlingi* has 25–35 ribs but is grossly oval and inequilateral.

The present specimen does not fall within the morphologic range of any previously described species and is judged new to science. It is most closely similar to *P. globiformis* from Norian–Rhaetian strata of Misool–Sumatra–Vietnam–southern China in outline, inflation, subequilateral form, slightly prosogyral beak, and fine concentric sculpture, differing only in the possession of a few more ribs than in Boettger's species.

Locality and age. This specimen comes from sample 11, dredge station 95/DRO12, located from 16° 29.16'S, 118° 59.55'E to 16° 28.76'S, 118° 3.64'E at 3625–4470 m on the ridge west of Clerke Canyon. *Palaeocardita* is a Middle and Upper Triassic genus, with most species of Carnian–Norian age. *P. globiformis* was described from the Rhaetian of Sumatra (Boettger, 1880), then found more widely in Norian strata of Sumatra (Musper, 1929), Misool and Vietnam. A Norian–Rhaetian age is likely for the present specimen.

?*Palaeocardita* sp. indet.

Material. 6–8 incompletely exposed valves in very hard cemented tuffaceous coarse bioclastic calcarenite.

Description. Valves small (up to 5 mm across), subcircular, moderately inflated, with at least 15 (probably c.20) strong radial ribs with granular or gemmate crests and rounded interspaces of comparable width.

Discussion. These valves are too poorly exposed to provide certain identifications, but in the features displayed they are comparable with the previous species and could be conspecific. However, none of the specimens seen shows the full outline, complete ornamentation or any internal characters and the identification must be tentative at best.

Locality and age. These valves occur in sample 9 from dredge station 95/DR012 and suggest a Middle to Late Triassic age.

Associated macrofauna and implications

In this section, other macrofossils occurring with the bivalves are briefly reported and for each sample or set of samples palaeoenvironmental and palaeogeographic implications are considered.

DR08/sample 4

The fragments of brownish-yellow oolitic calcarenite contain, in addition to many fragments and a few more complete valves of *Pseudopecten (P.) dugong* n.sp., a few crinoid stem ossicles. Some are pentagonal, some pentastellate, and there is also an articulated series of three cylindrical forms. They are partly decalcified, show indifferent preservation of articulation surface features, and no identification has been attempted, although crinoids are potentially helpful both in age and palaeogeographic considerations.

The macrofauna is thus

Isocrinida gen. et sp. (spp.) indet.

Gryphaeinae gen. et sp. indet.

Pseudopecten (Pseudopecten) dugong n. sp.

Pseudopecten is elsewhere a genus characteristic of oolitic and reefal deposits of shallow moderate-energy waters (Johnson, 1984), and the oolitic nature of sample DR08/04 and the gryphaeine oyster confirm a similar habitat for the new species.

The subgenus *Pseudopecten* has a dominantly European distribution and indicates Tethyan connections at the time. The lack of records of *Pseudopecten* from nearby southeast Asia may seem anomalous in this regard but, accepting the likelihood of an Early Jurassic age for the sample, it should be recalled that marine strata of this age are rare or missing over a wide region, with later marine Jurassic rocks unconformable on Triassic or Permian strata in, for example, eastern Indonesia (Sula: Sato et al., 1978; Misool: Pigram et al., 1982), Thailand (Meesook & Grant-Mackie, 1993) and Myanmar (Komalarjun & Sato, 1964). Marine Lower Jurassic strata occur in Vietnam (Hayami, 1984), but elsewhere in southeast Asia marine transgression did not occur until Toarcian times (see, e.g. Kobayashi, 1984).

This situation exists also nearby on the Australian landmass, with earlier Jurassic strata either absent or of non-marine origin (e.g. Skwarko, 1974). The shallow Tethyan nature of sample DR08/04 indicates, in the light of the above points, that whatever its precise age its deposition probably occurred in the same transgressive cycle that produced the marine Jurassic sequences of the on-land Perth and Carnarvon Basins and in Indonesia and mainland Southeast Asia.

DR12/sample 9

Apart from the small ?*Palaeocardita* valves and a number of unidentifiable organic remains this yellow-brown well-cemented calcarenite contains many small (c. 2 mm in diameter) circular structures which may be algal spheroids; the basal 6 mm of an echinoid spine which has a collar diameter of 3 mm and a shaft of 2 mm in diameter just above the collar; and the cross-section of the pedicle valve of a medium-sized (28 mm long) thick-shelled brachiopod with no radial sculpture. The cobble also shows what seems to be a partial calicular view of a small solitary coral 6 mm in diameter with two cycles of septa.

Few useful conclusions beyond what has been stated above concerning the bivalves can be drawn from this biota.

DR12/sample 10A

Apart from the limid and probable oyster, this hand specimen of yellow-brown moderately cemented quartzose calcarenite contains the scleractinian coral and spongiomorph reported on elsewhere (Stanley, 1994 — this issue) and the molluscs support the conclusions drawn there, *viz.* warm shallow seas, in fact reefal from the presence of the particular coelenterates.

DR12/sample 11

This single cobble of pale yellowish-brown muddy medium calcarenite has, in addition to *Palaeocardita* aff. *globiformis*, 6–7 cylindrical objects less than 1 mm diameter and up to 1.5 mm long with fine axial canals. These are interpreted as crinoid remains, either stem ossicles or cirral segments. The fauna is thus

Crinoidea indet.

Palaeocardita n. sp. aff. *P. globiformis* (Boettger)

Palaeocardita is a typically Tethyan genus more widespread than *Pseudopecten* in both time and space, and outside the Tethyan realm is known in the New Zealand Norian (Marwick, 1953; Kristan-Tollmann, 1987). This form shows a distinct eastern Tethyan affinity. *Palaeocardita* is a frequent member of neritic Triassic faunas and its occurrence in a lithology that is non-oolitic and finer-grained than those of DR08/04 or DR12/10A suggests a slightly more offshore site, possibly mid shelf.

Acknowledgements

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AGSO Journal of Australian Geology & Geophysics, 15/1, 127–133

Upper Triassic spongiomorph and coral association dredged off the northwestern Australian shelf

G.D. Stanley, Jr.[1]

Upper Triassic corals and spongiomorphs dredged during BMR Cruise 95 from the Rowley Terrace, offshore Canning Basin of northwestern Australia, indicate possible new occurrences of reef facies. These are comparable to counterparts known from the Northern Limestone Alps of central Europe. A branching spongiomorph, represented by the genus *Spongiomorpha* sp. and two coral taxa, *Pamiroseris rectilamellosa* (Winkler) and *Retiophyllia tellae* (Stoppani), are reported herein. Collectively, these fossils indicate a Late Triassic (Norian–Rhaetian) age. Although different in taxonomic composition, the fauna compares with one previously reported from a Late Triassic ODP reef site (site 764) on the Wombat Plateau, some 350 km to the west. The Rowley Terrace occurrences may represent an eastward extension of the Wombat reefs, developed along the rifted margin of Gondwana.

Introduction

Two specimens, a flat, light-colored, partially leached but highly fossiliferous limestone cobble (95/DR12/10A), approximately 10.0 x 8.5 cm, and a limestone block 10 x 8 cm (95/DR7/15) form the basis of this report. The cobble was sent initially to Prof. J.A. Grant-Mackie (University of Auckland) and kindly relayed to me by him for study. The block was provided by Neville Exon (Australian Geological Survey), thin sections of which have already been studied (Colwell et al., 1994 — this issue). Both samples were obtained from dredges of the BMR Cruise 95.

For such small samples, they contain a diversity of organisms — 95/DR12/10A: brachiopod, crinoid ossicles, a large coarse-ribbed oyster-like bivalve mollusc, a spongiomorph hydrozoan and several colonies of encrusting thamnasteroid colonial corals. 95/DR7/15: *in situ* phaceloid corals, bivalves, gastropods, and serpulid worm tubes. The thamnasteroid corals appear to be encrusting the shell of the large bivalve, while the phaceloid corals represent *in situ* constructional coral framework.

The spongiomorph and thamnasteroid corals were retrieved from two dredge stations off Rowley Terrace in the offshore Canning Basin, northwestern Australia (Fig. 1). 95/DR012/10A is from the ridge west of Clerke Canyon (16° 29′ South, 188° 02′ East, 4470–3625 m depth). The phaceloid corals are from dredge station 95/DR07/15A (16° 20′ South, 118° 23′ East, 4530–3900 m depth) from a site near the mouth of Mermaid Canyon (Fig. 2). Thin sections (95/DR07/15A and B) from this sample were studied by Dr. E. Kristan-Tollmann (unpublished sample descriptions) who reported *?Glomospirella*-type foraminifers, problematica, and *Alpinophragmium perforatum* Flügel. Collectively, these fossils indicated a Norian–Rhaetian age.

From the palaeogeographic point of view, these corals are very significant. They have direct affinities with distant counterparts from Norian–Rhaetian reef complexes of the Northern Calcareous Alps and appear to represent reef facies comparable to these deposits (Flügel, 1981).

As discussed below, the presence of *Spongiomorpha* sp. and *Pamiroseris rectilamellosa* (Winkler) most certainly indicates an Upper Triassic age for the sample. *Retiophyllia* (sample 95-DR07/15B) indicates an Upper Triassic (possibly Rhaetian) age. The presence of a massive-growing spongiomorph, phaceloid corals and abundant encrusting corals, brachiopods, molluscs, and crinoids could indicate the presence of an Upper Triassic reef complex.

Until recently, neither Upper Triassic carbonate rocks nor Triassic reef complexes were known to be associated with the Australian continent. However, ODP Leg 122 (site 764) on the Wombat Plateau (Exmouth Plateau) of northwestern Australia, some 240 km to the west of the present sites, reported a 200 m thick, Upper Triassic (Rhaetian) reef sequence with abundant sponges, spongiomorphs, corals and a variety of typical Tethyan reef microfacies types (Röhl, 1991). This reef is believed to have developed on the rifted continental margin of northeastern Gondwana (von Rad et al., 1989).

Microfossils dredged from the nearby Swan and Cygnet Canyons/Emu Escarpment, and the Wombat Plateau during the 1979 *Sonne* cruise 8 most likely also belong to this carbonate complex. They were studied by Kristan-Tollmann & Gramann (1992). Foraminifers, echinoderms and ostracodes were characteristic of Tethyan Alpine types and indicated a Norian to Rhaetian age.

Details of the Upper Triassic sponge and coral dominated communities (with associated hydrozoan–tabulozoan communities) obtained from ODP 122 cores were given by Sarti et al. (1992). These authors interpreted the ancient communities as belonging to a Rhaetian pinnacle-type "reef complex". The reported reef corals including *Astraeomorpha crassisepta, A. confusa, Margarosmilia charlyana*, and *Retiophyllia paraclathrata* and a branching spongiomorph, *Spongiomorpha ramosa* Frech. In species composition, all these taxa differ from those described herein. ODP Wombat cores were obtained from various sites at water depths of 2000 to nearly 3000 m on a tectonic horst structure which is steeply tilted to the north (Röhl et al., 1991). Although all of the samples were obtained from ODP cores drilled below a Tertiary sedimentary cover, the nature of the tilted structure does not preclude the possibility that some Upper Triassic carbonate rocks should lie exposed on the seafloor (e.g. along the southeastern side of the the Wombat Plateau and elsewhere on the North West Shelf margin).

The spongiomorphs and corals reported below provide additional evidence for the presence of reef development and typical Upper Triassic Tethyan reef organisms. The reef aspect of these samples from dredge sites on Rowley Terrace matches well with previous discoveries from ODP samples described from the Wombat Plateau, some 350 km

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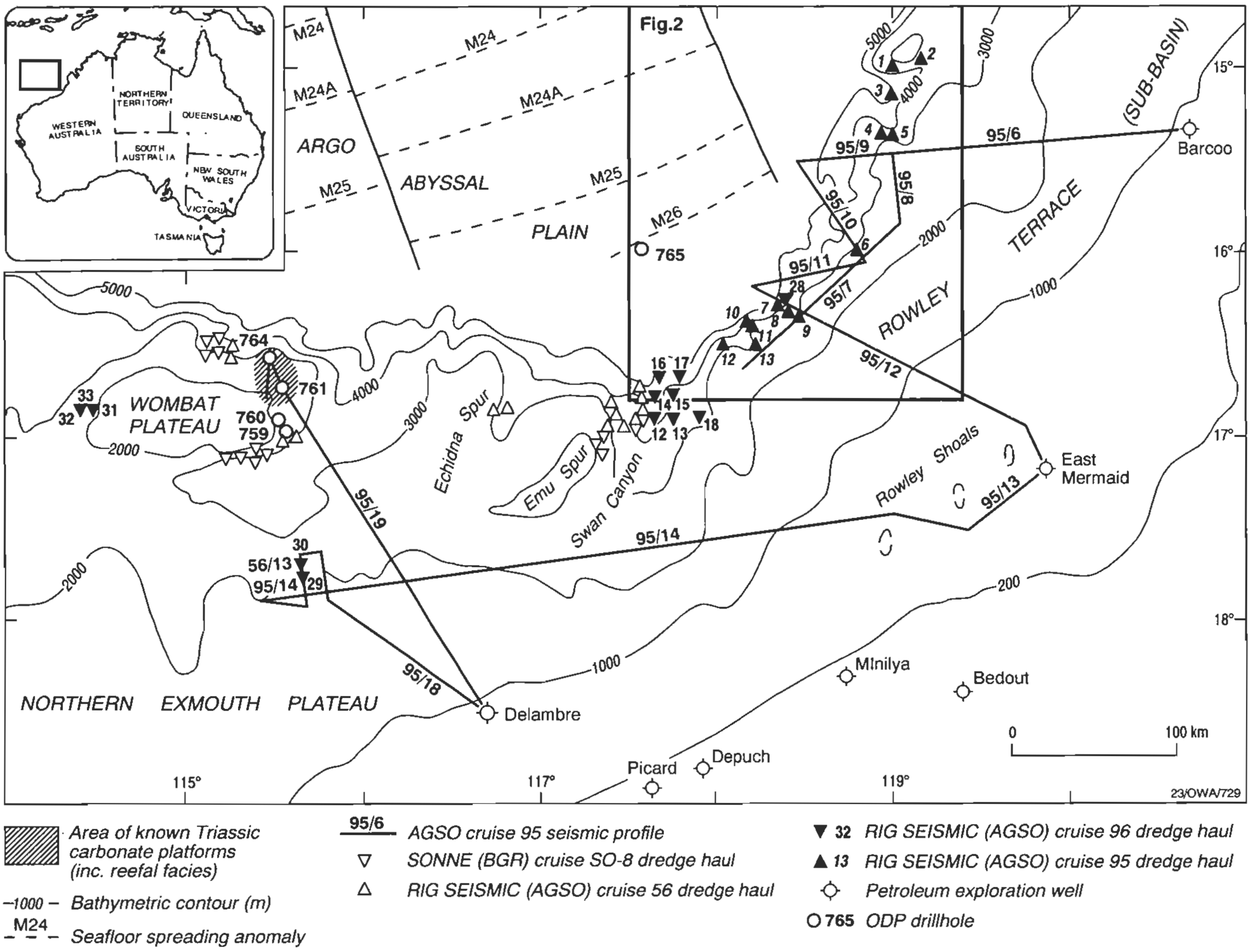

Figure 1. Location map of the northern Exmouth Plateau–Rowley Terrace margin showing locations of BMR Survey 95 and 96 dredge hauls and ODP sites on the Wombat Plateau. Area of Triassic reefs indicated. For details of the Rowley Terrace haul sites (inset), see Figure 2.

to the west. These observations indicate a possible eastward extension of reef facies, which is still present along the rifted margin of northwestern Australia.

Description of the fossils

Order SPONGIOMORPHIDA Alloiteau, 1952

Family SPONGIOMORPHIDAE Frech, 1890

Genus SPONGIOMORPHA Frech, 1890

Type species. *Spongiomorpha acyclica* Frech, 1890

SPONGIOMORPHA sp.

Fig. 3 A–B

Material. Australian Commonwealth Collection No. CPC 31429, one longitudinal surface exposure 25 x 20 mm in dimensions on dredge sample 95/DR12/10A.

Discussion. The coenosteum of the specimen is leached and poorly preserved, but it can be assigned to the genus *Spongiomorpha*. It bears some resemblance to *S. acyclica* Frech (1890) known from Norian–Rhaetian reef complexes of the Northern Calcareous Alps, but the quality of preservation of the Australian specimen precludes species assignment.

Distribution. The genus *Spongiomorpha* is greatly in need of revision. As defined by Frech (1890), this genus was restricted to the Upper Triassic, but it has since been described from rocks ranging in age from Early Jurassic to Early Cretaceous. The genus is widely distributed throughout the Tethys region and also occurs in western North America (Stanley & Whalen, 1989). Upper Triassic *Spongiomorpha* has been reported in ODP samples off the Wombat Plateau of northwestern Australia (Sarti et al., 1992).

Order SCLERACTINIA Bourne, 1900

Suborder ASTRAEOINA Alloiteau, 1952
(as Astraeoida)

Family PAMIROSERIIDAE Melnikova, 1975

Genus PAMIROSERIS Melnikova, 1971

Type species. *Thamnastraea meriani* Stoppani, 1860

PAMIROSERIS RECTILAMELLOSA (Winkler, 1861)

Fig. 1A, C–E.

Thamnastraea rectilamellosa Winkler, 1861, p. 487, pl. 8, fig. 7.

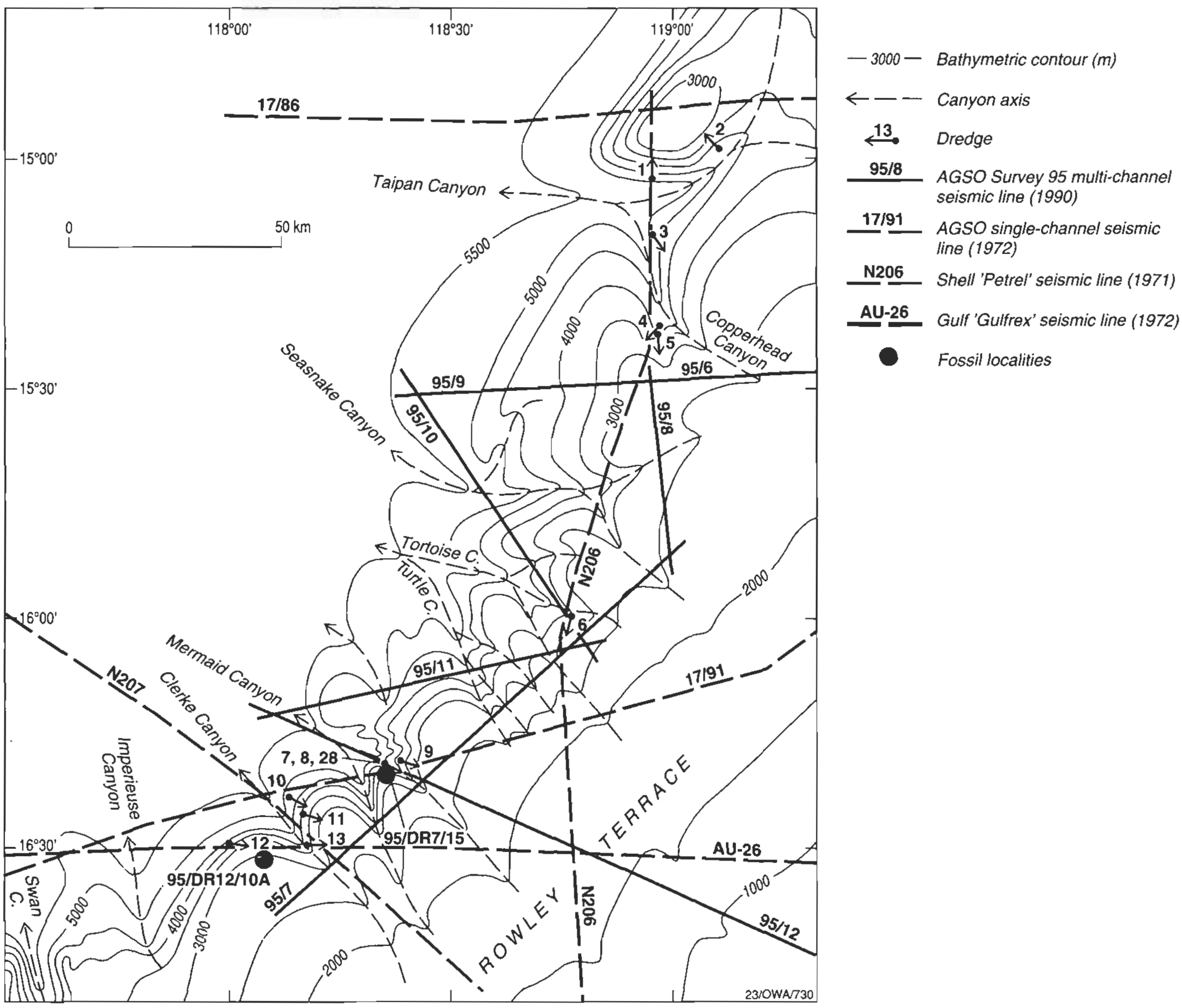

Figure 2. Detailed bathymetric survey of a portion of the Rowley Terrace showing locations of the samples described in this paper.

Pterastraea tenuis Reuss, 1865, p. 163, pl. 3, fig. 1.

Thamnastraea rectilamellosa, Frech, 1890, p. 60, text–fig. p. 61, pl. 16, figs. 1–15; Wurm, 1982, p. 218, pl. 34, fig. 5.

Stylina norica Frech, 1890, p. 33, pl. 11, figs. 1–1B.

Thamnasteria rectilamellosa var. *tibetana* Vinassa de Regny, 1932, pl. 18, figs. 7–8.

Thamnasteria rectilamellosa rectilamellosa Kristan-Tollman & Tollmann, 1964, p. 561, pl. 6, fig. 2; Kristan-Tollamnn et al., 1969, p. 16, pl. 3, figs. 1, 2.

Fungiastraea rectilamellosa pamirensis Melnikova, 1967, p. 24, pl. 2, fig. 1.

Thamnasteria rectilamellosa, Zankl, 1969, p. 36, fig. 34; Cuif, p. 151, text–fig. 27, pl. 17, fig. 1; Schäfer, 1979, p. 46, pl. 11, fig. 3; 1984, pl. 1, fig. 4.

Pamiroseris rectilamellosa, Gadźicki, 1974, pl. 18, figs. 1–4; Roniewicz, 1974, p. 114, pl. 8, fig. 3, pl. 10, figs. 1–3; Matzner, 1986, pl. 9, figs. 3, 4; Melnikova, 1986, p. 63, pl. 25, figs. 1, 2, pl. 26, fig. 1; Roniewicz, 1989, p. 111, pl. 34, figs. 3–5.

Pamiroseris rectilamellosa pamirensis, Melnikova, 1975, p. 130, pl. 30, fig. 7, pl. 33, fig. 1, pl. 34, fig. 1.

Thamnasteria rectilamellosa Kristan-Tollmann et al., 1980, p. 169, pl. 2, figs. 1–3; Stanley, 1979, pl. 1, fig. 12.

Thamnasteria (Thamnasteria) rectilamellosa delicata (Reuss), Yamagiwa & Taraz, 1981, p. 65, pl. , figs. 1–2.

Pamiroseris rectilamellosa rectilamellosa Fantini-Sestini & Motta, 1984, p. 353, pl. 29, fig. 1.

Material. Dredge sample 95/DR12/10A, Australian Commonwealth Collection CPC 31430.

Discussion. In all morphological features, the specimens agree well with Winkler's species as discussed, illustrated and revised by Roniewicz (1989). This distinctive species is widely distributed in Upper Triassic carbonate complexes of the Tethys. It tends to show notable morphologic variation over its geographic range of occurrences.

As far as known, this species belongs to the Norian to Rhaetian Stages of Europe and the lower Norian of western

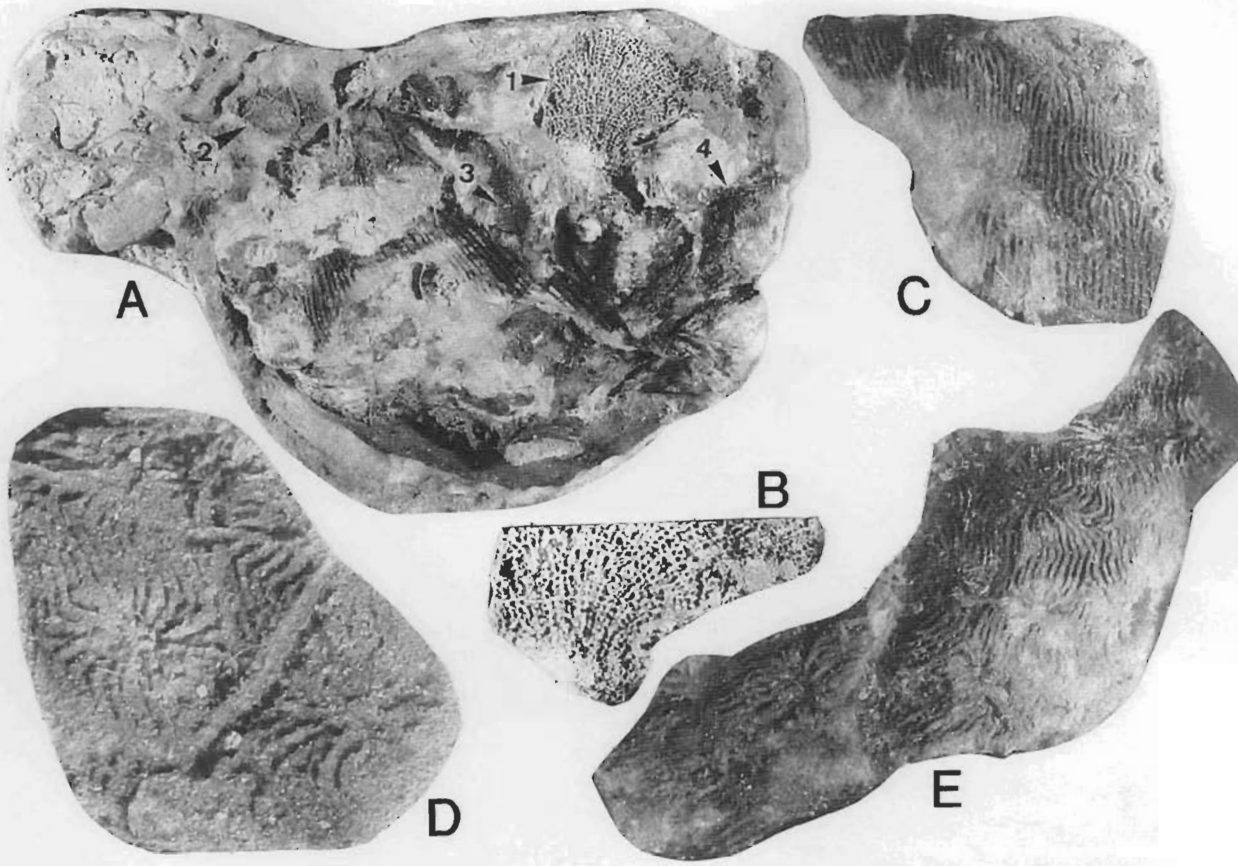

Figure 3. Fossils in dredged limestone sample 95/DR12/10A. A, one surface showing various fossils (arrow), X0.9; 1, *Spongiomorpha* sp., 2–4, *Pamiroseris rectilamellosa* (Winkler). B–E, details of the surface in 1A. B, leached coenosteum of *Spongiomorpha* sp. (CPC 31429), X2.7. C–E, *Pamiroseris rectilamellosa* (Winkler). C, showing several thamnasteroid corallites with confluent septa (CPC 31430a), X3. D, enlargement showing the distinctive thamnasteroid, straight to sigmoid septa and parietal columella, CPC 31430b, X5. E, CPC 31430c, illustrating the columella and confluent septa, X2.7.

North America. The possible "?Upper Carnian" occurrence mentioned by Roniewicz (1989, p. 113) is probably in error and at best questionable.

Distribution. Rhaetian of Northern Calcareous Alps; Norian–Rhaetian of the Caucasus Mountains, Iran and central Asia; Norian of the Pilot Mountains, Nevada, U.S.A.; Sonora, Mexico; Northwestern Australian Shelf.

Suborder CARYOPHYLLIINA Vaughan and Wells, 1943

Superfamily REIMANIPHYLLIOIDEA Melnikova, 1975

Family REIMANIPHYLLIIDAE Melnikova, 1975

Genus RETIOPHYLLIA Cuif, 1967

Retiophyllia tellae (Stoppani)

Fig. 4, A–F

Rhabdophyllia sellae Stoppani, 1862, p. 107, pl. 25.

Thecosmilia sellae Stoppani (Frech), 1890, p. 17, pl. 4, fig. 12.

Parathecosmilia sellae Stoppani (Roniewicz) 1974, p. 110, figs. 9–10, pl. 6, fig. 1–3, pl. 7, figs. 1–2.

Material. Australian Commonwealth Collection No. CPC 31617. Dredge sample 95/DR7/15.

Description. Large phaceloid colony 10 cm wide and partially recrystallized. Branches bifurcating at acute angles with commonly occurring anastomosing septal connections between adjacent corallites. Corallites nearly circular to slightly subcircular with diameters (in mm) ranging from 4.5–8.0 mm as given below:

Length	*Width*
4.5	5.0
6.0	5.0
6.0	6.0
7.0	7.0
8.0	7.0

Septa granular, numbering from 48–60. Recrystallization has obscured the septal microstructure, but first-order septa are thick and slightly fusiform in shape, reaching nearly to the corallite center. Subsequent cycles are much thinner and shorter, and some are zig-zag. Epitheca present, but very thin and not preserved on all corallites. Some curved dissepiments are present. In longitudinal section, endotheca composed of abundant subhorizontal and concave dissepiments, irregularly spaced at densities of 2–3 per mm of distance.

Discussion. The most distinctive aspect of this coral is

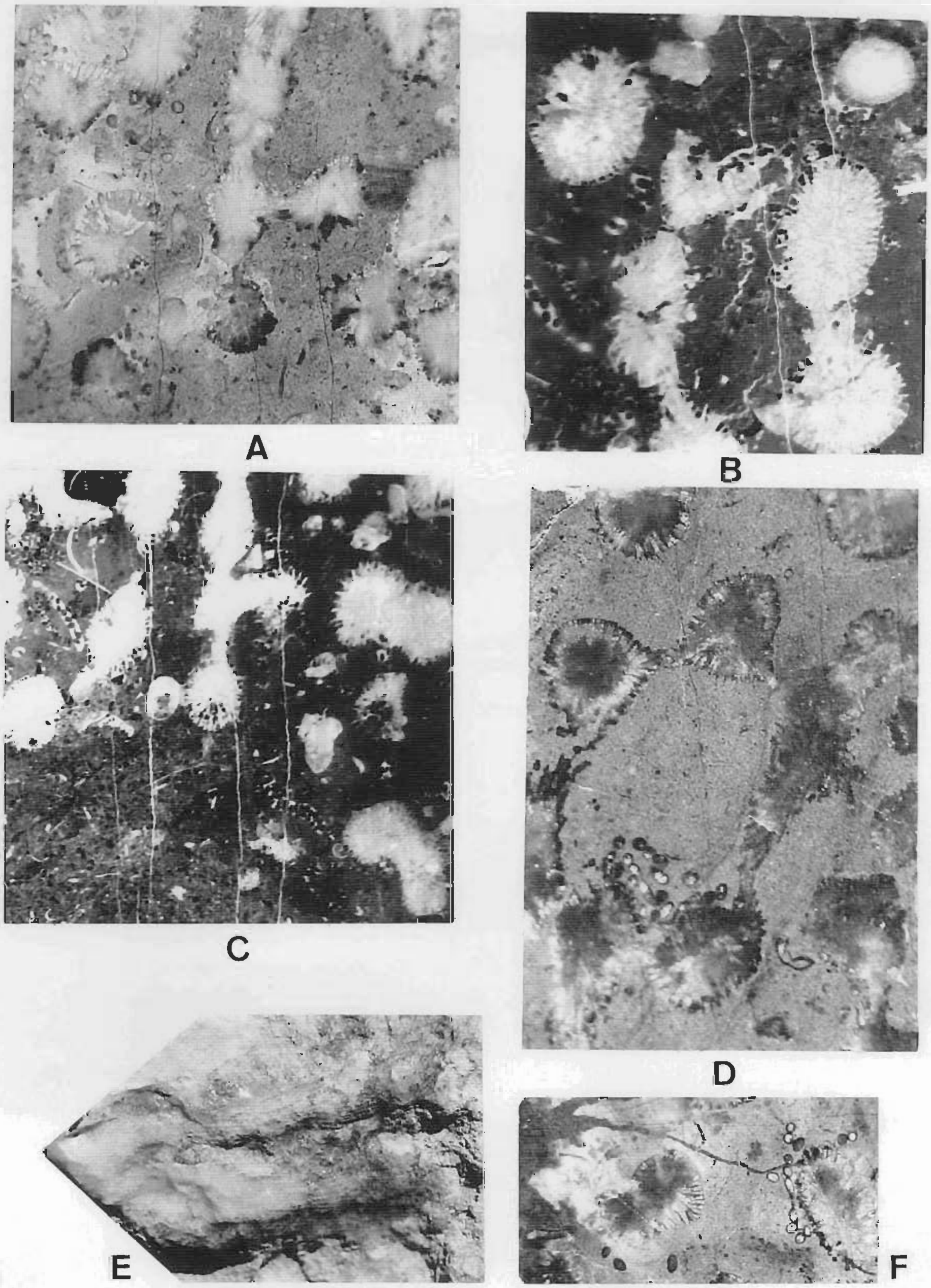

Figure 4, **A-F** ***Retiophyllia tellae*** **(Stoppani) Specimen no. CPC 31617). A. Polished surface view of recrystallized phaceloid corallites, some showing septal details. Note the numerous anastomosing connections between corallites x2. B. Thin section showing details of several fused corallites showing trabecular linkages, x 3. C. Thin section showing the frequent bifurcation and connections between corallites, x 1.6. D. Details of a polished surface showing six interconnected corallites, some incrusted by serpulid tubes (lower left). Interstitial, pellet-rich grainstone fills between corallites. x 2.5. E. Surface view of a branched individual showing acute angle of bifurcation, x1.5. F. Details of two corallites on polished surface, one on left incrusted by serpulid tubes, x2.**

the anastomosing peripheral regions where subadjacent corallites are fused to one another via trabecular linkages. Based on this and other features, it compares closely with *Retiophyllia sellae* (Stoppani) from the Lombardy Alps. Examples of this species were illustrated from Poland by Roniewicz (1974). She assigned them to *Parathecosmilia*, a genus now regarded as *Retiophyllia* Cuif (Roniewicz, 1989). Compared to the Australian specimen, Stoppani's type is relatively smaller in corallite diameters (4.0–6.0 mm). On the other hand, *Parathecosmilia sellae* from Poland ranges from 4.0–5.5 mm in diameter.

In addition to *Parathecosmilia sellae*, Roniewicz (1974, p. 111) also described from the same locality *Parathecosmilia* sp. This coral is almost identical with *Retiophyllia sellae*, differing from it only in the diameter of the corallites (6.0–8.0 mm). Since the Australian material is intermediate in size between both of these taxa, I assign it as well as Roniewicz's *Parathecosmilia* to Stoppani's species, attributing them both to the genus *Retiophyllia*.

Distribution. Upper Triassic (Norian–Rhaetian) Northern Limestone Alps, Austria; Rhaetian, Lombardy, Italy; Tatra Mountains (Rhaetian) Poland; the Northwestern Australian Shelf.

Conclusions

Discoveries resulting from ODP Leg 122, of Upper Triassic corals and reef facies developed on Wombat Plateau of northwestern Australia (von Rad et al., 1989), confirmed the expected continuation of the southern Gondwana reef trend postulated by Stanley (1988) and reproduced below in Figure 5. The additional Triassic spongiomorph and corals illustrated in this paper are congruent in age and palaeoecology with those reported from reef-building communities on the Wombat Plateau at ODP site 764 (Sarti et al., 1992). They are taken to indicate an eastward extension of the reef trend. The spongiomorph and corals, like other Upper Triassic fossils from the offshore Canning Basin, show close similarities with distant counterparts in the Alps of central Europe.

Acknowledgments

I thank Jack Grant-Mackie, University of Auckland, for encouraging me to undertake the study which resulted in this report. Neville Exon kindly supplied sample 95/DR7/15 for study. Photography in Figure 3 was carried out by Ms Louise Cotterall, University of Auckland. R.S. Nicoll and Barry Webby reviewed the manuscript and offered helpful comments.

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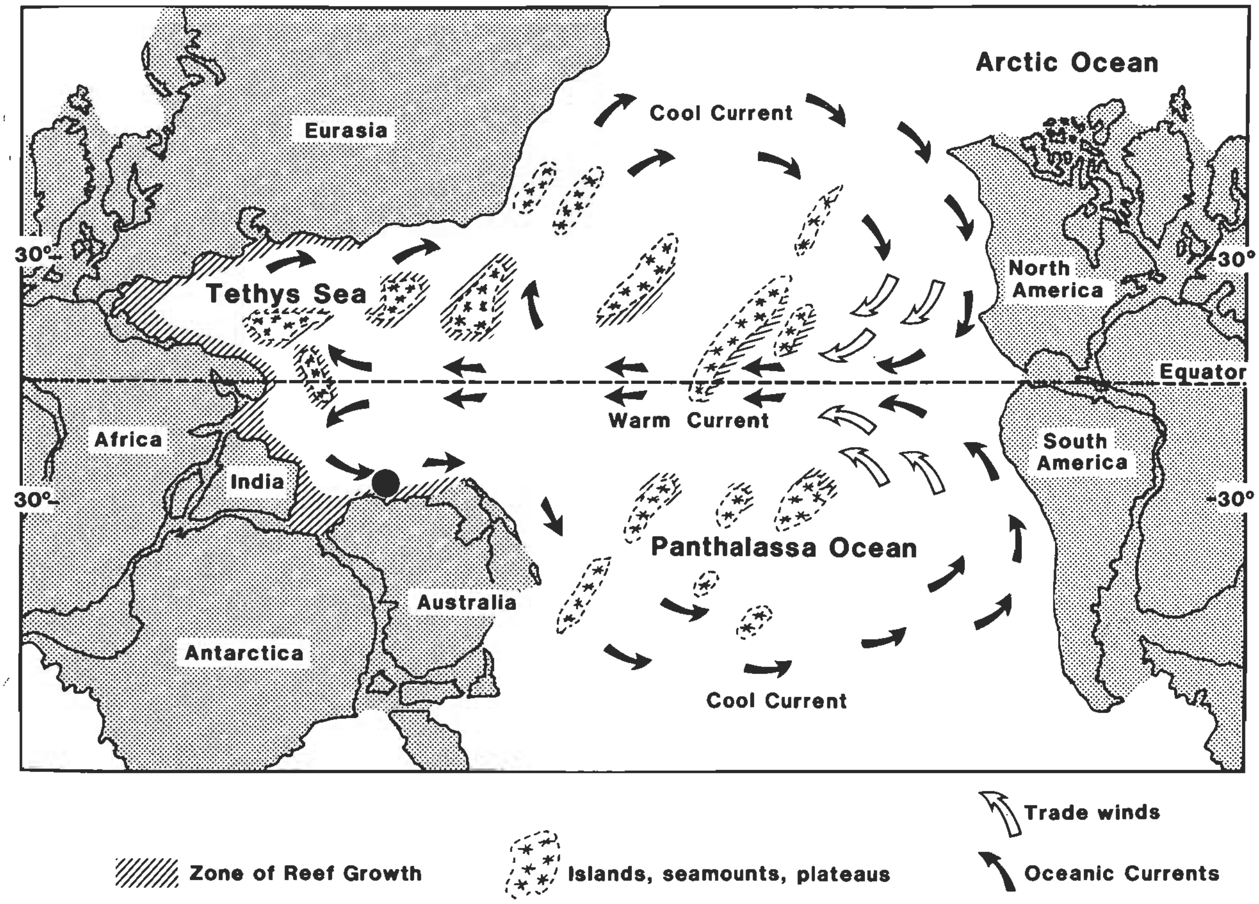

Figure 5. Palaeogeographic reconstruction of the Triassic world showing the Tethys and the postulated regions of reef development. Offshore Canning Basin reef site is indicated with dot. Palaeogeographic locations of islands of the ancient Pacific now making up the circum-Pacific terrane collage are not well established. Modified from Stanley (1988).

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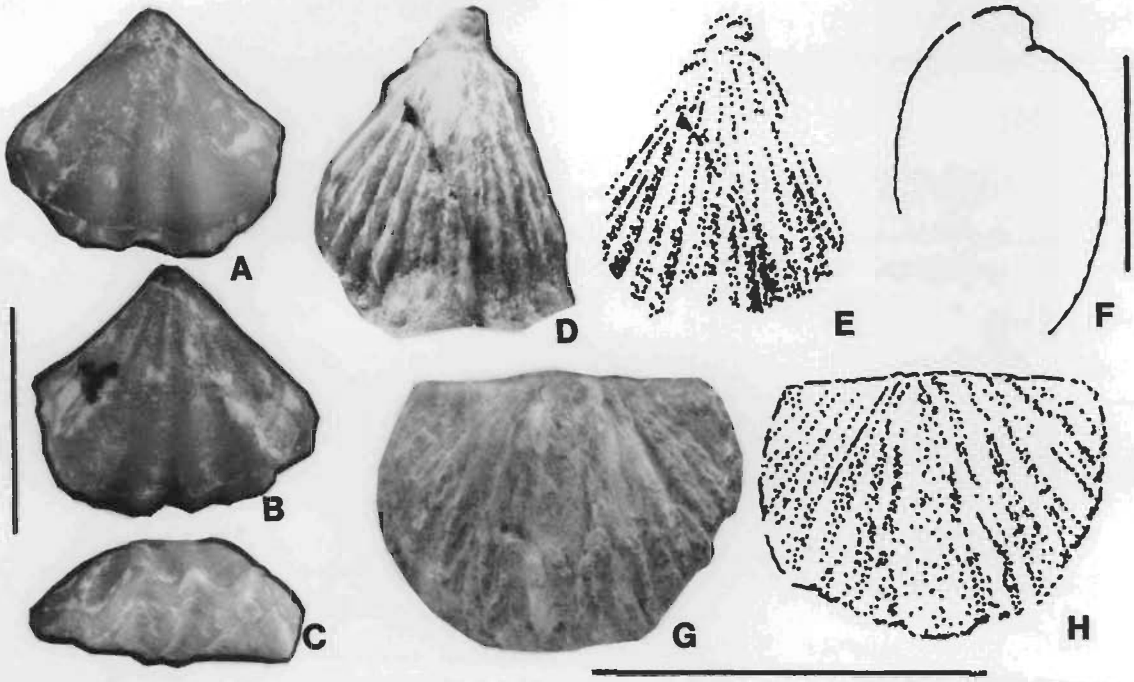

Figure 1. A, B, C: cf. *Trigonirhynchella* sp. (CPC 31600), dredge station 95/DR07/13A; Late Triassic; A: ventral view, B: dorsal view, C: anterior view, scale bar: 1.0 cm.
D,E,F: *Misolia* sp. (CPC 31599), dredge station 95/DR07/13A; Late Triassic; D: partially excorticated specimen viewed from the dorsal side, E: drawing of the same view, F: profile of both valves drawn from a photograph. Scale bar: 1.0 cm.
G,H: cf. *Zugmayerella* sp. (CPC 31601), dredge station 95/DR07/13B; Late Triassic; G: shelly specimen, dorsal exterior. H: drawing of the same view. Scale bar: 1.0 cm.

Valve convex, costate with about 9 strong costae. Shell finely punctate.

Remarks. Dorsal valves of the Family Laballidae and some other spiriferinaceans have generalised form, and association with *Zugmayerella* is only one of several possibilities. The family is especially characteristic of Triassic faunas (Campbell, 1991).

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AGSO Journal of Australian Geology & Geophysics, 15/1, 135–136

Late Triassic brachiopods from a dredge haul on the slope below Rowley Terrace, northwest Australia

J.D. Campbell[1]

Late Triassic brachiopods (*Misolia* sp., cf. *Trigonirhynchella* sp., and cf. *Zugmayerella* sp.) are described from mudstone dredged below the Rowley Terrace, northwest Australia.

Introduction

Two hand-specimen-sized samples from a dredging haul (BMR Cruise 95, locality DR07/13; 118°23'E, 016°18'S) were examined in this study. Each consists of fine-grained calcareous buff to grey mudstone in which brachiopods are the only macrofossils. The fossils are undistorted and shelly, occurring dispersed through the rock — they indicate a Late Triassic age for the sediment. The specimens have been deposited in the Commonwealth Palaeontological Collection, registered numbers CPC 31599–31601.

Systematic palaeontology

Order Rhynchonellida Kuhn, 1949

Superfamily Rhynchonellacea Gray, 1848

Family Wellerellidae Likharev, 1956

Genus *Trigonirhynchella* Dagys, 1963

Type species (original definition). *Trigonirhynchia trigona* Dagys, 1961. Late Triassic (Norian Stage), Northwest Caucasus.

cf. *Trigonirhynchella* sp. Fig. 1 A, B, C

Material. One double-valved shelly specimen CPC 31600 from sample 95/DR07/13A.

Description. Shell of modest size (L, 10.5; W, 12.5; T, 5.5mm), non-strophic, triangular with greatest width well forward. Apical angle about 96°. Beak erect, orthoclinal. Biconvex with costation arising early in ontogeny. About 7 costae of which 3 occur on the flat plication which deflects the anterior commissure. Shell impunctate. Foramen of moderate size, dethyrium covered.

Remarks. The specimen may not be fully mature as there is a single growth pause. Internal morphology cannot be observed and so the placing near to *Trigonirhynchella* is tentative. It should be pointed out that Moiseev's genus *Salgirella* (type species, *Rhynchonella albertii* Oppel, fide Ager, 1965) also has a triangular outline. Its plication may, however, be weaker.

Trigonirhynchella occurs in Late Triassic (Norian) sediments in Northwest Caucasus (Dagys, 1961; 1963).

Order Spiriferida Waagen, 1883

Suborder Retziidina Boucot, Johnson and Stanton, 1964

Superfamily Athyrisinacea Grabau, 1931

Family Athyrisinidae Grabau, 1931

Genus *Misolia* von Seidlitz, 1913

Type species (original definition). *Misolia misolica* von Seidlitz, 1913 Late Triassic, Misool, Indonesia.

Misolia sp. Fig. 1 D, E, F

Material. One double-valved shelly specimen CPC 31599, and several shell fragments from sample 95/DR07/13A, including an incomplete ventral valve from sample 95/DR07/13B, associated with CPC 31601.

Description. Modest-sized (L, 16.5; W, 13.00; T, 9.5mm) non-strophic, biconvex globular shell with prominent ventral beak. Foramen large, orthoclinal. Costae (18–20) strong, almost all persistent. Anterior commissure little deflected. Shell impunctate.

Remarks. Placing in *Misolia* relies on the external architecture of the shell. Internal morphology is not known in spite of an irregular longitudinal section being available (brachial skeletal parts are not preserved). External morphology suggests a placing near to *M. misolica* von Seidlitz and to *M. pinajae* Krumbeck. *M. lenticulata* Jefferies is less globular than the Canning Basin form.

Misolia is known from Indonesia (Misool, Seram, Sulawesi and Timor) where it may be an abundant member of Late Triassic invertebrate assemblages, and from India (Spiti) and Arabia (Oman) (Hudson & Jefferies; 1961). Within the Misool sequence it occurs with *Cochloceras* and *Rhabdoceras* variously cited as Rhaetian and Norian–Rhaetian indicators (J.A. Grant-Mackie & F. Hasibuan, pers. comm. 1993). It is a common fossil in the Bogal Formation, a calcareous deposit of the continental shelf (Hasibuan, 1990).

Superfamily Spiriferinacea Davidson, 1884

Family Laballidae Dagys, 1962

Genus *Zugmayerella* Dagys, 1963

Type species (original definition). *Spiriferina koessenensis* Zugmayer 1880. Rhaetian Stage, Austria.

cf. *Zugmayerella* sp. Fig. 1, G.H.

Material. One dorsal valve, shelly. CPC 31601 from sample 95/DR07/13B.

Description. Small shell (L, 6.05; W, 9.0 mm), strophic with greatest width slightly forward of the hinge-line.

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The nature of the Wallaby (Cuvier) Plateau and other igneous provinces of the west Australian margin

J.B. Colwell[1], P.A. Symonds[1] & A.J. Crawford[2]

Seismic reflection data and dredged rocks confirm that the Wallaby (Cuvier) Plateau off western Australia is underlain by a volcanic basement consisting of complex packages of dipping reflectors similar to those of other large, oceanic, igneous provinces (LIPs) of the eastern Indian Ocean, such as the Kerguelen Plateau. Wedges of seaward-dipping reflectors underlie the Wallaby Saddle, which separates the plateau from the adjacent upper continental slope. These are interpreted to be interbedded subaerial lava flows and volcaniclastics similar to those obtained by ODP drilling of like features beneath the margins of the North Atlantic Ocean. Volcanic rocks dredged from the Wallaby Plateau are altered transitional tholeiitic basalts with immobile element contents and ratios similar to basalts from the Naturaliste Plateau, eastern Broken Ridge, and the Bunbury Basalt of southern western Australia. Basalts from all of these regions show compositional features indicative of the involvement of sub-continental lithospheric mantle in their petrogenesis, and appear to have been erupted between 120 and 100 Ma as a result of an anomalous post-breakup heating event. A consistent explanation for the origin of the Wallaby Plateau, and similar large, post-breakup volcanic features (LIPs) of the eastern Indian Ocean, is that they are the product of a single, large hotspot (Kerguelen hotspot or mantle plume). However, for the Wallaby Plateau and its associated Sonne and Sonja Ridges there is another possible origin, that they are the products of convective partial melting following a ridge crest jump in the Cuvier Abyssal Plain. The distribution of other volcanic features of the western Australian margin can be readily explained by magmatism associated with dynamic rift processes rather than proximity to a plume.

Introduction

The Wallaby (Cuvier) Plateau[3] is one of several marginal plateaux that lie off the continental margin of western Australia (Fig. 1). It is separated from the western Australian continental shelf by the Wallaby Saddle (Fig. 2). The plateau ranges in water depth from approximately 2100 m at its crest on its northern part to about 4000 m at the base of the bounding slopes (Fig. 3), and has an areal extent within the 4000 m isobath of ~70 000 km^2.

The origin of the feature has been the subject of debate for many years. Prior to 1978, the plateau was regarded as a thinned continental fragment, similar to other marginal plateaux (e.g. Exmouth and Scott Plateaux) off the west Australian coast (Heezen & Tharp, 1965, 1966; Laughton et al., 1970; Symonds & Cameron, 1977). Subsequently, Veevers & Cotterill (1978) suggested that it is a thick accumulation of oceanic volcanics ('epilith') formed on oceanic crust after the start of spreading in the Cuvier Abyssal Plain area in the Early Cretaceous (Larson et al., 1979; Fullerton et al., 1989). The oceanic hypothesis for the plateau's origin was supported by dredging of a 1000 m thick sequence by the R/V *Sonne* on its eastern and southern flanks in 1979, which recovered lavas and volcaniclastic sediments apparently from a layered basement sequence beneath the main Neocomian unconformity (von Stackelberg et al., 1980). However, as noted by von Stackelberg et al. (op. cit.), volcanics from the plateau margins may simply represent marginal volcanism rather than reflecting the true nature (continental or oceanic) of the central part of the feature.

In this paper, we present previously unpublished seismic reflection profiles, as well as petrological–geochemical analyses of samples collected in 1990 during an AGSO program focussed on the plateau's central western flank. This latter work was aimed at sampling dipping basement reflectors in an area away from the plateau's 'volcanic' margins (Colwell et al., 1990). The results obtained (including the recovery of basalts, and basaltic conglomerates and sandstones) support a volcanic origin for the core of the feature and provide, for the first time, evidence of a genetic link to other large, oceanic, igneous provinces (LIPs; see Coffin & Eldholm, 1991) of the eastern Indian Ocean.

Regional setting

The physiography of the Wallaby Plateau was first systematically described by Falvey & Veevers (1974). The plateau consists of two main bathymetric highs. A large southeastern high, which was called the Wallaby Plateau by Veevers et al. (1985), forms the main part of the feature and contains two closures of the 2500 m isobath (Fig. 3). The largest closure occurs in the north and shallows to about 2100 m water depth. A smaller bathymetric high occurs in the northwest and was named the Quokka Rise by Veevers et al. (1985). The plateau is bounded in the north by the Cuvier Abyssal Plain (Basin), and to the south by the Perth Abyssal Plain. Its southern margin is formed by the steep, 2000 m high, NW-trending scarp of the Wallaby–Zenith Fracture Zone (Fig. 3). In the west, the Wallaby Plateau is separated from the Zenith Seamount (Plateau) (Fig. 1) by a 100–150 km wide, NNE-trending bathymetric trough. In the east, it is separated from the upper continental slope of the Australian margin (the Carnarvon Terrace) by the NNE-trending Wallaby Saddle (Figs 2 and 3).

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3 Although the term Wallaby Plateau is widely used in the Australian geoscience literature, on many bathymetric charts and on the AAPG tectonic map series the feature is referred to as the Cuvier Plateau. To add further to the confusion, on the AAPG maps the name Wallaby Plateau is applied to a feature farther to the west referred to by other workers (e.g. Tomoda et al., 1968; Veevers & Cotterill, 1978) as the Zenith Seamount and later by Veevers et al., (1985) as the Zenith Plateau. Part of this confusion stems from the original application of the term Wallaby Plateau(s) by Heezen & Tharp (1965, 1966) to a composite feature here taken to consist of the Zenith Plateau in the west and the Wallaby Plateau (the subject of this paper) in the east. Veevers et al. (1985) further subdivided the Wallaby Plateau into two features — a northwestern high that they called the Quokka Rise (Fig. 3) and a larger southwestern high, which they defined as the Wallaby Plateau. In this paper, we use the term Wallaby Plateau in its more general sense as referring to both of these features, unless otherwise specified.

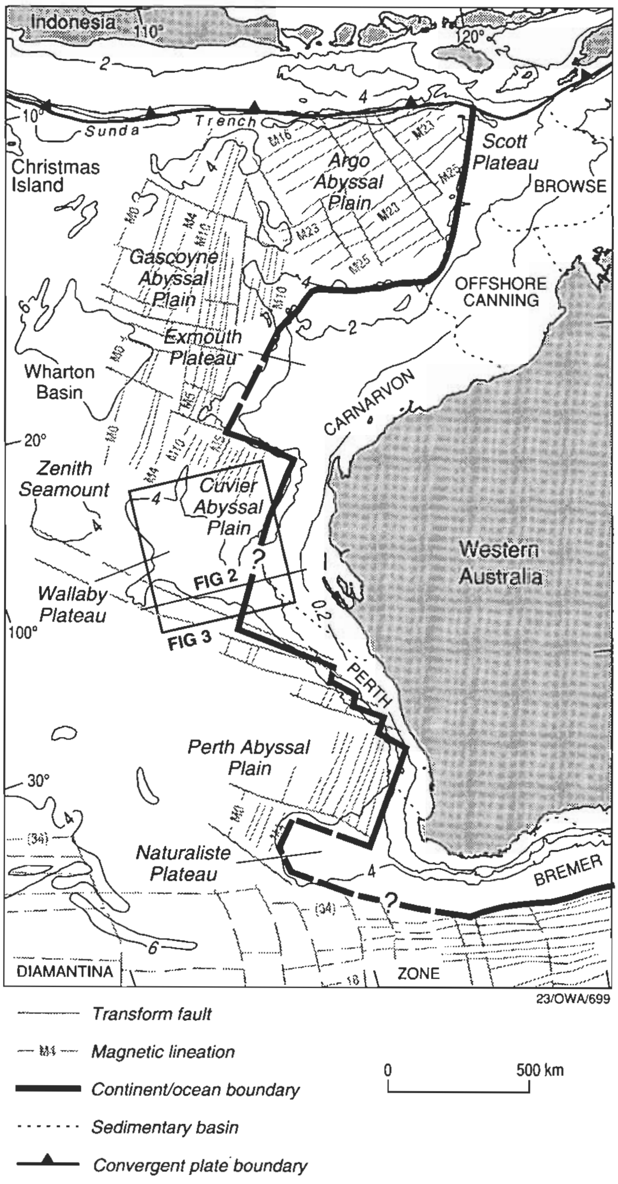


Figure 1. General location of the Wallaby Plateau and other features off the western Australian margin (after Falvey et al., 1990).

Two ridges trend NNE into the Cuvier Abyssal Plain from the northern margin of the Wallaby Plateau (Figs 1 and 3). The eastern ridge — the Sonne Ridge — is composed of several highs, which shallow to about 3500 m, and is generally considered to be an abandoned spreading ridge formed at about magnetic anomaly M4–5 time (Veevers et al., 1985; 126 Ma using reversal time scale of Kent & Gradstein, 1985). The western ridge — the Sonja Ridge — is composed of several highs which shallow to 2700 m, and is associated with anomaly M8–9 near the western edge of the early seafloor spreading episode (M10–M4) that formed the eastern Cuvier Abyssal Plain.

The stratigraphy and depositional history of the Wallaby Plateau/Cuvier Abyssal Plain area are poorly known and are based on one Deep Sea Drilling Project (DSDP) hole (Site 263), one petroleum exploration well just to the south on the adjacent continental shelf (Pendock ID-1), and scattered dredge and core samples (Fig. 3). Site 263 (Veevers et al., 1974) bottomed 100–200 m above the oceanic basement of the Cuvier Abyssal Plain in possibly allochthonous Early Cretaceous sediment (Veevers & Cotterill, 1978). Pendock ID-1 intersected mid-Cretaceous sedimentary rocks unconformably overlying Carboniferous to Devonian sedimentary rocks (Genoa Oil NL, 1970). Dredging undertaken by the German Bundesanstalt für Geowissenschaften und Rohstoffe (BGR) in 1979 in three areas on the southern and eastern flanks of the plateau (Fig. 3) recovered volcanic rocks and volcaniclastic sediments including variably altered tholeiitic basalts, tuffs, basalt breccias, and volcaniclastic sandstones and breccias (von Stackelberg et al., 1980; Table 1). A minimum Late Cretaceous K/Ar age of 89 Ma was obtained on a sample of altered basalt from site 170KD on the southern margin of the plateau. Sampling on the northern Sonne Ridge yielded basalt, volcanic breccia and tuffs (von Stackelberg et al., op. cit.).

Seismic reflection data in the region are sparse (Fig. 3). Two major unconformities have been identified throughout the region — one, 'C', corresponding to the base of the Santonian Toolonga Calcilutite, and the other, 'D', thought initially to be an intra-Neocomian unconformity separating Aptian–Cenomanian Winning Group sediments from Palaeozoic rocks (Symonds & Cameron, 1977; Fig. 2). The 'D' unconformity is now known to be the top of volcanic basement (von Stackelberg et al., 1980; this paper).

A NE-trending magnetic anomaly pattern encompassing anomalies M10–M5 has been identified in the Cuvier and adjacent Gascoyne Abyssal Plains (Larson, 1977; Larson et al., 1979; Johnson et al., 1980; Veevers et al., 1985; Fullerton et al., 1989). Breakup in the region began at M10 (about 130 Ma; Fullerton et al., 1989) or perhaps M11 (about 132 Ma; Veevers & Li, 1991), in the Valanginian, and a little later at anomaly M9 (about 129 Ma) in the Perth Abyssal Plain to the south (Veevers et al., 1991). Breakup is thought to be contemporaneous across the Cape Range Fracture Zone that separates the Cuvier and Gascoyne Abyssal Plains, and seafloor spreading is interpreted to be continuous to anomaly M4–5 (about 127 Ma — Hauterivian), when a westward jump in the Cuvier and northern Gascoyne Abyssal Plains left behind abandoned spreading ridges and associated symmetric anomaly patterns in these areas (Fig. 1).

New insights into the extensional and magmatic development of the region have arisen from the interpretation of two-ship seismic reflection and refraction data across the western and southern margins of the Exmouth Plateau, and the Cuvier margin of the Carnarvon Terrace. Mutter et al. (1989) examined the western rifted margin of the Exmouth Plateau and suggested that rifting occurred over a prolonged period and was initially dominated by detachment faulting, that passed laterally and temporally into an outer 100 km wide zone of pure shear deformation characterised by high-angle normal faults and magmatic underplating. Lorenzo et al. (1991) presented results for the Exmouth Plateau's southern transform margin indicating that the final rupture was associated with fault-block rotation and mafic intrusion. Significant underplating of the margin occurred only during the drift phase, and extended laterally to form a thickened oceanic layer 3 beneath the northern margin of the Cuvier Abyssal Plain. Hopper et al. (1992) interpreted a seaward-dipping reflector sequence beneath the lower slope of the Cuvier margin on a single deep seismic line, suggesting that this margin is a so-called "volcanic rifted margin". Veevers

TABLE 1. Details of 1979 *Sonne* SO-08 Wallaby Plateau Dredge Hauls (after von Stackelberg et al., 1980).

Dredge	*Position* S	*Position* E	*Water Depth* (m)	*Recovery* (kg)	*Lithology*
159KD	24°23.5′ 24°22.8′	109°43.1′ 109°41.8′	4470–4130	100	C. gr. volc. sst. with clay matrix and volc. & rare glauc. phosph. clasts
161KD	24°24.0′ 24°23.9′	109°45.0′ 109°44.3′	4470–4230	2	C. gr. volc. sst. with clay matrix, weathered and Mn crusted
165KD	24°23.7′ 24°23.7′	109°42.4′ 109°44.0′	4415–4240	5	Pebbly volc. sst., clasts mainly alt. glass; Mn polynodules
167KD	25°39.0′ 25°35.4′	108°36.5′ 108°35.1′	5340–4750	2	Pink-grey br. ?volc. clayst.; grey f.-c. vitric-lithic volc. breccia & qtz-rich silty tuff with glassy matrix; silicified volc. breccia; pale br. v.f. basalt or tuff; Mn polynodules
168KD	25°34.9′ 25°33.4′	108°34.3′ 108°35.0′	5100–4050	200	Alt. ± amyg. basalt; volc. siltst.; fissile f. tuff; volc. clayst.
170KD	25°31.6′ 25°31.0′	108°31.9′ 108°32.4′	4620–3970	120	Tholeitic basalt; basaltic breccia; volcanogenic mudst.; v.c. poorly sorted volc. sst.; volc. breccia; Mn polynodule
173KD	25°57.8′ 25°55.3′	109°05.8′ 109°04.4′	4980–3885	0.2	Reddish br. ferrug. (?rad.) vitreous qtz chert; highly altered tuff-volc. clayst.

et al. (1985) previously inferred a volcanic, 'epilithic' origin for this margin on the basis of the unusually shallow water depth of the magnetically defined continent/ocean boundary. Hopper et al. (op. cit.) suggested that rapid rifting in the Cuvier area resulted in the limited development of extensional structures and a larger volume of magma at the initiation of sea-floor spreading, and thus the initial emplacement of thicker oceanic crust.

The marked difference in the style of margin development between the Exmouth Plateau and Cuvier margins reflects differences in the rifting and magmatic history, and has important implications for tectonic evolution of the Exmouth–Cuvier–Wallaby region.

Seismic character of the Wallaby Plateau and Saddle

The one single-channel seismic line recorded by *Glomar Challenger* across the plateau in 1972 (Fig. 3) showed a number of thin, transparent-to-weakly-bedded sequences overlying 'acoustic basement' (Veevers & Heirtzler, 1974). Using the higher-quality data recorded by the AGSO Continental Margins Survey and Shell Petrel's multichannel-systems, Symonds & Cameron (1977) described variations in seismic character beneath the main Neocomian 'breakup' (D) unconformity, i.e. within the previously defined 'acoustic basement', which they ascribed to folding and faulting of Palaeozoic rocks, similar to the situation underlying the adjacent Carnarvon Terrace (Fig. 2). They also described a regional, Late Cretaceous 'C' unconformity in the overlying sedimentary sequence.

A re-examination of the seismic data reveals considerable variation in the seismic character beneath the main 'D' unconformity. Beneath much of the southern and central Wallaby Plateau it consists of sets of dipping, commonly diverging reflectors (see Figs 3–6). Individual reflectors are commonly over ten kilometres in length. In places, the dipping reflector packages are clearly offset by faulting (e.g. Figs 4A and 5). In other areas, the packages appear to splay out in different directions from an area of acoustic basement, or dip towards the centre of a 'basin' (Figs 3 and 4). These features are very similar to dipping reflectors seen within the basement complexes of large, oceanic, hotspot-related volcanic buildups such as the Kerguelen and Ontong Java Plateaux (Rotstein et al., 1990; Coffin et al., 1990; Hagen et al., 1993; Fig. 7). They also appear similar to features on Iceland, where the complex dipping sequences have been related to frequent shifts (jumps)

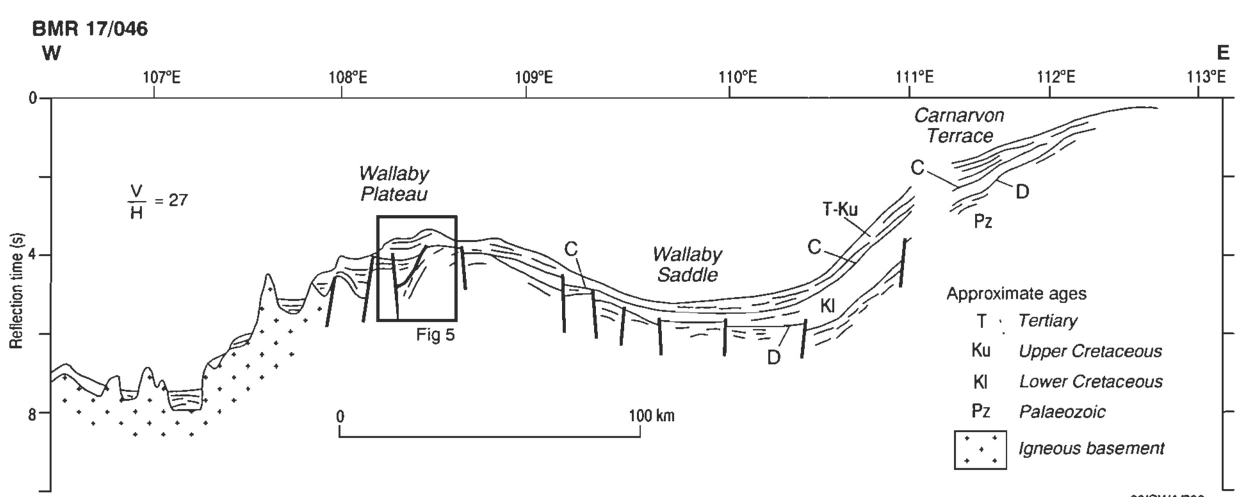


Figure 2. East-west section from the Carnarvon Terrace across the Wallaby Saddle and southern Wallaby Plateau to the Perth Abyssal Plain (after Symonds & Cameron, 1977). Location of the section shown in Figure 1.

of the active volcanic zone (Helgason, 1984). Drilling of basement beneath the Kerguelen and Ontong Java Plateaux as part of the Ocean Drilling Program (ODP) has recovered interbedded basalt flows and volcaniclastic sediments (Barron et al., 1989; Schlich et al., 1989; Kroenke et al., 1991). In the case of the Kerguelen Plateau, individual

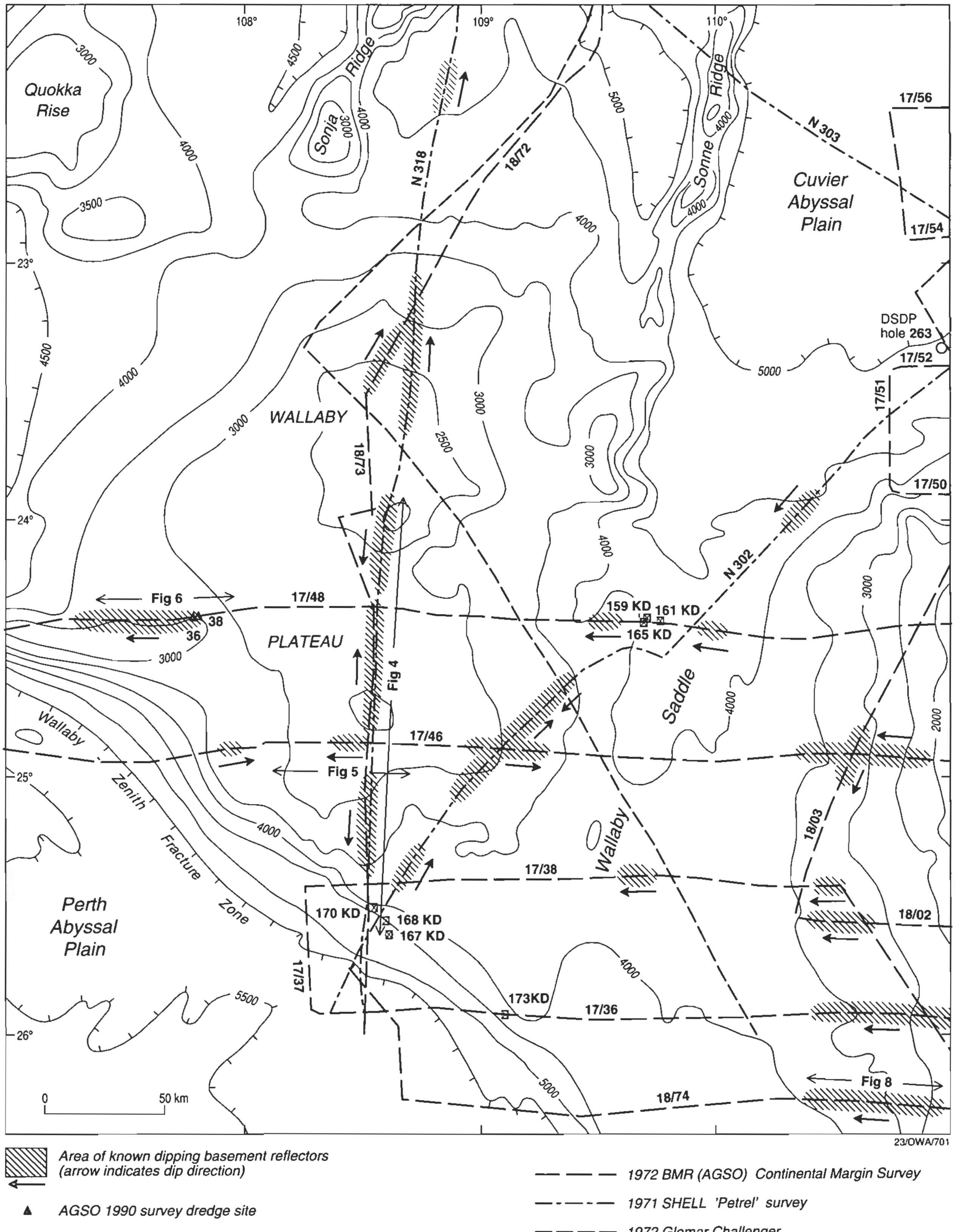


Figure 3. Bathymetric map of the Wallaby Plateau (after Jongsma et al., 1990, and GEBCO data) showing the location of seismic lines, dredge sites, DSDP drill Site 263, and areas of dipping, basement reflectors. Location of the map shown in Figure 1.

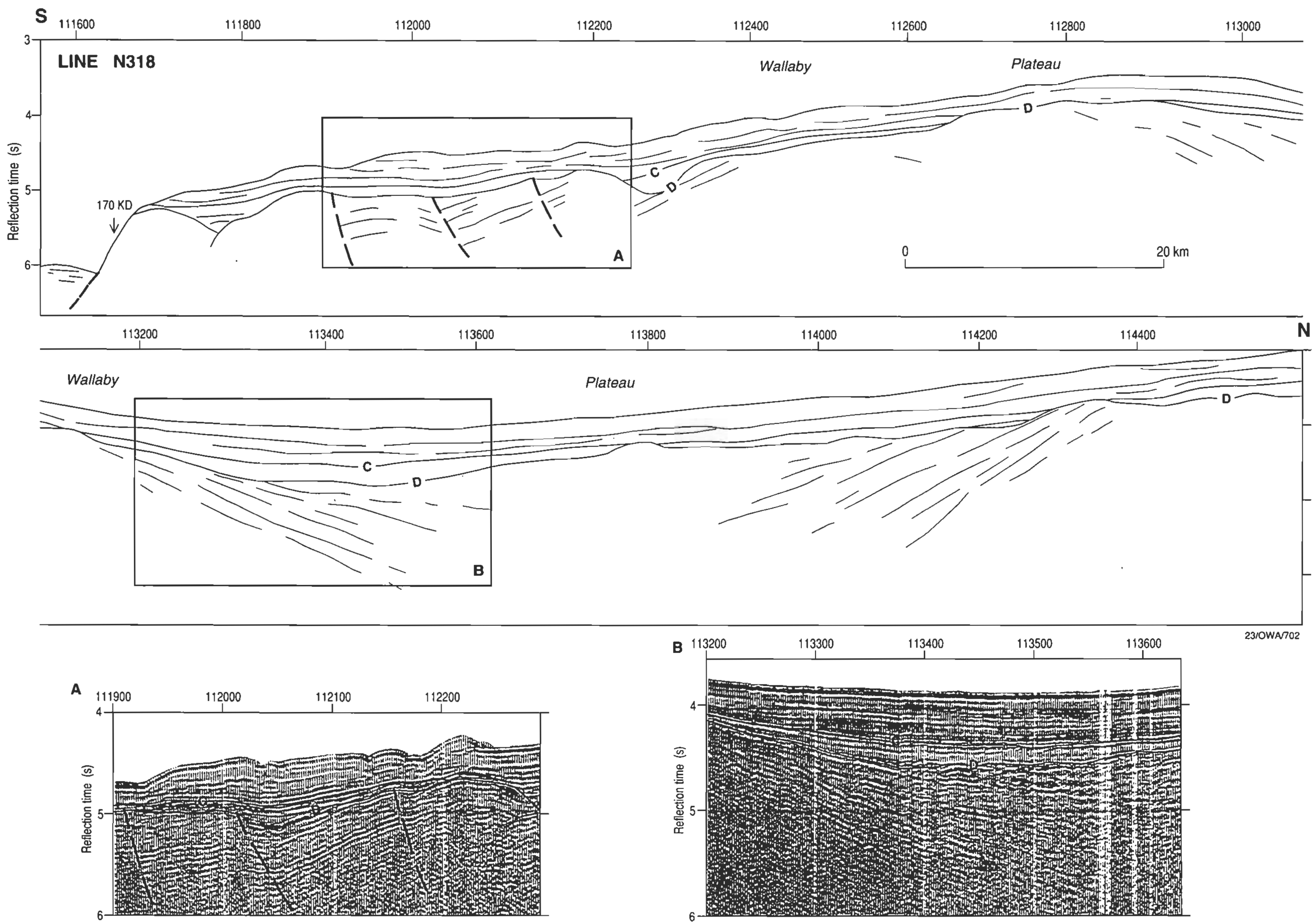


Figure 4. Line drawing of part of Shell *Petrel* Line N318 across the Wallaby Plateau showing dipping reflectors below the major, top basement, D unconformity. Note the switch in dip directions and the faulting affecting the dipping events. Inserts show parts of the single-channel seismic records. Location of the figure shown in Figure 3.

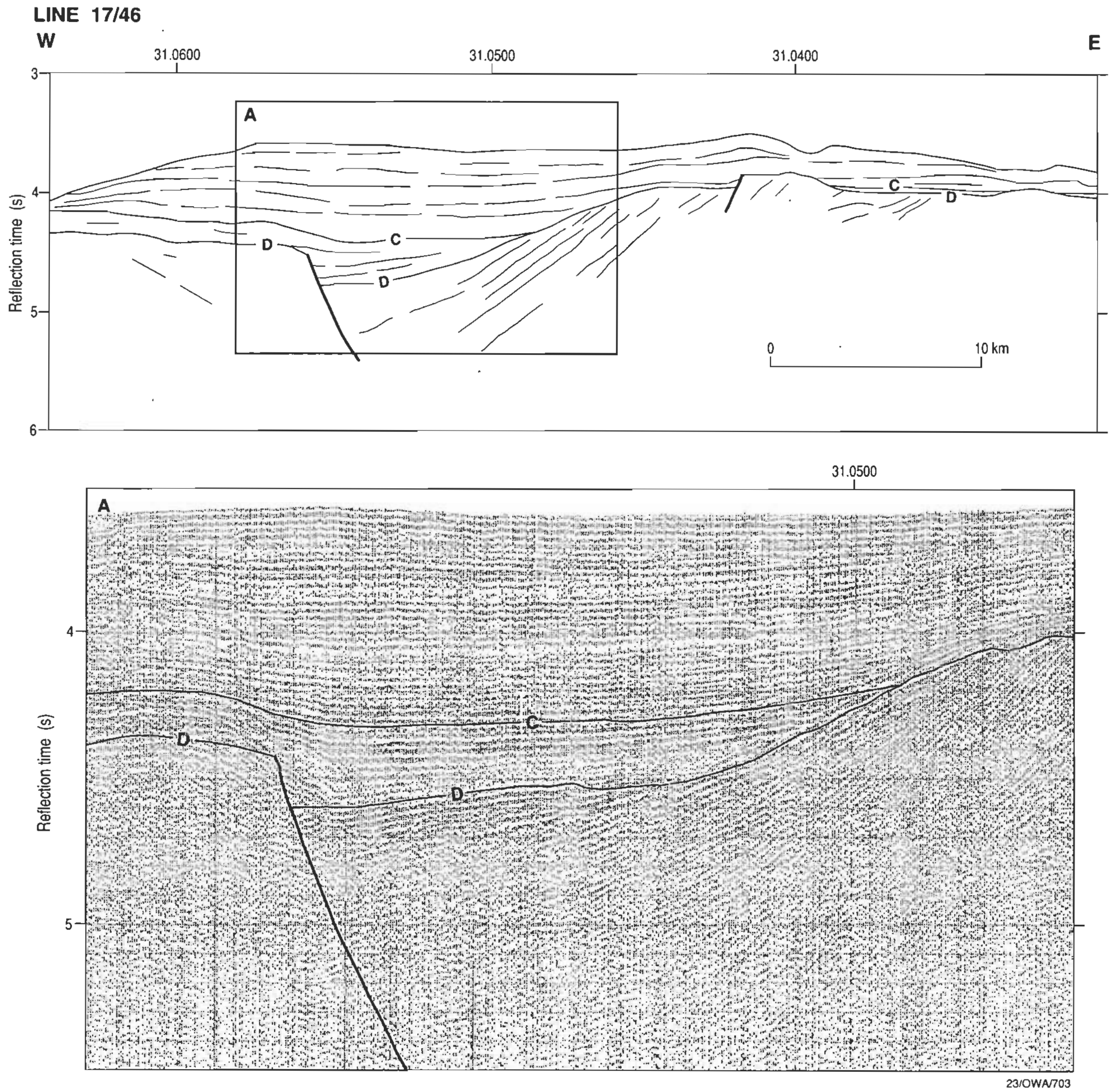


Figure 5. Line drawing of part of AGSO Line 17/46 from the southern Wallaby Plateau showing basement reflectors dipping west into a major fault. Insert shows part of the single-channel monitor record. To the west of the fault, basement displays little in the way of internal layering. Location of the figure shown in Figure 3.

basalt flows have a maximum thickness of several metres and commonly show evidence of subaerial eruption. At three of the Kerguelen Plateau 'basement' sites, isotopic data on basalts provide some evidence of continental lithospheric mantle involvement, whereas at a fourth, on the southernmost part of the plateau, lead isotope data strongly suggest contamination from underlying attenuated continental crust (Alibert, 1991; Storey et al., 1992). Geochemically, the Kerguelen Plateau lavas are transitional between mid-ocean ridge basalt (MORB) and ocean island (Kerguelen–Heard hotspot) basalt, whereas the Ontong Java lavas are high-degree tholeiitic melts similar to those occurring today around Iceland (Storey et al., 1992; Mahoney et al., 1993).

Wedges of seaward-dipping reflectors are common beneath eastern portion of the Wallaby Saddle (Figs 3 and 8). These are similar to those described from beneath several of the world's continental margins by Hinz (1981) and from beneath the outer Vøring Plateau off Norway by Talwani et al. (1981) and Mutter et al. (1982). Where drilled, for example during DSDP Leg 81 on the Rockall Plateau and ODP Leg 104 on the Vøring Plateau, such seaward-dipping complexes consist of subaerially erupted basalt flows and thin interbedded sediment layers (Roberts et al., 1984; Eldholm et al., 1987).

Dredge results

The 1990 AGSO geological program on the central west part of the plateau set out to sample the only known exposure of dipping basement reflectors within the plateau itself, i.e. not on its margins. Two dredge hauls were taken on an east-facing escarpment in 3120–2850 m of water using a chain-bag dredge (Figs 3 and 6; Colwell et al., 1990). Major rock types recovered were altered basaltic lavas, and basaltic conglomerates and sandstones (Table 2). The basaltic sediments are generally poorly sorted and contain rounded clasts up to 2 cm across. The petrology of the dredged rocks is discussed in detail in the Appendix. All samples are strongly altered. Consequently, geochemical analyses have been restricted to the two least-altered samples: samples 36/3 and 38/1 (Table 3). Sample 36/3 is an altered, trachytic-textured, sparsely plagioclase-phyric evolved basaltic to trachyandesitic lava, and sample 38/1 is an altered olivine+augite+plagioclase-phyric basaltic lava.

The intense seafloor alteration of the samples analysed is expressed compositionally by their very high loss on ignition values (around 8.5%). However, with due caution, several important conclusions can be drawn about the affinities and correlations of these lavas. For sample 36/3, despite its broadly basaltic levels of CaO (9%), Ni (98 ppm) and Cr (144 ppm), and its basaltic Ti/Zr (92) value, its SiO_2 content is 41.5%, and MgO is reduced to only 1.07%, whereas Fe_2O_3, P_2O_5 and Y contents are exceptionally high (20.7%, 2.35% and 75 ppm, respectively). This implies extensive depletion of MgO and SiO_2 from the abundant groundmass glass in this sparsely plagioclase-phyric lava, and addition of Fe_2O_3, P_2O_5, and Y. Identical compositional features are recorded in strongly altered basaltic lavas dredged from the Naturaliste Plateau (e.g. sample 12/3 of Coleman et al., 1982). The latter authors showed that the high P_2O_5 levels reside in altered groundmass glass, and the enhanced Y abundances are

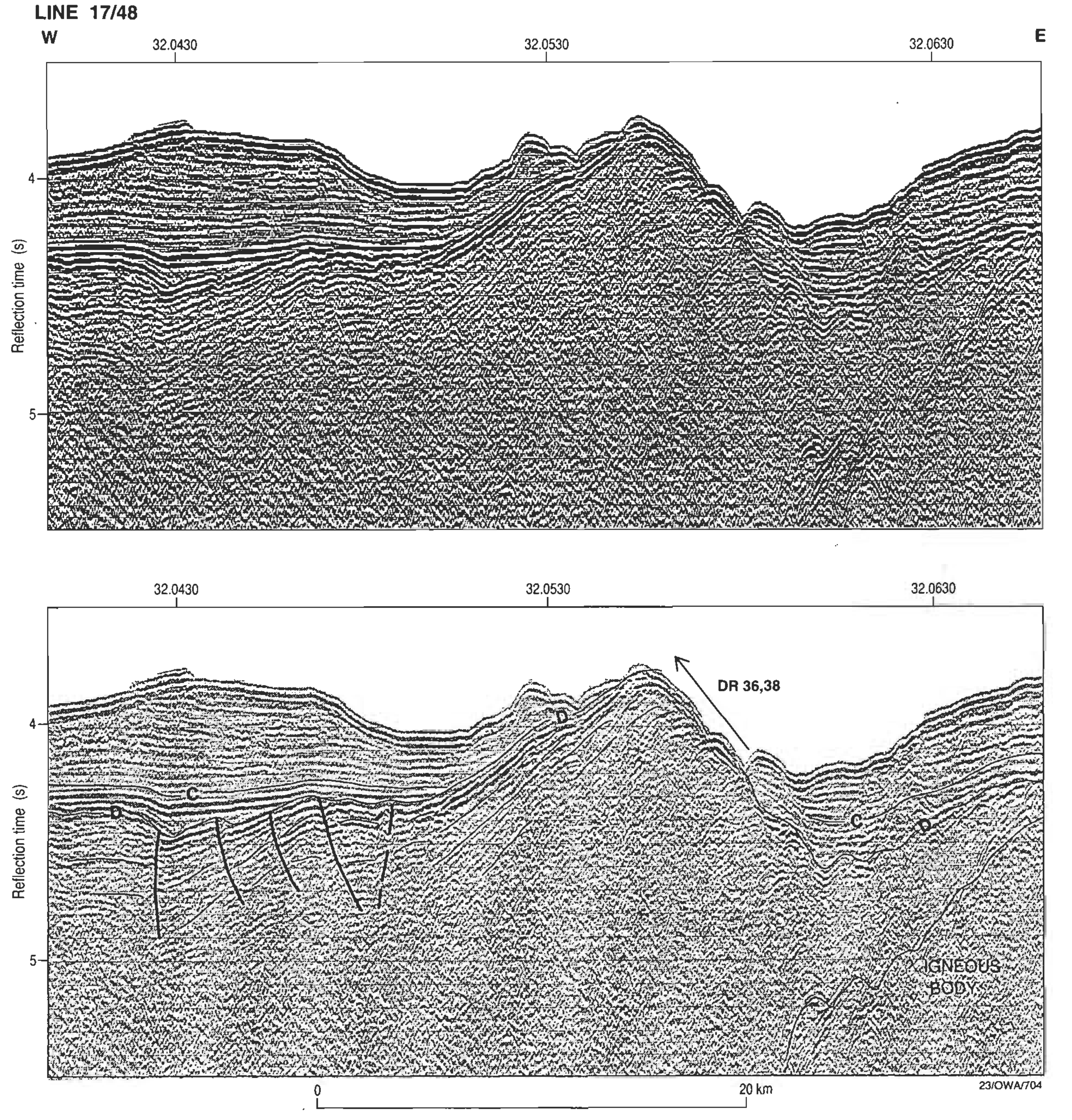


Figure 6. Portion of stacked data from AGSO Line 17/48 showing dipping and faulted reflectors within basement, and the location of the 1990 AGSO dredge hauls, DR 36 and 38. Location of the figure shown in Figure 3.

TABLE 2. Details of AGSO Survey 96 Dredge Hauls (after Colwell et al., 1990).

Dredge	*Position*		*Water Depth*	*Recovery*	*Lithology*
	S	*E*	*(m)*	*(kg)*	
DR36	24°21.81′ 24°22.13′	107°49.7′ 107°46.85′	3120–2850	30	(1)Basaltic conglomerate/sst.-mod. yellow. br.,poorly sorted, most clasts well rounded (2)Mn crusts; (3)trachyandesite/basalt-dk.yellow. altered, (4)?amygdaloidal basalt-mod. br., mod. altered.
DR38	24°21.9′ 24°21.6′	107°49.6′ 107°49.3′	3120–2980	5	(1)Basaltic conglomerate- dk. yell. orange-lt. br., bimodal, clasts subround, mod. consol.; (2)basalt- dk. yell. orange-grey, altered.

also probably located in this altered glass. They suggested that the ultimate source of the "excess" P_2O_5 was likely to have been biogenic.

Sample 38/1 is slightly less altered than 36/3, with the alteration being expressed petrographically as abundant black–brown clayey material in the groundmass, and compositionally as an abnormally high MnO content (2.7%) and perhaps a rather high Fe_2O_3 (16.1%), and low SiO_2 (46.0%) contents. Most other compositional features of 38/1 appear reasonable for a tholeiitic basaltic lava, except for the exceptionally high Pb (220 ppm) and Ni (511 ppm) contents, which almost certainly reflect adsorption of these metals onto the manganese oxides–hydroxides that make up most of the altered groundmass.

Geochemistry and implications

Mantle plume model

Over the years, a number of workers have suggested that certain large bathymetric features of the eastern Indian Ocean, such as the Kerguelen Plateau and Broken Ridge (Fig. 9), are the products of volcanic activity associated with a Kerguelen–Heard mantle plume or hotspot (e.g. Luyendyk & Rennick, 1977; Duncan, 1978, 1981; Morgan, 1981; Duncan & Storey, 1992). It has recently been pointed out that at the time of the main outpouring of lavas that formed the Kerguelen Plateau, the Indian Ocean was probably less than 700 km wide (Storey et al., 1989, 1992; Duncan & Storey, 1992). If the plume head

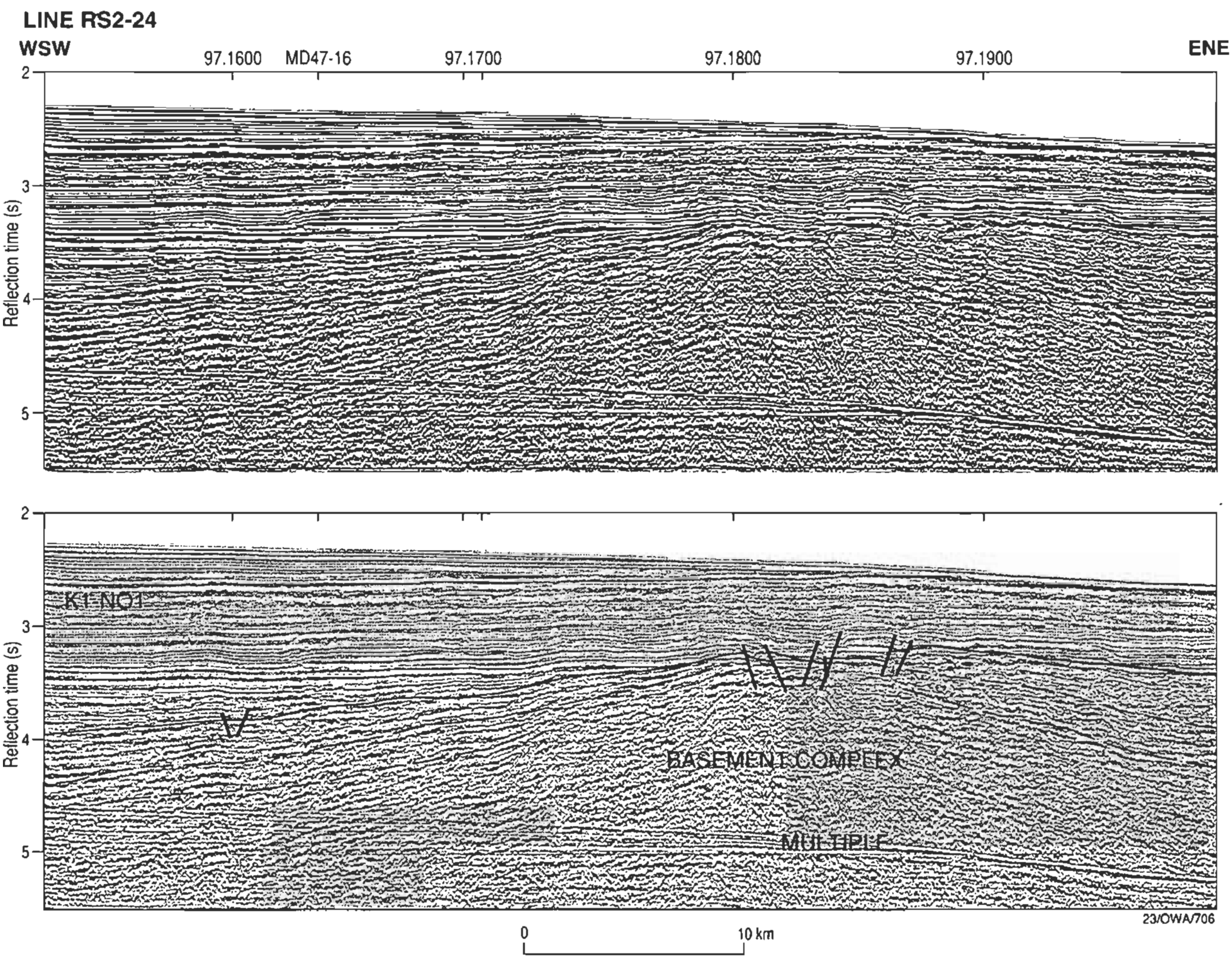


Figure 7. Portion of stacked AGSO seismic data from the southern Kerguelen Plateau (after Coffin et al., 1990) showing dipping basement reflectors similar to those observed beneath the Wallaby Plateau (cf. Figs 4–6).

associated with the Kerguelen plume was as large as has been hypothesized for major plumes (1000–2000 km, White & McKenzie, 1989; see also Duncan & Richards, 1991), then it is likely that some continental lithosphere from beneath Gondwana was mobilized, and involved in the immediately pre-breakup and breakup magmatism during the birth of the eastern Indian Ocean (Storey et al., op. cit.; Duncan & Storey, op. cit.; Mahoney et al., in press). Given the significant isotopic and key trace-element differences between sub-continental and sub-oceanic lithospheric mantle, it is possible to test for the involvement of sub-Gondwana continental lithospheric mantle in the generation of lavas during the early stages of Indian Ocean opening.

It has been convincingly argued by Mahoney et al. (in press) and Storey et al. (1992) that sub-Gondwana continental lithospheric mantle was involved in the petrogenesis of tholeiitic lavas from the Naturaliste Plateau (*Eltanin* Cruise 55), eastern Broken Ridge (*R.D. Conrad*, Cruise 27, Dredge 8), and southernmost Kerguelen Plateau (ODP Site 738). Petrogenesis of the latter suite may have involved contamination by underlying attenuated continental crust. These areas were all located close to continental blocks at the time of their formation. In contrast, the distinctive trace-element and isotopic compositions associated with a continental lithospheric mantle component (see below) are less well-developed in lavas from the main body of the Kerguelen Plateau and Broken Ridge, which were generated at greater distances from continental margins.

On this basis, it might be predicted that the Wallaby Plateau formed in an analogous manner to the Naturaliste Plateau, being a pile of 120–100 Ma tholeiitic basalts with key isotopic and trace-element fingerprints indicating involvement of sub-continental lithospheric mantle [in this case, of the West Australian Archaean craton(s)]. Storey et al. (1989, 1992) and Mahoney et al. (in press) have argued that a key compositional feature of subcontinental lithospheric-derived magma is a significant depletion in Nb (and Ta) relative to LREE, so that primitive mantle-normalized ratios of La/Nb [herein $(La/Nb)_N$] are significantly greater than 1. Values of this ratio for N-MORB and OIB (Sun & McDonough, 1989) are typically 1.1 and 0.8, respectively. Mahoney et al. (in press) show that the $(La/Nb)_N$ values for the eastern Broken Ridge Dredge 8 lavas are 1.4–1.6, for the Naturaliste Plateau 1.9–3.4, and for ODP Site 738 at the southern tip of the Kerguelen Plateau, $(La/Nb)_N$ values are ~2. These data provide strong evidence for the involvement of sub-Gondwana continental lithospheric mantle in the petrogenesis of these lavas.

Isotopic data for Pb-Nd-Sr provide even stronger support for continental lithospheric mantle involvement in the

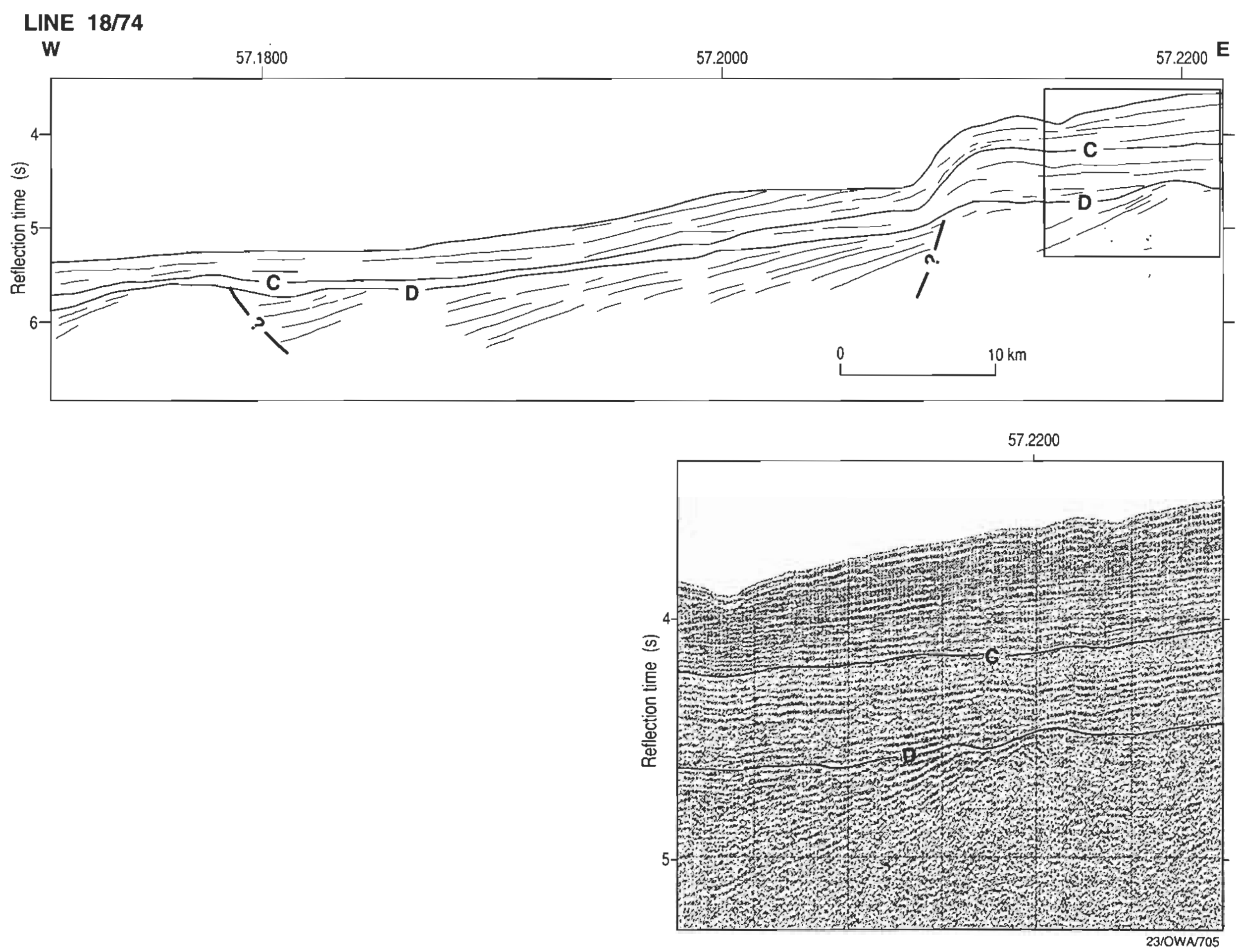


Figure 8. Line drawing of part of AGSO Line 18/74 showing seaward-dipping reflectors beneath the eastern Wallaby Saddle adjacent to the Carnarvon Terrace. Location of figure shown in Figure 3.

Table 3. Wholerock major and trace element analyses of Wallaby Plateau dredged basalts 36/3 and 38/1, and Bunbury Basalts BB1 and BB3 (this study), with comparative analyses from the Bunbury Basalt (BB2, 4, 5 and 7, from Storey et al., 1992), Naturaliste Plateau LREE-enriched basalts from Eltanin Dredge 55 (Mahoney et al., in press), eastern Broken Ridge basalts from Conrad Dredge 8, and two basalts from ODP Leg 119 Site 738 at the southeastern tip of the Kerguelen Plateau (Mahoney et al., in press).

	Rig Seismic	*Rig Seismic*	*Bunbury Bas.*	*Bunbury Bas.*	*Bunbury Bas.*	*Bunbury Bas.*	*Bunbury Bas.*	*Bunbury Bas.*	*Eltanin*	*Eltanin*	*Eltanin*	*Eltanin*	*Eltanin*	*Conrad*	*Conrad*	*Conrad*	*ODP Leg119*	*ODP Leg119*
	Cruise 96 36/3	Cruise 96 38/1	Bunbury Beach BB1	Gelorup Quarry BB3	Storey et al. BB2	Storey et al. BB4	Storey et al. BB5	Storey et al. BB7	Cruise 55 12–2	Cruise 55 12–8	Cruise 55 12–1	Cruise 55 12–21	Cruise 55 12–24	27–08 D8–B–4	27–08 D8–B–9	27–08 D8–E1b	738–33 R–2 65–68	738–34 R–4 30–33
SiO_2	41.5	46.0	51.8	52.1	52.2	51.7	52.7	53.1	59.9	54.7	50.3	51.3	50.6	49.4	49.1	49.9	52.3	51.5
TiO_2	2.34	1.18	1.64	1.97	2.05	1.77	1.96	1.31	1.8	1.88	0.63	0.74	0.99	2.22	3.11	1.6	1.97	1.69
Al_2O_3	17.7	16.1	15.8	14.4	14.6	15.6	15.2	14.7	15	18	20.6	19.8	19.6	13.9	14	13.8	15.8	15.4
Fe_2O_3	20.7	16.1	11.3	12.2	12.5	11.7	12.1	11.8	9.86	13.2	7.9	9.01	10.7	17.7	15.3	14.8	11.8	11.6
MnO	0.16	2.69	0.16.	0.18	0.19	0.18	0.19	0.20	0.12	0.20	0.11	0.12	0.14	0.37	0.42	0.24	0.10	0.13
MgO	1.07	4.41	5.70	5.55	5.71	5.71	5.57	6.54	4.04	4.98	5.24	4.86	4.32	2.48	5.16	6.04	5.70	6.18
CaO	9.03	9.54	10.20	9.45	9.67	10.10	9.77	10.50	4.02	3.87	10.65	9.51	8.90	7.45	9.33	10.40	7.73	9.21
Na_2O	3.31	2.75	2.88	3.35	3.12	3.09	3.22	2.83	3.25	2.75	3.74	3.31	3.40	3.61	2.90	2.49	3.23	3.09
K_2O	1.89	0.98	0.25	0.45	0.38	0.33	0.39	0.40	2.31	1.01	0.46	1.21	1.07	1.30	0.39	0.47	1.41	0.99
P_2O_5	2.35	0.23	0.18	0.24	0.22	0.18	0.21	0.13	0.59	0.24	0.06	0.08	0.11	1.01	0.38	0.18	0.22	0.19
Total	100	100	100	100	100.6	100.4	101.3	101.5	101	100.8	99.7	100	99.8	99.5	100	99.9	100.3	100
LOI	8.46	8.59	1.18	0.57							2.87	2.04	2.22	0.34	0.55	0.1	2.11	1.26
Ni	98	511	70	41	38	69	41	58	111	128	72	72	106	17	34	53	30	28
Cr	144	141	338	274							366	338	356		15	23	61	101
V	743	401	241	262							145	149	209		472	375	254	256
Sc	75	55	34	34	33	31	31	41	22	28	26	28	33	38	44	48	37	37
Zr	152	147	104	136	141	115	133	98	203	140	27	32	46	241	176	138	195	159
Nb	8.0	10.5	6.1	6.0	6.7	5.1	5.9	3.0	8.8	5	0.58	0.84	1.2	19.4	13.8	7.3	11.1	8.3
Y	75	47	35	41	39	35	37	32.2	27	28	12.7	14.7	16.4	48	31	38	27	28
Sr	243	272	228	227	235	231	247	163	132	141	187	151	163	338	297	127	271	301
Rb	27	25	4	8	4.3	5.6	5.7	8.9	61.2	38	5.4	13.7	20	9.7	3.5	6.7	34	13.1
Ba	113	251	55	102							14	58	30	471	160	125	213	303
Pb	3.3	220																
La	18.0	31.9		8.6	10.2	7.2	10.0	8.5	28.6	12.4	1.3	1.5	2.5	29.0	19.5	10.0	19.6	14.6
Ce	30.1	122.4		22.4	25.6	21.4	26.3	18.4	59.0	36.0	3.5	4.5	5.8	80.4	50.0	26.0	53.2	37.2
Nd	22.6	35.8		16.8					29.9	14.9	3.3	3.7	5.4	43.6	26.1	15.3	25.0	19.3
Sm		6.76		5.13	5.28	4.15	4.98	3.53	6.30	4.76	1.34	1.61	2.08	9.93	5.89	4.16	6.06	4.94
Eu		2.14		1.74	1.45	1.42	1.58	1.55	1.99	1.51	0.64	0.73	0.90	3.90	2.08	1.40	2.03	1.70
Tb		1.14		0.97					1.36	1.11	0.30	0.40	0.45	1.62	0.86	0.97	1.17	1.03
Ho		1.60		1.25														
Yb		4.08		3.10	3.50	2.86	3.21	2.84	4.41	2.87	1.26	1.42	1.60	4.38	2.87	3.97	2.72	2.62
Lu		0.61		0.44	0.46	0.38	0.37	0.42	0.52	0.28	0.20	0.22	0.23	0.63	0.43	0.65	0.37	0.36
Th		4.75		1.31	1.60	1.20	1.50	1.50	6.40	1.87				4.95	2.63	2.19	2.82	2.11
Ta		<0.5		0.59					0.62	0.34	0.06	0.09	0.09	1.19	0.80		0.61	0.50
Hf		2.76		3.40	4.50	3.80	5.00	3.20	5.50	3.70	0.71	0.84	1.23	5.91	4.07	3.21	4.53	3.79

Footnote: Major and trace elements excluding REE, Hf, Ta and Th by XRF analysis in Dept of Geology, University of Tasmania. Other elements by INAA at Becquerel Labs., Lucas Heights.

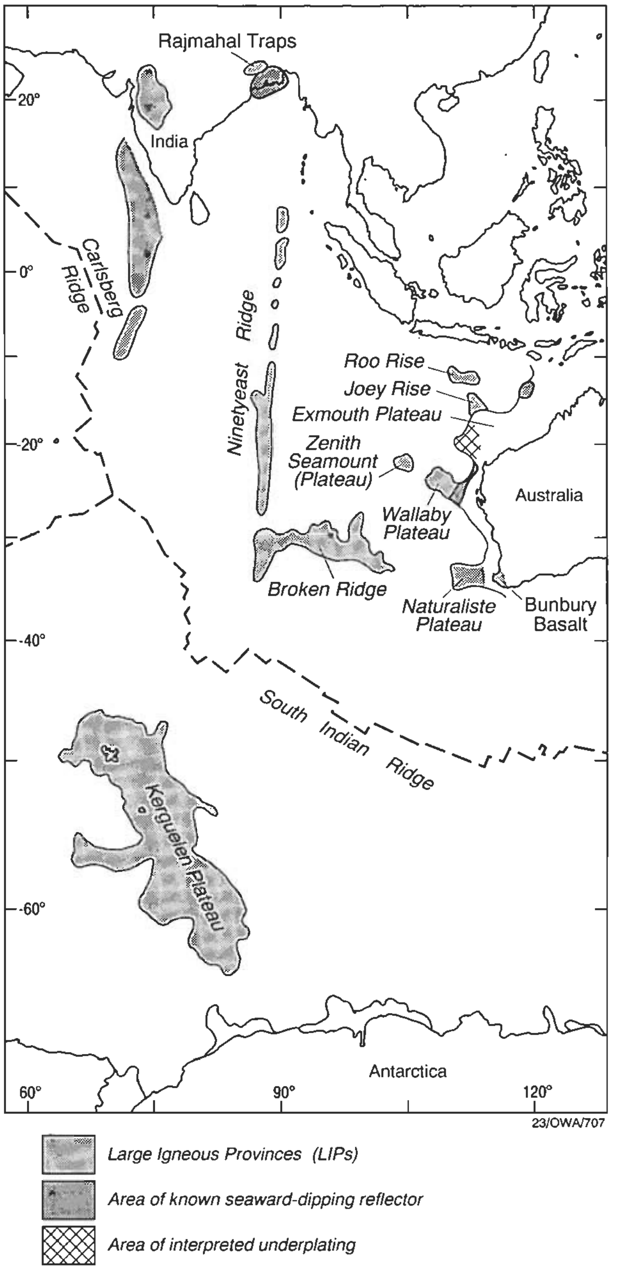


Figure 9. Distribution within the eastern Indian Ocean of areas of: (i) interpreted underplating; (ii) large igneous features (LIPs), and (iii) known seaward-dipping reflectors. Areas of underplating from Mutter et al. (1989) and Lorenzo et al. (1991); LIP distribution from Coffin & Eldholm (1991); and areas of seaward-dipping reflectors from this paper, Hinz (1981), Storey et al. (1992) and Hopper et al. (1992).

petrogenesis of lavas forming the eastern Broken Ridge, Naturaliste Plateau and the southernmost Kerguelen Plateau (Storey et al., 1989, 1992; Alibert, 1991; Mahoney et al., in press). These lavas show more extreme isotopic variations than recorded for the post-45 Ma Kerguelen plume magmas, with substantially more negative ε_{Nd} values, $^{87}Sr/^{86}Sr$ values extending to notably more radiogenic values, and lower $^{206}Pb/^{204}Pb$ values than the modern Kerguelen plume lavas.

Wallaby Plateau lavas

Abundances of the immobile elements Zr, Nb and the REE (Table 3) show that samples 38/1 and 36/3 from the Wallaby Plateau are probably evolved tholeiitic basalts. Both samples have Zr/Nb values (14–20) similar to most analyzed lavas from the eastern Broken Ridge and ODP Site 738 on the southern Kerguelen Plateau, and are just below values (23–28) for the non-plagioclase accumulative basalts from the Naturaliste Plateau reported by Storey et al. (1989). Similarly, sample 38/1 shows a strongly LREE-enriched REE pattern (with a pronounced alteration-induced positive Ce anomaly), very close to those reported by Storey et al. (1989) and Mahoney et al. (in press) for basalts from the other eastern Indian Ocean locations (Figs 10b, c, d). In the absence of isotopic data for the Wallaby Plateau samples, the high $(La/Nb)_N$ values of 3.1 for 38/1 and 2.33 for sample 36/3 (coupled with the <0.5 ppm Ta measured for 38/1), and the pronounced negative Nb anomalies on the multi-element variation diagrams in Figure 11a-d, provide substantial support for a broad correlation of the Wallaby Plateau basalts with those from the other eastern Indian Ocean locations (Naturaliste Plateau, southernmost Kerguelen Plateau, and eastern Broken Ridge).

Bunbury Basalt

Petrographic descriptions of two samples of the Bunbury Basalt, which has been suggested to be a continental flow basalt related to the Kerguelen plume (Davies et al., 1989; Storey et al., 1992), are given in the Appendix. They are plagioclase+sparsely augite-phyric rather coarse-grained basalts that show only very minor alteration of the glassy mesostasis. Data given in Storey et al. (1992) and our new data (Table 3) show that the Bunbury Basalt has $(La/Nb)_N$ values ranging from 1.46–2.94. REE patterns are less LREE-enriched (Fig. 10c) than those for the enriched lavas from the Naturaliste Plateau (Fig. 10b), from the eastern end of Broken Ridge (Fig. 10d), and from ODP Site 738 on the southern tip of Kerguelen Plateau (Fig. 10c), but normalized element variation patterns (Fig. 11b-d) still show pronounced negative Nb–Ta anomalies.

We provide here the first isotopic data for the Bunbury Basalt (Table 4). Data are plotted on Figure 12a-e (note initial ratios are plotted for Nd and Sr isotopes, but

Table 4. Isotopic data for Sr, Nd and Pb for Bunbury Basalts BB1 and BB3 (this paper). Initial Nd and Sr values recalculated for 110 Ma. Analyses performed by Ruth Lanyon at the Research School of Earth Sciences, Australian National University.

	Bunbury Beach BB1	*Gelorup Quarry BB3*
87Sr/86Sr	0.704128±27	0.70482±18
(87Sr/86Sr)i	0.704045	0.704741
143Nd/144Nd	0.512882±6	0.512741±7
(143Nd/144Nd)i	0.512604	0.512463
206Pb/204Pb	17.967	17.905
207Pb/204Pb	15.551	15.584
208Pb/204Pb	37.946	38.22

Footnote: Sr, Nd and Pb isotopes were collected using conventional ion exchange techniques and analysed using a Finnigan MAT 261 Mass Spectrometer operated in static multicollector mode. $^{87}Sr/^{86}Sr$ values are normalised to $^{86}Sr/^{88}Sr = 0.1194$; $^{143/144}$ Nd values are normalised to $^{146}Nd/^{144}Nd = 0.7219$. Sr and Nd 2σ precision refers to within-run statistics. Pb isotopes were analysed using the ^{207}Pb-^{204}Pb double spike technique (Hamelin et al., 1985). Mean ($\pm 2\sigma$) standard values obtained over this period of analysis include: NBS 987(n=39) - $^{87}Sr/^{86}Sr = 0.710209\pm35$; La Jolla (n=54) - $^{143}Nd/^{144}Nd = 0.512189\pm19$; NBS 981 (n=8) - $^{206}Pb/^{204}Pb = 16.939\pm10$, $^{207}Pb/^{204}Pb = 15.494\pm11$, $^{208}Pb/^{204}Pb = 36.712\pm31$.

measured values are plotted for Pb isotopes). It is evident from these plots that the Bunbury Basalt samples fall

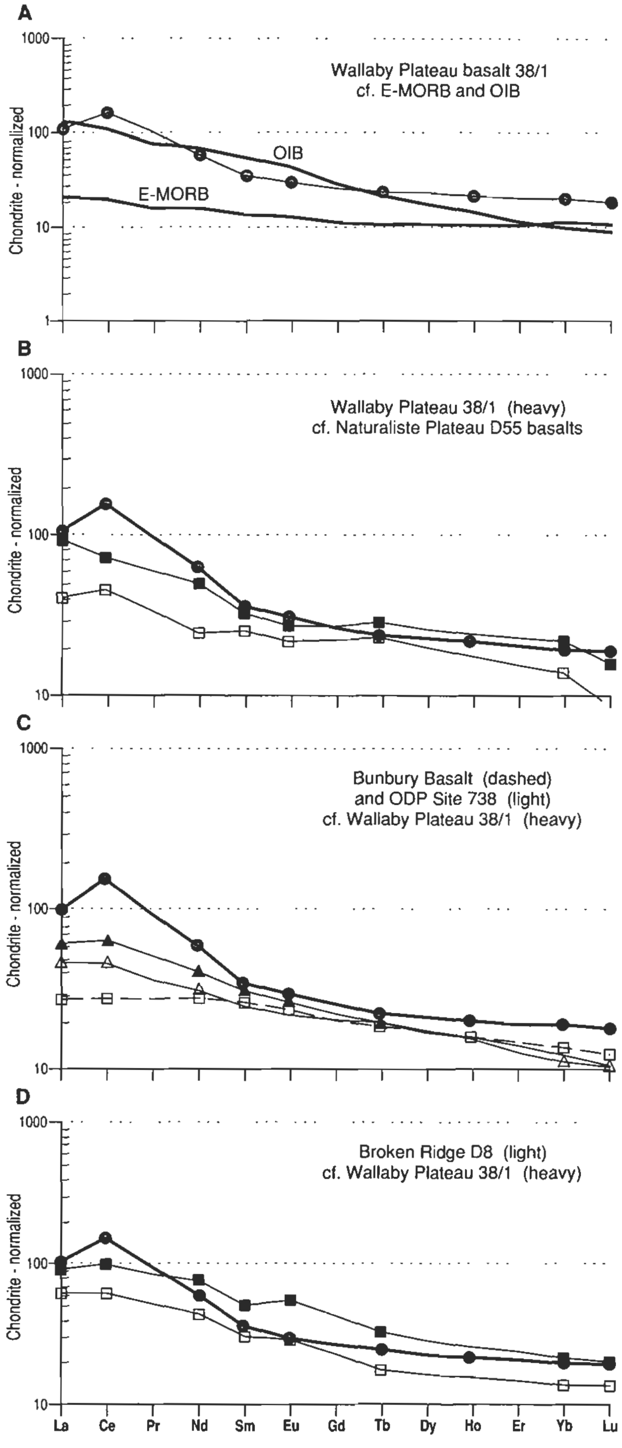


Figure 10. Chondrite-normalized REE pattern of dredged Wallaby Plateau basalt 38/1 (filled circles in Fig. 10a, heavy line in b, c, d) compared with those for: (a) the average E-MORB amd OIB (from Sun & McDonough, 1989); (b) two basalts dredged from the northern margin of the Naturaliste Plateau by USNS *Eltanin* at dredge site 55 (filled and open squares; data from Mahoney et al., in press); (c) Bunbury basalt (open squares, this paper) and two basalts from ODP Site 738 on the southeastern tip of the Kerguelen Plateau (filled and open triangles; data from Mahoney et al., in press); and (d) two basalts from the eastern end of Broken Ridge (filled and open squares; data from Mahoney et al., in press).

along the same broad trends in Pb–Sr–Nd isotopic space as do those samples from the Naturaliste Plateau and eastern Broken Ridge analyzed by Storey et al. (1992) and Mahoney et al. (in press). The enriched radiogenic

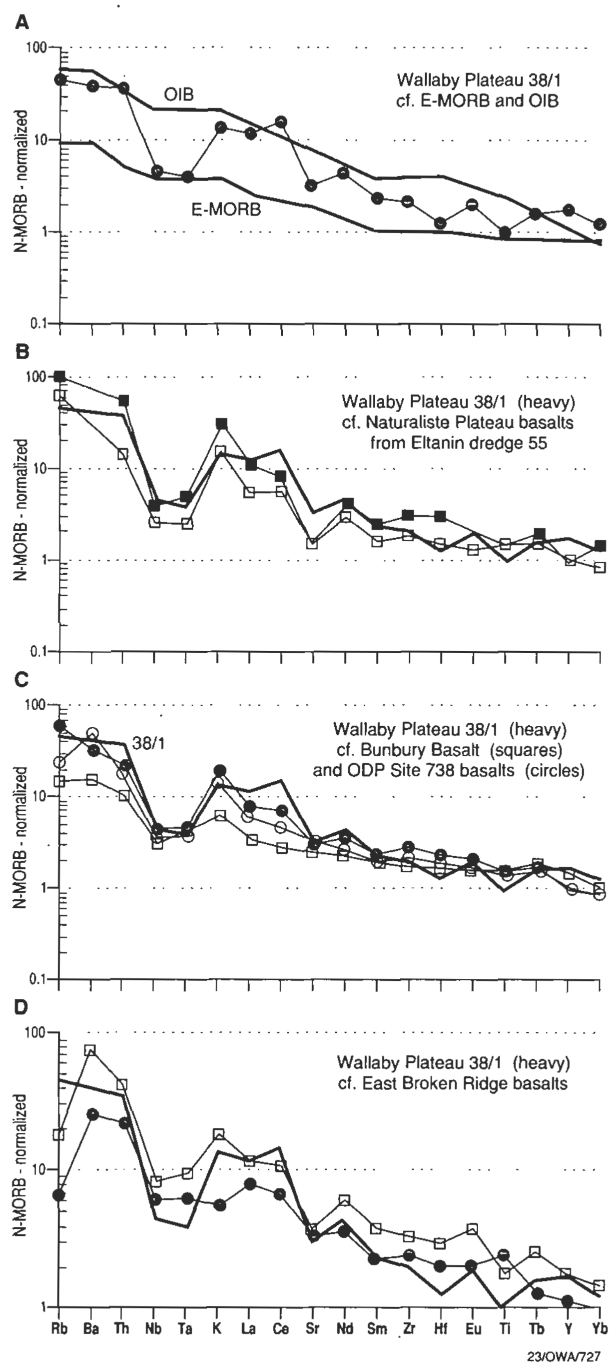


Figure 11. NMORB-normalized multi-element variation patterns of dredged Wallaby Plateau basalt 38/1 (filled circles in Fig. 11a, heavy line in b, c, d) compared with those for: (a) the average E-MORB and OIB (from Sun & McDonough, 1989); (b) two basalts dredged from the northern margin of the Naturaliste Plateau by USNS *Eltanin* at dredge site 55 (filled and open squares; data from Mahoney et al., in press); (c) Bunbury basalt (open squares, this paper) and two basalts from ODP Site 738 on the southeastern tip of the Kerguelen Plateau (filled and open circles; data from Mahoney et al., in press); and (d) two basalts from the eastern end of Broken Ridge (filled circles and open squares; data from Mahoney et al., in press).

mantle end-member involved in the petrogenesis of these lavas may be subcontinental lithospheric mantle, as hypothesized by the above authors. However, for the Bunbury Basalt, further study is required to determine whether the less-enriched mantle source component of these lavas was Kerguelen plume mantle, or typical Indian Ocean MORB asthenosphere.

In summary, key trace-element features, notably the very low Nb and Ta contents coupled with quite LREE-enriched REE patterns, suggest that two analyzed, altered basalts from the Wallaby Plateau are evolved tholeiites very similar to those sampled from the Naturaliste Plateau, the eastern part of Broken Ridge, and the southernmost part of the Kerguelen Plateau. Perhaps not surprisingly given their present positions with respect to the cratonic crust of Australia, the basalts sampled from the Wallaby Plateau, Bunbury and Naturaliste Plateau have compositional fingerprints suggesting the involvement of subcontinental lithospheric mantle in their petrogenesis.

Large igneous provinces off western Australia

The above analysis of seismic data and the new dredge samples from the Wallaby Plateau add new weight to previous studies that have suggested that the Wallaby Plateau is a volcanic feature, probably underlain by oceanic crust, and formed during or following breakup (Veevers & Cotterill, 1978; von Stackelberg et al., 1980; Veevers et al., 1985). This raises the question of how the Wallaby Plateau relates to other volcanic provinces that either once lay relatively near the west Australian margin at, or not long after breakup, or that now actually form part of the margin.

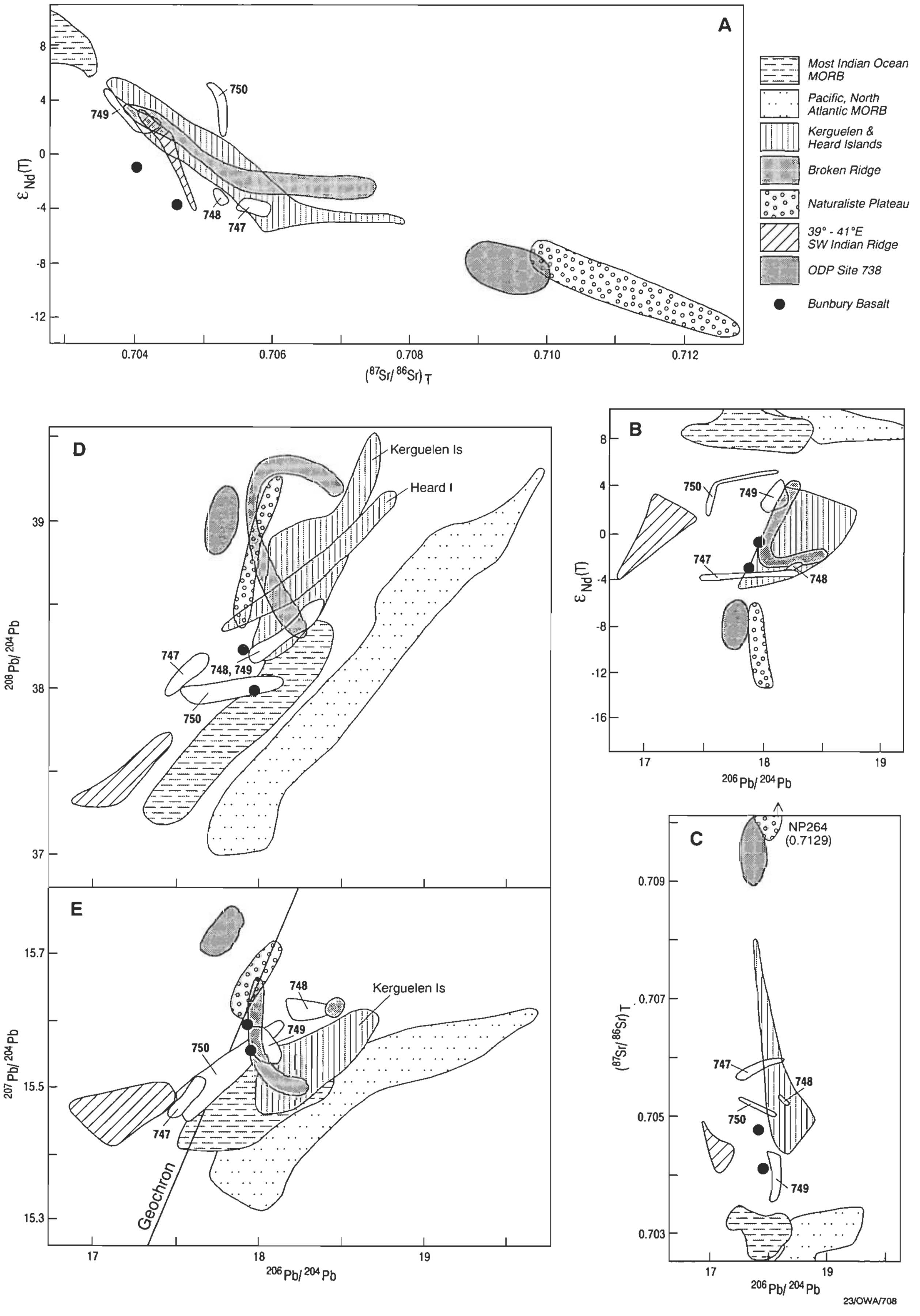
It is now well recognised that voluminous emplacement of mafic igneous rock via processes other than 'normal' seafloor spreading can produce features such as continental flood basalts (CFB) and associated intrusives; so-called 'volcanic passive margins'; oceanic plateaux (OP); submarine ridges (SR); ocean basin flood basalts; and seamount chains (Coffin & Eldholm, 1991, 1992). These large igneous provinces (LIPs) contain enormous volumes of lava that were emplaced very rapidly, and have many similarities in their temporal, spatial, and compositional characteristics (Coffin & Eldholm, op.cit.). Coffin & Eldholm (1991, 1992, in press) illustrate the distribution of LIPs in the eastern Indian Ocean, and summarise a variety of information on these features. Plate reconstructions, such as those of Davies et al. (1989) and Royer & Coffin (1992), indicate that six LIPs lay within or relatively close to the west Australian margin up until about 110 Ma, following breakup of Gondwanaland and the westward drift of Greater India. These are the Joey and Roo Rises (OP's); the Zenith and Wallaby Plateaux (OP's); the Bunbury Basalt (CFB — southwest Australia); the Naturaliste Plateau (?OP); the Rajmahal Traps (CFB — northeast India); the Kerguelen Plateau (OP); and Broken Ridge (SR) (Fig. 9). Several of these features have now been imaged with modern seismic data, and have been sampled during various DSDP and ODP legs in the Indian Ocean. The formation of most of these LIPs is interpreted to be related to volcanic activity associated with the Kerguelen–Heard mantle plume or hotspot. Davies et al. (1989) and Storey et al. (1992) include the Naturaliste Plateau and its possible on-land equivalent, the Bunbury Basalt, among the earliest products of the Kerguelen hotspot. This results in a hotspot track which Davies et al. (1989) hypothesized may date back to about 135 Ma and includes, in chronological order, the Bunbury Basalt, Naturaliste Plateau, Rajmahal Traps (India), Kerguelen/Broken Ridge, Ninetyeast Ridge, and finally the northeasternmost Kerguelen Plateau and the associated Tertiary–Recent volcanism of Kerguelen and Heard/McDonald Islands. Storey et al. (1992) in their figure 1 also suggest that volcanics of the Sylhet province (India), a seaward-dipping reflector sequence on the adjacent continental margin, the Damodar and Darjeeling Lamprophyres (India), and the Prince Charles Mountains Lamprophyres (Antarctica) are some of the early products of the Kerguelen hotspot.

We note here that an unpublished ^{40}Ar-^{39}Ar age of 115±2.0 Ma (R.A. Duncan, *pers. comm.*, 1994) for one of the Bunbury Basalt samples described in this paper puts formation of the Bunbury Basalt at the same time as the bulk of the Kerguelen Plume, and possibly the other compositionally similar, volcanic plateaux off western Australia (Naturaliste and Wallaby Plateaux and Broken Ridge). The Rajmahal Traps of eastern India have also commonly been linked to Kerguelen plume activity, and have yielded an ^{40}Ar-^{39}Ar age of 117±1 Ma (Baksi, 1986; Duncan & Storey, 1992). This whole volcanic province would have had a diameter of about 1500–2000 km, with the bulk of magma production occurring within the interval from about 120 Ma (earliest Kerguelen plume activity?) to about 110 Ma (Duncan & Storey, 1992). Such an extensive magmatic province fits with the 'mega plume' model of White & McKenzie (1989), in which massive igneous activity (volcanism and underplating) results from enhanced decompression melting of hot asthenospheric mantle related to a mantle plume, as it rises passively beneath stretched and thinned lithosphere. However, it is important to emphasize that the timing of the early Kerguelen plume activity appears be too young for the plume to be involved in the initiation of the passive margin of northwestern Australia, where breakup and accretion of oceanic crust occurred at ~155 Ma along the Argo Abyssal Plain and around 130 Ma on the Gascoyne and Cuvier Abyssal Plain sections.

Other magmatic provinces of the west Australian margin

Whereas the Naturaliste and Wallaby Plateaux are commonly referred to in discussions of LIPs of the eastern Indian Ocean (White & McKenzie, 1989; Coffin & Eldholm, 1992), it is not generally appreciated that seismic data, dredging and ODP sampling have also defined other large areas of magmatism (intrusives, extrusives and underplating) along the west Australian continental margin. Volcanics are commonly sampled along the margin and are described in numerous papers, for example by von Rad & Exon (1983), Veevers (1984), Mutter et al. (1988), Mutter et al (1989), von Rad et al. (1992), Exon & Buffler (1992), Hopper et al. (1992), Colwell et al. (1994 — this issue), and Crawford & von Rad (1994 — this issue). The styles of magmatism basically fall into four types:

1. Rift-related volcanics of possible Early Triassic (Scythian) to Middle Jurassic (Aalenian–Bathonian) age, but probably mainly focussed in the Late Triassic (Rhaetian; K–Ar age: 213 Ma) to Early Jurassic (Toarcian–Hettangian; K–Ar age: 190–206 Ma) sampled in several of the Browse Basin wells (landward of the Scott Plateau; Fig. 1), from the margins of the Wombat Plateau (northern part of the Exmouth


A
Most Indian Ocean MORB
Pacific, North Atlantic MORB
Kerguelen & Heard Islands
Broken Ridge
Naturaliste Plateau
39° - 41°E SW Indian Ridge
ODP Site 738
Bunbury Basalt
750
749
748
747
ε Nd(T)
(87Sr/ 86Sr) T
B
206Pb/ 204Pb
C
NP264 (0.7129)
D
Kerguelen Is
Heard I
748, 749
208Pb/ 204Pb
E
Geochron
207Pb/ 204Pb
23/OWA/708

Plateau, Fig. 1), and as dredged (but as yet un-dated) and interpreted from seismic data from the southwest margin of the Exmouth Plateau east of ODP Site 766 (Exon & Buffler, 1992). This suite of volcanics ranges from highly differentiated K-rich rhyolitic to trachytic rocks erupted under subaerial to very shallow-marine conditions (von Rad & Exon, 1983) to basaltic dykes and sills of broadly MORB composition (Buffler et al., 1992).

'Intrabasement' reflectors visible on deeper penetration seismic data from large areas of the Exmouth Plateau have been interpreted as detachments related to Permo-Triassic or Carboniferous–Permian extension (Mutter et al., 1989; Williamson et al., 1990); however, Exon & Buffler (1992) interpreted some of these features, particularly beneath the western plateau, as Middle–Late Jurassic sills intruded during the latest stage of margin rifting.

2. Probable Late Jurassic to Early Cretaceous volcanics (Hauterivian–Valanginian; K–Ar minimum age of 128–133 Ma) dredged from the margin of the Scott Plateau (Hinz et al., 1978; von Rad & Exon, 1983). These rocks are alkali basalts (?hawaiite) and are rather anomalous because, assuming that the K–Ar dating is not too inaccurate, they appear to have been emplaced some 20 million years after Argo Abyssal Plain breakup, dated at ~155 Ma (Ludden, 1992). In fact, the age of the Scott Plateau basalts corresponds better with the 130 Ma breakup age of the Gascoyne and Cuvier Abyssal Plains some 1000–1500 km to the southwest.

   Further evidence for widespread volcanism of this age is provided by the sampling of Berriasian–Valanginian bentonites (i.e. smectite-dominated, altered volcanic dacitic to rhyolitic ash layers) at ODP sites on the northern, western and southern margins of the Exmouth Plateau, and the southern Argo Abyssal Plain (von Rad & Thurow, 1992). Potential sources for the explosive volcanic activity that resulted in the deposition of proximal ash turbidites in the Wombat Plateau area include the Joey and Roo Rises (volcanics have been sampled, but are un-dated, and magnetic anomaly picks imply a Late Jurassic or younger age for these features — Fullerton et al., 1989), the Scott Plateau and the Wallaby Plateau/Cape Range Fracture Zone area.

   On the basis of dredging and seismic data, Ramsay & Exon (1994 — this issue) mapped a 'volcanic layer' beneath the northwestern Rowley Terrace, adjacent to the offshore Canning Basin (Fig. 1), covering an area of 25 000 km^2. They interpret this feature to be volcanic flows and volcaniclastics of possible Callovian–Oxfordian age, and suggested it was emplaced just prior to breakup in the Argo Abyssal Plain.

3. Seaward-dipping reflector sequences have been identified from seismic data beneath the northeast margin of the Argo Abyssal Plain (Hinz, 1981) adjacent to the Scott Plateau, and beneath the lower slope of the northern Carnarvon Terrace (Cuvier margin; Hopper et al., 1992) adjacent to the Cuvier Abyssal Plain (Figs 1 and 9). These sequences are interpreted as basaltic flows emplaced as a result of unusually voluminous subaerial volcanism around the time of breakup, i.e. of Callovian–Oxfordian age in the Argo Abyssal Plain, and Valaginian–Hauterivian age in the Cuvier Abyssal Plain. Hopper et al. (op. cit.) suggested that in the Cuvier area the seaward-dipping reflectors were formed during the early stage of oceanic crust formation (i.e.post-breakup) and resulted in an exceptionally thick oceanic crust adjacent to the margin.

4. Underplated crust defined by a layer with velocity 7.2–7.3 km/s on expanded spread seismic profiles beneath the western (Mutter et al., 1989) and southern (Lorenzo et al., 1991) Exmouth Plateau (Fig. 9). Mutter et al. (op. cit.) related the underplating beneath the western plateau to Middle to Late Jurassic lithospheric thinning, resulting in plutonic underplating, accompanied by large-scale normal faulting without a detachment. Lorenzo et al. (op. cit.), however, assigned the southern Exmouth Plateau underplating to melt generation during Early Cretaceous seafloor spreading in the Cuvier Abyssal Plain, and consequent transform motion along the Cape Range Fracture Zone.

Discussion

The distribution of volcanics along the western Australian margin appears to consist of two types. A zone of Triassic to Early Cretaceous volcanics occurs along the margin and includes areas of seaward-dipping reflectors, flows, sills and possible underplating. These features lie within a zone that parallels the margin, implying that igneous emplacement was controlled by dynamic rift/breakup processes and not necessarily by proximity to a plume. Such distributions can be explained by the convective partial melting model of Mutter et al. (1988), in which the lateral temperature contrasts caused by rifting and asthenospheric upwelling drive convection in the upper mantle. The resulting "convective melt" production supplements that available from the passive upwelling of the asthenosphere related to lithospheric extension and thinning. Another similar non-plume model was recently proposed by Holbrook & Kelemen (1993) to explain the widespread, margin-parallel zone of seaward-dipping reflectors observed on deep seismic data beneath the USA Atlantic margin. They argue that dynamic mantle upwelling during rifting can produce the melts that result in the accumulation of thick igneous crust when the

Figure 12. (a) Initial ε_{Nd} vs $^{87}Sr/^{86}Sr$ for two Bunbury Basalts (filled circles: this study, Table 4) compared with fields from Mahoney et al. (in press) for ODP Sites 738, 747, 748, 749 and 750 on the Kerguelen Plateau, LREE-enriched Naturaliste Plateau basalts, Broken Ridge basalts, modern Kerguelen-Heard plume basalts, and MORB from the Indian Ocean, excluding those from 39°–41°S on the SW Indian Ridge (field shown separately) and the central and western SW Indian Ridge.
(b) Initial ε_{Nd} vs present day $^{206}Pb/^{204}Pb$ for two Bunbury Basalts (filled circles: this study, Table 4) compared with fields from Mahoney et al. (in press) for suites noted in Figure 12a, plus part of field for Pacific–North Atlantic MORB.
(c) Initial $^{87}Sr/^{86}Sr$ vs present day $^{206}Pb/^{204}Pb$ for two Bunbury Basalts (filled circles: this study, Table 4) compared with fields from Mahoney et al. (in press) for suites noted in Figure 12a, plus part of field for Pacific- North Atlantic MORB.
(d) Present day $^{208}Pb/^{204}Pb$ vs $^{206}Pb/^{204}Pb$ for two Bunbury Basalts (filled circles: this study, Table 4) compared with fields from Mahoney et al. (in press) for suites noted in Figure 12a, plus part of field for Pacific–North Atlantic MORB.
(e) Present day $^{207}Pb/^{204}Pb$ vs $^{206}Pb/^{204}Pb$ for two Bunbury Basalts (filled circles: this study, Table 4) compared with fields from Mahoney et al. (in press) for suites noted in Figure 12a, plus field for Pacific–North Atlantic MORB.

asthenospheric upwelling rate exceeds the lithospheric extension rate. Presumably, the volcanics produced from the above processes, particularly in the immediate pre-breakup phase, could contain a strong signature of sub-continental lithospheric mantle, as appears to be the case for the Wallaby and Naturaliste Plateaux, and Bunbury Basalt. However, unlike these latter features, the volcanism would be expected to occur around the time of breakup.

The second style of volcanic distribution is reflected by the irregular volcanic buildups that form part of the margin, such as the Wallaby and probably Naturaliste Plateaux. These features have many of the characteristics of other eastern Indian Ocean LIPs that are distant from continental margins, such as Kergulelen Plateau and Broken Ridge. All appear to have formed some 10–20 million years after the 130 Ma breakup of the Gascoyne, Cuvier and Perth Abyssal Plains, by voluminous outpourings of plume-related magmas onto pre-existing, relatively young, oceanic crust.

Broad plume head models in which mantle plumes can impact on and heat the base of the lithosphere over a 2000 km diameter zone (e.g. White & McKenzie, 1989; Duncan & Richards, 1991) could explain the LIPs of eastern India, the southeast Indian Ocean (Kerguelen Plateau and Broken Ridge) and western Australia (the Wallaby and Naturaliste Plateaux, and Bunbury Basalt) as all of these lay within a 1500–2000 km diameter circle at this time. Plate reconstructions (e.g. Veevers et al., 1991; Royer & Coffin, 1992; Royer et al., 1992) indicate that from 120–110 Ma Greater India and Antarctica /Australia were separated by a 300–800 km wide zone of newly formed oceanic crust, and therefore the LIPs are all post-breakup features. Thus, the Wallaby and Naturaliste Plateaux are unlikely to have been formed by decompression melting associated with the stretching of continental lithosphere above a plume as suggested by White & McKenzie (1989).

The 119 Ma plate reconstruction of Royer et al. (1992) shows the spreading ridge between India and Antarctica/Australia intersecting the site of the future Kerguelen/Broken Ridge LIP, and passing by the sites of the future Naturaliste and Wallaby Plateau LIPs along major transform offsets. This implies that the distribution and form of the LIPs above a single plume head is controlled by processes occurring in the overlying lithosphere. That is, the plume is an essential pre-condition for the LIP as it provides the excess heat, but it is the thinning of the lithospheric cap that controls their location. As discussed by Saunders et al. (1992), plume heat may be channelled towards zones of thin lithosphere associated with features such as active spreading ridges and continental rifts. We speculate that the Wallaby and Naturaliste Plateau LIPs may have formed when the ridge tip associated with seafloor spreading between India and Antarctica/Australia passed along major transforms that separated it from recently accreted oceanic crust to the north. This would have resulted in re-heating, and perhaps rifting and thinning, of the pre-existing oceanic crust producing an episode of voluminous magmatism. If this model is correct, volcanism associated with the Wallaby Plateau LIP would have commenced at about 118 Ma (chron M0), some 10 Ma after the formation of the adjacent Cuvier Abyssal Plain crust. The Royer et al. (1992) plate reconstruction indicates that the Kerguelen/Broken Ridge LIP may have begun at the intersection of the spreading ridge with either a major transform, or the western end of the incipient spreading ridge between Antarctica and Australia. This location may have also been a locus of re-heating and rifting of pre-existing oceanic crust leading to the formation of a LIP.

The above discussion raises the question of when the postulated mantle plume responsible for the eastern Indian Ocean LIPs may have first begun to impact on the base of the lithosphere. Perhaps the ridge jump at about chron M4 time (123 Ma) in the Cuvier and Gascoyne Abyssal Plains was related in some way to this phenomenon.

Another intriguing possibility for the origin of the Wallaby Plateau may lie in the convective partial melting model of Mutter et al. (1988). They show an example of convective melting following an oceanic spreading ridge crest jump (Fig. 13 of Mutter et al., op. cit.) resulting in an oceanic upgrowth which is later split apart by normal seafloor spreading to form two parallel aseismic ridges. Such a model may explain both the Sonja and Sonne Ridges in the Cuvier Abyssal Plain, and their bracketing of the Wallaby Plateau. If a ridge jump occurred in the region at about magnetic anomaly M4 time, an oceanic upgrowth resulting from re-rifting of pre-existing oceanic crust and associated convective partial melting could have formed the Wallaby Plateau complex. North of a possible transform, the initial upgrowth was split by 'normal' seafloor spreading to form the Sonne and Sonja ridges, whereas to the south convective melting continued for some time, resulting in the Wallaby Plateau volcanic province. For this model to be valid, it would require some re-identification of the magnetic anomalies west of the Sonne Ridge. This would not appear to present a significant problem as, although the anomalies are clear, their form is not so distinctive that other options are not possible. Such a model could also explain several other observations concerning the Sonne and Sonja Ridges:

- The Sonja Ridge is normally associated with the western limb (about chrons M8–9) of the M10–M4 spreading episode that formed the eastern Cuvier Abyssal Plain; however, there is no comparable feature associated with the equivalent eastern limb.
- The Sonja Ridge is at least as significant a feature as the Sonne Ridge, which is normally considered to be an abandoned spreading centre.
- Both the Sonja and Sonne Ridges trend at 15–20°, whereas the seafloor spreading magnetic anomaly pattern that they are normally associated with is generally shown as trending about 30°. This difference in trends may reflect different episodes of seafloor spreading.

Conclusions

The Wallaby Plateau is underlain by a "basement" containing complex dipping seismic reflector sequences similar to those of large, oceanic, plume-related volcanic features (LIPs), such as the Kerguelen and Ontong Java Plateaux. ODP drilling at both the latter locations has confirmed that these sequences are composed of interbedded, subaerial basalt flows and volcaniclastic sediments. The Wallaby Saddle, which lies between the Wallaby Plateau and the upper continental slope, and the lower continental slope adjacent to the Cuvier Abyssal Plain, are underlain in places by seaward-dipping reflectors which appear similar to those drilled by the ODP on the

Rockall and Vøring Plateaux. These sequences were also shown to consist of subaerial basalt flows and interbedded sediments, and are thought to be the result of voluminous magmatism around the time of continental breakup.

Dredging of the margins and central part of the Wallaby Plateau obtained altered tholeiitic basalts, and basaltic conglomerates and sandstones, and supports a volcanic origin for the entire feature. Immobile element compositions of two Wallaby Plateau basalts provide strong support for a correlation with those from the Naturaliste and southernmost Kerguelen plateaux, and the eastern Broken Ridge. All these features have been suggested to have formed by plumehead-related volcanism, and all involved, to some degree, Gondwanan sub-continental lithospheric mantle in their petrogenesis. However, it is critical to note that both the basaltic marginal plateaux (the Wallaby and probably the Naturaliste plateaux), and related eastern Indian Ocean LIPs, all appear to be aged from 120-100 Ma, and therefore significantly post-date breakup and accretion of the earliest oceanic crust along the west Australian margin (155–130 Ma). Thus, the plumehead assumed responsible for the formation of these plateaux (and adjacent continental flood basalts, such as the Rajmahal traps and the Bunbury Basalt) cannot be implicated in the rifting and eventual breakup of this margin in the manner proposed by White & McKenzie (1989).

The extensive rift and breakup-related volcanism that occurs in a zone that parallels the west Australian margin appears to have been controlled by dynamic rift processes. It can be readily explained by models involving convective partial melting (Mutter et al., 1988) or dynamic mantle upwelling (Holbrook & Kelemen, 1993) without recourse to mantle plumes. Convective partial melting following a ridge crest jump in the Cuvier Abyssal Plain can also explain the formation of the Wallaby Plateau and probable related features, i.e. the Sonne and Sonja Ridges. However, although such an explanation can account for some of the details in the Wallaby Plateau area, it does not produce a consistent model for all the petrologically and seismically similar post-breakup LIPs of the eastern Indian Ocean.

We believe that the most likely explanation for the temporally and spatially complex arrangement of volcanic features along the west Australian margin and in the adjacent eastern Indian Ocean, involves a combination of non-plume, dynamic, rift-related volcanism leading up to and including breakup, and plume-related, plateau-forming, post-breakup volcanism.

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Appendix

Petrography of the dredged Wallaby Plateau volcanic/volcaniclastic rocks and of samples from the Bunbury Basalt.

Wallaby Plateau samples

Sample 36/4 (i.e. Dredge 36 rock type 4; Table 2) is a very strongly altered and oxidised, formerly plagioclase+olivine+augite-phyric, moderately vesicular basaltic lava. The most abundant phenocrysts are large, subhedral to euhedral, labradoritic plagioclase phenocrysts up to 7 mm long, often partially or totally altered to a very fine-grained clayey aggregate. Former olivine phenocrysts are slightly rounded and up to about 1 mm across, and make up only 1–2 modal % of the rock. They are now altered to colourless chlorite-serpentine with black rims of very fine-grained magnetite(?). Colourless augite phenocrysts and microphenocrysts to about 1 mm across are not uncommon (<5 modal %); the augite is still largely fresh, although many grains show marginal smectite–chlorite alteration. The sample's groundmass displays a highly-altered seriate–intergranular texture composed of small plagioclase laths and subordinate granular, anhedral fresh and altered augite crystals; interstitial glass and glassy mesostatsis are altered to almost-isotropic smectite–clay aggregates, mixed with black manganiferous oxides–hydroxides. Vesicles are lined by a narrow rim of analcime, mantled, in turn, by banded black manganiferous oxide–hydroxides.

Sample 36/3 is an altered, trachytic-textured, sparsely plagioclase-phyric evolved basaltic to trachyandesitic lava containing a few small former olivine phenocrysts altered to iddingsite, and <2 modal % of small (<0.5 mm long) euhedral intermediate plagioclase (An_{30-60}) phenocrysts. The groundmass of this sample was originally glass, charged with tiny plagioclase laths and microlites, but the glass has now been thoroughly altered to a murky, heterogeneous brown material, almost isotropic in places, that is probably composed largely of ultrafine-grained smectite/clay aggregates.

Sample 36/1 is a coarse-grained, framework-supported, volcaniclastic sandstone composed mainly of 1–3 mm sized well-rounded clasts of highly altered, dominantly basaltic lavas. A remarkably diverse array of textural types is present among the lava clasts, many of which are glassy and are now altered. The most common clasts in this rock are plagioclase + (former) olivine-phyric basalts, although aphyric lavas are also well represented. Olivine is always altered to iddingsite, and groundmass augite is always replaced by the same smectite–clay material that replaces glass. The matrix of this sandstone makes up approximately 20–30 modal % rock, and is a very fine-grained, recrystallized mud.

Dredge 38 yielded only one rock suitable for detailed study. Sample 38/1 is a conglomerate made up of clasts of an olivine+augite+plagioclase-phyric basaltic lava. Small former olivine phenocrysts (<4 modal %) are totally iddingsitized. Elongate tabular plagioclase phenocrysts (An_{50-70}) make up less than 2 modal % of the rock. Augite phenocrysts are small and fresh, but show marginal subophitic intergrowth with groundmass plagioclase laths. The groundmass is seriate-intergranular in texture, and dominated by laths of plagioclase separating subordinate small, granular, anhedral augite crystals. Small Fe Ti oxide crystals are common through the groundmass, and interstitial glassy mesostasis has been totally replaced by dark brown, clay–smectite aggregates.

Bunbury Basalt samples

Two samples of the Bunbury Basalt were studied, one (BB1, Table 3) from Bunbury Beach, and the other (BB3) from Gelorup Quarry. The rock is dark-grey, non-vesicular, and porphyritic with a glomeroporphyritic texture. Phenocrysts of plagioclase and occasional augite, and glomerocrysts of both plagioclase and clinopyroxene or both, are set within a finer-grained groundmass which comprises plagioclase, granular augite and Fe-oxides. Elongate plagioclase phenocrysts (An_{64-47}), up to 5 mm long, are both twinned and zoned and have occasional inclusions of fine-grained magnesian pigeonite ($Mg\#_{69-73}$), devitrified melt and groundmass material. Glomerocrysts range from 0.5 to 7 mm across. Some also contain patches of yellow clay (possibly after olivine), which also occurs as discrete anhedral to subhedral patches throughout the rest of the rock.

Clinopyroxene phenocrysts vary from elongate simply twinned crystals up to 2 mm long to euhedral crystals ~1 mm across, and they occasionally contain inclusions of plagioclase (An_{60}). They range in composition and may be predominantly either augite ($Mg\#_{69-78}$) or magnesian pigeonite ($Mg\#_{70-73}$, although one probed crystal has a core of subcalcic-augite ($Mg\#_{71}$) and a rim of augite ($Mg\#_{74}$).

The groundmass comprises elongate plagioclase laths (An_{61-58}) with an average length of 0.01 to 0.1 mm, many of which contain inclusions of clinopyroxene, twinned subhedral to euhedral clinopyroxene crystals (augite $Mg\#_{68-71}$) 0.1 to 0.05 mm in length, and skeletal to dendritic ilmenite and titaniferous magnetite grains, plus abundant interstitial devitrified brown glassy mesostasis.

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Geologic history of the eastern Argo Abyssal Plain based on ODP drilling and seismic data

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A long regional single-fold seismic line (Line 1, Plate 1F) connecting DSDP Site 261 and ODP Site 765 was collected during ODP Leg 123 while transiting the eastern Argo Abyssal Plain (AAP) (Fig. 1). Based on various velocity data collected at Site 765, an excellent correlation was obtained between the seismic data and the drilling results, which allowed for a regional extrapolation of the drilling results and a preliminary interpretation of seismic Line 1. This interpretation provides several new observations about the geologic history of the eastern AAP:

- A broad regional basement high progressively onlapped by sediments from the north and south has characteristics of an ocean spreading centre. This distribution, however, does not fit the interpreted magnetic anomalies, which show the crust getting progressively younger to the north.
- Large faulted basement highs in oceanic crust are interpreted to be manifestations of a major group of northwest-trending oceanic fracture zones that are interpreted to offset magnetic anomalies in the eastern AAP.
- Faults flanking these highs have been reactivated, indicating regional intraplate stresses have affected the oceanic areas at later times.
- A broad uplifted area in the northeastern AAP provides the tilted surface onto which Cainozoic turbidites progressively thin and onlap. This uplift represents a regional tectonic event of early Tertiary or, possibly, Early Miocene age. This event may be related to subduction along the Sunda–Java trench to the north.
- Early Cretaceous claystone deposition was generally quite uniform across the entire eastern AAP.
- Deep-sea depositional moats (cut-and-fill structures) along the flanks of basement highs within the Early Cretaceous section indicate strong deep-sea current systems had become established in the deep AAP, possibly signaling regional changes in palaeoceanography.
- Generally starved depositional conditions persisted across the eastern AAP for over 40 Ma from Late Cretaceous through Eocene.
- Major carbonate turbidite deposition sourced from the adjacent plateaus began in Middle Miocene in the southeastern part of the basin, onlapping the uplifted northern part of the area at site 261, where turbidite sedimentation was generally absent. A major change in turbidite sedimentation and source area took place in Late Miocene.

Introduction

A long regional singlefold seismic line (Line 1, Plate 1F), connecting DSDP Site 261 and the to be drilled ODP Site 765, was collected during ODP Leg 123 while transiting across the Argo Abyssal Plain (AAP) (Fig. 1). Based on various velocity data collected at Site 765, an excellent correlation can be made between the seismic data and the drilling results, allowing a regional extrapolation of the drilling results along seismic Line 1 between sites 765 and 261. The purpose of this report is to discuss: (1) the correlation of the drilling results at Site 765 with key horizons on ODP seismic Line 1; (2) the extrapolation of these dated horizons along Line 1 to Site 261; and (3) some observations and interpretations from Line 1 and their implications for the geologic history of the eastern Argo Abyssal Plain. (See Plate 1 in pocket.)

Background

The Argo Abyssal Plain is a deep ocean basin located off northwest Australia (Fig. 1). It is bordered on the south and east by the Exmouth Plateau and Scott Plateau, two large rifted and subsided continental blocks. On the west are the Joey Rise and Roo Rise, large oceanic volcanic structures. To the north, the basin is being subducted along the Java Trench (Fig. 1). The AAP is underlain by a generally flat-lying Late Jurassic and younger sedimentary package up to 1000 m thick overlying a Late Jurassic–Early Cretaceous oceanic crust.

The AAP ocean basin formed as a large continental block broke away from Australia and drifted northwestward during the Late Jurassic and Early Cretaceous (e.g. von Rad et al., 1989). This age and direction of rifting and seafloor spreading has been documented by the studies of marine magnetic anomalies (Larson, 1975; Heirtzler et al., 1978; Fullerton et al., 1989; Sager et al., 1992), which suggest initiation of seafloor spreading at magnetic

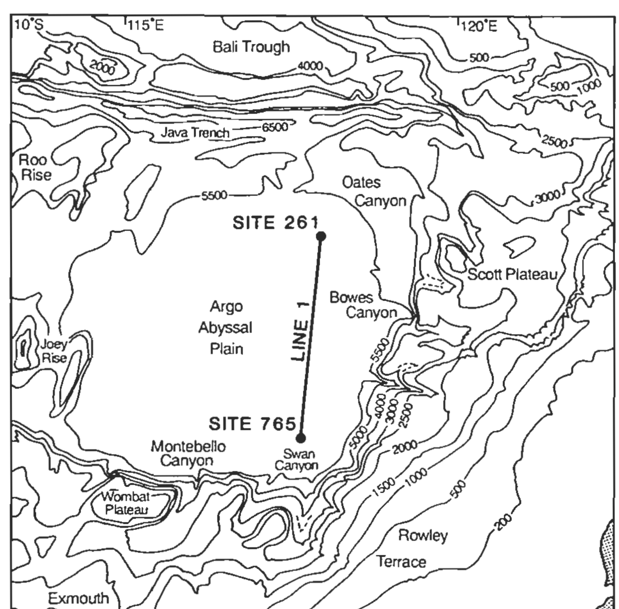

Figure 1. Bathymetric map of the Argo Abyssal Plain and surrounding areas showing major features and location of seismic Line 1 connecting DSDP Site 261 and ODP Site 765. Contour interval in metres. From Gradstein et al. (1992).

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anomaly M–26 (approximately 163 Ma) (Fig. 2). The sediments in the AAP were first sampled during DSDP Leg 27 at Site 261 (Fig. 1), where they consist of a thin section of Late Cainozoic ooze and clay unconformably overlying a thicker section of Late Jurassic–Cretaceous claystone (see left side of Plate 1) (Veevers et al., 1974). Site 261 bottomed in Late Jurassic oceanic crust (Fig. 2). Further investigations of the sediments underlying the AAP were made later in the 1970s based on regional singlefold seismic data tied to the drilling results, and a generalized seismic stratigraphy of the basin was developed (Cook et al., 1978; Heirtzler et al., 1978; and Hinz et al., 1978).

In 1988, ODP Leg 123 drilled Site 765 and sampled a more complete section of the sediments in the AAP as well as the underlying oceanic crust (Figs 1 and 3) (Ludden et al., 1990; Gradstein et al., 1992). The 931 m of sedimentary rocks include approximately 10 m of latest Jurassic claystone, 320 m of Early Cretaceous claystone, a condensed (starved) 100 m Late Cretaceous–Paleogene section of claystone and carbonate turbidites, and a thick section (400 m) of Neogene carbonate turbidites and debris flows (Fig. 3). The site was drilled approximately at magnetic anomaly M–25A (162 Ma, Oxfordian; Sager et al., 1992) (Fig. 2). This age and the age of celadonite veins in oceanic crust (156 Ma; Ludden, 1992) are older than the oldest sediments overlying oceanic crust (Tithonian, 150 Ma) (Fig. 3), suggesting a major hiatus and long period of sediment starvation in the newly formed deep basin.

Correlation with seismic line 1

Seismic Line 1 was collected by the *JOIDES Resolution* during the transit across the AAP enroute Site 765 (Fig. 1). It was begun at DSDP Site 261, so that a correlation could be established between the two sites. The interpreted line is presented as Plate 1F (in pocket). It is a singlefold line collected using two 80 cubic inch water guns at speeds ranging from 5–8 kts. The data were collected digitally and later processed by ODP at Texas A&M University. Details of the collection and processing of the seismic data are included with the line on Plate 1F.

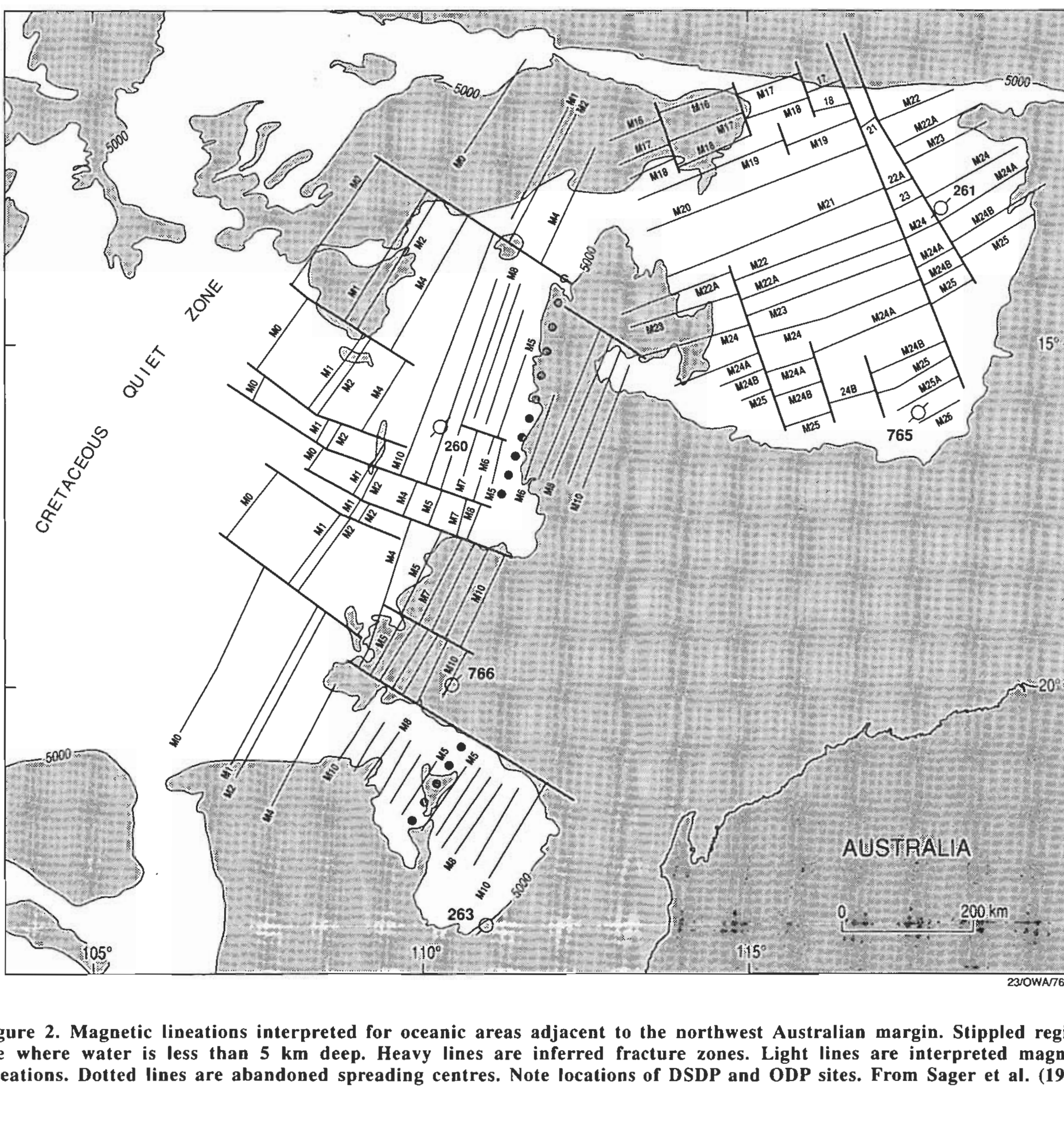

Figure 2. Magnetic lineations interpreted for oceanic areas adjacent to the northwest Australian margin. Stippled regions are where water is less than 5 km deep. Heavy lines are inferred fracture zones. Light lines are interpreted magnetic lineations. Dotted lines are abandoned spreading centres. Note locations of DSDP and ODP sites. From Sager et al. (1992).

Detailed velocity data were collected at Site 765 by a variety of methods, including shipboard measurements and velocity logs (Ludden et al., 1990), a VSP experiment (Bolmer et al., 1992), and sonobuoys (Lizarralde & Buffler, 1992). These data all showed a comparable velocity structure for the site (Bolmer et al., 1992; Lizarralde & Buffler, 1992). For the purpose of this report the results from the VSP experiment were used to correlate the drilling results with the seismic line (Fig. 4). Here the first arrival data from shots recorded at various depths in the hole (solid line) allow a direct conversion from depth to two-way travel time (Fig. 4). Both the seismic line and the VSP experiment were recorded using the same sound source (two 80 cubic inch water guns).

The seismic sequences shown on Figure 4 and Plate 1F (1–7) were determined from a preliminary analysis of the seismic data at Site 765 and were reported in the Initial Reports (Ludden et al., 1990). Their boundaries are used here for reference (Fig. 4; Plate 1F). For the purpose of correlating across Line 1 to Site 261 in this report, however, a new set of seismic reflections have been picked (A–F). These reflections are regionally continuous and they can be correlated easily across the entire seismic line (Plate 1F). They also represent major changes or breaks in sedimentation and geology. Ages have been assigned to each based on the correlation with the well (Fig. 4). Each reflection (A–F) is discussed briefly below.

A Approximately 189 m. Latest Miocene. Occurs at boundary between seismic sequences 2 and 3, which marks a downward change from alternating high/low amplitude, continuous seismic facies to a low-amplitude, continuous facies. This probably reflects a change from a more clayey Plio-Pleistocene section (Lithology Unit I) to a more uniform carbonate turbidite section (Lithology Unit II).

B Approximately 280 m. Middle Late Miocene. Occurs at boundary between seismic sequences 3 and 4, which correlates with a regional unconformity seen on nearby seismic lines and reflects a major change in turbidite source into the basin. It also corresponds to a major bedding break in carbonate turbidites (Simmons, 1992).

C Approximately 379 m. Middle Miocene. Corresponds to a change back to alternating high/low amplitude, continuous seismic facies, which probably reflects an increase in clay content (Lithology Unit IIA/IIB boundary).

D Approximately 500 m, near 5/6 sequence boundary. Oligocene by direct correlation (Fig. 4), but it may be equivalent to major gap/hiatus between Oligocene and Eocene strata at 520 m, which reflects the top of a major Late Cretaceous–Eocene condensed interval.

E Approximately 592 m. Early Cretaceous/Late Cretaceous boundary. Top of Early Cretaceous claystone section and base of Late Cretaceous–Eocene condensed interval, which consists of alternating claystone and carbonate turbidites/debris flows (alternating high/low amplitude facies).

F Approximately 931 m. Top Late Jurassic oceanic crust. Irregular and faulted high amplitude reflection onlapped by Late Jurassic–Early Cretaceous claystones.

Observations, interpretations and regional implications

Oceanic Crust

The top of oceanic crust (F) regionally is an irregular surface (Plate 1F), not unlike the top of oceanic crust in other ocean basins. It is overlain by a Late Jurassic–Early Cretaceous section that infills the lows and onlaps and thins against basement highs. The major highs fall into two categories:

- A broad regional high centred around 1200 h (6. September, 1988) onto which sediments gradually onlap and thin from both the north and south. This high has the characteristics of an oceanic spreading centre with sediments getting younger toward the crest. This observation, however, does not fit with the interpreted magnetic anomalies (Fig. 2), which show a crust getting progressively younger to the north.
- At least five large, high-relief, fault-bounded highs that occur along the northern half of the line between 2015 h (5. September, 1988) and 0700 h (6. September, 1988). The highs occur directly along the projection of a major group of northwest-trending oceanic fracture zones that offset magnetic anomalies in the eastern AAP (Fig. 2). These highs are interpreted to be manifestations of these fracture zones and represent original topography on the seafloor. Some of the highs, however, appear to have been uplifted at a later time, as evidenced by the faulted and uplifted overlying Cretaceous sediments (e.g. 2100 h; 5. September, 1988 and 2300–0000 h; 6. September, 1988). This implies a later reactivation of these fracture-zone faults due to later intraplate stresses that affected the oceanic areas. The age of the faulting is discussed later.

Elsewhere, the top of oceanic basement is offset by both minor and major faults. For example, a major fault offsetting basement and the overlying Lower Cretaceous rocks by about 300 m occurs just north of Site 765 (1600–1700 h; 6. September, 1988) (in a gap in the seismic record). The fault and Early Cretaceous sediments are truncated by reflection E, which dates the movement as mid-Cretaceous.

Along the north end of the line, basement rises regionally from a general depth of about 8.5 sec (two-way travel time, TWTT) to less than 8.0 sec near Site 261. This uplifted area, that includes the overlying Cretaceous sediments, provides the tilted surface onto which the Cainozoic carbonate turbidites progressively onlap. This is why much of the Cainozoic section is absent at Site 261. This uplift represents a regional tectonic event in the northern AAP, as similar relationships are observed on other regional singlefold seismic lines available in the area. This event is probably at least as young as early Tertiary in age, as Cretaceous rocks are uplifted. It may, however, be younger as discussed below. The distribution and origin of this uplift is the subject of ongoing studies using the available seismic data in the area.

Early Cretaceous

The Early Cretaceous claystone section (F–E) was drilled at both Sites 261 and 765 (Figs 3, 4; Plate 1), where it is approximately the same thickness (about 340 m). This suggests relatively uniform Early Cretaceous sedimenta-

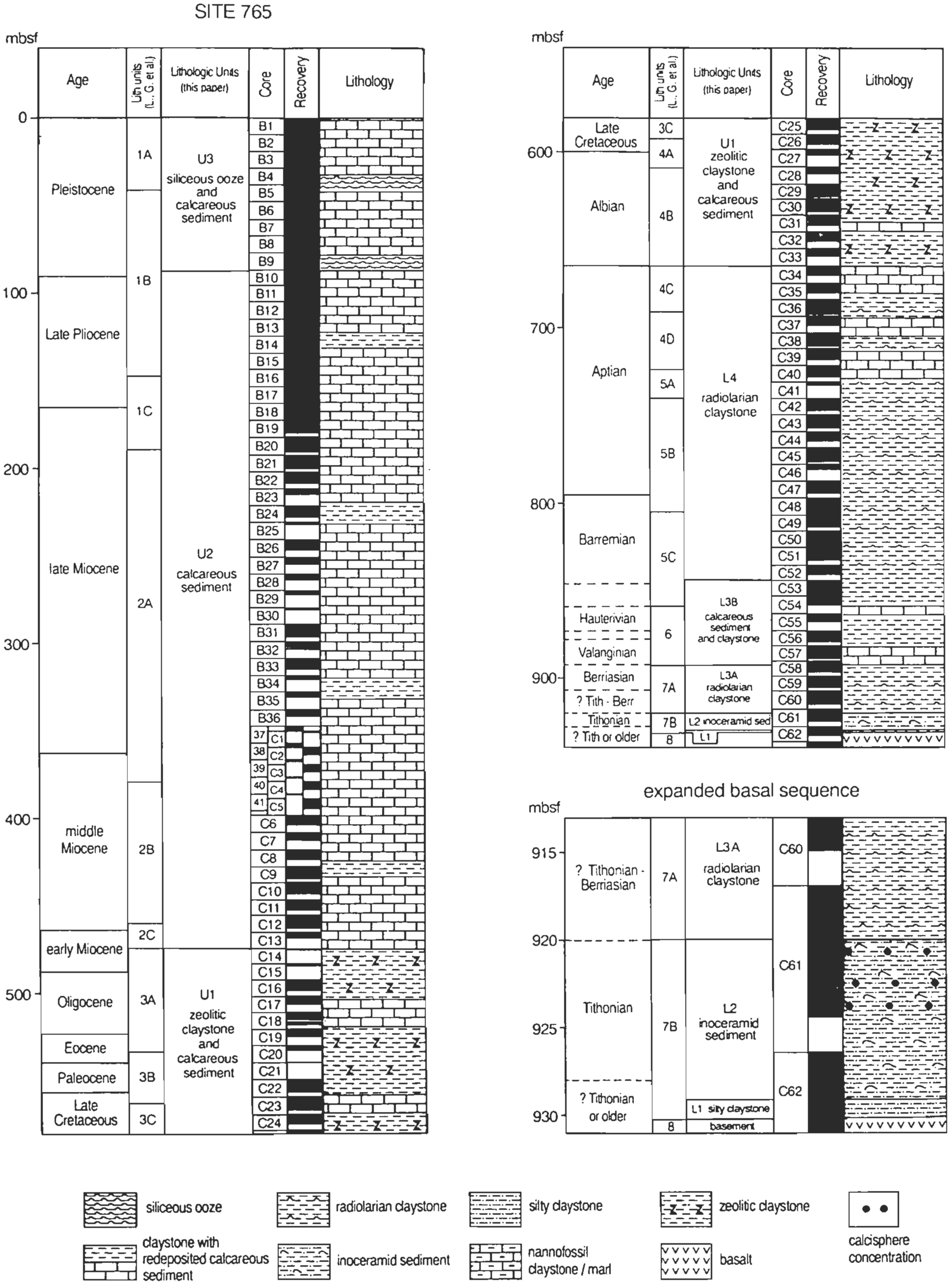

Figure 3. Stratigraphic section at Site 765 overlying oceanic crust. Depths are in metres below seafloor. From Gradstein et al. (1992). Note thin or condensed Late Jurassic section overlying inferred Oxfordian age oceanic crust based on magnetic anomalies (Fig. 2).

tion across the entire eastern AAP. The section thins depositionally by onlap fill onto basement highs. For example, the sequence thins to about 150 m over the broad regional high centered at 1200 h (6. September, 1988). It also fills in between and thins along the flanks of the large fracture-zone highs. In some places, the unit thins by fault uplift and truncation below unconformity E (e.g. along north end of line plus just north of Site 765, as discussed above). Also, as mentioned above, the unit has been uplifted and partially eroded from the tops of some of the reactivated fracture-zone highs (e.g. 2300–0000 h; 6. September, 1988).

The claystone lithology of the unit is reflected by the generally low-amplitude, continuous seismic facies typical of much of the line (e.g. near Site 261 and Site 765). In the middle of the line, however, the reflections have a more discontinuous, mounded and overlapping character with some higher amplitudes (0500–1100 h; 6. September, 1988). This may be due to local deformation or disruption of the sediments by decompaction or other tectonic events. Alternatively, this facies may represent a setting more proximal to the source, and represent coarser sediments deposited in channels and/or overlapping depositional lobes. Along the southern end of the line near Site 765, the upper part of the section is characterized by more high-amplitude continuous reflections, which may be reflecting the increase in carbonate turbidites identified at Site 765 (Fig. 3).

Perhaps the most interesting aspect of the Early Cretaceous section is the depositional moats observed along the flanks of several basement highs (e.g. 2015 h, 2100 h, 2230 h; 5. September, 1988). The one at 2015 h occurs in the middle of the section, while the other two occur toward the top of the section. These cut-and-fill features are observed in deep-sea sediments throughout the world and are generally attributed to strong and focused deep-sea current systems. Evidently, such currents became established along the flanks of the fracture-zone highs during the Early Cretaceous, and they may reflect some major change in palaeoceanography related to major plate reorganizations. The regional significance of these currents is being investigated further as part of a more regional study of this area.

Late Cretaceous–Eocene

The Late Cretaceous through Eocene section (E–D) represents an 80 m condensed interval consisting of alternating claystones and calcareous turbidites/debris flows. This lithology is reflected in the alternating high/low amplitude, continuous seismic facies. The unit is quite uniform in thickness and facies across the entire area, indicating quite uniform depositional conditions for a long period (over 40 Ma). The unit appears to onlap and thin depositionally along the flanks of the large fracture-zone highs and pinches out at approximately 0000 h (6. September, 1988). It is difficult to correlate across the high to the north, where it may form a low-amplitude, transparent unit that conformably overlies the deformed Early Cretaceous section. At Site 261, an 80 m? thick, Late Cretaceous brown claystone was drilled (see left side Plate 1F), suggesting that the unit is present here and has been uplifted along with the Early Cretaceous. This dates the regional uplift in the northern AAP as at least early Tertiary.

Oligocene–Middle Miocene

The Oligocene through Middle Miocene section (D–C) represents the real onset of carbonate turbidite deposition in the basin, which is mixed with background claystone deposition. This alternating lithology is reflected in the characteristic alternating high/low amplitude, continuous seismic facies of this unit. The Oligocene–Early Miocene lower part of the unit represents a more condensed interval with more clay, while the Middle Miocene upper part represents a major influx of the carbonate turbidites (Figs 3,4).

Regionally, the unit thickens slightly from about 140 m at Site 765 to about 180 m in the basins between two large basement highs at 0200–0330 h (6. September, 1988). This change occurs mainly due to an expansion of the Lower Oligocene–Early Miocene section. This lower part pinches out along the flanks of the two highs, while the upper part extends over the tops of the highs. Along the flank of the next high to the north, the unit completely pinches out (about 0000 h; 6. September, 1988). It appears again to the north of the high before pinching out entirely along the flank of the regionally tilted Cretaceous section. These turbidites, therefore, never reached the uplifted area at Site 261, or they may be represented by a thin pelagic condensed section that was not sampled.

There is one significant location that bears on the timing of the regional uplift along the north end of the line. At 0015 h (6. September, 1988), the lower Oligocene–Early Miocene part of the section appears to be tilted up and truncated along with the Cretaceous section, while the Middle Miocene onlaps over the truncated surface. This suggests that some of the uplift may be as young as Early Miocene, much younger than the early Tertiary suggested above. This regional uplift may be related to regional tectonic events, including subduction along the Java Trench just to the north, which could have generated intra-plate stresses. This subduction may have begun as early as latest Eocene (40 Ma) (Bellon et al., 1989) and continues today.

Middle–Late Miocene

The Middle to latest Miocene section (C–A) consists mainly of carbonate turbidites, which were sourced from the outer parts of the Scott and Exmouth Plateaus and fed into the deep basin through canyon systems along the margins. This depositional mode accounts for the unit's more uniform and low-amplitude, continuous seismic facies along most of the line. To the north, the facies becomes somewhat more higher amplitude in nature, possibly suggesting an increase in clay interbeds. The unit is flat-lying and maintains a relatively uniform thickness, thickening only slightly from about 200 m at Site 765 to about 250 m at the north end of the line, before pinching out against the regional high. Only the uppermost part of the unit is present at Site 261, where it is represented by only one sample of ooze (Plate 1F).

The unit is separated by a prominent basinwide reflection (B) that marks a regional unconformity in the southeastern AAP. The unconformity is conformable on Line 1, but is recognized by truncation and onlap on nearby singlefold lines. This unconformity suggests a major shift in the source of turbidite input into the basin. Details of this change await further study. This surface also corresponds to a major bedding change in carbonate turbidites recognized by Simmons (1992) at Site 765.

Late Miocene–Pleistocene

The latest Miocene through Pleistocene section (A-seafloor) represents an increase in clay content and carbonate debris flows, which are mixed with the carbonate turbidites. This change in lithology is reflected by the change back to alternating high/low amplitude, continuous seismic facies. The unit thins gradually across the basin from 190 m at Site 765 to about 100 m at Site 261. This thinning must also include an overall facies change from mainly turbidites at Site 765 to mainly claystone above

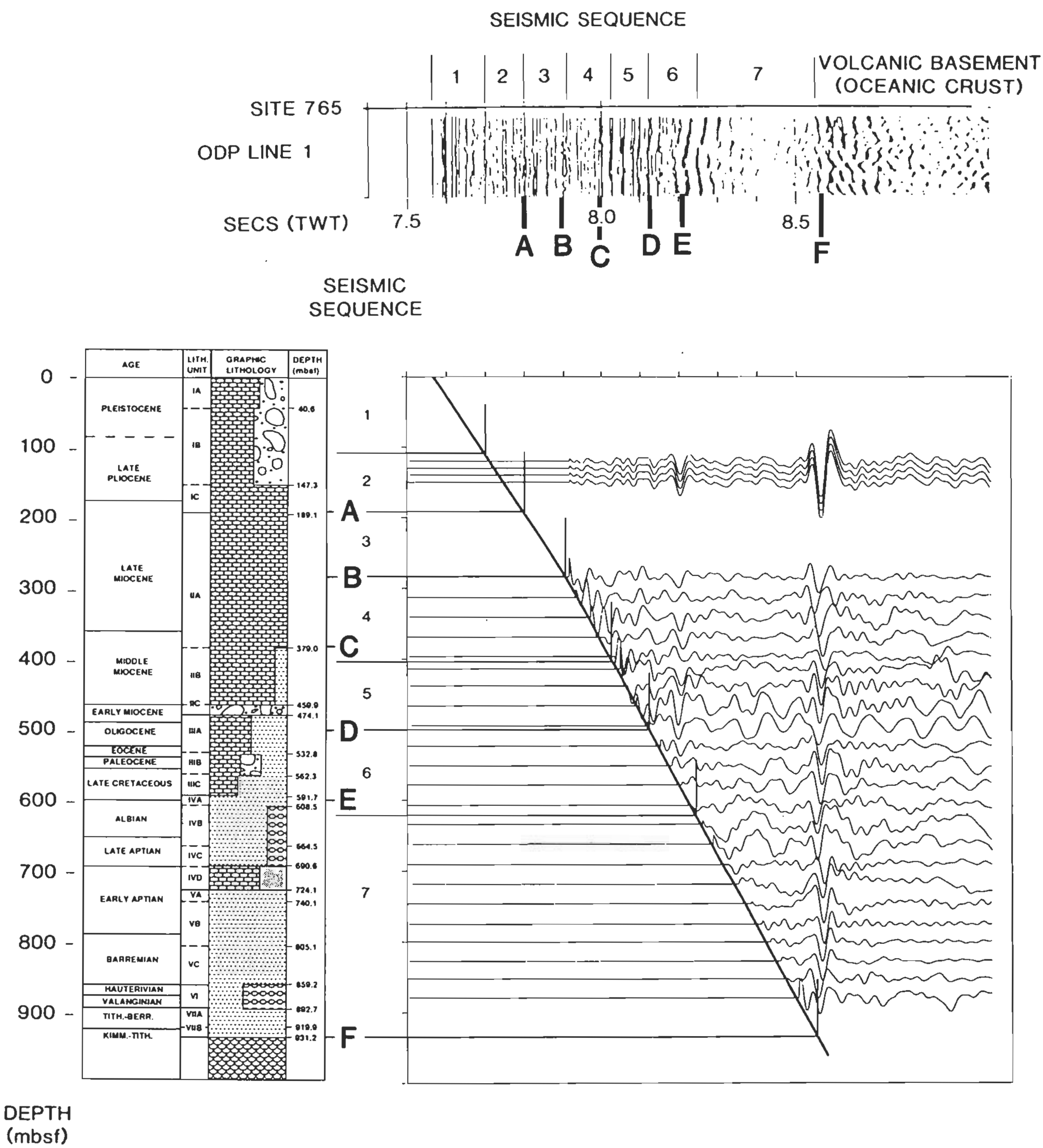

Figure 4. Diagram showing results of Vertical Seismic Profile (VSP) experiment at ODP Site 765. Stacked traces are from shots recorded at various locations within hole. Traces at the top of diagram are a composite of all the stacked traces. The trace of first arrivals (dark line) is used to correlate between drilling results (in depth) and portion of ODP seismic Line 1 at Site 765 (in two-way travel time). Sound source for both the VSP experiment and seismic line was the same (two 80 cubic inch water guns). From Bolmer et al. (1992). Letters designate major reflections and associated sedimentary changes used for correlation purposes along seismic Line 1 (Plate 1F).

the regional high at Site 261. This is not reflected in an obvious seismic facies change, but there is a subtle change at approximately 0000–0200 h (6. September, 1988) to a higher frequency, higher amplitude facies in the north. Thus, the regional high continued to exert a subtle influence on sedimentation.

Summary and conclusions

A long regional singlefold seismic line (Line 1, Plate 1F) connecting DSDP Site 261 and ODP Site 765 was collected during ODP Leg 123 while transiting across the eastern Argo Abyssal Plain (AAP) (Fig. 1). Based on various velocity data collected at Site 765, an excellent correlation was made between the seismic data and the drilling results, which allowed for a regional extrapolation of the drilling results and a preliminary interpretation of seismic Line 1. This interpretation provides several new observations about the geologic history of the eastern AAP:

- A broad regional basement high progressively onlapped by sediments from the north and south has characteristics of an ocean spreading center. This distribution, however, does not fit the interpreted magnetic anomalies, which show the crust getting progressively younger to the north.
- Large faulted basement highs in oceanic crust are interpreted to be manifestations of a major group of northwest-trending oceanic fracture zones that are interpreted to offset magnetic anomalies in the eastern AAP.
- Faults flanking these highs have been reactivated, indicating regional intraplate stresses have affected the oceanic areas at later times.
- A broad uplifted area in the northeastern AAP provides the tilted surface onto which Cainozoic turbidites progressively thin and onlap. This uplift represents a regional tectonic event of early Tertiary or, possibly, Early Miocene age. This event may be related to subduction along the Sunda–Java trench to the north.
- Early Cretaceous claystone deposition was generally quite uniform across the entire eastern AAP.
- Deep-sea depositional moats (cut-and-fill structures) along the flanks of basement highs within the Early Cretaceous section indicate strong deep-sea current systems had become established in the deep AAP, possibly signaling regional changes in palaeoceanography.
- Generally starved depositional conditions persisted across the eastern AAP for over 40 Ma from Late Cretaceous through Eocene.
- Major carbonate turbidite deposition sourced from the adjacent plateaus began in Middle Miocene in the southeastern part of the basin, onlapping the uplifted northern part of the area at site 261, where turbidite sedimentation was generally absent. A major change in turbidite sedimentation and source area took place in Late Miocene.

It must be remembered that these observations and interpretations of Line 1 represent only a 2-dimensional look at the eastern AAP, and must be considered somewhat preliminary. In the near future, as part of an ongoing regional study of the area by the University of Texas and AGSO, these interpretations will be further extrapolated throughout the region using all available seismic data. This will provide a more detailed and complete interpretation of the geologic history of the entire eastern AAP.

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Tectonic and sedimentary history of the Argo Abyssal Plain, eastern Indian Ocean

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Bathymetric and seismic reflection profiles totalling about 2800 km, and covering about 110 000 km^2 of the Argo Abyssal plain, have been analysed. The early emplacement of oceanic crust and volcanic edifices (Late Jurassic and Early Cretaceous) was followed by cooling, and marked subsidence until the Miocene. The principal structural features have been identified as the right-laterally sheared Wombat Graben in the south, the N–S trending Joey Rise in the west, the parallel sediment-filled Joey Graben to its east, a NW-trending eastward tilted and subsided volcanic massif to its east, a NW-trending linear volcanic massif on the old oceanic plate farther east, the deformed and uplifted sediments and basement in the north, and the central basin with 1.2 km thick sediments. The Joey, Wombat and Kivi Grabens are newly identified.

The N–S trending Joey Graben is 400–450 km long, parallel to and east of the Joey Rise, and it is filled with up to 1 km of sediment. Beneath it, the mantle has been identified at depths of 12–15 km and deepening westward. The other two grabens form moats hundreds of metres deep, at the foot of the Australian margin in the south. The right-lateral displacements in the axis of the Wombat Graben reflect the initial shearing and rifting of the thinned crust. The widespread faulting in the basement, and the persistence of some faults upward into sediments as young as early Miocene, indicate that tectonism was marked until then. The eastern and northern parts of the abyssal plain have been uplifted as they came within the peripheral fore-bulge south of the Sunda Trench.

Five sequences (A–E) overlie the oceanic crust and they have been correlated with the lithology of ODP Site 765 and DSDP Site 261. They probably represent Plio-Pleistocene, upper Miocene–lower Pliocene, middle Miocene, upper Oligocene–lower Miocene and uppermost Jurassic to Lower Cretaceous strata, respectively. The depositional trends in the Mesozoic and Miocene sediments are controlled by basement structure. Isopach maps of the Mesozoic and Cainozoic sediments reveal that their maximum thicknesses are about 0.6–0.8 km and 0.2–0.4 km, respectively, with the bulk of the sediments (calcareous turbidites and claystone) laid down in the Early Cretaceous and the middle Miocene to Pliocene. During the Oligocene to early Miocene period (37–16 Ma), a remarkable geologic event led to non-deposition or erosion of sediments.

Introduction

The Argo Abyssal Plain in the northeastern Indian Ocean is bounded by the Australian continental marginal plateaus (Exmouth/Wombat and Scott) in the south and east, the eastern Sunda Arc–Trench in the north and the Joey and Roo Rises in the west (Fig. 1). The plain's oceanic crust is the oldest, 108–156 Ma, in the Indian Ocean. Volcanic rises separate the Argo Abyssal Plain from the Gascoyne Abyssal Plain to the west. The plain is a quadrangle with water depths generally exceeding 5000 m, and is a depositional plain with a regional inclination down to the southwest. Important papers dealing with the plain include Cook et al. (1978), Heirtzler et al. (1978), Hinz et al. (1978), Powell & Luyendyk (1982), Veevers et al. (1985), Powell et al. (1988), Fullerton et al. (1989), Pillipenko & Sivuka (1990), Exon et al. (1982), and Gradstein & von Rad (1991). Other important papers are included in the DSDP and ODP volumes of Veevers et al. (1974); von Rad et al. (1992), and Gradstein et al. (1990, 1992).

Mesozoic M10–M25 magnetic anomalies are aligned N 70°E, and offset by northwest-trending fracture zones (Powell & Luyendk, 1982; Fullerton et al., 1989). The crust was drilled at Deep Sea Drilling Project (DSDP) Site 261 and Ocean Drilling Program (ODP) Site 765 (Figs 2 and 5), and the site investigations revealed the litho-, bio- and magneto-stratigraphy of the sediments, and the magneto-stratigraphy of the igneous basement. In the south, ODP Site 765 identified 155 Ma oceanic basement, as compared to the older age (160 Ma, pre-Oxfordian) of the basal sediments in the north, identified at DSDP Site 261. Thus, both site investigations confirm the Jurassic opening of the Indian Ocean, but some uncertainty about the exact age of initial opening of the plain remains. The present geomorphological and seismic investigations were done to map structural elements, the nature and temporal distribution of sediments, and the tectonics of the Argo Abyssal Plain region. Study of the sediments helps decipher the ongoing tectonic processes at the time of their deposition.

The study area of about 110 000 km^2 covers the Argo Abyssal Plain and lies in water depths exceeding 5000 m (Fig. 1). The various morphological units of the plain and the bordering continental margin (plateaus, rises and ridges) were identified from the bathymetric and seismic reflection studies. The data used in this study are six north–south and nine east–west profiles, totalling about 2800 line km (both multichannel and single-channel seismic) within the Argo Abyssal Plain, and carried out by the Soviet Research Vessels "*Issledovatel*", "*Yuzhmorgeologiya*", and "*17 Syezd Profsoyuzov*" during 1986–1988 under the auspices of the Trans Indian Ocean Geotraverse (Fig. 2). Seismic sequences were identified using the reflection patterns on seismic records (Figs 3 and 4), and correlated with the sequences in ODP Site 765. The chronostratigraphy of the sequences is interpreted by correlation with the DSDP and ODP results. Isopach and structure contour maps were prepared from the interpreted seismic profiles.

Regional distribution of seismic sequences in the Argo Abyssal Plain

Five seismic sequences have been recognised (A–E, Figs 3 and 4 A–D). Ages and average thickness are shown in Table 1. This study is complementary to that of Buffler (1994 — see this issue), who has correlated ODP Site 765 and DSDP Site 261 in the eastern Argo Abyssal Plain with key horizons and sequences on a seismic line that connects the sites. A correlation of Buffler's seismic sequences with ours is shown in Figure 5.

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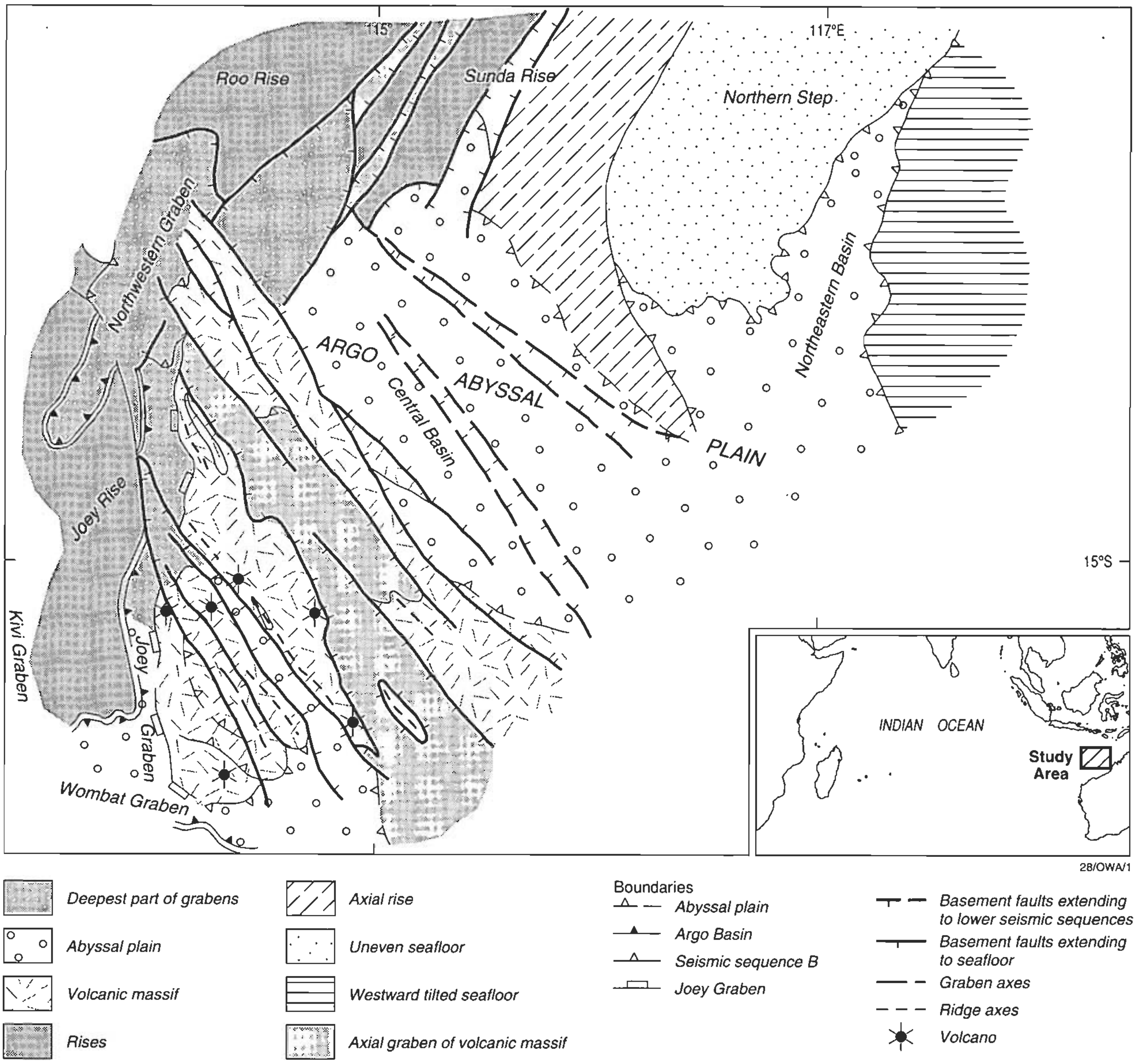

Figure 1. Geomorphology of the Argo Abyssal Plain.

Table 1. Seismic sequences.

Letter	*Biostratigraphic age*	*Average thickness in depocentres (m)*	*Seismic character*
A	L Pliocene — Recent	150	Well-stratified
B	L Miocene –E Pliocene	150	Well-stratified, semi-transparent
C	M Miocene	300	Well-stratified, semi-transparent
D	Oligocene –E Miocene	150	Well-stratified
E1	Palaeogene	200	Well-stratified, semi-transparent
E2	L Jurassic –Cretaceous	200	Hummocky clinoforms

The reflectors of the seismic sequences A to D are parallel laminated; the lower sequence E is subdivisible into a lower discontinous sequence E2 which has hummocky clinoforms, and an upper ubiquitous sequence E1 which is continuously stratified and acoustically semi-transparent. Several broad submarine depressions, filled with Late Miocene and Quaternary sediments (sequences A and B) were reported by Cook et al. (1978). The present study, of the morphology of basement (Fig. 6) and the surface of the plain (Fig. 1), delineates the major structural elements (Fig. 7): Wombat Graben, Joey Graben, northern and eastern steps; central basin; volcanic massif and Sunda Rise. The thickness of the various sequences in the different structural elements is summarised in Table 2.

The total sediment thickness within the plain is generally 1.0–1.2s two-way time (TWT) (Fig. 8). The Mesozoic and Cainozoic sediments are about 0.8s and 0.4s in maximum thickness (Figs 8 and 9). The major morphological units on the plain are described below.

The Wombat Graben

The newly defined sediment-filled Wombat Graben, at the northern foot of the Wombat Plateau of the northwestern Australian shelf base, is about 35 km long, 20 km

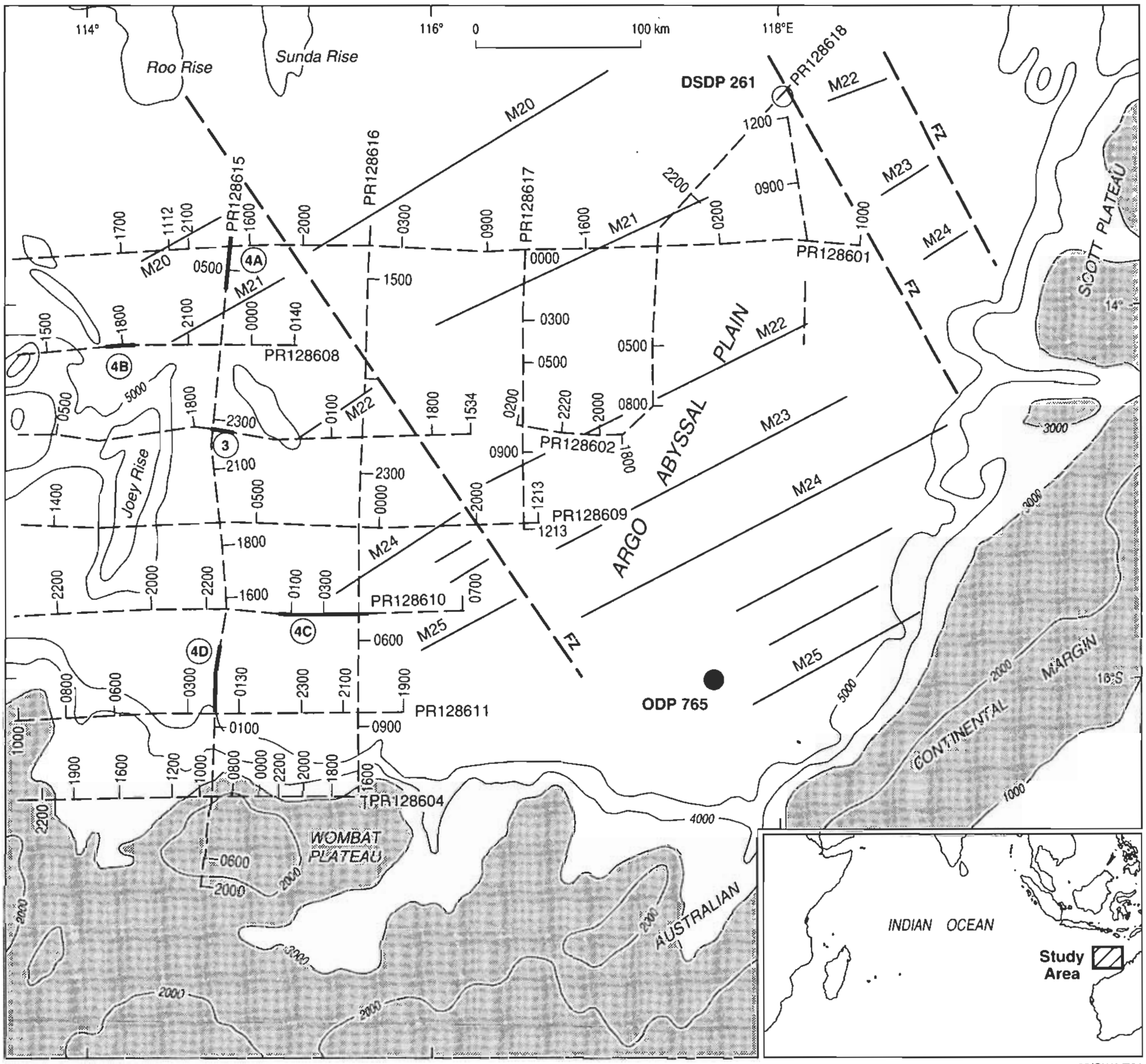

Figure 2. Generalised bathymetry (in metres) of the Argo Abyssal Plain area and index map of the study area. Tracks of seismic reflection and bathymetric investigations (dashed lines). Very thick lines refer to parts of seismic profiles shown in Figures 3 and 4. Thick dashed and continuous lines represent fracture zones and magnetic chrons respectively (after Powell & Luyendyk, 1982). Open and closed circles represent DSDP and ODP drill sites, respectively. Stippled areas mark the identified continental crust (after Veevers et al., 1985).

Table 2: Sediment thickness of Argo Abyssal Plain structures.

Structure	*Thickness of seismic sequences (in m)*					*Total sediment thickness (m)*
	[A]	*[B]*	*[C]*	*[D]*	*[E]*	
Wombat Graben	30–130	70–110	60–210	100–300	750–850	1200
Joey Graben	130–160	60–200	0	140–190	230–250	750–900
The western and eastern parts of volcanic massif	100–160	70–200	0–150	0–290	0–300	170–950
Axial graben	170–200	140–200	160–220	150–210	140–460	1140–1180
Central basin	120–240	220–270	170–270	150–200	120–500	1160–1200
Axial graben in central basin	100–160	210–220	150–160	120–220	0–250	610–650
Northern step	70	100	0	0	250	420
Eastern step	170–190	100	100–160	0–120	120–300	500–600
Average thickness in the grabens and depressions	140	130	220	145	376	921
On basement highs	116	145	130	110	125	626

Note: Thickness in metres estimated using average velocity of 2.2 km/s in sediments (after Bolmer et al., 1992).

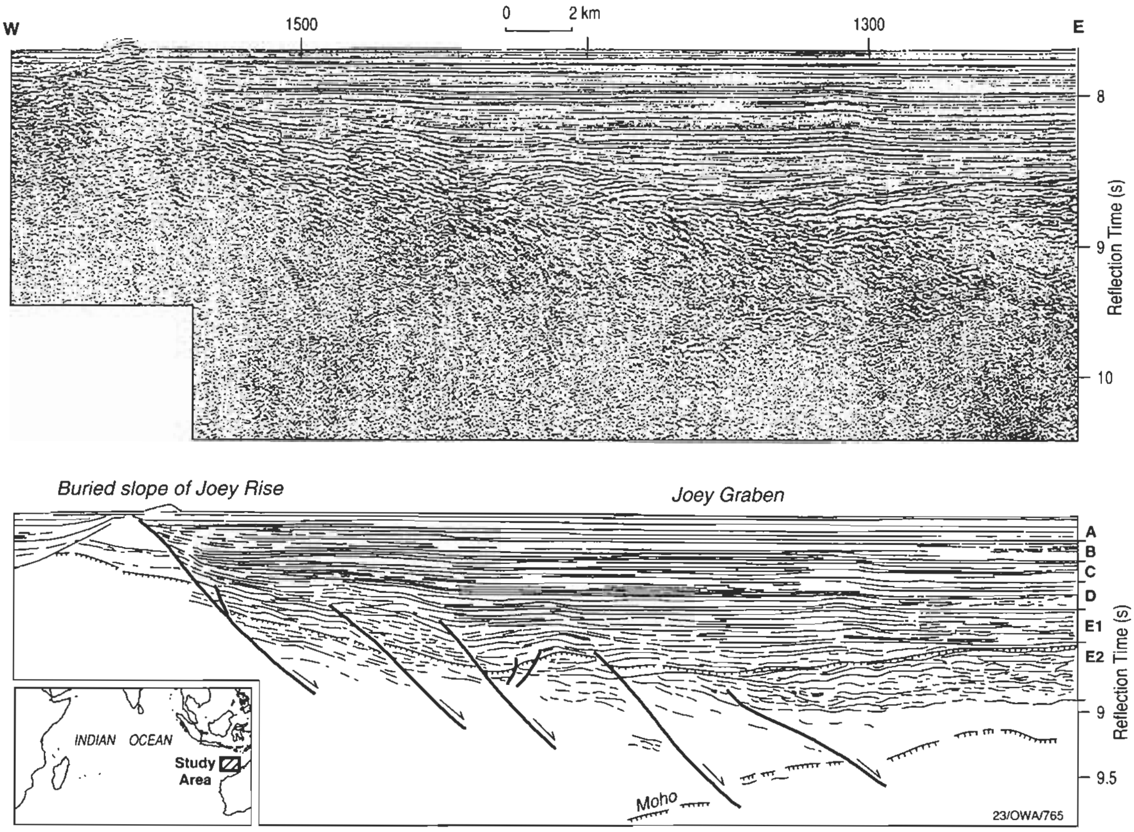

Figure 3. Multichannel reflection seismic profile 128602, east-west across the Joey Graben, and interpreted line drawing. Shows seismic sequences A, B, C, D, E_1, E_2; and the interpreted Mohorovicic discontinuity dipping westward toward the Joey Rise. Location in Figure 2.

wide and several hundred metres deep (Figs 4 and 9 D). It trends almost WNW close to the edge of the Argo Abyssal Plain, and is bounded by the Wombat Plateau in the south and a basement high in the north. There are several elongated, 10–12 km long, basement depressions reaching depths of 6800–7000 m, within the central graben and bounded by basement faults. These trend northwesterly, oblique to the graben edge, and are right-laterally displaced (Fig. 6). Gradual deepening and thickening of seismic sequence E occurs towards the north (Fig. 4D); its reflectors are nearly parallel and acoustically semi-transparent. Sequence D is absent in the axial graben, and reflectors in sequences C to A onlap underlying sequences and basement. The older sequence E gradually thins southward up the plateau slope (Fig. 4D). In the graben, axis sediments are up to 1.2s thick. The NNE-trending Kivi Graben (Figs 6 and 7), separated from the Wombat Graben by the Joey Rise, is also a newly defined similiar feature that lies NW of the Wombat Graben, north of the Australian margin and between the Joey Rise and a basement high farther west.

The Joey Graben

The newly defined Joey Graben is a north–south trending, 400–450 km long and 20–25 km wide feature, in 5400–5800 m water depths, with a smooth seafloor (Figs 3 and 4 B). It parallels and lies east of the Joey Rise and is oblique to the Mesozoic oceanic fracture zones (Figs 6 and 7). Multichannel seismic reflection profiles covering the eastern Joey Rise and part of the graben reveal intense reflectors dipping towards the Joey Rise at 9.0 to 10.5s (ca. 12–15 km) depth. The graben extends northward beyond the Joey Rise to terminate against the Roo Rise (Fig. 4B). It consists of local subgrabens trending west-northwest to north–south (Fig. 6), which have faulted basement blocks bounded by normal faults (Figs 3 and 4 B). The maximum sediment thickness within the graben is about 1.0s (TWT), and a maximum of about 0.6s of Mesozoic sediments is confined to the central graben close to the eastern slope of the Joey Rise. Miocene and Plio-Pleistocene sediments are about 0.2s thick in the graben and they thicken to the east.

The southern, north-trending, subgraben is about 150 km long and contains up to 1.0s of sediments. The northern, WNW-trending, subgraben is 100 km long and has sediments about 0.6s thick. Basement is generally about 6400 m deep (Fig. 6). In the northwest of the northern subgraben, upper Miocene (Sequence B) sediments are covered by Pliocene–Pleistocene (Sequence A) mass-wasting deposits from the Joey Rise (Fig. 4 B). Sequence C sediments are absent in the central Joey Graben. Numerous basement faults extend through the older sediments (Sequences D and E) and displace them. Sequences A, B, and C thicken eastward, while sequence D thins and dips eastward. The variable disposition of the depocentres implies changes in the tectonic regime with time.

Step-like structures

In the northeast of the plain, two steps within the basement are separated by a northeast-trending basin (Figs 7 and 8). Within the northeastern basin, small linear grabens separate the northern and eastern steps (Fig. 6). The northern step is broad and dips gently to the southeast and east (Figs 4 A and 6). It is down-faulted southward by up to 400 m, forming local horsts (around 6000 m

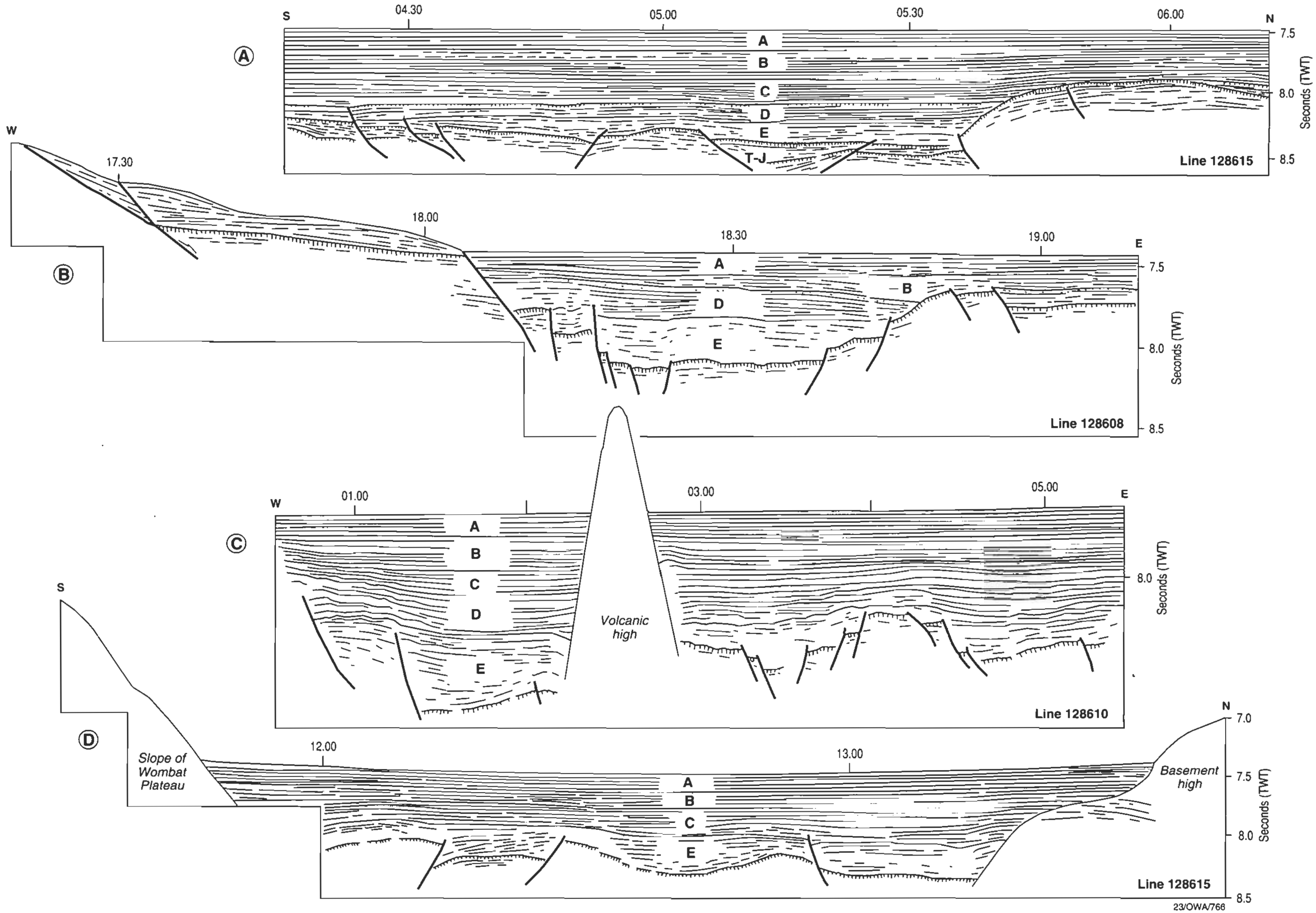

Figure 4. Line drawings of analogue seismic reflection records (TWT in seconds): A, B, C, D, and E are identified seismic sequences. A: Profile across the central basin and northern step; B: Profile across the Joey Graben; C: Profile across the volcanic massif; D: Profile across the Wombat Graben; note distinct block-faulted style of the basement on either side of the massif.

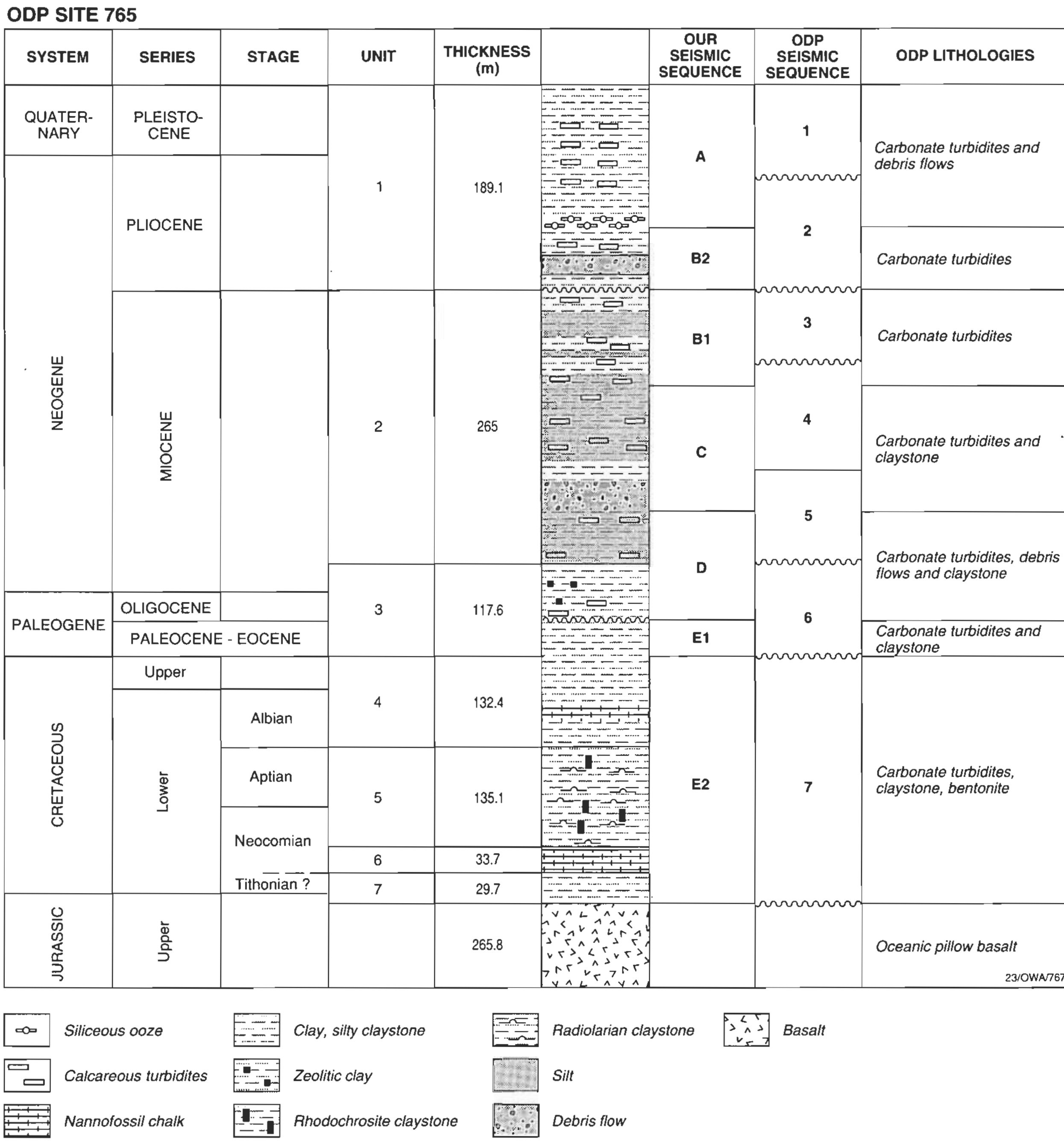

Figure 5. Lithological log of ODP Site 765 and correlation with our seismic sequences. ODP seismic sequences and lithological descriptions from Gradstein et al. (1990).

deep) and grabens (around 6400 m deep) about 5 to 6 km wide. The seismic sequences in the area can be grouped into two units. The lower sequences (C, D and E) fill the axial graben and are up to 0.7–0.8s thick. Sequence E sediments are up to 0.2s thick within the deepest parts of local grabens. The top of seismic sequence D is an unconformity which dips southward, with the lower reflectors of Sequence C overlapping it. All seismic sequences thin to the north toward the Sunda Rise (Fig. 4 A). Sequences E and D are dislocated by basement faults, resulting in local anticline-like structures. The sediments are between 0.25–0.8s thick on the northern step, and 0.45–0.85s on the eastern step. Up to 1.0s of sediment occurs at the intersection of the northern and eastern steps, where the basement is broken by numerous fractures (Fig. 6). Seismic sequence E is absent in the north (Fig. 4 A). Flexural deformation decreases upward. Local folds, in both acoustic basement and sediments, form a chain of subsurface hills in the northeast (Fig. 6).

Central basin

The central basin covers a large area of the plain (Figs 4A and 7). Sediment thickness varies from 1.0s to 1.2s (Fig. 8), and the thickest sediments are in the southeast where Mesozoic sediments of Sequence E are present. Sequence E is nearly twice as thick in the axial parts of the grabens within the basin as it is on basement rises. The entire central basin is covered with Cainozoic sediments 0.5–1.0s thick. Isopachs of the Plio-Pleistocene reveal a dominant northeast–southwest trend (Fig. 9 D), concurrent with the Mesozoic crustal features, but unlike the intervening sequences.

Flexural deformation of sequences E, D and C occurs above most of the fracture zones. Acoustically transparent zones, within parallel and horizontal seismic sequences, are widespread through sequences C, D and E and above local basement rises. Cook et al. (1978) identified them as mud diapirs, but the present multichannel seismic records show that the "diapirs" are faulted folds, and seismic reflectors continue across them. Seismic sequences A, B and C onlap onto them.

The graben northeast of the volcanic massif (Figs 4 C and 7) is 12–20 km wide and contains 1.0 to 1.2s of sediments, and basement there is 6850–7000 m deep. The lower sequences and basement are severely faulted against the eastern slope of the volcanic massif, and sequences D and E thin up its slope.

Volcanic massif

Southwest of the central basin are two lengthy NW-trending features: an oceanic basement rise and a volcanic massif farther west (Figs 4 C and 7). The volcanic massif itself consists of three linear basement ridges separated by two linear depressions or grabens. At a number of places basement ridges reach the sea floor (Fig. 8). Several local grabens, and submarine seamounts with complex structure, occur along the western part of the massif. Sediment thickness varies from zero on some of the volcanic seamounts to 1.0s in the grabens. In this area, the basement subsides in steps (Fig. 4A). West of the massif, the down-faulted basement consists of blocks 2–8 km wide, which are displaced 100–150 m by faults. A seismic section across the axial graben of the volcanic massif reveals downfaulted basement blocks, 70–80 km wide, at 6400–6000 m depth.

Thus, the eastern part of the volcanic massif (Fig. 4C)

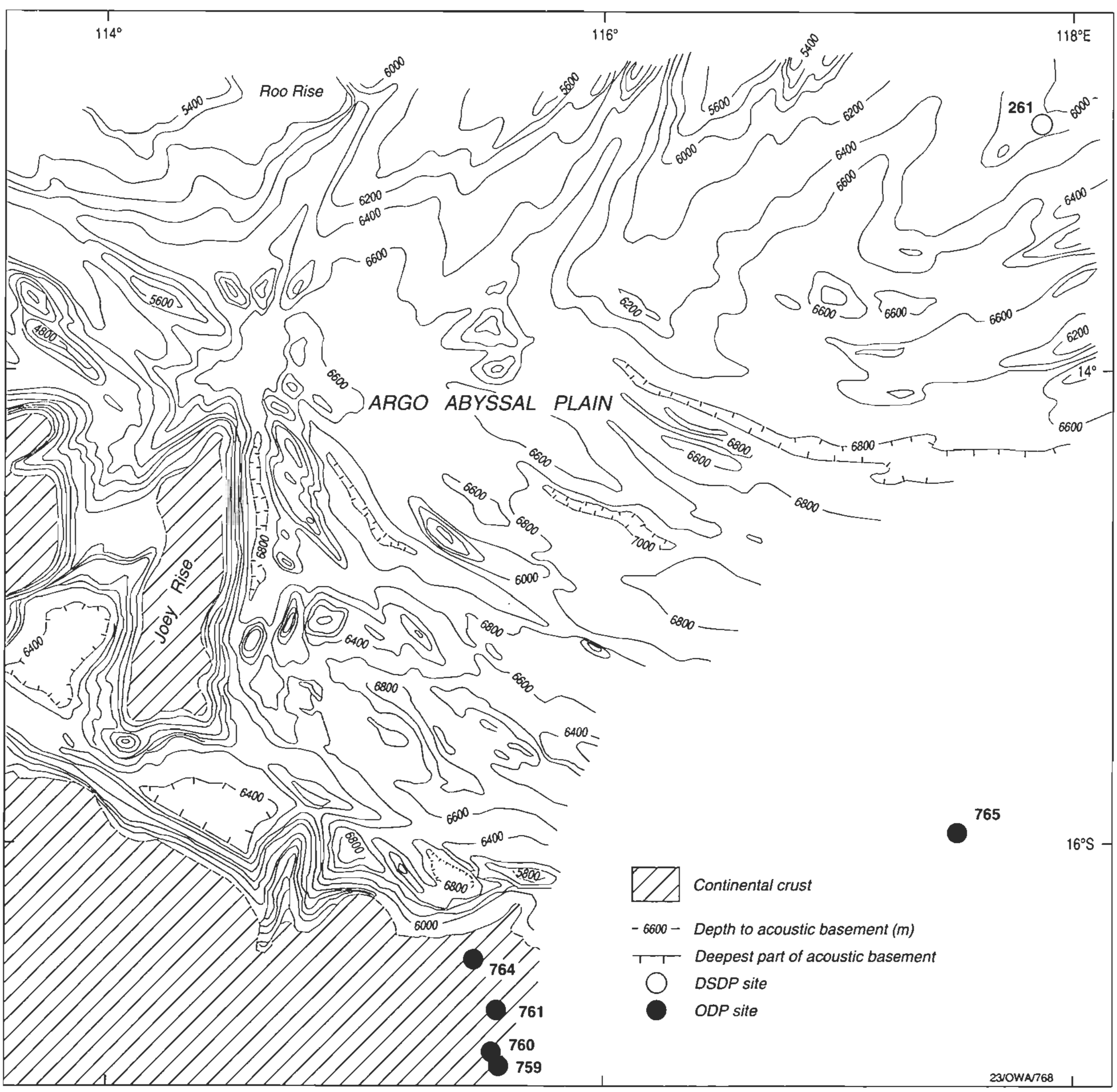

Figure 6. Map of depth to acoustic basement (depth in metres). Open and closed circles represent the DSDP and ODP drill well locations, respectively.

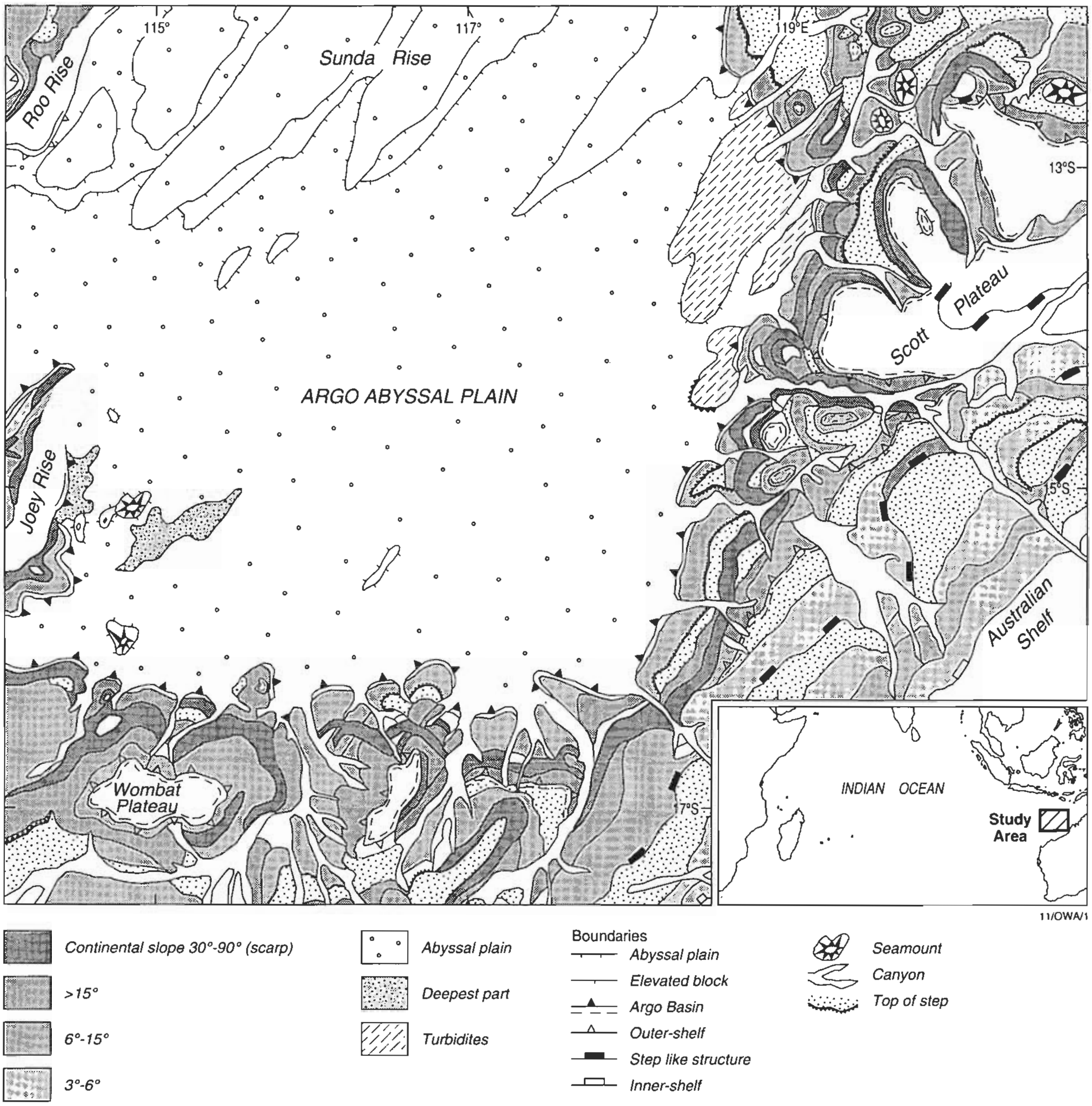

Figure 7. Map of the identified structural elements of the Argo Abyssal Plain. Index map on the right lower corner shows the study area.

is characterised by a less-rugged basement than the western part. Sequence E is thicker above deeper basement: 0.05–0.15s above the uplifted blocks, and 0.2–0.3s in depressions. It is everywhere overlain by about 0.1–0.2s of sequence D, and about 0.2s of sequence C. The C and B sequences unconformably overlie the older sequences on the eastern volcanic massif. Sequence B is 0.1 to 0.2s thick.

Crustal structure, sequences and tectonic evolution

Seismic sequences

The seismic sequences A, B, C, D and E are generally present throughout the Argo Abyssal Plain, excepting sequence C in the Joey Graben and sequence D in the Wombat Graben.

In large areas of the Argo Abyssal Plain, the Mesozoic (sequence E) and Early Cainozoic (sequence D) sediments occur in palaeo-deeps, in areas of downfaulted basement. DSDP site 261 revealed Cretaceous pelagic clays with glauconite in the northeastern plain (Veevers et al., 1974), while ODP Site 765 revealed hemipelagic clays and claystones overlain by calcareous turbidites in the south-east (Gradstein et al., 1990). This points to thick infill of pelagic sediments and turbidites in the grabens close to the continental margin, and pelagic sedimentation on the northeastern plain. Further, downfaulted basement, and faulting of the overlying Mesozoic to early Miocene (E and D) sediments in the central basin, the grabens of the volcanic massif, and the step-like structures, clearly indicate relatively intense tectonism till the end of the early Miocene. The confinement of the oldest sediments to the axial graben denotes early faulting, and its control of sedimentation. The regional tilt of the abyssal plain, with the basement deeper in the south and southwest, and higher in the west and in the north in the tectonic bulge south of the Sunda Arc, may have developed early, on the evidence of the thicker older sequences there.

The low frequency and high amplitude reflections of seismic sequence D mark near-bottom, high-energy sedimentation. DSDP site 261 revealed a sharp reduction of the sedimentation rate in the early Palaeogene, and palaeogeographic changes. A notable change in near-bottom sedimentation at the end of the early Miocene (sequence C) is inferred. Sequence D dips eastwards in the Joey Graben (Fig. 4 B) and across the adjacent volcanic massif, due to tectonism during the Oligocene/Miocene period. This may have caused erosion or non-deposition, as suggested by the absence of sequence C.

Gradstein & von Rad (1991) interpreted structural tilt northward of the Wombat Plateau into the Argo Abyssal Plain as having been caused by isostatic rebound due to subsidence of the Wombat half-graben (south of the plateau) along normal faults unloading it in the latest Jurassic, just before breakup. The faulted basement and the overlying Mesozoic sediments also suggest early tectonic activity of the area.

Increase in areal extent with time, and the thickness of sequences C to A, point to gradual migration of depocentres to the east, and increase in extent of sedimentation with time. The ubiquitous Plio-Pleistocene sediments of sequence A gradually thicken eastward. However, there are notable changes in sedimentation during the Oligocene–Miocene period (37–16 Ma) as shown by the absence of middle Miocene sediments (sequence C) in the Joey Graben; the presence of upper Oligocene–lower Miocene (sequence D) sediments northwest of the Wombat Graben;

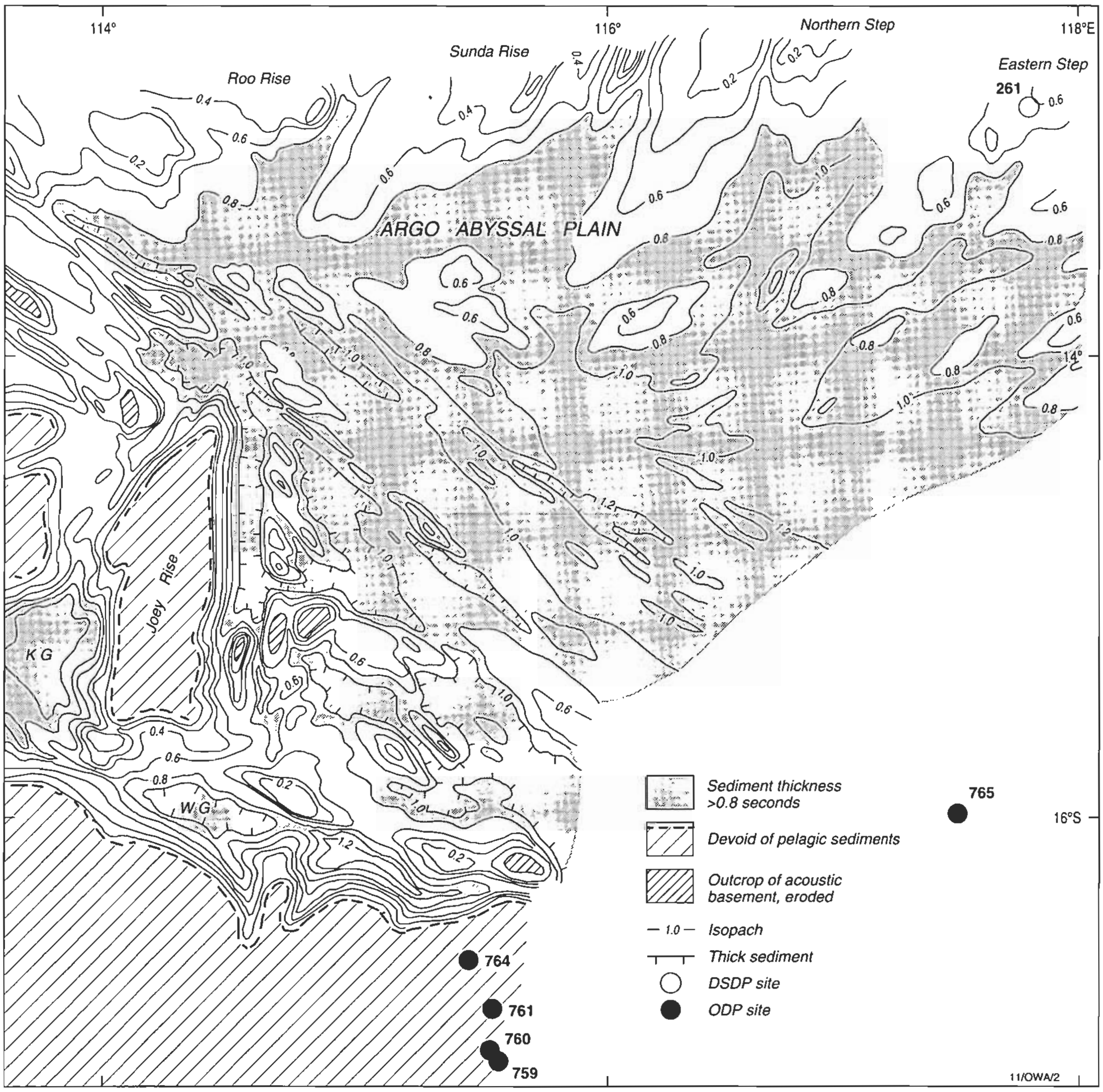

Figure 8. Isopach map of total sediments. Thickness values are in seconds (TWT). Open and closed circles are DSDP and ODP drill well locations, respectively. KG = Kivi Graben; WG = Wombat Graben.

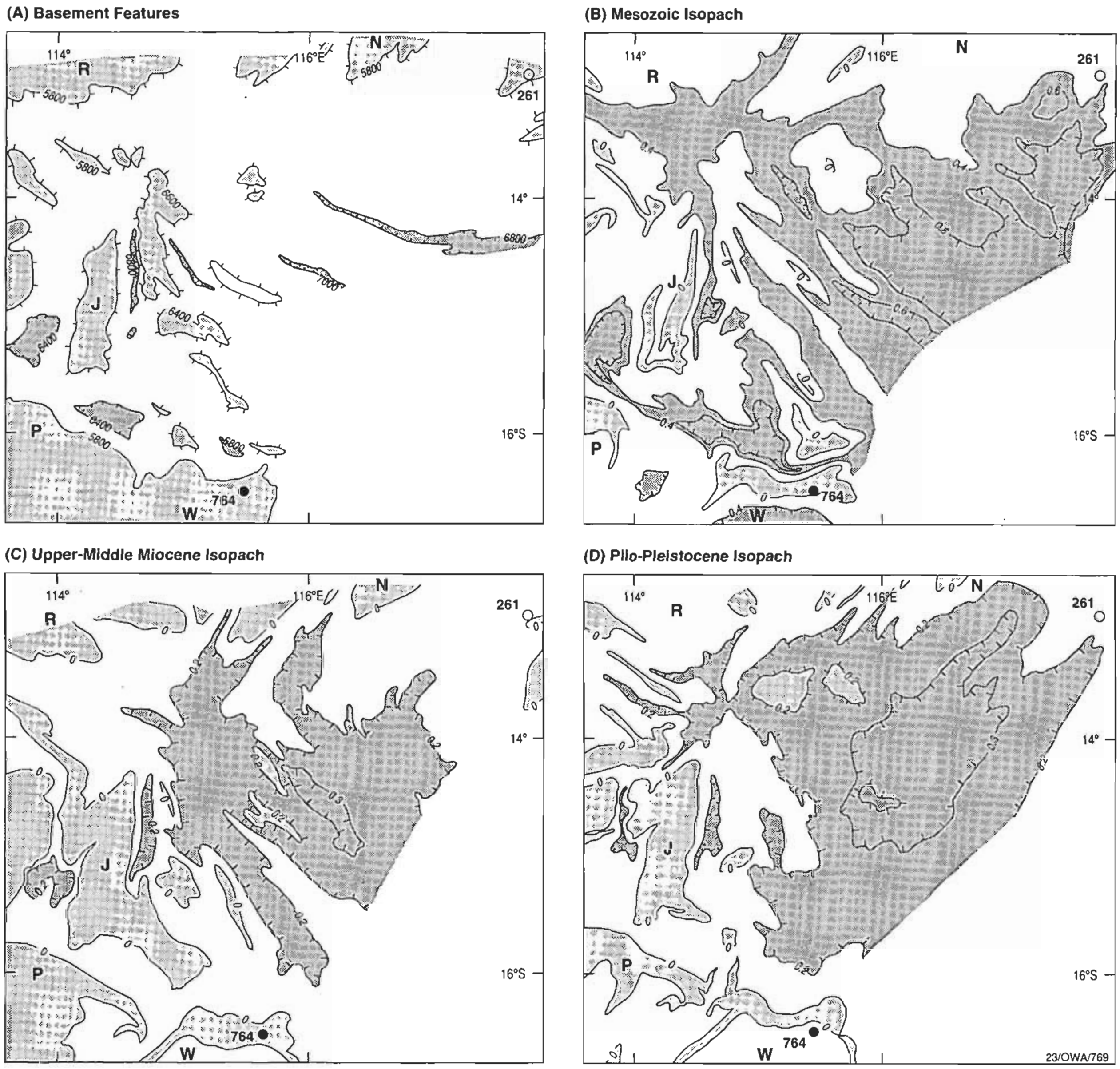

Figure 9. Maps of the Argo Abyssal Plain in seconds (TWT). A = basement features, B = Mesozoic isopach, C = upper to middle Miocene isopach, D = Plio-Pleistocene isopach. N = northern step, J = Joey Rise, P = Platypus Spur, W = Wombat Plateau, *261* = DSDP Site 261, *764* = ODP Site 764.

and the identification of the top of sequence D as an unconformity south of the step-like structures in the northern Argo Abyssal Plain. Bottom currents probably resulted in erosion or non-deposition.

Tectonics

The widespread fault-controlled deformation of earlier sediments, and the less-widespread deformation of younger sediments, suggest continued compressional stress, which is more pronounced in the northeast than the south. The basement in the northeast is broken into numerous local blocks 50–70 km across with vertical displacements of 250–300 m. Deformation continued until the early Miocene in the north, much later than in the south. Furthermore, there are altogether different structural and tectonic fabrics on either side of the volcanic massif (Figs 6 and 7). The presence of Upper Jurassic–Cretaceous and upper Oligocene–lower Miocene sediments along the slopes of the extinct volcano on the massif suggests that it originated in the Upper Jurassic. Downbending and increasing thickness of seismic sequences E and D toward the volcanic massif, with flat-lying sequences above them, mark its subsidence till the late Miocene (about 14 Ma). The increase of basement depth and earliest sediment thickness north of the Wombat Graben, and the right-lateral shift of the axial parts of the graben (Fig. 6) appear to be a manifestation of the initial shearing and rifting of the crust. Thick Mesozoic and Cainozoic deep-sea basin sediments surrounding the Australian continental margin in the Indian Ocean were also reported by Pillipenko & Sivukha (1990), who believed that they originated during an early stage of graben/basin evolution.

In the deep grabens, the basement contour map (Fig. 6) shows about 7000 m depth to acoustic basement, and the total sediment isopach map (Fig. 8) shows about 1200 m thickness of sediments. The individual isopachs of Mesozoic to upper Miocene sediments are closely controlled by the northwest structural trend of the Mesozoic ocean floor.

A comparison of the isopachs of the Mesozoic, upper to middle Miocene, and Plio-Pleistocene, with maps of basement features (Figs 6, 7 and 9), shows how sediment distribution has changed with time. The Mesozoic isopach (Fig. 9 B) represents the sediments that infilled the Late Jurassic to Early Cretaceous volcanic surface, and shows obvious control by the basement features. In the various areas, different basement features exert control: in the far north, NE trends; in the east, centre and northwest, NW trends; in the south the WNW-trending Australian continental margin, and in the west the north-trending Joey Rise. The sequence is absent only on local highs, and most of the plain is covered by more than 0.4s of section. There are local depocentres along the Australian margin where sediment was derived as turbidites from Australia (as first mapped off the Swan Canyon by Cook et al. 1978). There was little deposition between the Early Cretaceous and the Miocene.

The Miocene isopach (Fig. 9 C) shows that the area sedimented was smaller than in the Mesozoic, with little deposition near the Australian margin, and the depocentre was in the middle of the abyssal plain. NW-trends predominate in the central basin and NE trends in the north. The Plio-Pleistocene isopach (Fig. 9 D) is similar to the Miocene isopach, but NE-trends are prominent farther south, perhaps because of cracking of the oceanic crust in that direction, as it climbed the peripheral fore-bulge of the Sunda Trench. The tilt down to the south, and the deposition of unstable volcanic ash from the Indonesian Arc, caused the flow of sediment to the south.

Seafloor spreading started with early Mesozoic anomaly M26 (ca. 156 Ma). Spreading ridges trend NE and there are complementary NW-trending fractures zones. The trends of the major grabens like the Joey Graben (Fig. 7), oblique to the fracture/transform trends in the plain, point to further shearing and rifting of the crust in NNE to northerly directions. Seismic reflectors at 9.0–10.5s (12–15 km) depth near the Joey Rise are most likely from the upper mantle. The shallow mantle in the Joey Graben, and the right-laterally displaced axial parts of the Wombat Graben, suggest the presence of thin sheared and rifted crust in the grabens. In the southwest, too, the volcanic massif subsided till the Oligocene–Miocene, and there is early block-faulting of basement on either side of it. The regional tilt and increasing areal extent of the sediments on the accumulative plain to the south, and the shearing, rifting, subsidence and faulting of basement until the Miocene period, have been the keys to the present structural configuration of the Argo Abyssal Plain.

Conclusions

The Argo Abyssal Plain is limited by the Sunda Trench to the north, the Joey and Roo Rises to the west, the Scott Plateau and Rowley Terrace to the east, and the Exmouth Plateau to the south (Fig. 1). It started to form in the Late Jurassic as Gondwana broke up, and the Late Jurassic and Early Cretaceous oceanic crust should, according to magnetic mapping (e.g. Fullerton et al., 1989), be structured by fracture zones trending NNW and spreading ridges trending ENE. However, the basement structure of the western two-thirds of the plain mapped here does not show these trends at all (Fig. 6): the dominant trend is west to west-northwest in much of the area, northeasterly trends predominate in the north, and northerly and northwesterly trends in the north. This discrepancy could be the result of limitations in the mapping of the magnetic trends by various workers, or of the seismic mapping reported here, together with a complex tectonic history. In both cases, the line-spacing is open at 50 km or more, and perhaps we must await the release of GEOSAT images to be sure of the fabric. The volcanic edifices of the Joey and Roo Rises were built soon after the ocean floor formed.

The isopach maps show the broad changes in sediment pattern with time (Figs 8 and 9) and these can be related to the depositional pattern in ODP Site 765 (Fig. 5). The Mesozoic saw widespread sedimentation, about 400–600 m thick, of bentonite and clays, and carbonate and other types of turbidites (Fig. 5), derived largely from erosion of the Australian margin. Fan-like depocentres were present along the Australian margin, both in the area of the Wombat Plateau mapped here (Fig. 9 C), and in the area farther east off the Swan Canyon mapped by Cook et al. (1978). The trends in the isopachs mimic the basement trends in many areas, but northwesterly trends are more common in the isopach map than the basement map (Fig. 6), for unknown reasons.

The next major sequence is the Miocene (Figs 5 and 9 C), consisting of carbonate turbidites, debris flows and clay deposits. It is thinner than the Mesozoic, averaging 300 m thick, and is not as widespread around the margins of the abyssal plain. Trends in the isopachs are not dissimilar to those in the Mesozoic, with northeasterly trends prominent in the north, and northwesterly trends in the centre of the plain. The Plio-Pleistocene isopach (Fig. 9 D) shows a similar distribution and thickness to the Miocene isopach, but the northwesterly trends have largely been eliminated by sedimentation. This sequence consists of carbonate turbidites and debris-flow accumulations, much like the earlier ones (Fig. 5).

The structural and isopach maps, and the seismic sections, suggest that on-going movement on basement faults affected sedimentation through into the Miocene, but in the Plio-Pleistocene the only tectonism has been the uplift of the Sunda Trench peripheral fore-bulge as the plain moved north, and associated jostling of the northeasterly trending faults. The basement and the abyssal plain tilts southwestward down from the fore-bulge, and some seismic evidence suggests that turbidites are flowing in that direction. However, that tilt does not apply in the southwest, with its rugged basement relief.

The map of crustal structural elements (Fig. 7) summarises the tectonic situation. It shows the importance of northeast to north-northeast-trending structures in the north and west: Roo Rise, northwest graben, Sunda Rise, northern step, and northeastern basin. Northwesterly and northerly trends predominate in the southwest, with a series of NW-trending volcanic massifs and troughs, and the Joey Graben stretching about 400 km to the north, from east of Joey Rise to join the northwest graben. Along the WNW-trending Australian margin in the southwest (Fig. 6), the subsidence of the Australian margin has developed a moat up to 50 km wide, which extends eastward from the Kivi Graben of Figure 6. Figure 7, although a sketch, does indicate that the interplay of seafloor spreading, later volcanism, continental subsidence, and movement northward into the peripheral fore-bulge, has probably involved strike-slip as well as normal faulting. The seismic evidence of shallow mantle (12–15 km) below the Joey Graben suggests thinned crust that may have been readily sheared.

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Geological history of the outer North West Shelf of Australia: a synthesis

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This paper uses information contained in other papers in this issue of the *AGSO Journal* together with the scientific literature to produce a synthesis of the geological development of the outer continental margin of northwest Australia, and the adjacent plateaus and abyssal plains, with special emphasis on the northern Exmouth Plateau and Rowley Terrace. In the Palaeozoic, the outer North West Shelf formed part of the continental crust of East Gondwana. The region was stretched in the Late Palaeozoic, and subsequently subsided to form the Westralian Superbasin on the southern margin of Tethys. The superbasin filled with thick Triassic sediments, and variable thicknesses of Jurassic sediments, before progressive breakup in Callovian–Valanginian time.

The Late Triassic was dominated by fluvio-deltaic sedimentation but, on what ultimately became the outer shelf, carbonates were deposited. These included reefal buildups on what is now the northern Exmouth Plateau and Rowley Terrace. Rift volcanics were erupted in areas of future breakup, in the latest Triassic and earliest Jurassic. The Early Jurassic saw a similar sedimentary situation to the Late Triassic, with mixed siliciclastic and carbonate deposition. Restricted circulation allowed organic-rich, potential petroleum source-rocks to form in local depressions throughout the Jurassic. In the Middle Jurassic, shallow-marine conditions predominated and siliciclastic sedimentation was ubiquitous; coaly sequences were laid down in swamps on parts of what is now the northern Exmouth Plateau and outer Rowley Terrace. The late Middle Jurassic saw thermal uplift and erosion prior to breakup of Gondwana in the north, and a major period of faulting and rift volcanism. Callovian breakup led to the genesis of the Argo Abyssal Plain by seafloor spreading, but there was very little sedimentation on the plain until the Early Cretaceous. Late Jurassic and earliest Cretaceous sedimentation was shallow marine and siliciclastic.

The remainder of East Gondwana broke up in the Valanginian, and the Gascoyne, Cuvier and Perth Abyssal Plains were initiated as seafloor spreading began. The present-day physiography started to take shape. Volcanic buildups of the Wallaby Plateau and Joey Rise formed in the Early Cretaceous on oceanic crust. In the Neocomian, rift-style volcanics were erupted on the Scott Plateau, 30 m.y. after breakup in that area, and volcanic ash was a major source of sediment on and near the northern Exmouth Plateau. The Early Cretaceous was a period of transgression with siliciclastic shelf sedimentation on the subsiding continental crust gradually giving way to marls. In the Santonian, deposition of shelf carbonates became established, and thenceforth shelf and bathyal carbonate deposition characterised the margin. On the Argo Abyssal Plain, there was a steady rain of pelagic clay, and major influxes of carbonate turbidites and debris flows in the Early Cretaceous and the Mio-Pliocene. A tilt to the south developed in the Early Miocene, presumably as the fore-bulge south of the Java Trench affected the plain as it moved northward with the Australian Plate.

Introduction

This paper reviews the geological development of the outer North West Shelf, with special emphasis on the new results documented in this issue of the *AGSO Journal*, which concentrate on the Mesozoic sequences. The North West Shelf extends about 2400 km along the northwest margin of Australia out at least to the 2000 m isobath (Fig. 1), from near Darwin in the northeast to near Exmouth Gulf in the southwest (Purcell & Purcell, 1988). Our area of interest includes the Exmouth and Wallaby Plateaus and encompasses three major Phanerozoic sedimentary basins: the Browse Basin, adjacent to and extending beneath the Scott Plateau, the offshore Canning Basin, under and landward of the Rowley Terrace, and the offshore Carnarvon Basin, beneath and landward of the Exmouth Plateau and Carnarvon Terrace (Fig. 1). These basins are filled with superimposed sequences with shelf-wide unconformities between them, and are more than 15 km thick in places. Representative east–west regional cross-sections are shown in Figures 2 and 3; the Wallaby Plateau is clearly of quite different character from the Scott and Exmouth Plateaus. No plutonic igneous or metamorphic basement rocks of continental origin have been drilled or dredged along the outer margin in the area, the oldest rocks recovered being Upper Triassic sediments.

Our present understanding of the regional geology comes from a number of sources:

- Petroleum exploration on the Northwest Shelf, using high-quality seismic data sets and petroleum exploration wells. Many valuable papers were included in Purcell & Purcell (1988), the most valuable for our purposes being that of Barber (1988). More detailed information is available in company completion reports (see references in the specialist papers in this issue).
- Papers dealing with the breakup history of the margin, based largely on marine magnetic lineations. Some key papers are those of Larson (1975), Heirtzler et al. (1978), Veevers et al. (1985a,b), and Fullerton et al. (1989).
- Papers dealing with the continental margins and adjacent abyssal plains, based largely on reflection seismic data and seabed sampling. Key papers include those of Symonds & Cameron (1977), Hinz et al. (1978), Exon & Willcox (1978, 1980), von Stackelberg et al. (1980), Stagg & Exon (1981), Exon et al. (1982), von Rad & Exon (1983), Exon et al. (1988), Mutter et al. (1989), Exon & Ramsay (1990), Colwell et al. (1990), and von Rad et al. (1990).
- Reports dealing with the drilling results of the Deep Sea Drilling Project and the Ocean Drilling Program, which covered the Exmouth Plateau, the Wallaby Plateau and the abyssal plains. Major references are Veevers et al. (1974), Haq et al. (1990), von Rad et al. (1992), and Gradstein et al. (1990, 1992). Key papers arising from the drilling are those of Williamson et al. (1989), Exon et al. (1991), and von Rad et al. (1992). The most important drill sites for this paper were those on the Wombat Plateau (northern Exmouth Plateau), which fully cored the Upper Triassic sedimentary sequence for the first time (Fig. 4).
- The papers in this issue of the *AGSO Journal*, which will be referenced as appropriate.

Where geological ages are given in this paper, we use

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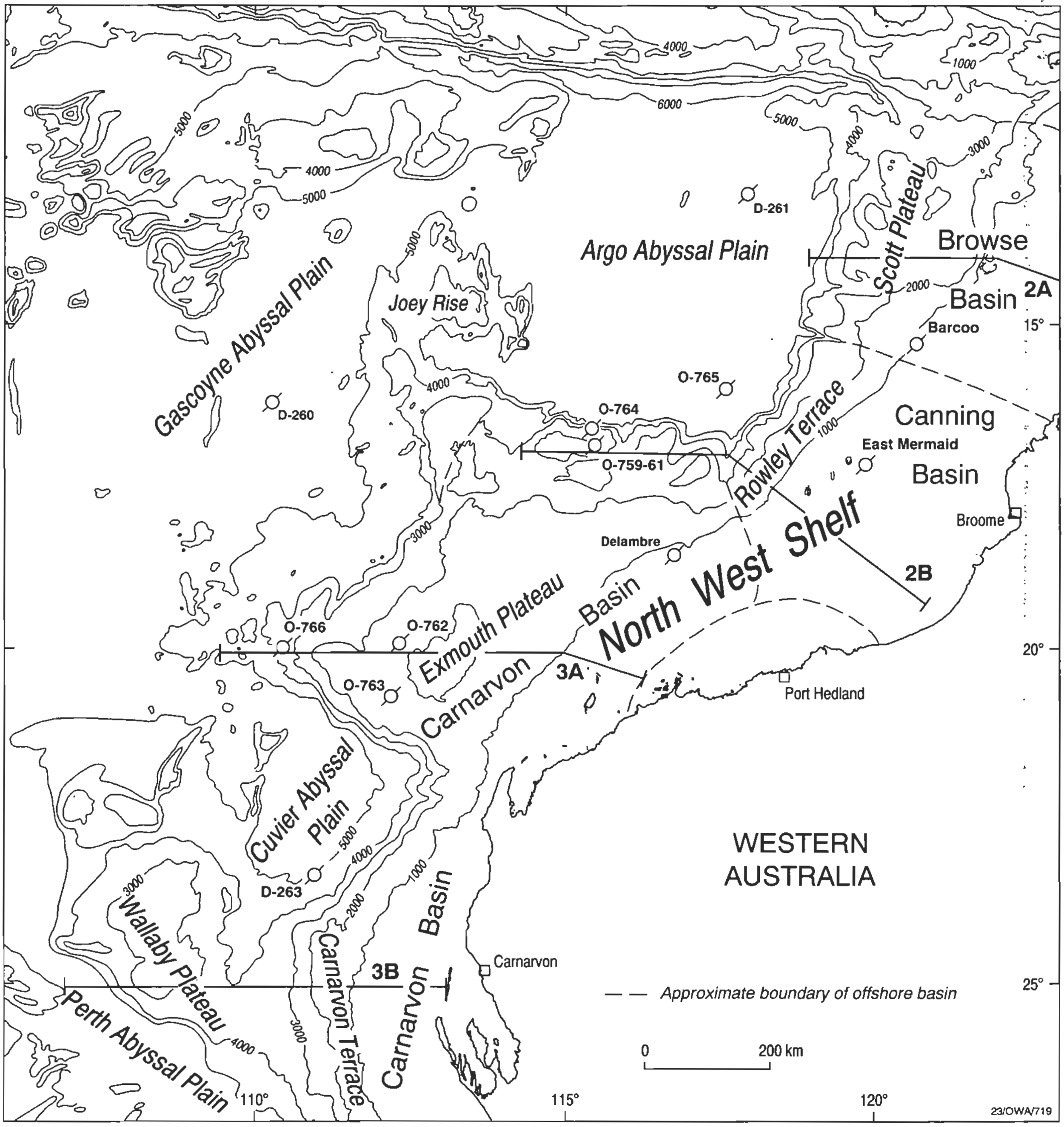


Figure 1. Bathymetric map of the North West Shelf and adjacent features, showing key topographic names, sedimentary basins, and wells referred to in this paper. The wells include those from the scientific Deep Sea drilling Project (DSDP) and Ocean Drilling Program (ODP), and several commercial wells studied during this investigation. Locations of cross-sections of Figures 2 and 3 shown.

the time scale of Haq et al. (1987).

Reflection seismic data, dredging and drilling information for the Exmouth Plateau is summarised in Figure 5, which shows the stratigraphy of the Exmouth Plateau in general, and the rather different sequences present on the northern plateau. The regional seismic reflector nomenclature (A-F) is that of Willcox & Exon (1976) — see Ramsay & Exon (1994 — this issue). Evidence of tilted fault blocks buried near the eastern margin of the Carnarvon Basin (Bentley, 1988) and deep-crustal seismic data from the Exmouth Plateau (Williamson et al., 1990) suggest that the continental crust was thinned in the Permo-Carboniferous. This period of stretching and thinning led to rapid and very widespread subsidence, leaving space for the deposition of several kilometres of Triassic fluvio-deltaic sediments. Exmouth Plateau stratigraphy is similar to that of the rest of the Carnarvon Basin east of the plateau, and of the Rowley sub-basin of the Canning Basin beneath the Rowley Terrace (Fig. 1). Figure 5 shows that the Triassic sequences are thick everywhere, whereas the Jurassic, Cretaceous and Cainozoic sequences are highly variable in thickness.

Yeates et al. (1987) have shown that Permian and thick Mesozoic sequences were laid down in the Westralian Superbasin below what is now the North West Shelf, on the southern edge of the Tethyan Ocean and the northern shores of Gondwana. When East Gondwana broke up in the Late Jurassic and Early Cretaceous, and abyssal plains formed to the west, the North West Shelf gradually took up its present configuration as the result of subsidence and deposition. The magnetic anomalies on the deep ocean floor (Fig. 6) and other evidence indicate that breakup came earlier north of the Exmouth Plateau (Late Jurassic) than it did in the south (Neocomian).

Geological development

The area under review was part of the northern margin of the continent of Gondwana and the southern margin of the Tethyan Ocean until breakup in Late Jurassic to Early Cretaceous times, and lay within the largely shallow marine–paralic Westralian Superbasin in the Permian, Triassic and Jurassic (Yeates et al., 1987). Only after

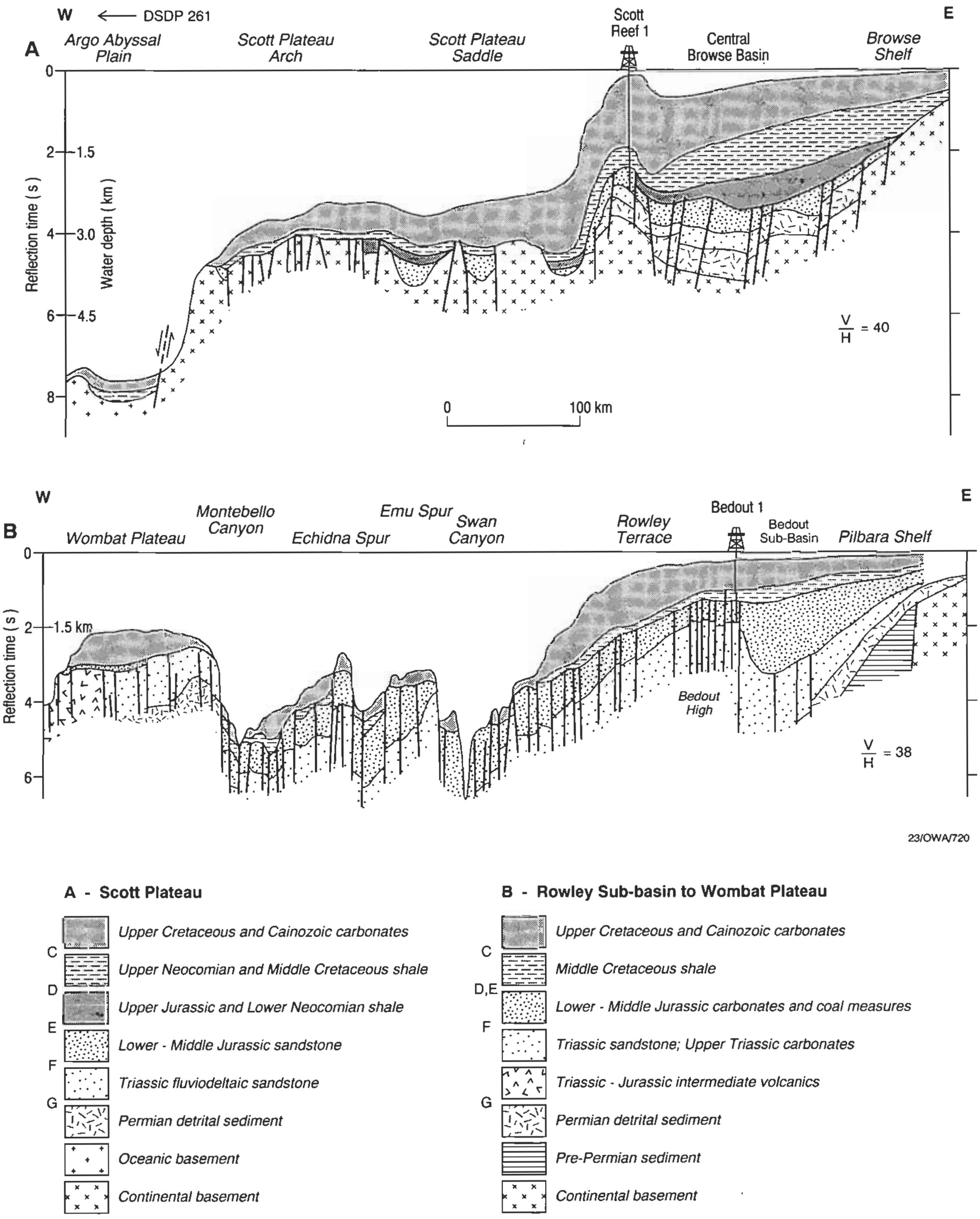


Figure 2. Generalised cross-sections across the Scott Plateau and Browse Basin (A), and across the northern Exmouth Plateau, Rowley Terrace and Bedout sub-basin of the offshore Canning Basin (B). After von Rad & Exon (1983). Locations in Figure 1.
Note that the high vertical exaggeration results in the faults appearing vertical or near vertical.

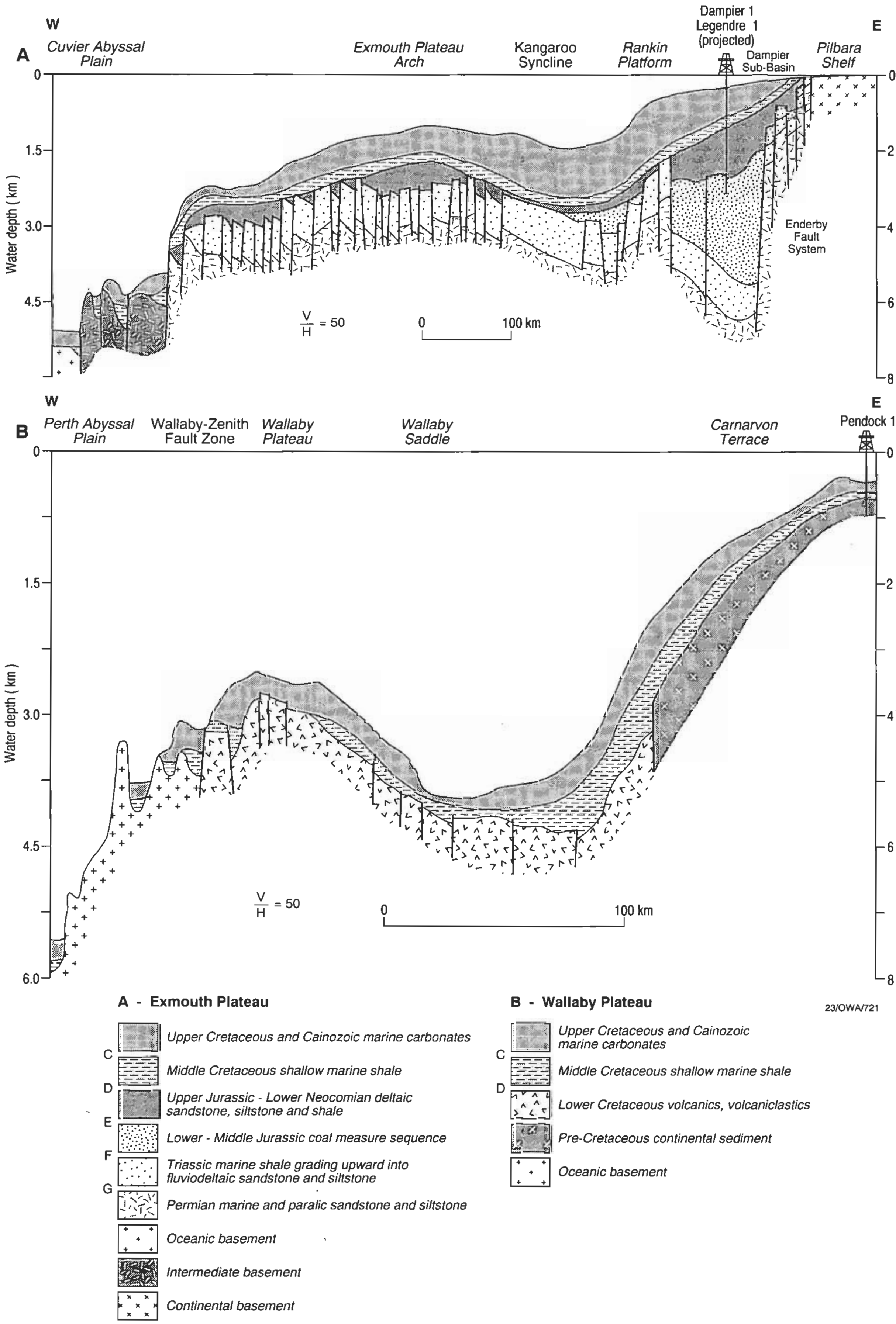


Figure 3. Generalized cross-sections across the central Exmouth Plateau and Dampier sub-basin of the Carnarvon Basin (A), and across the Wallaby Plateau and Carnarvon Terrace (B). After von Rad & Exon (1983). Locations in Figure 1. *Note that the high vertical exaggeration results in the faults appearing vertical or near vertical.*

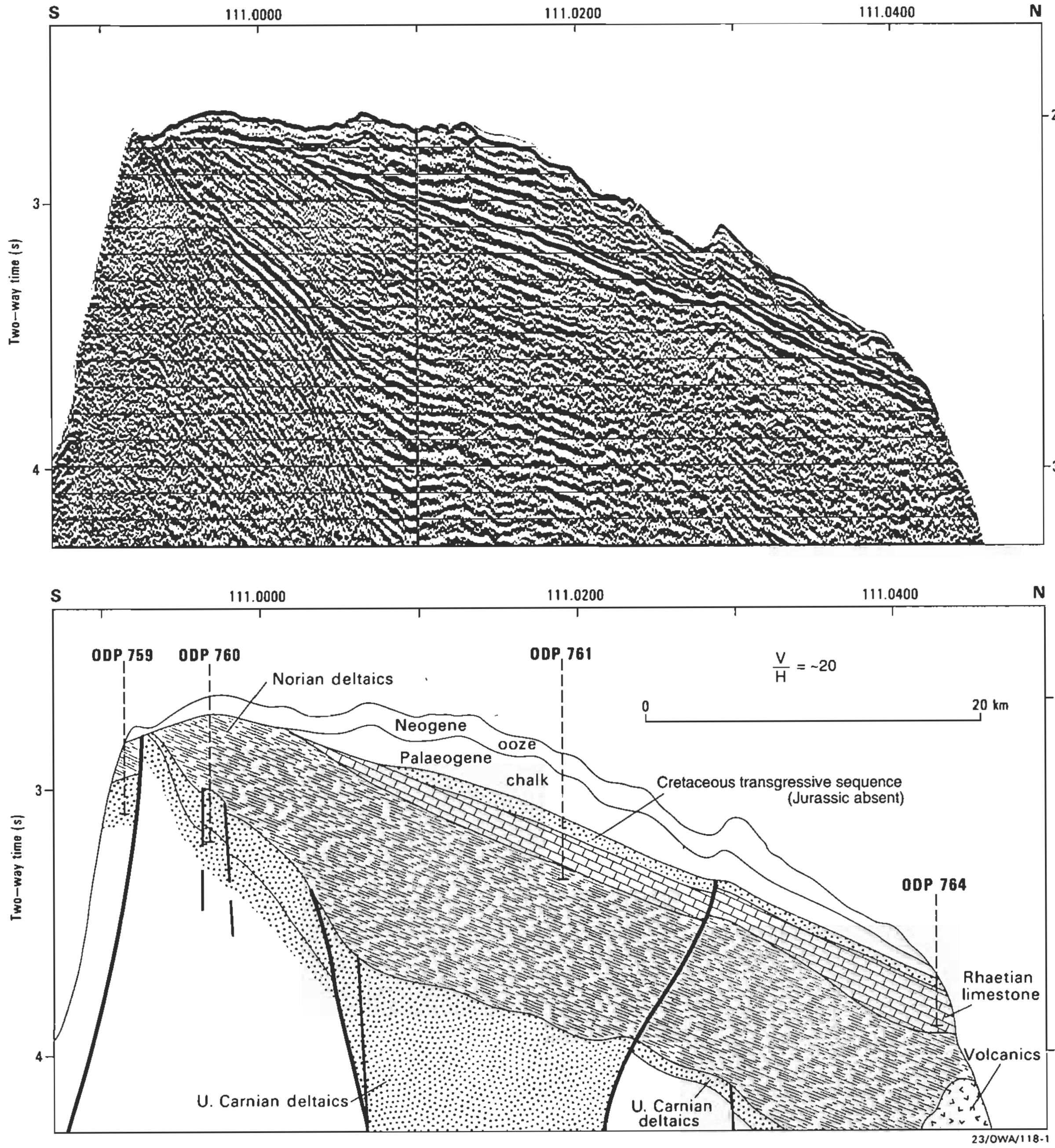


Figure 4. BMR seismic profile 56/13 across the Wombat Plateau, with four ODP drill sites projected on to it.

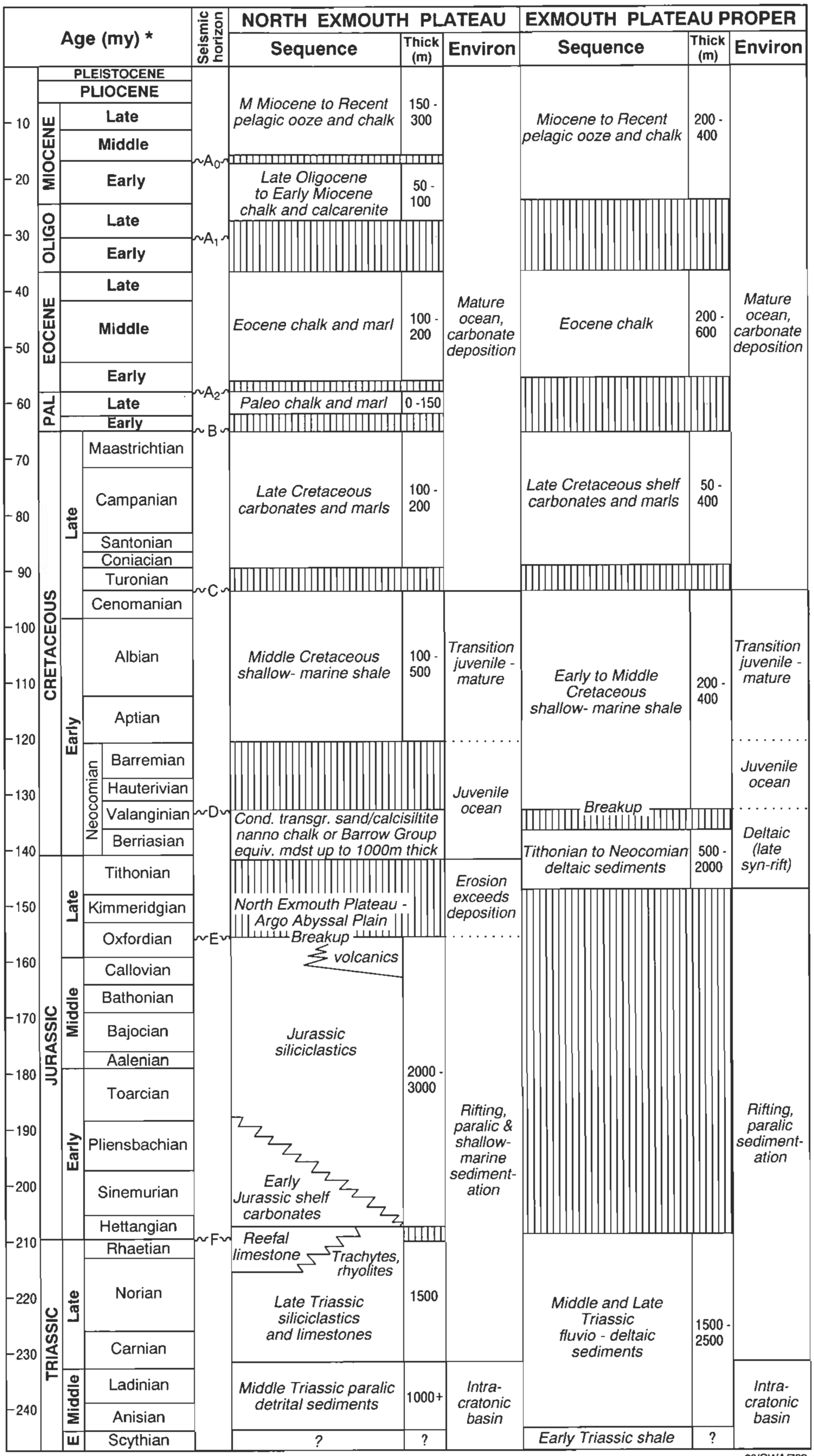
breakup did it start to take up its present configuration with the North West Shelf bounded by deep ocean basins. The region is dominated by sedimentary strata, but volcanic rocks are common in some places and at some times (Crawford & von Rad, 1994 — this issue).

Basement beneath the continental margin

Continental basement rocks have not been recovered from the outer margin by drilling, neither in commercial nor ODP wells, or by dredging. However, magnetic, gravity and seismic reflection data suggest that most of the outer margin is underlain at depth by continental basement rocks similar to those in the onshore Pilbara Block (e.g. Exon & Willcox, 1980; Stagg & Exon, 1981). Furthermore, deep crustal velocities from the Exmouth Plateau indicate the presence of thinned continental crust (Mutter et al., 1989). The exceptional area is that beneath the Wallaby Plateau and the adjacent part of the Wallaby Saddle (the saddle lies between the plateau and the Carnarvon Terrace; Fig. 1). Colwell et al. (1994b — this issue) have marshalled the existing information (seaward dipping reflectors, thick dredged volcanics, volcanic geochemistry) to suggest that basement beneath Wallaby Plateau and Saddle consists largely of oceanic volcanics, with some contamination from early, sub-continental lithosphere.

Triassic mega-sequence (pre-F reflector)

Extensive well and dredge information shows that Triassic sedimentary rocks are widespread along the continental margin (e.g. Fig. 5), except beneath the Wallaby Plateau (e.g. papers in Purcell & Purcell, 1988; Colwell et al., 1994a,b — this issue; Ramsay & Exon, 1994 — this issue). No older rocks have yet been obtained from the margin. On the basis of seismic lines, we believe that Triassic



* Tertiary timescale from Truswell et al. (1989); Mesozoic - from Gradstein et al. (in press)

Figure 5. Simplified stratigraphy of the Exmouth Plateau incorporating well, dredge and seismic data. The "North Exmouth Plateau" is defined here as north of 18°S.

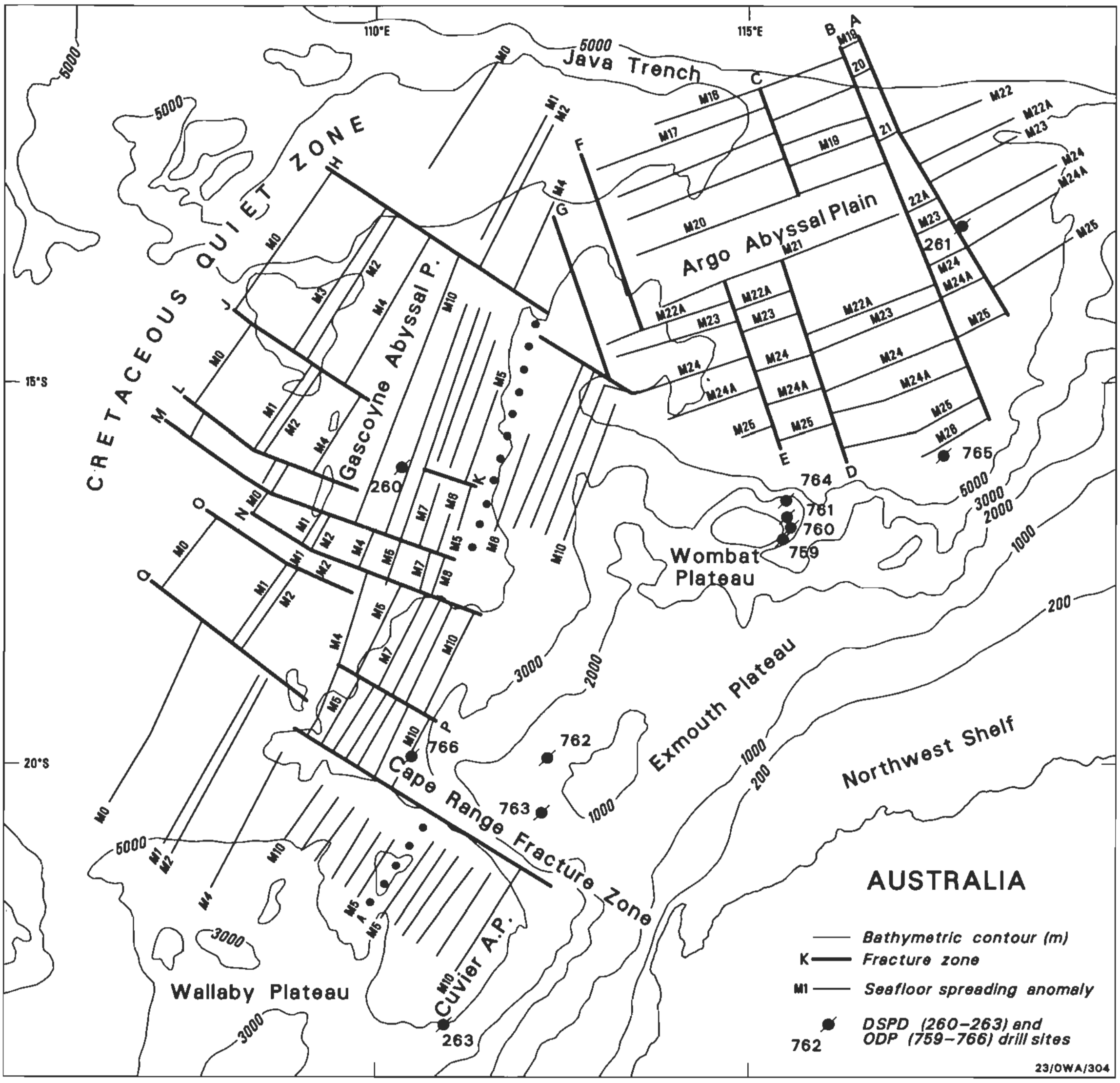


Figure 6. The northwest Australian continental margin with magnetic anomalies after Fullerton et al. (1989). Also shows simplified bathymetry, and ODP and DSDP drill sites.

sedimentary rocks exist nearly everywhere below the regional F unconformity on the continental margin. The margin is blanketed by several kilometres of Triassic rocks, and those sampled (Late Triassic in age) are mostly fluviodeltaic siliciclastic sediments (Fig. 7). However, ODP drilling (Haq et al., 1990) showed that Rhaetian reefs formed above deltaic sediments on the Wombat Plateau, and the literature indicates that Triassic shelf carbonates are common on the outer shelf, away from terrigenous sources (Barber, 1988; Williamson et al., 1989). We have no seismic images of reefal buildups in the Triassic, except on the Wombat Plateau (Fig. 8), where their seismic characteristics have been described by Williamson et al. (op. cit.).

Dredging on a number of cruises has recovered Upper Triassic rocks from the northern Exmouth Plateau and the Rowley Terrace: fluviodeltaic sediments, shallow-marine siliciclastics and shallow-marine carbonates. Many of these are described by Colwell et al. (1994a — this issue). They contain palynomorphs, shelly macrofossils including corals, and conodonts (Burger; Grant-Mackie; Campbell; Stanley; Nicoll & Foster; all 1994 — this issue). The comparison of conodonts and marine microplankton, from the same samples from dredge hauls and petroleum exploration wells, has led to a revision of the ages of the microplankton zones, and shows that they are generally younger than had been thought (Nicoll & Foster, op. cit.). Altogether, seven Norian and Rhaetian conodont zones were recognised for the first time on the North West Shelf, allowing far better correlation of the microplankton zones with the tightly controlled international conodont zones. Some microplankton zones seem to be facies equivalents rather than time successive. Stanley (op. cit.) identified Late Triassic reef-forming corals from two AGSO dredge hauls along the outer edge of the Rowley Terrace. He believes that one assemblage is probably from a reef complex, the first known on the North West Shelf outside the Wombat Plateau.

The seismic interpretation of Ramsay & Exon (op. cit.) indicates a number of unconformities within the Late Triassic; ODP drilling on the Wombat Plateau confirms that there are two unconformities within the Late Triassic

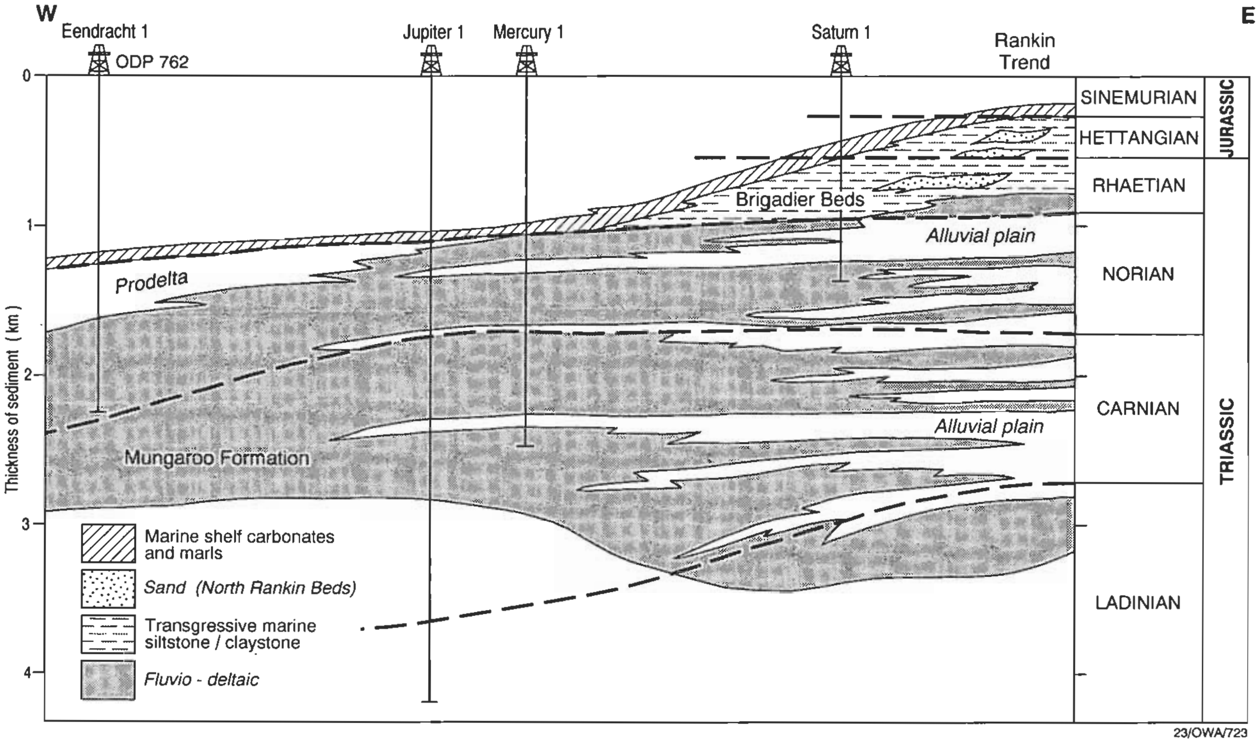


Figure 7. Depositional environments in the Upper Triassic and Lower Jurassic sequences on the Exmouth Plateau. After Barber (1988). Extends eastward from ODP Site 762 to the Rankin Platform (see Fig. 6).

sequence (von Rad et al., 1992). Most of the faulting affecting the Triassic sequence apparently occurred in the Jurassic. It is also clear that faulting was much more intense and widespread on the Exmouth Plateau and the northwestern edge of the Rowley Sub-basin than elsewhere. The large fault blocks involving Upper Triassic fluvio-deltaic sediments have been the main petroleum exploration targets on the North West Shelf, and reservoirs within them are the hosts to major gas/condensate fields such as North Rankin, Goodwyn and Gorgon. Such fault blocks on the northern Exmouth Plateau are commonly bounded by faults with throws of hundreds of metres, and are bounded on either side by Jurassic sediments, and sealed above by Early Cretaceous shales (Fig. 9). Rhaetian reefs, such as those found by ODP drilling on the Wallaby Plateau (Figs 4 and 8), offer an alternative target for petroleum exploration, particularly as they are hundreds of metres thick and highly porous and permeable (Williamson et al., 1989; Röhl et al., 1992).

Latest Triassic to earliest Jurassic rift volcanism

Dredging from the northern margin of the Wombat Plateau showed the presence of flows of rhyolite and undersaturated trachyte, three samples of which gave K/Ar ages of 213–192 Ma (von Stackelberg et al., 1980). These ages straddle the Triassic–Jurassic boundary and the volcanics appear, on the basis of their composition, to be of early-rift character (von Rad & Exon, 1983). Von Rad & Exon (op. cit.) also cited the presence of probably Middle Triassic volcanism in Bedout No. 1 well, just south of the Rowley Terrace. Exon & Buffler (1992) described, from seismic reflection profiles, a belt of generally reflector-free rock along the western and southwestern margins of the Exmouth Plateau, that lay at the same level as Upper Triassic sediments and showed a lateral transition into them landward (Fig. 10). They assumed that these were early-rift volcanics like those on the Wombat Plateau. The seismic sequence was later dredged, and vesicular trachybasalt, microlitic basalt, microgabbro, and olivine dolerite were recovered (Colwell et al., 1990; Crawford & von Rad, 1994 — this issue). On the basis of the above results, and the widespread undated but similar volcanics dredged along the margin, we believe that early-rift volcanics of latest Triassic to earliest Jurassic age will prove to be widespread along the outermost margin of the North West Shelf.

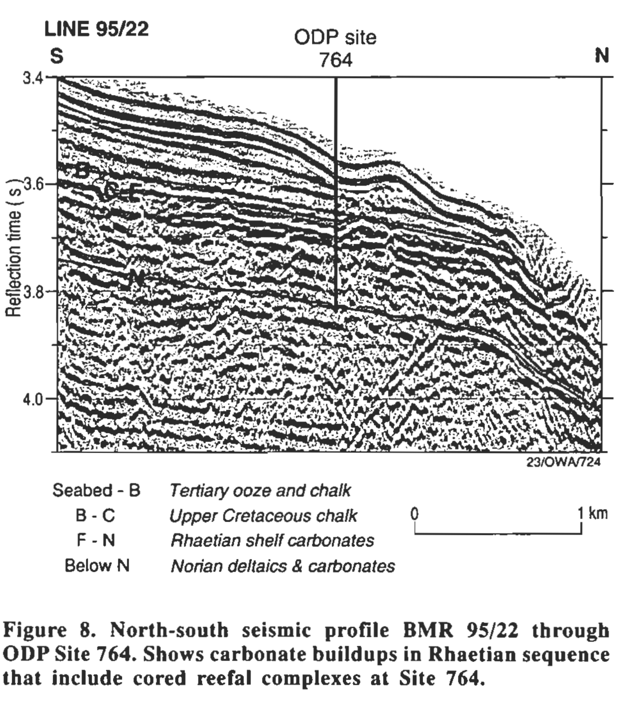


Figure 8. North-south seismic profile BMR 95/22 through ODP Site 764. Shows carbonate buildups in Rhaetian sequence that include cored reefal complexes at Site 764.

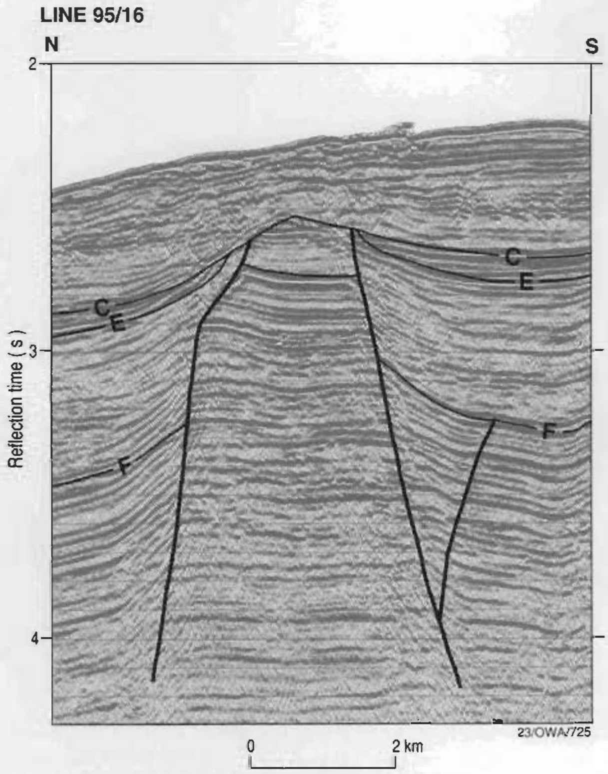


Figure 9. Faulted Lower Jurassic and Triassic sequences on seismic profile BMR 95/16 on the northern Exmouth Plateau. The top of a horst nearby has been dredged and proved to consist of Early Jurassic carbonate bank facies. An unresolved question is whether the banks formed on pre-existing horsts or whether they formed everywhere, on a level surface that was later faulted. In the former case the development of major Early Jurassic reefs is more likely than in the latter.

Lower and Middle Jurassic mega-sequence (F–E reflectors)

In much of the Carnarvon and offshore Canning Basins, the Lower and Middle Jurassic sequence (seismic sequence F-E) is 1000–2000 m thick (e.g. Fig. 5). However, on the unique high block of the Wombat Plateau, ODP drilling and dredging has not recovered any indisputably Jurassic rocks. Dredge and well evidence suggests that thinning is caused generally by non-deposition or erosion of the Middle Jurassic sequence (e.g. Colwell et al., 1994a — this issue; Exon, 1994 — this issue).

Dredge and seismic information indicates that there is a thick Lower Jurassic shelf carbonate sequence along the outer continental margin, beneath the northern Exmouth Plateau and the Rowley Terrace, overlain by a thick Middle Jurassic siliciclastic sequence (e.g. Colwell et al., 1994a — this issue; Ramsay & Exon, 1994 — this issue). Two seismic facies appear to correspond to these two lithofacies on the western end of BMR seismic line 95/12 (Ramsay & Exon, op. cit.). The lower, seismically relatively transparent facies is interpreted as consisting of homogeneous shelf carbonates and marls; and the upper, onlapping, seismically strongly reflective facies with internal normal faults, as consisting of detrital sediments and coal. On the central Exmouth Plateau, the Lower Jurassic shelf carbonates transgressed toward the continent with time (Fig. 7).

Dredge samples from the Rowley Terrace and northern Exmouth Plateau were examined in detail palaeontologically by Colwell et al. (1994a — this issue), Burger, Shafik, and Grant-Mackie (all 1994 — this issue). The samples include a great variety of shelf carbonates; marine, paludal and fluvial siliciclastic detrital sediments; coal; and ferruginized detrital sediments that were exposed during Middle to Late Jurassic pre-breakup thermal uplift of the margin. The foraminifers, ostracods, and bivalves in the carbonates are similar to those of other Tethyan margins, including the Northern Calcareous Alps in Europe. Nannofossils have been found in two assemblages only (Shafik, op. cit.). An early Toarcian assemblage appears to represent a brief holomarine ingression, whereas a younger early Bajocian assemblage represents a major regional transgression, in the sequence that is characterisised by coal measures. Two palynological assemblages from dredges from the Rowley Terrace are of Toarcian and Aalenian–Bajocian ages (Burger, op. cit.). The Toarcian assemblage suggests that paludal conditions preceded restricted marine conditions. The Aalenian–Bajocian assemblage again points to paludal and restricted marine environments.

Exon et al. (1991) identified presumed large carbonate buildups on horsts on the northern Exmouth Plateau in the Lower and Middle Jurassic sequence, using older seismic data and early versions of BMR Survey 95 seismic data. Dredging of one of these buildups yielded Lower Jurassic limestone of carbonate bank facies (Colwell et al., 1994 — this issue). From seismic data (Fig. 9), it is uncertain whether reefal buildups formed on high blocks or, alternatively, whether limestone banks were widespread on a flat Triassic surface, and subsequently cut by faulting into horsts and graben (Ramsay & Exon, op. cit.).

The Lower and Middle Jurassic sequence, like the Triassic sequence, is strongly faulted on the Exmouth Plateau and on the northwestern margin of the Rowley Sub-basin.

Tectonism and volcanism: late Early to early Late Jurassic

Seismic evidence from most of the southern North West Shelf shows that there was a major period of tectonism at or near Callovian time, represented most clearly by a major, frequently angular, unconformity (E). This unconformity has been dated palaeontologically as covering very variable time spans (commonly varying from a minimum of parts of the Callovian and Oxfordian, to as much as the entire Late Jurassic and Berriasian) and is related to thermal uplift preceding the breakup of Gondwana that formed the Argo Abyssal Plain.

The areas of most uplift, around the continental margin, were also extensively faulted, the faults showing both strike-slip and normal movement. Weathering and erosion followed, and these areas were eroded and the relief across the faults was reduced. Some areas were bevelled completely by wave action. Weathered (ferruginized) Jurassic sedimentary rocks have been widely dredged along the margin (e.g. Colwell et al., 1994a — this issue). Along the western Rowley Terrace, and along the high blocks on the northernmost Exmouth Plateau, the bevelled area is about 50 km wide, lying landward of the steep continental slope (Ramsay & Exon, op. cit.).

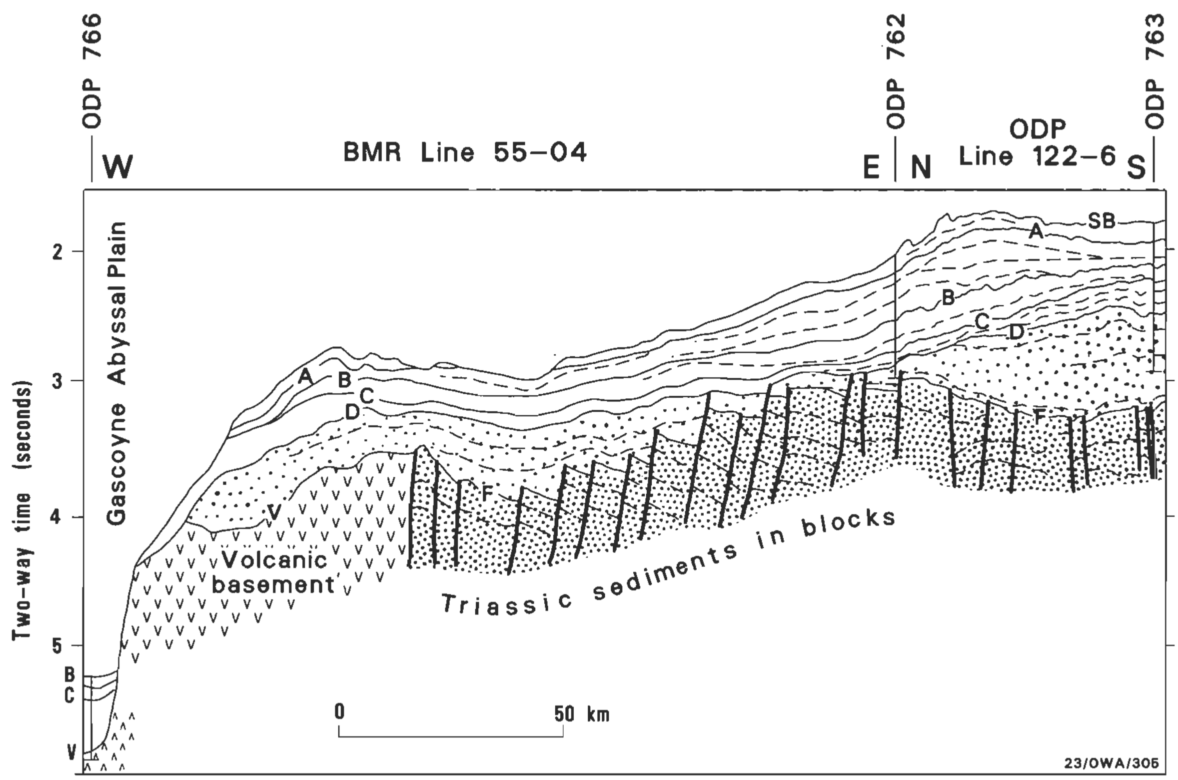


Figure 10. Line drawing from seismic profiles linking ODP Sites 762, 763, and 766, on the western Exmouth Plateau. After von Rad et al. (1992). Shows the Triassic sediments giving way oceanward to volcanic rocks along the western margin of the plateau. A variety of evidence suggests that the volcanic rocks are rift-related basaltic and trachytic flows and sills of Late Triassic and possibly earliest Jurassic age (Exon & Buffler, 1992). They are clearly not related to Valanginian breakup.

Dredging shows that some areas along the continental margin contain abundant volcanic rocks of possible Jurassic age, while others contain none or very few (Colwell et al., 1994a — this issue). Ramsay & Exon (op. cit.), using seismic characteristics, mapped a province where volcanic flows and volcaniclastics apparently underlie the E unconformity on the northwestern edge of the Rowley Sub-basin. The province extends eastwards from the continent/ocean boundary for roughly 150 km, and covers an area of about 25 000 km^2. They considered that most vents were in the western uplifted area, and that volcanics and volcaniclastics flowed down the slope to the east. Seismic stratigraphic evidence suggests that the age of the flows is immediately pre-breakup, i.e. Callovian–Oxfordian, and one dredged volcaniclastic grit was dated by forams as Oxfordian (David Lynch *in* Exon & Ramsay, 1990).

Callovian–Oxfordian breakup

Studies of the Argo Abyssal Plain show that breakup occurred at Callovian–Oxfordian time. The age of the oldest oceanic crust has been dated as 155 Ma ('Callovian' in the Haq et al., 1987, time scale) in ODP Site 765 (Ludden, 1992) where the oldest palaeontologically datable overlying sediments are Tithonian. Fullerton et al. (1989) identified the oldest magnetic anomaly as M26, in agreement with the drilling results. Thus, the age of breakup on the abyssal plain corresponds well with the dating of the E unconformity on the continental margin, indicating that the term 'breakup unconformity' is appropriate.

The position of the continent–ocean boundary (COB) was mapped by Veevers et al. (1985a,b) as corresponding with a prominent magnetic anomaly along the margin. Recent dredging (e.g. Colwell et al., op. cit.) indicates that continental rocks are present seaward of the Veevers et al. COB in places. Ramsay & Exon (op. cit.) identified the COB in the Rowley sub-basin as the outermost position of sedimentary reflectors similar to those on the high-standing part of the continental margin, and seaward of the Veevers et al. position. The position of the COB remains imprecise in most cases, since seismic quality is poor under the steeper and deeper parts of the margin.

Late Jurassic to earliest Cretaceous mega-sequence (E–D reflectors)

In this region, the Late Jurassic to Berriasian sequence (earliest Cretaceous) varies from absent in some areas, to as much as 2000 m thick where the Barrow delta and its equivalents are present. The equivalent E-D seismic sequence occurs intermittently away from the main areas of deltaic deposition, preserved in structural lows where it onlaps the bounding highs, showing that its deposition was governed by the pre-existing structure. Well evidence (e.g. Exon, 1994 — this issue) indicates that the succession consists of deltaic and shallow-marine detrital sediments, laid down as the margin started to subside following breakup. Palynological evidence from Late Jurassic dredge samples from the southern Rowley Terrace points to open shallow-marine conditions (Burger, 1994 — this issue).

Earliest Cretaceous (Valanginian) breakup and volcanism

Gondwana broke up along the western margin of Australia much later and on a different pole of rotation than it did north of the Exmouth Plateau (Fig. 6). The evidence of breakup age in the west comes from magnetic lineations on the Gascoyne, Cuvier and Perth Abyssal Plains (e.g. Fullerton et al., 1989), DSDP and ODP drilling on the abyssal plains, and the dating of a major unconformity in wells on the continental margin. All the evidence suggests that breakup occurred in the Early Cretaceous, in the Valanginian (at about 125 Ma, using the time scale of Haq et al., 1987), some 30 million years later than the 'Argo' breakup. The Valanginian 'breakup unconformity' is very widespread on the North West Shelf, even in areas where breakup occurred earlier, suggesting some sort of tectonic coupling. It is marked by our D horizon, which seldom shows much angularity, but does mark the onset of steadily increasing water depths in many areas, as the cooling continental margin and oceanic crust sank together. Below the western Exmouth Plateau, the latest Triassic to earliest Jurassic volcanic sequence was planated during marginal uplift, presumably caused by heating in the Early Neocomian before breakup (Exon & Buffler, 1992).

Breakup may have coincided roughly with the volcanism that formed the Wallaby Plateau (Colwell et al., 1994b — this issue). Whether the Wallaby Plateau formed from a mantle plume remains to be determined. Furthermore, this was a period of basaltic volcanism on the Scott Plateau in the north, as evidenced by K/Ar ages of 128–133 Ma (von Stackelberg et al., 1980). This suggests that volcanism (probably related to Valanginian breakup) may have affected widespread areas of the outer western Australian margin, even those that had broken up 30 million years earlier.

Continental history after breakup

Evidence from wells, for example the ODP sites on the Exmouth Plateau, indicates that a regional marine transgression started in the late Valanginian, immediately after breakup in the west (von Rad et al., 1992), with deposition on the E-D sequence as the continental margin and the new oceanic crust in the west, and the old oceanic crust in the north, cooled and sank. Marine erosion appears to have formed the Valanginian D unconformity, and the overlying sediments are wholly marine. The most spectacular result of the cooling and subsidence of the oceanic crust was the formation of the collapsed zone along the continental margin, between high-standing continental material and the COB, starting in the Late Jurassic north of the Exmouth Plateau and in the Neocomian farther south. This zone is tens of kilometres wide, and the total displacement of the pre-breakup sequence is of the order of 2000 m (Ramsay & Exon, 1994 — this issue). The collapse often occurred on a series of faults, many of which probably pre-dated collapse. The local character of the margin depends on whether breakup was dominated by rifting or shearing (Exon et al., 1982). Parts of the steep outer margin are cut by canyons, some controlled by fault structures. On the Argo Abyssal Plain, evidence of Early Cretaceous and Cainozoic calcareous pelagic turbidites from ODP Site 765 indicates that the canyons were active periodically from Berriasian times (Gradstein et al., 1992).

On the continental margin landward of the collapsed zone, there was little tectonic movement in post-Valanginian times. Shallow-marine detrital sediments and marls (D-C mega-sequence) were laid down over most of the region with a thickness of about 200–1000 m, with the exception of the highly structured parts of the northern Exmouth Plateau, where they onlap structural highs and are absent in places. Burger (1994 — this issue) describes Neocomian and Aptian palynomorph assemblages from the Rowley Terrace that point to open-marine shelfal environmements. The conformable transition to carbonate sedimentation in the Late Cretaceous led to the deposition of two similar mega-sequences of shelf and upper slope carbonates in the Late Cretaceous and Paleocene–Eocene, often separated by a disconformity (C-B and B-A1, respectively). Each sequence averages 100–200 m in thickness, except along the steep outer margin, where they are commonly absent because of non-deposition or erosion. By the Oligocene, the region had subsided considerably and deep-water pelagic carbonate sediments were being deposited, particularly on the outer margin. However, nearer land, the productivity of calcareous organisms was high, and prograding wedges of shelf carbonates built outward into deeper water. Ramsay & Exon (op. cit.) mapped three sequences, separated by unconformities or disconformities depending on the location, on the northern Exmouth Plateau and Rowley Terrace: Oligocene, early to middle Miocene, and late Miocene to Recent (A1-A0, A0-AP, AP-seabed, respectively). In deeper water, the combined thickness of the three sequences is less than 500 m. On the shelf, the three prograding sequences usually total 1000–2000 m, with the thickness of individual sequences varying systematically seaward.

History of the Argo Abyssal Plain

The history of the Argo Abyssal Plain (Figs 1 and 6) has been addressed by DSDP Site 261 (Veevers et al., 1974), ODP Site 765 (Gradstein et al., 1990, 1992), studies of magnetic lineations (e.g. Fullerton et al., 1989), and studies of marine seismic profiles (Hinz et al., 1978; Heirtzler et al., 1978; Rao et al., 1994 — this issue; Buffler, 1994 — this issue). According to Fullerton et al. (1989), the oldest magnetic lineation on the abyssal plain is anomaly M26, and the youngest is M16 (Fig. 6). Using the Haq et al. (1987) time scale, this means that the oldest oceanic crust is Oxfordian in age, and the youngest crust preserved south of the Java Trench is Berriasian. Magnetic lineations are mapped as trending ENE, and the complementary fracture zones as trending NNW.

The two drill sites revealed that the oldest oceanic crust, in ODP Site 765, was 155 Ma on the basis of K/Ar dating (Ludden, 1992), and that the oldest dated sediments are Tithonian in both ODP Site 765 and DSDP Site 261. This leaves a 10–15 m.y. hiatus, or highly condensed sediments, between basement emplacement and initial sedimentation. This lack of initial sedimentation may have been caused by eroded sediments being fed back inland from the uplifted and landward-tilted continental margin, as suggested by Ramsay & Exon (1994 — this issue) for the Rowley Terrace area. However, the lack of, or incredibly slow, early sedimentation remains an enigma — in ODP Site 765 only 1.5 m separates basement (Callovian) and the oldest dated sediments (Tithonian). The thickness of sediments on the plain is generally 600–1000 m, and the bulk of them are calcareous turbidites (containing bathyal foraminifers from the continental margin) and claystone. The thickest sequences are Lower Cretaceous and Middle Miocene to Pliocene.

Buffler (1994 — this issue) used a high-resolution ODP seismic line to correlate the drill results at ODP Site 765 with those of ODP Site 261. He showed that there was relatively uniform Early Cretaceous sedimentation of about 350 m of pelagic claystone across the eastern plain. Much of the Neocomian claystone was bentonite derived from ash falls (von Rad & Haq, 1992). Buffler pointed out that the Late Cretaceous to Eocene was a period of condensed sedimentation, with an average deposition of 80 m of alternating claystone and calcareous turbidites and debris flows. The Oligocene to middle Miocene, an average of 150 m thick, was characterised by the deposition of carbonate turbidites, but pelagic claystones were still important. The middle to latest Miocene saw the deposition of about 200 m of carbonate turbidites. From the latest Miocene onward, clay content increased and carbonate debris flows were interbedded with carbonate turbidites; the sequence thins northward from 190 m at ODP Site 765 to 100 m at DSDP Site 261.

The study of Rao et al. (1994 — this issue) is much broader than that of Buffler (op. cit.), being based on a grid of 2800 km of Soviet seismic profiles acquired over the western two-thirds of the Argo Abyssal Plain. Rao et al. (op. cit.) have mapped the principal structural features, and five seismic sequences that have been correlated with the drill sites. The structural grain of much of the plain is mapped as NW to WNW, with a N–S trend in the southwest around the eastern part of the volcanic Joey Rise, and a NE trend in the north on the slope up toward the fore-bulge of the Java Trench. These trends do not agree with those mapped by earlier workers such as Fullerton et al. (1989), and a final decision on structure must await the release of high-resolution satellite data. Structural depressions, with thicker sedimentary successions, have been identified in the oceanic crust along the Exmouth Plateau margin, and beside the eastern Joey Rise, and these must have formed because of down-bowing of the loaded crust. The early deposition was thickest near Australia, but Miocene and Plio-Pleistocene deposition was thickest in the middle of the plain.

In summary, eastern Gondwana broke up in the Callovian, and oceanic crust was formed from then at least until the Berriasian, about spreading axes probably oriented ENE, with fracture zones trending NNW (e.g. Fullerton et al., op. cit.) or NW to WNW (Rao et al., op. cit.). The abyssal plain contains an average of 500–1000 m of sediment, with consistently thin values in the north. The volcanic Joey Rise formed early in the history of the plain. The steadily sinking plain was filled with a mixture of pelagic clay and carbonate turbidites and debris flows; the more of the latter, the more rapid was sedimentation. There was little sedimentation in the Late Jurassic, but relatively rapid deposition (especially in structural depressions in the south and west) in the Early Cretaceous. Sedimentation was slow until the middle Miocene and, from then on, was relatively fast and concentrated in the middle of the plain. The onset of subduction in the Java Trench, possibly in the early Miocene, led to a regional tilt to the south (Buffler, op. cit.), the fracturing of the northern area by northeast-trending faults, and the concentration of sedimentation away from the forebulge of the trench.

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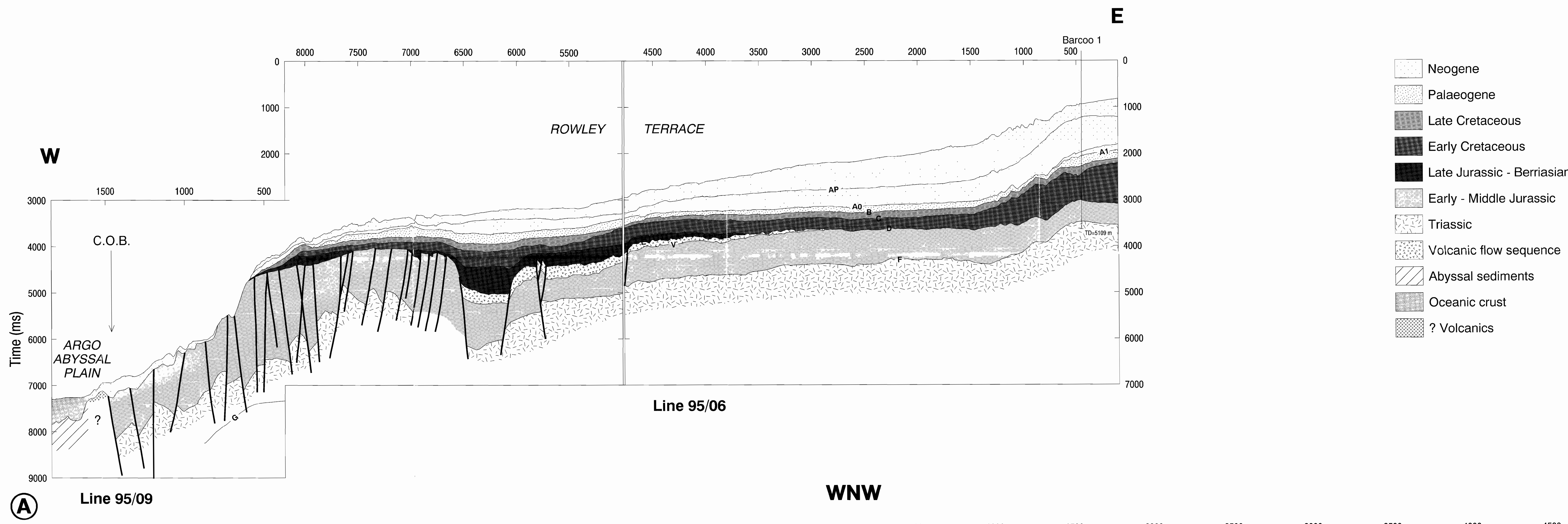

Plate 1A to C [*in* Ramsay & Exon, 1994. *AGSO Journal of Australian Geology & Geophysics*, 15 (1), 55–70]: Line drawings of BMR Cruise 95 regional reflection seismic profiles, across the Australian NW margin.

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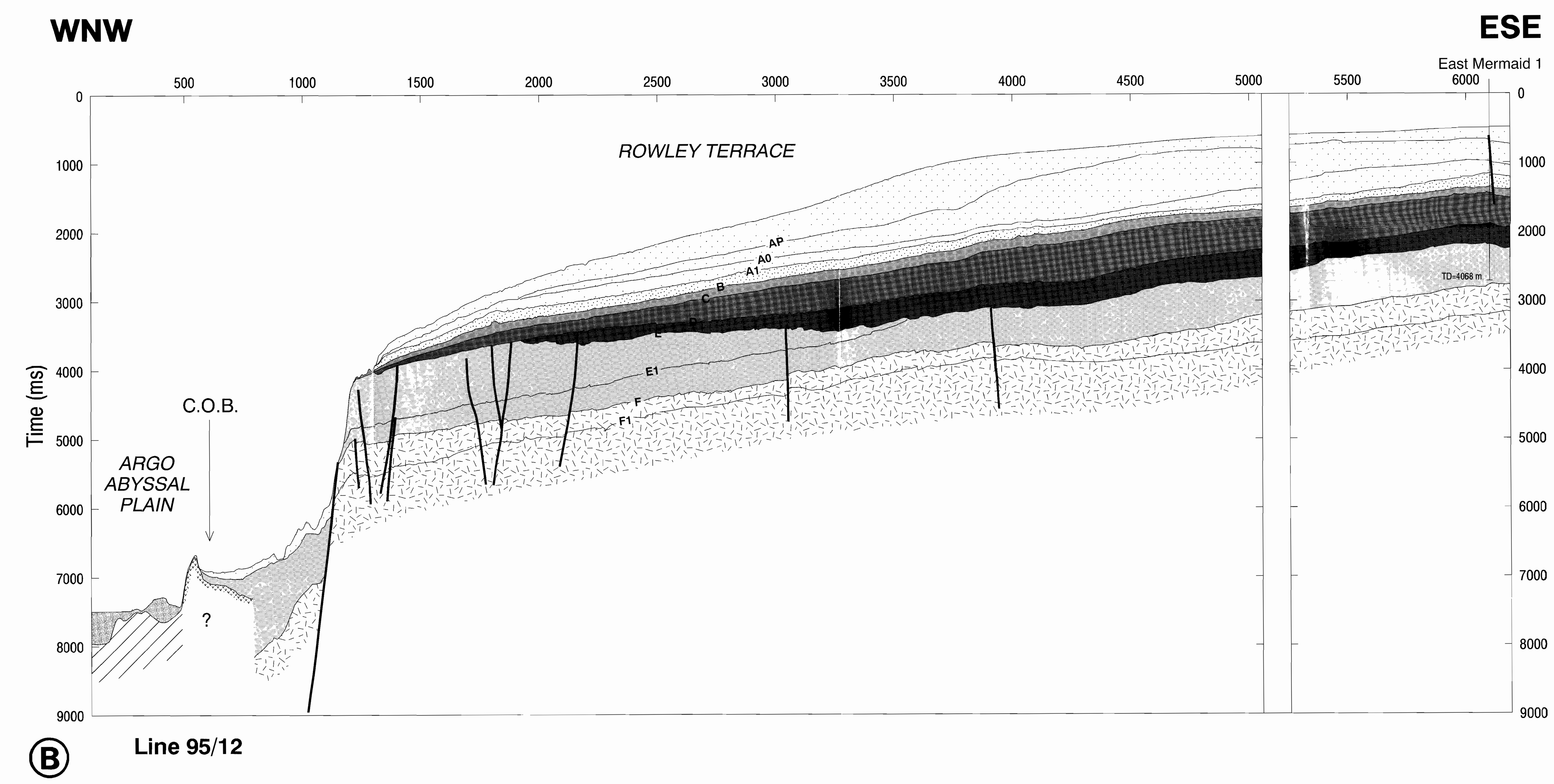


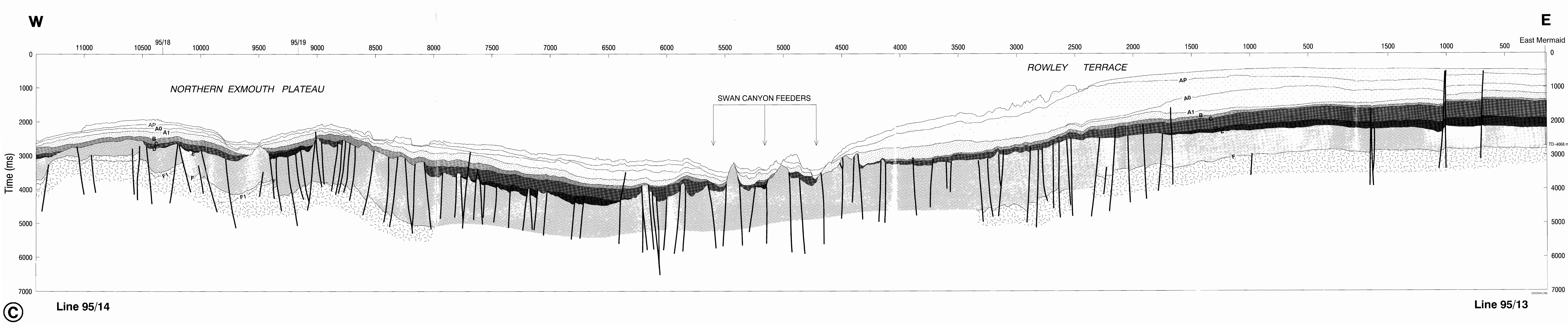
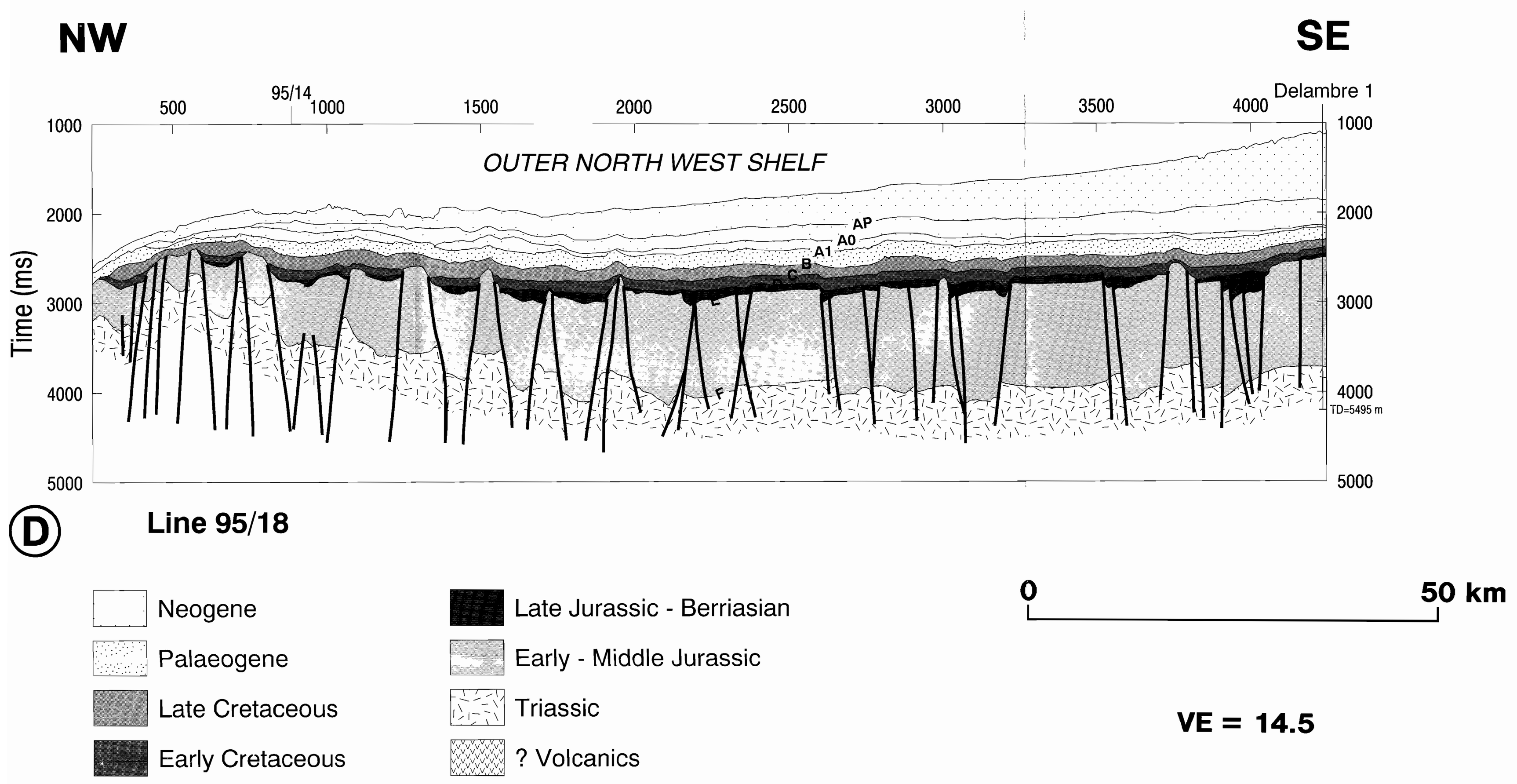
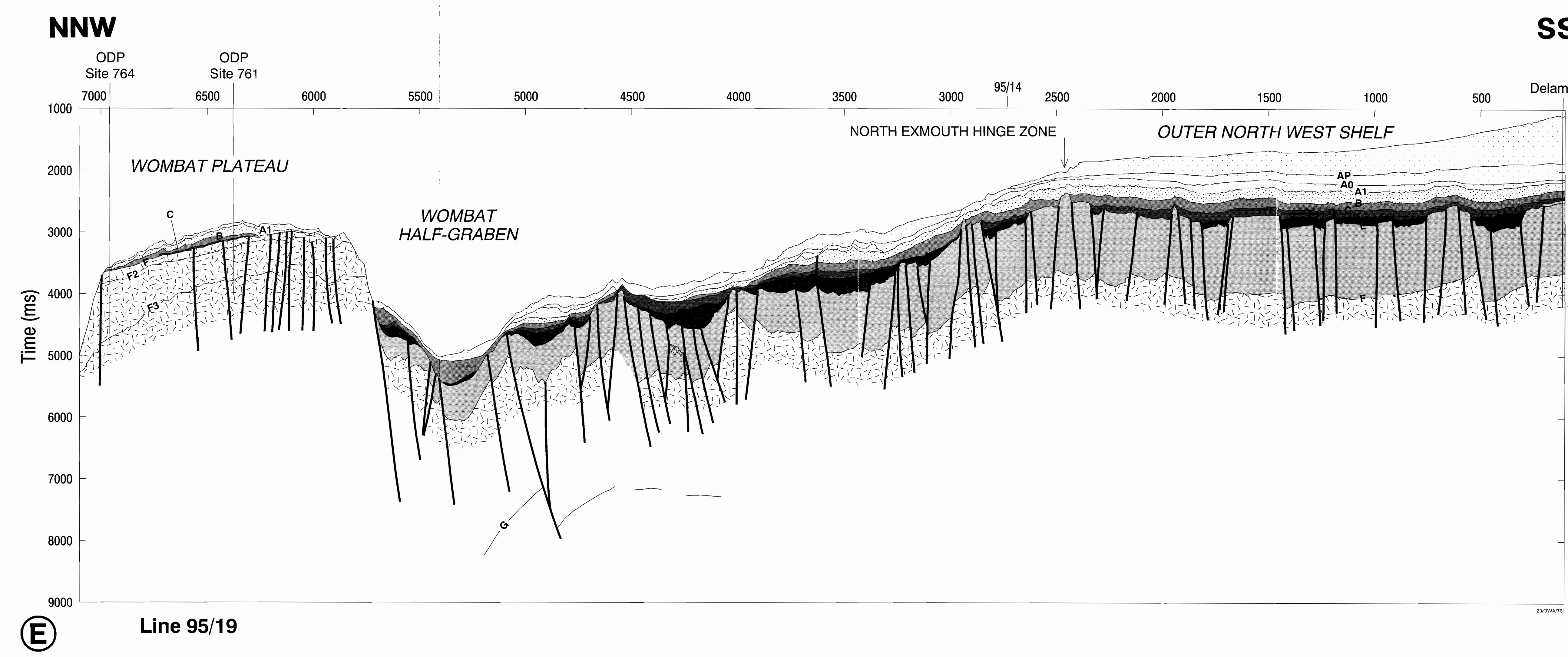


Plate 1D and E [*in* Ramsay & Exon, 1994. *AGSO Journal of Australian Geology & Geophysics*, 15 (1), 55–70]: Line drawings of BMR Cruise 95 regional reflection seismic profiles, across the Australian NW margin.





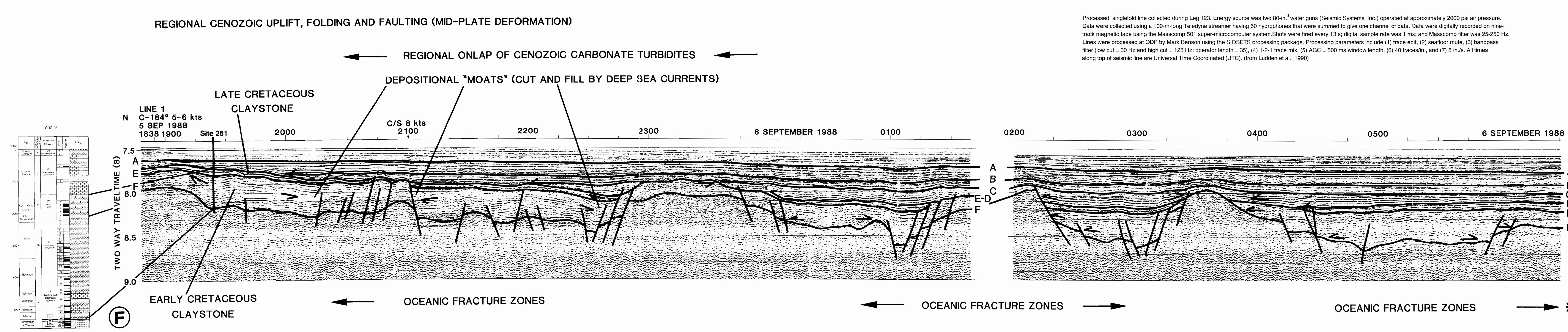
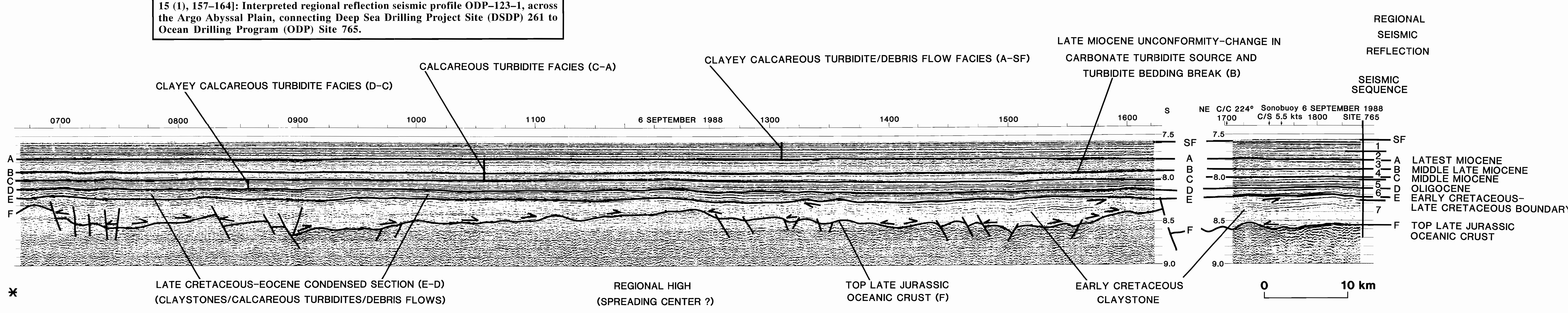


Plate 1F [*in* Buffler, 1994. *AGSO Journal of Australian Geology & Geophysics*, 15 (1), 157–164]: Interpreted regional reflection seismic profile ODP–123–1, across the Argo Abyssal Plain, connecting Deep Sea Drilling Project Site (DSDP) 261 to Ocean Drilling Program (ODP) Site 765.



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Neville F. Exon (Associate Editor)

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