The island of New Guinea is the mountainous margin of the Australian continent. Paleozoic and Proterozoic Australian craton extends northward beneath the shallow waters of the Arafura Sea to underlie the southern plains of New Guinea and, with overlying sediments, to form the dramatically sculpted southern slopes of the central range in a great fold and thrust belt. The fold and thrust belt marks the outer limit of the autochthon. Beyond, to the N, E and W, is an aggregation of terranes that have accreted since the Late Cretaceous, driven by oblique convergence between the Pacific and Indo-Australian plates. The terranes comprise continental fragments and blocks of oceanic volcanic arc and of oceanic crust and mantle origin, and include two great ophiolites. The plate boundary itself is a complex system of microplates, each with separate motion, and marked by every kind of plate boundary. In the E the opening of the Manus Basin is associated with rapid clockwise rotation of New Britain, and the opening of the Woodlark Basin causes extension of continental crust in the Papuan peninsula and islands. This has resulted in the development of low-angle extensional faults and domal structures in metamorphic rocks and the exhumation of Pliocene eclogite. Remarkably similar extensional structures and the exhumation of Pliocene eclogite are seen in the Bird’s Head area of western New Guinea (Wandamen Peninsula). Flat and shallow oblique subduction at the New Guinea Trench has caused the deformation of Plio-Quaternary sediments in the Mamberamo Basin, deformation and Pliocene igneous activity in the central range, and the southwestward motion of the Bird’s Head. The island has significant resources of economic minerals and hydrocarbons.

Introduction

The island of New Guinea is the alpine and, in part, Andean margin of the Australian continent. In plan view, the island resembles a great bird flying westward (Figure 1). It is the second largest island in the world (2,200 km long and up to 750 km wide) and one of the most mountainous, with peaks to c. 4.9 km above sea level (Figure 2).

Politically, the island is divided between the Independent State of Papua New Guinea (PNG) in the E and the Republic of Indonesia in the W, with a boundary that coincides, for the most part, with the 141°E meridian. The western half was known as Irian Jaya and is now known as Papua and Western Irian Jaya; Western Irian Jaya is the Bird’s Head and Neck.

The geology of New Guinea can be considered in three parts:
1. a western part that includes the Bird’s Head and Neck and adjacent islands;
2. a broad central part that adjoins the Australian continent (Figure 1); and
3. an eastern part that includes the Papuan peninsula and islands.

All three parts have a similar geometry with sedimentary basins on continental basement in the S and a hinterland of metamorphic and oceanic rocks including ophiolite and volcanic arc assemblages in the N.

Islands that lie to the NE and E, the Bismarck Archipelago and Solomon Islands, do not fit this pattern. They are thought to have formed solely by intra-oceanic volcanic arc activity and, in the case of the Solomon Islands, accretion of a mostly-submarine volcanic plateau (see below).

Geological maps of the New Guinea and adjacent islands at 1:1 million scale (Bain et al., 1972; Dow et al., 1986) and of PNG at 1:2.5 million scale (D’Addario et al., 1976) are available, as are map series maps at larger scales. A useful bibliographic data base is Van Gorsel (2011).

Geology of the Western Part: the Bird’s Head peninsula and islands

Pieters et al. (1983) discussed the geology of the Bird’s Head in terms of an oceanic province in the N, a continental province in the S, and a transitional zone between the two. The continental province occupies the greater part of the Bird’s Head and includes Misool Island (Figure 3). It is bounded on the N by the E-W Sorong Fault (SF in Figure 3) and on the E by a N-S fault that parallels the W coast of Cendrawasih Bay and connects to the Weyland Overthrust (WT in Figure 3).

The Continental Province

The Continental Province comprises sedimentary rocks over a Paleozoic basement. The basement is exposed in the mountains S of
the Sorong Fault and on Misool Island and comprises folded low-grade regionally metamorphosed turbidites that have been dated by Silurian graptolites and Devonian ostracods (Pieters et al., 1983). On the Bird’s Head, the sedimentary sequence is of Permian and Mesozoic platform sediments, Eocene to Mid-Miocene limestone, and Late Miocene to Pliocene and Quaternary siliciclastics that are part turbiditic and part non-marine (Pieters et al., 1983; Bailly et al., 2009). The Cenozoic sediments form the Salawati and Bintuni basins (SB and BB in Figure 3) that are separated by a N-S basement ridge.

The sedimentary sequence on Misool Island differs from the mainland. Here “a unique and almost continuous sequence of deep-water and shallow marine sediments extends from Triassic times to the present day” (Pieters et al., 1983). The older sediments were folded in the Late Triassic and Early Jurassic (Visser and Hermes, 1962).

In the SE part of the Bird’s Head the entire sedimentary section has been deformed and locally metamorphosed by contractional tectonics in the W-facing Lengguru fold belt (LFB in Figure 3). Bailly et al. (2009) interpreted the deformation to be the result of Late Miocene E-dipping subduction on the line of the present shoreline of Cendrawasih Bay. Miocene contraction was followed by Pliocene extension, the development of normal faults, and the unroofing of the Wandamen metamorphic core complex at 4–2 Ma (Bailly et al., 2009).

In the SW part of the Bird’s Head the Cenozoic carbonates have been arched upward to form the karstified limestone antiforms of the Onin and Kumawa peninsulas (Ratman, 1998). The antiforms trend NW towards Misool Island.

Oceanic Province

The Oceanic Province, N of the Sorong Fault includes Paleogene volcanic arc rocks and younger sediments, Triassic granitoids and, in the adjacent islands, ophiolite (Pieters et al., 1983).

The Transition Zone

The Transition Zone rocks E of the N-S fault include fault slices of Paleozoic (?) metamorphosed sediments intruded by Early Jurassic (197 Ma) granite and, on the Wandamen Peninsula, the Plio-Pleistocene metamorphic core complex with grades as high as eclogite (Bailly et al., 2009). Further to the SE, on the E-W part of the Bird’s Neck, Transition Zone rocks above the S-facing Weyland Overthrust include highgrade metamorphic rocks, ophiolite slices and Miocene diorite. The metamorphic rocks are pelitic and include staurolite-garnet-mica schist. Beneath and S of the thrust fault are footwall Paleozoic to Cenozoic sediments on continental basement; this is a westward arm of the Papuan Basin (Pigram and Panggabean, 1989).

Geology of the Central Part of New Guinea: 136–145°E

The central part of the island is made up of the Papuan Basin in
the S and a hinterland of mostly crystalline rocks in the N; the hinterland rocks are extensively overlain by Neogene sediments.

**Papuan Basin**

The Papuan Basin occupies all of autochthonous New Guinea – the southern part of the bird’s body (Figure 3). Sediments of the Papuan Basin underlie the southern plains and are exposed in the adjacent fold belt.

The basin is underlain by Australian craton of Precambrian age in the W and of Paleozoic age in the E; the boundary between Precambrian and Paleozoic basement is at around 141°E, with the exception that there is an exposure of Paleozoic basement at 140.3°E (Eilanden Metamorphics; Parris, 1996b). In the W the sedimentary section is 16 km thick and has late Proterozoic strata at base (Table 1). In the E the sedimentary section is 4 km thick and has Triassic and Jurassic sediments at base (Table 2). Hill et al. (2004) described the basin and its hydrocarbon potential.

**Western Papuan Basin**

The sedimentary sequence is known from exploration wells on the foreland platform (Kendrick and Hill, 2001) and from mapping of the fold and thrust belt, where the rocks are exposed in the eroded core of a frontal anticline (Mapenduma Anticline; Parris, 1994a). An almost complete Paleozoic and Mesozoic section is exposed along the Freeport Grasberg mine access road (Martodjojo et al., 1975; Parris, 1994b; Cloos et al., 2005) and the Cenozoic section is exposed near the mine (Quarles van Ufford and Cloos, 2005).

The rock units of Cambrian and older age, notably the Kariem, Nerewip and Awitagoh formations, are known only from isolated exposures and there is doubt about their inter-relationships (Table 1). However the younger rock units, beginning with Otonoma Formation, appear to be part of a paraconformable sequence that extends from Late Proterozoic or Cambrian to the Mid or Late Cenozoic.

The older sediments were deposited in a shelf environment. Break-up began in the Permian and continued in the Triassic and Early and Middle Jurassic (Pigram and Panggabean, 1989), and is recorded in the sediments of the Tipuma Formation (Parris, 1994a). Break-up was followed in Middle Jurassic and Cretaceous by the deposition of sag phase sediments of the Kembelangen Group (Table 1).

Carbonate sedimentation began in the Maastrichtian and Paleocene and persisted until mid-Miocene. An interval of clastic sedimentation (Sirga Formation) in the early Oligocene is correlated with a fall in sea level at the time of the first of the Cenozoic glacial maxima (Cloos et al., 2005). The later transition from carbonate to mixed pelitic and carbonate sedimentation at the beginning of the Late Miocene can be correlated with the fall in sea level at the time of the Late Miocene glacial maximum, though uplift associated with the first stages of mountain-building probably was a contributing factor. The emergence of the mountain mass in Late Miocene, Pliocene and Quaternary led to the rapid deposition of mostly molasse-type clastic sediments to S and N, notably in the Mamberamo Basin where total thickness may exceed 10 km (Visser and Hermes, 1962).

**Eastern Papuan Basin**

The Mesozoic–Cenozoic basin evolution and sedimentary sequence in the eastern Papuan Basin (Home et al., 1990) is similar to that in the W, though not identical (Table 2). Break up in Late Triassic and Early–Mid-Jurassic was followed by sag phase siliciclastic sedimentation through Late Jurassic and Cretaceous, carbonate sedimentation from Eocene–mid-Miocene, and development of the fold belt accompanied by molasse-type sedimentation and some volcanism in late Miocene, Pliocene and Quaternary.

A feature that is seen only in the eastern Papuan basin is the rift-related uplift at the end of the Cretaceous and resultant erosion of Cretaceous section. The rifting and uplift were precursors of the Paleocene opening of the Coral Sea basin. Another feature seen only in the E is the develop-ment of Pleistocene strato- and shield volcanoes.

The strike of the fold belt changes at the international border perhaps coincident with the transition from Paleozoic basement in the E to Precambrian basement in the W. Structural style changes at 142°E from a broad asymmetric S-facing basement-thrust-bounded anticline upon which are superimposed lesser structures in the W to a thin-skinned thrust belt of parallel thrust-bounded anticlines and valley-and-ridge topography in the E (Figure 4).

Oil and gas in the eastern basin are sourced from Jurassic Imburu Formation and have accumulated in uppermost Jurassic and lowermost Cretaceous (Neocomian) sands that developed in the mudstone environment during sea-level lowstands; the fluids migrated into structural traps in the Pliocene (Hill et al., 2004).

**Jimi-Kubor and Bena Bena blocks**

The Triassic to Cretaceous sedimentary rocks that are exposed in the Kabor Range and in the Jimi Valley, N of the Kabor Range, share some features in common with the Papuan Basin but there is much that is distinctive (Table 3). The distinctive character suggests that the Jimi-Kubor block is a terrane – probably a para-autochthonous terrane that broke from the Paleozoic Australian craton and was re-joined by collision in the late Paleocene or early Eocene (Davies et al. 1996; 1997). The basement that is exposed in the Kabor Range comprises metamorphosed Permian sediments intruded by Middle Triassic granitoids (Van Wyck and Williams, 2002).

Granitoids that intrude the Jimi-Kubor rocks are known as Kubor Granodiorite but form two populations, as indicated by K-Ar age, one c. 240 Ma and the other c. 220 Ma. Crowhurst et al. (2004) determined that the older suite is volcanic-arc-related and the younger is rift-related, as shown by Sr isotope and Sm-Nd data.

The Bena Bena terrane is a mountainous area of greenschist facies metamorphic rocks (Goroka and Bena Bena Metamorphics; Tingey and Grainger, 1976) in the eastern part of the area mapped as Jimi-Kubor in Figure 3. The protolith of the Bena Bena Metamorphics is part Late Triassic (221 Ma; Van Wyck and Williams, 2002) and the metamorphics are intruded by Jurassic gneissic granite (172 Ma; Page, 1976).

**The Hinterland of the Central Part of the Island**

The hinterland of the Papuan Basin extends from Cendrawasih Bay in the W (CB in Figure 3) to the Finisterre Range in the E (FR in Figure 3). The hinterland is entirely allochthonous, or may be para-autochthonous in part, and is made up of terranes that have accreted to the Australian craton in a succession of collisions beginning in the Late Cretaceous (Pigram and Davies, 1987; Davies et al., 1996).
Figure 3 Geological map of New Guinea. AB Aru Basin; AFB Auro fold belt; B Bougainville; BB Bintuni Basin; BK Biak; BT Bismarck Sea Transform; C Cyclops Mountains; CB Cendrawasih Bay; FR Finisterre-Saruwaged Range; G Gautier or Foja mountains; GR Grasberg Mine; KT Kilinailau Trench; L Lihir Island (mine); LFB Lengguru Fold Belt; M Manus; MB Manus Basin; MI Misool; MT Manus Trench; MU Manus; NB New Britain; OT Ok Tedi; P Porgera; PT Pocklington Trough; R Rabaul; SB Salawati Basin; SF Sorong Fault; ST Seram Trench; TT Tanim Trough; WC Weyland Thrust; W Wau; WA Waipora Basin; WB Woodlark Basin; WM Wamena; WN Wamakita Peninsula; WO Waigeo; WT Weyland Thrust; Y Yipen. (Map drawn by Randall Betuela).
Table 1 Stratigraphy of the Indonesian part of the Papuan Basin.

| Age                                      | Description and thickness                                                                 |
|------------------------------------------|------------------------------------------------------------------------------------------|
| Late Miocene–Quaternary                 | Molasse-type sediments derived from erosion of the emerging mountain mass. Mamberamo Group on N side of Central Range includes mid–late Miocene Makats Formation (1.8 km) and late Miocene to Pleistocene Mamberamo Formation (10 km; Visser and Hermes, 1962; as reported by Parris and Warren, 1996). |
| Late Miocene–Pliocene                   | Buru Formation: basal calcareous mudstone and sandy shale grades upward to well-bedded lithic sandstone and mudstone with a 200 m interbed of limestone; parts mapped as Iwoer (Iwur) Formation, Kau Limestone, and Birim Formation; conformable on New Guinea Limestone Group. Thickness up to 3 km. |
| Paleocene–Middle Miocene, locally lower limit is Maastrichtian | New Guinea Limestone Group. Gradational contacts with Ekmai Sandstone below and Buru Fm above; 1.6 km. Insksin Formation is deepwater equivalent (Pieters et al., 1983) |
|                                          |                                                                                          |
|                                          | **Cloos et al., 2005; Parris, 1994b**                                                      |
|                                          | **Pieters et al., 1983**                                                                  |
| Oligocene–Mid Mio                        | Kais or Ainod Fm, massive limestone, locally abund. fusulinids, forams, 1.1 km            |
| Early Oligocene (Te stage)               | Sirga Sandstone, quartz-rich; 100 m. Erosional disconformity at base.                   |
| Eocene (Ta–Tb stage)                     | Faumai Formation, well bedded arenaceous limestone, commonly muddy, 250 m (Pieters et al., 1983). Lower part of Yawee Limestone is equivalent of Faumai Formation. |
| Paleocene–Eocene; part Maastrichtian (Parris, 1996b) | Waripi Fm, well-bedded sandy oolitic calcarenite, 700 m, most has no fossils, Late Cretaceous age from forams (Parris 1996b). |
| Middle Jurassic–Late Cretaceous          | Kembelangen Group: grey variably argillaceous, glauconitic, calcareous, micaceous and pyritic sandstone and siltstone, black calcareous mudstone to limestone, quartz sandstone and orthoquartzite; 4.5 km. Kembelangen component rock formations are (Cloos et al., 2005; Parris, 1994b): |
|                                          | Late Cretaceous                                                                          |
|                                          | Ekmai Sandstone: pyritic and glauconitic quartz sandstone; 650 m.                         |
|                                          | Pinuya Mudstone: micaceous and glauconitic mudstone rare foraminifera; 700 m.              |
|                                          | Early–Late Cretaceous                                                                    |
|                                          | Woniwogi Sandstone: orthoquartzite with belemnites; 200–400 m.                           |
|                                          | Middle–Late Jurassic                                                                     |
|                                          | Kopai Formation: quartz sandstone, siltstone, mudstone, belemnites, gastropods, pelecypods, ammonites, limestone with star crinoids; shallow marine; 300 m. |
| Late Permian and Middle Jurassic; (alternatively Triassic) | Tipuma Formation. Maroon and green mudstone, lithic sandstone and pebble conglomerate, part non-marine. Comprises two rock units separated by disconformity, one late Permian and the other Middle Jurassic; ages by palynology (Parris, 1994a). Fossils in lower part include Glossopteris; thickness 2 km (Cloos et al., 2005). The disconformity may represent a rifting event (Parris, 1994a). Mapenduma Fm may be equivalent to Tipuma Fm; it is the field name given to thick sequence of grey turbidites in Wamena area, part Triassic and part probable Middle Jurassic from palynology (Parris, 1994a). |
|                                          | Devonian Formation                                                                        |
|                                          | Aiduna Formation. Lithic sandstone part feldspathic, part micaceous, interbedded with black shale, biocalcarenite, polymict conglomerate and coal; overlies Modio Fm and underlies Tipuma Fm; Permian age from brachiopods and plant fossils (Parris, 1994a); paraconformable on Modio Fm and grades upwards into Tipuma Fm (Pieters et al., 1983); up to 2.2 km (Parris, 1994b; Cloos et al., 2005). Part of Aifam Group in Bird’s Head basins (Pieters et al., 1983) |
| Devonian and possibly Silurian            | Modio Formation: fossiliferous dolostone and crinoidal grainstone with siliciclastic sediments towards the top; Devonian age (part Frasnian, part pre-Frasnian and possibly Silurian), age from corals (Oliver et al., 1995) and poorly preserved conodonts; 2 km approx. (Pieters et al., 1983; Parris, 1994b; Cloos et al., 2005). |
| Ordovician                                | Tuabu Formation: (136.6–138°E) coarse quartz sandstone with some conglomerate, interlaminated sandstone and siltstone, and at top red laminated siltstone and mudstone; overlies Otonoma and underlies Modio Fm; 1 km thick (Parris, 1994b; Pieters et al., 1983). Kora Fm (138.4–139.3°E) black slaty mudrocks, may be equivalent of Tuabu Fm, Ordovician age from nautiloids and graptolites (Parris, 1994a, 1996a). |
| Late Proterozoic–Cambrian                 | Otonoma Formation upward-coarsening turbidites exposed at 136.6–137.4°E, basal part weakly metamorphosed, one fission track zircon age is 675 Ma; thickness >3 km (Parris, 1994b). This suggests that the Otonoma Fm may be older than Nerewip Fm and therefore could be equivalent to Kariem Fm (Parris, 1994a). |
|                                          | Late Proterozoic–Cambrian                                                                |
|                                          | Awitagoh (140°E) and Nerewip (137°E) formations: basaltic lava, pillow lava and argillites 600 m thick. Nerewip Formation partly schistose and metamorphosed to greenschist facies. Awitagoh Fm overlies Kariem Fm; altered basalt has minimum K-Ar age of 486 ± 17 Ma (Parris, 1994a); may underlie Kariem Fm (Cloos et al., 2005). |
| Late Proterozoic > 820 Ma                | Kariem Formation: pyritic mudstone and dolomitic mudstone with dolerite sills, known only in the eastern central Range at 139.5–140.4°E, K-Ar ages 820 ± 21 Ma and 847 ± 5 Ma; thickness >2.5 km (Parris, 1996b). |
| Proterozoic                               | Basement (not exposed).                                                                   |
The terranes include one or more fragments of Paleozoic craton, ophiolites, a variety of metamorphic rocks, dioritic and granodioritic intrusives, island arc volcanic rocks and associated sediments, and oceanic crustal rocks. They are partly covered, unconformably, by as much as 10 km of sediment of Early and Middle Miocene and younger age.

**Derewo Metamorphics**

The Derewo or Ruffaer Metamorphics (Warren and Cloos, 2007; Weiland, 1999) extend for 550 km along the N side of the Central Range in West Papua. They include two distinct and probably unrelated metamorphic rock units: one is the great mass of contorted quartz-veined phyllitic and micaceous graphitic schists that comprise the bulk of the rock unit, and the other is a series of higher-grade metamorphic rocks including high-temperature amphibolite, blueschist and eclogite, that are found close to the contact with the ophiolite (Weiland, 1999; Parris and Warren, 1996). The phyllitic and graphitic schists are entirely greenschist facies but metamorphic grade increases northward (Warren and Cloos, 2007). They derive from and are transitional southward into unmetamorphosed Jurassic-Cretaceous Kembelangen Formation. In some places the contact is faulted (Warren and Cloos, 2007).

The amphibolites probably are part of the metamorphic aureole of the ophiolite. They appear to be similar to the high-temperature amphibolites and hornblende granulites that are found at and near the base of the Papuan Ultramafic Belt (PUB) ophiolite in PNG (Davies, 1971; Lus et al., 2004).

On the geological map (Figure 3) the Derewo Metamorphics appear to be truncated at the international border but this is not the case. Rather, the quartz-veined graphitic schists continue across the border where they were mapped as metamorphosed Jurassic Om Formation (Davies, 1982) and on Figure 3 are included within the area mapped as Papuan basin fold belt (rock unit 5). Metamorphic grade in metamorphosed Om Formation increases northwards and the schists have both gradational and faulted contacts with the unmetamorphosed protolith. The blueschists and eclogites of the Derewo Metamorphics are not continuous across the border but reappear 100 km E of the border at 142°E (Tau Blueschist; Ryburn, 1980; Davies, 1982).

**The West Papua Ophiolite**

Ultramafic and gabbroic rocks of the West Papua ophiolite are exposed for a length of 440 km on the N side of the central range, and...
as outliers within the Weyland Overthrust (Dow et al., 1986; Monnier et al., 2000). The ophiolite is in fault contact with Derewo Metamorphics to the S. Metamorphosed gabbro that is faulted against the metamorphics immediately W of the international border may be an outlier of the same ophiolite (Parris, 1996b; rock unit not shown in Figure 3).

The western sector of the ophiolite, W of 138.5°E, is mostly ultramafic rocks, and the eastern sector, E of 138.5°E, mostly mafic rocks (Dow et al., 1986; based on interpretation of aerial photographs). Weiland’s (1999) foot traverses and spot landings generally confirmed these boundaries but picked out an area of sheared serpentinite with blocks of blueschist (137.8°E) and eclogite (138.05°E), just W of the Gauttier Offset. (The Gauttier Offset is an ENE-trending left-lateral fault that may extend from the central ranges beneath Mamberamo Basin sediments to the eastern end of the Gauttier (Foja) mountains; Dow et al., 1988.)

Weiland (1999) described the ultramafic rocks as variably serpentinised peridotites. He searched for but did not find pillow lavas. Monnier et al. (2000) found harzburgite and dunite with minor wehrlite at two sites near 136.5°E, and gabbro with cumulus texture and some basalt near 138.7°E.

The ophiolite is thought to represent Late Cretaceous oceanic crust and mantle and to have been emplaced in the Late Cretaceous as indicated by ages of the high-temperature metabasites (68 Ma; Weiland, 1999; Bladon, 1988) or in the Eocene, 44 Ma, as indicated by ages of the high-temperature metabasites (68 Ma; Crust and mantle and to have been emplaced in the Late Cretaceous at two sites near 136.5°E, and gabbro with cumulus texture and some basalt near 138.7°E.

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Fragments of Paleozoic craton

The Border Mountains are an isolated block of Paleozoic metasediments and Triassic intrusive rocks that extend W from the international border at 3.75°S (the Idenburg Inlier of Dow et al., 1988). The metamorphic rocks include amphibolite gneiss, garnet-muscovite leucogneiss and ‘greenschist’. The intrusive igneous rocks include a layered mafic complex with layered ultramafic rocks, troctolite, gabbro, diorite, hornblende-quartz diorite and granodiorite, cut by andesitic dykes; most K-Ar ages are 250-240 Ma, and one is near 230 Ma (Davies, 1990; Parris, 1996c).

The fragment of Paleozoic craton extends E across the border to Amanab where metamorphic rocks are overlain unconformably by Late Cretaceous (late Campanian or Maastrichtian) limestone (Wilson et al., 1993) and may extend SSE to the Landslip Range, as is shown in Figure 3. There is no age evidence to confirm this but the Landslip Range has served as a barrier against which the generally westward tectonic trends have been sharply deflected to the N (Davies, 1982; Davies et al., 1997) and thus may represent older stronger lithosphere. Crowhurst et al. (2004) concluded that Sr and Nd isotopic values for the Amanab and Landslip metamorphic rocks indicate a mixed partly cratonic provenance.

Mamberamo Basin

The low-lying area N of the central range is occupied by the Mamberamo Basin, a sedimentary basin with as much a 10 km thickness of sediments. Middle Miocene turbidites are unconformably overlain by rapidly deposited late Miocene–Quaternary clastic sediments; these show extensive diapirism (Williams and Amiruddin, 1984). Mamberamo Basin sediments have been subjected to Pliocene-Quaternary N-S contractional and transpressional tectonics to produce folds and N-facing thrust faults. Isolated blocks of basement rocks protrude from the basin and partially deflect the deformation. The Gauttier or Foja mountain block is one such. It comprises ultramafic rocks with some Paleogene andesitic–basaltic volcanics, volcaniclastic sediments, and minor limestone and is unconformably overlain by E-W-trending Neogene sediments (P.E. Pieters, pers.comm., 2008).

Table 3 Stratigraphy of the Jimi-Kubor and Bena Bena blocks.

| Age            | Description                                                                 |
|----------------|-----------------------------------------------------------------------------|
| Late Cretaceous| Chim Formation and Asai Shale: massive finely laminated calcareous grey shale; some volcanics; Cenomanian–Maashrichtian and Early Paleocene; 3 km; (Davies, 1983). |
| Early Cretaceous| Kondaku Formation: tuff and volcanically-derived sandstone; cuesta-forming; 2.5 km; Aptian–Albian (Bain et al., 1975) |
| Late Jurassic  | Maril Shale: dark calcareous siltstone and shale; 2 km; Kimmeridgian bivalves (Davies, 1983). |
| Mid Jurassic   | BB Block: Karmantina gnissic granite, 172 Ma (Page, 1976), intrudes Bena Bena Metamorphics |
| Early Jurassic | Balimbu Fm: dark grey volcanolithic sandstone and siltstone; 300 m; Sinemurian–Plainsbachian age (Pigram et al., 1984). |
| Late Triassic  | “Kubor Granodiorite(2)” rift-related granitic intrusive rocks (Crowhurst et al., 2004); c. 220 Ma; includes younger parts of Kimil Complex. |
| Middle–Late Triassic | Kanim Complex intrusives: 3.5 km, rift-related, part contemporaneous with Kimil Complex. Protolith of Bena Bena Metamorphics 221 Ma (Van Wyck and Williams, 2002) |
| Middle Triassic| “Kubor Granodiorite(1)” volcanic-arc-related mafic granitic rocks (Crowhurst et al., 2004); 240 Ma |
| Late Permian   | Omung Metamorphics (Van Wyck and Williams, 2002) |

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The Efars and Sidoas mountain blocks, adjacent to the N, are of ultramafic rocks.

The Cyclops Mountains, near Jayapura, comprise an ophiolite assemblage and moderate to high grade metamorphic rocks: amphibolite, gneiss and schist (Baker, 1956; Monnier et al., 1999). Boninitic associated with the ophiolite has a K-Ar age of 43 Ma (Monnier et al., 1999). Immediately adjacent are Miocene sediments and Cenozoic volcanic arc rocks. Ultramafic rocks of the ophiolite may extend across the border as indicated by an exposure of sheared serpentinite in an erosional window just E of the border (Norvick and Hutchison, 1980).

**Sepik Complex 141–146 °E**

The basement rocks of the Sepik Valley are grouped together as Sepik Complex (Davies, 1990; Rogerson et al., 1987), an assemblage of arc-volcanic, ultramafic, metamorphic, dioritic intrusive and sedimentary rocks that is interpreted to have formed in the Cenozoic in two arc-continent collisions.

In the southern part of the Sepik Complex the volcanic rocks and associated sediments are of Mid-Eocene to basal Late Eocene age and have been mapped as “Salumei Formation”. Associated metamorphic rocks include blueschist and eclogite (Ryburn, 1980) and have K-Ar ages of 44-38 Ma and 27-23 Ma (Davies, 1990). The Salumei rock assemblage is thought to be a product of a late Eocene arc-continent collision. This suite of rocks underlies the Sepik Valley sedimentary basin and extends northward as far as the southern slopes of the northern ranges (Bewani-Torricelli ranges).

The northern slopes of the Bewani-Torricelli ranges expose arc-volcanic rocks and intrusives that are of generally late Eocene and Oligocene age. They have been mapped as Bliri Volcanics and are thought to have been emplaced by arc-continent collision at the end of the Oligocene. Oligocene Bliri and Eocene Salumei arc volcanic rocks are not readily distinguished from one another in the field and the distribution of each may be revised by future mapping. The Bliri Volcanics are overlain unconformably by Early Miocene and younger sediments. The prominent calc-alkaline intrusive rocks of the Bewani Torricelli ranges are mostly Late Eocene–Mid-Oligocene in age: 13 samples have K-Ar ages in the range 40–30 Ma; two others are Late Cretaceous and three are Early–Mid-Miocene (Hutchison, 1975).

**The youngest accretion events**

Pliocene MORB-type basalts that are exposed on Kairiru Island at 143.5°E (John, 2006); and Plio-Pleistocene siltstone on the adjacent mainland at 3.41°S, 143.42°E (Klootwijk et al., 2003) are the youngest of the accreted terranes. The siltstone had moved S from an initial position near the equator (Klootwijk et al., 2003).

**Cretaceous high-grade metamorphic rocks and associated ultramafics**

Cretaceous moderate to high-grade metamorphic rocks are known in the NE part of the Sepik Complex in the Prince Alexander Range at 3.4–3.5 °S, 143.0–143.5 °E (rock unit 7 in Figure 3). The rock types include amphibolite gneiss, orthogness and subordinate mica schist (Hutchison, 1975). Age from a number of K-Ar determinations is Aptian–Albian (c. 110 Ma). Other lower grade metamorphic rocks along strike to the E are exposed in fault contact with ultramafic rocks (Mt Turu Complex), and across the Sepik Valley to the S (Hunstein Range). These generally lower grade rocks were mapped as Ambuini Metamorphics. Their age is not known but Sr and Nd isotopes indicate mixed continental and oceanic provenance (Crowhurst et al., 2004) and the possibility remains that these too are Mesozoic or even late Paleozoic. The Mount Turu ultramafic rocks appear to be of igneous cumulate origin (Hutchison, 1975).

**Miocene granodiorite: the Maramuni arc**

Large bodies of Miocene granodiorite intrude the SE part of the Sepik Complex and extend to ESE in the Jimi-Kubor terrane. These intrusives are very likely related to Miocene arc volcanic rocks that form a major part of the cover sequence and the two together have been given the name Maramuni arc.

**Marum Ophiolite**

The Marum Ophiolite is a layered sequence of ultramafic and gabbroic rocks that dips NE beneath the plains of the Ramu River and on the SW side is faulted against low-grade metamorphics (5.5°S, 145°E; Jaques, 1981). Lateritic soils on the ultramafic rocks are currently being developed as a source of Ni and Co.

**Adelbert-Finisterre-Sarawaged ranges**

The Adelbert, Finisterre and Sarawaged ranges comprise Oligocene–early Miocene arc volcanic rocks overlain by Miocene and younger limestone (Jaques and Robinson, 1977; Abbott, 1995). Fault deformation of the volcanic rocks preceded deposition of limestone. The volcanics and limestone form a great S-facing thrust-based anticline marked by limestone dip-slopes on the N side and by faulted and rapidly eroding volcanic rocks on the S side.

**Geology of the Eastern Part of New Guinea: the Papuan Peninsula**

The tail of the bird comprises a SE-trending peninsula and chain of islands. The Owen Stanley Range extends along the peninsula as a mountainous spine though broken into two parts by an area of lower ground at 148.25–148.8°E; peaks in the NW part of the range rise to 4 km, and in the SE part to 3 km.

The main rock units of the peninsula are a great mass of generally low-grade metamorphics (Owen Stanley metamorphic complex), a major ophiolite (Papuan Ultramafic Belt (or PUB) ophiolite), a great mass of Late Cretaceous and Mid-Eocene submarine basalts with minor pelagic limestone, Oligo-Miocene sedimentary cover, Miocene and Pliocene dioritic and granodioritic and shoshonitic intrusive stocks, and two Quaternary stratovolcanic complexes (Lamington-Hydrographers and Victory-Trafalgar).

**Owen Stanley Metamorphic Complex**

The greater part of the metamorphic complex is made up of greenschist to low greenschist facies felsic rocks, some clearly derived from felsic volcanic detritus (Kagi Metamorphics; Pieters, 1978) with minor limestone and conglomerate. Along the NE margin of the Owen Stanley Range the felsic metamorphics are structurally overlain by a
Paleocene tonalite stocks and plutons in fine sediments associated with the basalt; one K-Ar age is 56 Ma thought to be Maastrichtian, c. 71–65 Ma, as indicated by foraminifers.

The association of exhumed metamorphic rocks and ophiolite intrude the gabbroic part of the ophiolite and mostly do not penetrate the basalt. Overlying the ophiolite are Middle Eocene arc-type volcanic rocks, middle Miocene and younger volcanics, rapidly deposited Miocene and younger clastic sedimentary rocks and some limestone, and Pliocene–Quaternary volcanic rocks, including those of the intermittently active major volcanoes, Lamington and Victory.

Metamorphic aureole of the ophiolite

At the base of the ophiolite there is intermittently exposed a 300 m thick thermal metamorphic aureole that comprises granulite and hornblende granulite grading to amphibolite away from the contact. The protolith is thought to be Emo Metamorphics. The cooling age of the contact metamorphic event is tightly constrained at 58.3 ± 0.4 Ma (Lus et al., 2004).

Owen Stanley Fault

The Owen Stanley Fault is an E and NE-dipping fault that separates the metamorphic rocks from the ophiolite. It is interpreted to be a former subduction system thrust fault and is now, for most of its length, an extensional fault that has allowed the emergence of the metamorphic rocks in Late Miocene–Quaternary time.

Suckling-Dayman Massif

East of 148.2°E, the Owen Stanley Range falls away to lower ranges (Musu Valley area) but resumes at 148.8°E as an E-W elongated antiformal range – the Suckling-Dayman Massif. The remarkable smooth arched surface of the range, on older maps mistaken for a stratovolcano, is interpreted to be an exhumed and antitclinally folded subduction thrust fault (Davies, 1980; Daczko et al., 2009). Greenschist facies metabasite at the frontal fault gives way to variably metamorphosed basalt and limestone with some high pressure indicator minerals away from the fault (Davies, 1980). The limestone is moderately metamorphosed but foraminifera of Maastrichtian age are preserved in places.

Submarine basalts

Much of the remainder of the Papuan peninsula, SE of the Suckling-Dayman Massif, is made up of a great volume of little-deformed submarine basalts with a thickness greater than 1 km. Much of the basalt is Mid-Eocene and some associated limestone shows low-angle extensional faulting. Mid-Miocene shoshonitic intrudes the basalts (Smith and Davies, 1976).

D’Entrecasteaux Islands, Louisiade Islands, Trobriand Islands and Muyua

The association of exhumed metamorphic rocks and ophiolite extends from the mainland NE to the D’Entrecasteaux Islands and ESE to Misima, Sudest and Rossel islands (Davies and Warren, 1988; Hill et al., 1992). Eclogite is preserved in the generally amphibolite facies metamorphic rocks of the northern D’Entrecasteaux Islands (Davies and Warren, 1992); some is coesite-bearing and some has been shown to have formed in the late Miocene or Pliocene (Baldwin et al., 2004, 2008; Little et al., 2011).
The Trobriand Islands are raised coral probably underpinned by Plio-Quaternary volcanic rocks. Woodlark (Muyua) Island comprises Quaternary limestone cover on Cenozoic volcanic arc rocks.

Small Ocean Basins

The Coral Sea Basin opened in the Paleocene (Weissel and Watts, 1979) leaving rifted fragments of Australian craton as submarine plateaus on the NW and SW margins of the basin (EP, PP, QP in Figure 1; Drummond et al., 1979). The Solomon Sea basin, N of the Trobriand Trough, opened in the early Oligocene, 34–28 Ma (Joshima et al., 1986) and has been subducted at both the New Britain Trench and the Trobriand Trough. The antiformal Solomon Sea plate can be traced WNW beneath mainland New Guinea for at least 400 km to 142°E (Cooper and Taylor, 1987). The Manus Basin (eastern Bismarck Sea) opened by asymmetric spreading in the last 3.5 Myr and Woodlark Basin in the last 6 Myr (MB and WB in Figure 3; Taylor, 1979; Taylor et al., 1995, 1999).

Bismarck Archipelago, Bougainville-Buka and the Ontong Java Plateau

The islands of the Bismarck Archipelago, together with the volcanic arc rocks of the Bewani-Torricelli and Adelpin-Finisterre-Saruwaged ranges, are thought to have a common origin and to possibly have formed as a single linear volcanic arc, referred to as the Bewani-Torricelli-Baining arc by Klootwijk et al. (2003). Each has a contact with the Solomon Islands and Bougainville-Buka in the Late Miocene (Mann and Asahiko, 2004). Faulted parts of the Plateau are exposed in the adjacent Solomon Islands.

Plate tectonic setting, evolution, neotectonics and volcanic activity

New Guinea is at the interface of the Pacific, Australian, Philippines Sea, SE Asian and Banda Sea plates (Hamilton, 1979; Hill and Hall, 2003; Bird, 2003). Convergence between the Pacific and Australian plates is at a rate near 110 mm/yr on azimuth near 070 degrees. The present plate boundary comprises a series of microplates (Figure 3), each demarcated by earthquake activity (Figure 6). Figure 6a shows clearly the Solomon Sea plate lithosphere dipping N beneath New Britain and extending W beneath the mainland as far as 142°E. Figure 6b, with fewer earthquakes, shows more clearly the NW-trending Mamberamo Fault (centred at 3°S, 140°E) and the SW-trending fault that runs through Nabire (3°S, 136°E) in south-eastern Cendrawasih Bay. Shallow earthquakes coincide with all of the trenches and with the front of the fold belt.

Convergence has led to a succession of collisions of the Australian craton with volcanic arcs, ocean crust and microcontinents including fragments of Australian craton that had rifted from the craton and then docked again (Figure 7; Pigram and Davies, 1987; Davies et al., 1997). Silver and Smith (1983) pointed to the remarkable parallelism between the accretion of terranes to the New Guinea margin in the Cenozoic and the accretion of terranes to the North American margin in the Mesozoic. In both cases the accreted basaltic lava flows over lain by 1 km of pelagic sediments. The lavas were emplaced in one remarkable magmatic event over a period of less than 7 Ma at c.122 Ma (Early Cretaceous, Aptian) with a lesser pulse at 90 Ma (Mahoney et al., 2001). The plateau is entirely submerged except for isolated atolls. WNW motion of the Pacific plate brought the plateau into contact with the Solomon Islands and Bougainville-Buka in the Late Miocene (Mann and Asahiko, 2004). Faulted parts of the Plateau are exposed in the adjacent Solomon Islands.
sediments within the Mamberamo Basin, the transfer of convergent upper plate also would explain the deformation of Plio-Quaternary (et al., 2005). A shallow-dipping slab that is partly coupled to the (Davies, 2009 a, 2010), rather than related to slab break-off (Cloos and Ok Tedi mineralised intrusive rocks, can be seen as slab-related igneous activity in the Papuan Basin fold belt, including the Grasberg (e.g., Pegler at al., 1995). If this interpretation is correct then the tomography model is partly supported by mapping of earthquake foci near the line of the S coast (Tregoning and Gorbatov, 2004). The shallow angle and to extend beneath the island of New Guinea to 1992). Seismic tomography shows the subducted slab to dip at a southwestward at the New Guinea Trench (see also Milsom et al., 2002) and with a seismic array experiment that showed crustal thinning investigated by Ocean Drilling Program Leg 180 (Taylor and Huchon, 2002) and with a seismic array experiment that showed crustal thinning investigated by Ocean Drilling Program Leg 180 (Taylor and Huchon, 2002). The same convergence causes the Finisterre-Saruwaged mountain mass to ride southward (Abbott et al., 1997) and results in loading and downwarping of the northern end of the Papuan peninsula, the coast of which is subsiding at a rate of 5 mm/ yr (Webster et al., 2004).

Sea-floor spreading in the Manus Basin in the eastern Bismarck Sea (MB in Figure 3; Taylor, 1979) and retreat of the New Britain Trench allow relatively rapid clockwise rotation of the island of New Britain (Tregoning et al., 1998; Wallace et al., 2004). Sea-floor mineralisation in the Manus Basin was investigated by the Ocean Drilling Program Leg 193 (Barriga et al., 2007).

The westward advance of sea-floor spreading in the Woodlark Basin (WB in Figure 3) has caused rifting and extension of the Papuan peninsula and adjacent islands and the development of domes and antiforms of layered metamorphic rocks (‘metamorphic core complexes’). The easternmost continental extensional structures were investigated by Ocean Drilling Program Leg 180 (Taylor and Huchon, 2002) and with a seismic array experiment that showed crustal thinning and extension within the upper mantle (Abers et al., 2002).

Neotectonics

In NW New Guinea the oblique convergence between the Australian and Pacific plates is accommodated partly by left-lateral strike-slip motion on an E-W fault system that connects the Bismarck Sea Transform Fault in the E with the Sorong Fault in the W; partly by left-lateral strike-slip motion on another E-W fault system on the S coast – the Tarera-Aiduna fault system that extends W from the Weyland Overthrust (Pubellier and Ego, 2002); partly by subduction at the New Guinea Trench; partly by transpressional and strike-slip faulting in the fold belt (Abers and McCaffrey, 1988) and by folding and thrust faulting in the Mamberamo Basin.

The lithosphere of the Caroline Sea or Pacific plate is subducted southwestward at the New Guinea Trench (see also Milsom et al., 1992). Seismic tomography shows the subducted slab to dip at a shallow angle and to extend beneath the island of New Guinea to near the line of the S coast (Tregoning and Gorbakov, 2004). The tomography model is partly supported by mapping of earthquake foci (e.g., Pegler at al., 1995). If this interpretation is correct then the igneous activity in the Papuan Basin fold belt, including the Grasberg and Ok Tedi mineralised intrusive rocks, can be seen as slab-related (Davies, 2009 a, 2010), rather than related to slab break-off (Closs et al., 2005). A shallow-dipping slab that is partly coupled to the upper plate also would explain the deformation of Plio-Quaternary sediments within the Mamberamo Basin, the transfer of convergent motion southward for a distance of 400 km from the line of the New Guinea Trench to the southern front of the fold belt and the observed WSW movement of the Bird’s Head.

The Bird’s Head is moving WSW at a rate of 86 ± 9 mm/yr relative to Australia (Stevens et al., 2002). This is slower than the motion of the Pacific Plate but has the same azimuth and is most likely explained by some degree of coupling between the Bird’s Head and the underlying subducted slab. This motion in turn requires subduction of Bird’s Head lithosphere at the Seram Trough (ST in Figure 3; Stevens et al., 2002). The WSW motion also has caused the opening of Cendrawashiy Bay, the development of the Waipona sedimentary basin, and the change from Miocene contractual to Pliocene extensional tectonism in the Wandamen Peninsula and Lengguru Fold Belt (Bailly et al., 2009).

In NE New Guinea, collision between the Bismarck volcanic arc and the leading edge of the island of New Guinea is causing uplift of the north coast of the Huon Peninsula at (averaged) rates of 1–3 mm/yr. (Study of raised coral terraces on the peninsula (Figure 8; Chappell, 1974) yielded a high-quality record of fluctuations in sea level during the Late Quaternary.) The same convergence causes the Finisterre-Saruwaged mountain mass to ride southward (Abbott et al., 1997) and results in loading and downwarping of the northern end of the Papuan peninsula, the coast of which is subsiding at a rate of 5 mm/yr (Webster et al., 2004).

Volcanic Hazards

Fourteen active and 22 dormant volcanoes are a danger to an estimated 250,000 people (Saunders and Itikarai, 2006). Of the active volcanoes, 6 have been categorized as high-risk volcanoes. Five of these are in the Bismarck volcanic arc which extends along the N coast of New Britain and WNW to the islands offshore from mainland New Guinea. The five are the island volcanoes, Manam and Karkar, and the New Britain volcanoes Pago/Witori, Ulawun and Rabaul (Figure 9). The sixth high risk volcano is Lamington volcano on the NE coastal plains of the Papuan peninsula, NE of Port Moresby.

The caldera collapse volcanoes Long Island, Dakataua, Witori and Rabaul have been the source of devastating eruptions in the past, including a remarkable sequence of major eruptions from the three New Britain caldera volcanoes in the 7th century (McKee et al., 2011). Ash emission from an eruption of Long Island in c.1665 was sufficiently voluminous to block out the light of the sun for several days, and is recalled in legend in the PNG highlands as a time of darkness (Blong, 1982).
Natural Resources

The island is richly endowed with natural resources. The Grasberg mine in Indonesian Papua contains the largest recoverable reserves of Cu and the largest Au reserve in the world and Papua New Guinea ranks eleventh in the world in the annual production of Au. In addition both PNG and the Indonesian provinces are about to become major exporters of LNG. In the Indonesian provinces oil is produced from Miocene sediments in the Salawati and Bintuni basins and gas for LNG from Jurassic sediments in the giant Tangguh field (Robertson, 2012).
In PNG hydrocarbons are produced as oil from the Kutubu field and will be produced as gas for LNG from the Hides anticline (142.8 °E) and other structures in the fold belt, and from Miocene reefs in the eastern fold belt. Copper and Au are produced from mines at Grasberg and Ok Tedi (Figure 10), Au from Porgera, Hidden Valley, and Lihir Island (locations shown on Figure 3) and, commencing in 2012, Ni and Cu will be produced from the Ramu lateritic deposit, SW of Madang. Mineral deposits likely to be developed in the near future are Cu-Au at Wafi-Golpu (SW of Lae), at Frieda River (NE of Ok Tedi), and at Yandera (SW of Madang: Cu and Mo), and Cu-Au-bearing sea-floor massive sulfides in the Manus Basin.

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