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The Queensland Plateau

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ABSTRACT

The Queensland Plateau, a large fragment of Australian continental crust lying more than 1000 m below sea level off northern Queensland, was covered by extensive marine geophysical traversing in 1971 and tested by stratigraphic drilling in the same year. This Bulletin presents the results of stratigraphic mapping of the Plateau using the acoustic stratigraphy available from seismic reflection profiling tied to the drill site. The sedimentary and structural history of the Plateau is deduced and compared with generalised models of rift-margin evolution; considerable departure from the 'norm' is demonstrated.

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SUMMARY

The Queensland Plateau is a large block of submerged continental crust embedded in the continental margin off northeast Queensland. The Plateau and its bounding troughs (Queensland and Townsville) were surveyed in 1971 by Compagnie Générale de Géophysique under contract to the Bureau of Mineral Resources. Seismic reflection profiling, gravity and magnetic sensing, and bathymetric profiling were conducted on a systematic grid of traverse lines. One of the lines approaches very closely a Deep Sea Drilling Project hole drilled on the outer edge of the Plateau.

The basement structure of the Queensland Plateau has been mapped and strongly resembles the typical structural framework of the onshore Tasman Geosyncline. The offshore structures have been sharply truncated on their eastern, or outer, edge by a steep slope with a linear northwest trend. The slope is interpreted as having resulted from a rifting event which split off the northeastern corner of the Tasman Geosyncline.

Between 500 and 1500 metres of sediment overlies basement on the Plateau. In order to decipher the depositional history of the Plateau, seismic reflection profiles of the Bureau of Mineral Resources were matched with results from Deep Sea Drilling Project drilling. Three acoustically different units were mapped on reflection profiles, and they are equivalent to the three lithological units encountered in the drill-hole. The units consist of middle Eocene largely terrigenous rocks and upper Eocene hemipelagic rocks separated by a regional unconformity from a sequence of Neogene pelagic sediments. By studying the distribution and structure of the units, the uppermost Mesozoic and Tertiary history of the Queensland Plateau has been determined.

The evolutionary history consists of an uplift leading to an erosional phase in the Late Cretaceous to middle Eocene, differential subsidence in the late Eocene and Oligocene, then uniform subsidence in the Neogene. Sediment characteristic of each phase in the scheme was deposited. The general trend is for increasing pelagic and decreasing terrigenous content through time.

With an evolutionary model for the Queensland Plateau determined, a complete history of plateau and basin evolution is proposed by relating the Plateau evolution to that of the adjoining Coral Sea Basin using results from Deep Sea Drilling Project drilling in the Basin. The model is then compared with Falvey's (1974) model of Atlantic continental margin formation and a significant departure is demonstrated. No rift-valley stage appears to have preceded continent break-up and ocean formation. This is explained by using two lithospheric thermal anomalies, one in the Late Cretaceous and one in the Palaeocene-Eocene, and by suggesting rapid rifting of the Coral Sea Basin associated with the younger heat source.

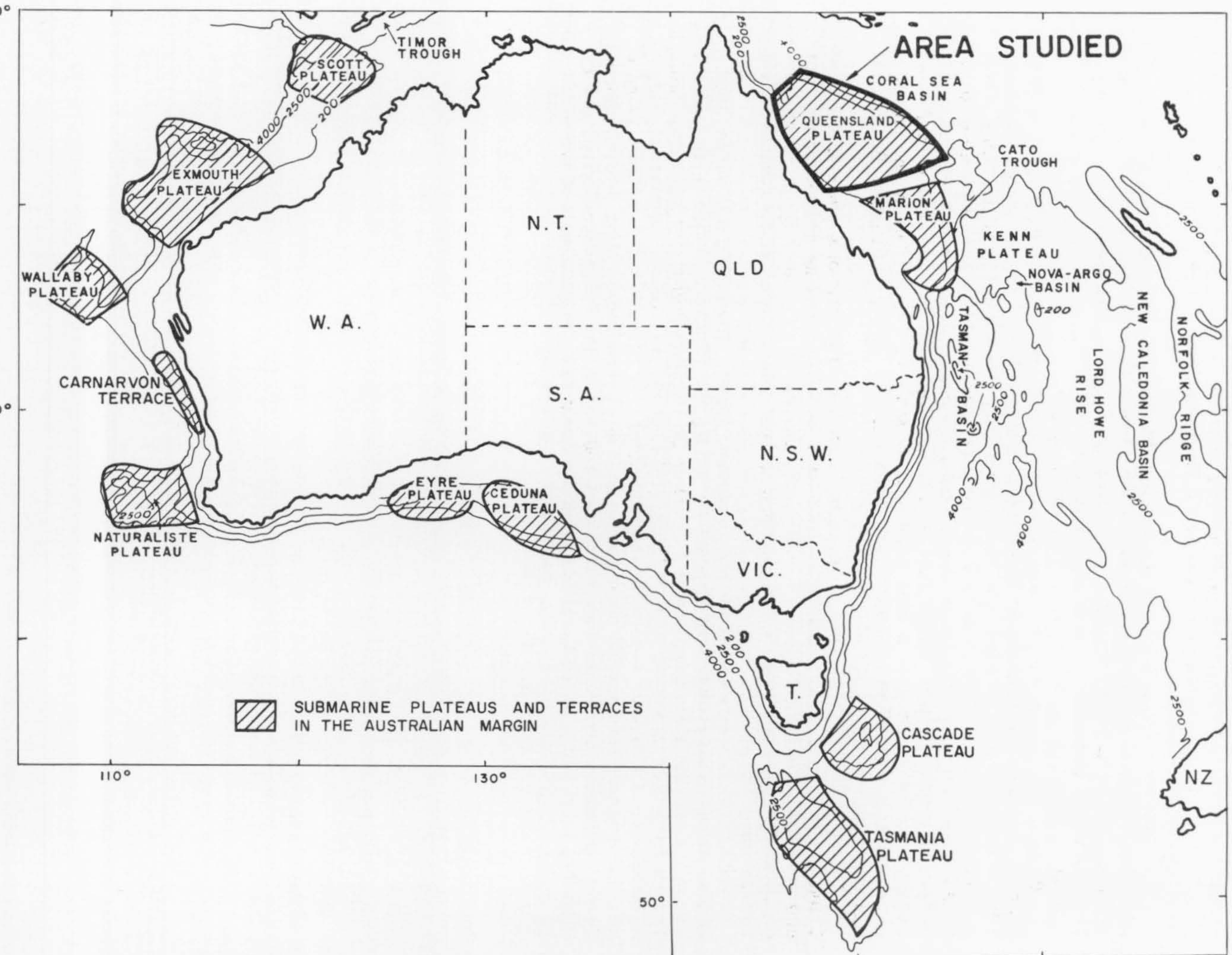


Fig. 1. Location diagram.

INTRODUCTION

Submarine plateaus and terraces are commonly found embedded in many inactive continental margins, including that of Australia. Many of these features, some of which occupy extensive areas of the sea-floor at depth intermediate between shelf and abyssal plain, have been investigated by geophysical methods or by drilling and have been shown to have a structural and genetic relation with the adjoining continental block. In general they have been formed by the action of relatively recent oceanic tectonism on ancient continental structures. That is, they constitute an integral part of the transition zone between continent and ocean.

The Queensland Plateau is the largest of eleven similar submarine structures identified in the Australian continental margin (Fig. 1). It is bounded by well-developed troughs, lies at a median depth of 1100 m below sea level, and is studded with numerous growing coral reefs. The extensive reef growth makes the Queensland Plateau almost unique and possibly points to a slower rate of subsidence than that of other plateaus.

Subsidence of a plateau is a manifestation of continent-ocean interaction along a developing continental margin. Tracing the history of subsidence of a marginal plateau should yield information on the processes in play at the continental edge. In the case of the Queensland Plateau the adjacent ocean basin has been classified by Karig (1971b) as a Western Pacific 'marginal basin'. Such ocean basins apparently do not form by continental splitting in the scheme described for Atlantic-type oceans by Dewey & Bird (1970) and more recently by Falvey (1974); they are believed to have formed by extensional processes acting behind island arc/trench systems. As such, their geological histories and that of the adjacent continental margin may differ greatly from their Atlantic counterpart.

Since the earliest days of oceanographic surveying, ships from many institutes have been investigating the Coral Sea region. Studies intensified in the 1960s with the development of modern surveying techniques and instrumentation. The two most significant advances in data gathering in the area were made in late 1971. They were the drilling of the ocean floor by the *Glomar Challenger*, and the systematic multi-sensor geophysical coverage of the continental margin off Queensland by the Bureau of Mineral Resources (BMR). Combination of these two distinctly different sets of data has enabled a thorough description of the Queensland Plateau. From this the geological history of the Plateau can be described, and from that the relations to surrounding regions can be assessed and implications made about the regional tectonics.

GEOLOGICAL SETTING

The Queensland Plateau lies on the continental edge of what Karig (1971b) has termed a Western Pacific 'marginal basin'. His definition is based on the 'tectonic position' of basins which lie behind volcanic chains of island arc systems. However, his definition is a combined geological and geographic one which he uses 'not only because most are marginal to continents but because they are produced by tectonic activity occurring along the margins of converging lithospheric plates'. Such an origin has not yet been demonstrated for the Coral Sea Basin and should not be assumed based only on analogy with similar basins. In the context of this discussion, the term 'marginal basin' is used in its geographic sense only. In fact, it would seem unwise to use a tectono/geographic term to describe a large group of basins, not all of which have been demonstrated to have the same tectonic history.

The Coral Sea Basin is an inactive marginal basin with normal heat flow in Karig's classification. It lies adjacent to a continental block on the southwest, where a relatively simple margin is formed by the linear outer edge of the Queensland Plateau (Fig. 2). Elsewhere, the basin margin is much more complex. It is delimited to the north by the Papuan Plateau and Louisiade Archipelago, to the east by a large plateau area south of Pocklington Trough, and to the south and southwest by the Mellish Rise and several low-relief ridges or saddles. The nearest island arc/trench system is the New Britain and Solomons zone which is over 500 km from the outer edge of the Coral Sea Basin. The trenches dip away from the continent and hence consume crust of the Indian plate. The Solomons and Woodlark Basins, plus several 'remnant arcs' (Karig, 1972) lie between the Coral Sea Basin and the

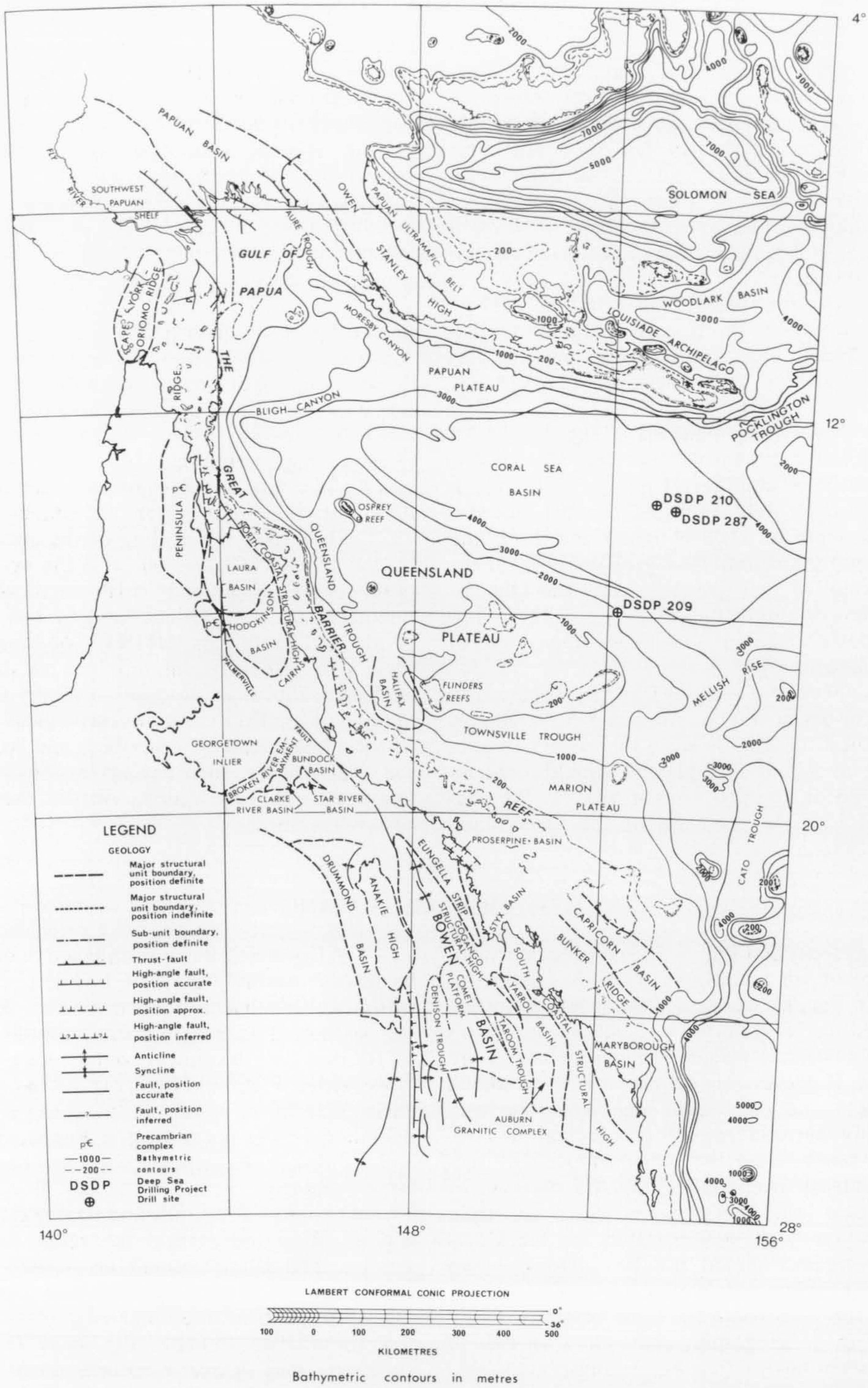


Fig. 2. Structure of the Coral Sea region.

consumption zone. The age relations of the three basins are not yet established.

The tectono/magmatic activity associated with basin evolution operating adjacent to a continental edge is probably closely linked with subsidence of the margin and formation of submarine plateaus such as the Queensland Plateau. Falvey (1972) has discussed this relation, and developed a generalized scheme for the evolution of a 'rifted, or Atlantic type' continental margin (Falvey, 1974). No similar rigorous analysis of continental margin evolution after marginal basin formation has yet been attempted.

The evolution of a marginal plateau or other submerged continental fragment, either attached to or separated from the mother continent, is governed by two major factors: evolution of the ocean basin adjacent to the plateau, and the structural framework of the continental block being modified by oceanic processes. For marginal plateaus of the Coral Sea Basin, events such as the interaction of the Pacific and Indian plates, which may have led to crustal genesis in the Coral Sea and the triggering of plateau subsidence, assume equal significance with the tectonic fabric of the Tasman Geosyncline as a controlling influence on the structural development of the plateau. Thus, to place the Queensland Plateau in its geological setting it is necessary to review the basic features of any peripheral tectonic units which may have had an influence on the development of the Plateau. The broadest sub-division which conforms with the small ocean basin concept yields the following three units:

Tasman Geosyncline

The Tasman Geosyncline is a north to northwest-striking Palaeozoic tectonic belt which occupies the eastern fifth of the Australian continent. It stretches from southern Tasmania to northern Queensland in numerous elongate structural troughs and highs (Packham, 1960; Brown, Campbell & Crook, 1968; Crook, 1969).

Twenty-seven units recognized by Crook fall into two major groupings: the Cambrian to Middle Devonian Lachlan group, and the Ordovician to Late Permian New England group. The younger New England group lies to the east of the Lachlan and occupies the eastern belt from central New South Wales to central Queensland. In Queensland both the Lachlan and New England groups are present and units within each intersect the coast.

The detailed geology and geological history of the geosyncline in Queensland is complex. Even the simplified picture presented in Figure 2 shows that many individual structural elements are present. These can, however, be divided into a simple scheme of volcanic and non-volcanic troughs and highs. This division was made by Crook and roughly corresponds to eugeosynclinal and miogeosynclinal belts defined by Voisey (1959) for the Australian region. Crook's belts are shown in Figure 3, and have been extended farther north and east than Crook's division. They form a system of sub-parallel troughs and basement highs creating the tectonic frame of the Queensland coastal strip.

Oversby (1971) described the Palaeozoic history of the Tasman Geosyncline in Victoria and New South Wales in terms of plate tectonic theory. His analysis was later expanded and modified by Solomon & Griffiths (1972) who postulated a long history of subduction and island-arc volcanism which has resulted in the eastward outgrowth of the continental margin at least since Ordovician time. They suggested that orogenic stages are related to the collision of crustal blocks, such as the Lord Howe Rise, with the developing margin. Packham (1973) also made the Tasman Geosyncline fit the plate tectonic model and described an easterly-migrating subduction zone active from the Ordovician to the Early Cretaceous.

These analyses demonstrate that modern theories of global tectonics can be applied to an ancient geosyncline. The fit to the theory must be regarded as tentative, however, because a complete Palaeozoic geological record is not generally present. To treat the analyses of Packham, Solomon & Griffiths, or Oversby as anything but a first approximation would be unwise, but treated as such they can be of value.

According to Packham (1973) the area which is now occupied by the Queensland Plateau was the site of a trench, fore-arc zone, and volcanic arc (from east to west) in the Early Devonian. As the trench moved eastward the arc zone became inactive and stabilized to form the tectonic framework of the present-day Plateau. This is equivalent to saying that the structural lineaments of the Tasman Geosyncline extend offshore onto the submerged continental blocks. Maxwell (1968), in describing the geological framework of eastern Queensland, extended his analysis, and the geosyncline, to the outer edge of the Queensland Plateau. His deductions were not based on a plate tectonic model, but stemmed from

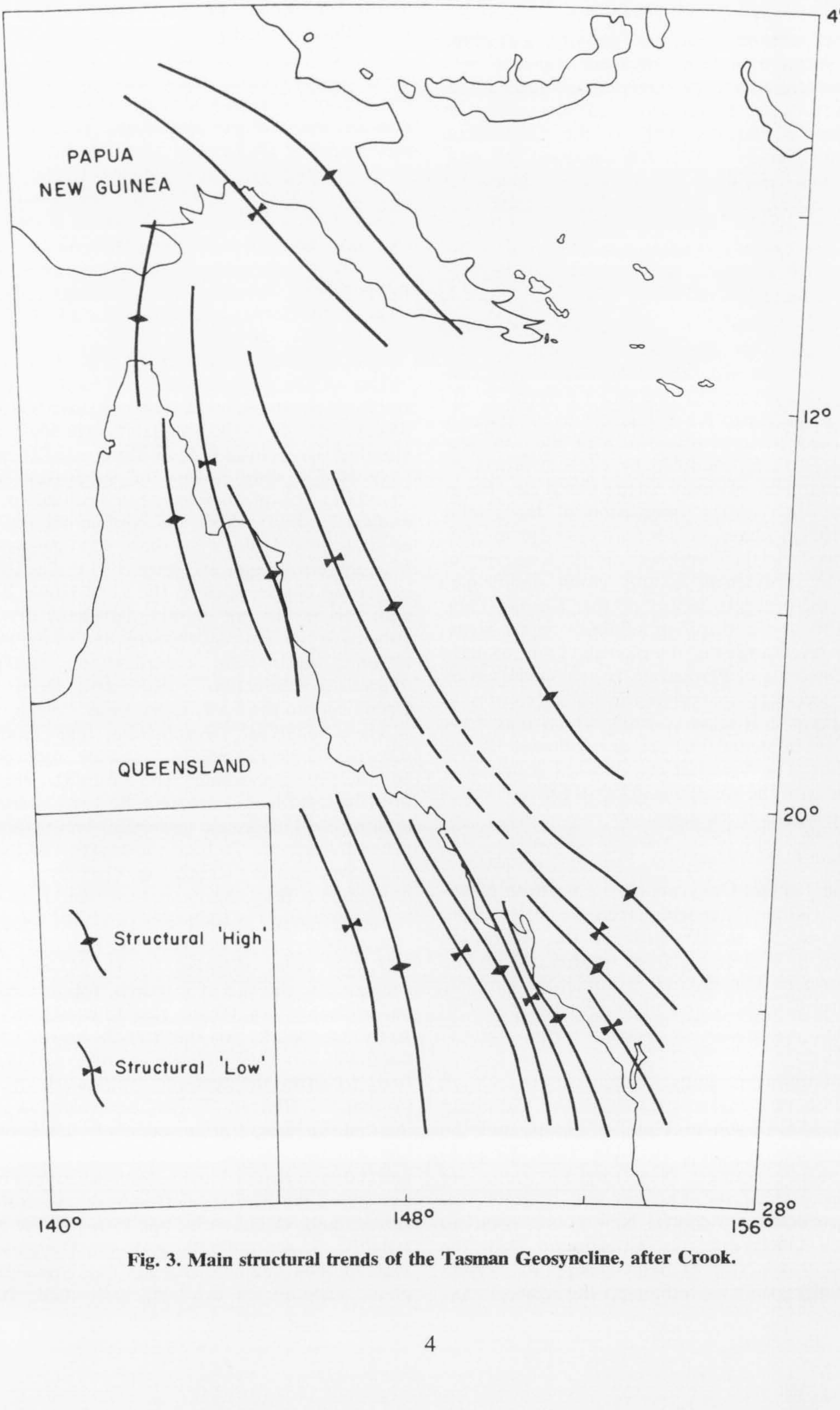


Fig. 3. Main structural trends of the Tasman Geosyncline, after Crook.

extrapolation of onshore geological trends. The proposed extension of the geosyncline beyond the coastline into ocean-covered areas represents an extension beyond strict geological observation into hypothesis. The hypothesized extension is plausible and raises no geological difficulties, so it is used here as a working assumption.

Coral Sea Basin

Small ocean basins in the Western Pacific are generally shallower than large ocean basins such as the Pacific or Indian (Packham & Falvey, 1971). In this aspect the Coral Sea Basin is unusual in that it is about 1 km deeper than would be expected for a basin of similar area. Its crustal structure, however, appears to be 'normal oceanic' (Ewing, Hawkins & Ludwig, 1970); the only unusual feature is a 7.3-km/s layer underlying the outer edge of the Queensland Plateau. This layer lies immediately above mantle and creates a fourth level in the crustal zonation. Hales (1975) noted that with the advent of the sonobuoy refraction technique, the fourth layer is being discovered in many oceans but most commonly near a continental edge. It may prove to be a characteristic of the continent-ocean transition zone.

The floor of the Coral Sea Basin lies at a median depth of 4500 m and occupies an area of about 7000 km^2. The abyssal plain slopes to the southwest at an angle of less than 1° (Gardner, 1970). Several major canyons and numerous minor submarine valleys drain the elevated areas surrounding the deep sea-floor, and Winterer (1970) deduced a complex history of structural development, downcutting, and infilling for some of the major canyons. Canyons lead into the Coral Sea Basin from the northwest (Moresby Canyon) and funnel sediments from Papua into the Basin. Canyons also rise in the northeast (Pocklington Trough), the west (Blight Canyon) and southwest, but few are noted in the south and southeast. This distribution may partly account for the southwest tilt on the basin floor as the greatest amount of turbidite input to the basin would be from the most heavily canyoned areas in the north and northeast.

Gardner (1970) presented the results of 16 core and grab samples from the floor of the Coral Sea Basin. He described a thin (40 cm) layer of pelagic material overlying a sequence of terrigenous turbidites with a relatively high silt content. He concluded that the pelagic sediments are Holocene, accumulated at a rate of 3.6 cm/1000 years, and he related the change from turbidite to pelagic deposition to changes in sea level after the last glacial epoch. The longest core Gardner described was only 11 m and probably did not penetrate past the Pleistocene. In contrast DSDP holes 210 and 287 in the Coral Sea Basin penetrated 711 and 252 m of section, and reached Eocene strata.

Drilling at these two sites has produced the most comprehensive cross-section of the Coral Sea Basin geology available to date and has made it possible to deduce the Basin history. At site 210 (Burns, Andrews et al., 1973), drilled in 4643 m of water on 31 December 1971, five lithologic units were encountered (Fig. 4). The scientific party aboard DSDP Leg 21 (Burns, Andrews et al., 1973) interpreted the following history for the basin:

Early to late Eocene:

Accumulation of detrital clay and biogenic pelagic sediment. The provenance of the detrital material was probably a sedimentary or low-grade metamorphic terrain, most likely the Australian continent.

Late Eocene to early Oligocene:

Formation of an angular unconformity by non-deposition in the interval, and folding of the lower beds.

Early to middle Oligocene:

Deposition of nannofossil ooze at a depth near to the carbonate compensation depth.

Early to middle Miocene:

The basin deepened and/or the carbonate compensation depth rose, resulting in a change in sediment lithology to abyssal clay. Provenance of terrigenous detritus changed from that of the Eocene section to a new area, probably New Guinea.

Late Miocene to late Pliestocene:

Massive influx of graded silt and clay with interbedded nannofossil ooze, all as turbidite deposits. Terrigenous turbidites are mineralogically similar to New Guinea rocks and there is little doubt that New Guinea was the source area. Biogenic deposits were probably derived from slumping and canyoning on the edge of the Coral Sea Plateau. The upper two-thirds of Unit 1 was accumulated at twice the rate of the lower one-third and this may relate to Pliocene folding and uplift in the Aure Trough.

Drilling at site 210 did not penetrate to basement so a further hole (287) was drilled in the Coral Sea Basin on Leg 30 in May 1973 (Andrews, et al., 1973). It was drilled on a basement high and cored an essentially similar, but much compressed section (Fig. 5). A

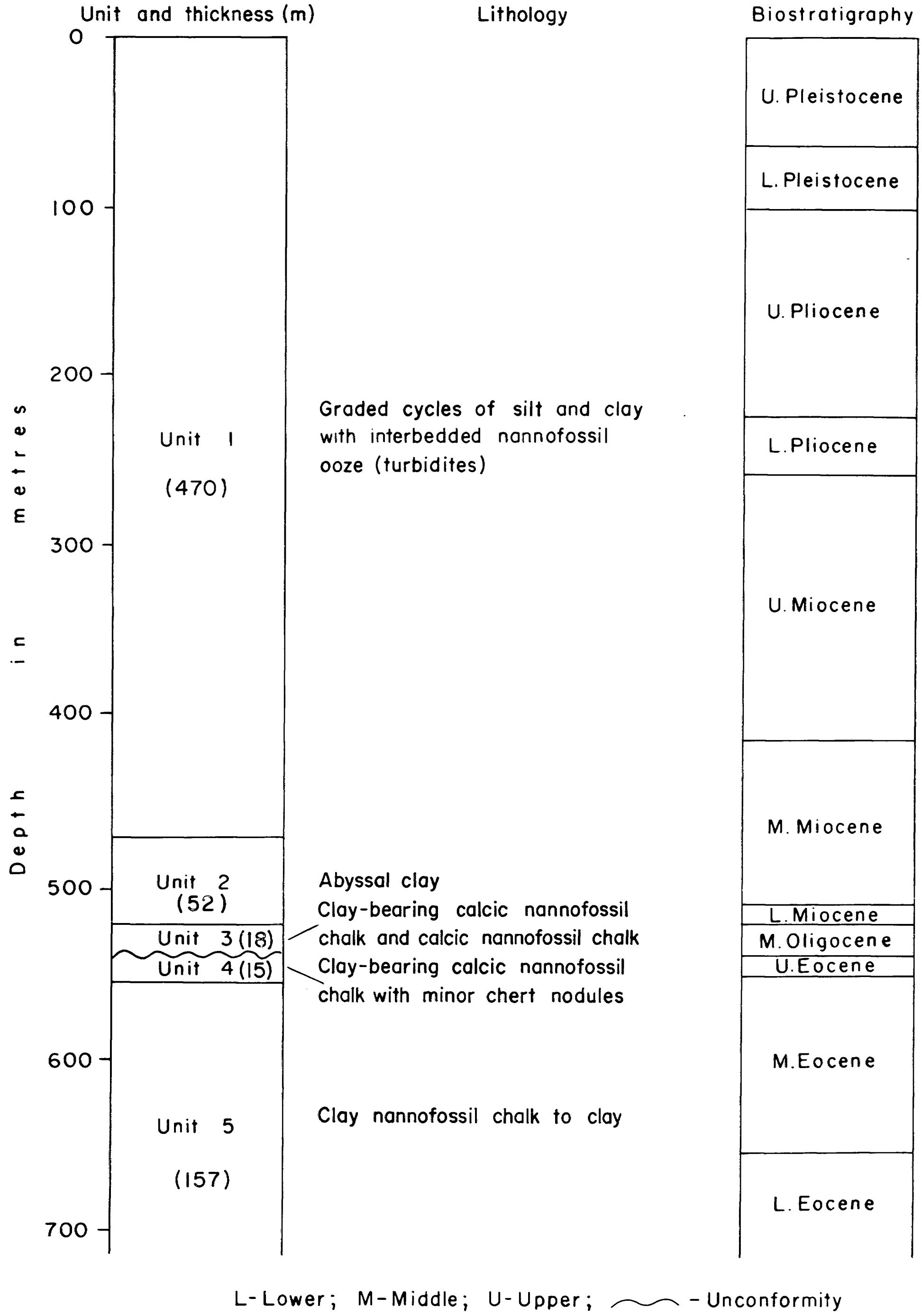


Fig. 4. Summary of results from DSDP 210.

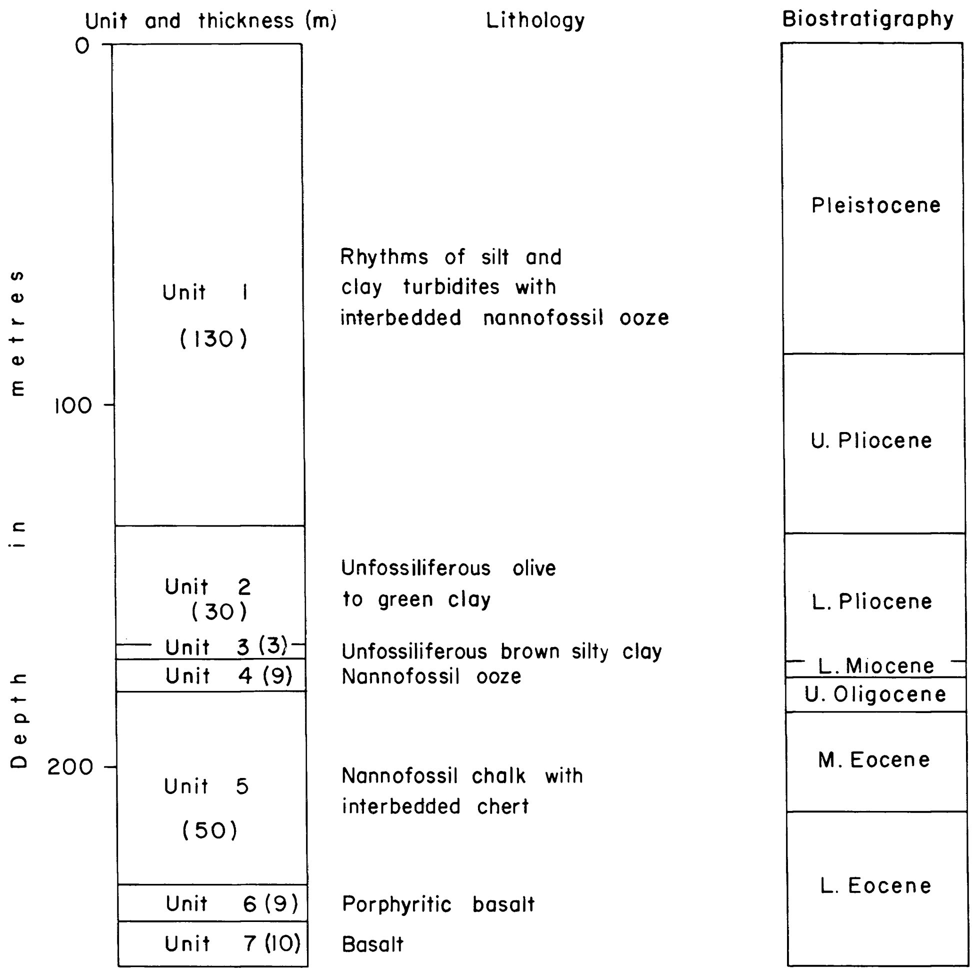


Fig. 5. Summary of results from DSDP 287.

basaltic basement was reached and fixed the age of oceanic crustal genesis at early Eocene. The Eocene/Oligocene unconformity was encountered, as it was in holes 206 to 210 of Leg 21. The unconformity is of regional importance and has been related to readjustments in oceanic circulation patterns after the break-up of Australia and Antarctica in middle Eocene time (Kennett et al., 1972).

Seismic reflection and refraction data from the Coral Sea Basin were used by Ewing, Houtz & Ludwig (1970) to describe the sediment distribution in the area. They found that the sediments were 'quite thick' and distinguishable into three layers with different acoustic properties. The two upper layers had velocities of 2.0 and 2.5 km/s and were 'weakly to moderately stratified'. Then followed a 'transparent' layer lying above a basement with velocities in the range 4.3 to 5.6 km/s. Ewing et al. (op. cit.) suggested that the basal sediments might be contemporaneous with 'much of the older (Queensland) plateau sediments' and that the upper two sequences were turbidites, one derived from erosion on the Queensland Plateau and the other from New Guinea. These suggestions were made before DSDP drilling and are not altogether consistent with the later results; the upper two layers are probably the middle Miocene to Pleistocene turbidite, and the lower transparent layer is probably the Eocene to Miocene pelagic sequence.

Geological interpretation of seismic reflection data in the Coral Sea Basin is made tenuous by the lack of good correlation between seismic reflection events and lithologic or physical property breaks; only one of five prominent reflections obtained on sections from the approach run of the *Glomar Challenger* coincides with a physical or stratigraphic break (Andrews, 1973). Three reflections originate within the turbidite sequence (Unit 1) and one marks the top of Unit 3 which is generally the acoustic basement. At slow traversing speeds a deeper reflector is found. The Eocene/Oligocene unconformity does not give rise to a reflection, and stratified and transparent zonation in seismic records does not correspond to a lithological zonation. Hence, even the very simplistic interpretation of the data of J. Ewing et al. may be in error.

Gardner (1970) and Davies & Smith (1970) have proposed dates and mechanisms for the opening of the Coral Sea but both conflict with the more recent DSDP dating. Karig's (1972) analysis of remnant arcs and marginal basin evolution in the New Guinea region avoids specific mention of the Coral Sea Basin. Generalized theories of marginal basin evolution such as those of Karig (1971b), Packham & Falvey (1971), or Carey (1958) could be made to fit the available facts. A correlation between the Coral Sea Basin and the Queensland Plateau evolution is presented later in this paper as an outcome of investigations into the structural and sedimentological development of the Queensland Plateau.

Melanesia

The islands of New Guinea, New Britain, Bougainville, the Solomons, and New Hebrides, together with their associated trench systems, mark the subducting junction of the Pacific and Indian lithospheric plates in the immediate vicinity of the Coral Sea region. In this region only the Coral Sea Basin was tested by DSDP drilling so it is the only submarine feature with a firm age dating. However, most of the surrounding island chain has formed within the period early Eocene to Present (Packham, 1973) and the small ocean basins lying south and west of the islands would be of approximately the same age if Karig's (1971b) evolutionary model is valid for this area.

The arc/trench system displays reversed polarity; i.e. the subduction zone dips away from the continent and marginal basins. A study of the tectonic framework of the New Hebrides suggested to Karig & Mammerickx (1972) that the arc had reversed its polarity between middle and late Miocene. Since the reversal, the trench has been consuming crust interior to the arc system. The crust being consumed, that of the several small ocean basins located between the trench system and the continent, must therefore be older than the date of the reversal. If marginal basin formation occurred as an outgrowth of extensional basins beginning at a palaeo-Australian continental margin, then the Coral Sea Basin with an early Eocene datum would be the oldest in the region. Hence the complex Melanesian basin and ridge province would have formed in an interval of about 45 m.y. between limits defined by Coral Sea opening and arc polarity reversal. This time span is more than adequate in view of the extension rate of 10 cm/year deduced by Karig (1971a) for active inter-arc basins in the Mariana region.

Present-day seismicity in the southwest Pacific sector of the Indian plate (Denham, 1973) indicates that almost all the tectonic activity in that area is occurring along the boundary of the plate. This modern tectonic pattern can have little effect on the Australian continental margin. However, if basin accretion began adjacent to the continental block, then at the inception of the formative process a subducting plate boundary must have been located near the edge of that block. The tectonic activity which would have been associated with marginal basin generation and the high heat flow resulting therefrom would have created conditions by which the old continental margin was modified into its present configuration of troughs and marginal plateaus.

GEOPHYSICAL AND GEOLOGICAL DATA ON THE PLATEAU

Numerous oceanographic vessels have crossed the Queensland Plateau as part of larger survey programs. These ships carried either seismic, gravity, or magnetic recording instruments and many were equipped for multisensor surveying. In addition, aeromagnetic surveys and underwater gravity readings have been made in the general region. The tracks of the major geophysical surveys are shown in Figure 6, and relevant information which can be deduced from them will be referred to.

The pre-1971 geophysical survey which produced the greatest amount of published information was a co-operative exercise between the University of New South Wales and the Lamont-Doherty Geological Observatory in

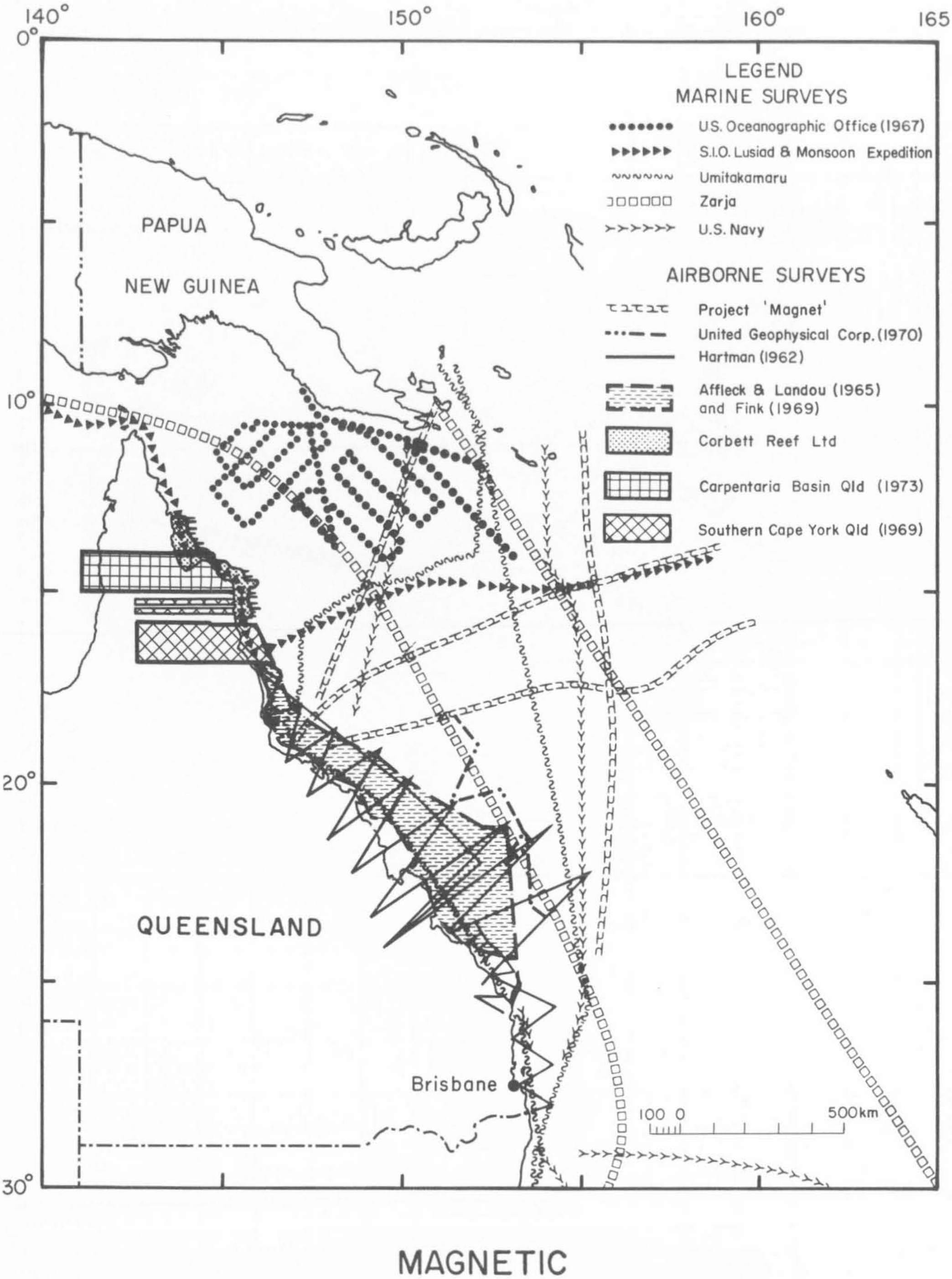
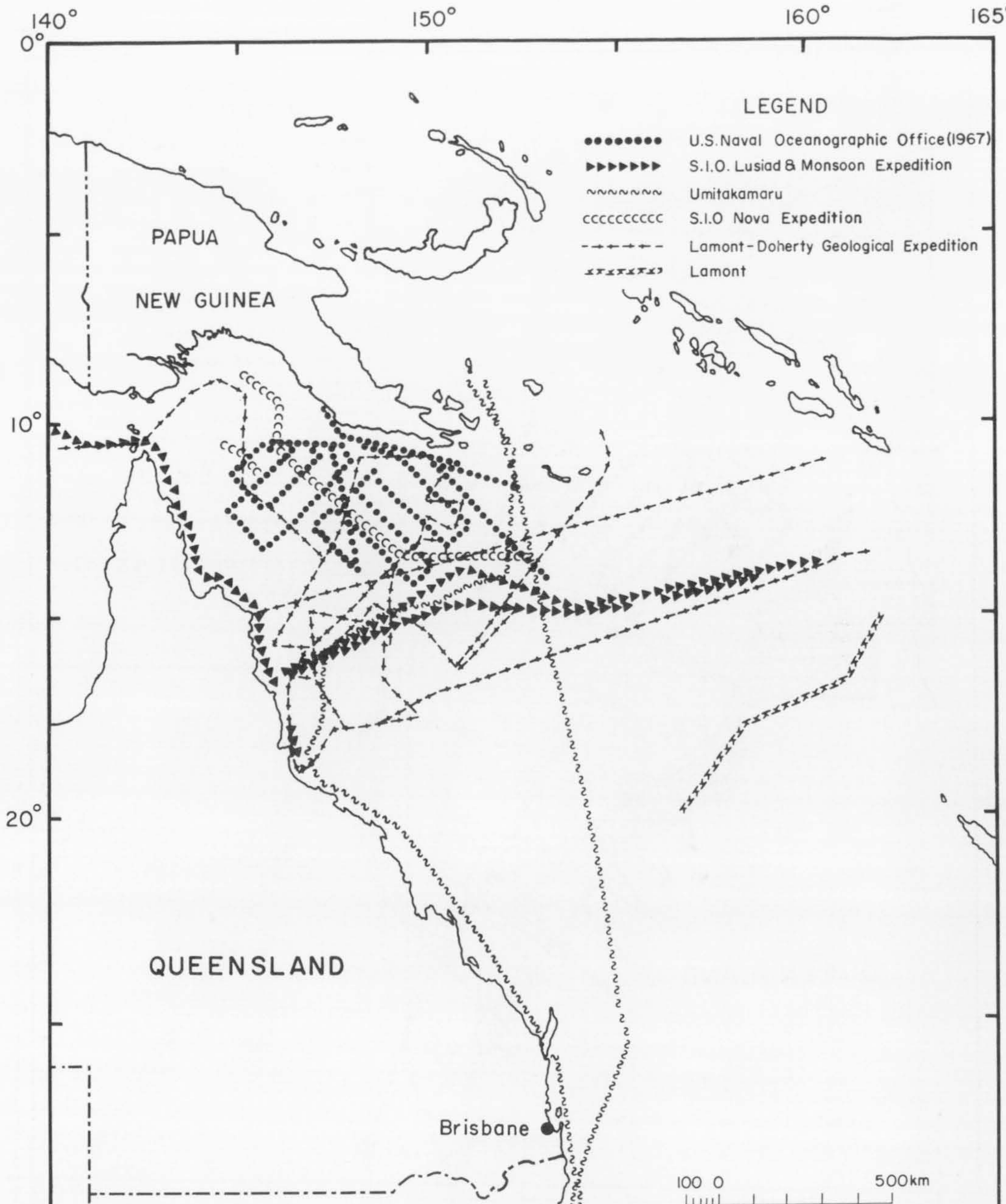
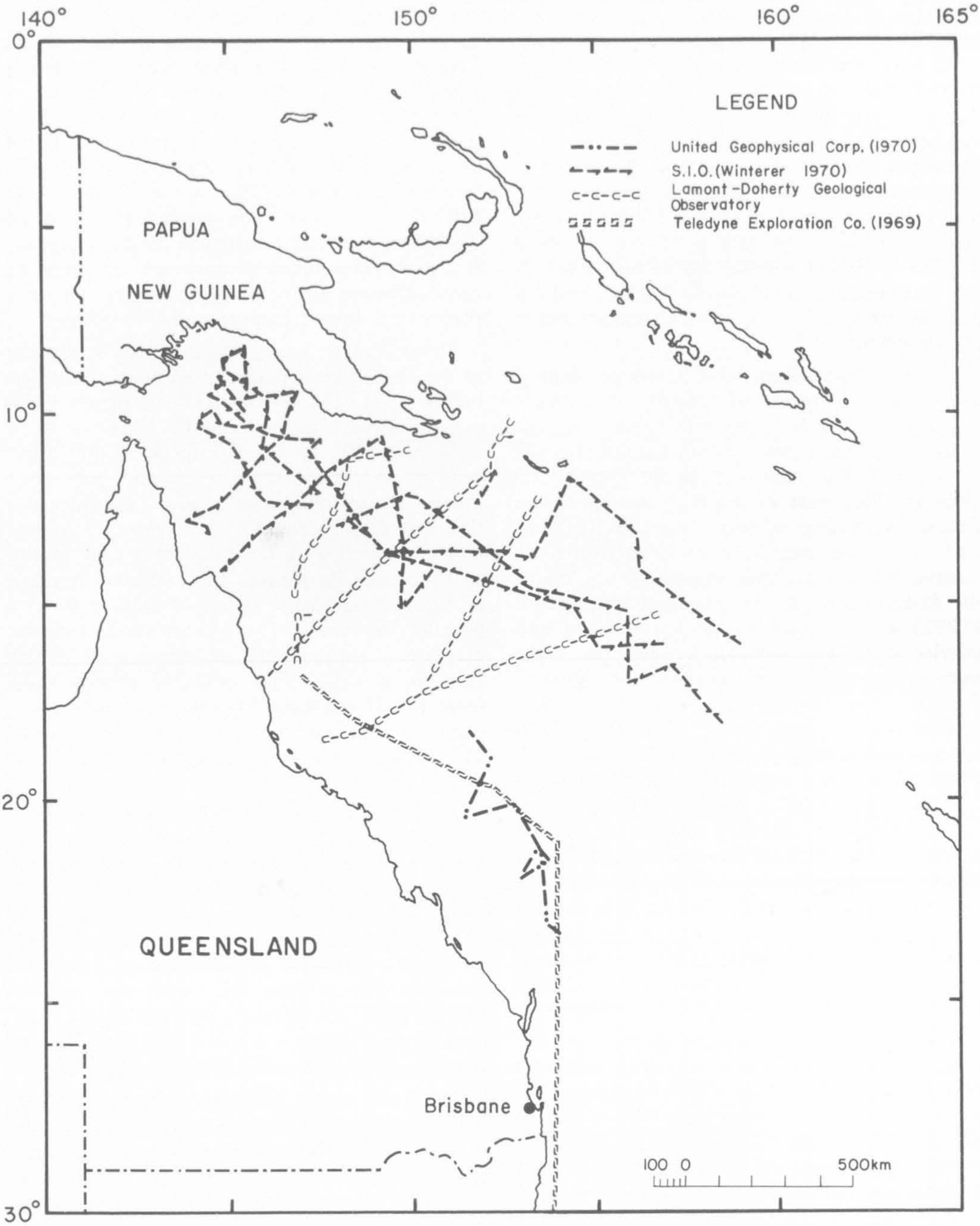


Fig. 6. A. Previous magnetic surveys.



GRAVITY

Fig. 6. B. Previous gravity surveys.



SEISMIC

Fig. 6. C. Previous seismic surveys.

1967 (Ewing, Houtz, & Ludwig, 1970; Ewing, Hawkins & Ludwig, 1970). Gardner (1970) and Falvey (1972) used these results in broad studies of the Coral Sea area. Falvey interpreted the structure of the Queensland Plateau and Basin across to the Louisaide Archipelago, and regarded the plateau as being underlain by crust of continental thickness (25 km). Two sedimentary layers, with velocities of about 2.3 km/s and about 5.4 km/s, overlie a continental basement with a velocity of about 6.3 km/s. He interpreted the eastern flank of the Queensland Trough to be faulted, and the plateau reefs to lie atop elevated structures in the basement.

Falvey indicated crustal thinning of about 5 km below the Queensland Trough. This amount of thinning would seem unnecessary as the crust should have sufficient mechanical strength to support the relatively small Queensland Trough. This part of Falvey's interpretation does not have deep seismic control, so the interpreted thinning must be considered with some reservation. Crustal thinning of 20 km below the Rockall Trough was proposed by Scrutton (1972) on the basis of deep refraction and gravity data. The Rockall Trough bears the same geographical relation to the Rockall Plateau as the Queensland Trough does to the Queensland Plateau, i.e. it separates a submarine plateau from a continent. It is, however, about three times wider than the Queensland Trough, some 1000 m deeper, and contains 5 km of sediment. Thinning toward an oceanic section in the Rockall Trough is therefore a reasonable proposition as it is a requirement for isostatic equilibrium. In the Queensland Trough, sediment thickness is not well known so a gravity interpretation is less constrained. Little or no crustal thinning is required if a sediment thickness of about 1 km is assumed.

The crustal thickness below the Queensland Plateau deduced by Falvey (1972) naturally only applies to the profile which he interpreted. It can be extended to the whole Plateau by use of the Bouguer anomalies shown in Figure 7. Simple infinite slab calculations give a crustal thickness of about 25 km, which agrees well with Falvey's computation, and the smoothness and relatively constant level of the Bouguer field over the Plateau suggest that this thickness applies to the whole Plateau.

Gardner (1970) reviewed the sedimentary data from the Queensland Plateau from core and grab samples and the N.S.W.-Lamont seismic reflection and refraction data. The data showed sediment thickness on the Plateau of 0.5 to 1.0 km. Cores generally recovered calcareous ooze which had accumulated at an average rate of 3.6 cm/1000 years in the Holocene, equal to the rate for the same period in the Coral Sea Basin. Hence a uniform recent pelagic veneer probably covers most of the area. Within the sedimentary sequence on the Plateau, Gardner (1970) and J. Ewing et al. (1970) noted a marked unconformity which was interpreted as erosional in origin; Gardner, from several sources of evidence, ascribed an early Miocene age to it, suggesting further that plateau subsidence commenced at that time.

No systematic gravity or magnetic surveying of the Queensland Plateau had been attempted before 1971. Dooley (1965) summarized the gravity surveys carried out by BMR between 1954 and 1960 on the Barrier Reef and adjacent coast. The 1954 survey also measured gravity on offshore coral reefs. This information was supplemented by underwater readings in 1958 which were later tied to a land survey in 1960. Results showed the expected increase in Bouguer anomaly values offshore generally parallel the coast; Dooley assumed that the increase was due to crustal thinning across the continental edge. The available gravity data from the Queensland Plateau have been combined into a free-air anomaly map by Falvey (1972) and used in his analysis of the structure of the Plateau.

Marine magnetic surveying of the Plateau has been unsystematic and generally incidental to major seismic reconnaissance. An aeromagnetic survey by Shell (1968) has, however, defined a deep basin, the Halifax Basin (Fig. 2), at the junction of the Queensland and Townsville Troughs and containing 5 km of sediment. Results from this survey also showed that a major fault exists in the continental shelf west of the Halifax Basin and that the Flinders Reefs lie on a high in the crystalline basement.

The BMR survey of offshore Queensland was carried out as part of a larger program of geophysical reconnaissance of the continental margin of Australia over a 2.5-year period from August 1970 to January 1973 when over 100 000 n miles of multisensor geophysical data was collected. Surveying in offshore Queensland occupied the period July to November 1971, when 20 000 n miles of combined seismic, gravity, and magnetic data was obtained on a systematic grid of east-west survey lines spaced 20 n miles apart (Fig. 8). The seismic source was a 120-kJ sparker, and a six-

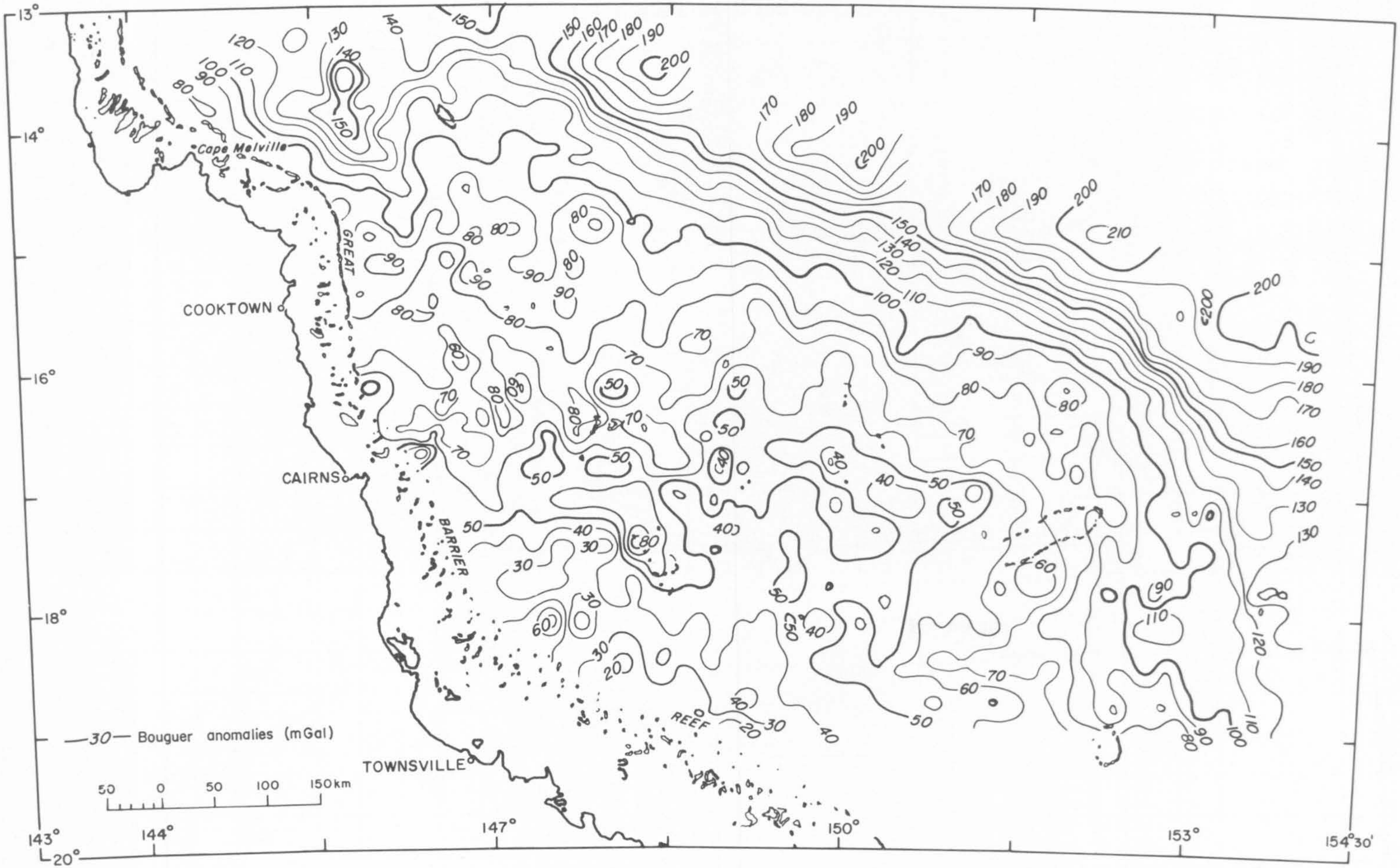


Fig. 7. Bouger anomalies.

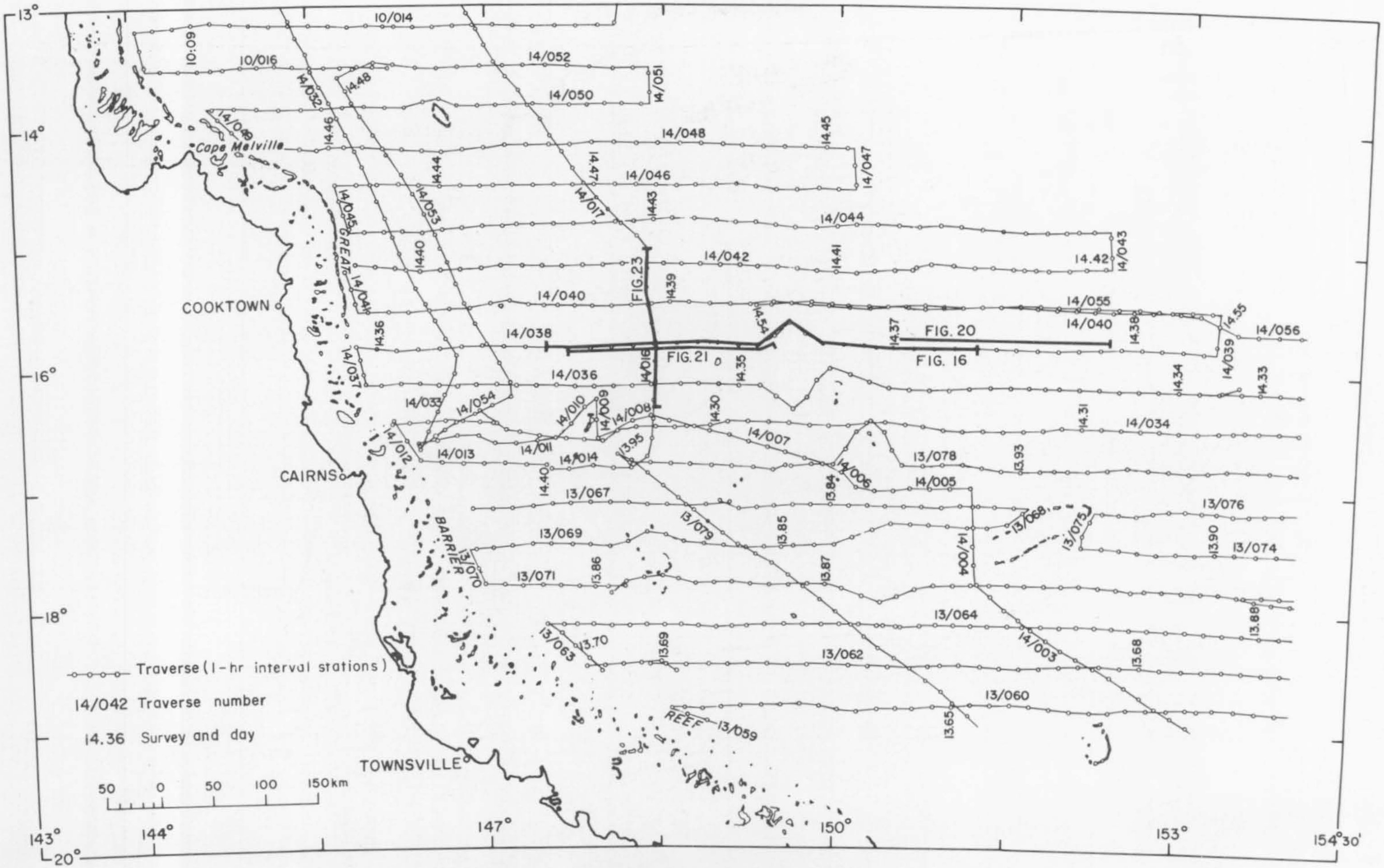


Fig. 8. BMR survey traverses.

channel cable received the return. A sonobuoy refraction probe using the same seismic source was made approximately once a day, given good bottom conditions. A LaCoste & Romberg stabilized platform gravity meter recorded the gravity field variation, and a Varian proton precession magnetometer measured the total field. The ports of Cairns, Port Moresby, Townsville, Gladstone, and Brisbane were entered during surveying, and at each a tie was made to the BMR network of gravity base stations (Barlow, 1970). The ship (M/V *Hamme*, later renamed M/V *Lady Christine*) was navigated using fixes from a U.S. Navy Transit satellite and a pulsed sonar Doppler dead-reckoning system. An analysis of the data quality giving statistics of errors at the intersection of tie lines (Mutter, 1974) shows that the data are generally of good quality.

The Bouguer anomaly, free-air anomaly, magnetic anomaly, and bathymetric contour maps presented in this paper (Figures 7, 11, 17, 19) were produced from the BMR data. A few minor additions have been made to the bathymetric data in reef areas where BMR data did not adequately define the physiography. The data points contoured were sampled at every hour point on the survey lines. The maps were contoured by a computer-contouring method which used linear interpolation on a triangular grid of points. The contours were later smoothed by hand. The use of hourly values creates a rigorous data base and computer contouring by strict linear interpretation provides an accurate, objective set of completely consistent, comparable maps. Naturally, shorter-wavelength features will not show on the hourly value contour maps or will be aliased to a wavelength of about 20 n miles. To overcome this shortcoming in the presentation, profiles obtained from one-minute sampling have also been drawn (Fig. 12).

The BMR survey represents the most comprehensive set of marine geophysical data collected on the continental margin off Queensland in recent years and complements the DSDP results. As with the Coral Sea Basin, the most revealing piece of information about the geology and history of the Queensland Plateau comes from DSDP hole 209, which was drilled in 1428 m of water on the outer edge of the Plateau in late December 1971. It cored three lithologic units dating back to middle Eocene (Fig. 9). Drilling did not reach basement and the oldest sediment cored was middle Eocene. The sequence of geological events deduced from the core data by the onboard scientific party (Burns, Andrews, et al., 1973) is as follows:

Late middle Eocene:

Deposition in shallow water (neritic), probably on the continental margin, of bioclastic sediment rich in foraminifera with subordinate silt and sand. Terrigenous detritus makes up 50 percent of the material in Unit 3, indicating near-shore conditions.

Late middle Eocene to late Eocene:

Gradual deepening of the site, reduction of terrigenous input, and development of pelagic ooze composed mainly of sand-sized planktonic foraminifera. Silt-sized foraminifera were possibly winnowed out by currents which transported larger echinoderms, molluscs, and Bryzoa to the site. Secondary silicification and replacement of ooze by chert.

Late Eocene to late Oligocene:

Non-deposition or very slight submarine erosion. The time width of the hiatus is 16 m.y.

Late Oligocene to late middle Miocene:

Further deepening of the site with deposition of almost pure foraminiferal ooze. Ocean currents transporting benthonic foraminifera to the site and winnowing out most nannoplankton.

Late middle Miocene to middle Pliocene:

Non-deposition or very slight submarine erosion. The time width of the hiatus is nearly 8 m.y.

Middle Pliocene to Recent:

Continued deepening of the site to mid-bathyal depths with reduction in the ocean currents, causing higher proportions of nannoplankton and planktonic foraminifera to be deposited.

Marked reduction in terrigenous material occurs at the site, from greater than 50 percent in Unit 3 to less than 5 percent in Unit 1 (Fig. 10). The initial rapid drop in terrigenous input during the late Eocene was suggested by the onboard scientific party to reflect the formation of the Queensland Trough as a barrier to sediment transport to the site from a western landmass.

Seven seismic reflecting horizons were observed on reflection records made on the *Glomar Challenger*. Three horizons correspond to velocity interfaces within lithologic units. The lowest corresponds roughly to the late Eocene/late middle Eocene boundary at 270 m and marks 'the lower level of chert stringers in Unit 2' (Andrews, 1973). The 'erosional unconformity' of J. Ewing et al. (1970) possibly

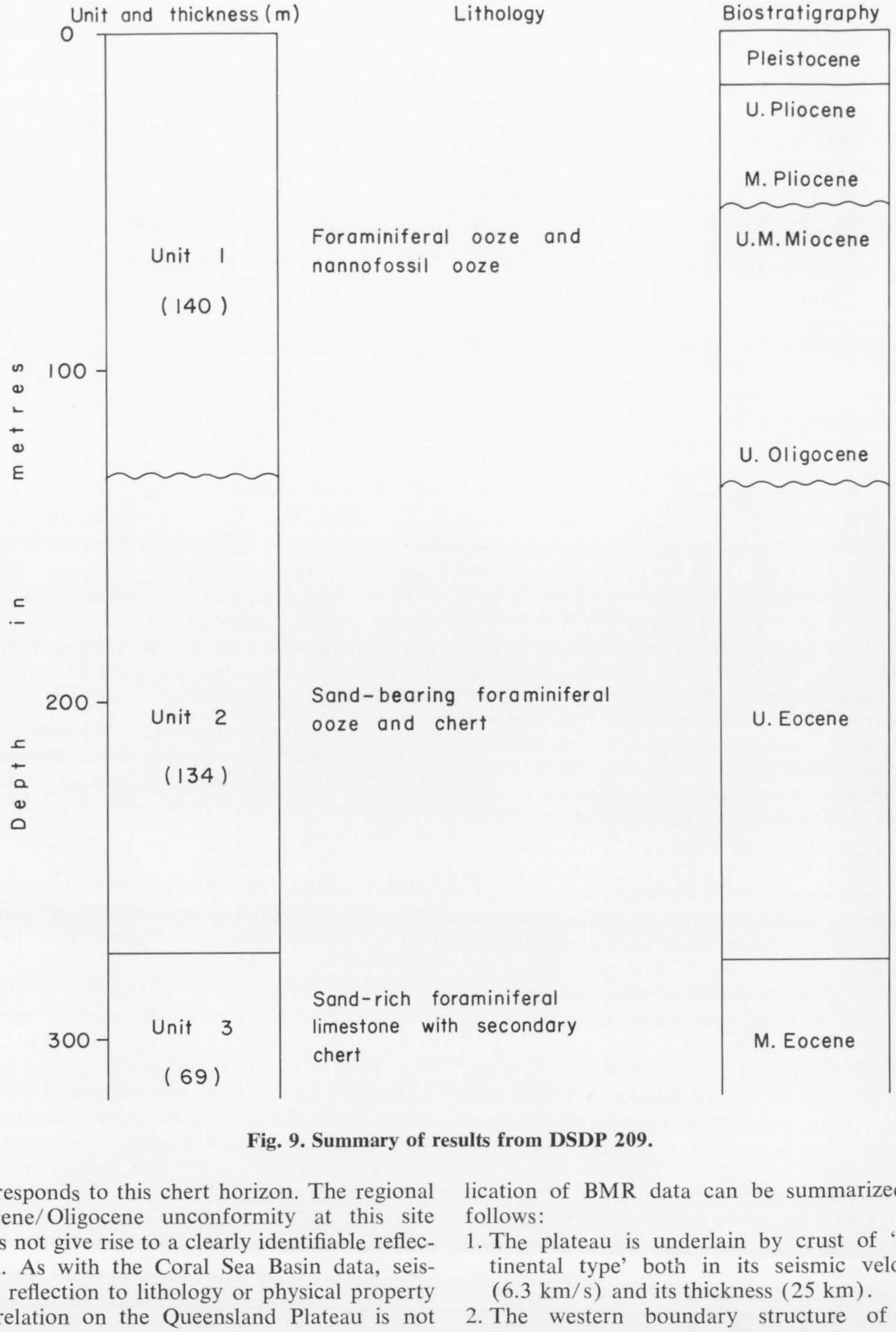


Fig. 9. Summary of results from DSDP 209.

corresponds to this chert horizon. The regional Eocene/Oligocene unconformity at this site does not give rise to a clearly identifiable reflection. As with the Coral Sea Basin data, seismic reflection to lithology or physical property correlation on the Queensland Plateau is not particularly good, and seismic interpretation must be treated with caution.

The major outcomes of analyses of the Queensland Plateau region made before publication of BMR data can be summarized as follows:

1. The plateau is underlain by crust of 'continental type' both in its seismic velocity (6.3 km/s) and its thickness (25 km).
2. The western boundary structure of the Plateau (the Queensland Trough) is a graben formed by faulting in the basement.
3. Some reefs on the plateau surface lie atop elevated structures in the crystalline base-

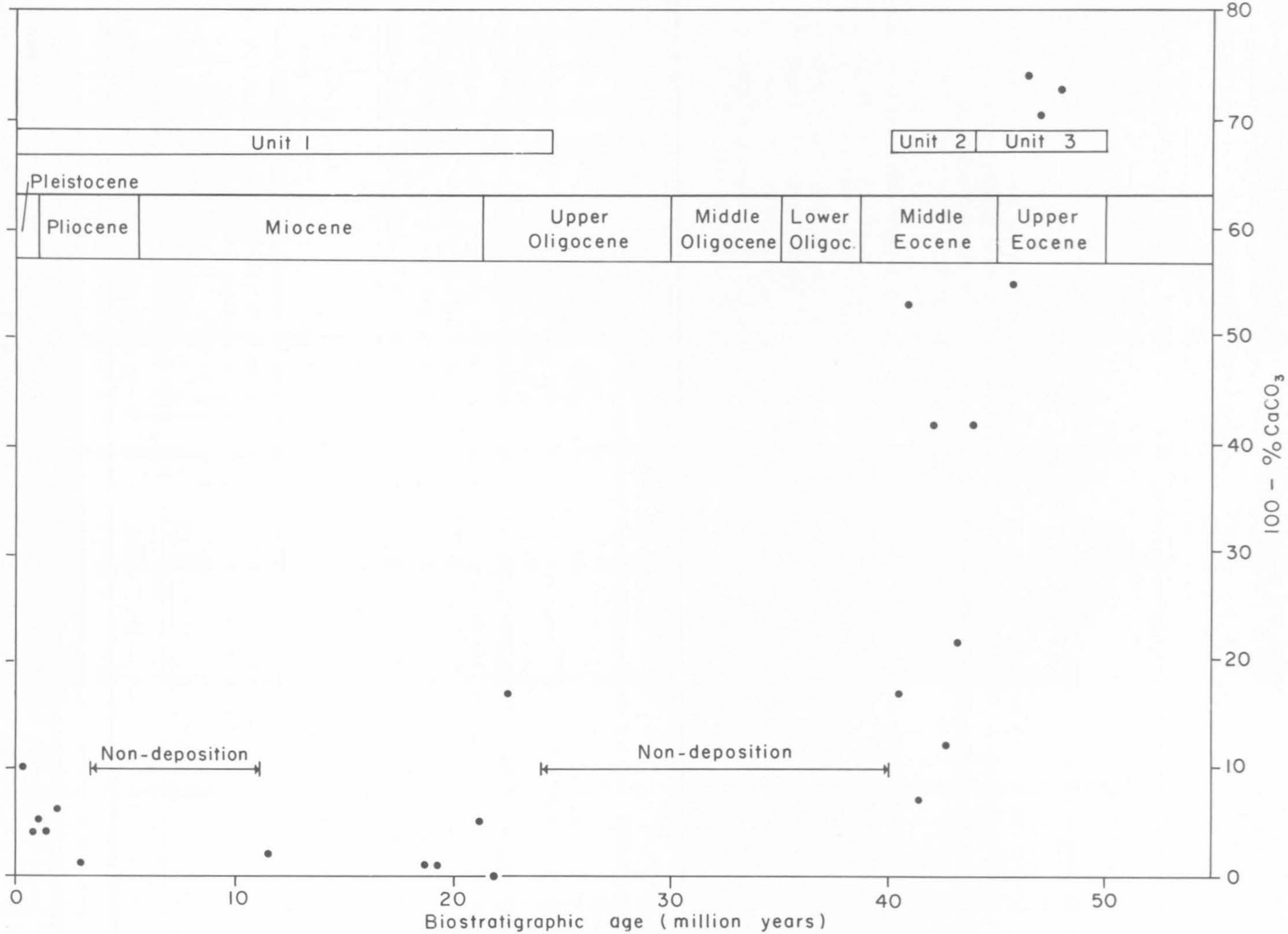


Fig. 10. Terrigenous content of cores from DSDP 209.

ment (Falvey, 1972).

4. Sediment cover is relatively thin, varying from 0.5 to 1.0 km on average and most of this material is probably middle Eocene to Pleistocene foraminiferal and nannofossil ooze. Deposits show the influence of bottom currents.
5. The outer edge of the Plateau has been continuously subsiding from neritic to mid-bathyal water depths since the late Eocene.
6. A regional Eocene/Oligocene unconformity is present on the Plateau but probably does not give rise to a reflection event. It does not correspond to the widespread 'erosional unconformity' described by J. Ewing et al. (1970).
7. The Queensland and Townsville Troughs may have formed at about early Oligocene time.

BATHYMETRIC DESCRIPTION

The Queensland Plateau is the largest of eleven major plateaus in the Australian continental margin. It has an area, within the 1000-m isobath of about 181 000 km^2, and 373 000 km^2 overall. The former figure is almost twice that for the next largest plateau, the Exmouth Plateau off the west coast of Australia. The Queensland Plateau is the only marginal plateau with well developed trough boundaries; most others have shallow troughs or assume the profile of a large terrace or wedge embedded in the continental slope. It lies on the edge of a marginal basin, but all others face into Atlantic-type oceans such as the Wharton Basin and Indian Ocean.

The Plateau is roughly triangular with its western margin striking north-northwest, its northeastern margin which faces the Coral Sea Basin striking northwest, and its southern margin striking east-west. The western and southern margins are both formed by linear troughs. Many valleys and canyons lead from the Plateau surface into the troughs and the Coral Sea Basin. Only those leading into the Coral Sea Basin were intersected during the BMR survey (Fig. 12).

The Plateau surface lies at a median depth of 1100 m (Fig. 11) and away from reef areas is generally very smooth and flat. There are 15 areas of major reef development, the total area of which occupies almost one-quarter that of the Plateau. The Plateau surface as a whole has a very gentle northwest tilt, its surface being most deeply submerged around Osprey Reef. This tilt suggested to Faibridge (1950) that the Plateau formed by tectonic rather than sedimentary processes. Fairbridge further observed that the Plateau reefs grow from as much as 1500 m below sea level, well beyond the normal ecological limit of reef growth. This led him to suggest that the Plateau had subsided to its present depth from an initial elevation close to sea level, reef growth keeping pace with subsidence.

At least two lineaments are present in the reefs (Fig. 11). The most striking is the line of reef development on the western edge of the Plateau, bordering the Queensland Trough. Reefs from Flinders at 17°50′S to Osprey at 13°55′S lies along a straight line trend striking north-northwest. Another lineament lies along a line of longitude at about 149°15′S between 16°S and 18°S and includes Moore Reef, Herald Cays, and Malay Reefs. The trend of both lineaments is very close to that of the Tasman Geosyncline and may reflect fundamental structures of the Geosyncline within the continental margin.

Other reef lineaments are not as striking, e.g. the trend from Diane Bank to Tregrosse Reefs, or from Malay through Tregrosse to Lihou Reefs. The structural significance of these trends has not been established.

The northeastern margin of the Queensland Plateau is linear for more than 600 kilometres between 13°30′S and 16°00′S with a slight change in trend at about 14°20′S. This suggests that tectonic influences have shaped the Plateau. This margin forms the lower continental slope leading down to the Coral Sea Basin. Slopes here are relatively steep, ranging from 1:25 to 1:35; a normal continental slope is about 1:40 (Shepard, 1948). The profile of this margin (Fig. 12) shows a convex shape whereas a continental margin formed by the process of seaward accretion of sediment normally has a concave profile. This may also point to a tectonic influence on plateau formation. Profiles of the northern part of the Plateau show that extensive canyoning has modified the shape of the outer margin of the Plateau. Considerable mass transport of sediment from the Plateau into the Coral Sea Basin could occur via this canyon system.

Immediately east of Lihou Reefs the plateau surface deepens in two steps; one adjacent to the reefs and one along longitude 153°15′E. Two small terraces have formed as a result;

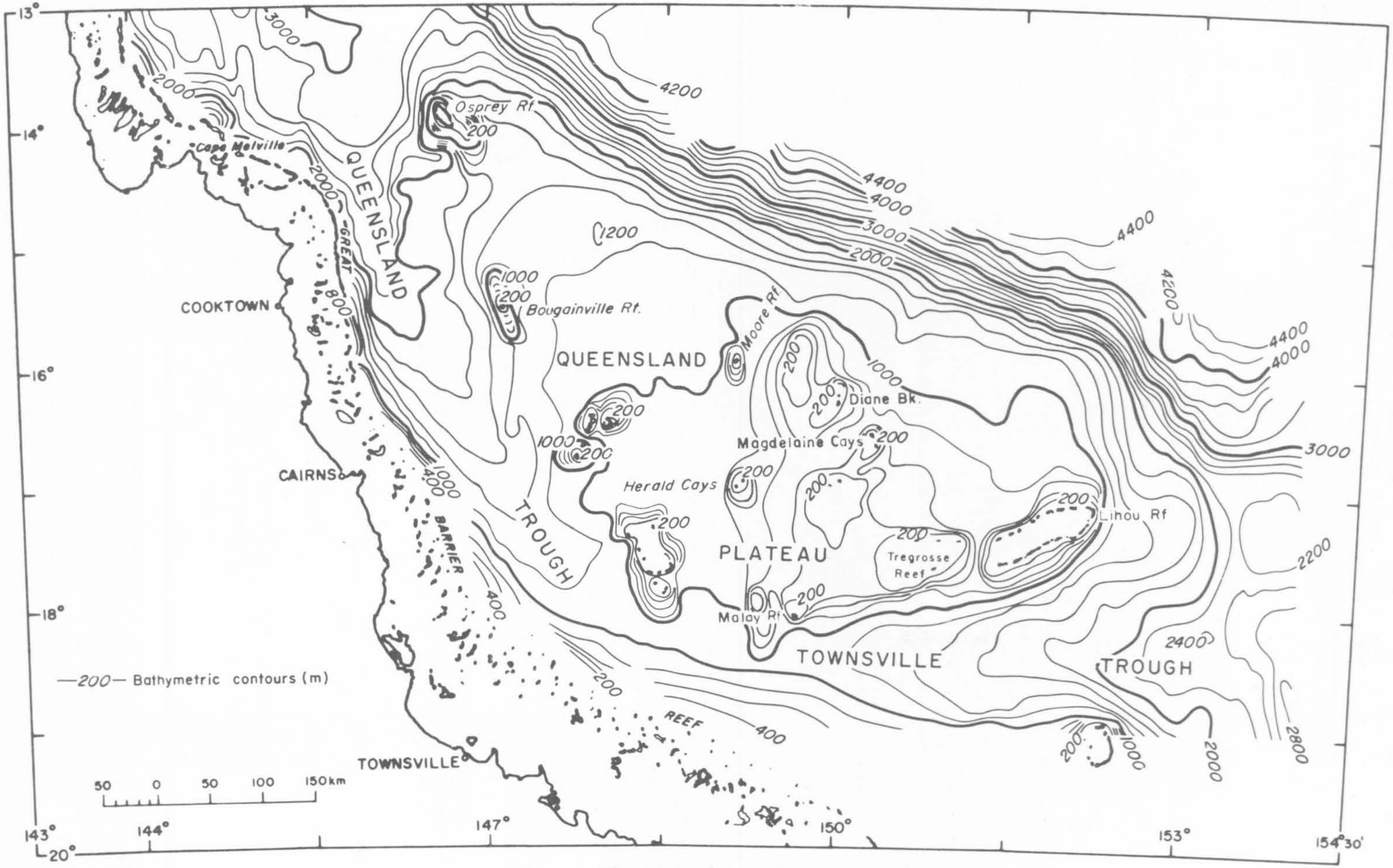


Fig. 11. Bathymetry.

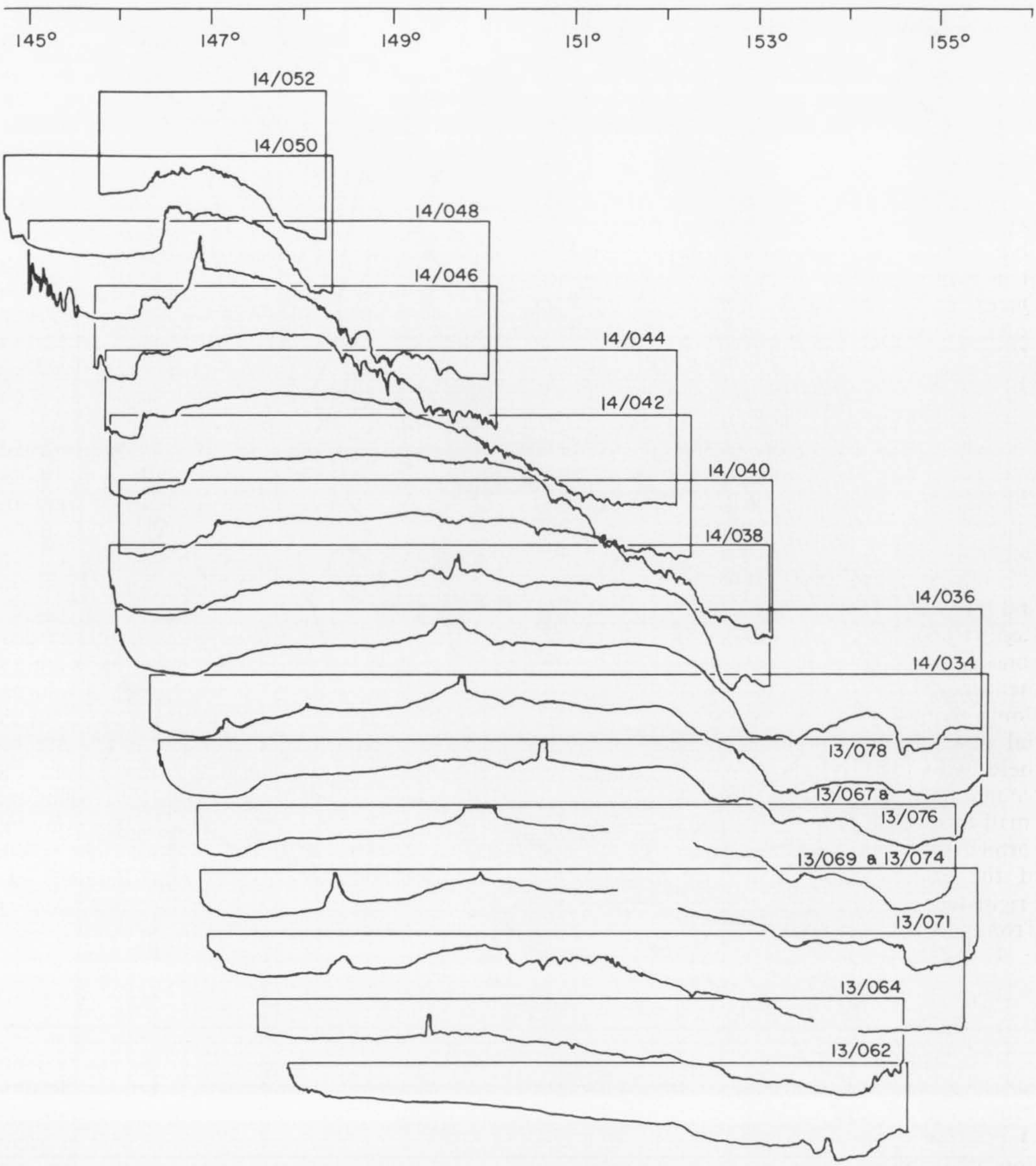


Fig. 12. Bathymetric profiles.

one at a depth of 1400 to 1600 m, and the deeper at 2200 m; the latter has a shallow trough on its western edge. No similar microstructures occur on the boundary between the Plateau and the Coral Sea Basin and this may point to a different genetic influence on these two areas.

The Queensland Trough occupies the region between the continental shelf of Queensland and the Queensland Plateau between 14°S and 17°30'S, adjacent to the Great Barrier Reef. Its western margin is much steeper than its eastern margin, with gradients up to 1:3 (at 15°S). The steepness of this slope implies a tectonic origin for the western margin.

The Trough has a smooth, flat floor which gently deepens to the north-northwest from about 1100 m. It joins an embayment region at a depth of about 3000 m between the Queensland Plateau and the Easter Fields Fan north of the area discussed here (Mutter, 1972). It is fed from both sides by canyons (Falvey, 1972).

In the southern part of the Trough, between Flinders and Bougainville Reefs, the profile is simple, but north of Bougainville Reefs the profile becomes more complex (Fig. 12). The eastern wall of the trough appears to rise in two steps; one along about 146°15'E where there is a relatively steep rise through about 500 m, and the second adjacent to the Flinders/Osprey Reefs trend. Gardner (1970) noted an 'abrupt constriction (in the trough) behind which sediment has ponded' at 15°S, between Osprey and Bougainville Reefs. He suggested that differential down-faulting along a lineament between these two reefs was responsible for forming the trough margin in that region. As a result, a small terrace at a depth of about 2000 m has formed south-southwest of Osprey Reef. Most of the eastern flank of the trough probably originated in fault movements, making the Trough a graben or half-graben. This agrees well with Falvey's (1972) interpretation of the Trough.

The strike of the Trough is that of the dominant structural grain of the Tasman Geosyncline in northern Queensland (Hill & Denmead, 1960). It mirrors the trend of, and has approximately the same strike extent as, the North Coast Structural High (Fig. 2). West of this high, structural depressions are the site of the Laura and Hodgkinson Basins. With such a structural framework onshore immediately west of the Queensland Trough it may be speculated that the Queensland Trough lies along a structural 'low' of the Tasman Geosyncline.

The Townsville Trough has no clear relation to any known structure onshore, being roughly perpendicular to the main geosynclinal trend. Falvey (1972) suggested that part of the trough 'is, in appearance, a continuation of the Devonian to Carboniferous Broken Embayment', but such a relation is not at all certain and it is equally possible that the trough reflects oceanic trends such as the Mellish Rise (Cullen, 1970). Solomon & Griffiths (1972) considered that the Broken River Embayment trend, extended east and offshore to the head of the Townsville Trough, is a Palaeozoic fracture zone. Evidence presented later in this paper shows that the Trough is a Cainozoic feature which may, however, lie along an older structural trend.

The Trough has a symmetric U-shaped profile which is maintained over most of its length. At its eastern end, at about 154°E, a bifurcation sends one branch into the Cato Trough, and the other winding sinuously north into the Coral Sea Basin. Sediment derived from the Queensland Plateau or Queensland mainland could reach the deep ocean areas via the Townsville Trough and its offshoots. The profile of the Trough may have been modified into its present gentle U-shape by scouring resulting from sediment transport down the trough.

STRUCTURES AND SEDIMENT DISTRIBUTION

Seismic reflection records from the BMR survey have been analysed to determine the structures underlying and marginal to the Queensland Plateau, and to define the distribution of sedimentary rocks covering the Plateau. Gravity and magnetic data, in the form of profiles and contour maps, have also been studied. The potential field data have been used primarily as an adjunct to the seismic reflection interpretation; a full analysis of the regional field variation was not attempted. Seismic refraction data are also available from the BMR survey but is of doubtful value in many places owing to poor quality. Sonobuoy refraction data given by J. Ewing et al. (1970) are considered preferable.

Seismic reflection records on the Plateau generally reveal a sedimentary cover which is poorly to strongly reflecting. In all but a very few areas a reasonably strong reflection marks the base of the 'sedimentary' reflections. This basal reflector is taken as the effective basement, as penetration into the underlying sequence was seldom achieved. When penetration did occur, such as in areas where the overlying sedimentary cover is thin, banded coherent, or stratified reflections characteristic of bedded sedimentary rocks were observed. In general, however, the 120-kJ sparker had insufficient energy to reveal the fine structure of the basement.

Hence, the acoustic basement found on BMR seismic reflection records is probably a sedimentary or metamorphic rock with a very high acoustic impedance. This observed layer will be referred to simply as 'basement', while the use of qualifying adjectives such as 'crystalline' will be reserved for indirectly observed or interpreted structures.

With a strong basement reflection present over most of the area, sediment thickness and basement topography can be easily mapped. The basement topography and the mapped fault pattern form the basis of the structural interpretation. The sediment thickness, subdivided into sequences according to age and lithology, provides the means of defining the sediment distribution on the plateau.

Structures

A basement elevation map (Fig. 13) has been compiled by scaling the time-depth of the basement reflection at hourly points and converting these values to metres below sea level before contouring the results. To convert from reflection time to metres the velocity of acoustic wave propagation in water was taken as 1500 m/s and that of the rocks lying between seabed and basement as 2000 m/s. The use of a constant 1500 m/s sea-water velocity is justified in the absence of salinity and temperature measurements and because the errors introduced by not applying the Matthews' corrections (Matthews, 1939) are negligible. The errors introduced by using a constant 2000 m/s sediment velocity will vary with the thickness of sediments. From the sonobuoy refraction data of J. Ewing et al. (1970) and from examination of BMR sonobuoy refractions, it would appear that the upper 300 m or so of Plateau sedimentation has a seismic velocity of 1700-1800 m/s and that the deeper strata have velocities of about 2500 m/s. Hence estimates of basement elevation will be too deep in areas where sediment cover is relatively thin, and too shallow in areas of thick sedimentation. As thin sedimentary sections are generally associated with basement uplifts and thick sediments with basement depressions, the effect of using a constant velocity in the middle of the range of true velocities is to creat a topographic map which is a smoothed version of the true topography; the map will be 'under-exaggerated'.

In Figure 14 a group of Matthews' corrections for zone 46 in the Southwest Pacific and a group of possible basement depth errors have been plotted. These clearly show that the assumed velocity contributes insignificant errors; a maximum of 10 m at 2400 m in water depth (or 0.4%), and 75 m at 2000 m in basement depth (or 3.75%).

Although the quality of the BMR sonobuoy refraction data was not good enough to allow determination of a good velocity function for the sediments it does provide a reasonable estimate of basement velocity. The reflection mapped as the basement marker in Figure 13 appears to lie at the top of a rock layer with a seismic velocity in the range 4.9 to 5.6 km/s. This layer varies in thickness from about 0.5 to 2.0 km and overlies a crustal layer with a velocity of about 6.3 km/s. The lower layer marks the true crystalline basement; the layer above probably consists of highly lithified or indurated Palaeozoic sediments of the Tasman Geosyncline. Hence it is the upper surface of Palaeozoic sediments, which have presumably been intensely deformed in the Geosyncline, which now marks the effective basement. A Palaeozoic sedimentary basement is in line with the observation that the acoustic basement in places shows banded reflectors characteristic of sedimentary rocks.

Basement elevation contours shown in Figure 13 define the topography of the surface upon which post-Palaeozoic sediments have accumulated. In the western Queensland Trough and Townsville Trough, basement is so deeply buried that it cannot be recognized. Other unmapped areas occur where large reefs prevented traversing.

The Queensland Plateau can be divided into four large structural provinces on the basis of the basement elevation contours. These have been given the following names for convenience of description only.

Province	*Origin of Name*
Holmes Complex	Holmes Reefs
Willis Platform	Willis Island
Diane Uplift	Diane Bank
Diamond Trough	Diamond Islets

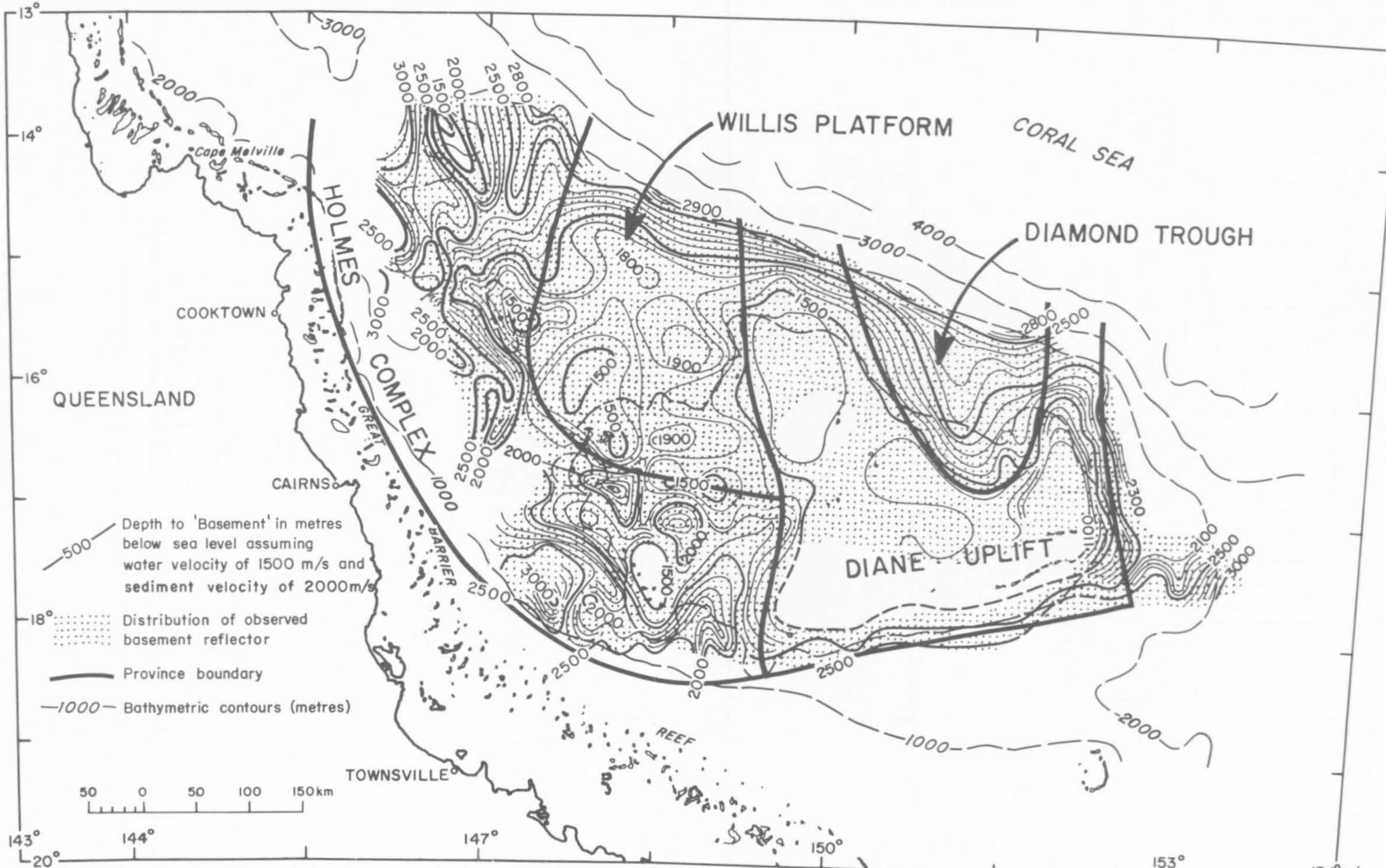


Fig. 13. Basement elevation and structural provinces.

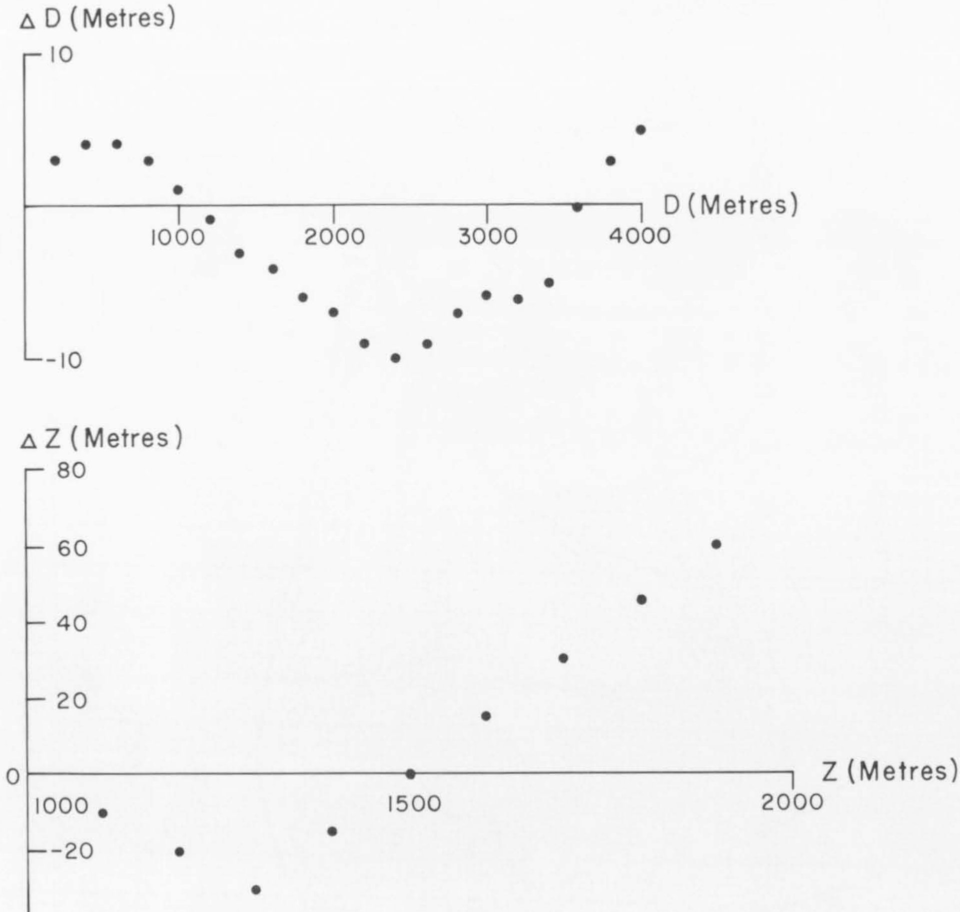


Fig. 14. Estimate of errors in water depth and basement elevation.

The *Holmes Complex* is an elongated zone of strong-relief basement features, which broadly corresponds with the eastern half of the Queensland Trough and western edge of the Queensland Plateau. Basement scarps have up to 1000 m relief. Deep troughs west of Osprey and Flinders Reefs cannot be recognized in the bathymetry (Fig. 11) as they are completely infilled with sediment. Two similarly infilled depressions occur north of Flinders Reefs. Isolated or elongated high areas underlie most of the active reefs along the Flinders-Osprey trend. That is, the basement structure beneath the trend is not continuous as might be suggested by the reef lineation. It is possible, however, that a continuous basement ridge once existed beneath the Flinders-Osprey trend and has been dissected by fault movement into its present discontinuous structure.

In Figure 15 the major faults have been identified. Faulting occurs most commonly in the northwest and west of the Queensland Plateau in the Holmes Complex. Thus this province is a zone of dissected basement, consisting of several basement ridges (some are horsts) and troughs (some are grabens), and normal faulting. These structures are generally associated with tensional tectonics, implying that the Queensland Trough formed by down-faulting along a series of faults on its eastern flank and possibly a single, very long fault on its western flank. The age of the fault movements cannot easily be established and some possibilities are discussed in a later section.

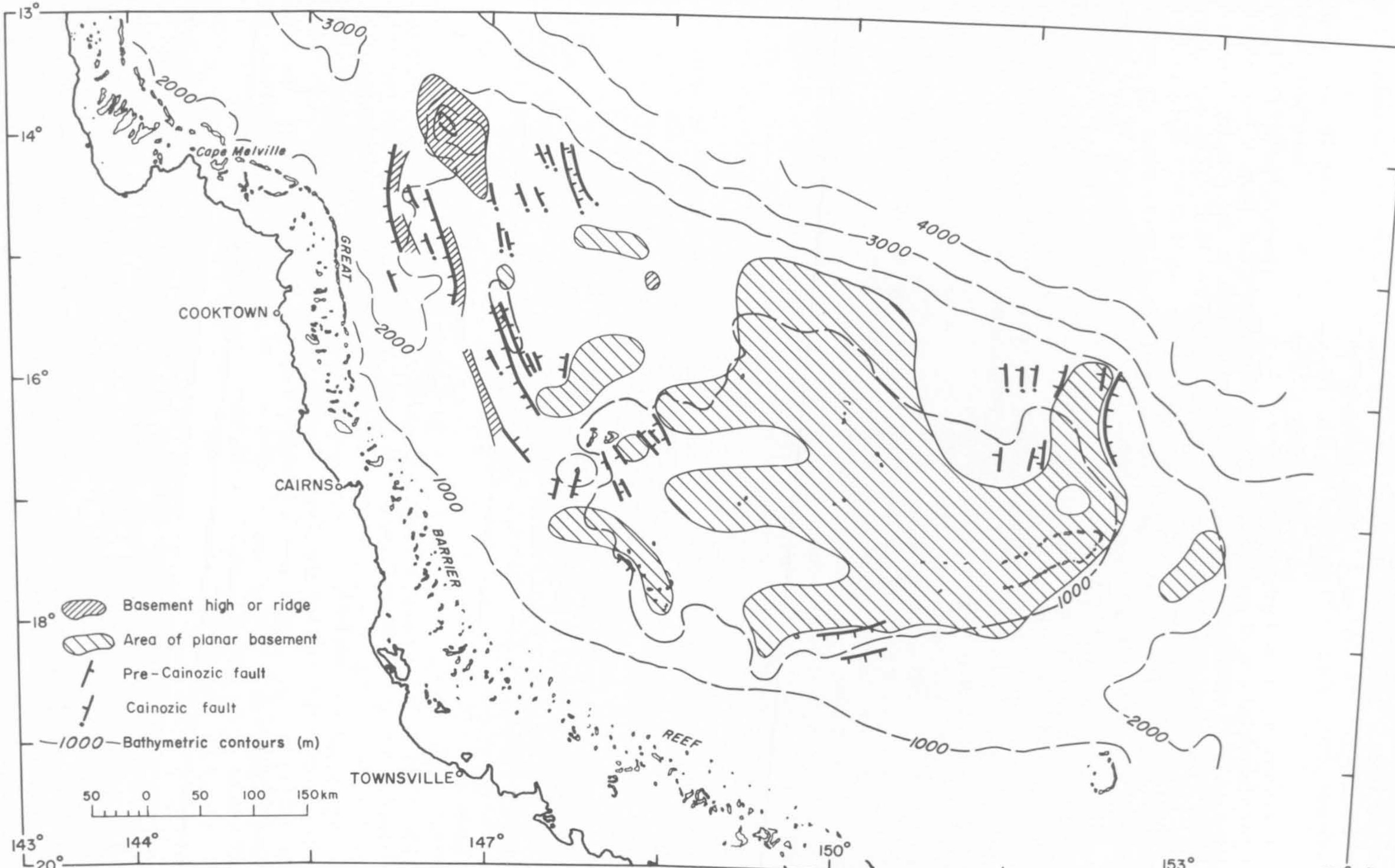


Fig. 15. Fault pattern and basement morphology.

Apart from the Queensland Trough the major basement troughs noted above do not appear to be fault-bounded and may have formed as ancient erosional channels cut into basement.

The *Willis Platform* is a broad flat-lying area of subdued basement relief east of the Holmes Complex and west of the central elevated area or Diane Uplift. The average depth to basement is about 1800 m. It is almost devoid of major faulting, suggesting that the area has been relatively stable for a considerable time.

The *Diane Uplift* is an area of relatively elevated basement, almost flat in its central area, occupying most of the eastern half of the Queensland Plateau. It consists of two limbs between which lies the Diamond Trough. The western limb reaches to within 1400 m of sea level and is elongated north-northwest. The eastern limb trends north and comes to within 1100 m of sea level. In the centre of the western limb and over nearly all of the southern part of the province, prominent reef growth has made it impossible to determine basement depth. It is possible, from rough estimates that can be made by extrapolation from adjacent regions, that the basement in the two reef areas reaches the same elevation as that on the eastern limb.

In Figure 15, areas of a distinctive basement character have been mapped. Within these areas the basement has a very flat or planar surface. By comparing Figure 15 with Figure 13 it can be established that in most areas where the basement lies within about 1500 m of sea level a planar surface has been developed. A characteristic seismic section across the centre of the Plateau (Fig. 16) illustrates the nature of the basement in these areas. It is suggested that the planar surface represents a denudation surface formed by erosion of the basement before submergence of the Plateau. Almost all of the area of the Diane Uplift would have been affected by this erosional episode.

If the erosion hypothesis is correct then the westward fingering of the planar surface on the western side of the Diane Uplift, and the breaking up of the surface into small isolated areas west of that may indicate an east to west directionality that may be explained by tilting of the entire basement surface caused by uplift of the eastern edge of the Plateau. The age of such an uplift cannot be determined but it must be considerably pre-Tertiary and could be related to uplift during the Permian Hunter-Bowen Orogeny (Brown, Campbell, & Crook, 1968) or the Maryboroughan Orogeny (Ellis, 1966).

The *Diamond Trough* may lie within the Diane Uplift, or form a unit completely separate from it depending on how one views its mode of formation. It is a large simple trough which rises in the planated basement area and deepens into the Coral Sea Basin along a north-northwest trend. It may therefore be a very large erosional feature which drained the uplands of the Diane Province, in which case it is part of the Diane Province. However, the trend of the Holmes Complex, both limbs of the Diane Uplift, and the Diamond Trough are all very close to that of the Tasman Geosyncline in north Queensland and it is more reasonable to suppose that the trends of the structural provinces reflect deep fundamental structures in the Geosyncline. The preferred alternative is that the Diamond Trough is a distinct structural unit.

The northern edge of the Willis Platform, the western limb of the Diane Uplift, and the Diamond Trough are all sharply truncated by a steeply dipping basement scarp which has a linear northwest trend. This strongly suggests rifting and supports a similar interpretation made from an examination of the water depths. Hence, Tasman Geosyncline trends have been abruptly terminated since their development by post-Palaeozoic tectonism which split off the northern part of the Geosyncline and caused it to migrate away from the Australian continent. The position of the rifted fragment is undetermined, but may lie in the New Guinea region.

The three eastern provinces are probably more representative of the style of deformation in the Tasman Geosyncline than is the Holmes Complex. They form a succession of gentle elongated swells and depressions somewhat similar to the general character of the Geosyncline onshore (Fig. 3), whereas the Holmes Complex shows an intensity of deformation not commonly observed onshore. This may indicate that the major tectonic event which dissected the basement in this province was not part of the Palaeozoic geosyncline building process but occurred as a more recent event, possibly linked with subsidence of the Queensland Plateau.

Figure 17 is a magnetic anomaly map which has been contoured from the total magnetic field data recorded during the BMR survey. The anomaly is computed by removing the International Geomagnetic Reference Field

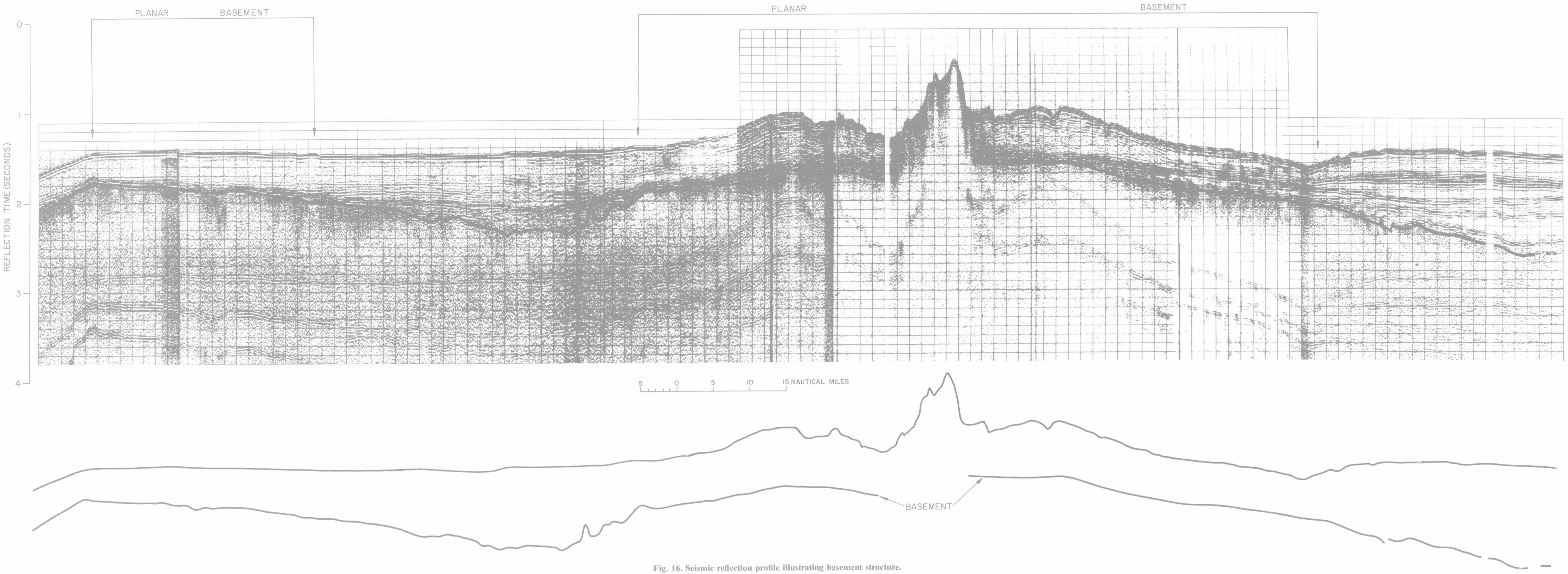


Fig. 16. Seismic reflection profile illustrating basement structure.

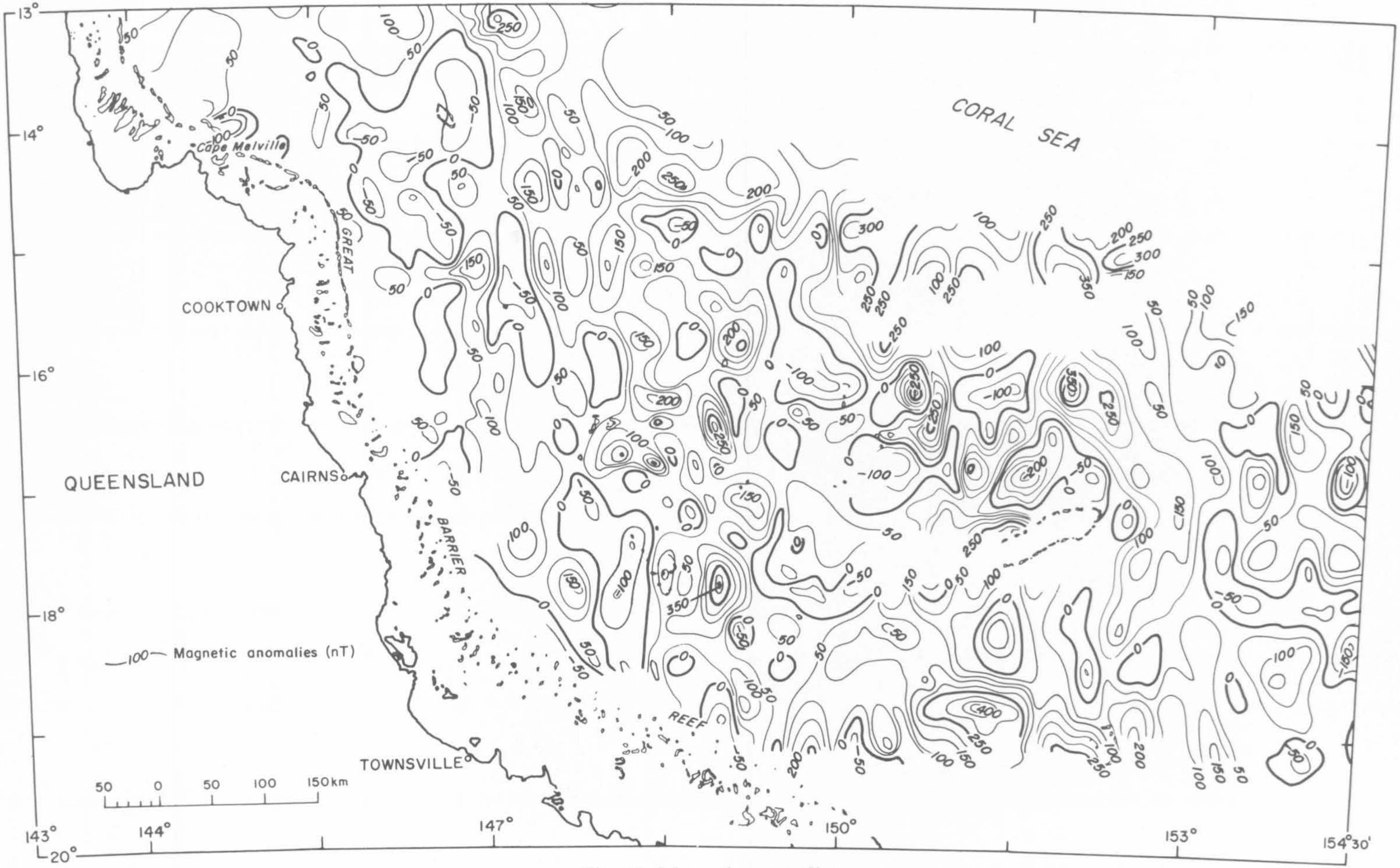


Fig. 17. Magnetic anomalies.

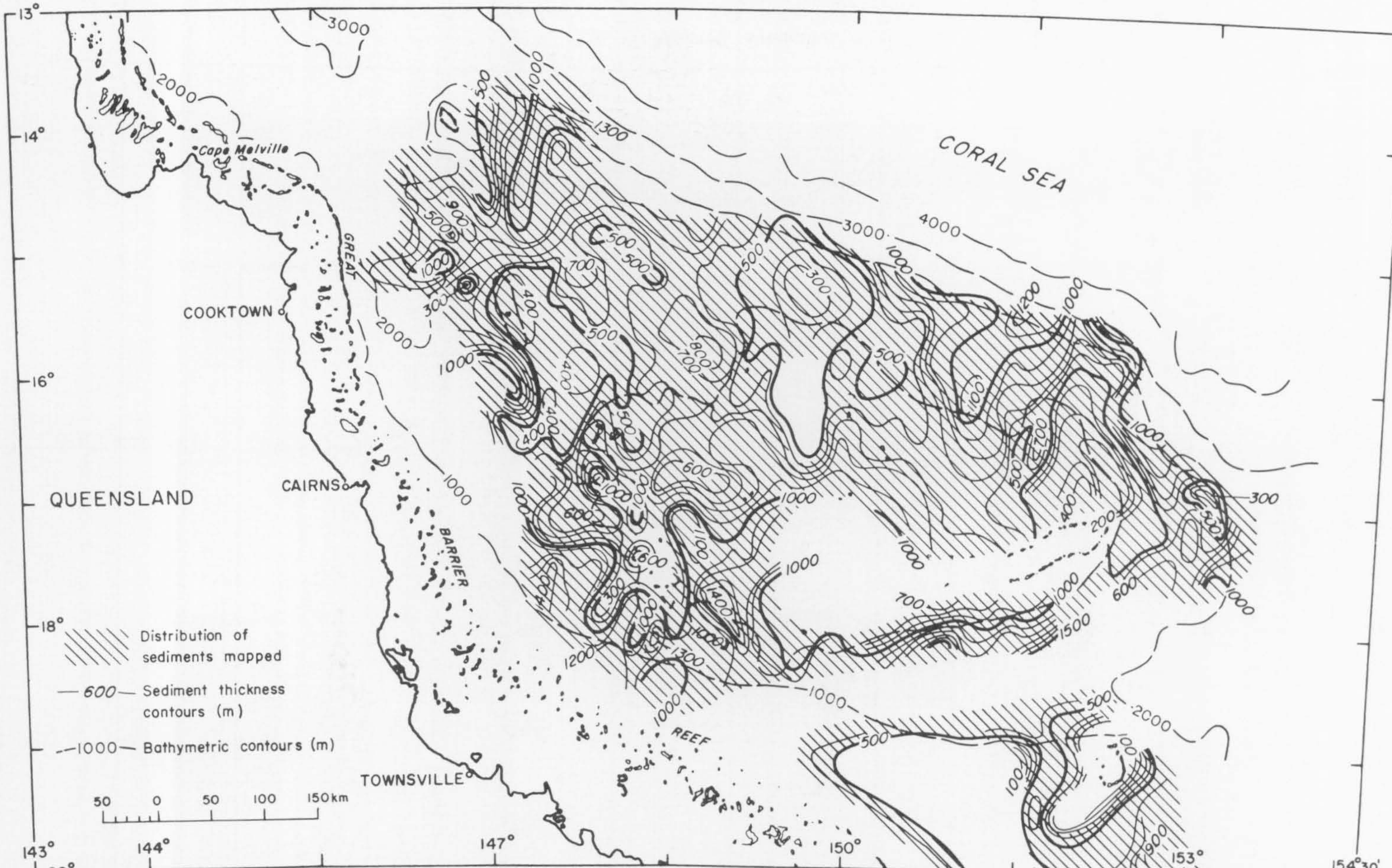


Fig. 18. Total sediment thickness contours.

(Cain et al., 1967). The pertinent deduction that can be made by comparing this map with the basement elevation map is that there is almost no direct correlation between them. That is, the source of magnetic anomalies is not the basement surface mapped. This gives further evidence that the mapped basement consists of highly lithified Palaeozoic strata as was suggested by seismic reflection and refraction data.

Although there is little direct correlation between basement structure and magnetic anomalies, there is an overall tendency for the anomalies to take up a north to northwest trend. This would suggest that the crystalline magnetic basement has the north to northwest trend of the Tasman Geosyncline, which is also exhibited by the gross Palaeozoic basement structural features. Clearly these trends are ancient, possibly lowest Palaeozoic.

Sediment distribution

A total sediment thickness contour map has been compiled for the Queensland Plateau (Fig. 18) by scaling off the seabed-to-basement reflection time at hourly points along the traverses and converting to thickness in metres using a seismic velocity of 2000 m/s. As noted earlier, this assumed velocity will give rise to over-estimates and under-estimates of sediment thickness according to whether the sedimentary layer is thick or thin; the error will rarely exceed 30 m, as before.

As with estimates of basement depth, it has not been possible to produce reliable values for reef-covered areas, or where the basement is deeply buried in the Townsville and western Queensland Troughs. Although it is not always shown in the contours, sediments thin very rapidly towards areas of reef exposure and it can be assumed that sediment thickness decreases to zero at the bases of these areas. In the vicinity of reef development, and in places at a considerable distance from them is an apron of reef-derived or reef-influenced sediment. Here the seismic reflection records show many diffractions and disturbances in the bedding, together with steeply inclined dips and local angular unconformities, which suggests a vigorous influence from the reefs in the form of slumping, mass transport, and the gradual shedding of detrital material. It is sometimes very difficult to recognize basement below such a sedimentary section and hence it is difficult to give an estimate of the thickness of the section. The assumed 2000 m/s sediment velocity could be very inaccurate in such areas.

Depositional maxima occur in the Diamond Trough (1200 m) and local infilled basement depressions in the Holmes Complex (up to 1400 m). The overall distribution can be related to the structural provinces defined above; variable thickness in the Holmes Complex, a fairly thin, even coverage of the Willis Platform, generally very thin except in the centre of the Diane Uplift, and forming a deep basin in the Diamond Trough. Basement elevation has clearly played a large part in controlling plateau sedimentation, resulting in deeply infilled depressions and thinly covered elevated areas. This distribution would imply that much of the sediment was derived from material transported by turbidity currents as these would flow from elevated to low-lying areas, causing the gradual infilling of depressions. It will be shown below that the nature of the sedimentation has changed progressively throughout the Cainozoic from a dominantly terrigenous (turbiditic) to a wholly pelagic type.

Correlation between sediment thickness or basement elevation and the free-air anomaly contours in Figure 19 is generally reasonable. Areas of elevated, thinly covered basement display relatively positive anomalies whereas deep, thickly covered areas show relatively negative anomalies. The whole of the Queensland Plateau shows regionally positive anomalies, while the Queensland and Townsville Troughs are almost entirely negative. This regional correlation probably reflects isostatic imbalance rather than sediment distributions. The generally good correlation of sediment thickness to free-air anomaly probably indicates that the main density contrast lies at the interface of sediment to Palaeozoic basement. Simple infinite slab calculations suggest that the contrast might be as high as 0.7 g/cm^3.

The sedimentary sequence of the Plateau can be broken into three units based on their acoustic properties and they can be correlated with the three lithological units in DSDP 209 (Burns, Andrews, et al., 1973). Although the correlation between seismic reflections and breaks in lithology or physical properties made by Andrews (1973) from *Glomar Challenger* seismic records was very poor for both holes 209 and 210, it is contended here that a very good tie can be made to BMR seismic reflection records and this has enabled the detailing of the sedimentary history of the Plateau through the Cainozoic.

In Figure 20 the seismic reflection profile recorded on the traverse which most closely

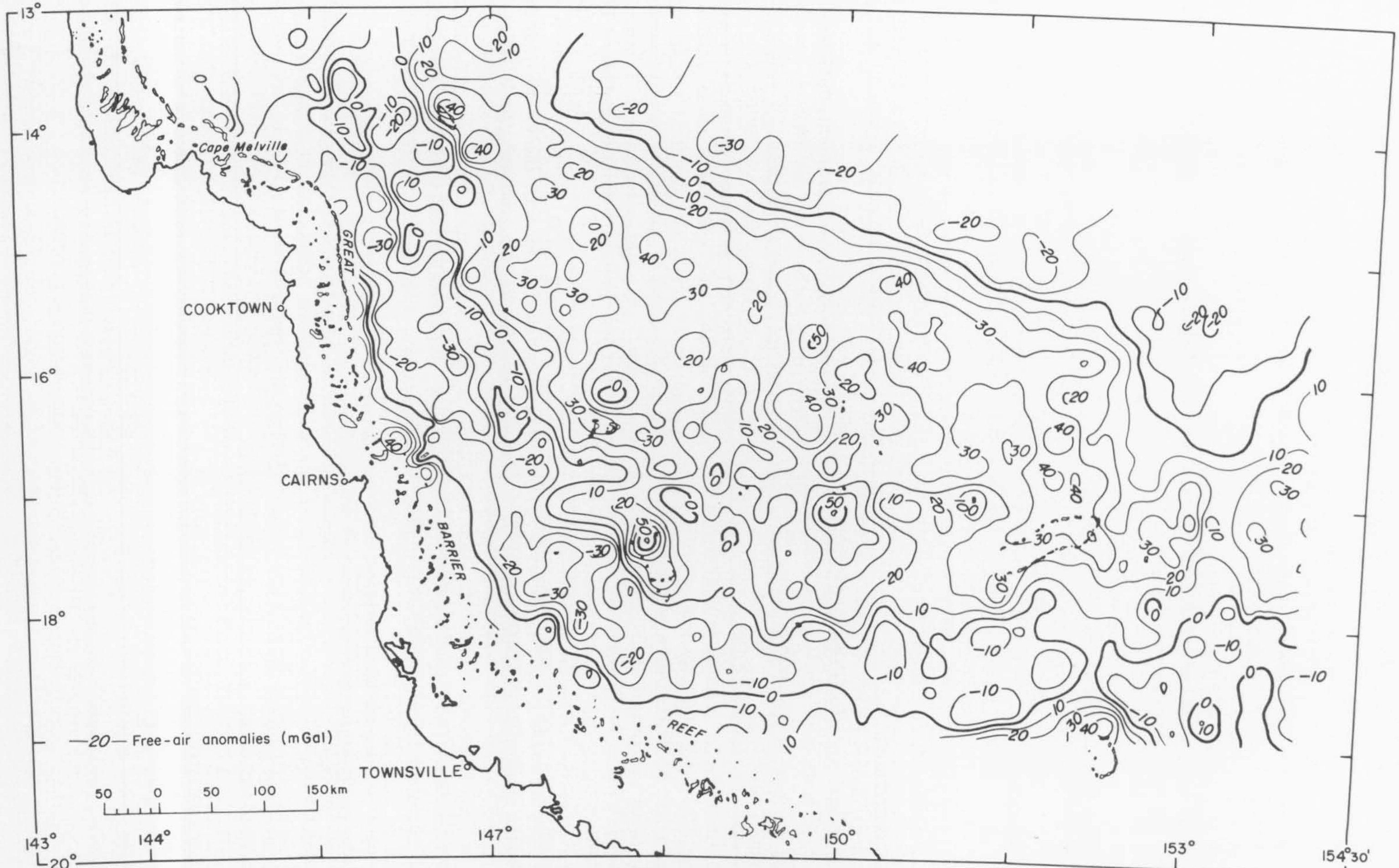


Fig. 19. Free-air anomalies.

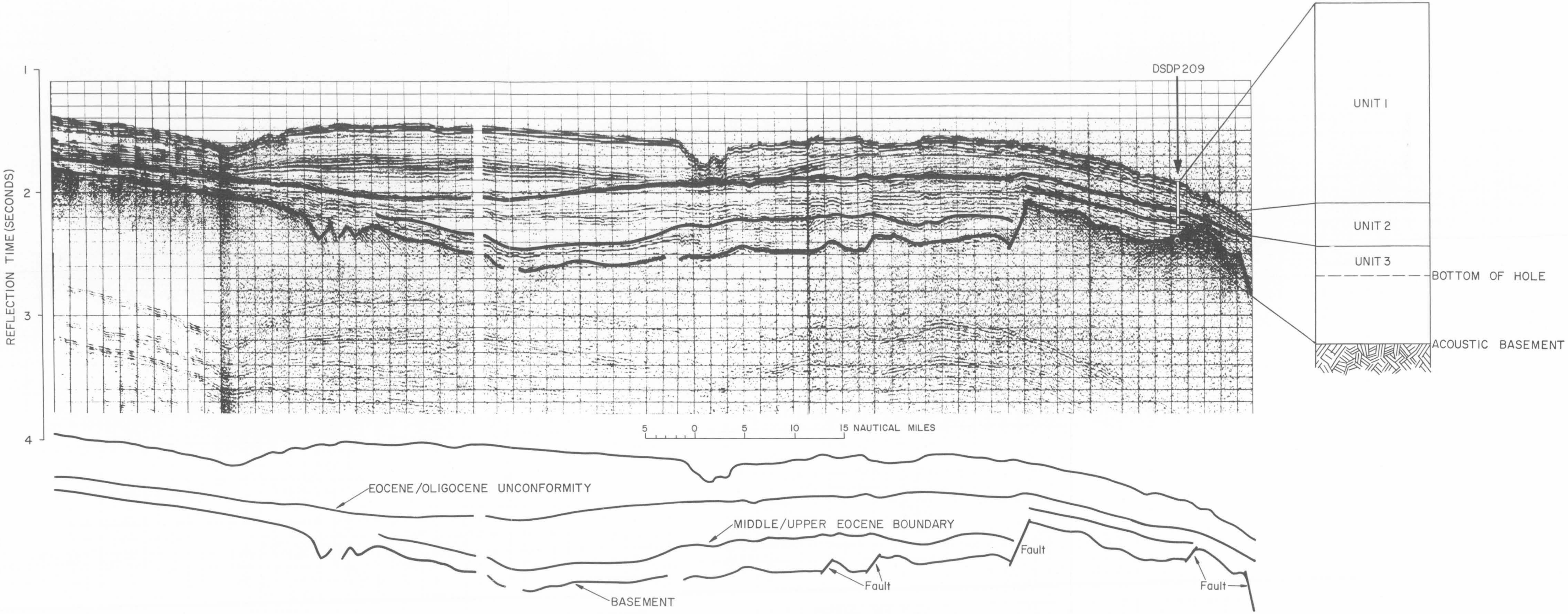


Fig. 20. Seismic reflection profile illustrating correlation with section from DSDP 209.

approached DSDP 209 is displayed. The drill site is about 4 n miles south of the reflection profile and the position shown on the profile is that point which has the same water depth as that of the drill site. If the interpolation onto the profile was made as a projection due north then the location would be about 1.5 n miles east of the position shown. As the BMR reflection profile does not precisely intersect the drill site there is some doubt about the best positioning of the comparison reflection section. In addition, navigational inaccuracy of greater than 1 n mile is expected when dead-reckoning between satellite fixes in deep-water regions. Hence the indicated location of DSDP 209 is an estimated position probably accurate to no better than plus or minus 2 n miles.

Alongside the reflection profile in Figure 20 a simplified version of the results from DSDP 209 is shown for comparison. The depth scale of the column has been converted to reflection time using the depth data of Burns, Andrews, et al. (1973), and sediment velocity data and seismic reflection descriptions of Andrews (1973). This allows a direct comparison between stratigraphic column and seismic reflection profile.

The position of the boundary between lithological units 1 and 2, and 2 and 3 have been projected onto the reflection profile in Figure 20. It is immediately clear that the uppermost boundary correlates well with a prominent reflection. This reflection has been identified in the line drawing of the reflection profile and as it corresponds to the first lithological boundary it marks the Eocene/Oligocene unconformity. The reflection is mappable over much of the Queensland Plateau and its structure is described below. The next-best correlation is that of the basement reflections. These match within about 20 ms or 17 m, and the mismatch could be removed by shifting the location of DSDP 209 on the profile about 0.7 n mile to the east. This is well within the positioning error. The boundary between units 2 and 3 is not correlated with a prominent reflection similar to the upper unit boundary or the basement. It is, however, very close to an anomalous high-frequency reflection which has been indicated on the line drawing below. This reflection event also has regional occurrences and it is its regional structure which provides the main argument for correlation with DSDP 209.

As exemplified by the seismic reflection profile of Figure 21, the lowermost acoustic unit over most of the Willis Platform, Diamond Trough, and eastern Holmes Complex, is a moderately disturbed but highly reflecting layer lying above basement and bounded on its upper surface by a high-frequency reflection (greater than 100 Hz, compared with the norm of about 50-60 Hz). Sediments above this layer are generally much less structured and commonly poorly reflecting. The lower layer pinches out against basement rises, and the level at which the upper bounding reflection meets the rising basement layer is almost immediately adjacent to an area of planated basement. In some places the lower layer onlaps the planated basement but in general the lower layer is found filling relatively depressed areas of basement adjacent to areas of relatively elevated, planated basement.

The simplest, most convincing explanation for the lower sequence just described is that it originated by the accumulation in low-lying areas of the products of the denudation of the now-planated basement areas. If this is so, then it probably correlates with the middle Eocene shallow-water sediments which form the lowermost unit in DSDP 209. It is contended here that the lower unit of highly reflecting strata on the Plateau is the same lithological unit as the lowermost sequence in DSDP 209 and that it can be mapped as a sedimentary sequence representing shallow-water deposits over most of the Plateau.

In Figure 22 the *Shallow-Water Sequence* has been mapped. It is found in the Diamond Trough where the depositional maximum occurs, over the Willis Platform where it is fairly evenly distributed, and on the eastern edge of the northern Holmes Complex. Over most of the Diane Uplift and in isolated areas on the Willis Platform and Holmes Complex the shallow-water sequence is not represented, having pinched out adjacent to rising basement surfaces. In the western and southern Holmes Complex the shallow-water sequence cannot be recognized in the acoustic zonation.

The shallow-water sediments show considerable structural modification. Structures are syn-depositional or post-depositional. Upturning of beds caused by differential compaction, some folding, and draping make up the first type; dislocation by faulting makes up the second type. Both types are represented in the seismic reflection profile in Figure 23. Post-depositional structures are important in that they reflect tectonic activity, probably associated with the early stages of plateau subsidence in the uppermost middle to late Eocene.

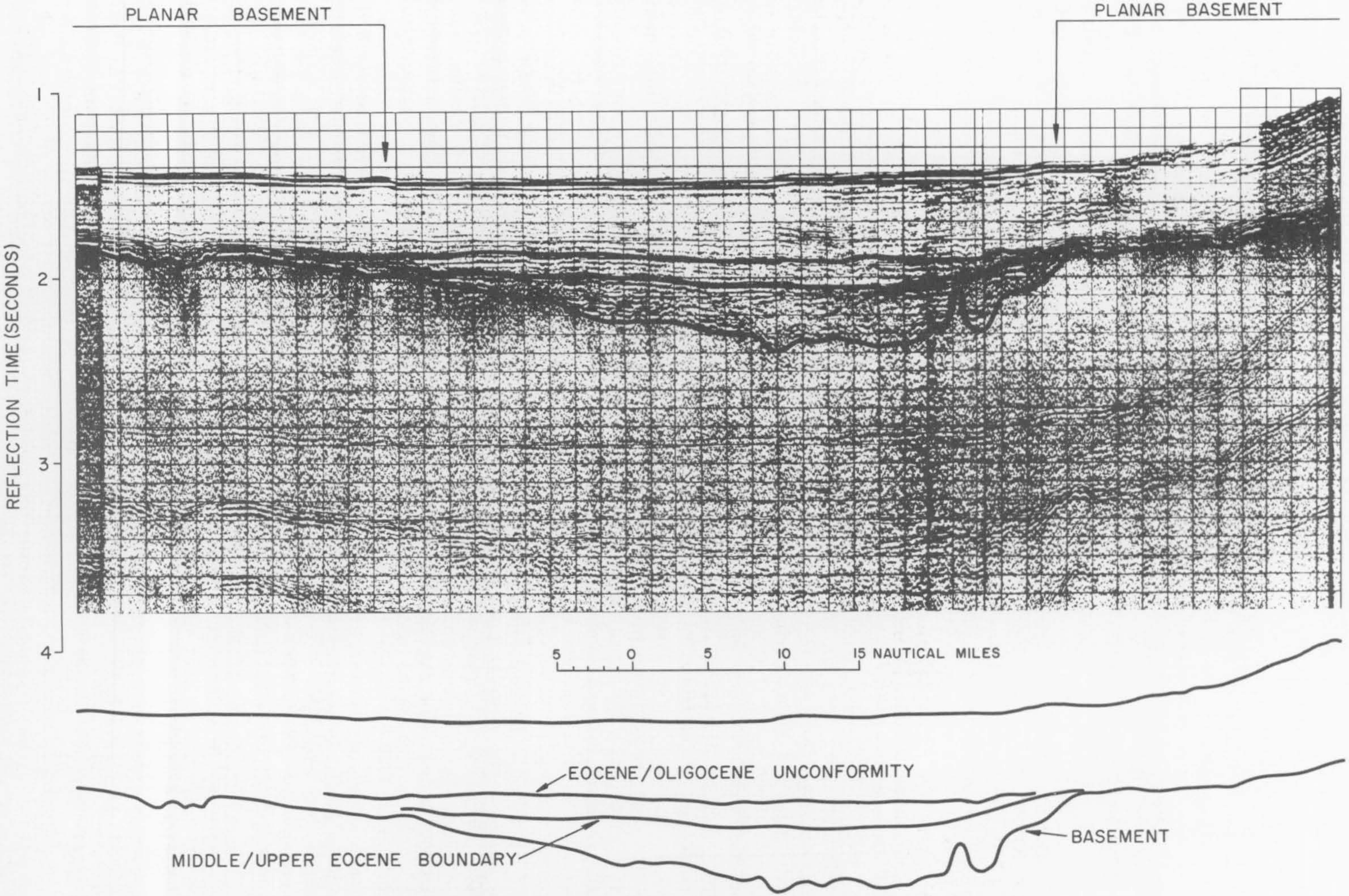


Fig. 21. Seismic reflection profile illustrating the distribution of shallow-water sediments.

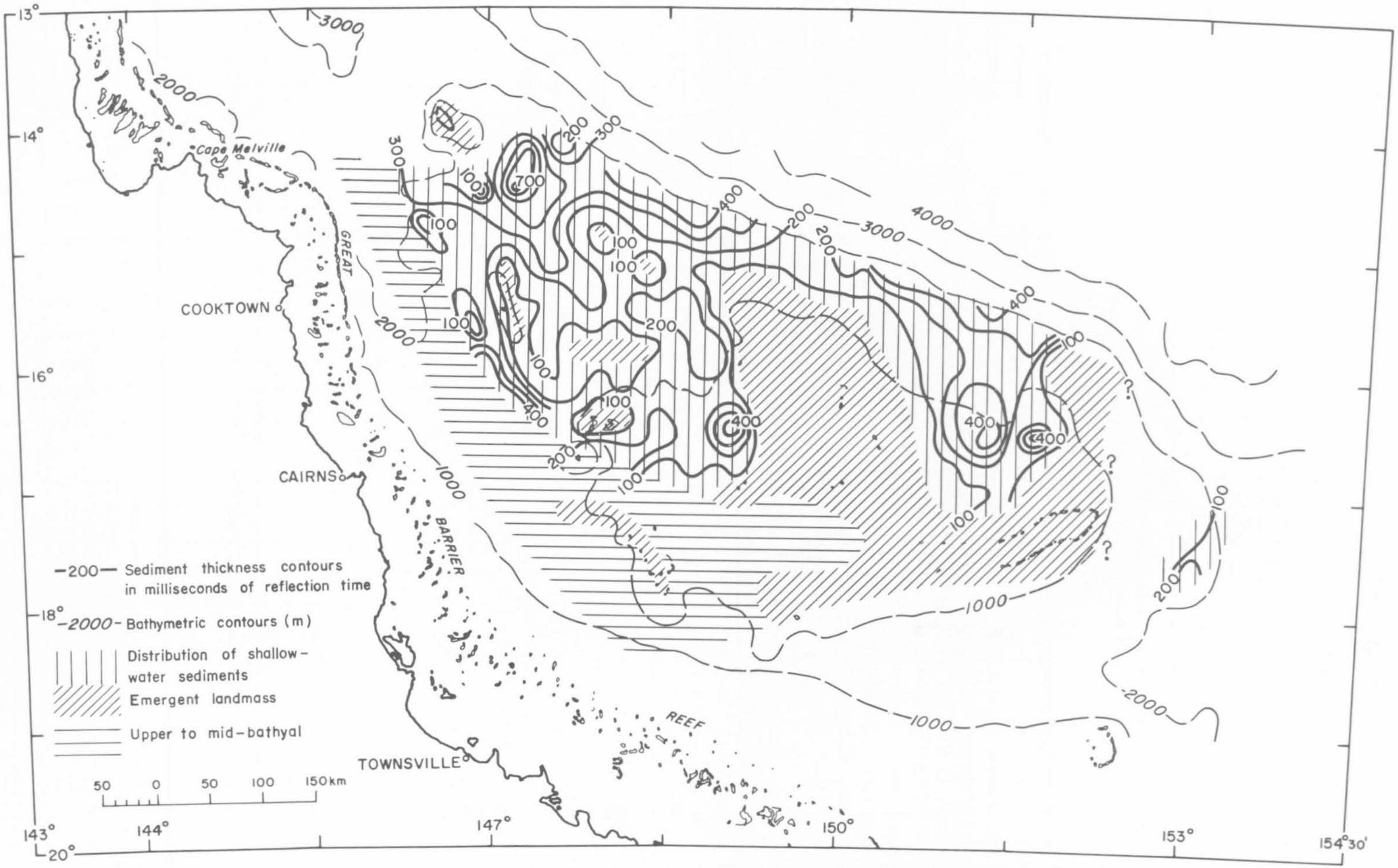


Fig. 22. Middle Eocene depositional provinces with distribution of shallow-water sediments.

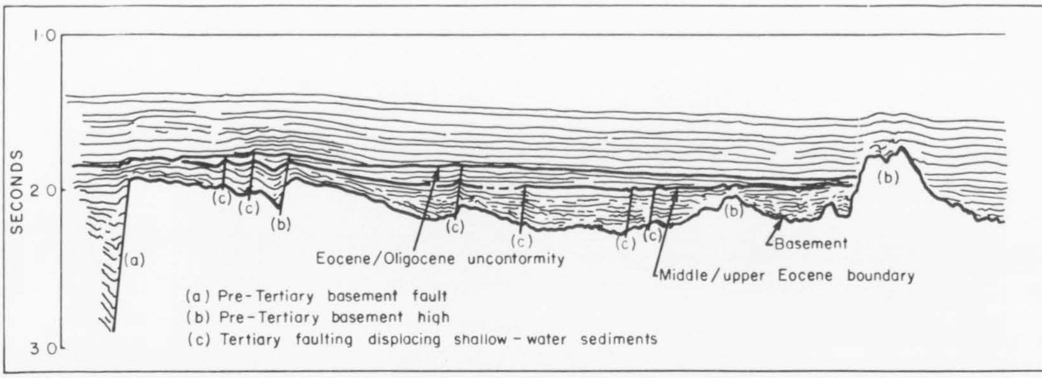


Fig. 23. Seismic reflection profile illustrating structure in the shallow-water sediments.

The distribution of shallow-water sediments has been interpreted in terms of palaeo-depositional environments for the middle Eocene. Areas where the shallow-water sequence is not present are interpreted as having been emergent at this time. The correlation between middle Eocene emergent areas and areas of planated basement is very good. As planated basement is generally found only in areas elevated to within 1500 m of sea level, this implies that 1500 m of subsidence has occurred since middle Eocene time. Areas of the western and southern Holmes Complex, where the shallow-water sequence could not be identified, are interpreted as having been moderate to deep-water areas during the time at which shallow-water conditions existed elsewhere in the submerged areas of the plateau. These areas correlate well with the regions where the basement is now most deeply buried.

The regional palaeogeography of the middle Eocene comprises a large emergent landmass of low relief, shedding material as erosion products into a surrounding shallow-water area with some islands to the west, and a relatively deep-water trough separating these areas from the mainland.

The depth to the top of the shallow-water sequence has been mapped in Figure 24. The descent into the Queensland Trough is steep. This is interpreted as representing post-depositional movement along faults in the Holmes Complex. The surface is very flat in the Willis Platform and generally northward-dipping in the Diamond Trough. If all the sediment deposited in the Diamond Trough during the middle Eocene is a genuine shallow-water sequence laid down in depths of 200 m or less, then the tilt on the sediments observed in the trough would be post-depositional as it is unlikely that compaction could account for the whole effect. This would indicate differential subsidence of the Diamond Trough relative to the Willis Platform during and after the middle Eocene. Faults mapped in Figure 15 on the eastern side of the Trough could be the results of subsidence or pre-existing structures along which differential subsidence occurred.

After deposition of the shallow-water sequence in the middle Eocene, results from DSDP 209 indicate a lithologic change to sand-bearing foraminiferal ooze and chert resulting from a change in depositional environment caused by deepening of the site. The top of the upper Eocene sequence is bound by an unconformity representing an hiatus in sedimentation between the late Eocene and the late Oligocene. Although Andrews (1973) could not identify a seismic reflection event correlative with this hiatus, a clear reflection is present on BMR reflection records at the correct depth in the DSDP hole.

In Figure 20 this unconformity is well developed. Reflections are conformable below, and make an angular contact above, the unconformity. This behaviour is characteristic and allows the unconformity to be mapped over the Plateau.

Sediments deposited after the cessation of shallow-water conditions and before the Eocene/Oligocene hiatus are a mixture of upper Eocene terrigenous and pelagic types. Terrigenous content has been reduced to 10-30 percent, and the deposits reflect a change from near-source terrigenous to open-marine pelagic sedimentation.

In Figure 25 the thickness of the upper Eocene sequence has been mapped. The sequence pinches out against shallow-water sediments or basement in the centre and south-east of the plateau and in two ill-defined areas to the west. In the Queensland Trough the

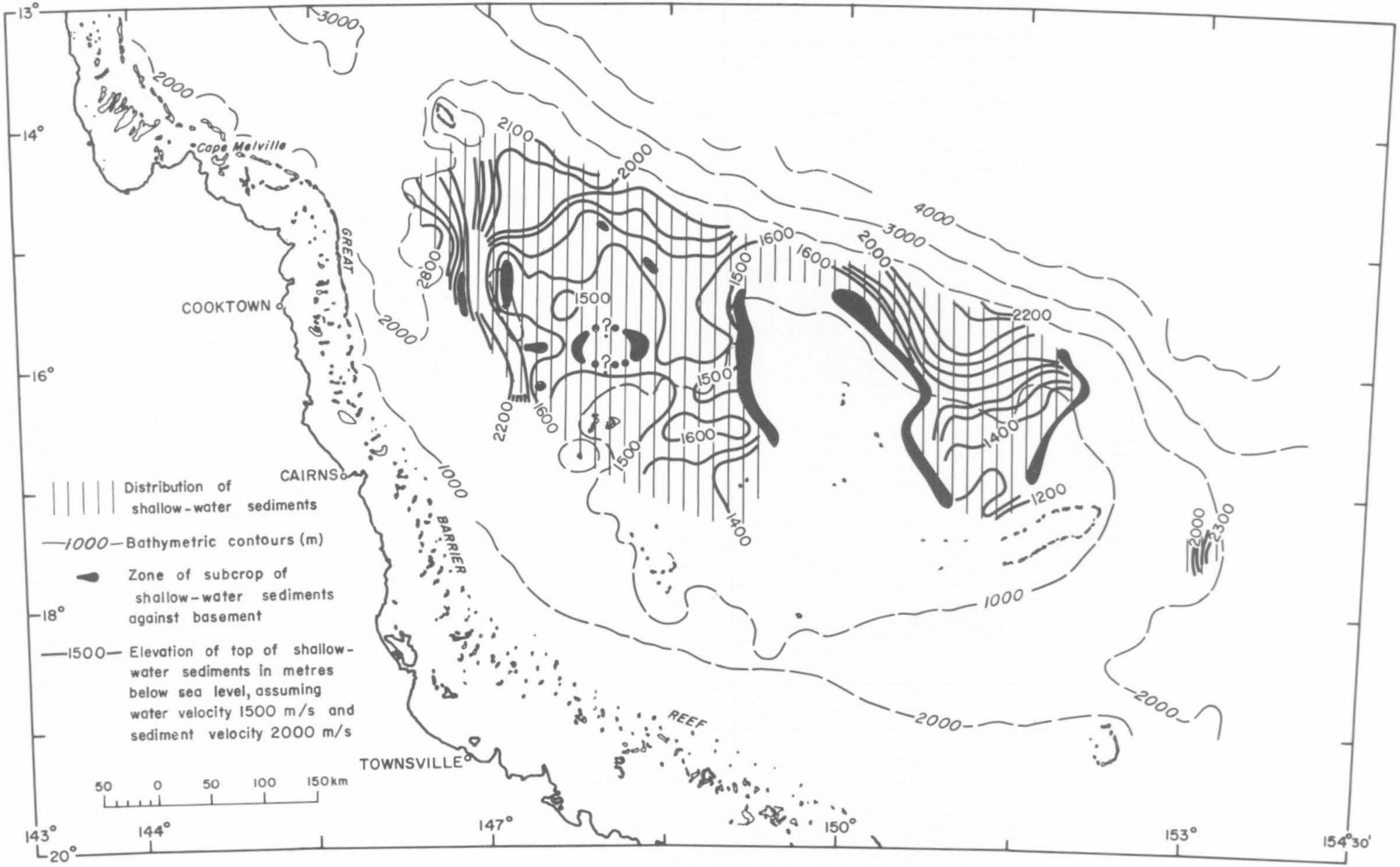


Fig. 24. Topography of the middle/late Eocene boundary.

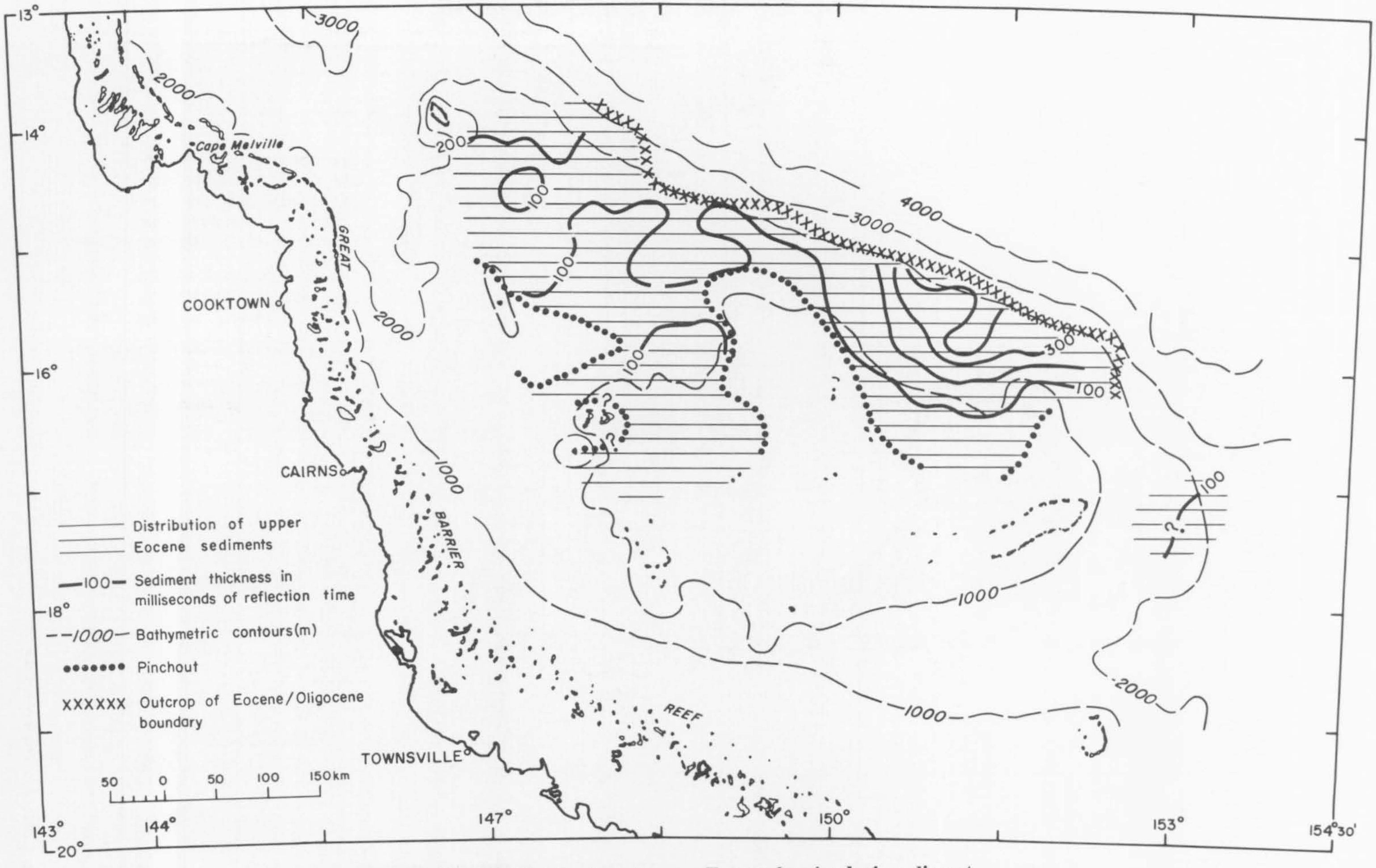


Fig. 25. Distribution of upper Eocene hemi-pelagic sediment.

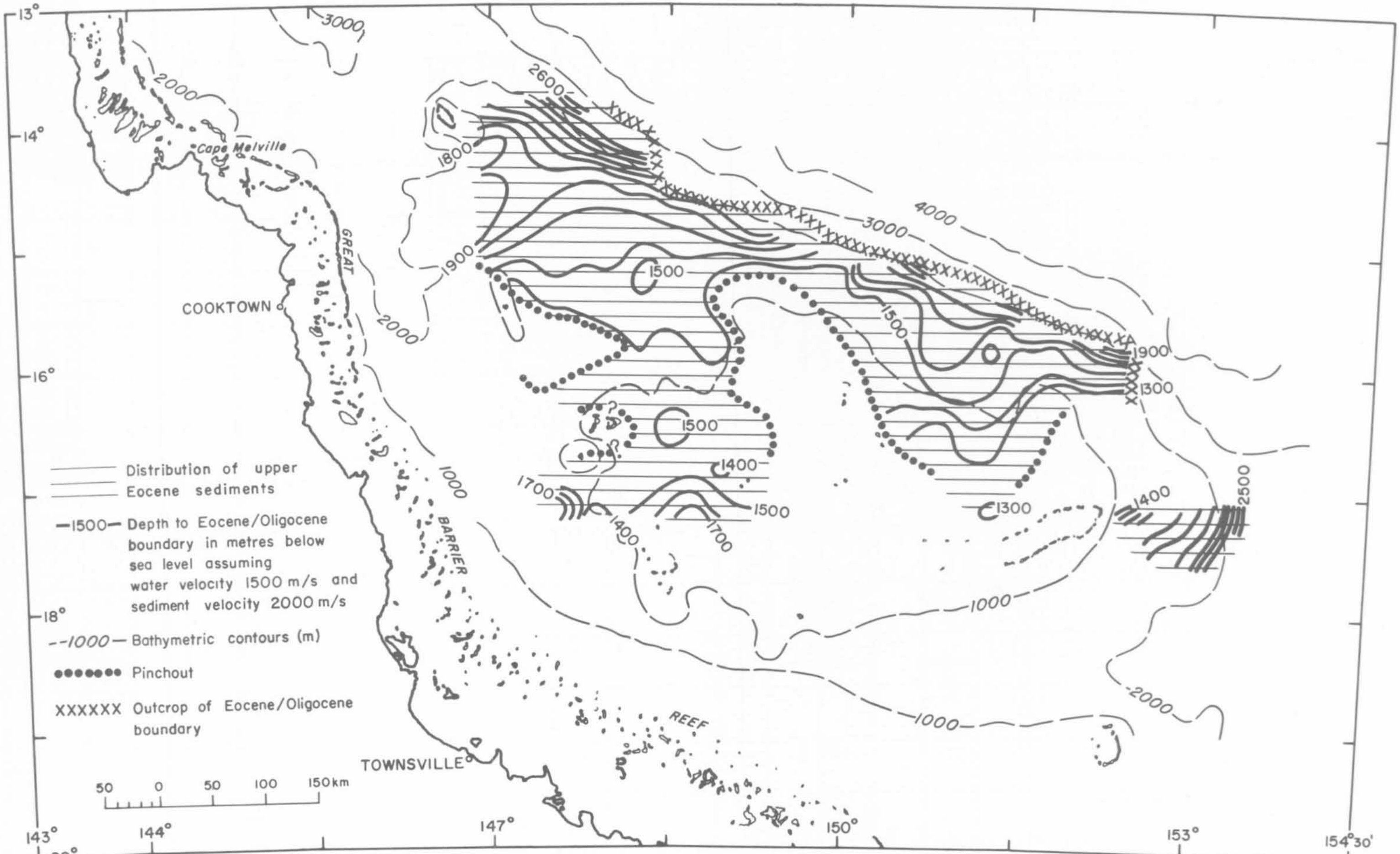


Fig. 26. Topography of the Eocene/Oligocene boundary.

Eocene/Oligocene unconformity could not be identified, but Pinchin & Hudspeth (1974) have attempted to map it from sonobuoy refraction data. On the outer edge of the Plateau the unconformity crops out about half-way down the slope.

Areas within the pinchout zones probably represent late Eocene emergent areas. The middle Eocene coastline had retreated considerably by the late Eocene, and the low islands in the Willis Platform were submerged. The coastline had retreated far more on the east of the western limb of the Diane Uplift than on its western side, and the northern part of the eastern limb was submerged. This may point to a tilted form of subsidence which affected eastern areas more than those to the west. As further evidence, the upper Eocene sediments are very thinly and uniformly distributed over the Willis Platform and are absent in some places. The depositional maximum is again in the Diamond Trough. Hence the differential subsidence which probably occurred after the middle Eocene deposition may have continued beyond the late Eocene.

Upper Eocene sediments are almost completely unstructured. Some of the faulting which disturbed the middle Eocene sequence has had a small affect on the upper Eocene deposits. This generally involves the propagation of drape structures into the overlying sequence. In general the amount of disturbance of upper Eocene deposits is small compared with that of the middle Eocene rocks. They may indicate a change through these times from orogenic to epeirogenic subsidence of the plateau.

The topography of the Eocene/Oligocene boundary has been mapped in Figure 26. Again the Willis Platform is a very flat area and the Diamond Trough contains sediments gently dipping to the north. The boundary reflector dips more steeply down the edge of the Plateau before it crops out on the more steeply sloping Plateau surface. As the upper Eocene sequence is largely pelagic, and there is little evidence of post-depositional movement, it is probable that the mapped surface formed as a result of deposition and has undergone little subsequent modification. As differential subsidence probably occurred during the formation of this surface it was probably achieved by tilting or regional warping rather than by fault movements.

When deposition recommenced after the Eocene/Oligocene hiatus, blanket pelagic sedimentation continued until Pleistocene time. This final stage of sedimentation is correlative within the upper lithologic unit in DSDP hole 209, which is a foraminiferal and nannofossil ooze. Terrigenous content in the sequence is very small at less than 5 percent (Burns, Andrews, et al., 1973). Some sedimentation occurred in the late Oligocene but most took place in the Miocene, Pliocene, and Pleistocene so the sequence will be referred to as the Neogene pelagic sediments.

The *Neogene pelagic sediments* are almost uniform in thickness over the plateau (Fig. 27). The Diamond Trough no longer shows a depositional maximum. The north-northeast-trending trough on the western side of the Diamond Trough is not a depositional feature but is the result of excavation by a large canyon rising near Magdelaine Cays (Fig. 11). The Neogene pelagic sediments could not be identified near major reef growth but this is due to lack of definition rather than the absence of the sequence. Similarly the sequence could not be resolved in the Queensland and Townsville Troughs, but there is no doubt that Neogene sediments exist in these areas.

The Neogene pelagic sediments are almost completely undisturbed by faulting. They were deposited in large swells which do not conform with the topography of the Eocene/Oligocene boundary surface. This has given rise to an inverted angular unconformity as seen in Figure 20. The affects on deposition of submarine currents during the late Oligocene to late middle Miocene were noted in DSDP 209 and these may have produced the broad swells in the Neogene pelagic sediments. The even distribution of the sediments would suggest uniform subsidence of the plateau during the Neogene, i.e. differential subsidence probably ceased at about the Eocene/Oligocene boundary.

As a means of checking the quality of the dating of acoustic units on the Queensland Plateau, sedimentation rates for each unit have been computed and compared with the average rates obtained in DSDP 209. These are listed in the table below.

TABLE 1. SEDIMENTATION RATES
(cm/1000 years)

Period	*Willis Plateau*	*Diamond Trough*	*DSDP 209*
Pre-late Eocene (shallow-water sediments)	0.6	1.1	1.6
Late Eocene	2	6	2.7
Neogene	1.7	2.1	0.9

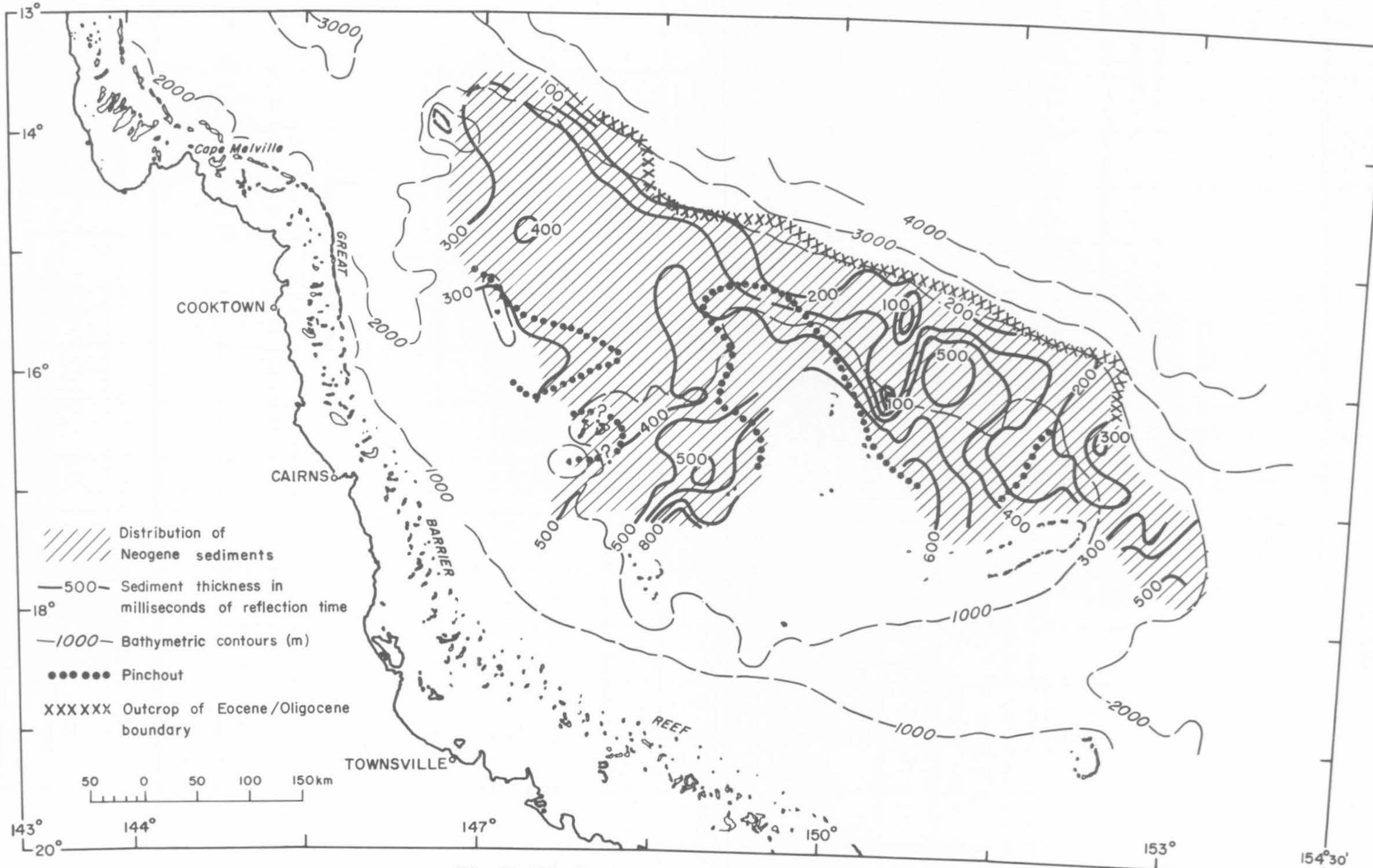


Fig. 27. Distribution of Neogene pelagic sediments.

Rates for the Queensland Plateau in the pre-late Eocene are computed assuming commencement of shallow-water sedimentation in the Cretaceous. The computed rates are slower than that observed for DSDP 209. This may be due to the fact that the drill site was located very close to a middle Eocene emergent area and may therefore have experienced a somewhat higher rate of sedimentation than the average. Alternatively, the Late Cretaceous commencement date may be incorrect. To equalize the computed rates with the DSDP rates a commencement date of about 60 m.y. B.P. would be required. This date corresponds to lowermost Cainozoic and is probably much too early to accommodate the period of denudation of basement highs which has resulted in shallow-water deposition in basement lows. Arguments for the age of this denudation cycle are given in the following section.

The late Eocene rates compare fairly well with the DSDP rate, being towards the lower end of the Queensland Plateau rates. Neogene rates on the Plateau are higher than that found in the DSDP hole. The generally lower rates in the DSDP hole for the late Eocene and Neogene may result from the fact that the hole was drilled on an elevated basement structure (Fig. 20) and hence the overlying column may be somewhat compressed with respect to the average.

Within reasonable limits of variability in sedimentation rates, the comparison between rates computed for acoustic layers in reflection profiles and that for DSDP lithologic units seems good and provides supporting evidence for the validity of the seismic interpretation.

GEOLOGICAL HISTORY

The Cainozoic history of the Queensland Plateau is difficult to establish as no direct sampling or observation of Mesozoic or earlier rocks have been made. One method of inferring pre-Cainozoic history is to extrapolate the onshore events into the adjacent ocean areas. This has been done by Pinchin & Hudspeth (1974) from the palaeogeographic maps of Brown, Campbell, & Crook (1968) and other data. Strucural development of the geosynclinal fabric of the Plateau began in the Ordovician (Packham, 1973) and was essentially complete by the Permian or Triassic. It resulted in the creation of subparallel, north to northwest-trending highs and lows which are recognizable on the Queensland Plateau in the structural provinces defined on the basis of basement elevation in the previous section (Fig. 13).

The latest pre-Cainozoic event recognizable in the subsurface of the Queensland Plateau is the erosion of geosynclinal highs and deposition of the erosion products into adjoining troughs. This presumably resulted from an uplift stage, the age of which cannot be pinpointed. It may be related to the Permian Hunter-Bowen Orogeny which produced considerable uplift onshore. If the uplift occurred in the Permian this would allow about 150 million years of erosion to have taken place before subsidence in the Early Cainozoic. Although transgressive and regressive cycles would probably have occurred on the Queensland Plateau after the Permian uplift, 150 million years seems too long. One might have expected considerable thicknesses of Mesozoic sediment to be deposited in low-lying areas as erosion products of the elevated regions. The thickness of shallow-water sediments, interpreted as sediments derived from the erosion process, rarely exceeds 400 m (Fig. 22) and in general is about 200 m. This figure seems to be too low to have been the result of such a long period of erosion.

A considerably younger event produced an erosional epoch in the Maryborough Basin. Ellis (1966) described 'a severe orogeny (which) followed the Aptian depositional phase'. The time of the orogeny is poorly dated, 'but by the Oligocene, the folded post-Jurassic succession had been eroded to base level'. General subsidence then occurred in the Oligocene-Miocene. Aptian ends at 106 m.y. B.P., allowing an erosional period with a maximum duration of about 70 million years before subsidence. If the post-Aptian orogeny caused uplift on the Queensland Plateau an erosional period of about 60 million years would have prevailed before Early Cainozoic subsidence. It is unlikely that a severe orogeny would have been restricted to the Maryborough Basin, but it may have affected eastern oceanic areas more than the western continental zone. If it is related to postulated sea-floor spreading and continental rifting (Packham, 1973) in the Late Cretaceous then one would expect the eastern zone to be more strongly affected; if not by orogeny, at least by thermal uplift associated with lithospheric heating at the spreading centre.

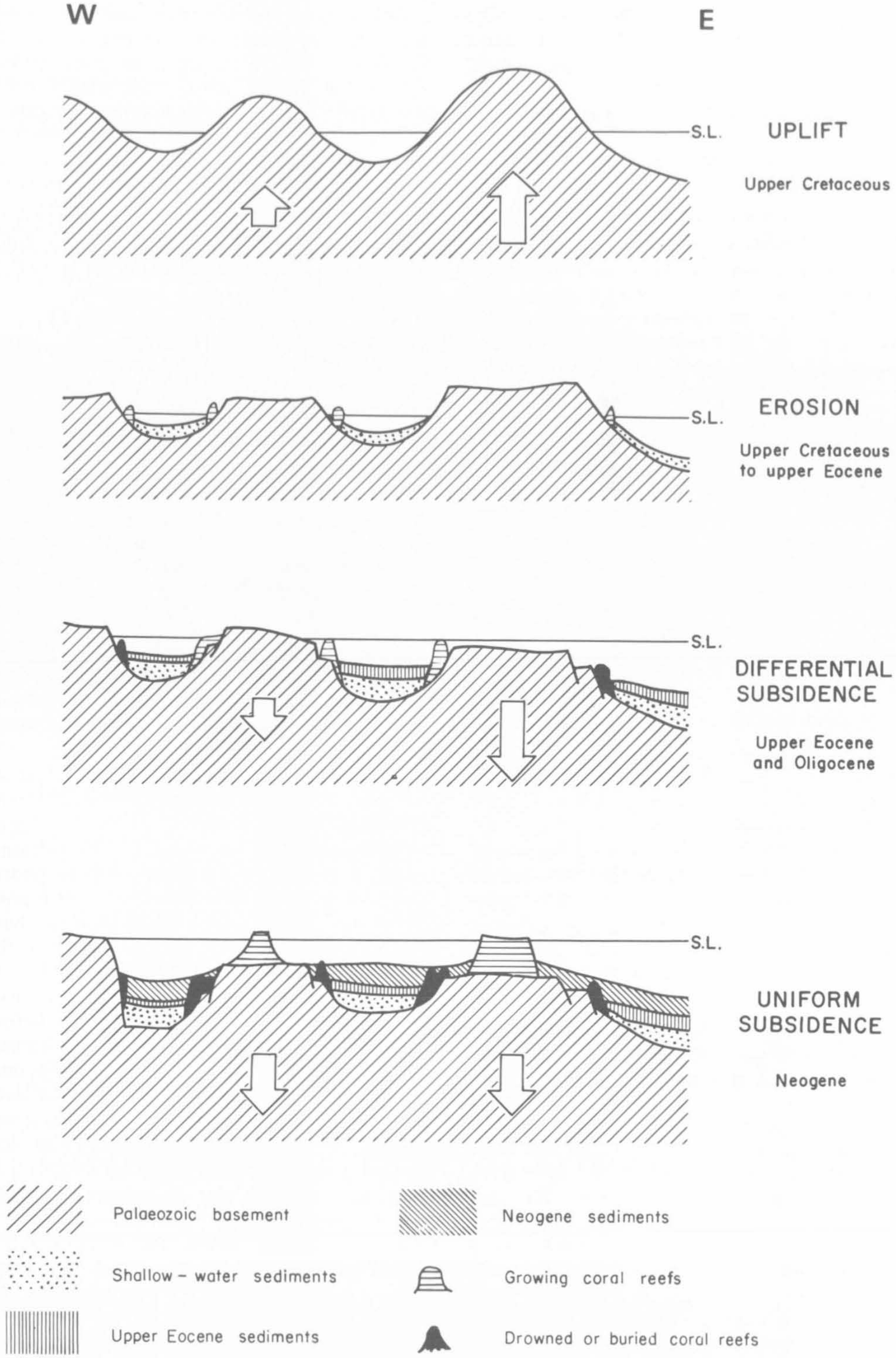


Fig. 28. Simplified geological history of the Queensland Plateau.

A *Late Cretaceous* time (say 80 m.y. B.P.) is preferred as the age of inception of uplift and erosion of the Queensland Plateau, which resulted in formation of the planar basement surface observed over much of the Plateau. The Permian Hunter-Bowen episode may have produced another erosional phase not recognizable on the Plateau.

Figure 28 is a schematic cross-section of the Coral Sea Plateau showing its development from the Late Cretaceous onward. Between the Late Cretaceous and the time of major subsidence, shallow to moderate marine conditions existed on the Queensland Plateau, with emergent areas as shown in the middle Eocene palaeogeographic map (Fig. 22). The site of the Queensland Trough was probably relatively deeper than adjoining areas. Conditions were stable, emergent areas being reduced by erosion and troughs being filled with the erosion products. The continent extended farther north and east than at present, the continental margin probably lying somewhere near the central Coral Sea Basin.

The next major tectonic event was the opening of the Coral Sea Basin. This probably occurred about *late Palaeocene* time and was probably responsible for rifting of the northeastern part of the Tasman Geosyncline and creation of the present linear northwest trend of the outer edge of the Queensland Plateau.

Subsidence of the Plateau did not begin immediately in response to formation of the new basins as shallow-marine conditions prevailed until the end of the middle Eocene. By this time the Coral Sea Basin was wide, open, and mature, receiving detrital clay and pelagic material. The relation between Coral Sea Basin and Queensland Plateau evolution is further investigated in a later section.

The Queensland Plateau began to subside in *late middle or early upper Eocene*. Subsidence was not uniform, affecting the outer edge of the Plateau much more than the inner edge, and was an orogenic stage resulting in faulting of the shallow-water sediments. On the outer edge of the Plateau, at DSDP site 209, terrigenous input rapidly diminished (Fig. 10) during the late Eocene. This is interpreted as indicating the loss by submergence of the central emergent landmass in the area of the Diane Uplift. By the end of the late Eocene, terrigenous content is about equal to that of the Pleistocene level, suggesting that the central massif ceased acting as a significant source area.

Differential subsidence continued with diminished orogenic activity through the *late Eocene*. Reefs, which began to grow sporadically in the Early Cainozoic, flourished during this stage. Figure 29 shows the location of drowned reefs, the buried parts of presently active reefs, and areas of reef-derived deposits; the map is incomplete as the BMR survey ship avoided known reef areas.

From *late Eocene to late Oligocene* time a progressive change from differential to uniform subsidence occurred. It is suggested that the Queensland and Townsville Troughs formed at about this time or a little earlier. The Queensland Trough probably formed along reactivated geosynclinal structures, although the Townsville Trough cuts across the geosynclinal trend. A possible reason for this is put forward in the next section.

With the bounding trough system established, the Queensland Plateau was able to subside more evenly. Uniform subsidence began about *early to middle Oligocene* time as the Plateau seems to have experienced uniform subsidence since the late Oligocene.

A stage of fairly rapid subsidence may have followed formation of the troughs and this could have been responsible for drowning the cluster of buried reefs in the southwest corner of the Queensland Plateau (Fig. 29). A shift in the centres of reef growth probably resulted, and the only surviving reefs are found on elevated areas of basement. That is, there is probably a progression in age of the reefs from oldest in the depressed areas of basement, to youngest in the elevated areas. The older buried reefs are probably those in the head of the Queensland Trough and they could be as old as uppermost Cretaceous, while the youngest would be at the base of Diane Bank and Willis Islets for example, and could be as young as late Oligocene.

Since the late Oligocene, or essentially during the *Neogene*, uniform subsidence lowered the Plateau surface, and pelagic sedimentation formed a blanket cover. Reef growth continued on the more elevated areas of the basement. By this time the subsidence was completely epeirogenic, being unaccompanied by faulting or folding. Regionally positive free-air anomalies on the Queensland Plateau indicate that it is out of isostatic equilibrium. The Plateau would need to subside as a complete crustal unit by about 300 m to achieve equilibrium. Epeirogenetic subsidence is probably still in progress on the Queensland Plateau,

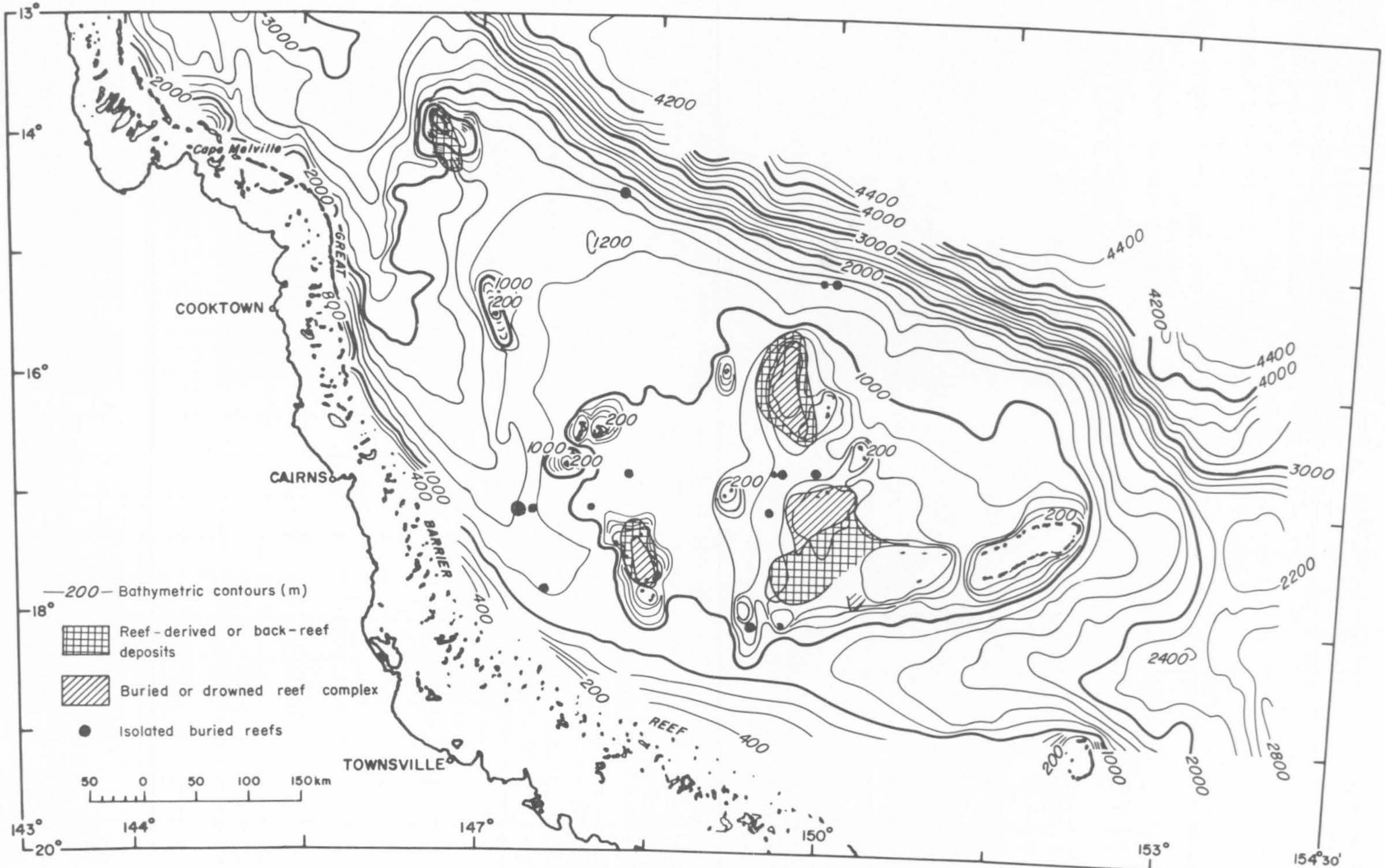


Fig. 29. Distribution of buried and drowned reefs.

causing the structure to tend toward the lower-energy state of isostatic equilibrium.

The most recent detectable phase in the geological development of the Plateau is canyoning of the surface. Correlation with results from DSDP hole 210 in the Coral Sea Basin suggests that this process may have begun about *middle Miocene* time (Burns, Andrews, et al., 1973). The later middle Miocene to middle Pliocene hiatus in Queensland Plateau sedimentation reported from DSDP hole 209 may have resulted from erosion caused by mass transport or slumping down the edge of the Plateau. The bathymetric profiles of Figure 12 show extensive canyoning of the outer margin of the Plateau, and seismic reflection records reveal a thick folded sequence of sediments at the base of the outer continental slope where it meets the Coral Sea Basin. Much of the thick sequence appears to have been derived from slumping of Queensland Plateau sediments.

EVOLUTION OF THE QUEENSLAND PLATEAU, CORAL SEA BASIN, AND NEW GUINEA

The two DSDP holes drilled in the Coral Sea Basin (210 and 287) have provided enough data to draw a fairly complete history of the evolution of the Basin. Although a single deep-sea drill hole provides data at one point, the gradual, predictable nature of lithological changes throughout an ocean basin makes it possible to deduce the history of the whole basin from one drilling. The predictability is an outcome of plate tectonic theory which postulates that ocean basins are formed by the creation of crust along a mid-ocean ridge. That is, the crust becomes progressively older away from the centre of an ocean basin. Western Pacific marginal basins apparently do not form by the same process but result from accretion of material in an extensional environment behind an active island arc/trench system (Karig, 1971a). This scheme of basin evolution is also regular and predictable although less so than the mid-ocean ridge model.

The two Coral Sea Basin holes lie close to the central axis of the Basin and were, therefore, probably drilled in some of the youngest crust in the Basin. Seismic reflection sections from the Scripp's Institution NOVA expedition show that the Basin is fairly uniform throughout in sediment thickness and structure. No central ridge is detectable, so the mid-ocean ridge-spreading model may not be applicable here. Nevertheless the DSDP holes probably give a good account of most of the Basin history.

Systematic geological mapping of New Guinea has been carried out by BMR since the 1950s. The coverage is now nearly complete, and detailed geological maps at 1:250 000 scale are available for most areas. Many authors have reviewed the geology and tectonics of the region (for example, Thompson, 1967; Australasian Petroleum Company, 1961; Rickwood, 1968; Pitt, 1966; Davies & Smith, 1970; and Bain, 1973). Attempts have been made to apply plate tectonics to the region (Davies & Smith, 1970; Johnson & Molner, 1972; Curtis, 1972). In general some difficulty has been found in applying modern theories to New Guinea; the existence of at least three sub-plates has been invoked by Johnson & Molner, requiring the presence of a quadruple junction near the eastern end of New Guinea. Karig's (1972) analysis using 'remnant arcs' is complicated and tenuous. Hence, although the geology is now well known, it has yet to fit the plate tectonic model in a convincing way.

If the tectonic units defined earlier form the elements of a Western Pacific marginal basin then facets of their geological evolution should be interlinked. The opening of the Coral Sea Basin is an event which should have left its signature on both the Queensland Plateau and the New Guinea region. Some relations are suggested.

Coral Sea Basin

The age of the oceanic second layer intersected in DSDP hole 287 was given as early Eocene (Andrews et al., 1973). The age of the second layer is taken to be that of the sediments lying immediately above, a nannofossil chalk with interbedded chert. The second layer consisted of porphyritic basalt and basalt and was drilled to a depth of 15 m. Hole 287 was drilled on a basement high to ensure that the second layer was reached, so it could be argued that the geology at that point is anomalous and does not represent that of the surroundings. However, the section at 287, though compressed, is very similar to that at 210 which was not drilled on an anomaly. In view of this similarity, and in the absence of complete testing of the Basin, an early Eocene age for second-layer genesis at site 287 is accepted.

TABLE 2. MAIN TECTONIC AND DEPOSITIONAL PHASES IN THE CORAL SEA REGION

	Queensland Plateau	*Coral Sea Basin*	*New Guinea*
LATE MESOZOIC	Shallow sea with large low-relief landmass shedding terrigenous material to form sediments in troughs and submerged platform areas	Continental crust of Tasman Geosyncline and W palaeo-Pacific Ocean	In approximate area, N (or leading) edge of Indian plate being subducted along N-dipping subduction zone. Indian plate contains Australian continental block which is receiving shelf deposits and turbidites in a N and E marginal trough
PALAEO-CENE	Continuation of the above regimes	Middle or late Palaeocene—early stages of opening of Coral Sea Basin with resultant rifting off of continental crust on NE edge of Australian continent	Late Mesozoic conditions continued through Palaeocene
EOCENE	Latest to middle Eocene—subsidence of Plateau, affecting outer edge more than inner. Emergent areas gradually sinking below sea level. Reef growth begins. Latest Eocene — almost all emergent areas submerged	Early Eocene—basin open and mature, receiving detrital material and pelagic sedimentation; detritus probably from Australia	Deep-water environment; deposition of chert and fine-grained limestone in area now the N and E Aure Trough. Deep-water basalt extruded onto deep-sea floor in SE. That is, deep-water conditions formed to S and E of Mesozoic sialic rocks
OLIGOCENE	Late Eocene to late Oligocene — non-deposition. Progressive change from differential to uniform subsidence. Formation of Queensland and Townsville Troughs. Drowning of some early reefs	Late Eocene to late Oligocene — non-deposition. Folding of basin sediments, creating unconformity after restart of deposition	Early to middle Oligocene—metamorphism, folding, and thrust-faulting of Mesozoic sialic rocks. In SE Papua, accompanied by overthrusting of wedge of oceanic crust. No record of sedimentation
MIOCENE TO PRESENT	Continued subsidence. Middle Miocene — probable initiation of canyoning on outer edge of Plateau.	Middle Miocene — massive influx of turbidite deposits, graded silts and clays	Middle Miocene — uplift of main range of New Guinea with rapid sedimentation in Aure Trough. Clastic sedimentation in Cape Vogel Basin. New Britain Trench probably formed about middle to late Miocene

As mentioned earlier, the two drill sites probably lie in one of the youngest parts of the Coral Sea Basin as they are located close to its central axis. If it is assumed that crustal accretion occurred along the axis then the oldest crust will lie along the outer flanks of the Basin. If accretion occurred by injection from a 'thermal diapir' then the opening rate could have been as high as 10 cm/year (Karig, 1971b). However, a more normal Pacific accretion rate is 2-3 cm/year. The Coral Sea Basin is about 400 km wide between 2000-m isobaths at its widest point, so it would take 10 million years to open at a rate of 4 cm/year. Hence the initial rifting of the Basin could have occurred in middle or late Palaeocene time.

In Table 2 the main tectonic and depositional events for the Coral Sea Basin, Queensland Plateau, and New Guinea have been summarized for the purpose of making comparisons. The Queensland Plateau data are deduced from DSDP 209 and this study, Coral Sea Basin events from DSDP 210 and 287, and New Guinea events from many published writings and D. B. Dow (pers. comm., 1974).

By comparing the histories of Plateau and Basin, several salient points can be recognized:

(1) Before the Cainozoic neither feature existed as it is recognized today. At least half of the Coral Sea Basin was probably occupied by continental crust.

(2) The opening of the Coral Sea Basin probably took place between the late Palaeo-

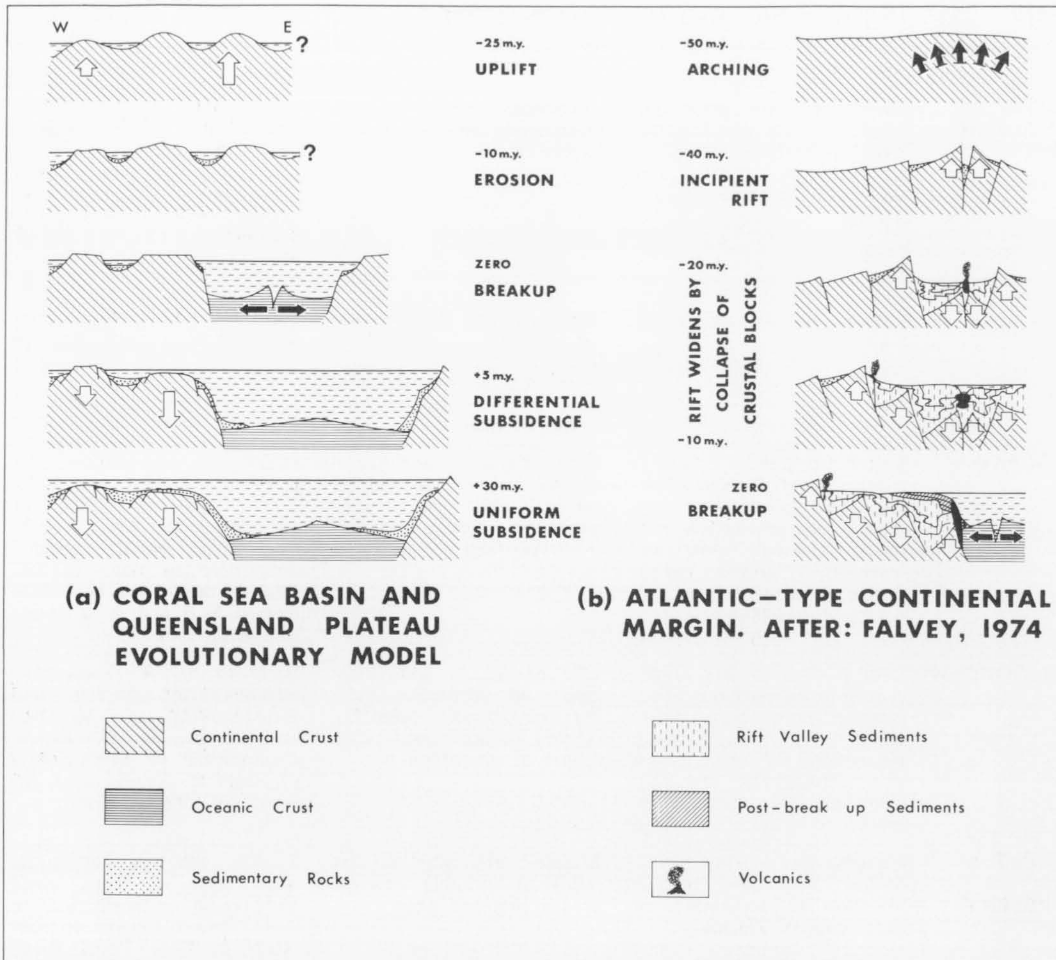


Fig. 30. Comparison between evolutionary models of Coral Sea and Atlantic-type continental margins.

cene and the early Eocene, before the subsidence of the Queensland Plateau.

(3) Subsidence of the Plateau began with orogenic differential downthrowing and passed into epeirogenic uniform subsidence after the Coral Sea Basin was completely open.

The lag between formation of the Basin and subsidence of the Plateau is probably the most important point to emerge. In the scheme put forward for development of an Atlantic-type continental margin by Falvey (1974), subsidence of continental blocks induced by deep crustal metamorphism begins up to 40 million years before break-up of the continents and formation of oceanic crust. In the development of the Queensland Plateau, subsidence did not occur until several million years after the break-up and ocean-forming stages were complete.

In Figure 30 the evolution of the Coral Sea Basin and Queensland Plateau are shown diagramatically together with the scheme for Atlantic-type continental margins taken from Falvey (1974). The main difference between the two models is the relation between, and duration of, the uplift and subsidence stages. In Falvey's model, subsidence of large crustal blocks occurs about 10 million years after uplift (arching), but in the Coral Sea model the lag is about 30 million years. In Falvey's model, a protracted rift-valley stage lasted some 40 million years before break-up, but in the Coral Sea model no rift valley stage is encountered. Figure 31 is an elevation-versus-time graph showing the relation between the two models.

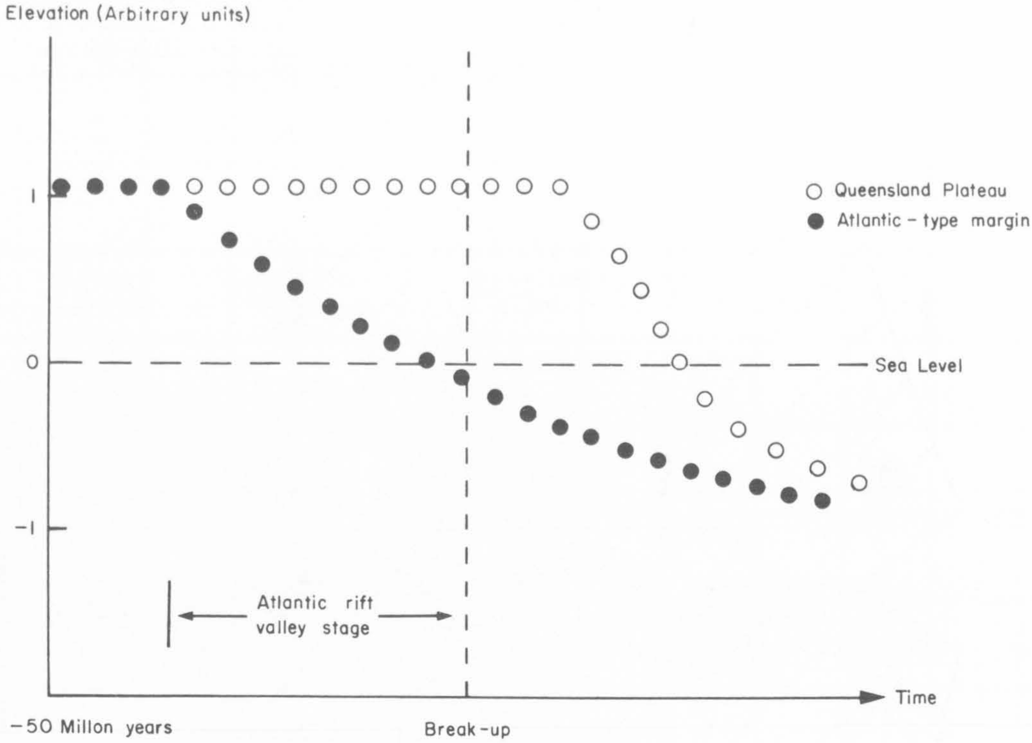


Fig. 31. Elevation-versus-time curves for Coral Sea and Atlantic-type margins.

The gross variation of crustal elevation with time over a period approaching 100 million years is probably directly related to the mechanism of margin formation. Such large-scale long-period variations are more likely to have their origin in deep crustal changes rather than in modifications to the surficial geology. Hence the observed differences in the elevation-versus-time curves for the Queensland Plateau and Atlantic margin may point to a different evolutionary mechanism in the two examples. As pointed out earlier, the Coral Sea Basin is a Western Pacific 'marginal' basin, i.e. it is not of the Atlantic type, at least under Karig's (1971b) classification. It is therefore possible that the elevation-versus-time behaviour observed for the Coral Sea Basin may be explained as the manifestation of a 'marginal' basin evolutionary scheme. However, in order to conform with Karig's model the Coral Sea Basin must have formed behind an island arc/trench system active during the Palaeocene and Eocene. If such a system existed at that time, then no evidence for it has been preserved. The expected indicators, such as andesitic volcanism, are absent in the rocks of southeast Papua and the Louisiade Archipelago. The absence of evidence is taken here as indicating that no such island arc system existed; the Coral Sea Basin must then be of a modified Atlantic type.

Some reasons for the departure from Atlantic-type behaviour will now be explored. In Falvey's (1974) model the effects of uplift (arching) and subsidence of large crustal blocks are induced by the same phenomenon. A thermal anomaly which causes heating and expansion of the lithosphere resulting in uplift of the crust also causes thermal metamorphism of the crust after a time lag of 10-15 million years. This increases deep crustal densities and thereby produces subsidence. This thermal anomaly later becomes the site of magma injection and the generation of oceanic crust. An uplift stage is present in the history of the Queensland Plateau; it is reasonable to assume that this was caused by a thermal anomaly, but it does not necessarily follow that that anomaly was at all times centred in the Coral Sea Basin. Uplift could be caused by an anomaly located a considerable distance from the centre of the Basin. If this were so then it may

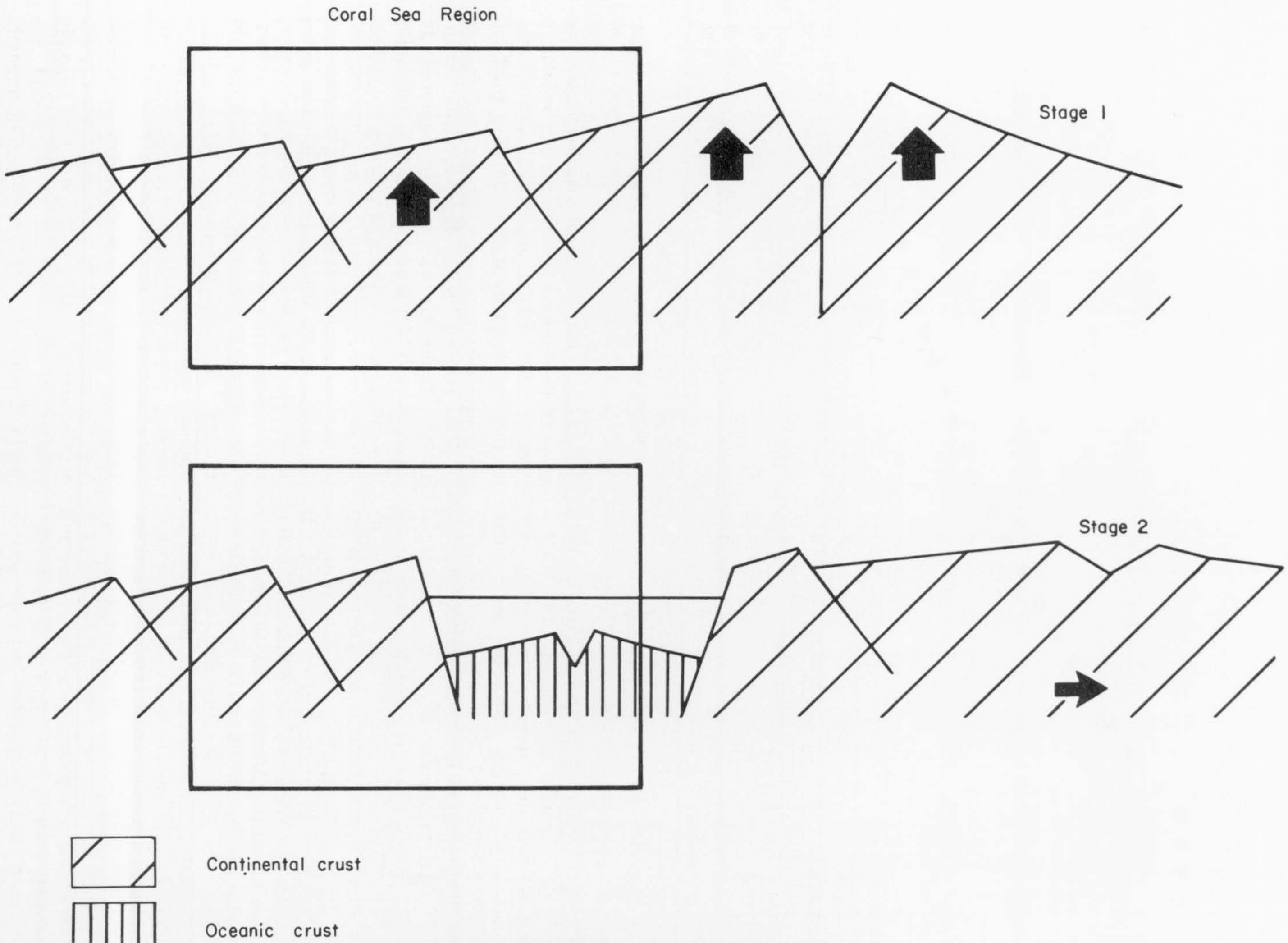


Fig. 32. Modified Atlantic margin development.

explain why subsidence did not follow uplift in the expected short period of 10 million years. A thermal anomaly located some distance east of the Coral Sea Basin could conceivably produce uplift of the Queensland Plateau, but as its intensity would be diminished with distance from the source it may not have produced a thermal metamorphic reaction and consequent subsidence. Falvey (1974, fig. 7) showed uplift occurring about 150 km away from the centre of the thermal anomaly and metamorphosed zone in his 'hogback uplift' stage of margin development. Packam (1973) postulated a spreading centre active in the Late Cretaceous (the time of uplift on the Queensland Plateau) and considerably east of the Coral Sea Basin. Such a centre of thermal activity would be adequate to produce the observed amount of uplift, yet would probably not produce thermal metamorphism in the region of the Queensland Plateau.

The uplift phase is shown diagramatically as stage 1 in Figure 32 where the Coral Sea region is shown on the flank of a normally developing Atlantic system. This is about equivalent to the 'incipient rift' (—40 million years) of Falvey's (1974) model in Figure 30. Subsidence of the Plateau followed the opening of the Coral Sea Basin and it is reasonable to assume that a causative relation existed. Subsidence was presumably brought about by rifting and creation of oceanic crust and had associated with it a second thermal anomaly. This anomaly would be centred along the axis of the Coral Sea Basin and may have been an offshoot or relocated younger equivalent of the Late Cretaceous anomaly. The absence of a rift valley stage in formation of the Basin simply means that the process occurred very quickly. Instead of the 50 million years which Falvey (1974) indicated, the entire period of basin formation probably took place in less than 5 million years. In such a short span the sedimentary facies which identify the rift-valley stage would not develop or would be so lacking in extent that they have not been recognized.

Formation of the Coral Sea Basin is shown as stage 2 in the scheme shown in Figure 32. The incipient rift in stage 1 died and was replaced by a secondary rift formed on the flank of the old system. The crust, weakened by uplift and fracturing in stage 1, allowed the secondary rift in stage 2 to occur very quickly. The now-extinct rift system would migrate away from the Coral Sea region; to the right in the diagram, the left side being taken as fixed. Although the period of formation was probably very short, the Coral Sea thermal anomaly could still have metamorphosed a thin strip of the deep crust along the outer edge of the Queensland Plateau. The increase in crustal density would give rise to loading of the outer edge of the Plateau and this could produce the initial differential orogenic stage of subsidence. Unequal loading of the Plateau would set up a stress regime which would tend to split off a section of continental crust equal in length to that of the metamorphosed zone. The length of this supposed zone cannot be determined but probably spanned at least the length of the Queensland Plateau. If the stress caused by unequal loading of the crust were complemented by cooling of the lithosphere in the Coral Sea Basin after cessation of spreading, then the result could be the splitting off of a slice of crust along faults in the Queensland and Townsville Troughs. The vertical loading of the crust caused by the metamorphosed zone, which would extend between the outlets of these two troughs, would give rise to vertical displacement, and lithospheric cooling in the Coral Sea Basin would produce a tensional system resulting in graben faulting along the fault zone. Hence it is suggested that formation of the transcurrent Townsville Trough represents the response to a stress regime set up by thermal metamorphism of the base of the crust along the outer edge of the Queensland Plateau combined with lithospheric cooling in the Coral Sea Basin. A diagrammatic cross-section illustrating the stresses affecting the continental margin off Queensland after formation of the Coral Sea Basin is shown in Figure 33.

After release of the stress on the Plateau, and after the graben troughs formed, differential subsidence would pass into uniform subsidence on the Queensland Plateau.

New Guinea

The relation between the evolution of the Queensland Plateau and the New Guinea region is indirect, being linked through the history of the Coral Sea Basin. It would be beyond the scope of this paper to attempt an analysis of the tectonic evolution of New Guinea but some comments can be made in the light of the recent results.

From the table of major tectonic and depositional events (Table 2) for the Coral Sea region some pertinent relations emerge:

(1) Before the Eocene, deposition in the New Guinea region consisted of turbiditic sedimentation in a trough which wrapped

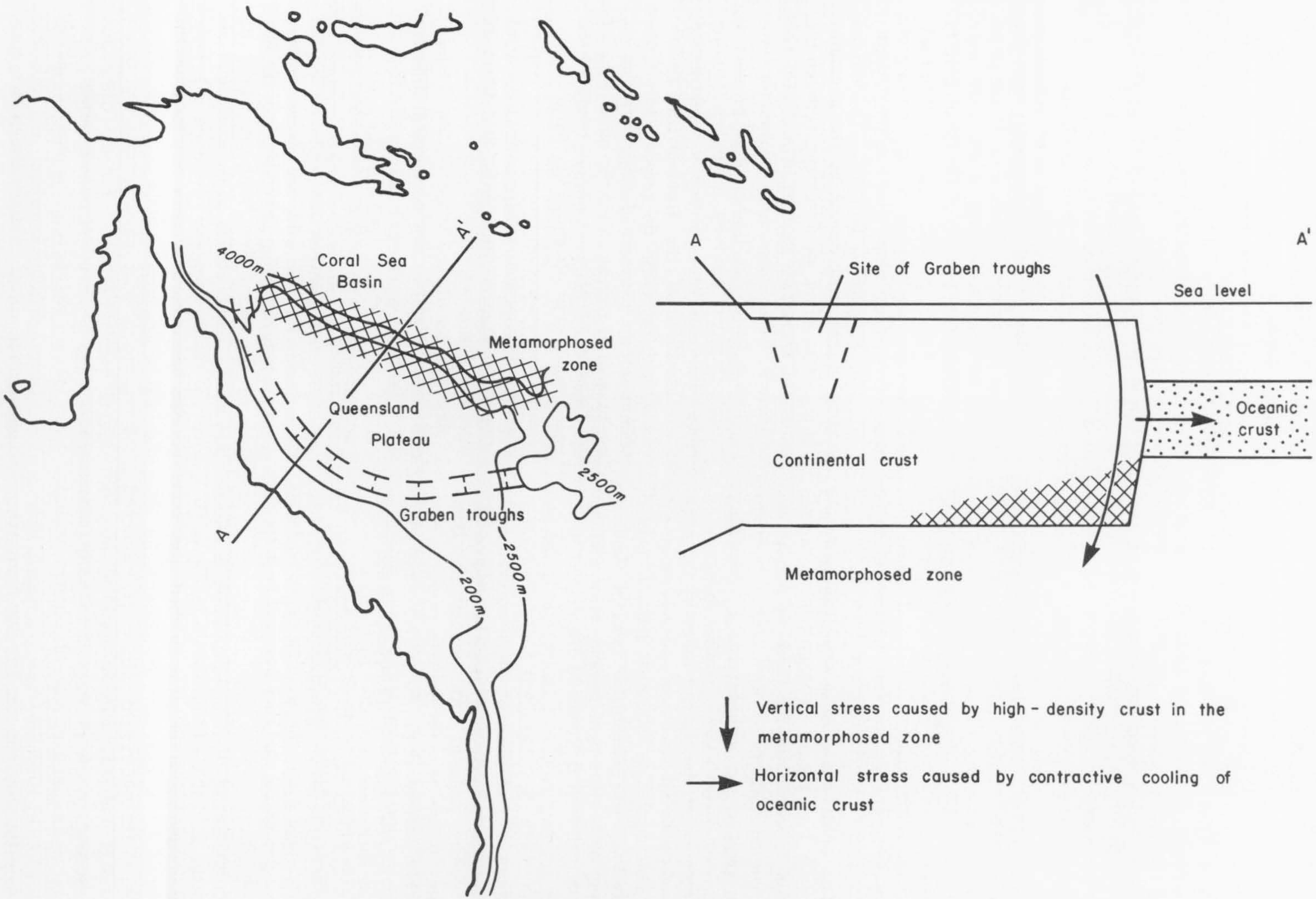


Fig. 33. Stress system and distribution of the metamorphosed zone.

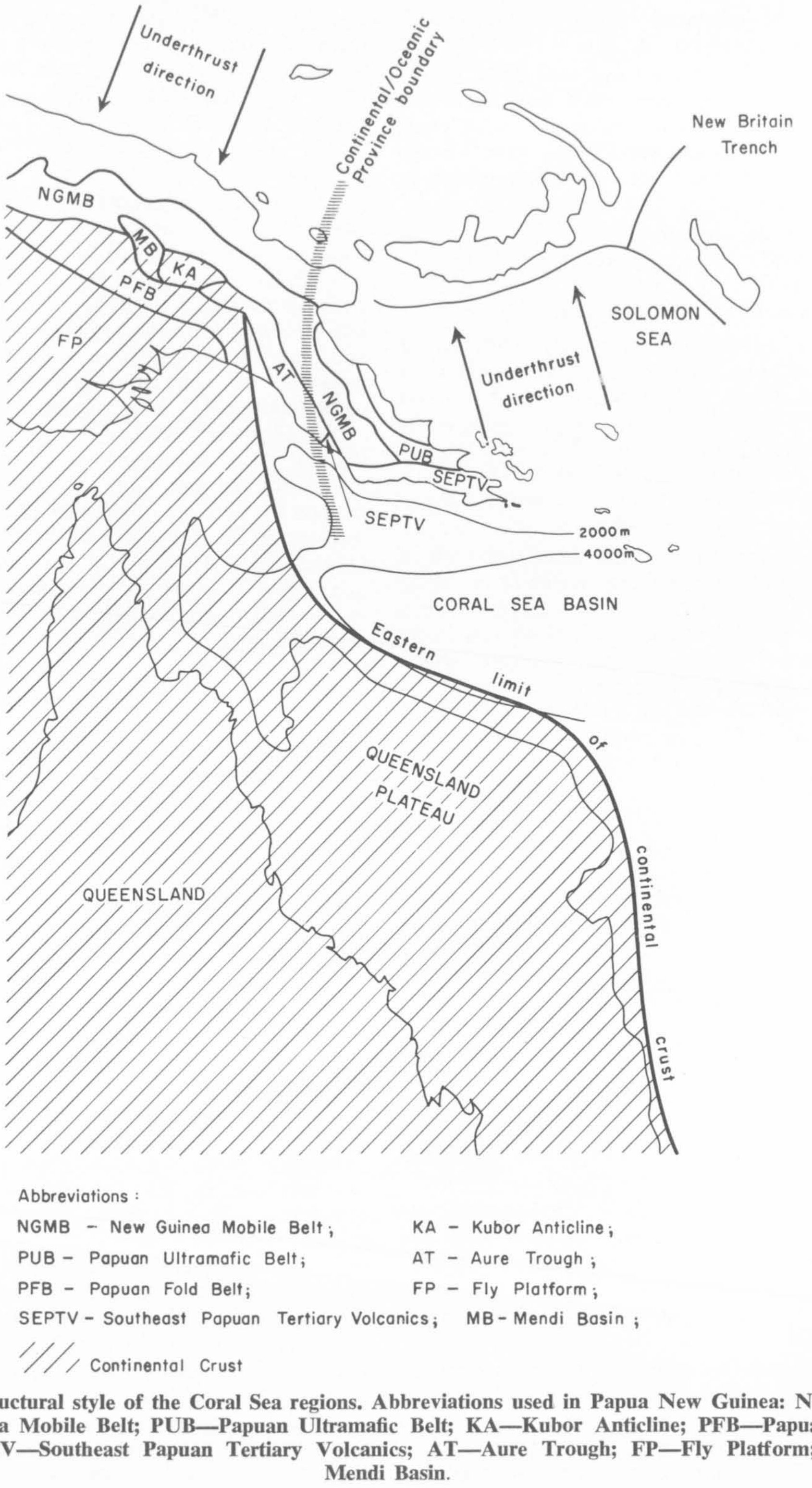


Fig. 34. Structural style of the Coral Sea regions. Abbreviations used in Papua New Guinea: NGMB—New Guinea Mobile Belt; PUB—Papuan Ultramafic Belt; KA—Kubor Anticline; PFB—Papuan Fold Belt; SEPTV—Southeast Papuan Tertiary Volcanics; AT—Aure Trough; FP—Fly Platform; MB—Mendi Basin.

around the northeast corner of the ancient Australian continent.

(2) Opening of the Coral Sea Basin is correlatable with formation of a deep-water environment accompanied by deposition of chert and fine-grained limestone in the Aure Trough region, and extrusion of deep-water basalts in the southeast.

(3) Metamorphism of the sialic sediments occurred after the opening of the Coral Sea Basin, not before as previously suggested (Davies & Smith, 1970; Curtis, 1972).

Figure 34 shows a schematic rendering of the structural elements of New Guinea in their relations to the Coral Sea region based on Bain (1973). The region has been split along the indicated province boundary by reference to the regional structural framework. The dividing line is not meant as a structural boundary of the type used by Bain.

West of the dividing line the distribution of shallow and intermediate earthquakes along a south-dipping zone (Denham, 1969) suggests that the crust north of New Britain is being underthrust. South of the seismicity zone, crustal shortening has taken place by intense folding and thrust-faulting (Bain, 1973). The approximate limit of continental crust is also shown, indicating that in the western province New Guinea lies on the northern edge of the Australian continental mass.

In the eastern province the Benioff zone dips north under New Guinea (Denham, 1969). Between the seismicity zone and southeast Papua lies the Solomon Sea, a possible subplate of dominantly oceanic crust. In southeast Papua crustal shortening has been accomplished by the massive overthrust of a wedge of oceanic crust (Davies & Smith, 1970). This is manifestly different from the method by which shortening was achieved in the western province. South of the overthrust zone deep-water sediments and oceanic basalt have been uplifted and exposed north of the Coral Sea Basin.

The eastern province is characterized by oceanic crust and tectonic features, while the western province has more affinity to continental rocks. The provinces will be referred to as the *Continental* and *Oceanic* provinces for convenience of description only.

The intense deformation and metamorphism in the New Guinea Mobile Belt, and the overthrust in southeast Papua is an early or middle Oligocene event (D. B. Dow, pers. comm., 1974). The major events which preceded and were presumably consequent on this orogeny were (1) movement of Australia north owing to opening of the ocean basin between Australia and Antarctica in the late early Eocene (Wiessel & Hayes, 1971), and (2) the opening of the Coral Sea Basin in late Palaeocene to early Eocene time.

Before the northward drift of Australia began, the northern boundary of the Indian plate probably lay north of New Guinea and was probably a subducting boundary (D. B. Dow, pers. comm., 1974). When northward movement of Australia began, continental crust and marginal sediments were moved into contact with the subduction zone, the buoyancy of the continental crust prevented its consumption, and severe deformation occurred by folding and thrust-faulting. When crustal shortening by folding and faulting could not accommodate the advance of the continent, an underthrust zone became active along the north New Guinea coast. This zone is active today and is presently taking up the northward movement of Australian continental crust.

Such a scheme, although it is probably oversimplified and does not describe all aspects of New Guinea geology, is probably a reasonable model for the evolution of the Continental province. It does not describe the structures in the Oceanic province, where the leading edge of the plate had been modified by the opening of the Coral Sea Basin before the movement of the plate. The opening split off a slice of continental crust which migrated to a position close to that presently occupied by southeast Papua.

This is in agreement with Manwaring's (1974) palaeomagnetic data which suggest an anti-clockwise rotation of New Guinea during the Early Cainozoic. The continental crust would have been covered by thick Mesozoic sediments which had formed on the northern and eastern margins of the former continental block.

When northward motion of the plate began, the response in the Oceanic province was somewhat different from that in the Continental province. When the slice of continental crust entered the hypothetical subduction zone a more successful attempt at consuming continental crust took place. As the slice was essentially detached from the main body of the continent it may have been less resistant to subduction than the continental crust in the western province. The more successful attempt may be manifest in the overthrusting of oceanic crust or underthrusting of continental crust in the Papuan Ultramafic Belt (Fig. 34). When the buoyancy of the continental fragment eventually terminated the overthrusting the location of the zone of crustal shortening and of the Indian/Pacific plate boundary shifted to their present position at the New Britain Trench. Here subduction of oceanic crust of the Solomon Sea is taking up the northward movement of the Indian plate.

Deformation by overthrusting in southeast Papua had an affect on the crust of the Coral Sea Basin. During the Eocene/Oligocene hiatus, Coral Sea Basin sediments were folded with the resultant creation of an angular unconformity. It is also notable that the change from differential to uniform subsidence of the Queensland Plateau occurred at about this time and it is possible that the massive orogeny in southeast Papua could have given the trigger for this change.

CONCLUSIONS

The Queensland Plateau has formed by modification of continental crust by the action of oceanic tectonism. The continental margin of the palaeo-Australian continent was subject to a series of orogenic and epeirogenic cycles beginning in Late Cretaceous time and continuing to the present. The locus of the causative tectonism probably changed through time as did the nature of the tectonism.

In the Late Cretaceous the Plateau area was uplifted and an erosion cycle began. This event is probably correlatable with the Maryboroughan Orogeny. Shallow to moderate marine conditions prevailed in basement depressions formed during the main geosyncline building period in the Palaeozoic. The source of the uplift was probably thermal expansion of the lithosphere induced by a heat source associated with a spreading centre to the east of the area. This orogenic stage is recognizable by areas of planated basement surface on the plateaus, and shallow-marine deposition.

Fairly stable conditions existed until middle to late Palaeocene time when the northeast corner of the geosyncline was split off as a consequence of formation of the Coral Sea Basin. This event is recognizable in the linear rifted margin of the Queensland Plateau.

Not until the ocean-building cycle was complete did the Plateau begin to sink. This phase was orogenic initially and involved differential subsidence of the outer edge of the Queensland Plateau relative to the inner, then passed into an epeirogenic phase after the Queensland and Townsville Troughs formed, after which uniform subsidence of the Plateau took place. The change to uniform subsidence is time-coincident with massive orogenesis in southeast Papua in early to middle Oligocene time.

The Plateau probably sank in response to excess loading of the crust formed by thermal metamorphism which raised lower crustal densities, and the effects of thermal contraction of the lithosphere after cessation of the ocean-building process. The heat source which caused the initial uplift stage was probably insufficient to cause deep crustal metamorphism as no subsidence occurred until the Coral Sea Basin was fully open, some 40 million years later.

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