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Front cover:

The continental margin of Australia can be mapped in terms of zones of oceanic crust generated during a number of separate seafloor-spreading episodes. The cover shows the present state of part of the margin, with the youngest crust next to an east-west spreading ridge, and older units further from it. The development of the whole continental margin is reviewed in a paper by Falvey & Mutter in this issue.

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Regional plate tectonics and the evolution of Australia's passive continental margins

David A. Falvey[1] *& John C. Mutter*[2]

The divergent, passive continental margins around Australia have evolved through the progressive dissection of eastern Gondwanaland in five separate seafloor-spreading episodes. The earliest episode occurred 155 m.y. ago off northwestern Australia. The latest episode started 55 m.y. ago south of Australia and continues on the Southeast Indian spreading ridge to the present. The geological and structural evolution of each passive margin was a protracted process. The onset of seafloor spreading (breakup) was preceded by more or less extensive sedimentary basin subsidence. Non-volcanic rift-grabens, with fairly clear boundary faults, began evolving 40-50 m.y. before breakup. Such rifting was often preceded by a broader intracratonic style of basin subsidence straddling the incipient continent-ocean boundary between 50 and 100 m.y. before breakup. Rift and infrarift sedimentation rates declined exponentially towards the breakup unconformity. Postbreakup subsidence was again rapid, but exponentially decreased with time. Immediate postbreakup sedimentation was usually interrupted by submarine erosion in the shallow, but rapidly subsiding, ocean basin. Prograding marine shelf conditions prevailed from about 20-40 m.y. after breakup along all margins.

The dominant driving mechanism of postbreakup subsidence seems clearly to be lithospheric thermal contraction caused by the removal, by seafloor spreading, of a thermal anomaly from beneath the continental margin. The driving mechanism of the rift-phase subsidence is far less clear. Simple surface observations do not unambiguously discriminate between sudden pull-apart rifting, finite rate stretching, and deep crustal metamorphic contraction. All processes may occur to at least some extent and dominate at different times. However, the apparent absence of both extensive rift volcanics, and transform/transcurrent faulting at rift offsets, plus the non-conformity of rift, breakup, and basement structural patterns, argue against stress-driven stretching as a crucial mechanism. Specific model predictions of crustal structure, heat flow and cumulative sedimentation pattern favour a dominantly metamorphic subsidence mechanism, although this fails to account fully for some thin continental crust near continent-ocean boundaries.

Introduction

Passive continental margins evolve through initiation of seafloor spreading and divergent plate motion within a pre-existing continent. There is specific evidence in the Australian region that this has occurred at least through the Mesozoic and Cainozoic. The kinematic evolution of a continental margin is a relatively straightforward process: at a discrete time, marked by the oldest seafloor-spreading magnetic anomaly, continental breakup begins, and an ocean basin forms at an active, spreading mid-ocean ridge. The geological and structural evolution of a continental margin is, however, an extremely complex and protracted process. While this has been recognised to greater or lesser extent ever since the concepts of continental drift were first enunciated, the majority of evolutionary schemes proposed have been largely qualitative and generally have not been based on studies of passive margins themselves (examples are Wegener, 1929; Du Toit, 1937; Heezen, 1960; Dewey & Bird, 1970).

Falvey (1974) envisaged an evolutionary scheme, based on the Australian southern margin, which involved three distinct phases.

- Initially, intracratonic basins and then rift basins developed as regions of major basement subsidence, generally in the vicinity of an incipient breakup axis. The rifting occurred some 50 m.y. before the initiation of a divergent plate boundary (the rift-valley phase). Subsidence and deposition rates generally decreased towards breakup time.
- During a 5-10 m.y. period near the time of breakup, the intensity of relative uplift and then subsidence increased (the breakup phase).
- After seafloor spreading had been active for a few million years, subsidence and deposition rates again generally decreased with sedimentation showing a marked increase in marine influences (the postbreakup phase).

These evolutionary phases appear to provide a generally satisfactory description of the development of most parts of Australia's rifted margins (details are discussed below).

The mechanism of continental margin formation has been described in terms of models which usually relate cycles of uplift and subsidence to the thermal evolution of continental and oceanic lithosphere (Dewey & Bird, 1970; Sleep, 1971; Falvey, 1974; Mutter, 1978). However, such tectonic models differ markedly, and appear strongly dependent on the type of continental margin chosen by a particular author for analysis. Thus, the emergence of a general continental margin model has been hindered by apparent geological and structural complexity, and also by the limited depth of penetration and resolution of relevant marine geophysical methods.

There are major advantages to the choice of the Australian region as a study area for continental margin problems (Fig. 1):

1. The progressive dispersal of Gondwanaland from around continental Australia has provided a

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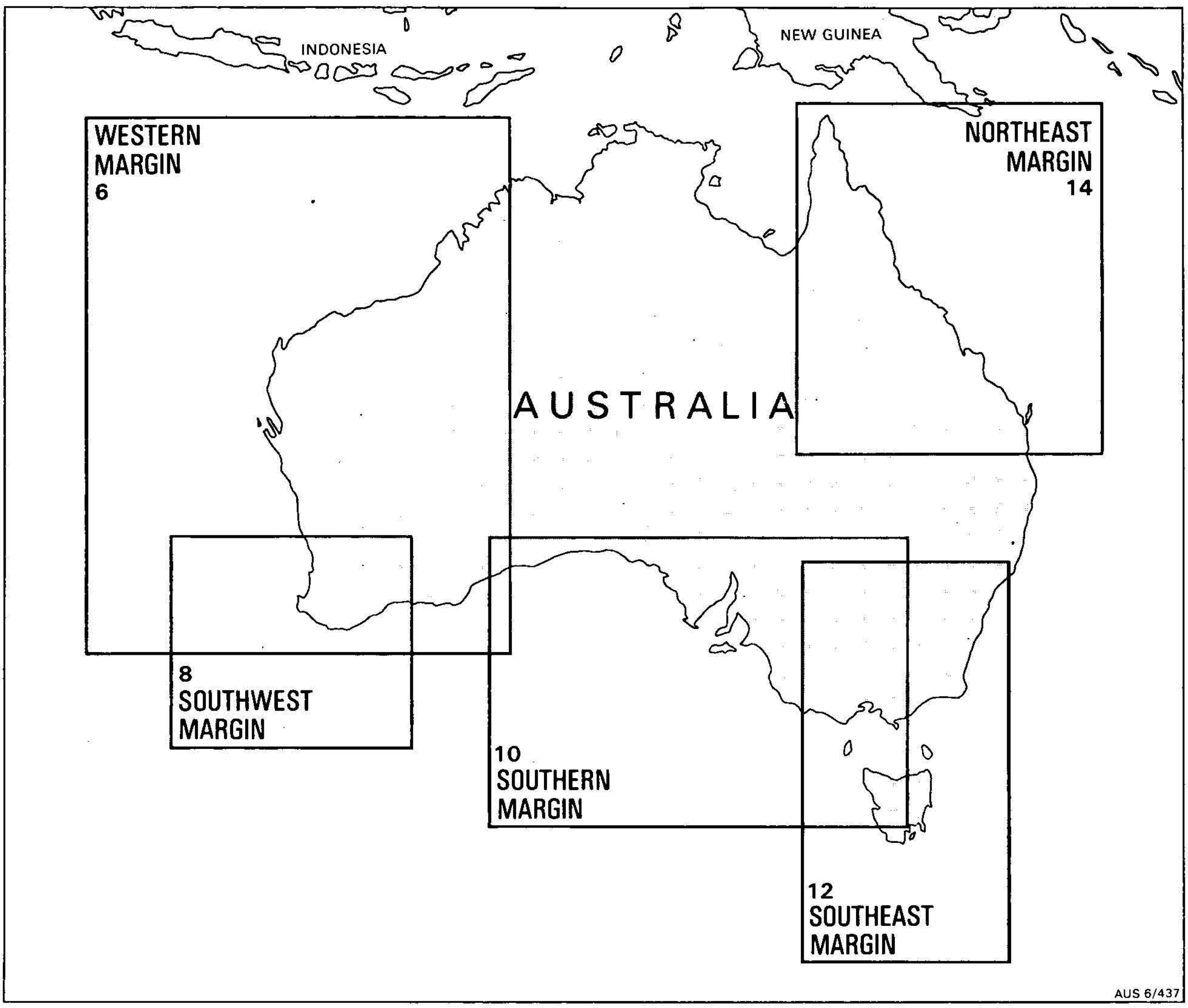

Figure 1. The framed areas cover the separate margin segments described in the text.
The number within each frame refers to Figure number of the structural maps with bathymetric contours, which have been compiled for each of the areas.

sequence of continental margins with diverse breakup times:

a. Northwest—Late Jurassic (155 m.y. B.P.)
b. Southwest—Early Cretaceous (120 m.y. B.P.)
c. Southeast—Late Cretaceous (80 m.y. B.P.)
d. Northeast—Early Palaeocene (65 m.y. B.P.
e. South—Early Eocene (55 m.y. B.P.)

2. The kinematic evolution of the ocean basins adjacent to these margins is fairly well understood (Figs. 2-5).
3. For various climatological and oceanographic reasons, postbreakup phase deposition on these continental margins consists dominantly of thin carbonates. This alleviates many problems involving seismic resolution and depth of penetration.
4. Fairly extensive oil search coupled with government policy on open file and public release of information has provided a comprehensive set of data on all these margins.

In this paper we review the kinematic, geological, and structural evolution of Australia's five continental margins, by interpretation of all available and relevant data. We describe and correlate key time-stratigraphic events and establish broad constraints as a basis for future tectonic models.

Regional plate tectonics and the breakup of eastern Gondwanaland

Major orogenic episodes of probable plate tectonic origin affected eastern Gondwanaland, including Australia, throughout the Palaeozoic. During the Mesozoic the southern supercontinent began to break up, with the formation of the Indian Ocean, Tasman and Coral Seas, and, finally, the Southern Ocean. These ocean basins formed at normal spreading ridges (divergent plate boundaries) in the following pattern.

(a) Northwest (Fig. 2)

At approximately 155 m.y. B.P. (Late Jurassic), seafloor spreading commenced in the Argo Abyssal Plain (anomaly M-25 time). The northeast-trending spreading pattern was initially described by Falvey (1972a). Basin age was established by DSDP drilling (Veevers, Heirtzler & others, 1974), and anomaly identification by Larson (1975). Recent work by Heirtzler & others (1978) has confirmed this pattern. Throughout the Jurassic, rift-graben tectonics affected the entire western margin.

(b) Southwest (Fig. 3)

At approximately 120 m.y. B.P. (Early Cretaceous), spreading began in the Cuvier and Perth Basins

Figure 2. Reconstruction of eastern Gondwanaland at 150 m.y. B.P., with respect to Australia.
Western Australia is reconstructed with India (after Falvey, 1972a, and Larson & others, 1979); southern Australia with Antarctica (after Deighton & others, 1976); and eastern Australia with east Papua (Taylor & Falvey, 1977); with the Lord Howe Rise (after Shaw, 1970); and with New Zealand (after Weissel & others, 1977). Major rift faults and thick rift-phase sedimentary accumulations (shaded) are also shown (see Figures 6, 7, 8, 9, and 10).

(Anomaly M-3 to M-10 time). Portions of the north-northeast-trending spreading pattern were initially described by Markl (1974) and Larson (1977), and have been integrated recently by Larson & others (1979). Details of the spreading pattern around the southwest corner of Australia may be complex (Markl, 1978), and have not been fully described. Through the earliest Cretaceous, late-stage rift-graben tectonics affected the southwest margin bordering India and Antarctica. Early Cretaceous rifting also affected eastern Australia—the Gippsland Basin, Lord Howe Rise, and Queensland Plateau. The earliest phase of rifting also began on the southern margin adjacent to Antarctica.

(c) Southeast (Fig. 4)

At approximately 80 m.y. B.P. (Late Cretaceous), spreading began on the Southwest Pacific spreading ridge (between Antarctica and the Campbell Plateau) and on the Tasman spreading ridge (Anomaly 34 time). This has been described by Hayes & Ringis (1973), Weissel & others (1977), and Shaw (1978). The New South Wales and southern Queensland margin were not affected by seafloor spreading until 65-70 m.y. B.P., when the ridge jumped to the western side of the Dampier Ridge (Ringis, 1973; Jongsma & Mutter, 1978; Shaw, 1979). Rifting affected the eastern margin through the late Cretaceous. Major rift-graben tectonics affected the region between Australia and Antarctica.

(d) Northeast and south (Fig. 5)

The Tasman spreading ridge propagated northwards, forming the Cato Trough and the Coral Sea Basin from 65 m.y. B.P. (Palaeocene; Anomaly 29 time). Seafloor spreading ceased on this Tasman-Coral Sea plate boundary at 55 m.y. B.P. The pattern was initally recognised by Falvey & Talwani (1969) and Taylor & Falvey (1977), and identified and integrated with the Tasman Sea teconic pattern by Weissel & Watts (1979) and Shaw (1979). Seafloor spreading between Australia and Antartica commenced at about 55 m.y. B.P. (Anomaly 22 time), a short time after spreading ceased in the Tasman and Coral Seas. The tectonic pattern was initially described by Weissel & Hayes (1972), and refined by Weissel & others (1977). The reconstruction adopted here is that of Deighton & others (1976). Seafloor spreading south of Australia also marks the onset of marginal basin spreading behind western Pacific Island arcs. Spreading in the New Hebrides Basin also began about 55 m.y. B.P. (Weissel & others, in press; Falvey, in prep.).

Structure and stratigraphy of Australia's rifted and passive margins

The following is a discussion of the structural and stratigraphic patterns on the five segments of Australia's passive margins, which experienced divergent tectonics at different times. The discussions are largely summaries of published literature: acknowledgement of which is given in the figure captions. For the sake of brevity, reference to source articles is avoided in the text. Veevers (1981—in press) has also produced a current review of Australian passive margins.

Western margin

The earliest phase of continental margin formation occurred off Western Australia (Figs. 2, 3, and 6). The deep continental margin comprises both normal lower

Figure 3. Reconstruction of eastern Gondwanaland at 110 m.y. B.P.
Seafloor-spreading pattern in the Indian Ocean is based on Larson & others (1979).

continental slope and a number of large marginal plateaus, separated from the shelf and upper continental slope by shallow troughs or saddles. The plateau margins reflect transform offsets in seafloor spreading. Intracratonic and fault-bounded (rift) sedimentary basins predating breakup occur beneath the shelf and continental margin, and parts of some basins are exposed in coastal plains. A broad, intracratonic phase of basin subsidence commenced at least as early as Permian times, and some related activity may have occurred before the Carboniferous. Sedimentation continued throughout the Jurassic in rift-grabens or asymmetric half-grabens bounded by normal down-to-basin faults. Major rift-basin width is generally 50-75 km, and Jurassic sedimentary sections are up to 8 km thick (Fig. 7(a)—profile 3). The margin as a whole can be seen as a complex ancient rift system which culminated in the breakup of eastern Gondwanaland.

Figures 6, 7a and 7b, illustrate typical margin structures and sedimentation, and show the continuity of major structural elements along the entire western margin. The structure profiles shown in Figure 7a are taken from published descriptions of sedimentary basins. They are reproduced here at the same scale (except profile 4, which has approximately twice the vertical exaggeration of other profiles, to better illustrate depositional relationships in the Dampier Sub-Basin).

All profiles in Figure 7a show a consistent style of development of the prebreakup continental margin basins. To varying degrees, each profile shows prebreakup basins bounded on their eastern sides by a major normal fault or fault system. The profiles have been stacked to align these principal boundary faults. To a lesser extent, the western limits of these basins are fault controlled. The thickness of pre-Cretaceous sediments

Figure 4. Reconstruction of eastern Gondwanaland at 70 m.y. B.P.
Seafloor-spreading pattern in the Tasman Sea is based on Shaw (1979); and in the southwest Pasific Ocean, on Weissel & others (1977). The Pacific-Lord Howe Rise convergent plate boundary is based on work by Falvey (in prep.).

Fig. 5. Reconstruction of Australian, Antarctic and Pacific Plates at 45 m.y. B.P.
Seafloor-spreading patterns south of Australia and New Zealand are based on Weissel & others (1977); in the Tasman Sea, on Shaw (1979); in the Coral Sea, on Weissel & Watts (1979) and Shaw (1979). The reconstruction of the Melanesian arc off New Caledonia is based on palaeomagnetic reconstructions by Falvey (in prep.).

changes by an order of magnitude from east to west across these faults, usually in less than 10 km distance. The rift basins north of 25°S form a sublinear en-echelon trend, which is clearly reflected in the gravity map of Australia (BMR, 1976).

At about 25°S the structural trend changes abruptly and the shallower parts of the margin and the coastal regions are occupied by intracratonic Palaeozoic basins. The Exmouth Sub-Basin (Fig. 7(a)) appears to cross from continental shelf to slope, then narrows and disappears. The Houtman Basin (Symonds & Cameron, 1977) occupies the upper continental slope south of 25°S, and trends nearly at right angles to the northern basins. The structural grain trends more nearly north-south through the Perth Basin; and is apparently similar to the onshore Palaeozoic basin north of the Perth Basin, in the region from 23-29°S.

Horst systems often dissect or bound the western side of the Jurassic rift basins. The type example is the Rankin Trend, forming the western margin of the Dampier Sub-Basin. These form classical structural-stratigraphic hydrocarbon traps. Each basin shows an outer high or horst block which resembles the Rankin structure. Figure 6 illustrates the continuity of the outer high structures along the entire northwest margin. The structural picture south of 25°S is less clear. Symonds & Cameron (1977) did not identify a major boundary fault limiting the Houtman Basin (though they suggested that one is probably present) nor did they describe an outer structure equivalent to the Rankin platform. However, the general structural picture may be similar to basins further north.

In most of the basins the outer horst is accompanied to the west by a secondary uplift which divides the

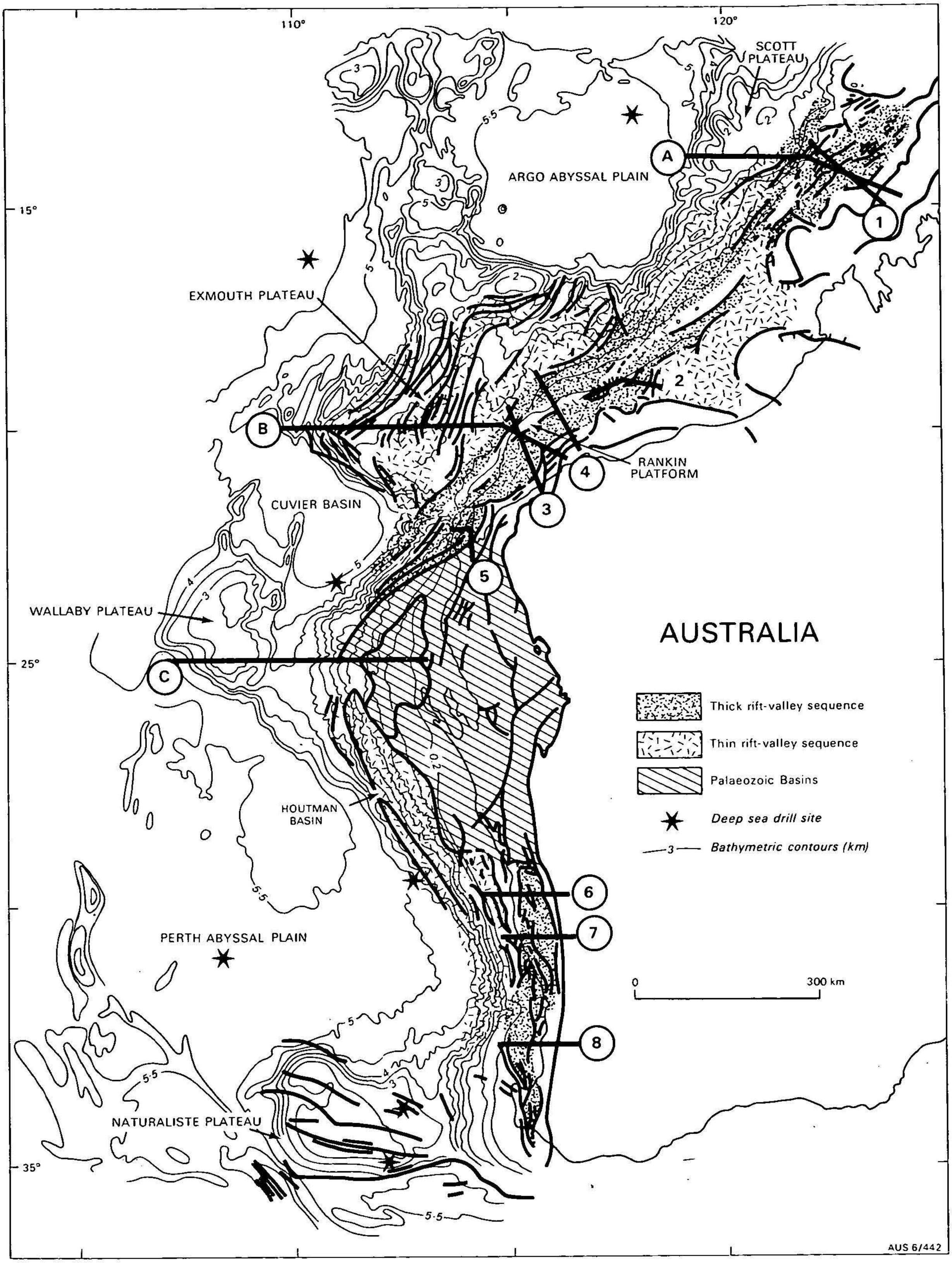

Figure 6. Major structural trends and depositional provinces on Australia's western continental margin.
Bathymetric contours are compiled from charts presented by Markl (1974), Falvey & Veevers (1974), and Heirtzler & others (1978). The structural trends include fault traces, fold axes, hinge lines and outlines of tectonic provinces as presented by the following authors: Allen & others (1978), Crostella (1976), Crostella & Chaney (1978), Exon & Willcox (1978), Halse (1976), Stagg (1978), Symonds & Cameron (1977), Thomas (1978), Thomas & Smith (1976), Warris (1976). The solid bars with circled numbers or letters alongside show the locations of structural profiles in Figures 7a and 7b.

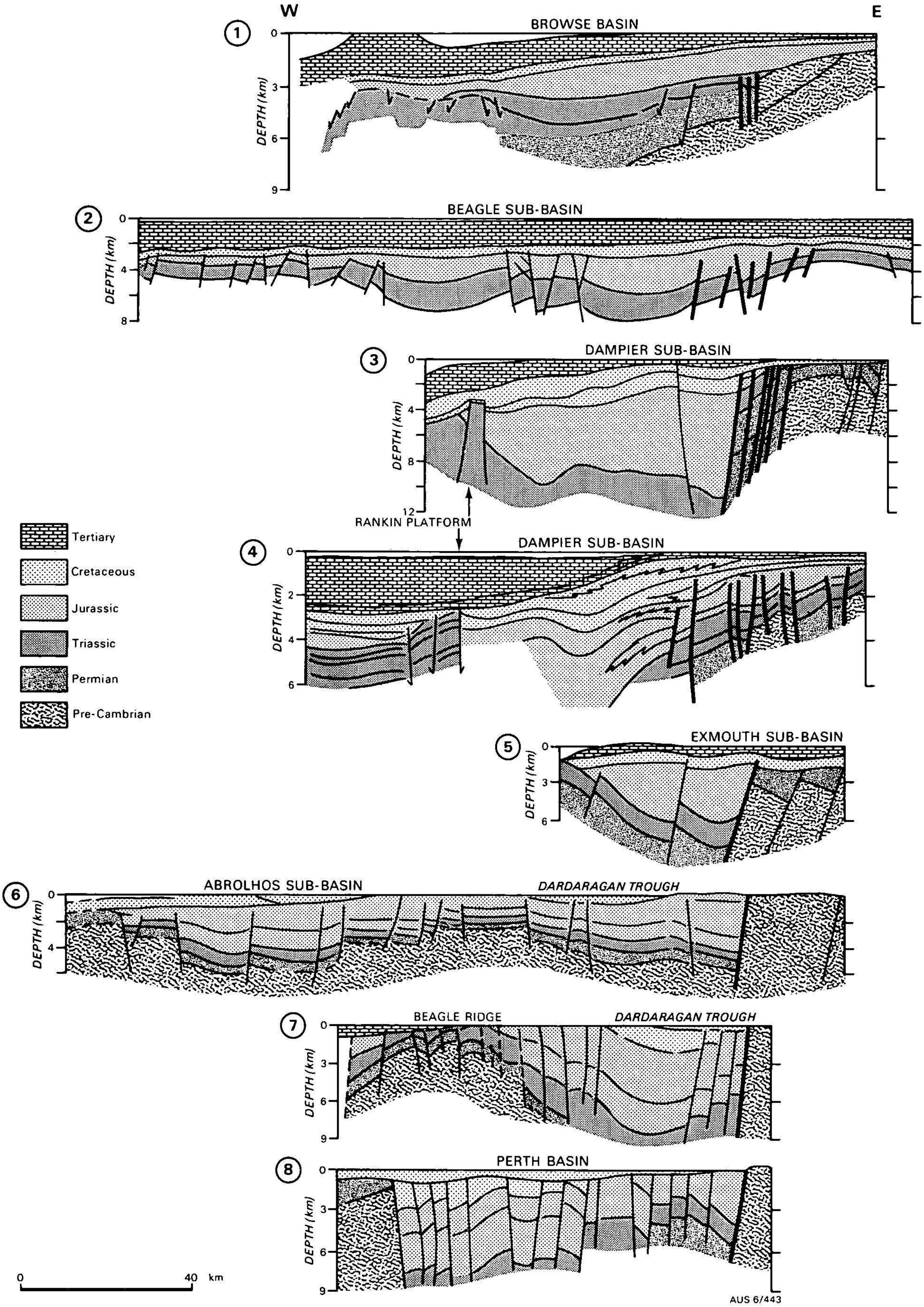

Figure 7(a). Structure profiles of sedimentary basins in the shallow margin off Western Australia.
They were obtained from the following sources: Profile 1—slightly modified after Crostella (1976) by reference to Allen & others (1978). Profile 2—from Halse (1976). Profile 3—from Thomas & Smith (1976). Profile 4—from Veevers (1974). Profiles 6, 7, and 8—from Jones (1976).

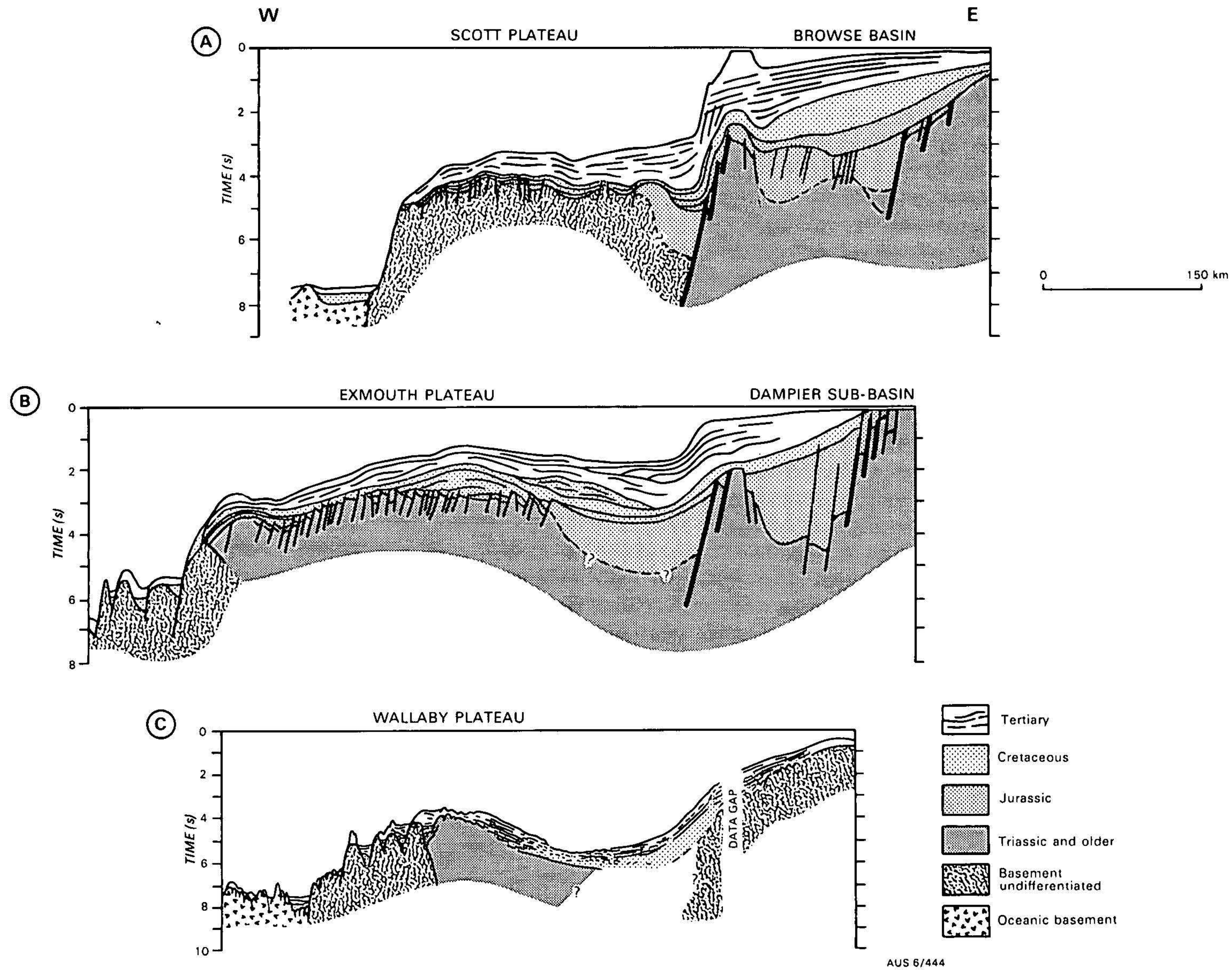

Figure 7(b). Structure profiles of the marginal plateaus and shallow continental margin off Western Australia.
Profiles A and B are based on drawings presented by Powell (1976). Profile A has been modified to make the stratigraphic interpretation consistent with that of Allen & others (1978) and Stagg (1978) for the Browse Basin. Profile B was modified to include the interpretation of Exon & Willcox (1978) for the Exmouth Plateau and to make the profile of the Dampier Sub-Basin the same as that given by Thomas (1978) and shown in profile 3. The deep-water parts of profiles A and B are line drawing interpretations of seismic reflection profiles obtained by the Bureau of Mineral Resources using a Sparker source. These data were also used to construct profile C. Petkovic (1975) presented a reproduction of the original reflection profile C, from which we have made a line drawing and included stratigraphic and structural boundaries to be largely consistent with the interpretation of the same profile given by Symonds & Cameron (1977).

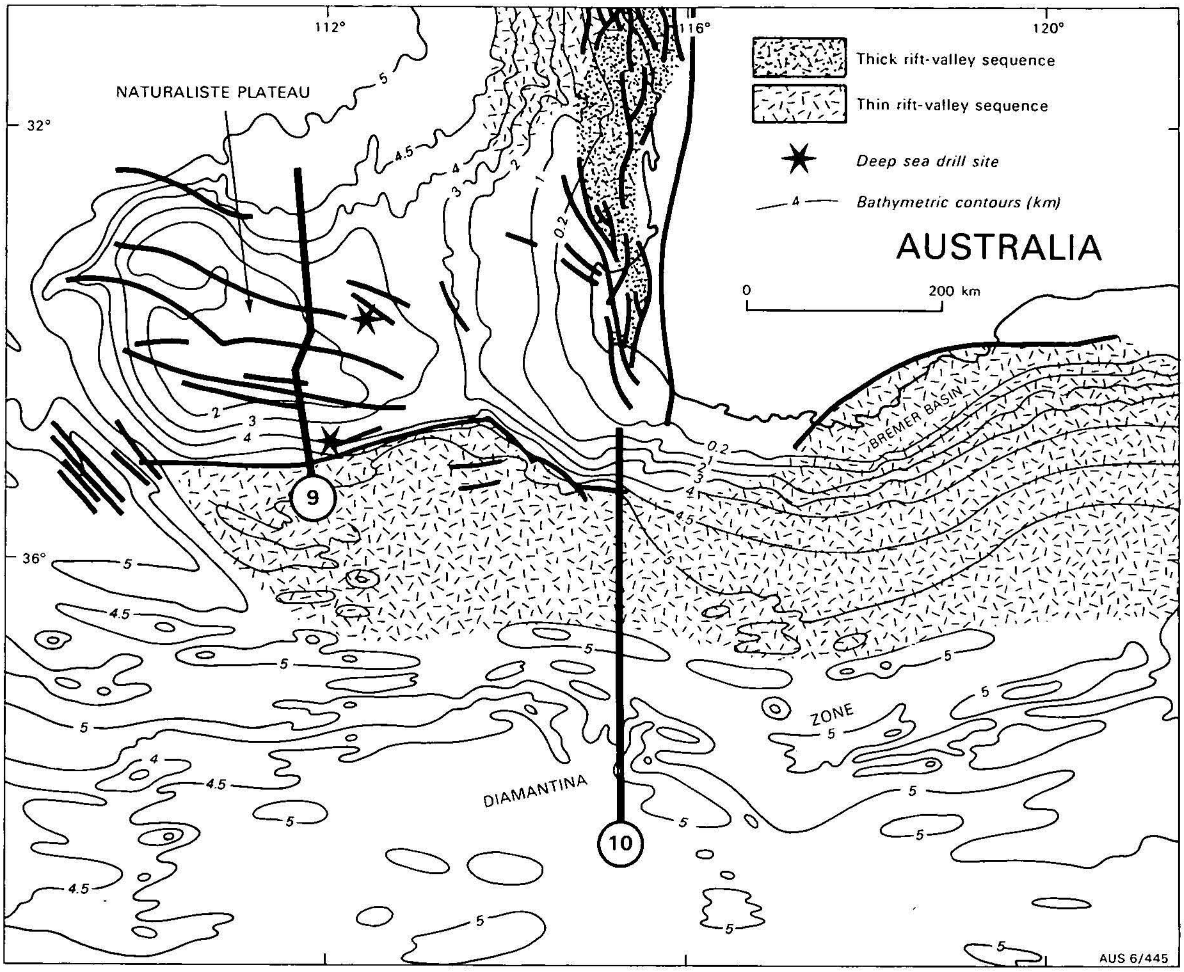

Figure 8. Bathymetry and principal structural and depositional provinces on the southwest margin of Australia. Bathymetry is from Hayes & Conolly (1972). Structural trends are from Jones (1976), Jongsma & Petkovic (1977), and Cooney & others (1974). The bars with circled numbers indicate the location of structure profiles shown in Figure 9.

basins into two parallel troughs (in some cases the troughs have been given separate names). The depocentre of the basins appears to have shifted landward as the outer high rose during the Early Mesozoic. The distribution of Permian and Triassic strata is not well well known in the central and outer parts of the basins. The Permo-Triassic sequences apparently formed in a broad shallow basin, largely confined to the west of the boundary faults, and the subsequent development of uplifts which delineate the mid-Mesozoic basins occurred within this broad basin, not along existing high trends. There is little evidence of re-activation of ancient structures, either in generating boundary faults or basin margin highs. The outer highs generally appear to be growth structures, having experienced repeated uplift during deposition in the adjacent basins.

The deep margin off Western Australia is dominated by three large marginal plateaus—the Scott, Exmouth, and Wallaby Plateaus. The structure and sediment distribution on these are fairly well known from seismic reflection profiling. The stratigraphy has been inferred from correlations with the shallow margin basins described above and ties of seismic reflection data to exploration drilling in those basins. The crustal structure of the majority of plateaus is generally considered continental (Falvey, 1972b).

The Exmouth Plateau has been extensively described (Falvey, 1972b; Falvey & Veevers, 1974; Veevers & others, 1974; Hogan & Jacobsen, 1975; Powell, 1976; Exon & others, 1975; Exon & Willcox, 1978; Willcox & Exon, 1976; Wright & Wheatley, 1979). Veevers & others (1974) have mapped the prebreakup fault pattern over most of the Exmouth Plateau (Figure 6), demonstrating that it closely parallels structural patterns of the shallow margin basins. The lower slope and rise below the plateau surface consist of a deeply faulted basement with little or no prebreakup sedimentary section. The thickest prebreakup sedimentary section underlies the physiographic trough which separates the plateau from the shallow margin. Here, Lower to Middle Jurassic rift-phase strata make up a large part of the total Mesozoic section. There are various views as to the timing of faulting and subsidence. We believe that the data reviewed provide compelling evidence for a relatively elevated central and outer plateau in all three cases. There is also inconclusive seismic reflection evidence of shallow-water sedimentary processes at work near the central Exmouth Plateau in the immediate post-breakup sedimentary section. Of course, the definition of breakup time varies depending upon which part of the plateau is considered (see Figs. 2 and 3).

Von Stackelberg & others (1980) have reported 30 dredge hauls from the outer slopes of the Exmouth Plateau. More than half contained Jurassic and Triassic prebreakup shallow-water sediments. Four dredges also contained immediate to acid volcanics dated at about the time of rift onset. This suggests limited con-

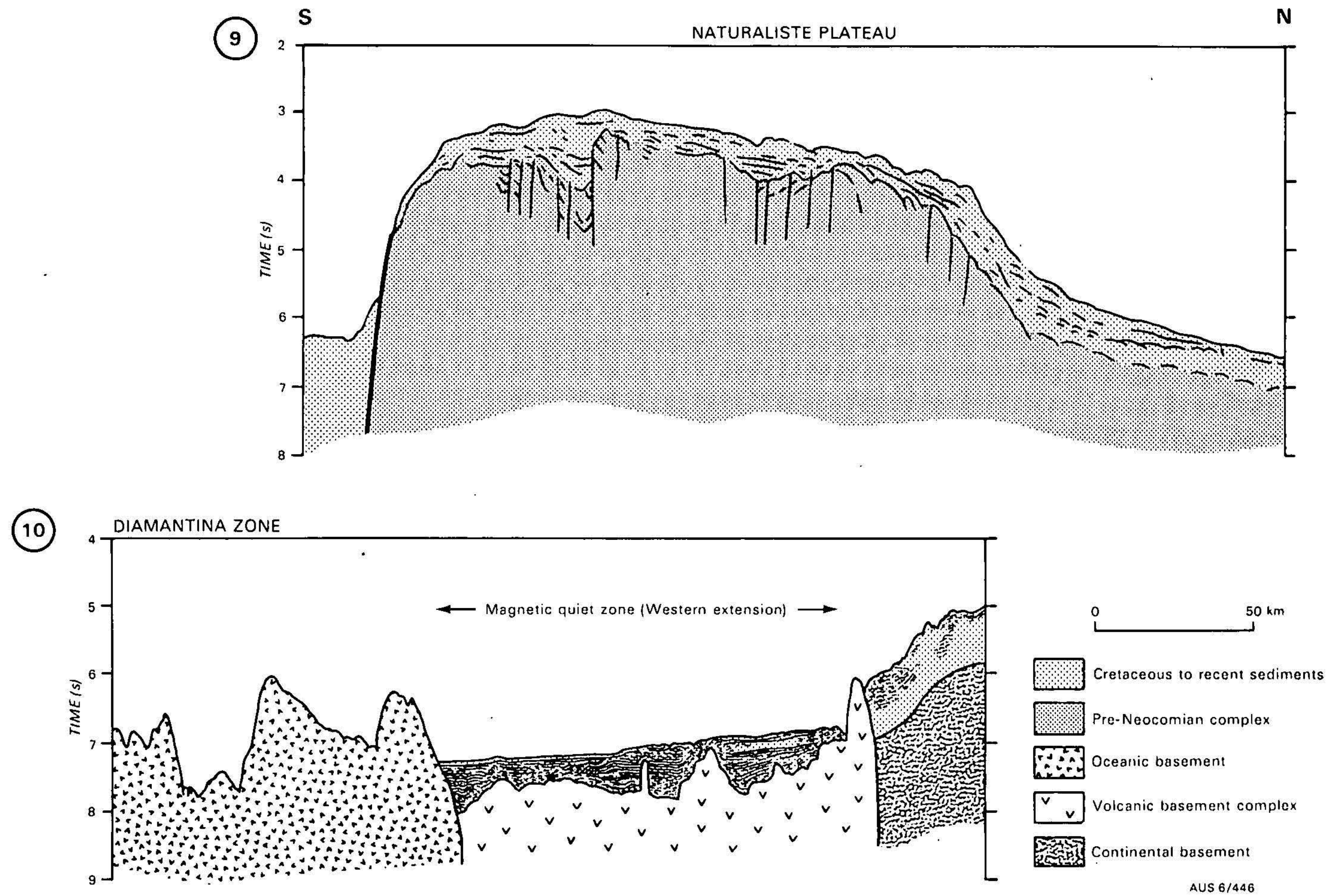

Figure 9. Structure profiles of the Naturaliste Plateau and the western part of the Australia's southern continental margin.
Profile 9 is taken directly from Jongsma & Petkovic (1977), profile 10 is from Mutter (1978). The horizontal and vertical scales are the same for both, and both are interpreted line drawings of single channel seismic reflection profiles obtained by the Bureau of Mineral Resources in 1972 using a sparker energy source. Stratigraphic divisions shown on profile 9 were made with reference to Deep Sea Drilling on the Naturaliste Plateau. The location of the profiles can be obtained from Figure 8.

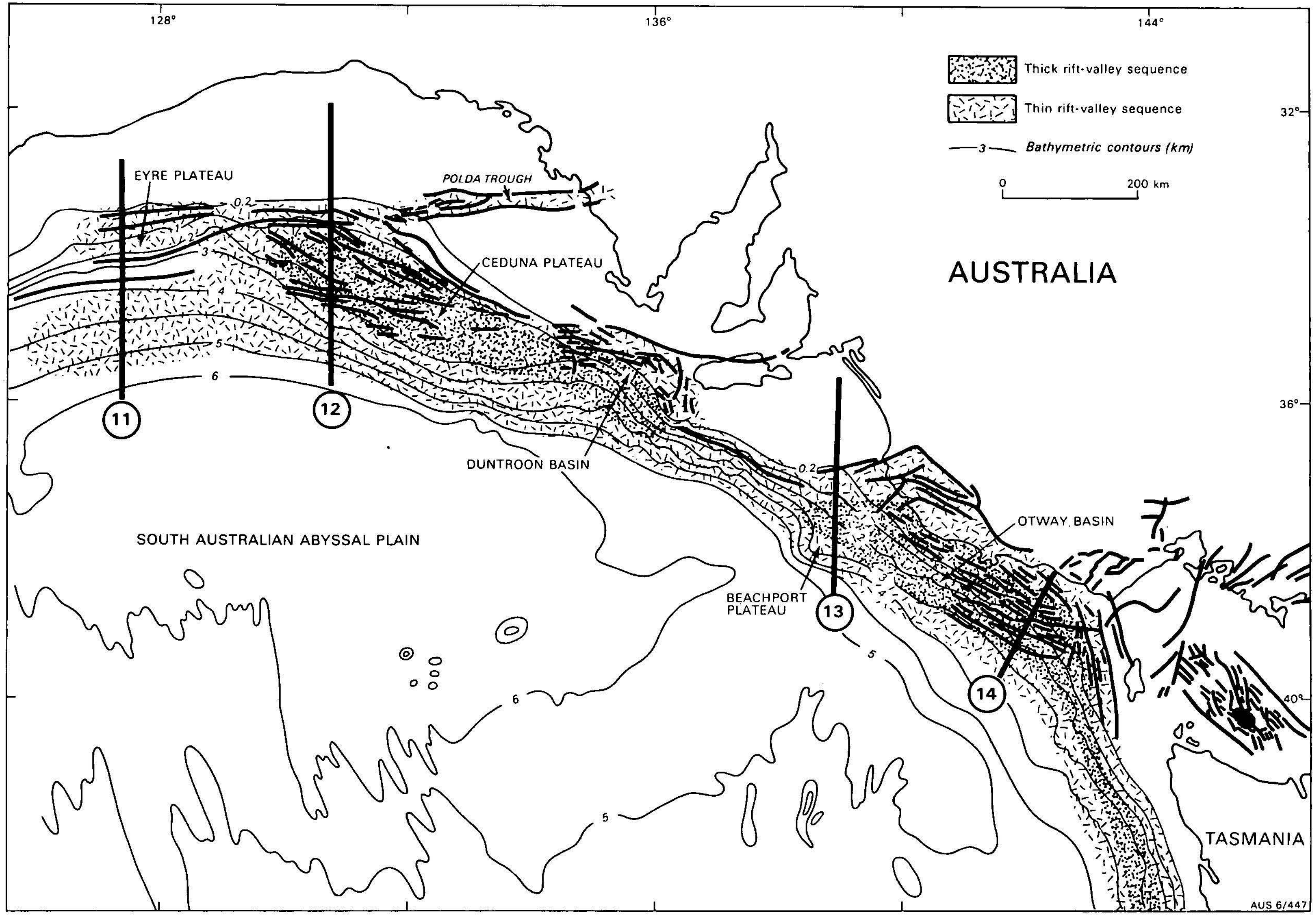

Figure 10. Bathymetry and principal structural and depositional patterns of Australia's southern continental margin.
The bathymetric contours are from Hayes & Conolly (1972). Structural trends include fault traces, fold axes and hinge lines and are compiled from the following sources: Abele & others (1976), Denham & Brown (1976), Douglas (1976), Brown (1976), Ellenor (1976), Fraser & Tilbury (1979), Hocking (1976), Kenley (1976), Mutter (1978), Pattinson & others (1976).

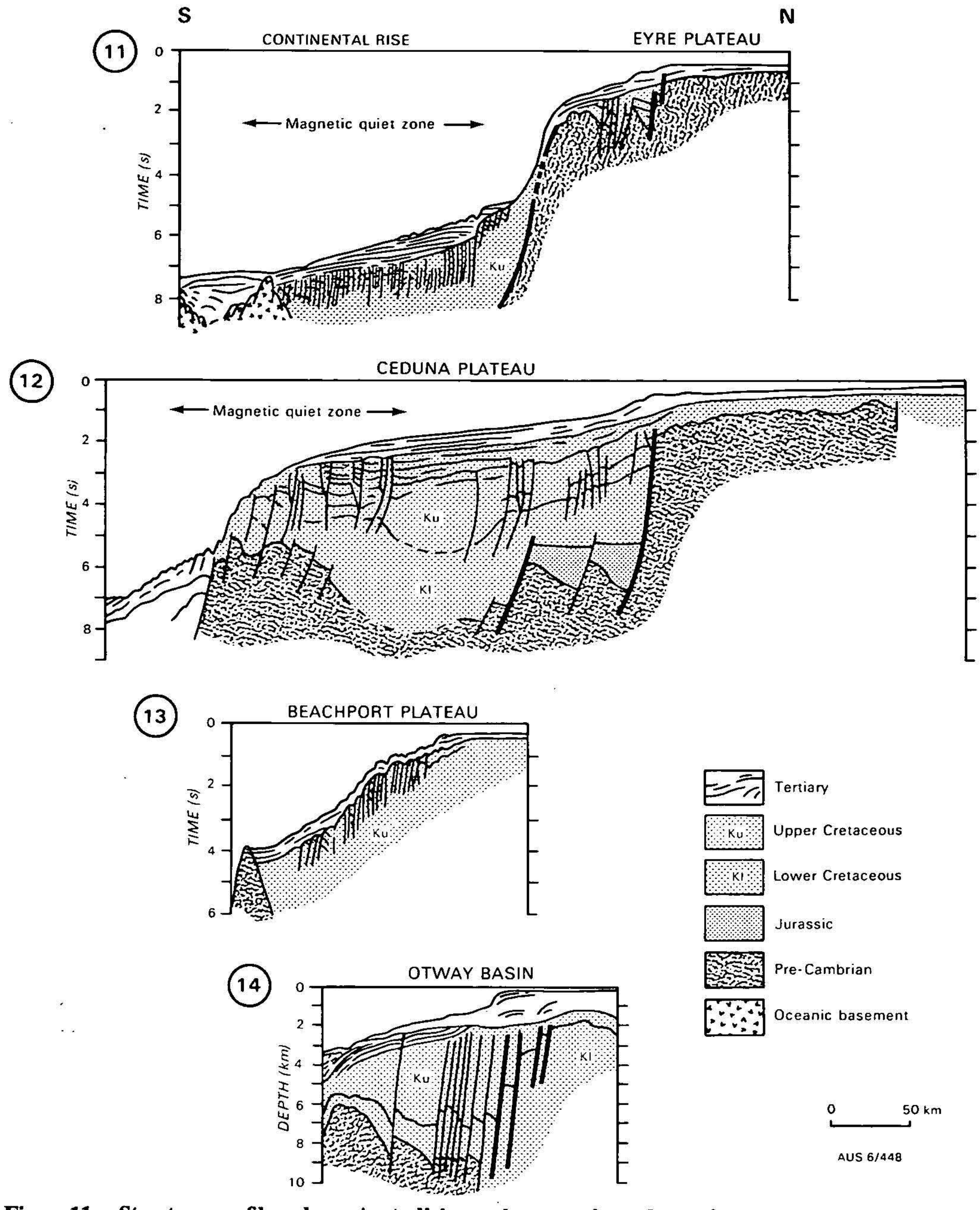

Figure 11. Structure profiles along Australia's southern continental margin.
The sources are as follows: Profile 11—original line drawing made by Talwani & others (1979), from which the extent of the magnetic quiet zone is also taken. Willcox (unpublished) has made a line drawing of the same profile with somewhat different structures and stratigraphic divisions. Profile 12—original line drawing was made by Willcox (unpublished)—has had stratigraphic divisions revised according to results presented by Talwani & others (1979) and the discussion of Mutter (1978), and includes Jurassic strata as suggested by Boeuf & Doust (1975) and supported by Fraser & Tilbury (1979). Profiles 11, 12, and 13 are interpreted line drawings of seismic reflection profiles obtained by the Bureau of Mineral Resources. The faults shown with stronger line weight are the boundary faults discussed in the text. The horizontal scale of each profile is the same, vertical scales vary somewhat. The locations of the profiles can be obtained from Figure 10. Profile 14—modified after Denham & Brown (1976).

tinental crustal anatexis very near to the incipient continent-ocean boundary. These data also indicate that it is not valid to interpret the occurrence of volcanics on a marginal plateau slope as definitive evidence of non-continental crustal structure.

Allen & others (1978) included a discussion of the Scott Plateau in their description of the Browse Basin, and Stagg (1978) has made a specific study of it. Both suggested that the inferred absence of Permian to Jurassic strata on the plateau could be attributed to its much greater elevation during that time. They also show a sedimentary trough seaward of the Scott Reef Horst, beneath the shallow physiographic trough which separates the Scott Plateau from the shallow continental margin. We therefore see structural asymmetry similar to the Exmouth Plateau. The wavelength of basement structures appears to be similar in the Browse Basin and Scott Plateau.

Hinz & others (1978) have reported dredging results from the Scott Plateau. Five of these were

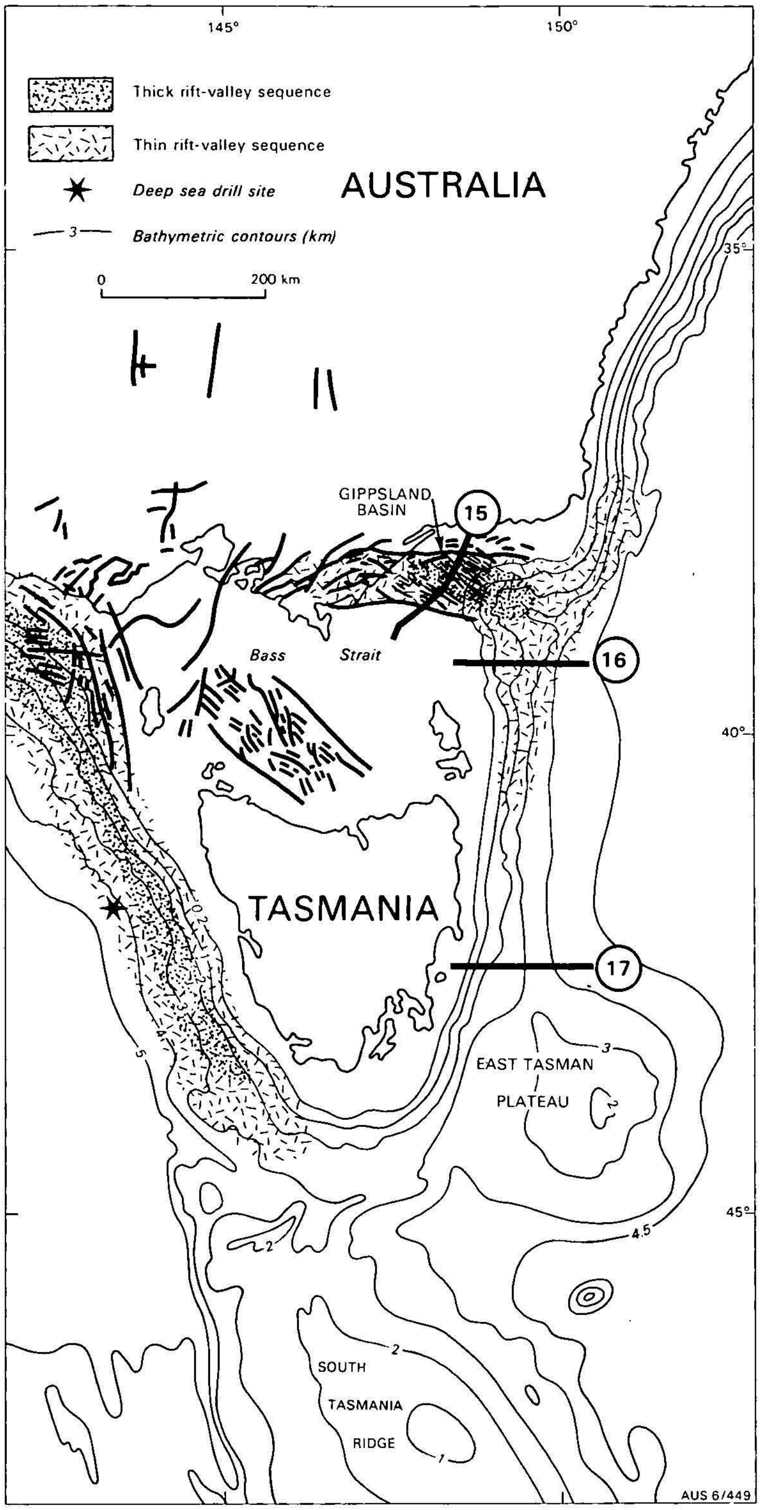

Figure 12. Bathymetry with principal structural and depositional patterns on the southeastern part of the continental margin of Australia. Bathymetric contours are taken from Hayes & Conolly (1972). Structural patterns are from Denham & Brown (1976) and Hocking (1976).

obtained from regions where Stagg (1978) had proposed that prebreakup sedimentary sequences crop out. At these sites basalts were obtained and no prebreakup sedimentary rocks were encountered. Reflection profiles indicate a thick stratified sequence is present in the area of dredging. The reflecting sequences probably comprise volcanogenic sediments and basalt flows which were deposited rapidly, shortly after the initiation of spreading.

Symonds & Cameron (1977) briefly described the Wallaby Plateau and the prominent trough which separates it from the continental margin. Despite poor stratigraphic control, it is proposed that there was relative elevation of the plateau region during deposi-

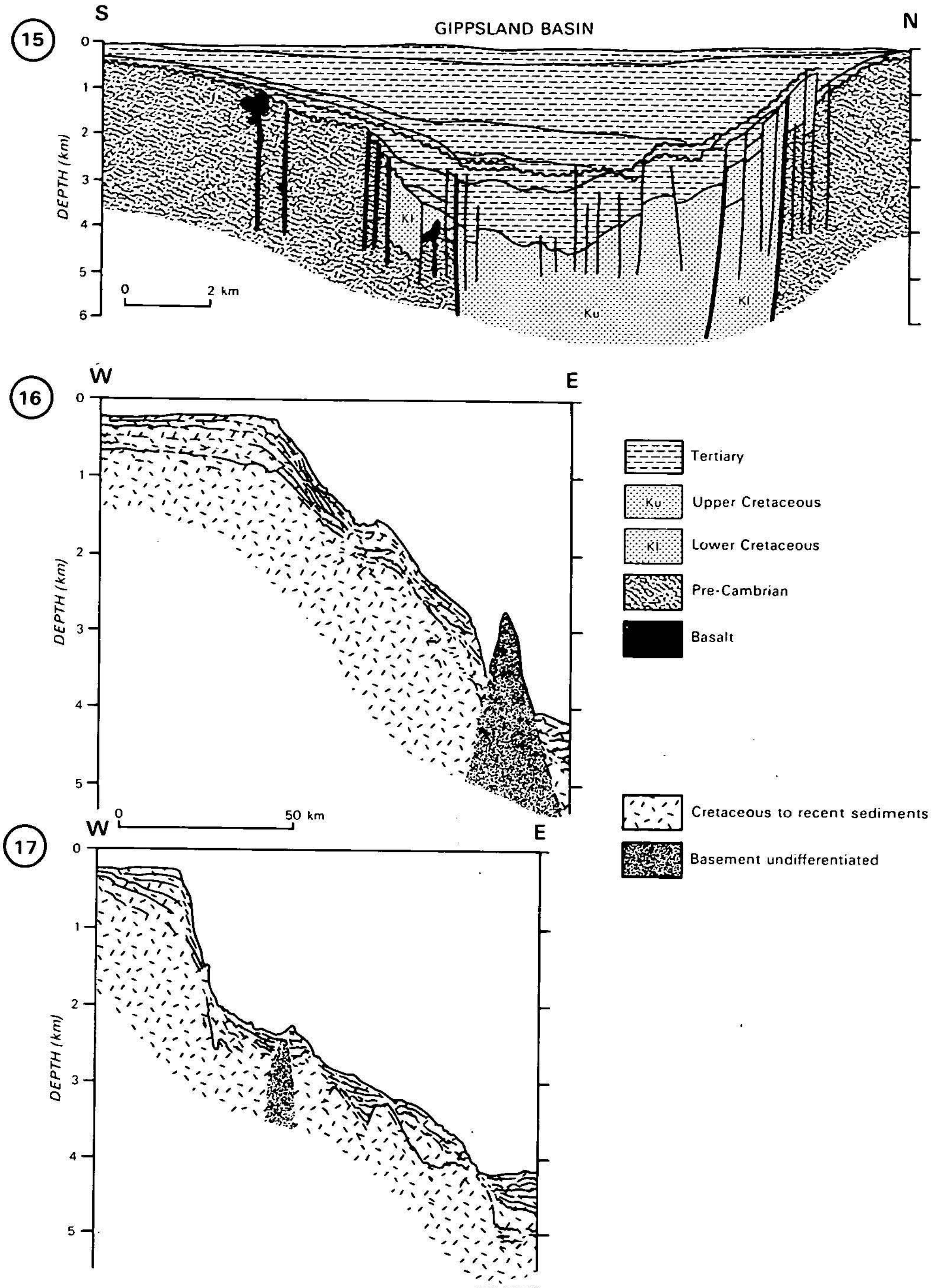

Figure 13. Structure profiles of the Gippsland Basin and east coast of Tasmania.
Profile 15 is taken directly from Hocking (1976) and is unchanged except for the emphasis on the boundary faults mentioned in text. Profiles 16 and 17 are line drawing interpretations of seismic reflection profiles obtained by the Bureau of Mineral Resources. The original drawings were made by Jongsma (unpublished data, 1978). The horizontal and vertical scales of profiles 16 and 17 are identical, but quite different from that of profile 15.

tion in the trough through most of the Mesozoic. Von Stackelberg & others (1980) reported dredging only breakup-time volcanics from this plateau.

Both Exon & Willcox (1978) and Stagg (1978) have suggested that the Exmouth and Scott Plateaus did not subside substantially below sea level until well after seafloor spreading had begun in the adjacent ocean basins. Exon & Willcox interpreted a postbreakup 'deltaic' sequence across the Exmouth Plateau as indicating shallow-water conditions just after breakup along the plateau's northern margin, together with shallow-water unconformities and erosional channels in younger sections. They proposed that major subsidence was delayed until the Miocene. Stagg cited lack of sediment accumulation and erosional unconformities to make a similar proposal. No similar propositions have been made for the Wallaby Plateau.

As will be discussed later, there is ample evidence in the sediments themselves, here and elsewhere, of submarine erosion and reworking of sediments. Comparison between postbreakup sedimentary velocities beneath the shelf and plateau led Veevers & others

(1974) to conclude that subsidence commenced soon after breakup.

Southwest margin

The southwest Australian continental margin (Figs. 8 and 9) includes the Naturaliste Plateau. It is elongate in an east-west direction, roughly perpendicular to the onshore Perth Basin. Plateau structures also strike east-west (Markl, 1978; Jongsma & Petkovic, 1977), quite unlike the northern plateaus. The southern margin of the plateau is formed by a steep escarpment bordering the Eocene and younger Southern Ocean. The northern margin grades into the Early Cretaceous Perth Abyssal Plain (profile 9, Fig. 9).

Several authors have discussed the nature of basement rocks beneath the Naturaliste Plateau, questioning whether they are continental or oceanic. Deep-sea drilling at two sites on the plateau (Veevers, Heirtzler, & others, 1974; Hayes, Frakes, & others, 1975) terminated short of basement. The oldest sediments en-

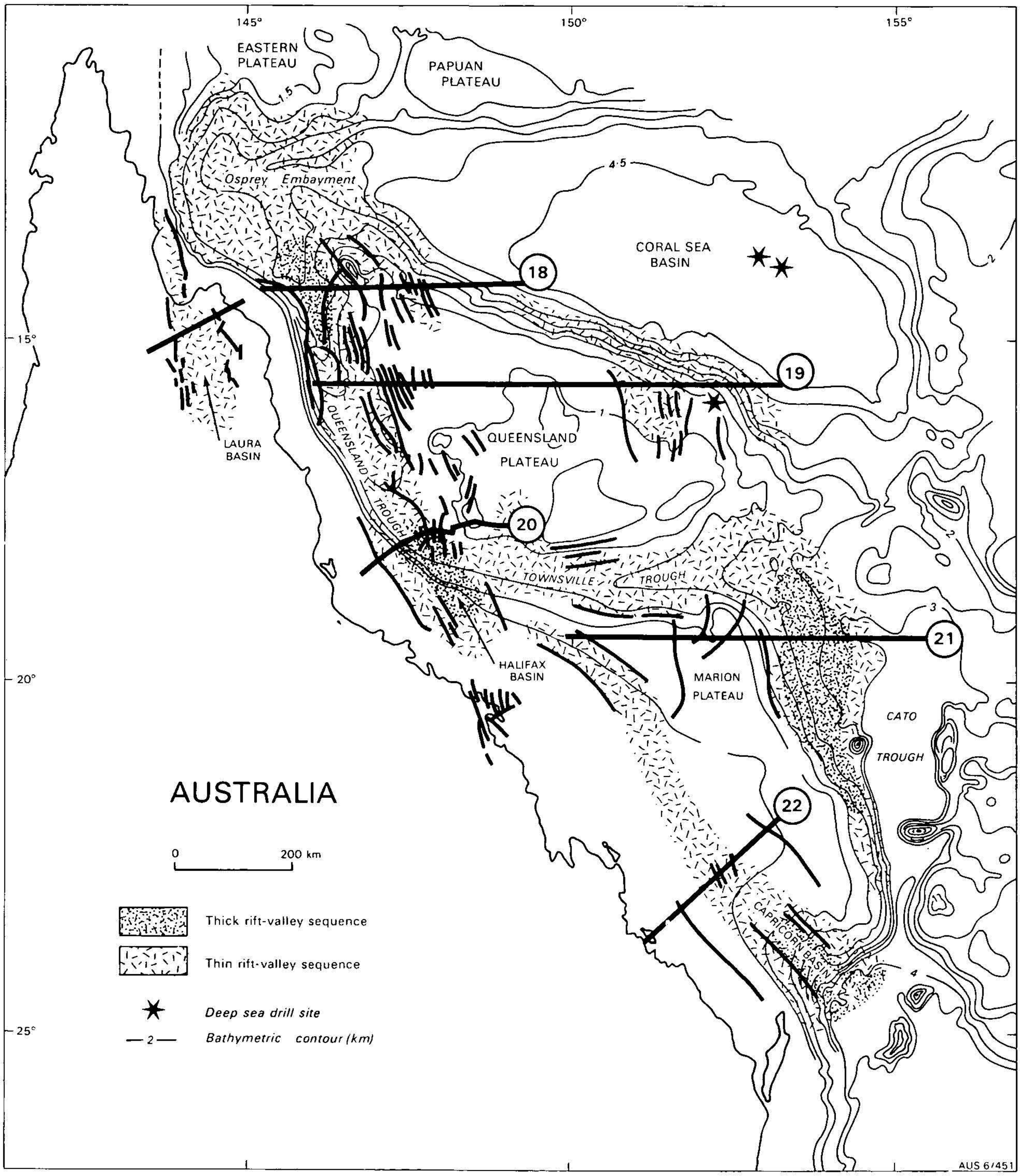

Figure 14. Bathymetry and principal structural and depositional patterns on the northeast margin of Australia. Bathymetric contours are from Mammerickx & others (1971), and Taylor (1977). Structure and depositional patterns are taken from the following sources: Ericson (1976), Day (1976), Mutter (1977), Mutter & Karner (1980), Pinchin & Hudspeth (1975), Rasidi & Smart (in press), Swarbrick (1976), Taylor (1977), Taylor & Falvey (1977). The heavy bars with circled numbers show the positions of structure profiles shown in Figure 15.

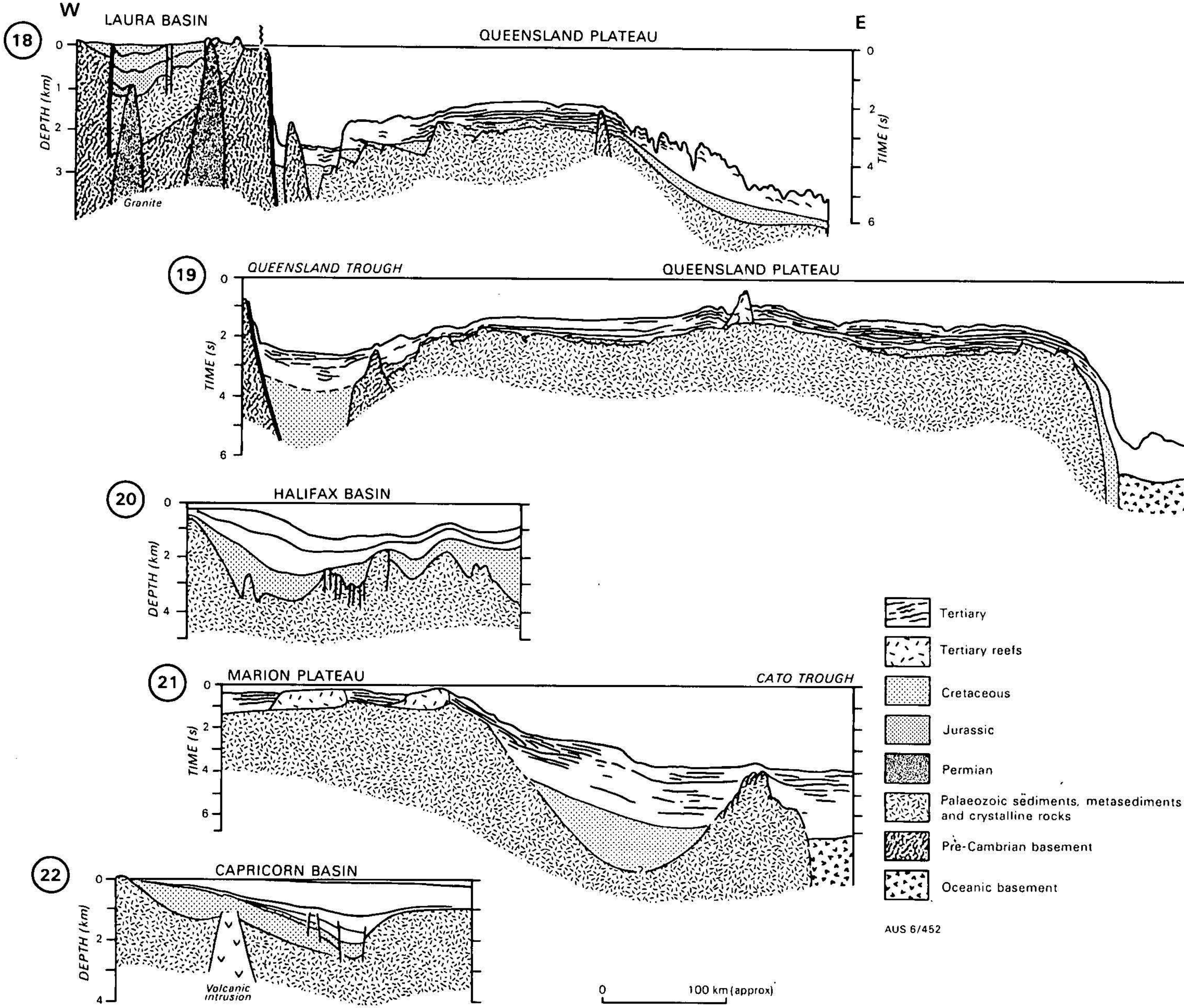

Figure 15. Structure profiles from the northeast margin of Australia.
Profile 18 combines a cross-section presented by Day (1976) for the Laura Basin with a line drawing interpretation of a seismic reflection profile, obtained by the Bureau of Mineral Resources, from Mutter (unpublished manuscript, 1974). Profile 19—line drawing interpretation of seismic reflection profile from the same source as the above. The stratigraphic divisions are made with reference to Deep Sea Drilling on the Queensland Plateau and exploration drilling in the Laura Basin. Profile 20—from Rasidi & Smart (in press). The profile is a combination of line drawing interpretations of seismic reflection data obtained by Gulf Oil and by the Bureau of Mineral Resources. The stratigraphic subdivisions are based on interpretation of sonobuoy velocity measurements. Profile 21—slightly modified after Mutter & Karner (1980). It is a line drawing interpretation of seismic reflection profile obtained by the Bureau of Mineral Resources. Profile 22—from Ericson (1976). It is based on exploration drilling data and seismic profiling. Horizontal scale on all profiles is the same, vertical scales vary somewhat. The locations of the profiles are shown in Figure 14.

countered were early Cretaceous (pre-Santonian) silts and sands underlying oceanic pelagic sediments. Jongsma & Petkovic (1977) traced reflection horizons (including one corresponding to the conglomerate layer) over the plateau for which they proposed ages based on correlation with the deep-sea drilling data. Their interpretation of one seismic line is shown in profile 9 (Fig. 9).

Coleman & others (in prep.) have studied the rocks from a dredge haul taken on the northern face of the Naturaliste Plateau, earlier reported by Heezen & Tharp (1973). The rocks were obtained from a basement exposure and are predominantly basic plagioclase tholeiitic basalt. A detrital mineral suite in minor proportion and with clear continental provenance was also present. The volcanic breccia at the base of DSDP 264, on the southern flank of the plateau, was described by Ford (1975) as being andesitic, but Nicolls & others (in press) and Coleman & others (in prep.) consider it more likely that Ford's analyses describe a weathered basaltic rock. The occurrence of volcanics, as discussed above, cannot be considered as definitive evidence of the non-continental origin of plateau basement. Indeed, there is no obvious structural discontinuity between Precambrian outcrop on the Leeuwin Block and basement underlying the plateau, and no compelling evidence for any dissimilarity.

The steep southern escarpment of the plateau resembles the major rift-valley boundary fault identified on Australia's southern continental margin (see discussion below). The east-west structural pattern of the plateau is also more like that on the southern margin than on the western margin, illustrating the Cretaceous rifting which dominates the southern margin. A speculative, thin Mesozoic rift-valley sequence is included in Figure 8. A wide continental rise is present south and east of the Naturaliste Plateau. South of the continental rise the Diamantina Zone forms a region of extraordinary rugged seafloor topography, which thus far has defied explanation. Talwani & others (1979) have described the eastward subsurface continuation of the Diamantina Zone, and suggested that it formed between seafloor spreading anomalies 19 and 22. They suggested that it represents an area of very anomalous crustal generation during early Australia-Antarctica seafloor spreading, and is thus of oceanic origin. Weissel & others (1979) have recently questioned the identification of lineations 19 to 22 in the Diamantina Zone and raised the possibility that they may instead be Late Cretaceous.

Basement rocks beneath the continental rise north of the Diamantina Zone appear to include a rift-valley volcanic sequence (Talwani & others, 1979), resulting in the very rugged appearance on reflection profiles (Fig. 9). Nicolls & others (in press) have described an upper mantle nodule dredged from the crest of an exposed basement peak immediately north of the Diamantina Zone. They suggest that it was derived from sub-oceanic upper mantle material and emplaced near the continent-ocean boundary at about the time of initiation of spreading. Boillot & others (1980) have described a somewhat similar rock from a structurally analogous location off the Iberian margin.

Southern margin

The southern continental margin of Australia extends from 110°E to 145°E, and commenced prebreakup development in the early Cretaceous. Seafloor spreading on the Southeast Indian Ocean Ridge began at anomaly 22 time (taken at 55 m.y. B.P.) along a latitudinal trend displaced by major transform offsets in the region of the present west coast of Tasmania. The Mesozoic structures and sedimentary basins are subparallel to the spreading ridge crest and to the broad coastal trend. Unlike the western margin with its complex system of large marginal plateaus, we consider the southern margin to be relatively simple.

Figures 10 and 11 show a major boundary fault or fault system forming the northern limit of the Mesozoic sedimentary basins. This is associated with a distinct magnetic anomaly minimum up to −500 nT. It also defines the northern limit of a magnetic quiet zone, which extends south to the first seafloor spreading anomaly (Weissel & Hayes, 1972; Talwani & others, 1978, 1979; König & Talwani, 1977). West of the Ceduna Plateau (profile 12, Fig. 11), low postbreakup sedimentation on the slope and continental rise have preserved a bathymetric expression of the Cretaceous rift immediately south of the steep continental slope (profile 11, Fig. 11; south of the Eyre Plateau). The structure and tectonic position of the Eyre Plateau are similar to the downfaulted half-grabens which occur on the western margin.

Rift subsidence beneath the continental rise south of the Eyre Plateau is as much as 8 km, based on sonobuoy measurements (Talwani & others, 1979) and structural/stratigraphic interpretations (Boeuf & Doust, 1975; Deighton & others, 1976; Mutter, 1978; Talwani & others, 1979). Profile 12 (Fig. 11) shows that the Ceduna Plateau consists of 9 km of prebreakup sediment overlain by about 1 km of postbreakup sediment. The continental rise is a small pelagic apron.

Boeuf & Doust (1975) and Mutter (1978) have discussed the continent-ocean crust boundary. They observed it on seismic reflection profiles as a simple junction between crustal types in the central and eastern parts of the margin. However, in the extreme west the continent-ocean boundary is not distinct. The basement beneath the continental rise appears to be rugged and volcanic in nature, and is almost indistinguishable from the adjacent oceanic crust.

Talwani & others (1978, 1979) and Mutter (1978) have discussed results of a sonobuoy survey of the continental margin west of 135°. They showed a variable velocity structure suggesting 'oceanic' and 'continental'-type structures lying close together. 'Oceanic'-type structures dominate in the extreme west of the margin. Sonobuoy measurements of crustal velocities also show that, west of 135°E, the change in physiographic profile results from the variable thickness of Mesozoic sediments. The Ceduna Plateau is a sedimentary prism loading a deep basement surface which is essentially the same as that across the Eyre Plateau (Fig. 11, profiles 11 and 12). Significant Cretaceous rift-valley sedimentation extends east through the Beachport Plateau and Otway Basin (Fig. 11, profiles 13 and 14).

At the eastern edge of the Otway Basin, fault trends change abruptly to south-southeast, parallel to the coast of western Tasmania. Griffiths (1971) suggested that the region contains a rift-valley system. The margin is characterised by a magnetic quiet zone (Weissel & Hayes, 1972), which, according to deep-sea drilling, is in part oceanic crust.

Southeast margin

Figure 12 shows the physiography and basic structural elements of the continental margin off Tasmania and the southeast margin of Australia. The more extensively studied region is the Bass Strait/Gippsland Basin. The remaining area is covered by reconnaissance seismic data.

The East Tasman Plateau and South Tasmania Ridge (or South Tasman Plateau) are probably subsided continental crust, according to the reconstructions of Weissel & others (1977) and Shaw (1978). They are separated from the narrow continental shelf and steep upper continental slope by troughs structurally similar to those bordering the major plateaus on the western margin of Australia.

The South Tasmania Ridge is elongate south-southeast and extends to 51°S. The northern surveyed portion shows flanking horst and graben structures with a central platform region. Sediments up to 3 km deep, but of unknown age, occur on the flanks. Deep-sea drilling on the South Tasmania Ridge has shown that postbreakup deposition on the plateau commenced in the Oligocene (Kennett, Houtz, & others, 1974).

The East Tasman Plateau has been poorly surveyed. Its physiography and structure are similar to the Walllaby Plateau (Figure 7, profile C). The troughs which bound the East Tasman Plateau and South Tasmania Ridge intersect at about 45°S 148°), where they are joined by a third deep trough (Fig. 12).

The Gippsland Basin strikes across the continental shelf, and Burke & Dewey (1973) suggested that it was the 'failed arm' of a three-branch rift system. The basin formed as a nearly symmetric graben (Figure 13, profile 15) in the Early Cretaceous.

North of the Gippsland Basin the continental margin is very narrow and steep. The continental slope is apparently almost devoid of prebreakup sediments. Postbreakup sedimentation forms a thin and narrow wedge at the shelf edge (Shaw, 1979). Jongsma & Mutter (1978) proposed that a rift sequence was developed, but that, because of asymmetric breakup, it separated from Australia and presently underlies the Dampier Ridge and Lord Howe Rise on the eastern side of the Tasman Sea.

Northeast margin

Figure 14 shows the structural trends and sediment distribution pattern of offshore Queensland. Figure 15 shows cross-sections of the margin, some rather schematic. The continental shelf is narrow north of about 18°S, and no sedimentary basins have been recognised on the shelf between 15°S and 18°S. North of 15°S, the shelf contains the offshore extensions of the Laura Basin and the southward extension of the Papuan Basin (not shown); the two may be interconnected (Oppel, 1970). The continental shelf south of 18°S broadens considerably and is underlain by the ill-defined Halifax Basin. This basin trends across the shelf edge near the Townsville Trough and lies partly beneath the deeper continental margin (Rasidi & Smart, in press). The Halifax Basin extends into the Queensland Trough, which itself extends northward into the Osprey Embayment (Mutter, 1977), a region about which very little is known.

The Capricorn Basin (Ericson, 1976) contains an Upper Cretaceous clastic sequence, considered by Taylor & Falvey (1977) to be rift-phase deposits. By correlating seismic velocities, Taylor & Falvey have interpreted the presence of a thick Cretaceous rift section in the Queensland and Townsville Troughs. For similar reasons, Rasidi & Smart (in press) suggest the existence of a Cretaceous section in the Halifax Basin. It is likely that the basins connecting the Halifax and Capricorn Basins also contain Cretaceous sediments. The Queensland Trough can be clearly recognised as a rift-graben, though the Halifax Basin appears to be a half-graben.

The tectonic position of the Jurassic-Cretaceous Laura Basin (Day, 1976) is less clear. The Maryborough Basin (not shown), lying adjacent to the western side of the Capricorn Basin, is also a Jurassic-Cretaceous depositional centre (Ellis, 1976), and its tectonic position is similarly unresolved.

The large marginal plateaus which dominate the continental margin have a thin Cainozoic sediment cover. Mesozoic strata are absent except on the outer flanks. On the outer Queensland Plateau, Taylor & Falvey (1977) proposed that a thick Mesozoic sequence was present, similar to that in the Queensland Trough. Their evidence came from sonobuoy refraction data and crustal models derived from gravity data. This interpretation has been supported recently by Cameron & others (1979).

The outer slope of the Marion Plateau exhibits a well-defined magnetic quiet zone, which Taylor & Falvey (1977) and Mutter & Karner (1980) have suggested is underlain by a thick Mesozoic rift sequence. The trend of the depocentre is nearly north-south, oblique to the depocentres on the shelf to the west.

Mutter & Karner (1980) have suggested that the unusual distribution of continental margin troughs and the geometry of the continent-ocean boundary may have resulted from the influence of up to three late Mesozoic three-branch rift systems (after Burke & Dewey, 1973), of which the Townsville and Queensland Troughs and Osprey Embayment represent 'failed arms'.

Tectonic and depositional correlations on Australia's rifted margins

We have outlined the basic physiography and structure of the five major segments of Australia's rifted continental margin, each corresponding to an ocean basin of a different age. They dissect a variety of existing geological structures on the continent, such that the margin basins are elongate at various angles to these structures. Formation of each margin was protracted, and, although the margin basins exhibit individual characteristics, similarities of structure and stratigraphy are clearly present. Characteristic depositional patterns are shown in the time-stratigraphic diagrams in Figures 16 and 17. These figures were prepared from published time-stratigraphic sections by the following authors: northwest—Powell (1976); southwest—Johnstone & others (1973); southeast—Partridge (1976), Shaw (1979); northeast—Taylor & Falvey (1977); southern —Falvey (1974), Boeuff & Doust (1975), Deighton (1977). Considerable data have also been drawn from the initial reports of the Deep Sea Drilling Project in each Basin. All data and summaries have been interpreted or reinterpreted in terms of the general format outlined by Falvey (1974). The general facies change beginning with breakup is evident on each cross-section. Thus, we have divided the tectonic and depositional history of the margin segments into prebreakup and postbreakup stages (Falvey, 1974). The time of breakup is defined as the onset of seafloor spreading. The term 'rifting' or 'initial rifting' is often, but incorrectly, used to describe this event. We find it preferable to restrict the use of the term 'rifting' to the intracon-

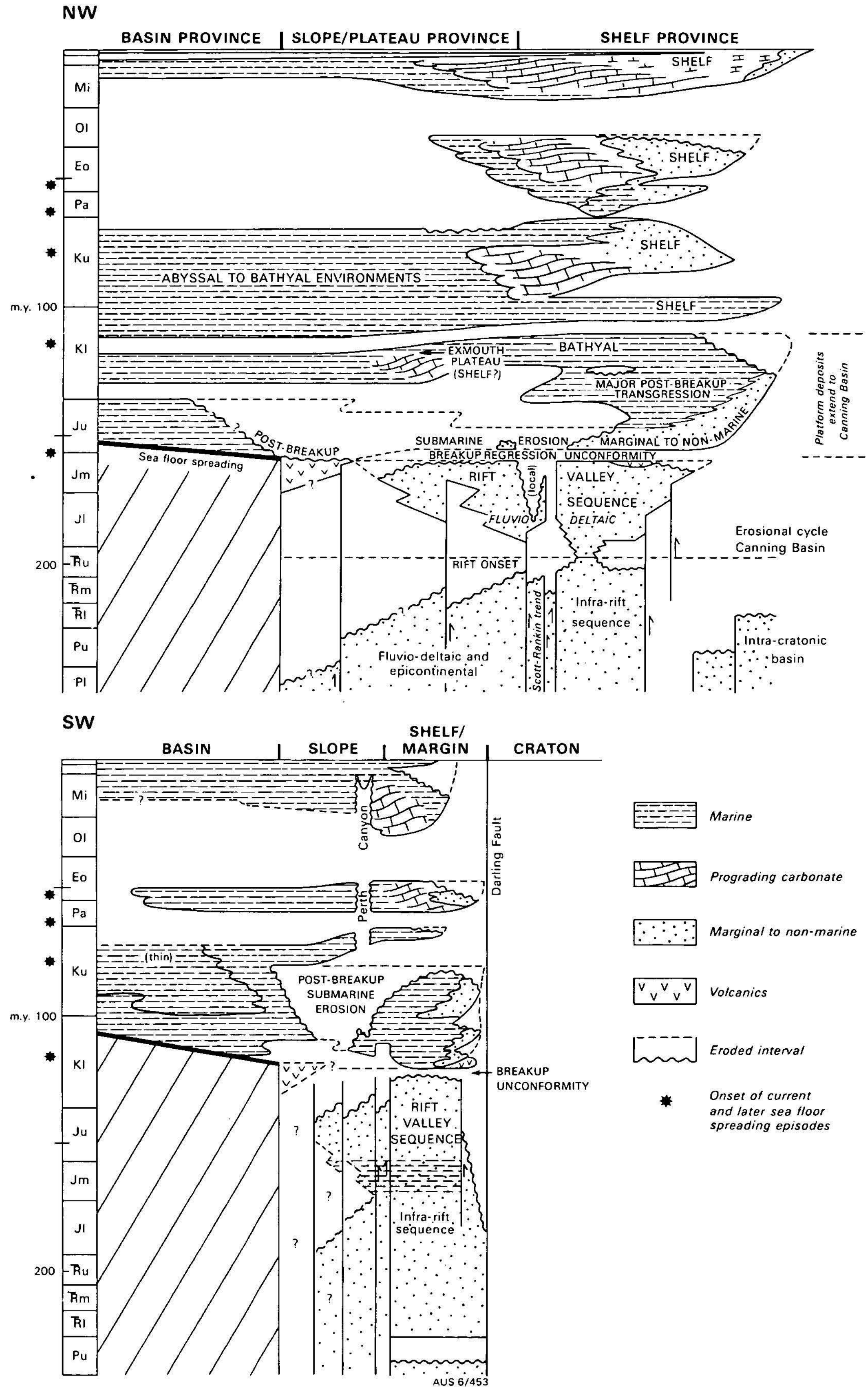

Figure 16. Generalised time-stratigraphic cross-sections of the Western Australian continental margin. The sections do not necessarily correspond to a specific profile, but are intended to represent the general range of environmental settings which may occur beneath each physiographic province. The sections have been derived by interpretation of data and compilations in Johnstone & others (1973), Veevers & Heirtzler (1974) and Powell (1976).

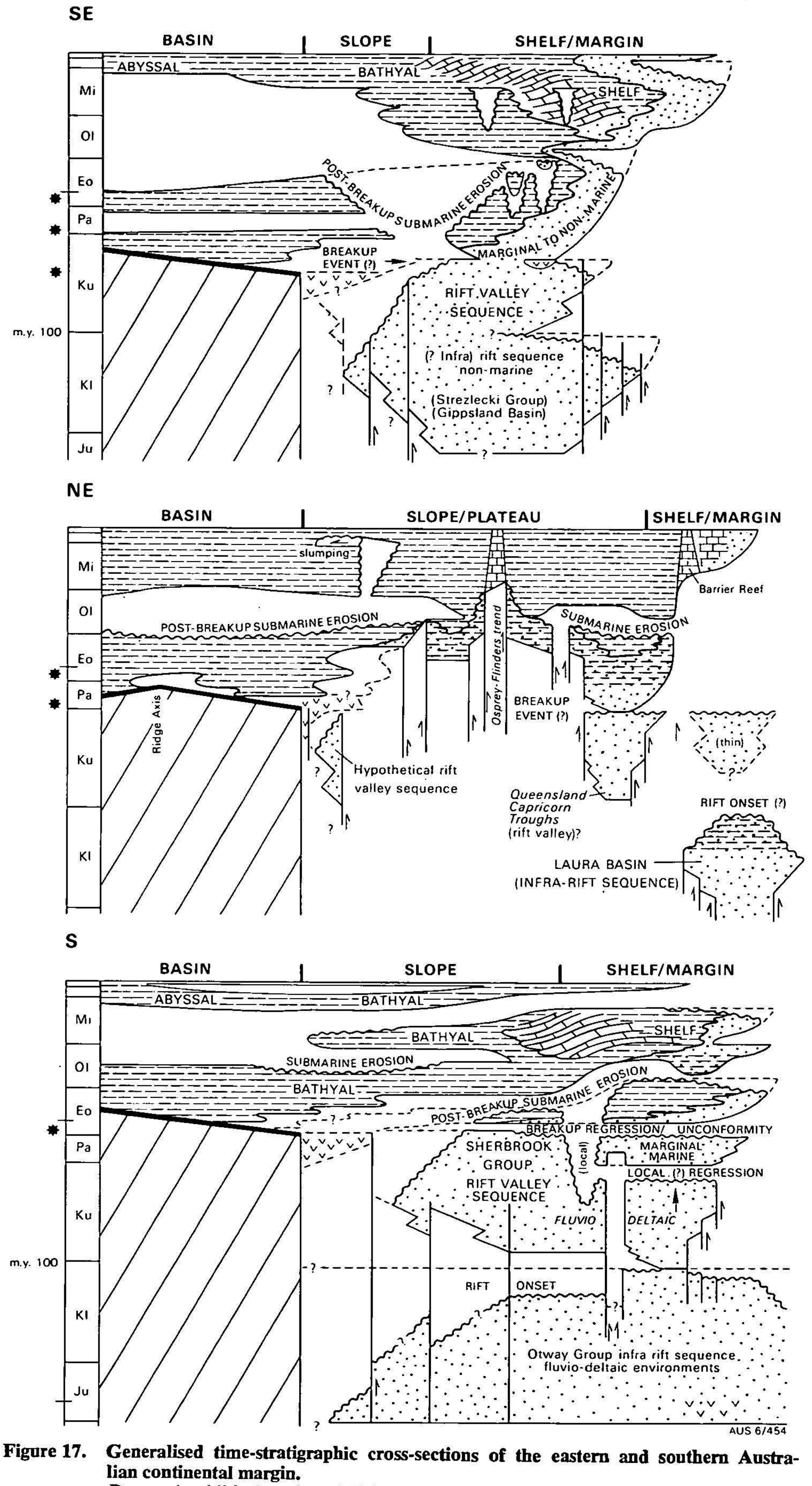

Figure 17. Generalised time-stratigraphic cross-sections of the eastern and southern Australian continental margin.

Data and published section of Falvey (1974), Boeuf & Doust (1975), Partridge (1976), Deighton (1977) and Taylor & Falvey (1977) have been re-interpreted or modified. The legend is the same as for Figure 16.

tinental fault-bounded subsidence which precedes the commencement of plate motion. This usage is consistent with the common description of major grabens such as the Rhinegraben, Gregory Rift, or Oslo Graben, as 'rifts', 'rift valleys', or 'rift-grabens'.

In all the margin segments described, the prebreakup stage consists of an extensive period of rift-valley tectonism and continental to restricted marine clastic deposition. Falvey (1974) regarded this stage as limited to about 50 m.y. He described 'prerift basins', which occupied broad downwarps in the same region as the rift valley, to account for known sedimentary sequences deep on some continental margins which predate this 50 m.y. period. Mutter (1978) argued that rift-valley and prerift basins were inseparable on Australia's southern margin. He proposed that tectonic activity began about 100 m.y. before breakup, with rapidly subsiding intracratonic-style basins forming along the incipient continental margin region. In Figures 16 and 17 we have adopted the term 'infrarift' sequence to describe these early-stage sediments in order to place this cycle of deposition unambiguously as part of the continental margin forming process, and not distinct from it. 'Infrarift'-stage sediments generally accumulated in broad elongate troughs, which show little evidence of faulted margins. Onset of the true rift stage involves major faulting, which causes depocentre shifts to narrower rift basins, and some uplift and erosion. The evolution from a simple narrow rift-graben to a broad complex graben, as envisaged by Dewey & Bird (1970), amongst others, does not seem applicable on the Australian margin. The rift stage is frequently marked by a rift onset unconformity, although the clear separation of 'infra' and 'true' rift-valley stages is not always possible.

Depositional rates (and hence basement subsidence rates) are frequently very high in the infra and rift-valley stages, often greatly exceeding postbreakup rates. Depositional environments are continental, paralic, fluvio-deltaic, and marginal marine. We know of no outer neritic or bathyal environments on a continental margin until well after breakup. Subsidence rates decrease towards the time of breakup; the general form of prebreakup subsidence shows an exponential decline in basement elevation, much like that of postbreakup subsidence described by Watts & Ryan (1976). Prebreakup and postbreakup cumulative subsidence from early Cretaceous to Present, for five southern margin exploration wells is shown in Fig. 18. Shell Platypus-1 most clearly shows exponentially declining prebreakup subsidence, followed by exponentially declining postbreakup subsidence.

In contrast to the early continental margin depositional sequences in the north Atlantic, North Sea, and other regions, Australia's prebreakup and postbreakup sedimentary sequences contain no evaporites or anhydrites. This is probably due to climate and the relative proximity of the Tethys Sea, and the Indian and Pacific oceans.

The basement structure that developed during the prebreakup rifting shows distinct similarities around the Australian margin. Landward boundary faults and fault systems are usually well defined and are clearly major crustal features. The seaward flank of the rift basins is generally a complexly faulted arch, horst block, or platform, rather than a single fault or fault system. Horst uplift appears more or less contemporaneous with deposition in the adjacent troughs. The strike of the horsts parallels that of the boundary faults (or fault system), and neither appears to be

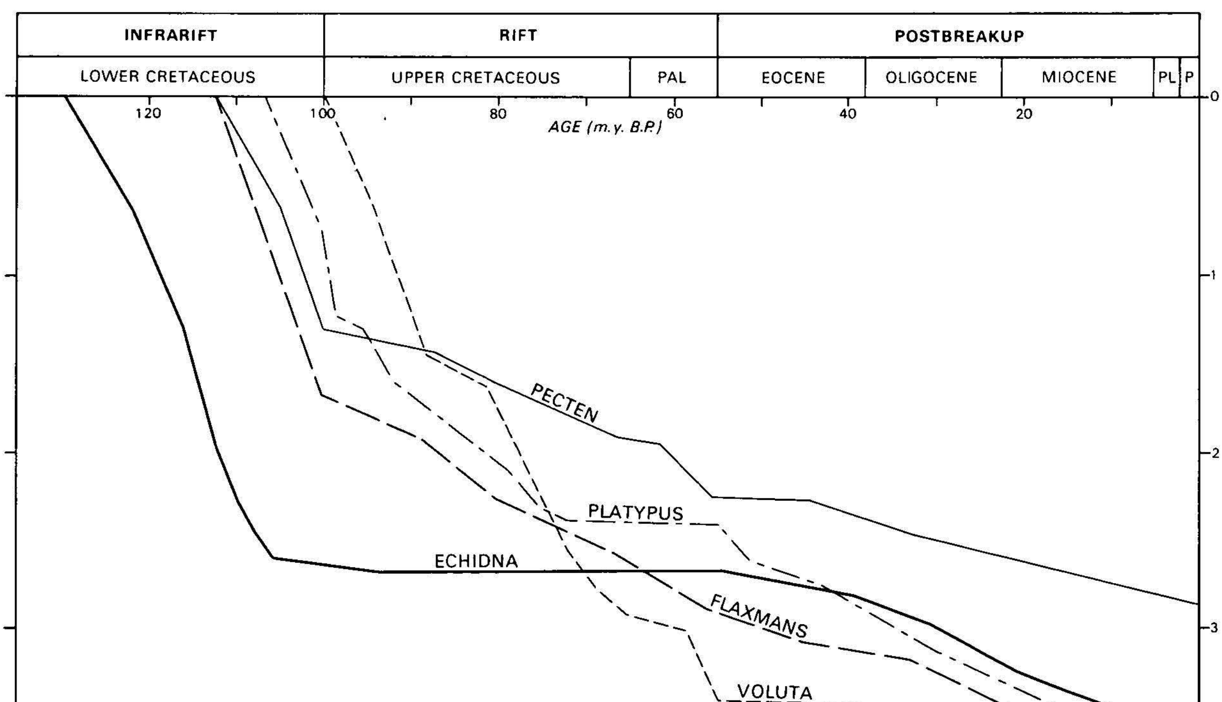

Figure 18. Cumulative subsidence or geohistory plots for 5 exploration wells from Australia's southern margin. Only well total depth (T.D.) is shown. Sedimentary formations were backstripped using a compaction correction algorithm described by Falvey & Middleton (1981). Sea-level and palaeowater depth are not shown.

dominantly influenced by observable structural grain in the continent. The strike of the horst-boundary fault system, while generally parallel to spreading trends, is sometimes noticeably oblique (southern margin west of 130°E).

At the end of the protracted rift stage in the development of a margin, the major structural fabric has been formed. Breakup entails separation of the oceanic lithosphere accretionary heat source from the margin, and thus the lithosphere beneath the margin begins to cool. This marks the onset of the postbreakup subsidence modelled by Sleep (1971) and Watts & Ryan (1976). On the southern margin, for instance, the Upper Cretaceous rift-valley stage, consisting of non-marine clastic sediments, is unconformably overlain by a marginal marine to marine section. Between 10 and 20 m.y. postbreakup, the trend towards a marine prograding carbonate shelf is dominant. A similar stratigraphic break is present on other parts of the margin (Figs. 16 and 17). Falvey (1974) described this hiatus as the 'breakup unconformity'. It has been recognised around much of Australia's margin, variously as an unconformity, disconformity, or paraconformity. Powell (1976) recognised a breakup unconformity or hiatus everywhere on the Northwest Shelf in the Callovian/Oxfordian. Locally, it may be represented only by a regressive to transgressive cycle or, more rarely, is not defined. This may be the case in the Gippsland Basin, but precise stratigraphic data are not available.

The complexities of postbreakup depositional environments were discussed in some detail by Deighton & others (1976) and Deighton (1977). They noted major postbreakup submarine erosion in the late Eocene to early Oligocene of the Otway Basin (Fig. 17). This was caused by ocean currents circulating at shallow and intermediate water depths in the young ocean basins. The timing of such an event would be governed by spreading rate, margin subsidence rate, and basin geometry and oceanography. Taylor & Falvey (1977) recognised a similar Oligocene submarine erosion event on the Queensland Plateau (Fig. 17).

We suggest that the effects of such ocean current circulation should exist on all new continental margins to a greater or lesser extent. We recognise it in the Tithonian/Neocomian of the northwest, and in the late Cretaceous of the southwest (Fig. 16). We also observe late Eocene/Oligocene and early Miocene events in the southeast (Fig. 17). It should be noted that this erosion may remove immediate postbreakup and prebreakup sediments, and widen and distort the definition of the breakup unconformity.

Continental shelf deposition following submarine erosion has been dominated by prograding carbonates. Locally, coral reefs mark Neogene margin subsidence in the northwest and northeast. Significant canyon development has been noted off Perth (Figs. 6, 8, 16), west of the Beachport Plateau (Fig. 10), and in the Gippsland Basin (Figs. 12, 13, 17). Major hiatuses in shelf deposition occur in the Palaeocene and Oligocene, depending on subsidence and sedimentation rates relative to sea level.

Mechanisms for passive margin subsidence

The preceding discussion has illustrated the remarkably consistent tectonic style of Australia's five major continental margin segments. This is particularly evident when the datum or origin of the structural or time stratigraphic cross section is fixed at the breakup time and/or the landward boundary fault. However, a comprehensive, quantitative theory of continental margin development does not yet exist. Rather, there are a number of theoretical mechanisms which have been put forward to explain all or some of the three principal subsidence phases. The diversity of opinion concerning the relation between vertical margin tectonics and plate divergence is in part due to the relative difficulty, on a global scale, of recognising those phases of subsidence. They are as follows (Figs. 16 and 17):

(1) Infrarift subsidence. The cycle begins with regional sedimentary basin subsidence, and deposition of mostly non-marine and non-volcanic sediments, preceded by erosion of basement or prerift sediments to various degrees. The basin is generally not fault bounded during deposition and straddles or closely flanks the incipient breakup. This phase is generally widespread, but is probably absent off the New South Wales margin.

(2) Rift subsidence. This phase consists of localised, narrow graben and half-graben subsidence in fault-bounded troughs, which most commonly flank the incipient breakup axis. Sediments are marginal to non-marine and non-volcanic, except, possibly, near to the incipient breakup. Grabens and bounding faults are en echelon and do not appear to be interconnected by transform or transcurrent cross faults. Rift subsidence is not usually the initial margin-forming event, nor is it generally preceded by erosion of infrarift sediments. Early deposition rates are rapid, exponentially declining towards breakup time. In graben depocentres, the transition from infrarift to rift-stage deposition is marked by essentially continuous sedimentation and sedimentation rate (Fig. 18).

(3) Postbreakup subsidence. Sedimentation is widespread from near-coastal regions and onlaps progressively younger oceanic crust. Depositional environments are marginal marine near shore and marine deep offshore, with the marine facies migrating gradually shorewards. Usually, sediments are reworked by ocean currents on the deeper parts of the margin. The postbreakup phase is also preceded by relatively little erosion of rift or prebreakup sediments, and is relatively undisturbed by faulting. Early deposition rates are rapid, exponentially declining with time, and strongly influenced by relative sea-level changes.

These general observations describe the development of Australia's passive margin basins independent of breakup time. When they are taken together with various regional geophysical data, it is possible to place some constraints on the various published subsidence models as they are applied to the three distinct subsidence phases. The following discussion draws on both general margin data, and the data specifically reviewed above.

Gravity loading models

The application of a sedimentary load to the top of the crust will produce subsidence of the crust until some form of local or regional isostatic equilibrium is achieved. Marine sediment deposited in a given water depth will accumulate to a maximum thickness approximately twice the initial water depth, assuming local Airy isostatic compensation. This model cannot solely be applied to prebreakup subsidence, since no significant palaeowater-depths (based on reliable benthonic microfauna) have ever been noted. In well-established postbreakup settings, on continental slopes and rises, such a mechanism is clearly relevant. This is easily demonstrated by the fact that isostatic gravity anomalies over continental margins are close to zero, and certainly do not reflect the positive load effect of sediments simply displacing water without compensating subsidence.

The long-term accumulation of shallow water to non-marine sediments, especially before breakup, implies an external driving mechanism for basement subsidence. However, it seems clear that isostatic loading multiplies the magnitude of this external mechanism. This isostatic multiplying factor is derived from mantle, water, sediment and air density contrasts, and is approximately 3, if the external mechanism is considered to operate in air, or 2, if in water. If sedimentation has a further feedback effect on the external driving mechanism, say, by altering the pressure and temperature conditions at a migrating metamorphic facies boundary (see below), then this isostatic multiplying factor is further increased.

The question as to whether isostasy operates as a local mechanism or in a regional (flexural loading) form governed by the strength of lithosphere is difficult to answer. The dominant form may vary with time during margin formation. Neither Sleep (1971) nor Watts & Ryan (1976) have been able to show that either model is preferable in describing postbreakup sediment accumulation. Most quantitative modelling is carried out on a local compensation assumption without apparently contradictory results. It is also difficult to decide, given the available data, whether flexural isostatic loading is of importance during the rift and infrarift stages of basin subsidence. Mutter (1978) has suggested that it plays a role in keeping the incipient breakup region free of sediment until breakup, but this is difficult to quantify. Such flexural arching is known to occur around seamounts as a response to the imposed load. In regard to rift-phase flexure, it is interesting to note that across one of Australia's simplest rifts, with little postbreakup sediment—the Capricorn Basin (Figures 14, 15)—there is no gravity expression of isostatic flexural effects (BMR, 1976). The real question, which remains unresolved, is to what extent isostatic flexural effect could be important as sedimentation responds to the various subsidence driving mechanisms (see below) affecting the three subsidence phases.

Lithosphere cooling

Sleep (1971) and Watts & Ryan (1976) have successfully argued that postbreakup sediment accumulation is a response to cooling and lithosphere contraction that follows the removal of the mid-ocean ridge heat source after breakup. These authors have modelled the driving mechanism of basin subsidence as an exponentially declining function of postbreakup time. This model explains fairly satisfactorily the postbreakup data presented above, given only that, at the time of breakup, most of the margin is not significantly above sea level. The common range of depositional environments either side of the breakup unconformity supports such an assumption.

McKenzie (1978) has suggested that sedimentary basins, analogous to rift and infrarift subsidence phases, also subside in response to cooling. He proposed that rapid pull-apart (continental stretching) leads to initial rapid subsidence (in the case of continental crust of normal thickness) and some hot asthenosphere upwelling, followed by cooling and exponentially declining basin subsidence and sedimentation. It is this last aspect of the model that might be presumed to correspond to observed rift and infrarift sedimentation patterns. However, the model fails on two critical points. No interrift transform faults are observed on any margin corresponding to plate pull-apart nor do rift orientations conform to pull-apart directions (initial transforms) or basement structures (See Figures 6, 8, 10, 12, 14); and the rapid subsidence phase, which should occur during the rapid pull-apart of continental crust, has not been identified, nor have any associated deeper-water marine sediments.

There is also a third arguable consequence of McKenzie's model. A number of Northwest Shelf rift sequences (Figure 7) are thick enough to have required very high levels of rift-phase heat flow, corresponding to large stretching factors. Yet, only about one percent of all the hundreds of exploration wells used in the various compilations of profiles shown in Figure 7 have encountered rift-phase volcanics or rift-volcanically derived sediment (e.g. Powell, 1976). On present evidence, we must discount this lithosphere cooling subsidence following crustal stretching.

Continental crustal stretching

Bott (1971) proposed that the crustal thinning necessary to drive continental margin subsidence isostatically was caused by stress-induced stretching or necking of the continental crust above the depth of isostatic compensation. McKenzie (1978) and Jarvis & McKenzie (1980) also have suggested that crustal attenuation may be caused by a stress-induced mechanical extraction of the crust. This model must be dissected in terms of prebreakup and postbreakup subsidence mechanisms.

Prebreakup subsidence is dominated by the rift-valley phase. From the data reviewed here, these rifts occur dominantly in en-echelon arrangement, flanking incipient continent-ocean boundaries (see Figures 2, 3, 4, 5). Whilst it is analytically possible to model the exponentially declining sediment accumulation of rift-phase deposition (Fig. 18) with gradual plate pull-apart, we note the following difficulties:

(a) Plate divergence (or strain) rate would need to decrease with time as the onset of breakup is approached, so as to match the commonly observed slowing of sedimentation rates just before breakup (and the breakup unconformity—Figures 16, 17, and 18).

(b) As noted above, transform faults connecting en-echelon rift segments are not known, and rift orientations do not conform everywhere to pull-apart directions.

(c) It is difficult to explain why a site of initial plate divergence (the rift valley) should so frequently be abandoned in favour of the site of incipient breakup (where rift-valley section, as opposed to half-grabens, is rare). In this context, it is worth noting that continental rifts (Rhinegraben, Ethiopian and Baikal rifts) are usually simple grabens.

(d) Stress differences above the depth of isostatic compensation that could induce rift formation by pull-apart (before breakup) are not easy to envisage operating over 50-100 m.y. before ocean crust formation.

(e) Necking by crustal extension involves the attenuation of continental rocks which provide a radiogenic component to observed heat flow. For a crustal heat production of 1.1 $\mu W\ m^{-3}$, observed heat flow should decrease by 5 $mW\ m^{-2}$ for every kilometre of water depth which is in isostatic response to such crustal thinning. Foucher & Sibuet (1979) have presented some high-quality data from the northern margin of the Bay of Biscay:

Oceanic crust	45 ± 9 mW m^{-2}
Slope 4 to 5 kms	41 ± 7 mW m^{-2}
Slope 3 to 4 kms	40 ± 3 mW m^{-2}
Slope 2 to 3 kms	42 ± 6 mW m^{-2} (ignores station Ch66-11)
Shelf to 2 kms	no data

The heat flow is essentially constant, where a variation of 10 mW m^{-2} might have been expected.

(f) The East African Rift is usually taken to be a tensional feature. However, the analogy noted frequently between the East African Rift and continental margin rifts may be overstated. While margin rift-phase sedimentation is dominantly non-volcanic (see above), the present Ethiopian-Gregory Rift is dominantly a volcanic province (McCall, 1967). The strong implication is that extension leads to volcanism; and that is not normally observed away from the incipient breakup axis (e.g. Powell, 1976).

(g) Notwithstanding this objection, it may even be reasonably argued that the East African Rift is not entirely dominated by extension between the Somali and African Plates. The apparent right-lateral transform offset between the Gregory-Rudolf Rift in Kenya and the Stefanie-Ethiopian Rift in southern Ethiopia is not represented on the ground by east-west transform faults (data in Davidson & others, 1976; also P. Purcell, pers. comm.). Further, the interpretation of the 7.1-7.5 km s^{-1} anomalous material underlying 6.3-6.4 km s^{-1} rocks beneath the Gregory Rift (Maguire & Long, 1976) as a crustal intrusion is not unique. It could also consist of the volatile-depleted, metamorphic equivalent of the overlying crustal rocks. Similar seismic refraction velocities are observed beneath most margins and prebreakup rifts (see, for example, Ewing & others, 1970; Taylor & Falvey, 1977). This will be discussed further below.

(h) The crustal refraction data available on Australia's southern margin (Talwani & others, 1979) also support the argument against large stretching. The thinnest crustal sections were encountered beneath the western margin. Following a finite stretching rate hypothesis, one would argue that the western rift should have been stretched by a greater amount than the eastern rift zone. However, the rift province on Australia's southern margin is almost constant in width, or slightly narrower, westward of about 135° E (Fig. 10), quite contrary to stretching predictions.

With these observations, we must downgrade finite rate crustal extension as the significant mechanism in prebreakup subsidence. However, at breakup and after breakup the situation may indeed change. Then, reasonable stress fields may exist, and strong listric faulting and some volcanism are known (e.g. Montadert & others, 1979) on the deepest parts of the continental margin.

The precise extent of the stretching-dominated component of subsidence landward of the continent-ocean boundary is rather difficult to specify. Some key seismic refraction profiles are unreversed (e.g. profile 10; Avedik & Howard, 1979). However, from the distribution of refraction solutions given by Talwani & others (1979) which contain, in part, clearly defined shallow mantle within the magnetic quiet zone south of Australia (Figure 11), it may be concluded that continental crustal extension contributes to subsidence on the outermost continental rise.

There is also evidence south of Australia for the progressive eastward disappearance of the earliest part of the seafloor-spreading anomaly sequence against the continental margin not accommodated at transform offsets. Talwani & others (1979) suggested that this might have resulted when opening by seafloor spreading did not initially extend across the whole rift valley such that opening occurred in the west contemporaneously with some stretching in the east. They believed that this would have taken place only in the first few million years after breakup.

Therefore, we conclude that stretching of the crust probably occurs close to the continent-ocean boundary at around the time of breakup, but plays a minor role, if any, in modifying crustal structure during the prebreakup, true rift-valley stage.

Deep crustal metamorphism

Falvey (1974) proposed that subsidence is driven by an increase in density, or contraction of the lower part of the continental crust. It was suggested that this density increase results from large-scale metamorphism of greenschist grade rocks to amphibolite grade or higher. The metamorphism is driven by heat from below the lithosphere moving into the lower crust. The reactions are pressure-temperature dependent and endothermic. The temperature of reaction is 400-500°C and is, therefore, below average rock melting point. Middleton (1978, 1980) has extended this hypothesis to the quantitative modelling of intracratonic basin subsidence.

This mechanism is not a significant factor in postbreakup subsidence. Clearly, the dominant components of the driving mechanism are cooling and contraction. However, the deep crustal metamorphism hypothesis may be applied to both phases of prebreakup subsidence. The strength of the argument lies in the model prediction that subsidence occurs during lithosphere heating. Middleton (1978, 1980) has shown that initiation of a sub-lithosphere thermal anomaly will produce exponentially declining sediment accumulation analogous to the cooling and contraction prediction. This is due to the similarity between the mathematical description of the heating and cooling cases. Falvey & Middleton (1981) have successfully modelled the cumulative subsidence pattern in Shell Platypus-1, using a single thermal anomaly at 60 km depth, of 300°C above ambient, and lasting from 140 to 53 m.y. B.P.

The effect of sediment accumulation during this process is to further increase the pressure and temperature conditions of rocks at the greenschist-amphibolite facies boundary. Thus, the grade of susceptible rocks will increase as a result of deposition, and add to the subsidence-driving mechanism. This effectively increases the isostatic multiplying factor for submarine sedimentation (depending on the greenschist-amphibolite density contrast, and the extent to which the geothermal gradient varies with time). Other important corollaries of this model are:

(a) Deep continental margin seismic refraction layers with velocity 7.0-7.6 km s^{-1} (Ewing & others, 1970; Talwani & others, 1979) may be interpreted as the deep crustal metamorphic equivalent of the overlying, lower-velocity crustal layer.

(b) This layer can be included in continental margin gravity models as a higher-density crustal layer (Taylor & Falvey, 1977), giving satisfactory computer-observed fits and relatively smaller isostatic load anomalies.

(c) Since the lower crust becomes denser, rift faults curve inwards at depth towards the centre of crustal contraction beneath the rift.

(d) Since no plate pull-apart is necessary to produce rift subsidence, there is no requirement for inter-rift transform fault segments.

(e) Since the metamorphism occurs at temperatures well below rock melting points, there need be no rift-phase volcanism of any significance away from the continent-ocean boundary region.

Falvey & Middleton (1981) have also successfully modelled the distribution of vitrinite reflectance values in Shell Platypus-1, using the thermal anomaly noted above. A stretching thermal model over-estimated the vitrinite reflectance at the base of the well by 10 percent.

The deep crustal metamorphism model has yet to be applied in cross-section to the various cycles of margin uplift and subsidence, but it does satisfy many of the sedimentary and structural constraints noted in this review.

Supracrustal erosion

A major difficulty with the deep crustal metamorphism subsidence model is—what happens when low metamorphic grade continental crust is not present, as might be the case beneath Archaean Shields? Clearly, the lithosphere heating cycle prior to breakup will cause significant uplift above sea level, since there is no metamorphic crustal contraction. Erosion of the land surface will cause crustal thinning, which will also bias the remaining continental crust to a higher average density. Postbreakup thermal contraction will produce marked submarine subsidence and a crustal structure in which the evidence for the thinning process is far from clear. The only evidence which we can cite, of prerift uplift and erosion, occurs where the Tasman Sea breakup has intersected the Permo-Triassic intracratonic Sydney Basin. In various parts of the basin bituminous coals occur at or near the surface, suggesting marked pre-Tertiary erosion.

It seems unlikely, however, that large-scale supracrustal erosion (leading to later subsidence) is a widespread occurrence. Indeed, the one place noted above, where there is direct evidence of erosion, is also notable for its marked lack of rift and infrarift sedimentary basins. In general, sedimentary basins have not been formed flanking incipient breakup or rift regions before the rift or infrarift stages of margin formation, as might be predicted.

Conclusions

The data reviewed in this paper lend weight to some of the published continental margin subsidence mechanisms. We believe it is most important to consider the three phases of margin subsidence as distinct, in terms of those mechanisms. Figure 19 is a generalised section across the continental margin, and illustrates where Australian data suggest these different mechanisms are probably dominant or at least operative.

It should be noted that, from observations of the structure of the basement and the distribution of sediments through time, it is not possible to make an unambiguous case in favour of any particular mechanism of margin formation. In particular, the

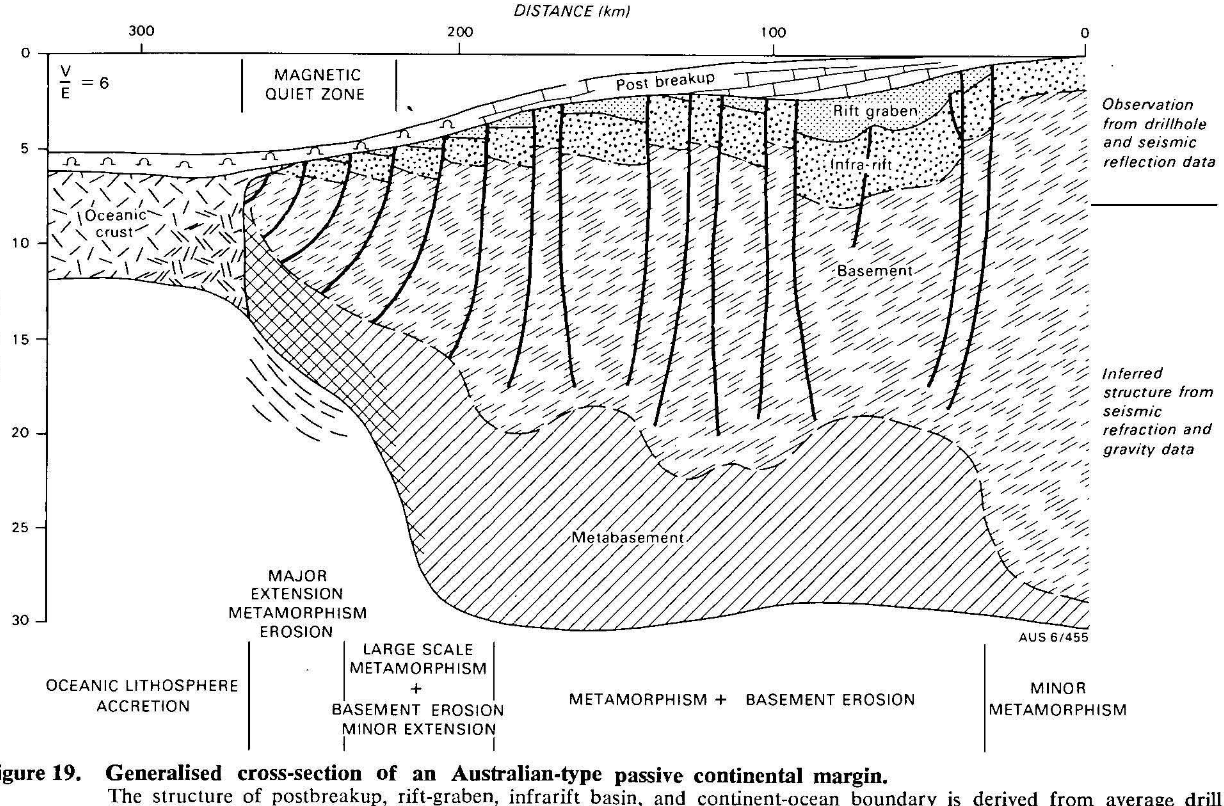

Figure 19. Generalised cross-section of an Australian-type passive continental margin.
The structure of postbreakup, rift-graben, infrarift basin, and continent-ocean boundary is derived from average drill hole and seismic reflection data summarised above. The structure of the metamorphic facies boundary and crust-mantle boundary is inferred from seismic refraction and gravity data. A magnetic quiet-zone plus a deep continental rise (of structural origin) may in part be underlain by distended and high-grade metamorphic continental crust, but should not be a zone of dyke injection. There should be little stretching landward of the rise. Cool mantle underplating may occur beneath the distended outer rise.

observations of exponentially declining sedimentation rates can be accommodated within several distinctly different models, although with varying degrees of difficulty. We find, however, that the style of development of rift-stage sedimentary basins is difficült to explain by a mechanism based on stress-driven continental stretching. This mechanism may be important in developing crustal structures at around the time of breakup, near the continent-ocean boundary.

The importance of obtaining precise deep crustal and upper mantle structural information is very clear. If, indeed, very thin continental crust overlies normal mantle beneath large parts of the margin, particularly in magnetic quiet zones, this would lend weight to a stretching hypothesis. Conversely, if a relatively normal thickness is present, but deep crustal layering is anomalous, support for a crustal metamorphic model would be provided. If a highly variable crustal structure was obtained, associated with the widespread presence of prebreakup volcanics, this could support the notion of a volcanically pervaded crust.

The thermal manifestations of mechanisms (i.e. palaeoheat-flow) which would lead to the development of these different types of crust are distinctly different. Palaeoheat-flow is fundamental to understanding levels of organic metamorphism and hydrocarbon maturation. Hence, if prediction of subsidence history and maturation depth are to become useful exploration tools on Australia's continental margins, a reliable passive-margin tectonic model is paramount.

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References

ABELE, C., KENLEY, P. R., HOLDGATE, G., & RIPPER, D., 1976—Otway Basin. *In* DOUGLAS, J. G., & FERGUSON, J. A. (editors), Geology of Victoria. *Geological Society of Australia Special Publication* 5, 198-229.

ALLEN, G. A., PEARCE, L. G. G., & GARDNER, W. E., 1978—A regional interpretation of the Browse Basin. *The APEA Journal*, 18(1), 23-33.

AVEDIK, F., & HOWARD, D., 1979—Preliminary results of a seismic refraction study in the Meriadzek-Trevelyan area, Bay of Biscay. *In* MONTADERT, L., ROBERTS, D. G., & others, *Initial Reports of the Deep Sea Drilling Project*, 48, 1015-23.

BMR, 1976—Gravity map of Australia, 1:5 000 000. *Bureau of Mineral Resources, Australia, Canberra.*

BOEUF, M. G., & DOUST, H., 1975—Structure and development of the southern margin of Australia. *The APEA Journal*, 15(1), 33-43.

BOILLOT, G., GRIMAUD, S., MAUFFRET, A., MOUGENOT, D., KORNPROBST, J., MERGOIL-DANIEL, J., & TORRENT, G., 1980—Ocean-continent boundary off the Iberian margin: A serpentine diapir west of the Galicia Bank. *Earth and Planetary Science Letters*, 48, 23-4.

BOTT, M. P. H., 1971—Evolution of young continental margins and formation of shelf basins. *Tectonophysics*, 11, 319-27.

BROWN, B. R., 1976—Bass Basin, some aspects of the petroleum geology. *In* LESLIE, R. B., EVANS, H. J., & KNIGHT, C. L. (editors), Economic Geology of Australia and Papua New Guinea, 3—Petroleum. *Australasian Institute of Mining and Metallurgy Monograph* 7, 67-82.

BURKE, K., & DEWEY, J. F., 1973—Plume-generated triple junctions; key indicators in applying plate tectonics to old rocks. *Journal of Geology*, 81, 406-33.

CAMERON, P. J., FRITSCH, J., HINZ, K., KARNER, G. D., MOAINA, R. B., ROESER, H. A., SCHLUTER, H. U., & BALFE, P. E., 1979—Papuan and Queensland Plateau margins, Coral Sea; geophysical results of Sonne Cruise SO-7. IUGG, XVII General Assembly, *Interunion Commission on Geodynamics, Abstracts and Timetable*, 5/6.

COONEY, P. M., EVANS, P. R., & EYLES, D., 1974—Southern Ocean and its margins. *In* VEEVERS, J. J. (editor), Deep Sea Drilling in Australasian Waters, *Challenger Symposium, Sydney*, 26-7.

CROSTELLA, A., 1976—Browse Basin. *In* LESLIE, R. B., EVANS, H. J., & KNIGHT, C. L. (editors), Economic Geology of Australia and Papua New Guinea, 3—Petroleum. *Australasian Institute of Mining and Metallurgy Monograph* 7, 194-9.

CROSTELLA, A., & CHANEY, M. A., 1978—The petroleum geology of the outer Dampier Sub-basin. *The APEA Journal*, 18(1), 13-22.

DAVIDSON, A., MOORE, J. M., DAVIES, J. C., SHIFERAW, A., TEFERRA, M., DEGEFFU, A., WOLDE RUFAEL, A., GELATTA, M., & HINTSA, N., 1976—Preliminary report on the geology and geochemistry of parts of Gemu Gofa, Kefa and Ilubabor Provinces, Ethiopia. *Ethiopian Ministry of Mines and Power, Omo River Project, Report* 2.

DAY, R. W., 1976—Laura Basin. *In* LESLIE, R. G., EVANS, H. J., & KNIGHT, C. L. (editors), Economic Geology of Australia and Papua New Guinea, 3—Petroleum. *Australasian Institute of Mining and Metallurgy Monograph* 7, 443-6.

DEIGHTON, I., 1977—The Paleogene stratigraphy of the southern Australian margin—an environmental approach to the general theory of rifted margin development. *M.Sc. Thesis, University of Sydney* (unpublished).

DEIGHTON, I., FALVEY, D. A., & TAYLOR, D. J., 1976—Depositional environments and geotectonic framework; southern Australia continental margin. *The APEA Journal*, 16(1), 25-36.

DENHAM, J. I., & BROWN, B. R., 1976—A new look at the Otway Basin. *The APEA Journal*, 16(1), 91-8.

DEWEY, J. F., & BIRD, J. M., 1970—Mountain belts and the new global tectonics. *Journal of Geophysical Research*, 75, 2625-47.

DOUGLAS, J. G., 1976—Gippsland Basin. *In* DOUGLAS, J. G., & FERGUSON, J. A. (editors), Geology of Victoria. *Geological Society of Australia Special Publication* 5, 158-63.

DUTOIT, A. L., 1937—Our wandering continents: An hypothesis of continental drifting. *Oliver & Boyd, London.*

ELLENOR, D. W., 1976—Otway Basin. *In* LESLIE, R. B., EVANS, H. J., & KNIGHT, C. L. (editors), Economic Geology of Australia and Papua New Guinea, 3—Petroleum. *Australasian Institute of Mining and Metallurgy Monograph* 7, 82-91.

ELLIS, P. L., 1976—Maryborough Basin. *In* LESLIE, R. B., EVANS, H. J., & KNIGHT, C. L. (editors), Economic Geology of Australia and Papua New Guinea, 3—Petroleum. *Australasian Institute of Mining and Metallurgy Monograph 7*, 446-50.

ERICSON, E. K., 1976—Capricorn Basin. *In* LESLIE, R. B., EVANS, H. J., & KNIGHT, C. L. (editors), Economic Geology of Australia and Papua New Guinea, 3—Petroleum. *Australasian Institute of Mining and Metallurgy Monograph* 7, 464-73.

EWING, M., HAWKINS, L. V., & LUDWIG, W. J., 1970—Crustal structure of the Coral Sea. *Journal of Geophysical Research*, 75(11), 1953-62.

EXON, N. F., & WILLCOX, J. B., 1978—Geology and petroleum potential of Exmouth Plateau area off Western Australia. *American Association of Petroleum Geologists Bulletin*, 62(1), 40-72.

EXON, N. F., WILLCOX, J. B., & PETKOVIC, P., 1975—A preliminary report on the regional geology of the Exmouth Plateau. *Bureau of Mineral Resources, Australia. Record* 1975/158 (unpublished).

FALVEY, D. A., 1972a—Sea-floor spreading in the Wharton basin (northeast Indian Ocean) and the breakup of eastern Gondwanaland. *The APEA Journal*, 12(2), 86-8.

FALVEY, D. A., 1972b—The nature and origin of marginal plateaux and adjacent ocean basins off Northern Australia. *Ph.D. Thesis, University of New South Wales* (unpublished).

FALVEY, D. A., 1974—The development of continental margins in plate tectonics theory. *The APEA Journal*, 14(1), 95-106.

FALVEY, D. A., & MIDDLETON, M. F., 1981—Passive continental margins: evidence for a prebreakup deep crustal metamorphic subsidence mechanism. *Oceanologica Acta*, 4(2), in press.

FALVEY, D. A., & TALWANI, M., 1969—Gravity map and tectonic fabric of the Coral Sea. *Geological Society of America, Abstracts with Programs*, Part 7.

FALVEY, D. A., & VEEVERS, J. J., 1974—Physiography of the Exmouth and Scott Plateaux, Western Australia, and adjacent northeast Wharton Basin. *Marine Geology*, 17, 21-59.

FOUCHER, J. P., & SIBUET, J. C., 1979—Thermal regime of the northern Bay of Biscay continental margin in the vicinity of DSDP sites 400 to 402. *In* MONTADERT, L., ROBERTS, D. G., & others, *Initial Reports of the Deep Sea Drilling Project*, 48, 289-96.

FORD, A. B., 1975—Volcanic rocks of Naturaliste Plateau, eastern Indian Ocean, site 264, DSDP Leg 28. *In* HAYES, D. E., FRAKES, L. A., & others, *Initial Reports of the Deep Sea Drilling Project*, 28, 821-33.

FRASER, A. R., & TILBURY, L. A., 1979—Structure and stratigraphy of the Ceduna Terrace region, Great Australian Bight. *The APEA Journal*, 19(1), 53-65.

GRIFFITHS, J. R., 1971—Continental margin tectonics and the evolution of south east Australia. *The APEA Journal*, 11, 75-9.

HALSE, J. W., 1976—Beagle Sub-Basin. *In* LESLIE, R. B., EVANS, H. J., & KNIGHT, C. L. (editors), Economic Geology of Australia and Papua New Guinea, 3—Petroleum. *Australasian Institute of Mining and Metallurgy Monograph* 7, 188-93.

HAYES, D. E., & CONOLLY, J. F., 1972—Morphology of the Southeast Indian Ocean. *In* HAYES, D. E. (editor), Antarctic Oceanology II: The Australian-New Zealand Sector. *American Geophysical Union, Antarctic Research Series*, 19, 125-45.

HAYES, D. E., FRAKES, L. A., & others, 1975—*Initial Reports of the Deep Sea Drilling Project*, 28.

HAYES, E. D., & RINGIS, J., 1973—Seafloor spreading in the Tasman Sea. *Nature*, 243, 454-8.

HEEZEN, B. C., 1960—The rift in the ocean floor. *Scientific American*, 203, 98-110.

HEEZEN, B. C., & THARP, M., 1973—USNS Eltanin Cruise 55. *Antarctic Journal of the United States*, 8, 137-40.

HEIRTZLER, J. R., CAMERON, P., COOK, P. J., POWELL, T., ROESER, H. A., SUKARADI, S., & VEEVERS, J. J., 1978—The Argo Abyssal Plain. *Earth and Planetary Science Letters*, 41, 21-31.

HINZ, K., BEIRSDORF, H., EXON, N. F., ROESER, H. A., STAGG, H. M. N., & VON STACKELBERG, U., 1978—Geoscientific investigations from the Scott Plateau off northwest Australia to the Java Trench. *BMR Journal of Australian Geology & Geophysics*, 3, 319-40.

HOCKING, J. B., 1976—Gippsland Basin, Stratigraphy. *In* DOUGLAS, J. G., & FERGUSON, J. A. (editors), Geology of Victoria. *Geological Society of Australia Special Publication* 5, 256-8.

HOGAN, A. P., & JACOBSEN, E. P., 1975—Geophysical results from the northwest continental margin of Australia. *Bureau of Mineral Resources, Australia, Record* 1975/101 (unpublished).

JARVIS, G. T., & MCKENZIE, D. P., 1980—Sedimentary basin formation with finite extension rates. *Earth and Planetary Science Letters*, 48, 42-52.

JOHNSTONE, M. H., LOWRY, D. C., & QUILTY, P. G., 1973—The geology of southwestern Australia—a review. *Journal of the Royal Society of Western Australia*, 56, 5-15.

JONES, D. K., 1976—Perth Basin. *In* LESLIE, R. B., EVANS, H. J., & KNIGHT, C. L. (editors), Economic Geology of Australia and Papua New Guinea, 3—Petroleum. *Australasian Institute of Mining and Metallurgy Monograph* 7, 109-26.

JONGSMA, D., & MUTTER, J. C., 1978—Non-axial breaching of a rift valley: evidence from the Lord Howe Rise and the southeastern Australian margin. *Earth and Planetary Science Letters*, 39, 226-34.

JONGSMA, D., & PETKOVIC, P., 1977—The structure of the Naturaliste Plateau. *The APEA Journal*, 17, 3-12.

KENLEY, P. R., 1976—Otway Basin, western part. *In* DOUGLAS, J. G., & FERGUSON, J. A. (editors), Geology of Victoria. *Geological Society of Australia Special Publication* 5, 147-52.

KENNETT, J. P., HOUTZ, R. E., & others, 1974—*Initial Reports of the Deep Sea Drilling Project*, 29.

KÖNIG, M., & TALWANI, M., 1977—A geophysical study of the southern continental margin of Australia, Great Australian Bight and western section. *Geological Society of America Bulletin*, 88, 1000-14.

LARSON, R. L., 1975—Late Jurassic sea-floor spreading in the eastern Indian Ocean. *Geology*, 3, 69-71.

LARSON, R. L., 1977—Early Cretaceous breakup of Gondwanaland off Western Australia. *Geology*, 5, 57-60.

LARSON, R. L., MUTTER, J. C., DIEBOLD, J. B., CARPENTER, G. B., & SYMONDS, P. A., 1979—Cuvier Basin. A product of ocean crust formation by early Cretaceous rifting of Western Australia. *Earth and Planetary Science Letters* 45, 105-14.

MCCALL, G. J. H., 1967—Geology of the Nakuru—Thomson's Falls—Lake Hannington area. *Geological Survey of Kenya, Report* 78.

MCKENZIE, D., 1978—Some remarks on the development of sedimentary basins. *Earth and Planetary Science Letters*, 40, 25-32.

MAGUIRE, P. K. H., & LONG, R. E., 1976—Structure on the western flank of the Gregory Rift (Kenya), Part I. The Crust. *Geophysical Journal of the Royal Astronomical Society*, 44, 661-75.

MAMMERICKX, J., CHASE, T. E., SMITH, S. M., & TAYLOR, I. J., 1971—Bathymetry of the Southwest Pacific. *Scripps Institution of Oceanography & Institute of Marine Resources*, Preliminary edition (unpublished).

MARKL, R. G., 1974—Evidence for the breakup of eastern Gondwanaland by the early Cretaceous. *Nature*, 251, 196-200.

MARKL, R. G., 1978—Basement morphology and rift geometry near the former junction of India, Australia and Antarctica. *Earth and Planetary Science Letters*, 39, 211-25.

MIDDLETON, M., 1978—Genesis of the Cooper and Eromanga Basins. *Ph.D. Thesis, University of Sydney* (unpublished).

MIDDLETON, M., 1980—A model of intracratonic basin formation, entailing deep crustal metamorphism. *Geophysical Journal of the Royal Astronomical Society*, 61, 1-14.

MONTADERT, L., ROBERTS, D. G., DECHARPAL, O., & GUENNOC, P., 1979—Rifting and subsidence of the northern continental margin of the Bay of Biscay. *In* MONTADERT, L., ROBERTS, D. G., & others, *Initial Reports of the Deep Sea Drilling Project*, 48, 1025-60.

MUTTER, J. C., 1977—The Queensland Plateau. *Bureau of Mineral Resources, Australia, Bulletin* 179.

MUTTER, J. C., 1978—An analysis of the Mesozoic-early Cenozoic rifting history of Australia and Antarctica. *M.Sc. Thesis, University of Sydney* (unpublished).

MUTTER, J. C., & KARNER, G., 1980—The continental margin off northeast Australia. *In* HENDERSON, R. A., & STEPHENSON, P. J. (editors), The Geology and Geophysics of North-eastern Australia. *Geological Society of Australia (Queensland Division), Brisbane*, 47-69.

NICOLLS, I. A., FERGUSON, J., JONES, H. A., MARKS, G. P., & MUTTER, J. C., in press—Ultramafic nodules from the ocean floor southwest of Australia. *Contributions to Mineralogy and Petrology*.

OPPEL, T. W., 1970—Exploration of the southwest flank of the Papuan Basin. *The APEA Journal*, 10, 62-9.

PARTRIDGE, A. D., 1976—The geological expression of eustacy in the Early Tertiary of the Gippsland Basin. *The APEA Journal* 16(1), 73.

PATTISON, R., WATKINS, G., & ABELE, C., 1976—Great Australian Bight Basin, South Australia. *In* LESLIE, R. B., EVANS, H. J., & KNIGHT, C. L. (editors), Economic Geology of Australia and Papua New Guinea, 3—Petroleum. *Australasian Institute of Mining and Metallurgy Monograph* 7, 98-104.

PETKOVIC, P., 1975—Origin of the Naturaliste Plateau. *Nature*, 253, 30-3.

PINCHIN, J., & HUDSPETH, J. W., 1975—The Queensland Trough: its petroleum potential based on some recent geophysical results. *The APEA Journal*, 15(1), 21-31.

POWELL, D. E., 1976—The geological evolution of the continental margin off northwest Australia. *The APEA Journal*, 16(1), 13-23.

RASIDI, J. S., & SMART, J. A., in press—Halifax Basin, Queensland. *Queensland Government Mining Journal*.

SHAW, R. D., 1978—Seafloor spreading in the Tasman Sea: A Lord Howe Rise—eastern Australia reconstruction. *Bulletin of the Australian Society of Exploration Geophysicists*, 9(3), 75-81.

SHAW, R. D., 1979—On the evolution of the Tasman Sea and adjacent continental margins. *Ph.D. Thesis, University of Sydney*. (unpublished).

SLEEP, N. H., 1971—Thermal effects of the formation of Atlantic continental margins by continental breakup. *Geophysical Journal of the Royal Astronomical Society*, 24, 325-50.

STAGG, H. M. J., 1978—The geology and evolution of the Scott Plateau. *The APEA Journal*, 18(1), 34-43.

SWARBRICK, C. F. J., 1976—Hodgkinson Basin. *In* LESLIE, R. B., EVANS, H. J., & KNIGHT, C. L. (editors), Economic Geology of Australia and Papua New Guinea, 3—Petroleum. *Australasian Institute of Mining and Metallurgy Monograph* 7, 384-6.

SYMONDS, P. A., & CAMERON, P. J., 1977—The structure and stratigraphy of the Carnarvon Terrace and Wallaby Plateau. *The APEA Journal*, 17(1), 30-41.

TALWANI, M., MUTTER, J., HOUTZ, R., & KONIG, M., 1978—The margin south of Australia—a continental margin paleorift. *In* RAMBERG, I. G., & NEUMANN, E. R. (editors), Tectonics and Geophysics of Continental Rifts. *Reidel Publishing Company, Holland*, 203-19.

TALWANI, M., MUTTER, J., HOUTZ, R., & KONIG, M., 1979—The crustal structure and evolution of the area underlying the magnetic quiet zone on the margin south of Australia. *In* WATKINS, J. S., MONTADVERT, L., & DICKERSON, P. W. (editors), Geological and Geophysical Investigations of Continental Margins. *American Association of Petroleum Geologists Memoir* 29, 151-75.

TAYLOR, L. W. H., 1977—The western Coral Sea; sedimentation and tectonics. *Ph.D. Thesis, University of Sydney* (unpublished).

TAYLOR, L. W. H., & FALVEY, D. A., 1977—Queensland Plateau and Coral Sea Basin: stratigraphy, structure and tectonics. *The APEA Journal*, 17(1), 13-29.

THOMAS, B. M., & SMITH, D. A., 1976—Carnarvon Basin. *In* LESLIE, R. B., EVANS, H. J., & KNIGHT, C. L. (editors), Economic Geology of Australia and Papua New Guinea, 3—Petroleum. *Australasian Institute of Mining and Metallurgy Monograph* 7, 126-55.

VEEVERS, J. J., 1974—Western margin of Australia. *In* BURK, C. A., & DRAKE, C. L. (editors), The Geology of Continental Margins. *Springer-Verlag, New York*, 605-16.

VEEVERS, J. J., 1981—Australian rifted margins. *In* SCRUTTON, R. A. (editor), Geodynamics of Seismically Inactive Continental Margins. *Final Report I.C.G. W.G.8, A.G.U., Washington*, in press.

VEEVERS, J. J., FALVEY, D. A., HAWKINS, L. V., & LUDWIG, W. J., 1974—Seismic reflection measurements off northwest Australian margin and adjacent deeps. *American Association of Petroleum Geologists Bulletin*, 58, 1731-50.

VEEVERS, J. J., & HEIRTZLER, J. R., 1974—Tectonic and palaeogeographic synthesis of leg 27. *In* VEEVERS, J. J., HEIRTZLER, J. R., & others, *Initial Reports of the Deep Sea Drilling Project*, 27, 1049-54.

VEEVERS, J. J., HEIRTZLER, J. R., & others, 1974—*Initial Reports of the Deep Sea Drilling Project*, 27.

VON STACKELBERG, U., EXON, N. F., von RAD, U., QUILTY, P., SHAFIK, S., BEIERSDORF, H., SEIBERTZ, E., & VEEVERS, J. J., 1980—Geology of the Exmouth and Wallaby Plateaus off northwest Australia: sampling of seismic sequences. *BMR Journal of Australian Geology & Geophysics*, 5, 113-40.

WARRIS, B. J., 1976—Canning Basin, off-shore. *In* LESLIE, R. B., EVANS, H. J., & KNIGHT, C. L. (editors), Economic Geology of Australia and Papua New Guinea, 3—Petroleum. *Australasian Institute of Mining & Metallurgy, Monograph* 7, 185-8.

WATTS, A. B., & RYAN, W. B. R., 1976—Flexure of the lithosphere at continental margin basins. *Tectonophysics*, 36, 25-44.

WEGENER, A., 1929—The Origin of Continents and Oceans. *Methuen, London*.

WEISSEL, J. K., & HAYES, D. E., 1972—Magnetic anomalies in the Southeast Indian Ocean. *In* HAYES, D. E. (editor), Antarctic Oceanology II. The Australian-New Zealand sector. *American Geophysical Union, Antarctic Research Series*, 19, 165-96.

WEISSEL, J. K., HAYES, D. E., & HERRON, E. M., 1977—Plate tectonic synthesis: the displacements between Australia, New Zealand and Antarctica since the late Cretaceous. *Marine Geology*, 25, 231-77.

WEISEL, J. K., & WATTS, A. B., 1979—Tectonic evolution of the Coral Sea Basin. *Journal of Geophysical Research*, 84, 4572-82.

WEISSEL, J. K., HAYES, D. E., MUTTER, J. C., & KONIG, M., 1979—New ideas on the early separation of Australia and Antarctica. *IUGG, XVII General Assembly, Inter-disciplinary Symposia Abstracts and Timetable*, 147.

WEISSEL, J. K., WATTS, A. B., & LATROUILLE, A., in press—Magnetic anomaly evidence for Late Paleocene to Late Eocene seafloor in the southern New Hebrides Basin. *Tectonophysics*.

WILLCOX, J. B., & EXON, N. F., 1976—The regional geology of the Exmouth Plateau. *The APEA Journal*, 16, 1-11.

WRIGHT, A. J., & WHEATLEY, T. J., 1979—Trapping mechanism and hydrocarbon potential of the Exmouth Plateau, Western Australia. *The APEA Journal*, 19, 19-29.

BMR Journal of Australian Geology & Geophysics, 6, 1981, 31-41

Geological factors in the development of sanitary landfill sites in the Australian Capital Territory

G. Jacobson & W. R. Evans

The application of geology to waste disposal problems is important with respect to site selection, design, and management. The initial site selection should take into account soils, and geomorphological and hydrological considerations, and site investigations should be based on detailed geological mapping. The seismic refraction technique is a powerful tool for the assessment of excavation conditions for feasibility study and design, and electrical resistivity techniques are potentially useful for monitoring changes in groundwater composition as a result of pollution. Increased awareness of the need for pollution control means that hydrogeological investigations and the establishment of groundwater monitoring systems are mandatory. These factors are illustrated by three case studies in the Australian Capital Territory.

Introduction

Waste disposal is a growing problem in modern society. Affluent industrial man generates 1-3 kg of solid waste per person per day, and as population increases and land available for disposal becomes scarcer, the problem will be compounded. The most common methods of solid waste disposal are open dumps, sanitary landfill, and incineration. In Australian cities sanitary landfill is generally considered the most suitable method at present. Sanitary landfill is an engineered method of disposal on land in which the solid waste is compacted into a fill in thin layers under carefully controlled conditions to avoid pollution (Fig. 1). The waste may be placed in an excavated trench, or it may be spread and compacted on the natural ground surface and used to fill in a depression or valley.

Canberra is estimated to generate solid waste at the rate of approximately 1.2 kg per person per day, and this is mainly domestic and trade waste (NCDC, 1975). The amount per person is probably less than in other Australian cities, because there is only a small amount of industrial waste generated in Canberra. Nevertheless, for a population of 210 000, it amounts to an annual total of 95 000 tonnes. Planning studies carried out for the National Capital Development Commission have confirmed that sanitary landfill is the most economical disposal method, although resource recovery techniques could be economical within a decade (Joint Committee on the Australian Capital Territory, 1976).

This paper summarises geological aspects of landfill-site development in the Australian Capital Territory. There are, at present, two sanitary landfill sites operating in the ACT. Over a period of several years BMR has carried out geological and geophysical investigations for the development of the West Belconnen landfill site, which serves North Canberra and Belconnen (Fig. 2). A second site, the Mugga South landfill site serves South Canberra, Woden, and Tuggeranong, and was selected following geological and geophysical investigation of a number of alternative sites. A third major landfill site, at Pialligo (Fig. 2), was closed in 1978, because it was considered a bird hazard to aircraft using the adjacent Canberra airport. It has been investigated as a potential pollution hazard.

Landfill sites and the geological environment

The environmental impact of landfill sites has led in recent years to increasing awareness of the geological factors in site selection, especially the potential for pollution. A number of reviews and case studies have been published in the United States (for example, Hagerty & Pavoni, 1973; Sendlein & Palmquist, 1975) and in Europe (for example, Heitfeld & others, 1976; UK Department of the Environment, 1978). In Australia, documented geological investigations of landfill sites include studies of the Hertha Road site in Perth (Bestow, 1977) and the Lucas Heights site in Sydney (Knight & others, 1978).

The main geological factors to be considered in landfill-site selection are the depth of soil and weathered rock, and the ease of excavation, the occurrence of groundwater and the potential for groundwater or surface water pollution by leachate developed in the landfill, and the availability of suitable cover materials. There are other, more general, planning constraints to the siting of landfills. These include com-

Figure 1. Compaction of solid waste at the Mugga South sanitary landfill site, ACT.

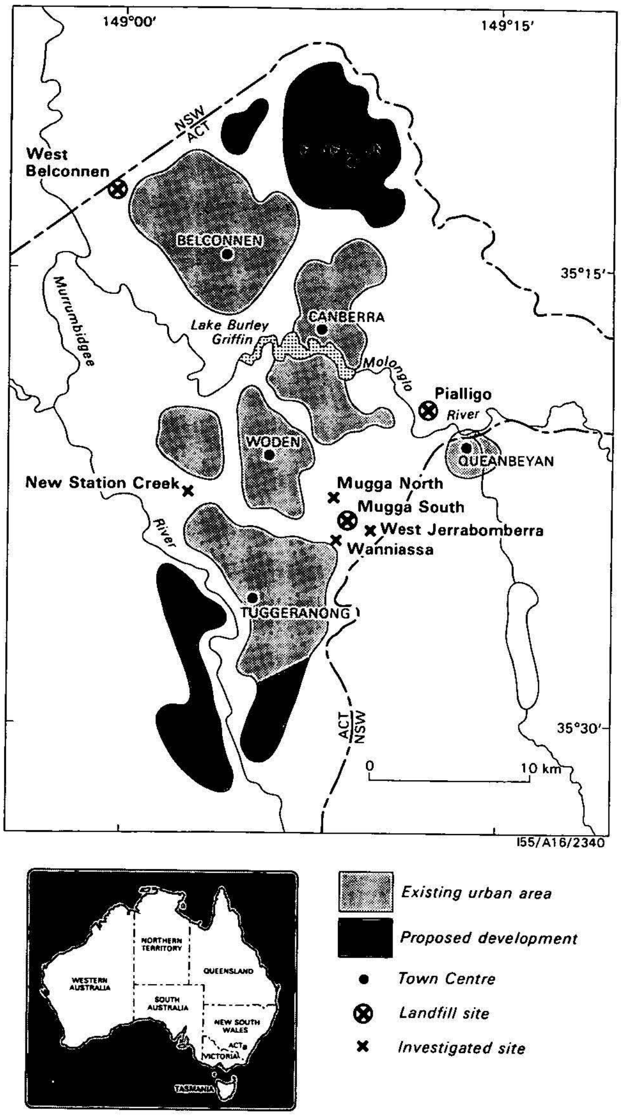

Figure 2. Location of sanitary landfill sites and sites investigated for such use.

patibility with existing or proposed land use of adjacent areas, location of the site relative to the sources of refuse, and possible reclamation and future land use of the site itself. Ideally, landfill sites should be selected at the time of initial urban planning, to enable a satisfactory resolution of geological as against other planning constraints.

Geological assessment is based initially on geological mapping and knowledge of the inter-relations of landscape, soils, and geology. Detailed evaluation of excavation characteristics can most readily be made by geological mapping, seismic surveys, and drilling. For any particular site, excavation depths might be variable, and the site has to be designed accordingly. The possibility of groundwater pollution can be assessed by a study of the groundwater regime, which involves drilling and testing of aquifers.

In the Australian Capital Territory, four main land surfaces have been identified by plains and benches at various elevations along the Molonglo River longitudinal profile (Van Dijk, 1959). They are characterised by deeper weathering than at other points in the landscape. Most urban development has been on the lowest of the four, the Yarralumla land surface, at an elevation of 565-580 m, extending up surrounding footslopes to about 620 m. The major landfill sites have been developed on the Yarralumla land surface or on the next highest land surface, the Yass-Canberra Tablelands, at an elevation of 670 m.

The depth of landfill is generally less than 10 m, except in the case of filling in old quarries. An understanding of soils and surficial geology is therefore necessary. In the ACT, alluvial, colluvial, and aeolian deposits of Cainozoic age overlie weathered Palaeozoic bedrock. The surficial geology is complex and related to cyclic soil-forming events, designated K_0 (youngest) to K_6 (oldest) by Van Dijk (1959). Bedrock geology is also complex, as the region comprises folded and faulted Ordovician to Devonian sedimentary and volcanic rocks, and granitic rocks (Strusz & Henderson, 1971). Depths of weathering vary according to the rock unit, and may be irregular.

Groundwater occurs in fractured rock aquifers and in thin alluvial and colluvial aquifers of the surficial sediments. Groundwater recharge is mainly by the infiltration of rainfall into fractured rock in hilly areas with skeletal soils, while discharge is mainly from the surficial sediments as springs and streams. The urban part of the ACT has an annual soil moisture deficit, with evapotranspiration greater than rainfall (Fig. 3). Recharge to the regional groundwater aquifers follows short intensive rainfall, and is seasonal, occurring mainly between May and November. However, for recharge to landfill sites, the time scale for consideration of moisture conditions should be shortened to days to account for individual rain events. Recharge to landfill sites can take place during the summer months, even though there is a moisture deficit when measured on a monthly basis.

Assessment based on these considerations has indicated that four geological units in the ACT are generally suitable for landfill. The various factors involved are shown in Table 1. Other geological units in the area have insufficient depth of soil, and are considered unsuitable.

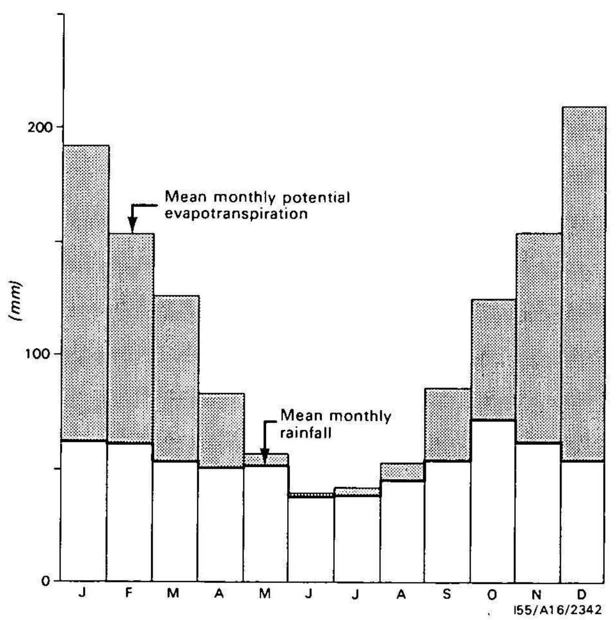

Figure 3. Mean monthly rainfall and potential evapotranspiration in the ACT.
The shaded area represents the excess of evapotranspiration over rainfall. Evapotranspiration is taken as 0.8 × pan evaporation.

Geological materials and approx. thickness	*Ease of excavation*	*Permeability of in situ material*	*Suitability of material for refuse cover*	*Position of highest water table*	*Permeability of material at base of excavation*	*Type of underlying aquifer*
Layered outwash fan deposits, up to 15 m. Includes clays, sands, gravels, aeolian sands; varying degree of induration	Variable excavation by mechanical means, depending on degree of induration	Variable, but mainly low; some sandy beds quite permeable	Permeability varies, but low permeability dominant; suitable	Generally above base of excavation; perched conditions common	Variable permeability; may require a seal	Fractured rock; aquifer may be confined and under pressure; minor water supply
Weathered volcanic rock	Readily excavated by mechanical means; difficult where irregularly weathered	Generally low	Permeability varies, but low permeability dominant; suitable	Generally below base of excavation	Generally low permeability; seal not usually required	Fractured volcanic rock; may be minor water supply
Deeply weathered sedimentary rocks to 15 m	Readily excavated by mechanical means	Generally low	Permeability low in fine-grained sediments such as shales and mudstone; suitable	Considerable seasonal fluctuation in low-permeability sediments; position of water table variable	Generally low permeability; seal not required	Fractured sedimentary rock; minor supply
Alluvium generally less than 6 m	Readily excavated by mechanical means	Variable from high to low	Permeability variable from low to very high: silty fraction may be suitable	Generally at or near base of alluvium	Permeability at base of excavation probably high; seal generally essential	Alluvial and or fractured rock; may be minor supply

Table 1. Suitability of geological materials in Canberra for landfill (adapted from Wilson, 1975).

The influence of geology on design— West Belconnen landfill site

The West Belconnen landfill site is about 50 ha in area, and is drained to the west by a tributary of the Murrumbidgee River. It is the only landfill site in Belconnen and North Canberra (Fig. 2), and serves a population of over 100 000. Planning and design of the site began in 1971, and operations, in 1976. Geology has been a considerable factor in design and development.

Initial geological studies of the site included augering and seismic surveys (Hill, 1972). These investigations showed that up to 4 m of residual clay overlies weathered Silurian sedimentary and volcanic rocks (Fig. 4). Groundwater occurs in a fractured rock aquifer and is confined by the overlying clay.

The aquifer was considered a potential pollution hazard and, as such, was a major planning constraint. To minimise the possibility of groundwater pollution, the designers, L. T. Frazer & Associates Pty Ltd, adopted a dry fill technique, whereby the solid waste is laid in trenches and sealed with a low-permeability clay layer. Surface water is directed around the landfill to minimise infiltration, and a clay layer is left in the base of the trenches to act as a low-permeability barrier between the solid waste and the underlying fractured rock aquifer. Leachate that drains from the solid waste and contaminated runoff are treated on site in storage ponds by anaerobic decomposition and oxidation.

Groundwater is regularly monitored to determine the effects of the landfill operation, and to assess future requirements for leachate disposal or treatment. To establish the groundwater monitoring program, 12 auger holes were drilled in 1976. Eight of these encountered groundwater, and were equipped with slotted plastic casing and surrounded by a concrete apron and protective fence to serve as monitoring bores.

Measurements of water level indicate that groundwater flow is generally to the southwest below the site. The potentiometric surface is generally 1-2 m below ground level on the slopes of the site, although seasonal fluctuations range up to 2.5 m in some bores.

Water quality is monitored every 6 months using Dissolved Organic Carbon (DOC) as a pollution index (Hughes & others, 1974). Background values of groundwater samples have been 1-3 mg/l DOC and for dam water, 8-20 mg/l DOC. The chemistry of the natural groundwater varies considerably with total

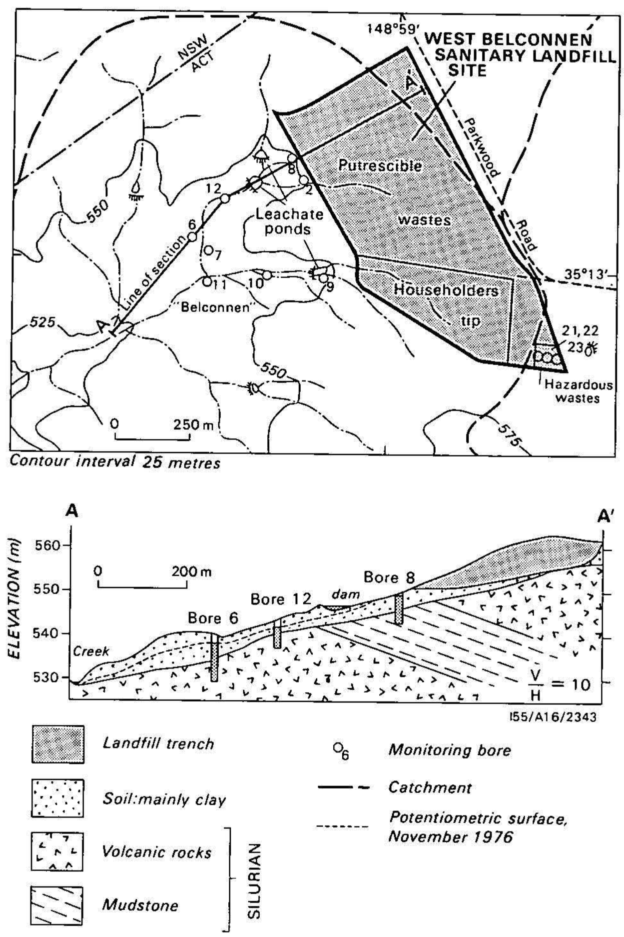

Figure 4. West Belconnen sanitary landfill site.

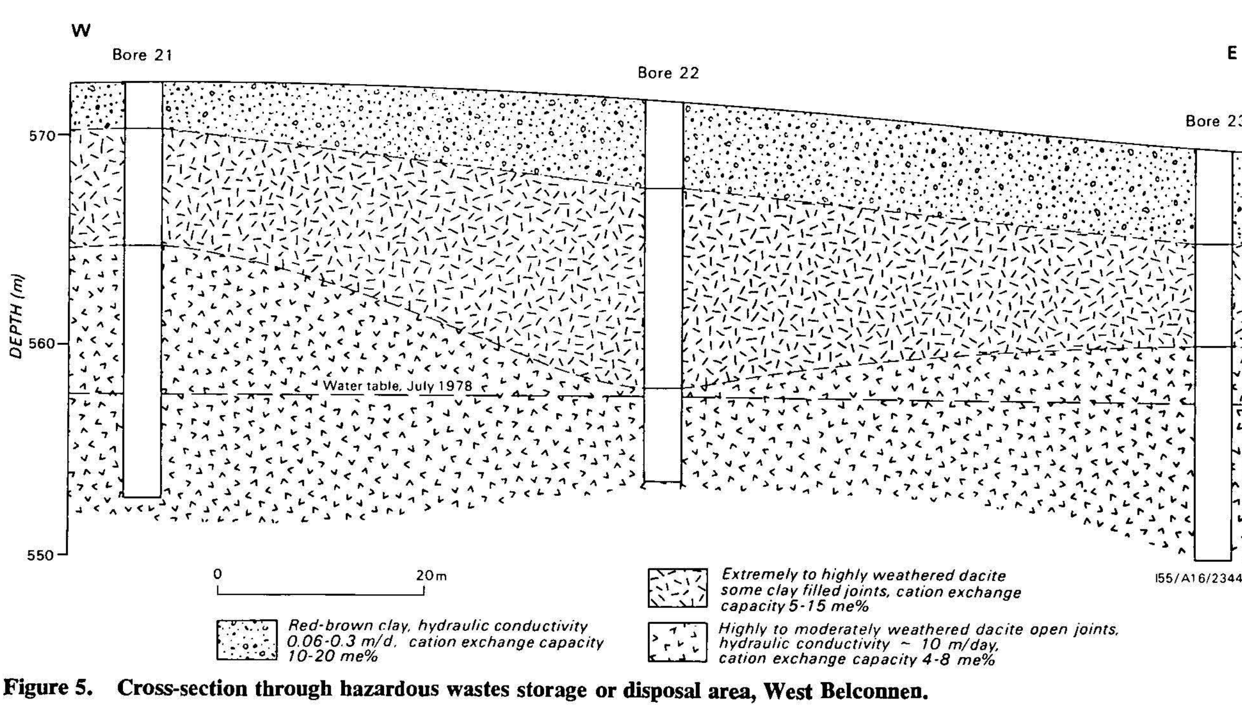

Figure 5. Cross-section through hazardous wastes storage or disposal area, West Belconnen.

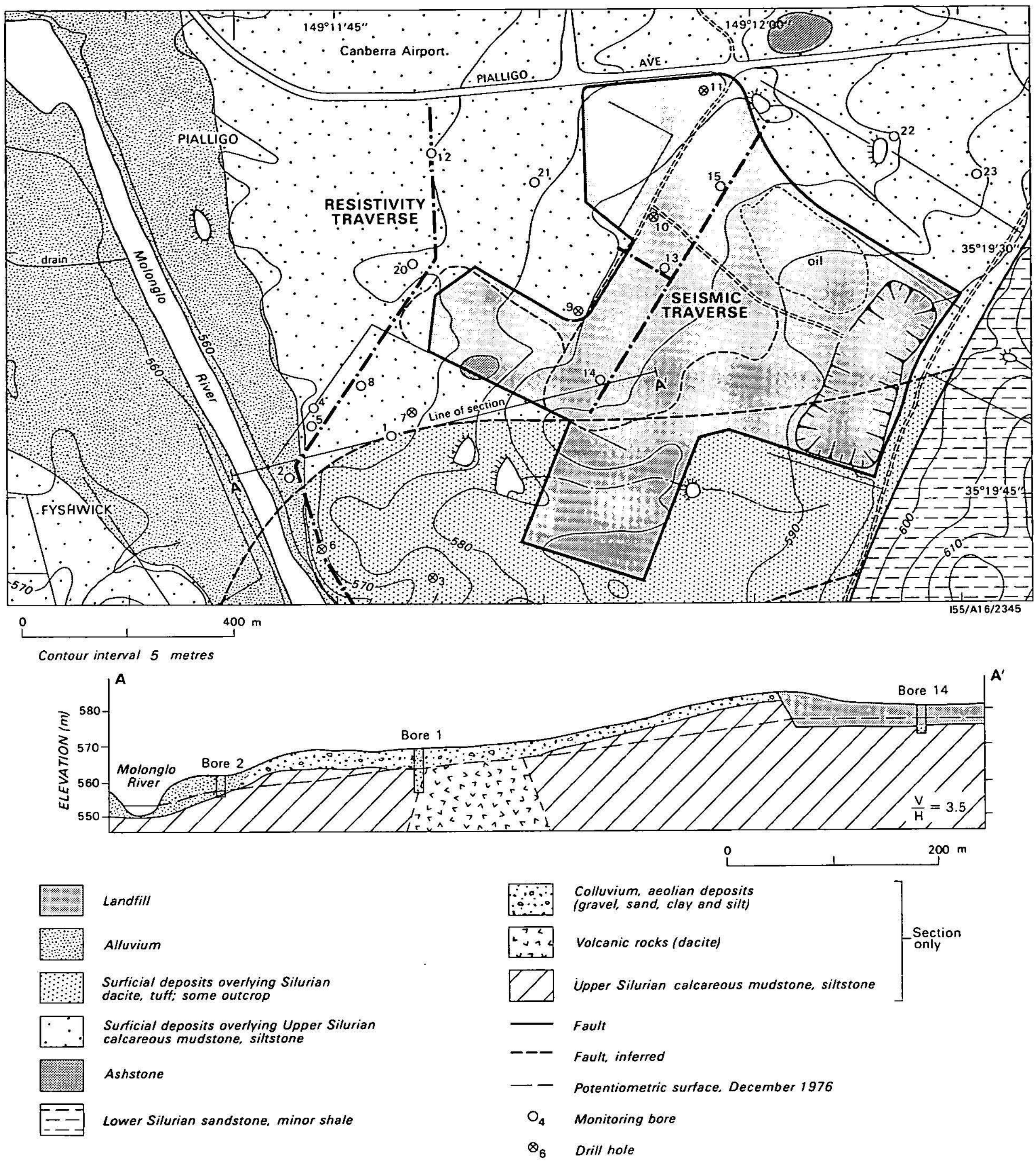

Figure 6. Pialligo landfill site.

dissolved solids ranging from 219 mg/l to 1334 mg/l over a small area. In 4 years of operation, no groundwater pollution has been detected.

Hazardous waste site

The southeast corner of the site has been designated as a possible storage or disposal area for hazardous wastes, and has been investigated in detail. A cross section of the storage area is shown in Figure 5. Several metres of clay overlies weathered and closely jointed porphyry. The fractured rock aquifer is unconfined at this location, and the water table is 11-14 m below the ground surface. The aquifer transmissivity was determined as 0.15 m^2/day by a series of pump tests. The frequency and openness of joints were measured in the drill core, and the rate of groundwater flow through fractures was determined at 0.13 m/day. From these data the travel time for pollutants passing through the rock to a discharge point one kilometre to the west could be estimated as 21 years.

Infiltration tests on the soils at the site indicate that about 14 days would be required for water to travel from the ground surface to the water table. A dye tracing experiment using rhodamine WT was also carried out, but the dye failed to appear in observation bores. The ion-exchange capacity of both soil and rock samples was analysed, and results indicate that the clayey soils have a greater capacity for ionic absorption of pollutants than has the weathered porphyry.

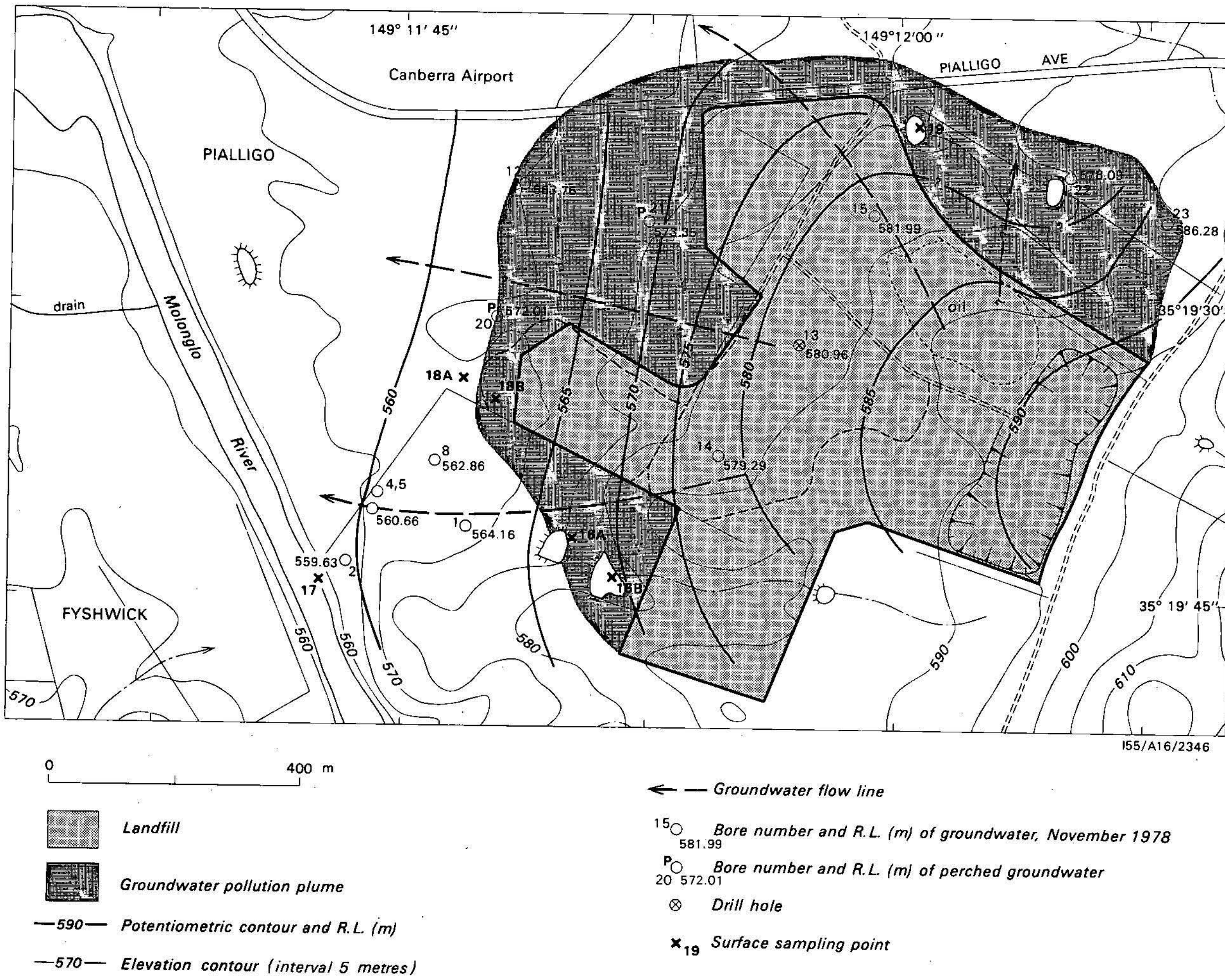

Figure 7. Groundwater regime, Pialligo landfill site.

	1* Rock	12* Surficial	12† Surficial	14* Landfill	15* Landfill	20† Surficial	21† Surficial	22† Rock
	Bore No. and aquifer type							
pH	7.4	6.9	7.0	5.1	5.3	6.4	6.7	7.8
T.D.S. calculated	301	852	906	9375	14213	111	457	677
T.D.S. evaporated residue	285	833	—	30662	18898	—	—	—
Electrical conductivity	576	1542	1760	16000	18799	219	819	1126
Calcium	71	63	74	1912	3500	6	42	17
Magnesium	21	42	49	575	1160	6	23	9
Sodium	17	214	212	945	1115	28	83	238
Potassium	3	2	2	475	392	6	11	6
Bicarbonate	319	530	573	8037	12041	102	264	540
Sulphate	9	53	30	259	634	7	13	24
Chloride	23	217	256	1256	1490	8	89	67
Fluoride	0.20	0.10	0.10	0.05	0.10	0.05	0.15	0.60
Phosphate	0.07	0.04	0.09	0.35	<0.01	0.17	0.09	0.34
Boron	<0.05	<0.05	<0.05	2.50	3.35	0.05	0.05	0.10
Total hardness	264	330	387	7140	13513	40	200	91
Total alkalinity	261	435	470	6588	9870	83	217	443
Cadmium (μg/l)	5	<3	<0.05	<3	<3	< 0.05	<0.05	0.3
Chromium (μg/l)	<3	<3	1	25	41	10	5	5
Copper (μg/l)	80	<10	<0.5	60	30	9.1	5.4	13.1
Iron	0.20	0.11	0.24	1037	765	10.9	0.24	6.30
Manganese	1.65	0.95	—	400	266	—	—	—
Nickel (μg/l)	15	15	10	1170	600	10	5	5
Lead (μg/l)	<30	<30	27	220	220	<0.5	3.4	7.2
Zinc (μg/l)	295	35	28	83500	23650	55	5.3	55
Petroleum spirit extract	<0.5	<0.5	—	98	53	—	—	—
Surfactants (μg/l)	50	<25	42	210	170	<50	51	<50
Chemical oxygen demand	18	42	—	45500	50000	—	—	—
Total nitrogen	0.65	1.1	—	1060	660	—	—	—
Ammonia	N.D.	N.D.	—	605	75	—	—	—

* July 1977 † February 1979

Table 2. Chemical analyses of groundwater and leachate, Pialligo, ACT (in mg/l; electrical conductivity in μs/cm).

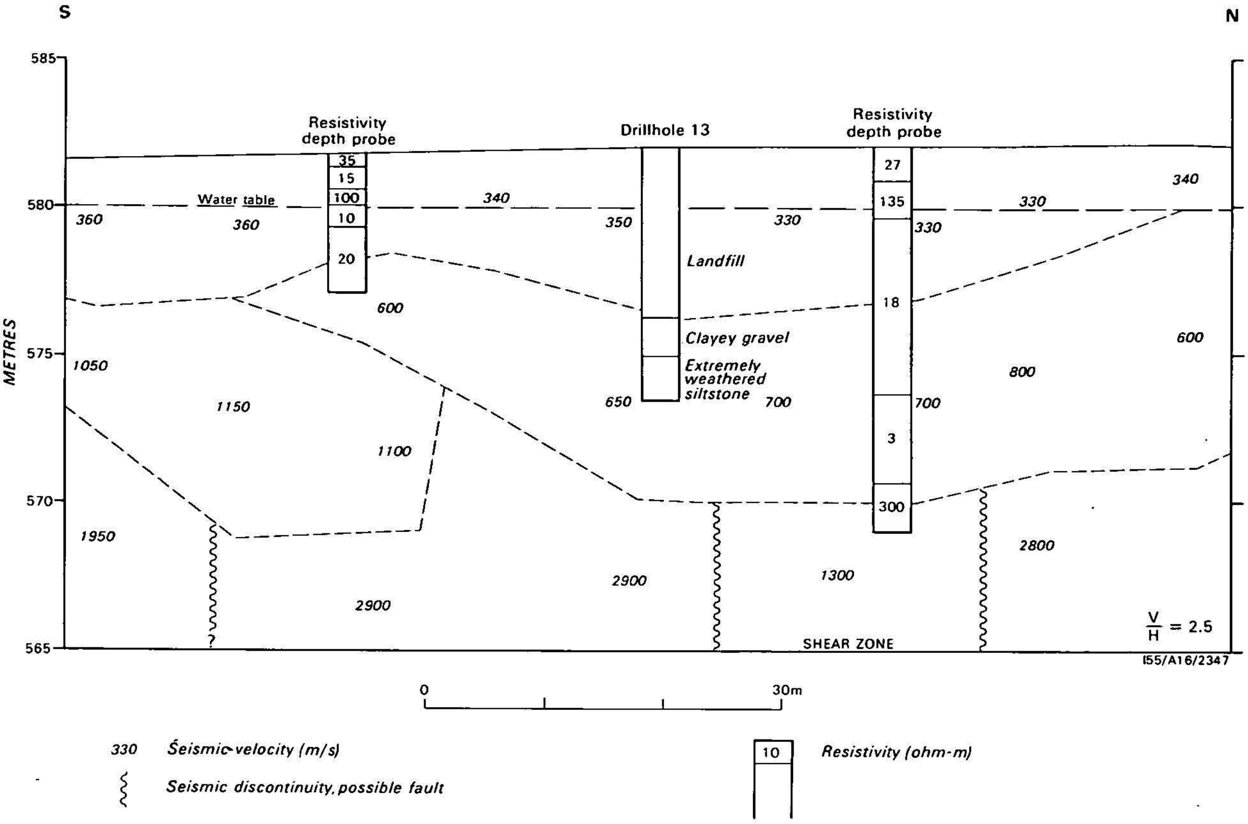

Figure 8. Part of a seismic traverse correlated with a drillhole, Pialligo landfill site.

Groundwater pollution by leachate—Pialligo landfill site

The Pialligo landfill site opened in the mid-1960s and was the main landfill site for Canberra's domestic waste until 1978. Most of the site was developed before the importance of geological factors in landfill operations was recognised. The area contained worked-out sand pits, and fill was placed in existing pits or in excavated trenches to depths of several metres and then covered with up to one metre of soil. A post-1976 extension of the site to the south was designed by consulting engineers, and was filled in compartmentalised trenches, 6m deep, and covered with compacted soil.

The Pialligo landfill site is on aeolian and colluvial slope and terrace deposits of the K1, K2, and K3 cycles (Van Dijk, 1959) between the Molonglo River to the west and a low bedrock ridge to the east (Fig. 6). Bedrock beneath the site consists of fault-bounded blocks of Silurian silstone, calcareous mudstone, and dacite. The rock is weathered to 10-15 m below the surface.

Geological investigations have been undertaken to determine and monitor the extent of groundwater pollution, (Evans & others, 1979). The investigation included 19 drillholes, of which 12 encountered groundwater and were equipped with slotted plastic casing to serve as monitoring bores.

Groundwater occurs in the landfill, in the surficial deposits, and in fractured rock. The potentiometric contours (Fig. 7) indicate that the landfill forms a groundwater mound which recharges the natural aquifers by lateral infiltration through spatially variable gravel lenses, and by vertical infiltration to the water table. Groundwater flows westward, towards the Molonglo River, and northward. The landfill itself is recharged by the infiltration of rainfall and surface water through the covering material.

Seasonal fluctuations in groundwater levels range up to 2 m, and some bores in perched surficial aquifers dry up seasonally, indicating that lateral infiltration ceases as the water level falls. Climatic data (Fig. 3) and water level hydrographs in observation bores indicate that groundwater recharge and the generation of leachate take place when the soil moisture content is near capacity, enabling steady infiltration to the water table immediately after rainfall.

Water quality is monitored every 6 months. Accepted background levels of DOC are 1-5 mg/l for natural groundwaters and 10-20 mg/l for dam waters. Pollution of surficial aquifers has been evident in bores 12, 21, and 23, where over 10 mg/l DOC has been determined, and pollution of the fractured rock aquifer is evident in bore 22. The probable extent of the pollution plume around the site is shown in Figure 7.

Chemical analyses of some groundwater and leachate samples are shown in Table 2. The leachate contains 9000-14 000 mg/l total dissolved solids, in contrast to slightly polluted groundwater which contains 400-900 mg/l, and apparently unpolluted groundwater, 100-300 mg/l. The electrical conductivity and chemical oxygen demand of the leachate are correspondingly high. The leachate is acid, with pH 5.1-5.3.

There is no extraction of groundwater at present in the immediate vicinity of the site, and no immediate pollution hazard to the Molonglo River. However, continued monitoring is necessary.

A seismic refraction survey was carried out at the site in order to measure the thickness and seismic

	Investigation						
	Seismic traverses	*Auger holes*	*Geology*	*Ease of excavation*	*Permeability of cover material when compacted*	*Groundwater occurrence*	*Permeability of basal material of landfill excavation*
Mugga North	6	14	Sandy colluvial soil, up to 5 m overlying weathered rhyodacite and mudstone	Can be excavated by mechanical means to average 6 m	Low to moderate	Colluvial sand aquifers at shallow depth encountered in two drillholes	Probably low if base of excavation in weathered rock
Mugga South	—	6	Sandy colluvial soil, 5-8 m deep overlying weathered volcanic rock	Readily excavated by mechanical means to at least 5 m	Low to moderate	Encountered in one drillhole	Not known
Wanniassa	2	7	Weathered volcanic rock, in part overlain by sandy alluvial and colluvial soils	Variable; rock outcrops require blasting and weathered profile is irregular	Variable from low to high	Alluvial aquifer with shallow water table in northeast part of site	Not known
West Jerrabomberra	—	3	Alluvium: interbedded sand, silt, and clay	Readily excavated by mechanical means to at least 7 m on S side	Variable from low to high	Alluvial aquifer with shallow water table encountered in southwest part of site	Variable
New Station Creek	1	3	Sandy colluvial soil 4-6 m deep overlying weathered volcanic rock	Readily excavated by mechanical means to 12-15 m	Low to moderate	Not encountered	Variable

Table 3. Alternative landfill sites for South Canberra and Tuggeranong.

Figure 9. Mugga South sanitary landfill site, ACT, developed on Cainozoic colluvial outwash fan deposits.

velocity of the fill and underlying material, and assess the degree of compaction of the fill. An example of part of a seismic traverse, correlated with a drillhole, is shown in Figure 8. The inferred materials corresponding to seismic velocity are: landfill—320-470 m/s; extremely weathered rock—600-800 m/s; highly weathered rock — 1000-1300 m/s; moderately weathered rock—1700-2000 m/s; slightly weathered rock—2500-3200 m/s. The seismic velocity of the fill layer did not increase significantly with age. In areas identified as landfill more than 10 years old, the velocity of the fill layer was 340-470 m/s. In the area identified as landfill 3-5 years old, the velocity was 320-380 m/s.

Resistivity depth probes were also recorded (Fig. 8), but correlation with drillhole data was poor, possibly because of the variable electrical properties of the fill. In general, the resistivity of the fill is low compared with measurements taken outside the area. Resistivity traverses were recorded along the perimeter of the site, in an attempt to identify permeable pathways draining leachate from the fill, and with a view to monitoring gross changes in groundwater composition with time.

Future land use of the Pialligo site may include building development or recreational activities. At present, the site is undergoing continued differential settlement, and cracks up to 5 cm wide and 30 m long have developed in the cover material. The bearing capacity may have to be increased by placing additional cover material. Alternatively, buildings will have to be founded on continuous concrete slabs or on piles or piers which transmit loads to bedrock. Because of the possibility of methane gas being generated in the fill, structures will require gas-tight membranes beneath the ground floor slab.

To reduce the generation of leachate, recharge to the landfill should be minimised. Planning of future land use should ensure adequate drainage of the site. The cover material should have a low infiltration capacity, and should be as thick as possible to allow for soil moisture storage rather than infiltration.

The influence of geology on site selection—Mugga South landfill site

With the impending closure of the Pialligo site, a geological evaluation of five possible landfill sites in South Canberra and Tuggeranong was undertaken in 1976-7 (Evans & others, 1978). The sites were initially selected on the basis of urban planning considerations and likely geological suitability (Table 1).

The five sites were investigated by geological mapping, power augering, and seismic survey to evaluate excavation conditions. From past experience in the Canberra area, material with seismic velocity up to 1200 m/s is considered to be rippable. Rock with seismic velocities of 1200-2000 m/s may be rippable, but will probably require blasting. The Gemco power auger generally failed to penetrate material with a seismic velocity greater than 1200 m/s.

A summary of the relevant engineering properties of the material at each of the five sites is given in Table 3. Comparative evaluation showed that the best sites for depth and ease of excavation, and suitability of cover materials were those designated Mugga South and New Station Creek. Two other sites, Mugga North and West Jerrabomberra, were also suitable for landfill, but the fifth site, Wanniassa, was unsuitable, because of the presence of rock close to the surface.

The decision to utilise the Mugga South site was based on general planning grounds as well as the assurance of geological suitability. Additional drilling and seismic surveys were made at the selected site to provide detailed information on excavation conditions for design (Evans & Bennett, 1978). The thickness of readily excavatable material was found to be more than 5 m over most of the site. The groundwater regime was invesigated, and a leachate monitoring system was instituted. Trenching and filling at the site began in 1979.

The Mugga South landfill site is being developed on dissected colluvial outwash fans that grade laterally into valley-floor alluvium (Fig. 9). These deposits correspond to the older K cycles (K5 and K6) of Van Dijk (1959). The underlying bedrock is highly to extremely weathered dacite and dacitic tuff. Groundwater occurs in both perched colluvial/alluvial aquifers and in fractured rock. The groundwater regime is shown diagrammatically in Figure 10. On the upper slopes, groundwater movement is mainly lateral, within the younger, more porous colluvium, and vertical infiltration is restricted by older, indurated colluvium. The downslope groundwater flow recharges the valley-

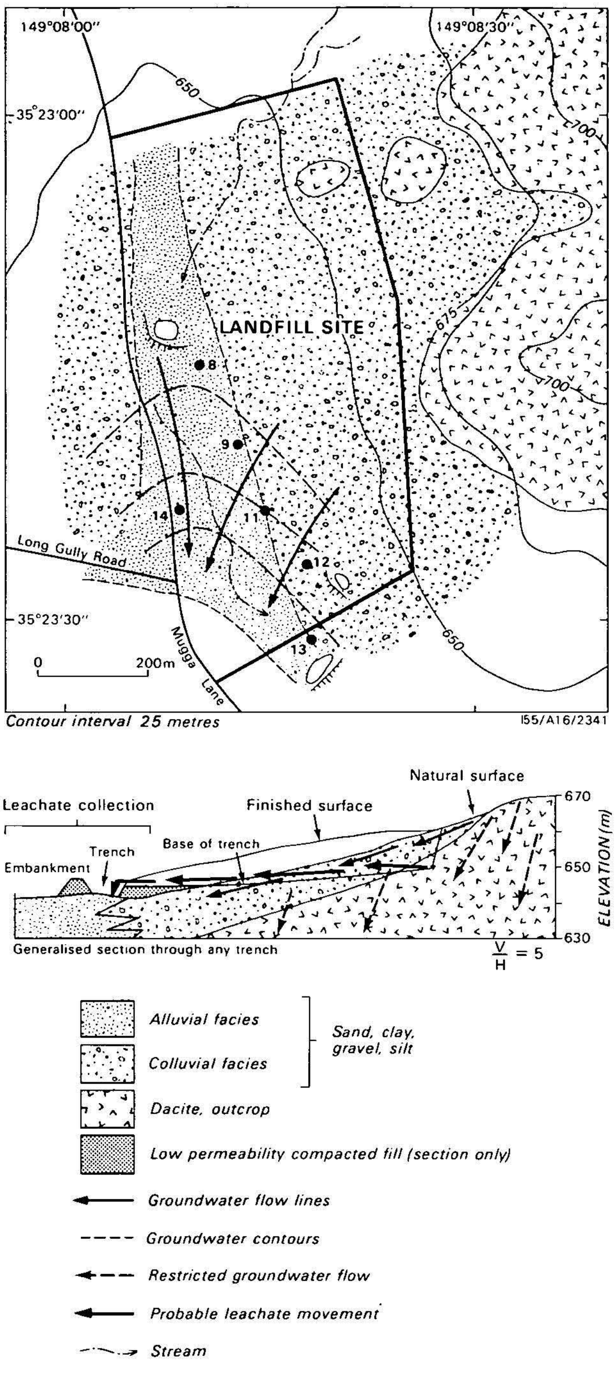

Figure 10. Mugga South sanitary landfill site.

floor alluvium. Groundwater levels are generally more than 4 m below surface. The landfill trenches are designed to be clay blanketed to minimise the entry of water into the fill and the infiltration of leachate into the groundwater.

As in the other sites the natural chemistry of the groundwater is variable, with total dissolved solids ranging from 70 mg/l in an alluvial bore to 1400 mg/l in a fractured rock bore. Background values of DOC have been 1-5 mg/l in groundwater and 14-50 mg/l in dam water.

Conclusions

At the West Belconnen landfill site, geology has influenced the design and management, which have been carried out in such a way as to minimise groundwater pollution and assist future reclamation of the site. The Pialligo landfill site is an example of a pollution problem, and possibly a reclamation problem, developing because of the lack of geological consideration during the early development of the site. The Mugga South landfill site is an example of the usefulness of rapid geological and geophysical assessment in landfill-site selection.

The feasibility and effectiveness of the sanitary landfill technique depends largely on geological conditions. The integration of waste-disposal geology with early planning of urban development is therefore highly desirable. As land-use pressures develop, the importance of detailed geological mapping and knowledge of the soils, geomorphology, and hydrology of urban and near-urban areas will increase. Geophysical techniques, especially seismic refraction, have a definite role in the rapid evaluation of sites. Modern standards for pollution control require detailed groundwater investigations and leachate monitoring systems.

Waste disposal in the ACT is concerned mainly with relatively harmless domestic wastes; the problems are bigger with more toxic and hazardous wastes, but the principles of containment are similar.

Acknowledgements

The investigations described in this paper were undertaken for the National Capital Development Commission and the Department of the Capital Territory. Some of the investigations were done in collaboration with S. Hollis of L. T. Frazer & Associates—the Commission's consulting engineers for waste disposal —and with M. J. Storey of the Department of the Capital Territory's Conservation and Agriculture Laboratory, who assisted with sampling and did the DOC analyses. The geophysical surveys at Pialligo and Mugga were made by D. G. Bennett and G. R. Pettifer. The West Belconnen site was originally investigated by P. H. Vanden Broek in 1971, and additional work was done by P. D. Hohnen and J. R. Kellett in 1978. Assistance was given with dye tracing by D. Ingle Smith, Centre for Resource and Environmental Studies, ANU, and with cation exchange tests on soils by H. J. Beatty, CSIRO Division of Soils. Chemical analyses of groundwater and leachate samples were done by AMDEL, Adelaide.

References

BESTOW, T. T., 1977—The movement and changes in concentration of contaminants below a sanitary landfill, Perth, Western Australia. *In* Effects of urbanisation and industrialisation on the hydrological regime and on water quality. Proceedings of the Amsterdam Symposium, October, 1977. *International Association of Hydrological Sciences, Publication* **123**.

EVANS, W. R. & BENNETT, D. G. 1978—Geological and geophysical investigation of the proposed Mugga South landfill site, A.C.T., 1977. *Bureau of Mineral Resources, Australia, Record* **1978/69** (unpublished).

EVANS, W. R., BENNETT, D. G. & JACOBSON, G., 1978—Geological and geophysical investigations of five alternative landfill sites, South Canberra and Tuggeranong, A.C.T., 1976. *Bureau of Mineral Resources, Australia, Record* **1978/68** (unpublished).

EVANS, W. R., JACOBSON, G. & BENNETT, D. G., 1979—Geological and geophysical investigation of the Pialligo landfill site, A.C.T. *Bureau of Mineral Resources, Australia, Record* **1979/77** (unpublished).

HAGERTY, D. J., & PAVONI, J. L., 1973—Geologic aspects of landfill refuse disposal. *Engineering Geology* **7**, 219-29.

HEITFELD, K. H., KRAPP, L. & OLZEM, R., 1976—Fundamentals to develop a programme for the optimum design of regional refuse deposits. *Bulletin of the International Association of Engineering Geology*, **13**, 99-103.

HILL, P. J., 1972—Belconnen refuse disposal area seismic refraction survey, A.C.T., 1971. *Bureau of Mineral Resources, Australia, Record* **1972/8** (unpublished).

HUGHES, J. L., ECCLES, L. A. & MALCOLM, R. L., 1974—Dissolved Organic Carbon (DOC), an index of organic contamination in groundwater near Barstow, California. *Ground Water* **12**, 283-90.

JOINT COMMITTEE ON THE AUSTRALIAN CAPITAL TERRITORY, 1976—Canberra City wastes—a long term strategy for collection and disposal. *Parliament of the Commonwealth of Australia, Report*, December 1976.

KNIGHT, M. J., LEONARD, J. G. & WHITELY, R. J., 1978—Lucas Heights solid waste landfill and downstream leachate transport—a case study in environmental geology. *Bulletin of the International Association of Engineering Geology*, **18**, 45-64..

NCDC, 1975—A possible refuse planning strategy for Canberra. *National Capital Development Commission, Canberra.*

SENDLEIN, L. V. A. & PALMQUIST, R. C., 1975—A topographic—hydrogeologic model for solid waste landfill siting. *Ground Water*, **13**, 260-68.

STRUSZ, D. L. & HENDERSON, G. A. M., 1971—Canberra City, A.C.T., 1:50 000 geological map and explanatory notes. *Bureau of Mineral Resources, Canberra.*

UK DEPARTMENT OF THE ENVIRONMENT, 1978—Cooperative programme of research on the behaviour of hazardous wastes in landfill sites. Final report of the Policy Review Committee. *HMSO, London.*

VAN DIJK, D. C., 1959—Soil features in relation to erosional history in the vicinity of Canberra. *CSIRO, Australia, Soil Publication* **13**.

WILSON, E. G., 1975—The influence of geology and hydrogeology on landfill disposal of solid waste, and its application in Canberra, A.C.T. *Bureau of Mineral Resources, Australia, Record* **1975/99** (unpublished).

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Sedimentological studies of Cainozoic sediments from the Exmouth and Wallaby Plateaus, off northwest Australia

J. B. Colwell & U. von Stackelberg[1]

Eighty-two cores of Quaternary and Quaternary/Tertiary sediments were taken by the R/V *Sonne* in the Exmouth and Wallaby Plateau areas off northwestern Australia in 1979. The Quaternary sediments show a systematic variation in composition with water depth which basically reflects change in the abundance and nature of certain biogenic components. These changes are mainly related to the positions of the aragonite and carbonate compensation depths, which presently lie in the region at about 800 m and between 4100 and 4800 m, respectively. The sediments form four major facies, ranging from relatively coarse carbonate sands on the continental shelf to siliceous clays on the abyssal plains.

Tertiary sediments, which were penetrated in four of the cores, mainly consist of Oligocene or Miocene foraminiferal nanno oozes or chalks. A volcaniclastic sandstone containing rounded phosphatic nodules was penetrated on the eastern margin of the Wallaby Plateau and is consistent with the proposed volcanic origin of the plateau.

Introduction

Early in 1979 the West German research vessel *Sonne* undertook a 6-week cruise (cruise SO-8) over the Exmouth and Wallaby Plateaus, off northwestern Australia. During this cruise, which was aimed primarily at sampling Mesozoic rocks on the flanks of the plateaus, 82 cores, totalling 196 m of Quaternary and Quaternary/Tertiary sediment, were taken in water depths ranging from 122 to 5676 m. Figure 1 shows the location of the coring stations; station and core recovery data are given in the unpublished cruise reports of Exon (1979) and von Stackelberg (1979a). This was the first comprehensive sampling of Cainozoic sediments undertaken in the region.

This paper presents the results of sedimentological studies made on these cores and, to a limited extent, on material from two cores collected in the area by the R/V *Robert Conrad* and R/V *Vema*. These studies were undertaken jointly at the Bureau of Mineral Resources, Geology & Geophysics, Canberra, by JBC (sediment descriptions and coarse-fraction analyses) and at the Bundesanstalt für Geowissenschaften und Rohstoffe, Hannover, by UvS (core descriptions and sedimentary structures), and were primarily aimed at determining the nature of Quaternary sedimentation in the region. Additional studies are in progress at Kiel University (Sarnthein and co-workers) on palaeoclimatic aspects of the Quaternary sedimentation. Other aspects of the scientific evaluation of the cruise are being studied by von Stackelberg & others (1980)—geology of the pre-Quaternary sequences; Zobel—Quaternary biostratigraphy; Quilty (1980)—Tertiary biostratigraphy; and Exon & others—development of the passive margins of the Exmouth Plateau.

The work undertaken by R/V *Sonne* on the Exmouth and Wallaby Plateaus follows work undertaken by R/V *Valdivia* on the Scott Plateau and Java Trench areas in 1977 (Hinz & others, 1978), and by R/V *Atlantis II* on the Argo Abyssal Plain in 1976 (Cook & others, 1978).

Regional setting

The Exmouth and Wallaby Plateaus (Fig. 1) lie off northwest Australia in water depths ranging from 800 m to 2000 m and 2100 m to 3500 m, respectively. They lie oceanward of the Northwest Shelf and Carnarvon Terrace, and are surrounded on their northern, western and southern sides by abyssal plains of the Wharton Basin, where depths exceed 5000 m. The physiography of the region is largely structurally controlled with a predominant northeasterly trend (Willcox & Exon, 1976). The Swan Canyon and the Montebello Canyon are major features which disrupt the

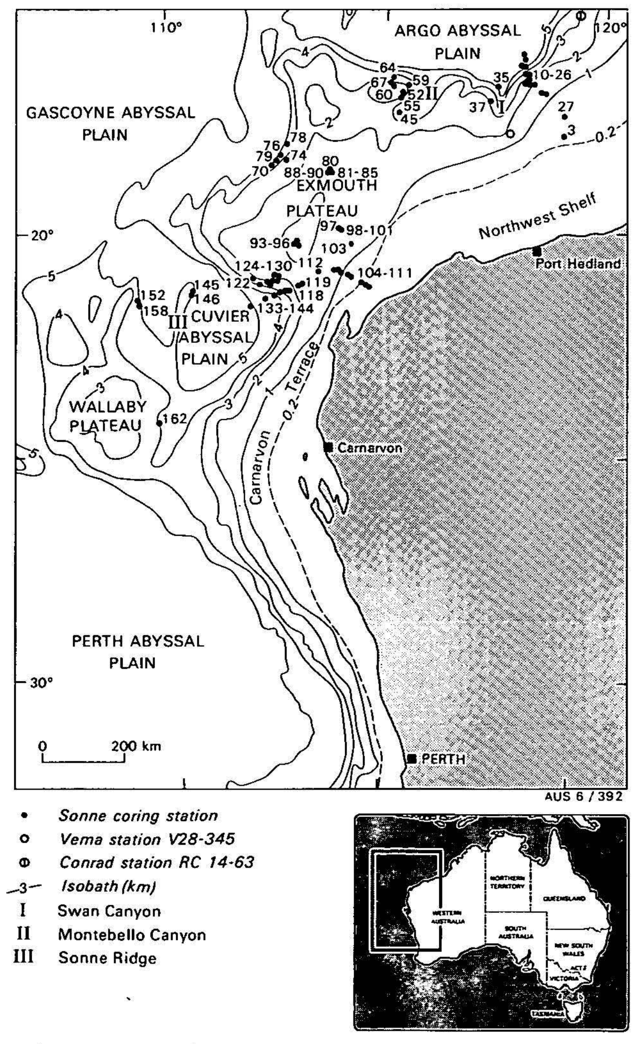

Figure 1. Location of successful coring stations.

1. **Bundesanstalt für Geowissenschaften und Rohstoffe, Hannover, F.R. Germany.**

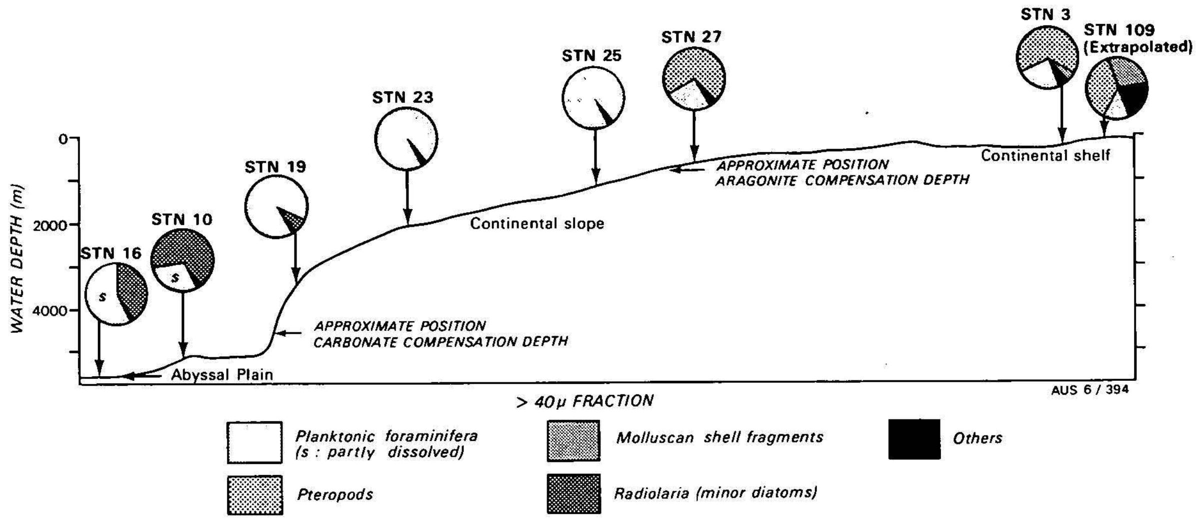

Figure 2. Profile across the continental shelf and slope north of the Exmouth Plateau, showing the variation in surface element composition with depth. Refer to Figure 1 for location.

northern margin of the Exmouth Plateau (Fig. 1); smaller canyons are common along the western margin of the Carnarvon Terrace. Ridges of volcanic rock extend northward from the Wallaby Plateau and probably represent former spreading ridges associated with the development of the Cuvier Abyssal Plain (Larson & others, 1979). A detailed description of the physiography of the region is given by Falvey & Veevers (1974).

Methods

Cores were obtained using either a 5 m or 10 m piston corer (outer diameter 80 mm or 105 mm), a 6 m gravity corer (outer diameter 120 mm), several 1.2 m free-fall corers (outer diameter 73 mm), either a 6 m or 12 m Kastenlot (box corer with a 300 x 300 mm cross-section), and a Kastengreifer short box corer (capacity about 30 l). The box and gravity corers were mainly used by scientists from Kiel University in several detailed transects over the continental slope northeast and south of the Exmouth Plateau (stations 3-26, 104-111 and 133-144). The results of preliminary examination of the sediments made aboard ship are given by von Stackelberg & others (1980). The results of a shipboard investigation by Quilty and Shafik of foraminifera and coccoliths were used as a basis for the stratigraphic subdivision used by von Stackelberg & others (1980) and in this paper. Biostratigraphic work is continuing at Hannover.

Lithological descriptions, X-ray and visual descriptions of sedimentary structures, semi-quantitative coarse-fraction analysis of the 40-63 µm and > 63 µm fractions, and smear-slide analysis of the < 40 µm fraction were undertaken onshore. X-ray photographs (radiographs) were made of slices cut 5 mm thick along the cores over a length totalling 33 m. The most intensive investigations were carried out on the BGR piston cores. In general, the cores were sampled at 10, 20 or 30 cm intervals, from the surface to the core catcher—the surface sediments being examined to give an indication of the patterns of sedimentation existing at present over the region. Calcium carbonate values were determination for representatives samples by the CO_2-evolved method at the Australian Mineral Development Laboratories, Adelaide. All cores are stored either at the BGR or Kiel University.

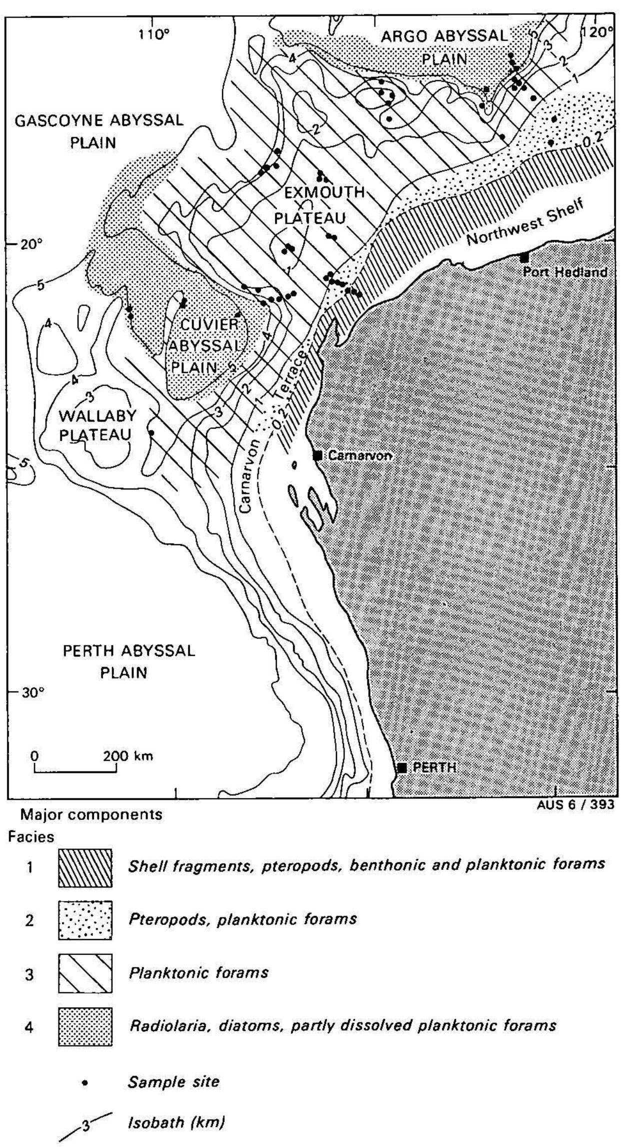

Figure 3. Composition of the surface sediments (>40 µm fraction).

Nature of the surface sediments over the region

The surface sediments show a systematic variation in composition with water depth, which reflects changes in the abundance and nature of certain biogenic components (Figs. 2 and 3). The most significant of these changes are the disappearance of pteropod material at about 800 m and the pronounced reduction in planktonic foraminifera, which occurs between 4100 m and 4800 m, corresponding to the carbonate compensation depth (CCD).

In broad terms, four major facies are represented: Facies 1 (<200 m water depth)—shelly carbonate sands; Facies 2 (200-800m)—foraminiferal oozes and sands containing pteropod material; Facies 3 (800-4500 m)—foraminiferal oozes and sands; and Facies 4 (>4500 m)—siliceous clays containing partially dissolved planktonic foraminifera. The foraminifera, which make up a volumetrically very important part of many of the sediments, typically belong to the tropical-subtropical eastern Indian Ocean assemblage described by Bé & Hutson (1977).

The distribution of the facies approximately parallel to the bathymetry (Fig. 3) reflects various physical, chemical, and biological conditions of the environment. The distribution of Facies 1 is controlled mainly by shallow-water conditions (water depth, substrate, turbulence etc.), and the distribution of the other facies, mainly by the positions of the aragonite and carbonate compensation depths. The aragonite compensation depth (ACD), which is marked by the disappearance of pteropod (aragonite) material from the sediments (change from Facies 2 to Facies 3), lies at about 800 m, which is similar to the postglacial ACD recorded by Berger (1977) in other areas, mainly in the west equatorial Pacific and North Atlantic.

Although the CCD lies somewhere between 4800 m and 4100 m, partially dissolved foraminifera and coccoliths occur below this depth range as a result of turbidity current activity and slumping. Coccoliths tend to persist to somewhat greater depths than foraminifera, owing to their slightly greater resistance to solution (McIntyre & McIntyre, 1971).

Description of the core material

In most of the 82 cores recovered during the cruise only one facies is represented. This reflects the relative consistency of the patterns of Quaternary sedimentation, and also, to some extent, the distribution of the cores with respect to water depth. Very few cores were taken close to the CCD, owing to the steep bottom gradients of the lower continental slope. The few cores reaching the Tertiary (64 KL, 78 KL, 132 KL and 142 KL) generally show a major change of facies at the Quaternary/Tertiary boundary, from an Oligocene or Miocene foraminiferal nanno chalk to Quaternary foraminiferal sands and oozes. Aspects of the cores are discussed below; analytical and descriptive data are summarised in Table 1.

Facies 1 cores

Six cores of Facies 1 sediments, consisting of shelly carbonate sands and sandy oozes, were recovered from the continental shelf and slope in water depths ranging from 122 m to 498 m (Fig. 4). The cores ranged in length from 19 cm to 353 cm, and averaged 132 cm. In general, over 60 percent of the sediment is sand, although this value decreases to less than 40 percent in cores on the upper slope. The high sand values reflect the hydrodynamic conditions on the shelf, which, as noted by Jones (1971), is largely a zone of winnowing and sediment by-pass rather than of deposition. Mollusc fragments, pteropod shells, and planktonic and benthonic foraminifera are the main skeletal components present.

Facies 2 cores

Thirty-four Facies 2 cores were recovered in water depths of 800 m to 1500 m from a wide area of the Exmouth Plateau (see Fig. 4). The sediments consist mainly of pelagic foraminiferal oozes and sands, containing in the 40-63 μm fraction, and to a lesser extent in the > 63 μm fraction, significant quantities of pteropod material. The presence of the pteropods distinguishes the cores from the Facies 3 cores, which occur on the deeper parts of the plateau below the ACD. In general, the planktonic foraminifera belong to the following genera—*Globorotalia* (*G. menardii* and *G. truncatulinoides* predominant), *Globigerinoides, Globigerina, Orbulina, Pulleniatina* and *Sphaeroidinella*, and are similar to the foraminifera occuring in the surface sediments. Full species determinations are being undertaken by B. Zobel at BGR.

In many of the cores, the pteropod component increases markedly with depth in the core, commonly from zero at the surface; this feature is particularly evident in cores 90 BL, 93 KA, 96 KA, 99 BL (Fig. 5) and 106 KAL. This increase probably reflects fluctuations in the ACD. According to Berger (1977), the ACD has fluctuated considerably during the last 20 000 years, culminating in a pteropod preservation spike at about 14 000 yr B.P., when the ACD was markedly lowered. The difference in the distribution of the Facies 2 sediments between the surface sediments (Fig. 3) and the cores (Fig. 4) reflects the rapid postglacial rise in the ACD recorded by Berger (1977). Further work, partly underway at Kiel University, needs to be undertaken on the present group of cores to further identify these fluctuations.

Radiographs covering the total length of a typical core (core 99 BL) reveal indistinct lenticular layering and intensive bioturbation. Horizontal and vertical burrows of 4-10 mm diameter are common throughout the core.

Facies 3 cores

Thirty Facies 3 cores, consisting mainly of foraminiferal oozes and sands, were recovered from the deeper parts and margins of the Exmouth Plateau as well as from the eastern margin of the Wallaby Plateau (Fig. 4). The cores ranged in length from 5 cm to 845 cm, and averaged 337 cm. The sediments generally contain 20-60 percent sand and 5-10 percent silt, both values being largely related to the content of planktonic foraminifera (calcium carbonate values generally range from 60 percent to 75 percent). In most cores the relatively high sand values probably reflect the winnowing of 'fines' (mainly nannofossil remains) from oozes by current action.

The >63 μm and 40-63 μm fractions typically consist of planktonic foraminifera, generally of the same genera as listed for the Facies 2 cores. Few of the foraminifera show significant signs of solution and most of the cores appear to have remained considerably above the CCD throughout the Quaternary.

Facies	Basic lithology	Surface distribution (water depth)	Skeletal components* >63 µm frac.	Skeletal components* 40-63 µm frac.	Typical grainsize[1] % Sand	Silt	Clay[2]	Core distribution (Core Nos)[3]	% $CaCo_3$[1]	Colour
1	Shelly carbonate sands	<200 m	M T P B (Te) (E) (S) (O)	T M P B (Te) (E) (S)	>60	5-15 but variable	10-20	3KA, 107KA, 108SL, 109KA, 110KA, 111SL	67-90 (80) N=13	Pale yellowish brown (10YR 6/2) to light olive grey (5Y 6/1)
2	Foraminiferal oozes and sands containing pteropod material	~200-800 m	P T in certain layers (B) (E) (S) (F)	P T (B) (S)	20-40 (22.9)	3-15 (5.1)	50-80 (72.0)	25KA, 26SL, 27KA, 80KA, 81-85BL, 88-90BL, 93KA, 94SL, 95KA, 96KA, 97KA, 98-101BL, 103SL, 104KA, 105SL, 106KAL, 112-117BL, 118KA, V28-345	60-89 (74) N=49	Greyish orange pink (10R 8/2) to light olive grey (5Y 6/1)
3	Foraminiferal oozes and sands	~800-4500 m	P (E) (B) (S) (F) (O)	P (S) (B) (E) (F) (R)	20-60 (27.1)	5-10 (7.3)	40-80 (65.6)	19KA, 20KA, 21SL, 22KAL, 23KA, 24SL, 52KL, 55KL, 59KL, 60KA, 64KL, 67KAL, 70KL, 74KL, 78KL, 79KL, 119SL, 122KL, 124KL, 125-128BL, 132KL, 138SL, 141KA, 142SL, 162KL, RC14-63	60-75 (68) N=20	Yellowish brown (10YR 5-6/2) to light olive grey (5Y 6/1)
4	Siliceous clays	>4500 m	Ps R D in some cores S (F) (Q)	Ps R D in some cores S (F) (Q) (Sc)	<8 (5.3)	<8 (5.9)	>85 (88.6)	10BL, 12KA, 13SL, 16SL, 18BL, 35KL, 133KA, 145KA, 146SL	Highly variable (47) N=14	Brown (10YR 4/4-10YR 6/4)
4A	Siliceous clays containing manganese micronodules	—	Mn R Z (F) (S) (Q) (D)	N.D.	<2 (1.5)	<1 (0.7)	>97 (97.8)	152KL, 158KL	N=0	Brown (10YR 4/2-7.5YR 6/6)

* M	Mollusc fragments	D	Diatoms	Mn	Manganese micronodules
R	Radiolaria	S	Sponge spicules	Os	Ostracods
E	Echinoid fragments	Sc	Silicoflagellates	Z	Zeolitic grains
Q	Quartz	B	Benthonic forams	O	Others (scaphopods, undiff. material, etc.)
Te	Undiff. terrigenous material	Ps	Partially dissolved planktonic forams	N.D.	Not determined
P	Planktonic forams	F	Fish debris		
T	Pteropod fragments				

Minor components shown in brackets.

1. Based on representative samples. Average values in parentheses. N = No. of samples analysed.
2. Mainly clays, siliceous organisms and coccoliths.
3. KA, BL etc. refer to the coring device used. KA: short box corer, KAL: long box corer, KL: piston corer, SL: gravity corer, BL: boomerang corer.

Table 1. Characteristics of the Quaternary sediment facies.

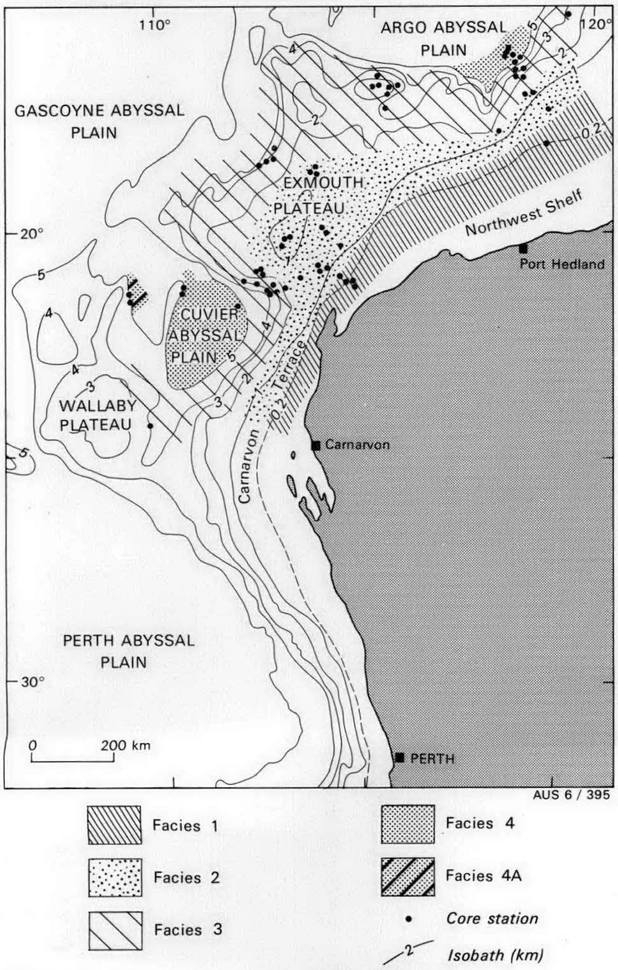

Figure 4. Distribution of the different types of cores.

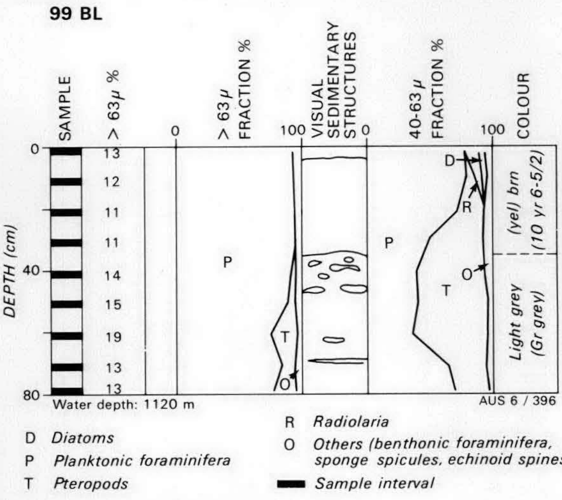

Figure 5. Typical core, Facies 2.

Cores 64 KL, 78 KL, 132 KL, and 142 SL differ from the other Facies 3 cores in that they penetrate pre-Quaternary sediments. Core 78 KL, which is fairly typical of this group of cores, consists of 86 cm of Quaternary foraminiferal ooze with indistinct lenticular layering and light, mainly horizontal, burrows (Fig. 6A), overlying early Miocene stiff light greyish brown, nanno ooze and chalk. A foraminifera-rich sand layer, 1 cm thick, occurs on the unconformity surface (Fig. 6B), below which the scarcity of calcareous tests and the occurrence of fish debris indicate severe solution.

Core 162 KL, on the edge of the Wallaby Plateau, penetrated typical Quaternary foraminiferal nanno ooze overlying a volcaniclastic clayey sandstone containing rounded phosphatic nodules up to 5 cm in diameter. The presence of the phosphatic nodules suggests low rates of sedimentation and the formation of lag deposits. The sandstone was dated on board, on the basic of nannofossils, as early Pleistocene, although the nannofossils possibly include some reworked material. The presence of the sandstone is consistent with the volcanic origin of the Wallaby Plateau proposed by Veevers & Cotterill (1978).

Facies 4 cores

Nine Facies 4 cores were recovered from the Argo and Cuvier Abyssal Plains in water depths ranging from 4947 m to 5676 m (Fig. 4). The cores, which range in length from 12 cm to 570 cm, consist mainly of brown slightly sandy siliceous clays, although thin foraminifera-rich sand layers occur in places (see Figs.

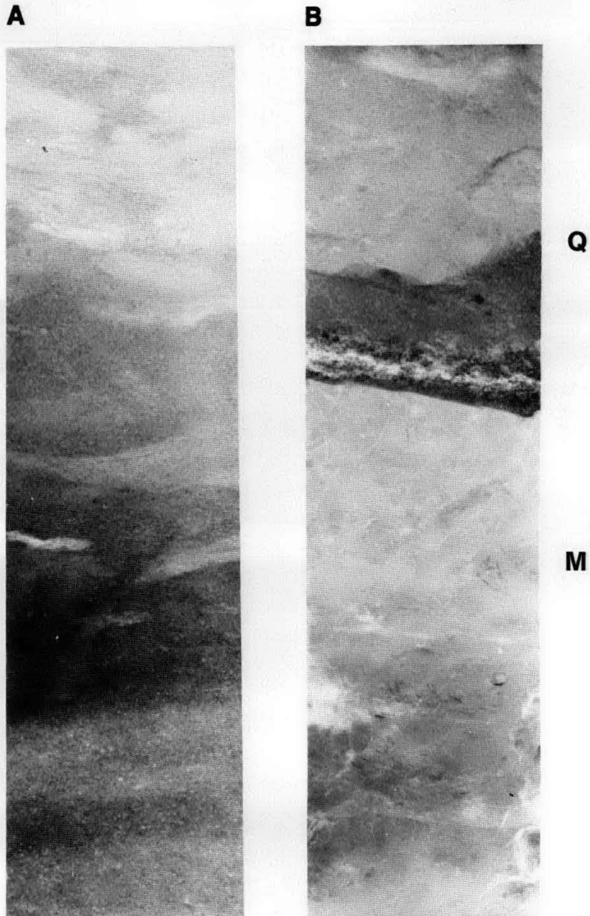

Figure 6. Radiographs (positive, x 0.8) of core 78KL.
(A) 60-76.5 cm, showing a foraminiferal ooze with indistinct lenticular layering and light mainly horizontal burrows, and (B) 80-96.5 cm, showing Quaternary sands and oozes containing abundant foraminifera (Q) overlying stiff early Miocene nanno ooze (M). The unconformity occurs at the base of the dark sand layer.

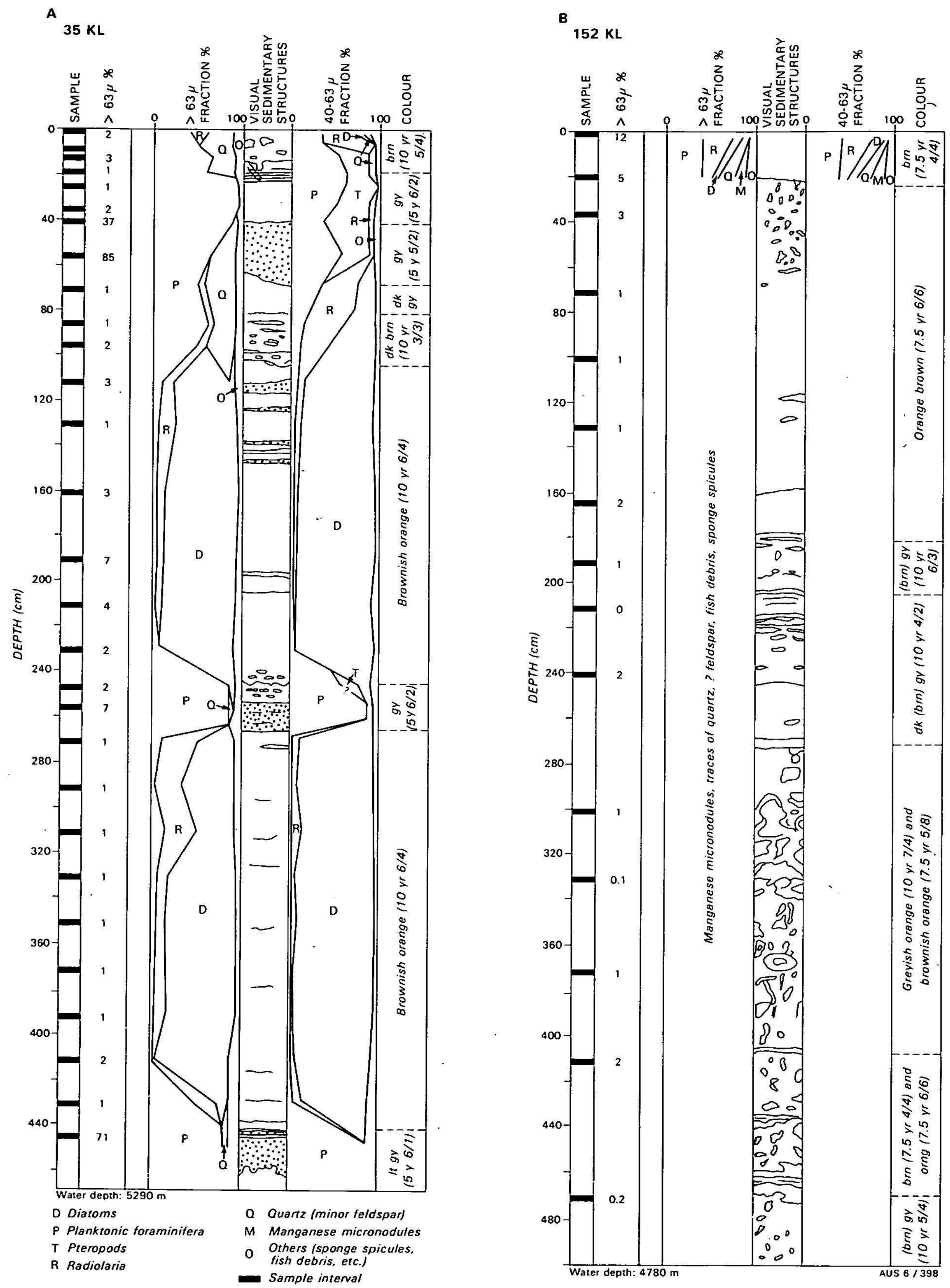

Figure 7. Typical cores, (A) Facies 4, (B) Facies 4A.

Figure 8. Radiograph (positive, natural scale) of core 35 KL (Facies 4).
418-434.5 cm, showing a fine-grained diatomaceous ooze overlying coarse-grained sand layers (turbidites). Black dots in the sand layers mark concentrations of quartz grains.

7A and 8), indicating input from turbidity currents. These layers commonly contain quartz, which is possibly derived from several sources, including windblown material (particularly in the case of iron-oxide stained grains), Mesozoic quartzose sandstones, which crop out on the northern margin of the Exmouth Plateau (von Stackelberg & others, 1980), and Tertiary and younger sands, which occur on the continental shelf (Quilty, 1974).

The sand and silt fractions of the sediments consist mainly of siliceous organisms (Fig. 7A). Planktonic foraminifera generally occur as a relatively minor component (except in the turbidite layers), and are generally partially dissolved.

Most of the Facies 4 sediment falls within the <40 μm fraction. This fraction typically consists of a mixture of siliceous organisms (radiolaria, diatoms, and silicoflagellates), coccoliths, and clays (mainly kaolinite and montmorillonite). Core 35 KL, which is fairly typical of the cores, contains fragments of the large diatom *Ethmodiscus rex*.

Sedimentary structures are dominated by burrows, which are particularly common in the sediments immediately overlying the turbidite layers. Sections between the turbidite layers commonly contain indications of a lenticular lamination, which may be the result of a fluctuating input of reworked material by turbid layer sedimentation.

Facies 4A

A subdivision of Facies 4 has been distinguished in two cores (152 KL and 158 KL) recovered in water depths of 4780 m and 4845 m, adjacent to the north-eastern margin of the Wallaby Plateau (Fig. 4).

A B

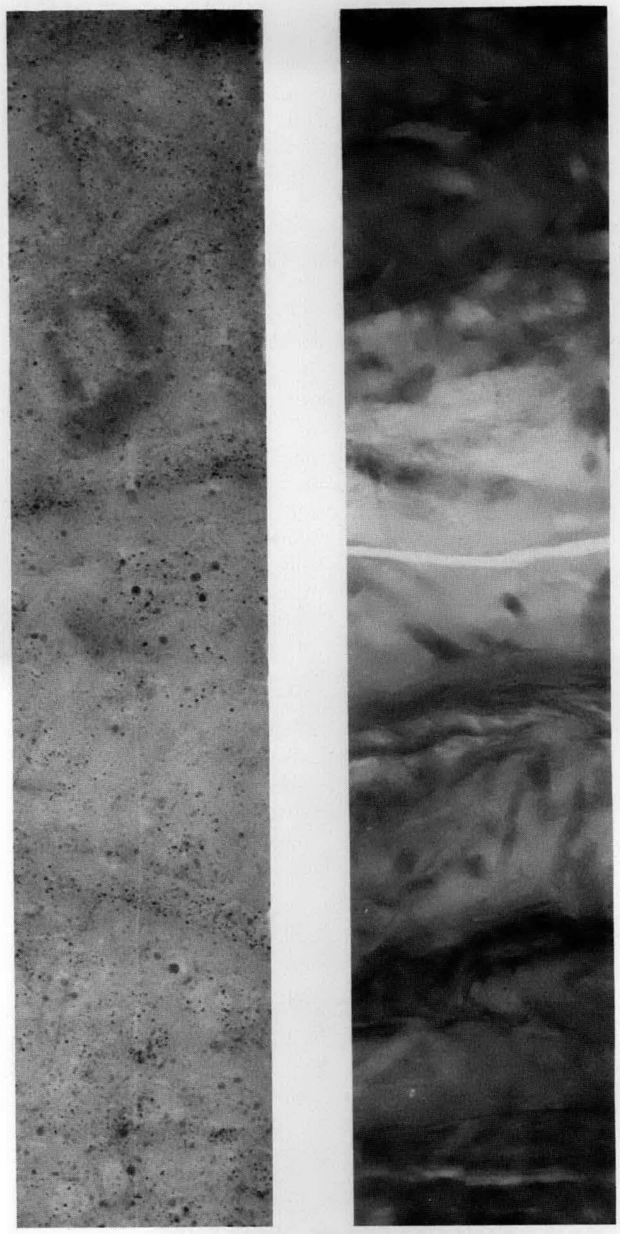

Figure 9. Radiographs (positive, natural scale) of core 152 KL (Facies 4A).
(A) 136-152.5 cm showing abundant manganese micronodules and zeolitic grains, and (B) 590-605.5 cm showing zones of intensive burrowing and of very fine lamination.

Although basically similar to the cores of Facies 4, they differ in that they contain manganese micronodules, have lower sand and silt contents, and generally lack turbidite sand layers, owing to their position away from the continental slope. The cores were 792 cm and 336 cm, respectively.

The sediments consist mainly of siliceous organisms, clays, and, in the upper 10-20 cm, partially dissolved planktonic foraminifera, and volcanic glass shards. Both cores appear to lie approximately at the CCD. The >40 μm fraction, which usually makes up less than 5 percent by weight of the sediment, typically consists of manganese micronodules, zeolitic grains, quartz (an aeolian component), fish debris, radiolarians, and sponge spicules (Fig. 7B).

The manganese micronodules range in size from 63 μm to 500 μm and display typical microbotryoidal textures. They are distributed fairly widely throughout the cores (Fig. 9A), a feature which suggests that the depositional environment has remained relatively unchanged for a considerable period. The presence of the nodules indicates low rates of sedimentation, oxygen-rich bottom currents, and the decay of organic substances in areas of abundant organic remains (Margolis, 1973; von Stackelberg, 1979b).

The sedimentary structures are characterised by intensive, mainly vertical, burrowing, alternating with a very fine lamination (Fig. 9B).

Comparison with the sediments of the Scott Plateau

A comparison between the cores taken in 1977 by R/V *Valdivia* on the Scott Plateau (approximately 600 km northeast of the Exmouth Plateau), in water depths of 3180 m to 3290 m (Hinz & others, 1978), and the cores taken in similar water depths on the Exmouth Plateau (Facies 3 cores of this report) shows that the Exmouth Plateau cores generally contain higher proportions of sand-size biogenic material. The northern Scott Plateau cores generally contain a significant terrigenous (mainly clay) fraction, which is probably derived from a number of sources: the Kimberley Block to the east, the Timor Sea to the northeast, and possibly, in the form of ash, from Indonesian volcanoes (Ninkovich & Donn, 1977). The Indonesian source is supported by the presence of a significant glass and opaque mineral fraction in the cores (von Stackelberg & others, 1978). In contrast, the Exmouth Plateau cores contain very little terrigenous material, possibly due to the plateau's relatively isolated position—being surrounded on three sides by abyssal plains and on its fourth side by the Montebello Trough, which effectively separates the plateau from the upper continental slope.

Acknowledgements

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References

Bé, A. W. H., & Hutson, W. H., 1977—Ecology of planktonic foraminifera and biogeographic patterns of life and fossil assemblages in the Indian Ocean. *Micropaleontology*, 23, 369-414.

Berger, W. H., 1977—Deep-sea carbonate and the deglaciation preservation spike in pteropods and foraminifera. *Nature*, 269, 301-4.

Cook, P. J., Veevers, J. J., Heirtzler, J. R., & Cameron, P. J., 1978—The sediments of the Argo Abyssal Plain and adjacent areas, northeast Indian Ocean. *BMR Journal of Australian Geology & Geophysics*, 3, 113-24.

Exon, N. F., 1979—Preliminary deep sea sampling results, *R. V. Sonne* geological cruises off Western Australia. *Bureau of Mineral Resources Australia, Record 1979/26* (unpublished).

Falvey, D. A., & Veevers, J. J., 1974—Physiography of the Exmouth and Scott Plateaus, Western Australia, and adjacent northeast Wharton Basins. *Marine Geology*, 17, 21-59.

Hinz, K., Beiersdorf, H., Exon, N. F., Roeser, H. A., Stagg, H. M. J. & von Stackelberg, U., 1978—Geoscientific investigations from the Scott Plateau off northwest Australia to the Java Trench. *BMR Journal of Australian Geology & Geophysics*, 3, 319-40.

Jones, H. A., 1971—Late Cenozoic sedimentary forms on the northwest Australian continental shelf. *Marine Geology*, 10, M20-6

Larson, R. L., Mutter, J. C., Diebold, J. B., Carpenter, G. B., & Symonds, P. A., 1979—Cuvier Basin: a product of ocean crust formation by Early Cretaceous rifting off Western Australia. *Earth and Planetary Science Letters*, 45, 105-14.

Margolis, St. V., 1973—Manganese deposits encountered during DSDP Leg 29 in sub-Antarctic waters. *In* Morgenstein, M. (editor), The Origin and Distribution of Manganese Nodules in the Pacific and Prospects for Exploration. *Proceedings of symposium organised by Valdivia Manganese Exploration Group and Hawaii Institute of Geophysics, Honolulu, 1973.*

McIntyre, A., & McIntyre, R., 1971—Coccolith concentrations and differential solution in ocean sediments. *In* Funnell, B. M., & Riedel, W. R. (editors), The Micropalaeontology of Oceans. *Cambridge University Press.* 235-61.

Ninkovich, D., & Donn, W. L., 1977—Cainozoic explosive volcanism related to east and southeast Asian arcs. *In* Talwani, M., & Pitman, W. C. (editors), Island Arcs, Deep Sea Trenches and Back-Arc Basins. *American Geophysical Union, Maurice Ewing Series*, 1, 337-47.

Quilty, P. G., 1974—Tertiary stratigraphy of Western Australia. *Journal of the Geological Society of Australia*, 21, 301-18.

Quilty, P. G., 1980—Tertiary Foraminiferida and stratigraphy, northern Exmouth Plateau, Western Australia. *BMR Journal of Australian Geology and Geophysics*, 5, 141-9.

Veevers, J. J., & Cotterill, D., 1978—Western margin of Australia: evolution of a rifted arch system. *Geological Society of America Bulletin*, 89, 337-55.

von Stackelberg, U., 1979a—Geologische Untersuchungen des NW—australischen Kontinentalrandes Forschungsfahrt SO-8 mit M.S. 'Sonne' vom 3.1. bis 15.2.1979. *Report of the Bundesanstalt für Geowissenschaften und Rohstoffe, Hannover* (unpublished).

von Stackelberg, U., 1979b—Sedimentation, hiatuses, and development of manganese nodules: Valdivia Site VA-13/2, northern central Pacific. *In* Bischoff, J. L., & Piper, D. Z. (editors), Marine Geology and Oceanography of the Pacific Manganese Nodule Province, *Plenum Publishing Corporation, New York.*

von Stackelberg, U., Belford, D. J., Burger, D., & others, 1978—Notes on the sedimentology and palaeontology of cores and dredged sediments obtained from the Scott Plateau and Java Trench by R. V. *Valdivia*, 1977. *Bureau of Mineral Resources, Australia, Record* 1978/50 (unpublished).

von Stackelberg, U., Exon, N. F., von Rad, U., Quilty, P., Shafik, S., Beiersdorf, H., Seibertz, E., & Veevers, J. J., 1980—Geology of the Exmouth and Wallaby Plateaus off northwest Australia: sampling of seismic sequences. *BMR Journal of Australian Geology & Geophysics*, 5, 113-40.

Willcox, J. B., & Exon, N. F., 1976—The regional geology of the Exmouth Plateau. *The APEA Journal*, 16, 1-11.

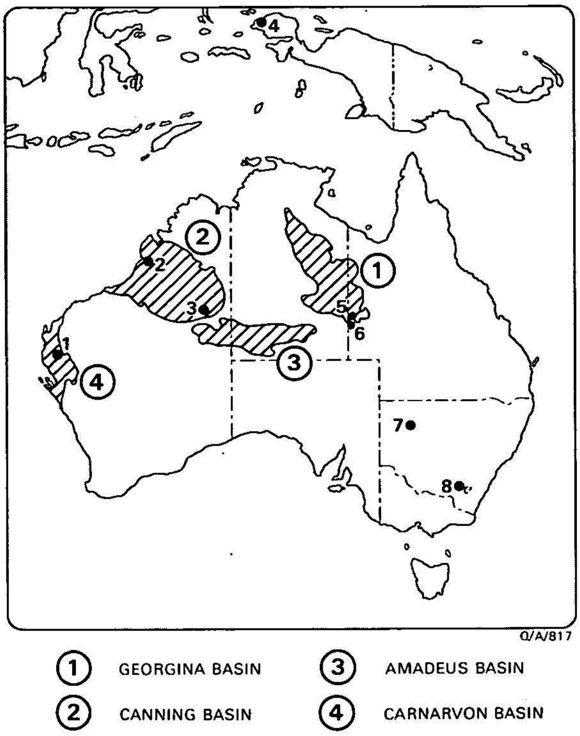

Figure 1. Distribution of reported occurrences of thelodont remains from the Australian Plate.
Localities, at present all Devonian in age, are numbered as follows: 1—Gneudna Formation, Williambury Station, Carnarvon Basin; 2—Thangoo Calcarenite, Roebuck Bay No. 1 well, Broome Platform, Canning Basin (based on a mis-identification, see page 65); 3—Tandalgoo Red Beds, Wilson Cliffs No. 1 well, Kidson Sub-basin, Canning Basin; 4—Kemum Formation (sample no. 78CP103/3), 80 km east of Sorong, Kepala Burung, Irian Jaya; 5—Cravens Peak Beds, shot holes west of Toko Range, Toko Syncline, Georgina Basin; 6—Cravens Peak Beds, southern Toomba Range, Toko Syncline, Georgina Basin; 7—Lower part of the Mulga Downs Group, Mount Jack Station, Cobar Basin; 8—Hatchery Creek Conglomerate, Taemas/Wee Jasper region, New South Wales.

scale has been referred to the Early Devonian (Dittonian) genus *Apalolepis* by Young (personal communication, 1979).

In this paper we describe the Early Devonian thelodont scales and associated fauna from all the Toko Syncline localities, redefine the Cravens Peak Beds and discuss their depositional environment and palaeogeographic implications, and consider the significance of the distribution of Australian Devonian thelodonts on a global scale.

Most of the figured specimens described in this paper are deposited in the Commonwealth Palaeontological Collection (prefix CPC), housed in the Bureau of Mineral Resources, Geology and Geophysics, Canberra. Other figured specimens are housed in the palaeontological collections of the Australian Museum, Sydney (prefix AM), and the British Museum (Natural History), London (prefix P).

The Cravens Peak Beds

Previous investigations

The Cravens Peak Beds were defined by Reynolds (*in* Smith, 1965) to include the sandstone and conglomerate of 'Silurian-Devonian' age that unconformably overlie the Middle Ordovician Mithaka Formation within the core of the Toko Syncline. As originally mapped, the beds comprised Pritchard's (1960) unit Om-11 in the Toko Range (Reynolds, 1965, 1968), an unnamed sandstone in the Toomba Range ('Middle Ordovician Undifferentiated' of Smith, 1963), and a conglomerate and sandstone sequence near Mithaka Waterhole (Smith, 1965).

Recent field mapping (Draper, 1980) has demonstrated that the Cravens Peak Beds, as discussed in Smith (1972) and shown on his 1:500 000 geological map of the Georgina Basin (Sheet 4), are composed of three separate units: (1) the Middle Ordovician Ethabuka Sandstone (the lowest part of Pritchard's unit Om-11); (2) the Devonian Cravens Peak Beds (*sensu stricto*); and (3) a post-Devonian valley-fill conglomeratic deposit.

The Ethabuka Sandstone (Draper, 1980) crops out as a low ridge in the southwestern corner of the Glenormiston 1:250 000 Sheet area (Reynolds, 1965), and

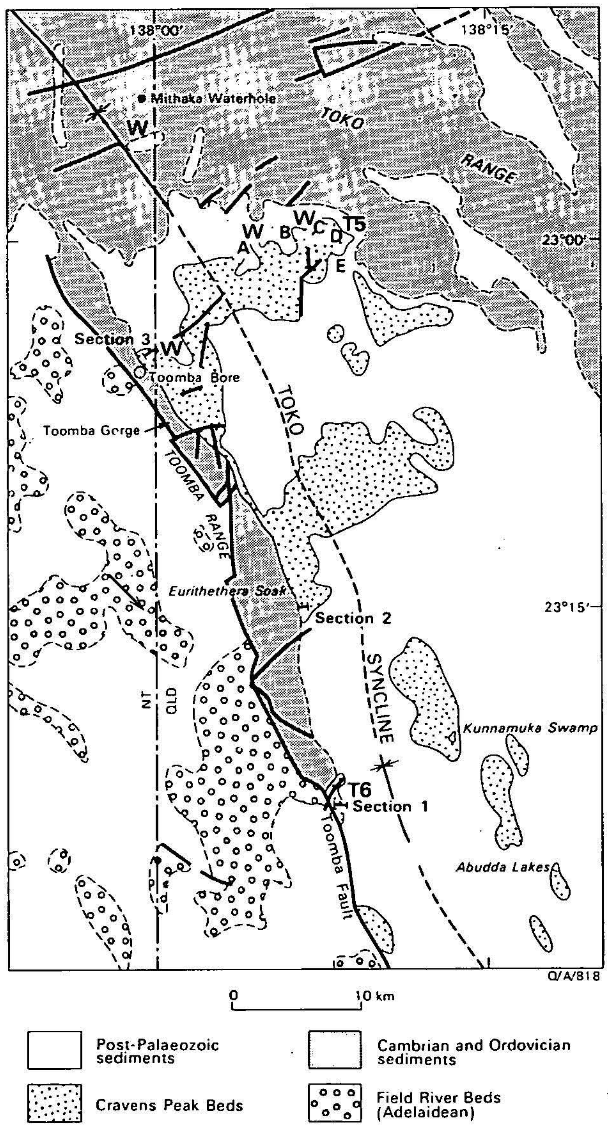

Figure 2. General geology of Toko Syncline showing positions of measured sections of the Cravens Peak Beds and the Devonian vertebrate localities with thelodonts (localities 5 and 6 of Fig. 1), and the Wuttagoonaspis fauna.
W—*Wuttagoonaspis* fauna; T—*Turinia* fauna; 5—Shot-point localities west of Toko Range: A—SP801, B—SP839, C—SP799, D—SP798, E—SP813; 6—Toomba Range locality GEO 65/28 (74710577).

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Early Devonian thelodonts (Agnatha) from the Toko Syncline, western Queensland, and a review of other Australian discoveries

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Thelodont scales recovered from the basal (calcareous) unit of the Cravens Peak Beds in the Georgina Basin, are referable to *Turinia australiensis* Gross, 1971, *T.* cf. *pagei* (Powrie, 1870), and *Gampsolepis* ? sp. undet. The thelodonts probably lived in a marginal marine environment (as evidenced from the associated ostracods and eridostracans) at about the same time as the placoderm *Wuttagoonaspis* sp. lived in the freshwater bodies, now represented by the sandstone and conglomerate facies of the Cravens Peak Beds.

Scales of *Turinia australiensis* Gross, 1971, associated with *Wuttagoonaspis* plates, from the lower part of the Mulga Downs Group in the Cobar/Wilcannia area of New South Wales, are at least as young as late Early Devonian (Emsian), because they post-date the Pragian age of the underlying Amphitheatre Group. By correlation, those parts of the Cravens Peak Beds (Georgina Basin) and the Tandalgoo Red Beds (Canning Basin) that also contain *Turinia australiensis* are approximately coeval.

After reaching Australia in Early Devonian time, the *Turinia* fauna began an adaptive radiation to give apparently younger (Middle Devonian) stocks that have survived longer in the Australian region than elsewhere, as the youngest known scales come from the Gneudna Formation (latest Givetian-earliest Frasnian) in the Carnarvon Basin, Western Australia.

Introduction

Thelodont remains were first recognised in Australia in 1963, when one of us (PJJ) noted (*in* Reynolds & Pritchard, 1964) the presence of coelolepidid (synonym thelodontid) scales in the lower part of the Cravens Peak Beds in the western Queensland part of the Georgina Basin (Johnstone & others, 1967; Gilbert-Tomlinson, 1968; Smith, 1972). This discovery, in the area east of the axis of the Toko Syncline and west of the Toko Range, was also the first recorded of agnathan remains from the Southern Hemisphere (Figs. 1 and 2; locality 5).

The original material was not formally described because the taxonomy of the entire Thelodonti at that time was poorly understood and in much need of revision. A general Late Silurian-Early Devonian age was suggested for the lower part of the Cravens Peak Beds (Jones *in* Reynolds & Pritchard, 1964; Johnstone & others, 1967), which corresponded to the total range of the Thelodonti known at the time. It was not until Gross (1967) published his important monograph on thelodont scales, based entirely on European material, that some precision was added to the classification, and the potential of these microfossils for biostratigraphic zonation was realised (for example, see papers by Karatajute-Talimaa, 1968, 1970; Mark-Kurik, 1969; Mark-Kurik & Noppel, 1970; Moskalenko, 1968; Ørvig, 1969a, 1969b, 1969c; Ritchie, 1968; Turner, 1973).

Soon after this, Gross (1971b) described scales of a thelodont; *Turinia australiensis*, of presumed Early Devonian age from Western Australia (Fig. 1; locality 3), and those from the Toko Range were tentatively identified by one of us (PJJ) as belonging to *T. pagei* (Powrie, 1870) (*in* Turner, 1973, p. 573). Then, in 1975, a second locality with the abundant thelodont scales described here was found in the Toko Syncline, on its western flank, in the southern part of the Toomba Range (Fig. 1, locality 6).

The *Turinia pagei* fauna is now accepted as a good indicator of the start of the Dittonian Stage of the Lower Old Red Sandstone in Europe (Turner, 1973; Karatajute-Talimaa, 1978), and, notwithstanding the problems of the definition of the Silurian/Devonian boundary faced by British stratigraphers (cf. Cocks & others, 1971; Westoll & others, 1971; Lawson, 1971; Halstead, 1971; and Turner, 1971), indicates an Early Devonian age. Devonian thelodont remains are now being reported from Iran (Blieck & Goujet, 1978; Turner & Janvier, 1979), Thailand (Blieck & Goujet, 1978), Indonesia, and more localities in Australia (Fig. 1).

In Western Australia they are known from the Canning and Carnarvon Basins. *Turinia australiensis* Gross, 1971, the first representative of the Thelodonti to be described from the Southern Hemisphere, is the only authenticated representative of the group in the Canning Basin. We regard the report of a thelodont scale from this basin by McTavish & Legg (1976), to be based on a misidentification (see page 65). Recently, Turner & Dring (1981) have described thelodont scales, which they refer to *Australolepis seddoni* gen. et. sp. nov., from the Gneudna Formation of the Carnarvon Basin (Fig. 1, locality 1). On conodont evidence (Seddon, 1969; 1970; Roberts & others, 1972), this taxon is of late Givetian-early Frasnian age, and the youngest known representative of the Thelodonti.

In eastern Australia, thelodont remains have been found in the lower part (Emsian-Eifelian) of the Mulga Downs Group at Mount Jack in the Cobar region of western New South Wales by Ritchie (personal communication, 1980) (Fig. 1, locality 7), and in the late(?) Eifelian part of the Hatchery Creek Conglomerate (Middle Devonian) of the Wee Jasper area, near Canberra (Young & Gorter, 1981; Fig. 1, locality 8). Other occurrences of Middle Devonian thelodont scales have been reported from New South Wales by Turner & Janvier (1979, p. 892).

Recently, thelodont remains have been found in Indonesia, in the Kepala Burung region of Irian Jaya (Fig. 1, locality 4), where a single small thelodont

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in the northern part of the Toomba Range in the northeastern corner of the Hay River 1:250 000 Sheet area (Smith, 1963), where it forms the lower 35 m of the original reference section of the Cravens Peak Beds (Smith, 1972, table 31). The post-Devonian valley-fill conglomeratic unit unconformably overlies the Ordovician Toko Group, and cuts across the major structure of the area.

The Cravens Peak Beds (*sensu stricto*) consist of a basal limestone and calcareous siltstone with Devonian thelodont scales, herein referred to as the *basal (calcareous) unit,* and a conglomerate and sandstone sequence (the dominant rock types) with the Devonian placoderm *Wuttagoonaspis* Ritchie, 1973. These two units correspond to the 'lower Cravens Peak Beds' and the 'upper Cravens Peak Beds' of previous authors (for example, Johnstone & others, 1967; Gilbert-Tomlinson, 1968), who have suggested that the apparent disparity between the ages of these units implies a possible hiatus. An unconformable relationship between these units was supported (Draper, 1976), until later mapping (by JJD, 1977) showed that their mutual contact is transitional. The basal (calcareous) unit is known from only two areas within the Toko Syncline—one on its northeastern limb, just west of the Toko Range, and the other on its southwestern limb, in the southern part of the Toomba Range. The spatial distribution of the Cravens Peak Beds, as clarified in this paper (Appendix), is shown in Figure 2, and three composite sections, showing variation along the Toomba Range, are shown in Figure 3.

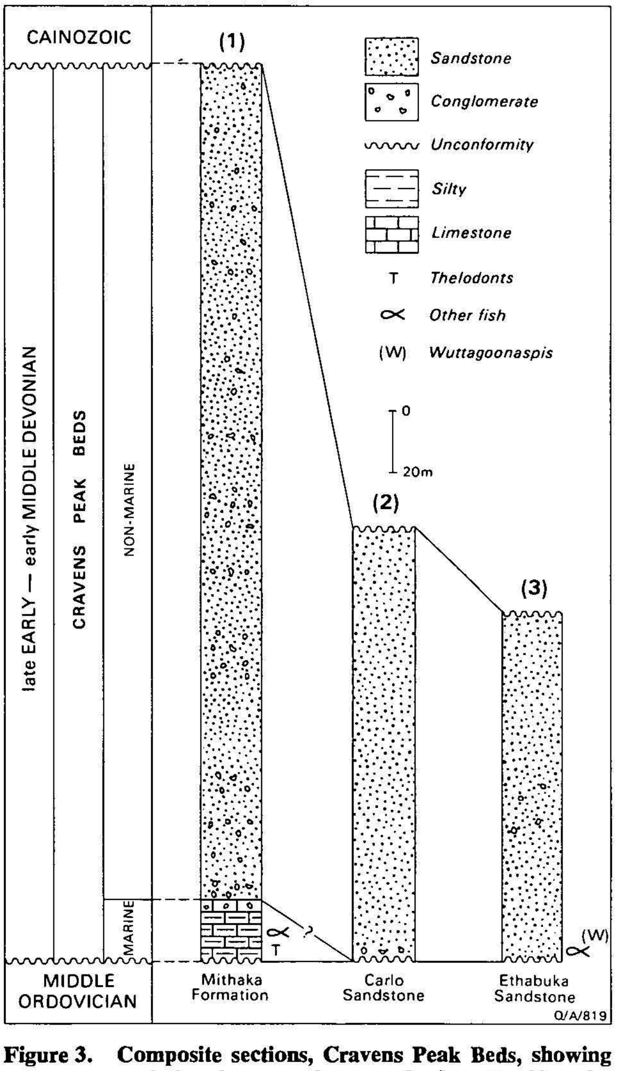

Figure 3. Composite sections, Cravens Peak Beds, showing variation from south to north along the Toomba Range.
1—southern part of the Toomba Range, about 16 km SSE of Eurithethera Soak; 2—near Eurithethera Soak; 3—100 metres E of Toomba Bore.

Material and localities

The original (1963) collection of thelodont scales was recovered from shot-hole samples (prefix SP) from a seismic reflection survey (conducted by Austral Geoprospectors Pty Ltd, on behalf of Phillips Petroleum Co. & Sunray Mid-Continent Co., in 1960) over the eastern limb of the Toko Syncline, in the areas covered by the southwestern part of the Glenormiston 1:250 000 geological sheet (Reynolds, 1965), viz., SP 798, 799, 801 and 839, and the northwestern part of the Mount Whelan 1:250 000 geological sheet (Reynolds, 1966), viz., SP 813. The approximate geographical co-ordinates of the shot-holes are SP 801 — 22°59′49″S 138°04′16″E, SP 839 — 22°59′25″S 138°05′46″E, SP 799 — 22°59′46″S 138°06′06″E, SP 798 — 22°59′47″S 138°06′48″E, and SP 813 — 23°00′31″S 138°08′30″E. The rocks sampled were siltstones and calcareous siltstones, which readily disintegrated in boiling water, yielding many thelodont scales, a few acanthodian scales, ostracods, and eridostracans. Samples SP 813 and 839 also yielded a few Middle Ordovician conodonts (simple drepanodontiform and cordylodontiform elements), which are similar to those found in the Mithaka Formation from other shot holes made during the same seismic survey, and this association is attributed to contamination.

The 1975 collection of thelodont scales was recovered from the insoluble residue of a recrystallised limestone sample, and abundant eridostracans and a few ostracods were extracted before it was dissolved in monochloracetic acid. The sample was collected from an isolated outcrop in the southern part of the Toomba Range—BMR locality GEO 65/28 (74710-577))—about 42 km south of the shot-point localities, on the western flank of the Toko Syncline (Mount Whelan 1:250 000 geological sheet) at approximately 23°23′47″S; 138°08′10″E. Here some 20 m of the basal (calcareous) unit is exposed, representing the only limestone of Devonian age known from central Australia (Fig. 4). The lower 6 metres consists of laminated calcareous siltstone terminated by a distinct stromatolite bed, followed by a covered interval of 4 metres. The upper 7 metres consists of minor conglomerate, calcareous siltstone, recrystallised oncolitic and phosphatic limestone, and calcareous sandstone, terminated by a covered interval of 3 metres.

The identified taxa (thelodont and acanthodian scales, eridostracan, and ostracods) and their distribution within the shot-hole localities and the Toomba Range locality are listed in Figure 5.

Depositional environment

The basal (calcareous) unit at the Toomba Range locality (GEO 65/28) is interpreted as representing an initially shallowing sequence passing from shallow subtidal conditions to beach. A slight deepening of water resulted in the development of offshore bars (Figure 4). Stromatolites, a few thelodont scales, and other fish fragments are present in the shallow-water, nearshore environment. Oncolites, numerous thelodont and acanthodian remains, and phosphatic pellets are

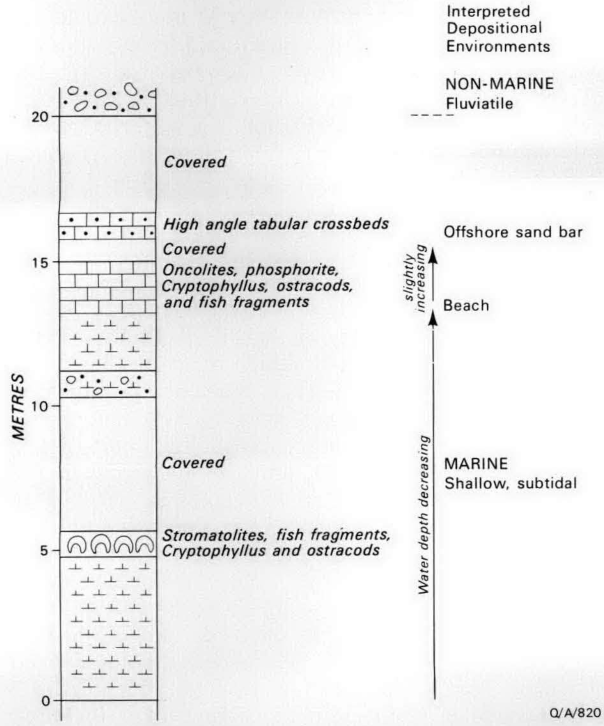

Figure 4. Basal (calcareous) unit of Cravens Peak Beds 100 metres N of section 1 in southern part of the Toomba Range.
Included within section 1 in Figure 3.

SPECIES \ LOCALITY	TOKO RANGE SP 801	SP 839	SP 799	SP 798	SP 813	GEO 65/28
Turinia cf. *pagei* (Powrie, 1870)						a
T. australiensis Gross, 1971	r	r	c	c	r	r
Gampsolepis ? sp. indet						r
Gomphonchus ? sp. indet						r
Nostolepis sp.			r			
Healdianella inconstans Polenova, 1974?			r			r
Baschkirina ? sp		r				r
Cryptophyllus sp.		r	r			a

Q/A/821

Figure 5. Determinations of fauna present in the basal (calcareous) part of the Cravens Peak Beds, Toko Syncline.
Frequency of specimens indicated as abundant (a), common (c), and rare (r).

present in the cross-stratified coarser-grained sediments. Higher energy conditions are confirmed by the petrography of the recrystallised limestone, which consists of skeletal grainstones in which the skeletal material is fragmented, and oncolitic limestone with oncolites forming around skeletal fragments and various mineral grains, and algal boundstones.

The microcrustacean evidence is compatible with the postulated marine depositional conditions. *Cryptophyllus,* from its association with marine fossils in other Australian occurrences (Jones 1962, 1968), is thought to be marine, and the ostracods tentatively referred to the genera *Healdianella* and *Baschkirina* probably indicate a marine or marginal marine environment. The presence of thelodont scales, however, provides conflicting evidence. Articulated specimens of *Turinia pagei* from Turin Hill, Scotland (the holotype) and from Mitcheldean in the Welsh Borderland, are thought to have lived in freshwater (Allen & others, 1968), but the species may have retained a marine phase during its life history (Turner, 1973; Halstead & Turner, 1973), and Goujet & Blieck (1977, 1979) have reported scales of *T. pagei* from marine sediments in Spitsbergen, Podolia, and northern France.

The overlying sandstone and conglomerate of the Toomba Range (Figure 3) were probably deposited under braided-stream conditions, similar to the environment figured by Allen (1965, figure 35B). Features in common with Allen's model are lack of argillaceous sediment, lenticular nature of the bedding, and the general lack of biologic activity. The placoderm fish remains may occur as isolated plates or fragments, but at some localities skull and trunk armour plates from a number of individuals are closely associated, suggesting that post-mortem transportation was minimal. At these localities the remains indicate fish of considerable size, which presumably inhabited permanent bodies of freshwater (G. C. Young, personal communication).

Thelodont fauna

Three thelodont species, as determined from their scales, are present in the Cravens Peak Beds: *Turinia* cf. *pagei* (Powrie, 1870), *T. australiensis* Gross, 1971, and *Gampsolepsis?* sp. undet.

Turinia cf. **pagei** (Powrie, 1870) (Figs. 6-8)

About 60 scales were recovered from the Toomba Range sample (GEO 65/28; Figure 5), of which 15 are figured here as *Turinia* cf. *pagei.* Most of them fall within the range of variation exhibited by the scales of *Turinia pagei* (Powrie, 1870), which have been previously described under various names (see Turner, 1976; Karatajute-Talimaa, 1978, for synonymies). Because only two or three articulated specimens are known, and none is well-preserved—the scales being water-worn, if preserved at all—the range of scale variation in this species is determined primarily from the uniformity of histological structure and a gradation of scale ornament (Gross, 1967; Ørvig, 1969a; Karatajute-Talimaa, 1964, 1978; Turner, 1973). Karatajute-Talimaa (1978) has subdivided *Turinia pagei* into *T. pagei* (*sensu stricto*) and *T. polita,* but some scales referred to the latter, we regard as *T.*

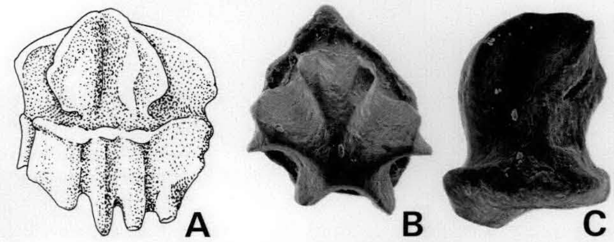

Figure 6. *Turinia* **cf.** *pagei* **(Powrie, 1870)—Head scales from Toomba Range locality GEO 65/28.**
A—CPC 20079/11, lateral view showing 'rootlets' from the basal rim, x 30; B—CPC 20079/18, crown view, x 30; C—CPC 20079/19, lateral view, x 35.

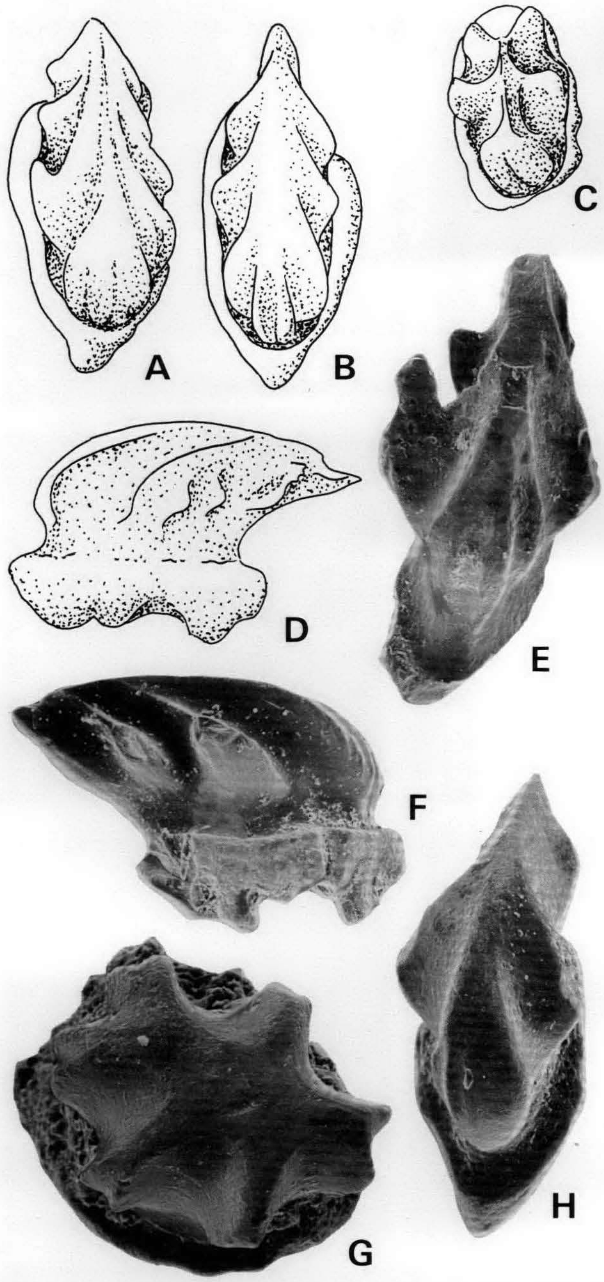

Figure 7. ***Turinia* cf. *pagei* (Powrie, 1870)—Transitional scales from Toomba Range locality GEO 65/28.** A—CPC 20079/1, crown view, x 30; B—CPC 20079/22, crown view, x 30; C—CPC 20079/5, crown view, x 50; D—CPC 20079/2, lateral view, x 30; E—CPC 20079/3, crown view showing lateral winglets, x 50; F—CPC 20079/10, lateral view showing rootlets at base, x 60; G—CPC 20079/16, crown view, x 70; H—CPC 20079/9, crown view, x 50.

pagei, and others as *T. australiensis*. Scales of *T. pagei* and *T. australiensis* are found together in the Dittonian of the Welsh Borderland and Podolia (Karatajute-Talimaa, 1964; Turner, 1973).

Many of the scales in the Toomba Range sample are semi-transparent, and, with the addition of a little anise oil, the internal structure can be seen. They show little difference in histological structure from the *Turinia pagei* scales described by Gross (1967). In general morphology, however, some of the scales have more ribs and 'side-lappets' than *T. pagei,* and provide sufficient reason to refer the material to *Turinia* cf. *pagei*. They range in size from 0.55 mm to 1.62 mm.

Head scales (Figs. 6a-c) are rare in the Toomba Range sample, as is often the case in samples from the Welsh Borderland. These scales have rounded crowns with a crenulated crown rim, or more-incised ridges rising to a central point on the crown. The bases are rounded, either with a large wide open pulp cavity or one to three pulp openings, typical of the genus *Turinia*. Occasionally, there are small 'rootlets' extending from the basal rim (Fig. 6a).

Transitional scales (Figs. 7a-h) come from the region between head and trunk and may cover a large area because, on the basis of articulated specimens, *Turinia* was a large thelodont, growing to about 35 cm long. Ørvig (1969a, fig. 2A) figured some typical scales of this type. They are large, elliptical scales with a crenulated crown edge. The crown may come to a posterior point and have three pairs of lateral ridges

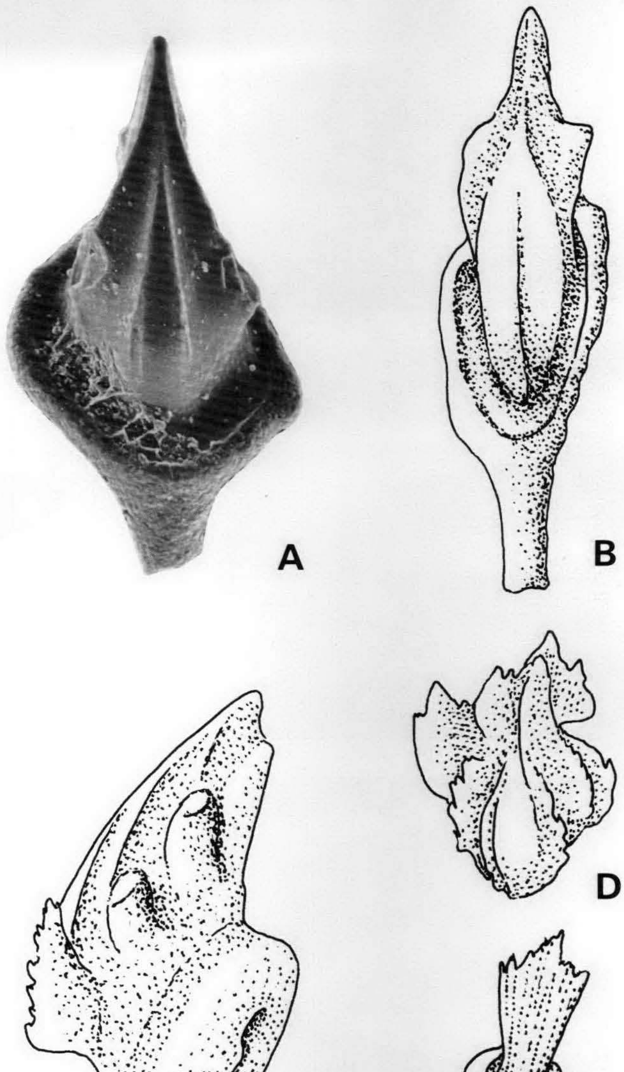

Figure 8. ***Turinia* cf. *pagei* (Powrie, 1870)—Body and specialised scales from Toomba Range locality GEO 65/28.**
A—CPC 20079/14, crown view, x 90; B—CPC 20079/6, crown view, x 90; C—CPC 20079/20, lateral view, x 50; D—CPC 20079/4, crown view of a small special scale, which may belong to *T. australiensis* Gross, 1971, x 50; E—CPC 20079/7, crown view of a small special scale, x 50.

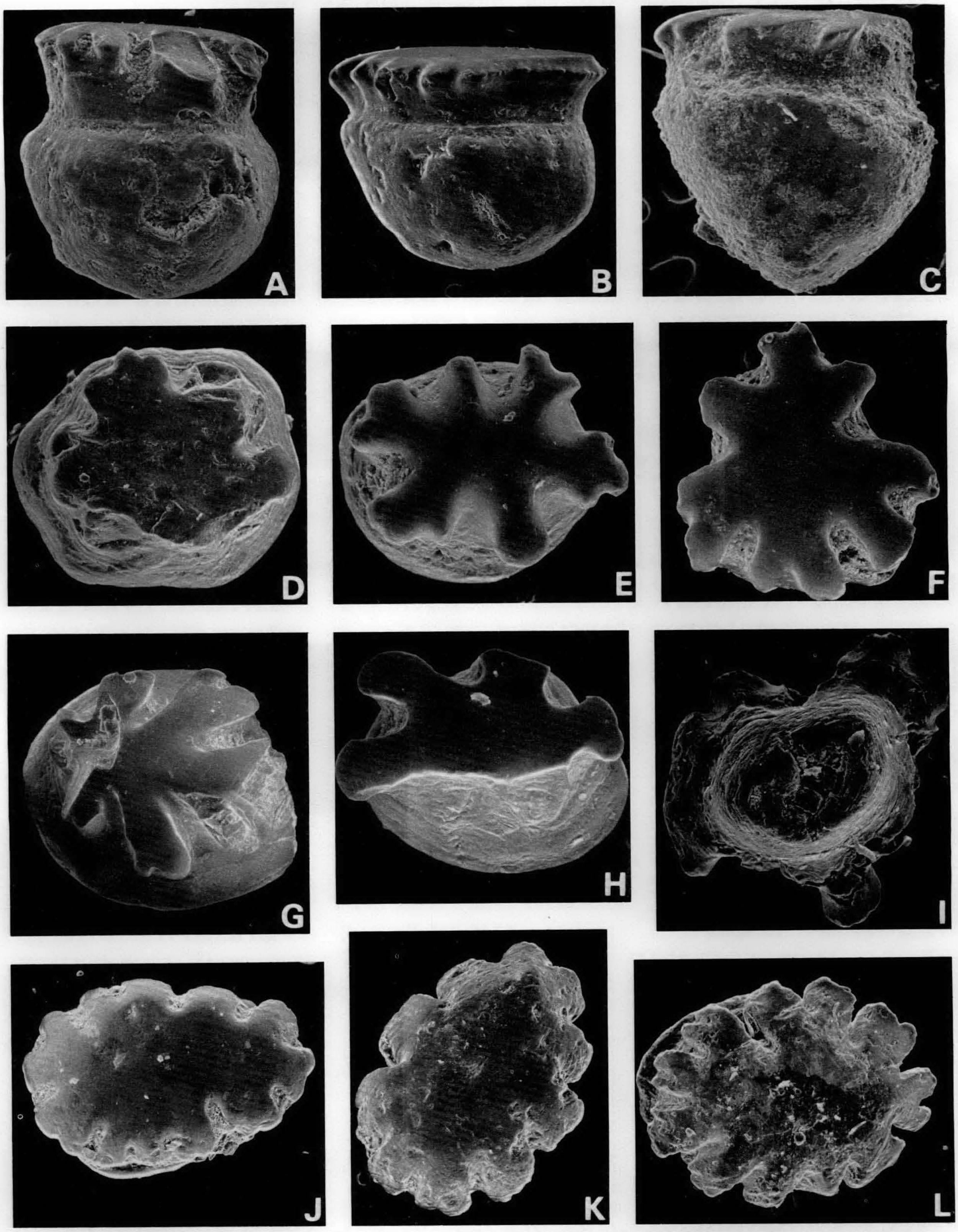

Figure 9. *Turinia australiensis* **Gross, 1971—Head (A-I) and head-transitional (J-L) scales from the Toko Range shot hole localities.**
A—CPC 20081/6, lateral view showing deep base, x 75, locality SP 798; B—CPC 20084/4, lateral view, x 55, locality SP 813; C—CPC 20084/6, lateral view, x 70, locality SP 813; D—CPC 20081/5, crown view, x 60, locality SP 798; E—CPC 20083/5, crown view, x 95, locality SP 801; F—CPC 20083/2, crown view, x 90, locality SP 801; G—CPC 20081/9, crown view, x 65, locality SP 798; H—CPC 20084/3, crown view, x 150, locality SP 813; I—CPC 20083/7, basal view, x 90, locality SP 801; J—CPC 20084/2, crown view, x 65, locality SP 813; K—CPC 20085/2, crown view, x 75, locality SP 839; L—CPC 20085/1, crown view, x 60, locality SP 839.

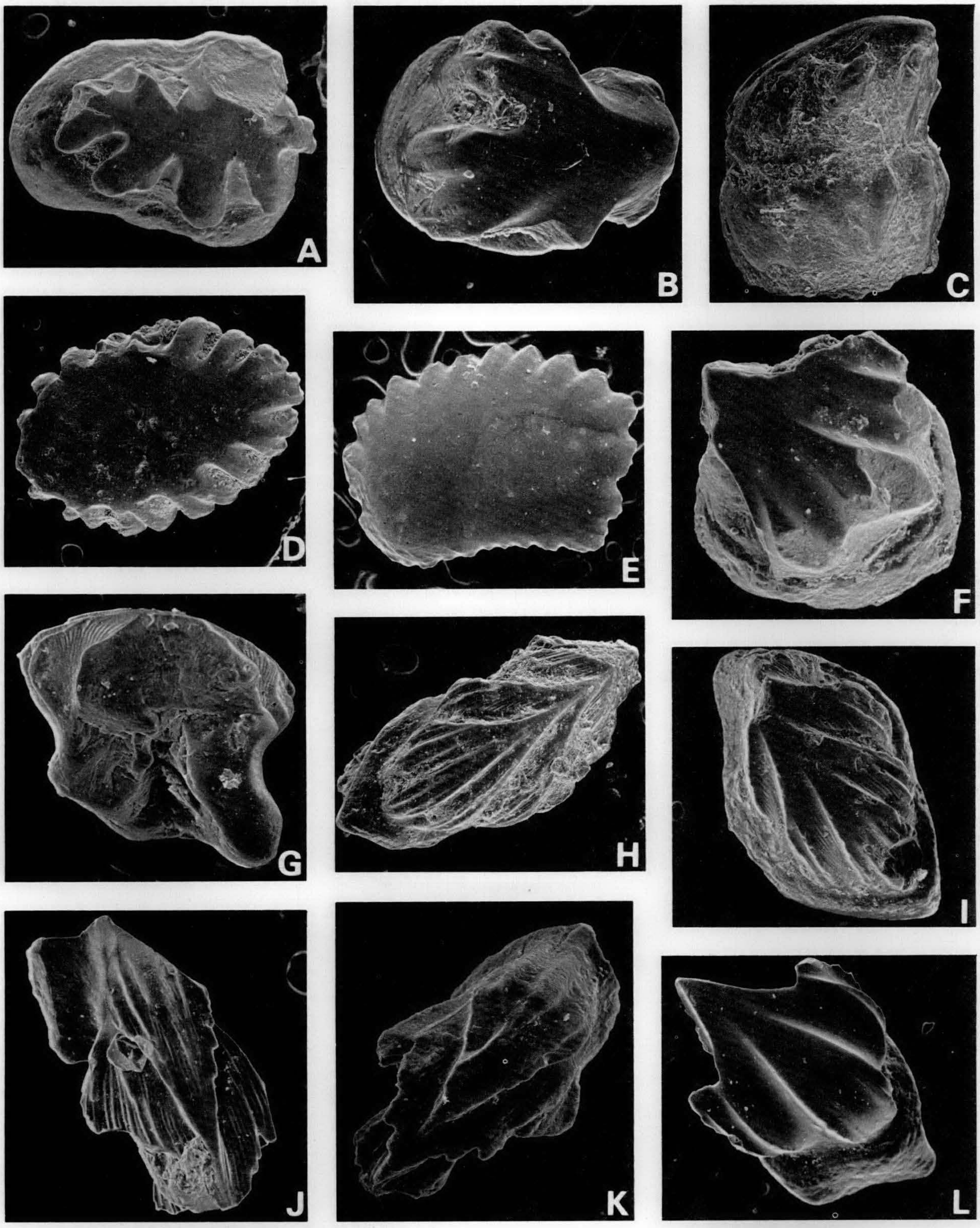

Figure 10. *Turinia australiensis* **Gross, 1971—Transitional (A-E) and other (F-I) scales from the Toko Range shot hole localities; body (J, K) and special (L) scales from the Toomba Range locality GEO 65/28.**
A—CPC 20071/8, crown view, x 60, locality SP 798; B—CPC 20082/9, crown view, x 90, locality SP 799; C—CPC 20082/3, lateral view, x 60, locality SP 799; D—CPC 20083/3, crown view, x 30, locality SP 801; E—CPC 20083/6, crown view, x 50, locality SP 801; F—CPC 20083/4, crown view of a special scale, x 90, locality SP 801; G—CPC 20084/5, lateral view of a fragmentary transitional/body scale, x 80, locality SP 813; H—CPC 20082/4, crown view of a body scale, x 55, locality SP 799; I—CPC 20083/1, crown view of a body or transitional scale, x 60, locality SP 801; J—CPC 20080/1, crown view, x 75, K—CPC 20080/3, crown view, x 100, L—CPC 20080/2, crown view, x 50.

with an anterior bifurcated ridge curving down to the neck. The extension of the lateral ridges into prongs or winglets is well displayed on some of the Australian scales (Fig. 7e). These scales are probably more posterior transitional scales showing a gradation to body scales on which this feature is more common.

The body scales (Figs. 8a-c) are generally large and elongated, navicular in shape with a flat central crown pointed posteriorly, often with one to three pairs of lateral flaps or winglets. The crown may rise at quite a steep angle from the base and be finely ribbed. The scales in the Toomba Range sample are often well preserved and show a complex development of lateral wings and ridges. The base is often wider than the crown, with a deep moat-like neck and an anterior spur. Several scales may be special scales that covered particular areas on the body, and these will not fit easily into a simple description until better preserved, articulated material is found (Figs. 8d, e).

Turinia australiensis Gross, 1971 (Figs. 9-12)

About 100 scales were recovered from the Toko Range shot hole localities (Fig. 5), of which 32 are figured here as *Turinia australiensis* Gross, 1971. Other figured specimens referred to this species are five scales (CPC 20080/1-5) recovered from the Toomba Range locality GEO 65/28.

Gross (1971b, p. 98, fig. 1, pl. 12, figs. 1-7) referred the Western Australian scales to the genus *Turinia,* on general morphology and histology, and differentiated them from those of *T. pagei* on their smaller size, simpler form, and the occasional presence of minor crossfluting on the ribs. Because this material lacked the typical body scales of *T. pagei* (that is, with a wedge-shaped unsculptured plain crown; cf. Gross, 1967, pl. 7, figs. 1-4) *T. australiensis* was defined on the basis of the head and transitional scales, without the knowledge of the body scales. In some instances, it is difficult to decide whether a scale belongs to *pagei* or *australiensis,* and certainly some of the Toko Range scales are large, and, at 1.6 mm long, extend into the size range of *pagei;* however, they still seem simpler than *T. pagei* scales. Gross surmised that *T. australiensis* is probably a separate species, as very small or finely ribbed scales have not been observed on any of the articulated specimens. It is possible that the scales of *T. australiensis* may be special scales of *T. pagei*

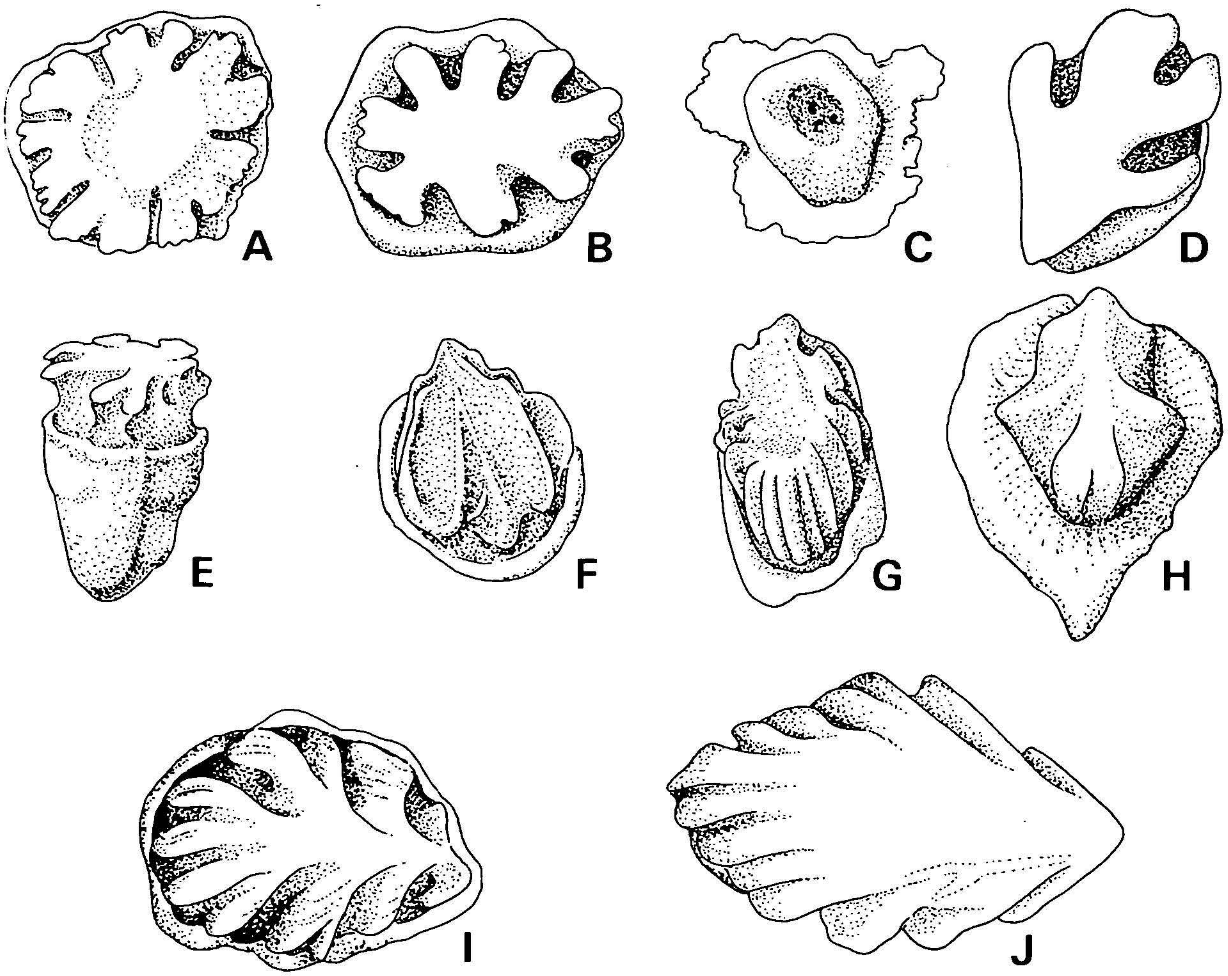

Figure 11. *Turinia australiensis* **Gross, 1971—Head (A-E), transitional (I, J), transitional/body (G) and special (F) scales from the Toko Range shot hole localities; and a head scale (H) from the Toomba Range locality GEO 65/28.**
A—CPC 20081/4, crown view, x 40, locality SP 798; B—CPC 20084/1, crown view, x 40, locality SP 813; C—CPC 20083/7, basal view, x 73, locality SP 801 (compare with Fig. 9I); D—CPC 20082/7, crown view, x 73, locality SP 799; E—CPC 20081/7, tilted lateral view, x 73, locality SP 798; F—CPC 20083/4, crown view, x 53, locality SP 801; G—CPC 20081/3, crown view, x 43, locality SP 798; H—CPC 20080/4, crown view, x 53; I/CPC 20081/1, crown view, x 43, locality SP 798; J—CPC 20081/2, crown view, x 43, locality SP 798.

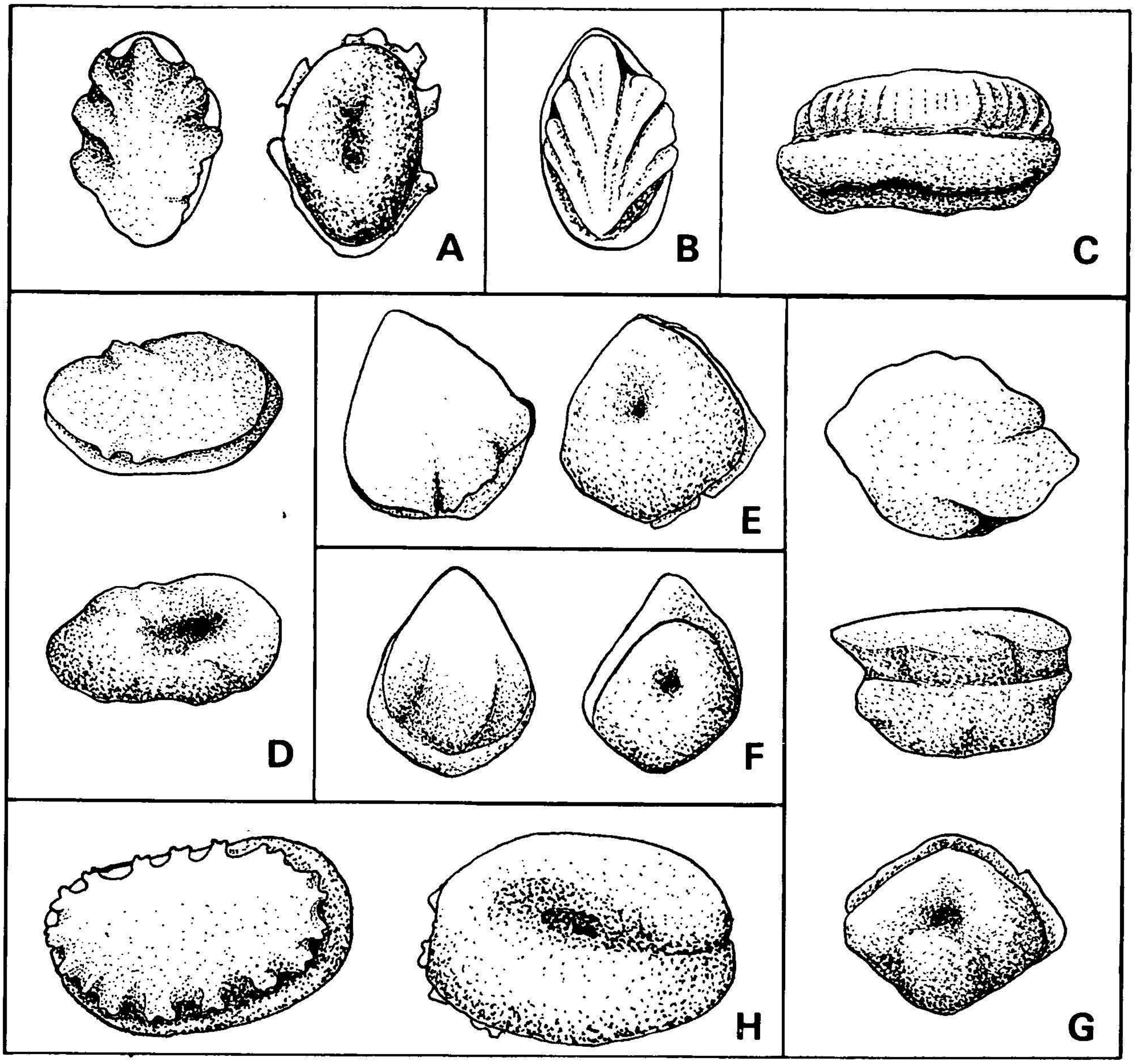

Figure 12. *Turinia australiensis* **Gross, 1971—Transitional (A, B) and head (C-H) scales from Earnstrey Hall Farm, Shropshire, from the Lower Dittonian** *Traquairaspis pococki* **'zone'.**
Deposited in the British Museum (Natural History), London (P60730-P60737), x 70.

(for example, those covering gill areas) or those of younger *T. pagei*, though this is less likely, as they can occur in large numbers and independently. Definite determination, as with most thelodont 'species', must await the discovery of good articulated material.

Many of the scales in the samples are head scales, or possibly special scales. Some bear a similarity to scales of *Thelodus sculptilis* Gross 1967. The head scales have a simple smooth flat crown, oval, ovate, or rhombic in shape, with more or less deeply crenulated edges (Figs. 9a-d; 11a, b, e). The crown perimeter is often smooth with a pair of notches anterior or midway, directed distally (Fig. 11d). Other special scales have a triangular crown with a tripartite division of ridges (Fig. 10f). The transitional scales have crowns which are more elliptical, with crenulations which are often bifurcated or trifurcated (Figs. 9j; 10a, c, e, g; 11i, j). Some have the minor ribbing or fluting on the crown. The body scales have crowns with more heavily incised ridges, again with minor ribs on them (Figs. 10g, h; 11c, g). They are similar in general shape to those of *T. pagei* of the Welsh Borderland, but are more complex in that they have a three-tiered crown with an anteriorly raised diamond-shaped portion; the crown may be smooth or finely striated with a lower central posterior point, some with minor cusps.

Typical head scales of *Turinia australiensis* are small, and have very deep rounded bases (Figs. 9a, c, e, h; Fig. 11e). Often the crowns are much smaller in area than the base (Fig. 11b), and the neck is shallow and unornamented. The scales referred to *T. australiensis* from the Toomba Range locality are also very small (0.5 mm to 1.0 mm) with deeply incised crenulations on the crown (Fig. 11h), whereas the scales of *T.* cf. *pagei* are larger (0.55 to 1.62 mm).

Some small head and transitional scales from the Dittonian deposits of the Welsh Borderland were so similar to those figured by Gross (1971b) that they were referred by one of us (Turner, 1973, p. 569, fig. 8f) to *T.* cf. *australiensis*, and later (Turner, 1976, p. 14) directly to this species. Karatajute-Talimaa (1978, p. 123) has since suggested that this form is synonymous with *T. polita* Karatajute-Talimaa, 1978,

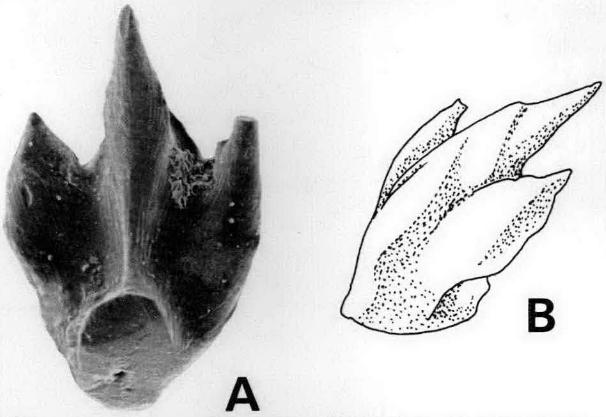

Figure 13. ***Gampsolepis*** **? sp. undet. (Family Nikoliviidae Karatajute-Talimaa, 1978)—a tricuspid scale CPC 20086 from Toomba Range, locality GEO 65/28.**
A—basal view; B—tilted lateral view; both x 60.

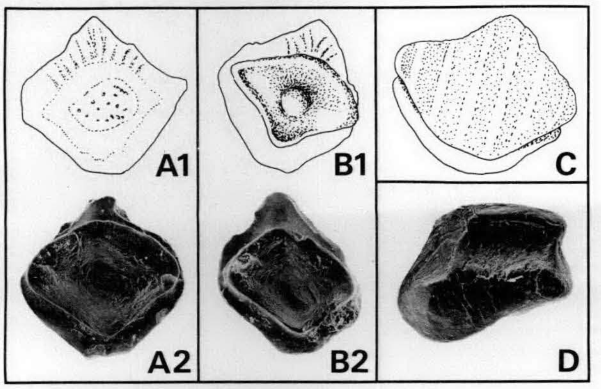

Figure 14. Acanthodian scales A-C—*Gomphonchus* ? sp. from Toomba Range locality GEO 65/28.
Crown (A1) and basal (A2) views of CPC 20087/1; basal (B1) and tilted basal (B2) views of CPC 20087/2; crown (C) view of CPC 20087/3; all x 43. D—*Nostolepis* sp. from locality SP 799, CPC 20088/1, lateral view, x 24.

and this may be correct in part. More scales from the Welsh Borderland are figured in this paper (Fig. 12) for comparison with those of *T. australiensis* Gross, 1971 and those referred to this species from the Toko Syncline.

Gampsolepis? sp. undet. (Fig. 13a, b)

One scale (CPC 20086) from the Toomba Range locality (GEO 65/28) is very different from those just described. It is tricuspid in shape, somewhat like the footprint of a wading bird. The crown rises at an angle of about 60° from a small rounded base, less than one third the total length of the scale, with a single pulp hole. Its median part rises to a posterior point, and in its anterior half two lateral wings, almost as long as the central part, are placed at an angle of about 10°.

This scale probably belongs in the family Nikoliviidae Karatajute-Talimaa, 1978, and resembles some described by Ørvig (1969b) as *Amaltheolepis winsnesi* (Wood Bay Group, late Emsian-early Eifelian, Vestspitsbergen), and others described by Karatajute-Talimaa (1978, and personal communication) as *Gampsolepis insueta* (Chortkov and Ivanev Horizons, Gedinnian, Podolia). With only one scale, no definite identification can be made, and histological analysis must await the discovery of further scales.

Associated fauna

Fish

Associated with the thelodont remains are two types of acanthodian scales. Those from the Toomba Range locality (GEO 65/28), CPC 20087/1,2, are small, about 0.5 mm, simple, and generally transparent, with a deep domed base and a flat smooth crown (Figures 14A, B). At least one scale shows the suggestion of 5 to 6 horizontal ribs on the crowned surface, aligned antero-posteriorly (Fig. 14C). These scales resemble those described by Gross (1967, 1971a) as *Gomphonchus gracilis,* but may eventually have to be referred to a new species.

Three larger and somewhat broken scales (CPC 20088) from locality SP799 (Fig. 14D) resemble *Nostolepis striata* (Pander) as redescribed by Gross (1971a). The genera *Gomphonchus* and *Nostolepis* commonly occur in Upper Silurian and Lower Devonian deposits in Europe.

Small curved, broken acanthodian spines with fine longitudinal ribbing have also been found. Bone fragments found in the samples are indeterminate remains of acanthodians or agnathans.

No other identifiable fish remains were found in the Cravens Peak Beds (in situ) until 1977. None of the fish fragments collected from the three localities noted by Gilbert-Tomlinson (1968) were found in place. Those from the locality 26 (incorrectly marked in Gilbert-Tomlinson, 1968, text-figure 5 as locality 24) were later determined by Ritchie (1973) as a species of *Wuttagoonaspis*—similar to, but larger than, *W. fletcheri* (the type species) from the lower part of the Mulga Downs Group of the Cobar region.

Young (BMR) is currently studying several collections of fossil fish remains which he made in 1977 from the Cravens Peak Beds. Placoderm (pterichthyodid antiarch) and onychodontid crossopterygian remains have been found in the basal (calcareous) unit with scales of *Turinia* cf. *pagei and T. australiensis* (Young's locality 11 at Toomba Range locality GEO 65/28), and *Wuttagoonaspis* sp., placoderms (including phlyctaeniids), crossopterygians (including onychodontids), and acanthodians have been recovered from the overlying conglomerate and sandstone sequence (Young & Gorter, 1981).

Crustacea

A small crustacean fauna, consisting of octracods and the problematic eridostracan *Cryptophyllus,* occurs together with the thelodont scales at locality GEO 65/28 in the Toomba Range, and in the shot hole localities SP 799 and 839, west of the Toko Range.

The Toomba Range sample contains about 100 specimens of *Cryptophyllus,* and 6 poorly preserved ostracod carapaces, probably belonging to the Kloedenellacea. Other ostracod specimens are present in both the Toomba and Toko Range samples as damaged isolated valves, broken steinkerns, and shell fragments. The valves are mostly smooth, lacking external macrosculpture, and internal details such as the adductor muscle scar pattern and the contact margin are obscured. Without such details, the following determinations, based mainly on the outline of the valves, must be open to question.

Healdianella inconstans Polenova, 1974? (Figs. 15A, B)

Six specimens from locality SP 799, of which two right valves (CPC 20089, length 1.55 mm, height

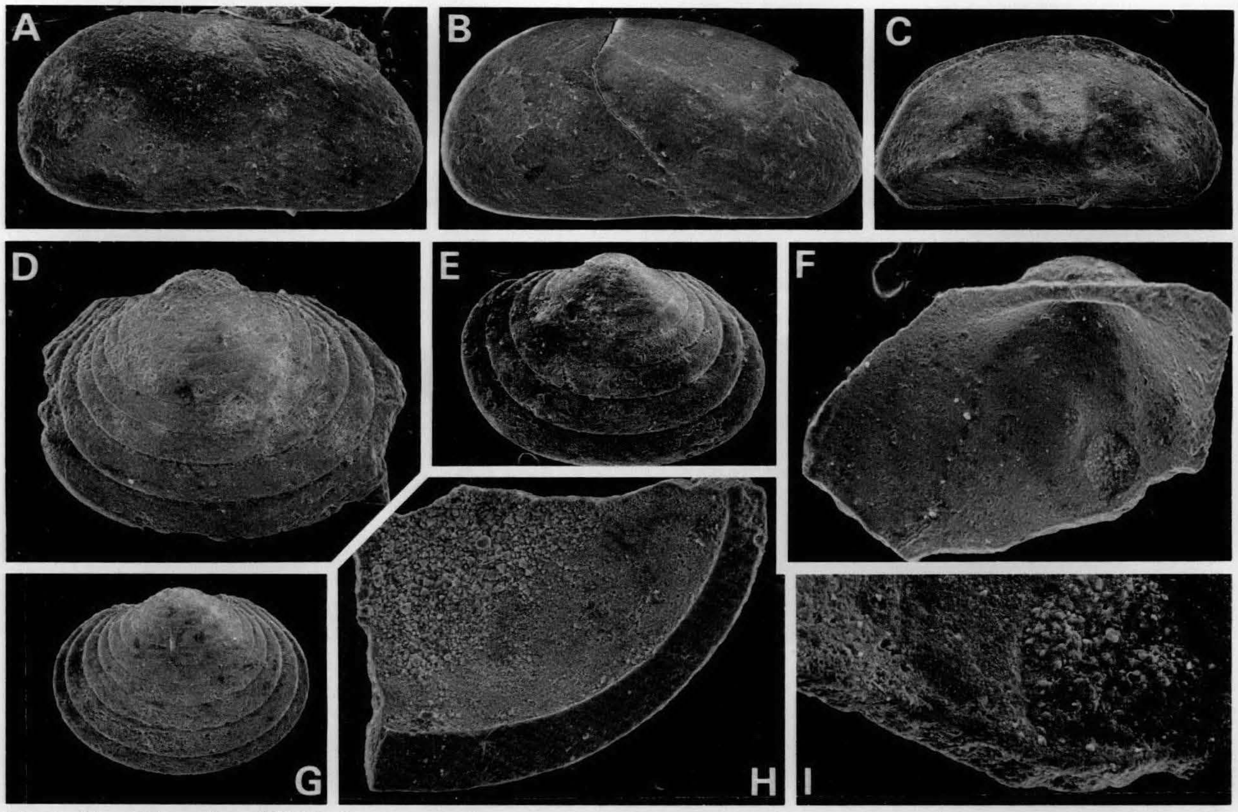

Figure 15. Ostracoda and Eridostraca.
A, B—*Healdianella inconstans* Polenova, 1974 ? from locality SP 799, x 40; A—right valve, CPC 20089; B—right valve, CPC 20090; C—*Baschkirina* ? sp., carapace, CPC 20091, from locality SP 839, x 40. D-I—*Cryptophyllus* sp. from the Toomba Range locality GEO 65/28, x 40, except where indicated. D—valve, CPC 20092; E—valve, CPC 20093; F—broken valve, CPC 20095, internal view showing position of adductor muscle scar and other muscle scars, x 90; G—carapace, CPC 20094; H—fragment of valve margin, CPC 20096, showing simple narrow duplicature, x 97; I—detail of adductor muscle scar on tilted valve, CPC 20095, x 270.

0.80 mm; CPC 20090, length 1.60 mm, height 0.80 mm) are figured here, appear to fall within the limits of variation of *Healdianella inconstans* Polenova, 1974. This species was originally described by Polenova (1974, p. 64, pl. 25, figs. 2-5, pl. 26, figs. 2, 3) from the Lower Devonian (Siegenian) lower Settedabanskii 'horizon' (beds with *Sibiritoechia lata*) in the Sette-Daban Ridge region of the northeastern part of the Siberian Platform. The Australian specimens may have had a calcified inner lamella, as evidenced by shell fragments and a broken steinkern, which raises the vexing question, beyond the scope of this paper, of the relationship between the two genera *Cytherellina* Jones & Holl, 1869 and *Healdianella* Posner, 1951 (see Adamczak, 1976, p. 341).

Baschkirina? sp.
(Fig. 15C)

A single carapace (CPC 20091, length 1.36 mm, height 0.68 mm, width 0.50 mm) from locality SP 839 (Fig. 15C) and an unfigured carapace from the Toomba Range locality GEO 65/28 have similar lateral and dorsal outlines as those of *Baschkirina densa* Polenova, 1974 from the Lower Devonian (Siegenian) rocks of Novaya Zemlya and of the Sette-Daban Ridge region (Polenova, 1974, p. 55, pl. 21, figs. 3-6). The Australian species differs, however, in that it has a few small rounded nodes in the centrolateral area of the valves, which are aligned with a rib-like bend at the posteroventral end. Both specimens have their posterior ends missing, and the poor preservation of the valves does not permit interpretation of the posterodorsal overlap; therefore, the species is questionably referred to *Baschkirina* Rozhdestvenskaya, 1959.

Cryptophyllus sp.
(Figs. 15D-I)

The eridostracan genus *Cryptophyllus* contains many species that are difficult to recognize on the basis of the published literature. Even with specimens for comparison, the problems of discrimination, owing to morphological variability, remain unresolved. Because the Toomba Range specimens present such a problem, they are, for the present, placed in open nomenclature.

From their external features, both figured (CPC 20092-94) and unfigured specimens appear to fall within the limits of variation for most of the external characters of *Cryptophyllus sp.* A. Jones, 1962. The ovate to slightly asymmetrical lateral outline, low umbo, and 8-11 wide growth bands, are all features common to this species, which was described from the Gneudna Formation in the Carnarvon Basin of Western Australia (Jones, 1962, p. 24, pl. 3, figs. 1-5). Specimens CPC 20093 and 20094 (Figs. 15E, G) have hinge lines which are slightly shorter (in relation to the total length of the shell) than those of the figured specimens of *Cryptophyllus* sp. A (CPC 4225, 4226, 4227), but this may be a matter of variation within the species. Unlike the specimens from the Gneudna Formation, they have well-defined cardinal angles, which appears to be a fairly constant character for the Toomba Range species.

The external features of specimens belonging to this species may be compared also with those referred in the literature to species of Early Devonian age in the Sahara (Le Fevre, 1963), North Spain (Becker & Sanchez de Posada, 1977), and Podolia (Abushik, 1968). For example, CPC 20094 (Fig. 15G) is comparable with *Cryptophyllus* sp. 1 Le Fevre, 1963 (pl. 8, fig. 127) from the middle Emsian part of the Tefergeunit Formation, and with *C.* sp. Abushik, 1968 (pl. 1, fig. 6) from the Gedinnian Borschov Horizon; CPC 20092 (Fig. 15D) is comparable with *C.* sp. A Becker & Sanchez de Posada, 1977 (pl. 14, figs. 8-11) from the late Emsian part of the Moniello Formation. However, none of these Old World Realm species appear to have the characteristic well-defined cardinal angles of the Toomba Range species.

On the internal surface of specimens of this species, the adductor muscle scar pattern (Fig. 15F, I) is similar to that recognized in other Australian species of *Cryptophyllus* (namely *C. diatropus* Jones, 1962; *C. platyogmus* Jones, 1962). SEM photographs indicate the presence of additional muscle scars (Fig. 15F), of as yet unknown significance, and a thickening of the free margin in the form of a simple narrow duplicature (Fig. 15H), not unlike that figured and described for *Cryptophyllus* sp. A Becker & Sanchez de Posada, 1977.

While this paper is not the place for a detailed discussion of the biological affiinties of *Cryptophyllus* and allied eridostracan genera, none of the features that Schallreuter (1977) uses in support of his argument that the genus is a true ostracod are *exclusive* to the Ostracoda. Few studies have been made of the adductor muscle scar patterns and contact margins in the Eridostraca, and there are no comprehensive comparative studies which indicate the position of this group within the Phylum Crustacea.

Age and correlation

Age relationships

Turner (1973) and Karatajute-Talimaa (1978) have reviewed the distribution of *Turinia pagei*. The scales of this species have been found in many Lower Devonian agnathan localities in Europe and the Soviet and Canadian Arctic, and are now being accepted as indicators of the beginning of the Devonian (that is, *Traquairaspis symondsi* Zone; Dittonian *sensu* Holland & Richardson, 1977). The type locality of *Turinia pagei* is in the Lower Garvock Group (of the Lower Old Red Sandstone) of the northern Midland Valley of Scotland, the age of which is Siegenian, based on spore evidence (J. B. Richardson, personal communication T. S. Westoll, 1977), and is no older than the base of the Dittonian (*sensu* White, 1950) based on the fish and eurypterids (Westoll, 1977). *Turinia pagei* is now known from the Dittonian (*sensu* Holland & Richardson, 1977) of the Welsh Borderland (Turner, 1973), the Baltic region, Brest Depression, Volynia-Podolia region, and the Arctic regions of Spitsbergen, Severnaya Zemlya (Karatajute-Talimaa, 1978), and possibly Prince of Wales Island (Turner, 1973). Recently, it has been described by Goujet & Blieck (1979) from the late Gedinnian of northern France, in association with the late Dittonian zone fossil *Belgicaspis crouchi* Lankester.

Gross (1971) regarded the scales of *Turinia australiensis* from Wilson Cliffs No. 1 well, Western Australia, as Dittonian, but there is no corroborative fossil evidence to confirm this age. Furthermore, despite the fact that some scales of the *T. australiensis* type are now known from Dittonian rocks of the Volynia-Podolia regions (under the name *Thelodus scoticus* Karatajute-Talimaa, 1964), and the Welsh Borderland (Turner, 1976, p. 14), we suggest that *T. australiensis* may extend into younger (Emsian-?Eifelian) deposits in Australia.

The family Nikoliviidae, represented in the Cravens Peak Beds by one scale, is an integral part of the Early (Dittonian) to Middle Devonian (Eifelian) succession of thelodont assemblages of Europe (Turner, 1973 Karatajute-Talimaa, 1978; Goujet & Blieck, 1979), and, if further nikoliviid scales are found, they may help to reduce the margin of error within this age range. The ostracod evidence is inconclusive for this purpose, because the specimens are poorly preserved and lack distinctive features. If the comparisons with *Healdianella inconstans* Polenova, 1974 and *Baschkirina densa* Polenova, 1974 are valid, they are more likely to indicate Early Devonian rather than a younger age. Despite a superficial likeness, the eridostracan *Cryptophyllus* sp. is probably not conspecific with C. sp. A Jones, 1962 from the Late Devonian Gneudna Formation of the Carnarvon Basin.

Thus, based on the age of the *Turinia pagei* fauna in the Old World Realm, the thelodont scales from the Cravens Peak Beds are no older than Dittonian, that is, no older than Gedinnian. The Devonian age of the basal (calcareous) unit of the Cravens Peak Beds is unequivocal, and the earlier consideration of a Late Silurian age now can be discarded (cf. Smith, 1972, pp. 131-32; Strusz, 1972, p. 449; Talent & others, 1975, p. 23).

The sandstone and conglomerate sequence which conformably overlies the basal (calcareous) unit contains *Wuttagoonaspis*, a placoderm at present unknown outside Australia (Young, 1974). The type species of *Wuttagoonaspis* (*W. fletcheri* Ritchie, 1973) is regarded as probably late Early to early Middle Devonian in age, on the basis of numerous small arctolepids associated with it in the lower part of the Mulga Downs Group of western New South Wales (Ritchie, 1973). Ritchie (1969, text-figs. 3, 4) pointed out that these forms show a close resemblance to Northern Hemisphere genera such as *Huginaspis, Heterogaspis*, and *Arctolepis*, and indicate a probable Emsian-Eifelian age, and he suggested (Ritchie 1973, p. 71) that the Mulga Downs Group (lower part) and the Cravens Peak Beds (also with *Wuttagoonaspis*) were approximately contemporaneous. A latex rubber impression of the concentration of skeletal debris found in association with *W. fletcheri*, kindly provided by Dr Ritchie, includes numerous scales of *Turinia*, which we refer to *T. australiensis* (Fig. 16). Thus, based on the combined thelodont, placoderm, and ostracod evidence, the total possible age for the Cravens Peak Beds appears to range from Gedinnian to Eifelian (that is Early Devonian to early Middle Devonian). Within this possible range, we emphasise the presumed Emsian-Eifelian age of the *Wuttagoonaspis* fauna (Ritchie, 1973), and suggest a similar age is possible for the *Turinia pagei* fauna, because the basal (calcareous) unit of the Cravens Peak Beds forms a gradational, conformable upper contact with the more

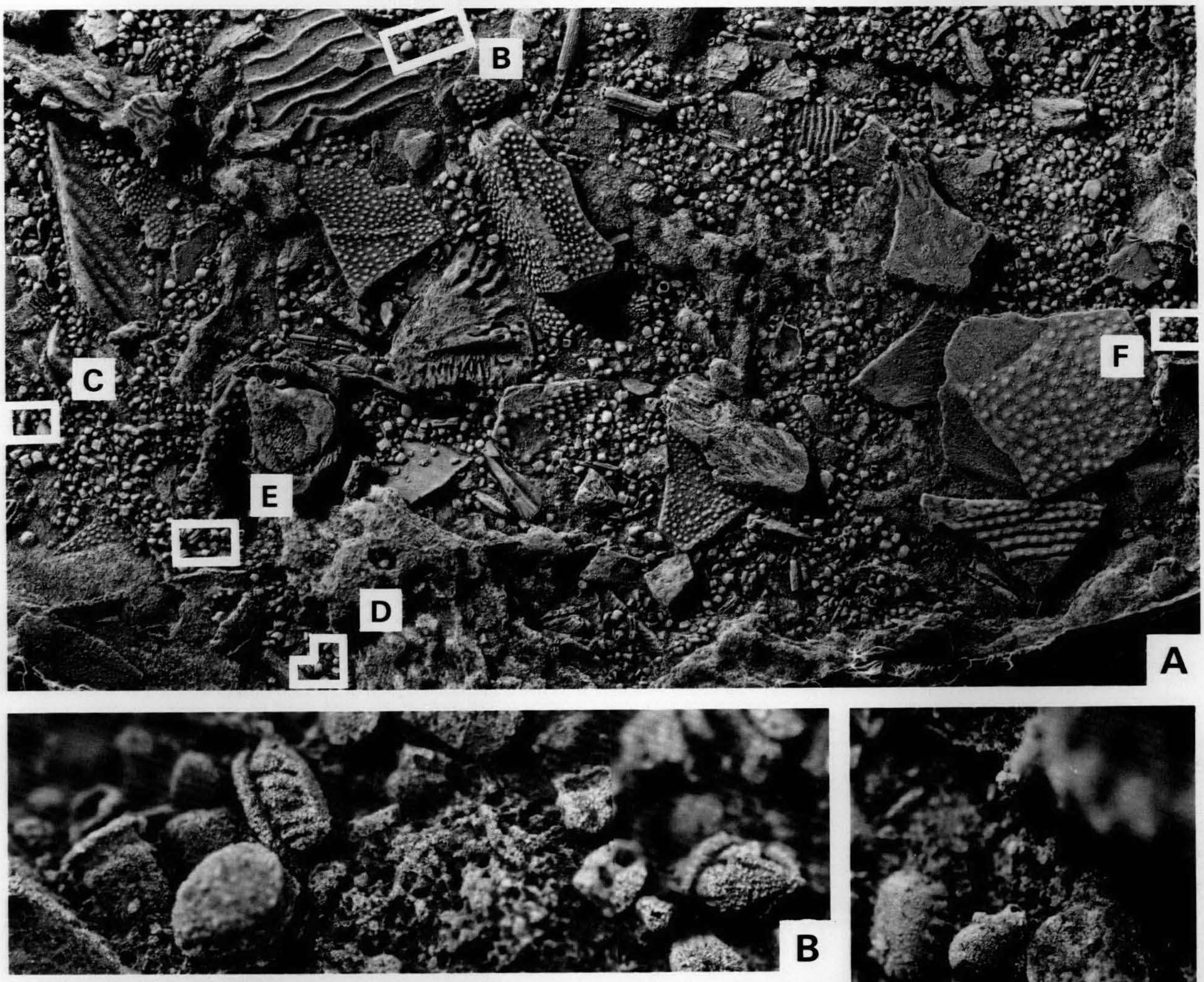

Figure 16. *Turinia australiensis* **Gross, 1971.**
A latex rubber cast prepared from a slab of fine grained quartzose sandstone (AM F 56264) collected by Dr A. Ritchie from the lower part of the Mulga Downs Group, exposed on Mount Jack Station (143°43'E 30°51'S) about 4 km E of homestead, and about 70 km NNE of Wilcannia in western New South Wales. A—view of specimen showing a rich concentration of thelodont scales in association with *Wuttagoonaspis* and other placoderm plate fragments, x 2. B-F—portions of specimen figures in A showing head and transitional scales of *Turinia australiensis*, x 22.

typical sandstones and conglomerates. These two vertebrate-faunas are probably biofacies equivalents of each other, and in the lower part of the Mulga Downs Group at least, they appear to be contemporaneous. Moreover, they are no older than Emsian in age, because Glen (1979) has recently shown that the Mulga Downs Group overlies, possibly paraconformably, the upper part of the Amphitheatre Group, which has been dated as Pragian on the presence of the brachiopod *Howellella jaqueti* (Dun).

Thus, we suggest that the age of the *Turinia pagei* fauna, which is Dittonian in Europe, is not necessarily contemporaneous elsewhere, and that, in Australia at least, this fauna is found in rocks at least as young as Emsian. Turner & Janvier (1979), for instance, have suggested that the Iranian species, *Turinia hutkensis* Blieck & Goujet, 1978, is probably Middle Devonian (Eifelian) rather than Early Devonian in age, and Young & Gorter (1981) have described a species that closely resembles *T. hutkensis* (in association with *Bothriolepis*) from late? Eifelian rocks in New South Wales.

Regional correlation

On the basis of the thelodont evidence, the Cravens Peak Beds appear to be broadly coeval with at least part of the Tandalgoo Red Beds in the Canning Basin of Western Australia, and with the lower part of the Mulga Downs Group in the Cobar Basin of north western New South Wales (Figure 17).

In the Canning Basin, scales of *Turinia australiensis* were recovered from Wilson Cliffs No. 1 well (22°16′39″S 126°46′55″E; Fig. 1, locality 3) (Gross, 1971), from core 5 (4439-4446 feet), which is situated within an interval (3590-5835 feet) identified by Creevey (1971) as Tandalgoo Red Beds. Because, as we have mentioned previously, scales of this thelodont species are associated with the *Wuttagoonaspis* fauna of central Australia (including western New South Wales), their presence and position within the Tandalgoo Red Beds suggest a late Early Devonian-

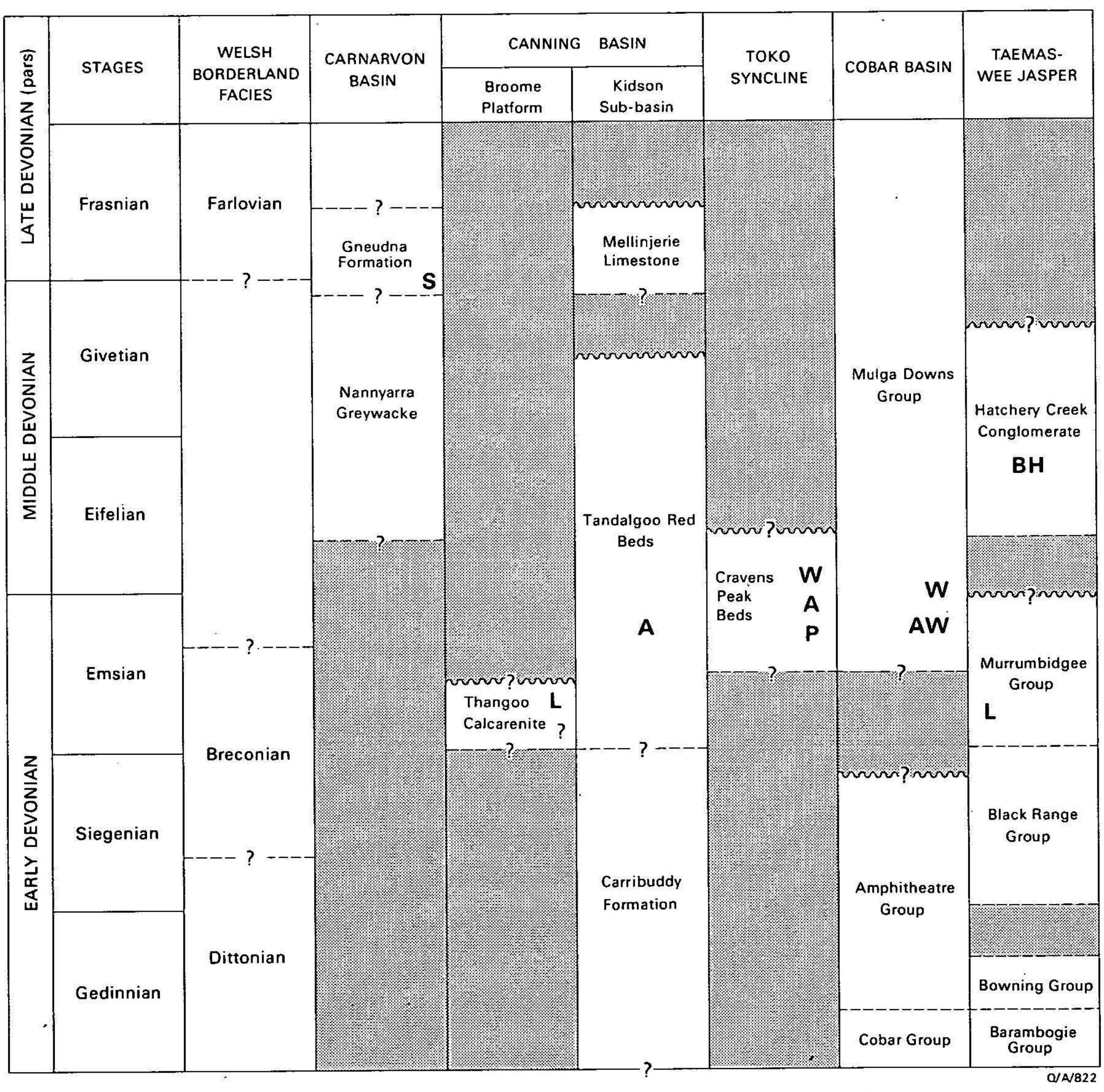

Figure 17. Correlation chart of Devonian sequences in Australia which have yielded thelodont scales and other fish remains mentioned in the text.
Key to symbols is as follows: P—*Turinia* cf. *pagei*; A—*T. australiensis*; H—*T.* ? cf. *hutkensis*; S—*Australolepis seddoni*; L—*Ligulalepis*; W—*Wuttagoonaspis* spp; B—*Bothriolepis* spp. Shaded areas within a column indicate a hiatus.

Middle Devonian age for the formation. Associated with the thelodont scales were indeterminate conodonts, and acanthodian fish remains, which Gross (1971) did not identify, but compared with *Diplacanthus longispinus* Agassiz from the Middle Devonian of Scotland. He found none of the acanthodian scales commonly associated with *Turinia pagei* in European localities. However, Turner (1973) has shown that acanthodian species are not very useful for correlation, having, in general, longer stratigraphic ranges than thelodonts. The conodonts, as Veevers (1976) pointed out, show that at least part of the Tandalgoo Red Beds is marine. The stratigraphy is complicated by the fact that the core which yielded these fossils also contained Permian spores (Gross, 1971b). Of the two possible solutions to this problem—(i) contamination of Devonian rocks by Permian spores, or (ii) reworking of Devonian fossils in Permian deposits—Veevers (1976, p. 190) favoured the second. We suggest, however, the first solution is the more plausible, because the sequence containing both Early Devonian fossils and Permian spores is overlain by the Mellinjerie Limestone (Creevey 1971), which elsewhere contains Middle Devonian spores.

A second report of thelodont remains in the Canning Basin was noted by McTavish & Legg (1976, p. 459), and based on a single scale from the upper part of the Thangoo Calcarenite (*sensu* McTavish, *in* Playford & others, 1975, p. 326) in Roebuck Bay No. 1 well (18°09'S 122°27'E) on the Broome Platform (Fig. 1, locality 2). It was identified, but not described or figured, as belonging to cf. *Apalolepis obruchevi* Karatajute-Talimaa, 1967, a species known from the Early Devonian (Dittonian) Chortkov Horizon of Podolia. Later examination of a photograph of this scale by one of us (ST) leads us to conclude that it came from a palaeoniscid cf. *Ligulalepis* Schultze, 1968, and not from a thelodont. On this basis, the Thangoo Calcarenite may be younger than Dittonian, because at present the only known occurrence of *Ligulalepis* is in the Taemas area, New South Wales, in the *'Spirifer' yassensis* limestone in the lower part of the Murrumbidgee Group, which is early Emsian (Zlichovian) in age (Pedder & others, 1970; Klapper & Johnson, 1980). Thus, the age of the Thangoo Calcarenite may not be much different to those parts of the Tandalgoo Red Beds, Cravens Peak Beds, and Mulga Downs Group which contain scales of *Turinia australiensis*.

The remains of *Turinia australiensis* and the *Wuttagoonaspis* fauna in the lower part of the Mulga Downs Group, are older than the recently described (Young & Gorter, 1981) fish fauna in the Hatchery Creek Conglomerate from the Taemas-Wee Jasper region of New South Wales (Fig. 1, locality 8). This fauna includes, amongst other species, a new species of *Bothriolepis*, with primitive characters indicative of an early representative (Eifelian) of the genus, and thelodont scales, which Young & Gorter refer to *Turinia?* cf. *hutkensis* Blieck & Goujet, 1978, from the Middle Devonian (Eifelian) of Iran (see Turner & Janvier, 1979).

Palaeogeography

Regional

The results of the present study have a bearing on some of the published palaeogeographic reconstructions of Devonian Australia (Johnstone & others, 1967; Gilbert-Tomlinson, 1968; Brown & others, 1968; Webby, 1972; Veevers, 1976). Our recognition of marginal marine deposition in central Australia during the Early Devonian supports the interpretation of Veevers (1976, p. 190), who showed a marine connection between the Georgina and Canning Basins via the Amadeus Basin. The presence of *Turinia* in both the Canning Basin and the Georgina Basin supports this contention, and its distribution in New South Wales (Fig. 1, localities 7 & 8) further suggests a connection with the Lachlan Fold Belt.

In most reconstructions of Middle and Late Devonian palaeogeography, the streams responsible for deposition of continental sediments in central Australia are assumed to flow ultimately into either southeastern or northeastern Australia. However, current direction measurements in the Cravens Peak Beds (a total of 33 measurements at six locations) gave an overall vector mean of 330°, indicating for Early Devonian time, a general direction of flow towards the northwest. The possibility of a river system flowing via the Wiso Basin and The Granites-Tanami area into either the Canning or Bonaparte Gulf Basin is worthy of further investigation. The restriction of *Wuttagoonaspis*, an apparently freshwater form, to central Australia (including western New South Wales) is a major obstacle to such a reconstruction. Many more current direction measurements from other Devonian sandstones, and a more detailed knowledge of fossil fish distribution patterns are required before a realistic Lower Devonian palaeogeographical reconstruction is possible.

Global

Halstead & Turner (1973) and Turner (1973) have posed the question of the whereabouts of Australia in Early Devonian times. Both Europe and Australia were apparently in tropical latitudes during the Devonian (Embleton, 1973; Smith & others, 1973; Heckel & Witzke, 1979), and Turner (1973) suggested that thelodonts may have dispersed via Arctic Canada, either to or from Australia.

Of the available palaeogeographic reconstructions for the Devonian, only the Heckel & Witzke map (1979, text-fig. 5) for the Middle Devonian shows the presumed current patterns and shorelines, and for that reason the map, based on palaeoclimatic indicators, is selected here as a base to show the known world distribution of thelodont faunas during the Devonian (Fig. 18).

The Lower Devonian occurrences of thelodont genera in northern Canada, Spitzbergen, northern Europe, and New South Wales suggest direct communication between Arctic North America and eastern Australia. Klapper & Johnson (1980, pp. 434-5) suggested a strong equatorial current for the dispersal of brachiopod larvae across the Proto-Pacific from North America to Australia in the Early Devonian, and the same current system explained the similarity of Cordilleran-Arctic conodont faunas with those of eastern Australia in the Early and Middle Devonian. We suggest that the thelodont faunas also may have been distributed by this current system.

It is tentatively postulated that the *Turinia* fauna reached Australia in the Early Devonian (Breconian?) and then began an adaptive radiation to give apparently younger (Middle Devonian) thelodont faunas that survived longer in the Australian region than elsewhere (Ørvig, 1969a, b, c; Karatajute-Talimaa, 1978; Turner & Janvier, 1979). The youngest known thelodont scales come from the Gneudna Formation (of latest Givetian-earliest Frasnian age) in the Carnarvon Basin, Western Australia.

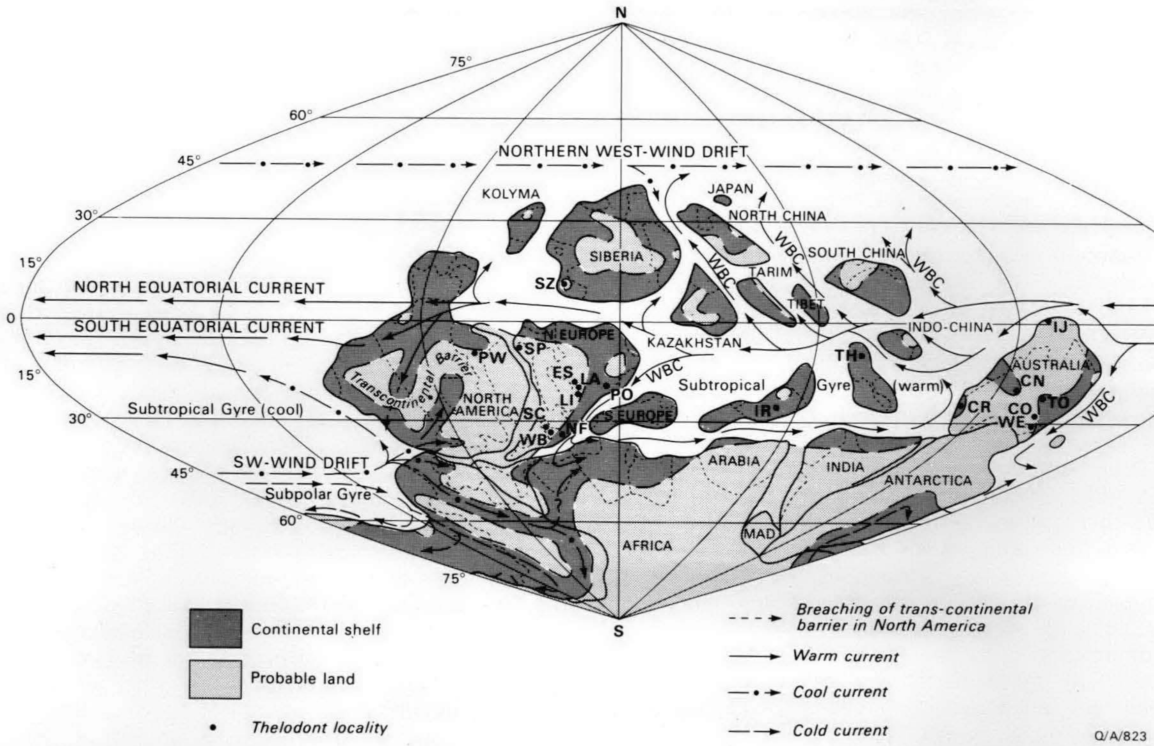

Figure 18. Middle Devonian world reconstruction, based on palaeoclimatic indicators, showing expected pattern of oceanic circulation, land and continental shelf, and distribution of Devonian thelodont localities; modified from Heckel & Witzke (1979, text-fig. 5).
Symbols for localities are as follows: CN—Canning Basin; CR—Carnarvon Basin; CO—Cobar Basin; ES—Estonia; IJ—Irian Jaya; IR—Iran; LA—Latvia; LI—Lithuania; NF—North France; PO—Podolia; PW—Prince of Wales Island; SC—Scotland; SP—Spitsbergen; SZ—Severnaya Zemlya; TH—Thailand; TO—Toko Syncline; WB—Welsh Borderland; WE—Wee Jasper.

Conclusions

1. Thelodont scales found in the basal (calcareous) unit of the Cravens Peak Beds belong to *Turinia australiensis* Gross, 1971. Some are similar to, but possibly not conspecific with, *Turinia pagei* (Powrie, 1870), and the family Nikoliviidae is probably represented by a single tricuspid scale, which may belong to the genus *Gampsolepis.*
2. Ostracods (e.g., *Healdianella inconstans* Polenova 1974?; *Baschkirina?* sp.) and eridostracans (viz., *Cryptophyllus* sp.) associated with the thelodont scales indicate a marine to marginal marine depositional environment for the basal (calcareous) part of the Cravens Peak Beds.
3. The basal (calcareous) part of the Cravens Peak Beds grades conformably upwards into an overlying sequence of sandstone and conglomerate (280 m) of braided stream origin. The sandstone contains a fish fauna with *Wuttagoonaspis* sp. that probably represents a freshwater biofacies of the *Turinia* fauna.
4. The presence of the *Turinia* fauna in the Cravens Peak Beds indicates an unequivocal Devonian age no older than Dittonian *sensu* Holland & Richardson, 1977), and the earlier suggestion of a possible Late Silurian age now can be discarded (cf. Smith 1972, pp. 131-32; Strusz, 1972, p. 449; Talent & others, 1975, p. 23).
5. The age of the Cravens Peak Beds is probably younger than Dittonian, because, in the lower part of the Mulga Downs Group in the Cobar area, New South Wales, scales of *Turinia australiensis* Gross, 1971 (figured in this paper) are associated with the *Wuttagoonaspis* fauna, which is at least as young as Emsian (Glen, 1979).
6. A tentative correlation is proposed between those parts of the Cravens Peak Beds, Tandalgoo Red Beds, and lower Mulga Downs Group that contain the *Turinia australiensis* fauna. Thus, although *Turinia australiensis* is present in the Dittonian of the Welsh Borderland, where it is commonly associated with *T. pagei,* the *T. australiensis* fauna in Australia appears to characterise younger deposits, of late Early to early Middle Devonian age.
7. During the (Emsian?) marine phase in the early part of the depositional history of the Cravens Peak Beds, a sea connection was possibly established between the Canning and Georgina Basins, via the Amadeus Basin, and perhaps with a southeastern extension into the Lachlan Fold Belt. With the onset of continental conditions, a river system may have flowed in a northwest direction via the Wiso Basin and The Granites-Tanami area into either the Canning or the Bonaparte Gulf Basins.
8. Globally, the thelodont faunas are thought to have been distributed by the system of current patterns envisaged by Heckel & Witzke (1979); it is tentatively postulated that the *Turinia* fauna reached Australia in Early Devonian (Breconian?) time, and then began an adaptive radiation to give apparently younger (Middle Devonian) thelodont faunas that survived longer in the Australian region than elsewhere; the youngest scales now known come from the Gneudna Formation (of latest Givetian-earliest Frasnian age) in the Carnarvon Basin, Western Australia.

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References

ABUSHIK, A. F., 1968—Ostrakody Borshchovskogo gorizonta Podolii (Ostracods from the Borshchov horizon of Podolia): *in* SOKOLOV, B. S., & IWANOWSKI, A. B. (Editors), Biostratigrafiya pogranichn'ykh otlozhenii Silura i Devona (Biostratgraphy of Silurian/Devonian boundary deposits) *Moskva Nauka*, 49-74.

ADAMCZAK, F., 1976—Middle Devonian Podocopida (Ostracoda) from Poland; their morphology, systematics and occurrence. *Senckenbergiana lethaea*, **57**, 265-467.

ALLEN, J. R. L., 1965—A review of the origin and characteristics of recent alluvial sediments. *Sedimentology*, **5**, 91-191.

ALLEN, J. R. L., HALSTEAD (TARLO), L. B. & TURNER, S., 1968—A Dittonian ostracoderm fauna from the Brownstones of Wilderness, Quarry, Mitcheldean, Gloucestershire. *Proceedings of the Geological Society of London*, **1649**, 141-53.

ALLIANCE OIL DEVELOPMENT AUSTRALIA N.L., 1975—Well completion report Ethabuka No. 1 (unpublished company report).

BECKER, G., & SANCHEZ DE POSADA, L. C., 1977—Ostracoda aus der Moniello-Formation Asturiens (Devon; N-Spanien). *Palaeontographica*, (A) **158**, 115-203.

BLIECK, A., & GOUJET, D., 1978—A propos de nouveau material de Thélodontes (Vertébrés Agnathes) d'Iran et de Thailande: apercu sur la repartition géographique et stratigraphique des Agnathes des 'regions gondwaniennes' au Paléozoique moyen. *Annales de la Société Geologique du Nord*, **97**, 363-72.

BROWN, D. A., CAMPBELL, K. S. W., & CROOK, K. A. W., 1968—The Geological Evolution of Australia and New Zealand. *Pergamon Press, London.*

COCKS, L. R. M., HOLLAND, C. H., RICKARDS, R. B., & STRACHAN, I., 1971—A correlation of Silurian rocks in the British Isles. *Journal of the Geological Society, London*, **127**, 103-36.

CREEVEY, K. J., 1971—Recent information and prospects in the Kidson Sub-basin. *The APEA Journal*, **11**, 53-8.

DRAPER, J. J., 1976—The Devonian rocks of the Toko Syncline, western Queensland. *Bureau of Mineral Resources, Australia, Record* **1976/29** (unpublished).

DRAPER, J. J., 1980—Ethabuka Sandstone, a new Ordovician unit in the Georgina Basin, and a redefinition of the Toko Group. *Queensland Government Mining Journal*, **81**, 469-75.

EMBLETON, B. J. J., 1973—The palaeolatitude of Australia through Phanerozoic time. *Journal of the Geological Society of Australia*, **19**, 475-82.

GILBERT-TOMLINSON, J., 1968—A new record of *Bothriolepsis* in the Northern Territory of Australia. *Bureau of Mineral Resources, Australia, Bulletin* **80**, 191-226.

GLEN, R. A., 1979—The Mulga Downs Group and its relation to the Amphitheatre Group, southwest of Cobar. *New South Wales Geological Survey, Quarterly Notes*, **36**, 1-10.

GOUJET, D., & BLIECK, A., 1977—La faune de Vertébrés de l'horizon 'Vogti' (Groupe de Red Bay, Spitsberg). Comparaison avec les autres faunes ichthyologiques du Dévonien inférieur européen. *Compte rendus des Seances de l'Academie des Sciences, Paris*, **284**(16) Série D 1513-5.

GOUJET, D., & BLIECK, A., 1979—Les Vertébrés de l'Assise des Schistes et Grès de Pernes (Devonien du Nord de la France). *Annales de la Société Geologique du Nord* **98**, 263-78.

GROSS, W., 1967—Über Thelodontier-Schuppen. *Palaeontographica*, (A) **127**, 1-67.

GROSS, W., 1971a—Downtonische und Dittonische Acanthodier-Reste des Ostseegebietes. *Palaeontographica*, (A) **136**, 1-82.

GROSS, W., 1971b—Unterdevonische Thelodontier-und Acanthodier-Schuppen aus Westaustralien. *Palaeontologische Zeitschrift*, **45**, 97-106.

HALSTEAD, L. B., 1971—The Siluro-Devonian boundary revisited—a personal recollection. *Journal of the Geological Society, London*, **127**, 630-2.

HALSTEAD, L. B., & TURNER, S., 1973—Silurian and Devonian Ostracoderms. *In* HALLAM, A. (Editor), Atlas of Palaeobiogeography. *Elsevier, Amsterdam*, 67-79.

HECKEL, P. H., & WITZKE, B. J., 1979—Devonian World Palaeogeography determined from distribution of carbonates and related lithic palaeoclimatic indicators. *In* HOUSE, M. R., SCRUTTON, C T., & BASSETT, M. G. (Editors), The Devonian System. *Special Paper in Palaeontology*, **23**, 99-123.

HOLLAND, C. H., & RICHARDSON, J. B., 1977—The British Isles: *In* MARTINSSON, A. (Editor), The Silurian-Devonian Boundary. *International Union of Geological Sciences*, Series A Number 5, 35-44.

JOHNSTONE, M. H., JONES, P. J., KOOP, W. J., ROBERTS, J., GILBERT-TOMLINSON, J., VEEVERS, J. J., & WELLS, A. T., 1967—Devonian of Western and central Australia. *In* OSWALD, D. H. (Editor), International Symposium on the Devonian System, *Alberta Society of Petroleum Geologists, Calgary*, **1**, 599-612.

JONES, P. J., 1962—The ostracod genus *Cryptophyllus* in the Upper Devonian and Carboniferous of Western Australia. *Bureau of Mineral Resources, Australia, Bulletin*, **62-3**, 1-37.

JONES, P. J., 1968—Upper Devonian Ostracoda and Eridostraca from the Bonaparte Gulf Basin, northwestern Australia. *Bureau of Mineral Resources, Australia, Bulletin* **99**, 1-93.

KARATAJUTE-TALIMAA, V. N., 1964—Dannye k stratigrafii Nizhnego Devona Yuzhnoi Pribaltiki (Contributions to the stratigraphy of the Lower Devonian of the Southern Baltic). *In* KARATAJUTE-TALIMAA, V. N., & NARBUTAS, V. V. (Editors), Voprosy stratigrafii i paleogeografii Devona Pribaltiki (Questions of stratigraphy and palaeogeography of the Devonian Baltic). *Geologii Institut Vilnius*, 21-39.

KARATAJUTE-TALIMAA, V. N., 1968—Novye telodonty, heterostraki i artrodiry iz Chortkovskogo gorizonta Podolii (New thelodonts, heterostracans and arthrodires from the Chortkov horizon in Podolia). *In* OBRUCHEV, D. V. (Editor), Ocherki po filogenii i sistematike iskopaemykh ryb i beschelyustnykh (Investigations on the phylogeny and taxonomy of fossil Agnathi and Fishes). *Moskva, Nauka*, 33-42.

KARATAJUTE-TALIMAA, V. N., 1970—Ikhtiofauna Dauntona Litvy, Estonii i Severnogo Timana (Ichthyofauna of the Downtonian deposits in Lithuania, Estonia, and north Timan). *In* Paleontologiya i stratigrafiya Pribaltiki i Belorussii (Palaeontology and Stratigraphy of the Baltic and Byelorussia) Sbornik **2**, *Vil'nyus*, 33-66.

KARATAJUTE-TALIMAA, V. N., 1978—Silurian and Devonian Thelodonts of the USSR and Spitsbergen. *Vilnius, Mokslas*, 1-334 (In Russian).

KLAPPER, G., & JOHNSON, J. G., 1980—Endemism and dispersal of Devonian conodonts. *Journal of Paleontology*, **54**(2), 400-55.

LAWSON, J. D., 1971—The Silurian-Devonian boundary. *Journal of the Geological Society, London*, **127**, 629-30.

LE FEVRE, J., 1963—Microfaunes de l'Emsien et du Dévonien moyen de la région Ougarta-Saoura (Sahara). *Diplome d'Études Supérieures-Université de Bordeaux, Pau.*

MCTAVISH, R A., & LEGG, D. P., 1976—The Ordovician of the Canning Basin, Western Australia. *In* BASSETT, M. G. (Editor), The Ordovician System: proceedings of a Palaeontological Association symposium, Birmingham, September 1974, Cardiff. *University of Wales Press and National Museum of Wales*, 447-78.

MARK-KURIK, E., 1969—The distribution of vertebrates in the Silurian of Estonia. *Lethaia*, **2**, 145-52.

MARK-KURIK, E., & NOPPEL, T., 1970—Additional notes on the distribution of vertebrates in the Silurian of Estonia. *Eesti NSU Tead. Akad Toimetised, Keemia, Geoloogia*, **19**, **2**, 171-3.

MOSKALENKO, T. A., 1968—Llandoveriiskie ostatki Agnatha v Sibiri (Llandoverian agnathan fossils in Siberia). *In* Ocherki po filogenii i sistematike iskopaemykh ryb i beschelyustnykh (Investigations on the phylogeny and taxonomy of fossil Agnathi and Fishes). *Moskva, Nauka*, 29-32.

MULREADY, J., 1975—Well completion report Ethabuka No. 1. *Alliance Oil Development Australia N.L. Company Report* (unpublished).

ØRVIG, T., 1969a—Thelodont scales from the Grey Hoek Formation of Andrée Land, Spitsbergen. *Norsk Geologisk Tidsskrift*, **49**, 387-401.

ØRVIG, T., 1969b—The vertebrate fauna of the *Primaeva* beds of the Fraenkelryggen Formation of Vestspitsbergen and its biostratigraphic significance. *Lethaia*, **2**, 219-39.

ØRVIG, T., 1969c—Vertebrates from the Wood Bay Group and the position of the Emsian-Eifelian boundary in the Devonian of Vestspitsbergen. *Lethaia*, **2**, 273-328.

PEDDER, A. E. H., JACKSON, J. H., & PHILIP, G. M., 1970—Lower Devonian biostratigraphy in the Wee Jasper region of New South Wales. *Journal of Paleontology*, **44**, 206-51.

PLAYFORD, P. E., COPE, R. N., COCKBAIN, A. E., LOW, G. W., & LOWRY, D. C., 1975—Phanerozoic. *In* Geology of Western Australia. *Geological Survey of Western Australia Memoir* **2**.

POLENOVA, E. N., 1974—Ostrakody Rannego Devona Articheskikh Raionov SSSR, (Early Devonian ostracods of the USSR Arctic Regions) *Akademiya Nauk SSSR, Sibirskoe Otdelenie, Instituta Geologii i Geofiziki Trudy*, **234**, *Moskva, Nauka*.

PRITCHARD, P. W., 1960—The Ordovician section in the Toko Range. *In* HILL, D., & DENMEAD, A. K. (Editors), The Geology of Queensland. *Journal of the Geological Society of Australia* **7**, 110-4.

REYNOLDS, M. A., 1965—Glenormiston, Queensland, 1:250 000 Geological Series. *Bureau of Mineral Resources, Australia, Explanatory Notes* **SF/54-9**.

REYNOLDS, M. A., 1968—Mount Whelan, Queensland, 1:250 000 Geological Series. *Bureau of Mineral Resources, Australia, Explanatory Notes* **SF/54-13**.

REYNOLDS, M. A., & PRITCHARD, P. W., 1964—The geology of the Glenormiston 1:250 000 Sheet area. *Bureau of Mineral Resources, Australia, Record* **1964/28** (unpublished).

RITCHIE, A., 1969—Ancient fish of Australia. *Australian Natural History*, **16**, 218-23.

RITCHIE, A., 1973—*Wuttagoonaspis* gen. nov., an unusual arthrodire from the Devonian of western New South Wales, Australia. *Palaeontographica*, (A) **143**, 58-72.

ROBERTS, J., JONES, P. J., JELL, J. S., JENKINS, T. B. H., MARSDEN, M. A. H., MCKELLAR, R. G., MCKELVEY, B. C., & SEDDON, G., 1972—Correlation of the Upper Devonian rocks of Australia. *Journal of the Geological Society of Australia*, **18**, 467-90.

SCHALLREUTER, R. E. L., 1977—On *Cryptophyllus gutta* Schallreuter. *Stereo-Atlas of Ostracod Shells*, **4**, 1-8.

SCHULTZE, H.-P., 1968—Palaeoniscoidea-schuppen aus dem Unterdevon Australiens und Kanadas und aus dem Mitteldevon Spitzbergens. *Bulletin of the British Museum (Natural History), (Geology)* **16**(7), 341-68.

SEDDON, G., 1969—Conodont and fish remains from the Gneudna Formation, Carnarvon Basin, Western Australia. *Journal and Proceedings of the Royal Society of Western Australia*, **52**, 21-30.

SEDDON, G., 1970—Frasnian conodonts from the Sadler Ridge-Bugle Gap area, Canning Basin, Western Australia. *Journal of the Geological Society of Australia*, **16**, 723-53.

SMITH, A. G., BRIDEN, J. C., & DREWRY, G. E., 1973—Phanerozoic World Maps. *In* HUGHES, N. F. (Editor), Organisms and Continents through Time. *Palaeontological Association, London, Special Papers in Palaeontology*, **12**, 1-42.

SMITH, K. G., 1963—Hay River, Northern Territory, 1:250 000 Geological Series. *Bureau of Mineral Resources, Australia, Explanatory Notes* **SF/53-16**.

SMITH, K. G., 1965—Tobermory, Northern Territory, 1:250 000 Geological Series. *Bureau of Mineral Resources, Australia, Explanatory Notes* **SF/53-12**.

SMITH, K. G., 1972—Stratigraphy of the Georgina Basin. *Bureau of Mineral Resources, Australia, Bulletin* **111**, 1-156.

SMITH, K. G., VINE, R. R., & MILLIGAN, E. N., 1961—Revisions to the stratigraphy of the Hay River, Huckitta and Tobermory 4-mile sheets, Northern Territory. *Bureau of Mineral Resources, Australia, Record* **1961/65** (unpublished).

STRUSZ, D. L., 1972—Correlation of the Lower Devonian rocks of Australasia. *Journal of the Geological Society of Australia*, **18**, 427-55.

TALENT, J. A., BERRY, W. B. N., & BOUCOT, A. J. (with contributions from PACKHAM, G. H., & BISCHOFF, G. C. O.) 1975—Correlation of the Siluran rocks of Australia, New Zealand, and New Guinea. *Geological Society of America Special Paper* **150**.

TURNER, S., 1971—Thelodonts and the Silurian-Devonian boundary. *Journal of the Geological Society, London*, **127**, 632-5

TURNER, S., 1973—Siluro-Devonian thelodonts from the Welsh Borderland. *Journal of the Geological Society, London*, **129**, 557-84.

TURNER, S., 1976—Thelodonti (Agnatha): *in* WESTPHAL, F. (Editor) Fossilium Catalogus 1: Animalia. *Junk, 's-Gravenhage*, **122**, 1-33.

TURNER, S., & DRING, R. S., 1981—Late Devonian thelodonts (Agnatha) from the Gneudna Formation, Carnarvon Basin, Western Australia. *Alcheringa* (in press).

TURNER, S., & JANVIER, P., 1979—Middle Devonian Thelodonti (Agnatha) from the Khush-Yeilagh Formation, northeast Iran. *Géobios*, **12**, 889-92.

VEEVERS, J. J., 1976—Early Phanerozoic events on and alongside the Australasian-Antarctic Platform. *Journal of the Geological Society of Australia*, **23**, 183-206.

WEBBY, B. D., 1972—Devonian geological history of the Lachlan Geosyncline. *Journal of the Geological Society of Australia*, **19**, 99-123.

WESTOLL, T. S., 1977—Northern Britain. *In* HOUSE, M. R., & OTHERS. A correlation of Devonian rocks of the British Isles. *Geological Society, London, Special Report* **8**.

WESTOLL, T. S., SHIRLEY, J., DINELEY, D. L., & BALL, H. W., 1971—The Silurian Devonian Boundary—Written contribution to the paper 'A correlation of Silurian rocks in the British Isles'. *Journal of the Geological Society, London*, **127**, 285-8.

WHITE, E. I., 1950—The vertebrate faunas of the Lower Old Red Sandstone of the Welsh Borders. *Bulletin of the British Museum (Natural History), (Geology)* **1**, 51-67.

YOUNG, G. C., 1974—Stratigraphic occurrence of some placoderm fishes in the Middle and Late Devonian. *Newsletters on Stratigraphy*, **3**, 243-61.

YOUNG, G. C., & GORTER, J. D., 1981—A new fish fauna of Middle Devonian age from the Taemas/Wee Jasper region of New South Wales. *Bureau of Mineral Resources, Australia, Bulletin* **209** (in press).

Appendix
Stratigraphic nomenclature

CRAVENS PEAK BEDS (Clarification of unit)

Derivation of name: From Cravens Peak Holding; named by Reynolds (*in* Smith, 1965).

Distribution: Glenormiston, Mount Whelan and Hay River 1:250 000 Sheet areas. The unit lies entirely along the Toomba Range and within the Toko Syncline (Fig. 2, this paper). It has been reported to occur in AOD Ethabuka No. 1 well (Mulready, 1975), a claim negated by palaeontological evidence, and at present, the Cravens Peak Beds have yet to be proved to occur south of 23° 31'S. Ordovician marine fossils, identified by Fleming (1975) from a core taken at 775 m of a sandstone unit encountered between 640 m and 1024 m in this well, indicate that this interval should be referred not to the Cravens Peak Beds, as suggested by Mulready (1975), but to the Ethabuka Sandstone (Draper, 1980). The basal (calcareous) unit is known from only two areas.

Lithology: Quartzose sandstone with mud pellets and conglomerate make up most of the unit, and minor red brown mudstones are present near its base. The conglomerates are restricted mainly to the southern portion of the Toomba Range. The basal (calcareous) unit, considered here as an informal member of the Cravens Peak Beds, comprises calcareous siltstone, limestone, calcareous sandstone, and conglomerate.

Reference section: In the reference section (Fig. 2, section 3), 100 m east of Toomba Bore, the base of the Cravens Peak Beds is recognised immediately above a 10 m wide ferruginised zone. A thick section with conglomerate (Fig. 2, section 1) is present in the southern part of the Toomba Range about 16 km SSE of Eurithethera Soak, and overlies the only known section of the basal (calcareous) unit.

Thickness: The maximum thickness known is 280 m, but nowhere is the top of the unit definitely exposed.

Contacts: The unit unconformably overlies Lower and Middle Ordovician rocks, and is overlain unconformably by Mesozoic and Cainozoic rocks and sediments. Smith (1972) suggested that it may also be overlain by Permian rocks, but the evidence for this is very tenuous, and the boulder scree, previously thought of as Permian, is probably a weathering product of the Cravens Peak Beds (Draper, 1976). Faulting and folding have resulted in complex contact relationships in the Toomba Range.

Fossils: Basal (calcareous) unit—Thelodont (*Turinia australiensis; T.* cf. *pagei; Gampsolepis?* sp. undet.), acanthodian (*Gomphonchus*? sp.; *Nostolepis* sp.), placoderm (pterichthyodid antiarch) and onychodontid crossopterygian remains; ostracods (*Healdianella inconstans*?; *Baschkirina*? sp., kloedenellaceans), and the eridostracan *Cryptophyllus* sp.

Conglomerate and sandstone sequence—*Wuttagoonaspis* sp., placoderm (including phlyctaeniids), crossopterygian (onychodontid) and acanthodian remains.

Age: late Early–? early Middle Devonian (Emsian–? Eifelian).

Discussion: Contrary to earlier investigations of the Cravens Peak Beds (Draper, 1976), the conglomerate and sandstone sequence forming the greater thickness of the unit is now known to conformably and transitionally overlie the basal (calcareous) unit.

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Discrimination of surficial and bedrock magnetic sources in the Cobar area, NSW

Peter R. Gidley

Known mineral deposits in the Cobar area of NSW are magnetic. Hence, the magnetic method is an important tool in the search for mineral deposits in this area. Unfortunately, widespread occurrences of near-surface magnetic mineral accumulations commonly produce aeromagnetic anomalies similar to those expected from bedrock sources. Work by BMR during 1978 and 1979 has shown that the near-surface response is caused by laterite, which contains concentrations of hematite/maghemite minerals with high magnetic remanence and/or susceptibility. These magnetic minerals usually occur in surface layers up to 10 m thick and exhibit large lateral and vertical variations in magnetic properties over small distances. Frequency domain analysis of both model and field magnetic data shows the spectral signatures of bedrock and near-surface sources to be distinctly different. Hence, spectral techniques may be used to separate the responses of bedrock and near-surface sources which occur at the same locality, if high resolution data are available. A comparison of the response of near-surface and bedrock sources by modelling indicates that, with careful survey design, a magnetic target such as Elura could be detected at a depth of 300 m, even in areas of severe surface magnetic noise.

Introduction

In the Cobar area of NSW, surficial deposits of magnetic laterite minerals are widely developed and produce prominent aeromagnetic and ground magnetic anomalies which may mask or be mistaken for anomalies from bedrock sources. During 1978 and 1979 the Bureau of Mineral Resources (BMR) investigated the application of methods able to discriminate between anomalies produced by surface and bedrock magnetic sources, and detect mineral deposits lying beneath surficial sources. Investigations undertaken by BMR included airborne, ground, and downhole magnetic surveys, with laboratory physical property measurements on core samples. Previous BMR work relevant to this problem has been described by Wilkes (1979), Gidley & Stuart (in press) and Gidley (in prep.).

Characteristics of airborne anomalies

Airborne magnetic surveys in the Cobar area are usually flown at 70-150 m above ground level. However, at these heights many surficial and bedrock sources cannot be distinguished by analysis of anomaly shape and amplitude.

Response to bedrock sources

Figure 1 shows the results of an aeromagnetic survey over the Elura lead-zinc deposit flown for the Electrolytic Zinc Co. of Australasia Ltd in 1974 (from Wilkes, 1979). The survey was flown at a height of 90 m above ground level with a line spacing of 300 m. The magnetic anomaly is located directly over the deposit and shows a near-circular anomaly with an amplitude of approximately 40 nT. The results of modelling a north-south aeromagnetic profile flown by BMR across the deposit at a ground clearance of 90 m are shown in Figure 2 (after Gidley & Stuart, in press). A good fit to the observed profile is provided by the response of a magnetic prism comprising a core of moderate susceptibility and a substantial amount of remanence parallel to the Earth's field, and an outer sheath of low susceptibility. This model is reasonably consistent with drilling and physical property measurements, which indicate the deposit contains an inner core of magnetic pyrrhotite with substantial remanence (Gidley & Stuart, in press).

It is important to note that a good fit to the Elura anomaly is also provided by an essentially surface model, as indicated by model 2 in Figure 2. In this model fifteen small three-dimensional blocks with a susceptibility of 150 000 x 10^{-6}SI are used to represent a shallow inhomogeneous source which approximates a surficial accumulation of magnetic laterite minerals. The similarity between the surface and bedrock source aeromagnetic responses is remarkable and illustrates the difficulty in distinguishing between near-surface magnetic sources and bedrock sources in the Cobar area.

Response of surficial sources

A typical example of a near-surface magnetic source occurs in the Shearlegs area, 40 km south of Cobar. The results of an aeromagnetic survey over this area at a ground clearance of 70 m and a line spacing of 400 m are shown in Figure 3 (from Wilkes, 1979). Although flight line spacing and data processing have contributed to the contour pattern in this case, the observed magnetic closures with amplitudes up to 65 nT could easily be interpreted as being due to a bedrock source. However, a substantial program of surface and downhole work (Gidley, in prep.) has shown that the anomalies are produced by upper soil horizons containing up to 40 percent by volume of magnetic minerals.

The results of modelling an aeromagnetic profile flown by BMR across the anomaly at a ground clearance of 90 m are shown in Figure 4. This profile was flown along the line A-B shown in Figure 3. The Shearlegs profile is similar in shape and amplitude to the anomaly observed over the Elura deposit and, again, an Elura-style bedrock source (model 1 in Fig. 4) can provide a good fit to the observed anomaly. The detailed structure of the surface model (model 2 of Fig. 4) was derived by forward modelling a series of twenty discrete magnetic cells to fit ground magnetic data acquired by BMR beneath the aeromagnetic profile along line A-B (see Fig. 3).

Aeromagnetic surveys over other areas with surficial concentrations of magnetic minerals also illustrate this

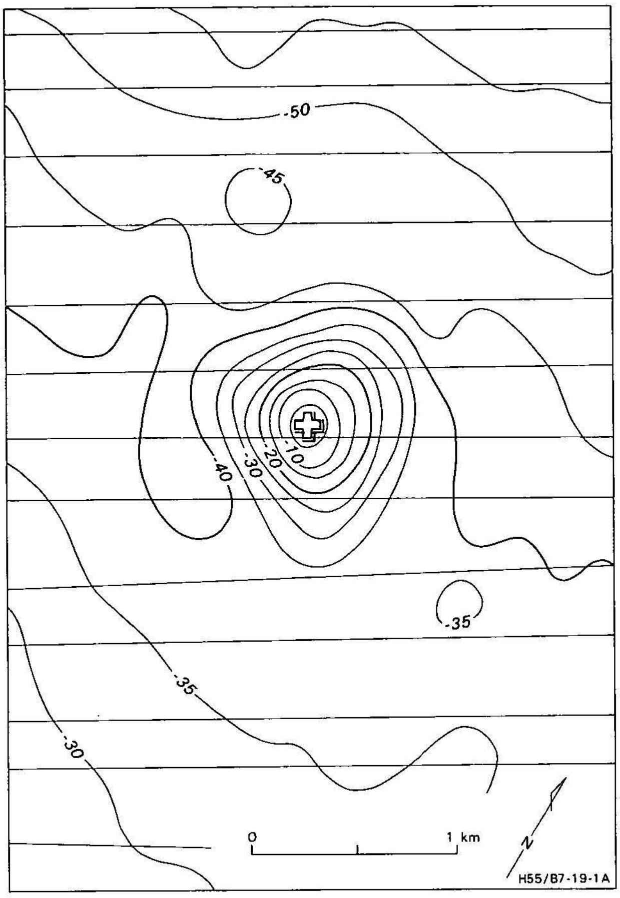

Figure 1. Aeromagnetic contour map of the Elura area obtained from a survey with 90 m ground clearance and 300 m line spacing (from Wilkes, 1979).

ambiguity. For instance, the results shown in Figure 5 (from Wilkes, 1979) were recorded at Burri, 52 km northwest of Cobar in a survey flown at 90 m ground clearance and 300 m line spacing. Interpretation of this near circular anomaly of amplitude 30 nT as having a bedrock source at a depth of approximately 400 m would be entirely reasonable (Wilkes, 1979). However, extensive drilling failed to locate a bedrock source, but indicated surficial lateritic concentrations similar to the Shearlegs case (Gidley, in prep.). A near-surface source, similar to model 2 of Figure 4, can be developed to model the observed data.

Characteristics of ground magnetic anomalies

Normally in mineral exploration, aeromagnetic anomalies are followed up by ground survey techniques. However, the ubiquitous accumulation of magnetic minerals in the soils of the Cobar area produces high-amplitude short-wavelength spikes which may distort or make unrecognisable the response of deeper sources.

The effect of surficial magnetic sources in the Cobar area is illustrated in Figure 6 which shows the results of eleven east-west BMR car-borne magnetometer traverses across the Elura deposit (from Gidley & Stuart, in press). The car-borne survey system consists of a Geometrics G803 proton precession magnetometer, and results are recorded digitally. The detector is carried 20-30 m behind the vehicle to reduce noise and a sampling rate of 1.0s gives a reading interval of approximately 0.3 m for a survey speed of 1.0 km per hr.

The stacked profiles shown in Figure 6 reveal the widespread presence of high-amplitude short-wavelength anomalies at Elura. However, the longer wavelength response due to the Elura deposit centred on 2450E/50800N is still discernible.

At sites such as Shearlegs, where magnetic laterites are strongly developed, very high-ampltiude short-wavelength anomalies are recorded on the ground. These make the detection of possible bedrock sources difficult, and survey design is very important if a true representation of the magnetic field is to be obtained. For example, the two profiles presented in Figure 7 (after Wilkes, 1979) indicate the difference in anomaly shape which may be recorded at Shearlegs when the measurement interval is increased from 5 m to 25 m. The coarse 25 m sampling interval records an apparent long-wavelength component, owing to the inadequate sampling of high-amplitude short-wavelength components. This problem will affect interpretation and further processing procedures.

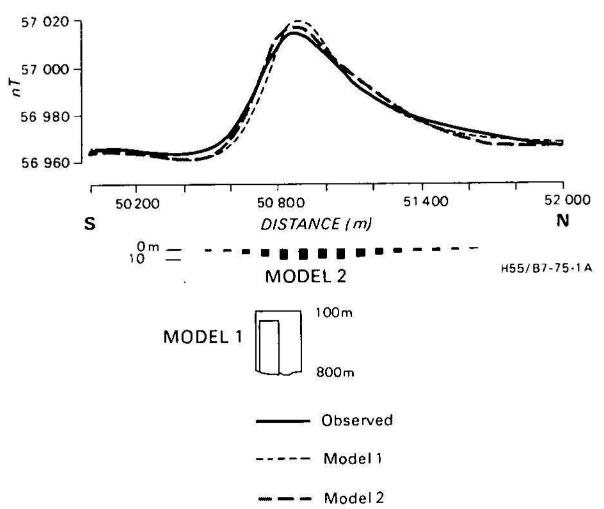

Figure 2. Complex source models for the Elura aeromagnetic anomaly (after Gidley & Stuart, in press).
The outer sheath in model 1 has a magnetic susceptibility of 1000 × 10^{-6} SI and no remanence. The inner prism has a susceptibility of 32 000 × 10^{-6} SI and Koenigsberger ratio of 3.8 with remanent vector parallel to the Earth's magnetic field.

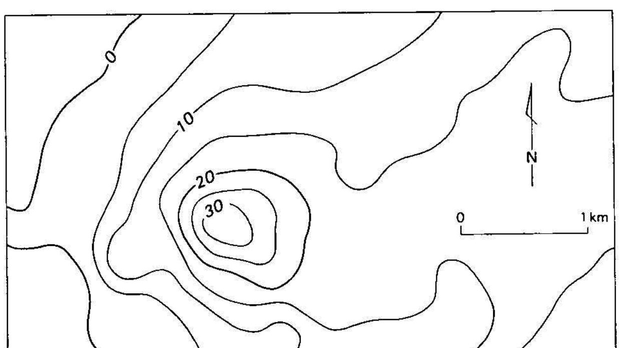

Figure 3. Aeromagnetic contour map of the Shearlegs area, obtained from a survey with 70 m ground clearance and 400 m line spacing (from Wilkes, 1979).

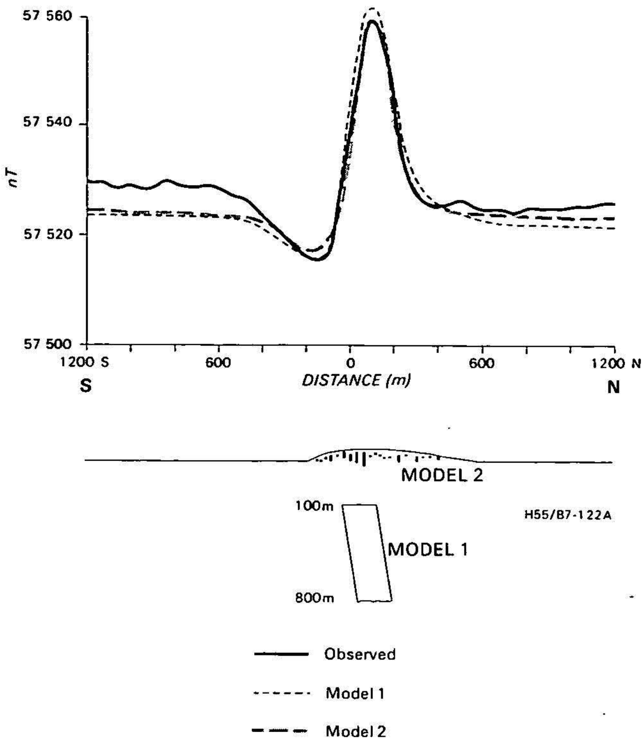

Figure 4. Complex source models for the Shearlegs aeromagnetic anomaly, in profile form along A-B for Fig. 3.
The bedrock source model is a two-dimensional dyke with 80° dip to the south and magnetic susceptibility of 30 000 × 10^{-6} SI.

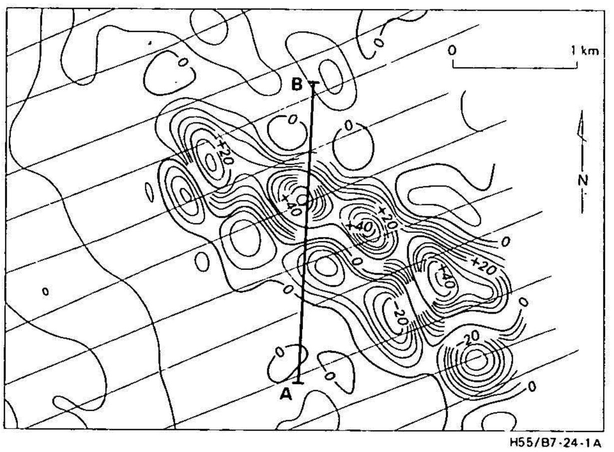

Figure 5. Aeromagnetic contour map of the Burri area obtained from a survey with 90 m ground clearance and 300 m line spacing (from Wilkes, 1979).

The geology and physical character of surficial sources

To study the geophysical behaviour and characteristics of surficial magnetic sources it was necessary to examine drilling results and laboratory analyses from sites where surface accumulations of magnetic minerals are extensively developed. Results of this work from Shearlegs are shown in Figure 8. Shearlegs lies in an area of undifferentiated Cobar Supergroup rocks (Pogson & Felton, 1978), consisting of red and yellow siltstone with some claystones, shales, and argillaceous and quartzitic sandstones. Eleven rotary air-blast holes were drilled by CRA Exploration at 50 m intervals over the main magnetically disturbed zone along the line A-B shown in Figure 3. The holes ranged in depth from 5.5 to 14 m and were logged, as shown in Figure 8, for lithology, magnetic susceptibility and the intensity of the vertical magnetic field.

The geological section reveals a 2-4 m thick layer of iron-rich soil lying upon an east-west trending ridge. BMR X-ray diffraction work indicates that the predominant mineral is quartz, with accessory minerals kaolinite, illite and mica (Gidley, in prep.). Of the iron minerals present, approximately 75 percent is hematite and the remainder is highly magnetic maghemite. To the south of the main maghemite development are low-susceptibility hematitic soils.

The magnetic susceptibility logs were obtained in the laboratory from drilling samples. The maximum susceptibility measured was 250 000 x 10^{-6}SI. The susceptibility in all holes decreases with depth to typical bedrock values of about 400 x 10^{-6}SI. The vertical field magnetic log was measured with a BMR-constructed

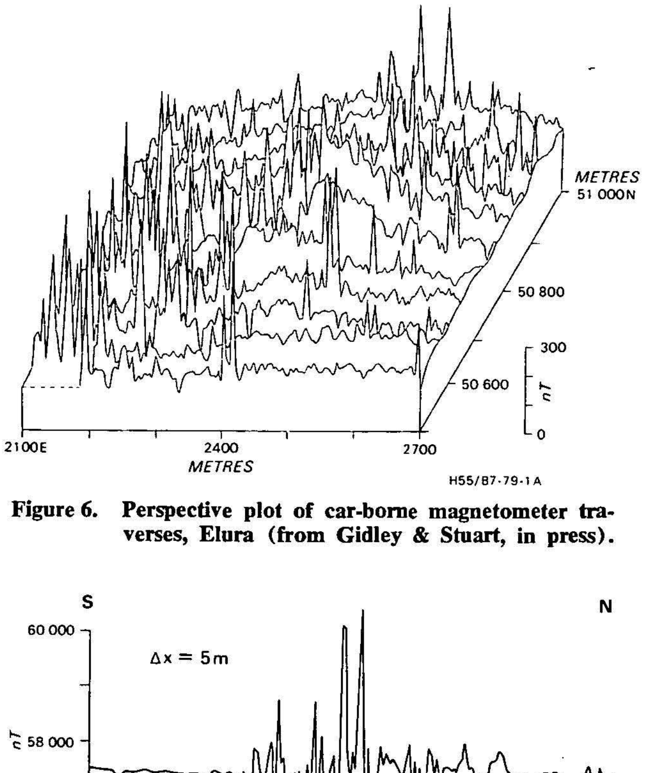

Figure 6. Perspective plot of car-borne magnetometer traverses, Elura (from Gidley & Stuart, in press).

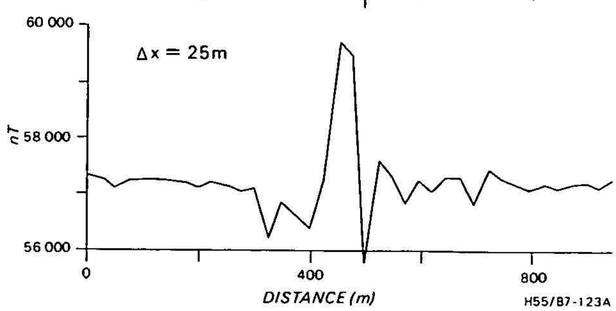

Figure 7. Total field magnetic traverses with 5 m and 25 m station intervals in the Shearlegs area (after Wilkes, 1979).

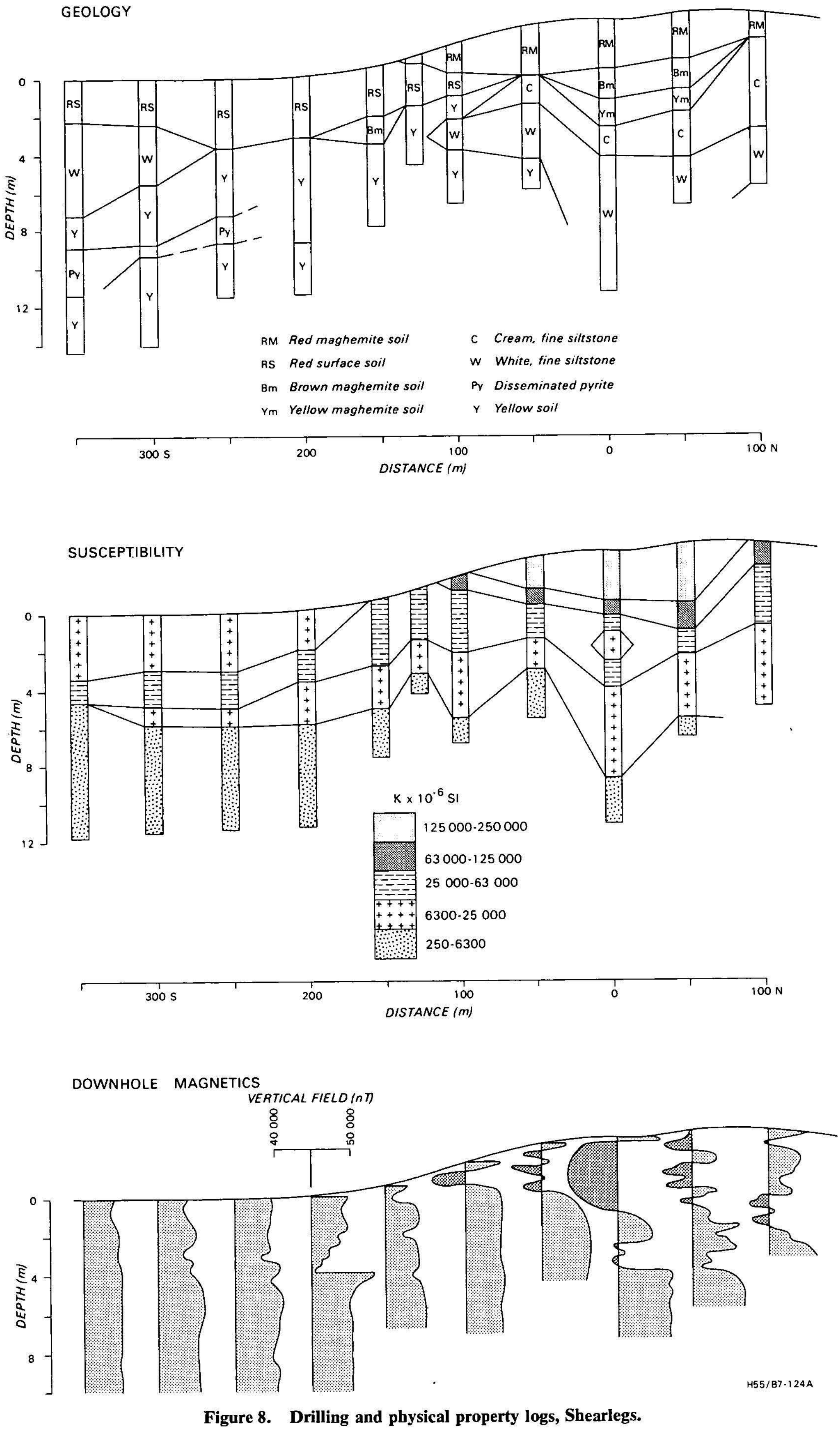

Figure 8. Drilling and physical property logs, Shearlegs.

downhole fluxgate magnetometer. Measurements were recorded at 0.25 m intervals and showed extremely high magnetic gradients over small distances, which indicate the rapid vertical variations in magnetic properties. A good correlation between boundaries observed in the susceptibility, geological and vertical magnetic field logs is demonstrated in Figure 8.

Remanence measurements made on soil samples collected 1-2 m below surface from auger sample holes resulted in Koenigsberger ratios up to 92, indicating that remanence, as well as high susceptibility, is very important in producing disturbed magnetic fields associated with the near-surface magnetic sources (Gidley, in prep.).

Based on the geological control afforded by these studies along line AB (see Fig. 3), a three-dimensional magnetic model was developed for the Shearlegs area. As shown in Figure 9 the response of this model at ground level, 20 m and 60 m above ground level fits closely to the results observed by BMR along AB. It is important to note that for a source of this type the high-frequency component is evident in airborne surveys below 20 m altitude. Above this altitude the individual elements of the model are not seen, and the long-wavelength bulk effects dominate the spectrum.

Spectral characteristics of sources

The results of ground and airborne studies indicate that the magnetic response of surface and bedrock sources, and consequently the spectral responses, are distinctly different at low flying altitudes provided sufficient resolution of the data is obtained. The one-dimensional spectral character can be obtained by transforming the magnetic data from the spatial domain to the frequency domain by Fourier transform. The result of this spectral analysis can be plotted as a power spectrum as shown in Figure 10 (from Gidley & Stuart, in press), where frequency is plotted in cycles per km

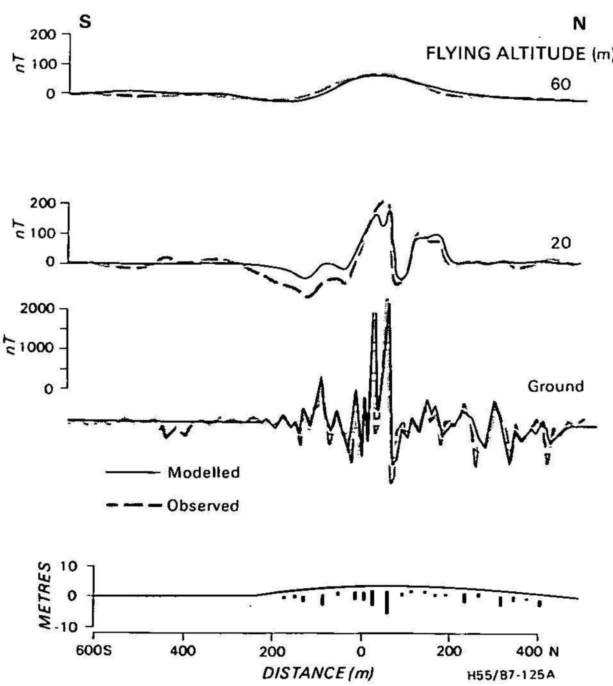

Figure 9. Surficial source magnetic model, Shearlegs. Twenty near-surface discrete bodies were used to model the ground profile and then the profiles were calculated for 20, 40 and 60 m altitudes.

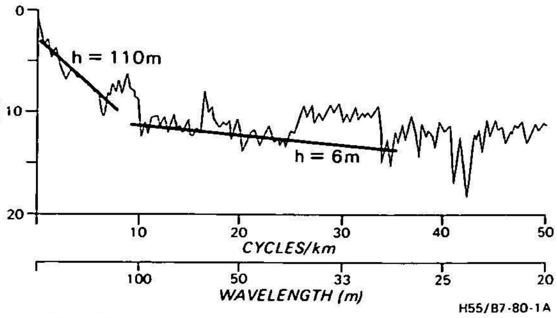

Figure 10. Spectral analysis of car-borne magnetic data, Elura (from Gidley & Stuart, in press).

along the horizontal axis and the normalised power in decibels is plotted on the vertical axis. Examination of power spectra presented in this form shows the relative contribution of different wavelength signals to the original magnetic data. The analyses of frequency domain data can provide information on depth to source, number of sources present, and, possibly, the size and spacing of magnetic cells. As explained by Spector (1968) and Spector & Grant (1970), the attenuation of spectral power with frequency depends on a number of factors including the depth of the source. For tabular sources the plot of log spectral power versus linear wavelength is commonly a straight line, the slope of which can be directly used to estimate the depth to the magnetic source. For spectral analysis to be carried out, the magnetic anomaly must be adequately sampled. As was illustrated in Figure 7, inadequate sampling of high amplitude, short wavelength components of the anomaly will introduce erroneous long wavelength components into the data.

An illustration of the use of spectral analysis is seen in the power spectra derived from Elura and Shearlegs data. The power spectrum of Elura obtained from car-borne magnetometer data is shown in Figure 10 (from Gidley & Stuart, in press) and suggests two gradients and, therefore, two magnetic ensembles. The first of these occurs in the low frequency part of the spectrum, from 0.5 to approximately 9 cycles per km, and is represented by a steep gradient indicative of a source approximately 100 m deep. The similarity between this estimate and the known depth of the Elura deposit is remarkable and indicates that this part of the spectrum is largely due to the magnetic field anomaly in this area caused by the Elura deposit. The second gradient is less well defined, but is observed between

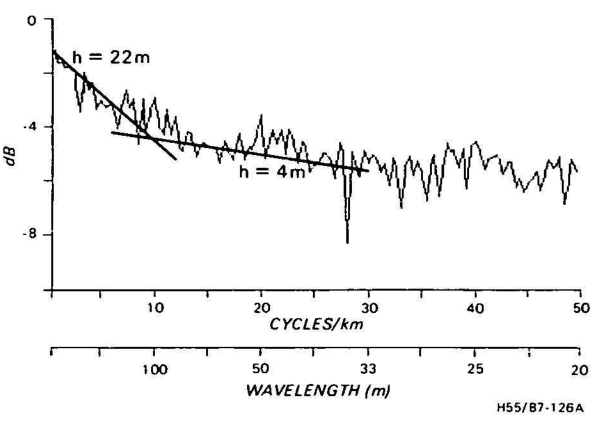

Figure 11. Spectral analysis of observed ground magnetic data, Shearlegs.

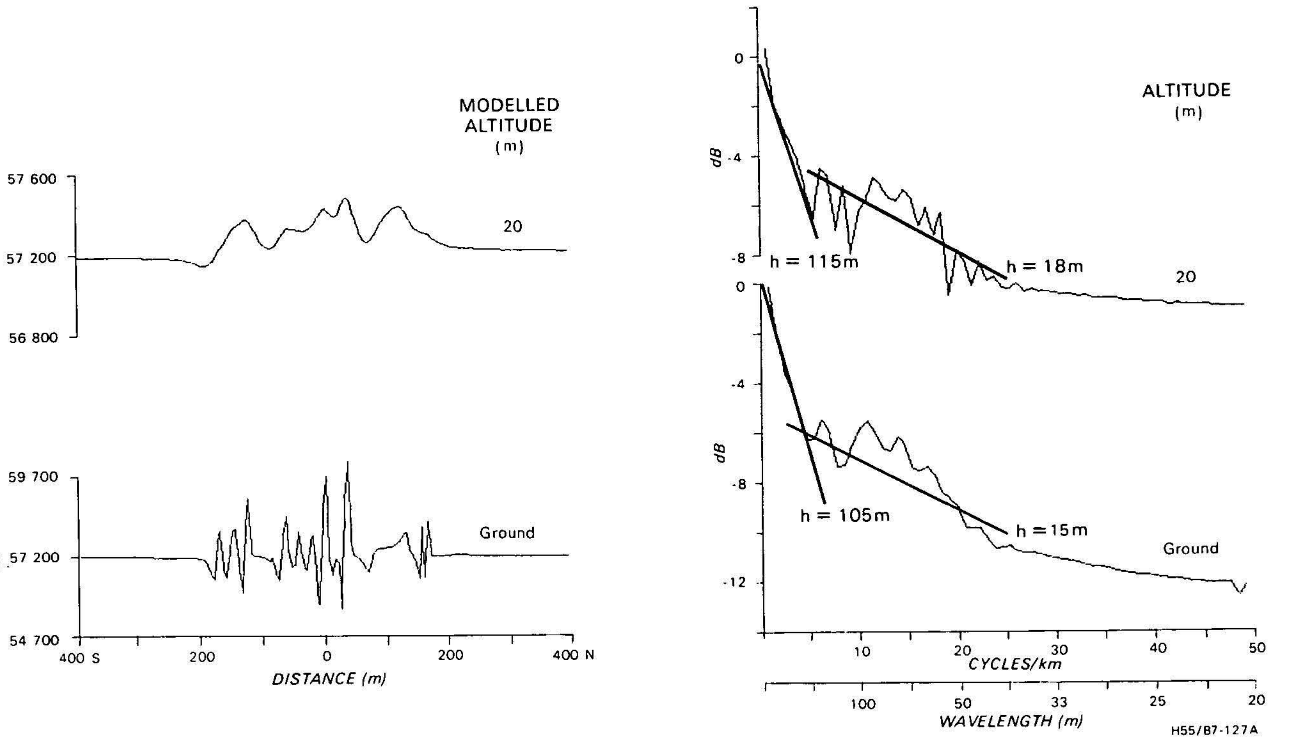

Figure 12. Profiles and spectral plots of the response of a Shearlegs surface model combined with an Elura-type bedrock model.

9 cycles per km and about 36 cycles per km. The depth to the source or sources producing this part of the power spectrum is certainly shallow, and is estimated to be approximately 6 m. Hence, this gradient would appear to characterise the near-surface source at Elura, and the peaks at 9, 17, and approximately 30 cycles per km may indicate the size of individual near-surface magnetic cells. Beyond 40 cycles per km the power spectrum degenerates into noise and cannot be meaningfully interpreted.

The Shearlegs ground spectrum presented in Figure 11 is indicative of a spectrum obtained from a combination of near-surface sources. The spectrum can be fitted by a series of straight lines whose gradients represent a depth range between 4 and 22 m which is consistent with the presence of a near-surface source. The advantage of making depth estimates in the frequency domain rather than the spatial domain is related to the fact that in the frequency domain all data are used simultaneously in the analyses. In the spatial domain, depth interpretation of long wavelength features is likely to be less reliable, owing to the need to remove high frequency components before analyses.

Discrimination of combined surface and bedrock sources

Model data

The spectral characters of anomalies due to surface and bedrock sources in the Cobar area appear to be markedly different. Hence, methods of spectral analysis may solve the problem which exists when a surface magnetic source overlies a bedrock source. An example of this problem is shown in Figure 12 where a surface model of the Shearlegs area is combined with a bedrock model of Elura centred at 0. At and below 20 m ground clearance the short wavelength components due to the presence of the surface magnetic sources dominate the spatial domain data, and it is almost impossible to identify the response of the bedrock source.

A spectral analysis of the combined sources for each modelled altitude is also presented in Figure 12. The spectra for ground and 20 m ground clearance data show a change in gradient, indicating two magnetic sources contribute to the data. For the ground spectrum, the gradient of the long wavelength feature, from 1 to 6 cycles per km, indicates a depth to source of approximately 105 m, which is similar to the depth to the modelled bedrock source. The second ensemble becomes prominent at about 7 cycles per km and the slope of this ensemble response indicates a near-surface source. At approximately 18 cycles per km the power due to the near-surface source decays rapidly to a smooth low-noise response.

As modelled height increases, the slopes of both ensembles increase and merge. The results of this analysis highlight the need for both high spatial resolution and low-level data for the successful discrimination of sources.

Field data with severe noise

The final test of this discrimination method was to apply the spectral analysis technique to field data. The observed profile and drilling evidence suggest that the Shearlegs area represents a concentration of maghemite as severe as found anywhere in the Cobar region. A spectral analysis was undertaken of the observed ground data from Shearlegs combined with a model of Elura. These results are shown in Figure 13. The spatial domain profiles are also shown for the ground traverse over Shearlegs, the Elura model, and for the sum of these two profiles.

The frequency domain spectra shown in Figure 13 reveal the nearly flat gradient of the Shearlegs ground data and suggest a combination of near-surface sources. The Elura model suggests a single source about 100 m deep.

The spectrum derived from the combined profiles shows a noticeable change in decay rate at 4 cycles per km and thus indicates two sources at different depths. Depth interpretation based on the two gradients indicates the source manifest in the low frequency part of the spectrum is about 100 m deep, while the other gradient indicates a shallow depth of approximately 6 m. This result demonstrates the capability of spectral analyses to discriminate between surficial and strong bedrock sources.

Discrimination of weak bedrock sources in the presence of strong surficial sources

The limit of detection of a bedrock source beneath a surface magnetic source by spectral analysis was determined by modelling the combined Shearlegs and bedrock models with different physical constraints of bedrock depth and susceptibility. In Figure 14 the three spectra shown were produced by combining the ground magnetic response for the Shearlegs near-surface model with the Elura-type bedrock model at depths of 100, 200, and 400 m. As depth to the bedrock source increases, both the slope of the bedrock response in the frequency domain plot and the relative power of the surficial source response increase. Finally, for a bedrock depth of approximately 400 m the two responses merge and interpretation of the data suggests that only a single shallow source is present.

Figure 15 shows a second set of spectra generated by combining the response of a Shearlegs-type surface model and an Elura-type bedrock model, where the depth to the bedrock source is kept constant at 100 m, but magnetic susceptibility varied from 30 000 x 10^{-6} SI. When the bedrock model has a susceptibility of 30 000 x 10^{-6} SI the existence of two sources is apparent. Such discrimination remains possible until the bedrock source susceptibility is reduced to about 10 000 x 10^{-6} SI. Note that as the susceptibility has been altered, the frequency content has remained constant and only the relative power of the two responses has changed.

Conclusions

Widespread occurrences of near surface magnetic mineral accumulations provide a source of ambiguity

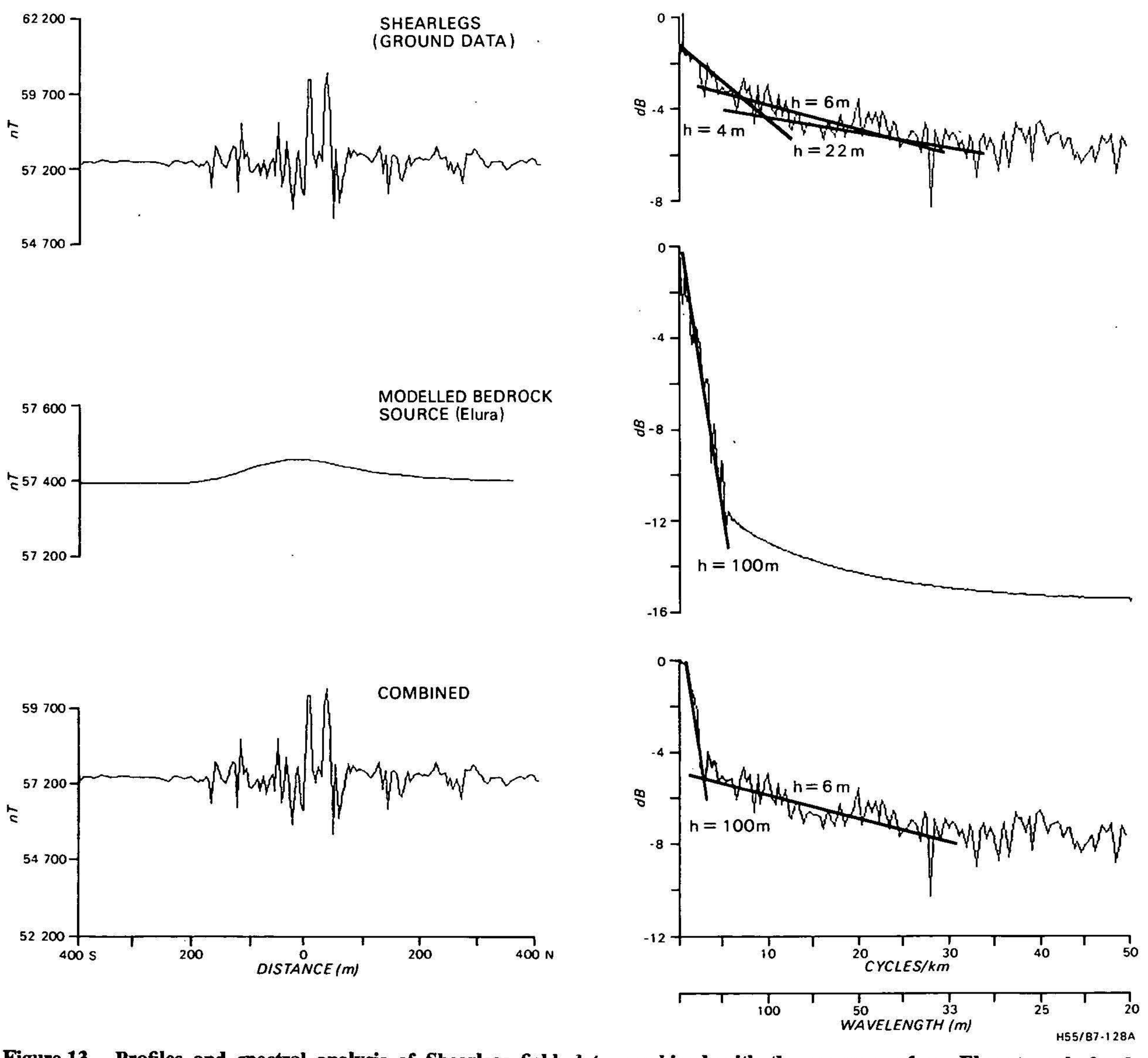

Figure 13. Profiles and spectral analysis of Shearlegs field data combined with the response of an Elura-type bedrock model.

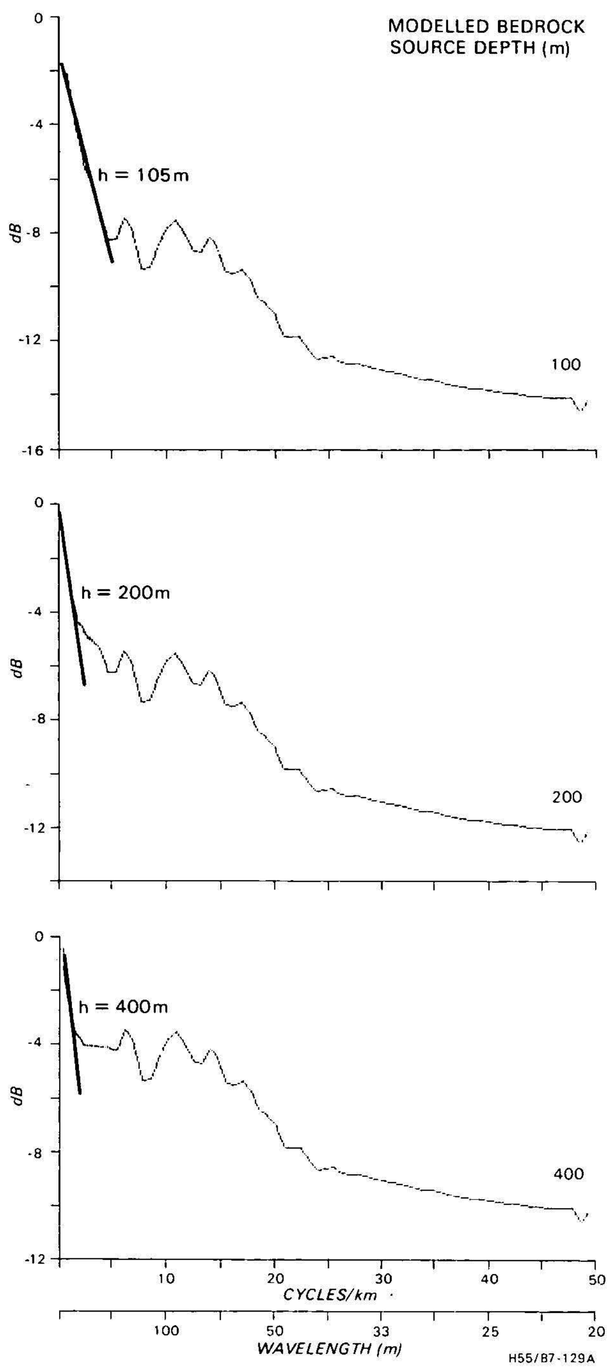

Figure 14. Spectral analysis of the response of a Shearlegs surface model combined with Elura-type bedrock model at depths of 100, 200 and 400 m.

in interpretation of magnetic data in the Cobar area. Airborne, ground, laboratory and downhole magnetic studies have defined the character of the surface magnetic sources and shown them to be due to accumulations of maghemite and hematite which have a high susceptibility and remanence. These magnetic sources usually occur in surface layers up to 10 m thick and exhibit rapid lateral and vertical variations in magnetic properties and concentration over small distances.

Frequency domain analysis shows the spectral signatures of bedrock and surface sources to be distinctly different. Hence, spectral analysis can be used to separate the response of bedrock and surficial sources. The results of these investigations suggest that a target such as Elura can be recognised amid even severe noise by the spectral analyses of high-resolution ground surveys or very low-level aeromagnetic surveys. However, limits to the applicability of the method do exist, and these may be determined by modelling. For example, it may not be possible to detect the presence of a bedrock source such as Elura beneath a near-surface magnetic source if the deposit were at a depth of 400 m. Similarly, a surface source could preclude the detection of an Elura-type bedrock source at a depth of 100 m if the bedrock source susceptibility was as low as 10 000 x 10^{-6} SI.

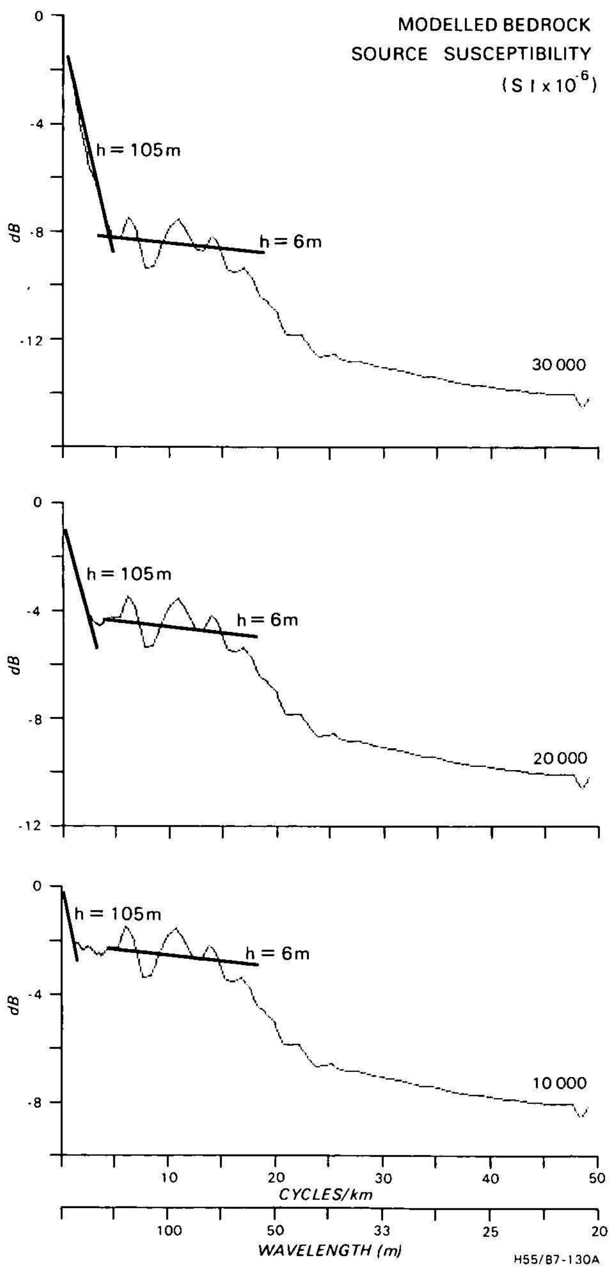

Figure 15. Spectral analysis of the response of a Shearlegs surface model combined with Elura-type bedrock model at depth of 100 m and having a susceptibility of 30 000, 20 000 and 10 000 × 10^{-6} SI.

References

GIDLEY, P. R. (in prep.)—Geophysical characteristics of some surficial magnetic sources near Cobar, NSW. *Bureau of Mineral Resources, Australia, Report.*

GIDLEY, P. R., & STUART, D. C., in press—Magnetic property studies and magnetic surveys of the Elura prospect, Cobar, NSW. *Bulletin of the Australian Society of Exploration Geophysicists.*

POGSON, D. J., & FELTON, E. A., 1978—Reappraisal of geology, Cobar-Canbelego-Mineral Hill Region, Central Western NSW. *Geological Survey of New South Wales Quarterly Notes,* **33**.

SPECTOR, A., 1968—Spectral analyses of aeromagnetic data. *Ph.D. Thesis University of Toronto, Canada* (unpublished).

SPECTOR, A., & GRANT, F. S., 1970—Statistical models for interpreting aeromagnetic data. *Geophysics,* **35**, 293-302.

WILKES, P. G., 1979—Characteristics of magnetic sources and guidelines for exploration in the Cobar area, NSW. *Bulletin of the Australian Society of Exploration Geophysicists,* **10**(1), 34-41.

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The Babbagoola Beds, Officer Basin, Western Australia: correlations, micropalaeontology and implications for petroleum prospectivity

M. J. Jackson & M. D. Muir[1]

Lithological and palaeontological comparisons suggest correlation between the Babbagoola Beds in Western Australia and the Observatory Hill Beds of South Australia, which are considered to be the major potential source of petroleum in the eastern Officer Basin. The presence of potential petroleum traps, such as salt domes and broad anticlines, and good potential reservoir beds, such as the Lennis Sandstone and Wanna Beds in Western Australia, is now complemented by the suggestion of a potential early Cambrian petroleum source bed. Further petroleum exploration and stratigraphic drilling are warranted.

Introduction

This paper elaborates on ideas presented at the 8th BMR Symposium in May 1979, where lithological and micropalaeontological comparisons between drillholes in the South Australian and Western Australian parts of the Officer Basin were used to upgrade the petroleum potential of the Western Australian part of the basin (Jackson, 1979). A synthesis of the regional stratigraphy of the basin in Western Australia is also presented, as none is yet available for this little known and inaccessible area. This will complement a recent review of the South Australian part of the basin, in which stratigraphic correlations, drilling, and petroleum prospects have been discussed (Pitt & others, 1980). A detailed description of the geology of the Western Australian part of the Officer Basin is given in Jackson & Van de Graaff (in press).

Regional stratigraphy

In Western Australia, the term 'Officer Basin' is used to describe the mainly sedimentary rocks, south of the Canning Basin and north of the Eucla Basin, that fill a deep depression between the Archaean Yilgarn Block and the Proterozoic Musgrave Block (Fig. 1). In it, a generally slightly folded and faulted sequence, several kilometres thick, of Precambrian and Early Palaeozoic rocks is overlain by relatively thin and horizontal Permian, Mesozoic, and Cainozoic sediments.

Despite the extensive cover of Cainozoic soils, the flat-lying nature of many outcrops, and lack of relief coupled with a lack of subsurface information, it is now possible to make a broad synthesis of the stratigraphic evolution of this large region of Western Australia, and attempt correlations with sequences in nearby basins.

In a composite stratigraphic column (Fig. 2) four rock sequences are recognised. Following Western Australian practice (as adopted by Jackson & van de Graaff, in press), the oldest is assigned to the basement and the other three are defined as forming the Officer Basin:

Late Palaeozoic and younger—thin flat-lying;

Early Palaeozoic—thin flat-lying;

Younger Proterozoic to earliest Cambrian—thick gently folded;

Older Proterozoic—thick unexposed, largely unknown, layered sequence (basement).

This basin definition contrasts with that adopted by the South Australia Department of Mines and Energy, who would omit the Late Palaeozoic and younger units from the Officer Basin and include them in the Canning Basin sequence with which they are continuous.

The distribution of these units, the regional structural setting of the basin, and the distribution of the formations that make up the younger Proterozoic to earliest Cambrian sequence are shown in a block diagram in Figure 3. Outcrops of some of the younger formations are shown in Figure 4.

The *older Proterozoic* rocks do not crop out, but have been identified on a Hunt-BMR seismic cross-

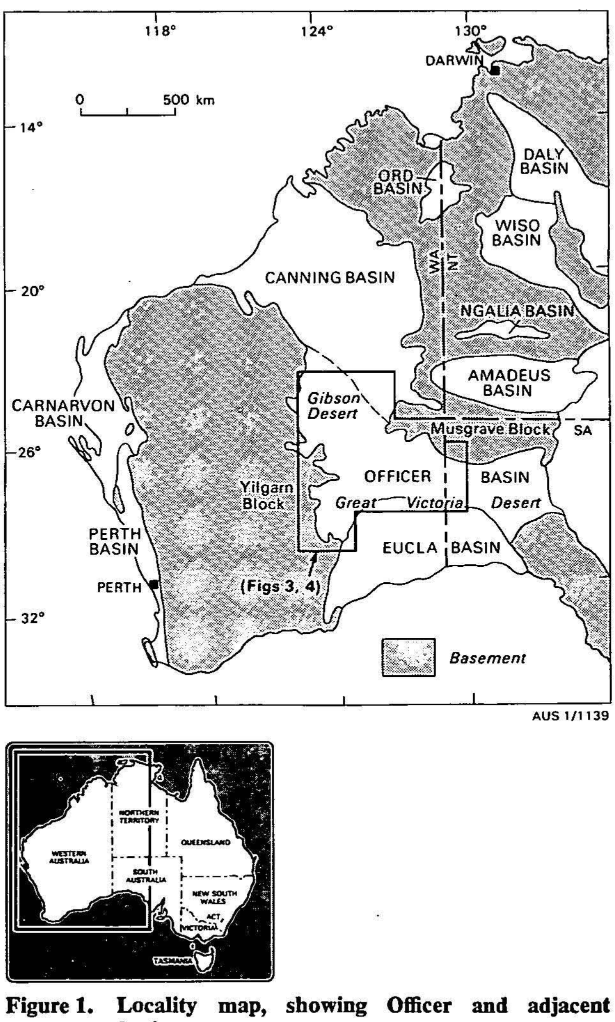


Figure 1. Locality map, showing Officer and adjacent basins.

1. **Present address: CRA Exploration Pty Ltd, P.O. Box 656, Fyshwick, ACT 2609.**

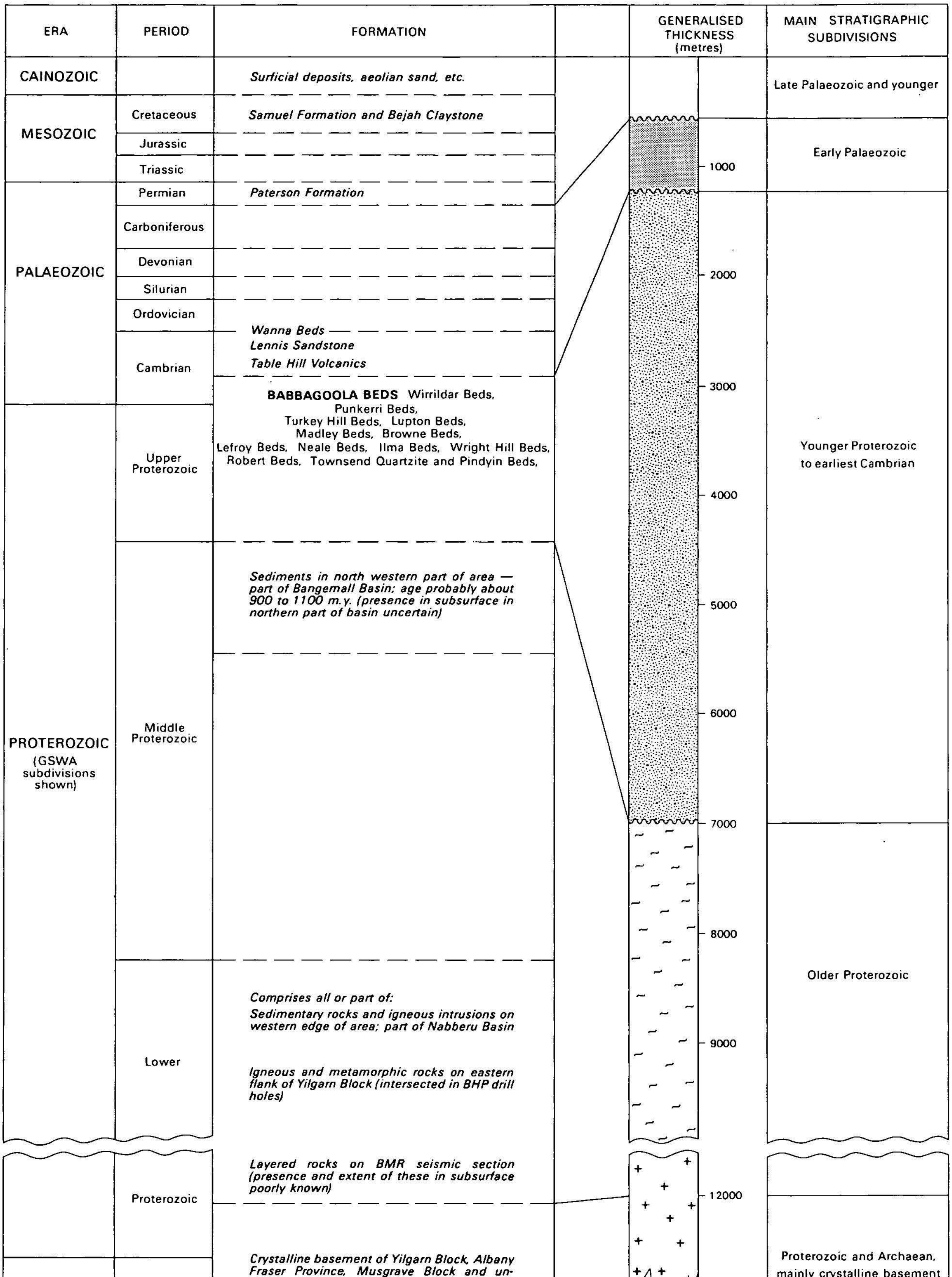


Figure 2. Composite stratigraphic column, showing ages and approximate thicknesses of the major stratigraphic subdivisions within the Officer depression.

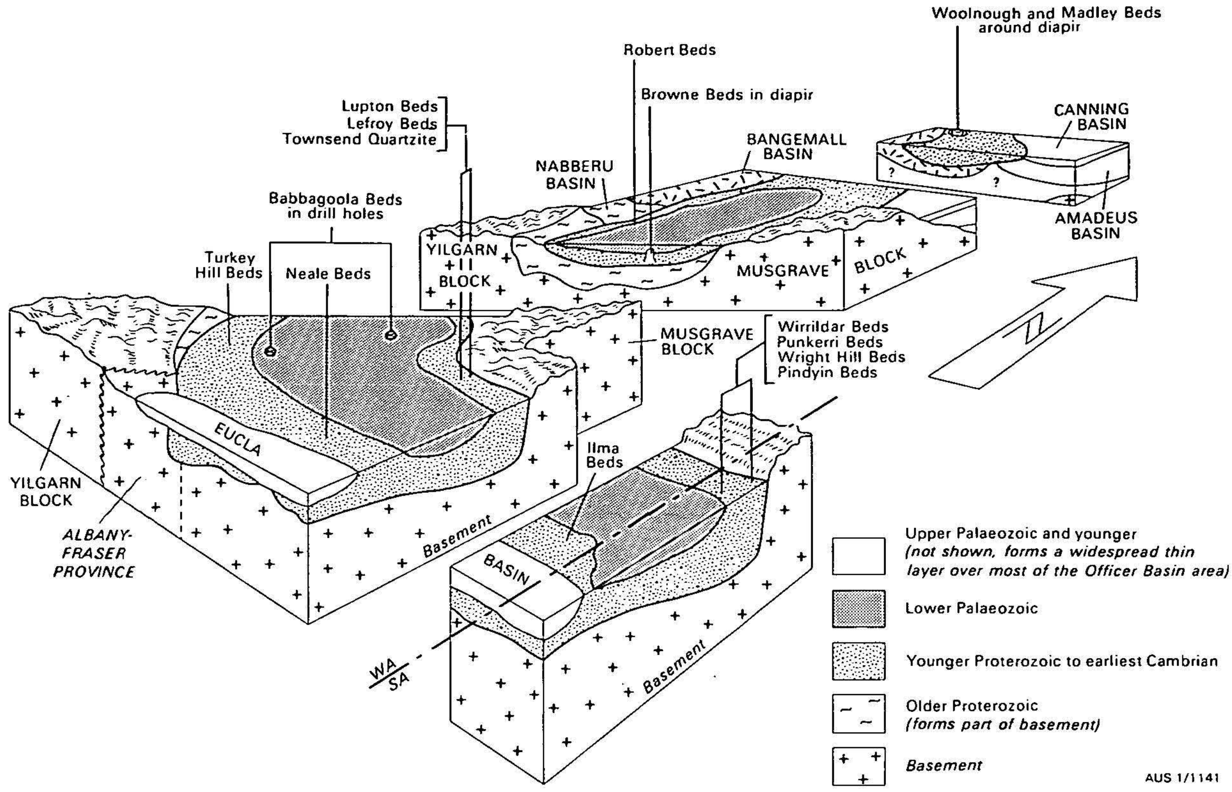


Figure 3. Interpreted distribution of major stratigraphic/structural subdivisions of the Officer Basin and adjacent blocks and basins.

section (A-B on Fig. 4) and on BMR seismic traverses along the Gunbarrel Highway in the north of the basin. The sequence is layered and includes igneous, metamorphic, and sedimentary rocks. Crystalline rocks were intersected at shallow depth in the BHP drillholes in the west of the area (Fig. 4). The sedimentary rocks are likely to be lateral equivalents of the Nabberu and/or Bangemall Basins, and hence are excluded from the Officer Basin.

The *younger Proterozoic to earliest Cambrian* sequence, up to about 6 km thick, with seismic velocities between 3.3 and 5.7 km/s overlies the older Proterozoic and probably forms the bulk of the Officer Basin sequence throughout Western Australia (Fig. 3). The layering within it is markedly discordant to that of the underlying older Proterozoic. On some seismic records the base of it marks a décollement surface from which diapiric structures appear to originate. Along the seismic cross-section the younger Proterozoic is thickest in the northeast and has a steep northeast boundary and more gentle southwest margin.

The lowermost unit of the younger Proterozoic to earliest Cambrian sequence is variously called the *Townsend Quartzite* (along the southern flank of the Musgrave Block in WA), the *Pindyin Beds* (in the identical tectonic setting in SA) and the *Robert Beds* (along the western margin of the basin) (Fig. 3). It is a shallow marine to deltaic basal quartz arenite, up to about 200 m thick, deposited unconformably on older rocks during a marine transgression. Though gradually subsiding, the area appears to have remained close to sea level for a considerable time, as succeeding units accumulated mainly in shallow-marine waters, occasionally under evaporitic conditions. The *Wright Hill Beds* (in the northeast), *Ilma* and *Neale Beds* (in the south) and *Browne* and *Madley Beds* (in the central and northern parts, respectively) contain shallow-water features such as stromatolitic and oolitic carbonates and intraclast beds, and the last two units contain evaporites. Some intervals of quieter conditions in deeper water are, however, indicated by the presence of well-bedded siltstone and shale (for example the *Lefroy Beds*).

A period of mild tectonism is indicated, at least in the vicinity of the Musgrave Block, as rocks from there and the Townsend Quartzite are incorporated in the overlying fluvioglacial *Lupton Beds*, which accumulated along the southern flank of the Musgrave Block in Western Australia. Similar fluvioglacial rocks and associated shallow-water clastics (*Turkey Hill Beds*) are poorly exposed close to the southwest margin of the basin.

After the glaciation, the area was again characterised by mainly shallow-water epeiric sedimentation, and a poorly known, apparently conformable sequence, 2000-3000 m thick, was laid down. The *Punkerri Beds* and *Wirrildar Beds* form the upper part of this sequence in the extreme east of Western Australia, but they are better exposed further east in South Australia. The Punkerri Beds contain minor carbonate, but consist mainly of shelf sands, with a possible Ediacara fauna indicating a latest Precambrian age. The uppermost part of the younger Proterozoic to earliest Cambrian subdivision in the central part of the basin is represented by the shales, dolomites, sandstones, and evaporites of the *Babbagoola Beds* intersected in drillholes Hunt Yowalga 2 and BMR Throssell 1, and described in detail below.

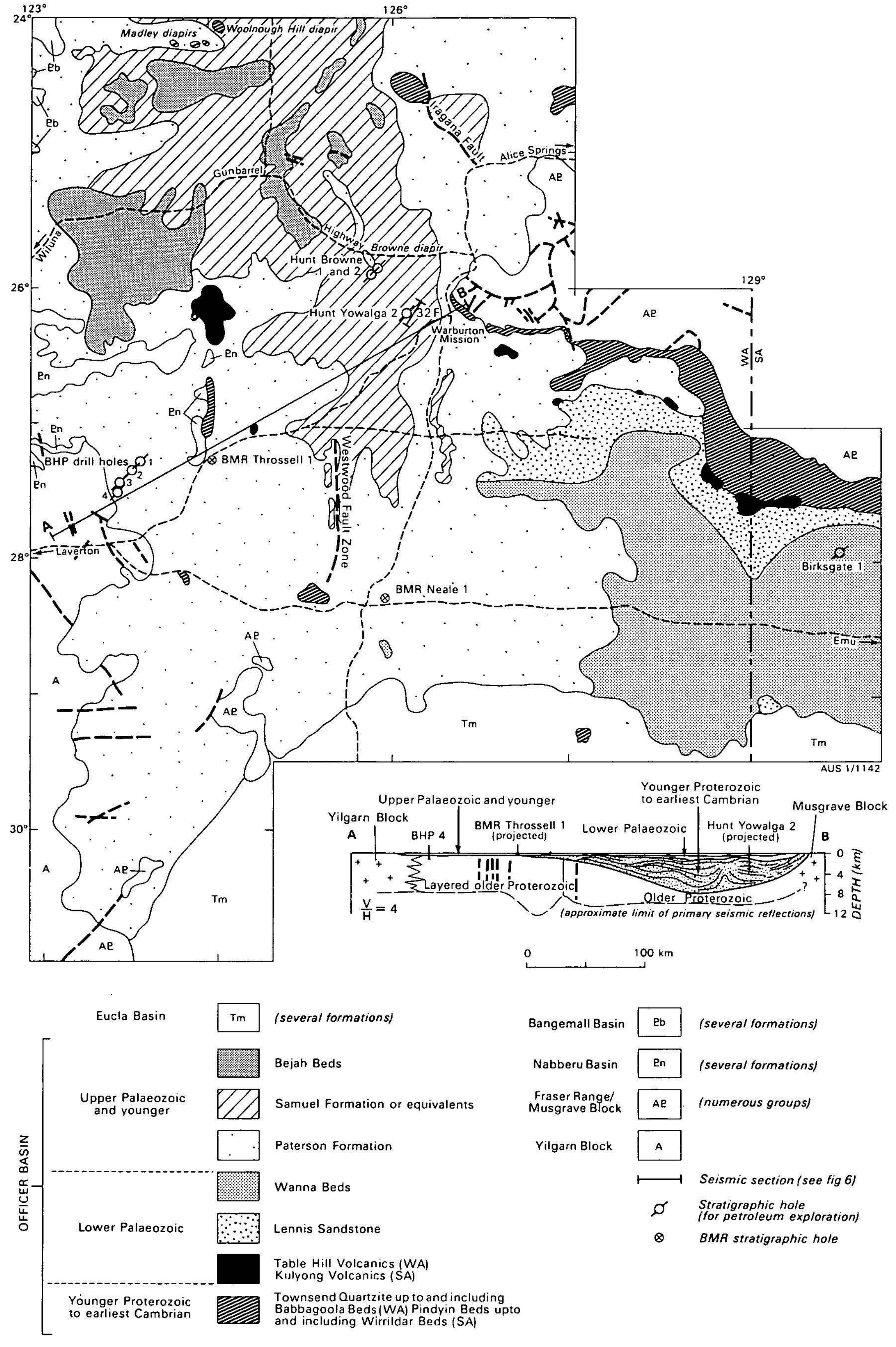


Figure 4. Simplified geological map (after Jackson & van de Graaff, in press), with locations of drillholes mentioned in text, and simplified section from Cosmo Newberry to Warburton (line A-B on map).

In general, the sequence described above comprises a basal quartzite, followed by fine clastics and carbonates, which are in places evaporitic, then glacigenic deposits and, finally, widespread thick shelf sand sheets, again with some carbonates and evaporites. It therefore partly mirrors the broad stratigraphic sequence seen in the upper Proterozoic in both the Amadeus Basin and the Adelaide Geosyncline, and indicates that the epeirogenic sedimentation characteristic of this period in central Australia extended into Western Australia. Although not well known, the Officer Basin sequence appears to be thinner and to contain longer breaks than the Adelaidean or Amadeus sequences.

Prior to the eruption of the overlying early Cambrian volcanics, most of the basin fill in Western Australia was gently folded and faulted, but more intense tectonism is evident along the southern edge of the Musgrave Block, where the basal Townsend Quartzite is more steeply folded. This folding is interpreted as the marginal effects of the Petermann Ranges Orogeny, which in this area is, therefore, of earliest Cambrian age. This tectonism may also have initiated diapiric intrusion from the lower part of the sequence. In the South Australian part of the basin, the Precambrian to Cambrian transition appears comformable.

The *Early Palaeozoic sequence* comprises three flat-lying formations, with a combined maximum thickness of about 600 m, that were deposited on a flat land surface. Two thin, but widespread, flows of tholeiitic basalt (*Table Hill Volcanics*) were extruded over most of the area southeast of 25°S in Western Australia. Equivalent basalts, the Kulyong Volcanics (Major & Teluk, 1967) are present in the extreme western part of the basin in South Australia. Considerable problems have been encountered in dating the volcanics (Compston, 1974), but radiometric ages of 475-485 ± 20 m.y. (*in* Major & Teluk, 1967) and 575 ± 40 m.y. (Compston, 1974) taken within the stratigraphic framework

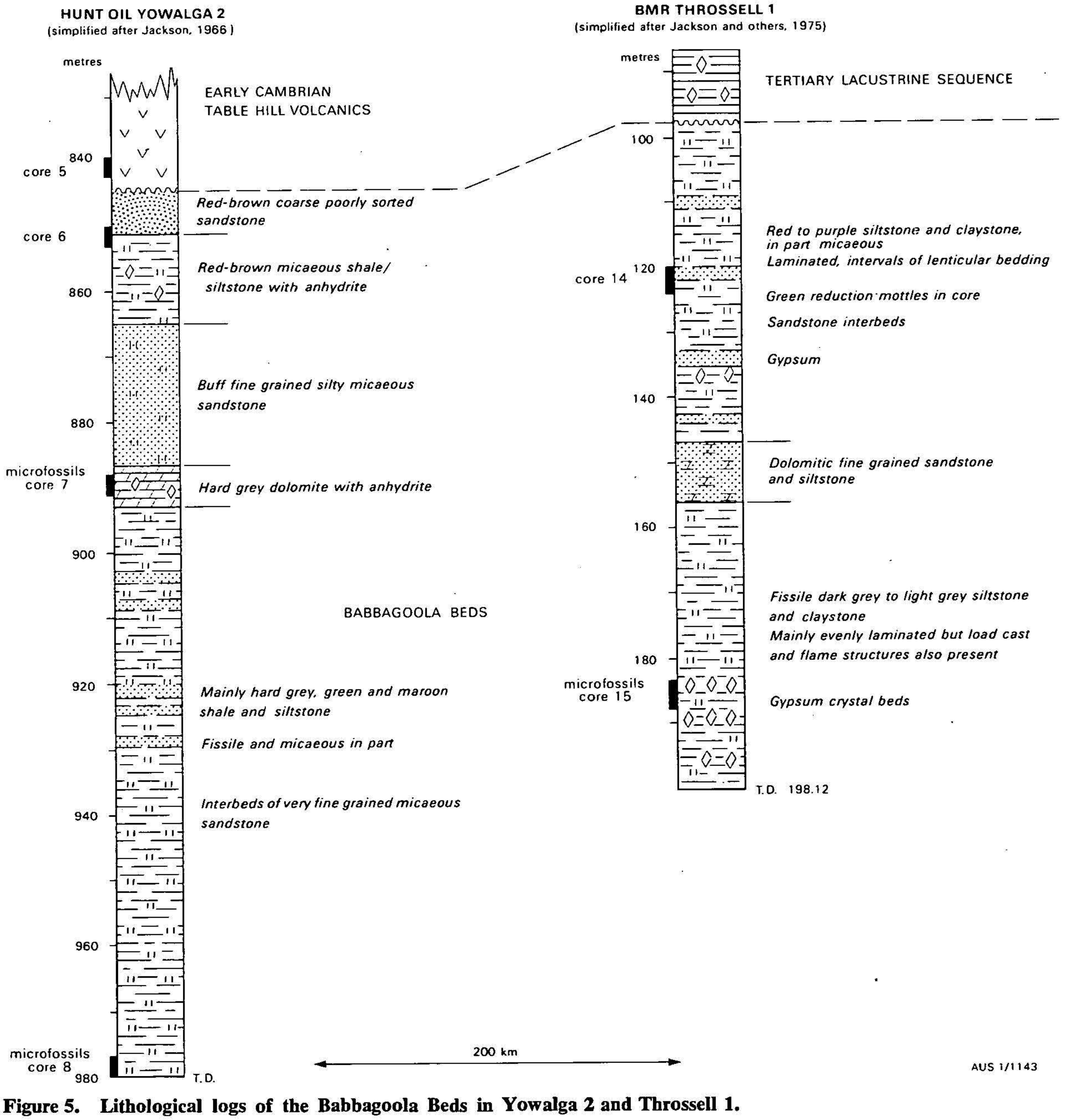


Figure 5. Lithological logs of the Babbagoola Beds in Yowalga 2 and Throssell 1.

indicate a probable earliest Cambrian age for their extrusion. Although cropping out only round the margins of the basin, the volcanics can be traced seismically throughout the basin at depths down to 1300 m. Disconformably overlying the volcanics are apparently unfossiliferous shallow marine subtidal to marginal marine sandstone units, the *Lennis Sandstone* and *Wanna Beds*, with a combined thickness of about 500 m, and probably also of Cambrian age.

Late Palaeozoic and younger units comprise flat-lying partly dissected Permian fluvioglacials (*Peterson Formation*) up to 450 m thick, Cretaceous marine units (*Samuel Formation* and *Bejah Claystone*) up to 100 m thick, and a variety of thin Cainozoic units, all continuous with and lithologically similar to formations in the Canning Basin to the north.

Babbagoola Beds

Distribution

The Babbagoola Beds are known only from Hunt Oil Yowalga 2 in the central part of the Officer Basin, where they are overlain by the Lower Cambrian Table Hill Volcanics, and BMR Throssell 1, 200 km to the southwest, near the basin's western margin (Fig. 4), where they are overlain by a poorly lithified Tertiary lacustrine sequence. However, these thin intersections of Babbagoola Beds are part of what appears to be a gently folded structurally conformable sequence, of late Proterozoic to earliest Cambrian age, which is several kilometres thick and appears to be widespread in the Western Australian part of the basin (Jackson & van de Graaff, in press).

Lithology

The formation can be divided (Fig. 5) into an upper unit of reddish brown poorly sorted sandstone and siltstone (more than 50 m thick), a thin middle interval of dolomite or dolomitic sandstone (less than 10 m thick), and a lower unit of fissile grey to green laminated siltstone and claystone (at least 100 m thick). Gypsum and anhydrite are common along fractures and as vein fillings in both holes.

Microfossils

Twelve samples of dark grey siltstone and shale were taken for palaeontological examination from core 15 in the lower subdivision of the unit in Throssell 1. Six samples of similar rock types were taken from core 8 at the bottom of Yowalga 2 (Fig. 5). Two additional samples from Yowalga 2 (one of dolomite in core 7 and another from core 8) were also examined. Thin sections were prepared of about half of the samples, and the remainder were macerated by normal palynological methods (i.e. dissolution of carbonate and silicate in HCl & HF, and separation of organic matter by flotation in heavy liquid). The preparations were studied using a Zeiss Photomicroscope III in reflected and transmitted light modes. All preparations (nos MFP 7119-7130) and negatives are stored at the Bureau of Mineral Resources.

The organic matter from Yowalga 2 is much lighter brown, and better preserved than that from Throssell 1. In Yowalga 2, microfossils range from dark yellow to light brown, and, despite compactional effects, their morphology is clearly visible. Amorphous and sheet-like organic matter is also pale-coloured and relatively undamaged physically. In Throssell 1, in contrast, the amorphous organic matter is black and opaque, and occurs in thick irregular lumps. Microfossils are rare, poorly preserved, and dark brown. Some spherical black opaque objects may have been microfossils, but could not be identified as such because of their high degree of carbonisation.

Most microfossils in the assemblage are acritarchs (Downie & others, 1963) with the Sphaeromorphitae most in evidence. Most sphaeromorphs have long stratigraphic ranges, and only those of limited range are illustrated here. Acanthomorph acritarchs (those forms with spines) are present, although they are not very abundant. They are, however, of considerable stratigraphic value. A species of *Tasmanites* reported from the Lower Cambrian of the Baltic region also occurs. The remaining microfossils are of environmental, but not stratigraphic, interest.

Of the Acanthomorphitae, the most frequent is *Micrhystridium lanatum* Volkova 1969 (Fig. 6 a-c). This small (4-10 μm) form with abundant short spines is very common in early Cambrian assemblages in Europe and Australia (see Table 1). It is morphologically very similar to *M. pallidum* Volkova 1969 and appears to have the same stratigraphic range. The two species may perhaps be synonymous. *Baltisphaeridium orbiculare* Volkova 1968 (Fig. 6d) occurs only as fragmentary pieces, whose estimated complete diameter is between 15 and 23 μm. *Deunffia flagellata* Jankauskas 1975 (Fig. 6 f, g) and *Alliumella* sp. (Fig. 6e) are identical to forms described from the early Cambrian of the Georgina Basin (Walter & others, 1979) and South Australian Officer Basin (Muir, 1979), as are the sphaeromorph acritarchs *Protosphaeridium densum* Timofeev 1966 and *P. muirii* Bliss 1979. The thin-walled *Leiopsophosphaera pelucidus* Schepelova 1962 (Fig. 6 h, i) and the porate *Tasmanites bobrowskae* Wazynska 1967 (Fig. 7a) are recorded for the first time in the Lower Cambrian of Australia. Two other acritarchs (not illustrated here, because of poor preservation) in the assemblage are *Granomarginata squamacea* Naumova 1961, and *Vulcanisphaera microspinosum* Tynni 1978.

None of the other illustrated forms has a useful stratigraphic range. *Strictosphaeridium* sp. (Fig. 6n) is identical with a form illustrated by Vidal (1979) from the Upper Riphaean and Vendian. It is much darker than the other microfossils in the assemblage, and morphologically is very similar to the probable blue-green algal form *Glenobotrydion aenigmaticus* Schopf 1968 from the Adelaidean Bitter Springs Formation of the Amadeus Basin. Both *Stictosphaeridium* sp. and *G. aenigmaticus* have morphological counterparts in older rocks (Roper Group—Peat & others, 1978), and similar forms exist among the blue green algae of the present day. The dark colour of the specimen in the Yowalga 2 core may indicate reworking from Adelaidean sediments, or the organism may have been washed into the depositional basin from some marginal area, where it may have suffered exposure to the atmosphere and partial oxidation, which can cause carbonisation (Mackowsky, 1968).

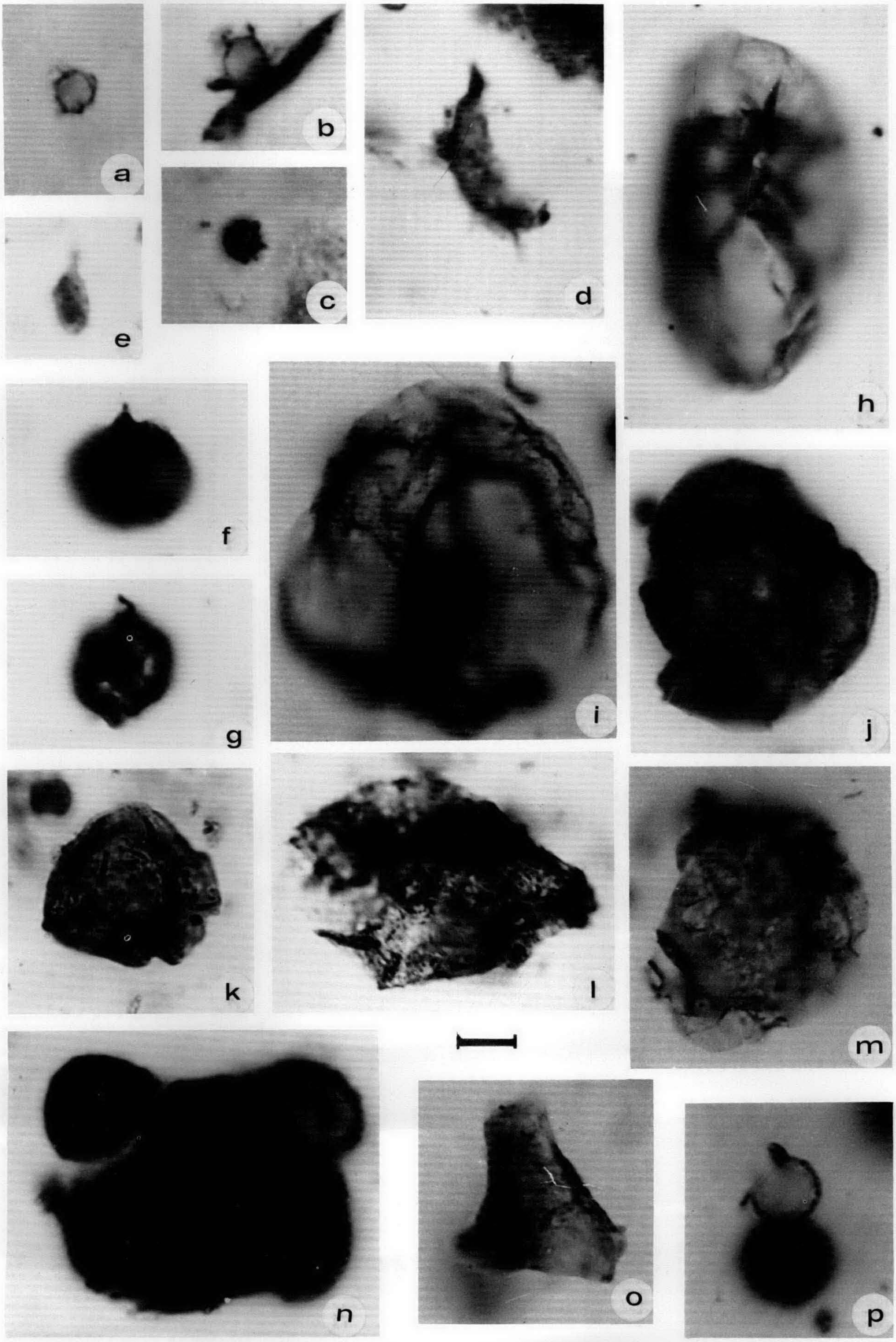
The triangular form *Triporina* Naumova 1953 (Fig. 6o) has been reported from a number of Phanerozoic horizons. It is believed to be algal in origin (Naumova, 1953), and has no stratigraphic value.

Figure 6p shows an organic film enveloping a pyrite framboid (black), and an upper empty vesicle. The inside of the vesicle is irregular with the impressions

Species	*Babbagoola Beds Yowalga 2*	*Babbagoola Beds Throssell 1*	*Observatory Hill Beds Wilkinson 1 (Muir, 1979)*	*Adam Shale Hay River 10 (Walter & others, 1979)*	*East Greenland (Vidal, 1979)*	*Finland (Tynni, 1978)*	*Estonia (Volkova, 1968)*	*Latvia (Birkis & others, 1970)*	*Poland (Wazynska, 1967)*	*Scotland (Bliss, 1977)*	*Stratigraphic range*
	Formations and Localities										
Micrhystridium lanatum	X	X	X	X		X*	X	X		X	Lower Cambrian
Baltisphaeridium orbiculare	X				X		X	X			Lower Cambrian
Alliumella sp.	X		X	X						X	Tommotian-Lower Cambrian
Deunffia flagellata	X	X		X						X	Tommotian-Lower Cambrian
Vulcanisphaera microspinosum	X					X	X				Lower Cambrian
Granomarginata squamacea	X					X	X	X			Tommotian-Lower Cambrian
Protosphaeridium densum	X	X	X	X						X	Tommotian-Lower Cambrian
Protosphaeridium muirii	X			X						X	Upper Riphaean-Lower Cambrian
Leiopsophosphaera pelucidus	X								X		Lower Cambrian
Tasmanites bobrowskae	X					X			X		Lower Cambrian
Solisphaeridium sp.		X	X	X						X	Lower Cambrian
Stictosphaeridium sp.	X(1)				X(2)						(1) Lower Cambrian (2) Upper Riphaean

* here described as *Micrhystridium pallidum* f. minor.

Table 1. Stratigraphic ranges of microfossils found in the Babbagoola Beds and Observatory Hill Beds, compared with other assemblages of similar age.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of small pyrite crystals, which were probably dissolved out of the organic film during maceration.

The sac-like microfossil (Fig. 7b) is similar to a number of such objects described by Bloeser & others (1977) and Vidal (1979) from late Riphaean and Vendian rocks. The microfossils are similar in morphology to the Chitinozoa, which range from Ordovician to Carboniferous, but Vidal (1979) dismisses the possibility that the 'sac-like microfossils' are real Chitinozoa, commenting that such a simple morphology could arise in any number of unrelated groups. In the Babbagoola Beds, the 'sac-like microfossils' are in the same state of preservation as the rest of the assemblage (with the exception of *Strictosphaeridium* sp.), and thus appear to be contemporaneous with the assemblage.

The other illustrated specimens from the Babbagoola Beds are the abundant nonseptate filaments (Fig. 7c-g). Those in Figure 7c-e are narrow filaments similar to *Gunflintia oehlerae* Muir 1976. These are narrow, ribbon-shaped filaments, which are abruptly twisted at intervals along their length. Although they range at least from about 1650 m.y. ago (Amelia Dolomite, Muir, 1976) until the earliest Cambrian, they were abundânt at this time in Australia (Walter & others, 1979; Muir, 1979) and the British Isles (Peat & Bland, unpublished data).

The other, larger, filaments (Fig. 7f, g) are illustrated in a petrological thin section. The filaments weave in all directions through the micritic grains of the sediment, and served to bind and stabilise the surface. None of the filaments observed in any preparations contains any cell residues, and the remains illustrated here are almost certainly empty blue green algal sheaths.

Correlatives

The nearest possible correlatives in South Australia are the Wirrildar Beds, which crop out in the central-western part of the Birksgate area (Fig. 3). Although they crop out very poorly, Major (1973) described the sequence as gently folded and comprising arkosic sandstone, micaceous sandstone and siltstone, and flaggy dolomite. They are unconformably overlain by the Kulyong Volcanics, which are the South Australian equivalent of the Table Hill Volcanics. Therefore, the lithology and structural setting of the Wirrildar Beds are similar to those of the Babbagoola Beds. Although it is thicker, the middle part of the unnamed sequence (505-1122 m) intersected in Continental Oil's Birksgate 1 (Henderson & Tauer, 1967) is lithologically very similar to that seen in Yowalga 2 and Throssell 1—red and brown calcareous siltstones and shales, overlying carbonates, which in turn overlie grey hard fissile shale and siltstone—and a direct correlation is suggested. The sequence of white and red sandstone above 505 m in Birksgate 1 can be correlated with the Wanna Beds and Lennis Sandstone that crop out in the eastern parts of Western Australia. A marked reduction in drill penetration rate at 505 m was interpreted by Henderson & Tauer (1967) as a possible unconformity. There is no evidence of the Kulyong Volcanics at this unconformity, although they crop out about 40 km to the northwest.

McKirdy & Kantsler (1980, fig. 2) and Pitt & others (1980, fig. 10) illustrate correlations between wells and outcrops in the South Australian part of the basin. McKirdy has correlated the middle part of Birksgate 1 drillhole sequence (described above) with the Observatory Hill Beds in Emu 1, Murnaroo 1, Wilkinson 1, and Byilkaoora 1 wells. Therefore, the Babbagoola Beds can be correlated with the Observatory Hill Beds.

These lithostratigraphic correlations of the Babbagoola Beds in Western Australia with the Observatory Hill Beds of South Australia are confirmed by palynological studies. Five samples of organic-rich laminated carbonate from the Observatory Hill Beds in SADME Wilkinson 1 (between depths of 314 m and 512 m) contain stratigraphically significant microfossils identical to those found in the Babbagoola Beds. By far the most common form is *Micrhystridium lanatum* Volkova 1969, although *Granomarginata squamacea* Naumova 1961 and *Protosphaeridium densum* Timofeev 1966 also occur. A form similar to *Solisphaeridium* sp. Bliss 1979 was found in the upper part of the sequence examined (at 314.47 and 461.72 m). There is no evidence of reworking in the assemblage and, therefore, the Observatory Hill Beds are Lower Cambrian and correlatable with the Babbagoola Beds.

The stratigraphic ranges of the species found in the Babbagoola Beds and Observatory Hill Beds are shown in Table 1 and compared with other assemblages of similar age.

Implications for petroleum prospectivity

Babbagoola Beds as source rocks

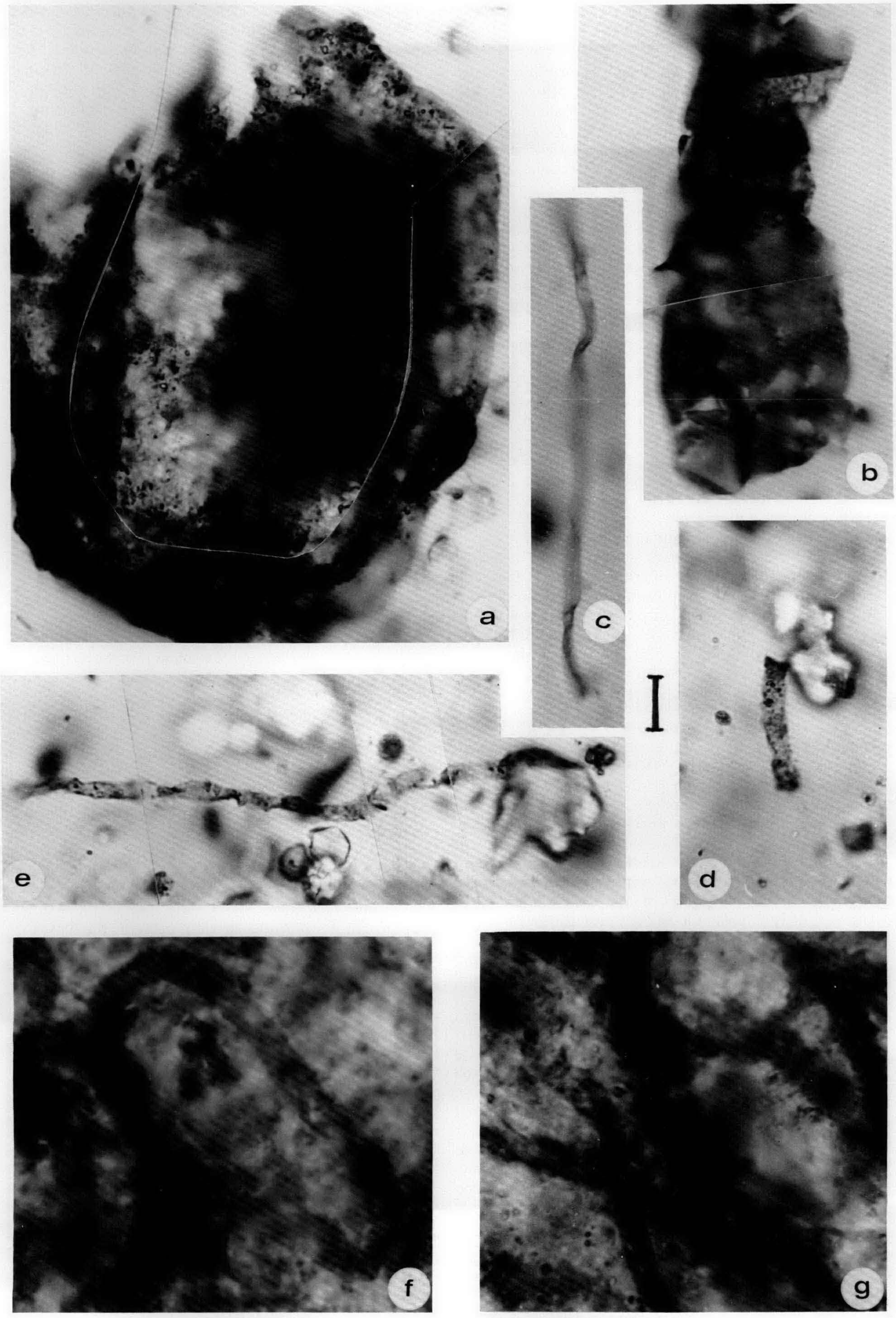
Until about 1978, the petroleum prospects of the Western Australian part of the Officer Basin were considered very poor. The basin fill is mainly Cambrian or older, and there were no known potential source rocks. Recent studies in South Australia, however, have indicated that good to excellent hydrocarbon source beds occur in the Cambrian Observatory Hill Beds, which, from their stratigraphic setting, lithology, and microfossils, are equivalent to the Babbagoola Beds. Although the Babbagoola Beds have only been intersected in two drillholes, regional mapping and seismic studies indicate that they are probably widespread in the subsurface of the basin in Western Australia. This fact alone significantly upgrades the potential of the Western Australian part of the basin, at least to the stage where investigation of the Babbagoola Beds is warranted to see if they have a source rock potential similar to that of their correlative in South Australia.

The better source rocks in South Australia are fine-grained carbonates deposited in marine sabkha environments (e.g. Wilkinson 1 and Wallira West 1 on the southeast margin of the Officer Basin). In contrast, the shale-siltstone lithofacies (for example, Birksgate) are characterised by lower organic carbon contents and hydrocarbon yields (McKirdy & Kantsler, 1980).

Reservoir rocks

There appear to be ample reservoirs: much of the Phanerozoic and Late Proterozoic sequence consists of

Figure 6. Microfossils from Hunt Oil Yowalga 2. Scale bar is 5 µm long. a,b,c—Micrhystridium lanatum; d—Baltisphaeridium orbiculare; e—Alliumella sp; f,g—Deunffia flagellata; h,i—Leiopsophosphaera pelucidus; j,k—Protosphaeridium densum; l,m—Protosphaeridium muirii; n—cf. Stictosphaeridium sp; o—Triporina; p—pyrite framboid with empty sac.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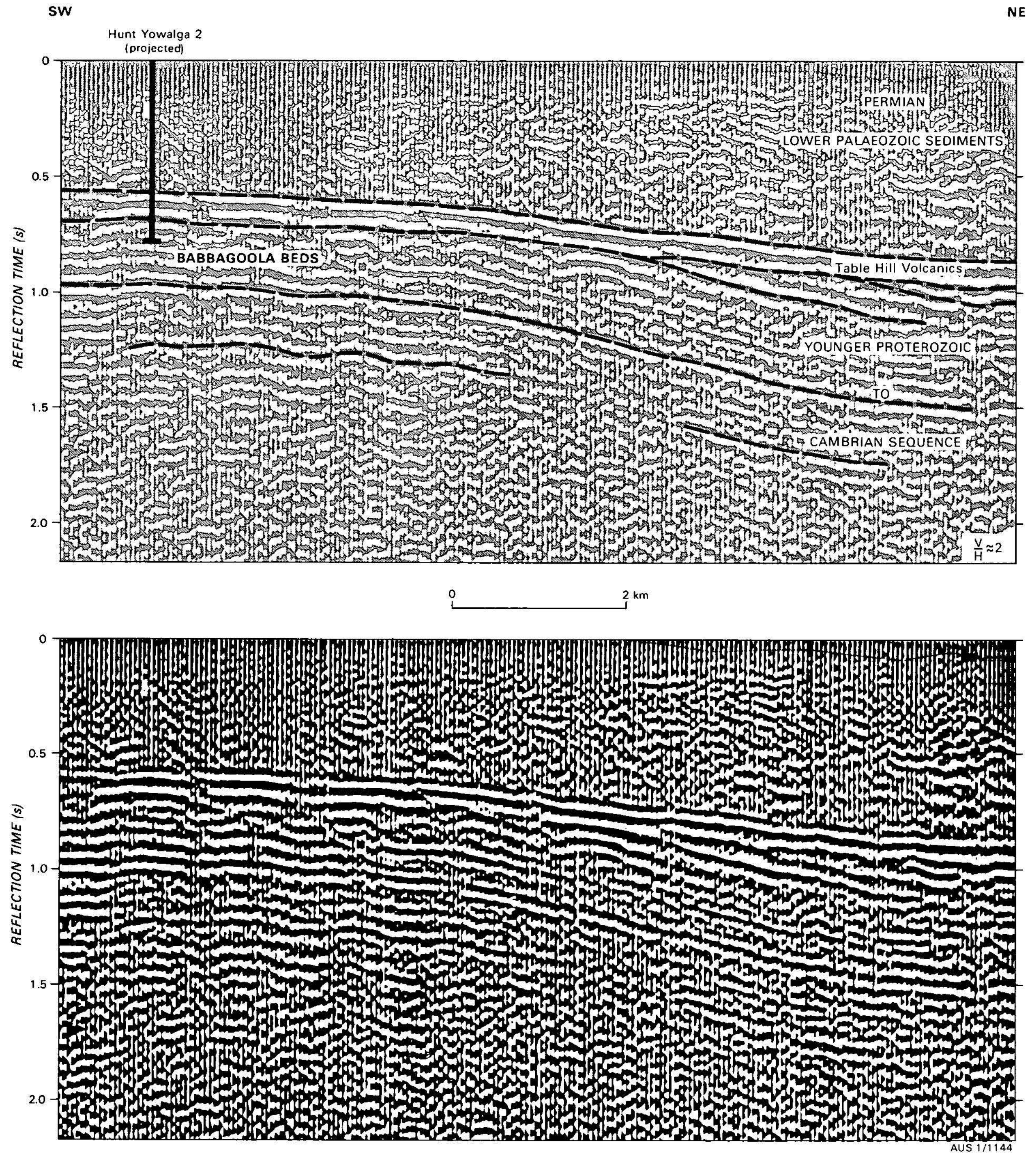

Figure 8. Part of seismic line Hunt Oil 32-F, showing unconformity at base of Table Hill Volcanics and gently folded Proterozoic to Cambrian sequence containing Babbagoola Beds.

coarse-grained clastic rocks (e.g. Punkerri Beds, Lennis Sandstone, Wanna Beds) with highly porosity and permeability. Core samples from the Lennis Sandstone and Wanna Beds in BMR Neale 1 show porosities ranging from 23 to 31 percent, and permeabilities in the range 62 to 390 millidarcies (Jackson & others, 1975). For comparison, oil and gas producing wells in the Pacoota Sandstone of the Amadeus Basin commonly have porosities in the range 5 to 12 percent and permeabilities between 1 and 300 millidarcies.

Structure

The broad anticlinal flexures mapped on the seismic sections (e.g. Fig. 8) in the younger Proterozoic sequence and the diapiric structures penetrating the basin sequence nearly to the surface are examples of classical types of hydrocarbon trap. Apart from the doming of strata, the diapirs could offer numerous potential traps in the form of fault closures, pinch-outs, and truncation of reservoir beds.

Figure 7. Microfossils from Hunt Oil Yowalga 2 (continued). Scale bar is 5 μm long; a—Tasmanites bobrowskae; b—sac-like microfossil; c,d,e—twisted filaments, c.f. Gunflintia oehlerae; f,g—filaments binding sediment.

Maturation

The light colour of the organic matter in Yowalga 2 is compatible with a maturation grade in the oil window, at the higher temperature level. The coalification of the organic matter in Throssell 1 is so complete that dry gas is the best that could be expected. Indeed, the material from Throssell 1 may be overmature, with no hydrocarbon potential at all. However, the possibility of post-maturation migration of liquid or gaseous hydrocarbons cannot be ruled out.

Hydrocarbon shows

There are no proven oil or gas seeps in the Western Australian part of the basin. Minor oil and gas shows were encountered during the drilling of Hunt Oil Browne 1 & 2 (Jackson, 1966). Gas cut mud, gas odours, and good fluorescent cuts were noted in well cuttings between 134 m and 275 m in Browne 1. Subsequent analysis showed traces of oil in core samples at 213 m, 214 m, 258 m and 259 m. Similar shows were encountered in Browne 2 between 259 m and 262 m and were also confirmed by core analysis. All of these shows were within the Browne Beds, which Jackson & van de Graaff (in press) correlate with the Bitter Springs Formation of the Amadeus Basin. They are, therefore, older than the Babbagoola Beds and lie near the base of the Late Proterozoic to earliest Cambrian sequence. Hunt Oil Browne 1 & 2 were drilled into the top of a diapir that has intruded some 4000-5000 m upwards from near the base of this sequence, through the Phanerozoic, almost to the surface. The source of the hydrocarbon shows is unknown. It could be from within the Browne Beds, which would imply an Adelaidean ?algal source for the material, but lateral migration into the top of the brecciated diapir from other potential sources (for example the Babbagoola Beds) is also a possibility.

Conclusions

With this establishment of a correlation between the little-known Babbagoola Beds in Western Australia and the Observatory Hill Beds in South Australia, which locally have excellent source rock potential, we consider that the Western Australian portion of the Officer Basin should be considered as an attractive area, warranting further petroleum exploration. It has potential petroleum traps in salt domes and broad anticlinal flexures, several sandstone formations that could make good reservoirs, and, now, a potential source rock sequence.

Although the late Proterozoic to early Cambrian sequence that contains the Babbagoola Beds is some thousands of metres thick and probably underlies much of the area shown in Figure 3, it has been penetrated in only four drill-holes. Additional stratigraphic drilling would be the most appropriate way to test further the petroleum potential of this vast onshore area.

References

Birkis, A. P., Brangulis, A. P., Volkova, N. A., & Rozanov, A. Yu, 1970—New data on the Cambrian Stratigraphy of Western Latvia. *Doklady of the Academy of Sciences of the U.S.S.R., Earth Science Sections*, 1971, **195**, 58-61.

Bliss, G. M., 1977 — The micropalaeontology of the Dalradian. *Ph.D. thesis, University of London* (unpublished).

Bliss, G. M., 1979—New microfossil evidence from the Dalradian and its bearing on the age of the Supergroup. *Journal of the Geological Society, London.*

Bloeser, B., Schopf, J. W., Horodyski, R., & Breed, W. J., 1977—Chitinozoans from the Late Precambrian Chuar Group of the Grand Canyon, Arizona. *Science*, **195**, 676-9.

Compston, W., 1974—The Table Hill Volcanics of the of the Officer Basin—Precambrian or Palaeozoic? *Journal of the Geological Society of Australia*, **21**(4), 403-12.

Downie, C., Evitt, W. R., & Sarjeant, W. A. S., 1963—Dinoflagellates, hystrichospheres, and the classification of the acritarchs. *Stanford University Publications, Geological Science*, **7**, 3.

Henderson, S. W., & Tauer, R. W., 1967—Birksgate No. 1 Well, South Australia, stratigraphic drilling project, well completion report. *South Australia Department of Mines, Open File Envelope* **768** (unpublished).

Jackson, M. J., 1979—Mineral potential of the Officer Basin, Western Australia. *In* Abstracts, 8th BMR Symposium. *Bureau of Mineral Resources, Australia, Record* **1979/13** (unpublished).

Jackson, M. J., van de Graaff, W. J. E., & Boegli, J.-C., 1975—Shallow stratigraphic drilling in the Officer Basin, Western Australia, 1972. *Bureau of Mineral Resources, Australia, Record* **1975/49** (unpubl.).

Jackson, M. J., & van de Graaff, W. J. E., (in press)—The Geology of the Officer Basin, Western Australia. *Bureau of Mineral Resources, Australia, Bulletin* **206**.

Jackson, P. R., 1966—Geology and review of exploration Officer Basin, Western Australia. *Hunt Oil Co.—Hunt Petroleum Corp—Placid Oil Co. Report* (unpublished).

Jankauskas, T. V., 1975—New Lower Cambrian acritarchs of the Baltic region. *Palaeontologicheskii Zhumal*, **9**(1).

McKirdy, D. M., & Kantsler, A. J., 1980—Oil geochemistry and potential source rocks of the Officer Basin, South Australia. *The APEA Journal* **20**(1), 68-86.

Mackowsky, M.-Th., 1968—European Carboniferous coalfields and Permian Gondwana coalfields. *In* Murchison, D., & Westoll, T. S. (Editors), Coal and Coalbearing Strata. *Oliver & Boyd, Edinburgh* 325-45.

Major, R. B., 1973—The Wirrildar Beds. *Geological Survey of South Australia Quarterly Geological Notes* **45**, 8-11.

Major, R. B., & Teluk, J. A., 1967—The Kulyong Volcanics. *Geological Survey of South Australia Quarterly Geological Notes*, **22**, 8-11.

Muir, M. D., 1976—Proterozoic microfossils from the Amelia Dolomite, McArthur Basin, Northern Territory. *Alcheringa*, **1**, 143-58.

Muir, M. D., 1979—Palynological examination of microfossils from the Observatory Hill Beds, Wilkinson No. 1 D.D.H., Officer Basin, South Australia. *Bureau of Mineral Resources, Australia, Professional Opinion* Geol 79/010 (unpublished).

Naumova, S. N., 1949—Spori nizhnezo Kembriya. *Izvestiya Akademia Nauk SSSR, Seria Geologia*, **4**, 49-56.

Naumova, S. N., 1961 — Sporovo-pyltseye Kompleksy Refeiskikh i nizhnekembriiskikh otlozhenii SSSR. Mezhdunarodny Geologicheskii Kongress, XXI Sessiya (1960). *Doklady Sovetskikh Geologov*, 109-117.

Peat, C. J., Muir, M. D., Plumb, K. A., McKirdy, D. M., & Norvick, M. S., 1978—Proterozoic microfossils from the Roper Group, Northern Territory, Australia. *BMR Journal of Australian Geology & Geophysics*, **3**, 1-17.

Pitt, G. M., Benbow, M. C., & Youngs, B. C., 1980—A review of recent geological work in the Officer Basin, South Australia. *The APEA Journal* **20**(1), 209-19.

Schopf, J. W., 1968—Microflora of the Bitter Springs Formation, late Precambrian, Central Australia. *Journal of Palaeontology*, **42**, 651-88.

Shepelova, Ye. D., 1962—Plant? fossils of unknown taxonomic position from the deposits of the Bavlinskaya Series in the Volga-Urals oil province. *Doklady of the Academy of Sciences of the U.S.S.R., Earth Science Sections*, 1964, **142**, 179-81.

TIMOFEEV, B. V., 1966—Micropalaeophytological research into ancient strata. *Nauka, Moscow.*

TYNNI, R., 1978—Lower Cambrian fossils and acritarchs in the sedimentary rocks of Soderfjarden, Western Finland. *Geological Survey of Finland, Bulletin* **297**, 41-81.

VIDAL, G., 1979—Acritarchs from the Upper Proterozoic and Lower Cambrian of East Greenland. *Gronlands Geologiske Undersogelse, Bulletin* **134**.

VOLKOVA, N. A., 1968—Acritarcha of Precambrian and Lower Cambrian deposits of Estonia. *Trudy Geological Institute, Academy of Sciences, SSSR,* **188**, 8-36 (In Russian).

VOLKOVA, N. A., 1969—Acritarchs of the northwest Russian Platform. *In* ROZANOV, A. YU, & OTHERS, Tommatian Stage and the Cambrian Lower Boundary Problem. *Trudy Geological Institute, Academy of Sciences, SSSR,* **206**, 2224-36 (In Russian).

WALTER, M. R., SHERGOLD, J. H., MUIR, M. D., & KRUSE, P. D., 1979—Early Cambrian and latest Proterozoic stratigraphy, Desert Syncline, southern Georgina Basin. *Journal of the Geological Society of Australia,* **26**, 305-12.

WAZYNSKA, H., 1967—Wstepne badania Mikroflorystijczne osadow sinianu : Kambni z obszam Bialoiviezy. *Kirartalnik Geologie,* **1**, 10-20.

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Intrusive felsic-mafic net-veined complexes in north Queensland

D. H. Blake

Several net-veined complexes, formed where mafic magma has intruded either felsic magma or melted felsic rock, are exposed in the Mount Isa region and elsewhere in north Queensland. The most typical features of these complexes are pillow-like inclusions of mafic igneous rock, with intricately crenulate to cuspate contacts, enclosed in and veined by granite or granophyre.

Introduction

Net-veined complexes, in which closely spaced rounded to angular inclusions of dark mafic rock are enclosed in and veined by pale felsic rock, are common in areas where both mafic and felsic intrusive igneous rocks are exposed. The purpose of this paper is to point out that, although such complexes have commonly been interpreted, in Australia and overseas, as granitic intrusions containing xenoliths of pre-existing mafic rocks, this apparently obvious explanation is probably incorrect in most cases. Close examination generally reveals that much or all of the mafic component is essentially the same age or younger, not older, than the granitic component. This is particularly clear in areas such as Iceland, where well-exposed net-veined complexes intrude flat-lying to gently dipping Tertiary volcanics (e.g. Blake, 1966), and all gradations are present from intrusive net-veined complexes to composite dykes and to composite lavas and pyroclastics formed of co-existing basaltic and rhyolitic magma (Blake & others, 1965; Blake, 1969).

Examples of net-veined complexes known to the author in north Queensland include several of Precambrian age in the Mount Isa region and some near Herberton, Georgetown, and Bowen that are late Palaeozoic (Fig. 1). These complexes range in size from bodies several kilometres across to zones a metre or so wide associated with mafic dykes intruding granite.

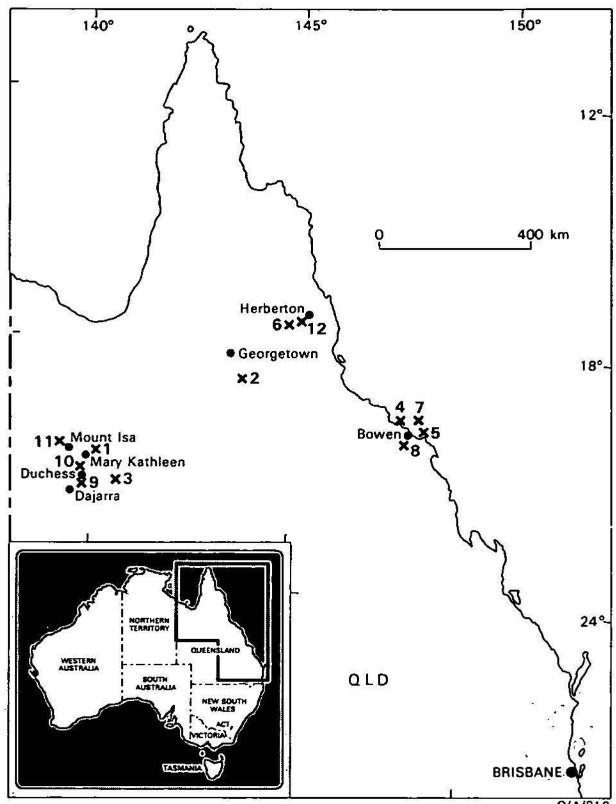

Figure 1. Locality map.
1—Burstall Granite; 2—Bagstowe Ring Dyke Complex; 3—Boorama Waterhole; 4—Cape Upstart; 5—Gloucester Island; 6—Gurrumba Ring Complex; 7—Holbourne Island; 8—Ida Creek; 9—Mount Erle Igneous Complex; 10—Myubee Igneous Complex; 11—Sybella Granite; 12—Silver Valley.

Characteristic features of felsic-mafic net-veined complexes (Figs. 2-6)

Net-veined complexes typically consist of two main intrusive rock types, one mafic (basaltic/gabbroic or andesitic/dioritic) and the other felsic (rhyolitic/granitic); both may be fine, medium, or coarse in grain-size. Subordinate heterogeneous hybrid rocks, of intermediate composition, and xenoliths of country rocks are also commonly present. The mafic rocks occur as inclusions, many of which have rounded forms resembling pillows in subaqueous basaltic extrusions. However, the contacts of the pillows in net-veined complexes are highly irregular, and in detail are crenulate or cuspate, with rounded protuberances of mafic rock alternating with pointed embayments of felsic rock. Contacts between the two rock types are generally sharp, but may be diffuse in places.

Individual mafic pillows can range from a few centimetres to several metres in maximum diameter. At some localities they show a progressive decrease in grain-size towards their margins, and in many places the margins are appreciably darker than the interiors. Most pillows are cut by felsic veins connected to the enclosing granitic rock. These veins range from sinuous to straight-sided and, even where only a millimetre or so thick, they may extend for many metres into large pillows. Some veins terminate within pillows; others, especially straight-sided veins, cut right across. Many pillows, and also angular mafic inclusions, are cut by net-works of felsic veins, and partly fragmented pillows are common.

The felsic rock immediately adjacent to the mafic pillows, and also that veining the pillows, is commonly more leucocratic than that away from the pillows, and may show distinct textural features. For example, in the Mount Isa area, the leucocratic felsic phase in some net-veined complexes is not foliated, whereas the 'normal' felsic rock shows a prominent foliation, and at one locality non-porphyritic leucogranite separates mafic pillows from porphyritic granite (Fig. 5).

Patchy heterogeneous hybrid rocks, intermediate in composition between the felsic and mafic components, commonly occur where there are concentrations of mafic inclusions. They may form inclusions within more leucocratic rock or part of the 'matrix' enclosing closely spaced mafic inclusions.

Many mafic inclusions in net-veined complexes are angular rather than rounded, and in some complexes angular inclusions appear to predominate. Such inclusions generally have straight sides, in marked contrast to the pillows, but some may have one side with irregular crenulate to cuspate contacts. Recognisable country rock xenoliths, where present, are angular to rounded; they do not have crenulate margins.

Figure 2. Net-veined complex of the Mount Erle Igneous Complex near Duchess, showing pillows of dolerite enclosed in and veined by granite.

Figure 3. Angular blocks of gabbro, probably representing fragments of mafic pillows, in granite of the Myubee Igneous Complex, north of Duchess. (Photo—R. J. Bultitude)

Interpretation of the features

It is now generally accepted that net-veined complexes of the type described here result from the commingling of felsic and mafic magmas or melts in relatively high-level intrusions (e.g., Blake & others, 1965; Walker & Skelhorn, 1966; Blake, 1966; Yoder, 1973; Wiebe, 1980; Taylor & others, 1980). The mafic component represents basaltic magma which either came into contact with felsic magma, or was intruded into pre-existing granitic rock, which was then partially melted and became remobilised. The basaltic magma, with a crystallisation temperature of 1100-1300°C, chilled against the felsic magma or granitic melt, of temperature 750-950°C (Sparks & others, 1977), and solidified, preventing widespread mixing of the two liquids. In such situations the adjacent felsic magma or melt adjacent to the mafic magma probably becomes superheated, so that it is less viscous than normal felsic magma and markedly less viscous than adjacent crystallising mafic magma; it also remains highly fluid long after the mafic magma has solidified, and hence is able to form thin veins intruding crystallised mafic rock.

The mafic magma forms pillows in the felsic liquid, as it does when erupted into water. The highly irregu-

Figure 4. Net-veined complex at Boorama waterhole, 99 km E of Duchess.

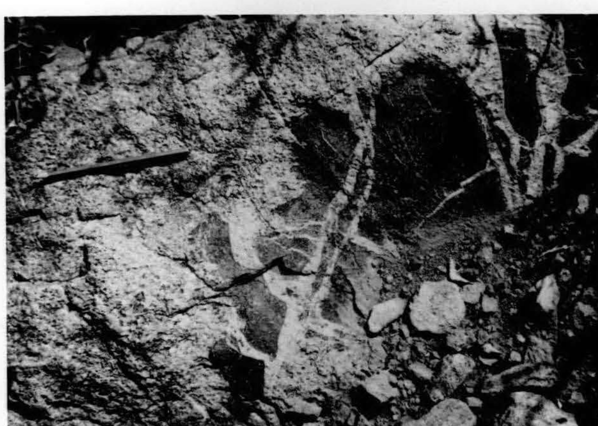

Figure 5. Partly fragmented pillow of Lunch Creek Gabbro enclosed in and veined by Burstall Granite in net-veined complex 9 km east of Mary Kathleen.

Figure 6. Intricately crenulate contact between a mafic pillow and Sybella Granite in a net-veined complex 14 km NW of Mount Isa. (Photo—R. J. Tingey)
The fine-grained granite in contact with the pillow probably represents a partial melt derived from the adjacent coarse-grained porphyritic granite.

lar contacts of the mafic pillows in net-veined complexes are the result of the mafic and felsic components being liquid simultaneously. Once formed, the larger pillows tend to become first fractured, then fragmented, and finally dispersed as small and mainly angular inclusions within the felsic liquid. All stages may be seen from pillows with a few felsic veins filling early fractures (Fig. 4), through partly fragmented pillows (Fig. 5), to scattered, mainly angular, mafic inclusions probably representing pieces of fragmented pillows (Fig. 3). Dark margins to pillows may be due partly to primary chilling against felsic magma, and partly to metasomatic alteration of the marginal pillow rock during crystallisation of the adjacent felsic component. Leucocratic felsic rock found adjacent to many pillows may represent the partial melt derived from the host granitic rock, more or less in situ, by the pillow-forming magma. Hybrid rocks may be formed either in situ or at deeper levels in the crust where minor mixing of melts has been able to take place.

Not all mafic igneous inclusions in a net-veined complex may represent mafic magma which was intruded into either felsic magma or melted felsic rock. Some, for example, may be xenoliths of older igneous country rocks, and some may be cognate xenoliths or 'restite' material. Although such xenoliths are commonly sufficiently different in composition from pillow-forming mafic rock to be readily distinguished, this may not always be the case. For example, a net-veined complex forming part of the Tuross Heads Tonalite, Moruya Batholith, and beautifully exposed on the coast at Tuross Heads, NSW, contains many good examples of mafic pillows with crenulate to cuspate margins, but may also contain some 'restite' xenoliths (Griffin & others, 1978).

North Queensland examples (Figure 1)

Mount Isa region

Mount Erle and Myubee Igneous Complexes. Bultitude & others, 1978). These Proterozoic igneous complexes intrude tightly folded and commonly scapolitic calc-silicate rocks of the Corella Formation near Duchess. The Mount Erle intrusion contains some copper ore-bodies, including that at the long abandoned Duchess copper mine. It crops out over about 30 km^2, and consists mainly of foliated, fine to medium-grained granite containing biotite, hornblende, and, commonly, scapolite, and non-foliated dolerite and gabbro, which locally contain primary olivine and pyroxene. In many places these rocks, and subordinate dioritic hybrids, occur together in net-veined complexes (Fig. 2). Much of the granite in these complexes is not foliated. The favoured interpretation of the field evidence is that the mafic component of the Mount Erle Igneous Complex was intruded as magma into pre-existing foliated granite, which was melted and remobilised by this magma, and later crystallised as non-foliated granite. Subsequently, the Mount Erle Igneous Complex was cut by faults, along which all rocks are highly sheared. The Myubee Igneous Complex is similar, but much smaller. It has a circular outcrop about 2 km across and consists of a non-foliated gabbroic core partly encircled by foliated granite, which forms a net-veined complex with the gabbro. Most of the gabbro fragments are angular (Fig. 3), but some with crenulate margins are present locally (R. J. Bultitude, personal communication, 1980), indicating that at least some of the gabbro probably represents mafic magma intruded into felsic magma or melted felsic rock. The gabbroic and granitic components could be correlatives of those forming the Mount Erle Igneous Complex.

Boorama Waterhole net-veined complex. At this locality bedded and brecciated calc-silicate rocks and associated amphibolite are intruded by a small, partly deformed, net-veined complex (Fig. 4) made up of fine to medium-grained leucogranite and dolerite (Donchak & others, 1979). The granite and dolerite here could be of similar age, although the dolerite is thought to be somewhat younger than the granite; both are Proterozoic.

Burstall Granite and Lunch Creek Gabbro. Contacts between these two Proterozoic units, which crop out east of Mary Kathleen, in the Marraba 1:100 000 Sheet area (Derrick, 1980), are marked by a well-developed net-veined complex containing good examples of partly fragmented pillows (Fig. 5). The Burstall Granite is generally foliated, except in the net-veined complex, and, in the writer's view, may be much older than the Lunch Creek Gabbro, which is not foliated. However, Derrick (1980)interprets the granite as being younger than the gabbro.

Sybella Granite west of Mount Isa. A small Proterozoic net-veined complex is exposed 14 km northwest of Mount Isa, where Sybella Granite has been melted and partly remobilised adjacent to a dyke-like dolerite intrusion. The dolerite forms a band of pillows, with chilled margins and intricately crenulated contacts (Fig. 5), enclosed in the granite. The granite in contact with the pillows is generally finer-grained and more leucocratic than that elsewhere, and probably represents rock which was melted by the younger pillow-forming magma. Another net-veined complex may be present a few kilometres to the north, east of May Downs homestead, as Joplin (1955) has described an exposure here with which consists of numerous mafic inclusions, some with crenulate margins (fig. 12 in Joplin, 1955; and plate 19, fig. 2 in Carter & others, 1961). Small net-veined complexes similar to that associated with the dolerite dyke northeast of Mount Isa are also present at a few of the many other localities in the Mount Isa Inlier where granite is intruded by mafic dykes.

Herberton district

Gurrumba Ring Complex (Blake, 1972). This is an elliptical structure up to 5.5 km across, consisting of an intrusive outer ring, presumed to be a ring dyke, surrounding a core of sedimentary and volcanic rocks. Most of the outer ring is a net-veined complex of pink porphyritic granophyre, medium to fine-grained gabbro and dolerite, and intermediate hybrid rocks, mostly of quartz diorite composition (Blake, 1972, plates 11 and 12). The granophyric and mafic components were probably emplaced at about the same time during the late Palaeozoic.

Silver Valley area, southwest of Herberton (Blake, 1972). Two dykes of pink porphyritic granophyre found in this area contain pillows of andesitic rock with crenulate to cuspate margins. The andesitic magma was probably injected into these dykes very shortly after the intrusion of the felsic magma, while the latter was still liquid, to form small net-veined complexes.

Georgetown area

A net-veined complex of diorite and granite occurs within the late Palaeozoic Bagstowe Ring Dyke Complex, about 110 km south-southeast of Georgetown. Here, near Bagstowe homestead, pink biotite microgranite of intrusive unit 8 contains rare pillows and abundant angular blocks of hornblende-biotite microdiorite belonging to intrusive unit 6 (Bain & others, 1980). Hybrid rocks and later cross-cutting pink rhyolite dykes are also present.

Bowen area

Several late Palaeozoic igneous intrusions near Bowen contain net-veined complexes of granitic, dioritic, and doleritic rocks. These are well exposed in coastal sections on Cape Upstart, Holbourne Island, and Gloucester Island (Paine & others, 1970, e.g. figs. 12, 13), and also inland near Ida Creek, south-southwest of Bowen (Paine & others, 1974, e.g. figs. 7, 8, 9, 17, 18).

Other felsic-mafic intrusions

Net-veined complexes in which felsic and mafic magmas were intruded more or less simultaneously are a type of composite intrusion (Blake & others, 1965). Another common type, also found in north Queensland, are intrusive sheets, mainly dykes, consisting of mafic margins and felsic central parts. In these intrusions, the mafic magma appears to have preceded and prepared the way for the felsic magma, which was injected before the mafic component had completely crystallised. In some cases, the felsic component contains rounded inclusions of the mafic component, especially near its margins. These inclusions and also the mafic margins of the intrusions may show chilled (finer-grained) and liquid-to-liquid (intricately crenulate) contacts with the adjacent felsic component. Two Proterozoic examples of this type of composite intrusion, still recognisable, although metamorphosed to granitic gneiss and amphibolite, are exposed in the Cameron River, 16 km north-northeast of Mary Kathleen, in the Mount Isa area. Another in the Mount Isa area may be represented by the Garden Creek Porphyry, northeast of Dajarra (Blake & others, 1978): this is a dyke-like body of porphyritic microgranite, intruding metasediments (Mount Guide Quartzite), which is almost invariably flanked by metadolerite.

Significance of net-veined complexes

Net-veined complexes are of two main types. In one of these, represented by most if not all of the Mount Isa region examples, the felsic component was crystalline rock when intruded and subsequently melted and mobilised by mafic magma, and it may be much older than the mafic component. In the other type, represented by Icelandic net-veined complexes and probably by some of those in northeastern Queensland, the felsic component was partly or entirely liquid when invaded by mafic magma, and the two components are, therefore, closely comparable in age, though they may be genetically unrelated. Both types of net-veined complex are formed in the upper crust, probably within a few kilometres of the surface.

Net-veined complexes of the first type indicate that in the upper crust mafic magma is commonly, perhaps usually, incapable of assimilating sufficient granitic rock to produce much hybrid material of andesitic or dacitic composition. Even where relatively large volumes of mafic magma have been involved, as in the Mount Erle Complex near Duchess, only very small amounts of intermediate hybrid rocks are present. The likelihood of voluminous andesitic and dacitic volcanic rocks resulting from assimilation of crystalline rocks or even glassy rocks, as suggested by Eichelberger (1974), by mafic magma in the upper crust seems remote.

Although many mafic dykes intrude granitic rocks, in only relatively few cases is there evidence of them causing some melting of these rocks. One reason for this may be that such melting is largely restricted to situations where the granitic rock, because of recent emplacement or metamorphism and tectonism, is already hot, perhaps within 100° to 200°C of its melting temperature, when intruded by the mafic magma. This would explain, for example, why no melting of granite has taken place where the Burstall Granite/Lunch Creek Gabbro net-veined complex east of Mary Kathleen, in the Mount Isa area, is cut by a large but relatively young dyke of Lakeview Dolerite.

The second type of net-veined complex may be formed where mafic magma is injected into felsic magma in a high-level magma chamber, perhaps beneath an active volcanic centre. The injection of mafic magma results in physical mixing of the two and may trigger an explosive eruption, as suggested by Blake (1969) and Sparks & others (1977). Sparks & others have shown that felsic magma is likely to become rapidly superheated when physically mixed with much hotter (by about 350°C) mafic magma, and vigorous convection may take place; if such is the case the felsic magma may become supersaturated in volatiles, resulting in vesiculation of the magma. Pressure in the magma chamber will increase, due to the injection of basic magma and to vesiculation, and may be sufficient to cause fractures to be developed in the overlying rock. Highly fluid felsic magma, charged with gases, can then escape through the fractures to reach the surface and erupt explosively. The products of such eruptions will consist mainly of felsic material, but are likely to contain some inclusions representing the denser mafic magma. This mechanism accounts for the widespread occurrence of mafic inclusions in felsic pyroclastic deposits. A good example is exposed 35 km north of Duchess, in the Mount Isa region, where numerous small ellipsoidal mafic inclusions are contained in foliated felsic meta-ignimbrite forming part of the Precambrian Corella Formation.

In some cases the two contrasting magmas may become so thoroughly mixed, perhaps as a result of particularly vigorous convection, that a homogeneous magma of intermediate composition is produced in the high-level magma chamber. Rocks attributed to such mixing commonly contain phenocrysts derived from both mafic and felsic magmas (e.g., Walker & Skelhorn, 1965; Eichelberger, 1978). However, in net-veined complexes mixing is much less thorough, and the mafic and felsic components retain their separate identities.

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References

BAIN, J. H. C., OVERSBY, B. S., MACKENZIE, D. E., WITHNALL, I. W., MCLEOD, R. L., DRAPER, J. J., & SCOTT, P. A., 1980—Georgetown project; *in* Geological Branch Summary of Activities 1979. *Bureau of Mineral Resources, Australia, Report* **222**, 175-89.

BLAKE, D. H., 1966—The net-veined complex of the Austurhorn intrusion, southeastern Iceland. *Journal of Geology*, **74**, 891-907.

BLAKE, D. H., 1969—Welded tuffs and the Maelifell caldera, Alftafjordur volcano, southeastern Iceland. *Geological Magazine*, **106**, 531-41.

BLAKE, D. H., 1972—Regional and economic geology of the Herberton/Mount Garnet area—Herberton Tinfield, north Queensland. *Bureau of Mineral Resources, Australia, Bulletin* **124**.

BLAKE, D. H., DONCHAK, P. J. T., & BULTITUDE, R. J., 1978—Precambrian geology of the Dajarra 1:100 000 Sheet area, northwestern Queensland—preliminary data. *Bureau of Mineral Resources, Australia, Record* **1978/46** (unpublished).

BLAKE, D. H., ELWELL, R. W. D., GIBSON, I. L., SKELHORN, R. R., & WALKER, G. P. L., 1965—Some relationships resulting from the intimate association of acid and basic magmas. *Quarterly Journal of the Geological Society of London*, **121**, 31-49.

BULTITUDE, R. J., BLAKE, D. H., & DONCHAK, P. J. T., 1978—Precambrian geology of the Duchess 1:100 000 Sheet area, northwestern Queensland—preliminary data. *Bureau of Mineral Resources, Australia, Record* **1978/112** (unpublished).

DERRICK, G. M., 1980—Marraba, Queensland. *Bureau of Mineral Resources, Australia, 1:100 000 Geological Map Commentary*.

DONCHAK, P. J. T., BLAKE, D. H., & JAQUES, A. L., 1979—Precambrian geology of the Mount Angelay 1:100 000 Sheet area (7055), northwestern Queensland—preliminary data. *Bureau of Mineral Resources, Australia, Record* **1979/93** (unpublished).

EICHELBERGER, J. C., 1974—Magma contamination within the volcanic pile: origin of andesite and dacite. *Geology* **2**, 29-33.

EICHELBERGER, J. C., 1978—Andesitic volcanism and crustal evolution. *Nature*, **275**, 21-7.

PAINE, A. G. L., GREGORY, C. M., & CLARKE, D. E., 1970—Geology of the Ayr 1:250 000 Sheet area, Queensland. *Bureau of Mineral Resources, Australia, Report* **128**.

PAINE, A. G. L., CLARKE, D. E., & GREGORY, C. M., 1974—Geology of the northern half of the Bowen 1:250 000 Sheet area, Queensland (with additions to the geology of the southern half). *Bureau of Mineral Resources, Australia, Report* **145**.

SPARKS, S. R. J., SIGURDSSON, H., & WILSON, L., 1977—Magma mixing: a mechanism for triggering acid explosive eruptions. *Nature*, **267**, 315-8.

TAYLOR, T. R., VOGEL, T. A., & WILBAND, J. T., 1980—The composite dikes at Mount Desert Island, Maine: an example of coexisting acidic and basic magmas. *Journal of Geology*, **88**, 433-44.

WALKER, G. P. L., & SKELHORN, R. R., 1966—Some associations of acid and basic igneous rocks. *Earth-Science Reviews*, **2**, 93-109.

WIEBE, R. A., 1980—Commingling of contrasted magmas in the plutonic environment: examples from the Nain anorthositic complex. *Journal of Geology*, **88**, 197-209.

YODER, H. S. Jr, 1973—Contemporaneous basaltic and rhyolitic magmas. *American Mineralogist*, **58**, 153-71.

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Shallow inter-reefal structure of the Capricorn group, southern Great Barrier Reef

P. J. Davies, J. F. Marshall, H. Hekel[1], *& D. E. Searle*[1]

276 km of continuous seismic reflection profiles have been obtained from the inter-reef areas of the Capricorn Reefs. Five reflectors were identified, three being widespread: reflector A, the shallowest, is found only close to modern reefs; reflector C is widespread and varies from flat-lying to irregular; the deepest reflector, D, is limited to the area north and west of Heron Island, where it is flat-lying. All the reflectors are assumed to represent erosional surfaces, and the ages of the intervening sequences are not known. Either reflector C, or the present sea bottom and reflector A, represent the surface of the Holocene transgression.

Introduction

The inter-reef areas of the southern Great Barrier Reef were investigated by the Bureau of Mineral Resources and Geological Survey of Queensland in April 1977, with the aim of defining the depth and configuration of Pleistocene unconformities in these areas.

Continuous seismic reflection profiling was accomplished with an EG & G 'UNIBOOM' system operating at an energy level of 200-300 joules. The filtered return signal was recorded on an E.P.C. recorder. Velocity data for time-depth interpretation were obtained from previous (Davies & others, 1977b, Harvey, 1977) and concurrent refraction profiles on the reefs. Positions were fixed with a Motorola R.P. system and radar distance triangulation. The location of the seismic profiles, which totalled 276 line km, is shown in Figure 1. Copies of the original seismic sections are contained in Searle & others (1978).

Previous work

Major seismic reflection surveys in the southern Great Barrier Reef have been conducted by Gulf Oil (1966, 1967) and by the Bureau of Mineral Resources (Marshall, 1977). A few high resolution boomer lines have been run by the University of Queensland (Jell & Flood, 1977). Limited resistivity surveys (Davies, 1974), and extensive refraction surveys (Davies & others, 1977b; Harvey, 1977) have also been carried out on the reefs. The refraction surveys have helped delineate a Holocene sequence with a seismic velocity of 1500 m sec^{-1} and a pre-Holocene sequence with a seismic velocity of 2400 m sec^{-1}. Petroleum exploration refraction data also indicate a velocity of 2400 m sec^{-1} for layers close to the sea bed. Sediment distribution studies have been conducted by Maxwell & Maiklem (1964), Maiklem (1967, 1970) and Marshall (1977). Drill holes at Heron Island (Richards & Hill, 1942) and Wreck Island (Humber, 1960) show a Pleistocene to Holocene sequence 140-190 m thick (Palmieri, 1971; Hekel, 1973). Major solution unconformities, probably representing subaerial erosion, have been recognised within the Heron Island sequence at −20 m, −35 m, and −96 m. (Davies, 1974).

Results

Seafloor morphology

Seismic records and echo-sounder profiles show the following features. Between the reefs, the seafloor is flat or nearly so. However, a large easterly-trending bank, informally named Isbell Shoal, connects Northwest, Wilson, and Broomfield Reefs, rising steeply from −45 m to −20 m near Wilson, but less steeply on the northern side (Fig. 2E). The bank is about 25 km long, more than 5 km wide, and 6-13 m high. Maiklem (1968) stated that it appears to be mainly hard rock with loose coral gravel.

Where the reefs are close together, the bottom is rough and often steep. Maiklem (1968) noted that in the channel between Heron and Wistari Reefs, the slope is as steep as 45°. In general, leeward slopes are gentler than windward slopes close to the reefs.

Well-defined sub-horizontal surfaces, terraces and 'nick' points occur on most traverses. Two prominent surfaces occur at −5 m to −7 m and around −15 m. The −5 m to −7 m surface occurs on the reefal slope of Heron, Erskine, Masthead, Polmaise, Wistari, Sykes, Wreck, and Tryon Reefs, and on Isbell Shoal. The −15 m surface is present on the reefal slope of Heron and Lamont Reefs, and on Rock Cod Shoal. A prominent terrace at 20-22 m is present on some profiles. Many others can not be correlated between traverses. The origin of such features is unclear.

Seismic reflectors

Line drawings from representative seismic reflection profiles are shown in Figures 2 and 3. Along most sections sub-bottom reflections are limited to areas where water is deeper than 10 m, and the inter-reefal seismic reflectors could not, therefore, be traced across the reefs. Five reflectors (A-E) have been identified, although reflector B was seen only west of Rock Cod Shoal (Fig. 1), and reflector E is a weak and transient event of uncertain significance. The distribution and depth variations of reflectors A, C, & D are shown in Figures 2 and 3. Depths to sub-bottom reflectors in this report are expressed in milli-seconds (ms) two-way travel time.

Reflector A, where present, is the shallowest event. It is often limited to the northern slope of Holocene reefs (Figs. 3A, 3B, 3D, 3E, 3F) but occurs on both the southern and northern sides of Isbell Shoal and Tryon Reef (Fig. 2E, 2F). It is absent in the lee of Irving Shoal. North of Polmaise, Masthead, Wistari, Heron, Tryon, and Fitzroy Reefs, and Isbell Shoal, reflector A slopes away from the reef into the inter-reefal areas. Its depth varies between 20 and 65 ms below sea level.

Reflector C is a strong event, which has been identified on all the lines shown in Figure 1. In most inter-reefal areas, it is flat-lying (Fig. 2B), but it is irregular in places (Figs. 2A, 2C, 2E). In two sections, it is stepped (Figs. 2D, 3D) similar to the present-day seafloor. Reflector C rises both towards known reefal areas

1. Geological Survey of Queensland (GSQ).

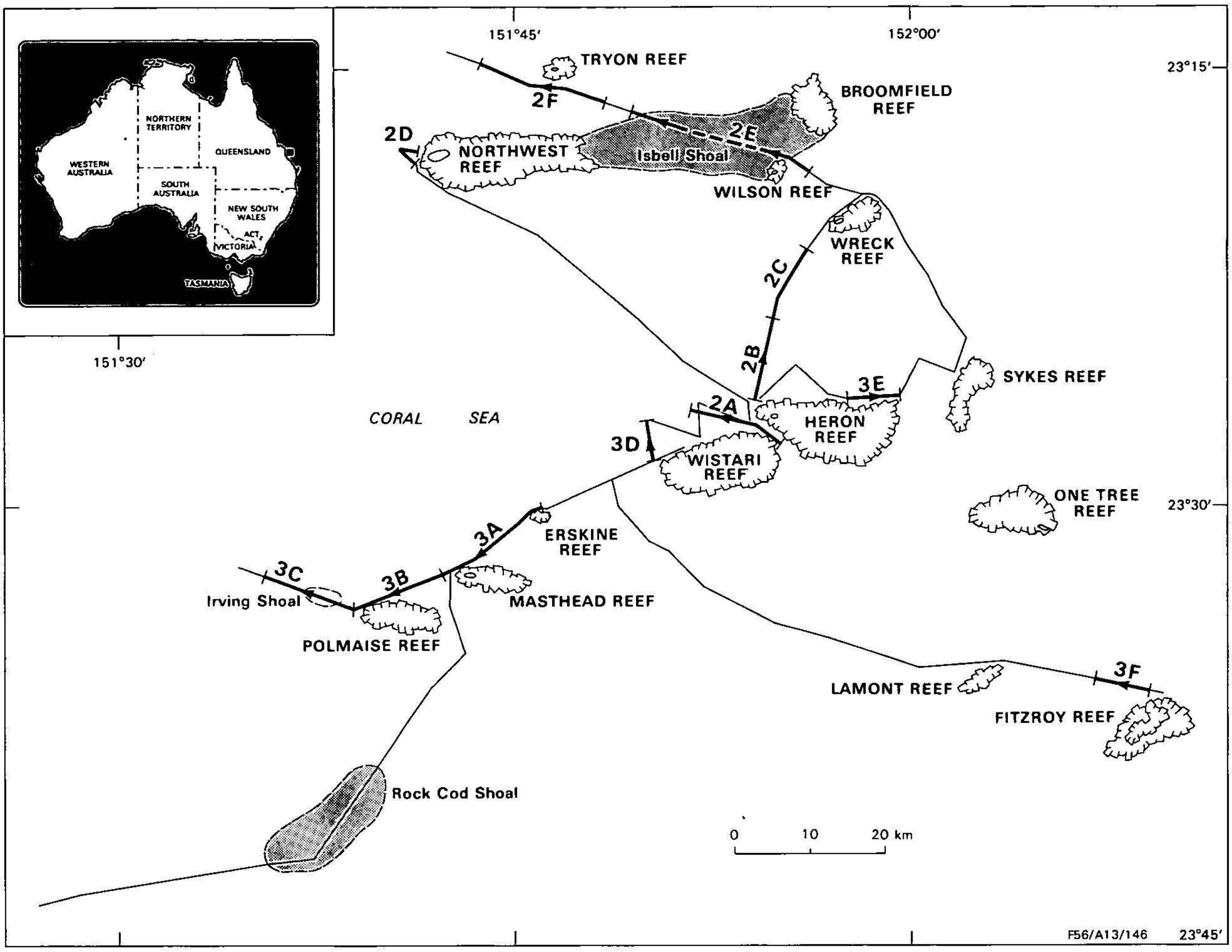

Figure 1. Locality map, showing position of seismic reflection profiles.
Numbered sections (full lines) refer to seismic sections shown in Figures 2 and 3. Shaded area between Northwest, Broomfield, and Wilson Reefs shows the approximate position of Isbell Shoal; shaded area south of Polmaise Reef shows the approximate position of Rock Cod Shoal.

(Figs. 2B, 2F, 3D and towards Isbell and Rock Cod Shoals. Its total depth range is 25-72 ms. It crops out to the west-southwest of Wistari Reef, and may be close to the seafloor south of Masthead Reef.

Reflector D is also a strong event, but is only seen northwest of a line joining Wreck and Heron Reefs. It is not identifiable on the section between Wistari and Fitzroy Reefs. In the inter-reefal areas, it is generally flat-lying (Figs. 2B, 2C) and sub-parallel to reflector C. Between Rock Cod Shoal and Masthead Reef, there is some evidence of erosion. Reflector D rises towards the reefs, is often irregular on the flanks (Fig. 2D), and is not visible below the reefs (Fig. 2B). In some inter-reefal (Fig. 2C) and marginal reef areas (Fig. 2B) it is truncated by reflector C. The depth of reflector D varies from 28-62 ms.

Ages of sequences between reflectors

Important questions arising from the seismic data are the ages of the units bounded by the seismic reflectors: in particular, the age of the units between the seafloor and reflector A, and the seafloor (or reflector A) and reflector C. Although there is no direct indication of age, some idea may be gained from a consideration of three possible alternatives:

1. All units below the seafloor are pre-Holocene.
2. The sequence between reflector C and the seafloor is Holocene.
3. Only the sequence between reflector A and the seafloor is Holocene.

The first alternative is unlikely in view of the abundant Holocene reef growth and resultant dispersal of sediment into the inter-reefal area. However, alternatives 2 and 3 can be considered in more detail.

The unit between reflectors C and A could represent diachronous shallow-water deposits laid down during the postglacial transgression. This would require correlation of reflector C with the first solution unconformity at −20 m in the Heron Island Borehole (Fig. 2B). If this is so, then 5-20 m of sediment was deposited over an area of at least 2000 km^2 in 10-12000 years. Considering that the sediment would probably consist of bioclastic reefal material, this amounts to an enormous production of calcium carbonate, at least an order of magnitude higher than normally produced by Holocene reefs (Davies, 1977). A radiocarbon date of 4950 ± 70 years B.P. (Maxwell, 1969) for a surface grab sample of inter-reefal sediment indicates extremely slow sedimentation in areas closely associated with reefs, at least since that time. It is also extremely unlikely that erosion rates during the last low sea level were high enough over large enough areas to have produced sufficient sediment for re-deposition during the postglacial transgression.

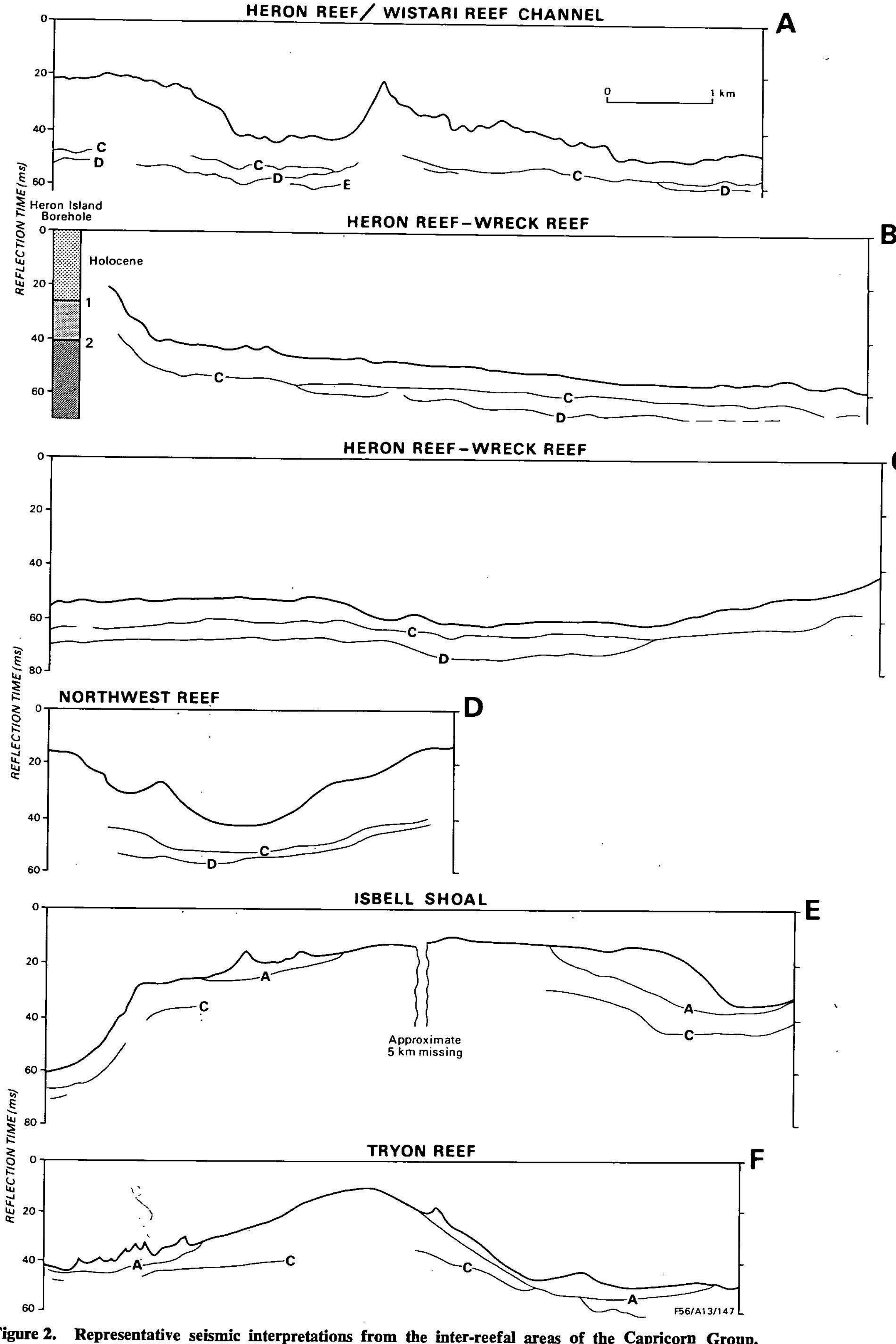

Figure 2. **Representative seismic interpretations from the inter-reefal areas of the Capricorn Group.**
Location of sections is shown in Figure 1. Section 2B shows the approximate depth of Holocene reef growth in the Heron Island borehole and the position of solution unconformities (numbered 1 and 2).

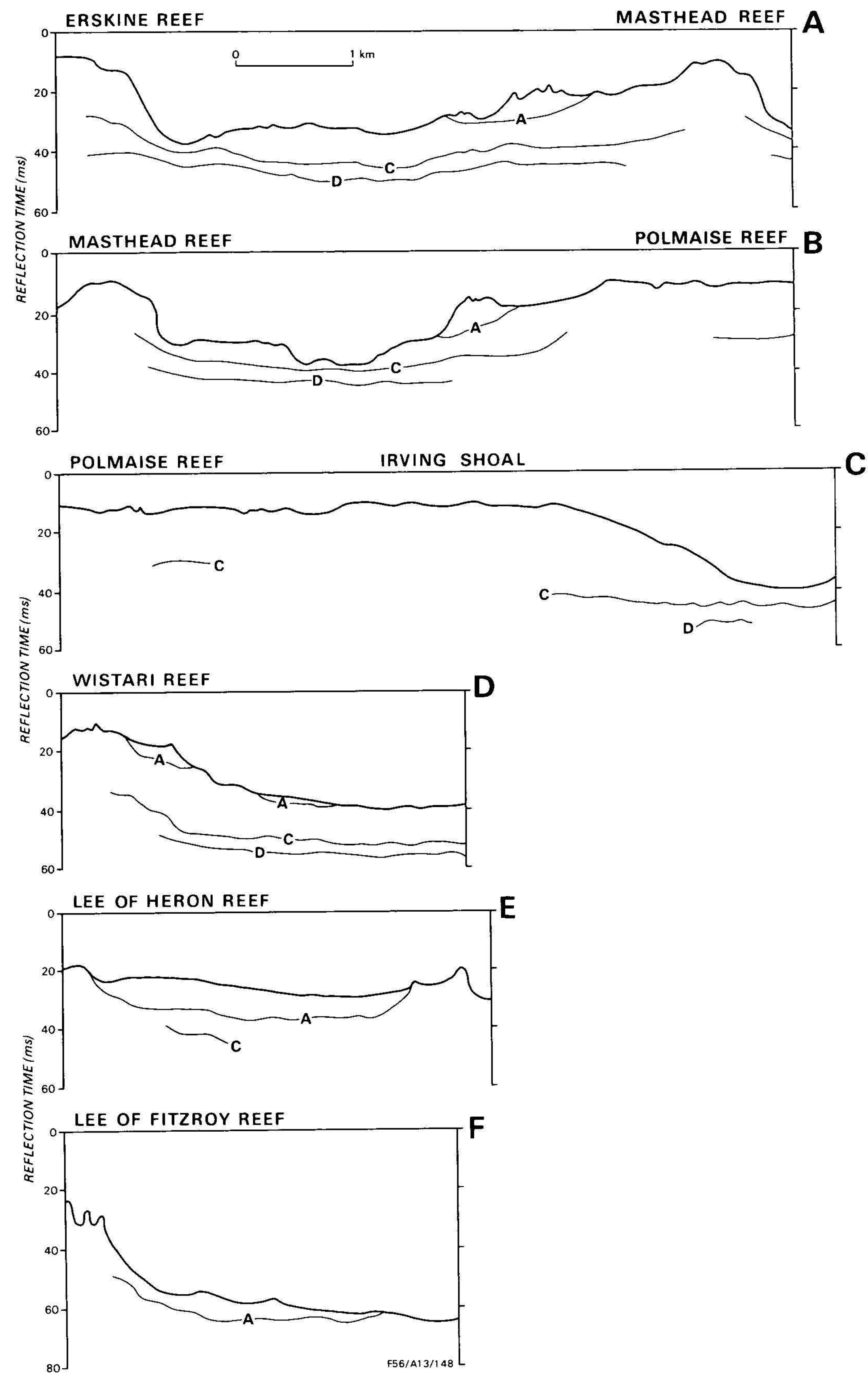

Figure 3. Representative seismic interpretations from the inter-reefal areas of the Capricorn Group. Location of sections is shown in Figure 1.

The third alternative limits Holocene sedimentation to the unit above reflector A, which is relatively thin and patchy, and generally limited to the leeward or lateral slopes of the present-day reefs. These characteristics are consistent with the sedimentation model invoking leeward growth, postulated by Davies & others (1977a,b), which proposes that leeward sand accretion becomes dominant only after the reef has reached sea level. Thus, sediments above reflector A do not appear to occur in the lee of Irving Shoal, a reef which has not yet reached sea level. More important, however, volumes of sediment in the leeward wedges above reflctor A are consistent with the sediment production rates on the reef (Davies & Kinsey, 1977) and the sedimentation rates on the leeward margins (Davies, 1977). The presence of reflector A on both sides of Isbell Shoal does, however, present a problem. There is little evidence of a modern reef, although Northwest, Wilson, and Bloomfield occupy parts of the shoal, and Jell & Flood (1977) referred to 'coral growth on the platform'. Maiklem (1968) described the top of the bank as mainly hard rock with loose coral gravel. The sediment above reflector A on Isbell Shoal may, therefore, be the product of reef growth and leeward accumulation during the Holocene or may represent a similar facies on a much older reef. Evidence of a low-amplitude sand-wave field on the leeward edge of Isbell Shoal (Jell & Flood, 1977) suggests the former. Presumably, sediment deposited on the shallow (−7 m to −11 m) crest of the shoal has been winnowed by currents and re-deposited on the flanks. A combination of tidal currents and dominant southeasterly swell has resulted in the bulk of this sediment being deposited on the leeward edge.

If reflector A, in shallow water, and present seafloor, in deep water, approximate the surface of the Holocene transgression, then 1) the inter-reef areas are floored by Pleistocene rocks eroded during the last sea-level low, although it is likely that much of the transgressed surface is mantled with a veneer of sediment too thin to be resolved by the seismic system used; 2) the Wisconsin erosion surface correlates with the highest solution unconformity in the Heron Island Borehole (Fig. 2B); and 3) Holocene reefs and associated sediments are not as extensively developed as their Pleistocene counterparts, i.e. not all Pleistocene reefal platforms are covered by Holocene reefs.

Acknowledgements

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References

DAVIES, P. J., 1974—Subsurface solution unconformities at Heron Island, Great Barrier Reef. *Proceedings, 2nd International Symposium on Coral Reefs, Brisbane.* 2, 273-78.

DAVIES, P. J., 1977—Modern reef growth—Great Barrier Reef. *Proceedings, 3rd International Symposium on Coral Reefs, Miami.* 2, 325-30.

DAVIES, P. J., & KINSEY, D. W., 1977—Holocene reef growth—One Tree Island, Great Barrier Reef. *Marine Geology*, 24, M1-M11.

DAVIES, P. J., MARSHALL, J. F., FOULSTONE, D., THOM, B. G., HARVEY, N., SHORT, A. D., & MARTIN, K., 1977a—Reef growth, southern Great Barrier Reef—preliminary results. *BMR Journal of Australian Geology & Geophysics*, 2, 69-72.

DAVIES, P. J., MARSHALL, J. F., THOM, B. G., HARVEY, N., SHORT, A. D., & MARTIN, K., 1977b—Reef development —Great Barrier Reef. *Proceedings, 3rd International Symposium on Coral Reefs, Miami,* 2, 331-7.

GULF (AUSTRALIAN GULF OIL CO), 1966—Swain Reefs seismic project, phase II, 1966. *Bureau of Mineral Retralia, Petroleum Search Subsidy Acts Report* 66/11022 (unpublished).

GULF (AUSTRALIAN GULF OIL CO), 1967—Swain Reefs seismic project, phase 11, 1966. *Bureau of Mineral Resources, Australia, Petroleum Search Subsidy Acts Report* 66/11093 (unpublished).

HARVEY, N., 1977—Identity of subsurface solution disconformities on the Great Barrier Reef, Australia, 15°-17°S, using shallow seismic refraction techniques. *Proceedings, 3rd International Symposium on Coral Reefs, Miami.* 2, 45-52.

HEKEL, H., 1973—Late Oligocene to Recent nannoplankton from the Capricorn Basin (Great Barrier Reef area). *Geological Survey of Queensland Publication* 359 (Palaeontological Papers, No. 33).

HUMBER (BARRIER REEF OILS PTY LTD), 1960—H.B.R. Wreck Island No. 1, Qld, well completion report. *Bureau of Mineral Resources, Australia, Petroleum Search Subsidy Acts Report* 62/1021 (unpublished).

JELL, J. S., & FLOOD, P. G., 1977—Guide to the geology of the Capricorn and Bunker Groups, Great Barrier Reef Province. *University of Queensland, Department of Geology, Papers,* 8(3).

MAIKLEM, W. R., 1967—Black and brown speckled foraminiferal sand from the southern part of the Great Barrier Reef. *Journal of Sedimentary Petrology* 37, 1023-30.

MAIKLEM, W. R., 1968—The Capricorn Reef Complex, Great Barrier Reef, Australia. *Journal of Sedimentary Petrology*, 38, 785-98.

MAIKLEM, W. R., 1970—Carbonate sediments in the Capricorn Reef Complex, Great Barrier Reef, Australia. *Journal of Sedimentary Petrology*, 40, 55-80.

MARSHALL, J. F., 1977—Marine geology of the Capricorn Channel area. *Bureau of Mineral Resources, Australia, Bulletin* 163.

MAXWELL, W. G. H., 1969—Radiocarbon ages of sediment: Great Barrier Reef. *Sedimentary Geology*. 3, 331-3.

MAXWELL, W. G. H., & MAIKLEM, W. R., 1964—Lithofacies analysis, southern part of the Great Barrier Reef. *University of Queensland, Department of Geology, Papers,* 5, 1-21.

PALMIERI, V., 1971—Tertiary subsurface biostratigraphy of the Capricorn Basin. *Geological Survey of Queensland Report* 52.

RICHARDS, H. S., & HILL, D., 1942—Great Barrier Reef bores, 1926 and 1937. Descriptions, analyses and interpretations. *Report of the Great Barrier Reef Committee*, 5, 1-122.

SEARLE, D. E., DAVIES, P. J., HEKEL, H., KENNARD, J., MARSHALL, J. F., & THOM, B. G., 1978—Preliminary results of a continuous seismic profiling survey in the Capricorn Group, Southern Great Barrier Reef. *Geological Survey of Queensland Record* 1978/46 (unpublished).

BMR Journal of Australian Geology & Geophysics, 6, 1981, 106-108

Evidence for former evaporites in the Carboniferous Moogooree Limestone, Carnarvon Basin, Western Australia

Bruce Radke & Robert S. Nicoll

The former presence of evaporites within skeletal carbonate lithofacies of the Moogooree Limestone, Carnarvon Basin, is inferred from the presence of quartz geodes which have formed by the replacement of early diagenetic anhydrite nodules, and patchy dedolomitisation of much of the sequence.

The Moogooree Limestone is a Lower Carboniferous formation that crops out near the Minilya River on the eastern margin of the Carnarvon Basin, Western Australia (Fig. 1). The formation comprises limestone, dolostone, and recessive lithologies*, probably calcareous siltstone and shale. Chert nodules occur in association with silicified fossils (Playford & others, 1975). Skeletal carbonates predominate, and there are minor micritic carbonates lower in the section. It has been interpreted as a shallow-water marine sequence (Condon, 1965) with strand-line shelly carbonates in the middle and upper parts of the sequence (Lavaring, 1979). Quartz geodes are present in an eleven metre thick interval of thick-bedded limestone in about the middle of the type section (Fig. 2).

Geodes

The geodes are crude oblate spheroids (Fig. 3a), 5-15 cm in diameter, with surfaces of second-order spheroidal knobs that have a radial grooved texture. Such geodes have elsewhere been called cauliflower cherts (Chowns & Elkins, 1974). None of the Moogooree geodes shows recognisable fossil morphology.

Internally, the geodes have a crude concentric pattern; some have hollow centres. The outermost zone is chalcedonic, consisting of agglomerations of quartzine (length-slow chalcedony) rosettes, and grades inwards into a zone of megaquartz crystals which have euhedral

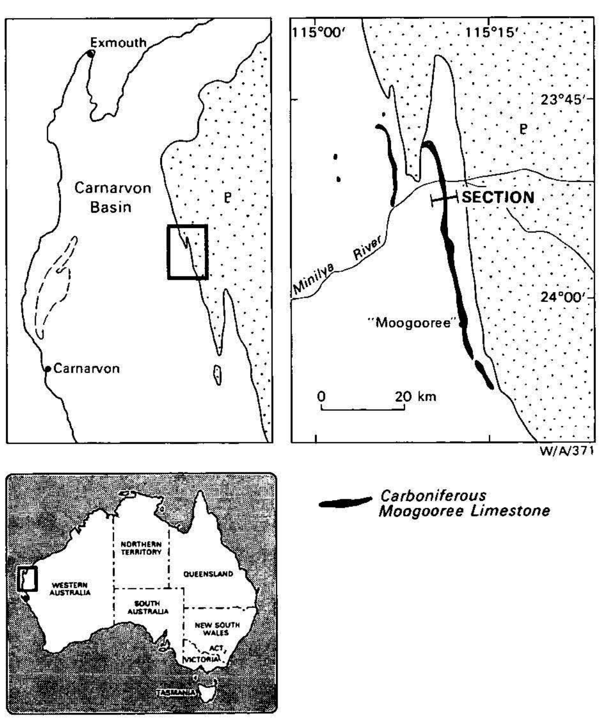

Figure 1. Extent of the Moogooree Limestone in outcrop, Carnarvon Basin, Western Australia.

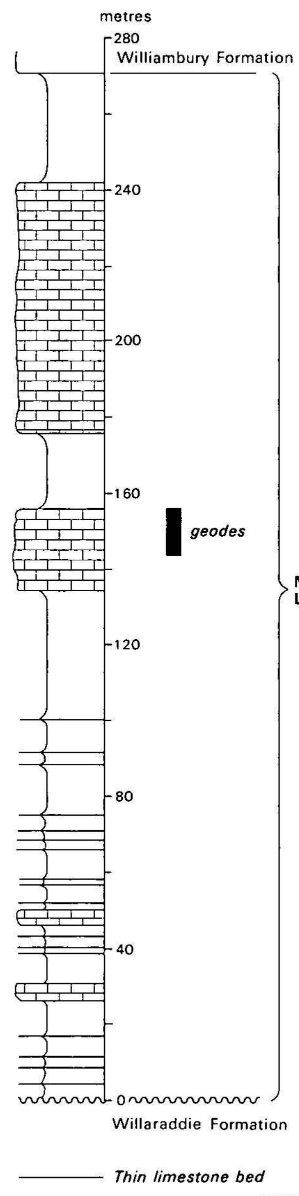

Figure 2. Generalised type section of Moogooree Limestone, Carnarvon Basin.
Geode horizon is indicated. Section is that indicated on Fig. 1.

* Recessive lithologies are those that tend not to crop out, as they are prone to rapid weathering and erosion.

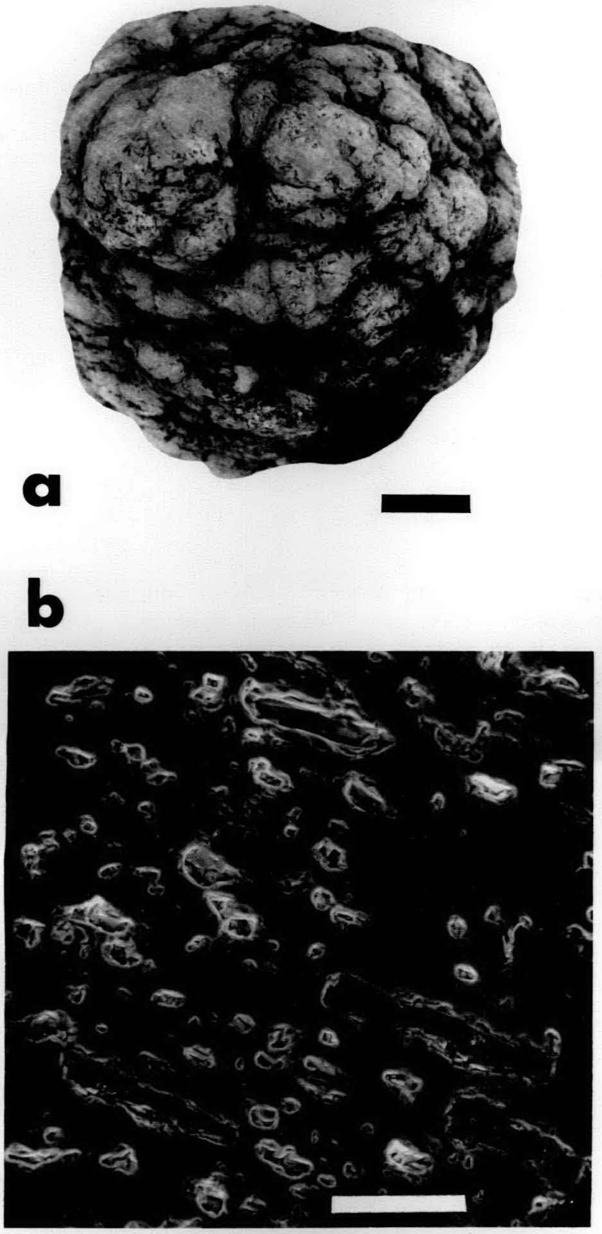

Figure 3. a. External view of geode, showing spheroidal shape with crudely felted external surface. (bar scale — 1 cm.)
b. Micro-texture in megaquartz crystals.
A relict felted-lath texture is preserved by numerous small anhydrite inclusions (bar scale—100 μm).

terminations in the central vug, if one is present. The crystal faces are commonly coated with botryoidal chalcedonite (length-fast chalcedony) or large calcite crystals. Some geodes have a crude geopetal structure in the vug, formed by the collapse of a delicate replacement silica sponge-work after solution of the sulphates. Felted-lath textures (Fig. 3b) are commonly preserved both in the outer quartzine rind and in the megaquartz, and are defined by patterns of inclusions, especially small relict anhydrite inclusions, which are 30-180 μm long.

Nodular pseudomorphs after anhydrite have previously been described and associated with former evaporites by Chowns & Elkins (1974) and Milliken (1979). Such nodules have been compared in origin to nodular anhydrite forming in the Holocene arid coastal sabkhas of the Persian Gulf by displacive growth within the carbonate sediments (Shearman, 1966; Kinsman, 1969).

Dedolomitisation

Carbonates hosting the geodes are pale orange-brown, with millimetre-sized crystalline textures. Relict primary textures of skeletal grainstone and packstone are preserved in places. Compostion is variable, from dolomitic limestone to calcareous dolostone. This irregular and patchy variation, as well as the ochreous colours that imply iron sesquioxide inclusions, is suggestive of dedolomite i.e. dolomite that has been altered to calcite (Von Morlot, 1947; de Groot, 1967) with oxidation and exsolution of iron. This process of incongruent dissolution of magnesium is most common in the weathering cycle of carbonates where sulphate has been present in the meteoric waters (Tatarsky, 1949; Yanat'eva, 1955; Lucia, 1961).

Summary

We have presented evidence for the former existence of early diagenetic sulphates in an interval of the Moogooree Limestone. The extent of evaporite overprint, however, has not been established. Relict carbonate textures and the faunal diversity of the sequence indicate that the host limestones probably accumulated under shallow-water normal-marine conditions (Condon, 1965), between wave base and strand line (Lavaring, 1979). With seaward progradation of this shoreline environment, suitable supratidal conditions probably became established with consequent growth of scattered sulphate nodules in the skeletal sediments.

To our knowledge, evaporitic overprints have not been previously reported from the Carboniferous of the Carnarvon Basin. It is probable that evaporites were more extensive in the Moogooree Limestone, especially in the recessive lithologies, than we have been able to document.

References

Chowns, T. M., & Elkins, J. E., 1974—The origin of quartz geodes and cauliflower cherts through the silicification of anhydrite nodules. *Journal of Sedimentary Petrology*, 44, 885-903.

Condon, M. A., 1965—The geology of the Carnarvon Basin, Western Australia, Part 1, pre-Permian stratigraphy. *Bureau of Mineral Resources, Australia, Bulletin* 77.

De Groot, K., 1967—Experimental dedolomitization. *Journal of Petrology*, 37, 1216-20.

Kinsman, D. J., 1969—Modes of formation, sedimentary associations and diagnostic features of shallow-water and supratidal evaporites. *American Association of Petroleum Geologists Bulletin* 53, 830-40.

Lavaring, I. G., 1979—Palaeoecological and palaeogeographic implications of *Rhipidomella michelini?* (Leveille) in the Carboniferous of the Carnarvon Basin. *Geological Survey of Western Australia, Annual Report* 1978, 89-92.

Lucia, F. J., 1961—Dedolomitization in the Tansill (Permian) Formation. *Geological Society of America Bulletin*, 72, 1107-10.

Milliken, K. L., 1979—The silicified evaporite syndrome —Two aspects of silicification history of former evaporite nodules from southern Kentucky and northern Tennessee. *Journal of Sedimentary Petrology*, 49, 245-56.

PLAYFORD, P. E., COPE, R. N., COCKBAIN, A. E., LOW, G. H., & LOWRY, D. C., 1975—Phanerozoic. *In* Geology of Western Australia. *Geological Survey of Western Australia, Memoir* 2, 277.

SHEARMAN, D. J., 1966—Origin of marine evaporites by diagenesis. *Institute of Mining and Metallurgy, Transactions,* 75, 208-15.

TATARSKY, V. B., 1949—Distribution of rocks in which dolomite is replaced by calcite. *Akad. Nauk. SSSR Doklady,* 69, 849-51.

VON MORLOT, A., 1947—Sur l'origine de la dolomite. *Acad. Sci. (Paris) Comptes rendus,* 26, 311-15.

YANAT'EVA, O. K., 1955—Effect of aqueous solutions of gypsum on dolomite in the presence of carbon dioxide. *Akad. Nauk. SSSR, Doklady,* 101, 911-12.

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Nannofossil biostratigraphy of the *Hantkenina* (foraminiferid) interval in the upper Eocene of southern Australia

Samir Shafik

Based on calcareous nannofossil evidence, the stratigraphically important interval with *Hantkenina (H.) alabamensis primitiva* Cushman & Jarvis in the upper Eocene of southern Australia is placed high in the foraminiferal Zone P. 16 of the tropics; previously it was correlated with the upper part of Zone P. 15. Consequently, the upper Eocene sediments above this interval at Browns Creek (Otway Basin) and Blanche Point (St Vincent Basin) are considered to represent an expanded section, with a very high rate of sedimentation. This is consistent with, and probably helps understanding of, the disjunct vertical distribution of several foraminiferal species reported previously in these sediments—particularly at Browns Creek.

Introduction

The biostratigraphic significance of the few known occurrences of the low-latitude foraminiferid *Hantkenina (H.) alabamensis primitiva* Cushman & Jarvis (Fig. 1) in the upper Eocene of southern Australia cannot be over-stressed, in view of the difficulties in correlating the foraminiferal succession, as noted by McGowran (1978, p. 89): "(1) The absence of any event that allows direct correlation with standard zones . . .; the faunas were extratropical. (2) Several species distributions are disjunct, and relative numbers of various species fluctuate enormously . . .". Correlation of these occurrences of *Hantkenina* with standard zones (e.g. Blow's P. Zones, 1969, 1979) should help in solving some of the difficulties inherent in the local biostratigraphy.

Forms assignable to *Hantkenina (H.) alabamensis primitiva* Cushman & Jarvis have been found in widely-separated areas in the St Vincent and Otway Basins (e.g. Parr, 1947; Glaessner, 1951; Lindsay, 1969; McGowran, 1973) and seem to represent a short-lived event, two important characteristics essential for regional and interregional correlations.

Discussion

Shortly after its discovery in Australia (Browns Creek section, Aire District, Otway Basin) by Parr (1947, as *H. alabamensis* subsp. *compressa*), this subspecies of *Hantkenina* was used in Australian biostratigraphy by Glaessner (1951), and has since been used by other investigators (e.g. Carter, 1958; Ludbrook, 1963, 1967; Wade, 1964; Lindsay, 1969; McGowran, 1973, 1978; Taylor *in* McGowran & others, 1971). It has also been used in correlation of foraminiferal schemes based on material from the Australian southern margin and New Zealand (e.g. McGowran, 1973).

Blow (1979) gave the range of *H. (H.) alabamensis primitiva* in the tropics as Zone P. 13 to earliest Zone P. 17; earlier he had suggested a younger level for its base, within Zone P. 15 (Blow, 1969). Associated nannofossils and foraminiferids indicate that the range of this taxon in southern Australia represents only part of its range in the tropics, but the question of which part seems difficult to answer with available foraminiferal data. McGowran (1973, p. 50) based his conclusion that the Browns Creek material correlated with Zone P. 15 on a view expressed by Blow (1969) that

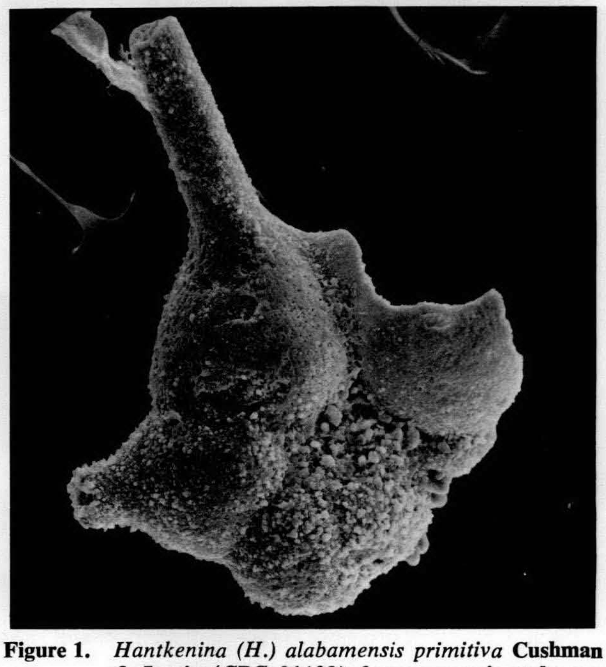


Figure 1. *Hantkenina (H.) alabamensis primitiva* **Cushman & Jarvis (CPC 21109) from approximately one metre above the base of the Blanche Point Marls, section at Uncle Tom's cabin near Blanche Point.**

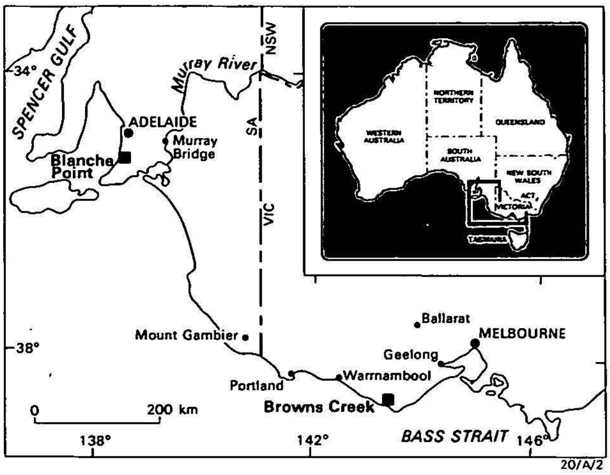


Figure 2. Locality map for the Blanche Point and Browns Creek sections.

this material represents the evolution of *Hantkenina primitiva* from *Pseudohastigerina*. Earlier, McGowran & others (1971) placed the *Hantkenina* interval as straddling the P.15/P.16 boundary. Recently, Blow (1979, p. 1162) stated that he "... has abandoned his earlier views that *primitiva* represented a phylogenetically isolated and separate development from a pseudohastigerine ancestor ...".

Aim of the study

The aim of this note is to investigate the calcareous nannofossil assemblages of *Hantkenina (H.) alabamensis primitiva*-bearing sediments from the St Vincent and Otway Basins to determine a) whether a correspondingly important nannofossil event exists in these sediments, and if so, b) its bearing on the evaluation of the Hantkenina interval in terms of the P. Zones.

Material studied

Samples studied are from immediately above the Tortachilla Limestone, at the base of the lower member of the Blanche Point Marls (Transitional Marls, section at Uncle Tom's Cabin near Blanche Point, St Vincent Basin, Fig. 2: see Reynolds, 1953 for stratigraphical details and location), and from the lower third of the Browns Creek Clays (Browns Creek section adjacent to the mouth of Johanna River, Otway Basin, Fig. 2; see Abele & others, 1976). *Hantkenina (H.) alabamensis primitiva* is restricted to an interval of about 60 cm, approximately one metre above the base of the Transitional Marl in the section at Uncle Tom's Cabin, but in the Browns Creek section it occurs in an interval of about 3 metres that includes the prominent *Notostrea* greensand and the topmost part of the underlying *Turritella* clays.

Figured specimens or their negatives are deposited in the Commonwealth Palaeontological Collection (CPC) in the Bureau of Mineral Resources, Canberra.

Calcareous nannofossil assmblages

The distribution of selected nannofossil taxa recovered from the material studied is shown in Table 1. Identification of these taxa was made mainly by optical microscopy, but illustration of some (see Figs. 3 & 4) was made by scanning electron microscopy.

Note on systematic palaeontology

Calcareous nannofossil taxa considered in this study are listed below; three new combinations are included. Bibliographic references of most of these taxa are provided by Loeblich & Tappan (1966, 1968, 1970a, 1970b, 1971, 1973); any not included in these publications are given in the references herein and identified by an asterisk.

Blackites amplus Roth & Hay in Hay, Mohler, Roth, Schmidt & Boudreaux, 1967
Blackites perlongus (Deflandre) n. comb. (basionym: *Rhabdolithus perlongus* Deflandre in Deflandre & Fert, 1954, Ann. Pal., vol. 40, p. 158, pl. 12, figs. 34-35; text-fig. 86).
Blackites spinosus (Deflandre) Hay & Towe, 1962
Blackites spinulus (Levin) Roth, 1970*
Blackites tenuis (Bramlette & Sullivan) Sherwood, 1974*
Blackites vitreus (Deflandre) n. comb. (basionym: *Rhabdolithus vitreus* Deflandre in Deflandre & Fert, 1954, Ann. Pal., vol. 40, pp. 157-158, pl. 12, figs. 28-29; text-figs. 83-84)
Braarudosphaera bigelowi (Gran & Braarud) Deflandre, 1947
Chiasmolithus altus Bukry & Percival 1971
Chiasmolithus oamaruensis (Deflandre) Hay, Mohler & Wade, 1966
Clathrolithus ellipticus Deflandre in Deflandre & Fert, 1954
Clausicoccus cribellum (Bramlette & Sullivan) Prins, 1979*
Coccolithus eopelagicus (Bramlette & Riedel) Bramlette & Sullivan, 1961
Coccolithus pelagicus (Wallich) Schiller, 1930
Corannulus germanicus Stradner, 1962
Cyclicargolithus floridanus (Roth & Hay) Bukry, 1971
Cyclicargolithus luminis (Sullivan) Bukry, 1971
Cyclicargolithus reticulatus (Gartner & Smith) Bukry, 1971
Cyclococcolithina formosa (Kamptner) Wilcoxon, 1970
Cyclococcolithina protoannula Gartner, 1971
Discoaster elegans Bramlette & Sullivan, 1961
Discoaster saipanensis Bramlette & Riedel, 1954
Discoaster tani tani Bramlette & Riedel, 1954
Discoaster tani nodifer Bramlette & Riedel, 1954
Discoaster tani ornatus Bramlette & Wilcoxon, 1967
Helicopontosphaera bramlettei Müller, 1970
Helicopontosphaera seminulum (Bramlette & Sullivan) Stradner, 1969
Isthmolithus recurvus Deflandre in Deflandre & Fert, 1954
Koczyia wechesensis (Bukry & Percival) Sherwood, 1974*
Lanternithus minutus Stradner, 1962
Micrantholithus attenuatus Bramlette & Sullivan, 1961
Micrantholithus truncus Bramlette & Sullivan, 1961
Neococcolithes dubius (Deflandre) Black, 1967
Orthozygus aureus (Stradner) Bramlette & Wilcoxon, 1967
Pemma basquensis (Martini) Báldi-Beke, 1971
Pemma papillatum Martini, 1959
Pentaster lisbonensis Bybell & Gartner, 1972*
Polycladolithus opersus Deflandre in Deflandre & Fert, 1954
Pontosphaera multipora (Kamptner) Roth, 1970
Pontosphaera plana (Bramlette & Sullivan) Haq, 1971
Reticulofenestra hampdenensis Edwards, 1973*
Reticulofenestra scissura Hay, Mohler & Wade, 1966

	Blanche Point Transitional Marl				Browns Creek 'Turritella' clays							greensand†	
	Base	0.6 m	0.9 m	1.5 m	0.9 m*	2.4 m	4.0 m	5.5 m	7.0 m	8.1 m	9.3 m	9.5 m	11.0 m
Blackites amplus	+				+		+				+	+	+
perlongus					+					+	+	+	
spinosus		+			+	+	+		+		+	+	+
spinulus	+	+	+	+	+	+	+	+	+	+	+	+	+
tenuis	+	+			+	+	+		+		+	+	+
vitreus					+	+	+	+	+		+	+	+
Braarudosphaera bigelowi			+		+	+	+			+	+	+	+
Chiasmolithus altus					+			+		?	?	+	+
oamaruensis		+			+	+	+	+	+	+	+	+	?
Clathrolithus ellipticus		+	+		+	+	+				+		
Clausicoccus cribellum	+				+	+	+	+	+				
Coccolithus eopelagicus	+	+			+	+	+	+	+	+	+	+	+
pelagicus	+	+	+	+	+	+	+	+	+	+	+	+	+
Corannulus germanicus	+					+	+						
Cyclicargolithus floridanus	+	+	+		+	+	+	+	+	+	+	+	+
luminis	+				+		+					?	
reticulatus	+	+	+		+	+	+	+	+	+	+		
Cyclococcolithina formosa	+				+	+	+		+	+	+	+	+
protoannula	+		+		+		+		+	+	+	+	+
Discoaster elegans						+	+		+	+	+		
saipanensis		+	+	+	+	+	+		+	+	+	+	+
tani nodifer					+				+	+	+	+	
tani ornatus					+	cf.			+	+	+		
tani tani	+	+		+	+	+		+			+	+	
Helicopontosphaera bramlettei							+				+	+	
seminulum					+	+				+	aff.	aff.	aff.
Isthmolithus recurvus		+	?	+				+	+	+	+	+	+
Koczyia wechesensis						+	+				+	+	
Lanternithus minutus	+	+	+	+	+	+	+	+	+	+	+	+	+
Micrantholithus attenuatus					+	+	+		+	+	+		
truncus						+	+					+	
Neococcolithes dubius	?				+	+							
Orthozygus aureus					+				+				
Pemma basquensis					+	+	+				+		
papillatum					+	+							
Pentaster lisbonensis						+							
Polycladolithus opersus					+			+					
Pontosphaera multipora					+	+	+		+	+	+	+	+
plana								+		+			+
Reticulofenestra hampdenensis	+				+	+	+	+	+		+	+	+
scissura	+	+	+	+	+	+	+	+	+	+	+	+	+
scrippsae	+	+	+	+	+	+	+	+	+	+	+	+	+
umbilica	+	+	+	+	+	+	+	+	+	+	+	+	+
Sphenolithus moriformis	+	+		+	+	+	+	+	+	+	+	+	+
Syracosphaera labrosa							+	+			+	+	
Transversopontis obliquipons				+	+		+	+	+	+	+	+	+
pulcher	+					+		+	+	+		+	
zigzag			+	+	+		+	+	+	+	+	+	+
Trochoaster simplex							+				+		+
Zygrhablithus bijugatus crassus	+	+	+	+	+	+	+	+	+	+	+	+	+
bijugatus crassus	+										+		+
Hantkenina (H.) alabamensis primitiva‡			+	+						+	+	+	+

* level from base of the '*Turritella*' clays; it represents the base of the marine/calcareous plankton upper Eocene section at Browns Creek.

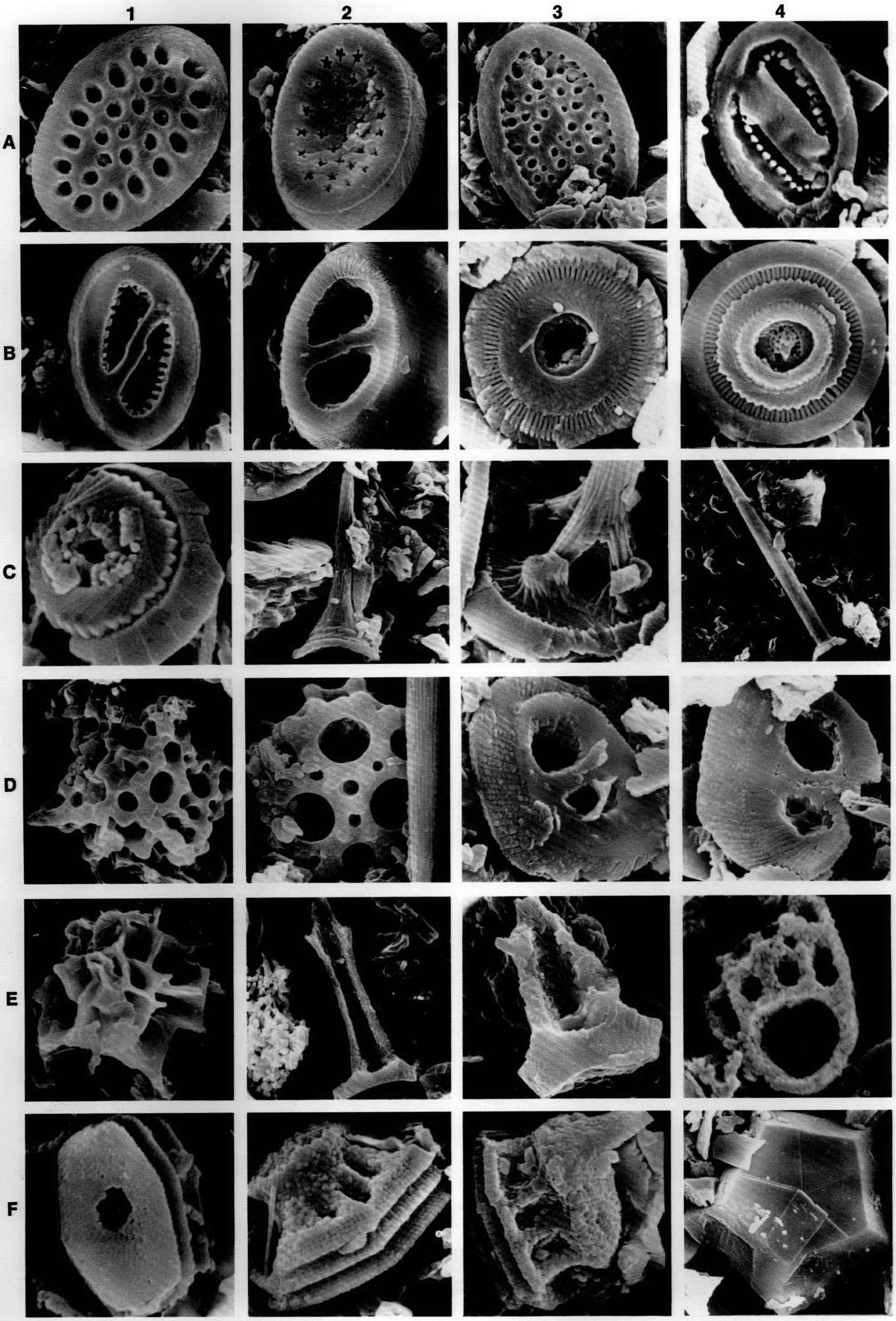
† greensand samples from near the base and top of the unit.

‡ this subspecies does not range above the tabulated samples.

Table 1. Distribution of selected calcareous nannofossil taxa and *Hantkenina (H.) alabamensis primitiva* in the lower parts of the Blanche Point and Brown Creek sections.

Figure 3. SEM micrographs of some calcareous nannofossil taxa from the Browns Creek section (magnifications are approximate).

A1—*Pontosphaera multipora* (Kamptner), distal view, CPC 21110, X8020; A2—*Pontosphaera multipora* (Kamptner), proximal view, CC 21111, X7600; A3—*Pontosphaera multivora* (Kamptner), distal view, CPC 21112, X4920; A4—*Transversopontis zigzag* Roth & Hay, proximal view, CPC 21113, X8780; B1—*Transversopontis zigzag* Roth & Hay, distal view, CPC 21114, X8350; B2—*Transversopontis obliquipons* (Deflandre), distal view, CPC 21115, X7600; B3—*Blackites spinosus* (Deflandre), proximal view, CPC 21116, X8730; B4—Blackites spinosus (Deflandre), distal view, CPC 21117, X8380; C1—*Blackites amplus* Roth & Hay, distal view, CPC 21118, X14 500; C2—*Blackites spinosus* (Deflandre), side view, CPC 21119, X4110; C3—*Blackites vitreus* (Deflandre), oblique distal, CPC 21120, X10 140; C4—*Blackites tenuis* (Bramlette & Sullivan), side view, CPC 21121, X2540; D1—*Trochoaster simplex* Klumpp, CPC 21122, X4510; D2—*Polycladolithus opersus* Deflandre, CPC 21123, X5240; D3—*Helicopontosphaera seminulum* (Bramlette & Sullivan), proximal view, CPC 21124, X6730; D4—*Helicopontosphaera seminulum* (Bramlette & Sullivan), distal view, CPC 21125, X6320; E1—*Sphenolithus moriformis* (Brönnimann & Stradner), CPC 21126, X7930; E2—*Zygrhablithus bijugatus bijugatus* (Deflandre), side view, CPC 21127, X3960; E3—*Zygrhablithus bijugatus bijugatus* (Deflandre), side view, CPC 21128, X6850; E4—*Orthozygus aureus* (Stradner), distal view, CPC 21129, X11 220; F1—*Lanternithus minutus* Stradner, proximal view, CPC 21130, X9250; F2—*Lanternithus minutus* Stradner, oblique distal, CPC 21131, X10 340; F3—*Lanternithus minutus* Stradner, oblique distal, CPC 21132, X9370; F4—*Braarudosphaera bigelowi* (Gran & Braarud), CPC 21133, X3390.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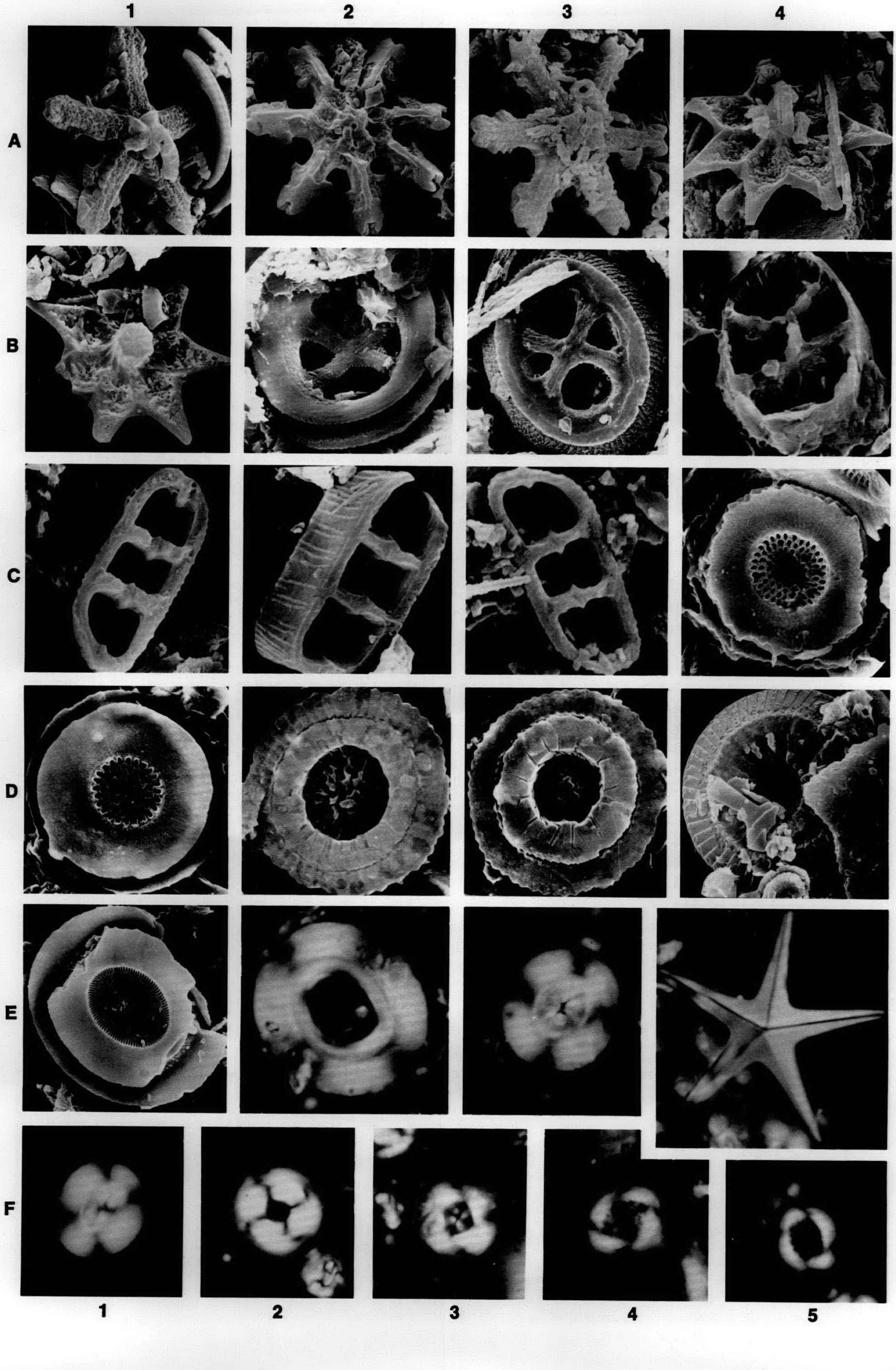
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Reticulofenestra scrippsae (Bukry & Percival) n. comb. (Basionym: *Dictyococcites scrippsae* Bukry & Percival, 1971, Tulane Stud. Geol. Paleont., v. 8(2), p. 128, pl. 12, figs. 7-8)
Reticulofenestra umbilica (Levin) Martini & Ritzkowski, 1968
Sphenolithus moriformis (Brönnimann & Stradner) Bramlette & Wilcoxon, 1967
Syracosphaera labrosa Bukry & Bramlette, 1969
Transversopontis obliquipons (Deflandre) Hay, Mohler & Wade, 1966
Transversopontis pulcher (Deflandre) Perch-Nielsen, 1967
Transversopontis zigzag Roth & Hay in Hay, Mohler, Roth, Schmidt & Boudreaux, 1967
Trochoaster simplex Klumpp, 1953
Zygrhablithus bijugatus bijugatus (Deflandre) Deflandre, 1959
Zygrhablithus bijugatus crassus Locker, 1967

Palaeoenvironmental analysis

During the *Hantkenina* ingression in southern Australia, warm conditions must have prevailed. Calcareous nannofossil evidence for this climatic warming in the two sections investigated is slender. However, species such as *Chiasmolithus oamaruensis* (Deflandre) and *C. altus* Bukry & Percival, whose geographic distribution suggests cooling, are extremely scarce or absent in samples containing *Hantkenina*. In contrast, the questionable warm-water species *Discoaster tani nodifer* Bramlette & Riedel, *D. tani ornatus* Bramlette & Wilcoxon, and *D. saipanensis* Bramlette & Riedel are frequent to common; *Discoaster barbadiensis* Tan, which occurs in coeval sediments deposited seemingly under warm-water conditions in the Carnarvon Basin, is absent or extremely rare in the material studied.

Calcareous nannofossils seem more diverse below the *Hantkenina* interval than within the interval itself. This is taken to suggest that some taxa, including *Clausicoccus cribellum* (Bramlette & Sullivan) *Corannulus germanicus* Stradner, and *Cyclicargolithus luminis* (Sullivan), preferred cool conditions.

The consistent occurrence of the fossil holococcoliths *Lanternithus minutus* Stradner and *Zygrahblithus bijugatus bijugatus* Deflandre, and the sporadic appearance of several other taxa, such as *Braarudosphaera bigelowi* (Gran & Braarud), *Clathrolithus ellipticus* Deflandre, *Micrantholithus* spp., *Pemma* spp., *Pontosphaera* spp., *Orthozygus aureus* (Stradner) and *Transversopontis* spp., suggest that deposition was in shallow marine, near-shore or shelf environments.

Biostratigraphy

For biostratigraphic usage, the assemblages recovered can be described as monotonous, except for the appearance of *Isthmolithus recurvus* Deflandre below the *Hantkenina* interval and the disappearance of *Cyclicargolithus reticulatus* (Gartner & Smith) within this interval. The disappearance of *Neococcolithes dubius* (Deflandre) below the appearance of *Isthmolithus recurvus* in the Browns Creek material (Fig. 5) is consistent with results from elsewhere in Australia (Shafik, unpublished data).

The geographic distribution of *Isthmolithus recurvus* and *Cyclicargolithus reticulatus* in Australia and elsewhere suggests that both thrived in the temperate belt, although *I. recurvus* is known to be prominent among cool-water assemblages, and *C. reticulatus* seems to have preferred warmer conditions. This can be taken as a measure of the biostratigraphic reliability of both species in high to mid-latitude areas such as southern Australia.

The disappearance of *C. reticulatus* after the appearance of *I. recurvus* in the material studied, and elsewhere in Australia (Shafik, 1973), is consistent with results from abroad (see e.g. Gartner, 1971).

Appearance of Isthmolithus recurvus. Since Hay & others (1967) proposed the earliest appearance of *I. recurvus* as a biostratigraphic criterion for defining the base of a zone bearing the same name, this biostratigraphic event has been used consistently by other investigators in upper Eocene zonations (e.g. Martini, 1970b, 1971; Gartner, 1971; Edwards, 1971). Martini (1970a) regarded *I. recurvus* as a non-tropical species, and Bukry (1975, 1977a) included it among the upper Eocene-lower Oligocene cool-water indicators; Gartner (1971) emphasised the importance of its earliest appearance by calling it a datum. *I recurvus* has been recorded from widely separated land-based and oceanic sections (e.g. Martini, 1971; Bukry, 1973).

Most nannofossil investigators (e.g. Martini, 1971; Gartner, 1971; Roth & others, 1971) seem to agree that the earliest appearance of *Isthmolithus recurvus* is within the younger part of Zone P. 15 of Blow (1969).

Extinction of Cyclicargolithus reticulatus. Although *Cyclicargolithus reticulatus* has been reported in upper Eocene sediments from widely separated areas, particularly in hemipelagic sediments (e.g. Gartner, 1971; Proto Decima & others, 1975; Bukry, 1973, 1977a-b), its extinction level has not been used as widely in the upper Eocene biostratigraphy of these areas. In southern Australia, this horizon is particularly useful and appears

Figure 4. SEM and OM micrographs of some calcareous nannofossil taxa from the Browns Creek section (magnifications of SEM micrographs are approximate); magnification of OM micrographs* is 2000.

A1—*Discoaster tani nodifer* Bramlette & Riedel, CPC 21134, X3790; A2—*Discoaster* sp. aff. *D. tani nodifer* Bramlette & Riedel, CPC 21135, X3410; A3—*Discoaster tani ornatus* Bramlette & Wilcoxon, CPC 21136, X3120; A4—*Discoaster saipanensis* Bramlette & Riedel, CPC 21137, X4050; B1—*Discoaster saipanensis* Bramlette & Riedel, CPC 21138, X4700; B2—*Chiasmolithus oamaruensis* (Deflandre), proximal view, CPC 21139, X3400; B3—*Chiasmolithus oamaruensis* (Deflandre), distal view, CPC 21140, X3190; B4—*Neococcolithes dubius* (Deflandre), CPC 21141, X8480; C1—*Isthmolithus recurvus* Deflandre, CPC 21142, X6280; C2—*Isthmolithus recurvus* Deflandre, CPC 21143, X7490; C3—*Isthmolithus recurvus* Deflandre, CPC 21144, X7980; C4—*Cyclicargolithus reticulatus* (Gartner & Smith), proximal view, CPC 21145, X5770; D1—*Cyclicargolithus reticulatus* (Gartner & Smith), proximal view, CPC 21146, X6500; D2—*Cyclicargolithus reticulatus* (Gartner & Smith), distal view, CPC 21147, X6980; D3—*Cyclicargolithus reticulatus* (Gartner & Smith), distal view, CPC 21148, X6610; D4—*Cyclococcolithina formosa* (Kamptner), distal view, CPC 21149, X3920; E1—*Reticulofenestra umbilica* (Levin), proximal view, CPC 21150, X3480; E2*—*Reticulofenestra umbilica* (Levin), CPC 21151; E3*—*Reticulofenestra scissura* Hay, Mohler & Wade, CPC 21152; E4*—*Micrantholithus attentuatus* Bramlette & Sullivan, CPC 21153; F1*—*Reticulofenestra scrippsae* (Bukry & Percival), CPC 21154; F2*—Cyclococcolithina formosa (Kamptner), CPC 21155; F3*—Cyclicargolithus reticulatus (Gartner & Smith), CPC 21156; F4*—*Reticulofenestra hampdenensis* Edwards, CPC 21157; F5*—*Reticulofenestra hampdenensis* Edwards, CPC 21158.

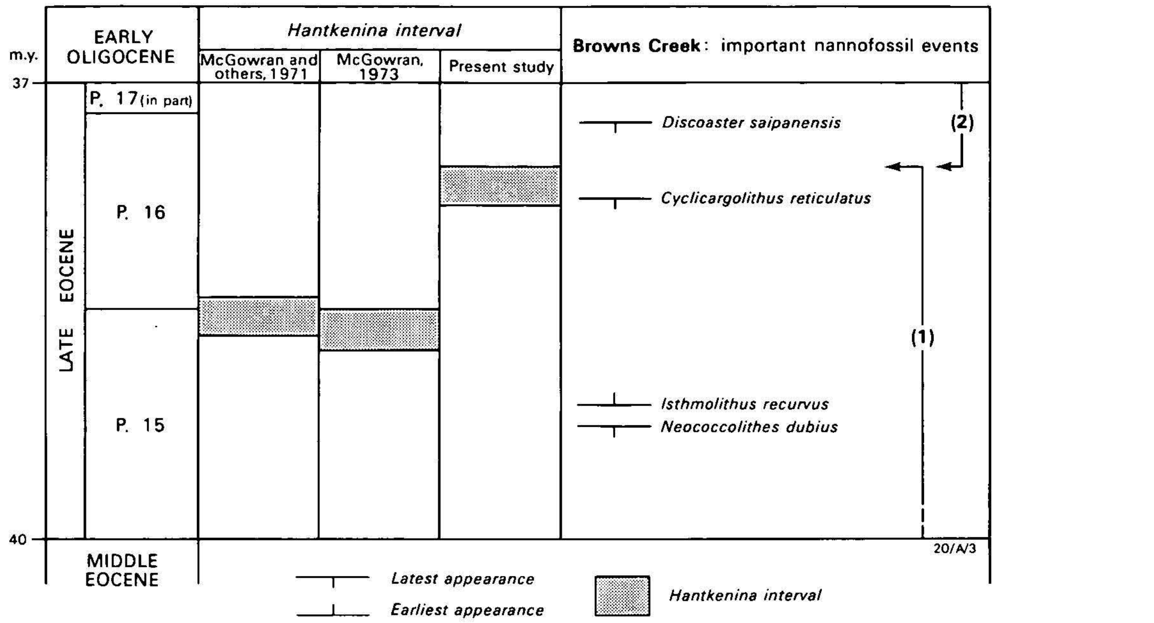

Figure 5. Important calcareous nannofossil events in the Browns Creek section, and position of the *Hantkenina* interval relative to the P. zonation.
(1) represents about 11 m of sediments; (2) represents about 28 m of sediments. P. Zones after Blow, 1969, 1979. Absolute dates after Hardenbol & Berggren, 1978.

consistently between the earliest appearance of *Isthmolithus recurvus* and the extinction level of *Discoaster saipanensis* (Shafik, 1973); the overlap in the ranges of *I. recurvus* and *D. saipanensis* indicates a late Eocene age. Distribution charts given by Gartner (1971) and Proto Decima & others (1975) put the extinction level of *C. reticulatus* high in Zone P. 16 of Blow (1969).

Evaluation of the Hantkenina interval. The present study indicates that the *Hantkenina* (*H.*) *alabamensis primitiva* ingression into southern Australia occurred during the late Eocene, after the introdcution of the nannofossil species *Isthmolithus recurvus,* which equates with the later part of Zone P. 15 of the tropics. This is consistent with results from New Zealand (Hornibrook & Edwards, 1971), where a similar (and most likely isochronous) ingression of *Hantkenina* occurred subsequent to the introduction of *I. recurvus.* Considering Blow's (1979) range for the subspecies, the incoming of *Hantkenina* (*H.*) *alabamensis primitiva* into southern Australia (and New Zealand) must have occurred either late in Zone P. 15 or sometime during the interval of Zones P. 16 to earliest P. 17. However, the extinction of the nannofossil species *Cyclicargolithus reticulatus* at a level high within Zone P. 16, during the *Hantkenina* ingression in southern Australia, seems more conclusive as a basis for evaluation.

Cyclicargolithus reticulatus has been regarded as a tropical to temperate species (Bukry, 1977b), and therefore its extinction within the *Hantkenina* interval must be interpreted as a true extinction, and not an extinction in response to the climatic warming evidenced by the *Hantkenina* ingression. The same species has been considered among the shallow-marine indicators (Bukry, 1974). Although this consideration means that the species is likely to disappear prior to its true extinction in a deepening environment, there is no evidence among the nannofossil assemblages to suggest that a substantial increase in water depth occurred during the *Hantkenina* ingression in southern Australia.

The indications are, therefore, that at the time of extinction of *Cyclicargolithus reticulatus,* during the later part of Zone P. 16, *Hantkenina* (*H.*) *alabamensis primitiva* was already in southern Australia. It it is assumed that the *H.* (*H.*) *alabamensis primitiva* ingression in southern Australia was a short-lived event (and there is no evidence to the contrary), it may be concluded that this ingression occurred essentially during the late part of Zone P. 16 (Fig. 5); the same conclusion may well be valid for the *H. alabamensis* ingression in New Zealand.

For this reason and because of the known duration of Zone P. 17, the upper Eocene sediments above the *Hantkenina* (*H.*) *alabamensis primitiva* interval in southern Australia (Lindsay, 1969; McGowran, 1978), and probably also those in New Zealand, must be regarded as an expanded section with a very high rate of sedimentation (Fig. 5). Over 28 m of sediments above the *Hantkenina* interval in the Browns Creek section have been assigned a late Eocene age on foraminiferal evidence by McGowran (1978): the nannofossil evidence does not disagree with this result. A similar thickness of upper Eocene sediments above the *Hantkenina* interval in equivalents of the Blanche Point section in the St Vincent Basin has been reported by Lindsay (1969). The observations made by McGowran (1978) on the foraminiferal succession above the *Hantkenina* interval in the Browns Creek section (which include, for example, disjunct vertical distribution of several species) strengthen the evidence for a very rapid sedimentation rate (during fluctuating climatic conditions; McGowran, 1978) subsequent to the *Hantkenina* ingression.

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Browns Creek results are based on two sets of samples: a set provided by Mr David J. Taylor and another set collected during a visit in 1977. Blanche Point samples, information about their content of *Hantkenina,* and the specimen illustrated in Figure 1 were supplied by Dr Richard J. F. Jenkins. I am grateful to both these colleagues.

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References

ABELE, C., GLOE, C. S., HOCKING, J. B., HOLDGATE, G., KENLEY, P. R., LAWRENCE, C. R., RIPPER, D., & THRELFALL, W. F., 1976—Tertiary. *In* DOUGLAS, J. G., & FERGUSON, J. A. (editors), Geology of Victoria. *Geological Society of Australia, Special Publication* 5, 177-273.

BLOW, W. H., 1969—Late middle Eocene to Recent planktonic foraminiferal biostratigraphy. *In* BRÖNNIMANN, P., & RENZ, H. H. (editors), *Proceedings of the First International Conference on Planktonic Microfossils, Geneva, 1967*, 1, 199-422.

BLOW, W. H., 1979—The Cainozoic Globigerinida. *E. J. Brill, Leiden.*

BUKRY, D., 1974—Phytoplankton stratigraphy, offshore East Africa, Deep Sea Drilling Project Leg 25. *In* SIMPSON, E. S. W., SCHLICH, R., & others, *Initial Reports of the Deep Sea Drilling Project*, 25, 635-46.

BUKRY, D., 1975—Coccolith and silicoflagellate stratigraphy near Antarctica, Deep Sea Drilling Project, Leg 28. *In* HAYES, D. E., FRAKES, L. A., & others, *Initial Reports of the Deep Sea Drilling Project*, 28, 709-23.

BUKRY, D., 1977a—Coccolith and silicoflagellate stratigraphy, South Atlantic ocean, Deep Sea Drilling Project, Leg 29. *In* SUPKO, P. R., PERCH-NIELSEN, K., & others, *Initial Reports of the Deep Sea Drilling Project*, 39, 825-39.

BUKRY, D., 1977b—Cenozoic coccolith and silicoflagellate stratigraphy, offshore Northwest Africa, Deep Sea Drilling Project Leg 41. *In* LANCELOT, Y., SEIBOLD, E., & others, *Initial Reports of the Deep Sea Drilling Propject*, 41, 689-707.

*BYBELL, L., & GARTNER, S., 1972—Provincialism among mid-Eocene calcareous nannofossils. *Micropaleontology*, 18(3), 319-36.

CARTER, A. N., 1958—Tertiary foraminifera from the Aire District, Victoria. *Geological Survey of Victoria, Bulletin* 55, 1-76.

EDWARDS, A. R., 1971—A calcareous nannoplankton zonation of the New Zealand Paleogene. *In* FARRINACCI, A. (editor), *Proceedings of the 2nd Planktonic Conference, Roma 1970*, 1, 318-419.

*EDWARDS, A. R., 1973—Key species of New Zealand calcareous nannofossils. *New Zealand Journal of Geology and Geophysics*, 16(1), 68-89.

GARTNER, S., 1971—Calcareous nannofossils from the JOIDES Blake Plateau cores, and revision of Paleogene nannofossil zonation. *Tulane Studies in Geology and Paleontology*, 8(3), 101-21.

GLAESSNER, M. F., 1951—Three foraminiferal zones in the Tertiary of Australia. *Geological Magazine*, 88, 273-83.

HARDENBOL, J., & BERGGREN, W. A., 1978—A new Paleogene numerical time scale. *American Association of Petroleum Geologists, Studies in Geology*, 6, 213-34.

HAY, W. W., MOHLER, H. P., ROTH, P. H., SCHMIDT, R. R., & BOUDREAUX, J. E., 1967—Calcareous nannoplankton zonation of the Cenozoic of the Gulf Coast and Carribean-Antillean area and transoceanic correlation. *Gulf Coast Association of Geological Societies, Transactions, 17*, 428-80.

HORNIBROOK, N. DE B., & EDWARDS, A. R., 1971—Integrated planktonic foraminiferal and calcareous nannoplankton datum levels in the New Zealand Cenozoic. *In* FARINACCI, A. (editor), *Proceedings of the 2nd Planktonic Conference, Roma 1970*, 1, 649-57.

LINDSAY, J. M., 1969—Cainozoic foraminifera and stratigraphy of the Adelaide Plains Sub-basin, South Australia. *Geological Survey of South Australia, Bulletin*, 42, 7-60.

LOEBLICH, A. R., Jr, & TAPPAN, H., 1966—Annotated index and bibliography of the calcareous nannoplankton. *Phycologia*, 5(2-3), 81-216.

LOEBLICH, A. R., Jr, & TAPPAN, H., 1968—Annotated index and bibliography of the calcareous nannoplankton II. *Journal of Paleontology*, 42(2), 584-98.

LOEBLICH, A. R., Jr, & TAPPAN, H., 1969—Annotated index and bibliography of the calcareous nannoplankton III. *Journal of Paleontology*, 43(2), 568-88.

LOEBLICH, A. R., Jr, & TAPPAN, H., 1970a—Annotated index and bibliography of the calcareous nannoplankton IV. *Journal of Paleontology*, 44(3), 558-74.

LOEBLICH, A. R., Jr., & TAPPAN, H., 1970b—Annotated index and bibliography of the calcareous nannoplankton V. *Phycologia*, 9(2), 157-74.

LOEBLICH, A. R., Jr, & TAPPAN, H., 1971—Annotated index and bibliography of the calcareous nannoplankton VI. *Phycologia*, 10(4), 315-39.

LOEBLICH, A. R., Jr, & TAPPAN, H., 1973—Annotated index and bibliography of the calcareous nannoplankton. *Journal of Paleontology*, 47(4), 715-59.

LUDBROOK, N. H., 1963—Correlation of the Tertiary rocks of South Australia. *Royal Society of South Australia, Transactions*, 87, 5-15.

LUDBROOK, N. H., 1967—Correlation of the Tertiary rocks of the Australasian region. *In* Tertiary correlations and climatic changes in the Pacific. *11th Pacific Science Congress, Tokyo, 1966, Symposium* 25, 7-19.

McGOWRAN, B., 1973—Observation Bore No. 2, Gambier Embayment of the Otway Basin: Tertiary micropalaeontology, and stratigraphy. *South Australia Department of Mines Mineral Resources Review*, 135, 43-55.

McGOWRAN, B., 1978—Early Tertiary foraminiferal biostratigraphy in southern Australia: a progress report. *Bureau of Mineral Resources, Australia, Bulletin*, 192, 83-95.

McGOWRAN, B., LINDSAY, J. M., & HARRIS, W. K., 1971—Attempted reconciliation of Tertiary biostratigraphic systems, Otway Basin. *In* WOPFNER, H., & DOUGLAS, J. G. (editors), The Otway Basin of southeastern Australia. *Geological Surveys of South Australia and Victoria, Special Bulletin*, 273-81.

MARTINI, E., 1970a—The upper Eocene Brockenhurst Bed. *Geological Magazine*, 107, 225-8.

MARTINI, E., 1970b—Standard Palaeogene calcareous nannoplankton zonation. *Nature*, 226, 560-1.

MARTINI, E., 1971—Standard Tertiary and Quaternary calcareous nannoplankton zonation. *In* FARINACCI, A. (editor), *Proceedings of the 2nd Planktonic Conference, Roma 1970*, 2, 739-85.

PARR, W. J., 1947—An Australian record of the foraminiferal genus *Hantkenina*. *Royal Society of Victoria, Proceedings*, 58, 45-7.

PROTO DECIMA, F., ROTH, P. H., & TODESCO, L., 1975—Nannoplancton calcareo del Paleocene e dell'Eocene della Sezione di Possagno. *Schweizerische Paläontologische Abhandlungen*, 97, 35-161.

*PRINS, B., 1979—Notes on Nannology-1. *Clausicoccus*, a new genus of fossil coccolithophorids. *INA (Newsletter, Proceedings of the International Nannoplankton Association)*, 1(1), N2-N4.

REYNOLDS, M. A., 1953—The Cainozoic succession of Maslin and Aldinga Bays, South Australia. *Royal Society of South Australia, Transactions*, 76, 114-46.

*ROTH, P. H., 1970—Oligocene calcareous nannoplankton biostratigraphy. *Eclogae geolgicae Helvetiae*, 63(3), 799-881.

ROTH, P. H., BAUMANN, P., & BERTOLINO, V., 1971—Late Eocene-Oligocene calcareous nannoplankton from Central and Northern Italy. In FARINACCI, A. (editor), *Proceedings of the 2nd Planktonic Conference, Roma 1970*, 2, 1069-97.

SHAFIK, SAMIR, 1973—Eocene-Oligocene biostratigraphy in the western and southern margins of Australia. *Abstracts, 45th Congress, Australia and New Zealand Association for the Advancement of Science*, Section 3, 101-3.

*SHERWOOD, R. W., 1974—Calcareous nannofossil systematics, paleontology, and biostratigraphy of the middle Eocene Weches Formation of Texas. *Tulane Studies in Geology and Paleontology*, 11(1), 1-79.

WADE, M., 1964—Application of the lineage concept to biostratigraphic zoning based on planktonic foraminifera. *Micropaleontology*, 10(37), 273-90.

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