

Neogene stratigraphy, sedimentation and petroleum potential of the Oiapu-Yule Island-Oroi Region, Papua New Guinea

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Abstract: The Oiapu-Yule Island-Oroi region is an onshore and offshore triangular area bounded approximately by 145°30'E 8°00'S, 146°30'E 7°45'S and 147°00'E 9°20'S. It lies within the Mesozoic to Cainozoic Papuan Basin. Geological mapping and stratigraphic/structural studies carried out since 1982 by members of the Geological Survey of Papua New Guinea and the University of Papua New Guinea Geology Department have led to a greater understanding and revision of the geological history and petroleum potential of this region. The Miocene sequence has been found to consist largely of bathyal mudstones and turbiditic sandstones (Aure association). However, convergent deformation initiated a phase of folding in latest Middle Miocene times, culminating with thrusting in the Late Miocene and creating the south-eastern Papuan Fold Belt. This deformation was associated with basaltic volcanism which gave rise to the pyroclastics and volcanoclastics of the Talama Formation. In the cores of some rising anticlines, neritic conditions were established giving rise to local carbonate buildups such as the Ou-Ou Limestone Member. Within anticlinal cores, rocks of the Aure association and Talama Formation were often eroded and unconformably overlain by the latest Miocene/Early Pliocene Orubadi beds.

During the Early Pliocene there was a general regression and an influx of coarser detritus from rising mountainous areas to the northeast, giving rise to the inner neritic and terrestrial Era beds. There was a second phase of thrusting in Late Pliocene/Early Pleistocene times.

Most of the thrusts and fold hinge surfaces formed in these two phases of deformation dip to the northeast, though some structures dip to the southeast. The fold-and-thrust structures provide potential traps for petroleum, and suitable reservoir rocks occur in the pyroclastics, volcanoclastics and carbonates of the Talama Formation. The southeastern part of the fold belt has a long history of unsuccessful exploration, largely because of complex mesoscale geology and because the macroscale structure was not understood. Consequently most exploratory wells were drilled off-structure and failed to penetrate their target horizons within areas of closure. Current knowledge of the regional palaeogeography and structure is a key factor upon which a more successful exploration programme might be based.

INTRODUCTION

The Oiapu-Yule Island-Oroi region (Figure 1) consists mainly of coastal plains and foothills, together with estuaries and adjacent parts of the continental shelf. Access is via the Hiritano Highway from Port Moresby.

The first geological work in this region was carried out by Stanley (1911, 1913) who investigated a number of gas seeps and concluded that the region was prospective for petroleum.

Subsequently the Anglo-Persian Oil Company conducted geological surveys on behalf of the Australian Government (Montgomery, 1930). Montgomery and his co-

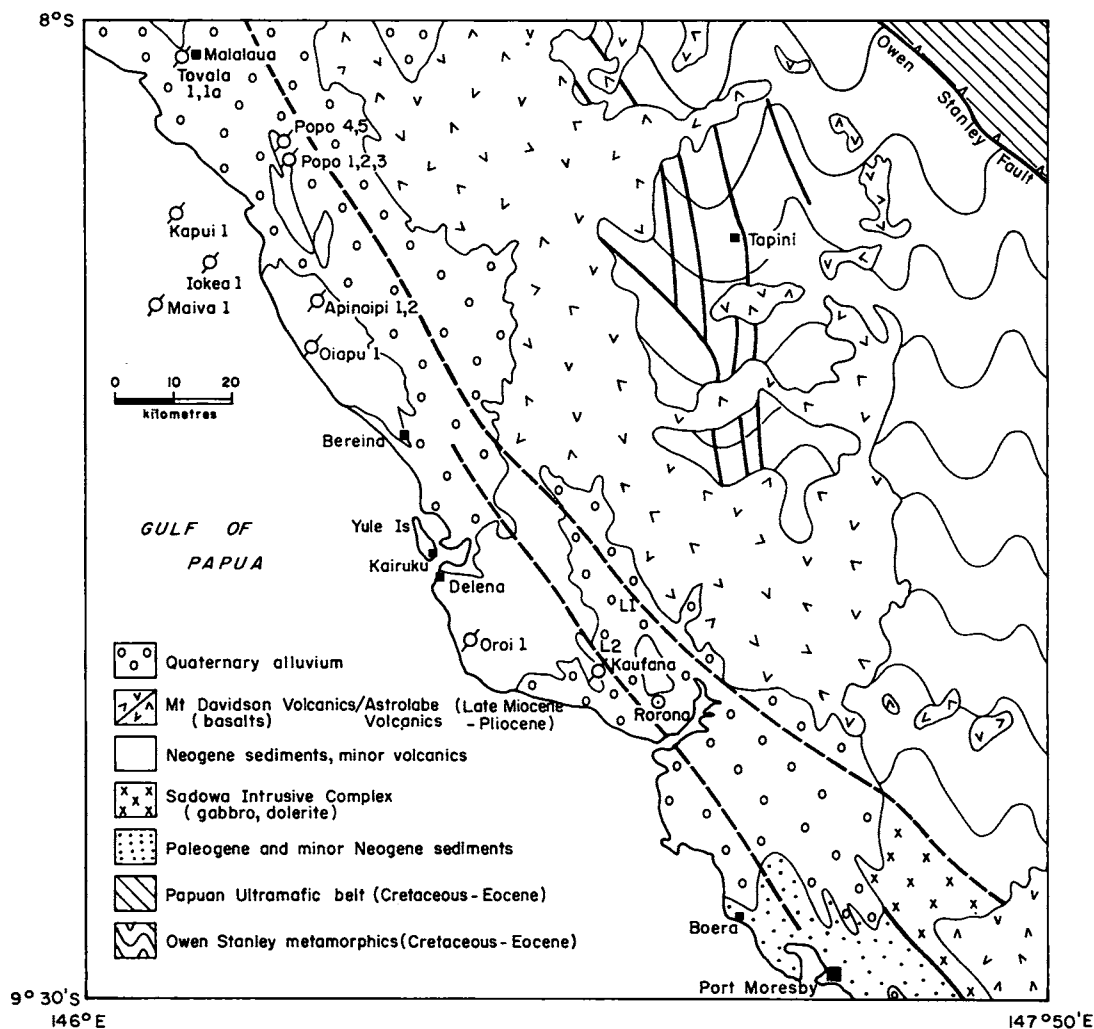


Fig. 1. Regional geology of the Oiapu-Yule Island-Oroi region, PNG (Modified from Brown, 1977).

workers prepared a 1:63,360 geological map of Delena and the southern part of Yule Island. In 1929, Anglo Persian drilled a series of shallow test bores in the Popo Anticline, which is located on the mainland about 60 km to the north of Yule Island. Drilling problems were caused by flowage in bentonitic Pliocene mudstones and the bores only encountered gas shows.

From 1938–1942, the Papuan Apinaipi Petroleum Company was active on the mainland about 35 km north of Yule Island. Two exploration wells and five shallow test bores were drilled on the Oiapu Anticline by the Papuan Apinaipi Co. in 1942 and Millward and Stach (1941) studied the stratigraphy of Yule Island and Delena. In 1949

the Australian Petroleum Company (A.P.C., 1949) drilled Oroï 1 within Permit 7 to a total depth of 1681 m in Miocene sediments. Only gas shows were encountered. APC surrendered the area in 1950, but Papuan Apinaipi continued exploration until 1960. Allen (1957) remapped Yule Island and Delena recognising the presence of northerly to northwesterly trending faults. Kaufana 1 was drilled on the Kaufana Anticline to the southeast of Delena, reaching a total depth of 1030 m in 1958. The well was a dry hole which terminated in early Middle Miocene greywacke and mudstone. Stanley (1960) made an extremely comprehensive and detailed review of the petroleum prospects of Permit 22, which included many specific recommendations on drilling targets. However Papuan Apinaipi ceased active exploration at this time and most of the prospects recommended by Stanley remain untested.

During the late sixties, offshore exploration was carried out by Phillips Australian Oil Company, which drilled the Iokea 1, Kapuri 1 and Maiva 1 wells about 50 km northwest of Yule Island. Kapuri 1 was drilled on a seismically defined pinnacle reef of probable Pliocene age, whereas the other two wells were drilled on seismically defined anticlines (Patterson, 1968a). Gas shows were encountered from 1500–240 m in Maiva 1. Both Maiva 1 and Iokea 1 terminated in Late Miocene volcanics which were considered to be economic basement.

In the same period, Marathon Petroleum and Basin Oil did onshore seismic surveys and drilled the Tovala 1 and 1A wells. Rapid changes of dip within the 2000–2700 m interval in Tovala 1A were thought to be associated with easterly dipping thrusts (Basin Oil, 1969). The well terminated within Middle Miocene mudstone at a total depth of 3207 m. Tovala 1 was abandoned at a depth of 1372 m because of problems caused by overpressured mudstone and gas pockets.

Recently, a marine seismic survey was conducted for Kina Oil and Gas. The northwestern part of the region lies within the southeastern extremity of Petroleum Prospecting Licence (PPL) 30 which is held by Kina. The areas to the north, east and southeast have not been under licence since 1970. The region has been mapped at 1:250,000 scale by the Geological Survey of Papua New Guinea (Brown, 1977) and the stratigraphic names used on that map were defined by Brown (1975).

The research which is reported in this paper commenced with large scale geological remapping and stratigraphic studies of Yule Island (Francis *et al.*, 1982) (Figure 2) and Delena (Perembo, 1983). Work is currently in progress on the remapping of the Bereina-Oiapu region and a reinterpretation of the geology of Oroï 1.

Biostratigraphy

In these recent studies, planktic foraminiferids have been used to biostratigraphically subdivide the succession. The zonation adopted here follows Blow (1969, 1979), except for the N12/N13 boundary which is positioned at the extinction level of *Globorotalia fohsi* s.l. rather than at the first appearance of *Sphaeroidinellopsis subdehiscens* (following Srinivasan and Kennett, 1981).

The correlation of the zones with stages and the palaeomagnetic epochs follows the work of Ryan *et al.* (1974) on type-sections of the various stages. It should be noted

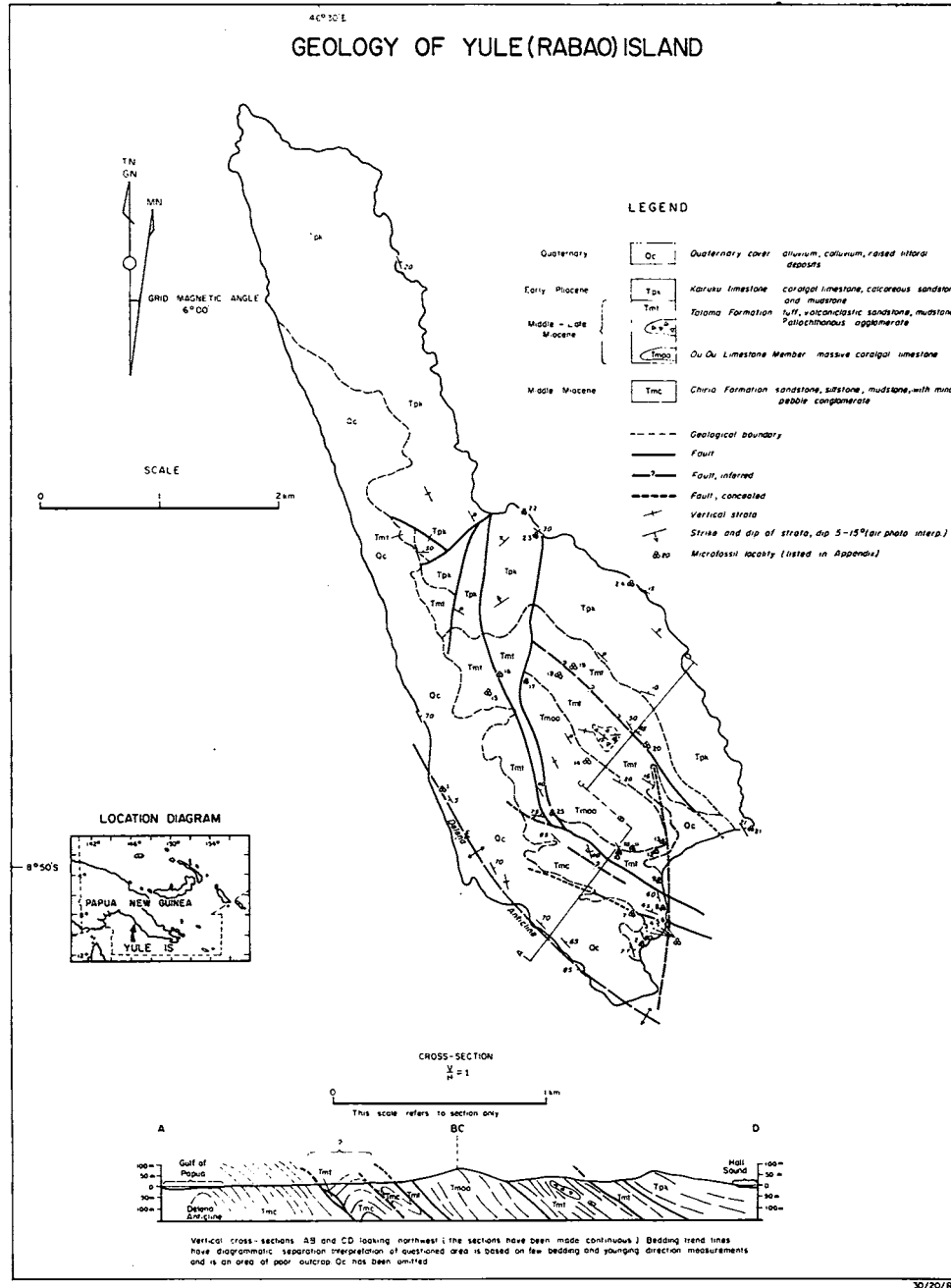


Fig. 2. Geology of Yule Island.

that the palaeomagnetic correlation differs somewhat (particularly for the lower and middle Miocene) from that made indirectly by Keller (1981) in the north Pacific. Absolute ages assigned to the palaeomagnetic epochs are taken from Lowrie and Alvarez (1981, Figure 4).

The ratio of planktic to benthic foraminiferids in an assemblage is related to water depth. The ratios found in studied samples have been compared to those reported from Recent neritic and bathyal sediments of northern Australia by Palmieri (1976) and Betjeman (1969).

Many of the fossil benthic foraminiferids recovered from samples can be identified with extant species. The modern depth distributions of these forms can therefore be used to determine palaeobathymetry. From the western Pacific region, modern distributions of selected species have been recorded by Cushman (1921, Philippine region), Frerichs (1970, Andaman Sea), Palmieri (1976, central Queensland shelf), and Hughes (1977, Honiara Bay, Solomon Islands).

Samples have been assigned to bathymetric zones following the classification of Ingle (1980) which is summarized below:

Inner Neritic zone:	low tide–50 m
Outer Neritic zone:	50 m–150 m
Upper Bathyal zone:	150 m–500 m
Upper Middle Bathyal zone:	500 m–1500 m

LITHOSTRATIGRAPHY

Pre-Neogene Lithostratigraphy

Little is known of the distribution of pre-Neogene rocks within this region, as there are no recorded outcrops of Mesozoic or Palaeogene rocks and all wells drilled terminated within Neogene strata. Therefore the nature of the pre-Miocene sequences can only be inferred from other regions some distance to the southeast, northeast and northwest such as Port Moresby, Tapini and Tauri. It is probable that much of the region is underlain by Jurassic to Cretaceous strata of the Wahgi Group.

In the Tapini region, a Late Cretaceous sequence of interbedded bathyal mudstone, siltstone and muddy sandstone was mapped by Brown (1977) and earlier workers as part of the Auga Beds. It now seems likely that this unit is the Chim Formation, which is the youngest unit in the Wahgi Group.

Within the Port Moresby region is a well developed but extensively thrust sequence of latest Cretaceous and Palaeogene lower bathyal to uppermost abyssal pelagic sediments (Rogerson *et al.*, 1981) which we refer to here as the Port Moresby association. Eocene cherts, siliceous mudstones and carbonates of this association occur as thrust slices in the Tapini and Tauri regions and were reported by Stanley (1960) to be the bottom hole formation in Rorona 1.

These rocks are often thrust against middle to upper bathyal mudstones,

turbiditic sandstones and occasional distal pyroclastics of mid-Oligocene to Miocene age which we refer to here as the Aure association. The latter association includes the Aure beds *sensu stricto*, the Omaura Formation of the Tauri and Kainantu regions (Rogerson *et al.*, 1982), the "Dokuna Tuff", "Boira Formation" and "Fairfax Formation" of the Port Moresby region (Rogerson *et al.*, 1981) and the Chiria Formation which is described below. This association consists of upper to middle bathyal turbiditic sediments.

Neogene Stratigraphy

Few data are available on Early Miocene strata within the Oiapu-Yule Island-Oroi region. Rocks of this age have been reported by Brown (1977) from the Kaufana Anticline, but the dating was based on transported benthic foraminiferids. Since derived Early Miocene benthics commonly occur in the Middle Miocene Chiria Formation, such reports must be regarded as suspect until planktic foraminiferal dating of the relevant strata has been carried out.

The nature of the post-Early Miocene sequence is summarised in Table 1 and in Figures 3 and 4.

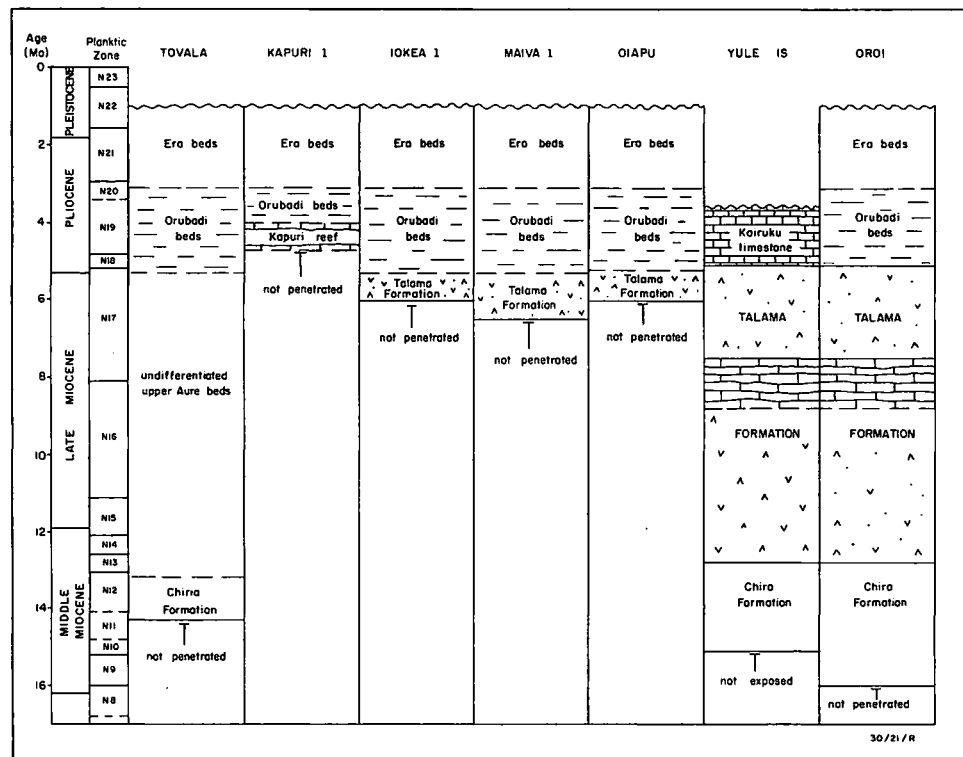


Fig. 3. Neogene time-stratigraphic diagram for Oiapu-Yule Island-Oroi region, PNG.

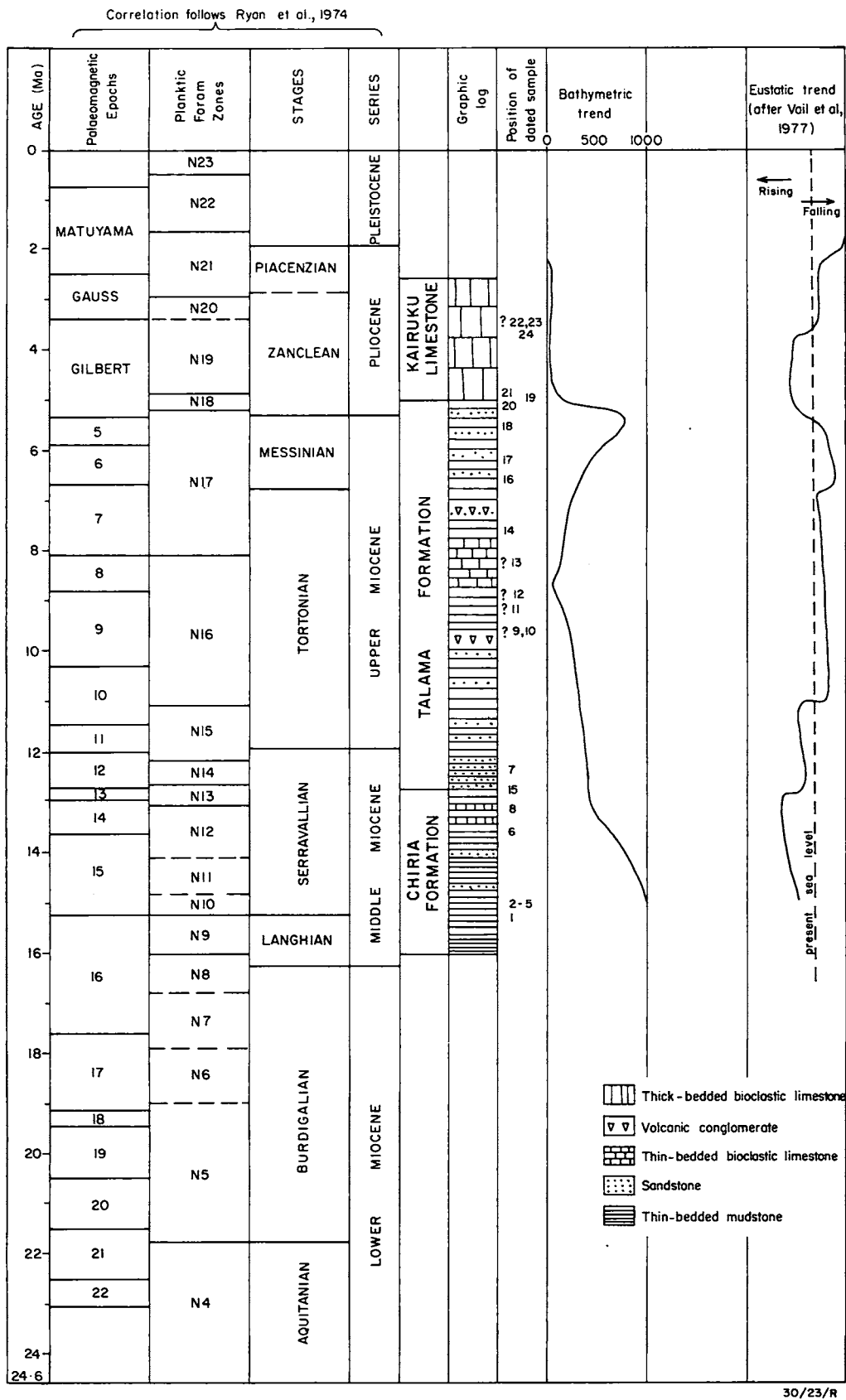


Fig. 4. Summary time/stratigraphic diagram for Yule Island incorporating palaeobathymetric data derived from foraminiferid population studies. Locations of dated samples are plotted on Figure 2.

TABLE 1
SUMMARY OF NEOGENE STRATIGRAPHY, DEPOSITIONAL ENVIRONMENTS AND PETROLEUM POTENTIAL

Formation	Age	Rock types	Thickness	Sedimentary environment	Petroleum potential
Era beds	no older than N19/N20	Calcareous and non-calcareous tuffaceous sandstone, pebble and cobble conglomerate, siltstone and mudstone; minor limestone	?2000m	Inner neritic to terrestrial	Immature source rocks. Some reservoir potential in sandstones lower in formulation overlain by mudstone.
Kairuku limestone	N18 (in part)	Medium to thick bedded yellow-brown to pinkish grey biocrinoid, biosparite and biomicrite with interbeds of calcareous mudstone and siltstone.	250m	Outermost to inner neritic; a regressive sequence.	Possibility of pinnacle reefs. Has high porosity.
Orubadi beds	mid N17-N19/N20	Bluish-grey mudstone with occasional interbeds of siltstone and	up to 700m	Upper bathyal to inner neritic; a regressive sequence.	Forms seal over Talama Formation.
Talama formation	N13-basal N18	Bluish-grey siltstone and mudstone, pinkish-grey to yellowish-brown muddy calcareous sandstone, lithic-crystal tuff, allochthonous agglomerate blocks. Biomicrudite and biomicrite lenses are also present.	400-1500m	Upper bathyal to inner neritic, variations in water depth due to synsedimentary tectonism.	Porous and permeable pyroclastics and reefal carbonates constitute potential reservoirs rocks.
Chiria formation	N9-13	Thinly bedded bluish-grey to greenish-grey carbonaceous mudstone and muddy sandstone with interbeds of grey calcareous siltstone, pebbly sandstone, muddy algal-foraminiferal biomicrite, biosparite and occasional 20cm coal seams.	300m	Foraminiferal assemblages indicate that the Chiria Formation was deposited in the bathyal zone.	Sandstones generally tight and argillaceous, mudstones a poor to fair gas scarce.

Chiria Formation

Synonymy: Chiria Formation and most of Pupunina Formation (Millward and Stach, 1941; Allen, 1957) Chiria Formation (Brown, 1975).

Type Section: The intertidal rocks platforms on Yule Island from GR 49492262 to GR 49572328 (Yule Island geological sheet of Francis *et al.*, 1982).

Lithology: The unit is composed of thinly bedded, bluish-grey to greenish-grey carbonaceous mudstone and muddy sandstone with interbeds of grey calcareous siltstone, pebbly sandstone, muddy algal-foraminiferal biomicrite, biosparite and occasional (< 20 cm) coal seams.

Age: Middle Miocene (N9–N13)

Thickness: Millward and Stach (1941) and Allen (1957) believed that the Chiria Formation exposed on Yule Island had a thickness of 600 m and that the overlying Pupunina Formation had a thickness of 350 m. We consider that there has been much repetition of strata through structural complications which these workers overlooked and that the actual thickness of this sequence is considerably less than the above estimate. Although it is very difficult to accurately estimate stratigraphic thicknesses in deformed sequences like this, we suspect that the Chiria Formation in the Yule Island-Oroi region is no more than 400 m thick. In Tovala 1A, the Chiria Formation occurs from 1800 m to 3207 m but there is repetition of strata by thrusting in this part of the well.

Stratigraphic Relationships: Within its type section, the Chiria Formation is in fault contact with N16 to lower N17 mudstones of the Talama Formation. At Delena there is a conformable and gradational contact between the Chiria Formation and the overlying Talama Formation. N14 mudstones from the lower part of the Talama Formation occur in a fault sliver a few hundred metres to the northwest of the Chiria Formation type section (micro-fossil locality 7, Figure 2) indicating that the absence of N14–N15 strata from most parts of Yule Island is due to faulting rather than to a depositional hiatus.

The base of the Chiria Formation is not exposed within the Yule Island-Delena region and Oroi 1 terminated within the Chiria Formation. It is probable that the Chiria Formation is underlain by the “Fairfax Formation” of Rogerson *et al.* (1981), a sequence of N6–N7 bathyal tuffaceous sandstones and mudstones with biosparite and micrite interbeds, which crops out in the Port Moresby region about 70 km to the southeast of Delena. The Chiria Formation is in part a lateral equivalent of the neritic “upper Tf”, or Tf₂ Kido Limestone of Pieters (1978), which occurs in the Redscar region to the southeast of Oroi.

Palaeoenvironment: The fine clastics of the Chiria Formation are dominated by planktic foraminiferids, with assemblages including Middle Miocene indices such as *Orbulina universa* and the *Globorotalia fohsi* group.

The foraminiferal assemblages indicate an upper to middle bathyal environment

with water depths generally exceeding 300 m (Figure 4). However, transported neritic-benthic foraminiferids such as *Lepidocyclina* are common, particularly in the turbiditic sands. The detrital grains are derived almost entirely from the Owen Stanley Metamorphics and from a basaltic to andesitic volcanic source. The presence of labradorite suggests that there was penecontemporaneous pyroclastic vulcanism within or adjacent to the basin, since calcic plagioclase is easily weathered and does not usually survive reworking from older rock units. The metachert is probably derived from Eocene rocks of the Port Moresby association. This inference is supported by the occurrence of chert clasts and limestone clasts containing Eocene foraminiferids within the Chiria Formation at Delena (Montgomery, 1930; Perembo, pers. comm.). Metachert can form through deformation under low grade metamorphic conditions and does not require medium or high grade regional metamorphism of the type which formed the Owen Stanley Metamorphics. The metamorphic detrital grains in the Chiria Formation are generally low grade, but Montgomery (1930) identified glaucophane and lawsonite which indicate source rocks of blueschist facies. This is important evidence for a derivation from the Owen Stanley metamorphics rather than from metamorphics in Northern Australia.

Within the upper part of the Chiria Formation on Yule Island there is a transition to lighter coloured carbonate turbidites containing coral fragments and shallow water benthic foraminiferids. These contrast greatly with the darker coloured pyritic and glauconitic sediments in the lower part of this unit which were deposited under poorly oxygenated conditions.

However, this is a change in provenance rather than a change in depositional environment, as the upper part of the Chiria Formation was also deposited under bathyal conditions and contains some darker mudstone interbeds similar to those in the underlying sequence. At Delena, the N12-N13 beds in the Chiria Formation consist almost entirely of dark grey mudstone. The neritic carbonate detritus has possibly been derived from the Kido Limestone of the Redscar region.

Talama Formation

Synonymy: Talama volcanics (Pratt and Whittle, 1938) upper Pupunina Formation and Lavao Formation (Millward and Stach, 1941) Lavao Formation and Talama Volcanics (Brown, 1975).

Type Section: Along the Angabunga River about 45 km northeast of Yule Island (Brown, 1975).

Lithology: On Yule Island the Talama Formation is composed of bluish-grey siltstone and mudstone, pinkish-grey to yellowish-brown muddy calcareous sandstone and lithic-crystal tuff with allochthonous basaltic agglomerate. The agglomerate is not persistent along strike and is probably an olistolith. There are interbeds of sandy pinkish-grey to yellowish-orange algal foraminiferal biomicrite and biosparite with some uncemented algal-coralline biomicrudites. Volcanolithic pebble conglomerate is prominent in some parts of the section. Within the type section to the northeast of Yule Island volcanics are more prominent and carbonates are absent (Brown, 1975).

Age: Middle Miocene to earliest Pliocene (N13 to basal N18). The basal horizons in the type section contain a Tf_{1-2} benthic foraminiferal assemblage (Terpstra, 1969). Although we have accepted the correlation made by Brown (1975) between the type Talama and the pyroclastics at Oiapu and Yule Island, it should be noted that this correlation needs to be confirmed by more precise dating of the type section.

Isotopic ages of 7.0 ± 0.5 m.y and $5.2 \pm$ m.y were determined for core samples of basalt from the Talama Formation in the Maiva 1 well (Patterson, 1968a). However the ages were not of stratigraphic order, suggesting either repetition of strata through faulting or radiogenic argon loss from the sample with the younger apparent age. The degree of alteration in these rocks (Armstrong, 1968) makes the latter possibility more likely. The Talama Formation at Delena consists almost entirely of inner neritic sediments which lack age-diagnostic foraminiferids.

Petrography: The tuffs contain 0.1–1.0 mm crystals of plagioclase (An 50–An 60) and clinopyroxene which is often partially replaced by epidote. There are also fragments with vitrophyric or intersertal textures containing microphenocrysts of plagioclase and clinopyroxene. These sometimes occur as lapilli up to 7 mm in diameter. The matrix of the tuffs is yellow-brown glass which is largely altered to chlorite and zeolites. The agglomerates contain clasts of olivine basalt and trachybasalt with intergranular or intersertal textures. Phenocrysts of olivine have been largely replaced by iddingsite or serpentine.

The brownish-green clinopyroxene phenocrysts may be slightly titaniferous, suggesting that the basalts are mildly alkaline or transitional types. The trachybasalts contain modal sanidine, indicating a higher K_2O/Na_2O ratio.

Sandstones and conglomerates within the Talama Formation are predominantly volcanogenic, with crystals and lithic fragments derived from the basaltic pyroclastics, variable amounts of vein quartz and metamorphic rock fragments similar to those in the Chiria Formation are also present. Detrital grains of glaucophane were found in sandstone of the Talama Formation at Delena. Both cleanly washed sandstones and muddy sandstones occur; the cement is usually finely crystalline sparry calcite.

In some cases there are abundant abraded algae, foraminiferid and pelecypod fragments.

Thickness: On Yule Island, the Talama Formation has a thickness of approximately 400 m, but Brown (1975) estimated its thickness in the type section as 1500 m.

Palaeoenvironment: Most of the Talama Formation on Yule Island was deposited in a bathyal environment with water depths exceeding 150 m. However, the Ou Ou Limestone Member was laid down in a shallower neritic environment, as was most of the Talama Formation at Delena (Perembo, pers. comm). The transition to neritic facies occurs within the N14 zone at Delena and within the N16 zone at Yule Island.

In the latter locality the regression was followed by a mid-N17/basal N18 transgression. These non-synchronous bathymetric changes suggest syndimentary tectonism.

Brown (1975) recognised two distinct formations, a clastic/carbonate Lavao Formation and a volcanic/volcaniclastic Talama Volcanics which interdigitated at localities such as Yule Island. We consider that the Late Middle Miocene and Late Miocene sequence is best regarded as a single formation with some distinctive sub-units such as the Ou Ou Limestone Member which are contained within it.

Stratigraphic relationship: The Talama Formation is both conformably and gradationally overlain by the Kairuku limestone at Yule Island. In Maiva 1 and Iokea 1, it is overlain by the Orubadi beds with local (?) disconformities.

The Talama Formation is a lateral equivalent of the upper Aure beds which overlie the Chiria Formation in Tovala 1A.

Ou Ou Limestone Member (Talama Formation)

Synonymy: Ou Ou Limestone Member (of Lavao Formation) (Millward and Stach, 1941).

Type Section: Along the road near Kairuku from GR 4930262 to GR 4814967 (Yule Island geological sheet of Francis *et al.*, 1982).

Lithology: This member is composed of pinkish-grey to pale orange algal-coralline-foraminiferal biomicrudite and biomicrite with subordinate intrasparrudite and biosparite; interbeds of 0.5 m bluish-grey muddy sandstone and mudstone occur in the basal part of the type section. The biomicrudites consist of largely unconsolidated algal and coralline fragments with a little interstitial lime mud. No sessile organisms were found in growth positions but the poorly consolidated nature of this carbonate means that few well exposed sections are available. The biomicrites and biosparites show 10–20% mouldic porosity formed by the dissolution of allochem grains.

Age: Late Miocene (N16 to lower N17)

Thickness: 280 m

Stratigraphic Relationships: The Ou Ou Limestone Member separates two pyroclastic-volcaniclastic sequences within the Talama Formation at Yule Island, Delena and Oro. It has conformable and gradational contacts with both of these units.

Kairuku limestone

Synonymy: Kairuku Limestone Member (of Lavao Formation) (Millward and Stach, 1941); Kairuku Formation (Francis *et al.*, 1982).

Type Section: D'Albertis Point (GR 50312384) (Yule Island geological sheet of Francis *et al.*, 1982).

Lithology: Medium to thick bedded yellow-brown to pinkish-grey biosparrudite, biosparite and biomicrite with interbeds of calcareous mudstone and siltstone. The allochem grains include corals, calcareous algae, foraminiferids, pelecypod, gastropods and echinoids. On Yule Island no sessile forms have been found in situ, suggesting that the unit represents near reef detritus. The biosparites on Yule Island are often sandy with 0.5–3.0 mm vein quartz and metamorphic rock fragments.

Age: Early Pliocene (N18 in part)

The upper parts this unit were deposited in extremely shallow water and lack age-diagnostic planktic foraminiferids. It is probable that the Kairuku limestone ranges above the N18 zone.

Thickness: 250m on the ridge to the northwest of the type section.

Stratigraphic Relationships: The Kairuku limestone overlies the Talama Formation with a conformable and gradational contact.

This transition is associated with a marked regression probably caused by syndimentary tectonism. The Kairuku limestone is much more areally restricted than the Talama and Chiria Formations. It is absent from Maiva 1 and Iokea 1, within which, its time equivalent is the fine bathyal clastics of the Orubadi beds. However a carbonate buildup here named the Kapuri reef, which is possibly equivalent to the Kairuku limestone, was penetrated in Kapuri 1.

It is likely that such carbonate buildups are restricted to latest Miocene structural highs such as the Delena Anticline.

Orubadi beds

Synonymy: Miaru Mudstone (Brown, 1975)

Reference Section: 1497–2426 m in Maiva 1

Lithology: This unit consists of bluish to greenish-grey mudstone with subordinate siltstone and argillaceous volcanoclastic sandstone. The mudstones are often calcareous and carbonaceous. Thin limestone interbeds occur at a few localities.

Age: Early Pliocene (N18 to N18/N20)

Thickness: Generally about 600 m, but difficult to estimate in areas of poor outcrop.

Stratigraphic Relationships: The Orubadi beds overlie the Talama Formation with local unconformities in structural highs. Elsewhere the sequence from the Talama

to the Orubadi is conformable. The Orubadi is in part a lateral equivalent of the Kairuku limestone.

Palaeoenvironment: This unit is a regressive one, which ranges from upper bathyal in the basal part to inner or mid—neritic near the top. Incomplete sequences are present where the unit is draped over structural highs.

Era beds

Synonymy: Apinaipi Formation (Brown, 1975; Francis *et al.*, 1982) Rim Group (Stanley, 1960).

Reference Section: 290–1497 m in Maiva 1

Lithology: The Era beds consist of interbedded mudstone, grey to brown tuffaceous sandstone, polymictic pebble and boulder conglomerate, with occasional interbedded tuff and corallgal limestone (including the Wedge Hill Limestone of Brown, 1975).

Age: Pliocene to (?) Early Pleistocene

Stratigraphic Relationships: The Era beds are conformable on the Orubadi beds.

Palaeoenvironment: This unit was deposited in an inner neritic to terrestrial environment. The sands display well developed cross-bedding, scour-and-fill and slump structures.

The Era beds have proven extremely difficult to date, because the shallow water facies is unsuitable for age-diagnostic foraminiferids, and there is extensive reworking of nannofossils derived from the Talama and Chiria Formations. The age assigned to this unit is thus based on its stratigraphic position.

Mount Davidson Volcanics

Lithology: This unit consists of basaltic and minor andesitic agglomerate, tuff and flows with some volcanolithic conglomerate and sandstone.

Age: (?) Pliocene (no fossil or isotopic ages determined).

Stratigraphic Relationships: Brown (1975) and earlier workers considered that these volcanics were paraconformable over the Talama Volcanics and were in part a lateral equivalent of the sequence referred to in this report as the Era beds. We have re-examined the “facies boundary” mapped by Brown about 20 km northeast of Yule Island and consider that this boundary is a thrust. Further work is necessary to reliably determine the age and stratigraphic relationships of the Mount Davidson Volcanics.

Quaternary coral and alluvium

Extensive deposits of Late Pleistocene to Holocene coral and alluvium unconformably overlie the Era beds and older strata throughout the region.

Some Problems of Regional Stratigraphy

Within the Papuan Basin there has been much unnecessary proliferation of stratigraphic nomenclature. This approach reached the height of absurdity in the earlier work on Yule Island where Millward and Stach (1941) recognised two distinct sets of members for the Lavao Formation, one set in central Yule Island and another in southern Yule Island. Such members would have an extent of less than 2 km along strike and would be terminated by arbitrary facies boundaries. Indeed, within the region which extends 25 km to the southeast of Yule Island, no less than 20 stratigraphic names have been employed for Neogene rocks.

We consider that a more economical approach is necessary and that if workers wish to refer specifically to facies of limited lateral extent, then informal names such as Kairuku limestone and Kapuri reef should be used.

In this paper we have largely overcome the proliferation of stratigraphic names by extending into the Oiapu-Yule Island-Oroi region the use of names such as Orubadi beds and Era beds which were originally employed in regions some distance to the northwest.

Previous biostratigraphic studies of the Neogene units described above have been based almost entirely on benthic foraminiferids. The work by Haig (1984) on the micropalaeontology of the sediments on Yule Island, and by Perembo (1983) on the sediments near Delena, has revealed some serious limitations of the earlier studies. The occurrence of benthic foraminiferids is very much controlled by the sedimentary environment and water depth. The boundary between the Tf₃ stage and the Tg stage in the level of *Lepidocyclina* (*Nephrolepidina*).

On Yule Island the highest occurrence of *Lepidocyclina* (*Nephrolepidina*) is in the N16 to mid-N17 Ou Ou Limestone Member. However at Delena, detailed sampling at 0.5 m intervals by Perembo (1983) has shown that this morphotype is absent from strata younger than N13. In the past, micropalaeontologists such as Stach (1941) have assigned Tg ages to strata purely on the absence of *Lepidocyclina* (*Nephrolepidina*). The evidence presented above shows that such reasoning is suspect. In fact the disappearance of *Lepidocyclina* (*Nephrolepidina*) from the sequence on Yule Island coincides with a change from outer neritic to bathyal conditions. In Fiji this morphotype ranges up into the Early Pliocene (Adams, in press) and thus its extinction level within Papua New Guinea might lie well above the top of the Ou Ou Limestone Member. For these reasons we consider that the Tg stage should not be used unless there is specific evidence that the absence of *Lepidocyclina* (*Nephrolepidina*) is not merely a reflection of the local palaeoenvironment.

STRUCTURE AND TECTONICS

Most authors (Dow, 1977; Brown *et al.*, 1980; Dow *et al.*, 1974) have considered that the rocks constituting mainland Papua New Guinea south of the Ramu-Markham lineament can be subdivided into: the more northerly New Guinea Mobile Belt, and the (Papuan) Platform (Dow, 1977); or the New Guinea Mobile Belt, a medial Papuan Fold Belt and a more southerly Papuan Platform (Brown, *et al.*, 1980). Figure 5

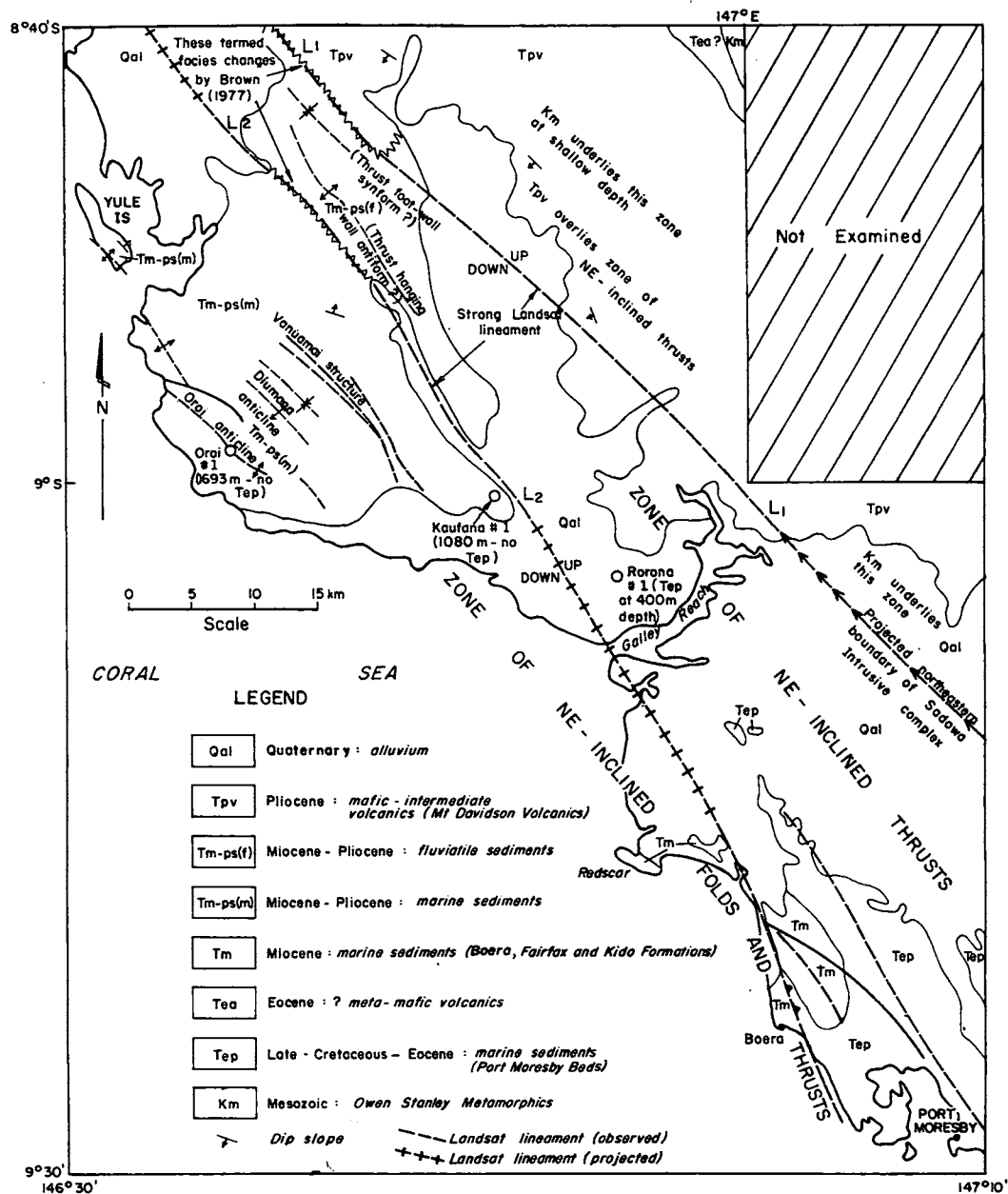


Fig. 5. Macroscale structural interpretation of the south-eastern part of the Papuan Fold Belt. Geology is based on Brown (1977) and Rogerson *et al.*, (1981). Landsat lineaments L1 and L2 were observed on colour composite of scene 103/66.

attempts to recognize difference in tectonic style within the S.E. Papuan Fold Belt (cf. Rogerson *et al.*, 1981). The boundaries between these subdivisions, however, are placed on an irregularly gradational change from thrust-dominated to fold-and-thrust dominated structural styles. The Oiapu-Yule Island-Oroi region lies within the fold-and-thrust zone of the Papuan Fold Belt, not far to the southwest of the boundary with the thrust dominated zone.

Macroscale structure of the Yule Island-Oroi Region

Figure 6 is a simplified interpretation of Landsat scene 103/66 which shows: Owen Stanley Metamorphics; a Neogene sequence divided into a dominantly volcanic part (marked by a well jointed southwest dipping ramp), and folded sedimentary part; and Quaternary alluvium. Spinose folds in the Neogene sequence trend 320° and doubly plunge. Of particular interest however, are two lineaments labelled L1 and L2, the northern most of which (L1) appears to be a positive topographic feature up to 600 m wide. Brown (1977) on the Yule 1:250 000 geological sheet shows Neogene "facies boundaries" corresponding to these lineaments.

Nevertheless in the Kubuna region 20 km northeast of Yule Island, L1 is a northeasterly dipping thrust or reverse fault along which Mount Davidson Volcanics have been thrust over Era beds. It seems likely that L2 is also a fault. L1 when projected southeast appears to be contiguous with the southwest boundary of the Owen Stanley Metamorphics. This is important, as this lineament now separates uncleaved rocks in the south from multi-deformed cleaved Owen Stanley Metamorphics in the north, suggesting north-east block-up motion.

Lineament L2 is an important regional structure, as it appears to separate rocks mainly affected by thrusts from rocks affected by both folds and thrusts. We have projected L2 to the southeast where macroscale folds first become important in the western part of the Port Moresby area (Rogerson *et al.*, 1981). Moreover, a possible 600 m displacement northeast block up motion occurs across this fault as Kaufan 1, with a T.D. of 1080 m, failed to intersect Port Moresby Beds, whereas Rorona 1 northeast of the fault, bottomed in Eocene cherts of the Port Moresby association at a depth of 372 m.

Of further interest are the two folds, one apparently a hanging wall anticline immediately northeast of L2 and the other, a footwall syncline immediately southeast of L1. These two folds suggest that L1 and L2 are northeasterly dipping thrusts.

An interpretation of regional structure from 1:120 000 Skaipiksa air photographs (Figure 7) is more complex and particularly, more faulted, than previous workers have proposed. Most anticlinal structures doubly plunge and are conspicuously associated with faults. Several other noteworthy features are:

- (i) the Oroi Anticline may have been continuous with the Delena Anticline prior to faulting;
- (ii) the Oroi Anticline appears to have been refolded at its southeastern end by a north trending fold, or some more complex structure (combination of faulting and folding) may be responsible;

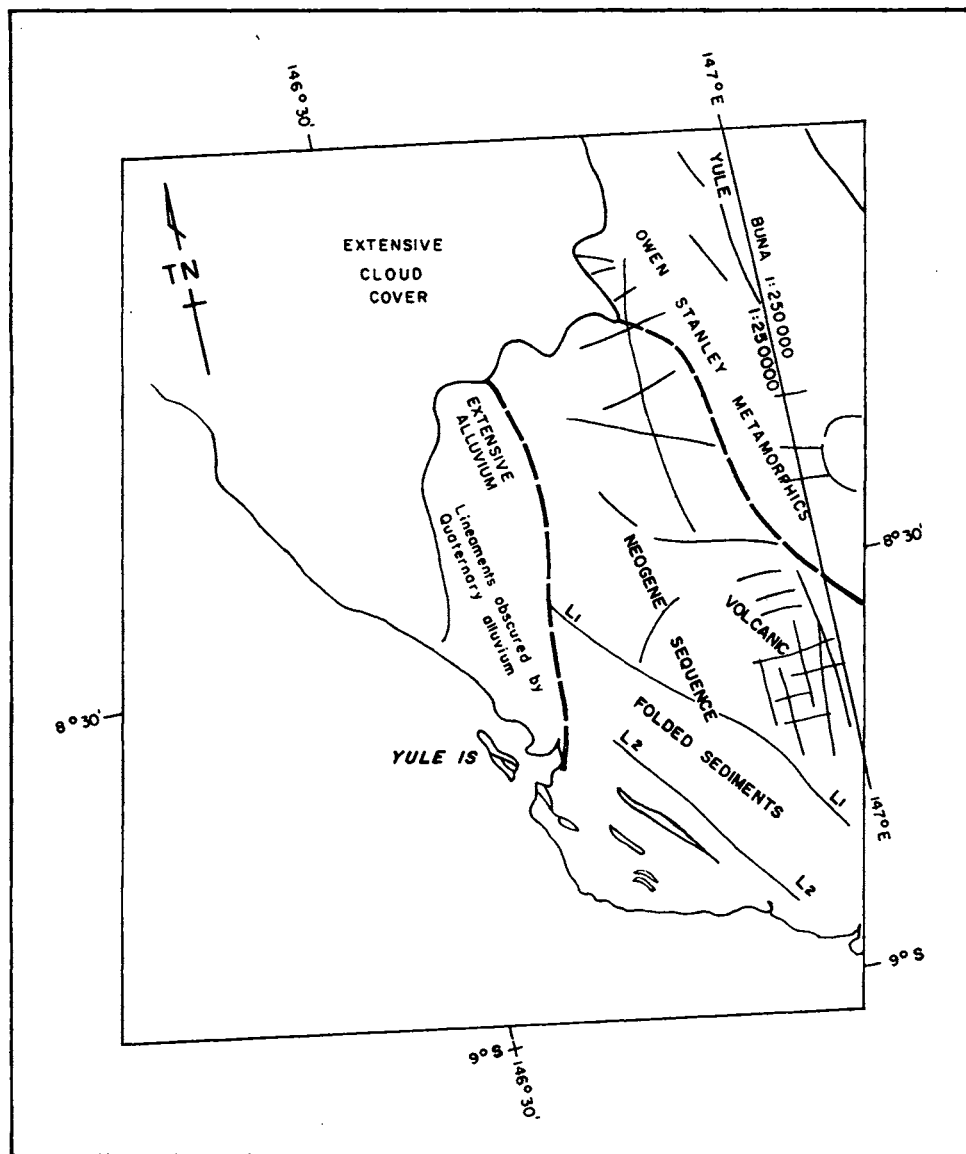


Fig. 6. Landsat lineaments observed on colour composite of scene 103/66.

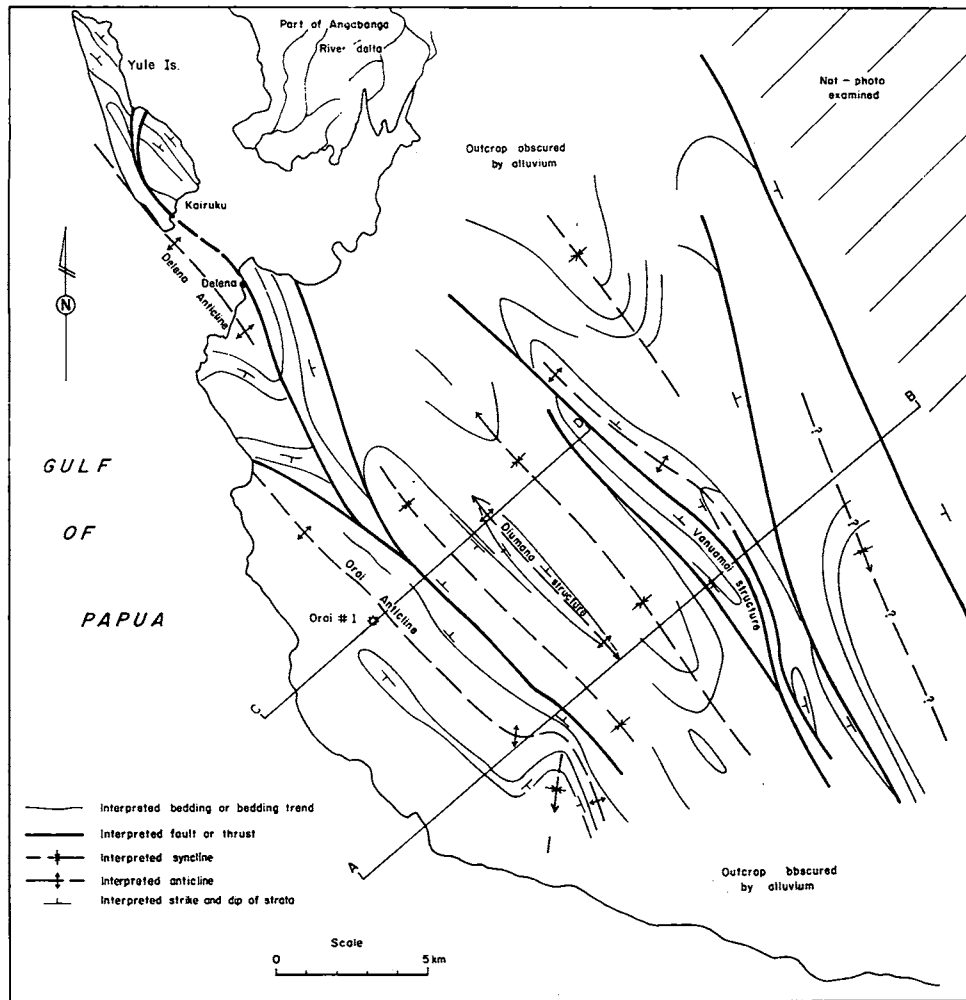


Fig. 7. Macroscale structure of the Yule Island-Oroi area as determined by interpretation of Skaipiksa series 1:120 000 airphotographs. Section AB and CD are shown in Figure 8. The Dumana structure may be more complex than shown. Structure means are taken from Stanley (1960).

(iii) one fault crosses from the mainland to Yule Island. This fault has been identified by R. Perembo (*pers. comm.*) during his Delena Headland mapping, where it is a northeast dipping thrust;

(iv) synclines appear to be relatively open compared to open to tight anticlines.

Although mesoscale structural data will be detailed below, it is pertinent at this stage to summarize some of the evidence for reverse rather than normal faulting.

(i) Faults L1 and L2 have a northeast block-up motion.

- (ii) High angle reverse faults and thrusts have been demonstrated in the Port Moresby area (Rogerson *et al.*, 1981), which is structurally continuous with the present area.
- (iii) Observable mesoscale northwest-trending faults are reverse faults (eg. Delena Headland fault).

The arguments outlined above have been used to construct Figure 8 which compares a section constructed by the present authors with a section constructed by Allen (1957). Except for the little known Diumana structure, the other two major structures in Section AB are northeasterly inclined anticlines with associated thrusts. The structure of the Oroí Anticline is discussed in more detail below, but besides vergence, section AB is structurally more complex than Allen's (1957) section CD, which is an example of the sections used to select exploratory oil well locations prior to 1970.

Rickwood (1968) and Jenkins (1974) have proposed gravity sliding to explain Papuan Fold Belt structural style whereas Davies (1981) has stressed involvement of basement in typical lateral compressive fold and thrust tectonics in the Wabag sheet area. Our work in the present area does not preclude either of these hypotheses but we suggest that fault L2, which displaces Eocene beds at least 600 m northeast block up, is a major structure which could reach and affect basement. Moreover, fault L1, if it is the continuation of the southwestern thrust boundary of the Owen Stanley Metamorphics (Figure 5), affects basement. However, it is possible that many smaller reverse faults do not reach basement.

Mesoscale Structure of the Yule Island-Oroí Region

Figure 9 shows poles to bedding for the study area. Yule Island data were collected by the present authors. Insufficient structural data were plotted to statistically analyse fold shape, but approximate π -planes and hence π -poles can be determined. The principal structures, the Oroí Anticline and Vanuamai structure, have extremely shallow to zero plunges toward the northwest. Yule Island has more variable bedding orientations.

Twenty-six bedding poles measured on Yule Island yield an extremely approximate π -pole of $O^\circ \rightarrow 302^\circ$, and all except two poles suggest the rocks occur on the northeasterly limb of an open anticline, named the Delena Anticline by Stanley (1960). Its hinge line probably passes just west of the island's western shoreline. Figure 9 shows more gentle dips on the eastern side of the island further away from the hinge area and in stratigraphically high Pliocene limestones. Moving westwards across the island, dips become steeper and every large outcrop on wave cut platforms, and small back-platform cliffs shows mesoscale deformation features; shear zones, small displacement faults and strike faults with unknown displacement. Dips and dip directions, particularly in the older Chiria Formation, are variable and at 490237, overturned beds occur, probably adjacent to a major thrust or high angle reverse fault. Mesoscale shallow plunging folds occur at 478243 adjacent to the Delena Anticline hinge. Section AB (Figure 2) is a vertical cross-section to only shallow depth, which shows our interpretation of the structure.

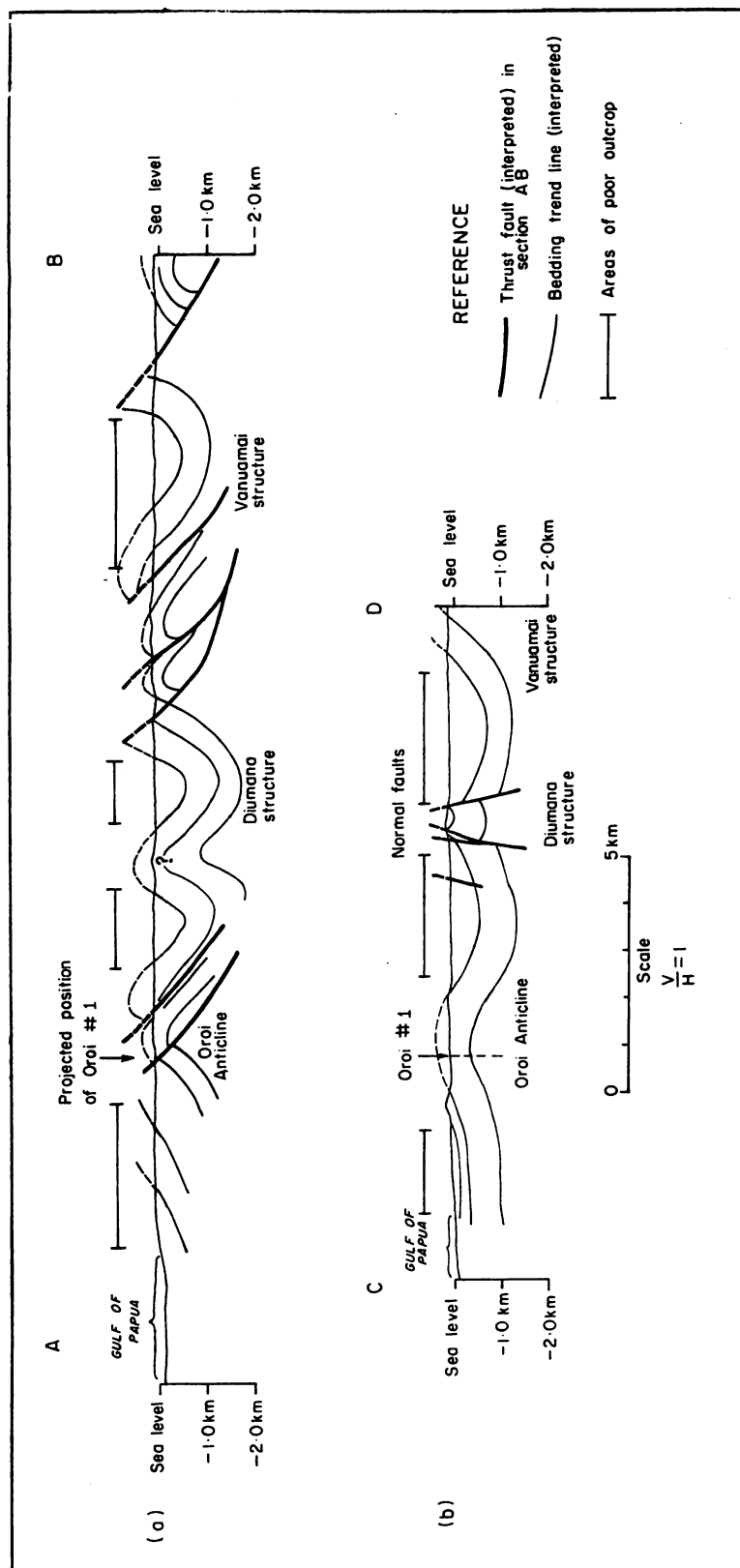


Fig. 8. Vertical cross-sections along AB from Figure 7 using bedding trend lines. (a) Approximate tectonic profile (looking N) showing favoured interpretation. (b) Vertical cross-section (looking N) showing structure as interpreted by Allen (1957).

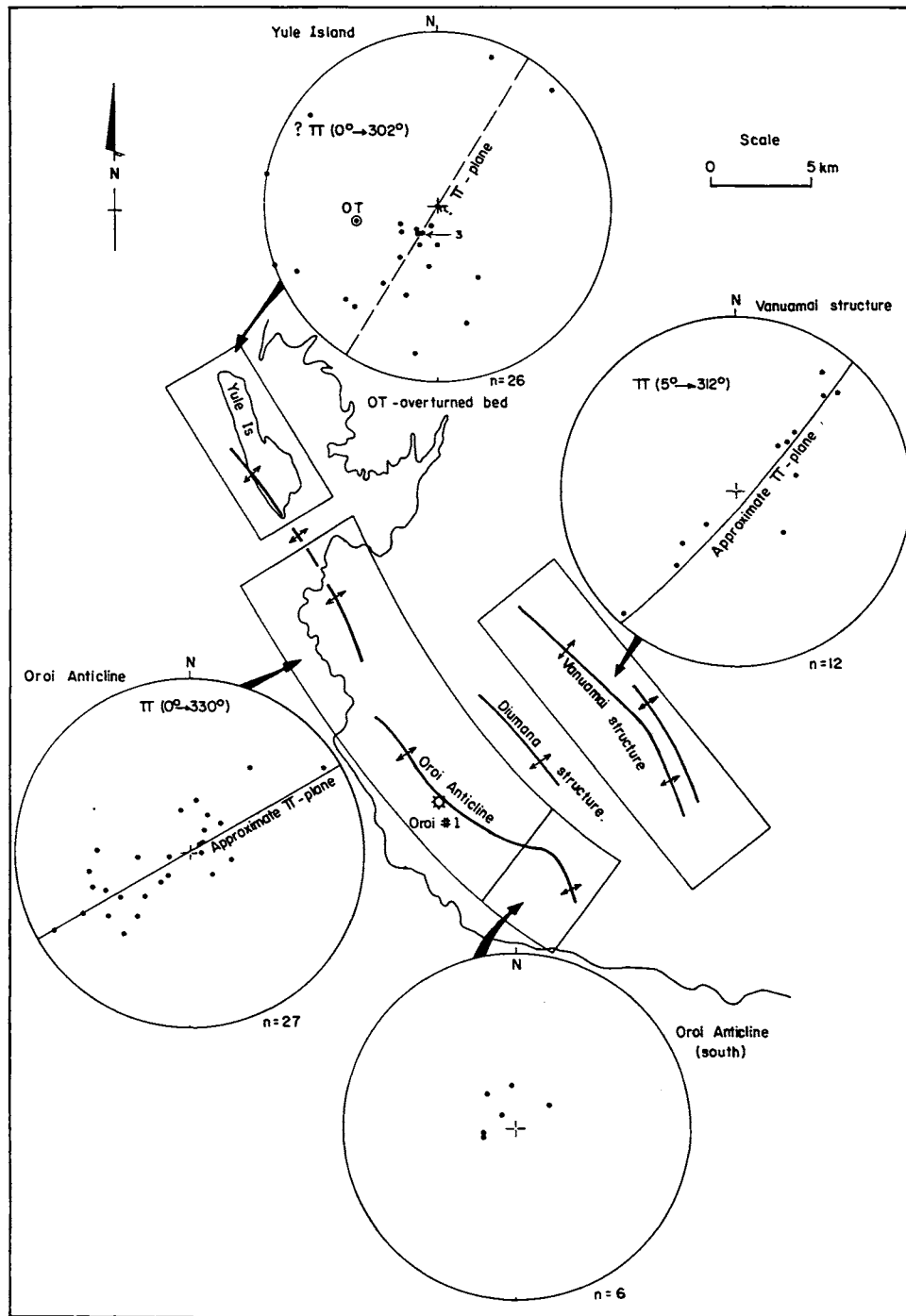


Fig. 9. Equal-area stereographic projections of bedding poles. Insufficient poles plotted for statistical analysis. Poles on the mainland have been determined by measuring bedding azimuths on Allen's (1957) map.

Mesoscale faulting is particularly noticeable in most large outcrops. Small displacement faults oblique to bedding with apparent right lateral motion (and unknown dip slip) trend 050° – 080° (10 measurements) whereas left lateral faults trend 110 – 130° (5 measurements). However, many strike faults striking northwest with unknown displacement also occur. Fault stepping from one bedding plane to another occurs frequently and beds affected often vary in dip markedly adjacent to the fault; slight overturning was noted near 496233. We believe these are reverse faults, similar in orientation to larger reverse faults striking northwest.

At island scale we suggest these structures are often present at formation boundaries and that the Ou Ou Limestone Member is partly bounded by one such fault. It is highly probable that the reverse fault discovered by Perembo (pers. comm.) on Delena Headland is represented on Yule Island. The 300° trending fault at the Yule Mission is an important structure which apparently causes the omission, except in fault slivers, of N14–15 strata on Yule Island.

Structure of the Oroí Anticline

Measurements of bedding orientations by Allen (1957) on this structure are plotted in Figure 9. Subsurface data on the Oroí Anticline are provided by the dipmeter results of Oroí 1 together with recent age determinations for core samples from the well by Haig (1984). The well drilled on what had been regarded as a relatively simple anticline (Allen, 1957), but penetrated a steeply dipping and faulted sequence of Chiria Formation. Both the micropaleontological data and observations on graded bedding by Thompson (in A.O.C., 1949) indicate that parts of the sequence are overturned (Figure 10).

The main structural features present are:

- (i) a fault at 470 m with N12 strata on the hanging wall and N9 strata on the footwall. This would be either a normal fault or a thrust running subparallel to bedding which steps down through the sequence.
- (ii) a zone isoclinal folding from 1100–1225 m with several intervals of overturned facing.
- (iii) a reverse fault or thrust at 1250 m with N10 beds on the hanging wall and probable N11 beds on the footwall. The steep northeasterly dip of strata in the hanging wall and steep southwesterly dip in the footwall suggest that the fault dips to the northeast.

This well illustrates the extremely complex mesoscale geology which is characteristic of the region. The macroscale structure is less complex, being characterised by thrusts and asymmetrical overthrust anticlines. Although the dominant dip of thrusts and fold hinge surfaces is to the northeast, some structures dip to the southwest.

The complex mesoscale maps produced by Mathews and Sturmfels (1957) and Power and Morgan (1957) on the Vanuamai structure and the Diumana Anticline, are

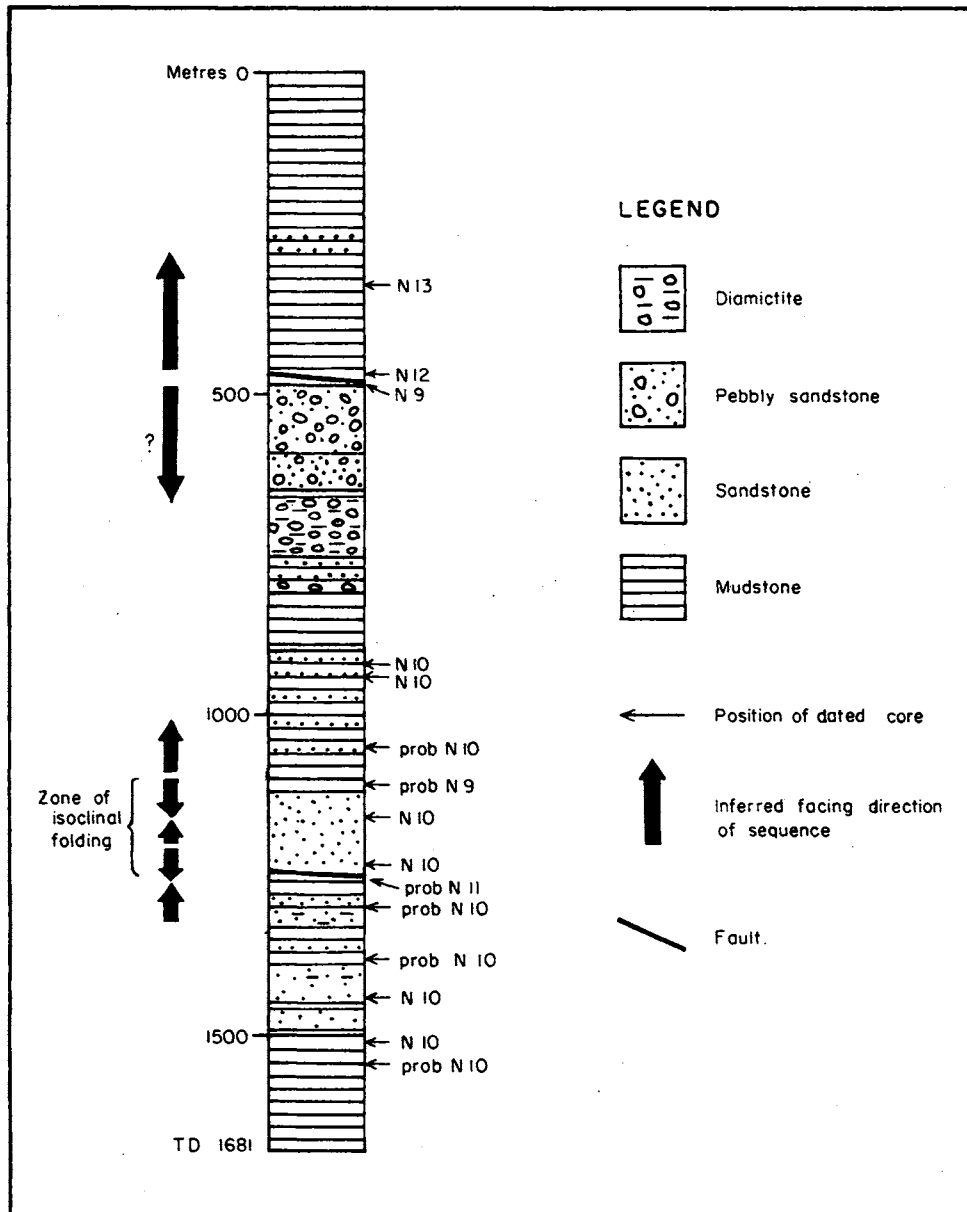


Fig. 10. Stratigraphic/structure diagram for Oroi 1 well based on micropalaeontological results in Haig (1984).

attempts to portray complex geology in an area of poor outcrop without sufficient palaeontological control. The criticisms made by Stanley (1960) of the maps produced by Mathews and Sturmfels (1957) and Power and Morgan (1957) were perhaps unjustified. The obvious conclusion which should have been reached by Stanley in his 1960 review was that these workers showed such areas to have complex structure, probably produced by pervasive faulting. More simplistic maps such as that of Millward and Stach (1941) do not portray that complexity.

Timing of Deformation

The tectonic evolution of the Papuan Peninsula has been reviewed in detail by Rogerson *et al.*, (1981). Two main models have been proposed:

- (i) Rifting of southeastern Papua from northern Australia and subsequent choking of a northerly dipping subduction zone in Early Eocene times, causing emplacement of Cretaceous oceanic crust (the Papuan Ultramafic Belt) in northeastern Papua (Davies and Smith 1971; Davies, 1978).
- (ii) Development of a Late Tertiary subduction zone complex on the southwestern side of a south-migrating arc (Hamilton, 1979).

Data from the Port Moresby region cannot easily be reconciled with either model (Rogerson *et al.*, 1981). Rogerson and his co-workers found that rocks as young as Early Miocene were affected by the thrusting, contrary to the suggestion by Davies (1978) that the thrusting occurred in Eocene times and that subsequent tectonic activity involved predominantly vertical movements. However, the absence of Middle Miocene to Early Pliocene strata from the Port Moresby region meant that only a general post-Early Miocene age could be determined for the thrusting.

In the Yule Island, Delena and Oiapu regions, the thrusting affects Pliocene strata of the Kairuku limestone, Orubadi beds and Era beds. However offshore seismic evidence indicates that a number of thrusts only affect Miocene strata of the Chiria Formation, Talama Formation and Aure beds, and are truncated by the Pliocene Orubadi beds. This suggests two main phases of compressive deformation, one in the latest Miocene and another in the Late Pliocene/Early Pleistocene.

The scarcity of reliable structural and stratigraphic data on the central cordillera of the Papuan Peninsula makes it difficult to correlate tectonic events in the Yule Island-Oroi region with events in regions to the northeast such as Tapini. Davies and Smith (1971) consider that near Tapini, unmetamorphosed sediments of Middle Eocene to Miocene age unconformably overlie the Owen Stanley Metamorphics indicating an Early Eocene age for the metamorphism and thrusting. Nevertheless, Macnab (1969) maintains that in this region sediments of Late Oligocene to Early Miocene (Te) age form thrust slices interstacked with Owen Stanley Metamorphics. If the latter interpretation is correct then the thrusting continued at least until the Miocene.

At Yule Island, glaucophane and lawsonite occur as detrital grains in Middle Miocene (N10) sediments, showing that by this time high pressure facies of the Owen

Stanley Metamorphics had been uplifted and exposed at the surface. It is possible that strong compressive deformation commenced in the Tapini region in Early Miocene times and migrated southwest to Yule Island by mid-Pliocene times. Further work on areas between Yule Island and Tapini is necessary to firmly establish time-space relationships for deformation.

PETROLEUM POTENTIAL

Source Rock

Little work has been done on source rock geochemistry in the Oiapu-Yule Island-Oroi region. Reiman and Dielwart (1975) analysed core samples from the Orubadi beds and Talama Formation in Maiva 1, and Robertson Research have recently completed a general geochemical study of the Papuan Basin, including the Maiva 1 and Tovala 1A wells (H. Alimi, pers. comm.).

The results suggest that the Chiria Formation, Aure beds and Orubadi beds in Tovala 1A are poor to fair gas sources, as are the Orubadi beds in Maiva 1. However Reiman and Dielwart (1975) found that carbonaceous mudstones of the Talama Formation in cores at 2871 m and 2950 m in Maiva 1 were fair to good oil sources but immature. Downward extrapolation of the maturation trend in this well suggests that the top of the oil generation zone lies at a depth of 3500 m.

No source rock studies have been carried out on Palaeogene rocks of the Port Moresby association. However good oil sources are known to be present in the Mesozoic Wahgi Group, particularly in the Jurassic Maril Shale (Alimi, pers. comm.). It is likely that this group underlies at least part of the Oiapu-Yule Island-Oroi region.

Reservoir Rocks and Seals

Porous and permeable rocks capable of forming petroleum reservoirs occur in the Kairuku limestone and Talama Formation. At Delena, the Ou Ou Limestone contains biolithites of finger corals with up to 40% porosity. The Pliocene pinnacle reef intersected in Kapuri 1 is composed of limestone possibly equivalent to the Kairuku Formation. Drill stem tests produced flows of 110–200 barrels of formation water per hour from this reef (Patterson, 1968a). The Kairuku limestone on Yule Island contains porous bioclastic limestones, as does the Ou Ou Limestone Member in the underlying Talama Formation. These limestones are often weakly cemented with significant porosity. An allochthonous biosparite from the Chiria Formation in Oroi 1 had a porosity of 28% (Stanley, 1960) and samples from the Chiria and Talama Formations examined in the present study had porosities of approximately 10–25%.

Much of this is mouldic porosity produced by the dissolution of allochem grains rather than primary intergranular porosity.

Many of the pyroclastics and some of the volcanoclastic sandstones in the Talama Formation possess excellent porosity and permeability, although Talama Formation volcanics have been regarded as economic basement by petroleum exploration companies in the past. A drill stem test of these rocks in the 1384 m–1475 m interval

within Iokea 1 produced 348 barrels of formation water per hour and a tuff cored in this interval had a porosity of 28.4% (Armstrong, 1968). However, muddy limestones and mudstones which occur as interbeds in the Talama Formation and the basal part of the Kairuku limestone probably form tight zones. An adequate seal over potential reservoirs in the Talama Formation and Kairuku Limestone would be provided by fine grained clastics of the Orubadi and Era beds.

CONCLUSIONS

The Oiapu-Yule Island-Oroi region has been affected by synsedimentary compressive deformation in late Neogene times. This has formed a belt of fold-and-thrust structures, with the dominant dip of thrusts and fold hinge surfaces being to the northeast. The structures provide potential traps for hydrocarbon accumulations, but exploration in the past has been greatly hampered by lack of structural understanding and lack of good quality seismic data.

Good potential reservoir rocks are known to exist in the Talama Formation and in localised Pliocene carbonate buildups such as the Kairuku limestone and the Kapuri reef.

Although the limited data available suggest that Neogene source rocks are generally gas-prone, there are some potential source rocks for oil in the Talama Formation, and oil sources may be present in the underlying Palaeogene and Mesozoic sequences.

Consequently further exploration is required to accurately assess the petroleum potential of the Oiapu-Yule Island-Oroi region.

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APPENDIX

MICROFOSSIL DETERMINATIONS, YULE ISLAND

Sample	Locality	Age	Bathymetry
1		N10	upper middle bathyal
2		N10	upper middle bathyal
3		N10	upper middle bathyal
4		N10	upper middle bathyal
5		N10	upper middle bathyal
6		N12	upper middle bathyal
7		N14	upper bathyal
8		N12	upper bathyal
9		N16/lower N17	upper bathyal
10		N16/lower N17	upper bathyal
11		N16/lower N17	upper bathyal
12		—	inner neritic
13		no older than N16	outer neritic
14		N17	outer neritic
15		N13	upper bathyal
16		N17	upper bathyal
17		N17	upper bathyal
18		N17	upper middle bathyal
19		—	inner neritic
20		N18	upper bathyal
21		N18	inner outer neritic
22		—	inner neritic
23		—	inner neritic
24		—	inner neritic
25		N13/N14	outer neritic

Sample localities are shown in Figure 2.