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CONTENTS

THE FORAMINIFERA AND STRATIGRAPHY OF THE CHIMBU LIMESTONE, NEW GUINEA

by J. H. C. BAIN and J. G. BINNEKAMP

SUMMARY

The stratigraphy and Foraminifera of the Chimbu Limestone are described. Foraminifera, common throughout the 300-m thick sequence, indicate ages ranging from Ta_3 (middle Eocene) to Tc (lower Oligocene). The faunal succession is discussed and the species figured and discussed. One new species, *Dictyoconus chimbuensis,* is described from the Eocene strata.

INTRODUCTION

The Chimbu Limestone is the basal unit of the Tertiary sequence in the south-western limb of the Yaveufa Syncline, central highlands of New Guinea. It forms a prominent strike ridge that extends 73 km southeastwards from near Kerowagi almost to the Asaro River north of Lufa.

From near Kerowagi to the Asaro River the Chimbu Limestone rests (with paraconformity?) on the upper Cretaceous *Chim Group*; in the Chimbu River gorge it is overthrust onto upper Jurassic *Maril Shale.* Lower Miocene (upper Te) clastic sediments and limestone (*Movi Beds*) overlie the limestone with slight disconformity and near Nambaiyufa they overlap the Chimbu Limestone cuesta. The first samples of Chimbu Limestone were collected in 1937 by N. H. Fisher, then Government Geologist in New Guinea, from the vicinity of Chimbu airstrip (Kundiawa).

I. Crespin examined the material and in 1938 recorded and figured the following Foraminifera, which she regarded as typically Eocene:

Valvulammina sp.
Rhapydionina sp.
Lacazina wichmanni
cf. *Chapmanina* sp.
Coskinolina sp.
Biplanispira mirabilis

A reconnaissance survey of the area was made by Noakes (1939), who measured the section along the Chimbu River. He recorded at least 600 ft (180 m) of Chimbu Limestone overlying a thick sequence of shales, sandstones, and mudstones. Some 1 700 m below the base of the Chimbu Limestone Noakes reported a limestone shown on his map as a lens about 300 m long and 50 m wide, which he believed to be at least very close to the horizon from which Fisher collected his samples. Edwards & Glaessner (1953) questioned whether this lens was in situ.

F. K. Rickwood (1955) recognized the limestone block near Chimbu airstrip as 'a residual rock-fall' from the Chimbu Limestone. Along the Chimbu River he measured 600 m of limestone which he mapped as Chimbu Limestone; he described the sequence from top to bottom as:

90 m of bluish grey to brownish cream, fairly pure limestone.
The fossils include *Miogypsina borneensis, Miogypsinoides* sp., *Cycloclypeus* sp., *Eulepidina* sp., *Heterostegina borneensis, Sporadotrema* sp., *Amphistegina* sp., *Globigerina* sp., derived *Discocyclina* sp., and *Lithothamnium* sp.
Age: undoubtedly e-stage, probably upper e-stage.

300 m of cream and purplish limestone containing *Nummulites intermedius, Heterostegina borneensis, Amphistegina* sp., *Operculina* sp., *Halimeda* sp., *Lithothamnium* sp., and bryozoan fragments.
Age: Oligocene.

60 m of rather fragmental cream sandy limestone containing *Lacazina wichmanni, Spiroloculina* sp., *Textularia* sp., *Halimeda* sp., and *Lithothamnium* sp.
Age: Eocene.

150 m of calcareous sandstones with occasional pebble beds and grey sandy limestone lenses containing *Alveolina* sp., *Quinqueloculina* sp., *Pyrgo* sp., *Triloculina* sp., *Rotalia* sp., and *Halimeda* sp.
Age: Eocene.

In 1968 a field party of the Bureau of Mineral Resources visited the area again (Bain, Mackenzie, & Ryburn, 1970) and numerous spot samples were collected. After examination it appeared that samples from what was then called Chimbu Limestone included samples of Ta-b-c (Eocene-Oligocene) and upper Te (Miocene) age. They showed that elsewhere in the Yaveufa Syncline the Miocene limestone occurs only in lenses in fine-grained clastics (Movi Beds). In 1970 J. H. C. Bain and D. J. Belford measured and sampled the Chimbu Limestone in the section of the Chimbu River Gorge. They found that the Miocene limestones are separated from the Eocene-Oligocene limestones by mudstones; only near Nambaiyufa does the Miocene limestone rest on Eocene-Oligocene limestone.

Bain et al. (1970) therefore proposed that use of the name Chimbu Limestone be confined to the Eocene-Oligocene beds. J. G. Binnekamp examined the faunas and is responsible for the palaeontological part of the present study. All specimens figured are stored in the Commonwealth Palaeontological Collections, Canberra; all other thin sections prepared from Chimbu Limestone samples are kept in the micropalaeontological collections of the BMR, Canberra.

We gratefully acknowledge the assistance of D. E. Mackenzie and R. J. Ryburn, who collected specimens of the Chimbu Limestone in 1968, and also D. J. Belford, who measured and sampled the type section in 1970.

Distribution, Thickness, and Lithology

The Chimbu Limestone is exposed in the southwestern limb of the Yaveufa Syncline; it forms a prominent scarp extending from near Kerowagi southeastwards almost to the Asaro River north of Lufa. Near Kundiawa it has been strongly deformed by folding and faulting; several fault wedges of the limestone occur in the Chimbu River gorge. Apart from a few outcrops in the Goroka valley, the limestone does not appear in the northeastern limb of the Yaveufa Syncline.

Approximately 300 m of calcarenite and limestone have been measured in the Chimbu River gorge. This appears to be the average thickness of the formation, although in places it is much thinner and at Mount Elimbari it exceeds 1 000 m.

The formation consists entirely of massive limestone and calcarenite. Throughout the formation Foraminifera, Algae, and Bryozoa form a large part of the rock. The lowermost 72 m in the type section in the Chimbu River gorge is composed of dark grey, coarse-grained calcarenite and a finer-grained brownish grey to buff-coloured limestone, largely composed of algal debris and numerous

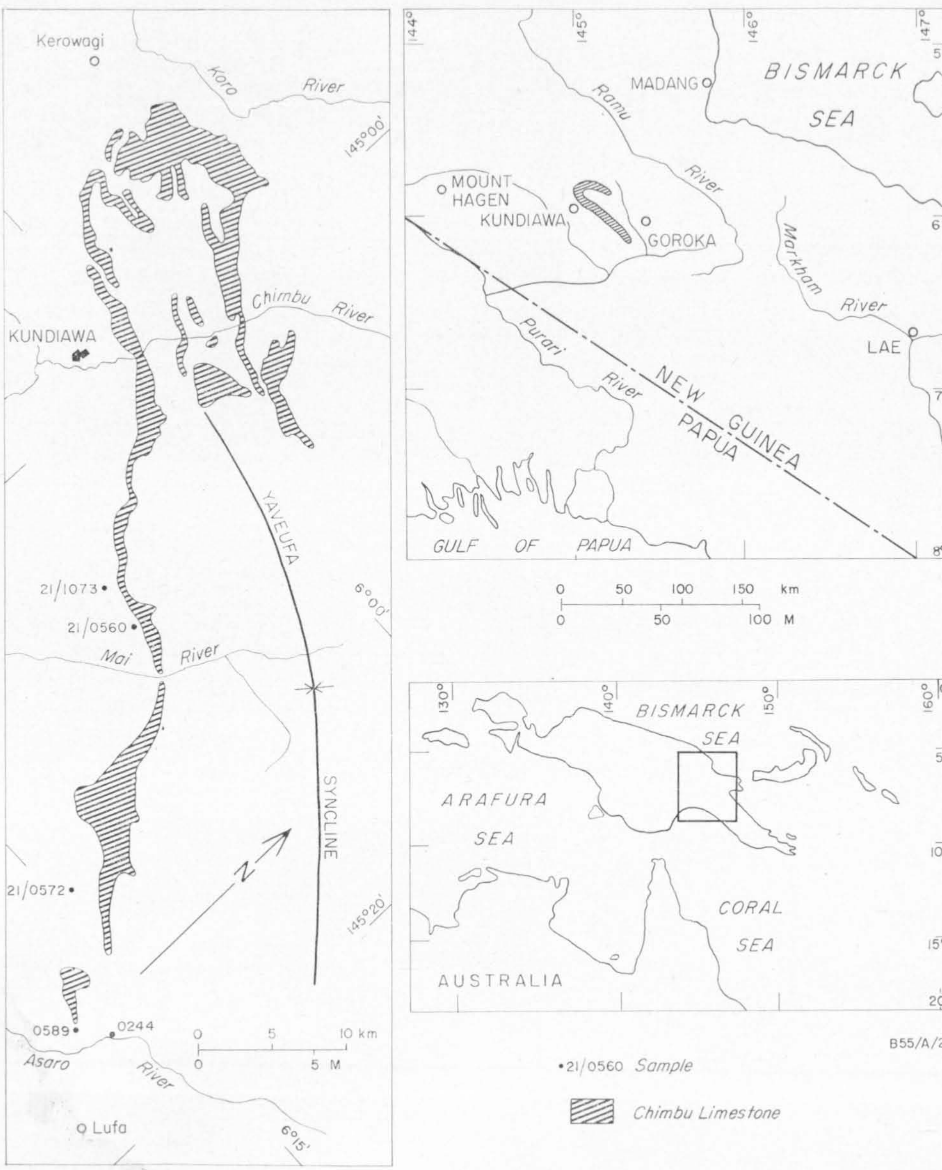

Fig. 1. Locality map.

Fasciolites and miliolids. Above these beds is 12 m of brown to buff-coloured algal calcarenite with abundant *Lacazinella.*

This is overlain by some 60 m of fine-grained light grey and buff-coloured algal limestone with few Foraminifera. This part of the section is poorly exposed and largely covered by rubble consisting of large boulders. The uppermost 130 m of the section is composed of light grey algal-foraminiferal and pure white limestones, formed almost entirely of *Nummulites,* heterosteginids, and operculinids, with a good deal of molluscan, echinoid, and algal debris. At many places in the buff-coloured Eocene beds (notably near Chuave) there is an abundance of well preserved, though difficult to collect, macrofossil material. This includes gastropods, belemnites, pelecypods, and echinoids. There is an unsubstantiated report that a complete large fish was collected from one of the quarries near Chuave.

Faunal Succession and Age

The only characteristic species of Foraminifera in sample 0090 from the base of the section is *Fasciolites* sp. cf. *F. elongata.* This species occurs throughout the lowermost 72 m (last occurrence in sample 0109). Some 14 m above the base (sample 0094) *Dictyoconus chimbuensis* appears and persists commonly for the next 60 m; its last occurrence is in sample 0110, 2 m above the last occurrence of *Fasciolites* sp. cf. *F. elongata.* Sample 0095, about 70 m up from 0094, yielded *Nummulites javanus. Lacazinella wichmanni* appears abruptly in sample 0110, just after the disappearance of *Fasciolites,* and occurs abundantly for some 12 m. *Fabiania cubensis* occurs sporadically throughout the lowermost 84 m of the succession.

All species occurring in this part of the formation are characteristic Eocene representatives. A marked change in the faunas occurs in the interval from 72 to 74 m above the base (samples 0109 and 0110); here *Fasciolites* and *Dictyoconus* disappear and *Lacazinella* appears. According to Adams (1965, p. 313) 'it is generally accepted that *Alveolina* s. str. (=*Fasciolites*) dies out in the Biarritzian (upper part of middle Eocene). *Alveolina* has never been described from beds of Tb age in Indonesia and the Pacific'. As there is little or no change in the lithology, it appears justifiable to assume that the disappearance of *Fasciolites* at this level of the succession reflects the extinction of the genus. Placing the boundary between Ta_3 and Tb at this level would restrict the occurrence of *Lacazinella wichmanni* to the Tb strata in this succession; however, in spot sample 0244 this species was tentatively identified in association with *Fasciolites.* The same association was recorded by Rutten (1936). Its range must therefore extend at least into the upper part of Ta_3.

Assuming that Tb begins after the disappearance of *Fasciolites,* this stage is represented by 12 m of limestone in the Chimbu gorge succession, and is characterized by *Lacazinella wichmanni* only. No *Pellatispira, Biplanispira,* or any species of *Nummulites* occurs. Above the *Lacazinella* beds an abrupt change in lithology coincides with the complete disappearance of Foraminifera. Such a sudden change strongly suggests a discontinuity, although no evidence of this was seen in the Chimbu gorge. Crespin (1938) recorded a fauna of *Lacazinella wichmanni* and *Biplanispira mirabilis* in samples of Chimbu Limestone from near Chimbu airstrip. If it is true that, as present evidence suggests, *Biplanispira* charact[illegible]s late Tb limestone and is absent from early Tb (Adams, 1970, p. 110), the beds of Chimbu Limestone near Chimbu airstrip might represent a younger horizon than the *Lacazinella* beds in the Chimbu River gorge.

Unfortunately such a sequence is not known from an undisturbed succession. Derived Eocene faunas, including *Lacazinella wichmanni,* cf. *Dictyoconus chim-*

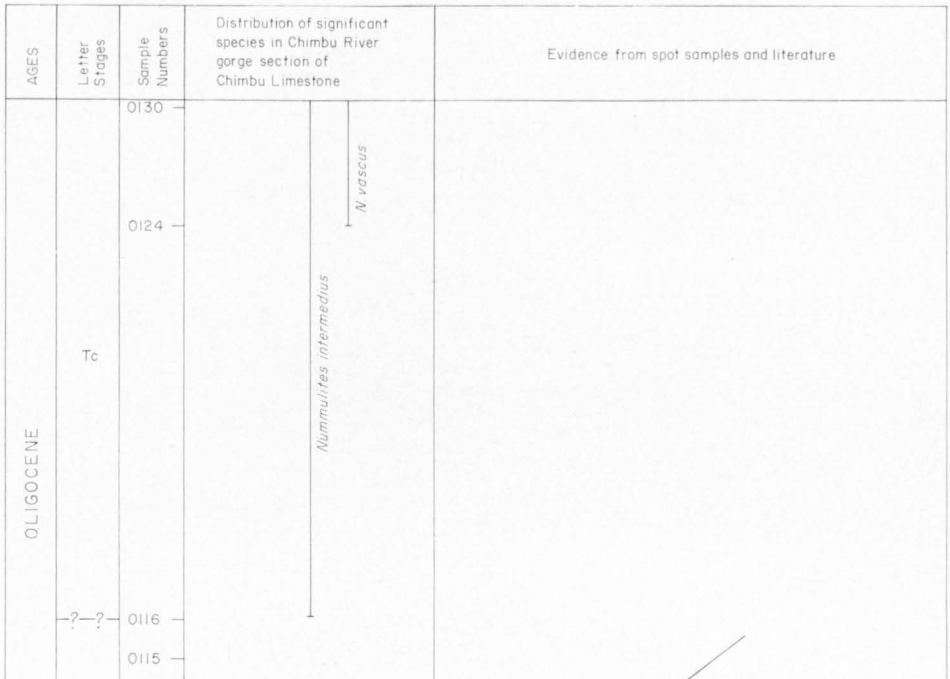

Fig. 2. Occurrence of Foraminifera in the Chimbu Limestone.

buensis, Discocyclina (s.l.) sp., *Pellatispira* sp., *Fabiania* sp., *Fasciolites* sp., and *Nummulites javanus,* are known from Td and Te strata outcropping southwest of the area presently occupied by Chimbu Limestone. Therefore there appears to be a strong possibility that a period of erosion took away part of the Tb strata of the Chimbu Limestone as originally deposited.

About 65 m above the last *Lacazinella* horizon the first *Nummulites intermedius* appears. From this level onwards this species occurs commonly to the top of the formation, some 130 m higher in the sequence. It is accompanied by unidentifiable heterosteginids and operculinids, often fragmented. In the top 30 m specimens identified as *N. vascus* are common. It is possible that, in the lower 100 m, earlier representatives of this group of closely related species are represented by small specimens, which are difficult to identify in the brecciated masses of foraminiferal tests. This part of the succession is entirely of Tc age.

NOTES ON FORAMINIFERA*

Order FORAMINIFERIDA
Suborder TEXTULARIINA
Superfamily LITUOLACEA
Family ORBITOLINIDAE
Genus DICTYOCONUS Blanckenhorn, 1900
DICTYOCONUS CHIMBUENSIS sp. nov.
(Pl. 4, figs 1-8)

Material: numerous specimens and fragments, mainly in random sections of limestone; some oriented sections prepared; numerous free specimens from one sample, many distorted.

Derivation of name: from the Chimbu River.

Holotype: specimen CPC 12087.

Type locality: Chimbu River gorge section of the Chimbu Limestone.

Level: middle and upper Eocene, Ta_3 and Tb in the East Indian letter classification.

Description: Free specimens are small, conical, diameter up to 2 mm, height up to 1.7 mm; diameter/height index about 1.2, base plano-convex, apex rounded, often broken, wall finely granular. Vertical sections show 6-12 marginal chambers undivided by horizontal partitions, others show partitions in some chambers, and in some specimens partitions occur in nearly all chambers.

Horizontal sections show a margin subdivided into chambers by vertical partitions forked at their inner ends; chambers are further subdivided by two sets of vertical plates.

Discussion: The vertical partitions of the marginal zone show the characteristic pattern of *Dictyoconus*. Other morphological characters, however, differ from those of many species of the genus, including its type species, particularly in the single, partly developed set of horizontal partitions and the apparently small size of the initial coil. Such differences make the generic assignment doubtful, but the author regards it as best placed in this genus for the time being (see also discussion in Adams (1965, pp. 335-337).

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* by J. G. Binnekamp.

D. melinauensis Adams, from Tc beds of Sarawak, Borneo, has identical structures but, although known only in random sections, it appears to have a lower conical form. The random section on which Cole (1958) based his new species *D. saipanensis* shows strong resemblance to some sections of *D. chimbuensis*. The characters of Cole's species, however, are insufficiently known to make any comparisons possible.

Suborder MILIOLINA
Superfamily MILIOLACEA
Family MILIOLIDAE
Subfamily FABULARIINAE
Genus LACAZINELLA Crespin, 1962
LACAZINELLA WICHMANNI (Schlumberger, 1894)
(Pl. 5, figs 1-2)

1962 *Lacazinella wichmanni* (Schlumberger); Crespin, p. 338, pl. 1, figs 1-9; pl. 2, figs 1-15; text-figs 2A, B.

Lacazinella wichmanni is known only from New Guinea and adjacent islands. Crespin gave a complete record to which little can be added.

In the Chimbu River gorge section it occurs for the first time in beds immediately above the last horizon with *Fasciolites* sp. cf. *F. elongata* and in association with *Fabiania* sp. cf. *F. cubensis* and *Dictyoconus chimbuensis*. A spot sample (21NG 2517) contains the same association with some rare specimens of *Halkyardia* sp. Crespin found it in association with *Biplanispira mirabilis* in the sample from near the Chimbu airstrip. Rutten (1936) recorded it together with *Fasciolites wichmanni* and *Nummulites bagelensis,* which is generally regarded as the megalospheric form of *Nummulites javanus*. In spot sample 21NG 0244 it was tentatively identified in association with *Fasciolites* sp., *N. javanus* (micro- and megalospheric forms), and *Discocyclina* (s.l.) spp.

Family ALVEOLINIDAE
Genus FASCIOLITES Parkinson, 1811
FASCIOLITES sp. cf. F. ELONGATA (d'Orbigny, 1828)
(Pl. 5, figs 3-5; Pl. 6, figs 1-4)

1960 *Alveolina elongata* d'Orbigny; Hottinger, p. 168, pl. 14, figs 10-14; text-figs 92, 93.
1962 *Alveolina* cf. *elongata* d'Orbigny; Adams, p. 52, pl. 3, figs 8-13.

Only megalospheric specimens were observed. They are elongate and rodlike; length/width ratio of 8 specimens varies from 3.3 to 5.4, the majority being between 4 and 5. The number of whorls seems fairly constant at 10 or 11. Proloculus is subspherical to somewhat elongate in the axial direction, and maximum size about 0.7 cm. All oriented sections prepared from samples from the bottom 72 m of the Chimbu Limestone in the Chimbu River gorge section contain only this species of *Fasciolites*.

This form compares well with the specimens described by Adams. Specimens figured by Hottinger are bigger and have a higher length to width ratio.

It would be worthwhile to revise all the occurrences of *Fasciolites* in the Indo-Pacific, attempting to apply Hottinger's elaborate classification of the genus. Most material of the Indo-Pacific has been insufficiently described and figured and many determinations appear to be based on not quite oriented sections. *F. wichmanni* Rutten is the most elongate species identified from this region. Both Bakx (1932) and Ritsema (1951) gave an average l/w ratio of about 3 for this species. Figures 26 and 27, Plate 4, by Bakx compare well with some specimens from the Chimbu Limestone. Rutten (1936) recorded l/w ratios between 3.5 and 5.7 and a maximum length of 0.8 cm for specimens identified as *F. wichmanni* from the island of Pisang.

In the Chimbu Limestone *Fasciolites* sp. cf. *F. elongata* occurs in the lower 72 m of the section in the Chimbu River gorge, together with *Nummulites javanus* and *Dictyoconus chimbuensis*. In a spot sample (21NG 0244) *Fasciolites* sp. was found in random sections with *N. javanus, Discocyclina* (incl. *Asterocyclina*) spp., and ?*Lacazinella wichmanni*.

Suborder ROTALIINA
Superfamily ROTALIACEA
Family NUMMULITIDAE
Subfamily NUMMULITINAE
Genus NUMMULITES Lamarck, 1801

NUMMULITES FICHTELI Michelotti, 1841
(Pl. 6, figs 5-6; Pl. 7, fig. 1)

1947 *Nummulites intermedius-fichteli* d'Archiac & Michelotti; Bursch, p. 19, pl. 1, figs 4-6, 26; pl. 2, figs 6, 7; pl. 5, fig. 5.
1965 *Nummulites fichteli* Michelotti (s.l.); Adams, p. 298, pl. 23, figs f-1.
1965 *Nummulites fichteli* Michelotti; Bozorgnia & Kalantari, p. 19, pl. 22, figs 7-11; pls 23, 24.

This widely distributed Tc and Td species occurs abundantly in the uppermost 130 m of the Chimbu River gorge section of the Chimbu Limestone, associated with *Nummulites* sp. cf. *N. vascus* and specifically indeterminable heterosteginids and operculinids.

NUMMULITES JAVANUS Verbeek, 1891
(Pl. 7, figs 2-5; Pl. 8, figs 1-3)

1891 *Nummulites javanus* vars α, β, γ, and δ, Verbeek, p. 105, figs 1-4 (microspheric form).
1891 *Nummulites bagelensis* I and II Verbeek, p. 107.
1896 *Nummulites javanus* Verbeek; Verbeek & Fennema, p. 1096, pl. 3, figs 45-57; pl. 4, figs 58-73, fig. 94.
1948 *Nummulites perforatus* (de Montfort); van Andel, p. 1014, figs 1-3.
1965 *Nummulites javanus* Verbeek; Adams, pl. 23, figs d, e.

Nummulites javanus, originally described from Java, is believed by some authors to be synonymous with *N. perforatus* from the upper part of the middle Eocene (Biarritzian) in Europe (see van Andel, 1948). Schaub (1963, pp. 291 and 294) described the European species as showing a typical threefold division of the equatorial section of the microspheric form: inner part a regularly expanding spiral with isometric chambers; median part with a wide spiral, thick and irregular marginal cord, and irregular chambers; and outer part, about half of the spiral, with a very narrow spiral and regular, low, and relatively long chambers. Neither the specimens figured from Indonesia and New Guinea nor the specimens from the Chimbu Limestone show such a division; the spiral expands regularly and the chambers are isometric.

In the Chimbu River gorge section this species occurs rarely in sample 20NG 0095, about 16 m above the base of the section, in association with *Fasciolites* sp. cf. *F. elongata*. In several spot samples (21NG 0244, 0572, and 0589) megalospheric and microspheric forms were found associated with *Discocyclina* (incl. *Asterocyclina*) spp., *Fasciolites* sp., and ?*Lacazinella wichmanni*.

NUMMULITES VASCUS Joly & Leymerie, 1848
(Pl. 8, fig. 4; Pl. 9, figs 1-2)

1961 *Nummulites vascus vascus* Joly & Leymerie; Montanari, p. 576, pl. 1, figs 2a-c, 6, 7a-b; pl. 2, fig. 14.
1962 *Nummulites vascus* Joly & Leymerie; Eames, Banner, Blow, & Clarke, pl. 1, figs A-B.
1965 *Nummulites vascus* Joly & Leymerie; Bozorgnia & Kalantari, p. 18, pl. 21, figs 1-10; pls 23, 24.

Nummulites vascus is a commonly reported species from the Middle East, where it occurs in strata of lower and middle Oligocene (Tc-d) age. Grimsdale (1952, p. 236) reported megalospheric forms assigned to the *Nummulites vascus* group, accompanied by two microspheric forms, in the Oligocene of Kirkuk: a larger and more compressed form referred to *N. vascus* s. str. and exceptionally stout and thick-walled variants referred to *N. vascus* var. *semiglobulus* (Doornink). Doornink's species was described from the upper Eocene (Tb) from Java and is regarded by Cole (1957) as a synonym of *N. pengaronensis* Verbeek, a common Tb form in the Indo-Pacific, once also reported in beds of Tc age in Borneo by van der Vlerk (1929).

N. pengaronensis appears to be more robust than *N. vascus* and also possesses an axial plug, which is not present in the forms figured by Bozorgnia & Kalantari and Montanari, but is very distinct in the specimen figured by Eames et al. Grimsdale's observations show that these differences could well be within the morphological variation of these forms.

Only megalospheric specimens were observed in the Chimbu Limestone. They attain bigger sizes than the specimens studied by Bozorgnia & Kalantari, but otherwise are identical. They are common in the top 30 m of the Chimbu River gorge section, where they are accompanied by *N. fichteli* and indeterminable heterosteginids and operculinids.

Superfamily ORBITOIDACEA
Family CYMBALOPORIDAE
Genus FABIANIA Silvestri, 1924
FABIANIA CUBENSIS (Cushman & Bermudez, 1936)
(Pl. 9, fig. 3; Pl. 10, figs 1-7)

1936 *Pseudoorbitolina cubensis* Cushman & Bermudez, p. 59, pl. 10, figs 27-30.
1953 *Fabiania saipanensis* Cole, p. 28, pl. 15, figs 1, 2.
1956 *Fabiania indica* Nagappa, p. 192, pl. 30, figs 1-9; pl. 31, figs 1-3.
1957 *Fabiania saipanensis* Cole; Cole, p. 767, pl. 245, figs 1, 2.
1958 *Fabiania saipanensis* Cole; Cole, p. 337, pl. 102, figs 7-9; pl. 118, fig. 8.

F. saipanensis, the commonly reported Tb species in the Indo-Pacific, is characterized by its irregular conical shape with a deeply excavated umbilicus. In 1957 Cole included in the synonymy of this species *F. indica* described from the Sylhet Limestone (Prang stage, middle Eocene, Central Assam, according to

Nagappa, 1959). Nagappa's illustrations of *F. indica* show a very variable form with specimens resembling both *F. saipanensis* and *F. cubensis.* Random sections of specimens from the Chimbu Limestone show a similar variation. They occur together with *Fasciolites* sp. cf. *F. elongata, Dictyoconus chimbuensis,* and *Lacazinella wichmanni,* in beds regarded as Ta_3 and Tb age.

List of samples

20NG 0090-0130	Limestones from section along Chimbu River gorge.
21NG 0560	Limestone from talus from bottom of scarp about 3 km northwest of Chuave Patrol Post.
21NG 0572	Limestone from talus from bottom of scarp about 6 km west of Nambaiyufa Mission along road to Chuave Patrol Post.
21NG 0589	Limestone from talus from bottom of scarp about 7 km south of Nambaiyufa Mission.
21NG 1073	Limestone about 5 km west-northwest of Chuave Patrol Post on the road to Kundiawa.
21NG 2517	Limestone from about 5 km east of Movi Mission.
21NG 0244	Limestone boulder in Asaro River 8 km south-southeast of Nambaiyufa Mission.

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Scarp of the Chimbu Limestone southeast of Chuave ; the highest point is Mt. Elimbari (2850 m above sea level).

PLATE 2

Lithology of Chimbu Limestone

Figures

1 Algal-foraminiferal calcarenite containing numerous *Fasciolites* sp. cf. *F. elongata.* x4. Sample 20NG 0094, slide 4 (CPC 12083).

2 Algal-foraminiferal calcarenite containing numerous *Lacazinella wichmanni.* x4. Sample 20NG 0112, slide 1 (CPC 12084).

Lithology of the Chimbu Limestone.

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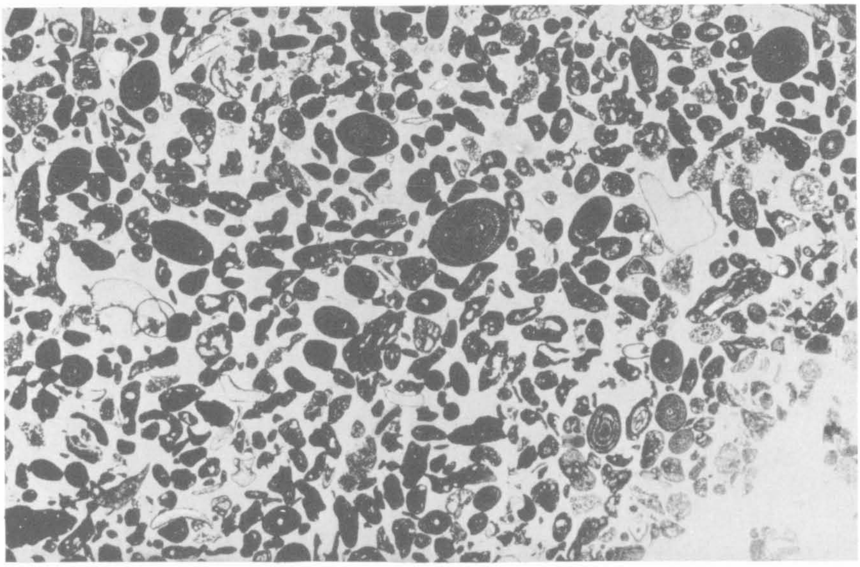

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PLATE 3

Lithology of Chimbu Limestone

Figures

1 Detrital algal calcarenite, x4. Sample 20NG 0113, slide 1 (CPC 12085).

2 Calcarenite largely composed of *Nummulites, Heterostegina,* and *Operculina,* x4. Sample 20NG 0127, slide 1 (CPC 12086).

Lithology of the Chimbu Limestone.

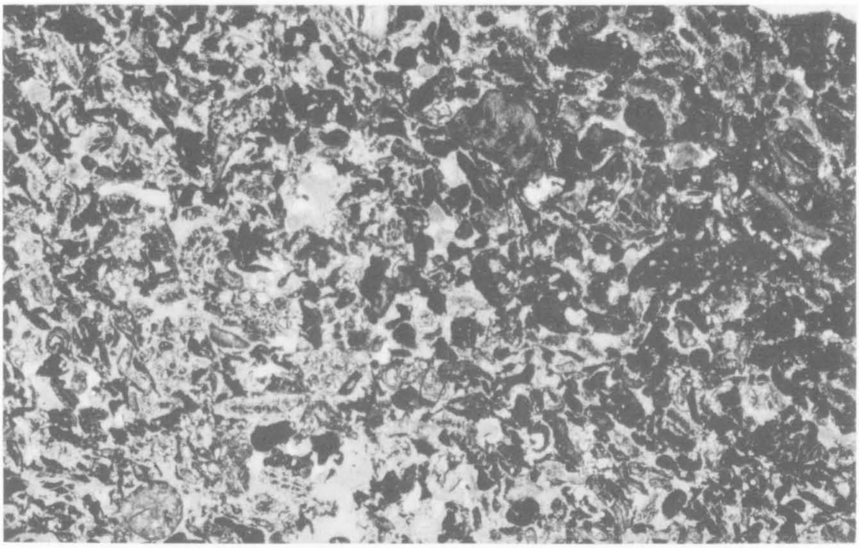

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PLATE 4

Dictyoconus chimbuensis sp. nov.

Figures

1 **Holotype,** vertical section x30. Some secondary horizontal partitions visible on right hand side. Sample 20NG 0098, slide 4 (CPC 12087).

2 Vertical section x30, slightly off centre; some secondary horizontal partitions visible on right hand side. Sample 21NG 0560a, slide 32 (CPC 12088).

3 Vertical section x30. This specimen shows no secondary horizontal partitions. Sample 21NG 0560a, slide 23 (CPC 12089).

4 Vertical section x30, slightly off centre; secondary horizontal partitions appear in all chambers. Sample 20NG 0100, slide 1 (CPC 12090).

5 Horizontal section x30. Sample 21NG 0560a, slide 16 (CPC 12091).

6 Detail x50 of specimen shown in fig. 5.

7 Part of horizontal section x30. Sample 21NG 2517, slide 4 (CPC 12092).

8 Detail x50 of specimen shown in fig. 7.

Dictyoconus chimbuensis sp. nov.

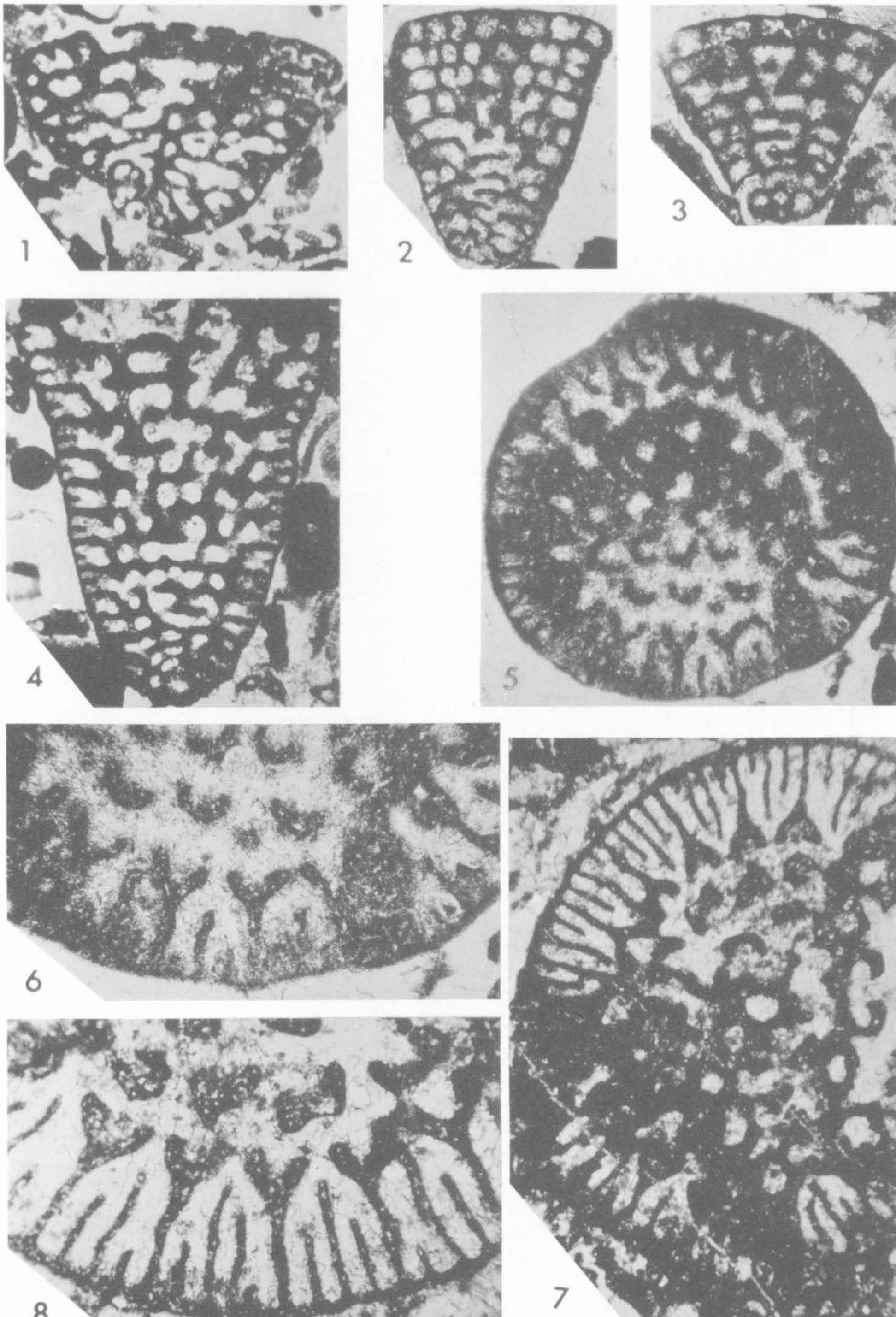


PLATE 5

Figures

1-2 *Lacazinella wichmanni* (Schlumberger)

1, horizontal section x30. Sample 20NG 0111, slide 1 (CPC 12093).

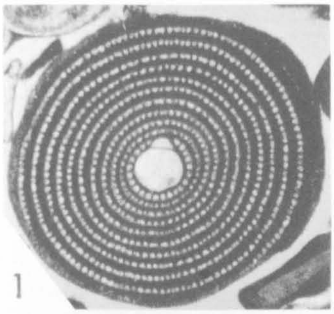
2, vertical section x30. Sample 20NG 0111, slide 2 (CPC 12094).

3-5 *Fasciolites* sp. cf. *F. elongata* (d'Orbigny)

3, axial section x20. Sample 20NG 0094, slide 1 (CPC 12095).

4, detail x50 of specimen shown in fig. 3.

5, axial section x20. Sample 20NG 0094, slide 2 (CPC 12096).



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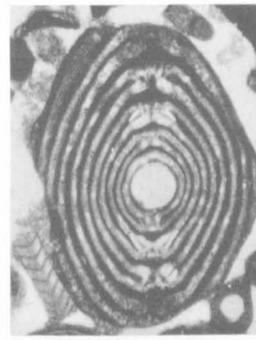

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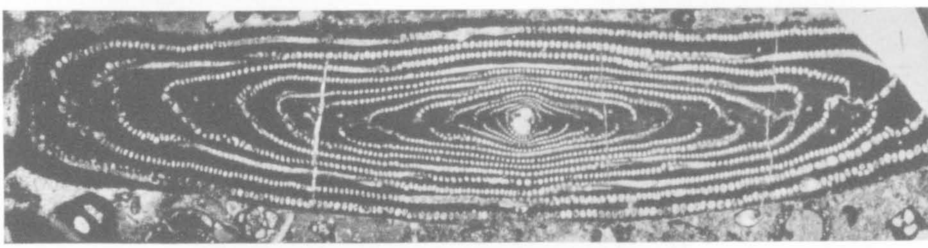

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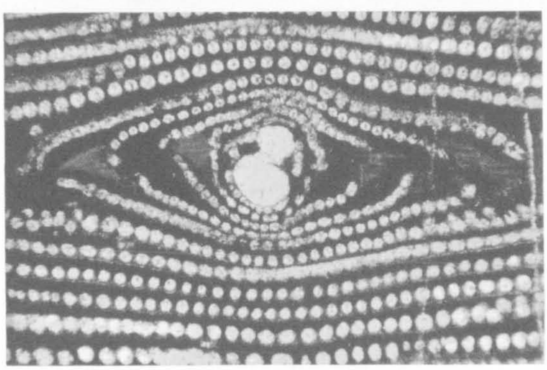

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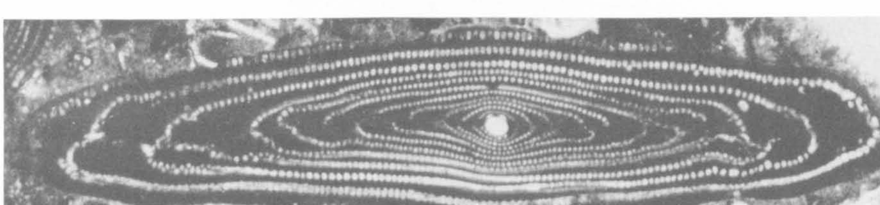

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PLATE 6

Figures

1-4 *Fasciolites* sp. cf. *F. elongata* (d'Orbigny)

1, detail x50 of specimen shown in Pl. 4, fig. 5.

2, equatorial section x30. Sample 21NG 0560a, slide 40 (CPC 12097).

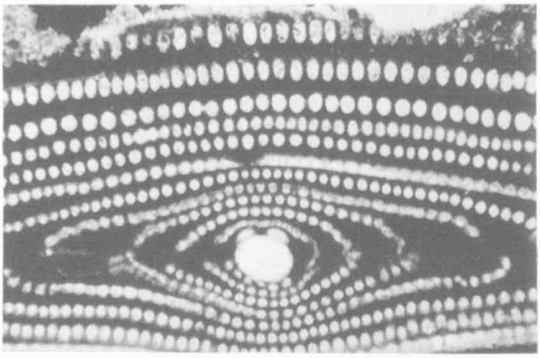
3, cross-section x30. Sample 20NG 0094, slide 4 (CPC 12098).

4, detail x70 of specimen shown in fig. 3.

5-6 *Nummulites fichteli* Michelotti

Vertical sections x20.

Sample 20NG 0127, slide 3 (CPC 12099) and slide 4 (CPC 12100).



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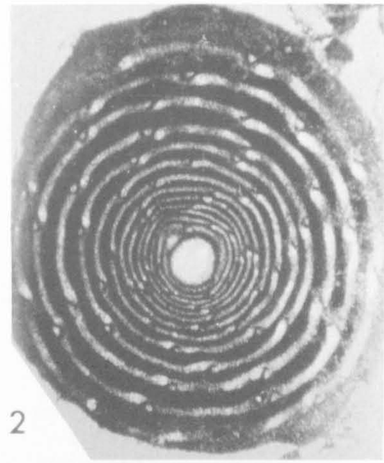

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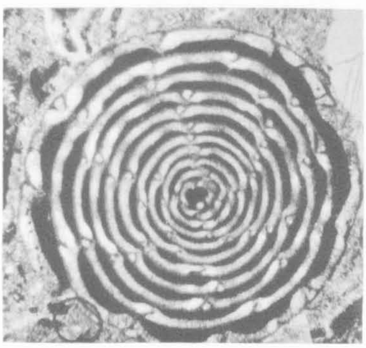

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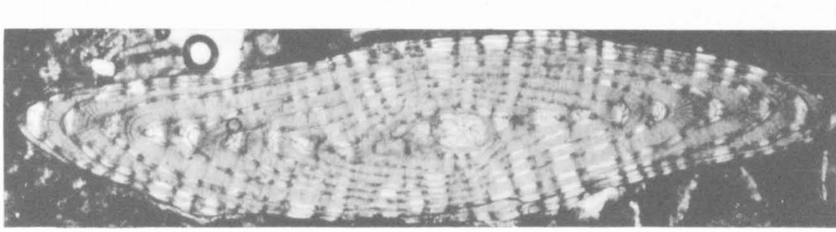

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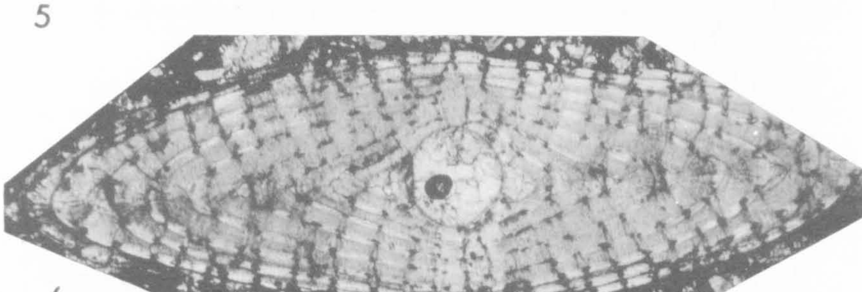

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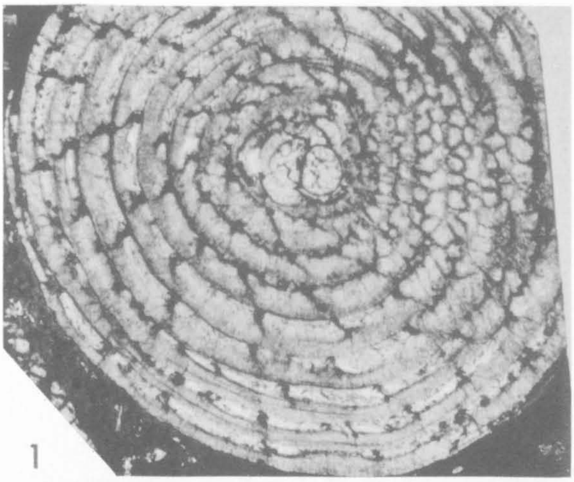
PLATE 7

Figures

1 *Nummulites fichteli* Michelotti
Horizontal section x20.
Sample 20NG 0127, slide 8 (CPC 12101).

2-5 *Nummulites javanus* Verbeek
Vertical sections x30 of megalospheric specimens.
2, sample 21NG 0244, slide 1 (CPC 12102).
3, sample 21NG 0572, slide 2 (CPC 12103).
4, sample 21 NG 0572, slide 2 (CPC 12104).
5, sample 21NG 0572, slide 2 (CPC 12105).



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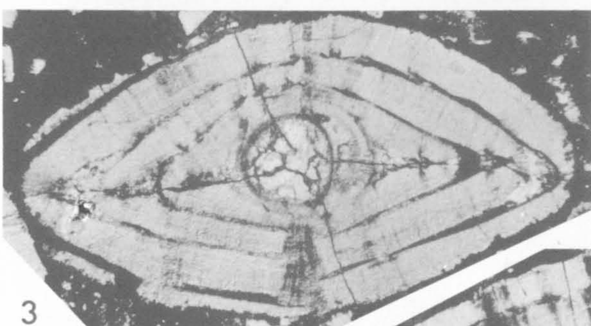

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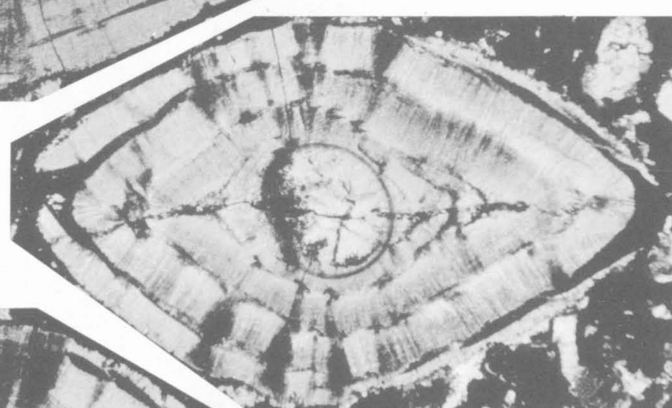

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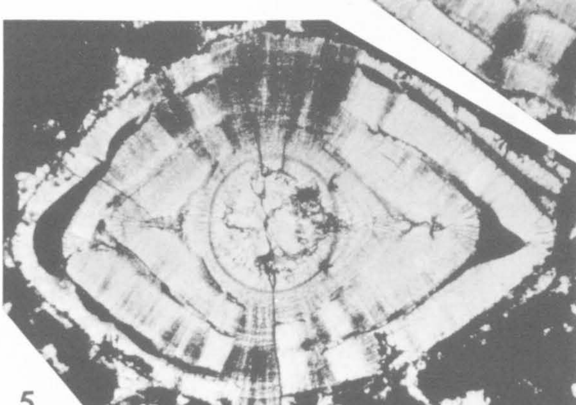

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PLATE 8

Figures

1-3 *Nummulites javanus* Verbeek

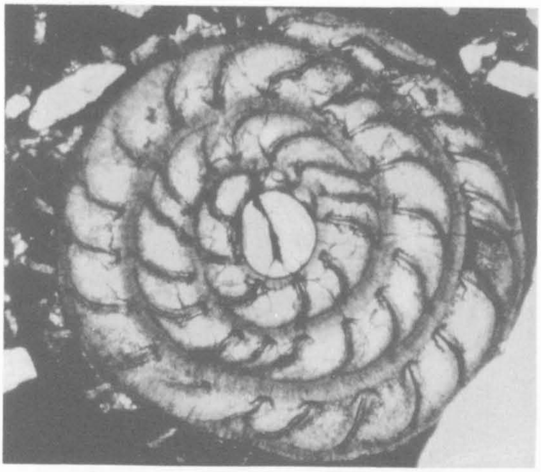
1, horizontal section x20 of megalospheric specimen.
Sample 21NG 0589a, slide 5 (CPC 12106).

2, vertical section x6 of microspheric specimen.
Sample 21NG 0572, slide 3 (CPC 12107).

3, part of horizontal section x8 of microspheric specimen.
Sample 21NG 0572, slide 6 (CPC 12108).

4 *Nummulites vascus* Joly & Leymerie

Vertical section x20.
Slightly off centre. Sample 20NG 0127, slide 1 (CPC 12109).



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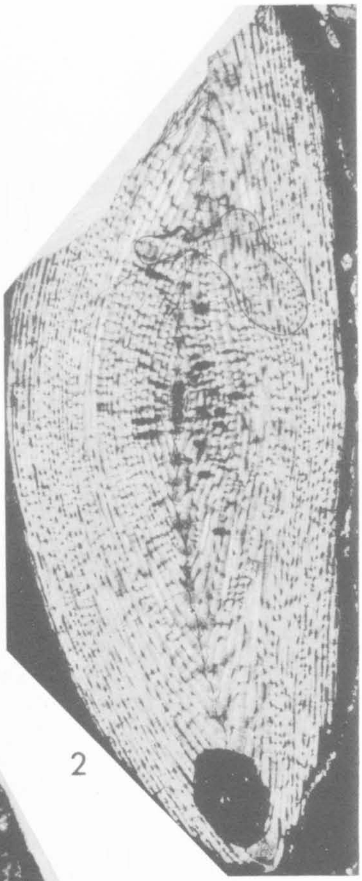

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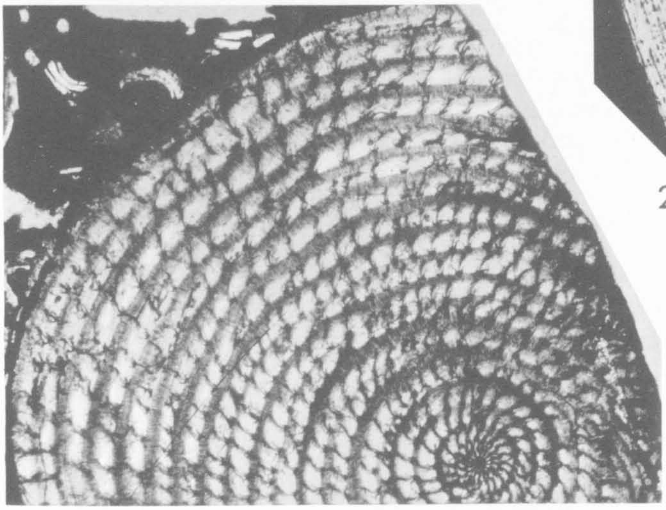

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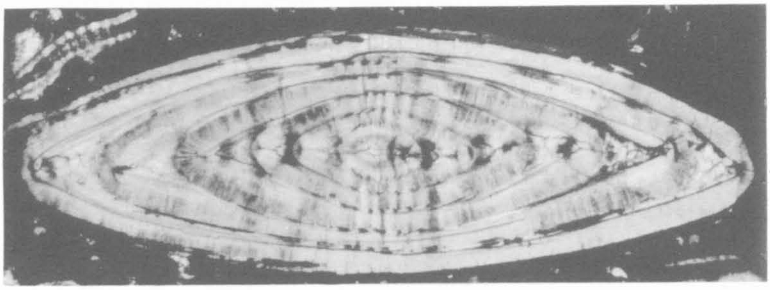

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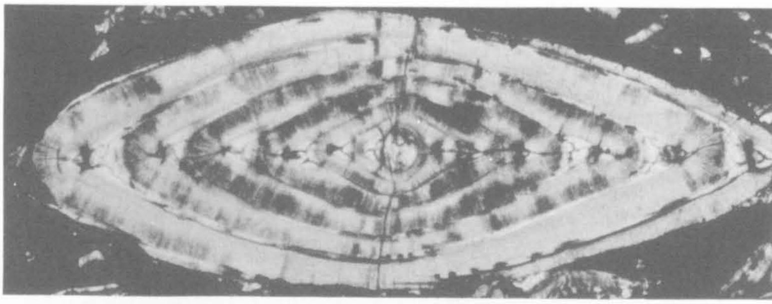
PLATE 9

Figures

1-2 *Nummulites vascus* Joly & Leymerie
1, vertical section x20. Sample 20NG 0127, slide 3 (CPC 12109a).
2, horizontal section x20. Sample 20 NG 0127, slide 7 (CPC 12110).

3 *Fabiania cubensis* (Cushman & Bermudez)
Random section x30. Sample 21NG 0560a, slide 5 (CPC 12111).



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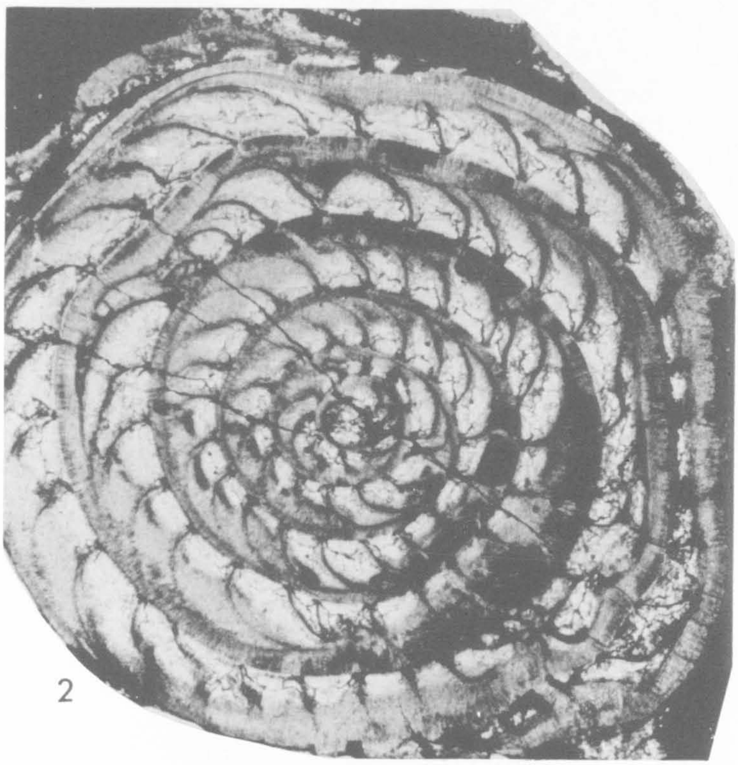

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PLATE 10

Fabiania cubensis (Cushman & Bermudez)

Figures

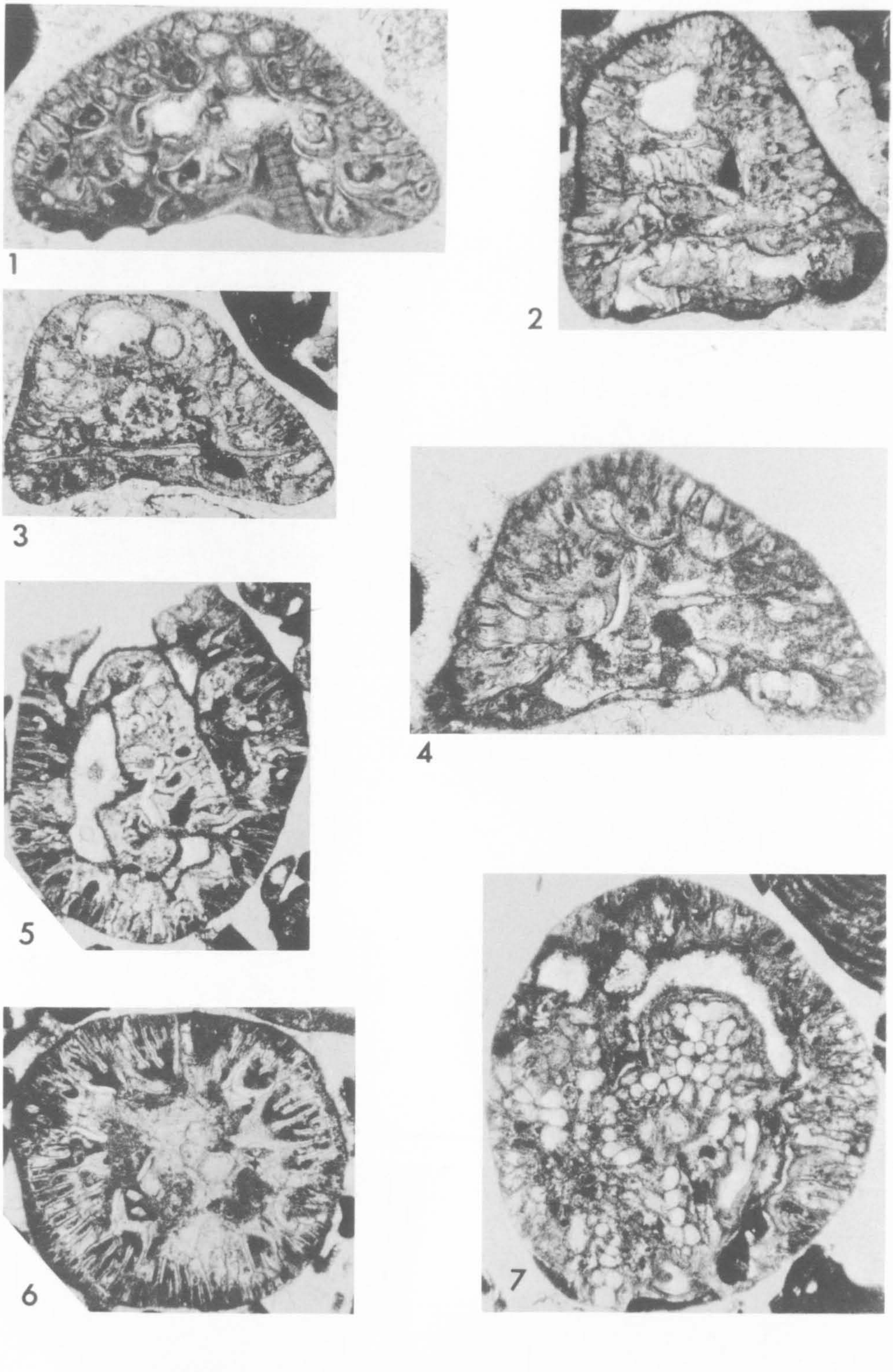
1-4 Random sections close to vertical plane, x30.
1, sample 21NG 1073a, slide 10 (CPC 12112).
2, sample 21NG 0560a, slide 27 (CPC 12113).
3, sample 21NG 1073a, slide 13 (CPC 12114).
4, sample 21NG 0560a, slide 4 (CPC 12115).

5 Random section x30. Sample 20NG 0111, slide 1 (CPC 12116).

6 Random section x30, close to horizontal plane.
Sample 20NG 0111, slide 1 (CPC 12117).

7 Random section x30 close to horizontal plane.
Sample 21NG 1073a, slide 12 (CPC 12118).

Fabiania cubensis



A REVIEW OF THE MIDDLE CAMBRIAN STRATIGRAPHY IN THE QUEENSLAND PORTION OF THE GEORGINA BASIN

by F. DE KEYSER

ABSTRACT

Recent detailed mapping of Cambrian sediments in the eastern part of the Georgina Basin indicates that the present scheme of stratigraphic nomenclature is unsatisfactory for the field geologist. This situation has arisen because of the mixture of biostratigraphic and lithostratigraphic units developed by earlier workers.

Use of the concept of magnafacies and parvafacies effectively clarifies the fundamental differences between the lithostratigraphic and the chronostratigraphic units. It is possible to recognize six distinctive mappable lithosomes which highlight the highly diachronous nature of the lithostratigraphic units.

INTRODUCTION

The Georgina Basin is a large Palaeozoic sedimentary basin extending from western Queensland into the Northern Territory, covering an area of approximately 300 000 sq km (Fig. 1). It contains sediments ranging in age from Cambrian to Devonian; Middle Cambrian sediments are particularly widespread.

The discovery of major phosphate deposits in the eastern part of the Georgina Basin, in 1966, prompted the Bureau of Mineral Resources to carry out a programme of detailed mapping of the Cambrian sediments, from 1967 to 1969. A number of geologists were involved in this project, including F. de Keyser (Party Leader), R. Thieme, J. H. Shergold, P. J. Cook, and C. G. Gatehouse (Bureau of Mineral Resources) and C. Murray (Geological Survey of Queensland).

This account summarizes the modifications to the stratigraphy resulting from this survey, and examines the application of the magnafacies-parvafacies concept to stratigraphic relationships in the Georgina Basin.

DEVELOPMENT OF THE CURRENT STRATIGRAPHIC CLASSIFICATION

Following the realization by Woolnough (1912) that sediments underlying the Barkly Tableland in the Northern Territory were Cambrian in age, it became apparent that so-called Devonian, Cretaceous, and Tertiary outcrops in north-western Queensland were in fact Cambrian. It was not until 1936 that Whitehouse made the first systematic stratigraphic and faunal subdivision of the Cambrian in

Queensland. He recognized a basal, noncalcareous 'Templeton Series' overlain by the 'Georgina Limestone', the clastic 'Pituri Sandstones', and the 'Ninmaroo Limestone' (Fig. 2A). In this classification, the 'Templeton Series' includes those Middle Cambrian units that are now known as the Beetle Creek Formation and its equivalents, and the Inca Formation; the 'Georgetown Limestone' includes all the late Middle and early Upper Cambrian limestone units as well as the Camooweal Dolomite; and the 'Pituri Sandstones' encompass such clastic units as the Split Rock Sandstone and the Steamboat Sandstone. Whitehouse did not recognize the presence of a basal carbonate unit (the Thorntonia Limestone), and placed the dolomites of the Barkly Tableland (the Camooweal Dolomite) well up in the sequence. Where he observed carbonate beds directly overlying Precambrian outcrops, he assumed an overlap of 'Georgina Limestone'.

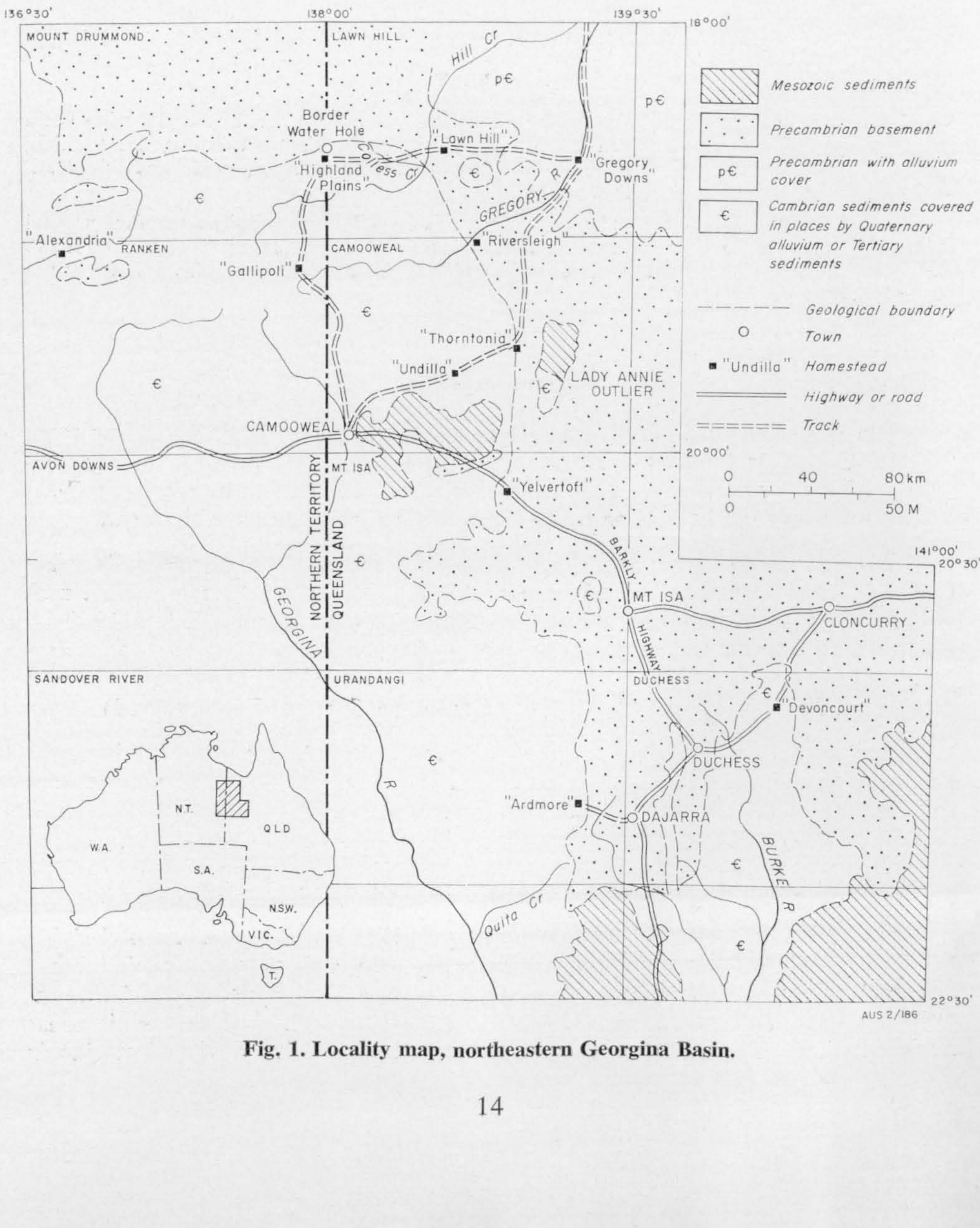


Fig. 1. Locality map, northeastern Georgina Basin.

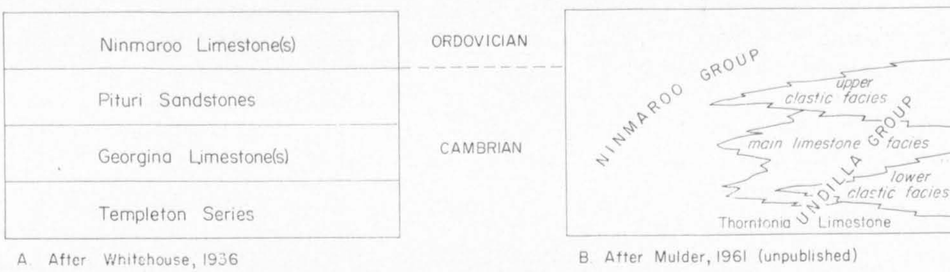


Fig. 2. Subdivision of the Cambrian section in the Georgina Basin, northwest Queensland, according to Whitehouse (A) and Mulder (B). Horizontal lines represent time lines.

Initially Whitehouse (1936) regarded his units as purely time-stratigraphic entities, but later (1939) he considered the 'Templeton Series' to be the time-equivalent of the lower part of the 'Georgina Limestone', regarding it as the leached residuum of an originally calcareous bed.

David & Browne (1950) adopted essentially the same subdivisions, recognizing the Templeton/Georgina 'facies' relationships. Unfortunately their nomenclature was not consistent as they indiscriminately used the terms 'Georgina Limestone', 'Georgina Series', 'Georgina Limestone Series', as well as 'Templeton Series' and 'Templeton Facies'.

Noakes & Traves (1954) introduced the term 'Barkly Group' for all Cambrian units of the Barkly Tableland, but this term, being rather meaningless, was later abandoned. An important phase of the investigation was begun about 1950 by the Bureau of Mineral Resources, particularly by Öpik, who completely revised the stratigraphy and introduced more than 40 new formation names based partly on lithology, partly on the fossil record (Öpik, 1956a, 1960, 1961). Some of Öpik's 'formations' are not mappable lithostratigraphic formations in the strict sense of the definition, but are biostratigraphic units. A controversial aspect of Öpik's work was his opinion that the Camooweal Dolomite is a late Proterozoic or Lower Cambrian unit, forming a basement to the Middle and Upper Cambrian units, and separating the Cambrian deposits in Queensland from those in the Northern Territory. This provided an explanation of the faunal differences between the two provinces, but was a contradiction of Whitehouse's concept that these dolomites are high up in the 'Georgina Limestone' section, and represent the youngest strata that are preserved in the central parts of the Cambrian basin. Subsequent investigations (Condon, 1961; Mulder, 1961; Randal & Brown, 1962; Smith & Roberts, 1963), gradually made it clear that, in fact, the Camooweal Dolomite is at least Middle to Upper Cambrian in age and perhaps Ordovician in places. Drilling results support this conclusion, e.g. Johnson, Nichols, & Bell (1964); Milligan (1963); and also more recent drilling by Mines Exploration Pty Ltd in the Northern Territory. Mapping by the writer in the Border Waterhole-Lawn Hill area (de Keyser, 1969b) showed conclusively that the Middle Cambrian Currant Bush Limestone in the headwaters of Colless Creek is overlain by Camooweal Dolomite, and that other outcrops of 'Camooweal Dolomite', resting directly on Precambrian basement, are actually part of the Thorntonia Limestone. The Currant Bush Limestone and the Border Waterhole Formation in the Colless Creek region both pinch out southwards, and Camooweal Dolomite directly overlies

Thorntonia Limestone in that direction. Although in outcrop the latter two formations cannot be lithologically distinguished, their contact can be fairly easily followed on air-photographs for many miles before it is lost under superficial Mesozoic sediments. One can only conclude that the Camooweal Dolomite (and the dolomite lithosome) probably ranges in age from lower Middle Cambrian to Ordovician.

Nichols (1966) and Brown (1968) suggest that these sediments were deposited on intertidal mud flats, in saline lagoons, and possibly in shallow ephemeral saline lakes, and that early diagenetic dolomitization occurred during periods of exposure. The Camooweal Dolomite thus formed a contemporaneous physical and chemical barrier separating the western faunal province of the Northern Territory from the eastern province in Queensland. The marine sector of the Georgina Basin in Queensland, present in the Yelvertoft-Thorntonia area, the Mount Isa-Quita Creek area, and the Border Waterhole area, must have formed a rather narrow sea arm no more than 80 km wide, at the most, and some 450 km long. The Burke River Outlier in the Duchess district formed another, much smaller, sea arm about 100 km long and 30 km wide. Both arms probably opened towards the ocean to the south and southeast.

In an unpublished report Mulder (1961) noted the marked facies changes in the Cambrian: a predominantly clastic facies at the edge of the basin is laterally replaced and overlain by a carbonate facies towards the centre. Using mainly photogeological interpretation he distinguished between an 'Undilla Group' made up of four lithological units, and a 'Ninmaroo Group' which included the Camooweal Dolomite as well as the Ordovician Ninmaroo Formation (Fig. 2B). The four units of the 'Undilla Group' can be closely compared with Whitehouse's units: (1) the Thorntonia Limestone; (2) a lower clastic facies (cf. 'Templeton Series' of Whitehouse); (3) a main limestone facies (cf. 'Georgina Limestone'); and (4) an upper clastic facies (cf. 'Pituri Sandstones'). Mulder's 'Ninmaroo Group' is characterized by the predominance of dolomite instead of limestone. It should be noted that all his units are facies units, in which considerable time transgressions are possible, whereas the equivalent Whitehouse units were thought to be time-bound divisions. Unlike Whitehouse, Mulder separated the dolomites from the limestones, regarding the Camooweal Dolomite as a dolomitic facies of the upper part of his 'Undilla Group'. He also considered that towards the north, the lower members of the 'Undilla Group' were developed in a dolomite facies so that it becomes difficult to draw a boundary between the dolomites of the Thorntonia Limestone and those of the Camooweal Dolomite.

A new impetus to the study of the Cambrian stratigraphy of the Georgina Basin was given by the discovery, in 1966, of phosphate deposits in lower Middle Cambrian formations (Russell, 1967). In the wake of this, the Bureau of Mineral Resources carried out a mapping programme during the period 1967-1969 (de Keyser, 1968, 1969a and b, 1972; de Keyser & Cook, 1972), in which the Geological Survey of Queensland initially participated; later, from 1968 onwards, the State Survey launched an independent programme of palaeoecological and environmental studies undertaken by G. Fleming. During all these investigations, the conclusions of Randal, Brown, Mulder, Smith & Roberts, etc. were generally confirmed and extended. Smith (1972) should be consulted for a comprehensive bibliography of Georgina Basin literature.

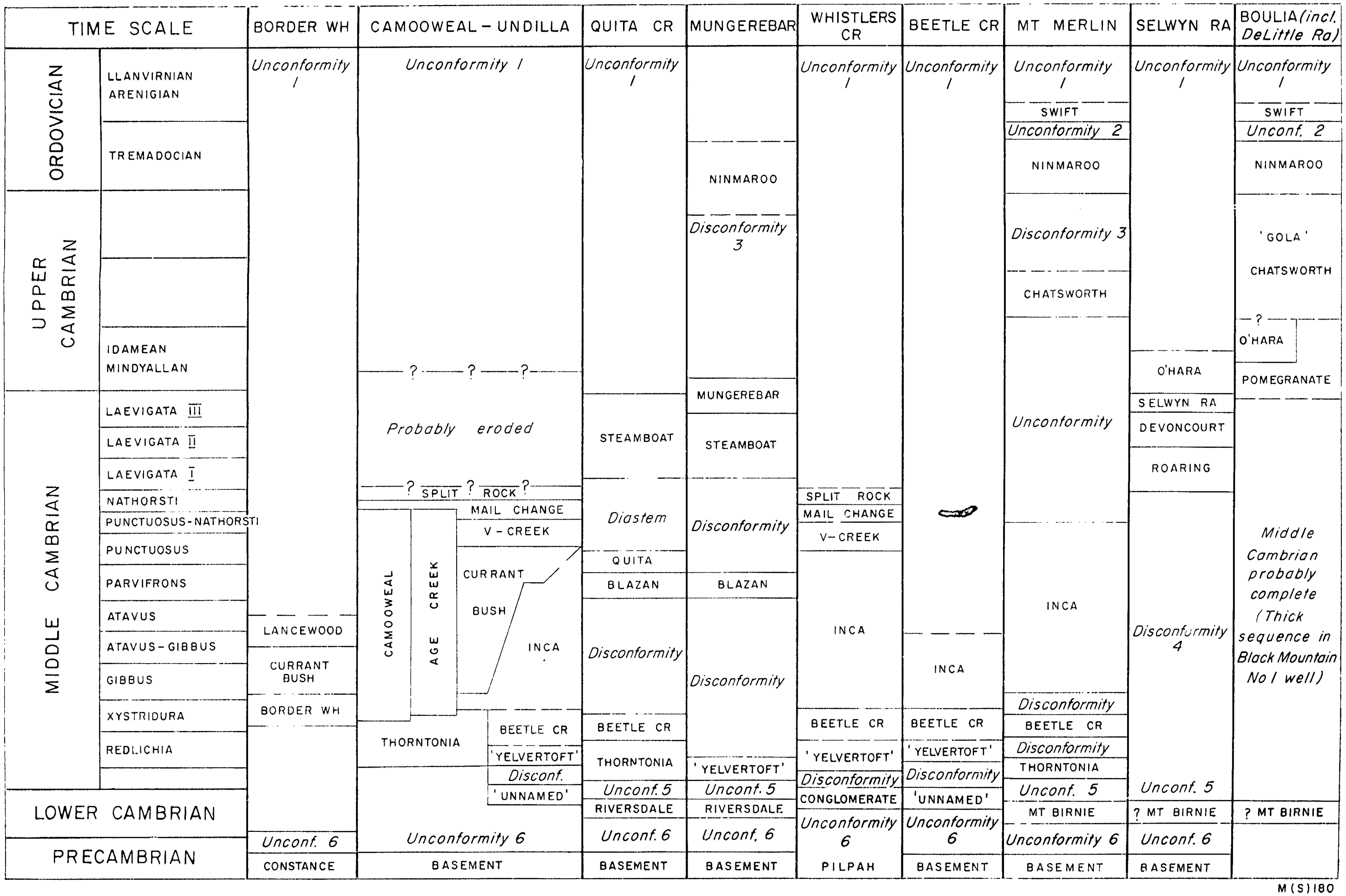


Fig. 3. Current classification and correlation chart of Cambrian (and Ordovician) deposits, Georgina Basin, northwestern Queensland. After Opik (1960) with minor modifications.

Problems of the Current Stratigraphic Nomenclature

According to the Australian Code of Stratigraphic Nomenclature, 'Formations are the mappable units which the field geologist normally defines, portrays on a map, and uses in his description of the rock succession in a particular district'. There are two main requisites for a formation: mappability, and lithological constancy (expressed by a single dominant rock type, or by a distinctive assemblage of several rock types). A formation is a rock-stratigraphic unit; its boundaries may transgress time planes. It may be characterized by any suitable physical factor that helps in distinguishing it from other formations. Where fossils are plentiful and distinctive they, too, may be used in the delineation of a formation. However (according to the American Stratigraphic Code), a 'unit distinguishable from the enclosing rocks only by its fossils shall not, in general, constitute a formation'. For most strata in which fossils are not abundant this rule is a sound one for the field geologist.

Time-stratigraphic units are defined by the fossil assemblages they contain. The units used in this case are therefore biostratigraphic and based on fossil zones, each of which is designated by the name of a characteristic fossil. Biostratigraphic units are bounded by time planes, and because rock-stratigraphic units (formations) may transgress time planes, biostratigraphic units and formations do not necessarily coincide.

Taking stock of the current stratigraphic subdivision of the Cambrian of the Georgina Basin, as built up over the years by Öpik and others (see Fig. 3), one comes to the conclusion that it is a mixture of rock-stratigraphic and biostratigraphic elements. Öpik (1954) started the stratigraphic classification correctly by stating, in an explanation of one of his charts, that 'the fossil ranges are explicitly not connected with the formations as such', but during subsequent work this principle was not always followed. In some instances the biostratigraphic and the lithostratigraphic boundaries may coincide, owing to the presence of diastems or disconformities, but there are many cases where the 'formations' shown on the chart (Fig. 4) are not formations *sensu stricto* (i.e. rock units), but are purely biostratigraphic units.

A good example of this can be seen in the Burke River Outlier, in the Duchess district, where the Inca Formation and Roaring Siltstone, although they are lithologically identical and occur in continuous outcrop, have nevertheless been given different names because of their (gradually) differing ages as indicated by their fossils. Another example may be seen in the Highland Plains homestead area, where a limestone in the upper section of the Border Waterhole Formation is lithologically indistinguishable from the adjoining younger Currant Bush Limestone. In general, most of the Middle and Upper Cambrian limestone sequences consist of lithologically identical, thin-bedded to medium-bedded, grey calcilutites and calcisiltites that can be dated only where fossils happen to be found. In areas where several of these units occur together, the existing biostratigraphic classification is unsatisfactory and completely impracticable for the purpose of field mapping.

In some cases, the situation is saved more or less fortuitously by the fact that the relevant units occur in geographically distinct outcrop areas, separated from each other by younger superficial sediments. This applies, for example, to the

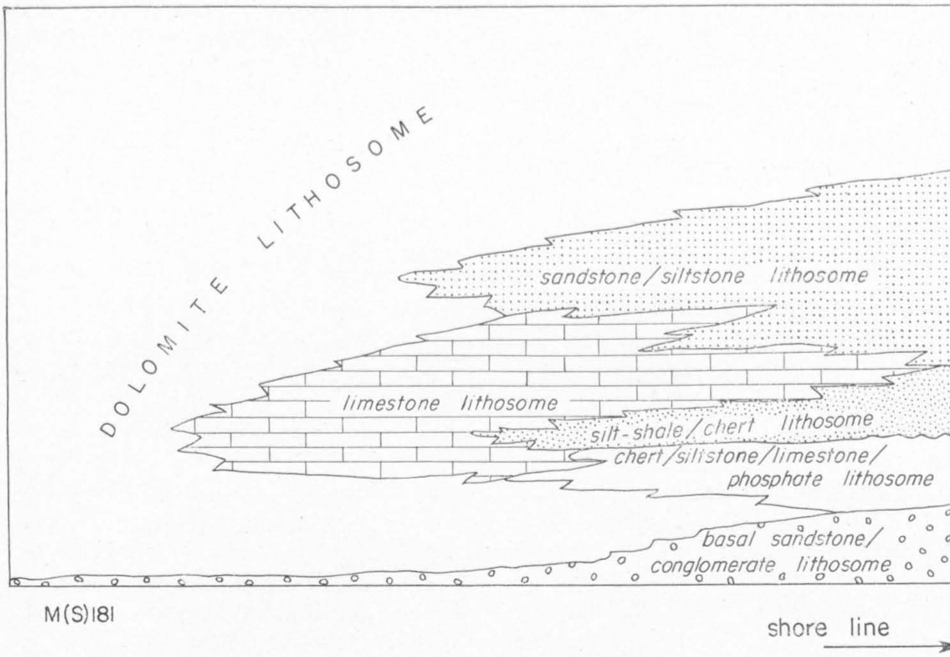


Fig. 4. The relationship between the six main lithosomes of the Georgina Basin, northwestern Queensland. Horizontal lines represent time lines.

Upper Cambrian Pomegranate Limestone in the Burke River Outlier, which is set apart from the lithologically identical Middle Cambrian or lower Upper Cambrian Devoncourt and Selwyn Range Limestones by an alluvial plain.

Apart from the mixture of biostratigraphic and lithostratigraphic nomenclature, a second point that has led to some complication is the application of different names to one and the same unit exposed in widely separated areas. This arose from the previous less complete state of geological knowledge, which interpreted the various outcrop areas as remnants of separate basins or sub-basins developed in a Lower Cambrian or upper Proterozoic land area. For instance, there is no lithological difference between the contemporaneous Beetle Creek Formation, Border Waterhole Formation, and over the border, the Burton Beds, Wonarah Beds, and even the Gum Ridge Formation as far west as Tennant Creek. Similarly, the Currant Bush Limestone and the Quita Formation are of the same age and appearance, and so are the Blazan Shale and the Inca Formation. The one reason for their distinction—their geographical separation—is fast becoming untenable as the more detailed mapping and drilling results of the past few years are gradually linking up these units.

Suggested Amendments to the Nomenclature

The existing stratigraphic classification can be improved by dropping unnecessary terms such as Blazan Shale, Quita Formation, and Border Waterhole Formation. It can, however, be made satisfactory without abolishing most of the current familiar biostratigraphic units, by overprinting a lithostratigraphic nomenclature on to the biostratigraphic one. Use is made of the concepts of magnafacies and parvafacies (see Krumbein & Sloss, 1963, p. 320). Several rock units probably are tongues of one large lithosome*, and the determination of their boundaries may conceivably be rather arbitrary in the field.

It is convenient to group the units into six different lithosomes (Fig. 4) each with its own lithological characteristics.

The manner in which this six-fold division may be applied to the stratigraphy of the Georgina Basin is shown in Figures 5-7.

1. *Basal sandstone-conglomerate lithosome*

This lithosome, of late Precambrian to Lower or Middle Cambrian age, includes the informal 'Mount Hendry Formation', named by Broken Hill South geologists in the Lady Annie area. It is possible that part at least of the Riversdale Formation, and perhaps also the upper part of the Mount Birnie Beds, may be the equivalent of the 'Mount Hendry Formation'. Elsewhere the unit is unnamed or has been included with the Beetle Creek Formation. It comprises conglomerate, basal breccia, grit and pebbly sandstone, and pockets of earthy siltstone and mudstone. The size of clasts, sorting, rounding, constitution of components, thickness, dolomite content of the matrix, and other physical factors, are extremely variable. Thicknesses up to 25 m have been measured. The sandstones are usually coarse or medium-grained (0.25-0.50 mm), and often quite immature. The basal breccias generally reflect the composition of the local Precambrian basement. Cross-bedding, with some individual beds up to 1.5 m thick, is especially well developed in the Ogilvie Range locality 20 km south-southwest of Yelvertoft homestead, where individual sets may be up to 2 m thick. Imbrication in conglomerate is common. There is little doubt that this lithosome represents a transgressive, littoral-marine, in places perhaps fluviatile, series of deposits forming the base of the sedimentary Cambrian succession.

2. *Chert-siltstone-limestone-phosphate lithosome*

Included in this lithosome are the Beetle Creek and Border Waterhole Formations, and in the Northern Territory the Burton Beds and Wonarah Beds, all of lower Middle Cambrian age. The cherts are very commonly silicified carbonate rocks; they are characteristically lumpy and nodular, and commonly crop out as loose rubbly layers representing the residue of leached chert-carbonate beds. Even-bedded chert also occurs. A very common chert type, typical of this lithosome, is the silicified microcoquinite.

The siltstone and shale are commonly phosphatic to some degree, and are usually bleached to white, cream, yellow, or pink colours. They are well bedded, and in places very fossiliferous.

The limestones are predominantly white or creamy biopelsparites, and almost invariably phosphatic. They are regarded as submarine bank and shoal deposits.

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* Wheeler & Mallory (1956, p. 2719) define a lithosome as 'a lithostratigraphic body which is mutually intertongued with one or more bodies of differing lithic constitution'. It is perhaps expressed a little more succinctly by Krumbein & Sloss (1963, p. 301), who regard lithosomes as 'masses of essentially uniform lithologic characters which have intertonguing relationships with adjacent masses of different lithology'.

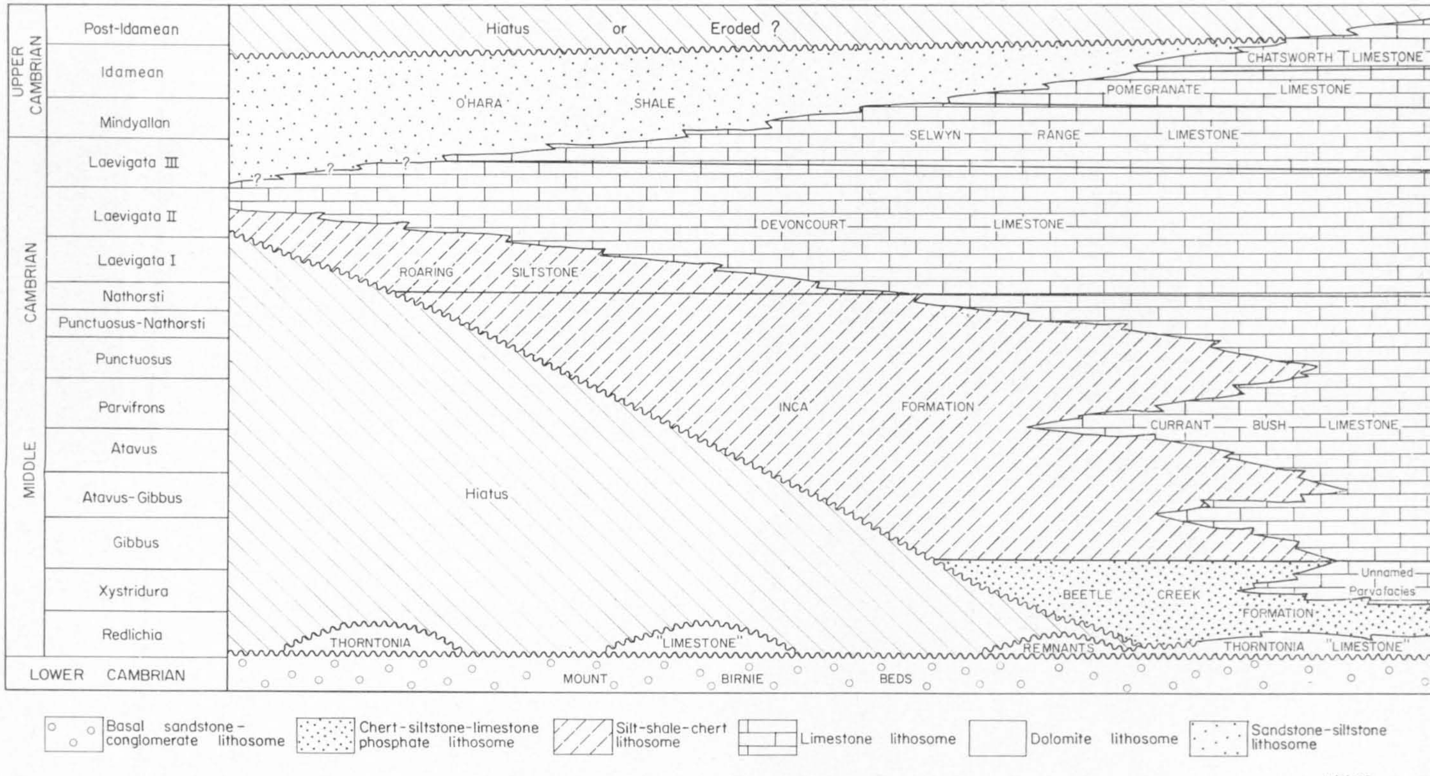

Fig. 5. Stratigraphic relationships in the Burke River Outlier, Duchess District, northwest Queensland. (Horizontal lines represent time lines. Lithostratigraphic units are capitalized.)

South of Thorntonia homestead, a few intercalations of coarsely crystalline (up to 2 mm) white dolomite beds are found, one of which forms the top of the formation. Grey, micritic limestone is very scarce in this lithosome.

The phosphorites include pelletal, fine-grained and silty, microphosphoritic, and secondary recrystallized varieties.

All these different lithological types are closely associated, and are therefore grouped together in the one lithosome. The lateral variability, the presence of diastems, the palaeogeographical configuration, and the clay mineralogy (Cook & Armstrong, 1970), all suggest that the lithosome is composed of strata that were deposited in very shallow water over a shelf platform characterized by shoals and an irregular coastline broken by embayments, lagoons, and estuaries.

3. *Silt-shale-chert lithosome*

This is represented mainly by the Inca Formation, but also includes the Blazan Shale, Lancewood Shale, and Roaring Siltstone. The lithosome, which can be equated with Whitehouse's 'Templeton Series', or Mulder's lower clastic facies, ranges in age from middle to upper Middle Cambrian.

Well bedded, thinly laminated siltstone, silt-shale, siliceous shale, very fine-grained sandstone, and intercalated thin-bedded chert are the normal constituents. Shingle-type superficial rubble of flat platy fragments of silt-shale and thin chert are rather typical of this interval. They can be distinguished from the Beetle Creek silt-shale rubble by their fossils (almost exclusively agnostids and sponge spicules instead of the larger trilobites of the Beetle Creek Formation), their thin chert fragments (instead of the nodular cherts and silicified microcoquinites of the Beetle Creek Formation), and the angularity of the platy fragments (whereas the Beetle Creek fragments tend to be slightly more rounded and perhaps a little thicker).

The siltstone is composed predominantly of quartz and sericite, with a little feldspar, chloritized biotite, and accessories including tourmaline and pyrite. Phosphate is absent, except locally near the base of the Inca Formation, where thin lenses (10 cm thick) of pelletal phosphorite (with some barite and a little fluorite) in an otherwise siliceous matrix possibly represent detrital accumulations of pellets derived from the Beetle Creek Formation.

The thin lamination and bedding and the consistency of the unit suggest that quiet conditions of sedimentation in a somewhat deeper marine environment prevailed. The environment was possibly somewhat toxic, as fossils are scarce and rather limited. Where they do occur they are commonly present in large numbers but restricted to a thin bed.

4. *Limestone lithosome*

This includes a large number of units, including the Currant Bush, V-Creek, and Mail Change Limestones, the Quita Formation, Mungerebar and Georgina Limestones, the Devoncourt, Selwyn Range, and Pomegranate Limestones, and also the top part of the Border Waterhole Formation. The age of this lithosome ranges from Middle to Upper Cambrian.

The limestones are calcilutites composed of regular thin- to medium-bedded micrites, pelmicrites, and fine-grained pelsparites, generally light grey to dark grey;

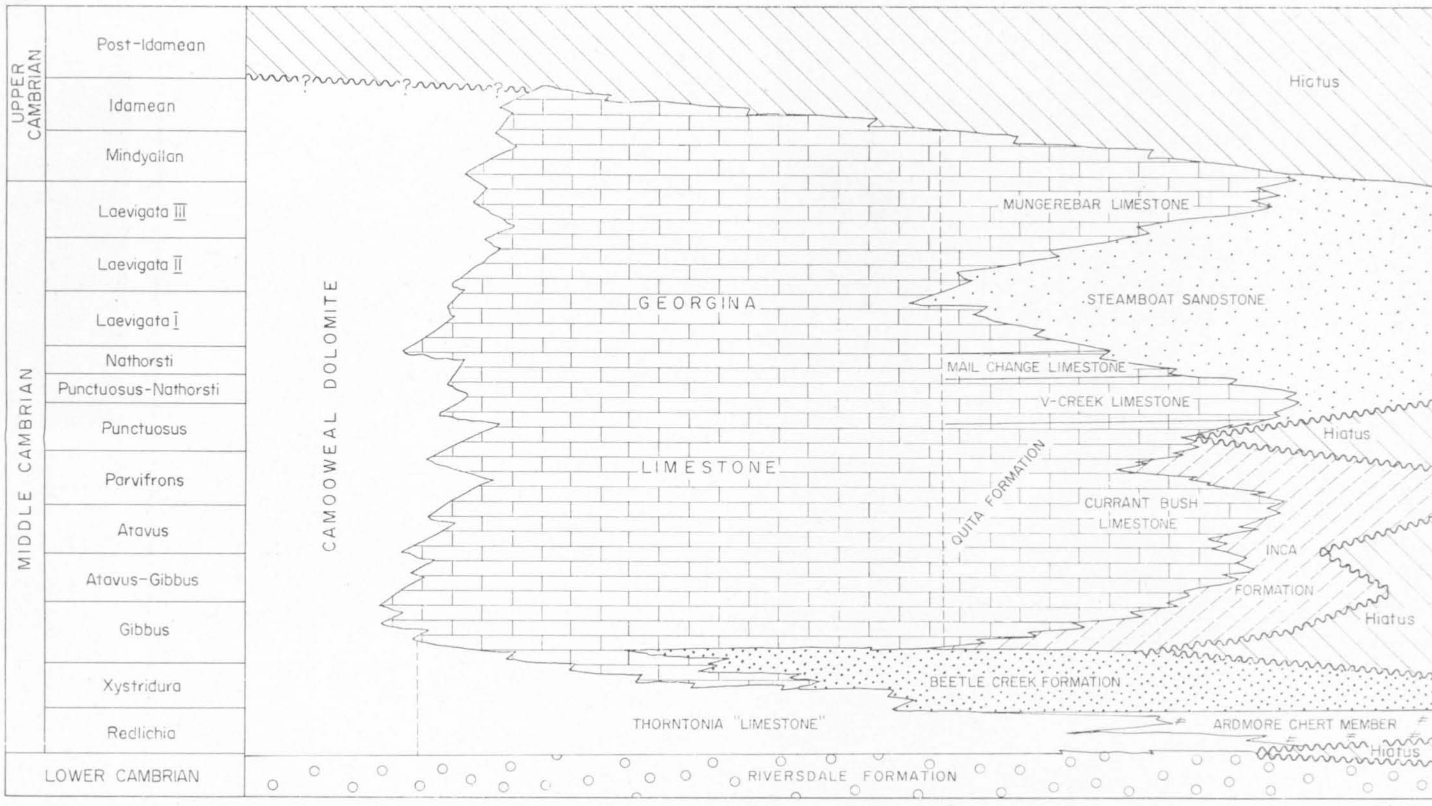

Fig. 6. Stratigraphic relationships in the Ardmore/Quita Creek area, northwestern Queensland.

they are commonly bituminous. Most are soft, but there are also hard, splintery, rather siliceous varieties. Chert is common as nodules and laminae. Intercalated with the limestones are marly and silty partings and interbeds. The Mail Change Limestone is distinguished by its thick beds and striking two-tone colour texture.

The grainsize of these calcilutites normally ranges from 0.005 to about 0.10 mm, but most commonly falls between 0.01 to 0.05 mm. Silt-sized grains of quartz, mica, some feldspar, iron oxides, and tourmaline are present in almost all members, often in sufficient quantity to make the rock a silty or fine sandy calcilutite. The coarser calcilutites usually contain a proportion of micrite pellets and grains of glauconite, and dolomite rhombs are locally fairly abundant. Sedimentary structures are rare, but include some small-scale cross-lamination.

The limestone lithosome replaces the silt-shale-chert lithosome towards the central parts of the basin, and disappears close to the old shoreline. The bituminous smell and the dark grey colours (when fresh) would suggest a somewhat euxinic environment of deposition, and the micritic texture is indicative of predominantly quiet-water sedimentation.

5. *Dolomite lithosome*

The dolomite lithosome has a thickness of up to 300-600 m and includes the Thorntonia Limestone (actually a misnomer as it consists largely of dolomite), the Age Creek Formation, the Camooweal Dolomite, and possibly the Ordovician Ninmaroo Formation. The lithosome includes thin-bedded to massive dolsparite, dolmicrite, and dolomitic limestone. They are buff, light brown, ochre yellow, cream, white, or mottled when fresh, but the exposed surface is commonly tarnished to a dark grey fragmented dolomite; micrite-pellet dolomite, oolitic dolomite, and dolarenite occur as intercalations. Chert nodules are abundant in some beds. Medium-scale cross-bedding is common in places. Stylolites are abundant. Thin sections usually show the dolomitic rhomb texture. Consequently any fossils originally present have been destroyed during dolomitization and recrystallization, although in places ghost remnants of fossil fragments are still recognizable and algal structures have been reported by many authors.

The dolomites as a whole can be distinguished from the limestone sequences by the presence of thick and massive beds and the dirty grey or ochre yellow colour, as opposed to the light grey colour of the limestone.

6. *Sandstone-siltstone lithosome*

This includes the late Middle Cambrian and Upper Cambrian clastic units grouped together as the 'Pituri Sandstones' by Whitehouse, and equivalent to Mulder's 'upper clastic facies'. Units involved are the Split Rock Sandstone in the Yelvertoft-Thorntonia area, the Steamboat Sandstone in the Ardmore-Quita Creek area, and the O'Hara Shale in the Duchess district. The first two units consist mainly of quartz sandstone. The last is a siltstone with interbedded chert layers, resembling the units of the silt-shale-chert lithosome, and physically probably connected with them.

The sandstone-siltstone lithosome represents a phase of increased clastic sedimentation and appears to regressively offlap the limestone lithosome.

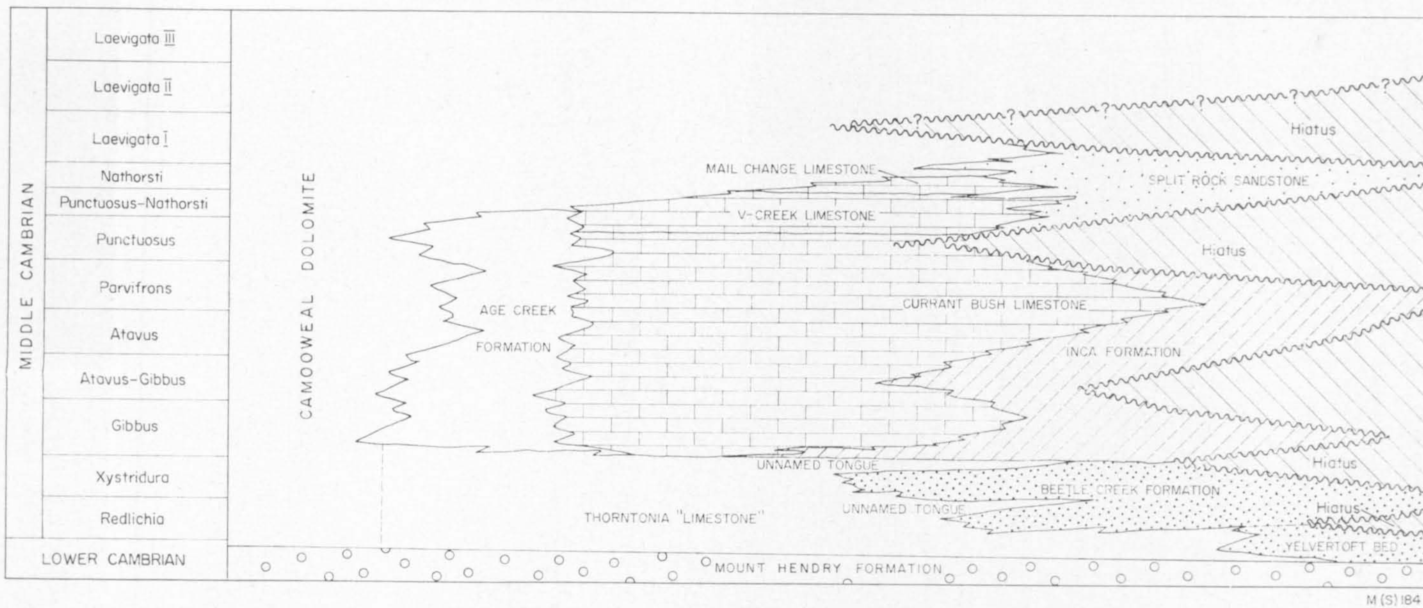

Fig. 7. Stratigraphic relationships in the Undilla Basin area, northwestern Queensland.

Conclusions

1. The stratigraphic nomenclature presently in use in the Georgian Basin is primarily biostratigraphic. It is, in many cases, unsatisfactory for the field geologist.

2. The concept of magnafacies and parvafacies can be applied successfully to the stratigraphic subdivision of the Georgina Basin and serves to highlight the diachronous nature of many of the lithostratigraphic units.

3. The many formally named units may be grouped into six lithosomes, each characterized by a distinctive lithology, and probably formed under unique environmental conditions. The six lithosomes are as follows:

   1. Basal sandstone-conglomerate lithosome (deposited during an initial transgressive littoral-marine phase).
   2. Chert-siltstone-limestone-phosphate lithosome (deposited in extremely shallow water, on a shelf characterized by shoals and a coastline broken by embayment, estuaries, and lagoons).
   3. Silt-shale-chert lithosome (deposited under conditions of quiet sedimentation in slightly deeper water).
   4. Limestone lithosome (deposited some distance off-shore in open marine conditions).
   5. Dolomite lithosome (deposited in intertidal mud flats, saline lagoons, and lakes).
   6. Sandstone-siltstone lithosome (deposited during a regressive clastic phase).

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SUMMARY OF BMR STUDIES OF THE ONSHORE BONAPARTE GULF BASIN 1963-71

by J. ROBERTS[1] and J. J. VEEVERS[2]

INTRODUCTION

The Bonaparte Gulf Basin is the northernmost Phanerozoic sedimentary basin on the northwestern margin of Australia. It extends from a small onshore area around the Joseph Bonaparte Gulf to a wide offshore area beneath the Timor Sea (see Fig. 1 and Veevers, 1967a). Audley-Charles (1965) and Veevers (1969a) have shown the geological continuity, at least from the Permian to the early Miocene, between northwestern Australia and Timor, and Veevers (1967a) has shown the continuity during the Phanerozoic between the Canning and Bonaparte Gulf basins.

The almost exclusively Palaeozoic onshore basin (Figs 2-3) is complemented by the Mesozoic and Cainozoic part of the offshore basin known from drilling and exemplified by the Ashmore No. 1 Well (Fig. 4). Accounts of the offshore basin are given by Caye (1968), Mollan, Craig, & Lofting (1970), and Veevers (1967, 1969a).

History of BMR work

BMR work completed before 1963 is summarized by Traves (1957), Noakes, Crespin, & Öpik (1952), and Thomas (1957). In 1963 a BMR party comprising Veevers (leader), Kaulback, and Roberts spent five months in the field, and in 1965 Veevers and Roberts again spent five months in the field. Extended visits to the party were made by E. C. Druce and P. J. Jones. In 1963 Australian Aquitaine Petroleum also fielded a party in the Bonaparte Gulf Basin; the groups freely exchanged information on the surface geology. A knowledge of subsurface geology came from a parallel campaign of deep drilling which was started onshore in 1963 (Table 1). Laboratory work by those mentioned above and others started in 1963 and drew to a close with the acceptance of manuscripts for publication.

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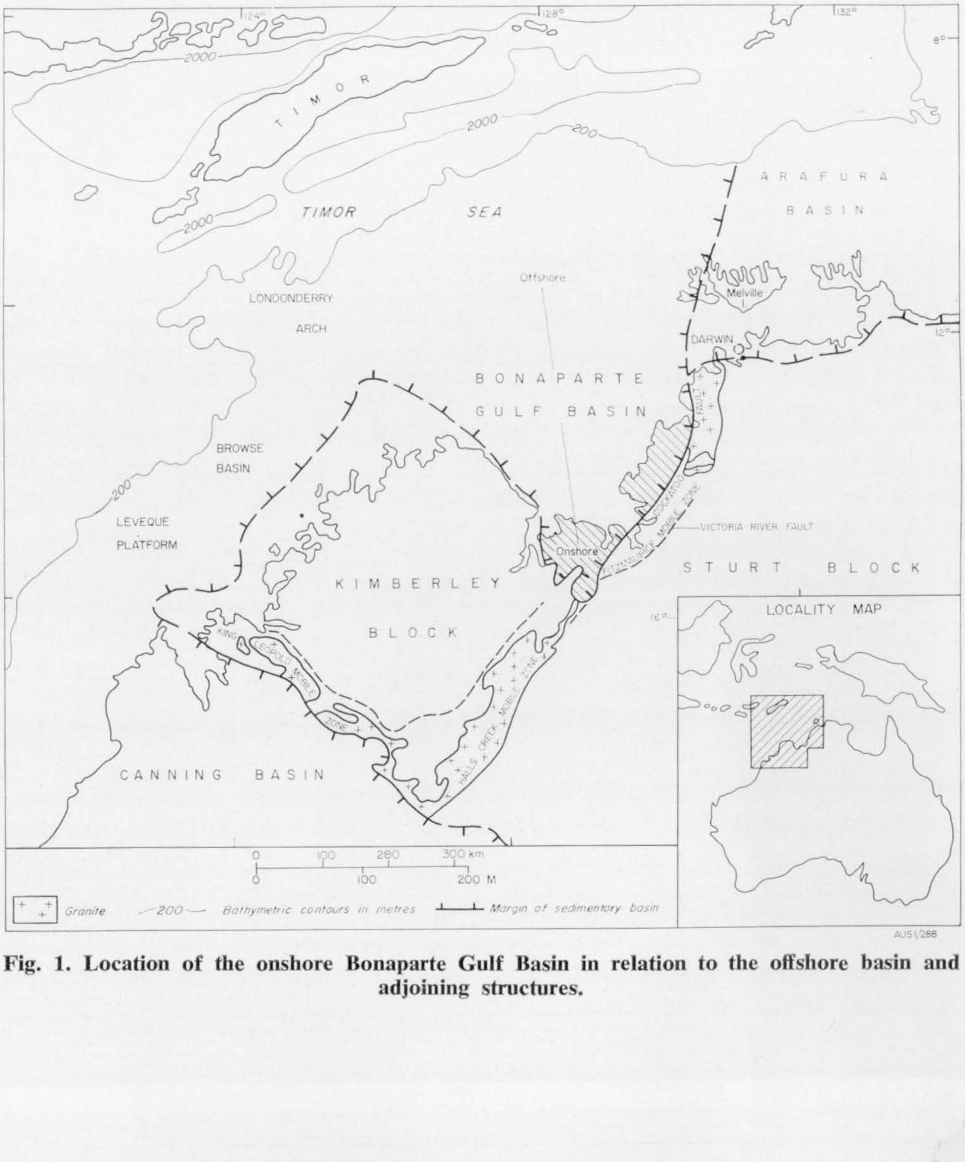

Fig. 1. Location of the onshore Bonaparte Gulf Basin in relation to the offshore basin and adjoining structures.

TABLE 1. ONSHORE AND NEARSHORE DEEP WELLS

Date	*Company*	*Name*	*Total Depth (feet)*	*Spudded into*	*Completed in*
1963	Alliance Oil Development	Bonaparte No. 1	10 530	Lower Permian	Upper Devonian
1964	Alliance Oil Development	Bonaparte No. 2	7 000	Upper Carboniferous	Lower Carboniferous
1965	Australian Aquitaine Petroleum	Kulshill No. 1	14 416	Mesozoic/Lower Permian	Upper Devonian
1966	Australian Aquitaine Petroleum	Kulshill No. 2	6 432	Mesozoic/Lower Permian	Lower Carboniferous
1966	Australian Aquitaine Petroleum	Moyle No. 1	1 767	Mesozoic/Lower Permian	Precambrian
1968	Australian Aquitaine Petroleum	Keep River No. 1	15 623	Lower Permian	?Precambrian
1969	Arco Ltd	Lacrosse No. 1	10 020	Tertiary-Quaternary	Lower Carboniferous
1969	Arco Ltd and Australian Aquitaine Petroleum	Petrel No. 1	13 057	Upper Paleocene	Upper Permian

Aim of this paper

The aim of this paper is twofold: firstly to provide a key to BMR publications on the basin since 1963, and secondly to provide a synopsis of the past eight years' work and in the process to update any material now superseded. This is not an appropriate place to introduce new information except where new material from drilling affects already published conclusions.

Publications

As reflected in the key to publications (Table 2), the main task of the survey was to study the Devonian-Carboniferous sequence. Besides a detailed stratigraphy (Veevers & Roberts, 1968) 20 publications, including eight monographs, are devoted to this topic. Regional accounts of the Cambrian-Ordovician and the Permian-Mesozoic sequences have been published, but there are few specialized works.

Work still in hand includes Carboniferous ostracods (P. J. Jones, BMR), Devonian and Carboniferous corals (D. Hill and J. S. Jell, University of Queensland), Devonian and Carboniferous fish (J. Gilbert-Tomlinson, BMR), Cambrian and Ordovician trilobites (J. Shergold, BMR).

Material in the BMR collection that warrants further attention includes Devonian and Carboniferous bivalve molluscs, Devonian and Carboniferous tracks and trails (ichnofossils), and Permian marine invertebrates. Detailed studies in the field and laboratory could profitably be made of the Devonian and Carboniferous carbonate rocks (including the calcareous algae and stromatolites which are excep-

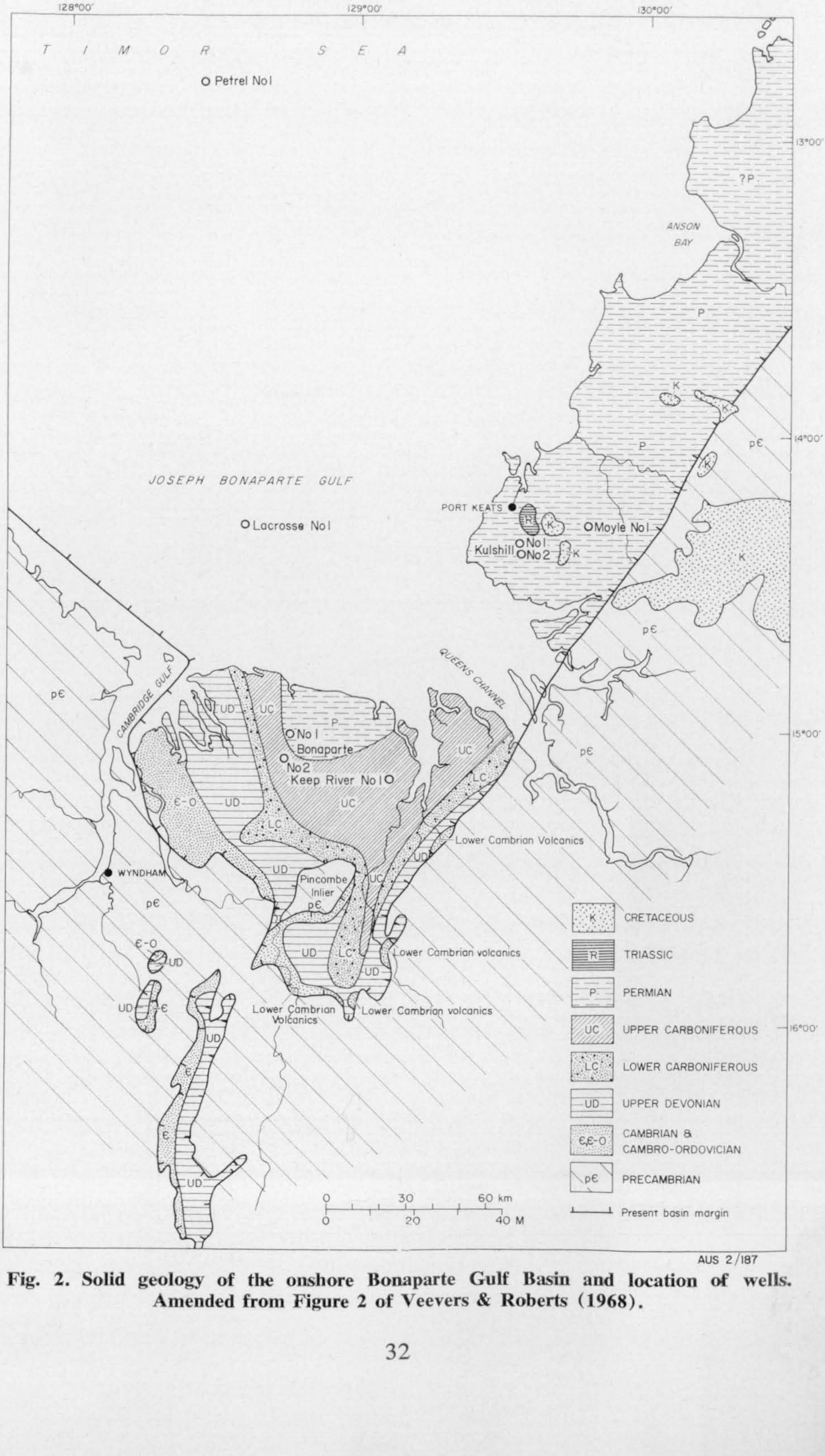

Fig. 2. Solid geology of the onshore Bonaparte Gulf Basin and location of wells. Amended from Figure 2 of Veevers & Roberts (1968).

TABLE 2. KEY TO BMR PUBLICATIONS

Sequence	*General stratigraphy*	*Specialized work*	*Remarks*
Permian-Mesozoic	Dickins, Roberts, & Veevers, 1972		
Upper Devonian-Carboniferous	Veevers & Roberts, 1968	A. *Palaeontological Monographs*	
		Belford, 1970	Systematic description of Upper Devonian & Lower Carboniferous endothyrid Foraminifera.
		Druce, 1969	Systematic description of Upper Devonian & Lower Carboniferous conodonts & correlation.
		Jones, 1968	Systematic description of Upper Devonian ostracods & correlation.
		Playford, 1971	Systematic description of Lower Carboniferous spores, particularly from late Viséan carbonate rocks.
		Roberts, 1971	Systematic description of Upper Devonian & Lower Carboniferous brachiopods & correlation.
		Thomas, 1971	Systematic description of brachiopods from the Septimus Limestone; complements Roberts (1971).
		Veevers, 1970	Description of Upper Devonian and Lower Carboniferous calcareous algae.
		B. *Sedimentological Monograph*	
		Veevers, 1969b	Platform carbonates & coarse terrigenous sediments.
		C. *Special Reports, Announcements*	
		Johnstone et al., 1967	Devonian geology of the western half of Australia.
		Jones & Druce, 1966	Conodont correlations.
		Mamet & Belford, 1968	Carboniferous Foraminifera.
		Playford, Veevers, & Roberts, 1966	Recognition of reefs.
		Roberts, Jones, & Druce, 1967	Upper Devonian correlation from marine invertebrates.
		Roberts & Veevers, 1971	Carboniferous geology—summary.
		Veevers, 1967a	Geology of northwest Australia.
		Veevers, 1968	Computer classification of reefs.
		Veevers, 1969c	Association in carbonate rocks.
		Veevers & Roberts, 1966	Littoral talus breccia.
		Veevers & Roberts, 1967	Upper Devonian geology—summary.
		Veevers, Roberts, Kaulback, & Jones, 1964	New information from 1963 field season.
Cambrian-Ordovician	Kaulback & Veevers, 1969	Jones, 1971	Systematic description of Lower Ordovician conodonts & correlation.
		Öpik, 1967	Cambrian biostratigraphy.

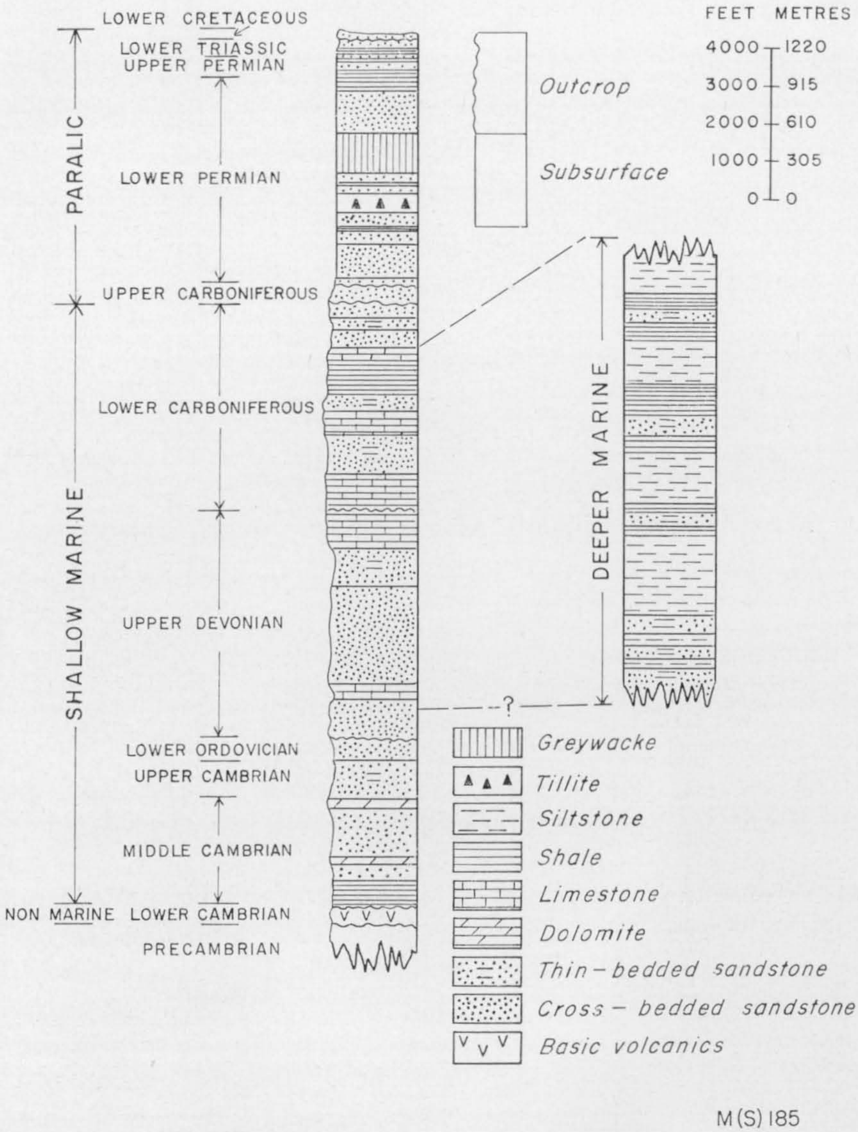

Fig. 3. Diagrammatic composite stratigraphical sections, onshore Bonaparte Gulf Basin. Figure 3 of Veevers & Roberts (1968).

tionally well preserved) and details of reef complexes remain to be worked out. The palaeoecology of selected formations, in particular the Point Spring Sandstone and the Enga Sandstone, would also be worth studying.

EPI-PRECAMBRIAN PALAEOGEOLOGY

Side by side with our study in the Bonaparte Gulf Basin the BMR investigated the bordering Precambrian terrain (Dow & Gemuts, 1969; Pontifex et al., 1968; Morgan et al., 1970; Gellatly et al., in prep.; Plumb, in prep.). We draw on the work of these parties for an account of the Precambrian geology. This is an important topic because the basin evolved as a result of movements along faults between Precambrian mobile zones and stable blocks, or by warping of the blocks.

Much of the southeastern part of the onshore Bonaparte Gulf Basin lies above the northeastern lineament of the Halls Creek and Fitzmaurice Mobile Zones, and the southwestern part lies above a presumed extension of the Kimberley Block. The southeastern margin is marked by the Cockatoo Fault, a splay off the Halls Creek Fault. Archaean metamorphics intruded by Lower Proterozoic and Carpentarian granite are present in the Halls Creek Mobile Zone and in the area southwest of Darwin, and Upper Proterozoic (probably Adelaidean) sedimentary rocks are present in the Fitzmaurice Mobile Zone and in parts of the Halls Creek Mobile Zone.

The chief structural elements are long faults which cut the zones and also separate them from the adjoining blocks and the Bonaparte Gulf Basin. The Kimberley Block contains mainly horizontal quartz sandstone, siltstone, and basic volcanics of Carpentarian age, and the Sturt Block contains nearly horizontal dolomite, chert, siltstone, sandstone, and glacigene sediments of Adelaidean age. The boundaries between the Precambrian and Phanerozoic rocks are everywhere faults except on the northwestern margin of the basin, which has subsided by warping of the Precambrian basement.

Thus at the end of the Precambrian, after a sequence of events that had lasted for some 2000 million years, the region was a fairly flat-lying area of platform sediments bisected by a dormant mobile zone in which the last plutonic episode had taken place some 100 million years before. As a potential source area for future Phanerozoic basins, the region was dominated by mature sediments, mainly quartz sandstone and shale.

CAMBRIAN AND LOWER ORDOVICIAN

On this surface a new regimen of deposition was initiated in the Cambrian and continued throughout the entire Phanerozoic. The mobile zone remained structurally unstable during the Palaeozoic and movements were registered in the sedimentary facies of the southeast Bonaparte Gulf Basin. In contrast, the southwestern depositional margin of the basin is obscure. The Lower Carboniferous littoral talus breccias of the Waggon Creek area are the only clear shoreline features in the southwest, and throughout most of the Palaeozoic the depositional basin extended a considerable but as yet unknown distance southwestwards, as indicated by the downfaulted outliers of Mount Rob, Gap Point, Dillon Spring, and Ragged Range.

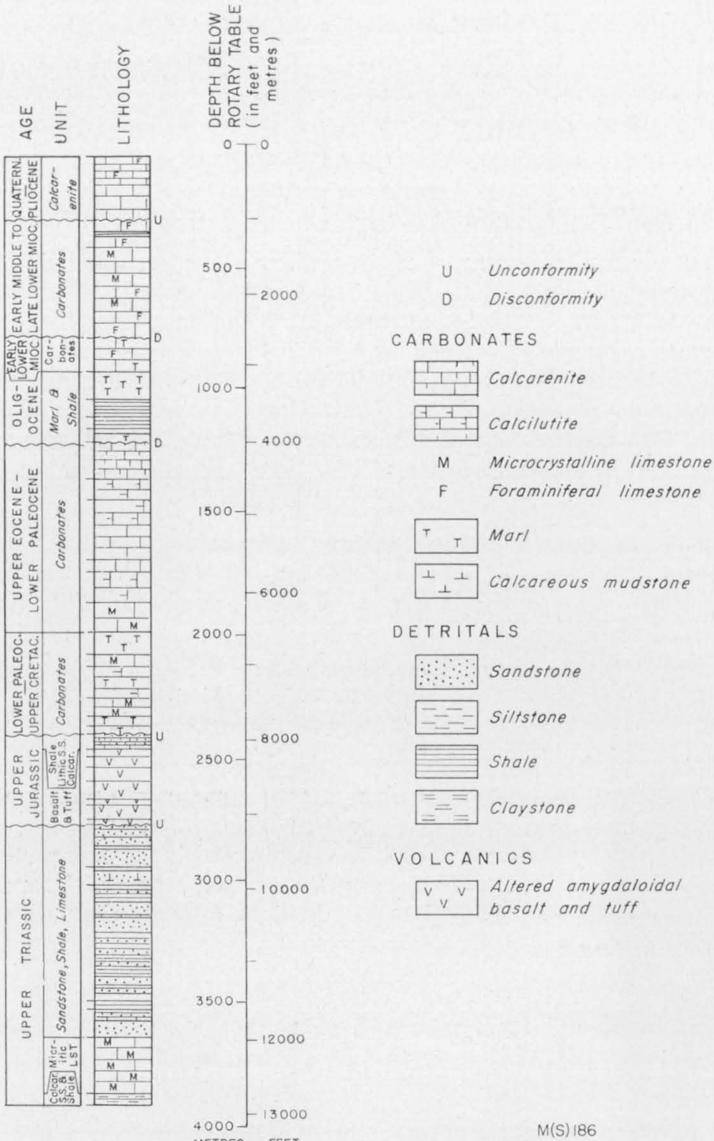

Fig. 4. Section through the offshore Bonaparte Gulf Basin, exemplified by the log of Ashmore No. 1 Well. From figure 2 of Veevers (1969a).

A sheet of basalt flows, tuff, and agglomerate with minor interbedded sandstone and up to 1000 m thick was deposited subaerially over an area of 100 000 km^2 of northern Australia in the Lower Cambrian. The withdrawal from the mantle of this volume of magma (of the order of 100 000 km^3) could conceivably have initiated crustal sagging concomitant with deposition in the Middle Cambrian. From the Middle Cambrian to the Lower Ordovician the area of basalt extrusion was covered by a shallow sea, as indicated by erosional remnants of Cambrian sediments in the Rosewood and Hardman structural basins, by Cambrian and

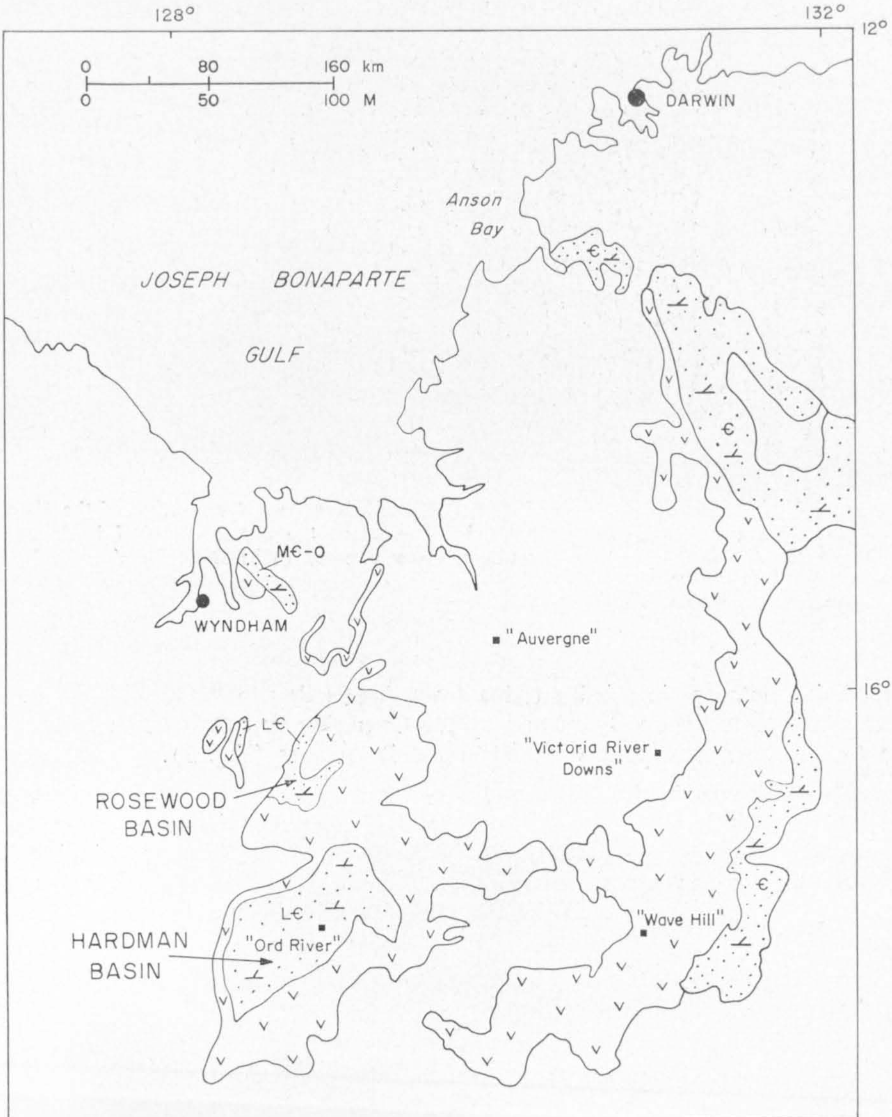

Fig. 5. Outcropping Cambro-Ordovician rocks of the Bonaparte Gulf Basin and its outliers (northwest), Rosewood and Hardman Basins (southward), and Wiso and Daly River Basins (eastern side). Antrim Plateau Volcanics denoted by Vs, sediments by dots. The rocks labelled L€ are now regarded as Middle Cambrian.

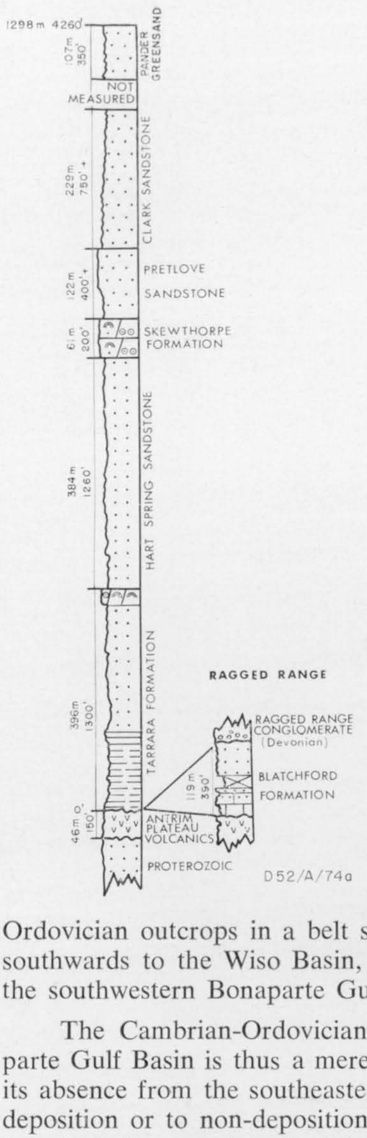

Fig. 6. Composite stratigraphical section of the Cambro-Ordovician rocks of the Bonaparte Gulf Basin and its outliers. From figure 2 of Kaulback & Veevers (1969).

Ordovician outcrops in a belt striking from Anson Bay to Wave Hill and farther southwards to the Wiso Basin, and by the Cambrian and Ordovician outcrops of the southwestern Bonaparte Gulf Basin and its outliers (Fig. 5).

The Cambrian-Ordovician sequence of the southwestern part of the Bonaparte Gulf Basin is thus a mere fragment of an originally widespread deposit, and its absence from the southeastern part of the basin is due either to stripping after deposition or to non-deposition. As elsewhere in this and later shallow seas, carbonate rocks were deposited during intervals of reduced supply of terrigenous sediment.

The Lower Cambrian *Antrim Plateau Volcanics* lie at the base of the sequence in the Bonaparte Gulf Basin (Fig. 6). Here they range in thickness from 30 to 150 m, and are poorly exposed and much altered. They overlie the Precambrian sequence at an erosional and angular unconformity and are overlain by Cambrian and younger (mainly Upper Devonian) sediments at a low-angle unconformity. Part at least of this unconformity is due to differential uplift of the volcanics along faults, as seen in Ragged Range.

Öpik (Appendix 3 *in* Kaulback & Veevers, 1969) divided the Cambrian-Ordovician sedimentary sequence into twelve parts on the basis of fossils, mainly trilobites. Conodonts from the upper part of the sequence have been used by Jones (1971) to correlate these rocks with well documented sections in western Queensland (Jones, Shergold, & Druce, 1971). By these means the sequence is dated as lowest Middle Cambrian* to late Tremadocian or early Arenigian.

The work of Kaulback & Veevers (1969) brought out two new points: first that the rhythmically deposited quartz sandstone, sandy dolomite, and stromatolitic dolomite in the Middle Cambrian *Skewthorpe Formation* are strand, lagoonal, and algal reef deposits of successive transgressions, and secondly that faunal breaks and changes in thickness in the Upper Cambrian *Pretlove Sandstone* indicate differential contemporaneous movement, probably along faults.

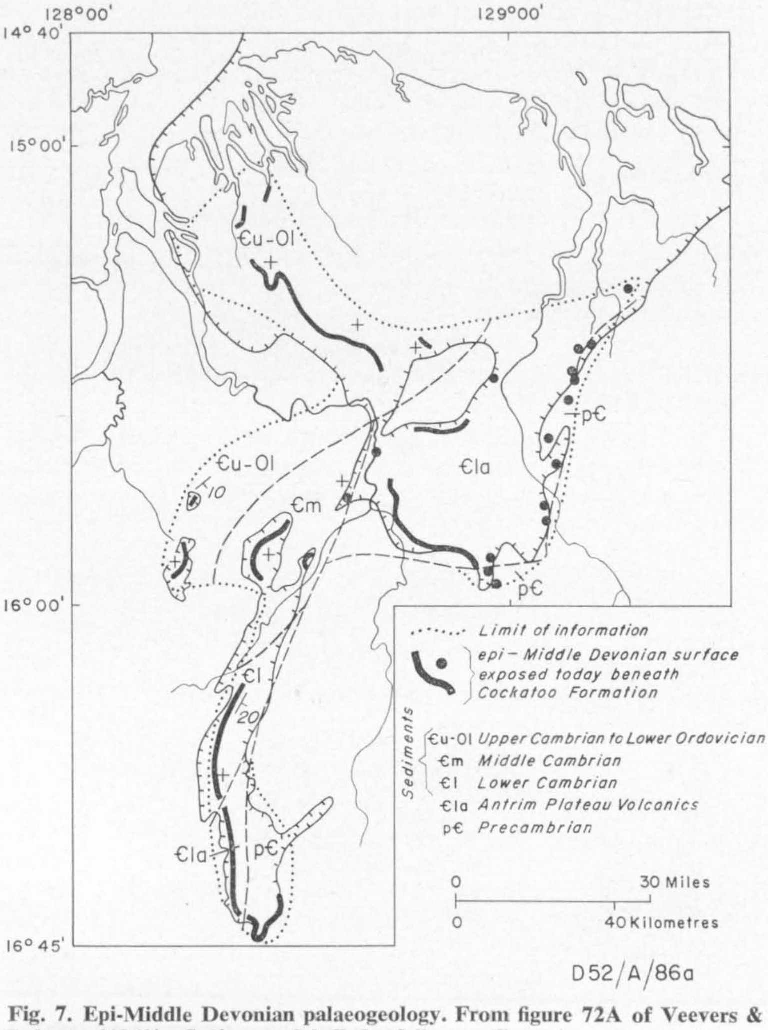

Fig. 7. Epi-Middle Devonian palaeogeology. From figure 72A of Veevers & Roberts (1968). Sediments labelled €l Lower Cambrian are now known to be Middle Cambrian in age.

* This supersedes the late lower Cambrian age earlier attributed to the Blatchford Formation (Kaulback & Veevers, 1969, P1).

Middle Ordovician to Middle Devonian

No record of this interval of about 100 million years is known in the basin. At one stage, in perhaps the Middle Ordovician, the basin was uplifted and eroded. By the end of the Middle Devonian the palaeogeology was as shown in Figure 7, which was drawn according to the information on rocks exposed today beneath the basal Upper Devonian *Cockatoo Formation*. The pre-Cockatoo Formation surface consisted of Precambrian rocks in the east and southeast and progressively younger rocks from Lower Cambrian to Lower Ordovician rocks to the northwest. Except in a few places the Cambrian and Ordovician sediments were subhorizontal. Because the thickness of the Cambrian and Ordovician sedimentary sequence exceeds 1200 m, structure dominated over morphology in determining this distribution. Whether the regional structure was a homocline with a low northwesterly dip, or a fault structure, or a combination of both is not known. In the east and southeast parts of the basin the absence of Middle Cambrian to Lower Ordovician sediments above the Lower Cambrian Antrim Plateau Volcanics and the absence in places of the volcanics are interpreted as indicating differential uplift and erosion between the Lower Ordovician and Upper Devonian.

Upper Devonian and Lower Carboniferous

In the early Upper Devonian, movement along faults caused renewed deposition in the basin and uplift in the Precambrian source areas (composed predominantly of quartzose sandstone) along the southern margin of the basin. Repetition of these movements, but with decreasing intensity, throughout the rest of the Upper Devonian and Lower Carboniferous, and simple differentiation of the supply of quartz sand and mud, led to the distribution of sediments shown in Figure 8. The marine shelf consisted of a shallow platform that bordered a deeper basin so that, as in the comparable Sahul Shelf of today (van Andel & Veevers, 1967), well washed quartz sands and carbonates accumulated on the platform and muds accumulated in the basin.

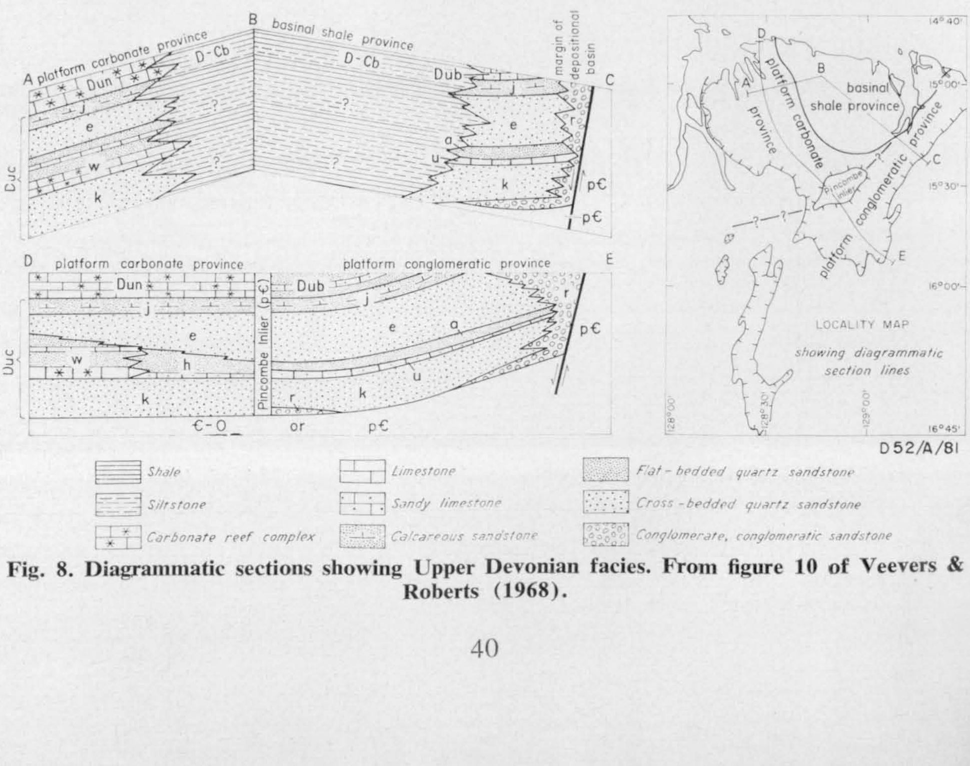

Fig. 8. Diagrammatic sections showing Upper Devonian facies. From figure 10 of Veevers & Roberts (1968).

In a little more detail, the Upper Devonian and Lower Carboniferous shelf sequence comprises four basic elements (Fig. 9):

1. Carbonates.
2. Flat-bedded quartz sandstone deposited along a shifting shore.
3. Cross-bedded quartz sandstone distributed across the platform by tidal currents.
4. Siltstone and shale deposited in the basinal part of the shelf.

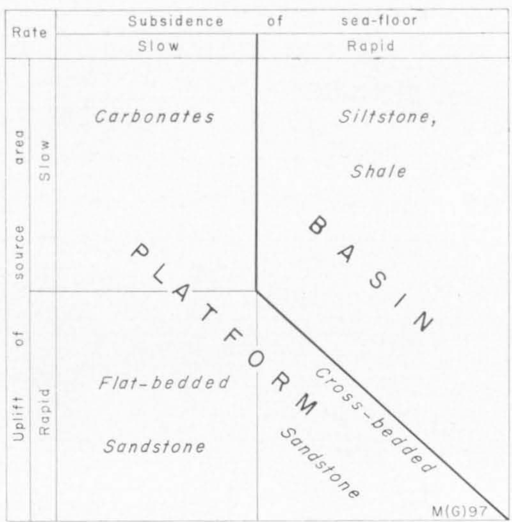

Fig. 9. Depositional response to vertical movements in the source and depositional areas. From figure 29 of Veevers (1969b).

The chief single factor governing the proportion and distribution of these elements was the rate of vertical movement in the source and depositional areas.

The second factor was distance from the shore; this determined:

1. The lateral terrigenous sequence from the shore to the open basin of conglomerate, pebbly quartz sandstone, quartz sandstone, shale, and siltstone (Fig. 10a).
2. The lateral reef facies sequence of inshore lagoon, offshore lagoon, back-reef, reef and inter-reef, fore-reef, and basinal shale and siltstone (Fig. 10b).
3. In the late Tournaisian and Viséan, the simpler lateral sequence (without reefs) of inshore quartzose skeletal limestone, offshore skeletal limestone, and basinal siltstone and shale (Fig. 10d).

The third factor which put an end to reef deposition was the mid-Tournaisian disruption of the pre-existing steady state by an intense uplift of the land and accompanying influx of detritus levelling out the sea floor (Fig. 10c). All the chief vertical and horizontal facies in the shelf sequence are accounted for by these three factors.

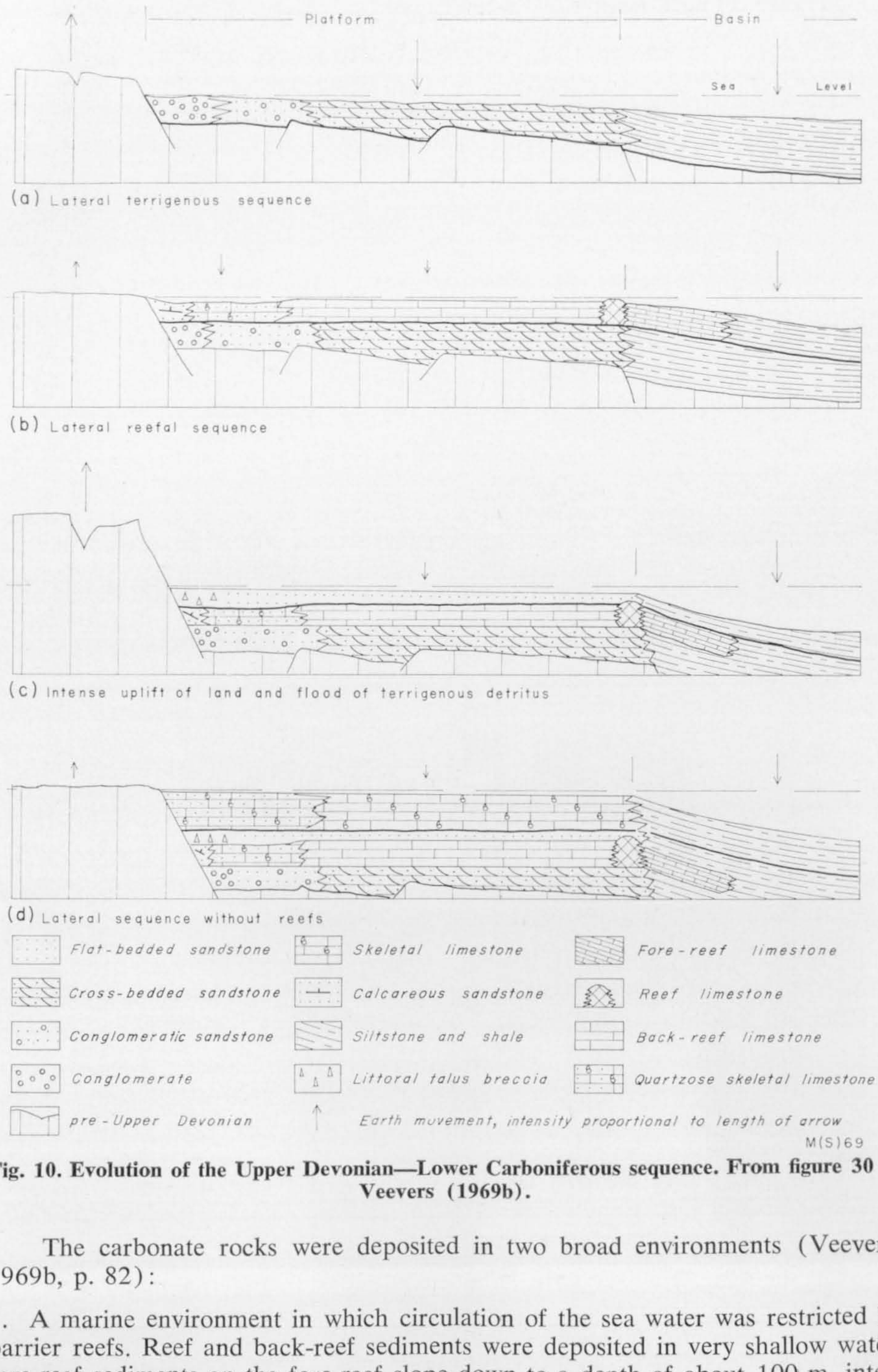

Fig. 10. Evolution of the Upper Devonian—Lower Carboniferous sequence. From figure 30 of Veevers (1969b).

The carbonate rocks were deposited in two broad environments (Veevers, 1969b, p. 82):

1. A marine environment in which circulation of the sea water was restricted by barrier reefs. Reef and back-reef sediments were deposited in very shallow water, fore-reef sediments on the fore-reef slope down to a depth of about 100 m, inter-reef sediments in channels that cut across the reef, and lagoonal sediments in water

deep enough to cover most of the lagoon even at low tide. Marine circulation was greatest over the fore-reef slope, less over the lagoon, and least over the back-reef. The salinity was lowest in parts of the inshore lagoon because of dilution from rivers, high in other very shallow parts, near normal over the rest of the lagoon and the fore-reef and inter-reef, and highest over the reef/back-reef. All but a few of the Frasnian to early Tournaisian carbonates were deposited in this environment.
2. A shallow marine environment open to the sea, in which the later Tournaisian and Viséan carbonates were deposited.

The remaining carbonates, with a considerable proportion of terrigenous material, were deposited close inshore. The basinal siltstone and shale were deposited in front of the fore-reef deposits and platform calcarenite and quartz sandstone, in water whose depth must have exceeded 100 m.

After the uplift of source areas and the marine transgression in the early Frasnian, 1500 m of terrigenous sediments consisting of quartz sandstone and conglomerate and minor carbonate rock (Cockatoo Formation) were deposited on two platform areas, one northwest of the Precambrian Pincombe Inlier characterized by carbonates and terrigenous sediments, and the other southeast of the Inlier containing predominantly terrigenous sediments.

In the southeastern part of the basin, in the vicinity of the Cockatoo Fault, wedges of conglomerate were deposited at the foot of the fault scarps. Within a few kilometres these conglomerates passed laterally into a basal member of parallel-bedded and cross-bedded quartz sandstone. This sandstone was deposited by tidal currents sweeping across a basin elongated in a northwesterly direction (Figs 11-12). Succeeding the basal member in the southeastern part of the basin are two thin members consisting of glauconitic sandstone and minor dolomite. Northwestwards along the platform carbonate region (Fig. 8), these members contain more carbonate rocks and pass through a transitional member into reef sediments near the mouth of Cambridge Gulf.

Fig. 11. Cross-bedding azimuths (a) of the early Frasnian Ragged Range and Kellys Knob Members, and (b) of the later Frasnian Cecil Member, with distribution of the conglomerate (circles), and postulated outlines of tidal inlet. After figure 17 of Veevers (1969b).

Renewed faulting later in the Frasnian rejuvenated the source area and resulted in the deposition of another thick quartz sandstone member. This set of movements anticipated the Tournaisian uplift and may have been responsible for a change in the axis of the basin from northwest in the early Frasnian to east-northeast from late Frasnian to late Viséan. This shift is indicated by the change in mean cross-bedding dip azimuths in the two main sand bodies in the Frasnian Cockatoo Formation (see Fig. 11 and Veevers, 1969b). The effect of the faulting was reduced by the end of the Frasnian and the quartz sandstone was succeeded by a sandy carbonate member.

Fig. 12. Early Frasnian.

Early in the Famennian an extensive reef complex became established on the platform areas. It is best observed on the northwestern platform, but it extended over the Pincombe Inlier into the southeastern part of the basin and northwards along the eastern margin (Fig. 13). Recent drilling by Australian Aquitaine Petroleum in Keep River No. 1 Well has confirmed the palaeogeographic reconstruction of Veevers & Roberts (1968, fig. 72), who postulated a reef complex along the eastern margin of the basin during the Famennian. The exposed reef complex (*Ningbing Limestone*) consists of 300 m of back-reef, a narrow reef, inter-reef, and fore-reef; thicker back-reef and reef sediments have been penetrated by Keep River No. 1 Well. Behind the main reef complex Veevers (1968) has recognized impure carbonate rocks deposited in inshore and offshore lagoon areas (*Buttons Beds*).

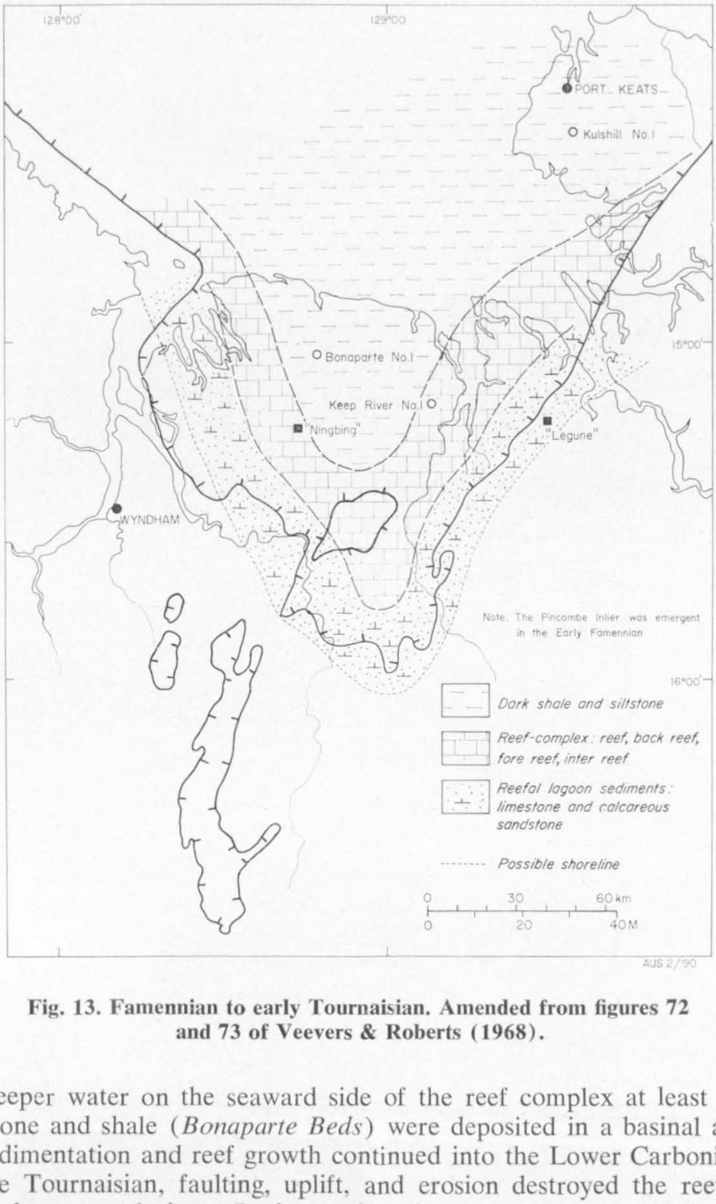

Fig. 13. Famennian to early Tournaisian. Amended from figures 72 and 73 of Veevers & Roberts (1968).

In deeper water on the seaward side of the reef complex at least 700 m of dark siltstone and shale (*Bonaparte Beds*) were deposited in a basinal area. Both basinal sedimentation and reef growth continued into the Lower Carboniferous. In the middle Tournaisian, faulting, uplift, and erosion destroyed the reef complex on the northwestern platform. In the south and east reef growth ceased at this time also but marine sedimentation continued without a break (Fig. 14).

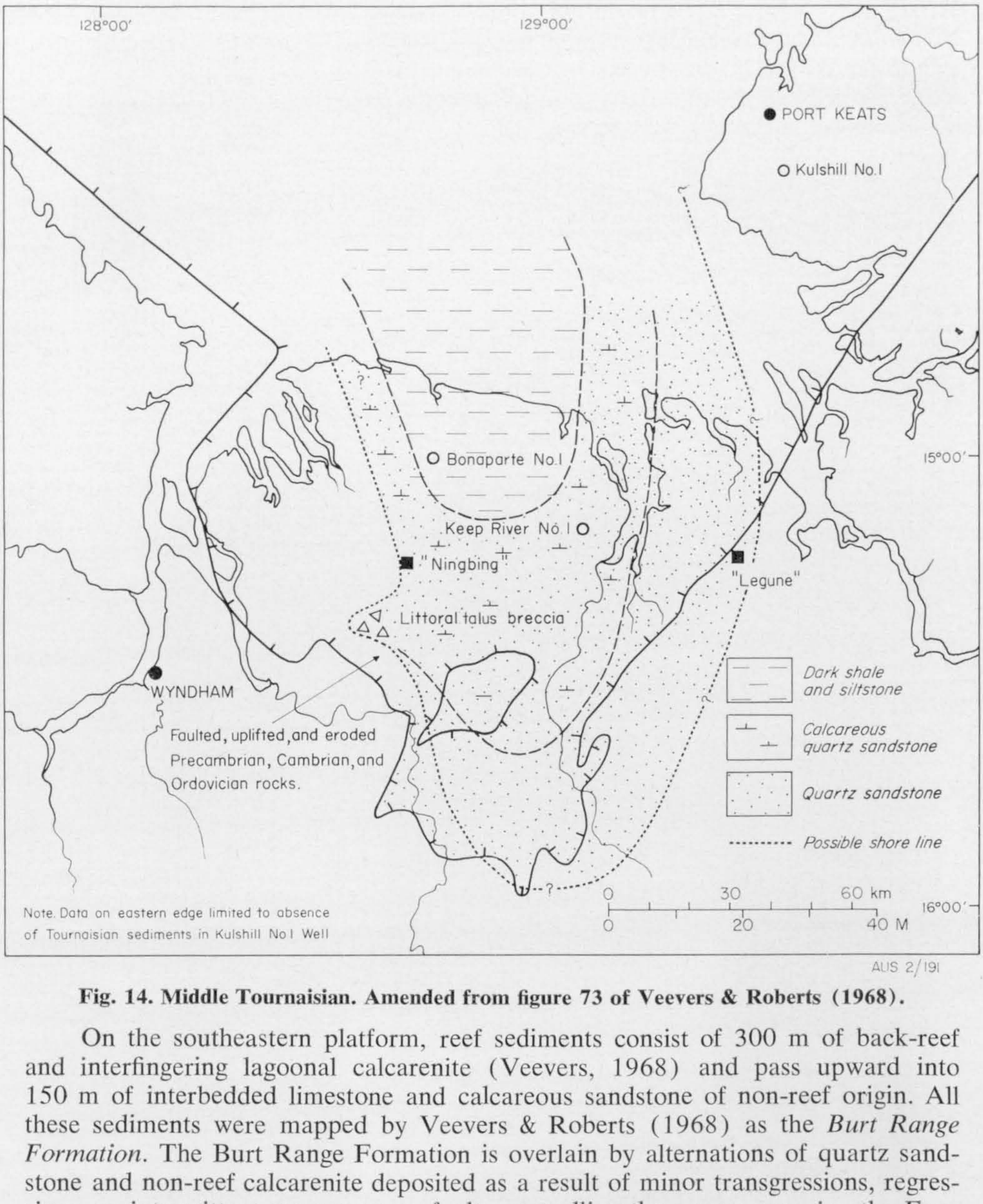

Fig. 14. Middle Tournaisian. Amended from figure 73 of Veevers & Roberts (1968).

On the southeastern platform, reef sediments consist of 300 m of back-reef and interfingering lagoonal calcarenite (Veevers, 1968) and pass upward into 150 m of interbedded limestone and calcareous sandstone of non-reef origin. All these sediments were mapped by Veevers & Roberts (1968) as the *Burt Range Formation*. The Burt Range Formation is overlain by alternations of quartz sandstone and non-reef calcarenite deposited as a result of minor transgressions, regressions, or intermittent movement on faults controlling the source area, i.e. the *Enga Sandstone* (160 m of white quartz sandstone), the *Septimus Limestone* (180 m of crinoidal calcarenite), and the *Zimmermann Sandstone* (140 m of brown to white quartz sandstone). The succession from the Burt Range Formation to the Zimmermann Sandstone is Tournaisian in age.

Seventy kilometres to the north, Australian Aquitaine Petroleum Keep River No. 1 Well, drilled in 1969, penetrated a sequence similar to that in the southeastern part of the platform (Caye, 1969, unpubl.). The succession (Fig. 15)

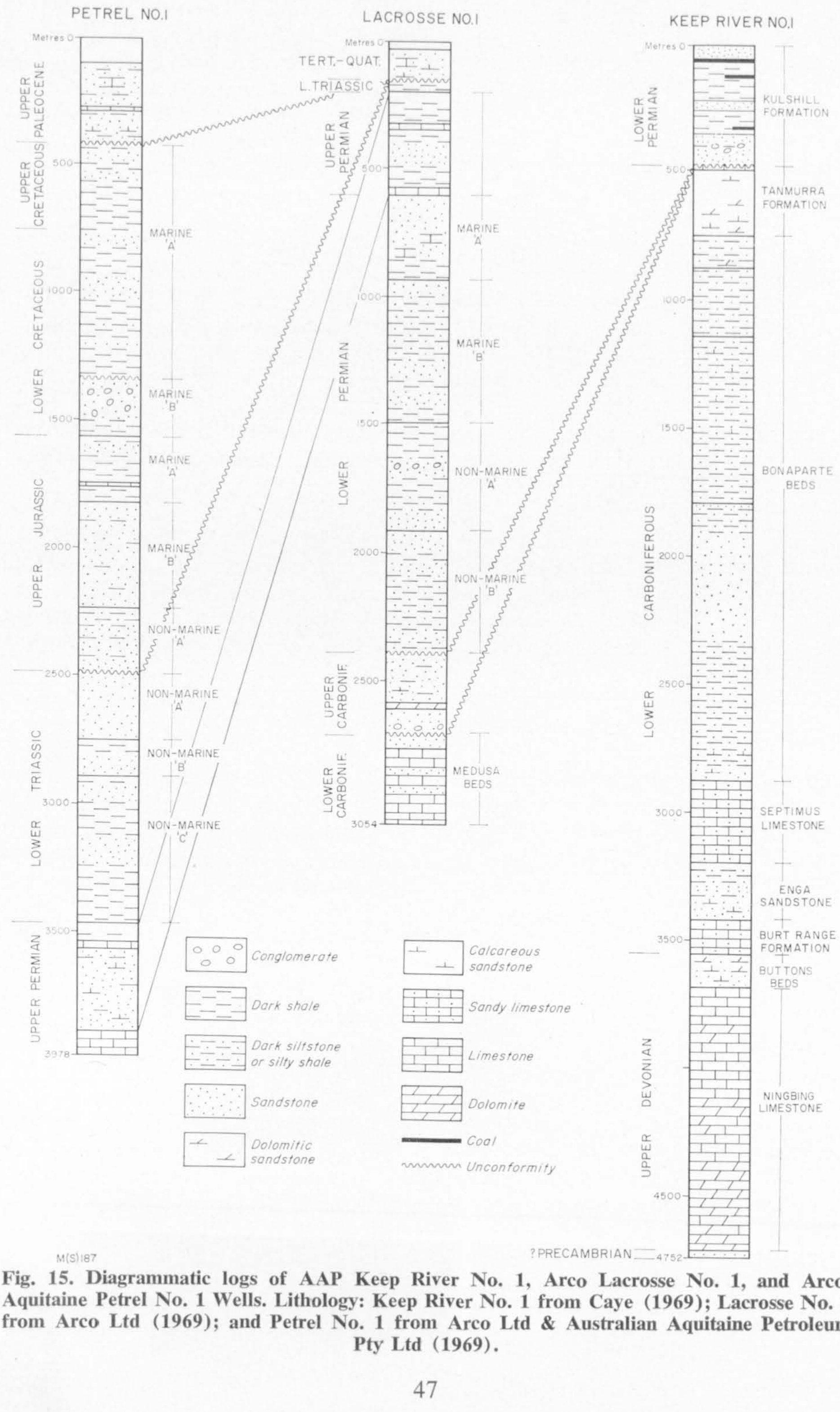

Fig. 15. Diagrammatic logs of AAP Keep River No. 1, Arco Lacrosse No. 1, and Arco-Aquitaine Petrel No. 1 Wells. Lithology: Keep River No. 1 from Caye (1969); Lacrosse No. 1 from Arco Ltd (1969); and Petrel No. 1 from Arco Ltd & Australian Aquitaine Petroleum Pty Ltd (1969).

commences at a depth of 15 623 feet (4 762 m) with 83 feet (25 m) of white to grey quartzite which is referred to either the Upper Devonian Cockatoo Formation or an unnamed Proterozoic formation. Overlying the quartzite are 2 440 feet (744 m) of dolomitic back-reef lagoonal sediments followed by 120 feet (280 m) of reef, all identified as Ningbing Limestone. Dolomitic lagoonal sediments 465 feet (142 m) thick succeed the reef limestone; they are dated by Caye (1969, unpubl.) as Tournaisian to Upper Devonian and are tentatively identified herein as Buttons Beds.

The lagoonal sediments are in turn overlain by the sequence identified in the Burt Range area as: 410 feet (125 m) of Burt Range Formation, 735 feet (224 m) of Enga Sandstone, and 1 060 feet (323 m) of Septimus Limestone. The Septimus Limestone is overlain by 7 030 feet (2 143 m) of dark silty shale with minor sandstone and limestone (Bonaparte Beds) of Viséan age, followed by 906 feet (276 m) of upper Viséan fine to medium-grained sandstone with traces of limestone (*Tanmurra Formation*). Above an unconformity are 1 575 feet (480 m) of Lower Permian silty shale and sandstone containing traces of coal (*Kulshill Formation*).

On the northwestern platform, as mentioned above, faulting, uplift, and erosion in the middle Tournaisian destroyed the reef complex. After a short period of erosion, transgressive seas deposited Tournaisian breccia, 30 m thick, on the shore of the Waggon Creek valley (Veevers & Roberts, 1966); the breccia has not been named. A short time later, in the early Viséan, dark shale was deposited in the central part of the valley and slightly later in the Viséan crinoidal calcarenite (*Utting Calcarenite*) with a thickness of at least 120 m accumulated on the platform north of the Waggon Creek valley.

In the basinal area, dark shale and siltstone approximately 1 800 m thick (*Bonaparte Beds*) continued to be deposited throughout the Tournaisian and most of the Viséan. During the early Viséan, shale was deposited on the platform areas; this was due either to the lowering of source areas, the absence of coarse terrigenous detritus, and conditions unfavourable for carbonate deposition, or to a widespread marine transgression (Fig. 16).

A regression over the whole basin in the late Viséan probably moved the basinal area seawards and resulted in the deposition of littoral sediments over parts of the old basinal area (Fig. 17). The littoral sediments consist of about 85 m of sandstone, shale, and gravelly limestone (*Burvill Beds*) and 56 m of shoreline breccia (*Waggon Creek Breccia*). They are overlain by 270 m of flat-bedded and cross-bedded quartz sandstone (*Point Spring Sandstone*). Farther northward these sediments are represented by about 300 m of sandstone and interbedded limestone (*Tanmurra Formation*). In the present offshore area, as indicated by Arco Ltd Lacrosse No. 1 Well, equivalent rocks with larger amounts of massive oolitic and algal limestone associated with sandstone and calcareous sandstone (Fig. 15) were probably deposited in a shallow shelf environment (Arco, 1969, unpubl.).

In the late Carboniferous the onshore and possibly the offshore parts of the basin were uplifted and eroded. Fluvial quartz sandstone, conglomerate, and siltstone (*Border Creek Formation*) were deposited disconformably over many of the

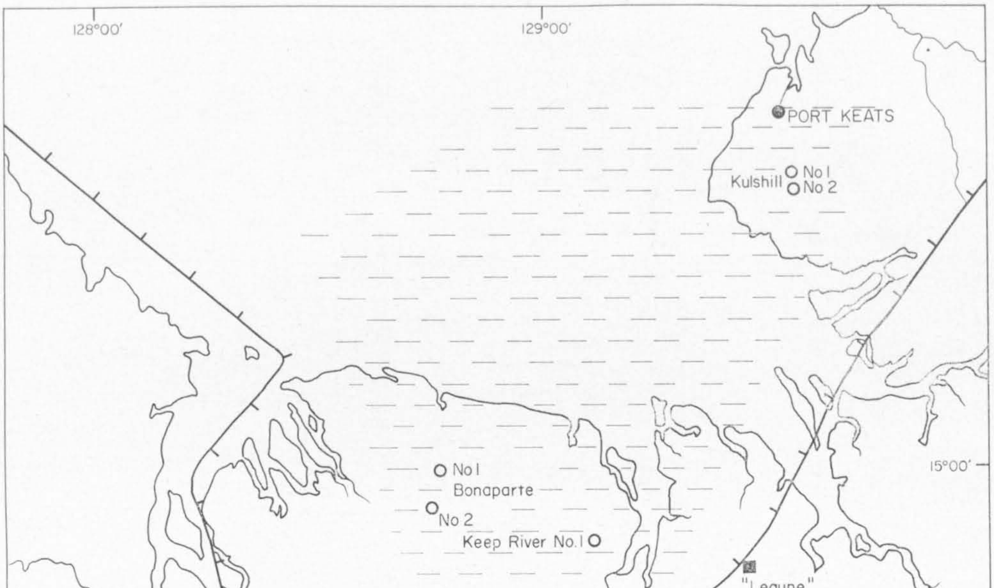

Fig. 16. Early Visean. Amended from figure 73 of Veevers & Roberts (1968).

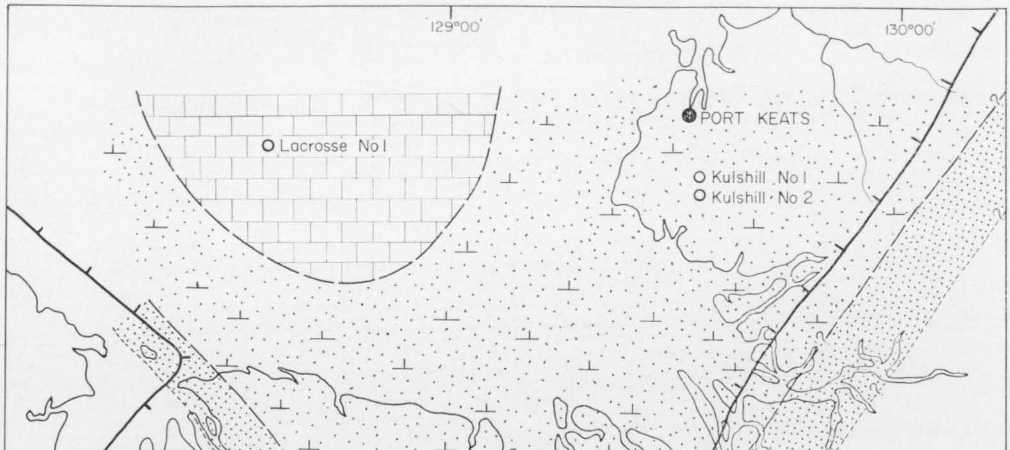

Fig. 17. Late Visean. Amended from figure 73 of Veevers & Roberts (1968).

marine early Carboniferous sediments in the onshore part of the basin (Fig. 18). Offshore in Lacrosse No. 1 Well, the disconformity between Upper and Lower Carboniferous sediments is delineated at 8 866 feet (2 702 m) by a change in dipmeter readings. The Upper Carboniferous sediments, which are unnamed, consist of 681 feet (209 m) of white to grey pyritic or calcareous quartz sandstone and interbedded siltstone and minor crinoidal dolomite and they overlie equivalents of the Tanmurra Formation. Balme (Appendix 1D *in* Arco, 1969, unpubl.), from a study of spores, considers the sediments to range in age from Namurian to Westphalian; possible brachiopod spines and a foraminifer from the Upper Carboniferous sediments suggest that they are marine (Lloyd, Appendix 1C *in* Arco, 1969, unpubl.). Stephanian rocks have not been identified in the Bonaparte Gulf Basin.

Fig. 18. Upper Carboniferous, showing the possible extension to the Border Creek Formation.

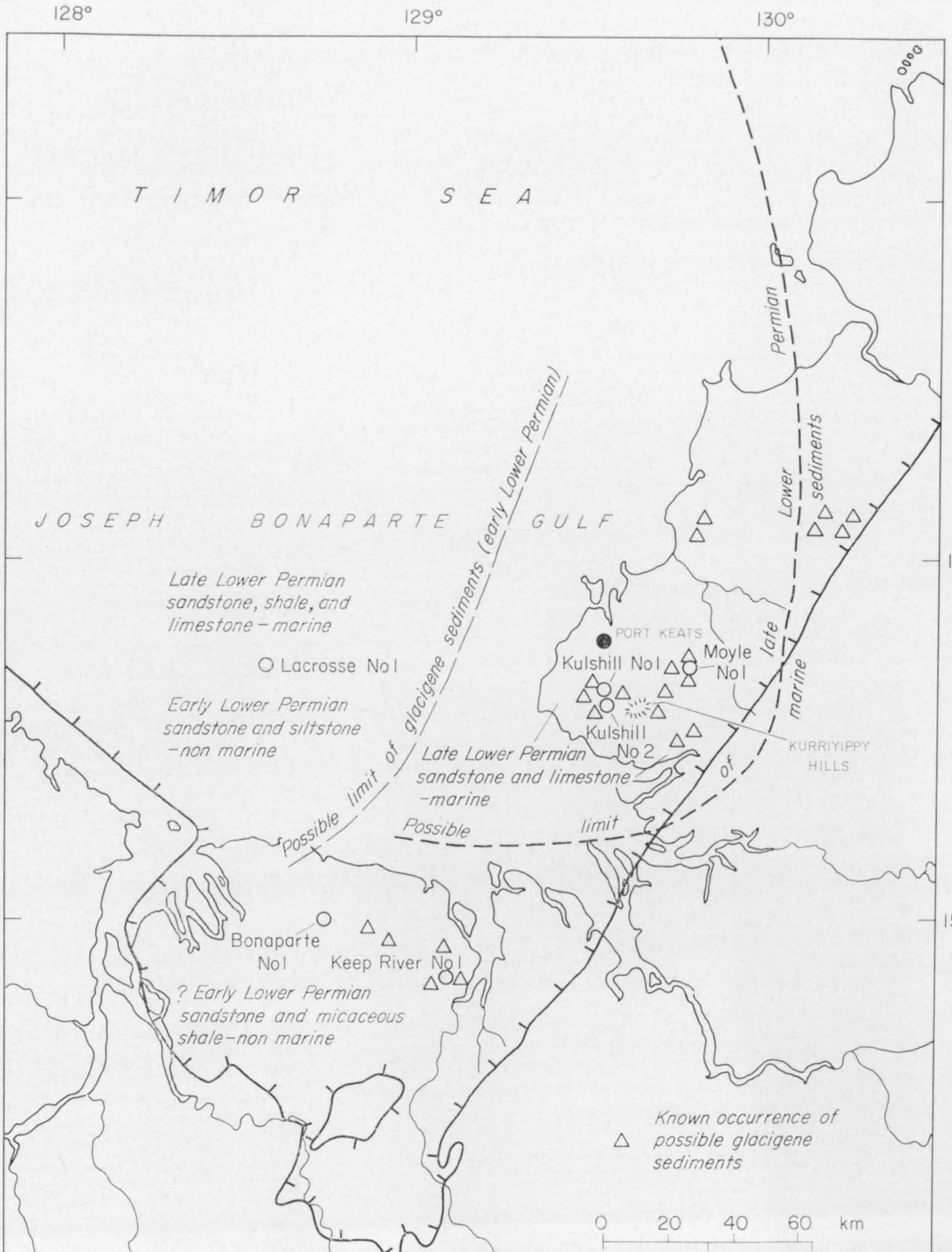

Fig. 19. Lower Permian.

Permian

A period of erosion, brought about by uplift and regression of the sea in the late Carboniferous, was followed by deposition of Lower Permian non-marine sediments on a low, and at times swampy, coastal plain or deltaic region (Dickins, Roberts, & Veevers, 1972). These conditions persisted until the late Lower Permian, when a marine transgression covered the plain (Fig. 19).

In some areas along the northeast margin of the basin possible Lower Permian sediments onlap the Precambrian basement with angular unconformity. Away from the Precambrian margin, for example near Port Keats, the Lower Permian disconformably overlies the Lower Carboniferous Tanmurra Formation, and at a greater distance from the margin in Lacrosse No. 1 Well it possibly disconformably overlies an unnamed Upper Carboniferous unit. The Lower Permian sediments consist of a lower unit up to 1 200 m thick consisting of silicified sandstone, probable tillitic rocks, and lithic sandstone and shale with minor coal horizons (*Kulshill Formation,* Caye, 1968); and at least 550 m of sandstone, shale, and limestone, the lower part of which is non-marine and coal-bearing, and the upper part is marine (*Sugarloaf Formation,* Guillaume, 1966). Possible tillitic sediments are present in several localities on the landward part of the basin, both in outcrop and in the subsurface; they are absent from Lacrosse No. 1 Well, and a tentative limit of their extent is given in Figure 19. Late Lower Permian marine sediments outcrop along the coast south of Port Keats in the Kurriyippi Hills and in the vicinity of Kulshill No. 2 Well.

A regression of the sea in the early Upper Permian reinstated the swampy environment in at least part of the northeast area (Fig. 20), but this was shortly followed by another marine transgression depositing sandstone and siltstone, as seen in the Port Keats area.

In Lacrosse No. 1 Well, Lower Permian units have a similar transition from non-marine to marine. The non-marine units consist of 1 536 feet (468 m) of light grey kaolinitic sandstone overlain by 1 387 feet (423 m) of grey argillaceous siltstone. The marine units are made up of 1 816 feet (554 m) of grey medium-grained quartz sandstone with occasional beds of limestone and dark shale. The Upper Permian, which according to Arco (1969, unpubl.) may unconformably overlie the Lower Permian, consists of 220 feet (67 m) of shale with minor sandstone and a limestone, 1 518 feet (463 m) of grey quartz sandstone with interbeds of shale, coal, and limestone, and 429 feet (131 m) of fine-grained sandstone, shale, and a conspicuous limestone.

Upper Permian sediments are also known from Arco Ltd-Australian Aquitaine Petroleum Petrel No. 1 Well drilled offshore 265 km west-southwest of Darwin (Fig. 20). They are marine, and between 11 366 and 13 040 feet (3 464 m and 3 975 m) consist of fine to medium-grained light grey sandstone, some of which is calcareous, and two fossiliferous limestone horizons. Both limestones can be traced from seismic records between Petrel No. 1 Well and Lacrosse No. 1 Well, and also over a wide area beneath the Timor Sea. Little is known of the section between 13 040 feet and 13 057 feet (3 975 m and 3 980 m) in Petrel No. 1 Well because of a blowout (Arco & Australian Aquitaine Petroleum, 1969).

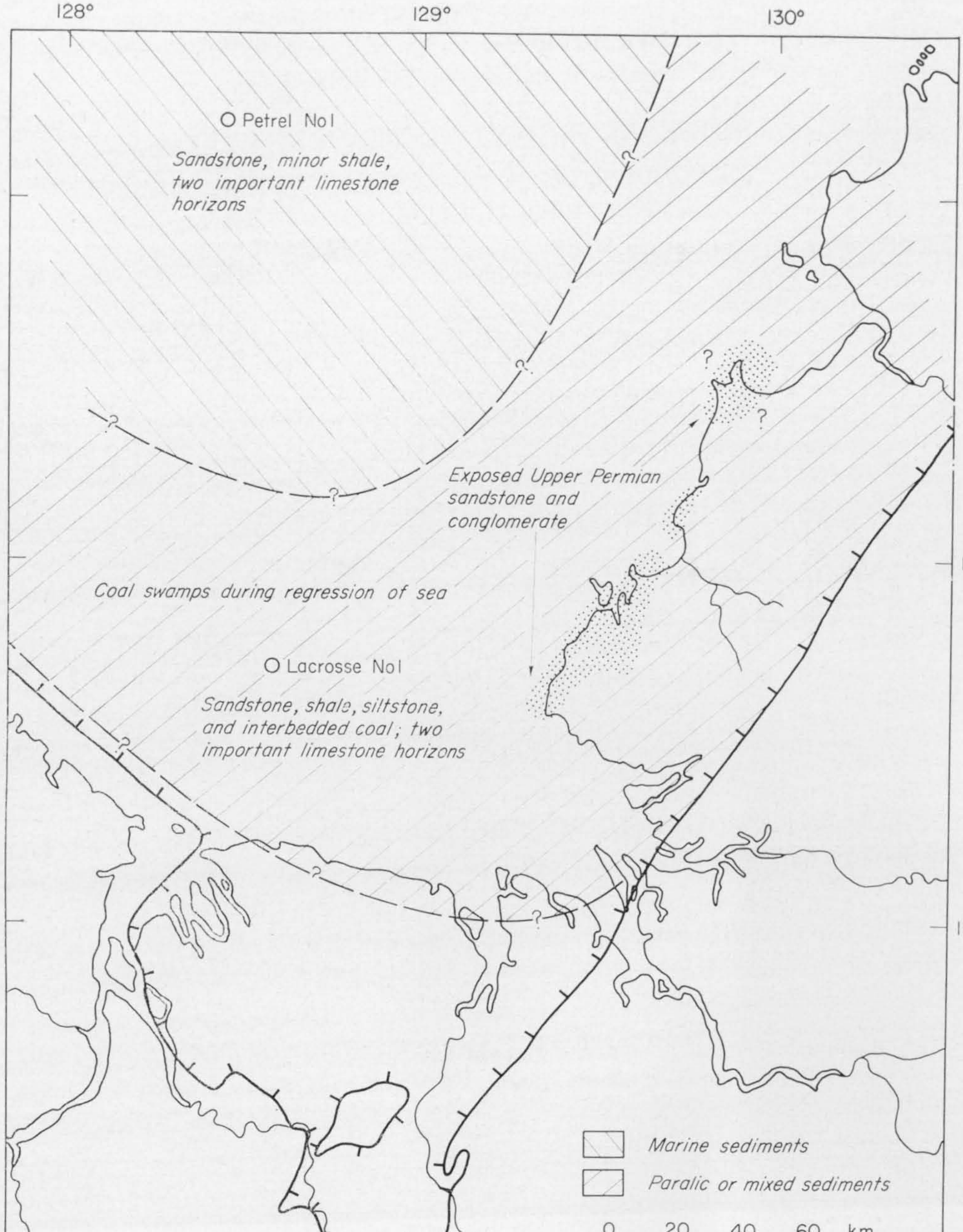

Fig. 20. Upper Permian.

TABLE 3. GENERALIZED DEPOSITIONAL AND STRUCTURAL HISTORY

Age	*Area*			
	Outliers	*Northwestern Area (Carlton Sub-Basin)*	*Southeastern Area (Burt Range Sub-Basin)*	*Port Keats*
CRETACEOUS				Marine transgression and widespread deposition of claystone
UPPER TRIASSIC				Uplift and erosion
LOWER TRIASSIC				Estuarine claystone and sandstone
UPPER PERMIAN		No record		Marine sandstone and siltstone. Non-marine sandstone, coal swamps. Uplift.
LOWER PERMIAN		Glacial erratics	No record	Marine sandstone and limestone. Non-marine glacigene sediments, coal swamps.
UPPER CARBONIFEROUS		Deposition of gravel, sand, silt by rivers. Uplift and erosion.		Uplift and erosion
LATE TOURNAISIAN-NAMURIAN	No record (probably land)	Subsidence, deposition of calcarenite, followed by littoral talus breccia and near-shore terrigenous sediments. Tournaisian-Viséan basinal shale.	Continued deposition of alternating terrigenous and carbonate (non-reef) sediments in Touraisian; Tournaisian-Viséan basinal shale	Viséan basinal shale, and calcareous sandstone. No record of Tournaisian sediments
MIDDLE TOURNAISIAN	Uplift	Uplift, tilting, erosion, and deposition of littoral talus breccia	Deposition of quartz sandstone	Uplift
UPPER DEVONIAN-EARLY LOWER CARBONIFEROUS (EARLY TOURNAISIAN)	Subsidence on faults along southeast margin; two cycles of terrigenous (sandstone) and carbonate (reef complex) deposition; continuous deposition in basinal area; probable uplift in late Frasnian of southwest part of the basin			No record of Tournaisian sediments. Upper Devonian variegated shale, sandstone, and shaley sandstone
MIDDLE ORDOVICIAN-MIDDLE DEVONIAN	Uplift and erosion			
MIDDLE CAMBRIAN-LOWER ORDOVICIAN	Subsidence and deposition on a shallow shelf of quartz sandstone and dolomite. Local breaks due to contemporaneous faulting		Non-deposition or deposition as elsewhere	
LATE LOWER CAMBRIAN	Differential uplift along faults, warping and erosion			
LOWER CAMBRIAN	Terrestrial deposition of basalt flows, agglomerate, and sandstone.			
LATE PRECAMBRIAN	Uplift and erosion.			

Mesozoic and Cainozoic

Late in the Permian the seas regressed from the northeast parts of the basin and Lower Triassic sediments were deposited in estuarine, brackish-water, or fresh-water environments. These sediments comprise laminated siltstone and fine-grained sandstone. No record of Jurassic sediments is found on the landward part of the basin. In the Cretaceous, marine and paralic fine sandstone and claystone were deposited over a wide area of northern Australia. Remnants of some Mesozoic rocks remain in the area around Port Keats in the northeast part of the basin. Offshore, there is a thick sequence of Mesozoic sediments exemplified by that in Petrel No. 1 Well (Fig. 15).

Except during a brief marine incursion in the Miocene (Lloyd, 1967) the onshore basin was dry during the Cainozoic, whereas deposition offshore, as indicated by B. O. C. Ashmore No. 1 Well (Fig. 4), was almost continuous.

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THE GEOLOGY OF THE CALVADOS CHAIN, SOUTHEASTERN PAPUA

by I. E. Smith

Abstract

The Calvados Island chain in the western part of the Louisiade Archipelago is composed mainly of low-grade schists which are thought to represent Mesozoic sediments metamorphosed during the Eocene. The schists are intruded by upper Tertiary basic and intermediate dykes. Miocene reef limestone and volcanics form the westernmost islands in the chain.

Introduction

The Calvados Chain is an elongate group of islands forming part of the Louisiade Archipelago in southeastern Papua (Fig. 1). The chain extends northwestward some 70 km from Nimoa Island (11°19′S, 153°15′E) to Panavaravara Island (11°7′S, 152°18′E) and is made up of about 38 small steep islands. The Tawa Tawa Mal reef with its scattered coral islands lies on the northern side of the chain; to the south only a discontinuous and partly submerged barrier reef lies between the Calvados Chain and the eastern Coral Sea.

Gibb Maitland (1892) recorded scattered geological observations from the western end of the Calvados Chain and Davies (1959) made geological observations on several of the islands in the eastern part of the chain. This paper presents a geology of the Calvados Chain and is based on field work carried out with P. E. Pieters (Geological Survey of Papua & New Guinea) during April 1969 (Smith & Pieters, 1969).

Geology

The dominant rock types in the Calvados Chain are low-grade schists of the Calvados Schist. They are intruded by dykes of intermediate composition and overlain by basic and intermediate volcanics. Shallow-water reef limestone forms some of the islands at the western end of the chain. Excellent exposures of unaltered rock occur around the coasts of almost all the islands, and wherever outcrop can be found in the dense bush which covers most of the islands the rocks are fresh. An exception is Hemenahei Island, which is relatively flat and low-lying, with a cover of sparse grassland and coastal sago swamp. The schists of Hemenahei Island are totally weathered to red and yellow clays, but the structures in the schists are commonly preserved.

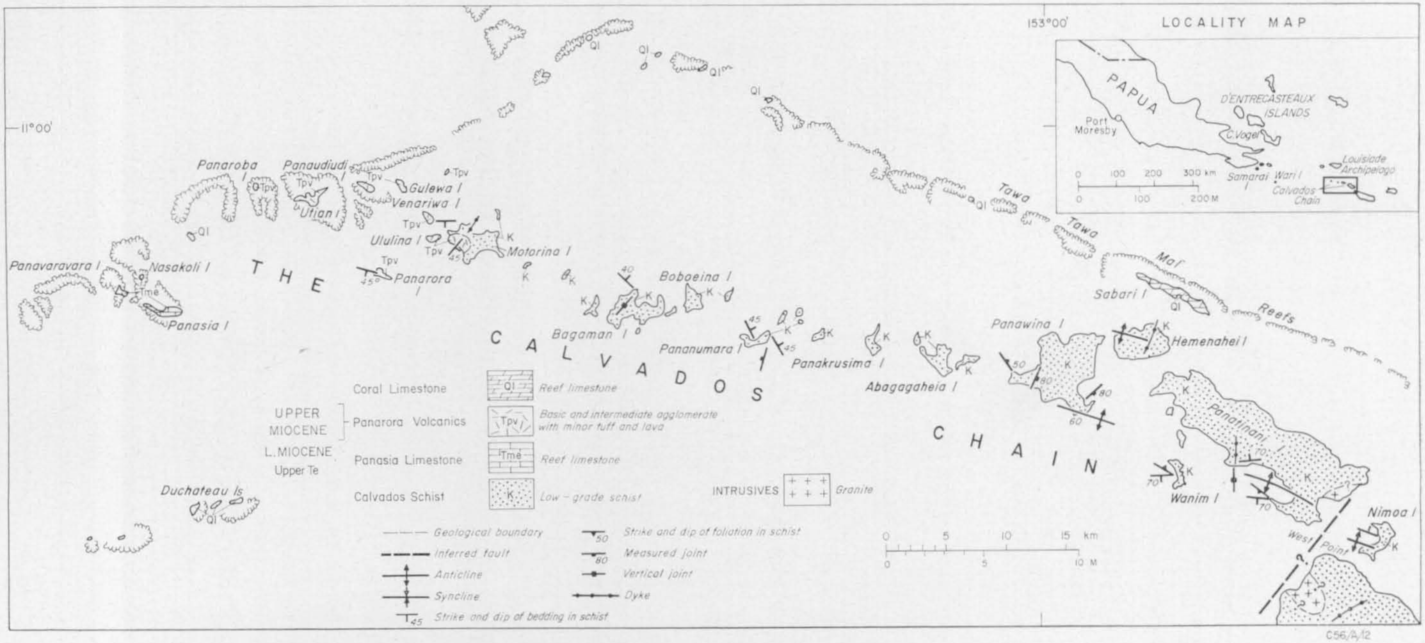

Fig. 1. Geological map of the Calvados Chain.

Calvados Schist

The name Calvados Schist was first used by Davies (1959) to describe the low-grade metamorphics in the Calvados Chain, on Sudest Island, and on eastern Misima Island. Calvados Schist was not used subsequently by de Keyser (1961) in his account of the geology of Misima Island, but Smith & Pieters (1969) used the name to describe low-grade schists in the Calvados Chain and on Rossel Island, Sudest Island, and the Renard Islands elsewhere in the Louisiade Archipelago.

The Calvados Schist consists of a well bedded series of pelitic siltstone, sandstone, and minor conglomerate which have been subjected to low-grade regional metamorphism. The maximum observed metamorphic grade is lowermost greenschist facies (quartz-albite-muscovite-chlorite subfacies of Turner & Verhoogen, 1960). The series is gently to moderately folded on a regional scale with measured dips ranging from 10° to 50°; fold axes generally strike east-west. Locally, tight isoclinal and recumbent folds of the order of 0.5 m to 2 m in amplitude have been observed. Primary sedimentary structures such as graded bedding and intraformational microslumping are preserved at some localities, but in others they are obscured by one or more metamorphic foliations.

In outcrop the schists generally have a poorly to well developed slaty cleavage or schistosity parallel to the original bedding. Where the bedding is on a fine scale the schists have a banded appearance. A second, and in some outcrops a third, metamorphic foliation may be superimposed on the primary metamorphic texture and thus gives rise to spectacular micro-folding (Pl. 11).

Figure 2 is a diagrammatic representation of metamorphic foliation measured at outcrops of the Calvados Schist throughout the Louisiade Archipelago. In the lower part of the diagram poles to metamorphic planes are plotted; in the upper part the corresponding strikes are plotted on a rose diagram as a percentage of the total. Principal trends are between 255° and 355°.

The age of the Calvados Schist is unknown. On the eastern Papuan mainland the Owen Stanley Metamorphics are thought to be late Mesozoic rocks metamorphosed during the Eocene (Davies & Smith, 1970), and the Calvados Schist is possibly the same.

There is no evidence for a Mesozoic source of sediment near the present-day position of the Louisiade Archipelago. However, if, as some authors suggest, the Coral Sea Basin has opened up since the beginning of the Tertiary (see e.g. Ewing, Hawkins, & Ludwig, 1970; Gardner, 1970), the sediments which formed the Calvados Schist could have been laid down in a basin adjacent to a Mesozoic landmass in the position of present-day northern Queensland.

Petrography. Texture in specimens from the Calvados Schist varies from fine-grained clastic with little sign of metamorphism to strongly schistose with platy minerals concentrated in subparallel planes. The banding and folding observed in outcrop are commonly visible on a microscopic scale.

The most common mineral assemblage is quartz, albite, sericite (muscovite), and chlorite, with minor epidote, zoisite, sphene, prehnite, and various amounts of opaque minerals. Small detrital grains of plagioclase, pyroxene, and hornblende

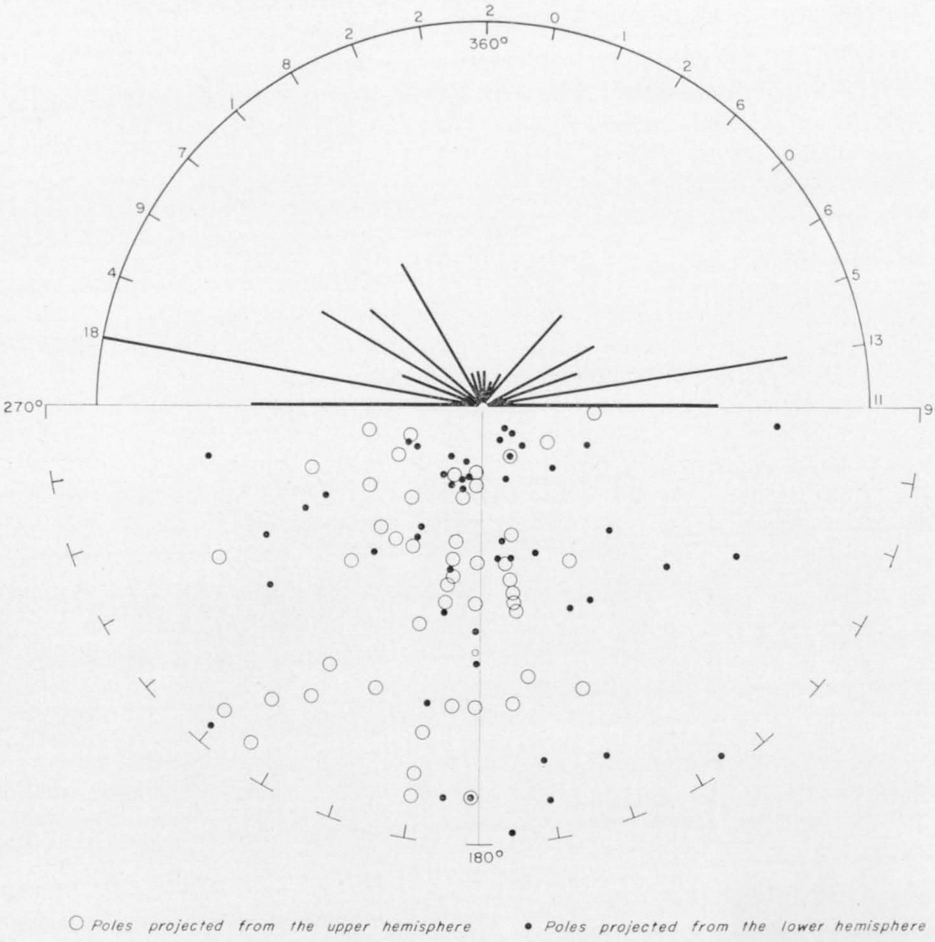

Fig. 2. Metamorphic foliation measured in the Calvados Schist. In the lower half of the diagram the poles of all the bedding and metamorphic planes are plotted on a stereographic projection. In the upper half of the diagram the corresponding strikes are represented as a percentage of the total.

are very minor constituents of some specimens. Quartz, albite, epidote, and zoisite occur as porphyroblasts up to 6 mm across in some of the schists. Calcite, albite, prehnite, and epidote are common vein minerals.

Panasia Limestone

The Panasia Limestone is a lower Miocene reef deposit which forms Panasia, Panavaravara, and Nasakoli Islands at the western end of the Calvados Chain. The limestone is medium to fine-grained and creamy yellow on broken surfaces; it consists of abundant microfossil tests in a matrix of fine-grained calcite. In outcrop the limestone has grey weathered surfaces and well developed vertical fluting.

A sample collected from Panasia Island (6952.3690) contained the Foraminifera *Lepidocyclina* (*Eulepidina*) spp., *L.* (*Nephrolepidina*) spp., *Miogypsinoides* sp. cf. *M. dehaarti, Amphistegina* sp., *Cycloclypeus* sp., and *Gypsina* sp., which indicate a lower Miocene (upper Te) age (D. J. Belford, pers. comm.). The limestone was apparently deposited in a shallow-water reef environment and shows that at the western end of the Calvados Chain there were areas of shallow water with no appreciable terrigenous sedimentation during the lower Miocene.

Panarora Volcanics

Volcanic rocks occur on Panarora, Utian, Panaudiudi, Gulewa, Tobaium, Venariwa, Ululina, Moturina, and Panaroba Islands, near the western end of the Calvados Chain. Lavas and consolidated ash are recorded from Utian Island (Gibb Maitland, 1892), but elsewhere the volcanics consist almost entirely of bedded volcanic agglomerate and minor tuff. The volcanics are named from Panarora Island where over 200 m of bedded, coarse, unsorted agglomerate dips 45° to the south. On Moturina Island a sheet of agglomerate dips 40° to 50° southeastward and overlies schists and intermediate intrusives. On Venariwa Island, massive, medium to coarse, moderately well sorted volcanogenic conglomerate occurs as beds 1-10 m thick, dipping 25° to the south. The conglomerate on Venariwa Island is more compacted and appears older than that observed elsewhere; it possibly represents an early phase in the volcanic activity of the area.

Total rock K-Ar dating of a sample of pyroxene andesite (6952.3689E) from the agglomerate on Panarora Island gave an age of 11.0 to 11.4 million years, which indicates that the Panarora Volcanics are middle Miocene or lower upper Miocene. This is older than the Pliocene or Pleistocene age previously guessed for the unit by Smith & Pieters (1969). Details of the K-Ar age determination are given in Table 1.

TABLE 1. TOTAL ROCK K/AR DATING OF SPECIMEN 6952.3689E, A PYROXENE ANDESITE FROM PANARORA ISLAND*

% K	*Radiogenic Ar^{40}*	*Radiogenic Ar^{40}/K^{40}*	*% atmospheric Ar^{40}*	*Age (m.y.)*
1.639	3.3402×10^{-11}	0.0006603	34.4	11.4±0.3
1.641	3.2134×10^{-11} (moles/g.)	0.0006439	40.3	11.0±0.3

* Dating carried out at the Australian Mineral Development Laboratories (AMDL), analyst A. W. Webb.

Petrography. On Panarora Island the agglomerate clasts include a variety of pyroxene and hornblende andesites. Specimens representative of the main types within the agglomerate are:

Vesicular pyroxene andesite (Specimen 6952.3689B): A porphyritic rock with dark pyroxene phenocrysts and vesicles in a deep red fine-grained groundmass. It consists of euhedral augite, andesine, minor hypersthene, and magnetite in a fine-grained red groundmass containing microlites of augite and plagioclase. Rare pseudomorphs of fine-grained secondary material may represent former olivine crystals.

Pyroxene andesite (Specimens 6952.3689A and E): medium to dark grey, dense, fine-grained rocks with small dark mafic crystals. The rocks consist of andesine, augite, minor hypersthene, and magnetite in a fine-grained plagioclase groundmass. Texture is slightly porphyritic and the plagioclase phenocrysts are oriented to form a weakly developed flow texture. Pseudomorphs of iddingsite after olivine and light brown hornblende crystals surrounded by an opaque reaction rim are rare constituents.

Hornblende andesite (Specimen 6952.3689C and D): medium grey with a patchy purple hue, lighter in colour than the pyroxene andesites. It is porphyritic with conspicuous dark mafic phenocrysts in a lighter groundmass. Euhedral brown hornblende and augite occur as phenocrysts in a fine-grained groundmass of sodic labradorite, minor hypersthene, brown biotite, augite, hornblende, and magnetite. In one specimen (6952.3689D) the groundmass contains minor glass.

Specimens from the *volcanogenic conglomerate* on Venariwa Island and the *agglomerate* on Moturina Island are fine-grained and have an altered appearance. In thin section they are either porphyritic or fine and even-grained and consist of augite, plagioclase, magnetite, and green interstitial mesostasis. Some specimens contain iddingsite after olivine. In the porphyritic varieties the phenocrysts are augite set in a fine-grained groundmass. These rocks are basaltic rather than andesitic in character.

Coral Reefs and Alluvium

Coral reefs are only locally developed around the islands of the Calvados Chain; this is due in part to the steep submarine slopes of the islands and in part to the sheltering effect of the barrier reef to the north. Reefs are more extensive at the western end of the chain where it meets the Tawa Tawa Mal barrier reef from the north. Small coral islets occur along the Tawa Tawa Mal reef; the most substantial of these is Sabari Island, which is composed entirely of coral limestone and rises to 55 m above sea level.

Because of the steep hilly nature and the small size of the islands in the Calvados Chain, areas of alluvium are limited to small coastal flats and beaches.

Intrusive Rocks

Microdiorite, porphyritic microdiorite, and andesite dykes intrude the schists in the Calvados Chain, especially those of Moturina Island near the western end of the chain. Some of these may be related to the Panarora Volcanics. A small outcrop of granite has been reported from the eastern end of Pana Tinani Island (W. Manser, pers. comm.).

Petrography. Diorite is medium-grained, grey, granular, and generally contains conspicuous dark mafic crystals. It is composed of slightly pleochroic (light green to brown) hornblende, plagioclase (oligoclase to andesine), and minor opaque minerals.

Microdiorite occurs as dykes on Panapompom Island and Moturina Island, and probably elsewhere in the Calvados Chain. It is generally medium to fine-grained with dark mafic crystals in a medium to dark grey matrix. Plagioclase (andesine to labradorite) is accompanied by pale green-brown slightly pleochroic hornblende and minor opaque minerals. Minor brown biotite and augite occur in some specimens and orthoclase is usually present. Chlorite, calcite, and fine-grained secondary material are common in some altered specimens. Textures are fine-grained and may be slightly porphyritic.

Porphyritic microdiorite has been observed as dykes intruding the schists on Pana Tinani Island and Moturina Island; it probably also occurs elsewhere in the Calvados Chain. The rocks are porphyritic, with dark mafic crystals in a fine-grained medium to dark grey matrix. In thin section the texture is strongly porphyritic, with euhedral phenocrysts of pale brown-green, slightly pleochroic hornblende and less commonly euhedral pseudomorphs of sericite after feldspar. The groundmass consists of hornblende and plagioclase with secondary fine-grained alteration products and calcite. Small augite phenocrysts are a minor constituent in a specimen from Pana Tinani Island.

A dyke of strongly *porphyritic andesite* intrudes schists on Moturina Island. It consists of phenocrysts of zoned, euhedral andesine up to 3 mm across, with smaller pale green hornblende and brown biotite phenocrysts (1-5 mm across) in a fine-grained groundmass of feldspar and mafics.

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Calvados Schist : Kink folding caused by deformation of the schistosity (dipping right) by a secondary metamorphic foliation (dipping left) (Wanim Island)

AN UPLIFTED WAVE-CUT TERRACE ON SUDEST ISLAND, SOUTHEASTERN PAPUA

by I. E. Smith

Abstract

The central mountain range of Sudest Island is flanked by an inclined platform of low relief and intricate drainage. This platform is interpreted as a wave-cut terrace, although no littoral or reef deposits have been found on its surface and such a terrace could not have developed behind the barrier reef surrounding the island at the present time. The presence of a wave-cut terrace on Sudest Island provides evidence for Recent uplift of at least 120 m.

Introduction

Sudest Island is a narrow island 70 km long (northwest) and up to 20 km wide; it is the largest island in the Louisiade Archipelago (Fig. 1). The island is composed almost entirely of low-grade mica schist with minor intrusives (Smith & Pieters, 1969; Smith, this vol.). An unusual feature of the topography of Sudest Island is the presence of a gently sloping area of low relief which surrounds the central mountain range. In this paper the topography of Sudest Island is described and the origin of the area of low relief is discussed. The work is based on field work carried out in early 1969 as a part of the regional geological mapping of the Louisiade Archipelago.

Physiography

A mountain range runs two-thirds of the length of Sudest Island from its northwestern end. The range has steep youthful topography and rises to 810 m above sea level at Mount Riu. The range is flanked by an inclined platform of low relief which slopes from a maximum elevation inland of about 120 m to elevations of 10-15 m at the coast. At the coast there is commonly an abrupt drop from the edge of the platform to sea level. The platform is most extensive at the southeastern end of the island, where it has been named the Siri Peneplain (French, 1966), and along the northern side of the mountains, where Smith & Pieters (1969) named it the Feori level to the west and the Manbari level to the east (Fig. 1). The main characteristics of the platform are a gently sloping surface, intricate deeply incised drainage pattern, and deep weathering of the underlying schists; these features contrast with the rugged topography, simple dendritic drainage, and fresh rock outcrops of the central mountains.

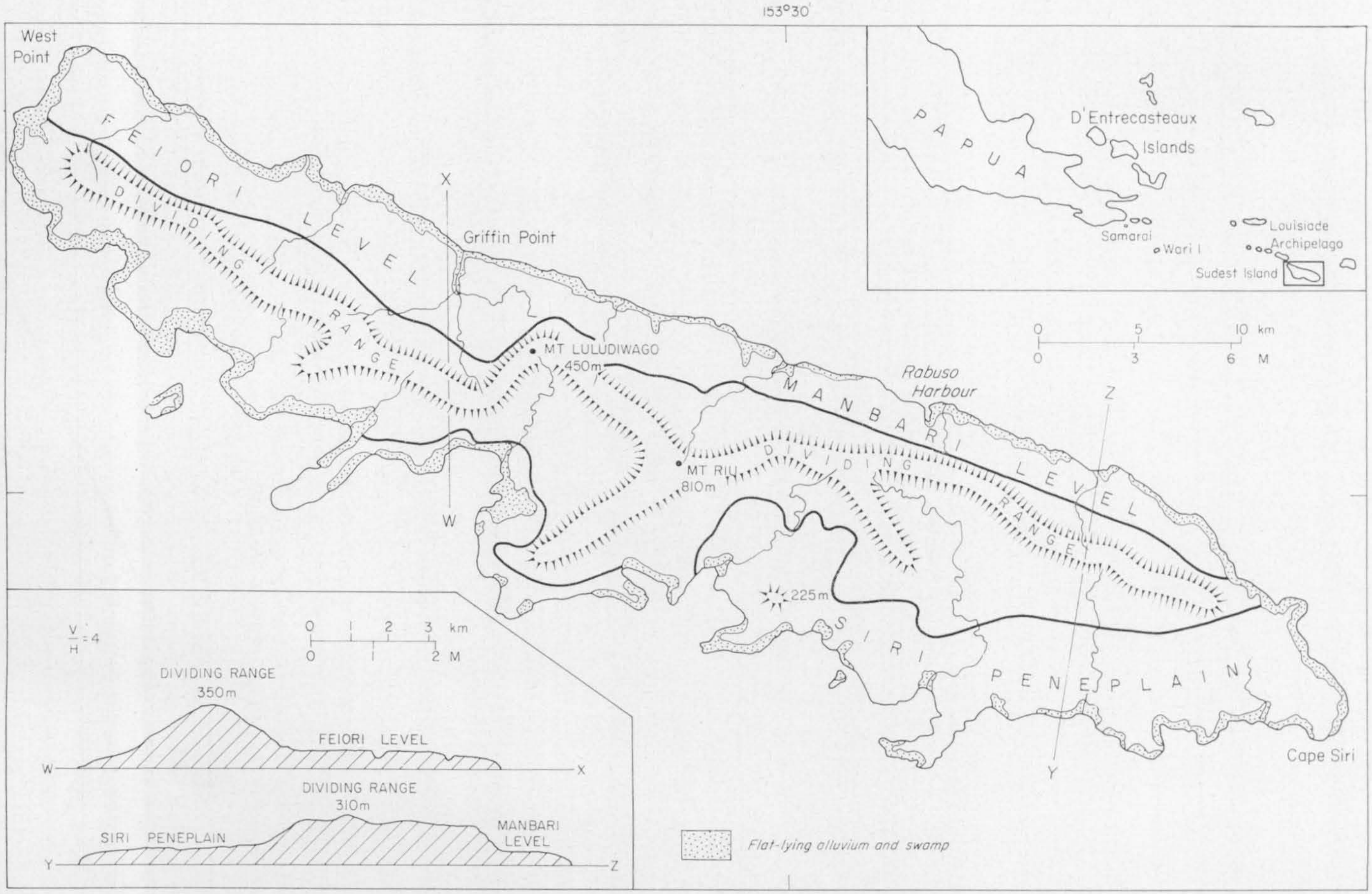

Fig. 1. The geomorphology of Sudest Island.

There is an abrupt change in slope between the central mountains and the platform (profile, Fig. 1). Streams which are swiftly flowing in the central mountains change character on the platform and have long, slowly flowing, reaches separated by short rapids or cataracts. Most of the major streams on the platform are deeply entrenched. The boundary line between the platform and the mountain range is sinuous and follows up the valleys of the major rivers.

Discussion

The platform on Sudest Island cannot be explained as a constructional feature: fault control is ruled out by the sinuous boundary with the mountain range, and field observations show that it is not due to lithological changes.

Smith & Pieters (1969) suggested that the platform is a wave-cut terrace. If this is the case it is puzzling that no littoral or reef deposits have been observed on the surface of the platform. This can be explained if they were stripped from the surface after a period of uplift, but it is also possible that littoral or reef deposits have been overlooked during field work on the island.

Sudest Island is at present surrounded by an extensive barrier reef; it is certain that an extensive wave-cut platform could not develop behind such a reef, and I suggest that at the time the platform developed the environment was unfavourable to reef-building organisms.

If we accept that the platform on Sudest Island is a wave-cut terrace it provides some evidence of Recent tectonic movements in the area. The difference in elevation of over 100 m between coastal and inland parts of the platform suggests that the island was sinking while the terrace was being cut. Since then, there has been an uplift of at least 120 m to raise the terrace to its present elevation. There is no evidence of comparable Recent uplift on immediately adjacent islands, but on Misima Island, 100 km to the northwest, a succession of raised coral platforms provides evidence for uplifts of up to 460 m (de Keyser, 1961).

References

French, D. J., 1966—The geology and geochemistry of Sudest Island. *Report to the Senior Resident Geologist, Port Moresby* (unpubl.).

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THE GEOLOGY OF THE DEBOYNE ISLAND GROUP, SOUTHEASTERN PAPUA

by I. E. SMITH and P. E. PIETERS*

ABSTRACT

Panaete, Panapompom, and Nivani Islands, in the Deboyne Group at the western end of the Louisiade Archipelago, consist of igneous and metamorphic rocks, some of which may be as old as Mesozoic. The basement rocks are metabasalt, chlorite schist, gabbro, and minor chert; they are intruded by microdiorite dykes. Coral limestone is extensively developed on Panaete Island and provides evidence for Recent uplifts of 3-4 m.

INTRODUCTION

The Louisiade Archipelago extends southeastward from the Papuan mainland to longitude 154°25′E. The archipelago contains numerous islands, islets, and extensive coral reefs, encompassing a total area of about 2 400 km^2. The Deboyne, Conflict, and Torlesse Island Groups at the western end of the archipelago are mainly low-lying coral islands, but on the three largest islands in the Deboyne Group there are outcrops of igneous and metamorphic basement, some of which may be as old as Mesozoic. This paper gives an account of the geology of the Deboyne Group and is based on field work done in early 1969 (Smith & Pieters, 1969).

PHYSIOGRAPHY

The Deboyne Group (10°47′S, 152°20′E) consists of larger islands, Panaete, Panapompom, and Nivani, and a number of small coral islands (Fig. 1). Panaete Island, the largest, lies in the north of the group, and from it reefs extend to the south and southeast to encircle Panapompom Island and Nivani Island. The coral islets occur mainly on the northeastern part of the encircling reef.

Panaete Island is about 12 km^2 in area and is mainly flat-lying with an average elevation of 3-4 m. A small conical hill at the western end of the island rises to approximately 260 m. Panapompom and Nivani Islands, although smaller (4 km^2 and 0.5 km^2 respectively), are steep and hilly. Panapompom Island has a number of hills between 130 m and 170 m in elevation and Nivani Island is a single hill rising to approximately 100 m.

* Geological Survey of Papua New Guinea.

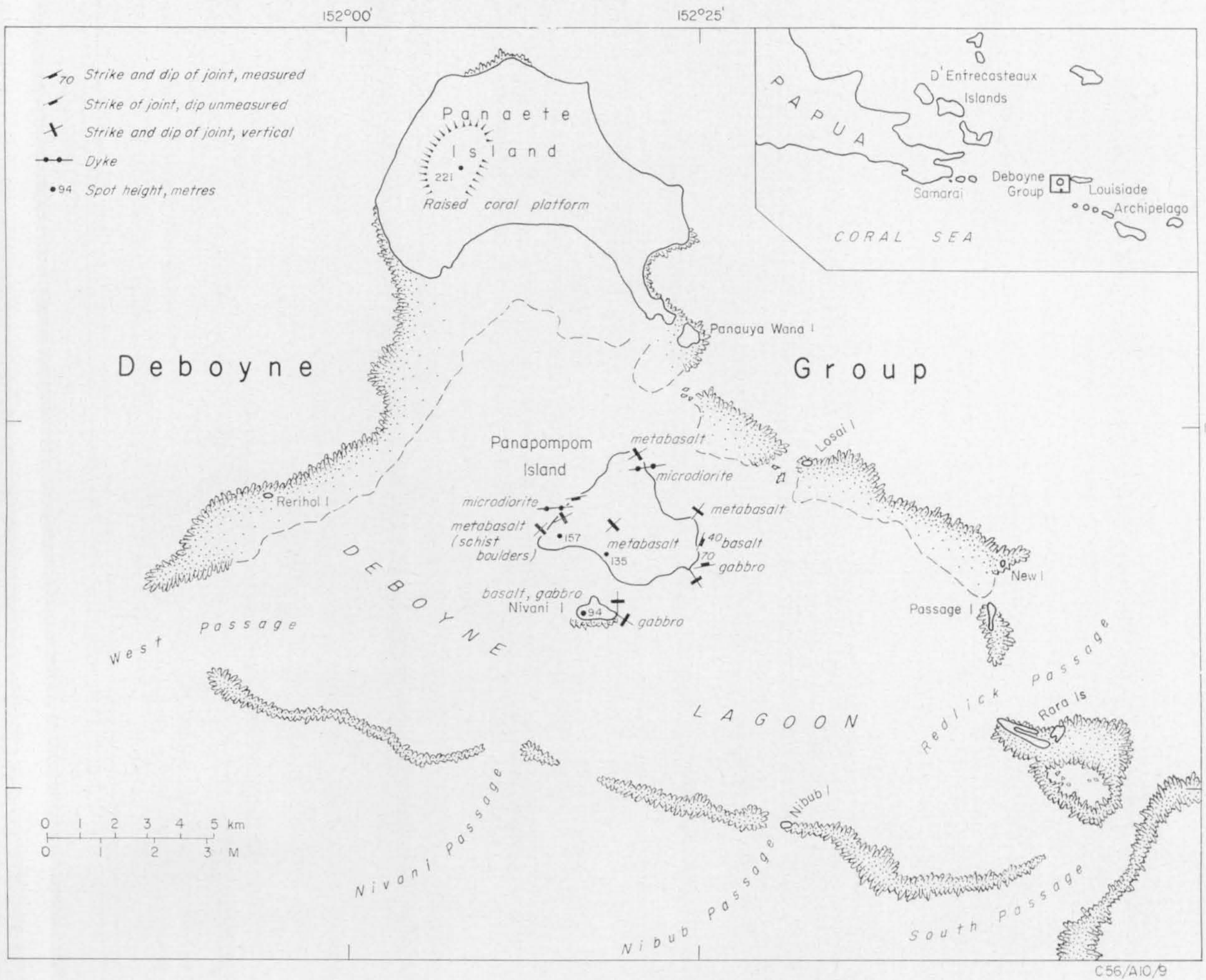

Fig. 1. Geological observations in the Deboyne Island Group.

Geology

The basement rocks in the Deboyne Group are metabasalt, chlorite schist, gabbro, and minor chert; they are intruded by microdiorite dykes on Panapompom Island. Coral limestone is extensively developed on Panaete Island.

Panaete Island consists mainly of coral limestone forming a raised platform 3-4 m in elevation. From the occurrence on the beach of pebbles of basic crystalline schist, Gibb Maitland (1892) concluded that the interior of the island was composed of schistose rocks. These rocks probably make up the hill at the western end of the island and are probably comparable with the basic schists and metabasalts of Panapompom Island.

Panapompom Island is made up of gabbro and metabasalt with subordinate chlorite schist, chert, and microdiorite. The dominant outcrop at the western end of the island is metabasalt, which is intruded by narrow dykes of microdiorite at the western and northern points of the island. Chlorite schist was collected from beach boulders on the southwestern side of the island. The fine-grained chert containing dark siliceous lenses, which was mentioned by Davies (1959), is probably associated with the metabasalt. Gabbro crops out mainly in the eastern part of the island; it is commonly altered and veined with calcite.

Jointing in outcrops of metabasalt and gabbro is steeply dipping to vertical and strikes generally northeastward; a second, less well developed joint set, observed at two localities, strikes southeastward. The dykes at the western end of the island strike 080° and are near vertical.

Nivani Island consists of gabbro. Outcrops at the eastern end of the island have a principal vertical joint set striking 310° and a secondary vertical joint set striking 360°.

Petrography

The chlorite schist is fine-grained, green-grey, and slightly schistose. In thin section the texture is weakly schistose; bands of very fine-grained interlocking quartz and albite crystals alternate with bands consisting mainly of fine-grained (0.1 mm average) chlorite and epidote. Chloritoid forms lenses and bands oriented parallel to the schistosity.

The metabasalt is uniformly dark grey to black and fine-grained. The texture shows only very weak orientation, which is not seen in hand specimens. Metamorphism has produced a retrogressive assemblage of brown-green amphibole, albite, epidote, and fine-grained unresolvable material. Relict augite is present in some specimens.

The gabbro is medium-grained and medium to dark grey; when altered it takes on a greenish colour. It consists of a primary assemblage of augite, labradorite, and opaque oxides. Augite forms larger crystals (2 mm) subophitically intergrown with smaller (1 mm) intergranular plagioclase. The augite is commonly extensively altered to amphibole and the plagioclase to sericite. Albite, epidote, and prehnite occur in veins and as discrete crystals.

Microdiorite is fine-grained (1-2 mm average) with equigranular texture and consists of plagioclase (oligoclase-andesine), pale brown hornblende, and subordinate opaque oxides.

Discussion

The metabasalt and schist in the Deboyne Islands are thought to be late Mesozoic on the basis of a correlation with rocks of similar metamorphic grade on the Papuan mainland.

Although the gabbros in the Deboyne Islands have suffered some retrogressive changes in mineralogy their textures are unaffected, and they are thought to have been emplaced after the metamorphic event which affected the metabasalts.

The coincidence of the strikes of the microdiorite dykes and jointing directions on Panapompom Island suggests that the emplacement of the dykes was linked to an episode of regional tectonic stress. Dykes of similar composition occur on Misima Island (de Keyser, 1961) and in the western Calvados Chain (Smith & Pieters, 1969), where they are thought to have been emplaced during the late Tertiary.

The presence of the raised coral platform on Panaete Island provides evidence for Recent uplifts of at least 3-4 m. This contrasts strongly with evidence of Recent uplifts of over 460 m on nearby Misima Island (de Keyser, 1961).

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NOTES TO ACCOMPANY A GEOLOGICAL MAP OF ROSSEL ISLAND, SOUTHEASTERN PAPUA

by I. E. Smith, P. E. Pieters*, and C. J. Simpson

Abstract

Rossel Island, at the eastern end of the Louisiade Archipelago, is made up of a metamorphic basement intruded by gabbro, dolerite, and minor microgabbro. The metamorphic basement consists of low-grade schists forming an anticlinal structure flanked by foliated gabbro and basalt. At the western end of the island foliated ultramafic intrusives form a sill intruding the schists. Structural trends on the island strike approximately east-west.

Introduction

Rossel Island (11°20′S, 154°10′E) lies at the eastern end of the Louisiade Archipelago in southeastern Papua (Fig. 1). It is one of the larger islands in the archipelago (about 300 km^2) and is mountainous and heavily forested; the highest point on the island is Mount Rossel, which rises to 850 m above sea level. European settlement is confined to a Catholic mission on the northern side and a small plantation on the southern side of the island.

The geological map of Rossel Island presented here (Fig. 1) is based on a photo-scale topographic map prepared by radial line plot from air-photographs taken during World War II. The geology was compiled at photo-scale from field observations and airphoto interpretation.

Geology

Calvados Schist

The main rock unit on Rossel Island is the Calvados Schist, named from the Calvados Chain in the western part of the Louisiade Archipelago. A more detailed description of the unit is presented elsewhere (Smith, this vol., pp. 58-64), and only the main features are given here.

The Calvados Schist consists of well bedded pelitic siltstone, sandstone, and minor conglomerate which have been subjected to low-grade regional metamorphism. The rocks have poorly to well developed slaty cleavage or schistosity parallel to the original bedding. A secondary and in places a third metamorphic foliation has been observed superimposed on the primary metamorphic texture.

* Geological Survey of Papua New Guinea.

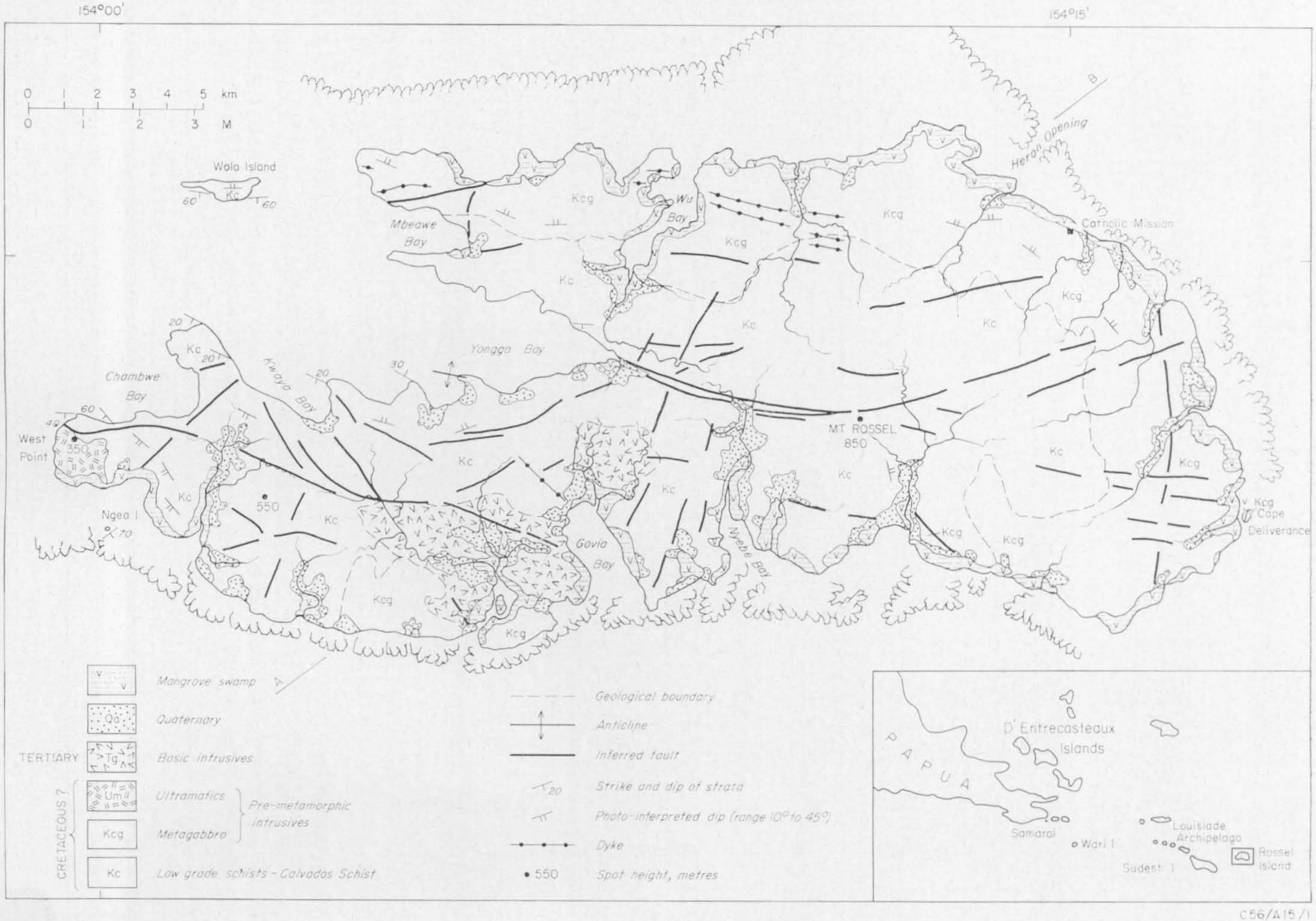

Fig. 1. Locality map and geological map of Rossel Island.

The schist is usually fine-grained; the most common mineral assemblage is quartz-albite-sericite (muscovite) and chlorite with minor epidote, zoisite, sphene, prehnite, and varying amounts of opaque oxides. Small detrital grains of plagioclase, pyroxene, and hornblende are minor constituents of some specimens. Quartz, albite, epidote, and zoisite may occur as porphyroblasts up to 6 mm across. Calcite, albite, prehnite, and epidote are common vein minerals.

The Calvados Schist is gently to moderately folded on a regional scale with measured dips ranging from 10° to 50°. Tight isoclinal and recumbent folds of 0.5-2 m amplitude are developed in places.

Pre-metamorphic Intrusives

Foliated ultramafic (pyroxenite and serpentinite) and mafic (metagabbro and basalt) rocks crop out on Rossel Island. They are thought to have intruded the Calvados Schist before metamorphism.

Ultramafic. Pyroxenite and *serpentinite* form a sheet approximately 150 m thick, which conformably overlies low-grade schists at the southwestern tip of Rossel Island. The joint planes in the ultramafics are subparallel to bedding and schistosity planes in the schist. The ultramafics are intruded by minor basalt and diorite near West Point.

The pyroxenite is medium to coarse-grained, granular and grey-green when fresh, tending toward green-grey with dark mafics in a light matrix when altered. Serpentinite is typically dark green to dark grey and fine-grained and commonly has a greasy lustre. The pyroxenite consists of anhedral augite crystals up to 3 mm across, with interstitial fibrous secondary tremolite and some fine-grained secondary vein material. The augite commonly shows strain effects such as bent cleavage planes and strain twinning; it is altered to varying degrees. In highly altered specimens only a little augite remains in a contorted matrix of tremolite with fine-grained secondary material, including albite and vein serpentine. Shearing is common.

Mafic. Foliated *metagabbro* and *basalt* form a discontinuous strip along the northern, eastern, and southern coasts of Rossel Island. The rocks are weakly to moderately foliated and are thought to have been intruded into the sedimentary sequence (Calvados Schist) before or at an early stage of metamorphism.

Hand specimens are grey-green, medium to fine-grained, and usually have a foliated texture. Augite, plagioclase, and opaque oxides are the primary minerals, but they have been substantially or completely altered to sericite, brown-green hornblende, tremolite-actinolite, chlorite, albite, and fine-grained unresolvable alteration products. Epidote, prehnite, and calcite are common secondary and vein minerals. Quartz occurs in variable amounts and is probably mainly secondary.

Post-metamorphic Intrusives

Gabbro and *dolerite* are exposed around Govia Bay on the south coast of Rossel Island. On the air-photographs these rocks show up as relatively depressed areas with an intricate drainage pattern which contrasts with the steep ridges and simple drainage pattern of the Calvados Schist.

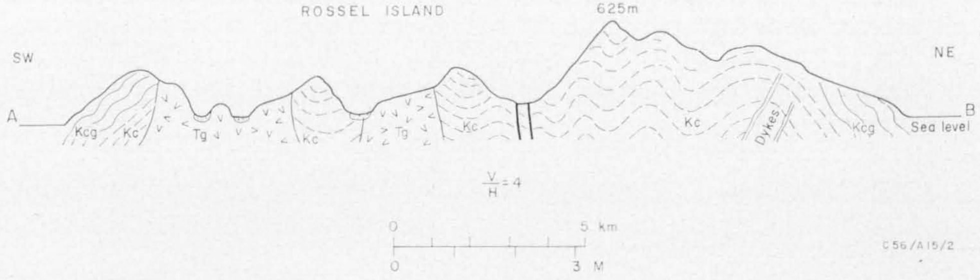

Fig. 2. Geological section.

The basic intrusives are fine to coarse-grained, medium to dark grey, with a granular texture. They consist of augite, plagioclase, and opaque oxides. The augite is generally partly or completely altered to sericite and fine-grained alteration products. Secondary albite and fine-grained interstitial mesostasis are common. Less altered specimens show a texture which is subophitic in part, and contain plagioclase tentatively determined as calcic andesine.

The diorite which intrudes the ultramafics at West Point is a medium-grained, grey, granular rock containing dark mafic crystals. It is composed of slightly pleochroic light green to brown hornblende, plagioclase (oligoclase to andesine), and minor opaque oxides. Epidote and albite are common vein minerals.

Structure

On Rossel Island the Calvados Schist apparently forms the core of an anticlinal structure flanked on either side by pre-metamorphic or early metamorphic foliated metagabbro and metavolcanics (Fig. 2). Structural trends, as interpreted from air-photographs and illustrated on the geological map, strike approximately east-west. They parallel the structural trends observed in the schists of the Calvados Chain (Smith & Pieters, 1969) but lie at an angle to the overall west-northwesterly trend of the Louisiade Archipelago as a whole.

Reference

Smith, I. E., and Pieters, P. E., 1969—The geology of the Louisiade Archipelago, T.P.N.G., excluding Misima Island. *Bur. Miner. Resour. Aust. Rec.* 1969/93 (unpubl.).

STRATIGRAPHY AND STRUCTURE OF THE CONTINENTAL MARGIN BETWEEN NORTH WEST CAPE AND SERINGAPATAM REEF, NORTHWEST AUSTRALIA

by J. J. VEEVERS[1]

SUMMARY

Shallow seismic profiles from the 1968 BMR marine geophysical survey of the southern Sahul Shelf, entire Rowley Shelf, adjacent North Australian Basin, and part of the Exmouth Plateau were made with a Spark-array source and a Subot Hydro-streamer receiver along 15 000 km of traverse. The profiles were interpreted in terms of the Cainozoic stratigraphic section penetrated in five offshore wells, and from their continuity with previously described profiles in the adjacent Timor Sea.

A wedge of sediments, dominantly carbonates, was built upward and outward during the Cainozoic. This wedge contains two prominent reflectors that are interpreted as unconformities representing the Oligocene and late Miocene. The unconformities crop out along the faulted upper continental slope that faces the North Australian Basin. Southward, adjacent to the Exmouth Plateau, the late Miocene unconformity crops out again along the upper continental slope, but the Oligocene unconformity extends beneath the Plateau. The parts of the Plateau traversed are very broadly folded, and thinning across the anticlinal axis indicates that the folding took place in the late Miocene.

The abundant profiles of the sea floor confirm an earlier view that the regional break in slope between shelf and slope is exceptionally deep at 550 m.

INTRODUCTION

The region studied (Figs 1, 2A) includes the entire Rowley Shelf and the southwest part of the Sahul Shelf, as defined by Fairbridge (1953), the east flank of the North Australian Basin, and the north part of the Exmouth Plateau. Together with the Timor Sea region studied by Veevers (1971a), this study encompasses the entire upper continental slope and edge of the Australian Shelf between North West Cape and Melville Island.

The older summaries of the offshore geology of the region by Fairbridge (1953), Carrigy & Fairbridge (1954), and Veevers & Wells (1961) have now been superseded by the current phase of marine geological (Jones, 1968, 1970, 1971) and geophysical (Whitworth, 1969) exploration by the Bureau of Mineral Resources, and by the geophysical and drilling exploration by B.O.C. (Australia) Ltd and associates in the search for petroleum (Mollan, Craig, & Lofting, 1970;

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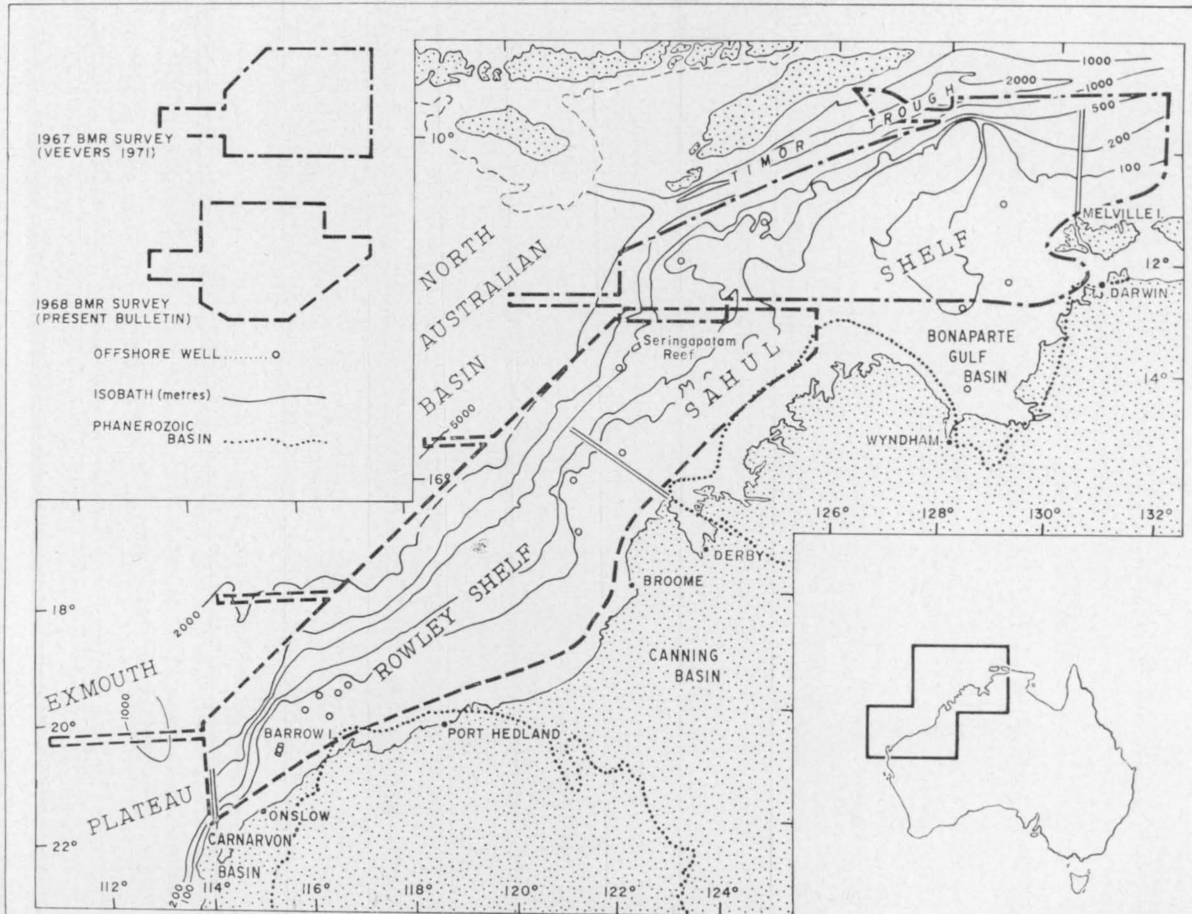

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Fig. 1. Location of study area and earlier study area (Veevers, 1971). Also shown is boundary between Rowley Shelf and Sahul Shelf.

Challinor, 1970; Halse & Hayes, 1971). To date (March 1971), eight offshore wells have been drilled in the area described here, and a ninth is now being drilled at Scott Reef. Information from the first five wells was available at the time of writing (Fig. 3). Barrow Island, with its producing oilfield (McTavish, 1965) also lies within the area.

Drilling has shown that, parallel with the situation in the Timor Sea (Veevers, 1971a), the thick dominantly Palaeozoic sequence in the onshore Canning Basin (Veevers & Wells, 1961) is complemented by an equally thick Mesozoic and Cainozoic sequence offshore. The wells on the Rowley Shelf (Fig. 3) reveal the following succession:

Quaternary to Pliocene carbonate

UNCONFORMITY (here labelled Reflector 1)

Lower to middle Miocene carbonate

UNCONFORMITY (R_2)

Paleocene to Eocene carbonate, claystone, and sandstone

UNCONFORMITY

Middle Jurassic to Upper Cretaceous claystone, siltstone, and sandstone

The Miocene sequence is thin at Barrow Island and is lacking in the Lacepede and Leveque Wells, probably because of the nearshore situation of these localities. Likewise, the Paleocene-Eocene sequence is missing in the Lacepede and Leveque Wells. The Madeleine Well alone contains lower and middle Oligocene strata.

This study of the shallow seismic sections from the 1968 BMR survey was undertaken in an attempt at unravelling the later depositional history of the continental margin. Facing the North Australian Basin in the north and the Exmouth Plateau in the south, this part of the margin provides contrasts within itself and with the region to the north (Veevers, 1971a), which faces the Timor Trough and Timor.

As in the Timor Sea, so here the clearest records come from the shelf edge and upper part of the continental slope. Poor results from the shallow part of the shelf are due chiefly to bottom multiples, so that very few records from shallow areas were usable (Fig. 2A).

Methods of Observation and Analysis

The survey was carried out under contract to BMR by the Ray Geophysical Division of Mandrel Industries Inc. from M.V. *Robray I* in the period 24 September to 17 December 1968, and about 28 000 km were traversed. Navigation was by a combination of the satellite doppler system, v.l.f. navigation system, sonar doppler, buoys, Raydist, and star fixes. The seismic equipment (Whitworth, 1969) was essentially the same as that used in the 1967 survey of the Timor Sea (Jones, 1969; Veevers, 1971a). The facsimile records used in this study were made on an E.G.G. Model 254 recorder with a sweep time of 0.8 s. The receiving system was a Subot Hydro-streamer (Geotech) of 27 hydrophones spread 0.5 m apart, and it was towed 146 m behind an E.G.G. Spark array acoustic source of 21 000 watt-seconds energy capacity. The firing rate was once every 4.4 s, and the ship steamed at 18.5 km/h.

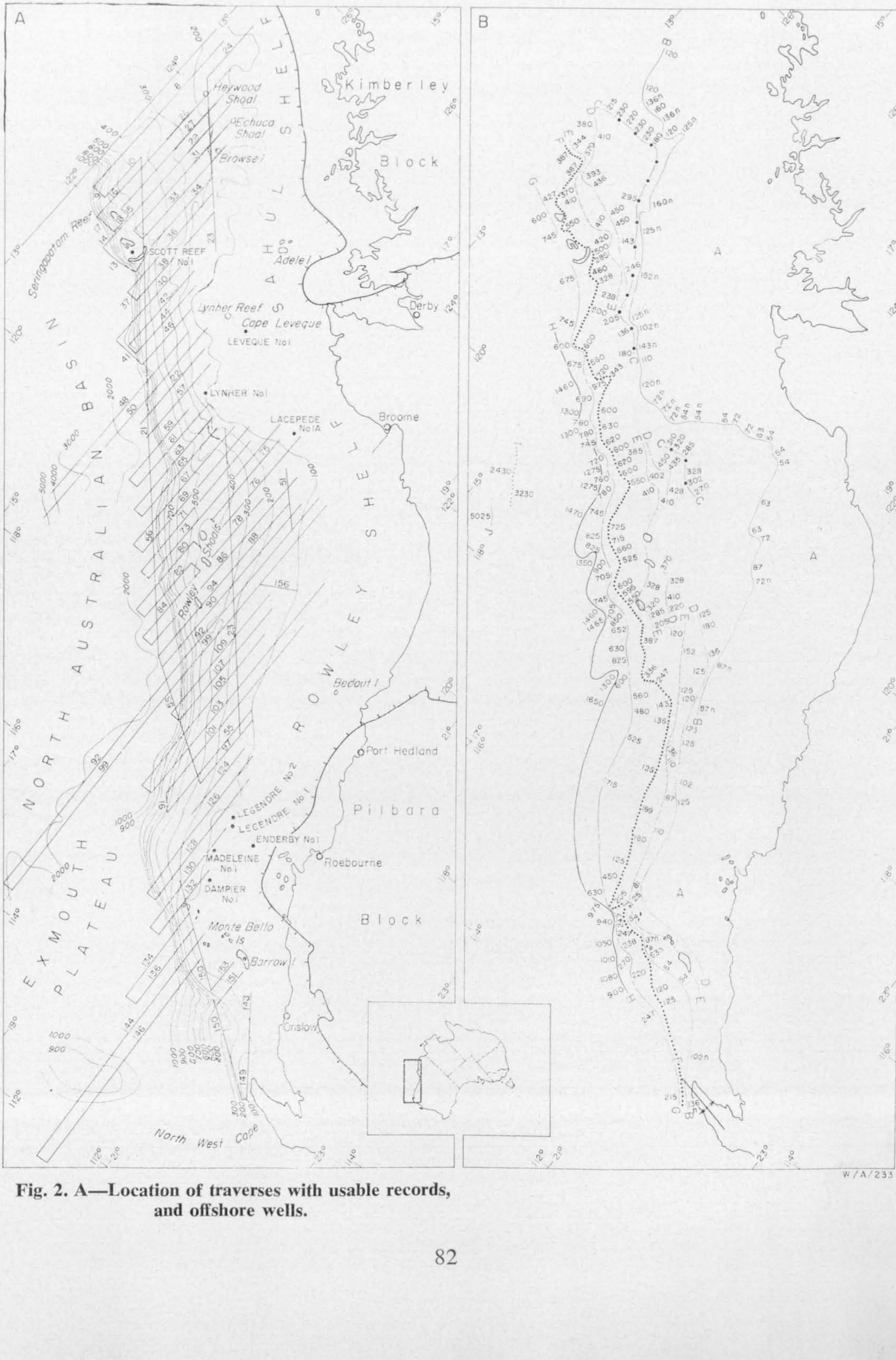


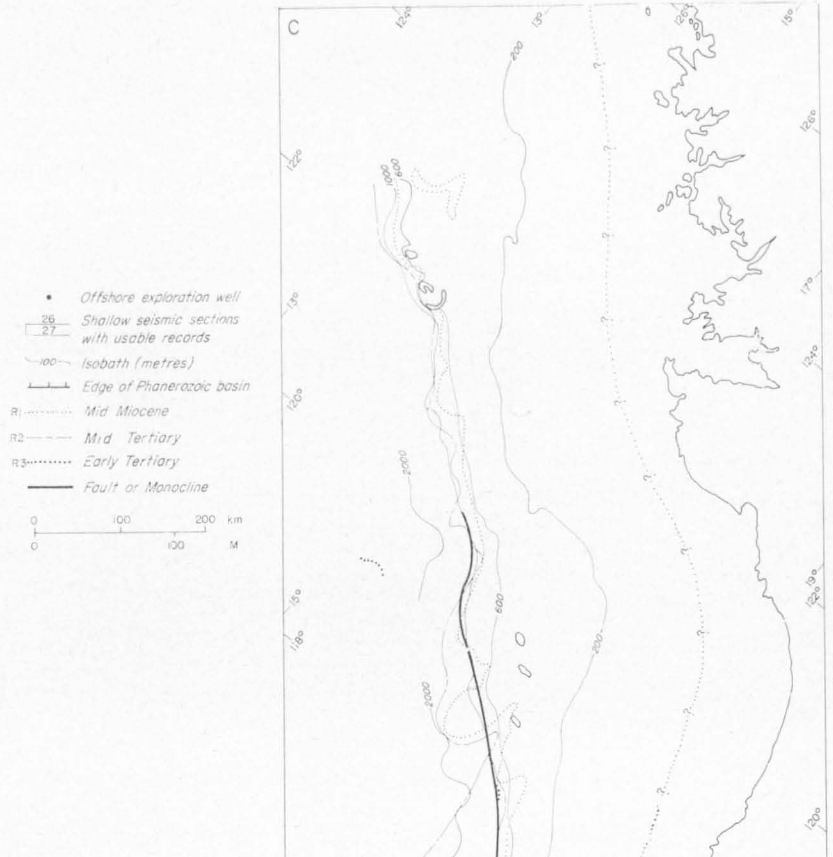
Fig. 2. A—Location of traverses with usable records, and offshore wells.

B—Depths of bathymetric features:

On the shelf-edge zone
(A) Shelf proper
(B) Shelf edge
(C) Notch
(D) Depression
(E) Rise

On the continental slope
(F) Change in gradient
(G) Strip of rough topography
(H) Change in gradient
(I) Intermediate terrace

At the foot of the slope
(J) Deep-sea floor



C—Outcrop of main reflectors.

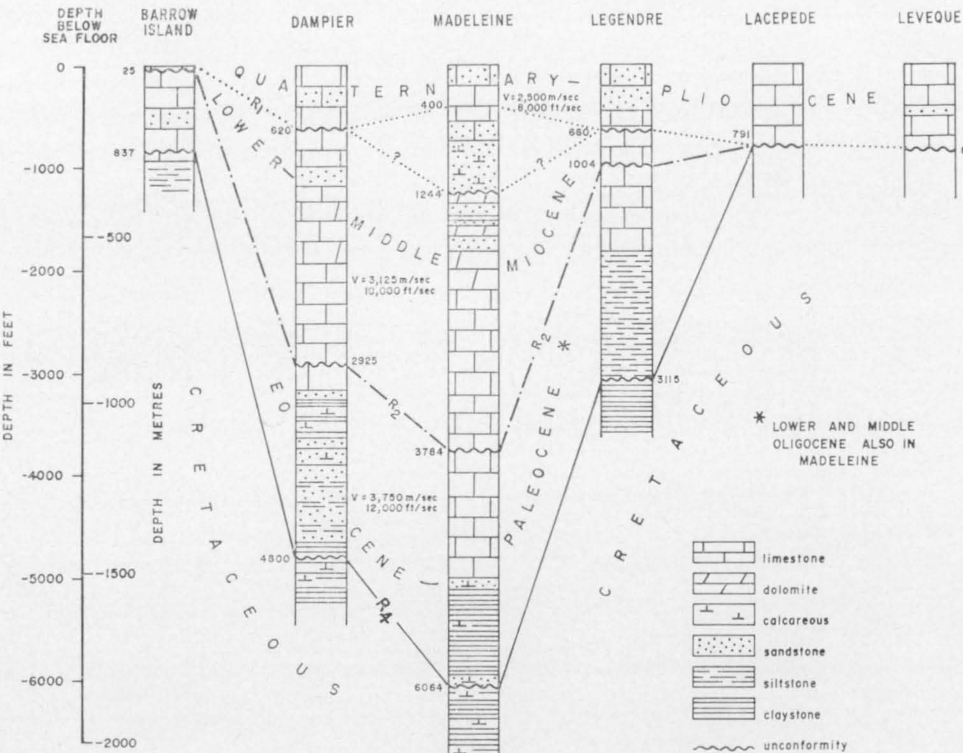


Fig. 3. Cainozoic sequences in offshore wells.

The facsimile records are identical in form with those from the 1967 survey. The average vertical exaggeration is 33:1, and the tracings of the sections below are not corrected for water depth. An improvement over the 1967 survey was the adjustability of the facsimile recorder to shift the range of the vertical scale by amounts less than 0.8 s. The seismic pulse was 40 ms long. Resolution in the near-surface layers was 60 m, and it improved at depth because of attenuation.

For the sake of the concurrent gravity-meter survey, traverses were run east-west in lines 16.5 km apart, with northeast-southwest tie-lines roughly 180 km apart. Ties were made into the 1967 survey of the Timor Sea. With rare exceptions only, records from traverses in water shallower than 100 m (Fig. 2A) were not usable. In water deeper than 100 m, virtually all the records were usable. Since the traverse lines extended almost to the coast, this meant that about half the surveyed area, or 240 000 km^2, yielded usable shallow seismic sections.

The interpreted records were reduced photographically and compiled into eight figures. Figures 4-10 cover the east-west traverses in order from north to south, and Figure 11 contains the long east-west sections and the southwest-northeast sections. Plate 12 shows selected records and their interpretation.

Sea Floor Morphology

Fairbridge (1953) was the first to point out the chief morphological features of the region of the Sahul and Rowley Shelves: an abnormally deep change of gradient from shelf to slope at about 300 fathoms (550 m), and 'a major "step" or break in the slope of this shelf . . . at about 55-60 fathoms' (100-110 m). Van Andel & Veevers (1967) and Jones (1968) failed to find the regional break in slope at 550 m, but abundant data available to me in both the Timor Sea (Veevers, 1971a) and the area described here confirm Fairbridge's work. The postulated shallower step or break in slope at about 110 m was also confirmed. Carrigy & Fairbridge (1954) made the change in slope at 60 fathoms (110 m) the division between an inner and an outer shelf, which had its edge at 550 m. A simpler classification is Jones's (1970) threefold division of the region into (a) shelf, from 0-120 m; (b) edge of shelf zone from 120-300 m; and (c) the continental slope. I prefer this classification, and use it below.

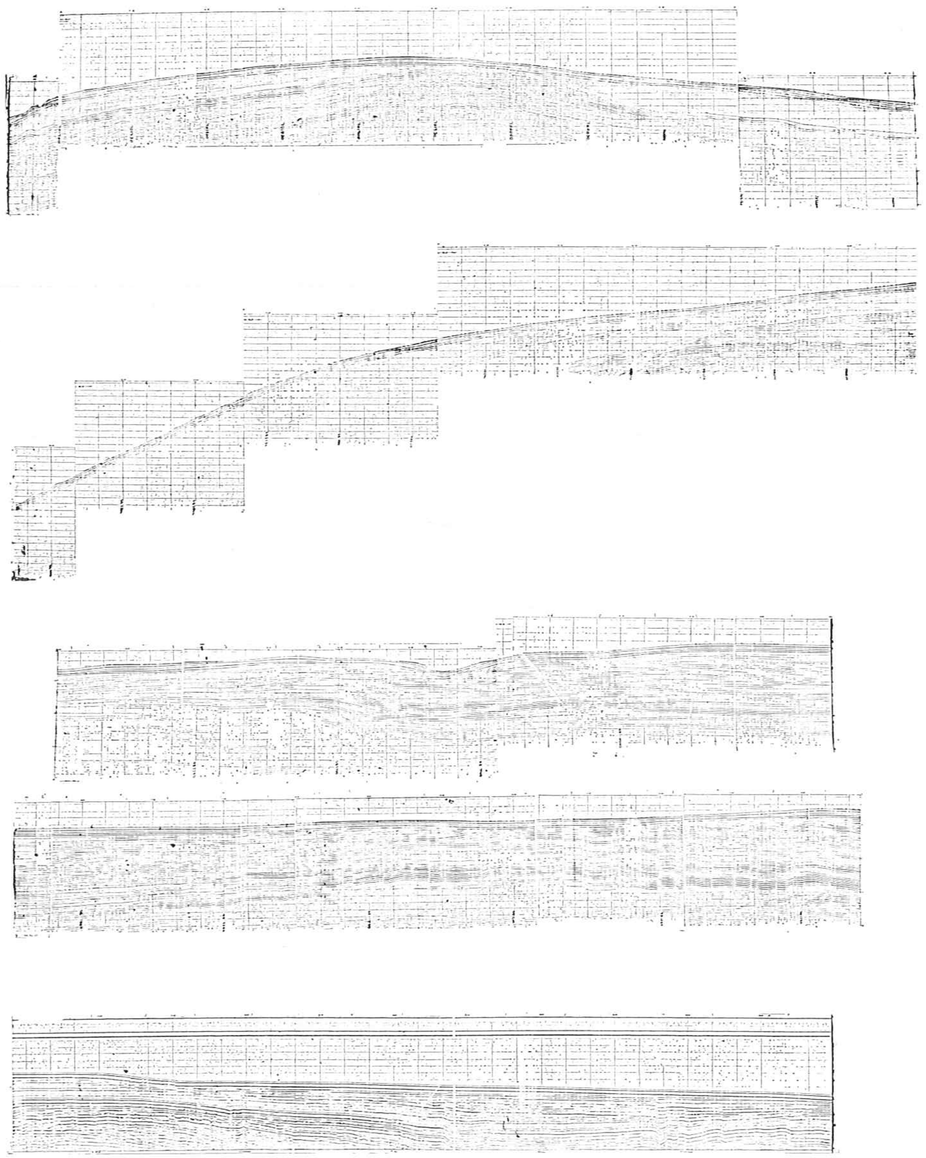
Half of the marine area shown in Figure 2A is shelf, but, owing to the poor seismic results in the shallow water of the shelf, only the westernmost edge of the shelf appears in the sections, most of which show the broad shelf-edge zone and upper continental slope. In what follows, the morphology is described in terms of breaks of slope, labelled A to J (Figs 2B, 12).

Shelf

Shelf Proper (*A*). The shelf proper ranges in width from 15 km off North West Cape to 200 km off Cape Leveque, and its almost flat surface (gradient 0°02′ to 0°03′) is broken by mounds and depressions of probable reef origin and by sand waves with a height of 5-10 m (Jones, 1970, pp. 16-17). The shelf is largely an area of winnowing and sediment transport, and it is mantled by coarse skeletal, pelletal, and oolitic calcarenite (Jones, 1970).

Shelf Edge (*B*). The shelf edge is marked by a notch or change in slope. In all but three crossings between latitudes 13°-16°S (traverses 8-57, Table 1, Fig. 2B), the shelf edge is marked by a notch that ranges in depth from 102-160 m (mean 130 m). From latitude 16°-19°S (traverses 61-55), the shelf edge is marked by a notch in eight of the nineteen crossings, and by a change in slope in the other crossings; its depth ranges from 54-87 m, with a mean of 67 m. From latitude 19°00′-19°45′S (traverses 122-130), the shelf edge is not detectable. Finally, from latitude 19°45′-22°00′S (traverses 132-148), the shelf edge is notched in four of the nine crossings and, except in the two southernmost crossings, ranges in depth from 54-87 m (mean 65 m). In the two southernmost crossings (both in traverse 148), the shelf edge lies at 102 m and 136 m.

If we follow the traditional interpretation of the shelf-edge notch as the indication of a single eustatic lowstand of sea level at the end of the Pleistocene, we must invoke warping of this once level surface to account for its present depth range of 54-160 m. Over the Sahul Shelf (van Andel & Veevers, 1967; Veevers, 1971a), as shown in the northern part of the present study area, the shelf edge lies generally at 130-140 m (75 fathoms), which is the mean depth of the shelf in most other parts of the world (Shepard, 1963). The shallow shelf edge south of 16°S (Rowley Shelf), at a mean depth of 66m, may therefore imply upwarping of some 70 m since the last eustatic lowstand. The deeper shelf edge near North West Cape (Traverse 148) suggests that North West Cape marks the southern border of the uplift. In summary, the Rowley Shelf seems to have been affected by movement since the last lowstand whereas the Sahul Shelf has remained steady.



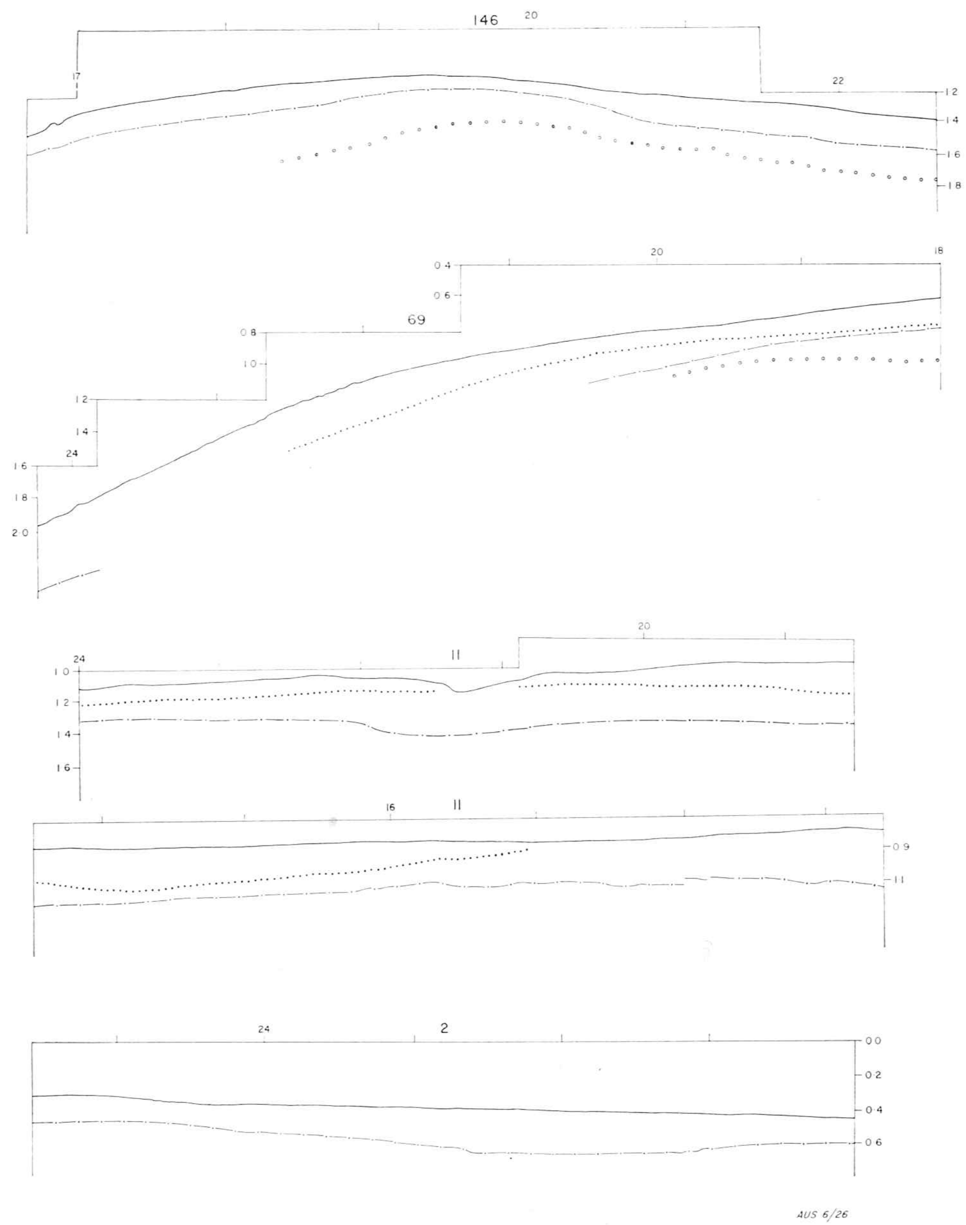


Plate 12. Photographic prints of selected seismic records, and (opposite page) tracings showing seafloor and reflectors R_1 (dots), R_2 (dots and dashes), and R_3 (circles). Horizontal scale in hours, related to ship's speed of 10 knots; vertical scale in seconds of two-way reflection time.

Traverse 146 shows a broad anticline at the crest of the Exmouth Plateau; Traverse 69 is a profile across the shelf edge zone and upper slope, and it shows a marked thickening of prograding strata above R_1; Traverse 11, broken into two parts for presentation here, is a profile along the shelf edge zone; it shows R_1 cropping out in three places, and complex small-scale structure beneath R_1; Traverse 2, also a longitudinal profile, shows the 40-millisecond seismic pulse, as seen in the direct wave, and the prominent R_2. For further details, see appropriate tracings in accompanying figures.

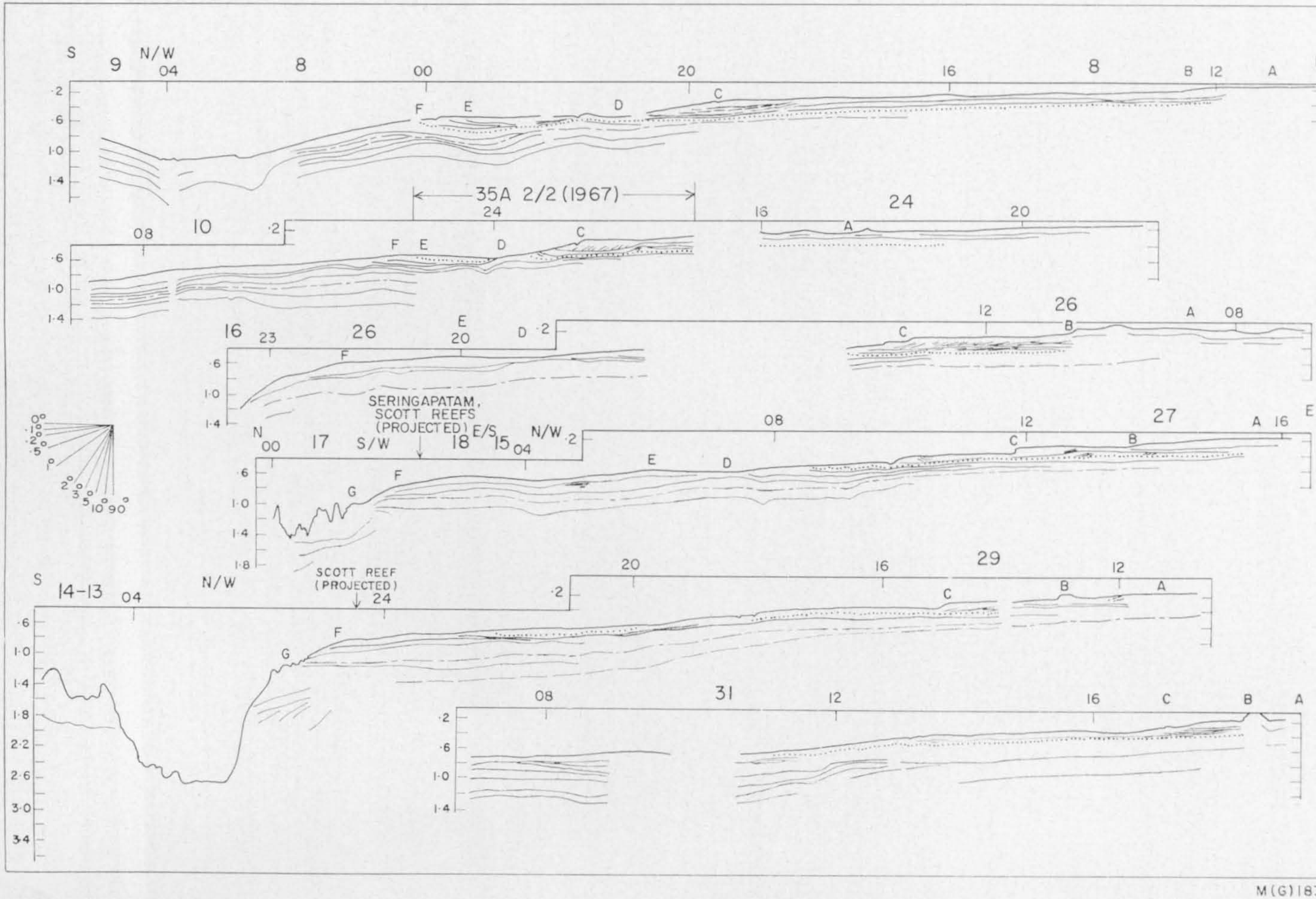


Fig. 4. East-west profiles (west on left-hand side). The vertical scale is the two-way reflection time in seconds, and the horizontal scale is indicated by hour marks, which relate to speeds of 18.5 km/h. True slopes indicated in diagram. Location of traverses shown in Figure 2A.

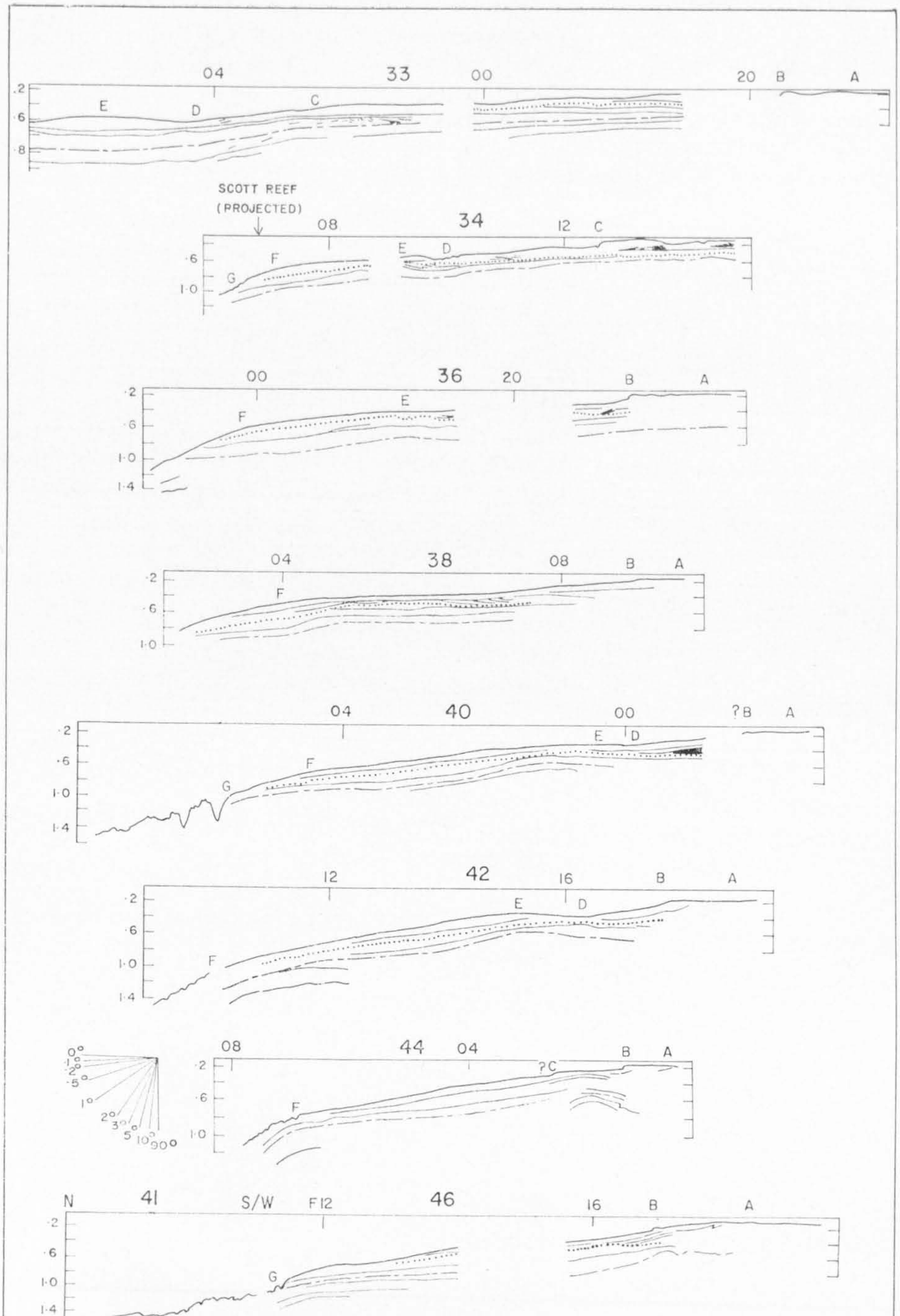


Fig. 5. East-west profiles (west on left-hand side). See caption of Figure 4 for explanation of symbols and scales. Location of traverses shown in Figure 2A.

Phipps (1966) mapped warped terraces on the shelf off Sydney similar to that described above. C. V. G. Phipps (pers. comm.) now regards these terraces as having been shaped originally earlier in the Pleistocene and mantled by younger sediments during successive transgressions. I am inclined to follow Phipps's view if only because the rate of warping of the Rowley Shelf would be inordinately high if the warping had been concentrated in the last 20 000 years.

The *Shelf-Edge Zone* extends from the shelf edge to the continental slope. Its inner side ranges in depth from 54 to 160 m, and its outer side from 215 to 900 m. Its average gradient is 0°07′. According to Jones (1970), it is a zone of deposition of fine-grained carbonates, including planktonic Foraminifera.

Notch (*C*). A prominent notch, up to 30 m high, lies from 13 to 150 km seaward of the shelf edge (B) between latitude 13° and 15°30′S, and 16°20′ and 17°10′S. In its northern occurrence, the foot of the notch ranges in depth from 136-230 m (except in traverse 34 at 295 m, a possible misidentification) and is 60-110 m deeper than the shelf edge, which is generally 130-140 m deep. In its southern occurrence, its depth range is 270-328 m, and it is 198-274 m deeper than the shelf edge, whose average depth is 66 m. This notch was first seen in the southern part of the 1967 survey area (Veevers, 1971a); here the foot of the notch lies at 256-311 m, some 119-174 m lower than the shelf edge, whose average depth is 137 m.

The notch is interpreted as an excavation made at the shore during an earlier Pleistocene stillstand. Using the depth of the present shelf-edge notch (B), itself a shore line feature, as datum we interpret the different levels of C and B as indicating differential subsidence in the interval between the formation of C and B. The amounts, in round figures, are:

100-200 m between latitudes 12½°-13° (Veevers, 1971a)
60-100 m between latitudes 13°-15½°
200-300 m between latitudes 16°20′-17°10′

This amounts to less than 200 m on the south Sahul Shelf and more than 200 m on the north Rowley Shelf. This is a reversal of direction of movement since the formation of the shelf-edge notch.

Depression (*D*) *and Rise* (*E*). Seaward from the notch (C) in the northern half of the area and seaward from the shelf edge in the southern half are a broad depression (D) and a rise (E). The relief of the rise above the depression is less than 50 m. The rise bears the Rowley Shoals on the shelf-edge zone in the middle part of the area, and the Rankin Shoals, Monte Bello Islands, and Barrow Island on the shelf proper in the south. From a general depth of 100 m in the south, the depression and rise deepen to a general level of 400 m in the north. Most of the deepening occurs along a zone near latitude 18°S across which the rise and depression are displaced 40 km to the left.

Jones (1971) first described the rise on the shelf south and southwest of the Rowley Shoals and concluded that it 'marks the zone of active sedimentation since deposition commenced on Surface 2; it is the site of deposition of sediments of somewhat pelagic aspect consisting of planktonic organisms from the waters of the shelf (*Globigerina* etc.), fine-grained material winnowed from the middle and inner shelf, and minor amounts of terrestrial silts and clay. Normally such

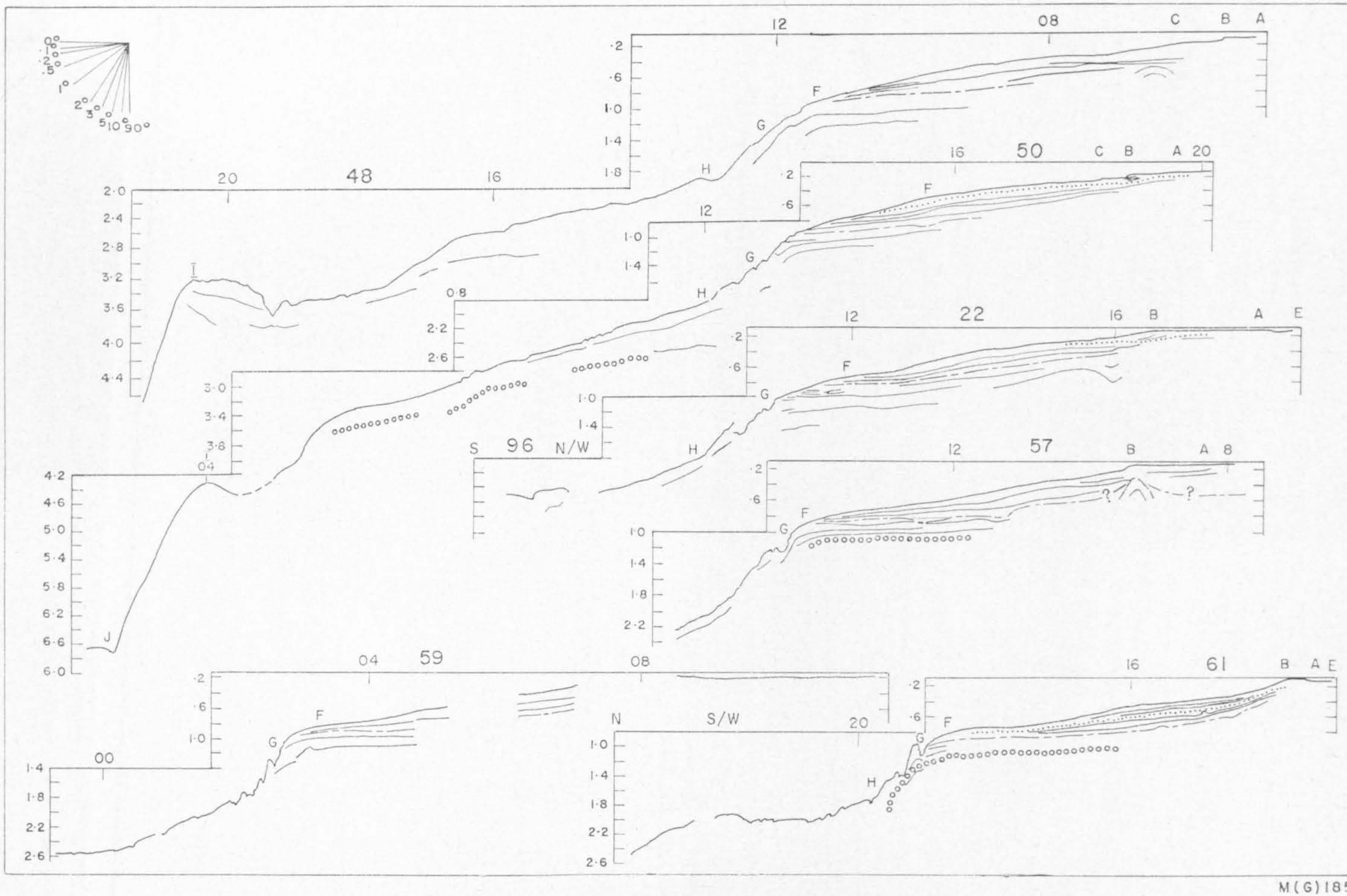


Fig. 6. East-west profiles (west on left-hand side). See caption of Figure 4 for explanation of symbols and scales. Location of traverses shown in Figure 2A.

sediments would be deposited seaward of the continental shelf and would contribute to the prograding of the continental slope . . . [but] the local regime of sediment transport and deposition and the great width of the shelf itself in this area has resulted in the depositional zone occurring on the shelf itself'. Jones further interpreted some buried structures as earlier Cainozoic equivalents of the rise.

The seismic sections show that the rise has three structural forms (Fig. 13). The biconvex form (a) occurs exclusively between 13°-16½°S, and the plano-convex (b) and concavo-convex (c) forms between 16½°-19½°S. The biconvex form is obviously produced by the erosion of a sediment-filled depression, the plano-convex from the piling up of sediment over a flat surface and subsequent erosion, and the concavo-convex form by deposition or erosion parallel to the initial layering.

Continental Slope

The continental slope has three parts: steep upper and lower parts and an intermediate terrace, including the Exmouth Plateau.

A change of gradient (F) marks the boundary between the shelf-edge zone and the continental slope. The change in gradient ranges in depth from 54 to 800 m. North of latitude 18°15′S (traverses 109-8), the range is 328-800 m (mean 535 m) and southward (traverses 107-151) 54-247 m (mean 142 m) (Fig. 14). The shallow change in gradient corresponds to the flank of the Exmouth Plateau, and the deep change to the flank of the North Australian Basin. The mean depth of 535 m of the change in gradient is a startlingly precise confirmation of Fairbridge's 300 fathoms (550 m).

Seringapatam and Scott Reefs lie immediately seaward of F in the northern part of the area.

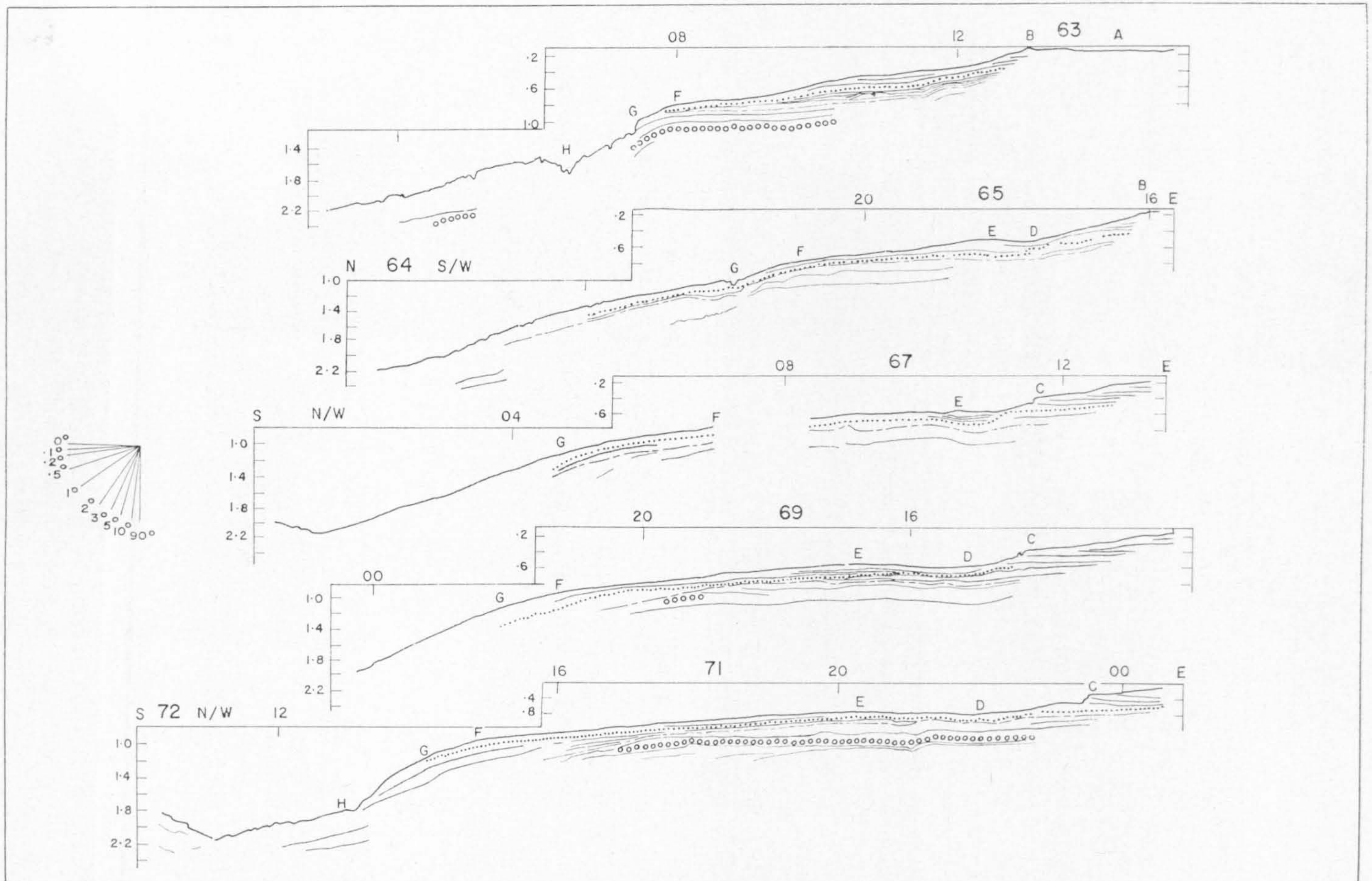
A strip of rough topography (G) along the upper continental slope marks another change in gradient. Between latitudes 15½° and 18½°S, this is the surface expression of a zone of strata disturbed by a fault or monocline with displacement down in a seaward direction.

The steep gradient of G flattens out at H, at a depth of 900-1 080 m in the south, and from 1 275 to 1 650 m in the middle part of the area.

An intermediate terrace (I), marked by changes of slope at H and I, is seen in the only two traverses that cross the slope, traverses 48 and 50. The Exmouth Plateau, which occupies the southwest part of the area, is probably an extension of the intermediate terrace. As seen in traverses 144 and 146, the plateau has an undulating surface of rises and depressions dissected on its north flank by canyons (traverses 92 and 99).

A change of slope (J) at 5 025 m, seen in traverse 50 only, marks the foot of the continental slope and the edge of the North Australian Basin.

The morphology described above, and summarized in Figure 12, is as it appears in east-west profiles. The long southwest-northwest profile (Fig. 11, traverses 150, 56, 11), most of which crosses the continental slope, reveals the canyons in the middle part of the area.



M(G) 186

Fig. 7. East-west profiles (west on left-hand side). See caption of Figure 4 for explanation of symbols and scales. Location of traverses shown in Figure 2A.

Stratigraphic Succession

The stratigraphic succession is illustrated in Figure 12. Two persistent reflections, called R_1 and R_2, have been traced over the entire area. The upper one (R_1) is the widespread mid-Miocene unconformity; the lower one (R_2) is the surface between the lower to middle Miocene and the lower Tertiary sequences. A third reflector of narrower extent (R_3) lies within the lower Tertiary. The base of the Tertiary is not visible in the shallow profiles, and the lowermost surface (R_4), near the base of the Tertiary shown in each of the three diagrammatic sections of Figure 12, is from Whitworth (1969).

Identification of Reflectors

R_1 is identified as the unconformity between the lower to middle Miocene sequence and the Pliocene-Quaternary sequence, and R_2 as the unconformity between the Miocene sequence and the lower Tertiary. Evidence for these identifications is provided by five offshore wells, B.O.C. Dampier, Madeleine, Legendre, Lacepede, and Leveque Wells (Fig. 3), by outcrop and subsurface data on Barrow Island and at North West Cape, and by continuity with the 1967 BMR geophysical survey of the Timor Sea (Veevers, 1971a). R_3 cannot be identified precisely. It lies between the Cretaceous and R_2, and hence lies within the Paleocene-Eocene sequence.

R_1 is an obvious unconformity in a few profiles only, viz. 11, 33, 61, 65, 67, 73, 75, 82, 109. Elsewhere, it is a paraconformity. R_2 is even more rarely seen to be an unconformity (traverses 22, 48, 57, 71, 86, 55, 126), probably because it is commonly the lowest reflector. The unconformity has low relief, except in traverse 126, where it has a relief of 0.4 s over a distance of 30 km. R_3 is a conformable surface in the few traverses in which it is visible.

All the subsurface well information comes from the shelf, whereas practically all the seismic information comes from the area beyond the shelf. The poor expression in the profiles of the unconformities found in drilling may indicate a succession beneath the slope and shelf edge that is more complete than that beneath the shelf proper.

Sequences Between the Unconformities

1. *Pliocene to Quaternary sequence above R_1*. In the five offshore wells for which information is available (Fig. 3), the sequence above R_1 consists predominantly of marine carbonates. The Lacepede section is wholly carbonate, and the Dampier, Madeleine, Legendre, and Leveque sections consist mainly of carbonates, with a greater or lesser amount of quartz skeletal sand. The carbonates are mainly biogenic calcarenite, with smaller amounts of coquina, skeletal dolomite, marl, and calcilutite, and they range from loose sediment to hard limestone or dolomite.

The thickness ranges from 189 to 257 m (620-843 ft) in four of the offshore wells. In the fifth well, Madeleine, the position of the unconformity is unknown. Chiefly by correlation with Dampier and Legendre, it is placed at 380 m (1 244 ft). But according to the seismic profiles studied here, the unconformity lies at about 0.1 s beneath the sea floor or at a depth of 122 m. Both interpretations are shown in Figure 3.

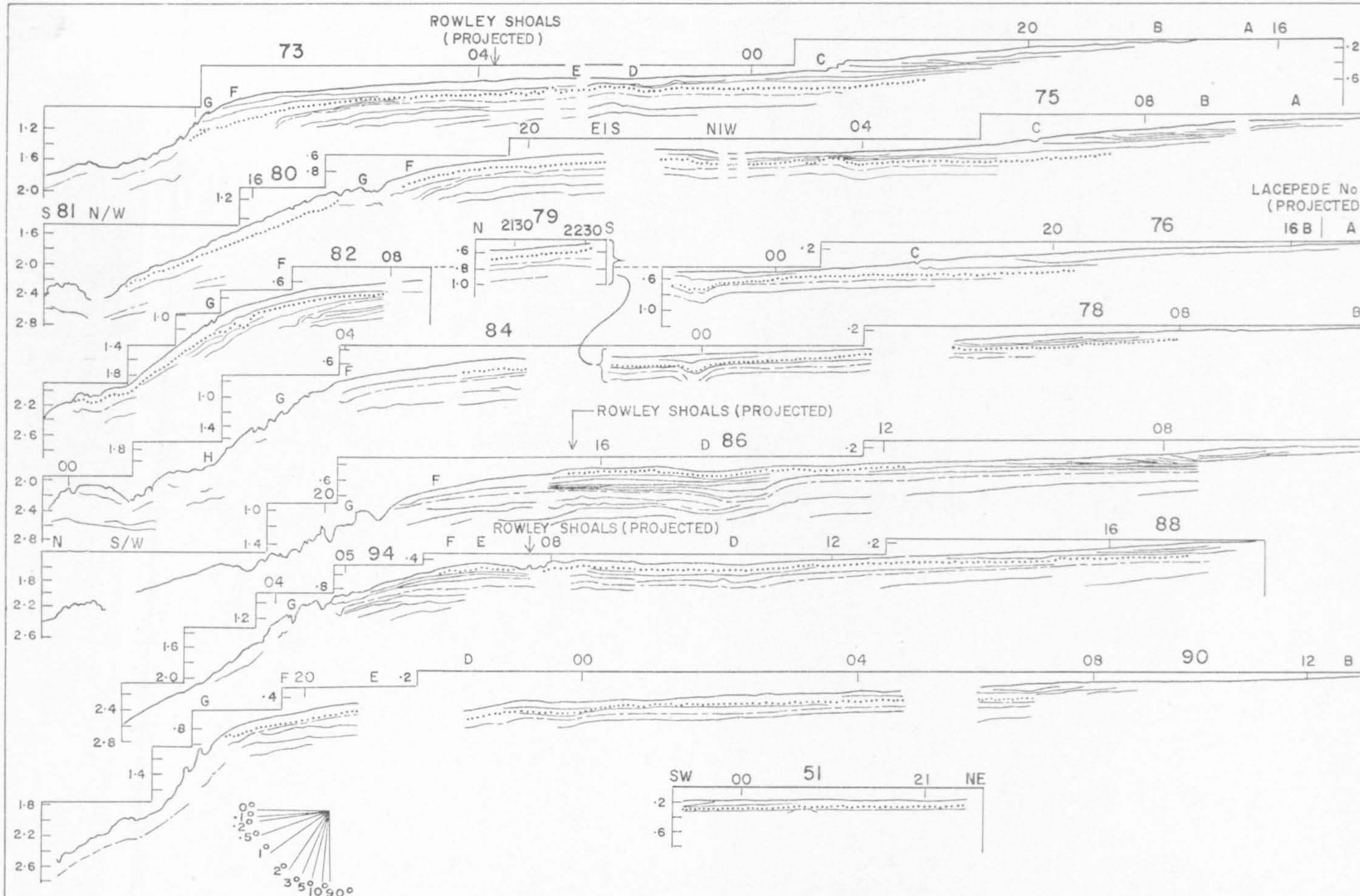


Fig. 8. East-west profiles (west on left-hand side). See caption of Figure 4 for explanation of symbols and scales. Location of traverses shown in Figure 2A.

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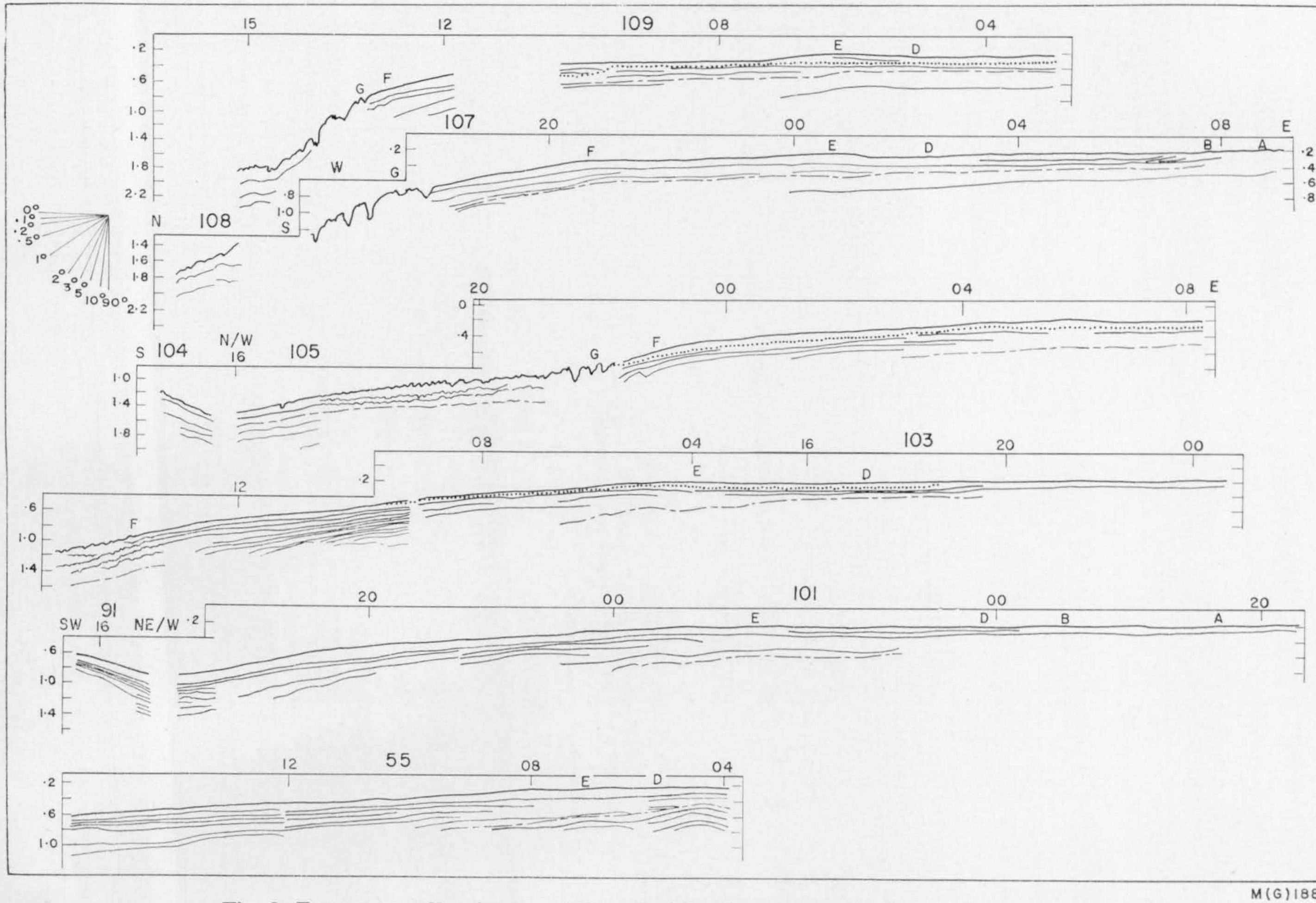

Fig. 9. East-west profiles (west on left-hand side). See caption of Figure 4 for explanation of symbols and scales. Location of traverses shown in Figure 2A.

The style of deposition of the Pliocene-Quaternary sequence in the southwest part of the Timor Sea, south of latitude 12½°S (Veevers, 1971a, fig. 14), extends through the northern part of the present study southward to latitude 15½°S (traverses 8-22, Fig. 12a). This area has a fairly uniform morphology which reflects the structure of the underlying Pliocene-Quaternary strata. Thus the shelf edge (B) is underlain by outbuilt layers roughly concordant with the slope of the adjacent shelf-edge terrace. The morphology and structure at C are essentially the same as at B, and indicate an earlier cycle of outbuilding during a lower stand of sea level.

From 15½° to 17½°S, the sequence above R_1 comprises horizontal upbuilt layers truncated laterally by the sea floor. Thus R_1 is locally 0.4 s deep beneath B, shallows seaward to the depression (D), and then parallels the sea floor at a fairly constant depth of 0.1 s to crop out at the top of the slope (G) or deeper. From 17½°S to the southern margin of the Rowley Shelf, the sequence above R_1 again comprises horizontal upbuilt layers, but they thicken immediately landward of the outcrop of R_1 at G. These three areas of Pliocene-Quaternary deposition reflect the shape of the continental margin, from a broad shelf-edge zone in the north to a steep upper continental slope in the south. Apparent inbuilt layers are seen in traverses 27, 34, and 38.

From twelve low-power sparker traverses in the area, Jones (1970) was able to find three surfaces of erosion above R_1, which corresponds with Jones's Surface 4.

'Surface 1 . . . probably represents the surface exposed during the last major low sea-level stand of the Pleistocene. It is likely that this surface and the underlying Surface 2 control the present-day sea floor over wide areas of the middle shelf where little or no Recent deposition has occurred. Elsewhere . . . 20-70 metres of Recent sedimentary fill has been deposited upon Surface 1.

'Surfaces 2 and 3 occur below Surface 1 and above the Mio-Pliocene Surface 4 [$=R_1$]. One or possibly two other planes of erosion may occur in this interval also. Surface 2 is a strong irregular reflector of regional extent which is overlain by up to 100 metres of sediment; it is normally separated from the less persistent and less prominent underlying Surface 3 by 35 metres or less of section and the two surfaces probably merge locally' (Jones, 1970, p. 26).

2. *Lower to middle Miocene sequence between R_1 and R_2*. Three offshore wells, Dampier, Madeleine, and Legendre (Fig. 3), show that the sequence between R_1 and R_2 comprises carbonate rocks of lower to middle Miocene age. The thickness of this sequence ranges from 774 m or possibly as much as 1 032 m in Madeleine to a preserved 7.5 m on Barrow Island, where it is called the Trealla Limestone. Some areas, such as those around the Lacepede and Leveque Wells, lack this sequence altogether. The Miocene sequence in Dampier No. 1 Well is more precisely dated than it is elsewhere, and it consists of calcarenite and calcilutite, dolomitic in places, quartz sandstone, and marl.

Except in a few traverses, the sequence between R_1 and R_2 is acoustically simple, with layers parallel to R_1 and R_2. In traverses 33 and 34, the sequence locally contains low-angle inclined sets with apparent dip landward, and in traverse 63 sets dipping seaward. The sequence between R_1 and R_2 generally thickens across the shelf to a maximum beneath the other part of the shelf-edge zone. The sequence thins across the axis of an anticline seen in traverses 40-57.

3. *Lower Tertiary sequence below R_2*. In three offshore wells, Dampier, Madeleine, and Legendre, and on Barrow Island, R_2 is identified as the unconformity between Paleocene-Eocene and Miocene sediments. A more complete Paleogene section in Madeleine includes lower and middle Oligocene sediments. The hiatus at R_2 thus encompasses the upper Oligocene in Madeleine, and the entire Oligocene in the other wells and at Barrow Island. In the other offshore wells, Lacepede and Leveque, the entire Paleogene (Paleocene-Oligocene) is missing.

R_2 is the most extensive and readily traced reflector or zone of reflectors over the area. In traverses 31, 22, 69, 71, 82, 86, 94, 55, 126, and 2, R_2 is seen to be an angular unconformity. Elsewhere it is a paraconformity.

The deepest prominent reflector or zone of reflectors (R_3) has a restricted occurrence, mainly beneath the shelf-edge zone, in traverses 57, 61, 63, 69, 71, 52, and 157 (all in the central part of the area) and in the deeper water of the Exmouth Plateau in traverses 134, 136, 144, and 146. R_3 is not recognized in wells. Its age lies within the interval Upper Cretaceous to Eocene. The lower limit of Upper Cretaceous is set, not by deeper reflectors found in this study (of which there are very few), but by a reflector very low in the Tertiary mapped by Whitworth (1969, pl. 10) and Challinor (1970, fig. 9) from deeper seismic records.

The Paleogene section consists dominantly of carbonate on Barrow Island and of shale, siltstone, and sandstone overlain by carbonate rocks in Dampier, Madeleine, and Legendre Wells. The shale, siltstone, and sandstone sequence, exceptional in the dominantly carbonate Cainozoic succession, reaches a thickness of 498 m in Dampier, and is interpreted as being 'derived from river outlets accumulated in lagoonal conditions behind a possible offshore bar complex' (Challinor, 1970, p. 86).

Folds and Faults

The only structures of regional extent are the monocline and fault that underlie the upper continental slope. These structures persist northward into the Timor Sea area. The monocline dominates in the area that faces the North Australian Basin, and the fault dominates opposite the Exmouth Plateau. In the rest of the area, the sea floor parallels the monocline except where faulting has produced a steeper slope.

Between traverses 40 and 57 the shelf-edge zone is underlain by an anticline in R_2 and older strata. Since the anticline is only weakly reflected in the strata above R_2, it helps reveal the unconformable contact between R_2 and the younger strata.

On the Exmouth Plateau, several great folds were found in traverses 92, 99, 144, and 146. The largest of these, in traverses 144 and 146 and first reported by Whitworth (1969, pl. 11), measures some 300 km from trough to trough and has an amplitude of 1.2 s. In its enormous size, this structure resembles that of the Ontong Java Plateau.

The folds seen in traverses 92 and 99 are smaller: the biggest is 130 km across with relief of 0.3 s. Because all these structures on the Exmouth Plateau are glimpsed from only two pairs of widely spaced traverses, the shape of complete structures must await the completion of a comprehensive survey.

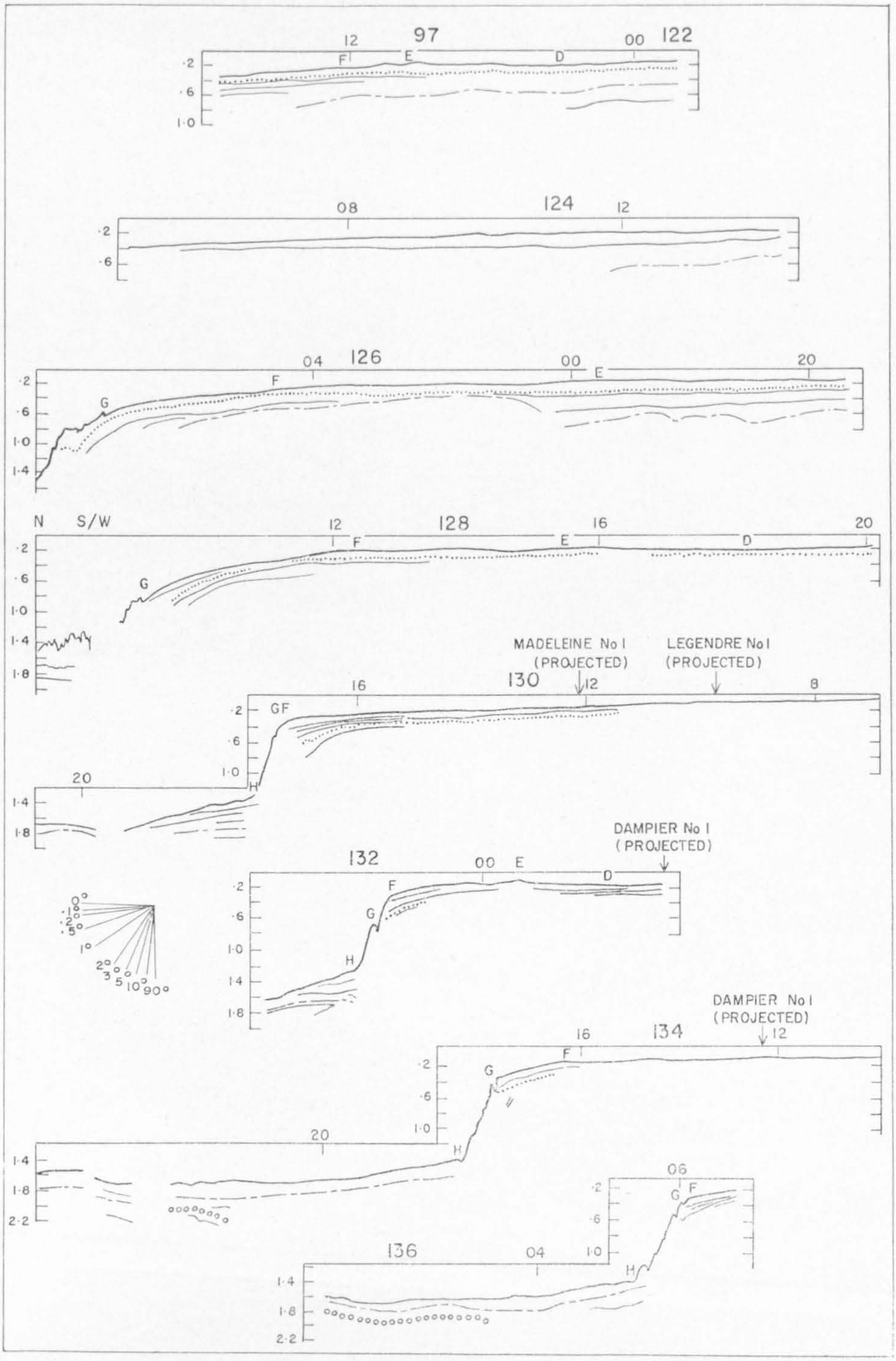

Fig. 10. East-west profiles (west on left-hand side). See caption of Figure 4 for explanation of symbols and scales. Location of traverses shown in Figure 2A.

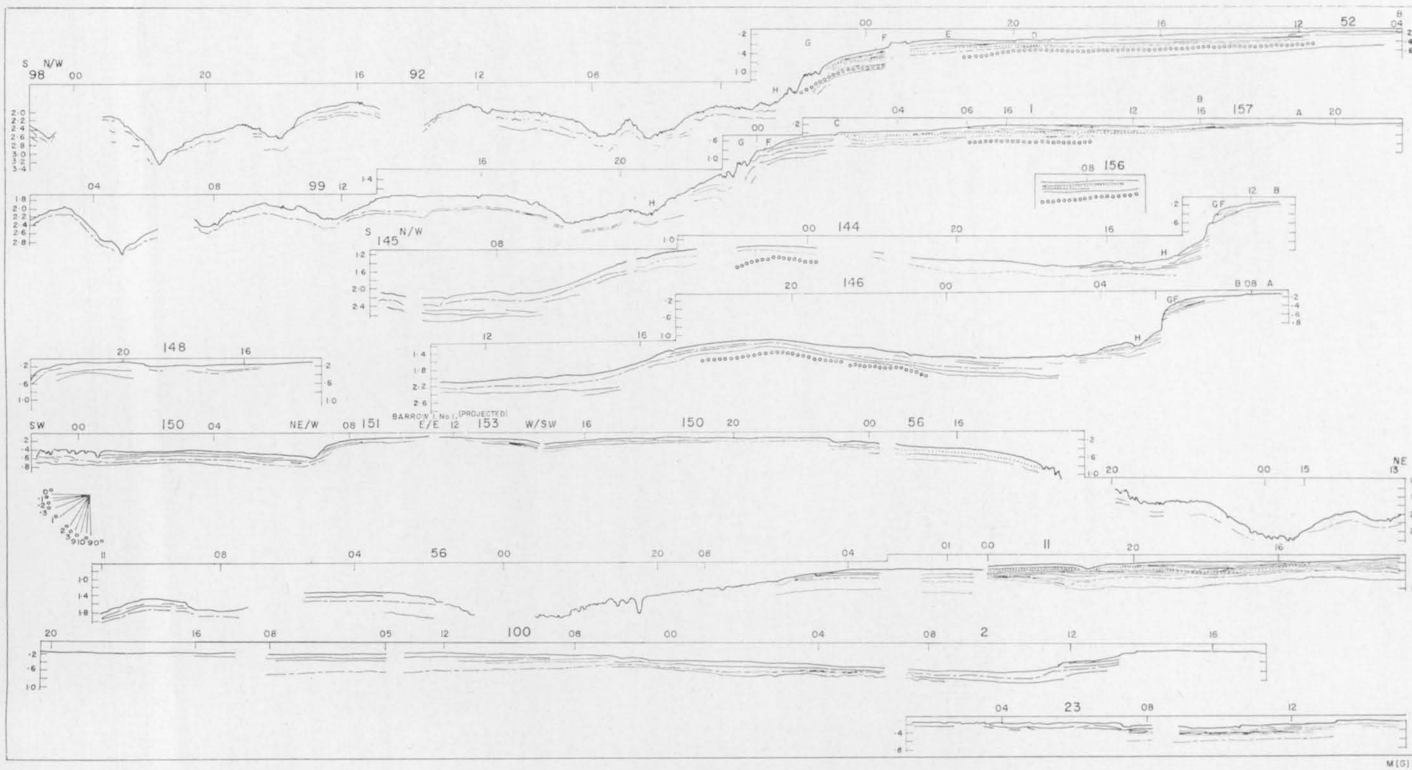

Fig. 11. East-west profiles (traverses 92, 99, 144, 146, west on left-hand side) and southwest-northeast profiles (southwest on left-hand side). See caption of Figure 4 for explanation of symbols and scales. Location of traverses shown in Figure 2A.

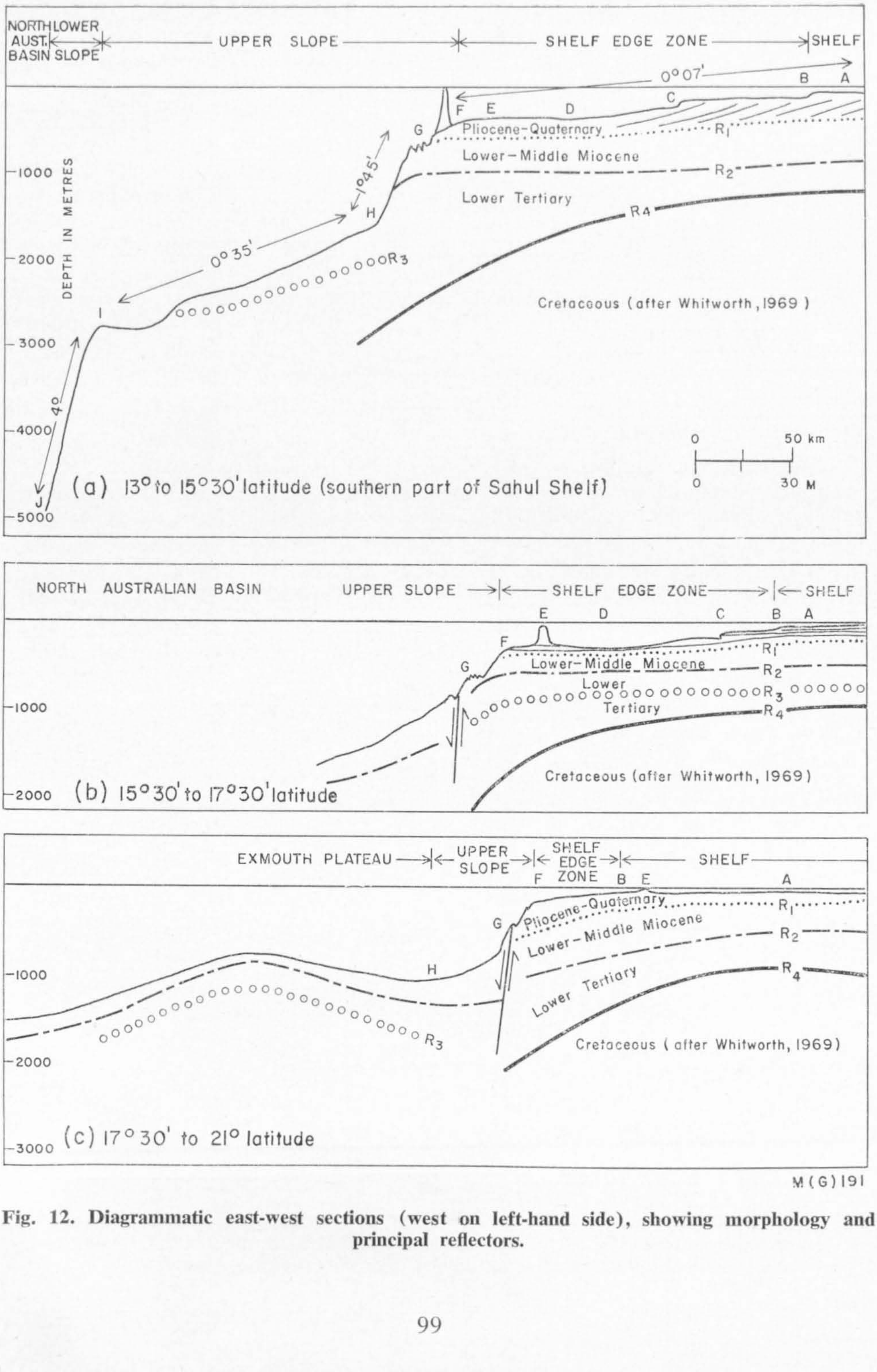

Fig. 12. Diagrammatic east-west sections (west on left-hand side), showing morphology and principal reflectors.

General Structure

The general structure is illustrated in Figure 12.

(a) *Northernmost area (Southern part of Sahul Shelf)*. The structure of the southern part of the Timor Sea area (Veevers, 1971a, fig. 14) is clearly traceable into the present study area as far southward as traverse 29, and less clearly as far as traverse 36. The structure is characterized by two sets of prograding layers of sediment above R_1: one set beneath the deep notch (C), and the other beneath the shelf-edge notch (B), implying a history of repeated subsidence and progradation. This kind of structure, modified beneath by an anticline in R_2, continues southward to traverse 50, which marks the southern edge of the Sahul Shelf.

(b) *Southern part of Rowley Shelf*. This part of the shelf adjoins the Exmouth Plateau, and strata thicken at the seaward edge of the shelf-edge zone in a monocline modified by a fault. The boundary between the slope and the shelf-edge zone is on the average 400 m shallower than it is farther northward.

Cainozoic Depositional History

The reflections studied here lie wholly within the Cainozoic sequence, which is an entity deposited in response to a regional event. According to Veevers, Jones, & Talent (1971) and Veevers (1971b), Australia and India broke away from each other in the late Cretaceous, and the offshore section of Cainozoic sediments, mainly carbonates, documents an episode of marginal subsidence. At Ashmore Reef (Veevers, 1971a), the Cainozoic section is complete except for a hiatus that coincides with the mid-Miocene diastrophism on Timor, which Veevers (1971b) interprets as indicating the collision of the Australian and Asian blocks. At the same time and probably linked with the collision, a set of vertical movements affected the area, as shown best by the broad fold of the Cape Range and by the regional occurrence of R_1.

The area studied consists of two parts: the continental margin of the Exmouth Plateau and adjacent shelf and upper slope, and the North Australian Basin and adjacent slope and shelf.

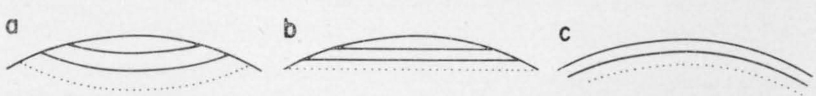

Fig. 13. Structural forms of the rise (feature E). Reflector 1 (R_1) is indicated by dotted line. a—biconvex; b—plano-convex; c—concavo-convex.

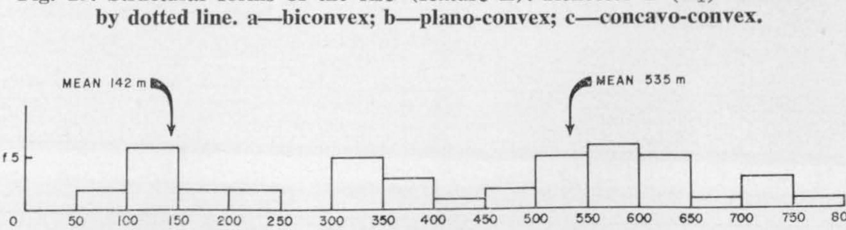

Fig. 14. Frequency distribution of depth of boundary between shelf-edge zone and continental slope. The shallow crossings (50-250 m) are flanked by the Exmouth Plateau, the deep crossings (300-800 m) by the North Australian Basin.

Exmouth Plateau

Conolly (1968) and Whitworth (1969) regard the Exmouth Plateau as a downfaulted or foundered continental block with as much as 3 000 m (Whitworth) or 6 000 m (Conolly) of sediments. Conolly identifies the sediments as mainly late Jurassic and Cretaceous deltaic sediments. Clear reflections are lacking across the zone between the plateau and adjacent shelf, so that the excellent reflectors beneath the plateau cannot be precisely identified. On the basis of the deep reflection work, Whitworth (1969, pl. 11) regards these reflectors as possibly Tertiary, and from its seismic character, I tentatively identify the chief reflector as R_2. Beneath the four traverses across the Exmouth Plateau (92, 99, 144, 146), R_2 lies at a depth of 0.05-0.3 s beneath the sea floor. In traverses 144 and 146 (Fig. 12C), R_2 generally parallels the smooth profile of the sea floor but shallows significantly across the axis of a broad rise, while a deeper reflector, probably R_3, parallels R_2. In traverses 92 and 99, which cross the northern edge of the plateau, R_2 again generally parallels the sea floor, which has a rugged profile due to many canyons. R_2 is deeper (0.3 s) beneath rises than it is beneath canyons, in some of which it is exposed.

In summary, the sequence between R_2 and R_3 has a fairly uniform thickness, but between R_2 and the sea floor the sequence thins beneath the axis of a broad rise. From this, I conclude (a) that the Exmouth Plateau was folded in the later part of the Miocene, and hence at roughly the same time as the folding of Cape Range; and (b) that subsequently the northern flank of the Exmouth Plateau was cut into canyons.

The Exmouth Plateau probably foundered initially in the late Cretaceous during the phase of marginal subsidence that followed continental dispersal. Faulting continued in the Tertiary along the marginal fault that extends from $15\frac{1}{2}°$-$18\frac{1}{2}°$. Landward of the plateau, a wedge of Tertiary sediments, initially made up of detrital sediments and then later of carbonates, was built upwards, and later exposed along marginal faults.

North Australian Basin

The margin facing the North Australian Basin was deposited by a process of upbuilding and outbuilding (Figs 12a-b). Along part of this margin, from 15°40′ to 18°30′S, the upper slope is marked by a down-to-basin fault or monocline. R_2 crops out along the upper slope (Fig. 2c), and in traverse 50 R_3 is seen to crop out on the lower part of the terrace between the upper and lower slopes.

Acknowledgement

Initial compilation of the drawings was the work of Mr D. R. P. Ellis.

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TABLE 1. POSITION AND DEPTH OF MORPHOLOGICAL FEATURES

Feature	B				C			D			E			F			G			H			I		
	Time	*Depth*		Notch	Time	*Depth*		Time	*Depth*		Time	*Depth*		Time	*Depth*		Time	*Depth*		Time	*Depth*		Time	*Depth*	
Traverse		*s*	*m*			*s*	*m*		*s*	*m*		*s*	*m*		*s*	*m*		*s*	*m*		*s*	*m*		*s*	*m*
8	1225	.20	120	—	1940	.31	215	2105	.51	380	2340	.48	344	2355	.52	387									
10/24/35	1530	?.20	120	?	0230	.33	230	2400	.55	410	1220	.50	370	1210	.52	387									
26	1050	.22	136	x	1315	.32	220	1905	.53	393	2010	.50	370	2145	.57	427	2250	.80	600						
27	1350	.25	160	—	1145	.33	230	0720	.58	436	0600	.55	410	0200	.73	550	0120	.95	720						
29	1310	.22	136	x	1500	.33	230	—	—	—	—	—	—	0040	.80	600	0120	1.00	745						
31	1820	.20	120	bank	1705	.27	180																		
33	1930	.21	125	x				0425	.60	450	0540	.55	410												
34					1230	.41	295	1000	.60	450	0930	.56	420	0720	.67	500	0640	.90	675						
36	1810	?.25	160	x							2150	.40	285	0010	.62	460									
38	0900	.21	125	x										0410	.45	327									
40								2400	.35	246	0020	.32	220	0420	.67	500	0550	1.0	745						
42	1735	.24	152	x				1615	.38	270	1505	.34	238	1005	1.07	800									
44	0120	.20	120	x	0240	.30	205							0700	.80	600	0700	.80	600						
46	1650	.21	125	x										1150	.75	560	1115	.90	675						
48	0530	.18	102	x	0615	.22	136							1130	.95	720	1215	1.3	975	1300	1.93	1460	2040	3.22	2430
50	1840	.23	143	x	1820	.27	180							1530	.47	343	1320	.92	690	1200	1.75	1300	0400	4.3	3230
22	1615	.19	110	—										1130	.80	600	1040	1.05	780	1830	1.75	1300			
57	0930	.20	120	x										1400	.84	630	1430	1.05	780						
59														0305	.82	620	0220	1.00	745						
61	1400	.16	72	x										1835	.80	600	1900	.95	720	1950	1.70	1275			
63	1255	.16	72	x										0800	.82	620	0720	1.02	760	0625	1.70	1275			
65	1605	.16	72	x				1750	.55	410	1820	.52	385	2100	.80	600	2155	1.05	780	0200	1.95	1470			
67	1345	.14	54	x	1130	.43	310							0800	.73	550	0500	1.00	745						
69	1130	.14	54	x	1425	.44	320	1520	.60	450	1640	.54	402	2120	.97	725	2200	1.1	825						
71					2330	.40	285	2150	.58	435	2015	.55	410	1440	.95	715	1405	1.1	825	1305	1.80	1350			
73	1800	.14	54	—	2305	.45	328	0130	.57	428	0230	.55	410	0730	.88	660	0750	1.2	900						
75	1145	.16	72	—	0620	.42	302							1840	.70	525	1755	.95	705						
76	1550	.16	72	—	2200	.38	270							0940	.80	600	1025	1.0	745	1155	1.93	1460			
78	1030	.15	63	—										0400	.78	595	0325	.95	705	0200	1.94	1465			
86	0100	.14	54	—				1440	.50	370	1545	.45	328	1830	.73	550	1920	1.13	850						
88	2300	.14	54	—				1030	.45	328	0700	.38	320	0650	.47	343	0450	.87	652						
90	1300	.14	54	—				2210	.55	410	2130	.50	370	2100	.47	343	1850	.84	630						
92	0415	.15	63	—				2040	.32	220	2200	.30	205	2330	.52	387	0120	1.1	825	0230	1.75	1300			
99	1605	.15	63	—	0215	.43	310	1320	.22	136	1500	.20	120				0000	.80	600	2100	2.2	1650			
109	2130	.16	72	—				0450	.21	125	0600	.20	120	1120	.46	336	1310	.75	560						
107	0750	.17	87	—				0230	.27	180	0020	.24	152	2100	.35	247	1810	.64	480						
105	0330	.16	72	x													2200	.7	525						
103								1700	.22	136	0400	.21	125	0530	.23	143	1340	?.95	715						
101	2315	.17	87	x				0020	.21	125	0210	.20	120	0100	.22	136									
55	0330	.17	87	x																					
122								0900	.21	125	1100	.19	110	1200	.21	125									
124														0800	.28	189									
126								2150	.18	102	2300	.17	87	0340	.25	160	0700	.60	450						
128								1815	.21	125	1555	.19	110	1215	.21	125	0910	.83	630						
130														1730	.30	205	1730	.40	285	1745	1.30	975			
132	2330	.16	72					0210	.17	87	0030	.13	25	2330	.16	72	2230	.35	247	2205	1.25	940			
134	1620	.14	54											1620	.14	54	1735	.35	247	1815	1.4	1050			
136																	0600	.34	238	0505	1.35	1010			
144	1120	.17	87	x													1300	.38	270	1430	1.45	1080			
146	0755	.15	63	x													0555	.32	220	0500	1.20	900			
151	1230	.14	54	—										1430	.20	120									
151	1000	.14	54	—										0740	.21	125	0715	.35	247						
148	1920	.18	102	x																					
148	2205	.22	136	x													2230	.31	215						

REVISED CORRELATIONS AND STRATIGRAPHIC NOMENCLATURE IN THE PROTEROZOIC CARBONATE COMPLEX OF THE McARTHUR GROUP, NORTHERN TERRITORY

by K. A. PLUMB and M. C. BROWN

SUMMARY

Detailed remapping of the McArthur Group in the Bauhinia Downs 1:250 000 Sheet area has shown that a previously postulated barrier reef complex is not present, and that supposed 'fore-reef' sediments are in fact younger than the supposed 'back-reef' deposits. New correlations have been made, and the stratigraphic nomenclature revised. The revised stratigraphy is important in the search for possible further bedded sulphide orebodies similar to the H.Y.C.

INTRODUCTION

The McArthur Group is a sequence of Proterozoic carbonate rocks up to 4 200 m thick, cropping out around the southwestern and western shores of the Gulf of Carpentaria from near the Queensland/Northern Territory border to Arnhem Land. It forms part of the type Carpentarian succession of Dunn, Plumb, & Roberts (1966). Its distribution was outlined initially in unpublished work by Mount Isa Mines Ltd during 1955 and later, in more detail, during regional mapping by the Bureau of Mineral Resources (BMR) between 1960 and 1962; 1:250 000 geological sheets of the region have since been published. This paper is principally concerned with the Bauhinia Downs (SE/53-3) 1:250 000 Sheet area (Fig. 1). Here the McArthur Group conformably overlies the Tawallah Group, and is unconformably overlain by the Roper Group.

The group contains the large McArthur River (or H.Y.C.) lead-zinc deposit, thought to be of syngenetic origin. The orebody at Mount Isa is of similar age (Dunn et al., 1966). The group was also believed to contain a large barrier reef complex similar to the Permian reef complex of the Delaware Basin, U.S.A. (Smith, 1964; Plumb & Paine, 1964; Plumb & Rhodes, 1964); brief descriptions have also appeared in Walpole (1962, p. 26) and Brown, Campbell, & Crook (1968, pp. 13-14). The reef was thought to be the oldest barrier reef known (about 1500 million years).

Various company geologists who have worked in the area have expressed doubts about the barrier reef hypothesis, but have not produced conclusive evidence to refute it. In 1968 M. C. Brown began a detailed study of the depositional

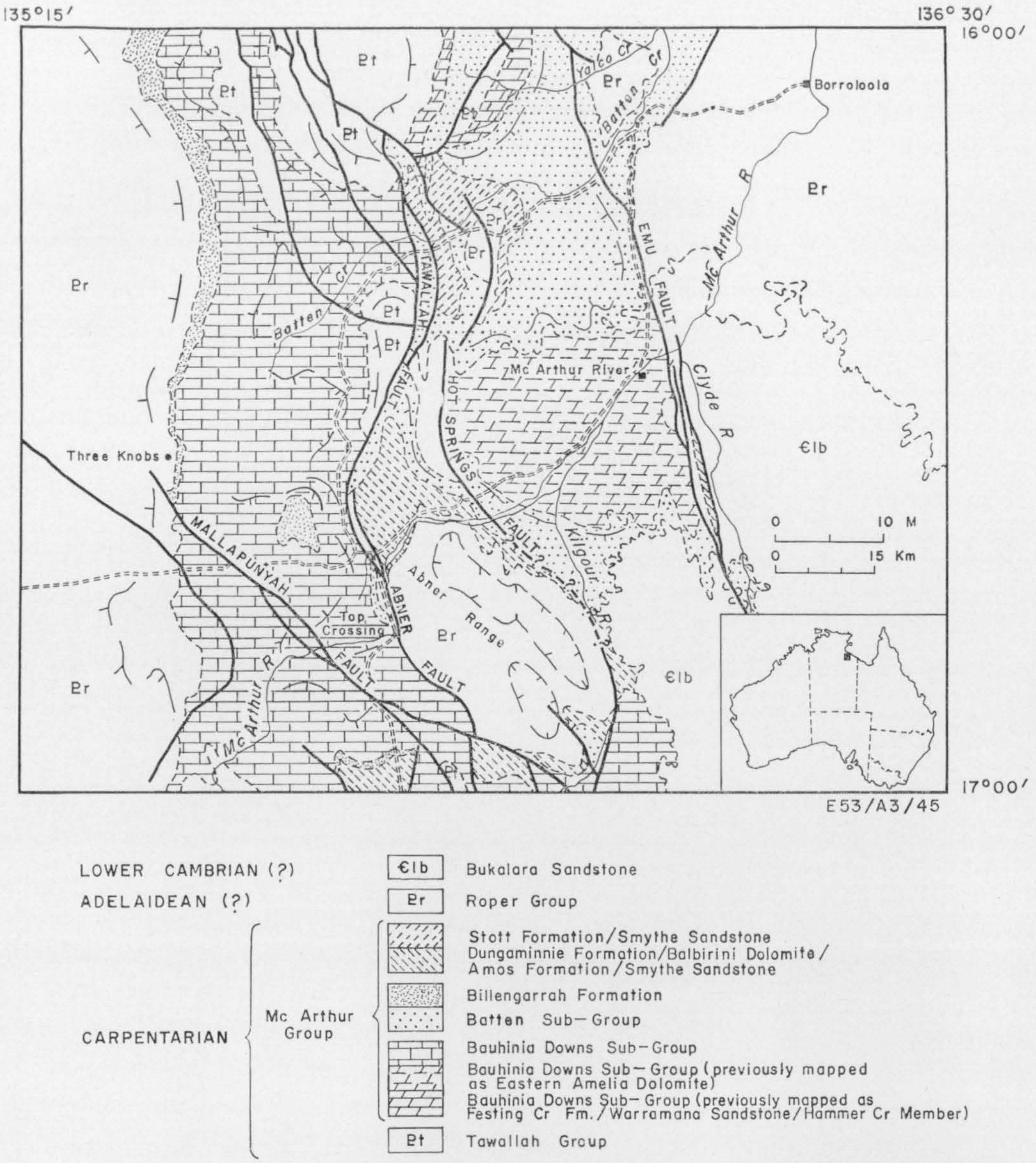

Fig. 1. Generalized geological sketch map, part of Bauhinia Downs Sheet area.

environment of the Barney Creek Formation and related units. As a prelude to this study we both examined some critical areas in 1967 and showed that the reef hypothesis was incorrect.

This paper outlines the revised stratigraphic interpretations and nomenclature of the McArthur River region, and compares them with the old interpretations and nomenclature. Formal definitions and descriptions of the units will appear in a later work by Dunn et al. (in prep.). Further work on the area is summarized in Brown (in prep.). The revised nomenclature has been approved by the Commonwealth Territories Divisional Sub-Committee of the Stratigraphic Nomenclature Committee of the Geological Society of Australia.

Acknowledgements

The BMR party that originally mapped the Bauhinia Downs Sheet area in 1960 consisted of J. W. Smith (Party Leader), H. G. Roberts, K. A. Plumb, and A. W. Webb. Further mapping was carried out in the McArthur Group in 1961, principally by Plumb and Roberts.

We are indebted to Carpentaria Exploration Company Pty Ltd (hereafter referred to as C.E.C.) for making the results of their detailed mapping and their diamond-drill cores available for study. J. A. Shaw and W. Murray (C.E.C.) collaborated with the authors in revising stratigraphic nomenclature of the Group.

BMR Investigations

Previously, two main facies were recognized in the McArthur River region, a back-reef facies (Bauhinia Downs Subgroup) in the west, and a fore-reef facies (Batten Subgroup) in the east (Fig. 2). These were separated by the Top Crossing Dolomite, interpreted as a barrier reef.

The 'back-reef', 'reef', and 'fore-reef' sediments were thought to be the same age and to have been deposited on an older unit, the Amelia Dolomite, which was very much thinner in the west than in the east. The Barney Creek Member of the Amelia Dolomite contained the major lead-zinc deposits of the area. The Amelia Dolomite in the east, although very thick, could not be subdivided at the detail of the regional mapping because of poor outcrop.

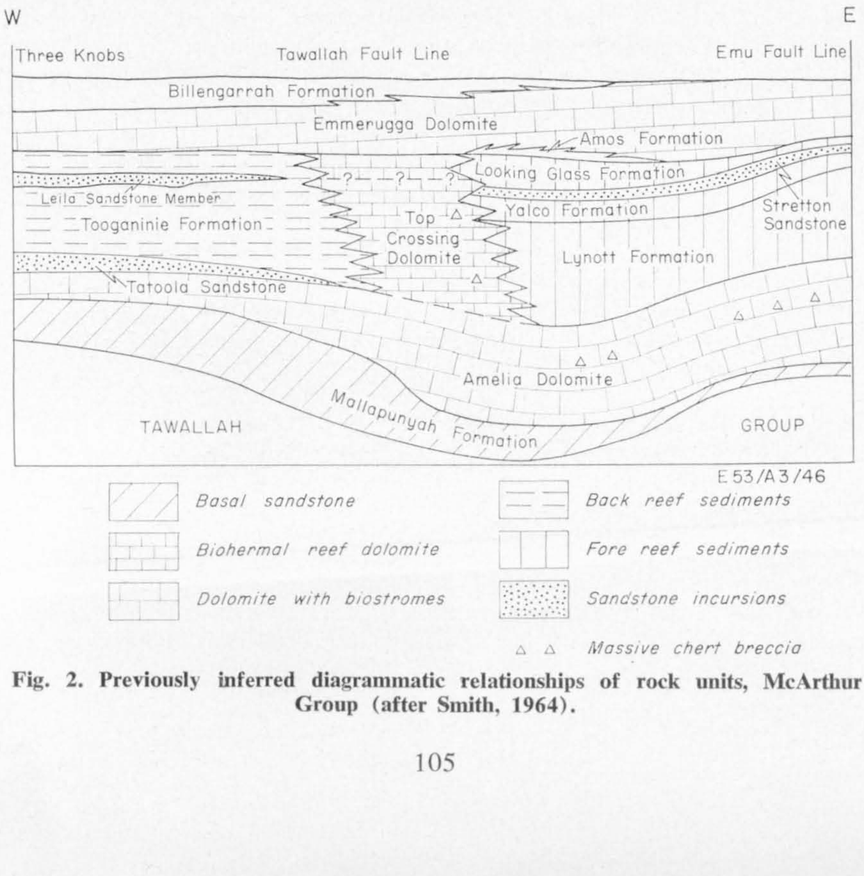

Fig. 2. Previously inferred diagrammatic relationships of rock units, McArthur Group (after Smith, 1964).

The 'back-reef' and 'fore-reef' facies were overlain by a dolomite unit, the Emmerugga Dolomite, which graded laterally into the Amos and Billengarrah Formations. The 'back-reef', 'reef', and 'fore-reef' facies were in contact in only one area, near Top Crossing. Elsewhere the Bauhinia Downs and Batten Subgroups were separated by a major fault, the Tawallah Fault, along which the reef was supposed to have developed.

Around Yalco Creek, in the northeastern part of the Bauhinia Downs Sheet area and east of the postulated barrier reef, a poorly exposed sequence beneath the Lynott Formation did not appear to be normal Amelia Dolomite. It was given new names, the Festing Creek Formation, Warramana Sandstone, and Hammer Creek Member (of the Lynott Formation), and the lower two were correlated with the Amelia Dolomite on stratigraphic position.

Company Investigations

Following a reconnaissance survey of the area, geologists of Australian Aquitaine Petroleum (Duchemin & Zolnai, 1964) expressed doubts as to the existence of the reef but subsequently (Caye & Haskins, pers. comm., 1967) reversed their ideas.

Since 1960 C.E.C. has carried out detailed mapping and diamond drilling around McArthur River, and subdivided the Amelia Dolomite in that area into informal units. Following a reconnaissance of the area west of the Tawallah Fault in 1966 they recognized a consistent lithological correlation between the BMR sequence of Amelia Dolomite, Tatoola Sandstone, Tooganinie Formation, and Emmerugga Dolomite in the west, and their subdivisions of the Amelia Dolomite near McArthur River (J. A. Shaw, pers. comm.; Munt, 1966). They also correlated the Billengarrah Formation with the Batten Subgroup, and recognized the Barney Creek Member at its base. Although this suggested very strongly that the BMR correlations were wrong, they did not map the critical Top Crossing area.

Present Investigations

During September 1967, we visited the McArthur River area together as a prelude to Brown's detailed sedimentological study of the Barney Creek Formation and related units. Study of C.E.C.'s work showed lithological similarities between the sections they correlated in the east and west. The critical Top Crossing area was therefore remapped.

Detailed mapping revealed several errors in the original photo-interpretation and lithological correlations. C.E.C.'s correlations are substantially correct.

Top Crossing Area

Mapping in 1967 showed that the Top Crossing Dolomite lies conformably on the Tooganinie Formation and is, in fact, Emmerugga Dolomite plus Teena Dolomite at the top. These are in turn overlain conformably by a couple of hundred metres of black dolomitic siltstone and dolomite previously mapped as Lynott Formation, but now identified as the Barney Creek Formation and Reward Dolomite.

These beds are disconformably overlain by a thinner than normal Batten Subgroup sequence, from which some of the formations are absent. The Batten Subgroup is then overlain by the Smythe Sandstone, the Amos Formation, the newly named Balbirini Dolomite (previously mapped as Emmerugga Dolomite but lithologically distinct), and finally the Dungaminnie Formation (previously mapped as Billengarrah Formation).

Therefore, the Batten Subgroup clearly overlies the old Bauhinia Downs Subgroup (renamed the Umbolooga Subgroup); they are not lateral equivalents.

Yalco Creek Area

Outcrops that had not been visited during the original mapping showed that the Umbolooga Subgroup sequence, from the Emmerugga Dolomite down, lies beneath the Lynott Formation. The correlations shown in Table 1 were established, and it is recommended that the new nomenclature be used in future.

TABLE 1. COMPARISON OF OLD AND NEW STRATIGRAPHIC NOMENCLATURE IN THE LOWER PART OF THE McARTHUR GROUP, YALCO CREEK AREA, NORTHEASTERN BAUHINIA DOWNS 1:250 000 SHEET AREA

Old Nomenclature (Smith, 1964)		*New Nomenclature*
Lynott Formation		Lynott Formation
Hammer Creek Member	(Chert breccia)	Emmerugga Dolomite
	(Chert, silt, etc.)	Tooganinie Formation
Warramana Sandstone		Tatoola Sandstone
Festing Creek Formation	(Fine sand and silt)	
	(Silicified dolomite)	Amelia Dolomite
	(Purple silt, etc.)	Mallapunyah Formation

Immediately to the north, in the Mount Young Sheet area, basic volcanics were found in the upper part of the Festing Creek Formation during the original mapping, and a relationship suggested with tuffs at H.Y.C. (Plumb & Paine, 1964). The revised correlations now show that they are quite unrelated to the H.Y.C. beds (W-fold Member).

OUTLINE OF REVISED STRATIGRAPHY

Table 2 summarizes the revised stratigraphy. The stratigraphic relationships are illustrated in Figure 3, and Figure 4 shows the inferred relationships of the upper part of the Bauhinia Downs Subgroup in more detail.

The changes made can be summarized as follows:

1. The Batten Subgroup overlies the old Bauhinia Downs Subgroup with local unconformity. The earlier hypothesis of a barrier reef is incorrect.
2. The Top Crossing Dolomite has been identified as Emmerugga Dolomite. The old name is no longer used.

TABLE 2. SUMMARY OF REVISED STRATIGRAPHY, McARTHUR GROUP, McARTHUR RIVER REGION

Rock Unit	*Thickness (in metres)*	*Lithology*	*Stratigraphic Relationships*	*Comparison with Previous Nomenclature*
Kookaburra Creek Formation	up to 300	Oolitic chert, banded chert, quartz sandstone, chert breccia, dolomitic siltstone, stromatolitic dolomite	Unconformably overlain by Roper Group. Conformably overlies Mount Birch Sandstone. Considered to be northwestern stratigraphic equivalent of Stott Formation.	Unchanged.
Mount Birch Sandstone	15 to 60	Feldspathic and quartz sandstone, abundant chert fragments. Chert and quartz conglomerate.	Conformably overlain by Kookaburra Creek Formation. Unconformably overlies Vizard Formation and units as low as Tooganinie Formation of Umbolooga Subgroup. Considered to be northwestern stratigraphic equivalent of Smythe Sandstone.	Unchanged.
Stott Formation	c. 750	Oolitic chert, massive chert, minor dolomite, quartz sandstone.	Unconformably overlain by Roper Group. Conformably overlies Smythe Sandstone. Considered to correlate with Kookaburra Creek Formation in northwest and with Amos Formation plus Balbirini Dolomite and Dungaminnie Formation in south.	Unchanged.
Dungaminnie Formation	c. 150	Siltstone, sandstone, oolitic and stromatolitic dolomite, chert, chert breccia.	Unconformably overlain by Roper Group. Unconformably overlies Balbirini Dolomite.	Previously incorrectly mapped as Billengarrah Formation in Abner Range area.
Balbirini Dolomite	579	Dolomite, generally flaggy, and laminated dolarenite, and dolomitic conglomerate. Some stromatolitic dolomite. Minor dolomitic sandstone, dolomitic siltstone, chert, chert breccia.	Conformably overlain by Dungaminnie Formation. Conformably overlies Smythe Sandstone.	Previously incorrectly correlated with, and mapped as, Emmerugga Dolomite in Abner Range area.
Amos Formation	350	Dolomite, siltstone, shale, local siderite-bearing mudstone.	Conformably overlain by Balbirini Dolomite. Conformably overlies Smythe Sandstone.	Unchanged. Overlying unit was previously wrongly correlated and now has a new name; underlying unit was not previously recognized.
Smythe Sandstone	0 to 180	Chert-quartz sandstone, chert conglomerate.	Overlain conformably by either Amos Formation or Stott Formation. Overlies Looking Glass Formation, apparently conformably but possibly disconformably. South of Abner Range, locally rests with strong disconformity on Mara Member of Emmerugga Dolomite. Correlated with Mount Birch Sandstone.	Unchanged, except previously not recognized beneath Amos Formation in Top Crossing area, where it was mapped as Looking Glass Formation.

TABLE 2—*continued*

Rock Unit	*Thickness (in metres,*	*Lithology*	*Stratigraphic Relationships*	*Comparison with Previous Nomenclature*
Vizard Formation	up to 1000	Dolomitic and cherty siltstone, chert, chert breccia, dolomite, stromatolitic dolomite, dolomitic feldspathic, or quartz-rich sandstone.	Unconformably overlain by Mount Birch Sandstone. Unconformably overlies Mount Reid Beds (Urapunga Sheet area). Correlated with Batten Subgroup but probably also includes equivalents of Umbolooga Subgroup; lower part of section generally not exposed.	Unchanged.
Billengarrah Formation	up to 1200	Chert, sandstone, dolomite, shale, siltstone, chert breccia.	Unconformably overlain by Roper Group. Unconformably overlies Umbolooga Subgroup. Correlated with Batten Subgrorp although individual units cannot be recognized. Locally eroded before deposition of Mount Birch Sandstone.	Essentially unchanged in reference area although, as previously mapped in Three Knobs - Bauhinia Downs homestead area, it included some Reward Dolomite and Barney Creek Formation. Incorrectly identified in Abner Range area.
BATTEN SUBGROUP		Constituent units: Looking Glass Formation Stretton Sandstone Yalco Formation Lynott Formation	Overlain, apparently conformably but possibly disconformably, by Smythe Sandstone. Overlies Umbolooga Subgroup with local unconformity. Correlated with Billengarrah Formation.	Constituent formations, boundaries, distribution, etc. unchanged. Stratigraphic relationships with Umbolooga Subgroup revised.
Looking Glass Formation	80 to 230	Chert, cherty siltstone, chert breccia, subsidiary sandstone.	Top unit of Batten Subgroup. Overlain conformably or possibly disconformably, by Smythe Sandstone. Overlies Stretton Sandstone conformably. Locally, in Top Crossing area, directly overlies Yalco Formation.	Unchanged except for modified stratigraphic relationships. Includes rocks previously incorrectly mapped as Stretton Sandstone in Top Crossing area. Rocks which were mapped there as Looking Glass Formation are mainly Smythe Sandstone.
Stretton Sandstone	30 to 245	Flaggy quartz sandstone, sometimes micaceous.	Overlain conformably by Looking Glass Formation. Conformably overlies Yalco Formation.	Unchanged, except no longer present in Top Crossing area as originally mapped.
Yalco Formation	c. 130	Chert, siltstone, shale, chert breccia.	Conformably overlain by Stretton Sandstone and conformably overlies Lynott Formation. Cannot be differentiated from Lynott Formation in Top Crossing area.	Unchanged.
Lynott Formation	520 to 760	Dolomitic siltstone, dolomitic sandstone, quartz sandstone, dolomite, shale, chert breccia, dolomite breccia, tuffaceous siltstone.	Lowest unit of Batten Subgroup. Overlies with local unconformity, Umbolooga Subgroup. Overlain conformably by Yalco Formation.	Unchanged except rocks previously mapped as Lynott Formation in Top Crossing area are now Reward Dolomite and Barney Creek Formation.
Donnegan Member (of the Lynott Formation)	105	Quartz sandstone, dolomitic sandstone.	Occurs at top of Lynott Formation.	Unchanged.

TABLE 2—*continued*

Rock Unit	*Thickness (in metres)*	*Lithology*	*Stratigraphic Relationships*	*Comparison with Previous Nomenclature*
UMBOLOOGA SUBGROUP		Constituent units: Reward Dolomite Barney Creek Formation Cooley Dolomite Member H.Y.C. Pyritic Shale Member W-Fold Shale Member Teena Dolomite Coxco Dolomite Member Emmerugga Dolomite Mitchell Yard Dolomite Member Mara Dolomite Member Tooganinie Formation Myrtle Shale Member Leila Sandstone Member Tatoola Sandstone Amelia Dolomite Mallapunyah Formation	Overlain with local unconformity by Batten Subgroup. Conformably overlies Tawallah Group.	Replaces old Bauhinia Downs Subgroup because definition contains new constituent units including units previously mapped as 'Eastern Amelia Dolomite'. Stratigraphic relationships with Batten Subgroup also revised. Contains a conformable sequence of rocks of related origins, distinctly different to overlying Batten Subgroup.
Reward Dolomite	c. 300	Dolomite with characteristic chert pellets, dolomitic sandstone and shale, tuffaceous quartz arenite, breccia, tuffs, local stromatolitic dolomite.	Upper unit of Umbolooga Sub-Group, overlain by Batten Subgroup. Conformably overlies Barney Creek Formation.	Previously mapped as part of the 'Eastern Amelia Dolomite'. Minor exposures mapped as Lynott Formation near Top Crossing and near Reward Prospect as lower part of Billengarrah Formation in west, and upper part of Emmerugga Dolomite in southwest. Corresponds to Reward Dolomite and Deep Creek Dolomite of Cotton (1965).
Barney Creek Formation	up to 530	Characteristically dolomitic, tuffaceous, bituminous and pyritic shales. Dolomite breccia and graded dolarenite.	Conformably overlies Teena Dolomite. Conformably overlain by Reward Dolomite, locally unconformably overlain by Batten Subgroup.	Corresponds roughly to previous Barney Creek Member of 'Eastern Amelia Dolomite' but does not include Laminated Dolomite of Cotton (1965). Previously, locally mapped as Lynott Formation in Top Crossing area, or in lower part of Billengarrah Formation or upper part of the Emmerugga Dolomite in west and south.
Cooley Dolomite Member (of Barney Creek Formation)	490	Massive and brecciated dolomite and dololutite, stromatolitic dolomite. Minor sandstone and mudstone. Complex of reef, lagoon, and basinal dolomites.	Conformably overlies W-Fold Shale Member. Conformably overlain by Reward Dolomite. Interfingers with H.Y.C. Pyritic Shale Member.	Newly discovered member; originally recognized in subsurface.

TABLE 2—*continued*

Rock Unit	*Thickness (in metres)*	*Lithology*	*Stratigraphic Relationships*	*Comparison with Previous Nomenclature*
H.Y.C. Pyritic Shale Member (of Barney Creek Formation)	0 to 490	Bituminous and tuffaceous pyritic shale, tuff, bituminous dolomitic shale, dolarenite, dolomite breccia, bedded sphalerite and galena.	Conformably overlain by Reward Dolomite. Conformably overlies W-Fold Shale Member. Only found in Bulburra Depression; lenses out elsewhere. Upper and lower boundaries probably diachronous. Interfingers with Cooley Dolomite Member near Emu Fault.	Contains H.Y.C. orebody. Previously mapped as part of Barney Creek Member. Corresponds to H.Y.C. Pyritic Shales of Cotton (1965).
W-Fold Shale Member (of Barney Creek Formation)	15 to 150	Green tuff; red potash-rich tuffaceous mudstone; dolomitic tuff; bituminous dolomitic shale; minor dolomite breccia, limestone breccia.	Conformably overlain by H.Y.C. Pyritic Shale Member or Cooley Dolomite Member. Only recognized in Bulburra Depression. Passes laterally into dolomitic shales of undifferentiated Barney Creek Formation. Conformably overlies Teena Dolomite. Upper and lower boundaries probably diachronous.	Previously mapped as part of Barney Creek Member. In H.Y.C. area corresponds to Green Vitric Tuffs of Cotton (1965).
Teena Dolomite	57 (reference section)	Laminated to thick bedded dololutite sometimes containing stromatolites, dolomite flake breccia, dolarenite, dolomitic sandstone, silty dolomite, rare halite casts, occasional potassium-rich mudstone.	Conformably overlain by Barney Creek Formation. Locally unconformably overlain by Batten Subgroup or Billengarrah Formation. Conformably overlies Emmerugga Dolomite.	New formation. Previously mapped as part of Emmerugga Dolomite. Transitional unit between Emmerugga Dolomite and Barney Creek Formation. Corresponds to Laminated Dolomite of Cotton (1965).
Coxco Dolomite Member (of Teena Dolomite)	15 to 70	Massive dololutite with occasional interbeds of potassium-rich mudstone. Characteristic pseudomorphs of radiating (?) aragonite needles.	Occurs at top of Teena Dolomite conformably overlying undifferentiated Teena Dolomite. Conformably overlain by Barney Creek Formation.	Newly recognized member.
Emmerugga Dolomite	300 (reference section)	Dolomite, with or without stromatolites. Minor breccia, dolomitic siltstone and sandstone, potassium-rich (?) tuffaceous mudstone, solution collapse dolomite breccias. Superficial chert breccias common in outcrop.	Conformably overlain by Teena Dolomite. Conformably overlies Tooganinie Formation.	Includes all the rocks except the Teena Dolomite previously mapped as Emmerugga Dolomite in the west. Locally previously mapped as Top Crossing Dolomite at Top Crossing and as part of 'Eastern Amelia Dolomite' or Hammer Creek Member in east.
Mitchell Yard Dolomite Member (of Emmerugga Dolomite)	15 to 120	Thick bedded clean dololutite. Very minor potassium-rich mudstone.	Top member of Emmerugga Dolomite. Conformably overlain by Teena Dolomite. Conformably overlies Mara Dolomite Member.	Newly recognized member. previously mapped as part of Emmerugga Dolomite in west and 'Eastern Amelia Dolomite' in east.

TABLE 2—*continued*

Rock Unit	*Thickness (in metres)*	*Lithology*	*Stratigraphic Relationships*	*Comparison with Previous Nomenclature*
Mara Dolomite Member (of Emmerugga Dolomite)	c. 240	Cherty dololutite with stromatolites and algal laminations. Minor flake breccia, dolarenite, dolomitic siltstone, dolomitic sandstone, potassium-rich mudstone, solution collapse dolomite breccias. Outcrops commonly altered to chert breccia.	Lower member of Emmerugga Dolomite. Conformably overlain by Mitchell Yard Dolomite Member. Conformably overlies Tooganinie Formation.	Newly recognized member, previously mapped as part of Emmerugga Dolomite in west and 'Eastern Amelia Dolomite' in east.
Tooganinie Formation	807	Dolomite, with chert bands and nodules, often stromatolitic, occasionally oolitic. Dolomitic sandstone, dolomitic siltstone, dolomitic shale, subsidiary ferruginous shale, intraformational conglomerate; halite casts common.	Conformably overlain by Emmerugga Dolomite. Locally unconformably overlain by Mount Birch Sandstone. Conformably overlies Tatoola Sandstone.	Corresponds to Tooganinie Formation as originally mapped in the west. In the east previously included in 'Eastern Amelia Dolomite' or in Hammer Creek Member.
Myrtle Shale Member (of Tooganinie Formation)	30 to 240	Dolomitic shale; minor dolomitic sandstone, silty dolomite.	Top member of Tooganinie Formation. Conformably overlain by Emmerugga Dolomite. Conformably overlies Leila Sandstone Member.	New member. Previously mapped as undifferentiated Tooganinie Formation overlying Leila Sandstone Member. 'Red beds' characteristic.
Leila Sandstone Member (of Tooganinie Formation)	c. 140	Dolomitic sandstone, sandy dolomite, quartz sandstone. Subsidiary dolomite, often stromatolitic.	Conformably overlain by Myrtle Shale Member. Conformably overlies undifferentiated Tooganinie Formation.	Corresponds to original Leila Sandstone Member.
Tatoola Sandstone	c. 140	Quartz sandstone, feldspathic sandstone, dolomitic sandstone. Minor dolomite and siltstone, more abundant near base.	Conformably overlain by Tooganinie Formation. Conformably overlies Amelia Dolomite.	Corresponds to Tatoola Sandstone as originally mapped in west. In east originally mapped as Warramana Sandstone (with which it was correlated) or as part of 'Eastern Amelia Dolomite'.
Amelia Dolomite	90 to 240	Bedded dolomite, commonly stromatolitic. Regular silicified beds. Some thin beds of dolomite breccia, green shale, white calcilutite.	Conformably overlain by Tatoola Sandstone. Conformably overlies Mallapunyah Formation.	Corresponds to Amelia Dolomite previously mapped in the west. In the east previously mapped as part of 'Eastern Amelia Dolomite', or Festing Creek Formation, with which it was correlated.
Mallapunyah Formation	30 to 750	Quartz sandstone, ferruginous sandstone, ferruginous shale, dolomite, dolomitic siltstone, chert, sideritic bands.	Lowest unit of McArthur Group and Umbolooga Subgroup. Conformably overlain by Amelia Dolomite. Conformably overlies Tawallah Group.	Corresponds to Mallapunyah Formation as previously mapped. In northeast, was included in Festing Creek Formation, with which it was correlated. 'Red beds' characteristic.

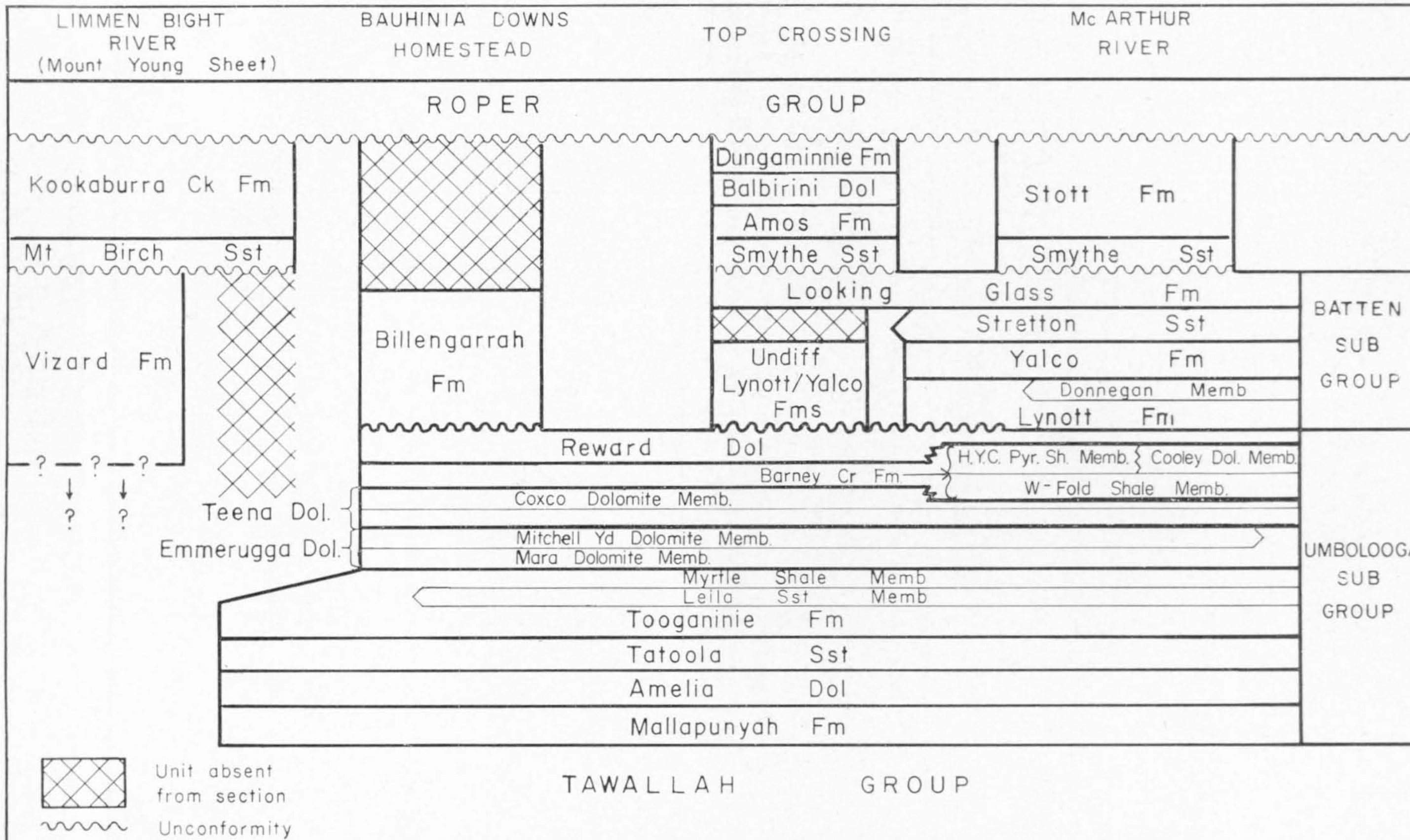

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Fig. 3. Stratigraphic correlation chart, McArthur Group, McArthur River Region.

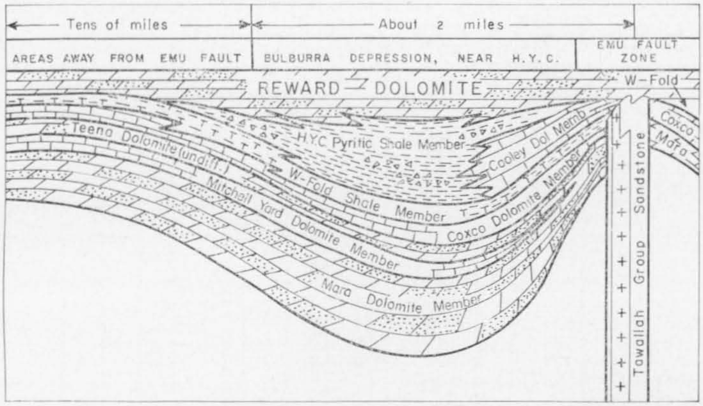

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Fig. 4. Diagrammatic relationship of rock units, upper part of Bauhinia Downs Subgroup.

3. The rocks previously mapped as Amelia Dolomite east of the Tawallah Fault correlate with the old Bauhinia Downs Subgroup plus Emmerugga Dolomite and Amelia Dolomite in the west.

4. The name Bauhinia Downs Subgroup is abandoned and a new subgroup, the Umbolooga Subgroup, is defined to include all the units previously incorrectly mapped as Amelia Dolomite in the east, i.e. the Reward Dolomite, Barney Creek Formation, Teena Dolomite, Emmerugga Dolomite, Tooganinie Formation, Tatoola Sandstone, and Amelia Dolomite (as mapped in the west) plus the Mallapunyah Formation. All these units can now be mapped both east and west of the Tawallah Fault.

5. The Barney Creek Member has been upgraded to formation status and its boundaries modified to suit the regional picture. A new formation, the Teena Dolomite, is separated out from the top of the old Emmerugga Dolomite and another new formation, the Reward Dolomite is recognized at the top of the Umbolooga Subgroup.

6. The Barney Creek Formation, the Teena Dolomite, and the Emmerugga Dolomite are subdivided into several members, which are of significance to the study of the regional setting of the H.Y.C. orebody.

7. The Billengarrah Formation is correlated with the Batten Subgroup. Dolomite and shale occurring at its base in some areas are now mapped as Reward Dolomite and Barney Creek Formation.

8. Units overlying the Amos Formation at Top Crossing, and previously mapped as Emmerugga Dolomite and Billengarrah Formation, are now renamed Balbirini Dolomite and Dungaminnie Formation respectively. They are still correlated with the Stott Formation (Fig. 3).

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